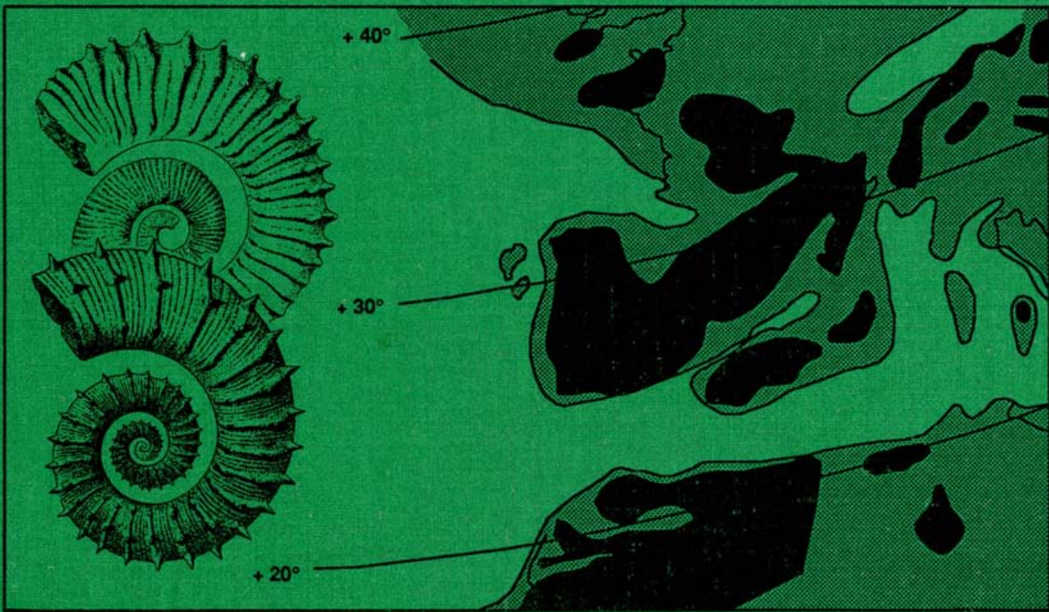
Cretaceous of the Western Tethys

Edited by
Jost Wiedmann



Proceedings of the 3rd
International
Cretaceous Symposium
Tübingen 1987

E. Schweizerbart'sche Verlagsbuchhandlung
(Nägele u. Obermiller) Stuttgart

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and on 5 foldouts

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Editor's Preface

In 1978, in Münster, the German Subcommission on Cretaceous Stratigraphy initiated a first International Symposium on the Cretaceous. The focus of the symposium was on Germany and Central Europe. The proceedings were published in 1979. The second symposium was held in 1982 in Munich and concentrated on the Alpine Cretaceous. The proceedings were published in 1983.

Published in this volume are proceedings of the third symposium, which took place in Tübingen in 1987 and whose topic was the western Tethys. The participants of the symposium unanimously decided to dedicate the volume published to TOVE BIRKELUND in response to the loss of our highly esteemed Danish colleague.

The great variety of topics presented during the Tübingen Symposium and offered in this volume can be grouped together in the following subjects: (A) The **Western Mediterranean**, including its transition to the North Atlantic; (B) The **Alps, Carpathians, Dinarids, and Caucasus**; (C) Cretaceous **Events**; (D) **Biostratigraphy, Correlation,** and **Paleogeography**; and (E) **Volcanism** and **Magnetostratigraphy**. I hope that this diversity will be appreciated by the members of the Cretaceous community.

The completion of this volume was only possible with the continuous assistance of many colleagues contributing in different ways such as reviewing, re-typing, or re-drawing the present papers. I would like to express my sincerest gratitude to all of them, but above all to my friend FRANZ ALLEMANN (Bern), who was also responsible for the success of the symposium excursions, as well as to HANS J. HANSEN and WALTER K. CHRISTENSEN (both Copenhagen), BRAD SAGEMAN (now Boulder, CO), KAI-UWE GRÄFE, JÖRG PROSS and WOLFGANG SCHWENTKE.

Costs of printing increase and financial support of proceedings such as this one become more and more difficult to obtain. I am deeply indebted to two organizations that contributed extensively to the printing costs, and thus the realization of this volume. This are the Deutsche Forschungsgemeinschaft and the Alexander von Humboldt Foundation. Last but not least, thanks are due to Dr. E. NÄGELE and his crew from Schweizerbart Editors for the layout of this volume.

October, 1989 JOST WIEDMANN

In memoriam TOVE BIRKELUND (1928–1986)

The first response to the first circular of the 3rd International Symposium on the Cretaceous came from TOVE BIRKELUND and was dated June 2, 1986. She wrote me "I would very much like to come, but my health does not allow me to travel any more, which is the reason why I cannot participate. I wish you a successful meeting, and send you my best wishes. Yours sincerely, Tove". It was her last letter. TOVE BIRKELUND died June 24, 1986 after a period of severe illness. We all have lost a good friend and a highly enthusiastic and qualified colleague. We all appreciate the efforts TOVE BIRKELUND made in transforming our Cretaceous community into a Cretaceous family during the two wonderful and highly successful symposia she organized at Copenhagen in 1979 and 1983.

TOVE BIRKELUND was born in Nordby, on the island of Fano, November 28, 1928. Her father was a schoolmaster and much interested in natural history. No wonder that TOVE, after having finished school in 1947, started studying natural history at the University of Copenhagen. In 1954 she graduated in geology with a master thesis, describing Upper Cretaceous belemnites from Denmark (BIRKELUND 1957). Her love for geology was most probably the result of her participation in excursions to the Nûgssuaq Peninsula, West Greenland in 1949 and 1952 under the guidance of Professor ROSENKRANTZ.

In 1954, TOVE BIRKELUND married cand. mag. SVEND ANDERSON.

In 1958, she was awarded the university's gold medal in appreciation of her monograph on the scaphitid ammonites from West Greenland. A monograph on the complete Upper Cretaceous ammonite fauna from West Greenland followed in 1965; this was her habilitation. The high quality of this monograph substantially contributed to her international reputation. Greenland became one of the favoured areas in TOVE's research.

TOVE became an assistant at the Mineralogical and Geological Institute and Mineralogical Museum after having graduated; in 1960 she became fully employed; in 1963 leader of the stratigraphic-phytopaleontological section of the museum; and in 1966 a Professor of Historical Geology. When the institute was split into 5 departments in 1967, she became chairman of the Institute for Historical Geology and Paleontology. It was her merit immediately to introduce democratic rules and to reorganize the institute. Later on she played a major role in the creation of the new Geologisk Centralinstitut in Øster Voldgade 5-7, and in Øster Voldgade 10 later on. At the same time she was very much involved in the elaboration of a new curriculum for students.

Parallel to her study of ammonites, TOVE published papers on Cretaceous belemnites from West Greenland (1956) and Denmark (1957). All of these early papers were confined to systematics, stratigraphy and paleogeography.

She began working with a second topic in 1967, namely the ultramicroscopic shell structures of cephalopods (BIRKELUND & H. J. HANSEN 1968, 1974, 1975, BIRKELUND 1981). After detailed research in this area she moved on to describe the last ammonites of Maastrichtian age (BIRKELUND 1979, 1982), and finally investigated the problem of mass extinction at the Cretaceous-Tertiary boundary. Together with her colleagues she organized the "Cretaceous-Tertiary Boundary Events Symposium" in 1979, which was accompanied by wonderful excursions and was a great success. Here, WALTER ALVAREZ (Berkely) first presented his ideas about a cosmic impact at this boundary. In 1982, TOVE BIRKELUND together with E. HÅKANSSON presented an original concept of multicausal events to explain the boundary extinctions. Now, instead of individual fauna the whole faunistic community affected by these events (sponges, bryozoa, brachiopoda, echinoderms, and others) was taken into consideration resulting in a very detailed and complex boundary scenario.

TOVE's research was, however, not restricted to the Cretaceous. She activated a very effective international working group to re-investigate the late Paleozoic to early Cretaceous of Jameson Land, Scoresby Land, and Milne Land, East Greenland, leading to numerous publications (BIRKELUND 1968, 1971, 1976, BIRKELUND & PERCH-NIELSEN 1969a, 1969b, 1976, SURLYK & BIRKELUND 1972, SURLYK et al. 1973, PERCH-NIELSEN et al. 1974, BIRKELUND et al. 1974, 1978, 1984, CALLOMON & BIRKELUND 1980, 1982, BIRKELUND & CALLOMON 1985, and many others). Papers on the Jurassic and Cretaceous of North Greenland followed (HÅKANSSON et al. 1981, BIRKELUND & HÅKANSSON 1983), as well as publications on the same age levels in Norway (BIRKELUND et al. 1978, AARHUS et al. 1986), on drill sites from the North Sea (BIRKELUND et al. 1983), and on the Upper Jurassic of England (BIRKELUND et al. 1984). TOVE organized and took part in a great number of expeditions to Greenland, the last one in 1985, one year before she died.

Accuracy, efficiency, and great diversity are synonyms for TOVE's oeuvre, which was to a large extent realized in collaboration with colleagues from Denmark and abroad. K. PERCH-NIELSEN, J. CALLOMON, F. FÜRSICH, and others were pleased to work with her over many years.

The broad international appreciation of TOVE's scientific and human personality resulted in her nomination for chairman of the International Subcommission of Cretaceous Stratigraphy in 1976. She made great efforts to reorganize this body and to make it as effective as possible. These efforts culminated in the realization of a second symposium focusing on the Cretaceous stage boundaries she organized at Copenhagen in 1983. This symposium (BIRKELUND et al. 1983) also became a great success due to the perfect preparation and organization of the Copenhagen working group around TOVE BIRKELUND. The proceedings of this symposium, published in the following year (BIRKELUND et al. 1984), can be considered a standard work on the Cretaceous stage boundaries and related problems.

TOVE BIRKELUND was not only an outstanding scientist, she was a charming woman, and not at least, a highly political person. As I learned from the obituary of WALTER K. CHRISTENSEN et al. (1987), she was chairman of the Danish Geological Society (1969-1971), a member of the Danish Natural Science Foundation (1970-1979), a member of the Danish Royal Society (since 1971), a member of the directory board of Carlsberg Foundation, and likewise of the United Breweries (since 1978). She was chairman of the Planning Committee for Research (1981-1984), a member of many other research and planning committees, e. g. Scientific Research in Greenland, Oil and Gas Research, and Teaching Natural Science, and at the same time chairman of a collegium of women (1974-1982). She worked for local church communities, the "Politiken" newspaper foundation, and many other groups. How TOVE BIRKELUND was able to manage all of these different tasks is impossible to imagine.

JOST WIEDMANN

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TOVE BIRKELUND's curriculum is largely adopted from

CHRISTENSEN, W. K., SURLYK, F. & SØRENSEN, H. (1987): TOVE BIRKELUND. - Dansk. Geol. Foren. Årsskr. **1986:** 83-91 (see also complete list of publications).

Contents

B. Alps, Carpathians, Dinarids, Caucasus

C. Events

D. Biostratigraphy, Correlation, Paleogeography

A. Western Mediterranean

Iberian Kinematics during the Cretaceous – Paleogeographic consequences[1]

JACQUES ANDRÉ MALOD, Paris

With 4 Text-Figures and 1 Table

MALOD, J. A. (1989): Iberian Kinematics during the Cretaceous - Paleogeographic consequences. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 3-16. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The Iberian plate behaved as a small independent plate during the Mesozoic. Its motion with respect to the major plates has induced synchronous events such as rifting and spreading all around its borders. During the late Jurassic and the earliest Cretaceous Iberia moved southwest with respect to Eurasia. At the time of magnetic anomaly M-10N (Valanginian), this direction changed to a southeast one. Then Iberia followed Africa in its eastward motion until the late Cretaceous. This kinematic history allows understanding of the formation of the sedimentary basins linked to the Iberian continental margins. Intraplate basins are also probably related to such plate kinematics, but a more specific model is needed to allow precise correlations.

Kurzfassung: Die Iberische Platte war während ihrer mesozoischen Geschichte eine kleine unabhängige Platte. Ihre Bewegungen - relativ zu den Großplatten - haben synchrone Ereignisse induziert, wie z. B. Rifting und Spreading entlang ihren Außenrändern. Im späten Jura und in der frühen Kreide bewegte sich Iberia - relativ zu Eurasia - nach SW. Zur Zeit der Magnetanomalie M-10N (Valangin) veränderte sich dies in eine SE-Richtung. Danach folgte Iberia der Afrikanischen Platte in ihrer E-Bewegung bis in die späte Kreide. Die kinematische Geschichte erlaubt es, die Bildung von Sedimentbecken an den Kontinentalrändern Iberias zu verstehen. Auch Intraplatten-Becken sind wahrscheinlich mit dieser Plattenkinematik verknüpft; allerdings bedarf es eines spezifischeren Modells, um präzise Korrelationen zu ermöglichen.

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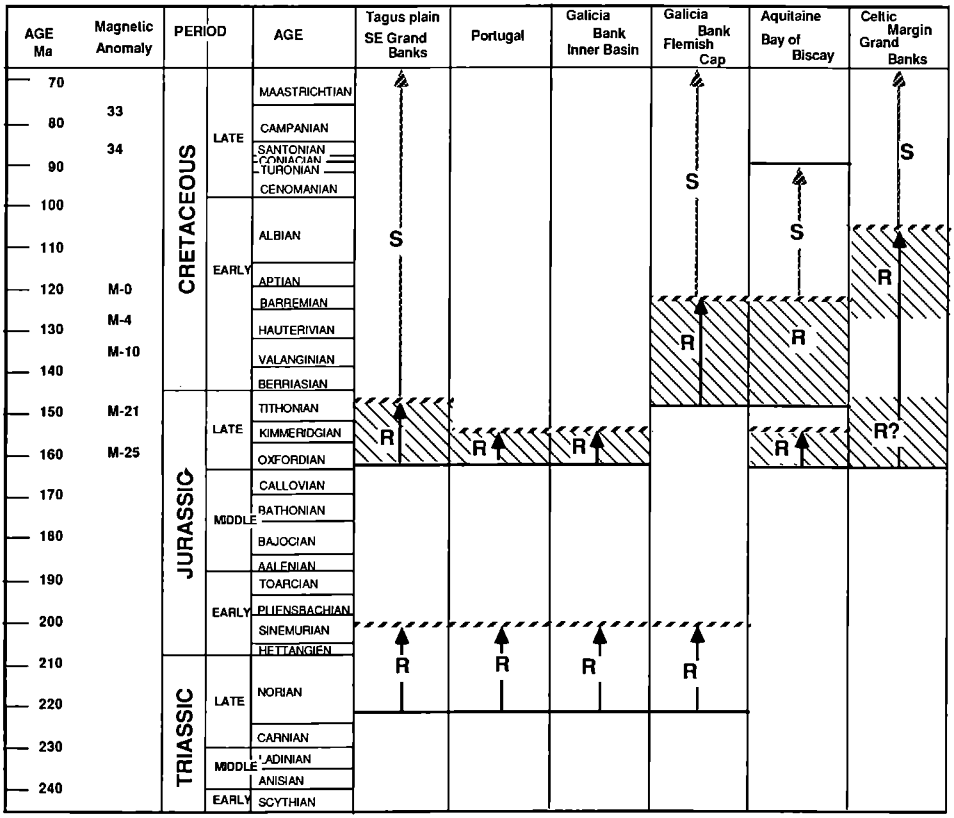
[1] Contribution No. 0416 of the Groupe d'Etude de la Marge Continentale et de l'Océan, Equipe UA 718, CNRS.

1. Introduction

The motion of Iberia must be described with respect to 3 major plates: North America, Eurasia and Africa. Thus, the Iberian continental margins and basins have followed distinct and separate geodynamic histories, depending on which major plates they were associated with. Geological events of different nature (faulting, continental rifting, oceanic spreading) have been under way during the same time all around the Iberian plate. But only the onset of these tectonic events or changes in their rates or directions can be considered as synchronous.

The motion of Iberia was the result of several phases of continental rifting and oceanic spreading (Table 1). During the rifting phases, the stretching of the lithosphere corresponded to motion of the plates apart from the rifting zone which can be considered as a diffuse boundary between the plates (VINK 1982). These early motions are difficult to specify since there are no oceanic magnetic lineations to record them. During the

Table 1. Chronology of the main rifting and spreading events around the Biscay triple junction during the Mesozoic. R is rifting phase and S is spreading phase.



oceanic spreading phases, the motion is more easily defined. However, in the case of the Iberian peninsula oceanic spreading took place mainly during the Cretaceous quiet zone in the Bay of Biscay. Thereafter the constraints are weak.

2. What is really established?

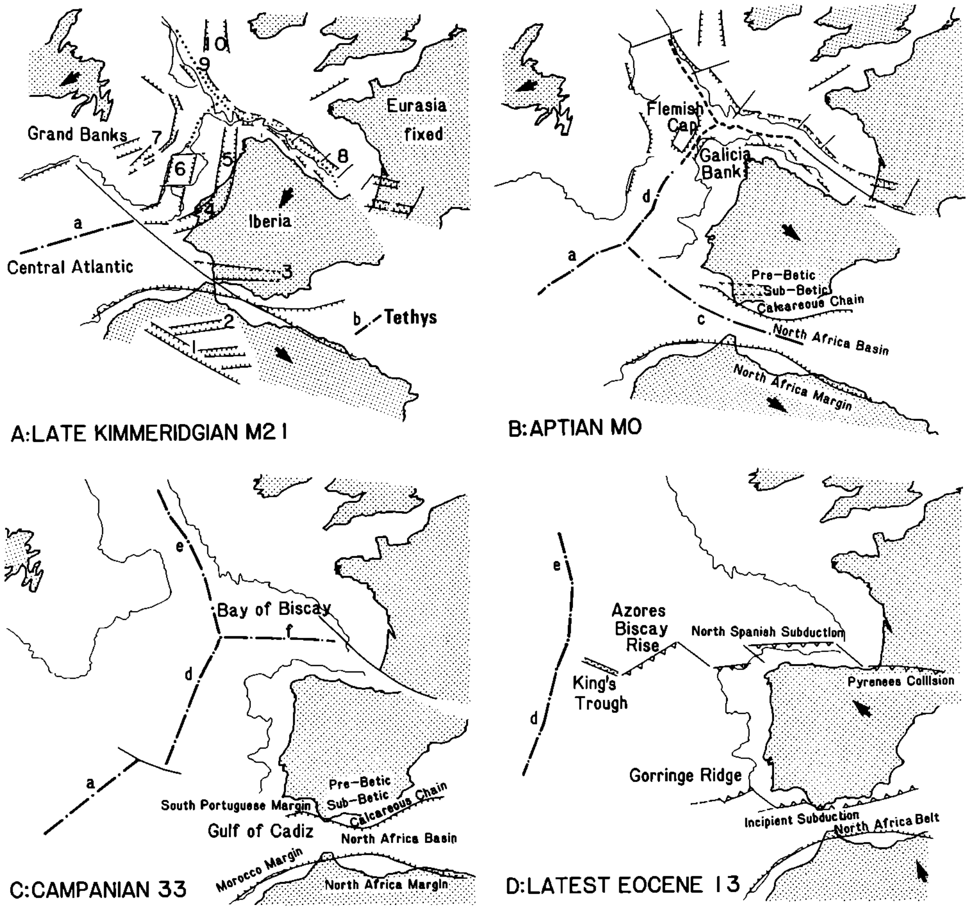
The opening of the Bay of Biscay is the result of a westwards motion and rotation of Iberia (LE PICHON & SIBUET 1971, OLIVET et al. 1984) which is well evidenced by paleomagnetic studies (VAN DER VOO 1969, STAUFFER & TARLING 1971). The total rotation of about 35°, measured between the Permian-Triassic and the Present time, seems to have occurred after the Middle Jurassic (SCHOTT et al. 1981) and mainly during the late Jurassic-early Cretaceous (GALDEANO et al., in press). The distribution in time of this rotation needs further paleomagnetic investigations.

Two oceanic magnetic anomalies M-0 (late Aptian) and 34 (Campanian) have been used to define the motion of Iberia. M-0 allows location of Iberia at the time of onset of oceanic spreading in the Bay of Biscay and between the Galicia Bank and Flemish Cap (Text-Fig. 1 B) (late Aptian-early Albian). However this location can be discussed since the exact location of the M-0 anomaly is still debated to the west of Iberia. Anomaly 34 gives the location of Iberia just before the end of the opening of the Bay of Biscay which takes place before the anomaly 33 (Campanian) (Text-Fig. 1 C). This latter reconstruction is probably the most accurate one. The Atlantic pattern of oceanic magnetic anomalies (KRISTOFFERSEN 1978) controls very well the position in longitude of Iberia, while the latitudinal location may be subject to some variation according to the definition of anomaly 34 in the Bay of Biscay. On the reconstruction of Text-Fig. 1 C (OLIVET et al. 1984) the Bay of Biscay is wider than presently and a marine basin underlain by stretched continental crust separates Iberia and Eurasia along the Pyrenean domain.

Between these two situations the kinematics can be discussed: The rotation pole between Iberia and Eurasia has been calculated using tectonic features which have been later deformed during the Tertiary compressive phases (for instance the North Pyrenean Fault) (LE PICHON & SIBUET 1971). As a result this pole needs to be revised in the future.

Going back in time, during the rifting event in the Bay of Biscay (before M-0), detailed kinematics remains highly speculative (HANISH 1984, MASSON & MILES 1984). Plate motions have been generally interpolated between the initial fit built from geological and geometric patterns and the location of Iberia at the time of the M-0 anomaly.

In conclusion, the eastward motion of Iberia with respect to Eurasia is definitively established, but a detailed kinematics is not yet available. This is especially true if one wants to use it to address geological problems at a local or regional scale. We will present briefly the most significant constraints on the motion of Iberia and try to integrate in the same comprehensive model the evolution defined by OLIVET et al. (1984) for the spreading phase and some new ideas concerning the continental rifting phase.

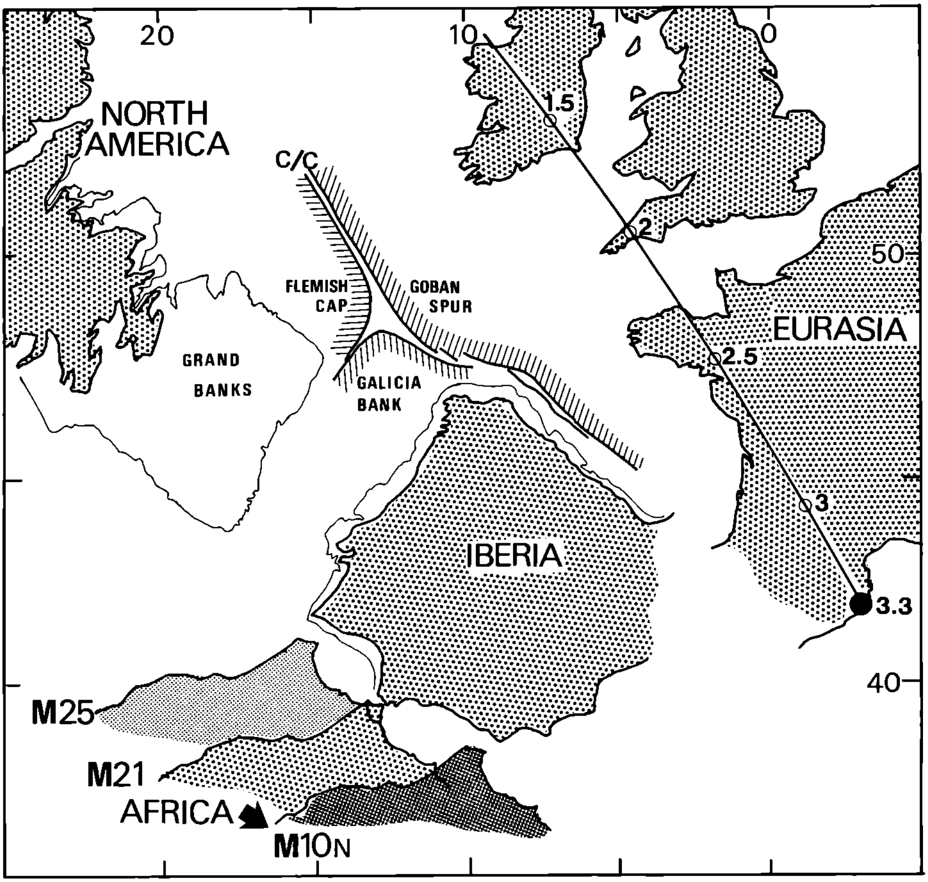


Text-Fig. 1. Plate tectonic evolution after KLITGORD & SCHOUTEN (1986), MALOD (1987) (A and B) and OLIVET et al. (1984) (C and D).
Fig. 1 A. Kimmeridgian - Magnetic anomaly M-21.
Fig. 1 B. Aptian - Magnetic anomaly M-0.
Fig. 1 C. Campanian - Magnetic anomaly 33.
Fig. 1 D. Latest Eocene - Magnetic anomaly 13.
The arrows show the directions of motions with respect to Eurasia.
Continental rifts: 1: South Atlas; 2: North Atlas; 3: Sub-Betic; 4: South Portuguese; 5: North Portuguese; 6: West-Iberian; 7: Grand Banks; 8: Bay of Biscay; 9: Celtic; 10: Porcupine.
Mid-Oceanic ridges: a: Central Atlantic; b: Tethys; c: North-Africa; d: West-Iberia; e: West-Ireland; f: Bay of Biscay.

3. Constraints on the motion of Iberia

3.1 The initial fit

One must take into account the geometric assemblage and the geological pattern of the margins and the most probable kinematics. In order to im-



Text-Fig. 2. Initial fit of the North American, Eurasian and Iberian plates: c/c is the continental boundary restored to its initial location before stretching of the margins.
The motion of Africa with respect to the North America-Eurasia-Iberia block is shown for the M-25, M-21 and M-10N magnetic anomalies.
The location of the pole of rotation of Iberia with respect to Eurasia is given by the curve with an indication of the angle of rotation of North America with respect to Eurasia (see text).

prove the previous continental fit (BARD et al. 1971, LEFORT 1980) the stretched border of the continents has been restored in its initial state by means of stretching rates derived from geophysical studies (MALOD 1986). As a result the initial reconstruction is much tighter than the previous one, especially to the west of Iberia between the Grand Bank and the Portuguese margin and around the Biscay triple junction between Goban Spur, Flemish Cap and the Galicia Bank (Text-Fig. 2).

3.2 The African plate motion

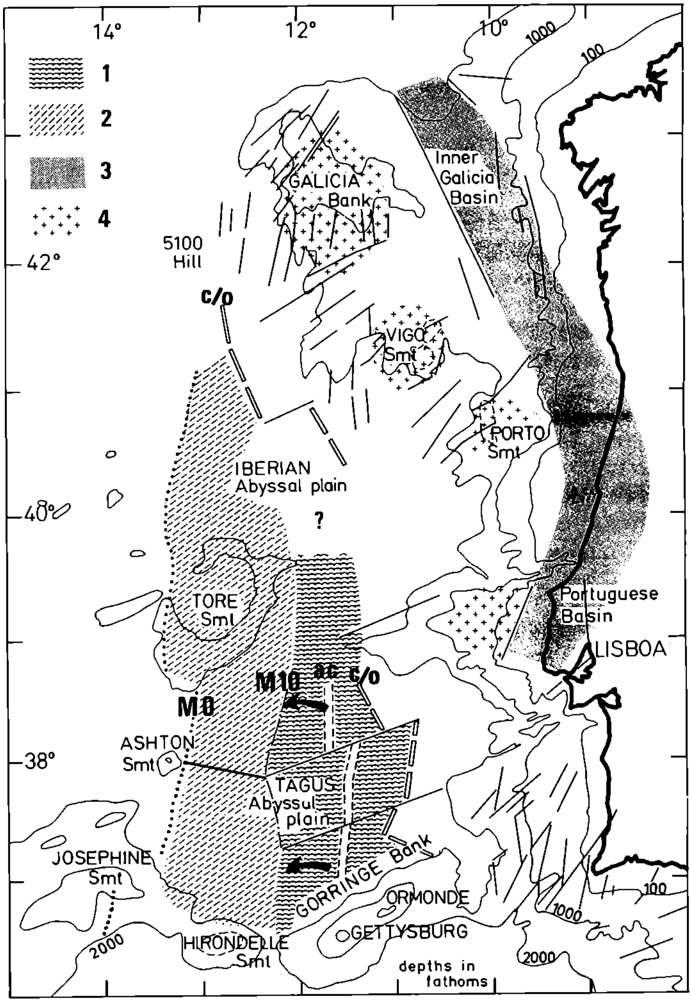
Given an initial fit of the continental block formed by Eurasia, North America and Iberia, which was made as close as possible by taking care of the stretched borders of the continents, we have drawn the location of Africa at different times (Text-Fig. 2). Africa is clearly responsible for the eastward motion of Iberia and thus Iberia may have begun to move as early as Africa. In the same way Iberia cannot have stayed fixed to the Eurasian block after the time at which Africa came in an eastward position with respect to it. In this case Iberia should later have moved faster than Africa to reach its M-0 situation (Text-Fig. 1 B). Text-Fig. 2 shows that Iberia's first motion is then older than M-21 (late Jurassic). This observation is dependent on the initial reconstruction, but a closer reconstruction between Grand Banks and Iberia would provide an older upper limit age for the beginning of the Iberian motion.

3.3 Evidence for a two phase kinematics during the late Jurassic - early Cretaceous

a - The oceanic domain between Iberia and North America

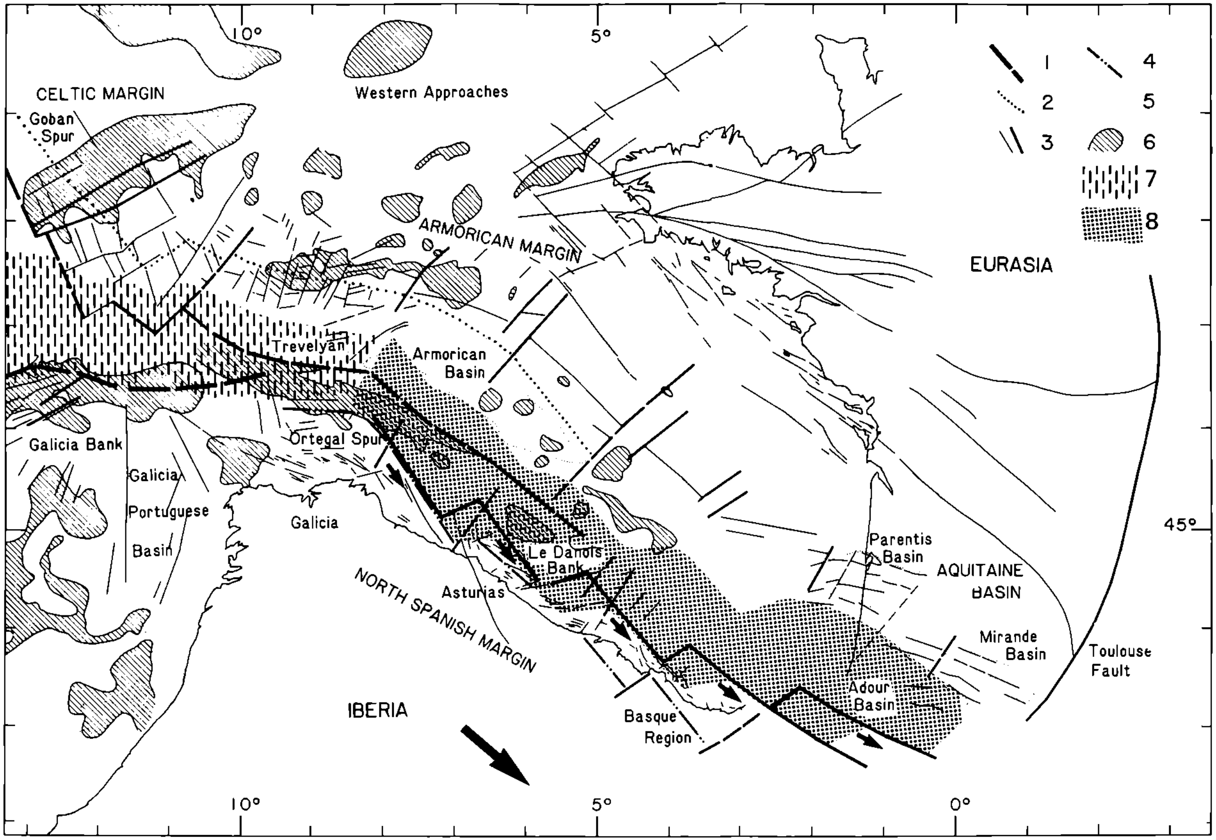
Between Iberia and North America, magnetic anomalies and spreading directions in the oceanic domain older than the M-0 anomaly can be used to constrain the motion of Iberia. In this way the Tagus abyssal plain is a very important clue in the understanding of the region. The age and the nature of the crust underlying this basin have been discussed by OLIVET (1978) who proposed a Permian or Triassic age. KLITGORD & SCHOUTEN (1986) explain the southern part of the Tagus abyssal plain by oceanic accretion during the late Jurassic-earliest Cretaceous opening of the central Atlantic. New multichannel seismic lines (MAUFFRET et al., in prep.) allow us to recognize the continent-ocean boundary (Text-Fig. 3). Within the Tagus abyssal plain itself, the oceanic crust displays structures which can be interpreted as an extinct rift axis associated with a linear magnetic anomaly. From seismic reflection stratigraphy, MAUFFRET et al. (in prep.) date this spreading phase in the late Jurassic-earliest Cretaceous. This spreading event may have followed continental rifting in the Portuguese Mesozoic basin which is mainly of Kimmeridgian age (GUERY 1985). Transform directions associated with this oceanic spreading ridge are WSW-ENE, close to the main direction of the Gorringe Bank and similar to the direction of extension in the south Portuguese basin (GUERY 1985, LEPVRIER, pers. comm.). Gorringe Bank and Hirondelle Bank have been interpreted as old fracture zones reactivated during the Tertiary phase of convergence between Africa and Iberia (AUZENDE et al. 1979).

Thus, the eastern part of the Tagus abyssal plain is an oceanic basin corresponding to a WSW-ENE direction of motion of North America with respect to Iberia during the late Jurassic. On the contrary, the western part of the Tagus abyssal plain is interpreted by a westward shift of the spreading axis before the M-0 magnetic anomaly inducing transform directions of spreading nearly W-E (Text-Fig. 3). They indicate a clear change in the kinematics between Iberia and North America. Following the results of KLITGORD & SCHOUTEN (1986), MAUFFRET et al. (in press) propose



Text-Fig. 3. Structural trends of the west Iberian margin.
1: Old oceanic crust with E-NE transform directions (M-21 to M-10); 2: Oceanic crust with W-NW transform directions (M-10 to M-0); 3: Area of continental rifting during the late Jurassic; 4: Continental blocks of the west Iberian margin. c/o is the continent–ocean boundary; ac is an abandoned spreading centre (see text); arrows show the shift of the spreading axis at the time of the magnetic anomaly M-10.

that this kinematic change and the spreading axis shift have occurred at the time of the magnetic anomaly M-10N (Valanginian) which is the time of a major reorganization in the North Atlantic.



Text-Fig. 4. The Biscay Aquitaine rift branch: Structural trends are plotted on the plate reconstruction at anomaly M-0. The large arrow on the Iberian plate gives the direction of motion of Iberia during the late Cretaceous.
1: ocean-continent boundary; 2: continent boundary before stretching; 3: fault; 4: Cretaceous north Spanish edge of the continental shelf; 5: sedimentary basin boundary; 6: magnetic anomalies greater than -50 gammas; 7: North west divergent Iberian margin; 8: North Spanish transform margin: area of pull apart basins. A schematic en echelon pattern of faults is indicated.

b - The Bay of Biscay: Directions of extension

We assume that the fault pattern on the present continental margin represents pre-existing zones of weakness reactivated during the rifting. In particular, fractures transverse to the margin represent, on average, a regional trend associated with rifting, and may have acted as a guide-line for plate motion at least during the early phase of rifting. Or more simply, they reflect approximately the direction of rifting.

This point of view has been discussed by several authors. MONTADERT et al. (1979) have shown that the listric faults which bound the tilted blocks are not continuous along the Celtic and Armorican margins. Faults perpendicular to the listric faults acted as transform faults between different tilted blocks (Text-Fig. 4). More recently, MONTADERT (1984) described an anastomosed listric fault pattern and less numerous transverse faults, with regard to studies on continental rifts (CHENET & LE TOUZEY 1983).

On the Armorican margin (OLIVET et al. 1984) as well as on the North Spanish margin (BOILLOT et al. 1987, TEMIME 1984) the transverse faults are mainly old Paleozoic structures reactivated during the rifting. We propose that the more prominent of these structures give an approximate direction for the plate motions especially during the beginning of the rifting.

We have examined the Bay of Biscay and the west of the Galicia Bank from this point of view. Following the previously discussed assumptions, the direction of the early relative motion of Iberia with respect to Eurasia is documented by the fault pattern perpendicular to the rift. Along the Armorican margin and in the Aquitaine basin, the main transverse directions are southwest (Text-Fig. 4). In its Aquitaine segment, the Toulouse fault has the same orientation and bounds the late Jurassic-early Cretaceous basins to the east (BOILLOT 1984). On the North Spanish margin several trends of similar direction have been described (BOILLOT et al. 1974) (Text-Fig. 2). This fault pattern has been assumed to indicate the direction of early Cretaceous rifting (DEREGNAUCOURT & BOILLOT 1982, MALOD 1982), in contrast to a southeast direction of motion of Iberia during the post-Aptian oceanic accretion in the Bay of Biscay.

This kinematics change during the continental rifting has more likely occurred at the same time as the kinematic change observed in the Tagus abyssal plain: Iberia changes from a southwest motion during the late Jurassic-earliest Cretaceous to a southeast motion after the Valanginian. This latter motion has been then continued during oceanic spreading.

c - The late Jurassic phase of Iberian motion

Rotation of Iberia during the late Jurassic implies a relative motion of North America and Eurasia. Otherwise, compressive tectonics would have occurred in the Pyrenean domain between Iberia and Eurasia. Using estimates of the pole of rotation between Iberia and North America and North America and Eurasia respectively (MALOD 1986), we have drawn on Text-Fig. 2 the location of the Iberian Eurasian pole with respect to the angle of opening between North America and Eurasia. In order to get a rotation of Iberia without compression along the Biscay Pyrenean domain and to minimize the motion between North America and Iberia, a tentative pole of rotation has been located eastwards (42° N, 3° E). This results in a scissors opening of the Bay of Biscay during this early phase.

4. Kinematics of Iberia

The kinematics of Iberia is presented in 5 stages:

We neglect here the late Triassic and early Jurassic phases of rifting which correspond to the first phase of the Iberian kinematics.

During the 2nd stage (late Oxfordian to Valanginian, anomaly M-25 to M-10N) Iberia moves in a southwest direction with respect to Eurasia (Text-Fig. 1 A). The entire motion can be described as a scissors opening of the Bay of Biscay and the Pyrenean domain. Rifting occurred also in the Portuguese Galicia rift and is followed during the earliest Cretaceous by oceanic spreading in the Tagus abyssal plain. To the south, between Iberia and Africa, rifting and oceanic spreading is predicted during the late Jurassic and the earliest Cretaceous.

During the 3rd stage (Valanginian to late Aptian, anomaly M-10N to M-0) (Text-Fig. 1 B) Iberia begins to move eastwards in connection with African motion. Oceanic spreading continues to the south and southwest of Iberia, while continental rifting shifts from the Inner Galicia basin to the place between the Galicia Bank and Flemish Cap. To the north, continental rifting is always under way in the Bay of Biscay.

The 4th stage corresponds to the post-Aptian propagation of oceanic spreading west of the Galicia Bank and into the Bay of Biscay (just after anomaly M-0 to anomaly 34) (OLIVET et al. 1984).

Finally, the 5th and last stage corresponds to the convergent motion of Iberia with respect to Eurasia (Text-Fig. 1 D) and will not be discussed here.

5. Paleogeographic consequences

Paleogeographic consequences of plate motions can be regarded from several points of view: At a large scale, the relations between the global kinematic events and the sedimentary events (particularly transgression and regression) are of first interest but are beyond the scope of this paper. I will concentrate here on the relation between the Iberian kinematics and the formation of basins either around the Iberian plate or within it.

5.1 Bay of Biscay and Pyrenees

The kinematic model for the first order motions of major plates (North America, Eurasia, and Africa) provides a frame for the relative second order motion of smaller plates like Iberia. Previous models consider North America and Eurasia attached together during the late Jurassic-early Cretaceous, while our model takes into account the relative motion of North America and Eurasia. For this reason, and instead of the model presented by MASSON & MILES (1984), our model explains fairly well the geological evolution of the Bay of Biscay and the Aquitaine Basin. Particularly, it does not predict any compression in the Pyrenees during the opening of the Bay of Biscay. The late Jurassic-earliest Cretaceous northeast extension explains the formation of sedimentary basins on the north Spanish margin (RAMIREZ DEL POZO 1969, VALENZUELA et al. 1986, PUJALTE 1981, WIEDMANN et al. 1983) and the occurrence of stretching in the Pyrenees (CANÉROT 1987). Subsequent southeastern motion of Iberia (Text-Fig. 4) corresponds to the Bay of Biscay margins formation and the development of the deep basins north of the Pyrenees. Because of the left lateral motion of Iberia with respect to Eurasia it is likely that pull apart basins developed along the Pyrenean domain. This could explain the rhomboidal shape of the Albian Cenomanian basins of the north Pyrenean zone (DEBROAS 1987).

5.2 Basins between Iberia and Africa

Our model provides also a different kinematic evolution between Africa and Iberia. Between Africa and Iberia, west or southeast trending marine basins begin as early as Triassic and Liassic. They are probably related to the eastward motion of Africa and may be interpreted as early aborted rifts or pull apart basins and may include the southern part of the Tagus abyssal plain (KLITGORD & SCHOUTEN 1986). At the end of Dogger and more specifically during the late Jurassic, two well defined continental margins are formed on the north and south borders of the east-west trending North African basin. During this time and until the earliest Cretaceous, calcareous flysches were deposited in this latter basin (WILDI 1983). Our model, which predicts a significant extension between Africa and Iberia during the late Jurassic rifting, is in very good agreement with this geological evolution (Text-Fig. 1 B).

5.3 The west Iberian domain

In this region continental rifting is mainly Triassic-Liassic and late Jurassic in age. The further evolution of the margin has taken place west of the present coast line and most of the early Cretaceous basins are now located at sea (Text-Fig. 3).

West of Iberia the two stage evolution corresponds to a change in the transform directions, but also clearly to a shift of the rifting zone from east to west and south to north: stretching events along the Portuguese trough and probably between the Galicia Bank and Galicia are Kimmeridgian in age, while the onset of oceanic spreading is latest Jurassic-earliest Cretaceous in the Tagus abyssal plain and Albian west of Galicia Bank.

5.4 Intraplate basins

During the separation between Iberia and Eurasia, several basins have formed within the Iberian plate itself. The most significant are perhaps those located within the Iberian domains. For example, Soria and Maestrazgo basins show several stages of extension along different directions, especially during the late Jurassic and the early Cretaceous (VIALLARD 1983, CANEROT 1985). These phases are probably related to the kinematics of Iberia. How we can relate them to the directions of motion is not obvious. Intraplate motions are likely of a lower order of magnitude than the plate motions. A better understanding of the kinematics of Iberia will be a very good guide line to the study of the evolution of these basins.

6. Conclusions

Finally, our attempt to give a kinematic history for late Jurassic-early Cretaceous rifting and to integrate it into the models given for the oceanic spreading phase leads us to a new interpretation of the geological evolution of the region situated around Iberia. The kinematic model presented here can serve as a general frame, but is probably not detailed enough to allow very precise correlation with the geological events at the scale of a sedimentary basin.

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The Iberian Cretaceous: Climatic Implications

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With 5 Text-Figures

RAT, P. (1989): The Iberian Cretaceous: Climatic Implications. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 17-25. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The major features of the Iberian Cretaceous (lithology, pattern) were governed by tectonics and changes in sea level. However, another group of factors was just as determining, but is more difficult to estimate: the climate.

- General features ensured some uniformity. In the first place, high temperatures (Cretaceous is known as a warm period; the Iberian Peninsula was in the tropical zone, as the prominent place of carbonates bears witness). In the second place, rainfall was important (as is proved by weathering and the large amount of mature siliciclastic deposits). As a result siliciclastic and carbonate sediments are often conflicting.
- Geographic characteristics. The various edges of the Iberian plate were probably subject to different conditions, as nowadays: E-W Tethyan currents on the SE side and high pressures or even upwelling on the West Atlantic side. The continental effect and the topographic shelter (rain shadow) may explain the evaporitic internal playas or littoral sebkhas.
- Variations in time in which the large extent of marine sheet certainly played an important part: contrasting climates with marked seasons during the Wealden facies; a certain uniformity due to the middle Cretaceous transgression: again a more continental climate at the end of the Cretaceous period.

Résumé: Les traits majeurs du Crétacé ibérique (lithologie, organisation) ont été commandés par la tectonique et les changements de niveau marin. Toutefois un autre ensemble de facteurs, plus difficile à apprécier, a été tout aussi déterminant: le climat.

- Les traits généraux, facteurs d'unité. En premier lieu, des températures élevées (le Crétacé est connu comme une période chaude: la péninsule ibérique était en zone tropicale; l'importance des calcaires en est un témoignage). En second lieu, une pluviosité importante (altérations, volume des apports siliciclastiques évolués). D'où le conflit fréquent entre siliciclastiques et carbonates.
- Les particularités géographiques. Comme aujourd'hui, il devait y avoir des différences entre les diverses façades de la plaque ibérique (courants téthysiens E-W du côté SE; hautes pressions atlantiques voire upwellings du côté W). L'effet continental, l'abri topographique (effet d'ombre) peuvent expliquer les sebkhas à évaporites internes ou littorales. . .
- Les variations au cours du temps pour lesquelles l'extension de la nappe marine a dû jouer un rôle important: climats contrastés, à effets sai-

sonniers du Wealdien; une certaine uniformisation par la transgression du Crétacé moyen; retour vers plus de continentalité au Crétacé final.

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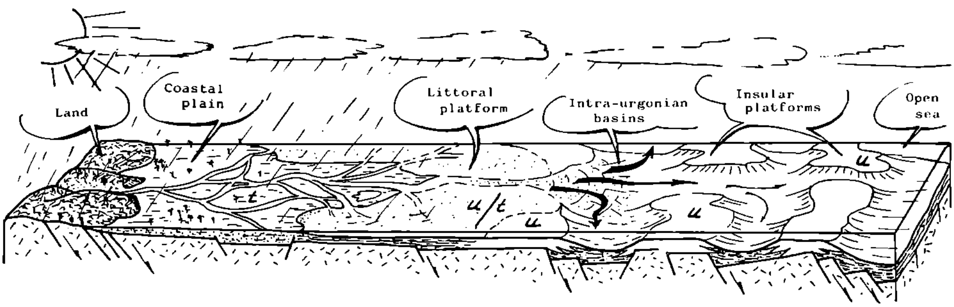
This contribution will probably raise more questions than solve problems. It is only an attempt to understanding the part of climatic factors towards the genesis, nature, structure and volume of a sedimentary body; the examples are from the Cretaceous of Spain.

1. The wet subtropical Urgonian climate

To begin with, the Urgonian period (Aptian to Middle Albian in the Basco-Cantabrian Basin) gives a well-known example: that of a warm and humid climate (RAT & PASCAL 1979, PASCAL 1985).

A warm temperature is deduced from:

- Biological records: ubiquitous occurrence of thick-shelled rudistids and large foraminifera (Orbitolinidae); corals are frequent as well as schizophytal structures. . .
- Sedimentological data: the huge volume of limestone, especially of carbonate mud keeping pace with the rapid subsidence of tilting blocks (Text-Fig. 1); the early lithification.



Text-Fig. 1. The Urgonian platforms in the Basco-Cantabrian Basin (Aptian to Albian): tectonic, climatic and biological control.
u - Urgonian limestone constructions with rudistids. u/t - Interdigitation or mixed terrigenous (allochthonous) and carbonate (autochthonous) material. t - Alluvial spreading.

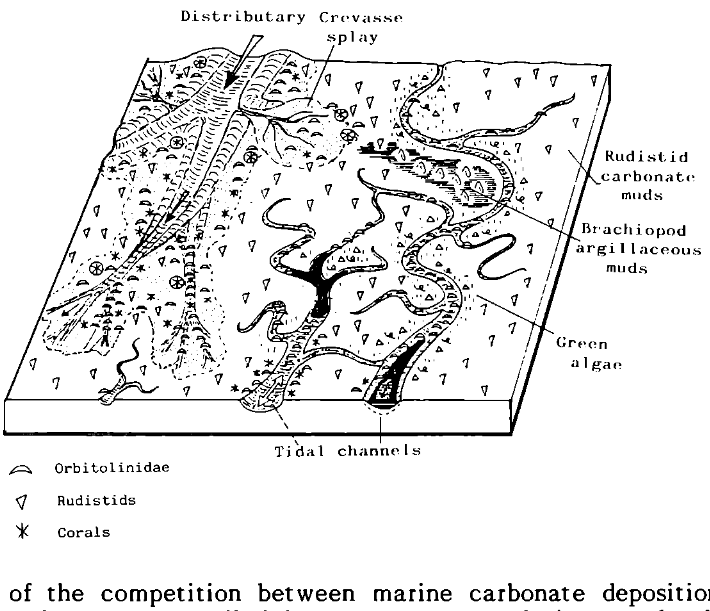
This inference is compatible with the intertropical location of the Urgonian area and also with the fact that temperatures on the earth's surface must have been higher during the Cretaceous than nowadays (BARRON & WASHINGTON 1982).

Rainy conditions are proved by:

- The lack of evaporites
- The occurrence of continental organic matter and lignite
- Brackish environments with charophytes
- The occurrence and great bulk of siliciclastic formations (sand and clay) with fluviatile or deltaic structures
- Residual minerals (chlorite, illite, potassic feldspars) and kaolinite testifying chemical weathering in humid, warm and well drained areas.

However, in this type of climate - which may be termed "wet subtropical" when compared to the present climatic zonation - A. PASCAL (1985) thinks it is possible to discern some variations during the involved period:

- A cooler episode (Lower Gargasian) can be deduced from the ingression of boreal ammonites and the production of glauconite. It could be correlated with a transgression which, at least partly, was due to a eustatic rise. The question is: Was it really a change in the zonal temperature or only a local variation in the course of ocean currents in relation with the greater depth of the Urgonian embayment?
- A more rainy episode (Upper Gargasian) is revealed by the expansion of deltaic bodies (i. e. La Lunada, Santander) and an important phase of dolomitization (caused by the antagonism between fresh and marine waters in very shallow areas) either associated with karstification or not. It is noteworthy that this episode seems to be correlated with the Upper Almaden sandstone of Portugal.
- Fluctuations in the river flow (Lower Albian) are evidenced (Text-Fig. 2) by the alternation of siliciclastic bodies (rainy periods) and rudistid micritic limestones (period of relative drought).



Text-Fig. 2. Detail of the competition between marine carbonate deposition (due to biological settlements controlled by temperature, relative sea level, tectonic stability) and terrigenous input (controlled by tectonic activity and rainfall regime). After PASCAL (1985).

2. Rainy climate with a dry season during the Wealden time

Taking into consideration a longer period of time, two major breaks appear in the continuity of the siliciclastic deposition; the first near the Jurassic-Cretaceous boundary when Wealden sand and clay sedimentation develops; the second in Albian time with the wide-spread Utrillas sand cover.

HALLAM (1984) notes "the sudden replacement of marginal carbonates by coarse paralic "Wealden type" siliciclastics over a huge area extending from the eastern North American margin via Europe and North Africa to the southern USSR and the Arabian Peninsula". Spain participated in this sedimentary change which is particularly well evidenced in the northern basins where "Wealden" fluvial sand plains and thick accumulations overlay "Purbeck facies" in which lacustrine limestones prevail.

Such a sedimentary change implies renewed erosion and fluvial transportation. Nevertheless, what were the major factors controlling this renewal?

HALLAM points out that "Insofar as this change has been discussed, it has usually been in terms of tectonic uplift provoking erosion and hence siliciclastic sedimentation".

It is true that this broad regional revolution affecting erosion and sedimentation coincided with the beginning of the North Atlantic rifting which caused the dislocation of the Atlantic front of Western Europe: Parentis and Adour tectonic basins in southwestern France, Basco-Cantabrian and Soria basins in northern Spain were initiated at that time.

However, the filling of these new-formed sedimentary traps did not begin with major detrital facies. In the Soria Basin, J. SALOMON (1982) described the first deposits (Kimmeridgian in their lower part) as a piedmont system: conglomerates and sandstones in some channels; limestone in permanent lakes; red iron-stained and variegated silts and marly limestones in flood plains. Clastic deposits seem to be located near supplying fault-scarps and do not originate from the denudation of a weathered mantle on wide land areas.

Then probably in Berriasian time, alluvial plains extended, overlain by a detrital system with fluviatile sandstones and conglomerates, lacustrine sandstones and clays, implying the removal of a weathering cover.

These data agree with HALLAM's conclusion: the controlling environmental change was essentially one to a more humid climate provoking an increase of runoff.

In fact, according to the biohexistasy theory (ERHART 1956) often in use in papers of French workers, evolution towards a more humid climate does not necessarily result in more active ablation. The thick vegetation cover that may develop is an efficient protection against removal of the loose weathered mantle by running waters. Increasing rainfall in a warm area with smooth reliefs can give rise to a denser forest, more weathering, a greater loss of material by solution but not to more mechanical ablation.

The seasonal distribution of rainfall can modify the vegetal cover and the erosion as much as or more than the mean annual rainfall. More aggressive are contrasting climates with a dry season which does not favour vegetation and heavy rainfalls occurring on a discontinuous vegetal cover.

For the Wealden formation, HALLAM records from ALLEN (1975), ALVIN et al. (1981) and HARRIS (1981), the presence of irregular growth rings in tree trunks and desiccation cracks which also indicate the existence of a significant dry season.

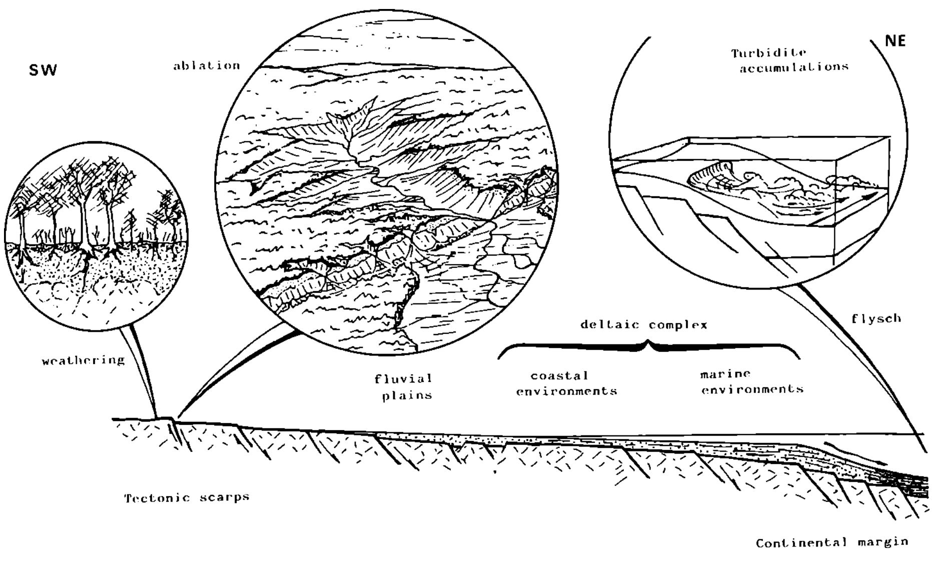
In Spain, as in England, frequent red colours in argillaceous parts of the Wealden succession led to conclude in favour of a dry influence. Hence, the idea of a contrasting climate with marked seasons seems compatible with chemical weathering, temporarily important runoff provoking ablation, transportation and detrital discharges on the one hand and with reddening oxidizing depositional environments or pedogenesis in the low alluvial plains on the other hand.

3. Less efficient rainfall regimes in Upper Cretaceous

The second major break in the continuity of the siliciclastic deposition occurred in Albian times (Text-Fig. 3) with a very large distribution of sands and clays which range from continental environments in the Iberian mountain ranges (Utrillas Formation) to deltaic (Valmaseda Formation) and turbidite accumulations (Black flysch) in the Basco-Cantabrian Basin (FEUILLÉE et al. 1983).

There were tectonic causes related to the oceanic opening of the Biscay Bay and to the framing of a passive, subsiding, North-Iberian continental margin. But such an explanation does not seem sufficient: the Utrillas Formation, the Valmaseda delta, the flysch of Biscay imply a broad river system, a renewed denudation and transportation by water. The extent and volume of the Albian Valmaseda delta can be compared with the Rhône delta today. Much water is necessary.

With the wide distribution of the Utrillas sands began a new sedimentary sequence - Albian to Santonian -; siliciclastics gave way to carbonate deposits. On the Basco-Cantabrian margin, thin hemipelagic limestones or thick carbonate turbidites (MATHEY 1987) followed the clastic Black



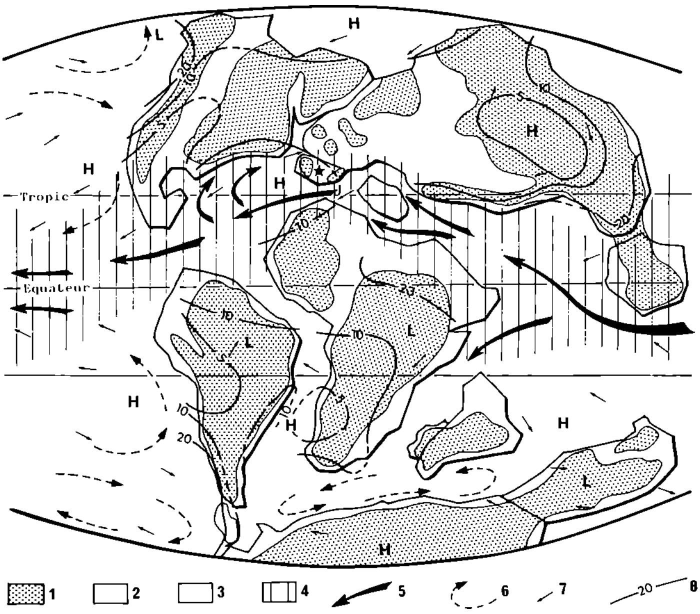
Text-Fig. 3. Utrillas sands and Black flysch system (Northern Spain) in Upper Albian time. Climatic and tectonic control.

flysch. Southwards, in a more proximal position on the Iberian plate (Iberian Ranges), rudistid limestones or dolomites overlay the Utrillas Formation. Such a gradual change may be compared to the change from Albian greensands to chalk in the Anglo-Parisian Basin and in wide parts of Europe north of the Alpine foldbelt.

The explanation by a period of tectonic stability is adequate for the chalk (HALLAM 1984) but it does not seem satisfactory here. We should not forget, there was a period of huge lava-flows along the Basco-Cantabrian margin. A climatic variation towards less erosive rainfall regimes seems necessary.

4. Geographic features

So far, we have assumed changes in precipitation and rainfall regimes within a warm climate of intertropical type during the span of time extending from late Jurassic to Santonian. The geographical aspects must now be taken into consideration (RAT 1982).



Text-Fig. 4. The Iberian plate in the general climatic conditions of the earth during mid-Cretaceous times.
1 - Emerged lands. 2 - Epicontinental seas. 3 - Oceans. 4 - Intertropical belt of warm water deduced from the distribution of large foraminifera and rudistid limestones. 5 - Warm oceanic drift currents. 6 - Cold currents. 7 - Atmospheric circulation in the northern winter: H. High-pressure cells, L. Low-pressure cells (from PARRISH & CURTIS 1982). 8 - Predicted distribution of rainfall. Numbers are only indicative, no units are implied: <5 = low rainfall, 5-10 = moderately low rainfall, 10-20 = moderately high rainfall, >20 = high rainfall (after PARRISH, ZIEGLER & SCOTESE 1982).

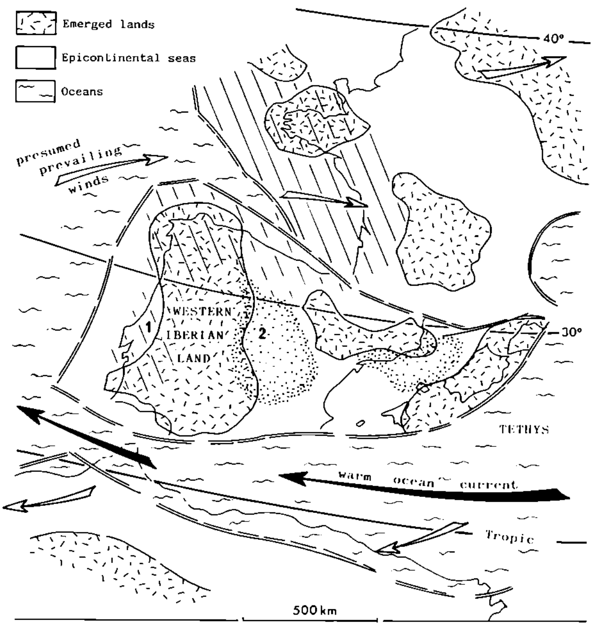
The first explanation for the warm temperatures prevailing in Spain during the Cretaceous is its latitudinal position about 10 to 12° further south than present, i. e. close to the tropic of Cancer (Text-Fig. 4). Moreover, the temperature of the earth's surface during the Cretaceous was most probably higher than today.

As regards precipitation, we must keep in mind:

- The location of the Iberian plate near the Tethyan seaway; a warm ocean belt round the world in the subtropics, with presumably westward currents. Possibly monsoon conditions related to seasonal migrations of the intertropical convergence.
- Variations of both the global marine surface and of the distribution of oceans and epicontinental seas, with two major aspects.

On the one hand, the opening of the North Atlantic and Bay of Biscay was initiated at the end of the Jurassic. According to HALLAM, "it seems very probable that the sharp Early Cretaceous increase of humidity from eastern North America to Middle East, is directly bound up with plate tectonic events creating the early northern Atlantic". Spain was directly facing the new-born ocean to the west and to the north.

On the other hand, the spreading of epicontinental seas as a result of the early and mid-Cretaceous eustatic sea-level rise could favour atmospheric humidity and precipitation. The ingression of a shallow northwest-southeast trending seaway during the Middle Cretaceous into the continental Iberian plate could also affect the local climate.



Text-Fig. 5. Zonal and local influences upon the Iberian climate in Upper Cretaceous times.
1 - Maximum rainfall. 2 - Rain shadow effect.

- In spite of the presumed low altitudes and smooth reliefs of Cretaceous land surfaces inherited from post-Variscan planations, the position of evaporites and frequency of dolomites in the Iberian Ranges could be interpreted by a rainshadow effect (Text-Fig. 5). It seems that the emerged epeirogenic reliefs of the Cretaceous Meseta (Hesperian Massif) were sufficient to shelter, perhaps according to seasons, the Iberian Upper Cretaceous seaway from heavy rainfall. A comparison can be made with the present day relatively dry climate of the interior of Spain.
- The extent of the land area west of the Iberian Upper Cretaceous seaway may have given rise to fluvial systems as important as the Iberian rivers of today: the Tagus (81.000 km^2) or the Douro (97.000 km^2); but it seems that the distribution of altitudes favoured the supply of the river system flowing toward the Basco-Cantabrian margin.

5. Conclusions

The Cretaceous seems to have been a period of weaker contrasts than the present: a reduced gradient of equator-to-pole surface temperature, smooth reliefs of land areas due to post-Variscan erosion and relative tectonic stability, a broad spreading of the seas (nearly three quarters of the earth's surface were occupied by oceanic or epicontinental marine waters). Some authors have even spoken of a "sluggish" atmospheric and oceanic circulation (BARRON & WASHINGTON 1982) though it is denied by others (KEMPER 1987).

In spite of the presumed equability of the Cretaceous climate, local data imply geographical differences even inside a limited area such as the Iberian plate. It can be assumed that the variety of Cretaceous climates (particularly the rainfall regime, the competition between precipitation and evaporation) was not determined by zonal distribution only, but that there really was a mosaic of regional climates as today.

Thus it is dangerous to generalize without caution the inferences from local sedimentological data.

Furthermore, palaeoclimatology can be approached in two very different ways implying different methods of investigation:
- On a large scale: that of the earth's system in search of global controls;
- On a regional scale, analysing available data and collecting precise information in order to draw a picture of local climates and their distribution pattern. Such a patient regional study is possible and open to further investigations.

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Two Transversals through the Cretaceous North African Continental Margin: The Tellian Units of the Western Rif (Morocco) and of the Babors (Algeria)

WOLFGANG KUHNT, Tübingen, and DANIEL OBERT, Paris

With 6 Plates, 14 Text-Figures and 2 Tables

KUHNT, W. & OBERT, D. (1989): Two Transversals through the Cretaceous North African Continental Margin: The Tellian Units of the Western Rif (Morocco) and of the Babors (Algeria). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 27-89. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Cretaceous facies types, paleogeography and subsidence history have been studied along two well outcropping and almost complete transversals through the Tellian units of the Mesozoic North African paleomargin. Sedimentologic observations and characteristic foraminiferal assemblages enabled estimations for Upper Cretaceous paleobathymetries. Both palinspastic reconstructions, and sedimentologic and biofacies analyses lead to the following results:

A. The morphology and evolution of the Cretaceous North African margin which in general represents a classic passive continental margin, were complicated by various factors, such as Upper Cretaceous compressional and lateral movements, the onset of (tectonically controlled?) diapirism in the Albian/Lower Cenomanian, and the existence of intramarginal highs and basins. The most prominent of them are preserved in the northerly domains of the Babors. This complexity has not yet been introduced in the large-scaled palinspastic reconstructions of the Tellian units.

B. The Cretaceous subsidence history of both areas can be divided into four stages:

1. Distension and subsidence of the margin (Lower Cretaceous).
2. A first compressional phase with uplift and slight metamorphism in the Albian/Lower Cenomanian, which affected mainly the northerly ("Haut Tellien") paleogeographic zones. In several of the paleogeographic zones along the North African paleomargin this stage is accompanied by first diapiric movements and resedimentation of Triassic saliferous material.
3. An Upper Cretaceous stage of subsidence (Cenomanian-Santonian).
4. A second compressional phase starts with the Campanian and is reflected by the formation of sedimentary klippes and olistostrome-complexes.

C. During Cenomanian time important intramarginal highs existed in the northerly paleogeographic zones (Brek Unit of the Babors, Tanger Unit in the region of Tangier). As a general trend, sedimentary basins deepened from south to north during Campanian/Maastrichtian giving rise to a characteristic succession of bathymetric zones which have been observed on both transversals. These zones comprise characteristic facies successions as well as foraminiferal assemblages which point to an almost continuous sequence from neritic to lower bathyal/abyssal environments along the western Rif-transversal (probably influenced by Atlantic margin subsidence). Bathymetries below the middle bathyal have not been observed in the Babors.

A comparison of this evolutionary history with the opening of the North Atlantic reveals that times of increased spreading rates in the North Atlantic probably were associated with compressional phases on the North African paleomargin of the Western Tethys.

Résumé: Les types de faciès du Crétacé, leur signification paléogéographique et l'évolution de la subsidence ont été étudiés le long de deux transversales, dans les Unités telliennes de la Paléomarge Mésozoique nord-africaine. Ces transversales ont été choisies en raison de la qualité des affleurements et du large éventail stratigraphique représenté. La reconstruction palinspastique et structurale, l'analyse des faciès, l'évaluation des paléobathymetries du Crétacé supérieur s'appuyant sur des observations sédimentologiques et l'étude des associations de Foraminifères, nous ont conduits aux résultats suivants:

A. La morphologie et l'évolution de la marge crétacée nord-africaine, considerée en général comme une marge continentale passive classique, sont, en fait, compliquées par la superposition d'événements structuraux multiples:

- Existence de diapirisme contrôlé par la tectonique à l'Albien-Cénomanien inférieur;
- Compressions et déplacements latéraux durant le Crétacé supérieur;
- Apparition de hauts-fonds et de bassins intramarges dont les plus remarquables sont préservés dans la partie septentrionale des Babors. L'intégration de toutes ces données n'avait pas encore été réalisée dans le cadre d'une large reconstruction palinspastique des Unités telliennes.

B. L'évolution de la subsidence peut être subdivisée en quatre phases:

1. Distension et subsidence de la marge au Crétacé inférieur.
2. Première phase compressive avec soulèvement et métamorphisme léger à l'Albien-Cénomanien inférieur, surtout marquée dans les zones paléogéographiques septentrionales (Haut-Tellien). Cette phase est accompagnée de diapirisme avec resédimentation de matériel salifère d'origine triasique dans plusieurs des zones paléogéographiques de la paléomarge nord-africaine.
3. Subsidence pendant le Crétacé supérieur (Cénomanien-Santonien).
4. Seconde phase compressive commencant au Campanien et marquée par la formation de klippes sédimentaires et de complexes olistostromiques.

C. Alors que dans les zones paléogéographiques septentrionales (Unité du Brek dans les Babors, Unité de Tanger dans le Rif), existent des zones élevées, importantes, intramarges; une tendance générale à l'approfondissement des bassins sédimentaires se développe du Sud vers le Nord pendant le Campanien-Maestrichtien. Pendant cette période une succession Sud-Nord de zones bathymétriques bien caractérisées a été observée sur les deux transversales. Ces zones montrent, tant par les successions de faciès caractéristiques que par les associations de foraminifères, que l'évolution paléogéographique est presque continue depuis le milieu néritique jusqu'au milieu abyssal le long de la transversale du Rif occidental (probablement sous l'influence de la subsidence de la marge atlantique), tandis que dans les Babors aucune bathymétrie inférieure au bathyal moyen n'a été observée.

Une comparaison de cette évolution avec l'ouverture de l'Atlantique nord révèle un certain synchronisme entre la période d'accroissement du taux d'expansion dans l'Atlantique nord et les phases compressives survenant sur la paléomarge nord-africaine de la Téthys occidentale.

Kurzfassung: Die kretazische Fazies- und Subsidenzentwicklung zweier Transversalen durch den spätmesozoischen nordafrikanischen Kontinentalrand mit relativ kompletten paläogeographischen Abfolgen wurde verglichen. Die palinspastische Abwicklung, Rekonstruktion der Subsidenzgeschichte, Faziesanalyse und Abschätzung der paläobathymetrischen Position einzelner Serien aufgrund sedimentologischer und mikropaläontologischer Daten läßt folgende Grundzüge der kretazischen Entwicklung dieses Kontinentalrandes erkennen:

A. Die Morphologie und Entwicklung des kretazischen nordafrikanischen Kontinentalrandes, der im allgemeinen einem klassischen passiven Kontinentalrand entspricht, wurden durch oberkretazische kompressive und Transversal-Bewegungen, den Beginn eines (tektonisch beeinflußten?) Diapirismus an der Wende Alb/Cenoman und die Existenz von intramarginalen Becken- und Hochgebieten vor allem in den Babors stark kompliziert. Diese Komplexität der Kontinentalrand-Morphologie blieb bei den bisherigen großräumigen palinspastischen Rekonstruktionen der Tell-Einheiten noch weitgehend unberücksichtigt.

B. In der kretazischen Subsidenzgeschichte lassen sich - mit nur wenigen lokalen Sonderentwicklungen auf beiden untersuchten Transversalen - vier Stadien unterscheiden:

1. Distension und deutliche Subsidenz bestimmten die Entwicklung während der Unterkreide.
2. Erste kompressive oder transpressive Bewegungen, verbunden mit der Heraushebung einzelner Blöcke und schwacher Metamorphose vor allem in den nördlichen ("Haut-Tellien") paläogeographischen Zonen, wurden in der höheren Unterkreide der Babors beobachtet. Diese Phase wurde von einsetzendem Diapirismus und Resedimentation triadischen Salinars in den meisten paläogeographischen Zonen entlang des gesamten nordafrikanischen Kontinentalrandes begleitet.
3. Erneute verstärkte Subsidenz erfolgte vom Cenoman bis ins Santon.
4. Eine zweite kompressive Phase begann mit dem Campan und wurde von gravitativen Massenbewegungen (sedimentären Klippen und Olisthostrom-Komplexen) in allen Abschnitten des Kontinentalrandes begleitet.

C. Während bedeutende intramarginale Hochgebiete mit neritischen Faziesbereichen vor allem während des Ober-Alb und Cenoman für die nördlichen paläogeographischen Zonen (Brek-Einheit der Babors, externe Tanger-Einheit im Raum Tanger) charakteristisch sind, kann während des Campan/Maastricht auf beiden Transversalen eine Abfolge von flacheren zu tieferen bathymetrischen Zonen von Süd nach Nord beobachtet werden. Dabei wurden in den distalen Bereichen der Rif-Transversale (Tanger-Einheit) bathymetrische Zonen des tiefen Bathyal bis Abyssal erreicht, während Fazies und Foraminiferen-Vergesellschaftungen der nördlichen Zonen auf der Babors-Transversale (Brek-Einheit) maximal mittel-bathyale Wassertiefen anzeigen.

Beim Vergleich dieser Subsidenzentwicklung mit der Öffnungsgeschichte des Nordatlantik und den Relativbewegungen zwischen Afrika und Eurasia lassen sich die am nordafrikanischen Kontinentalrand beobachteten kompressiven/transpressiven Phasen mit erhöhten Spreading-Raten im Atlantik (ab Unter-Apt, Anomalie M0) und der beginnenden Konvergenz zwischen Afrika und Eurasia (Subduktion am Nordrand Iberias ab Campan) korrelieren.

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1. Introduction

The Tellian units of the North African Maghrebidian foldbelt represent the sediments on the Mesozoic - early Tertiary North African continental margin of the transitional ocean between the North Atlantic and Mesogea (Text-Fig. 1). Due to Miocene overthrusting, this continental margin succession is rather incomplete and badly preserved in most parts of the foldbelt. Exceptions are found on the Rif and Babors transversals due to their particular structural position. In the overthrusted units of both these areas almost complete late Mesozoic sedimentary successions are outcropping. The palinspastic reconstruction of these units is possible due to a relatively low value of tectonic shortening. Thus, this area is favourable for studying the influence of tectonics on sedimentation (MITCHELL & READING 1978) along a late Mesozoic passive continental margin which is strongly influenced by transform movements.

2. Palinspastic Reconstruction

2.1 Babors (Text-Fig. 2)

The Babors domain (Algeria) belongs to the north-eastern part of the Tellian Zone and separates the "Petite" from the "Grande" Kabylie. The most internal parts of the Tellian realm are preserved in this particular

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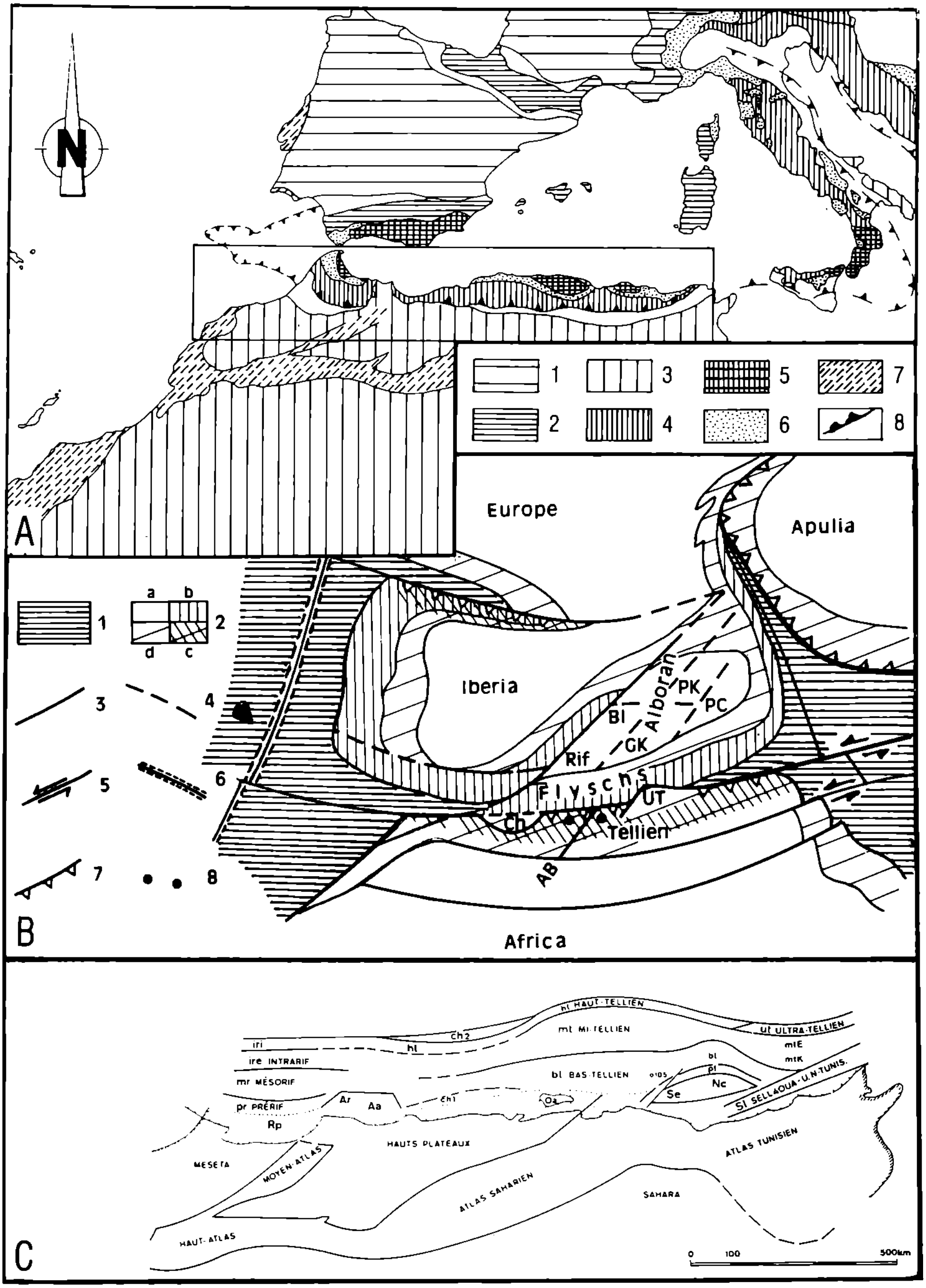
Text-Fig. 1. Structural setting and palinspastic reconstructions of the Cretaceous North African Margin.
A: Simplified structural map of the Western Mediterranean. Position of the Mesozoic North African Margin is indicated. 1. Stable Europe - 2. Deformed European continental margins - 3. Stable Africa - 4. Deformed North African Margin - 5. Microplates - 6. Deformed oceanic basins - 7. Mesozoic epicontinental seas - 8. Front of overthrusting.
B: Hypothetical sketch of the Western Mediterranean area in the Upper Cretaceous (modified from OBERT 1984). 1. Oceanic area - 2. Continental area (a. land and epicontinental sea, b. thinned continental crust with deep-water sedimentation, c. metamorphized zone, d. continental margin) - 3. Major fracture - 4. Future or supposed major fracture - 5. Transform fault - 6. Active spreading - 7. Compressive/transpressive margin (with subduction?) - 8. Magmatism. - Ch: Chouala Unit - UT: Ultra-Tellian Unit - AB: Babors fault (WILDI 1983). - The ALKAPECA microplate is, or will be, separated into different blocks: BI: Internal Betic domain - GK: Grande Kabylie - PK: Petite Kabylie - PC: Peloritano-Calabrais block.
C: Palinspastic reconstruction of the Cretaceous North African paleomargin (from WILDI 1983).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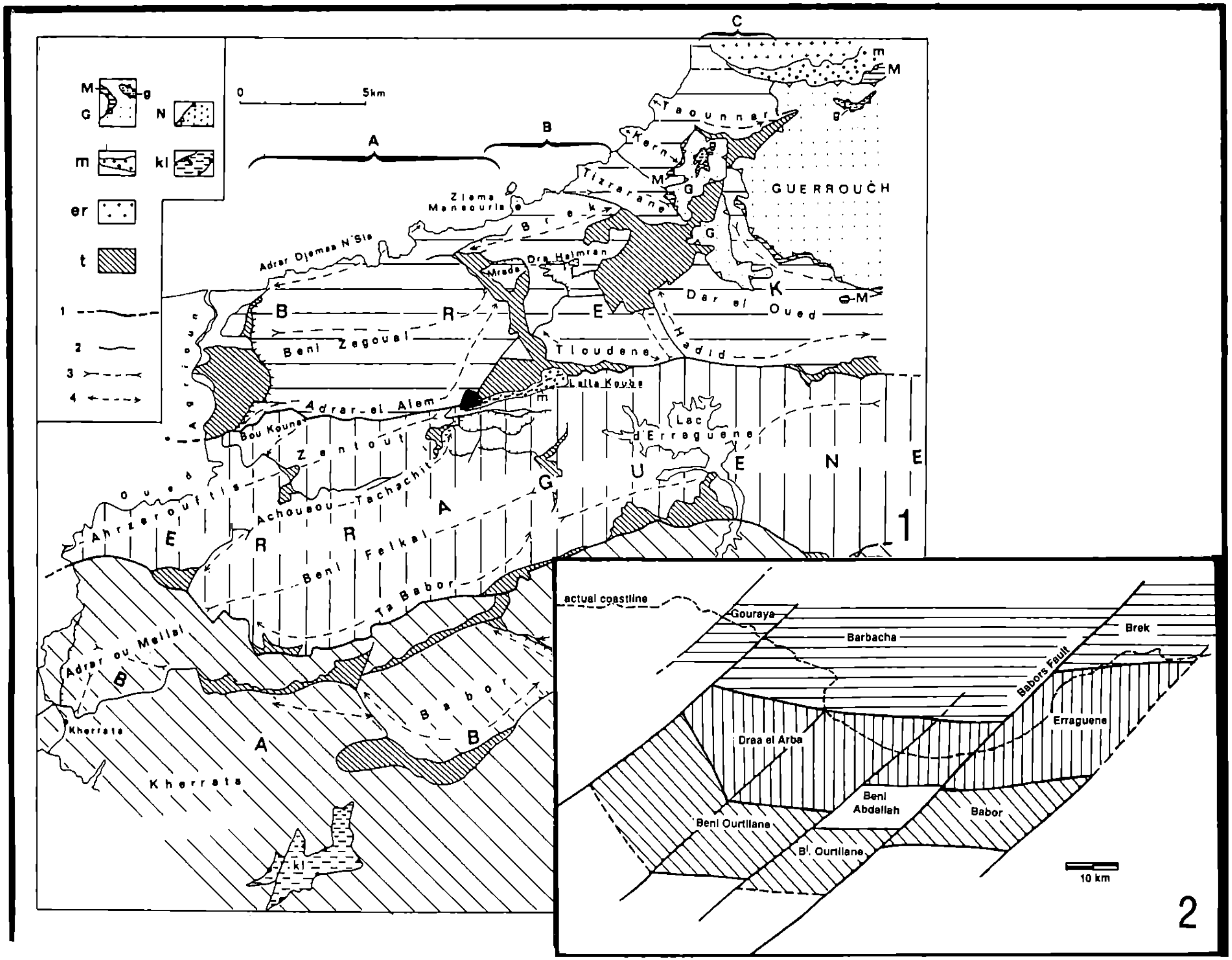

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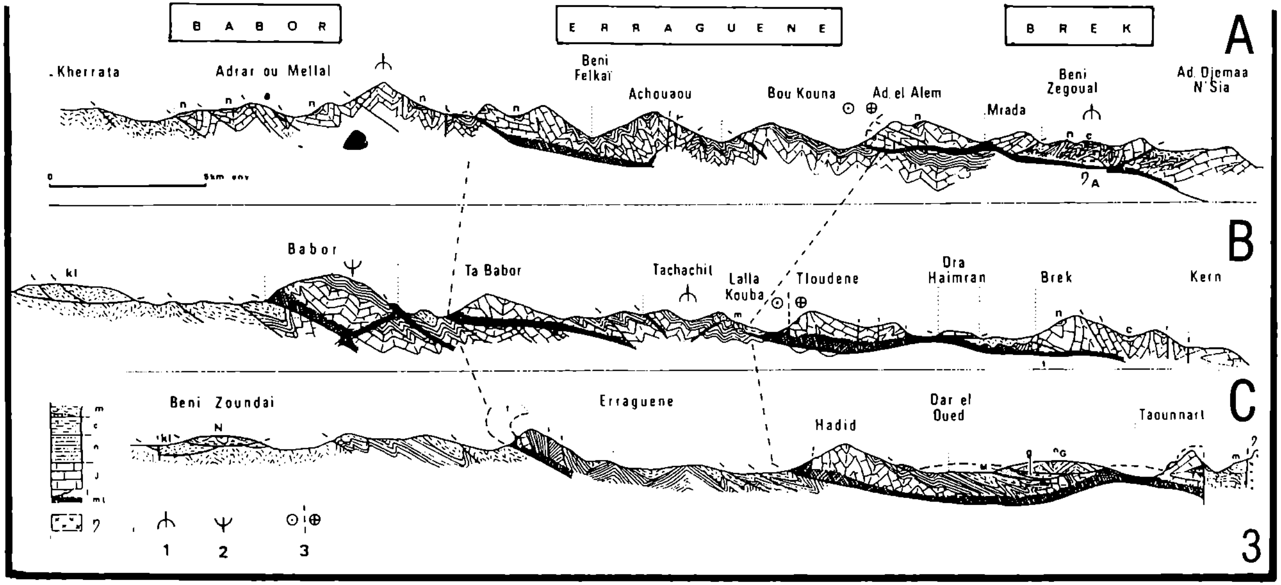


Text-Fig. 2. Simplified structural map, palinspastic reconstruction and cross sections of the Babors. 1. Structural map of the Eastern Babors with main tectonic units (from OBERT 1984). G: Guerrouch nappe (Mauretanian) - M: Massylian Flysch - g: Oligocene micaceous sandstone - N: Numidian Flysch - kl: Klippes with Upper Cretaceous-Eocene sediments - m: Miocene (Burdigalian-Langhian) - er: El Aouana magmatic complex - t: Melanges with resedimented Triassic material. - 1. Boundaries of main Tellian Units - 2. Main faults - 3. Syncline or synclinorial axes - 4. Anticline or anticlinorial axes.

2. Palinspastic reconstruction for the Upper Cretaceous of the main structural units of the Babors.

3. Synthetic cross-sections of the eastern Babors (trace of the sections is indicated in Text-Fig. 3/1). Magmatic rocks: massif of El Aouana. mt: Melanges with Triassic material and Mesozoic sediments originating from the Tellian Units of the Babors. This formation forms the base of overthrusted units, and it is injected into the main faults. - t: Triassic sediments - J: Jurassic - n: Lower Cretaceous - c: Upper Cretaceous-Eocene - m: Miocene. - 1. Westward overthrusting - 2. Eastward overthrusting - 3. Sinistral strike slip fault.

setting; from this it is possible to study the whole Mesozoic Tellian sequence. Tertiary compression results in thrusting of the entire domain and southward movement of Tellian nappes mainly using the Mid-Cretaceous discontinuity surface as shear plane. The lower part of the series (Jurassic and Lower Cretaceous) is sheared off in slices with southward vergency. This main tectonic deformation is followed by several weaker tectonic phases with limited back thrusting, reactivation of southward overthrusting and sinistral strike-slip faulting.

The Tellian of the Babors is mainly formed by three structural units, which are southward overthrusted and which are, in general terms, still in contact with their basement. These are from north (hanging-wall position) to south (foot-wall position): (1) Brek Unit or Brek-Gouraya Unit ("Haut-Tellien" according to the definition of WILDI 1983), (2) Erraguene Unit ("Mi-Tellien"), and (3) Babor Unit ("Bas-Tellien"). In accordance with the nomenclature used in the Gibraltar Arch area, the southerly, foot-wall units are also called "external", the northerly, hanging-wall units "internal". A characteristic feature is the presence of Triassic material ('mélange' of evaporites and pelites) in the main faults and thrust planes, which is either owing to injection, or faulting along the saliferous level.

In the north, the Brek Unit disappears beneath the Mediterranean Sea. The laterally, in Algeria observed "Ultra-Tellian Unit" perhaps forms the northern continuation of the Brek Unit.

In the south, the Babor Unit is slightly overthrusting the Djemila Nappe (VILA 1980), which is mainly formed by sheared off Upper Cretaceous material of the external Tellian zones from the Babors and adjacent areas.

The whole structural building of the Babors domain is cut by several, mainly SW-NE orientated sinistral strike-slip faults, which have been active since the Middle Jurassic and have still affected the nappe pile in the Middle Miocene.

The palinspastic reconstruction of this highly complicated deformational history (Text-Fig. 2, OBERT 1981, 1984) reveals a relative north-south directed shortening of the Tellian realm of about 50 % and a relative position of the different blocks, which was similar in the Upper Cretaceous and Present.

2.2 Western Rif and Gibraltar Arch area (Text-Fig. 3)

In the western and central Rif, the remainders of the Mesozoic North African continental margin are found in the external (= Tellian) Units. Three main structural units and corresponding paleogeographic zones can be distinguished (SUTER 1980a, b): (1) The northern and hanging-wall Intrarif units, (2) the Mesorif Zone, and (3) the foot-wall and most southern Prerif Zone. The external zones of the central and western Rif show an arch-like rotation of the structures due to Miocene collision of the Alboran block (OLIVIER 1984, LEBLANC & OLIVIER 1984). An east-west strike in the eastern part bends to a north-south strike close to the Strait of Gibraltar. This feature strongly complicates palinspastic reconstructions. In order to avoid misunderstandings, the structurally high Tellian units have been termed "internal", revealing distal Mesozoic facies and transitional features to the Massylian Flysch units. The proximal Prerif Zone with its more or less continuous contact to the foreland represents the most "external" part of the Tellian realm within the Gibraltar Arch area.

Text-Fig. 3 gives an overview of the palinspastic reconstruction of the main structural and facies units of the Tellian domain in the Rif. Units 2-6 belong to the Intrarif (= "internal") of SUTER (1965 and 1980a, b), Unit 7 represents the Mesorif, Units 8-10 are subdivisions of the Prerif ("external").

The present structural situation and the late Mesozoic paleogeography probably are strongly influenced by two important WSW-ENE and SW-NE orientated strike-slip faults, the Jebha-Chrafate-Loukkos Fault in the NW and the Nekor fault in the SE, respectively. These two faults limit the more or less autochthonous Mesorif Zone to the NW and SE. Their influence on Mesozoic paleogeography of the external "Tellian" realm is not yet very well understood, but at least for the Jebha-Chrafate-Loukkos Fault it seems to be of larger importance than previously assumed.

3. Facies types, foraminiferal assemblages and paleobathymetry

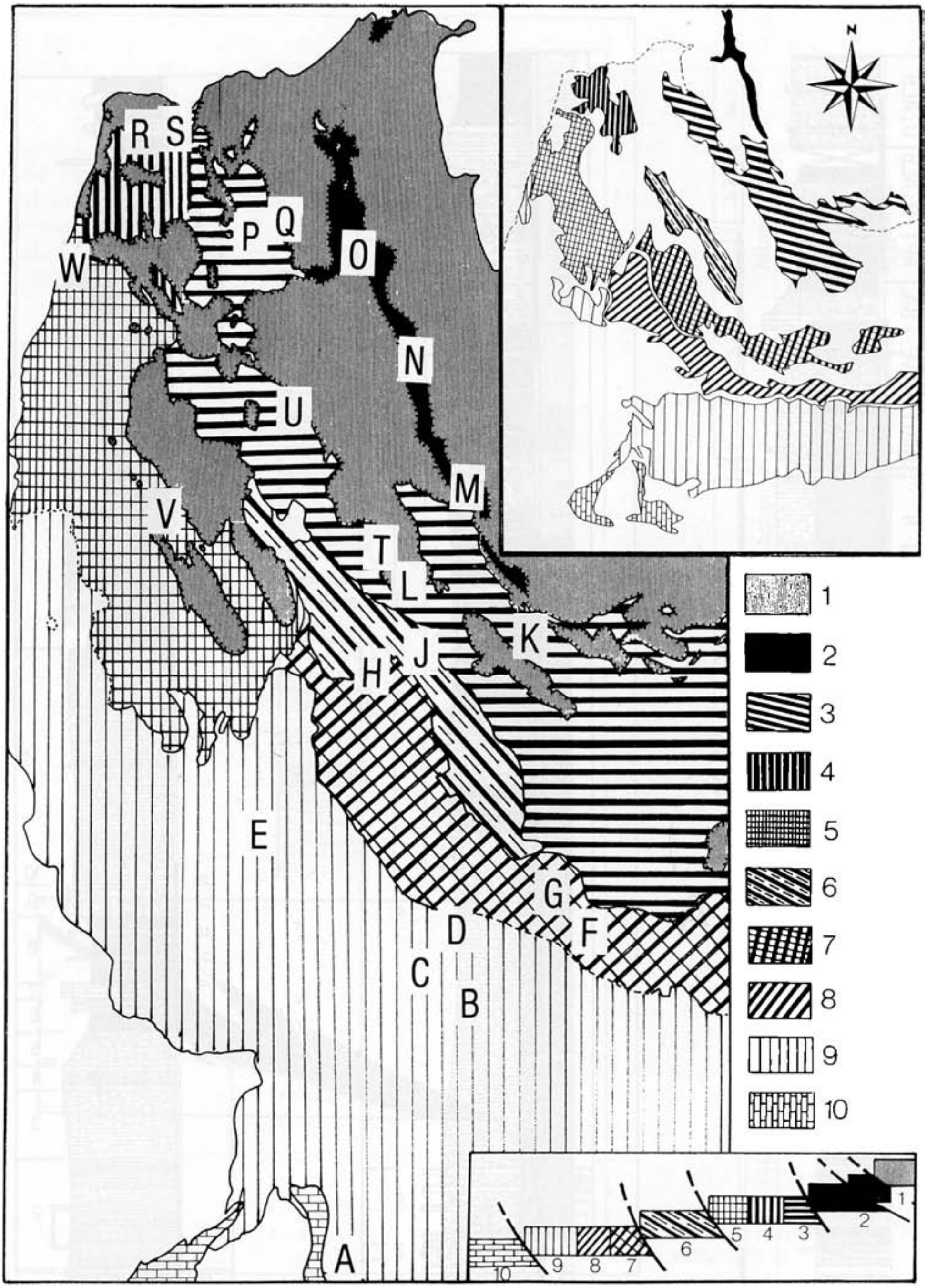
3.1 Western Rif Transversal

3.1.1 Sedimentology and facies distribution

The Cretaceous transgressive cycle of the **Rides Prérifaines** (locality A in Text-Fig. 3) commences with a calcareous breccia upon Bajocian sediments in the El Kelaa section (FAUGERES 1978). The first biostratigraphic marker found a few meters above the basal breccia (*Mortoniceras aequatoreale* KOSSMAT), indicates Upper Albian (KUHNT 1987). Thick lamellibranchiate shell-beds with abundant oysters characterize Upper Albian to Cenomanian sedimentation. The Turonian is represented by dolomitic limestones of about 15 m thickness, containing pelagic macrofaunas (*Inoceramus*, ammonites) and plankton-dominated foraminiferal assemblages. A comparison with the Turonian of the Moroccan Meseta (WIEDMANN et al. 1978) is possible. Coniacian and Santonian sedimentation of the Rides Prérifaines consists of a few tens of meters of light grey calcareous marls bearing planktonic and benthic foraminifers. Campanian and Maastrichtian has not yet been observed, probably due to emersion. This sequence is overlain by Paleocene sediments which contain reworked Upper Cretaceous microfossils (FAUGERES 1978).

Cretaceous sedimentation of the **Prerif and external Mesorif** Zones (localities B-G in Text-Fig. 3) is characterized by pelagic limestones with gradually increasing siliciclastic input during Berriasian-Barremian (Text-Fig. 4, A-C). The Aptian-Albian consists of pelitic and siliciclastic sediments, which reveal reduced sedimentation rates and hiatuses in the southern (shelf-) sequences of the Prerif and channelized slope sedimentation in the more northerly sequences of the Mesorif Zone (Text-Fig. 4, D). Mid-Cretaceous sediments of these zones are made up by pelagic limestones and marls, locally with biosiliceous and organic-rich spikes at the Cenomanian/Turonian boundary (Text-Fig. 5). Generally, Upper Cenomanian to Turonian sedimentation of these zones is characterized by stratigraphical gaps due to sediment gravity transport of the sediments to deeper parts of the basin. Marls and marl/limestone rhythms with high carbonate contents are typical for the depositional environment in the Upper Cretaceous. Sediment gravity movements such as slumps and slides can be observed locally. In the Mesorif these movements become an important factor in sedimentary deposition;

Text-Fig. 3. Geological sketch map with sampled areas, palinspastic reconstruction and simplified structural relations of the External (Tellian) Units in the Western Rif. Internids and those flysch units which are related to the Internids with respect ot their resediments (THUROW 1987) are combined. Nomenclature according to DURAND DELGA et al. (1962), SUTER (1965, 1980a, b), and KUHNT (1987).
1. Flysch Units and Internids - 2. Internal Tanger Unit, and parts of the Massylian Flysch Units which are not related to the Alboran Margin - 3. Bab Taza Sequence of the external Tanger Unit and Ketama Unit (mainly Lower Cretaceous which forms at least partly the stratigraphic base of the Tanger Units) - 4. External Tanger Unit in the Tangier region - 5. Nappe du Habt - 6. Loukkos Zone - 7. Mesorif Zone - 8. Prerif Zone (Internal part = "Ligne des Sofs", not differentiated on the geological sketch map) - 9. Prerif Zone (External part = Miocene Prerif-Olistostrome) - 10. Rides Prérifaines (parautochthon margin of the Moroccan Meseta).
Localization of type-sections for late Cretaceous formations (coordinates according to Carte du Maroc 1/50000 - Type 1922; for detailed description of the sections see KUHNT 1987):
A. Rides Prérifaines; section P3 (500,5 - 502,0/380,0 - 380,5), northern flank of the Oued Daya valley - B. Prerif; outcrops in the Kariat Mesnaoua area, W of Kariat Bab Sebt - C. Prerif; section P11 (503,500 - 503,600/441,200 - 441,400), N of Sidi Bousbeur - D. Prerif; outcrops in the Djebel Mguedrouz area - E. Prerif; section P2 (472,870/465,520), southwestern flank of Djebel Ouartilou, W of Ouezzane - F. Mesorif; section M91 (535,900 - 536,500/ 446,500 - 447,0), near Tazzarine, W of Rhafsai - G. Mesorif; section M3 (527,400 - 527,950/453,250 - 453,450), Oued Aoudour - H. Mesorif; section M83 (498,650/483,200), W of Mokhisset - J. Loukkos Zone; sections L19 (503,800/485,0), L22 (503,400/485,650), road sections between Mokhisset and Souk el Had - K. Bab Taza; section Fi6 (518,850/493,600), type section W of Bab Taza - L. Bab Taza; section M17 (500,900/499,700), southern flank of the Djebel Sougna - M. Tanger-Intern; section M41 (511,250 - 511,400/508,0 - 508,250), N of Chaouen - N. Tanger-Intern; sections M5 (502,750 - 503,200/526,450 - 526,650), M52 (502,500 - 502,900/530,050 - 503,300), near Souk el Arba des Beni Hessane - O. Tanger-Intern; section B27 (491,700/551,300), W of Tetouan - P. Tanger-Intern (?); section rh1 (474,450/565,650) - Q. Melloussa (Massylian); section M9 (476,150/566,750), type area W of Melloussa village - R. Massylian; section E27 (453,100/ 572,850), near Mediouna, W of Tangier - S. Tanger-Extern; sections M63 (461,100/574,700), village of Tangier, zi2 (458,0/569,950), near Sidi Rahhali, W of Tangier - T. Intermediate Ain Lahcen sequence; section R4 (498,250 - 498,500/503,900 - 504,0), type section W of the village Ain Lahcen - U. Intermediate Er Rkayah sequence; section M84 (484,600/529,650), type section at Er Rkayah - V. Nappe du Habt and intermediate Oued Mekhacen sequence; sections M51 (458,0 - 458,150/516,500), W of Souk es Sebt, M16 (463,500/513,800), M16A (463,500/513,200), Oued Mekhacen - W. Tanger-Extern; section [illegible]38 (445,650/547,150), southernmost outcrop at the Tangier-Larache main road.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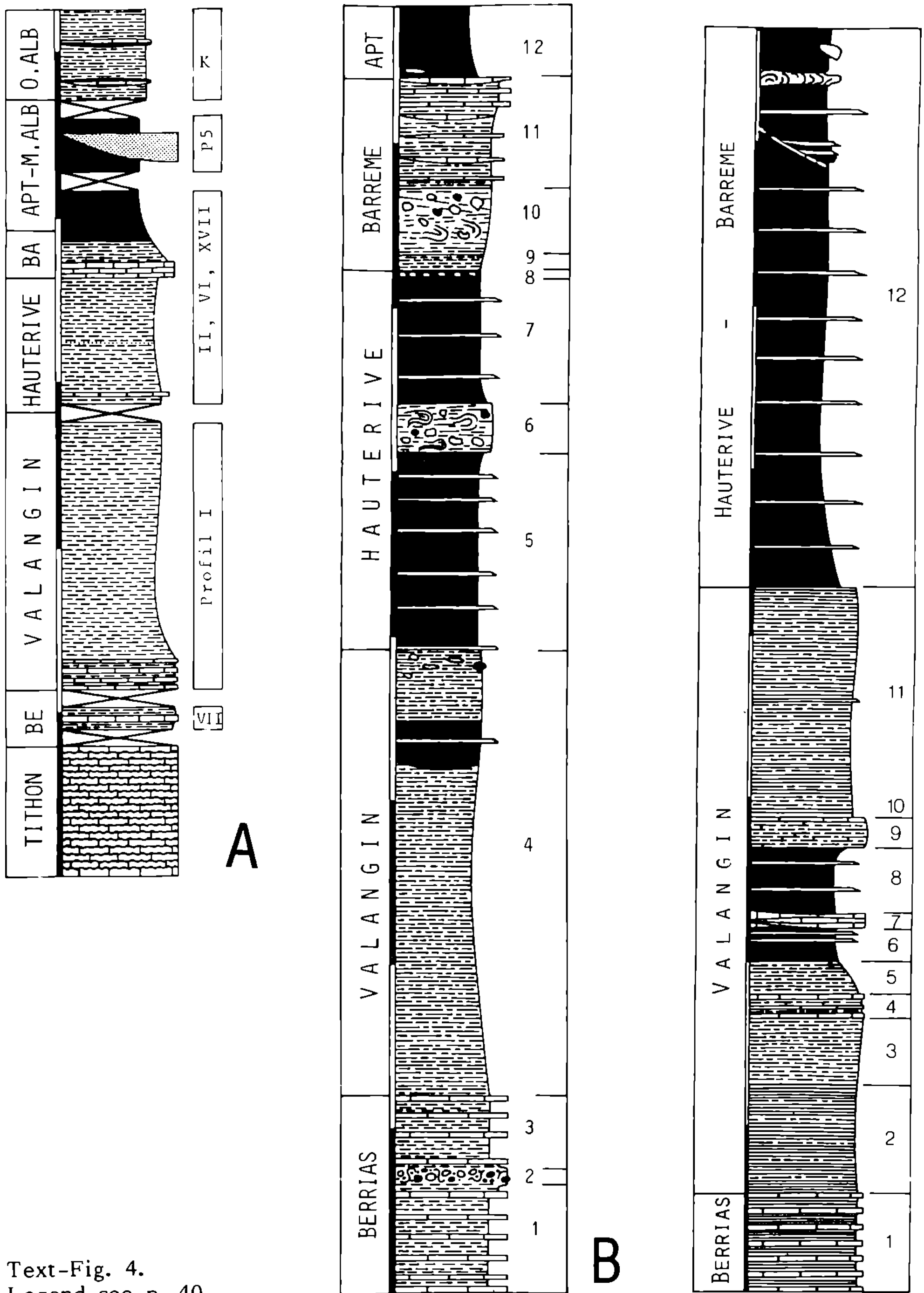
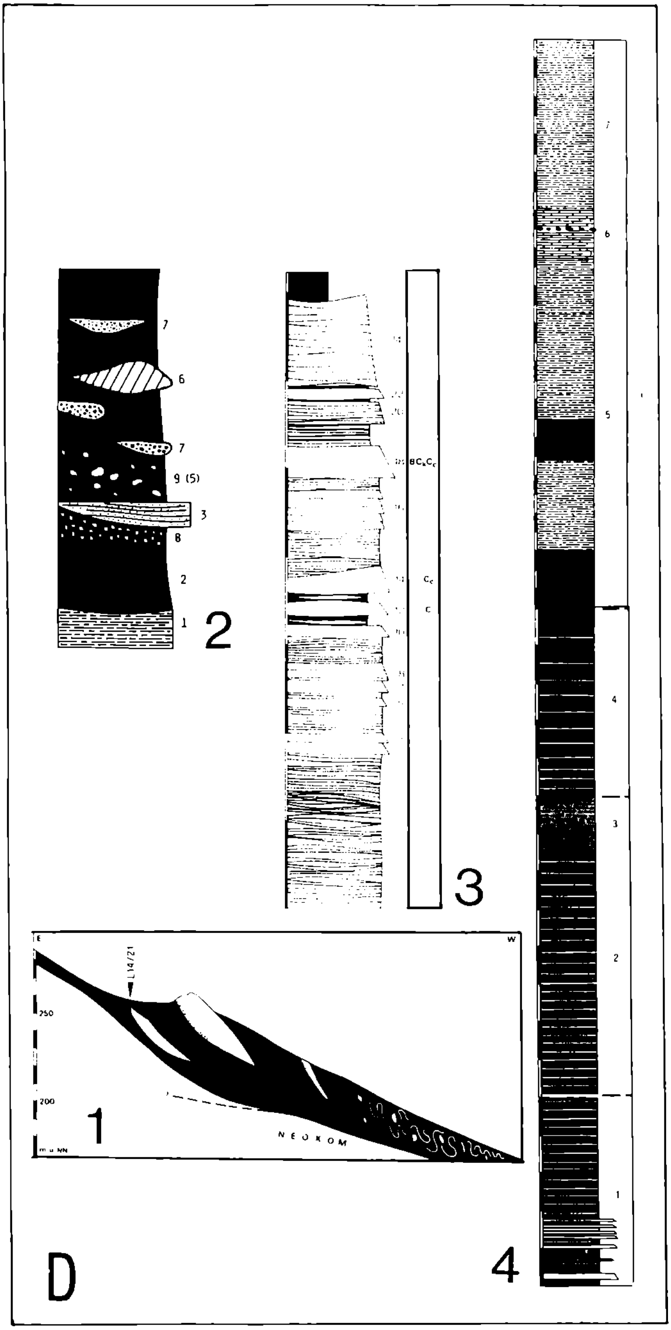


Text-Fig. 4.
Legend see p. 40.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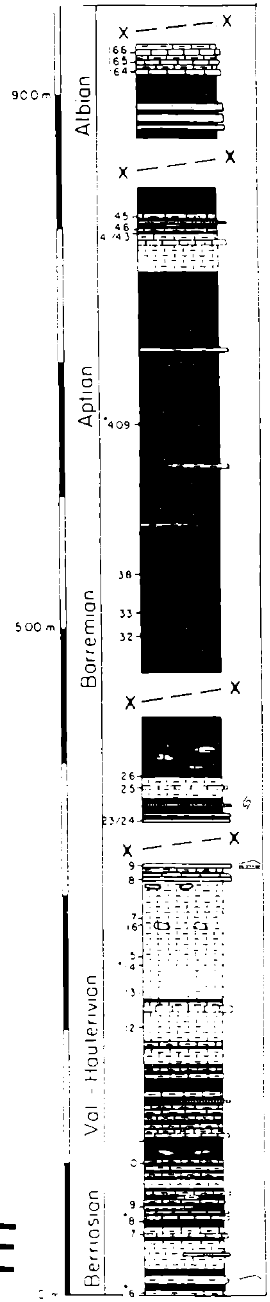
900 m
500 m
Albian
Aptian
Barremian
Val - Hauterivian
Berriasian

E

debris-flows and proximal calciturbidites are common. Although most of the outcrops are strongly tectonized and no complete section of the entire Upper Cretaceous is outcropping, sediment thickness is considerable in several of the outcrops studied (e. g. more than 300 m for the Maastrichtian of the Mesorif Zone, KUHNT 1987).

Sedimentation of the **Loukkos Zone** (locality J in Text-Fig. 3) begins above a tectonic contact with an Albian siliciclastic sequence followed by Upper Albian black laminated limestones and marls, and Cenomanian marls (Text-Fig. 6, B). Within the Loukkos Zone Turonian sediments have been observed in two different facies types (Text-Fig. 5): Limestones, marls and cherts, showing intense slump features and bearing abundant and well

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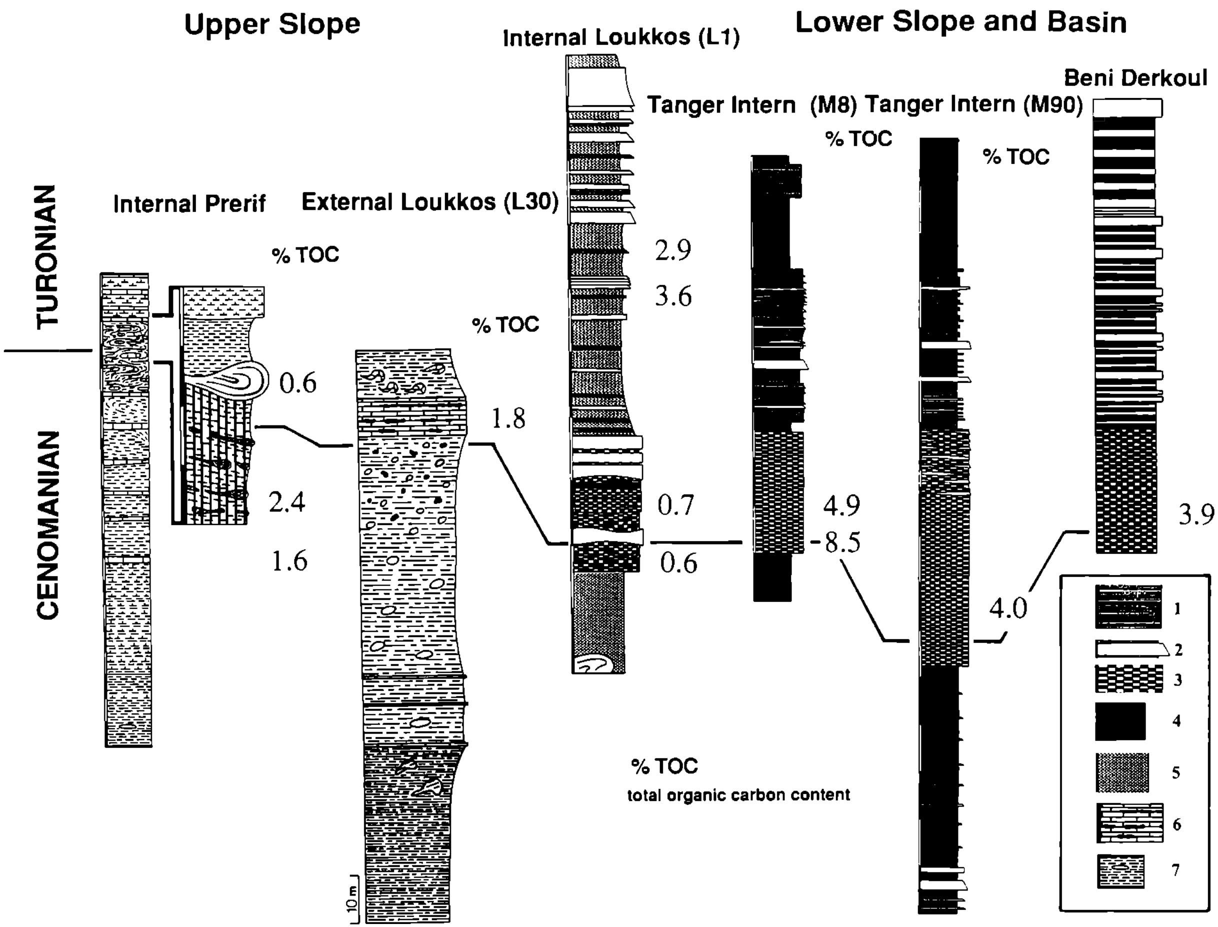
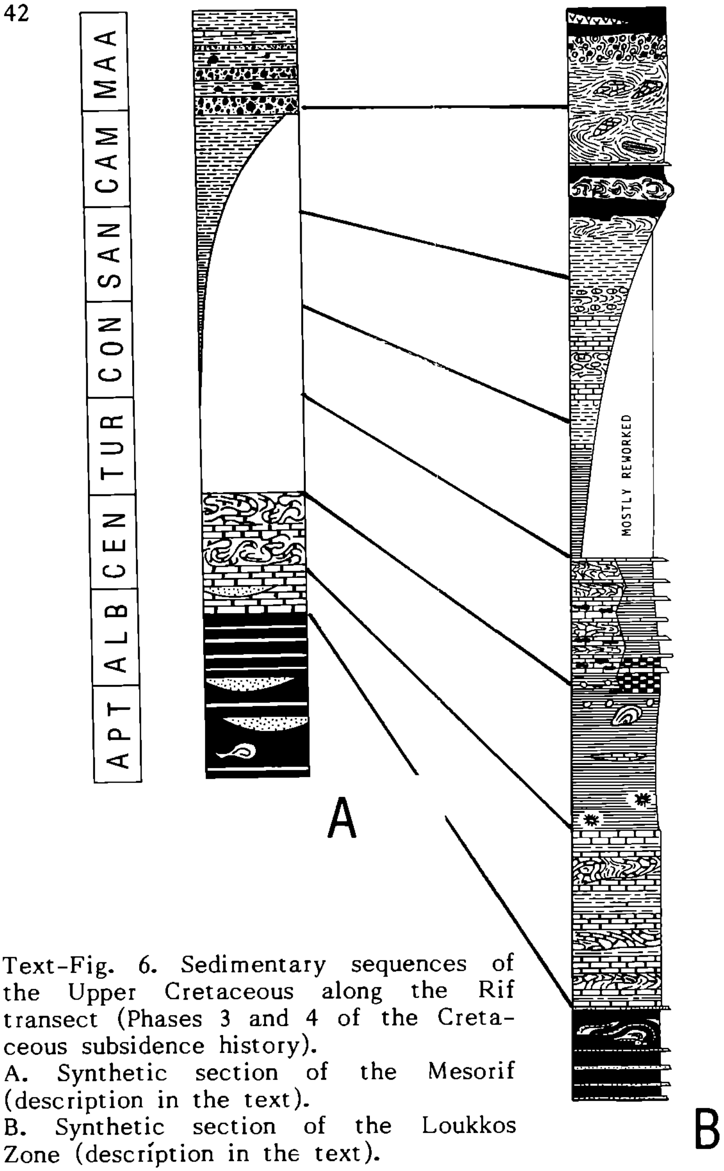
Text-Fig. 4. Sedimentary sequences of the Lower Cretaceous along the Rif transect (Phases 1 and 2 of the Cretaceous subsidence history).
A. Synthetic section of the Lower Cretaceous of the Prerif (Moyen Ouerrha region). Scale bar is 50 m.
B. Lower Cretaceous section at the limit of the Prerif and the Mesorif (Djebel Mguedrouz section). Scale is 50 m.
C. Lower Cretaceous of the Mesorif (Zaouia Amjot section south of Tabouda).
D. Aptian-Albian terrigenous sedimentary cycle in the Mesorif. - D1: Channelized slope facies (cross-section of the Djebel north of Ouezzane). - D2: Chaotic slope facies with redeposition of neritic biogenic *Orbitolina*-facies and Triassic sediments. Synthetic section of the Djebel Beni Izzou north of Rhafsai, not to scale: (1) Neocomian marly shales; (2) silty claystones, background sedimentation; (3) channel sedimentation with mixed terrigenous and calcareous biogenic detritus; (6) Triassic resediments, i. e. evaporites, ophitic volcanics; (7) allodapic limestones with neritic biogenic detritus; (8, 9) slumps and olistolites of pelagic limestones. - D3: Channel-margin facies in the Forêt d'Izzarene north of Ouezzane (terrigenous turbiditic sedimentation, corresponding to facies E of MUTTI & RICCI LUCCHI (1972) with the following characteristics: (1) high sand/shale ratio; (2) thin-bedded; (3) discontinuous bedding, i. e. flaser bedding, rapid lateral disappearance of single beds; (4) coarse-grained, badly sorted sandstones with layers of plant-debris. Scale is in meters. - D4: Transition of the Albian siliciclastic sequence to late Albian predominantly calcareous sedimentation. Azib ej Jaafra section, northeast of Taounate. Scale is in meters.
E. Synthetic stratigraphic section of the southern part of the Ketama Unit in the Dhar Souk-Tamchecht area (from ZÜBELI 1982). Scale bar is 100 m.

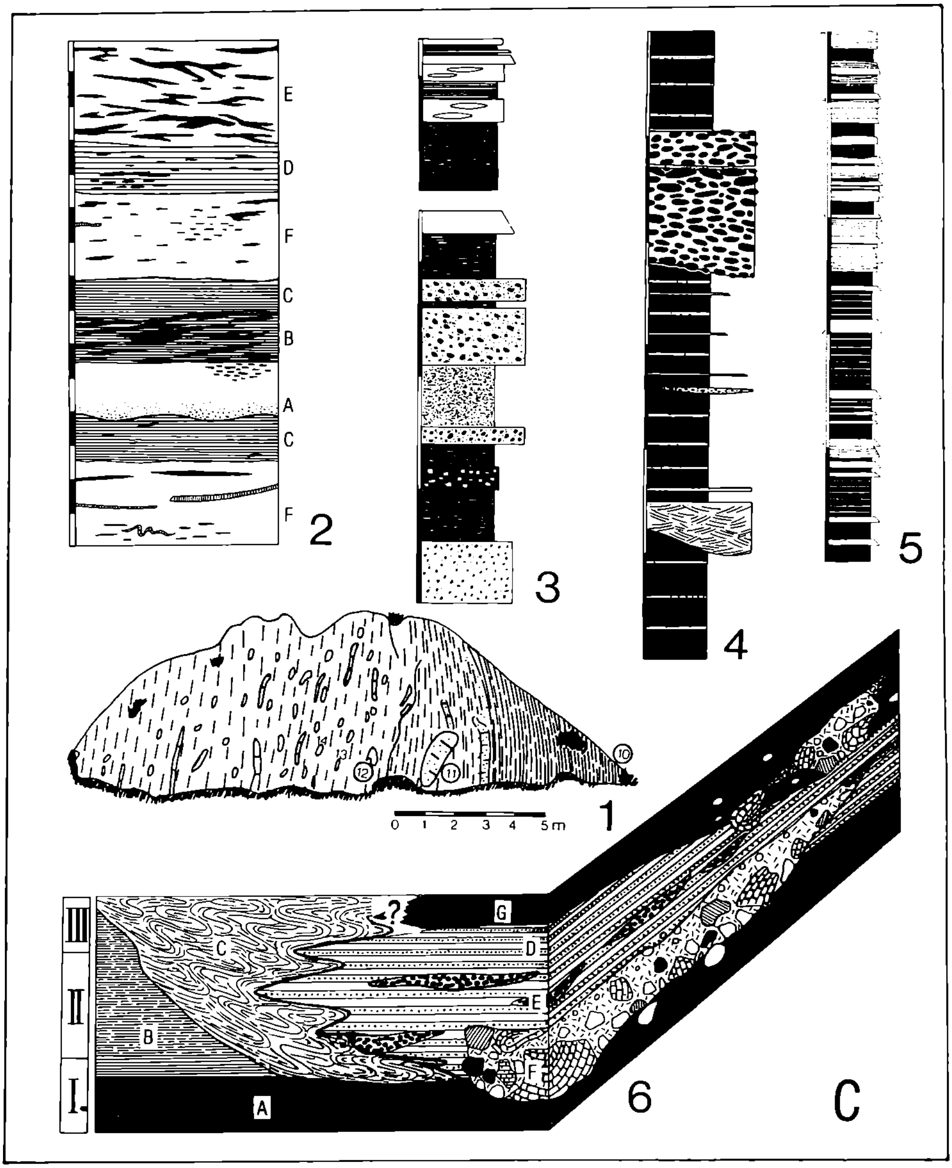
Text-Fig. 5. Siliceous organic carbon-rich sedimentation around the Cenomanian/Turonian boundary in the Rif.
1. Siliceous sediments: silicified pelites, radiolarites and silicified fine-grained turbidites - 2. Calcareous turbidites - 3. Black biosiliceous, organic carbon-rich sediments ("phthanites") - 4. Argillites (multicoloured claystones, characteristic background sedimentation of the basinal sequences) - 5. Hemipelagic mudstones (background sedimentation of the lower slope sequences) - 6. Pelagic limestones with chert and organic carbon-enriched intercalations - 7. Marlstones with slumping (characteristic background sedimentation of the upper slope sequences).



Upper Slope
Internal Loukkos (L1)
Lower Slope and Basin
Beni Derkoul
Tanger Intern (M8)
Tanger Intern (M90)
% TOC
% TOC
TURONIAN
CENOMANIAN
Internal Prerif
External Loukkos (L30)
% TOC
% TOC
2.9
3.6
0.6
1.8
2.4
1.6
0.7
0.6
4.9
-8.5
3.9
4.0
1
2
3
4
5
6
7
% TOC
total organic carbon content
10 m



Text-Fig. 6. Sedimentary sequences of the Upper Cretaceous along the Rif transect (Phases 3 and 4 of the Cretaceous subsidence history).
A. Synthetic section of the Mesorif (description in the text).
B. Synthetic section of the Loukkos Zone (description in the text).
C. Facies model and characterizing sedimentary sequences of the Tanger Unit (examples from the Tangier area). - C1: Passage of Santonian/Lower Campanian argillites (argillite formation 2, to the right) to Campanian/ Maastrichtian hemipelagic slope sediments (to the left). - C2: Typical sedimentary rhythm of Maastrichtian hemipelagic sedimentation of the Bab Taza Formation (Briquetterie of El Mrouhra Kebira, south of Tangier). (A) light grey marly limestone with sharp basal contact, sometimes showing micrograding, bioturbation is rare (mud turbidites?); (B) dark grey to black mudstone, intensely bioturbated; (C) dark grey to black mudstone, occasionally laminated, no or few bioturbation (oxygen-deficient environment); (D) grey mudstone, few bioturbation; (E, F) light grey marlstones, often intensively bioturbated, *Inoceramus* shells are common (oxic environment). Scale in



centimeters. - C3: Fine-grained debris-flow deposits of the Lower Maastrichtian (example from the Nappe du Habt west of Souk es Sebt). Scale bar is 50 cm. - C4: Channelized calciturbidite sequence (Lower Maastrichtian of the external Tanger Unit, village of Tangier). - C5: Calciturbidite sequence (Lower Maastrichtian of the external Tanger Unit, village of Tangier), levee sedimentation? - C6: Facies model for the external Tanger Unit south of Tangier. Stratigraphic scale: I. Santonian-Campanian; II. Lower-Middle Maastrichtian; III. Upper Maastrichtian-Paleocene. Sedimentary formations: A. Argillite Formation; B. Hemipelagic sequence of the Bab Taza Formation; C. Slumped hemipelagic sequence; D. Calciturbiditic sequence of the "faciès à microbrêche" with intercalated minor debris-flows and channels (E.); F. Olistostromes (chaotic facies with large sedimentary klippes); G. Argillite Formation III, locally with slumps and minor olistolites.

preserved planktonic microfaunas occur in the southern (external) part of the unit. Silicified calciturbidites, black shales, biosiliceous (radiolarian-bearing) layers and hemipelagic sediments characterize the more northerly (internal) facies type. Calcareous microfossils are partly dissolved, thus the depositional environment was probably below the Calcium Carbonate Lysocline. Upper Cretaceous sediments of the internal Mesorif Zone and Loukkos Zone (Text-Fig. 6, A-B) show massive mudstone facies and marl/limestone-cycles with important sediment gravity deposits such as pebbly mudstones, slumps, debris-flows, olistostromes and rare proximal turbidites. An overall-decrease in carbonate content is evident. Olive-green marly shales and shales are often intensely bioturbated and are typical for the pelagic sedimentation of these zones.

Sedimentation of the entire **Tanger Unit** is characterized by two argillite formations (argillite formation I: Albian/Cenomanian; argillite formation II: Turonian/Lower Campanian) which sandwich biosiliceous black shales of the Cenomanian/Turonian Boundary Event (CTBE, THUROW & KUHNT 1986) (Text-Fig. 5). Locally, Lower Cenomanian marly or limy intercalations with planktonic microfaunal assemblages can be observed. Calciturbidite sequences of Upper Cenomanian and Coniacian/Santonian ages occur in the internal Tanger Unit.

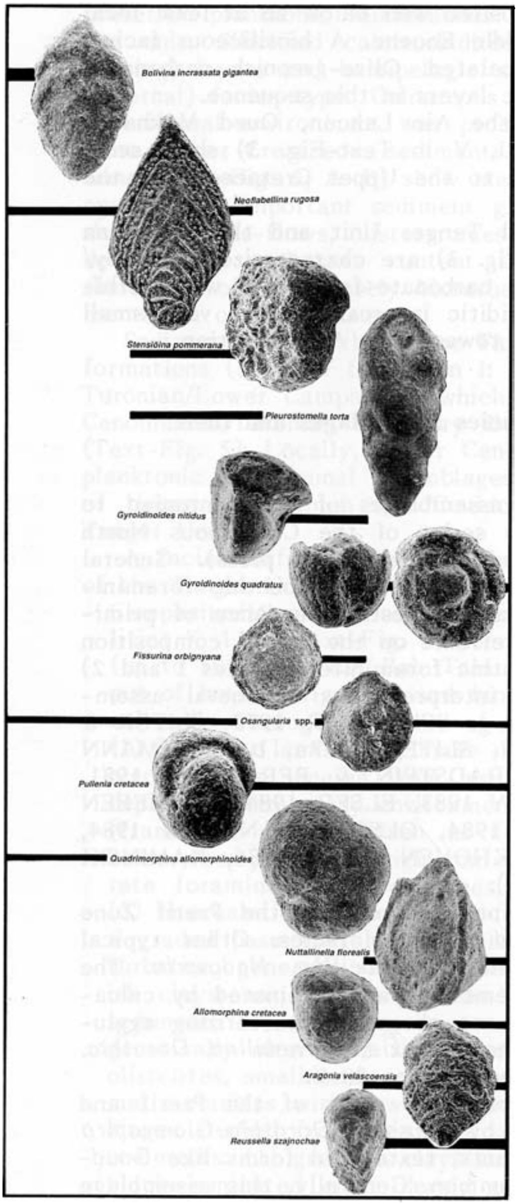
A facies differentiation can be observed in the Campanian/Maastrichtian of the Tanger Unit:

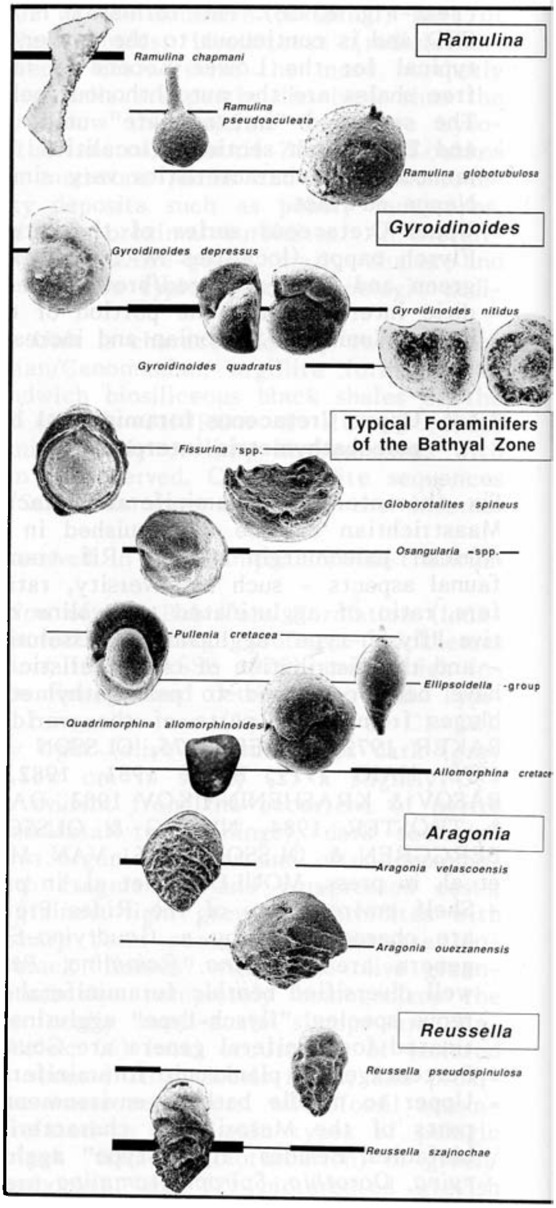
- Campanian/Maastrichtian sedimentation of the Bab Taza Formation (localities K-L in Text-Fig. 3) is characterized by high sediment thicknesses (up to 500 m in the Bab Taza region, LESPINASSE 1975) and a high input of both terrigenous and biogenic fine-grained detritic material. Typical sedimentologic features of this formation are (Text-Fig. 6, C1-C2): (1) irregular lamination, made up by light (olive greenish) and dark (grey black) layers ranging from mm to few cm in scale; (2) a slightly O_2-depleted sedimentary environment is evident from the occurrenc of pyrite (framboidal pyrite, often as foraminiferal test fillings), dark sediment colour, comparatively high contents in organic matter and often depauperate foraminiferal assemblages, with fragmented and compressed tests; (3) intercalation of extremely fine-grained, light grey calciturbidites with a graded base (silt-fraction) and sharp basal contacts; (4) intense bioturbation, generally recognizable in black "flames" within the olive greenish claystones; in some cases bioturbation completely homogenizes the sediment; (5) slumping is common, locally small-scale slumps are diagenetically altered to limestone nodules; (6) local occurrence of isolated olistolites, small-scale channel- (or scour-) fills, debris-flows, and proximal turbidites with reworked shallow-water sediments. Microfossil assemblages studied from this zone are related to two main types of pelagic sediments: (1) greenish clays with black "flames" (bioturbation) or green/black laminated clays devoid of carbonate, and (2) bioturbated greyish marls with variable amounts of carbonate and organic matter.
- A debris-flow sequence with abundant calciturbiditic redeposits of all grain sizes ("faciès à microbrèches") characterizes the Upper Campanian and Maastrichtian of the Nappe du Habt and the external Tanger Unit in the Tangier region (localities V, W and S in Text-Fig. 3; Text-Fig. 6, C3-C5). In Maastrichtian, the calciturbiditic formation grades into an argillaceous series through thinning and fining upward sequences. The lower part of the argillaceous formation partly forms the matrix of olistostromes and contains locally calciturbiditic layers and limestone-olistolites

(Text-Fig. 6, C6). This formation is deposited well below an at least local CCD and is continuous to the Lower/Middle Eocene. A biosiliceous facies, typical for the Lower Eocene is intercalated. Olive-greenish carbonate-free shales are the autochthonous pelagic layers in this sequence.
- The so-called "intermediate" units of the Ain Lahcen, Oued Mekhacen and Er Rkayah sections (localities T, U, V in Text-Fig. 3) show sedimentological characteristics very similar to the Upper Cretaceous of the Nappe du Habt.
- Upper Cretaceous series of the internal Tanger Unit and the Massylian Flysch nappe (localities M-Q in Text-Fig. 3) are characterized by grey/green and variegated red/brown-greenish carbonate-free clays with turbiditic intercalations. The portion of turbiditic intercalations is very small in the Coniacian/Santonian and increases towards the Maastrichtian.

3.1.2 Upper Cretaceous foraminiferal biofacies assemblages and their paleobathymetric interpretation

Six characteristic foraminiferal biofacies assemblages of the Santonian to Maastrichtian can be distinguished in the series of the Cretaceous North African paleomargin on the Rif transversal (KUHNT in press). General faunal aspects - such as diversity, ratio of planktonic to benthic foraminifers, ratio of agglutinated to hyaline calcareous tests, abundance of primitive "flysch-type" agglutinants, dissolution effects on the faunal composition - and the distribution of characteristic benthic foraminifers (Plates 1 and 2) have been compared to paleobathymetric interpretations of coeval assemblages from other parts of the world (e. g. SCHNITGER 1972, SLITER & BAKER 1972, SLITER 1975, OLSSON 1977, SLITER 1977a, b, BECKMANN 1978, HAIG 1979, BUTT 1981, 1982, GRADSTEIN & BERGGREN 1981, BASOV & KRASHENINNIKOV 1983, DAILEY 1983, ELSER 1984, HEMLEBEN & TRÖSTER 1984, NYONG & OLSSON 1984, OLSSON & NYONG 1984, BERGGREN & OLSSON 1986, VAN MORKHOVEN et al. 1986, KAMINSKI et al. in press, MOULLADE et al. in press):
- Shelf environments of the Rides Prérifaines and part of the Prerif Zone are characterized by a *Gaudryina-Frondicularia* biofacies. Other typical genera are *Gyroidina, Ramulina, Palmula, Neoflabellina, Nodosaria.* The well diversified benthic foraminiferal assemblages are dominated by calcareous species, "flysch-type" agglutinants are absent. Characterizing agglutinated foraminiferal genera are *Gaudryina, Tritaxia, Verneuilina, Dorothia.* Percentages of planktonic foraminifers reach up to 80 %.
- Upper to middle bathyal environments of lower parts of the Prerif and parts of the Mesorif are characterized by *Reussella-Dorothia-Glomospira* biofacies. Besides "flysch-type" agglutinants, textulariid forms like *Gaudryina, Dorothia, Spiroplectammina* are common. Generally, this assemblage shows optimal benthos-diversity. Plankton percentages exceed generally 90 %; few samples show a decrease in plankton percentages to about 50 %. The latter faunas contain also compressed and fragmented tests of planktonic foraminifers.
- Parts of the Mesorif and Loukkos zones, which are interpreted as representing a more or less middle bathyal bathymetric zone are characterized by a *Spiroplectammina-Glomospira* biofacies. "Flysch-type" forms (e. g. *Glomospira, Ammodiscus, Recurvoides, Hormosina, Paratrochamminoides, Rhabdammina*) dominate the agglutinants. The plankton percentages show



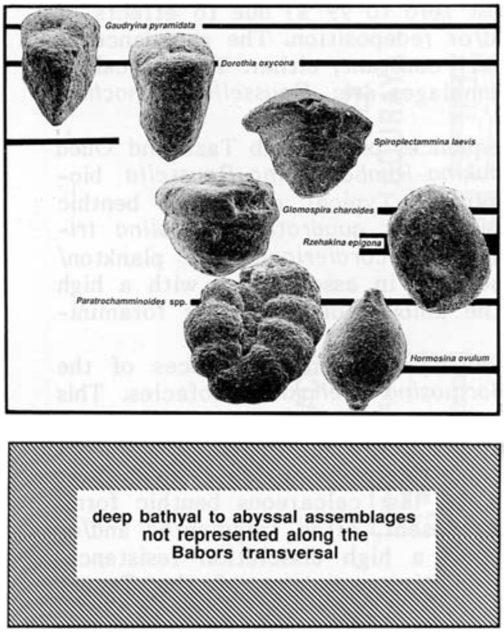


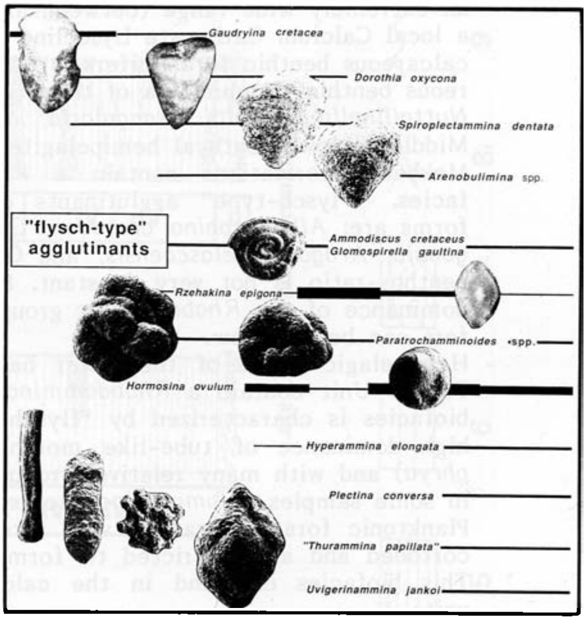
BABOR UNIT	ERRAGUENE UNIT	BREK UNIT proximal	BREK UNIT distal
outer neritic - upper bathyal	upper bathyal	upper to middle bathyal	middle bathyal

Stage 4 of the Cretaceous Subsidence History (MAASTRICHTIAN)

neritic	bathyal			abyssal
	upper	middle	lower	

Stages 3 and 4 of the Cretaceous Subsidence History (SANTONIAN - MAASTRICHTIAN)





BABOR UNIT	ERRAGUENE UNIT	BREK UNIT proximal	BREK UNIT distal
outer neritic - upper bathyal	upper bathyal	upper to middle bathyal	middle bathyal
Stage 4 of the Cretaceous Subsidence History (MAASTRICHTIAN)			

neritic	bathyal			abyssal
	upper	middle	lower	
Stages 3 and 4 of the Cretaceous Subsidence History (SANTONIAN - MAASTRICHTIAN)				

Plate 2

Distribution of benthic agglutinated foraminifers in the Campanian/Maastrichtian of the Babors (left column) and Rif (right column) transects.

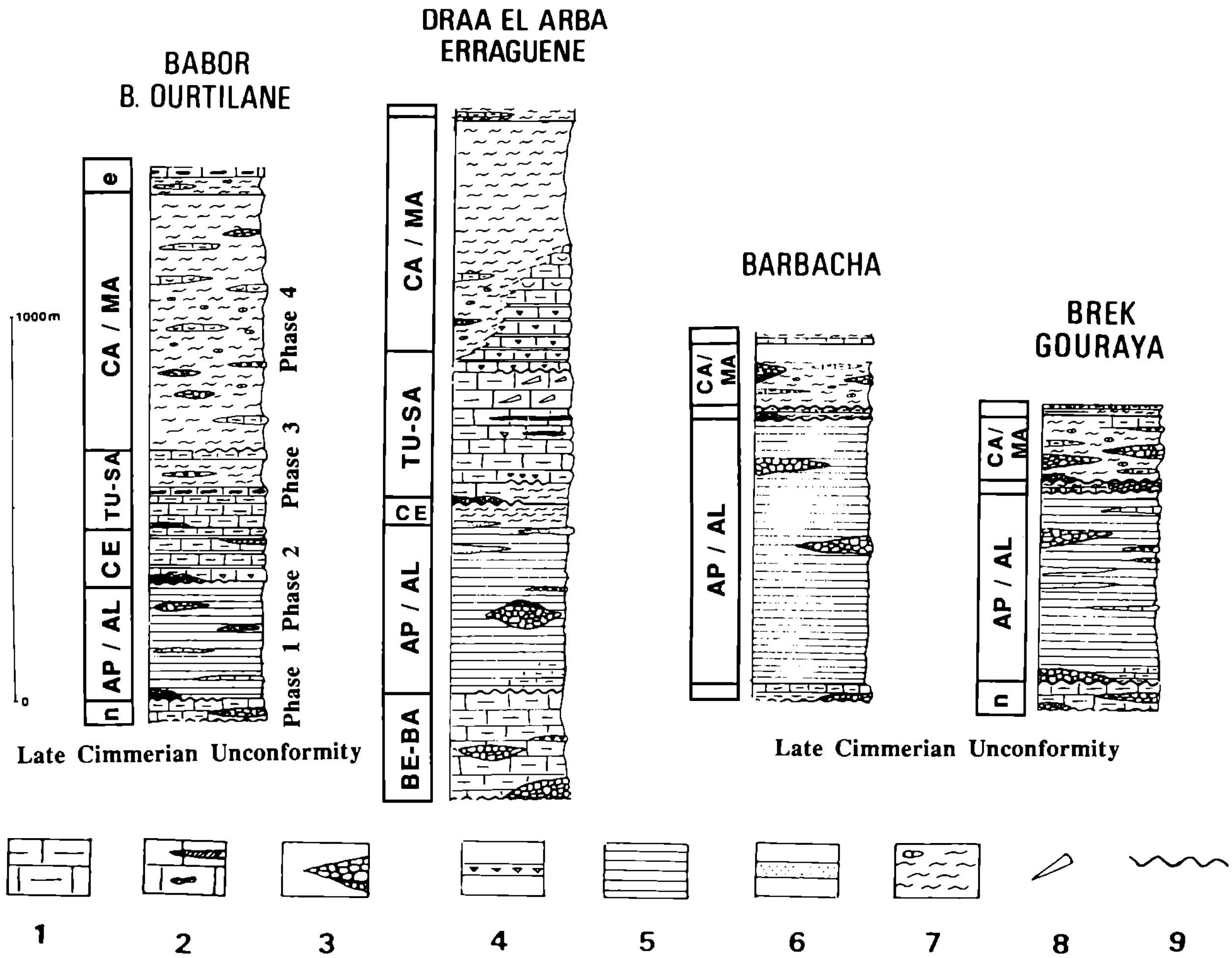
← **Plate 1**

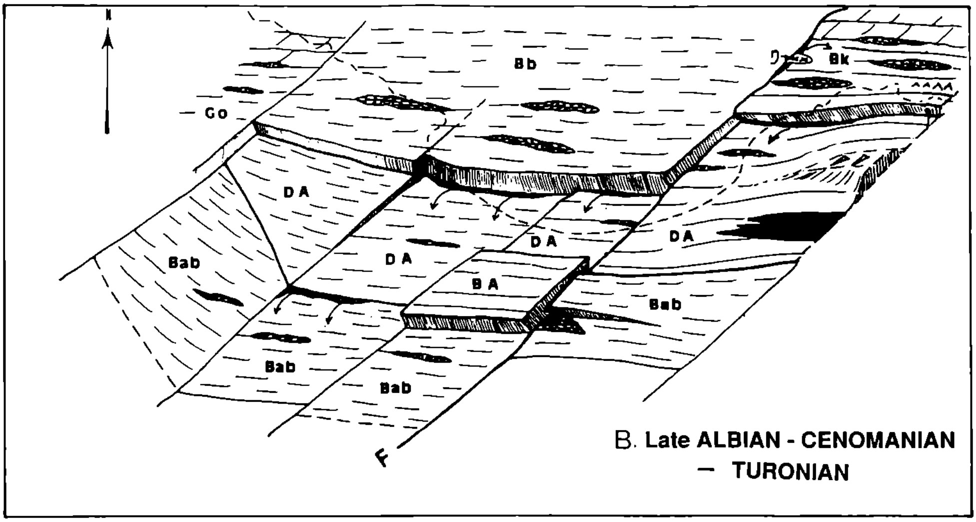
Distribution of calcareous benthic foraminifers in the Campanian/Maastrichtian of the Babors (left column) and Rif (right column) transects.

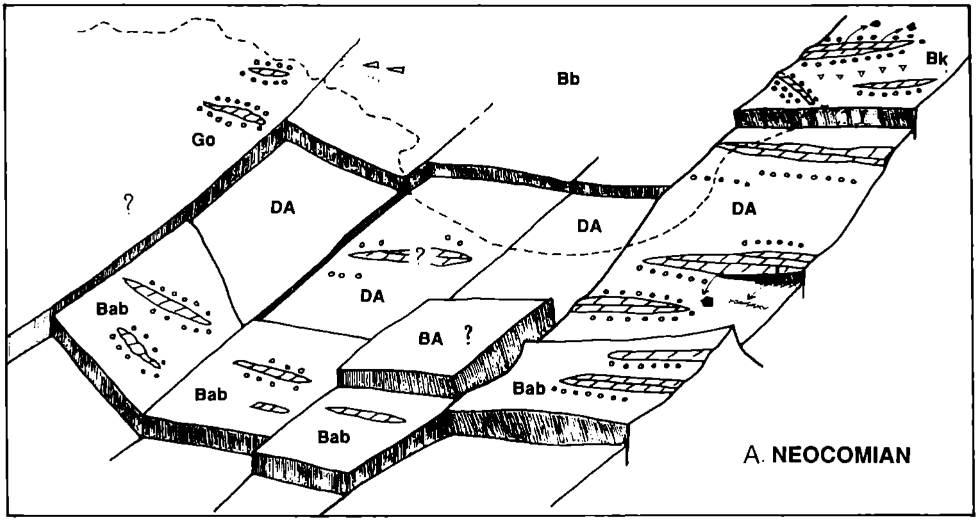
an extremely wide range (between almost zero to 99 %) due to effects of a local Calcium Carbonate Lysocline and/or redeposition. The abundance of calcareous benthic foraminifers is not very constant, either. Typical calcareous benthic foraminifera of these assemblages are: *Reussella szajnochae, Nuttallinella florealis, Osangularia.*

- Middle to lower bathyal hemipelagite sequences of the Bab Taza and Oued Mekhacen formations contain a *Rzehakina-Rhabdammina-Reussella* biofacies. "Flysch-type" agglutinants dominate. Typical calcareous benthic forms are: *Allomorphina cretacea, Gyroidinoides quadratus, Guttulina trigonula, Aragonia velascoensis,* and *Osangularia cordieriana.* The plankton/benthos-ratio is not very constant. Especially in assemblages with a high dominance of the *Rhabdammina* group the amount of planktonic foraminifers can be very low.
- Hemipelagic layers of the lower bathyal calciturbidite sequences of the Tanger Unit contain a *Rhabdammina-Hormosina-Rzehakina* biofacies. This biofacies is characterized by "flysch-type" agglutinated assemblages with high dominance of tube-like morphotypes (e. g. *Rhabdammina, Dendrophrya*) and with many relatively robust and relatively coarse-grained taxa. In some samples *Allomorphina* occurs as the last calcareous benthic form. Planktonic foraminifera are very rare or absent, often fragmented and/or corroded and are restricted to forms with a high dissolution resistance. This biofacies is found in the calciturbiditic sequences of the Tanger units.
- Abyssal environments of the distal parts of the Tanger and Almarchal units, and the Massylian Flysch units of the Rif and Campo de Gibraltar are characterized by a *Recurvoides-Rhabdammina-Hormosina* biofacies. "Flysch-type" agglutinated assemblages dominate without any autochthonous calcareous foraminifera (deposition below CCD). An increasing diversity of agglutinated forms can be observed in comparison to the *Rhabdammina-Hormosina-Rzehakina* biofacies of the more proximal turbiditic sequences. Other typical benthic foraminifers are: *Paratrochamminoides, Hyperammina, Recurvoides, Bathysiphon, Ammodiscus, Glomospira, Karreriella,* and *Uvigerinammina jankoi.* These biofacies assemblages are almost identical with coeval deep-sea assemblages, well-known from the Carpathian flysch trenches (MORGIEL & OLSZEWSKA 1982).

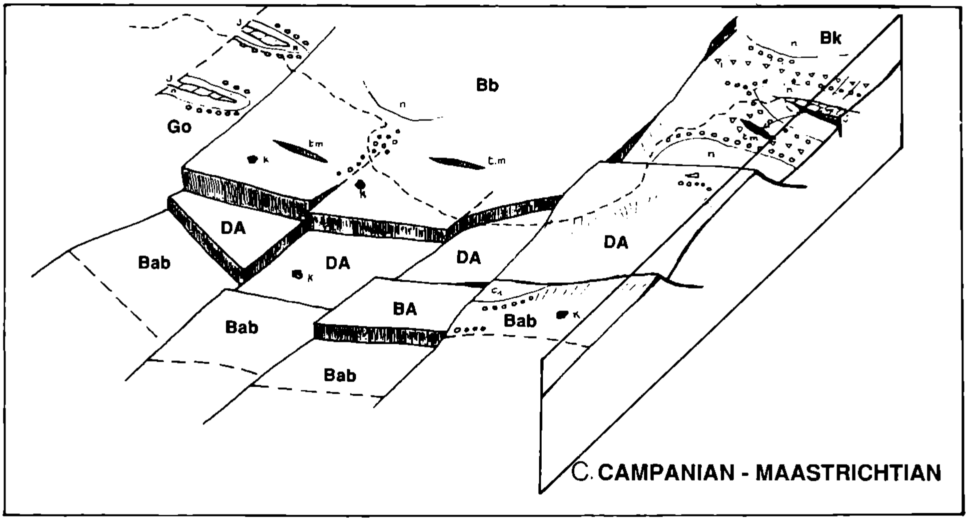
Text-Fig. 7. Sedimentary sequences of the Cretaceous along the Babors transect. Lithostratigraphic columns are set in the order of tectonic superposition of the units. Southern (lower) units to the left. - 1.-2. Top of Upper Jurassic to early Cretaceous carbonate sequence (1. marly limestone; 2. limestone with chert) - 3. Conglomerate (reworked shallow-water carbonates, mainly of Jurassic origin) - 4. Microbreccia (proximal turbidite or small-scale debris-flow) - 5. Shale and argillite - 6. Sandstone - 7. Marl with yellow nodules - equivalent of the hemipelagite sequence of the Rif transect - 8. Rudists (Rotalitidea) - 9. Unconformity.

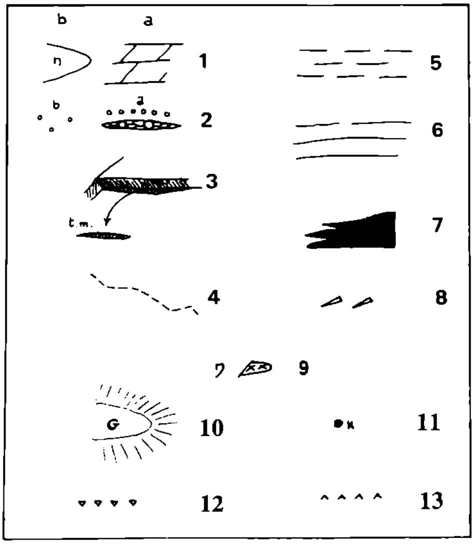

BABOR
B. OURTILANE
DRAA EL ARBA
ERRAGUENE
BARBACHA
BREK
GOURAYA
1000m
0
e
CA / MA
TU-SA
CE
AP / AL
n
Phase 4
Phase 3
Phase 2
Phase 1
BE-BA
CA/
MA
Late Cimmerian Unconformity
Late Cimmerian Unconformity
1
2
3
4
5
6
7
8
9





Text-Fig. 8. Paleogeographic sketches of the Babors during main phases in the evolution of the North African Continental Margin.
A. Lower Cretaceous subsidence phase (Neocomian) - B. Mid-Cretaceous compressional phase (i. e. Cenomanian) - C. Late Cretaceous compressional phase (Campanian/Maastrichtian).
The structural pattern of different blocks, which developed in the Lias (OBERT 1986), influences the whole Mesozoic evolution. The important tectonic phase of the late Lower Cretaceous is followed by an uplift of the whole region and erosion of parts of the sedimentary cover. The Ceno-





1. Unconformities
   a. on Jurassic limestones
   b. on Neocomian marls and shales
2. Redeposited shallow water limestones
   a. conglomerates (debris-flows)
   b.pebbles (olistolites)
3. Main fault, boundary of paleogeographic units
   F : Babors fault
4. Actual coast-line
5. Predominantly marly sedimentation
6. Predominantly marl-limestone sedimentation
7. Siliceous biogenic sediments (e.g. radiolarites)
8. Platform with rudists
9. Diorite of Bou Zazen
10. Swell with stratigraphic gap (emersion)
11. Sedimentary klippes
12. Turbidites
13. Gypsum , reworked from Triassic salinar

manian/Turonian transgressive period, accompanied by a new phase of subsidence gives rise to a slightly more calcareous sedimentation with an important facies differentiation between the single units, due to an important relief: Biosiliceous radiolaritic sediments occur in the eastern part of the Erraguene Unit, whereas the neighbouring highs are colonized by rudists. Triassic material is mobilized along important faults and redeposited in the nearby sedimentary basins. Reconstruction for the western part of the Babors under consideration of data from LEIKINE (1971).

3.2 Babors Transversal

3.2.1 Sedimentology and facies distribution (Text-Figs. 7, 8)

Cretaceous sedimentation of the southern **Babor/Beni Ourtilane Unit** begins unconformably after about 60 m of Upper Jurassic reddish pelagic marls and marly limestones with siliceous intercalations. A condensed marly limestone sedimentation follows. Locally, conglomerates occur mainly consisting of reworked Jurassic limestones. Hiatuses and sedimentary discordances are common. This Berriasian to Hauterivian sequence of less than 60 m thickness is followed by dark grey marly limestones bearing pyritic ammonites, and dark grey to black marls and shales of about 300 m of maximum thickness. This sequence was probably deposited under oxygen-deficient conditions. Conglomeratic channel-fills are locally important (Barremian-Middle/Upper Albian). Upper Albian and Cenomanian sedimentation is made up by about 130 m of marls and marly limestones. During Upper Cretaceous marls and limestones with inoceramids and marls with yellow concretionary nodules dominate the sedimentation. Sediment thicknesses are about 200 m for the Turonian-Santonian and increase to 650 m for the Campanian-Maastrichtian. Foraminiferal assemblages within this thick sediment pile generally indicate middle to outer neritic environments and locally oxygen-deficient conditions.

Cretaceous sedimentation within the **Draa el Arba and Erraguene units** begins also above an unconformity surface on the top of less than 100 m of Kimmeridgian-Tithonian pelagic sediments. Berriasian to Hauterivian sedimentation consists of about 270 m of marly limestones with pyritic ammonites. Barremian to Lower Albian is made up by 490 m of shaly sedimentation with detrital intercalations similar to those of the Babor Unit. The Upper Albian to Cenomanian of the Erraguene Unit is quite particular. The more southerly part of the Erraguene Unit is characterized by marls with pelagic microfaunas (mainly planktonic foraminiferal assemblages of Upper Albian to Lower Cenomanian ages) followed by biosiliceous limestones and partly radiolaritic sedimentation, which can be compared to the "phthanites" of the Cenomanian/Turonian boundary beds in the Rif. In the northern part of the unit neritic conditions persist with shallow water limestones, which, in its upper part, contain locally well preserved Turonian rudistid faunas. Approximate sediment thicknesses are 410 m for the Upper Albian-Cenomanian and about 200 m for the Turonian to Santonian. Campanian/Maastrichtian sedimentation of the Erraguene Unit is marly, about 400 m thick, and contains foraminiferal assemblages which indicate outer neritic and/or upper bathyal water depths.

Upper Jurassic and Neocomian sedimentation of the **Brek Unit** is similar to that of the Babor and Erraguene units, but conglomeratic intercalations with redeposited platform carbonates (biodetritic limestones and oolitic limestones) of Jurassic age are abundant. Thickness of the sediments is less than 100 m for the Kimmeridgian-Tithonian and less than 50 m for the Berriasian-Hauterivian. Barremian to Albian sediments are made up by about 500 m of dark grey argillites with abundant detritic intercalations (sedimentary klippes, olistolites, thick conglomerates, probably channel fills, debris-flows of different grain sizes, mainly microbreccia, and proximal calciturbidites), which mainly comprise reworked Jurassic limestones. In contrast to the southerly units detrital input of quartz-sandstones is important in the Barremian to Albian of the Brek Unit. Cenomanian/Turonian sediments in the Brek Unit occur only locally and consist of limestone breccia, which yield reworked Cenomanian neritic limestones (Plate 4, Fig. F). Maximum thickness of this intercalation is about 20-30 m. The Coniacian/Santonian-Maastrichtian consist mainly of hemipelagites with intercalated calciturbidites, and conglomeratic debris-flows, which can be interpreted as channel-fills. These resediments often contain important amounts of redeposited Triassic material. The autochthonous hemipelagic sedimentation is made up by dark-grey to olive-greenish mudstones, often slumped and containing large yellow concretionary nodules. The observed thickness of this sequence is very low (about 150 m, with a probable decrease of sedimentation rates towards the top). Foraminiferal assemblages are characterized by a great abundance of redeposited tests. Assemblages which can be regarded as autochthonous indicate bathyal environments.

3.2.2 Foraminiferal biofacies assemblages and their paleobathymetric interpretation

3.2.2.1 Lower Cretaceous

General distribution patterns of Lower Cretaceous foraminifers in the Babors (Tables 1 and 2) can be interpreted in terms of paleobathymetry. A clear difference in faunal composition can be detected between neritic facies of the Babor and Brek units containing larger foraminifers, which are typical for the photic zone, and the more pelagic assemblages of the Erraguene Unit. Lower Cretaceous planktonic and benthic foraminifers have been utilized by GUERIN (1981) to define paleoenvironments and paleobathymetric zones within pelagic environments of the Tethys and the Atlantic Ocean. In comparison with the paleobathymetric distribution for some important benthic foraminifers of the Lower Cretaceous (GUERIN 1981, MOULLADE

Table 1. Distribution of foraminifers during stage 1 of Cretaceous subsidence history in the Babors (Berriasian-Barremian). Note concentration of deep-water species in the Erraguene Unit.

	Babor	Erraguene	Brek
Lenticulina gibba	+		
Lenticulina nodosa		+	
Lenticulina ouachensis		+	
Gavelinella sp.		+	
Dorothia hauteriviana		+	
Epistomina sp.		+	
Hedbergella sp.		+	
Trochammina sp.	+	+	
Haplophragmoides sp.	+		+
Palaeodictyoconus gr. *arabicus*	+		
Pseudocyclammina sp.	++	resedimented (?)	
Orbitolina sp.	+		
Trocholina cf. *alpina*			+

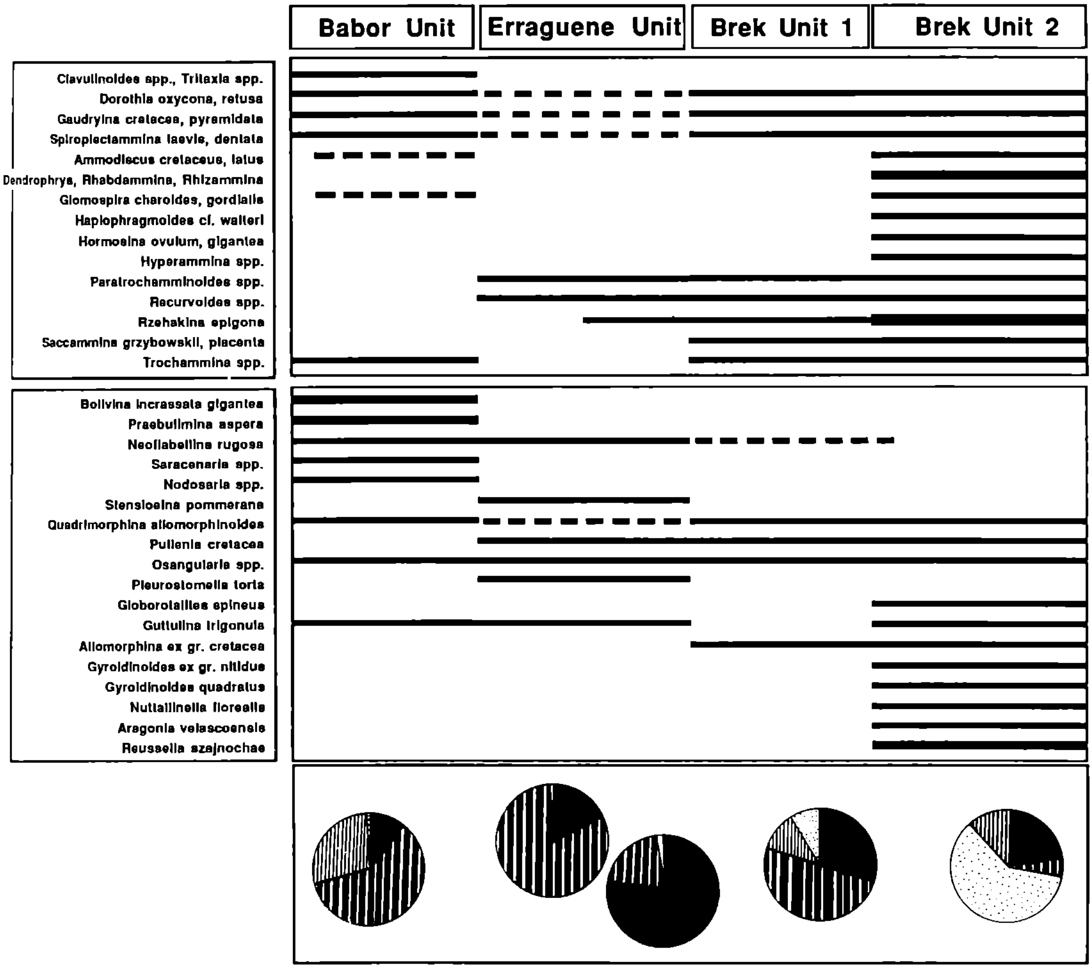
Table 2. Distribution of foraminifers in the late Lower Cretaceous (Aptian-Albian/Vraconian), compressional stage 2 of the margin evolution in the Babors. Note highly diversified assemblages with deep-water foraminifers in the Erraguene Unit, whereas Babor and Brek units are characterized by neritic species.

	Babor	Erraguene	Brek
Ammodiscus gaultinus		✧	
Trochammina gr.*vocontiana*		✧✧	
Trochammina sp.		✧✧	
Haplophragmoides sp.		✧✧	
Verneuilinoides sp.		✧	
Gaudryinella sp.		✧	
Eoguttulina sp.		✧	
Lenticulina oligostegia		✧✧	
Lenticulina subangulata	✧	✧	
Lenticulina gibba	✧		
Lenticulina sp.		✧	✧
Nodosaria sp.		✧	
Vaginulina sp.		✧	
Gavelinella berthelini		✧	
Gavelinella sp.		✧✧	
Gyroidinoides globosus		✧✧	
Hedbergella delrioensis		✧	
Hedbergella roberti	✧	✧	
Hedbergella infracretacea		✧✧	
Hedbergella sp.		✧	
Globigerinelloides sp.		✧	
Ticinella multiloculata	✧	✧✧	
Praeglobotruncana stephani		✧	
Rotalipora subticinensis		✧	
Pseudoglandulina humilis			✧
Pseudoglandulina sp.	✧		
Trocholina cf. *alpina*			✧
Palaeodictyoconus gr. *arabicus*	✧		
Mesorbitolina sp.			✧
Orbitolina sp.	✧	resedimented (?)	✧✧
Paracoskinolina sp.	✧		
Pseudocyclammina sp.	✧✧	resedimented (?)	

1984), upper bathyal environments can be assumed for the assemblages of the Barremian-Albian of the Erraguene Unit.

3.2.2.2 Upper Cretaceous (mainly Campanian/Maastrichtian)

Four biofacies assemblages have been observed in the Campanian/Maastrichtian sediments of the Babors (Text-Fig. 9, Plates 1 and 2):



Text-Fig. 9. Distribution of some characteristic benthic foraminifers on the Upper Cretaceous North African paleomargin in the Babors and their paleobathymetric implications.
Compared with the bathymetric model of the Rif transversal (KUHNT 1987) the three main units of the Babors correspond to the following zones: **Babor Unit:** Neritic assemblages of shelf basins (Prerif 1) - **Erraguene Unit:** Upper bathyal assemblages of slope and intramarginal basins (Prerif 2 - Mesorif 2) - **Brek Unit:** Upper/middle bathyal assemblages of slope, partly affected by a local Calcium Carbonate Lysocline (Mesorif 2 - Mesorif 3).
The deeper (lower bathyal to abyssal) environments of the Rif transversal are **not** represented in the Babors. The studied transect is indicated in the paleogeographic sketch for the Campanian/Maastrichtian in Text-Fig. 8 C.

Composition of microfossil assemblages

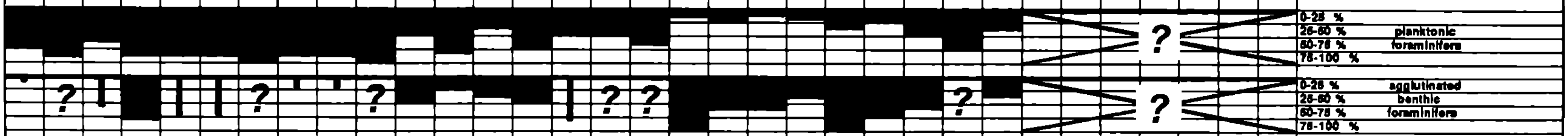
keeled planktonic foraminifers
non-keeled planktonic foraminifers
agglutinated benthic foraminifers
calcareous benthic foraminifers
radiolarians

- Marlstones of the Babor Unit are characterized by a *Bolivina incrassata-Neoflabellina-Nodosaria* biofacies. Plankton/benthos ratio is 2 : 1 or less; the use of the ratio seems to be of some value to distinguishing different environments within the group. These typical shelf assemblages are comparable to the biofacies assemblages of the Prerif Zone in the Rif (biofacies 1).
- Marlstones of the Erraguene Unit contain a *Neoflabellina* gr. *rugosa-Praebulimina-Pleurostomella* biofacies. Plankton/benthos ratio is in general very high (more than 9 : 1); plankton values of 98-99 % are common. The faunal composition is quite similar to assemblages of the Brek Unit, but is lacking typical bathyal species such as *Recurvoides, Hormosina ovulum, Rzehakina epigona, Aragonia velascoensis, Osangularia velascoensis, Nuttallinella florealis,* and *Gyroidinoides quadratus*. An outer neritic to uppermost bathyal environment can be assumed for this biofacies. Corresponding environments in the Rif have been found in the deeper part of the Prerif and in the external Mesorif Zone (biofacies 2).
- Grey and grey-green marlstones of the Brek Unit are characterized by a *Rzehakina epigona-Neoflabellina-Gavelinella* biofacies showing a high plankton/benthos ratio, and an often high benthos diversity. Other typical benthic species are *Gyroidinoides globosus/nitidus, Osangularia, Pullenia*. These assemblages of the Brek Unit represent an upper bathyal environment - characteristic samples are 91-1 (U.CA/L.MA), 215-2 (U.CA/L.MA) (Text-Fig. 10) - and are comparable to the biofacies 2-3 of the Rif (mainly corresponding to the Mesorif Zone).
- The deepest biofacies assemblages observed in the Tellian Units of the Babors occur in the hemipelagic facies which consist of grey-green hemipelagic marls and shales with yellow concretions into which proximal calciturbidites of the Brek Unit are intercalated. These assemblages are represented by a *Rzehakina epigona-Glomospira* biofacies. A characteristic feature is the dominance of agglutinated foraminifers. *Rzehakina epigona* is common to abundant in many of the samples, and the plankton/benthos ratio is low (important influence of carbonate dissolution). Further typical benthic species are agglutinated foraminifers like *Dorothia, Recurvoides, Ammodiscus, Hormosina ovulum*, tube-shaped forms of the *Bathysiphon/Rhabdammina/Dendrophrya* complex and the calcareous benthic forms *Osangularia, Nuttallinella, Guttulina trigonula, Globorotalites spineus, Quadrimorphina allomorphinoides* and *Gyroidinoides quadratus*. Characteristic samples are 177-1 (CA/MA) and 523-1 (CA) (Text-Fig. 10). The general faunal aspect is comparable to the biofacies 3-4 in the Rif, where they have been interpreted as middle bathyal.

Text-Fig. 10. Campanian/Maastrichtian foraminiferal assemblages of the Brek Unit (Babors). Assemblage types: 1 Planktonic foraminifers dominate, benthic forms characterize Upper Slope environments - 2 Planktonic percentages decrease, benthic forms characterize Mid-Slope environments; agglutinants often dominate - 3 Impoverished assemblages of probably oxygen-depleted environments - 4 Redeposited assemblages. - Symbols: · rare-few - C common - x redeposited.

ASSEMBLAGE TYPE	1	1	1	1	1	1	1	1	1	1	1-2	1-2	1-2	1-2	1-2	1-2	1-2
SAMPLE	671-6	215-2	130-1	92-2	92-1	91-1	90-6	90-3	89-8	83-1	6/1	86-4	866-1	866-2	138-1	86-2	31-3
AGE	U.C	L.M.		L.M	L.M	L.M	C	U.C	C	C	C/M	M	C/M	C	C	C	C/M
grey marls with yellow nodules											X						
resediments (turbidites, debris-flows)		X														X	
gray, or gray-green marls	X			X	X	X			X						X		X
gray, or yellow-green shales			X					X					X				
dark-gray or black shales and marls																	
gray and yellow limy clay							X			X				X			

ASSEMBLAGE TYPE	2	2	2	2	2	2	2-3	3	3	4	4	4	4	4	4	1-2
SAMPLE	3/9	823-1	177-1	186-4	136-8	80-2	82-1	78-6	32-1	4/3	889-3	252-1	184-2	177-2	144-4	89-6
AGE	L.C	C	C/M	M		U.C	C	C	U.C	C/M	U.C	U.M	U.C	C	C	L.M
grey marls with yellow nodules				X	X		X									
resediments (turbidites, debris-flows)										X	X	X	X	X	X	
gray, or gray-green marls		X	X													
gray, or yellow-green shales	X					X										X
dark-gray or black shales and marls								X	X							
gray and yellow limy clay																



SAMPLE	671-6	215-2	130-1	92-2	92-1	91-1	90-6	90-3	89-8	83-1	6/1	86-4	866-1	866-2	138-1	86-2	31-3
Ammodiscus cretaceus, planus				•	•						•		•	•			
Matanzia varians												•					
Dendrophrya, Rhabdammina, Rhizammina			•	C	•	•	•				•	•	•	•	•		
Dorothia oxycona, retusa											•	C		•			
Gaudryina cretacea, pyramidata											•	•				•	
Glomospira charoides, gordialis				•							•	•			•		
Glomospirella gaultina															•		
Haplophragmoides cf. walteri													•			•	
Hormosina ovulum, gigantea											•						
Hyperammina spp.											•		•		•		
Paratrochamminoides				•													
Recurvoides spp.											•		•				
Reophax spp.				•							•						
Rzehakina epigona		•			•	•		C				C	•	C	•	•	
Saccammina grzybowskii, placenta				•													
Spiroplectammina laevis, dentata									•		•	•					
Trochammina spp.																	
Allomorphina ex gr. cretacea					•				•								
Aragonia velascoensis											•						
Bolivinoides decoratus, draco																	
Dentalina spp., Nodosaria spp.	•			•	x			x	x		•x	•	x	•	x		
Fissurina orbignyana								•									
Gavelinella spp.	•				•			•			•	•	•				•
Globorotalites spineus																	
Guttulina trigonula			•	•	•	•			•		•	C	•	C	•		
Gyroidinoides ex gr. nitidus				•							C	•	•				
Gyroidinoides quadratus																	
Lenticulina spp., Marginulina spp.	•							•	x		x	x	x		x		
Neoflabellina rugosa											•	•	•				
Nuttallinella florealis		•									•						
Osangularia spp.				•	•						•	C	•				
Pleurostomella spp.											•	•					
Praebulimina spp.								•			C	•				•	•
Pullenia cretacea											•	•			•	•	
Quadrimorphina allomorphinoides											•						
Reussella szajnochae	•										•						
Ramulina spp.								•			•		•	•			
Ostracodes	•		•	•	•	•					•						

SAMPLE	3/9	823-1	177-1	186-4	136-8	80-2	82-1	78-6	32-1	4/3	889-3	252-1	184-2	177-2	144-4	89-6
Ammodiscus cretaceus, planus	•	•		•								x				
Matanzia varians				•												
Dendrophrya, Rhabdammina, Rhizammina	C	C	•		•			•								
Dorothia oxycona, retusa			•	•	•				•		x		x	x		
Gaudryina cretacea, pyramidata			•	•				•			x		x	x	x	•
Glomospira charoides, gordialis		C	•	•	•											
Glomospirella gaultina		•			•											
Haplophragmoides cf. walteri	•															
Hormosina ovulum, gigantea		C			•											
Hyperammina spp.					•											
Paratrochamminoides					•											
Recurvoides spp.	•	C	•		•	•										
Reophax spp.																
Rzehakina epigona	•	C	•		C	•										
Saccammina grzybowskii, placenta	•			•		•										
Spiroplectammina laevis, dentata								•								
Trochammina spp.				•			C	•	•	•						
Allomorphina ex gr. cretacea						•	•			•	x					C
Aragonia velascoensis																
Bolivinoides decoratus, draco				•								x				
Dentalina spp., Nodosaria spp.		x	x	x							x		x			
Fissurina orbignyana																
Gavelinella spp.								x	•	•				x	x	
Globorotalites spineus			C													
Guttulina trigonula		C		•			•	x	•	•	x			x	x	•
Gyroidinoides ex gr. nitidus				•						•				x		C
Gyroidinoides quadratus				•												
Lenticulina spp., Marginulina spp.		x		x			•				x	x	x	x	x	•
Neoflabellina rugosa			•										x		x	
Nuttallinella florealis			C	•												
Osangularia spp.		C	•	C												
Pleurostomella spp.													x			
Praebulimina spp.				C			•				x			x		C
Pullenia cretacea				•												
Quadrimorphina allomorphinoides		•														
Reussella szajnochae													x			
Ramulina spp.											x	x		x	.	
Ostracodes		x					•			•	x	x				C

3.3 Definition of Cretaceous type-formations which can be traced along the North African Mesozoic continental margin from the Gibraltar Arch area to eastern Algeria

Sedimentary sequences of similar lithology, age, faunal assemblages and petrographic composition have been observed on both transversals, thus permitting the definition of several Cretaceous sedimentary formations. These are in ascending order:

1. Cephalopod marls and marl/limestone cycles with spikes of terrigenous sedimentation (Valanginian-Barremian).
2. Shales with terrigenous turbiditic intercalations ("flysch albo-aptien"; Lower Aptian-Middle Albian).
3. Black laminated limestones and marl/limestone cycles (Upper Albian).
4. Neritic biogenic limestones (Upper Albian-Lower Cenomanian).
5. Pelagic marls and shales with low detrital input (uppermost Albian-Cenomanian): Argillite Formation I of the distal sequences.
6. Siliceous black shales around the Cenomanian/Turonian boundary (CTBE).
7. Pelagic limestones, marls, and shales with low detrital input (Middle Turonian-Santonian): Argillite Formation II of the distal sequences.
8. Hemipelagic sequence (often slumped marlstones and claystones, "faciès à boules jaunes"; Campanian-Maastrichtian).
9. Olistostrome/debris-flow sequence with abundant calciturbiditic resediments of all grain sizes ("faciès à microbrêche", Campanian-Maastrichtian).
10. Shales with rare turbidites, slumps and olistolites (Maastrichtian-Eocene): Argillite Formation III of the distal sequences.

3.3.1 Cephalopod marls and marl/limestone cycles with spikes of terrigenous sedimentation (Valanginian-Barremian; Plate 3, Fig. A)

This facies type can be observed in all external units of the North African orogenic belt where Lower Cretaceous sediments are preserved. In the Rif, the more southerly Prerif sedimentation is dominated by marls and marl-limestone cycles with abundant pyritic ammonites, whereas the more northerly Mesorif Zone is characterized by greater thickness of the sediment pile and a larger amount of fine-grained terrigenous input. In the Babors all units show a marly sedimentation with a decrease in pyritic ammonite abundance from southern to northern units. A characteristic first spike of terrigenous resediments (probably a reworked Wealden facies?) occurs in the Upper Valanginian in all units of the Rif transversal; similar intercalations are found throughout the Hauterivian. This first phase of terrigenous input terminates with the Barremian limestone-dominated facies.

The paleotectonic setting of this sedimentation has been a continuously subsiding North African passive continental margin (stage 1 of the Cretaceous subsidence history). Local occurrence of conglomerates and olistolites (mainly resedimented Jurassic carbonates, Plate 3, Fig. E) can be correlated with erosional features at margins of subsiding tilted blocks.

3.3.2 Shales with terrigenous turbiditic intercalations ("flysch albo-aptien"; Lower Aptian-Middle Albian; Plate 3, Figs. B-D)

The onset of this formation is marked by an abrupt decrease in carbonate content and a remarkable increase in terrigenous detrital input, which leads to the formation of several fan systems on the North African continental margin. The most important of which is reported from the region of the Ouarsenis in eastern Algeria (MATTAUER 1958, WILDI 1983). Sedimentation in the western Rif and in the Babors is mainly dominated by thick pelitic sequences and the siliciclastic influence is of minor importance. Nevertheless, some of the facies associations of MUTTI & RICCI LUCCHI (1972) can be distinguished in the Mesorif, Loukkos and Ketama units of the Rif (GÜBELI 1982, KUHNT 1987) which enable to reconstruct a facies model from channelized facies of the upper slope (southern Mesorif Zone) down to outer fan deposits (northern Ketama Unit, GÜBELI 1982). In the Babors, several conglomeratic debris-flows occur within this formation, mainly reworking the underlying Jurassic limestones. These mass-flows may be early signs of a compressional phase, affecting this segment of the North African Margin.

3.3.3 Black laminated limestones and marl/limestone cycles (Upper Albian; Plate 4, Figs. A, B)

A transgressive sequence of the Upper Albian is represented in the shallow shelf areas of the Rif transversal (Rides Prérifaines), where bivalve and oyster facies onlap Jurassic limestones. On the outer shelf and slope pelagic limestones and marls show a rapid decrease in terrigenous input. The limestones and marls continue into the Cenomanian. The occurrence of intercalated dark, laminated, carbonaceous marls and limestones, with up to 1.3 % TOC and kerogen of terrestrial or mixed marine and terrestrial origin is restricted to the Upper Albian, though. In the Babors realm a great part of Albian formations is missing due to erosional features.

The occurrence of O_2-depleted environments may be effected by an enhanced relief with several restricted basins created by first compressional movements. Just above this sequence the first resedimentation of diapiric Triassic saliferous material can be observed. It is thus possible that diapiric structures, mobilized by the compressional movements played an important role in the creation of an articulate topography on the North African Margin.

3.3.4 Neritic biogenic limestones (uppermost Albian-Lower Cenomanian; Plate 4, Figs. C-F)

This characteristic shallow-water facies occurs on both transversals in the northern units, at the limit of the Erraguene and the Brek units in the Babors and in the external Tanger Unit of the Rif transversal. In the Rif this facies type has only been observed as resediments. Sometimes huge sedimentary klippes more than 100 m large are found in Campanian/Maastrichtian deep-water facies. The age of these shallow-water limestones is Upper Albian/Vraconian (dated by the larger foraminifera *Orbitolina concava* and *Neoiraquia* sp. and ammonites, e. g. *Mortoniceras* sp.). In some outcrops

Plates 3 - 6

Characteristic facies types during the Cretaceous subsidence history of the North African paleomargin (examples from the Rif and the Babors)

Plate 3. Neocomian-Albian: First phase of subsidence.

Fig. A: Neocomian marl-limestone cycles (Prerif Zone, near Mjara).

Fig. B: Siliciclastic turbidite, distal part of the terrigenous sedimentation, which starts in the Middle Valanginian and ends at the base of the Upper Albian.

Fig. C: Channel-"megabed" of quartz-sandstone in the Lower Cretaceous pelitic sedimentation of the Rif (Mesorif Zone, Forêt d'Izzarene).

Fig. D1+2: Flaser-bedding in the terrigenous marginal channel sequences of the Mesorif Zone (Forêt d'Izzarene).

Fig. E: Debris-flow in the pelitic sequences of the Lower Cretaceous in the Babors (Erraguene Unit, near Lac d'Erraguene).

Plate 4. Upper Albian-Lower Cenomanian: First phase of compressional movements and uplift.

Development of an enhanced relief: black laminated limestones, partly biosiliceous sedimentation in the basinal areas, neritic biogenous facies on the highs.

Fig. A: Laminated black bituminous limestones of the Mesorif Zone (near Rhafsai).

Fig. B: Intensely slumped black limestones of the Mesorif Zone (near Rhafsai).

Fig. C-D: *Orbitolina* facies in the redeposited neritic Upper Albian/Lower Cenomanian in the external Tanger Unit (S of Tangier).

Fig. E: *Mortoniceras* sp. in the large huge olistolites of Upper Albian neritic facies in the Tangier area.

Fig. F: Cenomanian shallow-water redeposits in the Brek Unit of the Babors (typical shallow-water indicators are miliolid foraminifers and *Orbitolina* fragments).

Plate 5. Upper Cenomanian-Campanian/Maastrichtian: Second phase of subsidence.

Fig. A: Biosiliceous bituminous facies ("phthanites") around the Cenomanian/Turonian boundary of the Loukkos Zone (Loukkos river near Souk el Had).

Fig. B: Pelites and thin fine-grained calciturbidites (Argillite Formation II) in the Coniacian-Campanian of the Bab-Taza Unit (near Beni Ahmed).

Fig. C: Typical fibrous concretions ("bouses") in the Argillite Formation II of the Rif (Bab-Taza Unit near Beni Ahmed).

Fig. D: Thick pelitic series showing features of sediment gravity movements in the Upper Cretaceous (Campanian) of the Loukkos Zone.

Fig. E-F: Hemipelagic sedimentation of the Upper Cretaceous "faciès à boules jaunes" in the Babors: The intense bioturbation, the relative richness in organic material (flasers in Fig. F), locally the formation of framboid pyrite and the comparatively high sedimentation rates are characteristic features of this sequence. (Thin-section 196-2; magnification x 7.5, Neg.-Nr. 97/35 (E); magnification x 27, Neg.-Nr. 95/17 (F)).

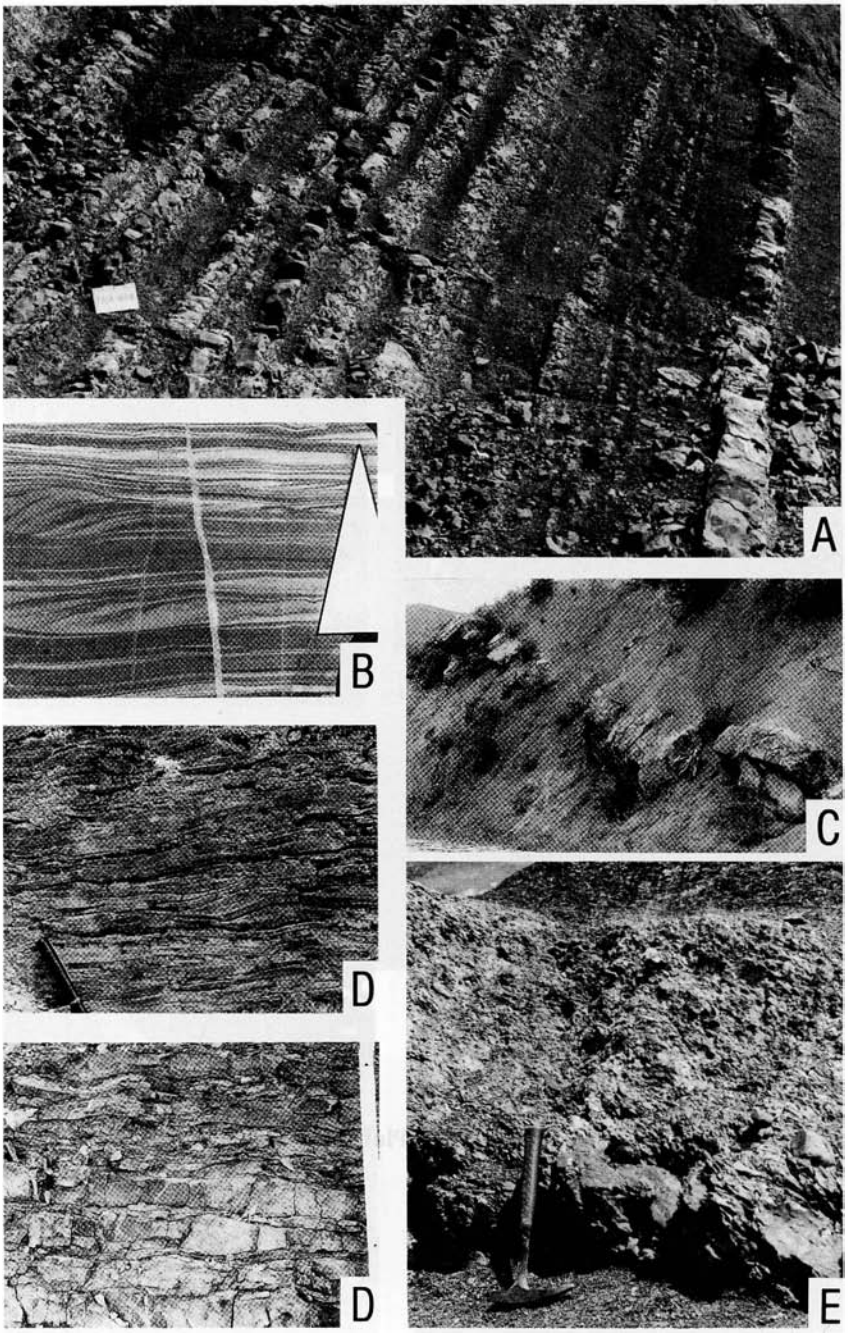


Plate 3

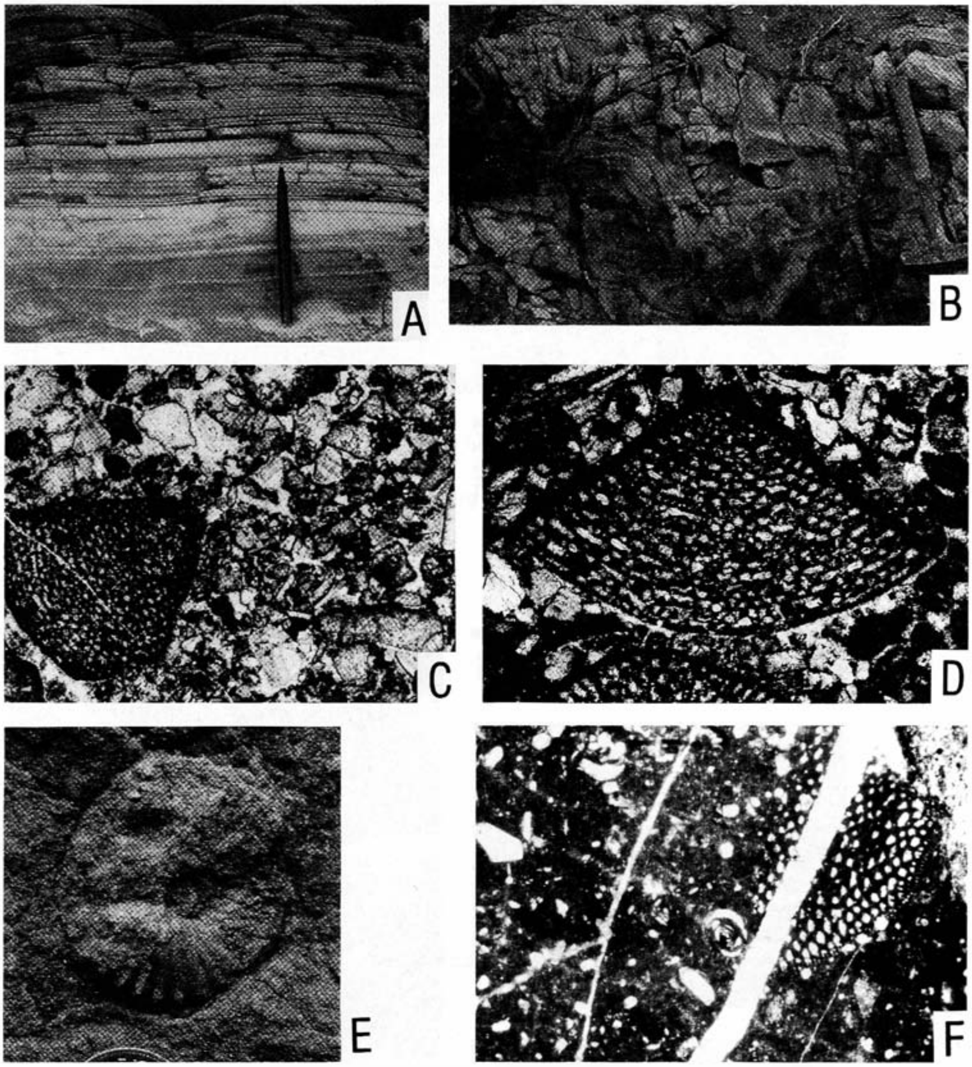


Plate 4

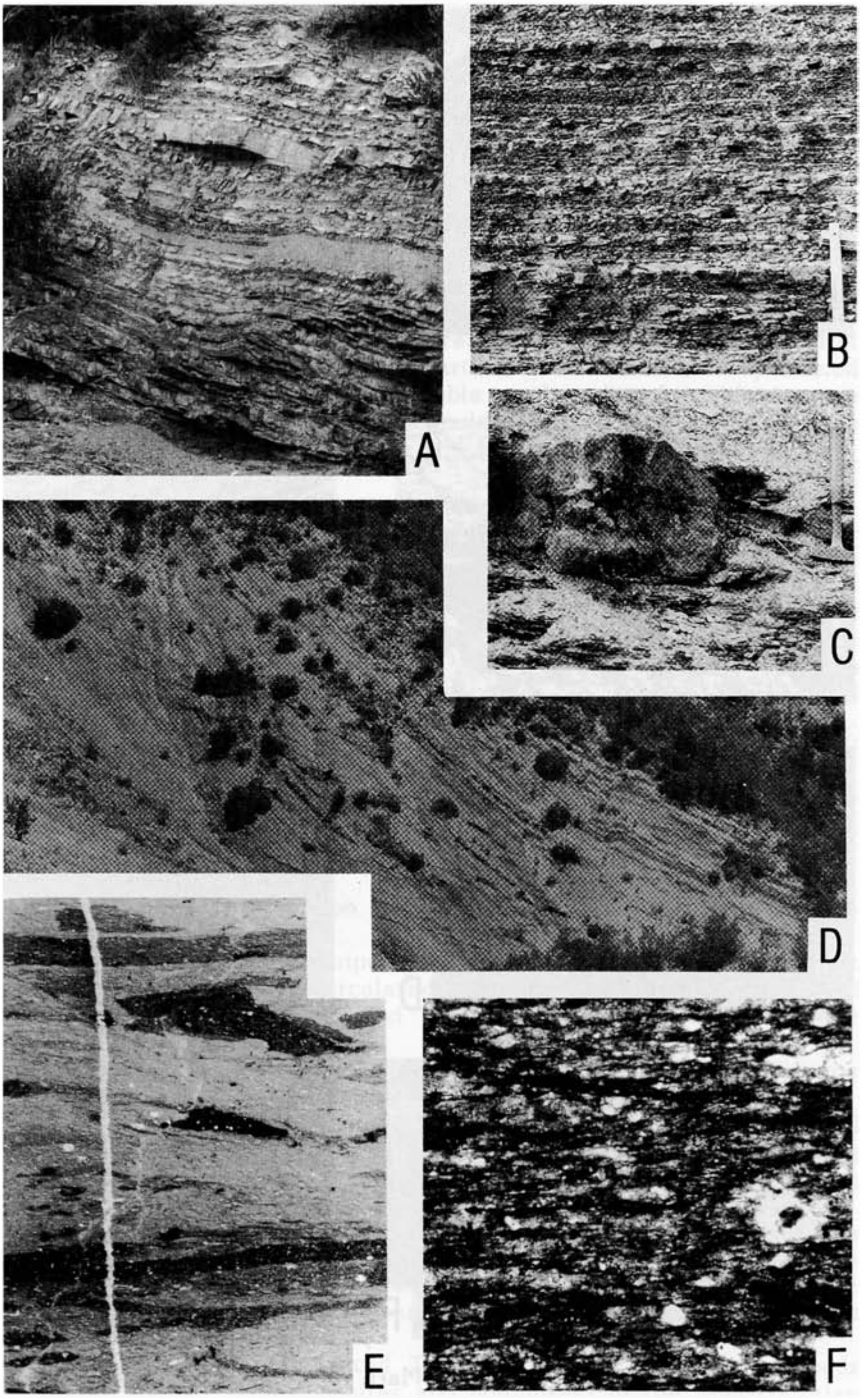


Plate 5

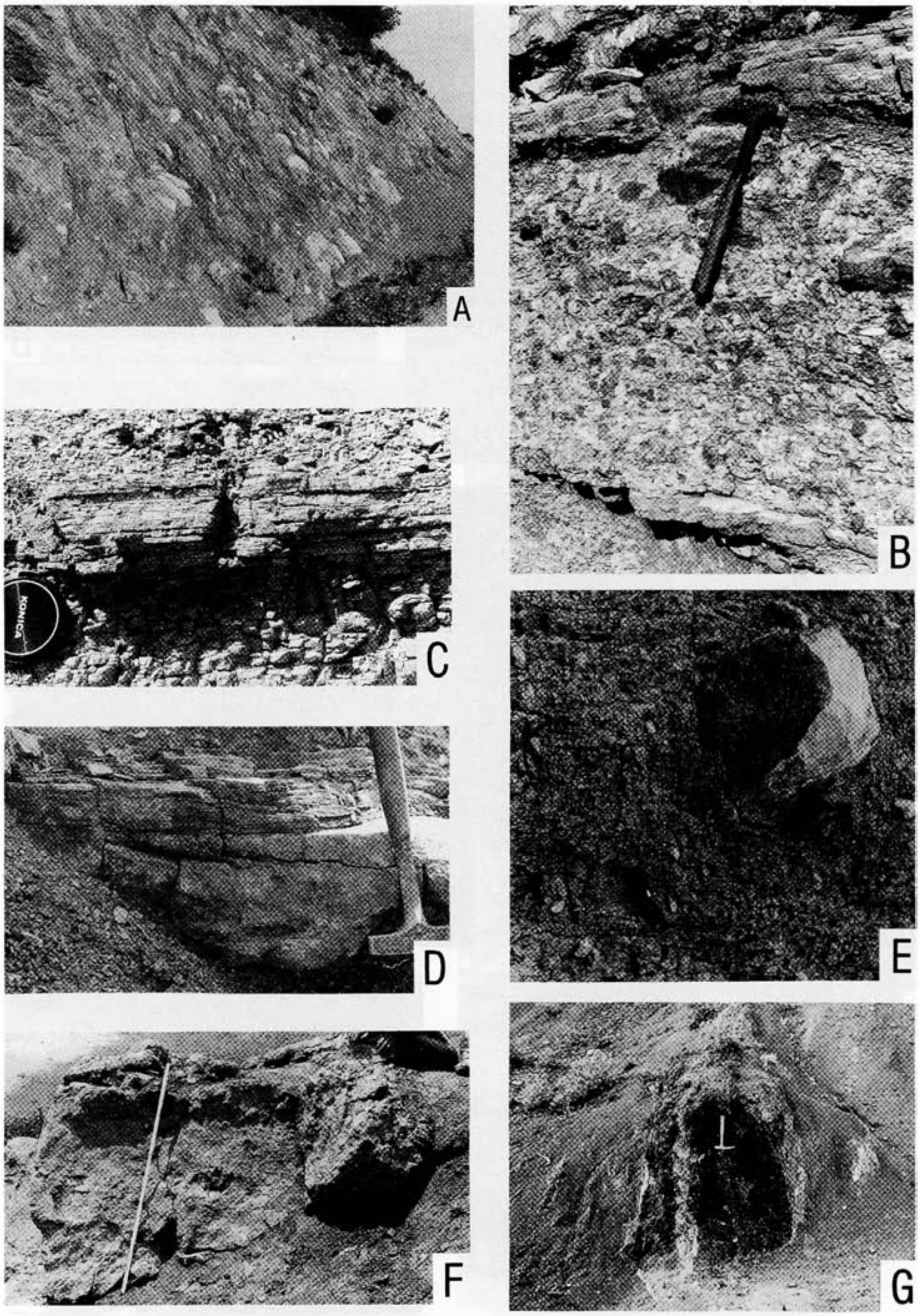


Plate 6

a first resedimentation of this shallow-water facies can be observed in Vraconian pelagic marls (dated by planktonic foraminiferal assemblages with the zonal marker species *Planomalina buxtorfi*). The underlying deposits of this shallow-water sequence are unknown. Even an analysis of the components of the Campanian/Maastrichtian olistostromes in the Tanger region did not yield any sediments older than Upper Albian/Vraconian.

The shallow-water facies in the Babors is partly (internal part of the Erraguene Unit) represented by rudistid limestones which seem to be in place. An enhanced relief can also be assumed for this region during Upper Albian/Lower Cenomanian, because of narrow distances between shallow-water rudistid platforms and pelagic radiolaritic sediments.

Both shallow-water facies are quite isolated within the pelagic environment of the internal part of the Tellian realm. The suggestion that first compressional movements in the northern part of the Tellian units produced uplifted areas seems to be the most probable explanation for this strange occurrence of neritic environments. This would signify a pre-Vraconian age for the first compressional movements on the North African paleomargin.

3.3.5 Pelagic marls and shales with low detrital input (Upper Albian-Cenomanian): Argillite Formation I of the distal sequences

Most of the series in the Rif area underly a rapid subsidence starting in Upper Albian/Cenomanian (begin of stage 3 of the Cretaceous subsidence history of the North African paleomargin). The following common features can be observed in most of the slope sequences on the Rif transversal (e. g. Internal Prerif, Mesorif, Loukkos, and Bab Taza zones):

Plate 6. Maastrichtian: Redeposition during the second phase of compressional movements.

Fig. A: Intensively slumped hemipelagites of the external Tanger Unit (near Beni Arouss) intercalated in this sequence are redeposited shallow-water sediments and Triassic material of an evaporite sequence.

Fig. B: Debris-flow and proximal turbidites in the Maastrichtian of the Mesorif Zone (near Tabouda).

Fig. C, D, F: Proximal turbidites, scour-fills and channel-fills of redeposited conglomeratic shallow-water material (Albian-Campanian) in a deep-water sequence of the Maastrichtian (Tanger Unit, region of Tangier).

Fig. E: Detail of a large olistostrome-like debris-flow body in the Maastrichtian of the Mesorif Zone near Tabouda. Typical resedimented components are Triassic evaporites and volcanics, Jurassic pelagic carbonates and Lower Cretaceous terrigenous sandstones and shales.

Fig. G: Resedimentation of injected Triassic material (salinar, red-beds of the Keuper facies, ophites and dolomites) in the Maastrichtian slope sequence of the Mesorif Zone (W of Mokhisset).

1. reduction of terrestrial or shallow-water biogenic detrital input,
2. a shift in the foraminiferal assemblages to a domination of planktonics in the southern (shelf and upper slope) series and to flysch-type assemblages in the northern (middle and lower slope) series; the latter suggest a position below an at least local CCD for the Bab Taza and Tanger series (Argillite Formation I), and
3. slumps and slides as common features accompanying the subsidence process in all units.

3.3.6 Siliceous blackshales around the Cenomanian/Turonian boundary (CTBE) (Text-Fig. 5; Plate 5, Fig. A)

Sequences of the upper slope (Prerif and external Loukkos zones) are characterized by a marly limestone sedimentation with a plankton-dominated foraminiferal microfauna. Intercalated are thin dark to black layers with up to 2.5 % TOC with kerogen mainly of marine origin (type II, HI 350/OI 40).

Black siliceous-carbonaceous shales and radiolarites ("phthanites") are the characteristic sedimentation of the deeper parts of the North African Slope during late Cenomanian and basal Turonian. The most important sedimentologic features of this formation, which represents the local expression of the global Cenomanian/Turonian Boundary Event (CTBE, THUROW & KUHNT 1986) are as follows:
- Main components of the sediments are silica, clay minerals and organic matter; in cases, plenty of radiolarians are recognizable
- Low sedimentation rates (< 1 m/m. y. - 10 m/m. y.)
- Fine lamination (often obscured by silicification)
- High TOC with increasing values towards the basin, where up to 13 % TOC (HI 400-650/OI 3-50, kerogen type II) have been observed in the basinal flysch units of the Rif (KUHNT et al. 1986).

3.3.7 Pelagic limestones, marls, and shales with low detrital input (Middle Turonian-Santonian): Argillite Formation II of the distal sequences (Plate 5, Figs. B-D)

Subsidence commenced with the late Albian/early Cenomanian (stage 3 of the subsidence history), continues throughout the Turonian-Santonian and is accompanied by a nearly detritus-free pelagic sedimentation. In the upper part of the continental slope (e. g. Prerif, Mesorif, and Loukkos zones of the Rif transversal and nearly all units of the Babors transversal) sedimentary gaps are common due to erosional features by slumping and sliding.

3.3.8 Hemipelagite sequence (often slumped marlstones and claystones, "faciès à boules jaunes"; Campanian-Maastrichtian; Plate 5, Figs. E, F, Plate 6, Fig. A)

Characteristic sedimentological features of this formation are:
1. Irregular lamination, made up by light (olive greenish) and dark (grey black) layers in the range of mm to few cm.
2. Occurrence of pyrite (framboidal pyrite, often as fillings of foraminiferal

tests), dark sediment colour, comparatively high contents in organic matter point to a slightly O_2-depleted sedimentary environment. This is in accordance with the common occurrence of depauperate foraminiferal assemblages which show dissolution of calcareous tests.

3. Intercalation of extremely fine-grained, light grey, turbiditic limestone beds with a graded base (silt fraction) and sharp basal contacts. Thickness of these "mud turbidites" is very variable.
4. Intense bioturbation, generally recognizable in black "flames" in the olive-greenish claystones. In cases, bioturbation completely homogenizes the sediment and produces mixed microfaunal assemblages of transported and autochthonous tests.
5. Slumping is common; locally small-scale slumps are diagenetically altered to limestone nodules (Plate 6, Fig. A).
6. Local occurrence of isolated olistolites, small-scale channel- (or scour-) fills, debris-flows, and proximal turbidites with reworked shallow-water sediments.
7. Sedimentation rate is generally high in comparison with the underlying series.

Facies patterns comparable to the hemipelagite series of the North African Margin are those of the hemipelagic rocks of the lower Monterey Formation (Early to Middle Miocene Sandholdt Member) on the Pacific Margin of North America. GARRISON & GRAHAM (1984) interprete these sediments as hemipelagic muds, probably deposited in dysaerobic to anaerobic environments which commonly excluded a burrowing infauna. The most likely environment was within or near the edges of the intersection of the oxygen-minimum zone with a submarine slope, an interpretation compatible with the upper to middle bathyal depths indicated by the benthic foraminifers.

Dolomite occurs in these laminated hemipelagic rocks as resistant lenticular beds and ellipsoidal concretions which are dark grey but weather to a conspicuous yellow-brown or ochre. The concretions are up to 3 m in maximum dimension, and lenses are hundreds of meters long and of varying thickness; both tend to be concentrated within specific stratigraphic intervals. The dolomites are of secondary origin and formed before the completion of compaction (GARRISON & GRAHAM 1984).

3.3.9 Olistostrome/debris-flow sequence with abundant calciturbiditic resediments of all grain sizes ("faciès à microbrèches", Campanian-Maastrichtian; Plate 6, Figs. B-G)

In the Campanian (or even Upper Santonian in the Babors) calciturbidites and debris-flows begin to intercalate in the pelagic marly or pelitic Upper Cretaceous series. This continuously increasing input of reworked marine sediments culminates in large-scale mass movements (thick debris-flows, olistostromes, sedimentary klippes) in nearly all external units of the Rif transversal (except the Prerif) during Lower Maastrichtian (KUHNT 1984). Even in the distal External and "Flysch" units an influence of these movements is visible in a remarkably increased sand/shale ratio and the occurrence of coarse-grained "proximal" turbidites.

There is a striking difference in the composition of the reworked material between the northern zones (Tanger Unit in the Tangier region) with predominantly reworked Upper Cretaceous shallow water facies and the southern zones (Mesorif and Loukkos units) where reworked Triassic evaporites and volcanic rocks (ophites) make up a remarkable part of the resediments. These Triassic resediments are evidential of important diapiric movements which occurred most likely along the listric faults separating single blocks of the subsiding margin.

These mass-movement features with a nearly coeval onset along the entire Tellian realm in Campanian/Lower Maastrichtian are interpreted to represent a second compressional phase which affected the North African paleomargin.

3.3.10 Shale with rare turbidites, slumps and olistolites (Maastrichtian-Eocene): Argillite Formation III of the distal sequences

At the beginning of the Maastrichtian an argillaceous pelitic series is developing in the Tanger region from the calciturbiditic formation through thinning and fining upward sequences. In its lower part this formation partly forms the matrix of olistostromes and contains locally calciturbiditic layers and limestone-olistolites. Upsection, this pelitic formation which is deposited well below an at least local CCD continues to the Lower/Middle Eocene with intercalations of a typical Lower Eocene siliceous facies. Within this sequence indications of compressional movements, e. g. olistostromes, turbidite sequences and resedimentation of Triassic material disappear and "normal" subsidence revives.

4. Subsidence history

Along the two studied palinspastic north-south cross-sections through the Tellian zones of the late Mesozoic North African paleomargin (Rif and Babors) three combined vertical profiles have been chosen to construct geohistory diagrams (VAN HINTE 1978, GUIDISH et al. 1985) for the Cretaceous time span (Text-Fig. 11).

The basis for the construction of these geohistory curves were synthetic sections of structural units, which are considered to have once formed a single block in the late Mesozoic. The units chosen on both transversals correspond to the Bas-, Mi-, and Haut-Tellian Units defined by WILDI (1983) and depicted in his palinspastic reconstruction of the Tellian zones. Reconstruction of thickness and paleobathymetry of Cretaceous sediments of these units is still only possible within a relatively wide range of error and the precision of the data base differs for different settings and time slices. Thus, corrections for compaction and sea level oscillation have not been made, because their effect falls into the range of exactness of the field data. Even when taking maximum and minimum assumptions for bathymetries and thickness of some strata into account, the comparison of subsidence in the different units does not differ very much from the average estimation used in Text-Fig. 11. This comparison of the subsidence curves on each transversal shows marked differences in subsidence history between north and south. Generally, the subsidence in the southern proximal (shelf) areas is minor compared to the more distal parts of the continental margin in the north. Listric faults between different subsiding blocks forming the late Mesozoic North African Margin seem to be responsible for the different subsidence history in the different studied units.

Studies on undeformed continental margins show that subsidence history in distal parts of a continental margin is closely related to crustal thinning in the transitional zone between continental and oceanic crust (WATTS & RYAN 1976, KEEN 1979). In the most distal parts of a continental margin the subsidence curve is similar to an ideal curve for oceanic crust (SCLATER et al. 1971), and subsidence decreases towards the continent (WATTS & STECKLER 1979, HARDENBOL et al. 1981). This general picture can also be observed on the Rif transversal, where the most distal parts show subsidence features of a remarkably thinned continental crust. The geohistory of the Babors transect appears more complicated due to some areas uplifted by compressional movements. It generally shows features of a subsiding normal continental crust, i. e. subsiding shelf areas.

A comparison of the Cretaceous subsidence curves with those compiled by FUNK (1985) for the Helvetic Shelf and the South Alpine realm (Monte Generoso) shows similarities between the slowly subsiding Helvetic Shelf and the proximal Babor and Prerif zones. Subsidence increased slightly in the Erraguene, Brek and Mesorif zones and is very high in the Ketama/Tanger units of the Rif transversal, which may point to a thinned crust and beginning influence of thermal subsidence in this distal zone of the North African paleomargin.

5. Influence of Cretaceous compressional movements and transversal tectonics on the sedimentary evolution of the margin

The complex tectonic evolution along the Rif and Babors transversals is linked with the particular position of these domains on the northern margin

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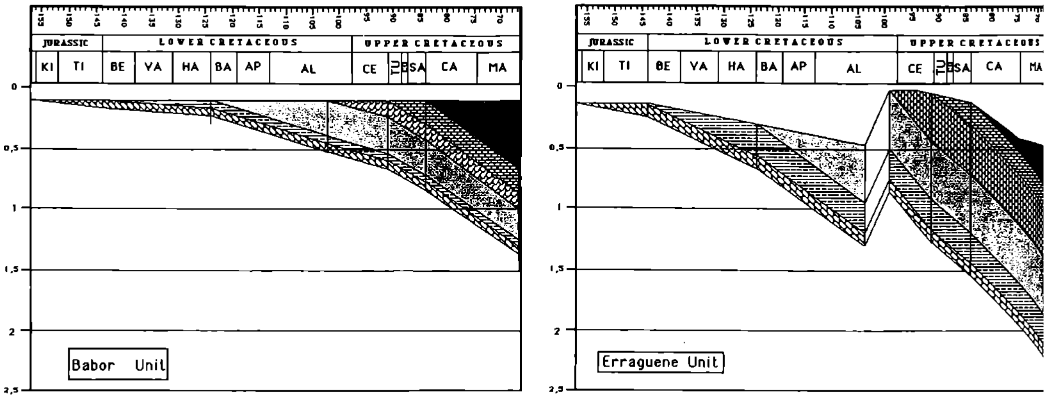
Text-Fig. 11. Calculation of burial history of the Cretaceous sedimentary wedge in selected sections from the Rif and Babors transects. Geohistory curves have been constructed for the late Jurassic and Cretaceous of "Bas"-, "Mi"- and "Haut"-Tellian units (WILDI 1983) on both transects. - Upper row: Babors transversal (sediment thickness according to OBERT 1981, 1984). - Lower row: Rif transversal (sediment thickness for the Rides Prérifaines according to FAUGERES (1978), KUHNT (1987); for the Prerif Zone according to data from LABUDE (1978) for the Kimmeridgian-Berriasian, KUHNT (1978) for the Lower Cretaceous, KUHNT (1987, and unpublished data) for the Middle and Upper Cretaceous; Mesorif Zone: data compiled from WILDI (1981, 1983), SUTER (pers. comm.), KUHNT (1978, 1987); sediment thickness for the Ketama and Bab Taza units based on data from the Djebel Tizerhine section of ANDRIEUX (1971) for the Upper Jurassic and Neocomian, from the Ketama area for the Aptian and Albian (ANDRIEUX 1971, GÜBELI 1982), and from the area between Bab Taza and Beni Ahmed for the Upper Cretaceous (LESPINASSE 1975, KUHNT 1987). Relative position of the single paleogeographic zones on a continental margin model is indicated. - Southern (shelf) units to the left, northern (slope) units to the right. Compaction not corrected. Uppermost curve indicates suggested water depth according to micropaleontologic data. Note different depth-scale for Rif and Babors!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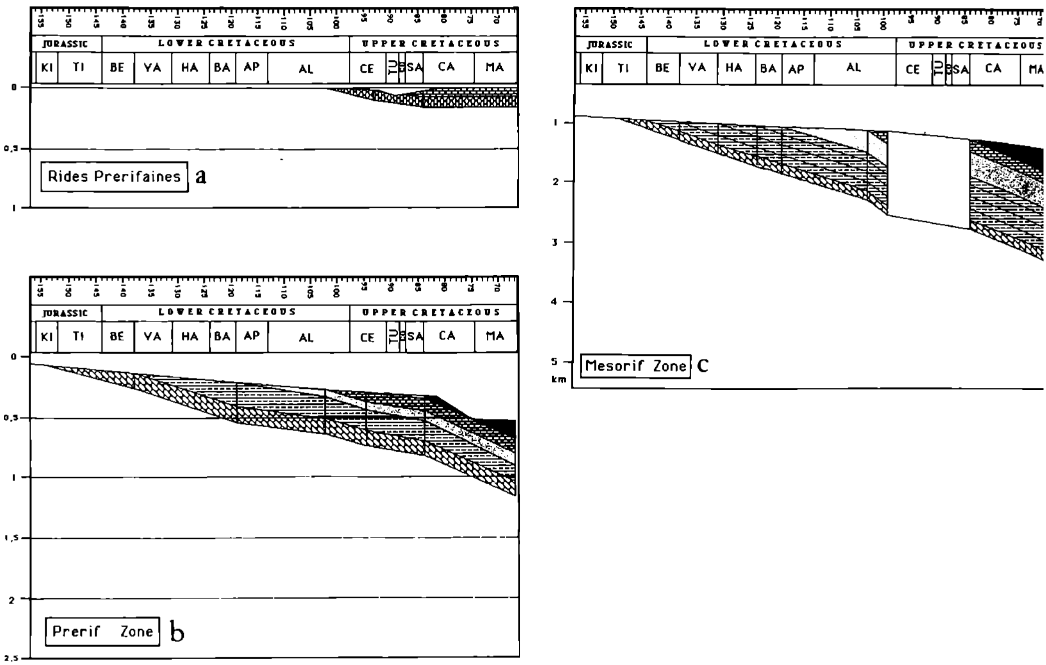
Erraguene Unit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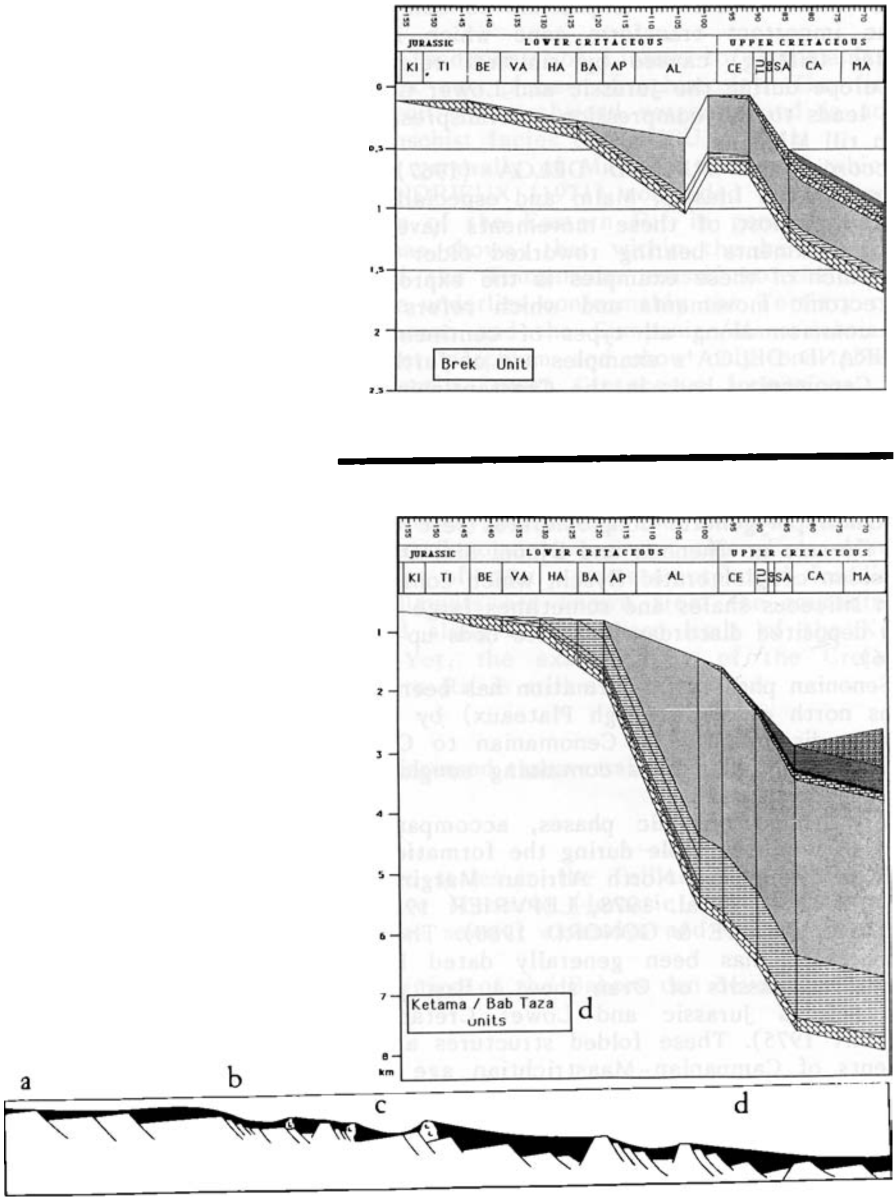
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KI TI BE VA HA BA AP AL CE SA CA MA
1
2
3
4
5
6
7
8
km
Ketama / Bab Taza units
d
a
b
c
d

of the African Plate. It is possible to comprise this paleotectonic evolution to the important transform zone which is characterized by distensional features (rifting) caused by sinistral relative movements between Africa and Europe during the Jurassic and Lower Cretaceous. The following convergence leads to the compression or transpression of this zone since the Upper Albian till Miocene.

According to DURAND DELGA (1967), independent local movements occurred in the Liassic, Malm and especially during several episodes of the Cretaceous. Most of these movements have been interpreted from conglomeratic sediments bearing reworked older formations, and it is not quite clear which of these examples is the expression of important compressional paleotectonic movements and which refers to simple resedimentation features common along all types of continental margins. Nevertheless, some of DURAND DELGA's examples are of further interest:

1. Cenomanian beds in the Constantinois area of the Tellian Atlas contain conglomerates of reworked Jurassic and rest discordantly on the Lower Cretaceous of the Western Kabylie des Babors.

2. Upper Cretaceous movements have been recorded from the Alboran margin as well as from the North African Margin:

- Senonian polygenetic conglomerates rest unconformably on the Liassic in the Dorsale at Chenoua and Djebel Sidi-Dris.
- Senonian conglomeratic flysch, which contains abundant reworked Cenomanian siliceous shales and sometimes large exotic blocks of Malm sediments was deposited discordantly across beds up to Albian in age (e. g. RAOULT 1966).
- A Senonian phase of deformation has been reported from the Hodna Mountains north of Msila (High Plateaux) by KIEKEN (1962) - Maastrichtian resting discordantly on Cenomanian to Campanian, and by BERTRANEU (1952) - Santonian beds containing conglomerates unconformably overlying Albian sediments.

Early Alpine tectonic phases, accompanied by metamorphism probably played an important role during the formation of metamorphic and schistose complexes along the North African Margin (OBERT 1974, DUBEL et al. 1977, CENTENE et al. 1978, LEPVRIER 1978, MALUSKI et al. 1979, FRIZON DE LAMOTTE & GONORD 1980). The first phase of schistosity and metamorphism has been generally dated between Albian and Campanian. The coastal massifs of Oran show a first phase of schistosity and folding, which affects Jurassic and Lower Cretaceous sediments (FENET 1975, GUARDIA 1975). These folded structures are overlain by unmetamorphozed sediments of Campanian-Maastrichtian age and Upper Cretaceous (Campanian) conglomerates, containing Triassic material, which give evidence of a pre-Campanian paleotectonic event (CENTENE et al. 1978, BANDET et al. 1979, FRIZON DE LAMOTTE & GONORD 1980). Syntectonic metamorphism of the greenschist facies affected Jurassic and Lower Cretaceous sediments in the region of the Chelif and Mouzaia-Blida. The age of this metamorphism has been dated as 85 m. y. (MALUSKI et al. 1979). This metamorphic phase has been accompanied by important folding with southward vergency. A second paleotectonic phase already affected these structures and was also of Cretaceous age (LEPVRIER 1978, MALUSKI et al. 1979).

In the Rif, schistosity has been observed in parautochthonous external units, which occur in windows below the flysch nappes to the north and in the external "Nappes Rifaines" to the south (MARÇAIS & SUTER 1966, ANDRIEUX 1971, 1973). These parautochthonous units can be subdivided into the Ketama Unit to the NW and the Temsamane Unit to the SE, both separated by the Nekor Fault.

The **Ketama Unit** from which sediments of Liassic to Cenomanian age are known is affected by at least two phases of schistosity. The first of these two phases shows folding with a southward vergency and is accompanied by metamorphism of greenschist facies (ANDRIEUX 1971).

The **Temsamane Unit** consists generally of Miocene sediments which are affected by schistosity. Thus, ANDRIEUX (1971) concluded a Miocene age for schistosity and metamorphism of the Eastern Rif in general. Recently FRIZON DE LAMOTTE (1982) has shown that within the half-window of the Nekor river (prolongation of the Temsamane massif to the west) a Campanian/Maastrichtian sequence underlies conformably the Tertiary of the Temsamane Unit. Both, the Miocene and the Campanian/Maastrichtian sequences are not affected by metamorphism and show only one phase of schistosity. They overly discordantly Lower Cretaceous formations, which are much more intensely deformed and affected by two phases of schistosity. The first paleotectonic phase, accompanied by metamorphism in the Eastern Rif, is thus of pre-Campanian age (FRIZON DE LAMOTTE & GONORD 1980). Proceeding on the assumption that the Cenomanian of the Ketama Unit has also been affected by schistosity and metamorphism, these authors (op. cit.) concluded a post-Cenomanian age for this phase. However, in the meantime several outcrops of nearly unmetamorphosed fossiliferous Vraconian/Lower Cenomanian sediments are known from the southern part (LEBLANC & WERNLI 1980) and along the northern limit of the Ketama Unit (ASEBRIY pers. comm.). Yet, the exact dating of the Cretaceous paleotectonic events in the Eastern Rif is still a challenge.

5.1 Observations on the two discussed transversals

5.1.1 Babors (Text-Fig. 12A)

The outcrop of complete Mesozoic series in the Tellian units of the Babors enables us to reconstruct the late Mesozoic (Jurassic and Cretaceous) depositional history and to recognize several extensive and compressive Alpine paleotectonic events.

Seven main phases in the evolution of the Babors can be discriminated:

1. Jurassic rifting

A carbonate platform is built up during Lower Lias. Distension is becoming more pronounced since the Lower Domerian. Subsidence, drowning, and faulting (OBERT 1986) of the carbonate platform continues to the Upper Jurassic. This stage of subsidence is locally accompanied by the formation of conglomeratic deposits, mainly in the northerly units (Brek-Gouraya).

2. Late Jurassic transpressional movements

First paleotectonic movements can be observed in the late Jurassic. Transversal deformations with N-S trending axes are reflected by the Upper Jurassic facies distribution: Terrigenous facies (comparable to the "Ferrysch" of WILDI 1981) occur in the western zones. Pelagic and calciturbiditic sediments characterize the central part of the Babors. Pelagic sedimentation, devoid of detrital input occurs in the eastern part. Additionally, folding with N-S trending axes has been observed in the Jurassic limestone massifs of the northerly units (OBERT 1981, 1984). Lower Cretaceous sediments overlie discordantly these late Jurassic folds.

3. Early Cretaceous subsidence

Distensional features and subsidence can be observed in all units during Neocomian.

4. Late Lower Cretaceous compression

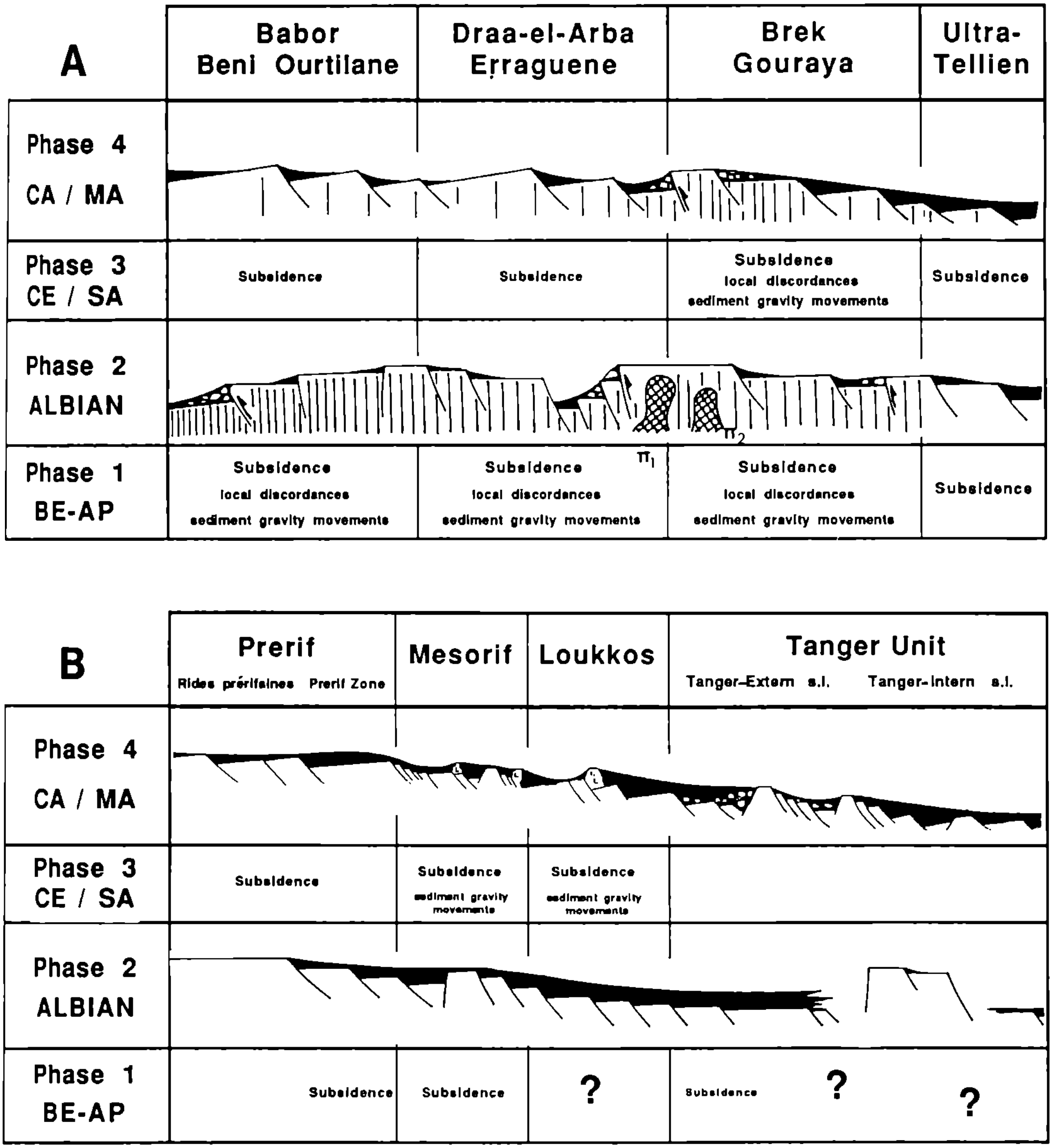
In the Aptian first deformations can be observed, which become more and more important and result in a paroxysmal Upper Albian tectonic phase, which is accompanied by magmatic events, by metamorphism of ante-Albian sediments in the greenschist facies (with chlorite, albite, epidote), and by schistosity. The first phase of deformation in the Aptian is characterized by the W-E to NW-SE folding and formation of sedimentary klippes and conglomerates, generally reworking Jurassic and early Cretaceous material. It is also notable that first resediments of Triassic evaporites have been observed within the Erraguene Unit. A post-Neocomian discordance is clearly visible in the zones of the Djebel Babor and the Djebel Brek. These movements continue progressively to the ante-Vraconian main compressional phase, which exhibits a first maximum of injective and redeposited Triassic material. However, in some areas (southern part of the Erraguene Unit) subsidence still continues.

5. Upper Cretaceous subsiding blocks

After the Upper Albian compressional phase, the northern part of the Babors (Brek-Gouraya and Erraguene units) forms an uplifted block, partly emerged and with erosional features in its central part and shallow water deposits on its margins (rudistid facies). These shallow water deposits are also found as resediments in the adjacent basins. This uplifted zone is progressively subsiding throughout Upper Cretaceous. Discordances, conglomeratic resedimentation, and resedimentation of Triassic material can be observed locally. These sediment-gravity movements, which indicate an enhanced relief might be due to different subsidence rates of the single blocks, and/or may indicate local compressional or transpressional movements.

6. Late Cretaceous compression

A supra-regional discordance can be observed in the early Campanian. The fold-axes of these deformations repeat the E-W directions of the late Lower Cretaceous folds. This folding is followed by a second generation of folding with variable (in general N-S) directions of the folding axes. This second folding is accompanied by imperfectly developed shear cleavages.



Text-Fig. 12. Comparison of the Cretaceous evolution of the two transects: evolution of the margin, subsidence history and paleotectonic events. Note the generally higher thicknesses of sediment pile (in black) and water depths along the Rif transversal. **Metamorphism:** Density of vertical lines is proportional to metamorphic intensity which has been observed in each unit. Metamorphism is anchizonal to epizonal during the Lower Cretaceous, and lower during the Senonian, causing only an aggradation to illitic pole of clay minerals. **Magmatism:** Located along the boundary of septentrional units. First, there are alkaline dolerites (π_1). Their precise dating is dubious: between Neocomian and Albian. Then, quartziferous diorite (π_2) metamorphizes Lower Albian schists. They are anterior to Senonian marls which are not metamorphized and overlay the schists.

The two directions of schistosity produce the typical "pommes frites" pattern of the early Cretaceous schists. Folding and schistosity can be observed throughout the Maastrichtian, but seems to be absent in the Cenozoic sediments, thus indicating a late Cretaceous compressional phase.

7. Tertiary overthrusting and formation of nappes

Towards the Upper Eocene (Priabonian?) the thrusting of the whole domain results in the formation of the Tellian nappes which sheared off along the middle Cretaceous discontinuity surface. The lower part of the series (Jurassic and Lower Cretaceous) is cut off in slices with southward vergency. This event is followed by several minor tectonic phases, i. e. limited back thrusting, revival of southward overthrustings, sinistral strike-slip faults.

5.1.2 Western Rif (Text-Fig. 12B)

The analysis of the late Mesozoic sedimentary history of the North African Continental Margin in the external zones of the Western Rif enables discrimination of the following evolutionary stages:

1. Early rifting of the Triassic to early Jurassic

An evaporitic lagoonal Triassic facies (red-beds, dolomites, multicoloured "Keuper" facies) is accompanied by basaltic "ophitic" volcanism. Enhanced subsidence to shallow water environments in the early Liassic leads to the formation of a carbonate platform.

2. Jurassic rifting

Increased subsidence results in drowning of the carbonate platform and deposition of pelagic limestones with rich Pliensbachian ammonite faunas and in the formation of a Bajocian "ammonitico rosso" facies in the northern zones. In the southern zones (Rides Prérifaines) subsidence rates decrease, followed by an emersion in the Upper Bajocian (FAUGERES 1981). A pelagic *Posidonia*-marl facies of the Bathonian is succeeded by a deltaic sedimentation, prograding from SE to NW and leading to submarine fan sedimentation in the deeper parts of the basin (Callovian-Upper Oxfordian "Ferrysch", WILDI 1981).

3. Late Jurassic basin and swell topography

Pelagic limestones of Kimmeridgian to Berriasian reveal a facies distribution, which is typical for a basin and swell topography ("ammonitico rosso" facies with hardgrounds, short-term emersion or disconformities on the highs, important debris-flow and calciturbiditic sequences on the margins and pelagic calpionellid limestone facies in the basins). This pronounced relief might be, in parts, the result of late Kimmerian tectonical movements.

4. Early Cretaceous subsidence

Rapidly increasing subsidence in the Berriasian/Lower Valanginian leads to a progressive deepening of the basin from south (Prerif) to north (Mesorif and Intrarif). Since Middle/Upper Valanginian, spikes of terrigenous turbiditic sedimentation can be observed in all zones with exception of parts in the Prerif.

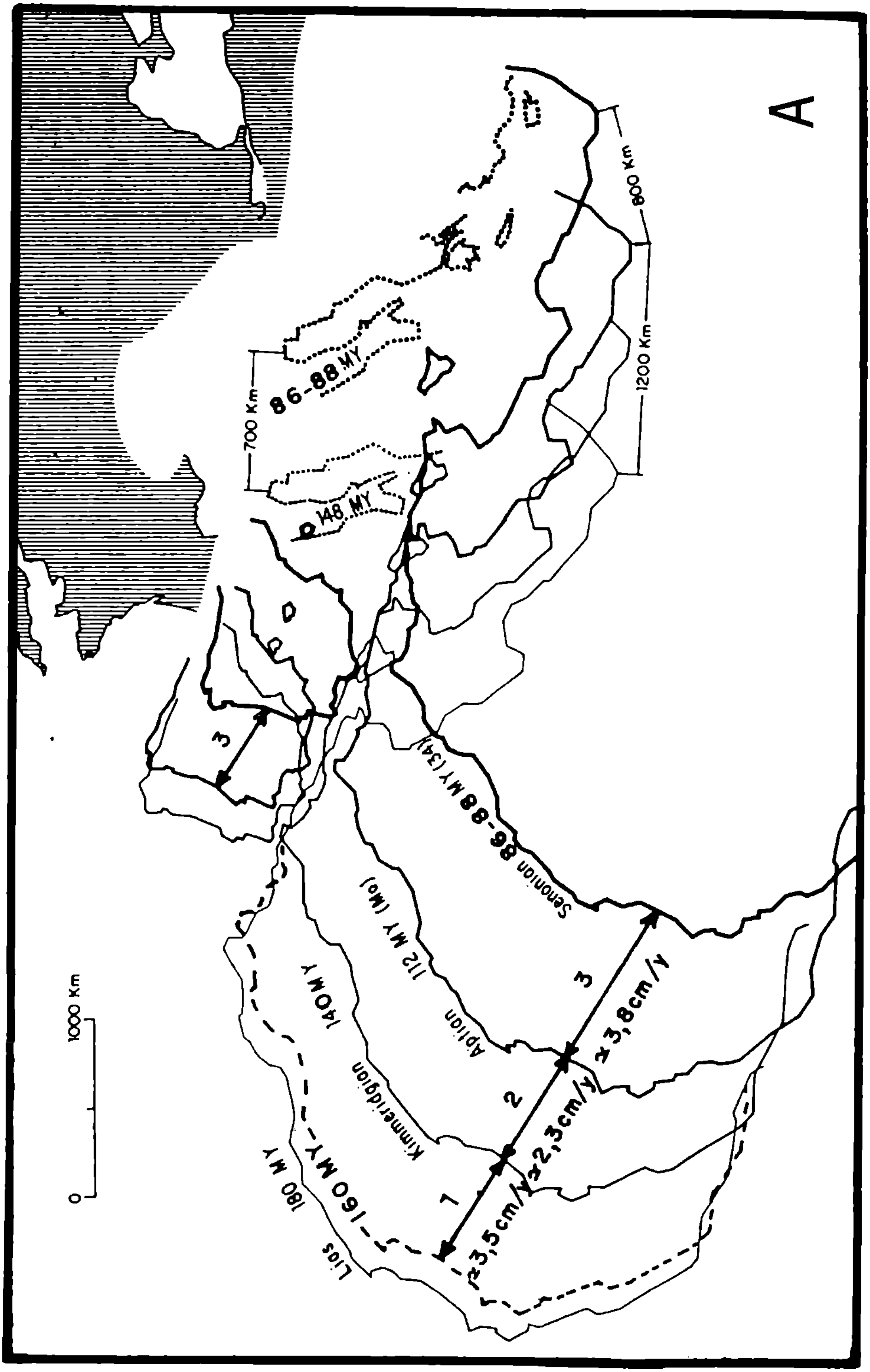
5. Channelized margin and terrigenous fan systems of the Aptian and Albian

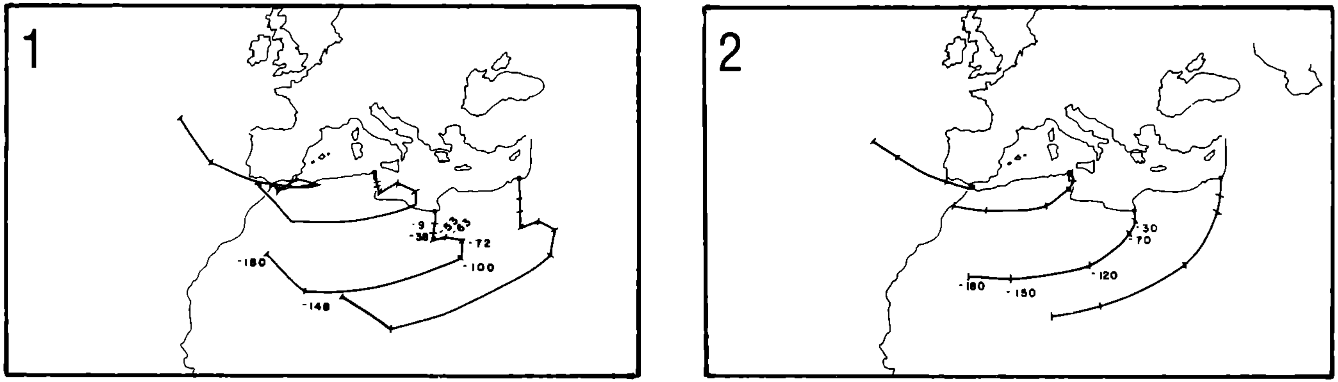
Since (Upper?) Aptian, the detritic input increases and governs the Aptian-Albian siliciclastic regime throughout the basin. Coeval steepening of the relief is reflected by channelized slope facies and redeposition of shallow-water biogenic facies in the deep-sea environment.

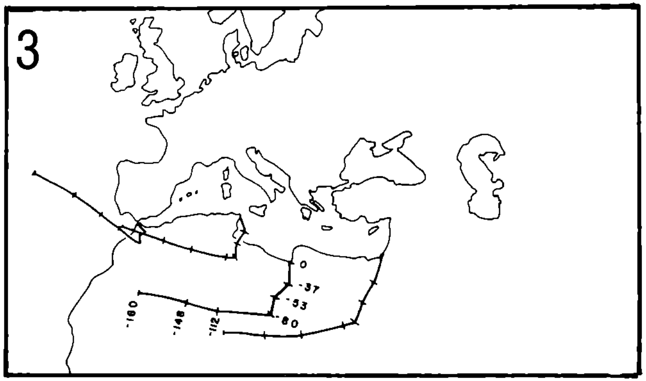
6. Upper Albian black shale event

A distinct lithologic change can be observed in the Upper Albian. Aptian/Albian clastics are followed by mainly marly or limy sedimentation along with strongly reduced terrigenous input. This lithologic change is accompanied by a widespread "black shale" sedimentation with black laminated limestones and marls, often showing slumping and sliding features. The formation of these organic-rich sediments (mixed terrestrial and marine organic matter) may be favoured by an enhanced relief and the formation of several isolated slope-basins. Occasionally, first resedimented Triassic evaporites and volcanics give evidence of diapiric activity. All these features are probably the expression of a paleotectonic (transpressive?) event of minor importance.

Text-Fig. 13. Mesozoic relative movements of Africa and Iberia compared with Eurasia.
A. Reconstruction according to magnetic anomalies in the Atlantic Ocean (after OLIVET et al. 1984). Note the faster relative eastward movement of Africa during Jurassic and Neocomian, leading to transtensive tectonics along the North African Margin. The more or less congruent eastward movement of Africa and Iberia since Aptian (Mo) coincides with first transpressional movements along the North African Margin.
B. Movement paths of Africa with respect to Eurasia (from OLIVET et al. 1984). 1: according to BIJU-DUVAL et al. (1977) - 2: according to TAPPONNIER (1977) - 3: according to OLIVET (1978) and OLIVET et al. (1984).
Numbers are in million years B. P. Note the different hypotheses on the timing of first compressional movements: abruptly at about 100 m. y. (late Albian) in figure 1, continuously between 120 m. y. (Barremian) and 70 m. y. (Maastrichtian) in figure 2, and abruptly at about 80 m. y. (late Santonian/early Campanian, depending on timescale) in figure 3.


A
800 Km
1200 Km
700 Km
86-88 MY
148 MY
3
Senonian 86-88 MY (34)
112 MY (Mo)
Aptian
140 MY
Kimeridgian
160 MY
Lias
180 MY
1
2
3
≥3,5 cm/y
≥2,3 cm/y
≥3,8 cm/y
0
1000 Km





CHANGE IN RELATIVE PLATE MOTION BETWEEN AFRICA AND EUROPE DURING THE CRETACEOUS

B

7. Upper Cretaceous subsidence

Increasing subsidence of northerly parts of the slope leads to autochthonous pelagic sedimentation below CCD. Pelagic sedimentation in the distal parts of the slope can be compared to facies successions known from the North Atlantic Ocean and from the flysch basins of the western Mediterranean orogen: (1) Cenomanian argillitic sequence (argillite series 1), (2) organic-rich, biosiliceous sedimentation at the Cenomanian/Turonian boundary (CTBE, THUROW & KUHNT 1986), (3) Turonian-Lower Campanian argillitic sequence (argillite series 2) which grades continuously into a late Cretaceous hemipelagic and/or calciturbiditic sequence. Locally, i. e. in the western part of the Loukkos Zone and in some parts of the external Tanger Unit neritic biogenic sedimentation persisted throughout most of the Upper Cretaceous, probably on top of either rotated fault-blocks or isolated tectonic highs on the slope. These neritic sediments (e. g. *Orbitolina*-limestones of late Albian/Cenomanian age, early Senonian oyster-shell beds and Campanian *Inoceramus*-limestones) are generally found as huge sedimentary klippes, olistolites and components of debris-flows within Maastrichtian slope sediments.

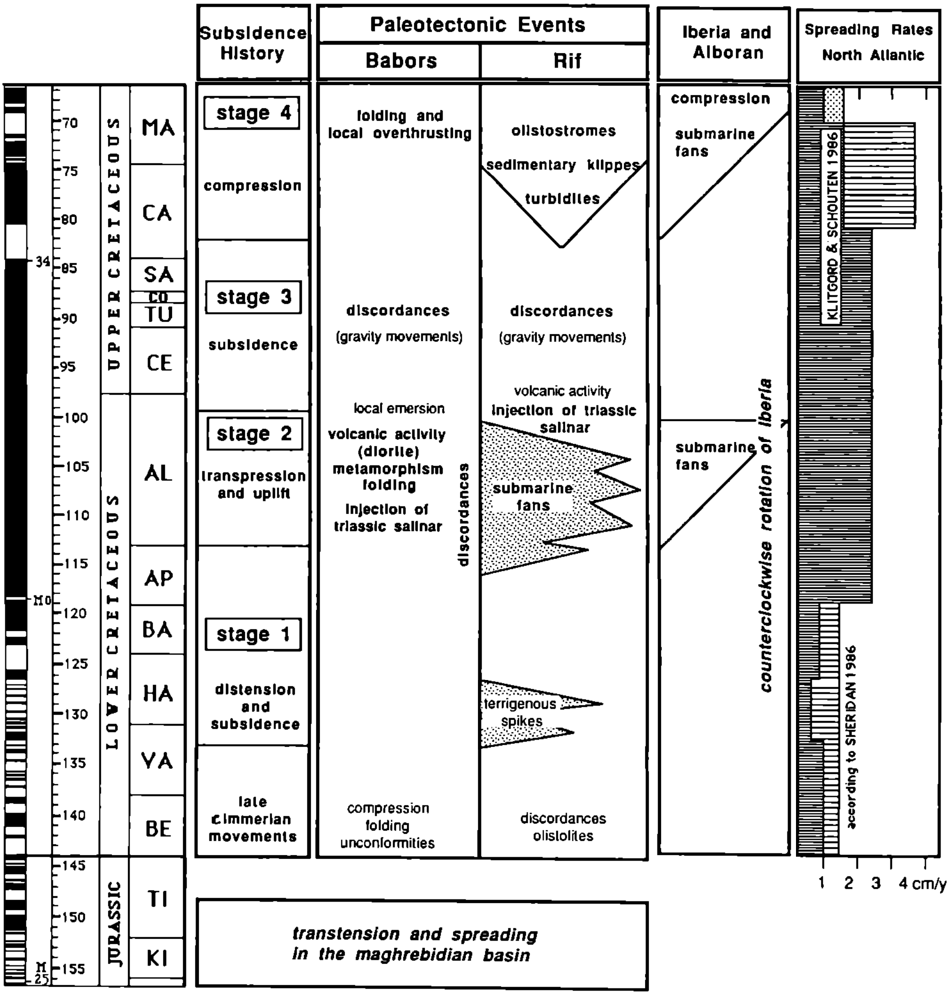
8. Late Cretaceous compressive or transpressive phase

Beginning in the Lower Campanian and culminating in the Lower Maastrichtian, slumping and sliding (huge sedimentary klippes, olistostromes, pebbly mudstones, debris-flows, and calciturbidites) are frequent. Resedimented Triassic evaporites and volcanics form gypseous polygenic breccias, and are frequent in the Mesorif and Loukkos Zones. These redeposits have been interpreted as the result of synsedimentary nappe-emplacement or strike-slip faulting (ASEBRIY 1984, ASEBRIY et al. 1987). In the Tanger Unit the resediments are dominated by Cretaceous neritic and slope sediments. In the Maastrichtian flysch-type sequence of the Ain Lahcen Unit typical Lower and Middle Cretaceous **deep-water** sediments occur as olistolites and as components of debris-flows: Lower Cretaceous quartzite-turbidites and pelitic sediments of the Cenomanian/Turonian with "phthanite" intercalations (radiolarites, siliceous bituminous sediments). These components give evidence of compressive movements and tectonic uplift, followed by thinning and fining upwards cycles of a calciturbidite sequence in the Lower/Middle Maastrichtian.

5.2 Comparison with the supra-regional plate tectonic pattern

From the opening history of the Atlantic Ocean three main phases influenced the North African Margin (Text-Figs. 13, 14):

1. Rifting and, since Callovian/Oxfordian, spreading in the Central North Atlantic south of the Azores transform fault leads to an eastward movement of the African Plate relative to Europe and the ALKAPECA microcontinent with subsequent transtensive movements along the North African Margin.
2. Spreading in the North Atlantic north of the Azores transform fault begins in the Lower Cretaceous (break-up unconformity at the Galicia



Text-Fig. 14. Cretaceous subsidence history and paleotectonics in the Babors and Rif, comparison with spreading events in the North Atlantic and movements between the European and African plates. Timescale and magnetic reversal scale according to PALMER (1983), movements of the Iberian plate after MALOD (1982). Spreading rates in the Atlantic Ocean according to SHERIDAN (1986) and KLITGORD & SCHOUTEN (1986).

Bank is Aptian in age, BOILLOT et al. 1985) and the relative motion of Africa and Europe becomes less significant. Consequently, the transtensive movements between Europe and Africa loose their significance and subsequently first signs of transpressional movements can be observed along the North African Margin.

3. Since Albian a general compressional trend between Europe and Africa can be observed, probably connected with spreading in the South Atlantic.

During all stages old NNE-SSW striking Hercynian lineaments on the North African Margin, e. g. the Babors fault, seem to play an important role as movement paths for transverse as well as for vertical movements.

6. Conclusion

Summing up, it may be said that the sedimentary fill of marginal basins along the two studied transects of the Mesozoic North African continental margin show a very similar depositional history. This is thought to reflect the uniform subsidence along this margin, as well as the influence of supra-regional paleoceanographic events and the regional character of tectonics.

Biostratigraphic data from both transects suggest that the succession of sedimentary formations might be influenced by periods of extensional or compressional stress as recorded in the sedimentary sequences near the outer edges of the African Plate by periods of tectonism. One pronounced difference between both transects is the restriction of the Babors transect to the shelf, and the basins and marginal highs on the upper to middle slope, while the Western Rif comprises a topographically complex but nearly complete bathymetric transect from the coastline to the deep sea. These differences may be partly caused by the more important role of late Jurassic and late Albian compressional phases in the Babors.

How the interfingering was established between the eastern North Atlantic and the western termination of the Mediterranean Tethys west of the Alboran Massif is still a matter of discussion. It is of interest to note, that the times of tectonic activity on the studied transects of the late Mesozoic North African Margin correlate perfectly with episodes of major tectonic activity on the Atlantic continental margins of eastern North America and NW Africa (JANSA & WIEDMANN 1982):

1. Late Kimmerian tectonics at the Jurassic-Cretaceous boundary: tilting and warping of marginal basins, creation of transverse extensional fractures, drowning of carbonate platforms and extinction of reef biotopes, rejuvenation of hinterlands as a source area of major clastic influx, and most probably initiation of salt motion in the future diapiric zones.
2. Austrian tectonics of the late Albian: uplift of the central portion of the Grand Banks and initial rise of the Avalon Uplift at the northeastern American margin (JANSA & WADE 1975); compressive tectonic activity has also been observed in the NW African Tarfaya-Aaiun Atlantic coastal basin (JANSA & WIEDMANN 1982).
3. Laramid tectonism of the late Upper Cretaceous: uplift of the northwestern African Atlantic coastal basins, compressional tectonics in the Senegal and Tarfaya-Aaiun Basin.

Although the general subsidence history pattern of the two marginal transects shows striking similarities, an important difference is seen in the more pronounced compressional movements in the Babors, especially during the Upper Albian compressional phase 2. The reason for this different evolution may be a more pronounced influence of the Alpine compressional movements (Austrian phase) on the eastern part of the margin, whereas the western part probably occupies an intermediate position between subsiding Atlantic margins with spreading in the oceanic basin, and western Mediterranean transpressional zones.

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Upper Cretaceous Deep-Water Agglutinated Benthic Foraminiferal Assemblages from the Western Mediterranean and Adjacent Areas

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With 6 Text-Figures

KUHNT, W. & KAMINSKI, M. (1989): Upper Cretaceous Deep-Water Agglutinated Benthic Foraminiferal Assemblages from the Western Mediterranean and Adjacent Areas. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 91-120. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: We investigated the biostratigraphic and paleoenvironmental distribution of Upper Cretaceous deep-water agglutinated benthic foraminiferal assemblages from 10 selected areas in the Western Mediterranean and adjacent areas in the North Atlantic and in the Alpine/Carpathian foldbelt. This distribution pattern, compared with paleoenvironmental data (e. g. paleobathymetry, oxygenation of the bottom waters, amount of terrigenous input and substrate disturbance) allows us to distinguish 12 general types of assemblages in which deep-water agglutinated taxa occur.

Assemblages consisting purely of agglutinated foraminifers

1. Flysch-type, high diversity (*Paratrochamminoides* Faunas)
2. Flysch-type, low diversity (*"Rhabdammina"* Faunas)
3. Low-latitude slope assemblages (*Rzehakina* Faunas)
4. "Low-oxygen" assemblages (*Glomospirella* Faunas)
5. High-latitude slope assemblages (*Glomospira* Faunas)
6. Abyssal assemblages ("Krasheninnikow"-type Faunas)

Mixed assemblages with a minor planktonic component

7. Mixed assemblages of slope basins
8. Mixed assemblages of the lower slope
9. Abyssal mixed assemblages

Plankton-dominated deep-water assemblages

10. "Scaglia-type" assemblages of Western Mediterranean deep-water limestones
11. Plankton-dominated assemblages of slope marls
12. Plankton-dominated assemblages in hemipelagic layers of turbiditic sequences

Low latitude slope faunas, present on the North African margins and in the Betic Cordillera of Southern Spain contain common calcareous ataxophragmids. In contrast, high latitude slope assemblages, recorded on the Labrador Margin, lack calcareous elements and are dominated by ammodiscids and lituolids. Flysch type (Type A) assemblages lack calcareous elements. This type includes assemblages dominated by coarsely agglutinated forms in proximal turbiditic environments and highly diversified assemblages in distal environments. Deep-water limestone assemblages include flysch-type forms, but also some slope forms and small smooth-walled agglutinates which can be compared to abyssal taxa. Abyssal multicolored or red clays of the North Atlantic, Pacific and Indian Ocean were populated by thin, smooth-walled varieties and are taxonomically distinct from others. Similar assemblages have been observed in the Rumanian Eastern Carpathians.

Biostratigraphic turnover in the taxonomic content of assemblages is observed in the Lower Turonian, mid-Campanian and in the late Maastrichtian to early Paleocene. These datum levels correspond to inter-regional and time-constant paleoceanographic events, which probably also affected the deep-water benthic biota. This allows us to use deep-water agglutinated foraminifers (DWAF) for biostratigraphy in the Western Mediterranean and to extend the geographic utility of currently used zonal schemes which have been established in the Carpathian and Alpine areas.

Kurzfassung: Biostratigraphie und Environment-Abhängigkeit sandschaliger benthischer Tiefwasser-Foraminiferen der Oberkreide wurden in 10 ausgewählten Gebieten des Westmediterranraumes sowie angrenzender alpiner und atlantischer Tiefseebereiche untersucht. Hierbei wurden die in den einzelnen Untersuchungsgebieten unterschiedlichen Paläoenvironment-Bedingungen (Wassertiefe, Sauerstoffverhältnisse am Ozeanboden, detritischer Eintrag und Beeinträchtigung des Substrats durch Strömungs- und Sedimentationsprozesse) zu der jeweiligen taxonomischen Zusammensetzung der agglutinierenden Benthos-Fauna in Beziehung gesetzt. Für den Zeitraum vom Turon bis zum Maastricht ließen sich zwölf charakteristische Vergesellschaftungen agglutinierender Tiefwasser-Foraminiferen unterscheiden.

Ausschließlich aus agglutinierenden Formen bestehende Faunen:

1. Flysch-Faunen hoher Diversität (*Paratrochamminoides*-Faunen)
2. Flysch-Faunen niedriger Diversität ("*Rhabdammina*"-Faunen)
3. Slope-Faunen niederer Breiten (*Rzehakina*-Faunen)
4. Vergesellschaftungen ungünstiger Sauerstoffverhältnisse (*Glomospirella*-Faunen)
5. Slope-Faunen hoher Breiten (*Glomospira*-Faunen)
6. Abyssale Faunen ("Krasheninnikov-Faunen")

Gemischt kalkschalige und agglutinierende Benthosfaunen mit geringem Planktonanteil:

7. Gemischte Faunen in Kontinentalrand-Becken
8. Gemischte Faunen des tieferen Kontinentalhanges
9. Abyssale gemischte Faunen

Plankton-dominierte Tiefwasser-Vergesellschaftungen:

10. "Scaglia-Faunen" der rötlichen pelagischen Oberkreidekalke des Westmediterran-Raumes
11. Plankton-dominierte Faunen in Kontinentalhang-Mergeln und Kalk/Mergel-Rhythmiten
12. Plankton-dominierte Faunen in hemipelagischen Abschnitten von Turbiditserien

Kontinentalhang-Vergesellschaftungen niedriger Breiten, die in den Kontinentalhang-Sedimenten Nordafrikas und des südspanischen Betikums auftreten, führen (im Gegensatz zu entsprechenden Faunen in hohen Breiten) reichlich kalkig agglutinierende Ataxophragmiden. Flysch-Vergesellschaftungen der westmediterranen Flysch-Einheiten enthalten keine kalkschalige Formen; sie werden von charakteristischen, grobkörnig agglutinierenden Formen in proximalen turbiditischen Environments dominiert, während in distalen turbiditischen Environments hochdiverse Vergesellschaftungen beobachtet wurden. Die Vergesellschaftungen der pelagischen Scaglia-Fazies enthalten Elemente der Flysch-Vergesellschaftungen, daneben aber auch Kontinentalhang-Formen und winzige glattschalige Formen, die Beziehungen zu distalen abyssalen Faunen aufweisen. Derartige abyssale Faunen spezieller taxonomischer Zusammensetzung wurden sonst nur in den bunten Tiefseetonen des Nordatlantik, Pazifik und Indik und in roten Tiefseetonen der rumänischen Ostkarpathen beobachtet.

Die biostratigraphische Gliederung der Tiefseesedimente des Westmediterranraumes mit Hilfe dieser Faunengruppe läßt sich weitgehend mit den Zonierungen, die in den Flyschzonen der Karpathen entwickelt wurden, korrelieren. Wichtige Faunenschnitte können im basalen Turon, im mittleren Campan und an der Kreide/Tertiär-Grenze beobachtet werden. Diese Zeitstufen sind durch überregionale, zeitgleiche paläoozeanographische Events charakterisiert, die wahrscheinlich auch die Evolution des Tiefsee-Benthos beeinflußt haben. Die Korrelation der Entwicklung agglutinierender Tiefsee-Foraminiferen mit diesen globalen, zeitkonstanten Events läßt die biostratigraphische Brauchbarkeit dieser Faunengruppe auch für überregionale Korrelation in einem neuen Licht erscheinen.

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1. Introduction

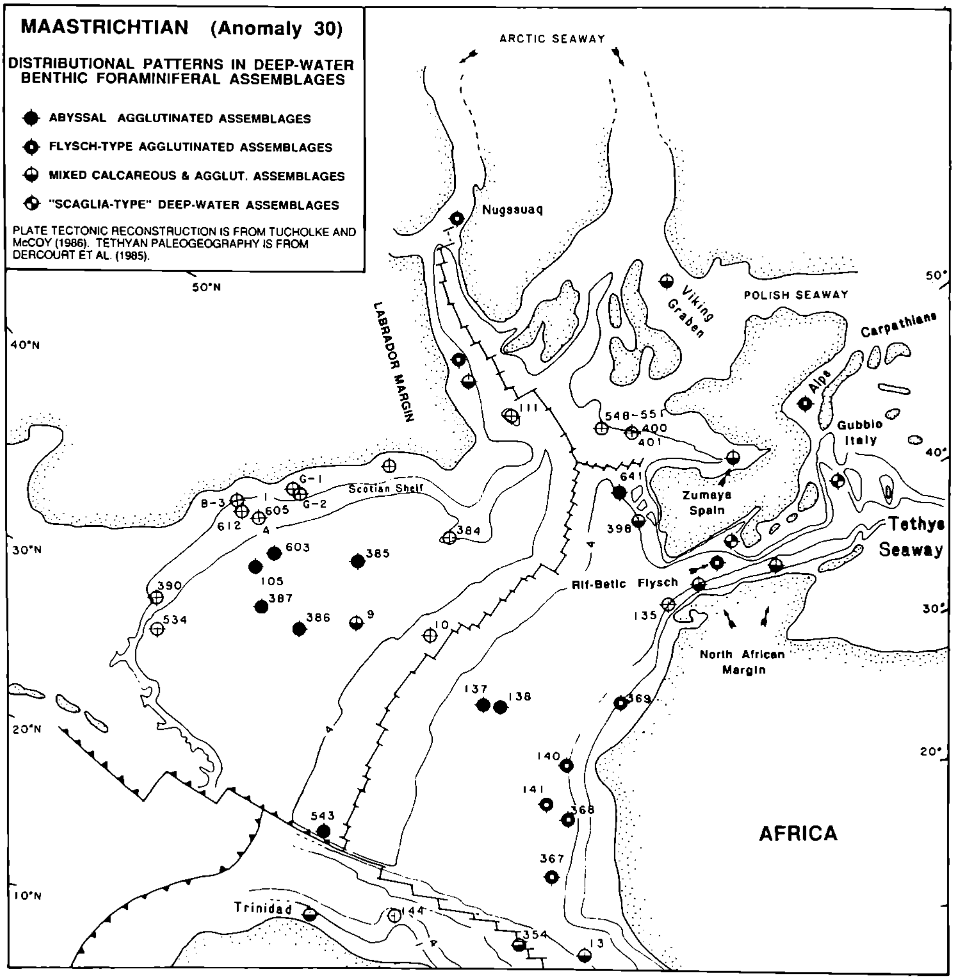
The sample base for this study is material from 10 localities in various depositional and paleogeographic settings in the Western Mediterranean.

1-2. Subbetic and Penibetic zones of the southern Iberian paleomargin (southern Spain). The paleoenvironment of these zones is characterized by bathyal water depths well above the CCD, little fine-grained clastic input, and well oxygenated bottom waters. The typical sediments are red pelagic marls and limestones.

3-6. North African paleomargin (Tellian Units in Northern Morocco and Algeria): We have compared assemblages derived from the upper slope, continental margin basin and deep bathyal to abyssal zones at the distal part of the margin. Here oxygen-deficient bottom-water conditions prevail which can probably be attributed to a high input of fine-grained terrigenous sediments.

7. Almarchal Unit of the Campo de Gibraltar flysch domain (Southern Spain). An Upper Cretaceous fine-grained calciturbidite sequence was deposited in this paleogeographic zone within a deep bathyal environment.

8-9. Rif-Betic Flysch (remnants of the Gibraltar-Tethyan seaway in Southern Spain and Northern Morocco): Deep bathyal to abyssal water depths (deposition below the CCD), oxygen levels at the seafloor and detrital overprint varied, depending on the relative position of the sections in



Text-Fig. 1. Paleogeographic map of the late Cretaceous North Atlantic and Western Tethys revealing the distribution of deep-water benthic foraminiferal assemblages.

relation to local fan systems. Two principally different paleogeographic domains can be distinguished: the Mauretanian and Numidian deep-water clastic fan systems and the Massylian distal pelagic deep-water realm.

10. Italy (Gubbio section, Umbrian Apennines): Bathyal water depths well above the CCD, sediment-starved, well-oxygenated; the typical sediments are red pelagic "Scaglia"-limestones.

The ranges and relative abundances of characteristic taxa have been compared to assemblages from various localities around the North Atlantic and in the Alpine-Carpathian mountain belt (Text-Fig. 1), including the Zumaya Section of Northern Spain, the high latitude locality of the Indian Harbour well on the Labrador Margin, slope environments in Trinidad, flysch deposits in the Romanian Eastern Carpathians, and the abyssal DSDP/ODP Sites 137, 141, 367, 368, 543, 603, and 641 in the North Atlantic.

2. Characteristic Deep-Water Benthic Foraminiferal Assemblages

Twelve different deep-water agglutinated benthic foraminiferal (DWAF) assemblages of Upper Cretaceous age from the Western Mediterranean and North Atlantic have been distinguished by comparing the environmental data and taxonomical composition of agglutinated assemblages. Main environmental features of these assemblages are as follows. The complete taxonomic composition of some examples of these assemblages is given in Text-Fig. 2.

Assemblages consisting purely of agglutinated foraminifers

1. Flysch type, high diversity (*Paratrochamminoides* Faunas)

General features:	Low faunal density, comparatively high diversity
Characteristic taxa:	*Paratrochamminoides* spp. *Rhizammina* *Uvigerinammina jankoi* (Lower Campanian and older) *Subreophax scalaris* *Hormosina excelsa* (small variety) *Hormosina ovulum*
Sediment:	Red carbonate-free claystones between coarse-grained calciturbidites and debris-flows, well oxygenated bottom water conditions
Examples:	Flysch Units of the Moroccan Rif: Turonian-Santonian argillites of the Massylian Units, Turonian-Maastrichtian of the Mauretanian Units, Maastrichtian of the Numidian Talaa Lakrah Unit, DSDP Sites 367, 368, 141

2. Flysch type, low diversity ("*Rhabdammina*" Faunas)

General features:	Assemblages mainly consisting of large, coarsely agglutinated forms, faunal density can be high, diversity is low, single species often dominate
Characteristic taxa:	*Dendrophrya excelsa* *Rhizammina* *Paratrochamminoides*

Aschemonella carpathica
Hormosina gigantea
Reophax duplex

Sediment: Gray-green bioturbated carbonate-free claystones intercalated in a succession of fine-grained calciturbidites and mud-turbidites, less oxygenated bottom water conditions

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Text-Fig. 2. Biogeographic and stratigraphic distribution of Upper Cretaceous benthic agglutinated deep-water foraminiferal species in the Western Mediterranean and adjacent areas. The following localities have been included into this study:

1. Sierra de Las Cabras section in the Western Subbetic Zone of the southern Iberian paleomargin (Southern Spain). Mixed calcareous and agglutinated assemblages from pelagic marls with high plankton/benthos ratio.
2. Hacho de Montejaque section (Penibetic Zone, Southern Spain). Turonian-Campanian "Scaglia-type" assemblages from HCl-residues and Campanian/Maastrichtian mixed assemblages from washing residues.
3. Prerif Zone (Northern Morocco), which formed in the late Mesozoic the upper part of the North African paleomargin. Mixed assemblages from pelagic marls with high plankton/benthos ratio.
4. Mesorif Zone (Northern Morocco), Maastrichtian of the M83 section. Mixed assemblages from pelagic marls with low plankton/benthos ratio.
5. Loukkos Zone (Northern Morocco), Campanian-Maastrichtian of the Souk el Had area. Exclusively agglutinated assemblages from greenish marls and claystones.
6. Tanger Unit (Northern Morocco), Campanian-Maastrichtian of the Tangier area (hemipelagic marls and greenish claystones, fine-grained calciturbidite sequences). Agglutinated assemblages (*Rzehakina* and "*Rhabdammina*" faunas).
7. Almarchal Unit of the Campo de Gibraltar flysch domain (Southern Spain). Campanian/Maastrichtian agglutinated assemblages of a fine-grained calciturbidite sequence in a deep bathyal environment.
8. Distal Tellian and Massylian Upper Cretaceous deep-water (abyssal) argillites of the Rif-Betic Seaway (Tangier area).
9. Deep-sea clays intercalated in Mauretanian and Numidian deep-water turbiditic sequences (Campanian/Maastrichtian of the Beni Ider and Talaa Lakrah units, Northern Morocco).
10. Gubbio section, Umbrian Apennines (Italy). HCl-residues from red pelagic "Scaglia"-limestones, deposited in a sediment-starved, well-oxygenated, bathyal environment well above the CCD: "Scaglia-type" agglutinated assemblages.

Assemblages from various localities around the North Atlantic and from the Alpine-Carpathian mountain belt are added for comparison: The Zumaya section of Northern Spain (11), a high latitude locality of the Indian Harbour well on the Labrador Margin (12), slope environments in Trinidad (13), flysch deposits in the Romanian Eastern Carpathians (14), DSDP Sites 141, 367, 368 with flysch-type assemblages (15), DSDP Site 543 with abyssal mixed assemblages (16), and DSDP/ODP Sites 137, 603, and 641 with abyssal agglutinated assemblages (17).

LOCALITY						1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
AGE	T	C	S	C	M	Ca-M	T-Ca	Ca-M	Ca-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	T-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	Ca-M	T-M
Ammobaculites agglutinans					---												X					
Ammobaculites aubertae					---												X					
Ammobaculites jarvisi					---				X							X	X	X				
Ammobaculites sp.3					---											X	X	X				
Ammobaculites sp.4					---												X					
Ammodiscus asperellus					---																X	X
Ammodiscus cretaceus	---	---	---	---	---		X		X	X	X	X	X	X	X	X	X	X	X	X	X	X
Ammodiscus glabratus					---		X								X	X	X	X		X		
Ammodiscus infimus	---	---	---					X											X			
Ammodiscus pennyi				---	---				X						X			X		X		X
Ammodiscus pennyi cf.				---	---		X													X		
Ammodiscus peruvianus				---	---				X		X	X				X	X	X		X		
Ammodiscus planus				---	---	X	X		X				X		X		X	X		X		
Ammodiscus sp.1				---	---										X					X		
Ammolagena clavata					---												X	X				
Ammosphaer. pseudopauciloculata		---	---	---	---		X		X						X	X	X	X	X	X		
Arenobulimina dorbignyi				---	---											X	X	X			X	
Aschemonella carpathica				---	---							X			X	X			X	X		
Aschemonella ex gr. grandis				---	---		X		X						X		X	X		X		
Bathysiphon spp.	---	---	---	---	---				X		X	X	X	X		X	X	X	X	X		
Bolivinopsis parvissimus			---	---	---																X	X
Budashevaella trinitatensis				---	---												X	X		X		
Clavulinoides aspera					---													X			X	
Clavulinoides eggeri				---	---		X									X						
Clavulinoides subparisiensis				---	---	X	X	X								X			X		X	
Cribrostomoides sp. 1				---	---		X								X							
Cribrostomoides trinitatensis					-		X		X							X	X	X	X	X		
Dendrophrya ex gr. excelsa	---	---	---	---	---				X	X	X	X	X	X		X		X	X	X		
Dendrophrya latissima				---	---							X						X		X		
Dorothia crassa trochoides				---	---	X		X	X			X				X		X	X		X	
Dorothia oxycona	---	---	---	---	---	X	X	X	X							X	X	X	X	X	X	
Dorothia retusa				---	---		X	X	X							X	X	X				
Dorothia sp.1 (coarse)					---												X					
Gaudryina ex gr. cretacea				---	---		X		X	X						X	X	X	X			
Gaudryina pyramidata				---	---	X	X	X	X							X		X	X		X	
Gaudryina sp. 1				---	---												X					
Glomospira charoides	---	---	---	---	---		X		X	X		X	X	X	X	X	X	X		X	X	X
Glomospira diffundens					---				X								X	X	X	X	X	
Glomospira gordialis	---	---	---	---	---		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Glomospira irregularis	---	---	---	---	---	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Glomospira serpens			---	---	---				X	X	X	X			X	X		X	X		X	
Glomospirella gaultina	---	---	---	---	---		X		X	X	X	X	X	X	X	X	X			X	X	X
Glomospirella grzybowskii					---				X							X				X	X	

LOCALITY						1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
AGE	T	C	S	C	M	Ca-M	T-Ca	Ca-M	Ca-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	T-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	Ca-M	T-M
Goesella rugosa				---	---	X	X		X							X			X			
Haplophragmium problematicus	---	---	-				X								X				X			X
Haplophragmoides bulloides	---	---	---																X			X
Haplophragmoides concavus cf.	---	---	---	---	---		X		X		X		X									X
Haplophragmoides eggeri					---				X										X			
Haplophragmoides fraudulentus		---	---	---	---																X	X
Haplophragmoides glabra cf.				---	---												X	X				
Haplophragmoides herbichi	---	---	-																X			
Haplophragmoides horridus				---	---													X		X		
Haplophragmoides kirki					---												X					
Haplophragmoides linki cf.		---	---																			X
Haplophragmoides menitens		---	---	---	---																X	X
Haplophragmoides multicamerus		---	---	---	---																	X
Haplophragmoides multiformis		---	---	---	---																	X
Haplophragmoides perexplicatus s.l.		---	---	---	---																X	X
Haplophragmoides pseudokirki				---	---																	X
Haplophragmoides retroseptus				---	---				X								X	X		X		
Haplophragmoides sp.1		---	---	---	---		X								X							X
Haplophragm. suborbicularis ex gr.				---	---										X		X	X				
Haplophragmoides walteri cf.		---	---	---	---		X		X	X	X	X	X	X	X	X	X	X			X	X
Hormosina crassa	---	---	---	---	---		X				X									X	X	X
Hormosina excelsa		-	-	-	---								X	X			X		X	X	X	
Hormosina gigantea				---	---		X		X	X	X	X					X		X	X	X	X
Hormosina ovuloides	-	-	-	---	---		X	X	X	X	X	X						X		X	X	X
Hormosina ovulum	-	-	-	---	---	X	X		X	X	X	X		X		X	X	X	X	X	X	X
Hormosina trinitatensis				---	---													X				
Hormosina velascoensis				---	---		X		X		X	X			X	X	X	X	X	X	X	
Hormosinella distans		---	---	---	---		X								X							X
Hormosinella sp. 141				-	-															X		
Hyperammina dilatata	---	---	---	---	---		X	X	X	X	X	X	X			X	X	X		X		X
Hyperammina elongata	---	---	---	---	---		X		X	X			X		X		X	X	X	X	X	X
Hyperammina subdiscreta	---	---	---	---	---		X								X					X	X	X
Kalamopsis dubia				---	---				X										X	X		
Kalamopsis grzybowskii			-	---	---		X		X					X	X	X	X	X	X	X	X	X
Karreriella conversa	---	---	---	---	---		X		X	X			X	X	X		X	X	X	X	X	X
Karreriella horrida					---									X			X	X		X		X
Labrospira inflata		---	---	---	---																X	X
Labrospira pacifica		---	---	---	---													X			X	X
Labrospira sp.1					---												X					
Lagenammina sp.1				---	---													X		X		
Lituotuba lituiformis	-	-	---	---	---		X		X				X		X	X	X	X		X		
Matanzia varians					---				X						X	X	X	X	X			
Paratrochamminoides acervulatus				---	---				X			X		X			X			X	X	X

LOCALITY						1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
AGE	T	C	S	C	M	Ca-M	T-Ca	Ca-M	Ca-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	T-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	Ca-M	T-M
Paratrochammin. heteromorphus				---	---		X	X				X		X	X	X			X			
Paratrochamminoides intricatus s.l.	---	---	---	---	---																X	X
Paratrochamminoides irregularis				---	---				X		X	X			X		X	X	X	X		
Paratrochammin. semipellucidus s.l.	---	---	---	---	---																X	X
Paratrochamminoides sp.1		---	---	---	---		X								X							
Paratrochamminoides sp.2		---	---	---	---		X								X					X		
Paratrochamminoides sp.3			---	---	---		X													X		
Paratrochamminoides spp.	---	---	---	---	---		X	X			X	X	X	X	X	X			X	X	X	X
Phenacophragma elegans					-													X				
Plectorecurvoides/Recurvoides spp.	---	---	---	---	---	X	X	X	X		X	X	X	X	X	X	X	X	X	X	X	X
Plectorecurvoides parvus				-	-																X	
Plectorecurvoides rotundus				-	-																X	
Praecystammina globigerinaeformis		---	---	---	---		X								X							X
Praecyst.(?) cf. globigerinaeformis		---	---	---	---		X								X					X	X	X
Psammosphaera fusca	-	-	-	---	---						X	X	X	X			X		X	X		
Psammosphaera scruposa	-	-	-	---	---				X		X	X		X		X	X	X			X	
Pseudobolivina cuneata		---	---	---	---																X	X
Pseudobolivina lagenaria		---	---	---	---		X								X					X		X
Pseudobolivina munda		---	---	---	---		X								X					X	X	X
Pseudobolivina sp.1	---	---																				X
Pseudobolivina sp.2	---	---																		X		X
Pseudobolivina sp.3				---	---		X								X							
Pseudobolivina sp.4				---	---		X													X		
Pseudobolivina spp.					---											X	X					
Recurvoides anormis					---				X									X		X		
Recurvoides deflexiformis					---				X								X	X		X		
Recurvoides gerochi					---												X	X		X		X
Recurvoides subturbinatus cf.				---	---													X		X		X
Recurvoides walteri				---	---				X								X			X		X
Reophax aff. dentaliniformis	---	---	---	-	-																	X
Reophax duplex				---	---				X		X	X	X			X	X	X	X	X		
Reophax globosus					---												X	X				
Reophax pilulifer				---	---				X		X	X					X		X	X		
Reophax sp.2		---	---	---	---		X								X		X	X		X		
Reophax sp.3				---	---										X							
Reophax sp.4			---				X															
Reophax sp.5					---										X							
Reophax subfusiformis					---												X	X				
Reophax subnodulosus cf.			-	-	-		X		X						X							
Rhabdammina spp.	---	---	---	---	---	X	X	X	X		X	X	X	X	X	X	X	X	X	X	X	X
Rhizammina algaeformis cf.	-	-	-	-	-		X								X					X	X	X
Rhizammina grzybowskii					---													X				
Rhizammina indivisa	---	---	---	---	---	X	X	X	X			X	X	X	X	X	X	X		X	X	X

LOCALITY						1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
AGE	T	C	S	C	M	Ca-M	T-Ca	Ca-M	Ca-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	T-M	Ca-M	Ca-M	Ca-M	T-Ca	Ca-M	Ca-M	T-M
Rzehakina epigona				-	---		X		X	X	X	X			X	X	X	X	X	X		
Rzehakina fissistomata					-				X		X								X	X		
Rzehakina inclusa				---	---				X		X	X			X				X	X		
Rzehakina minima				-	-													X				
Saccammina grzybowskii	---	---	---	---	---		X	X	X	X	X	X	X	X	X	X	X	X	X	X	X	X
Saccammina placenta				-	---		X		X			X		X	X	X	X	X				
Saccammina placenta cf.	---	---	---	---	---		X	X				X		X	X					X		
Saccammina sphaerica			---	---	---		X								X						X	X
Saccorhiza ramosa cf.				---	---										X							
Silicosigmoilina perplexa				-	-															X		
Spaerammina gerochi					---													X		X		
Spiroplectammina dentata aff.		-	-	---	---	X	X		X	X	X				X	X	X	X	X		X	
Spiroplectammina israelskyi				---	---		X		X						X							
Spiroplectammina israelskyi cf.				---												X					X	
Spiroplectammina laevis	-	-	-	---	---		X		X						X						X	
Spiroplectammina navarroana					---				X								X	X				
Spiroplectammina spectabilis aff.					---				X						X		X	X				
Spiroplectammina subhaeringensis			-	---	-	X		X				X							X		X	
Spiroplectinata (?) sp.1				-											X							
Subreophax guttifer	---	---	---	---	---		X								X	X						X
Subreophax pseudoscalaris					-											X	X	X		X		
Subreophax scalaris	---	---	---	---	---	X	X	X	X	X		X	X	X	X	X	X	X		X	X	X
Subreophax sp.1					---										X					X		
Subreophax splendidus				---	---							X			X	X	X		X	X		
Thurammina sp.		---	---	---	---							X	X					X		X		
Tolypammina sp. 1		---	---	---	---		X								X							
Tolypammina sp. 2			---	---	---		X								X							
Tolypammina sp. 3					-										X							
Trochammina altiformis	---	---	---	---	---												X	X		X		X
Trochammina bulloidiformis				---	---														X	X		X
Trochammina deformis				---	---		X					X			X	X	X			X		
Trochamm. globigeriniformis ex gr.				---	---								X			X	X			X	X	
Trochammina gyroidinaeformis		---	---	---	---																X	X
Trochammina sp.1 (coarse)					---				X													
Trochammina spp.	---	---	---	---	---	X		X	X		X	X	X	X	X		X	X		X		
Trochamminoides dubius					---	X			X	X		X			X	X		X	X	X		
Trochamminoides dubius cf.				---	---		X								X					X		
Trochamminoides proteus				---	---				X			X				X		X	X	X		
Trochamminoides proteus cf.			---	---	---		X								X	X				X		
Trochamminoides subcoronatus					---												X	X		X		
Turritellella sp.					-										X						X	
Uvigerinammina jankoi	---	---	---	-			X						X		X		X		X	X	X	X
Verneuilina cretacea				---	---		X									X					X	
Verneuilinoides polystrophus					---		X								X			X				

Examples:	Campanian-Maastrichtian of the Almarchal Unit (Campo de Gibraltar), Campanian-Maastrichtian of the Tanger Unit (Rif, Morocco)

3. Slope type (*Rzehakina* Faunas, "low oxygen" assemblages)

General features:	Low diversity, species with a finely agglutinated, silicified wall dominate
Characteristic taxa:	*Rzehakina epigona* *Glomospirella gaultina* *Recurvoides* *Paratrochamminoides irregularis* *Glomospira serpens* *Hormosina velascoensis*
Sediment:	Green-black laminated or bioturbated claystones, poorly oxygenated bottom water conditions
Examples:	Campanian-Maastrichtian of the Tellian Bab Taza and Loukkos Units (Rif, Morocco)

4. High latitude slope assemblages

General features:	Low to medium diversity, flysch-type agglutinated taxa dominate
Characteristic taxa:	*Glomospira charoides* *Cribrostomoides trinitatensis* *Bathysiphon* *Rhabdammina* *Recurvoides walteri* *Hormosina ovulum* *Karreriella horrida* *Uvigerinammina jankoi*
Sediment:	Greenish gray claystones, slightly oxygen-deficient bottom-water conditions
Examples:	Campanian-Paleocene of the Indian Harbour well (Labrador Margin)

5. Abyssal assemblages under well-oxygenated bottom water conditions ("Krasheninnikov-type")

General features:	High diverse agglutinated fauna, consisting of tiny, finely agglutinated species. These assemblages correspond to the abyssal agglutinated faunas first described from the Pacific and Indian Ocean (KRASHENINNIKOV 1973, 1974)
Characteristic taxa:	*Rhizammina* *Uvigerinammina jankoi* *Hormosina gigantea* *Haplophragmium problematicum* *Pseudobolivina munda* and *lagenaria* *Karreriella conversa* *Praecystammina globigerinaeformis* *Haplophragmoides fraudulentus, multiformis, menitens* *Labrospira inflata, pacifica*
Sediment:	Brown, varicolored zeolitic claystones, well oxygenated bottom-water conditions, slow pelagic sedimentation

Examples:	Turonian-Maastrichtian of DSDP/ODP Sites 137, 603, 641

6. Impoverished abyssal assemblages under oxygen-deficient bottom water conditions (biofacies B, *Glomospirella* Faunas)

General features:	A low-diversity agglutinated assemblage, consisting of often poorly preserved "primitive" tests, commonly associated with rich radiolarian assemblages
Characteristic taxa:	*Glomospirella gaultina* *Glomospira gordialis, irregularis* *Ammodiscus* *Rhizammina* *Haplophragmoides concavus* *Hormosina*
Sediment:	Greenish to dark-grey zeolitic claystones, poorly oxygenated bottom water conditions
Examples:	Lower Turonian and Lower/Middle Campanian of DSDP Site 603, Lower Turonian of ODP Site 641, Paleocene of DSDP Site 543

Mixed assemblages with a minor planktonic component

7. Mixed assemblages of slope basins

General features:	Highly diverse assemblages with large numbers of calcareous benthic foraminifers and calcareous agglutinating ataxophragmids
Characteristic taxa:	*Matanzia varians* *Recurvoides walteri, anormis* *Rzehakina epigona, inclusa* *Dorothia oxycona, retusa, crassa* *Haplophragmoides retroseptus* *Spiroplectammina dentata* *Goesella rugosa*
Sediment:	Grey-greenish marls and claystones, less oxygenated bottom water conditions
Examples:	Campanian-Paleocene of the Mesorif Zone (Rif, Morocco), Campanian-Paleocene of Trinidad, Paleocene of Zumaya (Spain)

8. Mixed assemblages of the lower slope

General features:	Agglutinated assemblages with only small numbers of calcareous benthic foraminifers and calcareous agglutinating ataxophragmids
Characteristic taxa:	*Dendrophrya excelsa* *Rhizammina* *Recurvoides* *Paratrochamminoides* *Rzehakina epigona* *Aschemonella carpathica*

Sediment: Light- and dark-gray "hemipelagites" consisting of fine-grained calciturbidites, mud-turbidites, and strongly bioturbated autochthonous layers, less oxygenated bottom-water conditions

Examples: Campanian-Maastrichtian of the Tellian Bab Taza and Tanger units (Rif, Morocco), Campanian of the Almarchal Unit (Campo de Gibraltar)

9. Abyssal mixed assemblages

General features: Assemblages with large numbers of calcareous benthic foraminifers and calcareous agglutinating ataxophragmids occur in addition to a highly diverse agglutinated fauna, consisting of tiny, finely agglutinated species. Deposition close to the CCD

Characteristic taxa: brown clay assemblages:
Rhizammina
Haplophragmoides perexplicatus, menitens, molestus
Subreophax scalaris
Bolivinopsis parvissimus
Labrospira inflata, pacifica
calcareous agglutinated assemblages:
Spiroplectammina subhaeringensis
Verneuilina cretacea
Tritaxia aspera
Dorothia crassa
Gaudryina pyramidata
Arenobulimina orbignyi

Sediment: Brown zeolitic clays and light gray marly intercalations, well oxygenated bottom water conditions

Examples: Campanian-Maastrichtian of DSDP Site 543

Plankton-dominated deep-water assemblages

10. "Scaglia-type" assemblages of Western Mediterranean deep-water limestones

General features: Well diversified assemblages with large numbers of rhizamminids and tiny, finely agglutinated species

Characteristic taxa: *Rhizammina*
Uvigerinammina jankoi
Subreophax scalaris
Tolypammina, and *Komoki*-like forms
Karreriella conversa
Trochamminoides
Spiroplectammina
Matanzia varians
Aschemonella carpathica
Ammosphaeroidina pseudopauciloculata
Praecystammina globigerinaeformis
Pseudobolivina cf. *munda, lagenaria*
Haplophragmium problematicus

Sediment:	White, yellow, rose colored to reddish biomicritic limestones and marly limestones, well oxygenated bottom water conditions
Examples:	Turonian-Campanian of the Penibetic Zone (Southern Spain), Turonian-earliest Paleocene of the Gubbio sequence (Umbrian Apennines, Italy)

11. Plankton-dominated assemblages of slope marls

General features:	Highly diverse assemblages with large numbers of calcareous benthic foraminifers and calcareous agglutinating ataxophragmids.
Characteristic taxa:	*Dorothia oxycona, crassa, retusa* *Gaudryina pyramidata* *Spiroplectammina dentata* *Paratrochamminoides* *Glomospira gordialis, irregularis*
Sediment:	Gray marls or marl-limestone couples, moderately oxygenated bottom water conditions
Examples:	Campanian-Maastrichtian of the Prerif Zone (Northern Morocco), Campanian-Maastrichtian of the Subbetic and Penibetic Zones (Southern Spain)

12. Plankton-dominated assemblages in hemipelagic layers of turbiditic sequences

General features:	Diverse assemblages with large numbers of calcareous benthic foraminifers and calcareous agglutinating ataxophragmids
Characteristic taxa:	*Dendrophrya excelsa* *Spiroplectammina dentata, subhaeringensis* *Recurvoides* *Dorothia oxycona, crassa* *Goesella rugosa*
Sediment:	Greenish or reddish calcareous marls, well oxygenated bottom water conditions
Examples:	Campanian-Maastrichtian of the Zumaya section (Northern Spain)

3. Paleoecology and Paleoceanography

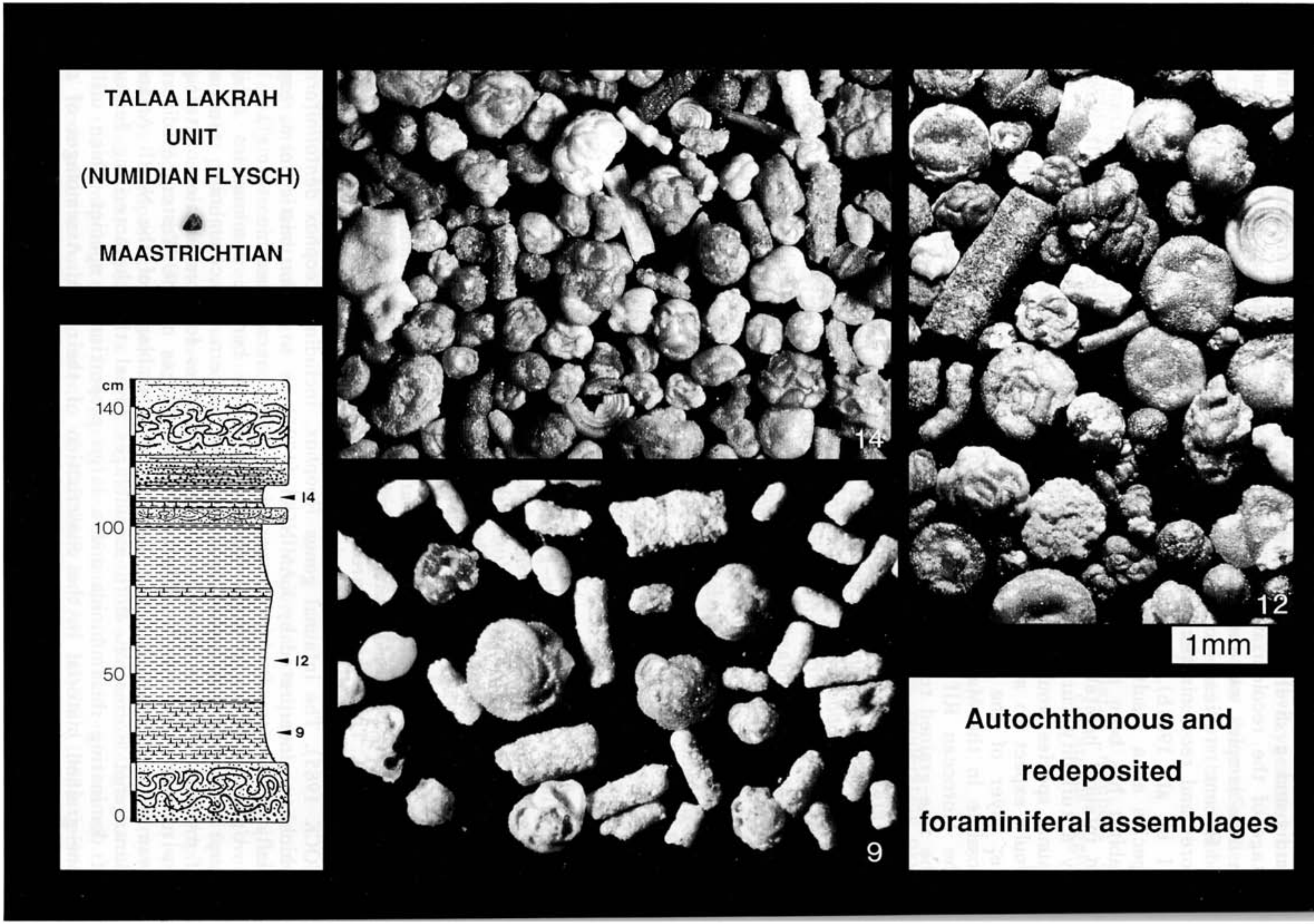
3.1 Detrital Input and Community Structure

Detrital input and substrate disturbance are important sedimentary factors affecting the community structure of deep-water agglutinated foraminiferal faunas. Successive recolonization of the sea floor after a turbiditic event has been postulated as the cause of small-scale vertical changes in foraminiferal assemblages in hemipelagic sediments above turbidites in Alpine flysch deposits (GRÜN et al. 1964, BUTT 1981). Vertical changes in assemblage composition from an astrorhizid-dominated assemblage directly above the coarse layer of a turbidite to a more diverse assemblage further up in the hemipelagite was interpreted as evidence of recolonization of the sea floor after a turbiditic event. VERDENIUS & VAN HINTE (1983) defined in the Norwegian-Greenland Sea a "frontier-area subfauna" of primitive forms and a diverse "mature subfauna" which was interpreted as a later stage of the recolonization. Dominance of frontier faunas or mature faunas in bulk samples was consequently attributed to turbidite intensity.

Recent studies, including data from experimental studies using spade cores and recolonization trays (SCHRÖDER 1986, KAMINSKI 1988, KAMINSKI et al. 1988b), have interpreted concentrations of "primitive" tubular species as a result of hydrodynamic sorting. In these studies an epifaunal habitat has been demonstrated for the modern *Dendrophrya*, in contrast to an infaunal habitat of rectilinear species of the genus *Reophax*. If erosion by a turbidity current effects mainly the flocculent surface sediment, epifaunal species would preferably be entrained by downslope currents and one would expect to see concentrations of *Dendrophrya* in the turbiditic Td or Te_t layer of the Bouma Sequence. Corresponding field observations were possible in the Maastrichtian Numidian Flysch of the Talaa Lakrah Unit in the Moroccan Rif (Text-Fig. 3), where redeposited assemblages found in the fine-grained turbiditic Te_t layers, can be distinguished from the autochthonous reddish sub-CCD sedimentation by its carbonate content and its light-gray color. These redeposited assemblages, presumably from a deep-distal slope environment, are made up by tubular agglutinated species of the genus *Rhabdammina* or *Dendrophrya*, accompanied by pelagic planktonic foraminifers (e. g. *Globotruncana falsostuarti*). In contrast, the autochthonous agglutinated assemblages of the red-clay facies are highly diverse and contain only few tubular species (typical *Paratrochamminoides* assemblages).

A generalized model about the effect of substrate disturbance on the structure of modern DWAF communities has been developed by KAMINSKI (1988). In tranquil areas covered by fine-grained pelagic sediment, which provide stable environments for benthic organisms, the agglutinated fauna is dominated by species of Komokiacea and Astrorhizidae. These forms have branching tubular tests and live in the flocculent surface layer, e. g. the suspension-feeding species *Rhizammina algaeformis*. Generally, the agglutinated assemblage is comparatively diverse and contains a large proportion of species which utilize fine-grained material for test construction. Disturbed environments are characterized by a coarse-grained substrate, consisting of coarse detrital quartz and/or carbonate sand and reworked planktonic and benthic foraminiferal tests. The in situ-agglutinated fauna commonly displays lower diversity, and contains a large proportion of species which utilize coarse-grained material in the construction of their tests, reflecting the coarse nature of their substrate. The fauna is dominated by robust, non-branching species of Astrorhizidae and a number of lituolids and trochamminids. These groups are epifaunal, and constitute the morphogroup of surface-dwelling herbivores, detritivores and amnivores (JONES & CHARNOCK 1985). The infaunal genus *Reophax*, including *Reophax dentaliniformis* which was interpreted by KAMINSKI (1985) as an opportunistic form, especially dominates after brief periods of disturbance ("benthic storms").

Upper Cretaceous deep-water agglutinated benthic communities exhibit surprising analogies with this distribution pattern and community structures of modern DWAF assemblages. Typical examples for communities of tranquil environments are the "Scaglia-type" assemblages of the Western Mediterranean pelagic limestones and the abyssal assemblages of the North Atlantic Plantagenet Formation. Both assemblages reveal the characterizing features of dominating rhizamminids and a large proportion of species which utilize fine-grained material in the construction of their test. Assemblages of the


TALAA LAKRAH
UNIT
(NUMIDIAN FLYSCH)
MAASTRICHTIAN
cm
140
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1mm
Autochthonous and
redeposited
foraminiferal assemblages

Western Mediterranean flysch zones exhibit a large variety of benthic agglutinated foraminiferal communities, including typical assemblages for areas with disturbed environments and a rapid succession of turbidites, and benthic storms. The distribution pattern of these benthic foraminiferal communities in time and space can provide valuable information about regional tectonic effects, such as the relationship between depocenters and continental breakup or collision.

3.2 Bottom-Water Anoxia and Paleoceanographic Events

The oxygenation of bottom water appears to be an important factor for the composition of deep-water agglutinated foraminiferal assemblages. During the mid- to late Cretaceous and Paleocene the world oceans underwent a period characterized by several periods of oxygen depletion in the deep oceanic realm. These periods of oxygen depletion (or oceanic anoxic events of authors) are characterized by benthic-free zones or distinct "low-oxygen" benthic foraminiferal assemblages and appear to coincide with significant faunal changes also observed in the deep-water agglutinated foraminifers. Three major global paleoceanographic events have been distinguished during the Upper Cretaceous:

1. Cenomanian/Turonian Boundary Event (CTBE)

The CTBE is characterized by a predominating biosiliceous sedimentation and shows distinct and marked anoxic facies in the deep-sea (HERBIN et al. 1986, KUHNT et al. 1986, THUROW et al. 1982 and in press). The event is accompanied by important taxonomic changes in deep-water benthic foraminifers (GEROCH & NOWAK 1984, KUHNT 1987, MOULLADE et al. in press). In the deep-water limestones of the Western Mediterranean and in the North Atlantic Plantagenet Formation the CTB is devoid of benthic foraminifers and overlying beds are characterized by rare and low-diverse agglutinated assemblages (mainly indeterminable "tubes" and species of the family Ammodiscidae). The HCl-residues consist almost completely of radiolarians and (secondarily) silicified planktonic foraminifers. In recent oceans the oxygen-minimum zone on continental slopes below high productivity surface waters appears rather barren of in situ foraminiferal fauna (ZOBEL 1973). This may be a possible model for the rare occurrence and low diversity of agglutinants in the beds overlying the CTB, which exhibit important chertification and slightly more reducing environments than the sequence upsection.

The re-occurrence of benthic foraminifers occurred in the early to Middle Turonian, and diversified benthic assemblages can be observed beginning in the Middle-Upper Turonian (P. helvetica and M. schneegansi Zones),

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Text-Fig. 3. Redeposition of agglutinated assemblages from deep-distal sources. Maastrichtian of the Talaa Lakrah Unit (Numidian Flysch), Rif, Northern Morocco.

with characteristic *Haplophragmium problematicus - Uvigerinammina jankoi* assemblages.

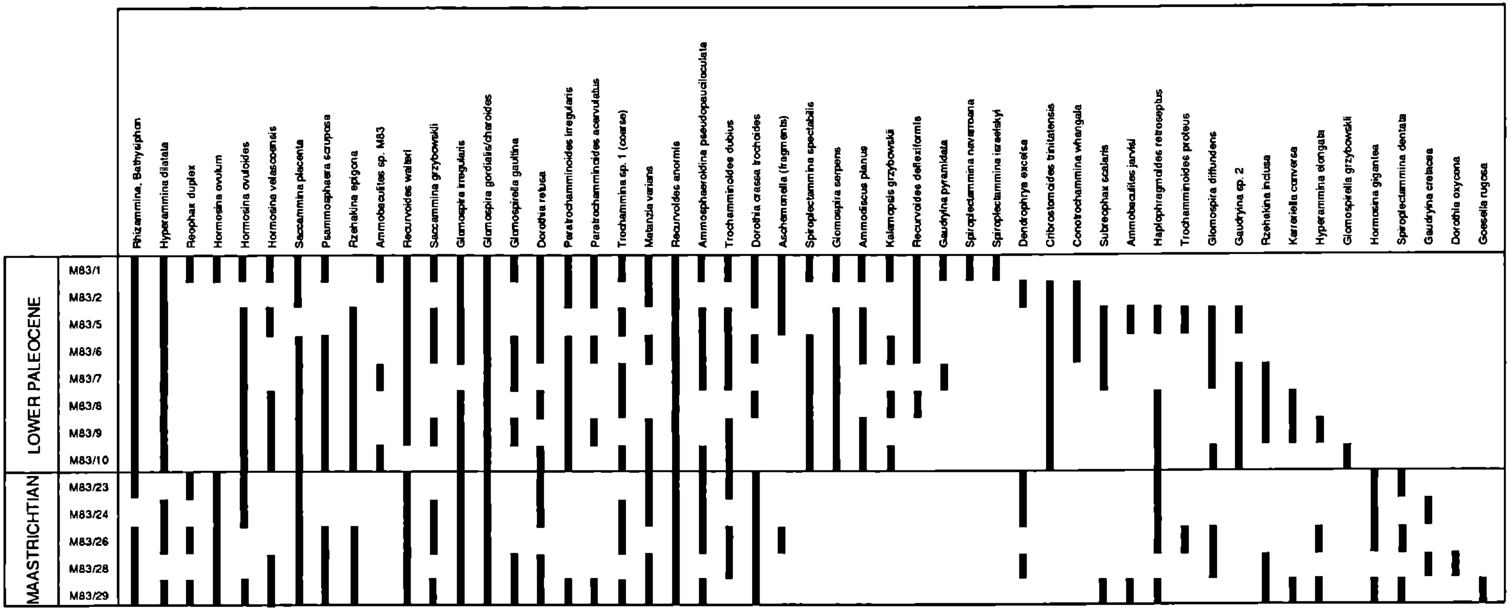
2. A Lower/Middle Campanian Event (LMCE)

This event is characterized by the intercalation of a biosiliceous facies in the Tethyan flysch basins (NEAGU 1968, HERM 1962, BUTT 1981, KUHNT 1987). A radiolarian biofacies with a peculiar taxonomic composition of the radiolarian assemblages prevails across this interval and has also been observed in the zeolitic clays (sub-CCD deposits of the Plantagenet Formation) of the North Atlantic Ocean (MOULLADE et al. in press, THUROW in press). In both cases this biosiliceous event coincides with a major faunal change in agglutinated foraminifers: The *Uvigerinammina jankoi* assemblage, which dominates the Turonian-Santonian sequences, is replaced by a *Hormosina gigantea* assemblage which characterizes Upper Campanian and Maastrichtian biofacies in the flysch series as well as in the zeolitic claystones deposited below the CCD. In the pelagic limestone sequences of the Gubbio area and the Penibetic Zone, no prominent biosiliceous signal has been observed in the Lower Campanian. However, a turnover in the composition of agglutinated foraminiferal assemblages is likewise observed at that locality.

3. The Cretaceous/Paleocene Boundary Event (KTBE)

At the end of the Cretaceous period (66.4 Ma), hundreds of species of oceanic calcareous phyto- and zooplankton suddenly became extinct. According to the scenario of ALAVAREZ et al. (1980), the catastrophic impact of an asteroid-sized bolide produced short-term climatic changes which resulted in the collapse of the shallow-water marine trophic structure. Stable oxygen and carbon isotopes provide evidence of temperature change and greatly reduced oceanic primary productivity coincident with the K/T boundary clay and during the first few hundred thousand years of the Paleocene (HSÜ & McKENZIE 1985, ARTHUR et al. 1987, KAMINSKI & MALMGREN in press). However, the KTB also exhibits in several localities a dark laminated boundary clay, indicating oxygen deficiency at the ocean floor (HSÜ 1986). The effects of the K/T boundary event on deep marine benthic organisms are more poorly understood. Deep water benthic foraminiferal species were certainly less strongly affected by this event than planktonic foraminifera (KELLER 1988). A number of researchers have pointed out that there were either few extinctions among deep-water benthic foraminifera at the end of the Cretaceous (BECKMANN 1960, HILLEBRANDT 1965) or extinction occurred not suddenly but over a longer period of time (WEBB 1973, DAILEY 1983). We studied the taxonomic composition of DWAF assemblages in a continuous section in a slope sequence of the North African paleomargin (Section M83, Mesorif Zone, compare KUHNT 1987). The stratigraphic distribution of 51 agglutinated taxa across the K/T boundary is recorded in this section (Text-Fig. 4). Thirteen of these species appear for the first time in the Lower Paleocene. Only 5 species have their last occurrence near the boundary and for 3 of these 5 species a last occurrence has been recorded already within the Upper Maastrichtian.

Recent research on the population dynamics of modern species of benthic foraminifera (KAMINSKI et al. 1988b) provides new insight into the problem



Text-Fig. 4. Range chart of DWAF across the K/T boundary in a slope section of the North African Margin (Section M83, Mesorif Zone, Northern Morocco).

of detecting possible mass mortality among benthic foraminifers. In an experiment using recolonization trays placed in the Panama basin, KAMINSKI et al. (1988b) were able to identify which morphogroups of benthic foraminifera are opportunistic and which groups show limited capability for dispersal. A remarkable pattern in the community structure of benthic foraminifera has been observed by KUHNT (submitted) in the Gubbio section of Central Italy. The morphogroup of benthic foraminifera interpreted as opportunistic by KAMINSKI et al. (1988b) was observed to increase in relative abundance in the two samples directly overlying the K/T boundary clay. The morphogroup which characterizes stable environments and displays limited capability for dispersal displays a dramatic decrease from the top of the Cretaceous to the base of the Paleocene. This pattern can be interpreted as evidence for mass mortality among benthic foraminifers across the K/T boundary.

MOULLADE et al. (in press) observed a significant faunal break in deep-water agglutinated foraminifers coinciding with the K/T boundary in the North Atlantic Plantagenet Formation. However, in the sites studied, the Paleocene portion of the sequence revealed only few and poor foraminiferal assemblages, and a confirmation of these findings requires additional study.

The available data is still ambiguous as to whether or not the deep-sea benthos had been strongly affected by the K/T boundary event. This information is crucial to discriminate between the different proposed models to explain the cause of this major event.

4. Paleobathymetry

The paleobathymetric distribution of Upper Cretaceous DWAF assemblages can be reconstructed for the Maastrichtian time slice from a transect on the thrusted North African Margin in the Tellian Units of the Moroccan Rif (KUHNT in press). These data are compared to DSDP/ODP sites on the Northwest African Margin, Galicia Margin and North American Margin (compare Text-Fig. 1).

Analysis of the taxonomic composition of benthic foraminiferal assemblages and comparison with general faunal trends such as plankton/benthos ratios, diversity and abundance of agglutinated species led to the evaluation of characteristic foraminiferal biofacies. These trends have been used to define five bathymetric zones along a composite transect of the Upper Cretaceous North African and Iberian continental margins:

1. Upper Slope Assemblages (200 - 500 m water depth): Besides first flysch-type agglutinants (e. g. *Glomospira*) calcareous agglutinating forms like *Matanzia, Verneuilina, Gaudryina, Dorothia*, and *Spiroplectammina* are common (corresponding to the *Marssonella* association of HAIG 1979). Generally, this assemblage shows optimal benthos diversity. The plankton/benthos ratio is above 9.

2. Middle Slope Assemblages (500 - 1500 m water depth): Flysch-type forms (e. g. *Glomospira, Ammodiscus, Hormosina, Paratrochamminoides, Rhabdammina*) become an important constituent of the agglutinants, and calcareous ataxophragmids of the "*Marssonella* association" are still common. The amount of calcareous forms, especially planktonics, is quite variable (from almost zero to 99 %) due to effects of a local carbonate lysocline and/or redeposition.

In an intermediate bathymetric zone a *Rzehakina - Rhabdammina - Reussella* assemblage has been observed between the typical middle slope and the lower slope assemblages: Flysch-type agglutinants dominate. The plankton/benthos ratio is quite inconstant. The amount of planktonic foraminifera can be very low especially in assemblages with a high dominance of the *Rhabdammina* group.

3. Lower Slope Assemblage (1500->2500 m water depth): *Rhabdammina* assemblage, flysch-type agglutinated assemblages with high dominance of tube-like morphotypes (e. g. *Rhabdammina, Dendrophrya*). Planktonic foraminifera are very rare or absent, often fragmented and/or corroded and are restricted to forms with a high dissolution-resistance.

4. Abyssal Assemblage ("Flysch-type", occurring in regions with significant detrital input): *Recurvoides - Paratrochamminoides* assemblage without any autochthonous calcareous foraminifera (deposition below CCD). The diversity of agglutinated forms is higher in comparison to slope assemblages.

5. Abyssal *Labrospira - Praecystammina* Assemblages ("Krasheninnikov-type"): A purely agglutinated assemblage consisting mainly of small smooth-walled forms which is typical for late Cretaceous deep oceanic basins with low detrital input. The diversity of this assemblage is generally lower than that of flysch-type assemblages.

Similar paleobathymetric distribution patterns, typical for Upper Cretaceous low-latitude areas, have been observed in Trinidad and the Caribbean Sea (HEMLEBEN & TRÖSTER 1984, KAMINSKI et al. 1988a), and on the Baltimore Canyon transect (NYONG 1983, NYONG & OLSSON 1984).

Paleobathymetric patterns in northern assemblages (e. g. Labrador Margin) differ from low-latitude assemblages by the lack of a shallow "*Marssonella* association" of calcareous ataxiophragmids. Shallow Maastrichtian assemblages contain abundant coarse tubular species and lituolids with only very rare calcareous ataxophragmids. Deeper assemblages contain more abundant *Glomospira, Hormosina, Paratrochamminoides, Ammosphaeroidina, Praecystammina* and finely agglutinated lituolids (*Cribrostomoides, Haplophragmoides* and *Labrospira*). Campanian deep assemblages contain abundant *Uvigerinammina*.

5. Biostratigraphy

Several species of agglutinated foraminifers appear to be stratigraphically useful and have correlative first or last occurrence levels in several basins of the Alpine-Carpathian Mountain belt (GEROCH 1959, NEAGU 1968, 1970, SANDLESCU 1973, MORGIEL & OLSZEWSKA 1981, GEROCH & NOWAK 1984, GEROCH & KOSZARSKI 1988). For several of these taxa similar stratigraphic ranges have recently been observed in the Gibraltar Arch area (KUHNT 1987), in the North Atlantic (MOULLADE et al. in press), in the Umbrian Apennines and the Betic Cordillera (KUHNT in press) and in Trinidad (KAMINSKI et al. 1988a).

Several species (e. g. *Hippocrepina depressa, Plectorecurvoides irregularis, Trochammina abrupta, Recurvoides imperfectus, Haplophragmoides gigas minor*) have their last occurrences near the Cenomanian/Turonian boundary. Their extinction is most probably caused by oxygen-depletion of the deep sea during the paleoceanographic event at the Cenomanian/Turonian boundary (CTBE). The characteristic well oxygenated red clay or limestone facies above the CTBE contains *Haplophragmium problematicum* as an indicator.

The total range of *H. problematicus* is given as Lower Cenomanian to lowermost Campanian (NEAGU 1970, KUHNT in press). Its partial range and optimum occurrence characterize the Turonian Haplophragmium problematicus Zone (= A. problematicus Zone of GEROCH & NOWAK 1984 and the H. lueckei Zone of MOULLADE et al. in press).

The species *Uvigerinammina jankoi* and *Hormosina gigantea* are perhaps the most distinctive Upper Cretaceous species in North Atlantic and Tethyan flysch-type assemblages, and are used as stratigraphic marker species in every zonal scheme. The first occurrence of *U. jankoi* is noted immediately above the benthic-free interval of the Cenomanian/Turonian boundary event. The highest occurrence of *U. jankoi* is from Middle-Upper Campanian of the Indian Harbor well on the Labrador Margin. The report of this species from the Upper Maastrichtian of the Labrador Margin (MILLER et al. 1982, GRADSTEIN & BERGGREN 1981) is probably based on insufficient biostratigraphic calibration. The species *H. gigantea* is restricted to the Middle-Upper Campanian and Maastrichtian. The first occurrence of this species is a reliable indicator of the Middle Campanian in flysch and deep-sea environments.

The last occurrence of *Praecystammina globigerinaeformis* is useful for determining a datum level close to the Coniacian/Santonian boundary in abyssal assemblages, but this species ranges into younger levels at bathyal depths.

One of the most interesting Upper Cretaceous lineages is the evolution of the typical *Hormosina excelsa* from ancestors, which may belong to the *Hyperammina dilatata* group in the Coniacian/Santonian. With respect to size, the evolution of this form parallels that of the *H. ovulum - H. gigantea* group. Upper Cretaceous specimens are typically small, but near the top of its stratigraphic range in the Upper Maastrichtian to Paleocene, a size increase is observed.

Several species have their first occurrence in the interval between the Lower/Middle Campanian and the K/T boundary (e. g. *Rzehakina inclusa, Rzehakina epigona, Hormosina velascoensis, Glomospira diffundens*). It still remains to be tested whether or not these first occurrences are coeval (at the resolution provided by planktonic foraminiferal zonations) benthic foraminiferal events in the different Tethyan basins.

GEROCH & NOWAK (1984) defined a Lower Campanian G. rugosa Zone in the Polish Carpathians. However, the utility of *Goesella rugosa* to determine the Santonian/Campanian boundary (and of *Matanzia varians* as an indicator of Middle Maastrichtian and younger strata) is limited to slope assemblages. Moreover, we have not observed *G. rugosa* on the western North Atlantic margin. These mainly calcareous agglutinated species are absent in sediments deposited below the CCD.

The genera *Paratrochamminoides, Haplophragmoides, Karreriella* and *Pseudobolivina* may also have regional stratigraphic importance in the Upper Cretaceous, but more detailed taxonomic work is required on these species.

A major faunal turnover at the Cretaceous/Tertiary boundary is reflected by the first occurrence of many new species in the early Paleocene in Trinidad (Text-Fig. 4). Characteristic new species in the Danian are: *Budashevaella* cf. *multicamerata, Clavulinoides amorpha, Clavulinoides globulifera, Clavulinoides paleocenica, Conotrochammina whangai, Dorothia indentata, Eggerella trochoides, Haplophragmoides lamella, Karreriella tenuis, Phenacophragma beckmanni, Reticulophragmium* spp., *Reticulophragmoides jarvisi, Spiroplectammina spectabilis* (acme), *Spiroplectammina excolata, Trochammina ruthven murrayi.*

Interestingly enough, in the Western Mediterranean and the North Atlantic major faunal turnovers in deep-water benthic agglutinated foraminifers coincide with global paleoceanographic events at the Cenomanian/Turonian boundary, in the Lower/Middle Campanian and at the Cretaceous/Tertiary boundary. These "benthic events" could be regarded as inter-regional and isochronous, i. e. reliable datum horizons for the biochronology of deep-sea sediments.

6. Benthic Foraminifers and the Subsidence History of Western Mediterranean Margins

The burial history of the sedimentary wedges along the North African and Southern European margins in the Western Mediterranean is strongly influenced by the tectonić history of the region. The geologic record on the North African Margin shows four phases of sedimentation during the Cretaceous (KUHNT & OBERT, this volume), which probably have influenced the entire Cretaceous Rif-Betic seaway and its paleomargins.

1. Distension and subsidence of the margins in the Lower Cretaceous.
2. A first transpressional phase with uplift and slight metamorphism in the Albian to early Cenomanian, which affected the northerly ("Haut Tellien") paleogeographic zones of the North African continental margin.
3. A late Cretaceous stage of subsidence (Cenomanian-Santonian).
4. A second compressional phase starting with the Campanian and reflected by the formation of submarine fans with turbiditic sequences, sedimentary klippes, and olistostrome complexes.

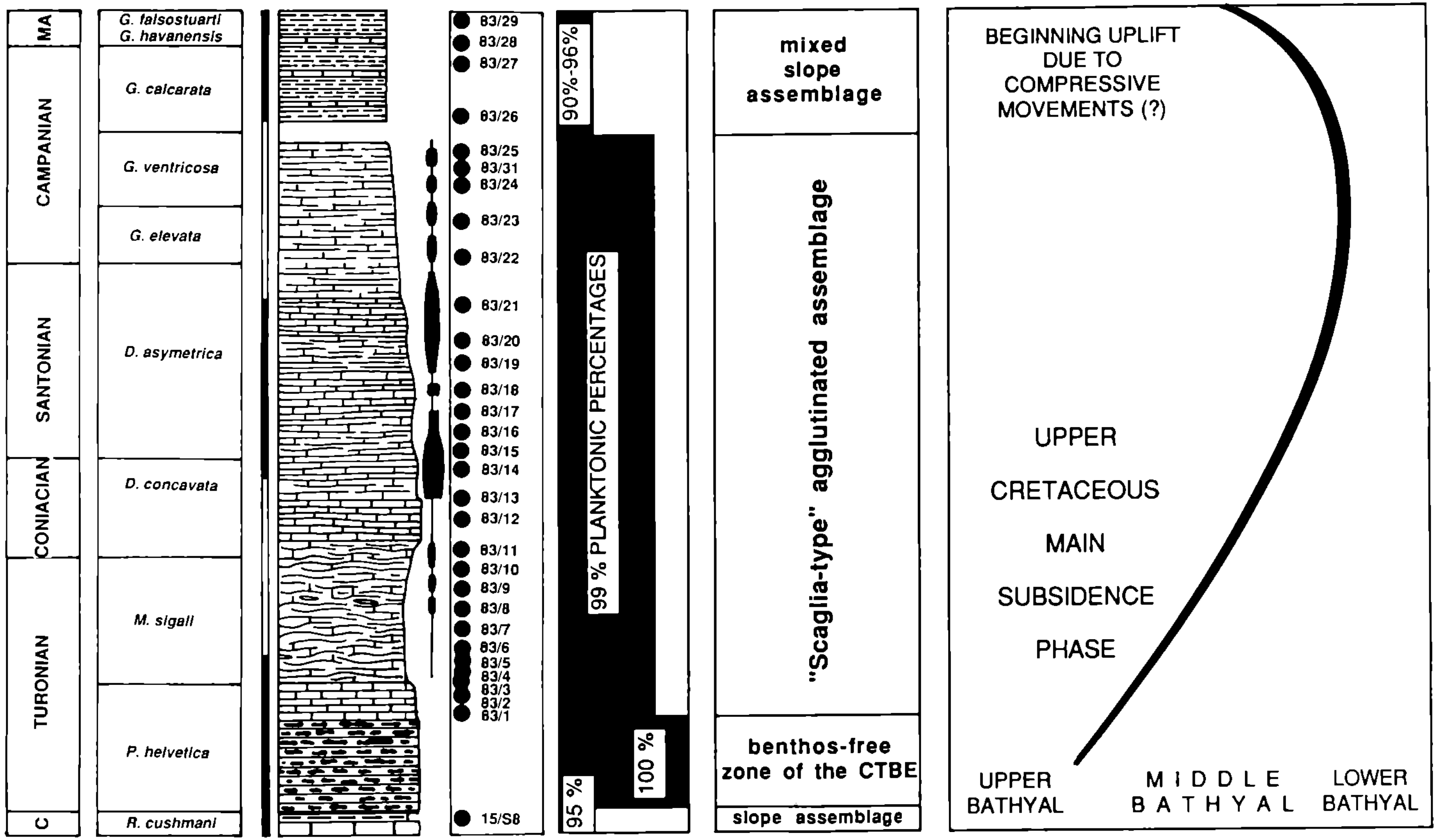
The temporal distribution of Upper Cretaceous deep-water agglutinated foraminiferal assemblages reflects this complex subsidence and sedimentation pattern. This relation can be demonstrated by two examples (Text-Figs. 5 and 6): the Upper Cretaceous of the Tanger Unit in the Moroccan Rif (North African Margin) and that of the Penibetic Zone in Southern Spain (European Margin of Iberia or the Alboran Block). In the Penibetic Zone (Text-Fig. 5), a shift from Albian-Cenomanian agglutinated assemblages, which are dominated by calcareous ataxophragmids towards "Scaglia-type" assemblages (KUHNT submitted) occurred during a Cenomanian-Turonian phase of strong subsidence, accompanied by a global sea-level rise. Since the late Campanian, when in the local paleotectonic pattern compressive movements and first uplift can be distinguished, ataxophragmid-dominated agglutinated assemblages re-occur in the Hacho de Montejaque section. In the Tanger Unit of the Moroccan Rif (Text-Fig. 6), well diversified flysch-

Text-Fig. 5. Cretaceous subsidence history and foraminiferal biofacies in the Hacho de Montejaque section (Penibetic Zone, Southern Spain).

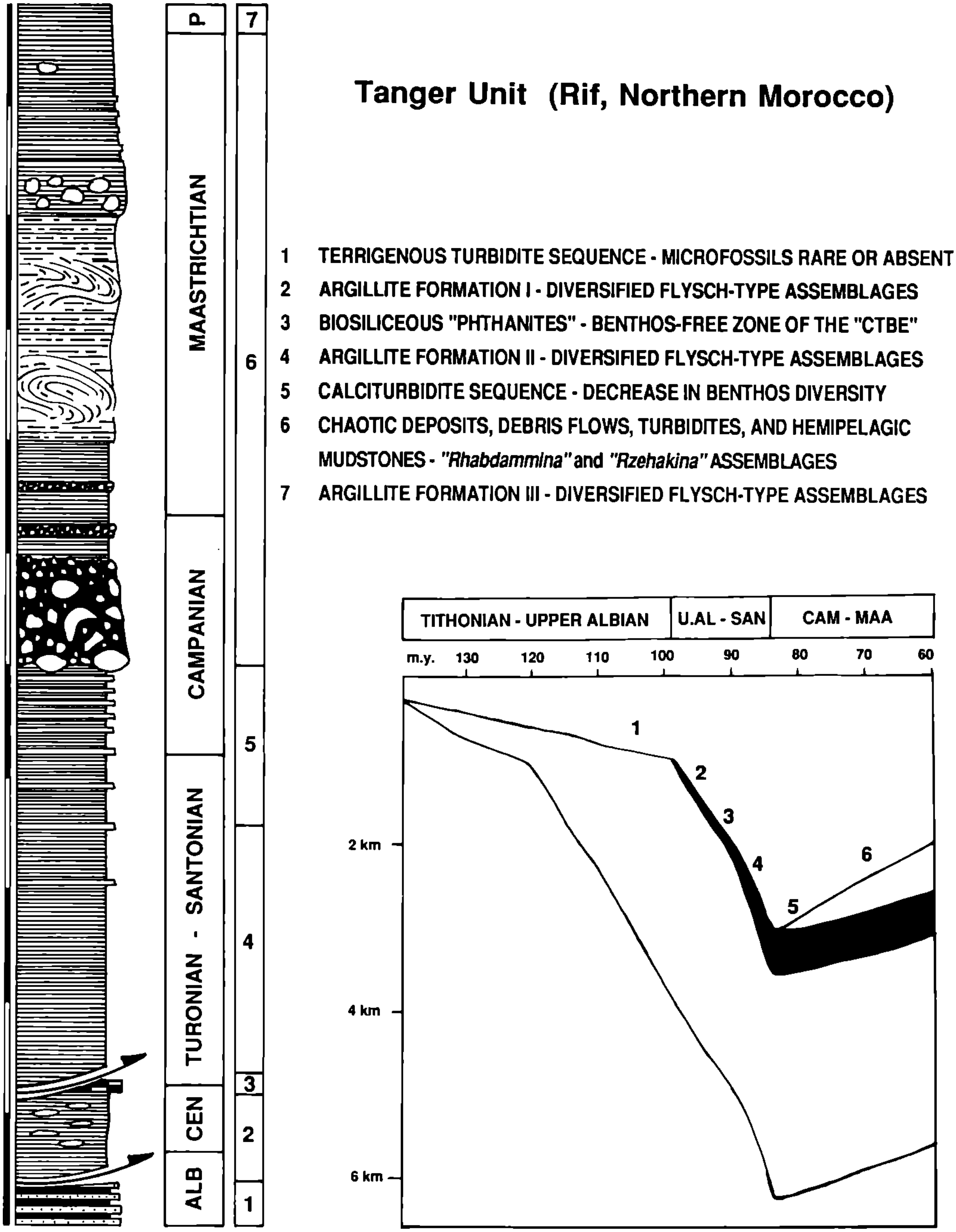
Text-Fig. 6. Upper Cretaceous to Paleogene in the Tanger Unit (Rif, Northern Morocco): Synthetic section, burial history (corrections for sea-level history and compactions are taken into account) and benthic foraminiferal biofacies.

Hacho de Montejaque Section

(Penibetic Zone, Spain)



Text-Fig. 5


P.
MAASTRICHTIAN
CAMPANIAN
TURONIAN - SANTONIAN
CEN
ALB
7
6
5
4
3
2
1
Tanger Unit (Rif, Northern Morocco)
1 TERRIGENOUS TURBIDITE SEQUENCE - MICROFOSSILS RARE OR ABSENT
2 ARGILLITE FORMATION I - DIVERSIFIED FLYSCH-TYPE ASSEMBLAGES
3 BIOSILICEOUS "PHTHANITES" - BENTHOS-FREE ZONE OF THE "CTBE"
4 ARGILLITE FORMATION II - DIVERSIFIED FLYSCH-TYPE ASSEMBLAGES
5 CALCITURBIDITE SEQUENCE - DECREASE IN BENTHOS DIVERSITY
6 CHAOTIC DEPOSITS, DEBRIS FLOWS, TURBIDITES, AND HEMIPELAGIC MUDSTONES - "Rhabdammina" and "Rzehakina" ASSEMBLAGES
7 ARGILLITE FORMATION III - DIVERSIFIED FLYSCH-TYPE ASSEMBLAGES
TITHONIAN - UPPER ALBIAN
U.AL - SAN
CAM - MAA
m.y.
130
120
110
100
90
80
70
60
2 km
4 km
6 km
1
2
3
4
5
6

Text-Fig. 6

type agglutinated assemblages, which characterize Albian to Santonian beds, were replaced by low-diversity *Rhabdammina* and/or *Rzehakina* faunas during the Campanian. These assemblages reflect the enhanced detrital input, the sedimentary filling of basins and probable tectonic uplift of certain regions during the Campanian-Maastrichtian compressive paleotectonic phase.

7. Conclusion

Deep-water agglutinated foraminifers (DWAF) provide valuable information about late Cretaceous slope and deep oceanic paleoenvironments in the Western Mediterranean area. Their community structure is mainly influenced by paleobathymetry, oxygenation of bottom waters, detrital input, substrate disturbance, and availability of calcium carbonate. The distribution pattern of these benthic foraminiferal communities in time and space can provide valuable information about regional tectonic effects, such as the relationship between depocenters and continental breakup or collision. The evolution of DWAF is strongly influenced by three paleoceanographic events: (1) at the Cenomanian/Turonian boundary, (2) in the Lower/Middle Campanian, and (3) at the Cretaceous/Tertiary boundary. The important faunal changes at these levels ("benthic events"), coeval with global paleoceanographic events, could be regarded as inter-regional and isochronous, i. e. reliable datum horizons for the biochronology of deep-sea sediments deposited below the CCD on more than a regional scale.

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Sedimentary Evolution of the Shallow-Marine Late Cretaceous in the Southern Basco-Cantabrian Basin (Northern Spain)

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With 3 Plates, 9 Text-Figures and 1 Table

MARTINS, U. P. (1989): Sedimentary Evolution of the Shallow-Marine Late Cretaceous in the Southern Basco-Cantabrian Basin (Northern Spain). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 121-144. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: During Senonian time, the Basco-Cantabrian Basin was divided into 3 main sedimentary belts of WNW-ESE extension. In the northern belt, thick flysch deposits reflect the strong subsidence of the Bay of Biscay. The central part is characterized by epicontinental marly limestones, whose thickness is reduced to 1/3 compared to the flysch trough in the North. A shallow marine environment developed in the southern belt. Mostly subtidal to intertidal sediments of very diverse facies were deposited throughout the Upper Coniacian to Santonian.

The Campanian is dominated by increasing clastic sedimentation which already begins in the Upper Santonian. Two locally active discharge funnels were established. As a result, carbonate sedimentation stopped in important parts of the studied area.

From Upper Campanian to Middle Maastrichtian time, the clastic sediment supply decreased in both deltas. As a consequence, in parts of the basin, carbonate platforms developed again. However, the environment was more restricted, and carbonate sedimentation, predominantly as primary dolomite, alternated with distal clastic fan lobes.

Towards the end of the Maastrichtian, terrigenous input increased in the central part of the southern Basco-Cantabrian Basin (Garumnium Facies), whereas shallow marine and sabkha sedimentation continued in the eastern and western parts.

This general scheme is complicated by important lateral facies variations, which are a consequence of the tectonic history. A framework of faults parallel and perpendicular to the direction of the main rift zone is responsible for the activation of a tilted-block system. Strong diapirism plays an additional role in facies differentiation.

Kurzfassung: Während des Senons war das Basko-Kantabrische Becken in drei parallele Fazieszonen mit WNW-ESE-Erstreckung unterteilt. Die nördliche Zone erfuhr starke Subsidenz, durch mächtige Flyschablagerungen belegt. In der mittleren Zone dominieren epikontinentale Kalk-Mergel-Serien mit erheblich geringerer Mächtigkeit (gegenüber dem Flyschtrog um zwei Drittel reduziert). Die südliche Zone dagegen wird von flachmarinen, subtidalen bis kontinentalen Sedimenten eingenommen.

Die Sedimentverbreitung im südlichen Basko-Kantabrischen Becken war während des Coniac und Santon mit subtidalen Plattform-Karbonaten, Dolomiten und Mergeln bereits uneinheitlich. Diese Tendenz verstärkte sich während des Unteren Campan und fand im Oberen Campan seinen extremen Ausdruck in der Schüttung zweier Deltasysteme. Damit wurde in bedeutenden Bereichen des Sedimentationsgebietes die Karbonatproduktion gestoppt.

Mit Ende des Campans ließ die Schüttungsintensität in den Deltas spürbar nach. Zögernd beginnende Kalkproduktion setzte erneut in allerdings eng begrenzten Gebieten ein. An anderen Stellen, vor allem im Südwesten, bildeten dolomitische Sabkhas weite Gürtel entlang der Küstenlinie, in denen aufgrund extremer chemischer Bedingungen keine normal-marine Sedimentation erfolgen konnte. Das gesamte Sedimentationsgebiet unterlag jedoch weiterhin schwachem klastischem Eintrag.

Im Verlauf des Maastricht wuchs der terrigene Sedimentanteil wiederum stark an, im zentralen südlichen Basko-Kantabrischen Becken bildeten sich lagunäre bis brackische Verhältnisse aus (Garumnium-Fazies). Die östlichen und westlichen Bereiche dagegen waren hauptsächlich von sehr flachmarinem Environment bestimmt (flaches Subtidal bis Sabkha).

Die beträchtlichen lateralen Unterschiede im Sedimentationsgeschehen sind auf die Bewegungen eines parallel sowie senkrecht zur Riffachse streichenden Störungssystems zurückzuführen, wodurch im Verlauf der Rotation der Iberischen Mikroplatte eine Reihe von Kippschollen reaktiviert wurde. Beträchtlicher Diapirismus beeinflußte die sedimentäre Entwicklung im Basko-Kantabrischen Becken während der gesamten Oberkreide.

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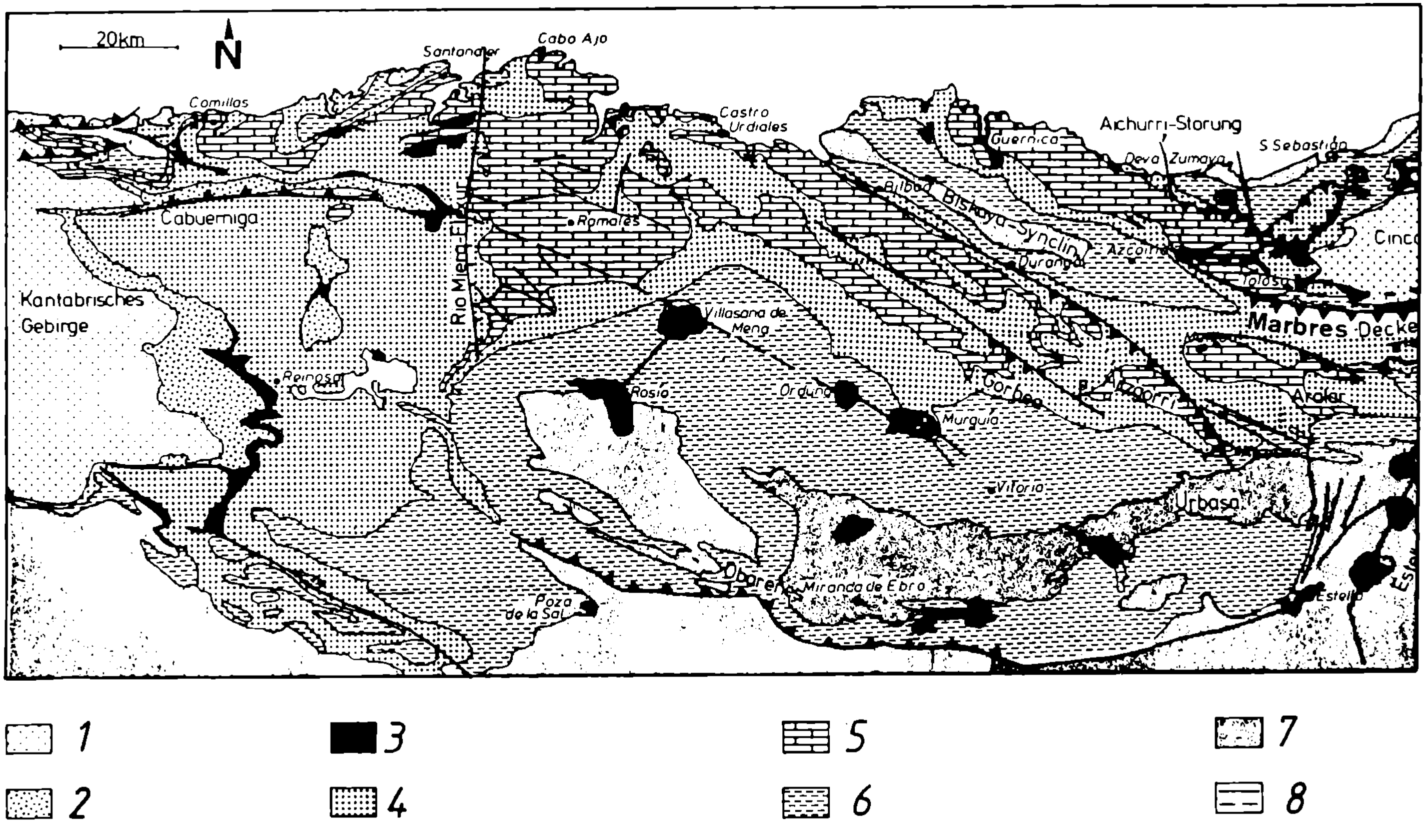
1. Introduction

The Basco-Cantabric Basin represents the eastern part of the southern margin of the Bay of Biscay, which developed as a branch of the Atlantic Ocean. Throughout Mesozoic time this basin was formed as a complex strike-slip pull-apart system by significant tectonic movements. During the middle and late Cretaceous the tectonic activity culminated in the establishment of a spreading ridge in the Biscay Ocean. This lead to a short-term subduction or underthrusting of approximately 100 km of crust material beneath the northern Iberian Continental Margin.

This effect, combined with the anti-clockwise rotation and northward drift of the Iberian Microplate (ENGESER & SCHWENTKE 1986) caused considerable compression and uplift. In consequence, sedimentary processes and environments in the entire basin changed significantly.

The Basco-Cantabrian Basin has a WNW-ESE extension (Text-Fig. 1) and is divided into 3 main sedimentary belts, due to the deepening of the basin towards the NNE. The northernmost belt represents the bathymetrically deepest environment and is mostly governed by turbiditic sedimentation throughout the whole Upper Cretaceous. The sediments of the central belt are epicontinental marly limestones and marlstones with a typical fauna rich in ammonites, echinids and planktonic foraminifera.

In contrast to this, the southern belt of the Basco-Cantabrian Basin was dominated by both shallow marine and fluviatile to deltaic sedimentation processes. The paleoshore was situated along this line - although significant fluctuations occurred through time.



Text-Fig. 1. Geological map of the Basco-Cantrabrian Basin (from ENGESER et al. 1984): 1 - Paleozoic; 2 - Permotriassic; 3 - Triassic; 4 - Jurassic-Clastic Lower Cretaceous; 5 - Urgonian and Basin Facies, Aptian; 6 - Upper Cretaceous; 7 - Tertiary; 8 - Faults.

Previous work has been presented by CIRY (1939), MANGIN (1960), LOTZE (1960, 1963), WIEDMANN (1962a, 1962b, 1964), KOPP (1964), RAMIREZ DEL POZO (1971), LAMOLDA et al. (1981), FLOQUET (1982), RAT et al. (1982), WIEDMANN (1982), WIEDMANN et al. (1983) and ENGESER et al. (1984).

CIRY (1939) identified the nature of the sediments in the southern Basco-Cantabrian Basin as having a shallow marine origin. The results of his and MANGIN's (1960) work provide a fundamental base of knowledge, which was very useful for the regional correlation of facies. PLAZIAT (1981), WIEDMANN (in LAMOLDA et al. 1981), FLOQUET (1982) and RAT et al. (1983) published the first paleogeographic interpretations for the Upper Cretaceous of the region studied. WIEDMANN et al. (1983), ENGESER et al. (1984), SCHWENTKE & WIEDMANN (1985), and ENGESER & SCHWENTKE (1986) and MESCHEDE (1987) elucidated the tectonic history of the Basco-Cantabric Basin and established models for the sedimentary evolution of the basin's northern and eastern parts. These models allow interpretation of basinal facies development in terms of modern plate tectonics and basin analysis.

The results given in this paper reflect the testing of the existing models in the shallow marine environment of the southern and western Basco-Cantabrian Basin and the reconstruction of the sedimentary response to tectonic movements. Three main sedimentary phases can be stressed:

1. Persistance of a vast belt of carbonate platforms along the paleoshore from the Upper Coniacian through Lower Campanian.
2. Development of two distinct delta systems, one from the South and one from the West, beginning in the Upper Campanian and ending in the Lower Maastrichtian.
3. Sabkha facies during the Upper Maastrichtian and Paleocene with minor clastic influence and rare marine incursions.

Methods

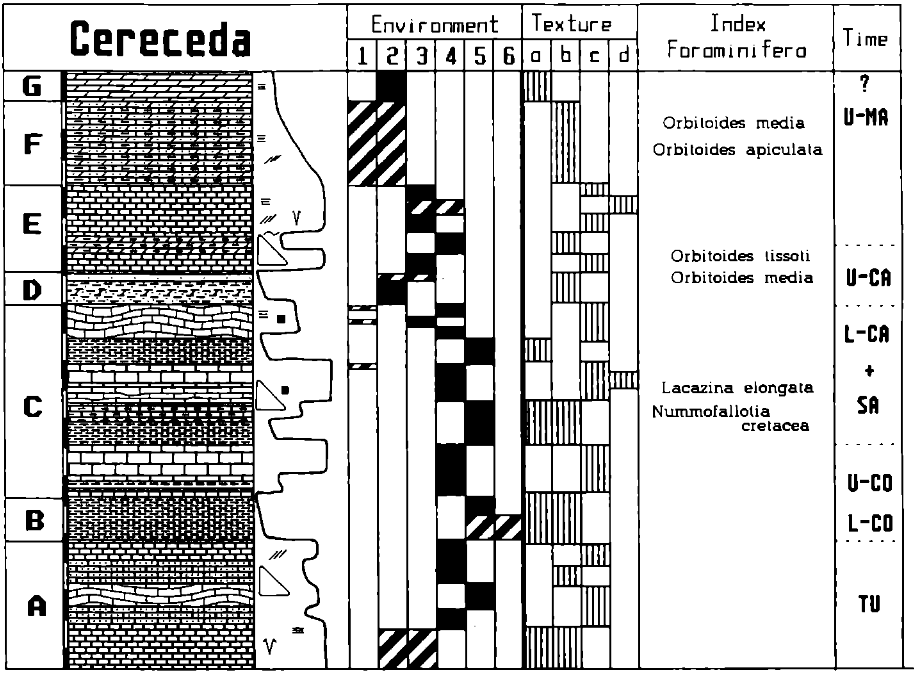
Description and measurement of various sections within the Basco-Cantabrian Basin yielded material for several hundred thin-sections. Detailed microscopic analysis was done by using the nomenclature of FLÜGEL (1982) and the current international references (CHOQUETTE & PRAY 1970, LONGMAN 1980, SCHNEIDERMANN & HARRIES 1985, REIJERS & HSÜ 1986, and others). The thin-sections containing dolomite were stained with Alizarin-S in order to differentiate between dolomite and calcite.

Mineralogic and geochemical analyses were done by X-ray diffractometry, electron-beam microprobe analysis with energy dispersive X-ray analyzer and scanning electron microscopy.

2. Lithologic description

Since the establishment of Walther's Law in 1894 (WALTHER 1894), facies analysis and correlation have evolved as the most valuable tool for the reconstruction of paleo-landscapes, usually designated as paleogeography.

The lateral (as well as the vertical) facies distribution in the Basco-Cantabrian Basin is highly diverse throughout the whole late Cretaceous. In Text-Fig. 2 a typical section of the central southern Basco-Cantabrian Basin



Text-Fig. 2. Detailed section of the late Cretaceous near Cereceda. Environments: 1 - continental/fluvial; 2 - supratidal; 3 - intertidal; 4 - shallow subtidal; 5 - platform; 6 - shelf. Texture: a - mudstone; b - wackestone; c - grainstone; d - floatstone. One unit of scale bar represents 20 m. (Patterns and symbols, see Table 1).

is presented. The section is located directly north of the small village of Cereceda (see Text-Fig. 3). For further explanation of the lithologic patterns used in the figures see Table 1.

The Cereceda section can be divided into several sedimentary units according to the general aspects of facies.

Unit A (90 m): Turonian facies are characterized by thin (maximum 1.40 m) sequences, whose bases (50 - 80 cm) consist of massive grainstones with various amounts of bioclasts (benthic foraminifera, debris of molluscan shells, lumps and peloids). The middle part (20 - 40 cm) of these sequences shows thin laminations, partly interrupted by vertical cracks, partly bowl shaped. The grain size within the laminations is extremely fine. Under the microscope, large amounts of small peloids (5 - 20 µm) can be observed. The tops of the sequences are sometimes lacking, but in most cases appear 20 - 30 cm thick layers of light to dark gray and brown coloured, fine-grained limestone with considerable dolomite content (up to 40 %). These layers show a generally gradual transition from the underlying laminites, but a sharp and often strongly undulating top. The fabric is governed by tubes and/or fine dikes usually filled with coarser material of lighter colours. In thin-section pisoids, argillans and strong dissolution features (Pl. 2, Fig. B) are observable.

Table 1. Lithologic patterns and symbols used in the text-figures in relation to the main facies zones of shallow marine environments defined by WILSON (1974, 1975).

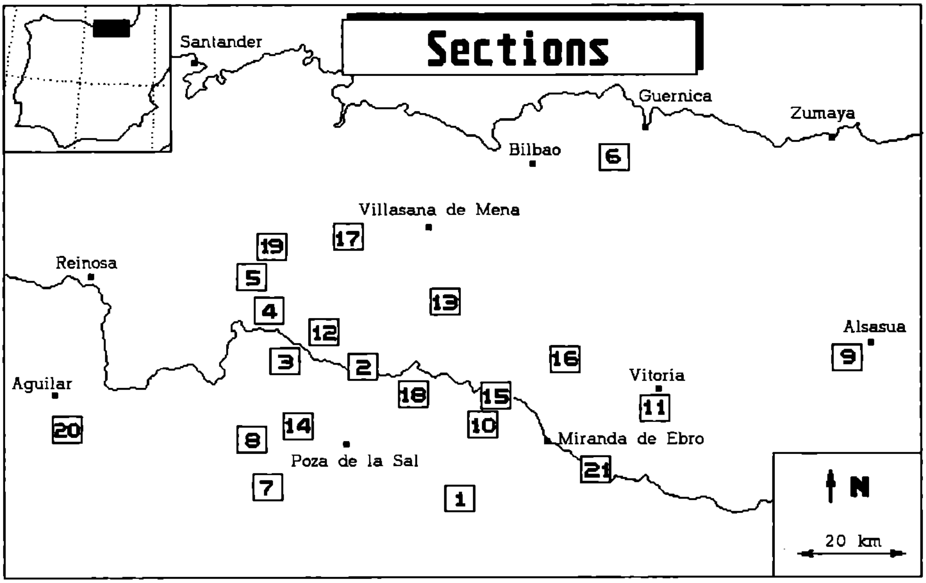
1	Limestone -> Platform (7)		
2	Limestone, rich in Rudists -> Platform Margin (5)		
3	Marl -> open Shelf (2)		
4	Arenitic Limestone -> Platform (7)		
5	Dolostone -> Tidal Flat to Sabkha (8 - 9)		
6	Arenitic Dolostone -> Platform to Sabkha (7 - 9)		
7	Calcareous and marly Sandstone -> Platform to Tidal Flat (7 - 8)		
8	Arenitic Marls -> Shelf to Foreslope (2 - 4)		
9	Graded Sandstone, Trough Cross Bedding -> Delta		
10	Dolomitic and marly sandstone -> Platfom to Sabkha (7 - 9)		
11	Dolomitic, arenitic, marly Limestone -> Shallow Platform (7 - 8)		
12	Dolomitic, arenitic Limestone -> Shallow Platform (7 - 8)		
13	Marly Dolostone -> Tidal Flat to Sabkha (8 - 9)		
14	Massive Limestone -> Platform (7)		
15	Limestone with undulating Bedding Planes -> Platform (7)		
///	Cross Bedding	V	Bioturbation
≡	Lamination	■	Coal
◺	Shallowing Upward Units	~	Ripple Marks

Unit B (30 m): The Coniacian strata mostly consist of marly limestones and marlstones of a light yellow colour. The fauna is not rich at the Cereceda location, but in other places in the area studied (Section Nidaguila, see Text-Fig. 3) strata of identical lithology and stratigraphic position have been described and dated as Upper Turonian to Upper Coniacian by WIEDMANN (1979).

Unit C (135 m): The Upper Coniacian to Lower Campanian is built up with massive limestone beds consisting of light gray to brown grainstones with variegated components, principally bioclasts. The strata display undulating bedding planes. An additional feature is the presence of beds containing reworked clasts, built up with material from the same environment, i. e., intraformational breccias (Pl. 1, Fig. A).

Towards the Upper Santonian some scarce, thin (a few centimeters) coal layers are intercalated, preceded by fine-grained and laminated limestone of similar appearance as described above from the Turonian. The Upper Santonian is characterized by the extraordinarily common miliolid foraminifera *Lacazina elongata* (Pl. 2, Fig. C). The transition from the Upper Santonian to the Lower Campanian cannot be recorded by paleontological means. Nevertheless, the lithologic patterns are comparable with other sections, where the Upper Santonian limestones are overlain by marly limestones of Lower Campanian age, which are dated with ammonites (*Neocrioceras risosi* from the section Quintanilla la Ojada, Nr. 13, Text-Fig. 3, see WIEDMANN 1962b; and *Eupachydiscus levyi* GROSS. from the section Nocedo, Nr. 8, Text-Fig. 3). Intercalations of coal layers and related facies are common in this level.

Unit D (23 m): The Upper Campanian is partly buried, but the exposed strata show the same features as in other nearby locations (section Quin-



1 Bureba
2 Cereceda
3 Condado
4 Incinillas
5 Manzanedo
6 Marquina
7 Nidaguila
8 Nocedo
9 Olazagutia
10 Portillo
11 Puerto de Vitoria
12 Quecedo
13 Quintanilla la Ojada
14 Quintanaloma
15 Sobron
16 Subijana
17 Torme
18 Trespaderne/Rio Oca
19 Tubilla/Rio Nela
20 Villaescusa de las Torres
21 Zambrana

Text-Fig. 3. Location of the sampled sections in the Basco-Cantabrian Basin.

tanaloma, Nr. 14, Text-Fig. 3), i. e., a limestone/dolostone with up to 50 % detrital quartz (grainsize 100 - 500 µm), comparable with the formation "Grès de Sedano" defined by CIRY in 1939. The strata do not contain any fauna which is stratigraphically relevant. KARRENBERG (1934) and WIEDMANN (pers. comm.) report the presence of crustacean fragments. At the Cereceda section no specific sedimentary features could be observed within the unit, but near Sedano, the Middle to Upper Campanian displays cross-bedding and low-angle sigmoidal bedding. Bright, yellow, tube-like structures are abundant within the unit (Pl. 1, Fig. B). The "Grès de Sedano" is very red in colour. The thickness at the Cereceda section is 3 m, whereas at the type locality near Sedano 25 m of this lithologic unit are found. Beneath the 3 m of calcareous and dolomitic sandstone lie about 20 m of arenitic marls with a high content of fine-grained dolomite.

Unit E (56 m): Above the red-bed like facies of the Upper Campanian appear 15 m of limestone, which are split into several fining upward cycles, each with an average thickness of 90 cm. The texture of the base of each cycle is that of a grainstone containing large amounts of bioclasts. The middle part is characterized by fine laminites with abundant open space structures such as bird's-eyes and stromatactis. The top of each cycle is marked by 10 - 20 cm thick marlstone layers of gray colour which are often bioturbated. The age probably is still Upper Campanian, based on the presence of *Orbitoides tissoti*.

Above the grainstones occur 12 m of red and yellow dolomitic marls with a high content of quartz (up to 15 %). The marls include a rare fauna of *Nautilus* sp., undeterminable lamellibranchiates and gastropods. Sedimentary structures are abundant: cross-bedding, flaser-bedding and ripple marks can be observed. The rock also displays a large number of vertical borrows.

The next higher member of unit E is 35 m of limestone, bedded in a 1 to 1.5 m rhythm. The texture is again mostly that of a coarse grain-

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Plate 1

Fig. A. Reworked Upper Santonian grainstones (intraformational breccia), Cereceda. Diameter of lens cap: 6.5 cm.

Fig. B. Colour mottling within the Middle to Upper Campanian "Grès de Sedano" as a product of intensive bioturbation, Quintanaloma-Sedano. Diameter of lens cap: 6.5 cm.

Fig. C. Small fluvial channel, Upper Campanian, Puerto de Vitoria. Hammer is 50 cm long.

Fig. D. Braidplane facies, consisting of alternances of coarse-grained sand-bodies with fine-grained sand/mud layers. Note large water-escape structures (tepees), Upper Campanian, Puerto de Vitoria section. Hammer is 50 cm long.

Plate 2

Fig. A. Rudist colonies, reworked, Upper Campanian, Quintanaloma. Diameter of lens cap: 6.5 cm.

Fig. B. Dissolution features in limestone, Turonian, Cereceda, SEM (GPIT 7871).

Fig. C. Miliolid foraminifera *Lacazina elongata*, Santonian through Campanian, PPL. Diameter of specimen 1 mm (GPIT 154/03).

Fig. D. Oyster fragment, imbedded in fine-grained dolomitic matrix, Lower Santonian, Villaescusa de las Torres, PPL. Scale bar 1 mm (GPIT 318/10).

Plate 3

Fig. A. Desiccation features in dolomite, Maastrichtian, Cereceda, PPL. Scale bar 1 mm (GPIT 1510/03).

Fig. B. *Siderolites calcitrapoides*, Upper Maastrichtian, Torme, PPL. Diameter of specimen 2 mm (GPIT 264/04).

Fig. C. Two different dolomite textures in Upper Santonian dolostone, Villaescusa de las Torres, SEM (GPIT 7870).

Fig. D. Foraminiferal test, partly replaced by idiotopic and zonated dolomite, Campanian, Trespaderne/Rio Oca, PPL. Scale bar 100 μm (GPIT 810/1).

Plate 1

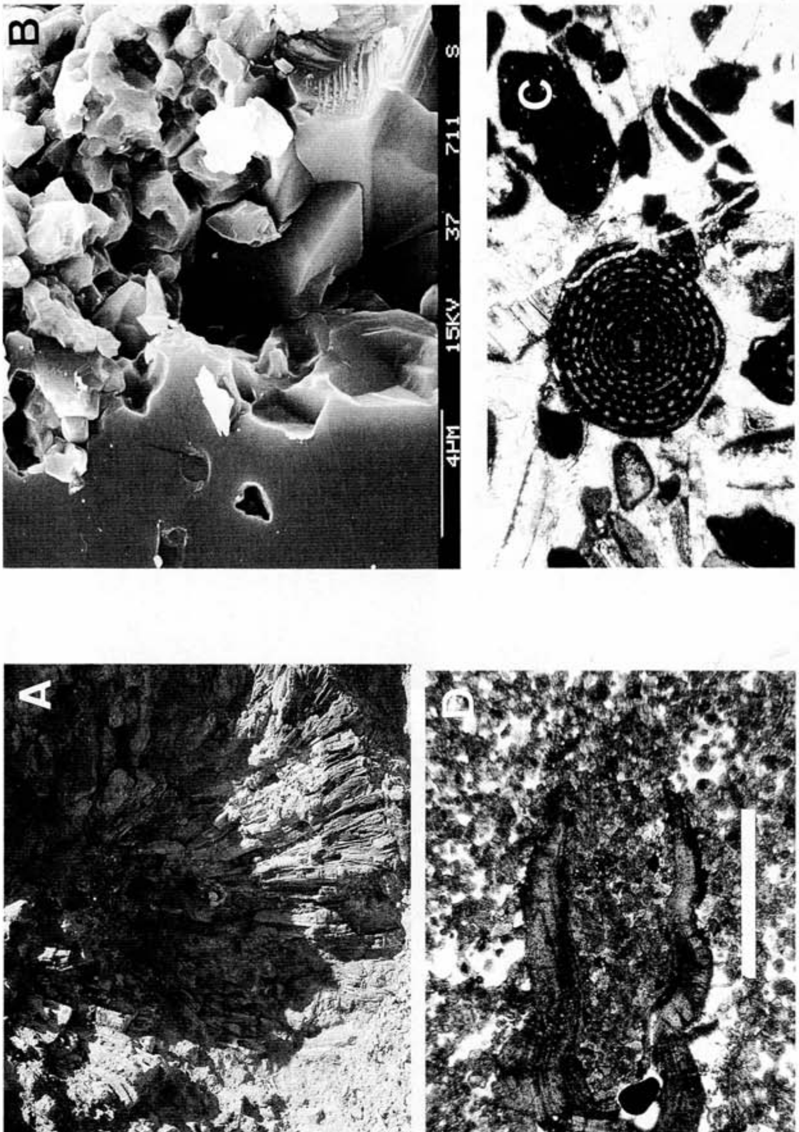


Plate 2

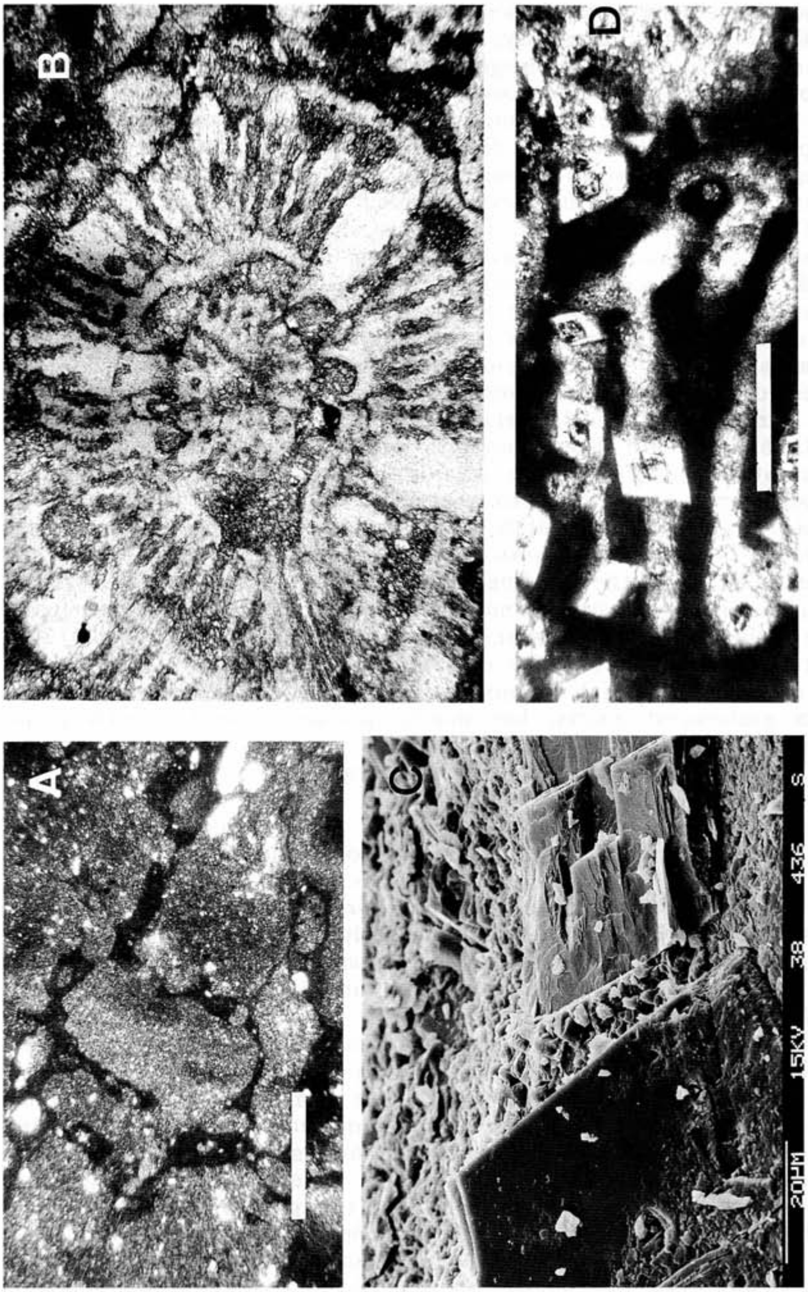
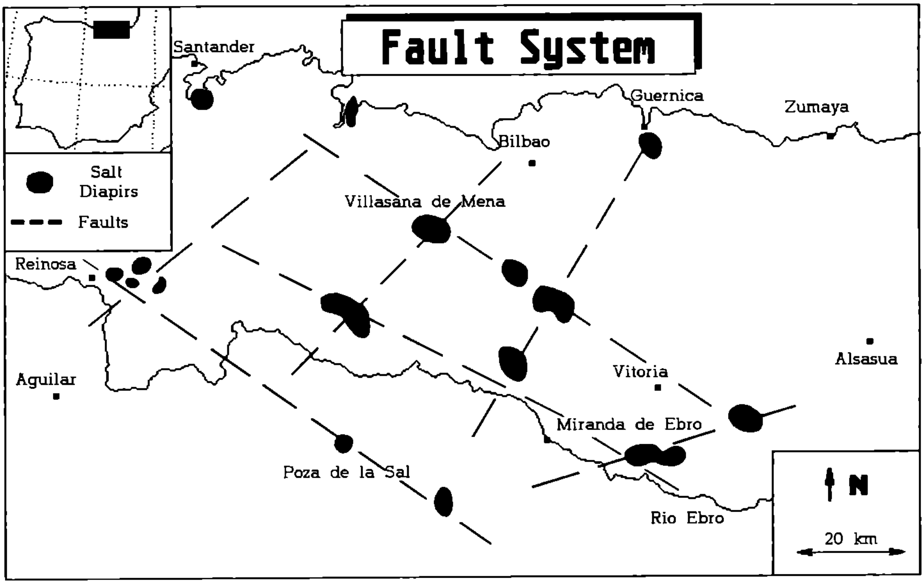


Plate 3



Text-Fig. 4. Tectonic framework with major salt diapirs (after REITNER 1986).

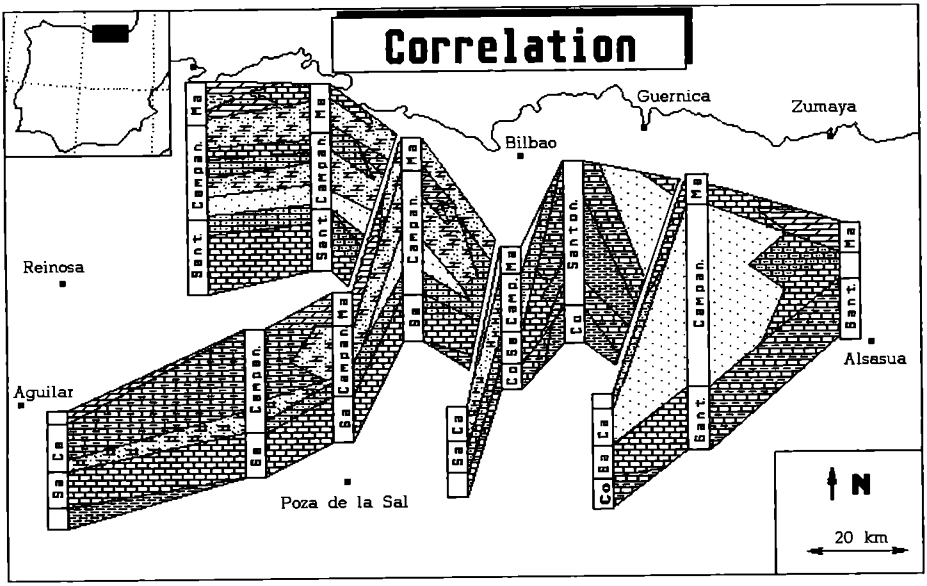
cant thinning of the sedimentary sequence (for example at the Zambrana section nearby the Penacerrada Diapir) and so complicating the patterns of coastal onlap. Due to this fact, the correlation of Cretaceous sealevel history in northern Iberia with the global record (HAQ et al. 1987) is not always satisfying.

The stratigraphic and lithologic correlation of selected sections is presented in Text-Fig. 5. Remarkable differences in the thickness of the sections can easily be explained by the influence of nearby saltdomes (Zambrana section) or the position with respect to the delta systems that were active during the Upper Campanian and Lower Maastrichtian. Sections situated in the vicinity of the discharge funnels are considerably thicker (Vitoria section) than those far from the sources of clastic sediment supply (Villaescusa de las Torres section).

4. Paleogeographic reconstruction

Using the results obtained by the analysis of lateral facies development in the Basco-Cantabrian Basin, a sequence of four different paleogeographical sketches (Text-Figs. 6-9) related to the four most significant periods within the Upper Cretaceous is presented.

These are the following: Upper Santonian, Lower and Upper Campanian and Upper Maastrichtian. Problems arise for the Upper Maastrichtian because the outcrops of this period are scarce and additionally, the biotopes seem to have been so unfavourable that paleontologic data are sometimes uncertain.



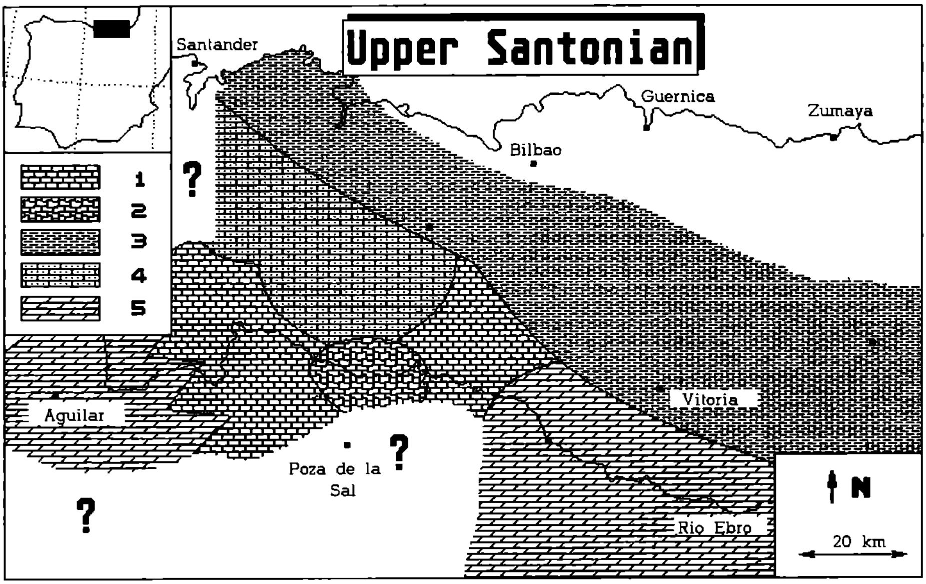
Text-Fig. 5. Facies correlation within the southern Basco-Cantabrian Basin, using significant sections in their geographic context (Patterns, see Table 1).

ENGESER et al. (1984) published several palinspastic reconstructions for the Basco-Cantabrian Basin. Their fig. 6 shows that the southwestern area did not suffer particular compression. Therefore, a palinspastic approach to the facies evolution of the southwestern Basco-Cantabrian Basin would not add to the information pertaining to normal paleogeographic maps.

4.1 Upper Santonian

Text-Fig. 6 represents the paleoenvironmental zonation of the Upper Santonian strata. The North and Northeast of the investigated area is dominated by marly limestones and marlstones. The transition from this epicontinental depositional environment to the near-shore zone follows a line parallel to the basin's main tectonic axis (WNW - ESE). Towards the South a threefold subdivision of the paleoenvironment is to be noted. Considerable amounts of dolomite occur in the Southeast. The middle sector shows bright, grainsupported limestones with only minor clastic components. Finally, the northwestern part shows more or less similar facies like those in the middle, but a distinct influence of clastic sedimentation can be noted (1 - 5 % quartz, mostly silt).

A relatively small area in the center of the area studied was colonized by rudist communities, mostly in the form of little "bouquets" (PHILIP 1972) and patches. None of the mostly elongated and cylindrical species could be found in-situ. They are entirely reworked and transported by the influence of storm events which can easily be recognized within the whole sequence.



Text-Fig. 6. Paleogeographic distribution of facies during Upper Santonian (Patterns, see Table 1).

Determination of the transition from one facies to another, especially in the region of the Villaescusa de las Torres section in the Southwest (5 km south of Aguilar), is very difficult because the Santonian to Maastrichtian strata do not crop out between Sedano/Tubilla del Agua (section Nr. 19, Text-Fig. 3) and the small syncline south of Villaescusa (section Nr. 20, Text-Fig. 3). Thus, the determination of the shoreline's movement through time is subject to further discussion.

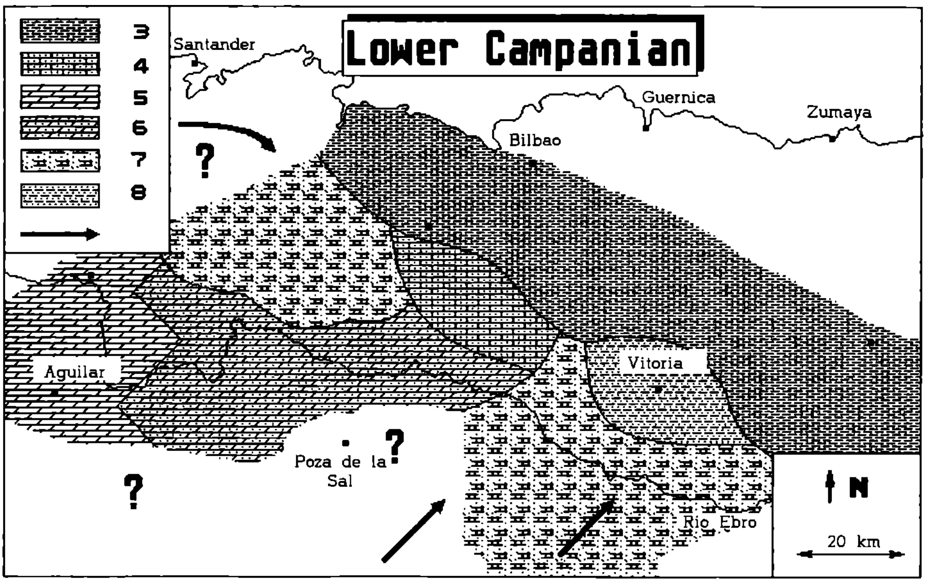
In the southwestern and southeastern parts of the investigated area none of the bioclastic, shallow subtidal platform carbonates are present that are found in all other places. Near Villaescusa de las Torres, Pancorbo (FLOQUET, in RAT et al. 1983) and Zambrana only the Lower Santonian shows bioclastic grainstones normally typical for the greater parts of the Santonian in the investigated area. In these places, the Upper Santonian is dominated by thick beds of dolomite, whose origin cannot be clearly determined.

Following the most recent literature on the dolomite problem (ZENGER et al. 1980, MACHEL & MOUNTJOY 1986, HARDIE 1987, GIVEN & WILKINSON 1987), the occurrence of thick beds of variable-grained, sometimes sucrosic dolomite which displays no clearly interpretable microscopic or macroscopic structure (for example, those found at the Villaescusa section) does not allow any reasonable statements as to whether the mineral is of primary, i. e., sabkha origin or only the consequence of changing physicochemical conditions during early or late diagenesis. Pl. 3, Fig. C shows two different textures of dolomite: fine-grained matrix and coarse-grained crystals replacing the matrix. The interpretation of the presumed Upper Santonian dolomite at the Villaescusa section does not provide fruitful results: the different textures could have resulted from replacement of former grainstones of the "normal" facies. This assumption is supported by the appearance of fragments and ghosts of former organisms (mostly oyster shells) in the Lower Santonian calcareous grainstones (Pl. 2, Fig. D).

Nevertheless, the presence of the dolomite itself indicates a notable difference in paleoenvironmental conditions around the southwesternmost sector of the Basco-Cantabrian Basin in comparison with all the other parts of the area. It seems very likely that the replacement of former aragonite by dolomite is a hint for restricted evaporitic or at least lagoonal conditions in the southwestern and southeastern sectors of the area. The desiccation features figured in Pl. 3, Fig. A, which occur within fine-grained dolomites are an additional aspect common in restricted environments with at least episodic or seasonal wetting and drying cycles, as in the intertidal to supratidal zone. This would coincide with the paleogeographical situation, bearing in mind that the coastline apparently follows the main direction of the basin's axis.

4.2 Lower Campanian

During the Lower Campanian the situation had profoundly changed. In Text-Fig. 7 two distinct areas influenced by strong clastic deposition can be seen. A transitional facies from the calcareous and marly sandstones or sandy limestones to an open marine environment (dominated by marls and marly limestones) could only be observed in the region around Vitoria. The facies south of a Northwest trending line between Vitoria - Santander is,

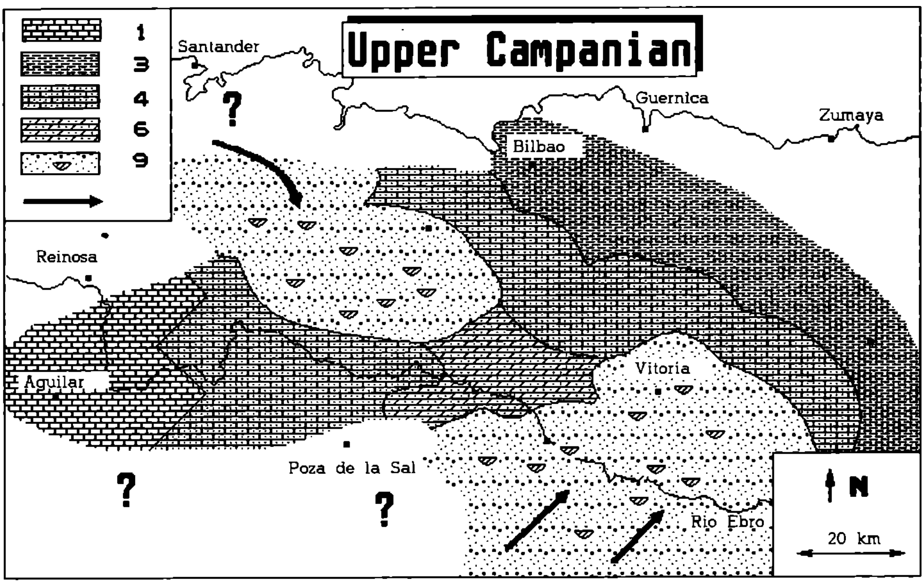


Text-Fig. 7. Paleogeographic distribution of facies during Lower Campanian. Arrows indicate paleocurrents (Patterns, see Table 1).

like the Upper Santonian, divided into three units related to different paleogeographic situations. The Southeast and Northwest are dominated by clastic sedimentation, still mostly marine, with abundant sigmoidal cross-bedding, borrows and other features typical of shallow subtidal to intertidal environments. Between these regions still persists a small bay with carbonate production on a subtidal platform, although strongly influenced by clastic input. Towards the West the third unit can be seen, illustrating a change in facies to a dolomite-dominated environment. The dolostones are always rich in detrital quartz (1 - 10 %). The presence of miliolid foraminifera (Pl. 3, Fig. D) nevertheless indicates marine influences during the Lower Campanian within the third unit.

Around Villaescusa de las Torres the dolomite is almost free of clastic components, probably because a morphologic barrier inhibited the supply of the detritus from the northern and eastern delta. This barrier was active at least until the Upper Campanian and Lower Maastrichtian: during this time the region around Villaescusa again experienced intensive subtidal carbonate production, which is indicated by the presence of slightly marly limestones with abundant fragments of oysters and other marine fauna (such as miliolids and scarce rudists).

Tilted block movements are responsible for the persistance of fully marine (although very shallow) conditions during the Lower and Upper Campanian in the vicinity of Villaescusa de las Torres. When the paleogeographic sketch (Text-Fig. 8) is compared with the framework of the fault system (Text-Fig. 4), the role of the tilted blocks becomes evident. These blocks, formed by the two main directions of faulting, acted as locally effective sedimentary traps or baffles.



Text-Fig. 8. Paleogeographic distribution of facies during Upper Campanian. Arrows indicate paleocurrents (Patterns, see Table 1).

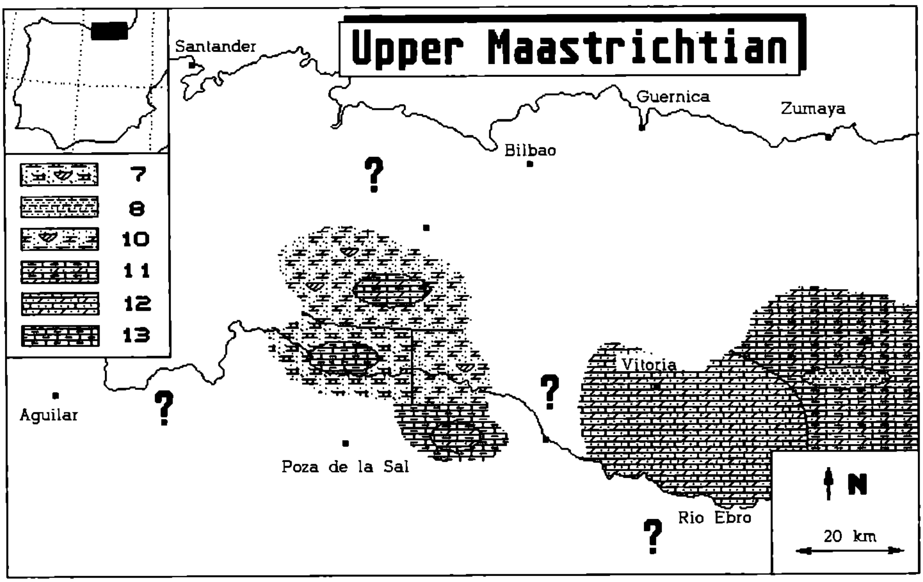
4.3 Upper Campanian

In this period the deltas play the major role. Almost the entire area studied is submitted to clastic influence, with the region neighbouring Villaescusa again as an exception. Pl. 1, Figs. C and D display distinct features of the facies. These features are already explained above and allow interpretation of the facies as a lateral transition, from South to North (first delta around Vitoria) or West to East (second delta between Reinosa and Villasana de Mena), beginning with continental braided river deposits and leading into a fluvial dominated delta front regime, which further to the North and East grades into a carbonate dominated shallow subtidal environment.

Pl. 2, Fig. A shows reworked rudist colonies from the Quintanaloma section, where the clastic influence is manifested only in the high amount of quartz detritus within a bedded limestone/marlstone sequence. The bed thickness ranges from 20 to 100 cm. The fauna is highly diverse (lamellibranchiates, rudists, corals, foraminifera, sponges, gastropods), which rules out the assumption of any kind of restricted or hypo/hypersaline conditions for the Upper Campanian sites located between the two deltas.

4.4 Upper Maastrichtian

For the Maastrichtian, especially the Upper Maastrichtian, problems arise not only because of the scarcity of a biostratigraphically relevant fauna (see also LEPPIG 1987), but also because of rare outcrops. The Lower Maastrichtian displays a facies distribution which is similar to the Upper



Text-Fig. 9. Paleogeographic distribution of facies during Upper Maastrichtian (Patterns, see Table 1).

Campanian. However, the deltas suffer a remarkable reduction of sediment supply and therefore facies change back to coastal and lagoonal carbonate producing settings.

During Upper Maastrichtian sedimentary history, the central region of the investigated area is governed by "Garumnium Facies", with features as described above (see also FLOQUET et al. 1982). Only rare marine ingressions can be noted. They appear as thin (1 - 2 m) limestone beds with a small lateral continuity, containing reworked rudist tests and miliolid foraminifera, as *Biloculina* sp. and *Quinqueloculina* sp. Towards the East, around Vitoria and Alsasua, marine conditions seem to be stable during this period. At the Torme section a highlight is a bed of 2 m thickness within the "Garumnium" and related facies, where Upper Maastrichtian orbitoids are preserved (*Siderolites calcitrapoides*, Pl. 3, Fig. B). This allows the conclusion that the shoreline was situated approximately along the line Vitoria - Villasana de Mena.

The sea-level changes responsible for the small-scale variations in the lithology of the Upper Maastrichtian generally cannot be correlated with the third order cycles established by HAQ et al. (1987). This has to be explained by the complicated interaction of the rifting in the Bay of Biscay and the rotation and translocation of the Iberian Plate.

5. Conclusions

1. The Basco-Cantabrian Basin, which formed as a consequence of the rifting of the Bay of Biscay during Triassic through Tertiary times is bordered on the South by a wide (40 - 60 km) belt of shallow marine to continental sediments.
2. This sedimentary belt suffered important lateral fluctuations through time. After a phase of carbonate platform sedimentation during Upper Santonian, the most important event was the development of two Gilbert-type delta systems during Upper Campanian.
3. The first delta had a general direction of discharge from Southwest to South and is situated around Vitoria; the second delta was established between Reinosa and Villasana de Mena, with a discharge direction from West to Northwest.
4. The Maastrichtian was governed by lagoonal to sabkha environments with short periods of marine flooding.
5. The sedimentary evolution of the southern Basco-Cantabrian Basin was controlled by the movement of a system of tilted blocks, dividing it into several subunits with a very distinct history. Salt diapirism plays an additional role.
6. The tilted blocks were the result of the intensive fracturing of the northern margin of the Iberian Peninsula, forming a rhombic network: listric faults parallel to the basin's main axis (WNW - ESE), cut by more or less southwest to northeast oriented normal faults. Additional tectonic activation of the tilted blocks is related to the anticlock-wise rotation and drift of the Iberian Microplate.
7. The rhythm of sea-level changes in the southern Basco-Cantabrian Basin during late Cretaceous coincides with worldwide events only in cycles of second order.

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Turonian integrated Biostratigraphy in the Estella Basin (Navarra, Spain)

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With 4 Text-Figures

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Abstract. This work deals with a simultaneous study of the planktonic foraminifera, inoceramid, and ammonite assemblages from the Ganuza and Ollogoyen sections (Estella Basin, Navarra, Spain) of Upper Cenomanian-Lower Coniacian ages. Almost all the macrofaunas have been found in the Lower Turonian Nodosoides Zone of Ganuza. This zone is characterized by the ammonite assemblage of *Mammites nodosoides* (SCHLÜTER), *Fagesia tevesthensis* PERON and *Choffaticeras pavillieri* PERVINQUIERE, while the inoceramids are represented from the base to the top of the section by the following succession of species: *Inoceramus (Mytiloides)* aff. *hercynicus* (PETRASCHECK), *I. (M.) submytiloides* (SEITZ)?, *I. (M.) goppelnensis goppelnensis* (BADILLET & SORNAY), *I. (M.) goppelnensis* (BADILLET & SORNAY) n. ssp.?, *I. (M.) mytiloides* (MANTELL), *I. (M.) transiens* (SEITZ) and *I. (M.) hercynicus* (PETRASCHECK). The Ganuza Middle Turonian macrofaunas and the Ollogoyen Upper Turonian ones are scarce and their vertical record is discontinuous. We find *Romaniceras kallesi* (ZÁZVORKA) and *Inoceramus (Mytiloides)* cf. *hercynicus* (PETRASCHECK) in the lower part of the Woollgari Zone, and *Romaniceras ornatissimum* (STOLICZKA) in the upper part. *Subprionocyclus neptuni* (GEINITZ) appears in the Upper Turonian.

Microfaunas collected continuously through the sections record Upper Cenomanian through Lower Coniacian planktonic foraminiferal zones.

Kurzfassung. Im Vergleich werden die Plankton-Foraminiferen, die Inoceramen und die Ammoniten des Obercenoman bis Unterconiac in den Profilen Ganuza und Ollogoyen (Estella-Becken, Navarra) studiert. Die Masse der Makrofauna, sowohl was die Zahl der Arten als auch die der Individuen angeht, konzentriert sich auf das Unterturon, u. zw. die Nodosoides-Zone von Ganuza. Das Vorkommen von *Mammites nodosoides* (SCHLÜTER), *Fagesia* aff. *tevesthensis* PERON und *Choffaticeras pavillieri* PERVINQUIERE spricht für eine Einstufung in diese Zone. Die Inoceramiden treten in folgender Abfolge auf (von unten nach oben): *Inoceramus (Mytiloides)* aff. *hercynicus* (PETRASCHECK), *Inoceramus (Mytiloides) submytiloides* (SEITZ)?, *Inoceramus (Mytiloides) goppelnensis goppelnensis* (BADILLET & SORNAY), *Inocera-*

mus (Mytiloides) goppelnensis (BADILLET & SORNAY) n. ssp.?, *Inoceramus (Mytiloides) mytiloides* (MANTELL), *Inoceramus (Mytiloides) transiens* (SEITZ) und *Inoceramus (Mytiloides) hercynicus* (PETRASCHECK).

Die Makrofaunen des Mittelturon von Ganuza und des Oberturon von Ollogoyen sind selten und zeigen eine spezifische Verteilung. Man findet *Romaniceras kallesi* (ZAZVORKA) und *Inoceramus (Mytiloides)* cf. *hercynicus* (PETRASCHECK) im unteren Teil der Woollgari-Zone, *Romaniceras ornatissimum* (STOLICZKA) im oberen Teil. *Subprionocyclus neptuni* (GEINITZ) kommt im Oberturon vor. Die planktonischen Foraminiferen des Zeitraums Obercenoman bis Unterconiac werden beschrieben.

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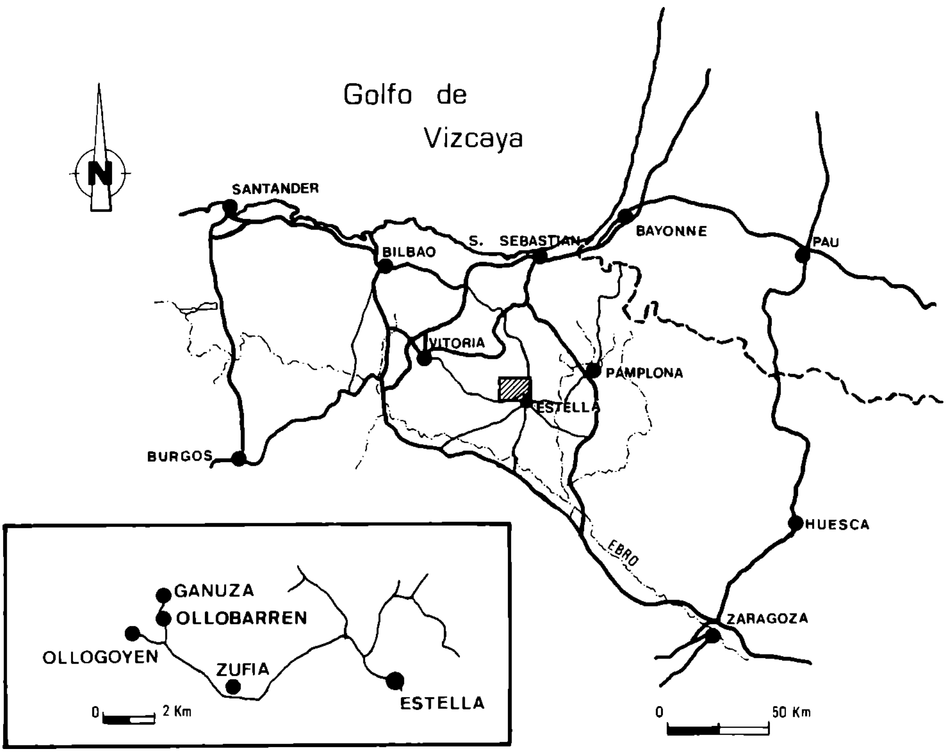
1. Introduction

The vast outcrops that surround the villages of Ganuza and Ollogoyen (Text-Fig. 1), NW of Estella, belong to the eastern part of the Navarro-



Text-Fig. 1. Location of the studied area showing the principal villages of the region.

Cantabrian Trough. They consist of basinal pelagic facies of alternating grey to blue marls and marly limestones with some interbedded calcarenitic levels. Sedimentation is continuous from the Cenomanian until the Lower Coniacian based on the complete record of established faunal zones.

These sediments yield abundant and regularly distributed planktonic and benthonic foraminifera in a variable ratio along the sections.

Among the macrofaunas with biostratigraphic significance we find in these sections ammonites and inoceramids, although they are abundant in some beds, their record is discontinuous.

The collection, bed by bed, of foraminifera, inoceramids and ammonites from the Ganuza and Ollogoyen sections (Text-Fig. 1) has allowed us to identify established planktonic foraminiferal zones from the Upper Cenomanian through the Lower Coniacian, as well as the Turonian ammonite zones. This has facilitated the interrelation between the two zonations at these localities.

The original work of COLOM (1952) in the Middle Cretaceous of the Allin Valley of North Spain pointed out the change of littoral facies in the Cenomanian to a pelagic one in the Lower Turonian. Later, FEUILLÉE (1967) restudied the foraminifera of this area in his work on the Basque-Cantabrian Cenomanian.

WIEDMANN & KAUFFMAN (1978) suggested an inoceramid biozonation for North Spain based on a material collected by WIEDMANN. Also in this work they proposed a correlation with the ammonites, studied long before, as well as another correlation with the foraminifera (mainly planktonic).

Afterwards, LAMOLDA (1981) suggested a Turonian biostratigraphic revision, in which all prior data on ammonites, inoceramids, echinoids, foraminifera and ostracods were compiled.

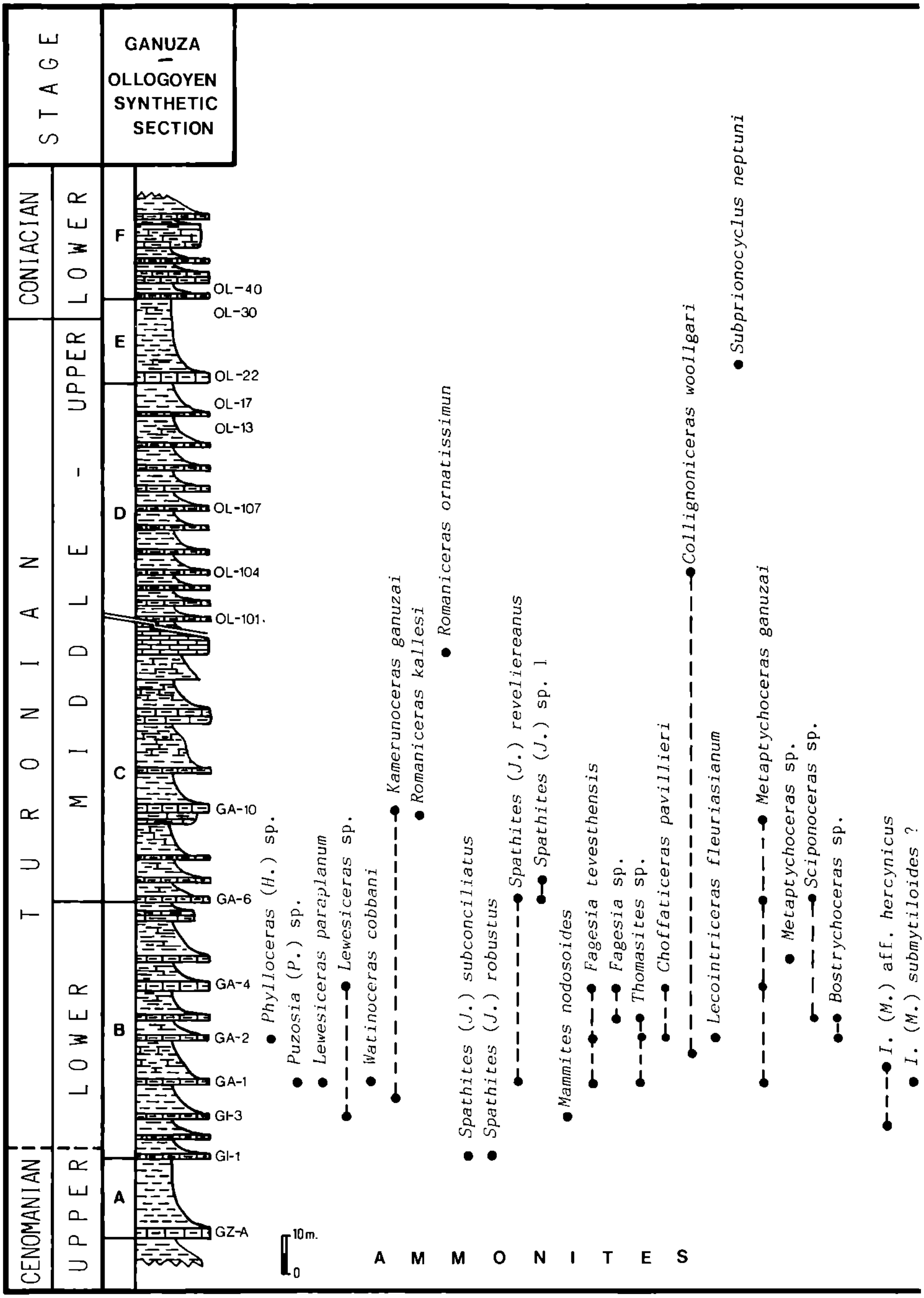
WIEDMANN (1980) and LAMOLDA et al. (1981) complete the study on macro- and microfaunas of this area. Two other recent works are: SCHWENTKE & WIEDMANN (1985) on Cretaceous regional geology and on the evolution of the Estella diapir, and LAMOLDA & PROTO-DECIMA (1986) on the Turonian-Coniacian passage, in Ollogoyen.

It should be noted that the macrofaunas in these outcrops were much more abundant in the past than they are now, due to intense collecting by geologists and private collectors. This fact has made it difficult to determine the precise location of some of the boundaries and events.

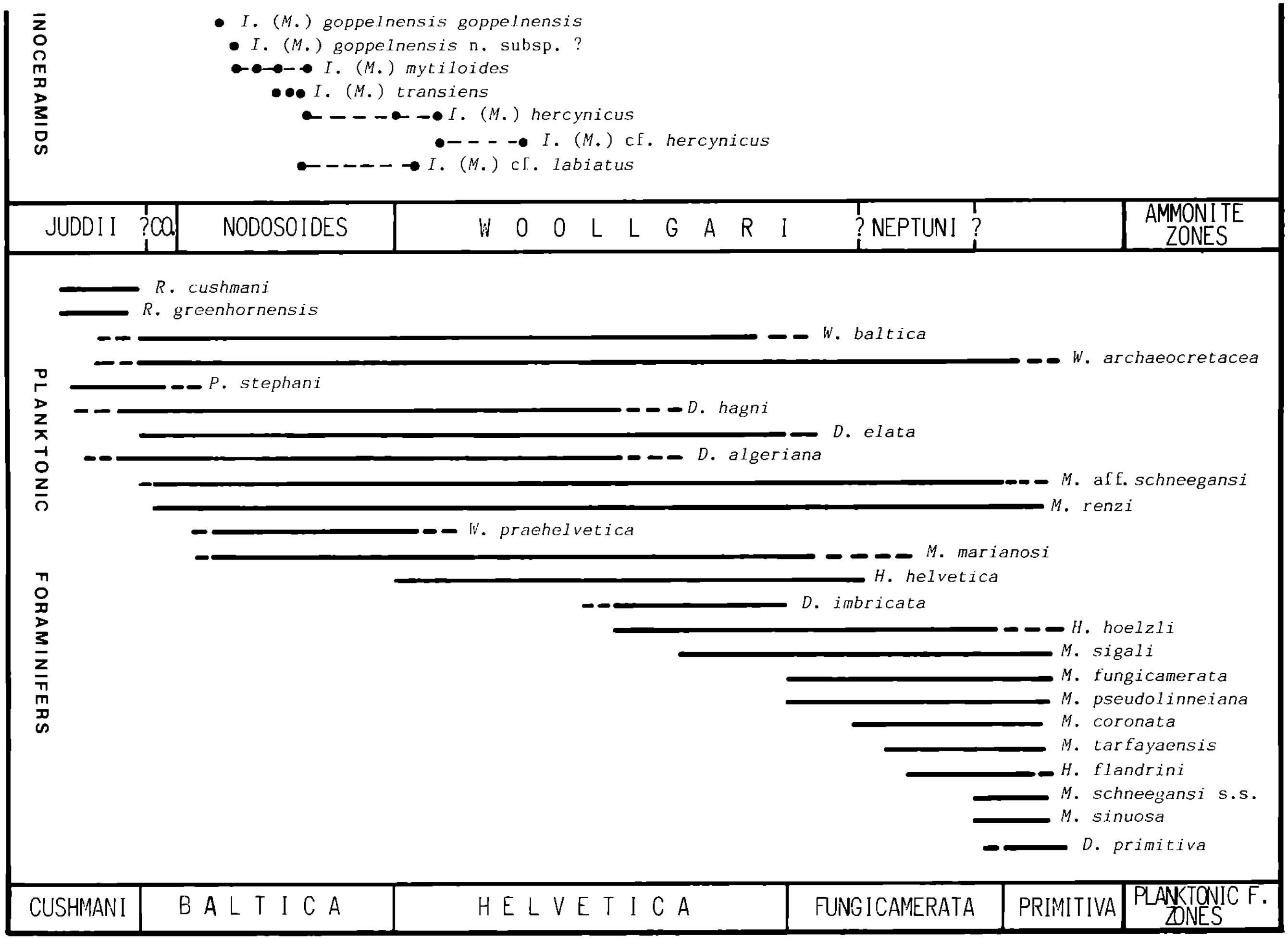
2. Stratigraphy

The Ganuza section includes the Cenomanian-Turonian boundary and the Lower and the Middle Turonian. At Ollogoyen the Middle and Upper Turonian are present. Text-Fig. 2 shows a composite section. Lithologically both sections are constituted by grey to blue coloured marls, with interbedded white marly limestones, 1 to 2 m thick. The Ganuza section was measured 400 m south of the village, beginning on the western side of the Ollobarren-Ganuza road and going to the west (Text-Fig. 1).

With a thickness of 156 m, three lithological intervals (A, B, C) have been identified. (A) has a thickness of 20 m. It consists of marls limited by limestone beds, 1 m thick. These beds contain the top of the Cushmani Zone, but the Cenomanian-Turonian boundary, based on planktonic foraminiferal assemblages, is found about 2 m above it.



Text-Fig. 2. Composite section and zonal distribution of the ammonites, inoceramids and planktonic foraminifera for the Ganuza-Ollogoyen area.


INOCERAMIDS
I. (M.) goppelnensis goppelnensis
I. (M.) goppelnensis n. subsp. ?
I. (M.) mytiloides
I. (M.) transiens
I. (M.) hercynicus
I. (M.) cf. hercynicus
I. (M.) cf. labiatus
JUDDII
?CO.
NODOSOIDES
W O O L L G A R I
? NEPTUNI ?
AMMONITE ZONES
PLANKTONIC FORAMINIFERS
R. cushmani
R. greenhornensis
W. baltica
W. archaeocretacea
P. stephani
D. hagni
D. elata
D. algeriana
M. aff. schneegansi
M. renzi
W. praehelvetica
M. marianosi
H. helvetica
D. imbricata
H. hoelzli
M. sigali
M. fungicamerata
M. pseudolinneiana
M. coronata
M. tarfayaensis
H. flandrini
M. schneegansi s.s.
M. sinuosa
D. primitiva
CUSHMANI
B A L T I C A
H E L V E T I C A
FUNGICAMERATA
PRIMITIVA
PLANKTONIC F. ZONES

(B) has a thickness of 65 m (Text-Fig. 2). The marls, some more calcareous than in (A), alternate with marly limestones, 0.5 to 1 m thick. The interval is Lower Turonian and probably Uppermost Cenomanian in age; the Baltica foraminiferal zone and Coloradoense and Nodosoides ammonite zones have been identified in it. The boundary between the two ammonite zones lies about 10 m above the base of the interval.

(C) has a thickness of 71 m. It consists also of marl units, 10 m thick and interbedded marly limestones. Towards the top there is an increase of marly limestone beds, becoming much more calcareous at the top. The lower boundaries of the Helvetica foraminiferal zone and the Woollgari ammonite zone have been identified in the interval (Text-Fig. 2).

The Ollogoyen section has been measured 300 m west of the village. It has a total thickness of 107 m distributed in three lithological intervals (D, E, F).

(D) has a thickness of 50 m (Text-Fig. 2). It consists of interbedded marls and marly limestones, 0.15 to 0.30 m thick. In this interval we find the upper part of the Helvetica and Woollgari zones and the lower part of the Fungicamerata foraminiferal zone and the Neptuni ammonite zone. The interval A of LAMOLDA & PROTO-DECIMA (1986) fits in the last 21 m of (D).

(E) has a thickness of 21 m (Text-Fig. 2). The basal 2 m of this interval are constituted by nodular limestones and marly limestones (in it WIEDMANN, 1980, places the Turonian-Coniacian boundary). The remainder are marls and marly limestones alternating with micritic limestones, 0.15 to 0.30 m thick (B and C intervals of LAMOLDA & PROTO-DECIMA 1986).

(F) is 25 m thick. Marly limestones and calcarenites are predominant over marls. These beds correspond to interval D of LAMOLDA & PROTO-DECIMA (1986) where the planktonic foraminifera Primitiva Zone of Lower Coniacian age was identified by these authors.

3. Planktonic foraminifera

In general, the microfaunal assemblages are rich and well preserved. The planktonic foraminifera ratio (planktonic forams/total forams) changes along the succession with a maximum at the Cenomanian-Turonian boundary (ratio 80 - 90 %). There is another maximum (ratios 60 - 70 %) in the middle part of the Upper Turonian, similar to that of the Upper Cenomanian. The smallest ratio of planktonic foraminifera occurs after the extinction of *Rotalipora* ssp. and in the Lower and Middle Turonian. The agglutinated benthonic foraminifera are dominant over the calcareous benthonic forms.

These associations have made possible the identification of biozones found in other localities of the Basque-Cantabrian area (LAMOLDA 1982), including Cushmani, Baltica, Helvetica, Fungicamerata and Primitiva zones.

Cushmani Zone: In this area the uppermost part of the zone is found, identified in the basal 20 m of the studied sections.

The assemblage is dominated by planktonic foraminifera (70 - 80 %), and among them by species of *Rotalipora*: *R. cushmani* (MORROW), *R. greenhornensis* (MORROW), *R. montsalvensis* MORNOD. Other species are: *Hedbergella delrioensis* (CARSEY), *H. portsdawnensis* (WILLIAMS-MITCHELL), *Whiteinella archaeocretacea* PESSAGNO, *Praeglobotruncana stephani* (GANDOLFI), *Dicarinella algeriana* (CARON), *D. hagni* (SCHEIBNEROVA). Species of *Tritaxia* and *Gaudryina* are the most common among the benthonic foraminifera assemblages, together with some Nodosariids and some species of *Gavelinella* and *Lingulogavelinella*.

Baltica Zone (= Archaeocretacea Zone): The extinction of *Rotalipora* species reduces the ratio of planktonic foraminifera to 60 % at the base of the zone. Then the abundance of species of *Whiteinella* raises this ratio to the 95 %. *Whiteinella* species include: *W. baltica* DOUGLAS & RANKIN, *W. aprica* (LOEBLICH & TAPPAN), *W. paradubia* (SIGAL), *W. archaeocretacea* PESSAGNO. This last species together with *Dicarinella elata* LAMOLDA, *D. hagni* (SCHEIBNEROVA) and *D. algeriana* (CARON) make up the basal assemblage of planktonic foraminifera which can be followed along this zone. Moreover, we find also *Praeglobotruncana stephani* (GANDOLFI), *P. gibba* KLAUS, *Dicarinella hilalensis* (BARR), *Marginotruncana* aff. *schneegansi* (SIGAL), *M. renzi* (GANDOLFI), etc. Although the genera *Gaudryina* and *Lenticulina* are found, it is the *Lingulogavelinella* and *Gavelinopsis* assemblage that are most typical for the lower half of the zone. When, in the upper half, *Whiteinella praehelvetica* (TRUJILLO), and *Marginotruncana marianosi* (DOUGLAS) are present, there is a ratio decline for the planktonic foraminifera (75 - 40 %), very similar to the one observed in the Menoyo area (LAMOLDA 1978); *Gavelinopsis* still persists in a similar ratio, while there is an agglutinated diversification with a higher ratio of *Ammobaculites* and *Placopsilina* species, although *Gaudryina* and *Tritaxia* are still prevailing.

Helvetica Zone: It has a thickness of about 100 m from the sample GA-6 at the Ganuza section to the sample OL-106/107 at the Ollogoyen section. The planktonic foraminifera constitute between 50 and 20 % of the total of foraminifera in this zone.

In addition to previously mentioned planktonic species *Helvetoglobotruncana helvetica* (BOLLI), *Hedbergella hoelzli* (HAGN & ZEIL), *Dicarinella imbricata* (MORNOD), *Marginotruncana sigali* (REICHEL) and *M. pileoliformis* LAMOLDA make up the complement of the Helvetica Zone. *M. sigali* and *M. pileoliformis* occur in the upper part, preceded by *D. imbricata* and *H. hoelzli*. While *H. helvetica* becomes dominant upward a decrease of the number and size of the *Whiteinella* species is observed. The benthonic foraminifera have, by now, a great diversity, mainly the agglutinated foraminifera mentioned in the Baltica Zone. The addition of the genera *Arenobulimina*, *Gavelinella*, and in the upper part *Recurvoides*, to the previously mentioned taxa constitutes the benthonic foraminiferal assemblage that is found in all higher zones of this section.

Fungicamerata Zone: Only found at Ollogoyen, it has a thickness of 57 m from the sample OL-106/107 to the OL-40. There is a rapid increase in the planktonic ratio up to values of 70 % with the addition of some *Marginotruncana* species. This ratio will fall sometimes to 50 % within the zone and decrease upward towards the boundary.

The lower half of the zone has an assemblage of planktonic species similar to that of the top at the Helvetica Zone, but with *Marginotruncana pseudolinneiana* PESSAGNO and *M. fungicamerata* (MARTIROSJAN). Upward, the species *H. helvetica* disappears and the species of the genus *Marginotruncana* dominate over the hedbergelliforms. Of these, a decrease in number of individuals and species is observed. At this level, *M. coronata* (BOLLI) and *M. tarfayaensis* (LEHMANN) are abundant. In the upper part *Marginotruncana paraconcavata* PORTHAULT, *M. undulata* (LEHMANN), *M. scorpionis* LAMOLDA and *Hedbergella flandrini* PORTHAULT appear successively. They are considered to indicate the proximity of the Turonian-Coniacian boundary. Following the suggestions of LAMOLDA & PROTO-DECIMA (1986), the first appearance of *Marginotruncana sinuosa* PORTHAULT and *M. schneegansi* s. s. characterize this boundary. The benthonic foraminifera assemblage is highly diversified, containing genera such as *Tritaxia, Dorothia, Arenobulimina* and *Gavelinella*. However, other genera such as *Gyroidinoides* and *Globorotalites* are also common in the upper part of this zone.

Primitiva Zone: The top boundary and upper part of this zone have not been identified, but the last 20 m of the Ollogoyen section belong to it. The ratio of planktonic foraminifera declines even more rapidly from 55 % to 40 %. *Marginotruncana* species prevail together with some globigeriniforms and some heterohelicids. *Dicarinella primitiva* (DALBIEZ) characterizes this zone. The benthonic foraminiferal assemblage is very similar to that seen in the Fungicamerata Zone, but in the upper part of the Ollogoyen section some species of the genus *Quinqueloculina* appear.

4. Ammonites

The outcrops of the uppermost Cenomanian and the lowermost Turonian along the road between Ollobarren and Ganuza have been spoliated of fossils by the collectors during the last two decades, and only few and badly preserved ammonites can now be found. Consequently we cannot recognize, by means of ammonites, the Cenomanian-Turonian boundary in these outcrops. However, the Lower Turonian (Ganuza section) is still rich in ammonites, though the record is discontinuous. The highest number of species and individuals is found in the calcareous beds. They appear generally deformed by lateral and equatorial compression and are not very well preserved. The Middle Turonian, also in Ganuza, is not so rich and the record is even more discontinuous. The Upper Turonian and the Lower Coniacian at the

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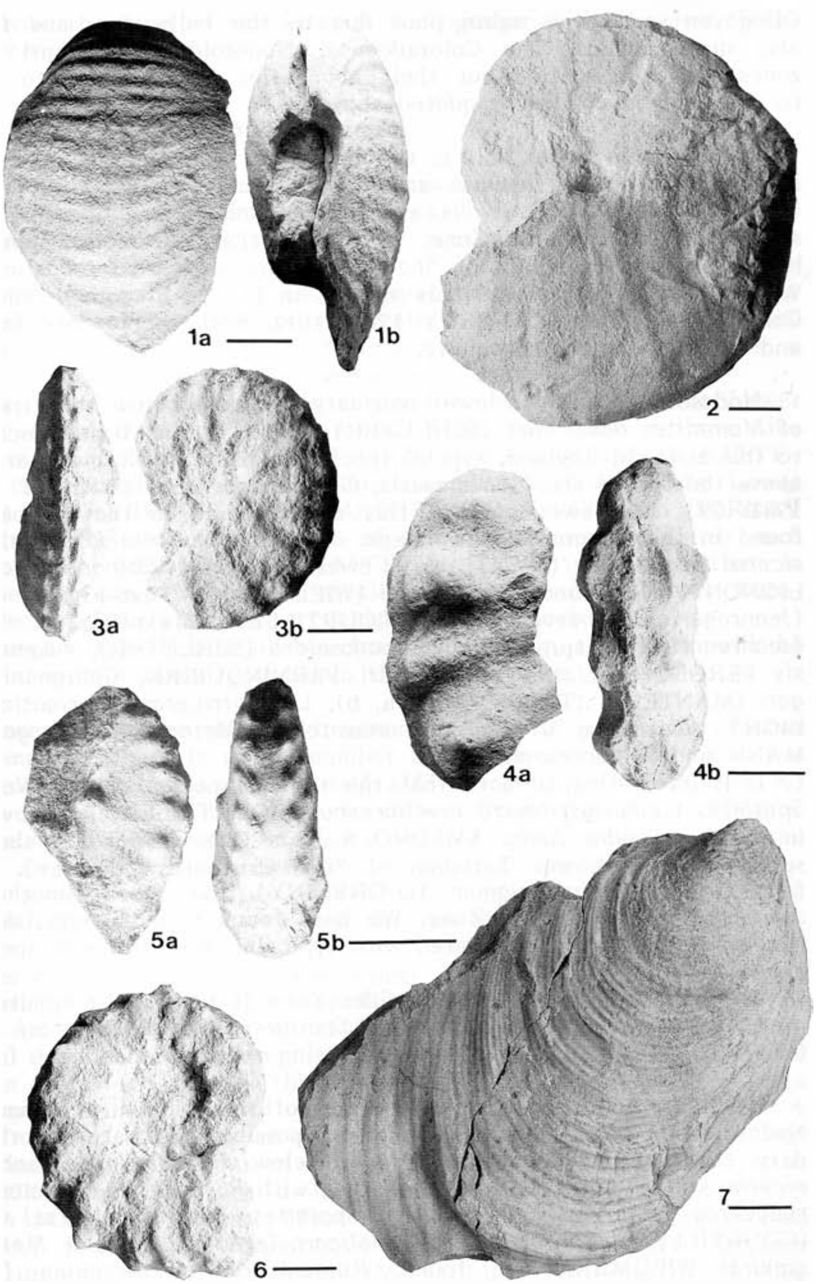
Text-Fig. 3. Some representative species of ammonites and inoceramids from Ganuza-Ollogoyen area. 1a, b: *Fagesia* sp. aff. *F. tevesthensis* PERON, ventral and lateral view, Lower Turonian of Ganuza. PAL.UAB. 12107. 2: *Choffaticeras pavillieri* PERVINQUIERE, lateral view, Lower Turonian of Ganuza. PAL.UAB. 12122. 3a, b: *Lecointriceras fleuriausianum* (D'ORBIGNY), lateral and ventral view, Lower Turonian of Ganuza. PAL.UAB. 12116. 4a, b: *Collignoniceras woollgari* (MANTELL), lateral and ventral view, Middle Turonian of Ollogoyen. PAL.UAB. 12124. 5: *Spathites (J.) reveliereanus* (COURTILLER) from the uppermost Lower Turonian of Ganuza. PAL.UAB. 12071. 6: *Kamerunoceras ganuzai* (WIEDMANN), lateral view, upper part of the Lower Turonian of Ganuza, PAL.UAB. 12051. 7: *Inoceramus (Mytiloides) mytiloides* (MANTELL), right valve, Lower Turonian of Ganuza. PAL.UAB. 38222. Bar = 10 mm. All specimens have been whitened with NH_4Cl and are housed in the Department of Geology (Paleontology) of the Universitat Autònoma de Barcelona.



Text-Fig. 3

Ollogoyen section, is again poor due to the collectors, and the record is also discontinuous. The Coloradoense, Nodosoides, Woollgari and Neptuni zones are so identified but their boundaries are difficult to establish due to the outcrop conditions quoted above.

Coloradoense Zone: This is the lowermost zone identified, and it is poor in both, number of species and in individuals. *Spathites (Jeanrogericeras) subconciliatus* (CHOFFAT) is a species found in the Juddii Zone, as well as in the Coloradoense Zone. In this section the Coloradoense Zone has been identified by means of *Spathites (Jeanrogericeras) robustus* WIEDMANN. WIEDMANN (1960) places this species in his Tu-III zone, equivalent to the Coloradoense Zone (KENNEDY 1985: 101). Both species are found scarcely and without vertical continuity.

Nodosoides Zone: Its lower boundary is taken below the first appearance of *Mammites nodosoides* (SCHLÜTER) which is the first species belonging to this zone. In England, typical specimens of *M. nodosoides* are found 2 m above the top of the Plenus marls, Geslinianum Zone (WRIGHT & KENNEDY 1981: 79). The lower part of this zone is rich in the number of species found in the ammonite assemblage, including: *Puzosia (Puzosia)* sp., *Lewesiceras peramplum* (MANTELL), *Lewesiceras* sp., *Watinoceras cobbani* COLLIGNON?, *Kamerunoceras ganuzai* (WIEDMANN) (Text-Fig. 3. 6), *Spathites (Jeanrogericeras) reveliereanus* (COURTILLER) (Text-Fig. 3. 5), *Spathites (Jeanrogericeras)* sp., *Mammites nodosoides* (SCHLÜTER), *Fagesia tevesthensis* PERON, *Choffaticeras pavillieri* PERVINQUIERE, *Collignoniceras woollgari* (MANTELL) (Text-Fig. 3. 4a, b), *Lecointriceras fleuriausianum* (D'ORBIGNY) (Text-Fig. 3. 3a, b), *Thomasites* sp., *Metaptychoceras ganuzai* WIEDMANN and *Sciponoceras* sp.

It is interesting to note that the typical species of the Woollgari Zone, *Spathites (Jeanrogericeras) reveliereanus* (COURTILLER) has now been found in the Nodosoides Zone. AMEDRO & HANCOCK (1986: 17) also found this species in the Lower Turonian of "Les Charentes" (France). Up to now, *Lecointriceras fleuriausianum* (D'ORBIGNY) has been another exclusive species of the Woollgari Zone. We have found it in Ganuza, down into the Nodosoides Zone and associated with typical Lower Turonian species as seen above.

The upper part of the Nodosoides Zone is poor in ammonites, with *Kamerunoceras ganuzai* (WIEDMANN), *Spathites (Jeanrogericeras) reveliereanus* (COURTILLER) and *Sciponoceras* sp. being the only specimens found.

Woollgari Zone: Due to the scarcity of the ammonite faunas around the Nodosoides-Woollgari boundary it is not possible to locate exactly this boundary. However, it will always lay well below the first appearance of *Romaniceras kallesi* (ZAZVORKA). Together with *R. kallesi*, specimens of *Kamerunoceras ganuzai* (WIEDMANN), *Spathites (Jeanrogericeras) reveliereanus* (COURTILLER), *Collignoniceras woollgari* (MANTELL), and *Metaptychoceras ganuzai* WIEDMANN are found. *Romaniceras ornatissimum* (STOLICZKA) appears together with *Collignoniceras woollgari* (MANTELL) about 40 m above the first appearance of *Romaniceras kallesi* (ZAZVORKA).

Neptuni Zone: This zone was identified at the Ollogoyen section, which is even poorer in macrofaunas than Ganuza. The lower boundary must be located below the sole specimen of *Subprionocyclus neptuni* (GEINITZ) found 2 m above the base of the interval E, together with the last fragments of *Romaniceras* spp.

5. Inoceramids

All the inoceramid faunas were found in Ganuza outcrops. The Lower Turonian contains abundant specimens, but with a low species diversity. The Middle Turonian is low in both, number of specimens as well as species diversity.

One hundred and sixty well preserved specimens and more than twice that number of broken ones have been classified. Identifications were based on their general shape and ornamentation. On undeformed specimens a quantification of the ontogenic variation has been made.

Up to now, there is still no confirmation for the validity of the Lower Turonian inoceramid zones at world scale. While for some authors like KAUFFMAN et al. (1978), KAUFFMAN (1977) and WIEDMANN & KAUFFMAN (1978) it is possible to establish inoceramid zonations for the Lower Turonian, for others like SEITZ (1934), TRÖGER (1967) and SORNAY (1982) the coexistence of index species of different zones invalidates such zonations. Moreover, WIEDMANN & KAUFFMAN (1978) point out that in the Ganuza section some inoceramid species considered to be Lower Turonian index species of different zones coexist in the same bed.

For this reason the inoceramid species identified in this work will be fitted in the ammonite zones recorded in the Ganuza area.

Coloradoense Zone: This is the lowest zone where inoceramid faunas have been identified; there is a low number of specimens as well as a low species diversity. *Inoceramus (Mytiloides)* aff. *hercynicus* (PETRASCHECK) (Text-Fig. 4. 5) has been identified. It differs from the PETRASCHECK (1903) species by its different ornamentation and minor growth along the perpendicular direction to the growth axis. *I. (M.)* aff. *hercynicus* (PETRASCHECK) is found associated to *I. (M.)* cf. *goppelnensis* (BADILLET & SORNAY). Both species are located near the top of the Coloradoense Zone.

Nodosoides Zone: *Inoceramus (Mytiloides) submytiloides* (SEITZ)? has been identified at the base of this zone. This species is referred by KAUFFMAN (1977) and KAUFFMAN et al. (1978) as characteristic of the transition between the Cenomanian and Turonian, while KELLER (1982) placed the species in the middle part of the Lower Turonian. There is also a large number of specimens of *I. (M.) goppelnensis goppelnensis* (BADILLET & SORNAY) (Text-Fig. 4. 2). 6 m above the first appearance of this last species we find more specimens of *Inoceramus (Mytiloides)* aff. *hercynicus* (PETRASCHECK), together with a large number of *I. (M.) goppelnensis* (BADILLET & SORNAY) n. ssp.? (Text-Fig. 4. 3). This is a very evolved form of *I. (M.) goppelnensis goppelnensis* (BADILLET & SORNAY) because it shows features of *I. (M.) mytiloides* (MANTELL) (Text-Fig. 3. 7). Coexisting with *I. (M.) goppelnensis* (BADILLET & SORNAY) n. ssp.?, we already find the first record of *I. (M.) mytiloides* (MANTELL), occurring along the middle part of the Nodosoides Zone. *I. (M.) transiens* (SEITZ) appears associated to the last specimen of *I. (M.) mytiloides* (MANTELL).

The assemblage of the upper part of the Nodosoides Zone is constituted by *I. (Mytiloides) hercynicus* (PETRASCHECK) and *I. (M.)* cf. *labiatus* (SCHLOTHEIM).

Woollgari Zone: This zone is poorly represented in number of species and specimens. For the lower part of this zone we identify the same assemblage as for the upper Nodosoides Zone, that is *Inoceramus (Mytiloides) hercynicus* (PETRASCHECK) and *I. (M.)* cf. *labiatus* (SCHLOTHEIM). The beds above the last appearance of *I. (M.) hercynicus* (PETRASCHECK) have supplied badly preserved specimens of *I. (M.)* cf. *hercynicus* (PETRASCHECK).

6. Conclusions

By means of the macrofauna recently collected by the authors, it is not possible to identify the Juddii-Coloradoense boundary zones, nor the Cenomanian-Turonian boundary. According to the planktonic foraminifera assemblage, this boundary lies about 2 m above the last appearance of *Rotalipora cushmani* and *Rotalipora greenhornensis* and thus, also above the Cushmani-Baltica zone boundary.

The base of the Nodosoides Zone will lay below the first appearance of *Mammites nodosoides* and also below *Inoceramus (Mytiloides) goppelnensis goppelnensis*. *Spathites (J.) reveliereanus* and *Lecointriceras fleuriausianum* have been found in the Nodosoides Zone, and so it is not possible to use them as indices for the Woollgari Zone.

From the base to the top of the Nodosoides Zone the following inoceramid succession has been identified: *Inoceramus (Mytiloides) submytiloides?*, *I. (M.) goppelnensis goppelnensis*, *I. (M.) goppelnensis* n. ssp.?, *I. (M.) mytiloides*, *I. (M.) transiens*, *I. (M.) hercynicus*.

The base of the Woollgari Zone is not well defined by means of the macrofauna, but will always lay below *Romaniceras kallesi*. This base seems to coincide with the base of the Helvetica Zone, defined by the first appearance of the species *Helvetoglobotruncana helvetica*.

The boundary between the Woollgari Zone and the Neptuni Zone is not traceable due to the scarcity of the ammonite faunas. The first appearance of *Marginotruncana fungicamerata* and *M. pseudolinneiana* characterizes the lower boundary of the Fungicamerata Zone, which lies in the Woollgari Zone. The last occurrence of *Helvetoglobotruncana helvetica* is near and below the known occurrence of *Subprionocyclus neptuni*, in accordance to the statement of LAMOLDA (1981). The macrofauna does not allow us to define the Turonian-Coniacian boundary, however the first appearances of

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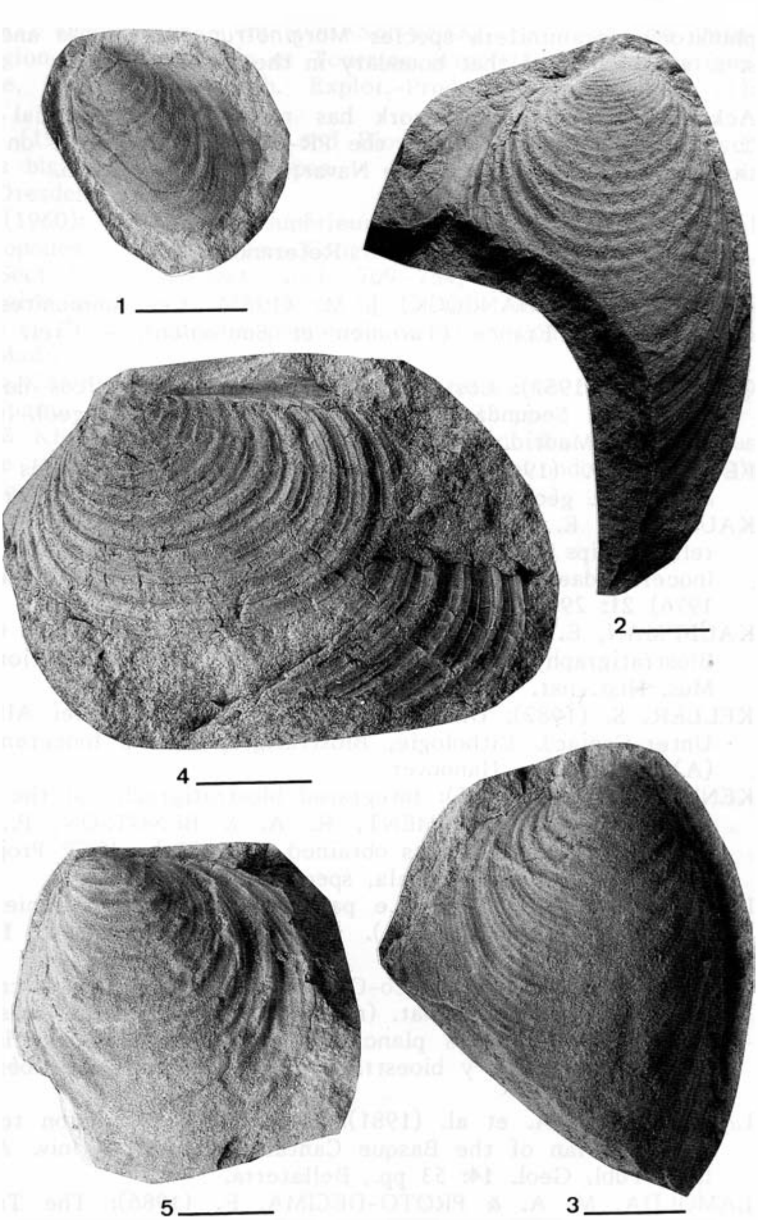
Text-Fig. 4. Some representative species of inoceramids from Ganuza area. 1: *Inoceramus (Mytiloides) submytiloides* (SEITZ)?, left valve, Lower Turonian. PAL.UAB. 38711. 2: *Inoceramus (Mytiloides) goppelnensis goppelnensis* (BADILLET & SORNAY), right valve, Lower Turonian. PAL.UAB. 38847. 3: *Inoceramus (Mytiloides) goppelnensis* (BADILLET & SORNAY) n. ssp.?, right valve, Lower Turonian. PAL.UAB. 38218. 4: *Inoceramus (Mytiloides) hercynicus* (PETRASCHECK), left valve, Middle Turonian. PAL.UAB. 38490. 5: *Inoceramus (Mytiloides)* aff. *hercynicus* (PETRASCHECK), left valve, lowermost Lower Turonian. PAL.UAB. 38764. Bar = 10 mm. All specimens have been whitened with NH_4Cl and are housed in the Department of Geology (Paleontology) of Universitat Autònoma de Barcelona.



Text-Fig. 4

planktonic foraminifera species *Marginotruncana sinuosa* and *M. schneegansi* s. s. are indices of that boundary in the Estella area.

Acknowledgements. This work has received the financial support of the C. A. I. C. Y. T., as a part of the PR-83/2693-C2 project on the Upper Cretaceous Biostratigraphy of the Navarro-Cantabrian Basin.

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Integrated Biostratigraphy of the Turonian-Coniacian Transition Interval in Northern Spain with Comparison to NW Germany

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With 4 Plates and 8 Text-Figures

KÜCHLER, TH. & ERNST, G. (1989): Integrated Biostratigraphy of the Turonian-Coniacian Transition Interval in Northern Spain with Comparison to NW Germany. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 161-190. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The litho- and biostratigraphy of the Upper Turonian and Lower Coniacian were studied in detail in seven selected sections in the eastern Barranca, the Estella Basin (Ollogoyen), and the Santander Basin (Liencres). Within the three structural units, the lithofacies, thicknesses, and faunal associations varied considerably.

The vertical ranges of ammonites, inoceramids, and echinoids, as well as foraminifera in part, are illustrated for each section. Furthermore, biozonal schemes were established using ammonites and inoceramids. Characteristic faunal associations allowed an exact correlation among these separate northern Spanish basins and comparisons with western Europe.

The base of the Upper Turonian is defined by the occurrence of *Romaniceras deverianum*. The Neptuni Zone is verified in northern Spain with large populations of *Subprionocyclus* gr. *neptuni-hitchinensis*. The first record of *Prionocyclus* cf. *germari* and significant inoceramids prove the existence of the Subprionocyclus normalis Zone in the uppermost Upper Turonian. *Romaniceras deverianum* ranges up into the Normalis Zone.

Within the Barranca, inoceramid assemblages of *I. costellatus, I. striatoconcentricus,* resp. *I. (? Cremnoceramus) waltersdorfensis* indicate the Neptuni Zone as well as the Normalis Zone. Interesting for far-reaching correlation is the first record of *Hyphantoceras, Scaphites geinitzii, Micraster (Eomicraster) michelini, Terebratulina* cf. *lata,* and others in the Upper Turonian of Ollogoyen.

The restricted Turonian/Coniacian boundary is discussed with reference to the proposals presented in Copenhagen (1983). In northern Spain this boundary can be best identified by an association containing *Didymotis, I. (? Cr.) waltersdorfensis,* and *I. (Cr.) rotundatus.* Two *Didymotis* events occur in the Santander Basin, just as in Bohemia and NW Germany.

Forresteria occurs first close above the postulated Turonian/Coniacian boundary within the first acme zone of *Micraster* gr. *cortestudinarium.* The uppermost Lower Coniacian can be subdivided based on the succession of

Peroniceras cf. *bajuvaricum* and *P. subtricarinatum*. The former enters together with *I. (Cremnoceramus) deformis*.

The congruity between the stratigraphic subdivisions of northern Spain and NW Germany is discussed.

Kurzfassung: Das Ober-Turon und Unter-Coniac wurde an sieben ausgewählten nordspanischen Profilen litho- und biostratigraphisch eingehender untersucht. Geographisch verteilen sich die Profile auf die E-Barranca, das Estella-Becken (Ollogoyen) and das Santander-Becken (Liencres). Lithofazies, Mächtigkeiten und Faunen-Assoziationen der drei strukturellen Einheiten zeigen beträchtliche Unterschiede.

Die Vertikalreichweiten von Ammoniten, Inoceramen, Echiniden sowie teilweise von Foraminiferen werden für die Einzelprofile aufgezeigt. Anhand von Ammoniten und Inoceramen werden Zonen-Gliederungen entwickelt. Typische Faunen-Assoziationen erlauben eine exakte Korrelation der separaten nordspanischen Becken und einen Vergleich mit dem westlichen Europa.

Die Basis des Ober-Turon wird mit *Romaniceras deverianum* gezogen. Die Neptuni-Zone wird erstmalig in N-Spanien durch reiche Populationen von *Subprionocyclus* gr. *neptuni/hitchinensis* verifiziert. Der Erstnachweis von *Prionocyclus* cf. *germari* und signifikanten Inoceramen weisen auf das Vorhandensein der Subprionocyclus normalis-Zone im höchsten Ober-Turon hin. *Romaniceras deverianum* reicht bis in die Normalis-Zone. In der Barranca indizieren die Inoceramen-Assoziationen von *I. costellatus, I. striatoconcentricus* resp. *I. (? Cremnoceramus) waltersdorfensis* und *I. websteri* sowohl die Neptuni- als auch die Normalis-Zone. Überregionales korrelatives Interesse besitzen ferner die Erstnachweise von *Hyphantoceras, Scaphites geinitzii, Micraster (Eomicraster) michelini* und *Terebratulina* cf. *lata* im Ober-Turon von Ollogoyen.

Der engere Turon/Coniac-Grenzbereich wird im Sinne der in Kopenhagen (1983) gegebenen Vorschläge diskutiert. In N-Spanien kann die Grenze am klarsten mit einer Assoziation von *Didymotis, I. (? Cr.) waltersdorfensis* und *I. (Cr.) rotundatus* gezogen werden. Im Santander-Becken sind, ebenso wie in Böhmen und NW-Deutschland, zwei *Didymotis*-Events nachweisbar. *Forresteria* tritt dicht über der postulierten Grenze innerhalb einer ersten Akme-Zone von *Micraster* gr. *cortestudinarium* auf. Das höhere Unter-Coniac kann durch die Abfolge *Peroniceras* cf. *bajuvaricum* und *P. subtricarinatum* untergliedert werden. Ersterer setzt gemeinsam mit *I. (Cremnoceramus) deformis* ein.

Die Übereinstimmungen der Gliederungsschemata von N-Spanien und NW-Deutschland werden diskutiert.

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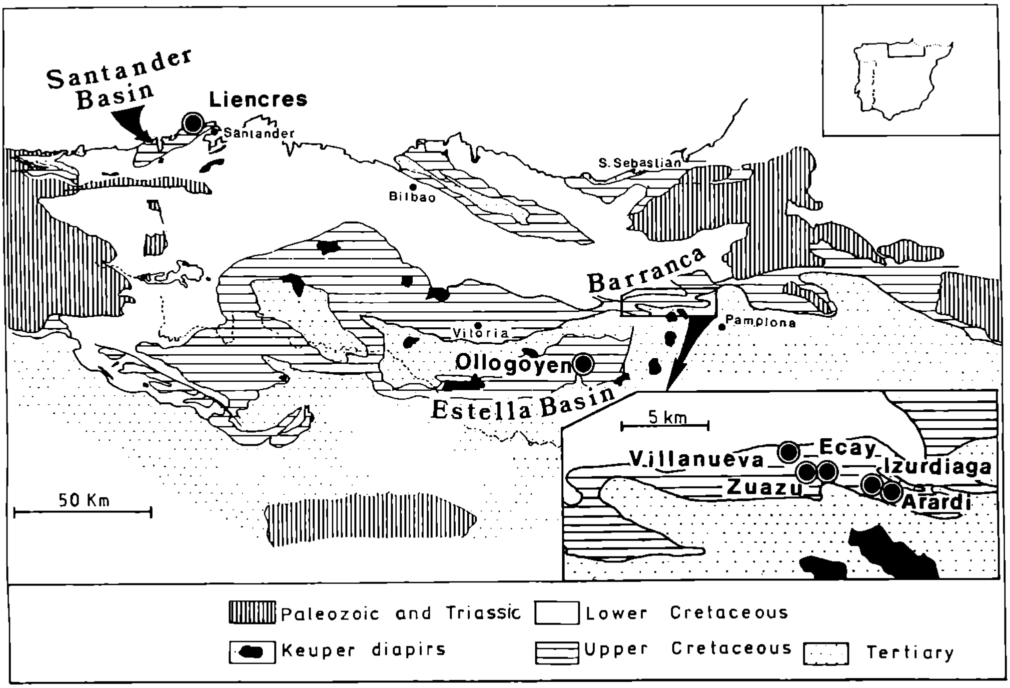
1. Introduction (Text-Fig. 1)

The stratigraphical topic described here is only a part of a larger project by the Berlin Working Group on the Upper Cretaceous of northern Spain (DEGENHARDT 1983, KÜCHLER 1983, THEUERKAUFF 1987, WOLZ 1985, ZANDER 1988, and others). In this paper, the Upper Turonian and Lower Coniacian with respect to the boundary interval is presented.



Text-Fig. 1. Locality map of the sections discussed in this paper (modified after GARCIA MONDEJAR & PUJALTE 1982).

Starting point of the investigation was the eastern part of the so-called Barranca in the Basque Province of Navarra. The faunal spectrum in this area is comparatively rich in stratigraphically important ammonites, inoceramids, and echinoids. Comparable sections were investigated in the western Barranca near Alsasua and in the eastern Vitoria Basin near the localities Eguino and Guevara. Outcrops in the latter areas are rather poor in stratigraphically useful macrofossils.

The correlation was continued to the northwest along strike direction of the Navarro-Cantabrian Basin (FEUILLÉE 1967) (= Vascogotic Trough sensu WIEDMANN & KAUFFMAN 1978) from the Barranca to the Santander Basin with the fossiliferous cliffs of Liencres (THEUERKAUFF 1987). In addition, the traditional Turonian to Coniacian standard section of Ollogoyen in the Estella Basin was critically examined.

The lithofacies, as well as the thicknesses of the different sections, are rather variable. In the eastern Barranca, the predominantly carbonate successions of the Upper Turonian to Lower Coniacian contain glauconitic hardground sequences, stratigraphic condensation, and hiatuses. In the western Barranca and in the eastern Vitoria Basin, the lowermost Coniacian is marked by intercalations of turbidites. In contrast, the Turonian of the Santander Basin is dominated by turbidites, whereas the Coniacian marl sequences are mostly autochthonous.

Stratigraphic work based mainly on foraminifera was carried out by RAMIREZ DEL POZO (1971) for the whole Cantabrian region. Most geological dating of the successions in our study area was previously based only on micropaleontological data (IGME, mapa geologico de España 1 : 50 000, hoja 34 Torrelavega, hoja 114 Alsasua, hoja 115 Gulina, hoja 140 Estella, and others). Specific research on selected sections and systematics of selected animal groups across the Turonian/Coniacian boundary include LAMOLDA & PROTO-DECIMA (1986), WIEDMANN (1977, 1979), and others. WIEDMANN & KAUFFMAN (1978) combined the ammonite and inoceramid stratigraphy. LAMOLDA (1982) summarized all available stratigraphic data on ammonites, bivalves, echinoids, foraminifers, and ostracods of the Turonian and its boundaries. We base our stratigraphic evidence on a systematic, detailed bed by bed collection.

2. Turonian/Coniacian boundary sequence in individual sections

Only two sections, the one from Liencres (Santander Basin) and the combined Izurdiaga-Arardi section (Barranca), encompass the stage boundary as used in this paper. The section from Ollogoyen (Estella Basin) covers up to the uppermost Turonian while the sections Zuazu-Ecay and Villanueva de Araquil (Barranca) cover the Lower Coniacian.

2.1 **Ollogoyen/Estella Basin** (Text-Fig. 2)

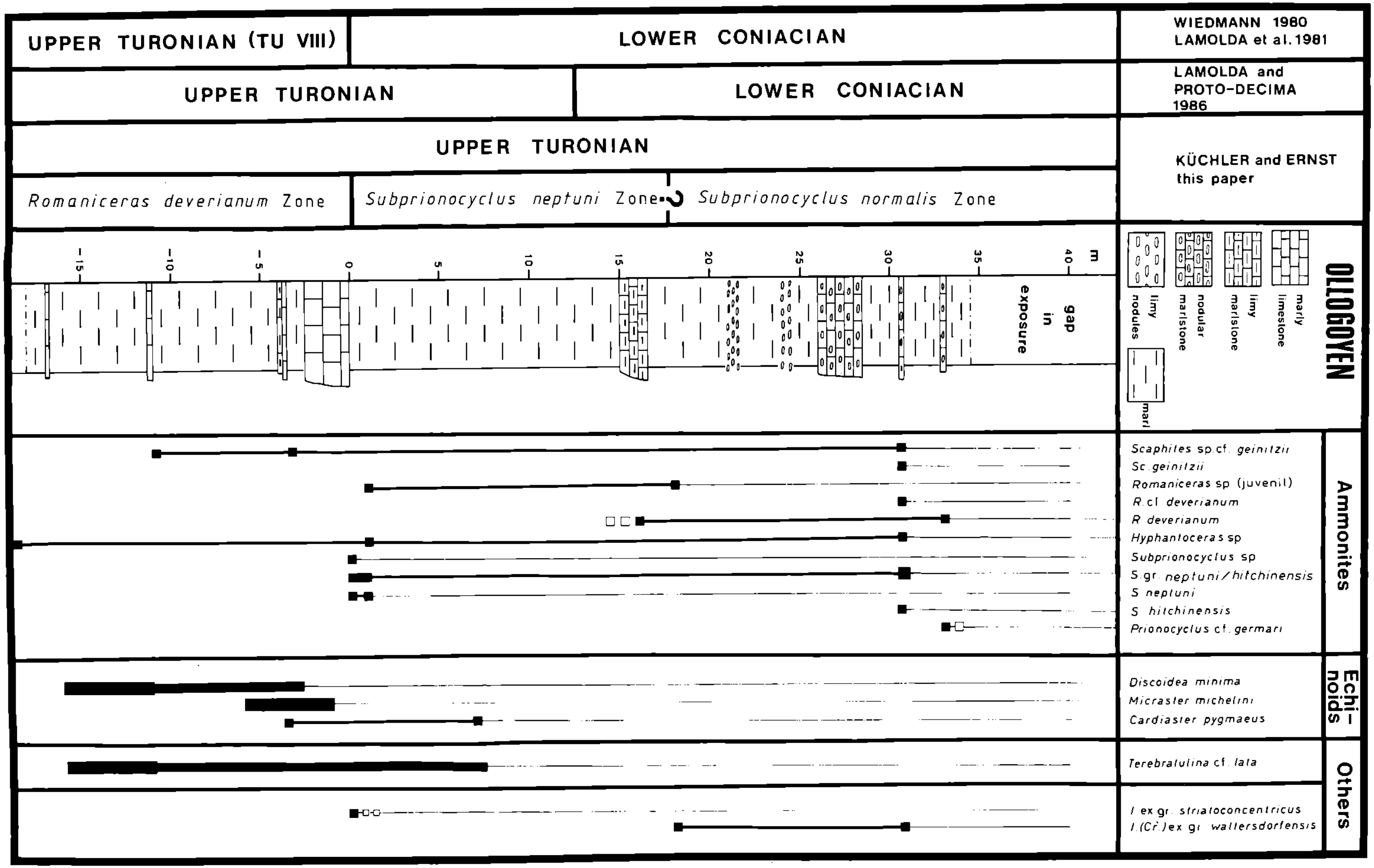
Location: The section is exposed 300 m west of the village of Ollogoyen (about 10 km west of Estella/Navarra).

The section contains the most complete sequence of the uppermost Turonian in northern Spain with a highly diverse fauna. The ammonites were studied by WIEDMANN (WIEDMANN & KAUFFMAN 1978, LAMOLDA, RODRIQUEZ-LAZARO & WIEDMANN 1981), and the micro- and nannofauna by LAMOLDA & PROTO-DECIMA (1986). Since then, this locality serves as a regional standard for the Turonian/Coniacian boundary in northern Spain.

LAMOLDA & PROTO-DECIMA already recognized that the boundary in foraminiferal terms lay at least 12 m above that postulated by WIEDMANN (1980). After our studies, Upper Turonian ammonite and inoceramid faunas still occur at least 33 m above WIEDMANN's boundary horizon (marked in Text-Fig. 2 by the 0 m level). Above the 33 m level until the beginning of massive limestones, the sequence formerly was completely covered by talus (compare Text-Fig. 2). In 1988 even we found *Romaniceras* sp. and *Subprionocyclus* sp. somewhat higher in newly exposed outcrops near to the 40 m level.

Unequivocal Coniacian index fossils are missing in our collection. It is still uncertain whether the quoted Coniacian ammonites of LAMOLDA & PROTO-DECIMA - *Tissotia (M.)* cf. *robini* THIOLLIERE and *Reesideoceras* cf. *camerounense* BASSE from the level of sample 40 - are actually from fallen blocks or slides.

Text-Fig. 2. Vertical ranges and abundance of faunal elements in the Upper Turonian part of the Ollogoyen section (Estella Basin). Instead of *Cardiaster pygmaeus* use the older synonym *C.* cf. *truncatus* (GOLDFUSS). Recent findings of ammonites and inoceramids are only considered in the text.


UPPER TURONIAN (TU VIII)
LOWER CONIACIAN
WIEDMANN 1980
LAMOLDA et al. 1981
UPPER TURONIAN
LOWER CONIACIAN
LAMOLDA and PROTO-DECIMA 1986
UPPER TURONIAN
KÜCHLER and ERNST this paper
Romaniceras deverianum Zone
Subprionocyclus neptuni Zone
?
Subprionocyclus normalis Zone
-15 -10 -5 0 5 10 15 20 25 30 35 40 m
gap in exposure
limy nodules
nodular marlstone
limy marlstone
marly limestone
marl
OLLOGOYEN
Scaphites sp. cf. geinitzii
Sc. geinitzii
Romaniceras sp (juvenil)
R. cf. deverianum
R. deverianum
Hyphantoceras sp
Subprionocyclus sp
S. gr. neptuni/hitchinensis
S. neptuni
S. hitchinensis
Prionocyclus cf. germari
Ammonites
Discoidea minima
Micraster michelini
Cardiaster pygmaeus
Echinoids
Terebratulina cf. lata
I. ex gr. striatoconcentricus
I. (Cr.) ex gr. waltersdorfensis
Others

During our investigation of the Turonian/Coniacian boundary, close attention was paid to the upper part of the section. The sequence is lithologically composed of marls intercalated with marly limestones. Characteristic is the concentration of macrofossils in a few dispersed layers. Their local vertical ranges are illustrated in Text-Fig. 2.

The highly diverse ammonite fauna consists of *Romaniceras, Subprionocyclus, Prionocyclus, Hyphantoceras*, and *Scaphites*. A few of the genera and species, according to our knowledge, are first records from northern Spain - for example *Hyphantoceras, Prionocyclus* cf. *germari* (REUSS) and *Scaphites geinitzii* D'ORBIGNY (see Pl. 1 and Pl. 2, Figs. 1-3). Of great stratigraphic interest is the existence of *Romaniceras deverianum* (D'ORBIGNY) associated with *Subprionocyclus* gr. *neptuni/hitchinensis* and *Prionocyclus* cf. *germari*.

In the upper part of the section, inoceramids are found in the same layers as the ammonites. Detailed studies of the inoceramids are not yet completed. Principally, the inoceramid fauna is poor and represented by the *striatoconcentricus* and *waltersdorfensis* groups, as well as the *lusatiae* and *glatziae* groups.

With the echinoids, the occurrence of *Micraster (Eomicraster) michelini* AGASSIZ and *Cardiaster* cf. *truncatus* (GOLDFUSS) is noteworthy. The former is characteristic of the uppermost part of the Turonian in NW Europe.

Furthermore for correlation purposes, the mass invasion of *Terebratulina* cf. *lata* is of interest, which corresponds to the similarly-aged occurrences in the Teutoburger Wald in Germany (compare chapter 4.4).

2.2 Barranca

2.2.1 Izurdiaga (Text-Fig. 3)

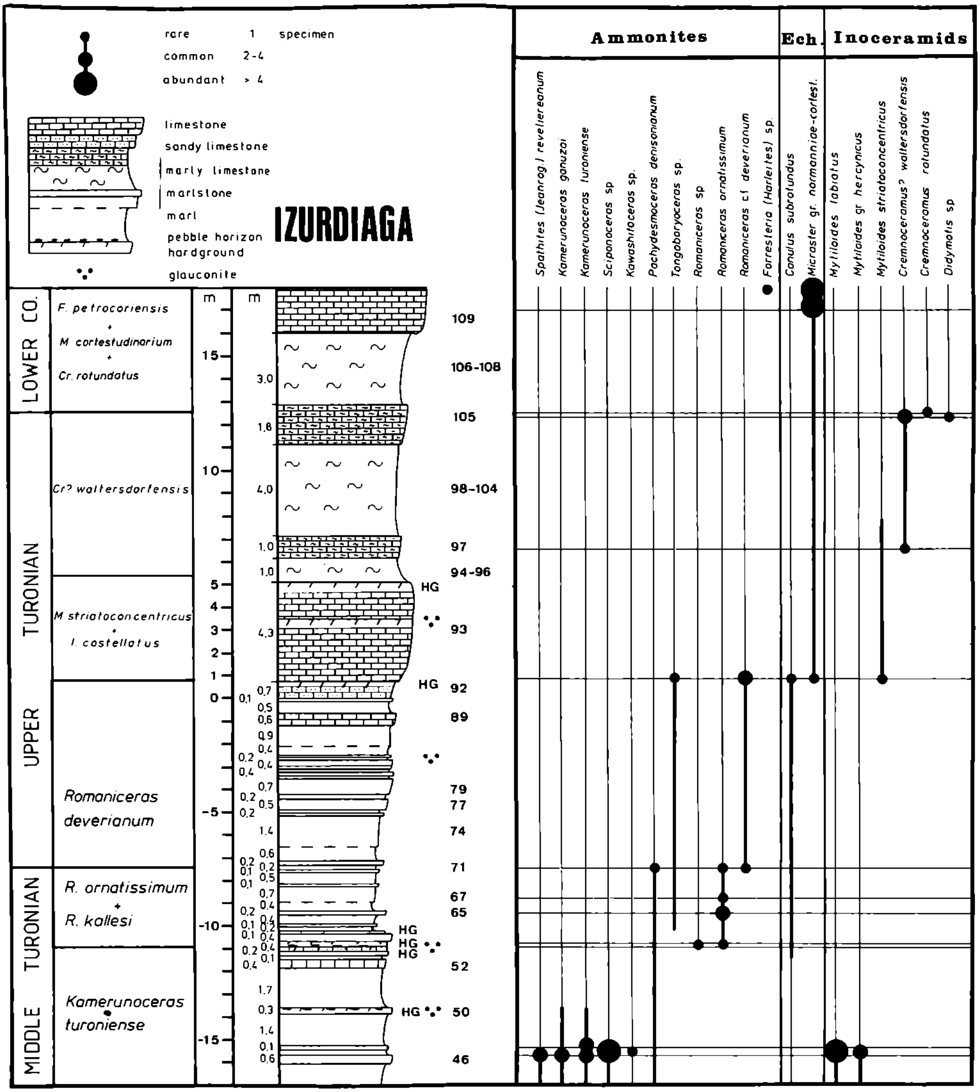
Location: This section lies about 1.5 km south of Irurzun, on the eastern edge of the village Izurdiaga.

The entire section which ranges in age from Turonian to Campanian was studied lithostratigraphically by DEGENHARDT (1983). The figured part of the section, comprising the Turonian and lowermost Coniacian, was investigated in detail by KÜCHLER (unpubl.).

Lithologically the section is composed of alternating layers of marl, marly limestone, and limestone. In the uppermost Middle Turonian and uppermost Upper Turonian, glauconitic hardground sequences are developed, representing condensation or stratigraphic gaps. With the uppermost hardground sequence the so-called Izurdiaga limestone begins, as incidentally mentioned by RADIG (1973).

Biostratigraphically important is the succession of *Romaniceras* species in the Izurdiaga section comparable with the succession recorded in France. *Romaniceras ornatissimum* (STOLICZKA) is replaced in the Upper Turonian by *Romaniceras* cf. *deverianum* (D'ORBIGNY).

The section is significant for the Turonian/Coniacian sequence because of the occurrence of characteristic index ammonites, inoceramids and echinoids, as well as the index bivalve *Didymotis*. The exact boundary of the Turonian/Coniacian, according to the authors, is characterized by the joint occurrence of *Inoceramus (? Cr.) waltersdorfensis* ANDERT, *I. (Cr.) rotundatus* FIEGE and *Didymotis* sp. (single specimen). A few meters above this boundary



Text-Fig. 3. Izurdiaga. Standard section of the Turonian/Coniacian boundary succession of the eastern Barranca with vertical ranges and abundance of macrofaunal elements.

layer, *Forresteria* sp. appears. This indicates the Coniacian according to ammonite stratigraphers. This entry coincides with the first acme of *Micraster cortestudinarium* (GOLDFUSS). Below this level, only the forerunners of this *Micraster* lineage - *M.* gr. *normanniae* - occur sporadically.

A comparable section lies about 2 km southeast of Izurdiaga on the northeast slope of Monte Arardi. By pooling the two sections, the vertical ranges of the macrofaunal elements are much easier to recognize.

2.2.2 Zuazu (Text-Fig. 4)

Location: This section is in a southern lateral valley (Irregueta) of the eastern Barranca, about 650 m south of the village of Zuazu.

The Turonian/Coniacian boundary sequence (sensu auctt.) is not exposed. Text-Fig. 4 only represents the lower part of the entire section which comprises a sequence from the middle part of the Lower Coniacian up to the Lower Santonian.

This locality is of great interest for Coniacian stratigraphy because it is possible to build a nearly complete standard succession for this area (KÜCHLER, in prep.). The macrofauna is characterized by a combination of inoceramids, ammonites, and abundant echinoids.

The evolutionary change within the genera *Peroniceras* allows a further division of the Lower Coniacian. *Peroniceras* cf. *bajuvaricum* (REDTENBACHER), which appears approximately 12 m under the base of the figured section, is replaced by *Peroniceras* cf. *subtricarinatum* (D'ORBIGNY) (see Pl. 3, Fig. 1) in the uppermost part.

Ecostratigraphically noteworthy is the peak occurrence of *Anagaudryceras* sp. in combination with inoceramids, which corresponds to the nearby section Villanueva de Araquil and other sections of the Barranca. The inoceramid fauna is generally typical for the Lower Coniacian with an association of *I. (? Cremnoceramus) waltersdorfensis* and *I. (Cr.) deformis* MEEK. The latter form constitutes ecological events within the Zuazu section (e. g. layer 146).

Most of the echinoids of the genera *Micraster, Echinocorys,* and *Cardiotaxis* are composed of species with a wide vertical range, so that the local stratigraphy is based on an increase in abundance in certain layers. Especially noteworthy is the occurrence of *Sternotaxis* aff. *placenta* (AGASSIZ), an index fossil of the uppermost Turonian and Lower Coniacian in NW Europe.

2.2.3 Ecay

Location: The Ecay section lies about 1 km east of the Zuazu section in another small lateral valley of the eastern Barranca.

Here the two prominent lower and upper members of the Izurdiaga limestone formation are truncated by a fault. The well-exposed part of the section contains the Subtricarinatum Zone of the Lower Coniacian. This part is ecostratigraphically, but not lithostratigraphically, comparable with the Zuazu section in many details.

The biostratigraphic results of both sections are assembled and illustrated in Text-Fig. 7. *Tissotioides haplophyllus* (REDTENBACHER) first appears in the uppermost part of the Lower Coniacian. A single specimen has been found in the Ecay section in the Subtricarinatum Zone. This is noteworthy

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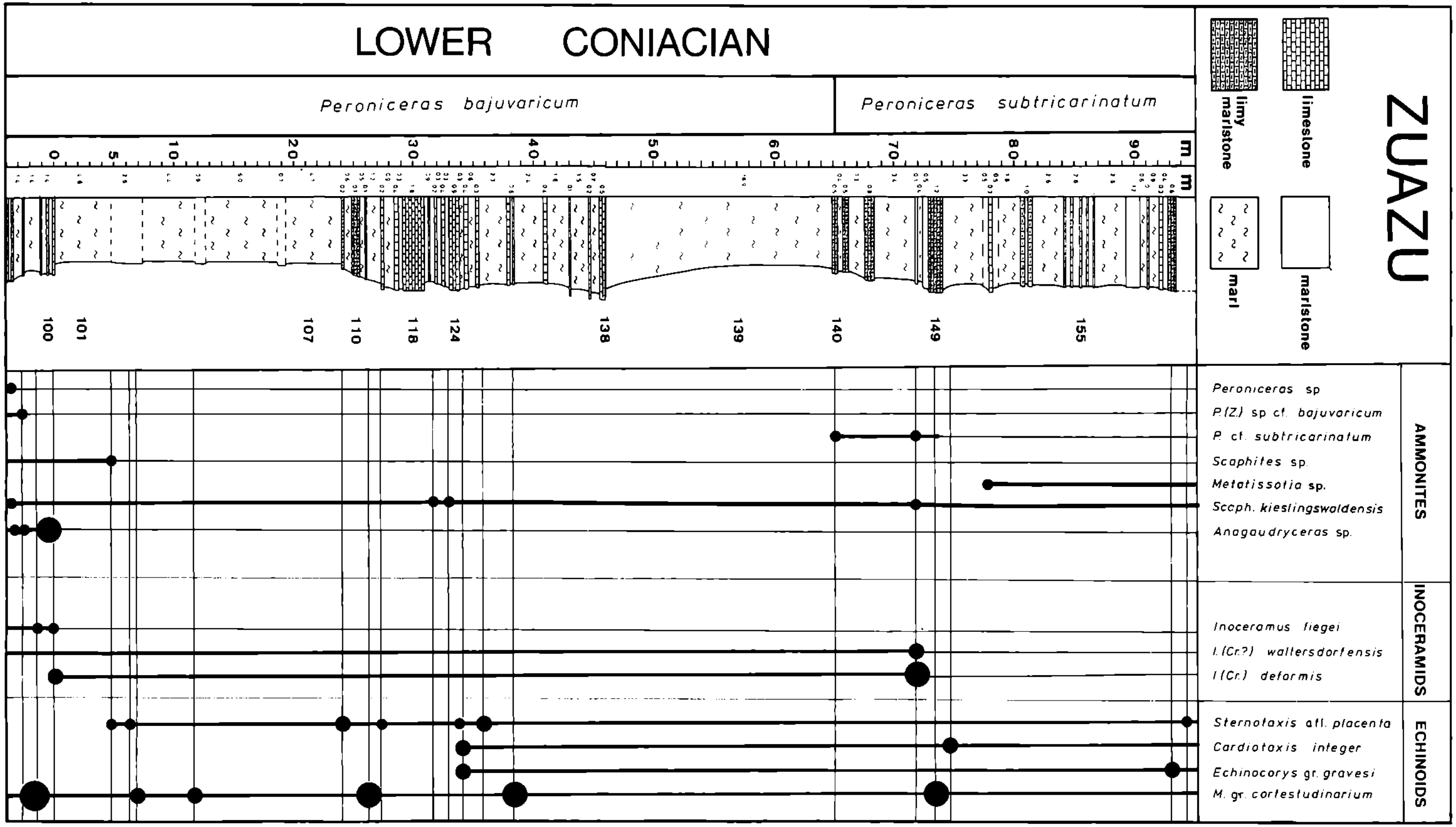
Text-Fig. 4. Middle part of the Lower Coniacian as exposed at Zuazu (Barranca), showing faunal distribution.


LOWER CONIACIAN
Peroniceras bajuvaricum
Peroniceras subtricarinatum
ZUAZU
limy marlstone
limestone
marl
marlstone
m
0 5 10 20 30 40 50 60 70 80 90
100 101 107 110 118 124 138 139 140 149 155
AMMONITES
Peroniceras sp
P.(Z.) sp cf. bajuvaricum
P. cf. subtricarinatum
Scaphites sp.
Metatissotia sp.
Scaph. kieslingswaldensis
Anagaudryceras sp.
INOCERAMIDS
Inoceramus fiegei
I. (Cr.?) waltersdorfensis
I. (Cr.) deformis
ECHINOIDS
Sternotaxis aff. placenta
Cardiotaxis integer
Echinocorys gr. gravesi
M. gr. cortestudinarium

because this species was used by WIEDMANN & KAUFFMAN (1978) as an index fossil for the lowest Coniacian in northern Spain.

2.2.4 Villanueva de Araquil (Text-Fig. 5)

L o c a t i o n: This section is found as a small isolated outcrop within alluvium in the middle of the Araquil valley about 500 m NE of Villanueva de Araquil. It is accessible along a knoll on which an iron cross is erected.

The sequence is extremely condensed, characterized by a series of hardgrounds. However, the hardgrounds, like the overlying marls, are extremely rich in echinoids, ammonites, and inoceramids. Collections were chiefly made in the marls immediately above the hardgrounds HG3 and HG4 (Text-Fig. 5).

The ammonites are generally preserved only as fragments. Stratigraphically important are two single specimens of *Forresteria* sp. cf. *petrocoriensis* (COQUAND) and *Forresteria nicklesi* (DE GROSSOUVRE) found in the lower portion of the section (see Pl. 2, Fig. 4). By means of *Peroniceras* the succession can be placed within the zone of *P. bajuvaricum* (see Pl. 2, Fig. 5). The section Villanueva is in the middle part of the Lower Coniacian, not in the basal portion as conjectured by WOOD, ERNST & RASEMANN (1984). An accurate comparison with the nearby Zuazu and Ecay sections is possible due to the peak zone of *Anagaudryceras* sp.

For the inoceramids, the joint occurrence of *I. (? Cr.) waltersdorfensis hannovrensis* HEINZ, *I. (Cr.) deformis* and related forms are typical for an association in the middle part of the Lower Coniacian (see Pl. 4, Figs. 2-5).

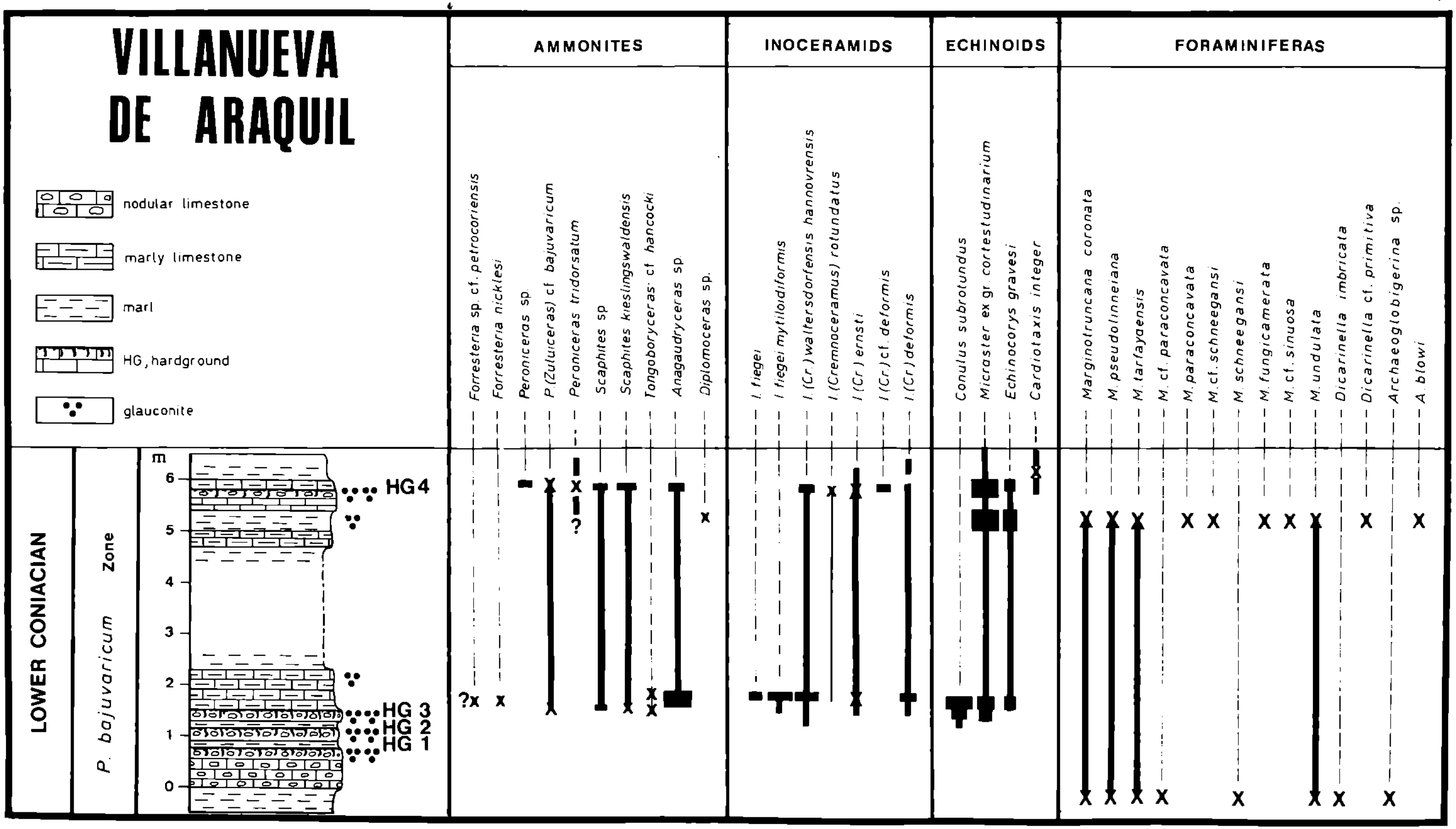
The hardground sequence which points to shallow marine conditions contains a very rich echinoid fauna. In the hardground layer HG3 and its overlying marl *Conulus subrotundus* (MANTELL), characteristic for such condensed sequences, is common. Below HG4 *Echinocorys* ex gr. *gravesi* develops especially tall tests. Within rich populations, *Micraster* ex gr. *cortestudinarium* likewise generates high gibbous forms.

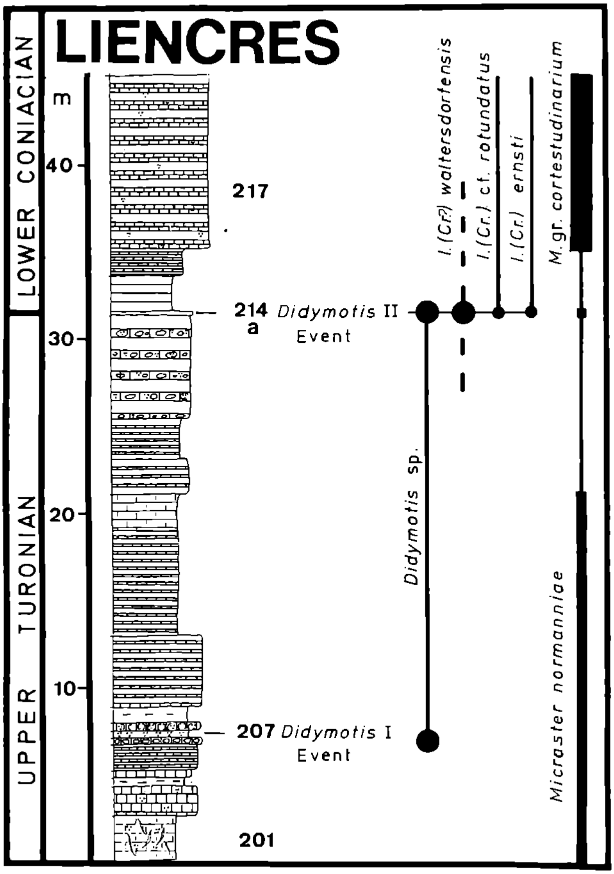
The foraminiferal fauna was kindly identified from two samples by M. LAMOLDA (Bilbao, correspondence from 13. 12. 83). The fauna is composed of a form-rich association with mostly long-range forms. The lower sample exhibits certain features which are characteristic for the Turonian/Coniacian boundary sequence (sensu LAMOLDA), while the upper sample is clearly from the Lower Coniacian (see Text-Fig. 5).

2.3 **Liencres/Santander Basin** (Text-Fig. 6)

L o c a t i o n: This section is exposed as a cliff along the Biscay coast about 10 km west of Santander and about 1 km NNW of the village of Liencres. It is exactly located on the ENE side of the Bay of Liencres

Text-Fig. 5. Condensed sequence of the middle part of the Lower Coniacian at the Villanueva de Araquil section. (Foraminifera data after LAMOLDA, pers. commun.).


VILLANUEVA DE ARAQUIL
nodular limestone
marly limestone
marl
HG, hardground
glauconite
LOWER CONIACIAN
P. bajuvaricum Zone
m
0
1
2
3
4
5
6
HG 1
HG 2
HG 3
HG 4
AMMONITES
Forresteria sp. cf. petrocoriensis
Forresteria nicklesi
Peroniceras sp.
P.(Zuluiceras) cf. bajuvaricum
Peroniceras tridorsatum
Scaphites sp.
Scaphites kieslingswaldensis
Tongoboryceras cf. hancocki
Anagaudryceras sp.
Diplomoceras sp.
INOCERAMIDS
I. fiegei
I. fiegei mytiloidiformis
I.(Cr.) waltersdorfensis hannovrensis
I.(Cremnoceramus) rotundatus
I.(Cr.) ernsti
I.(Cr.) cf. deformis
I.(Cr.) deformis
ECHINOIDS
Conulus subrotundus
Micraster ex gr. cortestudinarium
Echinocorys gravesi
Cardiotaxis integer
FORAMINIFERAS
Marginotruncana coronata
M. pseudolinneiana
M. tarfayaensis
M. cf. paraconcavata
M. paraconcavata
M. cf. schneegansi
M. schneegansi
M. fungicamerata
M. cf. sinuosa
M. undulata
Dicarinella imbricata
Dicarinella cf. primitiva
Archaeoglobigerina sp.
A. blowi



Text-Fig. 6. Turonian/Coniacian boundary sequence at the Liencres section (Santander Basin) with the two *Didymotis* events. (Highly modified after THEUERKAUFF 1987).

(Playas de Liencres). The section was measured and stratigraphically as well as sedimentologically interpreted by THEUERKAUFF (1987).

It ranges from the Middle Cenomanian - developed in hardground facies - to the uppermost Lower Coniacian. Only the Turonian/Coniacian boundary sequence will be discussed here. The section is especially important for correlation purposes because in addition to the abundant *Micraster*, inoceramids and *Didymotis* are present. Ammonites are extremely rare.

The northern seaward side of the section is lithologically composed of allochthonous sediments composed of turbidites and small slumps which are Lower and lower Middle Turonian in age. In the direction of the bay on the other side of the access road, after a gap in exposure, the section illustrated in Text-Fig. 6 is exposed. It comprises a more or less autochthonous series of strongly cemented marls and marly limestones. The sequence containing the two *Didymotis* events is developed in a nodular limestone facies containing some glauconite. Higher, near the 25 m level of the section, THEUERKAUFF (1987) found large *Thalassinoides*. These burrows, as well as the nodular limestone, could represent a stratigraphic condensation.

Didymotis was found only in two layers separated by 24 m. The specimens are rather small and badly preserved and resemble the species *D. uermoesensis* (SIMIONESCU) (Pl. 3, Figs. 2, 3, 5). This form is especially interesting for international correlation (chapter 4.3). It co-exists with *I. (? Cr.) waltersdorfensis* and *I. (Cr.) rotundatus* in the *Didymotis* II event and all these forms are characteristic for the stage boundary (sensu auctt.).

In the boundary sequence, *Micraster* forms are represented by late forms of the conservative *M. leskei* (DESMOULINS) and more common evolution members of the *precursor-normanniae-cortestudinarium* line. As in the eastern Barranca, distinct rich populations of *M. cortestudinarium* were found for the first time a few meters above the boundary layer (layer 230-234, from an unpublished report of P. WOLZ). From the restricted boundary area representatives of the *Gibbaster/Isomicraster* type also appear. The systematics of the *Micraster* fauna was first studied by LAMBERT (1920, 1922) without stratigraphic consideration and urgently needs revision.

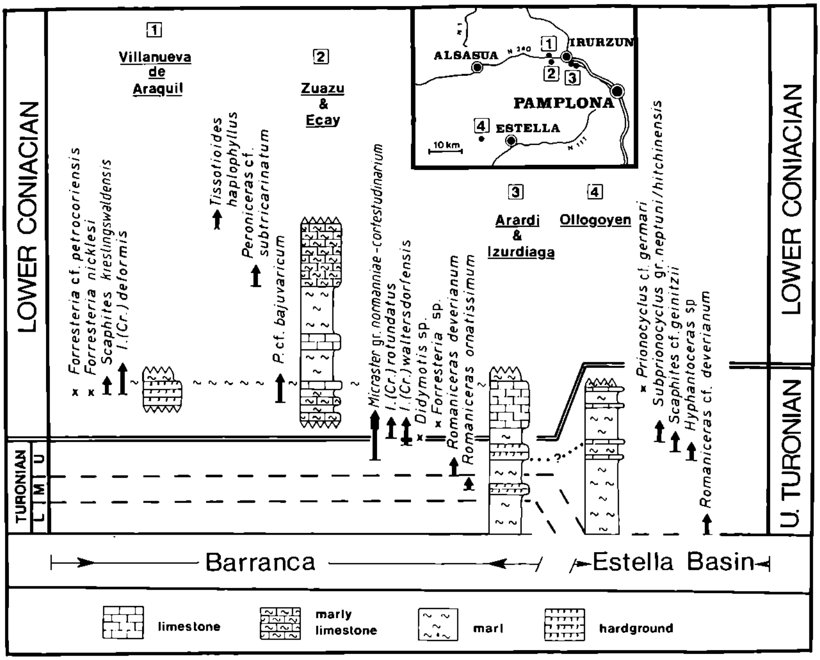
3. Correlation within Northern Spain (Text-Fig. 7)

In order to compare the Turonian-Coniacian boundary interval within northern Spain, we mainly used the section of Ollogoyen in the Estella Basin, Izurdiaga in the eastern Barranca, and Liencres in the Santander Basin. In Ollogoyen, a thick, almost complete Upper Turonian sequence is present with detailed stratigraphic data from ammonites and foraminifera. The Izurdiaga section comprises the Middle and Upper Turonian, which is condensed through two hardground sequences, and the non-condensed transition into the Coniacian, which is well-documented through inoceramids and *Didymotis*. The Liencres section complements the Izurdiaga section in respect to the restricted Turonian-Coniacian boundary sequence, with evidence from *Didymotis*, inoceramids, and echinoids.

In order to subdivide and refine the Lower Coniacian, the fossil-rich sections at Zuazu-Ecay and Villanueva de Araquil, from the eastern Barranca were also used. A few other sections in the western Barranca and in the Vitoria Basin, which were studied by others within our Cretaceous Working Group (e. g. WOLZ 1985, ZANDER 1988), contributed additional information in respect to the Upper Turonian and Lower Coniacian, but no information on the restricted boundary sequence. Text-Fig. 7 illustrates the most important sections from the eastern Barranca correlated with that from Ollogoyen.

3.1 Upper Turonian

The Upper Turonian can be divided through the ammonites *Romaniceras deverianum* and *Subprionocyclus neptuni* - similar to the Upper Turonian in France (AMÉDRO et al. 1982). In addition the upper part of the Neptuni Zone can be differentiated with a special fauna characteristic for the Subprionocyclus normalis Zone in NW Germany. In Izurdiaga, as well as in Ollogoyen, the vertical range of *R. deverianum* overlaps with its predecessor



Text-Fig. 7. Correlation of Upper Turonian and Lower Coniacian Barranca sections with the Ollogoyen section. The arrows mark the entry data of selected index fossils.

R. ornatissimum in the basal portion of the Deverianum Zone. Inoceramids are extremely rare in the two Romaniceras zones.

Clear evidence of a S. neptuni Zone can be recognized for the first time in northern Spain from rather rich populations of *Subprionocyclus* in several different horizons. Most forms exhibit characteristics from both *neptuni* and *hitchinensis* (especially the juvenile stages so that the two "types" can be interpreted to be closely related, or conspecific). It is stratigraphically significant that within the total range zone of *S. neptuni-hitchinensis, R. deverianum* rarely occurs.

In the eastern Barranca, the Neptuni Zone is strongly condensed. The zone is indicated by the presence of an inoceramid association composed of *I. costellatus* and *I. (M.) striatoconcentricus,* occurring together with *R. deverianum, Baculites undulatus* D'ORBIGNY, *Tongoboryoceras* sp., and *Kawoshitoceras* sp. cf. *multinodosum* (SCHLÜTER). Elements of this association are found at exactly the same level in Izurdiaga and in the parallel section at Arardi. In Ollogoyen the entry of *I. striatoconcentricus* coincides with the first appearance of *S. neptuni.*

Subprionocyclus normalis (ANDERSON), which is used, e. g. in Japan, the Teutoburger Wald (NW Germany) (KAPLAN 1986, 1988) in order to define the uppermost Upper Turonian, has not yet been demonstrated to be present in northern Spain. But the Normalis Zone is indicated by the single find of *Prionocyclus germari* in the Ollogoyen section. It was found together with late forms of *R. deverianum*.

Similarly-aged strata have also been tentatively determined for the section at Arardi. Here, the index inoceramids *I. (? Cr.) waltersdorfensis* and *I. websteri* (sensu WOODS non MANTELL) occur contiguously. The latter is characteristic of the faunal zone of *I.* aff. *frechi* and *I.* cf. *websteri* in NW Germany, which corresponds to the Normalis Zone (compare KAPLAN & BEST 1984).

The Normalis Zone is also observed in the Liencres section through the presence of the index bivalve *Didymotis* in the *Didymotis* I event. In fact, *Didymotis* I has been identified about 24 m under the postulated Turonian/Coniacian boundary recognized by THEUERKAUFF (1987) and ourselves (Text-Fig. 6). A few meters under the *Didymotis* I event, the evolution members *precursor-normanniae* of the main *Micraster* phylum appear. These forms are also characteristic of similar stratigraphic levels within the sections in the eastern Barranca.

3.2 Turonian/Coniacian boundary

WOOD, ERNST & RASEMANN (1984) proposed to place the boundary between the Turonian and Coniacian at the appearance of *I. (? Cr.) waltersdorfensis hannovrensis* in the *Didymotis* II event. In NW Germany, a first flood occurrence of *I. (Cr.) rotundatus* is found immediately above or in the same level. This characteristic combination of *Didymotis* with *I. (? Cr.) waltersdorfensis* and *I. (Cr.) rotundatus* was also found in northern Spain within the sections at Liencres and Izurdiaga, which comprise the Turonian/Coniacian transition. This widespread, well-documented manifestation strongly suggests that this position of the boundary is acceptable.

Forresteria petrocoriensis, proposed as a marker for the boundary by ammonite stratigraphers (BIRKELUND et al. 1984, KENNEDY 1984a) is rare in northern Spain as well as in England and NW Germany. A specimen of *Forresteria (Harleites)* sp. was found about 8 m above our stage boundary at the Izurdiaga section. Two other specimens of *Forresteria* were collected at Villanueva in the middle part of the Lower Coniacian.

French stratigraphers (POMEROL 1983, 1985) proposed *Micraster decipiens* (BAYLE) as an index fossil for the Turonian-Coniacian boundary. In our opinion, this echinoid is conspecific with *M. cortestudinarium* (GOLDFUSS), so that the latter name has priority (WOOD et al. 1984). This particular form develops a well-defined eco-maximum a few meters above our chosen boundary level in all measured sections (compare Text-Figs. 3, 6, 7).

According to foraminifera stratigraphy, the boundary is defined by the appearance of *Marginotruncana* from the *sinuosa* group, along with *Dicarinella* from the *primitiva-concavata* group (MARKS 1984). Both of these forms are present, though very rarely, in the upper portions of the section at Ollogoyen (LAMOLDA & PROTO-DECIMA 1986), where a typical *neptuni* fauna is present.

Objections against the use of *Marthasterites furcatus* as a boundary indicator were already raised by BAILEY et al. (1984) and WOOD et al.

(1984). Its first occurrence within the lower Upper Turonian remains a short-term event, while its second appearance in England directly marks the stage boundary sequence (sensu auctt.) (see BAILEY et al. 1984). It is therefore possible that the report of *M. furcatus* from the section of Ollogoyen by LAMOLDA & PROTO-DECIMA (1986) indicates its second occurrence near the boundary (40 m level in Text-Fig. 2). Recent findings in newly exposed outcrops prove, that *Romaniceras* sp. and *Subprionocyclus* sp. occur even in this stratigraphical level.

Inoceramus (Cr.) deformis, which some stratigraphers recommend as a Coniacian boundary indicator, is first clearly recognized above our designated boundary in the Peroniceras bajuvaricum Zone. This is true for all the sections measured in the Barranca.

3.3 Lower Coniacian

The Lower Coniacian is provisionally divided into three ammonite units; the eponymous index fossils for these zones are *Forresteria petrocoriensis*, *Peroniceras bajuvaricum*, and *Peroniceras subtricarinatum* (Text-Fig. 8). This scheme is only proven for the sections in the eastern Barranca.

In the western Barranca, the eastern Vitoria Basin, and the Santander Basin, this zone division is difficult to apply due to the lack of ammonites. Some parallels can be applied in France, where KENNEDY (1984b) divides the unit into two parts based on *F. petrocoriensis* and *P. tridorsatum*. In this paper we subdivide the Tridorsatum Zone into two units.

The Forresteria petrocoriensis Zone in the eastern Barranca is presently still poorly documented. The index fossil is unusually rare or missing. In the section at Izurdiaga, only a single badly-preserved specimen, which could only be assigned to *Forresteria (Harleites)* sp., was found. As mentioned, this example occurs in the first maximum occurrence of *Micraster cortestudinarium*, which is easy to follow as a peak zone and can also be found in the Santander Basin.

The index species of the Peroniceras bajuvaricum Zone is, contrary to *P. tridorsatum* more common. *P. tridorsatum* is represented by only a single specimen in this zone, as well as in the following one. The co-existing ammonites in the Bajuvaricum Zone include *Scaphites kieslingswaldensis*, *Tongoboryoceras* cf. *hancocki* KENNEDY, *Forresteria* sp. cf. *petrocoriensis*, and *F. nicklesi*. From inoceramid stratigraphic aspects, it is striking that *I. (Cremnoceramus) deformis* appears suddenly and develops an eco-maximum with *Anagaudryceras* sp., *I. (? Cr.) waltersdorfensis*, and *I. (? Cr.) w. hannovrensis*. A few meters above this level, the first occurrence of the globular *Sternotaxis* aff. *placenta* is characteristic of the section at Zuazu. Also, in reference to echinoid stratigraphy, the first appearance of the wide-ranging *Cardiotaxis integer* (AGASSIZ) and especially tall variants of *Echinocorys gravesi* (DESOR) in this zone are noteworthy.

The index form of the Peroniceras subtricarinatum Zone can be found in all the sections in the Barranca. A *Metatissotia* fauna (*Metatissotia ewaldi* (v. BUCH), *Metatissotia* sp.) occurs in the uppermost part of this zone. Also *Tissotioides haplophyllus* first appears in the upper part of this zone in the Ecay section. According to WIEDMANN & KAUFFMAN (1978) and WIEDMANN (1979), this species defines the base of the Coniacian in northern Spain.

The same inoceramids found in the Bajuvaricum Zone are also present in the Subtricarinatum Zone. Locally, a few *Echinocorys gravesi* events are stratigraphically important.

4. Intra-European comparison (Text-Fig. 8)

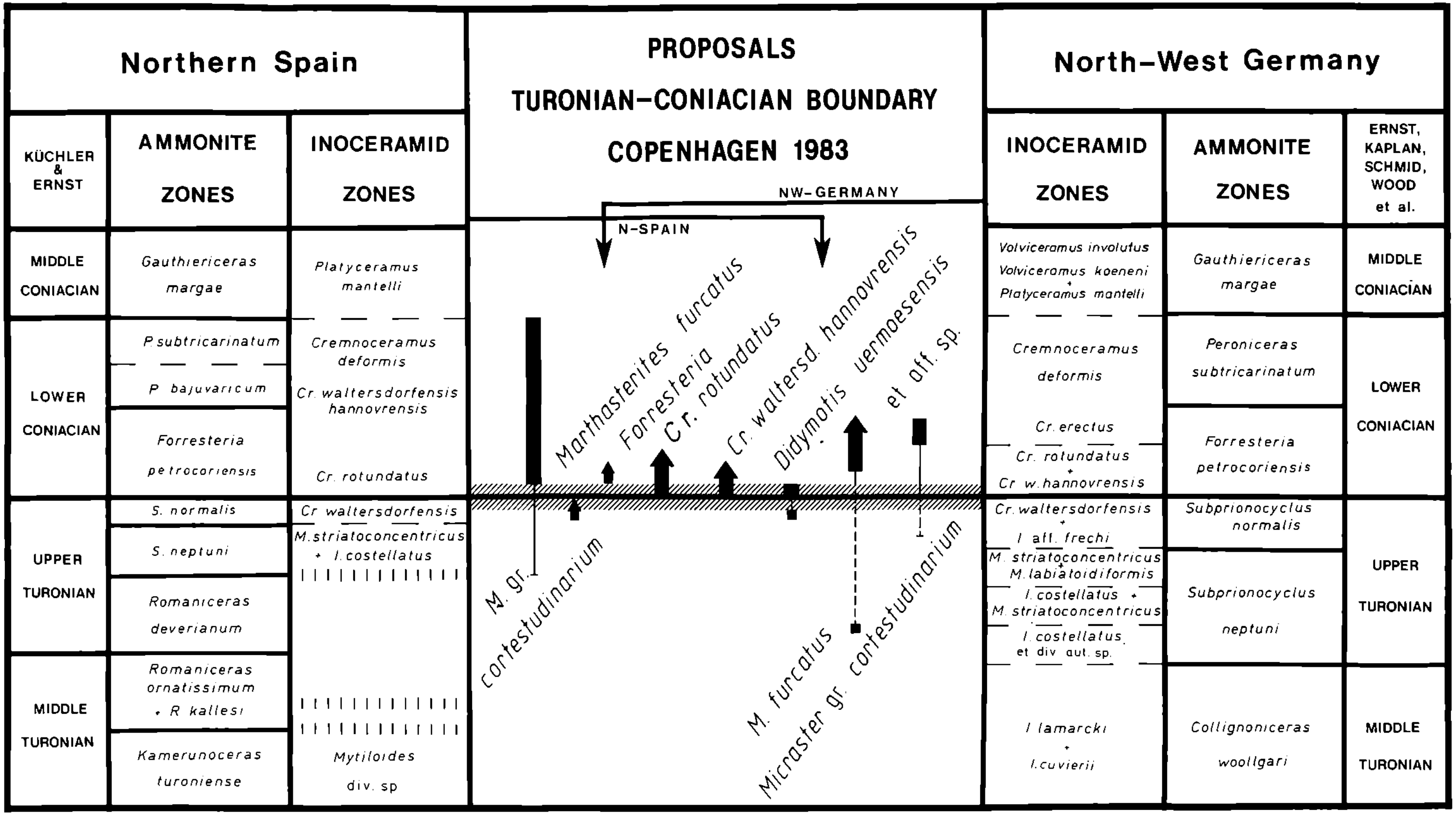
In this chapter the stratigraphically important assemblages of macrofossils between northern Spain and NW Germany are compared since these two regions are best known to the authors. Data from adjacent areas - England, France, Czechoslovakia and other countries - were used only incidentally.

Text-Fig. 8 displays the relevant results of the ammonite and inoceramid zonation developed within the Navarro-Cantabrian Basin and NW Germany. The most important difference is that the divisions in northern Spain are mainly based on ammonites while those in NW Germany on inoceramids. This basic distribution is due to the dominance of these fossil groups in their respective areas. A satisfactory correlation was first made possible with the successful completion of the inoceramid stratigraphy in Spain and ammonite stratigraphy in NW Germany. The latter stratigraphic scheme could only be developed with the consequent application of the event stratigraphy by KAPLAN (KAPLAN 1986, KAPLAN et al. 1987).

Accordingly, the arrangement of zones and boundaries in both regional schematics show deviations. For northern Spain we propose a three-fold division based on ammonites for the Upper Turonian as well as for the Lower Coniacian. A two-fold division based on ammonites has been used until now in Germany. The inoceramid division proposed by ERNST et al. (1983) and WOOD et al. (1984) for NW Germany - four assemblage zones in the Upper Turonian and three in the Lower Coniacian - cannot be recognized in such detail in northern Spain.

The substage boundary of **Middle/Upper Turonian** shown in Text-Fig. 8 does not correspond. According to AMÉDRO et al. (1982) (compare also WIEDMANN 1979) we place the substage boundary in northern Spain at the entry of *R. deverianum*. The Middle/Upper Turonian boundary in NW Germany is drawn at the first appearance of *S. neptuni*, which is isochronous with the so-called *costellatus/planus* event.

The **Lower/Middle Coniacian** boundary shows a general agreement between the two regional schemes, with regard to the joint appearance of *Gauthiericeras margae* (SCHLÜTER) and *I. (Platyceramus)* gr. *mantelli* MERCEY (BARROIS). For example, the two forms co-exist in the non-figured upper part of the Zuazu section (KÜCHLER, in prep.). A probable, similarly-aged frequency maximum of *I. (Platyceramus) mantelli* (until now without *G. margae*) is described by ZANDER (1988) from the western part of the Barranca near Alsasua in the Iturmendi I section. In Germany, *I. (Platyceramus)* gr. *mantelli* appears at about the same level as *I. (Volviceramus) koeneni* G. MÜLLER (TRÖGER 1981). In order to balance the division of the Coniacian, we believe that our sufficiently marked substage boundary is certainly more reasonable than the Lower/Middle Coniacian boundary proposed by KENNEDY (1984b).


Northern Spain
PROPOSALS
TURONIAN–CONIACIAN BOUNDARY
COPENHAGEN 1983
North–West Germany
KÜCHLER & ERNST
AMMONITE ZONES
INOCERAMID ZONES
NW–GERMANY
N–SPAIN
INOCERAMID ZONES
AMMONITE ZONES
ERNST, KAPLAN, SCHMID, WOOD et al.
MIDDLE CONIACIAN
Gauthiericeras margae
Platyceramus mantelli
LOWER CONIACIAN
P. subtricarinatum
P. bajuvaricum
Forresteria petrocoriensis
Cremnoceramus deformis
Cr. waltersdorfensis hannovrensis
Cr. rotundatus
UPPER TURONIAN
S. normalis
S. neptuni
Romaniceras deverianum
Cr. waltersdorfensis
M. striatoconcentricus + I. costellatus
MIDDLE TURONIAN
Romaniceras ornatissimum + R. kallesi
Kamerunoceras turoniense
Mytiloides div. sp.
Marthasterites furcatus
Forresteria
Cr. rotundatus
Cr. waltersd. hannovrensis
Didymotis vermoesensis et aff. sp.
M. gr. cortestudinarium
M. furcatus
Micraster gr. cortestudinarium
Volviceramus involutus
Volviceramus koeneni
+ Platyceramus mantelli
Cremnoceramus deformis
Cr. erectus
Cr. rotundatus
+ Cr. w. hannovrensis
Cr. waltersdorfensis
+ I. aff. frechi
M. striatoconcentricus
+ M. labiatoidiformis
I. costellatus +
M. striatoconcentricus
I. costellatus et div. aut. sp.
I. lamarcki
+ I. cuvierii
Gauthiericeras margae
Peroniceras subtricarinatum
Forresteria petrocoriensis
Subprionocyclus normalis
Subprionocyclus neptuni
Collignoniceras woollgari
MIDDLE CONIACIAN
LOWER CONIACIAN
UPPER TURONIAN
MIDDLE TURONIAN

4.1 Ammonites

The ammonite fauna of the studied section - situated within the Navarro-Cantabrian Basin corresponds well with those of more northern areas (France, NW Germany, and England).

Upper Turonian faunal comparison: The spectrum of genera in the Navarro-Cantabrian Basin and the NW European realm varies little. In regard to the abundance, *Romaniceras* dominates the Navarro-Cantabrian Basin while *Scaphites* is the dominant form in NW Germany.

Beyond that, the respective composition of the faunas was dependent on shelf position (near-shore or off-shore), water depth, and other ecological factors (compare WIEDMANN & KAUFFMAN 1978, DAHMER & ERNST 1986, KAPLAN 1988). So, *Collignoniceras* and *Subprionocyclus* represent a typical faunal composition in the Estella Basin and the Westphalian area. Both of these forms have not yet been found in the Barranca, where a community of *Romaniceras* with *Tongoboryoceras* is characteristic. There are affinities with the lowermost Upper Turonian in France (AMÉDRO et al. 1982, DEVALQUE et al. 1982).

The condensed Neptuni Zone in the Barranca contains elements (*R. deverianum, Tongoboryoceras* sp., *Baculites undulatus, Kawashitoceras* sp. cf. *multinodosum*) which are also described from the French Uchaux fauna (DEVALQUE et al. 1982), the English Chalk Rock fauna (WRIGHT 1979), and the NW German "Scaphiten-Schichten" fauna (KELLER 1982, KAPLAN 1986, 1988). Faunal elements missing from these communities in the Barranca can be found in the non-condensed Ollogoyen section, such as *Hyphantoceras* sp., *Eubostrychoceras* sp., *Subprionocyclus neptuni* (GEINITZ), *S.* gr. *neptuni/hitchinensis, Scaphites geinitzii* (D'ORBIGNY) (see Pl. 2, Figs. 1-2).

Lower Coniacian faunal comparison: Ammonites rarely occur in the Lower Coniacian of the studied sections. They are also rare in NW Germany and England. *Peroniceras* and *Scaphites* dominate the Spanish sections; the latter form is more common, just as in NW Germany. With the exception of France (KENNEDY 1984b) *Forresteria* is extremely rare in all the mentioned areas. Also, *Metatissotia* which is found in the uppermost Lower Coniacian of the Barranca, is rare in comparison with the rich *Metatissotia* fauna of the Burgos area (northern Spain).

4.2 Inoceramids

General differences between NW Germany and the Navarro-Cantabrian Basin concern the enormous abundance and high diversity of populations in the

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Text-Fig. 8. Comparison of zonal schemes in northern Spain and NW Germany. The middle portion shows the entry data and abundance of the boundary index fossils proposed in Copenhagen 1983. "Advanced" *Micraster* of the *normanniae-cortestudinarium* lineage have already entered in the Upper Turonian. Distinct *cortestudinarium* populations first appear somewhat above the boundary.

northern province. As with echinoids, the similarities between the two realms are more distinct for the Lower Coniacian than the Upper Turonian.

Upper Turonian faunal comparisons: The lowermost Upper Turonian (approximately Neptuni Zone) in NW Germany is dominated by associations with large forms of the *lamarcki-cuvierii* lineage, as well as by forms of the *costellatus, inaequivalvis, striatoconcentricus,* and *fiegei* groups.

The uppermost Upper Turonian (approximately Normalis Zone) contains a species-poor assemblage comprising *I.* aff. *frechi, I. websteri* and rare *I. (?Cr.) waltersdorfensis* (ERNST et al. 1983, KAPLAN & BEST 1984). A few meters below the stage boundary, the abundance and diversity of inoceramids again increases (WOOD et al. 1984).

Inoceramids have not yet been found in the Deverianum Zone in the studied sections in the Navarro-Cantabrian Basin. [Since completion of this typescript two not yet determined inoceramid specimens have been discovered in layers 73 and 74 from the Deverianum Zone of the Izurdiaga section (compare Text-Fig. 3).] On the whole, all representatives of the *lamarcki-cuvierii* group seem to be completely absent in this zone. Not until the Neptuni Zone appear rare specimens of the *costellatus* and *striatoconcentricus* group. Representatives of the *websteri* and *waltersdorfensis* groups enter still higher in the uppermost Upper Turonian in the Ollogoyen and Arardi sections. These forms indicate the stratigraphical level of the Normalis Zone.

Lower Coniacian faunal comparison: Besides the cremnoceramids of the *rotundatus-erectus-deformis* lineage, mass occurrences of *I. (?Cr.) waltersdorfensis, I. (?Cr.) w. hannovrensis,* and *I. (Cr.) inconstans* are characteristic for NW Germany.

All mentioned cremnoceramids are also identified from the Navarro-Cantabrian Basin. Their concentration within eco-events in the Lower Coniacian is noteworthy. Furthermore the associations yield representatives of the *fiegei-mytiloidiformis* lineage. The *I. (Pl.) mantelli* group is important for the definition of the Lower/Middle Coniacian boundary (as discussed above).

4.3 Other bivalves: *Didymotis* (Pl. 3, Figs. 3-6)

As already mentioned, the confidence sequence of the boundary between the Upper Turonian and Lower Coniacian is characterized by the short immigration of the bivalve *Didymotis*. Two well-defined flood occurrences of *Didymotis* separated by a unit nearly completely lacking this fossil are found in several localities in eastern Lower Saxony (the so-called *Didymotis* events I and II of WOOD et al. 1984). The lower event (D I) lies in the uppermost Upper Turonian, according to the associated inoceramids (and ammonites). The upper event (D II) marks the stage boundary (as defined in this paper) and is characterized by the sudden entrance of abundant *I. (?Cr.) waltersdorfensis hannovrensis*. Furthermore, in the same horizon or immediately above, the entry of *I. (Cr.) rotundatus* is noteworthy.

Nearly identical conditions were observed by ČECH (this vol.) in similar sequences in Czechoslovakia. The same is the case in the section of Liencres of the Santander Basin (Text-Fig. 6), although the inoceramids are less common in this region. Even the distance between the two *Didymotis* events in Salzgitter-Salder (Lower Saxony) and Bohemia is nearly equal (about 6 to 7 m), while it is about 24 m in Liencres.

Findings of *Didymotis*, mostly of the upper event, are also known from other European countries, e. g. Romania (SIMIONESCU 1899 and others), Poland (I. WALASZCZYK, pers. commun.), and NE England (WOOD et al.

1984). These ubiquitous occurrences make this form increasingly suitable as an ideal stage indicator, even without closer taxonomic revision of the different species determinations. Likewise the single specimen of *Didymotis* in the Izurdiaga section of the Barranca belongs to the same stratigraphic level. It remains to be seen whether the *Didymotis* findings in the Caribbean (KAUFFMAN 1978) correspond to an equivalent horizon or belong instead in the lowermost Lower Coniacian.

4.4 Brachiopods: *Terebratulina*

The brachiopod genus *Terebratulina* plays an important role in the subdivisions of the Turonian in England and France. A Terebratulina lata Zone is traditionally segregated in England. The boundaries of that zone are not clearly defined. The maximum occurrence lies in northern England according to C. J. WOOD, between the marker horizons (tephroevents) of the Deepdale Marls and North Ormsby Marl in the lower part of the Upper Turonian (WOOD & SMITH 1978 and pers. commun.). Likewise, *Terebratulina gracilis* D'ORBIGNY is used in France to separate a zone which extends from the base of the Middle Turonian to the Neptuni Zone (compare GASPARD 1982).

In reference to the widespread distribution of *Terebratulina lata* which perhaps is conspecific with the French *T. gracilis* it is of interest that this form occurs in abundance in the uppermost part of the Deverianum Zone and in the Neptuni Zone of the Ollogoyen section. *Terebratulina* obviously is absent in the uppermost Turonian (Normalis Zone) in the Spanish sections.

Terebratulina likewise was discovered in mass occurrences in the Hankenberge section in the Teutoburger Wald (NW Germany, unpubl. data). In this locality the peak zone is a few meters below the tephroevents F and G, that means in the uppermost Neptuni Zone.

The taxonomy of *Terebratulina* species urgently needs revision (GASPARD 1982). It is also unclear to what extend the morphology of this brachiopod was influenced by ecological factors.

4.5 Echinoids

Faunal province differences are generally rather clear in the Upper Turonian but slight in the Lower Coniacian. However, the assemblages in both provinces and substages are dominated by the irregular echinoid orders Holasteroida, Spatangoida, and Holectypoida.

Upper Turonian faunal comparison: The dominant forms in the Navarro-Cantabrian Basin include the taxa *Hemiaster, Micraster (Eomicraster),* and the small *Discoidea minima* (AGASSIZ). *Micraster* of the main lineage appears first in the latest Upper Turonian. *Cardiaster* cf. *truncatus* (GOLDFUSS), *Conulus subrotundus,* and others were only found locally. *Echinocorys* and *Sternotaxis* are missing in our sections. In contrast, the north temperate or boreal faunal province of NW Europe is dominated by the genera *Sternotaxis, Echinocorys, Infulaster,* and *Micraster.*

Lower Coniacian faunal comparison: Affinities between both faunal provinces exist in regard to the dominance of the main *Micraster phylum, Echinocorys* gr. *gravesi, Sternotaxis* gr. *placenta,* and, under special paleoecological conditions, in regard to *Conulus subrotundus.* A few southern elements - like *Hemiaster* div. sp., *Cardiaster integer*, and *Discoidea minima* - are not present at this time in NW Germany.

Using echinoids the Lower Coniacian of both regions is easier to correlate than the Upper Turonian. *Micraster* is the most important linking genus. The group has a certain tendency to develop geographical and ecological subspecies, but the general phylogenetic pattern (e. g. the transformation of the plastron and the subpetaloid ambulacra) is similar and isochronous throughout its distribution area.

"Advanced" *Micraster* of the early *bucailli* group, which seems to be the geographic modification of the *precursor-normanniae* group, invaded lithotopes in Lower Saxony and Westphalia rather suddenly in the sequence associated with ash layers F and G (i. e. in the boundary series of the Neptuni to the Normalis Zone), the so-called *Micraster* event (ERNST et al. 1983, WOOD et al. 1984). Above this in the uppermost Upper Turonian a gradual transformation occurs into *cortestudinarium* (= *decipiens*) populations. In the lowermost Lower Coniacian distinct populations of *M. cortestudinarium* come in at the so-called *cortestudinarium* event (WOOD et al. 1984). The evolutionary history of *Micraster* runs in a similar way in the Spanish, French, and English subprovinces (compare DRUMMOND & FOURAY 1983).

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Explanation of plates

All figures natural size, except Fig. 7 of Pl. 2 and Fig. 5 of Pl. 4. The material is housed in the collections of Institut für Paläontologie der FU Berlin (IPFU).

Plate 1

Fig. 1a, b. *Romaniceras deverianum* (D'ORBIGNY), Upper Turonian, Neptuni Zone, Ollogoyen (Estella Basin), 16.5 m level. IPFU-GE 89/1.

Fig. 2. *Romaniceras deverianum* (D'ORBIGNY), Upper Turonian, Normalis Zone, Ollogoyen (Estella Basin), 31 m level. IPFU-GE 89/2.

Fig. 3a, b. *Hyphantoceras* cf. *reussianum* (D'ORBIGNY), Upper Turonian, Deverianum Zone, Ollogoyen (Estella Basin), -19 m level. IPFU-GE 89/3.

Plate 2

Fig. 1. *Subprionocyclus* gr. *neptuni/hitchinensis*, Upper Turonian, Normalis Zone, Ollogoyen (Estella Basin), 31 m level. IPFU-GE 89/4.

Fig. 2. *Subprionocyclus* gr. *neptuni/hitchinensis*, Upper Turonian, Neptuni Zone, Ollogoyen (Estella Basin), 1 m level. IPFU-GE 89/5.

Fig. 3a, b. *Prionocyclus* cf. *germari* (REUSS), Upper Turonian, Normalis Zone, Ollogoyen (Estella Basin), 34 m level. IPFU-GE 89/6.

Fig. 4a, b. *Forresteria nicklesi* (DE GROSSOUVRE), Lower Coniacian, P. bajuvaricum Zone, Villanueva de Araquil (Barranca), HG3. IPFU-GE 89/7.

Fig. 5a, b. *Peroniceras (Zuluiceras)* cf. *bajuvaricum* (REDTENBACHER), Lower Coniacian, P. bajuvaricum Zone, Villanueva de Araquil (Barranca), 30 cm marl bed above HG3. IPFU-GE 89/8.

Fig. 6a, b. *Peroniceras tridorsatum* (SCHLÜTER), Lower Coniacian, P. subtricarinatum Zone, Satrustegui (Barranca), bed A3. IPFU-GE 89/9.

Fig. 7. *Scaphites kieslingswaldensis* LANGENHAN & GRUNDEY (x 1.1), Lower Coniacian, P. subtricarinatum Zone, Ecay (Barranca), bed SL. IPFU-GE 89/10.

Plate 3

Fig. 1a, b. *Peroniceras subtricarinatum* (D'ORBIGNY), Lower Coniacian, P. subtricarinatum Zone, Ecay (Barranca), bed P. IPFU-GE 89/11.

Fig. 2, 3, 5. *Didymotis* sp., Turonian/Coniacian boundary (Santander Basin), bed 214a (upper *Didymotis* ecoevent). IPFU-GE 89/12.

Plate 3 (continued)

Fig. 4, 6. *Didymotis uermoesensis* (SIMIONESCU), Turonian/Coniacian boundary, *Didymotis*-ecoevent (D II), Salzgitter-Salder section (Lower Saxony, NW Germany), bed MK 45. IPFU-GE 89/13.

Plate 4

Fig. 1. *I. (Mytiloides) striatoconcentricus* (GÜMBEL), Upper Turonian, Zone of I. costellatus/M. striatoconcentricus, Izurdiaga (Barranca), bed 93. IPFU-GE 89/14.

Fig. 2, 3, 4. *Inoceramus (?Cr.) waltersdorfensis hannovrensis* HEINZ, Lower Coniacian, P. bajuvaricum Zone, Villanueva de Araquil (Barranca), 30 cm marl bed above HG3. IPFU-GE 89/15.

Fig. 5. *I. (Cremnoceramus) deformis* MEEK (x 0.73), Lower Coniacian, P. bajuvaricum Zone, Villanueva de Araquil (Barranca), 30 cm marl bed above HG3. IPFU-GE 89/16.

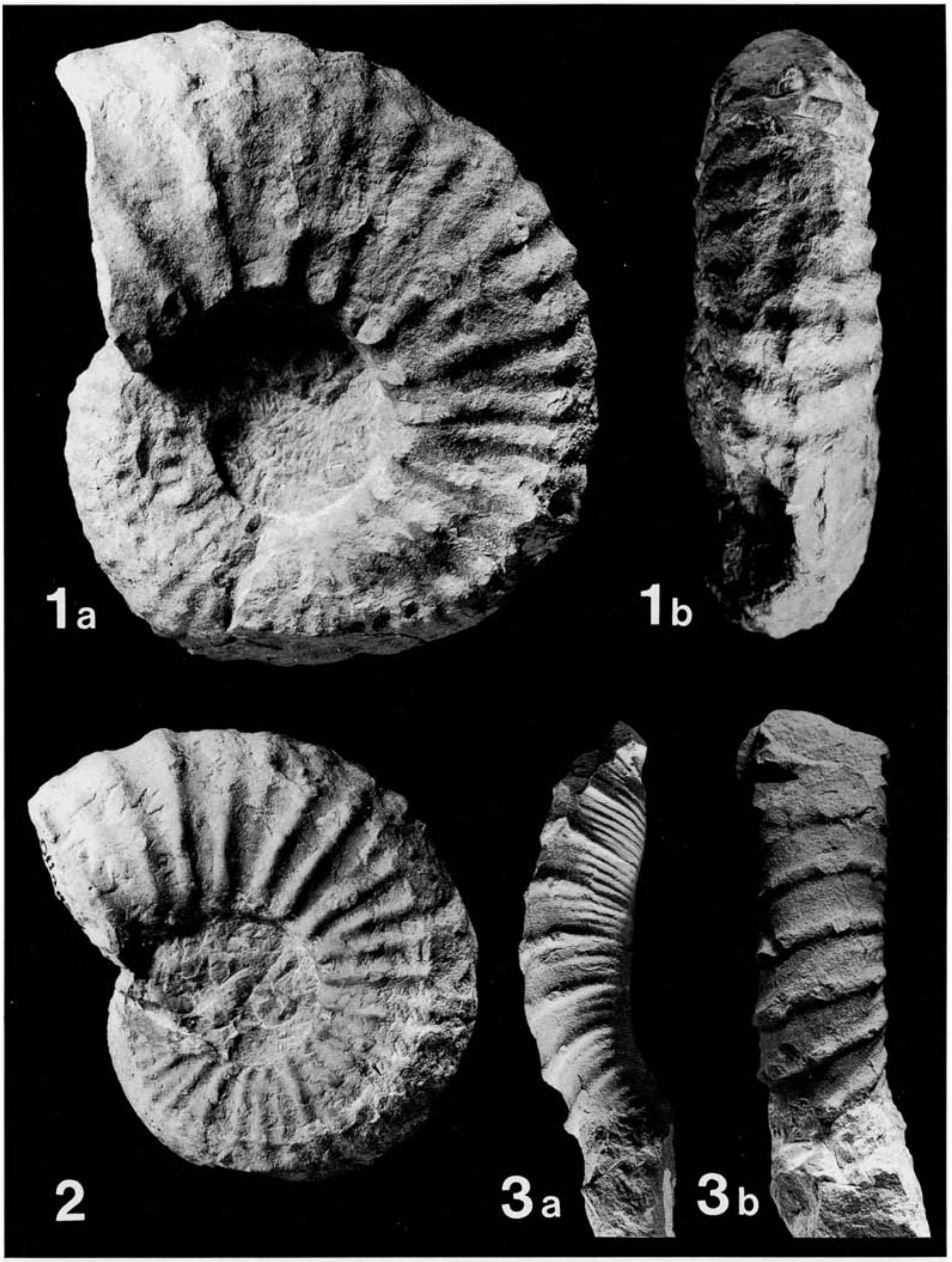

Plate 1

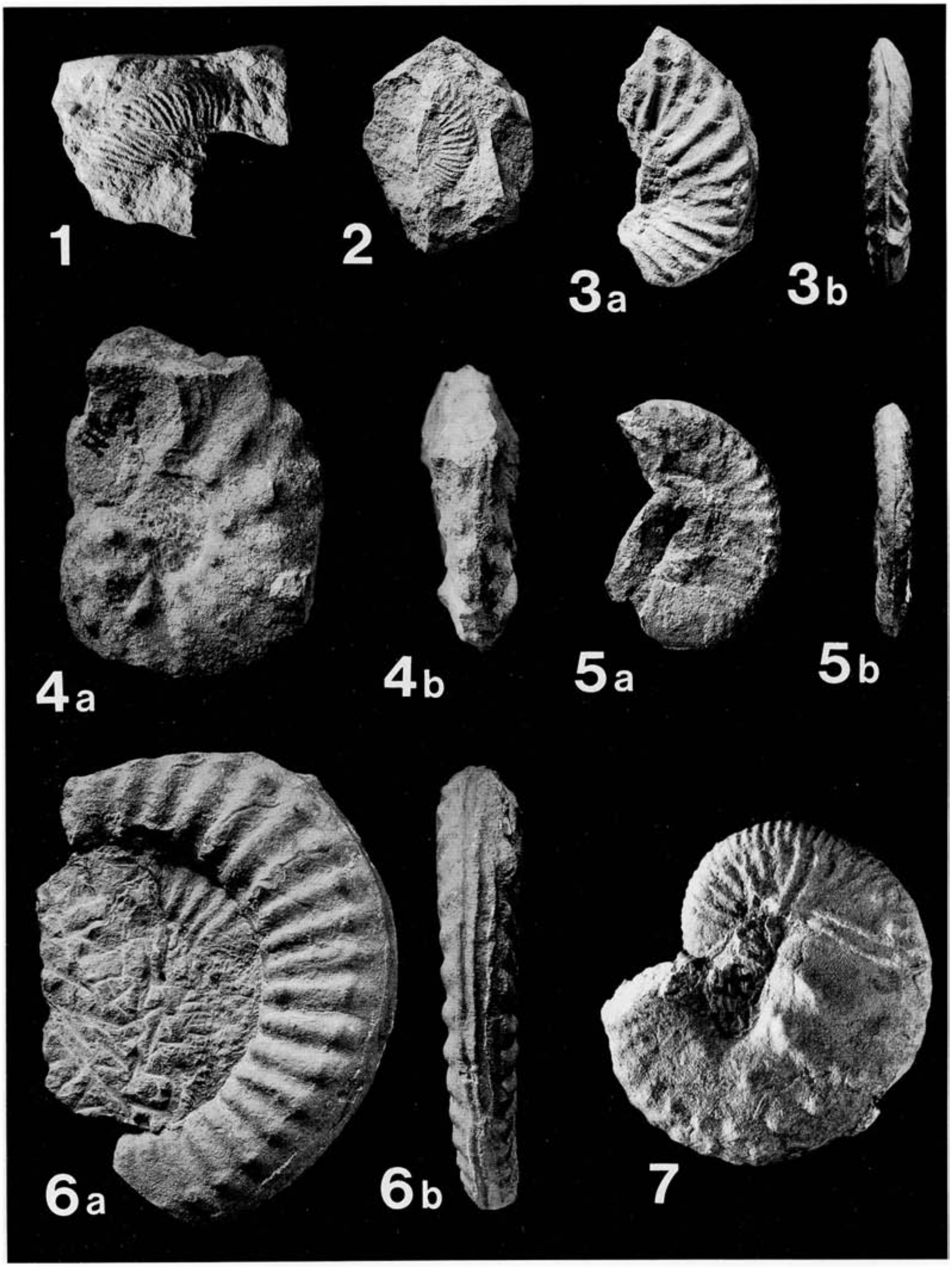

Plate 2

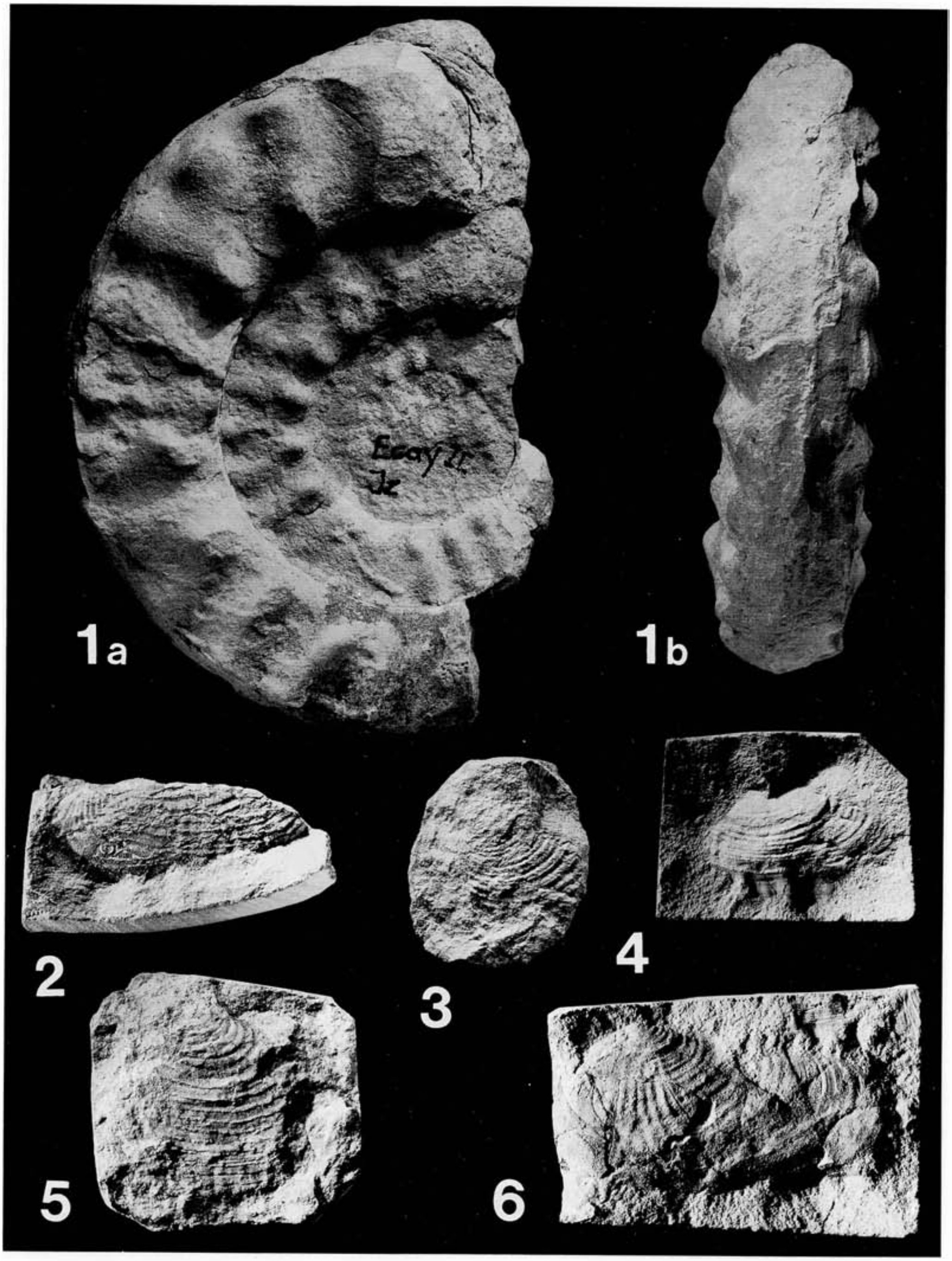

Plate 3

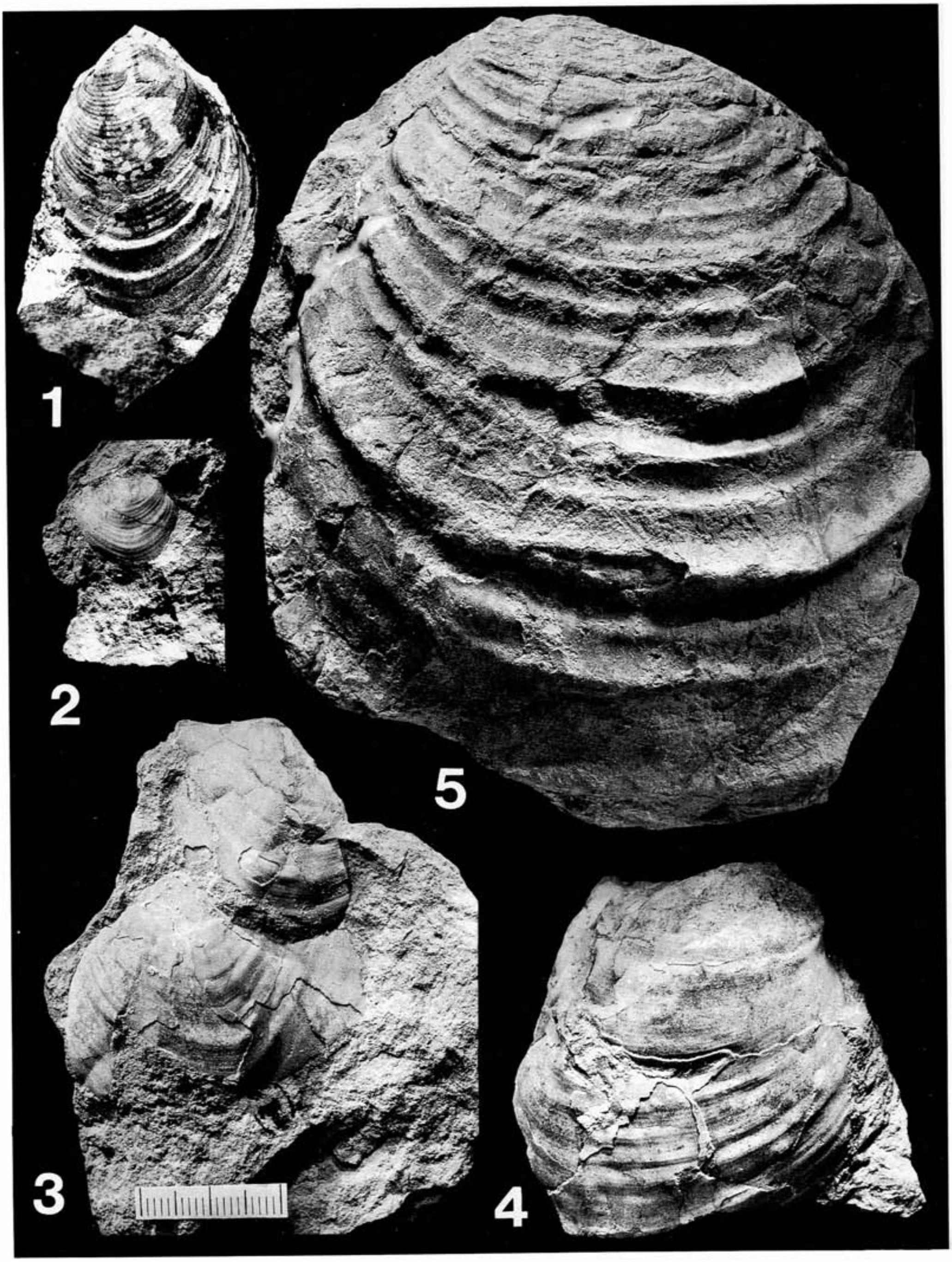

Plate 4

Biostratigraphie des Campan bis Unter-Maastricht der E-Barranca und des Urdiroz/Imiscoz-Gebietes (Navarra, N-Spanien)

Campanian - Lower Maastrichtian Biostratigraphy of the E Barranca and the Urdiroz/Imiscoz Area (Navarra, N Spain)

THOMAS KÜCHLER und ANDREAS KUTZ, Berlin

Mit 8 Text-Figuren

KÜCHLER, TH. & KUTZ, A. (1989): Biostratigraphie des Campan bis Unter-Maastricht der E-Barranca und des Urdiroz/Imiscoz-Gebietes (Navarra, N-Spanien) [Campanian - Lower Maastrichtian Biostratigraphy of the E Barranca and the Urdiroz/Imiscoz Area (Navarra, N Spain)]. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 191-213. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Lithologic successions for the E-Barranca and the area of Urdiroz/Imiscoz in N-Spain (Navarra) are presented and figured in composed standard and selected sections. Biostratigraphical schemes are established for the Campanian and Lower Maastrichtian, based upon ammonites and irregular echinoids (*Echinocorys, Offaster*). Faunal assemblages typical for the regional scheme of subdivision are described. The local subdivisions are correlated by faunal assemblages and bioevents.

Correlation with the "standard" ammonite zones for the Campanian and Lower Maastrichtian as recognized in Navarra is discussed.

The lowest Campanian contains an echinoid assemblage of *Offaster pomeli, M. (Isomicraster)* sp., and *Echinocorys scutata cincta*. Comparable assemblages are known from S-England and NW-Germany within the ammonite zone of Placenticeras bidorsatum. *Scaphites hippocrepis* as was formerly proposed for definition of the base of the Campanian is first recorded in N-Spain from the uppermost Lower Campanian.

The Lower/Upper Campanian boundary is drawn at the level of the first appearance of *Hoplitoplacenticeras marroti*. Late forms of *Sc. hippocrepis* have been quoted co-occurring with *H. marroti* and immediately below the first occurrence of *Trachyscaphites spiniger*. The first appearance of *Tr. spiniger* in association with *Pachydiscus haldemsis* lies above the entry of *H. marroti* into N-Spain.

We divide the traditional Polyplocum Zone by means of the recognizable succession of *Bostrychoceras* and *Didymoceras*, as supplemented by the change in assemblage of other ammonites.

The Campanian/Maastrichtian boundary is defined by the first appearance of *Pachydiscus neubergicus. Hoploscaphites constrictus* enters well above *P. neubergicus.*

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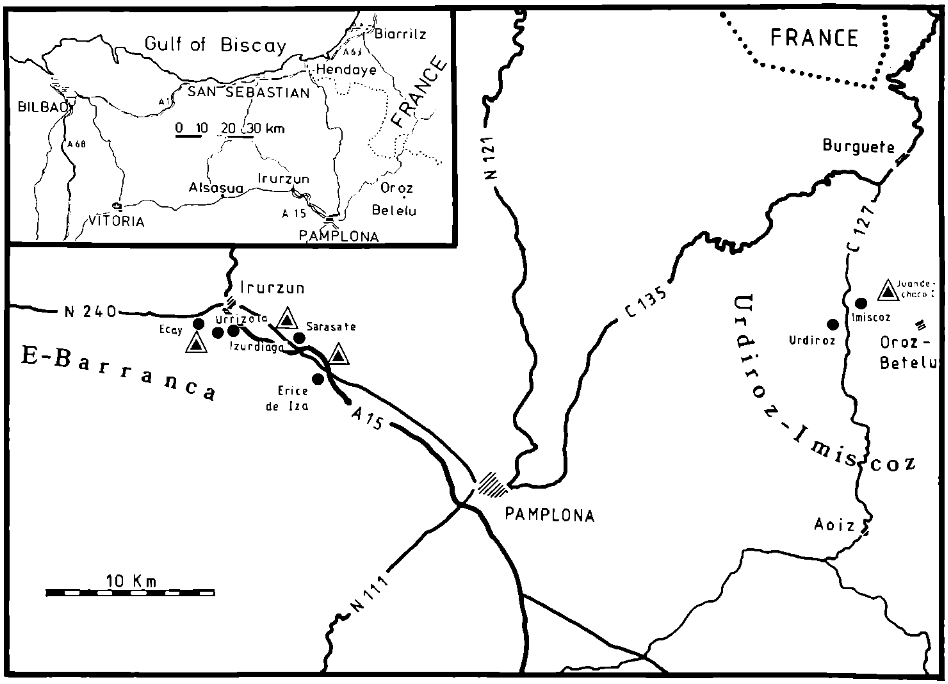
1. Einleitung

Die zwei untersuchten Gebiete (E-Barranca, Urdiroz/Imiscoz) liegen 20 km nordwestlich und 20 km nordöstlich von Pamplona (Text-Fig. 1). Das östliche Gebiet liegt im E-Teil der Barranca, dem schmalen E-W-orientierten Tal des Rio Araquil. Im N wird es von der Sierra de Aralar und im S durch die Sierra de Andia begrenzt.

Das Urdiroz/Imiscoz-Gebiet liegt im Vorland der W-Pyrenäen, nahe der französischen Grenze und des Oroz-Betelu-Massivs.



Text-Fig. 1. Übersichtskarte der untersuchten Gebiete (E-Barranca, Urdiroz/Imiscoz) mit den im Text erwähnten Lokalitäten (ausgefüllte Kreise) und Lage der abgebildeten Campan- und Unter-Maastrichtprofile (Dreiecke).
Text-Fig. 1. General map of studied areas (E Barranca, Urdiroz/Imiscoz), showing localities mentioned in the text (filled circles), with positions of figured Campanian and Lower Maastrichtian sections indicated by triangles.

Die bisherigen spanischen Arbeiten über die E-Barranca (RAMIREZ DEL POZO 1971, CARBAYO et al. 1978) und des Urdiroz/Imiscoz-Gebietes (VILLALOBOS VILCHEZ 1978) waren hauptsächlich mikrostratigraphisch ausgerichtet. Die Makrofauna blieb dabei weitgehend unberücksichtigt.

Eine paläontologische Bearbeitung der Ammoniten und eine erste biostratigraphische Gliederung wurde von WIEDMANN (1962, 1979) für N-Spanien vorgenommen. Aus den von uns bearbeiteten Gebieten beschrieb er bereits einzelne Arten.

RIOS et al. (1944), RIOS & HANCOCK (1961) geben Ammoniten- und Echiniden-Assoziationen aus dem Urdiroz/Imiscoz-Gebiet an, die stratigraphisch höheres Campan anzeigen.

RADIG (1973) untersuchte in der Barranca, im Urdiroz/Imiscoz-Gebiet und in der Mulde von Bôveda (Provinz Burgos) die horizontale und vertikale Verbreitung von Echiniden und gab eine paläontologische Beschreibung dieser Faunen.

Seine Sammelprofile dienten als Ausgangspunkt unserer eigenen Arbeiten (KÜCHLER 1983, KÜCHLER in Vorb., KUTZ 1987).

Die untersuchten Gebiete zeichnen sich durch eine individuen- und artenreiche Makrofauna aus, die sich aus Ammoniten, Echiniden und Inoceramen zusammensetzt. Die Fauna zeigt Übereinstimmungen mit der borealen Faunenprovinz.

Die lokalen biostratigraphischen Gliederungen für das Campan Navarras sind Ergebnisse umfangreicher Profilaufnahmen und horizontierter Fossilaufsammlungen. In der E-Barranca wurden sechzehn, in Urdiroz/Imiscoz sechs Einzelprofile bearbeitet.

In kontinuierlicher Abfolge liegt in der E-Barranca das Unter-Campan bis untere Ober-Campan vor. Die Abfolgen des höheren Ober-Campan und Unter-Maastricht wurden aus isolierten Profilen, die mehr oder weniger große Lücken aufweisen, rekonstruiert.

Westlich des Oroz-Betelu-Massivs ist bei Urdiroz und Imiscoz eine vollständige Abfolge von Ober-Campan bis Unter-Maastricht entwickelt, während das Unter-Campan fast vollständig ausfällt.

Korrelationsmöglichkeiten mit der E-Barranca ergeben sich daher erst mit dem Ober-Campan. Die Korrelation der Gebiete und der Vergleich der lokalen biostratigraphischen Gliederungen wird mit Hilfe übereinstimmender Faunenassoziationen und Bioevents vorgenommen. Für N-Spanien kann somit erstmals für das Unter-Campan sowie für das Ober-Campan bis Unter-Maastricht eine Zonengliederung, basierend auf der sukzessiven Abfolge von Ammoniten- und Echiniden-Faunen, vorgestellt werden.

2. Lithologie

2.1 E-Barranca

Untersuchungen wurden südlich Irurzun, in einem schmalen, NNW-SSE-orientierten Streifen nördlich der Sierra de Andia und Satrustegui durchgeführt. Die Profile sind hauptsächlich als Straßenanschnitte im westlichen Abschnitt der Autobahn A-15 (Irurzun - Pamplona) zwischen Irurzun und Erice de Iza aufgeschlossen. Weitere isolierte Aufschlüsse liegen bei den Ortschaften Izurdiaga, Urrizola und Ecay. Dieses Gebiet begrenzt das Ausstreichen der Sarasate-Formation (Sarasate-Glaukonitmergel, KÜCHLER 1983) in der E-Barranca.

Die **Sarasate-Formation** ist vom Campan bis ins Unter-Maastricht entwickelt. Sie besteht im wesentlichen aus siltigen, in Abschnitten stark glaukonitischen Mergeln im Wechsel mit dünnbankigen Mergelsteinen und erreicht eine Gesamtmächtigkeit von etwa 410 m. Typusregion ist das Gebiet um die Ortschaft Sarasate, 4,5 km südöstlich Irurzun.

Die Basis der Formation, welche mit der Santon/Campan-Grenze korrespondiert, ist definiert mit einem scharfen lithologischen Wechsel von weißgrauen Kalken zu blaugrauen Mergeln über einem erodierten Hartgrund bei Sarasate. Das Campan lagert über diesem Hartgrund mit einer schwachen Winkeldiskordanz. Die lithologischen und biostratigraphischen Befunde zeigen eine Transgression und Schichtlücke im basalen Campan an.

Im Gebiet von Izurdiaga/Urrizola werden Santon-Kalke diskordant von bis zu 60 cm mächtigen Konglomerathorizonten überlagert. Die Schichtlücke umfaßt dort, im Gegensatz zum Gebiet von Sarasate, größere Teile des Santon und fast das gesamte Unter-Campan.

2.1.1 Lithologie des Unter-Campan (Barranca)

Der basale Teil der Formation, welcher hohe Glaukonitanreicherungen aufweist, besteht aus einer alternierenden Folge von siltigen Mergeln mit dünnbankigen (0,1-0,2 m), größtenteils knolligen Mergelsteinen und bis zu 1-2 m mächtigen Mergelsteinbänken. Darüber folgen glaukonitfreie Kalkmergel.

Charakteristisch für die basale Hippocrepis-Zone sind dickbankige siltige Mergelsteine mit extrem hohem Glaukonitgehalt, der dem Gestein eine gelbliche, im Extremfall violettgraue Verwitterungsfarbe verleiht. Die Schichten zeichnen sich durch Kondensation bzw. Mangelsedimentation aus.

In der Brevis/Humilis-Zone dominieren wiederum Mergel.

Im höheren Teil der darauffolgenden Mergel/Mergelstein-Sequenz kommen in den dünnen Mergelsteinbänken Schwämme häufig vor. Dieser schwammreiche Abschnitt ist typisch für den Übergangsbereich Unter/Ober-Campan.

2.2 Urdiroz/Imiscoz

Das Gebiet liegt etwa 2 km westlich des "Baskischen Massives" von Oroz-Bételu, welches von Kalken des Devon und Buntsandstein-Sedimenten aufgebaut wird. Die einzelnen Profile wurden nahe der Ortschaften Imiscoz und Urdiroz sowie am Monte Juandechaco aufgenommen.

Die ältesten aufgeschlossenen Oberkreide-Schichten gehören ins Ober-Santon, zu der von AMIOT (1982) aufgestellten "Formación de Espinal-Oroz-Bételu". Sie besteht aus Dolomiten und Kalken. In den höheren Niveaus enthalten die Kalke *Lacazina elongata* MUNIER-CHALMAS.

Am Top der Formation ist ein ca. 30 cm mächtiger, glaukonitischer Hartgrund entwickelt. Der Übergang zum Campan zeichnet sich wie in der Barranca durch einen scharfen Fazieswechsel aus. Die Basis des Campan wird gekennzeichnet durch einen ca. 1 m mächtigen Horizont, der glaukonitumkrustete Kalkgerölle führt. Dieser wird überlagert von einer monotonen Wechselfolge aus Mergeln und Mergelsteinen von 445 m Mächtigkeit am Monte Juandechaco.

Die Wechselfolge umfaßt das Ober-Campan und ist bis ins Ober-Maastricht entwickelt. Das Unter-Campan fällt fast vollständig aus. Die unteren 270 m der lithologischen Einheit - Ober-Campan und Unter-Maastricht - wurden biostratigraphisch gegliedert.

2.3 Lithologischer Vergleich des Ober-Campan zwischen der E-Barranca und Urdiroz/Imiscoz

Das Ober-Campan im Urdiroz/Imiscoz-Gebiet und in der E-Barranca zeigt generelle lithologische Ähnlichkeiten. Zwei größere lithologische Einheiten lassen sich in beiden Gebieten unterscheiden.

Im tieferen Ober-Campan ist eine deutliche Mergel-Vormacht zu verzeichnen.

In der höheren Polyplocum-Zone tritt ein markanter sedimentologischer Wechsel zu rhythmischen siltig-sandigen Mergeln und Mergelsteinbänken ein. Dieser Fazieswechsel ist in der Barranca ausgeprägter als in Urdiroz/Imiscoz. In der Regel sind die Abfolgen außerordentlich reich an *Inoceramus* ex gr. *balticus/regularis*, sandschaligen Foraminiferen (*Navarella juaquini* CIRY & RAT) und *Pycnodonte vesicularis*.

In Urdiroz/Imiscoz nehmen im höheren Teil der Wechselfolge die Mergel gegenüber den Mergelsteinbänken an Mächtigkeit zu. Im Übergangsbereich Campan/Maastricht treten flaserige bis knollige Kalkmergel auf. Die bisher typischen Inoceramen-Lagen fehlen. In der Barranca ist die gesamte Abfolge durch mehr oder weniger stark glaukonitische Sedimente gekennzeichnet.

3. Biostratigraphie

Die Zonengliederungen der E-Barranca und des Urdiroz/Imiscoz-Gebietes basieren auf Ammoniten und irregulären Echiniden, deren relative Häufigkeit und lokale Vertikalreichweiten in Normalprofilen für die Gebiete gezeigt werden (Text-Fig. 2, 5a, b).

Die vier "Standard"-Ammoniten-Zonen für das Campan (Bidorsatum-, Hippocrepis-, Marroti- und Polyplocum-Zone) und die Neubergicus-Zone für das Unter-Maastricht sind in Navarra erkennbar. Bis auf die Placenticeras bidorsatum-Zone, welche durch eine für diesen Zeitabschnitt charakteristische Echiniden-Assoziation indiziert wird, sind die Zonen durch die Leitarten belegt.

Die Ammonitengliederung konnte hinsichtlich Hippocrepis- und Marroti-Zone durch Echiniden-Range- und Peakzonen sowie Ammoniten/Echiniden-Assemblage-Zonen verfeinert werden, wobei bei der Gliederung mit irregulären Echiniden Arten aus den Evolutionslinien der Gattungen *Echinocorys* und *Offaster* benutzt wurden.

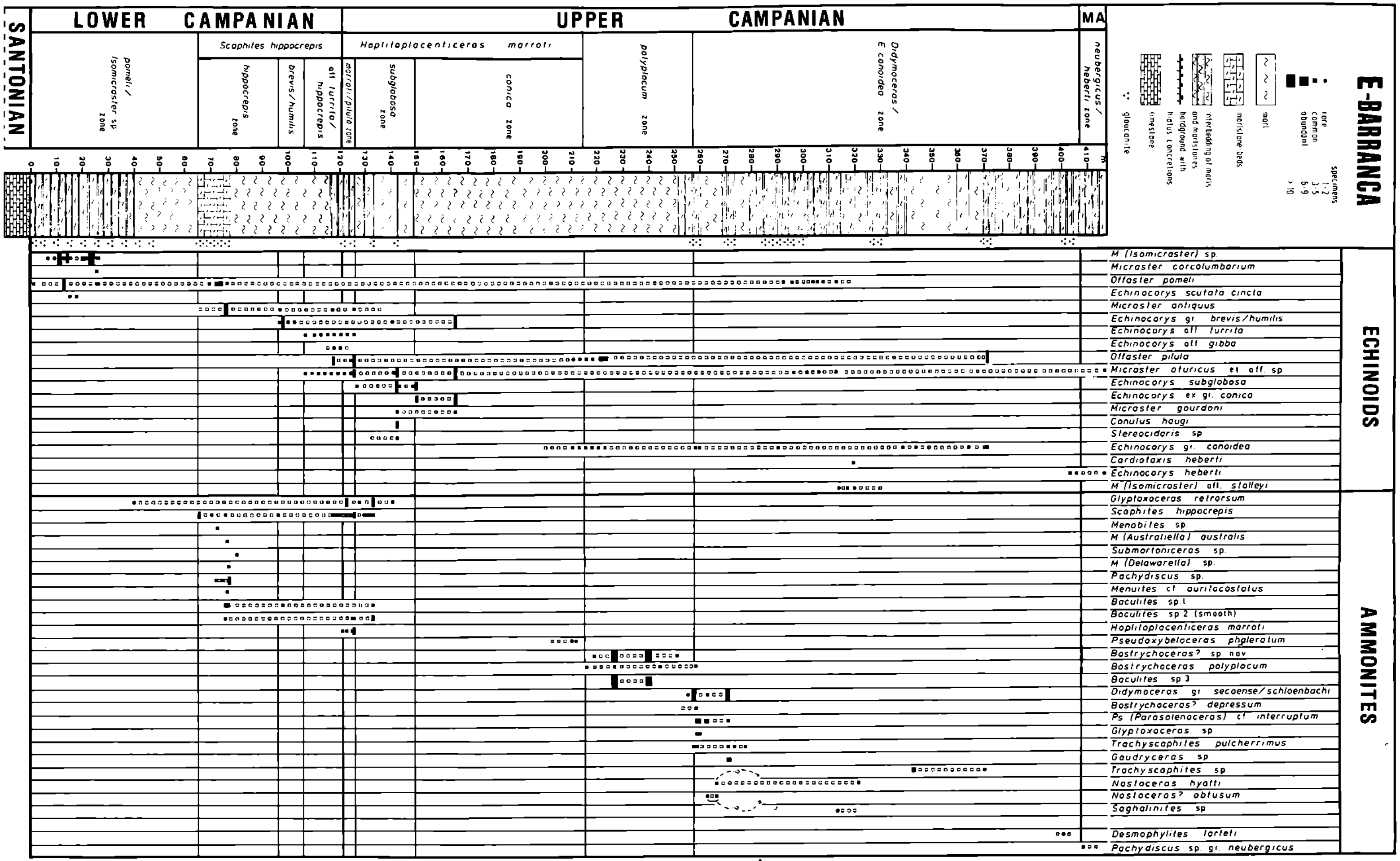
Die Verbreitung der Gattung *Offaster* scheint jedoch in Navarra auf die E-Barranca beschränkt. In diesem Gebiet ist die *Offaster*-Linie vom Unter-Santon bis ins höhere Ober-Campan belegt (KÜCHLER, in Vorb.). Im Gebiet von Urdiroz/Imiscoz sowie im westlichen Teil der Barranca wurden bisher keine *Offaster* nachgewiesen.

3.1 Unter-Campan in der E-Barranca (Text-Fig. 2)

3.1.1 Santon/Campan-Übergang

In der E-Barranca fallen stratigraphisch Teile des höheren Ober-Santon und das tiefste Unter-Campan aus. Die Santon/Campan-Grenze ist eine lithologische Grenze.

Akzeptiert man das Aussterben von *Marsupites testudinarius* SCHLOTHEIM als Marker für den Top des Santon, umfaßt die Schichtlücke im Campan der



SANTONIAN
LOWER CAMPANIAN
UPPER CAMPANIAN
MA
Scaphites hippocrepis
Hoplitoplacenticeras marroti
pomeli / Isomicraster sp zone
hippocrepis zone
brevis/humilis
aff turrita / hippocrepis
marroti/pilula zone
subglobosa zone
conica zone
polyplocum zone
Didymoceras / E conoidea zone
neubergicus / heberti zone
E-BARRANCA
specimens
rare 1-2
common 3-5
abundant 6-9
>10
marl
marlstone beds
interbedding of marls and marlstones
hardground with hiatus concretions
limestone
glauconite
ECHINOIDS
M (Isomicraster) sp.
Micraster corcolumbarium
Offaster pomeli
Echinocorys scutata cincta
Micraster antiquus
Echinocorys gr. brevis/humilis
Echinocorys aff turrita
Echinocorys aff gibba
Offaster pilula
Micraster aturicus et aff. sp
Echinocorys subglobosa
Echinocorys ex gr. conica
Micraster gourdoni
Conulus haugi
Stereocidaris sp
Echinocorys gr. conoidea
Cardiotaxis heberti
Echinocorys heberti
M (Isomicraster) aff. stolleyi
AMMONITES
Glyptoxoceras retrorsum
Scaphites hippocrepis
Menabites sp.
M (Australiella) australis
Submortoniceras sp.
M (Delawarella) sp.
Pachydiscus sp.
Menuites cf aurilocostatus
Baculites sp.1
Baculites sp 2 (smooth)
Hoplitoplacenticeras marroti
Pseudoxybeloceras phaleratum
Bostrychoceras? sp nov
Bostrychoceras polyplocum
Baculites sp 3
Didymoceras gr secoense/schloenbachi
Bostrychoceras? depressum
Ps (Parasolenoceras) cf interruptum
Glyptoxoceras sp
Trachyscaphites pulcherrimus
Gaudryceras sp
Trachyscaphites sp.
Nostoceras hyatti
Nostoceras? obtusum
Saghalinites sp
Desmophylites larteti
Pachydiscus sp. gr. neubergicus

Barranca, im Vergleich mit NW-Deutschland, vermutlich die Granulataquadrata- und die Lingua/Quadrata-Zonen.

In der E-Barranca ist das höhere Santon zur Zeit schlecht durch Makrofossilien belegt. Die wenigen Funde, wie *Texanites* sp. und *Cardiaster integer* (AGASSIZ), *Echinocorys scutata vulgaris* (BREYNIUS) und *Peroniaster* sp., werden von RADIG (1973) und DEGENHARDT (1983) angegeben. Die Fundschichten liegen in einem Profil bei der Ortschaft Izurdiaga etwa 45 m bzw. 15 m unterhalb der lithologischen Santon/Campan-Grenze.

3.1.2 Pomeli/Isomicraster sp.-Zone

Die Profile südlich Sarasate enthalten das am vollständigsten entwickelte Unter-Campan in der E-Barranca.

Bei Sarasate wird die provisorische Santon/Campan-Grenze mit dem scharfen sedimentären Wechsel und dem Erstauftreten von *Offaster pomeli* MUNIER-CHALMAS - unmittelbar über dem Tophartgrund des Santon - gezogen.

Offaster pomeli steht in engem verwandtschaftlichem Verhältnis zu *Offaster pilula* (LAMARCK). Letzterer ist in der borealen Kreide im Unter-Campan verbreitet und wird zur Zonengliederung der englischen und nordwestdeutschen Schreibkreide benutzt.

In der E-Barranca liefert die *Offaster*-Linie im Unter-Campan mit *Offaster pomeli* und im Grenzbereich Unter/Ober-Campan mit *Offaster pilula* brauchbare Leitformen und Events. Ein erster Peak von *Offaster pomeli* läßt sich etwa 13 m über der postulierten Stufengrenze nachweisen. Er fällt in die Akmezone von *Micraster (Isomicraster)* sp., in der gleichzeitig *Echinocorys scutata cincta* BRYDONE auftritt.

Die aufgeführte Echiniden-Assoziation ist von überregionaler Bedeutung, da sie eine Korrelation des nordspanischen Profiles mit denen von S-England und NW-Deutschland erlaubt. Ein Korrelationsdiagramm der nordwestdeutschen und südenglischen Zonengliederungen geben CHRISTENSEN & SCHMID (1987). In S-England tritt *Offaster pilula* gehäuft zusammen mit *E. cincta* in der höheren Offaster pilula-Zone auf, unterhalb der Schichten mit *Hagenowia blackmorei* (GASTER 1924, BAILEY et al. 1983).

Aus Lägerdorf (NW-Deutschland) gibt ERNST (1963) einen Peak von *Offaster pilula* zusammen mit *Offaster pomeli* am Top der Pilula-Zone an. Aus Niedersachsen berichtet ERNST (1970) über denselben Peak zusammen mit *M. (Gibbaster)* sp. in der Pilula/Senonensis-Subzone.

Diese Niveaus liegen in NW-Deutschland oberhalb des Erstauftretens von *Placenticeras bidorsatum* (ROEMER) (ERNST et al. 1979, Fig. 11) und unterhalb des gesicherten Einsetzens von *Scaphites hippocrepis* (DEKAY) so-

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Text-Fig. 2. Zusammengesetztes Normalprofil des Campan bis Unter-Maastricht in der E-Barranca, mit den Reichweiten von Ammoniten und Echiniden und einer integrierten Biozonierung.

Text-Fig. 2. Ammonite and echinoid range chart and integrated biozonation of the E-Barranca in a composed standard section of the Campanian and Lower Maastrichtian.

wohl in NW-Deutschland als auch in S-England (SCHMID & ERNST 1975, ERNST et al. 1979, BAILEY et al. 1983).

Trotz des Fehlens des Leitammoniten ist es daher möglich, die Schichten der E-Barranca der Bidorsatum-Zone zuzuordnen.

Die Verbreitung von *Placenticeras bidorsatum* scheint nach KENNEDY (1984, 1986) auf das nordwestliche Europa (Westfalen und Niedersachsen (D), Aquitaine (F)) beschränkt zu sein.

3.1.3 "Standard"-Zone des Scaphites hippocrepis

In den USA tritt *Scaphites hippocrepis* im tiefsten Unter-Campan auf (COBBAN 1969, GILL & COBBAN 1973). Aus Europa sind lediglich *Sc. hippocrepis* II-III nachgewiesen und - wie bereits von KENNEDY (1986: 13) ausführlich diskutiert - setzen diese "Subspezies" in Europa über *P. bidorsatum* ein.

In der E-Barranca, wo die Pomeli/Isomicraster sp.-Zone das tiefere Unter-Campan repräsentiert, ist *Scaphites hippocrepis* III ebenfalls erst aus dem höheren Unter-Campan nachgewiesen.

Die "Standard"-Zone des Sc. hippocrepis, deren Obergrenze in der E-Barranca durch das Einsetzen von *Hoplitoplacenticeras marroti* (COQUAND) definiert ist, läßt sich auf der Basis von *Echinocorys*-Arten detaillierter gliedern.

Die Hippocrepis-Zone in unserem Sinne repräsentiert nur den unteren Teil der "Standard"-Zone (vgl. Text-Fig. 4).

Die Basis der Zone ist definiert durch das Erstauftreten von kleinwüchsigen *Scaphites hippocrepis* III, welches mit einem Häufigkeitsmaximum der Spezies zusammenfällt. Über diesem Peak kommt die Art nur noch vereinzelt vor, zusammen mit seltenen Funden von *Menabites* sp., *Menabites (Australiella) australis* (BESAIRIE), *Delawarella* sp. und *Submortoniceras* sp., welche die Schichten bereits ins höhere Unter-Campan bzw. ins "Mittel-Campan" (im Sinne von WIEDMANN 1979) verweisen.

Die Zone zeichnet sich weiterhin durch die Häufigkeit von *Offaster pomeli, Micraster antiquus* COTTEAU, Baculiten und Pachydisciden, unter anderem *Menuites* cf. *auritocostatus* (SCHLÜTER), aus.

Dieser Zeitabschnitt dürfte den von WIEDMANN (1979) für N-Spanien ausgehaltenen Zonen des Sc. hippocrepis und der Delawarella delawarensis entsprechen. Im Vergleich mit Frankreich (KENNEDY 1986) zeigt die sukzessive Abfolge der Ammoniten Übereinstimmungen mit dem oberen Teil der P. bidorsatum- und Delawarella delawarensis-Zone.

Darüber schließt sich in der Barranca eine Echiniden-Peakzone der Echinocorys gr. brevis/humilis an.

Die "aff. turrita/Hippocrepis-Zone" enthält mit *Echinocorys* aff. *turrita* LAMBERT, *Echinocorys* aff. *gibba* (LAMARCK), *Micraster aturicus* SEUNES und *Offaster pilula* neue Faunenelemente. Der höhere Teil der Zone zeich-

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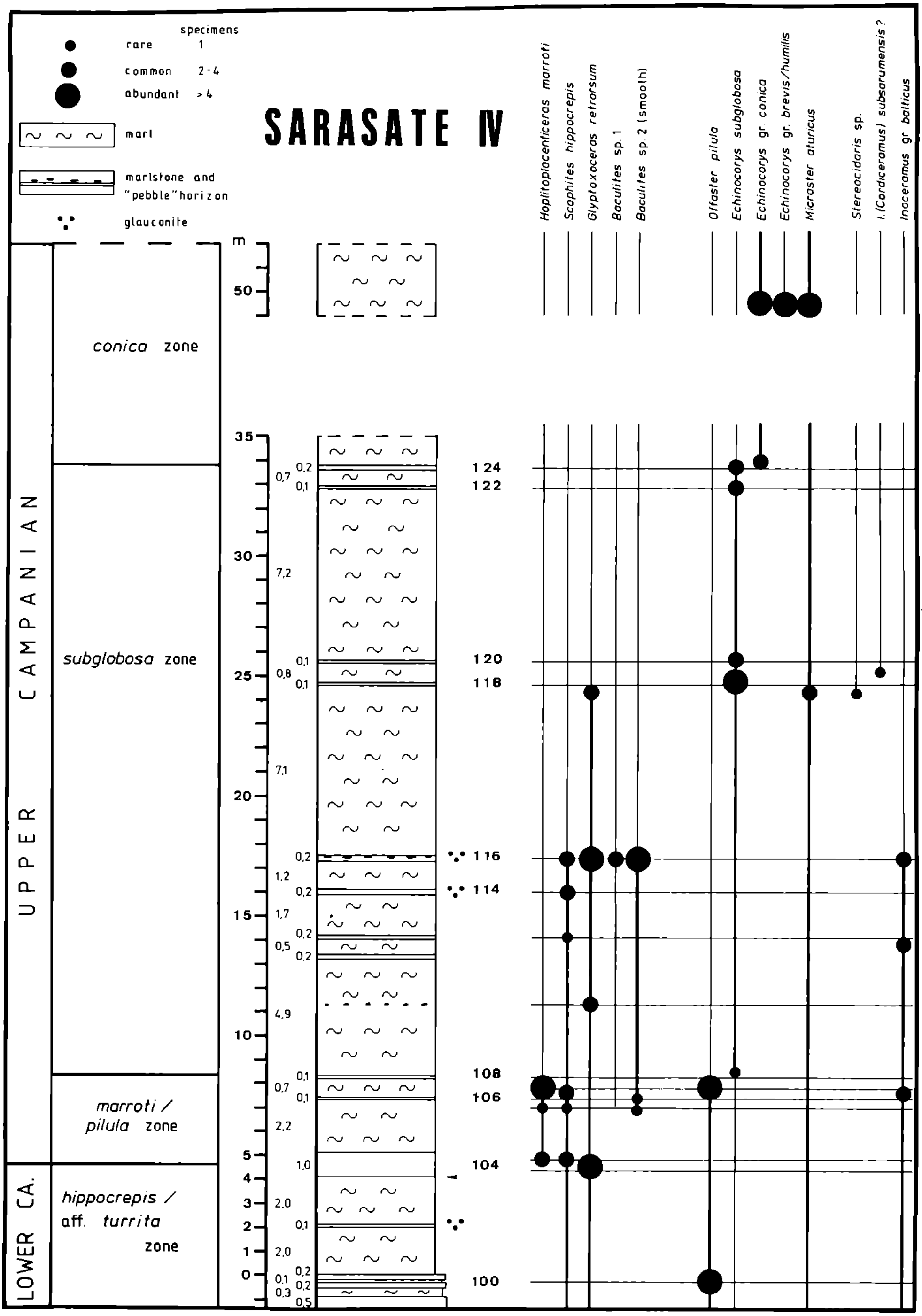
Text-Fig. 3. Sarasate IV (E-Barranca) - Profil des Unter/Ober-Campan-Grenzbereiches.

Text-Fig. 3. Sarasate IV (E Barranca) - section of Lower/Upper Campanian boundary.



specimens
rare 1
common 2-4
abundant > 4
marl
marlstone and "pebble" horizon
glauconite
SARASATE IV
Hoplitoplacenticeras marroti
Scaphites hippocrepis
Glyptoxoceras retrorsum
Baculites sp. 1
Baculites sp. 2 (smooth)
Offaster pilula
Echinocorys subglobosa
Echinocorys gr. conica
Echinocorys gr. brevis/humilis
Micraster aturicus
Stereocidaris sp.
I. (Cordiceramus) subsarumensis ?
Inoceramus gr. balticus
UPPER CAMPANIAN
LOWER CA.
conica zone
subglobosa zone
marroti / pilula zone
hippocrepis / aff. turrita zone
m
50
35
30
25
20
15
10
5
4
3
2
1
0
124
122
120
118
116
114
108
106
104
100

net sich durch die Häufigkeit von *Offaster pilula* und *Sc. hippocrepis* aus.

3.2 **Unterstufengrenze Unter/Ober-Campan** (Text-Fig. 3)

Das Einsetzen der Gattung *Hoplitoplacenticeras* ist in Europa weitgehend akzeptiert für die Unter/Ober-Campan-Grenzziehung auf der Basis von Ammoniten (BIRKELUND 1965, ATABEKJAN 1979, KENNEDY 1984, 1986, MARTINEZ 1982, BLASZKIEWICZ 1979, 1980).

Die Marroti-Zone kann erstmals in N-Spanien (Barranca) mit einer hochdiversen Ammoniten- und Echinidenfauna belegt werden. Die Basis wird mit dem Einsetzen von *Hoplitoplacenticeras marroti* (COQUAND) definiert und liegt in der Barranca etwa 2 m unterhalb eines sehr markanten Horizontes, welcher sich neben dem etwas häufigeren Auftreten von *H. marroti* durch das massenhafte Vorkommen von *Offaster pilula* auszeichnet.

Die Übergangsschichten enthalten Ammoniten, die in NW-Deutschland und S-England (GIERS 1964, SCHMID & ERNST 1975, GALE 1980) als Unter-Campan-Arten eingestuft werden, da sie im "Overlap-Bereich" von *Goniotheuthis quadrata* und *Belemnitella mucronata* auftreten.

In der E-Barranca wurden dagegen erstmals Spätformen von *Scaphites hippocrepis* und *Glyptoxoceras retrorsum* (SCHLÜTER) vergesellschaftet mit glatten und berippten Baculiten nachgewiesen, die die durch *H. marroti* definierte Unterstufengrenze überschreiten.

Ein einziger Fund von *Sc. hippocrepis* aus dem Urdiroz/Imiscoz-Gebiet ist gleichzeitig der höchste stratigraphische Nachweis dieser Art. Das Exemplar entstammt einem Horizont ca. 1 m unterhalb des exakt horizontierten Fundes von *Trachyscaphites spiniger* (SCHLÜTER).

Trachyscaphites spiniger wurde in N-Spanien (WIEDMANN 1979) und in Polen (BLASZKIEWICZ 1979, 1980) aufgrund des Fehlens von Hoplitoplacenticeraten als Indikator für das basale Ober-Campan benutzt.

Nach unseren Ergebnissen kennzeichnet *Tr. spiniger* in N-Spanien nicht das basale Ober-Campan. Im Vergleich beider Gebiete setzt *Tr. spiniger* über *H. marroti* ein und ist daher nicht für die Grenzziehung geeignet.

3.3 **Ober-Campan (Korrelation E-Barranca und Urdiroz/Imiscoz)** (Text-Fig. 4)

3.3.1 "Standard"-Zone des Hoplitoplacenticeras marroti

In der E-Barranca schließt die "Standard"-Ammoniten-Zone des H. marroti die Marroti/Pilula-, Echinocorys subglobosa- und die E. conica-Zonen ein.

Durch einen deutlichen Peak von *E.* gr. *conica* - in der Barranca im unteren Teil der Conica-Zone - läßt sich das Zonenschema mit dem von Urdiroz/Imiscoz korrelieren. Dort tritt dieser Peak an der Basis der Haldemsis/Conoidea-Zone auf (Text-Fig. 5a, b).

Der mittlere Teil des unteren Ober-Campan ist identisch mit der E. conica/subglobosa- und der basalen Haldemsis/Conoidea-Zone (Urdiroz/Imiscoz). Dies wird weiterhin indiziert durch das Vorkommen von: *Sc. hippocrepis* Spätformen, *E. subglobosa* (GOLDFUSS), *E.* gr. *conica* und *Micraster gourdoni* COTTEAU.

Die Unterstufengrenze als solche ist nur in der E-Barranca mit dem Auftreten von *H. marroti* dokumentiert.

Substages	Standard Ammonite zones	Local zone scheme of the E-Barranca	Local zone scheme of the Urdiroz/Imiscoz area
LOWER MAASTRICHTIAN	*neubergicus* zone	*P. neubergicus* + *E. heberti*	*P. neubergicus / E. heberti*
UPPER CAMPANIAN	*polyplocum* zone	*Didymoceras / E. conoidea*	*Didymoceras / E. conoidea*
		B. polyplocum	*B. polyplocum*
	marroti zone	*E. conica*	*P. haldemsis / E. conoidea*
		E. subglobosa	*E. subglobosa / E. conica*
		H. marroti / O. pilula	
LOWER CAMPANIAN	*hippocrepis* zone	*E. aff. turrita / Sc. hippocrepis*	? ?
		E. brevis / humilis	
		Sc. hippocrepis	
	bidorsatum zone	*O. pomeli / Isomicraster* sp.	
SANTONIAN			

Text-Fig. 4. Stratigraphisches Korrelationsdiagramm des Campan und Unter-Maastricht zwischen der E-Barranca und dem Urdiroz/Imiscoz-Gebiet.
Text-Fig. 4. Stratigraphic correlation diagram of the Campanian and Lower Maastrichtian between the E Barranca and the area of Urdiroz/Imiscoz.

In Urdiroz/Imiscoz ist die Grenze nicht exakt zu fassen (confidence zone of Lower/Upper Campanian boundary). *Trachyscaphites spiniger*, der in der E-Barranca fehlt, setzt in Urdiroz/Imiscoz an der Basis der Haldemsis/Conoidea-Zone ein. Im Vergleich mit der E-Barranca liegt sein Einsetzen oberhalb des von *H. marroti* (Text-Fig. 2, 4, 5a, 5b).

Die Fauna der Haldemsis/Conoidea-Zone (Urdiroz/Imiscoz) setzt sich im wesentlichen zusammen aus *Tr. spiniger*, *Pachydiscus haldemsis* (SCHLÜTER) und *E.* gr. *conoidea*. In der E-Barranca ist *E.* gr. *conoidea* erst im höheren Teil der äquivalenten Conica-Zone zusammen mit *Pseudoxybeloceras phaleratum* (GRIEPENKERL) nachweisbar.

Diese Ammoniten-Assoziation ist typisch für die Phaleratum-Zone in Polen (BLASZKIEWICZ 1980).

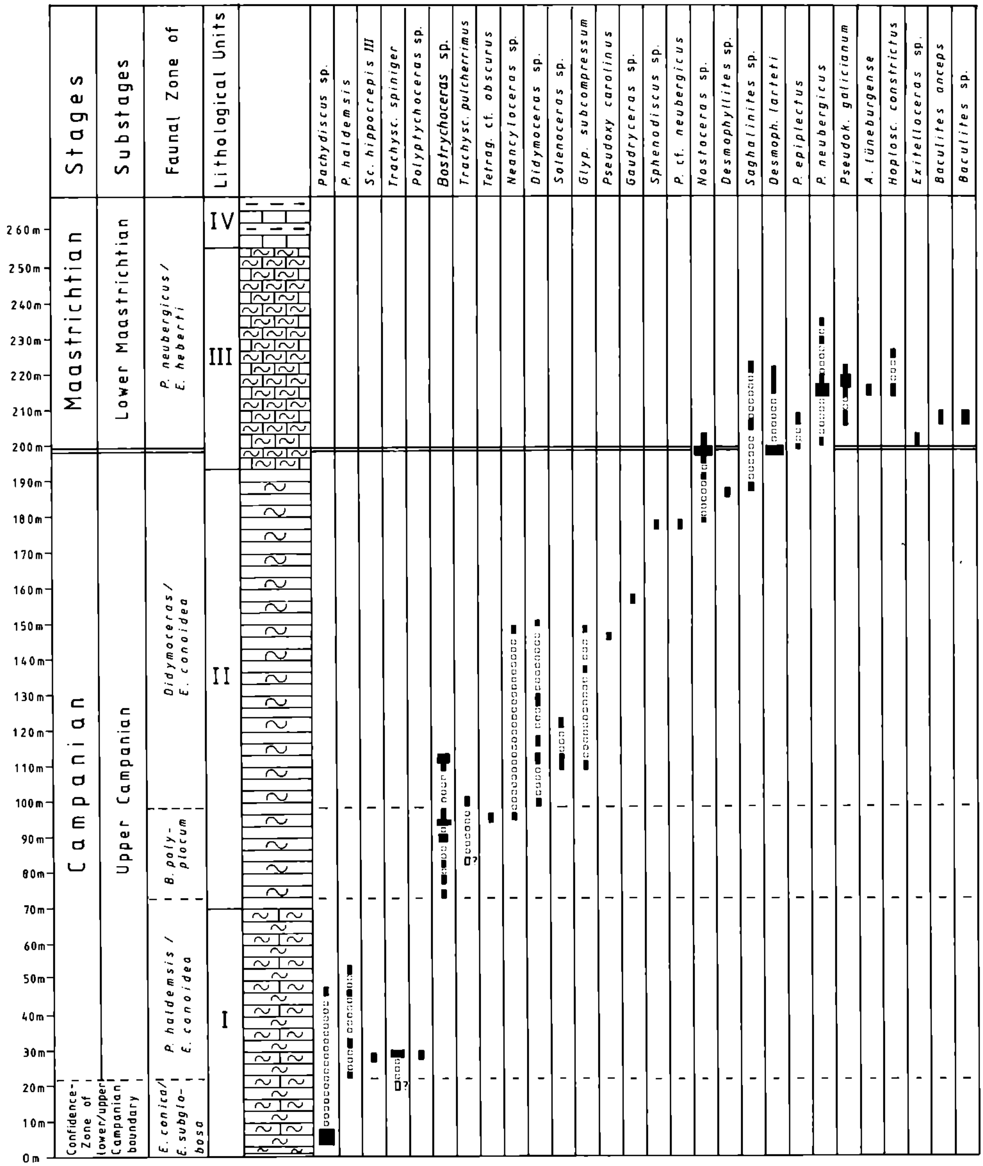
3.3.2 Oberes Ober-Campan

Die traditionelle Polyplocum-Zone konnte, ähnlich wie in Polen (BLASZKIEWICZ 1979, 1980), auf der Basis der sukzessiven Abfolge von *Bostrychoceras* und *Didymoceras* gegliedert werden.

Bei der Interpretation der Gattungen *Bostrychoceras*, *Didymoceras* und *Nostoceras* folgen wir weitgehend den Ansichten von BLASZKIEWICZ (1980: 19, 21).

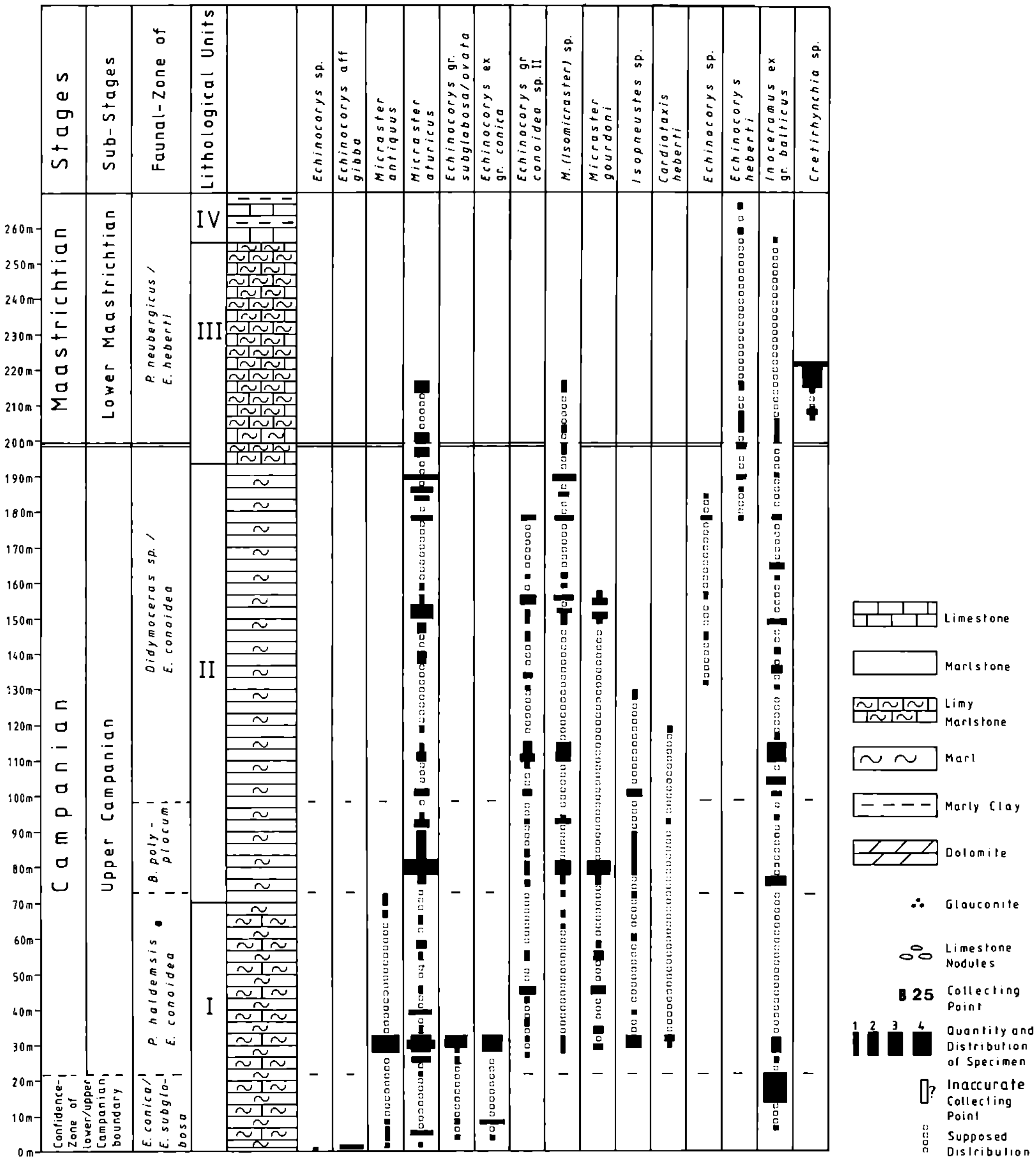
Während BLASZKIEWICZ eine Dreiteilung der traditionellen Polyplocum-Zone in eine Bostrychoceras polyplocum-, Didymoceras donezianum- und

Urdiroz / Imiscoz



Text-Fig. 5. Gliederung des Ober-Campan und Unter-Maastricht im Urdiroz/ Imiscoz-Gebiet.
5a) Stratigraphische Reichweite der Ammoniten.
5b) Stratigraphische Reichweite der Echiniden.

Urdiroz/Imiscoz



Text-Fig. 5. Subdivision of the Upper Campanian and Lower Maastrichtian in the area of Urdiroz/Imiscoz.
5a. Stratigraphic ranges of ammonites.
5b. Stratigraphic ranges of echinoids and inoceramids.

Nostoceras pozaryskii-Zonen vornimmt, läßt sich in N-Spanien vorerst nur eine Zweiteilung in Polyplocum- und Didymoceras/E. conoidea-Zone durchführen. Die Zonengliederung wird zusätzlich gestützt durch zwei unterschiedliche Faunen-Assoziationen in chronologisch aufeinanderfolgenden Niveaus. Die Abtrennung einer Nostoceras-Zone, wie in Polen, ist zur Zeit wegen zu geringer Funde nicht vertretbar.

3.3.2.1 Polyplocum-Zone

Die Basis der Zone, im Sinne der Autoren, ist gekennzeichnet durch das gleichzeitige Einsetzen der Index-Spezies *Bostrychoceras polyplocum* (ROEMER) und *Bostrychoceras?* sp. nov., einer von uns vorerst in offener Nomenklatur belassenen zweiknotigen Form, die KENNEDY (1983, pl. 3, fig. 8; 1986, pl. 15, figs. 3, 5, 6, 7?) noch zu *N. (Bostrychoceras) polyplocum* stellt und BLASZKIEWICZ (1980: 22, pl. 4, figs. 1, 2; pl. 5, fig. 7) als *Didymoceras* sp. beschreibt. Die von SCHLÜTER (1872, pl. 33, fig. 3, 4) und KENNEDY (1986, fig. 32A, B) abgebildeten Formen sowie *B. polyplocum polyplocum* (ROEMER) (BLASZKIEWICZ 1980, pl. 1, figs. 1-9; pl. 2, figs. 2, 3, 5, 6) und *B. polyplocum schlueteri* BLASZKIEWICZ (BLASZKIEWICZ 1980, pl. 2, figs. 1, 4, 9-11) entsprechen den von uns als *B. polyplocum* bezeichneten Formen.

B. polyplocum tritt in N-Spanien außerordentlich selten auf. Häufiger und damit markanter ist für N-Spanien und wahrscheinlich auch für Frankreich *Bostrychoceras?* sp. nov., dessen Reichweite sich mit der von *B. polyplocum* deckt.

So tritt *Bostrychoceras?* sp. nov. in der Barranca in mehreren Horizonten massenhaft auf, während *B. polyplocum* auf Einzelfunde beschränkt bleibt (Text-Fig. 6). Im Urdiroz/Imiscoz-Gebiet ist bisher nur *Bostrychoceras?* sp. nov. nachgewiesen worden.

Die Begleitfauna unterscheidet sich, mit Ausnahme von *E.* gr. *conoidea* und dem gehäuften Auftreten von *Inoceramus* ex gr. *balticus/regularis,* in Urdiroz/Imiscoz durch das Massenauftreten von *Micraster* div. sp. und in der Barranca durch das Vorkommen von grobberippten Baculiten und *Offaster pilula.*

3.3.2.2 Didymoceras/E. conoidea-Zone

Ein kontinuierlicher Übergang von der Polyplocum-Zone zur Didymoceras/E. conoidea-Zone ist im Gebiet von Urdiroz/Imiscoz, Profil Juandechaco I (Text-Fig. 7) entwickelt. Das Juandechaco-Profil kann als regionale Standard-Section für das höhere Ober-Campan und den Grenzbereich Campan/Maastricht betrachtet werden. Die Profile in der E-Barranca zeichnen sich

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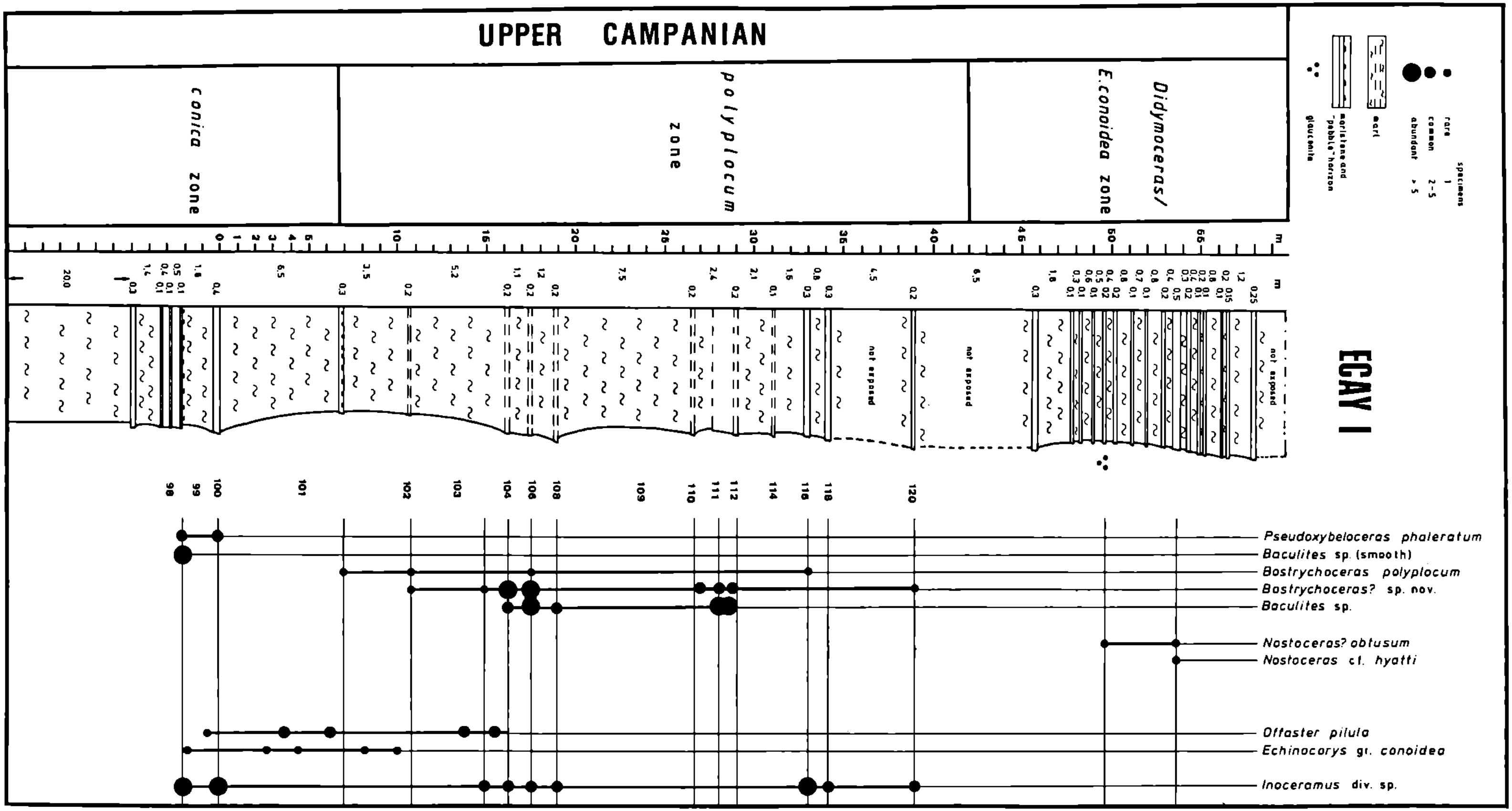
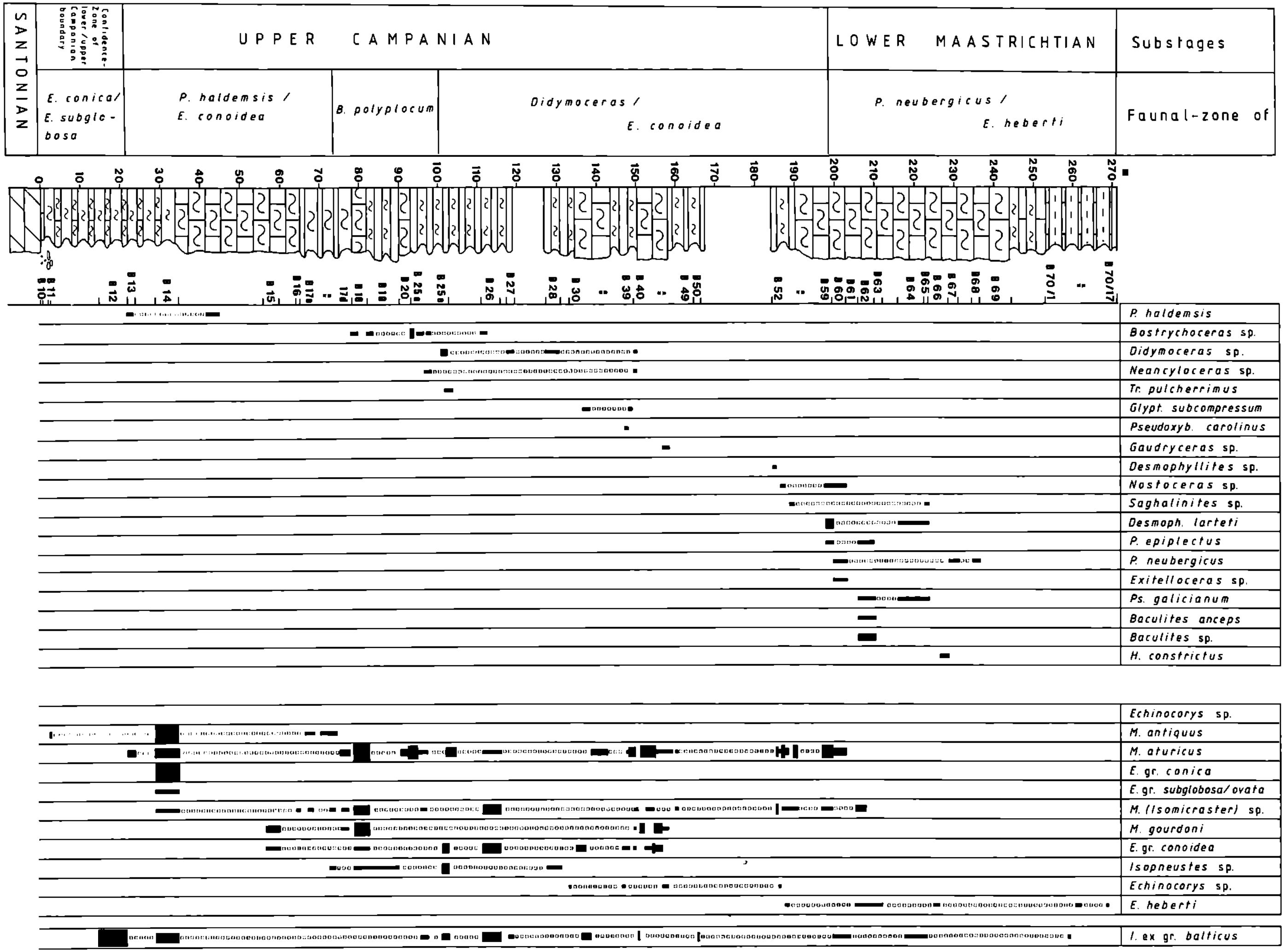
Text-Fig. 6. Ecay I (E-Barranca), Profil im Bereich der höchsten Conica-Zone, Polyplocum-Zone und der unteren Didymoceras/E. conoidea-Zone.
Text-Fig. 6. Ecay I (E Barranca), section in uppermost Conica Zone, Polyplocum Zone and lower Didymoceras/E. conoidea Zone.



UPPER CAMPANIAN
conica zone
polyplocum zone
E. conoidea zone
Didymoceras/
ECAY I
not exposed
not exposed
not exposed
rare
common
abundant
specimens
1
2-5
> 5
marl
marlstone and "pebble" horizon
glauconite
Pseudoxybeloceras phaleratum
Baculites sp. (smooth)
Bostrychoceras polyplocum
Bostrychoceras? sp. nov.
Baculites sp.
Nostoceras? obtusum
Nostoceras cf. hyatti
Offaster pilula
Echinocorys gr. conoidea
Inoceramus div. sp.

Juandechaco I



dagegen durch reichere Fossilfunde, zumindest in der Polyplocum-Zone und der basalen Didymoceras/E. conoidea-Zone, aus.

Markanter als das Aussetzen von *B. polyplocum* und *Bostrychoceras?* sp. nov., welche noch in einem Overlap-Bereich mit den ersten Didymoceraten auftreten, ist ein Faunenwechsel. Er äußert sich in dem Aufkommen von *Trachyscaphites pulcherrimus* (ROEMER), *Pseudoxybeloceras (P.)* cf. *interruptum* (SCHLÜTER), *Pseudoxybeloceras (P.) carolinus* (D'ORBIGNY), *Solenoceras* sp., *Glyptoxoceras* sp., *Glyptoxoceras subcompressum* (FORBES) und Didymoceraten.

Die Basis der Zone ist mit dem charakteristischen Faunenwechsel definiert. Dies korrespondiert in der E-Barranca mit dem ersten Massen-Auftreten von Didymoceraten (Text-Fig. 8) und in Urdiroz/Imiscoz mit deren Einsetzen.

Die Einzelfunde aus Urdiroz/Imiscoz und die in der E-Barranca auftretenden *Didymoceras*-Populationen zeigen im Berippungs- und Beknotungstyp, neben großer intraspezifischer Variation, eine große Variationsbreite der Merkmale zwischen Extremformen, die zu *Didymoceras secoense* (YOUNG) (YOUNG 1963, pl. 4, fig. 8), *Didymoceras archiacianum* (D'ORBIGNY) (siehe KENNEDY 1986, figs. C, H, I) gestellt werden können, und Formen, die bereits Merkmale von *Didymoceras schloenbachi* (FAVRE) aufweisen.

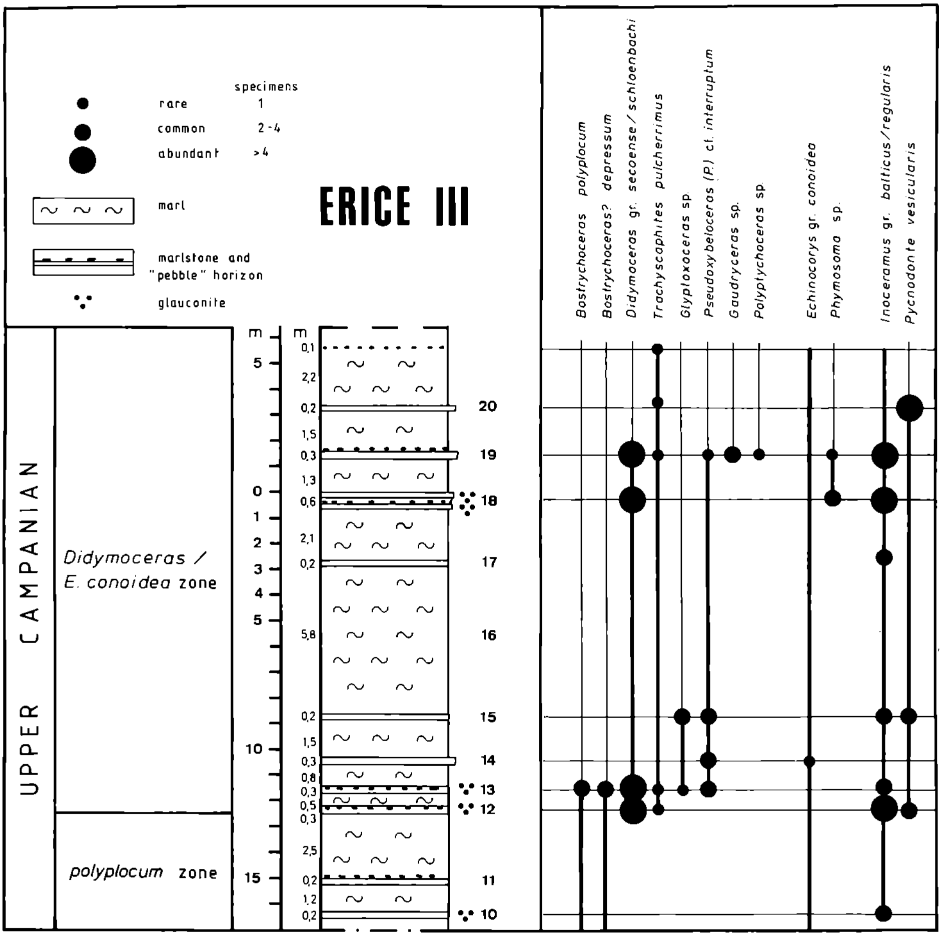
In der E-Barranca enthalten die Schichten oberhalb der *Didymoceras*-Akme vereinzelt *Nostoceras? obtusum* HOWARTH, *Nostoceras* cf. *hyatti* STEPHENSON und *N. hyatti* STEPHENSON. Während *Nostoceras? obtusum* erstmals in Europa belegt ist, können wir dem von KENNEDY (1986) aus Frankreich beschriebenen *Nostoceras hyatti* zwei weitere hinzufügen. KENNEDY gibt für das französische Exemplar, indiziert durch die Mikrofauna, ein Alter von Unter-Maastricht an, während *Nostoceras hyatti* in N-Spanien bereits im Ober-Campan vorkommt.

In Urdiroz/Imiscoz tritt die Gattung *Nostoceras*, wie in den USA (COBBAN 1974), Angola (HOWARTH 1965) und Israel (LEWY 1969, 1986), erst im Campan/Maastricht-Grenzbereich auf. Mit *Nostoceras* sp. sind in Urdiroz/Imiscoz *Sphenodiscus* sp., *Desmophyllites* sp., *Saghalinites* sp., *Desmophyllites larteti* (SEUNES) und *Pachydiscus epiplectus* (REDTENBACHER) vergesellschaftet. Die drei letztgenannten Arten überschreiten die Campan/Maastricht-Grenze.

3.4 Campan/Maastricht-Grenze

Während des "Symposium on Cretaceous Stage Boundaries" in Kopenhagen (1983) wurden für die Grenzziehung zwischen Campan und Maastricht unter anderem folgende drei Vorschläge in Betracht gezogen (BIRKELUND et al. 1984).

Text-Fig. 7. Juandechaco I (Urdiroz/Imiscoz). Profil 1 km östlich von Imiscoz mit einer kontinuierlichen Abfolge vom Ober-Campan bis Unter-Maastricht.
Text-Fig. 7. Juandechaco I (Urdiroz/Imiscoz). Section 1 km east of Imiscoz showing a continuous sequence from Upper Campanian to Lower Maastrichtian.



Text-Fig. 8. Erice III (E-Barranca). Profil der höchsten Polyplocum-Zone und der untersten Didymoceras/E. conoidea-Zone. Die Basis der Didymoceras/E. conoidea-Zone wird mit einem Faunenwechsel gezogen, der mit dem ersten massenhaften Auftreten von *Didymoceras* gr. *secoense/schloenbachi* korrespondiert.

Text-Fig. 8. Erice III (E Barranca). Section in uppermost Polyplocum Zone and lowermost Didymoceras/E. conoidea Zone. The base of the Didymoceras/E. conoidea Zone is taken by faunal change corresponding to the first mass occurrence of *Didymoceras* gr. *secoense/schloenbachi*.

Das Erstauftreten von:

1. *Belemnella lanceolata* (SCHLOTHEIM), welches gleichzeitig das Erstauftreten der Gattung *Belemnella* ist;
2. *Hoploscaphites constrictus* (SOWERBY);
3. *Pachydiscus neubergicus* (VON HAUER).

Zu 1: Weitgehend wurde in Kopenhagen akzeptiert, die Basis des Maastricht in die Nähe des Einsetzens von *Belemnella lanceolata* zu legen. Für die Grenzziehung in N-Spanien kommen Belemniten nicht in Betracht, da aus der oberen Oberkreide nur zwei Exemplare vorliegen (mdl. Mitt. SCHMID und MARTINEZ 1987).

Zu 2: *H. constrictus* galt lange Zeit als der typische Maastricht-Ammonit und wurde zur Grenzziehung herangezogen. Nachteilig sind sein seltenes Auftreten im tieferen Unter-Maastricht (BIRKELUND 1982, SCHULZ et al. 1984) und weitgehende geographische Begrenzung auf boreale Kreidevorkommen. Die südlichsten Funde stammen aus N-Spanien, aus den Provinzen Llerida und Navarra (MARTINEZ 1982, KUTZ 1987).

Zu 3: *Pachydiscus neubergicus* hat den Vorteil einer weltweiten Verbreitung. KENNEDY (1984) zählt die Verbreitungsgebiete von *P. neubergicus* zusammen mit den regionalen Faunen auf und gibt die daraus resultierenden Korrelationsmöglichkeiten an.

Die Campan/Maastricht-Grenze ist in kontinuierlicher und fossilreicher Abfolge nur in Urdiroz/Imiscoz erschlossen. Die Stufengrenze ist mit dem ersten sicher bestimmbaren *P. neubergicus* definiert. *H. constrictus* setzt erst 13 m über *P. neubergicus* ein.

Eine mikrostratigraphische Bearbeitung dieses Grenzbereiches steht noch aus, so daß auf die weiteren Kopenhagener Vorschläge z. Zt. nicht eingegangen werden kann.

3.5 Unter-Maastricht

Das Unter-Maastricht umfaßt als "Standard"-Ammoniten-Zone die Zone des P. neubergicus. In den echinidenreichen Profilen Navarras zeichnet sich diese Zone zusätzlich durch den großwüchsigen *Echinocorys heberti* SEUNES aus, der bereits im höchsten Ober-Campan einsetzt.

Die Ammoniten-Assoziation des Urdiroz/Imiscoz-Gebietes besteht neben den bereits im Ober-Campan auftretenden Arten aus: *Pseudokossmaticeras galicianum* (FAVRE), *Anagaudryceras lüneburgense* (SCHLÜTER), *Exitelloceras* sp. und *Baculites anceps* (LAMARCK).

In N-Europa wird die Reichweite von *P. neubergicus* von Unter-Maastricht bis unteres Ober-Maastricht angegeben, die von *H. constrictus* erstreckt sich über das gesamte Maastricht (BIRKELUND 1979, SCHULZ et al. 1984, KENNEDY & SUMMESBERGER 1986).

Erst das Auftreten der auf das Ober-Campan und Unter-Maastricht beschränkten Gattung *Pseudokossmaticeras,* zusammen mit *P. neubergicus* und *H. constrictus,* belegt sicheres Unter-Maastricht (siehe NAJDIN 1969, THIEDIG & WIEDMANN 1976, BLASZKIEWICZ 1980, KENNEDY et al. 1986, KENNEDY & SUMMESBERGER 1986). Mit *Ps. galicianum* liegt in Urdiroz/Imiscoz eine Art vor, die in Polen auf die N. pozaryskii- und die B. lanceolata-Zonen beschränkt ist (BLASZKIEWICZ 1980).

In der E-Barranca wird das Unter-Maastricht durch *Echinocorys heberti* und durch einen Einzelfund von *P.* sp. gr. *neubergicus* angezeigt. Die Foraminiferen-Fauna aus diesem 10 m mächtigen Bereich besteht aus *Globotruncanita stuartiformis* (DALBIEZ), *Globotruncana* gr. *lapparenti* BROTZEN, *Gltr.* cf. *ventricosa* WHITE, *Gltr. rosetta* (CARSEY) und *Stensioeina pommerana* und läßt keine eindeutige Datierung zu.

6. Zusammenfassung

Für die nordspanischen Gebiete E-Barranca und Urdiroz/Imiscoz (Navarra) werden lithologische und biostratigraphische Gliederungen für das Campan und Unter-Maastricht, basierend auf Ammoniten und irregulären Echiniden (*Echinocorys, Offaster*), sowie deren Reichweiten in Normal- und ausgewählten Einzelprofilen dargestellt. Die für das regionale Zonenschema typischen Faunen-Assoziationen werden beschrieben, die lokalen Gliederungen anhand der Faunen-Assoziationen und Bio-Events korreliert. Die Einhängung der regionalen Gliederung in die nachvollziehbare "Standard"-Ammoniten-Zonierung für das Campan und Unter-Maastricht wird diskutiert.

Das tiefste Campan enthält eine Echiniden-Assoziation von *Offaster pomeli, M. (Isomicraster)* sp. und *Echinocorys scutata cincta*. Vergleichbare Faunen-Zusammensetzungen sind in S-England und NW-Deutschland aus der Bidorsatum-Zone bekannt. In N-Spanien kennzeichnet *Sc. hippocrepis* nicht die Basis des Campan, sondern setzt erst im höheren Unter-Campan ein.

Die Unter/Ober-Campan-Grenze wird mit dem Einsetzen von *H. marroti* definiert. Ein Overlap-Bereich von *H. marroti* mit Spätformen von *Sc. hippocrepis* ist nachweisbar. *Trachyscaphites spiniger* setzt, vergesellschaftet mit *Pachydiscus haldemsis*, oberhalb von *H. marroti* ein.

Die traditionelle Polyplocum-Zone wird auf der Basis der chronologischen Aufeinanderfolge der Gattungen *Bostrychoceras* und *Didymoceras* zweigeteilt. Dies wird durch einen Wechsel der Ammoniten-Assoziation gestützt.

Die Campan/Maastricht-Grenze wird mit dem Einsetzen von *Pachydiscus neubergicus* gezogen. *Hoploscaphites constrictus* setzt deutlich oberhalb *P. neubergicus* ein.

Danksagung. Für die freundliche Unterstützung bei der Bestimmung zahlreicher Ammoniten danken wir Prof. Dr. J. WIEDMANN und Dr. W. J. KENNEDY. Bei der Bestimmung der Echiniden und Inoceramen waren uns C. J. WOOD und Prof. Dr. G. ERNST hilfreich.

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Rudist Horizons in the Montsec (South Central Pyrenees)*

ORIOL PASCUAL, JOSE MARIA PONS and ENRIC VICENS, Barcelona

With 3 Text-Figures

PASCUAL, O., PONS, J. M. & VICENS, E. (1989): Rudist Horizons in the Montsec (South Central Pyrenees). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 215-230. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The exceptional exposure of sediments in the Serra del Montsec provides one of the most complete sections of shallow facies deposited during Upper Cretaceous time in the Pyrenees.

Rudist horizons are abundant and diverse. Hippuritidae and Radiolitidae are the main components of rudist biota, while Plagioptychidae, Caprotinidae and Requieniidae are poorly represented.

Nine different assemblages of rudist species have been identified in the successive rudist horizons deposited during Upper Turonian to Lower Maastrichtian time. A biostratigraphical scale has been obtained that can be applied in other Pyrenean areas where the stratigraphical record is not as complete and well understood as it is in the Montsec area.

Consecutive steps have been recognized in a few single evolutionary lineages. The fact that these consecutive steps appear in a single section elucidates the relative stratigraphical position of some rudist species.

Kurzfassung: Die ausgezeichneten Aufschlußverhältnisse in der Serra del Montsec liefern eines der vollständigsten Profile der Flachwasserfazies, die im Laufe der Oberkreide in den Pyrenäen abgelagert wurde.

Rudistenhorizonte sind zahlreich und divers. Hippuritidae und Radiolitidae stellen die Hauptbestandteile der Rudistenfauna dar, während Plagioptychidae, Caprotinidae und Requieniidae nur schwach vertreten sind.

Neun verschiedene Rudistenvergesellschaftungen sind in den aufeinanderfolgenden Rudistenhorizonten identifiziert worden, die zwischen Oberturon und Untermaastricht abgelagert wurden.

Eine biostratigraphische Skala wurde erstellt, die in anderen Gebieten der Pyrenäen angewandt werden kann, wo die Stratigraphie weder so vollständig noch so deutlich ist wie im Montsec-Gebiet.

In einigen wenigen Entwicklungsreihen sind fortlaufende Entwicklungsschritte erkannt worden. Die Tatsache, daß diese Entwicklungsschritte in

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einem Einzelprofil gefunden worden sind, erklärt die relative stratigraphische Lage einiger Rudistenarten.

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1. Introduction

Some attempts have been made to integrate rudist biostratigraphical data concerning Pyrenees or other related western Tethyan areas by means of zonations or species distributions (PONS 1977, 1982, BILOTTE 1983, 1985, PHILIP & BILOTTE 1983, amongst others). Their results have been used with variable success.

It is well known that the occurrence of rudist horizons is related to facies control. Yet the precise species composition in each horizon is also closely linked to the environmental parameters. See a discussion of the problems on the use of rudists in biostratigraphy in GILI et al. (1986).

We have therefore considered it as very useful for our biostratigraphical work to have a true vertical succession of rudist species being as complete as possible. Knowing in detail in which facies they occur we could follow the depositional history within a single profile from base to top.

In order to study such a profile and its fossil content we have chosen the Serra del Montsec. For several reasons the central part, named Montsec d'Ager, was selected from among the three parts in which the ridge is divided by the rivers Noguera Pallaresa and Noguera Ribagorzana.

Upper Cretaceous sediments are very well exposed and very rich in fossil content. It is the type locality of many rudist species described by VIDAL (1878) and H. DOUVILLE (1895). There is a mountain road ascending from the village of Ager to the top of the Montsec and leading down to the Tremp Basin, crossing fresh outcrops along the whole section.

The area has been mapped at a scale of 1: 18.000, mainly to solve some small tectonic problems that arise both at the base of the profile and at the base of the main cliff. The geological map with the localization of the studied section appears in Text-Fig. 1.

2. Clot d'Olsi - Alsamora section

The section has been measured according to the Jacob's staff method and drawn originally at a scale of 1: 5.000 allowing us to differenciate levels of half a meter in thickness.

As can be seen in Text-Fig. 2, thickness, lithology, depositional texture, sedimentary structures and fossil content are represented and referred to

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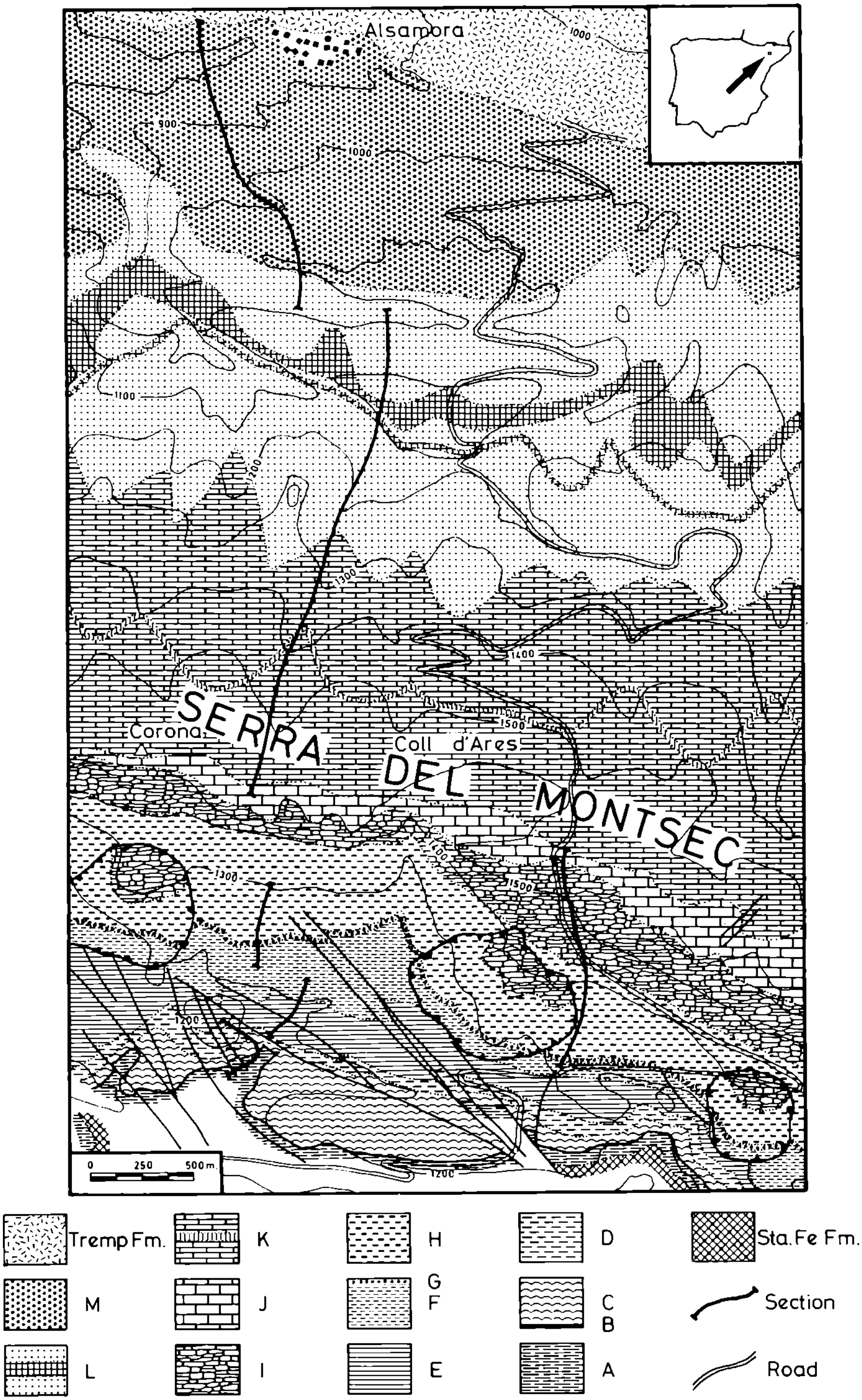
Text-Fig. 1. Geological map of the central part of Montsec d'Ager. Mapped units correspond to levels of the simplified section. M = 19; L = 18; K = 15, 16, 17; J = 13, 14; I = 9, 10, 11, 12; H = 7, 8; F, G = 6; E = 5; D = 4; C = 3; B = 2; A = 1.


Alsamora
1000
900
1000
1100
1200
1300
1400
1500
SERRA DEL MONTSEC
Corona
Coll d'Ares
1300
1600
1500
1200
1200
0 250 500 m
Tremp Fm.
K
H
D
Sta. Fe Fm.
M
J
G
F
C
B
Section
L
I
E
A
Road

each level. Some levels have been grouped in order to obtain a simplified profile in which the distribution of the rudist species is represented (Text-Fig. 3).

It has been possible to relate successive levels of the section to depositional sequences (third-order cycles of VAIL et al. 1977) recognized in south central Pyrenees by SIMO (1986), although not all the discontinuities used to limit them are clearly visible in the studied outcrops.

The base of the section is marked by the erosional surface that separates the Santa Fe sequence from the Congost sequence. The top is where continental red beds of the Tremp Formation appear.

3. Rudist horizons and assemblages

Rudist horizons in the section are very diverse in development, facies and taxonomic diversity.

As it is shown in Text-Fig. 3, nine rudist assemblages have been recognized; one of Turonian (Tu), one of Coniacian (Co), two of Santonian (Sa 1 and Sa 2), three of Campanian (Ca 1, Ca 2 and Ca 3) and two of Maastrichtian age (Ma 1 and Ma 2). Some of these assemblages correspond to a single horizon, but most of them to successive ones.

3.1 Assemblage Tu

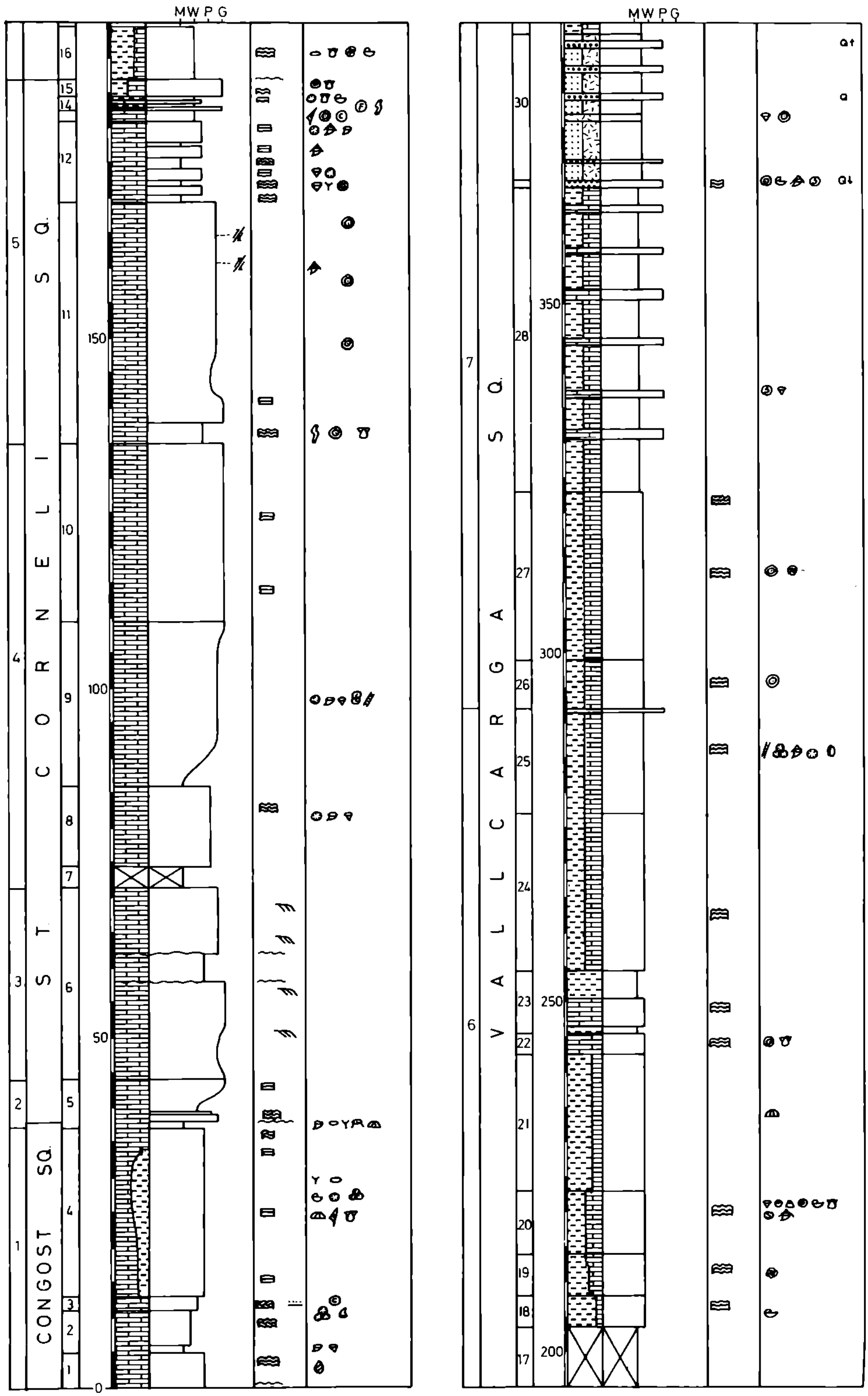
The oldest rudist horizon is located at the base of level 2 (see Text-Fig. 2) in a bed of nodulous aspect probably due to compaction; fossils occur in a mudstone-wackestone matrix. *Praeradiolites pailleteanus* (D'ORB.) and *Sphaerulites patera* ARNAUD are the most common species. *Vaccinites petrocoriensis* (DOUV.) and *V. rousseli* (DOUV.) are scarce and *Hippurites requieni* MATH. develops small thickets.

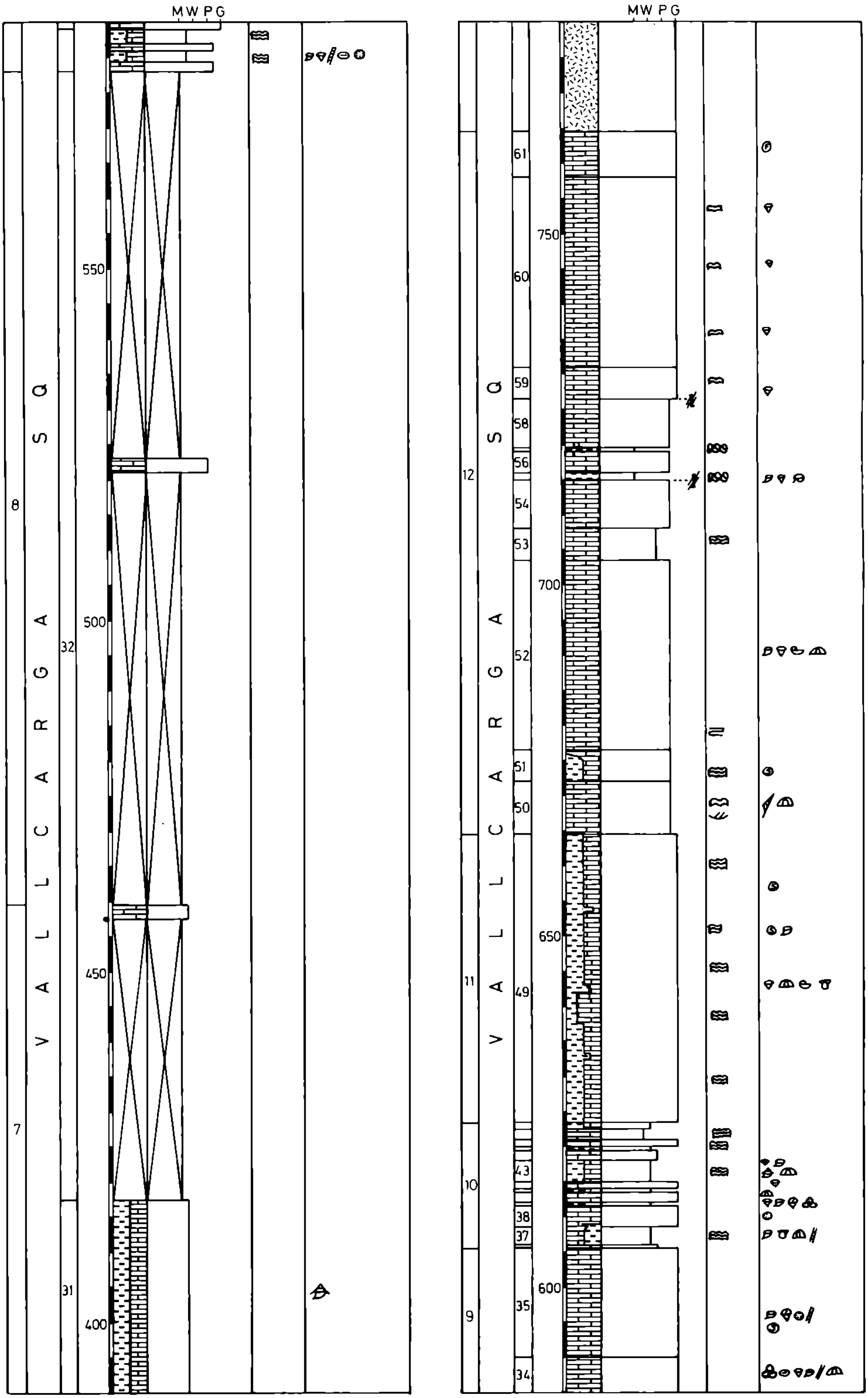
The age of this assemblage is clearly Upper Turonian.

3.2 Assemblage Co

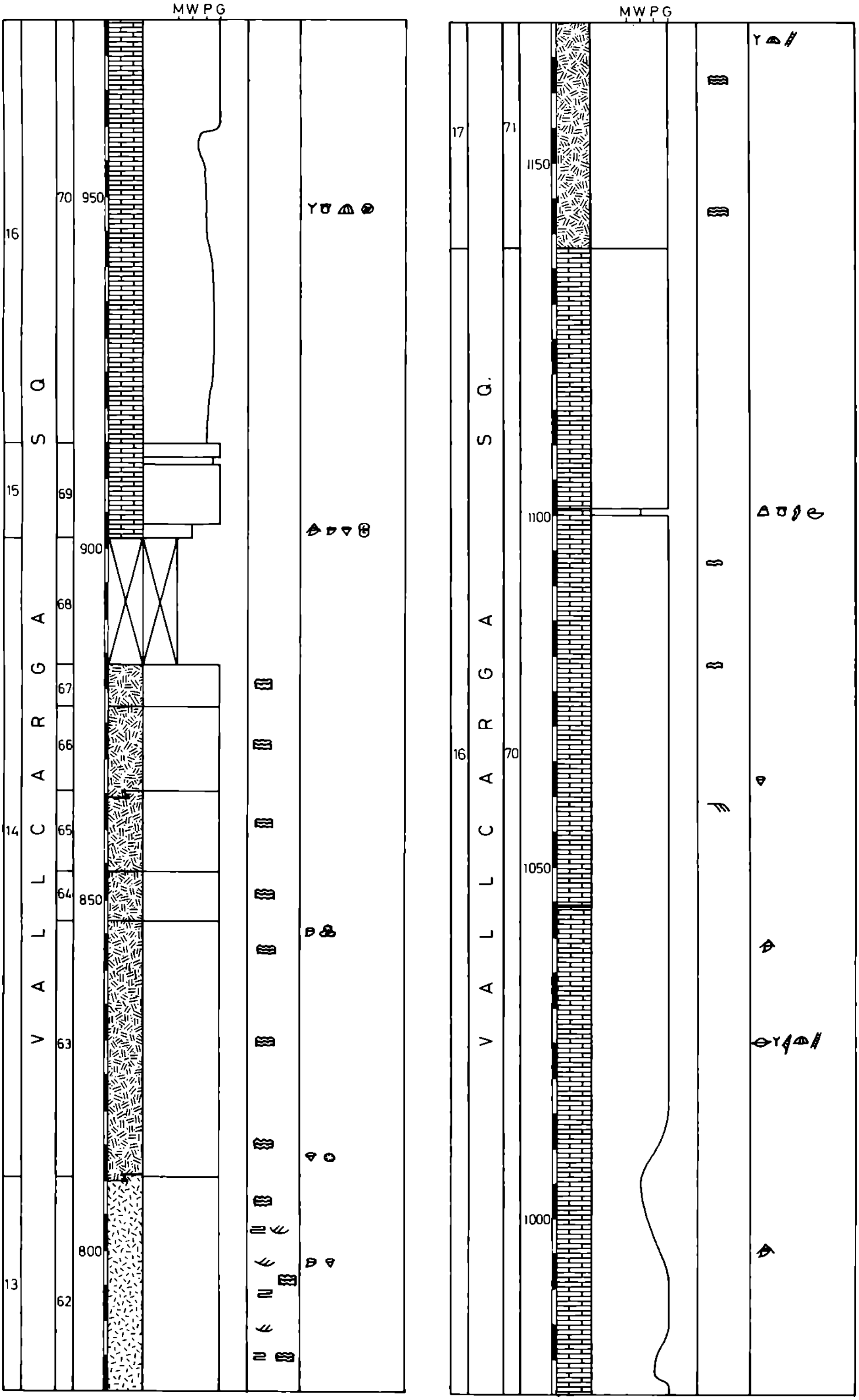
There is a very well developed thicket at the base of level 5, which we consider as the base of St. Corneli sequence (see Text-Fig. 2). Very long and slender hippuritids with very poor mudstone matrix are the main components: *Hippuritella praetoucasi* (TOUCAS), *Hippurites resectus* DEFRANCE and *H. socialis* DOUV. Yet, some radiolitids are also present; these are *Biradiolites praefissicostatus* TOUCAS, *Praeradiolites requieni* (D'HOM.-FIRM.), *Radiolites radiosus* D'ORB., *R. praegalloprovincialis* TOUCAS and *R. vallispetrosae* ASTRE. Four of these species are still found in levels 8 or 9 (see Text-Fig. 3) and the first appearance of *Vaccinites moulinsi* (D'HOM.-FIRM.) and *Radiolites douvillei* TOUCAS is in level 8.

Text-Fig. 2. Clot d'Olsi - Alsamora section. Stratigraphical profile in Turonian to Maastrichtian sediments in Montsec d'Ager, south central Pyrenees.


MWPG
CONGOST SQ.
ST CORNELI SQ.
VALLCARGA SQ.
0
50
100
150
200
250
300
350
MWPG



Text-Fig. 2, continued



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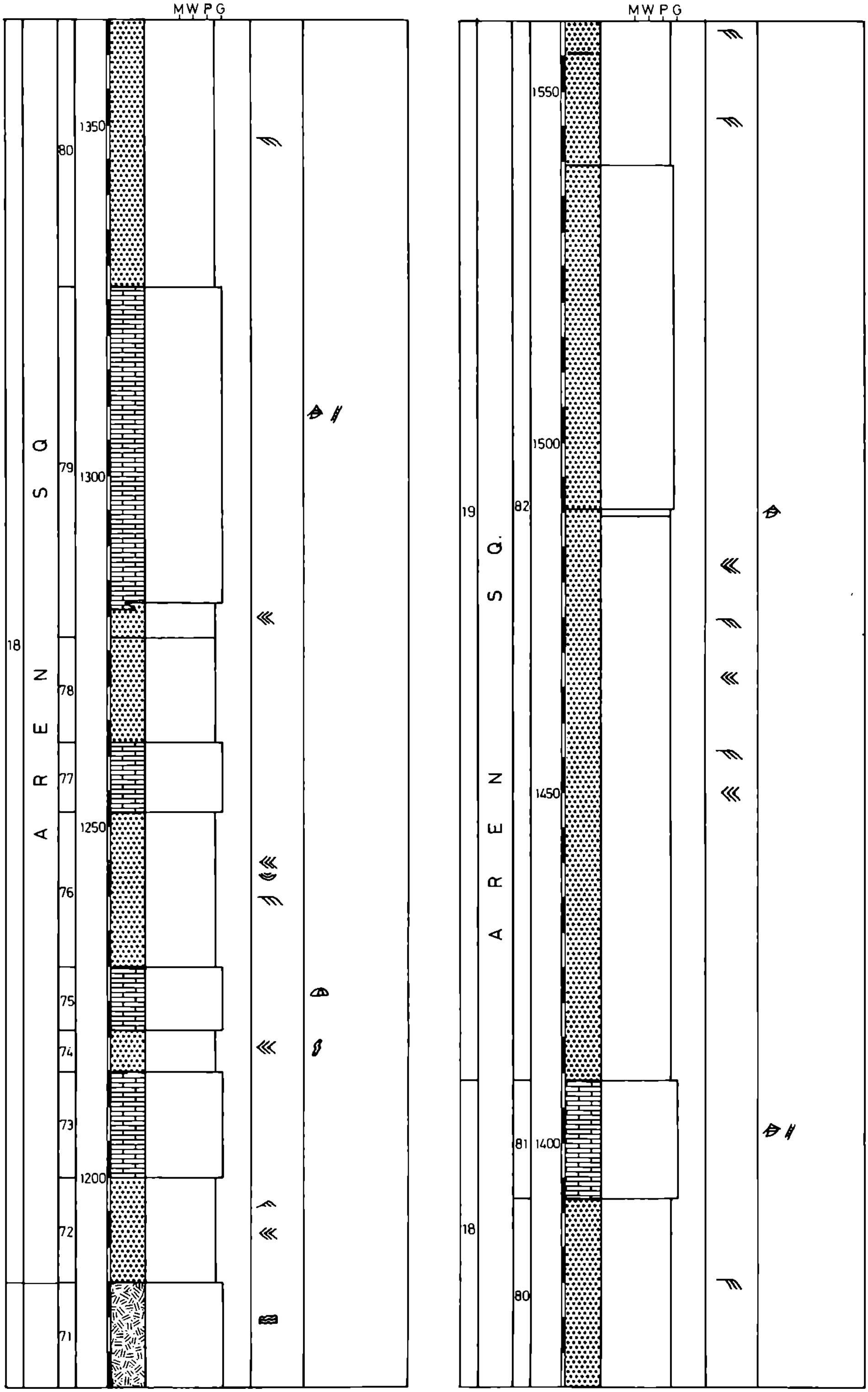


Text-Fig. 2, continued

The age of this assemblage is considered to be Coniacian.

3.3 Assemblage Sa 1

We include in this assemblage (1) those incipient bioconstructions with hippuritids and radiolitids developed in packstones of nodular aspect in level 9 (some of their species were also poorly represented in level 8), (2) isolated and fragmented specimens in the packstone-grainstone facies of level 11, and (3) those successive thickets, mainly consisting of radiolitids in level 12 (see Text-Fig. 2).

Species present (see Text-Fig. 3) are *Hippuritella sarthacensis* (COQ.), *Hippurites matheroni* DOUV., *H. praecessor* DOUV., *H. sublaevis* MATH., *Vaccinites beaussetensis* TOUCAS, *Biradiolites canaliculatus* D'ORB., *B. angulosissimus* TOUCAS, *B.* gr. *fissicostatus, B.* sp. 1, *Fossulites undaesaltus* ASTRE, *Radiolites sauvagesi* (D'HOM.-FIRM.), *R. subradiosus* TOUCAS, *R. squamosus* D'ORB., *R. gastaldianus* PIRONA, *R.* sp. 1, *"R." laciniatus* VIDAL and *Plagioptychus* sp. 1.

The age is Santonian.

3.4 Assemblage Sa 2

There are a few rudist horizons in the marly facies of the lower part of the Vallcarga sequence (see Text-Fig. 2). The first one of which developed in level 20. Level 25 is a complex bioconstruction of rudists, corals and algae with an abundant accompanying fauna. As can be seen in Text-Fig. 3, this level alone is responsible for the bulk of the enormous species diversity recorded in this assemblage. In levels 28 to 32 some isolated rudists still occur.

Species recorded are *Hippuritella maestrei* (VIDAL), *H. carezi* (DOUV.), *Hippurites microstylus* DOUV., *H. cristatus* DOUV., *H. canaliculatus* ROLL. DU ROQ., *H. turgidus* ROLL. DU ROQ., *Vaccinites beaussetensis* TOUCAS, *V. galloprovincialis* (MATH.), *Biradiolites carezi* TOUCAS, *B. praeingens* TOUCAS, *B. angulosissimus* TOUCAS, *B. ibericus* (VIDAL), *B. acuticostatus* (D'ORB.), *B. fissicostatus* D'ORB., *Praeradiolites sinuatus* (D'ORB.), *P. plicatus* LAJ. NEG. TOUL., *P. sarladensis* TOUCAS, *P. toucasianus* (D'ORB.), *P. coquandi* (BAYLE), *Radiolites galloprovincialis* MATH., *Sphaerulites boreaui* TOUCAS, *Plagioptychus toucasi* MATH., *P.* sp. 2 and *Monopleura minuta* VIDAL.

The age of this assemblage is Santonian.

3.5 Assemblage Ca 1

Some rudist horizons are found in the limestones forming the base of the main cliff of the Montsec and thus correspond to section levels 33 to 49 (Text-Fig. 2). These are developed in alternating shale and nodular beds, the latter displaying wackestone-mudstone texture. Isolated specimens also appear in packstone-grainstone beds.

The faunal assemblage is composed of *Hippuritella variabilis* (MUN.-CHAL.), *Hippurites crassicostatus* DOUV., *Biradiolites* aff. *ibericus, Praeradiolites aristidis* MUN.-CHAL., *P. sinuatus-hoeninghausi* and *Radiolites angeoides* P. DE LAP.

The age is Campanian.

3.6 Assemblage Ca 2

The upper part of the main cliff of the Montsec is formed by limestones (packestone-grainstone) and calcarenites which coincide with section levels 50 to 67 (Text-Fig. 2). Isolated rudists or small clusters of them occur in limestones and there mainly in beds with a nodulous aspect. In turn, they also occur in calcarenites and it is important to notice that greatest species diversity is present in coarse calcarenites (see Text-Fig. 3).

Rudist species of this assemblage are *Hippuritella variabilis* (MUN.-CHAL.), *Hippurites serratus* DOUV., *H. heberti* MUN.-CHAL., *H. lamarcki quintanalomensis* CIRY, *Vaccinites archiaci* (MUN.-CHAL.), *Biradiolites orbignyi* TOUCAS and *Praeradiolites sinuatus-hoeninghausi.*

The age is Campanian.

3.7 Assemblage Ca 3

A complex bioconstruction is developed in wackestone beds at the base of level 69 (see Text-Fig. 2). It is displayed rather evident where the mountain road reaches the top of the Montsec. Some of the species there present also occur in isolated specimens within several grainstone beds at level 70.

The species recognized are *Hippuritella variabilis* (MUN.-CHAL.), *Hippurites vidali* MATH., *Vaccinites archiaci* (MUN.-CHAL.), *Biradiolites orbignyi* TOUCAS, *B. leychertensis* TOUCAS, *Lapeirousia* sp. 1, *Praeradiolites subtoucasi* TOUCAS and *Apricardia* sp. 1.

The age is Campanian.

3.8 Assemblage Ma 1

Although sediments of the Areny sequence are mainly formed by sandstone in the Montsec, there are still some limestone levels where rudist horizons are poorly developed (see Text-Fig. 2).

The fauna is composed of *Hippuritella lapeirousei* (GOLDF.), *Lapeirousia* sp. 2, *L.* sp. 3, *Praeradiolites boucheroni* (BAYLE), *Radiolitella pulchellus* (VIDAL) and *Apricardia sicoris* ASTRE.

The age of this assemblage is Lower Maastrichtian.

3.9 Assemblage Ma 2

In the studied section, the sandstones of the Areny sequence are topped by a charophytes bearing bed and then followed by red continental sediments without rudists. Yet, only 8 km to the east, still in Montsec d'Ager, some rudist horizons are developed between the sandstones and the red continental sediments.

The fauna is very characteristic and made up of *Hippuritella castroi* (VIDAL), *Biradiolites chaperi* (BAYLE), *B. ara* PONS, *Praeradiolites boucheroni* (BAYLE), *"Radiolites" moroi* VIDAL, *"R." sellesi* BAUDELOT and *Monopleura moroi* (VIDAL).

The age is Lower Maastrichtian.

4. Paleontological considerations

In Text-Fig. 3, those rudist species recognized in the section are listed due to their first appearance, in order to show clearly their grouping in successive assemblages. Some considerations can be obtained, however, if we consider either all species of each genus or one of those groups of species recognized by TOUCAS (1903-4 and 1907-9), that are supposed to represent single phylogenetic lineages.

4.1 Genus *Hippuritella* DOUVILLE

There is a good register of the *Hippuritella variabilis* phylogenetic lineage. *H. sarthacensis* (COQ.) occurs in Co and Sa 1 assemblages and *H. maestrei* (VIDAL) in Sa 2. *H. variabilis* (MUN.-CHAL.) itself is recorded in Ca 1, Ca 2 and Ca 3, *H. lapeirousei* (GOLDF.) in Ma 1 and *H. castroi* (VIDAL) in Ma 2.

The only representatives of the *Hippuritella toucasi* group are *H. praetoucasi* (TOUCAS) in Co and *H. carezi* (DOUV.) in Sa 2. *H. toucasi* (D'ORB.) has not been found among species of the Sa 1 assemblage as would be expected.

4.2 Genus *Hippurites* LAMARCK

Interpretation of the *Hippurites canaliculatus* group has always been controversial. In the study area an almost complete record is existing: *H. requieni* MATH. in Tu, *H. resectus* DEFRANCE in Co, *H. matheroni* DOUV. in Sa 1, *H. cristatus* DOUV. and *H. canaliculatus* ROLL. DU ROQ. in Sa 2, *H. crassicostatus* DOUV. in Ca 1, *H. serratus* DOUV., *H. heberti* MUN.-CHAL. and *H. lamarcki quintanalomensis* CIRY in Ca 2 and *H. vidali* MATH. in Ca 3. It is important to notice the relative position of *H. serratus* DOUV. since the Montsec is mainly considered to be its type locality. The last species of the group, *H. radiosus* DESMOUL., does not appear in the studied section, but it has been recorded eastward (LIEBAU 1984), also in Montsec d'Ager in a level corresponding with our assemblage Ma 1.

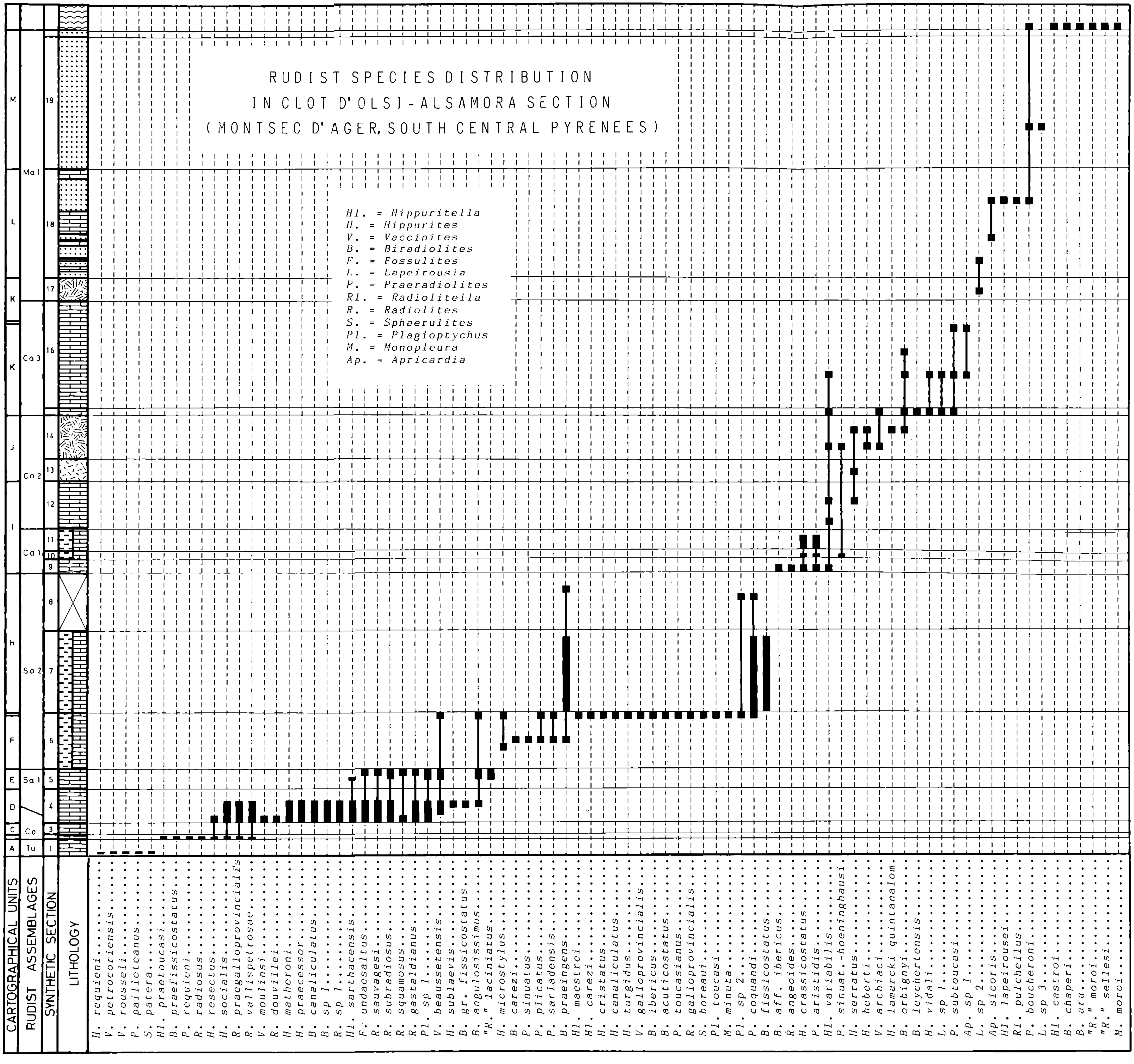
The *Hippurites socialis* group is represented by two species, *H. socialis* DOUV. in Co and Sa 1 assemblages and *H. microstylus* DOUV. in Sa 2.

The group of *Hippurites bioculatus* is only represented by *H. praecessor* DOUV. in Sa 1.

The *Hippurites turgidus* group is represented by two species, *H. sublaevis* MATH. in Sa 1 and *H. turgidus* ROLL. DU ROQ. in Sa 2.

4.3 Genus *Vaccinites* FISCHER

Unlike north of Conca de Tremp, the genus *Vaccinites* FISCHER is very poorly represented in Montsec. *V. petrocoriensis* (DOUV.) in the Tu assemblage is the only species of the *Vaccinites dentatus* group. *V. galloprovincialis* (MATH.) in Sa 2 is the only one of its group and *V. archiaci*



Text-Fig. 3. Rudist species distribution referred to a simplification of the Clot d'Olsi - Alsamora section of Text-Fig. 2.

(MUN.-CHAL.) in Ca 2 and Ca 3, is the only one of the *Vaccinites sulcatus* group.

The *Vaccinites moulinsi* group is represented by *V. rousseli* (DOUV.) in the Tu assemblage, *V. moulinsi* (D'HOM.-FIRM.) in Co, and *V. beaussetensis* TOUCAS in Sa 1 and Sa 2.

4.4 Genus *Biradiolites* D'ORBIGNY

Biradiolites carezi TOUCAS from Sa 2 assemblage is the only recorded species of the *Biradiolites lombricalis* group.

The group of *Biradiolites angulosus* is better represented; *B. angulosissimus* TOUCAS is present in Sa 1 and Sa 2 assemblages, *B. ibericus* (VIDAL) in Sa 2 and *B. leychertensis* TOUCAS in Ca 3. In Ca 1 appears a form of this group, quite similar to *B. ibericus* (VIDAL) but very different from the one that CIRY (1940) named *B. ibericus* mut. *campaniensis;* for that reason we have designated it as *B.* aff. *ibericus.*

The *Biradiolites acuticostatus* group is represented by *B. acuticostatus* (D'ORB.) in Sa 2 and *B. orbignyi* TOUCAS in Ca 2 and Ca 3 assemblages. A form we have designated as *B.* sp. 1 and that attains considerable size seems to have characteristics of this group and occurs in Sa 1.

From the *Biradiolites canaliculatus* group are recorded: *B. canaliculatus* D'ORB. in Sa 1 and *B. chaperi* (BAYLE) in Ma 2.

Biradiolites praeingens TOUCAS in the Sa 2 assemblage is the only registered species from the *Biradiolites ingens* group.

The *Biradiolites fissicostatus* group is represented by *B. praefissicostatus* TOUCAS in Co and *B. fissicostatus* D'ORB. in Sa 2. A form we have designated *B.* gr. *fissicostatus* and that shows intermediate characteristics between both mentioned species appears in Sa 1.

4.5 Genus *Fossulites* ASTRE

This genus is represented by its type species *Fossulites undaesaltus* ASTRE in the Sa 1 assemblage.

4.6 Genus *Lapeirousia* BAYLE

Three different species are present in the section, which we have named *Lapeirousia* sp. 1, *L.* sp. 2 and *L.* sp. 3. The first one appears in Ca 3 and the other two at different horizons in Ma 1. It is necessary to achieve a better understanding of the species of this genus since they will be very useful for the biostratigraphy of Upper Campanian and Maastrichtian in the Pyrenees.

4.7 Genus *Praeradiolites* DOUVILLE

The group of *Praeradiolites ponsi* is represented by *P. toucasianus* (D'ORB.) in Sa 2, *P. subtoucasi* TOUCAS in Ca 3 and *P. boucheroni* (BAYLE) in Ma 2. The great morphological variability is typical for all three species.

From the *Praeradiolites hoeninghausi* group we have found *P. sinuatus* (D'ORB.) in Sa 2 and a form with intermediate characteristics between *P. sinuatus* (D'ORB.) and *P. hoeninghausi* (DESMOUL.) in Ca 1 and Ca 2 assemblages.

The group of *Praeradiolites cylindraceus* is represented by *P. requieni* (D'HOM.-FIRM.), *P. plicatus* LAJ. NEG. TOUL. and *P. aristidis* MUN.-CHAL. in Co, Sa 2 and Ca 1 assemblages, respectively.

The group of *Praeradiolites pailleteanus* is represented by *P. pailleteanus* (D'ORB.) in the Tu assemblage and *P. sarladensis* TOUCAS in Sa 2.

4.8 Genus *Radiolitella* DOUVILLE

Only the species *Radiolitella pulchellus* (VIDAL) is present in Ma 1.

4.9 Genus *Radiolites* LAMARCK

The group of *Radiolites lusitanicus* is represented only by the one species *R. douvillei* TOUCAS in Co and Sa 1 assemblages, as is the group of *Radiolites sauvagesi* by *R. sauvagesi* (D'HOM.-FIRM.) in Sa 1.

The group of *Radiolites radiosus* seems to be very diversified in Lower Santonian. It is represented by *R. radiosus* D'ORB. in Co and by *R. subradiosus* TOUCAS, *R. squamosus* D'ORB. and *R. gastaldianus* PIRONÁ in Sa 1. *R. squamosus* and *R. gastaldianus* have been proposed as synonyms; the fact that both species appear and are clearly differentiated in the same level suggests that the synonymy is not so clear.

The *Radiolites angeoides* group is represented by *R. praegalloprovincialis* TOUCAS and *R. vallispetrosae* ASTRE in Co and Sa 1 assemblages, *R. galloprovincialis* MATH. in Sa 2 and *R. angeoides* P. DE LAP. in Ca 1. Differences between *R. praegalloprovincialis* and *R. vallispetrosae,* a still imperfectly known form, could just be a matter of growth form pattern.

It is important to notice that the youngest *Radiolites* that appears in the section is lowermost Campanian.

4.10 Genus "*Radiolites*"

We have used this name for species that were formerly incorrectly assigned to the genus *Radiolites* LAM. or also attributed later to *Agriopleura* KÜHN. Characteristics they share in common are very distinctive and certainly do not correspond either with *Radiolites* LAM. nor *Agriopleura* KÜHN. These species are "*R.*" *laciniatus* VIDAL, present in the Sa 1 assemblage and "*R.*" *moroi* VIDAL and "*R.*" *sellesi* BAUDELOT in Ma 2. We are actually revising these species.

4.11 Genus *Sphaerulites* LAMARCK

This genus is represented by *Sphaerulites patera* ARNAUD in Tu and by *S. boreaui* TOUCAS in Sa 2 assemblages.

4.12 Genus *Plagioptychus* MATHERON

Three species are represented, *Plagioptychus* sp. 1 in Sa 1 and *P. toucasi* MATH. and *P.* sp. 2 in Sa 2. The last species shows a very poor development of the canal layer and we suspect that it is the same species that was named *Monopleura montsecana* by VIDAL (1878). We are now revising all type specimens from the VIDAL collection, kept in Museu de Geologia de Barcelona.

4.13 Genus *Monopleura* MATHERON

Two species are recorded, *Monopleura minuta* VIDAL in Sa 2 and *M. moroi* (VIDAL) in Ma 2.

4.14 Genus *Apricardia* GUERANGER

Two species are recorded, *Apricardia* sp. 1 in the Ca 3 assemblage and *A. sicoris* ASTRE in Ma 1.

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Correlation of Larger Benthic and Planktonic Foraminifera of the Late Cretaceous in the South-Central Pyrenees

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With 4 Text-Figures

CAUS, E. & GOMEZ-GARRIDO, A. (1989): Correlation of Larger Benthic and Planktonic Foraminifera of the Late Cretaceous in the South-Central Pyrenees. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 231-238. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: New correlations between sequences of shallow platform sediments containing larger benthics, and deeper deposits rich in planktonics are established in the Tremp area (S. Pyrenees) with the help of sedimentologic criteria. This correlation permits the recognition of sedimentary cycles simultaneously reflected in both facies realms. Thus, the range in age of selected late Cretaceous larger foraminifera can now be defined in terms of the standard plankton zonation.

Kurzfassung: In der Zone von Tremp (S-Pyrenäen) werden Flachwasser-Sedimente, die reich sind an benthischen Großforaminiferen, mit tieferen Ablagerungen korreliert, die reiches Plankton enthalten. Zur Korrelation werden sedimentologische Kriterien herangezogen, nämlich die Entwicklung synchroner Sedimentzyklen, die beiden Faziesbereichen gemeinsam sind. Damit kann die stratigraphische Reichweite ausgewählter Großforaminiferen in die Plankton-Standardzonenfolge eingebunden werden.

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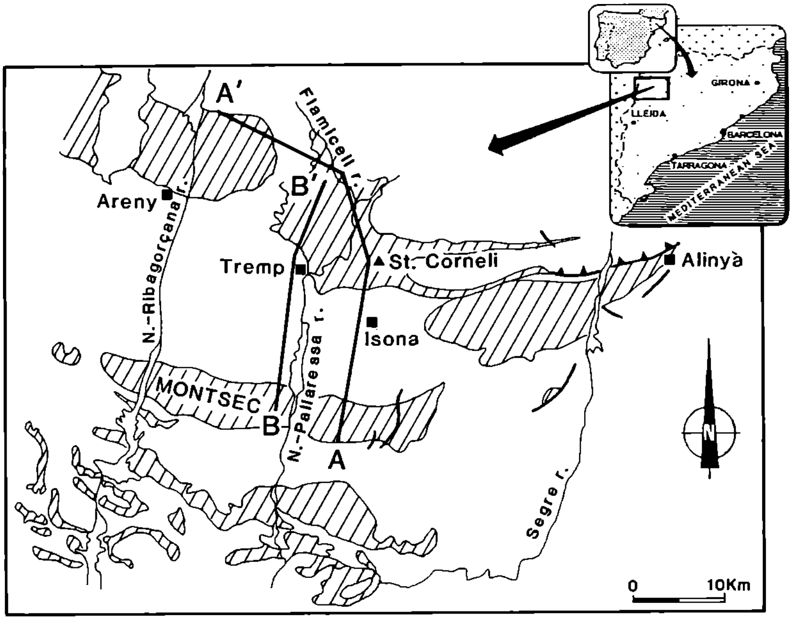
1. Introduction

Stratigraphic correlations between shallow facies containing larger foraminifera and deeper ones, rich in planktonic foraminifera and calcareous nannoplankton, are always difficult. New, detailed sedimentological work (SIMO 1986) and associated paleontological studies (larger foraminifera and planktonic foraminifera, this paper) in the Upper Cretaceous Pyrenean Basin showed the time relations between plankton zones and their biostratigraphic equivalents in shallow water deposits. Thus, the range in age of selected late Cretaceous larger foraminifera can be fixed in terms of plankton zone units, at least for the Pyrenean Basin.

Field and research work have been supported by CAICYT (PB85-0156).

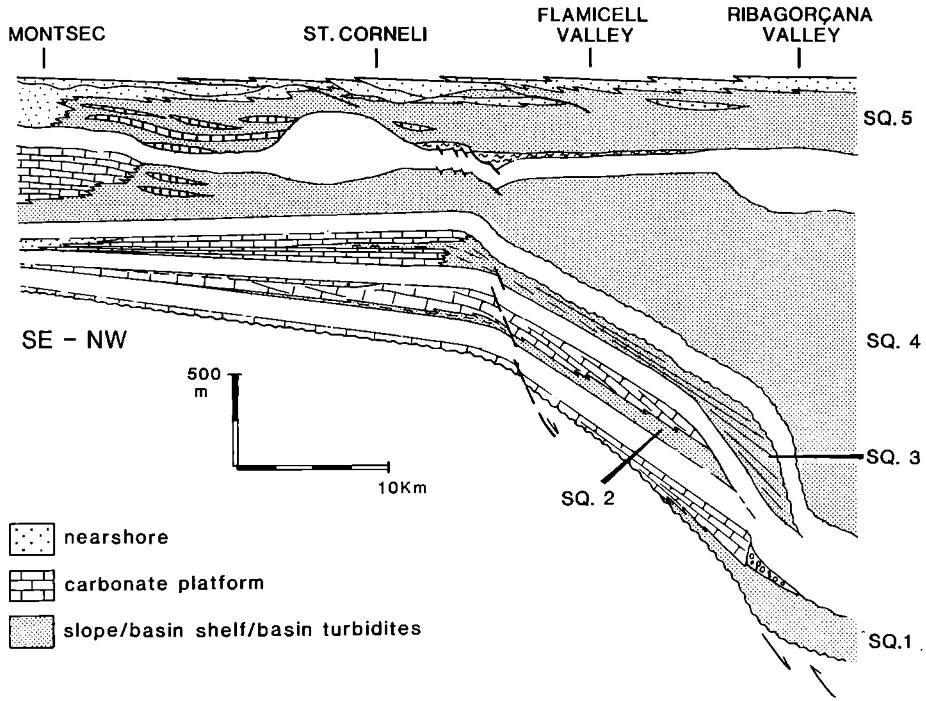


Text-Fig. 1. Outcrops of late Cretaceous sediments in the southern Pyrenees. A-A' Location of section, Text-Fig. 2. B-B' Location of section, Text-Fig. 4.

2. Distribution of outcrops of late Cretaceous sediments

In the central part of the southern Pyrenees, late Cretaceous sediments are visible in outcrops aligned in three parallel belts oriented from E to W (Text-Fig. 1). The southernmost series of outcrops occur in the Marginal Sierras. While in the northern part of this belt the sediments crop out in continuity, the southern part is composed of discontinuous, small tectonic units containing late Cretaceous sediments. The middle belt of outcrops is constituted by the Montsec Mountains continuing westwards in the Tolva and Mediano anticlines and eastwards in the Aubens-Turb mountains. The northern belt of outcrops is quite continuous in its western part but is interrupted towards the east by the Abella-Boixols thrust-sheet.

As the Pyrenean Basin was approximately oriented along an east-west axis, its southern shore line followed the same general direction. Thus, the three belts of outcrops oriented parallel to the basin axis document three different depositional environments. While the sediments in the Marginal Sierras and in the Montsec were deposited on an interior platform, the outcrops in the northern belt between the Segre and Noguera-Pallaresa valleys document different facies types including the platform margin and deeper parts of the basin.



Text-Fig. 2. Sedimentary sequences (SQ.), from the Cenomanian to the end of the Cretaceous of the carbonate platform bordering the Central Pyrenean Basin on its southern margin (after SIMO 1986).

3. Carbonate platform evolution

On the carbonate platform bordering the southern margin of the Central Pyrenean Basin five major sedimentary sequences have been recognized by SIMO (1986). They represent strata deposited from the Cenomanian to the end of the Cretaceous, and are numbered 1-5 from bottom to top, summarized by Text-Fig. 2 and described below:

1. The platform sequence 1 starts with the progradation of limestone deposits over deeper marls. The carbonate facies, mainly pelletoidal wackestones and limy mudstones, are rich in *Praealveolina cretacea, P. simplex, Ovalveolina ovum, Chrysalidina gradata, Charentia cuvillieri, Daxia cenomana, Cuneolina* spp., etc. Shales and glauconitic marls with slump scars represent the basinal facies and contain planktonic foraminifera from the Rotalipora cushmani Zone (Middle?-Upper Cenomanian).

2. Sequence 2 starts with *Pithonella* limestones documenting the drowning of the shelf margin in the Pyrenean Basin. This depositional sequence is shallowing and coarsening upward in every locality. Larger foraminifera are absent. Deeper fans consist mostly of nodular marls alternating with shales rich in planktonic foraminifera belonging to the Marginotruncana schneegansi Zone (late Turonian-early Coniacian). According to SIMO (1986), the ending

of this sedimentary sequence resulted from a tectonic tilting of the basin combined with a drop of relative sea level. This enhanced siliciclastic sedimentation on the platform by increased erosion in the hinterland and the triggering of megabreccias on the slope margins.

3. Shallow, carbonate platform deposits in the Montsec and St. Corneli areas, slope marls in the Flamicell- and Ribagorçana valleys and basinal, ribboned limestones in the Esera valley belong, according to SIMO (1986), to a third sequence. The platform carbonates are very rich in larger foraminifera forming extremely diversified assemblages (CAUS & CORNELLA 1983, level 4) typical for restricted lagoons (*Vidalina, Cyclogyra* etc.) and protected platforms (discoidal, agglutinant larger foraminifera, larger miliolids and rotaliids). The corresponding open shelf and slope facies are rich in planktonic foraminifera from the Dicarinella concavata Zone (late Coniacian-early Santonian).

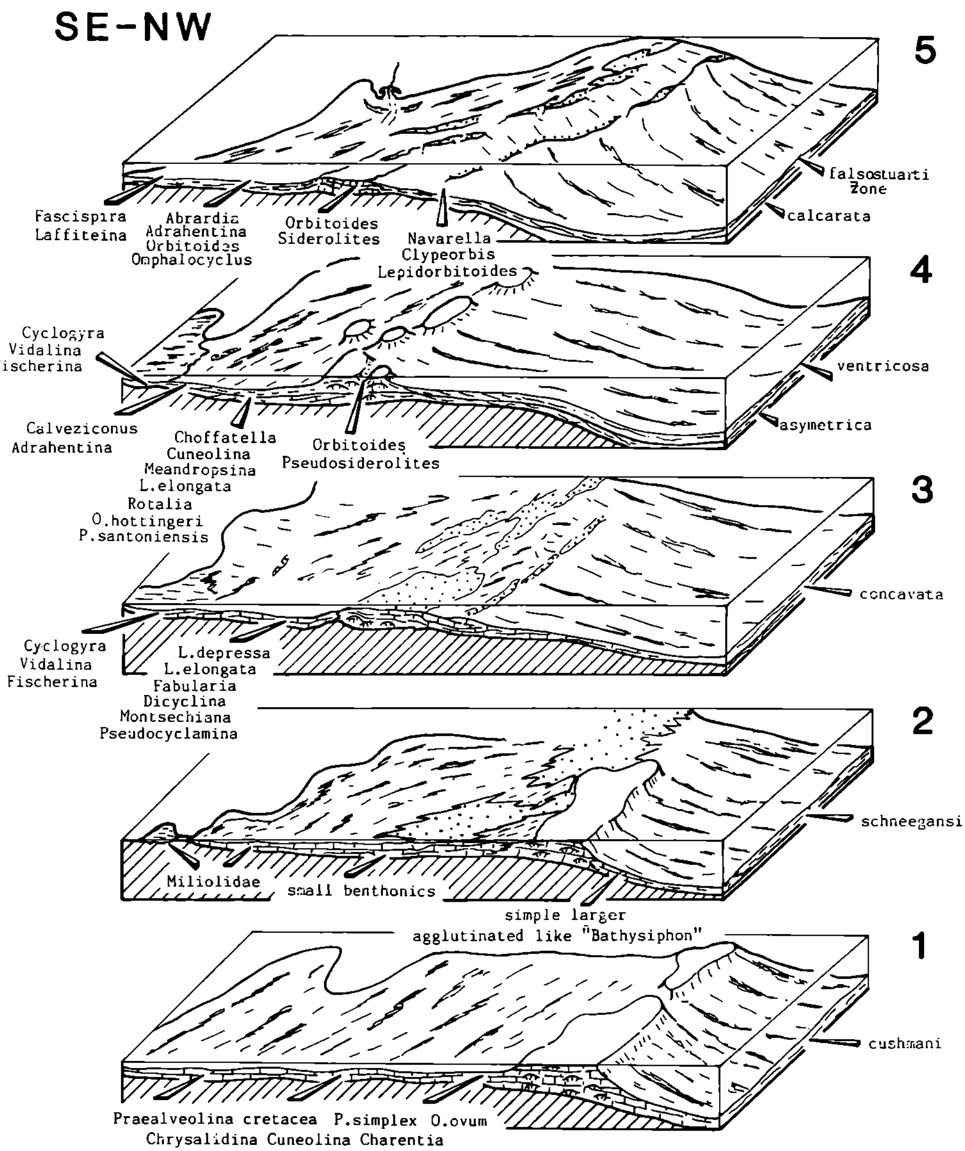
4. During the largest extension of the late Cretaceous sea, the sediments of a fourth sequence were deposited: turbidites on the slope and in the basin (Flamicell valley), shales on the shelf, rudist reefs (Collades) and carbonate platform fans (Montsec). The depositional geometry permits reconstruction of a shallow platform in the south deepening gradually northwards to where shelf shales and shallow water limestones were deposited. A deeper basin located northwestwards was filled with turbidites. The platform sediments constitute a carbonate ramp composed of very elongate limestone-bodies with large-scale crossbedding prograding north and northwestwards (CAUS, CORNELLA & GOMEZ-GARRIDO 1982). In the various shallow water environments the larger foraminifera were very abundant and diverse (CAUS & CORNELLA 1983, levels 5 and 6). Outer shelf and turbiditic facies are rich in planktonic foraminifera from the Dicarinella asymetrica, Globotruncanita elevata and Globotruncana ventricosa zones of late Santonian and Campanian age.

5. During Campanian (at the end of the G. ventricosa Zone) the exposure of the Pyrenean Basin to tectonics changed from extension to compression. The beginning of compressional movements is reflected by the collapse of the shelf edge leading to the deposition of slumped nodular limestones, debris flows and conglomerates (SIMO 1986). The olistostrome at the foot of the escarpment is composed of mixed deposits derived from siliciclastic and carbonate neritic source areas, and imbedded in basinal sediments. It reflects therefore, all by itself, the history of the different parts of the basin.

This sequence represents the final stage of the late Cretaceous marine basin fill. Its predominantly siliciclastic lithology reflects a shallowing upward cycle admitting at the top sedimentary structures made by tidal and wave currents. Larger foraminifera are abundant in all shallow-facies types from restricted to open environments. By direct correlation they belong to the Globotruncanita calcarata (level of the olistostrome) and Globotruncana falsostuari zones of upper late Campanian and early Maastrichtian ages.

4. Comments on plankton - shallow benthos correlations

The identification of depositional sequences in shallow and deeper environments helps to establish time correlations between deposits and their characteristic faunas from the interior, restricted platform to the open, outer platform and/or the basin (Text-Fig. 3). The distribution of planktonic



Text-Fig. 3. Time correlations between deposits and their characteristic fauna from the interior, restricted platform to the open, outer platform and/or the basin. 1. Upper Cenomanian. 2. Upper Turonian-Lower Coniacian. 3. Upper Coniacian-Lower Santonian. 4. Upper Santonian-Middle Campanian. 5. Upper Campanian-Lower Maastrichtian.

foraminifera in each sedimentary sequence dates its period of deposition and thus also the faunal content in its shallower equivalents. Consequently, the Pyrenean *Ovalveolina-Praealveolina* assemblage lived in the late Cenomanian as they are found in depositional sequence number I dated by Rotalipora cushmani Plankton Zone.

In the Turonian sequence (SQ. 2), no larger, complex foraminifera occur. The role (in ecology and as evolutionary stock) of larger agglutinated forms with a simple structure in deeper waters (*Bathysiphon* etc.) is far from clear and has to be studied in the future.

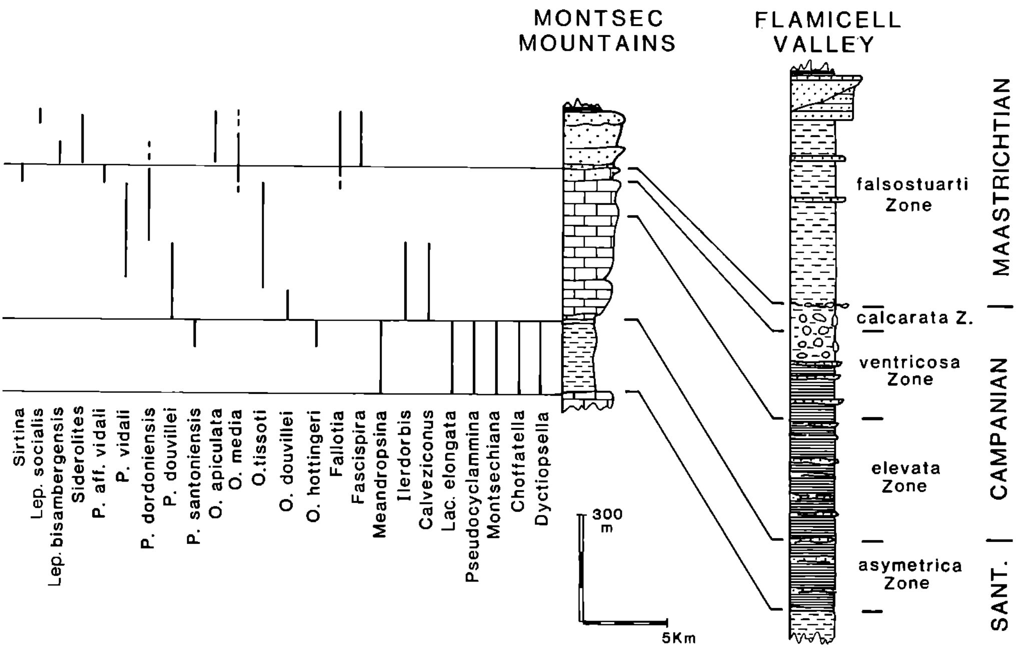
During the late Coniacian D. concavata Zone (SQ. 3), the first diversified assemblage of larger, complex shallow foraminifera appears. While in restricted lagoons Ophtalmiidae s. l. are common, the open, shallow environments yield discoidal, agglutinated foraminifera, meandropsinids, complex miliolids and large rotaliids.

Thus, the famous Pyrenean fabulariid assemblage dominated by *Lacazina depressa, Periloculina zitteli, Montsechiana, Cyclopsinella, Dicyclina* and *Orbitokathina* starts in the late Coniacian.

The environmental changes reflected by the transition from sequence 3 to 4 are responsible for the appearance of new assemblages, where discoidal agglutinant forms are substituted by conical ones. Larger spherical or flattened miliolids like *L. depressa* give way to elongated *L. elongata* (CAUS 1988). At the same time, the first Siderolitinae (*Praesiderolites santoniensis*) and the first orbitoids (*O. hottingeri*) appear. They constitute the root or first link in chains of long phylogenetic development. However, similarity of their habit to *Calcarina calcar* and *Planorbulinella,* as well as the type of bioclastic, washed-out sediments providing these fossils, indicate a local increase in water energy during this sedimentary cycle.

During deposition of the fifth sedimentary sequence, starting in the late Campanian, the development of consecutive species in the lineages of *Orbitoides, Lepidorbitoides* and in the Siderolitinae occurred. *Orbitoides tissoti* associated with *Pseudosiderolites vidali* correlates with the zones of G. elevata and G. ventricosa. *O. media* occurs in the upper part of the G. ventricosa and G. calcarata zones. Together with *O. media* we find also a stratigraphic successor of *P. vidali, Pseudosiderolites* aff. *vidali* in CAUS, RODES & SOLE (1988). These forms are much flatter and smaller than the earlier nominal species and seem to be confined to some platform facies of considerable depth. We do not know at present if the flattening trend represents an ecologic response to greater depth, as in recent larger foraminifera (HALLOCK & HANSEN 1979, REISS & HOTTINGER 1984), a phylogenetic change with time, independent from local facies changes, or both, i. e. an adaptation to deeper environments with time realized by genetic change.

The early Maastrichtian correlations of planktonic and shallower benthic foraminifera are particularly useful. *Fascispira* and *Laffitteina* from lagoons, orbitoids and siderolitids from the inner platforms and the *Lepidorbitoides* (*bisambergensis* and *socialis*) from outer platform bars can all be correlated with the plankton zonation. In particular, we have to point out that *Lepidorbitoides socialis* occurs in the G. falsostuarti Zone, its age being thus early Maastrichtian. Nevertheless, there are differences in the morphology between the *L. socialis* from the Tremp Basin and the topotypes of the species (CAUS, GOMEZ-GARRIDO & RODES 1988).



Text-Fig. 4. Distribution of selected larger foraminifera during deposition of the forth and fifth sedimentary sequence.

5. Conclusion

The distinction of 5 sedimentary sequences in the late Cretaceous deposits on the southern margin of the Central Pyrenean Basin permits correlation of assemblages of larger foraminifera from various shallow environments with the zonation of planktonic foraminifera elaborated in deeper areas. From these correlations it must be concluded that the well-known and widespread associations of "microfacies" in the sense of CUVILLIER (1951) do not correspond to chronostratigraphic stage units as traditionally used in the study of microfacies. However, the chronostratigraphic adjustments and precisions given to many classical microfacies types (such as the *Lacazina depressa* limestones dated traditionally as "Santonian") are minor and confirm, after all, the biostratigraphic value of such microfacies units as used in practical geology, i. e. for mapping. In fact, this biostratigraphic value reflects simply the short, generic time-range of the larger foraminifera, highly specialized, rapidly evolving k-strategists with a complex internal structure easy to identify from random sections in the slides (CAUS 1981).

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Paleoenvironmental Reconstruction of the Weald around Uña (Serranía de Cuenca, Cuenca Province, Spain)

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With 23 Text-Figures

GIERLOWSKI-KORDESCH, E. & JANOFSKE, D. (1989): Paleoenvironmental Reconstruction of the Weald around Uña (Serranía de Cuenca, Cuenca Province, Spain). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 239-264. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract. The depositional paleoenvironment for the Weald (Lower Cretaceous) around Uña in the Serrania de Cuenca (Cuenca Province, Spain) was analysed. A coal deposit and the presence of early mammal and reptilian remains makes these Weald deposits especially interesting. Two main sedimentary facies were recognized: Carbonate and Clastic. Each facies was divided into subfacies on the basis of sedimentological features. These facies were interpreted to represent an alluvial plain-lake system. A semi-arid, warm climate with seasons is postulated for the Weald around Uña. In order to explain the preservation of coal, subsidence, sediment supply, and basin hydrology were named as important variables in maintaining the necessary conditions for plant preservation within carbonates.

Kurzfassung. Vorgestellt wird ein Modell des Ablagerungsraumes des Wealden (Unter-Kreide) in der Umgebung von Uña in der Serrania de Cuenca (Prov. Cuenca, Spanien). Besonders berücksichtigt wurde das Profil von Uña mit einem Kohlevorkommen, das gut erhaltene Reste von frühen Säugetieren und anderen Wirbeltieren enthält. Es treten kalkige und klastische Faziestypen auf, die anhand sedimentologischer Faktoren wiederum in mehrere Subfazies-Typen unterschieden und beschrieben werden. Diese Subfazies-Typen repräsentieren Sedimente einer einem Hochland vorgelagerten alluvialen Ebene mit ganzjährig aktiven Flüssen und Seen. Ein semiarides, warmes Klima mit wechselnden Jahreszeiten wird angenommen. Voraussetzung für Entstehung und Erhaltung der Kohle von Uña waren Subsidenz, Sedimentzufuhr und ein relativ stabiler Grundwasserspiegel.

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1. Introduction

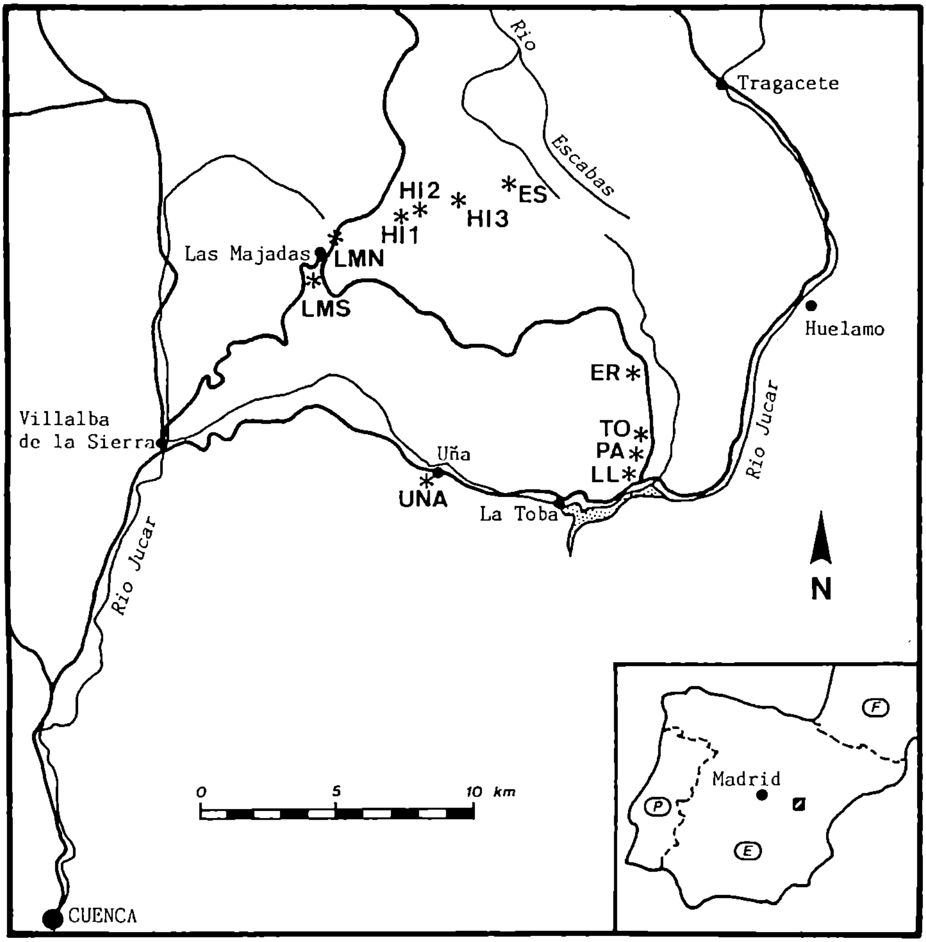
In the Serrania de Cuenca of the Southwestern Iberian Ranges of Spain, the Lower Cretaceous Weald is exposed as NW-SE trending basins (RAMIREZ DEL POZO & MELENDEZ HEVIA 1972). These sedimentary basins may have formed as stepped graben structures in strike-slip basins associated with the rotation of the Iberian plate and the tectonic activity related to the Tethys and the Biscayne seaway during the late Hauterivian to early Barremian (VILAS et al. 1983). The Weald near Uña unconformably overlies Upper Jurassic marine limestones and is conformably overlain in some areas by the Upper Cretaceous fluvial Utrillas Sandstone (RAMIREZ DEL POZO et al. 1975). The exact age of the Weald in the area around Uña is uncertain; age determinations based on ostracodes and charophytes (BRENNER 1976) range from the Valanginian to the Aptian while pollen analysis at Uña by MOHR (1987) resulted in an early Barremian age.

The Weald in the Serrania de Cuenca is generally recognized by its mixture of carbonates and clastics with minor coal seams (RAMIREZ DEL POZO & MELENDEZ HEVIA 1972). Especially noteworthy is the presence of a sub-economic coal deposit at Uña. Within the coal, mammal remains along with fish, frogs, lizards, turtles, crocodiles, and dinosaurs are present (KÜHNE & CRUSAFONT 1968, HENKEL & KREBS 1969, HENKEL 1970). Also present are charophytes, blue-green algae, ostracodes, gastropods, and bivalves in the Wealdian limestones of Uña and the surrounding area (BRENNER 1976, RAMIREZ DEL POZO et al. 1975, MELENDEZ HEVIA et al. 1975b, and this paper).

The paleoenvironment of the Weald around Uña had been interpreted to be "fluvial to brackish to marine" (RAMIREZ DEL POZO et al. 1975, MELENDEZ HEVIA et al. 1975a, b). Recent sedimentological work in the Cretaceous southeast of Uña (MELENDEZ 1983, MAS et al. 1982) indicate that within these grabens, continental conditions alternated with marine features throughout the Lower Cretaceous. However, it appears that the area around Uña remained continental throughout the Lower Cretaceous.

1.1 The Problem

An attempt has been made here to further refine the paleoenvironmental model of the Weald around Uña in order to add information to the knowledge of the paleoecology of the early mammals. Sedimentological methods employed in this study included the identification of the different facies and subfacies through field observations, limited field analysis of facies relationships, and some thin section analyses. This facies analysis was carried out in the area around Uña, Las Majadas, and Pantano de la Toba (Text-Fig. 1). This area is about 30 km north of Cuenca which is about 200 km east of Madrid. The longer, more continuous measured sections are schematically shown in Text-Fig. 2. Correlation among sections is impossible since the facies changes are rapid and no detailed work on the ostracode and charophyte stratigraphy has yet been done. Even the lower contact with the Upper Jurassic limestones may not be diachronous. An early Barremian age for the Uña coal is established by MOHR (1987) through the use of pollen and spores. And, according to MELENDEZ (1982, 1983, and personal communication), the Weald around Uña may correspond to the 2nd sedimentary cycle of the 1st Tecto-sedimentary Phase which is continental in character (La Huérguina Formation) and is late Hauterivian to early Barremian in age.

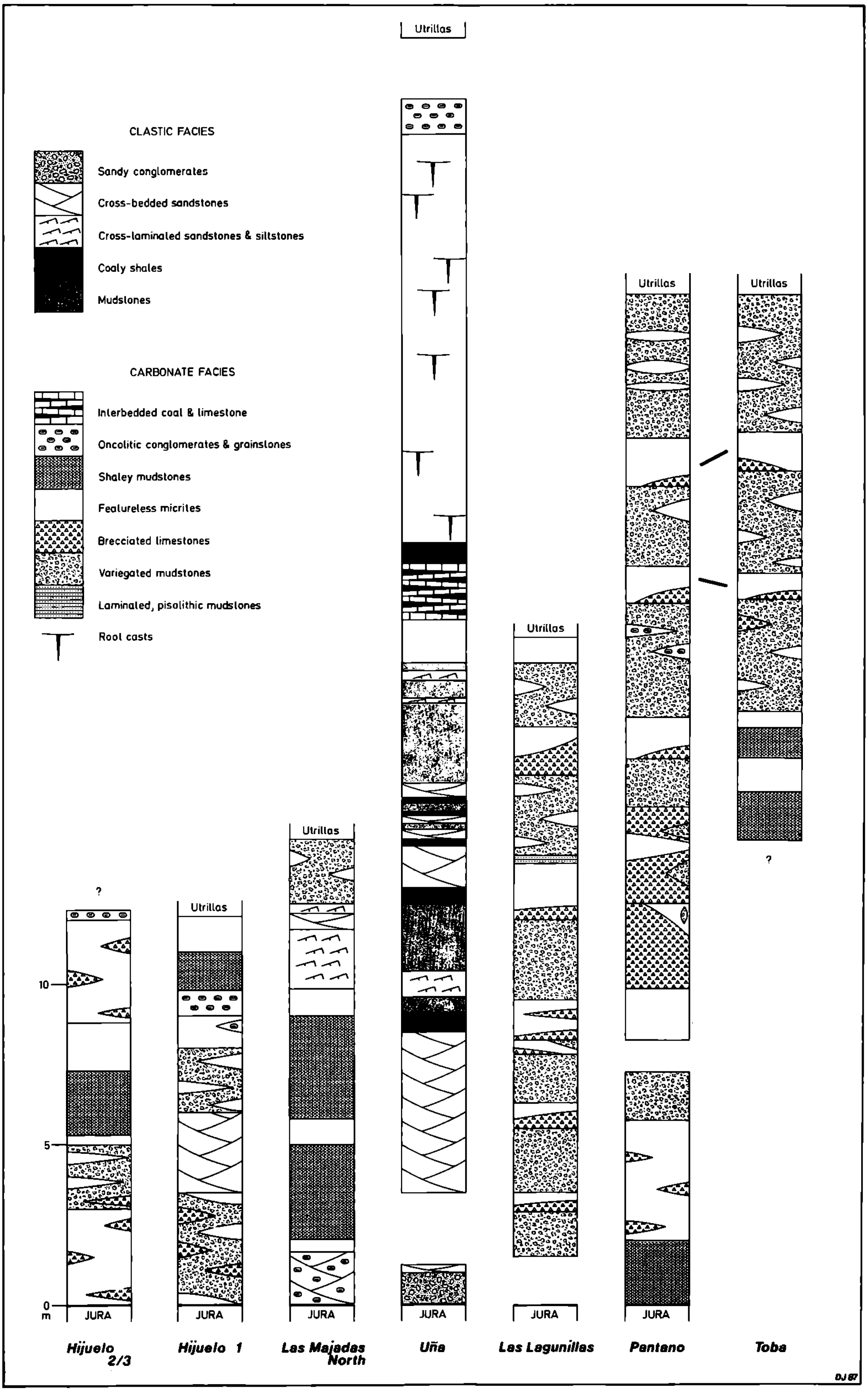


Text-Fig. 1. Locality map of the study area. Sections: UNA: Uña; LL: Las Lagunillas; PA: Pantano; TO: Toba; ER: Erquilla; LMS: Las Majadas South; LMN: Las Majadas North; HI: Hijuelo; ES: Espiño.

2. Facies Description

Two main facies were recognized in the Weald around Uña: the Carbonate Facies and the Clastic Facies (see Text-Fig. 2). Separate subfacies were recognized within both facies. The Clastic Facies is composed of:

(1) Sandy conglomerates
(2) Cross-bedded sandstones
(3) Cross-laminated sandstones and siltstones
(4) Coaly shales
(5) Mudstones



CLASTIC FACIES
Sandy conglomerates
Cross-bedded sandstones
Cross-laminated sandstones & siltstones
Coaly shales
Mudstones
CARBONATE FACIES
Interbedded coal & limestone
Oncolitic conglomerates & grainstones
Shaley mudstones
Featureless micrites
Brecciated limestones
Variegated mudstones
Laminated, pisolithic mudstones
Root casts
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Las Majadas North
Uña
Las Lagunillas
Pantano
Toba
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The Carbonate Facies contains:

(6) Interbedded coal and limestone
(7) Oncolitic conglomerates and grainstones
(8) Shaley mudstones
(9) Featureless micrites
(10) Brecciated limestones
(11) Variegated mudstones
(12) Laminated, pisolithic mudstones

Facies identification was based on general sedimentary features at outcrop scale, not on a microfacies level. Thin sections only supplemented interpretations. Simplification was deemed necessary due to the large variations within subfacies.

2.1 Descriptions of the Subfacies and Their Interrelationships

2.1.1 Subfacies of the Clastic Facies

Only at Uña (Text-Fig. 2) is the entire spectrum of the Clastic Facies exposed. Everywhere else in the study area only cross-bedded and cross-laminated sandstone lenses can be found, seemingly interrupting the "background" carbonate sedimentation. Evidence from the Weald southeast of Uña (MELENDEZ 1983) indicates that the clastic and carbonate depositional systems may have been physically separated within the early Barremian sedimentary basin(s). The separation appears to take the form of proximal clastic deposits versus distal carbonate deposits. This may be due to the hydrodynamic differences between clastic and carbonate detrital particles. Evidence will be presented here as to the detrital nature of the carbonate depositional system. Also, biological and/or chemical precipitation of carbonates occurs. Subfacies description of the Clastic Facies follows.

2.1.1.1 Sandy conglomerate subfacies

A single bed is present at Uña with a thickness of half a meter. The sizes of the cobbles range in size from 10 - 20 cm long. These cobbles float in a coarse quartz sandstone matrix. No sedimentary structures were visible due to the limited exposure.

2.1.1.2 Cross-bedded sandstone subfacies

The character of these sandstones varies across the study area. The grain size of the sandstone components varies from coarse to very coarse quartz grains to intermixed granules and pebbles of red Triassic mudstones, yellow Jurassic limestones (found at Las Majadas South), and Wealdian limestone

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Text-Fig. 2. Schematic sections of the Weald illustrating subfacies distribution in the area surrounding Uña. For section locations, see Text-Fig. 1.

fragments and oncolites. Cobbles and boulders of Wealdian limestones are rarely found within a predominately very coarse quartzose sandstone at Espiño. Bedding units appear to range within a decimeter scale except at Uña where a coarse to granular quartzose sandstone bed is about 3 m in thickness. A lens morphology for the sandstone beds is commonly observed when exposure allows it.

The most common sandstone of this subfacies found throughout the study area is a very coarse to granular quartzose sandstone containing pebble to cobble-sized oncolites (Text-Fig. 3). The cores of the oncolites are various rock fragments of clastic or carbonate origin.

Low-angle trough cross-stratification with decimeter scale cross-sets is the ubiquitous sedimentary structure linking these sandstones together in this subfacies. Paleocurrent measurements from these sandstone subfacies in Uña, Hijuelo 1, Las Majadas North, and Las Majadas South indicate paleocurrent directions from the north and west. This subfacies is gradational in composition to the oncolitic conglomerate and grainstone subfacies.

2.1.1.3 Cross-laminated sandstones and siltstones

This subfacies is only found at Uña and Las Majadas North. Fine quartzose sandstones to siltstones are poorly exposed at Uña while fine to medium-grained quartzose sandstones are found at Las Majadas North. At the latter exposure, depositional units are on average about 10 cm in thickness and are interbedded with medium to coarse-grained trough cross-bedded quartose sandstones. These clastic beds contain carbonaceous plant material, especially decimeter long stem pieces.

2.1.1.4 Mudstones and coaly shales

These two subfacies are only found at Uña and are poorly exposed. Unit thicknesses are generally on the decimeter scale. The contact between the

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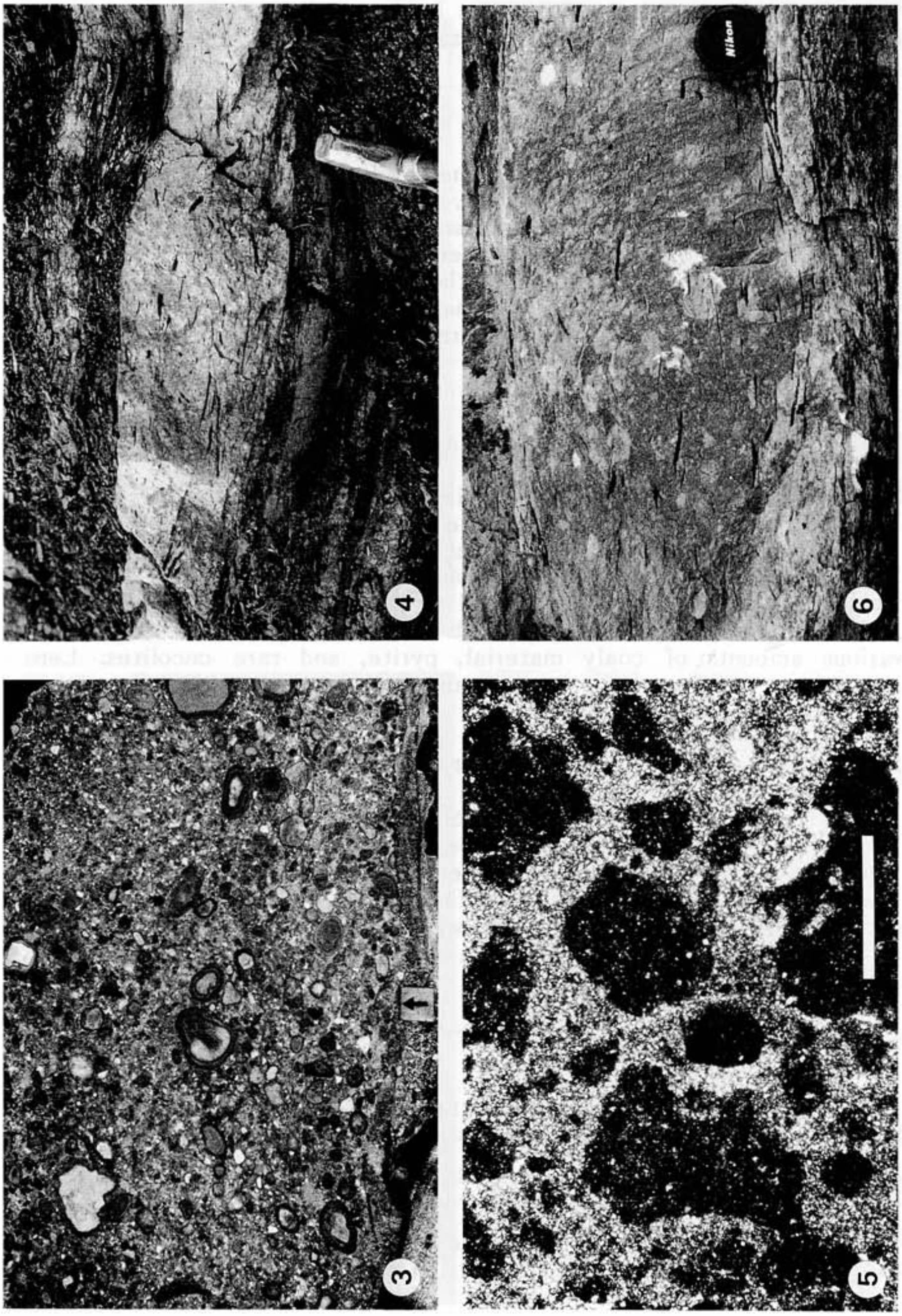
Text-Fig. 3. A polished rock slab of the oncolitic sandstone from Erquilla. Note the oncolitic envelope surrounding the rock fragments; many rock fragments are Wealdian in origin. Scale: 1 cm^2 with arrow indicating correct orientation.

Text-Fig. 4. Outcrop photo of a limestone lense within the coal layers at Uña. Coalified plant material is distributed throughout the unit. Hammer for scale.

Text-Fig. 5. Micrite and macrosparite texture within the featureless micrites of Pantano (PA 7). Thin section. Scale: 0.5 mm.

Text-Fig. 6. Outcrop photo of featureless micrites at Uña. Note the differential layering due to coalified plant material density. The lower layer is almost shaley from horizontally oriented plant material while upper thick limestone layer contains widely dispersed material. Lens cap is 5.5 cm in diameter.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Clastic and Carbonate Facies occurs in the mudstone subfacies and is gradational with a silty featureless micrite.

2.1.2 Carbonate Facies

This facies dominates most of the localities while it encompasses the upper part of the section at Uña. The carbonate subfacies are generally gradational to one another with most beds found to be lensoid in character when followed over a distance of tens of meters. This is schematically represented in Text-Fig. 2. Some of the subfacies contain substantial amounts of clastic material; also a gradational spectrum between oncolitic sandstones and oncolitic conglomerates and grainstones can be recognized. Subfacies descriptions follow.

2.1.2.1 Interbedded coal and limestone

Approximately 2.5 m of this unit is exposed at the Uña section. The coal is described as brown coal or lignite (HENKEL & KREBS 1969, MELENDEZ HEVIA et al. 1975). No petrographic analysis has been done to date. The main coal body with a minimal amount of limestone is approximately 65 cm in thickness and is known to extend at least 300 m northeast into the local abandoned mine. The limestone beds are lenses of featureless micrite with various amounts of coaly material, pyrite, and rare oncolites. Lens sizes range from centimeter to meter scale thickness (Text-Fig. 4).

2.1.2.2 Featureless micrites

This white to grey-coloured subfacies is found at all localities with bed thicknesses in centimeter to meter scales. The micrites are interbedded and gradational with the brecciated limestone, variegated mudstone, and the oncolitic and grainstones subfacies. The featureless micrite beds in general also have a lensoid shape which can be observed to be up to tens of meters in length. Fresh broken surfaces reveal a homogeneous tan to

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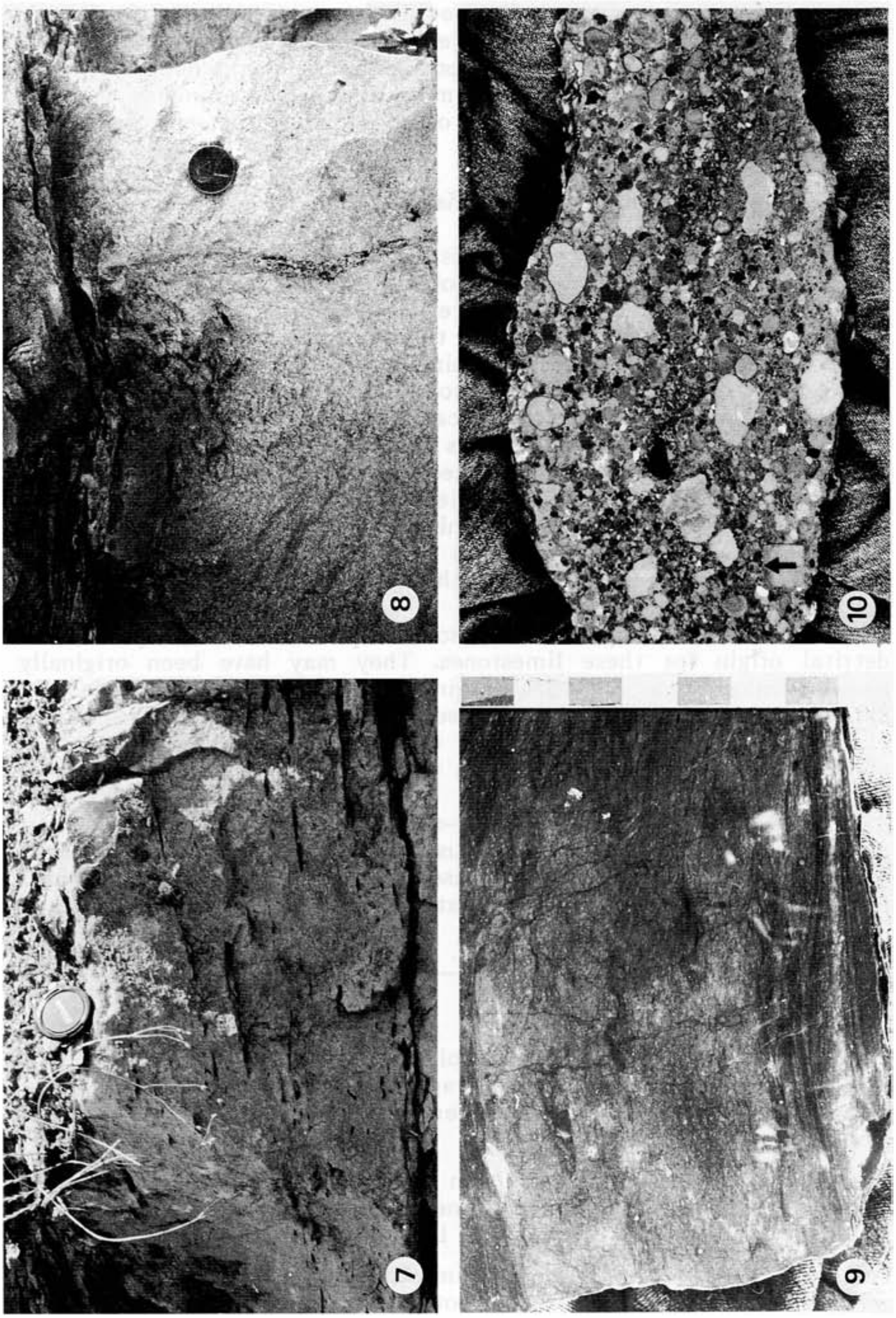
Text-Fig. 7. Relict trough cross-stratification as defined by coalified plant fragments. Uña. Outcrop photo. Lens cap is 5.5 cm in diameter.

Text-Fig. 8. Outcrop photo of featureless micrites at Uña. Note the carbonized rhizolith projecting downward. Penny for scale.

Text-Fig. 9. Rock slab of carbonaceous micrite. Pantano. Note lamination at the bottom while the upper portion is disturbed by bioturbation and/or rhizoliths (PA 4A). Centimeter scale; correct orientation. The colour is dark brown to black but picture was underexposed to show detail; white spots at the bottom are reflections.

Text-Fig. 10. Rock slab of a conglomeratic grainstone. Toba (TB 4). Scale: 1 cm^2 with arrow indicating correct orientation.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grey limestone with common sparite-filled specks and tiny fossil fragments. Though no sedimentary features are immediately obvious, thin sections reveal that fabrics are variable. A spotted micritic and microsparitic texture, as shown in Text-Fig. 5, is common. Charophyte remains and ostracode shells plus their fragments also occur singly dispersed or assembled in batches within a micritic matrix. Also, gastropod and bivalve shells are rarely found in this unit.

The featureless micrites of Uña are uniquely different from those in the other localities because these micrites contain substantial amounts of coalified plant material (as perhaps transitional to the coal and limestone unit). Charophytes, ostracodes, oncolites, and their fragments can be found among the carbonaceous plant material. The beds appear to be defined by the amount of coal material. The thicker units (ranging from a decimeter to half a meter in thickness) contain dispersed plant material and alternate with layers up to 10 cm thick containing mostly aligned coalified plant material almost shaley in appearance (Text-Fig. 6). In rare instances (Text-Fig. 7), the plant material appears to define cross-bedding with decimeter scale cross-sets. Also, carbon-filled tubules resembling root casts which commonly branch can be found projecting vertically from the tops of various cross-sets (Text-Fig. 8). These "rhizoliths" (KLAPPA 1980) are circular to ovoid in cross-section extending up to half a meter in length. The calcitic tubule wall is perhaps 3 to 4 mm thick and probably contains the carbonized remains of plant material.

All this evidence plus the lensoid shape of the beds points toward a detrital origin for these limestones. They may have been originally tractionally-deposited as sand-sized grains and then subsequently altered through diagenesis and/or biological processes. More evidence follows. Confirmation of in situ biogenic precipitation of calcite is demonstrated by the presence of microstromatolites and algae in thin section within this subfacies in all localities.

Also, a similarly weathered, sub-subfacies is found within the featureless micrites. It contains a carbonaceous clastic clay component (perhaps up to 40 % by volume) and in some cases parallel lamination with bioturbation features or soil development (Text-Fig. 9). These carbonaceous micrites

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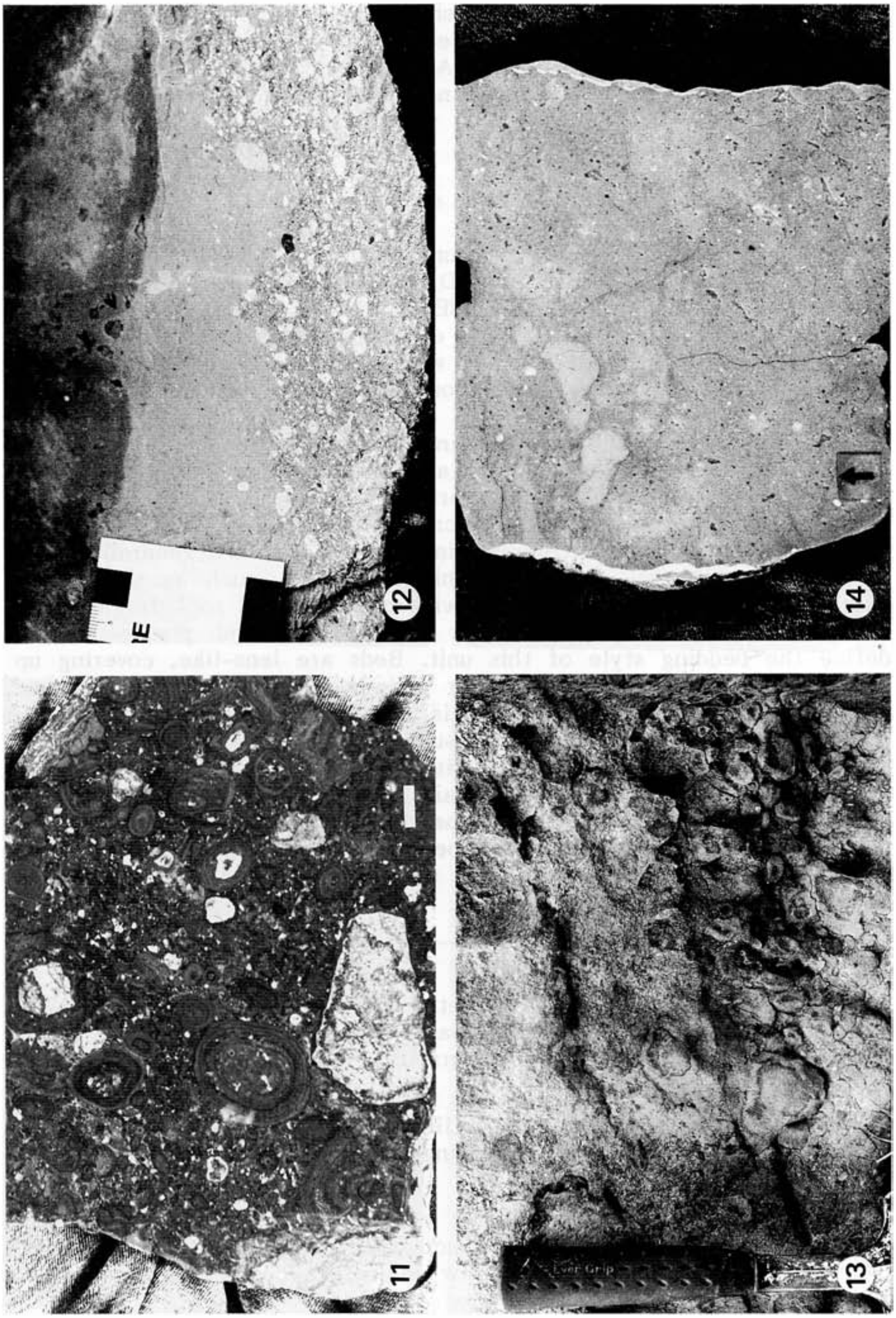
Text-Fig. 11. Rock slab of an oncolitic conglomerate from Hijuelo 1. Note oncolitic envelopes around rock fragments which are mainly Wealdian in origin (HI 1-1B). Scale in the lower right hand corner is 1 cm long. Up direction is toward the right.

Text-Fig. 12. Outcrop photo of an angular contact between a featureless micrite and conglomeratic grainstone. Float boulder - orientation unknown. Top of scale is in centimeters. Las Lagunillas.

Text-Fig. 13. Outcrop photo of limestone unit with indistinct outlines of oncolites and/or cobbles in the bottom portion with featureless micrites above. Between Las Lagunillas and Pantano. Hammer for scale.

Text-Fig. 14. Rock slab of a featureless micrite showing vague outlines of relict rock fragments. Las Lagunillas (LL 10). Scale: 1 cm² with arrow indicating correct orientation.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are found in association with the shaley mudstone subfacies and elsewhere (Pantano and Uña) and are considered to be gradational with that subfacies in composition. According to DARABI (1981), these carbonaceous micrites can be found in lenses up to 2 m in thickness throughout the area between Pantano and Uña.

2.1.2.3 Oncolitic conglomerates and grainstones

This subfacies is not commonly found in the measured sections or in the area between Uña and Pantano (DARABI 1981). But, along the Escabas River valley east of Espiño (KRAUSE 1980), and along the river valley containing Erquilla (BUCHELT 1976), oncolitic conglomerates are much more common. Unfortunately correlation with those measured sections is not yet possible since their stratigraphic position in relation to the basal Jurassic marine limestones is unknown.

These grey to tan units contain varying proportions of oncolites, carbonate rock fragments, and minor amounts of clastic rock fragments down to sand size (mostly Wealdian in origin) (Text-Figs. 10, 11). The oncolites range in size from small (~ 1 - 2 cm in diameter) to large (10 - 12 cm in diameter) and are round to ovoid in shape. Cores are generally rock fragments. Also, in some cases, biogenic fragments, such as shell pieces and charophyte remains, can be mixed with oncolites and rock fragments.

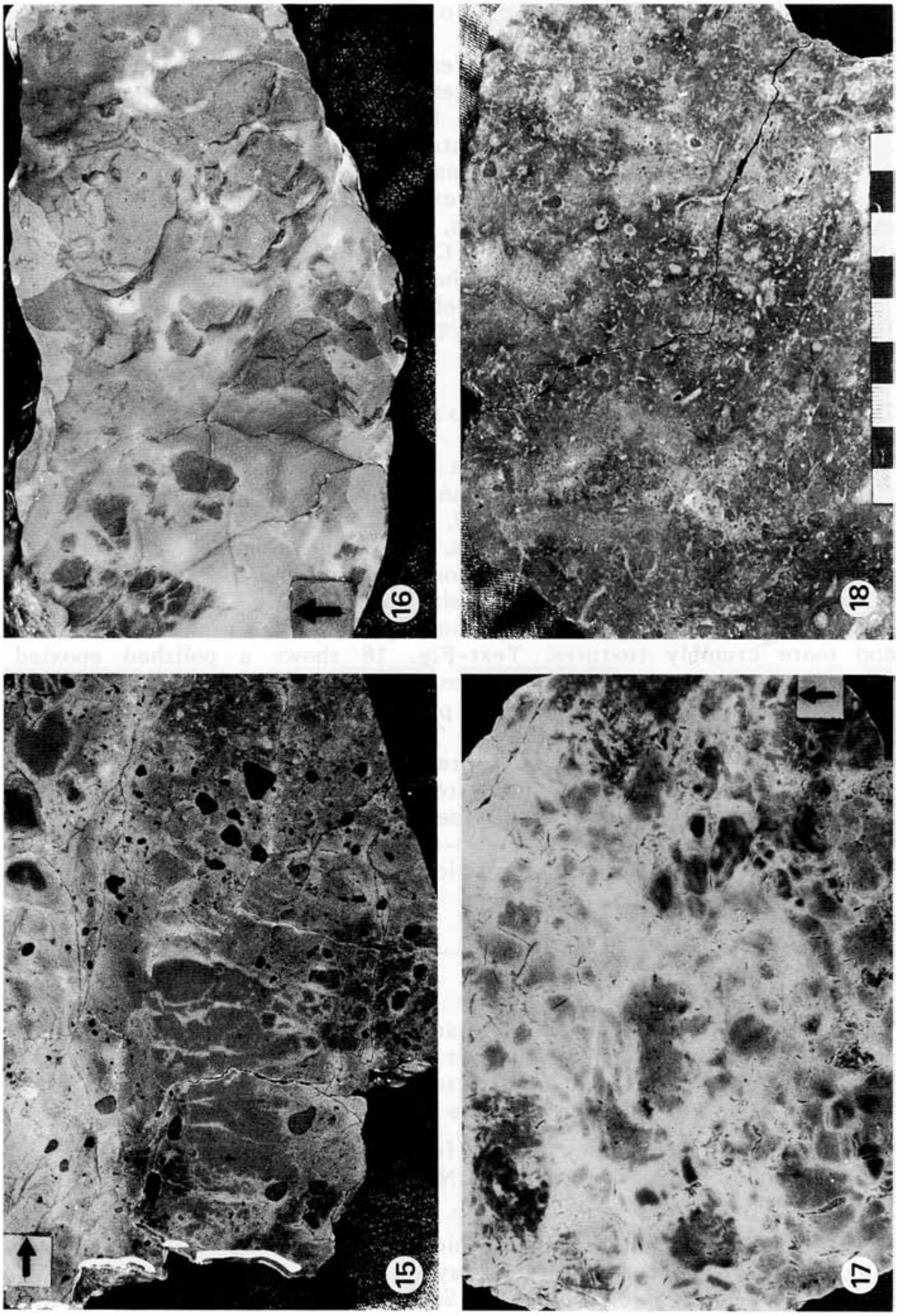
Cross-stratification, both planar and trough, and planar stratification define the bedding style of this unit. Beds are lens-like, covering up to a few meters in length. Paleocurrent directions appear to have been from the northwest to the northeast. This subfacies occurs with the featureless micrites and variegated mudstone subfacies. Its relationship with the featureless micrites appears varied with bottoms of micrite units containing sand-sized carbonates and quartz grains, or with abrupt, uneven, sharp-edged contacts (Text-Fig. 12). Also, vague, seemingly oncolitic to conglomeratic units (Text-Fig. 13) are found to be featureless micrites with vague remanents of rock fragments when fresh samples are taken (Text-Fig. 14).

Text-Fig. 15. Rock slab of brecciated limestone from Pantano. Note carbonate grains floating in micritic matrix (PA 2C). Scale: 1 cm^2 with arrow indicating correct orientation. Colours include red and yellow.

Text-Fig. 16. Rock slab of another type of brecciated limestone from Las Lagunillas. Note the indistinct outlines of relict rock fragments (LL 5). Scale: 1 cm^2 with arrow indicating correct orientation. Colours include yellow and tan.

Text-Fig. 17. Rock slab of another type of brecciated limestone from Las Lagunillas. Note absence of distinct rock fragments; a splotchy texture (LL 1B). Scale: 1 cm^2 with arrow indicating correct orientation. Colours include yellow, red, and tan.

Text-Fig. 18. An epoxied rock slab of the variegated mudstone facies from Las Majadas North. This reddish rock contains tiny, fine, branching lines with isolated rounded rock fragments (LM 14). Centimeter scale; up is toward the right.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2.1.2.4 Brecciated limestones

This unit is contiguous with the featureless micrites or occurs separately with the variegated mudstones as lenses just like the featureless micrites. Thicknesses vary but average about half a meter. Some weathered surfaces appear to be composed of brecciated material. Carbonate grains can be found floating within a micritic matrix (Text-Fig. 15) or angular grains with vague outlines also occur (Text-Fig. 16). Other weathered surfaces appear featureless, like the micrites, and only fresh surfaces (Text-Fig. 17) indicate subfacies differentiation. Colours on fresh surfaces include red, yellow, blue, brown and tan splotches; separate grains are not discernable. This colourful mottled feature is referred to as "marmorization" or "rubefaction" in the literature (cf. FREYTET & PLAZIAT 1982, VOGT 1984a).

2.1.2.5 Variegated mudstones

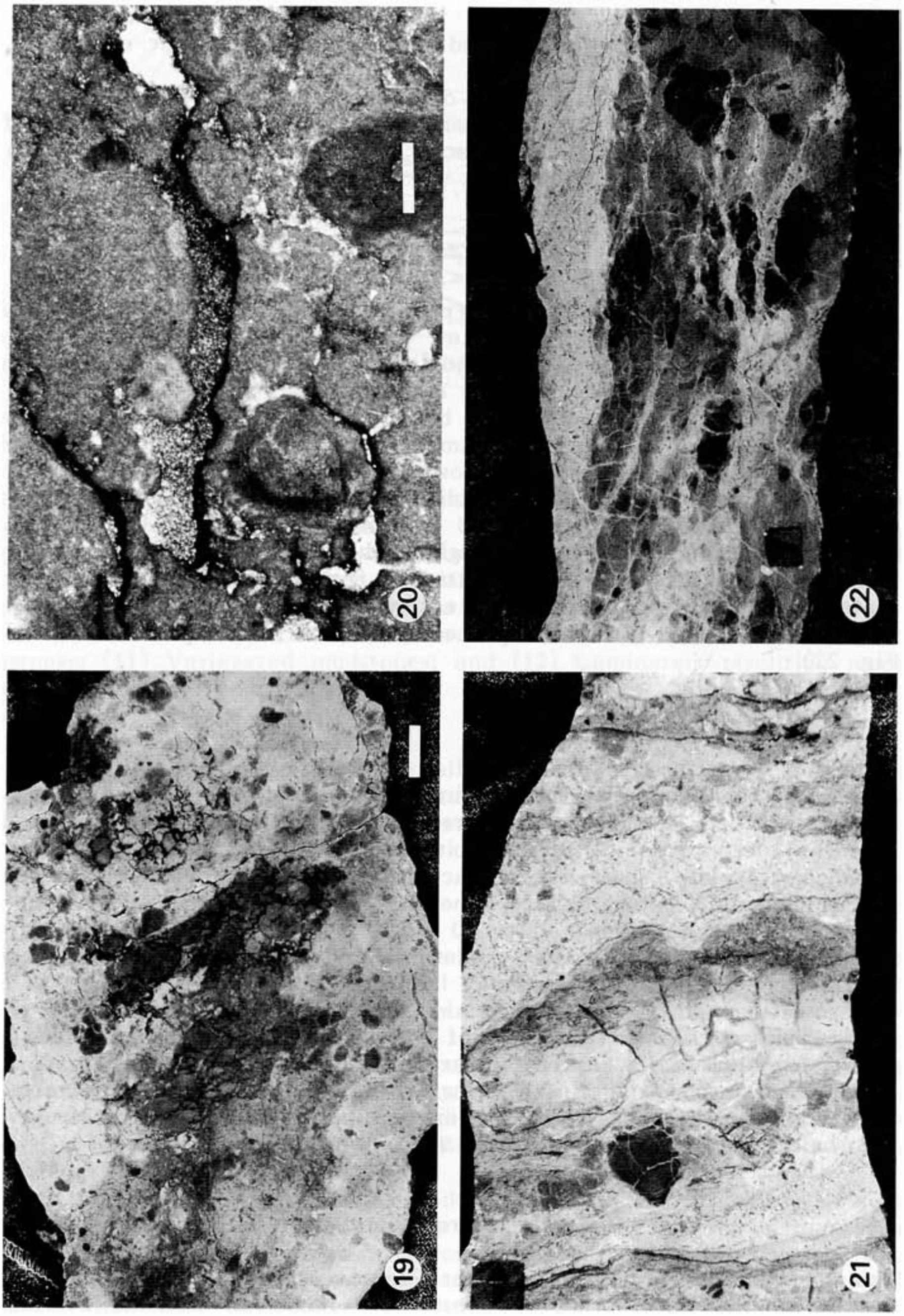
This subfacies is not found at Uña and appears to dominate the sections at Las Lagunillas, Pantano, Toba, and the area eastward from them. These friable carbonate mudstones are red, white, and yellow in colour and gradational with the featureless micrites. This relationship is best seen at Pantano where a 3 m cliff of limestone (containing featureless micrites and brecciated limestones) is seen lensing out into variegated mudstones with varying stages of "degradation" from light-coloured limestones into redder and more crumbly textures. Text-Fig. 18 shows a polished epoxied slab from the variegated mudstone facies. Note the splotchy texture with some dispersed breccia pieces and the presence of tiny, branching, fine lines filled with white calcitic material. Text-Fig. 19 exhibits a rock slab from a transitional zone between the brecciated limestone subfacies and variegated mudstones subfacies. The brecciation outlines appear more diffuse and darker, and clay material outlines or coats many of the breccia blocks and also seems to fill the micro-cracks found dispersed throughout the slab. A thin section of this transition region between subfacies (Text-Fig.

Text-Fig. 19. Rock slab from a transitional zone between brecciated limestones and variegated mudstones at Pantano. Clay material outlines many breccia blocks and fills many micro-cracks (PA 3H). Scale in the lower right is 1 cm long; up is toward the right.

Text-Fig. 20. Thin section photo from Pantano zone between featureless micrites and variegated mudstones. Note the clay cutans lining the sparite-filled voids (PA 3A). Scale: 0.5 mm.

Text-Fig. 21. Rock slab of pisolithic, laminated mudstone from Las Lagunillas. Note pisolites in various layers, floating Wealdian rock fragment, and vertical micro-cracks (LL 7). Scale: 1 cm^2 with arrow indicating correct orientation.

Text-Fig. 22. Rock slab of brecciated pisolithic, laminated mudstone from Las Lagunillas (LL 7). Scale: 1 cm^2 with arrow indicating correct orientation.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20) shows that the dark material is aligned clay which lines voids and cracks. These clays apparently lined voids which were later cemented with microsparite.

Also important is the presence of vertically elongated holes up to half a meter in length within the limestone cliff at Pantano which are filled with the variegated mudstone subfacies. More discussion will follow in the environmental analysis section.

2.1.2.6 Laminated, pisolithic mudstones

This subfacies is very rare and is presently only found at Las Lagunillas. The bed is lensoid in shape and a few decimeters in length with a maximum thickness of 30 cm. Its weathered surface is also similar to the featureless micrites which may explain its rare appearance.

This unit in parts appears to be crudely laminated on a centimeter scale with undulatory and uneven lamination contacts (Text-Fig. 21). Colours include white, grey, and tan to brown laminations. Close inspection of the laminae components indicates a pisolithic texture containing millimeter-sized roundish particles. Carbonate rock fragments up to 1.5 cm in size are apparently randomly dispersed throughout the laminae. Also, vertical to subvertical microcracks can be seen to cut through some layers. This interpretation of cracking is substantiated by other unit samples in which the brecciation of this subfacies proceeded before final preservation (Text-Fig. 22).

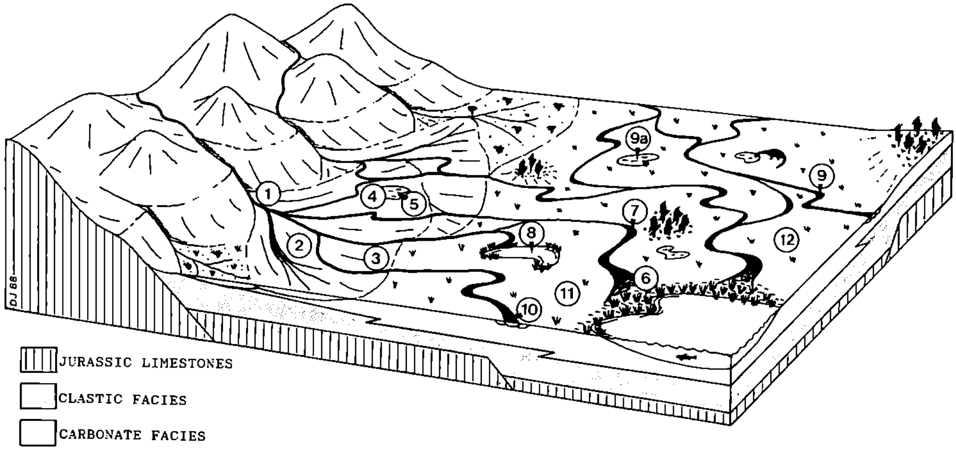
2.1.2.7 Shaley mudstones

This subfacies dominates the section at Las Majadas North and is also found at Hijuelo and Toba. This grey unit is generally poorly exposed, up to a few meters in thickness, and pinches out within tens of meters lateral distance. Its platy, shale-like character is attributed to a large amount of clastic material, especially clay minerals. The siliciclastic mud and silt-sized components reach perhaps up to 60 to 70 % by volume. The shaley mudstones are considered to be gradational in composition with the siltier featureless micrites. Discontinuous laminations of lignite are rarely found in association with this subfacies. Also, ostracodes are present.

At Toba, a 10 cm thick lens of a white carbonate is preserved within the shaley mudstone unit. This layer is composed of a micrite matrix with clastic and carbonate grains floating within it. Also present are vertebrate bone fragments. Since compaction did not seem to affect this bed within the shaley mudstone, an early diagenetic cementation is postulated.

3. Paleoenvironmental analysis

The continental origin of the Lower Cretaceous Weald around Uña is evidenced by the invertebrate and vertebrate fauna as well as by tectonic and stratigraphic considerations (MELENDEZ 1983). Coal deposits within continental basins are known to develop within lakes and/or marshes associated with fluvial, alluvial fan, and lake delta systems (McCABE 1984, CABRERA & SAEZ 1987). The coal of Uña and its surrounding deposits



Text-Fig. 23. Schematic block diagram depicting the depositional paleoenvironment of the Weald surrounding Uña. Labelled subfacies: (1) Sandy conglomerates; (2) Cross-bedded sandstones; (3) Cross-laminated sandstones and siltstones; (4) Coaly shales; (5) Mudstones; (6) Interbedded coal and limestones; (7) Oncolitic conglomerates and grainstones; (8) Shaley mudstones; (9) Featureless micrites; (9a) Carbonaceous micrites; (10) Brecciated limestones; (11) Variegated mudstones; and (12) Laminated, pisolithic mudstones. See section 3.0 for discussion.

are interpreted to be part of an alluvial plain-lake-marsh depositional system influenced by a seasonal climate (Text-Fig. 23). This section outlines the evidence supporting this sedimentological model.

3.1 Alluvial floodplain system

The subfacies within the Clastic Facies are interpreted to be a transition from alluvial fan to alluvial floodplain. The basal sandy conglomerate with matrix-supported cobbles could have been the remains of debris flow deposits (HOOKE 1967, RUST & KOSTER 1984). The lens-shaped cross-bedded sandstones interbedded with cross-laminated sandstones and siltstones, coaly shales, and mudstones may represent an alluvial floodplain containing channels separated by a floodplain containing small lakes or marshes (ETHRIDGE et al. 1981, GERSIB & McCABE 1981, FLORES 1981, RUST et al. 1984).

The transition from a siliciclastic-dominated system to a carbonate-dominated system at Uña could be the result of a change of source area and/or a widening of the basin through successive block faulting (cf. MONTY & MAS 1981) with the resultant displacement of the depositional center of the basin toward the area around Uña. This displacement could cause the change of depositional milieu from proximal clastics to distal carbonates. Due to the limited number of exposures and specific tectonic data, this is pure speculation. Additional evidence for contemporaneous tectonic activity is the seemingly sporadic influx of coarse sandstone lenses into a carbonate milieu. Climatic effects cannot be ruled out, but these sandstone deposits are limited in size and extent. Correlation among these sandstone bodies and to the basin edge to substantiate tectonic influence is presently impossible.

The Carbonate Facies is interpreted to be deposits of detrital carbonate alluvial plain-lake system. In other words, carbonates were traction-deposited and then subsequently chemically and/or biologically reworked in multiple cycles within a complex continental depositional system. Within the channel and extrachannel deposits of the alluvial plain, post-depositional processes included calichification, karstification, gleying (hydromorphic soils), pedogenesis, micritization, dissolution, recrystallization, cementation, bioturbation, or any combination. Sedimentation events alternating with subaerial exposure, each of varying duration, altered the carbonate deposits by different degrees into a complex mosaic of facies. A wide range of diagenetic grades is evident among and within carbonate facies. A short discussion of each facies follows.

The oncolitic conglomerates and grainstones probably represent the least unaltered or lowest diagenetic grade of carbonates. Interpreted as channel deposits, its detrital nature is confirmed by grain-supported cross-stratification features. A similar interpretation for this type of oncolitic facies is given by MELENDEZ (1983) and MONTY & MAS (1981) for the Weald exposed to the southeast toward Valencia, which is postulated to be the southern part of the same basin. Oncolites are known to occur in river channels, marshes, lakes, and floodplains of the past and present (GLAZEK 1965, MONTY & MAS 1981, ANADÓN & ZAMERREÑO 1981, NICKEL 1983, ORDÓÑEZ & GARCIA DEL CURA 1983).

The close interrelationship of the oncolitic conglomerates and grainstones with the featureless micrites suggests that the micrites were similarly deposited but had subsequently underwent intense diagenesis. The relict representation of cross-stratification by plant material (Text-Fig. 7) in the micrites, the presence of lag deposits on the bottoms of micrite beds, general lens morphology, and an angular "diagenetic" contact between micrite and oncolitic facies (Text-Fig. 12) confirms this supposition.

The diagenetic processes within the featureless micrites may have included cementation, recrystallization, dissolution, and micritization. Water movement through carbonate sediments can promote early lithification (VOGT 1984a). The micrite and microsparite texture (Text-Fig. 5) may indicate dissolution and cementation of voids, recrystallization, or micritization. Micritization is suggested in thin section by the presence of indistinct fossil remains and diffuse grain and void boundaries (Text-Figs. 13, 14). KAHLE (1977) has suggested that clotted textures within Holocene calcareous crusts is due to sparmicritization caused by chemical reactions on crystals through pore water chemistry and bacterial decomposition of fungi or algal material. Blue-green algae, which are found dispersed throughout the micrites, utilize dissolved CO_2 for photosynthesis and precipitate carbonate as a by-product with water as the chemical medium (GOLUBIC 1973). The homogeneous fabric of some of the micrites could also be due to bioturbation (FREYTET 1973, 1984).

The brecciated limestones probably also were originally channel deposits but subsequently altered through gleying and calichification. The marmorization or rubefication features (Text-Figs. 15, 16, 17) represent the presence of periodic hydromorphic soils or pseudogley soils (cf. BUURMAN 1980, FREYTET & PLAZIAT 1982). Immersion of sediments during sedimentation events by water allows the mobilization of iron which then becomes oxidized and fixed upon subaerial exposure (FREYTET 1973, 1984). This gleying process is better catalysed by the presence of microorganisms (VOGT 1984a) with organic matter as the iron source. Also, a perched water table can impregnate sediments during depositional flooding (FREYTET & PLAZIAT 1982). Another possibility for water saturation is accumulative hydromorphy (BOWN & KRAUS 1987) where older pedogenic horizons are buried through sedimentation and eventually move below the water table.

Calichification is defined as a carbonate degradation or erosion mechanism (ESTEBAN & KLAPPA 1983, WRIGHT & WILSON 1987) due to subaerial exposure and as a soil process (calcrete formation) (REEVES 1976, GOUDIE 1983). Many possible phases and textures are possible. Floating grains within a micritic matrix (Text-Fig. 15) are a characteristic but not diagnostic fabric for calichification (ESTEBAN & KLAPPA 1983, GOUDIE 1983). Evidence for calichification processes can also be found as Weald rock fragments and breccias within the oncolitic conglomerates and grainstones (Text-Fig. 11), laminated, pisolithic mudstones (Text-Fig. 21), and cross-bedded sandstones (Text-Fig. 3).

The pisolithic, laminated mudstones subfacies (Text-Figs. 21, 22) is interpreted to be laminar caliche which can also be affected by brecciation processes. Its formation is attributed to carbonate accumulation due to precipitation in a caliche profile (GILE et al. 1965, 1966) or, alternatively, due to algal trapping of detrital and aeolian particles (VOGT 1984b). Both theories acknowledge that a semi-arid to arid climate is needed to produce this type of caliche.

Karstification, as the other end of the spectrum of carbonate erosion processes (ESTEBAN & KLAPPA 1983, CHOQUETTE & JAMES 1988), is also a process that affected the carbonate Weald alluvial floodplain. Evidence entails the presence of elongate holes within the limestone cliff at Pantano which are filled with variegated mudstones (sec. 2.1.2.5). These may represent small dissolutional paleocaves filled with fluvial sediments (FORD 1988).

The variegated mudstones are interpreted to be extrachannel deposits upon the Wealdian alluvial floodplain - probably originally deposited out of suspension as carbonate fines with varying proportions of clastic clays, then subsequently affected by pedogenic processes. Such soils develop on the drier portions of the flood basin and/or where sedimentation rates were lower (BOWN & KRAUS 1987, KRAUS 1987). The branching fine lines (Text-Figs. 18, 19) probably designate root voids. The pieces of aggregate or rock fragments dispersed throughout may represent the remnants of brecciation as percursors to soil development (ESTEBAN & KLAPPA 1983, WRIGHT & WILSON 1987). The "pedogenetic maturity of extrachannel deposits" or the extent of degradation is dependent upon the distance from the active channel margin and local patterns of sedimentation and erosion (KRAUS 1987), assuming constant climatic conditions (BOWN & KRAUS 1987).

The best evidence for soil development is shown in the thin section from a transition rock between the variegated mudstones and brecciated limestones (Text-Fig. 20) which exhibits the dark clay material. These are interpreted to be cutans (BREWER 1976). These are oriented clays which are the product of illuviation processes within soils. They line voids and cracks and also coat aggregate grains (clay skins). Cutans are considered to be diagnostic criteria for the presence of paleosols (YAALON 1971). The presence of clay minerals is common in calcrete (GOUDIE 1983).

In thin section different generations of cements and cutans can be discerned and illustrate the complex exposure and immersion history of a continental carbonate deposit in the Weald. The sedimentation and diagenetic history of terrestrial carbonates is indeed complex and can completely alter the original depositional fabric (FREYTET & PLAZIAT 1982); this makes paleoenvironmental interpretation very difficult. More detailed diagenetic studies have yet to be done.

A modern analog for this paleoenvironment is described by BRANNER (1911) from the semi-arid to subarid limestone plains of Bahia in northeast Brazil. This limestone floodplain contains "soft and marly fresh deposits" and "rock hard" old deposits. The ground surface is pitted with depressions and holes caused by underground dissolution. Temporary shallow lakes form in the wet season and mollusc shells are strewn about. Though BRANNER attributes most of the carbonate to precipitation processes, the ubiquitous presence of "angular fragments and water-worn boulders of all kinds of rocks" suggests also detrital, traction-deposited carbonates.

3.2 Lacustrine to marsh system

While the alluvial plain facies exhibit evidence for periodic subaerial exposure, the featureless micrites at Uña and the shaley mudstones and carbonaceous micrites at other localities represent conditions of continual immersion. These facies represent the lacustrine to marsh system of the Weald around Uña. The shaley mudstones can be a few meters thick and contain the fine fraction of both carbonates and clastics. Most probably these were deposited from suspension in a standing body of water. Compaction has eliminated all sedimentary structures except for one lens of carbonate mudstone containing matrix-supported rock fragments, quartz grains, and bone fragments; it was probably cemented before compaction (sec. 2.1.2.7). This layer is interpreted to be a gravity flow deposit in a small lake on the Weald alluvial floodplain. The matrix-supported components point toward a slump or debris flow in a body of water.

The carbonaceous micrites may represent small depressions or marshy pools on the alluvial plain. The fine lamination suggests that this deposit also was deposited from suspension. The preservation of carbonaceous material could be the result of either relatively quick deposition and/or deposition in stagnant, quiet water. The presence of bioturbation and/or rhizoliths indicates that conditions were not totally anoxic and actually allowed bacterial activity to proceed and prevent the formation of coal.

The featureless micrites at Uña rarely contain the pedogenic and desiccation features found at all other localities. Algal remains are common here along with ostracodes and charophytes. The sedimentary structures suggest fluvial deposition; however, the preservation of lignitic coal points toward a waterlogged paleoenvironment (TEICHMÜLLER & TEICHMÜLLER 1982). The presence of carbonized rhizoliths (Text-Fig. 8) with minimal calichification features and the lack of hydromorphic soil development due to an oscillating water table probably indicate a continually marshy paleoenvironment with some rooted plants. Also, the relatively minor number of rhizoliths may imply a high rate of sedimentation with limited time between depositional events for soil or caliche development.

With all the above evidence, a marshy, deltaic deposit at the edge of a larger lake is postulated for the carbonates containing the coal at Uña.

Rapid deposition coupled with subsidence could be the mechanism by which coal could be preserved (AYERS & KAISER 1984, LI SITIAN et al. 1984). A modern analog can be found in a temperate marl lake in North America (TREESE & WILKINSON 1982). Allochthonous plant fragments are being deposited as proximal lake delta deposits at the mouth of the inflowing creek which drains a vegetated marsh. Peat preservation appears to be linked to burial due to delta progradation.

Deep lake deposits are not well represented at Uña, except perhaps at the Clastic/Carbonate Facies contact where some laminated claystones and carbonaceous fine micrites can be found. However, the presence of a diverse vertebrate fauna and the 25 m thick succession of trough cross-stratified carbonate units could indicate the presence of a large lake nearby, possibly eroded away from within the adjacent river valley.

Coal depositional models are used to explain and to predict the location of deposits. According to CABRERA & SAEZ (1987), coal formation is linked to tectonosedimentary and climatic factors, which basically determine groundwater levels within a basin. FIELDING (1987) proposes that the subsidence regime and, less important, sediment supply are the main factors in coal distribution, not sedimentary conditions. The coal distribution at Uña may suggest that subsidence and sediment supply are the main factors in lignite preservation in this Wealdian basin. A high subsidence rate is suggested by the relatively thick section at Uña and a high sediment supply by the sedimentary features. Carbonaceous micrites are found on the alluvial plain surrounding Uña, but coal could not develop because of oxic conditions conducive to plant decay. According to TEICHMÜLLER & TEICHMÜLLER (1982), low acidity or basic conditions, as would be expected in carbonate-rich waters, enhances bacterial activity and leads to extreme plant decomposition. Rapid burial coupled with subsidence appears to have been necessary to preserve plant material for coal formation in the carbonates at Uña.

3.3 Climatic evidence

Besides subsidence, sediment supply, and basin hydrology (groundwater levels) as controlling factors in coal deposition and formation, CABRERA & SAEZ (1987) also suggest climate as an important variable. They link water table levels to climatic conditions. This does not appear to be the case for the coals at Uña, assuming that all the measured sections were deposited at about the same time. The presence of permanent lakes on the alluvial plain and the immersion conditions at Uña argue for "wet" conditions, while the evidence for exposure and immersion within the fluvial channels point toward alternating "wet" and "dry" phases. Climatic evidence may help to solve this dilemma.

RAT (1982) believes that the climate during the Cretaceous in Spain was generally hot and humid due to paleogeographic and sedimentary evidence. He concedes, however, that the climate during Wealdian deposition could have been drier and more contrasting. KEMPER (1983, 1987) theorizes that the Cretaceous climate in the northern latitudes had cold and warm cycles. A warm cycle is postulated for the Hauterivian to Lower Aptian. ZIEGLER et al. (1983) place Uña at about 30° N during the early Cretaceous, which is at a dry, divergent subtropical zone, according to the climatic zone patterns of today. However, ZIEGLER et al. (1987) postulate that the early Cretaceous divergence/convergence climatic zones were different due to weaker polar fronts so that seasonal climate (wet and dry periods) probably occurred in the tropical to subtropical zones. Therefore, the deposits of Uña may have been deposited in a generally warm and semi-arid climate with seasons.

The basin could have been hydrologically open allowing higher groundwater levels and smaller fluctuations in Uña lake levels even during dry times. The decreased fluvial input during dry periods into the basin would allow various parts of the alluvial plain to undergo pedogenesis, calichification, and karstification to different degrees. Increased fluvial input would allow gleying, cementation, micritization, etc. to further alter the alluvial deposits. This possibility of reduced fluvial input with no sedimentation punctuated by high fluvial input with substantial sedimentation can explain many of the Wealdian sedimentary and diagenetic distribution patterns. This is, of course, assuming that the Weald deposits in the study area are diachronous.

Coal, however, requires sufficient and constant rainfall (ZIEGLER et al. 1987) in order to be preserved. Or, in other words, it needs to be constantly covered by water (TEICHMÜLLER & TEICHMÜLLER 1982). Though the Lower Cretaceous climate is characterized as warm and semi-arid but seasonal, this does not fit into classic warm and wet coal depositional models. Climate may not be an important factor in coal formation and preservation; subsidence, sediment supply, and basin hydrology appear to be the controlling factors in the area around Uña during the early Cretaceous.

4. Summary

(1) The Weald exposed around Uña (Cuenca Province, Spain) is characterized by clastics and carbonates. A brown coal or lignite deposit is found interbedded with the limestones at Uña. The age of the deposits has been determined to be late Hauterivian to early Barremian.

(2) Two main sedimentary facies were recognized within the Weald of Uña and its surrounding area: the Clastic Facies and the Carbonate Facies. These facies were divided into subfacies. The Clastic Facies contained (1) sandy conglomerates; (2) cross-bedded sandstones; (3) cross-laminated sandstones and siltstones; (4) coaly shales, and (5) mudstones. The Carbonate Facies comprised (6) interbedded coal and limestone; (7) oncolitic conglomerates and grainstones; (8) shaley mudstones; (9) featureless micrites; (10) brecciated limestones; (11) variegated mudstones; and (12) laminated, pisolithic mudstones.

(3) The environment of deposition was interpreted to be a continental basin with perhaps clastic proximal deposits and distal carbonate deposits. A mainly carbonate alluvial plain was envisioned with fluvial channels and extrachannel deposits of well-drained soils and small marshes and lakes. The coal deposit at Uña is postulated to have formed in a marsh, deltaic environment at the edge of a large lake.

(4) Diagenetic and sedimentary features of the alluvial plain carbonates indicated that pedogenesis, calichification, gleying, karstification, cementation, micritization, recrystallization, dissolution, and bioturbation processes were at work during the Weald. An oscillating water table due to seasonal weather may have promoted this diagenetic and sedimentary paleoenvironment.

(5) The climate of the Weald around Uña is characterized as warm and semi-arid, but seasonal. Coal formation and preservational models do not fit into this proposed climatic model at Uña. Other controlling factors such as subsidence, sediment supply, and basin hydrology were named as important in aiding the accumulation and burial of sufficient plant material and in maintaining the necessary submersed conditions for coal formation within the carbonates of the Weald at Uña.

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Middle Cretaceous (Upper Albian – Turonian) in the Central Sector of the Iberian Ranges (Spain)*

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With 5 Text-Figures and 1 Table

CARENAS, B., GARCIA, A., CALONGE, A., PEREZ, P. & SEGURA, M. (1989): Middle Cretaceous (Upper Albian - Turonian) in the Central Sector of the Iberian Ranges (Spain). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 265-279. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Two main sedimentary stages may be established in the central sector of the Iberian Range during the Middle Cretaceous. The first one (Upper Albian - Middle Cenomanian) represents a shallow shelf environment open to the Tethys. The second one (Upper Cenomanian - Turonian) represents a pelagic shelf setting becoming progressively shallower during the Turonian and being open to the Proto-Atlantic. These sedimentary stages are separated by stratigraphic disconformities related with Middle-Upper Albian, Middle-Upper Cenomanian and earliest Senonian eustatic events. The occurrence of active fractures conditioned the variations in thickness and the distribution of different facies through tectonic activity. This activity decreased during the Middle Cretaceous producing a lower subsidence rate and differential interaction between the different blocks.

Resumen: Durante la sedimentación del Cretácico medio en el Sector Central de las Cadenas Ibéricas se pueden establecer dos grandes etapas sedimentarias. La primera, de edad Albense superior - Cenomanense medio, representa una plataforma somera abierta al Tethys. La segunda, de edad Cenomanense superior - Turonense, representa la sedimentación de una plataforma pelágica abierta hacia el Proto-Atlántico que progresivamente se hace más somera. Estas dos etapas sedimentarias están limitadas por discontinuidades estratigráficas, en relación con eventos eustáticos del Albense medio-superior, Cenomanense medio-superior y de la base del Senonense. Se ha detectado también la existencia de fracturas activas que condicionan el espesor y la distribución de facies, fruto de una actividad tectónica semejante a la descrita en el Cretácico inferior y que a lo largo del Cretácico medio disminuye, siendo menor la subsidencia y el juego diferencial de los distintos bloques.

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1. Introduction

The Mesozoic sediments which constitute the Eastern Iberian Peninsula were deposited in four ranges with distinctive features (Text-Fig. 1) and folded during the Alpine Orogeny (Betics, Iberian Ranges, Pyrenees and Catalanides). The two main continental basins left in between them are being filled at present with Cenozoic deposits.

Among the different Alpine orogenic belts, the Iberian Ranges have been considered as an intermediate range located within the Iberian plate and bordered by the Iberian and Ebro massifs. The former is made up of Paleozoic material, and the latter is buried at present by the Tertiary sediments that fill this basin.

ALVARO et al. (1979) considered this basin as an aulacogen, as defined by HOFFMAN et al. (1974) and identified the following stages: a graben stage during the Lower-Middle Triassic, a flexural stage throughout the Jurassic and Cretaceous and a Cenozoic tectogenetic stage. This pattern was later modified by VILAS et al. (1983) pointing out two individual evolutionary stages of graben and flexure formation. The first one comprising the Triassic (graben) and Jurassic (flexural), and the second one comprising the Lower Cretaceous (graben) and the Middle-Upper Cretaceous (flexural).

In this part of the Iberian plate, Mesozoic sedimentation took place in continental or shallow shelf environments along with frequent interruptions. Full marine conditions were achieved only during some Jurassic and Upper Cretaceous events.

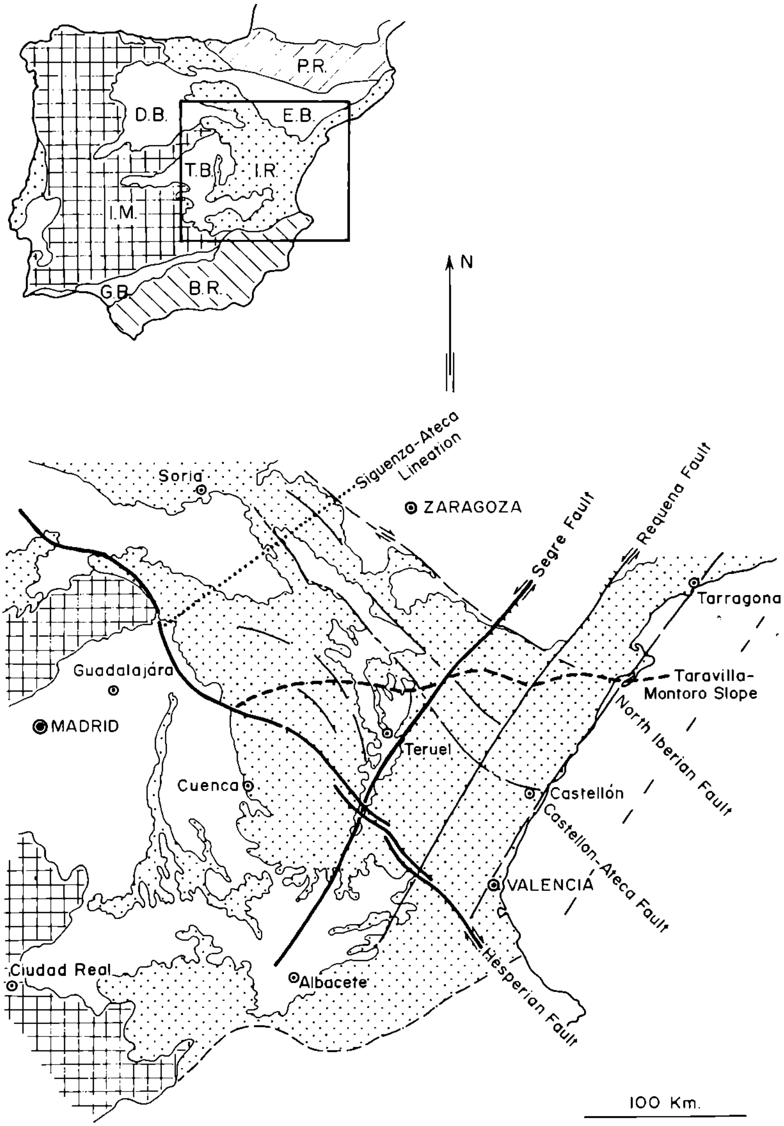
The thickness of the deposited materials and the distribution of the different sedimentary environments seemed to be conditioned by the relative assemblage of the basement blocks which are bounded by late-Hercynian fractures that were active in Mesozoic time.

The Cretaceous of the Iberian Range is formed by six sedimentary cycles. Two belong to the Lower Cretaceous (Hauterivian-Barremian and Upper Barremian-Lower Albian; MAS et al. 1982), another two belong to the Middle Cretaceous (Upper Albian-Middle Cenomanian and Upper Cenomanian-Turonian; GARCIA et al., in press) and the last two belong to the Upper Cretaceous (Coniacian-Santonian and Santonian-Maastrichtian; FLOQUET 1982). Each cycle extended more widely across the Iberian plate being each time less related with the tectonic activity.

The features and distribution of the Middle Cretaceous cycles in the central and southwestern area of the Iberian Range are thoroughly studied in this paper.

2. The Upper Albian-Middle Cenomanian cycle

An important sedimentary interruption caused by a regression - comprising at least part of the Lower and Middle Albian - is followed by a new



Text-Fig. 1. Studied area position and localization of the principal tectonic elements that condition the sedimentation during Middle Cenomanian: I.M. - Iberian Massif (Hercynian Domain); B.R. - Betic Ranges (Alpine building); P.R. - Pyrenees Range (Secondary Alpine Chain); I.R. - Iberian Range; E.B.- Ebro Cenozoic Basin; D.B. - Duero Basin; T.B. - Tajo Basin; G.B. - Guadalquivir Cenozoic Basin.

Tethys-derived transgression. Sedimentation then recommenced in this area of the Iberian Range. A great cycle that comprises Upper Albian to Middle Cenomanian begins with this transgression and ends with an important new regression that probably took place partly during Middle and Upper Cenomanian.

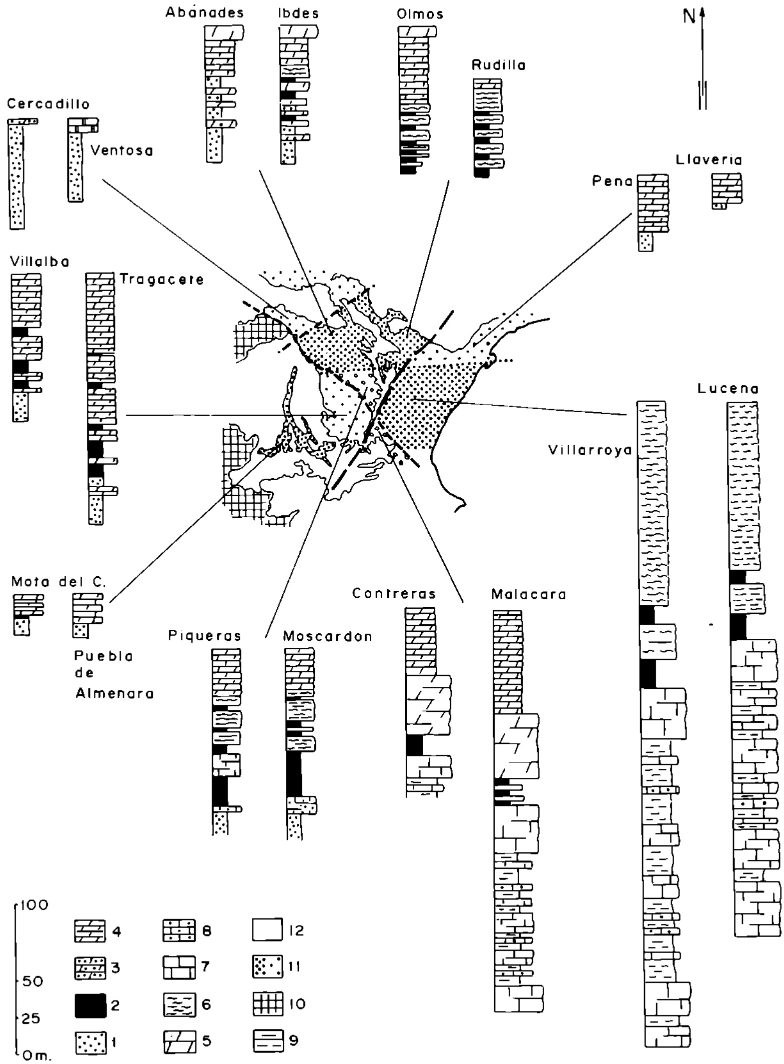
This complex transgressive-regressive cycle (MAS et al. 1982) is composed of several consecutive and more extensive pulses (GARCIA et al., in press). In this cycle, the Tethyan sea limit advanced to the west covering the eastern margin of the Iberian Massif and producing, in this sector of the Iberian Range, the occurrence of shallow marine environments (tidal flats, lagoons, shallow shelves), which evolved from mixed ones at the bottom to carbonate ones at the top. Most of the marine facies (represented by shallow marine shelf deposits) are Vraconian and Lower Cenomanian in age.

This cycle shows homogeneous deposits in the whole basin composed of sands and sandstones at the base ("Arenas de Utrillas" Formation and "Calizas, Margas y Areniscas de Sacaras" Formation, Table 1) passing to the top and to the East into an alternation of marls, limestones and dolomites ("Calizas de Aras de Alpuente" Formation, "Margas de Chera" Formation and "Dolomías de Alatoz" Formation or "Calizas y Margas de Mosqueruela" Formation, Table 1). These, in turn, passed upsection (Middle Cenomanian) into a finely bedded dolomitic facies body with small, locally interlayered marl and limestone levels ("Dolomías tableadas de Villa de Vês" Formation or "Calizas Dolomíticas de Nuêvalos" Formation, Table 1, in GARCIA et al. 1982).

These deposits present a complex facies distribution when studied in detail. As a result, different succession types as well as a complex stratigraphy developed in each of the Iberian Range areas (Text-Fig. 2):

Table 1. Lithostratigraphic units of Mid-Cretaceous (Upper Albian-Turonian) in the Iberian Ranges.

CENTRAL SECTOR	ARAGONESE BRANCHE	CASTILIAN BRANCHE AND VALENCIA RIDGE	MAESTRAZGO RIDGE
Bioclastic Limestone of Muñecos Formation	Marls and Limestones of Alarcón Formation		Dolomites of Barranco de los Degollados Formation
Marls of Picofrentes Fm.	Bioclastic Limestones of Jarabo Formation	Dolomites of the Ciudad Encantada	
	Nodular Limestones of Monterde Fm.	Limestones and Marls of Casas Medina Formation	
Sands, Marls and Limestones	Dolomitic Limestones of Nuévalos Fm.	Thin-bedded Dolomites of Villa de Ves Formation	
	of Sta. Maria de Hoyos Fm.	Dolomites of Alatoz Fm.	Limestones and Marls of Mosqueruela Formation
Sands of		Marls of Chera Formation	
	Utrillas	Limestones of Aras de Alpuente Formation	
	Formation	Limestones, Marls and Sandstones of sa'caras Formation	Sandstone of Maestrazgo Formation



Text-Fig. 2. Most indicative stratigraphic successions for each paleogeographic domain of the Upper Albian-Middle Cenomanian: 1. Terrigenous facies; 2. Green marls; 3. Sandy dolomites; 4. Thin-bedded dolomites; 5. Thick-bedded dolomites; 6. Marls and limestones; 7. Limestones; 8. Sandy limestones; 9. Limestones and marls; 10. Hercynian Massif; 11. Different paleogeographic domains; 12. Cenozoic depressions.

- in the Maestrazgo, more complete and thicker series have been found comprising Upper Albian to Middle Cenomanian (Lucena and Villarroya profiles, PEREZ DEL CAMPO, Ph.D. Thesis, in course). This area with well developed Upper Albian and Vraconian shows open marine facies within the overall carbonate sediments. The Middle Cenomanian displays either lagoonal or shallow carbonate shelf facies.

- in the Valencia region the series also belong to the Upper Albian-Middle Cenomanian, but they are less thick and shallow marine. They are made up of Upper Albian mixed sandstones, marls and sandy limestones which are succeeded by Middle Cenomanian dolomitic sediments deposited in carbonate tidal flats (Malacara and Contreras profiles).

The Upper Albian deposits disappear to the west of the Segre Fault (Text-Fig. 1), leaving only the Vraconian in marginal facies or as continental deposits (Utrillas Facies). Thus:

- an alternation of marls and dolomitic limestone levels belonging to the Lower and part of the Middle Cenomanian overlie the Vraconian terrigenous facies of the "Montes Universales" region. They represent mixed tidal flat and carbonate lagoon deposits, underlying a well-bedded dolomitic series (Middle Cenomanian) related to carbonate tidal flats (Piqueras and Moscardon profiles, SEGURA et al. 1983).

The lagoonal facies are distinctive of the basal Middle Cenomanian and disappear west of the Hespérica Fault (Text-Fig. 1). Thus:

- a marly and dolomitic bedded series was deposited in mixed or carbonate tidal flats which existed during the whole Lower-Middle Cenomanian (Villalba and Tragacete profiles, GARCIA et al. 1984). These overlie the Vraconian terrigenous facies in the "Serranía de Cuenca" region.

- the Lower and Middle Cenomanian dolomitic deposits of the La Mancha region, which more to the west, pass into sands and thus only a few meters of these deposits are left in the Puebla de Almenara and Mota del Cuervo profiles.

A similar situation may be encountered towards the north when passing the Siguenza-Ateca lineament where:

- the Lower and Middle Cenomanian of the Siguenza region is represented by sands of the Utrillas Facies deposited in fluvial or coastal environments (Cercadillo and Ventosa profiles, SEGURA et al. 1985). Only a small calcareous level remains at the Ventosa profile and seems to correspond to the Cantabrian basinal facies via a connection established in the last stretches of this series.

The Taravilla-Montoro slope, situated in the northern part of the studied area, was only covered by sediments after early Cenomanian (SEGURA & GARCIA 1985). This left:

- a terrigenous basal unit (Utrillas Facies sands) deposited in continental or coastal environments, and consisting of alternating limestone or dolomite and marls, and a dolomitic and limestone series found at the top. All of them were deposited in very shallow environments in the Molina de Aragón region (Ibdes profile). To the west (Abanades profile), a similar series may be observed in which the marly levels pass into sands.

- a basal unit formed by alternating dolomitic and marly limestone underlies a dolomitic and bedded dolomitic limestone series deposited in lagoons of mixed or carbonate tidal flat environments. The thickness decreases towards the northeast in the Aragonesa branch of the Iberian Range (Olmos and Rudilla profiles).

- a very small terrigenous basal unit (Utrillas Facies) was deposited in shallow marine or continental environments and outcrops in the lower Aragón and the Catalánides border. An upper unit follows being composed of bedded dolomites and deposited in carbonate tidal flats. They probably represent only Middle Cenomanian, as the Cenomanian itself is even thinner (Pena and Llavería profiles).

Three lower rank transgressive sedimentary cycles may be recognized from the detailed analysis of these sedimentary sequences exhibiting marls or sands at the bottom and dolomites at the top. Their ages were determined by means of benthic foraminifera (conjointly studied with Prof. R. SCHROEDER, Frankfurt) as Upper Albian, Vraconian, Lower and Middle Cenomanian, respectively.

Each of these three sedimentary cycles is progressively more extensive than the previous one. Even lower rank cycles may be noticed in some units of these cycles (CARENAS et al., in press). These are formed by terrigenous deposits at the bottom and limestones or dolomites at the top and are explained as lower order transgressive pulses.

3. Intra-Cenomanian disconformity

The Upper Albian-Middle Cenomanian cycle ends in this whole sector of the Iberian Range with an important stratigraphic disconformity, the identification of which has been difficult because it is located within carbonate sediments and it yields neither a sharp disconformity nor a clear erosive mark. Yet, the occurrence of paleofractures (AZNAR et al. 1983) or gentle erosive unconformities has been described locally (GARCIA et al., in press).

This disconformity was first outlined by biostratigraphic data (VILLENA & RAMIREZ 1975, MELENDEZ et al. 1975). It was then identified both in the lithological record as an unconformity (ALONSO & MAS 1981, SEGURA 1982) and in the basin evolution (GARCIA et al. 1985, GARCIA et al., in press).

When accurately studied, the unconformity is always seen as an irregular contact still difficult to be traced since the strata is often dolomitized. In several places, the uppermost deposits seem to fill underlying cavities (paleokarsts?) or show gentle unconformity. Sometimes local erosive disconformities are created (SEGURA & GARCIA 1984) or (in some profiles of some correlation panels) even the underlying upper member is absent thus involving the presence of an intra-Cenomanian erosion (GARCIA et al., in press).

This stratigraphic disconformity also suggests an important sedimentary and paleogeographic basin change in this sector of the Range. During the Upper Cenomanian-Turonian cycle sedimentation was first related with the Tethyan basin and then with the proto-Atlantic basin by developing an extended furrow opened to the NW (SEGURA & GARCIA 1984). This change in sedimentary basin configuration also requires important paleogeographic variations: the shallowest sedimentary areas, which were situated to the NW of the studied area and even emerged during the Tethyan cycle, shifted to the SE during proto-Atlantic sedimentation.

The age of the disconformity has not yet been determined accurately because: (1) the immediately overlying sediments contain Upper Cenomanian alveolinids (CALONGE, A., Ph.D. Thesis in course), (2) the underlying sediments are always dolomitized at the top making it impossible to date them, and (3) Middle Cenomanian alveolinids have been identified below the latter. This disconformity is therefore located between the Middle and the Upper Cenomanian, comprising parts of either one epoch.

4. The Upper Cenomanian-Turonian cycle

After an important and abrupt intra-Cenomanian regression a new and rapid proto-Atlantic derived transgression flooded the Iberian furrow from NW to SE leading to a new sedimentary cycle in this sector of the Iberian Range.

In this transgressive-regressive cycle the deposits corresponding to the regressive stage predominated. These are formed by:

- nodular limestones and marls with plenty of echinoids, pelecypods and ammonites that have allowed the dating of these deposits as Upper Cenomanian-Lower Turonian (WIEDMANN 1975, MOJICA & WIEDMANN 1977, SEGURA & WIEDMANN 1982) and, among others, they constitute in some zones the "Margas de Picofrentes", "Calizas nodulares de Monterde" and "Calizas y Margas de Casas Medina" formations (Table 1).

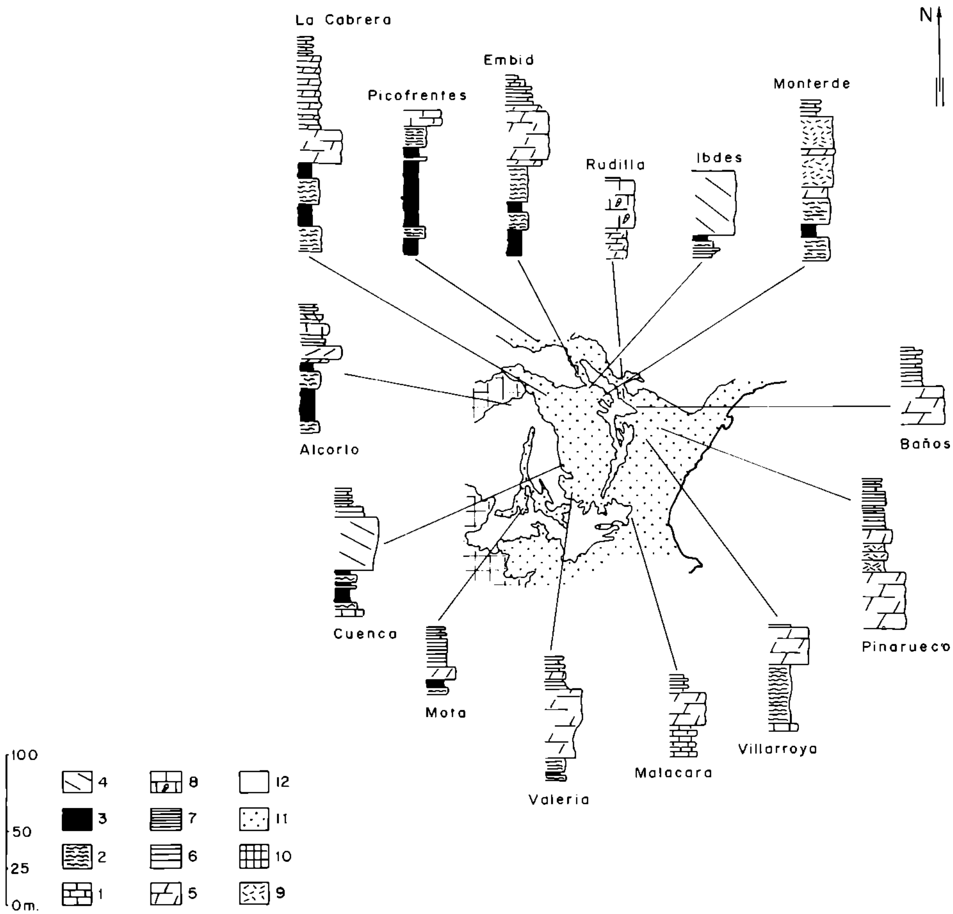
- a thick and rarely stratified dolomitic unit ("Dolomías de la Ciudad Encantada" Formation), in which large-scale cross-bedding may be locally identified showing characteristic weathering features. It is very troublesome to date this unit as it is strongly dolomitized and the results are contradictory: MELENDEZ et al. (1975) dated the top with microfauna as Lower Turonian at the Serranía de Cuenca, and PONS (in FERNANDEZ 1981) dated the top of the unit with rudists as Lower Turonian in the southern sector of the Sierra de Altomira (the same species are considered to be late Cenomanian by BERTHOU & PHILIP 1972). These data apparently disagree with the datations of the underlying units. Disagreement is even more evident if it is considered that the top of the strata, which overlie these dolomites in the southern part of the studied area, is dated as Upper Cenomanian owing to the occurrence of *Crysalidina gradata* (GIMENEZ 1987); even the overlying formation ("Margas de Alarcón") is dated as Upper Cenomanian because it contains *Praealveolina simplex*.

- well stratified decimetric dolomite beds with shadows of algal lamination gradually passing into the "Margas de Alarcón" Formation at the top.

These roughly described lithosomes show an assemblage of features which, considered in detail, change within the studied area from one sector to another (Text-Fig. 3). This assemblage has been identified in most of the marine sediments north of the area studied (Picofrentes profile; WIEDMANN 1979). Gray marls, plenty of ammonites and several nodular limestone (biomicrites) interlayers constitute the base and the top. They are interpreted as clearly marine deposits with an important mass of water over them (outer shelf). This cycle ends with a limestone bed (calcarenite) that is identified as a shallower shelf deposit.

A decrease in marl levels along with an increase in the amount of nodular limestones and a minor decrease in total thickness is noticed to the southeast within the studied area and across the Siguenza-Ateca lineament (Embid, La Cabrera and Alcorlo profiles). A better developed dolomitic unit above the marly levels produces a thick bed in the Embid profile, a thinner bed in the La Cabrera profile and an even thinner bed in the Alcorlo profile. A thick and finely bedded dolomitic series with algal laminations and several limestone (pelmicrites) levels outcrops over this bed which may be considered as deposited in shallow marine and carbonate environments (lagoons, tidal flats, . . .).

Southeast of this lineament (Ibdes section), an important reduction of the marls is noticed at the bottom. It changes to nodular limestone and marls with ammonites. Furthermore, an increase in thickness is observed in



Text-Fig. 3. Most characteristic stratigraphic successions for the Upper Cenomanian-Turonian: 1. Limestones with praealveolines; 2. Nodular limestones; 3. Marls; 4. Dolomites with large-scale cross-bedding; 5. Thick-bedded dolomites; 6. Marls and well stratified dolomites; 7. Thin-bedded dolomites; 8. Rudists bearing limestones; 9. Splintery limestones; 10. Hercynian Massif; 11. Alpine Chain; 12. Cenozoic basins.

the dolomitic stretch which displays large-scale cross-bedding and which is interpreted as the prograding deposits of a carbonate shelf.

The Monterde column is somewhat different; a thick, splintery dolomitic series with massive dolomite interlayers outcrops above nodular limestones and marls, and ends with thinly bedded dolomites considered as littoral and shallow shelf deposits.

In the Cuenca section a similar scheme to that of Ibdes is observed. Marls with nodular limestone interlayers occur at the base and are reduced in thickness. These are followed by thick cross-bedded dolomites and bedded dolomites which differ from the Ibdes section.

A similar pattern is encountered towards the south in the "La Mancha" region, but with a pronounced decrease in thickness. Nodular limestones and marls are met at the base, succeeded by cross-bedded dolomites - which may be considered as prograding shelf deposits (GARCIA et al. 1985) - rudist buildups, and well-developed bedded dolomites deposited on a shallow carbonate shelf.

These lithotypes vary towards the south (Valeria and Malacara profiles). First of all, the large-scale cross-bedded dolomites change to very thick-bedded dolomites with massive appearance and which are considered as shallow carbonate shelf deposits (Valeria column). The basal nodular limestones and marls then disappear and are replaced by thick-bedded limestone deposited on a shallow open shelf. Only the thick-bedded, top-forming dolomites remain and are increasingly reduced in thickness, suggesting a shallower sedimentation zone within this sector.

The initial sedimentary pattern changes to the east and northeast where the cross-bedded dolomites, the nodular limestones and the marly bottom segment are replaced by thick-bedded massive dolomites or by rudist-bearing layers. These represent shallow carbonate shelf sedimentation thus suggesting the occurrence of shallower environments in this area.

Thus, to the east, in the Puerto de Villarroya section, thick nodular limestones occur and are overlain by thick-bedded massive dolomites. Yet, this bottom segment is missing at the Pinarueco profile where the massive dolomites rest directly upon the underlying cycle. Splintery dolomites and thin-bedded dolomitic layers may be found topping these deposits in this last column.

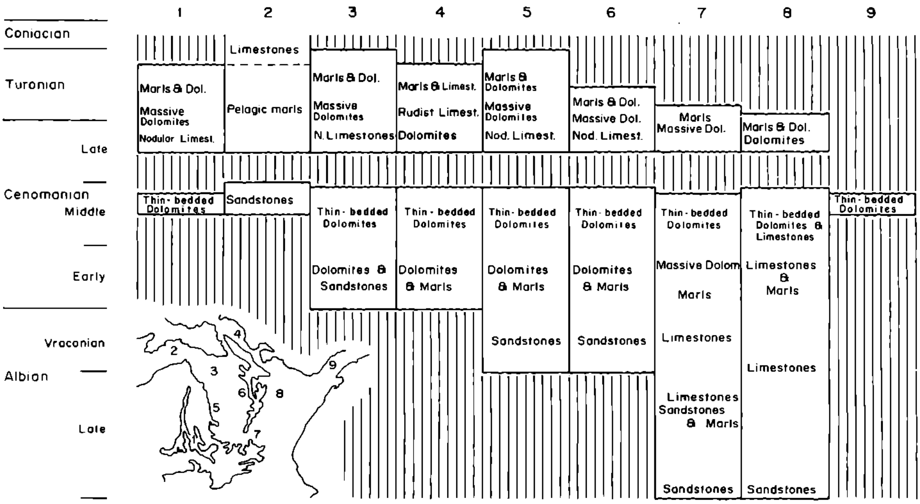
Finally, in the northeast, this scheme is repeated but reduced in thickness in the "Baños de Segura" section where thick-bedded strata is followed by a thin dolomitic unit with massive beds and then by a thin-bedded dolomitic segment. The "Puerto de Rudilla" column is somewhat different, for dolomite levels outcrop at the base and are overlain by rudist limestones and a green marl level revealing paleosoils.

From these data, and with the restrictions imposed by the fact that we are dealing with strongly dolomitized rocks which do not outcrop continuously, a sedimentological model has been rebuilt which handles the different sedimentary environments and their evolution along this cycle (GARCIA et al. 1985). This model can now be precised more accurately (Text-Fig. 4).

For the Upper Cenomanian, an important proto-Atlantic derived transgression - the features of which strongly differed from those of the preceding Tethyan ones - took place in this sector of the Iberian Range. The transgression was much faster and set a thick water sheet in this area.

A marl and limestone level deposited in an outer shelf environment (pelagic or pseudopelagic?) was left in northern and central areas during this transgression. Neritic carbonate shelves also developed in the marginal zones represented by rudist-bearing limestones and thick-bedded massive dolomites.

An initially slower but progressively accelerating regression followed (Text-Fig. 5) leading to a prograding shelf on which the neritic deposits



Text-Fig. 4. Temporal distribution of the stratigraphic successions: 1. Altomira Ridge; 2. Indentation zone of the Central and Iberian Ranges; 3. Castilian branch of the Iberian Range; 4. Aragonese branch of the Iberian Range; 5. Cuenca Ridge; 6. Albarracín Ridge; 7. Valencia Ridge; 8. Maestrazgó Ridge; 9. Southern area of the Catalan Range.

rest upon the outer shelf deposits. These, in turn, are overlain by littoral deposits which terminated with a general sedimentary interruption in this sector manifested by the thin overlying marly levels. A flooded shelf developed, as proposed in READ's model (1982), resulting in a sedimentary setting in which the formations are strongly diachronous.

Some of the oblique beddings observed in the large-scale cross-bedded dolomitic lithosome are interpreted as the slope deposits between the inner and outer shelves. A structure developed like the one described by BOSELLINI (1984) for the Triassic of the Dolomitic Alps, although on a smaller scale.

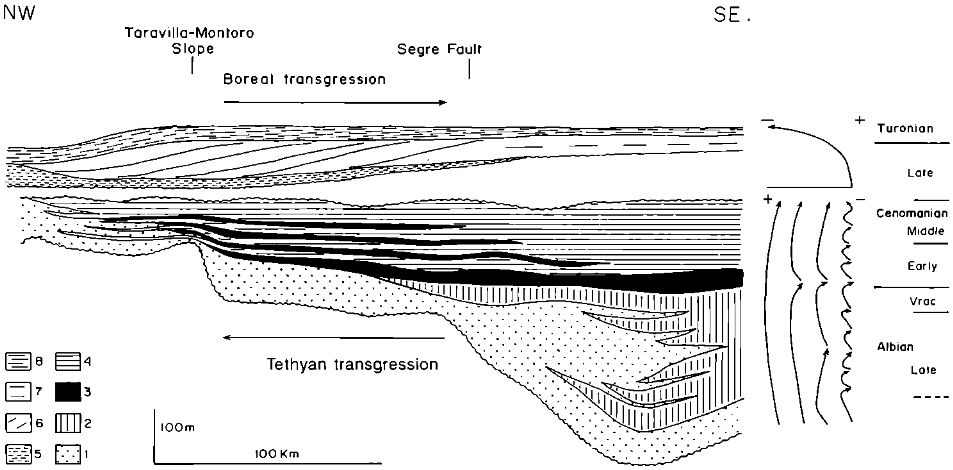
North of the studied area a continuous sedimentation took place in marine pelagic environments paralleling this structural development as seen in the Picofrentes section. This led to open sea conditions (proto-Atlantic) in this sector of the Iberian Range.

5. Eustatism

These sea-level variations have an eustatic origin relying on the present available data (GARCIA et al., in press). The lower rank cycles may also have been caused in this way.

The most important Middle Cretaceous eustatic event is the intra-Cenomanian regression.

It is an eustatically caused regression at which an important sea-level fall took place followed by a rapid transgression during which the sea water level reached its Cretaceous maximum. This led to open sea conditions



Text-Fig. 5. Sedimentary cycles and stratigraphic architecture of the Upper Albian, Middle Cenomanian and Upper Cenomanian-Turonian: 1. Terrigenous facies; 2. Limestones; 3. Green marls; 4. Dolomites and limestones; 5. Marls and nodular limestones; 6. Massive dolomites; 7. Thick-bedded dolomites; 8. Marls and well stratified dolomites.

(proto-Atlantic) in this sector of the Iberian Range. In several zones characteristic deep shelf deposits are separated only by a few centimetres from the tidal flat sediments deposited in the Tethyan sea. This rapid and nearly catastrophic rise of the sea-level concurs, because of its features and age, with other eustatic events described for other areas. We are currently studying the wideness of this event.

The Upper Albian-Middle Cenomanian cycle corresponds to a complex transgression during which a general sea-level rise is interrupted by regressive intervals. This generates a complex curve of sea-level oscillations for which the peaks are produced by stabilization or fall intervals at that level (Text-Fig. 5). This eustatic rise is more important in the beginning (Upper Albian cycle) and is being progressively weakened with each pulse (Vraconian, Lower Cenomanian and Middle Cenomanian) thus decreasing the curve slope up to the top.

Accurate analysis of the sedimentary record shows the presence of other second order cycles, which are weaker as we approach the Upper Cenomanian (GARCIA et al., in press). They are also equally reflected in the curve as lower rank oscillations, yielding a very complex curve which suggests more than one cause for the origin of these eustatic oscillations.

As was pointed out before, the Upper Cenomanian-Turonian cycle shows a rapid sea-level rise at its bottom followed by a progressively speeded up sea-level fall which may be interrupted by momentary sea-level rises as deduced from the smaller cycles that seem to be present in these deposits (Text-Fig. 5).

6. Tectonics

The occurrence of active fractures may be predicted from the comparison of the sedimentary record in different places (Text-Fig. 1). These fractures condition the distribution of sedimentary depositional environments within this sector of the Iberian Range during the Middle Cretaceous and they are reflected in the Alpine structure.

A NE-SW trending fracture set occurs in the east of the area (Segre fault) bounding an open graben to the Tethys that probably formed the western margin of a furrow, the opposite edge of which could be the Corcega-Sardia massif. The best known of these fractures is the Segre fault which limits the penetration of the Upper Albian cycle to the west causing the development of thick series over its eastern block.

The Hesperica fault, over which deposits of variable thickness developed, belongs to a group of faults with Iberian direction (NW-SE). This fault separates an area with Cenomanian deposits of very homogeneous facies from an eastern area with more varied facies.

Another tectonic element which conditions the sedimentation is the Taravilla-Morrón E-W trending lineament, although it is not reflected in the present structure of the Iberian Range. The occurrence of another fracture with the same direction, but located north of this paleogeographic element (Noribérica Fracture) has only been mentioned by VIALLARD (1973) and CANEROT (1974). This last fracture would limit the penetration of the Upper Albian and Vraconian cycles in the north, being surpassed in the Lower Cenomanian.

Studying its general behaviour, the occurrence of a synsedimentary tectonic activity may be suggested. Although it is less important than the Lower Cretaceous one, this activity is important as it produces uplifts and depressions that, in general, are the same as those described for the Lower Cretaceous. It may also be noticed that this activity decelerates during Upper Albian and Cenomanian, yielding laxer facies changes and less significant reductions in thickness.

Finally, it may be noted that throughout the Upper Albian-Middle Cenomanian the northward shift of the depositional centres - probably due to a tilting of the Iberian plate in the same direction - preludes a paleogeographical change that would be produced when the sedimentation resumed in the Upper Cenomanian-Turonian cycle.

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The Sardinian Early Cretaceous Bivalves and their Paleobiogeographic Affinities

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With 5 Text-Figures and 2 Tables

DHONDT, A. V. & DIENI, I. (1989): The Sardinian Early Cretaceous Bivalves and their Paleobiogeographic Affinities. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 281-297. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract. The paper presents the first inventory of bivalve species from the Lower Cretaceous deposits of Eastern Sardinia. A total of 73 species belonging to 49 genera have been identified. Five chronologically subsequent faunas are discussed. The most extensive are of latest Valanginian-Lower Hauterivian and of Upper Albian ages. The paleobiogeographic affinities of the former are closest with the northern Tethys margin and with the eastern Paris Basin: it is best described as a "Jura-fauna" (MIDDLEMISS 1979). The affinities of the Albian fauna seem to be closest to temperate faunas of western and central Europe.

Kurzfassung. Erstmals werden die Bivalven-Arten der ostsardinischen Unterkreide vorgestellt. Identifiziert wurden 73 Arten, die 49 Gattungen angehören. Sie lassen sich in fünf Faunen-Vergesellschaftungen gliedern, die einander ablösen. Die reichsten Faunen wurden im obersten Valangin/Unter-Hauterive und im Ober-Alb gefunden. Die erstgenannte Fauna zeigt Beziehungen zum östlichen Pariser Becken und zum Nordrand der Tethys; nach ihrer Verbreitung läßt sie sich am besten mit dem Begriff "Jura-Fauna" umschreiben (MIDDLEMISS 1979). Die Alb-Fauna ist dagegen den west- und zentraleuropäischen Faunen der "warm-temperierten" Zone am ähnlichsten.

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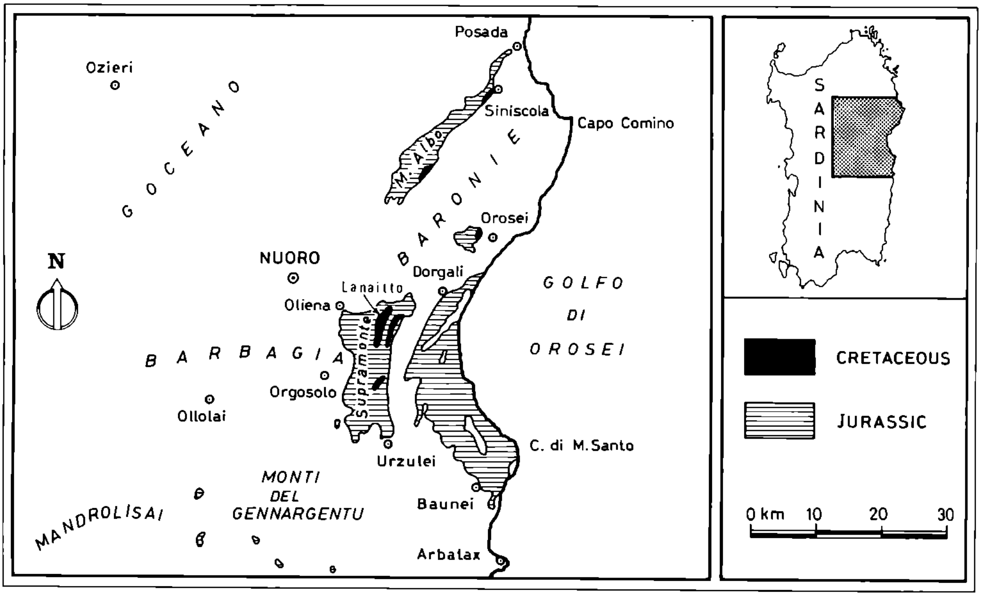
1. Introduction

Systematic geological fieldwork in Eastern Sardinia has resulted in the recognition and study of the Cretaceous formations around the Gulf of Orosei (DIENI & MASSARI 1985, 1987) (Text-Fig. 1).

Biostratigraphic papers on the Lower Cretaceous sequence have been published by DIENI & MASSARI (1963, 1966), WIEDMANN & DIENI (1968), DIENI et al. (1975), AZEMA et al. (1977). The Lower Cretaceous fossils



Text-Fig. 1. Location of Jurassic and Cretaceous formations in central Eastern Sardinia.

collected by DIENI are being studied. Taxonomic paleontological research has so far been completed for Foraminifera (DIENI & MASSARI 1966), ammonites (WIEDMANN & DIENI 1968), brachiopods (DIENI, MIDDLEMISS & OWEN 1975) and hydrozoans (DIENI & TURNSEK 1980). Bivalves also occur frequently in Lower Cretaceous sediments of Eastern Sardinia. Despite this, except for late Cretaceous rudists from NW Sardinia, the only Cretaceous bivalve species previously described from the island were "*Exogyra Couloni* DEFR." and "*Pecten* cf. *crassitesta* ROEM." in DENINGER (1907). It is the aim of the present paper to present a list of the early Cretaceous bivalves collected in Eastern Sardinia and to discuss their paleobiogeographic affinities. A detailed taxonomic inventory is being published elsewhere (DHONDT & DIENI 1988).

2. Bivalve Faunas from Eastern Sardinia

Bivalves are frequent in all Lower Cretaceous sediments of Eastern Sardinia and have been collected in great numbers. In the Lower Hauterivian they are more diverse than any other macrofossil group. In all, 73 species representing 49 genera have so far been identified. It is certain that further fieldwork aimed exclusively at the collecting of fossils would increase the number of taxa found.

The preservation of the specimens is variable. The Berriasian-Hauterivian material is generally preserved as internal or composite moulds, which is the usual state of bivalve fossils of that age in southwest Europe. It makes precise identification sometimes impossible, especially in heterodonts. Albian specimens are fairly well preserved but are rare for some species.

Berriasian-early Valanginian deposits of Eastern Sardinia have only yielded rare and generally poorly preserved bivalves in the Lanaitto area (Oliena) and in the Monte Albo area (Text-Fig. 1). The following taxa have been identified:

Gervillaria cf. *sowerbyana* (MATHERON)
Pterotrigonia sp.
"Lucina" sp.
"Cardium" gillieroni PICTET & CAMPICHE
"Cardium" sp.
Matheronia (Matheronia) rougonensis MONGIN.

Early late Valanginian macrofossils are uncommon and badly preserved. The following bivalve species have been collected in Orosei:

"Pecten" astierianus D'ORBIGNY
Astarte (Astarte) numismalis D'ORBIGNY
Pholadomya gigantea (SOWERBY).

Latest Valanginian-early Hauterivian strata[1] have a wide extension and outcrops have been found everywhere in Eastern Sardinia where the Lower Cretaceous is present. These beds contain numerous bivalves:

Nucula (Leionucula) planata DESHAYES in LEYMERIE
Grammatodon (Nanonavis) securis (LEYMERIE)
Cucullaea (Noramya) gabrielis LEYMERIE
Cucullaea (Noramya) "gresslyi" (DE LORIOL)
Pinna (Pinna) robinaldina D'ORBIGNY
Pinna (Pinna) sulcifera DESHAYES in LEYMERIE
Pinna (Stegoconcha?) hombresi PICTET & CAMPICHE
Gervillaria alaeformis (SOWERBY)
Gervillaria sowerbyana (MATHERON)
Gervillella anceps (DESHAYES in LEYMERIE)
Inoceramus ex gr. *neocomiensis* D'ORBIGNY
Isognomon (Isognomon) ricordeanus (D'ORBIGNY)
Entolium orbiculare (SOWERBY)
Camptonectes cottaldinus (D'ORBIGNY)
Chlamys? archiaciana (D'ORBIGNY)
Chlamys cf. *elongata* (LAMARCK)
Mimachlamys robinaldina (D'ORBIGNY)
Neithea (Neithea) atava (ROEMER)
Neithea (Neithella) valangiensis (PICTET & CAMPICHE)
Plicatula placunaea LAMARCK
Spondylus roemeri DESHAYES in LEYMERIE
Acesta dorbignyana (MATHERON)
Acesta cf. *subrigida* (ROEMER)
Ctenoides? undatus (DESHAYES in LEYMERIE)

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[1] Uppermost Valanginian and Lower Hauterivian occur in the same facies in Eastern Sardinia. Therefore the species from both levels are listed together: generally, species found in latest Valanginian beds also occur in early Hauterivian beds of the same formation.

Limaria? dubisiensis (PICTET & CAMPICHE)
Aetostreon latissimum (LAMARCK)
Ceratostreon boussingaulti (D'ORBIGNY)
Ceratostreon cf. *tuberculiferum* (KOCH & DUNKER)
Rastellum "rectangulare" (ROEMER)
Trigonia (Trigonia) carinata AGASSIZ
Quadratotrigonia nodosa (SOWERBY)
Thetis minor SOWERBY
Sphaera corrugata SOWERBY
Astarte (Astarte) numismalis D'ORBIGNY
Eriphyla gigantea (DESHAYES in LEYMERIE)
Ptychomya (Ptychomya) plana AGASSIZ
Pleuriocardia? voltzii (LEYMERIE)
Protocardia impressa (DESHAYES in LEYMERIE)
Protocardia peregrina (D'ORBIGNY)
Integricardium deshayesianum (DE LORIOL)
Proveniella bernensis (LEYMERIE)
Flaventia brongniartina (LEYMERIE)
"Venus" cornueliana D'ORBIGNY
"Venus" thurmanni DE LORIOL
"Venus" vendoperana (LEYMERIE)
Gastrochaena cf. *sinuosa* PICTET & CAMPICHE
Panopea gurgitis (BRONGNIART in CUVIER)
Pholadomya gigantea (SOWERBY)
Platymyoidea rostrata (AGASSIZ).

Late Aptian and **late Albian** bivalves have been found only in small outcrops in the Orosei area. Late Aptian beds have yielded only two species:

Spondylus gibbosus D'ORBIGNY
"Lucina" valdensis PICTET & CAMPICHE.

Late Albian (including Vraconian) bivalves are represented by the following species:

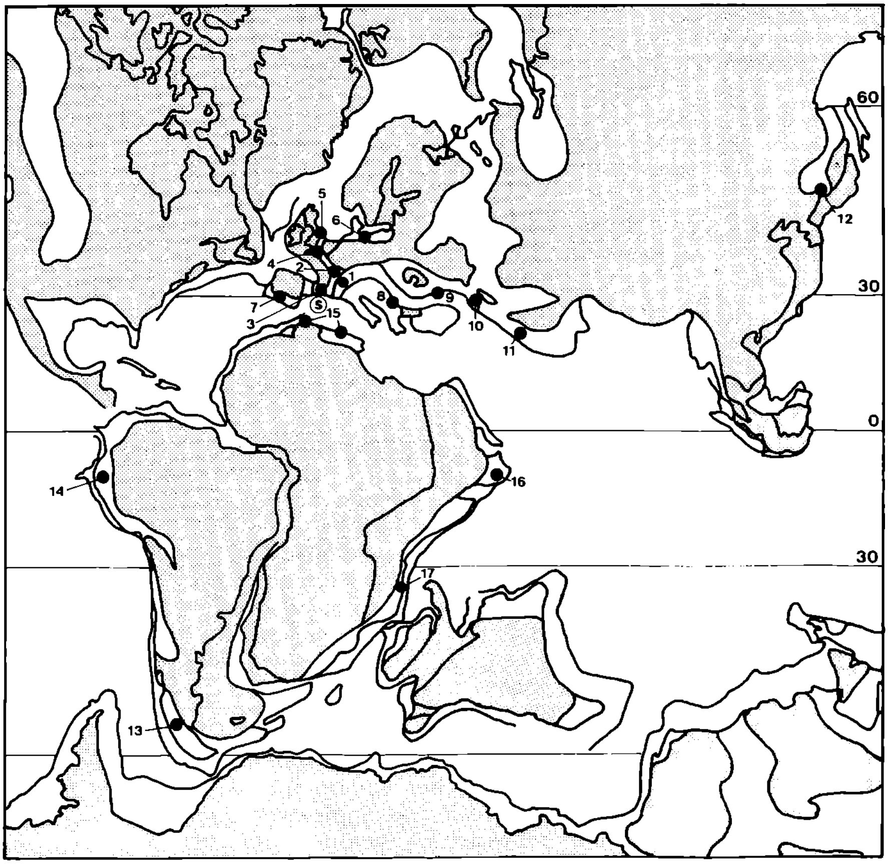
Nuculana (Jupiteria?) phaseolina (MICHELIN)
Cucullaea (Idonearca) glabra PARKINSON
Inoperna flagellifera (FORBES)
Gervillella cf. *sublanceolata* (D'ORBIGNY)
Pseudoptera anomala (SOWERBY)
Birostrina concentrica (PARKINSON)
Birostrina subsulcata (WILTSHIRE)
Birostrina sulcata (PARKINSON)
Propeamussium ninae (KARAKASCH)
Acesta subovalis (SOWERBY)
Limaria? elongata (SOWERBY)
Plagiostoma globosum (SOWERBY)
Pleuriocardia? rauliniana (D'ORBIGNY)
Proveniella crassicornis (AGASSIZ)
Gastrochaena cf. *brevis* PICTET & CAMPICHE.

Late Hauterivian to early Aptian strata are found in Eastern Sardinia in a very hard limestone facies. They contain bivalves but the material extracted from them is too incomplete for identification.

3. Paleobiogeographical Affinities

The geographical distributions of the early Cretaceous bivalve species (listed above) collected in Eastern Sardinia have been analysed based on personal research and published literature. Personal research refers to the critical reinterpretation by A. V. D. of the figured specimens studied in museum collections (list of these collections is added in appendix).

In the Upper Valanginian-Lower Hauterivian and Upper Albian bivalves were numerous and relatively easy to collect from the Eastern Sardinian outcrops of that age. These faunas are extensive and diversified enough to compare them with those of similar age in other areas.



Text-Fig. 2. Map at 120 million years after BARRON et al. (1981); location of the areas from where bivalve faunas are compared with those of latest Valanginian-early Hauterivian age in Sardinia; localities (1-17) are explained on Table 1; S = Sardinia.

Sardinian Species	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
Nucula planata	X	X		X	X	X			X	X							
Grammatodon securis	X	X	X		X	X		X	X	X		X	X		X		
Cucullaea gabrielis	X	X	X			X	X	X	X	X			X	X		X	X
Pinna robinaldina	X	X	X		X	X	X	X	X	X		X	X		X	X	
Pinna sulcifera	X	X						X									
Pinna hombresi	X		?	X		X		X									?
Gervillaria alaeformis	X	X	?	X		X	X	X	X	X	X	X			X	?	X
Gervillaria sowerbyana	X		X			X		X							X	X	?
Gervillella anceps	X	X	X					X	X	X			X				
Inoceramus gr. *neocomiensis*	X	X		X		X		X	X	X	X						
Isognomon ricordeanus	X	X	X	X		X		X	X	X		X	X			X	
Entolium orbiculare		X	X	X		X		X	X		X						
Camptonectes cottaldinus	X	X		X				X	X	X	X						
Chlamys ? archiaciana	X	X	X					X	X	X	X				X		
Chlamys sp. cf. *Chl. elongata*	X	X	X	X		X	X										
Mimachlamys robinaldina	X	X	X	X		X			X	X	X	X				X	
Neithea atava	X	X	X			X	X	X	X	X	X	X			X	X	X
Neithella valangiensis	X		X				X		X	X	X						
Plicatula placunaea	X	X	X			X	X	X	X	X					X		
Spondylus roemeri	X	X	X			X		X	X	X							
Acesta dorbignyana	X	X	X	X				X	X	X	X						
Acesta sp. cf. *A. subrigida*	X	X	X		X	X		X	X								
Ctenoides ? undata	X	X	X			X		X	X	X							
Limaria ? dubisiensis	X	X	?			X			X	X	X						
Aetostreon latissimum	X	X	X	X	X	X	X	?	X	X	X		X		X	X	X
Ceratostreon boussingaulti	X	X	X			X	X	?	X	X	X	?	X	X	X	X	X
Ceratostreon cf. *tuberculiferum*	X	X	X	X		X	X	?	X	X	X						
Rastellum "rectangulare"	X	X	X	X	X	X	X	?	X	X	X				X	X	X
Trigonia carinata	X	X	X	X		X	X	X	X	X	X		X	X	X		
Quadratotrigonia nodosa	X	X	X	X	?	X	X	X	X	X	X		X				
Thetis minor	X	X	?	X		X			X	X	X						
Sphaera corrugata	X	X	X	X			X	X	X	X	X			X	?	X	X
Astarte numismalis	X	X	?	X		X	?	X	X	X	X				X	X	X
Eriphyla gigantea	X	X	X					X	X	X							
Ptychomya plana	X	X	?	X				X	X	X	X	?	X	X	X		X
Integricardium deshayesianum	X					X		X	X	X							
Pleuriocardia ? voltzii	X	X	?			?			?								
Protocardia impressa	X	X	X				X	X	X	X							
Protocardia peregrina	X	X							X	X							
Proveniella bernensis	X	X	X						X	X					X		
Flaventia brongniartina	X	X								X					?		
"Venus" cornueliana	X	X															
"Venus" thurmanni	X																
"Venus" vendoperana	X	X	?				X										
Panopea gurgitis	X	X	X	X		X	X	X	X	X	X	X	X		X	X	X
Pholadomya gigantea	X	X	X	X		X	X	X	X	X	X		X	X	X	X	X
Platymyoidea rostrata	X	X	X			X		X	X	X						X	
Number of Species: 47	46	42	38	21	7	31	19	34	39	37	22	9	12	6	18	15	13
Percentage of common Species	98	89	81	45	15	66	40	72	83	79	47	19	26	13	38	32	28 %

Text-Fig. 3. Latest Valanginian-early Hauterivian bivalves from Eastern Sardinia and their occurrence in other areas; for each locality the number of species in common with Sardinia is indicated, and the percentage of the Sardinian fauna this represents; localities (1-17) are explained on Table 1.

Table 1. Origin of data used in the comparison of Valanginian-Hauterivian faunas.

	Areas	Stratigraphy	Authors	P.R.
(1)	Neuchâtel - Vaud (CH) & Mt. Salève (F)	Valanginian - Hauterivian	AGASSIZ 1840, 1842-45, PICTET & CAMPICHE 1864-1871, DE LORIOL 1861-1863, 1868, etc.	- + +
(2)	"Aube" (F) & Auxerre (Yonne) (F)	'Neocomian'	LEYMERIE 1841-1842, D'ORBIGNY 1844-1847	- +
(3)	Marseille (F)	Valanginian - Hauterivian	MATHERON 1843	+
(4)	Isle of Wight (G. B.)	'Aptian'	WOODS 1899-1913	+
(5)	Lincolnshire & Yorkshire (G. B.)	Volgian - Valanginian	KELLY 1984, WOODS 1899-1913	- +
(6)	Hannover - Braunschweig (D) & NW Westphalia	'Neocomian'	BRAMER 1967, HARBORT 1905, WEERTH 1884, WOLLEMANN 1896, 1900	- + -
(7)	NE Spain	'Aptian'	COQUAND 1865-1866	+
(8)	Bulgaria	'Neocomian' - Aptian	DIMITROVA 1969, 1974	+
(9)	Crimea (USSR)	Neocomian	KARAKASCH 1907, MUROMTSIEVA & JANIN 1964	+ +
(10)	N. Caucasus (USSR)	Neocomian	KARAKASCH 1897, MUROMTSIEVA & JANIN 1964	+ +
	S. Caucasus (USSR)	Neocomian	ERISTAVI 1955, 1957, KOTETISHVILI 1965, 1970	+ +
(11)	'Central Asia' (USSR)	Neocomian	BOGDANOVA 1961, 1966, PROZOROVSKII 1961	+ +
(12)	Japan	Neocomian - Aptian	HAYAMI 1965a, b, 1975	+
(13)	S. Chile & Argentina	'Neocomian'	BURCKHARDT 1903, WEAVER 1931	- -
(14)	Columbia & Peru	'Neocomian'	BENAVIDES-CACERES 1956, D'ORBIGNY 1842a, b	- +
(15)	NW Africa	Neocomian - Aptian	COQUAND 1862, PERVINQUIÈRE 1912	+ -
(16)	Somalia & Ethiopia	'Neocomian'	COX 1935, TAVANI 1948, 1949	+ -
(17)	Tendaguru (Tanzania)	Neocomian - Aptian	DIETRICH 1933, KRENKEL 1910, LANGE 1914, MUELLER 1900	- - + -

(P.R.: personal research)

The fauna from the Upper Valanginian-Lower Hauterivian beds is particularly rich (49 species). Several of these occur over a wide geographical area, often ranging from the W coast of South America to Japan via southern Europe and East Africa. A few examples of such wide distributions are found for: *Pinna (Pinna) robinaldina* D'ORBIGNY, the bakevelliid *Gervillaria alaeformis* (SOWERBY), the pectinid *Neithea (Neithea) atava* (ROEMER), the oyster species *Aetostreon latissimum* (LAMARCK) and *Ceratostreon boussingaulti* (D'ORBIGNY), and the fimbriid *Sphaera corrugata* SOWERBY, and *Pholadomya gigantea* (SOWERBY). Detailed distributions with maps can be found in DHONDT & DIENI (1988).

Faunal comparison of the Eastern Sardinian bivalve species with those from other areas have been attempted. The geographical distribution of these areas is shown on Text-Fig. 2. Table 1 lists the areas and indicates the origin of the faunal data. Text-Fig. 3 lists the late Valanginian-early Hauterivian bivalves from Eastern Sardinia and their occurrence in the areas listed on Table 1. At the bottom of Text-Fig. 3 the percentages of common species between the faunas from these areas are shown. These figures only have a relative value since the data used for comparison are not always completely comparable: in some areas bivalve faunas have not been studied recently or have been studied in less detail than in others. In most instances at least parts of the material listed per locality has been studied by one of us (A. V. D.). From the data available it is obvious that the **latest Valanginian-early Hauterivian fauna from Eastern Sardinia:**

1. shows the highest correlation with bivalve species in strata of the same age in western Switzerland;
2. is very similar to the faunas of the Eastern Paris Basin and of the region around Marseille;
3. and is also similar to the faunas of Crimea, the Caucasus and Bulgaria;
4. has two thirds of its species also occurring in northern Germany;
5. but only a few species (less than 15 %) also occur in northern England;
6. has a distribution extending along the northern Tethys margin as far as Central Asia.

Many of the species extend stratigraphically into the Aptian and have a geographical distribution which is typically Tethys-Temperate, from the southern tip of South America to Japan (illustrated by the localities on Text-Fig. 2). Virtually none of the species has been found in the well preserved and recently studied Siberian and Canadian Neocomian faunas (ZAKHAROV 1966, 1970, etc.). The paleobiogeographic distribution of these latest Valanginian-early Hauterivian bivalves is very similar to that found for other fossil groups from Eastern Sardinia. WIEDMANN & DIENI (1968) have shown that Sardinian ammonites contain no northern elements: they are restricted to the fauna which KOTETISHVILI (1983) designated as "Mediterranean". Brachiopod distribution has been discussed by MIDDLEMISS (1979, 1981): a combination of faunal elements from the extended Paris Basin and from the northern Tethys margin is found in brachiopods, as it is in bivalves. To distinguish this distribution from a cold Temperate or Boreal type (sensu CASEY & RAWSON 1973) and from a typical Tethys distribution, MIDDLEMISS coined the term "**Jura fauna**" which applies very well in both instances. Selected foraminifera have been shown by CHERCHI (1987) to have a similar distribution. For the Valanginian-Hauterivian, the paleobiogeographic concept of MIDDLEMISS seems closer to the faunas than that of KAUFFMAN (1973) in which the Temperate and Tethys faunas are considered as two distinct entities with virtually no transitional elements. On the data available it seems probable that for many groups the "**bipolar distribution**" suggested by CRAME (1986) is, on a global scale, a better concept than the "Boreal" fauna. Bipolar faunas contain some temperate elements especially in the southern hemisphere. Possibly the "**Jura fauna**" which is intermediate between Tethys and Boreal, could also be found in the southern hemisphere and contain those elements which have a very wide geographical distribution in the Neocomian-Aptian interval.

The fauna from the Upper Albian deposits in Eastern Sardinia is less extensive than that from the Upper Valanginian-Lower Hauterivian outcrops (15 species). Text-Fig. 4 lists the geographic distribution of the species. Text-Fig. 5 shows the areas from which faunas have been compared. Table 2 indicates the origin of the data used in these faunal comparisons.

Based on the relatively limited data shown in Text-Fig. 4 the Sardinian Albian bivalves have affinities with Temperate faunas of the same age in the extended Paris Basin, in Germany and in Poland. An extension of these faunas into Moldavia (USSR), Crimea (USSR) and further eastwards is prob-

Sardinian species	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Nuculana phaseolina	X	X	X	X										
Cucullaea glabra	X	X	X	X	X	X		X		X	X	X	X	
Inoperna flagellifera	X	X				X	X	X	X					
Gervillella cf. *lanceolata*	X		X	X	X									X
Pseudoptera anomala	X				X	X								X
Birostrina concentrica	X	X	X	X		X	X	X	X		X	?	X	X
Birostrina subsulcata			X	X	X			X	X	X	X			
Birostrina sulcata	X	X	X	X		X	X	X	X		X	X	X	
Propeamussium ninae											X	X		
Acesta subovalis	X				X	X	X	X		X				
Limaria ? elongata	X	X	X	X		X	X	X			X	X		X
Plagiostoma globosum	X			X		X		X		X		X		
Pleuriocardia ? rauliniana	X		X											
Proveniella crassicornis	X	X	X					X						
Number of Species: 14	12	6	9	8	5	8	5	9	4	4	6	6	3	4
Percentage of common Species	86	43	64	57	36	57	36	64	29	29	43	43	21	29 %

Text-Fig. 4. Late Albian bivalves from Eastern Sardinia and their occurrence in other areas; for each locality the number of species in common with Sardinia is indicated, and the percentage of the Sardinian fauna this represents; localities (1-14) are explained on Table 2.

Table 2. Origin of data used in the comparison of Albian faunas.

	Areas	Stratigraphy	Authors	P.R.
(1)	Paris Basin	Albian	D'ORBIGNY 1844-47	+
(2)	SE France:	Albian	D'ORBIGNY 1844-47,	+
	Escragnolles		PICTET & CAMPICHE 1864-71	+
(3)	W Switzerland	Aptian -	PICTET & ROUX 1847-53,	+
	('Grês verts')	Albian	PICTET & CAMPICHE 1864-71	+
(4)	'Gault' in S. England	Albian	WOODS 1899-1913	+
(5)	'Upper Greensand' in S. England	Albian	WOODS 1899-1913	+
(6)	'Meule' de Bracquegnies (B)	Albian	-	+
(7)	Hannover (D)	Albian	-	+
(8)	Poland	Albian	CIESLINSKI 1960,	-
			KOKOSZYNSKA 1949	-
(9)	Bulgaria	Albian	DIMITROVA 1974	-
(10)	Moldavia (USSR)	Albian - Cenomanian	SOBETZKI 1961, 1977	+
(11)	Crimea (USSR)	Albian	KARAKASCH 1907,	+
			MUROMTSIEVA & JANIN 1964	+
(12)	Caucasus (USSR)	Albian	KARAKASCH 1897,	+
			MUROMTSIEVA & JANIN 1964,	+
			KOTETISHVILI 1977	+
(13)	Transcaucasus &	Albian	KARAKASCH 1897,	+
	Central Asia (USSR)		KOTETISHVILI 1977	+
(14)	N Spain	Albian	-	+

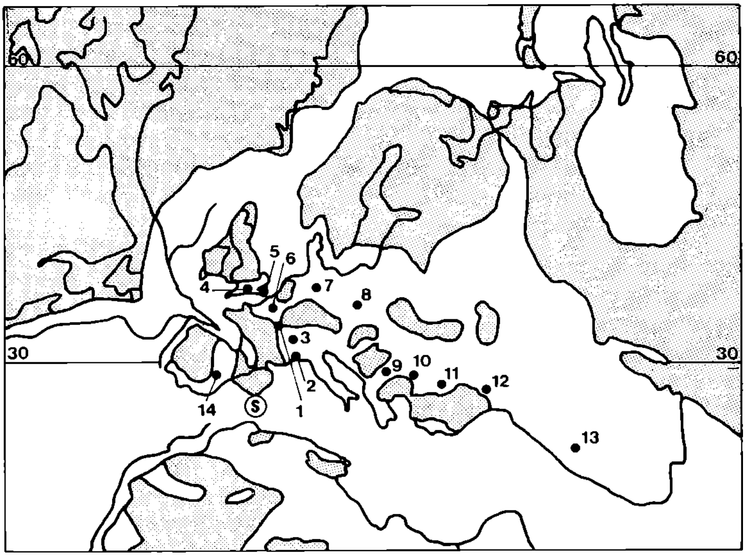
(P.R.: personal research)

able. The small fauna known from Sardinia contains few or no Albian Tethyan bivalves.

4. Conclusions

The first study on the bivalves from Lower Cretaceous deposits in Eastern Sardinia has shown the presence of five faunal assemblages: Berriasian-early Valanginian (6 species), early late Valanginian (3 species), latest Valanginian-early Hauterivian (49 species), late Aptian (2 species), late Albian (15 species). Because of poor preservation identification of species is sometimes only tentative or in open nomenclature. The total number of species identified is 73, belonging to 49 genera, and representing all major taxonomic bivalve groups existing in the Lower Cretaceous.

Attempts were made to discover the paleobiogeographic affinities of two of the Eastern Sardinian bivalve faunas. The presence-absence of the taxa found in Eastern Sardinia was verified for a series of well-known coeval faunas. This verification was based on literature data and on personal study of museum collections and of figured types.



Text-Fig. 5. Map at 100 million years after BARRON et al. (1981); location of the areas from where the faunas are compared with those of late Albian age in Sardinia; localities (1-14) are explained on Table 2; S = Sardinia.

The Sardinian bivalve fauna from the latest Valanginian-early Hauterivian interval has more than 70 % (in one case as many as 90 %) species in common with deposits from western Switzerland, the Paris Basin, Bulgaria, Crimea, the Caucasus, and the area around Marseille. This is considered as indicating a definite biogeographic affinity. The fauna contains species with a Temperate and Tethys margin distribution. For such a fauna the term "**Jura fauna**" of MIDDLEMISS (1979) is most appropriate.

The late Albian bivalve fauna from Eastern Sardinia is smaller. Comparing the presence-absence of its species in coeval faunas has been undertaken, but does not give significant results. The only certainty reached is that this fauna shows affinities with the extended Paris Basin.

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APPENDIX

List of the collections studied and used in the faunal comparisons

Type collections

Author	Museum	P	L	O	o	a
BOGDANOVA 1961, 1966	VSEGEI	+	+	+	+	-
COQUAND 1862, 1865	MAFI	+	+	+	+	+
COX 1935	BM (NH)	+	+	+	+	-
DIMITROVA 1974	U. Sofia	+	+	-	-	-
ERISTAVI 1955, 1957	TGU	+	+	+	+	-
HAYAMI 1965a, b, 1975	U. Kyushu	+	+	+	+	+
	U. Sendai	+	+	+	+	+
	U. Tokyo	+	+	+	+	-
KARAKASCH 1897, 1907	LGU	+	+	+	+	-
KOTETISHVILI 1965, 1970, 1977	TGU	+	+	+	+	-
LANGE 1914	H. M. B.	+	+	-	-	-
DE LORIOL 1861, 1863, 1868	M. Basel	+	+	+	-	-
	M. Genève	+	+	+	-	-
MATHERON 1843	M. Mars.	+	+	+	+	+
MUROMTSIEVA & JANIN 1964	MGU	+	+	+	+	-
D'ORBIGNY 1842a, b, 1844-1847	MHN Paris	+	+	+	+	+
PICTET & CAMPICHE 1864-1871	M. Genève	+	+	+	-	-
	M. Lausanne	+	+	+	-	-
PICTET & ROUX 1847-1853	M. Genève	+	+	+	-	-
	M. Lausanne	+	+	+	-	-
PROZOROVSKII 1961	LGU	+	+	+	+	-
SOBETZKI 1961, 1977	PIN	+	+	+	+	-
WEERTH 1884	Hann.	+	+	+	-	-
WOODS 1899-1913	BM (NH)	+	+	+	+	-

(P: pectinids, L: limids, O: Ostreacea, o: other bivalves, a: all types of that author)

Museum collections

	Coll.	T
Basel: Naturhistorisches Museum	+	+
Berlin: Naturkunde Museum, Humboldt-Univ. (H. M. B.)	+	+
Brussels: Koninklijk Belgisch Instituut voor Natuurwetenschappen, Palaeontology Dept.	+	–
Budapest: Magyar Allami Földtani Intézet (MAFI)	+	+
Cambridge: Sedgwick Museum (SM)	+	+
Fukuoka: Kyushu University, Geology	+	+
Geneva: Muséum d'Histoire naturelle	+	+
Hannover: Niedersächsische Landesanstalt (Hann.)	+	+
Lausanne: Musée géologique	+	+
Leningrad: Leningrad State University (LGU)	-	+
All Union Geological Survey (VSEGEI)	-	+
London: British Museum (Natural History) [BM (NH)]	+	+
Marseille: Musée d'Histoire naturelle (M. Mars.)	+	+
Moscow: Moscow State University, Geological Museum (MGU)	-	+
Palaeontological Institute of the USSR Academy of Sciences (PIN)	+	+
Paris: Muséum national d'Histoire naturelle, Institut de Paléontologie (MNH Paris)	+	+
Sendai: Tohoku University, Dept. Geology and Paleontology	+	+
Sofia: University Kliment Ohridski, Geology	+	+
Tbilisi: Tbilisi State University (TGU)	–	+
Tokyo: The University of Tokyo, University Museum	-	+

(T: type collections, Coll.: general collections)

The Campanian-Maastrichtian Boundary in the El Kef Section, Tunisia

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With 6 Text-Figures and 2 Tables

SALAJ, J. & WIEDMANN, J. (1989): The Campanian-Maastrichtian Boundary in the El Kef Section, Tunisia. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 299-315. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The Campanian-Maastrichtian boundary can now be traced in the El Kef section, Tunisia. Due to the presence of *Pseudokossmaticeras brandti* (REDTENBACHER) in the upper portion of the "marly alternation" between the two "barres calcaires d'Abiod", this part of the section has to be regarded as the base of the Maastrichtian. 160 m below this point, the top of the first "barre calcaire d'Abiod" contains *Bostrychoceras polyplocum* (ROEMER), and is thus late Campanian in age. Therefore, most of the "marly alternation" is considered to belong to the Campanian also, despite the fact that only its lowermost portion has been found to contain *Globotruncanita calcarata* spp. The middle portion is regarded as belonging to the Zone of Globotruncana stephensoni and is also of late Campanian age. For the uppermost portion of the "marly alternation" and the second "barre calcaire d'Abiod", the Zone of Archaeoglobitruncana kefiana has been proposed. This zone may correlate with the Zone of Pseudokossmaticeras brandti as proposed by WIEDMANN (1988). These lower Maastrichtian zones range up to the first appearance of *Globotruncana falsostuarti* SIGAL and *Pachydiscus neubergicus* (REDT.).

Even if the uncommon globotruncanid sequence found in the El Kef section is only of ecological-regional value, the proposed Campanian-Maastrichtian boundary is of importance, since this section was proposed as a hypostratotype for the Campanian and Maastrichtian stages.

Kurzfassung: Im Profil von El Kef/Tunesien kann die Campan/Maastricht-Grenze nun mit einiger Sicherheit gezogen werden. Entscheidend hierfür ist der Nachweis von *Pseudokossmaticeras brandti* (REDTENBACHER) im höheren Teil der "Mergel-Wechselfolge" zwischen den beiden "Kalk-Barren von Abiod". *Ps. brandti* wird als Vertreter des Unter-Maastricht betrachtet. Etwa 160 m tiefer, am Top der "1. Kalk-Barre von Abiod" wurde *Bostrychoceras polyplocum* (ROEMER) gefunden, der Ober-Campan-Alter angibt. Damit muß auch der größere Teil der "Mergel-Wechselfolge" noch ins Campan gestellt werden. Nur der unterste Teil ist allerdings noch durch *Globotruncanita calcarata* spp. charakterisiert. Der mittlere Teil der Wechselfolge wird als Zone der Globotruncana stephensoni ebenfalls noch dem Campan

zugerechnet. Für den höchsten Teil der Wechselfolge und die "2. Kalk-Barre von Abiod" kann eine Zone der Archaeoglobitruncana kefiana ausgeschieden und mit der Ammoniten-Zone des *Pseudokossmaticeras brandti* korreliert werden (WIEDMANN 1988). Diese Zonen entsprechen damit Unterem Maastricht und reichen bis zum Einsetzen von *Globotruncana falsostuarti* SIGAL und *Pachydiscus neubergicus* (WIEDMANN 1986, 1988).

Auch wenn diese ungewöhnliche Globotruncanen-Folge des El Kef-Profils regional-ökologische Gründe haben mag, verdient die hier vorgeschlagene Grenzziehung doch Interesse, da dieses Profil als "Hypostratotyp" für Campan und Maastricht vorgeschlagen ist.

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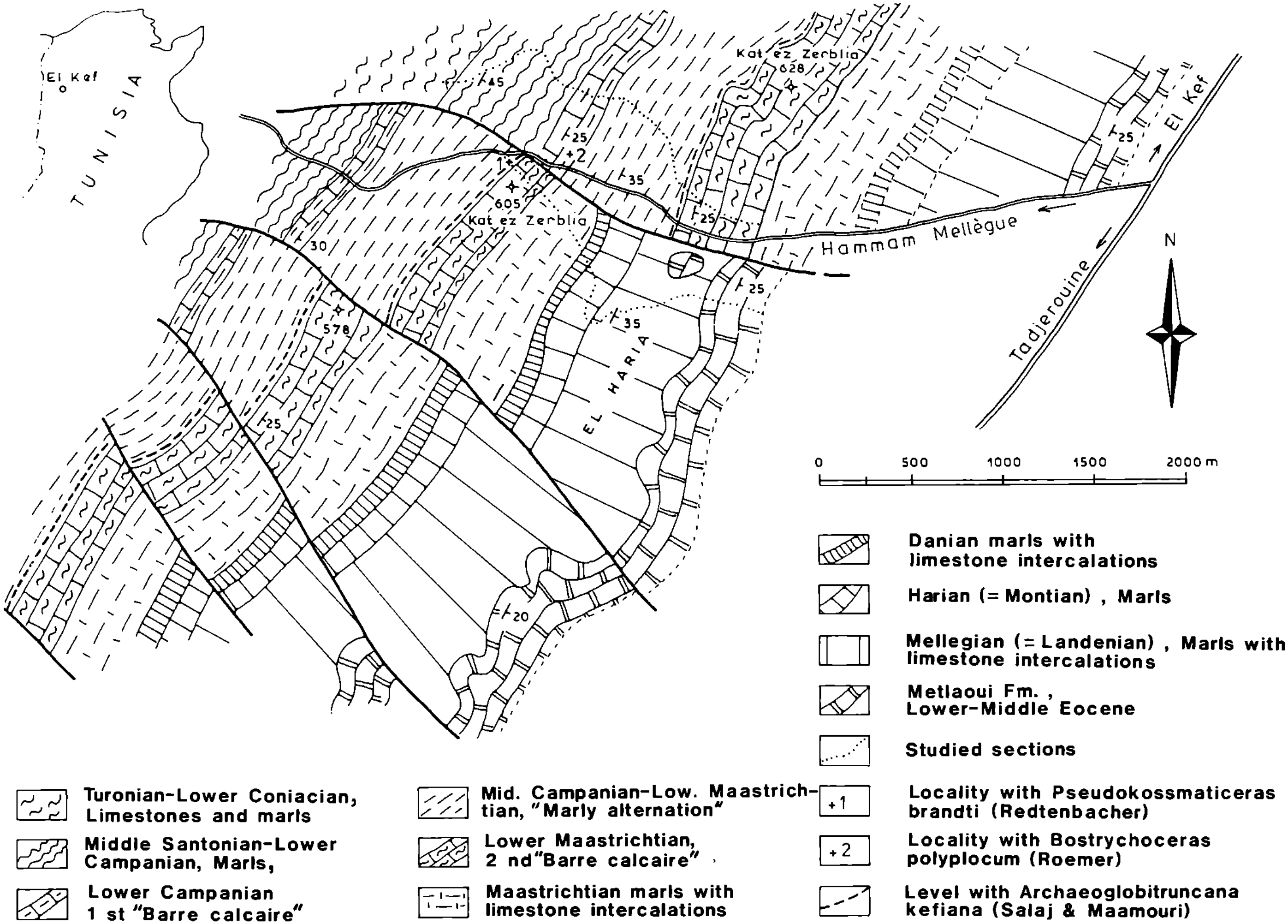
1. Introduction

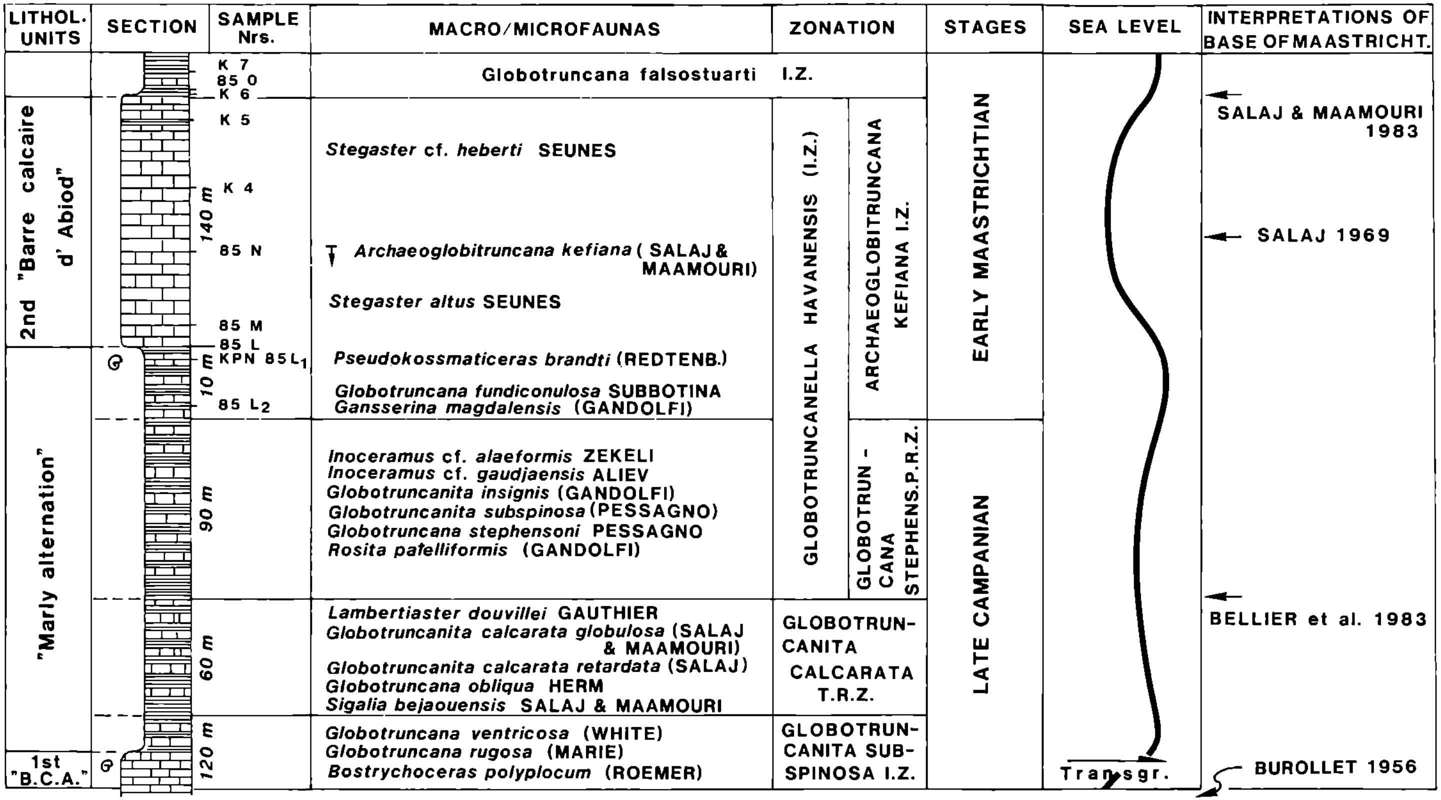
The famous El Kef section in northwestern Tunisia (Text-Fig. 1) has been proposed as a Tethyan hypostratotype for the Campanian and Maastrichtian stages (SALAJ 1973, 1974b, 1978, 1980, 1983, VERBEEK 1976, 1977, SISSINGH 1977, 1978, DONZE 1980, SALAJ & MAAMOURI 1982, 1983, 1984, and BELLIER et al. 1983) as well as for the Paleocene (SALAJ 1974a, b, 1980, POZARYSKA 1976, SALAJ et al. 1976).

However, the position of the Campanian-Maastrichtian boundary was problematic mainly due to the incompleteness of the European stage stratotypes (VAN HINTE 1965). In earlier zonations based on foraminifera, the first attempt to draw this boundary in the El Kef area was made by BUROLLET (1956: 130); he placed the boundary in the upper portion of the first "barre calcaire d'Abiod" and referred the top of the sequence with *Bostrychoceras polyplocum* (ROEMER) to the early Maastrichtian (base of Text-Fig. 2). SALAJ (1969, 1974a) moved the boundary to the central portion of the second "barre calcaire d'Abiod", above the last occurrence of *Stegaster altus* (SEUNES) and the disappearance of *Globotruncanita calcarata* CUSHMAN. Later, SALAJ & MAAMOURI (1982) realized that the *calcarata* specimens of the second "barre calcaire" had to be separated as a different species, *Archaeoglobitruncana kefiana*.

In accordance with the generally accepted Tethyan boundary position, BELLIER et al. (1983) proposed to shift the Campanian-Maastrichtian boundary at El Kef to the base of the Globotruncanella havanensis Zone, the lower portion of which was treated as a Globotruncana orientalis Subzone. This is the equivalent of the Globotruncana stephensoni Zone as established by SALAJ (1983). Finally, SALAJ & MAAMOURI (1983, 1984) suggested placing the stage boundary on top of the second "barre calcaire" at El Kef, thereby including the Globotruncanella havanensis Zone in the Campanian and starting the Maastrichtian with the appearance of *Globotruncana falsostuarti* SIGAL. On the basis of the first ammonite occurrence at the

Text-Fig. 1. Location map of the El Kef area, Western Tunisia, and the described Campanian-Maastrichtian transitional beds.


El Kef
TUNISIA
Kat ez Zerblia
628
605
578
Kat ez Zerblia
Hammam Mellègue
El Kef
Tadjerouine
EL HARIA
N
45
25
35
30
25
25
35
20
0 500 1000 1500 2000 m
Danian marls with limestone intercalations
Harian (= Montian), Marls
Mellegian (= Landenian), Marls with limestone intercalations
Metlaoui Fm., Lower-Middle Eocene
Studied sections
Locality with Pseudokossmaticeras brandti (Redtenbacher)
Locality with Bostrychoceras polyplocum (Roemer)
Level with Archaeoglobitruncana kefiana (Salaj & Maamouri)
Turonian-Lower Coniacian, Limestones and marls
Middle Santonian-Lower Campanian, Marls,
Lower Campanian 1 st "Barre calcaire"
Mid. Campanian-Low. Maastrichtian, "Marly alternation"
Lower Maastrichtian, 2 nd "Barre calcaire"
Maastrichtian marls with limestone intercalations



Text-Fig. 2. Campanian-Maastrichtian transitional section at El Kef, Tunisia, the main faunal occurrences, proposed biostratigraphic subdivision, and sea level changes.

Campanian-Maastrichtian transition at El Kef, the confusing boundary problem has to be reconsidered.

2. Definition of the boundary

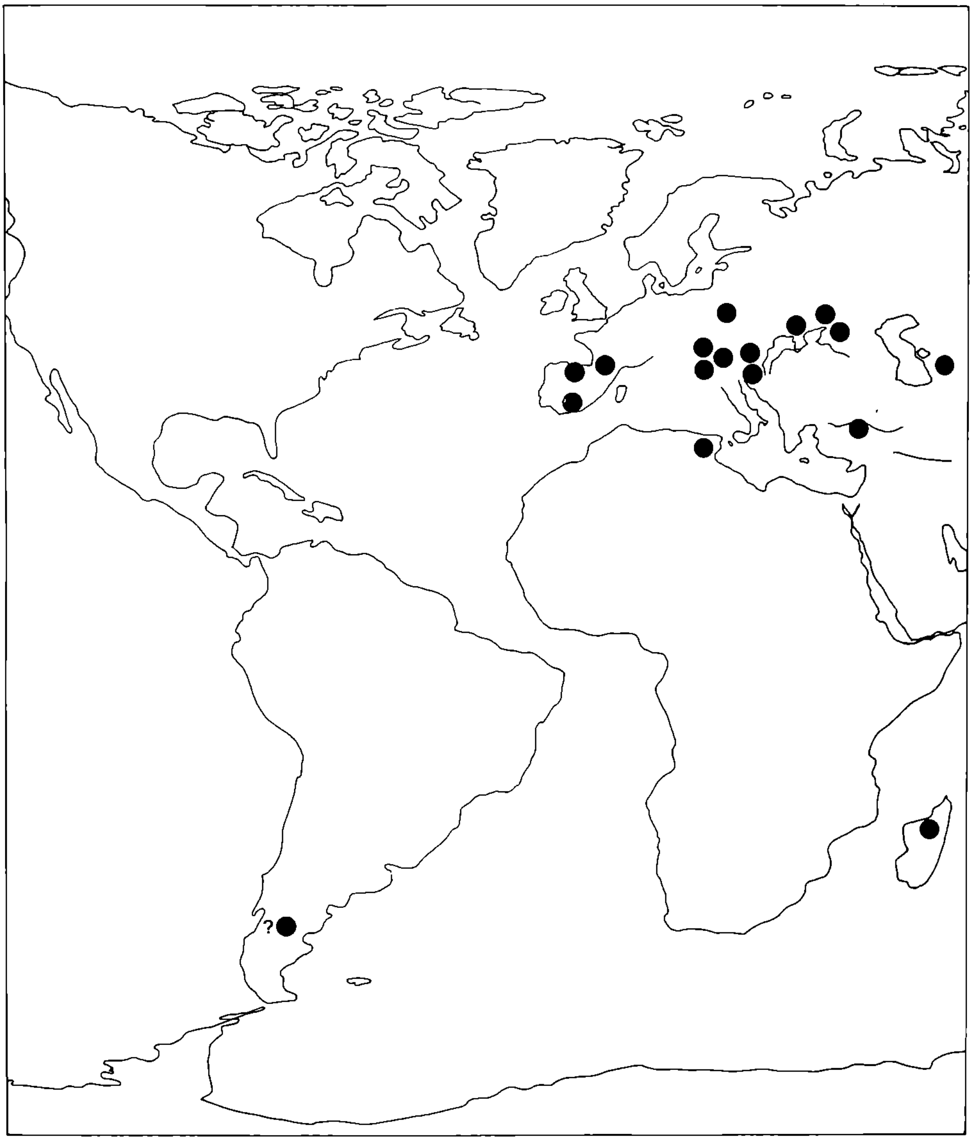
Important for establishing the Campanian-Maastrichtian boundary was the discovery of *Pseudokossmaticeras brandti* (REDTENBACHER) (Text-Fig. 3) in the marly alternation below the second "barre calcaire" (Text-Fig. 2) and thus near the first appearance of *Archaeoglobitruncana kefiana* (SALAJ & MAAMOURI) together with *Neoflabellina praereticulata* HILTERMANN.

In a paper on the Carinthian Gosau beds, THIEDIG & WIEDMANN (1976) revised the European pseudokossmaticeratids and confirmed a Lower Maastrichtian age for *Pseudokossmaticeras brandti* (REDT.) and allied forms. A basal Maastrichtian age can now be reconfirmed for this species group from recent investigations at the famous Zumaya section in northern Spain (WIEDMANN 1986, 1988), another candidate for a Tethyan Maastrichtian hypostratotype, and from the Pyrenees (BILOTTE 1984 -1986), Poland (BLASZKIEWICZ 1980), and the western and southern USSR (MIKHAILOV 1951, NAJDIN & SHIMANSKY 1959, ATABEKJAN & AKOPJAN 1970, AKOPJAN, ATABEKJAN & SHIMANSKY 1974, NAJDIN 1974, KENNEDY & SUMMESBERGER 1987). The problem of defining this stage and its boundaries has

Text-Fig. 3. *Pseudokossmaticeras brandti* (REDT.), lateral view of hypotype, GPIT coll. no. 1658. El Kef, level "KPN 85 L2". 1/1.

been stressed by BIRKELUND et al. (1984: 16f.) and is, to a large extent, due to the incompleteness of the "type" sections near Maastricht and the lack of Tethyan faunas within these sections, and vice versa, absence of Temperate faunas in Tethyan sections (KENNEDY 1984).

Pseudokossmaticeras



Text-Fig. 4. Paleobiogeography of early Maastrichtian pseudokossmaticeratids.

Foraminiferal data indicate that there is a considerable sedimentary thickness in the El Kef section between the Upper Campanian Zone of Globotruncanita calcarata and the first appearance of *Globotruncana falso-stuarti* SIGAL. As mentioned above, two more intermediate zones were here separated in between, i. e., a lower Zone of Globotruncana stephensoni and an upper Zone of Archaeoglobitruncana kefiana. There is still much controversy about their inclusion in the Campanian or Maastrichtian, and moreover about their restricted endemic or cosmopolitan value. But since *Pseudokossmaticeras brandti* (REDT.) has been discovered at the base of the Kefiana Zone, these discussions can now be abandoned. The Kefiana Zone is now considered to characterize the lowermost Maastrichtian, while the Stephensoni Zone - having not yet yielded any ammonite remains - is included in the Campanian.

Due to the favourable conditions of the El Kef section, which contains large numbers of foraminifera and calcareous nannoplankton, both zones can be perfectly defined by their faunal content (Tables 1, 2); the most important species are figured in Text-Figs. 5 and 6.

Table 1. The Foraminifera of the Globotruncana stephensoni Zone in the El Kef section.

Bolivinopsis rosula (EHRENBERG)
Spiroplectammina navarroana CUSHMAN
Haplophragmium grande CUSHMAN
Verneuilina triquetra (MUENSTER)
Gaudryina glabrella (CUSHMAN)
Gaudryina pyramidata CUSHMAN
Gaudryina supracretacea HOFKER
Heterostomella faveolata (MARSSON)
Tritaxia capitosa (CUSHMAN)
Tritaxia trilatera (CUSHMAN)
Dorothia bulleta (CARSEY)
Dorothia conula (REUSS)
Dorothia indentata (CUSHMAN & JARVIS)
Dorothia oxycona (REUSS)
Clavulina clavata CUSHMAN
Plectina conversa (GRZYBOWSKI)
Plectina watersi CUSHMAN
Neoflabellina aff. *efferata* (WEDEKIND)
Neoflabellina rugosa (D'ORBIGNY)
Neoflabellina leptodisca (WEDEKIND)
Neoflabellina permutata KOCH
Praebulimina acuta (REUSS)
Praebulimina reussi (MORROW)
Bolivina incrassata crassa VASILENKO & MJATLIUK
Bolivina incrassata incrassata REUSS
Bolivina plaita CARSAY
Bolivinoides delicatus CUSHMAN
Bolivinoides granulatus HOFKER
Bolivinoides miliaris HILTERMANN & KOCH
Bolivinoides decoratus decoratus (JONES)
Bolivinoides praelevigatus BAAR
Cibicides beaumontiana (D'ORBIGNY)
Cibicides bembix (MARSSON)
Cibicides voltziana (D'ORBIGNY)
Quadrimorphina camerata (BROTZEN)
Pullenia cretacea CUSHMAN
Globorotalites michelianus (D'ORBIGNY)
Gyroidinoides globosa (HAGENOW)
Gyroidinoides obliquasepta (MJATLIUK)
Gyroidina umbilicata D'ORBIGNY
Eponides haidingeri D'ORBIGNY
Anomalina welleri laevis VASILENKO
Gavelinella costata (BROTZEN)
Gavelinella costulata (MARIE)
Gavelinella monterelensis (MARIE)
Gavelinella whitei (MARTIN)

Table 1, continued.

Stensioeina excolata (CUSHMAN)
Stensioeina pommerana BROTZEN
Globotruncana arca arca (CUSHMAN)
Globotruncana arca rugosa (MARIE)
Globotruncana bollii GANDOLFI
Globotruncana mariei BANNER & BLOW
Globotruncana obliqua HERM
Globotruncana stephensoni PESSAGNO
Globotruncana ventricosa WHITE
Globotruncanella havanensis (VOORWIJK)
Globotruncanita smithi (SALAJ)
Globotruncanita stuartiformis (DALBIEZ)
Globotruncanita subspinosa (PESSAGNO)
Archaeoglobigerina globulosa (SALAJ & MAAMOURI)
Rosita fornicata (PLUMMER)
Rosita manaurensis (GANDOLFI)
Rosita patelliformis (GANDOLFI)
Rosita scutilla (GANDOLFI)
Rugoglobigerina rugosa (PLUMMER)
Hedbergella aspera (EHRENBERG)
Globigerinelloides volutus (WHITE)
Heterohelix globulosa (EHRENBERG)
Heterohelix ultimatumida (WHITE)
Heterohelix striata (EHRENBERG)
Heterohelix pulchra (BROTZEN)
Pseudoguembelina excolata (CUSHMAN)
Pseudotextularia elegans (RZEHAK)
Planoglobulina carseyae (PLUMMER)
Ventilabrella glabrata CUSHMAN
Sigalia bejaouensis SALAJ & MAAMOURI
Gublerina reniformis (MARIE)

Table 2. The Foraminifera of the Archaeoglobitruncana kefiana Zone in the El Kef section.

Ammodiscus cretaceus (REUSS)
Bolivinopsis rosula (EHRENBERG)
Heterostomella faveolata (MARSSON)
Heterostomella mexicana CUSHMAN
Gaudryina cretacea (KARRER)
Gaudryina pyramidata CUSHMAN
Dorothia conula (REUSS)
Dorothia crassa (MARSSON)
Dorothia oxycona (REUSS)
Clavalina clavata CUSHMAN
Pyramidina szajnochae (GRZYBOWSKI)
Pyramidina cristata (MARSSON)
Rzehakina epigona (RZEHAK)
Lagena sulcata REUSS
Gyroidina globosa HAGENOW
Gyroidina umbilicata D'ORBIGNY
Frondicularia aff. *jarvisi* CUSHMAN
Eponides haidingeri D'ORBIGNY
Globulina lacrima REUSS
Pullenia quinqueloba (REUSS)
Pullenia reussi CUSHMAN & TODD
Pullenia cretacea CUSHMAN
Pleurostomella subnodosa REUSS
Stensioeina exsculpta (REUSS)
Stensioeina pommerana BROTZEN
Anomalina welleri laevis VASILENKO
Gavelinella costata (BROTZEN)
Gavelinella monterelensis (MARIE)
Cibicides beaumontiana (D'ORBIGNY)
Cibicides excavata BROTZEN
Cibicides voltziana (D'ORBIGNY)
Praebulimina carseyae (PLUMMER)
Bolivina incrassata incrassata (REUSS)
Bolivina plaita CARSEY
Bolivinoides decorata (JONES)
Bolivinoides draco miliaris HILTERMANN & KOCH
Neoflabellina efferata WEDEKIND
Neoflabellina numismalis WEDEKIND
Neoflabellina rugosa D'ORBIGNY
Neoflabellina praereticulata HILTERMANN

Table 2, continued.

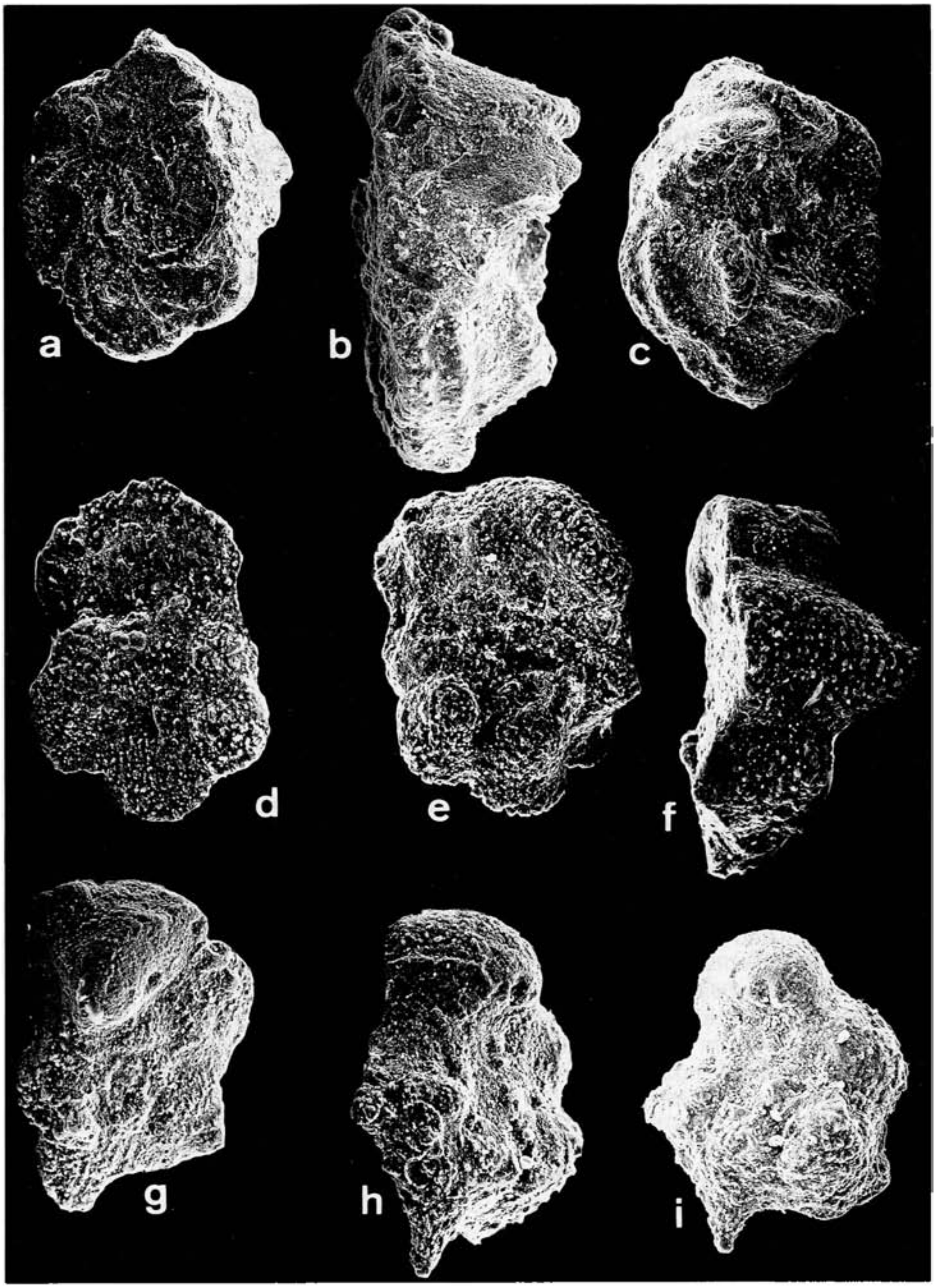
Globigerinelloides volutus (WHITE)	*Gansserina magdalensis* (GANDOLFI)
Globigerinelloides multispinus (LALICKER)	*Gansserina wiedenmayeri* (GANDOLFI)
Hedbergella planispira (TAPPAN)	*Rosita fornicata* (PLUMMER)
Hedbergella crassa (BOLLI)	*Rosita plummerae* (GANDOLFI)
Rugoglobigerina rugosa (PLUMMER)	*Rosita patelliformis* (GANDOLFI)
Archaeoglobitruncana kefiana (SALAJ & MAAMOURI)	*Rosita scutilla* (GANDOLFI)
Archaeoglobigerina bulloides (VOGLER)	*Heterohelix globulosa* (EHRENBERG)
Globotruncana arca arca (CUSHMAN)	*Heterohelix moremani* (CUSHMAN)
Globotruncana arca rugosa (MARIE)	*Heterohelix navarroensis* LOEBLICH
Globotruncana fundiconulosa SUBBOTINA	*Heterohelix planata* (CUSHMAN)
Globotruncana obliqua HERM	*Heterohelix pulchra* (BROTZEN)
Globotruncana stephensoni PESSAGNO	*Heterohelix ultimatumida* (WHITE)
Globotruncana trinidadensis GANDOLFI	*Pseudoguembelina costata* (CUSHMAN)
Globotruncana ventricosa WHITE	*Pseudoguembelina excolata* (CUSHMAN)
Globotruncanella havanensis (VOORWIJK)	*Sigalia bejaouensis* SALAJ & MAAMOURI
Globotruncanita insignis (GANDOLFI)	*Gublerina reniformis* (MARIE)
Globotruncanita smithi (SALAJ)	*Pseudotextularia deformis* (KIKOINE)
Globotruncanita subspinosa (PESSAGNO)	*Pseudotextularia elegans* (RZEHAK)
	Planoglobulina brazoensis MARTIN
	Planoglobulina carseyae (PLUMMER)
	Ventilabrella glabrata CUSHMAN

It should be stressed that the globotruncanid species *Gansserina magdalensis* (GANDOLFI) (Text-Figs. 5 i and 6 d, e) and *Globotruncana fundiconulosa* SUBBOTINA (Text-Figs. 5 d-f and 6 a-c) appear with *Archaeoglobitruncana kefiana* (SALAJ & MAAMOURI) (Text-Fig. 6 f-i). Further important species are *Rosita patelliformis* (GANDOLFI) (Text-Fig. 5 a-c), *Globotruncana bollii* GANDOLFI (Text-Fig. 5 g, h), *Gansserina wiedenmayeri* (GANDOLFI), and *Neoflabellina praereticulata* HILTERMANN. As seen in Table 1, *Globotruncanita smithi* (SALAJ), *Gl. subspinosa* (PESSAGNO), and many others disappear at the same time.

A break in calcareous nannoplankton sequence, i. e., the disappearance of *Quadrum trifidum* (STRADNER), occurs in the central part of the Kefiana Zone (sample no. 210 of SISSINGH 1977, fig. 9) and thus cannot be used for boundary definition.

3. Paleoenvironment

Considering the total Campanian (530 m) and Maastrichtian (320 m) sedimentary thicknesses in the Kef area (SALAJ & MAAMOURI 1982) and the absolute time ranges of both stages (after ODIN & KENNEDY 1982), i. e., 11 and 7 my, respectively, we get comparatively high average sedimentation rates of 48.2 m/my for the Campanian and 45.7 m/my for the Maastrichtian. This is one of the reasons for the eventual use of the El Kef section as a Tethyan hypostratotype.



A second reason is that the region of the "sillon tunisien" and the El Kef area exhibit epibathyal environments at the time of the Campanian-Maastrichtian transition. Based on ostracods, DONZE et al. (1982) postulated a depositional depth of about 400 to 500 m. The same can be said from the benthic and planktonic foraminiferal associations of late Campanian (Table 1) and early Maastrichtian ages (Table 2, and SALAJ & MAAMOURI 1982, SALAJ 1988).

In the lowermost portion of the Kefiana Zone, which belongs to the marly alternation and yields *Pseudokossmaticeras brandti* (REDT.), the planktonic foraminifera of all four bathymetric zones of HART & BAILEY (1979) are recognized. The first zone of surface waters is represented by globular forms of the genera *Hedbergella, Rugoglobigerina* and *Heterohelix,* and by *A. kefiana* (SALAJ & MAAMOURI). The second bathymetric zone is represented by flat globotruncanids like *Globotruncana obliqua* HERM and *Archaeoglobigerina bulloides* (VOGLER) as well as by *Sigalia bejaouensis* (SALAJ & MAAMOURI), gublerinids, and pseudoguembelinids. The third zone is represented in the Kef section by biconvex forms like *Globotruncana arca* (CUSHMAN), *Gl. stephensoni* PESSAGNO, *Gl. trinidadensis* GANDOLFI, and *Rosita fornicata* (PLUMMER), as well as by ventrilabellids, planoglobulinids, and racemiguembilinids. Finally, the fourth bathymetric zone is proven by globotruncanids of spiro-convex and ventro-convex shape: *Globotruncanita insignis* (GANDOLFI), *Gl. smithi* (SALAJ), *Globotruncana bollii* (GANDOLFI), and *Rosita patelliformis* (GANDOLFI).

Compared with the Recent (BÉ 1977), the first zone of surface waters occupies the upper 50 m, the second and third zones extend to water depths of 100 m, while the fourth bathymetric zone corresponds to water depths of several hundred meters, with adults generally restricted to water depths exceeding 100 m (zone III of BÉ 1977).

Obviously the lower part of the Kefiana Zone thus coincides with the maximum transgression/subsidence in the Kef area (Text-Fig. 2). Decreasing subsidence and/or regression can, however, be observed during the times of deposition of the two "barres calcaires", e. g. by rare occurrences of reworked orbitoid foraminifera in the first "barre calcaire" and by the presence of echinoids and inoceramids in the second "barre calcaire". Moreover, in the uppermost portion of the marly alternation are shell beds with

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Text-Fig. 5. a-c: *Globotruncana fundiconulosa* SUBBOTINA. Sample KPN-85 L1, Kat ez Zerblia, near El Kef. Archaeoglobitruncana kefiana Zone with *Pseudokossmaticeras brandti* (REDT.).

a. Spiral view of DSIGB coll. no. T20, 55 x. b. Lateral view, 75 x. c. Umbilical view, 60 x.

d, e: *Gansserina magdalensis* (GANDOLFI). Sample KPN-85 L1. Same level and locality.

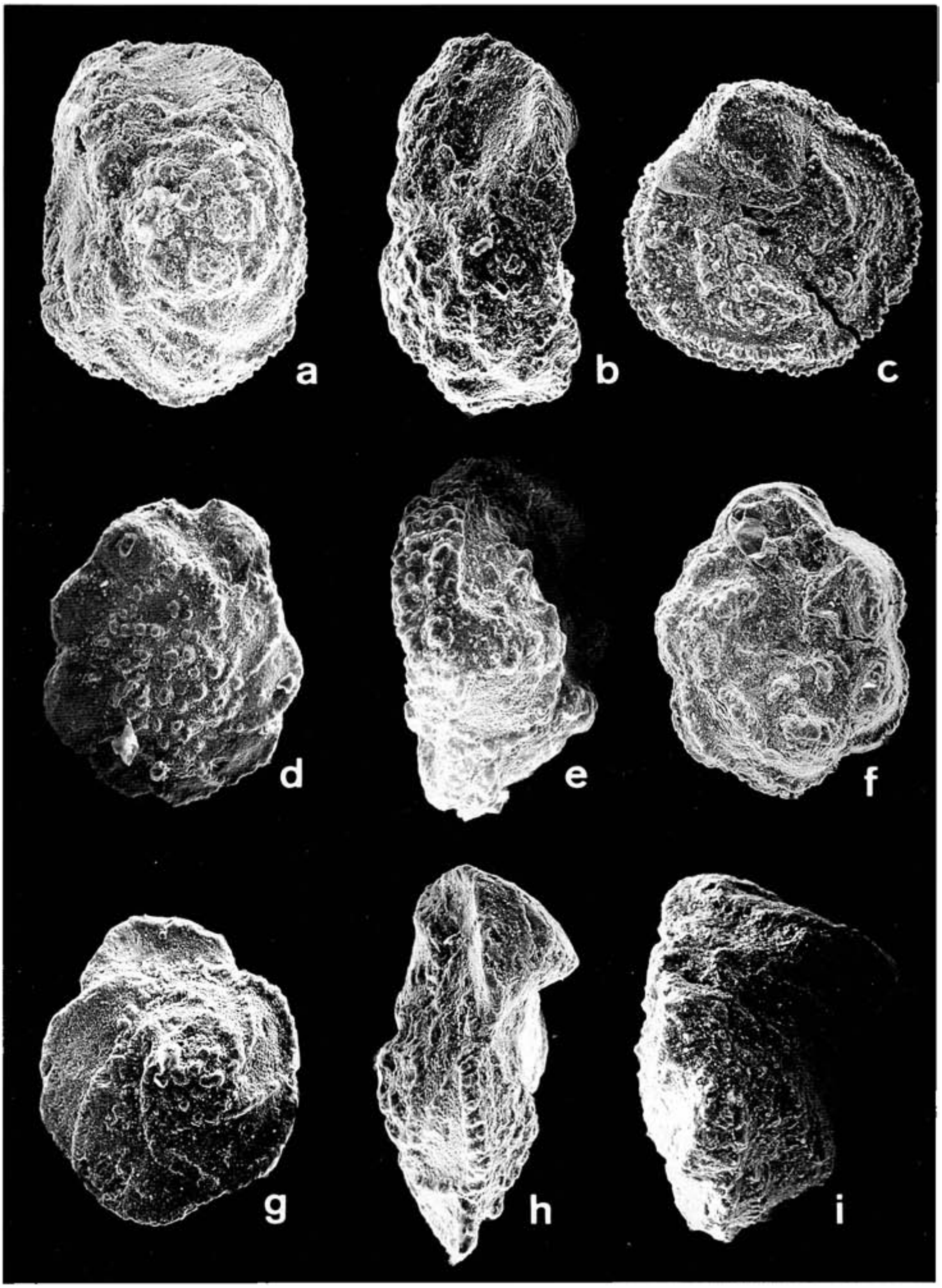
d. Spiral view of DSIGB coll. no. T21, 85 x. e. Umbilical view, 85 x.

f-i: *Archaeoglobitruncana kefiana* (SALAJ & MAAMOURI). Sample KPN-85 L1. Same level and locality.

f. Lateral view of DSIGB coll. no. T22, 100 x. g. Spiral view of DSIGB coll. no. T23, 90 x. h. Lateral view of DSIGB coll. no. T24, 85 x. i. Umbilical view, 85 x.

debris of rudists, oysters, and echinoids associated with small brachiopods indicating the existence of a nearby platform area.

Worth mentioning are, moreover, redeposited microfaunas in the Abathomphalus mayaroensis Zone of the late Maastrichtian (SALAJ 1980), as



well as the coeval nearly complete absence of the typical planktonics of the fourth bathymetric zone of HART & BAILEY (1979), i. e., ventro- and dorsoconical forms like *Globotruncana ventricosa* (WHITE), *Gl. fundiconulosa* SUBBOTINA, *Gansserina gansseri* (BOLLI), *Globotruncanita stuarti* (DE LAPPARENT), or *Rosita contusa* (CUSHMAN).

Despite the fact that these were short-term episodes, oceanic oscillations and changes in bathymetry can thus be recognized in the El Kef sedimentary sequence (Text-Fig. 2).

4. Conclusions

The Campanian-Maastrichtian boundary beds were investigated in the El Kef section, western Tunisia. Rich associations of foraminifera were studied between the Upper Campanian Calcarata Zone and the Falsostuarti Zone of Lower Maastrichtian age. Two zones can be distinguished within this interval, a lower Zone of Globotruncana stephensoni and a higher Zone of Archaeoglobitruncana kefiana. The co-occurrence of *Pseudokossmaticeras brandti* (REDT.) with the latter infers that the Kefiana Zone is lowermost Maastrichtian. The Stephensoni Zone is better included in the late Campanian, but no ammonites have been found so far from this part of the section.

The paleoenvironment of that time was epibathyal. No oxygen depletion can be recognized. The marly alternation between the two "barres calcaires d'Abiod" (BUROLLET 1956), in which the Campanian-Maastrichtian boundary is now placed, coincides with maximum subsidence due to the prevalence of 4th zone planktonics in the El Kef area (SALAJ & MAAMOURI 1982: 469).

The known paleogeographic distribution of representatives of the *Pseudokossmaticeras brandti* group (i. e. *Ps. tercense* (SEUNES), *Ps. galicianum* (FAVRE), and *Ps. cerevicianum* (PETHÖ), *P. muratovi* MIKHAILOV, and *P. tchihatcheffi* BÖHM) can now be extended from Russia, Poland, Turkey, Rumania, Bulgaria, Yougoslavia, Italy, Austria, southern Germany, southern France and Spain to northern Africa (Text-Fig. 4). *Pseudokossmaticeras* sp.

Text-Fig. 6. a-c: *Rosita patelliformis* (GANDOLFI). Sample no. 19a, Kat ez Zerblia, near El Kef (for details see SALAJ & MAAMOURI 1982: 464). Archaeoglobitruncana kefiana Zone.
a. Spiral view of DSIGB coll. no. T15, 70 x. b. Lateral view, 70 x. c. Umbilical view, 60 x.
d-f: *Globotruncana fundiconulosa* SUBBOTINA. Sample no. 19a. Same level and locality.
d. Spiral view of DSIGB coll. no. T16, 50 x. e. Lateral view, 75 x. f. Umbilical view, 50 x.
g, h: *Globotruncana bollii* GANDOLFI. Sample no. 19a. Same level and locality.
g. Spiral view of DSIGB coll. no. T17, 60 x. h. Lateral view of DSIGB coll. no. T18, 75 x.
i: *Gansserina magdalensis* (GANDOLFI). Sample no. 19a. Same level and locality. DSIGB coll. no. T19, 70 x.

aff. *brandti* in COLLIGNON (1973, pl. 647, fig. 2399) from the Lower Maastrichtian of Madagascar may be better included in *Brahmaites*. The group of *Pseudokossmaticeras pacificum* (STOLICZKA) seems to be restricted to the Lower Maastrichtian of southern India and Madagascar. *(?)Pseudokossmaticeras hauthali* (PAULCKE) from southern Patagonia (PAULCKE 1906) remains included with doubts in the present genus. The lack of pseudokossmaticeratids in NW Europe and N America is due to a widespread sedimentary gap at the Campanian-Maastrichtian boundary rather than to provinciality.

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B. Alps, Carpathians, Dinarids, Caucasus

Mobile Belts and Early Cretaceous Orogenic History of the Mediterranean Region – A Review

TODOR G. NIKOLOV, Sofia

With 4 Text-Figures

NIKOLOV, T. G. (1989): Mobile Belts and Early Cretaceous Orogenic History of the Mediterranean Region - A Review. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 319-328. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The development of the Lower Cretaceous is shortly described in the western and eastern Mediterranean Region. The Lower Cretaceous developed in mobile zones, which are usually inherited from the Jurassic. These zones are characterized through ophiolitic magmatism, strong subsidence and the distribution of thick flysch sediments. The basic orogenic events in the Mediterranean Region during early Cretaceous times are discussed.

Kurzfassung: Die Entwicklung der Unterkreide des westlichen und östlichen Mediterrangebietes wird kurz umrissen. Die Unterkreide entwickelte sich in mobilen Zonen, die bereits im Jura vorhanden waren. Diese Zonen sind gekennzeichnet durch ophiolitischen Magmatismus, starke Subsidenz und die Verbreitung mächtiger Flyschsedimente. Die wichtigsten unterkretazischen orogenen Ereignisse im Mediterrangebiet werden diskutiert.

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1. Introduction

The early Cretaceous evolution of the Tethys is connected with considerable tectonic activity known in the central parts of the Mediterranean Region (DEWEY et al. 1973, AUBOUIN 1973, BAUMGARTNER & BERNOULLI 1976, LAUBSCHER & BERNOULLI 1977).

The author considers the Mediterranean Region as part of the open Jurassic-Cretaceous oceanic space between Africa and Eurasia (NIKOLOV 1987). In contrast to the Jurassic and late Cretaceous the early Cretaceous in this region was a relatively quiescent time. This is probably the reason why in many works scanty information is given about the early Cretaceous evolution of the Tethys. As a matter of fact, the fossil record of the Lower Cretaceous in the Mediterranean offers enough information to draw an interesting picture of the tectonic diastrophism of the Central Tethys.

This review is based on several assumptions:

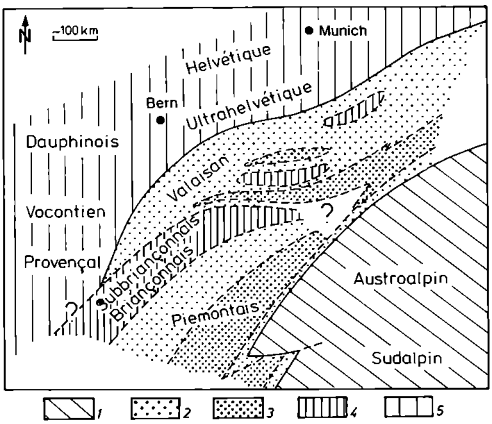
(1) Early Cretaceous mobile belts are inherited middle zones of the ocean in the Jurassic.
(2) These mobile belts developed into a complex plate tectonic network which is part of the global rift system (DEWEY et al. 1973, LAUBSCHER & BERNOULLI 1977, DEBELMAS & SANDULESCU 1987).
(3) The mobile belts include the central parts of the Mediterranean zone together with spreading ridges, zones of ophiolite magmatism and zones of active subsidence with thick flysch deposits.
(4) The mobile zones are surrounded by carbonate platforms on top of the northern and southern slopes of the Tethys (DERCOURT et al. 1985, NIKOLOV 1987).
(5) The early Cretaceous evolution of the Tethys is connected with the opening of the Atlantic Ocean (DERCOURT 1971, DEWEY et al. 1973, WIEDMANN 1982).

2. Oceanic Areas and Transform Zones

2.1 **Data** (Text-Figs. 1-4)

2.1.1 Western Tethys and Alps

The palinspastic reconstruction of the ancient Mediterranean Region should begin with the Alps as a key zone of the Tethys.



Text-Fig. 1. Tentative palinspastic sketch map of the Western Tethyan Ocean and its continental margins in the Alpine-Apennine area (approximately early Cretaceous) (after TRÜMPY 1980, modified according to CARON et al. 1981).
1 - Austro-Subalpine margin; 2 - Trough with thin continental crust, possibly oceanic to the east; 3 - Trough with oceanic sediments, Bündnerschiefer and ophiolites (oceanic crust); 4 - Microplate with thin continental crust; 5 - European margin.
Valaisian = North Penninic zone; Briançonnais = Middle Penninic zone; Piemontais = South Penninic zone.

Many authors noting the remnants of oceanic crust in the Alps (LAUBSCHER 1969, DEWEY & BIRD 1970, BIJU-DUVAL et al. 1977, LAUBSCHER & BERNOULLI 1977, DEBELMAS et al. 1980), called this zone the Ligurian-Piemont Zone. It is marked by the development of Jurassic ophiolites, which can be observed from the Mediterranean Sea in the south up to Grisons in the north and north-east, where they disappear under the Austroalpine nappes. As indicated by some authors (CORTESOGNO et al. 1981, BARRETT 1982), ophiolites and associated pelagic sediments of late Jurassic and early Cretaceous age are found in the Ligurian Apennines. They are compared to mid-oceanic ridge formations. In the Grisons area (Eastern Switzerland) ophiolites and associated pelagic sediments are the major components of the South Penninic nappes (WEISSERT & BERNOULLI 1985). These authors note that the Jurassic ophiolites form the basement which is overlain by the Upper Jurassic to Lower Cretaceous oceanic sediments: pillow-basalts, radiolarites, schists with chert, pelagic limestones - known as *Calpionella*- or *Aptychus*-limestones - black shales, and turbidite arenites. Generally speaking, the Lower Cretaceous sediments in the Ligurian-Piemont (or South Penninic) Zone constitute a pelagic cover of the Jurassic ophiolites.

The shales, widely known as Schistes lustrés or Bündnerschiefer, whose correlation with the South Penninic pelagic sediments remains unclear, present a special problem. The Bündnerschiefer formations were deposited in the Penninic Zone in the interval between the early Jurassic and the Cenomanian (BOLLI & NABHOLZ 1959). Opinions concerning the kind of deposition of the Bündnerschiefer are contradictory. Some authors define them as typical deep-sea formations or "deposited below the CCD" (ISLER & PANTIC 1980), but quite rightfully KELTS (1981) argues that Bündnerschiefer are hemipelagic formations, including sediments deposited on the basin slopes. In the time interval from the Middle Jurassic to the early Cretaceous sills of mafic rocks were injected into the hemipelagic oozes of Bündnerschiefer, while serpentinite elements and/or gabbros are wedging out lengthwise along transform faults (KELTS 1981).

The presence of thick metabasaltic (tholeitic) layers in association with serpentinite in the Bündnerschiefer formations is considered by some authors to represent "local openings" of the Valais Trough (DIETRICH & OBERHÄNSLI 1975). The Valais or North Penninic Trough is identical with the North Apennine Transform Zone of DEBELMAS & SANDULESCU (1987).

2.1.2 Eastern Tethys and Carpathians

The trough or rift of the Outer Dacides (DEBELMAS & SANDULESCU 1987) is an analogue of the Valais "rift" in the east of the Carpathians. In fact, according to SANDULESCU (1975), the Dacides include a structural assembly deformed during the Cretaceous and occupying the inner part of the Carpathians.

The character of the Upper Jurassic and the "Lower Units" (nappes of the black flysch of Gehlau in the Eastern Carpathians, as well as the Sinaia Beds which are in some places associated with ophiolites (Azuga Beds), and the nappes of Severin in the Southern Carpathians support the idea that this rift is a remarkable tectonic fissure (DEBELMAS & SANDULESCU 1987, NIKOLOV 1987).

The main structures of the Mid-Dacides (the Central and Eastern Carpathic nappes, Getic and Supragetic nappes) separate the Outer Dacides from the other oceanic areas of the Carpathians in the Metalliferous Mountains (Southern Apusenes). This situation is analogous to the separation of the Valais Trough from the Ligurian-Piemont or South Penninic Zone throughout the Briançonnais or Middle Penninic area (DEBELMAS & SANDULESCU 1987).

The presence of oceanic crust is marked by widespread development of ophiolites and volcano-sedimentary units of a basite type in the Southern Apusenes (the southern part of the Metalliferous Mountains). This zone continues to Eastern Serbia where Serbian geologists describe volcano-sedimentary rocks with diabases of the so-called Vratarnica Series (ANJELKOVIC 1975, GRUBIC 1975). It is possible that the Vratarnica Series is an analogue of the Azuga Beds in the Southern Carpathians (NIKOLOV 1987). In any case, the oceanic areas of the Carpathians do not continue into the Balkanides as it is incorrectly suggested on map 3 and 4 of DERCOURT et al. (1985). This interpretation is based on the idea of DEWEY et al. (1973) that Moesia and the Rhodopes are separated by oceanic crust. As HSÜ (1977) points out, this supposition of DEWEY et al. (op. cit.) is unacceptable.

During the early Cretaceous the Fore-Balkan was a basin of intensive subsidence filled mainly with up to 5 km thick flysch sediments (NIKOLOV 1987). The crust is thin, but of continental type. Cretaceous oceanic crust in Bulgaria developed during the late Cretaceous in the Srednogorie region.

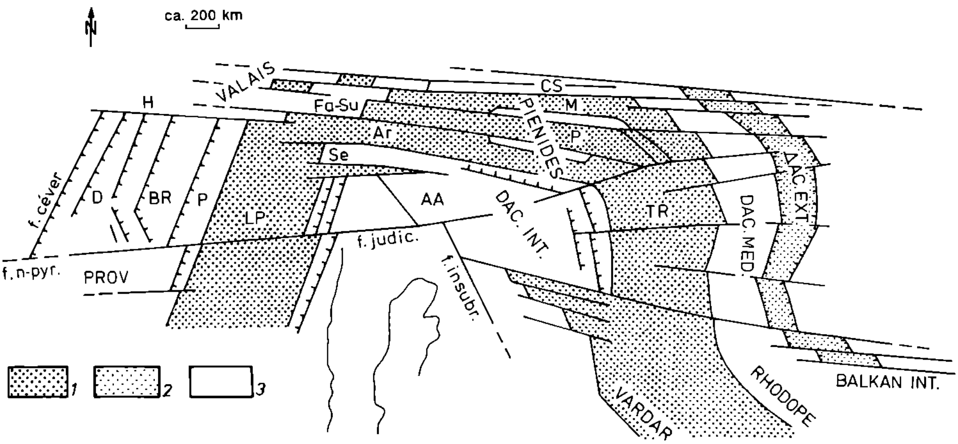
In Vilani and Meczek (the Pannonian depression), the Valanginian is connected with volcano-sedimentary formations, i. e. the trachyte-dolerite unit (FÜLÖP 1966). The emergence of this unit is determined by a transform fault (DERCOURT 1971). To the south of this fault there are traces of oceanic crust in the Vardar Zone, to the south of the Rhodopes and to the east. During the early Cretaceous oceanic crust also developed in the area between the Greater Caucasus and the Dzirulla Massif, where there was intensive ophiolite magmatism (MILANOVSKI & KHAIN 1963).

The Lower Cretaceous stratigraphic record of the area in the south-west of the Ligurian-Piemont Zone reveals a continuation of the mid-oceanic ridge in the Western Mediterranean and in the Central Atlantic (NIKOLOV 1987). For this reason, DE GRACIANSKY's et al. (1981) suggestion about the existence of a Mediterranean megatransform-zone in the Ligurian-Piemont region seems acceptable (Text-Fig. 4). According to DE GRACIANSKY et al. (op. cit.) this zone separates the Western Tethys from the North Atlantic.

2.1.3 Magmatic activity in the Western and Eastern Tethys

The magmatic activity in the Mediterranean Region is of great interest. Early Cretaceous basaltic magmatism is manifested in the continental slope of northwestern Africa and the Iberian Peninsula (UCHUPI et al. 1976, HINZ et al. 1982). In the Pyrenees near Bestiak, southeast of Tarascon, there is a locality of Aptian-Albian peridotites. In this area the Aptian-Albian sediments are slightly metamorphosed (PEYBERNES 1976).

During the Albian-Cenomanian there was a considerable magmatic activity, represented by a series of deposits from basalts to trachytes west of Garonne River. This magmatism is connected with considerable faulting of



Text-Fig. 2. Tentative palinspastic sketch map of the oceanic Tethyan domain and of its continental margins in the Alpine-Carpathic area in the early Cretaceous (after DEBELMAS & SANDULESCU 1987).
1 - oceanic crust; 2 - thinned continental crust, locally of oceanic type; 3 - continental crust.
The Austroalpine domain is replaced in its palinspastic position after a clockwise rotation of about 60°, according to paleomagnetic data (WESTPHAL 1976).
AA - Austroalpine domain; Ar - Arosa; BR - Briançonnais domain; CS - Silesian Cordillera; D - Dauphiné domain; Dac.int. - intern Dacides; Dac. med. - median Dacides; Dac.ext. - extern Dacides; f. - fault; Fa - Falknis; H - Helvetic domain; LP - Ligurian-Piemont domain; M - internal Magura zone; P - Pieniny Trough; Prov. - Provençale domain; Se - Sesia zone; Su - Sulzfluh; TR - Transylvanides.

the crust along the North Pyrenean fault (AUTRAN & DERCOURT 1980). Alkaline intrusions (nepheline syenites) of Albian age can be observed in the Corbières, S France.

The development of ophiolites, volcano-sedimentary complexes and radiolarites in the South Penninic (Ligurian-Piemont) and North Penninic zones, in the Pannonian depression (Meszek), the Carpathians between the Greater Caucasus and Dzirulla and in the Lesser Caucasus was mentioned above.

A number of intrusions composed mainly of quartz granodiorite developed in places of gabbro differentiation in the Lesser Caucasus during the early Cretaceous. The age of most of the intrusions is estimated at 100-120 my (Barremian-Albian), the age of the Zav Massif is 132-135 my (Berriasian) (MILANOVSKI & KHAIN 1963).

There was a considerable early Cretaceous ophiolite magmatism in Trodos (Cyprus) and in Kizil-Dag (Turkey). The age of the Kizil-Dag dolerites is 80-100 my, being formed most intensively during the early Cretaceous (Valanginian-Aptian). Part of the Kizil-Dag gabbro has also an early Cretaceous age (100-130 my, Valanginian-Albian).

The Trodos ophiolites are mainly late Cretaceous with the exception of part of the dolerites formed in the interval of 120-130 my (Valanginian-Hauterivian).

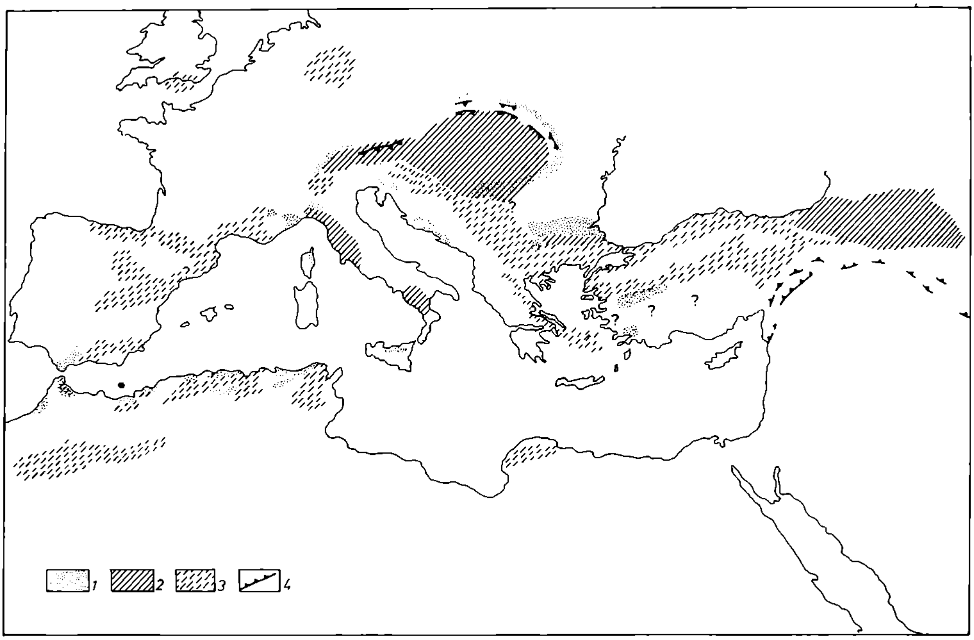
The flysch sediments are another characteristic depositional element in the Mediterranean Region. Their distribution coincides with the main arcs in the Mediterranean: the Gibraltar, the Alpine, and the Carpathian-Balkan arcs in the north, and the Calabrian-Sicilian and the Dinaric-Aegean arcs in the south (DURAND DELGA 1980, NIKOLOV 1987). There are flysch sediments also in several lineaments, e. g. in the Pyrenean, the Kraistides-Lujnitsa and in the Caucasus geosynclines. Quite often the flysch sediment-zones are bound to active continental margins thus forming flysch wedges in the subduction zones (AUBOUIN 1973).

Pelagic sediments and/or carbonate platforms were deposited in vast marginal areas between the mobile zones.

Pelagic limestones, often mixed with radiolarites, developed in the Vocontian Trough, the Alps (especially the Southern Alps), the Ionian zone, the Dinarides, and the Hellenides. All of these localities have inherited the ophiolite zones (WEISSERT 1981, WEISSERT & BERNOULLI 1985, NIKOLOV 1987).

2.2 **Orogenic events: A summary** (Text-Fig. 3)

The early Cretaceous orogenic events in the Mediterranean Region are ex-



Text-Fig. 3. Sketch map of the development of the early and middle Cretaceous flysch and folding in the Mediterranean Region (after ARGIRIADIS 1974).
1 - flysch; 2 - significant tectogenesis; 3 - moderate tectogenesis; 4 - nappe front.

amined in detail in a recent study (NIKOLOV 1987). For this reason, only a short summary is given here.

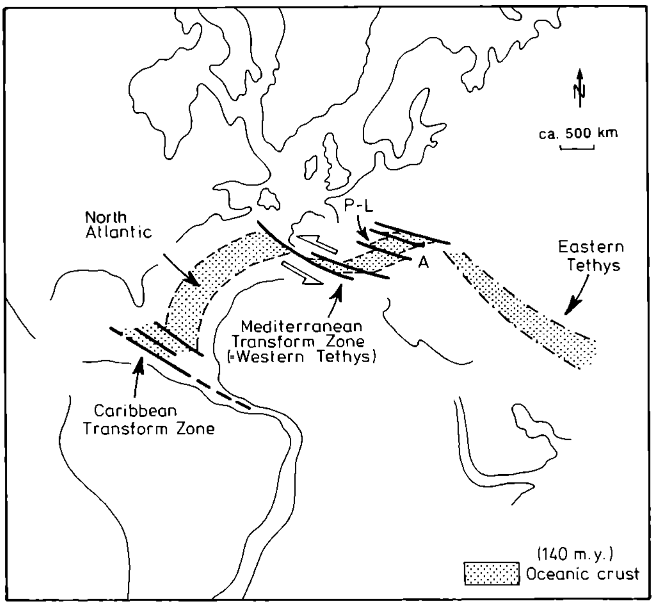
The Lower Cretaceous stratigraphic record of the Mediterranean Region reveals a number of regional unconformities. In some zones the first important orogenic event manifested itself as early as the beginning of the Barremian. The second orogenic event occurred at the beginning of the Gargasian, followed by post-Aptian, Albian and pre-Cenomanian foldings that affected vast areas of the Tethys. Most intensive are the Aptian-Albian and pre-Cenomanian events (ARGIRIADIS 1974, NIKOLOV 1987).

3. Early Cretaceous Evolution of the Mediterranean Region and its Connections with the Atlantic: Some Speculations (Text-Figs. 2-4)

The development of Lower Cretaceous units in the Mediterranean Region supports the reconstruction model of DEWEY et al. (1973).

There are two orogenic stages in the early Cretaceous evolution of the Mediterranean: Berriasian-Barremian and Aptian-Albian.

During the Berriasian-Barremian stage there was gradually increasing activity with a Hauterivian-Barremian paroxysm. There were active movements along the faults between the small Iberian plate and the European plate. The early Cretaceous ophiolites in the Alps and the Apennines were formed during the same stage. At the same time were also generated the



Text-Fig. 4. Tentative palinspastic sketch of the late Jurassic - early Cretaceous Tethys (after LEMOINE 1980, BERNOULLI & LEMOINE 1980; modified by DE GRACIANSKY et al. 1981).
A = Apulia; P-L = Piemon-Ligurian (South Penninic) oceanic domain.

Lower Cretaceous part of the Maiolica and Biancone formations, which can be traced from the central part of the Atlantic through the Southern Alps, the Lombardia zone, and the Central Apennines up to the Ionian Zone (BERNOULLI 1972, WEISSERT 1981, DE GRACIANSKY et al. 1981). The obduction of the Othrys ophiolites from the Vardar Zone took place during the Barremian (DEWEY et al. 1973, RICOU & MARCOUX 1980).

The second stage (Aptian-Albian) coincides with a strong activity in the Mediterranean Region. At the beginning of the Aptian the rate of spreading increased about three times compared to the beginning of the early Cretaceous. During the second stage a number of subduction zones were outlined: the South Penninic zone in the Alps, the Carpathians, Eastern Serbia, the Vardar Zone, the West Pontides (DEWEY et al. 1973, DERCOURT et al. 1985). The Aptian-Albian stage coincides with the black shale episode as mentioned by DE GRACIANSKY et al. (1981).

The Tethys in the Mediterranean Region was reduced gradually in size due to tectonic activity (subduction of the South Penninic and related zones) during the early Cretaceous parallel to the opening of the Atlantic Ocean.

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Sedimentology and Mineralogy of the Valanginian and Hauterivian in the Stratotypic Region (Jura Mountains, Switzerland)

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With 13 Text-Figures

ADATTE, T. & RUMLEY, G. (1989): Sedimentology and Mineralogy of the Valanginian and Hauterivian in the Stratotypic Region (Jura Mountains, Switzerland). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 329-351. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: With two profiles situated in the stratotypic region of Valanginian and Hauterivian (near the town of Neuchâtel, Jura mountains, Switzerland), the sedimentary development of the early Cretaceous (Berriasian to Hauterivian) is described. Some major discontinuities are defined and allow us to determine the main sedimentary sequences. A mineralogical study of the insoluble residue (clay and accessory minerals) shows the variation of the mineralogy all along our profiles and demonstrates the relation between the mineralogical assemblages and the sedimentary environments.

Kurzfassung: An zwei Profilen der Typlokalitäten des Valanginium und des Hauterivium (Region Neuchâtel, Schweizer Jura) wird die sedimentäre Entwicklung der tieferen Kreide (Berriasium bis Hauterivium) beschrieben. Mehrere größere Diskontinuitäten wurden erkannt. Sie erlauben es, die wichtigsten Sedimentabfolgen zu unterscheiden. Mineralogische Untersuchungen des karbonatfreien Restsediments (Tonminerale und Akzessorien) zeigen deutliche Schwankungen der Mineralzusammensetzung in den Profilen. Die verschiedenen Mineralassoziationen stehen in engem Zusammenhang zu wechselnden Ablagerungsbedingungen.

Résumé: Deux coupes proches des stratotypes du Valanginien et de l'Hauterivien (région de Neuchâtel, Jura suisse), nous permettent d'illustrer l'évolution sédimentaire du Crétacé inférieur du Berriasien à l'Hauterivien. Plusieurs discontinuités majeures sont reconnues; elles définissent les séquences sédimentaires principales. Parallèlement, une étude minéralogique du résidu insoluble (minéraux argileux et accessoires) est effectuée; elle décrit les variations de la minéralogie le long de nos profils et met en évidence la relation entre les assemblages minéralogiques et les milieux de dépôt.

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1. Introduction

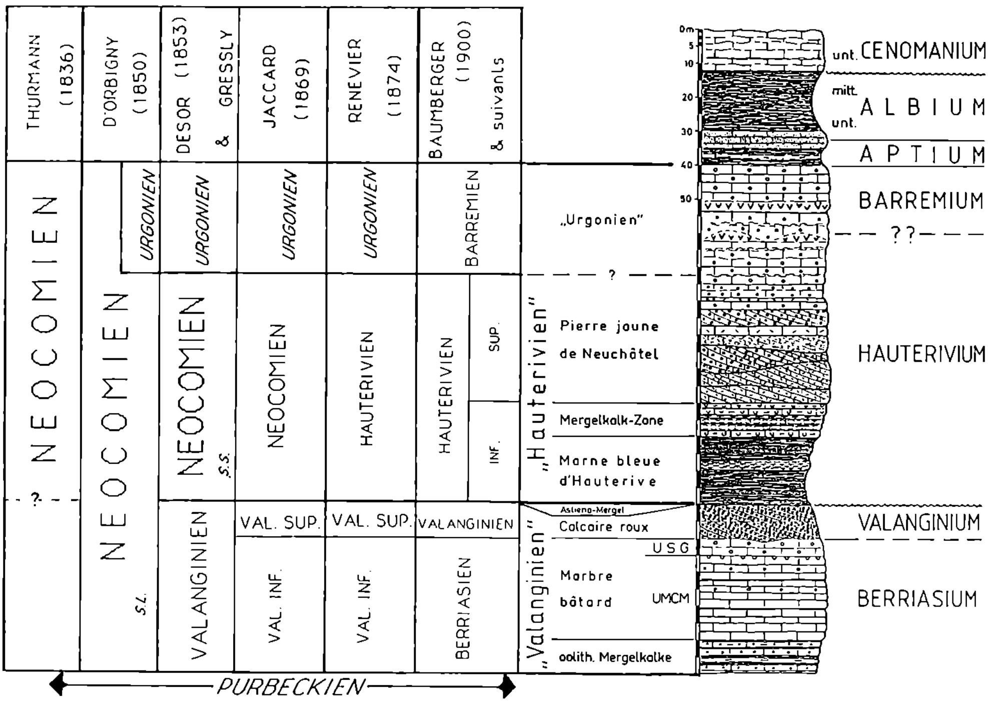
The Swiss Jura Mountains, near the town of Neuchâtel, are the place where two historical stratotypes (Valanginian and Hauterivian) were defined. Text-Fig. 1 resumes the evolution of these historical concepts since the last century.

The name Neocomian takes its origin from Neuchâtel (from Latin, Neocomium, THURMANN 1836) and was used to define the lowest Cretaceous marine layers.

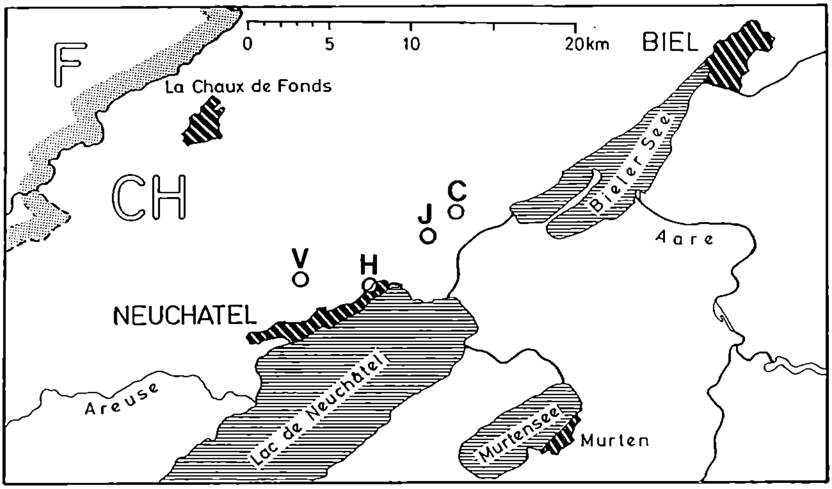
The Valanginian Stage was defined by DESOR (1853), in order to characterize the lower part of the Neocomian.

The Hauterivian Stage (from Hauterive, village east of Neuchâtel) was proposed by RENEVIER (1874) in order to avoid a confusion between the Neocomian s. l. and the Neocomian s. s. of DESOR & GRESSLY (1859) (Text-Fig. 1).

BAUMBERGER (1901) introduced the Berriasian (of COQUAND 1871) for the early Valanginian in the sense of JACCARD (1869). This was based on chronostratigraphic considerations. Therefore, the old Neocomian was divided into three independant stages, leaving aside the Barremian-Urgonian problem.



Text-Fig. 1. Historic interpretations of the early Cretaceous in the stratotypic region.



Text-Fig. 2. General map of the region (with the localization of the profiles and the historical stratotypes; J: Juracime quarry, C: Cressier marl pit, V: Valangin, H: Hauterive).

With two profiles situated near the stratotypes, we will illustrate the sedimentary development of the early Cretaceous, from the Purbeckian to the Hauterivian, and compare it with the mineralogy of the insoluble residue. The first profile is the quarry of Juracime in Cornaux in which we have a complete section from the Purbeckian to the Calcaire Roux, and the second is the marl pit of Cressier which shows a section from the Calcaire Roux to the Pierre Jaune de Neuchâtel (see localizations, Text-Fig. 2).

2. Lithostratigraphic description of the regional early Cretaceous

The stratigraphic succession of the early Cretaceous of the region of Neuchâtel can be described with five main lithostratigraphic subdivisions, according to the historical definition (Text-Fig. 1).

From the bottom to the top, above the Purbeckian facies, one finds the "Unité inférieure oolithique" (1) and the Marbre Bâtard (2), both these two formations belonging to the Berriasian. Above them follows the Calcaire Roux (3) belonging to the Valanginian. It is followed by the Marnes Bleues d'Hauterive (4) and the Pierre Jaune de Neuchâtel (5), which have an Hauterivian age, and precede the Urgonian facies.

(1) The lacustrine and marine facies of the "Purbeckian" are followed by the fully marine facies of the "Unité inférieure oolithique (U.I.O.)" which is a marly, black and grey limestone with ooids, echinoderms and glauconite.

(2) The Marbre Bâtard is a well stratified, grey-white limestone with beds of 0.2 to 2 m and more. Two parts can be distinguished: the lower part is a massive micritic limestone called "Unité moyenne calcaire massive (U.M.C.M.)", and the upper part a limestone containing a rather great

amount of quartz, the "Unité supérieure gréseuse (U.S.G.)". In this unit, we observe the association of *Keramosphera allobrogensis* and *Pseudotextulariella courtionensis* which is found in the whole Swiss Jura, and gives a late Middle Berriasian age. A stratigraphic gap exists between the U.S.G. and the next formation, the Calcaire Roux, which has a Valanginian age.

(3) The Calcaire Roux is a typical calcarenite with cross-stratification, iron-ooids and echinoderm remains. The top of the Calcaire Roux is formed by a hard-ground, on which it is sometimes possible to find a very thin deposit of yellow marls (about 20 cm), the "Marnes à *Astieria*". An ammonite, *Saynoceras verrucosum*, characteristic of the basal Upper Valanginian, was found in these marls at Valangin. It is therefore generally admitted that the whole Upper Valanginian is represented by these 20 cm of marls.

(4) 20 m of "Marnes Bleues d'Hauterive" follow in continuity and contain a rather rich fauna of ammonites, brachiopods, echinoderms, etc. The common occurrence of *Acanthodiscus radiatus* in the first meters of the Marnes Bleues proves the early Hauterivian age of this formation.

(5) The passage to the Pierre Jaune de Neuchâtel is gradual through a zone which is called zone marno-calcaire. This terminology tends to be abandoned because of the thickness variations of this zone from place to place. The Pierre Jaune is again a limestone with cross-stratification, ooids, remains of echinoderms, bryozoans, but without a fauna which would allow a precise age determination. The Hauterivian-Barremian limit can therefore not be determined in the Swiss Jura, yet.

3. Sedimentology

3.1 Berriaso-Valanginian: quarry of Juracime

3.1.1 General description

This large quarry, in which carbonate rocks are exploited for cement, is situated NW of the small village of Cornaux (Text-Fig. 2). It shows in a very good way the early Cretaceous from Purbeckian facies (Berriasian) to Calcaire Roux (Lower Valanginian).

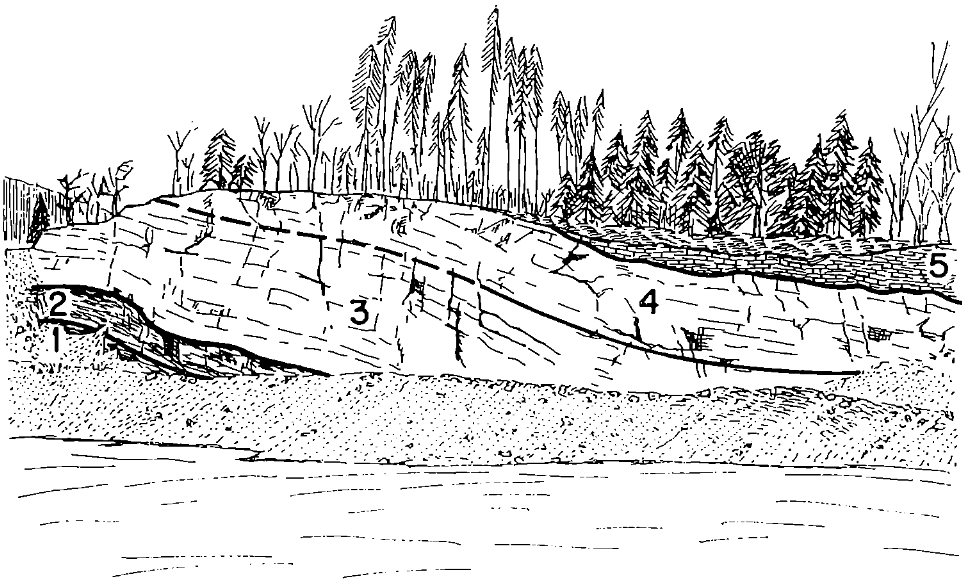
From the bottom to the top, we can observe (Text-Fig. 3):

a) the end of Purbeckian facies
b) the "Unité Inférieure Oolithique" (8 m, U.I.O.)
c) the "Unité Moyenne Massive" (9 m, U.M.M.)
d) the "Unité Supérieure Gréseuse" (10 m, U.S.G.)
e) the "Calcaire Roux" (3 m).

To simplify matters, we used the subdivisions of STEINHAUSER & CHAROLLAIS (1971). These remain appropriate in this quarry, but, in other outcrops, the U.S.G. is completely different, containing only carbonates and no quartz as is the case in the Juracime.

a) Purbeckian

Because of a strong disharmonic tectonic it is very difficult to observe the succession of individual beds. It is composed mainly of grey, black, some-



Text-Fig. 3. General view of the Juracime quarry.
1: Purbeckian facies, 2: Unité inférieure oolithique, 3: Unité calcaire massive, 4: Unité supérieure gréseuse, 5: Calcaire Roux (from STEINHAUSER & CHAROLLAIS 1971).

times green marls, limestones and breccia, which represent a brackish to lacustrine environment. Characeans are abundant, especially in green marls.

b) U.I.O.

At the outcrop, it is difficult to separate this unit from the Purbeckian. It is an alternation of blue to black marls and oolithic limestones. The presence of glauconite, echinoderms, bryozoans, remains of rudistids and some foraminifera (like *Feurtillia frequens*) proves an open marine environment.

c) U.M.M.

This unit is composed of thick beds of white limestones with blue algae and many benthic foraminifera. It represents a lagoonal environment.

d) U.S.G.

It is a succession of red to brown limestones. Its very high content of detrital quartz is the main characteristic. This unit is terminated by a great erosional surface (Text-Fig. 3). These limestones indicate a more external environment, and contain the characteristic association *Keramosphera allobrogensis* and *Pseudotextulariella courtionensis* (Text-Fig. 5).

e) Calcaire Roux

It is a cross-bedding unit which overlies the U.S.G. with a great unconformity, and cuts up to 2 m into the underlaying unit. It is composed of ferrugineous ooids and remains of echinoderms, bryozoans and bivalves, in a sparitic cement.

3.1.2 Listing of the observed microfacies (Text-Figs. 4 and 5)

3.1.2.1 Slope and external platform

Microfacies T1

- Biosparite with very external intraclasts and big remains of echinoderms, pelecypods, brachiopods, bryozoans. Foraminifera are rare (Miliolidae, Nodosaridae). It is possible to observe few ooids; quartz is not common.
Texture: grainstone.

Microfacies Pe1

- Biosparite with small elongated remains of bivalvia, echinoderms, bryozoans (more seldom). Ooids are more abundant. Foraminifera are rare (mainly Nodosaridae). Quartz is not common, glauconite is sometimes present. The elements are well sorted.
Texture: grainstone.

Microfacies Pe2

- Biosparite with elongated remains of bioclasts and intraclasts. The ooids are abundant and show sometimes a micritized rim. Quartz and glauconite are present. Large remains of echinoderms. Foraminifera are more frequent: *Nautiloculina*, Nodosaridae, large agglutinants, *Feurtillia*, Miliolidae, *Trocholina*.
Texture: grainstone.

Microfacies Pe3

- Biomicrite with elongated remains, rudistids, echinoderms and a few ooids. Quartz and glauconite are sometimes present. Foraminifera are more or less the same as in Pe2.
Texture: wackestone to packstone.

Microfacies Pe4

- Biosparite with remains of echinoderms, rudistids, Madreporaria. Quartz may be abundant (especially in the U.S.G.). There is still a little glauconite. It is possible to have the same facies with micrite. Foraminifera are more diversified: Nodosaridae, Miliolidae, *Nautiloculina*, Polymorphinidae and a lot of *Trocholina*. In the U.I.O. this facies shows a few transported gyrogonites of *Chara*. Ooids are frequent.

Texture: grainstone or packstone.

Microfacies Pe5

- Biosparite or biomicrite with remains of rudistids, echinoderms and madreporaria. Ooids are absent in this facies. Quartz can be present with high value. Foraminifera are diversified: Miliolidae, Nodosaridae, *Nautiloculina*, Polymorphinidae, and a lot of *Trocholina* and *Pseudocyclammina*.

Texture: grainstone to packstone.

3.1.2.2 Internal platform

Microfacies Pi1

- Biopelsparite with small micritized remains, echinoderms and foraminifera. Quartz is not common. Foraminifera are more common: frequent *Pseudocyclammina*, a few *Trocholina*, seldom *Pfenderina* and Miliolidae. *Lenticulina* is still frequent. As algae, *Cayeuxia* is sometimes found.

Texture: grainstone.

Microfacies Pi2

- Biointramicrite, sometimes sparite with rolled and micritized remains. The echinoderms tend to decrease. Bivalvia and brachiopods are common. Representants of *Lenticulina* are less frequent. The most abundant foraminifera are especially *Trocholina*, *Pseudocyclammina*, *Nautiloculina*, seldom *Pfenderina* and sometimes *Pseudotextulariella*. Algae, like *Cayeuxia* may be present.

Texture: packstone.

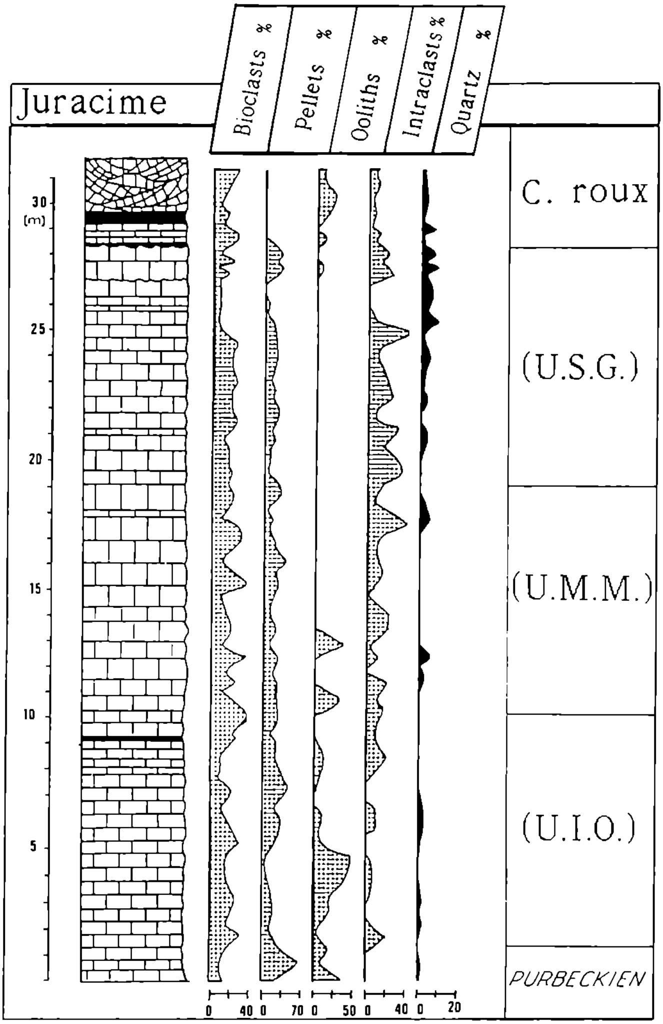
Microfacies Pi3

- Biomicrite or biosparite with diversified fauna and with large foraminifera. The echinoderms tend to disappear. Algae (Dasycladales) are present, too. The remains are micritized. The foraminifera are *Veurneuilla* cf. *polonica*, *Nautiloculina*, large Miliolidae, *Pseudotriloculina*, *Glomospira*, *Pseudocyclammina*, large agglutinants and a few *Pseudotextulariella*. Nodosaridae become more seldom. Quartz is present with variable content.

Texture: wackestone to packstone, grainstone.

Microfacies Pi4

- Biomicrite with algae and frequent *Trocholina*. The other foraminifera are Miliolidae, *Nautiloculina*, *Pyrgo*, *Glomospira*, *Pseudotriloculina*, and sel-



Text-Fig. 4. Percentage of the principal elements observed in microfacies.

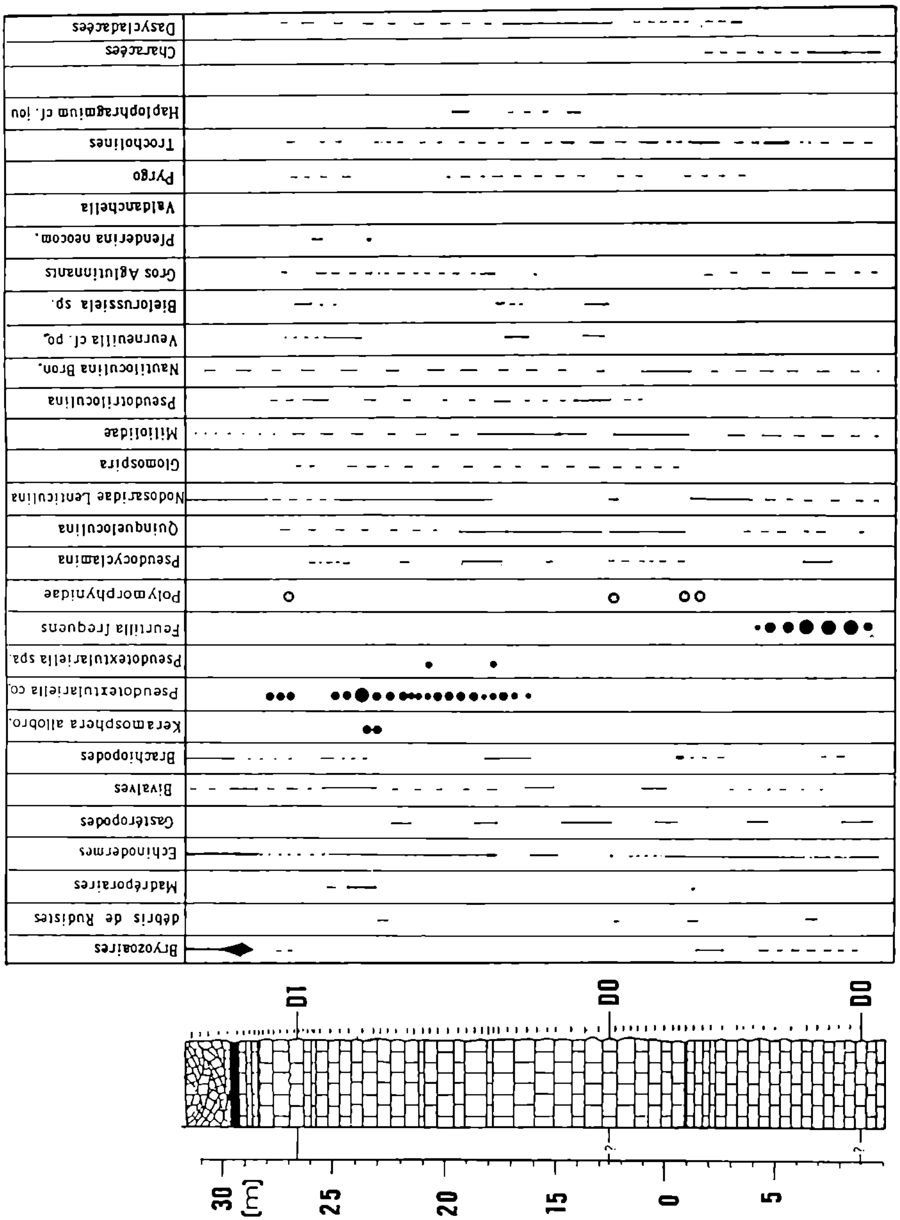
dom *Pseudotextulariella*. The algae are Dasycladales (*Clypeina*). Echinoderms are absent. Quartz is not common.

Texture: wackestone to packstone.

Microfacies Pi5

- Biosparite or biomicrite with oncoids. Echinoderms are absent. Foraminifera are mainly *Trocholina* and Miliolidae, *Glomospira*, *Pseudotriloculina*, *Bielorussiella*, *Nautiloculina*, seldom *Verneuilla*, *Haplophragmium* and *Pseudotextulariella*. Quartz is not common.

Texture: packstone to grainstone.



Text-Fig. 5. Repartition of the foraminifera and bioclasts.

Microfacies M1

- Micrite with *Pseudotriloculina*. The fauna is extremely poor. The foraminifera are *Pseudotriloculina, Bielorussiella, Pyrgo*. The other bioclasts are only ostracods, rare bivalvia and seldom algae. Birds-eyes are very common.
Texture: mudstone.

3.1.2.3 Mixed facies

Microfacies FT

This microfacies corresponds to the "Faciès de Transgression" of A. ARNAUD-VANNEAU (1980), and occurs often at discontinuities. It is a mixture of internal and external elements, but can also be a mixture of facies (for example M1 and slope facies). It reveals very external conditions. In our profile we have the following microfacies:

- Biopelmicrite with thin rolled remains. Echinoderms are very common. The more external foraminifera are small Polymorphinidae, mixed with more internal like *Bielorussiella, Pseudotriloculina, Pyrgo,* Miliolidae, and *Pseudotextulariella.* Nodosaridae are very common. Algae are present. Other constituents are small pellets mixed with large ooids. Quartz is also characteristic of this facies type and is present in a higher content (until 20 %). Glauconite is often present.

It is evident that the position of one or two microfacies on the theoretic profile of a carbonate platform is a little bit arbitrary. But we are obliged to order these if we want to use the results of mineralogy correctly.

3.1.3 The evolution curve of microfacies

Text-Fig. 6 shows the evolution of microfacies and principally three discontinuities.

The discontinuity D0 (end of Purbeckian facies) corresponds to the installation of open marine conditions, which causes a great change in fauna. This discontinuity was first recognized in the meridional Jura by DARSAC (1983).

The discontinuity D0' suddenly breaks the internal platform sedimentation that follows the more external facies of the U.I.O. It could correspond to the D0' which has been recognized by BOISSEAU (1987) in SE France.

The discontinuity D1 corresponds to a major event which can be correlated from here to the region of the Bourget Lake.

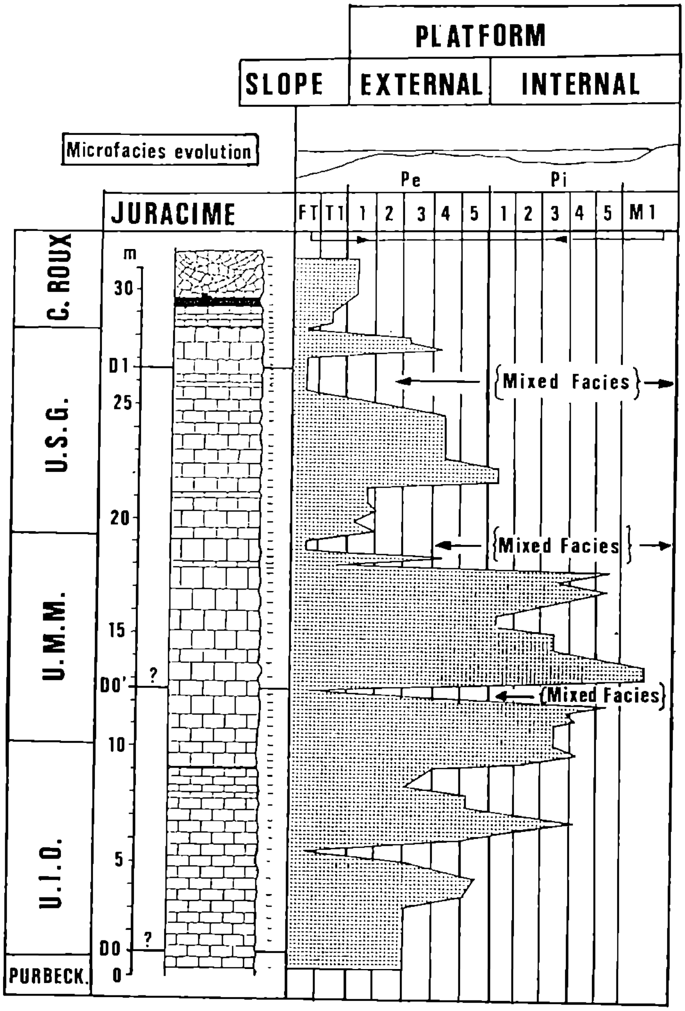
All these discontinuities correspond to mixed facies.

3.2 Hauterivian: marl pit of Cressier

3.2.1 General description

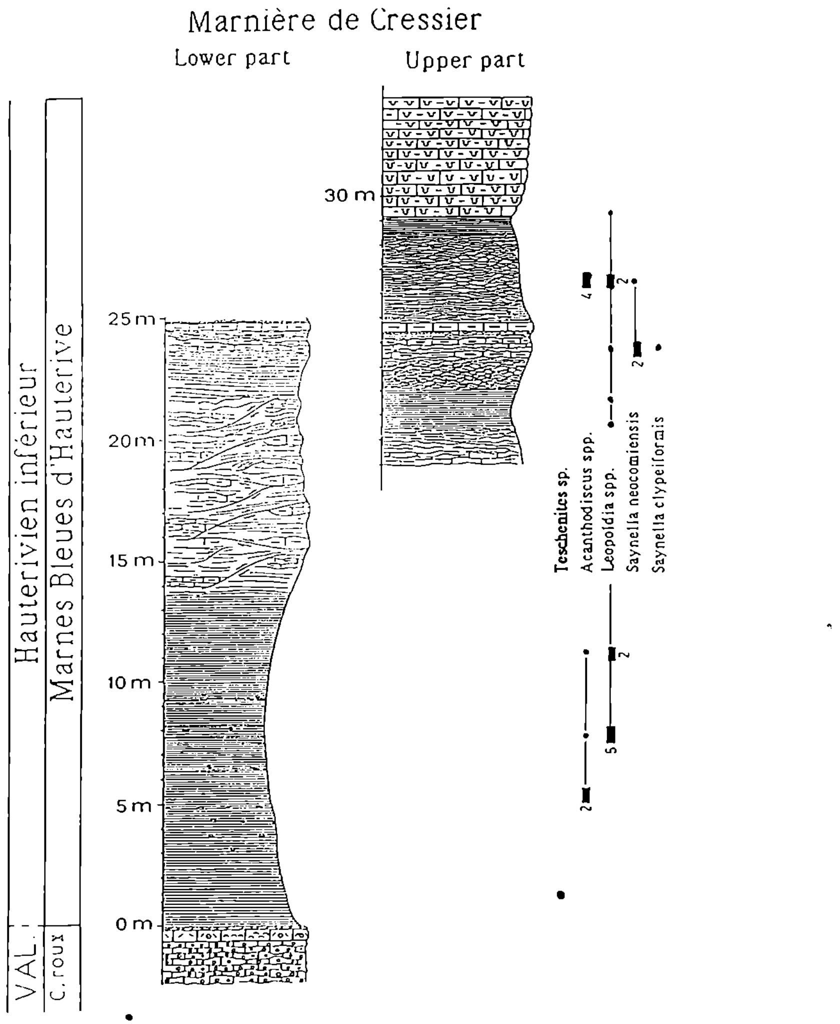
This outcrop is situated north of the village of Cressier (Text-Fig. 2), and the marls were exploited here until the beginning of the century. Some inconveniences are present: tectonic complications, vegetation, frequent fall of rocks and accumulation of the debris prevent us from having a good view of the series. *These facts and the absence of a good stratification (there are almost no clearly recognizable beds) make a precise measurement and description very difficult. However, this is the only place where we can see so much of the Marnes Bleues d'Hauterive and their contacts* with the lower and upper units.

The lithology given in Text-Fig. 7 was surveyed by REMANE (1982). He measured the profile from the top of the Calcaire Roux. This involves, of course, an inaccuracy for the total thickness on account of the absence of a good stratification.



Text-Fig. 6. Microfacies evolution of the Juracime profile.

In the northern part of the marl pit, the contact between the Calcaire Roux and the Marnes Bleues d'Hauterive is exposed. There are no Marnes à *Astieria*, here. The last bed of the Calcaire Roux is a grey calcarenite ending with a hard-ground; a typical Marne Bleue follows immediately. About 3 to 4 m above is the main fossiliferous bed of the marl pit; it is a marl with calcareous nodules. In the following meters, no beds can be clearly distinguished. Only between 24 and 25 m, REMANE described a more calcareous bed with burrows, the surface of which looks like a hard-ground. The upper part of the Marnes Bleues is a hard calcareous marl showing no stratification. The transition to the Pierre Jaune de Neuchâtel is very rapid; the first calcareous beds are grainstones, sandwiched by



Text-Fig. 7. Lithology and repartition of ammonites of the Marnière de Cressier (paleontological data from BUSNARDO & THIEULOY, in prep.).

some marly levels. The typical Pierre Jaune with cross-bedded stratification appears a few meters above.

3.2.2 Paleontology and stratigraphy

The Marnes à *Astieria* do not exist in Cressier. This means that the Upper Valanginian is not represented here. Thus, except for the Calcaire Roux, the entire marl pit series belongs to the Lower Hauterivian.

The Marnes Bleues show a typical open-sea fauna, where ammonites are rather often found. Among them, the most frequent genera are *Acanthodiscus* and *Leopoldia*. Some *Saynella* and one *Teschenites* were also found by BUSNARDO & THIEULOY (in prep.) (Text-Fig. 7). All these ammonites belong to the first zone of the Hauterivian, the Acanthodiscus radiatus Zone. This implies that certainly no great gap is within the Marnes Bleues d'Hauterive.

Otherwise, the fauna consists of many bivalves, brachiopods, echinids (*Toxaster*) and some nautilids.

The microfauna consists mainly of foraminifera. Among them the most frequent are *Lenticulina*, *Gaudryina* and some Nodosaridae.

In the Pierre Jaune, the fauna is only represented by debris of brachiopods, bivalvia, echinoderms, bryozoans and crinoids. The most frequent foraminifera are *Lenticulina* and *Gaudryina*.

3.2.3 Sedimentology

Above the hard-ground at the top of the Calcaire Roux - a biosparite with echinoderm and bryozoan remains - begins the sedimentation of the Marnes Bleues, which are characteristic of a rather deep open-sea environment. These marls are biomicrites with many remains of echinoderms, some annelids and bryozoans. They contain frequently glauconite and quartz.

Except for the hard-ground mentioned above, no significant change appears in the sedimentation of the Marnes Bleues. The hard-ground certainly does not involve a great gap, as is seen in the paleontological part.

At the top of the marls, the sedimentological environment passes gradually to an oolitic and/or bioclastic facies, the Pierre Jaune de Neuchâtel. This grainstone shows either ooids or debris of bioclasts, which are sometimes of very big size. Glauconite and chlorite are frequent. In other profiles, a major discontinuity can be seen in the Pierre Jaune. This discontinuity is marked by a hard-ground with some pebbles. This discontinuity might be correlated with "discontinuity B" as defined by VIEBAN (1983) in the meridional Jura and the Subalpine Chains.

4. Mineralogy

4.1 Introduction to the mineralogical study of carbonate rocks in relation with sedimentology

A mineralogical study was carried out for all our samples in each profile, in order to show the relation between the mineralogical association and the sedimentary environment. This method was first applied by VIEBAN (1983) and by DARSAC (1983) in Grenoble under the direction of Mr. B. KUEBLER and Mrs. and Mr. A. and H. ARNAUD.

In this method, we consider essentially the clay and the accessory minerals. Each sample is decarbonated, separated in two fractions (less than 2 and between 2 and 16 µm), and then analysed by X-ray diffraction. The main minerals recognized are:

- quartz, iron sulphides and oxydes (pyrite and goethite), titanium oxydes, sometimes dolomite, plagioclase and feldspar for the accessory minerals;

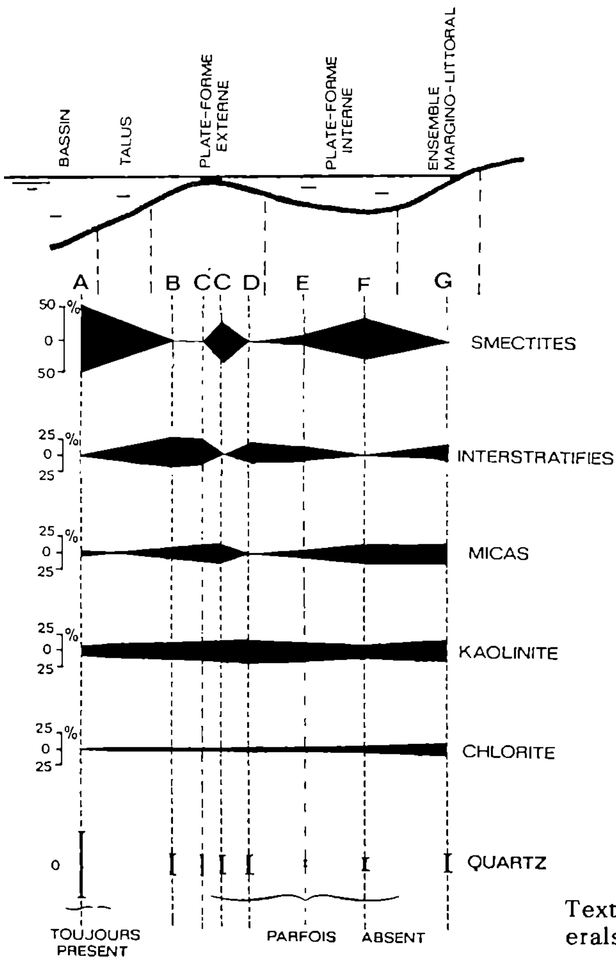
- detritic micas, kaolinite, chlorite, smectites and mixed-layers, for the clay minerals.

We assume that all these minerals have a detrital origin. Either with the absolute intensities, or with relative percentage (only for phyllosilicates) we can observe the distribution of each mineral all along our profiles.

In order to draw a parallel between mineralogy and microfacies, we arrange the diffractograms in groups according to their morphological characteristics. Then we search the corresponding microfacies for each sample and finally obtain the repartition of roentgenofacies (= classes of diffractograms) on a theoretical profile of a carbonate platform.

Knowing the mineralogical association of each roentgenofacies, we are able to determine the repartition of minerals on the platform (Text-Fig. 8).

Finally, we can draw the evolution curve of roentgenofacies by the same way as for microfacies. The Juracime profile will show the similarity between the two curves (Text-Fig. 11).



Text-Fig. 8. Repartition of minerals on the platform.

4.2 Berriaso-Valanginian

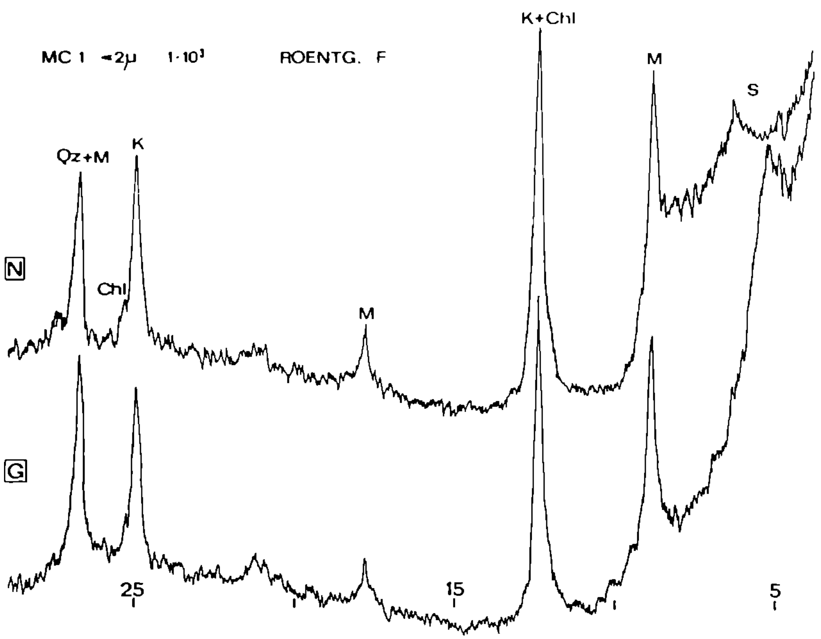
Text-Fig. 10 shows the distribution of the main clay minerals which have been identified in our profile. There are, as said before, smectite, mixed layers, mica, kaolinite and chlorite.

The smectite is not present in all the samples. It seems to correspond mainly to micritic samples. If it is present, it composes the greater portion of the clays assemblage (with mica).

Mixed layers (mixture of smectite and mica) seem to have an inverse behaviour. They occur mainly in sparitic samples. Mica are ubiquitous, but show large variations in quantity. Kaolinite and chlorite have the same behaviour. They appear together at the end of the U.I.O., in the facies of internal platform. With the method of the roentgenofacies, we will try to explain these and variations in quantity.

4.2.1 Roentgenofacies (= classes of diffractograms)

As it has been explained in the introduction, the clay minerals have been arranged (like microfacies) in eight groups. An example of a diffractogram is given for the roentgenofacies F (Text-Fig. 9).



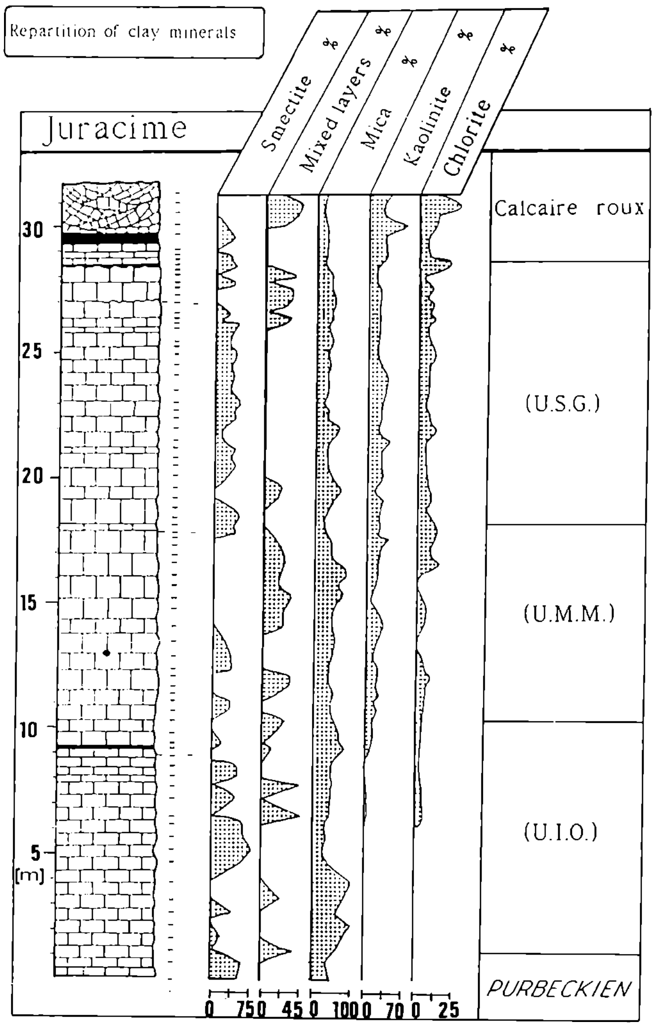
Text-Fig. 9. Typical diffractogram of the roentgenofacies F.

Roentgenofacies A

This class is composed of:

- smectites, generally the major part
- micas
- kaolinite and chlorite, in moderate quantity
- thin quartz with highest content.

For this class, smectite and mica are often badly crystallized; the relation between <2 μm and 2-16 μm fractions shows that thin clays and thin quartz are in greater proportion.



Text-Fig. 10. Repartition of the clay minerals in the Juracime profile.

Roentgenofacies B

This class is composed of:

- mixed layers
- micas, well crystallized
- kaolinite and chlorite with low quantity
- quartz, very seldom.

The main characteristic is the very low intensities of all minerals. This induces an artefact in regard to relative percentage. The intensity of kaolinite does not reflect its true abundance. The relation <2 µm/2-16 µm shows that there are lots of big particles in this class.

Roentgenofacies C

This class is composed of:

- well-crystallized smectite, in variable quantity
- well-crystallized micas
- kaolinite, chlorite and quartz are seldom.

The proportion of thinner clay is greater in this class.

Roentgenofacies C'

This class is composed of:

- mixed layers
- very well crystallized micas
- more kaolinite and chlorite
- quartz is seldom.

The ratio <2 µm/2-16 µm is inversed with regard to C.

Roentgenofacies D

This class has the same content as the group C', but the intensities of the minerals are lower. There is only more kaolinite and chlorite; quartz is almost absent.

Roentgenofacies E

This class is composed of:

- mixed layers but morphologically near of smectites
- micas
- a lot of kaolinite
- chlorite
- a few quartz.

In this class, the quantities of <2 µm and 2-16 µm are almost equal.

Roentgenofacies F

This class is composed of:

- smectites within medium quantity
- well-crystallized micas
- well-crystallized kaolinite in high content
- chlorite
- quartz is seldom.

The thinner fraction is the most important in this class.

Roentgenofacies G

- mixed layers with high content
- the better crystallized micas
- high content of kaolinite
- chlorite
- quartz is common again.

This is the class where kaolinite and mica are the best crystallized. There are more particles sized 2-16 µm than <2 µm in this class. The quartz is absent in the <2 µm fraction and shows a maximum in the 2-16 µm fraction.

4.2.2 Comparison between the evolution curve of microfacies and the evolution curve of roentgenofacies

These curves show the same trends (Text-Fig. 11). The discontinuities correspond always to a peculiar mineralogic assemblage. The method of roentgenofacies is most certainly not as accurate as is the microfacies method, but it reveals the link that exists between mineralogy and environment.

Smectites are especially present in external facies, where hydrodynamics are low (micritic facies). They are present in internal facies, too, but more rarely. If we have sparitic facies (external and internal platform), the mixed layers take the place of the smectites. This fact could be caused by the size of the particles. The analysis of the insoluble residue by Coulter counter (particles counter) shows that the roentgenofacies A is characterized by very thin particles and the roentgenofacies G or D by larger particles. It seems that the external platform is used as a barrier which catches the larger particles. Thus, for example, there is more and better crystallized mica and kaolinite in the internal platform, and less in more external facies. The absence of kaolinite in the U.I.O. (Text-Fig. 10) may have two explanations: there is no kaolinite because of the too external facies or because the sea is transgressing on Purbeckian strata (there is

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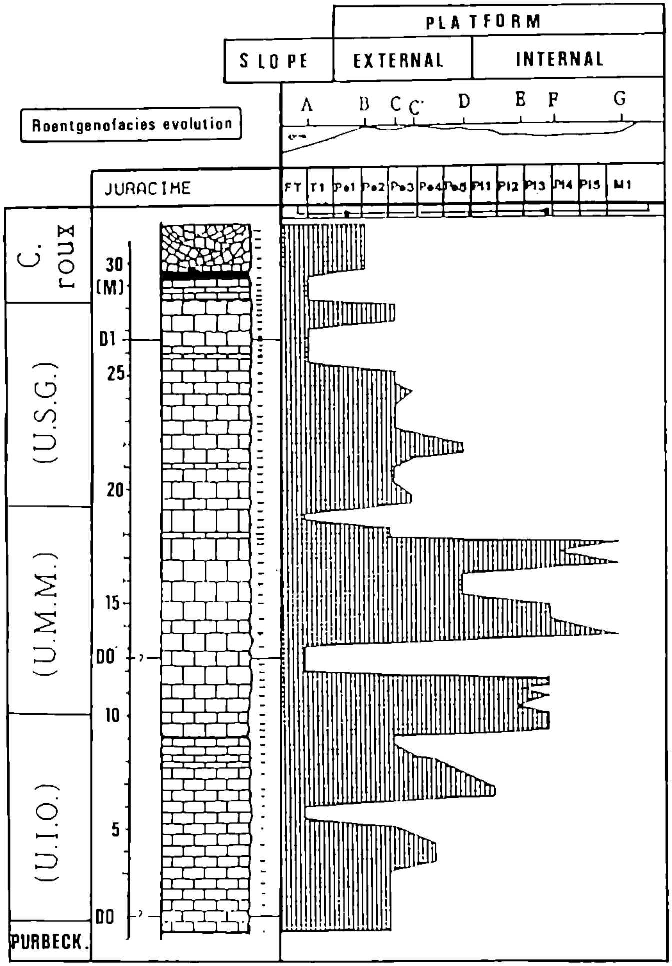
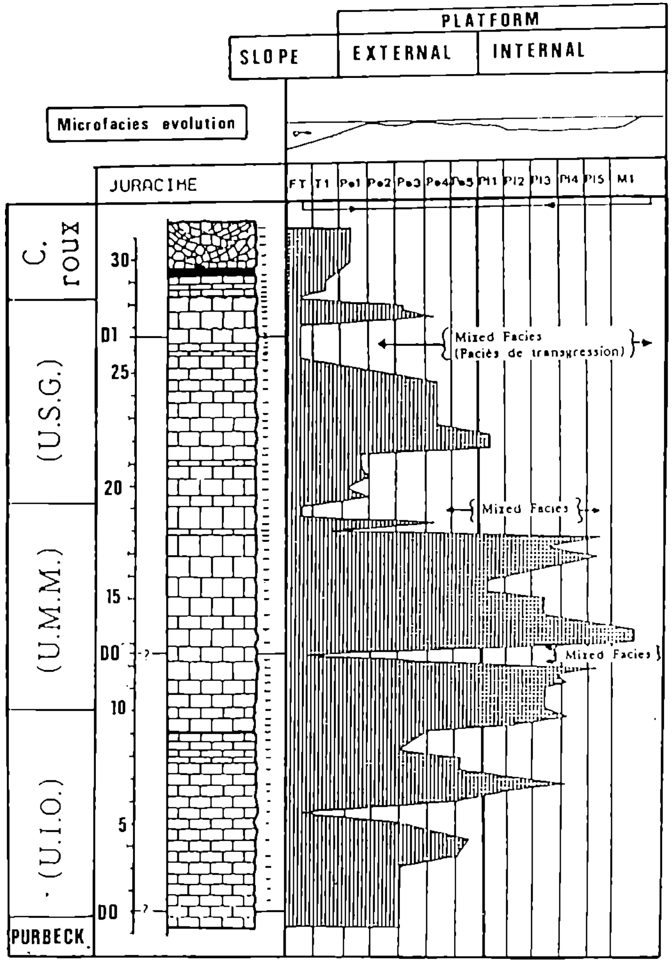
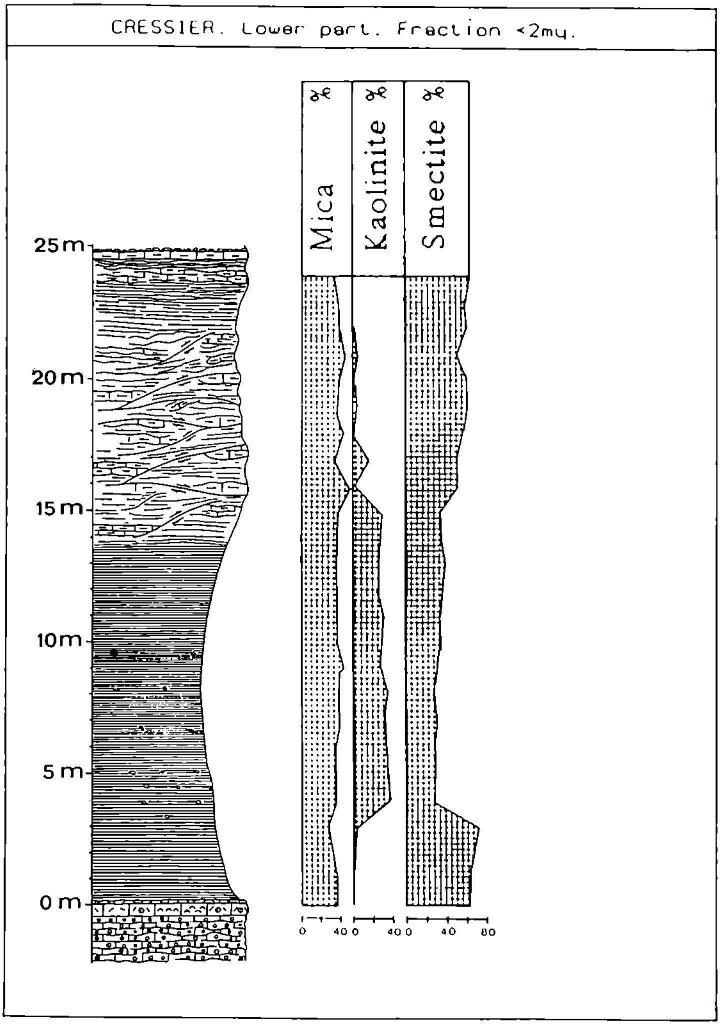
Text-Fig. 11. Comparison between the roentgenofacies evolutionary curve and the microfacies evolutionary curve of the profile of Juracime.


PLATFORM
SLOPE
EXTERNAL
INTERNAL
A B C C' D E F G
Roentgenofacies evolution
JURACIME
FT T1 Pe1 Pe2 Pe3 Pe4 Pe5 Pi1 Pi2 Pi3 Pi4 Pi5 M1
C. roux
(U.S.G.)
(U.M.M.)
(U.I.O.)
PURBECK.
30 (M)
D1
25
20
15
D0'
10
5
D0


PLATFORM
SLOPE
EXTERNAL
INTERNAL
Microfacies evolution
JURACIME
FT T1 Pe1 Pe2 Pe3 Pe4 Pe5 Pi1 Pi2 Pi3 Pi4 Pi5 M1
C. roux
(U.S.G.)
(U.M.M.)
(U.I.O.)
PURBECK
30
D1
25
20
15
D0'
10
5
D0
Mixed Facies
(Faciès de transgression)
Mixed Facies
Mixed Facies

no kaolinite in Purbeckian facies). Regarding other profiles, e. g., the Chambotte (DARSAC 1983), the first explanation could be the best. Moreover, the kaolinite appears in significant quantity, when internal platform facies are attained. The same remarks are valid for chlorite. To find kaolinite (with great content, sometimes) in the mixed facies is not so strange, because these facies show widespread reworking. Micas are relatively constant all along the profile, but are better crystallized in internal facies than in external facies.



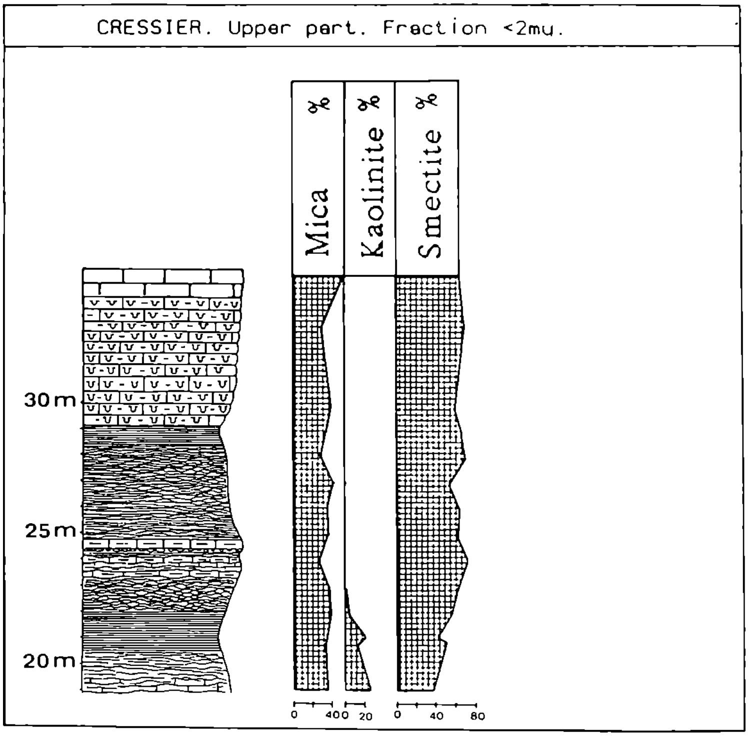
Text-Fig. 12. Repartition of clay minerals in the Cressier profile, lower part.

All these facts prove that it is very dangerous to make some correlations with mineralogy, especially with clays, because they are too much linked with the environment (e. g., kaolinite).

4.3 Hauterivian

Text-Figs. 12 and 13 show the relative percentage for clay minerals of the fraction inferior than 2 microns. The main tendencies are the same as those of the Berriasian-Valanginian. We can make the following remarks:

- The smectites constitute the major part of the mineralogy. They are present in a rather constant ratio all along the profile.
- The detritic micas show the same behaviour.
- The kaolinite appears 4 m above the Calcaire Roux, and is present, in a rather great amount (up to 40 %), along about 10 m, and then disappears almost completely (except 2 or 3 samples). This behaviour is until now unexplained because the presence of kaolinite in such a sedimentological environment is not a common fact.
- The chlorite is very rarely present and always in very small quantities. This behaviour is not significant.



Text-Fig. 13. Repartition of clay minerals in the Cressier profile, upper part.

These data are, of course, not sufficient to give an interpretation of the mineralogy. But, if we consider the mineralogical results of other profiles, we may give a first conclusion.

The presence of smectite as a major constituent of the clay minerals in the micritic facies is a general fact. Higher in the Pierre Jaune, in more energetic environments, the smectite is replaced by mixed-layer minerals. Thus, the distribution of the clay minerals in the sedimentological environment depends certainly on hydrodynamics.

Concerning the presence of kaolinite in the Marnes Bleues d'Hauterive, further investigations are necessary.

The method of roentgenofacies can also be applied for this profile; but, because the variations observed are not significant enough, we think there is not a great interest to present it here.

5. Conclusions

This study shows that the Berriasian-Valanginian as well as the Hauterivian series include important stratigraphic gaps.

The observation of microfacies allows us to distinguish two different sedimentary systems; the Berriasian-Valanginian sediments have typical carbonate platform facies, but this platform seems to disappear in the early Hauterivian. The facies of bioclastic accumulation of the Pierre Jaune de Neuchâtel gives evidence of a connection with the open sea. Finally, the mineralogical study shows clearly the link between mineralogy and sedimentary environment.

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Paleoecology and Environmental Analysis of the Lower Cretaceous Shallow-Marine Drusberg and Schrattenkalk Facies of the Gottesacker Area (Allgäu/Vorarlberg)

DORTE SALOMON, Berlin

With 13 Text-Figures

SALOMON, D. (1989): Paleoecology and Environmental Analysis of the Lower Cretaceous Shallow-Marine Drusberg and Schrattenkalk Facies of the Gottesacker Area (Allgäu/Vorarlberg). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 353-375. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The Urgonian Facies of the Helvetic Alps documents the southward progradation of a carbonate platform during Upper Barremian and Lower Aptian times. Due to this diachronous facies evolution, the sedimentary rocks of the basin (Drusberg-Schichten) interfinger laterally and vertically with those of the platform (Schrattenkalk). Within the marly basin sequences the influence of the platform is indicated by channel-fill sediments.

The Schrattenkalk platform consisted mainly of carbonate shoals interfingering with rudist mounds and coral-sponge bioherms. A quantitative analysis of faunal elements, especially of the microfauna, makes it possible to differentiate several biofacies.

Major environmental changes terminated the platform sedimentation on the Helvetic shelf by the end of Bedoulian time. This event included the development of hardgrounds and the change from benthic to planktonic foraminiferal assemblages. A connection of these occurrences to tectonic reorganizations within the Atlantic Realm is suggested.

Kurzfassung: Die Urgonfazies der Helvetischen Alpen ist vom Oberbarreme bis Unterapt durch eine südwärts progradierende Karbonatplattform gekennzeichnet. Durch diesen diachronen Fazieswechsel verzahnen sich die Sedimentgesteine des Beckens (Drusberg-Schichten) lateral und vertikal mit denen der Plattform (Schrattenkalk). Der Plattform-Einfluß macht sich innerhalb der mergeligen Beckenabfolgen durch Rinnensedimente bemerkbar.

Die Schrattenkalk-Plattform war größtenteils aus Karbonatsandbarren mit eingeschalteten Rudistenmounds und Korallen-Schwamm-Bioherme aufgebaut. Durch eine quantitative Analyse der Fauna, insbesondere der Mikrofauna, ist es möglich, verschiedene Biofazies zu unterscheiden.

Durch einen markanten Fazieswechsel wurde die Plattform-Sedimentation im ausgehenden Bedoule abrupt beendet. Die veränderten Bedingungen führten zur Entwicklung von Hartgründen und zu einem Wechsel von benthonischen zu planktonischen Foraminiferen-Gemeinschaften. Ein Zusammenhang dieser Ereignisse mit tektonischen Umgestaltungen im Atlantischen Raum wird vermutet.

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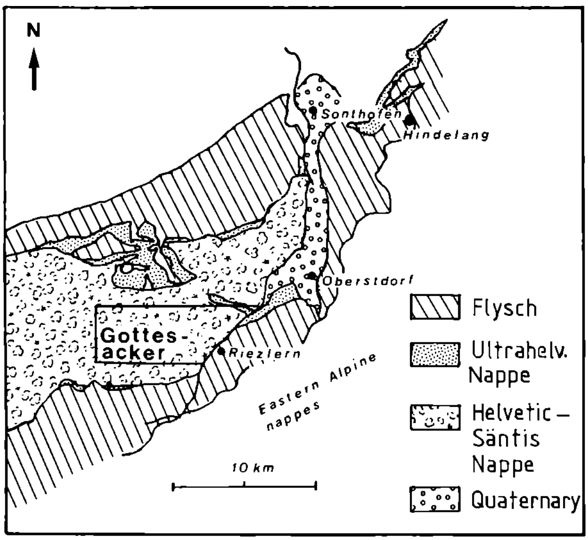
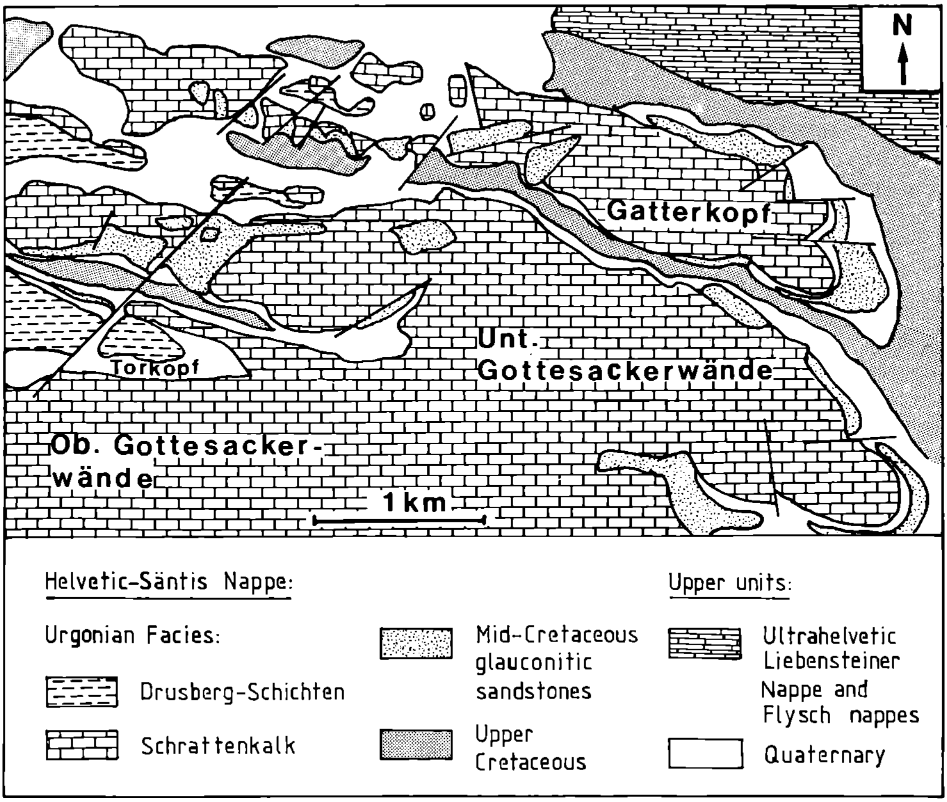
1. Introduction

The sedimentary rocks of the Helvetic Zone represent the northern continental shelf of the Penninic Ocean during Upper Jurassic and Cretaceous times. Today the Helvetic realm comprises a major portion of the Western Swiss Alps and the French Dauphiné Zone. Within the Austrian and Bavarian Alps the Helvetic Säntis-Nappe is exposed as a tectonic window within the Penninic flysch nappes (Text-Fig. 1).



Text-Fig. 1. Geological map of the Allgäu and Vorarlberg area. Modified after HÖFLE (1972). The framed area of the Gottesackerwände is shown in detail in Text-Fig. 2.

During Barremian and Lower Aptian times a widespread carbonate platform of Urgonian facies type developed on the Helvetic shelf (described by ZACHER 1973, FUNK & BRIEGEL 1979, ARNAUD-VANNEAU 1980, SCHOLZ 1984, and others). Generally, the sedimentation patterns of this period document a regressive sequence. The Lower to Middle Barremian is represented by a marly sequence, the so-called **Drusberg-Schichten.** During the Upper Barremian it was replaced by Urgonian shallow-water carbonates of the **Schrattenkalk** sequence. Towards the southern Helvetic realm the Schrattenkalk interfingers with and is ultimately replaced by the marly rocks of the Drusberg type. This vertical and lateral interfingering of Schrattenkalk and Drusberg-Schichten, as documented by several authors (e. g. BRIEGEL 1972 and FUNK & BRIEGEL 1979), is a result of the gradual southwards progradation of the platform during Upper Barremian to Lower Aptian times.



Text-Fig. 2. Geological map of the Gottesackerwände area (in Klein-Walsertal at the German-Austrian border).

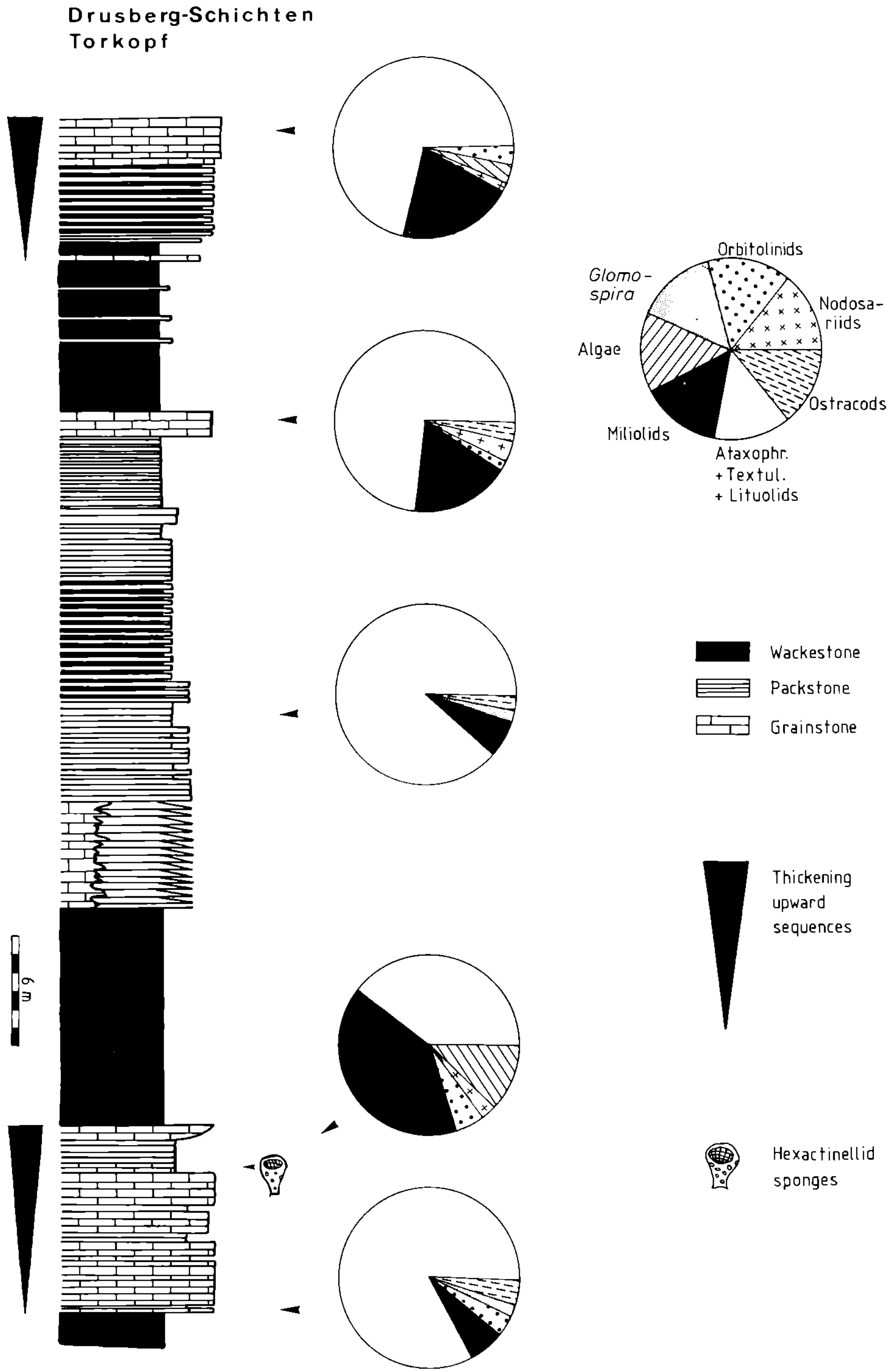
The area of study was the Gottesacker-Wände in Klein-Walsertal at the German-Austrian border (Text-Fig. 1). Within this area the Urgonian Facies is well exposed, and the vertical transition between Schrattenkalk and Drusberg marls can also be studied within the anticlines (Text-Fig. 2).

2. Problems of correlation within the Urgonian Facies

Due to the absence of adequate index fossils, the correlation of sections within the Urgonian Facies is extremely difficult. Large foraminifera, e. g. Orbitolinids are mostly badly preserved, and even in case of good preservation, they only allow a general differentiation between the Barremian and Aptian portions of the Schrattenkalk sequence. For this reason, several attempts at a lithostratigraphical correlation have been made. ZACHER (1973) divided the Upper Schrattenkalk into four E-W extending zones distributed successively from N to S in the Allgäu and Vorarlberg areas. This paleogeographical model was revised by SCHOLZ (1984). However, the models of ZACHER and SCHOLZ both require an approximate isochrony of the erosive top surface of the Schrattenkalk. Taking into account the diachronous facies change of a prograding platform system, every lithofacies correlation between sections seems highly controversial.

Paleogeographic reconstructions in a carbonate platform environment can be done only through the establishment of isochronous surfaces. Such a stratigraphy based on transgressive events has been developed by ARNAUD-VANNEAU (1980, 1986, and others) in order to correlate the platform sediments of the Grenoble area (French Dauphiné Zone) with deposits of the Vocontian Basin. However, this stratigraphic concept can hardly be applied to the Urgonian Facies of the Helvetic Nappes where the lateral transition from platform into basin facies is normally not exposed (except in the southern Helvetic realm). Furthermore, the correlation of platform and basin deposits is not always reliable because of the very different mechanisms of sedimentation within the two environments. A "mixed facies", for instance, which is often used as indicative of a transgressive event, might be a result of storm events on a platform. In a slope environment, a similar "mixed facies" may be caused by debris flows. Still, there would be no logical reason to correlate these two events. Also, discontinuities and abrupt facies changes are not unusual in a carbonate platform environment.

Text-Fig. 3. Profile of the Drusberg-Schichten with a quantitative evaluation of the microfauna. The increase of miliolid foraminifera, algae, and Orbitolinids within the grainstone sequences indicates the platform origin of these sedimentary rocks. In comparison the packstones and wackestones of the Drusberg-Schichten contain about 90 % small arenaceous agglutinating foraminifera.


Drusberg-Schichten
Torkopf
6 m
Orbitolinids
Glomo-
spira
Nodosa-
riids
Algae
Ostracods
Miliolids
Ataxophr.
+ Textul.
+ Lituolids
Wackestone
Packstone
Grainstone
Thickening
upward
sequences
Hexactinellid
sponges

Normally such facies fluctuations, which include changes of microfacies and fauna, like alterations within the benthic foraminiferal assemblages, do not necessarily indicate a major regional event.

For these reasons the correlation problem of the Urgonian Facies is considered to be still unsolved, at least in the Allgäu/Vorarlberg area. Therefore, the facies model presented in this paper cannot aim for paleogeographic accuracy. It merely describes the biological assemblages and lithofacies characteristics of the Schrattenkalk platform and their interaction with the basin facies of the Drusberg-Schichten.

3. Description of facies types

3.1 Drusberg-Schichten

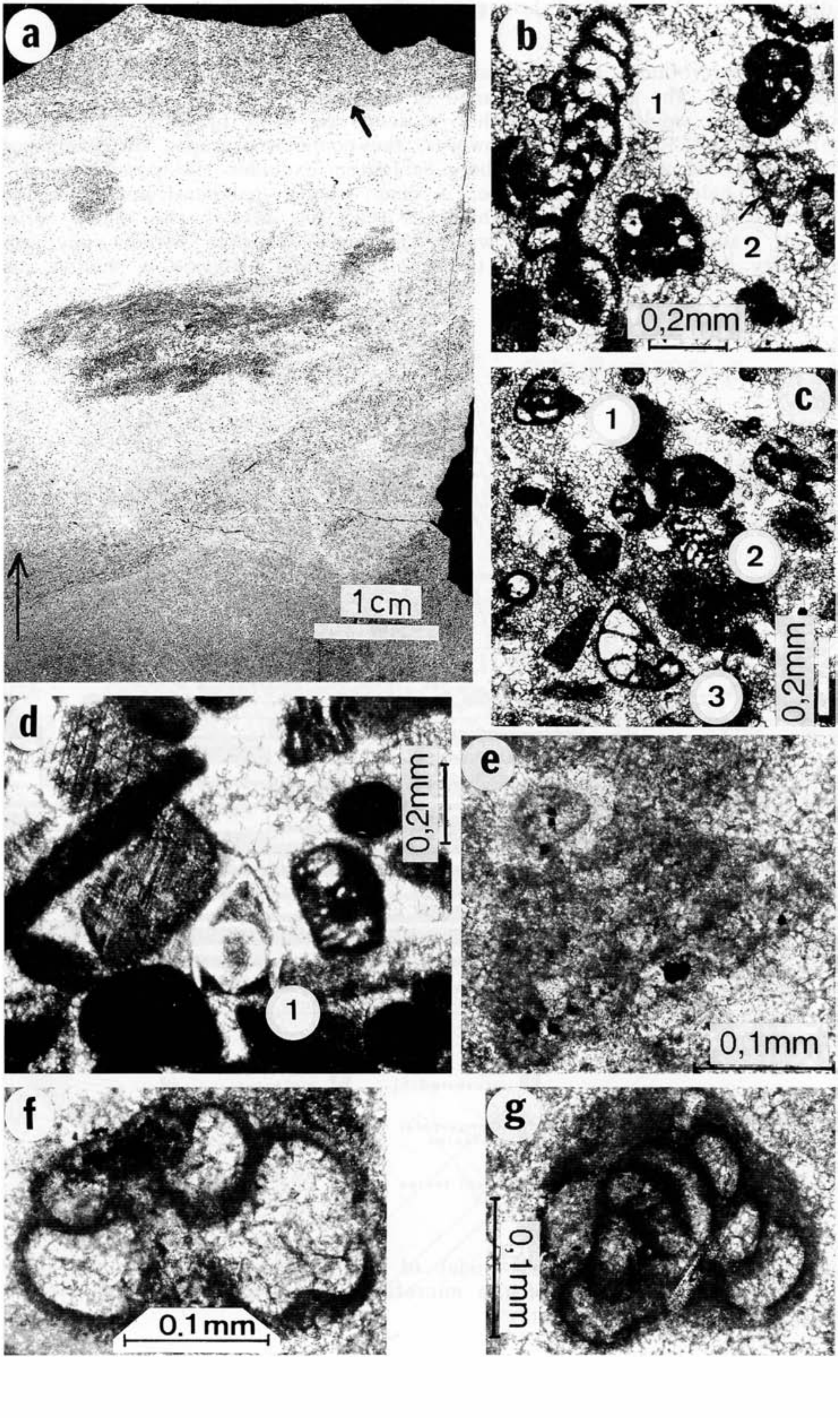
The marly Drusberg facies at the base of the Schrattenkalk platform generally shows an alternation between two different rock types: foraminifera wacke-/packstones and biodetrital grainstones (Text-Fig. 3).

a) Foraminifera wacke-/packstones (Text-Figs. 4a-c, 4e-f)

These are thin-bedded (5 - 15 cm) marly limestones of dark-grey colour, often strongly bioturbated. Microscopically they can be classified as packstones and wackestones containing small benthic foraminifera and sponge spicules. Within some of the packstone layers sedimentary structures can be observed: the base of these beds is erosional, and the upper part is sometimes laminated. They show a "micro-grading" with sponge spicules at the bottom and foraminifera at the top. Such beds are considered to be small turbidite layers (Text-Fig. 4a).

The microfauna of these marly beds is characterized by small arenaceous foraminifera, mainly Ataxophragmiids (especially *Gaudryina, Dorothia, Nezzatinella, Patellovalvulina,* and *Valvulammina*). Also Lituolidae, such as *Haplophragmoides,* the complex form *Sabaudia,* and Ammodiscidae like *Glomospira* are common. The arenaceous foraminifera comprise 80 - 90 % of the microfauna. In comparison, the thin-walled miliolids amount to only 5 - 10 %. Planktonic foraminifera, as described from the Drusberg-Schichten by FUCHS (1971) from the Hohenems area (Vorarlberg), have not been found.

Text-Fig. 4. Thin sections from the Drusberg-Schichten: a) Small turbidite layers within the packstone sequences: one graded bed is cut erosionally by another one (see arrow) (negative print). b) Foraminifera wackestone with *Thomasinella* (?) sp. (1), and *Tritaxia* sp. (2). c) Foraminifera wackestone with miliolid form (1), *Dorothia* (?) sp. (2), and *Nezzatinella* sp. (3). d) Grainstone with *Lenticulina* sp. (1). e) *Sabaudia minuta* (HOFKER). f) *Valvulammina* sp. g) *Glomospira* sp.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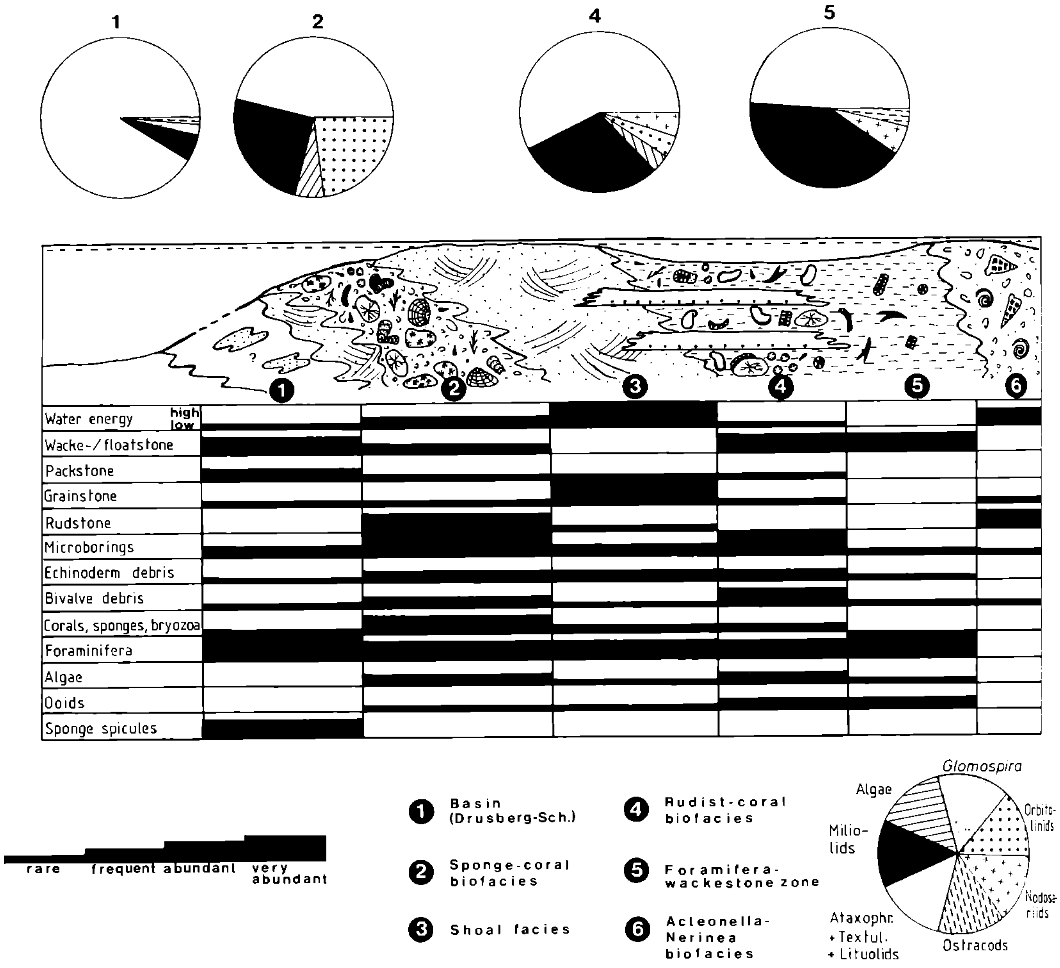
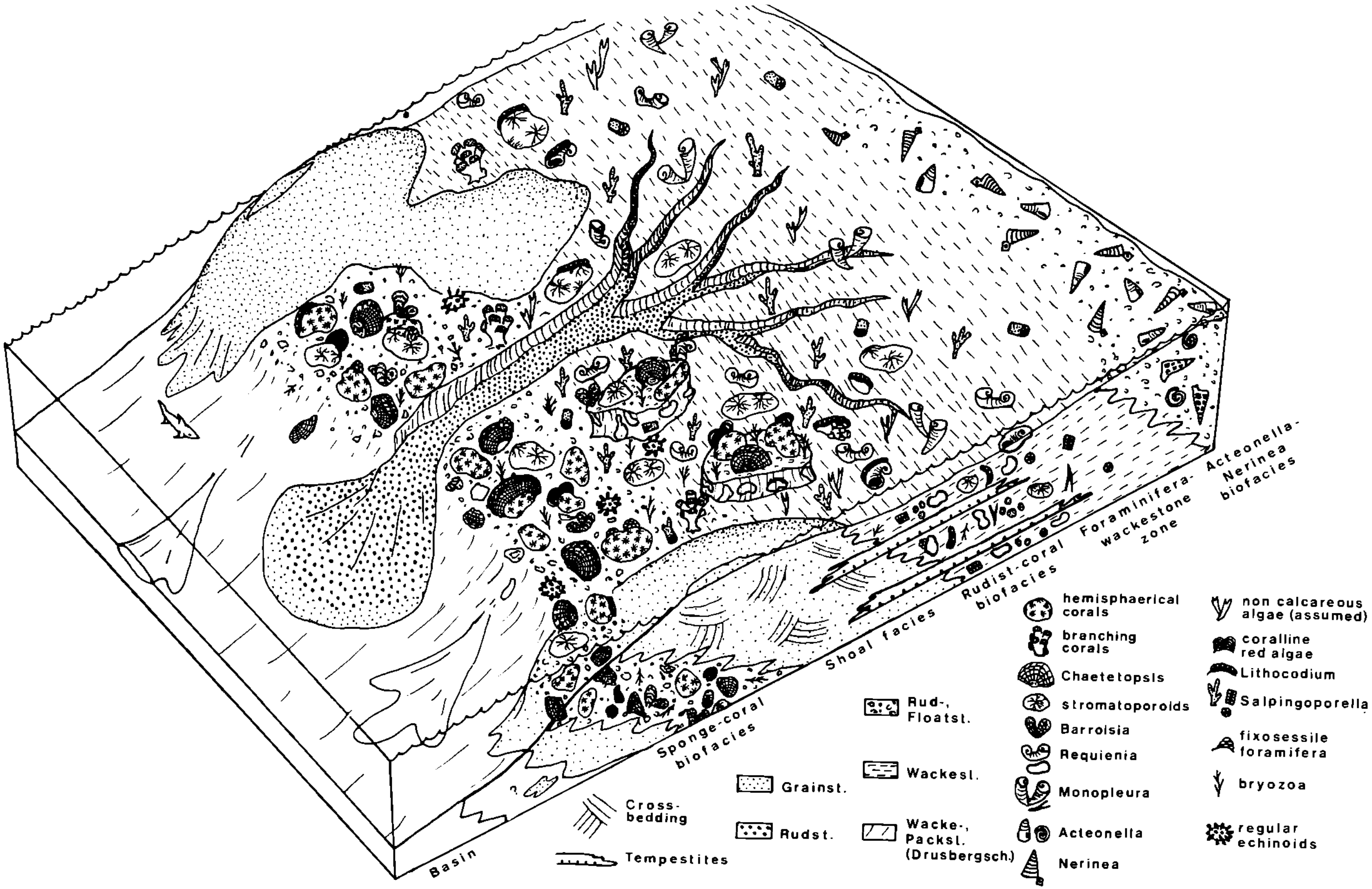
This microfauna cannot easily be compared to Recent foraminiferal assemblages. The strong dominance of agglutinating forms and the absence of Rotaliina would indicate either a deep sea or a brackish lagoonal environment (MURRAY 1973). However, the common presence of brachiopods and echinoids within the Drusberg-Schichten excludes the possibility of a brackish paleoenvironment. Also, a deep basinal paleoenvironment is not probable due to the limited thickness (ca. 100 m) of the marly series, which grade into the shallow water carbonate sequence without any sedimentary gaps. Thus, according to sedimentological evidence, a shallow sub-



Text-Fig. 5. Hypothetical facies model of Schrattenkalk and Drusberg-Schichten with a quantitative evaluation of the microfauna and some microfacial parameters. For further explanation, see text.



Text-Fig. 6. Hypothetical facies model of the Helvetic Urgonian Platform. For further details, see text.

tidal foreshore paleoenvironment is postulated. The presence of *Quinqueloculina* and similar miliolid foraminifera also discounts a deep sea paleoenvironment where *Pyrgo* is normally the only miliolid form (MURRAY 1973).

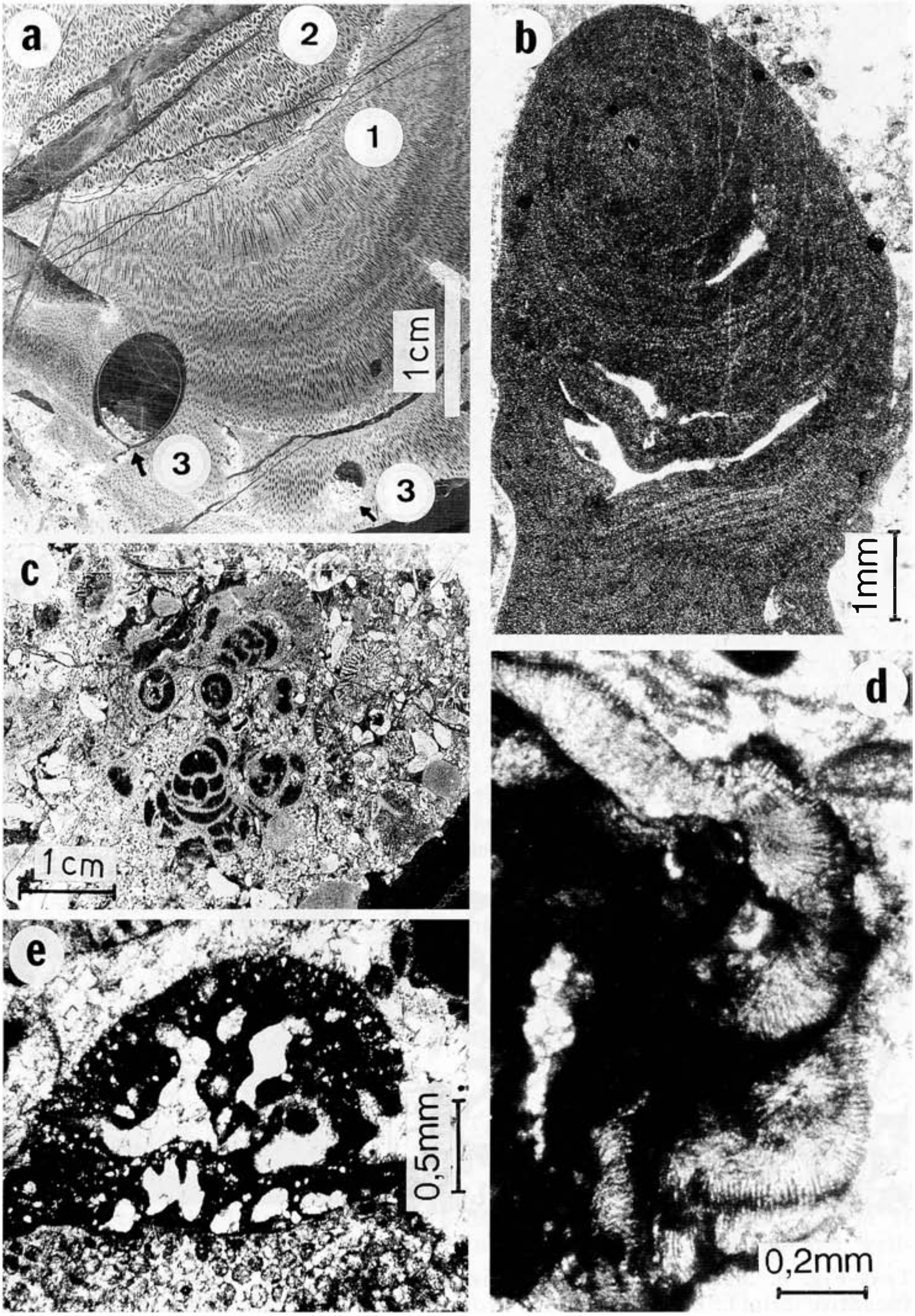
It appears that some shallow subtidal and platform biotopes now occupied by Rotaliina, e. g. Eponididae, Amphisteginidae, Rotaliacea, Elphidiidae, and Discorbidae (ROSE & LIDS 1977), were dominated by Ataxophragmiids during Lower Cretaceous time. This problematic dominance of arenaceous foraminifera may be due to unknown paleoecological circumstances (see the ecological studies of WEIDICH 1984). However, also a selective diagenetic solution of calcareous foraminifera (BERGER 1979) must be considered to explain the dominance of arenaceous forms. This especially applies to the Robertinacea, because of their aragonitic tests (BRASIER 1980).

b) Biodetrital grainstones (Text-Fig. 4d)

Typically, these grainstones are found as thickening upward sequences of well-bedded, light grey limestone layers of varying thickness (10 cm to about 2 m). A lateral pinching out and interfingering of these grainstone sequences with the packstone layers can commonly be observed. Their microfacies differs significantly from that of the marly series. They consist of well-sorted grainstones with abundant bryozoa and echinoderm debris; commonly they also contain ooids and green algae fragments, especially *Bouenia pygmea*. The microfauna is characterized by orbitolinids and by a larger amount (about 40 %) of miliolid foraminifera. This is a typical association of the Helvetic Urgonian Platform as described by ARNAUD-VANNEAU (1980). Because of this similarity, the light grey limestone sequences are interpreted as channel fill sediments linked with the carbonate platform. They document the progradation of the Schrattenkalk platform into the shallow basin of the Drusberg-Schichten.

This evidence suggests that a carbonate platform must have existed contemporaneously with the deposition of the Drusberg-Schichten. Therefore, it is necessary to incorporate the Drusberg-Schichten into the platform facies model.

Text-Fig. 7. Fossils of the sponge-coral biofacies: a) *Chaetetopsis favrei* (DENINGER) (1) encrusting on a coral (*Stylina* sp.) (2). Geopetal sediment within the boring mollusks (3) is opposite to the growth direction of *Chaetetopsis*, showing that the fossils were not in living position. (negative print). b) *Archaeolithothamnium* sp. c) Rudstone with *Barroisia* cf. *helvetica* (DE LORIOL) (negative print). d) *Ethelia alba* PFENDER. e) *Coscinophragma* cf. *cribosum* (REUSS).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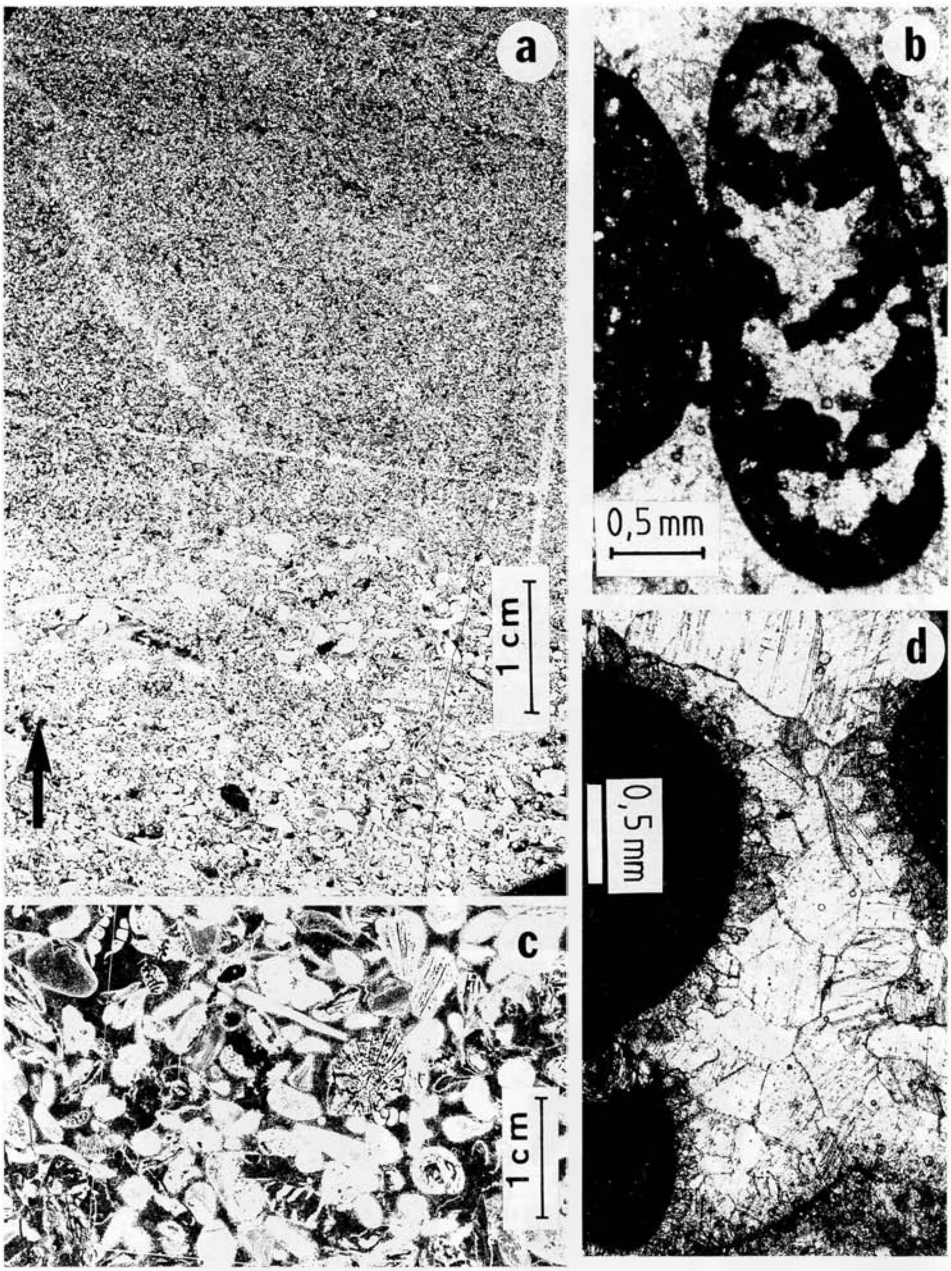


Text-Fig. 8. Shoal and channel facies: a) Graded bed, probably storm layer (negative print). b) *Reophax* sp. from shoal facies. c) Rudstone interpreted as lag sediment. Finer grain sizes are missing (negative print). d) Detailed close-up of c) showing the consolidating "dog tooth cements".

3.2 Schrattenkalk

For the analysis of the Schrattenkalk platform, a quantitative evaluation of the faunas and the microfacies types was performed. This analysis identified several characteristic biological assemblages and corresponding microfacies which were integrated into a hypothetical facies model (Text-Figs. 5, 6). The biological assemblages of micro- and macrofauna, as well as the corresponding microfacies types, confirm the results achieved by ARNAUD-VANNEAU (1980) from the Grenoble area.

However, as it has not yet been possible to develop an adequate stratigraphy in order to correlate different Schrattenkalk sections (as explained in paragraph 2.), accurate paleogeographic reconstructions are not possible.

The main facies types are: 1) basin (Drusberg-Schichten), 2) sponge-coral biofacies, 3) shoal and channel facies, 4) rudist-coral biofacies, 5) foraminifera-wackestone facies, 6) *Acteonella-Nerinea* facies. These will be described in the following sections.

3.2.1 Sponge-coral biofacies (Text-Fig. 7)

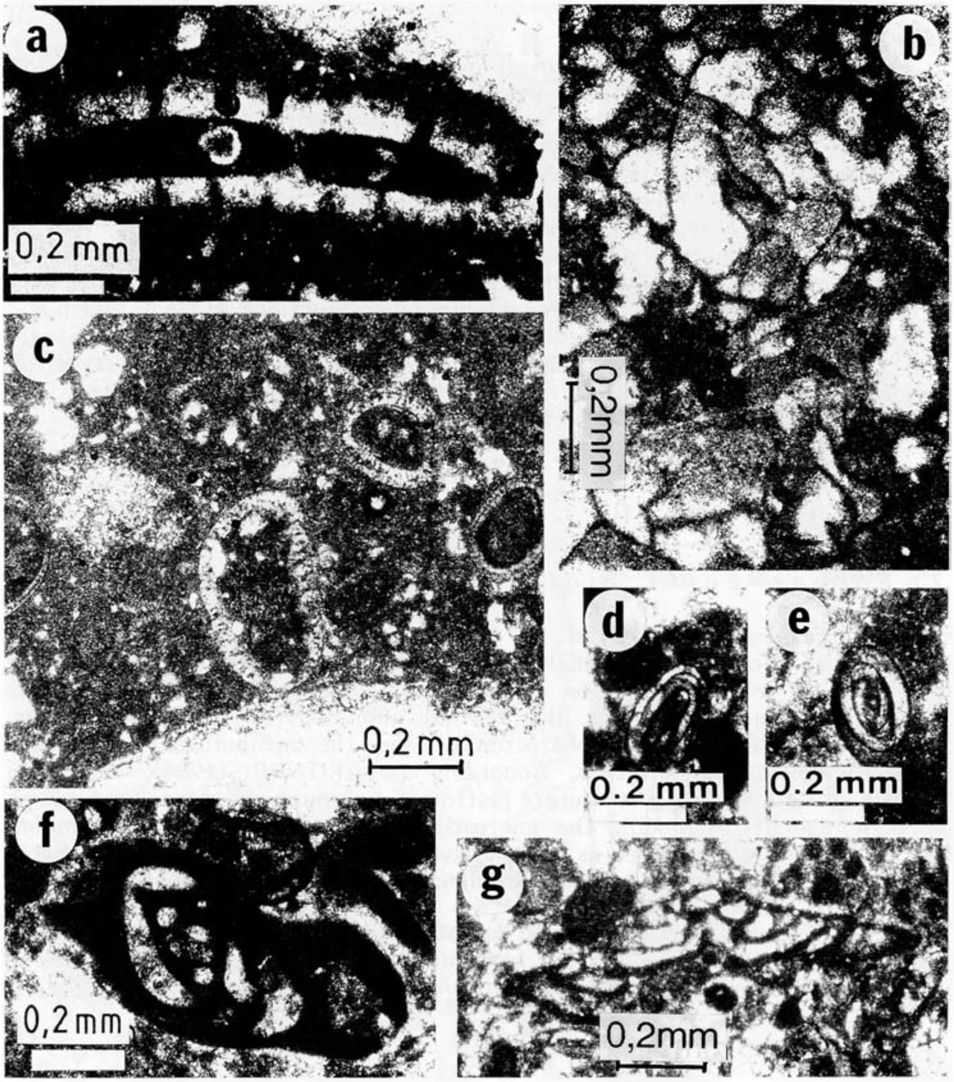
This biofacies is characterized by the maximum diversity of orgânisms (Text-Fig. 7). Fossils are normally parautochthonous, sometimes corals are even found in a living position (boundstones). Representative macrofossils include hemisphaerical corals like *Stylina* and *Eugyra*, coralline sponges, such as *Chaetetopsis favrei*, pharetronids, and the sphinctozoid calcareous sponge *Barroisia* cf. *helvetica*. According to REITNER (1984), the form *Barroisia* is limited to the outer platform environment. Algae exhibit a very high diversity, including the encrusting red alga *Archaeolithothamnium*. This encrusting alga indicates that wave resistant patch reefs probably existed. Other forms such as *Ethelia alba*, "*Bacinella*", *Salpingoporella*, and *Cayeuxia* are also common.

The microfauna is characterized by a maximum of orbitolinids in association• with large *Reophax*-type foraminifera, miliolids with heavy tests, and encrusting foraminifera like *Coscinophragma criposum*.

3.2.2 Shoal and channel facies (Text-Fig. 8)

The well-bedded, cross-stratified shoal facies comprises the major part of the Schrattenkalk. Thickness of the beds is very variable (about 10 cm to 2 m). Microfacially these rocks are grainstones with well-rounded and well-sorted components.

Cross-stratified beds are sometimes erosionally cut and replaced by graded layers (mostly of 30 to 50 cm thickness), which are interpreted as storm deposits (Text-Fig. 8a). Within such graded layers, it is possible to test the control of grain size on the distribution of microfossils.



Text-Fig. 9. Internal Rudist-coral and Foraminifera-wackestone biofacies: a) *Salpingoporella* cf. *muehlbergii* (LORENZ).- b) *Bacinella irregularis* ELLIOT. c) Wackestone with single layered, radial fibrous "stillwater ooids". d)-e) *Pseudotriloculina* sp. f) Miliolid foraminifera with heavy test. g) *Patellovalvulina* sp.

It was observed that the ratio of miliolids to arenaceous foraminifera hardly changed from the bottom to the top of such beds, whereas orbitolinids accumulated at the bottom. Thus, when applied to the well-defined environment of a carbonate platform (as done to the Bahama Platform by ROSE & LIDS (1977) and to the Helvetic Urgonian Facies by ARNAUD-VANNEAU (1980) and in this report) the miliolid abundance seems to be a reliable criterion for differentiating between restricted and open environments.

Banks of rudstones (thickness about 50 cm) consolidated by radial "dog tooth cements" are also present (Text-Fig. 8c-d). Finer grain sizes are missing within these beds, and they are interpreted as lag sediments, which would most likely be deposited in tidal channels located between the shoal-areas. HALLEY et al. (1983) described from the Bahama platform such 1-3 km wide tidal channels with a maximum depth of only 2-7 m. Identifying such channels in a fossil record ought to be difficult because of these proportions, and this might be the reason why they have never been described from the Helvetic Schrattenkalk.

3.2.3 Rudist-coral biofacies (Text-Fig. 9)

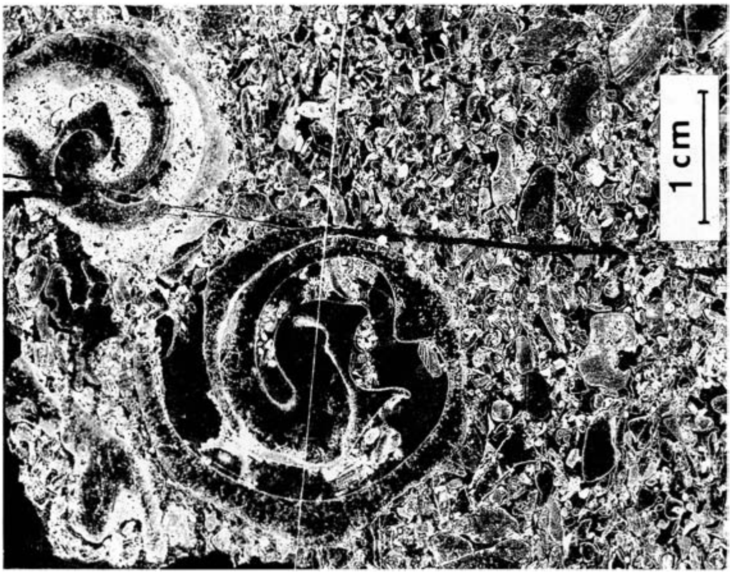
Within this facies mainly float- and wackestones occur indicating low-energy conditions. The diversity of organisms has decreased remarkably. Requinid rudists in association with branching corals and stromatoporoid sponges comprise the major part of the macrofauna. Algae are represented only by the problematic blue-green algae *Bacinella irregularis, Girvanella,* and a few dasycladaceans (mainly *Salpingoporella muehlbergii* and *Salpingoporella hasi*) (Text-Fig. 9a-b). The microfauna is similar to that of the Foraminifera-wackestone facies.

3.2.4 Foraminifera-wackestone facies (Text-Fig. 9)

Faunal diversity is very low within this biofacies. Macrofauna mainly consists of monopleurid rudists (*Argriopleura*) and the algae are represented exclusively by *Salpingoporella*. Calcareous foraminifera comprise about 50 % of the microfauna. Among the larger foraminifera, the form *Orbitolinopsis* is dominant compared to *Palorbitolina*. Ooids are abundant in all thin sections of these wackestones. These ooids are always of the radial fibrous, single layered type (Text-Fig. 9c). According to CAROZZI (1960) and SANDBERG (1975), they are considered to be stillwater ooids characteristic of an environment of slightly increased salinity. The foraminifera-wackestone biofacies probably represents the most restricted parts of the platform. A pure *Pseudotriloculina* facies, as described by ARNAUD-VANNEAU (1980) from the French Dauphiné Zone, has not yet been observed, although *Pseudotriloculina*-morphotype miliolids do occur (Text-Fig. 9d-e).

3.2.5 *Acteonella-Nerinea* facies (Text-Fig. 10)

This facies is characterized by rudstones in beds of 0.3-0.5 m thickness with an accumulation of large gastropods (Text-Fig. 10). These rudstones are always strongly affected by carbonate solution. Due to bad preservation, a quantitative evaluation of faunas was not possible.



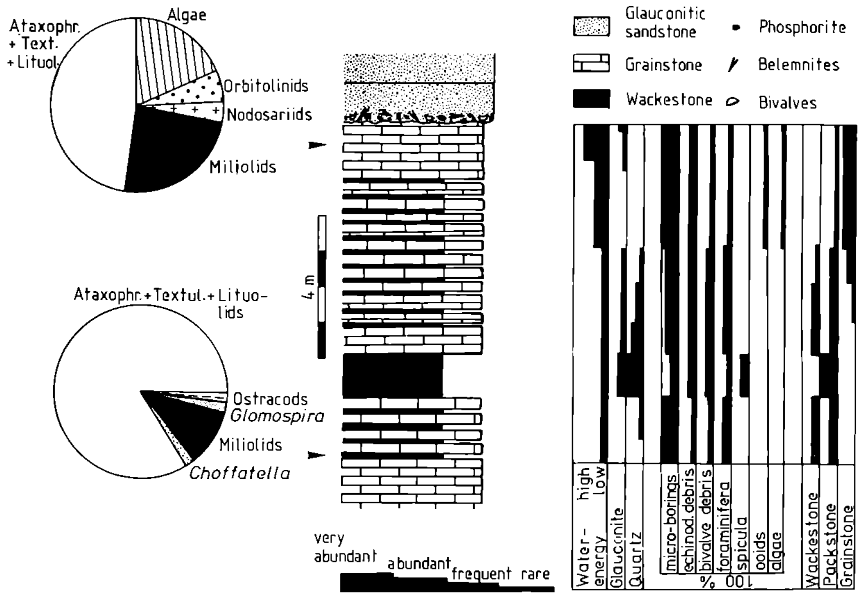
Text-Fig. 10. *Acteonella-Nerinea* rudstone (negative print).

4. The extinction of the Helvetic Urgonian Facies

4.1 **"Schönhaldenkopf-Facies"** (Text-Figs. 11, 13)

In the Gottesackerwände area, the uppermost Schrattenkalk is normally represented by a marly facies containing glauconite, quartz, and numerous orbitolinids. These beds, called "Schönhaldenkopf-Schichten" (LIEDHOLZ 1959), correspond to the "Couche à Orbitolines" of the Dauphiné Zone

(ARNAUD-VANNEAU 1980). It is a shallowing upward sequence as documented by microfacies and foraminiferal assemblages (Text-Fig. 11). This sequence probably represents the infilling of smaller lagoons, perhaps produced as a result of synsedimentary tectonics during the Lower Aptian.

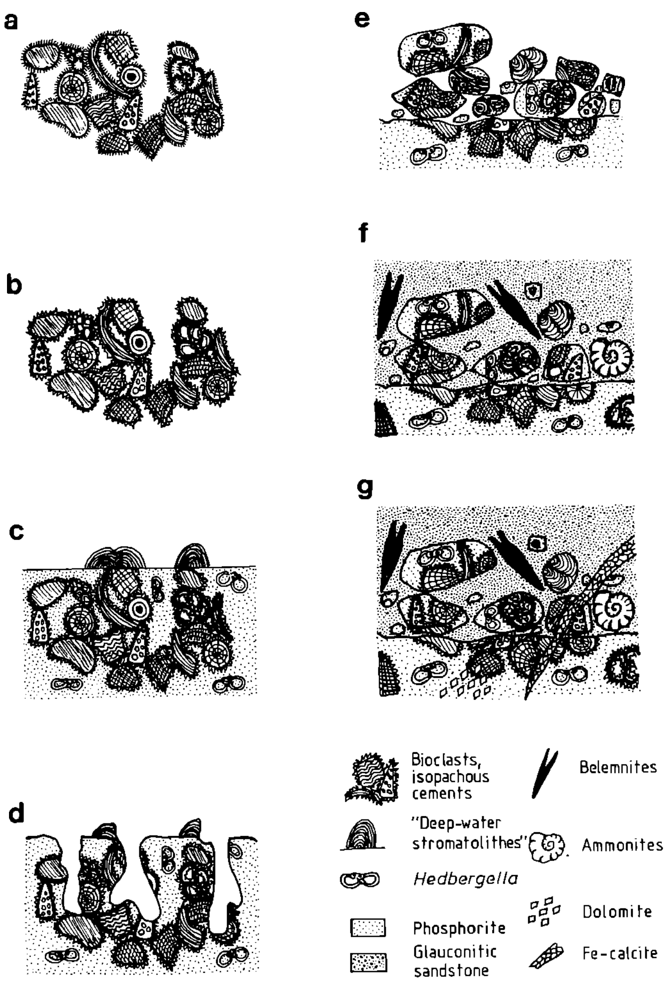


Text-Fig. 11. Representative profile of the uppermost Schrattenkalk, the so-called "Schönhaldenkopf-Schichten". Microfauna and microfacial parameters indicate a shallowing upward sequence.

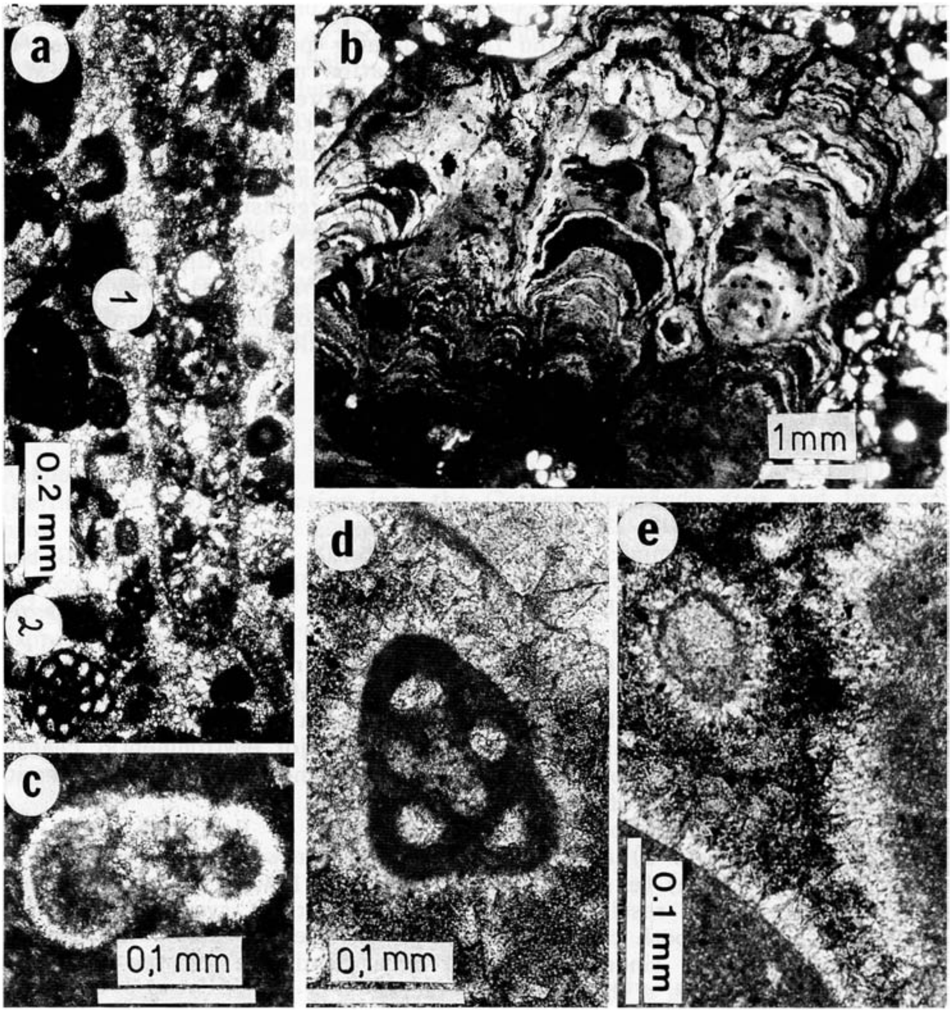
4.2 The terminating hardground

The uppermost Schrattenkalk bed is terminated by a hardground marking a major stratigraphic gap. Above this Urgonian sequence, deltaic glauconitic sandstones ("Brisi Sandstone") of Clansayes and Albian ages follow. Normally the terminating hardground is totally reworked, but in some sections autochthonous relicts can be found. This makes it possible to reconstruct the development of the phosphoritic horizon (Text-Fig. 12):

1) The sediment of the Upper Schrattenkalk was consolidated by early diagenetic marine phreatic cements.
2) These cements were slightly altered into bladed calcite cements (Text-Fig. 13d-e) due to a possible short-term meteoric influence.



Text-Fig. 12. Development of the hardground terminating the Urgonian Facies. a) Sediment of the uppermost Schrattenkalk was consolidated probably by radial fibrous, marine phreatic cements. b) These cements were slightly altered into bladed calcite cements, and c) the remaining pore space was filled by a phosphoritic matrix containing planktonic foraminifera. d-e) The hardground was colonized, bioturbated, and reworked. f) Finally the phosphorite pebbles were redeposited in a glauconitic sandstone together with cephalopods of Clansayesian age. g) The so-called "Luiterezug Fossilschicht" as it is found today. Sometimes autochthonous relicts of the terminating hardground can be observed under the layer of phosphorite pebbles. This figure is not true to scale. For better illustration, the planktonic foraminifera had to be magnified compared to the macrofossils.



Text-Fig. 13. a) Lower part of the "Schönhaldenkopf-Schichten" with *Choffatella decipiens* SCHLUMBERGER (1) and *Haplophragmoides* sp. b) Reworked phosphoritic "deep water stromatolite" from the "Luiterezug-Fossilschicht". c) *Hedbergella* sp. in phosphoritic matrix of the terminating hardground. d)-e) Bioclasts of the uppermost Schrattenkalk bed consolidated by radial bladed calcite cements. The pore space between the components is filled with a phosphoritic matrix.

3) Due to a major environmental change at the end of Bedoulian time, a phosphatization occurred and the remaining pore space was filled up by a phosphoritic matrix. This phosphatization must have occurred at a rather early stage of diagenesis, as the pore space was still open. Thus, hardground development is closely connected in time with the end of the carbonate sedimentation. Within the phosphoritic matrix, for the first time in this area, planktonic foraminifera appeared (Text-Fig. 13c).
4)-5) The hardground was bioturbated by boring organisms and reworked.
6) Finally the phosphorite pebbles were redeposited in a glauconitic sandstone together with non-phosphatized cephalopods of Clansayes age (e. g. *Nolaniceras nolani* and *Neohibolites* aff. *strombecki*). This sediment of reworked phosphoritic material and cephalopods is normally called the "Luiterezug-Fossilschicht" (JACOB & TOBLER 1906).

5. Conclusions

The youngest rocks of the Urgonian Facies described from the Helvetic Alps are of Lower Aptian age. Within the entire Helvetic realm the Urgonian platform with its highly diverse benthic fauna was terminated at the end of Bedoulian time. It is assumed that some major environmental changes caused this facies extinction. This event included the genesis of phosphorite and the first appearance of planktonic foraminifera.

To explain these abrupt changes at the end of Bedoulian time, two theories are proposed: 1) a major transgression (drowned platforms due to eustatic sea level rise) (e. g. ARNAUD-VANNEAU 1980, 1986), 2) influence of cooler waters as a result of changed oceanic current systems (BERGNER et al. 1982). Maybe both circumstances were involved. A major transgression is doubtful, however, since the deltaic Brisi Sandstone of the uppermost Aptian still documents a shallow water facies.

The Bedoulian/Gargasian Boundary Event certainly brought about widespread environmental changes. These changes can be documented on the Iberian Plate as well: In the Bay of Biscay an important faunal extinction occurred due to the influx of cooler waters (REITNER 1984). The Urgonian platform development was terminated by the end of Bedoulian time except for some isolated islands on the southern Biscay shelf, which were protected by the "Biscay High". At the same time the sedimentation in the basin changed to black shale deposits (GRACIANSKY et al. 1979). These events coincide with the first spreading of the North Atlantic Ocean as calculated by MONTANDERT et al. (1979) by magnetic anomaly fits.

The Bedoulian/Gargasian events appear to have had a strong influence mainly within the northern and western Mediterranean realm. The development of the Urgonian facies of the West Carpathian Tatric Zone is similar to the Schrattenkalk, and, at least within the Manin unit, platform development stopped in Lower Aptian time (CSASZAR et al., this vol.). This cessation of Urgonian facies and the development of phosphatic hardgrounds in the Bay of Biscay and within the Helvetic and West Carpathian regions may indicate at least a temporary connection of these realms, maybe through the Vocontian Basin. The Barremian-Albian within the southern Tethyal realm, e. g. in Hungary, differs from the Alpine-Carpathian development: Within the Villány Mts. Zone the platform sedimentation seems to be continuous throughout Aptian times (Nagyharsány Limestone Fm.), and within the Transdanubian Central Range Zone real Urgonian sedimentation was not established until the Albian (Környe and Zirc Limestone Fms.) (CSASZAR 1984, 1986, and CSASZAR & HAAS 1979).

Thus the platform extinctions at the Bedoulian/Gargasian boundary may be, at least to some degree, related with transformations within the Atlantic Realm. Probably these changes were linked with tectonic events during the first stages of spreading in the North Atlantic Ocean.

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The Schrattenkalk of Vorarlberg: An Example of Urgonian Sedimentation

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With 12 Text-Figures and 2 Tables

CSASZAR, G., OBERHAUSER, R. & LOBITZER, H. (1989): The Schrattenkalk of Vorarlberg: An Example of Urgonian Sedimentation. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 377-401. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Two characteristic profiles of Urgonian sedimentation from the Helvetic Zone of Vorarlberg have been studied in detail and compared with other localities of "Schrattenkalk"-type sediments. The road cuts at Ill-Gorge in Feldkirch, on the one hand, are considered to represent sediments which were redeposited from a carbonate platform towards a gentle inclined foreslope respectively outer shelf environment. The Schrattenkalk of Rhomberg quarry at Unterklien, on the other hand, shows the diversified environments of a typical carbonate platform sedimentation. The arrangement of facies belts within the Urgonian of the Helvetic Zone suggests the existence of a basin which separated the platform from the shore. The termination of Schrattenkalk deposition is explained by the probable establishment of a connection with the Boreal Realm.

The Urgonian facies of various East Alpine zones is reviewed. The paleogeographic reconstruction reveals, that the Tatric Zone - including the Urgonian formations of the West Carpathians - corresponds to the Central Penninic Ridge. Its central part represents the source area of the Tristelschichten, which were deposited in the peripheral parts of the Central Penninic Ridge. The northern Helvetic shelf, which is attached to the deep-sea Penninic Realm, shows Schrattenkalk sedimentation. The southern East Alpine shelf of the Thiersee Syncline is characterized by Urgonian facies as well (HAGN 1982).

Kurzfassung: Zwei Profile aus der typischen Urgon-Entwicklung von Vorarlberg werden im Detail beschrieben und mit anderen Gebieten mit Sedimentation vom Typ "Schrattenkalk" verglichen. Das Profil der Illschlucht in Feldkirch wird als eine von einem inneren Schelfbereich auf einen flachen Plattformabhang bzw. in den äußeren neritischen Bereich umgelagerte Bildung betrachtet. Der Schrattenkalk des Rhomberg-Steinbruchs in Unterklien bei Dornbirn wird hingegen einer typischen Urgon-Karbonatplattform zugeordnet. Das Ausklingen der Schrattenkalk-Sedimentation wird mit der Öffnung zum borealen Ablagerungsraum in Verbindung gebracht.

Ferner wird ein Überblick über die Urgonentwicklung verschiedener ostalpiner Zonen gegeben. Die Urgonbildungen der Westkarpaten mit dem Tatrikum werden als Äquivalente des Mittelpenninischen Rückens interpretiert, der im zentralen Abschnitt als Abtragungsgebiet und im peripheren Teil als Sedimentationsraum für die Tristelschichten diente. Zur gleichen Zeit bildet sich im Bereich des helvetischen Schelfgebietes, das an das tiefmarine Penninikum im Norden anschließt, Schrattenkalk und gleichzeitig am südlich des Penninikum-Troges gelegenen ostalpinen Schelfbereich - z. B. in der Thierseer Mulde in Tirol (HAGN 1982) - eine analoge Urgonfazies aus.

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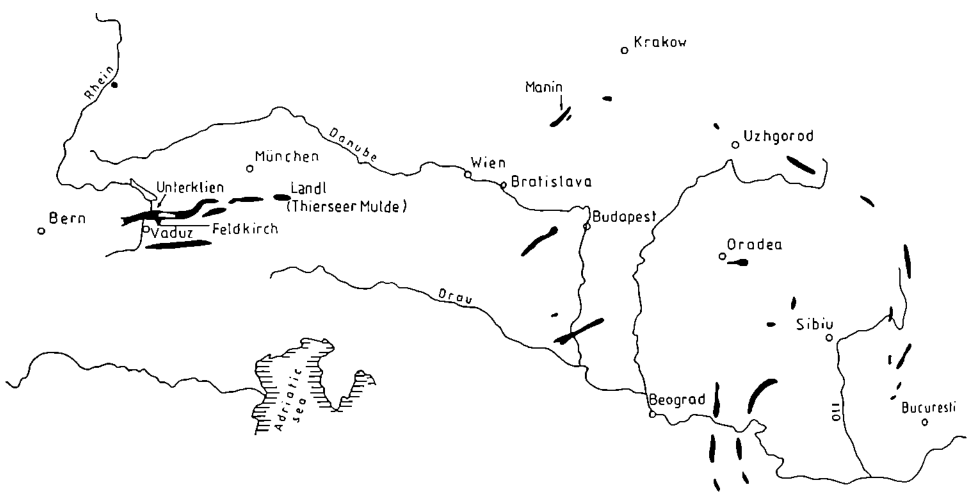
1. Introduction

After completing part of the synthesis of the widespread and diversified Hungarian Urgonian deposits the necessity for a comparison with the adjacent regions became evident. That was the reason for G. CSASZAR to start fieldwork on the Schrattenkalk in western Austria together with H. LOBITZER and later on the Urgonian in Slovakia (Text-Fig. 1). R. OBER-



Text-Fig. 1. Urgonian occurrences in the East Alpine - Carpathian system.

HAUSER contributed his knowledge of regional geology and stratigraphy of Vorarlberg and Tyrol.

The present work is mainly based on a detailed study of the sections of Rhomberg quarry at Unterklien and of the road cut at Ill-Gorge/Feldkirch. These sections show different types of Urgonian sedimentation. The understanding of spatial relationships and paleoecology was made easier by a comparative study of several other localities of Urgonian facies in Vorarlberg and their respective relations to the under- and overlying formations.

2. History of Research

The history of Urgonian research goes back to the pioneer time of Alpine geology (D'ORBIGNY 1847). The nice landscape along with richness of fossils in the Western Alps attracted earth scientists up to the present time. GÜMBEL (1861) subdivided the Urgonian sequence into three units, namely the basal "*Caprotina ammonea* group", the "bryozoan group" in the middle and the "*Orbitolina lenticularis* group" at the top. KAUFMANN (1867) distinguished "Unterer Schrattenkalk", "Orbitolinenschichten" and "Oberer Schrattenkalk".

The first detailed lithological and paleontological study of the Cretaceous formations of Vorarlberg - including also a thorough description of the Schrattenkalk - has been performed by VACEK (1879). His faunal lists of the *Ostrea*-bearing Lower Urgonian member of the Klien locality comprises seven taxa of pelecypods, six of brachiopods, three of echinoids and five taxa of each hydrozoans and corals. From another outcrop there he listed four rudist and one brachiopod species.

At Unterklien, HEIM & BAUMBERGER (1933) also distinguished five members within the sequence. A 7 m thick *Alectryonia* bed and an *Orbitolina* bed are furthermore mentioned. In the Feldkirch profile they distinguished a transitional unit to the underlying Drusberg Beds (25-30 m) and a 30 m thick Schrattenkalk unit.

The work by LIENERT (1965) pioneers the *Orbitolina*- and microfacies studies of the Säntis Urgonian, respectively Drusbergschichten. The basic study by ZACHER (1973) comprises sedimentological and (micro)paleontological data of a large number of well documented profiles and led to an impressive paleogeographic and tectonic synthesis. FELBER & WYSSLING (1979) introduced the name Mittagspitz Formation, which they consider as a southern basinal facies equivalent of the Schrattenkalk.

The synthesis and paleogeographical reconstruction of the underlying formations was carried out by WYSSLING (1986), and that of the overlying formations by FÖLLMI (1986). The Urgonian of eastern Switzerland was monographed by BRIEGEL (1972) and by FUNK & BRIEGEL (1979).

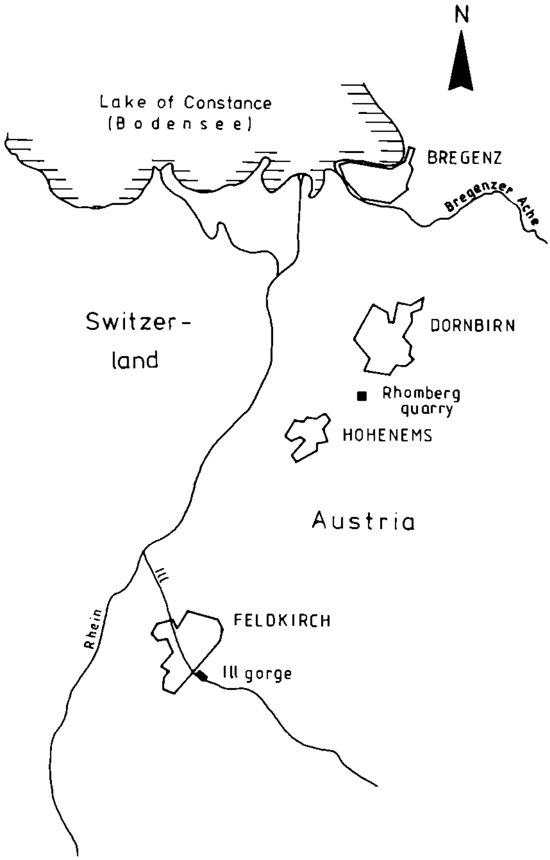
H. SCHOLZ (1984) provides a paleontological synopsis of the easternmost occurrences of Schrattenkalk in the Allgäu region. Special attention is given to the genuine bioherms.

During the last decade manifold new results on Urgonian facies and stratigraphy from different regions have accumulated, which led to stimulating paleogeographical reconstructions (ARNAUD-VANNEAU 1979, PEYBERNES 1979, ARNAUD 1981, SALOMON 1987, a. o.).

3. Geological Setting

The Schrattenkalk is one of the conspicuous northernmost formations of Urgonian facies-type. It represents calcareous shallow-water sedimentation of an inner shelf of the European plate comprising the Barremian-Aptian period. As pointed out by GWINNER (1978), the Schrattenkalk is underlain in all litho-tectonic units by the Middle Barremian-Lower Aptian Drusberg Formation of outer neritic facies (OBERHAUSER 1986). After a hiatus of varying time span, the Schrattenkalk is overlain by the Garschella Formation (FÖLLMI 1986).

Although the two sections studied (Text-Fig. 2) are quite similar in lithology by showing limestone as the prevailing rock type, they certainly must be considered as two rather extreme types in relation to their textural features, fossil content and position within the sedimentary cycle.



Text-Fig. 2. Location map of the studied sections.

4. Description of Sections

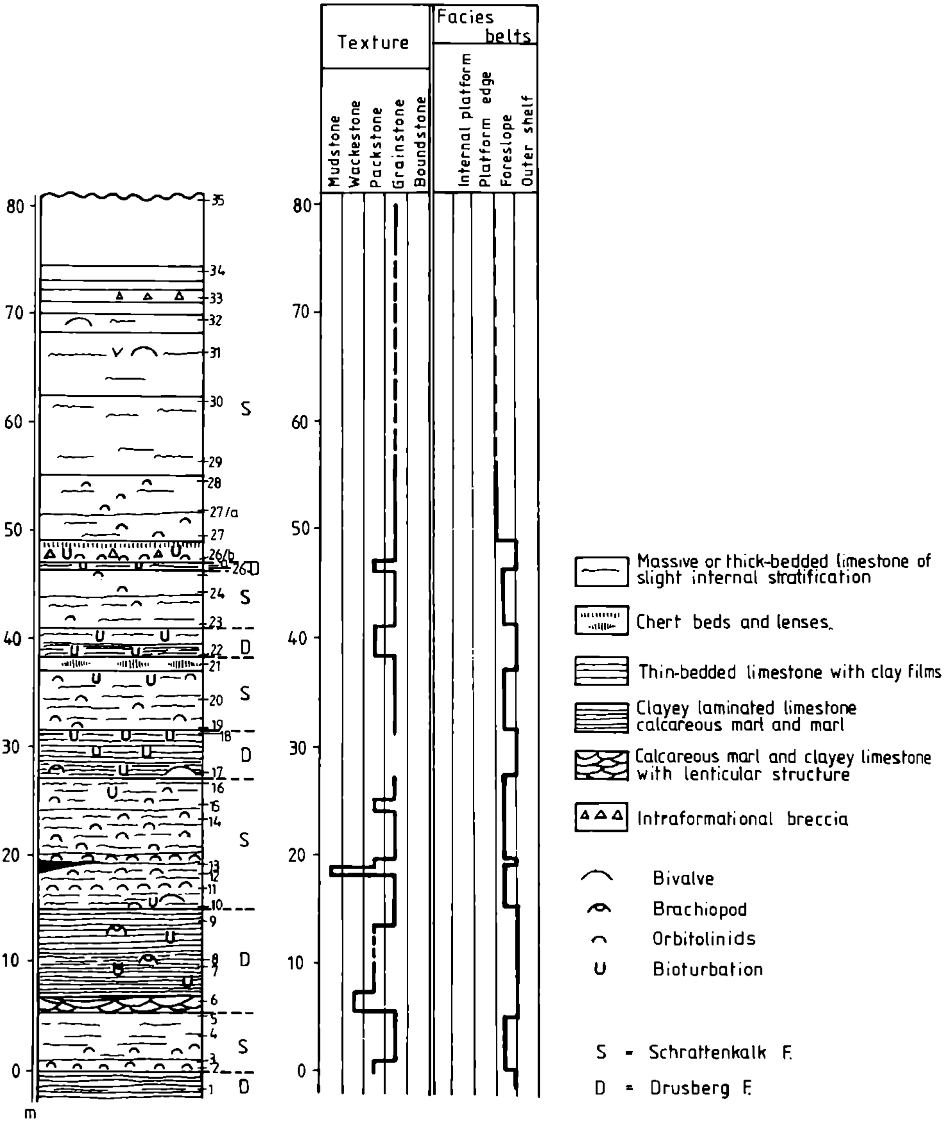
4.1 Upper Ill-Gorge Section at Feldkirch

The section comprises more than 80 m of predominantly medium- to coarse-grained calcarenites. The beds of varying thickness show sparry cemented light grey and dull brownish-grey packstones and grainstones. Megafossils are extremely rare, only small echinoderm debris is abundant as well as *Orbitolina* in part of the beds. Bioturbation is also a rather common phenomenon, including burrowing and horizontal traces of ichnofossils. The lower half of the profile is characterized by the recurrence of the Drusberg Formation with upward decreasing thickness of these incursions and is less pronounced from facies. These pelitic intercalations are dark grey, frequently flasered, thin-bedded or lenticular in structure. Graded beddings and cross-beddings are not characteristically developed. A silicification recognizable even to an unaided eye (thin layers or lenses) can be encountered in the middle of the section (Beds 21 and 27). In the upper third of the section, the original textural-structural features become unrecognizable as a result of recrystallization.

On the basis of the texture in thin sections the section can be divided into two major units. The lower 55 m (Samples 2-28) are characterized by an intrabiosparry texture of grainstone type, where 20 to 30 % sparry matrix has cemented 15-20 % intraclasts, 30-50 % bioclasts and 2-10 % pellets. Ooids are common, but never exceed 3 %. Silicification is frequent, but does not surpass more than 1-2 % in total amount, characterizing mainly the lower part of the section. This seems to be responsible for the authigenic formation of feldspar (dia- or epigenic) which in some places amounts to 4 %. Among the lower 15 samples some biomicritic, biomicrosparitic (wackestone-type) texture can also be observed. These texture types derive almost exclusively from intercalations typical for the Drusberg Beds. The amount of terrigenous clastics (up to 50 %) and in one case even of extraclasts is considerable, too.

Because of the extremely strong recrystallization the original texture of the upper third (Samples 29-30) has been changed beyond recognition. Along with the cement, the allochems have also been distorted by recrystallization. Accordingly, the biogenic components vary from 1 to 40 %. Both the terrigenous clastics and the authigenic plagioclase are of less significance in this profile interval.

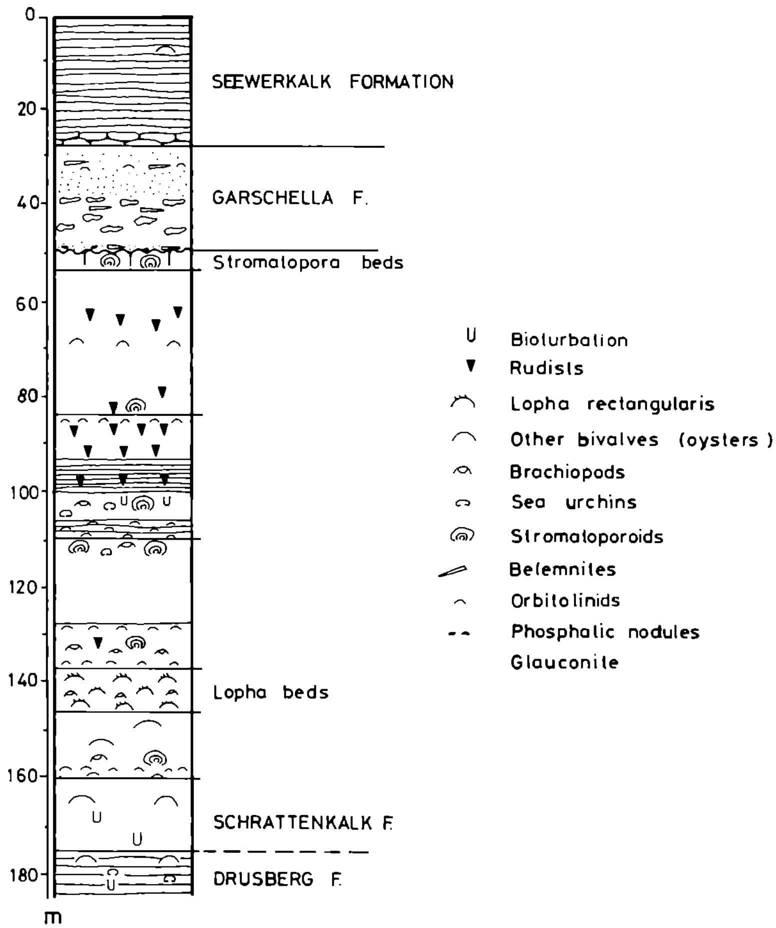
The terrigenous grains have an average diameter of 0.1-0.2 mm, but even the maximum does not exceed 0.5 mm. The largest grain size occurs in the middle third of the section which coincides very well with the culmination of the number of quartz grains per 1 cm^2. In spite of the difference in lithology, the composition and distribution of the fossils recognizable in thin sections testify the uniformity of the geological features of the section (Text-Fig. 3). The most abundant fossil group are foraminifera among which the calcareous benthonic forms play the leading role. As shown by I. KOVACS-BODROGI's results the genus *Nummoloculina* and especially the species *N. heimi* (BRÖN.) show the most striking abundance. Giant specimens of Miliolidae are typical for the Beds 20-25. In terms of frequency, the calcareous benthonic forms are followed by the arenaceous ones, among which the role of the genus *Orbitolina* is striking (Table 1).



Text-Fig. 3. Columnar section of the Ill-Gorge road cut, Feldkirch.

Highly abundant biogenic detritus throughout the section is represented by fragments of echinodermata, mainly echinoids, though a few crinoidea are present, too. Bivalve shell fragments vary widely in abundance. Among them, regardless of Sample 14, the rudist shell detritus is very scarce. The high abundance of bryozoans is conspicuous, whilst the hermatypical organisms such as corals and hydrozoans as well as sponges are quite scarce, too. The dasycladaceae fragments present are only poorly preserved.

In spite of the considerable number of fusite and xylite fragments with flaked rim, the analysis of pelites for sporomorphs remained unsuccessful (personal communication by M. JUHASZ).



Text-Fig. 4. Columnar section of the Rhomberg quarry, Unterklien, Vorarlberg.

4.2 Rhomberg Quarry Section, Unterklien

The profile of Unterklien shows an overturned stratigraphic sequence. It belongs to the northernmost Säntis Nappe or the Axen (or Hohenems) Nappe. In the quarry a sequence is exposed extending from the Drusberg Formation at the top to the Seewerkalk at the base (Text-Fig. 4).

HEIM & BAUMBERGER (1933) partly confused the stratigraphy by misinterpreting an echinoid fauna from the overlying marls as Hauterivian. OBERHAUSER (1969) reported the presence of *Conorotalites bartensteini intercedens* in these marls at Oberklien which proves Middle Barremian age (cf. OBERHAUSER 1986).

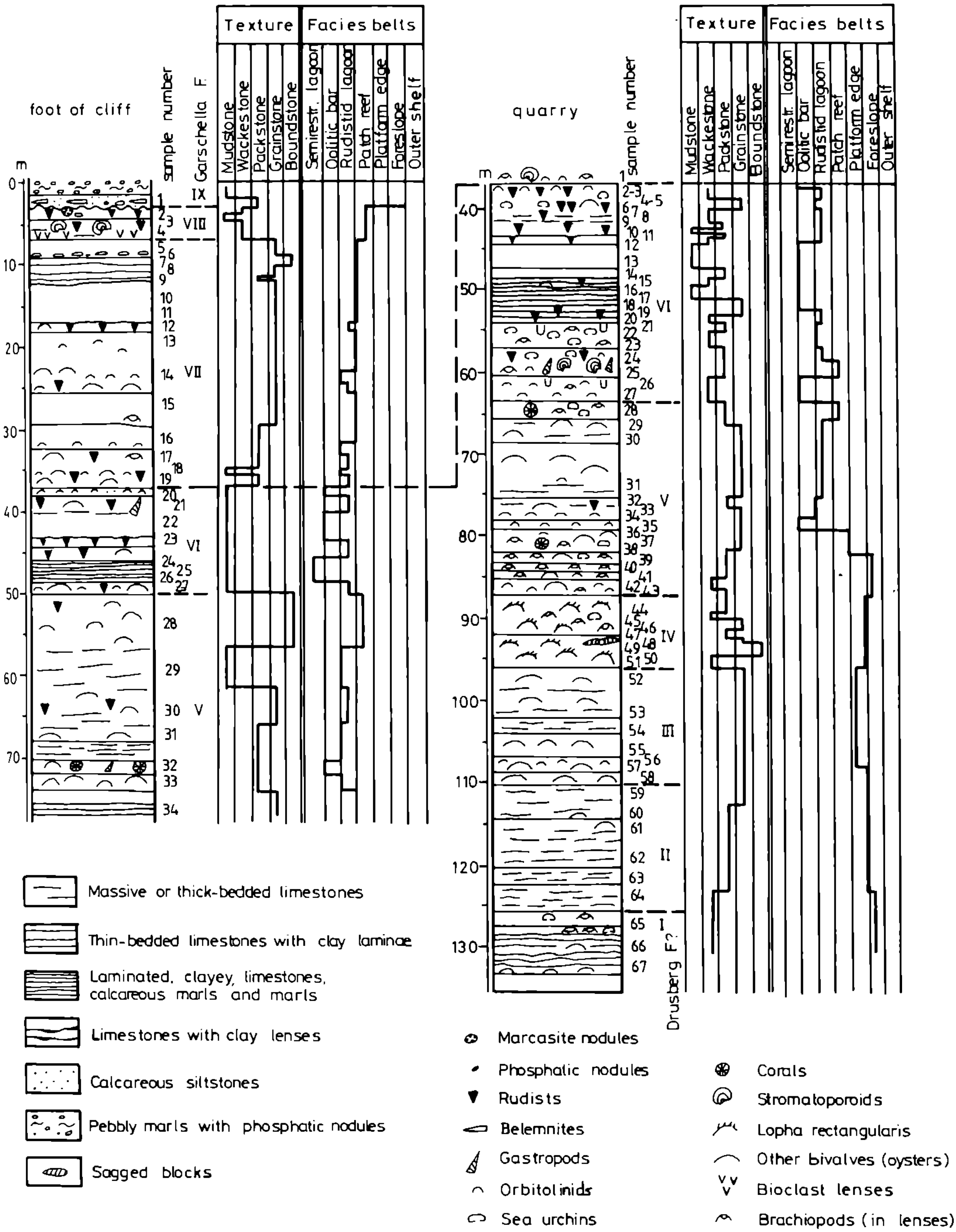
The block of the quarry is separated toward the bulk of the Emsrütti mountain segment by a fault plane of strike-slip origin. The fault plane of nearly vertical position is dotted with tiny and m-order-of-magnitude furrows. Similar slickensides which are one order of magnitude smaller can be seen on the bedding surface of the Drusberg Formation, testifying NE-oriented slips of blocks which might be compared to the movement of a pack of cards.

The sequence was studied in two sections partly overlapping each other (Text-Fig. 5) and comprises nearly one hundred metres starting from the southwestern corner of the quarry up to the *Lopha* bed. That part of the profile extending up to the Garschella Formation can be divided, lithologically and megafaunistically, into nine units.

Unit I is constituted by dark grey fine-grained, bedded, flasered limestone with thin marl and claymarl lenses or layers along the bedding planes. The lenses contain small brachiopods, high amounts of dispersed, thin-walled echinoid cross-sections and thick-bedded *Ostrea* valves exhibiting a biostrome pattern. Thin sections exhibit a biomicritic rock texture of wackestone-type. The most typical fossils recognizable include ostracoda, *Cadosina* and echinoderm detritus. The unit forms a transition between the Drusberg Formation and the Schrattenkalk. The typical Drusberg Formation occurs at a higher level at the waterfall farther to the southwest.

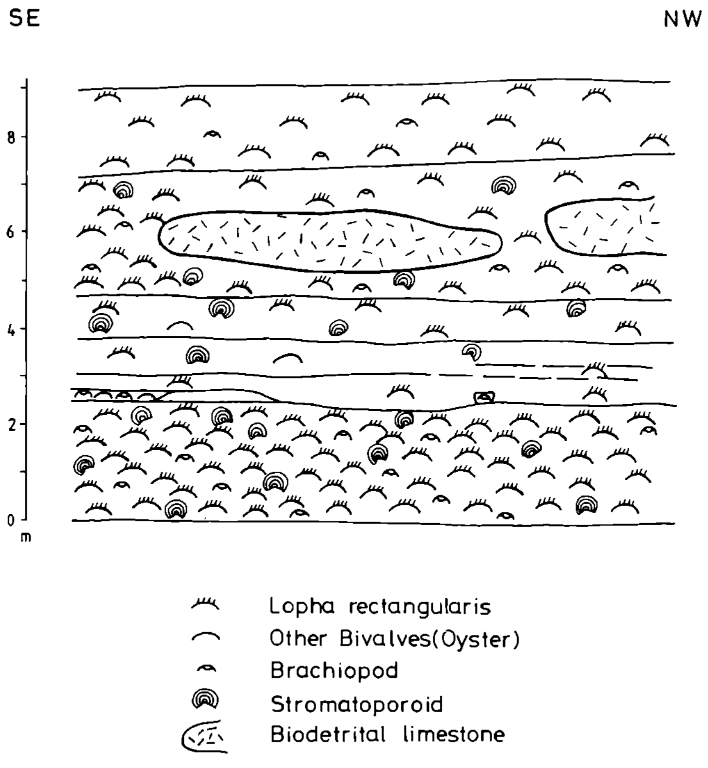
Unit II, unambiguously belonging to the Schrattenkalk, is 15 m thick and megaloscopically similar to the preceding unit, though lacking the flasered clay-intercalated bedding features. It is heavily bioturbated, and contains only a few *Ostrea* and *Lopha* valves. Thin sections show a predominantly biopelmicrosparitic packstone texture. Toward the younger beds, it contains increasing amounts of both calcareous (including Miliolidae) and arenaceous benthonic foraminifera. Planktonic foraminifera are absent as practically in the whole profile. The bioclast-dominated calcareous sandstone shows occasionally a graded structure.

Unit III is 14 m thick, light grey, thick-bedded or massive, often stylolitic, sometimes constituting ooidal limestone with small or medium grained bioclasts. Its texture of grainstone type varies between biointrasparry and biopelsparry types. The bioclast-dominated limestone shows occasionally graded-bedding structures. The proportion of coated grains has become considerable. *Ostrea* valves are present in a biostrome structure as well as *Orbitolina* and a few hermatypical algae. The foraminifera vary in abundance. Usually the calcareous benthonic forms are more frequent, but in a small interval arenaceous benthonic forms (mainly *Orbitolina*) are dominating. Bryozoans increase in abundance.



Text-Fig. 5. Columnar section and facies changes of the Schrattenkalk at the foot of the cliff and in Rhomberg quarry, Unterklien.

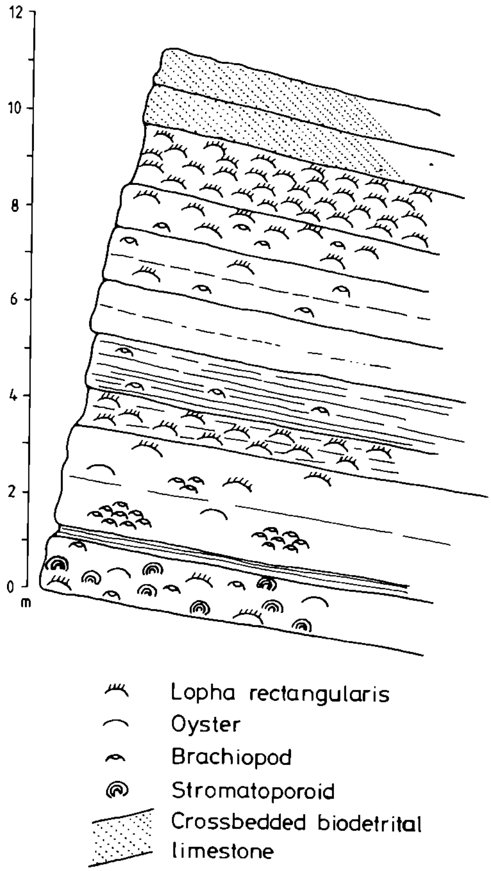
The most characteristic element of the Schrattenkalk is **Unit IV**, attaining a total of only 9 m thickness. It is composed of limestone and calcareous marl with *Lopha rectangularis* (HEIM's: "*Alectryonia*"), is of dark grey or black colour, varies in clay content, and is thick-bedded and composed overwhelmingly of valves of the name-giving fossil (Text-Fig. 6). In addition, frequent fossils include brachiopods, *Stromatopora* forming minor bioherms, hermatypical algae and sphinctozoan sponges. The taxa present seem to correspond to the forms described by H. SCHOLZ (1984). The thin section texture varies from wackestone to boundstone. The bioclast varieties recognized in thin sections show, for the first time, the predominance of bivalve shell detritus at the expence of echinoderms. In the *Lopha*-bearing parts of the unit there are rock bodies of varying size (0.2 x 0.3 m, 7.0 x 1.6 m, and 10 x 2.2 m) sharply differing both in lithology and fossil content from the bulk of the unit. Regardless of the smaller one these bodies are of moderate to coarse bioclastic texture, containing a considerable



Text-Fig. 6. Sagged blocks within *Lopha* beds in Schrattenkalk in SW corner of Rhomberg quarry, Unterklien.

amount of ooids. The grain size becomes coarser towards the younger beds. The predominant faunal elements are echinoid spines with which some brachiopods, pelecypod shell detritus, bryozoans and - near the present day upper boundary of the unit - also *Ostrea* are associated. The texture is of grainstone type.

Doubts as to the origin of the rock bodies will be convincingly eliminated, when the largest rock body is studied with more scrutiny. Quite easily distinguishable within this bioclastic block is the stratification which is confirmed by the presence of a two-generation fill in one brachiopod shell: 265/55°, in contrast with the 140/8° dip of the sequence bedding. An important circumstance is, furthermore, that this rock body is traversed by a fissurefill consisting of *Ostrea* coquinas with a dip of 130/37°. The third smallest rock body is constituted by small *Ostrea* and brachiopod shells. These blocks are regarded as having sagged during deposition.



Text-Fig. 7. *Lopha* beds in the NE corner of Rhomberg quarry, Unterklien.

At the same time, the unit, consisting overwhelmingly of *Lopha*-biostromes, is an excellent example for intraformational variability of facies. The thickness of the unit measured in the NE part of the quarry (Text-Fig. 7) is 1 m greater than in the profile of the SW part of the quarry. However, the difference in the character of bedding subdivision and the distribution of the megafauna is much more conspicuous. In the profile along the cliff the thickness decreases to 7 m towards the SW. In the steep cliff of Neu-Ems Castle further to the SW, this facies cannot be traced any more.

Unit V is 23.5 m thick and shows fossil assemblages strikingly different from the preceding unit. Its lower half is lighter grey in colour and coarsely bioclastic in texture. The upper part shows brachiopods as predominant megafossils, which are locally enriched mainly in lenses and frequently calcite-filled (Table 2). Both above and below the brachiopod biostrome, thick-walled *Ostrea* valves occur in considerable number. Two *Orbitolina*-dominated biostromes, *Stromatopora* and tiny sphinctozoans (mostly *Barroisia helvetica*) can be observed, too. Rudists and small coral bioherms appear for the first time. Among the microfossils arenaceous (mainly *Orbitolina*) and calcareous benthonic foraminifera are strikingly frequent. The bryozoans attain their maximum abundance in the profile here. The predominant thin section texture of the rock is intrabiosparitic-pelbiosparitic, i. e. grainstone, with few (a maximum of 5 %), but constant, ooidal grains. Here, the coated grains reach their maximum occurrence within the profile. At the base of the unit and at its top, the texture is of packstone-type.

According to a lithology-based correlation, Unit V was found to have a thickness of about 40 m in the section along the rock wall which is another excellent proof of the rapid lateral variability of the facies.

Unit VI is 26.5 m thick in the quarry profile, in the section along the cliff, however, only 13 m and represents the most pelitic unit of the formation. It consists of an alternation in two rhythms of dark grey laminated and thin-bedded calcareous marls and argillaceous limestones with medium-grey, massive, poorly sorted, bioclastic and slightly argillaceous limestones. *Orbitolina*-forming biostromes dominate at both the base and at the top (Table 1). Here, the rudists appear in highest frequency. Mrs. L. CZABALAY identified the frequent and relatively small forms as *Toucasia carinata* and a few larger specimens as *T. lonsdalei*. Brachiopods are characterized by a considerable frequency (Table 2).

In addition, the frequent sea-urchin spines, the *Stromatopora* bioherms and the gastropods (which appear here for the first time) are worth mentioning. Occasionally, the degree of bioturbation is high, too. The thin section texture is extremely varied which is in harmony with the megaloscopic lithologic features. The most common texture type is wackestone, though packstone and, for the first time in the profile, mudstones are frequent, too, whilst grainstones appear only occasionally. Almost all varieties of the texture types distinguished by FOLK are encountered. The frequency of terrigenous quartz grains per 1 cm^2 is strikingly high. The quantity of coated grains, however, has become largely reduced. The benthonic foraminifera, both arenaceous and calcareous, are common throughout the unit. The frequency of ostracods and dasyclad algae has increased considerably and *Cadosina* is frequent again, whilst the bryozoans have almost completely disappeared.

Table 1. Orbitolinids from various Schrattenkalk occurrences, Vorarlberg (determined by E. KÖHLER).

	Sample No.	*Orbitolinopsis cuvillieri* MOULLADE	O. cf. *cuvillieri* MOULLADE	*O. debelmasi* MOULLADE & THIEULOY	*O. kiliani* SILIVESTRI	*O. pygmaea* ARNAUD-VANNEAU	*O.* sp.	*Palorbitolina lenticularis* (BLUMENBACH)	*P.* cf. *lenticularis* (BLUMENBACH)	*Paracoskinolina maynci* (CHEVALIER)	*P. sunnilandensis* (MAYNC)	*P.* sp.
Unterklien (Rhomberg)	1b			x								
	2										x	
	3										x	
	4	x				x						
	12		x									
	13	x								x		
	27							x		x	x	x
	29						x	x			x	
	33							x				
Feldkirch, upper Ill-Gorge	11								x			
	34						x	x				
	35	x								x		
Alploch -	9							x			x	
Göfis					x			x		x		

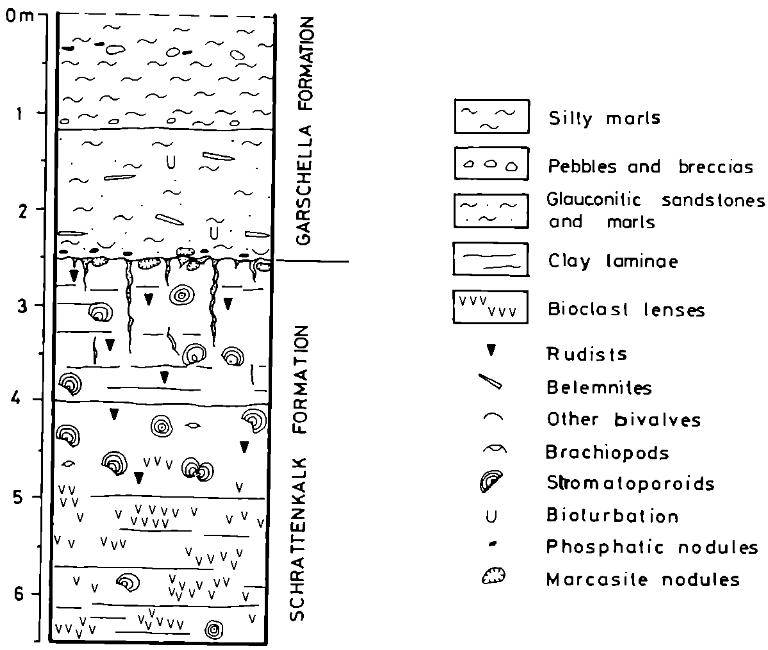
Table 2. Brachiopod specimens from the Schrattenkalk, Rhomberg quarry, Unterklien (determined by A. VÖRÖS).

	Lamellaerhynchia renauxiana (D'ORB.)	*L. gillieroni* (PICT.)	*L.* cf. *multicostata* BUR.	*Sellithyris* cf. *sella* (SOW.)	*Loriolithyris* cf. *russillensis* (LOR.)	*Symphythyris?* sp.	*Tamarella* cf. *tamarindus* (SOW.)	*Rugitela?* sp.	Terebratulida indet.
Sample No. 2		2	2	3					
Sample No. 23	2								
Sample No. 24	1								
Sample No. 39		124		12	5	1	7	2	13

From the rock wall counterpart of Unit VI, the older *Orbitolina*-bearing horizon could not be identified. Only a couple of cm-thick younger *Orbitolina* biostromes have increased here to 1 m in thickness. Their faunal content, besides of the extremely thin-walled rudists, is poorer than in the preceding unit. Another substantial difference is that the texture is represented by mudstone throughout the unit.

The 30 m thick **Unit VII** was studied only at the foot of the cliff. The limestone is again light grey in colour, bedded, peloidal, with fine- and medium-grained bioclasts. The megafauna is poor: rudists and less frequent other pelecypod valves, a few brachiopods and in the lower and upper parts of the unit *Stromatopora* bioherms. In the middle of the unit *Orbitolina* is more frequent. The foraminiferal assemblage is abundant and varied, too; large Miliolidae predominate. Echinoderm skeletal detritus, ostracods, *Cadosina* and dasyclad algae are common, too. The thin section texture is overwhelmingly of grainstone, subordinately of packstone type.

Attaining only a total of 4 m in thickness, **Unit VIII** is represented by bioclast-lens-dotted, flasered clay-intercalated and sometimes nodular limestone. The abundance of both *Stromatopora* and several decimeter large algal bioherms are also conspicuous as well as the less frequent rudist bivalve cross-sections. The upper boundary is unconformable. The glauconitic marls of the Garschella Formation can be found as fissure- and cavity-fills (Text-Fig. 8). Along the boundary Schrattenkalk debris is accompanied by marcasite nodules of cm-size. The thin section texture of the limestone is biomicritic wackestone or of grainstone type.

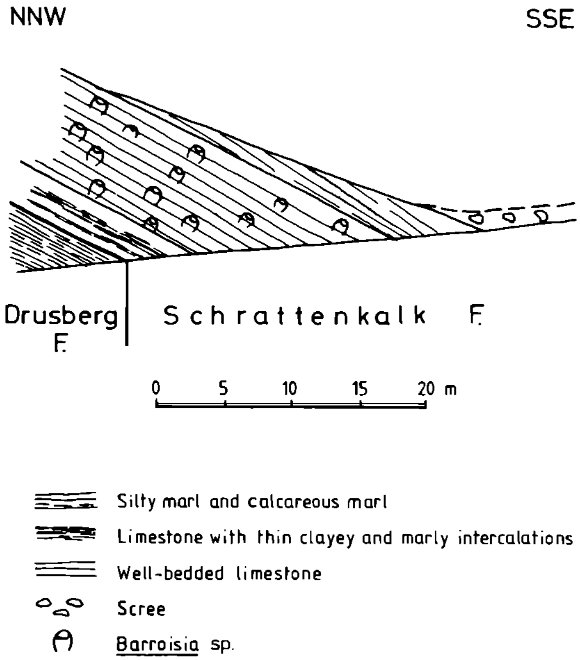


Text-Fig. 8. Contact between Garschella and Schrattenkalk formations in Rhomberg quarry.

Unit IX, the Garschella Formation, is made up of silty and glauconitic marl with some exterraneous detritus of pebble size. Radiometric studies confirm that these pebbles represent the oldest rock debris ever found in the Alps (FÖLLMI 1986). It is worth mentioning that along with glauconite appear muscovite and apatite in considerable amounts. These features together suggest changes in the source area topography and, in addition, in the sea currents regime. The K_2O content here is 4.10 %, the P_2O_5 content 2.50 %.

5. Depositional Environment

Comparing the curves representing the variation of the depositional environments reconstructed for the two sections, the marked difference is obvious. The section of Feldkirch (Text-Fig. 3) represents deposits accumulated on an extremely flat foreslope and perhaps outer shelf environment where megafossils are extremely scarce. The brachiopods, however, may constitute an autochthonous element. The fossil detritus as observed in thin sections



Text-Fig. 9. Road cut at the southern end of Übersaxen.

has derived almost entirely from the internal platform. This opinion is supported by a weak, but overall silicification visible even to an unaided eye, by some low-angle cross-bedding and by some graded bedding. Orbitolinids are common and very often in overturned position being evidence for their transportation. In spite of this fact it cannot be excluded that they might have lived, in part, in this environment.

The trend-like variation of beds of silty marl, pelitic limestone and calcareous sand facies is probably not primarily the result of eustatic sea-level changes, but is rather due to a temporarily decreasing rate of paleotransport from the internal platform. This opinion is supported by the Frutzkopf section on the road to Übersaxen (Text-Fig. 9) and other exposures where the transition between Drusberg and Schrattenkalk formations is rapid, without alternation of facies.

The very diversified depositional environments that can be reconstructed from the Ill-Gorge and Rhomberg-quarry section (Text-Fig. 5) vary from a flat-lying foreslope up to a semirestricted lagoon (Text-Fig. 10). The curve indicates a shallowing upward trend. The idea suggesting for the Drusberg Formation a flat foreslope paleoenvironment beneath the base of wave action (Text-Fig. 10) is quite consistent with the abundance of brachiopods and the frequent occurrence of thin-walled, but unbroken sea-urchins. However, the fact that they occur together with thick-walled *Ostrea* may be contradictory to this interpretation.

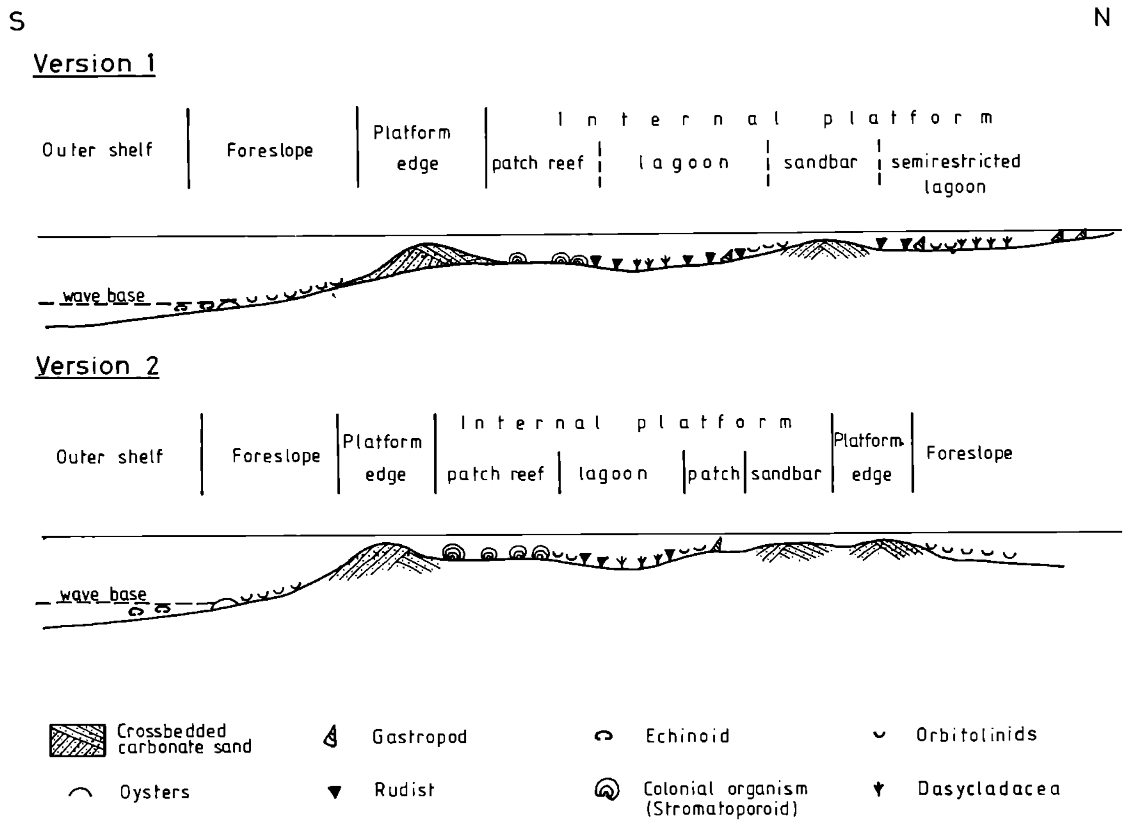
Small skeletons of siliceous sponges (mostly hexactinellids) were found in varying abundance in two profiles, namely in a roadcut south of Übersaxen through an 8 m sequence and also along an old road between Furx and Laterns in an approx. 3.5 m thick interval. This type of Schrattenkalk is supposed to represent the most distal facies.

There are at least three possibilities for the occurrence of oolitic sand. They can be transported basinward away from the platform where they originated. In such a case oolitic beds are usually not cross-bedded (see Feldkirch section). On the contrary, the oolitic sands both on the platform edge and on the internal platform are mostly cross-bedded.

Cross-bedded limestones of calcareous sand origin occur in the NE part of Rhomberg quarry in two beds above the *Lopha* biostrome (Unit IV, Text-Fig. 7) as well as in Nofels quarry, at a distance of about 15 m below the boundary with the Garschella Formation. Both are supposed to be deposited in an internal-platform environment. In the latter case it is directly adjacent to, or occasionally even concurrent with small sphinctozoan and *Stromatopora* bioherms. The abundance and commonness of rudist biostromes and algal species belonging to the Dasycladales are clearly an indication of an inner-platform paleoenvironment, whereas the similarly frequent orbitolinids occur both on the flat-lying foreslope and on the internal platform. As shown by CUGNY (1979), there is a kind of differentiation among the species even with respect to the depositional environment.

Since no true reef front is known from the Schrattenkalk, the most specific well-interpretable evidence of a shallow-water depositional environment is provided by the occurrence of bioherms of varying size constituted mainly by stromatoporoids and occasionally by chaetetids and, less frequently, by hermatypical corals. They can be observed on several points of some sections but their most frequent and largest occurrences are connected with the uppermost beds of the formation. This phenomenon bears witness to a shallowing-upward trend and to progradation.

The largest bioherms of *Stromatopora* and of the occasional corals can be observed in Unit VIII of the Unterklien section (Text-Fig. 8). Similar colonies can be found near Ebnit, at the bifurcation of Kobel Ache and in the vicinity of one of the tunnels of the road along Dornbirner Ache. In the latter case, nearly 50 % of the particular beds are constituted by the afore-mentioned bioherms. In this context, the possibility of the presence of a platform-margin reef may be suggested, too. A few meters below, biogenic or intraclastic pebbles of 0.5 - 1.0 cm in diameter can be observed. This is the highest energy horizon of the complete formation. In a similar stratigraphic position there are *Stromatopora* and *Chaetetopsis*-like bioherms in the Churfirsten Range, being perhaps more spectacular than even the most beautiful Vorarlberg occurrences and being associated with ramose sponge colonies tens of cm in size. In the about 80 m thick Schrattenkalk of the Allgäu the presence of hermatypical organisms is most typical for the middle segment of the formation (H. SCHOLZ 1984). This phenomenon and the continuous sedimentation between Schrattenkalk and Garschella formations seems to be inconsistent with the overall regression. Therefore, there is no evidence for subaerial erosion at the end of the Schrattenkalk sedimentation - in spite of the fact that there are some glauconitic pockets and fissure fills in the uppermost bed of the formation. What was the reason for the cessation of the Schrattenkalk sedimentation? Agreeing with



Text-Fig. 10. Possible versions of arrangements of the facies belts in the Helvetic Zone during the Schrattenkalk sedimentation.

SALOMON's and FUNK's opinions (pers. comm.) it could be a direct marine connection with the Boreal Realm. This connection could have come into existence in early Lower Aptian time.

6. Paleogeography

In terms of ZACHER's reconstruction of the Schrattenkalk (1973), the Helvetic Realm included three E-W-striking zones. In the north, there was a narrow oolitic facies, in the middle there was a bioarenite and intrasparite facies, whilst in the south there existed a large source area of biogenic detritus. In the light of the results of studies in the Allgäu area, H. SCHOLZ (1984) introduced some changes into the afore-mentioned model. Accordingly, the source of the large bioherms and biostromes would have existed within a wide bioarenite zone. At the same time he found no solution for eliminating the contradiction that the wave action, that is reflected by the oolitic sands, implies heavy agitation and could not proceed but through the zone of high biomass productivity.

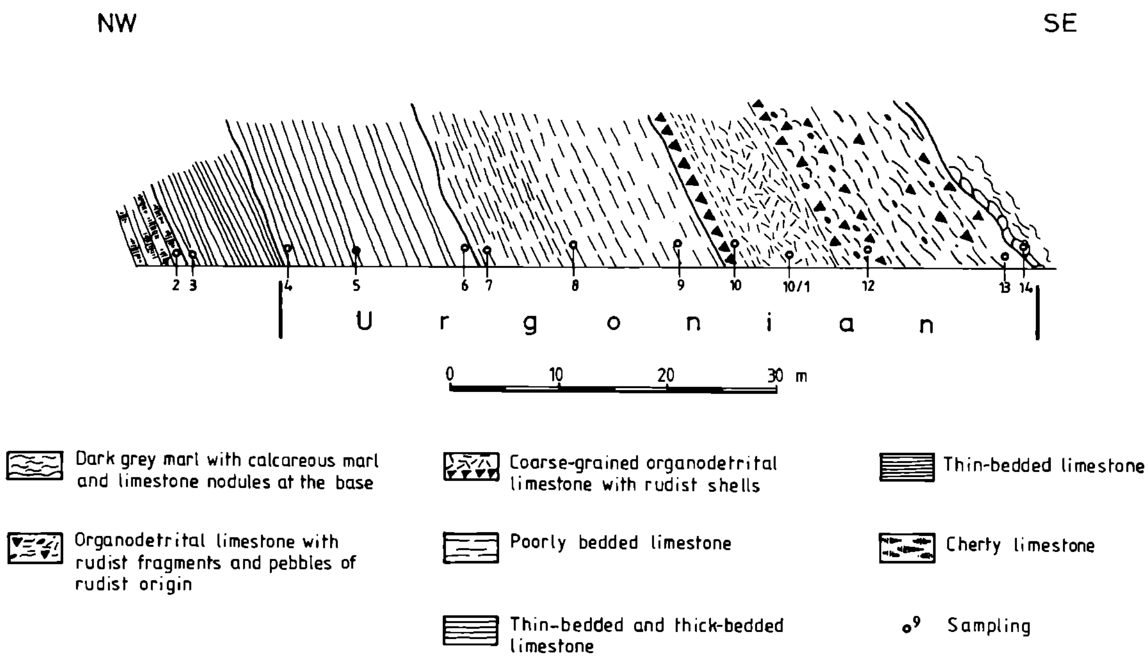
As a result of these investigations, there are two points in this context that are worth to emphasize:

- The absence of both the terrigenous detrital fraction and palynomorphs in the Schrattenkalk bears testimony to a considerable distance which separated the Schrattenkalk from the shoreline to the north. Therefore, a smaller perhaps semi-restricted basin is supposed to have existed on the platform and thus heavy wave action coming from the north may have been responsible for the oolitic facies.
- At the time of Schrattenkalk deposition, the Rankweil slope (FÖLLMI 1986) did not yet exist. The platform slope was still further to the south and was even more gentle. For this reason, the calcareous sands introduced into the foreslope or basin are gradually pinching out in a southern direction.

The calcareous sand facies of the Schrattenkalk differs from the facies known from the Falknis-Tasna nappes of the Vorarlberg Flysch Zone, where the bioclastic, often argillaceous and heavily compressed limestone of predominantly turbiditic origin is called "Tristelschichten" (SCHWIZER 1984, OBERHAUSER 1983, 1986). It is striking that no rudist shell debris are known among the bioclasts (SCHWIZER 1984). This allodapic limestone seems to have been removed from the Central Penninic ridge and washed into the flysch basin to the north.

According to HAGN (1982), Urgonian facies of Upper Barremian age with *Orbitolina* also exists in the Thiersee Syncline in the Northern Limestone Alps of Tyrol. Clasts of Urgonian limestones are found as pebbles in conglomerates of Cenomanian, Gosavian and Upper Eocene ages in this region. Non-Urgonian facies of Barremian-Aptian and (?) Lower Albian ages of Roßfeldschichten-type occur further east, predominantly in the Northern Limestone Alps of Salzburg and southeastern Bavaria. The Lackbach-Schichten of DARGA & WEIDICH (1986) can be, according to LOBITZER's opinion, considered a particular facies development of the Roßfeldschichten.

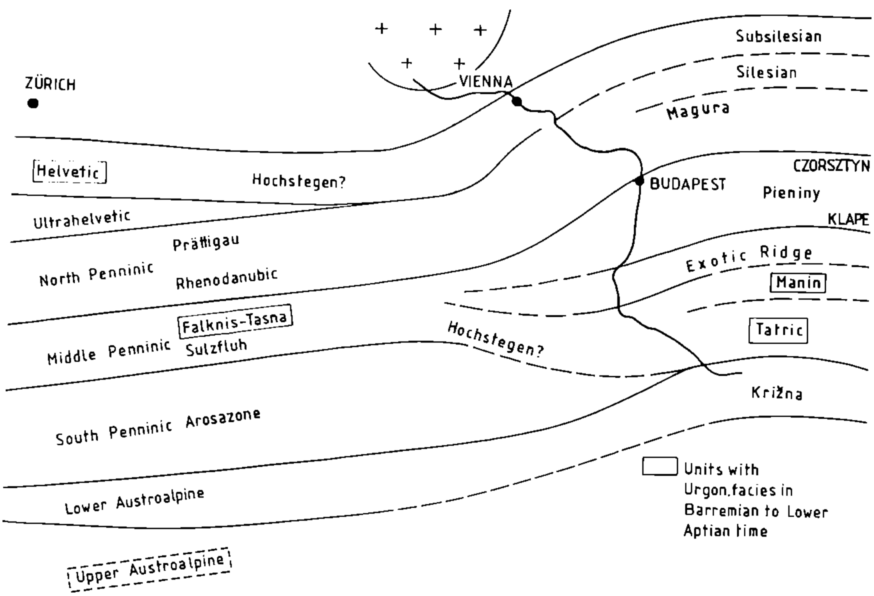
The Helvetic Zone including the Schrattenkalk becomes gradually narrower to the NE and finally disappears completely. The nearest occurrences of the Urgonian facies within the Carpathians are in the Tatric Zone (Manin Unit and the High Tatra). The Manin Formation is coeval with the Schrat-



Text-Fig. 11. Profile sketch of Urgonian facies in the Manin gorge, Slovakia.

tenkalk, i. e. Upper Barremian-Lower Aptian (RAKUS 1977, MISIK 1957), whilst the Vysoká-Turnia Formation in the High Tatra is Barremian-Aptian (LEFELD et al. 1985). Both the type-sections exposed in the Manin-gorge (Text-Fig. 11) and the Tatra-Mts. occurrences (LEFELD 1968, 1974) suggest that the formation of Urgonian facies in these regions develops gradually from the lower part of the Lower Cretaceous with a more pronounced shallowing trend than that observed in the Schrattenkalk. The heavily re-crystallized mostly clastic sequence with well rounded pebbles comprises hardly 80 m thickness in the Manin gorge. Thick-walled rudists and a few other pelecypod valves are only present in the upper half of the section. In the middle part of the High-Tatra section true coral and hydrozoan reefs are also encountered and even endogenic breccias are present (LEFELD et al. 1985). The overlying rock can be readily compared with the Schrattenkalk. The Aptian-Albian (Manin) and Albian (High Tatra) marls start with a phosphate-bearing, glauconitic basal layer.

It is evident from this general paleogeographic reconstruction (Text-Fig. 12), that in spite of the correspondence in both, age and geological features the paleogeographic situation was substantially different in the two regions. Whereas the Helvetic Zone lies on the northern margin of the European plate, the Manin and High-Tatra occurrences are situated in the Tatric Zone being several zones farther to the south (BIRKENMAJER 1986), being in correlation with the Central Penninic Zone, i. e. with the source area of the Tristelschichten.



Text-Fig. 12. Paleogeographic reconstruction of the East Alpine - West Carpathian region (early Cretaceous).

7. Conclusions

Regarding both, the lithological and the faunistic features on the one hand and the environments of sedimentation on the other, the two Vorarlberg sections studied are considerably different from each other. The sequence of the Ill-Gorge section was deposited on a very flat-lying "foreslope" or possibly in an outer shelf environment, while the section of Unterklien suggests a wide range of environments varying from a low-angle foreslope up to a semi-restricted lagoon.

In the block of the Rhomberg quarry, the NE-orientated dislocation of the topmost beds as compared to the ones underneath was proved by conclusive evidence.

The Schrattenkalk-sedimentation zone was probably separated in NW direction by a minor sedimentary basin from the shore-line. The cessation of Schrattenkalk sedimentation was provoked by a marked decrease in the temperature of the sea-water, probably by the establishment of a direct communication with the Boreal Realm.

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Foraminiferen, Kalkalgen und die Biostratigraphie des Schrattenkalks von Vorarlberg (Österreich)

ILONA BODROGI, Budapest

Mit 4 Tafeln und 7 Text-Figuren

BODROGI, I. (1989): Foraminiferen, Kalkalgen und die Biostratigraphie des Schrattenkalks von Vorarlberg (Österreich). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 403-429. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Microfaunal and calcareous algal biostratigraphy of three Schrattenkalk sections in Vorarlberg (Rhomberg quarry at Unterklien, section 1 and 2; Upper Ill-Gorge at Feldkirch) were studied within IGCP Project 262 (Tethyan Cretaceous Correlation).

78 taxa of the foraminifera fauna belong to 31 species and 50 genera, respectively. They include two new *Orbitolinopsis* species: *Orbitolinopsis* sp. 1 and *Orbitolinopsis* sp. 2, which will be described separately.

Most of the genera and species are attributed to agglutinating forms, while most of the specimens belong to miliolinids. Planktonic foraminifera are represented by a few specimens of *Hedbergella planispira* (TAPPAN) in section 1 at Unterklien and at Feldkirch.

The sequence representing the Schrattenkalk Member of Schrattenkalk Formation at Unterklien belongs to the Orbitolinopsis cuvillieri-kiliani Subzone of the Lenticularis Zone (stratigraphic subdivision after MOULLADE et al. 1985), since *Palorbitolina lenticularis* (BLUMENBACH) occurs in it.

The middle unit of the Member, the *Orbitolina* Marls form an especially characteristic, isochronous marker level suitable for regional correlation. They start with the base of the Lower Aptian with the first occurrence of *Palorbitolina lenticularis lenticularis* (BLUMENBACH). The *Orbitolina* Marls can be identified and correlated in both profiles of the Rhomberg quarry (section 1, sample No. 20; section 2, sample Nos. 26-27). The lowermost part of the Schrattenkalk Formation belongs presumably to the uppermost Barremian.

The stratigraphical subdivision is substantiated by a Dasycladaceae flora (*Salpingoporella muehlbergi, S. hasi, S. melitae, Cylindroporella benisarensis*), too.

The Schrattenkalk Formation gradually develops from the Drusberg Formation. The age of the latter one is early Middle Barremian in the Ranzenberg section near Hohenems (FUCHS 1971, WIEDMANN 1978).

The section of Upper Ill-Gorge is a slope deposit. The resedimented sequence belongs to the Palorbitolina lenticularis Zone. Its age is Upper Barremian to Lower Aptian.

Kurzfassung: Drei Schrattenkalk-Profile in Vorarlberg (Steinbruch Rhomberg bei Unterklien, Profil 1 und 2; Obere Ill-Schlucht bei Feldkirch) wurden im Rahmen des IGCP-Projektes 262 (Tethyan Cretaceous Correlation) biostratigraphisch untersucht.

Die Biostratigraphie stützte sich dabei vor allem auf benthische Foraminiferen und Kalkalgen. 78 Foraminiferen-Taxa verteilen sich auf 50 Gattungen und 31 Arten, darunter 2 neue Arten der Gattung *Orbitolinopsis*: *Orbitolinopsis* sp. 1 und *Orbitolinopsis* sp. 2, deren Beschreibung einer späteren Arbeit vorbehalten bleibt. Unter den Gattungen und Arten herrschen die agglutinierenden Formen vor, während miliolide Foraminiferen hinsichtlich der Individuenzahl dominieren. Planktonische Foraminiferen sind in Unterklien-1 und Feldkirch nur mit der Art *Hedbergella planispira* (TAPPAN) vertreten.

Die Biostratigraphie wird durch die Dasycladaceen-Flora ergänzt, in der die Arten *Salpingoporella muehlbergi, S. hasi, S. melitae* und *Cylindroporella benisarensis* charakteristisch sind.

Der Schrattenkalk bei Unterklien gehört zur Subzone von Orbitolinopsis cuvillieri-kiliani der Lenticularis-Zone (MOULLADE et al. 1985), wie das Auftreten von *Palorbitolina lenticularis* (BLUMENBACH) zeigt.

Die mittlere Einheit des Schrattenkalkes bei Unterklien - die Orbitolinen-Mergel - bilden einen charakteristischen, isochronen Horizont, der für regionale Korrelation genutzt werden kann. Die Orbitolinen-Mergel können der Basis des Apt zugeordnet werden, wie das Auftreten von *Palorbitolina lenticularis lenticularis* (BLUMENBACH) zeigt.

Der unterste Teil des Schrattenkalkes in den bearbeiteten Profilen gehört vermutlich zum obersten Barrême.

Der Schrattenkalk entwickelt sich allmählich aus den Drusberg-Schichten (Mittelbarrême, FUCHS 1971, WIEDMANN 1978).

Die Sedimente des Profils Ill-Schlucht bei Feldkirch werden als Slope-Ablagerungen interpretiert. Die resedimentierte Folge gehört zur Lenticularis-Zone (Oberbarrême bis Unterapt).

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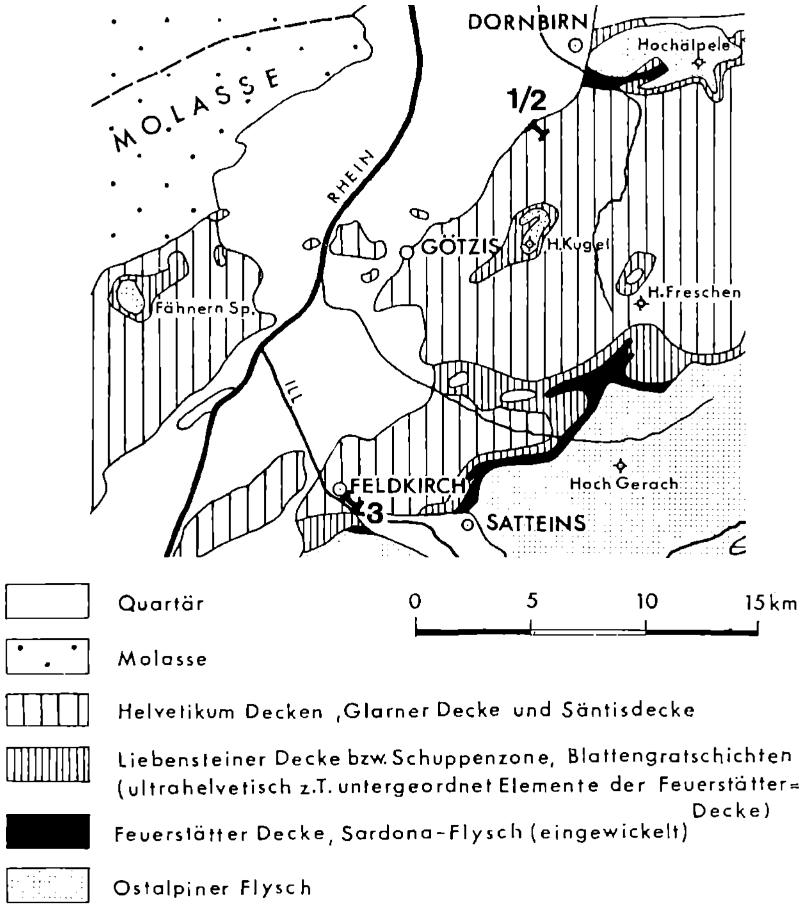
1. Einleitung

Mikrofauna und -flora von drei Profilen der zum nordöstlichen Teil der Säntis-Decke gehörenden Oberbarrême-Unterapt Schrattenkalk-Formation in Vorarlberg (Unterklien 1, Unterklien Rhomberg 2 und Hintere Ill-Schlucht, Feldkirch) kamen im Rahmen der österreichisch-ungarischen Zusammenarbeit 1986/87 zur Bearbeitung. Die Proben stammen aus Aufsammlungen von G. CSASZAR (1984/85).

Den Mikrofaunen- und Florenuntersuchungen lagen etwa 140 Dünnschliffe und einige Schlämmrückstände zugrunde, die in ihrer Mehrheit verhältnismäßig gut erhaltene Assoziationen enthielten. Berücksichtigt wurden Foraminiferen, Kalkalgen, Metazoen-Bruchstücke und Reste incertae sedis.

1.1 Historisches

Die erste grundlegende Arbeit stammt von VACEK (1879), der die regionalen Faziesänderungen des Schrattenkalks erkannt hat.



Text-Fig. 1. Die Lage der Profile Unterklien (1/2) und Ill-Schlucht (3), nach PREY (1968).

Die Aufmerksamkeit von MESSMANN (1925) galt den tektonischen und stratigraphischen Zusammenhängen beiderseits des Rheins. Dann legte die reiche Fossilsammlung von FUSSENEGGER den Grundstein zur Stratigraphie (Sammlung der Naturschau in Dornbirn).

Aufgrund der Änderungen des Makrofossilgehalts gliedern HEIM & BAUMBERGER (1933) die Schichtenfolge von Unterklien in fünf Einheiten. Die neueren Ergebnisse der Mikropaläontologie wurden von OBERHAUSER (1958, 1963) zusammengefaßt. LIENERT (1965) beschäftigt sich mit der Stratigraphie der Drusberg- und Schrattenkalk-Formation auf der Grundlage der Orbitolinen.

Als basales Schichtglied der Drusberg-Formation wurden von FUNK (1969, 1971) die Altmann-Schichten beschrieben.

CONRAD (1969) bearbeitete die Mikrofauna, die Mikrofazies und die Stratigraphie des Urgonkalkes der Umgebung von Genf sowie 1977 dessen

Kalkalgen. FUCHS (1971) bearbeitete die Mikrofauna der Drusberg-Schichten des Gebietes von Hohenems und bestimmte das Alter der ältesten Schichten der Drusberg-Formation bei Ranzenberg aufgrund von Foraminiferen als älteres Mittelbarrême. Dieses Alter wird durch Ammoniten-Nuklei aus Schlämmproben bestätigt (WIEDMANN 1978).

BRIEGEL (1972) behandelte die Drusberg- und Schrattenkalk-Formation der östlichen Alviergruppe. ZACHER (1973) faßte die erzielten Ergebnisse zusammen und beschäftigte sich mit faziellen und stratigraphischen Fragen des Schrattenkalks.

Von ARNAUD-VANNEAU und ARNAUD wurden die Urgonkalksteine des Barrême-Bedouliens Südostfrankreichs intensiv und aus verschiedenen Blickwinkeln studiert. Die mikrofaunistische Bearbeitung und die stratigraphische Gliederung wurden von ARNAUD-VANNEAU (1975, 1980), ARNAUD-VANNEAU et al. (1976) erstellt. Die Kalkalgen-Flora des Vercors wurde von MASSE (1976) bearbeitet.

ARNAUD-VANNEAU & ARNAUD (1978) korrelierten die Plattform- und Beckenbildungen (Chaînes Subalpines Septentrionales - Bassin Vocontien), ARNAUD-VANNEAU (1979) beschäftigte sich mit den Foraminiferen-Assoziationen und den Paläo-Milieus der Urgon-Plattformen.

FELBER & WYSSLING (1979) behandelten stratigraphische und tektonische Fragen des Hinteren Bregenzer Waldes. Sie führten als laterales Äquivalent des Schrattenkalks den Begriff der Mittagspitz-Formation ein und beschrieben das Typusprofil. FUNK & BRIEGEL (1979) setzten sich mit faziellen Fragen der Schrattenkalk-Formation der Osthelvetischen Decke auseinander. H. SCHOLZ (1979) bearbeitete die Bioherme und Biostrome des Allgäuer Schrattenkalks in stratigraphischer Hinsicht. Stratigraphische Fragen der ostalpinen Urgonplattformen des Barrême-Bedoule und des Vocontischen Beckens wurden von ARNAUD (1981) erörtert. ARNAUD-VANNEAU & DARSAC (1984) beschäftigten sich mit der Evolution der benthonischen Foraminiferen und mit ihrer stratigraphischen Rolle (Alpes du Nord, Frankreich).

ALEXANDER et al. (1984) fassen die nicht publizierten Diplomarbeiten der Studenten der Münchner Technischen Universität zusammen. WYSSLING (1986) beschäftigt sich mit Fragen der Stratigraphie, der Sedimentologie und der Paläogeographie der älteren Unterkreide-Bildungen des Helvetischen Schelfs. Er behandelte die mit dem untersten Schichtglied der Drusberg-Schichten altersgleichen Altmann-Schichten (FUNK 1969, 1971) mit einer reichen Ammonitenfauna. Diese werden in die Pulchellia compressissima-Zone des Unteren Barrême eingegliedert.

CSASZAR (1986) beschrieb den lithostratigraphischen Aufbau der in der vorliegenden Abhandlung behandelten drei Vorarlberger Profile.

GEBHARDT (1985), FÖLLMI (1986) und FÖLLMI & OUWEHAND (1987) setzten sich mit den Grünsanden (Garschella-Formation) auseinander, die den Schrattenkalk überlagern. In Vorarlberg wurde in verschiedenen Profilen *Deshayesites* sp. und *Dufrenoyia furcata* (SOWERBY) gefunden (FÖLLMI 1986). Diese Formen weisen auf das mittlere und spätere Früh-Apt hin. Die Untergrenze der Luitere-Schichten ist heterochron und wird von Ost nach West im allgemeinen jünger (FÖLLMI & OUWEHAND 1987).

Die Entwicklung des distalen Osthelvetischen Schelfs im Barrême und Früh-Apt (Drusberg-, Mittagspitz- und Schrattenkalk-Formation) in Vorarlberg und im Allgäu wurden von BOLLINGER (1988) in seiner Dissertation studiert. Mit den faziellen Verhältnissen des Schrattenkalks der drei hier bearbeiteten Profile haben sich jüngst CSASZAR (1986) und BOLLINGER (1988) eingehend auseinandergesetzt.

1.2 Problemstellung

Im Rahmen der österreichisch-ungarischen geologischen Zusammenarbeit und des IGCP-Projekts 262 "Tethyan Cretaceous Correlation" wurden in Ergänzung zu den lithofaziellen Profilaufnahmen von G. CSASZAR drei Schrattenkalk-Profile in Vorarlberg mit Hilfe von Foraminiferen und Dasycladaceen stratigraphisch bearbeitet.

In einer weiteren Projektphase ist ein Vergleich der Schrattenkalk-Profile von Unterklien und der Stratotypus-Profile der Nagyharsány Kalkstein-Formation Süd-Ungarns vorgesehen.

2. Allgemeine Charakterisierung der Mikrofauna und -flora

Der Fossilinhalt des Schrattenkalks setzt sich vor allem aus Bivalven, Echinodermen, Foraminiferen, Resten incertae sedis sowie Kalkalgen zusammen.

2.1 Foraminiferenfauna

Von den 78 isolierten Taxa können 50 lediglich generisch und 31 artlich identifiziert werden, wobei zwei *Orbitolinopsis*-Arten neu und einer späteren Beschreibung vorbehalten sind (Text-Fig. 4-6).

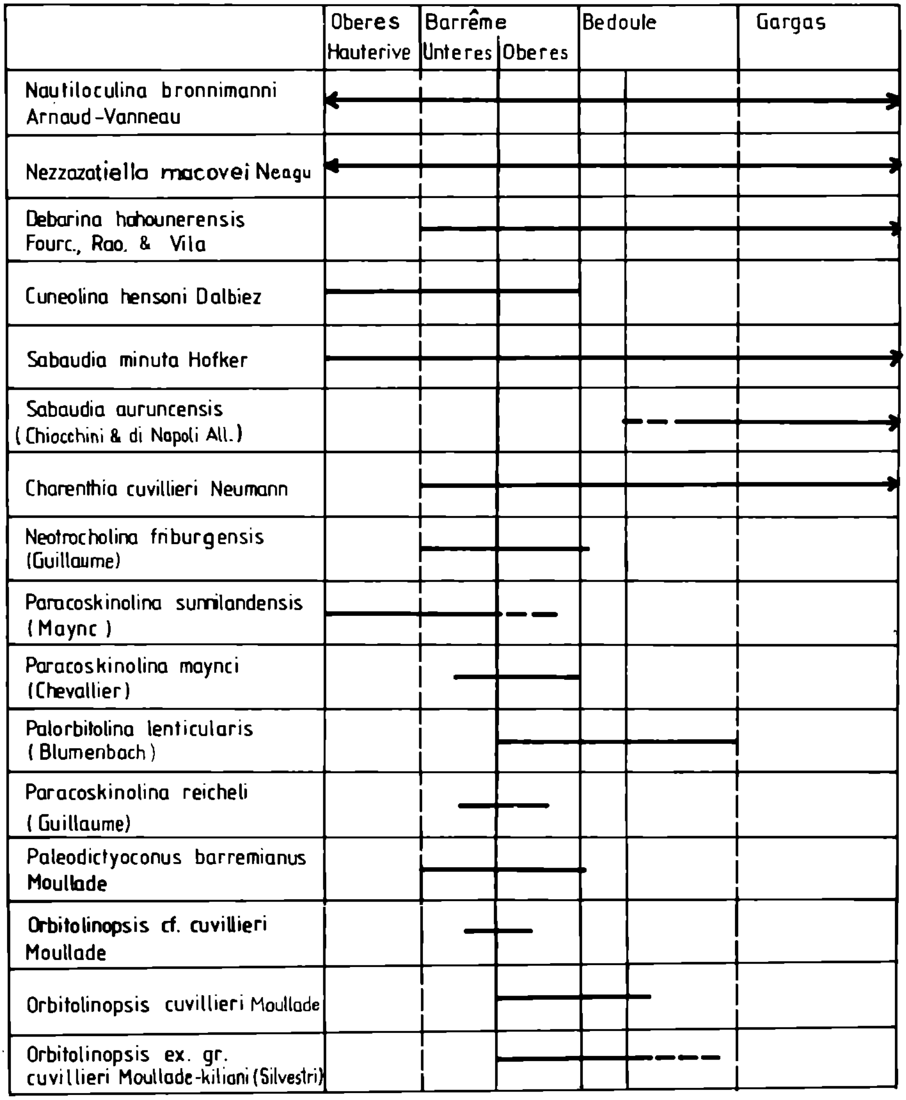
Die Foraminiferenfauna besteht fast ausschließlich aus benthischen Formen, wobei hinsichtlich der Gattungs- und Artenzahl das agglutinierende Benthos vorherrscht. Hier sind 28 Taxa nur generisch und 18 artlich bestimmbar. Vom kalkschaligen Benthos waren 21 Taxa nur generisch und 11 artlich identifizierbar, darunter sind die Miliolinen in der Mehrzahl der Proben häufig bis massenhaft vertreten.

Das Plankton ist durch schlecht erhaltene Exemplare von 2 Arten der Gattung *Hedbergella* (*H.* cf. *infracretacea*, *H. planispira*) sporadisch vertreten.

Verteilung der Foraminiferenfauna:

	Gattung		Art	
	Taxazahl	%	Taxazahl	%
Plankton	1	2	2	6,5
Kalkschaliges Benthos	21	42	11	35,5
Agglutinierendes Benthos	28	56	18	58,0
Insgesamt	50	100	31	100

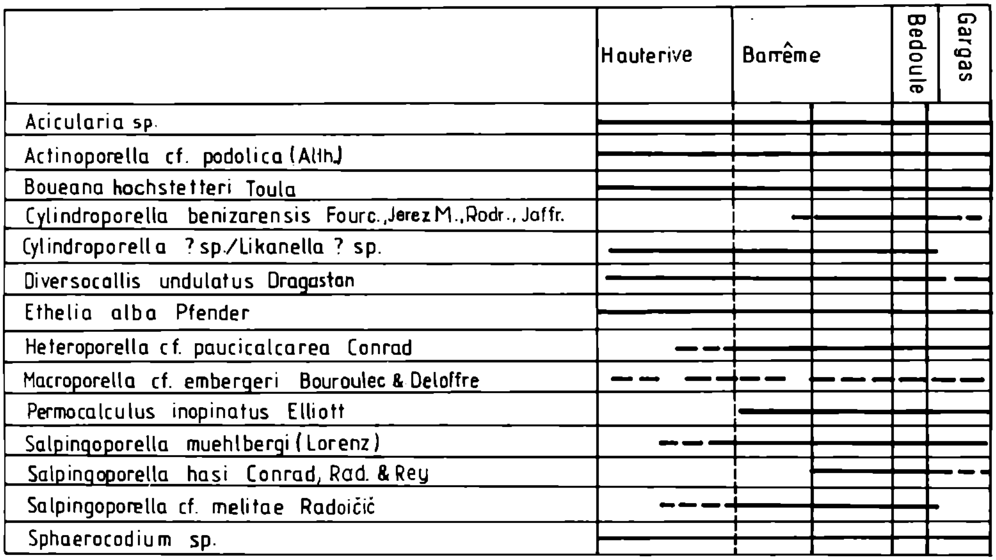
Die chronologische Verbreitung der wichtigsten Foraminiferen-Arten wird in Text-Fig. 2 dargestellt.



Text-Fig. 2. Chronologische Verbreitung der wichtigsten Foraminiferen-Arten (nach LIENERT 1965, SCHROEDER et al. 1968, CONRAD 1968, 1969, SCHOLZ 1979, 1984, ARNAUD-VANNEAU 1980, 1981, ARNAUD-VANNEAU & DARSAC 1984, MOULLADE et al. 1985).

2.2 Kalkalgen

Von den 16 isolierten Taxa können 13 generisch und lediglich 12 artlich bestimmt werden (Text-Fig. 4-6). Die chronologische Verbreitung der Mehrheit der Arten wird in Text-Fig. 3 dargestellt.



Text-Fig. 3. Die chronologische Verbreitung der Kalkalgen (nach CONRAD 1969, 1970, 1978, PEYBERNES & CONRAD 1970, MASSE 1976).

3. Beschreibung der Profile

3.1 Unterklien, Steinbruch Rhomberg 1 (Text-Fig. 4)

In der Foraminiferenfauna des 77,5 m mächtigen Profils (34 Proben) herrscht das agglutinierende Benthos vor. Aufgrund der mikrofaunistischen Änderungen kann das Profil in drei Einheiten gegliedert werden:

Einheit I

Von 0,0 bis 17,5 m (Proben 34-29); - als Untergrenze gilt die älteste Schicht (34) des Aufschlusses, die Obergrenze wurde mit dem Letztauftreten von *Trocholina friburgensis* gezogen. Laut den Daten von ARNAUD-VANNEAU (1980) überschreitet diese Art in Südostfrankreich die Obergrenze des unteren Apt nicht. Nur in der ältesten Probe sind die Miliolinen und Valvulamminen massenhaft vertreten.

Begleittaxa:

Glomospira/Glomospirella
Nezzazatiella macovei NEAGU
Sabaudia minuta (HOFKER)
Trocholina friburgensis (GUILLAUME & REICHEL)
Marssonella praeoxycona MOULLADE
Nautiloculina bronnimanni ARNAUD-VANNEAU
Textulariidae

Orbitolinenarten:

Orbitolinopsis ex gr. *cuvillieri-kiliani*
Orbitolinopsis sp.
Palorbitolina lenticularis praecursor (MONTANARI)

Kalkalgenarten:

Diversocallis undulatus DRAGASTAN
Ethelia alba PFENDER
Salpingoporella hasi CONRAD, RADOIČIČ & REY
Salpingoporella sp.
Sphaerocodium sp.

Metazoen:

Vorherrschend sind Bivalven-Bruchstücke, untergeordnet Bruchstücke von Echinodermaten, Kalkschwämmen und Gastropoden. Daneben treten Acicularien und Holothurien auf.

Einheit II

Von 17,5 bis 55,0 m (Proben 28-12); - Untergrenze der Einheit: Letztauftreten von *Trocholina friburgensis* (GUILLAUME & REICHEL); Obergrenze der Einheit: Einsetzen von *Sabaudia auruncensis* CHIOCCHINI & DI NAPOLI ALL.

Palorbitolina lenticularis lenticularis (BLUMENBACH) tritt mit schlecht erhaltenen Exemplaren erst in den Orbitolinen-Schichten auf (Probe 20) und setzt in Schicht 15 wieder aus. Nach Angaben von ARNAUD-VANNEAU (1980) und BOLLINGER (1986) setzt sie erst im früheren Unterapt ein.

Orbitolinenassoziation:

Orbitolinopsis sp.
Orbitolinopsis cuvillieri MOULLADE
Orbitolinopsis ex gr. *cuvillieri-kiliani*
Palorbitolina lenticularis lenticularis (BLUMENBACH)
Palorbitolina lenticularis praecursor (MONTANARI)
Paleodictyoconus barremianus MOULLADE
Paracoskinolina maynci (CHEVALLIER)
Paracoskinolina sunnilandensis MAYNC
Palorbitolina sp. I (Taf. 1, Fig. 1)

Andere agglutinierende Großforaminiferen:

Choffatella decipiens SCHLUMBERGER
Pseudocyclammina lituus YOKOYAMA
Pseudocyclammina hedbergi MAYNC
Pseudocyclammina sp.
Pseudotextulariella sp.

Miliolinen sind massenhaft vertreten.

Algenflora:

Cylindroporella benizarensis FOURC., JER. M., RODR. & JAFR.
Cylindroporella sp.
Camptocotylodon fontis PATRULIUS
Dasycladaceae
Macroporella cf. *embergeri* BOUR. & DELOFFRE
Salpingoporella muehlbergi (LORENZ)

Salpingoporella melitae RADOIČIĆ (Taf. 4, Fig. 38).
Salpingoporella sp.
Permocalculus inopinatus ELLIOTT

Metazoen:
Wie in Einheit I finden sich untergeordnet noch Kalkschwämme und Hydrozoen-Bruchstücke.

Weiter finden sich in der Begleitfauna *Aeliosaccus* sp., *Bacinella irregularis* RAD., *Cadosina*, Ostracoda, Holothurioidea und incertae sedis "Pseudostracoda".

Einheit III

Von 55,0 bis 77,5 m (Proben 11-1); - es herrschen nach wie vor die agglutinierenden Foraminiferen vor, hinsichtlich der Individuenzahl dominieren die Milioliden, in den jüngeren Schichten nimmt die Diversität rasch ab.

Es ist auffallend, daß in der Orbitolinen-Assoziation die großwüchsige *Palorbitolina lenticularis praecursor* und von Probe 4 an die Kalkalgen eine vorherrschende Rolle spielen.

Algenflora:
Camptocotylodon fontis PATR.
Permocalculus inopinatus ELLIOTT
Salpingoporella muehlbergi (LORENZ)
Salpingoporella hasi CONR., RAD. & REY
Salpingoporella melitae RADOIČIĆ
Salpingoporella sp.

Orbitolinenarten:
Orbitolinopsis cuvillieri MOULLADE
Paracoscinolina sunnilandensis MAYNC

Bei den Metazoen dominieren die Bivalven, sporadisch kommen Gastropoden und Kalkschwämme vor.

Die Taxa incertae sedis sind sporadisch durch *Aeliosaccus* und *Cadosina* vertreten.

3.2 **Unterklien, Steinbruch Rhomberg 2** (Text-Fig. 5)

Im Steinbruch bei Unterklien ist eine wohlgebankte inverse Schichtfolge ausgezeichnet aufgeschlossen, die von der Drusberg-Formation über den Schrattenkalk bis zur Seewerkalk-Formation reicht. Die Schichtfolge umfaßt also Sedimente vom Barrême bis in das Turon.

3.2.1 Drusberg-Formation

Von 0,0 bis 8,0 m (Proben 67-65); - im Steinbruchbereich ist als ältestes Gestein der oberste Teil der Drusberg-Formation, ein dunkelgrauer flaseriger und bioturbater Kalkstein in einer Mächtigkeit von 8 m aufgeschlossen.

Mikrofaunen-Assoziation

Sie wird durch eine benthische Foraminiferenfauna von geringer Diversität, mit verhältnismäßig wenigen Arten und Individuen repräsentiert. Die Foraminiferenfauna besteht aus Miliolinen, Glomospirellen, Valvulamminen und Textulariiden, unter denen kleinwüchsige agglutinierende benthische Formen überwiegen. Auffallend ist das Fehlen von planktonischen Foraminiferen, Großforaminiferen (Orbitolinen) und Kalkalgen. Außer Foraminiferen enthält dieser Teil des Profils nur wenige Cadosinen, Globochaeten und Pieninen.

Kennzeichnende Arten sind:

Debarina hahounerensis FOURC., RAO. & VILL.
Marssonella praeoxycona MOULLADE
Nezzazatiella macovei NEAGU

Begleitfauna:

Miliolinen
Arenobulimina? sp.
Glomospirella sp.
Dorothia sp.
Spiroplectammina sp.
Valvulammina sp.
Textularia sp.

Die Metazoen sind durch Mollusken-Schalenbruchstücke, Echiniden-Stacheln und Crinoiden-Bruchstücke vertreten.

Eine exakte stratigraphische Einstufung ist nicht möglich.

3.2.2 Schrattenkalk-Formation

Aus den Drusberg-Schichten entwickelt sich allmählich, unter Aufhören des flaserigen Gefüges, die etwa 120 m mächtige Schrattenkalk-Formation.

Aufgrund von Änderungen in der Makrofossilführung und der lithologischen Merkmale wurde sie von G. CSASZAR (1986) in 7 Einheiten unterteilt. Die Mächtigkeit der Einheiten ist - infolge von Faziesänderungen - in hohem Maße unterschiedlich.

Aufgrund der Änderungen der Mikrofauna - vor allem der Foraminiferenfauna - kann sie in drei biostratigraphische Einheiten gegliedert werden, innerhalb welcher noch weitere Horizonte unterschieden werden können (Text-Fig. 5).

Einheit I

Von 8,0 bis 44,0 m (Proben 64-45); - als Hauptkennzeichen gilt das Auftreten von Orbitolinen. Diese treten in Probe 61 sporadisch mit schlecht erhaltenen Exemplaren von mikritisierten Schalen auf. In Probe 58 erscheint erstmals *Palorbitolina lenticularis praecursor* (MONTANARI), in Probe 55 treten die ersten Orbitolinopsiden auf. Die Orbitoliniden erreichen ihre größte Häufigkeit in Probe 56.

In der ausschließlich aus benthischen Formen bestehenden Foraminiferenfauna nimmt die Zahl der Arten allmählich zu, es überwiegen die agglutinierenden Arten. Hinsichtlich der Individuenzahl herrscht das kalkschalige

Benthos und innerhalb desselben die Miliolinen-Arten (häufig-massenhaft) mit mittel- und großwüchsigen Exemplaren vor. Von Probe 59 an tritt *Nummoloculina heimi* (BONET) massenhaft auf.

Charakteristische Taxa der kalkschaligen benthischen Foraminiferen sind (Taf. 2, Figs. 13-23; Taf. 3, Figs. 24-35):

Trocholina friburgensis (GUILLAUME & REICHEL)
Nezzazatiella macovei NEAGU
Cyclogyra? sp.
Sabaudia minuta (HOFKER)

Agglutinierende benthische Taxa sind vertreten durch:

Choffatella decipiens (SCHLUMBERGER)
Debarina hahounerensis FOURC., RAO. & VILL.
Glomospirella sp.
Textulariidae
Placopsilina cenomana (D'ORBIGNY)
Pseudocyclammina lituus (YOKOYAMA)
Valvulammina aff. *picardi* HENSON
Valvulammina sp.
Orbitolinopsis sp.
Orbitolinidae
Palorbitolina lenticularis praecursor (MONTANARI) (Taf. 1, Fig. 2)

Eine mittlere Individuenzahl wird nur von den Textulariidae (*Textularia, Spiroplectammina, Dorothia*), von den *Valvulammina*-Arten und den Orbitolinidae erreicht. *Cuneolina hensoni* DALB. kommt nur mit wenigen Exemplaren in Probe 45 vor.

Die ersten Kalkalgen erscheinen in der Mitte der Einheit (Proben 56 und 55): *Ethelia alba* (PFENDER), *Acicularia* sp.; in Probe 53 treten sporadisch die ersten Salpingoporellen mit *Salpingoporella muehlbergi* (LORENZ), *Salpingoporella hasi* CONRAD, RADOICIC & REY auf.

Aufgrund der Dominanzverhältnisse der Mikrofauna können 3 Horizonte unterschieden werden:

1. Eine Miliolinen-Glomospirellen-Valvulamminen-Textulariiden-Assoziation mit verhältnismäßig vielen Bivalven-Bruchstücken, jedoch ohne Kalkalgen (Proben 65-60). Die Diversität ist mittelgroß, die dominanten Miliolinen sind nur in den Proben 64 und 62 häufig bis massenhaft vertreten. Die Orbitolinen treten in den Proben 61-60 sporadisch auf.

2. Eine Miliolinen-Sabaudien-Orbitolinen-Textulariiden-Assoziation mit Kalkalgen, zahlreichen Bivalven-, Crinoiden- und wenigen Echinoiden- und Sphinctozoen-Bruchstücken (Proben 60-53). Massenhaft sind die großwüchsigen Miliolinen, unter diesen *Nummoloculina heimi* BONET; Orbitolinen sind mit einer höheren Individuenzahl auf Probe 56 beschränkt.

Sabaudia minuta (HOFKER) ist in diesem Horizont am häufigsten. Weitere begleitende Foraminiferenarten:

Trocholina friburgensis (GUILLAUME & REICHEL)
Nezzazatiella macovei NEAGU
Valvulammina aff. *picardi* HENSON

Die Diversität der Textulariiden ist klein, sie erreichen in der Probe 54 eine mittlere Individuenzahl, wo sie gemeinsam mit *Choffatella decipiens* vorkommen.

In diesem Horizont treten die ersten Kalkalgen auf:

Salpingoporella muehlbergi (LORENZ)
Salpingoporella hasi CONRAD, RADOIČIĆ & REY
Ethelia alba PFENDER

Andere, sporadisch auftretende, aber durchgehend anzutreffende Formen sind:

Globochaete
Pieninia oblonga MISIK
Holothurien-Sklerite
Serpula
Spiroserpula

3. Eine Miliolinen-Trocholinen-Textulariiden-Assoziation mit wenigen Salpingoporellen-Bruchstücken (Proben 52-45). Eine starke Abnahme der Faunen-Diversität und -Dichte der Foraminiferen ist für sie kennzeichnend, ohne das Auftreten von neuen Arten.

Dominante Elemente der Metazoen-Bruchstücke sind Bivalven und Bryozoen sowie in untergeordneter Zahl Crinoiden, Brachiopoden, Inozoen und Sphinctozoen. In den Proben 49-48 kann *Barroisia helvetica* (DE LORIOL) erkannt werden. Dieser Profilabschnitt entspricht dem makroskopisch bestimmten Horizont mit *Lopha rectangularis* ("Alectryonien-Schichten" auct.).

Einheit II

Von 44,0 bis 74,0 m (Proben 45-26); - sie kann als eine Orbitolinen-Miliolinen-Assoziation mit vielen Bivalven-, Crinoiden- und Bryozoen-Bruchstücken angesehen werden. In der diversen benthischen Foraminiferenfauna sind hinsichtlich der Individuenzahl die Miliolinen-Arten *Quinqueloculina robusta*, *Q. danubiana*, *Quinqueloculina* sp., *Spiroloculina cretacea*, *Massilina* und *Nummoloculina heimi* vorherrschend. Unter diesen ist die letztere am häufigsten.

Daneben treten die folgenden kalkschaligen benthischen Foraminiferen auf:

Trocholina friburgensis (GUILLAUME & REICHEL)
Nezzazatiella macovei NEAGU
Sabaudia minuta (HOFKER)
Meandrospira sp.

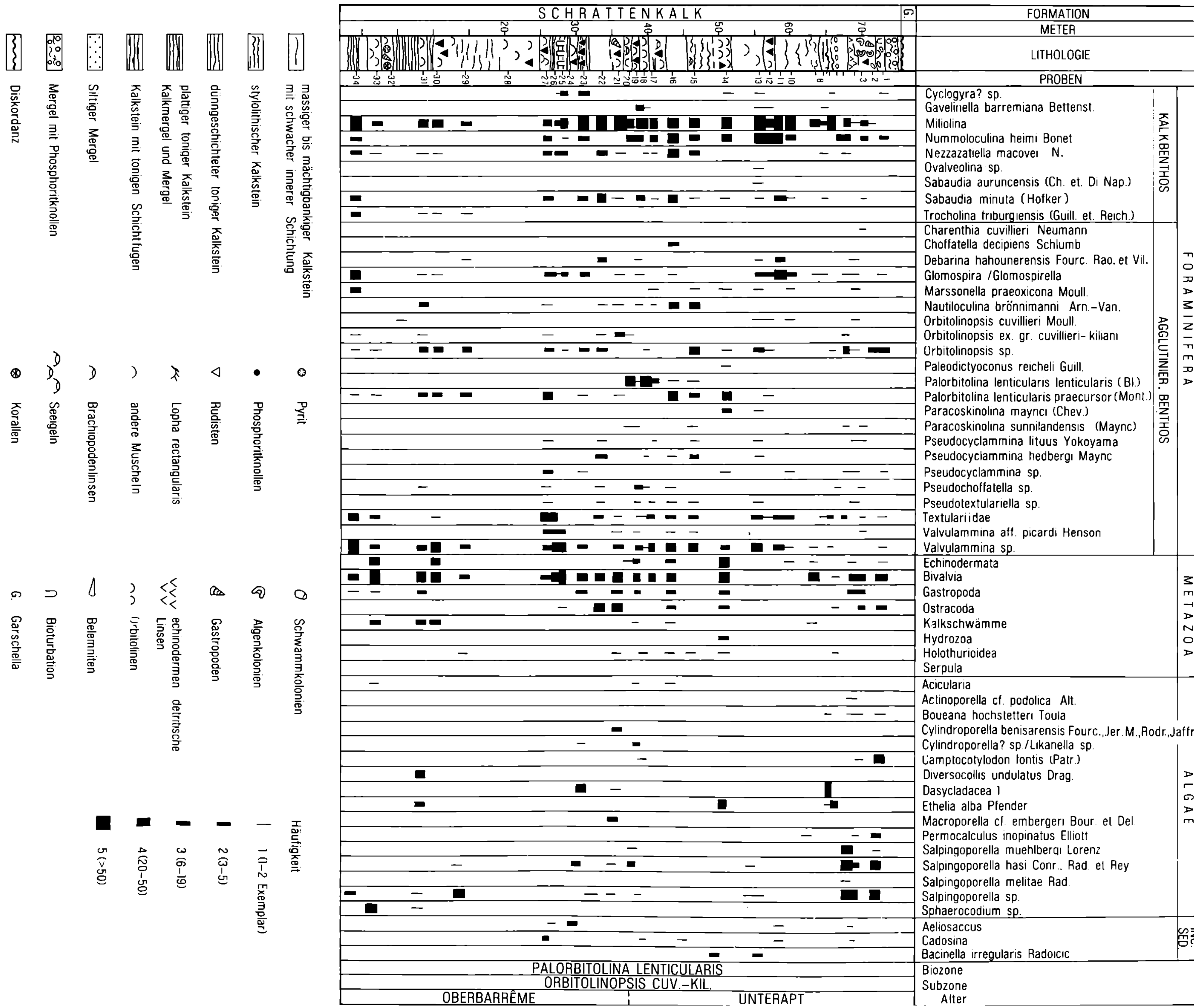
Orbitolinen-Vergesellschaftung:

Orbitolinopsis cuvillieri MOULLADE
Orbitolinopsis sp.
Palorbitolina lenticularis praecursor (MONTANARI)
Palorbitolina lenticularis lenticularis (BLUMENBACH)
Paracoskinolina sunnilandensis MAYNC
Paleodictyoconus barremianus MOULLADE

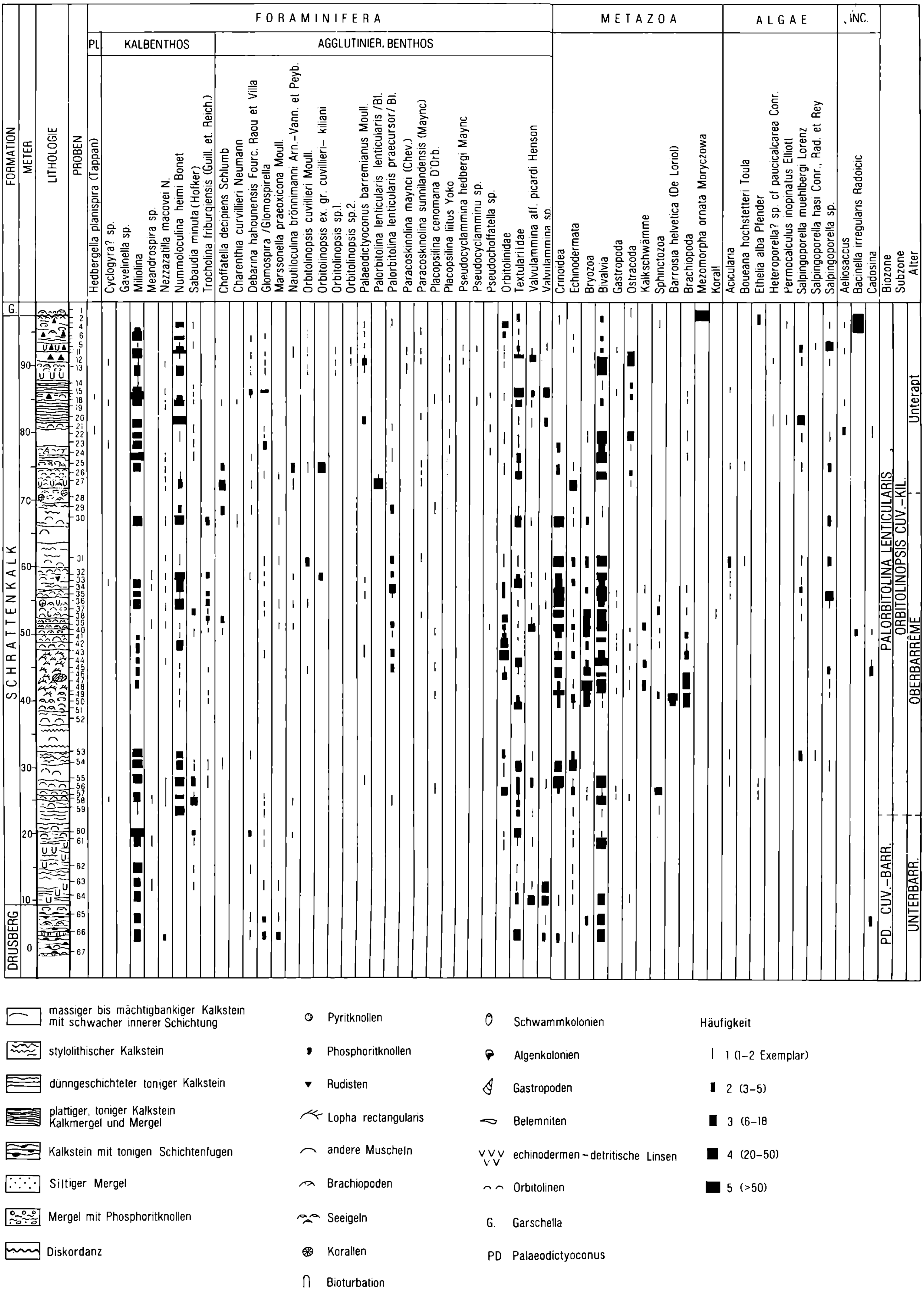
Orbitolinen sind nur in Probe 43 häufig anzutreffen. *Palorbitolina lenticularis praecursor* (MONT.) erscheint in der ältesten Schicht der Einheit (Probe 45) mit *Orbitolinopsis cuvillieri* zusammen. In Probe 26 ist *Palorbitolina lenticularis lenticularis* (BLUMENBACH) häufig und *Choffatella decipiens* relativ häufig.

Andere kennzeichnende Begleitarten der agglutinierenden Foraminiferenfauna sind:

Choffatella decipiens SCHLUMBERGER
Marssonella praeoxycona MOULLADE
Nautiloculina bronnimanni ARNAUD-VANNEAU



Text-Fig. 4. Profil Unterklien, Steinbruch Rhomberg 1.



Text-Fig. 5. Profil Unterklien, Steinbruch Rhomberg 2.

Placopsilina cenomana D'ORBIGNY
Valvulammina aff. *picardi* HENSON
Pseudochoffatella sp.

Die Kalkalgen sind durch *Salpingoporella muehlbergi* (LORENZ) in den ältesten Proben der Einheit mit einer niedrigen Individuenzahl vertreten (Taf. 4, Figs. 36, 37, 41). Die Metazoenfauna setzt sich meist aus Bivalven- und Crinoiden-Bruchstücken zusammen, untergeordnet treten Bryozoen- (Taf. 4, Fig. 39), Echiniden-, Gastropoden-, Kalkschwamm-, Sphinctozoen-, Brachiopoden- und Korallen-Bruchstücke auf. Ostracoden (Taf. 4, Fig. 40), Cadosinen, *Aeliosaccus* sp. und *Bacinella irregularis* RADOICIC kommen selten vor.

CSASZAR (1986) hat die orbitolinenreichen Schichten 27-26 dieses Profils mit der Orbitolinen-Mergelschicht 20 des Profils 1 korreliert.

Charakterisierung der Schichten 27-26:

Die älteste Schicht besteht aus einem Orbitolinen-Kalkmergel, auf welchen, mit einem Übergang von bioturbaten Kalkmergeln (Schicht 26), ein Kalkstein mit Seeigelstacheln und Brachiopoden folgt.

In der Foraminiferenfauna ist die Dominanz der Großforaminiferen auffällig:

Palorbitolina lenticularis lenticularis (BLUMENBACH)
Choffatella decipiens SCHLUMBERGER
Nummoloculina heimi BONET

Es fällt außerdem das Fehlen von klein-/mittelwüchsigen Miliolinen und Kalkalgen auf. Andere kennzeichnende Begleitelemente:

Orbitolinopsis sp.
Paracoskinolina sunnilandensis MAYNC

Nach vorliegenden Untersuchungen setzt *Palorbitolina lenticularis lenticularis* in Profil 1 in Probe 15 aus. Die Orbitolinen-Schichten sind daher von Schicht 20-15 repräsentiert.

Einheit III

Von 74,0 bis 98,7 m (Proben 25-1); - die Grenzen dieser Einheit wurden mit dem Auftreten der neuen *Orbitolinopsis*-Arten gezogen.

Untergrenze: Auftreten von *Orbitolinopsis* ex gr. *cuvillieri-kiliani*
Obergrenze: Auftreten der *Orbitolinopsis* sp. 1, *Orbitolinopsis* sp. 2 und *Paracoskinolina maynci* (CHEVALLIER)

Die Verteilung des Benthos ist ähnlich wie in Einheit II, es treten aber auch neue Taxa hinzu:

Pseudocyclammina hedbergi MAYNC
Pseudocyclammina sp.

Gesteinsbildend sind große Miliolinen, vor allem *Nummoloculina heimi* BONET.

Die Kalkalgenflora setzt sich zusammen aus:

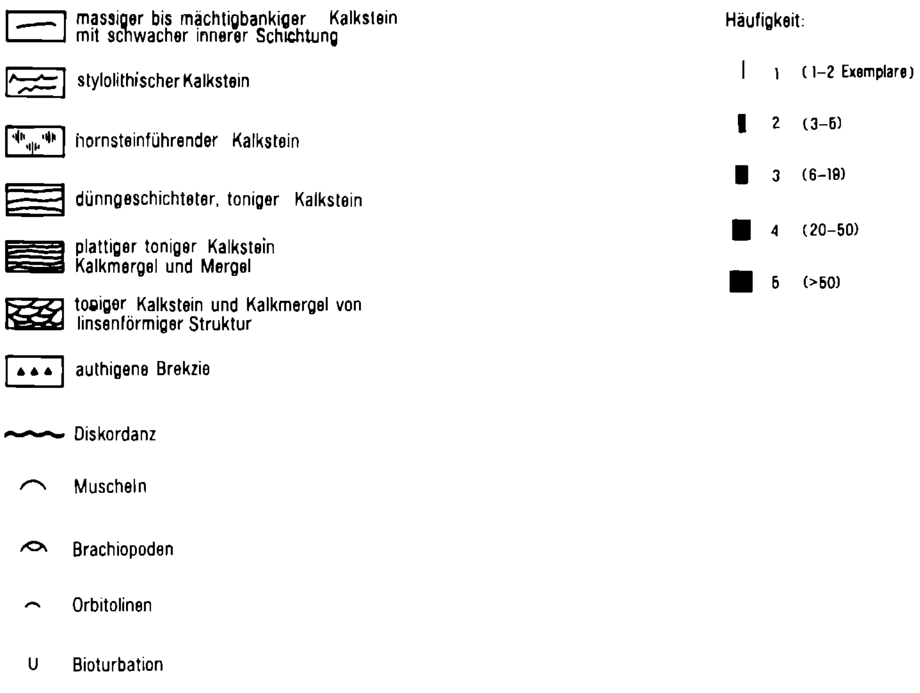
Salpingoporella muehlbergi (LORENZ)
Salpingoporella sp.
Heteroporella sp. ?*paucicalcarea* CONRAD
Permocalculus inopinatus ELLIOT und
Boueana hochstetteri TOULA

Die Metazoenfauna besteht zum überwiegenden Teil aus Bruchstücken von Bivalven, Crinoiden, Bryozoen, untergeordnet Platten und Stacheln von Echiniden, Brachiopoden und Gastropoden. Selten vorkommende Komponenten sind Cadosinen, Ostracoden, *Bacinella irregularis* RADOIČIĆ und *Aeliosaccus* sp. Die oberste Probe der Einheit (Probe 1) besteht aus einem Korallen-Bruchstück (*Mezomorpha ornata* MORYCZOWA).

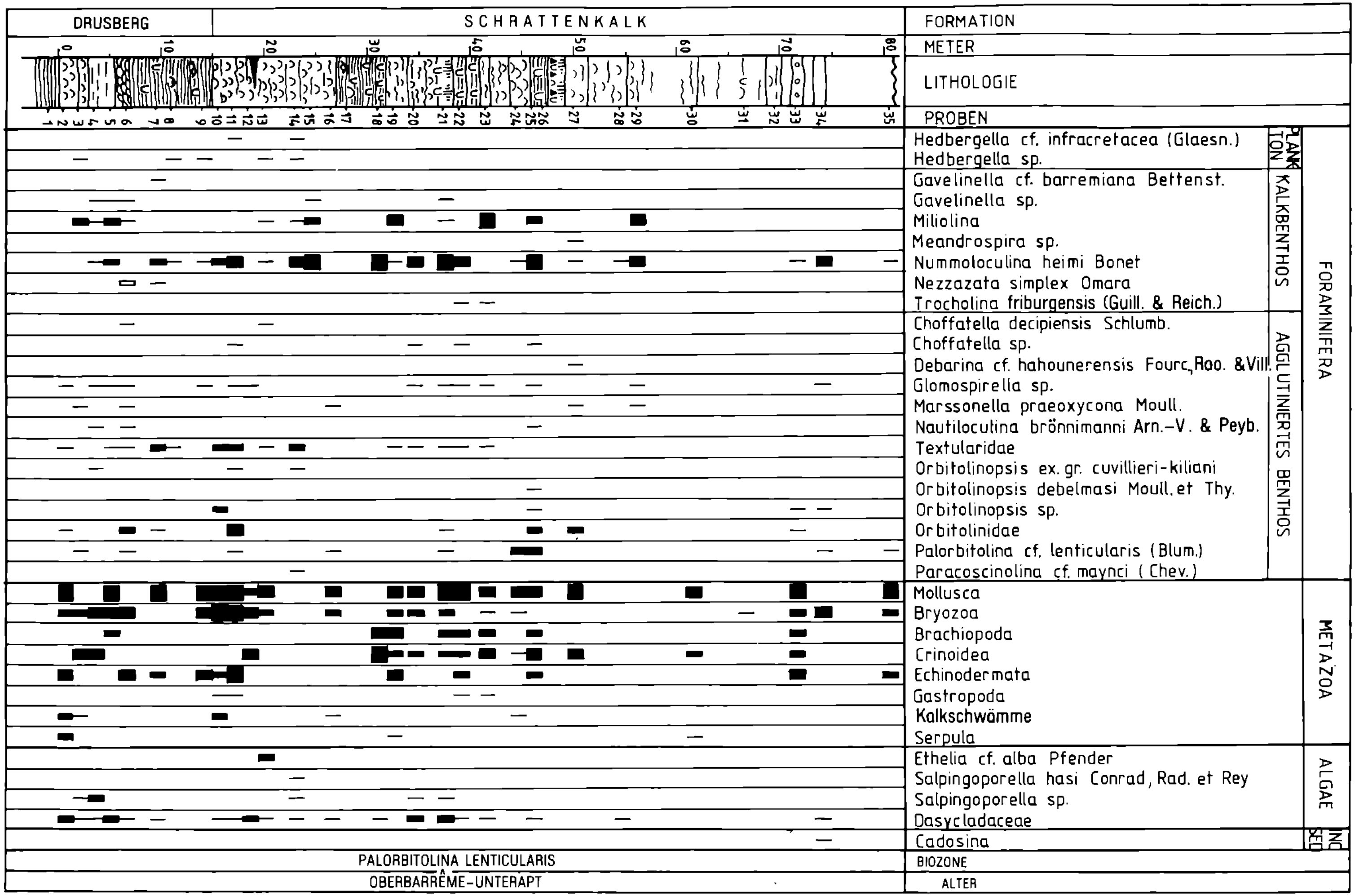
3.3 Hintere Ill-Schlucht bei Feldkirch

Im SE von Feldkirch ist am Ostufer des Flusses Ill längs der alten Straße Schrattenkalk aufgeschlossen. Hier ist erkennbar, daß sich der Schrattenkalk allmählich aus der Drusberg-Formation entwickelt (CSASZAR 1986). Seine Gesamtmächtigkeit beträgt ca. 80 m.

Der Schrattenkalk, der überwiegend in einer Kalksand-Beckenrand-Fazies ansteht, setzt sich zum großen Teil aus mittel- und grobkörnigen Litho- und Bioklasten zusammen. Vorherrschend ist ein sparitischer Zement. Die lithostratigraphische Beschreibung des Profils erfolgte durch CSASZAR (1986).



Text-Fig. 6. Profil Hintere Ill-Schlucht, Feldkirch.


DRUSBERG
SCHRATTENKALK
FORMATION
METER
0 10 20 30 40 50 60 70 80
LITHOLOGIE
PROBEN
1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26 27 28 29 30 31 32 33 34 35
Hedbergella cf. infracretacea (Glaesn.)
Hedbergella sp.
PLANKTON
Gavelinella cf. barremiana Bettenst.
Gavelinella sp.
Miliolina
Meandrospira sp.
Nummoloculina heimi Bonet
Nezzazata simplex Omara
Trocholina friburgensis (Guill. & Reich.)
KALKBENTHOS
Choffatella decipiensis Schlumb.
Choffatella sp.
Debarina cf. hahounerensis Fourc., Rao. & Vill.
Glomospirella sp.
Marssonella praeoxycona Moull.
Nautiloculina brönnimanni Arn.-V. & Peyb.
Textularidae
Orbitolinopsis ex. gr. cuvillieri-kiliani
Orbitolinopsis debelmasi Moull. et Thy.
Orbitolinopsis sp.
Orbitolinidae
Palorbitolina cf. lenticularis (Blum.)
Paracoscinolina cf. maynci (Chev.)
AGGLUTINIERTES BENTHOS
FORAMINIFERA
Mollusca
Bryozoa
Brachiopoda
Crinoidea
Echinodermata
Gastropoda
Kalkschwämme
Serpula
METAZOA
Ethelia cf. alba Pfender
Salpingoporella hasi Conrad, Rad. et Rey
Salpingoporella sp.
Dasycladaceae
ALGAE
Cadosina
INC SED
PALORBITOLINA LENTICULARIS
BIOZONE
OBERBARRÊME–UNTERAPT
ALTER

In der oberen Hälfte des Profils ist eine oft starke Stylolithisierung kennzeichnend. Das Vorkommen von Hornsteinlinsen oder -schichten ist auf den mittleren Abschnitt (Schichten 27-21) des Profils beschränkt.

Zum Zweck von Mikrofaunen-Untersuchungen wurde ein Schlämmrückstand aus den liegenden Drusberg-Schichten bearbeitet, der aber keine Mikrofauna enthielt. Aus der Schrattenkalk-Formation wurden zwei Schlämmrückstände (Proben 8 und 3) sowie 32 Dünnschliffe untersucht.

3.3.1 Schrattenkalk-Formation

LIENERT (1965) hat aus dem Aufschluß die folgenden Orbitolinen-Arten identifiziert: *Orbitolinopsis cuvillieri, Orbitolinopsis flandrini, Orbitolinopsis miliani, "Coskinolina" sunnilandensis, Dictyoconus reicheli, "Dictyoconus" arabicus, "Iraqia" hensoni.*

Die Dünnschliffe und Schlämmrückstände enthielten eine aus 33 Taxa bestehende, in ihrer Mehrheit stark mikritisierte, reiche Foraminiferenfauna von mittlerer bis schlechter Erhaltung. In dieser konnten 2 planktonische, 10 kalkschalige und 15 agglutinierende benthische Formen separiert werden (Text-Fig. 6). Unter diesen sind die nur sporadisch auftretenden Gattungen *Lenticulina, Dentalina, Nodosaria, Dorothia* und *Spiroplectammina* nicht angeführt, die Textulariidae wurden zusammengezogen.

Häufig sind Metazoen-Bruchstücke, hauptsächlich von Mollusken, Bryozoen, Brachiopoden und Echinodermen, untergeordnet auch von Gastropoden, Kalkschwämmen und Serpeln.

Die Kalkalgenflora besteht vorherrschend aus schlecht erhaltenen Resten von Salpingoporellen und anderen Dasycladaceen.

Stratigraphisch interessante Orbitolinenarten sind:

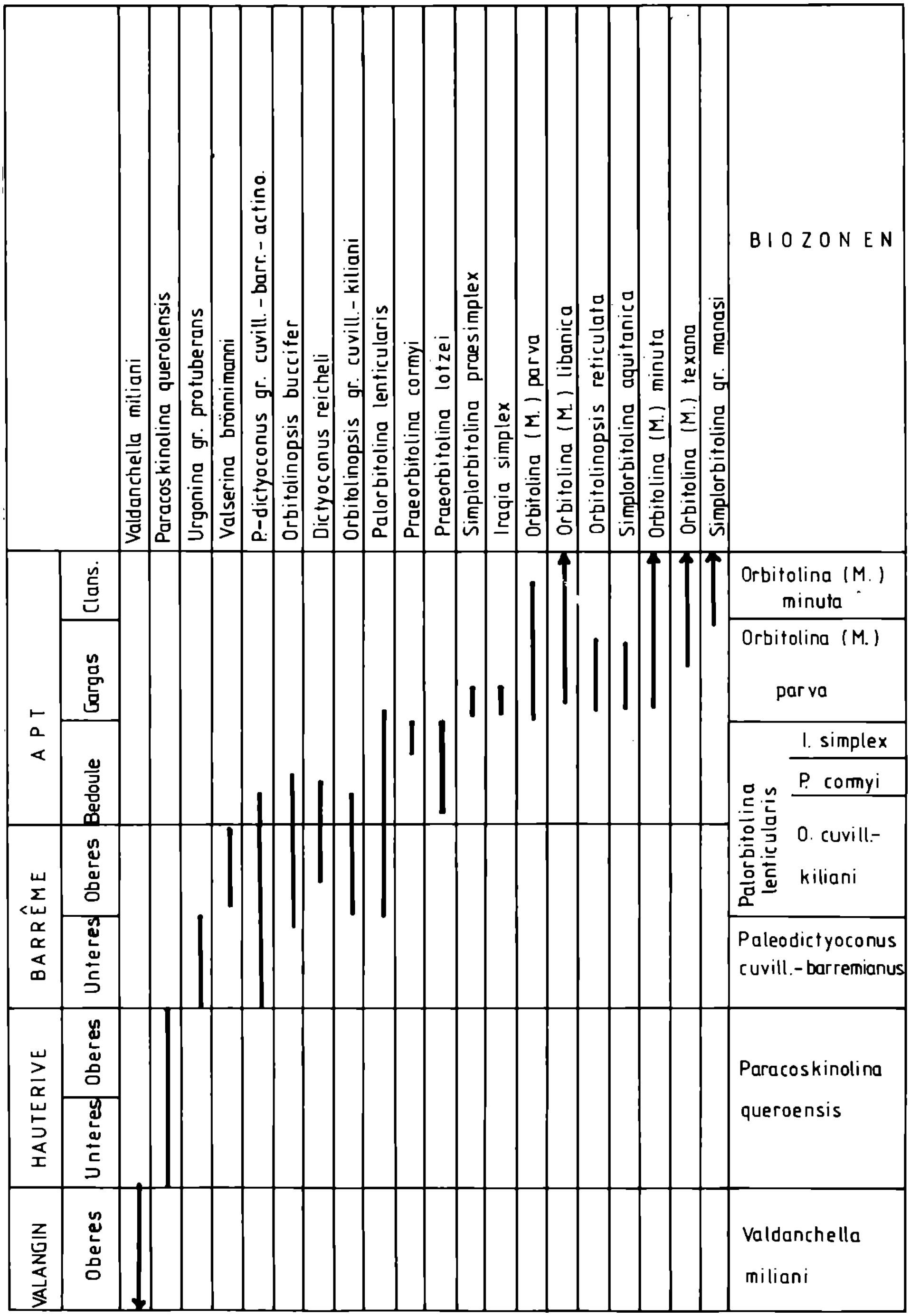
Orbitolinopsis ex gr. *cuvillieri-kiliani*
Palorbitolina cf. *lenticularis* (BLUMENBACH)
Palorbitolina lenticularis (BLUMENBACH) (Taf. 4, Fig. 40)
Paracoskinolina cf. *maynci* (CHEVALLIER) (Taf. 1, Fig. 10)
Orbitolinopsis cuvillieri MOULLADE

Die Zusammensetzung der Mikrofauna und der Kalkalgenflora ist jener des Steinbruchs Rhomberg ähnlich. Aufgrund dieser Orbitolinen und der Anwesenheit von *Trocholina friburgensis* gehört das Plattform-Äquivalent der Schichtfolge, von dem das Material erodiert und an den Beckenrand geliefert worden ist, zur Orbitolinopsis cuvillieri-kiliani-Subzone der Palorbitolina lenticularis-Zone des Oberen Barrême/Unteren Apt.

4. Stratigraphische Position des Schrattenkalks

Der Schrattenkalk kann als eine "regressive" Faziesentwicklung angesehen werden, die zwischen Drusberg-Schichten im Liegenden und Garschella-Formation im Hangenden eingeschaltet ist. Hervorzuheben ist, daß die Luitere-Schichten als untere Fazies-Entwicklung der Garschella-Formation regional heterochron einsetzen und damit unterschiedlichen stratigraphischen Umfang haben (FÖLLMI 1986, FÖLLMI & OUWEHAND 1987). Das Alter der Luitere-Schichten in Vorarlberg entspricht mit Funden von *Deshayesites* sp. und *Dufrenoyia furcata* (SOWERBY) mittlerem und spätem Unterapt (FÖLLMI 1986).

Ferner ist mit einer starken regionalen Gliederung des Osthelvetischen Schelfs gegen Ende des Unteren Barrême zu rechnen (WYSSLING 1986).



Text-Fig. 7. Die stratigraphische Verbreitung der Orbitolinen-Arten (MOULLADE, PEYBERNES, REY & SAINT-MARC 1985).

Aufgrund reicher Ammonitenfaunen entsprechen die Altmann-Schichten im Liegenden der Pulchella-Zone des Unter-Barrêmes (WYSSLING 1986). Das Alter der unteren Drusberg-Schichten entspricht bei Ranzenberg, zwischen Hohenems und Unterklien aufgrund der reichen Foraminiferenfauna (vor allem *Epistomina hechti* BETTENSTÄDT, BARTENSTEIN & BOLLI) Mittlerem Barrême (FUCHS 1971). Die pyritisierten Ammoniten-Nuclei desselben Fundortes bestätigen dieses Alter (WIEDMANN 1978).

Ein Grundproblem der stratigraphischen Gliederung der Plattformkarbonate ergibt sich aus dem Mangel an Ammonitenfunden. Die im Laufe der letzten 20 Jahre erarbeitete Orbitolinen-Stratigraphie ermöglicht eine Gliederung der urgonen Plattformkarbonate, ebenso wie eine Korrelierung der Beckenrand- und der Plattformsedimente.

Hier wurde die Gliederung von MOULLADE et al. (1985) übernommen (Text-Fig. 7), kombiniert mit den Ergebnissen, die an der Barrême-Bedoule-Schichtfolge der Urgon-Plattform (Süd-Vercors) und des Vocontischen Beckens in SE-Frankreich gewonnen wurden (ARNAUD-VANNEAU 1980, ARNAUD 1981).

4.1 Stratigraphie der Vorarlberger Profile

Die biostratigraphische Gliederung entspricht den Orbitolinen-Zonen des Urgons, wie sie aus der Umgebung von Genf von SCHROEDER et al. (1968), SCHROEDER & CONRAD (1968) und von CONRAD (1969) erarbeitet wurden. Von letzterem wurden fünf Orbitolinen-Zonen unterschieden, von denen die Zone I nach Cephalopoden älterem Barrême entspricht. Das Indexfossil der Zone III, *Palaeodictyoconus reicheli* GUILLAUME, setzt an seiner Typlokalität bei Fribourg im Oberen Barrême ein. Die Grenze Barrême/Apt verläuft in der Zone IV. Die Zone V gehört aufgrund des Auftretens von *Deshayesites* ex gr. *weissi* (NEUMAYR & UHLIG) schon zum älteren Unterapt.

Als Ergebnis der vorliegenden Untersuchung der Orbitolinen und kennzeichnender Begleit-Assoziationen kann festgehalten werden, daß in den Profilen von Unterklien die Zone III wahrscheinlich und die Zone IV und der Unterteil der Zone V von CONRAD (1969) sicher nachgewiesen werden können (Text-Fig. 4 und 5).

Es ist hervorzuheben, daß in den Profilen von Unterklien in den Zonen IV und V nur Vertreter der *Orbitolinopsis cuvillieri-kiliani*-Gruppe, nicht aber *Orbitolinopsis kiliani* selbst auftreten. Schließlich erscheinen am Top der Schichtfolge neue *Orbitolinopsis*-Arten mit *Orbitolinopsis* sp. 1 und *Orbitolinopsis* sp. 2. *Valserina*-Arten kommen in unseren Proben nicht vor.

H. SCHOLZ (1979, 1984), der die Bioherme und Biostrome der Schrattenkalk-Formation des Allgäus bearbeitet hat, bestimmt das Alter dieser Formation aufgrund des Auftretens von *Palorbitolina lenticularis, Salpingoporella muehlbergi, S. hasi* und spezifischen Molluskenarten als wahrscheinliches Unter-Bedoule. Im Laufe unserer Geländearbeit im Jahre 1987 hat H. LOBITZER ähnliche Bioherme in Straßenaufschlüssen bei Übersaxen erkannt, die ähnlich wie die Aufschlüsse im Allgäu den segmentierten Kalkschwamm *Barroisia helvetica* enthalten.

Beim Vergleich der Profile des Süd-Vercors (Zone de Borne, Le Devoluy; ARNAUD-VANNEAU in ARNAUD 1981, ARNAUD-VANNEAU et al. 1982) und der Unterklien-Profile ergibt sich, daß diese mit hoher Wahrscheinlichkeit der oberen Zone "Bi6", den Zonen "Bs1", "Bs2", "Bs3", "BsAi" und teilweise der Zone "Ail" entsprechen dürften. Diese Zonen entsprechen der Palorbitolina lenticularis-Zone (Subzone von Orbitolinopsis cuvillieri-kiliani). Danach ist ein Oberbarrême-Unterapt-Alter für die beiden Profile von Unterklien wahrscheinlich.

Das Plattform-Äquivalent des Profils Hintere Ill-Schlucht/Feldkirch gehört ebenfalls zur Palorbitolina lenticularis-Zone.

4.2 Kalkalgen

Die Kalkalgen der Umgebung von Genf wurden von CONRAD (1969, 1970) bearbeitet, die Dasycladaceae der Urgon-Kalksteine des Hauterive und Barrême wurden von PEYBERNES & CONRAD (1970), die der Urgon-Plattformen von SE-Frankreich von MASSE (1976) studiert.

Die zeitliche Verbreitung der Kalkalgen-Assoziationen des Schrattenkalks des Steinbruchs Rhomberg ist in Text-Fig. 2 zu sehen. Die Mehrheit der Arten sind Durchläufer bzw. besitzen lange stratigraphische Reichweiten. Hinsichtlich der stratigraphischen Gliederung kommen die *Salpingoporella*-Arten in Betracht (*S. muehlbergi, S. hasi, S. melitae*), ferner *Cylindroporella benisarensis*. Das gemeinsame Auftreten dieser Art und die begleitende Kalkalgen-Assoziation ergibt - ähnlich den Foraminiferen - ein Oberbarrême-Unterapt (Bedoule)-Alter der Profile.

5. Zusammenfassung

Im Steinbruch Rhomberg (Vorarlberg) wird der Schrattenkalk dem Oberbarrême/Unterapt zugeordnet; das obere Unterbarrême hingegen kann bislang nur vermutet werden. Diese stratigraphische Zuordnung erscheint durch das Vorkommen von *Palorbitolina lenticularis praecursor* (MONTANARI) und *P. lenticularis lenticularis* (BLUMENBACH), *Orbitolinopsis cuvillieri* MOULLADE sowie durch diverse Übergangsformen der *Orbitolinopsis cuvillieri-kiliani*-Gruppe abgesichert. Diese stratigraphische Einstufung wird auch durch eine vergleichsweise artendiverse Dasycladaceenflora (*Salpingoporella muehlbergi, S. melitae, S. hasi, Cylindroporella benisarensis*) unterstützt.

Im höchsten *Orbitolinopsis*-Horizont konnten zwei neue *Orbitolinopsis*-Arten festgestellt werden, die einer späteren Beschreibung vorbehalten sind und vorläufig mit *Orbitolinopsis* sp. 1 und *Orbitolinopsis* sp. 2 bezeichnet werden.

Dank. M. A. CONRAD (Genf) danke ich für die Überprüfung der Kalkalgen-Bestimmungen. P. RAT (Dijon), F. ALLEMANN (Bern) sowie R. OBERHAUSER (Wien) bin ich für Diskussionen und Ratschläge während der Abfassung dieses Manuskriptes dankbar. A. ARNAUD-VANNEAU danke ich für die Überprüfung der Foraminiferen-Bestimmungen. Herrn H. LOBITZER (Wien) möchte ich Dank sagen für die Führung im Gelände sowie für die Durchsicht dieses Manuskriptes.

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Tafeln 1-4

Rh-1	Steinbruch Rhomberg, Profil 1
Rh-2	Steinbruch Rhomberg, Profil 2
F/HI	Feldkirch/Hintere Ill-Schlucht
D =	Dünnschliff-Nr.

Tafel 1

Fig. 1. *Palorbitolina* sp. 1. Rh-1, D = 27, 53 x.
Fig. 2. *Palorbitolina lenticularis praecursor* (MONTANARI). Rh-2, D = 45, 34 x.
Fig. 3. *Orbitolinopsis* sp. 1. Rh-2, D = 9, 40 x.
Fig. 4. *Orbitolinopsis* sp. 1. Rh-2, D = 10, 34 x.
Fig. 5. *Orbitolinopsis cuvillieri* MOULLADE. Rh-2, D = 11, 34 x.
Fig. 6. *Orbitolinopsis* sp. 2. Rh-2, D = 12, 40 x.
Fig. 7. *Paracoskinolina sunnilandensis* (MAYNC). Rh-2, D = 27, 40 x.
Fig. 8. *Sabaudia minuta* (HOFKER). Rh-2, D = 62, 136 x.
Fig. 9. *Orbitolinopsis* cf. *cuvillieri* MOULLADE. Rh-1, D = 10, 53 x.
Fig. 10. *Paracoskinolina* cf. *maynci* (CHEVALLIER). F/HI, D = 14, 53 x.
Fig. 11. *Orbitolinopsis* sp. Rh-2, D = 9, 53 x.
Fig. 12. *Paracoskinolina maynci* (CHEVALLIER). Rh-1, D = 13, 53 x.

Tafel 2

Fig. 1. *Nummoloculina heimi* BONET. Rh-2, D = 12, 40 x.
Fig. 2. *Quinqueloculina* cf. *danubiana* NEAGU. Rh-2, D = 12, 136 x.
Fig. 3. *Spiroloculina cretacea* REUSS. Rh-2, D = 62, 136 x.
Fig. 4. *Trocholina friburgensis* (GUILLAUME & REICHEL). Rh-2, D = 45, 40 x.
Fig. 5. *Trocholina* aff. *friburgensis* (GUILLAUME & REICHEL). Rh-2, D = 34, 40 x.
Fig. 6. *Trocholina* sp. Rh-2, D = 39, 40 x.
Fig. 7. *Pseudochoffatella* sp. Rh-2, D = 54, 40 x.

Tafel 2, Fortsetzung

Fig. 8. *Quinqueloculina robusta* NEAGU, mikritisiert. Rh-2, D = 8, 40 x.
Fig. 9. Miliolina/*Derventina filipescui* NEAGU. Rh-2, D = 9, 34 x.
Fig. 10. *Massilina* sp. Rh-2, D = 45, 40 x.
Fig. 11. *Placopsilina cenomana* (D'ORBIGNY). Rh-2, D = 58, 40 x.

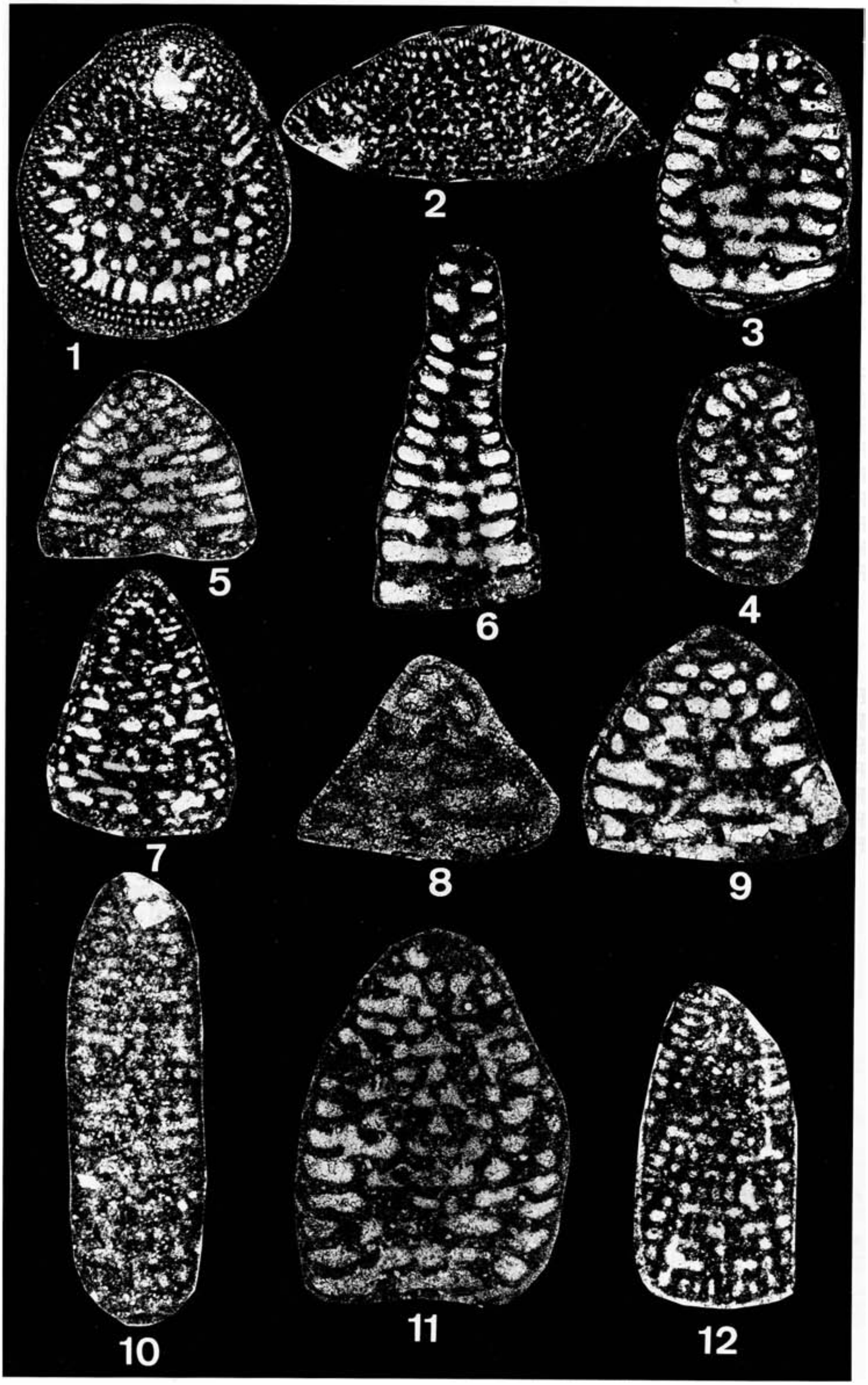
Tafel 3

Fig. 1. *Choffatella decipiens* SCHLUMBERGER.. Rh-2, D = 34, 40 x.
Fig. 2. *Choffatella decipiens* SCHLUMBERGER. Rh-2, D = 13, 32 x.
Fig. 3. *Aeliosaccus* sp. (*Erlandia ?conradi* ARNAUD-VANNEAU). Rh-2, D = 9, 40 x.
Fig. 4. *Charenthia cuvillieri* NEUMANN. Rh-2, D = 12, 40 x.
Fig. 5. *Glomospirella* sp. Rh-2, D = 62, 136 x.
Fig. 6. *Nautiloculina* sp. Rh-2, D = 58, 136 x.
Fig. 7. *Marssonella praeoxycona* MOULLADE. Rh-2, D = 62, 136 x.
Fig. 8. *Pseudocyclammina hedbergi* MAYNC. Rh-2, D = 13, 26 x.
Fig. 9. *Nautiloculina bronnimanni* ARNAUD-VANNEAU. Rh-2, D = 11, 32 x.
Fig. 10. *Valvulammina* sp. Rh-2, D = 58, 136 x.
Fig. 11. *Glomospirella urgoniana* ARNAUD-VANNEAU. Rh-2, D = 45, 40 x.
Fig. 12. *Valvulammina* sp. Rh-2, D = 45, 136 x.

Tafel 4

Fig. 1. *Salpingoporella muehlbergi* (LORENZ). Rh-2, D = 12, 40 x.
Fig. 2. *Salpingoporella* cf. *muehlbergi* (LORENZ). Rh-2, D = 9, 136 x.
Fig. 3. *Salpingoporella melitae* RADOIČIĆ. Rh-1, D = 2, 54 x.
Fig. 4. Bryozoa. Rh-2, D = 57, 40 x.
Fig. 5. Ostracoda. Rh-2, D = 60, 136 x.
Fig. 6. *Heteroporella* sp. cf. *paucicalcarea* CONRAD. Rh-2, D = 12, 40 x.
Fig. 7. *Palorbitolina lenticularis* (BLUMENBACH). F/HI, D = 34, 136 x.

Tafel 1



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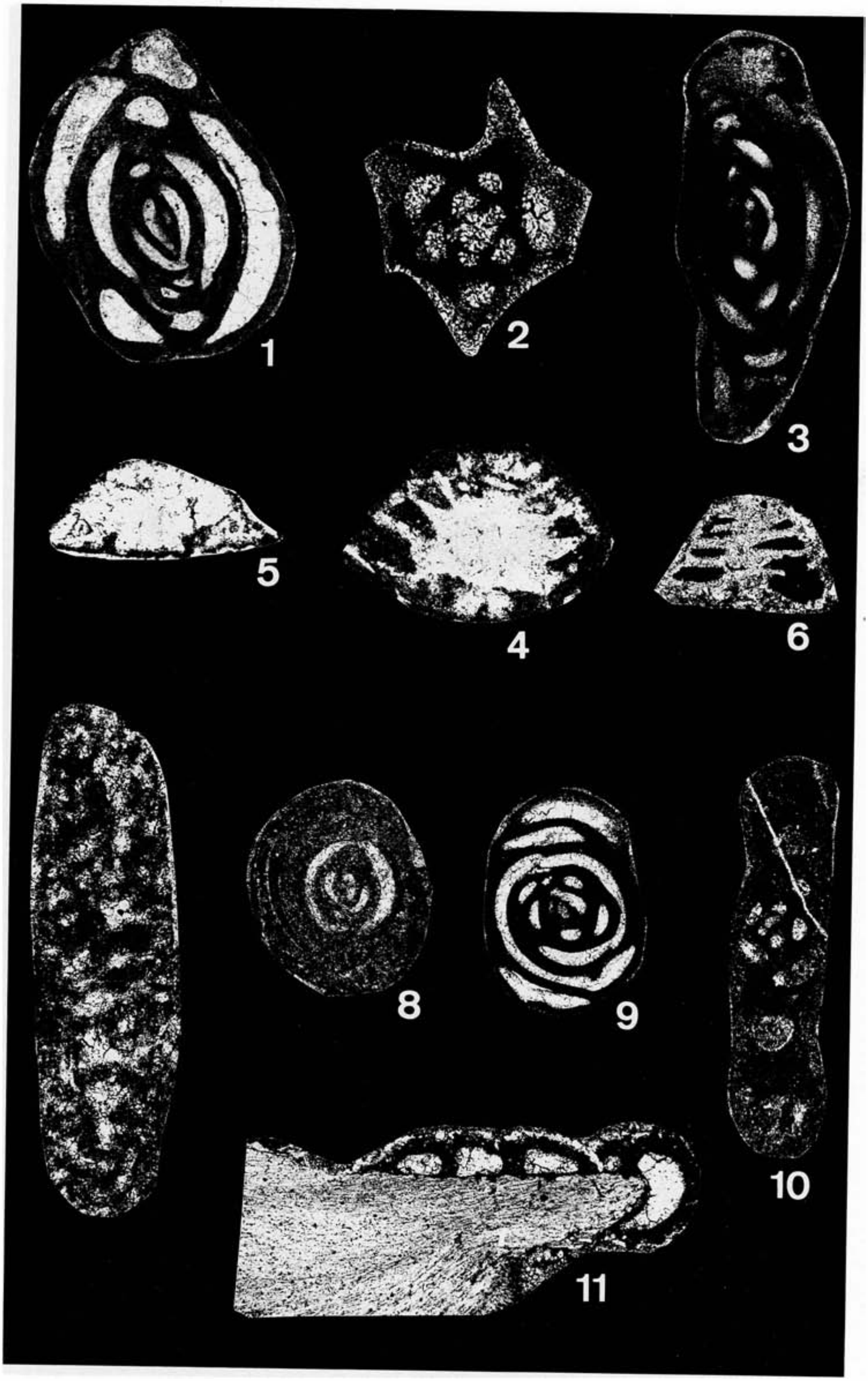

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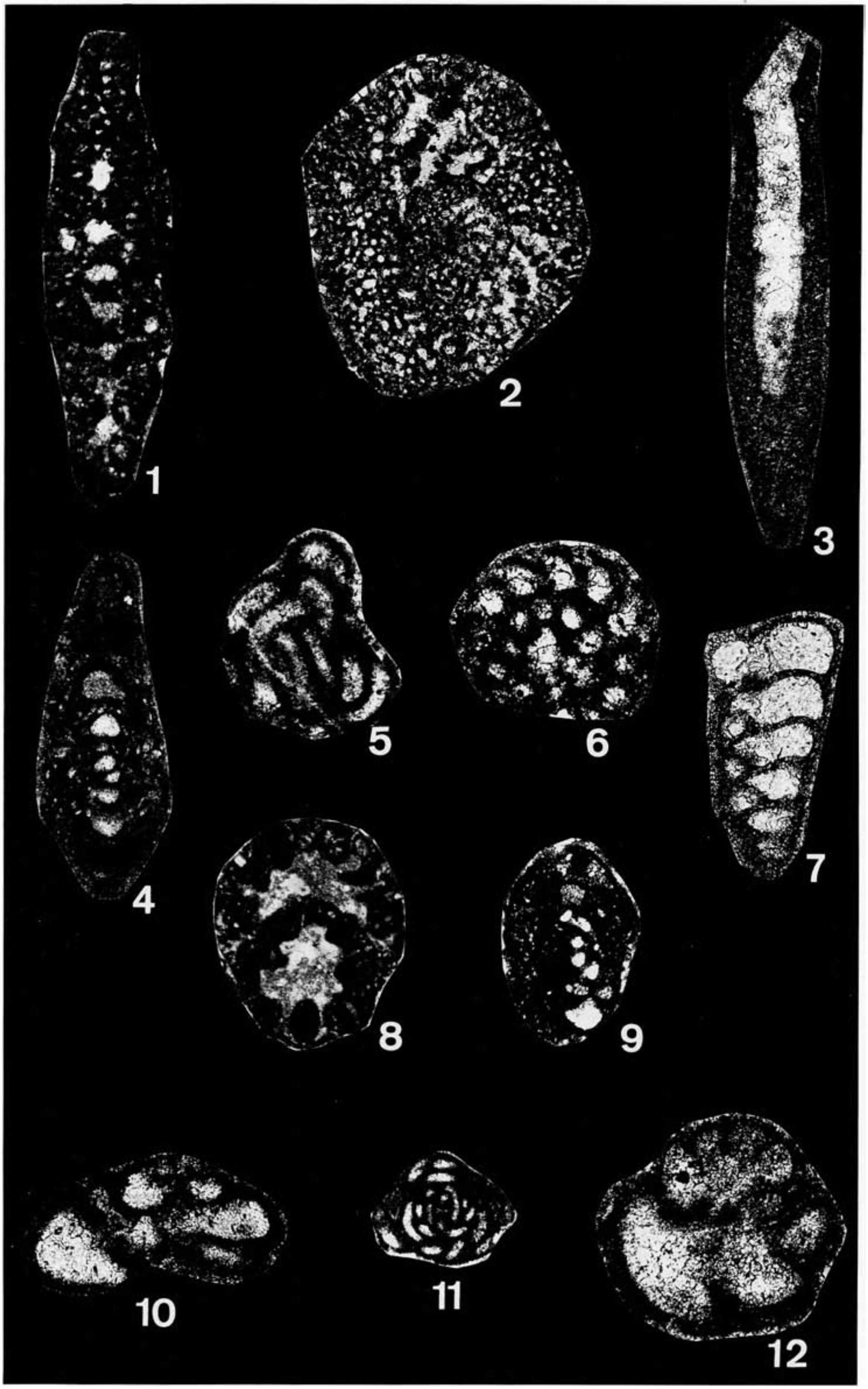
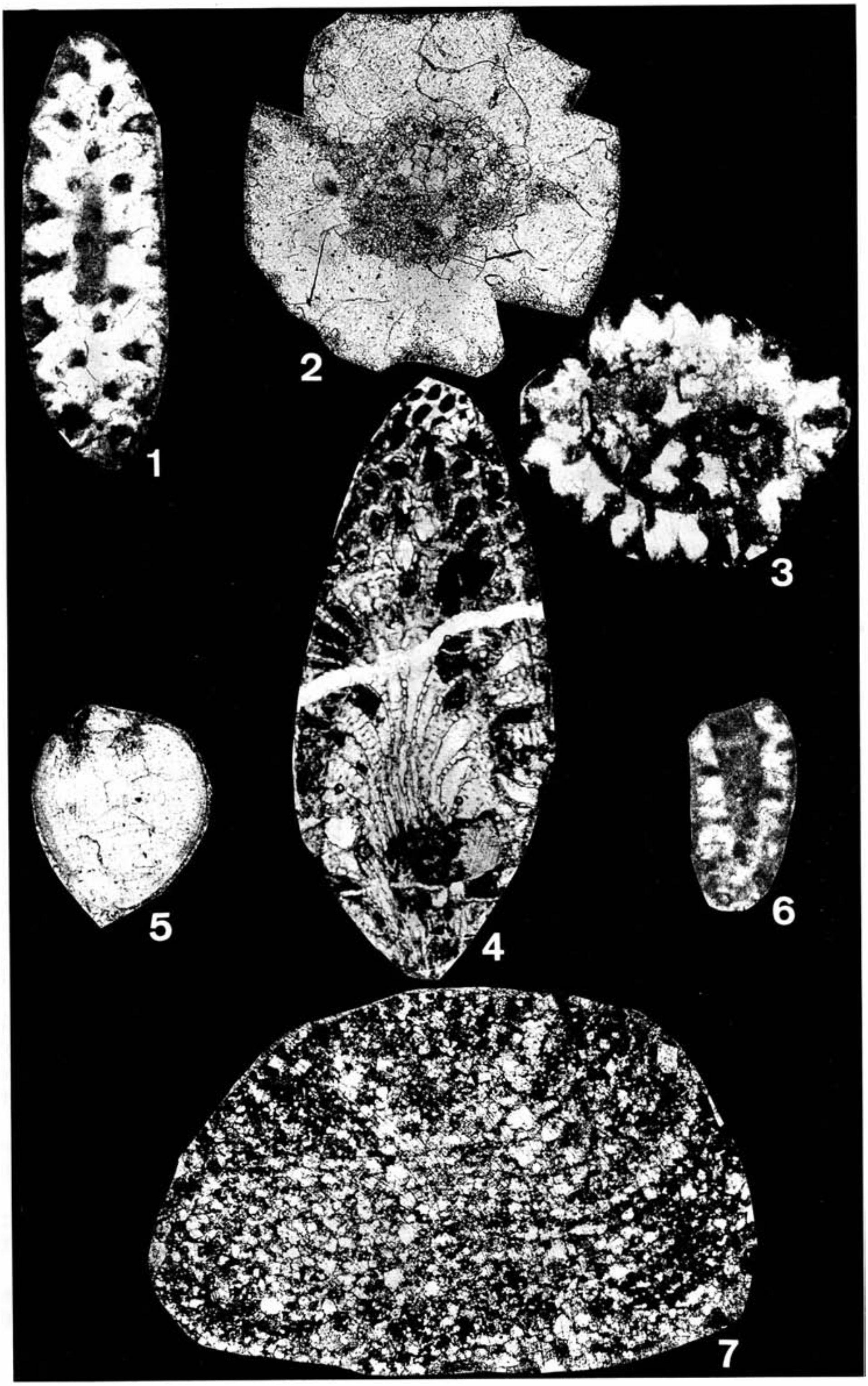


Tafel 4



Korrelation der Pachyodonten-Faunenzonen des Urgons der westlichen Tethys

LENKE CZABALAY, Budapest

Mit 11 Text-Figuren

CZABALAY, L. (1989): Korrelation der Pachyodonten-Faunenzonen des Urgons der westlichen Tethys. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 431-451. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Bivalve and gastropod faunas from Urgonian facies were studied at various localities and through time, i. e. in Vorarlberg (Schrattenkalk; Upper Barremian-Lower Aptian), at the foothills of the Vêrtes Mountains (Környe Limestone Formation; Lower-Middle Albian), and in the northern and southern Bakony Mountains (Zirc Limestone Formation; Middle-Upper Albian).

On the basis of the Pachyodonta-*Ostrea*-*Chondrodonta*-*Nerinea* fauna, faunal zones were identified within the Urgonian sedimentary sequence. These were primarily correlated with the fauna of the West-European Subprovince. Faunas from other zoogeographic subprovinces were also included in the analysis, in order to find out resemblances, deviations or age identities.

1. The Upper Barremian-Lower Aptian (Schrattenkalk, Mittagspitz) Toucasia lonsdalei-Aetostreon couloni faunal zone can be correlated with faunas from deposits of similar age and facies in the West European zoogeographic subprovince in Eastern Switzerland (Helvetic Zone), Germany (Allgäu), SE France (Orgon, Vercors, Marseille, Toulon) and Northern Spain (Pyrenees, Basco-Cantabrian Ranges). In spite of the different Pachyodonta fauna, it can be compared with coeval faunas occurring in the Carpathians (Manin Zone, High Tatras) and the Balkan zoogeographic subprovince (Bulgaria, Romania, Serbia).

2. The fauna of the Lower-Middle Albian (Környe Limestone Formation) Eoradiolites murgensis-Pseudotoucasia santanderensis-Chondrodonta munsoni faunal zone belongs to the West European Subprovince. Faunistically it can be correlated with the faunas of SE France, Northern Spain (Pyrenees, Basco-Cantabrian Ranges) and Portugal (Estremadura, Algarve).

3. The Middle-Upper Albian (Zirc Limestone Formation) in the N Bakony represents the Toucasia carinata-Pseudotoucasia santanderensis faunal zone (Olaszfalu Member).

In the S Bakony, the Eoradiolites davidsoni-E. hungaricus faunal zone is characteristic (Padragkut). In the Urkut area occurs a very rich *Chondrodonta hantkeni* fauna being a back-reef variety of the fore-reef faunal zone.

The Pachyodonta faunal zone of the northern Bakony resembles very much the *Toucasia*-*Eoradiolites* faunas of S France, N Spain, Portugal, and Italy. In addition, it is very much related to the West European zoogeographic subprovince (SE France, N Spain) and to the southern faunas of the Apulian/Preapulian zoogeographic province.

Kurzfassung: Im vorliegenden Beitrag werden die Muschel- und Schneckenfaunen des Urgons vergleichend untersucht. Im Detail handelt es sich um den Zeitraum Oberbarreme/Unterapt von Vorarlberg, Österreich, Unter/Mittelalb des Vorlandes des Vértesgebirges, Ungarn (Környe-Kalk-Formation) und Mittel/Oberalb des nördlichen und südlichen Bakony, Ungarn (Zirc-Kalk-Formation). Die Arbeit stellt einen Versuch dar, die Pachyodonten-Ostreiden-Chondrodonten-Nerineen-Assoziationen und Faunenzonen des Urgons zu definieren und mit der europäischen und anderen Faunenprovinzen zu korrelieren.

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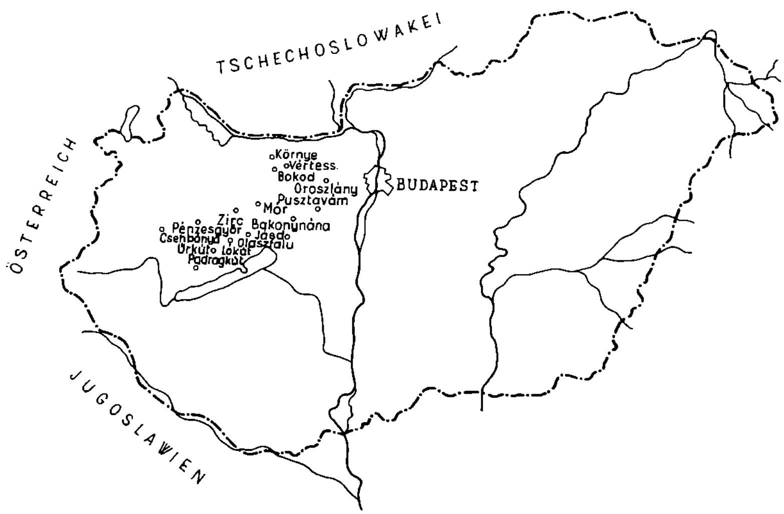
1. Einleitung

Es wird die Mollusken-Fauna (Muscheln, Gastropoden) aus folgenden Bohrungen und Übertageprofilen untersucht (Text-Fig. 1): Schrattenkalk - Profile 1-2 des Steinbruchs Rhomberg, Vorarlberg; Környe-Kalk-Formation - Vorland des Vértesgebirges, Bohrungen Bokod 1828, Oroszlány 1822, 1825, 1891, Környe 26, 27, Mór 7, 15; Zirc-Kalk-Formation - im nördlichen und südlichen Bakony, Bohrungen Urkut 254, 421, Pénzesgyör 5, Jásd 42, Padragkut 7 (Referenz-Bohrung) und Übertageprofile Jásd 1-2, Olaszfalu Eperkéshegy-1.



Text-Fig. 1. Mittelkreide-Lokalitäten der Transdanubischen Mittelgebirge.

In Vorarlberg wurden die Profile des Schrattenkalkes von G. CSASZAR (1985, 1986) bearbeitet. In Ungarn wurden die geologischen Verhältnisse der Környe-Kalk- und Zirc-Kalk-Formationen von G. CSASZAR (1986) zusammengefaßt.

Die Arbeiten wurden teilweise im Rahmen des IGCP-Projekts 262 durchgeführt und erste Ergebnisse über die Zirc-Kalk-Formation beim Kreide-Symposium Budapest 1982 vorgelegt (CZABALAY 1984).

2. Charakterisierung der Mollusken-Assoziationen und Faunenzonen und ihre Verbreitung

2.1 Der Schrattenkalk

In Vorarlberg wurden außer den Molluskenfaunen der beiden Profile des Steinbruchs Rhomberg und den Proben aus dem Profil von Unterklien auch jene aus den älteren Sammlungen von S. FUSSENEGGER und A. HEIM revidiert.

Aufgrund der Molluskenfauna in den beiden Profilen des Steinbruchs Rhomberg wurden mehrere Faunen-Assoziationen bestimmt und in die von CSÁSZÁR (1985, 1986) aufgenommenen Profile eingetragen.

Das beiliegende Profil ist in überkippter Schichtenfolge aufgenommen und hier in normaler Lagerung dargestellt (Text-Fig. 2).

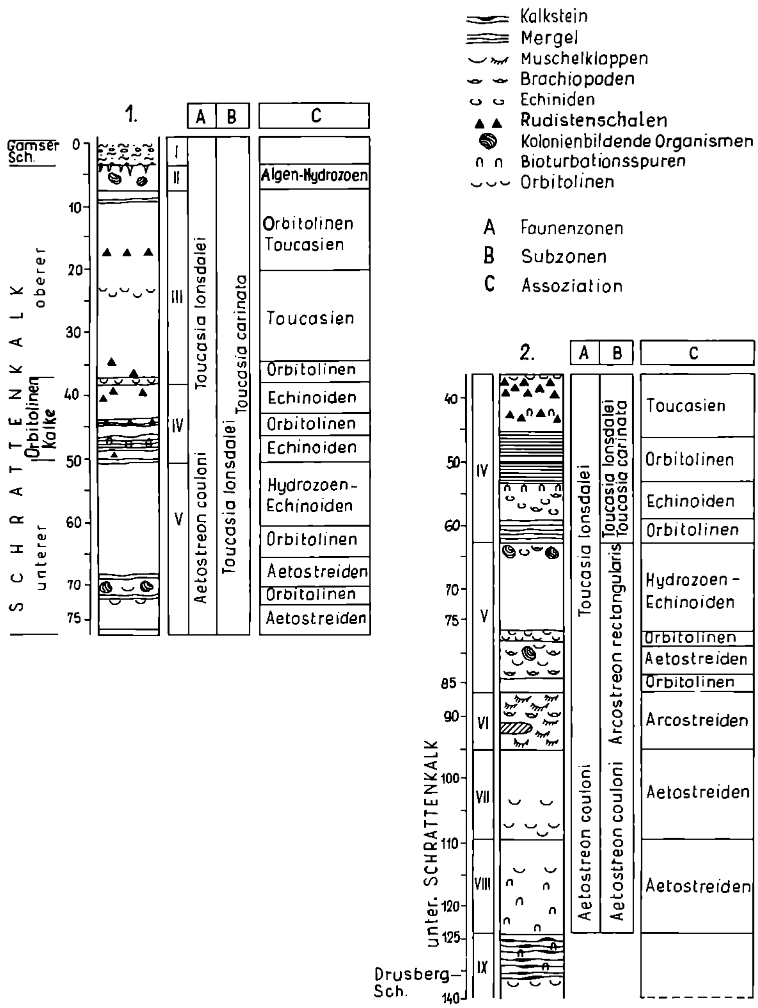
Für den Schrattenkalk ist ein Wechsel innerhalb der Hydrozoen-Korallen-Stromatoporen-Orbitolinen-Pachyodonten-Ostreen-Nerineen-Echinoiden-Faunen-Assoziationen kennzeichnend. In den festgestellten sieben Faunen-Assoziationen dominieren zum Teil die Mollusken. Im Sediment sind die Mergel-Zwischenlagen recht häufig, in denen die Echinoiden-Brachiopoden- und Orbitolinen-Faunen in den Vordergrund treten. In den Kalken sind die Pachyodonten-Stromatoporen-Hydrozoen-Korallen-Faunen häufig.

In der Sedimentfolge des Schrattenkalks alternieren Biostrome mit Biohermen. Die Intensität der Wasserbewegung wird periodisch stärker; dies stand wahrscheinlich mit der häufigen Bewegung des Meeresuntergrundes im Zusammenhang. In der Begleitfauna ist eine Algenanreicherung mit einer charakteristischen Benthos-Mikrofauna häufig.

Für die Molluskenfauna sind im oberen Teil großwüchsige Toucasien-Requienien-Arten kennzeichnend, stellenweise die Vergesellschaftung von Agriopleuren-Arten mit großwüchsigen Schnecken (*Nerinea, Leviathania, Harpagodes*). Die Muschelarten sind fast ausnahmslos vom endobiotischen, zementierten Typ, während die großwüchsigen Schnecken in den littoralen Zonen des Schelfs meist unter den Pflanzen lebten.

Im unteren Abschnitt des Schrattenkalks bilden die Ostreen-Arten Faunen-Assoziationen, stellenweise bauen sie ganze Bänke auf (z. B. *Arcostreon, Aetostreon*). *Arcostreon rectangularis* (ROEM.) als endobiotischer Typ hat sich periodisch auf dem Untergrund oder auf der Schale eines anderen Individuums mit Hilfe von Byssus-Fäden befestigt. Sie lebten in den littoralen Zonen auf sandigem schlammigem Boden. Sie werden durch die Art *Aetostreon couloni* (D'ORB.) abgelöst, eine ebenso endobiotische Art, die aber ihre riesige Schale permanent befestigt und zementiert hat.

Hier ist bereits eine Verstärkung der Wasserbewegung bemerkbar. Der Salzgehalt des Wassers war normal, die Temperatur war nicht hoch, sie lag um 18-20 °C. Die Wassertiefe dürfte um 10-15 m betragen haben. Es kann eher von gemäßigten Klimaverhältnissen als von subtropischen gespro-



Text-Fig. 2. Profile 1 und 2 des Steinbruchs Rhomberg, Vorarlberg.

chen werden. Dies wird auch von den Faunenelementen angezeigt, die großenteils im ganzen Tethys-Gebiet vorkommen, zu einem kleineren Teil aber gerade für dieses Gebiet kennzeichnend sind.

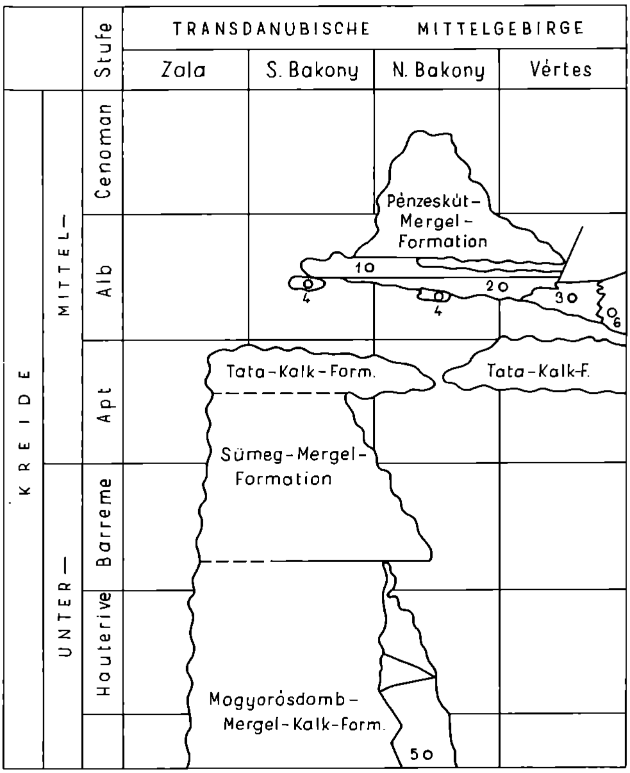
Die Sedimente des Schrattenkalks gehören einer einzigen Faunenzone (Text-Fig. 2) an, der Aetostreon couloni-Toucasia lonsdalei-Faunenzone. Innerhalb dieser Zone können zwei Subzonen unterschieden werden, eine untere Aetostreon couloni-Arcostreon rectangularis-Subzone und eine obere Toucasia lonsdalei-Toucasia carinata-Subzone.

2.2 Környe-Kalk-Formation

Im Vorland des Vêrtesgebirges zieht sich die Tés-Mergel-Formation in NE-SW-Richtung; in NE kann sie über die Zwischenschaltung des Riffs der Környe-Kalk-Formation bis zur Ortschaft Oroszlány verfolgt werden; sie geht in die Vêrtessomló-Aleurit-Formation über.

Im Vêrtes-Vorland sind die Sedimente der Környe-Kalk-Formation und Tés-Mergel-Formation teilweise miteinander verzahnt (Text-Fig. 3). Die Faunenänderungen sind in S-SW-Richtung, also senkrecht zur ehemaligen Lagune, gut verfolgbar.

Die Sedimente der Környe-Kalk-Formation enthalten eine reiche Fauna von Pachyodonten (*Toucasia, Pseudotoucasia, Agriopleura, Eoradiolites*), Gastropoden (*Nerinella, Plesioptyxis, Nerinea*), von anderen Muscheln (*Chondrodonta, Liostrea*), von Algen, Orbitolinen und Korallen. Die Flanke des



1 o Zirc-Kalk-Formation
2 o Tés-Tonmergel-Formation
3 o Környe-Kalk-Formation
4 o Alsópere-Bauxit-Formation
5 o Borzavár-Kalk-Formation
6 o Vértessomló-Aleurit-Formation

Text-Fig. 3. Stratigraphie der Mittelkreide-Formation der Vêrtes- und Bakony-Gebirge.

Riffs ist in Richtung gegen die Ortschaft Vértessomló von Vorriff-Charakter; hier sind die Algen und Korallen innerhalb der Pachyodonten-Nerineen-Chondrodonten-Assoziation häufig. Weiter nach W-SW nimmt die Zahl der Chondrodonten-Korallen-Algenarten auf dem Riffkörper ab, es überwiegen immer stärker die Nerinellen und Nerineen, zusammen mit den Pachyodonten (*Agriopleura, Toucasia*). Im Raum von Bokod (Bohrung Bokod 1828), wo das Riff ausdünnt, finden wir Gemeinschaften mit Toucasien, Nerineen-Toucasien und Agriopleuren. In den Sedimenten der Környe-Kalk-Formation und in den Faunen-Assoziationen mit Pachyodonten erscheinen bereits die ersten Liostreen. Die Faunen-Assoziationen spiegeln eine Mischung der Faunen-Assoziationen der Környe-Kalk- und Tés-Mergel-Formationen wider. Zwischen die Faunen-Assoziationen der Back-Reef-Entwicklung bei Pusztavám (Bohrung Pusztavám 980) mit Toucasien-Agriopleuren schalten sich die Faunen-Assoziationen mit Liostreen ein.

Text-Fig. 8 gibt einen Überblick über die Verbreitung der Eoradiolites murgensis-Toucasia carinata-Chondrodonta munsoni-Faunenzone und andere Faunen-Assoziationen im Vorland des Vértes.

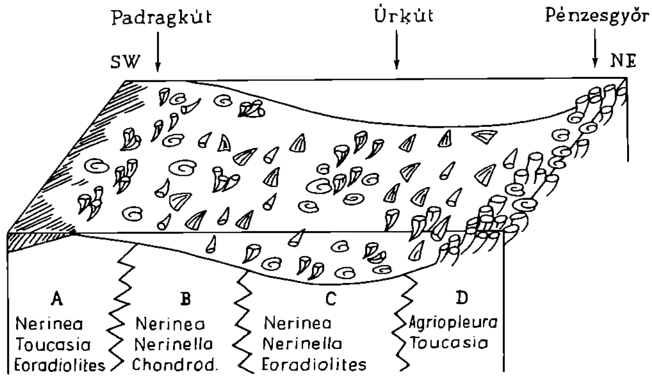
Es kann gefolgert werden, daß die Fauna der Környe-Kalk-Formation in warmem Wasser (über 20-25 °C), in einer subtropischen Umgebung, unter normalen Salinitätsverhältnissen, im Fore-Reef- und Back-Reef-Bereich lebte. Der Riffkörper war nicht sehr breit, so daß der Sedimentationsraum starken marinen Einwirkungen ausgesetzt war.

2.3 Zirc-Kalk-Formation

Die Zirc-Kalk-Formation ist im Süd- und Nord-Bakony unterschiedlich ausgebildet (Text-Fig. 9). Die Sedimente der Zirc-Kalk-Formation im Nord-Bakony sind in drei Schichtglieder zu untergliedern, und zwar das Pachyodonten-Member (Eperkéshegy-Member), das Orbitolinen- und das Plattenkalk-Member. Eperkéshegy-Member und Teile des Orbitolinen-Member und ihre Faunen entwickelten sich im Fore-Reef. Im Plattenkalk-Member kommen Faunenelemente vor, die im schlammigen Meeresboden der littoralen Zone lebten. Die geographische Verbreitung dieser Arten (*Avellana, Pecten*) überschreitet die Grenzen des mediterranen Geosynklinal-Gebiets; sie kommen auch in den Faunenprovinzen der gemäßigten Zonen Westeuropas vor.

Für den Süd-Bakony sind Nerinellen-Nerineen-Toucasien-Eoradioliten-Chondrodonten-Liostreen- und Algen-Assoziationen charakteristisch. In der Umgebung von Urkut wurden fünf Assoziationen festgestellt, in denen die Chondrodonten überwiegen. In Padragkút wurden sechs Mollusken-Assoziationen definiert; hier sind schnellere ökologische Veränderungen feststellbar, einige Faunen-Assoziationen wiederholen sich mehrmals.

Die Faunen-Assoziationen der Zirc-Kalk-Formation im Süd-Bakony werden in die Eoradiolites davidsoni-Eoradiolites hungaricus-Faunenzone gestellt; in der Umgebung von Úrkút ist sie durch Chondrodonten charakterisiert und in der Umgebung von Padragkút durch *Pseudotoucasia santanderensis* (Text-Fig. 9). Es war möglich, über gleiche Pachyodonten-Arten (*Pseudotoucasia santanderensis* DOUV. und *Toucasia carinata* MATH., *Eoradiolites murgensis* TORRE) die Pseudotoucasia santanderensis-Toucasia carinata-Faunenzone aus dem Nord-Bakony mit dem unteren Abschnitt der Faunenzone des Süd-Bakony zu korrelieren (Text-Fig. 4).



Text-Fig. 4. Verbreitung der Mollusken-Assoziationen im Süd-Bakony.

3. Biostratigraphische Auswertung und Korrelation der Faunenzonen des Urgons

Im allgemeinen wird die stratigraphische Datierung der Urgon-Bildungen mit Ammoniten, Algen und der Mikrofauna (Orbitolinen) durchgeführt. Hier gilt besondere Aufmerksamkeit den Pachyodonten als kennzeichnendem Faunenelement des Urgon, ebenso wie auch den begleitenden anderen Muschel- (Liostreen, Chondrodonten) und Schneckenarten (Nerineen).

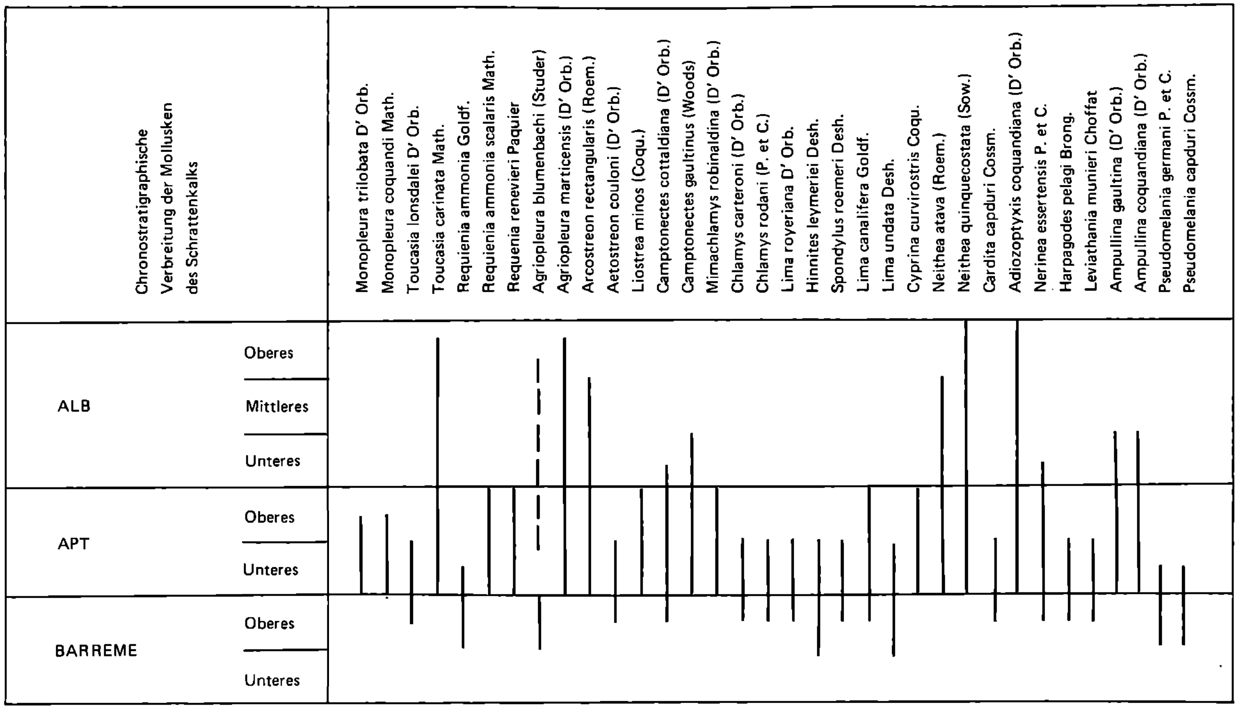
3.1 Schrattenkalk

Die Verbreitung einzelner Pachyodonten-Arten des Schrattenkalks ist auf den Zeitraum Oberes Barreme/Unteres Apt beschränkt, wie z. B. im Fall von *Requienia ammonia* GOLDF., *Agriopleura blumenbachi* (STUDER), *Toucasia lonsdalei* (D'ORB.), *Aetostreon couloni* (D'ORB.) und den Schneckenarten *Harpagodes pelagi* BRONG., *Leviathania munieri* CHOFFAT (Text-Fig. 5). Die Arten *Agriopleura marticensis* (D'ORB.), *Toucasia carinata* MATH. erscheinen im Unteren Apt, doch kommen sie auch im Oberen Apt vor.

Die Fauna des Schrattenkalks wird der westeuropäischen Subprovinz zugerechnet. Faunenzonen und -subzonen lassen sich unverändert in die Ostschweiz (Helvetische Decke) verfolgen (FUNK & BRIEGEL 1979). Im unteren Abschnitt des Schrattenkalks findet man hier ebenso die Ostreen-Subzone (*Arcostreon rectangularis*-*Aetostreon couloni*). Im oberen Drittel des unteren Abschnittes erscheinen die Requienien-Toucasien-Arten. Die Orbitolinen-führenden Schichten des oberen Schrattenkalks sind das Ergebnis einer kleinen warmen Fluktuation.

Auch in NE-Richtung, ins Allgäu, setzt sich der Schrattenkalk mit sehr ähnlichen Faunen-Assoziationen wie in Vorarlberg (Rhomberg) fort.

Auch in SW-Richtung, in der Umgebung von Genf, findet man gleichaltrige Pachyodonten-Nerineen-Orbitolinen-Faunen-Assoziationen (CONRAD 1969). Weiter in SW-Richtung, in Frankreich, im Dauphiné-Becken, kann man ebenfalls ähnliche Faunen in den Urgon-Kalken des Oberen Barreme/Unteren Apt beobachten. Hieran schließt sich das Randgebiet der Cévénole an (Text-Fig. 10).



Text-Fig. 5. Chronostratigraphische Verbreitung der Mollusken des Schrattenkalks.

In der Provence (Toulon, Marseille) gibt es im Oberen Barreme charakteristische Pachyodonten und Nerineen führende Urgon-Kalke; diese werden von Orbitolinen-Mergeln des Unteren Apt (Bedoule) abgelöst. Bei La Fare/ Mt. Ventoux hat die Gesamtfolge nur Bedoule-Alter (ARNAUD-VANNEAU et al. 1982).

Eine weitere, zur westeuropäischen Subprovinz gehörende Urgon-Fauna ist von West-Sardinien bekannt (MASSE & ALLEMANN 1982) mit Requienien- und Agriopleuren-Arten, die denen des unteren Schrattenkalks entsprechen.

Auf dem Gebiet des Karstes von Frioul (Jugoslawien, Slovenien) enthalten die Urgon-Kalke eine reiche Requienien-Fauna (O.-Barreme/U.-Apt); dieser schließen sich für die Apulische Faunenprovinz kennzeichnende Caprotina- und Monopleura-Arten an (VELIĆ et al. 1979). Damit weicht diese Entwicklung mit ihrer Fauna vom Schrattenkalk stark ab, ist aber altersgleich.

Die im Gebiet der Karpaten (Slowakei, Rumänien, Sowjetunion), der Hohen Tatra (Polen) und der Dinariden (Serbien, Bulgarien) vorkommenden Urgon-Sedimente mit Pachyodonten-Nerineen-Algen-Faunen-Assoziationen gehören bereits zur Balkanischen Subprovinz.

3.2 Kōrnye-Kalk-Formation

Aus der biostratigraphischen Auswertung der Pachyodonten-Nerineen-Chondrodonten-Fauna der Környe-Kalk-Formation kann gefolgert werden, daß einige Arten dieser Fauna altersspezifisch sind. Zu einer chronostratigraphischen Verwendung sind aber eher die Faunen-Assoziationen (Faunenzonen und Subzonen) geeignet.

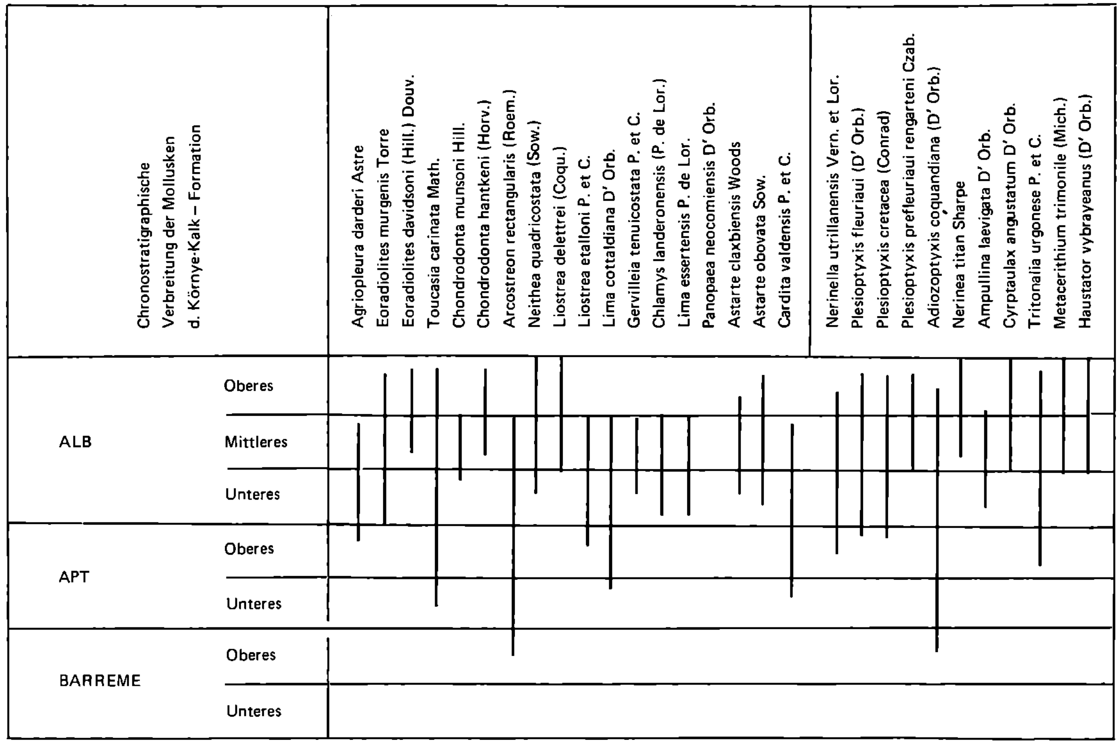
Die für die Pachyodonten sehr kennzeichnende Art *Agriopleura darderi* (ASTRE) ist in Südfrankreich (Provence) aus Oberem Apt und Alb bekannt (MASSE & PHILIP 1981); in den Pyrenäen (ASTRE 1932) und in Sizilien (CAMOIN 1983) wurde sie aus Riffbildungen des Mittleren Alb beschrieben.

Die Art *Eoradiolites murgensis* TORRE ist in Italien (Latinum-Gebirge, Murge Baresi; TORRE 1965) aus dem Unteren Alb und in Sizilien aus dem Mittleren Alb (CAMOIN 1983) beschrieben. *Toucasia carinata* MATH. ist vom Unteren Apt bis ins Obere Alb bekannt. Auch die biostratigraphische Verbreitung der Nerineen und Nerinellen ist sehr groß (Text-Fig. 6). *Plesioptyxis cretacea* (CONRAD), *Pl. fleuriaui* (D'ORB.), *Nerinea titan* SHARPE wurden in Portugal im Mittleren und Oberen Alb gefunden (BERTHOU & TERMIER 1972). Unter den Chondrodonten-Arten ist *Chondrodonta munsoni* HILL aus dem Oberen Apt und Alb von Mexiko und Kalifornien bekannt (STANTON 1947).

Die Schnecken- und Muschelfauna bestätigt im Vergleich, daß die Környe-Kalk-Formation ins Mittlere Alb gehört. Text-Fig. 11 gibt den Versuch einer Korrelation der Környe-Kalk-Formation mit anderen Gebieten wieder.

Im Vorland des Vértes sind die faunistischen Verbindungen der Környe-Kalk-Formation in S-SW-Richtung zu suchen; hier erfolgt eine Vermischung von Elementen der Westeuropäischen Subprovinz, der Jugoslawisch-Präapulischen und der Apulischen Faunenprovinzen. Natürlich ist bei diesem Vergleich auch auf die Endemismen der Faunen- und Subfaunenprovinzen Rücksicht zu nehmen.

Das Mittel-Alb der Provence ist im Raum Toulon und Marseille durch Korallen-Algen-Orbitolinen-Kalke repräsentiert. Da diese Kalke nur Caprini-



Text-Fig. 6. Chronostratigraphische Verbreitung der Mollusken der Környe-Kalk-Formation.

den enthalten, muß die Parallelisierung mit der Környe-Kalk-Fauna mit Hilfe der Mesorbitolinen erfolgen. In anderen Gegenden der Provence wurden die Alb-Sedimente des Oberen Urgons abgetragen (La Fare, Mt. Ventoux; ARNAUD-VANNEAU et al. 1982). In Slowenien, im Raum Loganski Planoti, enthalten die Alb-Kalke eine reiche Requienien-Fauna, und zwar vor allem eine *Toucasia-Caprotina-* und *Eoradiolites*-Fauna (ŠRIBAR 1979). In W-Slowenien, auf dem Gebiet des Präkarst von Frioul (VELIĆ et al. 1979) ist das Alb zum Teil durch Orbitolinen-Kalke repräsentiert, die denen des Vértes-Vorlandes entsprechen.

In S-Jugoslawien (Äußere Dinariden) besteht die Rudisten-Fauna hauptsächlich aus Caprotinen, von Eoradioliten-Arten begleitet. Sie enthält eine verhältnismäßig reiche Nerineen-Fauna. Diese Fauna ist für das Mittel-Alb charakteristisch (VELIĆ et al. 1979).

In den französischen Nord-Pyrenäen wurden von ASTRE (1932) Faunen-Assoziationen mit *Toucasia-Agriopleura-Eoradiolites* erwähnt. Aufgrund der in diesen Faunen sich befindenden *Pseudotoucasia santanderensis* DOUV. und *Eoradiolites*-Arten ist sie mit der Fauna der Környe-Kalk-Formation gut vergleichbar.

PHILIP (1981) erwähnt aus dem Aquitaine-Becken eine Faunen-Assoziation des Unteren Alb mit Riffcharakter, und zwar mit Hexacorallen, Pachyodonten und Nerineen. Die von ihm beschriebene Eoradioliten- und Nerineen-Fauna zeigt ebenfalls große Ähnlichkeit mit der Fauna der Környe-Kalk-Formation.

In Spanien kann man die Riffbildungen des Urgons (RAT 1959) mit Toucasien-Agriopleuren-Nerineen-Assoziationen, stellenweise mit Liostreen-Zwischenlagerungen, weiter verfolgen; sie sind von Apt/Alb-Alter.

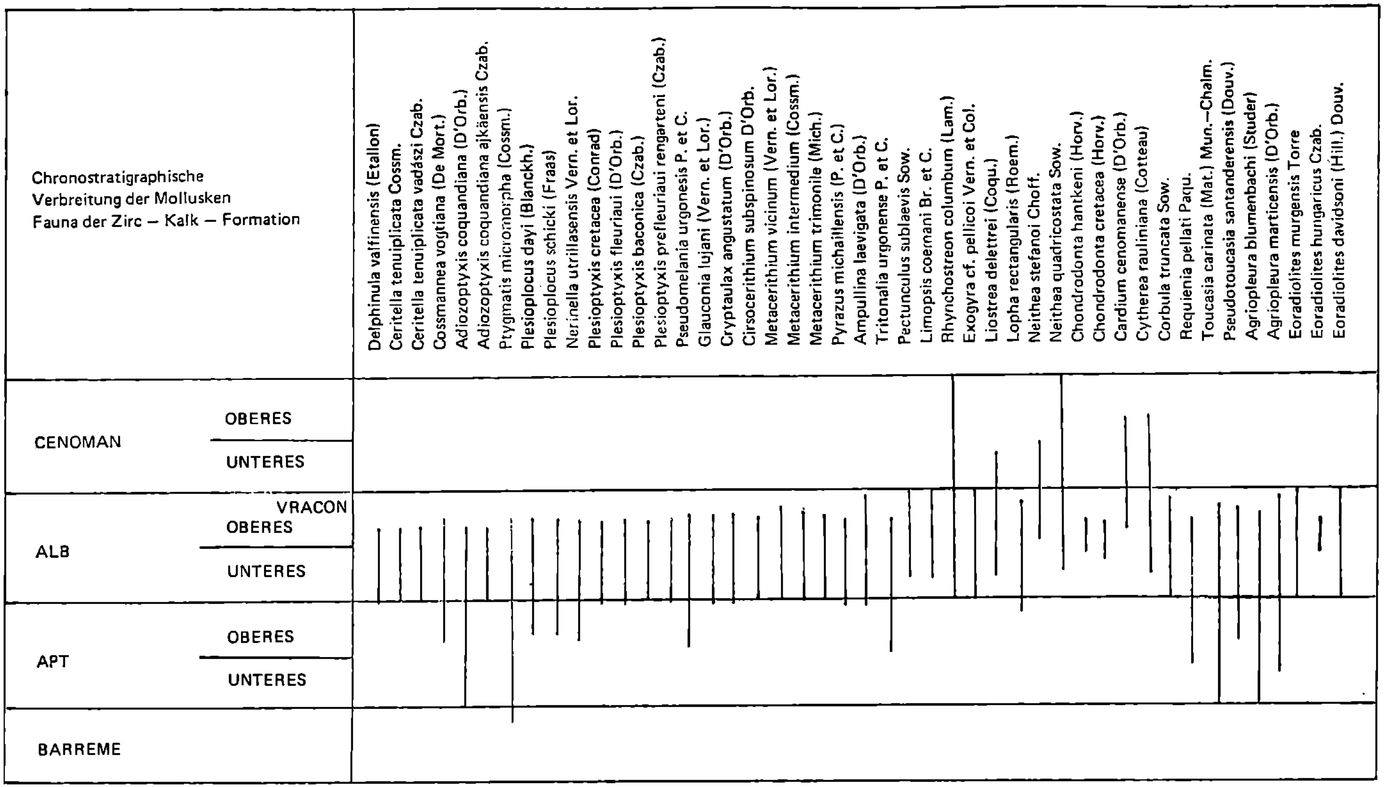
In den Pyrenäen (Navarro-Languedoc) können in der an die Iberische Platte angrenzenden Südzone die Mittel-Alb-Kalke des Urgons mit ihrer Fauna von *Pseudotoucasia santanderensis* DOUV., *Toucasia carinata* MATH. und *Eoradiolites murgensis* TORRE mit der Fauna unserer Faunenzone gut korreliert werden. Die Faunen-Assoziation kann auch in der Nordzone weiter verfolgt werden, die mit der europäischen Platte in Verbindung steht; in der Südzone setzen sich diese Assoziationen bei Lobe Montigri fort. Demgegenüber kommen in der Umgebung von Lobe Figueras nur Urgon-Bildungen des Barreme-Apt vor (PEYBERNÈS 1979).

Nach Feststellung von WIEDMANN et al. (1983) existiert im Alb der Basko-Kantabrischen Ketten eine Riff-Formation, deren reiche Rudisten- und Nerineenfauna mit der Fauna der Környe-Kalk-Formation altersmäßig gut übereinstimmt.

In der Estremadura Portugals (BERTHOU & LAUVERJAT 1978) wurde eine dem Vértes-Vorland ähnliche Orbitolinen-Algen-Nerineen-Pachyodonten-Fauna beschrieben. Ein Teil der in dieser Fauna enthaltenen Eoradioliten- und Nerineen-Arten stimmt mit der Fauna der Környe-Kalk-Formation gut überein. Die Chondrodonten fehlen hier, und es erscheinen Caprotinen des Alb. Eine ähnliche Faunen-Assoziation ist aus Südportugal (Algarve) bekannt.

In Italien (S. Polo Matese, Campobasso) hat MAINELLI (1983) im Zeitraum Barreme - Cenoman mehrere Rudisten-Zonen aufgestellt. Er charakterisiert u. a. die von ihm bestimmte "Zone D" durch die Art *Eoradiolites rousseli* (TOUCAS). Auch diese Mittel-Alb-Fauna zeigt gute Übereinstimmung mit unserer Fauna.

In Murge Baresi (Latinum-Gebirge) entsprechen nur die unteren Toucasien-Eoradioliten-Schichten der Környe-Kalk-Formation. In Sizilien wurden von CAMOIN (1983) Faunen von ähnlicher Entwicklung und Zusammensetzung



Text-Fig. 7. Chronostratigraphische Verbreitung der Mollusken der Zirc-Kalk-Formation.

aus dem Profil des Allaura-Berges und aus dem Becken von Imerese beschrieben, mit *Eoradiolites*, *Agriopleura* und Orbitolinen. Im Alb fehlen hier die Chondrodonten-Arten, die im Oberapt noch erwähnt sind. In der "Allaura-Formation" hat CAMOIN (1983) eine Faunenzone mit den Arten *Agriopleura darderi* ASTRE und *Eoradiolites davidsoni* (HILL), die - wie in Murge Baresi - mit Unterem und Mittlerem Alb datiert werden kann.

Zusammenfassend kann gesagt werden, daß die Fauna der Környe-Kalk-Formation älter ist als die der Zirc-Kalk-Formation, sie gehört wahrscheinlich ins Unter/Mittel-Alb.

3.3 Zirc-Kalk-Formation

Wie aus den Faunenlisten (Text-Fig. 6, 7) hervorgeht, hat die Zirc-Kalk-Formation unter den Nerinellen-Nerineen-Arten mehrere gemeinsame Formen mit der Környe-Kalk-Formation, im Unterschied aber erscheinen große Eoradioliten (*Eoradiolites davidsoni* (HILL), *E. hungaricus* CZAB.), die für das Obere Alb kennzeichnend sind. Die stratigraphische Verbreitung der urgonen Nerineen-Nerinellen-Fauna erstreckt sich auf das gesamte Apt-Alb (Text-Fig. 7); lediglich *Plesioptyxis cretacea* (CONRAD), *Pl. fleuriaui* (D'ORB.), *Pyrazus michaillensis* (P. & C.) persistieren in das Untercenoman.

Im Süd-Bakony dominieren in mehreren Faunen-Assoziationen die Chondrodonten-Arten. Ähnliche Faunen kommen nach PARONA (1909) in Nord-Italien im Friauler Gebirge vor, wo sie von einer Nerineen-Nerinellen-Fauna begleitet werden. Diese Fauna gehört zum Untercenoman und ist jünger als die Zirc-Kalk-Formation.

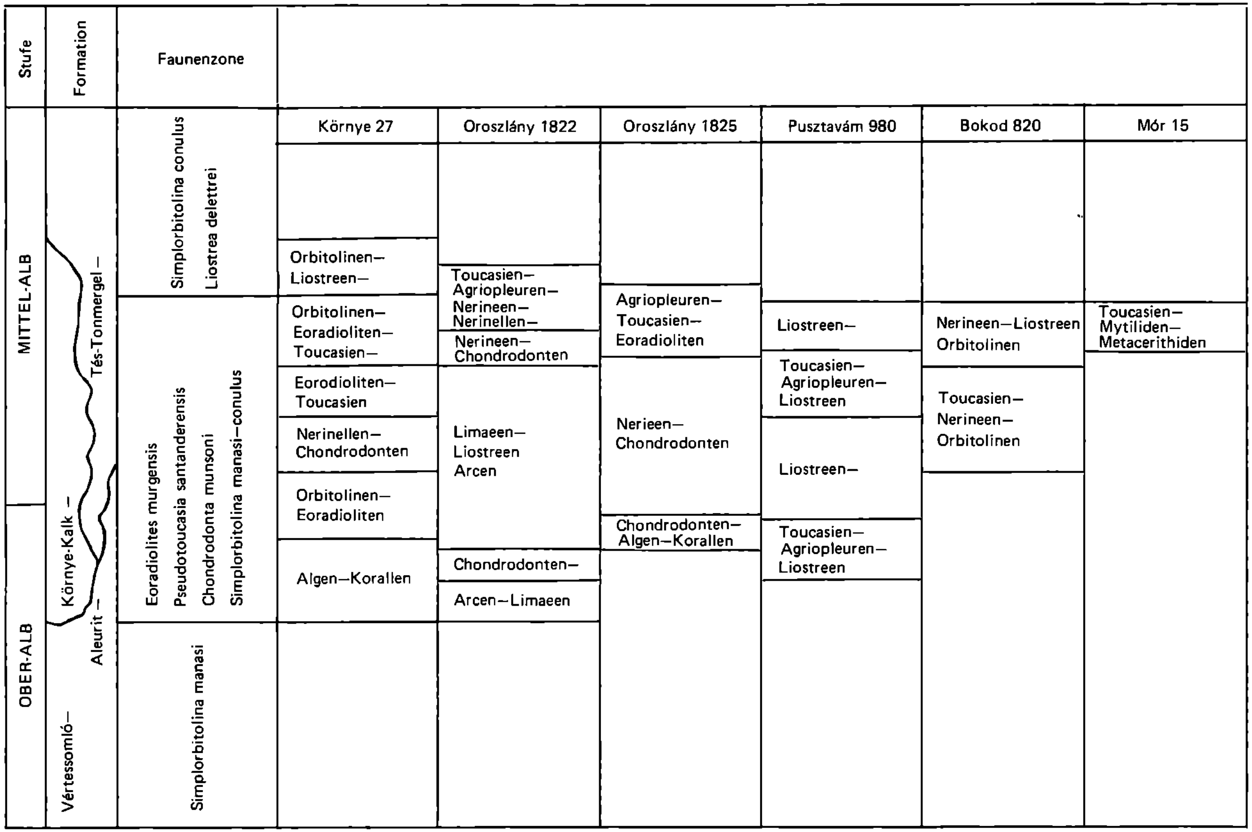
Die Chondrodonten-Arten des Süd-Bakony, *Chondrodonta hantkeni* und *Ch. cretacea*, stehen der Art *Ch. desioi* am nächsten. Die Entwicklungshöhe der Arten - obwohl endemisch - spricht für mittleres Ober-Alb (Text-Fig. 7).

Die Beziehungen der Zirc-Kalk-Fauna sind wie im Fall der Környe-Kalk-Formation in S-SW-Richtung zu suchen (Text-Fig. 11).

In den Pyrenäen kann sie mit der Biozone der Mesorbitolina conulus identifiziert werden (PEYBERNES 1979). Im südspanischen Präbetikum (Sierra Cazorla und Segura) zeigt GARCIA HERNANDEZ (1979) im "zweiten Sedimentzyklus" Riffbildungen mit einer ähnlichen Pachyodontenfauna. Aus den Basko-Kantabrischen Ketten (Umgebung von Albéniz-Eguino) wurden Supra-Urgon-Kalke mit einer reichen Pachyodonten-Korallen-Orbitolinen-Algen-Fauna beschrieben (WIEDMANN et al. 1983), die mit der Fauna der Zirc-Kalk-Formation gut korrelierbar sind. Der abweichende Charakter liegt in der reichen Caprotinen-Fauna. Im Biskaya-Synklinorium und im Estella-Becken wird sie durch Orbitolinen-Kalke des Oberen Alb repräsentiert.

In Portugal (Estremadura und Algarve) kann die Fauna der höheren Nerineen-Pachyodonten (Eoradioliten)-Orbitolinen-Kalke (Oberes Alb) mit der Fauna der Zirc-Kalk-Formation gut korreliert werden.

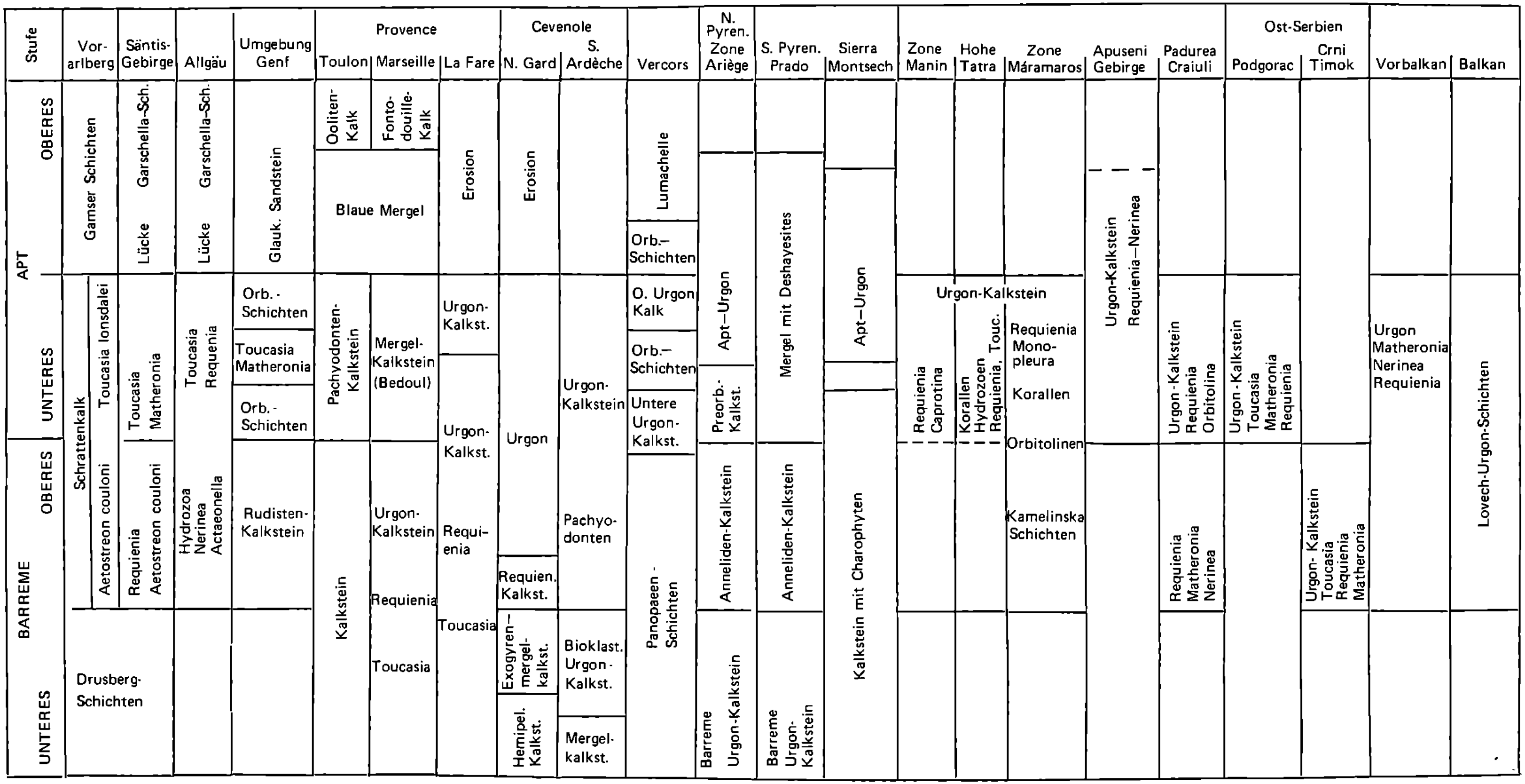
Auch in Italien, und zwar im Murge Baresi (TORRE 1965), in Sizilien (CAMOIN 1983), im Allaura-Berg und im Imerese wird das Obere Alb durch ähnliche Eoradioliten-Nerineen-Faunen-Assoziationen vertreten. In Nord-Italien ist im Monte D'Ocre (PARONA 1909) eine Fauna mit sehr ähnlichem Charakter, mit einer reichen Nerineen- und Chondrodontenfauna bekannt, die jedoch schon ins Untercenoman gestellt wird. Die neuen Chondrodonten-Arten des Bakony-Gebirges stehen - wie bereits angedeutet - der Art *Chondrodonta desioi* DOUV. nahe, die im Ober-Alb/Cenoman häufig ist.



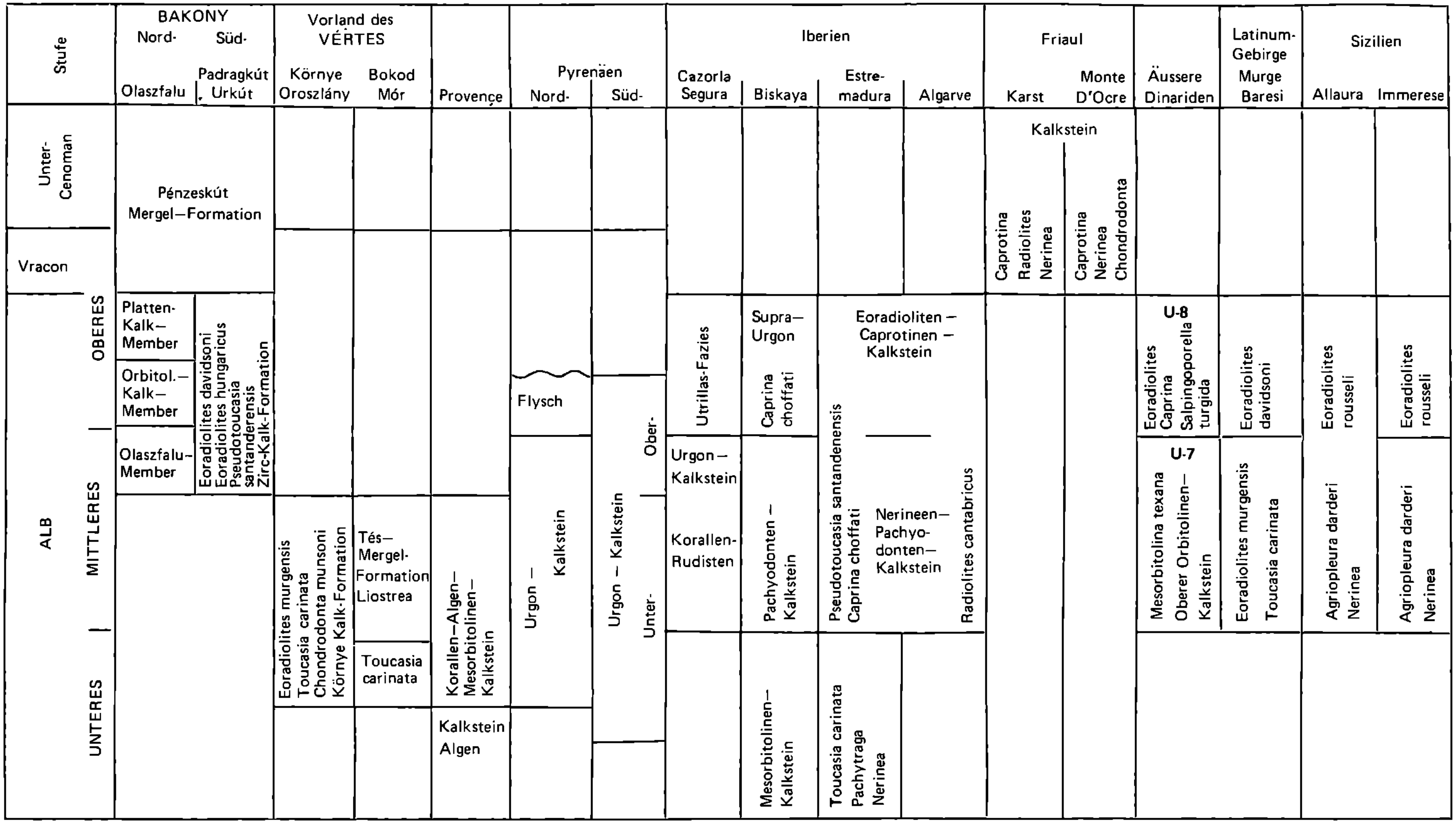
Text-Fig. 8. Die Verbreitung der Faunenzonen und Faunen-Assoziationen der Környe-Kalk-Formation.

Stufe	Formation	Member	Nord-Bakony		Süd-Bakony Padragkút		Urkút	
			Faunenzone	Assoziation	Faunenzone	Assoziation	Faunenzone	Assoziation
OBER-ALB	ZIRC-KALK-	Platten-Kalk	Orbitolina ex gr. texana Orbitolina concava	Orbitolinen–Mollusken	Eoradiolites davidsoni Eoradiolites hungaricus	Toucasien–Pseudotoucasien **A** Kleinw. Gastropoden **B** Nerinellen–Nerineen–Toucasien **C** Toucasien–Pseudotoucasien **A** Kleinw. Gastropoden **B** Algen–Nerineen–Toucasien–Eoradioliten **D** Kleinw. Gastropoden **B** Algen–Eoradioliten **E** Kleinw. Gastropoden **B**	Eoradiolites davidsoni Pseudotoucasia santanderensis	Nerinellen–Nerineen–Toucasien **A** Nerinellen–Nerineen Kleinw. Gastropoden Chondrodonten **B** Nerinellen–Nerineen–Toucasien **A**
		Orbitolinen-Kalk	Rhynchostreon Plesioptyxis laconica Orbitolina concava	Rhynchostreen–Plesioptyxiden Orbitolinen	S E N O N Eoradiolites davidsoni Chondrodonta hantkeni	S E N O N Nerinellen–Nerineen–Toucasien **C** Toucasien–Pseudotoucasien **A** Algen–Nerineen–Toucasien Eoradioliten **D**	Nerinella utrillasensis Chondrodonta hantkeni Plesioptyxis fleuriaui	Nerinellen–Nerineen–Chondrodonten **C** Nerinellen–Nerineen–Chondrodonten **D**
MITTEL-ALB		Olaszfalu-	Pseudotoucasia santanderensis Toucasia carinata	Toucasien–Pseudotoucasien Agriopleuren Nerineen	Eoradiolites murgensis Pseudotoucasia santanderensis Toucasia carinata	Kleinwüchsige Gastropoden **B**	–Algen–	Nerinellen–Nerineen Kleinw. Gastropoden Chondrodonten **B**
					Liostrea delettrei	Liostreen–Kleinw. Gastropoden **F**	Liostrea delettrei Arcostreon rectangularis	Liostreen–Arcostreen **E**

Text-Fig. 9. Die Verbreitung der Faunenzonen und Faunen-Assoziationen der Zirc-Kalk-Formation.



Text-Fig. 10. Korrelation des Schrattenkalks.



Text-Fig. 11. Korrelation der Környe-Kalk- und Zirc-Kalk-Formationen.

In Jugoslawien, in den Äußeren Dinariden, wird eine reiche Pachyodonten-Nerineen-Hydrozoen-Korallen-Assoziation in der "Faunenzone U-8" beschrieben (VELIC et al. 1979). Sie kann in ihrem Alter mit der Fauna der Zirc-Kalk-Formation verglichen werden; zwar fehlen dort die sog. perimediterranen Elemente, dagegen dominieren südliche und endemische Elemente.

Die Fauna der Környe-Kalk- und Zirc-Kalk-Formationen gehört zur Europäischen Faunenprovinz, und zwar insbesondere zur Westeuropäischen Subprovinz; untergeordnet verbinden sie einzelne Arten auch mit der Apulischen (Präapulischen) Faunenprovinz. Zur Balkanischen Subprovinz bestehen dagegen nur geringe Beziehungen, wobei der Grund in der zeitlichen Verschiedenheit liegt. Es ist anzunehmen, daß die Faunen der südlichen Gebiete Ungarns (Villány-Mecsek) enger mit dem Jugoslawisch-Ungarischen Block im Bereich der Balkanischen Subprovinz verbunden waren.

4. Zusammenfassung

Untersucht wurden die Muschel- und Schneckenfaunen des Urgons des Ober-Barreme/Unter-Apt (Schrattenkalk, Vorarlberg), Unter/Mittel-Alb (Környe-Kalk-Formation, Vorland des Vértes-Gebirges) und Mittel/Ober-Alb (Zirc-Kalk-Formation, nördliches und südliches Bakony-Gebirge).

Aufgrund der Pachyodonten-Ostreen-Chondrodonten-Nerineen-Fauna konnten in der Schichtfolge des Urgons Faunenzonen aufgestellt werden, die mit den Faunen der Westeuropäischen Subfaunenprovinz verglichen werden.

1. Die Faunenzone der Toucasia lonsdalei-Aetostreon couloni des Ober-Barreme/Unter-Apt (Schrattenkalk, Mittagspitz) wurde in zwei Subzonen gegliedert, die des Aetostreon couloni-Arctostreon rectangularis und der Toucasia lonsdalei-Toucasia carinata. Die Faunenzone kann mit den synchronen Faunen der Westeuropäischen Subfaunenprovinz der Ostschweiz (Helvetische Decke), Deutschlands (Allgäu), der SW-Schweiz (Umgebung Genf), SE-Frankreichs (Orgon, Vercors), S-Frankreichs (Provence) und N-Spaniens (Basko-Kantabrisches Gebirge) verglichen werden.

Im Gebiet der Karpaten (Tschechoslowakei, Rumänien, Sowjetunion), der Hohen Tatra (Polen) und des Balkans (Ost-Serbien, Bulgarien) kommen auch Urgon-Sedimente mit einer Pachyodonten-Algen-Orbitolinen-Hydrozoen-Korallen-Assoziation vor, die aber der Balkanischen Subfaunenprovinz angehören.

2. Die Faunenzone der Toucasia carinata-Eoradiolites murgensis-Chondrodonta munsoni des Unter/Mittel-Alb (Környe-Kalk-Formation) gehört ebenfalls zur Westeuropäischen Subfaunenprovinz. Ein faunistischer Zusammenhang mit den Faunen von S-Frankreich, N-Spanien (Pyrenäen, Basko-Kantabrisches Gebirge) und Portugal (Estremadura, Algarve) ist festzustellen. Es wurde dabei eine merkwürdige Ähnlichkeit mit den Faunen der Jugoslawisch-Präapulischen Subfaunen- und Apulischen Faunenprovinz (Präkarst, Latinum-Gebirge, Sizilien) festgestellt.

3. Das Urgon des Mittel/Ober-Alb (Zirc-Kalk-Formation) ist im Nord-Bakony mit drei Schichtgliedern vertreten (Olaszfalu-Orbitolinen-Plattenkalk). Das untere wurde als Faunenzone der Toucasia carinata-Pseudotoucasia santanderensis genannt, die mit dem unteren Profil-Abschnitt des Süd-Bakony-Gebirges gut korrelierbar ist.

Im südlichen Bakony ist die Faunenzone des Eoradiolites davidsoni-E. hungaricus kennzeichnend. Im Padragkút überwiegen die Fore-Reef-Faunen (mit *Pseudotoucasia santanderensis, Toucasia carinata*), demgegenüber im

Úrkút die Back-Reef-Sedimente mit einer reichen Chondrodonten-Fauna. Die faunistische Verbindung der Fauna der Zirc-Kalk-Formation kann in S-SW-Richtung gesucht werden. Die Faunen der Környe-Kalk- und Zirc-Kalk-Formationen gehören zur Europäischen Faunenprovinz mit einer engen Beziehung zur Westeuropäischen Subfaunenprovinz; sie sind jedoch durch viele Merkmale mit der Apulischen (Präapulischen) Faunenprovinz im Süden verknüpft.

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Redeposited Blocks of Seewen Limestone and Facies Differentiation in the Helvetic Amden Formation (Santonian-Campanian), German Alps*

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With 5 Text-Figures and 1 Table

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Abstract: Occurrences of Seewen Limestone (Turonian-Coniacian) in the marls of the Amden Formation (Santonian-Campanian) in the Alpine Helvetic Zone of the Allgäu and Kleinwalsertal (South Germany, Austria) have been studied and were interpreted as dislocated blocks in a lower slope facies of the newly defined Ifen structure. The 1-30 m wide and 0.5-6 m thick blocks were increasingly rounded and finally lens-shaped during their transport through the marls. Details of their genesis and transport described are based on characteristic exposures. At the Ifen structure the Cenomanian-Turonian is not preserved and the lower 80-100 m of the Amden Fm. were deposited by phacoidal debris flows overlain by laminated marls. The dislocated blocks are characteristic of a marginal facies belt with bioturbated or homogeneous marls (mud/debris flows). They are succeeded distally by bioturbated marls. The transport of the blocks towards the NNE over tens to hundreds of metres (but less than 4 km) in a highly viscous matrix is opposite to the general dip of the Helvetic shelf. The Ifen structure was generated by differential compaction of the early Cretaceous carbonate platform sediments.

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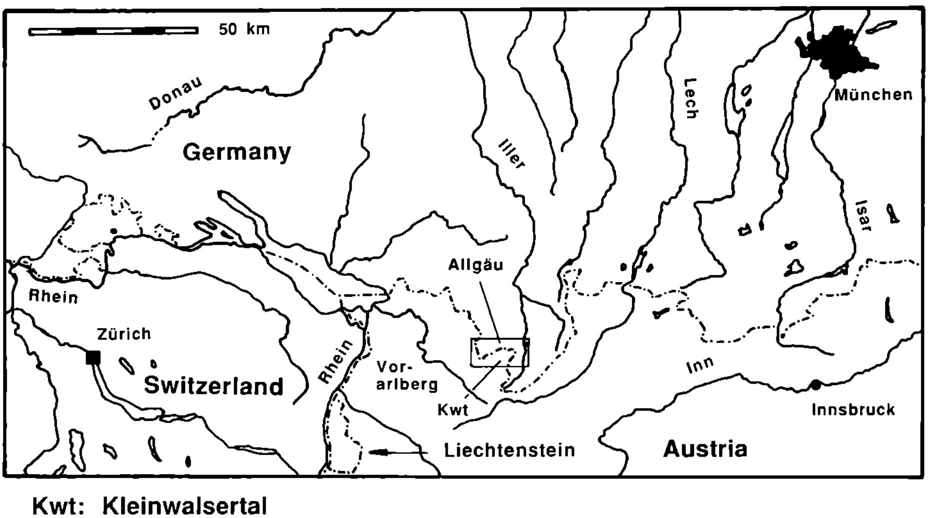
1. Introduction and stratigraphy

The Helvetic Zone of the northern Alps consists of sediments of the southern European shelf deposited prior to the Alpine orogenesis. In the Allgäu/Kleinwalsertal area (Text-Fig. 1) west of the rivers Breitach and Iller the Helvetic sediments are preserved and exposed in the Säntis Nappe. Most of the post-Santonian deposits, however, were eroded prior to the Eocene or truncated by the overthrust of the North Penninic Flysch nappes. The purpose of this paper is to report on evidence for pre-orogenic erosion and redeposition of sediment during the Santonian, related to local structural highs.

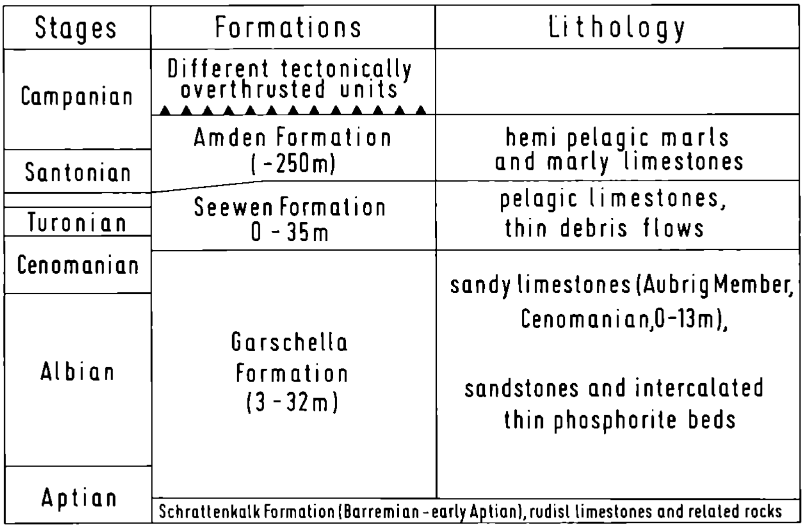
The Helvetic sequence dealt with in this paper (Text-Fig. 2) starts above the thick Schrattenkalk Formation ("Urgonian Facies", Barremian-early Aptian carbonate platform sediments). The Aptian-Albian (Garschella Formation, FÖLLMI & OUWEHAND 1987) is characterized by low sedimentation rates, stratigraphic condensation, and consists of glauconite sandstones and thin phosphorite beds. The Aubrig Member (late Albian-Cenomanian) consists of sandy glauconite limestones and glauconite sandstones.

The pelagic Seewen Limestone starts with the Fugenschicht which is in general a thin bed of redeposited glauconitic limestone (3-50 cm) mostly deposited during the Cenomanian-Turonian boundary interval in our working area (LIEDHOLZ et al. 1983, WEIDICH et al. 1983, WEIDICH 1984). This bed is overlain by the pelagic Seewen Limestone (Turonian-early Santonian, LIEDHOLZ et al. 1983, WEIDICH et al. 1983, WEIDICH 1984) but may start in the late Albian(?)-Cenomanian elsewhere (FÖLLMI & OUWEHAND 1987).

The succession above the Seewen Fm. was named as Amdenermergel by HEIM & OBERHOLZER (1907). The Amden Formation was introduced by



Text-Fig. 1. Geographical position of the Allgäu and Kleinwalsertal. Rectangle outlines approximate extend of the Helvetic Zone in the study area. For comparison with Text-Fig. 3: Oberstdorf is marked by closed circle at the eastern margin of the study area.



Text-Fig. 2. Stratigraphy of the Aptian-Campanian of the Alpine Helvetic Zone of the Allgäu and Kleinwalsertal.

OBERHÄNSLI-LANGENEGGER (1978) according to modern lithostratigraphic criteria and includes the late Santonian-early Campanian hemipelagic marls and marly limestones of the Alpine Helvetic Zone. Redeposition of older foraminifera and mixing of faunal elements of different depositional environments was reported from Vorarlberg and Switzerland by OBERHÄNSLI-LANGENEGGER (1978). She has also reported on contourites and occasionally turbidites which we have subsequently identified in the Allgäu and Kleinwalsertal.

The late Campanian-Maastrichtian Wang Formation is only known from a few localities. It is probably derived tectonically from the south Helvetic facies zone which is not exposed in our working area (middle Helvetic facies zone). The Amden Fm. is probably the youngest Cretaceous formation which escaped the post-Santonian and pre-Lutetian phases of erosion.

2. Earlier recognitions of "erratic" Seewen Limestone in the Amden Fm.

ARN. HEIM (1919) described an 8-9 m thick succession of Seewen Limestone intercalated into the Amden Fm. from the Breitachklamm section (Kleinwalsertal, LIEDHOLZ et al. 1983) which was a unique lithostratigraphic anomaly. OBERHÄNSLI-LANGENEGGER (1978) reported on comparable occurrences of Seewen Limestone and less frequent blocks derived from the Schrattenkalk and Garschella Formation. One of these sections (Rudachbach, Vorarlberg/Austria) was studied by FÖLLMI (1981) who suggested that the Seewen Limestone as a slide block was redeposited in the time of the Amden Formation.

In the Breitachklamm section (right bank of river) the autochthonous Seewen Limestone (Turonian, LIEDHOLZ et al. 1983) is overlain by 14-15 m of marls of the Amden Formation. The 8-9 m of Seewen Limestone intercalated into the Amden Fm. include an up to 50 cm thick marl seam which wedges out approximately westwards towards the Breitach river. This marl seam and the lower marl contain planktonic foraminifera of the late Santonian D. asymetrica Zone (LIEDHOLZ et al. 1983, WEIDICH 1984). Stratigraphic uncertainty arose from the recognition of *Dicarinella* cf. *concavata* (BROTZEN) and *Archaeoglobigerina cretacea* (D'ORBIGNY) in thin sections of the Seewen Limestone by HAGN (Munich, pers. comm.; samples from immediately below the marl seam taken by LIEDHOLZ). These planktonic foraminifera suggest the Coniacian-Santonian boundary interval. They co-occur with *Dicarinella hagni* (SCHEIBNEROVA) and *Dicarinella imbricata* (MORNOD) which are reworked Turonian species. However, based only on marl samples WEIDICH (1984) explicitly stated a late Santonian age (D. asymetrica Zone) for the interbedded Seewen Limestone and postulated local pelagic deposition (facies differentiation) in the hemipelagic Amden Formation. In the Seewen Limestone above the marl seam, HAGN (pers. comm.) found Coniacian species accompanied by reworked Turonian foraminifera. This agrees with our recent observations of matrix-supported rounded pebbles and indistinct flow structures in this interval which are indicative of a debris flow deposit. In addition we have recognized species of *Rotalipora* (Cenomanian) in this bed.

Our recent stratigraphic observations are summarized in Table 1. Following BOLLI (1944) we used large polished slabs of new closely spaced samples from the Seewen Limestone. The area of thin sections used for previous biostratigraphic studies of the Breitachklamm sections (up to 5 x 5 cm, i. e. 25 cm^2) is not large enough to permit the recognition of *D. asymetrica* unequivocally. According to our tests an area of at least 100 cm^2 should be inspected. Our samples are larger than 200 cm^2; the biostratigraphic dating is based on the inspection of about 500 determinable

Table 1. Litho- and biostratigraphic correlation in the Breitachklamm section (R 3592650, H 5250600). The autochthonous Coniacian Seewen Limestone was probably reworked at this locality during the late Santonian.

Lithostratigraphy		Biostratigraphy
Amden Formation		late Santonian (*D. asymetrica* Zone)
Seewen Limestone (2.2 m)	*	Coniacian incl. reworked Cenomanian-Turonian
injected marl (0 - 50 cm) seam derived from Amden Fm.		late Santonian (*D. asymetrica* Zone)
Seewen Limestone (6.1 m)	*	Coniacian - early Santonian
Amden Formation (14 - 15 m)		late Santonian (*D. asymetrica* Zone)
Seewen Limestone (2.3 m)	* *	Turonian

* "erratic" Seewen Limestone intercalated into Amden Formation; Interpreted as slide block derived from top of autochthonous section.

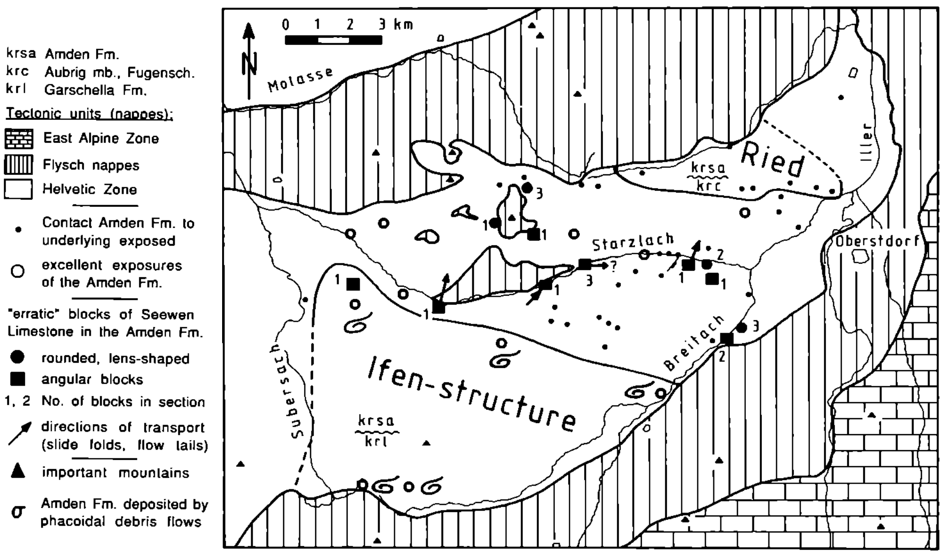
* * Seewen Limestone in normal stratigraphic position.

planktonic foraminifera. Foraminifera from the marls were studied from washed samples.

The 0-50 cm thick marl seam in the intercalated Seewen Limestone is injected material of the Amden Fm. which is confirmed by field observations and the microfauna. Additionally, we believe that it was unlikely that the debris flows should not have incorporated marl material of the Amden Fm. but reworked Cenomanian and Turonian limestones. We therefore reject the former assumptions of local pelagic sedimentation and facies differentiation at this locality by LIEDHOLZ et al. (1983) and WEIDICH (1984). The interbedded Seewen Limestone must have been derived from the normal stratigraphic succession below the Amden Formation. The debris flow deposits reflect redeposition of Seewen Limestone prior to the deposition of the Amden Formation. We have explored the Allgäu and Kleinwalsertal for more occurrences of "erratic" blocks in the Amden Formation. In addition, we have studied sites of tectonic deformation regarding the possible role of tectonic processes which could explain the dislocation of the Seewen Limestone.

3. Tectonic deformation of the Amden Formation

At various sites on the steeply inclined northern flanks of folds intense shear is indicated by a more or less irregular network of abundant calcite-filled joints. Although the primary bedding cannot be recognized due to



Text-Fig. 3. Geological map of the Allgäu and Kleinwalsertal area with locations of "erratic" blocks of Seewen Limestone in the Amden Formation. At the Ifen structure, the Amden Fm. was deposited by phacoidal debris flows. The Ried structure could have been passive during deposition of the Amden Fm.

rotation and shredding of the marly material, the three-dimensional character of sedimentary structures (e. g. ichnofossils) is not totally obliterated.

In the extreme northwest of our working area (Text-Fig. 3) tectonic reduction in thickness of the Seewen Limestone occurs through dislocation of beds from the top of the successions to their lower parts. Tectonic reduction therefore leads to tectonic dilatation at other nearby places. The tectonic transport of blocks occurs over distances of approx. 100 m but in no case we have observed injection of Seewen Limestone blocks into the Amden Fm. in the large and numerous exposures of that area.

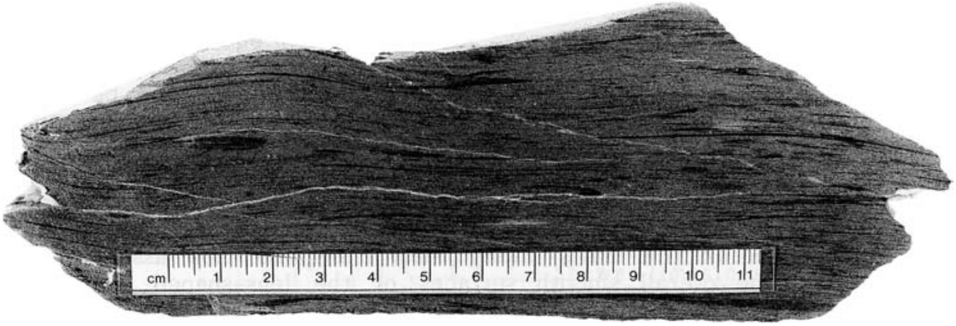
In some exposures of the contact between the Amden Fm. and the overthrusted Feuerstätter Flysch Nappe the rocks have suffered only little deformation during the overthrust. Others are characterized by drag folds and intense tectonic mixing of the overthrusted rocks in up to 10-15 m of vertical distance from the contact of the nappes. The preservation of the distorted sedimentary structures is three-dimensional. We have neither observed a flattening of ichnofossils nor a deformation which was not accompanied by calcite-filled joints. The original bedding was always distorted and the injection of flysch rocks into the Amden Fm. was always accompanied by strong deformation (folds, chaotic faulting, intense shear). We have never observed a mixing of different rock types which may have occurred if they were not cemented during their transport.

Thus, we cannot assume tectonic transport of the limestone blocks into the Amden Fm. although tectonic dislocation is common and intense. The features observed in relation to "erratic" Seewen Limestone exhibit markedly different characters.

4. New findings of "erratic" Seewen Limestone"

After detailed inspection we have identified 18 blocks of Seewen Limestone at 10 localities in the Allgäu and Kleinwalsertal (Text-Fig. 3). These include two above mentioned blocks in the Breitachklamm section which are separated by the 0-50 cm thick marl seam (Table 1). Three blocks in two sections (incl. the two in the Breitachklamm) are insufficiently exposed regarding that only the basal contact between the limestone and the Amden Fm. can be studied at these localities. In only one locality we have found a block of thin phosphorite and sandstone beds (repeated three times by folding). The exposure, however, does not permit unequivocally evaluation of its relation to the marls and was therefore excluded from our study. The following characteristics are common to all blocks of Seewen Limestone intercalated into the Amden Fm.

1. The marl/limestone contact is always very sharp. There are no ichnofossils penetrating from the limestone into the underlying or overlying marls.
2. The "erratic" blocks occur in a bioturbated or homogeneous matrix at a vertical distance of 3-30 m above the base of the Amden Fm.
3. Ichnofossils in the marls adjacent to the limestone within a distance of 10-50 cm are extremely flattened, squeezed, plastically deformed, and locally slightly folded (Text-Fig. 4) but never sheared or shredded. In homogeneous marls a shear lamination is observed.
4. Tectonically generated features, such as calcite-filled joints and faults are rare or absent from sites with limestones in the Amden Fm.



Text-Fig. 4. Flattened and squeezed ichnofossils in marls of the Amden Fm. close to the contact to an allochthonous block of Seewen Limestone. This ichnofossil preservation indicates high viscosity of the marls and laminar flow during transport of the blocks because otherwise the ichnofossils would have been destroyed. Rocks of this type are found 10-50 cm adjacent to the blocks. In the Walserschanz section they are indicators of their pathways through the bioturbated marls. At sites where the Amden Fm. is represented by homogeneous marls (mud flow; no ichnofossils), a shear lamination occurs in the Amden Fm. in comparable position.

5. There are "unconformities" between the bedding in the limestone blocks and the marls which range from about 10°-90°. The limestone beds are truncated at their margins, so that the contact to the marls is at least approximately conformable to their bedding.
6. Smaller blocks are generally lens-shaped and well rounded. Their size ranges between a width of 1-6 m and a thickness of 0.5-2 m.
7. Larger blocks are more angular. Their size ranges between a width of 6-30 m and a thickness of 2-6 m. They exhibit erosional features at their outer surface. The erosional products are pebbles (mm to dm-sized) and mixtures of completely disaggregated Seewen Limestone and marl exhibiting flow structures.
8. There is a tendency towards the dominance of rounded smaller blocks towards the NNE of our working area (Text-Fig. 3).

Shear lamination and flattened ichnofossils are not present near to the normal stratigraphic contact between the Seewen Limestone and the Amden Fm., even if it is tectonically overprinted. At sites of tectonic distortion the ichnofossils may be shredded or otherwise distorted but their fragments are still three-dimensional. This difference in preservation suggests plastic behaviour of the marl when the limestone blocks were transported and indicates a sedimentary process. Further evidence shall be given based on observations from three localities.

4.1 The Walserschanz section

This exposure is situated about 500 m away from the Breitachklamm section in the extreme SE of our working area. We have identified three blocks of

Seewen Limestone at this locality (Text-Fig. 3). The two lower blocks are lying close together in one horizon and overlap according to their lens-shape. They are separated by approx. 50 cm of marls. Both are well rounded, 1 m thick, 3.5 m and 8 m wide, respectively, and fully exposed. They are embedded in a matrix with squeezed ichnofossils which can be traced almost parallel to the bedding of the Amden Fm. in the approx. 3 m thick interval occupied by the blocks and up to a distance of 10 m from their margins. We suggest that this horizon was the pathway of the blocks through the marls. It grades into beds with three-dimensional ichnofossil preservation at its base and top, and indicates that the blocks were completely embedded in the marls during their transport. This assumption is confirmed by the overall smooth surfaces of the lens-shaped blocks and their orientation approximately parallel to the bedding of the marls, although the internal bedding in the limestones is inclined to it ("unconformity" of about 20°) and truncated at the contact between the limestone and the marl.

4.2 The Bärnloch section

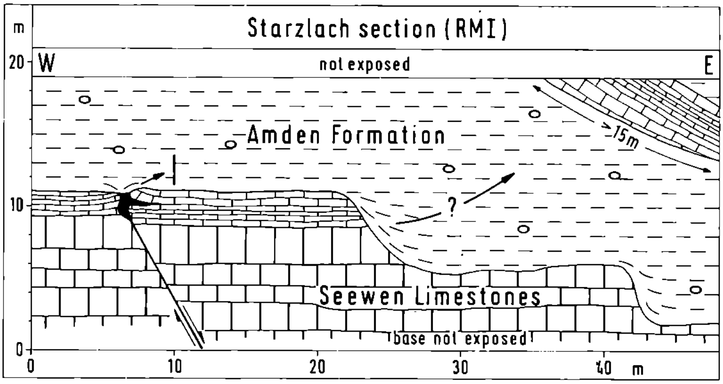
This section is located close to the centre of our working area and is marked by the eastward direction of transport as indicated in Text-Fig. 3. It is situated approx. 500 m SW of Rohrmoos in the Starzlach river. Three blocks of Seewen Limestone were identified in this section.

The lowermost block is slightly folded along its long axis and its thickness (2-3 m) varies due to truncation of beds. Most remarkable is the orientation of the long axis of the limestone block (approx. 15 m), viz. perpendicular to the bedding of the Amden Fm. It would be difficult to understand how this extreme "unconformity" of about 90° could have been generated by tectonic shear. The block and the adjacent marls do not exhibit faults or tectonic distortion as would be expected if the limestone would have been sheared and rotated by tectonic processes. More plausible is that the block was fully embedded and thus supported by the highly viscous marls in which rotation could have occurred without too much internal shear.

The exposure proves that the Seewen Limestone was firm and plastic but not cemented during its transport. Its outer surface exhibits small injections of marl into the limestone, and generation of pebbles 1-20 cm in diameter which fit into small grooves and hollows at the surface of the block. The rounding of the pebbles occurs rapidly over distances of transport through the marl of less than 1 m by erosion from the front of the pebbles. The material is then mixed with the marly matrix opposite to the direction of movement ("flow tails"). The direction of transport given in Text-Fig. 3 for the lower Bärnloch block reflects measured orientations of the flow tails which are suspected to have been modified by local current patterns adjacent to the block.

4.3 The Starzlach section

This exposure is situated approx. 3 km east of Rohrmoos in the Starzlach river. It is the easternmost section in which we were able to determine directions of transport (Text-Fig. 3). A sketch of the geologic situation is given in Text-Fig. 5.



Text-Fig. 5. Schematic sketch of the Starzlach section discussed in the text. Injected marls of the Amden Fm. and dikes in the uppermost beds of the autochthonous Seewen Limestone (west) are bound to syndepositional normal fault. In the east a relief at the top of the Seewen Limestone is compensated by the Amden Fm. and indicates erosion. Marls contain small rounded limestone pebbles and are homogeneous. Internal bedding of floating Seewen Limestone block (east) is truncated at the contact to the marl. The block may have been derived from the upper Seewen Limestone of this locality.

The autochthonous Seewen Limestone is exposed to the west. Its top is locally distorted by normal faults with displacements of less than 50 cm. These faults are covered by the Amden Fm. (D. asymetrica Zone) and are thus older. The marly material is injected into the faults and penetrates into the Seewen Limestone along the bedding planes forming up to 15 cm thick and up to 3 m long dikes. Approximately 50 cm thick successions of beds were uplifted by the injected marls (Text-Fig. 5) but are still in contact with the autochthonous section (Text-Fig. 5). This may reflect a very early stage for the origin of "erratic" Seewen Limestone in a phase prior to the separation from the underlying limestone.

A few metres further eastwards (Text-Fig. 5) the autochthonous Seewen Limestone cuts off and lies adjacent to the Amden Fm., however, faults are not observed. The relief caused by two approximately listric planes is compensated by the overlying marls. About 20 m further eastwards an approx. 4 m thick block of Seewen Limestone (exposed over approx. 15 m) is floating within the marls. Probably, it originated from the uppermost autochthonous Seewen Limestone at this locality and was transported for about 20 m.

Tectonic dislocation of the limestone is excluded on the evidence that the undistorted beds of the Amden Fm. can be traced throughout the exposure and the perfect compensation of the relief at the top of the Seewen Limestone. Small limestone pebbles with diameters of about 1 cm are found occasionally in the homogeneous marls of the Amden Formation. This may reflect deposition by debris or mud flow processes at this locality.

5. Facies differentiation of the Amden Fm. related to structural highs

In the southern part of our working area the late Santonian Amden Fm. rests on the Aptian-Albian Garschella Fm. (Text-Figs. 2, 3). Relicts of Seewen Limestone are found as components of thin debris flows or in thin mud flow channels at their top. This indicates that the absence of the Seewen Limestone is due to erosion rather than non-deposition. In this area the lower part of the Amden Fm. was deposited by numerous phacoidal debris flows which may sum up to 80-100 m thick successions and contain small blocks (< 1 m) derived from the Garschella Formation. The debris flow deposits are overlain by laminated grey marls with abundant slide folds. Based on the hiatus and extensive redeposition of sediment in this area, we have mapped a syndepositionally active structural high, which is referred to as the **Ifen structure** in Text-Fig. 3. It probably originated by differential compaction of the Schrattenkalk Fm. comparable to the situation reported from Vorarlberg by FÖLLMI & OUWEHAND (1987).

The "erratic" blocks of Seewen Limestone are restricted to an area NNE of the Ifen structure where the marls of the Amden Fm. are more or less continuously bioturbated, except for local occurrences of homogeneous mud flow deposits. The blocks are found in both, the bioturbated and the homogeneous facies and were detached from the autochthonous successions. While the Seewen Limestone was eroded by debris flows at the Ifen structure it was probably mobilized by local mud flows on its lower slope. In zones of weakness of the Seewen Limestone, e. g. where syndepositional faults occurred, the marls were injected into the underlying limestone leading to the detachment of the blocks which were subsequently transported away fully embedded in the highly viscous marl matrix. They were compacted and stable, however, not cemented during this process as is evident from abrasion and formation of pebbles. The limestones became increasingly rounded while they were transported revealing flat and lens-shaped disks in their late stage.

This interpretation is only valid if forces were available to move the blocks into the higher parts of their matrix during the mud flow process. A possible mechanism could have been the "BERNOULLI-Effect" introduced by FISHER & MATTINSON (1968). According to BERNOULLI's principle, the pressure in a fluid increases with decreasing flow velocity. Due to friction the velocity in a fluidized sediment decreases with depth. A large floating body will therefore be pushed upwards until the pressure difference at its base and top is in equilibrium with the weight pressure of the block. The weight of the block is significantly lowered by buoyancy in the marl matrix during this process. However, if the sediment has stopped to move there will be no forces except the shear strength of the matrix compensating for the weight of the limestone block. It will consequently start to sink downwards, however, due to its lens-shape there will also be a lateral component of movement. They could have penetrated into the adjacent bioturbated marls (which were not involved in the mud flow process) between the mud flow deposits. A possible pathway was identified in the bioturbated marls of the Walserschanz section.

In the northernmost part of our working area dislocated blocks of Seewen Limestone or other indications of extensive redeposition were not observed. The Amden Fm. consists of continuously bioturbated marls and is rarely distorted by small slump moulds.

In Text-Fig. 3 we have identified an area where the Amden Fm. rests on the Fugenschicht or the Aubrig Member (Text-Fig. 2) and have named it the **Ried structure** after the nearby village Ried. We have not identified redeposited sediments comparable to those of the Ifen structure in the Amden Fm., however, there are only small exposures available in this area. The Ried structure could have been passive in the Santonian.

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Planktonic and Benthonic Foraminiferal Zonations of the Lower Cretaceous of the Northern Calcareous Alps

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With 1 Plate and 1 Table

WEIDICH, K. F. (1989): Planktonic and Benthonic Foraminiferal Zonations of the Lower Cretaceous of the Northern Calcareous Alps. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 465-468. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

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The Munich working group on multistratigraphical subdivision of the uppermost Jurassic to Lower Cretaceous of the Northern Calcareous Alps (NCA) has almost fully completed its research work. The Lower Cretaceous ammonite stratigraphy and taxonomy have been published (IMMEL 1987), and the foraminiferal monograph is ready for printing (WEIDICH 1987). Studies of calpionellids, nannoconids, and Upper Jurassic ammonites (KAISER-WEIDICH & SCHAIRER in prep.), and of radiolarians (STEIGER in prep.), will follow.

The monograph on Lower Cretaceous foraminiferal faunas of the NCA is based on 42 sections and sample points from the Allgäu region (Germany) in the west to Vienna (Austria) in the east.

More than 400 washing samples were evaluated in order to achieve a detailed planktonic and benthonic foraminiferal stratigraphy. The samples were taken from Neocomian Aptychi Beds, Schrambach Beds, Roßfeld Beds, Lackbach Beds (DARGA & WEIDICH 1986), Thiersee facies (new described lithology, WEIDICH 1987), Tannheim Beds and Losenstein Beds.

The planktonic foraminiferal zonation (Table 1) is the same as known from other Tethyan regions. 55 species and subspecies are present.

The benthonic foraminiferal zonation (Table 1) is a local stratigraphy for the NCA. This zonation is a result of stratigraphic ranges of 392 species and subspecies. Zonal species derive from both groups, calcareous as well as agglutinated foraminifera.

These zonal species seem to be guide-fossils not only in the Alps but also in parts of Northern America, SE France, the Carpathians, the Crimea, Northern Caucasus and even as far as Western Siberian Lowlands (from literature and partly from comparative material).

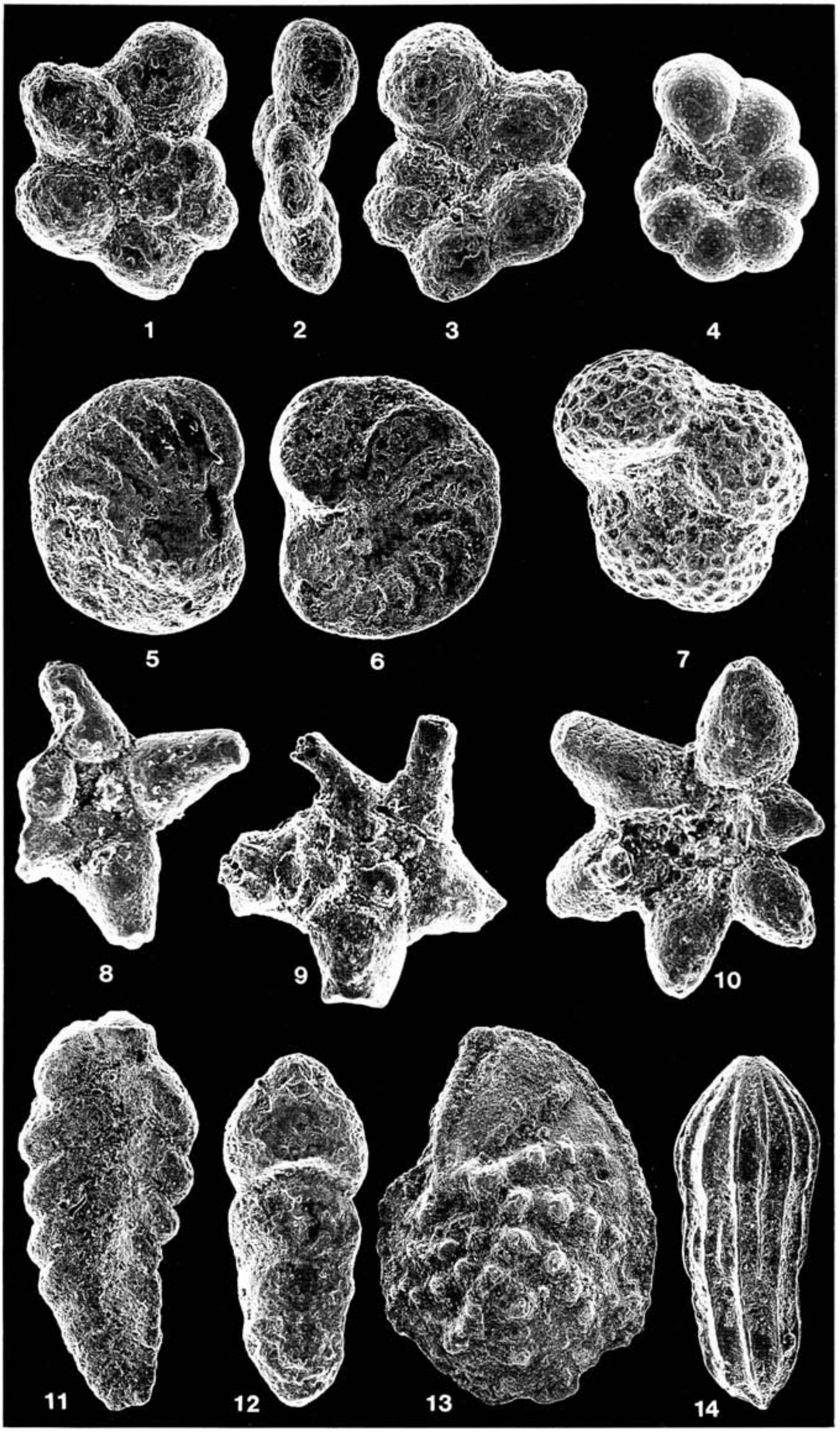
In the main part of the monograph, the foraminifera are described and figured on 62 plates. Eight species and subspecies are described for the

Table 1. Planktonic and benthonic foraminiferal zonations of the Lower Cretaceous of the Northern Calcareous Alps (from WEIDICH 1987, Ms.).

		PLANKTONISCHE FORAMINIFEREN ZONEN	PLANKTONISCHE FORAMINIFEREN SUBZONEN	BENTHONISCHE FORAMINIFEREN ZONEN	
OBER-KREIDE	CEN	*reicheli*			CEN
		13 *brotzeni*		10 *cenomanica/ cretosa*	
UNTER-	ALB	12 *appenninica*	12 b *appenninica/buxtorfi* 12 a *appenninica/ticinensis*		ALB
		11 *subticinensis/ ticinensis*		9 aff. *nitida/ macfadyeni*	
		10 *raynaudi breggiensis*		8 *berthelini/ imperfectus*	
		9 *primula*			
		8 *planispira*		7 *schloenbachi/ nonioninoides*	
		7 *gorbachicae*			
	APT	6 *algerianus*	6 b *algerianus/cheniourensis* 6 a *algerianus/ferreolensis*		APT
		5 *ferreolensis*		6 *intermedia/ dividens*	
		4 *cabri*			
KREIDE		3 *blowi/ similis*			
	BAR	2 *sigali*		5 *barremiana/ praedividens*	BAR
	HAU	1 *hoterivica*		4 *heiermanni/ vocontianus*; 1 *Lenticulina/Z. Spirillina-*	HAU
	VAL			3 *eichenbergi*	VAL
	BER			2 *nodosa/ kummi*	BER
OBER-JURA	TIT				TIT

Plate 1

Figs. 1-3. *Ticinella raynaudi digitalis* SIGAL. - Upper Albian; x 107.
Fig. 4. *Ticinella primula* LUTERBACHER. - Upper Albian; x 107.
Figs. 5-6. *Osangularia schloenbachi* (REUSS). - Lower Albian; x 107.
Fig. 7. *Favusella washitensis* (CARSEY). - Upper Albian; x 107.
Fig. 8. *Leupoldina pustulans* (BOLLI). - Aptian; x 214.
Fig. 9. *Leupoldina reicheli* (BOLLI). - Aptian; x 138.
Fig. 10. *Hedbergella bollii* LONGORIA. - Aptian; x 214.
Fig. 11. *Spiroplectinata lata* GRABERT. - Upper Albian; x 57.
Fig. 12. *Lingulina furcillata* BERTHELIN. - Aptian; x 138.
Fig. 13. *Lenticulina eichenbergi* BARTENSTEIN & BRAND. - Barremian; x 57.
Fig. 14. *Nodosaria paupercula* REUSS. - Upper Albian; x 57.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first time. In particular, emphasis is given to morphological variation, to fine structure and to phylogenetical developments.

Concluding chapters deal with the paleogeography of the Lower Cretaceous of the NCA, with the paleobiogeographical connections of the foraminiferal faunas to the west and to the east and with the paleoecology of those foraminiferal faunas.

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Zur Lebensweise der Ammoniten der Pénzeskut-Formation (Alb-Cenoman), Ungarn

ANNA HORVATH, Budapest

Mit 5 Tafeln und 2 Text-Figuren

HORVATH, A. (1989): Zur Lebensweise der Ammoniten der Pénzeskut-Formation (Alb-Cenoman), Ungarn. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 469-482. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The ammonite fauna of the Pénzeskut Formation (Lower Vraconian, Bakony Mts., Hungary) is investigated in the sections Jásd-1, Bakonynána-1 and the borehole Jásd-42. The fauna of the shallow-marine marls of Jásd-1 is reduced compared to the fauna of Bakonynána-1. Despite the general reduction of fossils in the section Jásd-1, the number of species and individuals of hysterocerids and discohoplitids is increased. To the basin center (borehole Jásd-42), this process will be reinforced.

It was supposed, that these differences (HORVATH 1985) are due to the influence of short-term oscillations, changes in temperature and salinity. In this paper, however, a different mode of life in different bathymetric levels is supposed. The fauna, which is enriched in heteromorphs and globulate species, probably lived near the place of embedding in a shallow-marine bottom-related environment (Bakonynána-1). For the hoplitids, a nektonic mode of life in an open neritic environment is suggested (Jásd-1 and Jásd-42). Biotope and area of burial were the same.

Kurzfassung: Ein Vergleich der Untervracon-Ammonitenfauna (Jásd-1 und Bakonynána-1, Pénzeskut-Formation, Bakony-Gebirge) zeigt, daß die Fauna der flachmarinen Mergelablagerungen von Jásd-1 gegenüber der Fauna von Bakonynána-1 stark verarmt ist. Gleichzeitig nimmt die Arten- und Individuenzahl der Hysteroceraten, zu denen sich auch Discohopliten gesellt haben, im Profil Jásd-1 beträchtlich zu. Zum Beckenzentrum hin (Stratotyp-Profil Jásd-42) wird dieser Prozeß weiter verstärkt.

Während diese Unterschiede früher (HORVATH 1985) auf den Einfluß von kurzfristigen Meeresspiegelschwankungen und auf Änderungen der Temperatur und Salinität zurückgeführt wurden, wird hier eine unterschiedliche Lebensweise in unterschiedlicher Wassertiefe angenommen. Wahrscheinlich hat die an Heteromorphen und globulösen Gehäuseformen angereicherte Fauna am Ort ihrer Einbettung im flachneritischen Bereich und bodenbezogen gelebt, während die schlanken Hopliten gute Schwimmer des offenen Neritikums gewesen sein dürften. Sedimentations- und Lebensraum dürften sich entsprochen haben.

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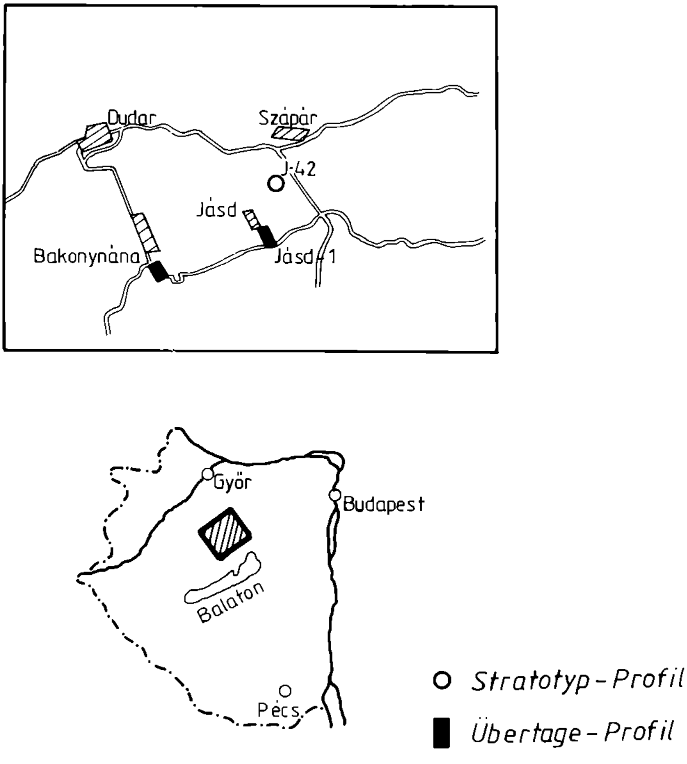
1. Einführung

Mit Hilfe von horizontiert gesammelten Ammoniten wurden die Ablagerungsbedingungen der Pénzeskut-Formation in zwei Übertageprofilen (Bakonynána-1, Jásd-1) und der Bohrung Jásd-42 (Stratotypenprofil) untersucht (Text-Fig. 1). Die Übertageprofile gehören nach den Ammonitenfaunen in das höchste Alb, und zwar in die Blancheti-Subzone der Dispar-Zone (Unteres Vracon). Demgegenüber entspricht die 474,5 m mächtige Schichtfolge der Bohrung Jásd-42 der gesamten Dispar-Zone (Blancheti- und Bergeri-Subzonen), der Zone des Mantelliceras mantelli (Unteres Cenoman) und eventuell im obersten faunenarmen Teil noch der Basis des Mittleren Cenomans (CSASZAR 1985a, 1985b, HORVATH 1985).

Ein auffallendes Merkmal der Profile ist ihr Ammonitenreichtum (Densität und Diversität): Insgesamt 1076 Exemplare können 38 Gattungen (und Untergattungen) bzw. 98 Arten (und Unterarten) zugeordnet werden. Außerdem sind in der Fauna divers vertreten und zum Teil bereits bearbeitet die Mikrofaunen (SIDO 1971, BODROGI 1985), Belemniten, Gastropoden, Bivalven (BENKÖ-CZABALAY 1965), Echiniden (SZÖRENYI 1955) und Brachiopoden.



Text-Fig. 1. Lageskizze der untersuchten Profile.

2. Schichtfolge und Ammonitenführung

Die transgressiven Serien des Vracon sind Ablagerungen eines epikontinentalen Schelfmeeres, mit stark reduzierter Temperatur. Diese Annahme wird u. a. durch eine starke Zunahme des bis dahin nur sporadisch auftretenden Glaukonitgehaltes belegt. Außerdem gelangte das Gebiet des Bakony-Gebirges mit der "Cenoman"-Transgression in den Einzugsbereich der borealen Faunenprovinz aus W/NW-Richtung (SZÖRENYI 1955, BENKÖ-CZABALAY 1965, KNAUER 1966). Für die hier untersuchten beiden Übertageprofile ebenso wie für das Untervracon der Bohrung Jásd-42 ist ein hoher Glaukonitgehalt kennzeichnend, der beim letztgenannten Profil nach oben allmählich abklingt (CSASZAR 1985a, 1985b).

2.1 Aufschluß Bakonynána-1

Das etwa 5 m mächtige Profil besteht aus einer Wechsellagerung von knolligen, glaukonitischen Kalken, dolomitischen Mergeln mit Kalksteinlinsen und Mergeln. Neben der Ammonitenfauna führt die Serie wenige limonitisierte Pflanzenreste, eine Ichnofauna und Echiniden (CSASZAR 1985a, HORVATH 1985). Die Ammoniten sind in etwa 1,5 m im mittleren Teil des Profils konzentriert.

Von den auf 16 Gattungen verteilten 44 Arten sind 18 Arten (und Unterarten) Vertreter der Heteromorphen, was 54,2 % der gesamten Ammonitenfauna entspricht. Es wurde wiederholt vermutet, daß die Heteromorphen dem vagilen Benthos zuzurechnen sind (TELEGDI-ROTH 1953: 712, WIEDMANN 1969: 590); insbesondere die Gehäuseform der Turriliten ist der einzelner Gastropodengruppen (u. a. der Turritellidae) so ähnlich, daß an eine vergleichbare Lebensweise zu denken ist. Auch für Formen mit geblähten Gehäusen, wie z. B. *Desmoceras (D.) latidorsatum,* Pervinquierien und ähnlichen Formen (Taf. 1, Fig. 1-10) kann eine stärker bodenbezogene Lebensweise angenommen werden.

2.2 Aufschluß Jásd-1

Die lithologische Beschaffenheit des Übertageprofils Jásd-1 (Unteres Vracon) ist der des Profils Bakonynána ähnlich. Obwohl die Schichtfolge hier mächtiger und die Verteilung der Ammoniten gleichmäßig ist, beträgt die Individuenzahl nur ein Drittel der Fauna von Bakonynána (HORVATH 1985: 153). Gleichzeitig ist aber die Gattungs- und Artenzahl nahezu gleich. Die insbesondere im Hinblick auf den Mächtigkeitsunterschied auffallende Abnahme der Individuenzahl ist ganz extrem bei den heteromorphen Formen (Lechiten, Anisoceraten, Turriliten, Scaphiten) und den Desmoceraten. Früher wurden als Ursache hierfür Lückenhaftigkeit der Sedimentation im Profil von Jásd, aber auch periodische Schwankungen des Salzgehalts vermutet (HORVATH 1985: 153). Heute wird die unterschiedliche Bathymetrie beider Sedimentationsräume für die Faunendifferenzierung verantwortlich gemacht. Außer der generellen Faunenverarmung erscheinen bei Jásd-1 als neues Element der Faunenassoziation die ersten feinberippten *Hyphoplites (Discohoplites)*-Arten mit ihren abgeflachten Gehäusen und demzufolge wahrscheinlich mit besserer Schwimmfähigkeit. Gleichzeitig kommt es bei Jásd-1 zu einer Vermehrung der Hysteroceraten mit einem mäßig ornamentierten, abgeflachten Gehäuse, und zwar ebenso an Arten wie an Individuen (Taf. 2, Fig. 1-8).

Tafel 1: Unter-Vracon-Ammoniten von Bakonynána-1

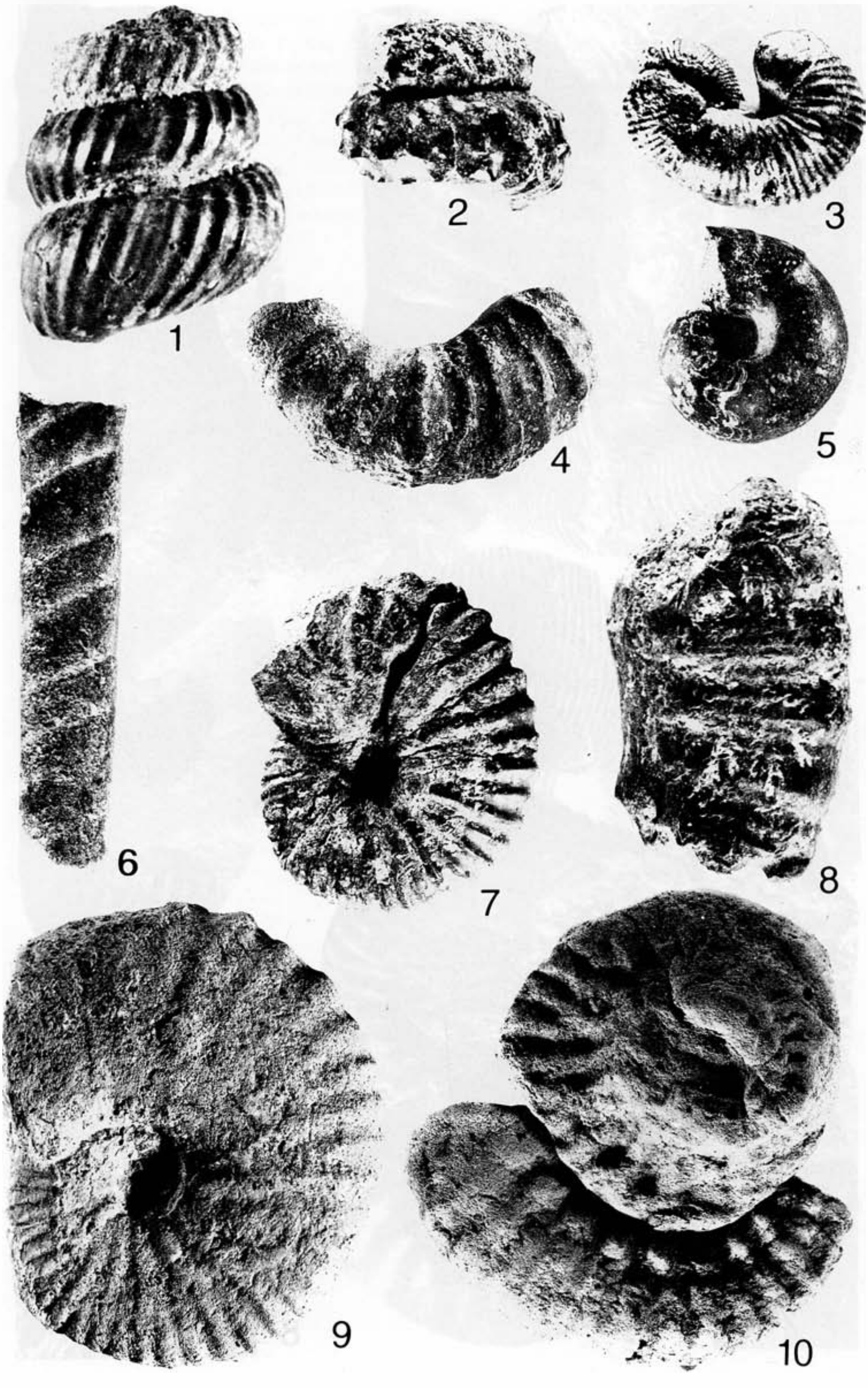
Fig. 1. *Turrilites (Turrilitoides) hugardianus* D'ORBIGNY. MAFI K-13781 Seitenansicht. 3/1.
Fig. 2. *Turrilites (Paraturrilites) escherianus* PICTET f. *cantabrigensis*. MAFI K-13787. Seitenansicht. 3/1.
Fig. 3. *Scaphites (Scaphites) hugardianus* D'ORBIGNY. MAFI K-13809. Lateralansicht. 2/1.
Fig. 4. *Anisoceras (Anisoceras) armatum* (SOWERBY). MAFI K-13771. Lateralansicht. 1/1.
Fig. 5. *Desmoceras (Desmoceras) latidorsatum* (MICHELIN). MAFI K-13887. Lateralansicht. 2/1.
Fig. 6. *Lechites gaudini* (PICTET & CAMPICHE). MAFI K-13746. Lateralansicht. 2/1.
Fig. 7. *Stoliczkaia dispar blancheti* (PICTET & CAMPICHE) f. *notha*. MAFI K-13904. Lateralansicht. 1/1.
Fig. 8. *Anisoceras (Anisoceras) pseudoelegans* PICTET & CAMPICHE. MAFI K-13770. Lateralansicht. 3/1.
Fig. 9. *Stoliczkaia dispar dispar* (D'ORBIGNY). MAFI K-13901. Lateralansicht. 1/1.
Fig. 10. *Pervinquieria (Pervinquieria) stoliczkai* SPATH. MAFI K-13885. Lateralansicht. 1/1.

Tafel 2: Unter-Vracon-Ammoniten von Jásd-1

Fig. 1. *Hamites (Hamites)* cf. *virgulatus* BRONGNIART. MAFI K-13737. Lateralansicht. 2/1.
Fig. 2. *Anisoceras (Anisoceras) armatum* (SOWERBY). MAFI K-13769. Lateralansicht. 1/1.
Fig. 3. *Hysteroceras (Cantabrigites) cantabrigense helveticum* (RENZ) f. *"valdense"*. MAFI K-13872. Lateralansicht. 1,5/1.
Fig. 4. *Scaphites (Scaphites) meriani* PICTET & CAMPICHE. MAFI K-13812. Ventralansicht. 1,5/1.
Fig. 5. *Hysteroceras (Cantabrigites) cantabrigense helveticum* (RENZ), Übergang zu f. *cantabrigense minor* f. *subsimplex*. MAFI K-13871. Lateralansicht. 1/1.
Fig. 6. *Turrilites (Turrilitoides) hungardianus* D'ORBIGNY. MAFI K-13779. Seitenansicht. 2,5/1.
Fig. 7. *Hysteroceras (Cantabrigites) cantabrigense* (SPATH) s. l. und *Pervinquieria (Pervinquieria)* cf. *pachys* (SEELEY). MAFI K-13889. Lateralansichten. 2/1.
Fig. 8. *Hysteroceras (Cantabrigites) cantabrigense* (SPATH) s. l. MAFI K-13868. 1/1.

Tafel 3: Unter-Vracon-Ammoniten von Jásd-42

Fig. 1. *Hyphoplites (Discohoplites) coelonatus coelonatus* (SEELEY) f. *subfalcatus*. MAFI K-13834. Lateralansicht. 2/1.
Fig. 2. *Hyphoplites (Discohoplites) coelonatus densecostatus* (RENZ). MAFI K-13835. Lateralansicht. 1/1.
Fig. 3. *Lechites moreti* BREISTROFFER. MAFI K-13754. Lateralansicht. 1,5/1.
Fig. 4. *Hyphoplites (Discohoplites) coelonatus densecostatus* (RENZ). MAFI K-13836. Lateralansicht. 1,5/1.
Fig. 5. *Hysteroceras (Cantabrigites) cantabrigense minor* (SPATH) f. *subsimplex*. MAFI K-13877. Lateralansicht. 2/1.
Fig. 6. *Hysteroceras (Cantabrigites) cantabrigense* cf. *minor* (SPATH). MAFI K-13874. Lateralansicht. 2/1.
Fig. 7. *Anisoceras (Anisoceras) armatum* (SOWERBY) f. *perarmatum*. MAFI K-13772. Lateralansicht. 1/1.
Fig. 8. *Lechites communis* SPATH. MAFI K-13752. Lateralansicht. 2/1.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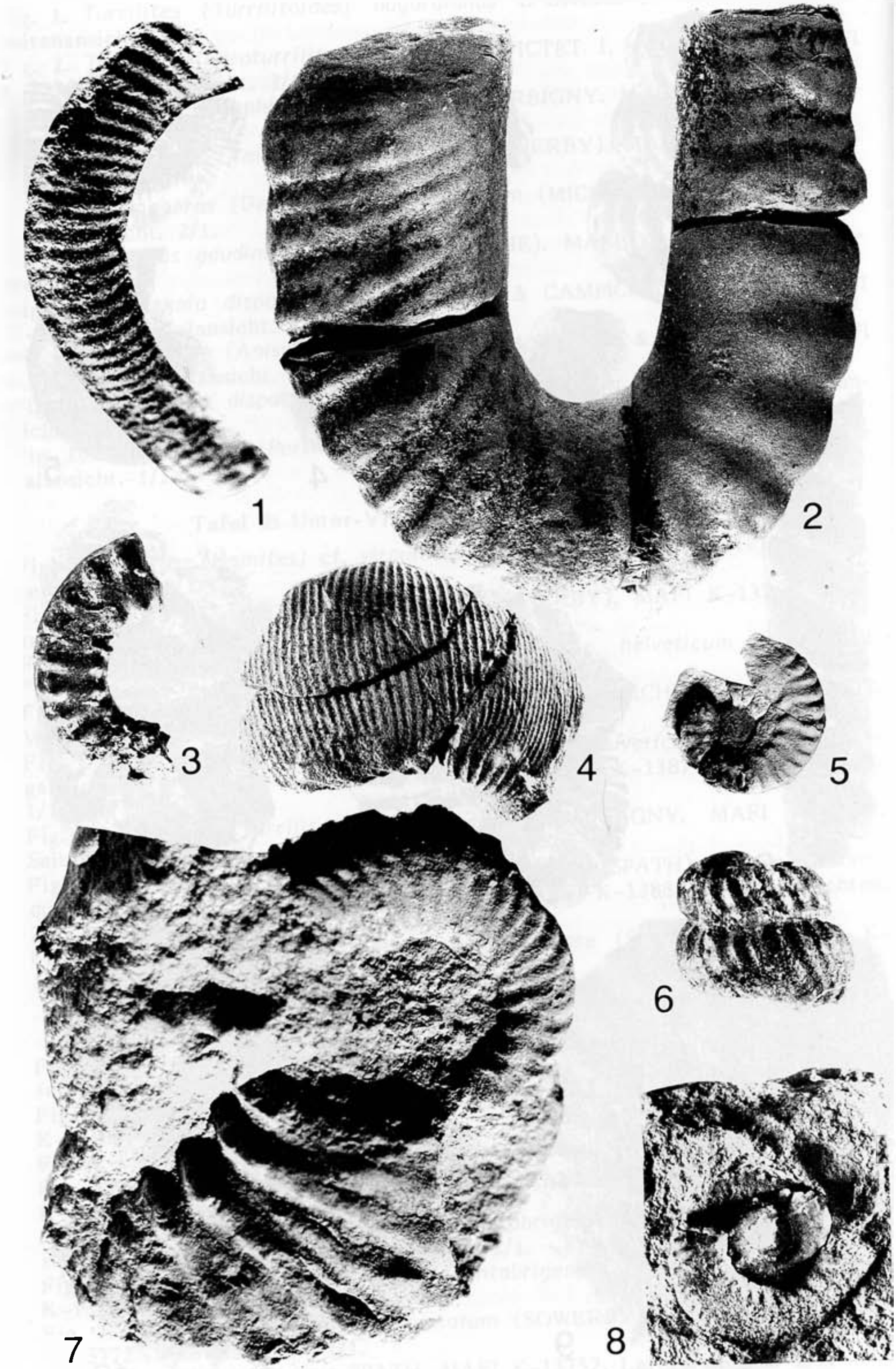
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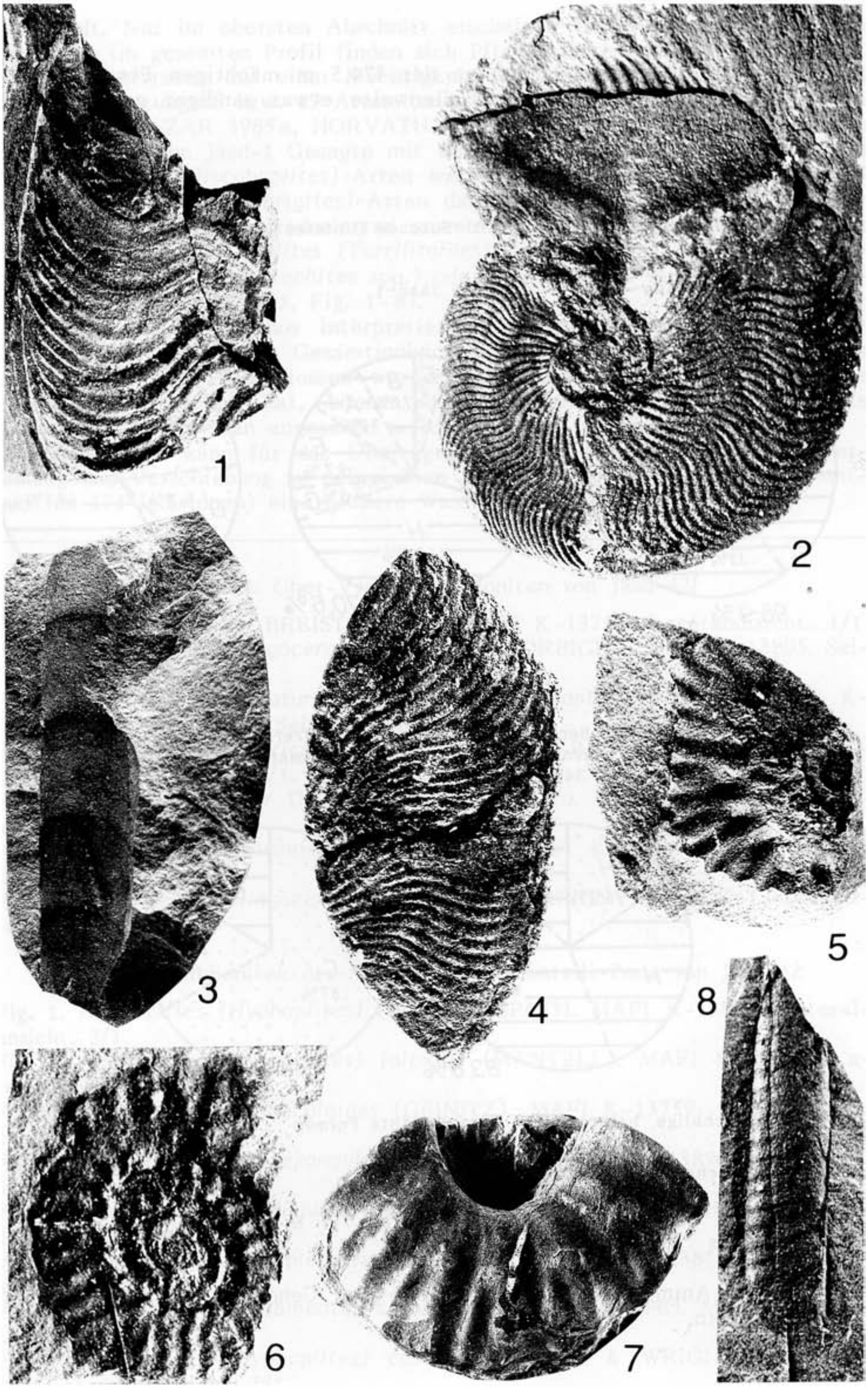
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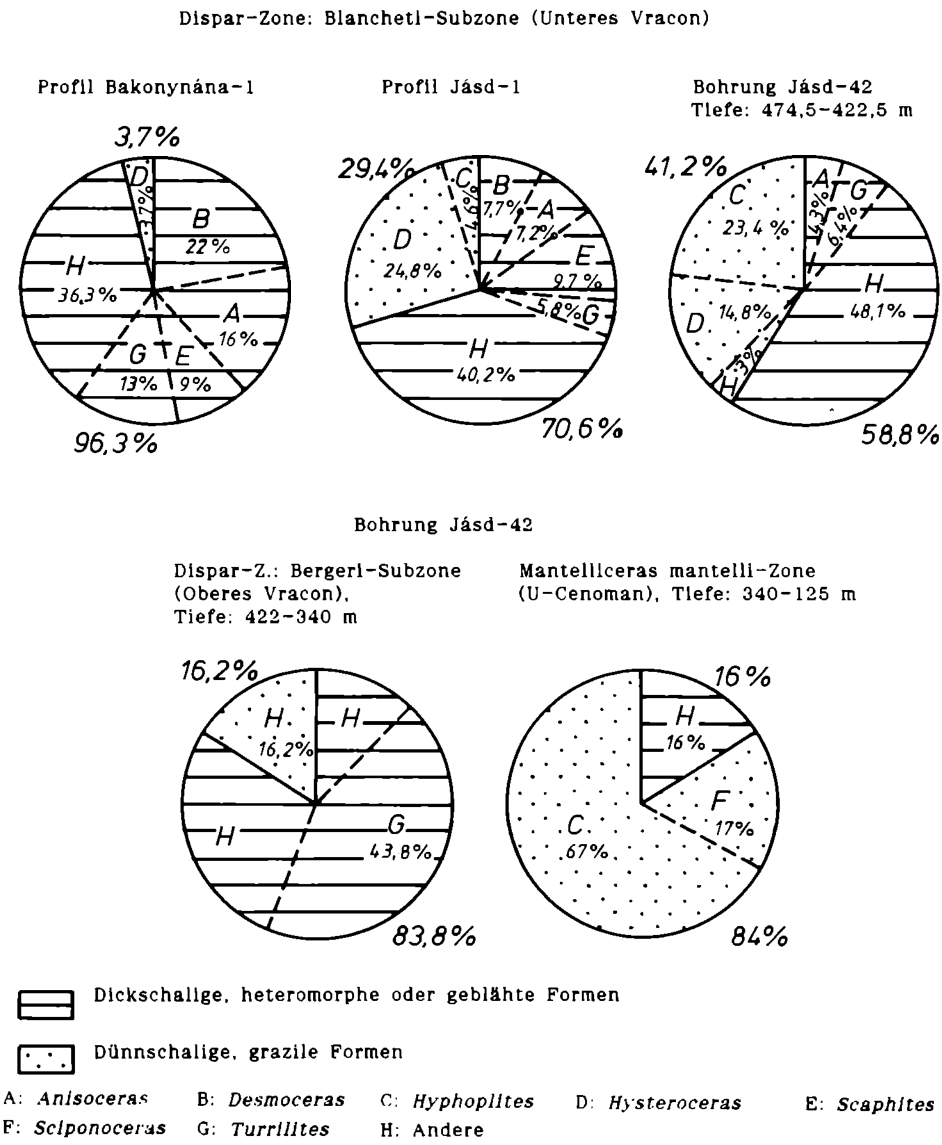
2.3 Bohrung Jásd-42

Die ammonitenführende Schichtfolge der 474,5 m mächtigen Pénzeskut-Formation ist hier als homogener, stellenweise etwas sandiger grauer Mergel



Text-Fig. 2. Ammonitenspektren des Alb und Cenoman der Pénzeskut-Formation, Ungarn.

entwickelt. Nur im obersten Abschnitt erscheinen feiner- und gröbersandige Schichten. Im gesamten Profil finden sich Pflanzenreste und Ichnofaunen.

Das Untervracon mit einer Mächtigkeit von 52 m enthält 12 Gattungen (und Untergattungen) bzw. 19 Arten (und Unterarten) mit insgesamt 47 Exemplaren (CSASZAR 1985a, HORVATH 1985). Für sie gilt das bereits beim Übertageprofil von Jásd-1 Gesagte mit der Ergänzung, daß die Individuenzahl der *Hyphoplites (Discohoplites)*-Arten weiter zugenommen, die Individuenzahl der *Hysteroceras (Cantabrigites)*-Arten dagegen abgenommen hat. Die schon im Übertageprofil stark reduzierten vermutlichen Benthonten sind entweder ganz verschwunden (*Turrilites (Turrilitoides) hugardianus, Desmoceras (Desmoceras) latidorsatum, Scaphites* spp.) oder nur noch in verminderter Individuenzahl vorhanden (Taf. 3, Fig. 1-8).

Die als vagiles Benthos interpretierten Formen stellen im Gebiet von Bakonynána-1 96,3 % der Gesamtindividuenzahl (562) dar (Text-Fig. 2). Da aus den Sedimenten geschlossen werden kann, daß hier auch die geringste Wassertiefe bestanden hat, können die Ammonitenspektren als wertvolle Bathymetrie-Indikatoren angesehen werden.

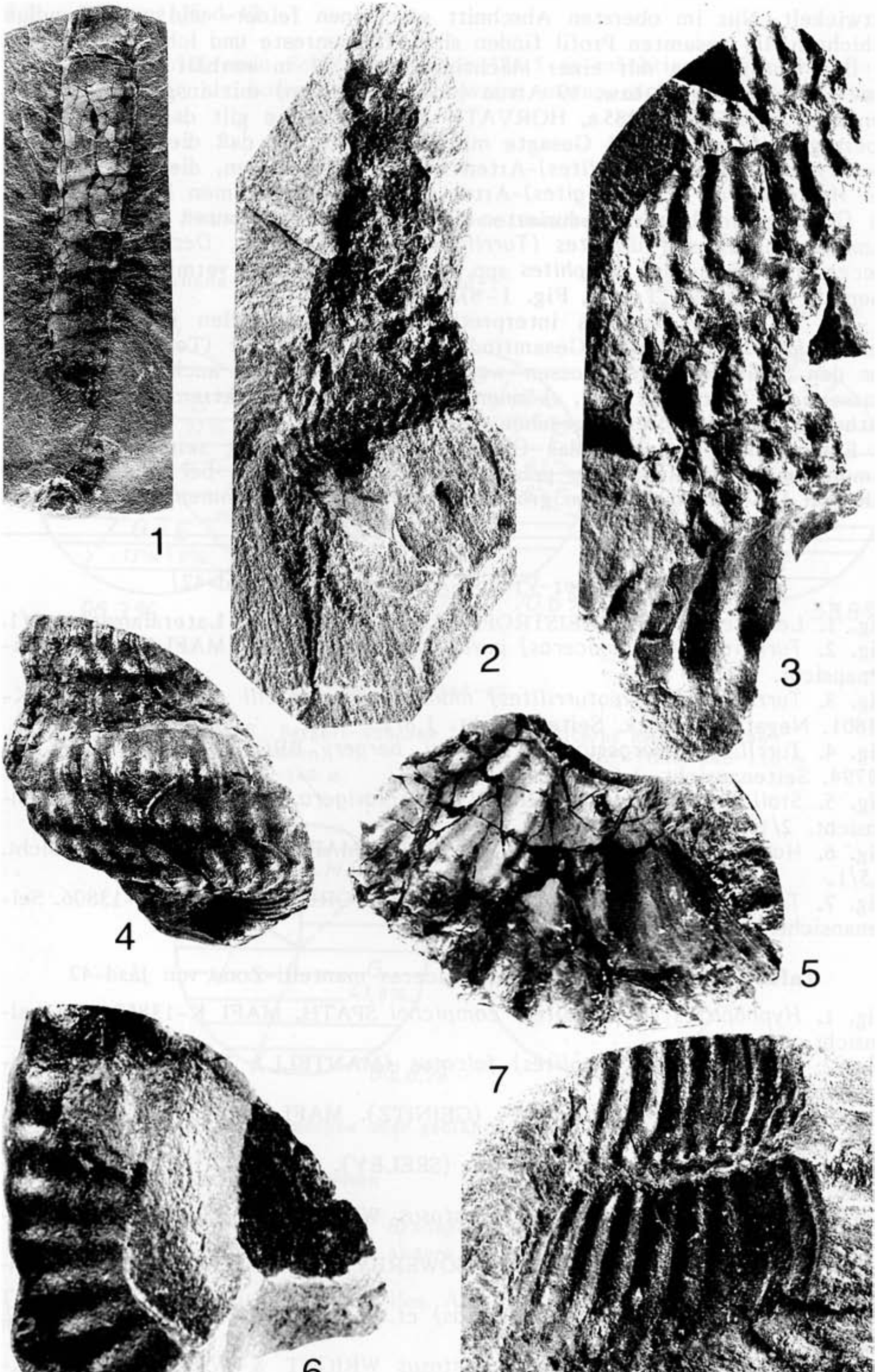
Entsprechend kann für das Übertageprofil Jásd-1 mit seiner Faunenverarmung und Verschiebung zu pelagischen Formen (29,4 % bei einer Gesamtzahl von 173 Individuen) eine größere Wassertiefe angenommen werden.

Tafel 4: Ober-Vracon-Ammoniten von Jásd-42

Fig. 1. *Lechites moreti* BREISTROFFER. MAFI K-13755. Lateralansicht. 1/1.
Fig. 2. *Turrilites (Ostlingoceras) puzosianus* D'ORBIGNY. MAFI K-13805. Seitenansicht. 2/1.
Fig. 3. *Turrilites (Eohypoturrilites) mantelli submantelli* SCHOLZ. MAFI K-13801. Negativ-Abdruck. Seitenansicht. 1,8/1.
Fig. 4. *Turrilites (Bergericeras) bergeri bergeri* BRONGNIART. MAFI K-13794. Seitenansicht. 1,5/1.
Fig. 5. *Stoliczkaia dispar* D'ORBIGNY f. *clavigera*. MAFI K-13899. Seitenansicht. 2/1.
Fig. 6. *Hamites* aff. *intermedius* SOWERBY. MAFI K-13744. Lateralansicht. 1,5/1.
Fig. 7. *Turrilites (Ostlingoceras) puzosianus* D'ORBIGNY. MAFI K-13806. Seitenansicht. 2/1.

Tafel 5: Ammoniten der Mantelliceras mantelli-Zone von Jásd-42

Fig. 1. *Hyphoplites (Hyphoplites) campichei* SPATH. MAFI K-13857. Lateralansicht. 2/1.
Fig. 2. *Hyphoplites (Hyphoplites) falcatus* (MANTELL). MAFI K-13838. Lateralansicht. 2/1.
Fig. 3. *Sciponoceras subbaculoides* (GEINITZ). MAFI K-13759. Seitenansicht. 1/1.
Fig. 4. *Neophlycticeras sexangulatus* (SEELEY). MAFI K-13896. Lateralansicht. 2/1.
Fig. 5. *Hyphoplites (Hyphoplites) costosus* WRIGHT & WRIGHT. MAFI K-13851. Lateralansicht. 2/1.
Fig. 6. *Anahoplites* cf. *splendens* (SOWERBY). MAFI K-13832. Lateralansicht. 2/1.
Fig. 7. *Mantelliceras (Submantelliceras)* cf. *saxbii* (SHARPE). MAFI K-13912. Lateralansicht. 1/1.
Fig. 8. *Hyphoplites (Hyphoplites) costosus* WRIGHT & WRIGHT. MAFI K-13847. Lateralansicht. 1/1.
Fig. 9. *Hyphoplites (Discohoplites) coelonatus transitorius* SPATH. MAFI K-13837. Lateralansicht. 2/1.



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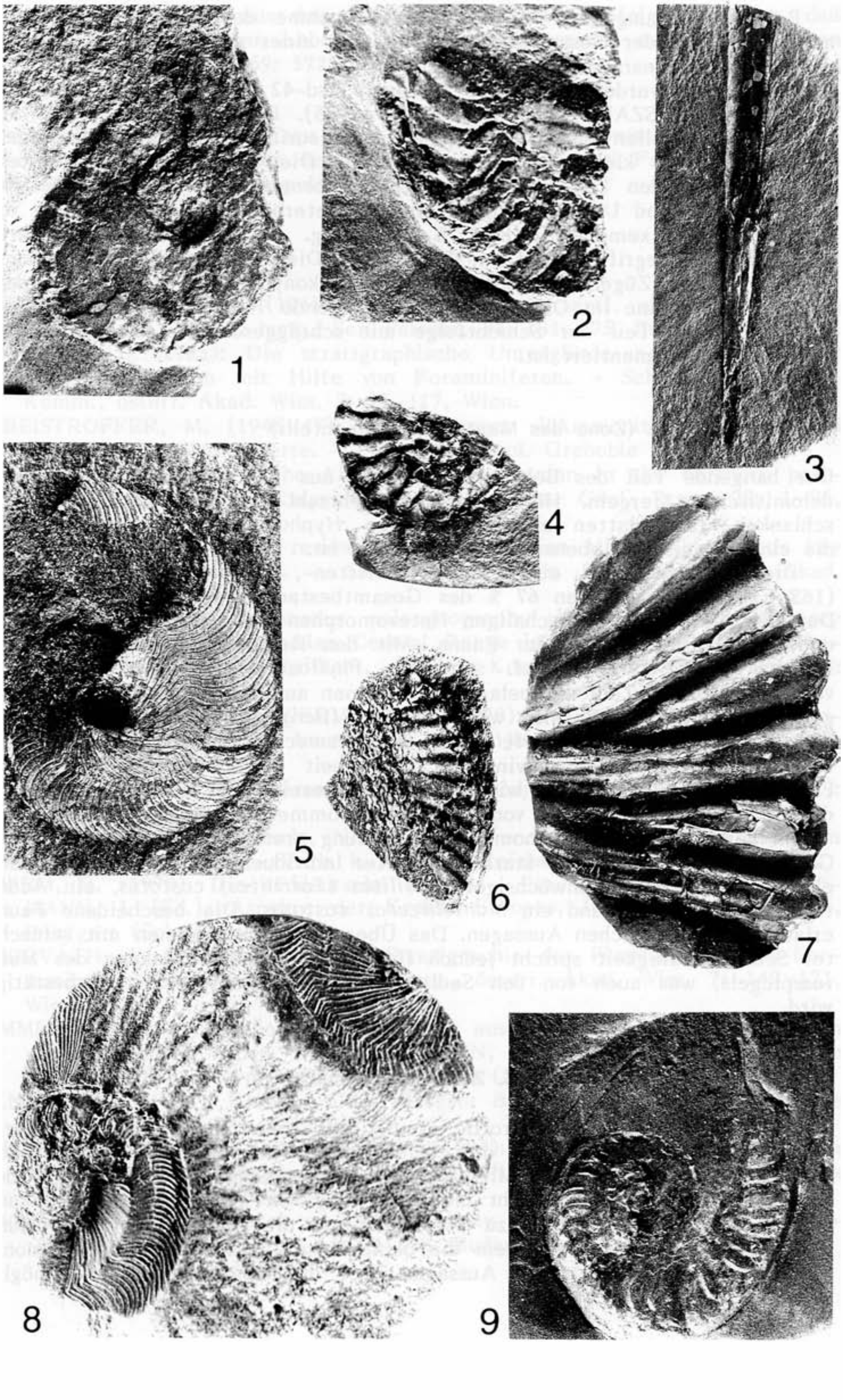
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Bei der Bohrung Jásd-42 dürfte die Zunahme der pelagischen Formen mit nun 41,2 % der Gesamtzahl (Text-Fig. 2) einer weiteren Tiefenzunahme des Sedimentationsraumes entsprechen.

Obervracon wurde nur in der Bohrung Jásd-42 (Teufe 422,5 - 340,0 m) untersucht (CSASZAR 1985a, HORVATH 1985). Das Sediment besteht aus einer grauen, stellenweise dolomitischen und aleurithaltigen Kalk/Kalkmergelschichtfolge mit kleinen Mengen Glaukonit. Die Ammonitenfauna besteht zum überwiegenden Teil aus heteromorphen Formen. Von den 19 Arten der 16 Gattungen (und Untergattungen) sind 12 heteromorphe Arten mit 45 von insgesamt 54 Exemplaren (83,8 %, Text-Fig. 2), andere ornamentierte Ammoniten einbegriffen (Taf. 4, Fig. 1-7). Diese Vergesellschaftung kann damit in großen Zügen mit der des Profils Bakonynána-1 verglichen werden. Dies wird auf eine im Oberen Vracon erfolgende Regression zurückgeführt, die im höheren Teil der Schichtfolge mit schräggeschichteten sublittoralen Sandsteinen dokumentiert ist.

Unteres Cenoman (Zone des Mantelliceras mantelli)

Der hangende Teil des Bohrprofils besteht aus 215 m mächtigen, grauen dolomitischen Mergeln. Hier wird die Mehrzahl der Ammonitenfauna von schlanken, fast glatten bis feinberippten *Hyphoplites*-Arten gebildet, für die eine pelagische Lebensweise anzunehmen ist.

Diese Formen, mit einer niedrigen Arten-, aber hohen Individuenzahl (163 Exemplare), machen 67 % des Gesamtbestandes der Fauna (241) aus. Daneben bilden die glattschaligen Heteromorphen *Sciponoceras subbaculoides* und *S. baculoides* 15 % der Fauna. Mit den Reliktformen *Neophlycticeras sexangulatus, Anahoplites* cf. *splendens, Phylloceras (Hypophylloceras)* cf. *velledae* wird die Zahl der pelagischen Formen auf 84 % erhöht. Die weniger gut schwimmfähigen Formen wie *Turrilites (Bergericeras) bergeri quadrituberculatus, Idiohamites, Mantelliceras* und *Desmoceras* stellen mit einem Anteil von 16 % eine verschwindende Minderheit dar (Text-Fig. 2, Taf. 5, Fig. 1-9). Als Wassertiefe wird für diesen Bereich des Bakony-Schelfs für diesen Zeitraum ein Betrag von 200 m angenommen.

Im vermuteten Mittelcenoman der Bohrung treten neben Vertretern der Gattung *Sciponoceras* (mit stark reduzierter Individuenzahl) an neuen Faunenelementen auf: ein kleinwüchsiger *Turrilites (Turrilites) costatus*, ein *Acanthoceras*-Bruchstück und ein *Mantelliceras costatus*. Die bescheidene Fauna erlaubt keine statischen Aussagen. Das Überwiegen von Formen mit schlechter Schwimmfähigkeit spricht jedoch für eine erneute Absenkung des Meeresspiegels, was auch von den Sedimenten dieses Bohrabschnittes bestätigt wird.

3. Zusammenfassung

Die drei hier untersuchten Profile ermöglichen eine Gliederung des ungarischen Untervracon-Meeres in drei verschiedene Fazieszonen. Auch die dem Zeitraum Untervracon bis Mittelcenoman entsprechende Schichtfolge der Tiefbohrung Jásd-42 ermöglicht die Ermittlung von Meeresspiegelschwankungen in diesem Zeitraum. Hierzu wird nicht nur die sedimentäre Entwicklung herangezogen, sondern vor allem die ökologische Auswertung von Ammonitenspektren, die damit auch Aussagen über die Paläobathymetrie ermöglichen. Außerdem kann aus den erzielten Ergebnissen gefolgert werden, daß sich bei den Ammoniten Lebens- und Einbettungsraum entsprochen haben dürften (WIEDMANN 1969: 572).

Dank. Die Autorin dankt Frau Dr. I. BODROGI für die von ihr durchgeführten Aufsammlungen in den beiden Übertageprofilen und Frau Dr. L. PELLERDY für die Anfertigung der Photographien.

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An Outline of the Cretaceous of Albania

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With 9 Text-Figures

PEZA, L. H. (1989): An Outline of the Cretaceous of Albania. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 483-504. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Cretaceous deposits in Albania have a relatively wide distribution and are developed in neritic, pelagic and flysch facies. In this paper typical Cretaceous sections of the different tectonic zones are described.

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1. Introduction

Neritic facies are encountered in the zones of Sazan, Kruja, Albanian Alps and Mirdita (Text-Fig. 1), where limestones are predominant, whereas conglomeratic limestones and clastic formations are less prominent. Urgonian facies has a wide distribution in the zone of the Albanian Alps and that of Mirdita.

Pelagic facies are met in the Ionian and Krasta-Cukali zones and in the Valbona subzone, where platy limestones with chert lenses and layers, marly limestones and marls are predominant.

Flysch facies are met in the zones of the Albanian Alps, Kelmendi, Krasta-Cukali, Mirdita and Korabi.

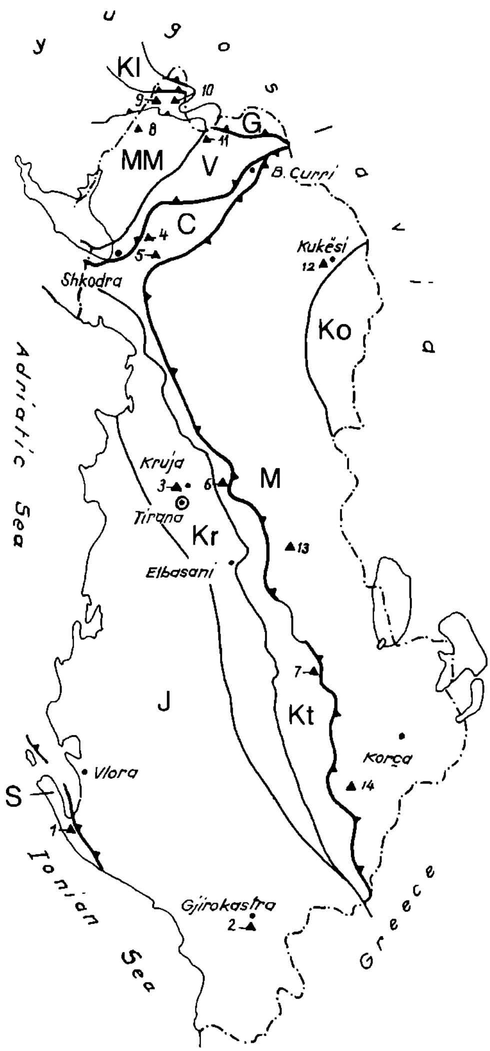
During the Cretaceous different tectonic phases repeatedly caused subaerial exposure of formations, accompanied with stratigraphic breaks, which are quite evident in the Mirdita Zone. The Upper Jurassic-Cretaceous boundary is gradual everywhere. The boundary between Cretaceous and Paleogene appears as a stratigraphic continuity in general, but in some areas stratigraphic gaps are evident.

Cretaceous deposits in Albania have a relatively wide geographic distribution and are encountered almost in all tectonic zones, except for the zone of Gashi where only Triassic is present. The formation of the Cretaceous deposits is closely linked with the evolution of the Mediterranean. As a consequence, they bear a great resemblance to the Cretaceous sediments of other Mediterranean countries and especially to the neighbouring countries Yugoslavia and Greece.

The Cretaceous is mainly built up of carbonate rocks. These are developed in neritic facies (in some cases reefs with rudists) in the zones of Sazan, Kruja, in the subzone of Malësia Madhe (Zone of the Albian Alps), and in the Mirdita Zone. The facies of the pelagic platy limestones with lens- and chert-stratification is developed in the Ionian Zone, in the subzone of

Valbona (Zone of the Albanian Alps), in some regions of the Mirdita Zone and in that of Korabi. Flysch facies, though less widespread in the Albanides, are encountered in the Zone of Krasta-Cukali, the subzone of Valbona, Kelmendi Zone, in some regions of Mirdita Zone and in that of Korabi.

Different phases of movement have played a very important role in the formation of Cretaceous deposits. These movements caused in some zones great changes in the sedimentary regime and sea level. This phenomenon is typical especially of inner zones (mainly in the Zone of Mirdita) and the subzone of Valbona.



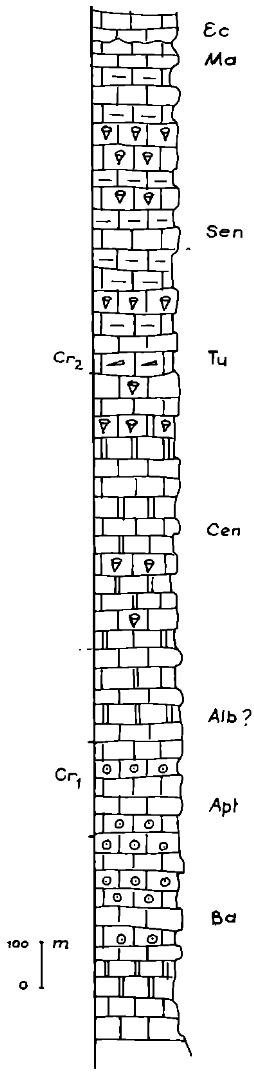
Text-Fig. 1. Tectonic sketch map of Albania with the locality sections of Cretaceous deposits.
Tectonic zones: External Albanides: S - The zone of Sazani; J - Ionian Zone; Kr - Kruja Zone; C, Kt - Krasta-Cukali Zone (C - Cukali subzone, Kt - Krasta subzone); MM, V - The zone of the Albanian Alps (MM - the subzone of "Malësia Madhe", V - the subzone of Valbona); Kl - Kelmendi; G - Gashi; M - Mirdita; Ko - Korabi.
Sections: 1 - Karaburun; 2 - Mali Gjërë; 3 - Makaresh; 4 - Ura Shtrejtë; 5 - Kroni madh; 6 - Derja; 7 - Kosore; 8 - Selcë; 9 - Jeshnicë; 10 - Budaca; 11 - Kollata; 12 - Kukës; 13 - Polis; 14 - Novosela.

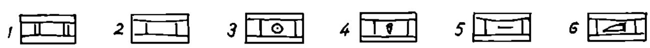
2. Regional descriptions

We first present Cretaceous deposits from the external zones of the Albanides, e. g. the Sazani, Ionian, Krasta-Cukali zones (with both subzones: Cukali and Krasta), Albanian Alps (with both subzones: Malësia Madhe and Valbona), and Kelmendi, then we proceed with those of the internal zones: Mirdita and Korabi (Text-Fig. 1).

2.1 The external Albanides

2.1.1 Sazani Zone





Text-Fig. 2. Section of Kanali Mountain. 1 - Dolomites; 2 - Micritic limestones; 3 - Oolitic limestones; 4 - Limestones with rudists; 5 - Clayey limestones; 6 - Limestones with *Chondrodonta*.
Ba - Barremian; Ap - Aptian; Alb - Albian; Cen - Cenomanian; Tu - Turonian; Sen - Senonian; Ma - Maastrichtian; Ec - Eocene.

This zone is an equivalent to the Apulia Zone in Italy and Paksos Zone in Greece, covering the western part of Albania and the isle of Sazan. It is built up of neritic carbonate deposits of Lower and Upper Cretaceous age upon which Upper Eocene limestones lie discordantly.

The lower part of the Lower Cretaceous deposits does not crop out at the surface, but is encountered in deep oil wells. Upper Jurassic is represented by limestones with *Clypeina jurassica* while Neocomian is made up of limestones with *Campbelliella striata* and *Salpingoporella annulata*. Upsection there are surface outcrops showing (Text-Fig. 2):

- 450 m of oolitic limestones alternating with micritic limestones; *Bacinella irregularis, Salpingoporella dinarica, Marginula torosa,* and *M. nuda* are encountered.
- 230 m of alternating biogenic limestones with dolomites in which *Quinqueloculina* sp., *Triloculina* sp., and *Textularia* sp. (DALIPI et al. 1966) are found.
- 450 m of oolitic limestones alternating with micritic limestones rich in: *Bacellina irregularis, Salpingoporella dinarica, Marginula torosa, M. nuda, Stensiöina* sp., and *Spiroloculina* sp.
- 200 m of alternating micritic and dolomitized limestones which often pass into dolomites. The following organisms were encountered in the limestones: *Salpingoporella dinarica, Iraqia hensoni, I. valentina, I. minima, Nezzazata simplex* and others.

All these levels belong to the Barremian-Aptian which, in the upper part, may pass into the Albian.

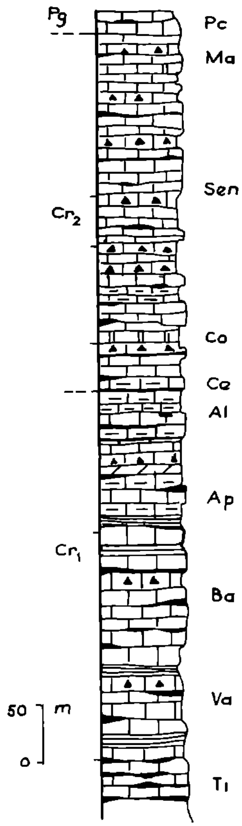
- 600 m of an alternation of dolomites and micritic limestones in which are encountered: *Nomuloculina heimi, Aeolissacus kotori, Thaumatoporella parvovesiculifera, Cuneolina* sp., *Praealveolina* sp., *Iraqia* sp., *Nezzazata* sp., and rudists. This sequence represents the Cenomanian.
- 10-15 m of white micritic marly limestones with *Chondrodonta, Vaccinites inferus* and *Plesioplocus futtereri*, which prove a Lower Turonian age (PEZA 1987).
- 50-70 m of white marly limestones very rich in *Plesioptygmatis* shells (so-called *Plesioptygmatis* horizon) among which are found: *P. requieni, P. bassani, P. armenica, P. myslimi, P. albanica, P. hasani,* confirming the Turonian (PEZA 1970, 1987).
- 700 m of an alternation of micritic and white clayey limestones with rudist biostromes, in which occur *Biradiolites stopani, Durania arnaudi, Sauvagesia sauvagesi, Radiolites mamillaris* and *Batolites organisans* confirming the Senonian (DALIPI et al. 1966).

The upper part of the Cretaceous deposits consist of micritic limestones containing *Orbitoides* sp., *Stomiosphaera sphaerica, Globotruncana* sp. and others (Maastrichtian).

Above a sedimentary gap caused by transgression follow discordantly biogene limestones with *Nummulites aturicus, Discocyclina scalaris, D. sella, Asterodiscus cuvillieri, Pellatispira* sp. and rare *Globigerina*, which confirm the Upper Lutetian-Priabonian (Gjeologjia e Shqiperise 1982).

2.1.2 Ionian Zone

Cretaceous deposits have a wide extension in the southern part of this zone, ranging from Vlora city up to the Greek borders. The Lower Creta-



Text-Fig. 3. Section of "Mali Gjërë". 1 - Pelagic limestones with chert lenses; 2 - Turbiditic limestones; 3 - Phosphatic limestones; 4 - marl-shales; 5 - marl-limestones.
Ti - Tithonian; Va - Valanginian; Al - Albian; Co - Coniacian; Pc - Paleocene; other abbreviations see Text-Fig. 2.

ceous is, in its lower part, represented by white porcellaneous limestones, rarely alternating with turbiditic limestones and thin strata of marly shales. Limestones, which partly belong to the Upper Jurassic, contain lens- and chert-stratification and generally reach up to 150-475 m in thickness. Further upsection limestones and mainly marls follow which alternate with clay-shales; especially in the lower part, turbiditic limestones and chert-stratification, reaching a general thickness that varies from 160 m up to 250 m.

Within these Lower Cretaceous deposits of the "Mali Gjërë" section (near the city of Gjirokastra; Text-Fig. 3) the following levels are distinguished (Gjeologjia e Shqipërisë 1982):

- the level with *Calpionella alpina, C. elliptica, Tintinnopsella carpathica, T. cadischiana, T. longa, Calpionellites darderi* (Upper Tithonian-Valanginian).
- the level with *Crioceratites duvali, Macroscaphites yvani, "Globigerina" hauterivica, "G." infracretacea, Hedbergella trochoidea* (Barremian-Aptian).

- the level with *Ticinella roberti*, which is often encountered in Albian deposits.

The Upper Cretaceous deposits in the lower part of the section consist of marly and turbiditic limestones. The limestones are about 60 m thick and contain lens- and chert-stratification. The upper part of these deposits is made up of micritic limestones alternating with turbiditic limestones; marly limestones are rare. These, in turn, rarely contain lens- and chert-stratification and reach a general thickness ranging from 100 m up to 250 m.

The Upper Cretaceous deposits in the eastern Ionian Zone are characterized by Coniacian phosphatic limestones, which reach an industrial concentration level in the region north of the city of Gjirokastra. In the western part of the zone they are replaced by rose-coloured micritic limestones.

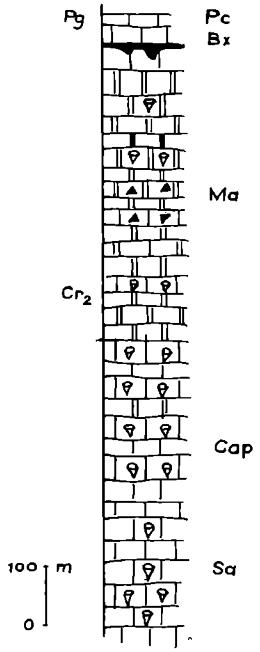
In the Upper Cretaceous limestones which are 165-250 m thick, the following levels are distinguished:

- the level with *Rotalipora appenninica, R. greenhornensis, Globotruncana helvetica, G. lapparenti, G. sigali, Praeglobotruncana stephani* and others, dating the Cenomanian passing into the Turonian (KONDO 1966).
- the level of phosphatic limestones containing *Globotruncana imbricata, G. lapparenti, G. coronata, G.* cf. *schneegansi, G. sigali, G.* cf. *fornicata;* an association which dates the Coniacian (SERJANI & YLLI 1984).
- the level with *Globotruncana stuarti, G. arca, G. conica, G. contusa, G. calcarata, G. bulloides,* and others which belong to the Upper Senonian.
- the level with *Orbitoides media, Lepidorbitoides socialis, Siderolites calcitrapoides, Globotruncana contusa,* and *Omphalocyclus macroporus* dating the Maastrichtian.
- the level with *Globigerina daubjergensis, G. compressa* var. *bulloides, G. linaperta, Globorotalia velascoensis, Gl. trinidadensis, Discocyclina seunesi,* and redeposited *Globotruncana* sp. and *Siderolites* sp. This association testifies to the existence of Paleocene.

It must be stressed, that in some regions stratigraphic gaps are marked within the Cretaceous of the Ionian Zone. Thus, in the southern part of the zone (Bogaz), Tithonian - Neocomian limestones rest transgressively upon Liassic limestones (XHOMO et al. 1968). In the sector between the cities of Vlora and Gjirokastra, Eocene deposits overlie those of the Barremian - Aptian and the latter overlie those of the Middle Jurassic. The same occurs in the region of the Cika Unit (near Vlora), where Albian deposits overlie those of Tithonian - Neocomian age and these, in turn, overlie those of the Upper Jurassic. In Pilur (near the Himara), though, Eocene deposits transgressively overlie those of the Senonian age (Gjeologjia e Shqipërisë 1982).

2.1.3 Kruja Zone

This zone is the equivalent of the Dalmatian Zone in the Dinarides and of the Gavrova Zone in the Hellenides. The Upper Cretaceous deposits are the only ones that crop out at the surface and stretch up to the Greek border as a thin belt. At the base they are mainly made up of thick limestones with rudist biostromes and alternate rarely with dolomitic strata. They are 400-600 m thick and rich in rudists and echinoids. Thick dolomites with rare limestones continue further upsection.



Text-Fig. 4. Kruja section. 1 - Limestones with rudists; 2 - Dolomites; 3 - Breccia-limestones; 4 - Bauxites. Sa - Santonian; Cap - Campanian; other abbreviations see Text-Figs. 2 and 3.

In the Kruja section (Text-Fig. 4) the following stratigraphic succession is found (PEZA 1967, 1977, 1982):

- 500 m of grey limestones, often with rudistid biostromes which rarely alternate with dolomite strata. In these limestones are encountered: *Biradiolites acuticostatus*, *Radiolites* sp., *Hippurites cornuvaccinum*, *Hippuritella* sp., *Durania* sp., *Inoceramus* ex gr. *salisburgensis*, echinoids as well as *Accordiella conica*, *Dicyclina schlumbergeri*, *Cuneolina pavonica*, *Thaumatoporella parvovesiculifera*, *Aeolissacus kotori*, the occurrence of which indicates the presence of Santonian-Campanian.
- 600-700 m of thick grey dolomites occasionally alternating with rudist-bearing limestones and breccia-limestones. In the limestones of this interval two biostratigraphic levels are distinguished:
  (a) level with *Lepidorbitoides socialis*, *L. minor*, *Orbitoides media*, *Clypeorbis* cf. *mamillata* and
  (b) level with *Rhapydionina liburnica*, *Cuneolina pavonia*, *Dicyclina* sp., *Microcodium* sp.
  Both these levels belong to the Maastrichtian and the second can be assigned to the Upper Maastrichtian.

After a short interval in non-deposition, which is accompanied by bauxites, follow limestones and then flysch deposits. In the limestones are found: *Miscellanea* cf. *miscella*, "*Rotalina*" *cayeuxi*, *Quinqueloculina*, *Triloculina*, which date the Paleocene, while the flysch deposits further upsection contain nummulites and belong to the Eocene (PEZA 1967).

Apart from this section, which is characteristic for the entire Kruja Zone, the situation in the Tomorri Mountain (southern Albania) is different.

- The lower part of the section consists of massive dolomites and breccia-limestones with rudists: *Biradiolites samniticus, B.* cf. *lumbricoides, Vaccinites* ex gr. *taburni, Bournonia* cf. *bournoni* which belong to the Santonian-Campanian.
- The upper part of the section is made up of turbiditic limestones, which contain *Globotruncana contusa, G. arca, G. lapparenti* (Maastrichtian).

Above these limestones follow Paleocene deposits without bauxites and without a stratigraphic gap.

2.1.4 The Krasta-Cukali Zone

This zone is the analogue to the Budva Zone in the Dinarides and to the Pindi Zone in the Hellenides. Cretaceous deposits consist of pelagic limestones and flysch facies and are rarely distributed throughout the territory of its extension. These deposits show different features in the two subzones; hence we shall consider them separately.

2.1.4.1 Cukali subzone

The Cretaceous is built up by rather condensed limestones and chert-lenses.

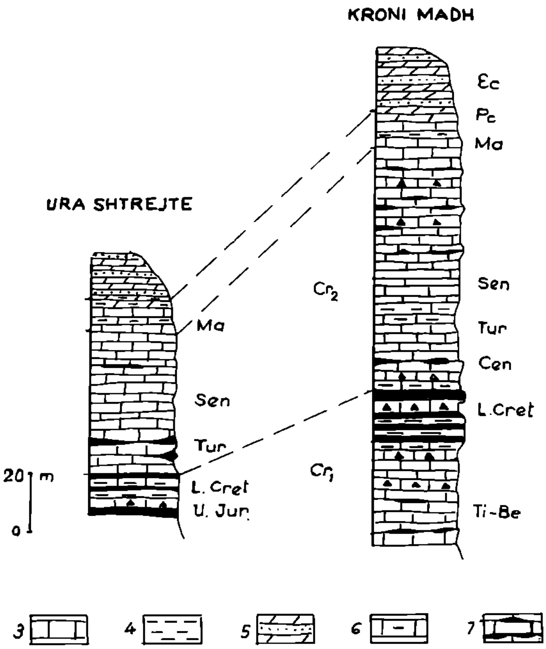
In the section of "Ura Shtrejtë" (Text-Fig. 5), the Lower Cretaceous is part of the lowest portion of the radiolaritic formation. This consists of red radiolaritic strata alternating with thin argillaceous chert layers and rare platy breccia-limestones which are supposed to belong to the Upper Jurassic - Lower Cretaceous. Continuing upwards, there are grey and beige platy micritic limestones rich in *Globotruncana* among which the following levels can be separated (XHOMO et al. 1975, THEODHORI et al. 1978):

- 15 m of limestones with *Rotalipora appenninica, R. cushmani, Praeglobotruncana stephani, Globotruncana helvetica,* dating the Cenomanian-Turonian.
- 15 m of limestones with *Globotruncana marginata, G. fornicata, G. tricarinata, G. lapparenti,* indicating the Lower Senonian.
- 16 m of limestones with *Globotruncana caliciformis, G.* cf. *citae, G. arca, G. stuarti, G. conica, Heterohelix globulosa,* dating the Campanian-Maastrichtian.
- 13 m of rose platy limestones alternating with thin marly strata (the so-called transitory strata towards flysch) containing *Globotruncana contusa, G.* cf. *citae* which date the Maastrichtian.

Above them follow limestones with *Globorotalia velascoensis, Globigerina pseudobulloides* and others, which indicate the Danian. Further up, this succession passes on to the Xhani Flysch (NOPCSA 1929) which consists of marls and sands, which in some regions contain an Eocene fauna with nummulites (SHEHU et al. 1966).

In the upper part of the "Kroni Madh" (Text-Fig. 5), radiolarites alternate with platy micritic limestones. The faunal content encountered is as follows: *Crassicollaria intermedia, Calpionella alpina, C. elliptica, Tintinnopsella carpathica* and *T. cadischiana,* which date the Tithonian-Berriasian.

Higher up in the section beige platy limestones with *Globotruncana* succeed. The following levels are distinguishable:



Text-Fig. 5. The sections of the Cukali subzone. 1 - Radiolarites with argillaceous chert-shales; 2 - Breccia-limestones; 3 - Platy limestones; 4 - Argillaceous shales; 5 - Xhani Flysch (marls and sands); 6 - Marly limestones; 7 - Limestones with chert-lenses.

- the level with *Praeglobotruncana stephani* and *Ticinella* sp. (Albian-Cenomanian);
- the level with *Praeglobotruncana stephani, Rotalipora appenninica, Globotruncana helvetica,* and *Globotruncana lapparenti,* which belong to the Cenomanian-Turonian;
- the level with G. *lapparenti,* G. *bulloides,* G. *tricarinata,* G. *fornicata;*
- the level with G. *lapparenti,* G. *coronata,* G. *tricarinata,* and G. *marginata.* Both these levels belong to the Turonian-Campanian;
- the level with G. *lapparenti,* G. *tricarinata,* G. *rosseta,* G. cf. *ventricosa,* G. cf. *arca,* and G. cf. *calcarata,* which still characterize the Senonian;
- the level with G. *lapparenti,* G. *tricarinata,* G. *bulloides,* G. *stuarti,* G. *conica,* which indicate the Upper Senonian.

The following transitional Maastrichtian-Paleocene strata are overlain by the Xhani Flysch just as in the "Ura Shtrejtë" and in other regions of the Cukali subzone (XHOMO et al. 1975).

It is worth stressing that in the region of Shllum of Merturi and Merturi mountain (between the towns of Puka and B. Curri), which constitute the eastern sector of the Cukali subzone, Cretaceous deposits have a peculiar setting. They lie transgressively upon limestones of Ladinian (Shllumi Merturit) and Upper Triassic age (Mertur) and consist of bioclastic and turbiditic limestones more than 35 m thick. These deposits belong to the Cenomanian-Senonian over which transitory strata continue towards the Xhani Flysch (Upper Maastrichtian) rich in *Globotruncana* (THEODHORI et al. 1978). Of course, it is probable that this sector lay adjacent to the Valbona subzone and has therefore been more influenced by its development.

In the western part of the Cukali subzone in Lisen and Spiten (near the towns of Lezha and Miloti) Cretaceous deposits begin in the upper part of radiolarites with limestones containing *Calpionella* of Tithonian-Neocomian age. These are succeeded by 25 m of grey and rose platy limestones in which we distinguish the levels:

- with *Hedbergella infracretacea* (Barremian-Aptian);
- with *Ticinella roberti* (Albian);
- with *Rotalipora* and *Praeglobotruncana* (Cenomanian-Turonian);
- with *Globotruncana lapparenti, G. calcarata, G. havanensis, G. elevata, G. stuarti* (Senonian-Maastrichtian).

These are overlain by transitory strata of the Upper Maastrichtian-Paleocene and by the Xhani Flysch (XHOMO 1966).

2.1.4.2 Krasta subzone (Text-Fig. 6)

Upper Cretaceous sediments are distributed throughout, but they are rather tectonized. The most complete sections are those of Kosore and Derja (LULA et al. 1981).

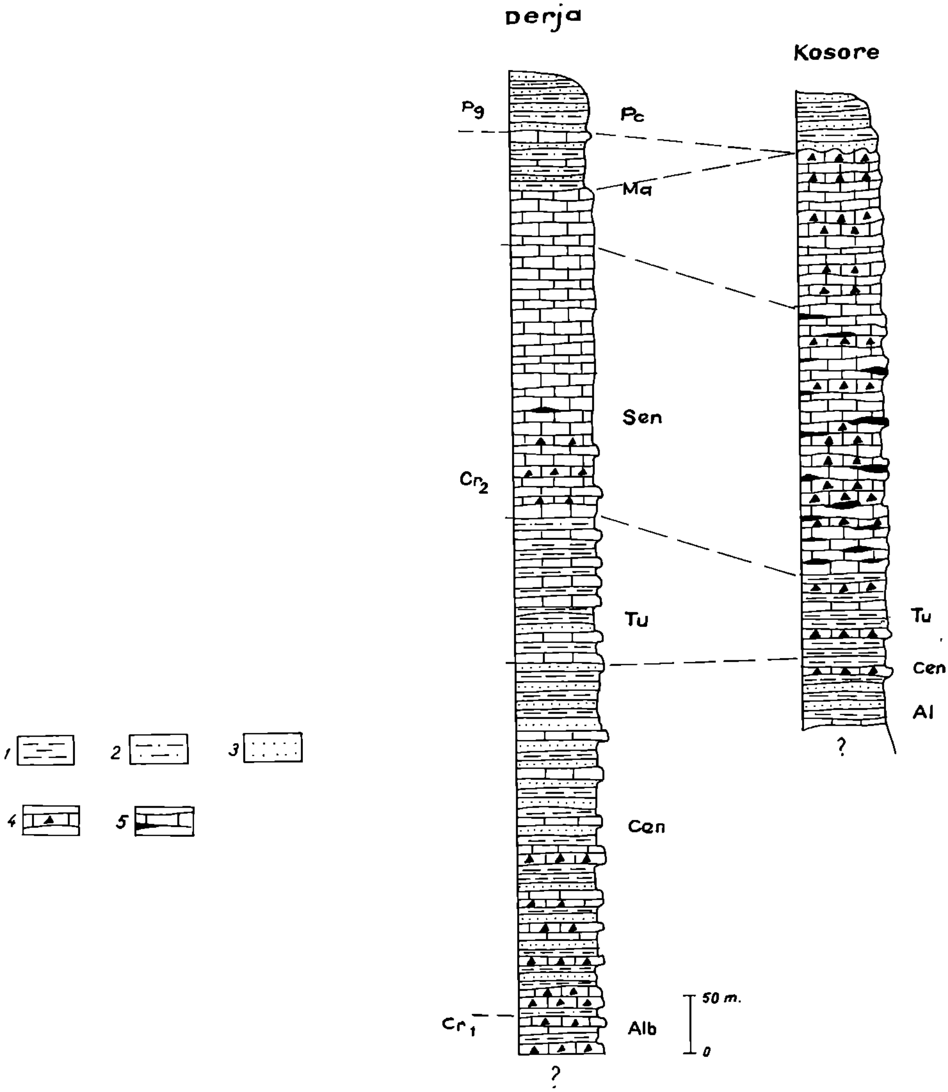
The lowest part is represented by flysch deposits consisting of thin alternations of clays, sands and some limestones, which are altogether up to 140-500 m thick. In these deposits rare fossils are encountered: *Ticinella roberti, Pithonella ovalis, P. sphaerica, Rotalipora appenninica, Praeglobotruncana stephani, P. reicheli, Orbitolina* sp., and radiolaria, which belong to the Albian-Cenomanian.

Platy limestones continue upsection alternating with thick breccia-limestones, chert-lenses and -stratifications. In these deposits are encountered: *Globotruncana lapparenti, G. concavata, G. fornicata, G. elevata, Rugoglobigerina rugosa, Guembelina* sp., *Orbitoides media, O. apiculata, Lepidorbitoides minor, L. socialis, Omphalocyclus macroporus, Siderolites calcitrapoides;* an association which characterizes the Senonian, the upper part belonging to the Maastrichtian. In the breccia-limestones fragments of rudists are abundant.

Transitory flysch deposits of about 50-60 m thickness follow and consist of the following levels:

- the lower level with *Globotruncana contusa, G. stuarti, G. mayaroensis, G. falsostuarti, G. gansseri,* which date the Upper Maastrichtian;
- and the upper level which contains *Globotruncana pseudobulloides* and *G. trinidadensis* which date Paleocene. Eocene Xhani Flysch continues further upsection.

Apart from these complete sections within the Krasta Zone, there are regions with stratigraphic gaps, e. g., North of Tirana (Cudhi etc.) a part of the Senonian deposits is absent and in this case those of Maastrichtian lie unconformably upon Albian-Cenomanian deposits. East of Tirana and in some regions south-east of Korca city transitory flysch strata of Maastrichtian-Paleocene age are absent and in these cases, Paleocene flysch deposits overlie unconformably Maastrichtian carbonate deposits (LULA et al. 1981).



Text-Fig. 6. The sections of the Krasta subzone. 1 - Argillites; 2 - Aleurolites; 3 - Sands; 4 - Turbiditic limestones; 5 - Platy limestones with chert-lenses.

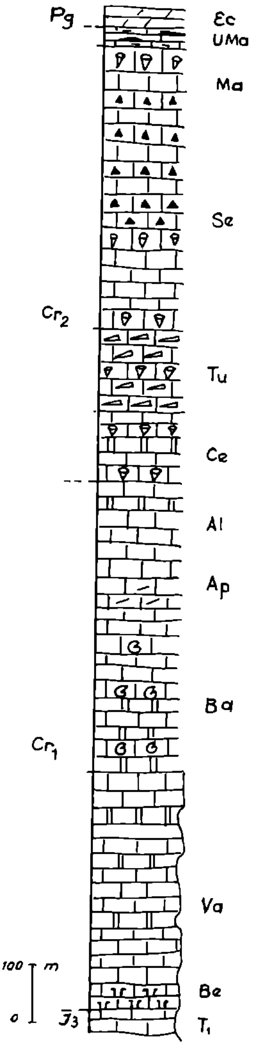
2.1.5 The Zone of Albanian Alps

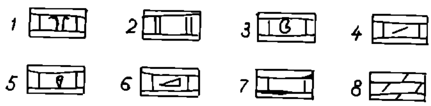
Cretaceous deposits are especially distributed in its northern part. They have, together with Jurassic deposits, enough facies differences which enable discrimination of the Valbona subzone (XHOMO et al. 1969). This subzone was also traced in Yugoslavia (BLANCHET 1970) and was separated from that of "Malësia Madhe" consisting of neritic Cretaceous sediments deposited in a relatively calm environment.

2.1.5.1 The subzone of Malësia Madhe

This subzone is the analogon of the Zone of High Karst in the Dinarides. Cretaceous sediments have a wide distribution in its northern part where all stages are met.

In the section of Selca (Text-Fig. 7) which is typical of this subzone, Cretaceous deposits overlie dark micritic limestones alternating with dolosparites very rich in *Clypeina jurassica, Thaumatoporella parvovesiculifera,*





Text-Fig. 7. Section of Selca. 1 - Limestones with *Campbelliella striata;* 2 - Dolomites; 3 - Limestones with requieniids; 4 - Limestone with *Salpingoporella dinarica;* 5 - Limestones with rudists; 6 - Limestones with *Chondrodonta;* 7 - Limestones with chert-lenses; 8 - Marls.

Salpingoporella annulata, Textularia, ostracodes, and nerineids such as *Itieria moreana, Nerinea jeanjani, N. tuberculosa, N. defrancei* var. *posthuma, Trochoptygmatis carpathica,* and *Ptygmatis pseudobruntrutana* which characterize the Tithonian (PEZA et al. 1973, PEZA 1981).

Within the overlying succession we distinguish:

- 20-75 m of micritic limestones and occasionally grey ruditic limestones which are less marly containing *Campbelliella striata, Favreina salevensis, Salpingoporella annulata, Clypeina jurassica, Thaumatoporella parvovesiculifera, Trocholina alpina,* Textulariidae, Trochominidae, Verneuillinidae, Codiacea and charophytes, which belong to the Berriasian (PEZA 1981);
- 95-395 m (according to different regions) of thick micritic and oncolitic limestones occasionally associated with dolomites, in which we encounter *Salpingoporella annulata, Thaumatoporella parvovesiculifera, Nautiloculina oolithica, Favreina salevensis, Cayeuxia* cf. *annee, Aeolissacus* sp., *Textulariella* sp., *Eggerella* sp., *Macroporella* sp., and charophytes, which indicate the Valanginian-Hauterivian;
- 200-270 m of thick micritic limestones sporadically alternate with dolomites in which we encounter *Salpingoporella dinarica, S. militae, S. cemi, Thaumatoporella parvovesiculifera, Favreina salevensis, Cuneolina* sp., *Palaeodictyoconus arabicus, Bacinella irregularis, Carpathoporella occidentalis, Requienia* sp., *Textulariella* sp., *Nerinea* sp., *Pseudolituonella* sp., *Aeolissacus* sp., *Nautiloculina oolithica* and others which belong to the Barremian-Aptian (PEZA 1983);
- 45-80 m of thick micritic limestones poor in fossils, in which we have come across only of *Cuneolina* sp., *Thaumatoporella parvovesiculifera,* miliolids and ostracodes, which in our opinion belong to the Albian;
- 50-185 m of dolosparites alternating with micritic limestones and passing into platy limestones showing *Cisalveolina fallax, Cuneolina pavonica, Biplanata peneropliformis, Pseudodomia vialli, Biconcava bentori, Nezzazata simplex, Dicyclina schlumbergeri, Nummofallotia apula, Trochospira avnimelechi, Pseudolituonella reicheli, Spiroloculina* sp., *Vidalina* sp., *Textulariella* sp., and many ostracodes and others which confirm the Cenomanian;
- 95-180 m of black micritic limestones rather rich in *Chondrodonta* shells containing also *Actaeonella* sp., *Plesioptygmatis* sp., rudist biostromes, *Dicyclina schlumbergeri, Trochospira avnimelechi, Nezzazata simplex, Moncharmontia appenninica, Nummoloculina heimi, Valvulammina picardi, Bacinella irregularis, Gavelinella* sp., *Praesorites* sp., *Cuneolina* sp., *Terquemella* sp., *Trochospira* sp., miliolids and ostracodes, which belong to the Turonian. Both Turonian and the younger deposits have a more restricted distribution than those described above, due to erosion.
- 190-310 m of massive limestones in the lower section and breccia-limestones with rudist biostromes higher upsection. Here we have encountered: *Hippurites cornuvaccium, Radiolites spinulatus, Medella (Fossulites)* sp., *Radiolites* sp., *Trochactaeon crisminensis, Actaeonella* sp., as well as *Biconcava bentori, Trochospira avnimelechi, Dictyopsella kiliani, Abrardia mosae, Moncharmontia appenninica, Murciella cuvillieri, Reticunella reicheli, Coxites* cf. *zubaivensis, Stensiöina surretina, Dicyclina schlumbergeri, Bacinella irregularis, Thaumatoporella parvovesiculifera, Nummofallotia apula, Valvulammina picardi, Accordiella conica, Nezzazata simplex, Rotorbinella scarsellai, Minouxia lobata,* and others which confirm the Coniacian up to Santonian;
- 80-150 m of breccia- and massive limestones which are encountered only in the Selca region with *Dicyclina schlumbergeri, Stensiöina surretina,*

- 25 m of brecciated and conglomeratic limestones with limestone-pebbles of the Upper Triassic in which we encountered *Orbitolina* sp., *Trocholina friburgensis, Bacinella irregularis* and rudists which indicate a Lower Cretaceous (Barremian-Aptian) age;
- 11 m of calcarenitic limestones with chert-lenses bearing *Rotalipora* sp., *Globotruncana lapparenti* and in the upper part *G. coronata, G. bulloides,* Heterohelicidae, Globigerinidae, Calcisphaerulidae, and rudist fragments, indicating Lower and Upper Cretaceous ages;
- 40 m of calcarenitic and platy, conglomeratic limestones with *G. tricarinata, G.* cf. *angusticarinata,* Globigerinidae and rudist fragments;
- 50 m of conglomeratic and calcarenitic limestones with *Orbitoides* cf. *media, Globotruncana* cf. *conica,* Globigerinidae and rudist fragments, which belong to the Senonian-Maastrichtian;
- more than 400 m of alternating clay-carbonatic shales, sands, marls and less frequently limestones which constitute the Vermoshi Flysch, these including *G. contusa, G. gansseri, G. stuarti* which indicate the Maastrichtian.

2.1.6 The Kelmendi Zone

This zone was established by PEZA et al. (1988) and is encountered in the northernmost extremity of Albania, in the Vermoshi valley (Kelmendi region) and constitutes the analogon of the Boshnjak Zone of the Dinarides. The deposits of this zone belong to the Upper Jurassic-Cretaceous, developed in flysch facies, which for a long time have been confused with the deposits of the Vermoshi Flysch they are in tectonic contact with.

These deposits consist of a 400-500 m thick alternation of mainly shales with thin marly limestones, calcarenitic and breccia-limestones, sandstones, and occasionally microconglomerates. All the rocks are rather compressed and folded, because of the overthrusting Gashi Zone further north.

In the earliest limestones of the section poorly preserved tintinnids (Tithonian-Neocomian) are encountered.

Above them there is the level with *Orbitolina* sp., *Camptocampylodon fontis,* Miliolidae, Textulariidae, Codiacea and rudist fragments, which are of Barremian-Aptian age.

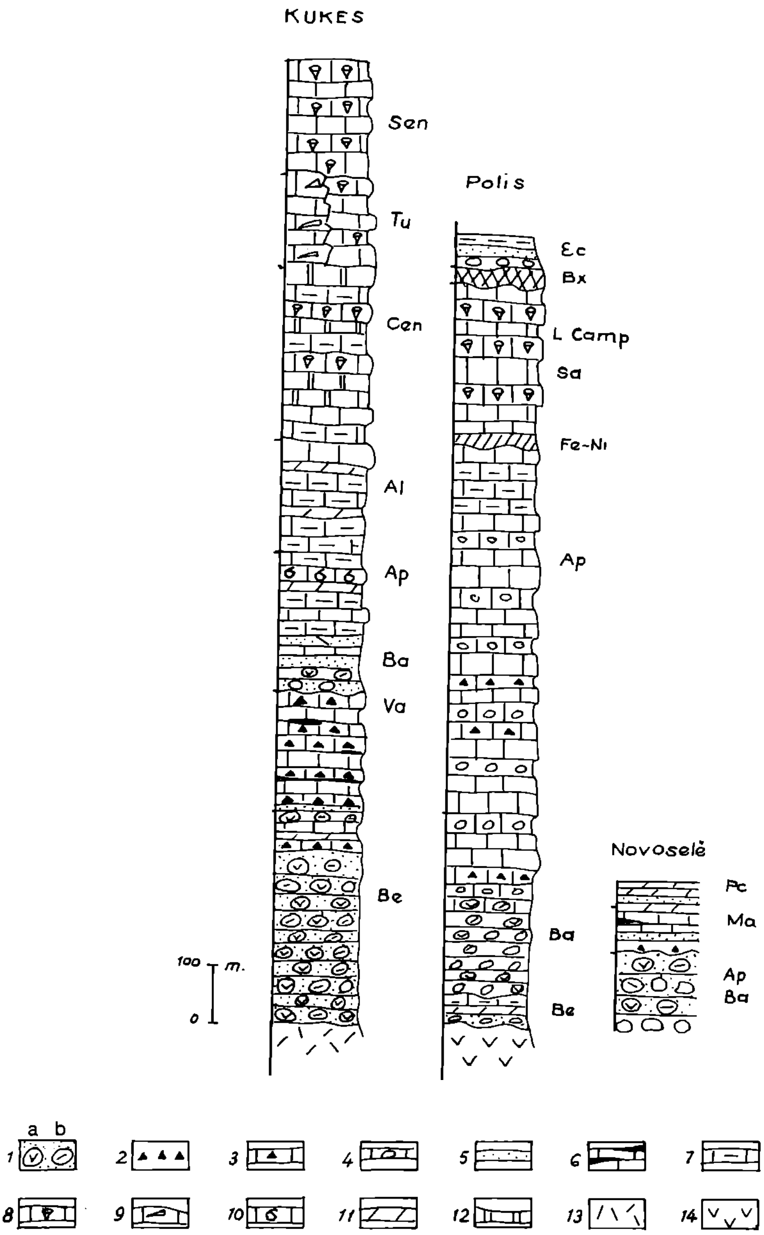
Years ago, a level was met in these deposits with *Ticinella* sp., *Planomalina* sp., *Pithonella ovalis,* and *Rotalipora* which belong to the Albian and which may pass over to the Cenomanian as well (PAPA et al. 1977).

Work is under way to complete the stratigraphy of the Kelmendi Zone deposits, which without doubt will contribute new data to complete its framework.

2.2 The internal Albanides

2.2.1 The Mirdita Zone (Text-Fig. 9)

This zone is the analogon of the Serb Zone in the Dinarides and of the Subpelagonian in the Hellenides. In this zone the Cretaceous has a wide geographic distribution and is developed in many facies. Until recent years, the sections of Cretaceous deposits have been considered to lack stratigraphic gaps, but the latest studies brought into evidence many stratigraphic



Text-Fig. 9. Sections of the Mirdita Zone. 1 - Conglomerates, a - Ophiolite pebbles, b - Limestones; 2 - Breccias; 3 - Breccia-limestones; 4 - Conglomeratic limestones; 5 - Sandstones; 6 - Limestones with chert-lenses; 7 - Marly limestones; 8 - Limestones with rudists; 9 - Limestones with *Chondrodonta;* 10 - Limestones with *Requienia;* 11 - Marls; 12 - Dolomites; 13 - Volcanics; 14 - Ultramafics.

In the region of Velivari and Kercuna Mountain, south of the town of Peshkopia, flysch deposits are encountered with reworked tintinnids of Upper Tithonian-Neocomian age, resembling those of the Mirdita Zone. Overlying them tectonically there are limestones with *Globotruncana lapparenti* and *G. conica*, over which Maastrichtian-Paleocene flysch continues (Gjeologjia e Shqipërisë 1982, MELO 1982).

South of Peshkopia, there is (according to Dr. V. KICI) a sequence of flysch again with reworked Tithonian-Neocomian calpionellids and a presumed continuation of grey and red clay occasionally alternating with sandstone strata. These contain globotruncanids of the Albian-Cenomanian and are overlain by:

- 150-200 m of micritic limestones with globotruncanids of Senonian-Maastrichtian age;
- 400 m of alternating micritic platy limestones and marls which pass into marly flysch with occasional strata of limestones and sands. In their lower part globotruncanids of the Maastrichtian appear, while the upper part passes into the Paleocene and Eocene.

3. Conclusions

The main conclusions for the Cretaceous deposits are as follows:

1. Cretaceous deposits have a wide distribution and are met in almost all tectonic zones. An exception comprises the Zone of Gashi (in the northern part of Albania), where Cretaceous deposits are not encountered due to late erosion.
2. Cretaceous deposits in Albania are developed in three main facies types: a - neritic facies, b - pelagic facies, and c - flysch facies.
   a. In neritic facies, Cretaceous deposits are mainly represented by thick strata of massive limestones, conglomeratic limestone deposits of both Urgonian and Gosau facies. Cretaceous deposits of neritic facies are present in the tectonic zones of Sazani, Kruja, Albanian Alps, Mirdita and Korabi.
   b. Cretaceous deposits of pelagic facies are represented by pelagic limestones with chert lenses, stratified marly limestones and marls. The deposits of this facies are located in the zones of Jonian Krasta-Cukali and Valbona.
   c. Flysch deposits are developed in the subzone of Valbona, Kelmendi Zones, Krasta-Cukali Zone, Mirdita Zone, and Korabi Zone.
3. The formation of Cretaceous conglomerates in the Albanides was mainly influenced by different phases of tectonic movements. Owing to these movements depositional facies changes occurred within the basin as well as did various stratigraphic gaps. These phenomena are more prominent in the Mirdita Zone, where several phases of Alpine orogenesis are documented.

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Lower Cretaceous Stratigraphy and Paleogeography of the Czechoslovakian Western Carpathians

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With 3 Plates, 2 Text-Figures and 3 Tables

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Abstract: The West Carpathian area, split by the evolving Penninic (= Vahic) Zone, belonged in the Mesozoic to the northern part of the Mediterranean Tethys. Pelagic sediments with abundant microplanktonic, nannoplanktonic and macrofaunal remains (mainly cephalopods) were deposited here during early Cretaceous time.

The best documented Outer Carpathian sequences (Berriasian to lowermost Cenomanian) are known from the Beskidic Zone. Valanginian and Lower Hauterivian pelosideritic marls contain mixed Boreal and Mediterranean ammonite faunas. On the other hand, the composition of younger ammonite associations is exclusively of the Mediterranean type.

The present Central Carpathian sequences accumulated on an extensive sialic fragment, separated from the European shelf by Penninic rifting. Monotonous marly limestones ("Neocomian Facies") prevailed here. Although a late Hauterivian Boreal belemnite (*Aulacoteuthis*) has been found in a peripheral Tatric Unit, the fauna in all the remaining sections is of pure Mediterranean character. The development of carbonate basins and adjacent ramps was truncated by Barremian to early Albian progradation of carbonate platforms. After the Lower Albian, carbonate sedimentation suddenly stopped, being substituted by bathyal pelites.

Kurzfassung: Das Mesozoikum der Westkarpaten, das im Bereich des Pennikums abgelagert wurde, gehört zum Nordteil der mediterranen Tethys. Die pelagischen Ablagerungen der Unterkreide führen eine reiche Nanno-, Mikro- und Makrofauna (besonders Cephalopoden).

Die bestbelegten Schichtfolgen der Äußeren Karpaten (Berrias bis unterstes Cenoman) sind aus dem Beskidikum (Schlesische Einheit) bekannt. Die toneisenhaltigen Mergel (Valangin - Unterhauterive) enthalten gemischte boreale und mediterrane Ammonitenfaunen. Die jüngeren Ablagerungen führen nur noch Ammonitenfaunen der mediterranen Provinz.

Die Schichtfolgen der Zentralkarpaten entstanden auf einem großen sialischen Fragment, das vom europäischen Schelf durch das penninische Riftsystem abgetrennt wurde. Hier überwiegen monotone mergelhaltige Kalksteine der "Neokom-Fazies". Die faunistischen Funde haben in allen Profilen ausschließlich mediterranen Charakter, mit Ausnahme eines Belemniten der Gattung *Aulacoteuthis* (oberes Hauterive), der in einer Randeinheit des Tatrikums gefunden wurde. Die Entwicklung von Becken und benachbarten Plattformen, in denen vorwiegend Karbonate abgelagert wurden, änderte sich durch Progradation der Karbonatplattformen vom Barreme bis frühen Alb. Die Karbonatsedimentation endete plötzlich nach dem Unteralb und wurde durch bathyale Pelite ersetzt.

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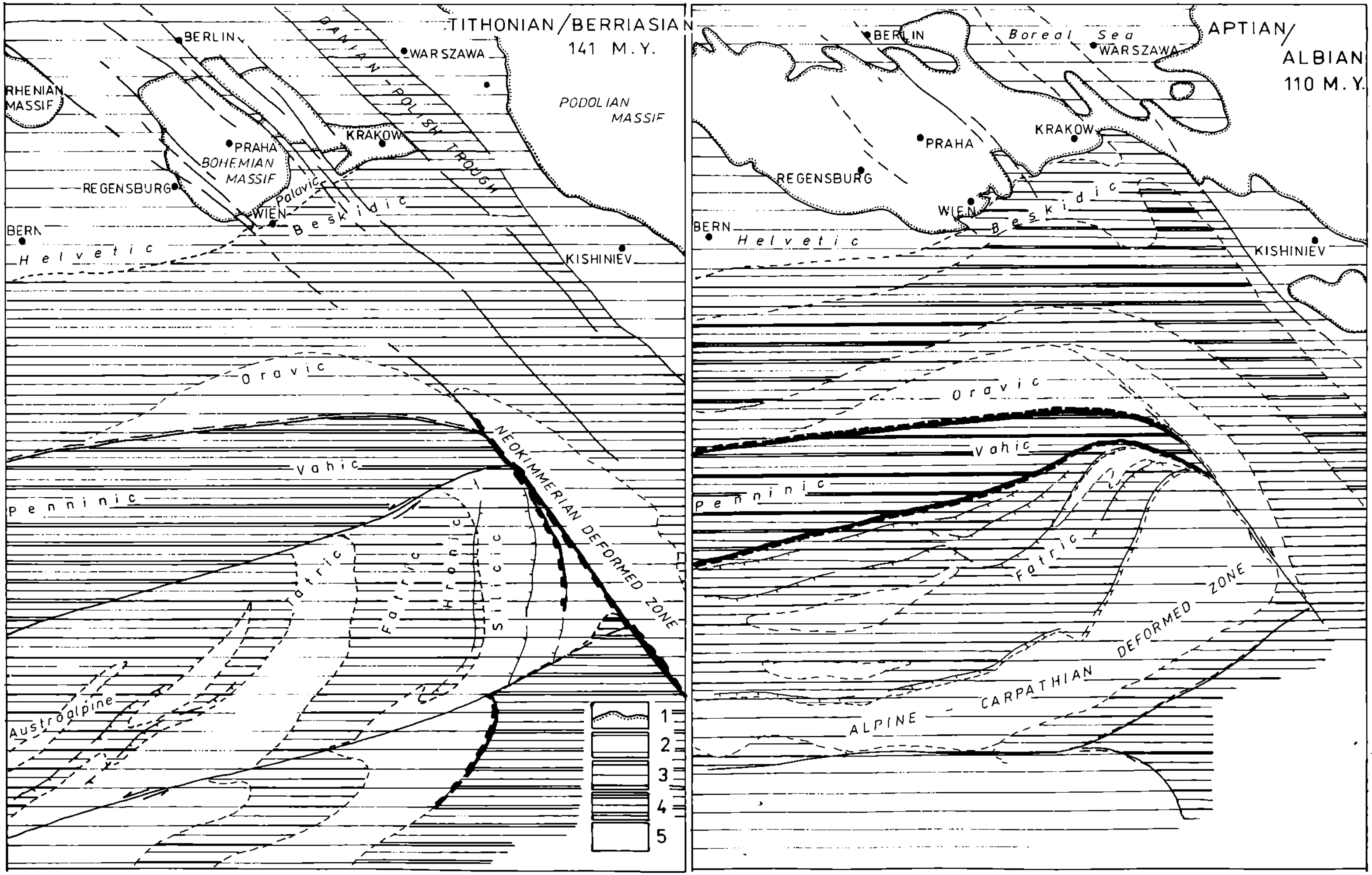
1. Introduction

Lower Cretaceous pelagic ("Neocomian") carbonates form an important part of the Mesozoic sedimentary sequences in the West Carpathian area (BORZA et al. 1980, 1984, 1987, MICHALÍK et al. 1987, VAŠÍČEK et al. 1983). We divided the Tithonian to Hauterivian sequences on the basis of the distribution of microplanktonic organisms (BORZA & MICHALÍK 1986, 1987). In addition, we considered the applicability of French and Bulgarian ammonite biostratigraphic schemes for the division of the Carpathian Hauterivian and Barremian sections (ADAMIKOVÁ et al. 1983, MICHALÍK & VAŠÍČEK 1984, 1987, VAŠÍČEK 1972, 1979a, b, VAŠÍČEK & MICHALÍK 1981, 1986, 1988). Finally, we tried to apply our results in some palinspastic reconstructions, and in reconstructions of interrelations in ancient faunal assemblages (MICHALÍK & VAŠÍČEK 1980, MICHALÍK & KOVÁČ 1982, MICHALÍK 1987, VAŠÍČEK & MICHALÍK 1981). In this report, we conclude some results of Lower Cretaceous ammonite biostratigraphy obtained during study of our fossiliferous sections and we use them in reconstruction of West Carpathian early Cretaceous development.

In numerous measured sections, we collected fossil macrofauna by bed-by-bed method (cf. VAŠÍČEK et al. 1983). The majority of profiles was sampled in metric intervals. We obtained several thousands of specimens. Ammonites and belemnites belong to the most frequent fossils. Brachiopods, echinoids and crinoids are less abundant, while gastropods, bivalves, sponges, corals and fish teeth occur rarely. The microfauna has been studied in thin sections. The evaluation of faunal distribution has been published in fore-mentioned papers.

Text-Fig. 1. Paleogeographical sketch of the West Carpathian territory during Tithonian-Berriasian boundary interval (left) and during Aptian-Lower Albian compressional events, illustrating change from Neokimmerian to Paleoalpine structural pattern.
Explanations: 1 - coast line, 2 - shallow marine sedimentation, 3 - basinal sedimentation, 4 - deep basins (partially oceanized?) and oceanic rifts, 5 - land.


TITHONIAN/BERRIASIAN
141 M.Y.
BERLIN
WARSZAWA
RHENIAN MASSIF
DANIAN-POLISH TROUGH
PODOLIAN MASSIF
PRAHA
BOHEMIAN MASSIF
KRAKOW
REGENSBURG
Palavic
WIEN
Beskidic
BERN
Helvetic
KISHINIEV
Oravic
NEOKIMMERIAN DEFORMED ZONE
Vahic
Penninic
Tatric
Fatric
Hronic
Silicic
Austroalpine
1
2
3
4
5
APTIAN/ALBIAN
110 M.Y.
BERLIN
Boreal Sea
WARSZAWA
KRAKOW
PRAHA
REGENSBURG
WIEN
Beskidic
BERN
Helvetic
KISHINIEV
Oravic
Vahic
Penninic
Fatric
ALPINE - CARPATHIAN DEFORMED ZONE

2. Paleogeographical setting

The problems connected with paleogeographical interpretations of the West Carpathian territory are caused both by the complicated structure of the mountains, as well as by several primary reasons. The orogen at present consists of several parts (Outer, Central and Inner Carpathians) evolving independently before the Alpine collision. Moreover, the evolution of all the Carpathians has been influenced by the immediate proximity of the triple junction between the shelves of both the northern and eastern European platforms and the Tethyan Penninic domain, the latter being subsequently subducted below the Alpine-Carpathian Orogene System (MICHALÍK & KOVÁČ 1982).

The development of the Outer Carpathians is intimately connected with the easternmost part of the North European Platform shelf in the proximity of the mouth of the Danian-Polish Trough (Text-Fig. 1). Episodical opening of this sea-way offered the opportunity of faunal exchange between the Tethyan and Boreal realms (VAŠÍČEK 1972, 1979b), while trends in sedimentation recorded the paleotectonic lability of this region.

The Central Carpathians were a part of the Austroalpine-Carpathian sialic block, which belonged to the European shelf during the Triassic. This is why the Central Carpathian sequences are similar to many East Alpine ones. Some differences have been caused by particularities of the eastern termination of this shelf block, which moved southeastwards during the Jurassic, co-acting together with the Adriatic microplate in Kimmerian convergence in the Eastern Carpathians, Balcanides, Dinarides and Hellenides. The late Kimmerian collision also indirectly influenced the development in the Western Carpathians (but in the East Alpine sector, too). During the early Cretaceous, the paleogeographical pattern gradually changed. The Kimmerian System of meridionally oriented structures disintegrated, being gradually superimposed by latitudinally arranged submarine troughs and ridges. This reorganization was caused by change in movement of the African continental plate, which led to slow subduction of the Penninic suboceanic (?) crust, and to a compression in the Alpine-Carpathian area as well.

3. West Carpathian area at the beginning of early Cretaceous

After the mid-Jurassic deep-sea stage, a new, gradual Malm-aged shallowing caused a considerable diversification of sedimentary environments, which affected the development of plankton. Planktonic communities, reacting sensitively to minor climatic and other environmental fluctuations, certainly

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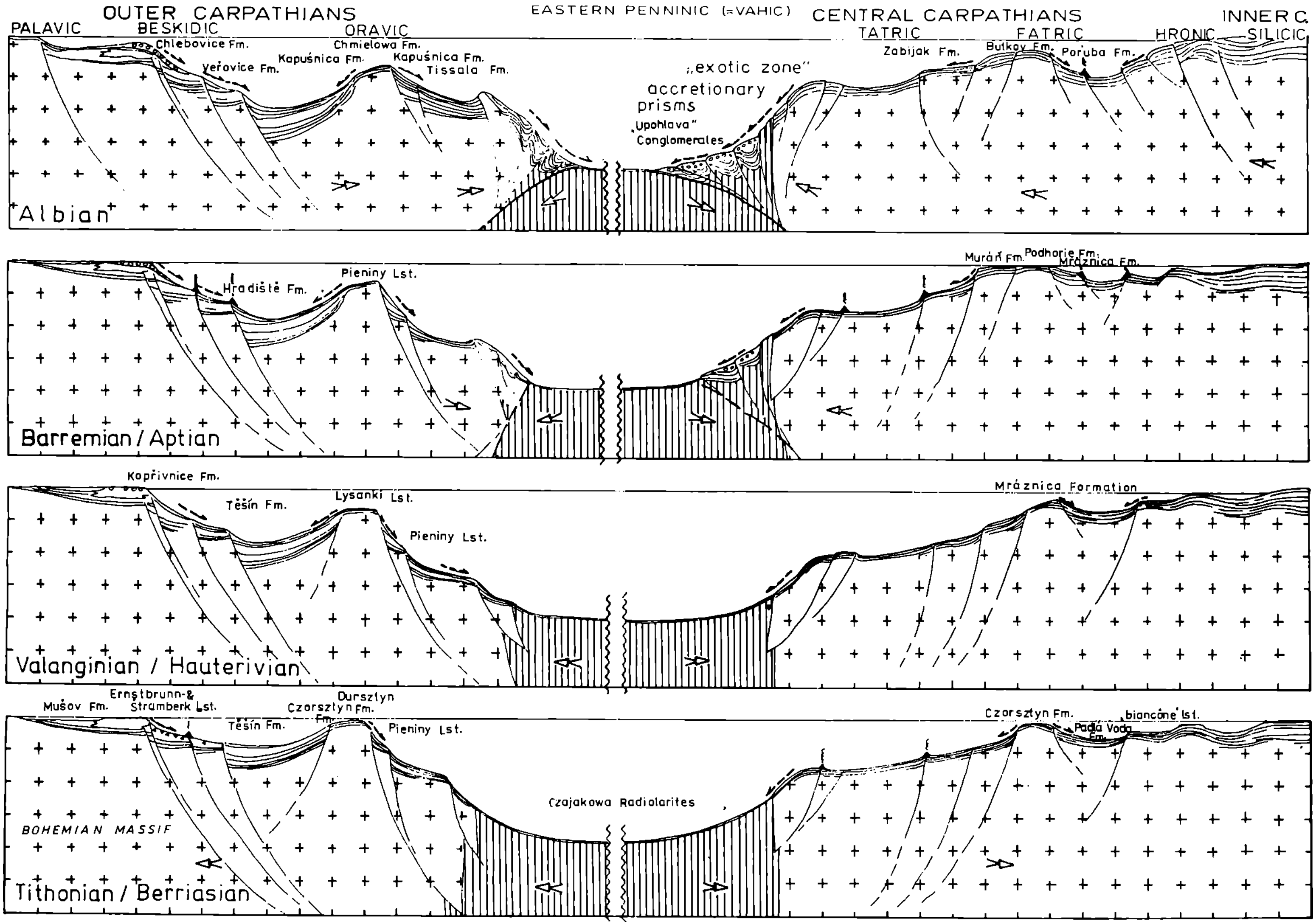
Text-Fig. 2. Simplified paleotectonic sections through the West Carpathian territory characterizing the development of the Tethyan northern margin during Lower Cretaceous. Thickness of sediments exaggerated. Crosses - continental crust, vertical hatching - suboceanic (?) crust, small circles - coarse clastics, black triangles - volcanics, white triangles - breccias, broken arrows - submarine transport (turbidites and debris flows), thick arrows - direction of tectonic movements.



OUTER CARPATHIANS
EASTERN PENNINIC (=VAHIC)
CENTRAL CARPATHIANS
INNER C.
PALAVIC
BESKIDIC
ORAVIC
TATRIC
FATRIC
HRONIC
SILICIC
Chlebovice Fm.
Veřovice Fm.
Kapuśnica Fm.
Chmielowa Fm.
Kapuśnica Fm.
Tissala Fm.
„exotic zone"
accretionary prisms
„Upohlava" Conglomerates
Zabijak Fm.
Butkov Fm.
Poruba Fm.
Albian
Hradiště Fm.
Pieniny Lst.
Muráň Fm.
Podhorie Fm.
Mráznica Fm.
Barremian / Aptian
Kopřivnice Fm.
Těšín Fm.
Lysanki Lst.
Pieniny Lst.
Mráznica Formation
Valanginian / Hauterivian
Mušov Fm.
Ernstbrunn-& Stramberk Lst.
Těšín Fm.
Czorsztyn Fm.
Dursztyn Fm.
Pieniny Lst.
Czorsztyn Fm.
biancone lst.
Padla Voda Fm.
Czajakowa Radiolarites
BOHEMIAN MASSIF
Tithonian / Berriasian

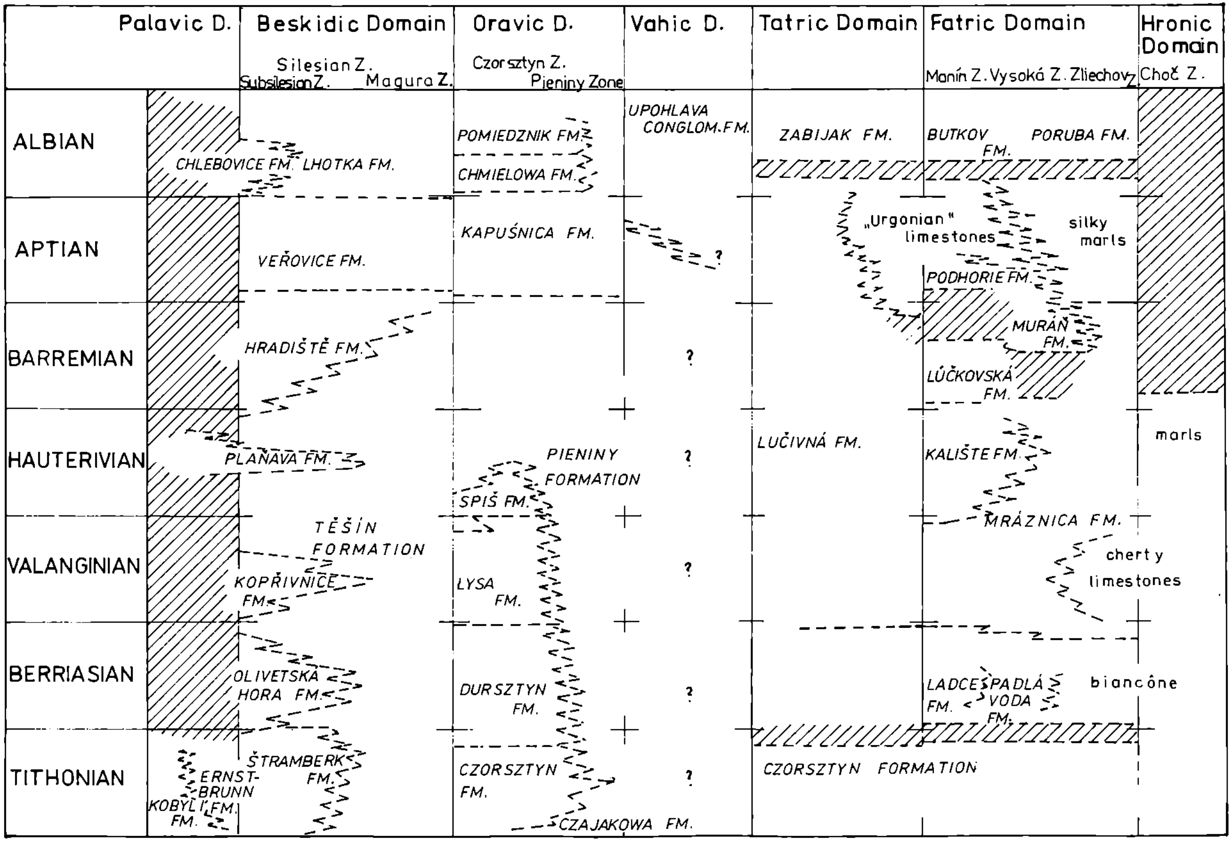
differ from the relatively monotonous and poorly developed Dogger associations. The Kimmeridgian, Tithonian and Berriasian pelagic formations are well distinguishable according to vertical changes in associations of planktonic microfossils (mostly tintinnids, stomiosphaeras, cadosinas, but also nannocones and nannoplanktic remnants), which constitute rock-forming accumulations.

Sediments of the Jurassic/Cretaceous transition beds are well represented in the Western Carpathians. Lagoonal complexes of neritic limestones and clastics (ELIÁŠ 1981, MICHALÍK et al. 1987) arose in a wide nearshore belt on the Bohemian Massif shelf. A belt of carbonate platforms and bioherms (Ernstbrunn, Štramberk Limestones), adjacent to the Těšín Formation basin (carbonate flysch with olistoliths of reefal limestones) evolved along the southern border of the neritic zone. The zone of elevated fault blocks dividing the Těšín Formation basin from the suboceanic Penninic Trough (lower panel of Text-Fig. 2 and Table 1) was characterized by red nodular limestones of Ammonitico Rosso facies (Czorsztyn Limestone). They interfingered with red marls on the slopes of the elevated areas. The tops of the elevations were covered by white crinoidal limestones, that became the dominating lithofacies during late Kimmerian movements (Dursztyn Limestone Formation with breccia intercalations, BIRKENMAJER 1977). Pelagic radiolarian-nannocone limestones of Maiolica type (Pieniny Limestone Formation), with an increasing share of siliceous nodules, were deposited deeper along the periphery of the Penninic (= Vahic, MAHEL' 1979) Trough. The Penninic sea floor was below the CCD and thus, sediments are characterized by radiolarites similar to the more widespread Oxfordian-Kimmeridgian Czajakowa Radiolarites. No information is available on the character of the Penninic substrate crust, with the exception of the presence of numerous small basic volcanic bodies (těšínites and limburgites) in the adjacent basins.

Red nodular Czorsztyn Limestone was also deposited on an elevated fault block on the southern side of the Penninic "Ocean", the area that became the Central Carpathians (Text-Fig. 2). Pelagic pink-gray biomicrital limestones of Biancône type were deposited on the slopes of this fault block. In contrast, shallow-marine limestone pebbles have been described (MIŠÍK & SÝKORA 1980) from the southernmost parts of the Carpathian block (Inner Carpathians). The late Kimmerian movements slightly affected some zones in the Central Carpathians during the late Tithonian and (mainly) during the Berriasian, when brecciated limestone intercalations originated (Nozdrovice Breccia, BORZA et al. 1980). They indicate an extensive destruction of both contemporary and Tithonian sediments.

The reorganization of paleogeographic conditions, and probably also sea-currents, caused a new quantitative boom in plankton development during Berriasian and Valanginian. Nannocone biomicrites rapidly covered the area of former sea floor facies. As a result, Berriasian and Valanginian sedimentary formations originated in an oxidizing eupelagic environment (Ladce, Padlá Voda formations, as well as other Biancône type limestones). The representatives of benthos are practically absent in the fossil collections; nektonic organisms are rare, i. e. aptychi (*Lamellaptychus mortilleti-noricus* TRAUTH, *Punctaptychus* sp.), corroded ammonite remnants (*Berriasella* cf. *picteti* JACOB), as well as several broken belemnite rostra. Plankton is characterized by the presence of nannocones and calpionellids (*Calpionella alpina* LORENZ, *C. elliptica* CAD., *Tintinnopsella carpathica* MURG. & FIL., *Remaniella cadischiana* (COLOM)); other microfossils are foraminifers, radio-

Table 1. A simplified Lower Cretaceous lithostratigraphical scheme of the Outer and Central West Carpathian sequences. The major stratigraphical gaps are diagonally hatched.



larians, sponge spicules, and ostracodes. The top part of the formations contains associations of both, *Calpionellopsis* and *Calpionellites* zone microfossils.

4. Valanginian and Hauterivian development

During Valanginian, tensional tectonics continued in basins along the northern Penninic periphery. The resulting subsidence affected not only the basins, but also elevated zones. Step-like diversification of bottom relief activated slumping and turbidity currents (Text-Fig. 2). On the other hand, emergence of the Bohemian Massif shelf (Palavic Domain, MAHEL' 1986) led to accumulation of terrigene clastics (HOUŠA 1983). Black flysch sedimentation continued in the Silesic Unit of the Beskidic Domain. Mediterranean types of ammonites dominate the Valanginian and Lower Hauterivian ammonite fauna of the Těšín Formation (VAŠÍČEK 1979b) over less frequent Boreal species (*Platylenticeras heteropleurum* NEUM. & UHLIG and *Endemoceras amblygonium* NEUM. & UHLIG). The late Hauterivian and early Lower Barremian ammonite fauna is very poorly represented by Mediterranean species.

The elevated faulted block zone of the Oravic Domain (MAHEL' 1986) became less expressive at this time. Organodetrital limestones (Lysa Limestone Formation) with sedimentary dykes and erosive grooves filled by crinoidal limestone (Spiš Limestone Formation) were deposited in the shallowest Czorsztyn Zone. The slopes of these elevated faulted blocks have been rimmed by monotonous nannocone limestones of Maiolica type (Pieniny Limestone Formation).

The Central Carpathian region was also affected by tensional stress. It is characterized by marly limestones with cherts in depressions. Micrital, intensively bioturbated limestones of the Mrâznica Formation with fluxoturbiditic bioclastic limestone intercalations were deposited on the slopes of the elevated faulted blocks (Text-Fig. 2). Reworked clasts of shallow-marine organisms indicate the existence of local shallows. In the Tatric Domain, the finding of *Aulacoteuthis* aff. *compressa* MUTT. in a green

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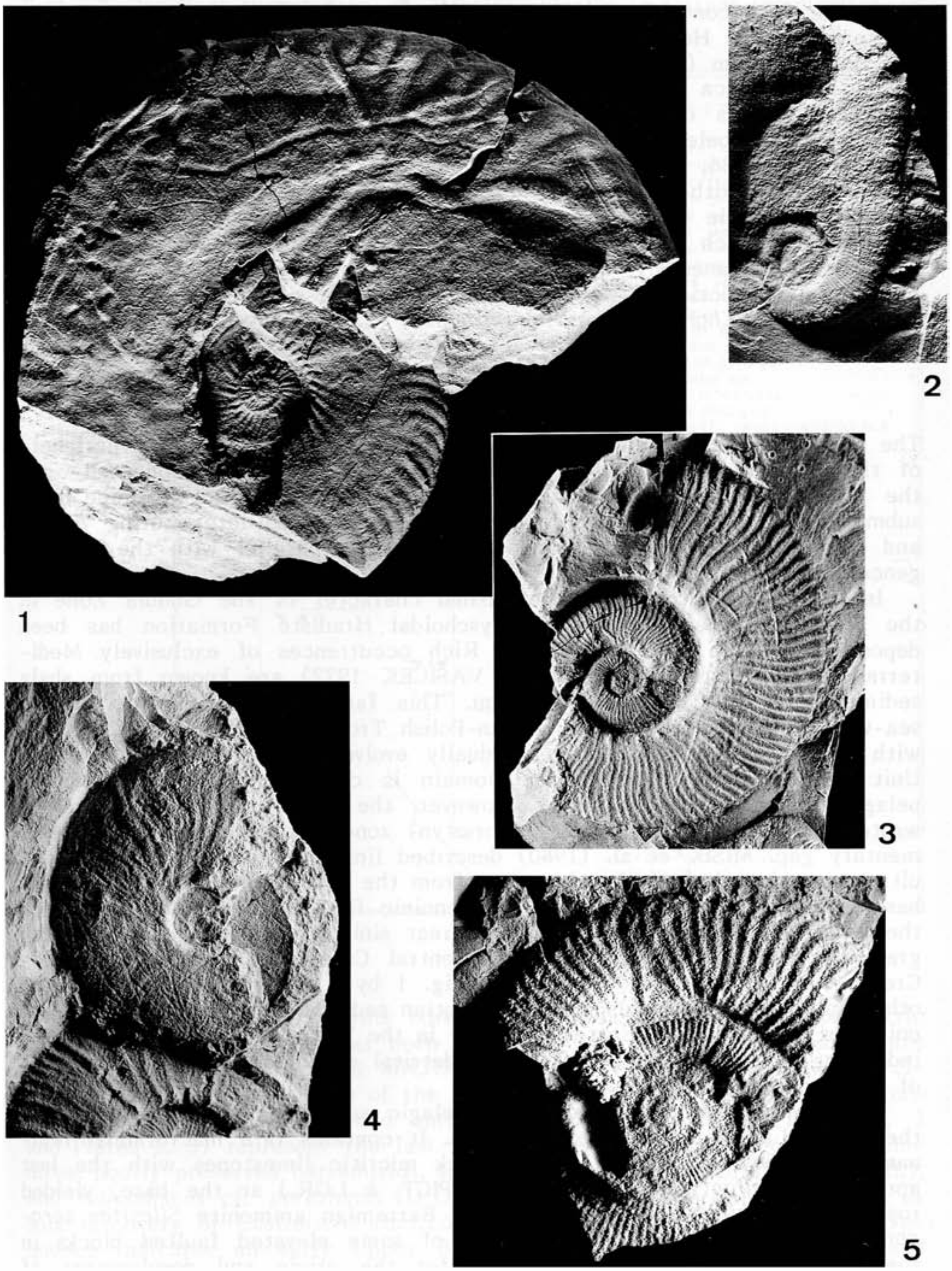
Plate 1

All specimen figured in natural size.

Fig. 1. *Busnardoites campylotoxus* (UHLIG 1902). Late Lower Valanginian. Butkov quarry (Strážovské Vrchy Mts.), prospecting gallery 10.94 m.
Fig. 2. *Acrioceras (Acrioceras) seringei* (ASTIER 1851). Lower Barremian. Polomec quarry (Strážovské Vrchy Mts.), 5th level, 250 m.
Fig. 3. *Crioceratites binelli* (ASTIER 1851). Lower Barremian. Polomec quarry, 5th level, 250 m.
Fig. 4. *Barremites (Barremites) dimboviciorensis* BRESKOVSKI, 1966. Lower Barremian. Polomec quarry, 5th level, 500 m, underlier of bed no. 32.
Fig. 5. *Silesites seranonis* (D'ORBIGNY 1841). Upper Barremian. Bralo quarry (Malá Fatra Mts.), 2nd level.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marlstone bed containing belemnites (Lučivná locality) indicates a late Hauterivian age. However, this genus has hitherto been known only from the Boreal Realm (GLAZUNOVA 1969, MUTTERLOSE 1983). On the other hand, the Mráznica Formation in the Fatric Domain contains rich Mediterranean associations of microfauna, nannoflora (BORZA et al. 1980, 1984, 1987), aptychi, belemnites and ammonites (VAŠÍČEK et al. 1983, VAŠÍČEK & MICHALÍK 1986, 1988). Dark marlstones and marly limestones in Polomec Hill at Žilina with *Acrioceras mulsanti* (AST.) and *Euptychoceras borzai* VAŠ. & MICH. lie above late Hauterivian turbidite beds. The abundance of benthic fauna such as echinoids, crinoids and brachiopods (MICHALÍK 1987) arose contemporaneously with the appearance of early Hauterivian ammonites in many sections. The most frequent ichnofossil genera were *Zoophycos, Planolites, Chondrites,* or *Helminthopsis.*

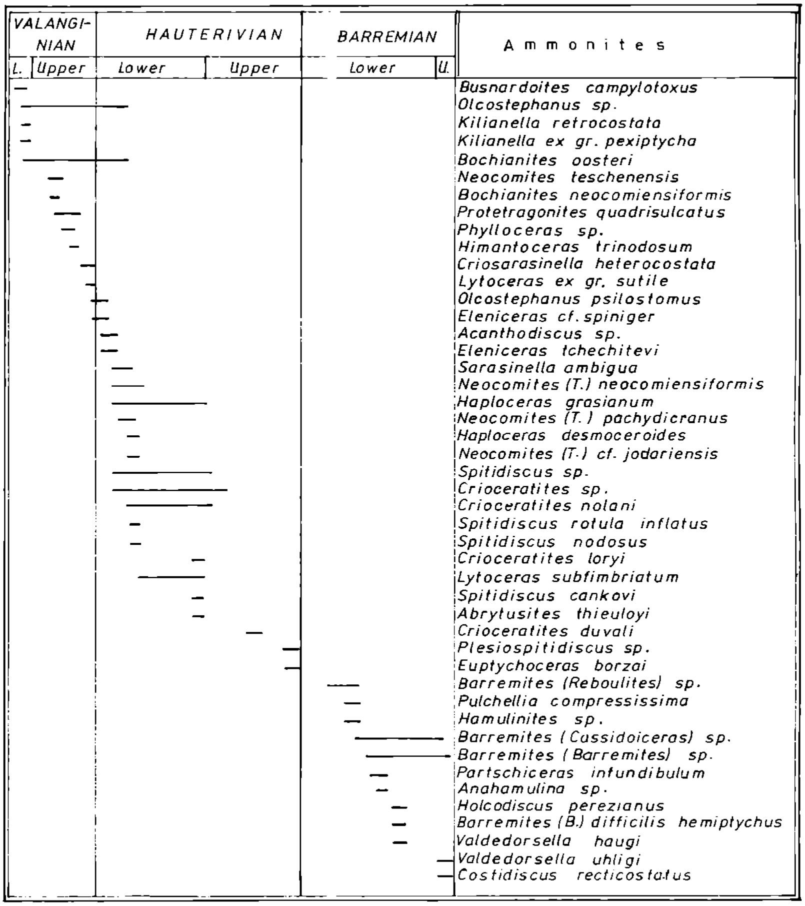
5. Barremian development

The first inexpressive compressional events ran along the southern periphery of the Penninic Domain at this time, and were probably connected with the start of subduction in this belt. However, tension accompanied with submarine basic volcanism (hyaloclastites), accentuated later during Aptian and early Albian time, continued in depressions parallel with the convergence zone.

In the Outer Carpathians, the basinal character of the Godula Zone in the Beskidic Domain, where the flyschoidal Hradiště Formation has been deposited, became better expressed. Rich occurrences of exclusively Mediterranean ammonites (UHLIG 1883, VAŠÍČEK 1972) are known from shaly sediments of the Hradiště Formation. This fact indicates closing of the sea-way in the mouth of the Danian-Polish Trough. A deep back-arc basin with black flysch sedimentation gradually evolved in the southern Beskidic Unit (Magura Zone). The Oravic Domain is characterized by monotonous pelagic Pieniny Limestone facies. However, the Barremian sequence in the western part of the shallowest (Czorsztyn) zone is interrupted by a sedimentary gap. MIŠÍK et al. (1980) described limestone pebbles with detrital ultrabasic minerals (spinels, chromite) from the Oravic Domain. Thus, ultrabasic bodies in the proximity of the Penninic Domain were uncovered since the Barremian (most probably along a great sinistral strike-slip fault, which gradually joined both the Outer and Central Carpathian areas during mid-Cretaceous time; it is shown on Text-Fig. 1 by a heavy dark line). On the other hand, detrital spinels appear in Aptian sediments in Central Carpathian units, as well as in the Eastern Alps, or in the Transdanubian Central Range, indicating a younger age of ultrabasic detrital support on the southern side of the Penninic Domain.

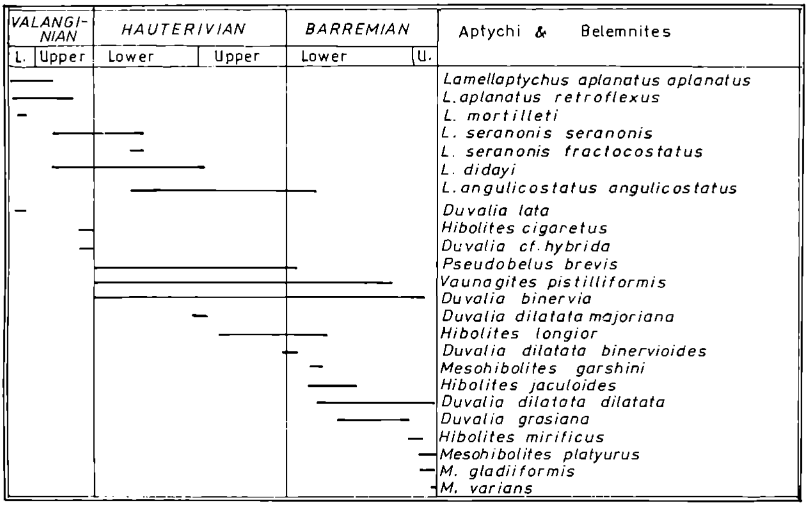
In the Tatric Domain, Barremian pelagic sediments are represented by the cherty Lučivná Limestone Formation. It contains rare macrofaunal remnants only. Several hundred meters thick micritic limestones with the last aptychi (*Lamellaptychus angulicostatus* PICT. & LOR.) at the base, yielded rostra of *Mesohibolites* sp. and the late Barremian ammonite *Silesites seranonis* (D'ORB.) (Plate 1). The surface of some elevated faulted blocks in the Tatric Domain yielded conditions for the origin and development of "Urgonian" carbonate platforms, which prograded during the Barremian, Aptian and early Albian across marginal slopes.

Table 2. Vertical distribution of stratigraphically important ammonites in the Lower Cretaceous sequence of Butkov quarry.



Similar carbonate platforms covered also areas marginal to the Tatric Domain. This development has been described and documented by ammonite biostratigraphy in VAŠÍČEK & MICHALÍK (1986) from the Manín Unit. Yellowish weathering limestones of the Lúčkovská Formation with intercalations of marls and with a rich and entirely Mediterranean fauna (Tables 2, 3 and Plates 2, 3) represent the last hemipelagic sediments. Frequent belemnites, poorly preserved barremitids, single *Pulchellia compressissima* (D'ORB.), and *Holcodiscus* cf. *perezianus* (D'ORB.) prove for early Barremian age. A sole discovery of *Costidiscus recticostatus* (D'ORB.) at the top of the sequence indicates an early Upper Barremian age. Brecciated horizons and submarine sliding phenomena indicate sedimentation in a slope environment. These sediments are covered, with probable sedimentary gap, by detrital slope material (Podhorie Formation) of a prograding carbonate platform,

Table 3. Vertical distribution of aptychi and belemnites in the Lower Cretaceous sequence of Butkov quarry.



containing reworked belemnite rostra and clasts of underlying as well as the contemporaneous sediment. The youngest clasts in the basal breccia are of early Upper Aptian age.

The prevailing part of the Fatric Domain had a basinal character. Deposition of spotted Mráznica Limestone occurred on adjacent slopes from the Hauterivian to the Barremian. The Hauterivian/Barremian boundary is well recognizable in this sequence by a conspicuous crioceratitid ammonite shell bed. It contains numerous *Crioceratites* sp., *Pseudothurmannia shankariae* SARKAR and *Acrioceras seringei* (AST.) at the base (Plate 1). Several representatives of *Anahamulina* HYATT, *Crioceratites, Pseudothurmannia* (including the zonal index species *P. angulicostata* (D'ORB.) and the subzonal

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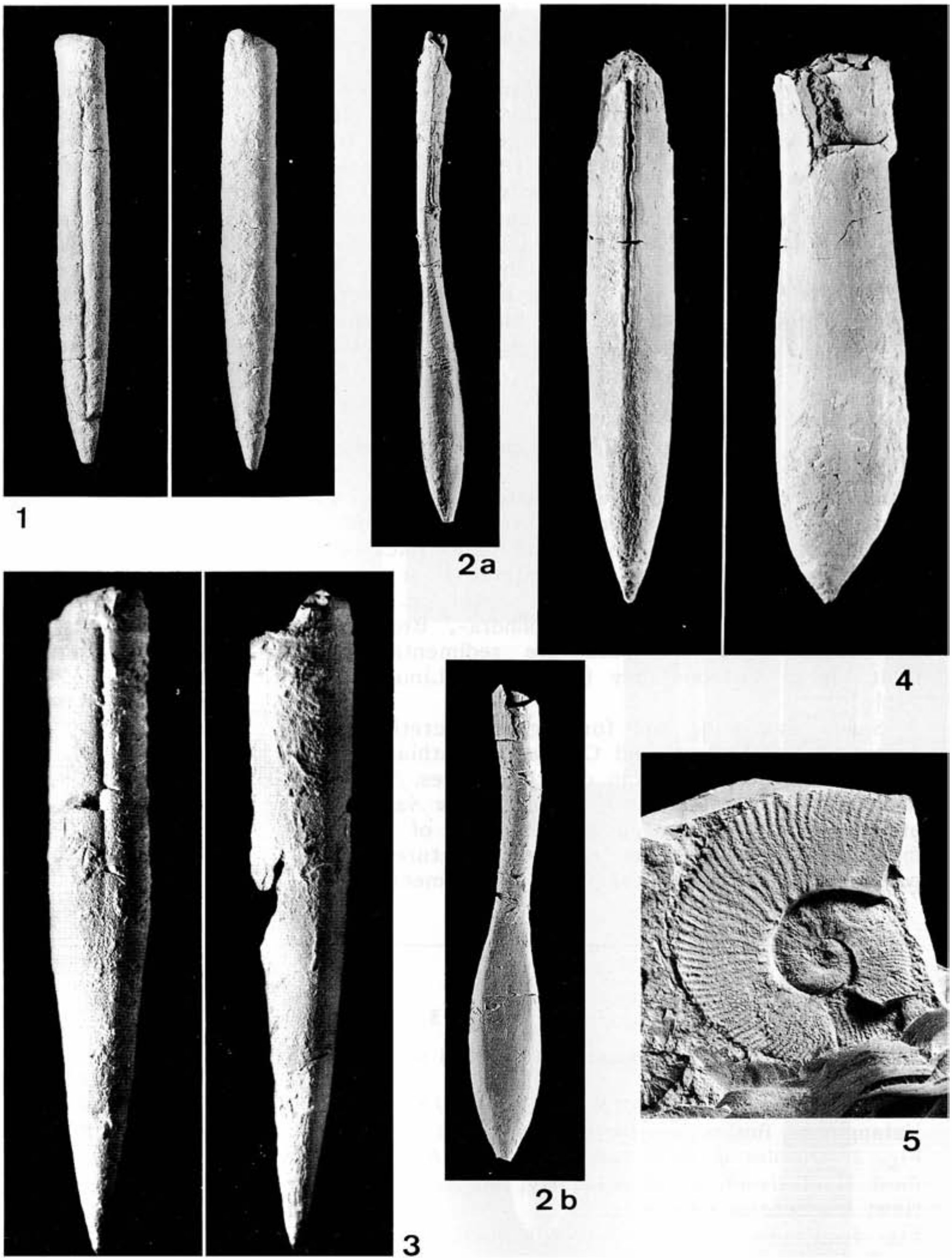
Plate 2

All specimens figured in natural size.

Fig. 1. *Aulacoteuthis* aff. *compressa* MUTTERLOSE, 1983. Probable Upper Hauterivian. Bralo quarry, marly layer at the base.
Fig. 2. *Duvalia binervia* (RASPAIL 1829). Lower/Upper Barremian boundary interval. Butkov quarry, 8th level, 70 m; a - dorsal; b - lateral view.
Fig. 3. *Mesohibolites garshini* STOYANOVA-VERGILOVA, 1965. Lowermost Barremian. Butkov quarry, 8th level, 250 m. Left: ventral; right: lateral view.
Fig. 4. *Duvalia grasiana* (DUVAL-JOUVE 1841). Lower Barremian. Butkov quarry, 8th level, 120 m. Left: dorsal, right: lateral view.
Fig. 5. *Pseudothurmannia mortiletti* (PICTET & LORIOL 1858). Lower Barremian. Polomec quarry, 5th level, 250 m, bed no. 10.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2a
2b
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index species *P. mortilleti* (PICT. & LOR.)), barremitid ammonites, and the last aptychi *Lamellaptychus angulicostatus* (PICT. & LOR.) become abundant again.

The Barremian part of the Mráznica Limestone contains a rare ammonite fauna dominated by *Barremites (Barremites)* sp. and *B. (Reboulites)* sp., together with rare *Holcodiscus* cf. *perezianus* (D'ORB.), *Hamulinites parvulus* (UHL.) and *Acrioceras tabarelli* (AST.). The upper part of the limestone contains the late Barremian index species *Silesites seranonis* (D'ORB.). The upper boundary of this limestone complex is of diachronous nature; the ammonite *Costidiscus recticostatus* (D'ORB.) in Malá Fatra Mts. (Medziholie section) dates it as late Barremian. The early Aptian *Deshayesites* ex gr. *involutus* SPATH occurs in the top limestone layer of the Horná Poruba section (Strážovské Vrchy Mts.) of the same formation. Towards the deeper part of the basin, the Mráznica Limestone passes into dark cherty limestones.

6. Aptian/Albian paleogeographical changes

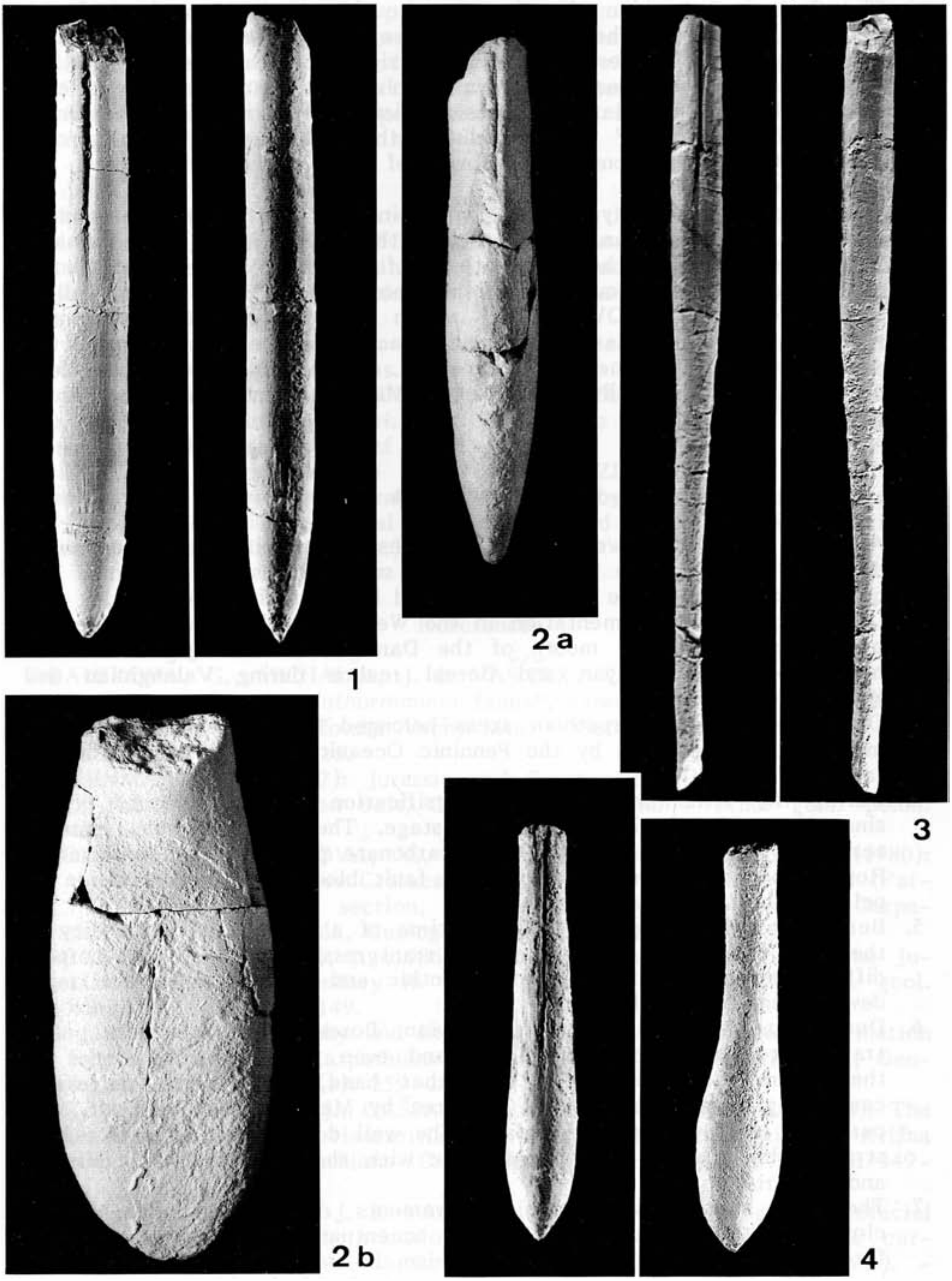
The Western Carpathians were affected during the Aptian by diastrophic events, which are indicated by breccias, turbidite beds, paraconglomerates (MICHALÍK & VAŠÍČEK 1984) and basic volcanics and their tuffs. Blocking of marine currents by a dissected sea-floor topography caused enlargement of the area of anoxic black shale deposition. Pelitic distal flysch sedimentation (Veřovice-, Kapušnica-, Koňhora-, Brodno formations) dominated in depressions, while the carbonate sedimentation was limited to elevated fault blocks surfaces only (Chmielowa Limestone Formation, Text-Fig. 2, Table 1).

Space shortening and forming of accretionary prisms characterized the contact zone of Outer and Central Carpathians. Coarse conglomerate masses contain clasts derived from deformed zones. These pebbles are the only evidence for the existence of several, now vanished, subducted areas (MIŠÍK et al. 1976, 1980). Complicated mobility of this zone, as well as the fragmentary knowledge of its intricate structure make the understanding of its paleogeography and paleotectonic development difficult.

Plate 3

All specimens figured in natural size.

Fig. 1. *Hibolites cigaretus* STOYANOVA-VERGILOVA, 1965. Uppermost Valanginian. Butkov quarry, 8th level, 510 m.
Fig. 2. *Duvalia dilatata majoriana* STOYANOVA-VERGILOVA, 1965. Uppermost Hauterivian. Polomec quarry, 4th level, 40 m, bed no. 17; a - dorsal view; b - lateral view.
Fig. 3. *Mesohibolites gladiiformis* (UHLIG 1883). Upper Barremian. Butkov quarry, 4th level, 120 m (with epirostrum preserved in the lower part). Left: ventral view; right: lateral view.
Fig. 4. *Hibolites mirificus* STOYANOVA-VERGILOVA, 1965. Lower Barremian. Butkov quarry, 3rd level, 210 m. Left: ventral view; right: lateral view.


1
2 a
2 b
3
4

The Central Carpathian situation was roughly similar to that of the Outer Carpathian one. The depressions have been filled by dark pelites with intercalations of limestone breccia, terrigene and volcanogene clastics. On the other hand, carbonate platforms evolved uninterrupted on the elevated fault blocks. The platform cores consisting of pale organogene limestones of the "Urgonian" type, produced thick slope debris (Haligovce-, Podhorie-, Muráň formations). The growth of bioherms continued until the end of the early Albian.

The most explicit bathymetrical change in the Central Carpathian area occurred in the mid-Albian. A collapse of the Tatric and Fatric Domains caused submergence of the carbonate platforms deep below the photic zone. Removal of shallow-marine barriers exposed the area to cold upwelling currents (MICHALÍK & KOVÁČ 1982), which stopped carbonate sedimentation in all the Western Carpathians and adjacent Europe. Subsequently, the formerly diversified sedimentary area was covered by monotonous deep marine pelitic marls and siltstones (Zabijak Marls, Butkov Formation, Poruba Formation) (Table 1).

7. Conclusions

1. All the preserved Lower Cretaceous West Carpathian sequences were deposited in a marine environment on several crustal fragments. Sequences of oceanic type are not preserved in the surface outcrops.
2. Lower Cretaceous sedimentation in the West Carpathians was influenced by a proximity of the mouth of the Danian-Polish Trough which connected both the Tethyan and Boreal realms during Valanginian and Hauterivian.
3. Outer and Central Carpathian areas belonged to different crustal segments, being separated by the Penninic Oceanic Domain until mid-Cretaceous.
4. During the Tithonian the first diversification of environments occurs since the mid-Jurassic deep-marine stage. The Tithonian was characterized by the origin of short-lived carbonate platforms, by Ammonitico Rosso type sedimentation on elevated fault blocks and by deposition of pelagic micrites in depressions.
5. Berriasian and Valanginian were the time of a new facies uniformity in the West Carpathians connected with transgressions. The lack of bottom differentiation caused scarcity of benthic and nektonic life, but rapid development of the plankton.
6. During late Valanginian and Hauterivian, Boreal faunal elements penetrated into the Outer Carpathians, and even into peripheral zones of the Central Carpathians. On the other hand, a moderate regression caused rapid colonization of all the area by Mediterranean nektonic and, partially, also by benthic organisms. The well documented ammonite biostratigraphic scale is well correlatable with those elaborated in France and Bulgaria.
7. The first inexpressive tectonic movements during Barremian caused closing of the Danian-Polish Trough, accentuation of bathymetric differences and start of submarine volcanism in the West Carpathians. Carbonate platforms occur on some elevated fault blocks (Tatric and Fatric Domains).
8. During the Aptian and Albian, the West Carpathians were affected by

diastrophism. Accretionary prisms of clastic material started to form in the contact zone of the Outer and Central West Carpathians, while black shale sedimentation prevailed in the basins.

9. The most explicit bathymetric change happened in the mid-Albian. Carbonate sedimentation stopped, distal flysch and pelitic terrigene sequences were deposited in the basins.

Acknowledgements. We dedicate this paper to the memory of our friend and collaborator Dr. KAROL BORZA (deceased in December 1985). He substantially contributed to refining of Lower Cretaceous microbiostratigraphic scale based on tintinnids, cadosinas, stomiosphaeras and colomisphaeras in the Western Carpathians.

Writing of this paper was stimulated by Prof. Dr. JOST WIEDMANN, who gave us valuable comments.

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The plate photographs are made by K. MEZIHORÁKOVÁ, Ostrava. All specimens are figures in natural size, whitened by ammonium chloride prior to photographing. The material will be deposited in the Slovakian National Museum (Natural History Dept.) in Bratislava.

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Age and Facies of the Calpionellid Formations from the South Carpathians

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With 3 Text-Figures

POP, G. (1989): Age and Facies of the Calpionellid Formations from the South Carpathians. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 525-542. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The sedimentary cover of several tectonic units from the South Carpathians, derived from the deformed European continental margin, includes late Tithonian to Hauterivian sediments. The sedimentary record consists of pelagic limestones of the Biancone type with allodapic calciturbidite interbeds in their lower part, of marly formations with interbeds of argillaceous pelagic micrites and of arenitic-calcareous flysch. All sediments of this time-span bear calpionellids.

Detailed correlation of formations is based on both calpionellid zones and subzones, such as: Crassicollaria intermedia and Cr. brevis subzones in the Crassicollaria Zone; Calpionella alpina, Remaniella and C. elliptica subzones in the Calpionella Zone; Calpionellopsis simplex, Cs. oblonga and Praecalpionellites murgeanui subzones in the Calpionellopsis Zone; and Calpionellites darderi and Tintinnopsella subzones in the Calpionellites Zone.

The carbonate and marly sediments were formed in subsiding basins, particularly in the basin slopes, between the compensation depths for calcite and aragonite. The pelagic sediments consist mainly of calcareous nannoplankton. Since the late Berriasian, the argillaceous influx suggests a relative sea-level fall and the diminution or cessation of the differential subsiding movements.

The basinal deposition includes also the resedimentation of the shallow water-derived carbonate sediments and the pelagic ones especially by turbidity currents and slumping. The shallowing facies of the overlying Hauterivian and Barremian formations suggests the first Cretaceous regional compressional tectonics preceding the Austrian Phase.

The sandy-calcareous flysch (Sinaia Formation) indicates an important intracontinental trough, developed on a stretched crust and bordered by emerged areas and shallow water carbonate platforms furnishing the very thick terrigenous and carbonate sediments. During the late Tithonian and early Berriasian the flysch sedimentation was accompanied by a weak basic volcanic activity.

Kurzfassung: Die sedimentären Deckschichten einiger tektonischer Einheiten der Süd-Karpathen, die vom deformierten europäischen Kontinentalrand stammen, enthalten Sedimente des Tithon bis Hauterive. Die Sedimente be-

stehen aus pelagischen Kalksteinen des Biancone-Typs, in deren tieferen Teil allodapische Kalkturbidite eingeschaltet sind, aus mergeligen Formationen, die mit tonigen pelagischen Mikriten wechsellagern, und aus sandig-kalkigem Flysch. Alle Sedimente dieses Zeitraums führen Calpionellen.

Die detaillierte Korrelation dieser Formationen basiert auf Calpionellen-Zonen und -Subzonen, und zwar den Crassicollaria intermedia- und Cr. brevis-Subzonen in der Crassicollaria-Zone; den Calpionella alpina-, Remaniella- und C. elliptica-Subzonen in der Calpionella-Zone; den Calpionellopsis simplex-, Cs. oblonga- und Praecalpionellites murgeanui-Subzonen in der Calpionellopsis-Zone und den Calpionellites darderi- und Tintinnopsella-Subzonen in der Calpionellites-Zone.

Die Karbonate und mergeligen Sedimente wurden in Becken mit rascher Subsidenz abgelagert. Die Ablagerung erfolgte vor allem im Bereich des Beckenhanges zwischen der ACD und CCD. Die pelagischen Sedimente bestehen vor allem aus kalkigem Nannoplankton. Für das späte Berrias kann aus der tonigen Sedimentation auf eine Absenkung des Meeresspiegels und eine Verminderung oder ein Ende der Subsidenz geschlossen werden.

Die Beckenablagerungen umfassen auch Resedimente von Flachwasser- und pelagischen Karbonaten. Die Resedimentation erfolgte vor allem durch Turbidite und Slumping. Die überlagernden Formationen des Hauterive und Barreme zeigen einen zunehmend flachmarinen Einfluß; sie lassen auf erste kompressive Bewegungen in der Kreide schließen, die der Austrischen Phase vorangegangen sind.

Der sandig-kalkige Flysch (Sinaia-Formation) weist auf einen intrakontinentalen Trog hin. Dieser entwickelte sich auf einer ausgedünnten Kruste und war durch herausgehobene Gebiete und Karbonatplattformen begrenzt, die das Liefergebiet für die sehr mächtigen klastischen und karbonatischen Trogsedimente bildeten. Während des Oberen Tithon und frühen Berrias wurde die Flyschablagerung von einem schwachen basischen Vulkanismus begleitet.

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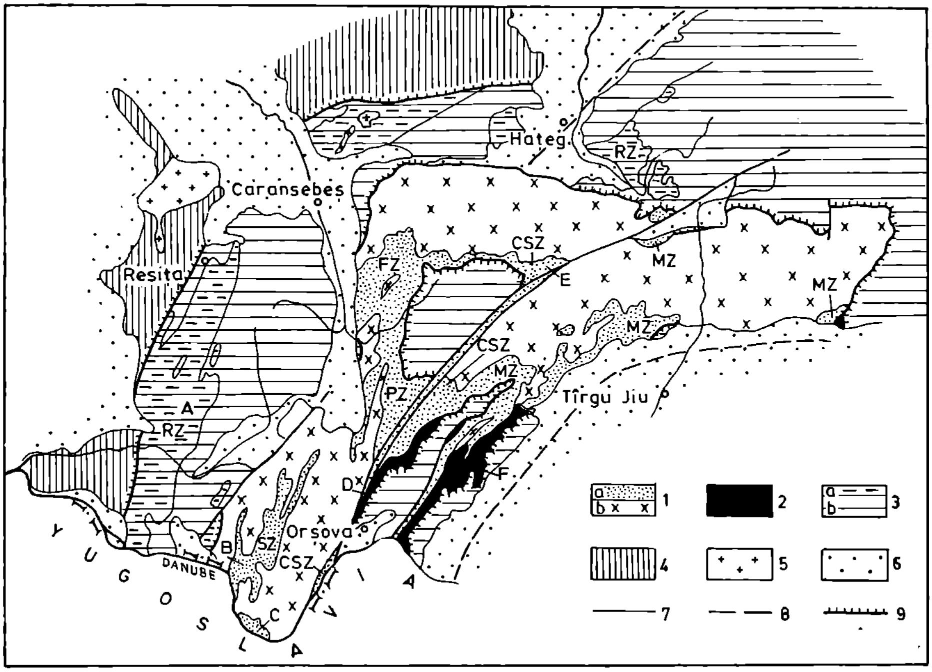
1. Introduction

The calpionellid-bearing deposits from the South Carpathians, representing the Upper Tithonian - Valanginian and possibly also part of the Hauterivian, are of Biancone or Maiolica type. In some areas their upper part is replaced by marly formations with more or less argillaceous limestone interbeds, except for the Sinaia Formation which is built up by a sandy, partly calcareous flysch. It is worth noting that the basinal carbonate deposits in this area, especially those in the Reşiţa and Sirinia sedimentation zones, are the most favourable for the study of calpionellids and their stratigraphic distribution in the whole area of the Romanian Carpathians.

2. Calpionellid-bearing formations

In the South Carpathians, these formations occur in the sedimentary cover of some Alpine tectonic units, where the successions of the Jurassic and Lower Cretaceous deposits show some facies peculiarities which led to their



Text-Fig. 1. Distribution of the Jurassic and Cretaceous sedimentary formations in the South Carpathians (central and western parts).
1 - Danubian Domain (a - Jurassic and Cretaceous sedimentary cover; b - crystalline basement and Paleozoic sedimentary formations). 2 - Severin Unit including the Sinaia Formation. 3 - Getic Domain (a - Jurassic and Cretaceous sedimentary cover; b - crystalline basement and pre-Liassic sedimentary formations). 4 - Supragetic Units. 5 - Banatites (Upper Senonian and Lower Paleogene magmatites). 6 - Tertiary sedimentary formations. 7 - Fault. 8 - Supposed fault. 9 - Overthrust.
RZ - Reşiţa Zone; SZ - Sirinia Zone; FZ - Feneş Zone; PZ - Presacina Zone; CSZ - Cazane-Stanuleţ Zone; MZ - Mehedinţi Zone.
Location of the examined sequences: A - Marila Limestones and Crivina Marls; B and C - Murguceva Limestones; D - Bîrza and Cerna formations; E - Scorota Limestones; F - Sinaia Formation.

grouping in the so-called "sedimentation zones" (CODARCEA 1940). Thus the calpionellids occur in the Marila Limestones and the Crivina Marls from the Reşiţa Zone (Getic Nappe or Median Dacids - according to SANDULESCU 1984), the Murguceva Limestones from the Sirinia Zone, the Bîrza Limestones and the Cerna Formation of the Dubova Unit as well as the Scorota Limestones in the Cazane-Stanuleţ Zone (Text-Fig. 1). All of them belong to the Danubian Domain (Marginal Dacids). Calpionellids also occur in the Sinaia Formation of the Severin Unit (External Dacids) (Text-Fig. 1). Although calpionellids were noticed in the pelagic limestones of the Presacina Zone (Danubian Domain, NASTASEANU 1979a), they have never been illustrated or compared with standard zones, so that they cannot be treated in this paper.

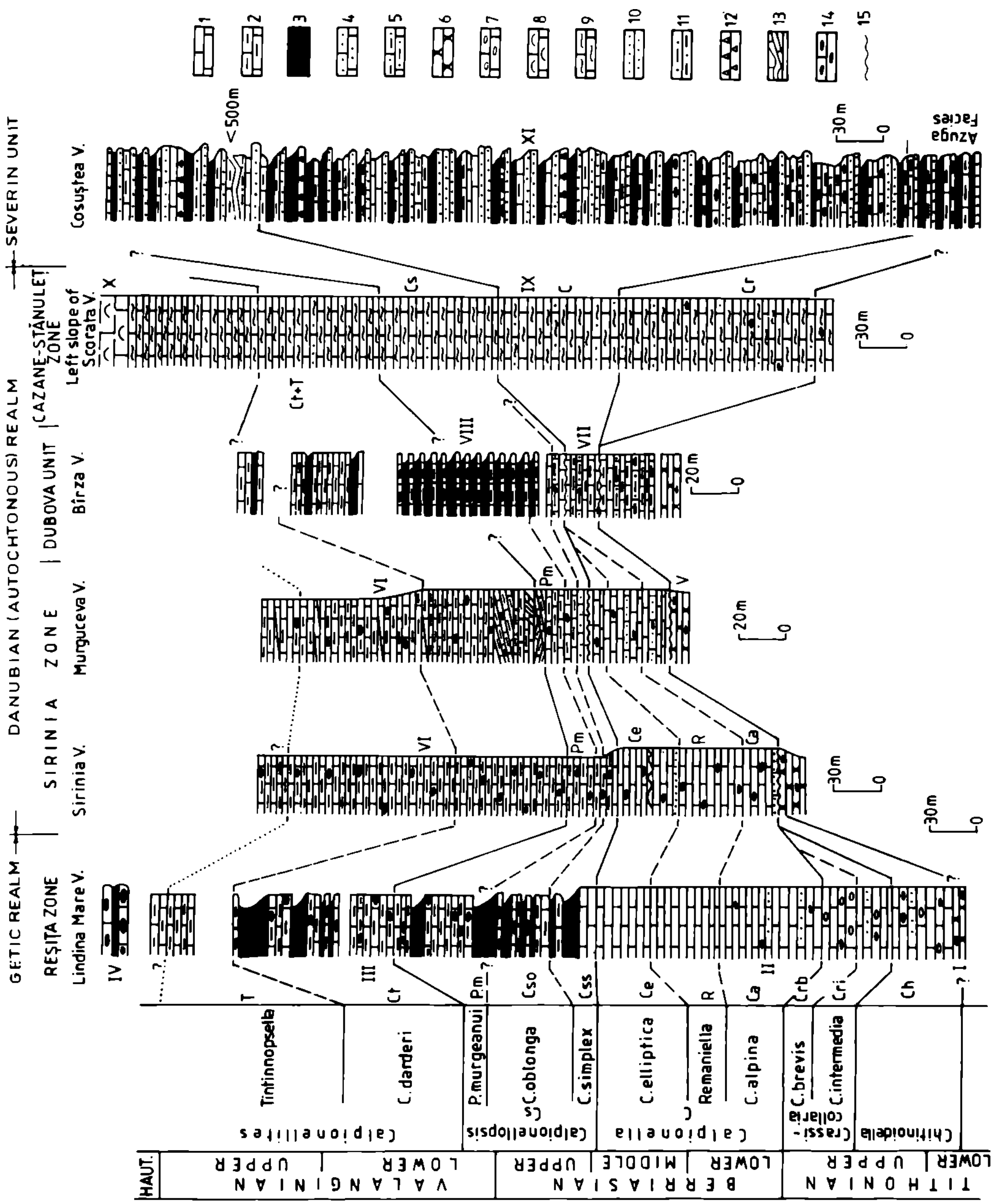
2.1 Marila Limestones and Crivina Marls

These two successive formations appear in the Reşiţa Zone and cover the Upper Tithonian-Valanginian interval, possibly also the basal Hauterivian (RAILEANU et al. 1957, NASTASEANU 1979b) (Text-Fig. 2).

The Marila Limestones (100-200 m thick) overlie conformably the Upper Kimmeridgian-Lower Tithonian nodular limestones (Bradet Limestones) and underlie accordingly the Crivina Marls. These limestones, showing a typical Biancone facies are essentially formed of whitish and cream-coloured micrites and biomicrites, with rare nodular cherts, representing partly resedimented pelagites. The micritic mass consisting of neomorphosed skeletal calcareous nannoplankton includes calcitized radiolaria, *Globochaete alpina,* calpionellids, calcisphaerulids, rare aptychi, thin-shelled bivalves, foraminifers, belemnites and, in their basal part, *Saccocoma* sp. The micritic mosaic is variously neomorphosed in the lower part of these limestones, in the Crassicollaria Zone inclusively, and more homogeneous and finer crystalline in their median and especially upper parts where the micrites are slightly argillaceous. In this latter part the Marila Limestones comprise occasionally centimetre-thick marly interbeds. In the lower part of these limestones, the micrites contain micritic intraclasts. If these clasts are more frequent, they form interbedded subnodular and rarely nodular limestones (intramicrudites), which are sometimes weakly dolomitized. The same part is marked also by the occurrence of rare allodapic peloid and/or bioclastic calcarenite-interbeds (calciturbidites) of grainstone, packstone and rarely wackestone type. The granular components consist of peloids, fragments of crinoids, neomorphosed bivalves, algal crusts, calcareous algae, corals and rarely bryozoans (often as cortoids), rare foraminifers, thin-shelled bivalves, oncoids and ooids. The components were originated in a shallow subtidal environment. Micritic intraclasts containing pelagic microorganisms are subordinately associated.

The lithological monotony of the Marila Limestones is generally uniformly stratified, the medium thickness of the beds being of 20-30 cm. Locally these limestones form slump folds (POP 1976b).

Text-Fig. 2. Correlation of the calpionellid-bearing formations of the South Carpathians.
I - Bradet (nodular) Limestones; II - Marila Limestones; III - Crivina Marls; IV - limestones with large ellipsoidal cherts and marls; V - nodular limestones; VI - Murguceva Limestones; VII - Bîrza Limestones; VIII - Cerna Formation; IX - Scorota Limestones; X - Urgonian type limestones (Barremian - Aptian); XI - Sinaia Formation (calpionellid-bearing sequence).
1 - pelagic micrites and biomicrites; 2 - argillaceous micrites and biomicrites; 3 - siltic marls and clays; 4 - allodapic calcarenites and calcirudites; 5 - argillaceous calcarenites and calcirudites; 6 - more or less argillaceous nodular micrites and biomicrites; 7 - subnodular micrites (intramicrudites); 8 - Urgonian-type limestones (anchimetamorphosed); 9 - anchimetamorphosed basinal limestones; 10 - sandstones; 11 - argillaceous sandstones and siltstones; 12 - breccias (mainly debris-flows); 13 - slumps and slump-folds; 14 - cherts; 15 - hiatus.


GETIC REALM
REŞIŢA ZONE
Lindina Mare V.
DANUBIAN (AUTOCHTONOUS) REALM
SIRINIA ZONE
Sirinia V.
Murguceva V.
DUBOVA UNIT
Bîrza V.
CAZANE-STĂNULEŢ ZONE
Left slope of Scorota V.
SEVERIN UNIT
Cosuştea V.
Azuga Facies
<500m
30m
20m
TITHONIAN
LOWER
UPPER
BERRIASIAN
MIDDLE
VALANGINIAN
HAUT.
Chitinoidella
Crassicollaria
Calpionella
Calpionellopsis
Calpionellites
C.intermedia
C.brevis
C.alpina
Remaniella
C.elliptica
C.simplex
C.oblonga
P.murgeanui
C.darderi
Tintinnopsella

The Crivina Marls (200-300 m thick) consist of an irregular alternation of argillaceous or sometimes silty marls (hemipelagites) and more or less argillaceous micrites (pelagites - hemipelagites). They form decimetre-thick beds that may be single or grouped in intervals. In places there is a cyclic alternation of marls and micrites.

Argillaceous micrites, sometimes containing nodular cherts, are formed of skeletal calcareous nannoplankton particles including calcitized radiolaria, calpionellids, calcisphaerulids, rare foraminifers, thin-shelled bivalves, and ammonite opercules (POP & IONESCU 1972, unpubl. rep.). Locally, more or less sandy allodapic calcarenites (calciturbidites) occur, which contain carbonate particles originated in the areas of the adjacent carbonate platforms. The calcarenites contain lamellibranchs, brachiopods, bryozoans, corals, hydrozoans, calcareous algal fragments, benthonic foraminifers, ooids, and oncoids. Calcareous sandstones are also occurring, bearing similar carbonate grains. Sometimes these sandstones include micritic intraclasts (BUCUR & OROS 1987).

The Crivina Marls underlie conformably a carbonate complex reaching maximum thickness of 200-250 m. It consists of argillaceous bioclastic and peletal micrites or biomicrites with large nodular cherts and interbedded marls assigned to the Hauterivian (NASTASEANU & DINCA 1962). It shows a transitional facies between basinal deposits and the Urgonian limestone (Barremian - Lower Aptian), a typical carbonate platform facies (POP & IONESCU 1972, unpubl. rep.).

2.2 Murguceva Limestones

In the Sirinia Zone the deposits containing calpionellids are represented by the Murguceva Limestones assigned to the Upper Tithonian (pro parte) - Hauterivian interval. These limestones - reaching 60 to about 300 m in thickness - overlie the Upper Kimmeridgian - basal Upper Tithonian nodular limestones and underlie the Upper Hauterivian - Lower Barremian marly Svinița Formation (RAILEANU 1953, AVRAM 1976, POP 1986a).

The Murguceva Limestones exhibit a Biancone-type facies and consist mainly of micrites and biomicrites, especially pelagites, which are whitish and cream-coloured in the lower part and gray, becoming gradually darker and even more argillaceous, in their upper part. Here, they include thin marl interbeds (hemipelagites) and show frequently bioturbation traces. In their lower part, calcarenites and calcirudites of grainstone, packstone and rarely wackestone type (calciturbidites) appear, sometimes with graded bedding and/or parallel or oblique lamination. They consist of carbonate grains formed in an area of a subtidal, shallow carbonate platform. The calciturbidites contain pelagic intraclasts. In places, carbonate debris-flows also occur in the lowermost part of this formation (POP 1986a). The micritic limestones and the calcarenites are petrographically very comparable to those known in the Marila Limestones.

2.3 Bîrza and Cerna formations

In the sedimentary cover of the Dubova Unit, the calpionellid-bearing deposits are represented by the Bîrza Formation (Tithonian - Upper Berriasian) and the Cerna Formation (Upper Berriasian - ?Lower Hauterivian) initially described as the "Grey Argillaceous Formation" (POP 1988).

The Bîrza Formation (about 35 m thick) overlies the Toplet Formation of black-shale facies (Oxfordian - Kimmeridgian) and underlies the Cerna Formation. Lithologically, the Bîrza Formation consists of grey and yellowish thin-bedded more or less argillaceous micrites and subordinately biomicrites (pelagites - hemipelagites) including interbedded calcarenites (calciturbidites) with nodular, lenticular and stratiform cherts. At the base of this formation (the first 8 m-thick interval), the argillaceous micrites and biomicrites exhibit a subnodular facies. Generally the limestones of the Bîrza Formation are made up of the same carbonate particles as those of the Marila Limestones.

The Cerna Formation consists of grey marly, argillaceous and silty deposits with subordinate thin-bedded or lenticular (boudined) argillaceous micrites, grouped sometimes in metre-thick intervals.

2.4 Scorota Limestones

In the Cazane-Stanuleţ Zone from the Retezat Mountains the limestones contain very scarce calpionellids. The carbonates form an about 500 m thick sequence within an over 1000 m thick succession of Upper Jurassic - Hauterivian basinal limestones, here named Scorota Limestones. They overlie conformably the sandy limestones assigned to the Middle Jurassic and underlie the prevailingly algal limestones (Barremian - Aptian) of Urgonian type (MORARIU & MORARIU 1977, POP 1986c). The basinal limestones are lithologically monotonous, most of them having evolved to "pseudomicrosparites" through Alpine dynamic anchimetamorphism. Aggrading recrystallization of these limestones was very inhomogeneous, so that calpionellids, calcisphaerulids, *Saccocoma* and aptychi fragments could be identified at some stratigraphic levels. In these limestones of Biancone type usually medium- to coarse-bedded allodapic calcarenites and calcirudites (calciturbidites) occur, containing carbonate grains originating in a shallow carbonate platform (possibly the Mehedinţi Threshold), among which echinoderms and peloids prevail. Carbonate breccias indicating debris flows are uncommon.

2.5 Sinaia Formation

In the Severin Unit, the calpionellid-bearing interval is represented by the lower part of the Sinaia Formation (over 1000 m thick) assigned to the Tithonian - Neocomian. These deposits, showing a flysch facies, crop out in the external part of the Danubian Domain. They overlie tectonically the Upper Cretaceous terrigenous formations and underlie in the same manner the Precambrian metamorphic Sebeş-Lotru Group of the Getic Nappe.

The Sinaia Formation consists of very thin- to very thick-bedded, grey, cleavaged argillaceous or marly siltites, arkose and lithic arenites, sublitharenites, feldspathic and lithic graywackes, calcareous arenites and graywackes, sandy calcarenites and more or less argillaceous or siltic grey to blackish micrites, which contain in its lower parts nodular cherts. Breccias (debris-flows) with pebbles mainly consisting of gneiss and quartzites are found at several stratigraphic levels (Text-Fig. 2). Except for some micrites representing pelagites/hemipelagites, the other rock types are almost exclusively turbidites including calciturbidites.

A peculiar facies occurs at the base of this formation, known as the Azuga Beds, characterized by the presence of partly silicified phyllite-like, greenish, grey and reddish argillaceous siltstones and of some basic rocks as stratiform bodies (basalts, spilites) associated with interbeds of sandstones and limestones of the above-mentioned types.

These usually diaclased deposits are intensely deformed so that the reconstitution of the normal succession of this formation is not possible. In the Coşuştea valley (outer Danubian Domain), where this formation reaches its maximum thickness, an about 50 m thick basal interval can be distinguished, exhibiting the Azuga Bed facies. It is followed by a thicker (200 m) interval where the argillaceous micritic interbeds are more frequent and contain cherts. This sequence is overlain by another 700-1000 m thick interval with fewer micrite interbeds and breccia levels (STANOIU 1978). The uppermost part of the Sinaia Formation, reaching several hundreds of metres in thickness consists almost exclusively of sandstones. The micritic intercalations contain calcitized radiolaria, rare foraminifers, calcisphaerulids, calpionellids, thin-shelled bivalves, *Globochaete alpina,* and very rare aptychi.

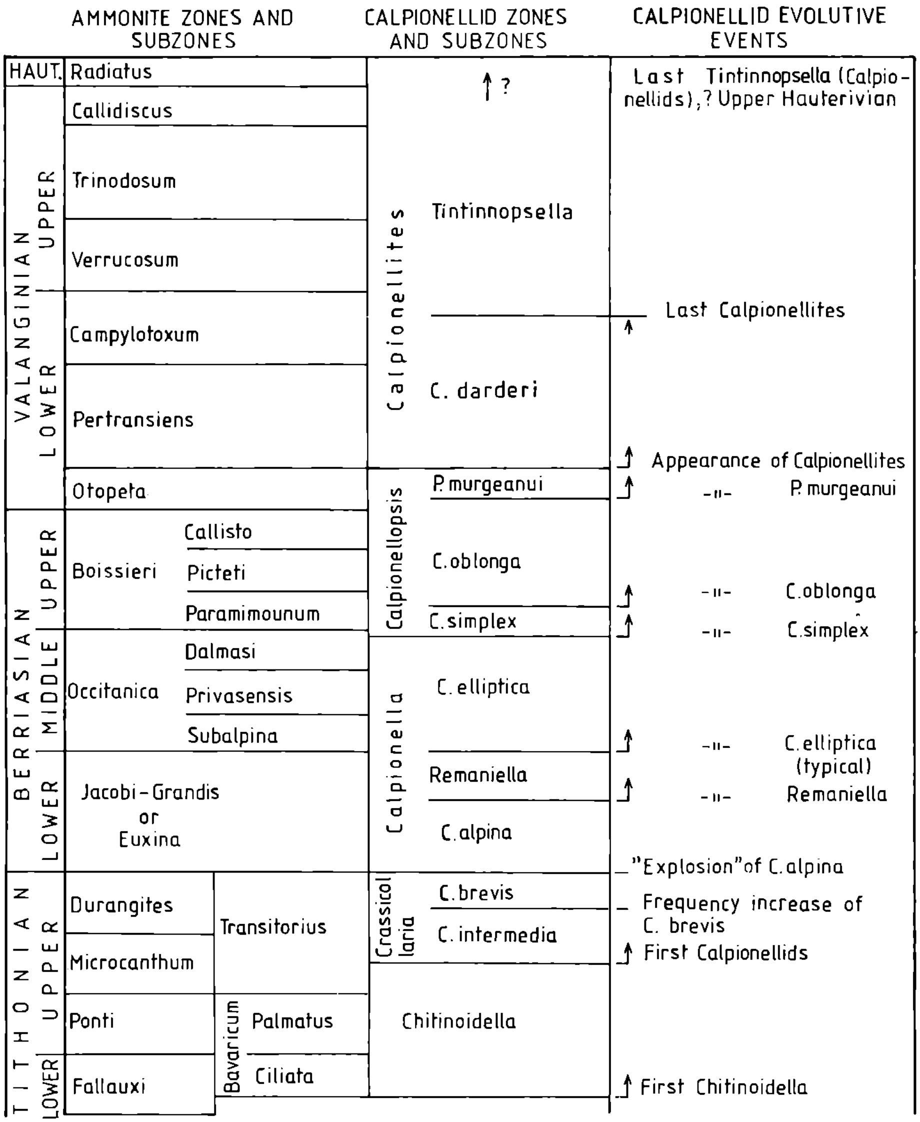
3. Zonal calpionellid distribution

The morphology and evolution of the calpionellids in the Tithonian-Hauterivian formations from the South Carpathians as well as their stratigraphic distribution are comparable to those known in other areas of the Tethyan Realm, showing the same biochronological significance.

Within these formations only few ammonite levels and intervals are known up to now, confirming the presence of some ammonite zones or subzones. Consequently, the author returns on the ammonite-calpionellid biozonal correlation, which was elaborated in formations of the same age in other Mediterranean areas (REMANE 1963, 1985, 1986, LE HÉGARAT & REMANE 1968, LE HÉGARAT 1971, ALLEMANN et al. 1971, 1975, ENAY & GEYSSANT 1975, BUSNARDO & THIEULOY 1979, ALLEMANN & REMANE 1979, HOEDEMAEKER 1982, 1987, REMANE et al. 1986, ZEISS 1986).

The standard calpionellid zones were first identified in the Marila Limestones and Crivina Marls (POP 1974) and subsequently in the same as well as in other formations (AVRAM 1976, 1984, MORARIU & MORARIU 1977, STANOIU 1978, NASTASEANU 1979a, POP 1980, 1986a, 1986b, 1988, ANTONESCU & AVRAM 1980, POP & MORARIU 1981, BUCUR et al. 1982, AVRAM et al. 1987, BUCUR & OROS 1987). Attempts for subzoning the standard zones were made by POP (1974, 1976, 1980, 1986a, 1986b, in press) such as: Crassicollaria intermedia and Cr. brevis subzones in the Crassicollaria Zone, Calpionella alpina, Remaniella and C. elliptica subzones in the Calpionella Zone, Calpionellopsis simplex, Cs. oblonga and Praecalpionellites murgeanui subzones in the Calpionellopsis Zone, and Calpionellites darderi and Tintinnopsella carpathica subzones in the Calpionellites Zone (Text-Fig. 3).

The previous data concerning the presence of the calpionellids in the involved formations are given in the above-mentioned papers.



Text-Fig. 3. Correlation of the ammonite and calpionellid zones and subzones.

3.1 Crassicollaria Zone (Upper Tithonian pro parte)

In this standard zone, defined by the occurrence of the calpionellids in the Microcanthum Zone and the base of the Calpionella Zone, two subzones are distinguished (POP 1974) probably representing two chronostratigraphic intervals of different time-span: the Crassicollaria intermedia Subzone in the lower part, corresponding approximately to the A_1 and A_2 subzones of REMANE (1963), and the Crassicollaria brevis Subzone in the upper part, possibly designating the A_3 Subzone of the same author.

The Crassicollaria Zone is completely represented only in the Marila Limestones. There, however, the calpionellids occur sometimes badly preserved due to more intense neomorphism in micrites. In the first basal 35-40 m of these limestones rare *Chitinoidella boneti* DOBEN and *Chitinoidella* sp. occur, indicating the Chitinoidella Zone. The next 40-45 m thick interval represents the Crassicollaria Zone.

In the Cr. intermedia Subzone, the index species is associated with *Crassicollaria parvula* REMANE, *Cr. massutiniana* (COLOM), very rare *Calpionella alpina* LORENZ, small *Tintinnopsella carpathica* (MURG. & FIL.) and, in the upper part, with *Crassicollaria brevis* REMANE.

The Cr. brevis Subzone is marked by the persistence of the same species, but a slight increase of the index species and the occurrence of *Crassicollaria colomi* DOBEN are noticed.

In the Murguceva Limestones, the Crassicollaria Zone occurs incompletely and only locally in an interval of 2 m maximum thickness at thê base of this formation that overlies the nodular or subnodular limestones containing in places *Chitinoidella* sp. The Crassicollaria Zone is represented by an assemblage corresponding probably to the basal part of this zone. The same zone occurs in a 125 m thick interval within the Scorota Limestones (MORARIU & MORARIU 1977) and probably in the basal part of the Sinaia Formation. According to the available data, this zone is missing in the Bîrza Limestones.

3.2 Calpionella Zone (Lower - Middle Berriasian)

This zone is determined by the "explosion" and morphological diversification of *Calpionella alpina* LORENZ at the base of the Jacobi-Grandis Zone and the appearance of the *Calpionellopsis* genus. The zone can be subdivided into the C. alpina, Remaniella and C. elliptica subzones by the first occurrence of *Remaniella* which is very close to the boundary between the Jacobi and Grandis subzones and the first occurrence of *C. elliptica* near or at the base of the Occitanica Zone (POP 1974, REMANE et al. 1986).

The Tithonian - Berriasian boundary (i. e. the Jurassic - Cretaceous boundary) is drawn here at the base of the Calpionella Zone.

The Calpionella Zone is wholly or partially present in all the described successions, but in intervals of different thickness. On the other hand, its subzonation is possible, especially in the Marila, Murguceva and Bîrza formations.

In the Marila Limestones, the C. alpina Subzone interval (50-65 m thick) includes the index species (abundant) associated with *Crassicollaria parvula* REMANE, *Cr. massutiniana* (COLOM), very rare small *Tintinnopsella carpathica* (MURG. & FIL.) and, in its lower part, with scarce *Crassicollaria intermedia* (DURAND-DELGA), *Cr. brevis* REMANE, and *Cr. colomi* DOBEN.

In the Murguceva Limestones, the C. alpina Subzone (5-20 m thick interval) occurs locally and incompletely following an important erosive hiatus. A similar case has been noticed also in the Bîrza Limestones, where the interval of this subzone is 5 m thick. It contains the same calpionellid species and overlies the lower micritic part (17 m thick) with *Parastomiosphaera malmica* (BORZA) assigned to the Lower Tithonian. In the Sinaia Formation, the calpionellids are generally rare and badly preserved and sometimes show anomalous assemblages due to the frequent redeposition. Consequently only some assemblages, corresponding to both the C. alpina and Remaniella subzones, were identified (*Calpionella alpina* LORENZ prevalent, *Crassicollaria parvula* REMANE, *Cr.* aff. *massutiniana* (COLOM), *Remaniella* aff. *cadischiana* (COLOM), small *Tintinnopsella carpathica* (MURG. & FIL.)).

In the Remaniella Subzone interval (25-45 m thick) from the Marila Limestones occur rare *Remaniella cadischiana* (COLOM) together with abundant *Calpionella alpina* LORENZ, rare small *Tintinnopsella carpathica* (MURG. & FIL.), and *Crassicollaria parvula* REMANE. Very scarce *Cr.* aff. *massutiniana* (COLOM) are noticed in the lower part of this subzone.

The same assemblage containing also *Remaniella dadayi* (KNAUER) occurs in the Murguceva Limestones, where the interval of this subzone (5-20 m thick) in places overlies directly the Tithonian nodular limestones. This subzone cannot be determined in the Scorota Limestones and Sinaia Formation and is missing in the Bîrza Limestones.

Moulds of *Pseudosubplanites grandis* (MAZENOT) are present in a centimetre-thick marly intercalation within the Marila Limestones (east of Ciclova Montana). They correspond to the upper part of the Grandis Subzone.

The Calpionella elliptica Subzone is evident in a 25-45 m thick interval in the upper part of the Marila Limestones, where the index species (frequent) is associated with *Calpionella alpina* LORENZ (prevalent), medium to large-sized *Tintinnopsella carpathica* (MURG. & FIL.), *Remaniella cadischiana* (COLOM), *R. dadayi* (KNAUER), and small *Crassicollaria parvula* REMANE. In the uppermost part of the Elliptica Subzone occur *Tintinnopsella longa* (COLOM), *Lorenziella plicata* REMANE, and *Lorenziella* sp. with the above-mentioned fauna. Comparable assemblages have been noticed in the Murguceva Limestones (5-40 m thick interval), including sometimes slumps from the subjacent subzones. In the Bîrza Limestones were recorded similar assemblages, following an erosive hiatus which at least corresponds to the Remaniella Subzone. In the Sinaia Formation, the Elliptica Subzone also occurs.

3.3 Calpionellopsis Zone (Upper Berriasian - basal Valanginian)

This zone is delimited by the first occurrence of the genera *Calpionellopsis* and the one of *Calpionellites* near the base of the Boissieri Zone and at the base of the Pertransiens Zone respectively, and includes three subzones: Calpionellopsis simplex, Calpionellopsis oblonga, and Praecalpionellites murgeanui. These subzones are separated by the appearance of typical forms of *Calpionellopsis oblonga* (CADISCH) in the upper Paramimounum Subzone and the one of *Praecalpionellites murgeanui* (POP) approximately in the medium part of the Otopeta Zone (POP 1986a). The first calpionellid subzone of Calpionellopsis simplex probably corresponds to the D_1 Subzone of REMANE (1963).

The Calpionellopsis Zone occurs in all the studied successions. Its best representation is in the Murguceva Limestones, where the three subzones are separable.

The Calpionellopsis simplex Subzone, found in relatively thin intervals (3-16 m thick), includes the index species associated with *Calpionellopsis simplex/oblonga*, typical *Tintinnopsella carpathica* (MURG. & FIL.), rare *T. longa* (COLOM), *Remaniella cadischiana* (COLOM), *R. dadayi* (KNAUER), *Calpionella alpina* LORENZ, *C. elliptica* CADISCH, a small variety of *Crassicollaria parvula* REMANE and very rare *Lorenziella plicata* REMANE. In the interval of this subzone (right slope of the Murguceva valley) *Ptychophylloceras ptychoicum* (QU.) and *Spiticeras* ex gr. *S. polyptroptychum* (QU.) are recorded (ANTONESCU & AVRAM 1980). A similar calpionellid assemblage occurs either in the Marila Limestones (30 m thickness, Valea Mîndrişagului), or in the terminal part of the same limestones (10 m thick interval) as well as in the basal part of the Crivina Marls (18 m thick interval, Lindina Mare valley). This points to a heterochronous boundary between the two formations. In the terminal levels of the Marila Limestones, east of Ciclova Montana, *Pseudosubplanites grandis* (MAZENOT), *Subthurmannia* cf. *boissieri* (PICTET) and *S. latecostata* (KILIAN) have been identified (MUTIHAC 1959). The last two species suggest that this lower part of the Boissieri Zone is also confirmed in a similar level in the Beu Sec valley (BUCUR et al. 1982). This subzone is also evident within several metres of the upper part of the Bîrza Limestones.

The Calpionellopsis oblonga Subzone occurs in 2-8 m thick intervals of argillaceous micrites in the Murguceva Limestones where the typical index species is associated with *Calpionellopsis simplex* (COLOM), *Tintinnopsella carpathica* (MURG. & FIL.), *T. longa* (COLOM), *T. (Amphorellina) subacuta* (COLOM) (very rare), *Remaniella dadayi* (KNAUER), *R. cadischiana* (COLOM), *Lorenziella plicata* REMANE, *L. hungarica* KNAUER & NAGY and accidentally small *Calpionella alpina* LORENZ. The basal part of this subzone has been noticed in the uppermost interval (the last 2-3 m) of the Bîrza Limestones. In the lower part of the Crivina Marls (maximum 110 m thick), the same subzone could not be separated from the next subzone. In the terminal part of the Marila Limestones from the Valea Minişului (Crivina) - probably corresponding to the base of the Oblonga Subzone - *Spiticeras (Negreliceras) paranegreli* was identified (AVRAM et al. 1987), a species known from the Boissieri Zone.

The Praecalpionellites murgeanui Subzone, indicated by the distribution of calpionellids within the Murguceva Limestones (12-20 m thick interval) and also found in comparable formations of Cuba (POP 1976a, 1986a, 1986b), comprises the index species (usually rare) associated with *Tintinnopsella carpathica* (MURG. & FIL.), *T. longa* (COLOM), *T. subacuta* (COLOM), *Remaniella dadayi* (KNAUER), *R. cadischiana* (COLOM), *Lorenziella hungarica* KNAUER & NAGY, *L. plicata* REMANE, *Calpionellopsis oblonga* (CADISCH) and sometimes *Cs. simplex* (COLOM).

3.4 Calpionellites Zone (Valanginian - Hauterivian pro parte)

This zone, defined by the occurrence of the index genus and the disappearance of the family Calpionellidae BONET, has been divided in two subzones: Calpionellites darderi and Tintinnopsella (TREJO 1975, POP 1980). These subzones are delimited by the disappearance of the genus *Calpionellites* approximately within the Campylotoxum Zone. The upper boundary of the Tintinnopsella Subzone is still in discussion. According to some data, it might be placed in the basal Upper Hauterivian (VAŠÍČEK et al. 1983, MICHALÍK & VAŠÍČEK, this vol.). Although the Calpionellites Zone is presented here according to the original definition (ALLEMANN et al. 1971), it is worth noting that the two mentioned subzones contain calpionellid assemblages which are different enough to allow not only a separation but they might also be considered as standard zones, namely, a Calpionellites Zone subdivided into two subzones (Calpionellites darderi and Ct. major subzones) and a Tintinnopsella Zone. The latter zone has been separated also by TREJO (1975) and BORZA (1984).

The Calpionellites Subzone is found in all the presented successions, except for the Sinaia Formation which essentially consists of sandstones in the corresponding interval.

The Calpionellites darderi Subzone from the Murguceva Limestones (10-30 m thick interval) includes *Calpionellites darderi* (COLOM), *Ct. major* (COLOM), *Ct. coronata* TREJO, *Ct. caravacensis* ALLEMANN, *Tintinnopsella carpathica* (MURG. & FIL.), *T. longa* (COLOM), *T. subacuta* (COLOM), *Praecalpionellites murgeanui* (POP), and *P. siriniaensis* POP. It contains in its lower part *Remaniella dadayi* (KNAUER), *R. cadischiana* (COLOM), *L. hungarica* KNAUER & NAGY, *L. plicata* REMANE, and sometimes extremely rare *Calpionellopsis oblonga* (CADISCH). The same *Calpionellites* and *Tintinnopsella* species were found also in the uppermost carbonate levels of the Cerna Formation from the Toplet area. In the Crivina Marls, this subzone contains only very rare *Calpionellites darderi* (COLOM) and species of the other above-mentioned genera.

In the middle of the Ct. darderi Subzone interval (50 m thick) from the Murguceva Limestones (right slope of the Murguceva valley) *Kilianella* aff. *roubaudiana* (D'ORB.) and *K. roubaudiana retrocosta* SAYN were found, indicating a Lower Valanginian age. Near its upper boundary *Olcostephanus* cf. *catulloi* (RODIGHIERO) was noticed (AVRAM 1976, 1984, ANTONESCU & AVRAM 1980). In the Crivina Marls from the Beu Sec valley, which probably represent the same subzone, *Thurmanniceras* cf. *pertransiens* (SAYN), *Neocomites pycnoptychus* UHLIG and *Kilianella bochianensis* SAYN, known in the Lower Valanginian, were identified (BUCUR et al. 1982, AVRAM et al. 1987). The same biochronologic interval is supported also by the presence of *Neocomites neocomiensis* (D'ORB.) in the ?middle part of the Crivina Marls from the Glava hill (MUTIHAC 1959).

The Tintinnopsella Subzone, the poorest one in calpionellids, comprises only *Tintinnopsella carpathica* (MURG. & FIL.), very rarely accompanied by *T. longa* (COLOM).

Within the interval of this subzone (45 m thick) of the Murguceva Limestones (right slope of the Murguceva valley) several ammonite species were noticed: *Olcostephanus* cf. *sayni* (KILIAN) at a level placed 5 m above the lower limit of the subzone. It occurs very close to the level where *Colomisphaera echinata* (NOWAK), and *Neocomites (Teschenites) pachydicranus* THIEULOY are present. These species indicate probably the Lower Hauterivian, about 18 m above the said boundary. *Crioceratites duvali* LEV., *C. matsumotoi* (SARKAR), *C. nolani* (KILIAN), *C. mandovi* nom. nov. (= *C. villersianum* var. *bituberculata* SARKAR), *C. majoricensis* NOLAN, *Paraspinoceras pulcherrinum* (D'ORB.), and *Spitidiscus* cf. *incertus* (D'ORB.) occur in the uppermost 5-10 m. This ammonite assemblage might represent the Middle Hauterivian, the Sayni Zone included (AVRAM 1976, 1984, ANTONESCU & AVRAM 1980).

In the Crivina Marls (?middle part) of the Glava hill, MUTIHAC (1959) reported also *Bochianites neocomiensis* (D'ORB.), *Kilianella biformis* SAYN and small *Olcostephanus* specimens (species revised by PATRULIUS & AVRAM, in: PATRULIUS et al. 1976) indicating especially Upper Valanginian.

In the limestones containing large ellipsoidal cherts which overlie the Crivina Marls from the Beu Sec valley, *Acanthodiscus* cf. *radiatus* (BRUG. in D'ORB.) was identified (AVRAM et al. 1987). This is an index species for the base of the Hauterivian.

4. Interpretations and conclusions

In the South Carpathians several alpine (laramic) tectonic units are formed by the deformation of the European continental margin of the Tethys. The Upper Tithonian-Hauterivian deposits are represented by pelagic limestones of the Biancone type and marls with basinal carbonate interbeds, except for the Sinaia Formation (Severin Unit) which shows a sandy, partly sandy-calcareous flysch.

The study of the zonal distribution of calpionellids within these formations allows a relatively detailed correlation and consequently the corresponding dating of some depositional events. The correlation is based on both calpionellid zones and subzones, such as: Cr. intermedia and Cr. brevis subzones in the Crassicollaria Zone, the C. alpina, Remaniella and C. elliptica subzones in the Calpionella Zone, the Cs. simplex, Cs. oblonga and Praecalpionellites murgeanui subzones in the Calpionellopsis Zone, and the Ct. darderi and Tintinnopsella subzones in the Calpionellites Zone.

According to these data, the Biancone-type limestones occur in the time intervals corresponding either to the Upper Tithonian - Upper Berriasian (Marila and Bîrza Limestones) or to the Upper Tithonian - Hauterivian (Murguceva and Scorota Limestones). These basinal limestones overlie the Upper Kimmeridgian - Lower Tithonian reddish, greenish or grey nodular limestones, except for the Scorota Limestones which show a uniform facies in this interval. The Biancone-type limestones underlie marly formations (Crivina and Cerna formations) assigned to the Upper Berriasian (pro parte) - Hauterivian (pro parte), or to the Upper Hauterivian - Lower Albian (Svinita Formation), or to the Barremian - Aptian prograding Urgonian limestones (Cazane-Stanuleţ Zone). The Crivina Marls underlie a Hauterivian carbonate complex of shallowing facies which is overlain in their turn by the Barremian - Aptian prograding Urgonian limestones.

The Upper Tithonian to partially Upper Berriasian sequences of the pelagic limestones include allodapic calciturbidites, several erosive synsedimentary hiatuses, slumps and redeposited pelagic sediments, whose frequency obviously decreases toward the upper part. From the Cs. simplex Subzone (Upper Berriasian) on, these limestones are more and more argillaceous or replaced by marly formations, except for the anchimetamorphosed Scorota Limestones which are only slightly argillaceous. In these sequences the allodapic calciturbidites are extremely rare.

As a consequence of rifting at the beginning of the Jurassic, the block-faulting of the South Carpathians' continental margin took place leading to a system of synsedimentary hosts and grabens delimited by normal faults, and to longitudinal sedimentation areas or zones (BERNOULLI 1972). During the Middle Jurassic and/or the earliest late Jurassic the same block-faulting coupled with a regional but differential subsidence. The relative rise of the sea level determined the transition of the shallow-marine or continental sedimentation zones to subsiding deep basins with pelagic and hemipelagic deposition and shallow-water carbonate platforms and even emerged areas.

The increase of the calcareous nannoplankton during the Upper Tithonian -Hauterivian time interval created an abundant source of skeletal particles of carbonate oozes accumulated at depths between the carbonate compensation limits ACD and CCD. The high pelagic carbonate sedimentation rate determined the dilution of the siliceous sediments, provided especially by still frequent radiolaria, and of the argillaceous ones. In the same time-span, an additional source of carbonate oozes were the fine-grained essentially aragonitic and high-magnesian calcite sediments formed in the areas of carbonate platforms. The same areas furnished the medium- to coarse-grained carbonate sediments forming interbedded calcarenites and calcirudites. The source area of the shallow-water derived calcareous turbidites interbedded in the Marila Limestones and Scorota Limestones may at least partially be located in the platform areas of the Semenic and Mehedinţi thresholds bordering the Reşiţa and Cazane-Stanuleţ basins, respectively (POP 1973, 1976b).

As from the Upper Berriasian the influx of argillaceous sediments gradually increased in the basinal areas leading to the dilution of the calcareous oozes and the accumulation of some prevailing marly sediments. The argillaceous supply, suggesting a relative lowering of the sea level, coinçided with the distinct decrease or the absence of shallow-water carbonate sediment influx, except for Cazane-Stanuleţ area where the input of such sediments probably remained important.

The accumulation of carbonate and argillaceous sediments reaching important thicknesses took place especially in the areas of the basin slopes by pelagic sedimentation, resedimentation and possibly normal bottom currents. The redistribution of the sediments by the interaction of sliding, slumping, debris-flows and turbidity currents is evident particularly in the Upper Tithonian-Berriasian sequences and - at the same time - responsible for the hiatuses appearing in these limestones.

The basinal carbonate sedimentation was controlled by an active and differential subsidence probably dominated by still distensional tectonics. Beginning with the Upper Berriasian, the argillaceous sediment influx can be interpreted not only as a fall of relative sea level but also at least as a diminution of the differential subsiding movements. The shallowing facies of the Hauterivian carbonate complex in the Reşiţa Zone and the Upper Hauterivian-Lower Albian Sviniţa Formation in the Sirinia Zone generally indicate the cessation of the differential subsiding movements as a consequence of the first Cretaceous regional compressional tectonics. Subsequently the shallowing environments in the area of the former basins (Reşiţa and Cazane-Stanuleţ basins) caused the progradation of the Barremian-Aptian (Urgonian) carbonate platform.

The Sinaia Formation suggests a building up in a wide intracontinental trough with a stretched crust bordered by emerged areas and shallow carbonate platforms, furnishing an important bulk of terrigenous and carbonate sediments accumulated as terrigenous and calcareous flysch (SANDULESCU 1984). Beginning with the Upper Berriasian, all pelagic carbonate oozes were diluted by the terrigenous supply. The presence of basic rocks in the lower part of this formation shows a volcanic activity during the late Tithonian and the ?early Berriasian.

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Biofacies Characteristics of Lower Cretaceous Deposits of Georgia

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With 3 Text-Figures

KOTETISHVILI, E. V. (1989): Biofacies Characteristics of Lower Cretaceous Deposits of Georgia. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 543-550. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: An analysis of the areal distribution of the Lower Cretaceous ammonites allows to establish two ammonite assemblages characterizing different bathymetric levels of the early Cretaceous sea in Georgia. The difference between ammonite assemblages of the epineritic and infraneritic zones is as follows: In the lower part of the Lower Barremian the epineritic zone is characterized by *Paracrioceras*, the infraneritic zone by the genera *Holcodiscus, Spitidiscus, Astieridiscus*; in the upper part of the Lower Barremian shallow marine *Pulchellia* and *Heinzia* are replaced by deep-water *Subpulchellia*; at the top of the Upper Barremian shallow marine groups of *Colchidites intermedius* and *C. colchicus* do not occur together with the deep-water group of *Colchidites shaoriensis*. In the lower part of the Albian *Leymeriella* and *Douvilleiceras* are related to shallow marine facies.

Similar deep-water ammonite assemblages are known, e. g. from Armenia and Algeria.

Kurzfassung: In der Verbreitung der Unterkreide-Ammoniten Georgiens lassen sich zwei Ammonitenvergesellschaftungen erkennen, die verschiedene Bathymetrien des Unterkreide-Meeres charakterisieren können. Die Ammoniten der epineritischen Zone unterscheiden sich von den Ammoniten der infraneritischen Zone folgendermaßen: Im frühen Barreme ist die epineritische Zone durch die Gattung *Paracrioceras* charakterisiert, während die infraneritische Zone durch die Gattungen *Holcodiscus, Spitidiscus* und *Astieridiscus* gekennzeichnet ist; im höheren Unterbarreme werden die Flachwasserformen *Pulchellia* und *Heinzia* von der Tiefseegattung *Subpulchellia* abgelöst; im höchsten Oberbarreme besiedelten die Flachwasserarten *Colchidites intermedius* und *C. colchicus* und die tiefmarine Gruppe des *C. shaoriensis* getrennte Areale. Im unteren Alb sind die Gattungen *Leymeriella* und *Douvilleiceras* auf flachmarine Fazien beschränkt.

Ähnliche tiefmarine Ammonitenvergesellschaftungen sind in Armenien und in Algerien zu beobachten.

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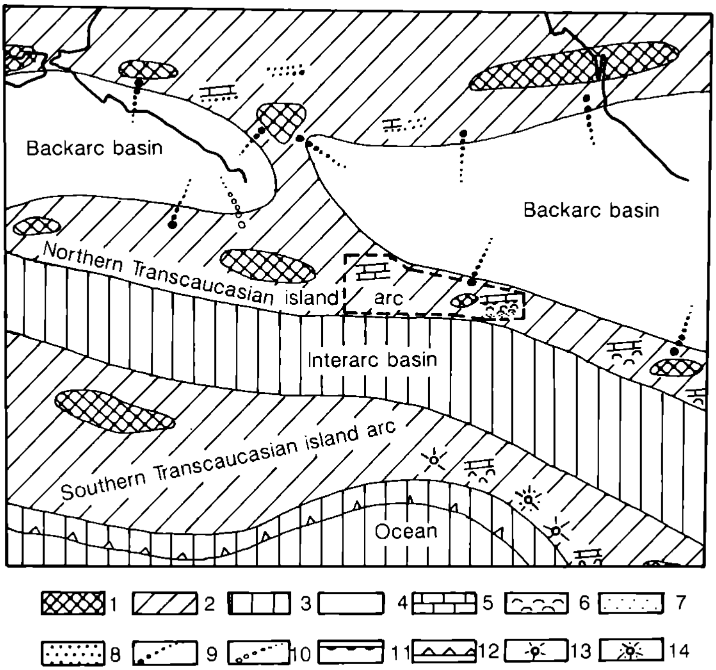
Correlation of heterofacies deposits belongs to the major stratigraphic problems. Pelagic deposits with ammonites are usually correlated with zoogenic or other sediments, devoid of ammonites and characterized by other shallow marine faunal groups. The possibility of correlating deep and shallow marine deposits comprising the same faunal groups is rare. Such a possibility exists in Georgia, where ammonite assemblages of different bathymetric levels occur in several horizons of the Lower Cretaceous deposits.

The Lower Cretaceous deposits of Georgia are characterized by facies variability, conditioned by complex geological structures typical of fold-belts.

The early Cretaceous paleogeographical units of Georgia from north to south are: The Great Caucasian marginal sea with flysch deposits and the Transcaucasian island arc with the Hercynian crystalline core overlain by the volcano-sedimentary subplatform cover (Georgian territory comprises



Text-Fig. 1. Barremian paleogeographical map of the Caucasus.

1 - Emerged land	8 - Conglomerates
2 - Shallow marine basin	9 - Siliciclastic turbidites
3 - Deep marine basin	10 - Calciturbidites
4 - Backarc flysch basin	11 - Hard-ground
5 - Limestones	12 - Oceanic subduction
6 - Reefal limestones	13 - Volcanism with lava
7 - Sandstones	14 - Volcanoclastics

Study area: strengthened contour.
Zigzag-lines: contours of actual Black and Caspian Seas.

mainly the north-western part of the island arc, known in the geological literature as the Georgian Block).

This paper is aimed to analyse paleogeographical environments of the island arc. The latter comprises a shallow marine basin, situated around salients of the Paleozoic basement and a deeper basin extending along its northern margin. The flysch basins of the Greater Caucasian marginal sea being the deposits of the northern part of Georgia, are not considered here (Text-Fig. 1).

In the northern part of the Georgian Block, in a deep marine basin, all stages of the Lower Cretaceous are represented and characterized by ammonites. Near Gagra, the Berriasian argillaceous and pelitomorphic limestones (40-50 m) contain *Pseudosubplanites* sp. in the lower part; at higher levels occur *Negreliceras negreli* (MATH.), *Euthymiceras* cf. *transfigurabilis* (BOG.) and *Protetragonites quadrisulcatus* (D'ORB.). In another locality, within a succession built up of alternating sandy limestones and thin-bedded sandy marls the following ammonites are known: *Fauriella incomposita* (RET.), *F. shipkovensis* (NIK. & MAND.), *Dalmasiceras* cf. *crassicostatum* DJAN. (see KVANTALIANI et al. 1981). From the Valanginian pelitomorphic limestones (35-40 m) *Thurmanniceras thurmanni* (PICT. & CAMP.) is known, whereas in lithographic limestones (8-10 m) *Kilianella* cf. *pexiptycha* (UHL.), *Thurmanniceras* cf. *campylotoxum* (UHL.), and *Neocomites* aff. *trezanensis* (LORY) occur (GAMKRELIDZE et al. 1952). *Crioceratites duvali* (LEV.), *Leopoldia bargamensis dubisiensis* (RASP.) are found in Hauterivian dolomitized limestones and limestones with flint concretions (35-200 m) (ERISTAVI 1952): in the lower part of well-bedded limestones with flint concretions (28.5) occur *Crioceratites nolani* (KIL.), *C. duvali* (LEV.), *Speetoniceras versicolor astarte* GLAS., and *S. inversum* (M. PAVLOV). From the upper part, *Pseudothurmannia (Pseudothurmannia) mortileti* (PICT. & LOR.), *P. (P.) renevieri* (SAR. & SCHÖND.), *P. (Balearites) balearis* (NOL.), *Acrioceras (Hoplocrioceras) pulcherrimum* (D'ORB.) (KAKABADZE 1980), *Olcostephanus* cf. *jeannoti* (D'ORB.), and *Rogersites* cf. *atherstoni* (SHARPE) are found in laminated pelitomorphic limestones (10 m) (GAMKRELIDZE et al. 1952).

The Barremian densely laminated limestones (50-235 m) at the lower levels comprise *Paracrioceras rondishiense* KAKAB., *P. dolloi* SARK., *Spitidiscus andrussowi* (KAR.), *Holcodiscus caillaudi* (D'ORB.), *H. gastaldi* (D'ORB.), and *Barremites difficilis* (D'ORB.). They are overlain by *Pulchellia* beds with *Subpulchellia brevicostata* KOTET., and *S. plana* KOTET. Laminated limestones and marls (25 m) with *Imerites* ex gr. *giraudi* (KIL.), *Colchidites longicostatus* KAKAB., and *Heteroceras* sp. follow upsection. In the uppermost part of the Barremian the laminated pelitomorphic limestones (10 m) contain *Colchidites ellipticus* (ROUCH.), *C. lakhephaensis* (ROUCH.), and *Imerites favrei* (ROUCH.).

Thus, along the northern margin of the island arc we observe environments characteristic of an infraneritic zone. This type of deposits was formed in the marginal part of the block, delimited by a deep marine basin. This deepening is considered to correspond to the lower circumlittoral (hemipelagic facies).

On the Georgian Block itself, the Cretaceous starts with a major transgression. The Upper Jurassic lagoonal deposits of the central part of the block are overlain by the basal formation of the Cretaceous system: quartz-arkose sandstones and conglomerates, younging from NNW (Berriasian) to SSE (Barremian) and following the slow propagation of the transgressive front. In the Valanginian and Hauterivian most of the Georgian Block is covered by a littoral sea. The existence of dinosaur footprints on two levels of the Valanginian limestones near the mountain Sathaplia testifies to repeated drainage and sea level fall. Littoral environments are achieved during Middle Hauterivian in the onset of Urgonian sedimentation throughout the block. Almost everywhere, Urgonian sedimentation is terminated in the late early Barremian. Further subsidence resulted in the deposition of ammonite bearing limestones. This marks the beginning of epineritic, i. e. infralittoral environments (8-50 m deep). The end of Urgonian sedimentation is heterochronous. Its earliest termination is established in sections where the Urgonian is overlain by laminated limestones with *Paracrioceras rondishiense* KAKAB. Owing to their stratigraphic position, they correspond to the Holcodiscus caillaudi Zone, but representatives of the family Holcodiscidae are missing in the epineritic zone on the Georgian Block; they are distinctly restricted to the infraneritic zone. The beds with *Paracrioceras* are overlain by *Pulchellia* beds, but only the representatives of the genera *Pulchellia* and *Heinzia* occur there; consequently, representatives of the genus *Subpulchellia* are also restricted to the infraneritic zone. Text-Fig. 2 illustrates the difference between equivalent deep marine and shallow marine Pulchellias. In the Upper Barremian, beds with representatives of *Hemihoplites* and overlying layers with *Imerites* occur in both infra- and epineritic zones. Yet, the first appearance of *Colchidites* in the uppermost part of the Barremian is marked by differences in depth of the marine basin; the group of *Colchidites shaoriensis* comprises a well developed discoïdal part of the shell, small helix and no uncoiled part, and occurs predominantly in the deeper part of the sea. In its shallow part, representatives of the groups of *Colchidites intermedius* and *C. colchicus* are present, showing large helix, a less developed discoidal part and a well developed uncoiled part. Text-Fig. 3 illustrates the difference between deeper and shallow marine *Colchidites*.

In the Aptian, the bathymetry seems to be quite the same all over the arc. Lithofacies are uniform in the northern and central parts of the arc and consist of marly limestones and marls of similar thickness with abundant organic remnants: *Deshayesites, Cheloniceras, Epicheloniceras, Kutatissites, Colombiceras, Megatyloceras, Ptychoceras, Costidiscus, Pseudohaploceras, Tetragonites, Australiceras, Tropaeum, Acanthohoplites, Diadochoceras, Hypacanthohoplites.*

Differences in bathymetry are again observed in the Albian. Representatives of the early Albian genera *Leymeriella* and *Douvilleiceras*, as well

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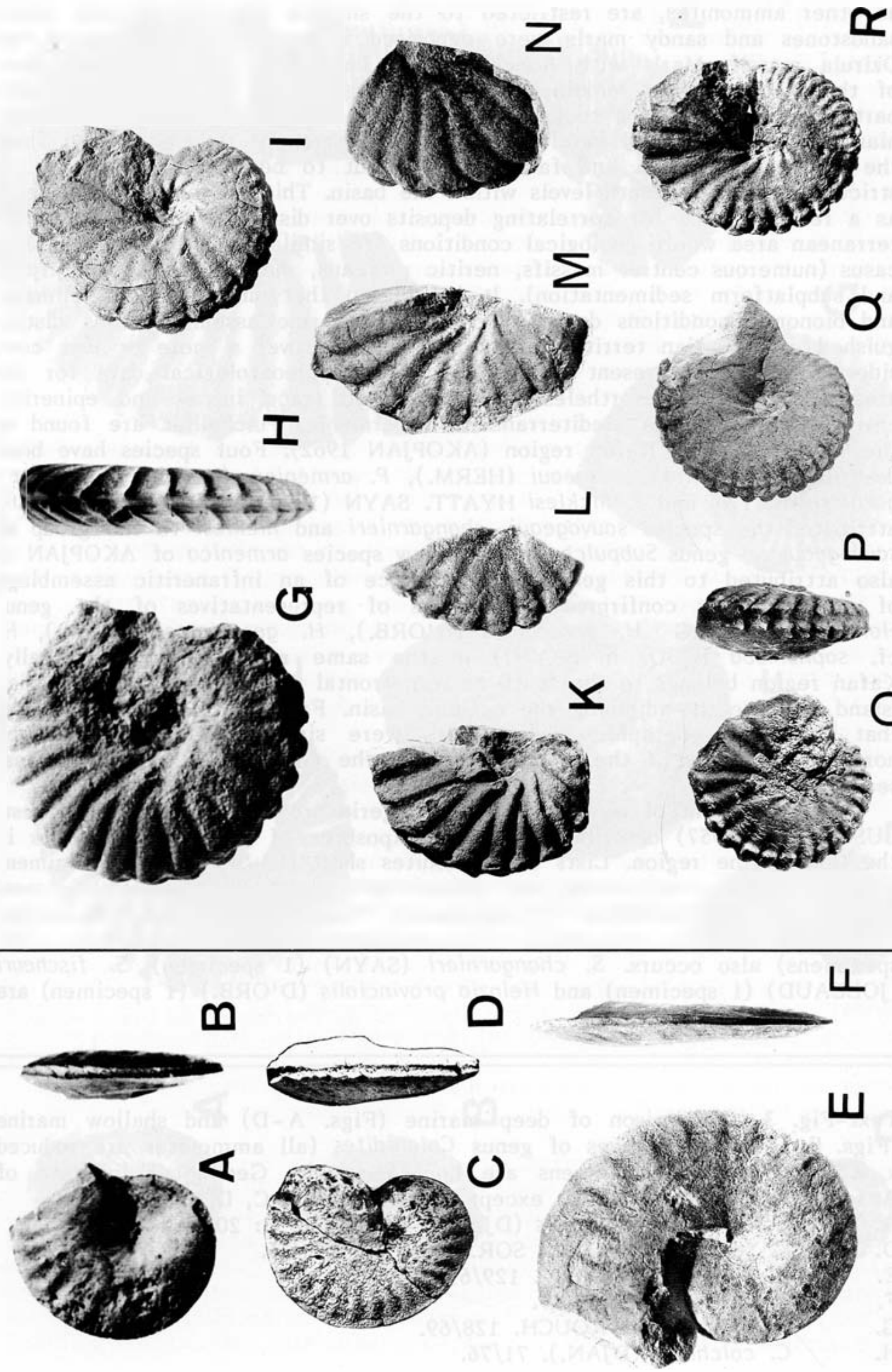
Text-Fig. 2. Comparison of deep marine (Figs. A-F) and shallow marine (Figs. G-R) representatives of family Pulchelliidae (all figures nat. size).

A, B. *Subpulchellia plana* KOTET. 153/95.
C, D, E, F. *S. brevicostata* KOTET. C, D: 7-250. E, F: 38/78.
G, H. *Pulchellia galeata* (BUCH). 131/69.
I. *Heinzia (Heinzia)* aff. *veleziensis* HYATT. 108/95.
K, L. *Heinzia (Carstenia) lindigi* (KARST.). K: 164/69. L: 145/95.
M. *Pulchellia* cf. *riedeli* BÜRGL. 14/95.
N. *Pulchellia* sp. ind. 182/95.
O, P, Q, R. *Heinzia (Heinzia) matura* HYATT. O, P: 20/95. Q: 23/95. R: 29/95.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as other ammonites, are restricted to the shallow part of the sea where sandstones and sandy marls were deposited along the peripheries of the Dzirula massif. Marls with Aucellinas and belemnites are deposited north of the block. In the Middle Albian the environments are similar in both parts, whereas emersion took place at the end of the Albian. The Vraconian fauna is preserved in aleurolites of the strongly reduced basin. Thus, the number of genera and families turns out to be stenobathyal and restricted to distinct depth levels within the basin. This phenomenon can serve as a reliable basis for correlating deposits over distances within the Mediterranean area where geological conditions are similar to those in the Caucasus (numerous central massifs, neritic plateaus, marine basins with flysch and subplatform sedimentation). It is implied that under similar climatic and bionomic conditions deeper and shallow marine assemblages as distinguished on Georgian territory, may be preserved over a more or less considerable area. At present the geological and paleontological data for the area are limited. Nevertheless, it is tried to trace infra- and epineritic environments into the Mediterranean. In Armenia, Pulchellias are found in the SE part of the Kafan region (AKOPJAN 1962). Four species have been described: *Pulchellia sauvageaui* (HERM.), *P. armenica* (HACOB.), *P. changarnieri* (BAYN) and *P. nicklesi* HYATT. SAYN (1890) and GIGNOUX (1920) attributed the species *sauvageaui, changarnieri* and *nicklesi* to the group of *sauvageaui* or genus *Subpulchellia*. The new species *armenica* of AKOPJAN is also attributed to this genus. The presence of an infraneritic assemblage of ammonites is confirmed by findings of representatives of the genus *Holcodiscus* UHLIG *H. perezianus* (D'ORB.), *H. geronimae* (HERM.), *H.* cf. *sophonisba* (COQ. in SAYN)) in the same association. Geologically, Kafan region belongs to the south-eastern frontal part of the Transcaucasian island arc directly adjoining the oceanic basin. Faunal associations indicate, that the paleogeographic environments were similar with those of the northern periphery of the arc adjacent to the Greater Caucasian marginal sea (Text-Fig. 1).

From this point of view the data on Algeria are of considerable interest. BUSNARDO (1957) describes numerous exposures of Barremian deposits in the Constantine region. Lists of ammonites show the number of specimens for each species. As it turns out, *Subpulchellia sauvageaui* (HERM.) is most frequent among Pulchellias (32 specimens). This species is always observed in association with *Holcodiscus* (51 specimens). *Heinzia ouachensis* (13 specimens) also occurs. *S. changarnieri* (SAYN) (1 specimen), *S. fischeuri* (JOLEAUD) (1 specimen) and *Heinzia provincialis* (D'ORB.) (1 specimen) are

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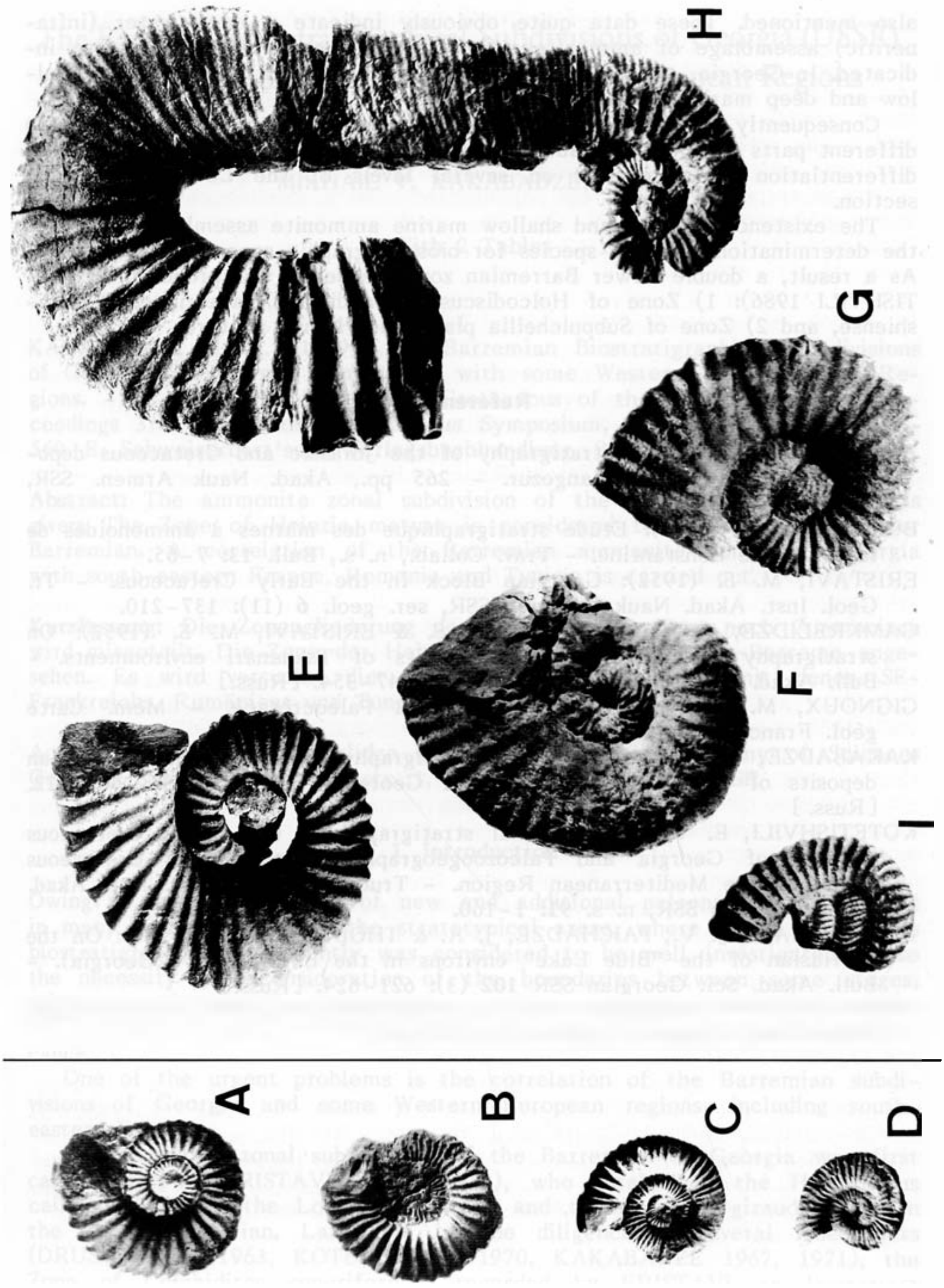
Text-Fig. 3. Comparison of deep marine (Figs. A-D) and shallow marine (Figs. E-I) representatives of genus *Colchidites* (all ammonites are reduced x 0.5). All figured specimens are housed at the Geological Institute of Academy of Sciences, Tbilisi, except for Text-Fig. 2 C, D.

A, B, C. *Colchidites shaoriensis* (DJAN.). A: 207/76. B: 202/76. C: 203/76.
D. *C. securiformis* (SIM., SOR. & BAC.). 243/76.
E. *C. kakabadzei* KOTET. 129/69.
F. *Colchidites* sp. 127/69.
G. *C. latecostatus* ROUCH. 128/69.
H. *C. colchicus* (DJAN.). 71/76.
I. *C. emerici costatus* ROUCH. 390/1096.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also mentioned. These data quite obviously indicate a deep water (infraneritic) assemblage of ammonites. However, the genus *Heinzia* is also indicated. In Georgia, there are no data on simultaneous occurrence of shallow and deep marine Pulchellias.

Consequently, the presence of various paleogeographic environments in different parts of the Transcaucasian island arc is indicative of bathymetric differentiation of ammonites on several levels of the Lower Cretaceous section.

The existence of deep and shallow marine ammonite assemblages requires the determination of index-species for biostratigraphic zones of both realms. As a result, a double Lower Barremian zonal sequence was proposed (KOTETISHVILI 1986): 1) Zone of Holcodiscus caillaudi and Paracrioceras rondishiense, and 2) Zone of Subpulchellia plana and Heinzia matura.

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The Barremian Biostratigraphical Subdivisions of Georgia (USSR) and Comparison with some Western Mediterranean Regions

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With 2 Tables

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Abstract: The ammonite zonal subdivision of the Barremian of Georgia is given. The Zone of Heinzia matura is considered to belong to the Upper Barremian. A correlation of the Barremian ammonite zones of Georgia with south-eastern France, Romania and Tunisia is carried out.

Kurzfassung: Die Zonengliederung des Barreme Georgiens nach Ammoniten wird mitgeteilt. Die Zone der Heinzia matura wird als Ober-Barreme angesehen. Es wird versucht, die Ammonitenzonen Georgiens mit denen SE-Frankreichs, Rumäniens und Tunesiens zu korrelieren.

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1. Introduction

Owing to the accumulation of new and additional paleontological material in many regions, including the stratotypical areas, where Lower Cretaceous biostratigraphy until recently was considered to be well investigated, arose the necessity of reconsideration of the boundaries between some stages, substages and zones. This phenomenon is closely connected with the problem of interregional zonal correlation which, at present, acquires special significance.

One of the urgent problems is the correlation of the Barremian subdivisions of Georgia and some Western European regions, including south-eastern France.

The ammonite zonal subdivisions of the Barremian of Georgia was first carried out by ERISTAVI (1951, 1952), who established the Holcodiscus caillaudi Zone in the Lower Barremian and the Imerites giraudi Zone in the Upper Barremian. Later on, by the diligence of several researchers (DRUSHTCHITZ 1963, KOTETISHVILI 1970, KAKABADZE 1967, 1971), the Zone of Colchidites securiformis, regarded by ERISTAVI as lowermost Aptian, was transferred to the uppermost Barremian. Most recently, KOTETISHVILI (1979, 1986) established two zones - that of Heinzia matura-Subpulchellia plana and that of Hemihoplites khwamliensis - between the Holcodiscus caillaudi and Imerites giraudi zones; the Lower/Upper Barremian boundary was drawn between these two new zones. Based on the analysis of all known and new paleontological and stratigraphical data, the Barremian zonal scheme (see Table 1) has been altered lately (KAKABADZE 1981, 1987). The substantiation of the mentioned alterations is given below when characterizing the Barremian zones of Georgia.

In the terminal part of this paper an attempt is carried out in correlating the Barremian zones of Georgia, south-eastern France, Romania and Tunisia.

2. Definition of Barremian Zones

2.1 Lower Barremian

2.1.1 The Zone of Holcodiscus caillaudi

In Georgia the base of the Holcodiscus caillaudi Zone marks the base of the Barremian stage.

The polemics on the Hauterivian-Barremian boundary, which arose at the beginning of this century (KILIAN 1907-1913, HAUG 1911, MÜLLER & SCHENK 1943) are still going on. At present, as it was mentioned at the symposium in Copenhagen (BIRKELUND et al. 1984) there are two good candidates for the Hauterivian-Barremian boundary: 1) the base of the *Pseudothurmannia* beds (BRESKOVSKI 1973, AVRAM 1983, VASICEK et al. 1983, etc.) or 2) the base of the *Holcodiscus* beds, i. e. the top of the *Pseudothurmannia* beds (DEBELMAS & THIEULOY 1965, EGOJAN 1968, THOMEL 1980, KAKABADZE 1981, ARNAUD-VANNEAU et al. 1982, etc.).

The Lower Cretaceous biostratigraphical data of Georgia, as well as that of other southern USSR regions show that the Zone of Pseudothurmannia mortilleti is closely connected with the underlying Upper Hauterivian Zone of Subsaynella sayni-Speetoniceras inversum by the presence of common genera, such as *Balearites, Pseudothurmannia, Acrioceras, Speetoniceras, Simbirskites, Crioceratites, Biasaloceras*. On the other hand, the Holcodiscus caillaudi Zone is characterized by the first appearance of the genera *Holcodiscus, Barremites* (s. str.), *Paracrioceras, Subpulchellia, Pulchellia, Anahamulina, Silesites,* whereas the typical Upper Hauterivian genera such as *Pseudothurmannia, Balearites, Speetoniceras, Simbirskites, Craspedodiscus* are not met in this zone. Among the transitional genera (from Hauterivian to Barremian) are to be mentioned *Crioceratites, Acrioceras, Biasaloceras* and *Phyllopachyceras*.

The following species are characteristic for the Holcodiscus caillaudi Zone of Georgia: *Holcodiscus caillaudi* (D'ORB.), *H. fallax* (COQ.), *H. gastaldi* (D'ORB.), *H. uhligi* (KAR.), *H. perezi* (D'ORB.), *Spitidiscus fallacior* (COQ.), *S. seunesi* (KIL.), *S. vandenheckei* (D'ORB.), *Astieridiscus morleti* (KIL.), *Acrioceras (A.) muckleae* SARK., *A. (A.) tabarelli* (ASTIER), *Barremites charrierianus* (D'ORB.), *Anahamulina picteti* (EICHW.), *Subpulchellia plana* KOTET.

2.2 Upper Barremian

2.2.1 The Zone of Heinzia matura

Considering this zone as Lower Barremian, KOTETISHVILI (1970, 1980) noted the following: a) the Pulchelliids in Georgia are widely distributed and all their representatives are timed to two layers (common thickness 0.6 - 0.9 m). For all that, the Pulchelliids are quantitatively predominant over the other groups. Together with them *Costidiscus recticostatus* (D'ORB.), *Paracrioceras* ex gr. *dolloi* SARK., *Barremites* sp. ind., *Hemihoplites* sp., *Mesohibolites beskidensis* (UHLIG), *Cymatoceras neocomiense* (D'ORB.), *Toxaster argilaceous* D'ORB. and bivalves (*Panope, Cucullaea, Camptonectes, Spondylus, Amphidonta*), as well as gastropods are found; b) the deposits of the Heinzia matura Zone are overlain by the layer with *Hemihoplites khwamliensis* being Upper Barremian; c) the deposits below the layer with Pulchelliidae are characterized by fossils of the Holcodiscus caillaudi Zone; d) comparison of the Barremian of Georgia with the stratotypical region (taking into account the data by BUSNARDO 1965) shows that both regions are characterized by the binominal subdivision of the Lower Barremian.

The cited arguments are not decisive in attributing this zone to the Lower Barremian, as they do not include: a) the biostratigraphical analysis of the species of the investigated zone, and b) the interregional correlative analysis.

First of all, it must be noted that there is no reason to conclude that all genera and all species of Pulchelliidae in Georgia are timed just to the Heinzia matura Zone. The indication by ERISTAVI (1952) on occurrence of Lower Barremian *Holcodiscus uhligi* (KAR.), *H. gastaldi* (D'ORB.), together with *Pulchellia compressissima* (D'ORB.) (redefined as *Subpulchellia plana* KOTET.) in the Racha region was confirmed by us. Moreover, in western Abkhasia representatives of Pulchelliidae were found also in the Lower Barremian, together with *Spitidiscus fallacior* (COQ.) and *Acrioceras (A.)* cf. *tabarelli* (AST.). Thus, we can infer that the first appearance of representatives of Pulchelliidae in Georgia is in the Lower Barremian Zone of Holcodiscus caillaudi.

As to the faunistical association, the biostratigraphical analysis shows that in the other Mediterranean regions ammonites of this zone are distributed either: a) only in the Upper Barremian (*Heinzia (H.) provincialis* (D'ORB.), *H. (H.) ouachensis* (COQ.), *Paracrioceras dolloi* SARK., *Hemihoplites* sp.); b) only in the Upper Barremian and Lower Aptian (*Costidiscus recticostatus* (D'ORB.)), or c) in both Barremian substages (*H. (Carstenia) lindigi* (KARST.), *Barremites* sp.). Only the new species *Paracrioceras rondishiense* KAKAB. is noted from the Lower Barremian of the Crimea (DRUSHTCHITZ 1960). As to the species *Pulchellia galeata* (BUCH.) and *Heinzia (H.) matura* HYATT, they are noted from the Barremian of south-eastern France and Carpathians, and the species *H. (H.) lorioli* (NICKL.) is reported from the Barremian of Spain.

Some Pulchelliids of the Heinzia matura Zone of Georgia are distributed also in the Middle Barremian of Colombia. It is remarkable that the Upper Barremian index species of the Mediterranean Realm - *H. (H.) provincialis* (D'ORB.) - is noted in Colombia from the upper part of the Middle Barremian; but two other species (*P. riedeli* BUERGL and *P. multicostata* RIEDEL) are noted from the whole Middle Barremian.

The endemic forms of Pulchelliids, met in the Heinzia matura Zone, are also rather various - *Heinzia (H.) tenuicostata* KOTET., *H. (H.)* aff. *veleziensis* HYATT, *H. (Carsteria) densecostata* KOTET., *Subpulchellia brevicostata* KOTET., *S. plana* KOTET., *Pulchellia* aff. *compressissima* (D'ORB.), *P.* aff. *fasciata* (GERH.) - but, naturally, they are not of use to solve the question under review.

Side by side with the ammonites, the biostratigraphical age of the rest of the faunal complex (belemnites, bivalves, echinids) also must be taken into account in order to solve the stratigraphical position of the Heinzia matura Zone. In other regions the representatives of these groups are distributed either: a) only in the Upper Barremian and Aptian (*Mesohibolites beskidensis* (UHL.), *Toxaster argilaceous* D'ORB., *Cucullaea glaba* SOW.); b) in both Barremian substages (*Amphidonta subsinuata* LEYM.), and c) in the Hauterivian-Aptian (*Cymatoceras neocomiense* (D'ORB.), *Panope gurgitis* BRONGN., *P. prevosti* LEYM.). It is remarkable that there are no genera or species in the Heinzia matura Zone, which are distributed only in the Lower Barremian.

Thus, the biostratigraphical analysis of both the ammonites and accompanied fauna of the Heinzia matura Zone obviously favours its Upper Barremian age.

2.2.2 The Zone of Hemihoplites soulieri

The lower boundary of this zone is clearly drawn by the first appearance of some species (see below) of *Hemihoplites*, as well as of *Audouliceras*.

This zone was separated as the Zone of *Matheronites (= Hemihoplites) khwamliensis* (KOTETISHVILI 1979) which was later (KAKABADZE 1981) renamed as *Hemihoplites soulieri*, considering the following data: the species *H. khwamliensis* (ROUCH.) is an endemic one, and, moreover, in Georgia it is also met in the two (upward) following Upper Barremian subzones of Imerites giraudi and Colchidites securiformis (KOTETISHVILI 1970, KAKABADZE 1981, etc.). As to the species *Hemihoplites soulieri* (MATH.), its specimens were found in the various geotectonic zones of Georgia only at the level under review. It is remarkable that in some West European regions, including the stratotypical area, this species is also distributed in the lower part of the Upper Barremian.

The following species are characteristic of the Hemihoplites soulieri Zone: *Hemihoplites soulieri* (MATH.), *H. khwamliensis* (ROUCH.), *H. feraudi* (D'ORB.), *Paracrioceras barremense* (KIL.), *Audouliceras collignoni* SARK., *"Acanthodiscus" amadei* (UHL.), *Barremites strettostoma* (MATH.), *Costidiscus microcostatus* (SIM.), *C. recticostatus* (D'ORB.), *Protetragonites crebrisulcatus* (UHL.), *Eulytoceras phestum* (MATH.), *Euphylloceras tethys* (D'ORB.).

2.2.3 The Zone of Imerites favrei-Heteroceras astieri

This zone is subdivided into two subzones: 1) the Imerites giraudi and 2) the Colchidites securiformis Subzone.

The base of the Imerites favrei-Heteroceras astieri Zone is defined by the first appearance of the genera *Heteroceras, Imerites, Eristavia, Arg-*

vethites and *Colchidites* (group of *Colchidites intermedius*). Among the genera transitional from the Hemihoplites soulieri Zone, the following must be mentioned: *Barremites, Hemihoplites, Protetragonites, Audouliceras, Costidiscus, Phyllopachyceras* and *Eulytoceras*. All the noted genera are distributed in both subzones, but the upper Colchidites securiformis Subzone is characterized by the first appearance of the genera *Paraimerites, Pseudocrioceras, Colchidites (Epicolchidites)* and by the abundance of various species of *Colchidites* (s. lato).

The ammonite species association of the Imerites giraudi Subzone consists of: *Heteroceras astieri* (D'ORB.), *H. bifurcatum* (D'ORB.), *Argvethites lashensis* (ROUCH.), *Imerites favrei* (ROUCH.), *I. sparcicostatus* (ROUCH.), *I. giraudi* (KIL.), *Eristavia dichotoma* (ERIST.), *E. tvishiensis* KAKAB., *Colchidites (C.) kutatisiense* KAKAB., *Costidiscus recticostatus* (D'ORB.), *Hemihoplites khwamliensis* (ROUCH.), *Protetragonites crebrisulcatus* (UHL.), *Eulytoceras phestum* (MATH.), *"Acanthodiscus" amadei* (UHL.), *Phyllopachyceras infundibulum* (D'ORB.), *Barremites strettostoma* (UHL.), *B. subdifficilis* (KAR.).

It is remarkable that, with the exception of *I. giraudi* (KIL.), *H. bifurcatum* (D'ORB.), *E. tvishiensis* KAKAB. and *C. (C.) kutatisiensis* KAKAB., the whole ammonite species association mentioned passes into the Colchidites securiformis Subzone. Moreover, the following species are characteristic of this subzone: *Colchidites (C.) colchicus* DJAN., *C. (C.) sarasini* ROUCH., *C. (C.) gamkrelidzei* ROUCH., *C. (C.) intermedius* DJAN., *C. (C.) bethleviensis* KAKAB., *C. (C.)* spp., *C. (Epicolchidites) securiformis* (ŠIM., BAC. & SOR.), *C. (E.) shaoriensis* DJAN., *C. (E.) tenuicostatus* KAKAB., *C. (E.) veleurensis* KAKAB., *C. (E.)* spp., *Paraimerites semituberculatus* (ROUCH.), *P. planus* (ROUCH.), *P. katsharavai* (ROUCH.), *P. densecostatus* (RENNG.), *P. tsholashensis* (ROUCH.), *P.* spp., *Costidiscus microcostatus* (SIM.).

The mentioned subdivision by Heteroceratidae is visible only in some sections of western Georgia and Kopetdag. Yet, in the majority of the southern USSR regions (north and north-western Caucasus, Minor Caucasus, Dagestan, Crimea, Tuarkir, Great and Minor Balchan) the binominal subdivision by Heteroceratidae in the Upper Barremian is not possible. Taking into account this circumstance, as well as the mentioned great generic and specific resemblance of the ammonites of these two "Imeritic" and "Colchiditic" layers, we came to the conclusion (KAKABADZE 1983) that they must be regarded as subzones and united into one Zone of Imerites favrei-Heteroceras astieri (Table 1).

3. Correlation

The described Barremian stratigraphical levels of Georgia rather clearly correlate with the Barremian subdivisions of south-eastern France (stratotypical region), Romania and Tunisia. The biostratigraphical schemes of all these regions were significantly detailed (see Table 2) during the last years (BUSNARDO 1984, AVRAM 1983, MEMMI 1981).

In the stratotypical area, the base of the Barremian is defined by the first appearance of the genera *Holcodiscus, Nicklesia* and by the flourishing of the genera *Barremites, Hamulina, Silesites, Spitidiscus*. According to the new Barremian scheme by BUSNARDO (1984) three zones are distinguished in the Lower Barremian, e. g. that of a) Spitidiscus hugii, b) Pulchellia

Table 1. Biostratigraphic subdivisions of the Georgian Barremian.

Stage	Sub-stage	Zone		AMMONITES	
BARREMIAN	Upper	Imerites favrei - Heteroceras astieri	Colchidites securiformis	*Heteroceras elegans, Colchidites (C.) colchicus, C. (C.) sarasini, C. (C.) intermedius, C. (C.)* spp., *C. (E.) securiformis, C. (E.). shaoriensis, C. (E.) tenuicostatus, C. (E.)* spp., *Paraimerites planus, P. densecostatus, P.*spp., *Costidiscus microcostatus*	*Heteroceras astieri, H. vermiforme, Argvethites lashensis, Imerites favrei, I. sparcicostatus, Eristavia dichotoma, Costidiscus recticostatus, Hemihoplites khwamliensis, Protetragonites crebrisulcatus, Eulytoceras phestum, "Acanthodiscus" amadei, Phyllopachyceras infundibulum, Barremites strettostoma, B. subdifficilis*
			Imerites giraudi	*Imerites giraudi, Eristavia tvishiensis, Heteroceras bifurcatum, Colchidites (C.) kutatisiensis*	
		Hemihoplites soulieri		*Hemihoplites soulieri, H. khwamliensis, H. feraudi, Paracrioceras barremense, Audouliceras collignoni, "Acanthodiscus" amadei, Barremites strettostoma, Costidiscus microcostatus, C. recticostatus, Protetragonites crebrisulcatus, Eulytoceras phestum, E. thethys*	
		Heinzia matura		*Heinzia (H.) matura, H. (H.) provincialis, H. (H.) ouachensis, H. (H.) lorioli, H. (H.) tenuicostata, H. (Carstenia) lindigi, H. (C.) densecostata, Pulchellia galeata, P. multicostata, P. riedeli, Subpulchellia plana, S. brevicostata, Costidiscus recticostatus, Paracrioceras dolloi, P. rondishiense, Hemihoplites* sp., *Barremites* sp.	
	Lower	Holcodiscus caillaudi		*Holcodiscus caillaudi, H. fallax, H. gastaldi, H. uhligi, H. perezi, Spitidiscus fallacior, S. seunesi, S. vandenheckei, Astieridiscus morleti, Acrioceras (A.) muckleae, A. (A.) tabarelli, Barremites charrierianus, Anahamulina picteti, Subpulchellia plana*	

Table 2. Correlation of the Barremian zonal subdivisions of Georgia, SE France, Romania and Tunisia.

Stage	Substage	SE France (BUSNARDO 1984)	Romanian Carpathians (AVRAM 1983)		Georgia (this paper)		Tunisia (MEMMI 1981)
BARREMIAN	Upper	*Colchidites* sp.	*Silesites seranonis*	*Parancyloceras?* sp.	*I. favrei - H. astieri*	*Colchidites securiformis*	*Leptoceras puzosianum*
		Heteroceras astieri		*I. giraudi - E. dichotoma*		*Imerites giraudi*	*Heteroceras astieri*
		Hemihoplites feraudi		"*C.*" ex gr. *barremense*	*Hemihoplites soulieri*		*Hemihoplites feraudianus*
		"*Emericiceras*" *barremense*		*Heinzia provincialis*	*Heinzia matura*		
	Lower	*Moutoniceras* sp.	*Holcodiscus caillaudi*	*Pulchellia compressissima*	*Holcodiscus caillaudi*		*Holcodiscus caillaudi*
		P. compressissima					
		Spitidiscus hugii		*P. changarnieri*			
HAUTERIVIAN	Upper	*Pseudothurmannia angulicostata*	*Pseudothurmannia picteti*		*Pseudothurmannia mortilleti*		*Angulicostatus - Balearis*

compressissima, and c) Moutoniceras sp. In the Romanian Carpathians the Lower Barremian Zone of Holcodiscus caillaudi with the two subzones of a) Pulchellia changarnieri and b) Pulchellia compressissima is established above the Zone of Pseudothurmannia picteti. It is remarkable, that in other Mediterranean regions (Spain, Bulgaria, Tunisia) the Lower Barremian fauna is also rich in ammonite species, but only one zone without subzonal division is established above the Pseudothurmannian beds. The picture is similar in Georgia. Taking into account the existence of Lower Barremian ammonites, similar to those of south-eastern France, Romania etc. (*Holcodiscus caillaudi* (D'ORB.), *H. fallax* (COQ.), *H. gastaldi* (D'ORB.), *H. perezi* (D'ORB.), *Spitidiscus seunesi* (KIL.), *S. vandenheckei* (D'ORB.), *Acrioceras (A.) muckleae* SARK.), we can only conclude that the entire Lower Barremian Holcodiscus caillaudi Zone of Georgia and of Tunisia corresponds to the Lower Barremian zones and subzones of these regions noted above.

The Upper Barremian in Georgia, as it was shown, begins with the Zone of Heinzia matura, which is characterized by the first appearance of the Upper Barremian *Heinzia (H.) provincialis* (D'ORB.), *H. (H.) ouachensis* (COQ.), *Costidiscus recticostatus* (D'ORB.), Gr. of *Paracrioceras barremense,* etc. This species association allows to conclude that the Zone of Heinzia matura of Georgia corresponds to the lowermost Upper Barremian Zone of "Emericiceras" barremense of SE France and to the Heinzia provincialis Subzone of Romania. Owing to the ammonite species association and to the stratigraphical position, the second Upper Barremian zone (of Hemihoplites soulieri) of Georgia correlates with the Hemihoplites feraudi Zone of the stratotypical area and with the "Crioceratites" ex gr. barremense-orbigny Subzone of Romania; among the species common of this level *Hemihoplites feraudi* (D'ORB.) and *H. soulieri* (MATH.) are to be mentioned first. Taking into account the ammonite species composition and the stratigraphical position of the Zone of Hemihoplites feraudianus of Tunisia, we consider that it correlates with both the Heinzia matura and Hemihoplites soulieri Zones of Georgia.

As to the correlation of subdivisions of higher Upper Barremian levels being based on the stratigraphical position and faunistical composition (e. g., resemblance of Heteroceratids), we conclude that the Heteroceras astieri and Colchidites sp. Zones of the stratotypical area correspond to the Imerites giraudi-Eristavia dichotoma and Parancyloceras sp. Subzones of Romania, to the Imerites giraudi and Colchidites securiformis Subzones of Georgia, as well as to the Heteroceras astieri and Leptoceras puzosianum Zones of Tunisia (see Table 2).

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C. Events

Stratigraphy and Mineralogy of the Selli Level (Early Aptian) at the Base of the Marne a Fucoidi in the Umbro-Marchean Apennines (Italy)*

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With 12 Text-Figures and 2 Tables

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Abstract: Detailed, lithostratigraphical, biostratigraphical and mineralogical studies on the "ichthyolithic-bituminous-radiolaritic" regional marker-bed (named Selli Level), located just above the lower boundary of the Marne a Fucoidi Formation (early Aptian-late Albian) in the Umbro-Marchean Apennines, were carried out. This distinctive organic-rich level is 1-3 m thick. It consists of mudstones alternating with radiolarian silty and sandy layers. Radiolaria are the exclusive components of the microfaunal assemblage (belonging to the Stichocapsa euganea Zone) found in the level. The age of the level is Lower Aptian (Bedoulian).

The main mineralogical composition of the Selli Level is characterized by the absence of calcite together with a high quartz content. Smectite, illite and illite-smectite mixed layers dominate the clay mineral fraction with some chlorite, chlorite-vermiculite and occasional kaolinite. Based on its mineralogical composition, the level is distinctly different from the underlying and overlying sediments of the Marne a Fucoidi.

The Selli Level is similar, in some respects, to the Bonarelli Level (dated close to the Cenomanian-Turonian boundary) which characterizes the uppermost part of the Scaglia Bianca in the Umbro-Marchean Apennines. Both the depositional sequences which include the above-mentioned marker-beds show a similar vertical trend. The Selli and Bonarelli Levels probably deposited under analogous palaeoceanographic conditions.

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Kurzfassung: Detaillierte lithostratigraphische, biostratigraphische und mineralogische Untersuchungen des "ichthyolithisch-bituminös-radiolaritischen" regionalen Leithorizonts (Selli-Horizont) werden beschrieben. Der Horizont liegt unmittelbar über der Basis der "Marne a Fucoidi-Formation" (frühes Apt-spätes Alb) im Umbrisch-Märkischen Apennin, ist 1-3 m mächtig und markant durch seinen hohen Gehalt an C_{org}. Er besteht aus Mudstones, die mit sandigen und siltigen Radiolarien-führenden Lagen wechsellagern. Die Mikrofaunen-Vergesellschaftung dieses Niveaus besteht ausschließlich aus Radiolarien (der Stichocapsa euganea-Zone); ihr Alter ist Unterapt (Bedoule).

Das mineralogische Hauptmerkmal des Selli-Horizonts ist das Fehlen von Kalzit gemeinsam mit einem hohen Quarz-Gehalt. Smektit, Illit und Illit-Smektit führende Mixed-Layers dominieren innerhalb der Tonminerale; untergeordnet treten Chlorit, Chlorit-Vermikulit und gelegentlich Kaolinit auf. Aufgrund seiner mineralogischen Zusammensetzung ist der Horizont deutlich verschieden von den unter- und überlagernden Sedimenten der Marne a Fucoidi.

In gewisser Hinsicht ist der Selli-Horizont mit dem Bonarelli-Horizont (Cenoman/Turon-Grenze) vergleichbar, der den höchsten Teil der Scaglia Bianca im Umbrisch-Märkischen Apennin kennzeichnet. Beide Folgen zeigen den gleichen vertikalen Trend des Ablagerungsmilieus. Es wird angenommen, daß sie unter analogen paläo-ozeanographischen Bedingungen gebildet wurden.

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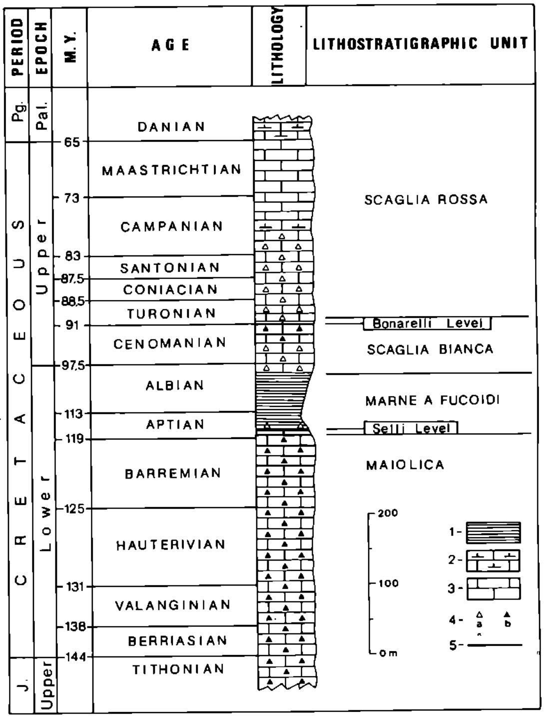
1. Introduction

Recent studies (WEZEL 1985, COCCIONI et al. 1987) have shown the presence of a radiolaritic-bituminous-ichthyolithic marker-bed (named Selli Level) located just above the lower boundary of the Marne a Fucoidi Formation (early Aptian-late Albian) in the Umbro-Marchean Apennines (see Text-Fig. 1). By means of the analysis of 21 representative sections (see Text-Figs. 2 to 5), this paper gives further stratigraphical and mineralogical data of this level, and confirms its regional characteristics and considerable geological and stratigraphical importance in the Umbro-Marchean succession.

2. Stratigraphy

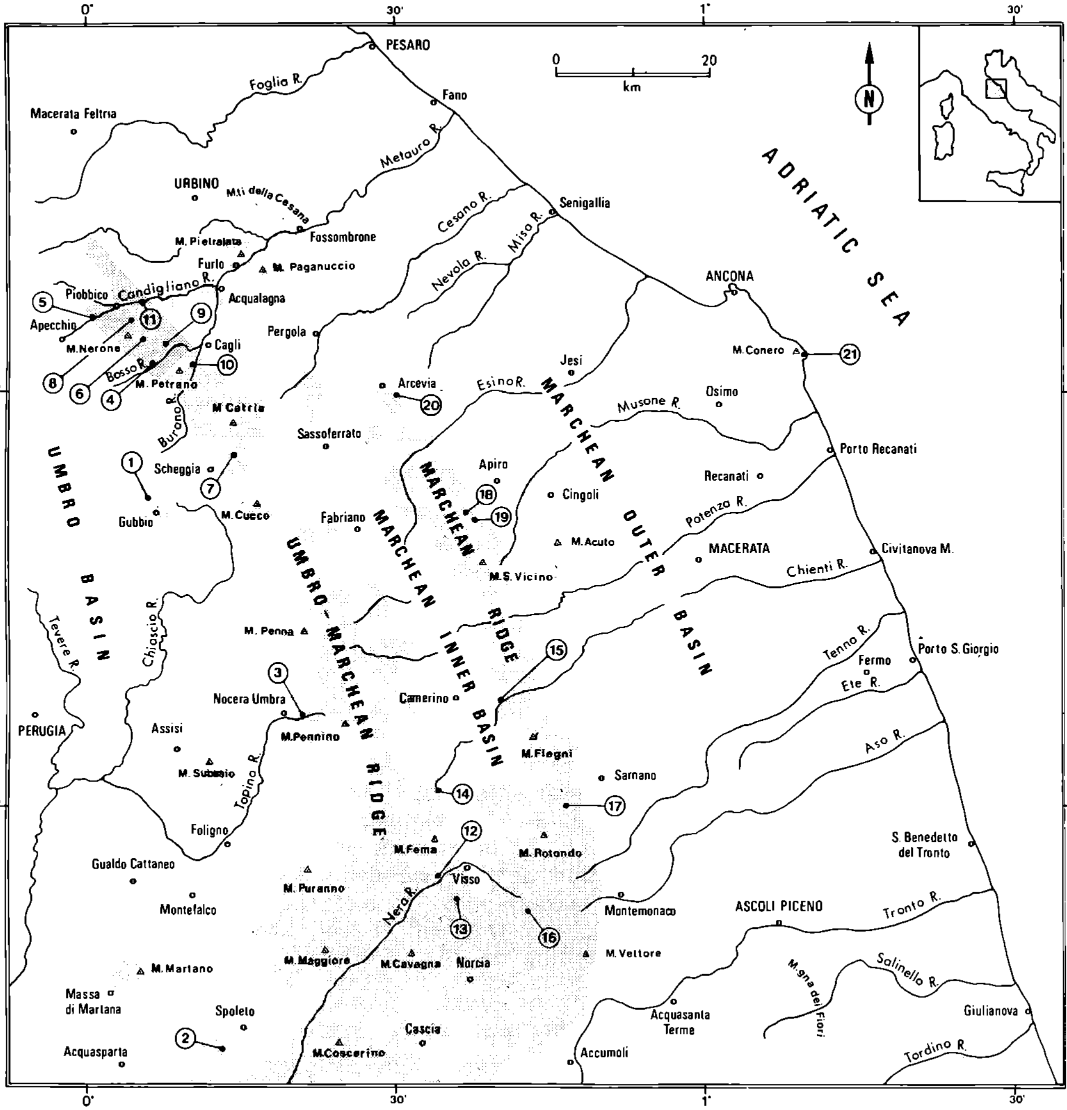
In the Marne a Fucoidi Formation of the Umbro-Marchean sequence six members are recognizable (COCCIONI et al. 1987, 1988) named from the bottom to the top: 1. "Greenish-grey cherty Member"; 2. "Lower reddish marly Member"; 3. "Brownish clayey Member"; 4. "Greenish marly Member"; 5. "Upper reddish marly Member"; 6. "Whitish marly limestone Member". As far as the lithostratigraphical, micropaleontological and mineralogical description of these members is concerned, we refer to COCCIONI et al. (1988).



Text-Fig. 1. Schematic average thickness litho- and chronostratigraphy of the Cretaceous sequence in the Umbro-Marchean Apennines. Legend: 1 = clay and marl; 2 = marly limestone; 3 = limestone; 4 = greenish, greyish to reddish (a) and dark-grey to black (b) chert beds and nodules; 5 = black-shale layer. The stratigraphic position of the two distinctive "ichthyolithic-bituminous-radiolaritic" regional marker-beds (i. e., the Bonarelli, see ARTHUR & PREMOLI SILVA 1982, and the Selli Level, see COCCIONI et al. 1987) is shown. Time scale according to HARLAND et al. (1982).

The lower boundary of the Marne a Fucoidi is placed by COCCIONI et al. (1987) at the top of the calcareous bed which includes the last layer of black chert of the Maiolica Formation. On the basis of the lithostratigraphic studies being carried out at the present, this layer has regionally isochronous significance.

The Selli Level occurs within a sharply contrasting sequence of greenish-grey and very light to medium grey fine-grained, cherty limestone and marly limestone ("Greenish-grey cherty Member") (see Text-Fig. 6). A very light to medium grey limestone underlies the Selli Level and is punctuated by dark-grey to black chert nodules and by interbedded layers of fissile black marlstone. Just above the Selli Level, greenish-grey and light olive-grey chert beds and nodules occur. The limestone is micritic and contains Radiolaria, calcispheres, nannofossils and nannofossil fragments together with sparse planktonic Foraminifera (i. e., small hedbergellids and globigerinelloids), and very rare calcareous benthic Foraminifera. Radiolaria



Text-Fig. 2. Location of the measured sections ordered from the inner part to the outer part of the Umbro-Marchean Ridges. Locations numbered are (1) Contessa section; (2) Monte li Rossi section; (3) le Cese section; (4) Fiume Bosso section; (5) S.S. Apecchiese-km 32.800 section; (6) il Cimaio section; (7) Valdorbia section; (8) Presale section; (9) Poggio le Guaine section; (10) Fiume Burano section; (11) Gorgo a Cerbara section; (12) i Molini section; (13) Preci section; (14) Madonna di Caspreano section; (15) S.S. 77-km 53.500 section; (16) Spina di Gualdo section; (17) Pizzo di Meta section; (18) Poggio San Vicino section; (19) Frontale section; (20) Avacelli section; (21) Monte Conero section.

often occur in discrete nodules and thin layers. A few metres above the top of the Selli Level, the greenish-grey cherty limestone becomes pale to moderate red ("Lower reddish marly Member") and even greyish-red. This change possibly reflects a higher oxidation state and a resulting decrease in organic matter and pyrite content.

The Selli Level is typically found within the calcareous sequence (see Text-Figs. 3 to 5) referred to as the "Greenish-grey cherty Member" of the Marne a Fucoidi. According to COCCIONI et al. (1987), this member can be informally subdivided from the bottom to the top as follows: Lower Unit, Selli Level and Upper Unit.

The Selli Level is 1 to 3 m thick. It consists of laminated to bioturbated greyish-yellow, olive-grey, greenish-grey, moderate brown to brownish-black and dark-grey to black mudstone and shale alternated with greyish-yellow, olive-grey, brownish-grey and medium to dark grey radiolarian sandy and silty layers 0.5 to 8 cm thick (Text-Figs. 7 and 8). In such layers, which are found throughout the study area, Radiolaria constitute the entire abundant microfauna. The greenish-grey mudstone and the radiolarian layers are typically bioturbated. The trace fossils recognized are referred to deformated *Planolites* and small *Chondrites*. The light yellow, olive-green, brown, very dark grey and black mudstones are typically laminated and unbioturbated. The laminated, greyish-yellow mudstone is phosphatic and contains well preserved fish fossils (see Text-Figs. 9 and 10). The lamination is exclusively parallel and prevalently fine (0.05-0.1 mm). Sometimes, thicker beds were also found (up to 3 mm). Contacts between the laminae, similar to those between the various types of mudstone, are generally sharp. Lamination is due to the alternation of dark-coloured layers (organic matter-rich) with light coloured laminae (siliceous and/or phosphatic material-rich) (see Text-Fig. 8 C). The siliceous material is unstructured and probably consists of fragments of crushed and altered radiolarian tests. The phosphatic material is constituted by fish remains.

According to COCCIONI et al. (1987), the Selli Level can be informally subdivided into two intervals (the "green interval" and, above it, the "black interval") that are exclusively differentiated on the basis of colours (see Text-Figs. 3 to 5). The passage between the two lithological intervals is transitional; the thickness of the "green interval" ranges from 28 (Fiume Burano section) to 190 cm (le Cese section). The "black interval" is 50 (Monte li Rossi section) to 110 cm (le Cese section) thick. Usually the

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Text-Figs. 3 to 5. Lithostratigraphical correlation between the measured sections. Legend: 1 = bioturbated, occasionally laminated, very light to medium grey limestone; 2 = bioturbated, greenish-grey limestone; 3 = bioturbated, greenish-grey marlstone and marly limestone; 4 = radiolarian sandy/silty layer; 5 = bioturbated, greenish-grey mudstone; 6 = laminated olive-grey mudstone; 7 = laminated, moderate brown to brownish-black and dark-grey to black mudstone, shale and fissile marlstone; 8 = laminated, greyish-yellow phosphatic mudstone; 9 = slumped interval; 10 = dark-grey to black chert beds (a) and nodules (b); 11 = greenish-grey and light olive-grey chert beds (a) and nodules (b); 12 = pyrite nodules; 13 = fossil fishes; 14 = fault; N.E. = not exposed. The total organic carbon content measured in the "black interval" of the Poggio le Guaine section is reported (as %wt).

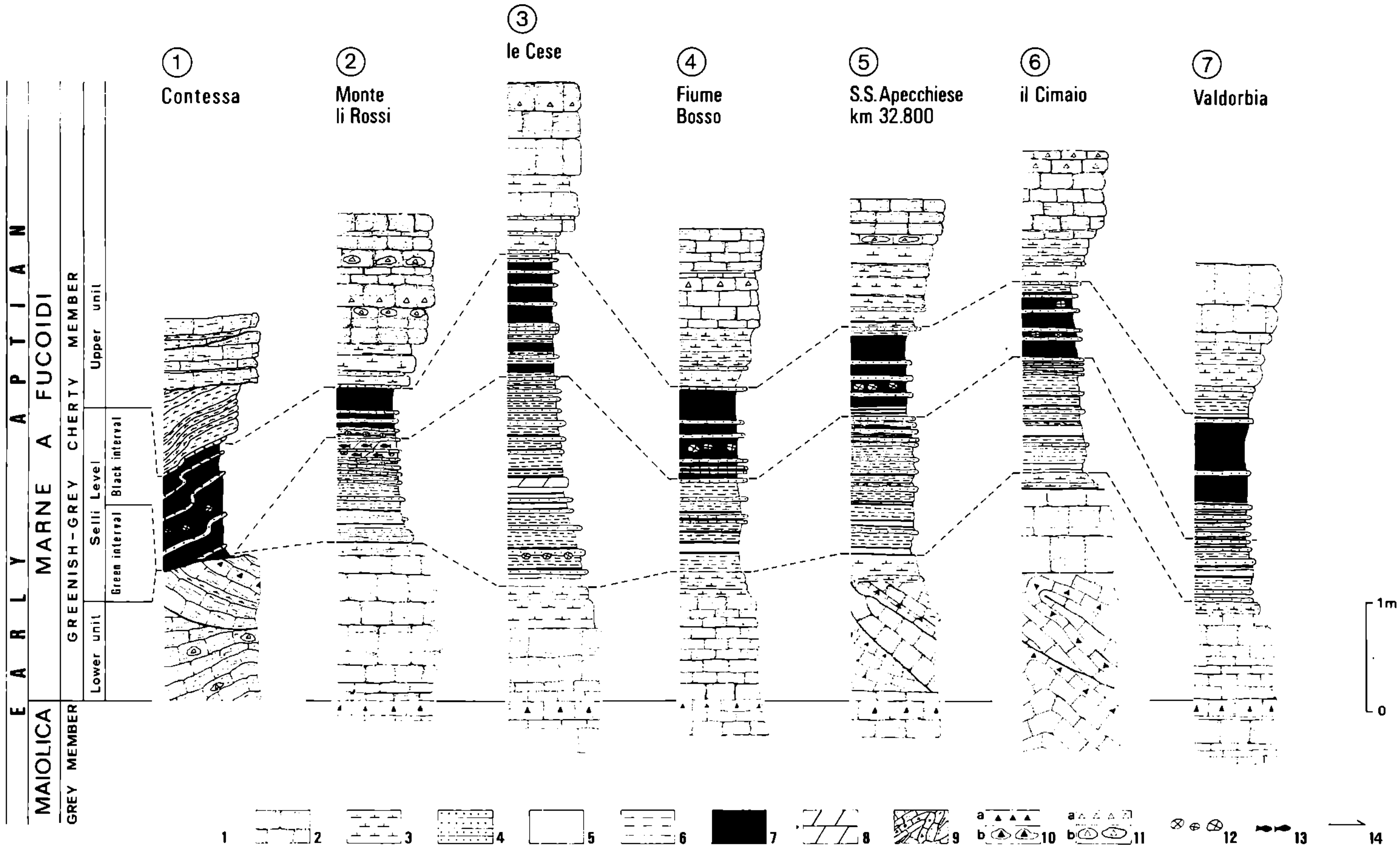
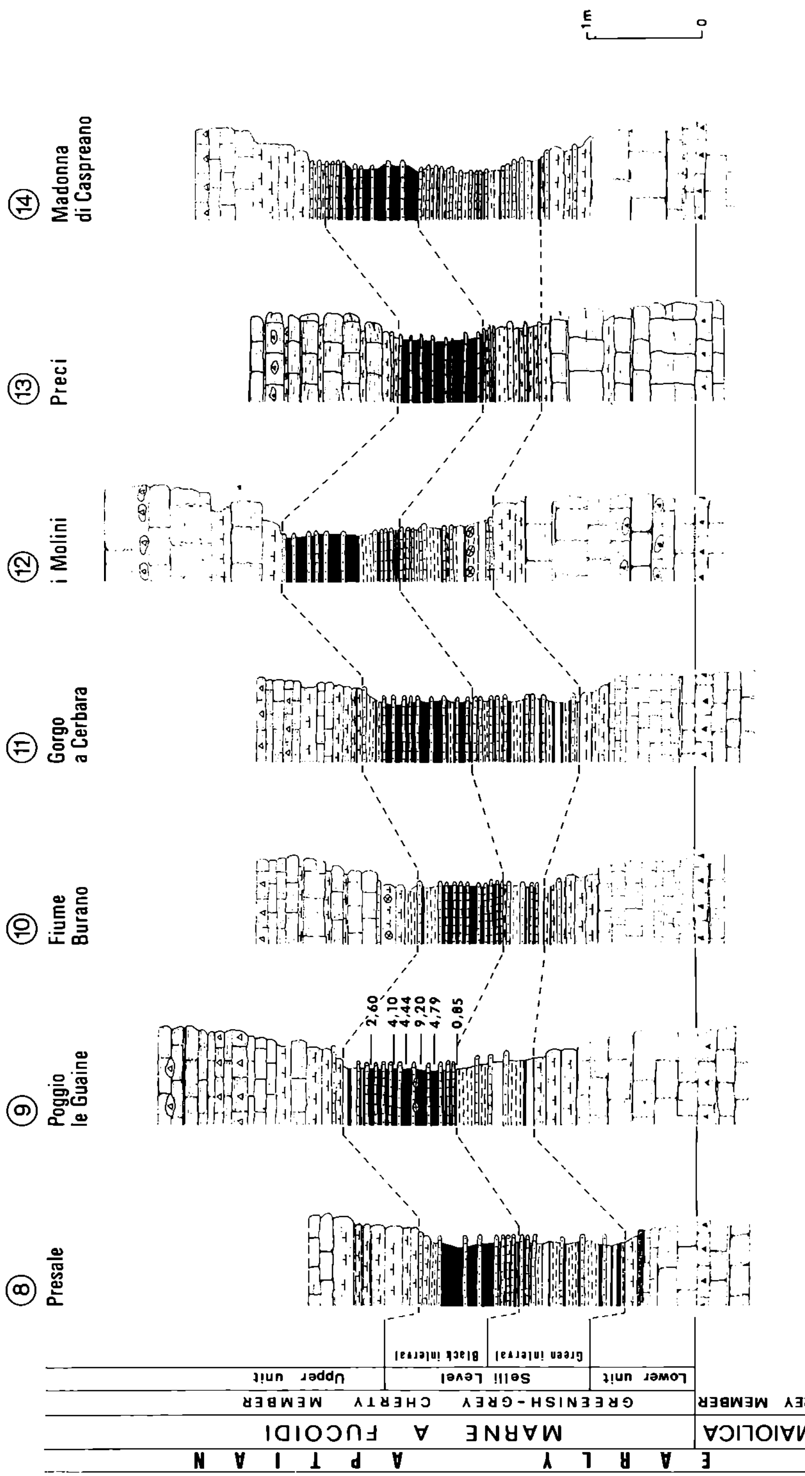

EARLY APTIAN
MAIOLICA
MARNE A FUCOIDI
GREY MEMBER
GREENISH-GREY CHERTY MEMBER
Lower unit
Upper unit
Selli Level
Green interval
Black interval
1 Contessa
2 Monte li Rossi
3 le Cese
4 Fiume Bosso
5 S.S. Apecchiese km 32.800
6 il Cimaio
7 Valdorbia
1m
0
1
2
3
4
5
6
7
8
9
a
b
10
a
b
11
12
13
14

Fig. 3.



1m
0
(8) Presale
(9) Poggio le Guaine
(10) Fiume Burano
(11) Gorgo a Cerbara
(12) i Molini
(13) Preci
(14) Madonna di Caspreano
2,60
4,10
4,44
9,20
4,79
0,85
Lower unit
Selli Level
Upper unit
Green interval
Black interval
GREY MEMBER
GREENISH-GREY CHERTY MEMBER
MAIOLICA
MARNE A FUCOIDI
EARLY APTIAN

Fig. 4.

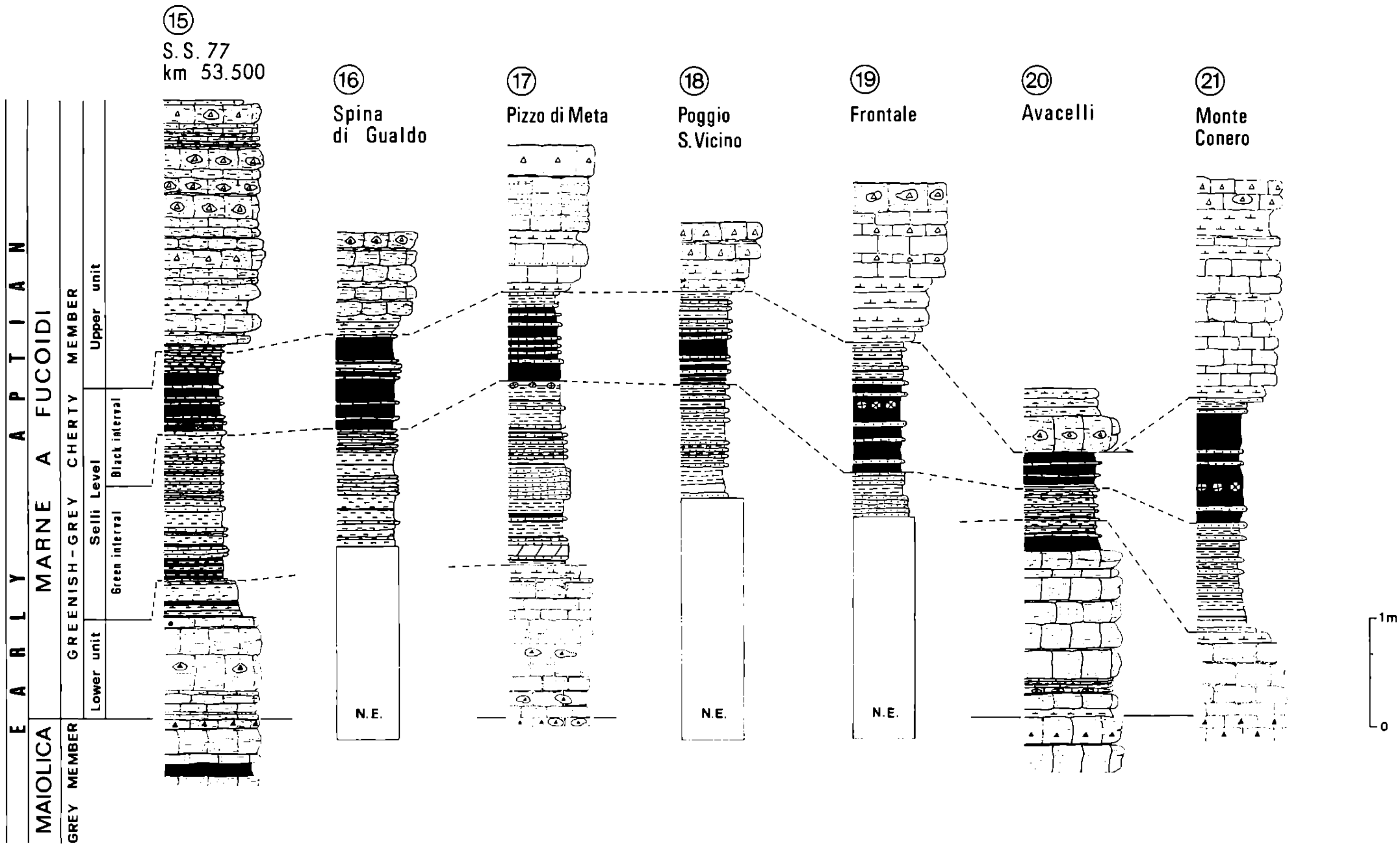

15
S.S. 77
km 53.500
16
Spina
di Gualdo
17
Pizzo di Meta
18
Poggio
S.Vicino
19
Frontale
20
Avacelli
21
Monte
Conero
EARLY APTIAN
MARNE A FUCOIDI
GREENISH-GREY CHERTY MEMBER
Upper unit
Selli Level
Black interval
Green interval
Lower unit
MAIOLICA
GREY MEMBER
N.E.
N.E.
N.E.
1m
0

Fig. 5

thickness of the "green interval" is greater than that of the black one. The different thickness of the "green" and "black" intervals together with both the thickness and the number of the radiolarian silty and sandy layers could be related to the morphostructural characteristics of the paleobasin at the time of deposition.

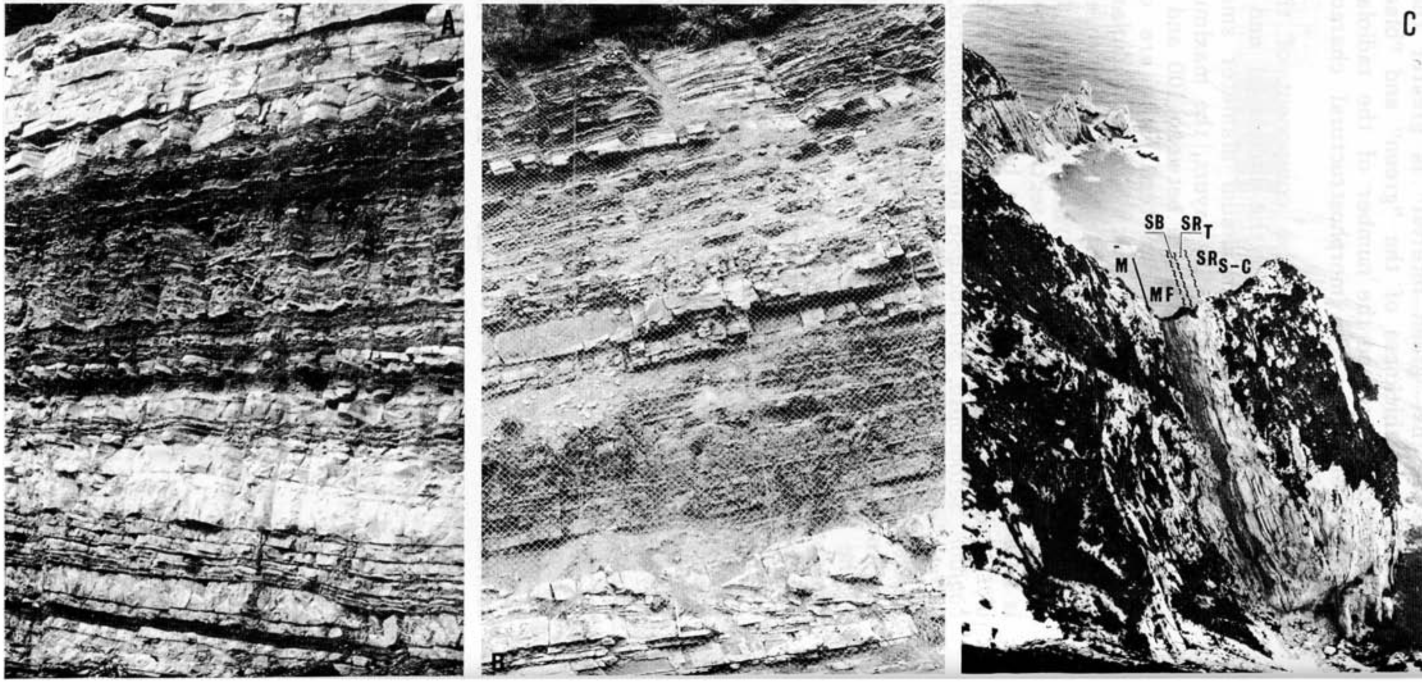
Radiolaria are the exclusive component of the microfaunal assemblages found in the Selli Level. In the laminated and bioturbated mudstone, the radiolarian tests have a maximum diameter smaller than 250 microns. In the radiolarian sandy and silty layers, the maximum diameter found is 1 mm but the most frequent sizes are between 100 and 300 microns. In the olive-grey mudstone the radiolarian test walls are often replaced with pyrite which sometimes fills the tests. In the radiolarian silty and sandy layers the radiolarian tests are constantly filled with radiating masses of chalcedony. The number of such layers is between 10 (Fiume Burano section) and 24 (Madonna di Caspreano section), and the thickness of these layers is between 1 and 8 cm. Radiolaria belong to the following genera: *Archaeodictyomitra, Cenosphaera, Dictyomitra, Eucyrtis, Hagiastrum, Holocriptocanium, Sethocapsa, Stichocapsa* and *Spongodiscus.*

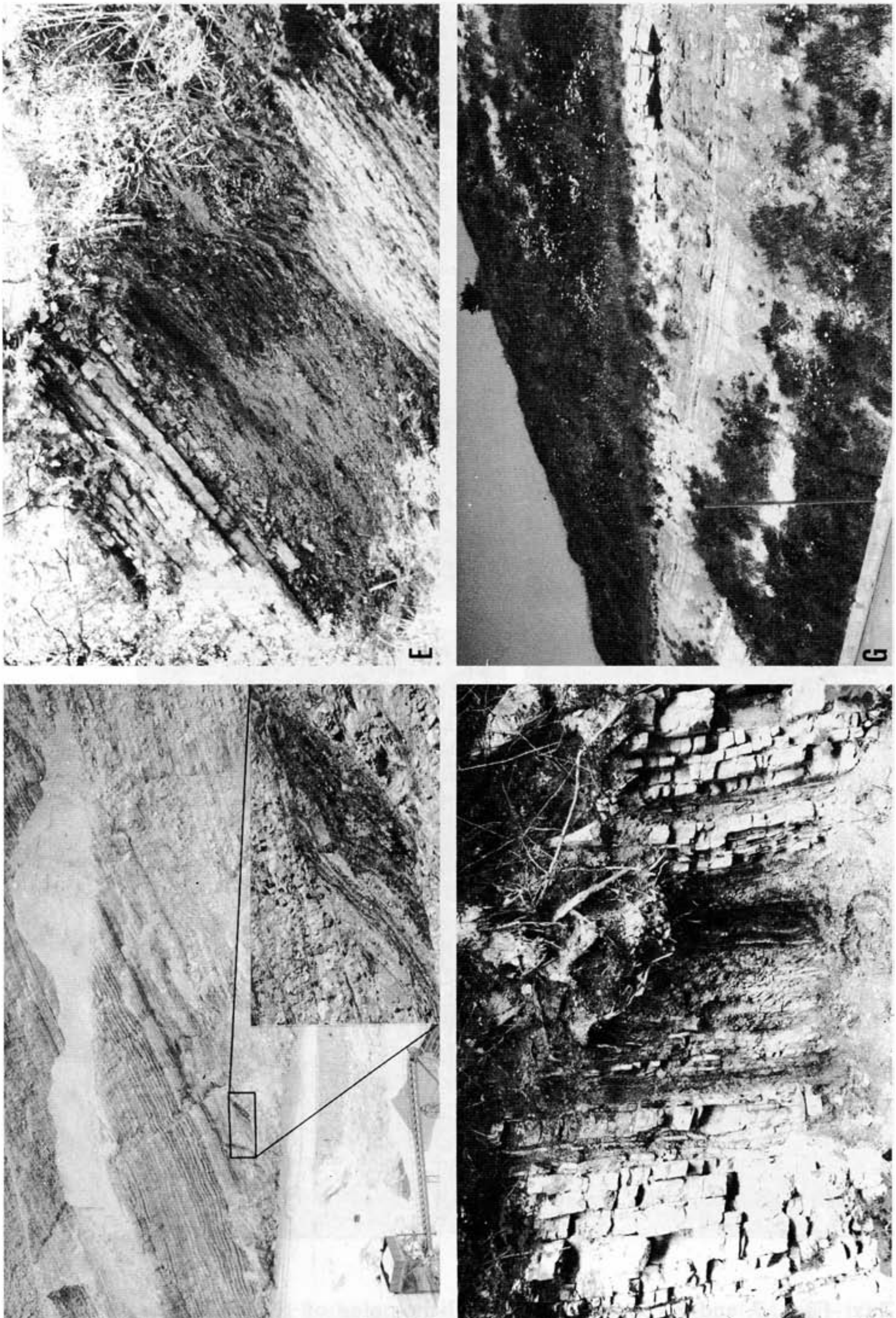
The Selli Level contains abundant fish remains (Text-Figs. 9 and 10) which occur mainly as scales and vertebrae, commonly along parting in the mudstone and shale. Entirely unbroken remains rarely occur. On the basis of the radiolarian assemblages, the Selli Level is placed within the Stichocapsa euganea Zone of SCHAAF (1981). Based on calcareous nannofossils and planktonic Foraminifera which occur above and below it, the Selli

Table 1. Main mineralogical composition (%) of some samples coming from the Selli Level (1) and from the sediments just above and below it (2). Key to symbols: Q = quartz, KF = alkaline feldspar, P = plagioclase, C = calcite, F = phyllosilicate + accessories.

			Q (%)	KF (%)	P (%)	C (%)	F+a (%)
Fiume Burano section	1	$\bar{x}$	37.5	2.0	1.5	-	59.0
		δ	2.5	2.0	1.4	-	4.0
	2	$\bar{x}$	19.2	2.0	1.0	31.2	46.5
		δ	5.0	1.2	1.0	9.5	4.5
Monte li Rossi section	1	$\bar{x}$	42.4	1.0	0.6	-	56.0
		δ	6.5	1.2	1.2	-	5.4
	2	$\bar{x}$	19.0	2.5	1.5	33.5	43.5
		δ	3.0	0.5	1.2	3.5	1.5
Pizzo di Meta section	1	$\bar{x}$	42.0	2.2	1.0	-	54.7
		δ	9.7	1.3	1.0	-	9.6
	2	$\bar{x}$	12.5	1.5	0.5	48.7	36.7
		δ	2.1	0.8	0.0	6.0	5.5
Poggio le Guaine section	1	$\bar{x}$	42.4	1.0	0.6	-	56.0
		δ	6.5	1.2	1.2	-	5.4
	2	$\bar{x}$	13.3	0.4	0.7	61.0	20.3
		δ	4.7	0.8	1.8	10.4	12.8

Text-Fig. 6. Some Selli Level outcrops. A = S.S. 77-km 53.500 section; B = S.S. Apecchiese-km 32.800 section: note the slumped zone in the upper part of the Maiolica Limestone; C = Monte Cònero section: photograph shows the sequence spanning from the uppermost part of the Maiolica Limestone (M) to the Scaglia Rossa Formation (SR); this sequence is not complete but punctuated by some hiatuses. According to CRESCENTI (1969), the lowermost part of the Marne a Fucoidi (MF) (including the Selli Level) is overlain by the Scaglia s. l. According to COCCIONI & LAGHI (in prep.), the base of the Scaglia s. l. is here represented by two centimetric layers of late Albian (Vraconian) age and then referred to the Scaglia Bianca s. s. (SB). Some layers, for a total thickness of 60 cm, overlap these levels. They are of late Turonian age (Scaglia Rossa s. s.). Finally, sediments (SRS-C) dated close to the Santonian-Campanian boundary overlie the sequence described so far. D = Contessa section: the Selli Level is involved in a slumped (or back-thrusted?) zone at the contact between Maiolica Limestone and overlying Marne a Fucoidi; E = Monte li Rossi section; F = Fiume Burano section: the calcareous beds on the left represent the basal portion of the Marne a Fucoidi; G = Pizzo di Meta section.




D
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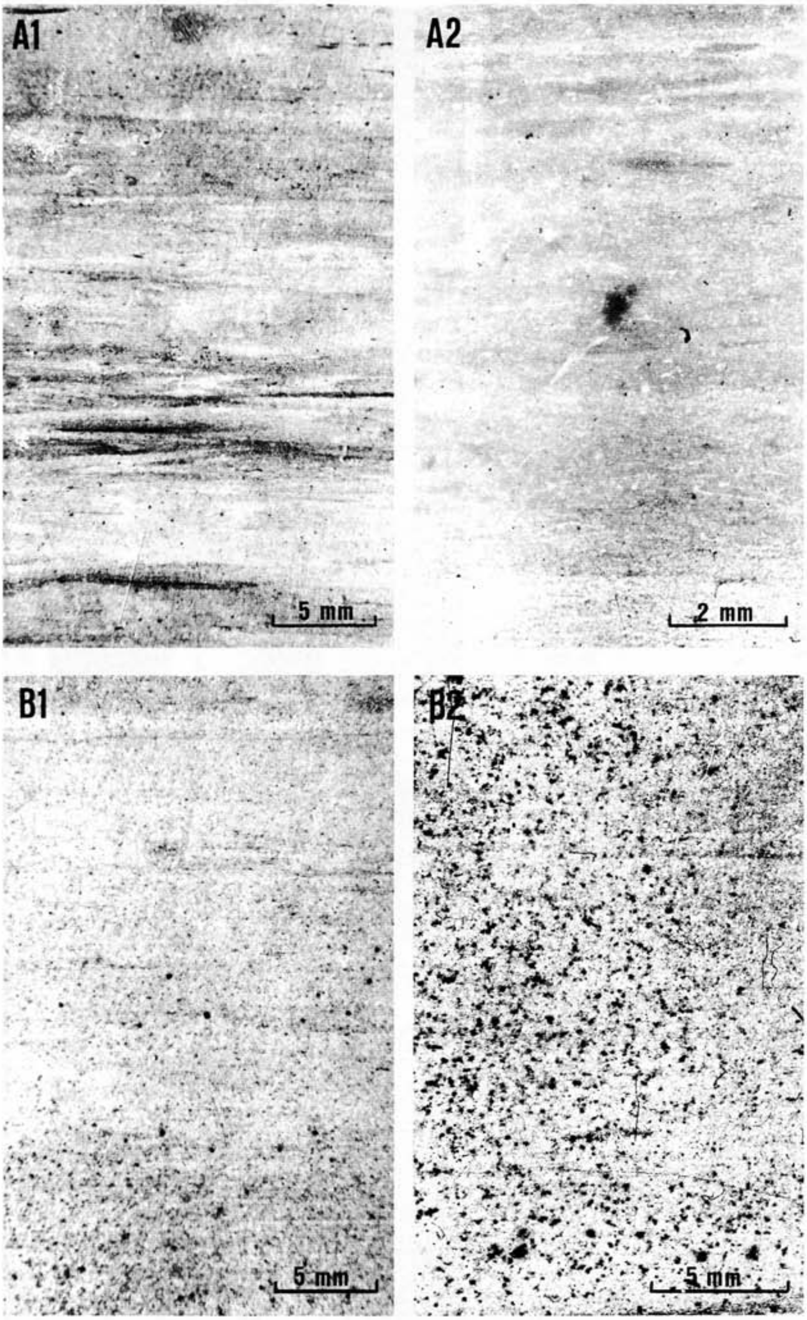
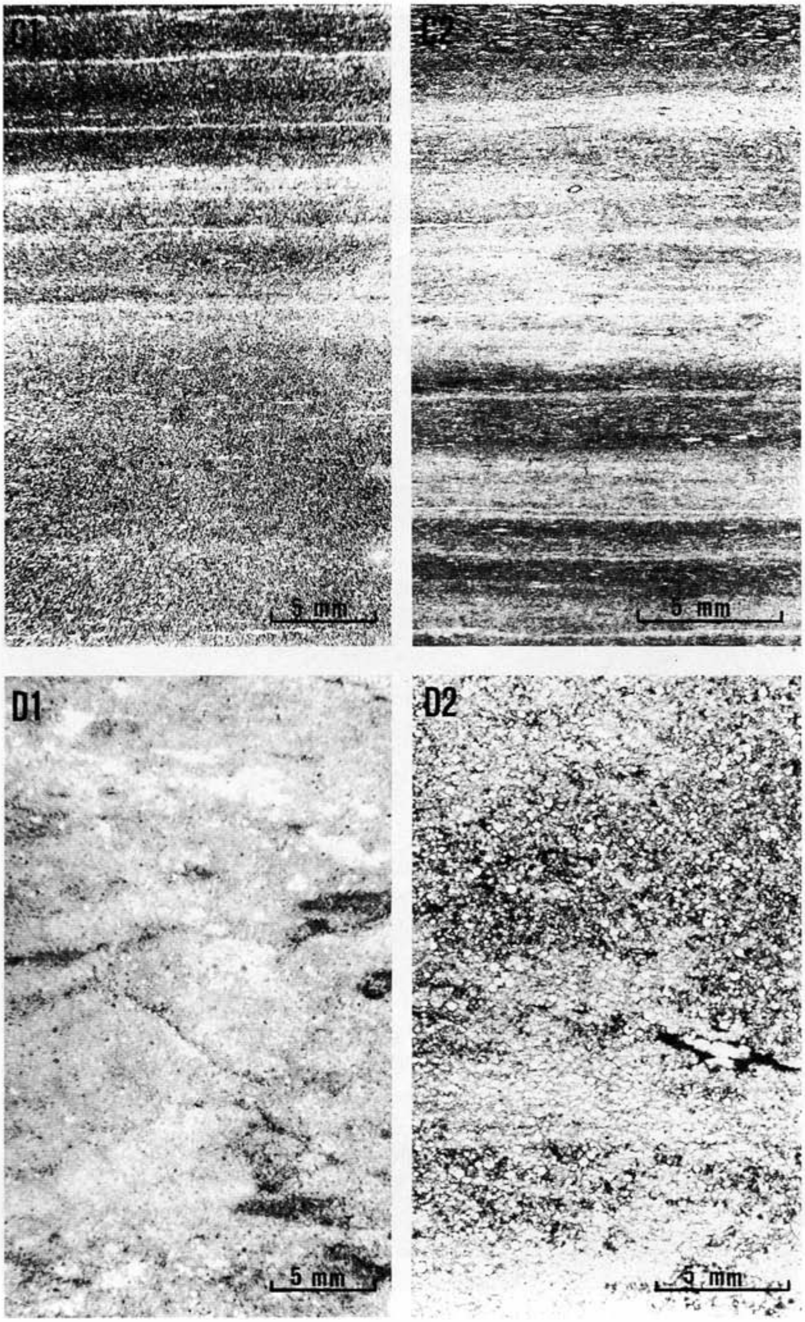
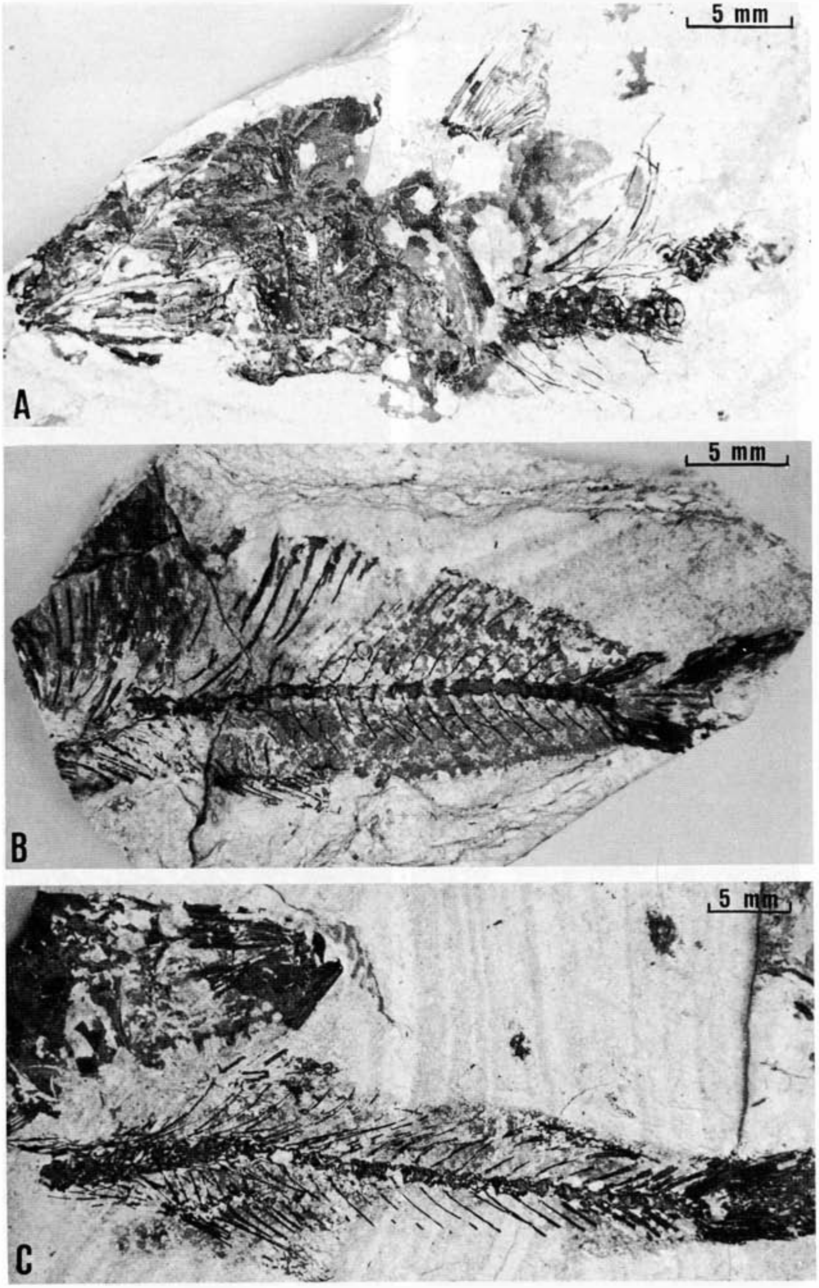
F
G



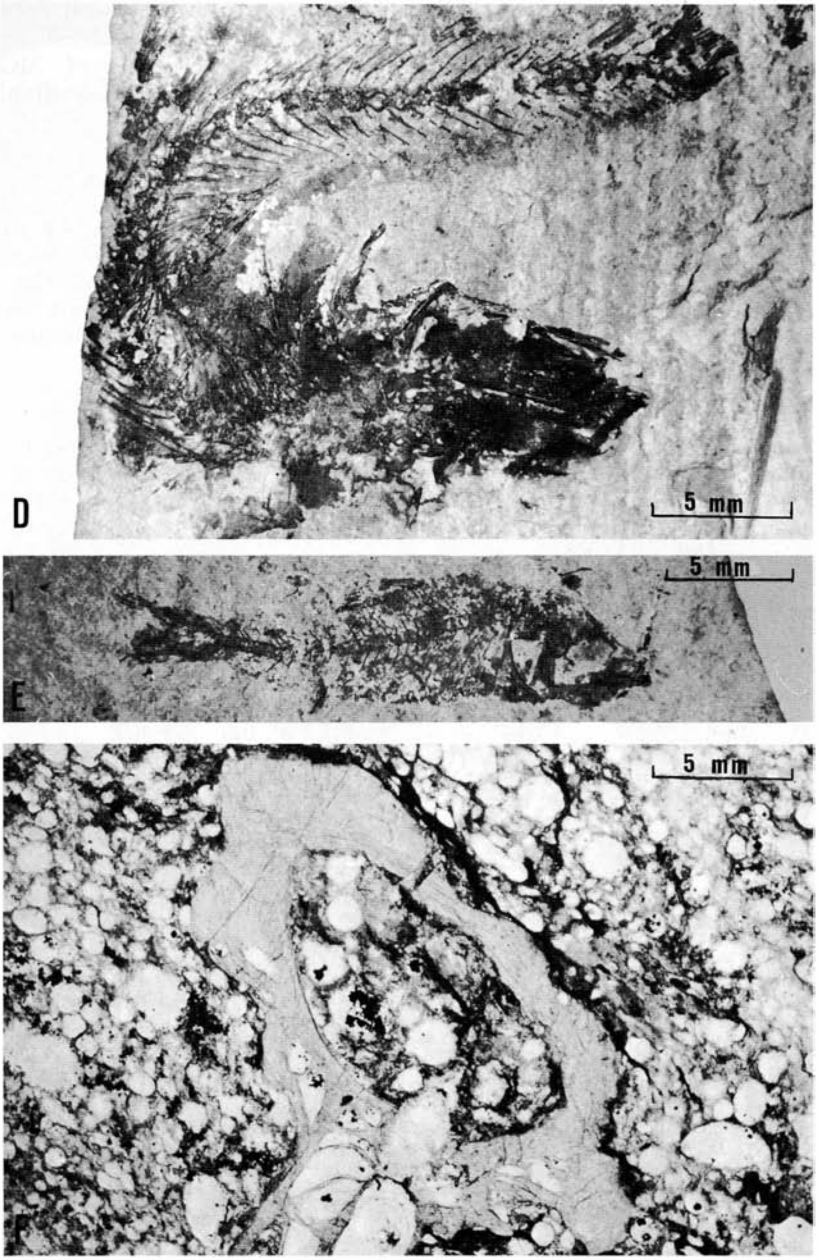
Text-Figs. 7 and 8. Some distinctive lithologies of the Selli Level. A = bio-turbated, greenish-grey mudstone; *Planolites* and *Chondrites* are the trace fossils recognized. B = laminated, olive-grey mudstone; pyritized radiolaria



occur. C = laminated black-shale. D = radiolarian-rich layer. (A1, B1, C1, and D1 = polished-slab photographs; A2, B2, C2, and D2 = thin section photographs of same sample).



Text-Figs. 9 and 10. A-E = fish remains on laminae surface of a laminated, greyish-yellow phosphatic mudstone from the Selli Level of the Monte li



Rossi section; F = thin section of fish vertebra in a radiolarian-rich layer from the same section.

Level is placed within the Chiastozygus litterarius nannofossil Zone of THIERSTEIN (1973) and in the Globigerinelloides gottisi/G. duboisi and/or Globigerinelloides maridalensis/G. blowi foraminiferal Zone of SIGAL (1977). Therefore, the age of the Selli Level is Lower Aptian (Bedoulian).

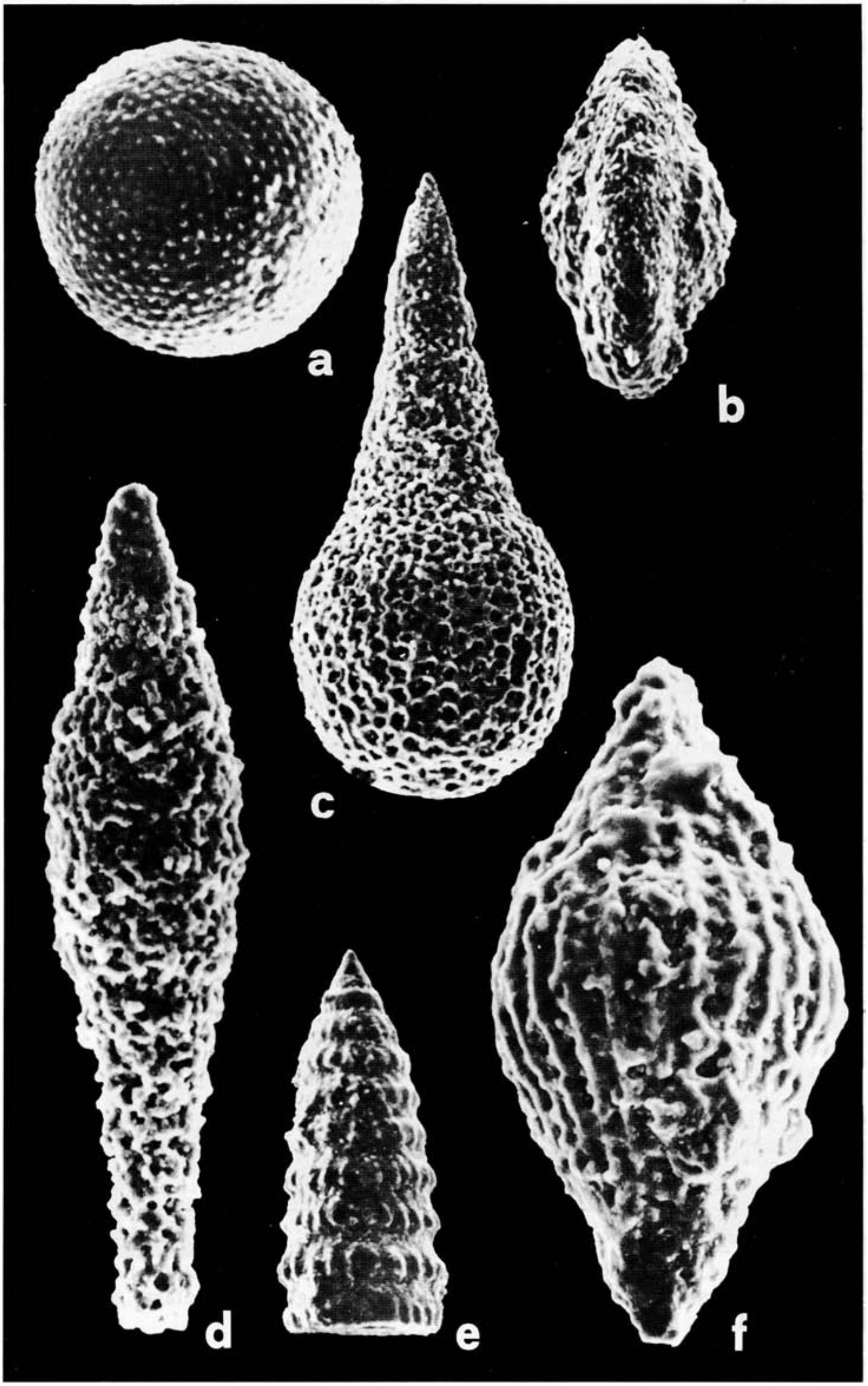
3. Mineralogy

A detailed investigation was carried out on the Selli Level in the Fiume Burano, Monte li Rossi, Pizzo di Meta, and Poggio le Guaine sections (Tables 1 and 2). Several samples coming from the Selli Level and the immediately adjacent overlying and underlying layers (either bituminous or not) were studied by using X-ray diffractometry and optical and scanning electron microscopy.

Table 2. Mineralogical composition (%) of the clay mineral fraction of some samples coming from the Selli Level (1) and from the sediments just above and below it (2). Key to symbols: K = kaolinite, I = illite, IS = illite-smectite, S = smectite, ClV = chlorite-vermiculite, Cl = chlorite.

			K (%)	I (%)	IS (%)	S (%)	ClV (%)	Cl (%)
Fiume Burano section	1	$\bar{x}$	5.0	32.5	12.5	37.5	5.0	7.5
		δ	5.0	2.5	2.5	2.5	5.0	2.5
	2	$\bar{x}$	-	33.3	20.0	46.6	-	-
		δ	-	4.7	0.0	4.7	-	-
Monte li Rossi section	1	$\bar{x}$	-	22.5	17.5	33.7	11.2	15.0
		δ	-	4.3	2.5	4.1	2.1	3.5
	2	$\bar{x}$	-	32.5	22.5	45.0	-	-
		δ	-	2.5	2.5	5.0	-	-
Pizzo di Meta section	1	$\bar{x}$	6.2	30.0	16.2	26.2	10.0	11.2
		δ	4.2	0.0	2.1	2.1	0.0	2.1
	2	$\bar{x}$	-	48.3	20.0	31.6	-	-
		δ	-	4.7	4.1	2.3	-	-
Poggio le Guaine section	1	$\bar{x}$	8.3	30.0	17.5	33.3	2.5	8.3
		δ	2.3	5.0	4.7	3.7	3.8	3.7
	2	$\bar{x}$	-	28.8	19.1	51.9	-	-
		δ	-	8.5	4.7	8.3	-	-

Text-Fig. 11. SEM photomicrographs of some distinctive radiolaria from the Selli Level. a = *Cenosphaera* sp., Poggio le Guaine section, x 150; b = *Spongodiscus* sp., Poggio le Guaine section, x 150; c = *Stichocapsa euganea* (SQUINABOL), Poggio le Guaine section, x 150; d = *Eucyrtis tenuis* (RUST), Poggio le Guaine section, x 300; e = *Pseudodictyomitra carpatica* (LOZYNIAK), Poggio le Guaine section, x 150; f = *Archaeodictyomitra lacrimula* (FOREMAN), Monte li Rossi section, x 300.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e
f

The main mineralogical composition, as well as that of the clay mineral fraction of the Selli Level, differs from that of the samples coming from the sediments just above and below it and, generally, from that of the Marne a Fucoidi as a whole because of the high quartz content but, above all, of the absence of carbonate minerals (see COCCIONI et al. 1988). This appears to be independent of the presence or absence of organic matter in the different strata making up the level. The assemblage of clay minerals is characterized by the presence of smectite, illite-smectite and illite (an assemblage typical of the Marne a Fucoidi), and of chlorite-vermiculite, chlorite and kaolinite. Petrographical observations by optical microscopy have pointed out the presence of Radiolaria, usually filled with chalcedony and at times representing up to 80 % of the entire sample, with variable granulometry without sorting. More or less closely-packed laminations are evident. In the bituminous levels, above all, spherical masses of pyrite, often of considerable size, are present. The terrigenous, non-argillaceous fraction is always present with a very fine, non-resolvible, granulometry. Very angular quartz grains and some lamina of muscovite are seldom found. By using optical microscopy, we observed clasts of irregular form, at times convoluted, and similar to fluidal structures. In parallel nicols their colours vary from light brown to reddish-brown, while in crossed nicols they present a low birefringence. Because of the difficulty in recognizing these clasts with optical microscopy, the same thin sections were analysed by microanalysis in conjunction with the SEM. From this analysis it resulted that the above-mentioned clasts are composed primarily of calcium and phosphorus (see Text-Fig. 12). Studies of large samples have confirmed the presence of phosphate compounds. Using X-ray diffractometry, it has been possible to assign these phosphate compounds to the hydroxyapatite class.

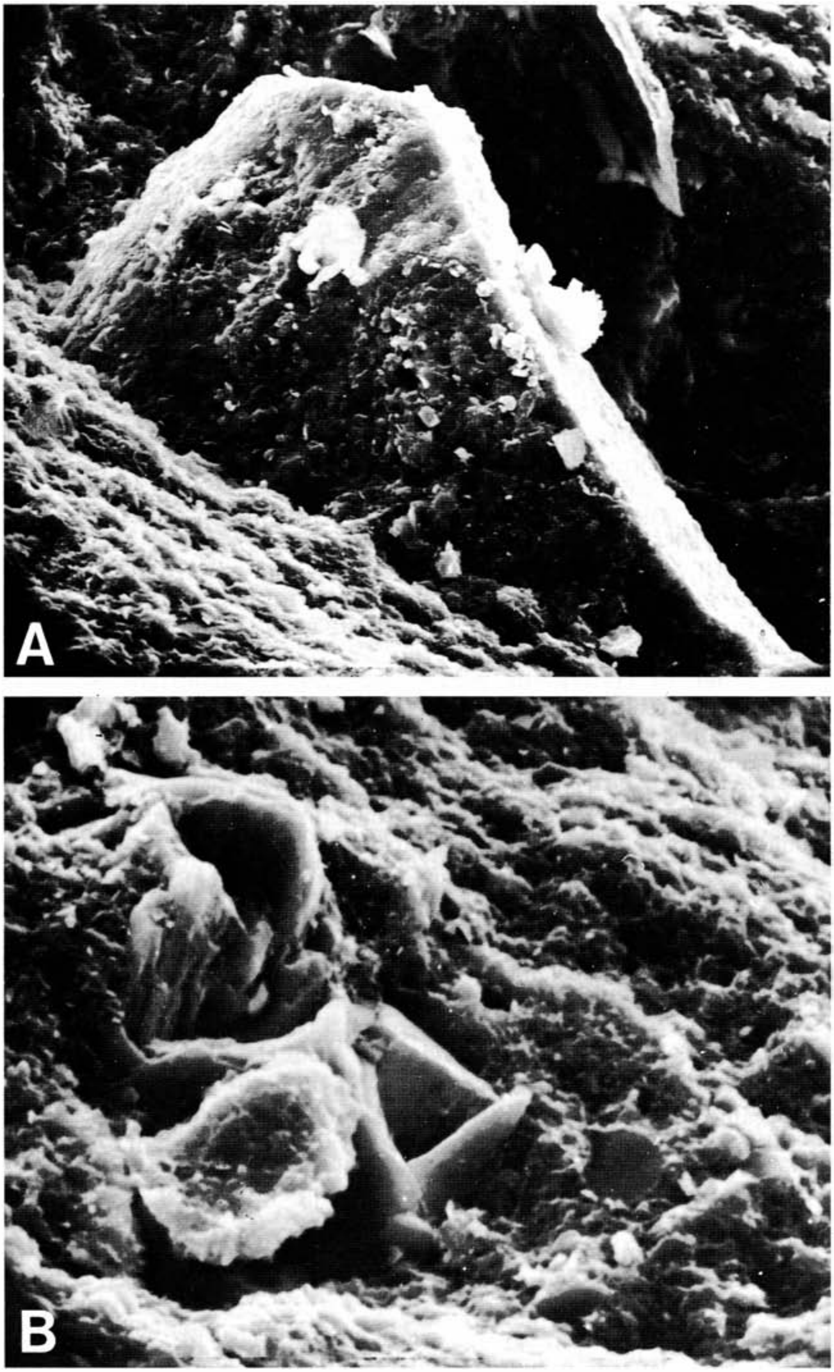
The total organic carbon content of some samples coming from the Selli Level of the Poggio le Guaine section (see Text-Fig. 4) was measured; it ranges from 0.85 to 9.20 % (COCCIONI et al. 1987).

4. Conclusions

The stratigraphical, sedimentological and mineralogical data suggest that the widespread Selli Level is a short-lived, anoxic event. There is no evidence of any redeposition.

The Selli Level is closely similar, because of fossil content, litho- and sedimentological characteristics and high total organic carbon (PRATT & KING 1986, COCCIONI et al. 1987), to the well-known Bonarelli Level (dated close to the Cenomanian-Turonian boundary) which characterizes the uppermost part of the Scaglia Bianca Formation in the Umbro-Marchean Apennines (see Text-Fig. 1; for more information see ARTHUR & PREMOLI SILVA 1982). Moreover, the depositional sequences of both Selli and Bonarelli Levels show a very similar vertical trend. Schematically, this trend is as follows (from the bottom to the top):

Text-Fig. 12. A: SEM photomicrograph of phosphatic clast (mean chemical composition: Ca = 68 %, P = 30 % and Si = 1.5 %), x 1200; Selli Level at Monte li Rossi. B: SEM photomicrograph of aggregate clast, x 600; Selli Level at Fiume Burano.


A
B

Selli Level	Bonarelli Level
1. Greyish limestone with black chert beds and nodules and interbedded black marlstone;	1. Greyish limestone with black chert beds and nodules and interbedded black marlstone;
2. Selli Level;	2. Bonarelli Level;
3. Greenish-grey limestone with greenish-grey and light olive-grey chert beds and nodules;	3. Greenish-grey limestone with greenish-grey and light olive-grey chert beds and nodules;
4. Reddish marlstone.	4. Reddish limestone with greyish-red and reddish-brown chert beds and nodules.

Consequently, because of such a similarity, the Selli and the Bonarelli Levels probably deposited under analogous palaeoceanographic conditions.

Several authors tried to explain what mechanism was responsible for the deposition of the Bonarelli Level (JENKYNS 1980, ARTHUR & PREMOLI SILVA 1982, DE BOER 1982 and 1986, HERBIN et al. 1986, KUHNT et al. 1986, ARTHUR, SCHLANGER & JENKYNS 1987, DE GRACIANSKY et al. 1987, SCHLANGER et al. 1987). However, we are far from a univocal response except that the Bonarelli Level formed during a maximal world-wide transgression (see HAQ, HARDENBOL & VAIL 1987). Such a relationship is unknown for the Selli Level.

In order to clarify the palaeoceanographic conditions which lead to the deposition of this level, research is in progress involving further more detailed studies. First results emerging from the semiquantitative analysis of the calcareous nannofossil assemblages (COCCIONI et al. 1988, PREMOLI SILVA, ERBA & TORNAGHI 1988) indicate that the sedimentation just below and above the Selli Level, probably occurred under moderate primary productivity and warm water conditions.

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Mid-Cretaceous Platform Drowning, Current-Induced Condensation and Phosphogenesis, and Pelagic Sedimentation along the Eastern Helvetic Shelf (Northern Tethys Margin)

KARL B. FÖLLMI, Zürich

With 7 Text-Figures

FÖLLMI, K. B. (1989): Mid-Cretaceous Platform Drowning, Current-Induced Condensation and Phosphogenesis, and Pelagic Sedimentation along the Eastern Helvetic Shelf (Northern Tethys Margin). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 585-606. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: In mid-Cretaceous times, the European northern Tethys margin experienced an important evolutionary phase during which a) a Barremian to Lower Aptian carbonate platform drowned, b) a current-dominated sedimentary regime, persistent through Aptian, Albian, and early Cenomanian, led to retarded sediment accumulation, condensation, phosphogenesis and sediment reworking, and c) from latest Albian onward, a pelagic regime blanketed the shelf with pelagic micrites. This general deepening-upward process was punctuated by several short phases of reinforced erosion, transport, and redeposition of eroded sediments, which in part correlate to orogenic phases along the Tethyan eo-Alpine collision front (at or near the Aptian-Albian, Cenomanian-Turonian, Turonian-Coniacian, and Santonian-Campanian boundaries). The development of the mid-Cretaceous Helvetic triad, carbonate platform, phosphatic sediments, and pelagic carbonates, was in phase with a worldwide observed episode of platform drowning, formation of detrital-rich, phosphatic and/or glauconitic sediments, and subsequent deposition of pelagic sediments. This synchronization of sedimentary regimes on a global scale is attributed to the mid-Cretaceous episode of enhanced volcanism and increased spreading rates.

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1. Introduction

1.1 Mid-Cretaceous times and sediment patterns

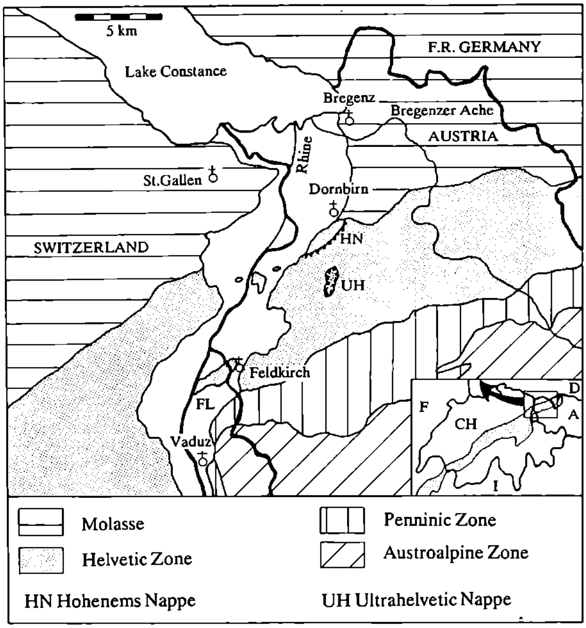
The mid-Cretaceous period (i. e., Aptian to Turonian) is generally considered as a tectonically active phase, during which spreading rates increased and intense off-ridge volcanism occurred (SCHWAN 1980, SCHLANGER et al. 1981, ARTHUR et al. 1985, REYMENT & BENGTSON 1986). Directly or indirectly related to this phase is the well-known and pronounced eustatic sea-level rise and highstand (late Aptian to Turonian times), and a modeled high atmospheric CO_2 rate and strong elevation of the Calcite Compensation Depth (CCD) (THIERSTEIN 1979, HAQ et al. 1987, BERNER & LASAGA 1989). During the mid-Cretaceous, the Atlantic started to open between West Africa and northern South America, as well as between Iberia and North America. As a result, the African and European plate systems collided, inducing the gradual closure of the oceanic Tethys (SAVOSTIN et al. 1986, ZIEGLER 1987, 1988, LE PICHON et al. 1988).

The mid-Cretaceous paleoceanographic and tectonic episode was accompanied by a remarkable synchronization of global sedimentary patterns. Aptian and Albian times are noted for the worldwide demise of shallow-water carbonate platforms (SCHLAGER 1981), and for the widespread deposition of terrigeneous sediments, commonly in association with phosphates and glauconites (referred to as "Gault" in many occurrences). In the late Albian and Cenomanian, pelagic sedimentary regimes spread across continental margins and epicontinental seas ("Cenoman-Transgression" of SUESS 1883). A typical (but certainly not omnipresent) mid-Cretaceous sequence from tropical and temperate continental margins and epicontinental seas would thus consist of shallow-water carbonates, overlain by terrigeneous, phosphatic and glauconitic sediments, overlain by calcareous pelagic sediments. Such a sequence developed along the southern European northern Tethys margin, traceable from southern Spain, via southeastern France and the Helvetic Alps, to the western Carpathians (OUWEHAND 1987, DELAMETTE 1988a).

1.2 Mid-Cretaceous sediments in the western Austrian Helvetic Alps

The western Austrian Helvetic Alps offer a good insight to the evolution of the mid-Cretaceous sequence, carbonate platform - terrigeneous and phosphatic sediments - pelagic carbonates, for they include a well-developed transect through the former Helvetic shelf including distal areas of the carbonate platform (here referred to as inner shelf), the platform margin (inner shelf margin), and proximal shelf areas beyond the platform margin (outer shelf). This transect is exposed in a single nappe tract, known as the Säntis Nappe (Text-Fig. 1). Parts of the proximal inner shelf and the distal outer shelf are preserved in the Hohenems Nappe, and in the Ultrahelvetic Nappe remnant of the Hohe Kugel area, respectively (Text-Fig. 1).

This paper makes an attempt to give an interpretation of the mechanisms, which contributed to the development of the sequence, carbonate platform - terrigeneous and phosphatic sediments - pelagic carbonates, in the eastern Helvetic area of the northern Tethys margin. It includes a brief introduction to the different sediment configurations which are met



Text-Fig. 1. Distribution of Helvetic sediments representing the former passive margin along the northern Tethys in western Austria.

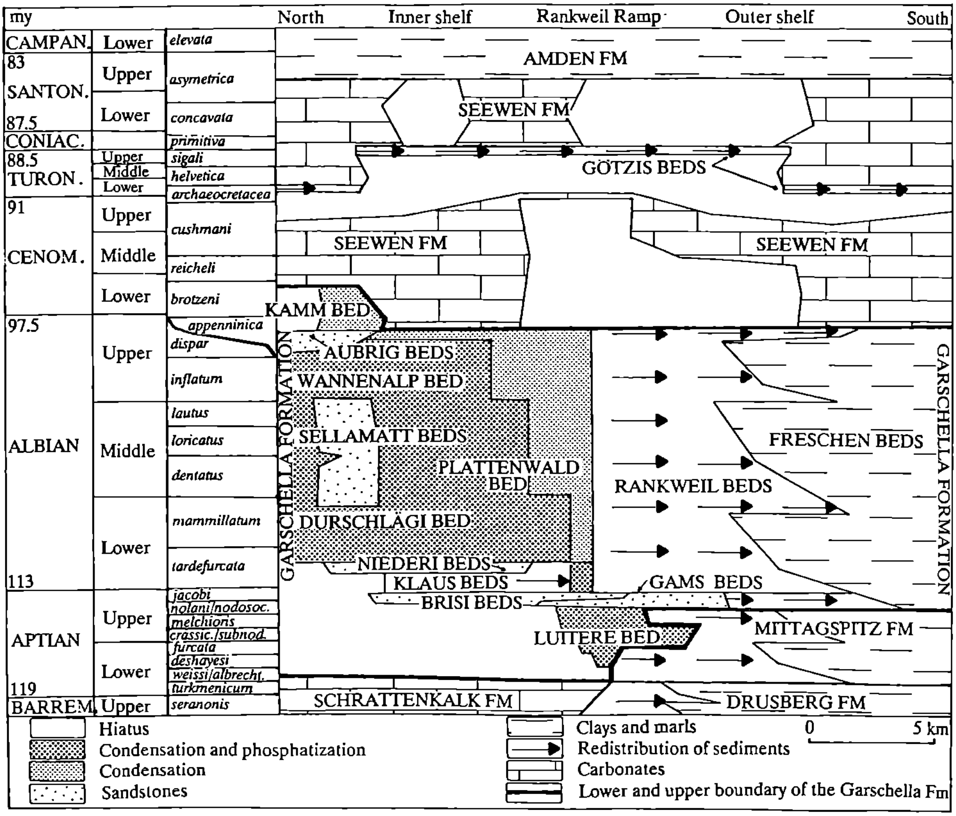
in the western Austrian mid-Cretaceous Helvetic transect, and focusses on a) the process of carbonate drowning, b) the genesis and development of condensed phosphatic beds in a current-dominated sedimentary regime, and c) the possible impact of eo-Alpine orogenic phases on the Helvetic sediment configuration.

Inferences made in this work are based on a detailed surveyance of the western Austrian Helvetic Alps during 1979-1986, in the course of which approximately 300 sections have been measured and sampled. Biostratigraphy is based on ammonoids and inoceramid bivalves (Aptian-Albian), and on globotruncanid foraminifera (from late Albian onward; Text-Fig. 2). Detailed stratigraphic descriptions are found in FÖLLMI (1986) and FÖLLMI & OUWEHAND (1987); a taxonomic review of ammonoids collected during fieldwork is given in FÖLLMI (1989a), and a general synthesis will appear in FÖLLMI (1989b).

2. Sediment patterns on the western Austrian Helvetic shelf during Aptian - Santonian times

2.1 Platform carbonates and outer shelf sediments

During Barremian time, a carbonate platform prograded along the southern European northern Tethys margin. Platform construction and progradation was controlled by a) biological sediment producers such as commonly patch-



Text-Fig. 2. Sediment distribution diagram through time and space. "Rankweil Ramp" corresponds to the inner shelf margin (from FÖLLMI 1989b).

reef bounded communities of red and green algae, stromatoporoids, scleractinean corals, and rudistid bivalves, b) dynamo(bio)chemical processes, forming large ooid shoals, and c) high-energy events (probably storms), distributing sediments (documented in the Barremian to Lower Aptian **Schrattenkalk Formation**; Text-Figs. 2, 3; e. g., SCHOLZ 1979, BOLLINGER 1986).

Episodically, platform sediments were transported into the muddy, marly outer shelf, as is documented in thick and monotonous limestone-marl alternations (Barremian to Lower Aptian **Drusberg Formation**; Text-Figs. 2, 3). A broad transition zone between the platform and the outer shelf, and

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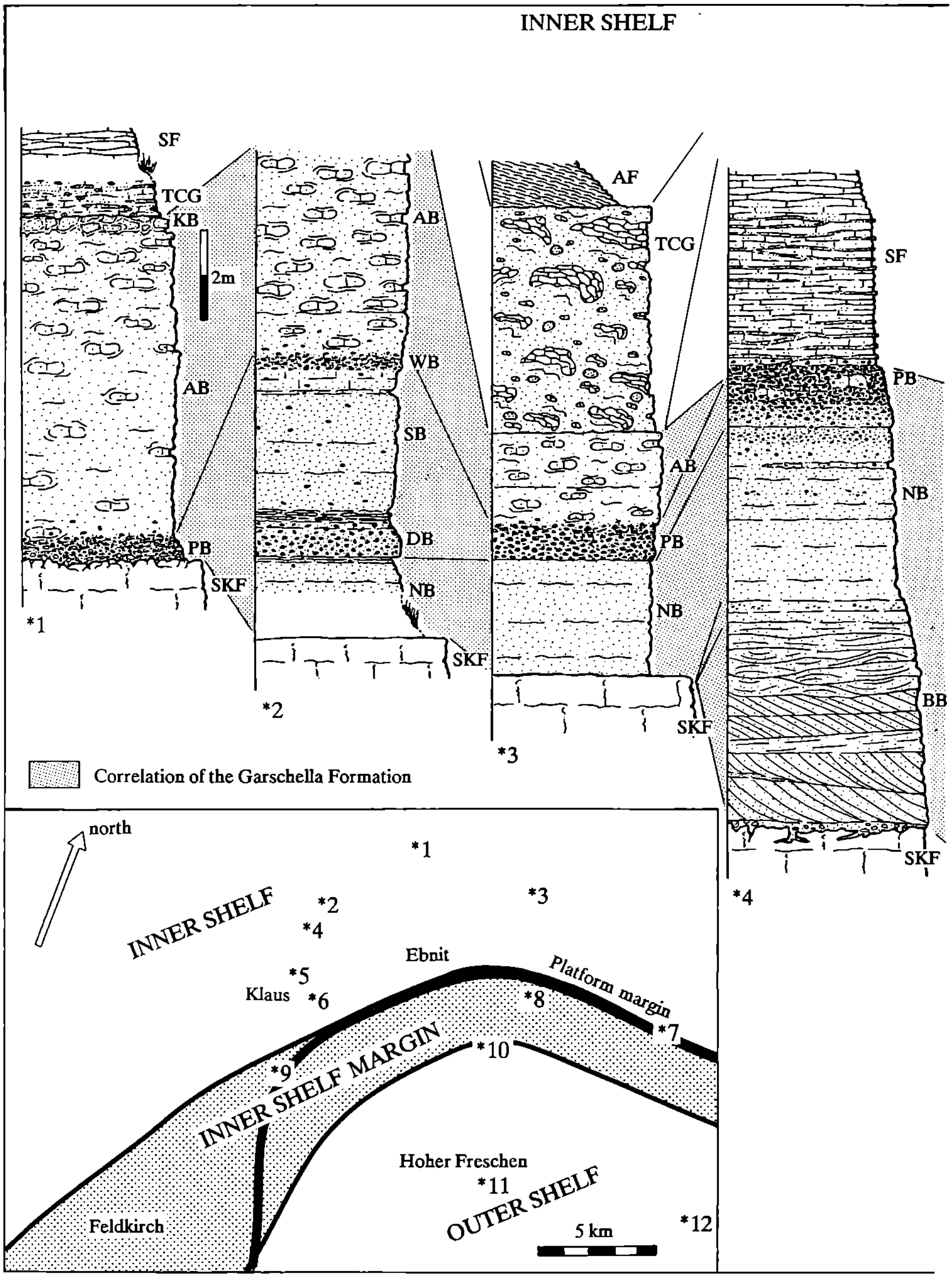
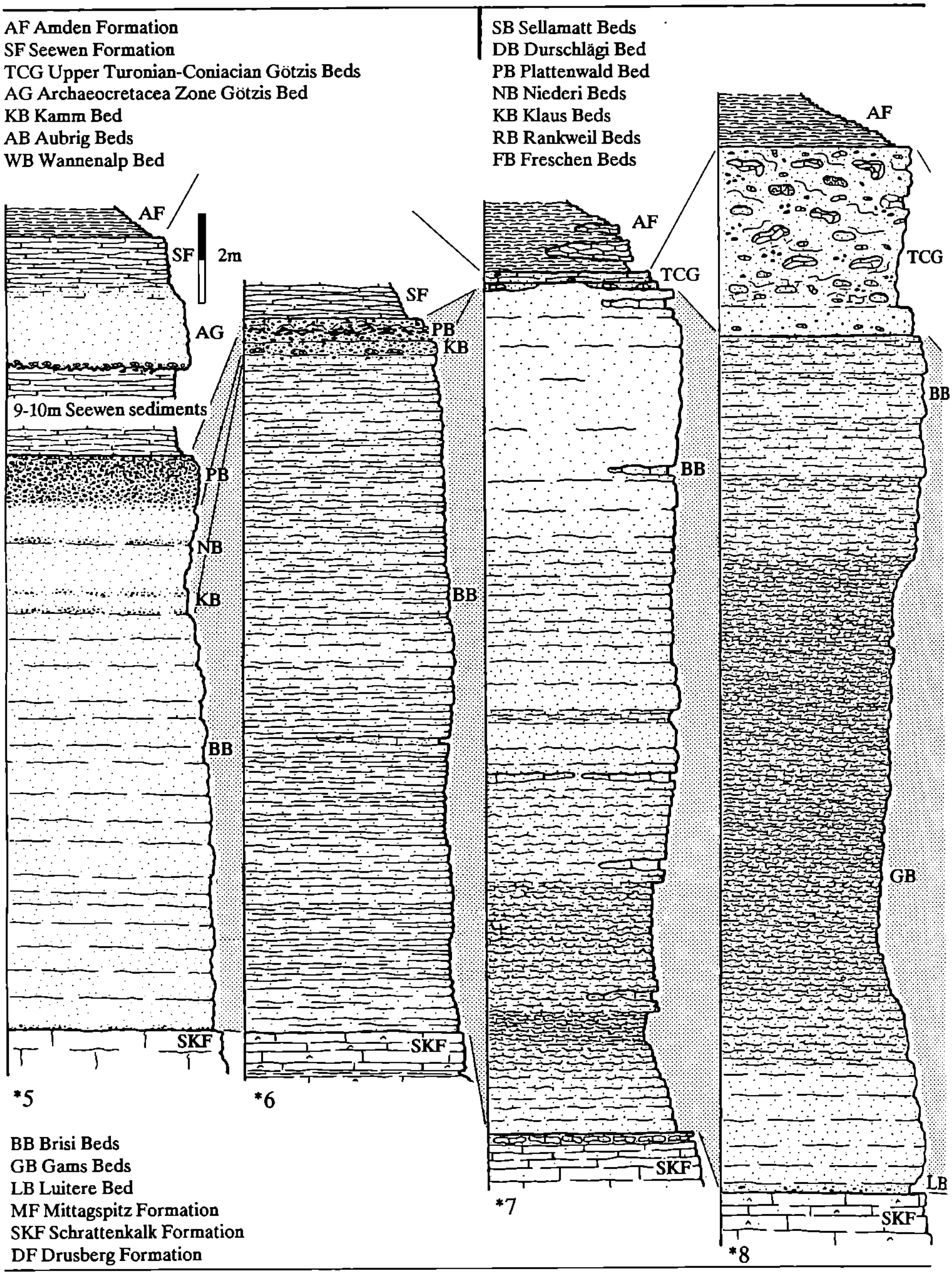
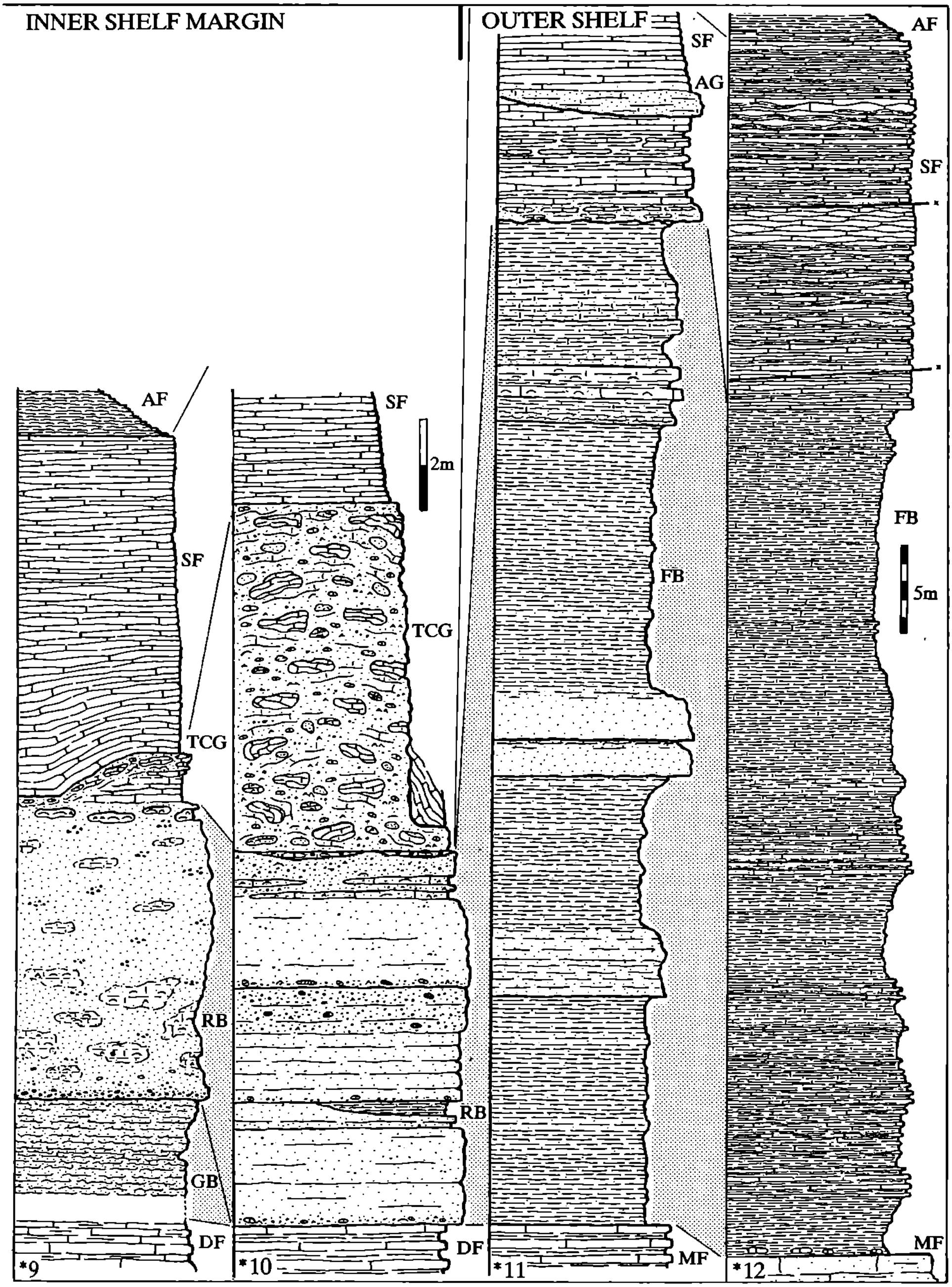
Text-Fig. 3. Representative mid-Cretaceous sections from the western Austrian Helvetic Alps (Säntis Nappe), arranged in a proximal-distal direction. Their position is indicated in a simplified palinspastic map (after FÖLLMI 1986).


INNER SHELF
SF
TCG
KB
2m
AB
PB
SKF
*1
AB
WB
SB
DB
NB
SKF
*2
AF
TCG
AB
PB
NB
SKF
*3
SF
PB
NB
BB
SKF
*4
Correlation of the Garschella Formation
north
*1
*2
*3
*4
INNER SHELF
Ebnit
Platform margin
*5
Klaus
*6
*8
*7
*9
*10
INNER SHELF MARGIN
Hoher Freschen
*11
OUTER SHELF
Feldkirch
5 km
*12


AF Amden Formation
SF Seewen Formation
TCG Upper Turonian-Coniacian Götzis Beds
AG Archaeocretacea Zone Götzis Bed
KB Kamm Bed
AB Aubrig Beds
WB Wannenalp Bed
SB Sellamatt Beds
DB Durschlägi Bed
PB Plattenwald Bed
NB Niederi Beds
KB Klaus Beds
RB Rankweil Beds
FB Freschen Beds
AF
SF
2m
AG
9-10m Seewen sediments
PB
NB
KB
BB
SKF
*5
SF
PB
KB
BB
SKF
*6
AF
TCG
BB
SKF
*7
AF
TCG
BB
GB
LB
SKF
*8
BB Brisi Beds
GB Gams Beds
LB Luitere Bed
MF Mittagspitz Formation
SKF Schrattenkalk Formation
DF Drusberg Formation



INNER SHELF MARGIN
OUTER SHELF
AF
SF
TCG
RB
GB
DF
*9
*10
2m
AG
FB
MF
*11
*12
5m

the absence of large erosive channels, indicate the presence of a gentle, homoclinal ramp along the distal platform margin (READ 1985).

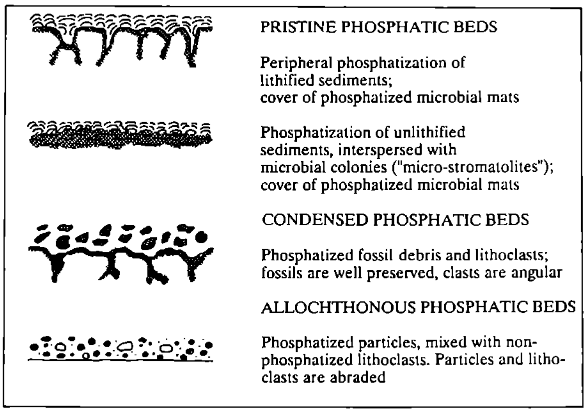
2.2 Condensation, phosphatization and sediment reworking on the inner shelf, impinging Oxygen Minimum Zone (OMZ) on the outer shelf

2.2.1 Inner platform-based shelf

In early Aptian time, a gradual change in sedimentary patterns occurred on the Helvetic shelf, most probably induced by the slow landward shift of a Tethyan current system onto the shelf (sections 3.1, 3.2; DELAMETTE 1985, 1988a, b, FÖLLMI 1986, 1989b, OUWEHAND 1987). Apparently incompatible with the presence of this current system, platform-carbonate production ceased diachronously in an east-west and distal-proximal trend (section 3.1; Text-Fig. 5).

Between early Aptian and early Cenomanian, the inner shelf was dominated by low net sediment-accumulation rates, erosion, redeposition of eroded sediments, and phosphatization, documented in the widespread occurrence of strongly condensed phosphatic beds (accumulation rates 2-20 cm/my; within the Lower Aptian to Lower Cenomanian **Garschella Formation**; Text-Figs. 2, 3). These beds display a framework of phosphatized particles and/or crusts, stratified in different types (Text-Fig. 4) forming typically thin (> 1 m) and complex layers.

1. **Pristine phosphates:** surficial phosphatization of previously lithified, rugged surfaces (carbonates and calcareous sandstones) overgrown by phosphatized, well-preserved microbial mats, in which siliciclasts or pelagic carbonates are trapped. Pristine lamellar phosphates occur as well on top of siliciclastic and glauconitic sandstone beds, which obviously were not lithified prior to phosphogenesis. The sandstone matrix includes abundant "micro-stromatolites", which probably represent microbial colonies, trapping and stabilizing the sands. The phosphatic lamellae are commonly overgrown by phosphatized, well-preserved microbial mats. Pristine phosphates have been excluded from sedimentary reworking processes, after phosphogenesis.



Text-Fig. 4. Overview of different stratification types within the phosphatic beds (from FÖLLMI 1989b).

2. **Condensed phosphates:** consist of thin packages of phosphatized, angular particles. Commonly, these particles include several accretionary phosphate generations, which are discernable by means of differing contents and grain sizes of included detritus, and sharp generation interfaces. These are occasionally coated by iron oxyhydroxids and encrusted by sessile foraminifera, bryozoa, or serpulids. Phosphatized fossils in the condensed nodular beds consist of an intimate mixture of representatives of different time zones and ecological habitats.

In the case of stronger condensation, packages of autochthonous crusts formed, commonly including large timespans (FÖLLMI et al. in prep.).

Multi-event condensed beds resulted from repetitive **Baturin cycling**, i. e., cycles of burial, phosphatization, winnowing and erosion, reexposure, burial, etc. (Text-Fig. 7; BATURIN 1971, KENNEDY & GARRISON 1975, BRANDT 1985, MULLINS & RASCH 1985).

3. **Allochthonous phosphates:** transported, commonly abraded phosphatic diaclasts (terminology from A. SEILACHER, pers. comm. 1989), usually mixed with non phosphatized lithoclasts, present in different kinds of gravity-flow deposits, or as fine-grained fractions in winnowed beds.

The Helvetic phosphatic beds usually include a combination of the above described stratification types, due to variations in hydrodynamics and sediment properties (section 3.2.2).

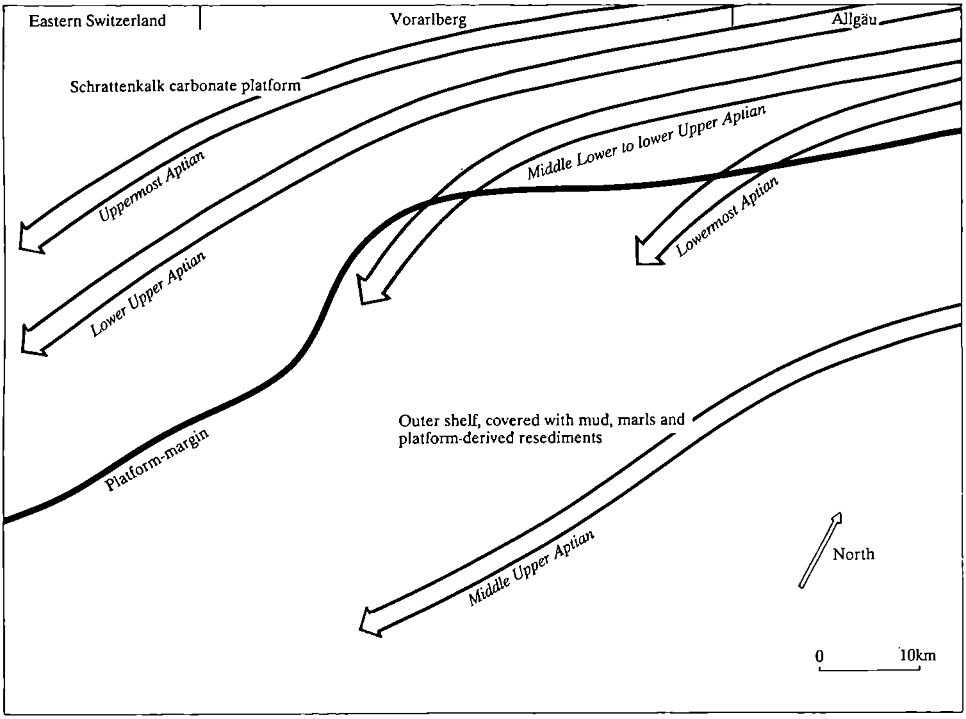
The western Austrian inner shelf phosphatic beds display the following trends in distribution and composition (Text-Fig. 6):

1. Non phosphatized sediments included in the phosphatic beds tend to become depleted in siliciclasts in a distal direction.
2. The degree of complexity of included, non phosphatized sediments decreases toward marginal, distal occurrences.
3. Toward the inner shelf margin, phosphatic beds tend to bundle into one bed; beds become generally thinner, and include larger timespans (e. g., latest Aptian to earliest Turonian in approximately 30 cm thick beds on the marginal inner shelf). Beds consist of multi-event winnowed nodular beds in proximal areas, predominantly of winnowed stacks of autochthonous phosphatic crusts in distal areas (Text-Fig. 6).
4. Although condensation processes continued on the entire inner shelf at least until late Albian, phosphatization ceased a) in the middle of the Tardefurcata Zone (early Albian) along the inner-shelf margin, b) at the end of the Mammillatum Zone (boundary Lower to Middle Albian) within an area adjacent to the margin, c) at the end of the Loricatus Zone (in the middle of Middle Albian) in intermediate inner-shelf areas, d) at the end of the Inflatum Zone (middle Upper Albian) in proximal inner-shelf areas, and e) at the end of the Brotzeni Zone (early Cenomanian) in most proximal inner-shelf areas. A strong correlation in time is observed between phosphatization and the presence of siliciclasts.

These phenomena indicate a) a diachronous termination of phosphogenesis along the inner Helvetic shelf in proximal directions, and b) a stronger and more persistent high-energy environment in distal inner-shelf areas (section 3.2; Text-Fig. 6). This latter observation is also consistent with the distribution of siliciclastic sandbodies on the inner, platform-based shelf.

During Aptian and Albian times, the inner, platform-based shelf experienced four phases of glauconitization and siliciclastic sand replenishment:

1. During the Nolani-Nodosocostatum Zone (and probably parts of Melchioris and Jacobi Zones: middle Upper Aptian), coarse-grained siliciclasts and calcareous bioclasts were distributed in approx. 50 m

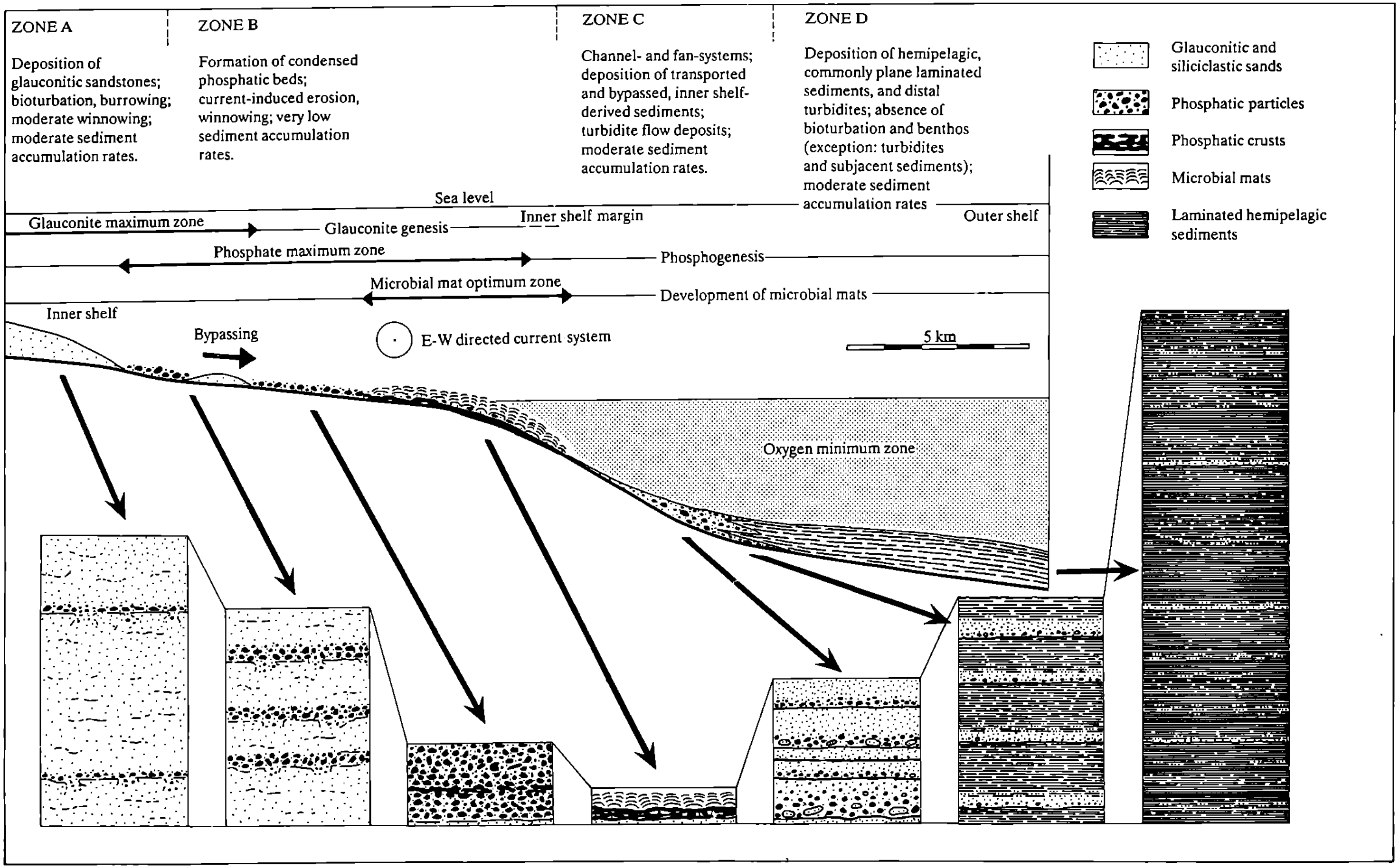


Text-Fig. 5. Drowning patterns of the Schrattenkalk carbonate platform, indicated by the position of the current axis (simplified palinspastic map; from FÖLLMI 1989b).

thick packages onto the entire inner shelf (**Brisi, Gams Beds**; Text-Figs. 2, 3). The high rates of detrital influx entirely overwhelmed the existing system of condensation and phosphogenesis for the timespan of the middle late Aptian, possibly also because of a short-term seaward current shift into outer shelf areas. A subsequent current shift onto proximal areas of the inner shelf not only terminated the influx of siliciclasts, but eroded major proximal portions of the sands as well.

2. Three phases of sand replenishment occurred in Middle Tardefurcata Zone (Lower Albian; **Niederi Beds**; Text-Figs. 2, 3);

Text-Fig. 6. Schematic diagram of a transect through the western Austrian distal inner shelf, inner-shelf margin and proximal outer shelf. Position of zones A - D is determined by the distal inner-shelf position of the current system. Vertical scale 1 : 40 exaggerated. Suite of representative and approximately time-equivalent sections in lower half of diagram (from FÖLLMI 1989b).



ZONE A
Deposition of glauconitic sandstones; bioturbation, burrowing; moderate winnowing; moderate sediment accumulation rates.
ZONE B
Formation of condensed phosphatic beds; current-induced erosion, winnowing; very low sediment accumulation rates.
ZONE C
Channel- and fan-systems; deposition of transported and bypassed, inner shelf-derived sediments; turbidite flow deposits; moderate sediment accumulation rates.
ZONE D
Deposition of hemipelagic, commonly plane laminated sediments, and distal turbidites; absence of bioturbation and benthos (exception: turbidites and subjacent sediments); moderate sediment accumulation rates
Glauconitic and siliciclastic sands
Phosphatic particles
Phosphatic crusts
Microbial mats
Laminated hemipelagic sediments
Sea level
Inner shelf margin
Outer shelf
Glauconite maximum zone
Glauconite genesis
Phosphate maximum zone
Phosphogenesis
Microbial mat optimum zone
Development of microbial mats
Inner shelf
Bypassing
E-W directed current system
5 km
Oxygen minimum zone

late Mammilatum to Lautus Zones (upper Lower to Middle Albian, **Sellamatt Beds**; Text-Figs. 2, 3); late Inflatum to early Dispar Zones (middle Upper Albian, **Aubrig Beds**; Text-Figs. 2, 3). During these phases, the current-induced sedimentary regime was **not** overpowered. Instead, the sands were shaped into elongated sand-bodies parallel to the inner-shelf margin, possibly sandwiched by current-offshoots (Text-Fig. 6). Larger amounts of sand were eroded after deposition and bypassed via distal inner-shelf areas of condensation and phosphatization into inner-shelf margin and proximal outer-shelf areas (preserved in the Freschen and Rankweil Beds; Text-Figs. 2, 3, 6).

Whereas the middle Upper Aptian phase of replenishment is documented over the entire inner shelf, the three Albian phases are only known from proximal and intermediate inner-shelf areas, with a proximality trend through Albian which is probably related to the general subsiding trend of the Helvetic shelf and rising sea levels (HAQ et al. 1987).

2.2.2 Inner-shelf margin

The inner-shelf margin, generally corresponding to the former carbonate-platform margin, experienced profound steepening and accentuation during the Aptian, especially during the latest Aptian (Jacobi Zone). The different subsidence patterns are most probably related to different compaction rates between the inner, platform-based shelf (early diagenetic lithified carbonates) and the outer, muddy and marly shelf, as well as to a probable tectonic phase at the end of the Aptian (section 3.3). The platform margin transformed from a gentle homocline during Barremian and early Aptian into a typical rampzone. This is documented by the covering sediments:

During Aptian time, the margin represented a broad transitional zone, mediating between the inner shelf area of condensation and phosphogenesis, respectively sand distribution (middle Upper Aptian), and the outer-shelf area, where hemipelagic marls and clays accumulated. Small and mobile channel and fan systems developed, in which eroded inner-shelf sediments were redistributed into proximal outer-shelf areas.

In latest Aptian, deep cutting channels formed on the marginal area. These channels differed from their Aptian precessors by a larger size (several 100 m wide, 20 to 30 m deep eroded into the middle Upper Aptian Gams Beds, and approximately 5 to 7 km long) and by their stability in position (throughout the Albian). Winnowed and bypassed, inner shelf-derived lithoclasts were transported along these channels (**Rankweil Beds**; Text-Figs. 2, 3). The margin represented a distinct facies boundary during latest Aptian and Albian, marking the distal boundary of the inner-shelf-condensed phosphatic beds.

2.2.3 Outer shelf

During Aptian and Albian, hemipelagic carbonates, marls and muds were deposited on the outer shelf. The hemipelagites interfinger proximally with channel and fan-system deposits, built up by inner shelf-derived sediments (Lower to lower Upper Aptian **Mittagspitz Formation**, and middle Upper Aptian to uppermost Albian **Freschen Beds**, latter included in the Garschella Formation; Text-Figs. 2, 3).

Outer shelf-background sediments, especially in the Freschen Beds, are characterized by the virtual absence of endo- and epibenthic organisms, by the presence of sediment laminations, as well as by abundant pyritized and phosphatized *Hedbergella*-type planktonic foraminifera. This indicates the presence of oxygen-depleted bottom waters, which is probably reflective of an Oxygen Minimum Zone (OMZ), impinging on the outer shelf during the late Aptian and entire Albian (Text-Fig. 6).

2.3 Pelagic carbonates and gravity flows

In late Albian and early Cenomanian, the current-dominated regime was gradually replaced by a pelagic regime, during which the heterogeneous sediments of the Garschella Formation were blanketed by monotonous calcareous oozes (Upper Albian to Campanian **Seewen Formation**; Text-Figs. 2, 3). The onset of pelagic deposition was diachronous and becomes younger in a proximal inner-shelf direction (outer shelf: Appenninica Zone [latest Albian]; distal inner shelf: Appenninica Zone; intermediate inner shelf: Brotzeni Zone [early Cenomanian]; proximal inner shelf: Reicheli Zone [Middle Cenomanian]). Probably due to its steepness and to contouring currents, the inner-shelf margin remained sediment-barren during Cenomanian. The first Seewen sediments from the rampzone date from the Archaeocretacea Zone.

Accumulation of the western Austrian pelagic Seewen sediments was dominated by **remobilization processes** such as slumping and gravity flows, which followed primary deposition, especially from the Turonian onwards. Two types of redeposition are distinguished: a) (pebbly) mudflows, characterized by the inclusion of wholesale allochthonic, integer foraminifera assemblages, and b) turbidites, characterized by mixed, disintegrated foraminifera assemblages. Two episodes of remobilization are especially important because they included eroded sediments of the Garschella and older formations.

1. Near the **Cenomanian-Turonian boundary**, an approx. 2 m thick and uniform bed of coarse-grained glauconitic sands was deposited, punctuating the pelagic Seewen sediments (Archaeocretacea Zone **Götzis Bed**, Text-Figs. 2, 3). This bed is preserved in the entire inner shelf and proximal outer shelf areas in the western Austrian and easternmost Swiss Helvetic Alps. It is interpreted as a high-density turbidity flow, which originated in innermost shelf areas (probably from a line source of reexposed Garschella Formation).

2. In **latest Turonian and early Coniacian**, the central part of the inner shelf and the inner shelf margin experienced a large-scale failure event, at which multiple debris flows and "mega" turbidites (BOUMA 1987) developed, incorporating eroded sediments from the Seewen, Garschella, Schrattenkalk, and Drusberg Formations (up to 30 m thick Upper Turonian and Coniacian **Götzis Beds**; Text-Figs. 2, 3). During this event, large portions of the Garschella Formation became remobilized; in large areas along the inner-shelf margin, the Garschella Formation disappeared.

Deposition of Seewen sediments ceased diachronously as well. The youngest representatives on the inner and proximal outer shelf date from the Upper Santonian, whereas their distal outer-shelf counterparts date as young as Campanian (WEIDICH 1987).

Sediments overlying the Seewen Formation consist of silty and fine-grained sandy muds and marls (Upper Santonian to Upper Campanian **Amden Formation**; Text-Figs. 2, 3; OBERHÄNSLI 1978). The presence of a sharp, erosive and often unconformable boundary toward the underlying Seewen sediments and the frequent inclusion of meter to decameter-sized slides of older sediments in basal Amden sediments suggest the presence of large-scale, olistostrome-type gravity-flow deposits in the basal Amden Formation.

3. Events and mechanisms

3.1 Drowning of the carbonate platform

Drowning patterns of the Barremian to Lower Aptian carbonate platform are reflected in time by the maximum age of superjacent condensed phosphatic beds, as well as by the minimum age of orbitolinids, included in the uppermost platform beds. They suggest that platform demise started near the Barremian-Aptian boundary for the easternmost Helvetic shelf (southeastern F. R. Germany; GEBHARD 1985), prograded to the eastern Helvetic shelf (western Austria) in middle Lower Aptian, and affected the entire distal and intermediate Helvetic platform area in early Upper Aptian. Platform drowning trended from east to west and from distal to proximal platform areas (Text-Fig. 5).

The Aptian-Albian condensed phosphatic beds were generated along the axis of a bottom-hugging Tethyan current system (section 3.2.1); their position on the platform-based inner shelf therefore reflects the position of the current axis (Text-Figs. 5, 6). This relates the landward shift of the Tethyan current system directly to the platform demise during Lower and early Upper Aptian, depicted by the presence and position of the **Luitere** condensed phosphatic bed (Text-Figs. 2, 5). The current system most likely imported eutrophic waters onto the platform, inhibiting and terminating the carbonate production of the delicate, oligotrophic platform communities (section 3.2.1).

The Lower to early Upper Aptian landward current shift was probably related to a slow relative sea-level change, indicated by the facies differences of the inner-shelf sediments during this period.

In middle Upper Aptian, a siliciclastic sand-replenishment phase imported larger amounts of calcareous bioclasts and lithoclasts as well (**Brisi Limestone**; typically cross-stratified). The calcareous components include eroded and reworked, micritized Schrattenkalk components, suggesting reexposure of Schrattenkalk platform sediments in proximal inner-shelf areas, but also abundant bioclasts of crinoids, bryozoa, brachiopoda, and bivalves, suggesting the regeneration of an impoverished platform biocommunity in innermost shelf areas. Hermatypic organisms are absent, which may be due to deterioring climatic conditions (e. g., KEMPER 1987). This final phase of platform production came to an end during a terminal Aptian event, during which a **rapid** relative sea-level rise occurred, and the Tethyan current system shifted from the outer shelf to the proximal inner shelf (FÖLLMI 1986, 1989b).

Platform drowning was, therefore, a two-step process; a first and most important phase occurred during early and early late Aptian; a second and final phase occurred in latest Aptian and served only as "coupe de grâce" to the already limited platform production in proximal areas.

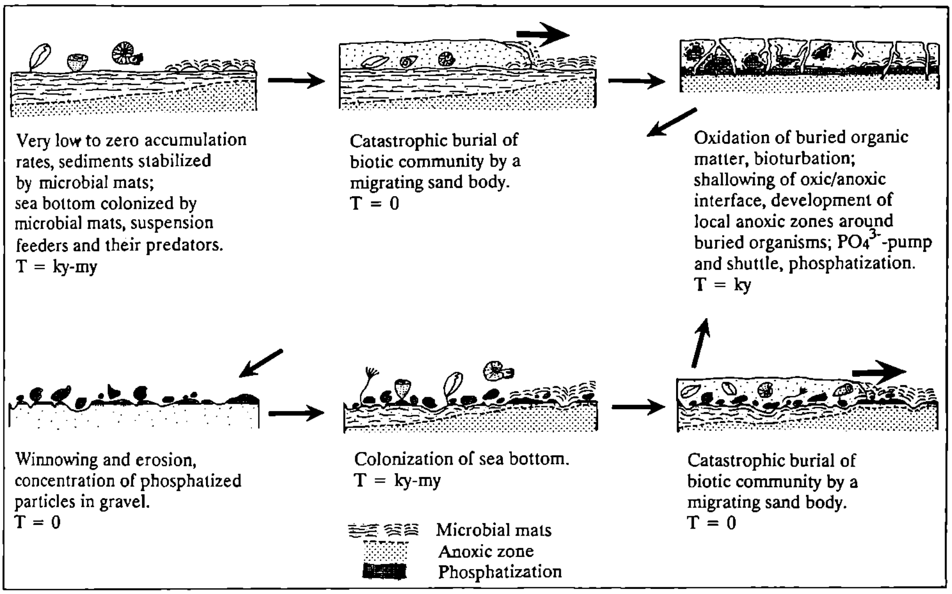
The platform demise appeared to be induced by a combination of following factors: a) landward shift of a probably nutrient-rich Tethyan current-system onto the platform, b) relative rise of sea levels, inducing the landward current shift, and c) climate deterioration during Upper Aptian. An important influence was probably also exerted by the high atmospheric CO_2 content and the rising CCD during Aptian times (e. g., ARTHUR et al. 1985).

3.2 Current-induced condensation and phosphogenesis

3.2.1 The Tethyan current system

Indications for the presence and influence of the Tethyan current system, referred to above, are preserved in the condensed phosphatic beds and in their position in regard to the overall Aptian and Albian sediment configuration.

1. The condensed phosphatic beds consist, to a larger extend, of winnowed beds, as is indicated by the superposition of multiple phosphate generations within the phosphatic crusts or diaclasts, and the intimate mixtures of phosphatized fossils of different time zones and ecological habitats. The "ultra"-low sediment accumulation rates of 2-20 cm/my over several million years are interpreted as the result of Baturin cycling along the axis of a



Text-Fig. 7. Hypothetical diagram of phosphogenesis and subsequent mechanical concentration of phosphatic particles. Current-induced Baturin cycling is indicated by thick arrows (from FÖLLMI 1989b).

bottom-hugging current system (by variations in current velocity and position; Text-Figs. 6, 7).

2. The phosphatic beds show bundling into one bed and a maximum in rates of condensation in distal inner-shelf areas (Text-Fig. 6). In proximal inner-shelf areas, the phosphatic beds include less time, and are separated by thicker, less condensed glauconitic sandstone beds, indicating generally higher accumulation rates. The distal inner-shelf area of intense and persistent condensation is situated in a zone parallel to the inner-shelf margin, and is interpreted as the "fingerprint" of the Tethyan current system, being located within the erosive current axis (Text-Fig. 6).

The bottom-hugging current system flowed in a westward direction, as indicated by the shape of the Albian siliciclastic sandbodies and by the discovery of a dropstone: an angular gneiss boulder of Precambrian age of closure which may be correlated in age and composition to occurrences of the eastern Moldanubicum (FÖLLMI 1986, 1989b). General geophysical considerations (Coriolis effect) support this flow direction.

Shifting positions of the condensed phosphatic beds on the inner shelf indicate that the current system shifted laterally through time as well.

Aptian-Albian inner shelf siliciclastic sediments are generally burrowed and bioturbated, whereas the outer-shelf sediments display plane lamination and lack preserved benthic organisms (section 2.2.3). Apparently, the outer-shelf environment was influenced by the presence of an impinging OMZ, whereas the proximal inner shelf was flooded with better oxygenated bottom waters. The Tethyan current system may, therefore, have limited the upward extension of the OMZ (similar to the modern northbound undercurrent along the upper OMZ boundary offshore central California; VERCOUTERE et al. 1987). The close relationship between the upper OMZ boundary zone and the Tethyan current system and the possibility to strip off nutrient-rich waters may, therefore, explain, why the current had such a negative impact on the oligotrophic carbonate producers during Aptian. The presence of nutrient-rich current waters is also indicated by the presence of abundant microbial mats and suspension-feeding organisms, preserved in the condensed phosphatic beds.

3.2.2 Phosphogenesis, glauconite genesis

Aptian and Albian phosphogenesis along the Helvetic shelf was almost entirely limited to the zone of current-induced condensation. Subordinate phosphogenesis occurred within the more proximal siliciclastic sands, and in the outer shelf laminated marls and muds. Yet, due to the absence of Baturin cycling, subsequent concentration of these phosphates could not take place.

One intriguing problem, posed by the mid-Cretaceous Helvetic strongly condensed phosphatic beds, is the reconciliation of the presence of phosphates and of well-preserved phosphatized fossils on one hand, and the very low sediment-accumulation rates on the other hand. Given the virtual absence of any sediment input, and the abundance of microbial mats and suspension feeders, one might expect to find a self-contained, balanced biochemic system on the seafloor along the current axis; a system in which most organic material is recycled, and thus the change of fossilization and phosphatization is minimal.

In this context, the notion of a strong correlation in time between the presence of siliciclastic sands and phosphatization is valuable (section 2.2.1). The presence of redeposited siliciclastic sands along the inner-shelf margin, commonly mixed with phosphatic particles (throughout Aptian and Albian times; Mittagspitz Formation, Rankweil Beds; Text-Figs. 2, 3, 6) points to a bypassing transport of siliciclasts via the zone of condensation and phosphatization. It is assumed that bypassing occurred in the form of episodically moving sandbodies and of bedload transport, and in the case of sand depletion, only by bedload transport (from undepleted, more proximal shelf areas). It is assumed as well that rates of bypassing maximized episodically during both lateral shifts in current position and episodes of high current velocities (water "jets"). These two assumptions are based on observations in modern current-dominated shelves, in which bypassing of siliciclastics occurs: e. g., eastern coast of South Africa and the Agulhas Current (FLEMMING 1988); Celtic and North Sea and tidal currents (HAMILTON et al. 1980, SMITH 1988, STRIDE 1988).

Regarding the above observations and assumptions, phosphatization may have taken place along the following route: a) catastrophic burial of faunal communities by bypassing palimpsest sands, remobilized by an increase in current velocity, or by a lateral shift in the current axis, b) reworking of the sandsheet by burrowing and scavenging organisms, c) creation of local anoxic zones around buried organisms and general shallowing of the redoxcline, d) concentration of dissolved PO_4^{3-} along oxic-anoxic interfaces, probably via physicochemical cycles of iron and manganese (O'BRIEN & HEGGIE 1988, FROELICH et al. 1988), e) precipitation of phosphates around and within the buried organisms, f) winnowing and erosion of the palimpsest sandsheet, until a gravel of phosphatic particles and/or crusts remains, followed by repetition of a - f (Text-Fig. 7).

This rather simple scenario may explain, why a) mainly benthic and benthos-related nektic organisms (ammonoids) are preserved, whereas remains of active nekton are rarely present within the condensed phosphatic beds, b) the preservation of phosphatized fossils (e. g., microbial mats) is excellent ("Pompeii scenario"; abraded surfaces on the phosphatized fossils are commonly due to subsequent reworking), and c) how Baturin cycles worked.

Glauconitic particles are particularly abundant in the siliciclastic sandsheets and bodies, which were formed during the four phases of sand replenishment (section 2.2.1). On the other hand, authigenic glauconites are rare in the zone of phosphogenesis and maximum condensation, and beyond it, in distal shelf areas. Glauconitization was favoured on the proximal inner shelf, within the siliciclastic sands; i. e., in areas, where the detrital influx was moderate to high, winnowing moderate, and bottom waters better oxygenated. Phosphogenesis was optimal in distal inner-shelf areas, where accumulation rates approached zero, winnowing was intense, detrital influence minimized, and bottom waters oxygen poorer (Text-Fig. 6).

3.3 Tectonic events

In the evolution of the mid- and early Upper Cretaceous Helvetic shelf, four short phases of reinforced erosion and redistribution of sediments punctuated the general buildup of the sequence, platform carbonates - phosphatic sediments - pelagic carbonates. All four events are correlatable to regional or global phases of reinforced sediment reworking, and/or tectonism.

1. During the **Aptian-Albian boundary interval** (Jacobi Zone and early Tardefurcata Zone), significant changes occurred in the Helvetic sediment configuration and shelf topography: a) a rapid relative sea-level rise is indicated in the presence of an uppermost Aptian condensed phosphatic and micritic bed, resting on top of middle Upper Aptian shallow-water Brisi sands and calcareous bioclasts in proximal inner shelf areas (**Twäriberg Bed**; present in eastern Swiss occurrences; in western Austria only in reworked, transported sediments of the **Klaus Beds**; FÖLLMI & OUWEHAND 1987), b) due to this relative sea-level rise, the Tethyan current shifted from an outer-shelf position to a proximal inner shelf position (Text-Fig. 5), c) in western Austria, the proximal inner shelf-based current eroded the entire middle Upper Aptian Brisi Limestone, most of the proximal Brisi Beds, as well as the uppermost Schrattenkalk sediments; in eastern Switzerland, erosion was less intense and a condensed phosphatic bed was formed (OUWEHAND 1987), d) eroded sediments were redeposited as debrites in distal inner-shelf areas, and in newly formed channel- and fan systems along the inner-shelf margin (Klaus, Rankweil Beds; Text-Fig. 2), e) the inner-shelf margin experienced a phase of rapid steepening at the beginning of this episode (section 2.2.2), f) minor faulting occurred in distal inner-shelf areas (though dating is insecure) and is probably related to margin transformation.

This episode correlates to a global 115-110 my tectonic event of many authors (SCHWAN 1980), and is probably even more directly related to a phase of collision between the Adriatic promontory and the European Plate (DERCOURT et al. 1986, ZIEGLER 1987, 1988, LE PICHON et al. 1988).

2. During the **Cenomanian-Turonian boundary interval**, a single widespread sandy and erosive turbidite bed accumulated. This phase does not correlate to a tectonic "event", but it correlates with a period of widespread anoxia, preceded by major erosion in many localities (DE GRACIANSKY 1986). ARTHUR et al. (1988) considered a latest Cenomanian tectono-eustatic "pulse" as possible trigger, influencing circulation patterns (anoxic sediments from this interval are not known from the Helvetic shelf).

3. During the **late Turonian-Coniacian interval**, a large-scale failure event occurred along the inner-shelf margin (section 2.3). This event was accompanied by strong erosion and redeposition of eroded sediments, as well as small-scale faulting. Similar failure events are known from other localities within the Helvetic shelf (FÖLLMI 1989b). They probably correlate to a widespread orogenic phase along the eo-Alpine collision front (ZIEGLER 1987, FRANK 1987).

4. Inner-shelf accumulation of pelagic sediments was terminated by a **late Santonian or early Campanian** episode of deposition of large-scale gravity flows, of faulting, and erosion. This "event" probably correlates to a further important orogenic event within the "eo-Alpine" realm (SAVOSTIN et al. 1986, FLÜGEL et al. 1987).

The Aptian-Albian boundary episode portrays the onset of rapid deepening along the Helvetic shelf, probably because of a collision-induced transpressive regime within the Tethyan Realm (ZIEGLER 1987). The event terminated the formation of the Barremian-Lower Aptian shallow-water platform, the final member within a cyclic suite of Lower Cretaceous shallowing-upward carbonate platforms, which commonly are topped by thin and condensed deepening-upward sequences (FÖLLMI 1989b).

The Turonian-Coniacian boundary event terminated the deepening-upward process in the Helvetic Realm. A post-Turonian shallowing-upward trend is indicated in the biofacies of the Seewen sediments (increasing benthic/planktonic foraminifera ratios; progradation of calcisphaerulids into the outer shelf), and persists throughout the remainder of the Cretaceous (TRÜMPY 1982).

4. Conclusions

The evolution of the mid-Cretaceous sequence, carbonate platform - terrigeneous and phosphatic sediments - pelagic carbonates, along the eastern Helvetic shelf appears to have been determined by:

1. the presence of a westbound Tethyan current system, which defined the boundary between the surficial mixed-water layer and an underlying body of oxygen-depleted waters impinging on the shelf. Residence of the current system on the inner, platform-based shelf owed to elevated sea levels during most of the Aptian to early Cenomanian interval, and caused a) a demise of the carbonate platform, probably because of the import of nutrient-rich bottom waters onto the platform, b) condensation and erosion along the current axis, c) bypassing of siliciclasts into outer-shelf environments beyond the platform, and therefore d) phosphogenesis, probably triggered by the catastrophic burial of benthic faunal communities with bypassing palimpsest sands;
2. a first-order sea-level rise, which caused the current shift onto the Helvetic shelf, and, being persistent, the subsequent installation of a pelagic sedimentary regime;
3. coupling of 1 and 2, which limited the low influx of coarse terrigeneous material to four discrete phases of sand replenishment;
4. tectonic events, which introduced a general transpressional regime into the Tethyan Realm, and probably reinforced subsidence patterns of the Helvetic shelf from latest Aptian onwards. Stronger compressional forces probably inverted this trend in late Turonian, and induced a persistent shallowing-upward trend for the remainder of the Cretaceous. Tectonic events were probably also responsible for four phases of reinforced erosion and sediment redistribution along the eastern Helvetic shelf (during the Aptian-Albian, Cenomanian-Turonian, Turonian-Coniacian, and Santonian-Campanian boundary intervals).

The evolution of the mid-Cretaceous Helvetic sequence mirrors global patterns of platform drowning, deposition of terrigeneous and phosphatic beds, and subsequent "pelagization". This is probably due to the fact that the mid-Cretaceous embodied a period of increased tectonic activity, with a global impact on sea levels and atmospheric CO_2 content, and, therefore, on general climatic, paleoceanographic, bio- and geochemical conditions.

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The Organic Matter of "Mid-Cretaceous" Deposits of the Median Prealps

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With 10 Text-Figures and 2 Tables

DUPASQUIER, CH., LIGOUIS, B. & CARON, M. (1989): The Organic Matter of "Mid-Cretaceous" Deposits of the Median Prealps. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 607-635. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The "Mid-Cretaceous" of the Median Prealps Nappe (Subbriançonnais, northern margin of the Tethys) consists principally of black or dark shaly marls and marls which alternate with limestones or siliceous beds containing radiolaria. The fauna is pelagic, and the planktonic foraminifera indicate Aptian to mid-Turonian ages.

The organic content has been studied by Rock-eval pyrolysis (Rep) and organic petrography to define source, depositional environment and maturation level. Two types of organic material have been recognized: 1) terrestrial organic matter deposited in low oxic to oxic environments; 2) a mixture of marine and terrestrial organic matter deposited in low oxic to anoxic environments.

The sediments are mature (oil window); the vitrinite reflectance associated with other field data indicate a transported metamorphism.

Kurzfassung: Die "Mittlere Kreide" der alpinen Medianes-Decken (Subbriançonnais, nördlicher Tethys-Rand) besteht hauptsächlich aus Wechsellagerungen von schwarzen bis dunklen Mergeln und Kalken oder kieseligen Lagen mit Radiolarien. Die Fauna ist pelagisch. Die gefundenen Biozonen reichen vom Apt bis ins Mittlere Turon.

Der organische Inhalt wurde mit der Rock-Eval-Pyrolyse (Rep) und der organischen Gesteinspetrographie bestimmt, um den Ablagerungsraum und die organische Maturität zu definieren. Zwei Typen von organischer Materie sind erkannt worden: 1. terrestrisches organisches Material, das in schwach oxischem bis oxischem Milieu zur Ablagerung kam; 2. ein Gemisch aus marinem und terrestrischem Material, aus einem schwach oxischen bis anoxischen Milieu stammend.

Die Sedimente sind reif (Öl-Fenster); die Vitrinitreflexion, assoziiert mit weiteren Feldbeobachtungen, indiziert eine transportierte Metamorphose.

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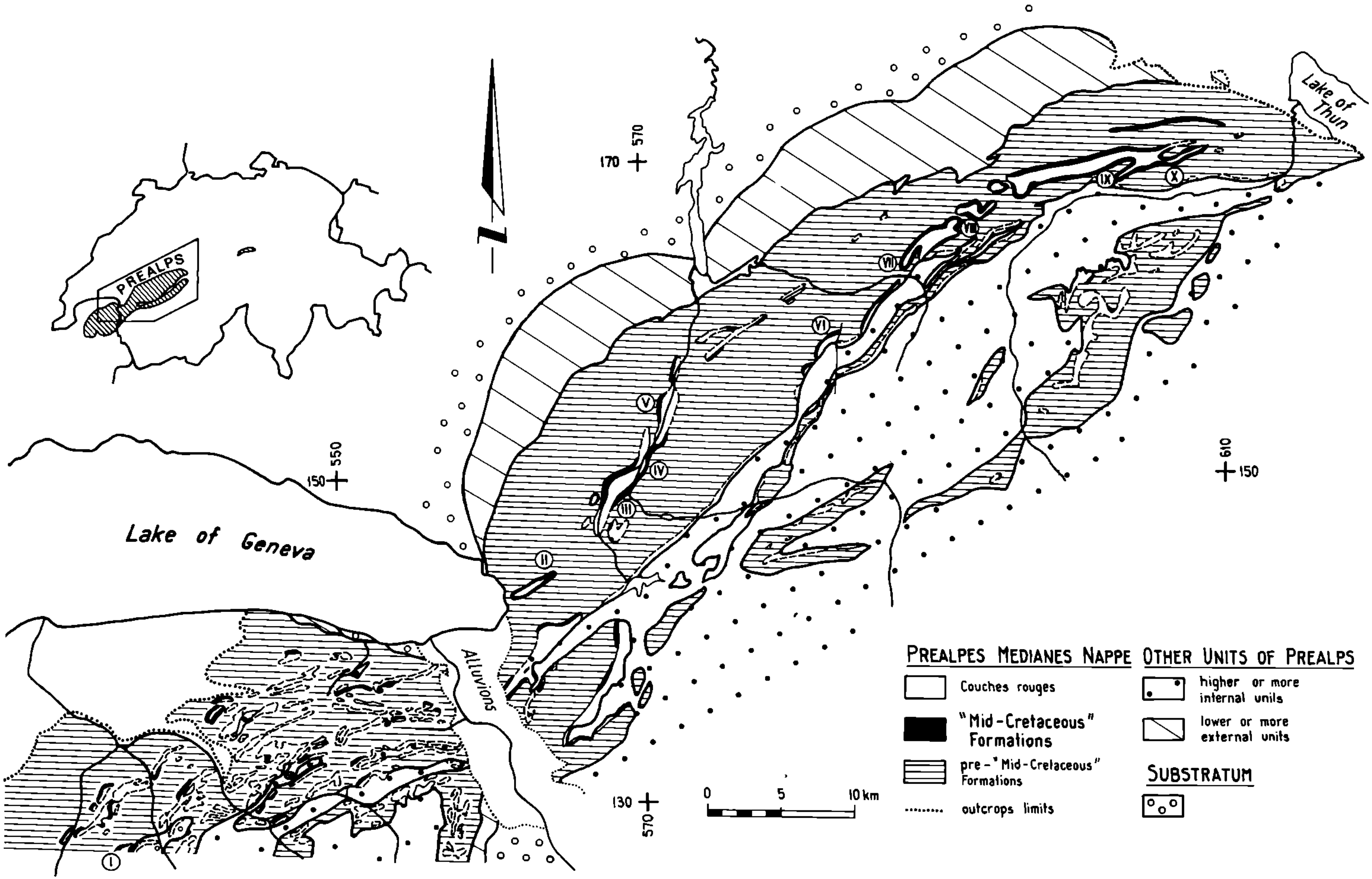
Lake of Thun
Lake of Geneva
Alluvions
170
570
550
150
610
150
130
570
I
II
III
IV
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VI
VII
VIII
IX
X
0
5
10 km
PREALPS
PREALPES MEDIANES NAPPE
Couches rouges
"Mid-Cretaceous" Formations
pre-"Mid-Cretaceous" Formations
outcrops limits
OTHER UNITS OF PREALPS
higher or more internal units
lower or more external units
SUBSTRATUM

1. Introduction

The Median Prealps nappe is a decollement nappe, originally belonging to the Penninic domain (Briançonnais s. l.) and situated on the northern margin of the Tethys. It lies south of Lake Geneva (Chablais Prealps) and between there and Lake Thun (Romandes Prealps) (Text-Fig. 1). Facies analogies with corresponding formations in the French Alps equate the northwestern part (Médianes plastiques) with the Subbriançonnais belt having complete stratigraphic successions from Upper Trias to Tertiary, and the southeast area (Médianes rigides with reduced series) to the external part of the Briançonnais proper (TRUEMPY 1980).

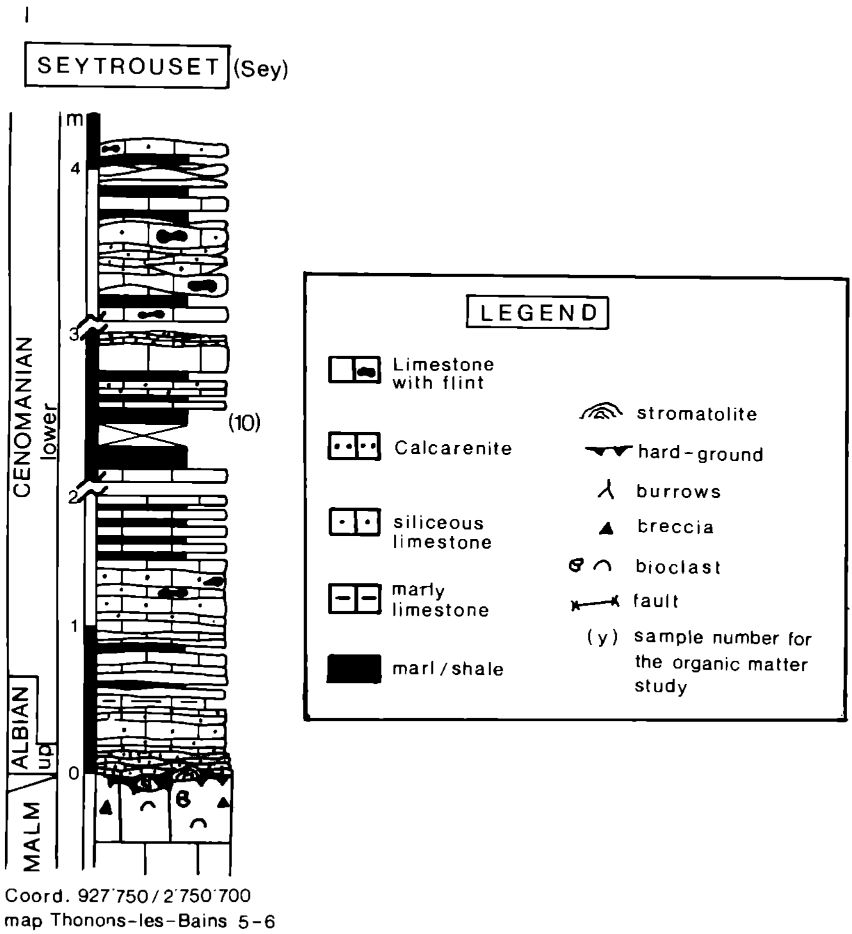
The "mid-Cretaceous" deposits (Aptian to mid-Turonian) of the Médianes plastiques (N-NE) are called the Complexe Schisteux Intermédiaire (CSI, SCHWARZ-CHENEVART 1945). They are essentially composed of about 70 m of pelagic black, grey, sometimes red or green limestones and marls (mudstones to packstones with planktonic foraminifera and/or radiolaria) or calcarenites (packstones to grainstones with in situ or reworked material). Towards the S-SW (Médianes rigides), the sequence is reduced to a stromatolitic hard-ground covered by few centimeters to meters of sometimes phosphatic and glauconitic calcarenites and siliceous limestones, or as fillings of neptunian dykes (mid-late Albian to mid-Turonian); these lie on the more or less deeply eroded surface of the Malm (Calcaires massifs Formation, HEINZ 1985).

Stratigraphic and sedimentological data (BAUD & SEPTFONTAINE 1980, BOLLER 1963, ISENSCHMID 1983, HEINZ 1985) from the Médianes Romandes indicate that the NE area (Médianes plastiques) was a basin situated probably between the ACD and the CCD during this time interval; the internal area (Médianes rigides) appears to have been a submarine high with a complex paleogeography. Within this context, ten localities were studied for their organic matter (O.M.), sampled in dark marly levels (external domain: sections II to IX; internal domain: sections I, X, Text-Figs. 2-5). The sections show three thicker black shaly levels of about 50 cm to 1.5 m thickness dated from the Barremian-Aptian boundary, the late Albian, and late Cenomanian-early Turonian; this facies is less pronounced at mid-Turonian age. Ages correspond with those of Black Shales studied in Atlantic (ARTHUR & PREMOLI SILVA 1982) and with the oceanic anoxic events described in the Tethys (JENKYNS 1980, 1985) (Barremian-early Aptian, latest Aptian-middle Albian, Cenomanian-Turonian boundary).

2. Lithological nature and stratigraphy

The black shales are present over the NW area (corresponding to the external part of the penninic domain) with sections of early Aptian to mid-Turonian age. They are less represented in the internal part which shows condensed sections of late Albian - mid-Turonian age.

Text-Fig. 1. Distribution of the "Mid-Cretaceous" Formations in the Préalpes Médianes Nappes.



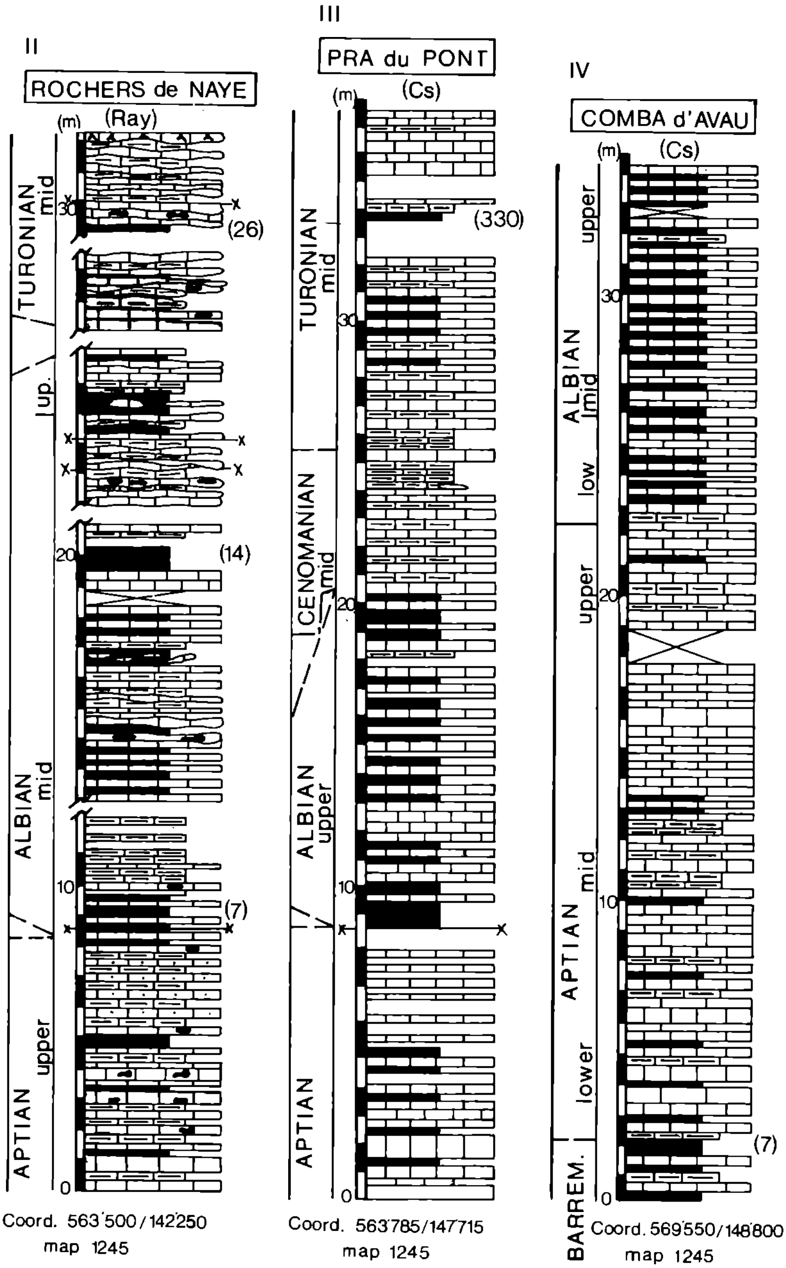
Text-Fig. 2. Profile I.

2.1 External domain

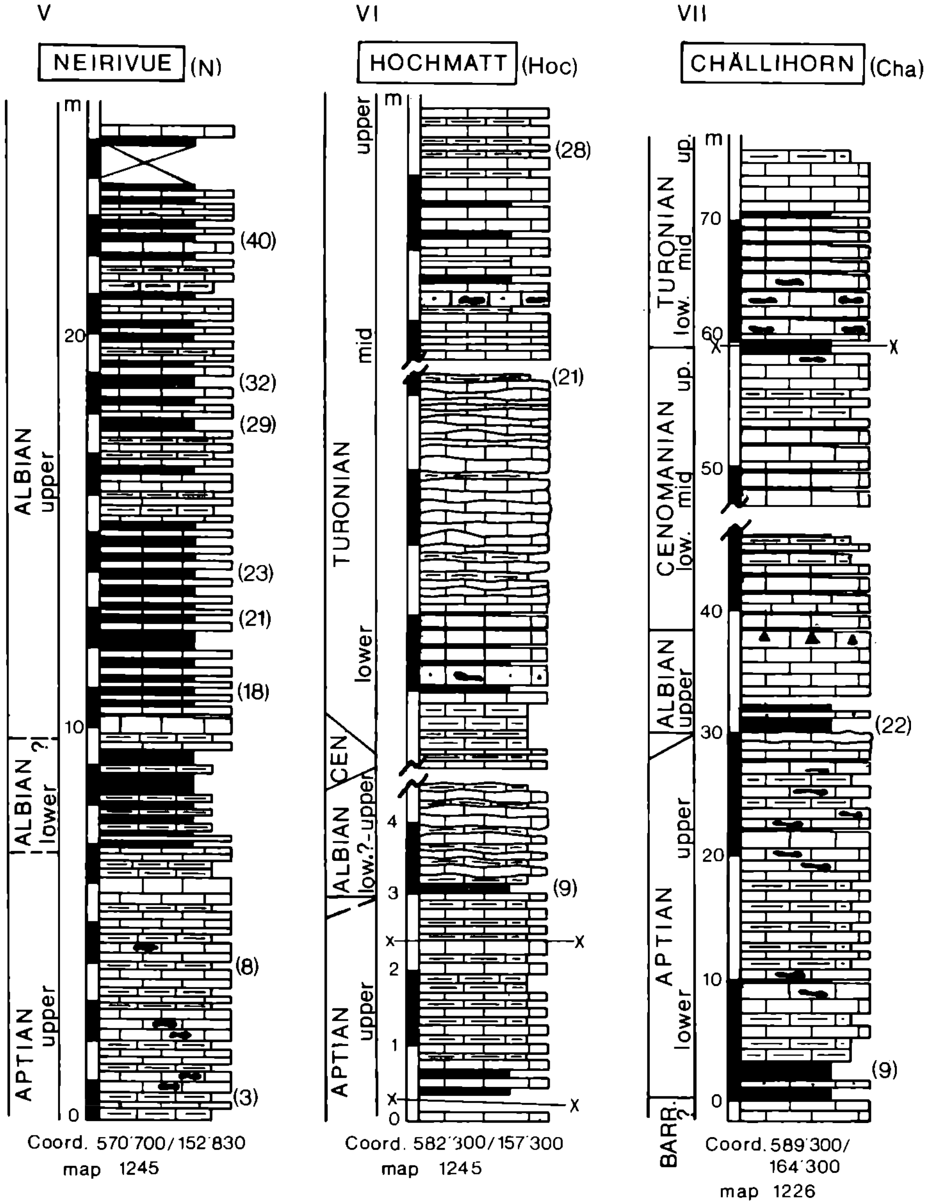
a) Limestone-marl-chert facies of the front of the nappe (Plastiques externes, BAUD & SEPTFONTAINE 1980)
(Sections II, III, IV, V, VI, VII, IX, Text-Figs. 3, 4 and 5)

These are 15 to 70 m of Aptian to mid-Turonian deposits, forming the "mid-Cretaceous" Complexe Schisteux Intermédiaire (CSI) (SCHWARZ-CHENEVART 1945) and being in stratigraphic continuity with the preceding formation (Calcaires plaquetés, SPICHER 1966). The section Chällihorn (VII) is the more complete and representative one.

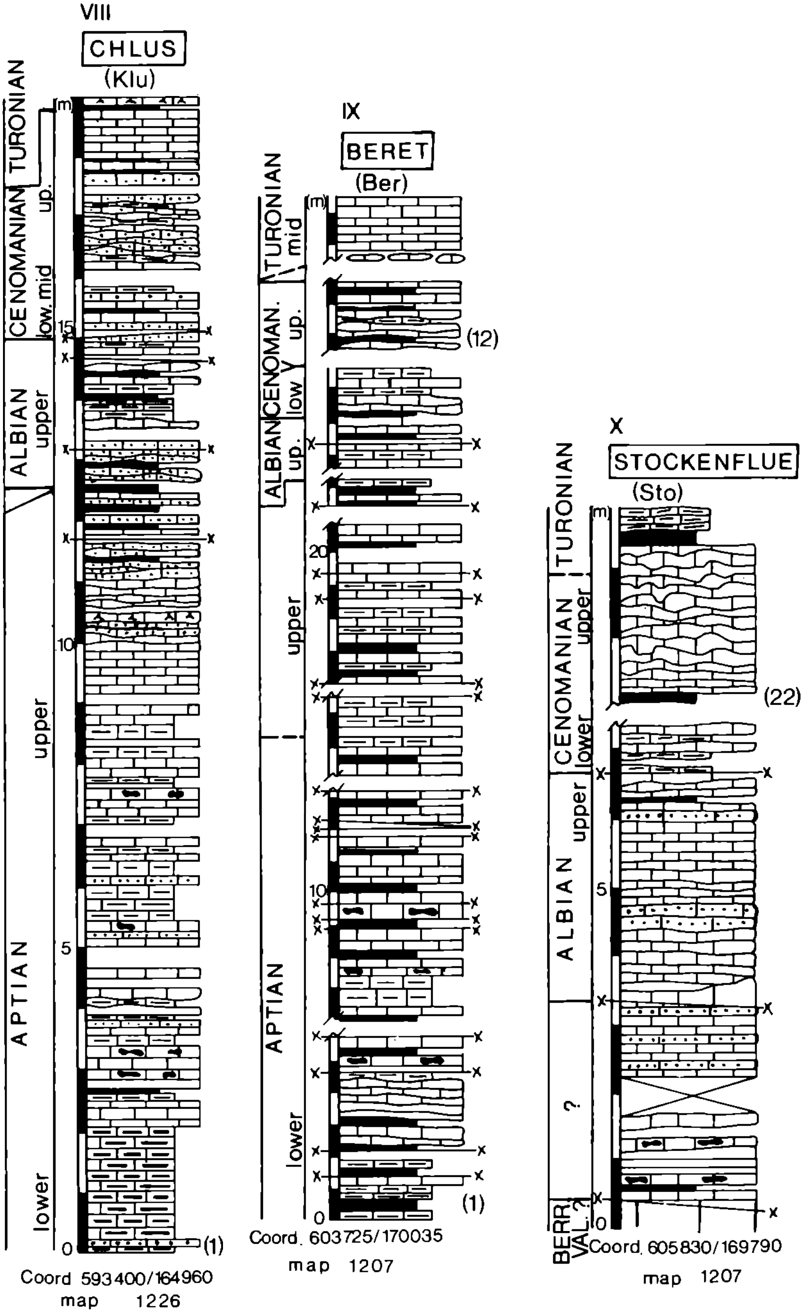
The base of the CSI (lower part of early Aptian?) is made up of black and bioturbated shales (50 cm thick) with pyrite, interbedded with thin



Text-Fig. 3. Profiles II-III-IV.



Text-Fig. 4. Profiles V-VI-VII.



Text-Fig. 5. Profiles VIII-IX-X.

parallel laminated radiolaria-rich black shales. The levels above are cherty limestones alternated with marls (Cabri Zone, early Aptian). Black marls either thicken up-section or disappear laterally to be replaced by red or green siliceous limestones (Algeriana Zone, late Aptian).

Then, green wavy-bedded limestones 1 to 3 cm thick (packstones with abundant large *Ticinella*) precede a second level of black shales (Breggiensis Zone to the lower part of Appenninica Zone, late Albian). Green and/or red bioturbated limestones, marly limestones or marls which follow are wackestones, with abundant inocerams and planktonic foraminifera (early Cenomanian). A recurrence of black limestones and bioturbated shales precedes thin beds of white limestones and siliceous beds (late Cenomanian - Cushmani Zone). The latter are frequently absent in our sections (sections II, III, VI) due to a regional hiatus.

The third level of black shales and siliceous laminated limestones is dated as late Cenomanian-early Turonian (Archaeocretacea Zone). Thin beds of light grey limestones, occasional siliceous limestones and subordinate grey-green shales precede thicker beds (20 cm) with abundant cherts. The top of these levels is composed of black limestones and laminated shales (mid-Turonian - Helvetica Zone).

These levels are overlain in stratigraphic continuity by (or separated from a hiatus with burrows) the red marly limestones of the Couches Rouges Group (late Turonian - Sigali Zone, GUILLAUME 1986).

b) Calcarenite facies

The section (VIII, 17 m) presents levels of black or grey wavy-bedded calcarenites (packstones with crinoids, inocerams, planktonic foraminifera and lithoclasts) interbedded with green or black marls and white limestones (mudstones to packstones with planktonic foraminifera). Two levels of black marls are present (early Aptian, early mid-Albian?).

Interpretation

These facies have been deposited over the Lower Cretaceous in the subsiding external domain, corresponding to a relatively deep basin, between the ACD limit and the lysocline of the calcite. Although planktonic foraminifera (calcitic tests) are well preserved, there are no ammonites and their aragonitic tests have probably been dissolved.

The limestone-marl-chert facies (Text-Figs. 3, 4, 5) are the autochthonous deposits, mixed with fine terrigenous clastics and they mark the deepest setting of the basin.

The calcarenites (Text-Fig. 5, section VIII) are resediments with neritic and pelagic components. They characterize an intermediate area between the basin (N-NE, external domain) and the submarine high (S-SW, internal domain).

2.2 Internal domain (condensed sections)

Section I: Seytrouset (Text-Fig. 2)

This sequence lies on the Calcaires massifs Formation (SPICHER 1966; late Tithonian, Zone B). Several centimeters of stromatolites and glauconitic packstones contain abundant *Ticinella* and *Rotalipora* (Appenninica Zone, late Albian). Grey to dark grey micritic limestones with planktonic foraminifera alternate with calcarenites, dark siliceous limestones with radiolaria, or pelites (early Cenomanian - Brotzeni Zone). Beds are wavy and show compensation structures and gutter casts.

Section X: Stockenflue (Text-Fig. 5)

The sequence, less than 10 m thick, lies on an early Cretaceous condensed level of biomicrite and crinoidal calcarenites with cherts (Zone D/Berriasian and Valanginian?).

Thin beds of crinoidal calcarenites alternate with red or violet condensed levels (packstones with phosphatic pebbles). Planktonic foraminifera give a late Albian age (Breggiensis to Appenninica Zone). The top of the section (strongly tectonized) is formed by 3 m of biomicrite (wackestone to packstone with planktonic foraminifera; Brotzeni to Archaeocretacea Zone - early Cenomanian to early Turonian).

Interpretation

The sedimentary structures of section I (Text-Fig. 2) are indicative of tempestites (BRENCHLEY 1985, GUILLOCHEAU 1983). These levels and condensed pelagic deposits of section X (Text-Fig. 5) suggest an environment of pelagic seamounts with outer-shelf and condensed sediments. Sedimentation was probably inhibited by currents, and there may have been a phase of emersion during the early Cretaceous.

The platform subsided slowly and uniformly from place to place up to early Turonian. The sedimentation starts again at mid-Turonian, which corresponds with the Helvetica Zone, being present over the whole domain.

3. Samples and methods

The sediments containing the O.M. are generally of dark colour; twenty-three samples have been selected from outcrops in the dark to black commonly bioturbated and pyrite-rich, shaly marls. The outcrop has been cleaned as much as possible to avoid any contamination or oxydation of the O.M. Samples have been investigated by Rep and organic petrography to determine facies and diagenetic stage of the O.M. Many other samples have been analysed by Rep, but only the results of the 23 studied by the two methods are described.

Petrographic analysis was conducted with a Leitz MPV-II microscope photometer interfaced to an HP-86B microcomputer. A total magnification of 625 x with oil immersion was used. A 100 W halogen lamp was used for white-light observation and reflectance measurements. The reflectance of organic particles was measured under oil immersion for the wavelength of 546 nm (details of the methods see in STACH et al. 1982 and in ICCP handbook 1963, 1971, 1975).

In order to simplify, the term vitrinite in this study is also used for corresponding particles showing less than 0.5 % reflectance although these should be called huminite. Blue-light observations were made using a 100 W ultra-high pressure mercury lamp, a 350-450 nm excitation filter (BG 12) and a 510 nm barrier filter.

Petrographic composition was determined from the observation of the microscopic preparations using reflected white light and reflected fluorescent light. The amounts are given in conditional units (Table 1), that is to say we have summarily quantified the constituents by the estimation of their frequency (for example: abundant = 5, very frequent = 4, common = 3, . . .).

Two types of microscopic preparations have been used for the petrographic study:

1) a polished rock mounted in an epoxy resin. The polished surface is prepared on the block perpendicular to the bedding. This type of preparation allows observation of the organic/inorganic relationship in reflected white light and in reflected fluorescent light;
2) a polished organic concentrate. The organic matter was concentrated by physical procedure. The rock sample is crushed to 200 µm and centrifugated in heavy liquid with a specific gravity of 1.7. The suspended organic matter is filtered on a micropore filter which is then dissolved on a plexiglass plate. The adhesion of the organic particles on the plexiglass plate is increased by a coating of epoxy resin. This type of preparation is used for the reflectance measurements and allows observation of the organic material both in reflected white light and in reflected fluorescent light.

Rock-eval pyrolysis and organic carbon determination were performed according to the methods described by ESPITALIE et al. (1977, 1985, 1986) and ESPITALIE (1986). The parameters obtained and used in the present study are:

- The Total Organic Carbon (TOC) content, calculated from the sum of residual organic carbon and pyrolysed organic carbon; expressed in %.
- The pyrolysis temperature (T_{max}) at which a maximum amount of hydrocarbon compounds coming from the thermal degradation of kerogen is released; expressed in °C.
- The Hydrogen Index (HI), as the S_2/TOC ratio, expressed in mg per g of TOC (S_2 peak: hydrocarbon compounds resulting from cracking of kerogen).
- The Oxygen Index (OI), as the S_3/TOC ratio, expressed in mg of CO_2 per g of TOC (S_3 peak: CO_2 released from cracking of kerogen).

4. Results

4.1 Quantity of organic carbon

Table 1 shows the organic carbon contents of the 23 samples. Values range from 2.69 to 0.05 %. Organic carbon contents for 17 of 23 samples are less than 1 %; for 6 of 23 samples less than 0.5 %. The higher values were found in the Barremian, early Aptian and late Albian.

4.2 Type and composition of organic matter

a) Pyrolysis data

- T_{max} diagram (Text-Fig. 6): 15 samples fall below the kerogen type III evolution pathway indicating a terrestrially derived organic matter with low hydrogen indexes. Five samples lie just above the kerogen type III evolution pathway indicating a mixture of type II and type III kerogen. Three of them (samples: Ber 1, Cha 22, Klu 1) have low hydrogen index due to a higher maturation level confirmed by the vitrinite reflectance. The sample CS 7 has a hydrogen index of 284 HC/g TOC indicating a type II kerogen.
- HI-OI diagram (Text-Fig. 7): this figure demonstrates that samples fall below the kerogen type III evolution pathway, the samples N 21 and Ray 26 being close to the boundary with the type II kerogen (the sample CS 7 is not plotted). The oxygen index values are variable and some are high (up to 70 mg CO_2/g TOC). This seems to be the result of either alteration processes which have affected the O.M., or of its detrital and terrestrial nature, or both.

b) Petrographic data

The petrographic compositions are compiled in Table 1.

- **The petrographic constituents** identified are:

Vitrinite

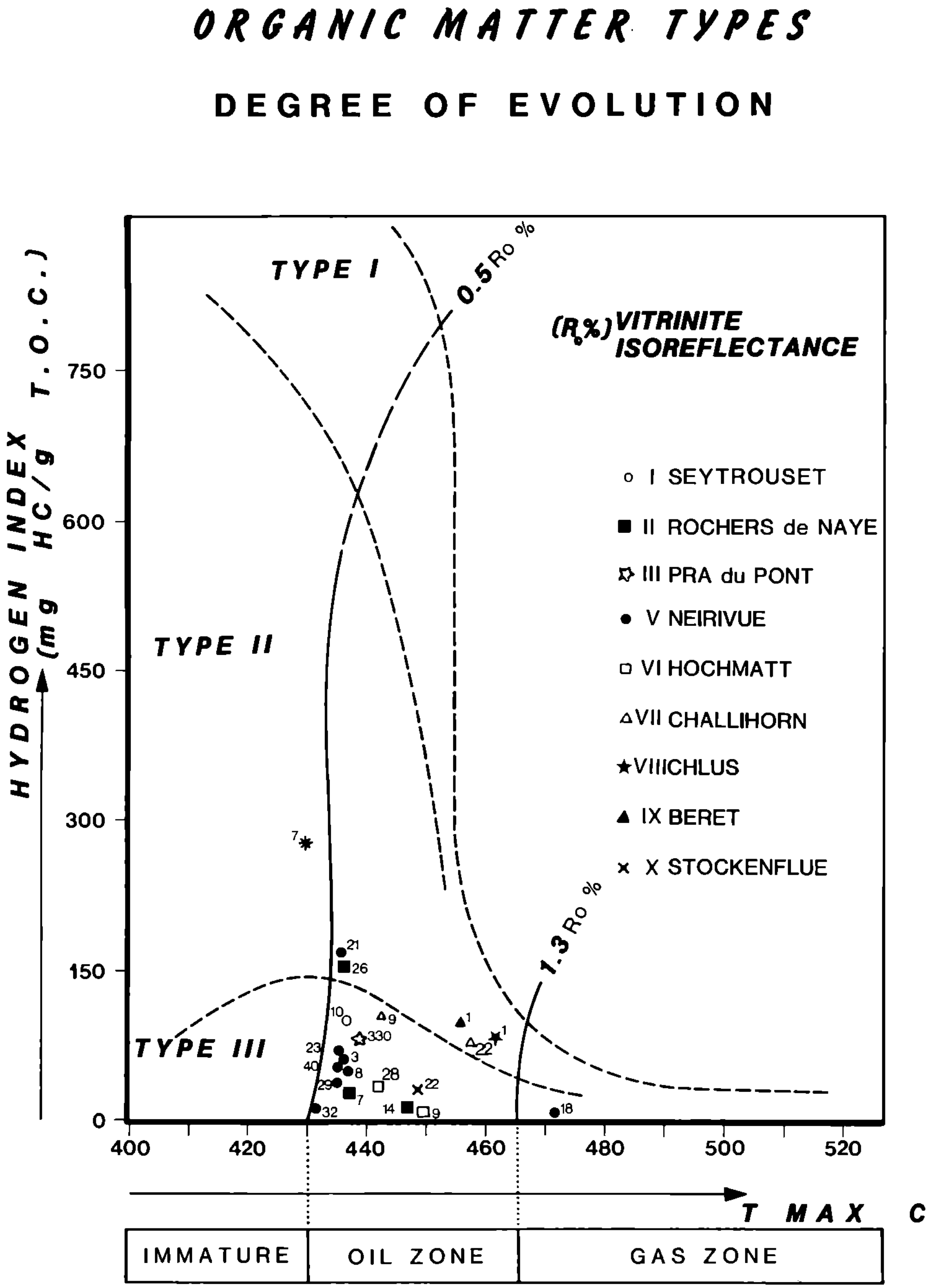
Two types of vitrinite have been observed. Grey particles elongated (in polished rock mounts) and occasionally with telinitic cell structures, correspond to the autochthonous vitrinite (fresh vitrinite) and angular to rounded white particles constitute recycled vitrinite (reworked and/or oxydized). In the samples N 3, N 8, N 29, N 32, Ray 7, Ray 14, Sey 10, reworked particles occur particularly as bimacerite grains (vitrinite associated with inertinite or liptinite), and trimacerite grains (vitrinite associated with liptinite and inertinite). In most of the samples analysed, the recycled vitrinite dominates and the autochthonous vitrinite is present in very small amounts. The reflectance values of the autochthonous vitrinite range from 0.44 to 1.20 % and those of recycled vitrinite are between 0.95 and 2.06 %.

Inertinite

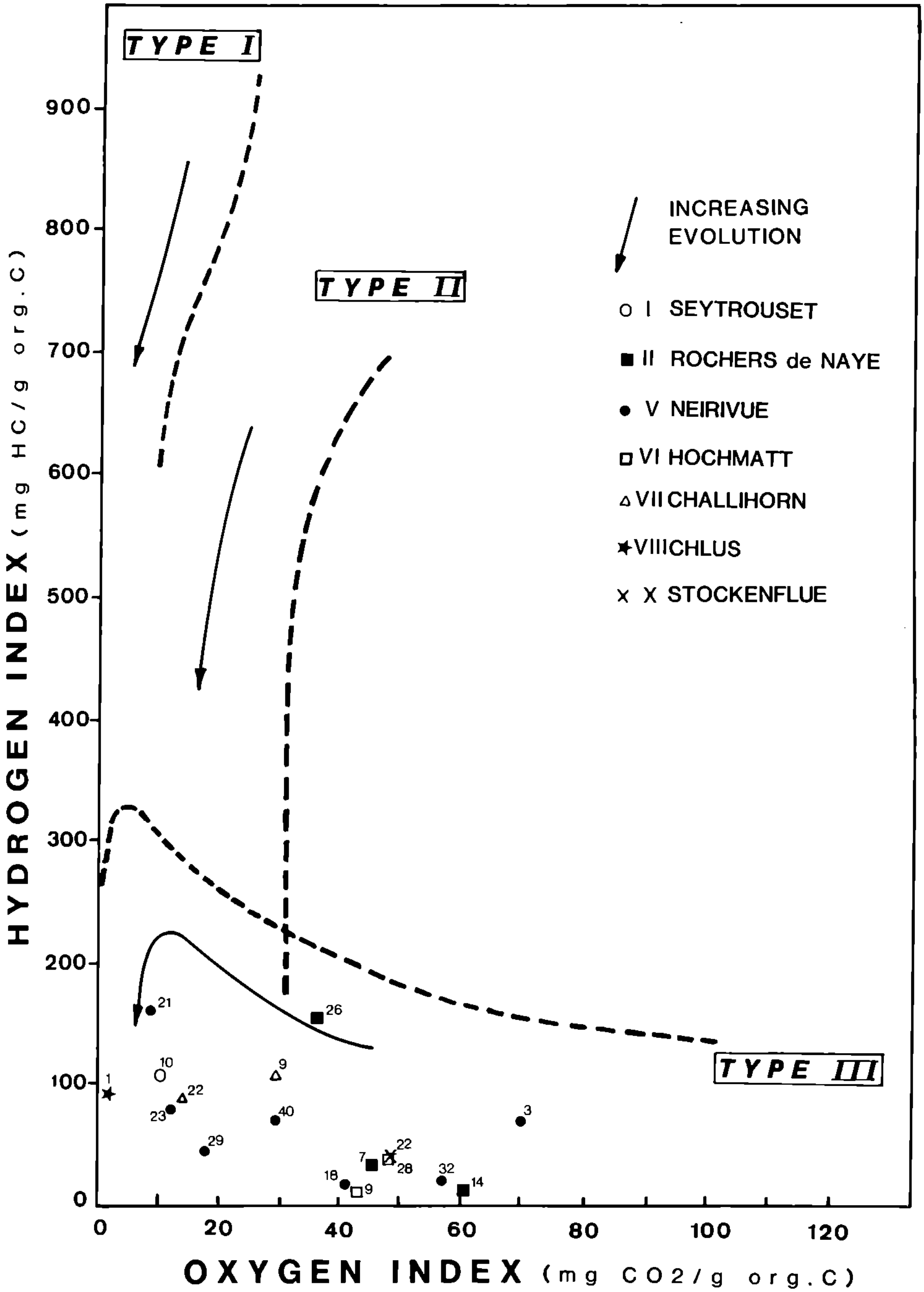
Inertinites are present in most samples. They occur as inertodetrinite and as splinters of fusinite and semifusinite.

Liptinite

The liptinite content of the samples is generally low and consists of a mixture of terrestrial (**sporinite, resinite**) and marine (**alginite, bituminite**) liptinites.



Text-Fig. 6. Results of Rock-eval pyrolysis, displayed in an HI-T_{max} diagram (ESPITALIE et al. 1984) for sediment samples studied.



Text-Fig. 7. Results of Rock-eval pyrolysis, displayed in an HI-OI diagram (ESPITALIE et al. 1985) for sediment samples studied.

ORIGIN (outcrops)	SAMPLE	AGE	TOC (%)	HYDROGEN INDEX (mg HC/g TOC)	OXYGEN INDEX (mg CO_2/g TOC)	VITRINITE		INERTINITE		LIPTINITE					BITUMEN	MICRINITE	ZOOCLASTS	FLUORESCENCE OF MIN.MATRIX		PYRITE	TOTAL OIL POTENTIAL S1 + S2 mg HC/g rock
						Autochthonous	RECYCLED	FUSINITE/ SEMI-FUSINITE	INERTO-DETRINITE	SPORINITE	RESINITE	ALGINITE	LIPTO-DETRINITE	BITUMINITE				COLOUR	INTENSITY		
Préalpes du Chablais																					
- Seytrouset	Sey A10*	Early Cenomanian	0.75	107	16	4	5	2	1	1	1	1	--	2	0	1	--	--	--	--	0.83
Préalpes Romandes																					
- R. de Naye	Ray 26**	Mid-Turonian	0.82	157	39	2	4	2	1	3	0	3	1	3	1	1	4	d.g.	low	3	1.39
	Ray 14*	Mid-Albian	0.36	22	61	2	4	2	1	1	1	0	--	1	0	0	--	--	--	--	0.11
	Ray 7 *	Mid-Albian	0.67	31	46	3	5	1	1	1	1	1	--	2	0	0	--	--	--	--	0.23
- Pra du Pont	CS 330**	Mid-Turonian	0.97	89	--	2	5	2	2	3	1	2	1	2	0	1	4	d.g.	low	2	0.89
- Comba d'Avau	CS 7 **	Barremian	2.02	284	--	3	5	2	1	2	1	3	4	2	0	1	1	d.g.→ l.g.	low to medium	4	5.93
- Neirivue	N 40 **	Late Albian	0.83	65	30	4	5	1	2	2	0	1	1	1	0	1	4	d.g.	low	2	0.54
	N 32 **	Late Albian	0.40	20	57	1	4	2	2	2	1	1	1	1	0	1	4	d.g.	low	3	0.08
	N 29 **	Late Albian	0.80	45	17	2	4	2	2	2	1	2	2	2	0	1	4	d.g.	low	3	0.36
	N 23 **	Late Albian	0.90	75	12	3	5	2	1	3	1	2	1	1	0	1	3	d.g.	low	3	0.69
	N 21 **	Late Albian	1.86	160	8	2	5	2	2	3	1	3	3	1	0	1	3	d.g.→ l.g.	low	4	3.01
	N 18 **	Late Albian	0.63	15	42	5	3	2	1	1	0	1	0	1	1	1	2	d.g.	low	3	0.10
	N 8 *	Late Aptian ?	0.38	57	(105)	3	4	2	1	2	1	1	--	1	0	1	--	--	--	--	0.23
	N 3 **	Late Aptian ?	0.57	73	70	3	5	2	1	1	1	1	1	1	1	1	2	d.g.	low	2	0.42
- La Hochmatt	Hoc 28*	Late Turonian	0.56	39	48	1	5	1	0	1	1	0	--	1	1	1	--	--	--	--	0.22
	Hoc 21*	Early Turonian	0.05	--	--	2	3	0	0	0	0	0	--	1	0	0	--	--	--	--	0.03
	Hoc 9 *	Early Albian	0.49	16	43	2	4	1	0	1	0	0	--	1	0	0	--	--	--	--	0.10
- Challihorn	Cha 22**	Late Albian	1.04	88	14	3	5	2	3	2	0	1	2	1	0	3	2	d.g.	low	3	1.21
	Cha 9 *	Early Aptian	2.69	113	30	4	5	2	1	0	1	0	--	1	2	1	--	--	--	--	3.28
- Chlus	Klu 1 **	Early Aptian	2.18	89	2	3	5	2	2	0	0	0	3	1	0	2	1	d.g.	low	4	2.36
- Beret	Ber 12*	Late Cenomanian	0.16	--	--	1	5	2	1	0	0	0	--	0	0	0	--	--	--	--	0.35
	Ber 1 **	Early Aptian	1.55	103	--	2	5	2	2	0	0	0	3	1	0	2	1	d.g.	low	3	2.01
- Stockenflue	Sto 22*	Late Cenomanian	0.77	38	47	3	5	2	1	0	0	0	--	1	1	1	--	--	--	--	0.31

Table 1. Organic carbon, hydrogen index and petrographic composition of the sediment samples studied. The petrographic composition is estimated from the observation of two types of microscopic preparations (**) and a polished organic concentrate (*).

5: abundant	d.g.: dark green
4: very frequent	l.g.: light green
3: common	— : no data
2: some	HC : hydrocarbons
1: rare	
0: absent	

In any samples with type II kerogen (CS 7) and mixed type II-type III kerogen (N 21, Ray 26), the higher liptinite content corresponds to an increase in sporinite, alginite, bituminite and particularly in liptodetrinite.

Sporinite: The sporinite is mainly represented by spores; the pollen seems to be rarer. Spores and pollens are often well preserved. They show a fluorescence of yellow and orange-brown colours with weak or more commonly medium intensity.

Resinite: Resinite is rare or absent or extremely difficult to recognize (particularly in concentrates). The bodies which have been assigned to the resinite group are rounded or oval. In reflected white light, these bodies appear rust brown with red internal reflections, dark grey or black. They mostly have a granular surface, contain pyrite and sometimes have a zoned structure or are porous. The reflectance of the resinite ranges from 0.08 to 0.27 %. Resinite shows yellow to green-yellow fluorescence mostly with medium intensity and sometimes with a zonation of colours (pale yellow to light brown to yellow).

Alginite: Alginite is mostly rare or absent except in samples CS 7 (type II kerogen) and N 21, Ray 26 (mixed type II-type III kerogen), where it is more common. Several types of algae are represented including:

- Acritarchs: They have been observed on polished rock mounts of samples CS 330, Ray 26 and Cha 22 in reflected fluorescent light. They show an intense to medium fluorescence of green colour.
- *Tasmanites: Tasmanites* algae are present in the samples Ray 7, N 8, N 23, CS 330, Sey 10, but always in very small numbers. They have a strong green-yellow fluorescence.
- Algae of filamentous shape (5-15 μm) with cell walls and showing a strong yellow-green to green-yellow fluorescence are the best represented algae (samples N 18, N 21, N 23, N 29, N 32, CS 7, CS 330). These algae could tentatively be assigned to the Schizophyceae described by MAEDLER (1968) and TEICHMÜLLER & OTTENJANN (1977) in the Lias of West Germany.
- A well preserved *Botryococcus* algae, yellow-green in reflected fluorescent light was observed in the sample CS 7.
- Each of these alginites is only slightly distinguishable (diffuse light brown colour) from the mineral matrix in reflected white light indicating a low to medium maturation level.

Liptodetrinite: Liptodetrinite was observed in practically all samples and particularly in the samples CS 7 (type II kerogen), N 21, Klu 1, Ber 1 (mixed type II-type III kerogen). Its observations were made on polished rock mounts and in reflected fluorescent light. Liptodetrinite appears mostly scattered in the mineral matrix and shows orange, yellow and green-yellow fluorescence colours. Liptodetrinite consists of fragments and fine degradation remains of liptinite group constituents (STACH et al. 1982) and a part probably derived from tiny unicellular algae (TEICHMÜLLER & OTTENJANN 1977).

Bituminite: Bituminite is a decomposition product of algae, faunal plankton and bacteria (STACH et al. 1982). It is generally rare and is best represented in the samples Ray 26 (mixed type II-type III kerogen), CS 7 (type II kerogen) and Ray 7, Sey 10, CS 330 (type III kerogen). In the present study two types of bituminites were identified.

The first type occurs as oval and rounded bodies with homogeneous finely granular and porous surfaces. These surfaces are sometimes scarred in polished rock mounts. In reflected white light, the bituminite bodies are dark-grey, brown-grey and rust-brown in colour mostly with reddish internal reflections. Their reflectance values range from 0.09 to 0.25 %. They show brown, orange-brown and yellow fluorescence of variable intensity. This bituminite may correspond to the bituminite II described by TEICHMÜLLER & OTTENJANN (1977).

The second type, which is rarer, occurs as lenses of rounded and oval bodies with fine granular structure (in polished rock mounts). In reflected white light, these bodies are of grey or dark grey colours. Their reflectance values range from 0.10 to 0.70 %. They show no fluorescence and may correspond to the bituminite III described by TEICHMÜLLER & OTTENJANN (1977).

- Secondary macerals or products

These constituents are generated during the maturation process. Two kinds of constituents have been observed:

Bitumen: Bitumen has been observed in many samples, but is typically rarer. Bitumen has geometric outlines, homogeneous surfaces and most shows cracks and, sometimes, small rounded cavities. Their reflectance values range from 0.17 to 0.57 %. Bitumen with less than 0.32 % of reflectance shows yellow fluorescence.

Micrinite: It was observed in most samples in very small amounts. Micrinite occurs finely granular in bituminite and in the mineral matrix; it is sometimes associated with large zooclasts and diagenetic carbonate crystals. A great part of this micrinite is developed from the bituminite as described by TEICHMÜLLER & OTTENJANN (1977). This was particularly visible in the samples Cha 22, Klu 1 and Ber 1, which had higher maturation levels.

- Zooclasts

Foraminifera, radiolaria, fish remains, fragments of calcitic macrofossil shell material and various organic faunal relics have been recorded on polished rock mounts.

Foraminifera are frequent in the samples N 21, N 29, N 32, N 40, CS 330 and Ray 26. Some foraminifera tests show no fluorescence and others have slightly yellow-green fluorescence. Radiolaria were found only in sample N 3. Some fish remains are present in samples N 32 and N 3; they are low in reflectance (0.15 - 0.20 %) and show yellow-green fluorescence. Frequent fragments of calcitic shell material occur in the sample N 32.

Several samples (Ber 1, N 18, N 21, N 23, N 29, N 40, Ray 26) contained various faunal relics impossible to assign. In reflected white light, these fragments have irregular shapes, almost structureless, with a fine surface granulation and show brown-grey and dark-brown colours. Their reflectance ranges from 0.10 to 0.25 %. Some fragments show no fluorescence and others show pale green to yellow-green fluorescence.

- Mineral matrix

Blue light microscopy of polished rock mounts reveals that mineral matrix fluorescence is mainly dark-green (Table 1). This fluorescence colour is characteristic of non-bituminous rocks and indicates very low or no adsorption of lipoid substances on the minerals. However, in the samples CS 7 and N 21, the mineral matrix fluoresces light green locally, with a slight increase in intensity. This could indicate some amount of lipoid substances adsorbed to the mineral matrix.

In most samples, some oil expulsions from fissures are visible in reflected fluorescent light if oil immersion on the polished rock mount is carried out during several days. These oil expulsions occur as droplets of less than 1 µm to 2 µm size and as diffused halos along fissures. They have strong fluorescence of green to green-yellow colours. Droplet formation seems to be due to an interface effect between oil (exsudate) and oil immersion and attests to the presence of oil in most samples.

Finally, the mineral matrix contains framboidal and fine crystalline pyrite, especially in samples CS 7, N 21 and Klu 1.

4.3 Maturity of organic matter

T_{max} values and mean vitrinite reflectance were used to assess the degree of kerogen thermal maturation. T_{max} values (Text-Fig. 6), ranging between 431 °C and 461 °C, indicate that samples are situated in the oil-formation zone. The reflectance data of all samples show three distinct populations of reflecting particles (Table 2, Text-Figs. 8-10):

- The first population (R1) with a reflectance from 0.17 to 0.57 % corresponds to the bituminite and bitumen.
- The reflectance of the second population (R2) ranges from 0.44 to 1.20 % and corresponds to the autochthonous vitrinite which is taken to

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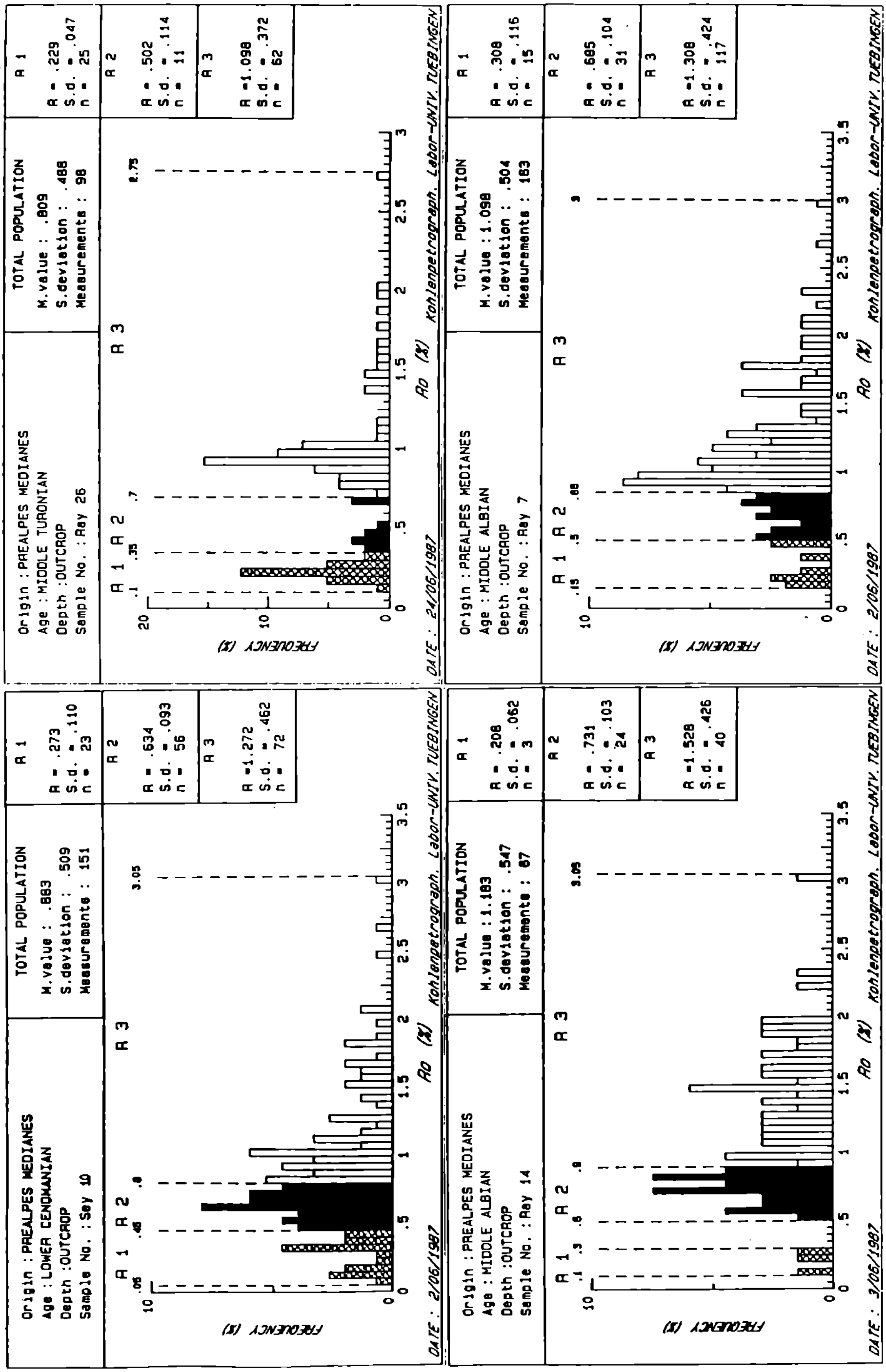
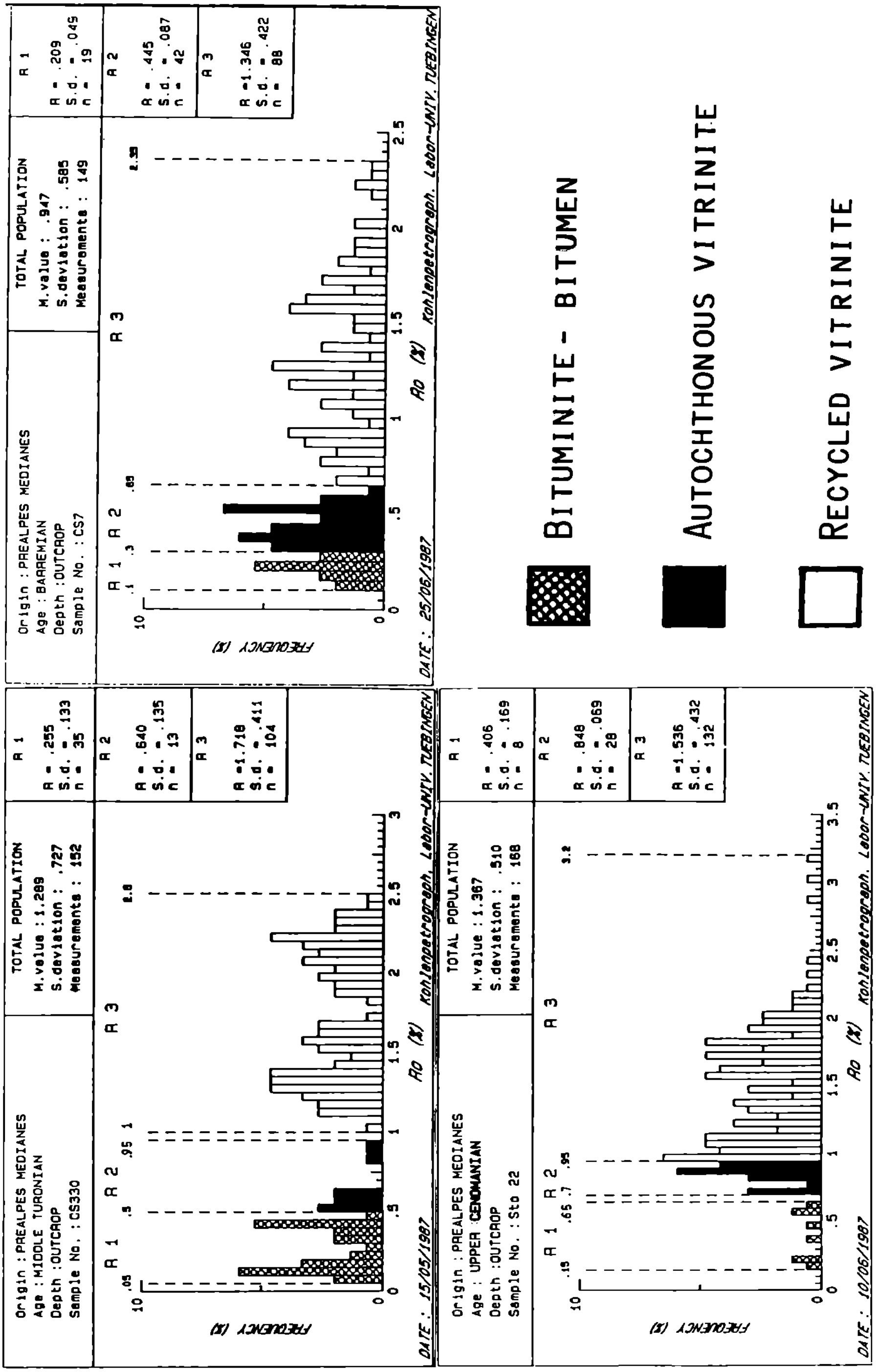
Text-Fig. 8. Reflectance histograms for the following sections: Seytrouset (Sey), Rochers de Naye (Ray), Pra du Pont (CS 330), Comba d'Avau (CS 7) and Stockenflue (Sto).

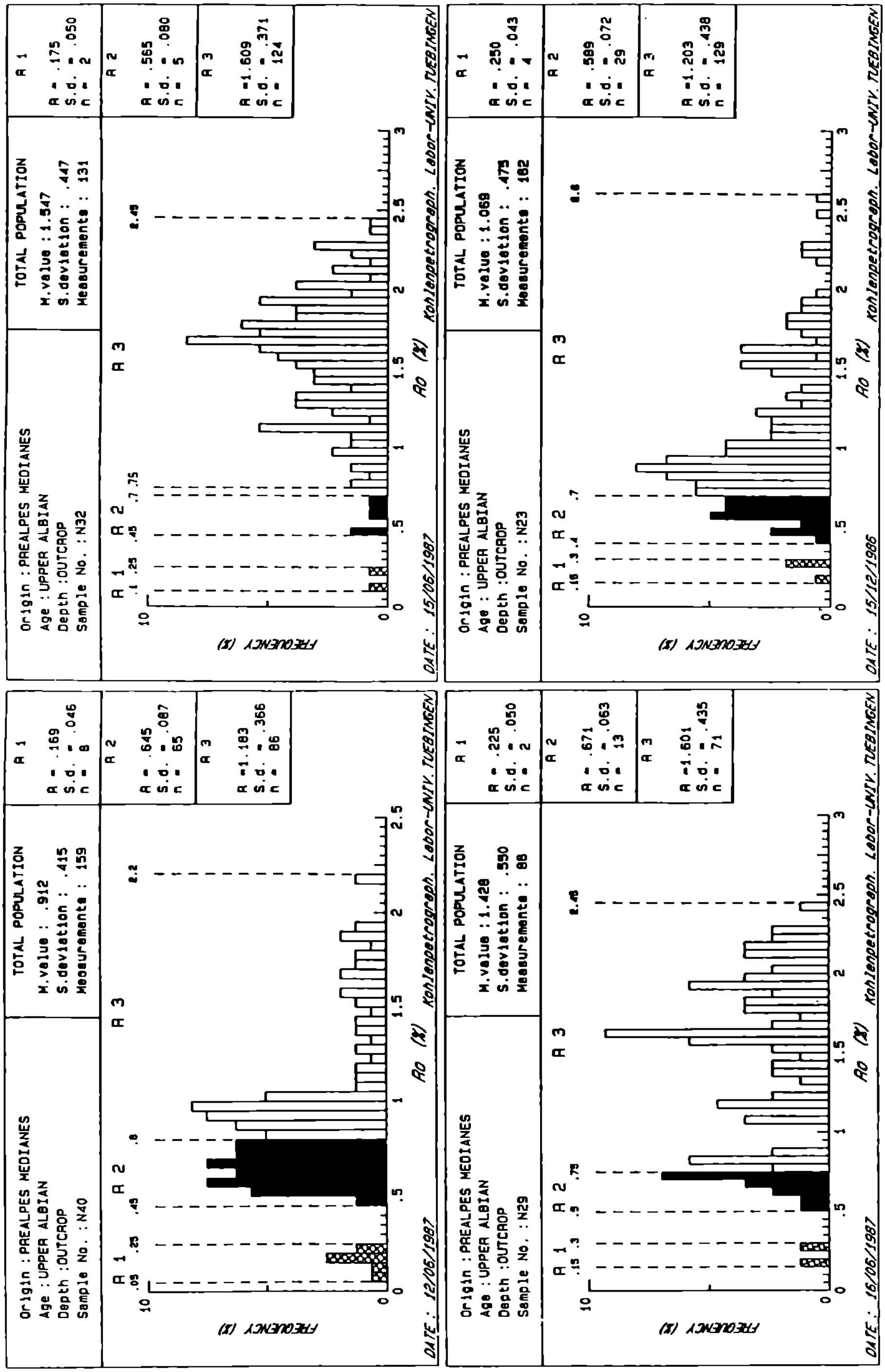
Text-Fig. 9. Reflectance histograms for the Neirivue Section (N).

Text-Fig. 10. Reflectance histograms for the following sections: Hochmatt (Hoc), Chällihorn (Cha), Chlus (Klu), and Béret (Ber).

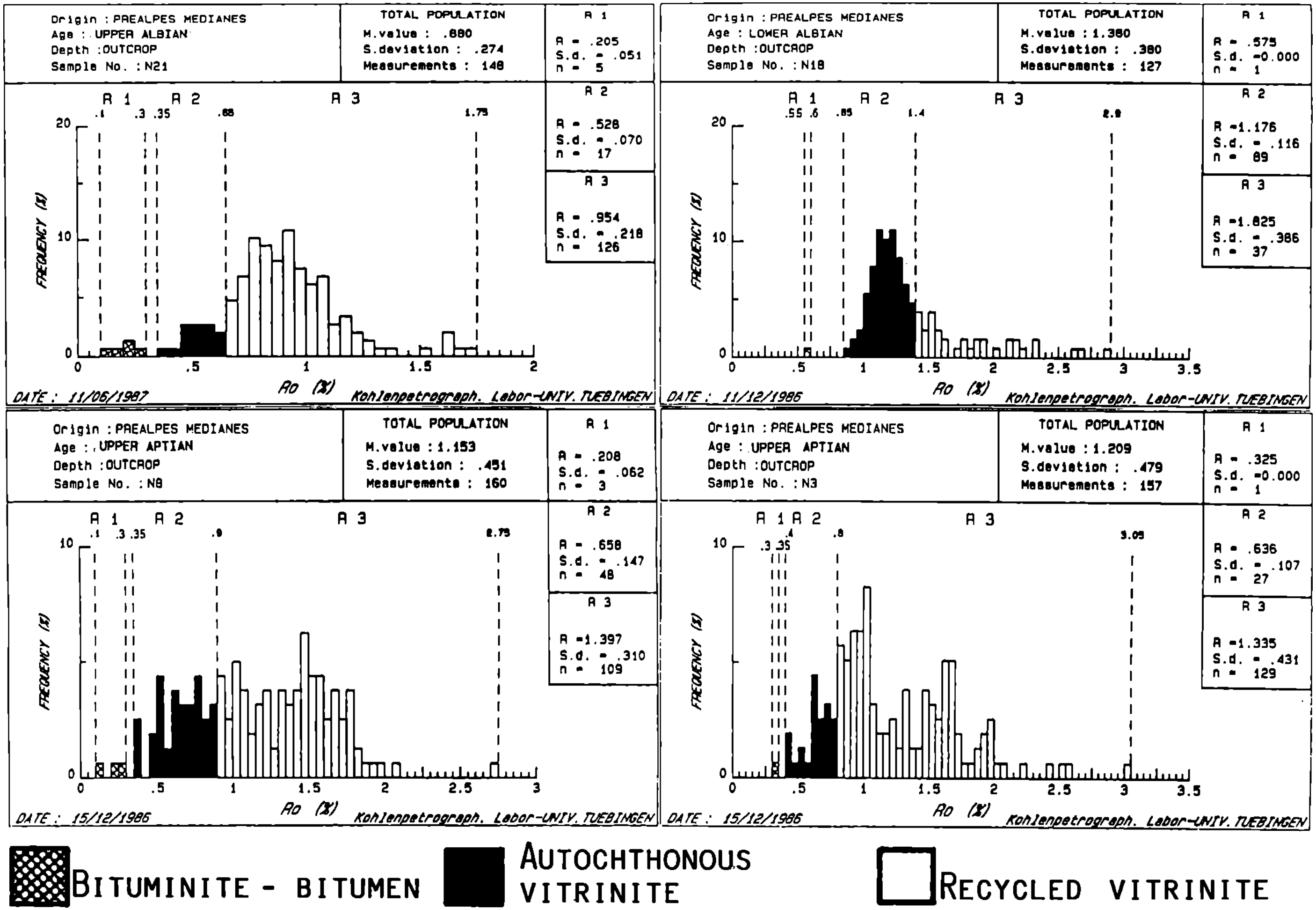

Origin : PREALPES MEDIANES
Age : LOWER CENOMANIAN
Depth : OUTCROP
Sample No. : Sey 10
TOTAL POPULATION
M.value : .883
S.deviation : .509
Measurements : 151
R 1
R = .273
S.d. = .110
n = 23
R 2
R = .634
S.d. = .093
n = 56
R 3
R =1.272
S.d. = .462
n = 72
FREQUENCY (%)
Ro (%)
DATE : 2/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : MIDDLE TURONIAN
Depth : OUTCROP
Sample No. : Rey 26
TOTAL POPULATION
M.value : .809
S.deviation : .488
Measurements : 98
R 1
R = .229
S.d. = .047
n = 25
R 2
R = .502
S.d. = .114
n = 11
R 3
R =1.098
S.d. = .372
n = 62
FREQUENCY (%)
Ro (%)
DATE : 24/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : MIDDLE ALBIAN
Depth : OUTCROP
Sample No. : Rey 14
TOTAL POPULATION
M.value : 1.183
S.deviation : .547
Measurements : 67
R 1
R = .208
S.d. = .062
n = 3
R 2
R = .731
S.d. = .103
n = 24
R 3
R =1.528
S.d. = .426
n = 40
FREQUENCY (%)
Ro (%)
DATE : 3/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : MIDDLE ALBIAN
Depth : OUTCROP
Sample No. : Rey 7
TOTAL POPULATION
M.value : 1.098
S.deviation : .504
Measurements : 163
R 1
R = .308
S.d. = .116
n = 15
R 2
R = .685
S.d. = .104
n = 31
R 3
R =1.308
S.d. = .424
n = 117
FREQUENCY (%)
Ro (%)
DATE : 2/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN


Origin : PREALPES MEDIANES
Age : BARREMIAN
Depth : OUTCROP
Sample No. : CS7
TOTAL POPULATION
M.value : .947
S.deviation : .585
Measurements : 149
R 1
R = .209
S.d. = .049
n = 19
R 2
R = .445
S.d. = .087
n = 42
R 3
R =1.346
S.d. = .422
n = 88
FREQUENCY (%)
RO (%)
DATE : 25/06/1987
Kohlenpetrograph. Labor—UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : MIDDLE TURONIAN
Depth : OUTCROP
Sample No. : CS330
TOTAL POPULATION
M.value : 1.289
S.deviation : .727
Measurements : 152
R 1
R = .255
S.d. = .133
n = 35
R 2
R = .640
S.d. = .135
n = 13
R 3
R =1.718
S.d. = .411
n = 104
DATE : 15/05/1987
Origin : PREALPES MEDIANES
Age : UPPER CENOMANIAN
Depth : OUTCROP
Sample No. : Sto 22
TOTAL POPULATION
M.value : 1.367
S.deviation : .510
Measurements : 168
R 1
R = .406
S.d. = .169
n = 8
R 2
R = .848
S.d. = .069
n = 28
R 3
R =1.536
S.d. = .432
n = 132
DATE : 10/06/1987
BITUMINITE - BITUMEN
AUTOCHTHONOUS VITRINITE
RECYCLED VITRINITE

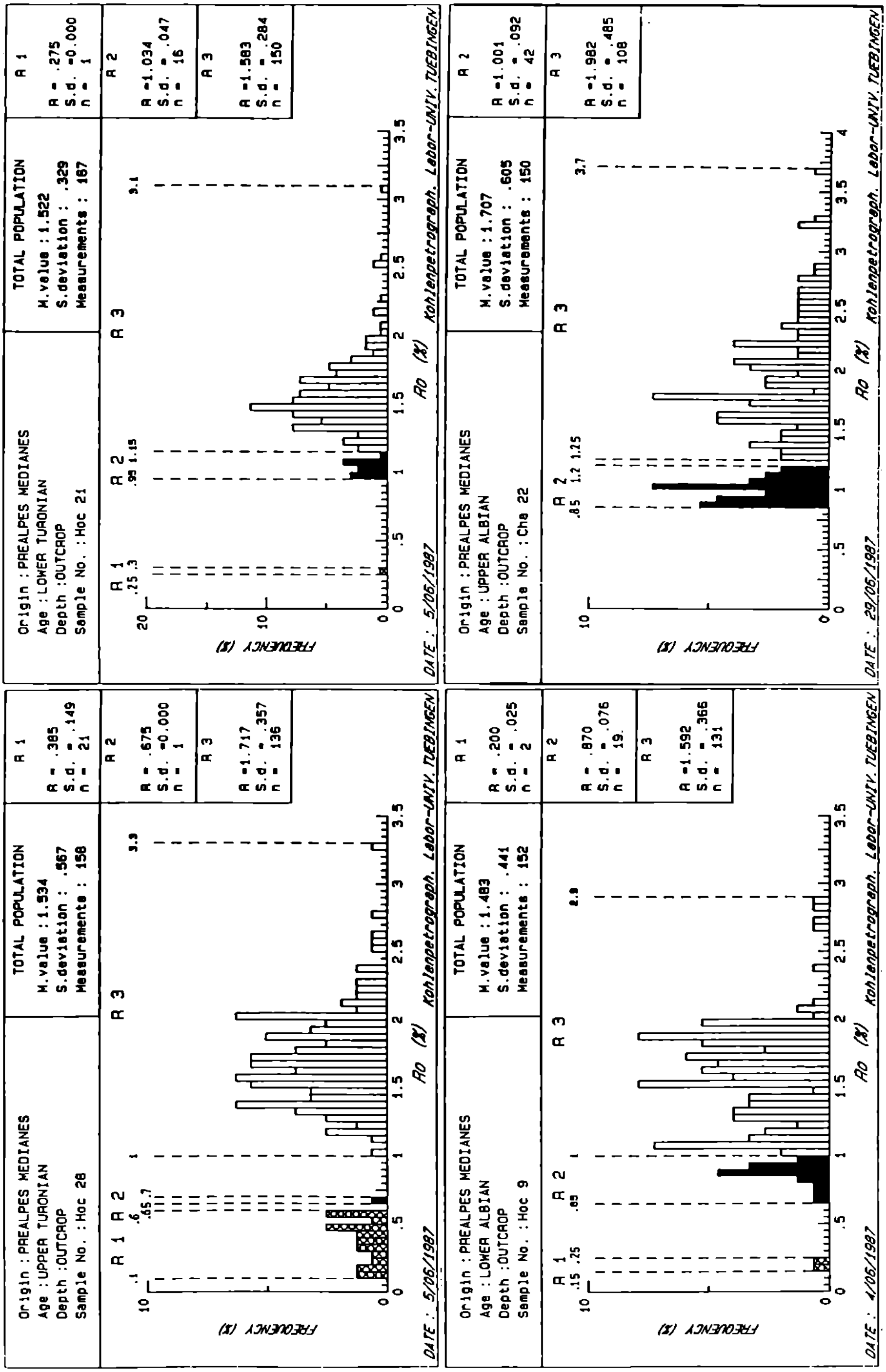
Text-Fig. 8


Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth :OUTCROP
Sample No. : N40
TOTAL POPULATION
M.value : .912
S.deviation : .415
Measurements : 159
R 1
R = .169
S.d. = .046
n = 8
R 2
R = .645
S.d. = .087
n = 65
R 3
R =1.183
S.d. = .366
n = 86
FREQUENCY (%)
Ro (%)
DATE : 12/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth :OUTCROP
Sample No. : N29
TOTAL POPULATION
M.value : 1.428
S.deviation : .550
Measurements : 86
R 1
R = .225
S.d. = .050
n = 2
R 2
R = .671
S.d. = .063
n = 13
R 3
R =1.601
S.d. = .435
n = 71
FREQUENCY (%)
Ro (%)
DATE : 16/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth :OUTCROP
Sample No. : N32
TOTAL POPULATION
M.value : 1.547
S.deviation : .447
Measurements : 131
R 1
R = .175
S.d. = .050
n = 2
R 2
R = .565
S.d. = .080
n = 5
R 3
R =1.609
S.d. = .371
n = 124
FREQUENCY (%)
Ro (%)
DATE : 15/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth :OUTCROP
Sample No. : N23
TOTAL POPULATION
M.value : 1.069
S.deviation : .475
Measurements : 162
R 1
R = .250
S.d. = .043
n = 4
R 2
R = .589
S.d. = .072
n = 29
R 3
R =1.203
S.d. = .438
n = 129
FREQUENCY (%)
Ro (%)
DATE : 15/12/1986
Kohlenpetrograph. Labor-UNIV. TUEBINGEN

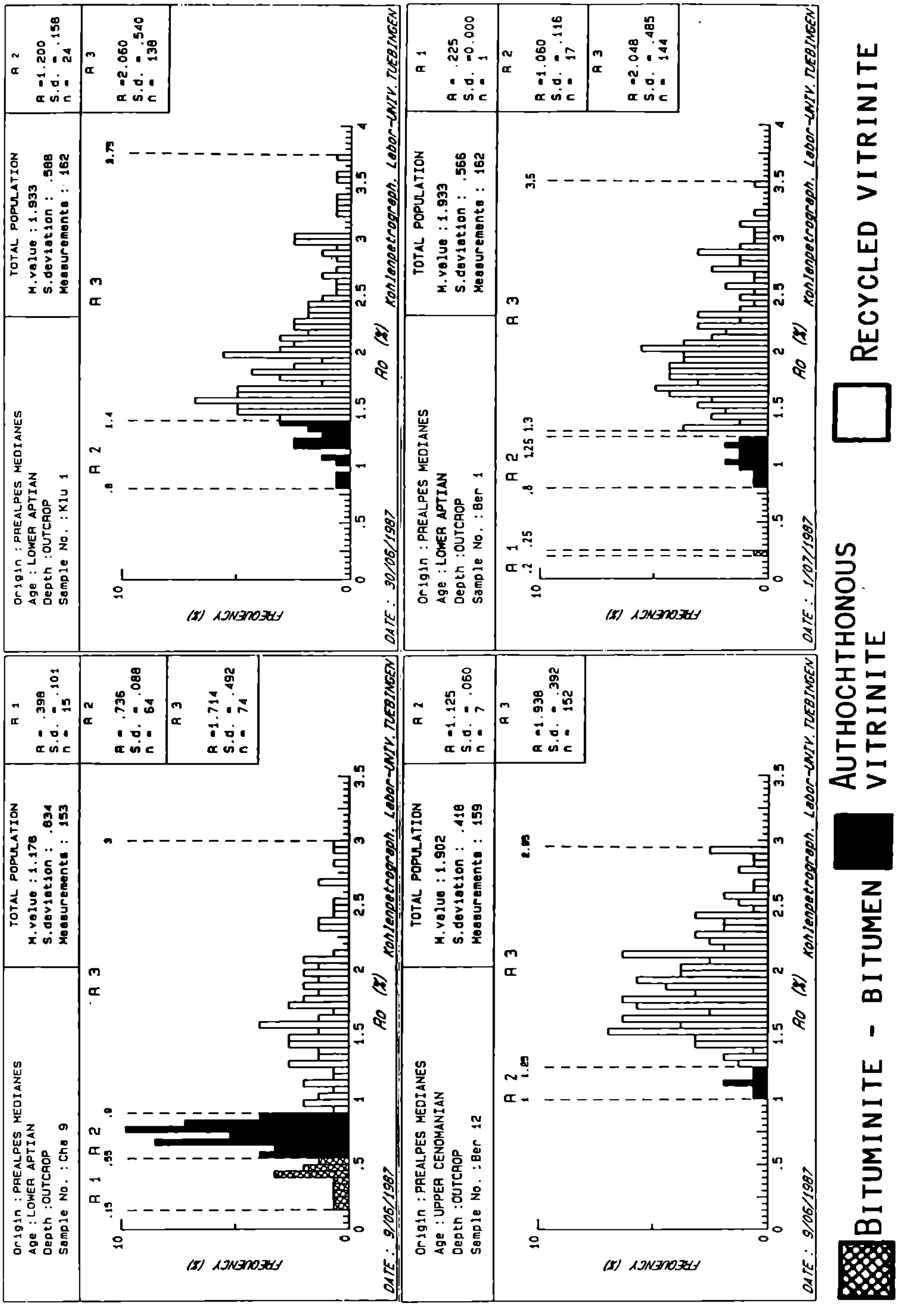
Text-Fig. 9


Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth :OUTCROP
Sample No. : N21
TOTAL POPULATION
M.value : .880
S.deviation : .274
Measurements : 148
A 1
A = .205
S.d. = .051
n = 5
A 2
A = .528
S.d. = .070
n = 17
A 3
A = .954
S.d. = .218
n = 126
FREQUENCY (%)
Ro (%)
DATE : 11/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : LOWER ALBIAN
Depth :OUTCROP
Sample No. : N18
TOTAL POPULATION
M.value : 1.380
S.deviation : .380
Measurements : 127
A 1
A = .575
S.d. =0.000
n = 1
A 2
A =1.176
S.d. = .116
n = 89
A 3
A =1.825
S.d. = .386
n = 37
DATE : 11/12/1986
Origin : PREALPES MEDIANES
Age : UPPER APTIAN
Depth :OUTCROP
Sample No. : N8
TOTAL POPULATION
M.value : 1.153
S.deviation : .451
Measurements : 160
A 1
A = .208
S.d. = .062
n = 3
A 2
A = .658
S.d. = .147
n = 48
A 3
A =1.397
S.d. = .310
n = 109
DATE : 15/12/1986
Origin : PREALPES MEDIANES
Age : UPPER APTIAN
Depth :OUTCROP
Sample No. : N3
TOTAL POPULATION
M.value : 1.209
S.deviation : .479
Measurements : 157
A 1
A = .325
S.d. =0.000
n = 1
A 2
A = .636
S.d. = .107
n = 27
A 3
A =1.335
S.d. = .431
n = 129
DATE : 15/12/1986
BITUMINITE - BITUMEN
AUTOCHTHONOUS VITRINITE
RECYCLED VITRINITE

Text-Fig. 9


Origin : PREALPES MEDIANES
Age : UPPER TURONIAN
Depth : OUTCROP
Sample No. : Hoc 28
TOTAL POPULATION
M.value : 1.534
S.deviation : .567
Measurements : 158
R 1
R = .385
S.d. = .149
n = 21
R 2
R = .675
S.d. = 0.000
n = 1
R 3
R = 1.717
S.d. = .357
n = 136
FREQUENCY (%)
RO (%)
DATE : 5/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : LOWER TURONIAN
Depth : OUTCROP
Sample No. : Hoc 21
TOTAL POPULATION
M.value : 1.522
S.deviation : .329
Measurements : 167
R 1
R = .275
S.d. = 0.000
n = 1
R 2
R = 1.034
S.d. = .047
n = 16
R 3
R = 1.583
S.d. = .284
n = 150
FREQUENCY (%)
RO (%)
DATE : 5/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : LOWER ALBIAN
Depth : OUTCROP
Sample No. : Hoc 9
TOTAL POPULATION
M.value : 1.483
S.deviation : .441
Measurements : 152
R 1
R = .200
S.d. = .025
n = 2
R 2
R = .870
S.d. = .076
n = 19
R 3
R = 1.592
S.d. = .366
n = 131
FREQUENCY (%)
RO (%)
DATE : 4/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN
Origin : PREALPES MEDIANES
Age : UPPER ALBIAN
Depth : OUTCROP
Sample No. : Cha 22
TOTAL POPULATION
M.value : 1.707
S.deviation : .605
Measurements : 150
R 2
R = 1.001
S.d. = .092
n = 42
R 3
R = 1.982
S.d. = .485
n = 108
FREQUENCY (%)
RO (%)
DATE : 29/06/1987
Kohlenpetrograph. Labor-UNIV. TUEBINGEN

Text-Fig. 10


BITUMINITE - BITUMEN
AUTHOCHTHONOUS VITRINITE
RECYCLED VITRINITE

Text-Fig. 10

Table 2. Origin, stratigraphy, pyrolysis parameter T_{max} and reflectance data (reflectance) for sediment samples studied. In some samples, very rare (v) bit and bituminite have been seen only in polished rock mounts.
std. dev. = standard deviation.

ORIGIN (outcrops)	SAMPLE	AGE	TOC (%)	T_{max} (°C)	BITUMINITE + BITUMEN		AUTOCHTHONOUS VITRINITE		RECYCLED VITRINITE		TOTAL POPULATION	
					R1%	n	R2%	n	R3%	n	R%	n
Préalpes du Chablais												
- Seytrouset	Sey A10	Early Cenomanian	0.75	436	0.27	23	0.63	56	1.27	72	0.88	151
Préalpes Romandes												
- R. de Naye	Ray 26	Mid-Turonian	0.82	434	0.23	25	0.50	11	1.10	62	0.81	98
	Ray 14	Mid-Albian	0.36	446	0.21	3	0.73	24	1.53	40	1.20	67
	Ray 7	Mid-Albian	0.67	437	0.31	15	0.68	31	1.31	117	1.10	163
- Pra du Pont	CS 330	Mid-Turonian	0.97	436	0.25	35	0.64	13	1.72	104	1.29	152
- Comba d'Avau	CS 7	Barremian	2.02	432	0.21	19	0.44	42	1.35	88	0.95	149
- Neirivue	N 40	Late Albian	0.83	436	0.17	8	0.64	65	1.18	86	0.91	159
	N 32	Late Albian	0.40	431	0.17	2	0.56	5	1.61	124	1.55	131
	N 29	Late Albian	0.80	434	0.22	2	0.67	13	1.60	71	1.43	86
	N 23	Late Albian	0.90	434	0.25	4	0.59	29	1.20	129	1.07	162
	N 21	Late Albian	1.86	435	0.20	5	0.53	17	0.95	126	0.88	148
	N 18	Late Albian	0.63	(470)	0.57	1	1.18	89	1.82	37	1.36	127
	N 8	Late Aptian ?	0.38	437	0.21	3	0.66	48	1.40	109	1.15	160
	N 3	Late Aptian ?	0.57	435	0.32	1	0.64	27	1.33	129	1.21	157
- La Hochmatt	Hoc 28	Late Turonian	0.56	441	0.38	21	(0.67)	1	1.72	136	1.53	158
	Hoc 21	Early Turonian	0.05	(465)	0.27	1	1.03	16	1.58	150	1.52	167
	Hoc 9	Early Albian	0.49	447	0.20	2	0.87	19	1.60	131	1.48	152
- Challihorn	Cha 22	Late Albian	1.04	457	--	v	1.00	42	1.98	108	1.71	150
	Cha 9	Early Aptian	2.69	444	0.40	15	0.74	64	1.71	74	1.18	153
- Chlus	Klu 1	Early Aptian	2.18	461	--	--	1.20	24	2.06	138	1.93	162
- Beret	Ber 12	Late Cenomanian	0.16	(420)	--	v	1.12	7	1.94	152	1.90	159
	Ber 1	Early Aptian	1.55	456	0.22	1	1.06	17	2.05	144	1.93	162
- Stockenflue	Sto 22	Late Cenomanian	0.77	447	0.41	8	0.85	28	1.54	132	1.37	168

determine the maturation level. We can see in Table 2 and Text-Fig. 6 that a quite good correlation exists between T_{max} and autochthonous vitrinite reflectance R2%.

- The third population (R3) corresponds to the recycled vitrinite with populations between 0.95 and 2.06 % reflectance and shows broad and polymodal distributions (Text-Figs. 8-10).

4.4 Hydrocarbon potential

Total oil potential (S_1 + S_2) is obtained from Rock-eval analysis and expressed in mg HC/g of rock (ESPITALIE et al. 1986). The oil potential is lower than 0.9 mg HC/g of rock for samples of type III kerogen, the sample Cha 9 being one exception with 3.28 mg HC/g of rock (Table 1). The samples of mixed type II-type III kerogen have an oil potential between 1.21 and 3.01 mg HC/g of rock. The sample CS 7 (type II kerogen) shows a fair oil potential with 5.93 mg HC/g of rock. It is possible, that the samples situated in the oil-formation zone, have already released hydrocarbons, and then the sum S_1 + S_2 represents the residual oil potential. This is probably true for the samples Cha 9, Cha 22, Klu 1, Ber 1, which are near the boundary type II-type III kerogen and of high maturation level.

5. Discussion

The environment of the black shales studied may be discussed with reference to the lithological, paleontological or/and petrographical data. They were laid down in a period (mid-Cretaceous) characterized by low oxygen content in the marine waters of the oceans and seas relative to the present and correspond to the OAE_1 and OAE_2 of JENKYNS (1980). SCHLAGER & JENKYNS (1976) proposed that these OAE's correlated with times of equable climate and with transgressive pulses. The petrographic analysis and the pyrolysis data do not allow to distinguish between the two events; the samples corresponding to the OAE_2 (global anoxic event of upper Cenomanian-early Turonian according to JENKYNS 1980, 1985) have no more marine components than these of OAE_1, contrary to the suggestions of ARTHUR (1979) and ARTHUR & NATLAND (1979), and the organic matter content does not vary greatly.

The samples studied for the organic matter are mostly laminated or slightly bioturbated. They sometimes show current lamination with graded radiolaria and/or planktonic foraminifera, caused by sea-bottom currents sweeping organisms from swell to basin (BAUMGARTNER 1988). The blooming of radiolaria precedes and is interbedded in the black shales levels; they can reflect either selective early preservations of siliceous relative to carbonate microfossils (PREMOLI SILVA et al. 1987), or higher productivity due to important nutrients (JENKYNS 1980, BAUMGARTNER 1988) both of which favour oxygen consumption and the apparition of reducing environments. The low content of O.M. (0.5-2.7 %) containing a high amount of vitrinite (especially recycled) and liptinite of terrestrial origin (sporinite, resinite) suggest a temporarily oxygenated environment; some samples (generally corresponding to Barremian-early Aptian) reflect more reducing conditions, which is attested to by lipid-rich marine organic matter. This is destroyed easily under oxidizing conditions (MUKHOPADHYAY et al. 1985, DEMAISON & MOORE 1980). These observations suggest, that the black shales were deposited at the limit of the dysaerobic-anaerobic zone (THOMPSON et al. 1985). Between periods of black shales deposition, the basin returned progressively to more aerobic conditions (grey and homogenized or bioturbated marls and limestones). These terminate in red or green and condensed levels (packstones with planktonic foraminifera). The three facies build up small 5-10 m thick sequences (CARON & DUPASQUIER 1988) analogous to PAC's (GOODWIN & ANDERSON 1985). An oxygen deficient basin model (JONES 1983) reinforced by tectonic barriers (WEISSERT 1981, SAUNDERS et al. 1973, TUCHOLKE & VOGT 1979), such as the Cordillière tarine to the North (ANTOINE & BARBIER 1978) and the Briançonnais to the South, would help to explain these periods with development of oxic or anoxic conditions throughout the basin. These conditions existed up to early Turonian, corresponding to the last calcarenites and (or) condensed levels on the submarine high; the deposits become uniform over the basin and the submarine high at mid-Turonian, with a last low oxic event. This corresponds to the highest sea level period of VAIL et al. (1984) and coincides with the change of sedimentation on the Briançonnais platform.

The maturity of organic matter, as observed from the mean vitrinite reflectance data and temperatures of maximum pyrolysis yields, range into the oil window. The onset of the oil window corresponds to vitrinite reflectances of 0.4 to 0.5 % Rm and 0.5 to 0.6 % Rm respectively, for type II and type III kerogen (ESPITALIE 1986). In basins affected by a normal geothermal gradient, the vitrinite reflectance reaches 0.5 % Rm for a minimum burial depth of 3000 m (ROBERT 1985). In our study, the superposed units are less than 300 m thick, and this is too little to give this maturation. According to MULLIS (1983) and MOSAR (1988), the diagenetic grade is a transported "metamorphic" signature, due to the superposition of the thick Nappes Supérieures over the Préalpes Médianes (CARON 1972).

In vertical sections, the vitrinite reflectance does not present any evolution. Some samples (N 18, Hoc 21) show abnormally high values of PRV that could be due to local tectonics.

6. Conclusions

The black shale deposits, present in the external domain of the Préalpes Médianes (Barremian-early Aptian, late Albian-early Cenomanian, Cenomanian-Turonian boundary of the Subbriançonnais), are the result of several concomitant factors:

- the low oxygen content of mid-Cretaceous oceanic waters,
- the paleotopography of this domain, temporarily barred between two tectonically higher areas (to the North and the South) with consequences on water circulation and its stratification,
- a higher productivity (radiolaria-rich levels) which contributed to oxygen consumption.

The depositional environment of these black marls corresponded with low oxic to oxic conditions, attested to by the low organic carbon content and the dominance of terrestrial O.M. (type III kerogen: vitrinite, inertinite and minor liptinite). During Barremian-early Aptian, local more anoxic conditions are suggested by higher organic carbon content and the increase in marine O.M. (type II kerogen with predominance of marine liptinite and mixed type II-type III).

The maturation of these levels judged by vitrinite reflectance and pyrolysis data does not correspond with the stratigraphic overburden; it corresponds to a transported metamorphism.

From the standpoint of hydrocarbon potential, all sediments of type III kerogen, except one early Aptian sediment, are very poor in quality. The sediments of mixed type II-type III kerogen and especially the Barremian sediments (type II kerogen) are able to generate hydrocarbons.

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Correlations between Mid-Cretaceous Vocontian Black Shales and Helvetic Phosphorites in the Western External Alps

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With 2 Plates and 4 Text-Figures

BRÉHÉRET, J.-G. & DELAMETTE, M. (1989): Correlations between Mid-Cretaceous Vocontian Black Shales and Helvetic Phosphorites in the Western External Alps. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 637-655. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The first correlations (at the scale of ammonite zones) are proposed between phosphorites of the Helvetic Shelf and organic-rich black marls and shales of the Vocontian Basin deposited during mid-Cretaceous time. The major intervals of organic carbon accumulation appear to be coeval with phosphoritic condensed beds on the shelf: in the late Lower Aptian (Goguel Event), the early Lower Albian (Paquier Event), and the late Upper Albian (Breistroffer Event). Thus, these two kinds of deposits seem to be the expressions of the same oceanographical events. Analysis of depositional sequences on the shelf leads to propose that such deposits belong to transgressive system tracts. These transgressive pulses caused starvation on the shelf and weakening of detrital input in the basin, favouring authigenesis as well as anoxia. However, this model does not account for any case since, for example, the phosphatic event on the Helvetic Shelf at the beginning of the Upper Albian does not correspond to an important black shale episode in the Vocontian Basin.

Kurzfassung: Erstmals wird eine Korrelation (auf der Basis von Ammonitenzonen) zwischen den Phosphoriten des Helvetischen Schelfs und den C_{org}-reichen schwarzen Mergeln des Vocontischen Beckens vorgeschlagen, beide während der Mittelkreide abgelagert. Die Haupt-Etappen der Akkumulation organischen Kohlenstoffs und der phosphoritischen Kondensation auf dem Schelf scheinen zeitgleich zu sein: im späten Unterapt (Goguel-Event), im frühen Unteralb (Paquier-Event) und im späten Oberalb (Breistroffer-Event). Danach scheinen diese beiden sehr unterschiedlichen Sedimenttypen das Ergebnis gleicher ozeanographischer Ereignisse gewesen zu sein. Die Analyse der Ablagerungsfolgen auf dem Schelf erlaubt es, diese Bildungen transgressiven Zyklen zuzuordnen. Diese transgressiven Pulse erzeugten Hungersedimentation auf dem Schelf und einen Rückgang des terrigenen Inputs im Becken, dabei Authigenese und Anoxia begünstigend. Allerdings trifft dieses Modell nicht in jedem Falle zu, da z. B. das Phosphorit-Ereignis auf dem Helvetischen Schelf zu Beginn des Oberalb keinem wichtigen Schwarzschiefer-Ereignis im Vocontischen Becken entspricht.

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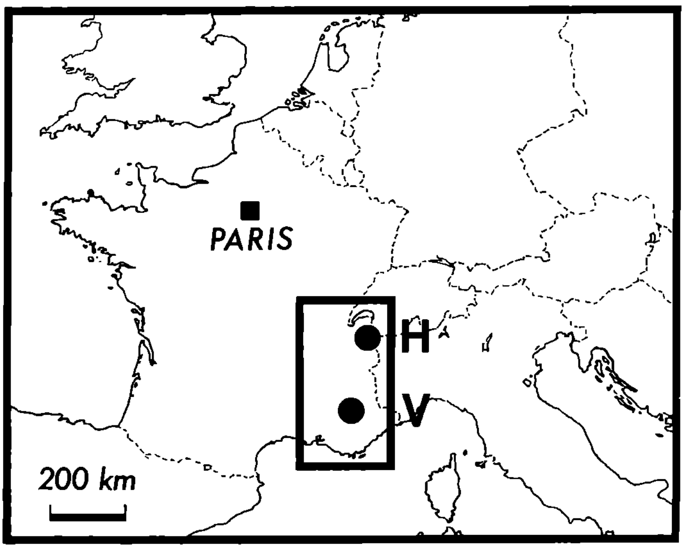
1. Introduction

Mid-Cretaceous black organic-rich lithologies were pointed out some years ago in a number of basins, especially in the Tethyan Realm (SCHLANGER & JENKYNS 1976, JENKYNS 1980). Approximate correlations with phosphoritic and glauconitic occurrences on the adjacent shelf were suggested. More recently, regional studies have been carried out in the basinal domain (BRÉHÉRET et al. 1986, BRÉHÉRET 1988) as well as on the shelf (DELAMETTE 1986). Detailed lithostratigraphy was established for both domains and a biostratigraphical frame was drawn up (mainly with the data of ammonites and planktonic foraminifera). Some organic-rich events have been shown to occur at the scale of the Alpine Sea (BRÉHÉRET 1985, 1988) and shelf phosphorite deposits were shown to be widespread as well (DELAMETTE 1988). The opportunity here is good enough to try to sketch a correlation between these two facies - basinal organic-rich layers and shelf phosphoritic layers - in order to set up their spatio-temporal relations.

The two areas that we are comparing here are the **Vocontian Basin** and the **Delphino-Helvetic** Domain located in the south-east of France (Text-Fig. 1). The first one, located east of the river Rhône and south of the river Drôme, corresponds to the Diois, Baronnies regions and Verdon valley that belong to the **Southern Subalpine Range.** The second one, situated in the Haute-Savoie area, south-east of the Léman lake, covers the Haut Giffre, Platé and Borne massifs, that belong to the **Northern Subalpine Ranges.** By mid-Cretaceous time both areas corresponded to the western part of the northern margin of the Alpine-Tethys Sea. The Vocontian



Text-Fig. 1. Location of the studied area. H: Helvetic Shelf, V: Vocontian Basin.

Basin was surrounded by shoals that were characterized, mostly for the Vercors and Ventoux areas, by the relief of the Lower Cretaceous Urgonian carbonate platform, which developed until the Lower Aptian (ARNAUD-VANNEAU et al. 1984). By mid-Cretaceous time, the Helvetic domain constituted shoals comparable to Vercors.

2. The mid-Cretaceous deposits of the Vocontian Basin

2.1 Introduction

After the prevailing limestone-marl cyclic sedimentation during Lower Cretaceous time, an abrupt change occurs at the end of the Upper Aptian. Up to 750 m of black marls and shales are deposited in the pelagic central part of the basin: the Marnes Bleues Formation. Only scarce limestone beds may be distinguished in the monotonous succession of black marls and shales. Yet, in spite of its homogeneous aspect, the section presents a cyclic sedimentation throughout (Text-Fig. 2 and Pl. 1, Fig. 1). This is due to the fluctuations of the planktonic carbonate flux - mostly planktonic foraminifera and calcareous nannofossils - and the changes in the colour (dark and light banding, because of iron sulphide abundance) and burrowing intensity as a consequence of redox parameter. In some instances, dark lithofacies are completely devoid of bioturbation, thus presenting a discrete lamination of planktono-euxinic type. Several may be described as paper (calcareous) shales. For the major part of the formation, continuity of the horizons may be stated at the basin scale by the means of correlations from layer to layer. However, siliciclastic turbidites, and slumps break this autochthonous record, mainly by Aptian time.

2.2 **Lithology** (Text-Fig. 2)

Black to blue marls occur upon a Lower Aptian hardground, slump or debris flow surface. After several meters of greenish calcareous shales follows a succession of numerous black laminated thin horizons of paper shales alternating with homogeneous dark layers for about 20 m: the Goguel Level (BREHERET 1988). A lot of thin turbidites (at a centimeter or millimeter scale) are interbedded. Some of them are calciturbidites, and several are very rich in phosphoclasts. The relative scarcity of planktonic fauna (radiolaria and foraminifera) and the almost entire absence of benthic fauna characterize this level as a result of strong anoxia. However, a good preservation of organic matter is observed. In the southern part of the basin, a decimeter thick glauconitic marly layer occurs instead of the black shales.

The succession becomes more limy upsection, and is marked by the setting up of limestone-marl alternations. Intense burrowing is a characteristic feature together with the occurrence of macro- and micro-benthic organisms (bivalves, gastropoda, foraminifera). This episode represents the late Lower Aptian, based on ammonite data. It corresponds to the extent of the planktonic foraminifera *Schackoina cabri*.

Then a clayey-rich dark cyclic sedimentation again takes place at the base of the Upper Aptian. Only rare horizons are laminated; bioturbation is developed throughout. *Chondrites* and even *Planolites* may be found in


MARNES BLEUES
T. O. C.
CaCO3
smectite / illite
quartz
1 2 3 %
50 %
1
30 %
CENOMANIAN
UPPER ALBIAN
M. ALB.
LOWER ALBIAN
UPPER APTIAN
LOWER APTIAN
B.l.
P.l.
K.l.
J.l.
w.l.
G.l.
100 m

the darker lithofacies. Benthic foraminifera never disappear completely. The restriction is by far less intense than in the first event.

Then sedimentation becomes more and more limy up to a limestone bundle: the so-called Faisceau Fromajet (BRÉHÉRET & DELAMETTE, in press) that marks the middle part of the uppermost Aptian ("Clansayesian") deposits. The limestone-marl cyclic sediments are rich in benthic fauna (*Aucellina*, foraminifera). They are characterized by abundant radiolaria and nannoconids. Several meters of black shale facies rich in ammonites with preserved aragonite - the Jacob Level (BRÉHÉRET 1983) - are interbedded in the upper part of this unit. Its top is marked by an almost azoic black shale layer, 1 m thick: the Kilian Level (ibid.).

The Lower Albian interval displays a black shale facies relatively poor in carbonate (8 to 25 %); planktonic foraminifera are then rare and small. More or less laminated horizons alternate with black homogeneous marls or shales, comprising scarce bioturbation. Well laminated horizons often are clustered in doublets or triplets (i. e. a superposition of two or three first order cycles) (Pl. 1, Fig. 5). A particular level that is known in other parts of the Alpine Tethys Sea may be distinguished as a key bed: the Paquier Level (ibid.) (Pl. 1, Fig. 1-4) which is a paper shale very rich in ammonites with crushed aragonite shells, particularly *Leymeriella tardefurcata* (D'ORB.) (Pl. 1, Fig. 3). This level is richer in carbonate (up to 55 %) due to abundant (but minute) planktonic foraminifera and *Nannoconus* (Pl. 1, Fig. 4). In the mid-Albian and at the base of the Upper Albian, the carbonate phase increases because of the abundance and size of planktonic foraminifera. Ammonites and bivalves may be abundant in some laminated layers. *Birostrina concentrica* and *B. sulcata* are frequent in the upper part of this unit. The total thickness of this interval is approximately 200 m.

About 100 m of a limestone-marl alternation follow. The fairly burrowed sediment is rich in pyritic moulds or imprints of ammonites, and shells of *Aucellina* and Pectinaceae. Planktonic foraminifera are abundant.

With the uppermost Albian ("Vraconian") a slightly more argillaceous sedimentation takes place with again some laminated layers. The main group of them, the Breistroffer Level (BRÉHÉRET 1988) contains numerous crushed aragonite shells of ammonites. Planktonic foraminifera are abundant with the typical *Rotalipora* group. The succession becomes very monotonous upsection. This kind of marly sedimentation continues in the lowermost Cenomanian, but without any laminated layer.

Text-Fig. 2. Lithology of the Aptian-Albian Marnes Bleues Formation on a synthetic composite section, with total organic carbon (T.O.C., values obtained by Rock Eval analysis), $CaCO_3$ (calcimetry), smectite/illite ratio, and quartz data (based on X-ray diffractometry semi-quantitative estimations); G.L.: Goguel Level; w.l.: white level; J.L.: Jacob Level; K.L.: Kilian Level; P.L.: Paquier Level; B.L.: Breistroffer Level.

2.3 Analytical data (Text-Fig. 2)

Organic matter. Data on organic matter have been obtained by Rock-Eval pyrolysis, theory and application of which can be found in ESPITALIE et al. (1977, 1985a, 1985b, 1986). Total organic carbon (T.O.C.) and the type of organic matter (O.M.) may so be characterized. A sketch of T.O.C. values is given on Text-Fig. 2. Detailed considerations are in BREHERET (1988). Variations in T.O.C. concentrations clearly mirror the lithology. Thus well calcareous bioturbated light facies contain between 0.3 and 0.5 % T.O.C. and darker equivalents between 0.5 and 0.7 %. Black homogeneous layers contain between 0.7 and 1.2 %, and black laminated horizons contain 1 to 2 % T.O.C. The well laminated fissile layers (paper shales) show more than 2 % and even up to 6 % T.O.C. The richer layers contain hydrogenous organic matter of marine planktonic origin (type II), beside that of continental origin (type III). It is preserved as a consequence of restricted environment. Major accumulation periods of organic matter

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Plate 1. Mid-Cretaceous black shale deposits of the Vocontian Basin.

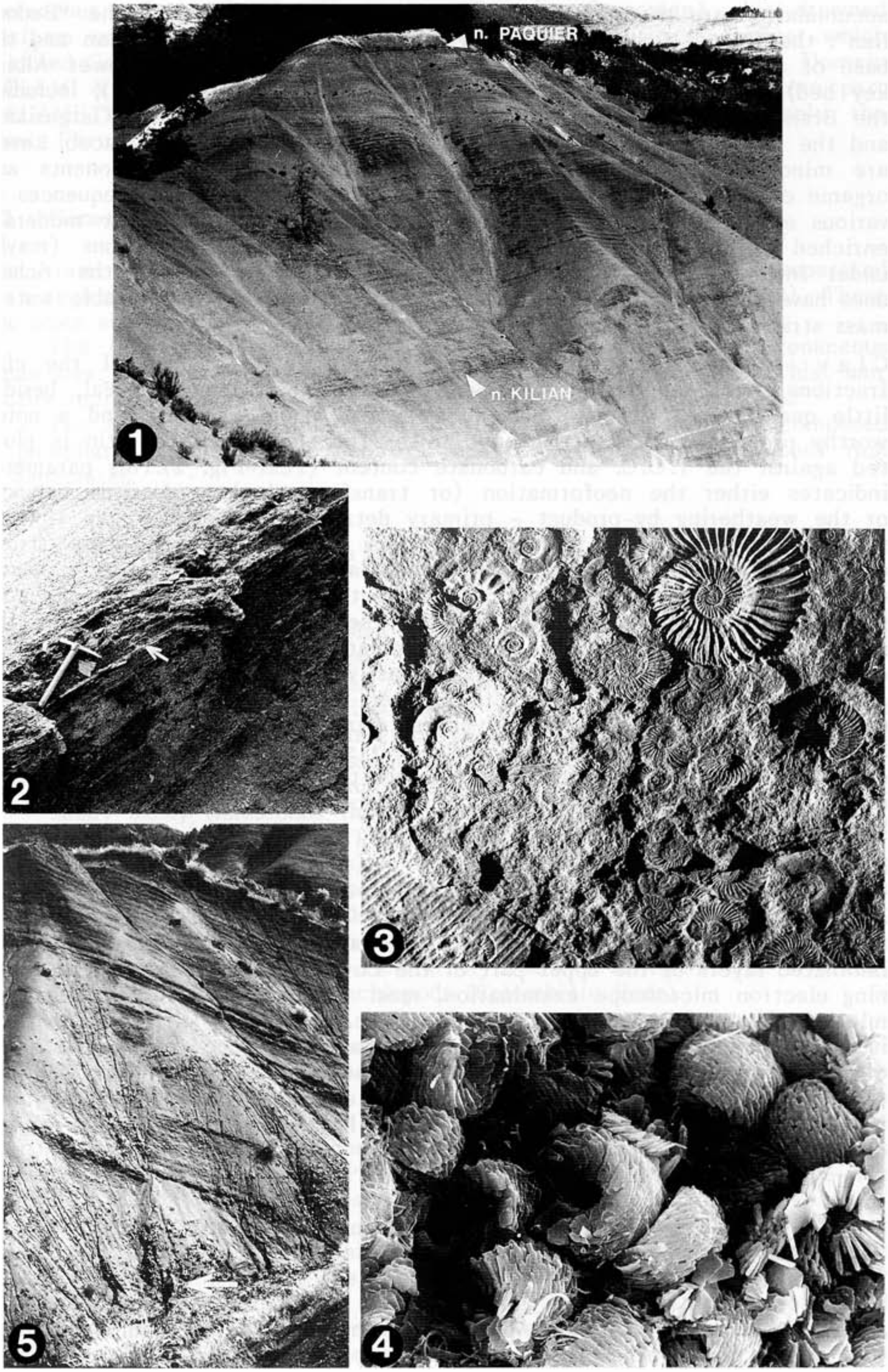
Fig. 1. Lowermost Albian beds. Les Briers, near Saint-André-les-Alpes.
The Kilian level marks the top of the Aptian beds characterized by strongly bioturbated light-coloured marls. Above, the lowermost Albian black marls and shales display a fairly expressed rhythmic bedding. Some scarce darker beds appear. Due to the stronger effect of compaction on these laminated layers, and their anisotropy they are thinner than non-laminated ones. The Paquier Level located at the top of the section is a paper shale that represents a key bed in the basin and the Alpine Tethys.

Fig. 2. The Paquier Level. Palluel Pass, near Rosans (Hautes Alpes).
The level is constituted of a set of cm-scale sequences composed of (1) a well laminated fissile lithofacies, and (2) a bioturbated or faintly laminated moderately fissile lithofacies. The arrow points out a calcareous layer very rich in *Nannoconus*.

Fig. 3. Biofacies of the Paquier Level. Pré-Guittard Pass, near Arnayon (Drôme).
The layers of this paper shale contain a very rich ammonite fauna, largely dominated by the genus *Leymeriella* in which the *tardefurcata* group is the best represented. The numerous aragonitic phragmocones are crushed.

Fig. 4. Nannofacies of the Paquier Level. Combe de Bagna, la Farette (Drôme).
Numerous laminae and calcitic lenses are disposed in the layers of the paper shale. They are mostly formed by an accumulation of the nannofossil *Nannoconus truitti*. This constitutes an anomaly because of their quasi-absence in the surrounding marls and shales.

Fig. 5. Upper part of the Lower Albian and Middle Albian beds. Pré-Guittard Pass, near Arnayon (Drôme).
The cyclicity of fissile and non-fissile lithofacies is well developed. The most fissile organic-rich layers are grouped by two or three units. The person (arrowed) gives scale.


n. PAQUIER
n. KILIAN
1
2
3
4
5

accumulation are the uppermost Lower Aptian (upper part of the "Bedoulian": the Goguel Level), the Lower Albian plus the Middle Albian and the base of the Upper Albian (including the Paquier Level, the Lower Albian key bed), and the upper part of the Upper Albian ("Vraconian"), including the Breistroffer Level. The middle part of the Upper Aptian ("Gargasian") and the upper part of the Upper Aptian (the "Clansayesian" Jacob Level) are minor events. Lithological characteristics, biological components and organic content indicate that the organic-rich deposits are consequences of various environmental conditions. However, it can be stated that moderate enriched horizons have been deposited under hypoxic conditions (maybe under influence of strong oxygen minimum zone?), whereas the richest ones have been deposited under anoxic conditions generated by stable water-mass stratification (DELAMETTE et al. 1986a).

C l a y m i n e r a l o g y. X-ray diffractometric studies of the clay fractions have been carried out. Semi-quantitative estimates reveal, besides little quantities of chlorite and mixed-layers, some kaolinite and a noteworthy proportion of smectites and illite. The smectite/illite ratio is plotted against the T.O.C. and carbonate content (Text-Fig. 2).This parameter indicates either the neoformation (or transformation) - detritus balance, or the weathering by-product - primary detritus balance. The gap towards the base of the column corresponds to a section localized under strong thermic influence due to overthrusting (Gaubert); it generated a nearly complete diagenetic transformation of smectites into chlorite. Strong correlations are clearly established between the quantity of smectites in the sediment and the organic matter accumulation episodes. They correspond with depletions in the carbonate curve, as well. The most prominent correlations are the Goguel Level, the Lower Albian, and the upper part of the Upper Albian. An attempt to characterize smectites has been made by means of saturation tests by Li^+ and K^+. It shows that the capacity of exchange is variable. Both neoformation and transformation may occur. Low charge beidellite constitutes in some instances an important component of the clay assemblage.

P h o s p h o r i t e c o n c r e t i o n s. Scarce phosphatic concretions occur in shales and marls. Most of them are mineralizations of burrows and some kinds of trace fossils. Others are centered on remains of ammonites, bivalves and fish. Phosphoritic lenses are particularly abundant in the laminated layers of the upper part of the Lower Albian. As stated by scanning electron microscope examination, most of them are the by-product of microphytobenthos activity near the interface. The X-ray diffraction pattern is that of francolite at various stages of diagenesis according to the origin of concretions in the basin. Their concentration in P_2O_5 is comprised between 11 and 33 %, whereas that of the enclosing beds is usually lower than 0.5 %. Their occurrences are clearly linked to the extension of the organic matter accumulation episodes.

3. The mid-Cretaceous deposits of the Helvetic Shelf

3.1 Introduction

At the end of the Lower Aptian, the shallow-water Urgonian platform (Barremian-Lower Aptian) drowned. Before the deposition of the Upper

Cretaceous pelagic Seewen limestones (Turonian-Santonian), this drowned platform was covered by a thin (5 to 100 m) glauconitic sand sheet which is called Garschella Formation in the northern part of the Helvetic Domain (FÖLLMI & OUWEHAND 1987) and Aravis Formation in the southern part (DELAMETTE et al. 1986b, DELAMETTE 1986). Furthermore, only the Aravis Formation will be treated (Pl. 2, Fig. 1).

3.2 General stratigraphic framework of the Aravis Formation

The Aravis Formation is composed of two stratigraphical units separated by a regional disconformity (late Upper Aptian disconformity: D 2). These two units are:

- The Bossetan Member (Upper Aptian) composed of calcarenaceous sandstones rich in large benthic fauna (especially bryozoans). This unit usually represents more than 80 % of the thickness of the formation.
- The Platé Member (late Upper Aptian to mid-Cenomanian) composed of phosphate-rich glauconitic deposits with several interstratified beds rich in ammonites.

a) The Upper Aptian deposits of the Bossetan Member

These deposits overly an extensive hardground which covers the top of the Urgonian limestones (early Upper Aptian disconformity: D 1). This hardground, often bored by bivalves, is usually phosphatized and ferruginized and sometimes bears phosphatized small ammonites (*Epicheloniceras* and *Colombiceras*) indicating the base of Upper Aptian. Except for some areas where the deposits of the Bossetan Member are missing or truncated by the late Upper Aptian unconformity, the Upper Aptian sequence includes:

- shelly sandy limestones rich in sponges (DELAMETTE et al. 1986b), bryozoans (DELAMETTE & WALTER 1984), belemnites and with some phosphatized ammonites (Aujon Beds).
- argillaceous fine sandstones containing only few remains of bivalves (mainly Limidae) (Borderan Beds). In some more argillaceous facies, ostracods and foraminifera have been collected (CHAROLLAIS et al. 1971).
- calcarenaceous sandstones rich in bryozoans (DELAMETTE & WALTER 1984), serpulids, oysters and brachiopods (Colombière Beds).

b) The Albian-Cenomanian deposits of the Platé Member

Starting over a regional disconformity bearing phosphate and uppermost Aptian ammonites (D 2), the Platé Member is composed of phosphate-rich glauconitic deposits which represent a vertical evolution from a neritic sand-rich facies bearing phosphorites and ammonites (uppermost Aptian to Upper Albian) to hemipelagic glauconite-rich foraminiferal mudstones (uppermost Albian to mid-Cenomanian). The deposits are characterized by:

- a great lateral facies change passing from authigenic-dominated facies with microbial encrustations in the hypercondensed successions (Pl. 2, Fig. 3-5) to sand-dominated facies in the more dilated successions.

- several fossiliferous phosphatic conglomerates which represent condensed and/or reworked horizons (Pl. 2, Fig. 2-4).

Plate 2. Condensed mid-Cretaceous deposits of the Helvetic Shelf.

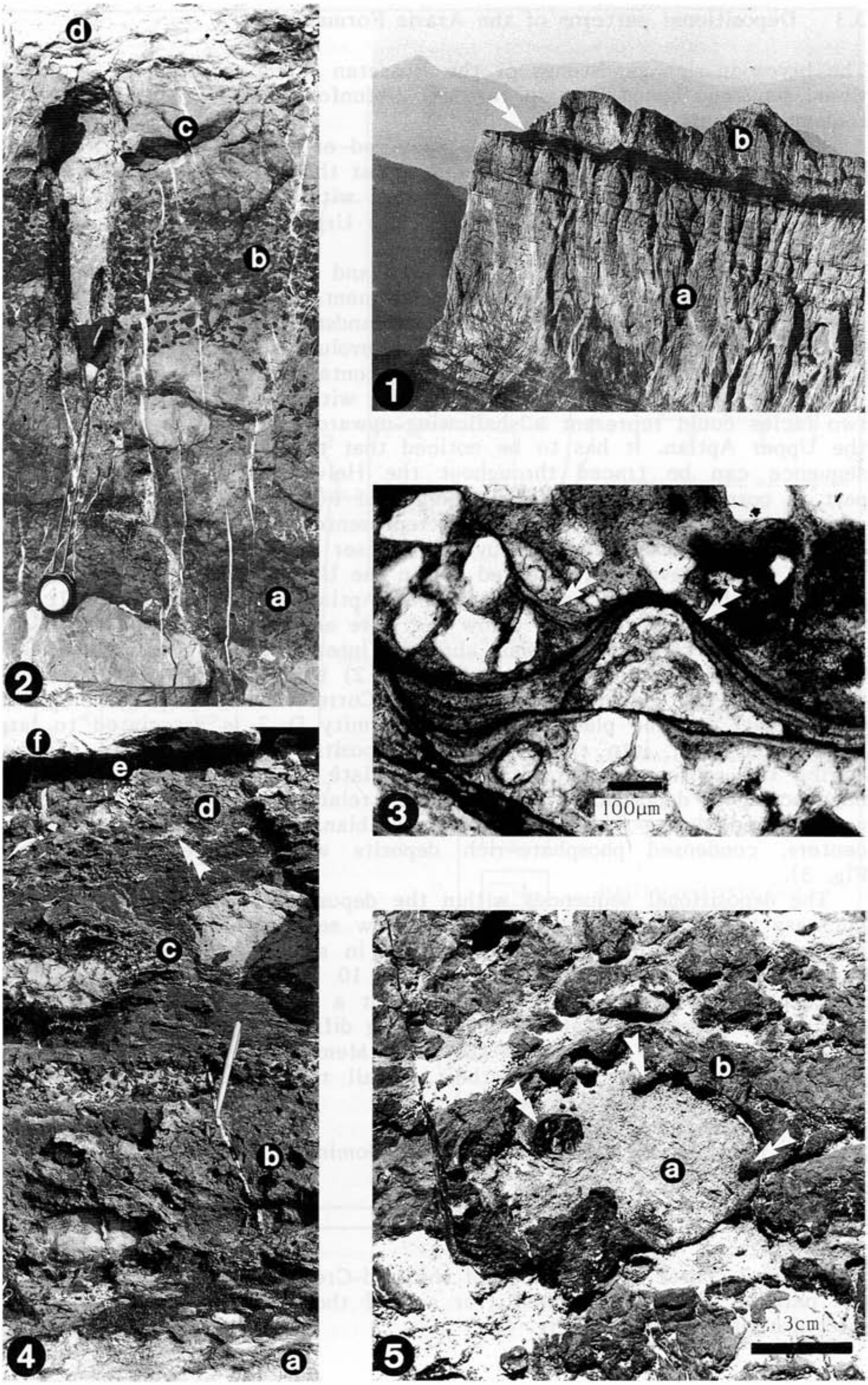
Fig. 1. The Aravis Formation (white arrow) in the southern part of the Helvetic Domain. Platé Massif, Haute Savoie.
This formation is represented by thin siliciclastic phosphate-rich deposits between the shallow-water Urgonian limestones (a) and the pelagic foraminiferal Seewen limestones (b).

Fig. 2. The upper part of the Aravis Formation in a relatively dilated succession. Lindars, Platé Massif, Haute Savoie.
a: abundant glauconitic bioturbations (*Thalassinoides* ichnosp.) associated to the disconformity D 5 (see Text-Fig. 3). This disconformity separates the Middle Albian argillaceous very fine sandstones from the Upper Albian phosphate-rich medium sandstones. b: phosphatic conglomerate mainly composed of Upper Albian ammonites (condensed section C 5: see Text-Fig. 3). c: hemipelagic glauconite-rich foraminiferal mudstone with late Upper Albian ammonites (*Mariella bergeri, Stoliczkaia* gr. *dorsetensis*), Lower Cenomanian ammonites (*Schloenbachia varians, Neostlingoceras carcitanense*) and Lower to Middle Cenomanian planktic foraminifera (Brotzeni and Reicheli zones) (condensed section C 6: see Text-Fig. 3). d: pelagic foraminiferal mudstone of the Seewen limestones formation (Helvetica Zone) (scale: the altimeter is 5 cm in diameter).

Fig. 3. Phosphatic microbial films (arrows) incrusting a sessile agglutinated foraminifera. Thin slide MD 410b (for origin see under Fig. 5).
This kind of microbial encrustation has only been detected in the Lower-Middle Albian deposits of the hypercondensed successions (see Fig. 2).

Fig. 4. Hypercondensed succession composed of stacked phosphate-rich beds. Rocher des Fiz, Platé Massif, Haute Savoie.
a: top of the Upper Aptian deposits (Bossetan Member). b: late Upper Aptian and early Lower Albian massive phosphorite (C 2-3, Text-Fig. 3). c: late Lower Albian and early Middle Albian sandy argillaceous phosphorites with carbonated nodules (C 4). The white arrow points to discrete glauconitic bioturbations which represent the disconformity D 5 (compare with Fig. 2). d: upper Albian phosphatic conglomerate (C 5). e: thin condensed level (not exposed on this picture) composed of micritic stromatolites encrusting Lower Cenomanian phosphatized ammonites (C 6). f: Upper Cenomanian pelagic foraminiferal mudstones of the Seewen Formation. Note that the base of this formation belongs to the Cushmani Zone instead of to the Helvetica Zone as seen on the Fig. 1 (scale: pen is 13 cm in length).

Fig. 5. Residual nodule (a) perforated by bivalves *Gastrochaenolites* ichnosp. (white arrow) and encrusted by phosphatic crust (b). Fenêtre à Grappins, Haut Giffre Massif, Haute Savoie.
This nodule was found just above the late Upper Aptian disconformity (D 2 on Text-Fig. 3). Its facies is that of the Colombière Beds which here underlay the phosphate-rich deposits of the Platé Member.



1
2
3
4
5
100μm
3cm

3.3 Depositional patterns of the Aravis Formation

The bryozoan-rich sandstones of the Bossetan Member represent a depositional sequence bound by two regional disconformities (D 1 and D 2). This sequence comprises:

- a transgressive system tract composed of a condensed section (Aujon Beds) overlying the marine (?) hardground at the top of the Urgonian limestones. This hardground may be associated with a hiatus, but no biostratigraphical data coming from the top of the Urgonian limestones are actually available.
- a highstand system tract (Borderan and Colombière Beds) which present a coarsening-upward siliciclastic sedimentation (argillaceous fine sandstones to medium-grained calcarenaceous sandstones). Perhaps, this coarsening-upward trend may reflect a regressive evolution. In fact, in some places the upper part of the Colombière Beds contain unfossiliferous well-sorted sandstones and calcarenaceous sandstones with cross-stratification. These two facies could represent a shallowing-upward evolution during the end of the Upper Aptian. It has to be noticed that the Upper Aptian depositional sequence can be traced throughout the Helvetic Shelf. In the northern part, it corresponds to the Brisi Member of FÖLLMI & OUWEHAND (1987) with the transgressive system tract represented by the Luitere-Zug Beds and the highstand system tract by the Gamser Beds and the Brisi Beds.

The vertical evolution detected within the Upper Aptian deposits is interpreted as the building up of an Upper Aptian bio-siliciclastic shelf in a storm-influenced environment below the wave action level.

This shelf construction was abruptly interrupted at the end of the Aptian. A major regional disconformity (D 2) bearing phosphate and ammonites covers the Upper Aptian deposits. Correlation between outcrops has shown that, in some places, this disconformity D 2 is associated to large scale scours cut into the underlying deposits of the Bossetan Member. During the subsequent deposits of the Platé Member, these scours have been acting as depositional traps in which relatively dilated sand-rich facies accumulated during Lower to Middle Albian. Besides these depositional centers, condensed phosphate-rich deposits were sedimented (see Text-Fig. 3).

The depositional sequences within the deposits of the Platé Member are not easy to detect because of the very low sedimentation rate (<1 m/M.Y. and usually <20 cm/M.Y.). For instance, in some hypercondensed successions, the whole Albian (i. e. more than 10 M.Y.) is represented by no more than 1 m of phosphatic deposits. But a careful study allowed to detect several stratigraphical levels containing different faunas (Text-Fig. 3).

Nevertheless, the deposits of the Platé Member represent a transgressive stratigraphical unit because of their overall tendencies to deepen upward. This vertical evolution is marked by:

- a clastic fining-upward trend,
- a facies change from silicoclastic-dominated sedimentation to foraminiferal limestones,

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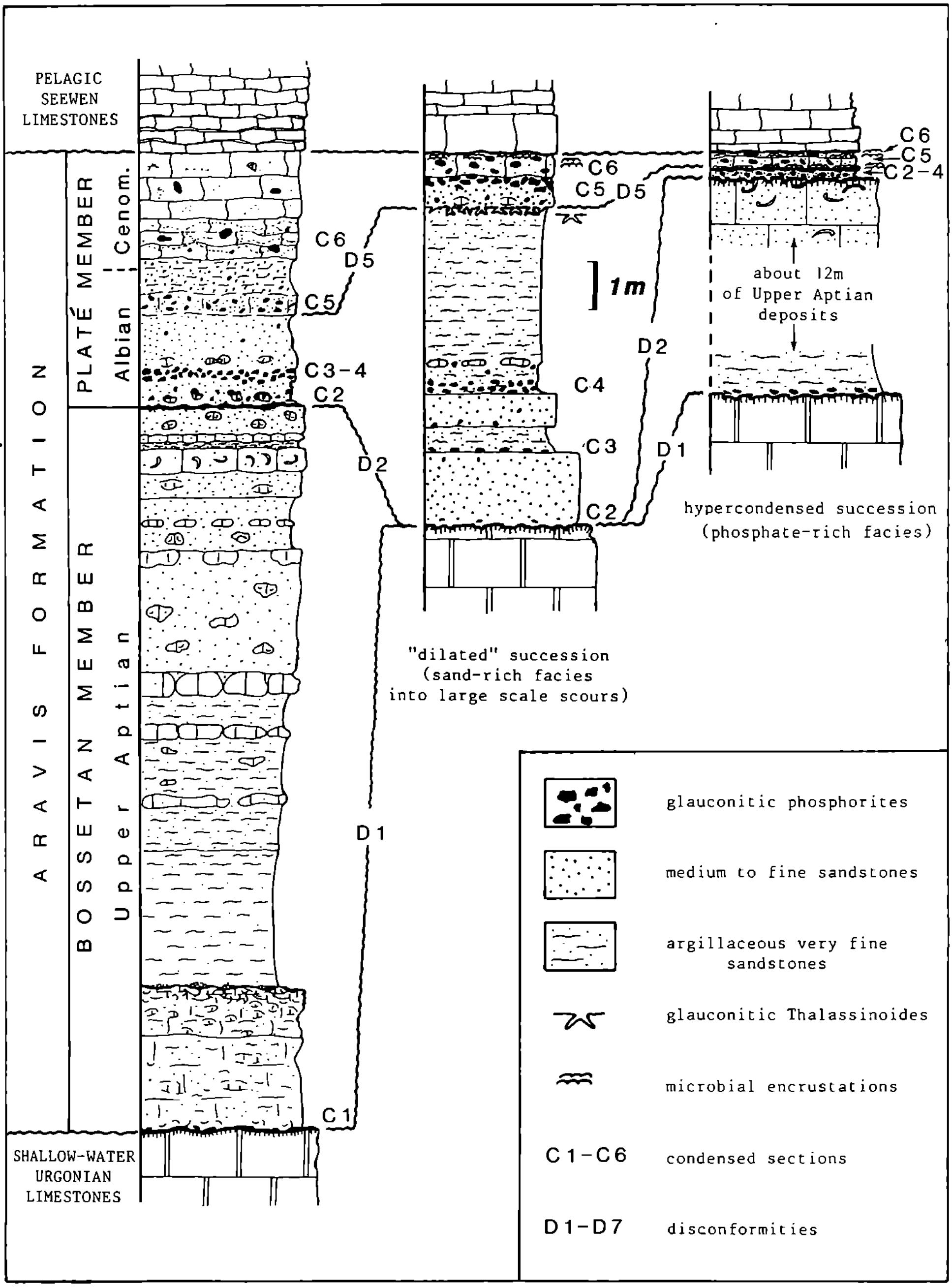
Text-Fig. 3. Three selected logs of the mid-Cretaceous series on the southern part of the Helvetic Shelf (for age of the condensed sections and the disconformities, see Text-Fig. 4).


PELAGIC SEEWEN LIMESTONES
ARAVIS FORMATION
PLATÉ MEMBER
Albian
Cenom.
BOSSETAN MEMBER
Upper Aptian
SHALLOW-WATER URGONIAN LIMESTONES
C1
C2
C3-4
C5
C6
D1
D2
D5
C3
C4
1m
"dilated" succession (sand-rich facies into large scale scours)
C2-4
about 12m of Upper Aptian deposits
hypercondensed succession (phosphate-rich facies)
glauconitic phosphorites
medium to fine sandstones
argillaceous very fine sandstones
glauconitic Thalassinoides
microbial encrustations
C1-C6 condensed sections
D1-D7 disconformities

- a faunal change from benthic faunas dominating the associations to pelagic associations composed of ammonites and planktonic foraminifera.

This deepening-upward evolution was not uniform and continuous. The study of more than 150 outcrops allowed to detect at least five disconformities (D 1 to D 5, Text-Fig. 3) within the deposits of the Platé Member. These disconformities are marked by one or more of the following features:

- extensive highly burrowed horizons,
- abrupt changes in facies or in the granulometric composition,
- intraformational conglomerate,
- extensive microbial encrustations.

Associated with these disconformities, condensed sections (C 2 to C 6) occur as fossiliferous conglomerates rich in ammonites often with faunal mixing. In accordance with VAIL et al. (1984) and LOUTIT et al. (in press), we used the disconformities and the condensed sections to subdivide the Albian deposits of the Platé Member into four depositional sequences (S 2 to S 5, Text-Fig. 3), each related to relative variation of the sea-level.

4. Correlation between the Helvetic Shelf deposits and the Vocontian Basin deposits

Owing to new biostratigraphical studies in the mid-Cretaceous deposits of the Helvetic shelf (DELAMETTE 1986) and the Vocontian Basin (DELAMETTE & BRÉHÉRET, work in progress), it is possible to present a tentative correlation between shelf phosphatic deposits and basinal black shales (Text-Fig. 4).

When looking at Text-Fig. 4, the following remarks can be expressed. First, good correlations exist between periods of phosphatic accumulation detected in the Helvetic Shelf and periods of organic carbon accumulation in the Vocontian Basin. It is best illustrated by the "Paquier Event" (early Lower Albian) which corresponds to the strongest phosphatic accumulation on the Helvetic Shelf (formation of thick phosphatic crusts, up to 15 cm, see Pl. 2, Fig. 4) and the strongest organic carbon accumulation in the Vocontian Basin.

Second, the accumulations of siliciclastic deposits on the Helvetic Shelf correspond to a period with no (or little) accumulation of organic matter in the Vocontian basin. This antagonism is well illustrated during the two following periods: upper part of the Upper Aptian and in the interval between the Middle Albian and the mid Upper Albian.

Third, all the periods of phosphatic accumulation on the Helvetic Shelf do not have a counterpart of organic matter accumulation in the Vocontian Basin. Thus, the widespread phosphorite accumulation known during the beginning of the Upper Albian (Inflatum to mid-Dispar Zones) cannot be correlated with a salient accumulation of organic matter in the Vocontian area. At that period, there was in this last named area an increasing siliciclastic influx leading to the deposition of silty marls (see the quartz abundance on Text-Fig. 2). This "silty interval" commenced near the limit

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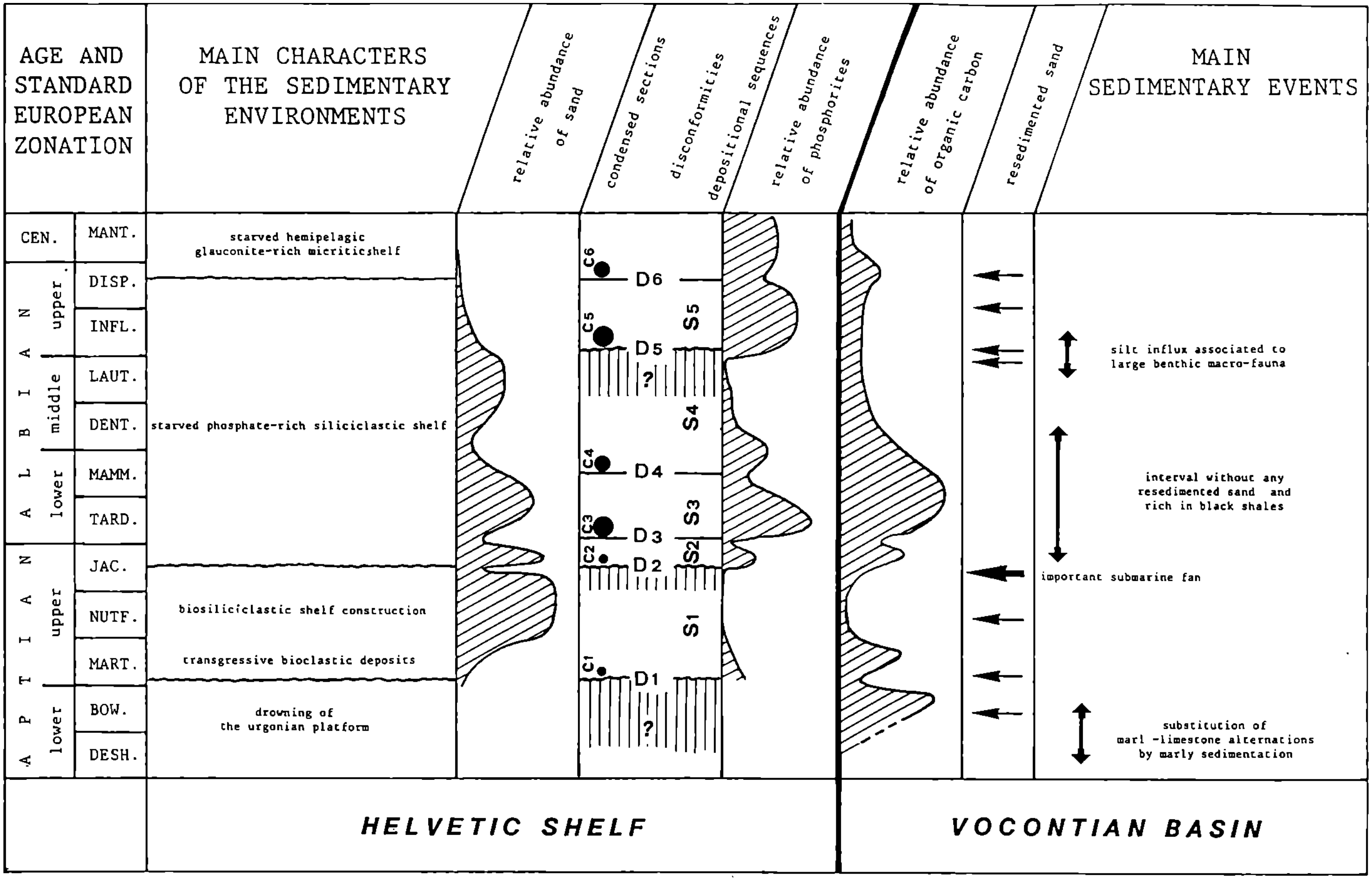
Text-Fig. 4. Correlations between the phosphate-rich facies of the Helvetic Shelf and the organic carbon-rich facies of the Vocontian Basin.


AGE AND STANDARD EUROPEAN ZONATION
MAIN CHARACTERS OF THE SEDIMENTARY ENVIRONMENTS
relative abundance of sand
condensed sections
disconformities
depositional sequences
relative abundance of phosphorites
relative abundance of organic carbon
resedimented sand
MAIN SEDIMENTARY EVENTS
CEN.
MANT.
ALBIAN
upper
DISP.
INFL.
middle
LAUT.
DENT.
lower
MAMM.
TARD.
APTIAN
upper
JAC.
NUTF.
MART.
lower
BOW.
DESH.
starved hemipelagic glauconite-rich micriticshelf
starved phosphate-rich siliciclastic shelf
biosiliciclastic shelf construction
transgressive bioclastic deposits
drowning of the urgonian platform
C6
D6
C5
D5
S5
?
S4
C4
D4
C3
D3
S3
C2
D2
S2
S1
C1
D1
?
silt influx associated to large benthic macro-fauna
interval without any resedimented sand and rich in black shales
important submarine fan
substitution of marl -limestone alternations by marly sedimentation
HELVETIC SHELF
VOCONTIAN BASIN

between Middle Albian and Upper Albian and continued until the end of early Upper Albian. This detritic influx does accompany an important change in the faunal association leading to the development of a macrobenthic fauna composed of large bivalves (mainly inoceramids such as *Birostrina concentrica* and *B. sulcata*). This fauna is unusual in the Vocontian Basin and involves particular conditions not yet understood very well (important decrease of the depth, adaptation to hypoxic conditions?).

5. Conclusions

In the Helvetic Shelf, the identification of condensed sections together with the disconformities leads to subdivide the deposits into depositional sequences related to relative variations in the sea-level.

With the help of biostratigraphical data, correlations were made between the basinal black shales of the Vocontian area and the phosphorites of the Helvetic Shelf. These correlations suggest that most of the phosphate on the shelf was accumulated contemporaneously with the organic matter in the basin. Thus, the three main organic carbon-rich intervals detected in the Vocontian (i. e. late Lower Aptian "Goguel Event", early Lower Albian "Paquier Event", and late Upper Albian "Breistroffer Event") can be correlated to the transgressive system tracts detected on the Helvetic Shelf.

The Goguel Event corresponds to the drowning of the Urgonian carbonate platform and to the development of a condensed section bearing phosphate (Aujon Beds).

The Paquier Event (BRÉHÉRET 1983, 1985, DELAMETTE et al. 1986) occurred during an important transgressive pulse documented on the shelves by the development of a widespread phosphate-rich condensed section covering the late Upper Aptian disconformity (D 2). In the Vocontian area, this transgressive pulse followed an important sea-level fall marked by a submarine fan (RUBINO 1984, FRIES 1986). This deep-water sand body is covered by black marls with numerous laminated organic-rich horizons. This late Upper Aptian-early Lower Albian transgressive pulse was also marked by an increasing faunal exchange between the European province (e. g. Anglo-Paris Basin) and the Mediterranean province. This exchange corresponded to a southward migration of the ammonite genus *Leymeriella* which is particularly abundant in the black shales of the Paquier Level (Pl. 1, Fig. 3).

The Breistroffer Event is also associated with an important transgressive pulse marked in the Helvetic area by the substitution of a neritic siliciclastic sedimentation by a hemipelagic micritic sedimentation. During this last event, hoplitinid ammonites (*Hyphoplites*) are represented in the Vocontian Basin for the first time.

The genesis of organic-rich beds is confined to periodic anoxia in the basin as a result of density stratification of waters. Indeed, an upwelling system (and a subsequent oxygen minimum zone development, cf. JENKYNS 1980) fails to explain the widespread occurrence of such a facies. The main anoxic events are effectively recognized in other countries such as the Umbrian-Marchean domain (Italy) (COCCIONI et al. in press) that was located on the southern margin of the Alpine-Tethys sea by mid-Cretaceous time. A genetic hypothesis has been proposed (DELAMETTE et al. 1986a) to explain the water mass stratification for the Paquier Event, i. e. slightly more saline waters of mid-latitude shallow epeiric seas (e. g.

Anglo-Paris basin), generated under a warm equable climate penetrated into the Vocontian Basin as into the Alpine-Tethys sea. They caused a restriction of vertical water-mass exchanges, thus inducing a stratification of the water column. However, such a model cannot be applied to all organic-rich events, since each episode displays peculiar features (particularly in faunal content) that suggest the diversity of genetic conditions. Phosphorite occurrences are widespread too; they are known from Betic-Cordillera to the Western Carpathians (DELAMETTE 1988). The present study shows their relation with anoxic episodes as a result of transgressive pulses. The starvations induced by these oceanographic conditions on the shelves (as on pelagic swells) lead to the development of microbial structures and favour phosphatic mineralization. In the same way the terrigenous input in the basin is not sufficient enough to hinder the development of early diagenetic processes, neoformations and/or transformations as is the case for smectites.

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Upper Cretaceous *Didymotis* Events from Bohemia

STANISLAV ČECH, Praha

With 1 Plate and 4 Text-Figures

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Abstract: In the Bohemian Cretaceous Basin (Czechoslovakia), the bivalve species *Didymotis costatus* (FRIČ 1893) is widely distributed, especially in the Teplice Formation where two peaks of occurrences were recognized. Both maxima of *Didymotis* may be considered as bioevents and are useful for intrabasinal and international correlations. *Didymotis* events and problems concerning the Turonian-Coniacian boundary are discussed.

Kurzfassung: Im Böhmischen Kreidebecken ist die Bivalven-Art *Didymotis costatus* (FRIČ 1893) weit verbreitet. Vor allem in den Teplice-Schichten sind zwei Kulminationspunkte der Verbreitung erkennbar. Beide Maxima können als wichtige Leitniveaus für regionale und überregionale Korrelation angesehen werden. In der Arbeit werden die Verbreitung der Gattung *Didymotis* und die damit verbundenen Probleme der Turon-Coniac-Grenze diskutiert.

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1. Introduction

During the last decade more attention has been paid to the cosmopolitan *Didymotis* (Bivalvia, Posidoniidae) and its occurrence near the Turonian-Coniacian boundary (KAUFFMAN in HERM et al. 1979, WOOD et al. 1984). ERNST et al. (1983) recognized in northwestern Germany two *Didymotis* ecoevents.

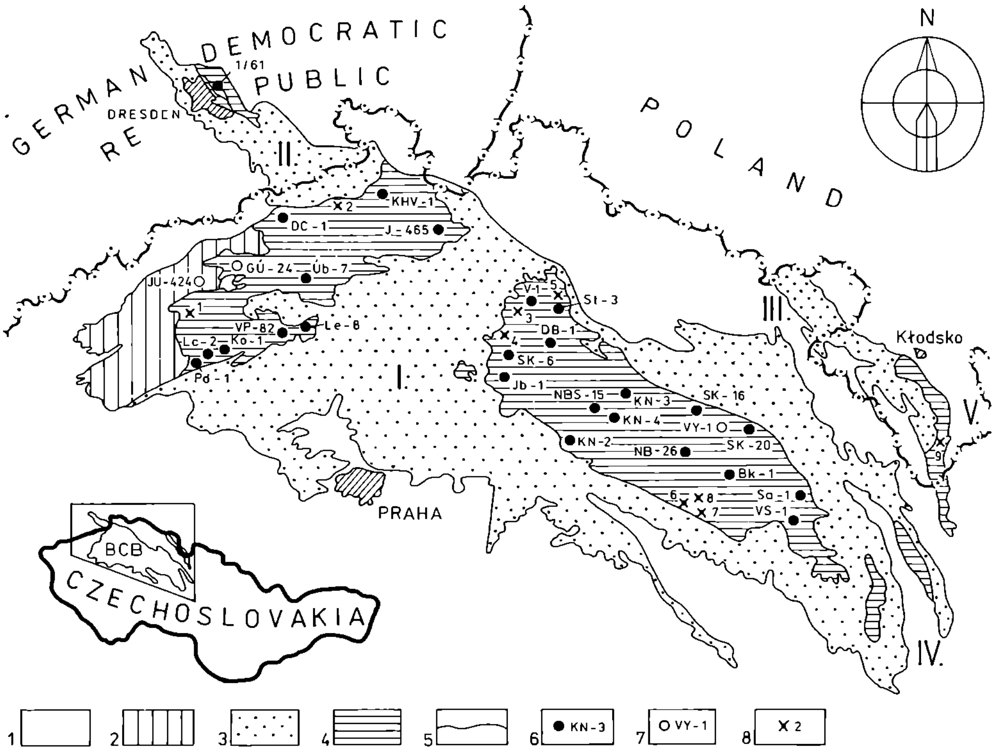
The genus *Didymotis* was described from the Bohemian Cretaceous Basin (BCB) as *Inoceramus planus* by REUSS (1846) and by FRIČ (1893), as *Inoceramus planus* var. *costata* by FRIČ (1893, textfig. 123), and as *Inoceramus costatus* and *I. costatus fritschi* by MACÁK (1967a). Most recently this species was described from BCB by ZÁZVORKA (1979) as *Sphenoceramus costatus*. The true systematic relevance of this false *Inoceramus* was recognized only recently.

This paper represents a complex synthesis of the biostratigraphical research on the Turonian - Coniacian boundary in the Bohemian Cretaceous Basin during the last two decades and also represents a contribution to the Project no, 216 of "Global Bioevents".

2. *Didymotis* events in the Bohemian Cretaceous Basin

Old and new collections from surface outcrops and from boreholes show that *Didymotis* always occurs in marly or clayey and chalk-marly facies of the Teplice and Březno formations (sensu ČECH et al. 1980) in the BCB (Text-Fig. 1). These facies are developed in the two areas: in the central and in the NW part of the BCB, including Upper Cretaceous deposits in Saxony.

Thanks to the boreholes drilled during the last two decades in the BCB it was documented that first occurrences of the species *Didymotis costatus*



Text-Fig. 1. Map of the horizontal distribution of *Didymotis costatus* (FRIČ) in the BCB.
1 - Pre-Cretaceous bedrock rocks, 2 - Cretaceous covered by younger Tertiary deposits, 3 - Peruc-Korycany to Jizera formation (Cenomanian - Upper Turonian), 4 - Teplice and Březno formations (Upper Turonian - Santonian) in clayey and chalk marly facies, 5 - limits of the Cretaceous basins, 6 - occurrences of *Didymotis costatus* FRIČ in boreholes, 7 - other boreholes, 8 - occurrences of *Didymotis costatus* in surface outcrops, BCB - Bohemian Cretaceous Basin, I - Bohemian Cretaceous Basin, II - Saxonian part of the BCB, III - Police Basin, IV - Vysoké Mýto and Orlice-Ustí synclines, V - Nysa-Graben.

(FRIČ) fall into the middle part of the Teplice Formation where two distinct maxima of occurrence may be distinguished - a lower and an upper one. Both levels (or acmes) are separated by 3 to 10 m of sediments in different boreholes (e. g. KN-2 Dlouhopolsko 3 m, SK-6 Plazy 3.8 m, Sa-1 Šachov 4.0 m, V-8 Vršovice 4.5 m, Ub-7 Rýdeč 5.6 m, J-465 Brniště 5.8 m, SK-16 Sadová 6.4 m, Jb-1 Jabkenice 6.5 m, Pd-1 Březno 10.2 m and Lc-2 Vršovice 10.8 m).

Both peaks of *Didymotis* occurrence, designated as *Didymotis* event I and II by WOOD et al. (1984), are characterized by mass occurrences of *Didymotis costatus* including articulated (15 % closed and 20 % open shells) and disarticulated shells (65 %) of normal size (average height 1.57 cm, n = 75). Shells of *Didymotis* are accumulated in only a few centimetre thick levels in both acme-horizons. This may be the reason why *Didymotis* was not discovered in many other boreholes, due to lower degree of paleontological documentation. In some older boreholes (e. g. Ko-1 Koštice, Bk-1 Borek ...) only one level of *Didymotis* was recognized for the same reason.

The larger specimens of *Didymotis costatus* (average 3.0 cm in height) occur sparsely above *Didymotis* events I and II up to the Rohatce Member (e. g. Nemošická stráň, JAHN 1896 and borehole Be-2 Březno, depth 10.8 m) and the lower part of the Březno (Priesen) Formation (e. g. Březno Formation type locality, bed 3 of FRIČ 1893, textfig. 122 and Lukovna-Počáply, JAHN 1905).

In most cases a distinct change in the composition of foraminiferal assemblages is within or little below the *Didymotis* maximum I, mainly in the central part of the BCB. At this level some foraminifera became extinct (e. g. "*Globotruncana renzi* GANDOLFI" and "*Globotruncana stephani* GANDOLFI", both sensu HERCOGOVÁ) and some appear for the first time in the BCB (such as *Praebulimina hofkeri* (BROTZEN), *Arenobulimina subsphaerica* (REUSS), *Gaudryina laevigata* FRANKE and "*Globotruncana cretacea* D'ORBIGNY" sensu HERCOGOVÁ). This change of assemblages is interpreted by HERCOGOVÁ (1967a, b, 1976b) as Middle-Upper Turonian boundary (see boreholes KN-2 Dlouhopolsko, KN-4 Hlušice and Pd-1 Březno).

In general, contemporaneous with *Didymotis* event II are first occurrences of *Cremnoceramus waltersdorfensis* (ANDERT) and *C. w. hannovrensis* (HEINZ) (e. g. in boreholes above *Didymotis* event II KHV-1 Kytlice 0.3 m, Ub-7 Rýdeč 0.6 m, SK-16 Sadová 1.8 m, Pd-1 Březno 1.9 m, KN-4 Hlušice 3.2 m, KN-2 Dlouhopolsko and J-465 Brniště 3.5 m, Sa-1 Šachov 5.3 m, DC-1 Malé Chvojno 5.7 m, KN-3 Chotělice 6.2 m, Jb-1 Jabkenice 6.5 m). At the *Didymotis* maximum II a minor acme of *Placenticeras orbignyanum* (GEINITZ) was observed in boreholes: Nová Ves-Komárov V-1 by SOUKUP (1963: 102), DB-1 Dolní Bousov (PRAŽÁK 1984, also together with "*I.*" *waltersdorfensis*) and SK-16 Sadová.

Both maxima of *Didymotis* may be considered as bioevents since they are widely distributed throughout the BCB (Text-Fig. 1) and represent only a very short time span (Text-Figs. 2-3). Regional geologic data and paleoecological interpretation of paleobathymetry (previously based on patterns of diversity of fauna and flora, see POKORNÝ 1971, ČECH et al. 1987) indicate a deepening of the basin during sedimentation of the Teplice Formation and show that the *Didymotis* events were associated with the maximum of transgression of the Cretaceous sea over the Bohemian Massif.

Biostratigraphic control shows that each *Didymotis* event is isochronous throughout the BCB. For this reason, *Didymotis* events can be used as reference levels for practical intrabasinal correlation. For example, in transverse sections across the BCB (Text-Figs. 2-3), *Didymotis* events document heterochroneity of the bed Xa (usually called "Coprolite layer" or "Glauconitic Contact layer"), which forms a boundary between Jizera and Teplice formations. This or these time transgressive horizon(s) (omission surface) reach the *Didymotis* events in marginal parts of the sedimentation area of the Teplice and Březno formations (e. g. boreholes J-465 Brniště, KN-2 Dlouhopolsko, Jb-1 Jabkenice and in sections in Vysoké Mýto and Orlice-Ústí synclines, see Text-Figs. 1, 3). Diachronous character, hiatus or condensation at these horizons was assumed earlier by SOUKUP (1949, 1955), MACÁK (1963) and recently by ČECH et al. (1987).

3. Occurrences of *Didymotis* outside of Czechoslovakia

In the Saxonian part of the BCB, *Didymotis* has been also found at two levels (depth 113.5 and 116.4 m) in the borehole 1/61 Dresden-Marienhof. According to K. A. TRÖGER (Bergakademie Freiberg, pers. comm. 1980), *Cremnoceramus waltersdorfensis* (ANDERT) - *C. rotundatus* (FIEGE) - *Mytiloides lusatiae* (ANDERT) appear for the first time 3.5 m higher than the latest occurrence of *Didymotis* (Text-Fig. 1).

Mass occurrences of the species *Didymotis costatus* in the BCB may be compared with two occurrences of *Didymotis* recognized as two ecoevents in the Salzgitter-Salder Quarry (Lower Saxony, NW Germany) by ERNST et al. (1983) and by WOOD et al. (1984). Rock thickness between *Didymotis* event I and II is 7 m in the Salzgitter-Salder Quarry and is comparable with thicknesses in the BCB. Nevertheless, these two occurrences of *Didymotis* in NW Germany cannot be interpreted as a total range zone (ERNST et al. 1983) since the species of *Didymotis costatus* occurs in the BCB also sparcely in younger sediments of the Rohatce Member and in lower parts of the Březno Formation (in both, the C. inconstans and C. deformis zones). In rocks equivalent to the Rohatce Member, *Didymotis* was recorded in Poland in the Nysa Graben at Wilkanow (RADWANSKA 1962, for location see Text-Fig. 1) and described under the name *Pecten britannicus* WOODS. *Didymotis uermoesensis* (SIMIONESCU) probably conspecific with *D. costatus* (FRIČ 1893) is known from Romania (SIMIONESCU 1899). The *Didymotis/Inoceramus rotundatus* event was also recognized by WOOD et al. (1984) in NE England. SUMMESBERGER (1985) noted *Didymotis* sp. together with true *Barroisiceras haberfellneri* HAUER from the Gosau Group in Austria. ERNST & KÜCHLER (this vol.) discussed their most recent observation of similar occurrence of *Didymotis* in northern Spain.

Other occurrences of the genus *Didymotis* are known from Japan (MATSUMOTO 1984), North and South America documenting its cosmopolitan distribution. Occurrences of *Didymotis* in the Western Interior of the United States (KAUFFMAN in HERM et al. 1979) associated with Lower Coniacian ammonites (*Forresteria* and *Barroisiceras*) and inoceramids (*C. deformis* and *C. erectus*) are in higher position than *Didymotis* events I and II in the BCB and in NW Germany (see also ERNST et al. 1983).

4. *Didymotis* events and the Turonian - Coniacian boundary

4.1 Ammonites

According to the concept of the Turonian - Coniacian boundary proposed at the Colloquium on the Turonian in Paris 1981 (ROBASZYNSKI 1983), and at the Symposium on Cretaceous Stage Boundaries in Copenhagen 1983 (BIRKELUND et al. 1984), the base of the Coniacian is placed on at the first occurrence of the ammonite *Forresteria (Harleites) petrocoriensis* (COQUAND) (= *Barroisiceras haberfellneri* HAUER, sensu DE GROSSOUVRE). Consequently, according to KENNEDY (1984), species of the genus *Peroniceras* do not indicate the base of the Coniacian because they appear above *Forresteria (H.) petrocoriensis.* For several reasons, however, the definition of the Forresteria (H.) petrocoriensis (Total Range) Zone in the BCB is somewhat difficult.

The first occurrence of *Forresteria (H.) petrocoriensis* (?) was reported by FRIČ (1893: 27, sub *Ammonites dentatocarinatus* from Vršovice-Černodoly, a type locality of the species *Didymotis costatus* FRIČ 1893, text-fig. 123), from an unknown horizon, but somewhere from the member Xc in the middle part of the Teplice Formation (see MACÁK 1967b, and VÁNĚ 1979).

Rare occurrences of *Forresteria (H.) petrocoriensis* (?) are known from the lower part of the Rohatce Member (Xd), in the NW part of the BCB, from Vinice hill near Koštice (FRIČ 1893: 28, sub *Amm. dentatocarinatus*), in the central part of the BCB from Oškobrch Hill (SOUKUP 1949: 722, sub *Barroisiceras haberfellneri*) and from Činěves (SOUKUP 1955: 643, for location see borehole KN-2).

It seems that the genus *Peroniceras* appears rather higher in section than *Forresteria. Peroniceras subtricarinatum* (D'ORBIGNY) is well known from the upper part of the Rohatce Member from Keblice (FRITSCH & SCHLOENBACH 1872, pl. 1, figs. 1-3) in the NW part of the BCB (see borehole Ub-7 Rýdeč on Text-Fig. 2).

Peroniceras subtricarinatum occurs also in the lower part of the Březno (Priesen) Formation type locality, beds 3-4 of FRIČ (1893, text-fig. 48), together with *Peroniceras (Zuluiceras) bajuvaricum* (REDTENBACHER) (JAHN 1896, text-figs. 1-2) and also with *Forresteria (H.) petrocoriensis* (FRITSCH & SCHLOENBACH 1872, pl. 14, fig. 3a, b; pl. 16, figs. 1-2; FRIČ 1893, text-fig. 51 and KENNEDY 1984, pl. 6, figs. 3-5; pl. 9, figs. 1-2).

On the other hand, an index Turonian ammonite *Subprionocyclus neptuni* (GEINITZ) has been described by GEINITZ (1872-75, pl. 36, fig. 4) from the Strehlener Schichten in Dresden (= lower part of the Teplice Formation, member Xb in the NW part of the BCB). Other occurrences referred to this species in the BCB need a systematic revision. Ammonites such as *Lewesiceras mantelli* WRIGHT & WRIGHT, *Scaphites geinitzii* D'ORBIGNY and other "Chalk Rock" ammonites which characterize the Neptuni Zone are known also from the lower part of the Teplice Formation in the NW part of the basin, e. g. from Koštice, Čízkovice and Hudcov (Hundorf) (FRIČ 1889, HOUŠA 1967), from borehole Pd-1 Březno (MACÁK 1968), and in surrounding areas from upper parts of the Jizera Formation (FRIČ 1883, SOUKUP 1964, HOUŠA 1967).

Actinocamax strehlensis (FRITSCH & SCHLOENBACH), an index Turonian belemnite, occurs rarely in the lower part of the Teplice Formation, e. g. in Koštice (FRITSCH & SCHLOENBACH 1872, pl. 16, fig. 17a; FRIČ 1889, text-fig. 46), Strehlen in Dresden (FRITSCH & SCHLOENBACH 1872, pl. 16, figs. 10-12) in the NW part of the BCB, and in Lâny na Dûlku (SOUKUP 1949, p. 718) in the central part of the BCB (for location see outcrop no. 6, Srnojedy on Text-Fig. 1).

Rare occurrences of *S. neptuni* are overlain by *Prionocyclus germari* (REUSS), *Placenticeras orbignyanum* (GEINITZ) and by heteromorph ammonites in member Xc of the Teplice Formation. First occurrences of *P. germari* and *P. orbignyanum* together with *Lewesiceras lewesicense* (= *L. mantelli* according to WRIGHT 1979) are known from the "Gastropod horizon" in Lenešice Brickyard, at the base of the member Xc (FRITSCH & SCHLOENBACH 1872, pl. 10, fig. 4a, b; pl. 14, figs. 1-2; FRIČ 1893, text-figs. 50, 53; HOUŠA 1967, pl. 8, figs. 1-7; ZAHÁLKA 1938). The true stratigraphic position of this outcrop has been recognized by MACÁK (1967b).

Occurrences of *P. germari* represent a distinct zone in the lower part of the member Xc (e. g. 19.8 m below *Didymotis* event in borehole Ko-1 Koštice, HOUŠA 1965, 6.2 m in borehole Lc-2 Vršovice, 2.3 m in borehole Jb-1 Jabkenice) and reach the *Didymotis* event I and II or little above it in boreholes KN-2 Dlouhopolsko (SOUKUP 1967a) and Pd-1 Březno (MACÁK 1968). *Peroniceras tricarinatum* recorded from borehole Pd-1 Březno, at a depth of 38.6 m, i. e. between *Didymotis* event I and II by MACÁK (1968) and by VÁNĚ (1979) is *Prionocyclus* sp. (according to the determination of W. J. KENNEDY, Oxford University, pers. comm. 1987).

Prionocyclus germari, according to ammonite specialists, indicates Upper Turonian, while *Placenticeras orbignyanum* (= *P. fritschi* of some authors) characterizes the Middle Coniacian Peroniceras tridorsatum Zone (see KENNEDY 1984), but this species appears in the BCB earlier than in the type areas in France.

An overlap of *Forresteria (H.) petrocoriensis* with some species of *Peroniceras* is recognized in the BCB and is in contrast with the distribution of ammonites in France.

4.2 Inoceramids

Inoceramids are more abundant than ammonites in the BCB, and for this reason represent one of the important faunal groups for detailed biostratigraphy.

SEITZ's (1952) and TRÖGER's (1967) concept of the Turonian - Coniacian boundary was accepted by MACÁK & MÜLLER (1968). KAUFFMAN (1978) and SEIBERTZ (1979) documented that the Turonian - Coniacian boundary sensu SEITZ and TRÖGER is higher than elsewhere in the world. From that time on the boundary "shifts" year by year lower and lower, e. g. in NW Germany it is well illustrated in papers by TRÖGER (1981), KELLER (1982), ERNST et al. (1983) and WOOD et al. (1984). On the other hand, SEIBERTZ (1986) recently goes "back up" to the concept of TRÖGER (1981) (Text-Fig. 4).

According to KAUFFMANN (1978), the Turonian - Coniacian boundary in the BCB lies at or below the Inoceramus schloenbachi - I. inconstans Zone of KLEIN & SOUKUP (1966). However, SOUKUP (1955, 1959) and KLEIN & SOUKUP (1966) were incorrect to correlate this zone with the whole Teplice Formation (Xa, b, c). A more detailed study of many sections in the BCB has shown that *I. schloenbachi* BOEHM (= *C. deformis* in this paper) appears for the first time in the upper part of the Rohatce Member (Xd) where it co-occurs with *Peroniceras subtricarinatum,* and disappears in the lower part of the Březno Formation. *Cremnoceramus inconstans* is abundant near the base of the Rohatce Member and is associated with *C. inconstans woodsi* (FIEGE), *C. waltersdorfensis* (ANDERT) and *C. rotundatus* (FIEGE). While the lower limit of the C. inconstans Zone (Interval Zone) is somewhere indistinct, the base of the C. deformis Zone (Total Range Zone) is widely traceable throughout the BCB. In addition, the inoceramid zonation of KAUFFMAN (1978) of the BCB cannot be used for practical stratigraphy due to several mistakes in determination of species and incorrect interpretation of the stratigraphical position of some outcrops or boreholes.

For this paper, a provisional inoceramid zonation of the BCB (see Text-Figs. 2-4) may be used. This zonation is based on a study of all outcrops and boreholes mentioned in this paper.

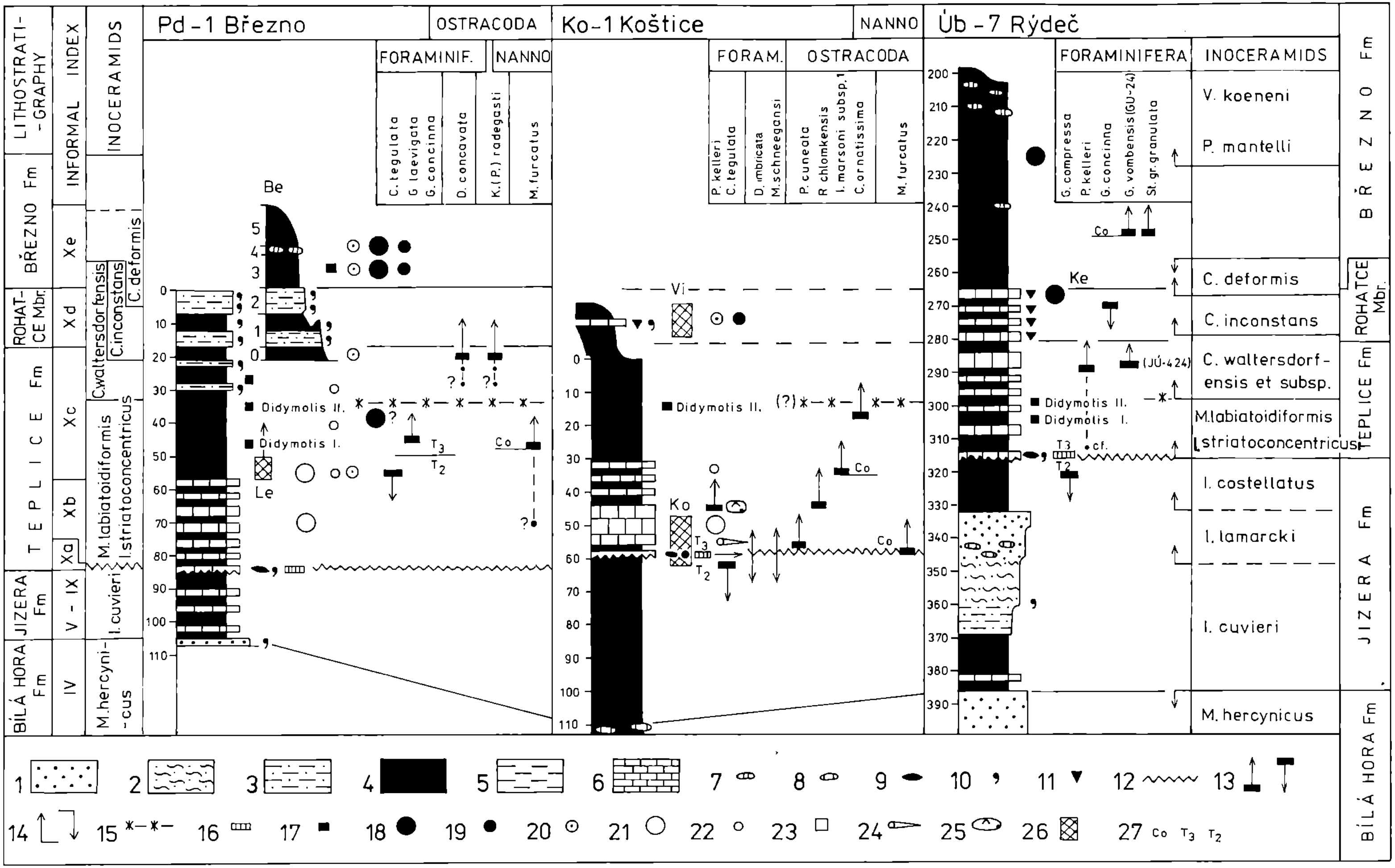
Together with Turonian index ammonites, inoceramids of the *Mytiloides* lineage such as *M. labiatoidiformis* (TRÖGER 1967, pl. 10, figs. 5-6), *M. labiatoidiformis* TRÖGER sensu KELLER (1982, pl. 5, fig. 5), *M. mytiloidiformis* (TRÖGER), *M. fiegei* (TRÖGER) (= *M. incertus* of some authors), *M. dresdensis* (TRÖGER) and inoceramids of the *Inoceramus* lineage with *I. striatoconcentricus* et subsp., *I. costellatus* WOODS, *I. frechi* FLEGEL, *I. inaequivalvis falcatus* (HEINZ) and fragments of prismatic layer of large *I. lamarcki stuemckei* (HEINZ) and *I. lamarcki cuvieri* sensu WOODS (1910-11, text-figs. 75 and 77) occur abundantly in the upper part of the Jizera Formation (except for the Ohře region in the vicinity of the boreholes Pd-1, Ko-1 and JU-424 where the upper part of the formation is missing) and in the lower part of the Teplice Formation.

There is only one important change in inoceramid evolution between the Rohatce Member and the lower part of the Teplice Formation. *Inoceramus-Mytiloides* dominated assemblages of inoceramids are replaced by the genus *Cremnoceramus* at the *Didymotis* event II in the middle part of the Teplice Formation. The first occurrence of *C. waltersdorfensis* (ANDERT) and *C. w. hannovrensis* (HEINZ) is little above the *Didymotis* event II as it has been shown above. The appearance of these subspecies is widely recognizable over the entire basin. *Cremnoceramus waltersdorfensis* et subspecies are abundant throughout the upper part of the Teplice Formation including the overlying Rohatce Member.

The appearance of *C. waltersdorfensis* et subsp. together with the second maximum of *Didymotis* was recommended for the base of the Coniacian and considered as more practical than with ammonites at the Symposium on the Cretaceous Stage Boundaries in Copenhagen 1983 (BIRKELUND et al. 1984).

I have some sympathy with this view, but there appears a little difficulty from this. *C. waltersdorfensis* et subsp. was found together with the Coniacian ammonite *Forresteria (H.) petrocoriensis* in SE England (BAILEY et al. 1984) and in Spain (WOOD et al. 1984), whereas in the Aube area in N France these inoceramids are noted by AMEDRO et al. (1982) together with *Collignoniceras woollgari* (MANTELL), *Subprionocyclus neptuni* and *Lewesiceras mantelli*. All these occurrences are associated with hardground complexes.

In Boulonnais, also in a hardground complex, ROBASZYNSKI et al. (1980) has regarded *C. waltersdorfensis* et subsp. and *C. schloenbachi* (op. cit. fig. 19) to belong to the Turonian Neptuni Zone which contrasts to the



LITHOSTRATIGRAPHY
INFORMAL INDEX
INOCERAMIDS
Pd-1 Březno
OSTRACODA
FORAMINIF.
NANNO
C. tegulata
G. laevigata
G. concinna
D. concavata
K.(P.) radegasti
M. furcatus
Ko-1 Koštice
NANNO
FORAM.
OSTRACODA
P. kelleri
C. tegulata
D. imbricata
M. schneegansi
P. cuneata
R. chlomkensis
I. marsoni subsp. 1
C. ornatissima
M. furcatus
Úb-7 Rýdeč
FORAMINIFERA
INOCERAMIDS
G. compressa
P. kelleri
G. concinna
G. vombensis (GU-24)
St. gr. granulata
BŘEZNO Fm
ROHATCE Mbr.
TEPLICE Fm
JIZERA Fm
BÍLÁ HORA Fm
Xe
Xd
Xc
Xb
Xa
V - IX
IV
C. deformis
C. inconstans
C. waltersdorfensis
M. labiatoidiformis
I. striatoconcentricus
I. cuvieri
M. hercynicus
V. koeneni
P. mantelli
C. waltersdorfensis et subsp.
I. costellatus
I. lamarcki
Be
Le
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Didymotis II.
Didymotis I.
Co
T_3
T_2
cf.
(JÚ-424)
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indications on figs. 16 and 18, since there is a gap in ammonite record between the *S. neptuni* and *P. tridorsatum* occurrences.

There are no index ammonites between the occurrences of *P. germari* and *F. H. petrocoriensis* (KAPLAN 1986, fig. 4; this paper Text-Fig. 4) even in areas with continuous sedimentation at the Turonian - Coniacian transition, such as NW Germany and the Bohemian Cretaceous Basin. *C. waltersdorfensis* first appears 15 m below *F. (H.) petrocoriensis* in NW Germany (SUMMESBERGER 1985) and about 17 m below in the BCB. KELLER (1982) considers *C. waltersdorfensis* as Upper Turonian in age, while WOOD et al. (1984) used the appearance of this inoceramid to define the base of the Coniacian.

4.3 Foraminifera

The distribution of planktonic foraminifera at the Turonian - Coniacian boundary provides the basis for discussion (see MARKS 1984), since stratotype sections do not contain diagnostic planktonic foraminifera. In the BCB, ŠTEMPROKOVÁ (in POKORNÝ et al. 1983) recorded from the Březno Formation type locality the occurrence of *Dicarinella concavata* (BROTZEN), widely accepted as Coniacian species; but none of *D. concavata* occurs in the lower part of the Teplice Formation in Koštice. Thus, the lower limit of this species could be somewhere within the member Xc.

Benthonic foraminifera are more useful for microbiostratigraphy of the BCB. The recent foraminiferal stratigraphic concept of the BCB (HERCOGOVÁ in KLEIN 1983, HERCOGOVÁ 1985) is derived from that of KOCH (1977). According to KOCH, the appearance of the subspecies *Stensioeina granulata granulata* OLBERTZ marks the base of the Coniacian. This subspecies has been found in the BCB in the uppermost part of the Rohatce Member, but more frequently in the overlying lower parts of the Březno Formation, e. g. in boreholes KN-4 Hlušice (HERCOGOVÁ 1967b, 1976b), VY-1 Všestary (HERCOGOVÁ 1965), RP-15 Bříza (HERCOGOVÁ 1986) and

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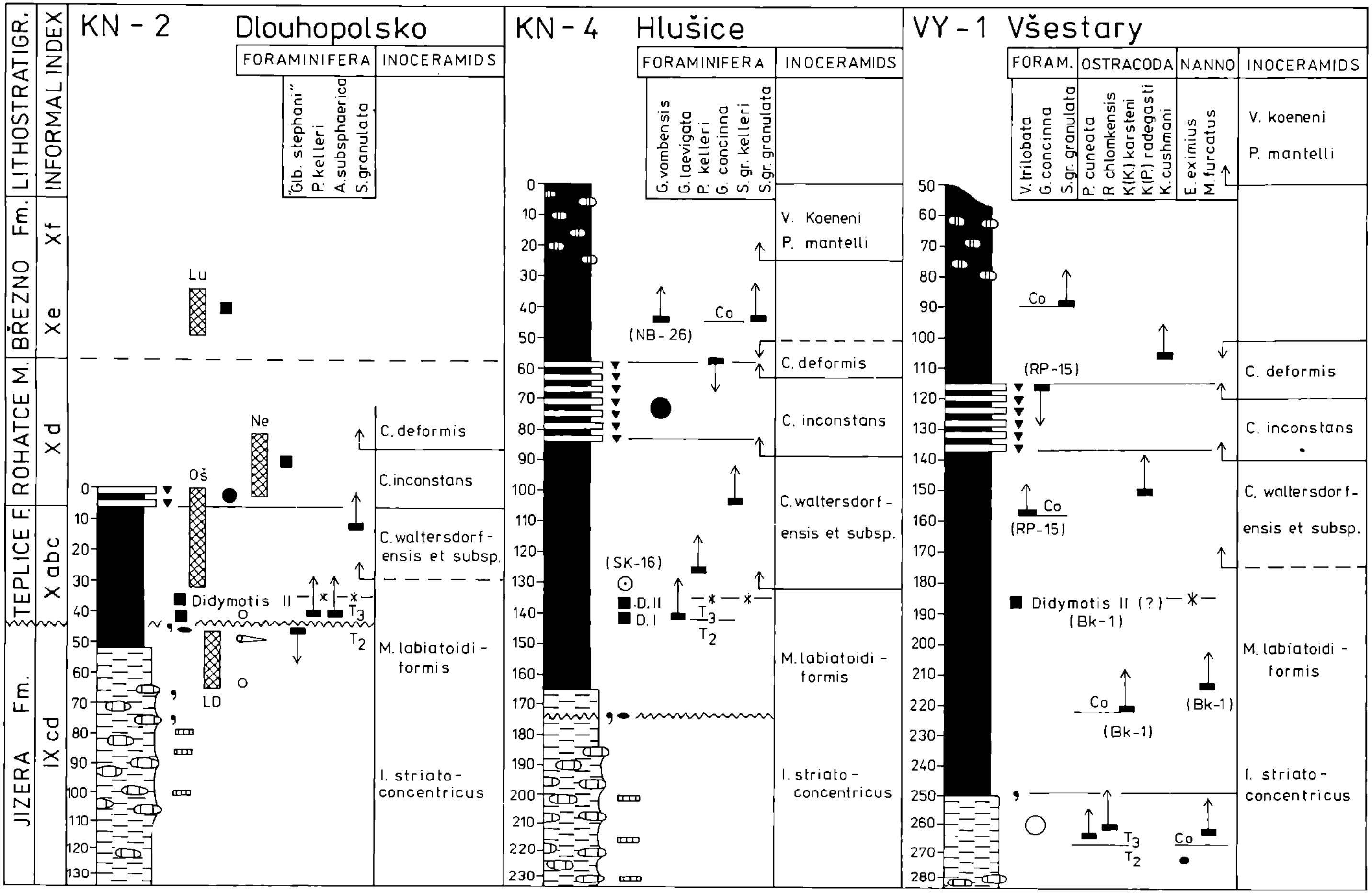
Text-Fig. 2. Lithology and vertical distribution of *Didymotis costatus* and selected guide fossils in the NW part of the BCB.
1 - quartzose sandstones, 2 - clayey sandstones, 3 - siltstones, 4 - claystones, 5 - marlstones, 6 - micritic limestones, 7 - pelosideritic nodules, 8 - carbonate nodules, 9 - phosphatic nodules, 10 - glaukonite (> 10 %), 11 - silicification, 12 - erosion, omission surface, 13 - first and last occurrence or local acme, 14 - first and last occurrence of inoceramids, 15 - reference level of the *Didymotis* event II, 16 - fragments of prismatic layer of large inoceramids, 17 - *Didymotis costatus*, 18 - *Peroniceras*, 19 - *Forresteria (H.) petrocoriensis*, 20 - *Placenticeras orbignyanum*, 21 - *Lewesiceras mantelli*, 22 - *Prionocyclus germari*, 23 - *Subprionocyclus neptuni*, 24 - *Actinocamax*, 25 - *Micraster leskei*, 26 - surface outcrops: Be - Březno (Priesen) Formation type locality, Ke - Keblice, Ko - Koštice, LD - Lány na Důlku, Le - "Lenešice Gastropode horizon", Lu - Lukovna-Počáply, Ne - Nemošická stráň, Oš - Oškobrh, Vi - Vinice hill, 27 - Co-Coniacian, T_3 - Upper Turonian, T_2 - Middle Turonian of micropaleontologists.


LITHOSTRATIGR.
INFORMAL INDEX
JIZERA Fm.
IX cd
TEPLICE F.
X abc
ROHATCE M.
X d
BŘEZNO Fm.
Xe
Xf
KN - 2 Dlouhopolsko
FORAMINIFERA
INOCERAMIDS
"Glb. stephani"
P. kelleri
A. subsphaerica
S. granulata
Lu
Ne
Oš
LD
Didymotis II
T3
T2
C. deformis
C. inconstans
C. waltersdorf-ensis et subsp.
M. labiatoidi-formis
I. striato-concentricus
KN - 4 Hlušice
G. vombensis
G. laevigata
P. kelleri
G. concinna
S. gr. kelleri
S. gr. granulata
V. Koeneni
P. mantelli
(NB-26)
Co
(SK-16)
D. II
D. I
VY - 1 Všestary
FORAM.
OSTRACODA
NANNO
V. trilobata
G. concinna
S. gr. granulata
P. cuneata
R. chlomkensis
K(K.) karsteni
K(P.) radegasti
K. cushmani
E. eximius
M. furcatus
V. koeneni
(RP-15)
Didymotis II (?)
(Bk-1)

Rýdeč Ub-7 (HERCOGOVÁ 1967c). At the same level a distinct change in foraminiferal assemblages appears (HERCOGOVÁ 1973, 1976b, 1985 and HERCOGOVÁ in KLEIN et al. 1982); there is also a second acme of *Pyramidina kelleri* (VASILENKO) while *Gaudryinella concinna* (REUSS) became extinct. This level has a high value for correlation within the BCB.

Nevertheless, KOCH used SEITZ's concept of the Turonian - Coniacian boundary (see chapter 4.2: Inoceramids), but at the present day the base of the Coniacian should be placed lower. There are several levels which may be considered in the BCB:

1) The first appearance of *Stensioeina granulata kelleri* KOCH sensu HERCOGOVÁ (HERCOGOVÁ in KLEIN et al. 1982, pl. 12, figs. 2-4); elsewhere *St. gr. levis* KOCH is recorded at this stratigraphic level. Rare occurrences of this subspecies were discovered by HERCOGOVÁ (1967a, b, 1976b) in the upper part of the Teplice Formation in the central part of the BCB, e. g. boreholes KN-2 Dlouhopolsko and KN-4 Hlušice (Text-Fig. 3). In the Anglo-Paris Basin, *St. gr. levis* was found little above *Forresteria (H.) petrocoriensis*, i. e. above the "Top Rock" (Navigation Hardground of BAILEY et al. 1984).

At or near this level *Gaudryina pyramidata* CUSHMAN and *Vaginulina trilobata* (D'ORBIGNY) occur abundantly and most recently HERCOGOVÁ (1986) regards this level as the base of the Coniacian, e. g. in borehole RP-15 Bříza in the central part of the BCB (for location and stratigraphic position, see borehole VY-1 Všestary on Text-Figs. 1 and 3).

2) At the change of foraminiferal assemblages recognized by HERCOGOVÁ (1973, 1976b, 1986) and by HERCOGOVÁ in KLEIN et al. (1982) in the central part of the BCB. This change near the *Didymotis* event I - see above - is considered by HERCOGOVÁ the Middle-Upper Turonian boundary. The position of the Middle - Upper Turonian boundary shows large discrepancies in different sections in terms of the foraminiferal stratigraphy in the NW part of the BCB. This change is situated at the *Didymotis* event I, e. g. in the borehole Pd-1 Březno between depth 45.0 and 47.0 m (HERCOGOVÁ revised for this paper in 1987) whereas elsewhere, e. g. in boreholes Ko-1 Koštice (HERCOGOVÁ 1967d), Ub-7 Rýdeč (HERCOGOVÁ 1967c) and GU-24 Lochočice (HERCOGOVÁ 1985), this change in foraminiferal assemblages occurs much lower at the bed Xa (Coprolite layer) at the base of the Teplice Formation (Text-Figs. 2, 4). The difference is about 40 m. Similar differences (up to 19.5 m) were documented by HERCOGOVÁ (1967d) in this area.

POMEROL (1985) mentioned *Gavelinella vombensis* (BROTZEN) and *Pyramidina kelleri* (VASILENKO) at the base of the Senonian in its stratotype section in N France.

In most cases, *G. vombensis* co-occurs with *St. gr. granulata* in the BCB, e. g. in boreholes GU-24 Lochočice (HERCOGOVÁ 1985) and NB-26 Žizkovec (HERCOGOVÁ 1976a, HERCOGOVA in KLEIN et al. 1982, figs. 1-3). Most recently *G. vombensis* has been discovered by HERCOGOVÁ (in

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Text-Fig. 3. Lithology and vertical distribution of the *Didymotis costatus* and selected guide fossils in the central part of the BCB. For explanation see Text-Fig. 2.

prep.) in much lower levels in the middle of the Teplice Formation in the borehole JU-424 Jenišův Újezd (Text-Figs. 1 and 2).

Pyramidina kelleri (HERCOGOVÁ in KLEIN et al. 1982, pl. 9, fig. 7) abundantly occurs little above the *Didymotis* event II (first acme of the *P. kelleri*), e. g. in the borehole KN-4 Hlušice. Yet, the first rare occurrence of the species was recorded by HERCOGOVÁ (1967a, 1973) at the *Didymotis* event I, e. g. in borehole KN-2 Dlouhopolsko. In the NW part of the BCB first sparse occurrences of *P. kelleri* are recorded from the upper part of the member Xb of the Teplice Formation, e. g. in Koštice (ŠTEMPROKOVÁ in POKORNÝ et al. 1983) and in borehole JU-421 Jenišův Újezd near JU-424 (HERCOGOVÁ in prep.). A first maximum occurrence of the species in this area appears approx. 25 m higher and above the *Didymotis* event II, e. g. borehole Ub-7 Rýdeč (HERCOGOVÁ 1967c, revised by HERCOGOVÁ in 1987).

BAILEY et al. (1984) documented an inconsistent appearance of *Pyramidina kelleri* and *Gavelinella vombensis* and their co-occurrence with Turonian ammonites of the Neptuni Zone in SE England.

4.4 Ostracoda

POKORNÝ (1978) established eight ostracod assemblage zones in the BCB. The scheme was based on a study of ostracod distribution in boreholes Ko-1 Koštice (POKORNÝ 1966a), Bk-1 Borek and Vy-1 Všestary (POKORNÝ 1966b).

According to POKORNÝ, the first appearance of species/subspecies *Karsteneis (Prosteneis) radegasti* POKORNÝ, *Karsteneis (Karsteneis) karsteni karsteni* (REUSS), *Imhotepia marssoni* subsp.-1 POKORNÝ, and *Cythereis ornatissima ornatissima* (REUSS) refers to the base of Coniacian (C_a ostracode assemblage Zone). The lower limit of this zone is placed by POKORNÝ to the member Xc of the Teplice Formation (Text-Figs. 2-4).

Karsteneis (P.) rad. radegasti appears above the *Didymotis* events and below the Rohatce Member at a comparable level with *Stensioeina* gr. "*kelleri*" in the central part of the BCB. In the NW part of the basin this subspecies is known from the Březno Formation type locality (POKORNÝ in POKORNÝ et al. 1983) and probably appears for the first time below the Rohatce Member.

Other above mentioned ostracode boundary markers occur below the *Didymotis* events near the base of the member Xc, at the "Lenešice Gastropod horizon", and belong to the Turonian ammonite or inoceramid zones. Some ostracodes, e. g. *Cythereis ornatissima, Phacorhabdotus semiplicatus* (REUSS) and *Krithe* n. sp. have quite different distributions in NW and central parts of the BCB (POKORNÝ 1966a, b, 1978).

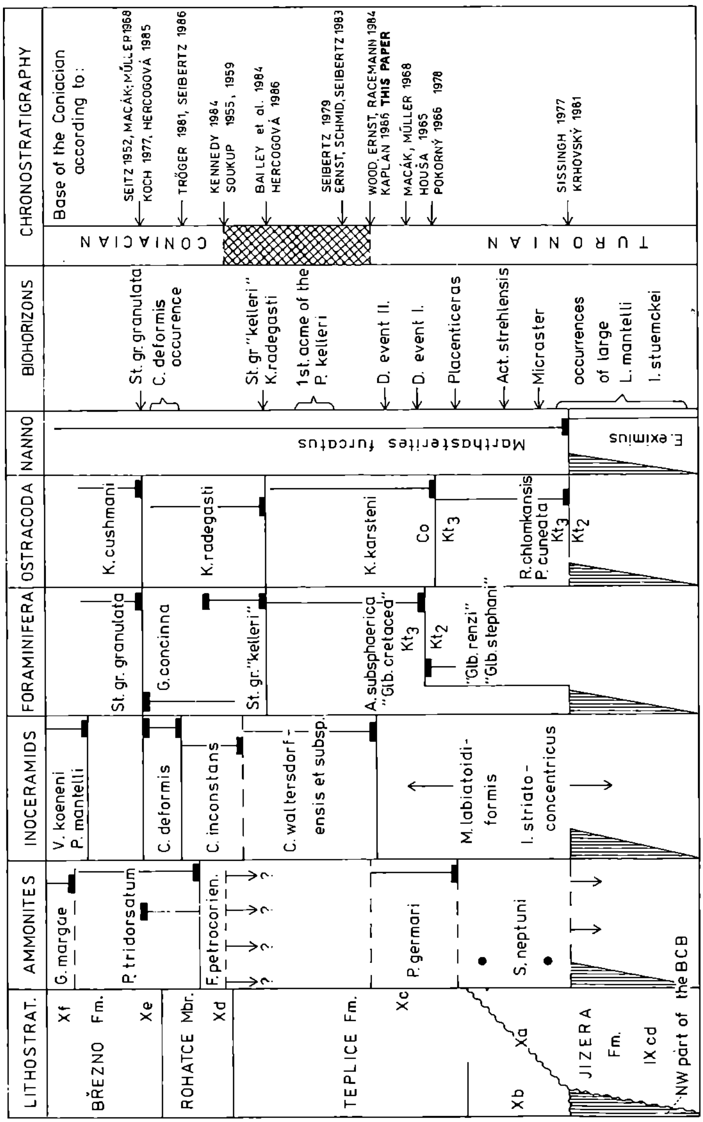
Somewhat below, *Planileberis cuneata* KAFKA and *Rehacythereis chlomkensis* POKORNÝ appear in a comparable biostratigraphic level in the lower parts of the Teplice Formation and in uppermost parts of the Jizera Formation in NW and central parts of the BCB, respectively, thus documenting a heterochronous boundary between both formations (Text-Fig. 4).

4.5 Calcareous nannoplankton

Recently, remarks concerning the distribution of calcareous nannoplankton at the Turonian - Coniacian transition in the BCB were made by KRHOVSKÝ

(1981) and ŠVÁBENICKÁ (1983). KRHOVSKÝ (1981) recognized the first occurrence of the "Coniacian key species" *Marthasterites furcatus* (DEFL.) DEFLANDRE in the lowermost part of the Teplice Formation (Xa, Coprolite layer) in several outcrops (including Koštice) in the NW part of the BCB. In contrast to this, ŠVÁBENICKÁ (1983) and ŠVÁBENICKÁ (in POKORNÝ et al. 1983) recorded the lack of *M. furcatus* in both the Coprolite layer and the member Xb of the Teplice Formation in the same area.

KRHOVSKÝ (1981) reports the occurrence of *M. furcatus* somewhere from the member Xc in Lenešice Brickyard. In 24 samples taken to study



Text-Fig. 4. Integrated biostratigraphy at the Turonian - Coniacian transition in the BCB.

the distribution of calcareous nannoplankton in borehole Pd-1 Březno (except sample from depth 71.0 m, probably contaminated) ŠVÁBENICKÁ (Central Geological Survey, Prague) prooved the first occurrence of *M. furcatus* at a depth of 44.6 m within the *Didymotis* event I (see Text-Fig. 2). A similar distribution of *M. furcatus* was observed by ŠVÁBENICKÁ (pers. comm. 1987) in the borehole SK-16 Sadová at depth 106.0 m (*Didymotis* event I is here at depth 100.5 m).

The first occurrence of *M. furcatus* associated with Turonian macro- and microfaunas was recognized by KRHOVSKÝ (1981) and by ŠVÁBENICKÁ (1984 and pers. comm. 1987) considerably below the *Didymotis* events in the lower part of the Teplice Formation or even in the uppermost part of the Jizera Formation, e. g. boreholes Bk-1 Borek at depth 195.0 m, VY-1 Všestary at depth 262.5 m, DB-1 Dolní Bousov at depth 135.5 m and VS-1 Týništko at depth 118.4 m.

For this reason, *M. furcatus* cannot be used as a marker for the Turonian - Coniacian boundary in the BCB. The same conclusion was made in SE England (BAILEY et al. 1984) and in NW Germany (WOOD et al. 1984).

5. Conclusions

Two *Didymotis* events are widely distributed throughout the BCB and are comparable with those in NW Germany and in N Spain.

Text-Fig. 4 summarizes the data on the distribution of some selected macro-, microfaunas and microfloras in the BCB near the Turonian - Coniacian transition. There are several difficulties to define the boundary between Turonian and Coniacian stages. At first, the comparison with the type areas in France is difficult because of discontinuous sedimentation and unfavourable preservation of the fossils camouflaging the faunal and floral successions. Subsequently, because of anarchy in the boundary concept. For ex-

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Plate 1

Specimens were slightly coated with ammonium chloride before photography.

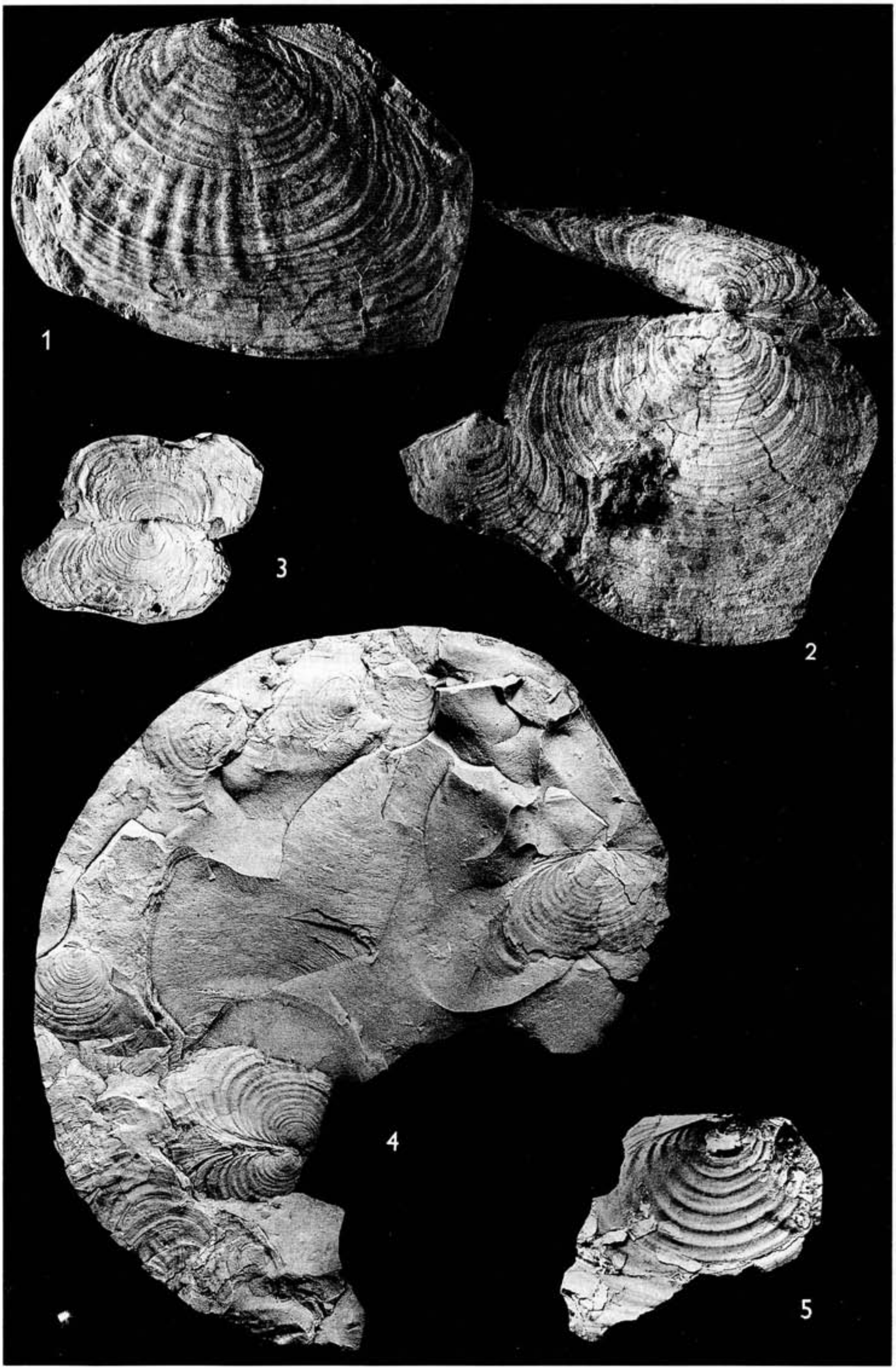
Fig. 1. *Didymotis costatus* (FRIČ), lateral view of left valve with radial ribs, holotype (FRIČ 1893, text-fig. 123), Vršovice-Černodoly, Teplice Fm., National Museum, Praha no. O 4289, x 2.

Fig. 2. *Didymotis costatus* (FRIČ), lateral view of right valve, original of FRIČ 1893, text-fig. 122, Březno, bed no. 3, lower part of the Březno Formation, National Museum, Praha, no. O 4288, x 1.6.

Fig. 3. *Didymotis costatus* (FRIČ), dorsal view, *Didymotis* event II, borehole KN-3 Chotělice, depth 91.3 m, Teplice Formation, x 1.

Fig. 4. *Didymotis costatus* (FRIČ), mass occurrence of the species within *Didymotis* event II, borehole KN-3 Chotělice, depth 91.3 m, Teplice Formation, x 1.

Fig. 5. *Didymotis costatus* (FRIČ), lateral view of right valve, *Didymotis* event I, borehole Lc-2 Vršovice, depth 20.6 m, x 1.5.


1
2
3
4
5

ample, inoceramid specialists (e. g. SEITZ 1952) and calcareous nannoplankton specialists (e. g. SISSINGH 1977) differ in regard of the Turonian-Coniacian boundary by a sequence of 175 m in the BCB (see borehole Vy-1 Všestary on Text-Fig. 3).

In terms of ammonite orthostratigraphy, apart from some differences in distribution of ammonites in the Bohemian Cretaceous Basin and in NW Europe, there are some reasons to suppose that *Forresteria (H.) petrocoriensis* appears in the BCB earlier than in the Rohatce Member Xd, which has been related to the base of the Coniacian by SOUKUP (1955, 1959) and by KLEIN & SOUKUP (1966). For example, *"Ammonites dentatocarinatus"* was cited to occur somewhere in the member Xc of the Teplice Formation in Vršovice-Černodoly (FRIC 1893); *Forresteria (H.) petrocoriensis* was also collected below the lower limit of *Stensioeina granulata levis* in SE England; *St. gr. "kelleri"* occurs at the same stratigraphic level in the BCB.

There are two possible boundary levels which fall into an unnamed interval between the known occurrences of the Turonian *Prionocyclus germari* and the Coniacian *Forresteria (H.) petrocoriensis* in the BCB (Text-Fig. 4):

1) At the first occurrence of rare *Stensioeina gr. "kelleri"* and *Karsteneis (P.) radegasti*.

2) The base of the Coniacian may be located where *Cremnoceramus waltersdorfensis* appears together with *Didymotis* event II but here are a few meters of overlap with the occurrence of *P. germari*. Nevertheless, *Didymotis costatus* and *C. waltersdorfensis* co-occur here abundantly. In the BCB, this level occurs in the Teplice Formation within a monotonous sequence of claystones or marlstones and limestones. The *Didymotis* event II is close to the maximum transgression over the Bohemian Massif, while a regression is recorded contemporaneously in the type areas in France. This indicates that the Turonian-Coniacian transition in the BCB occurs within a complete stratigraphic succession and is not interrupted by any lithological change, hiatus or hardgrounds, which would be detrimental to the proper establishment of a stratotype.

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The K/T Boundary and "Yellow Clay" Layers in the Gosau Group, Northern Calcareous Alps, Austria

ROMAN LAHODYNSKY, Wien

With 7 Text-Figures

LAHODYNSKY, R. (1989): The K/T Boundary and "Yellow Clay" Layers in the Gosau Group, Northern Calcareous Alps, Austria. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 677-690. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: By comparison of lithofacial series and with the help of nannoflora analyses (det. H. STRADNER) the author could locate three K/T boundary sections in the Flyschgosau of the Eastern Alps. In the Gosau "basin" of Gosau/Russbach two K/T boundary sites are situated within a series of interchannel deposits of the Zwieselalm Beds. In the Gosau area of Gams the K/T boundary was found within the slope association of the Nierntal Beds. The K/T boundary layers are enriched in Ca-smectite, montmorillonite and kaolinite. Iridium coincides with aluminium and elemental carbon. In both Gosau areas the main Ir-peak levels occur in different layers. This suggests diachronous events. Field observations and analyses point towards a gradual lithological change across the Maastrichtian/Danian transition. Some major element oxides decrease within the uppermost marly limestone layer (Gosau) or calcareous marl (Gams). The characteristic boundary clay layers as well as several other yellow clay intervals are associated with calcite layers, sheared zones and a sequence of brownish marls. Many of the observed microfractures in grains presumably have been formed as a result of shear displacement. Mineralogical analyses support a volcanic source for the element enrichments.

Kurzfassung: In der Flyschgosau der Ostalpen konnten vom Verf. mit Hilfe lithofazieller Serienvergleiche und Nannoflorenbestimmungen (det. H. STRADNER) drei Stellen lokalisiert werden, an denen die Kreide/Tertiär-Grenze aufgeschlossen ist. Im Gosau-"Becken" von Gosau/Russbach (Typlokalität) liegt die K/T-Grenze in einem Zwischenrinnenbereich; darüber erstreckt sich die Rinnenfazies der Zwieselalmschichten mit Unterbrechungen von der NP2-Zone bis in die Zone NP7. In der Gamser Gosau hält die Hangfazies der Nierntaler Schichten bis in die Zone NP2 an. Die Sedimentation der Zwieselalmschichten erreicht in diesem Gosauvorkommen ihre gröbste Schüttung in den Zonen NP3 bis NP5 und setzt sich mit einer Tonmergelserie bis in die Zone NP9 fort. Die Grenztone im Übergang Maastricht/Dan führen Smektit, Montmorillonit und Kaolinit und treten im Verband mit schichtparallelen Scherzonen und Kalzitlagen auf. Ein Zusammenhang zwischen Mikrorissen in Quarzen und den Scherflächen wird vermutet.

Bereits in der obersten Mergelkalk- (Gosau) bzw. Kalkmergelbank (Gams) des Maastrichts ist eine graduelle Abnahme des Karbonatgehaltes und verschiedener Oxide festzustellen. Die Verteilungskurven für Iridium, Aluminium und elementaren Kohlenstoff stimmen überein. Die Ir-Spitzenwerte treten diachron auf. Das sich über mehrere tausend Jahre erstreckende Ereignis im Übergang Kreide/Tertiär ist deshalb eher mit terrestrischen Ursachen (z. B. einem Hot Spot-Vulkanismus) zu erklären.

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1. Introduction

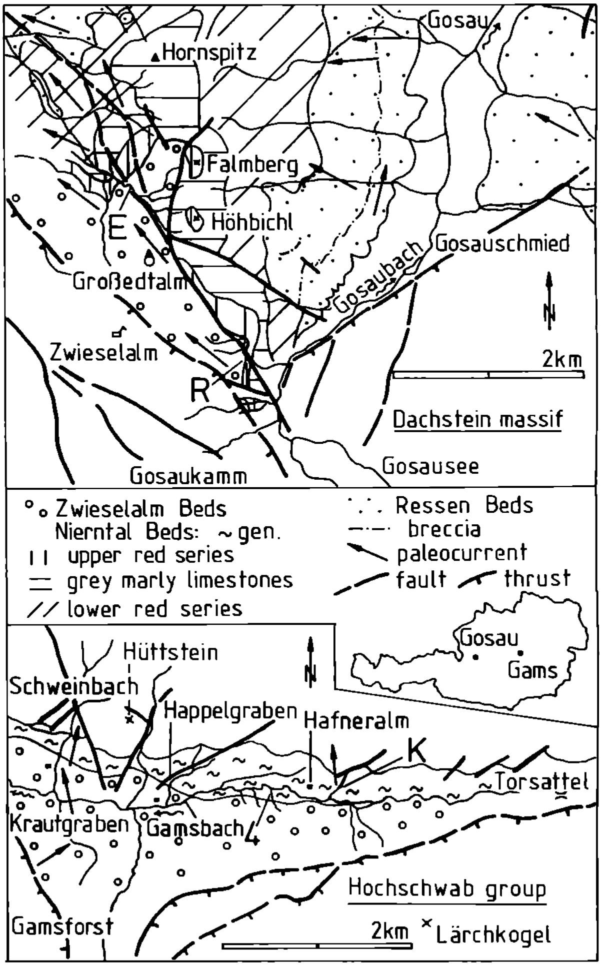
A K/T transition containing a boundary clay and geochemical anomalies is exposed only within the Upper Campanian - Lower Eocene flyschoid upper complex of the Gosau Group. The East-Alpine sections which have been studied in detail are Wasserfallgraben/Lattengebirge, Bavaria (HERM et al. 1981, PERCH-NIELSEN et al. 1982), Elendgraben/Dachstein, Salzburg (STRADNER et al. 1985, PREISINGER et al. 1986), Knappengraben/Hochschwab, Styria (STRADNER et al. 1987) and Rotwandgraben/Dachstein, Upper Austria (LAHODYNSKY 1987 and this paper). Results of mineralogical analyses and paleomagnetic measurements on the Elendgraben section have been reported by PREISINGER et al. (1986) and MAURITSCH (1986). It is the purpose of this paper to discuss the K/T boundary and other "yellow clay" layers in the Austrian sections within their rock framework.

2. Lithostratigraphical position and sedimentology

The Gosau Formation was deposited upon mid-Cretaceous nappe-structures and can be roughly divided into a lower shallow-water complex with reefs and a higher complex termed Flyschgosau. Sections wherein the K/T transition crops out have been found in the Gosau areas of Gosau (Elendgraben, Rotwandgraben - Text-Fig. 1) and Gams (Knappengraben - Text-Fig. 2).

Micropaleontological investigations by GANSS & KNIPSCHEER (1954), KÜPPER (1956) and WILLE-JANOSCHEK (1966) remained unsuccessful to locate early Danian because of sampling across and close to the Rotwandgraben fault. Yet KÜHN (1960) insisted on a Danian age for the Zwieselalm Beds and presumed a reworking of Maastrichtian marls by coarse grained layers.

2.1 In the area of Gosau (type locality of the Gosau Group) the deep sea-fan sequences of the Flyschgosau comprise Ressen-, Nierntal- and Zwieselalm Beds. Breccias and sandstones of the Ressen Beds mark the beginning of a coarse clastic development. Thickening and coarsening-upward cycles can be interpreted as a mid fan sequence. The Ressen Beds range from Lower Campanian to Upper Campanian. The thickness of this complex and of single turbidites decreases towards its distal part in the west. The overlying Nierntal Beds (Upper Campanian - Lower Maastrichtian) consist of red and grey marlstones with slump structures, interbedded thin sandstone layers and some mass flow deposits, and are interpreted as slope facies. Their disconformable deposition upon Triassic substratum in wide



Text-Fig. 1. Geological map of the southern part of the Gosau area with location of the K/T boundary sections (Elendgraben E and Rotwandgraben R) and mean directions of paleocurrent in the flyschoid Gosau formations.

Text-Fig. 2. Eastern Gams "basin" with K/T boundary Knappengraben K, outcrop "4" of WICHER & BETTENSTÄDT (1956) and paleoflow directions.

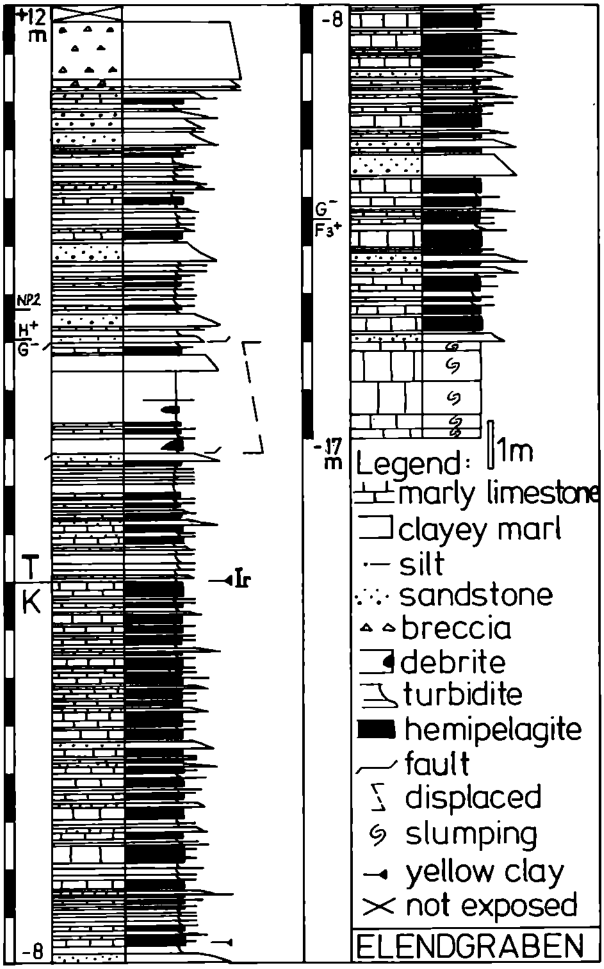
areas indicates deep subsidence in this part of the East-Alpine trough. In the Zwieselalm Beds (in Gosau from Upper Maastrichtian to Upper Paleocene) the turbiditic sedimentation continues with graded breccias and sandstones containing bioclastic and phyllitic rock fragments. Thin-bedded turbidite sequences with grey hemipelagic intervals (interchannel deposits) are often disrupted by synsedimentary slumps of mostly red marlstones. The Elendgraben site is located north of the Edtalm (Text-Fig. 1) on the western flank of a ravine immediately upstream of a sharp bend of the

Elendgraben torrent. The whole section across the K/T boundary is exposed along 5 m and owes its protection from erosion to the nearby passing Rotwandgraben fault which preserved the uppermost Maastrichtian and Paleocene sediments in the down thrown block. The K/T boundary clay is situated within a thin-bedded sandstone-siltstone/grey marl and marly claystone cyclic sequence above a series of grey Upper Maastrichtian marly limestone layers (Text-Figs. 3, 6).

Within the Tertiary (from NP2 to NP7, nannoflora det. H. STRADNER), sequences of coarse-grained breccias are intercalated, interpreted as channel fill association. Flute casts, imbrication of metamorphic rock fragments and slump-fold axis prove a paleostream towards northwest in general.

2.2 The Flyschgosau in the area of Gams is restricted to the Krautgraben valley east of the Noth-gorge. In the small torrents north of the Gamsbach the Upper Campanian and Maastrichtian marls of the Nierntal Beds dip downhill. An olistostrome at the Campanian/Maastrichtian transition contains olistoliths of Lower Campanian age (KRISTAN-TOLLMANN & TOLLMANN 1976). Breccias and sandstones of the Zwieselalm Beds (NP3 - NP5) follow further downslope. South of the Gamsbach a subsequent marly sequence reaches up to the NP9 zone (det. H. STRADNER). East of the Happelgraben the Gamsbach runs along the K/T transition and the Tertiary Nierntal Beds crop out in the ravines south of the Gamsbach torrent. In one of these the assumed K/T boundary of WICHER & BETTENSTÄDT (1956) is exposed (Text-Fig. 2). KOLLMANN (1964) mapped the Gams area - also a fault-bounded basin - with micropaleontological biostratigraphy and traced the Maastrichtian/Paleocene transition from the Hüttstein forest road west of the Schweinbach along the Gamsbach eastward into the Krimpenbach valley. Comparison of lithologic sequences together with nannofossil analyses by H. STRADNER enabled to find the K/T boundary in the Knappengraben (Text-Figs. 2, 4, 5). The outcrop is located east of the Hafneralm along the scarp of the Saugraben forest road. The Maastrichtian Nierntal Beds comprise alternating thin turbidites consisting of dark grey clayey marls (often without a sandstone layer) and light grey hemipelagic calcareous marls. Above the K/T boundary thin, fine-grained sandstones and greybrown-brownish red pelitic intervals with intercalated chaotic debrites comprise the Paleocene part of the Nierntal Beds. Here, pieces of the boundary clay layer are exposed on top of Maastrichtian olistoliths (Text-Fig. 5, lower part of the center column).

2.3 Previously the preservation of a K/T boundary in the vicinity of the Rotwandgraben (Text-Figs. 1, 6) was not assumed because of dilution within coarse-grained deposits in upward direction of the paleoslope. Although the K/T boundary could not be detected by means of nannoflora determination (which yielded NP6 due to sinter deposits) the comparison of sedimentological profiles effected the location of the boundary clay. Here too, no aftereffects of the hypothetical sudden event (tsunami deposits) could be found below the boundary clay (LAHODYNSKY 1987) but chaotic deposits occur in stratigraphically lower and higher positions in all of the investigated sites. Text-Figs. 3 and 4 show complete K/T boundary sections (biostratigraphical zonation after H. STRADNER, pers. comm. and in PREISINGER et al. 1986; magnetostratigraphy after BECKE & MAURITSCH, in

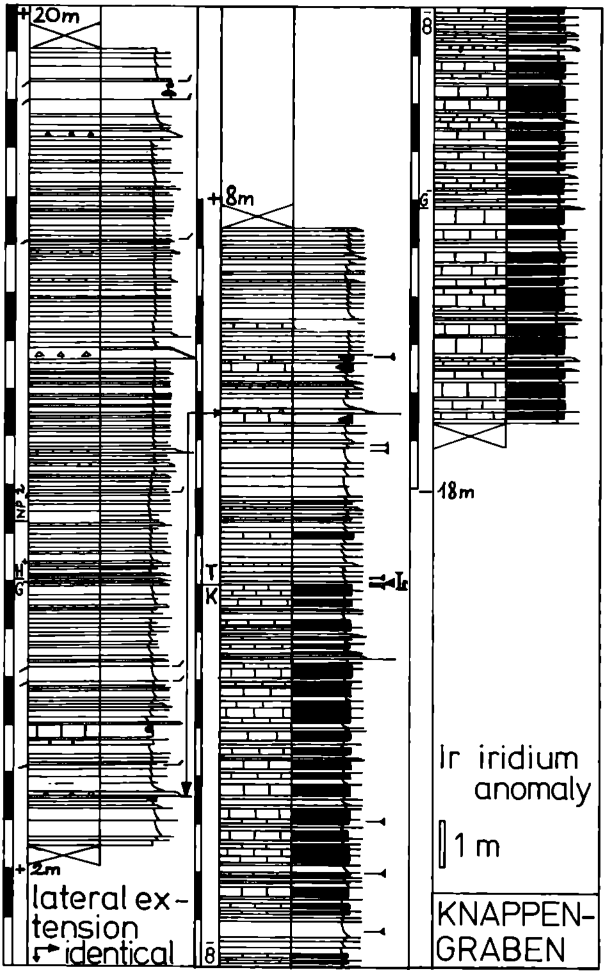


Text-Fig. 3. Complete lithostratigraphic profile across the K/T boundary Elendgraben, Abtenau, Salzburg (scale bar = 1 m).

PREISINGER et al. 1986 and MAURITSCH & ZEISSL 1987). Text-Figs. 5, 6 depict details of the lithology of three Austrian K/T transitions and of sections with other yellow clay layers.

3. Mineralogy

Below the K/T boundary the amount of $CaCO_3$ decreases gradually within the uppermost few cm of the Maastrichtian marly limestone or calcareous marl. Fe-oxides occur as concretions, crack fills and joint coatings. In all of the sites the K/T boundary layers consist of a white or greenish-grey



Text-Fig. 4. Lithostratigraphic profile across the K/T boundary Knappengraben, Gams, Styria (scale bar = 1 m). Upper part of the section in left column.

basal clayey marl, a yellow or rust-brown clay enriched in montmorillonite and pyrite-octahedrons (SURENIAN 1987) in the middle part (together less than 1 cm), and of a dark grey kaolinitic clay with coarser pyrite agglomerates on the top. This sequence is a characteristic feature of the K/T boundary clay but several other thin yellow clay layers could be observed in the Maastrichtian and in the Danian as demonstrated in Text-Figs. 5, 6.

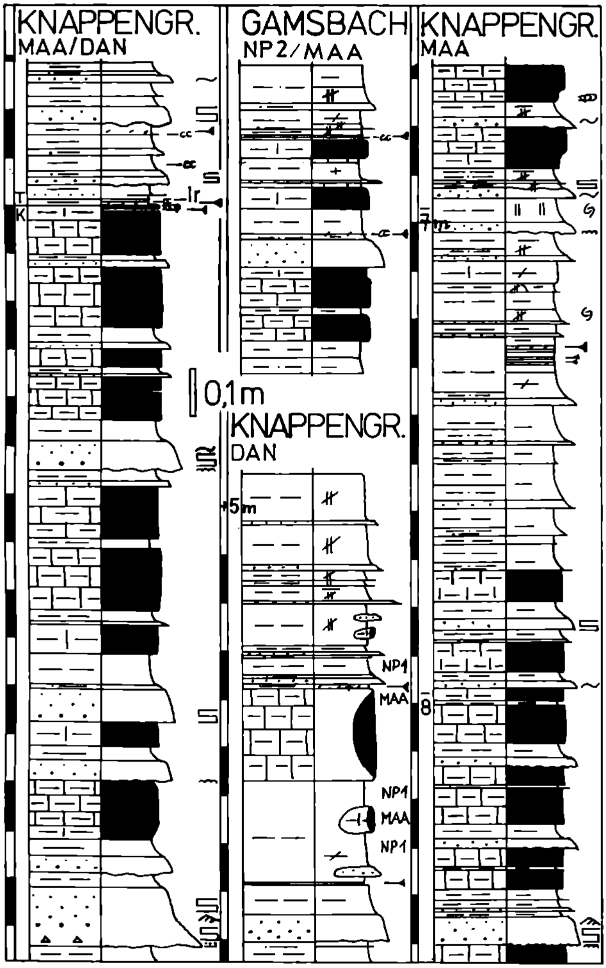
After NAIDIN (1987) visible hardgrounds, hidden omission surfaces and thin clay intercalations are an important feature of Campanian and Maastrichtian carbonate successions. The quantity of planktonic foraminifera increases beneath these intercalations. Associated with the increase in biological productivity is an accumulation of organic matter on the sea floor in excess. The following sharp decrease of planktonic foraminifera at the base of the clay is explained by carbonate dissolution caused by biogeochemical CO_2-formation. NAIDIN suggests that an end-Maastrichtian maximum biological productivity was the reason for an ecological catastrophe that led to mass mortality.

The unique K/T boundary clay and some of the other clayey intervals in the Gosau sections are linked to calcite layers and mylonitic zones. In most cases a succession of grey-brown or red-brown marls follows further upward (Text-Figs. 5, 6). The presence of sheared zones and calcite plates at the K/T boundary in the Gosau sections (but also at Zumaya) suggests that some of the observed microscopic fractures in grains occur adjacent to shear fractures and faults.

The clay minerals of the boundary clay (RAMPINO & REYNOLDS 1983, PREISINGER et al. 1986) may represent glassy volcanic ash that has been diagenetically converted to Ca-smectite and mixed layer clays. Alteration of volcanic glass particles to montmorillonite takes place by reactions that involve incongruent dissolution. The titanomagnetite can be derived from a basaltic volcanism. COURTILLOT & CISOWSKI (1987) favour the Deccan volcanism as an internal cause of the K/T event, but volcanic eruptions at the K/T transition happened elsewhere too, for instance in Japan. IIJIMA (1972) reported on end-Cretaceous and Paleocene violent volcanic activity which supplied acidic tuff, tuffaceous sediments and welded tuff to coal swamps. The K/T boundary therein is marked by a "grey mottled brown claystone and grey claystone with pyritiferous spots".

Legend of Text-Figs. 5, 6

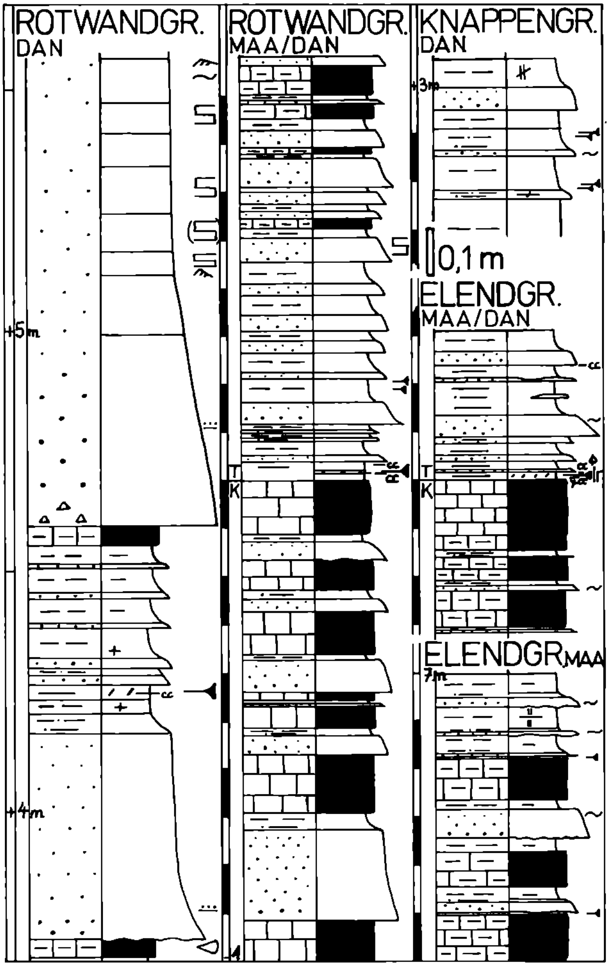
marly limestone	red colour
calcareous marlstone	red-brown
marl	yellow
clayey marl	grey-brown
clay	grey
sandy pelite	greenish
silt	Zoophycos
sandstone	Chondrites
breccia	plant detritus
debrite	displaced
hemipelagite	lensing
turbidite	crossbedding
slumping	wavy lamination
calcite layer	planar lamination
yellow clay horizon	graded division
iridium anomaly	flute cast
	erosive bed base



Text-Fig. 5. Lithostratigraphic profiles in the Gams area (scale bar = 0.1 m). Left column: K/T section Knappengraben. Center column: Yellow clay layers in Gamsbach section west of outcrop "number 4"; boundary clay on top of olistolith more than 4 m above K/T Knappengraben. Right column: Yellow clay layers in the Knappengraben section more than 7 m below K/T boundary.

4. Geochemical anomalies

Iridium enrichments were found within the K/T boundary layers and also within the turbidites immediately above at the Elendgraben site and the Knappengraben site. No measurements have been carried out on the Rotwandgraben samples until now. GRASS (1987, pers. comm.) measured a low bulk value of less than 1 ppb but titanomagnetite with an Ir content of 48 ppb in the rusty layer of the Elendgraben and a relatively high Ir



Text-Fig. 6. Lithostratigraphic profiles predominant in the Gosau area (scale bar = 0.1 m). Left column: Yellow clay layer in the Rotwandgraben section more than 4 m above K/T boundary. Center column: K/T boundary in the Rotwandgraben section, Gosau, Upper Austria. Right column: Thin brown clay layers in the Knappengraben section, 3 m above K/T boundary; Elendgraben section, K/T boundary and yellow clay layers more than 7 m below.

value of 14 ppb within the first sandy layer above the K/T boundary. Later a peak value of 14.5 ppb in the rusty layer of the Elendgraben section was reported (PREISINGER et al. 1986). This awaits verification by further measurements - the Elendgraben site was resampled together with M. NAZAROV in 1985 and A. HILLEBRAND in 1987. In the Knappengraben section the Ir content approximates 7 ppb within the rusty layer (STRADNER et al. 1987) but reaches its peak value of 11.4 ppb within the overlying dark kaolinitic layer (det. F. GRASS 1987 and pers. comm.). From these measurements it follows that at Gosau and Gams different layers contain the main Ir peak levels (Text-Figs. 5, 6). Therefore diachronous events which are possibly separated by hundreds or even thousands of years are strongly suggested. The high Ir values seem to be associated with the clay partings in the limestone sequence. In both areas a second peak of Ir concentration corresponds to the sandy divisions of very thin turbiditic layers. Ir strongly correlates with elemental carbon and with aluminium oxide (HANSEN et al. 1986: fig. 2, PREISINGER et al. 1986: fig. 4).

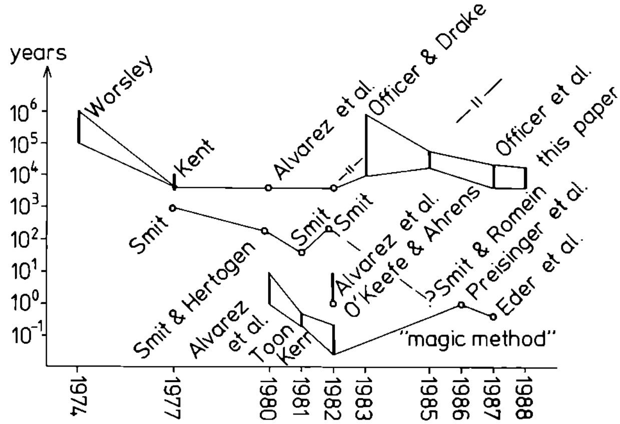
Ir enrichments are known from clay layers rich in ilmenite, magnetite and hematite and can be concentrated in sulfide rich sediments derived from stagnant environments. NAZAROV et al. (1982) point out that the observed Ir distribution could be connected with a difference in reducing conditions during sedimentation.

Recently significant Ir levels were measured in emissions from Kilauea (OLMEZ et al. 1986) which proves that hot spot volcanism could release sufficient Ir to account for the anomaly found at the K/T boundary. The measured higher Co, Ni, As and Sb levels before the Ir peak can be assigned to a volcanic origin (OFFICER & DRAKE 1985). This was confirmed by the Co and Ni measurements in the Elendgraben site (PREISINGER et al. 1986). CLOETINGH (1986) associates the dramatic end-Maastrichtian sea level fall with increased compression in the lithosphere. Could these stresses be linked to a deep source volcanism?

5. Timing of the K/T event

The estimated time interval for deposition of the boundary layers was subject of many debates and changed rapidly not only from author to author but also from page to page in the same publication (ALVAREZ et al. 1982: 8 - 1 year, 1982: 30 - 5000 years). From Text-Fig. 7 one may assume the development of a mysterious stratigraphical high precision technique in 1980. Its fundamental law was explained by ALVAREZ et al. (1982: 309): "... if one accepts the evidence from geochemical data that the iridium layer resulted from an impact, then it follows that the iridium layer represents an interval on the order of one year". For those who cannot accept this dogma and therefore will remain among "the few incurable skeptics" (HSÜ 1986) - who disbelieve in the success of shifting radiometric data to and fro to fit a periodicity of impacts - it is probably more realistic to enter again through ALVAREZ's backdoor (Text-Fig. 7). Although the precision of the timing was "enhanced" from 1000 years (SMIT 1977) through 200 years (SMIT & HERTOGEN 1980) to "very short" (SMIT & ROMEIN 1985) and even "shranked from the ominous 3 year period of darkness to a more comfortable 3 months or so" (KERR 1981) measurable geologic time remains on the order of thousands of years.

Estimation of sediment accumulation rates in the vicinity of the K/T transition depends on the duration of the magnetic time scale interval chron 29R which broadly varies (a comparison of different investigations of time intervals is given by OFFICER & DRAKE 1983: tab. 1). Previous calculations for the Elendgraben site (PREISINGER et al. 1986, EDER & PREISINGER 1987) were based on a flaw because the marly upper divisions of turbidites were added to the hemipelagic layers. These consist of calcareous marls and marly limestones in the Gosau "basins" deposited above the car-



Text-Fig. 7. Time estimation for the K/T boundary event.

bonate compensation depth (CCD) in a paleogeographically more southern position (HESSE & BUTT 1976, FAUPL 1979). Recalculated values around 1.7 cm/ka for the Cretaceous and 0.3 cm/ka for the Tertiary part of chron 29R coincide with calculations by SMIT & HERTOGEN (1980) for Caravaca, ARTHUR & FISCHER (1977) for Gubbio and for DSDP-site 465 (reported in OFFICER & DRAKE 1983).

6. Conclusions

A sudden event cannot be the cause of the observed phenomena such as the gradual decrease of $CaCO_3$, SiO_2, TiO_2, Fe_2O_3 and Na_2O within the uppermost limestone layer or calcareous marl below the boundary clay (PREISINGER et al. 1986: fig. 4) and the highly selective and gradual extinctions at the K/T transition. While only ichnofossils are abundant in the northern alpine Flyschgosau (*Zoophycos* MASSALONGO, *Chondrites* STERNBERG), the ammonites in the centralalpine Krappfeld-Gosau show a continuous decline throughout the Cretaceous (THIEDIG & WIEDMANN 1976). A duration of several 10^3 years and probably more than 10^4 years can be assumed for the deposition of the K/T boundary layers. The occurrence of the iridium peak level within different boundary layers indicates diachronous events - possibly separated by thousands of years.

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Faunal and Floral Distribution in Late Hauterivian Rhythmic Bedded Sequences and their Implications

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With 10 Text-Figures

MUTTERLOSE, J. (1989): Faunal and Floral Distribution in Late Hauterivian Rhythmic Bedded Sequences and their Implications. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 691-713. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The Lower Cretaceous sediments of the Boreal Realm are characterized by an obvious marl-clay rhythmically bedded sequence, which is present both in the marginal areas and in the basin. The differences of the bed-interbed rhythms are caused by an increased content of calcium carbonate of the pale marly layers respectively by an increased amount of detrital material and organic carbon in the dark, clayey beds. Certain faunas and floras of presumably Tethyan origin, which are restricted to the pale beds, indicate warm water conditions. This idea is supported by data gathered from Lower Cretaceous sections in France, which are represented by marl-limestone rhythms with an overall higher amount of calcium carbonate. Consistant with these observations and interpretations, the marl-clay rhythms of the Boreal Realm can possibly be explained by orbital forcing (Milankovitch cycle) postulated recently by KEMPER (1987).

Kurzfassung: Die Sedimente der borealen Unterkreide sind durch eine deutliche Ton/Mergelwechselfolge gekennzeichnet, die sowohl in den Randbereichen als auch im Beckenzentrum entwickelt ist. Die Bankungsunterschiede sind auf einen erhöhten Kalziumkarbonatanteil in den hellen, mergeligen Bänken bzw. auf einen erhöhten Anteil an detritischem Material und C_{org} in den dunklen, tonigen Bänken zurückzuführen. An die hellen, karbonatreichen Lagen sind Faunen und Floren mutmaßlich tethyaler Herkunft gebunden, die darauf hinweisen, daß es sich bei den Bänken um Warmwassersedimente handelt. Dieses Bild entspricht den Beobachtungen an französischen Unterkreide-Profilen, die, bedingt durch eine im Vergleich zu NW-Deutschland stärkere Karbonatführung, als Kalk/Mergelwechselfolge ausgebildet sind. In Übereinstimmung mit diesen Befunden wird die Ton-Mergel-Folge des Boreals auf die von KEMPER (1987) postulierten Schwankungen des Orbitalparameters zurückgeführt (Milankovitch-Zyklen).

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1. Introduction

The Lower Cretaceous sediments of both the Tethyan and Boreal realms are characterized by rhythmic bedded sequences. The Tethyan sequences consist of limestone-marl alternations, while the Boreal ones show a marl-clay alternation, generally poorer in calcium carbonate. The bed-scale alternating limestones and marls of the Vocontian Trough and the French Subalpine Ranges have been described and discussed most recently by DARMEDRU et al. (1982), COTILLON (1984), COTILLON & RIO (1984), FISCHER et al. (1985) and FERRY & RUBINO (1987a, b). COTILLON et al. (1980) explained the lithologic differentiation by periodic changes of the pelagic setting linked with changes of the detrital input. It seems plausible that global causes, either climatic fluctuation or eustatic changes, were controlling these alternations. COTILLON (1984, 1987) suggested a worldwide correlation of Lower Cretaceous strata using limestone-marl rhythms.

The NW European marlstone-clay alternations (Hell-Dunkel-Bankung, pale-dark-rhythms or p-d-rhythms), which are less obvious than the Tethyan ones, have been studied by SCHNEIDER (1963, 1964), ALIMIRZAIE (1972) and more recently discussed in detail by KEMPER (1987: 64). The colour changes are caused by rhythmical variation of the calcium carbonate content, the input of detrital material and the organic carbon content. According to KEMPER (1987) the marls represent warm phases of higher organic calcium carbonate production, while the clays were deposited during periods of high precipitation and increased production of suspended material. Similar interpretations have been made by PRATT (1983) for the marl-limestone cycles of the Greenhorn Cyclothem (Cenomanian) of Colorado.

If the assumption that the Lower Cretaceous p-d-rhythms are caused by either climato-eustatic or tectono-eustatic sea level changes is correct, these changes should be reflected in the composition of the fossil assemblages.

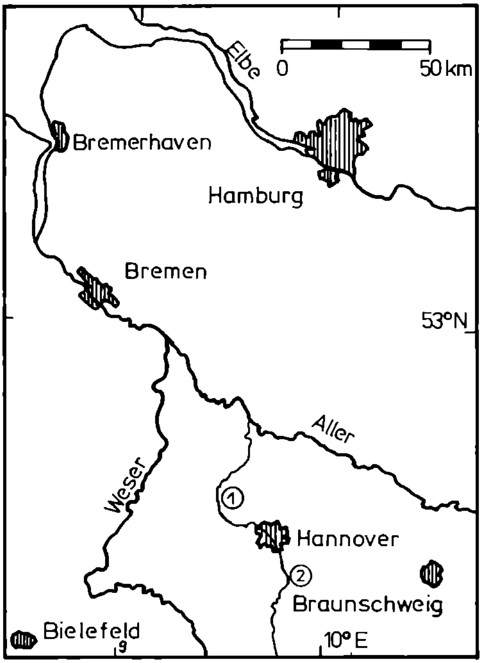
Recently two sections, which exposed very clearly the p-d-rhythms in a sequence of latest Hauterivian age, have been measured and analysed by the author.

2. Localities and material

Two localities in NW Germany exposing Upper Hauterivian (Discofalcatus Zone) sediments have been examined (Text-Fig. 1):

Frielingen: TK 25 Garbsen, no. 3523, R: 35 34 275, H: 58 17 125.

This now abandoned clay pit, Ziegelei Oltmann, was opened near Frielingen, about 20 km north-west of Hannover. About 15 m of fossiliferous clays are exposed, forming a rhythmic sequence of pale and dark beds, occasionally intensively bioturbated (Text-Fig. 2). Quite important is the occurrence of thin horizons rich in organic carbon. These beds are a forerunner of the Barremian "Blätterton"-facies, deposited under anoxic conditions. The strata, of latest Hauterivian age (Discofalcatus Zone), yielded a rich fauna: ammonites, belemnites, brachiopods, solitary corals, serpulids, shark teeth, crustaceans. Palaeogeographically the pit is situated in the easternmost



Text-Fig. 1. Map showing discussed localities. 1 = Clay pit Oltmann's near Frielingen; 2 = clay pit Gott near Sarstedt.

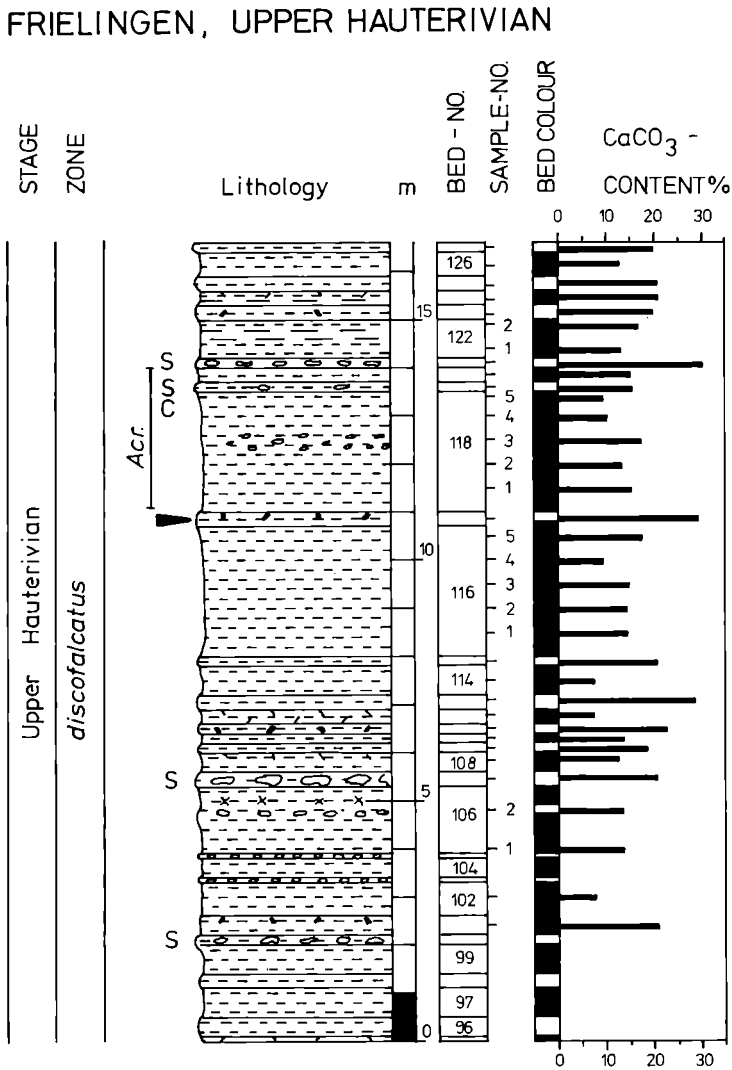
part of the central area of the Lower Saxony Basin, representing sediments of the basin facies.

Sarstedt: TK 25 Sarstedt, no. 3725, R: 35 60 400, H: 57 90 650.

The Gott clay pit, which is still worked, lies on the outskirts of Sarstedt, about 30 km south-east of Hannover. About 74 m of Upper Hauterivian, Barremian and Upper Aptian clays are exposed. The lowermost 11 m of the section, covering the Discofalcatus Zone (latest Hauterivian age), consist of a rhythmically bedded p-d sequence, yielding a rich fauna (Text-Fig. 3). These beds were deposited close to the Hildesheimer peninsula, indicating a shallow-water environment.

3. Lithostratigraphy

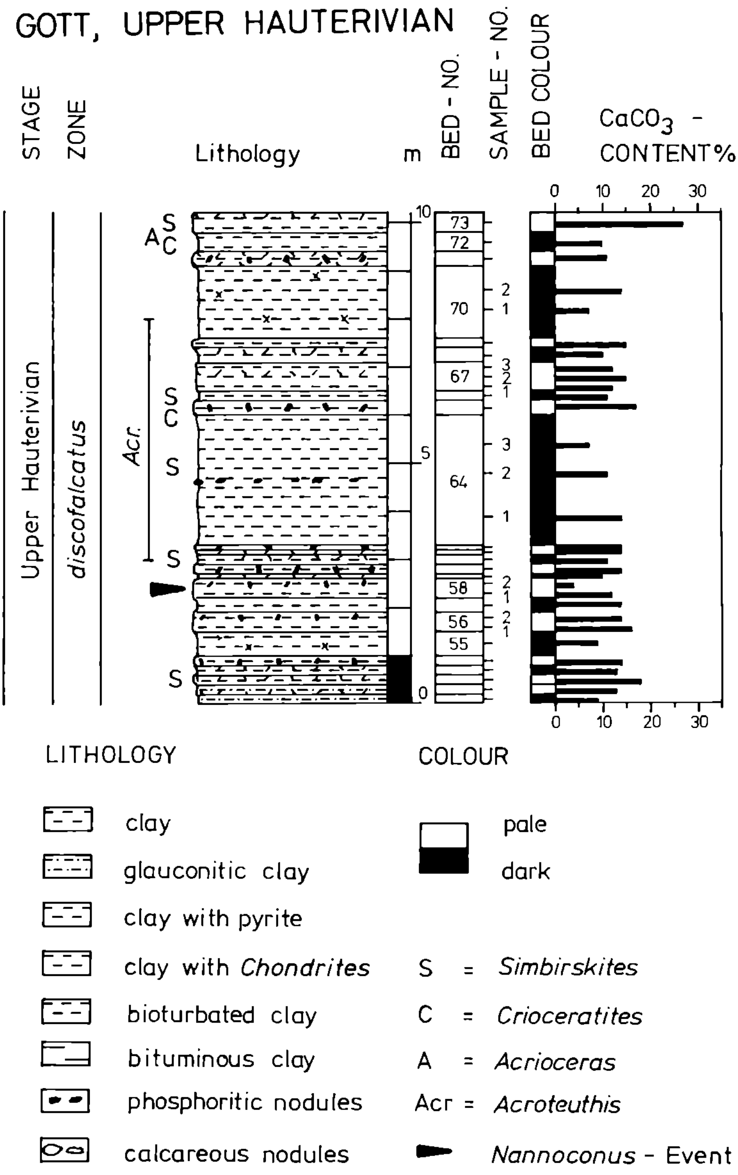
The marine Valanginian-Albian strata of the NW German Basin are characterized by a macroscopically more or less obvious p-d-rhythmic bedded sequence (Text-Fig. 4). According to SCHNEIDER (1963, 1964) this bed-scale p-d alternation is also present, though less clearly visible, in the



Text-Fig. 2. Lithology and biostratigraphy of the Frielingen section. Explanations see Text-Fig. 3.

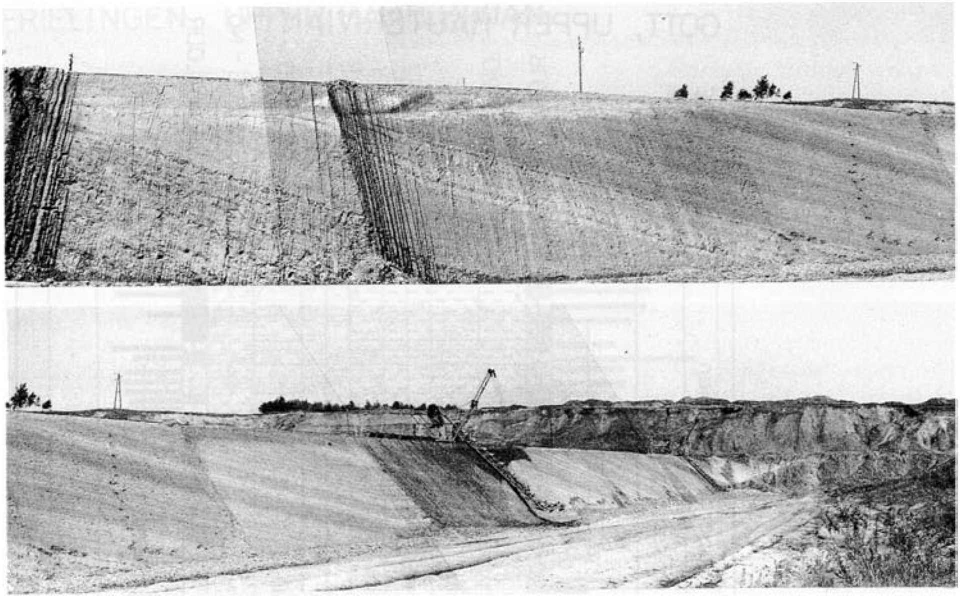
central part of the basin; it has, however, never been observed by the author. It becomes more obvious in the shallow-water sediments of the marginal facies. The p-d alternation can be observed throughout the Valanginian-Albian, although it is most clearly developed in the Upper Hauterivian. If the p-d rhythms have been correctly observed in the Valanginian, sediments of this age are only preserved in the basin facies; the p-d alternation is overprinted by high sedimentation rates. In the Barremian it is overshadowed by the palaeogeographic situation, favouring the deposition of the anoxic "Blätterton".

The thickness of the p-d-rhythms varies throughout the Lower Cretaceous. SCHNEIDER (1964) describes the thickest p-d-rhythms, up to 4 m,



Text-Fig. 3. Lithology and biostratigraphy of the Gott section.

for the Valanginian and Upper Aptian from the central part of the basin. Thinner rhythms are known from the Hauterivian and Barremian, since most of the Hauterivian and Barremian sections are to be found in a marginal position. The difference in colour is caused by variations in calcium carbonate content within the rhythms, which is greatest in the pale beds. The average calcium carbonate content in the Valanginian (Barremian) is



Text-Fig. 4. Clay-marl rhythms in the Barremian part of the Gott section, 1968. Foto det. Dr. F. SCHMID.

9.7-11 % (1.5-21 %) for the pale beds and 5.7-7.4 % (0.2-11 %) for the dark beds (SCHNEIDER 1964). The dark beds have a higher organic carbon content. This resulted from a higher input of suspended material of terrigenous origin.

The p-d-rhythms are not only restricted to the Lower Cretaceous of the Lower Saxony Basin, but have also been observed at Speeton/NE England. Strata of Hauterivian and early Barremian age, which are well exposed, show p-d-rhythms very similar to that of the NW German Basin (RAWSON 1971, RAWSON & MUTTERLOSE 1983). Since the p-d-rhythms become most obvious both in the marginal and basin facies in the latest Hauterivian, two sections exposing these rhythms have been studied in further detail. Both sections clearly show an increased calcium carbonate content for the pale beds (compare Text-Figs. 2 and 3). The calcium carbonate content of the pale beds varies from 15-53 % (average 24 %) for Frielingen and 11-27 % (average 14 %) for Gott. The dark beds contain between 7-22 % calcium carbonate (average 12 %) in Frielingen and 7-14 % (average 10 %) in Gott. This contradicts the above statement that the calcium carbonate content is generally higher in the marginal facies. The Gott section, which was closer to the coast, shows reduced thicknesses and reduced calcium carbonate values for both dark and pale beds in comparison to the Frielingen section. The input of detrital material was obviously higher in near-shore environments, though more continuous sections have to be analysed.

For the 16.5 m of the Frielingen section, 17 p-d-rhythms have been recognized, giving an average thickness of 1 m for each p-d-sequence. The 10.2 m of the Gott section consisted of 11 rhythms with an average thickness of 0.9 m. The thickness of the p-beds varies between 10-50 cm (average 22 cm) for Frielingen and 17-60 cm (average 32 cm) for Gott. The d-beds vary between 10-270 cm (average 77 cm) for Frielingen and 8-280 cm (average 60 cm) for Gott. The p-beds in both sections occasionally contain carbonate concretions, which may even form a solid carbonate layer. Statistical analysis of the data does not show any correlation between the bed thickness and the calcium carbonate content for either the pale or dark beds.

4. Biostratigraphy and correlation

Both sections yielded well preserved ammonites (*Simbirskites (Craspedodiscus) discofalcatus* (LAHUSEN), *Simbirskites (Simbirskites) toensbergensis* (WEERTH), *Simbirskites (Craspedodiscus) juddii* RAWSON, *Simbirskites (Simbirskites) picteti* (WEERTH)), placing these sequences in the Discofalcatus Zone (latest Hauterivian). In addition, the Tethyan derived *Crioceratites strombecki* (v. KOENEN) occurs in the upper part of both sections.

A remarkable biostratigraphic marker is the belemnite genus *Acroteuthis*, which is restricted to a horizon of a few metres in both sections. While *Acroteuthis (Boreioteuthis) stolleyi* PINCKNEY, a stout form, occurs in the lower part, the upper part is characterized by *Acroteuthis (Boreioteuthis) rawsoni* PINCKNEY which is much more slender and hastate in form. It seems likely that the latter, which closely resembles *Praeoxyteuthis pugio* and the genus *Aulacoteuthis*, is probably the ancestor of the Oxyteuthinae (*Praeoxyteuthis*).

This *Acroteuthis* bearing horizon enables a biostratigraphic correlation of both sections. Within the Frielingen section beds 118-121 yielded *Acroteuthis*, in the Gott section beds 61-70 (base). First *Crioceratites* occur in the pale beds above the *Acroteuthis* horizon. Quite striking is the fact that a pale bed rich in *Nannoconus* spp. occurs within both sections. Due to the abundance and diversity of *Nannoconus*, this bed (Frielingen, bed 117; Gott, bed 58) can be clearly correlated. The immigration of *Nannoconus* culminates in this bed. An accurate bed-by-bed correlation is more difficult,

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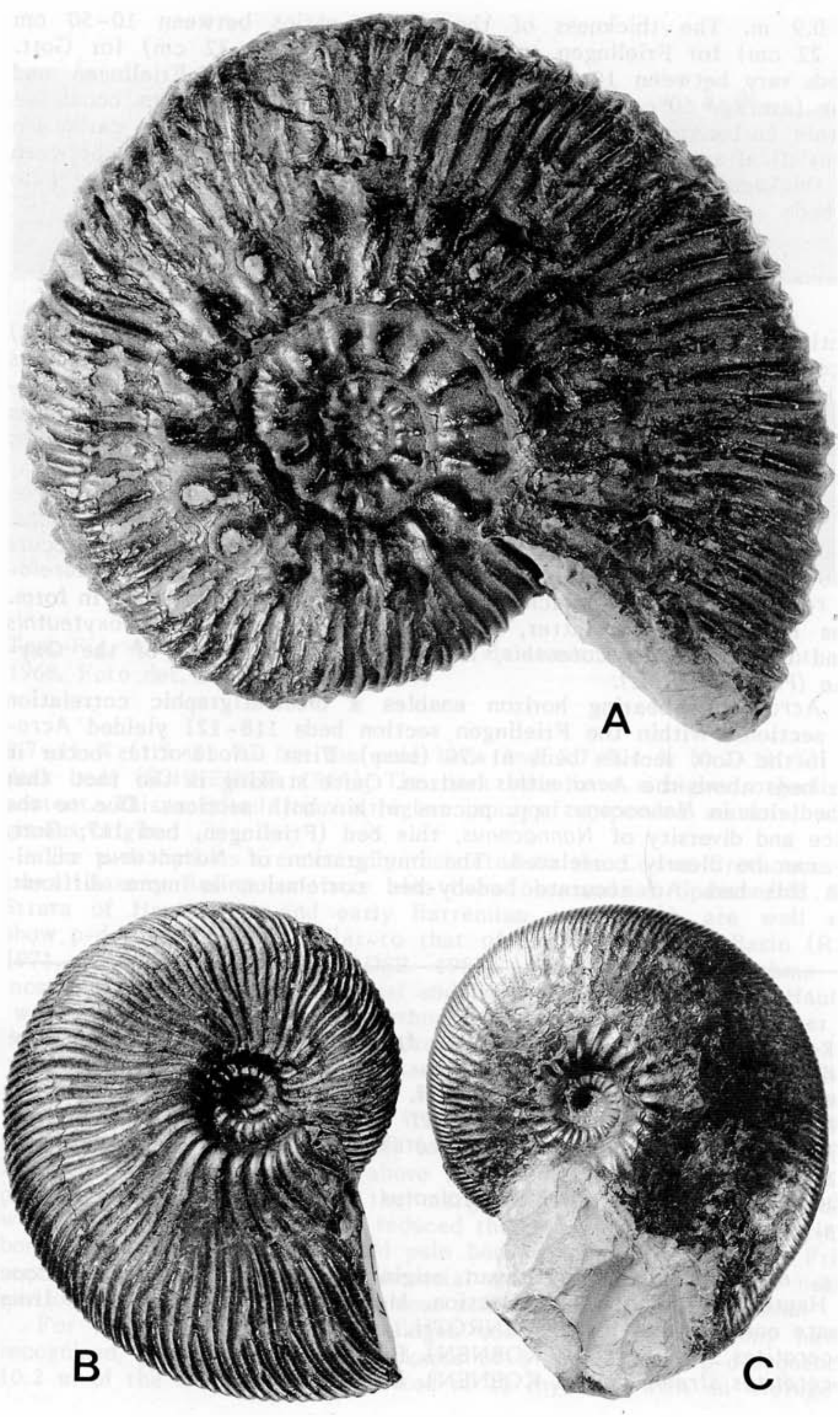
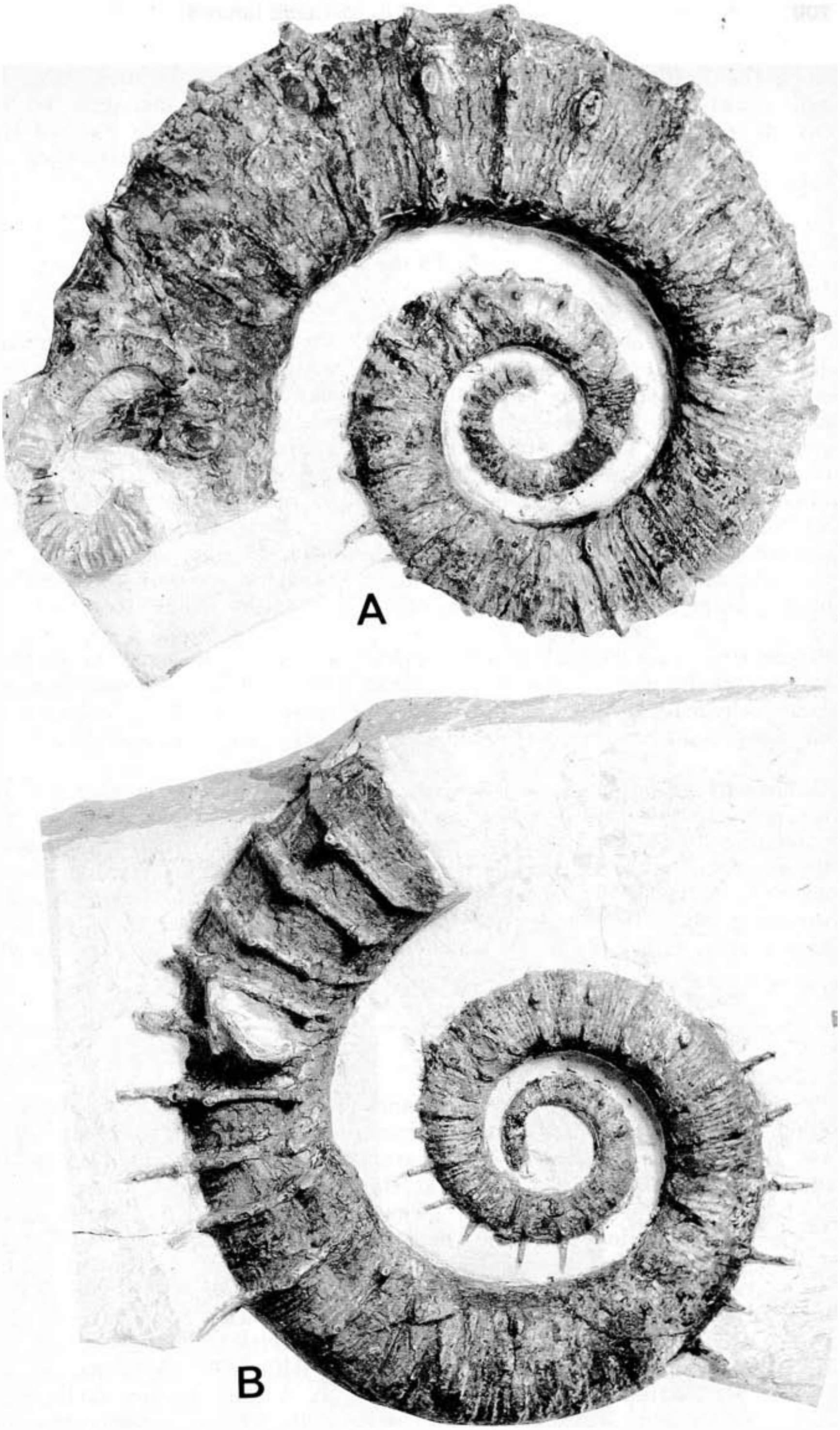
Text-Fig. 5. Ammonites of Boreal origin from the Discofalcatus Zone (Upper Hauterivian) of the Frielingen section. Material has been examined from the private collection of K. WIEDENROTH.
A *Simbirskites (Simbirskites)* ex gr. *picteti* (WEERTH). Bed no. 100, 1 x.
B *Simbirskites (Craspedodiscus) discofalcatus* (LAHUSEN). Bed no. 100, 1.6 x.
C *Simbirskites (Craspedodiscus) discofalcatus* (LAHUSEN). Bed no. 100, 0.5 x.

Text-Fig. 6. Ammonites of Tethyan origin from the Discofalcatus Zone (Upper Hauterivian) of the Gott section. Material has been examined from the private collection of K. WIEDENROTH.
A *Crioceratites strombecki* (v. KOENEN). Bed no. 73, 0.5 x.
B *Crioceratites strombecki* (v. KOENEN). Bed no. 73, 0.5 x.



Text-Fig. 5



Text-Fig. 6

since the field observations are controlled by varying factors, e. g. weathering, water content, etc. SCHNEIDER (1963, 1964) managed to correlate two of his sections. According to my own observations in two sections only 2 km from one another, a correlation of very characteristic pale and dark beds is possible (MUTTERLOSE 1983: 18).

5. Fauna and flora

Ammonites: The discussed interval yielded two ammonite genera of different origin: the Boreal genus *Simbirskites* (Text-Fig. 5) and the Tethyan derived genus *Crioceratites* (Text-Fig. 6). The vertical distribution of *Simbirskites,* which is represented by several species, may be described as follows. It is present both in the pale and dark horizons, although in Frielingen it is restricted to the basal part (beds 96-114) and top part (beds 120-127) of the section. No specimens have been found in the middle part (beds 115-119). Similar observations have been made for the Gott section, only beds 50-54 and 61-73 yielded *Simbirskites. Crioceratites strombecki,* which occurs in certain horizons of the Upper Hauterivian is thus far restricted to the pale beds and is only present in the upper part of the section (Frielingen, beds 118-121; Gott, beds 60-70).

Belemnites: Two species of *Acroteuthis,* a belemnite genus of Boreal-Arctic origin, are limited to a certain horizon, present both in Frielingen and Gott. The Tethyan-derived *Hibolithes jaculoides* SWINNERTON is common throughout both sequences, occurring in pale and dark beds (Text-Fig. 7).

Calcareous nannofossils: The evaluation of the calcareous nannofossils shows very clearly that the diversity and richness of the nannoflora has a positive correlation with the lithology. In general pale beds yielded rich and diverse floras; dark ones are impoverished. For the Frielingen section the Shannon diversity varies between 1.8-2.2 for the pale beds and between 1.2-1.9 for the dark beds. In average the diversity is 28 species per pale bed and 21 species per dark bed, the abundance varies between an average of 90 indivi-

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Text-Fig. 7. Belemnites of Boreal and Tethyan origin from the Discofalcatus Zone (Upper Hauterivian). All specimens have been coated with NH_4Cl and are figured in natural size. The material is housed in the Institut für Geologie und Paläontologie, Universität Hannover (GPIH).

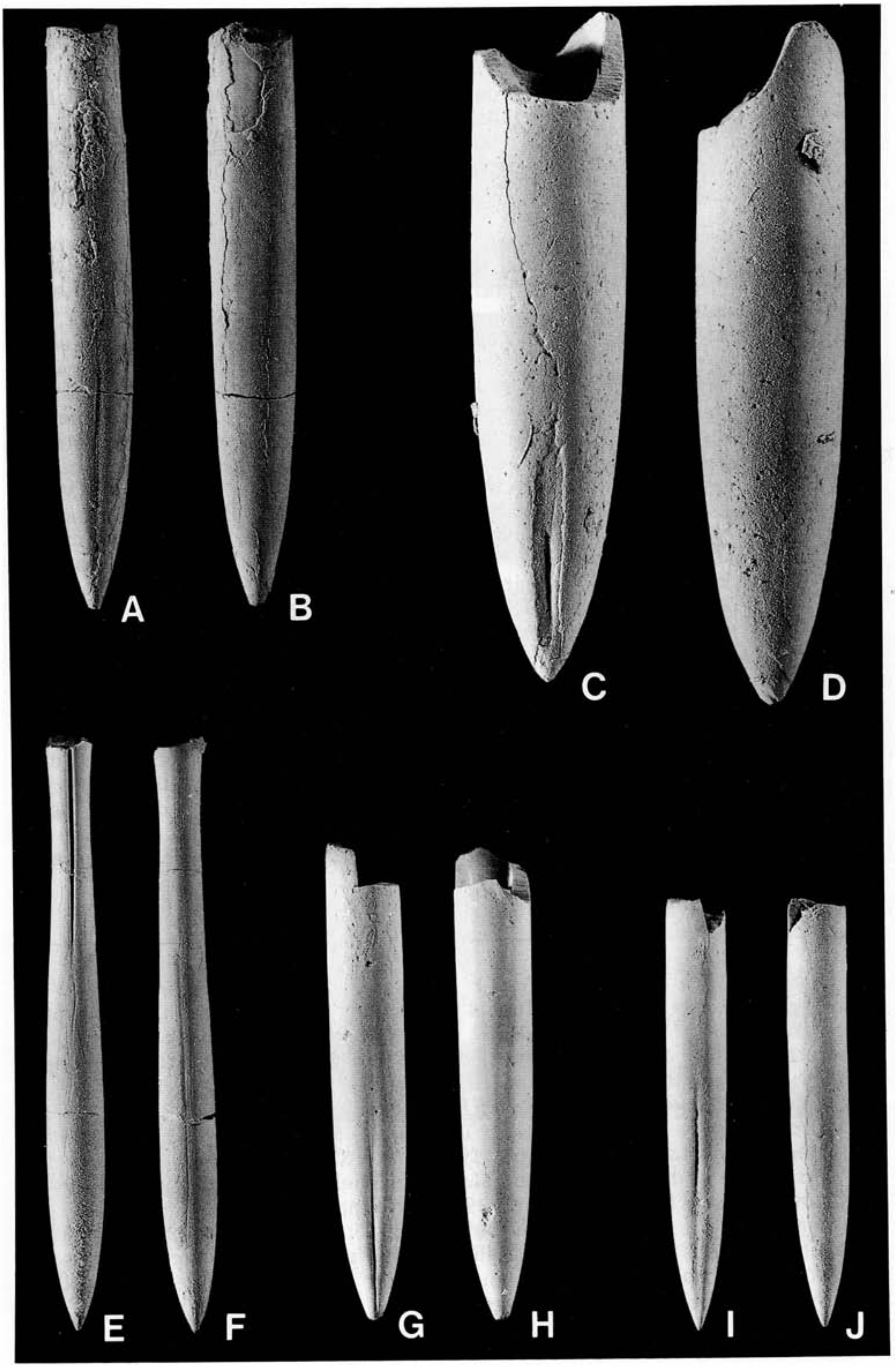
A, B *Acroteuthis (Boreioteuthis) rawsoni* PINCKNEY. A ventral -, B lateral view. Frielingen section, bed no. 118. Boreal species. GPIH 1989 I-1.

C, D *Acroteuthis (Boreioteuthis) stolleyi* PINCKNEY. C ventral -, D lateral view. Gott section, bed no. 65. Boreal species. GPIH 1989 I-2.

E, F *Hibolithes jaculoides* SWINNERTON. E ventral -, F lateral view. Gott section, bed no. 60. Tethyan species. GPIH 1989 I-3.

G, H *Acroteuthis (Boreioteuthis) rawsoni* PINCKNEY. G ventral -, H lateral view. Frielingen section, bed no. 120. Boreal species. GPIH 1989 I-4.

I, J *Acroteuthis (Boreioteuthis) rawsoni* PINCKNEY. I ventral -, J lateral view. Frielingen section, bed no. 120. Boreal species. GPIH 1989 I-5.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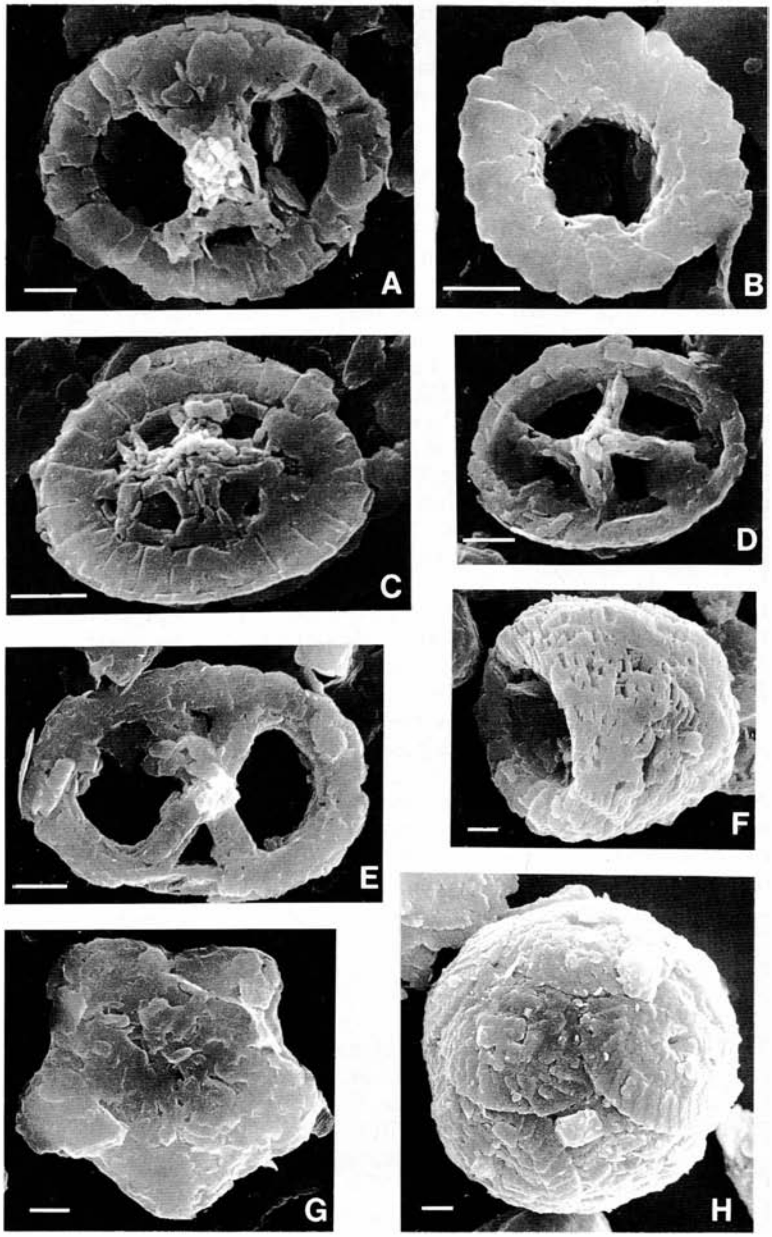
duals per mm² (pale beds) and 40 individuals per mm² (dark beds). In the Gott section the diversity varies between an average of 31 species (pale beds) and 25 species (dark beds), the richness between an average of 121 individuals per mm² (pale beds) and 66 individuals per mm² (dark beds). It should be noted, however, that those beds which are extremely rich in calcium carbonate (Frielingen, bed no. 121, 52 %) yielded only a very impoverished flora, neglected in the statistical analysis.

In addition to these distribution patterns, certain genera (*Nannoconus, Micrantholithus, Conusphaera*) are restricted to the pale horizons (Text-Figs. 8, 9). In the Frielingen section various species of *Nannoconus* KAMPTNER (*N. kamptneri* BRÖNNIMANN, *N. circularis* DERES & ACHERITEGUY, *N. globulus* BRÖNNIMANN, *N. minutus* BRÖNNIMANN, *N.* aff. *circularis*) occur occasionally in the pale beds (101 - 0.9 %, 107 - 1.5 %, 109 - 0.3 %, 111 - 0.3 %, 115 - 0.9 %, 117 - 9.9 %, 123 - 0.3 %, 125 - 0.3 %, 127 - 0.3 %; bed number - per cent of *Nannoconus*). *Nannoconus* is normally not present in the dark beds, it occurs only in bed 116. A similar positive correlation of pale colour, calcium carbonate content and the presence of *Nannoconus* has already been described from the Middle Aptian of NW Germany (MUTTERLOSE 1987, 1988). The vertical distribution of *Nannoconus*, however, is not uniform within the pale beds. The abundance of the nannoconids increases upwards in the pale beds of the section, reaches its peak in bed 117 (9.9 %) and then decreases upwards in the pale beds. Associated with *Nannoconus* are *Micrantholithus hoschulzii* (REINHARDT) and *M. obtusus* STRADNER, which are rare in the dark beds and more frequent in the pale beds. Both species comprise up to 5 % of the total assemblage in 117. Finally rare specimens of *Canusphaera rothii* (THIERSTEIN) have been observed only in the pale beds.

The Gott section shows a similar distribution pattern, with *Nannoconus* mainly restricted to the pale layers, although it occasionally occurs in dark beds as well. Normally *Nannoconus* makes up between 0.3-0.9 % of the whole nannofossil assemblage in the pale beds. However, in bed 58 it comprises up to 5 %. These two horizons (Frielingen bed 117; Gott bed 58) can be correlated. A more detailed statistical analysis of the data gathered still has to be done.

Text-Fig. 8. Calcareous nannofossils from the Discofalcatus Zone (Upper Hauterivian) of NW Germany. A, B, C specimens from a dark bed (57) of the Gott section. D-H specimens from a pale bed (117) of the Frielingen section. The bars in the left hand corner of the photographs = 1 µm.

A *Grantarhabdus meddii* BLACK 1971; Gott, 57/1.
B *Diazomatolithus lehmannii* NOEL 1965; Gott 57/1.
C *Retecapsa angustiforata* BLACK 1971; Gott 57/1.
D *Vekshinella stradneri* ROOD et al. 1971; Frielingen 117/1.
E *Chiastozygus octiformis* KÖTHE 1981; Frielingen 117/1.
F *Nannoconus* cf. *minutus* BRÖNNIMANN 1955; Frielingen 117/1.
G *Micrantholithus obtusus* STRADNER 1963; Frielingen 117/1.
H *Watznaueria barnesae* (BLACK in BLACK & BARNES 1959) PERCH-NIELSEN 1968; coccosphere; Frielingen 117/1.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6. Palaeobiogeography and migration patterns

The Lower Cretaceous sediments of NW Germany were deposited in a basin extending 280 km in E-W and 80 km in N-S direction. This basin formed the southernmost extension of the North Sea and the Boreal-Arctic Basin. Throughout the Hauterivian and Barremian this basin had marine connections to the Boreal Arctic Basin via the North Sea. On the other hand there was a direct open seaway towards the Tethys via Poland during the Valanginian and Hauterivian (KUTEK et al., this vol.). Tethyan faunas and floras immigrated via this Carpathian seaway into the Boreal Realm. In latest Hauterivian/earliest Barremian time this seaway was closed, the NW German Basin and the North Sea had only a small seaway towards the north. The closure of the Carpathian road is indicated by freshwater sediments of Barremian age in Poland, the last marine sediments yielding *Simbirskites (Craspedodiscus) gottschei* (MAREK & RACYNSKA 1979). This closure of a direct seaway towards the Tethys is clearly reflected by palaeontological and lithological data. There is an obvious increase of endemic taxa in NW Europe and the occurrence of anoxic sediments ("Blätterton"). The endemic belemnite species *Hibolithes minutus* replaces *H. jaculoides* at the Hauterivian/Barremian boundary. *Nannoconus abundans* and *N. borealis* PERCH-NIELSEN, which are restricted to the Barremian of the North Sea area, become the dominant *Nannoconus* species, probably evolving from *N. minutus*. Linked to these endemic faunas/floras is the deposition of anoxic sediments ("Blätterton") caused by the restriction of the Barremian sea (Text-Fig. 10). The first anoxic sediments are known from the highest Discofalcatus Zone in NW Germany (Text-Fig. 10), indicating the trend towards a separated basin. P-d rhythms are still present in the Barremian, though these are superimposed by the deposition of the anoxic "Blätterton".

Ammonites: The ammonite and belemnite data described above are somewhat contradictory. The ammonite faunas are dominated by various species of *Simbirskites*, a genus of Boreal-Arctic origin. Two waves of Simbirskitinae migrated from the Arctic Basin into the North Sea Basin in late Haute-

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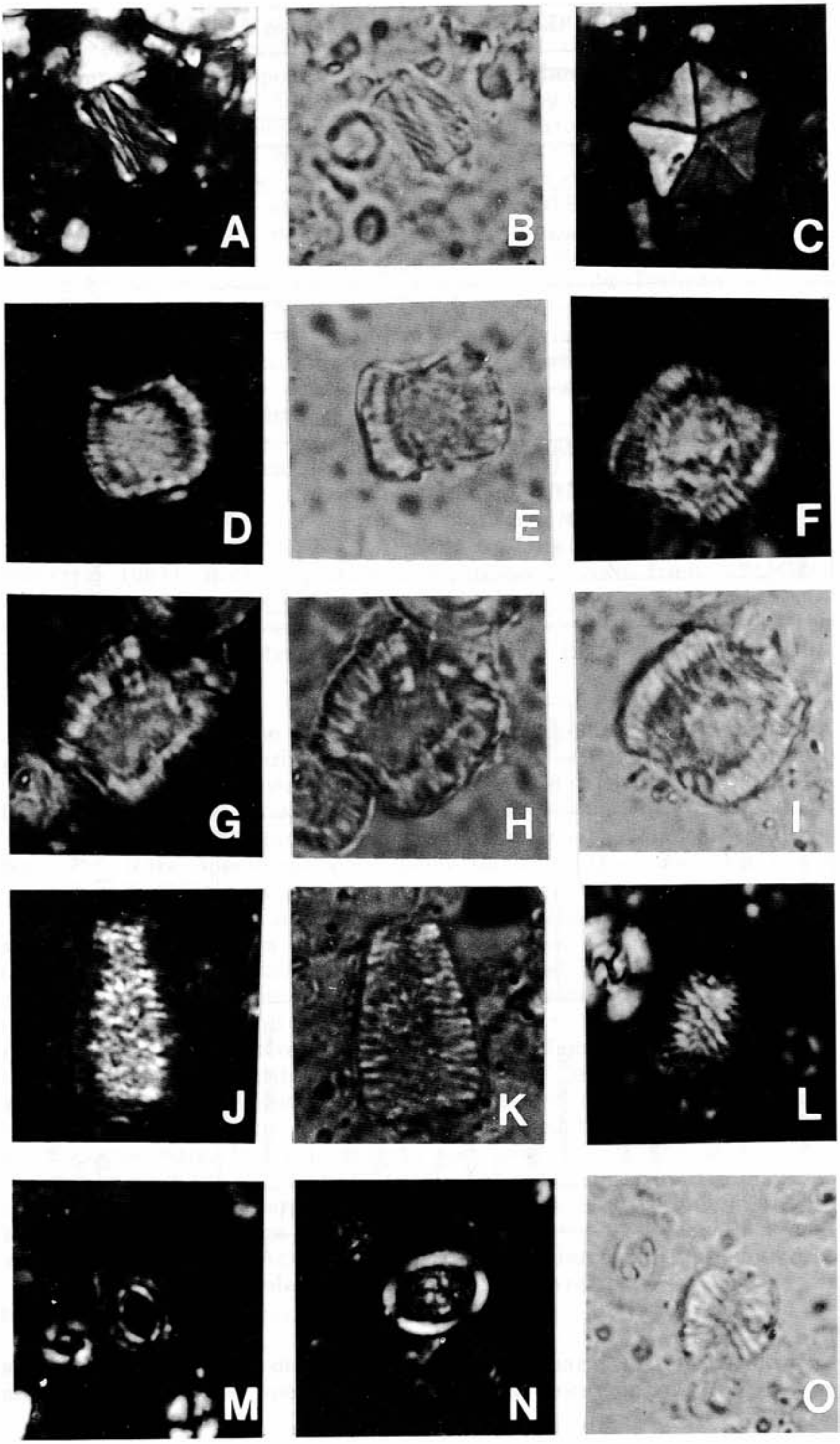
Text-Fig. 9. Calcareous nannofossils from the Discofalcatus Zone (Upper Hauterivian) of the Frielingen section. Assemblage from a pale bed (Frielingen 117/1). All magnifications x 2200.

- A, B *Conusphaera rothii* (THIERSTEIN 1971) JAKUBOWSKI 1986; A: x-nicols; B: bright-field, same specimen.
- C *Micrantholithus obtusus* STRADNER 1963; x-nicols.
- D-I *Nannoconus globulus* BRÖNNIMANN 1955; D: x-nicols; E: bright-field, same specimen as D. F: x-nicols; I: bright-field, same specimen as F. G: x-nicols; H: bright-field, same specimen as G.
- J, K *Nannoconus kamptneri* BRÖNNIMANN 1955; J: x-nicols; K: bright-field, same specimen as J.
- L, O *Nannoconus minutus* BRÖNNIMANN 1955; L: x-nicols; O: bright-field, same specimen as L.
- M *Sollasites horticus* (STRADNER et al. 1968) BLACK 1966; x-nicols.
- N *Rhagodiscus asper* (STRADNER 1963) REINHARDT 1967; x-nicols.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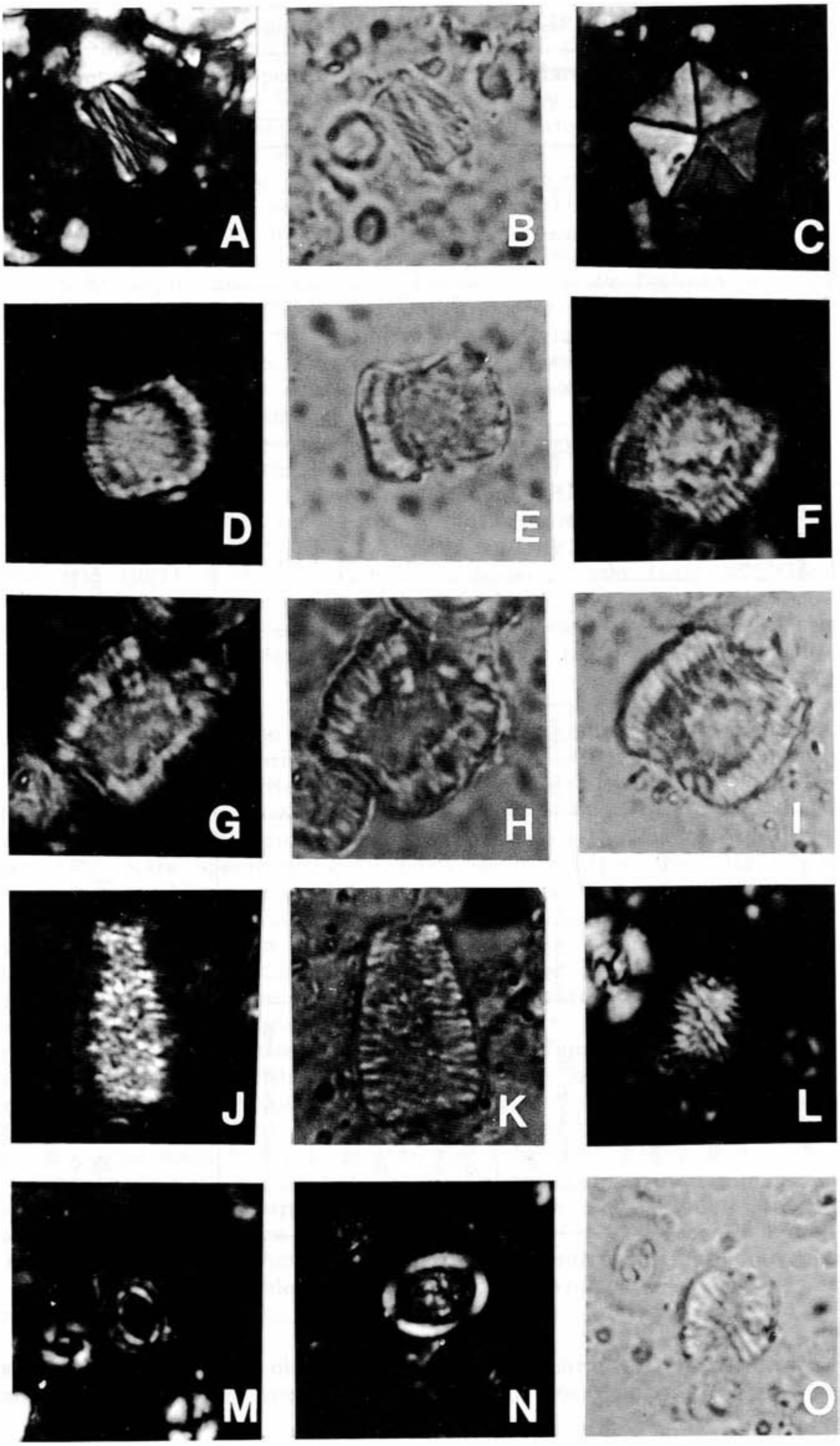
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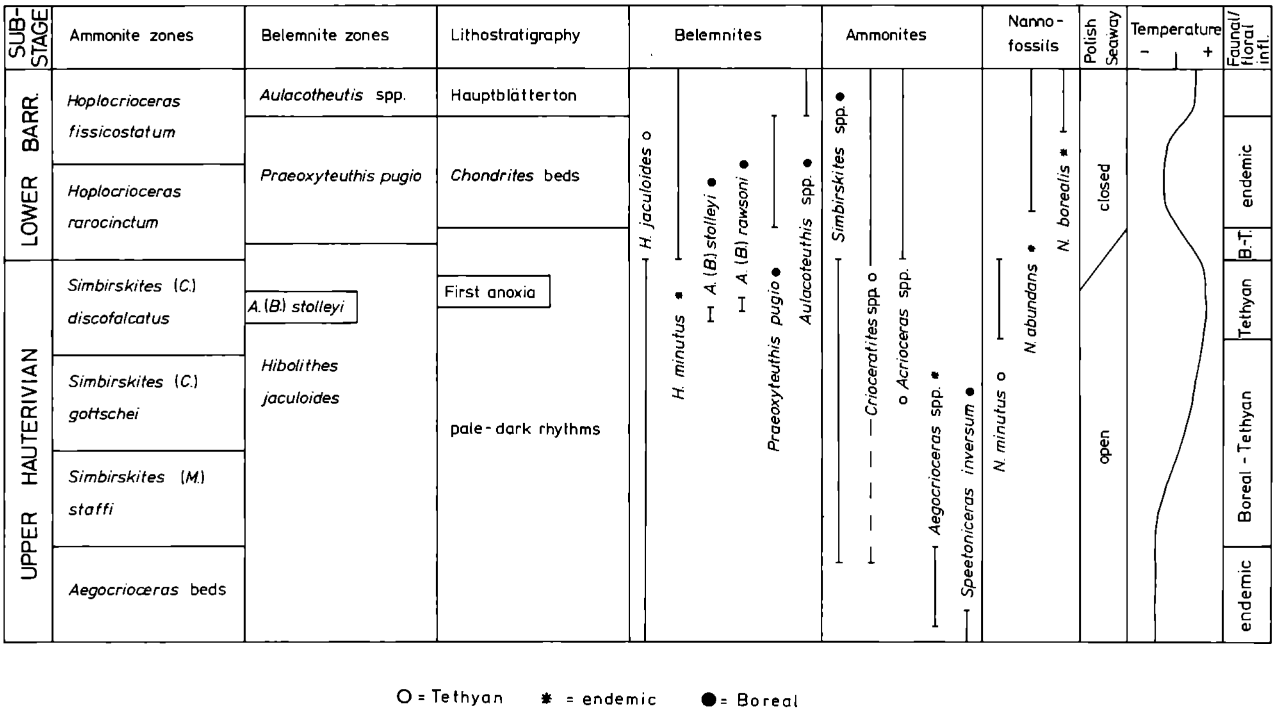

Text-Fig. 10. Faunal and floral distribution in the Upper Hauterivian and Lower Barremian of NW Germany. H. = *Hibolithes*, A. = *Acroteuthis*, B. = *Boreioteuthis*, N. = *Nannoconus*, C. = *Craspedodiscus*, M. = *Milanowskia*.

rivian time. The second migration wave was characterized by *Simbirskites* s. str., which is common throughout the remainder of the Upper Hauterivian of NW Europe and only disappeared at the Hauterivian/Barremian boundary. This southward migration of *Simbirskites*, forced by a regression in the Boreal-Arctic Basin, is accompanied by an adaptation to more southern living conditions (KEMPER & WIEDENROTH 1987, KEMPER, MUTTERLOSE & WIEDENROTH 1987). Since *Simbirskites* is known from the Hauterivian of France (THIEULOY 1977) and the Caucasus (KOTETICHVILI 1988) this genus was obviously able to migrate into the Tethyan Realm. OBATA & MATSUKAWA (1988) even mentioned *Simbirskites (Milanowskia)* sp. from the Barremian of Japan, associated with faunas of mainly Tethyan or Submediterranean affinities. This occurrence of a Boreal-Subboreal genus in a clearly Tethyan setting might be explained by a migration from the Boreal Sea to Japan via northeastern Siberia. This observation stresses the idea that *Simbirskites*, an originally Arctic genus, became more and more adapted to southern settings with time.

Crioceratites, an ammonite genus of Tethyan origin, migrated into the southernmost part of the Boreal Realm (= Subboreal) in two major waves. The *C. duvali* group entered the NW European basin in the early late Hauterivian and the *C. thiollieri/spathi* group immigrated in the latest Hauterivian (IMMEL 1979, KEMPER, RAWSON & THIEULOY 1981, KEMPER & WIEDENROTH 1987). Both migration horizons are known from NW Germany and NE England (Speeton). The latter of these migration horizons coincides with the described occurrence of *Crioceratites strombecki* in the Discofalcatus Zone. *Crioceratites (P.) spathi* is known from bed C2C (= Discofalcatus Zone) at Speeton.

Belemnites: *Acroteuthis* is the dominant belemnite genus in sediments of late Volgian to earliest Hauterivian age, although it was almost completely replaced early in the Hauterivian by the immigration of *Hibolithes* (MUTTERLOSE et al. 1983). *H. jaculoides*, which is of Tethyan origin, has been first recorded from the Valanginian and persists up to the Hauterivian/Barremian boundary. Rare specimens have been recorded from the Barremian (IMMEL & MUTTERLOSE 1980). The described pattern indicates a constant input of Tethyan belemnite faunas throughout the Hauterivian. Since none of the other belemnite genera common in the Tethys occur in the Boreal Realm, the presence of *H. jaculoides* can be explained by a wide ecological tolerance of this species. When the open seaways towards the Tethys ceased the endemic species *H. minutus* evolved.

The occurrence of a relatively rich *Acroteuthis* fauna in the *Hibolithes* beds has recently been described by MUTTERLOSE et al. (1987). These authors suggest an in situ evolution of the *Acroteuthis (Boreioteuthis) conoides - rawsoni - stolleyi* group. This idea is confirmed by recent bed-by-bed collections discussed here. In both sections the stout *A. (B.) stolleyi* is restricted to the lower part of the *Acroteuthis* horizon and the more elongate, cylindrical *A. (B.) rawsoni* characterizes the upper part. This indicates very clearly an evolutionary lineage from *A. (B.) stolleyi* to *A. (B.) rawsoni*. Therefore the occurrence of *Acroteuthis* in the Discofalcatus Zone does not indicate an immigration of belemnites from Boreal-Arctic seas, but rather an in situ evolution.

Calcareous nannofossils: The discussed interval is characterized by cosmopolitan and Tethyan derived species. The Tethyan floras mainly consist of

Nannoconus spp., while the cosmopolitan floras consist of the bulk of species. Endemic species (*N. abundans, N. borealis*) become common in the Barremian only. *Tegulalithus septentrionalis*, which is supposed to be a Boreal species, has been observed only once.

Since dissolution susceptible species like *Corollithion* and *Sollasites* are common both in the dark and pale beds, the distribution patterns described above cannot be explained by dissolution. A second important fact, which supports the idea that the floral differences in between marls and clays are primarily controlled by ecological differences, corresponds to the overall distribution. In both sections the percentage of *Nannoconus* steadily increases, culminates in one pale bed and declines. Due to the high percentage of *Nannoconus* this particular horizon can be correlated over 50 km (compare chapter "Fauna and Flora"). This indicates that the same palaeoceanographic conditions prevailed during the deposition of this bed.

7. Discussion

Lithologic changes in rhythmic bedded sequences (periodites) may be caused by various factors. EINSELE (1982) differentiates between rhythms caused by diagenetic factors and those caused by primary factors. Since diagenetic factors do not explain the lithological, faunal and floral differences between the p-d-beds, only primary factors (dissolution, dilution, productivity) are to be considered.

Dissolution has been used by various authors (e. g. ROTH & KRUMBACH 1986) to explain the different composition of floral assemblages in calcium carbonate-rich and -poor strata. However, since the preservation of calcareous nannofossils is good both in dark and pale beds, dissolution did not influence the composition of the floral assemblages. Dissolution susceptible species (*Corollithion geometricum, Sollasites horticus*) have been found both in p- and d-beds. Thus the major factor controlling the deposition of the p-d-rhythms are periodic changes in the detrital supply (dilution cycle) and/or fluctuations of the carbonate production (productivity cycle). For a detailed discussion and interpretation see KEMPER (1987). The phenomena described above can be explained by either climato-eustatic or tectono-eustatic changes.

The restriction of the Tethyan derived groups (*Crioceratites, Nannoconus, Conusphaera, Micrantholithus* (?)) suggests that the detrital input and the carbonate sedimentation were primarily controlled by climatic variation. The pale marls yielding Tethyan derived faunas and floras were deposited during dry and warm phases, the dark clays during wet and cool phases. The immigration of Tethyan faunas and floras is enforced during warm phases, while these elements retreat southward during cooler phases. Small-scale climatic fluctuations do not only explain the p-d-rhythms, but also the fact that certain beds can be correlated due to their floral composition. Comparable observations have been made by NOEL (1968) in the Upper Barremian limestone-marl rhythms of the Basses Alpes. The limestones are dominated by well preserved *Nannoconus*, while coccoliths are less abundant. The marls on the other hand have a different coccolith-*Nannoconus* ratio; coccoliths are outnumbering *Nannoconus*. This is clearly related to the pollen/dinoflagellate ratio. The limestones are dominated by dinoflagellates, the marls by pollen. NOEL (1968, 1978) and DUFOUR & NOEL (1978) explain this pattern by different ecological conditions required by *Nannoconus* in respect to the coccoliths. *Nannoconus* blooms were favoured by clean/clear water, while coccoliths tolerated a higher amount of suspended material. The difference of the pollen/dinoflagellate relation can be easily explained by the higher input of detrital material in the marls. This distribution pattern corresponds with observations made for the p-d-rhythms of the Boreal Realm, though the overall amount of calcium carbonate is reduced. The marl-clay rhythms of the Boreal Realm are equivalent to the limestone-marl rhythms of the Tethys. The dark beds of the p-d-rhythms have an increased organic carbon content. The organic carbon content of a dark bed of early Albian age varies between 0.9-1.4 %, that of a pale bed is 0.6 % (KUHN 1975, cited in KEMPER 1987). The increased organic carbon values of the dark beds imply higher detrital influx, although it may also indicate higher preservation of marine organic matter.

It is still open to discussion whether these rhythms are purely caused by dilution effects, assuming a constant calcium carbonate production, or by direct fluctuations of the carbonate production. It is not likely, however, that the restriction of *Crioceratites* and certain planktonic nannofossil genera to the pale beds, which is correlated with diverse and rich foraminifera assemblages (SCHNEIDER 1964), can be explained solely by dilution.

Explaining the p-d-rhythms by tectono-eustatic changes would suggest a more neritic, shallow-water regime for the deposition of the pale marls. Various authors (e. g. WISE et al. 1987) suggest a neritic character for nannoconids and micrantholiths. According to ROTH & BOWDLER (1981) and ROTH & KRUMBACH (1986) *Nannoconus* spp. and *Braarudosphaera* spp. are indicators of a continental margin, shallow plateau and epicontinental sea conditions. It does, however, not seem likely that the faunal and floral distributions described above, are mainly controlled by waterdepth. All the sections discussed here were deposited under shallow-water inner shelf conditions, both sections are near-shore sequences situated about 20-50 km off the coast. Within this area p-d-rhythms are to be found both in the basin and in the marginal facies. In both examples only the pale beds yielded nannoconids and micrantholiths. The great number of rhythms (approx. 700 for the Hauterivian-Albian interval) requires constantly changing epeirogenetic movements, causing shallowing and deepening effects.

KAUFFMAN et al. (1983) and BARRON et al. (1985) explained the cyclic bedded limestone-clay sequence of the Greenhorn Cyclothem (Cenomanian-Turonian) by a variation of the precipitation controlled by orbital periodicity. This periodicity was caused by changes in obliquity or axial tilt. These authors calculated an average of 40.000 years for each cycle, which corresponds to Milankovitch cycles. Various authors (COTILLON et al. 1980, COTILLON 1984, DE BOER 1982, FISCHER et al. 1985, FISCHER 1986, KEMPER 1987) tried to relate cyclic bedded sequences to climatic variations, explaining the changes by orbital forcing (Milankovitch cycles). COTILLON (1984) quoted several authors who calculated the duration of limestone-marl cycles in pelagic early and Middle Cretaceous sediments. According to these data the duration varies between 14.000-38.000 years (FERRY et al. 1981) to 30.000-100.000 years (ARTHUR 1979).

Interpolations for the Hauterivian rhythms gave a duration of 33.000-67.000 years. Though the interpolation of the data is speculative, relying on highly subjective field observations and on absolute data with a low tolerance, the calculated data lie within the given ranges.

Superimposed on these small-scale variations which are probably caused by orbital forcing are first and second order eustatic shifts. The second order is equivalent to the megacycles of KEMPER (1987) which lasted approx. 2 million years. Similar overprints of eustatic shifts have been described by FERRY & RUBINO (1987b). A first order shift is evident within the discussed interval. In both sections the Tethyan derived flora is characterized by an increase-decrease cycle, which is not clearly reflected in the lithological record. A similar trend is reported by MICHAEL (1979) who records an absolute maximum for dextrally coiled species of *Epistomina* in the latest Hauterivian. It seems possible that dominances of dextral-coiled planktonic and benthonic foraminifera indicate warm-water conditions (compare discussion by MICHAEL 1979: 319).

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D. Biostratigraphy, Correlation, Paleogeography

The Wąwał Section, Central Poland – An important Link between Boreal and Tethyan Valanginian

JAN KUTEK, RYSZARD MARCINOWSKI, Warsaw,
and JOST WIEDMANN, Tübingen

With 2 Plates, 6 Text-Figures and 2 Tables

KUTEK, J., MARCINOWSKI, R. & WIEDMANN, J. (1989): The Wąwał Section, Central Poland - An important Link between Boreal and Tethyan Valanginian. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 717-754. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The Neocomian section at Wąwał in Central Poland, in which, despite differing previous opinions, only Valanginian strata are represented, is reinterpreted stratigraphically in terms of Tethyan and NW European subdivisions. Two disconformities have been recognized, one at the base of *Platylenticeras* Beds, the second corresponding to the *Polyptychites* Beds.

Ammonites of the genus *Karakaschiceras,* including its hitherto unknown earliest representatives of *pertransiens* and pre-*pertransiens* age, are described, and two new species, *K. subgibbosum* and *K. pruszkowskii,* are established. The occurrence of *Karakaschiceras,* together with other Tethyan or Tethyan-derived ammonites, at three stratigraphic levels in Poland is due to three successive transgressions, (1) relative to the base of the Upper Berriasian, (2) to the base of the *Platylenticeras* Beds, and (3) to that of the Upper Valanginian (Verrucosum Zone).

Kurzfassung: Das Neokom-Profil von Wąwał in Zentralpolen, in dem - trotz anderslautender früherer Auffassungen - nur Schichten des Valangin vertreten sind, wird hier stratigraphisch revidiert und sowohl mit der tethydischen als auch mit der NW-europäischen Zonengliederung verknüpft. Zwei Diskordanzen sind im Profil erkennbar, die erste an der Basis der *Platylenticeras*-Schichten, die zweite den *Polyptychites*-Schichten entsprechend.

Ammoniten der Gattung *Karakaschiceras* werden beschrieben, einschließlich ihres bisher ältesten Vertreters von *pertransiens*- und prä-*pertransiens*-Alter. Zwei neue Arten werden aufgestellt, *K. subgibbosum* und *K. pruszkowskii*. Das Vorkommen von *Karakaschiceras* und anderen Tethys-Ammoniten in drei verschiedenen Niveaus des polnischen Valangin wird mit drei tethydischen Transgressionen in Verbindung gebracht, (1) an der Basis des Oberen Berrias, (2) an der Basis der *Platylenticeras*-Schichten und (3) an der Basis des Oberen Valangin (Verrucosum-Zone).

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1. Introduction

The section at Wąwał in central Poland can be regarded as a classic section of the Polish Neocomian. LEWIŃSKI's (1932) important paper "Das Neokom in Polen und seine paläogeographische Bedeutung" was chiefly based on sections in the region of Wąwał, and the clay-pit at this village still provides the sole exposure of Neocomian strata in Poland beyond the Carpathians. A new stratigraphic interpretation of the Wąwał section is given in the present paper, together with paleontological descriptions of some representatives of the genus *Karakaschiceras* important to connect Boreal and Tethyan realms.

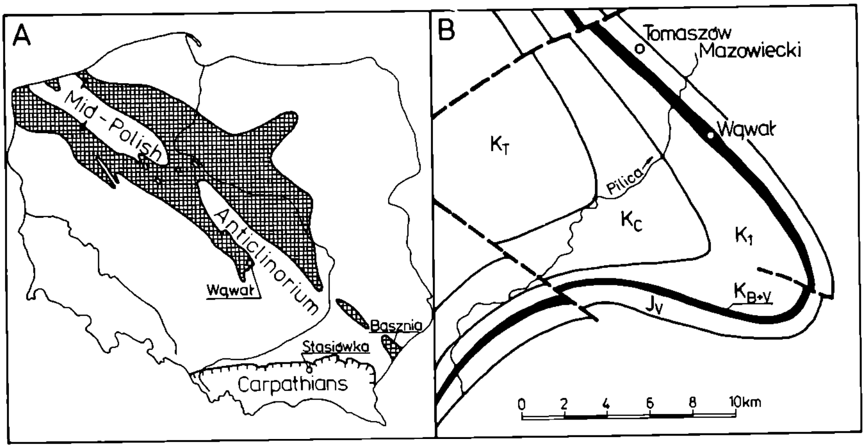
In this paper, the paleontological section has been written by J. WIEDMANN, its other sections by J. KUTEK and R. MARCINOWSKI.

Thanks are due to J. PRUSZKOWSKI for his data from Wąwał dealt with in this paper. Several specimens of *Karakaschiceras* here described have been offered to the authors by Prof. A. RADWAŃSKI, which is gratefully acknowledged. The authors are also indebted to Dr. E. PORĘBA for having made available unpublished data from the Tomaszów Syncline.

This study has been partly carried on within the Polish Project CPBP 03.04.

2. Available data

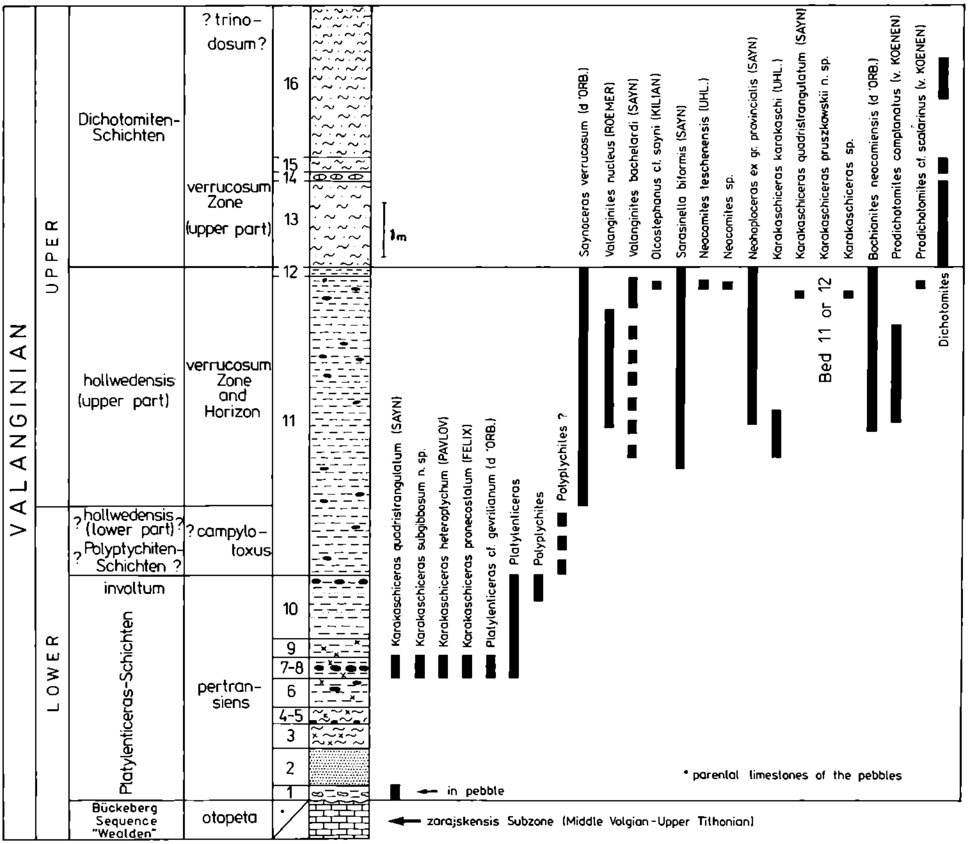
Wąwał is a village located a few kilometres southeast of the town of Tomaszów Mazowiecki on the Pilica River (Text-Fig. 1B). The Neocomian strata exposed at Wąwał belong to the northeastern limb of the Tomaszów (Tomaszów Mazowiecki) Syncline, a structure of the Mid-Polish Anticlinorium (Text-Fig. 1).



Text-Fig. 1. A: Distribution of Berriasian and Valanginian sediments in platformic Poland (modified after MAREK & RACZYŃSKA 1979a). - B: The Tomaszów Syncline (modified after MROZEK 1975); J_V - Volgian, K_{B+V} - Berriasian and Valanginian, K_l - higher Lower Cretaceous, K_C - Cenomanian, K_T - Turonian.

The clay-pit at Wąwał is still worked, but the lower part of the Neocomian section had been exposed here only for a few years, and the visible part of this section is now restricted to bed "11" and some higher strata (Text-Fig. 2). Therefore, the reinterpretation of the section presented below is largely based on evaluation of the data found in papers of previous authors (cf. Text-Fig. 3).

LEWIŃSKI's (1932) classic paper was based on data from exploring shafts carried out at Wąwał, and also at Nieborów, a village at the northern outskirts of Tomaszów Mazowiecki. Also the important paper by KOKOSZYŃSKA, published in 1956, chiefly refers to data from exploring shafts in the Wąwał region; her paper includes only a brief description of some upper part of the Neocomian section provided by the clay-pit of the Wąwał brickyard, then already in existence. A more complete section exposed in this pit has been described in a Master of Science paper by PRUSZKOWSKI (1962) and in a publication by WITKOWSKI (1969). Some additional borehole data can be found in papers by KOBYŁECKI (1948a, b).



Text-Fig. 2. Stratigraphic sequence and distribution of ammonites in the section at the Wąwał clay-pit. For lithology see Text-Fig. 3.

3. The lithologic sequence at Wąwał

The lithologic subdivision of the Neocomian section at the Wąwał clay-pit (Text-Fig. 2) is virtually that included in the unpublished paper by PRUSZKOWSKI (1962). This subdivision, in comparison to that in the paper published by WITKOWSKI (1969), allows for a more precise location of some ammonites in the section.

At the Wąwał clay-pit, Jurassic limestones of the Zarajskensis Subzone (lower Middle Volgian, lower Upper Tithonian) are disconformably overlain by a conglomeratic layer up to ca. 30 cm thick (bed "1", Text-Fig. 2). It consists of partly phosphatized calcareous pebbles and clayey-silty matrix. The pebbles, up to 5 cm in size, are very well polished and more or less well rounded. The pebbles consist of a micritic limestone with an admixture of silty quartz and, sporadically, also glauconitic grains. Some of the pebbles are bored by Tallophyta, indicating an age of the phosphatization of the pebbles posterior to the borings.

Because of a superficial lithologic similarity with the underlying Middle Volgian limestones, these pebbles were interpreted by earlier authors as derived from the Jurassic substrate. However, they display a different microfacies, and this is consistent with the finding of a post-Volgian *Karakaschiceras* in one of the pebbles (see below).

A simplified lithologic description of the succeeding Neocomian strata is given below (see Text-Fig. 2):

2. Clayey sand rich in bivalve shells. 0.8 m.
3. Clayey silt with ferruginous ooids. 0.5 m.
4. Silt with ferruginous ooids and concretions of limonite. 0.05 m.
5. Silt with ferruginous ooids. 0.3 m.
6. Clay with ferruginous ooids and red (hematized) sideritic nodules (clay-ironstones). 0.6 m.
7. Clay with ferruginous ooids and sideritic nodules (clay-ironstones). 0.15 m.
8. Red sideritic nodules (clay-ironstones) densely packed in clay with ferruginous ooids. 0.3 m.
9. Clay with ferruginous ooids. 0.35 m.
10. Clay with sandy streaks and numerous small bivalves; a characteristic layer of sideritic nodules at the top. 1.3 m.
11. Clay with scattered sideritic nodules. 6.25 m.
12. Clay with lenses of sand. 0.2 m.
13. Silt with numerous bivalves. 1.8 m.
14. A characteristic horizon of calcareous nodules, up to 20 cm thick. This horizon is also indicated in the section presented by KUBIATOWICZ (1983, tab. 3).
15. Silt with numerous bivalves. 0.3 m.

    The clay content of the interval encompassing PRUSZKOWSKI's beds "13" and "15" is stressed in the descriptions of the Wąwał section by WITKOWSKI (1969) and KUBIATOWICZ (1983).

16. The paper by PRUSZKOWSKI (op. cit.) includes a description of sediment overlying bed "15", over 1 m thick. Due to the fact that some younger Neocomian deposits were exposed at the Wąwał section after PRUSZKOWSKI's investigations had been finished, these sediments are described in the paper by WITKOWSKI (1969) and KUBIATOWICZ (1983) only. All

the Neocomian sediments overlying bed "15", about 5 m thick, are indicated in Text-Fig. 2 as bed "16". They are developed as clays and silts, and contain glauconite at some levels.

Several Neocomian ammonites were collected by LEWIŃSKI (1932) and KOKOSZYŃSKA (1956) in exploring shafts carried out at Wąwał, at distances of a few hundred metres from the clay-pit, and also at Tomaszów Mazowiecki. To some extent, it is possible to correlate lithologically the sections of those shafts with that of the Wąwał clay-pit. The best marker level, of which the boundary between beds "12" and "13" in the clay-pit is a part (Text-Figs. 2, 3), is at a boundary between clays and overlying silts and contains a characteristic bivalve assemblage; this lithologic boundary was taken as the boundary between the Valanginian and the Hauterivian by KOKOSZYŃSKA (1956) and WITKOWSKI (1969). The distribution of ammonites relative to that marker level, or to the base of the Neocomian, is also of some correlation value. Thus, it is possible to incorporate some of the paleontological data of LEWIŃSKI (1932) and KOKOSZYŃSKA (1956) into the section of the Wąwał clay-pit.

4. The ammonite sequence

The ammonites hitherto described from Wąwał are indicated in Text-Fig. 2. With the exception of the specimens of *Karakaschiceras* described and illustrated in this paper, these are mostly ammonites figured by KOKOSZYŃSKA (1956) and WITKOWSKI (1969); some of them have been renamed, or re-interpreted taxonomically (Table 1).

The distribution of the ammonites in the section needs some comment. A specimen of *Karakaschiceras quadristrangulatum* (SAYN) (Pl. 1, Fig. 4) was found by PRUSZKOWSKI (1962) in situ in a pebble, in the basal Neocomian conglomerate (this specimen was previously identified as *Neocomites* sp.) (cf. Text-Figs. 2, 3). A few other ammonites have been found as moulds or imprints of nuclei or small fragments of whorl, in other pebbles in the same conglomerate. All these specimens belong to Perisphinctaceae, but they cannot be identified precisely on a generic level; in any case, however, some of them do not represent *Karakaschiceras*. This gives some impression about the diversity of the ammonite assemblage in the parental limestones of the pebbles.

Specimens of *Platylenticeras* were found by LEWIŃSKI (1932: 261) in exploratory pits in three intervals (Text-Fig. 3). The lower interval (the third in LEWIŃSKI's section), which yielded *Platylenticeras* cf. *gevrilianum* (D'ORBIGNY), is ca. 0.40 m thick and consists of clays with ferruginous ooids and limonite concretions. This interval is comparable with the beds "7" and "8" of the clay-pit section. The next-higher interval "4", 0.80 m thick, consisting of clay with some sand and numerous small bivalves, also yielded *Platylenticeras* cf. *gevrilianum* (D'ORBIGNY), as well as *Polyptychites* sp. The interval "5", 2.25 m thick, is characterized as clay with some admixture of sand, containing numerous gypsum crystals and some nodules of clay-ironstone. The presence of numerous small bivalves and crushed ammonites with preserved shells is characteristic of the lower part of this interval; *Platylenticeras* cf. *gevrilianum* (D'ORB.), *P.* cf. *marcousianum* (D'ORB.) and *Polyptychites* sp. were collected here. A somewhat different development of the upper part of the interval "5" is indicated by LEWIŃSKI (op. cit.), who

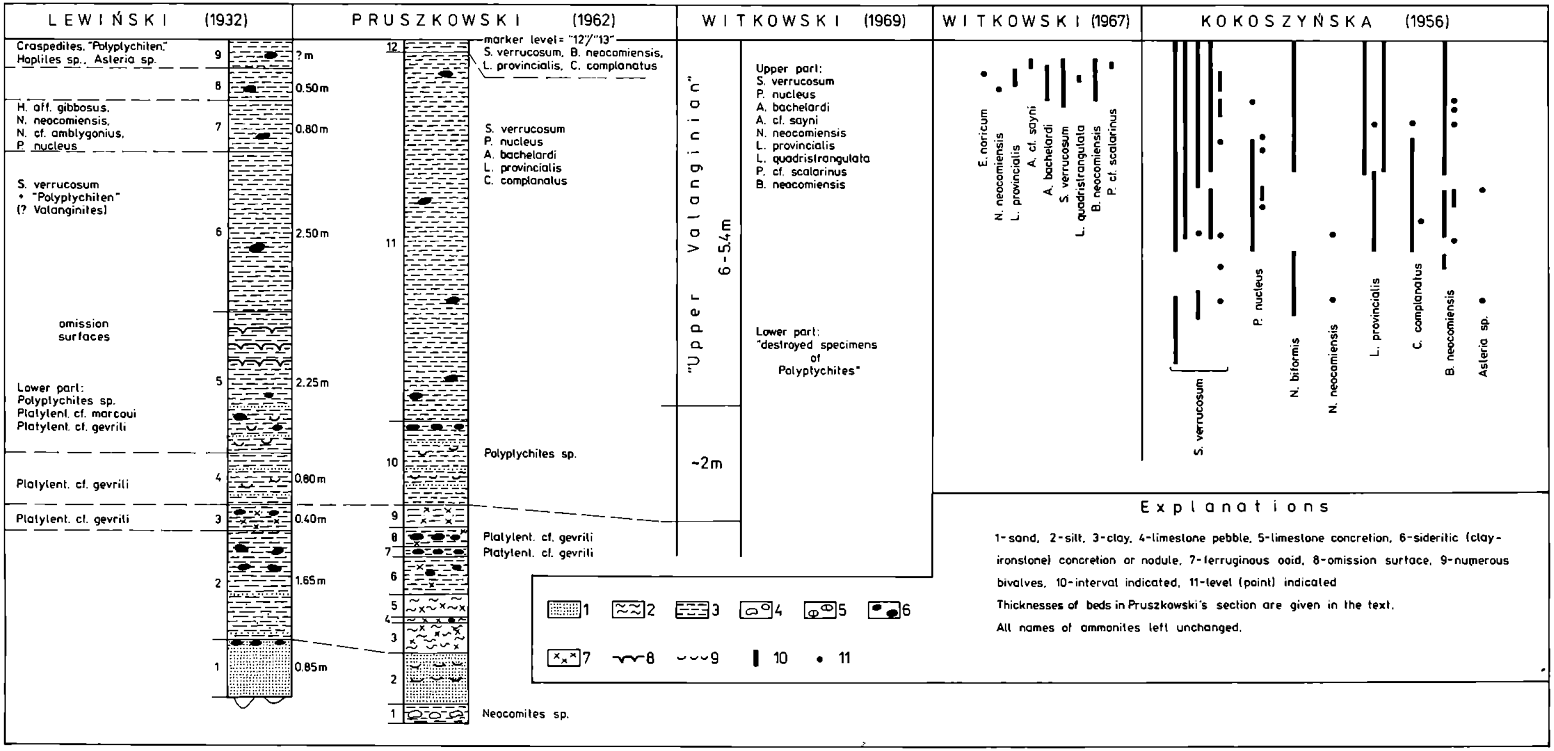
Table 1. List of ammonites revised and renamed.

KOKOSZYŃSKA 1956	WITKOWSKI 1969	THIS PAPER
Bochianites neocomiensis (d'Orb.) Pl. 4. figs. 15-16	Bochianites neocomiensis d'Orb. Pl. 19. fig. 4	Bochianites neocomiensis (d'Orb.)
Platylenticeras cf. gevrili (d'Orb.) Pl. 4. fig. 14	- - - - - - - -	Platylenticeras (Platylenticeras) cf. gevrilianum (d'Orb.)
Platylenticeras cf. marcoui (d'Orb.)	- - - - - - - -	Platylenticeras (Tolypeceras) cf. marcousianum (d'Orb.)
- - - - - - - -	Astieria cf. sayni Kil. Pl. 20. fig. 2	Olcostephanus cf. sayni (Kilian)
Saynoceras verrucosum (d'Orb.) Pl. 3. figs. 9-11	Saynoceras verrucosum (d'Orb.) Pl. 19. fig. 6	Saynoceras verrucosum (d'Orb.)
- - - - - - - -	Polyptychites cf. scalarinus Koen. Pl. 21. fig. 9	Prodichotomites cf. scalarinus (Koenen)
Polyptychites nucleus Roem. Pl. 4. figs. 12-13	Polyptychites cf. nucleus Roem. Pl. 21. figs. 7-8	Valanginites nucleus (Roemer)
- - - - - - - -	Astieria bachelardi Sayn Pl. 20. fig. 1	Valanginites bachelardi (Sayn)
- - - - - - - -	Dichotomites bidichotomus Leym. Pl. 22. figs. 3-7	Dichotomites evolutus Kemper
Craspedites complanatus Koen. Pl. 1. figs. 1-2	Craspedites complanatus Koen. Pl. 19. fig. 5	Prodichotomites complanatus (Koenen)
- - - - - - - -	Endemoceras noricum Roem. Pl. 21. figs. 2-4	Neocomites teschenensis (Uhlig)
- - - - - - - -	Endemoceras noricum Roem. Pl. 21. figs. 5-6	Neocomites sp.
Neocomites biformis Sayn Pl. 1. fig. 4	- - - - - - - -	Sarasinella biformis (Sayn)
Leopoldia (Hoplitides) provincialis Sayn Pl. 2. figs. 5-7; Pl. 3, fig. 8	Leopoldia provincialis Sayn Pl. 20. figs. 3-4	Neohoploceras gr. provinciale (Sayn)
Neocomites neocomiensis (d'Orb.) Pl. 1. fig. 3	- - - - - - - -	Karakaschiceras karakaschi (Uhlig)
- - - - - - - -	Neocomites neocomiensis d'Orb. Pl. 21. fig. 1 (left), (?right)	Karakaschiceras sp.
- - - - - - - -	Leopoldia quadristriangulata Sayn Pl. 20. figs. 5-6	Karakaschiceras gr. quadristrangulatum (Sayn)

also suggests the presence of two omission surfaces in that part of the interval. The bulk or lower part of LEWIŃSKI's interval "5" can be correlated with our bed "10", which is closely comparable lithologically, and in which *Polyptychites* sp. was recorded by PRUSZKOWSKI (1962). The highest part of LEWIŃSKI's interval "5" possibly corresponds to the lowest part of the bed "11" of the clay-pit section (cf. Text-Figs. 2, 3).

From the data presented above, the conclusion can be drawn that in other sections at Wąwał and Tomaszów Mazowiecki the recognized stratigraphic range of *Platylenticeras* encompasses an interval corresponding to the beds "7" - "9" of the clay-pit section, and a higher interval, corresponding to bed "10". In the latter interval *Platylenticeras* is associated with *Polyptychites* (Text-Fig. 2).

The specimens of *Platylenticeras* cf. *gevrilianum* (D'ORB.) and *P.* cf. *marcousianum* (D'ORB.) found by KOKOSZYŃSKA (1956) cannot be located in the Wąwał section with desired precision.



Text-Fig. 3. Correlation chart showing distribution of ammonites as indicated by different authors.

The specimens described in this paper as *Karakaschiceras heteroptychum* (PAVLOV), *K. subgibbosum* n. sp., *K. quadristrangulatum* (SAYN) and *K. pronecostatum* (FELIX) (Pl. 1, Figs. 1-3, 5; Pl. 2, Figs. 2-3) were found by Professor RADWAŃSKI in the Wąwał clay-pit, in spoil consisting of rock material characteristic of the lower part of the section, beneath bed "10". Moreover, the mode of preservation of these specimens precludes their location in any part of the section above bed "9", and in beds "1" - "2". On the other hand, their mode of preservation is closely comparable with that of the specimens of *Platylenticeras* found in beds "7" and "8". From this it follows that their position in the section falls most probably in the interval of beds "7" and "8" (Text-Fig. 2). Their location in a still lower part of the section (beds "6" - "3"), where no ammonites have been found up to date, cannot be totally excluded, but is much less justified.

The specimens of *Saynoceras verrucosum* (D'ORB.), *Neohoploceras* ex gr. *provinciale* (SAYN) and *Bochianites neocomiensis* (D'ORB.), found by PRUSZKOWSKI, are located with desirable precision in the thin bed "12". This is not the case with the thick bed "11", from which ammonites were not collected level by level in the Wąwał clay-pit. The paper by PRUSZKOWSKI (1962) provides information that *Saynoceras verrucosum* (D'ORB.), *Neohoploceras* ex gr. *provinciale* (SAYN), *Valanginites bachelardi* (SAYN), *Prodichotomites complanatus* (KOENEN) and *Bochianites neocomiensis* (D'ORB.) are present in bed "11", beneath bed "12". According to WITKOWSKI (1969: 22; names of ammonites unchanged), *Bochianites neocomiensis* D'ORB., *Craspedites complanatus* KOENEN, *Asteria bachelardi* SAYN (very common), *A.* cf. *sayni* KIL., *Leopoldia provincialis* SAYN, *L. quadristrangulata* SAYN, *Saynoceras verrucosum* SAYN (very common), *Polyptychites* cf. *nucleus* ROEM., *P.* cf. *scalarinus* KOEN., *Neocomites neocomiensis* D'ORB. and *Endemoceras (Lyticoceras) noricum* ROEM. occur in the upper part, and only rare, badly preserved ammonites of the genus *Polyptychites*, in the lower part of his Upper Valanginian (which comprises our bed "12" and the bulk of bed "11", with the exclusion of its lowest part, ca. 1 m thick; Text-Fig. 3).

An ammonite from PRUSZKOWSKI's collection, which is described in this paper as *Karakaschiceras pruszkowskii* n. sp. (Pl. 2, Fig. 1), comes from interval encompassing beds "11" and "12".

Some additional data on the distribution of ammonites in the discussed section are provided in the papers by LEWIŃSKI (1932) and KOKOSZYŃSKA (1956), and by an earlier paper by WITKOWSKI (1967). These data are compiled in Text-Fig. 3, and briefly reviewed below. The names of ammonites are left unchanged, when referring to the data of the authors.

Above LEWIŃSKI's (1932: 261-262) interval "5" mentioned above, there follow in his sections four successive clayey intervals that are closely comparable lithologically with our beds "11" and "12" (Text-Fig. 3). These are the intervals "6" (2.5 m), "7" (0.80 m), "8" (0.50 m) and "9"; unfortunately the thickness of interval "9" is not indicated. This may introduce some error into the correlations presented in Text-Fig. 3, because lithologically comparable sediments may have not exactly the same thickness in particular sections of the Wąwał region.

The following ammonites were found by LEWIŃSKI (1932):

"6". *Saynoceras verrucosum* D'ORB. and "numerous polyptychitids, badly preserved" (the last denomination may refer to *Valanginites*).

"7". *Hoplitides* aff. *gibbosus* v. KOENEN *(Karakaschiceras)*, *Neocomites neocomiensis* D'ORB. *(?Karakaschiceras)*, *N.* cf. *amblygonius* NEUM. & UHLIG (probably *Neocomites*), *Polyptychites nucleus* ROEMER *(Valanginites)*.

"8". No ammonites.
"9". *Craspedites* sp. *(Prodichotomites)*, *Hoplites* sp. *(?Neohoploceras, ?Karakaschiceras)*, *Astieria* sp. *(?Valanginites, ?Sarasinella)*.

The paper by KOKOSZYŃSKA (1956) makes it possible to establish the stratigraphic position at which several ammonites were found, relative to the marker level corresponding to the boundary of our beds "11" and "12". These data are indicated in Text-Fig. 3. The ammonites described by KOKOSZYŃSKA (op. cit.) as *Asteria* sp. probably belong to *Valanginites bachelardi* (SAYN), a species represented by a large number of specimens in the Wąwał collections.

Ammonites belonging to 10 species have been located with great accuracy by WITKOWSKI (1967, fig. 2) at the Wąwał clay-pit in an interval about 1 m thick, just below the same marker level (Text-Fig. 3). Two of these species, *Prodichotomites* cf. *scalarinus* (KOENEN) and *Olcostephanus* cf. *sayni* (KILIAN), are represented in WITKOWSKI's (1969) collection by one specimen each, thus these specimens are located with exceptional precision in the section.

The stratigraphic ranges of particular species indicated in Text-Fig. 2 are those that could be ascertained with reasonable certainty. In the case of several species, they undoubtedly represent only part of the total range of occurrence of these species in the Wąwał section. In this context it is noteworthy that *Saynoceras verrucosum* (D'ORBIGNY) is by far the most common ammonite species in the discussed part of the Wąwał section. KOKOSZYŃSKA (1956) collected 112 specimens of this species. WITKOWSKI's (1969) collection comprised 170 specimens, and, in addition, a great number of fragments of whorls. Moreover, *Saynoceras verrucosum* is already common at a relatively low stratigraphic level. For instance, KOKOSZYŃSKA (1956: 18) collected 27 specimens of this species in an interval 3.85 - 5.30 m below the marker level at the top of our bed "12". This may account for the fact that only the range of occurrence of *Saynoceras verrucosum* could be ascertained with some precision in a relatively lower part of bed "11".

The data compiled in Tables 1 and 3 indicate that the other species from beds "11" and "12" shown in Text-Fig. 2 occur within the stratigraphic range of *Saynoceras verrucosum*, and that this species ranges down to a relatively deep level within bed "11". Thus only an interval of about 1 m is left where polyptychitids not associated with *Saynoceras verrucosum* may occur in the lowest part of bed "11".

The presence of *Dichotomites bidichotomus* (LEYMERIE) was indicated by PRUSZKOWSKI (1962) in beds "13" and "15", and by WITKOWSKI (1967, 1969) in an interval 4 - 4.8 m above the marker level at the base of bed "13", and thus within our bed "16". At these levels there occur numerous ammonites, belonging mostly, if not exclusively, to the genus *Dichotomites* s. l. These ammonites are crushed, and it is very difficult to extract them undamaged from the soft rock (so that cores from boreholes seem to provide better specimens). For this reason not much is known about the paleontology of these dichotomitids. The marked contrast between the ammonite assemblage with *Saynoceras verrucosum* in beds "11" and "12" and the *Dichotomites* assemblage from the next-higher beds is worth note.

5. Systematic paleontology of the ammonite genus *Karakaschiceras*

Superfamily Perisphinctaceae STEINMANN 1890
Family Berriasellidae SPATH 1922
Subfamily Neocomitinae SPATH

Subfamily Neocomitinae SPATH is considered to include also subfamily Endemoceratinae SCHINDEWOLF 1966.

Genus *Karakaschiceras* THIEULOY 1971

Type species. *Hoplites biassalensis* KARAKASCH 1889, by original designation.

Discussion. The genus *Karakaschiceras*, proposed by THIEULOY (1971: 2299) constitutes an important late neocomitid lineage linking the Tethyan with the North and South(?) Temperate realms (KEMPER et al. 1981: 282). Morphologically, it is intermediate between *Neocomites* UHLIG s. str. and *Neohoploceras* SPATH, with a pronounced convergency towards *Leopoldia* MAYER-EYMAR. In early ontogeny (≈ the phragmocone), it has a hexangular whorl section with steep umbilical wall and flattened venter. With age (≈ living chamber), the venter becomes rounded. On the phragmocone the gently projected ribs multifurcate at different levels on the lateral sides. They start with umbilical bullae near the umbilical border and they disappear near a tubercle-like inflation on the marginal border, and thus do not cross the smooth siphonal band. Constrictions of variable number are more or less distinct at that stage. At present two subgroups can be distinguished with regard to the development of sculpture with age:

A. Group of *K. quadristrangulatum* (SAYN)
Small-sized, with persistent sculpture up to the adult(?), and more primary ribs than umbilical bullae.
B. Group of *K. biassalense* (KARAKASCH)
Large-sized, with sculpture weakening at an early stage, and - in some cases - finally disappearing. All primary ribs with umbilical bullae.

Not too much is known, however, about *Karakaschiceras* suture lines. BAUMBERGER (1906, figs. 30 - 33) figures external suture lines of *K. biassalense* (KAR.). This is the reason to reproduce on Text-Fig. 5 those parts of the external suture lines of the Polish specimens being visible. It can be summarized that the *Karakaschiceras* suture line is much denticulated, the lobe L is deep and asymmetric, E and U_2 are distinctly shorter. The saddles are asymmetrically subdivided. The umbilical suture line is curved backward. Thus, in general aspect, the *Karakaschiceras* suture line has much in common with that of *Endemoceras*.

Quite a number of neocomitid species are actually included in *Karakaschiceras* (KEMPER et al. 1981: 282). All of them were very near to one another and are, in most cases, linked by transitions. To make sure whether they are real species or merely variations of one or a few species, the study of variability in karakaschiceratids is needed.

Karakaschiceras is thought to represent a homogeneous stock which is restricted to the middle portion of the Valanginian (upper part of Zone of Thurmanniceras? campylotoxum and lower part of Zone of Saynoceras verrucosum).

The material described herein from Wąwał, Tomaszów Syncline, Central Poland, adds important new information about the systematic and stratigraphic extent of the genus, its origin and migration routes.

Systematically, two new species, *K. subgibbosum* n. sp. and *K. pruszkowskii* n. sp. can be added. Stratigraphically, the range of the genus can be extended to the lowermost Valanginian. This fact may imply that *Karakaschiceras* originated in Central Poland out of the neocomitid main stock, and spread from this area both to the North Temperate as well as to the Tethyan realms. Moreover, the origin of *Neohoploceras* SPATH can be traced back to species like *Karakaschiceras subgibbosum* n. sp.

A. Group of *K. quadristrangulatum* (SAYN)

Karakaschiceras quadristrangulatum (SAYN)
Pl. 1, Figs. 2, 4, 5; Text-Figs. 4A, 5B

1907 *Leopoldia quadristrangulata* SAYN, p. 56, pl. 3(7), fig. 21; pl. 5(9), fig. 20.
1962 *Sarasinella quadristrangulata* SAYN. - COLLIGNON, p. 49, pl. 193, figs. 879, 880 (non 881).
Sarasinella quadristrangulata SAYN. - COLLIGNON, p. 52, pl. 194, fig. 886.
? *Leopoldia biassalensis* KAR. - COLLIGNON, p. 52, pl. 194, fig. 888.
1969 *Leopoldia quadristriangulata* SAYN. - WITKOWSKI, p. 96, pl. 20, figs. 5, 6.
1981 *Karakaschiceras* cf. sp. b. - KEMPER et al., p. 286, pl. 41, figs. 12, 13.
? *Karakaschiceras* sp. b. - KEMPER et al., p. 286, pl. 41, figs. 1-3, 8, 9.

Holotype. The specimen figured by SAYN 1907, pl. 3(7), fig. 21 and pl. 5(9), fig. 20, from Beaumagne, SE France.

Material. 1 complete specimen (IGPUW/A/20/1) from a basal pebble in the conglomeratic bed "1" of Wąwał section (previously described by PRUSZKOWSKI 1962 as *Neocomites* sp. - cf. Text-Fig. 3), presumably equivalent to the Otopeta Zone. Fragmentary specimens from beds "7/8" equivalent to the Pertransiens Zone (IGPUW/A/20/2-3), all from Wąwał section, Central Poland (Text-Fig. 2). The specimens described by WITKOWSKI (1969) are from the Verrucosum Zone, bed "11" or "12".

Measurements.

	D	H	W	UD	H/W
IGPUW/A/20/1:	45 mm	23 mm (.51)	17 mm (.38)	9 mm (.20)	1.35
	about 20 primary ribs (PR) : 55 secondary ribs (SR).				
IGPUW/A/20/2:	ca. 60	28.5 (.48)	24 (.40)	ca. 15 (.25)	1.19
	about 18 PR : about 60 SR.				
IGPUW/A/20/3:	-	21	21	-	1.00
	about 18 PR : about 55 SR.				
Holotype in SAYN:					
	32	16 (.50)	13 (.41)	8 (.25)	1.23
	21 PR : 60 SR.				

	D	H	W	UD	H/W
COLLIGNON 1962, fig. 879:					
	41	22 (.54)	16 (.39)	8 (.20)	1.37
KEMPER et al. 1981, pl. 41, figs. 12, 13:					
	27	13.2 (.49)	11.5 (.43)	7 (.26)	1.15
	17 PR : 48 SR.				

Description. Small-sized(?) species with narrow umbilicus. Hexagonal whorl section higher than it is wide (Text-Fig. 4A), maximum width at the umbilicus. Umbilical wall steep, lateral sides slightly converging towards the flat venter.

Slightly biconcave and projected ribs which multifurcate at different levels of lateral sides, and some intercalated ribs. Most primary ribs are provided with umbilical bullae where they bi- or trifurcate. All ribs finish with small marginal tubercles; the venter remains smooth. Constrictions are irregularly spaced on the lateral sides. The ribs following these constrictions are more pronounced as are their umbilical bullae. About 20 primary ribs split into 50-60 secondary ones.

Discussion. *K. quadristrangulatum* is only known from smaller specimens and fragments. Whether these are adult or inner whorls only, is unknown due to the generally poor preservation. Up to the known maximum diameter of 60 mm, no smoothening of sculpture can be observed. This facilitates distinction from all members of the group of *K. biassalense* (KAR.). *K. heteroptychum* (PAVLOW) is similar to the present species, but the umbilicus is slightly larger, the whorl section as wide as it is high, ribbing is coarser and constrictions seem to be absent. *K. karakaschi* resembles the latter, but whorls are much higher than wide, and on some ribs lateral tuberculation can be observed. *K. subgibbosum* n. sp. can easily be separated due to its strong sculpture and inflated whorls.

Occurrence. *K. quadristrangulatum* is known from the middle part of the Valanginian from southern France (mainly Verrucosum Zone), from the same level of Madagascar ("Zone of Olcostephanus schenki"), and is doubtfully reported from the Hollwedensis Zone of northern Germany.

It is now described from the lowermost Valanginian Otopeta Zone(?), the Pertransiens Zone and the lower Verrucosum Zone of Wąwał, Central Poland. This is therefore the oldest known *Karakaschiceras* occurrence, and we can assume that Central Poland was the starting point and the migration center of all later karakaschiceratids.

Karakaschiceras heteroptychum (PAVLOW)
Pl. 1, Fig. 1; Text-Figs. 4B, 5E

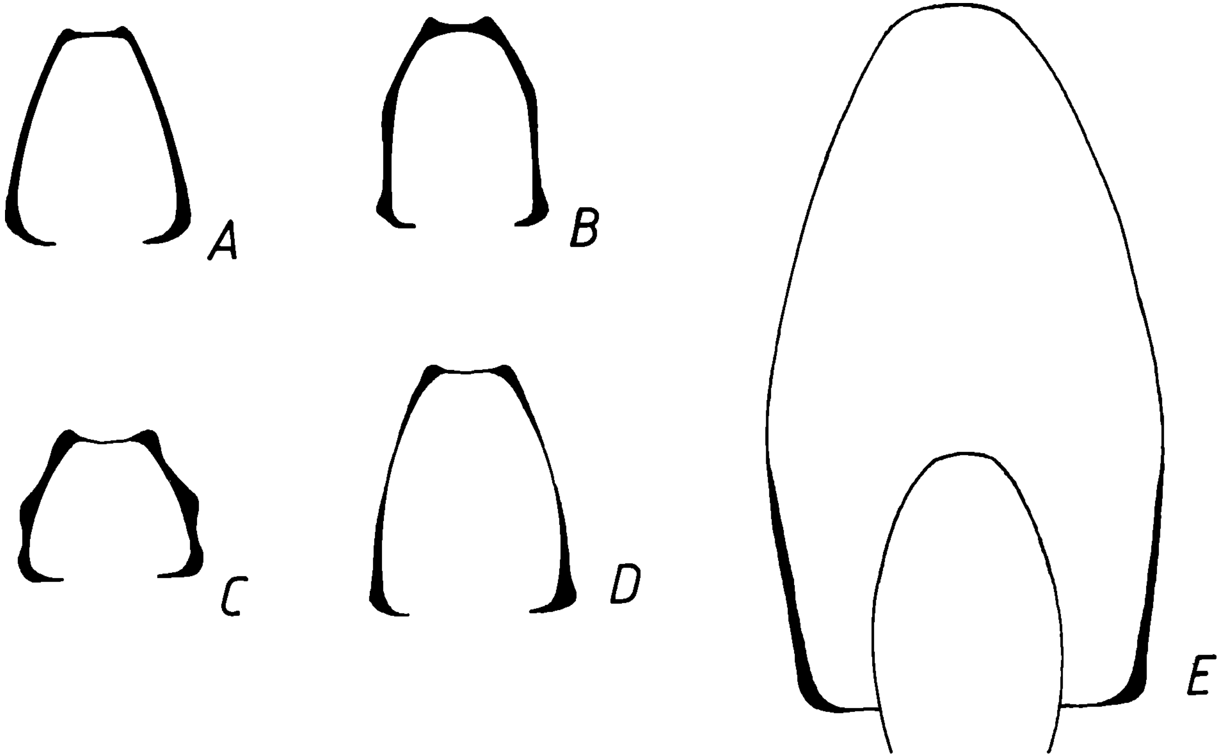
1892 *Hoplites heteroptychus* PAVLOW in PAVLOW & LAMPLUGH, p. 109, pl. 18(11), fig. 22.

non 1902 *Hoplites heteroptychus* PAVLOW. - v. KOENEN, p. 217, pl. 7, fig. 10.

non 1906 *Hoplites? heteroptychus* PAVLOW. - DANFORD, p. 106, pl. 14, fig. 2.

1981 *Karakaschiceras heteroptychum* (PAVLOW). - KEMPER et al., p. 284, pl. 41, figs. 10, 11, 16, 17.

1986 *Karakaschiceras heteroptychum* (PAVLOW). - KVANTALIANI & SAKHAROV, p. 65, pl. 2, fig. 11.



Text-Fig. 4. Whorl sections of A. *K. quadristrangulatum* (SAYN). IGPUW/A/20/2. - B. *K. heteroptychum* (PAVLOW). IGPUW/A/20/4. - C. *K. subgibbosum* n. sp. IGPUW/A/20/5. - D. *K. pronecostatum* (FELIX). IGPUW/A/20/6. - E. *K. pruszkowskii* n. sp. IGPUW/A/20/7. - All figures natural size.

Material. 1 specimen (IGPUW/A/20/4) from the Zone of Thurmanniceras pertransiens at Wąwał, Central Poland (Text-Fig. 2).

Holotype. The specimen figured by PAVLOW 1892, pl. 18(11), fig. 22, from Speeton, E England.

Measurements.

	D	H	W	UD	H/W
IGPUW/A/20/4:	60 mm	27 mm (.45)	>22 mm (.37)	17 mm (.28)	≈1.23
	about 20 PR : 60 SR.				
Holotype:	45	21 (.47)	19 (.42)	13.5 (.30)	1.11
	16 PR : 46 SR.				
KEMPER et al. 1981, pl. 41, figs. 16, 17:					
	95	36.5 (.39)	≈32 (.34)	30 (.32)	1.14
KVANTALIANI & SAKHAROV 1986, pl. 2, fig. 11:					
	32.5	14.3 (.44)	12.5 (.38)	10 (.31)	1.14

Description. Slightly evolute *Karakaschiceras* with inflated whorl section, nearly as wide as it is high. Lateral sides are parallel at first, and converge towards the flattened venter only on the outer part of lateral sides (Text-Fig. 4B). Ribs are multifurcating at different levels, or irregularly intercalated. About 20 primary ribs start at the umbilical border, half of them provided with umbilical bullae. They split into 60 secondary ribs which terminate in small marginal tubercles or clavi. Ribs do not cross the smooth venter. No distinct constrictions can be observed. Moreover, no weakening of sculpture can be seen up to the maximum diameter known of 100 mm.

Discussion. *K. heteroptychum* (PAVLOW) is morphologically near to *K. quadristrangulatum* (SAYN) and *K. karakaschi* (UHLIG). It can be distinguished only by minor characters, such as the whorl section, the umbilical width, course and strength of sculpture and absence of distinct constrictions.

Plate 1

Fig. 1. *Karakaschiceras heteroptychum* (PAVLOW). Hypotype IGPUW/A/20/4. Bed "7/8", Zone of Thurmanniceras pertransiens; Wąwał, Central Poland. A: lateral view, B: ventral view. 1/1.

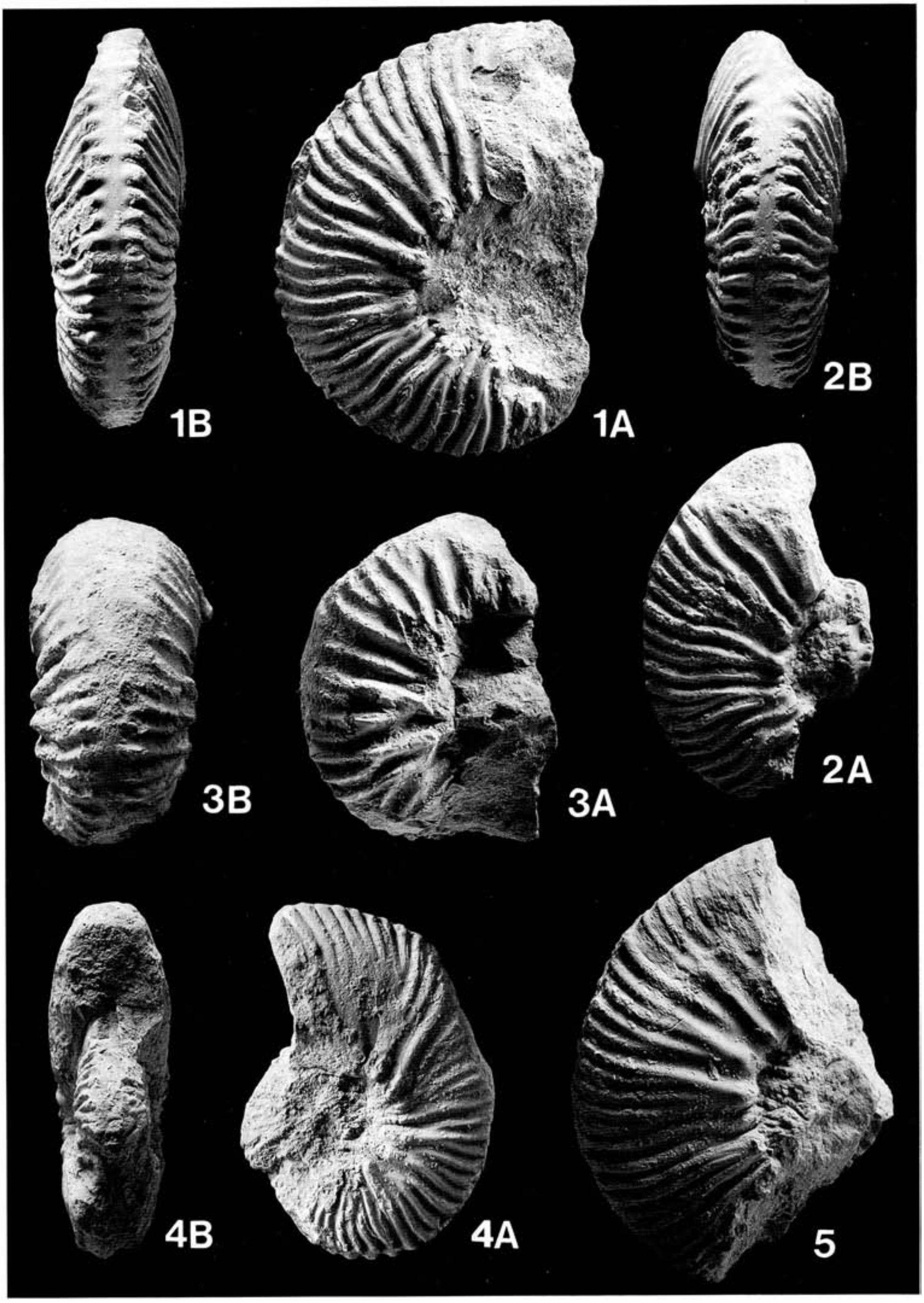
Figs. 2, 4, 5. *Karakaschiceras quadristrangulatum* (SAYN).

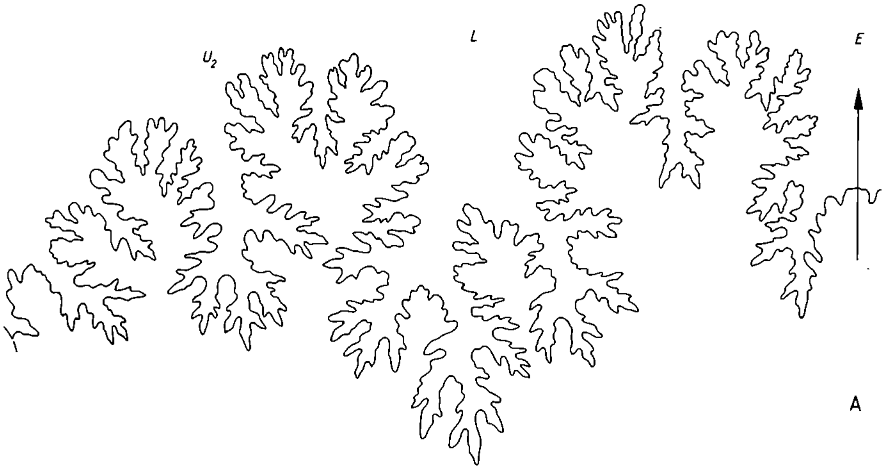
2. A, B: Hypotype IGPUW/A/20/3. Same horizon and locality. Lateral and ventral view. 1/1.

4. A, B: Hypotype IGPUW/A/20/1. Bed "1" (in pebble), Zone of Thurmanniceras otopeta; Wąwał, Central Poland. Lateral and frontal view. 1/1.

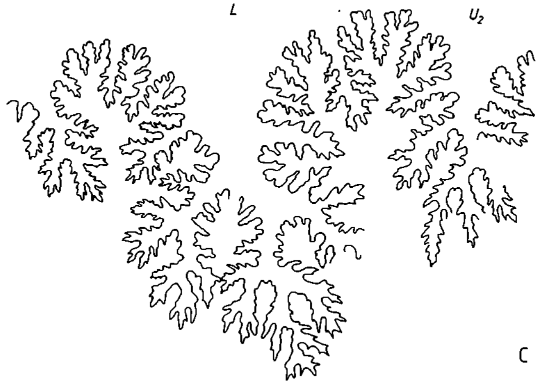
5. Hypotype IGPUW/A/20/2. Bed "7/8", Zone of Thurmanniceras pertransiens; Wąwał, Central Poland. Lateral view. 1/1.

Fig. 3. *Karakaschiceras subgibbosum* n. sp. Holotype, IGPUW/A/20/5. Bed "7/8", Zone of Thurmanniceras pertransiens; Wąwał, Central Poland. See also Pl. 2, Fig. 2. A: lateral view, B: ventral view. 1/1.

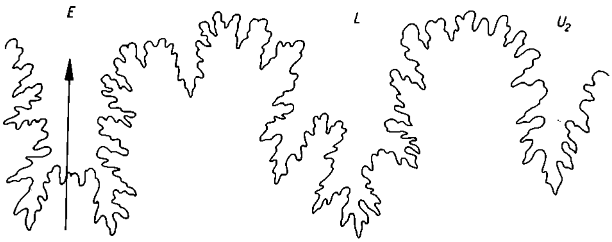

1B
1A
2B
3B
3A
2A
4B
4A
5



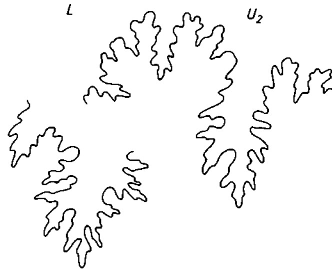
Text-Fig. 5. External suture-lines of A. *K. pronecostatum* (FELIX). IGPUW/A/20/6. At whorl height 25 mm, 5/1. - B. *K. quadristrangulatum* (SAYN). IGPUW/A/20/3. At whorl height 20 mm, 5/1. - C. *K. pruszkowskii* n. sp. IGPUW/A/20/7. At whorl height 50 mm, 3/1. - D. *K. subgibbosum* n. sp. IGPUW/A/20/5. At whorl height 20 mm, 5/1. - E. *K. heteroptychum* (PAVLOW). IGPUW/A/20/4. At whorl height 20 mm, 5/1.



C



D



E

O c c u r r e n c e. The present species is described from the Hollwedensis Zone of northern Germany and their equivalents in Yorkshire, England. It is also known from the Verrucosum Zone of the northern Caucasus, and is now described from the Pertransiens Zone of Lower Valanginian of Wąwał, Central Poland.

Karakaschiceras subgibbosum n. sp.
Pl. 1, Fig. 3; Pl. 2, Fig. 2; Text-Figs. 4C, 5D

?1907 *Leopoldia* aff. *L. gibbosa* v. KOENEN. - SAYN, p. 57, pl. 3(7), fig. 24.

H o l o t y p e. Specimen IGPUW/A/20/5, Zone of Thurmanniceras pertransiens at Wąwał, Central Poland (Text-Fig. 2).

M e a s u r e m e n t s.

	D	H	W	UD	H/W
IGPUW/A/20/5:	ca. 48 mm	19 mm (.40)	25 mm (.52)	12.5 mm (.26)	0.76
	about 20 PR : 40 SR.				

D i a g n o s i s. Inflated species with whorl width superior to whorl height. Sculpture consists of strong, subradiate and irregularly bifurcating ribs. 20 primary ribs start at the steep umbilical wall, half of them bearing umbilical bullae and bifurcating at these bullae. Ribs terminate at the marginal border forming tiny tubercles. Tendency to lateral tuberculation. No constrictions. Suture line unknown.

D e s c r i p t i o n. *K. subgibbosum* n. sp. belongs to the group of *K. quadristrangulatum* with persisting sculpture, and is the most inflated species of this group. The whorl section is broad-hexagonal with the maximum width near the center of lateral sides. These lateral sides only slightly converge towards the middle-sized umbilicus, but more pronounced towards the flattened and smooth venter (Text-Fig. 4C). The sculpture consists of strong, subradiate and irregularly bifurcating ribs, 20 of them starting at the umbilical border. Only 10 of these primary ribs bifurcate at strong umbilical bullae, the rest are untubercled and remain simple. There is a slight tendency to form lateral tubercles in some of the primary ribs. Altogether these ribs split into 40 secondary ribs which disappear at the outer margin with weak tuberculation. No constrictions can be observed.

D i s c u s s i o n. The present species can easily be distinguished from all other *Karakaschiceras* due to the inflated whorl section and strong sculpture. "*Leopoldia* aff. *L. gibbosa*" in SAYN (1907, pl. 3(7), fig. 24) may be identical with the new species.

Neohoploceras ambikyense COLLIGNON (1962, pl. 162, fig. 874) approaches likewise the present species due to its inflated whorl section; but it has more pronounced primary ribs on the inner part of the lateral sides, with umbilical and lateral tubercles, and has therefore to be included in *Neohoploceras* SPATH.

For morphological and stratigraphical reasons, an origin of *Neohoploceras* in karakaschiceratid species like *K. subgibbosum* can be assumed.

O c c u r r e n c e. *K. subgibbosum* n. sp. is here described from the Pertransiens Zone of Wąwał, Central Poland (Text-Fig. 2). SAYN's specimen from Arnayon, SE France, is doubtfully included in the present species.

B. Group of *K. biassalense* (KARAKASCH)

Karakaschiceras pronecostatum (FELIX)
Pl. 2, Fig. 3; Text-Figs. 4D, 5A

1860 *Ammonites neocomiensis* D'ORBIGNY. - PICTET & CAMPICHE, p. 247, pl. 33, figs. 1-3.
1891 *Hoplites pronecostatus* FELIX, p. 184.
1896 *Hoplites Leenhardti* KILIAN & ZÜRCHER, p. 996.
1901 *Hoplites Leenhardti*. - SARASIN & SCHÖNDELMAYER, p. 76, pl. 9, fig. 7.
1907 *Hoplites pronecostatus* FELIX. - KARAKASCH, p. 87, pl. 10, fig. 10; pl. 11, fig. 1.
1937 *Leopoldia biassalensis* (KAR.). - TZANKOV, p. 60, pl. 2, fig. 1 (non 3).
?1939 *Thurmanniceras* sp. ind. cf. *pronecostatus* (FELIX). - SPATH, p. 81, pl. 10, fig. 6.
?1962 *Leopoldia pronecostata* FELIX. - COLLIGNON, p. 52, pl. 194, figs. 889, 890.
1967 *Leopoldia pronecostata* (FELIX). - DIMITROVA, p. 126, pl. 58, fig. 2; pl. 62, fig. 2.

L e c t o t y p e (proposed herein)[1]. The specimen figured by PICTET & CAMPICHE 1860, pl. 33, fig. 1 from the Valanginian of Sainte-Croix.

M a t e r i a l. 1 chambered specimen (IGPUW/A/20/6), Zone of Thurmanniceras pertransiens, Wąwał, Central Poland (Text-Fig. 2).

M e a s u r e m e n t s.

	D	H		W		UD		H/W
IGPUW/A/20/6:	70 mm	33 mm	(.47)	27 mm	(.38)	16.5 mm	(.24)	1.22
Lectotype:	85	40	(.47)	29	(.33)	25	(.21)	1.39
	15 PR : 75 SR.							
KARAKASCH 1907, p. 87, spec. I:								
	81	39	(.48)	27	(.33)	22	(.27)	1.44

D e s c r i p t i o n. Large-sized *Karakaschiceras* of the group of *K. biassalense* (KAR.). *K. pronecostatum* has a middle-sized umbilicus and a high-oval whorl section. Whorls are much higher than wide. Maximum width of whorls is placed near the umbilical border (Text-Fig. 4D).

The sculpture consists of 15 primary ribs, all of them bearing umbilical bullae. The ribs multifurcate at different levels of lateral sides. Ribs are subradiate at first, but prorsiconcave on the outer lateral sides. They terminate with a strong projection towards the marginal border and with small marginal tubercles. The venter remains smooth. Primary and secondary ribs (about 60 per whorl) are primarily connected; at a shell diameter of 60 mm the sculpture disappears on the central portion of lateral sides, making primary and secondary ribs disconnected. No constrictions can be observed.

[1] DIMITROVA (1967: 126) proposed erroneously one of the specimens figured by KARAKASCH as type of the species.

D i s c u s s i o n. *K. pronecostatum* differs from the type of the genus and of the present subgroup, *K. biassalense* (KAR.), in preserving its sculpture up to a shell diameter of 60 mm. It also differs from the similar *K. pruszkowskii* n. sp. in whorl section and in a lower number of secondary ribs. *K. inostranzewi* (KAR.), however, has a much stronger sculpture and differing whorl section also, while *K. gibbosum* (v. KOENEN) and *K. brandesi* (v. KOENEN) differ in much rougher sculpture and umbilical width.

COLLIGNON's (1962, pl. 194, figs. 889, 890) Madagascan hypotype is different from the Russian type material (e. g. H/W = 1.71), and therefore only doubtfully included in the present species. Also the Salt Range specimen (SPATH 1939) is near to, but probably not identical with the present species.

O c c u r r e n c e. *K. pronecostatum* is a widespread species of late Lower and early Upper Valanginian ages; it is known from Crimea, Bulgaria[2], Switzerland, SE France, and with some hesitation from the Verrucosum Zone (= "Zone of Olcostephanus schenki") of Madagascar and the Salt Range. It is herein described from the Zone of Thurmanniceras pertransiens of Wąwał, Central Poland.

Karakaschiceras pruszkowskii n. sp.
Pl. 2, Fig. 1; Text-Figs. 4E, 5C

H o l o t y p e. Specimen IGPUW/A/20/7 from bed "11" or "12", Zone of Saynoceras verrucosum, Wąwał, Central Poland (Text-Fig. 2).

M a t e r i a l. The holotype.

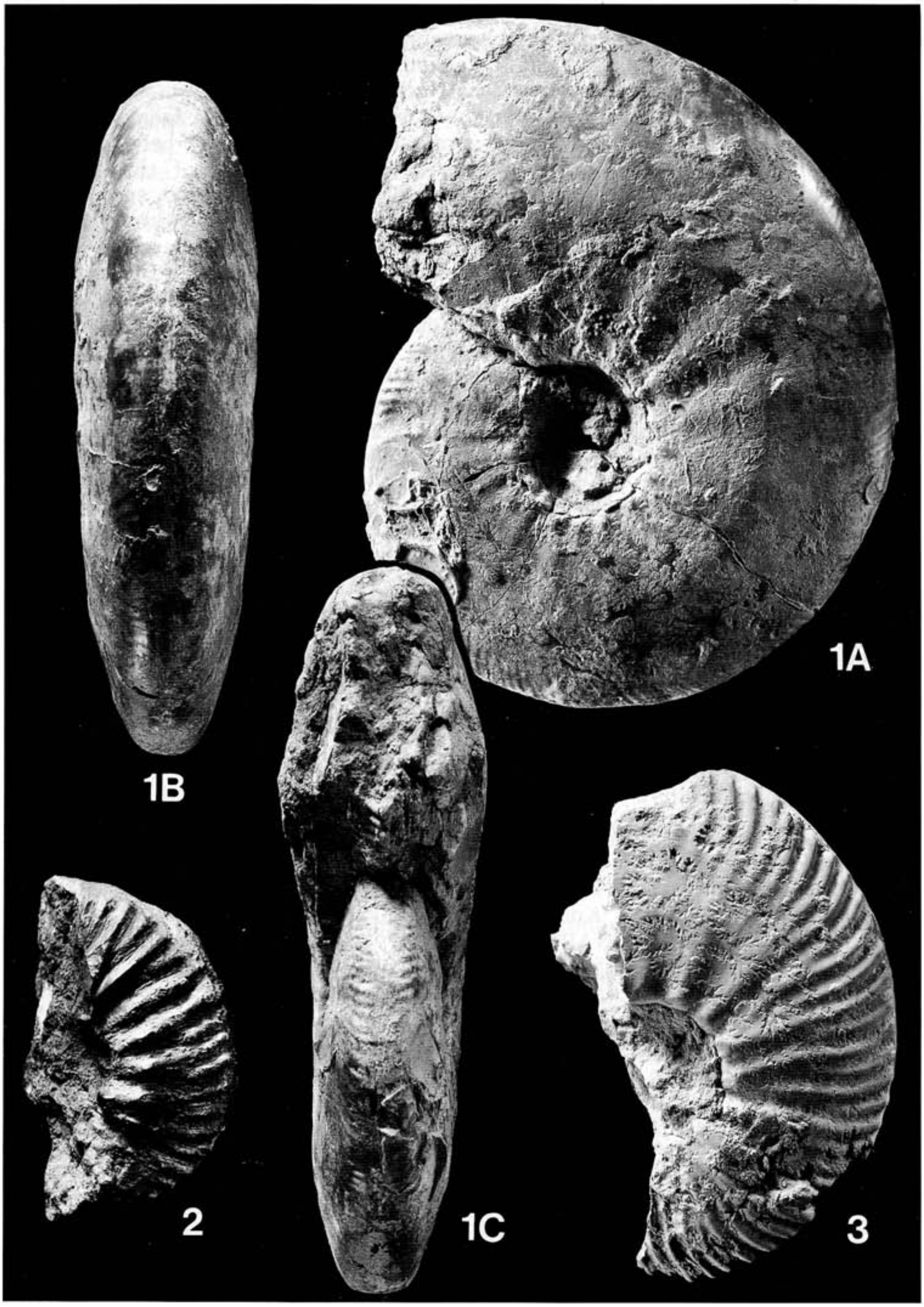
M e a s u r e m e n t s.

	D	H	W	UD	H/W
IGPUW/A/20/7:	196 mm	93.5 mm (.47)	54.3 mm (.28)	40 mm (.20)	1.72
	18 PR : 120 SR.				

[2] The Bulgarian hypotypes (DIMITROVA 1967) are erroneously attributed to the Lower Hauterivian.

Plate 2

Fig. 1. *Karakaschiceras pruszkowskii* n. sp. Holotype, IGPUW/A/20/7. Bed "11" or "12", Zone of Saynoceras verrucosum, Wąwał, Central Poland. A: lateral view, B: ventral view, C: frontal view. All 1/2.
Fig. 2. *Karakaschiceras subgibbosum* n. sp. Holotype, IGPUW/A/20/5. Bed "7/8", Zone of Thurmanniceras pertransiens, Wąwał, Central Poland. Lateral view. 1/1.
Fig. 3. *Karakaschiceras pronecostatum* (FELIX). Hypotype, IGPUW/A/20/6. Same level and locality. Lateral view. 1/1.


1A
1B
2
1C
3

Diagnosis. Large-sized *Karakaschiceras* of the *biassalense* group. Small umbilicus. Whorl sections subhexagonal, much higher than wide. Fine and dense ribbing (18 PR : 120 SR), weakening at early diameters. Flattened venter becomes rounded with age. Constrictions unknown.

Description. *K. pruszkowskii* n. sp. is known only from the holotype which is at a diameter of 196 mm still chambered. It belongs to the group of *K. biassalense* due to its pronounced smoothening of the sculpture at an early stage. The umbilicus is relatively small, the whorl section subhexagonal (Text-Fig. 4E). Whorl height is much higher than width; maximum whorl thickness is placed in the center of lateral sides where a kind of lateral keel can be observed. From this line the lateral sides converge towards the steep umbilical wall and towards the flattened venter. At a diameter of 125 mm, the venter becomes rounded.

The sculpture of the outer phragmocone consists of 18 projected primary ribs, all of them starting with small umbilical tubercles. Multifurcation of ribs on the lateral sides becomes progressively obsolete. Only the strongly projected, prorsiconcave outer secondary ribs can be observed on the last whorl and, finally, disappear (at a shell diameter of 150 mm). They form tiny marginal swellings and disappear on the flattened venter, leaving a smooth siphonal band in between. No constrictions can be seen.

Discussion. *K. pruszkowskii* n. sp. is easily distinguished from *K. biassalense* (KAR.) from which it differs by a slightly larger umbilicus, higher and differing whorl sections, and the longer persistance of its sculpture. In the density of ribbing it is also distinguishable from *K. pronecostatum, K. inostranzewi, K. gibbosum* and *K. brandesi.*

Occurrence. Up to now, *K. pruszkowskii* n. sp. is only known from bed "11" or "12", Zone of Saynoceras verrucosum, Wąwał, Central Poland.

6. Stratigraphic and geographic distribution of the genus *Karakaschiceras* THIEULOY 1971

The presence of the genus *Karakaschiceras* has been recognized in western Europe in the upper Lower and lower Upper Valanginian (BUSNARDO et al. 1979, KEMPER et al. 1981, COMPANY 1987). *Karakaschiceras* gr. *quadristrangulatum* (SAYN) and *Karakaschiceras* sp. from the Campylotoxus Zone of southeast France (BUSNARDO et al. 1979) are the earliest representatives of *Karakaschiceras* hitherto found in a well-recorded section in western Europe. *Karakaschiceras* was brought to northwest Germany and England by a transgression at the onset of the late Valanginian. All the records of this genus in Germany are from the Hollwedensis Zone (KEMPER et al. 1981).

However, the possibility cannot be ruled out that some earlier representatives of *Karakaschiceras* are present in some Tethyan sections in western Europe. For instance, in the Jura Mountains, *K. pronecostatum* (FELIX) has been found together with *Platylenticeras* and *Thurmanniceras,* but not precisely located stratigraphically in the Calcaire Roux, the stratigraphic range of which corresponds to the Pertransiens and Campylotoxus zones (BUSNARDO et al. 1979). Thus a *pertransiens* age of the genus *Karakaschiceras* from Calcaire Roux is a possibility. A more diversified assemblage of *Karakaschiceras,* comprising *K. biassalense* (KARAKASCH), *K. karakaschi* (UHLIG) and *K. pronecostatum* (FELIX) is reported from the "Marnes à *Astieria*", a Valanginian unit whose lithostratigraphic position in the Jura Mts. is higher than that of the Calcaire Roux.

In Bulgaria, *Karakaschiceras biassalense* (KARAKASCH) and *K. pronecostatum* (FELIX) were reported from the Lower Hauterivian (DIMITROVA 1967). The first species is precisely located in the Bulgarian lowest Hauterivian Meneghini-Cryptoceras Zone (MANDOV 1976). This zone, however, may correspond to the French latest Valanginian Callidiscus Zone (BUSNARDO et al. 1979).

In Romania, *Karakaschiceras biassalense* (KARAKASCH) and *K. inostranzevi* (KARAKASCH) have been collected from a middle part of the Valanginian in Dobrudja; the latter species has been found associated with *Saynoceras verrucosum* in the Banat region (Dr. E. AVRAM, pers. comm.).

Karakaschiceras seems to be a common genus in the Crimea highland, from where *K. biassalense* (KARAKASCH), *K. inostranzevi* (KARAKASCH), *K. pronecostatum* (FELIX) and *K. karakaschi* (UHLIG) have been reported (KARAKASCH 1889, 1905, 1907; DRUSHTCHIZ 1960). Their exact stratigraphic position relative to the Valanginian zonal schemes is not known. The combined data included in the papers by KARAKASCH (1907: 340) and DRUSHTCHIZ (1960: 62) indicate that, in the important Biassala section, specimens of *Karakaschiceras* occur as redeposited fossils in an Hauterivian conglomerate.

A recent paper by KVANTALIANI & SAKHAROV (1986) includes descriptions and illustrations of specimens referred to *Karakaschiceras heteroptychum* (PAVLOV) and *K. trezanense* (LORY). However, these specimens have been found, together with other Valanginian ammonites, redeposited at the base of the Hauterivian.

"*Leopoldia*" *biassalensis* KARAKASCH has been reported to occur in the Peri-Caspian Syneclise; however, its stratigraphic position is unclear (KOLTYNIN et al. 1986).

From this review it follows that the specimens of *Karakaschiceras* from the beds "1" and "7/8" of the Wąwał section, for which a pre-*campylotoxus* age is implied (see below and Text-Fig. 2), are the earliest representatives of the genus *Karakaschiceras* hitherto reported from well-recorded sections.

7. Biostratigraphic interpretation of the section

The biostratigraphic interpretation of the Neocomian section at Wąwał can be largely based on the Boreal zonal scheme of the Valanginian Stage as established in Lower Saxony, but for several reasons it is also important to interpret the section in terms of Tethyan Valanginian zones. A correlation chart, simplified after KEMPER et al. (1981), of the Boreal Valanginian zonal scheme of Northwest Germany and the Tethyan one established in Southeast France, is given in Table 2. For evidence for such a correlation, and other data, see THIEULOY (1977), BUSNARDO et al. (1979), and KEMPER et al. (1981); see also HOEDEMAEKER (1987).

The interval with *Saynoceras verrucosum* (D'ORBIGNY) at Wąwał (bed "12" and bed "11", with the exclusion of its lowest part), can be interpreted biostratigraphically with best precision. It can be equated with the Verrucosum Zone, and more precisely with the *verrucosum* Horizon at its base, to which the occurrences of *S. verrucosum* are restricted in French sections (the base of this zone and horizon is also, by definition, the base of the Upper Valanginian Substage). *Valanginites nucleus* (ROEMER) is another notable species restricted to the *verrucosum* Horizon in which there also occur some other species associated with *S. verrucosum* at Wąwał, such

Table 2. Correlation chart of the Tethyan and NW German subdivisions of the Valanginian Stage (simplified after KEMPER et al. 1981).

	TETHYAN ZONES	NORTH-WEST GERMANY ZONES	NORTH-WEST GERMANY
UP. VALANGINIAN	callidiscus		"ASTIERIEN"-SCHICHTEN
		tuberculata	"ARNOLDIEN"-SCHICHTEN
	trinodosum	bidichotomoides	DICHOTOMITEN-SCHICHTEN
		triptychoides	
		crassus	
	verrucosum	polytomus	
		hollwedensis	
LOWER VALANGINIAN	campylotoxum	sphaeroidalis	POLYPTYCHITEN-SCHICHTEN
		clarkei	
		multicostatus	
		pavlowi	
	petransiens	involutum	PLATYLENTICERAS-SCHICHTEN
		heteropleurum	
		robustum	
	otopeta		"WEALDEN" (PARS)

as *Neohoploceras provinciale* (SAYN), *Neocomites teschenensis* (UHLIG), *Sarasinella biformis* (SAYN) and *Bochianites neocomiensis* (D'ORBIGNY) (cf. THIEULOY 1977, BUSNARDO et al. 1979, HOEDEMAEKER 1984). The *Saynoceras* interval at Wąwał can also be correlated with the upper part of the Hollwedensis Zone of Northwest Germany (cf. THIEULOY 1977, KEMPER et al. 1981; see also Table 2).

The ammonite assemblage with *Saynoceras verrucosum* from Wąwał is clearly a Tethyan-related ammonite assemblage, including *Prodichotomites complanatus* (KOENEN) and *P.* cf. *scalarinus* (KOENEN) as the only notable Boreal faunal elements.

The strata with *Dichotomites* from Wąwał (beds "13 - 16" in Text-Fig. 2) should be correlated with some (probably some lower) part of the NW German *Dichotomites*-Schichten, with the exclusion of most of the Hollwedensis Zone. No precise stratigraphic conclusions can be drawn from the assignment (PRUSZKOWSKI 1962, WITKOWSKI 1969) of some dichotomitids from Wąwał to *Dichotomites bidichotomus* (LEYMERIE) (cf. KEMPER 1978: 201, 206). The specimen illustrated by WITKOWSKI (1969, pl. 22, fig. 3) as *D. bidichotomus* and reinterpreted by KEMPER (1978: 208) as *Dichotomites (Dichotomites) evolutus* KEMPER, and WITKOWSKI's (1969, pl. 22, fig. 2) specimen named *Polyptychites* sp. juv. *gradatus* KOENEN, which has been compared by KEMPER (1978: 208) also with *D. (D.) evolutus*, were found in boreholes in the region of Tomaszów Mazowiecki. *D. (D.) evolutus* is known to occur in the Hollwedensis, Polytomus and Crassus zones of Northwest Germany (KEMPER et al. 1981). No ammonites indicative of a later age have been found in the *Dichotomites*-bearing strata in the region of Wąwał and Tomaszów Mazowiecki. There is also no paleontological evidence for an Hauterivian age of any Neocomian strata exposed in the Wąwał clay-pit (the Hauterivian *Endemoceras noricum* (ROEM.) of WITKOWSKI 1969 = *Neocomites teschenensis* (UHL.) - see Table 1).

Because of the striking contrast between the Tethyan-related ammonite assemblage with *Saynoceras verrucosum* in beds "11" and "12", and the boreal *Dichotomites* assemblages in the immediately succeeding beds, the presence of a stratigraphic gap might be suspected at the junction of beds "12" and "13". There are some slight differences in the lithological development of beds "12" and "13", and a diversified bivalve assemblage with ostreids is characteristic of bed "13". Nevertheless, no evidence has been found for a significant stratigraphic gap at the discussed junction. The change in the ammonite assemblages should possibly be explained, in accordance with the suggestions expressed by HOEDEMAEKER (1984), as a biogeographical manifestation of a global sea level lowstand in late *verrucosum* time.

If there is no significant stratigraphic gap at the junction of beds "12" and "13", thus taking into account that *Saynoceras verrucosum* ranges up to the top of bed "12", some overlying strata must be included in the upper part of the Verrucosum Zone above the *verrucosum* Horizon. This part is characterized by the absence of *S. verrucosum* according to THIEULOY 1977 and BUSNARDO et al. 1979. Any evidence for the presence of the next-higher Trinodosum Zone in the highest Neocomian strata exposed in the Wąwał clay-pit is lacking, although such a possibility cannot be ruled out.

A correlative of the German *Polyptychites*-Schichten and the lower part of the Hollwedensis Zone must be restricted in the Wąwał section to an interval between the base of the strata with *Saynoceras verrucosum* and the top of the strata with *Platylenticeras* (Text-Fig. 2; see also KEMPER et al. 1981). Thus only an interval about 1 m thick, very poorly known paleontologically, is left in the section in the lowermost part of bed "11" (and perhaps also in the highest part of bed "10") to accomodate such a correlative. If there is a stratigraphic continuity in the section, this clayey interval should correspond to the four zones of the *Polyptychites*-Schichten (Pawlowi, Multicostatus, Clarkei and Sphaeroidalis) and the lower part of the Hollwedensis Zone (cf. Table 2). This makes it highly probable that there is a stratigraphic gap in the Wąwał section corresponding to a part, or the total, of the interval of the *Polyptychites* beds, a conclusion which finds strong support in borehole data from the Tomaszów Syncline (see below). As a consequence, also a total or partial absence of the Campylotoxus Zone should be accepted in the Wąwał section (cf. Text-Fig. 2, Table 2).

The interval comprising beds "7" - "10", where the genus *Platylenticeras* has been found, must correspond to some part of the German *Platylenticeras*-Schichten (which in turn, corresponds to the Pertransiens Zone - Table 2; THIEULOY 1977, BUSNARDO et al. 1979, KEMPER et al. 1981, HOEDEMAEKER 1987). The occurrence of *Platylenticeras* in association with *Polyptychites* in the interval corresponding to some upper part of bed "10" (Text-Fig. 2) is indicative of an *involutum* age for this interval (Table 2; KEMPER et al. 1981). An unexpectedly early age for the specimens of *Karakaschiceras pronecostatum* (FELIX), *K. quadristrangulatum* (SAYN), *K. heteroptychum* (PAVLOV) and *K. subgibbosum* n. sp. (here described and illustrated Pl. 1, Figs. 1-3, 5; Pl. 2, Figs. 2-3) is indicated by their position in an interval (beds "7/8") corresponding to the *Platylenticeras*-Schichten and the Pertransiens Zone (Text-Fig. 2).

No specimens of *Platylenticeras* from Wąwał and Tomaszów Mazowiecki have been identified with certainty at specific levels (cf. Text-Figs. 2, 3, Table 1, and MAREK et al. 1984b), and several specimens, which never have been described nor illustrated, were lost. Thus, the interpretation given below is subject to conjuncture. The occurrence of *Platylenticeras gevrilia-*

num (D'ORBIGNY) seems to be concentrated in the upper part of the Pertransiens Zone (THIEULOY 1977). This would suggest a relatively high position in the Pertransiens Zone of beds "7/8" at Wąwał, from which *P.* cf. *gevrilianum* was recorded. On the other hand, no precise stratigraphic interpretation can be based on LEWIŃSKI's (1932) specimens, referred to as *Platylenticeras* cf. *marcoui* (D'ORBIGNY), which were not identified on subspecies level. The specimens were found in LEWIŃSKI's interval "5", the position of which in the section, together with the occurrence of *Platylenticeras* in association with *Polyptychites* (Text-Figs. 2, 3), is consistent with an *involutum* age.

In the beds "1 - 6" of the Wąwał section no ammonites indicative of their age have been found. These beds can be correlated with some lower part of the German *Platylenticeras*-Schichten, and with some lower part of the Pertransiens Zone, on the following grounds. First, the lower limit of the stratigraphic range of the genus *Platylenticeras* in Poland has not been precisely recorded at any section, so there are no data precluding the interpretation advanced above. Second, it is commonly accepted that *Platylenticeras,* which appears at the base of the German *Platylenticeras*-Schichten, was brought to Northwest Germany from the Tethys via Poland as a result of an early Valanginian transgression. The basal conglomerate from Wąwał (bed "1" in Text-Fig. 2) is a manifestation of that very transgression. Third, a suitable interval must be left to accomodate the limestone which yielded the pebble with *Karakaschiceras quadristrangulatum* (SAYN) found in this conglomerate (Pl. 1, Fig. 4), and the stratigraphic gap inferred by the erosion of this limestone.

This interpretation, in turn, implies a pre-*pertransiens* age of that limestone. On the other hand, it can be inferred from regional data from Poland (see below) that this limestone must have been included in a marine sequence encompassing the Upper Berriasian and ranging up into the Valanginian. Thus, the age of the discussed limestone, and that of the specimen of *K. quadristrangulatum* found in the pebble in bed "1", must fall within an interval encompassing the Upper Berriasian, and the lowest Valanginian Otopeta Zone. An *otopeta* age is tentatively suggested for this ammonite (Text-Fig. 2), which, having been found in an unexpectedly low stratigraphic position, is now the earliest known representative of the genus *Karakaschiceras.* Thus, an assignment of this ammonite to the Upper Berriasian would mean a yet stronger enlargement of the hitherto known time range of this genus.

8. The Berriasian and Valanginian in the Tomaszów Syncline

Most of the data concerning the Neocomian of the Tomaszów Mazowiecki Syncline come from boreholes, the Wąwał pit being the sole exposure of the Neocomian sediments of the region (Text-Fig. 1B). Several data can be found in a large paper by WITKOWSKI (1969) and in a recent paper by PORĘBA (1987); some unpublished data from several boreholes were kindly made available to the authors by Dr. E. PORĘBA.

The Neocomian sediments of the Tomaszów Syncline were penetrated by a rather large number of wholly-cored boreholes, but the stratigraphic interpretation of particular sections with desirable precision is often difficult. The Neocomian sequences are highly variable, and their biostratigraphic dating is poor; ammonites have only been found at some levels in a few boreholes. Moreover, some stratigraphic correlations in the paper by WIT-

KOWSKI (1969), which were based on the assumption of stratigraphic continuity within several Berriasian and Valanginian sequences, and of the presence of Berriasian strata in the Wąwał section, need reinterpretation. Nevertheless, the following conclusions can be drawn from available evidence:

1. The Neocomian sediments are underlain by Middle Volgian limestones (upper part of the Zarajskensis Subzone) in some part of the Tomaszów Syncline, and by Middle Volgian shales (lower part of the Zarajskensis Subzone) in other parts (KUTEK & WITKOWSKI 1963, WITKOWSKI 1969, KUTEK & ZEISS 1974). These Middle Volgian strata correspond to some lower part of the Upper Tithonian (KUTEK & ZEISS 1974, ZEISS 1983, KUTEK & WIERZBOWSKI 1986).

2. Ammonites representing the genera *Riasanites, Neocosmoceras,* and possibly also *Dalmasiceras* have been found in three boreholes (WITKOWSKI 1969), providing evidence for a late Berriasian age of the lowest marine Neocomian sediments in some parts of the Tomaszów Syncline. The sediments ascribed to the Berriasian are only a few meters thick. Their lower part at least is largely developed as shales, but in some sections there is a thin basal conglomerate consisting of limestone pebbles. The presence of an erosional discontinuity at the top of the Berriasian sediments in the south-western part of the syncline has been pointed out by WITKOWSKI (1969).

3. Calc-dolomitic silts have been encountered in a few boreholes as thin intercalations between marine Middle Volgian and marine Neocomian sediments (WITKOWSKI 1969). The age of these silts, which yielded some sporomorphs, but no marine fossils, is difficult to assess.

4. No Neocomian conglomerates have been encountered in some of the sections provided by boreholes, but they have been found at the base of Neocomian sequences in several other sections (WITKOWSKI 1969, PORĘBA 1987). In some cases, there is reasonably good indirect evidence for their Berriasian age (WITKOWSKI 1969); in others, the lithostratigraphic context rather suggests an early Valanginian age of such conglomerates. Significantly, in some other boreholes, single conglomeratic layers appear within the Neocomian sequences, up to 4 m above their base, and, finally, in still other sections there are two conglomerates: a lower one at the base of, and a higher one within the Neocomian sequence (PORĘBA 1987). This makes it possible to interpret the lower conglomerates as parts of basal late Berriasian conglomerate and the higher conglomerates as forming a basal conglomerate of the *Platylenticeras* Beds, equivalent to bed "1" of the Wąwał section (cf. Text-Fig. 2).

5. So far, no limestones bearing Neocomian ammonites have been found in the Tomaszów Syncline. However, layers of limestones, some dozens of centimetres thick, have been encountered within Neocomian sediments near their base in some boreholes, and the earliest strata penetrated by some other boreholes are developed as grey limestones and marls, at least a few metres thick (Dr. PORĘBA, pers. comm.). In some of the latter sections, the top surface of the limestones is clearly erosional, and capped by a sedimentary breccia consisting of fragments of the underlying rock. Such limestones can be interpreted as the parental rock for the pebbles occurring in the conglomeratic bed "1" of the Wąwał section (cf. Text-Fig. 2).

6. In large parts of the Tomaszów Syncline the Jurassic rocks are directly overlain by an Upper Valanginian sequence, the Berriasian and Lower Valanginian sediments occurring there only as a few scattered relics (PO-

RĘBA 1987). This picture is consistent with the concept of a stratigraphic gap corresponding to a part, or the total, of the *Polyptychites* Beds, as suggested by the Wąwał section.

Thus there is evidence for three stratigraphic gaps in the Tomaszów Syncline, connected with slight unconformities: one at the base of the late Berriasian marine sediments, the second at the base of the *Platylenticeras* Beds (Pertransiens Zone), and the third corresponding to the *Polyptychites* Beds (Campylotoxus Zone).

The variable, discontinuous and very thin (up to 20 m - PORĘBA 1987) Berriasian to Upper Valanginian sequences of the Tomaszów Syncline are suggestive of slight rifting, coupled with oscillatory crust movements and/or oscillations of the global sea-level. Such a picture is consistent with the position of the Tomaszów Syncline within the Mesozoic Mid-Polish Rift, and, on the other hand, on the southern slope of the Polish extension of the Central-European Basin.

In the Tomaszów Syncline, a discontinuous and poorly dated early Cretaceous sequence, some 150 m thick, is intercalated between Valanginian and Upper Albian strata (Text-Fig. 1B, see also PORĘBA 1987). This sequence will not be discussed in this paper.

9. Tectonic framework of Neocomian paleogeography and sedimentation

During the Mesozoic, the Central-European (Northwest-European) Basin extended from the North Sea area eastward across North Germany, including a large part of platformic Poland. This basin was separated from the Carpathian basins by an arch, called the Meta-Carpathian Arch (KUTEK & GŁAZEK 1972). At some times this arch formed a paleogeographical barrier; at other times, it was only individualized as a zone of lesser subsidence. During its Triassic development, this arch can be regarded as a part of the well-known Beskid-Vindelician Arch.

A notable feature in Poland's regional geology is the Mid-Polish Anticlinorium, which extends from the Baltic Sea southeast across Poland and Western Ukraine, its southernmost part plunging beneath the Carpathian nappes (Text-Fig. 1A). This anticlinorium is the result of the Laramide inversion of the Mid-Polish Rift (POŻARYSKI & BROCHWICZ-LEWINSKI 1978). The north-western part of this rift was in existence at least from the Permian (as an axis of subsidence) and the Triassic (as a fault-bounded structure). However, the development of the southeastern part of the Mid-Polish Rift (situated southeast from the present Holy Cross Mountains, more precisely, southeast of the Holy Cross Lineament) dates only back to the late Middle Jurassic (KUTEK 1980). This coincides with strong Middle and Upper Jurassic distension tectonics in the Carpathian region, so that the southeastern part of the Mid-Polish Rift can be interpreted as an aborted branch of the Jurassic Carpathian rift system.

During the Mesozoic, the axis of the Meta-Carpathian Arch was usually situated south of the present northern front of the Carpathian nappes; however, there was a notable exception in the late Jurassic and early Cretaceous. In the late Middle Jurassic, the position of the axis of this arch was in Poland about the same as that of the front of the present-day Carpathian nappes. However, further development of this arch was interrupted for some time in the late Jurassic as a result of strong downwarping of the Carpathian foreland. Oxfordian and Lower Kimmeridgian sediments over 1 km thick accumulated in this region in southern Poland, attaining a significantly larger thickness than that of the Polish Lowland. This area, at the crossing of the Mid-Polish Rift with the axis of the Central-European Basin, was usually the site of strongest subsidence in platformic Poland during the Mesozoic.

The Meta-Carpathian Arch re-appeared towards the end of the Jurassic and was clearly individualized in the Neocomian, this time, however, in an unusual northern position, its axis being situated in the region of the Holy Cross Mountains. The arch acted then as the "Vistula Swell", the development of which has been depicted in an excellent paper by CIEŚLINSKI (1976). Such tectonic conditions virtually persisted to the Albian.

The present-day distribution of early Cretaceous and, in particular, Berriasian and Valanginian sediments in platformic Poland (Text-Fig. 1A) is a consequence of the tectonic pattern outlined above. Significantly, relics of Neocomian sediments are preserved in Southern Poland only in the border zones of the Mid-Polish Rift. On the other hand, continuous or nearly continuous Berriasian and Valanginian sequences attain thicknesses over 300 m in Central and Northern Poland (MAREK 1983a), within the Central-European Basin. In the region of the Tomaszów Syncline, which corresponds to the northern slope of the Meta-Carpathian Arch (the Vistula Swell), the thicknesses of discontinuous Berriasian to Valanginian sequences are less than 20 m (PORĘBA 1987). Farther south no Neocomian sediments are preserved, and still farther south, in regions corresponding to the southern slope of the arch, there occur thin sequences representing but a few stratigraphic members of the Neocomian.

The following paleogeographic and biogeographic consequences can be deduced from this paleotectonic pattern:

1. The existence of the Central-European Basin could facilitate faunal exchange between its subbasins, e. g. between Northwest Germany and Central Poland.

2. The Mid-Polish Rift provided for marine connections between the Carpathian Tethys and the Central-European Basin in Neocomian time, enabling faunal migrations from Mediterranean regions into that basin. It is hard to say how much successive Neocomian transgressions overstepped the borders of the Mid-Polish Rift where it crossed the Meta-Carpathian Arch in Southern Poland and Western Ukraine. On the other hand, intermittent slight uplifts of this arch or falls of the global sea-level, may have resulted in removal of previously accumulated Neocomian sediments, and in severing faunal migrations.

3. The Mid-Polish Rift had its southeastern extension in the sedimentary region of the Eastern Carpathians (for details see POŻARYSKI & ŻYTKO 1980). This makes probable some faunal exchange between the Polish part of the Central-European Basin and the basins of Crimea and Caucasus. In this context it may be of significance that the genus *Riasanites*, known from the Russian Platform but also from the Crimea and Caucasus, seems to be quite frequent in the marine Berriasian of the Polish Lowland. The statement in the paper by GORBACHIK & KUZNETSOVA (1984: 134) that the Berriasian foraminiferal assemblages from the Polish Lowland have constituents also found in assemblages from the Crimea and Caucasus, points in the same direction.

10. Tithonian to Valanginian sedimentation and ammonite succession in platformic Poland

Comprehensive information about the Tithonian (Volgian), Berriasian and Valanginian stages in Poland can be found in papers by DEMBOWSKA (1973), DEMBOWSKA & MAREK (1975), MAREK & RACZYŃSKA (1973, 1979a) and MAREK (1983a), as well as in the large volume of "Geology of Poland" - Stratigraphy, Mesozoic (SOKOŁOWSKI, Ed., 1976). Below, only some data relevant to the subject of this paper will be pointed out. It should be borne in mind that all the data reviewed and commented below have been obtained from boreholes, except for some outcrops of Volgian strata and the Valanginian sediments near Wąwał. The Tithonian to Volgian sediments occurring in Central and Northern Poland and thus within the Central-European Basin and its peripheries, will be reviewed first.

The latest Jurassic ammonites have been found in platformic Poland in strata which represent the Zarajskensis Subzone (the upper subzone of the Scythicus Zone, the lowest zone of the Middle Volgian Substage), and thus some lower part of the Upper Tithonian (KUTEK et al. 1973, KUTEK & ZEISS 1974, KUTEK & WIERZBOWSKI 1986). There is good evidence for faunal exchange with some peri-Tethyan regions persisting as late as *zarajskensis* time (KUTEK & ZEISS 1974, KUBIATOWICZ 1983, KUTEK in press).

In large parts of the Polish Lowland these Middle Volgian strata are overlain by a sequence without ammonites, which straddles the Jurassic/Cretaceous boundary (DEMBOWSKA 1973, MAREK et al. 1984a). It is subdivided into the ostracod zones A-F, lettered from top to bottom. This sequence begins with marine sediments, which are followed by sediments containing sulphates (gypsum and anhydrite), then by brackish and fresh-water sediments, and, finally, by marine-brackish sediments (zone A). The evaporitic event can be explained by a slight uplift of the Meta-Carpathian Arch, inducing evaporitic conditions in the Polish part of the shallow Central-European Basin as a result of restricted water circulation with the oceanic Tethys. Further uplift of the arch resulted in fresh-water sedimentation, and its subsequent downwarping in a return to brackish and marine conditions. Such an interpretation is consistent with independent stratigraphic and structural evidence from Poland. To some extent, the sequence of events could also be influenced by global sea-level changes.

The sequence A-F is overlain in the Polish Lowland by a marine sequence which is usually referred to as the Upper or marine Berriasian (MAREK & RACZYŃSKA 1979a, MAREK 1983a, MAREK et al. 1984a). This sequence yielded some ammonites which have been referred to Boreal genera such as *Surites, Subcraspedites* and *Praetollia* (MAREK 1967, 1983a, DEMBOWSKA & MAREK 1975, MAREK & RACZYŃSKA 1979a, MAREK & RAJSKA in MAREK et al. 1984b), but whose taxonomic status is still uncertain or debatable (see MAREK 1967: 160-161; CASEY 1973: 214; SHULGINA 1985: 26). Furthermore, this sequence yielded in addition to *Riasanites,* also several clearly Mediterranean ammonites, such as *Berriasella, Fauriella, Euthymiceras, Malbosiceras* and *Neocosmoceras* (MAREK 1967, 1983a, DEMBOWSKA & MAREK 1975, MAREK & RACZYŃSKA 1973, 1979a, MAREK & RAJSKA in MAREK et al. 1984b). The discussed sequence is usually equated (MAREK & RACZYŃSKA 1979a, MAREK 1983a, MAREK et al. 1984a) with the total of the Upper Berriasian Boissieri Zone sensu LE HEGARAT (1971), but, as remarked by HOEDEMAEKER (1987), the

Mediterranean ammonites found in this sequence are only indicative of the Paramimouna and Picteti Subzones (the two lower subzones of the Boissieri Zone, the Callisto Subzone being its highest subzone). In several parts of the Polish Lowland there is evidence for an unconformity at the base of the marine Berriasian sediments, these resting on different members of the sequence A-F, or on older rocks (see e. g. DEMBOWSKA & MAREK 1975, fig. 2).

The Valanginian sediments of the Polish Lowland are subdivided into the Lower Valanginian *Platylenticeras* Beds and *Polyptychites* Beds, and the Upper Valanginian Beds with *Dichotomites* and *Saynoceras* (MAREK 1983a). This subdivision is based on paleontological evidence, but in several regions the boundaries of these three Valanginian units may not correspond strictly to biostratigraphic boundaries.

The *Platylenticeras* Beds, which are also called the Beds with *Platylenticeras* and *Neocomites* (MAREK et al. 1984a), yielded *Platylenticeras*. In addition, *Polyptychites* and the belemnite genus *Acroteuthis* have been found in the upper part of these beds; the latter genus has also been found in the lowest parts of intervals ascribed to the *Polyptychites* Beds (BŁASZKIEWICZ 1963, MAREK 1968, 1969, MAREK & RACZYŃSKA 1979a). There is no doubt that a large part of the *Polyptychites* Beds in the Polish Lowland is coeval with the *Platylenticeras*-Schichten of Northwest Germany and thus with the Pertransiens Zone sensu BUSNARDO et al. 1979 (see KEMPER et al. 1981, HOEDEMAEKER 1987). However, it is highly probable that in some regions of the Polish Lowland the sediments ascribed to the *Platylenticeras* Beds range down to a deeper stratigraphic level than the Northwest German *Platylenticeras*-Schichten, thus including strata referable to the lowest Valanginian Otopeta Zone or even to the latest Berriasian Callisto Subzone.

An example supporting this interpretation is provided by data from the Wypychów T-75 borehole (cf. MAREK 1967, 1969). Here, in Central Poland, in a continuous marine sequence (*sinuata* Shales), a specimen of *Riasanites rjasanensis* has been found at the depth 82.9 m, and a specimen referred to as *Neocomites neocomiensis* D'ORB. var. *premolica* SAYN at depth 82.0 m; the base of the *Platylenticeras* Beds has been drawn between these two levels. A sedimentary unit known to contain *Platylenticeras* begins in the section at a level over 30 m higher than those at which these ammonites have been found. In Tethyan regions *Neocomites premolicus* appears near the base of the Alpillensis Subzone (which is also the top of the Picteti Subzone) and ranges up into the Trezanensis-Pexiptychum Subzone (HOEDEMAEKER 1982, 1983). Using, instead of HOEDEMAEKER's zonal scheme, those of LE HEGARAT (1971) and THIEULOY (1977), it can be said that *N. premolicus* ranges from the Callisto Subzone into the Pertransiens Zone (cf. HOEDEMAEKER 1982, 1987). In this context HOEDEMAEKER's (1987) suggestion is worth remembering, that the known occurrences of the genus *Riasanites* are restricted to intervals beneath a level corresponding to the top of the Picteti Subzone.

No intra-Neocomian stratigraphic gaps at the base of the *Platylenticeras* Beds have hitherto been recognized in the Polish Lowland beyond the Tomaszów Syncline but there are many instances of a larger geographic extent of these beds relative to the "Upper Berriasian" in several regions (MAREK 1977, 1983a, b, RACZYŃSKA 1987). In general, available evidence is consistent with the concept of a transgression at the beginning of *Platylenticeras* time (in the sense of the time span corresponding to the *Platylenticeras*-Schichten of Northwest Germany), following regressive tendencies and erosion in some parts of platformic Poland in earliest Valanginian (and latest Berriasian? time).

The *Polyptychites* Beds have been separated out in the Valanginian sequences of the Polish Lowland because of their intermediate position between the *Platylenticeras* Beds and beds with *Dichotomites* and *Saynoceras*, for which there is paleontological evidence. The *Polyptychites* Beds themselves have not yielded good diagnostic fossils; only a few fragments of *Polyptychites* have been found in their lower part. It is of interest that some part of the *Polyptychites* Beds of the Polish Lowland are developed as non-marine sediments, thus testifying to a late Lower Valanginian regression (MAREK 1983a).

The marine sediments of the Beds with *Dichotomites* and *Saynoceras* overstep earlier Valanginian sediments in several parts of the Polish Lowland, indicating a marine transgression at the beginning of the late Valanginian (MAREK 1983a).

Higher early Cretaceous sediments will not be discussed in this paper. A good review of this topic is provided by a paper by RACZYŃSKA (1979a).

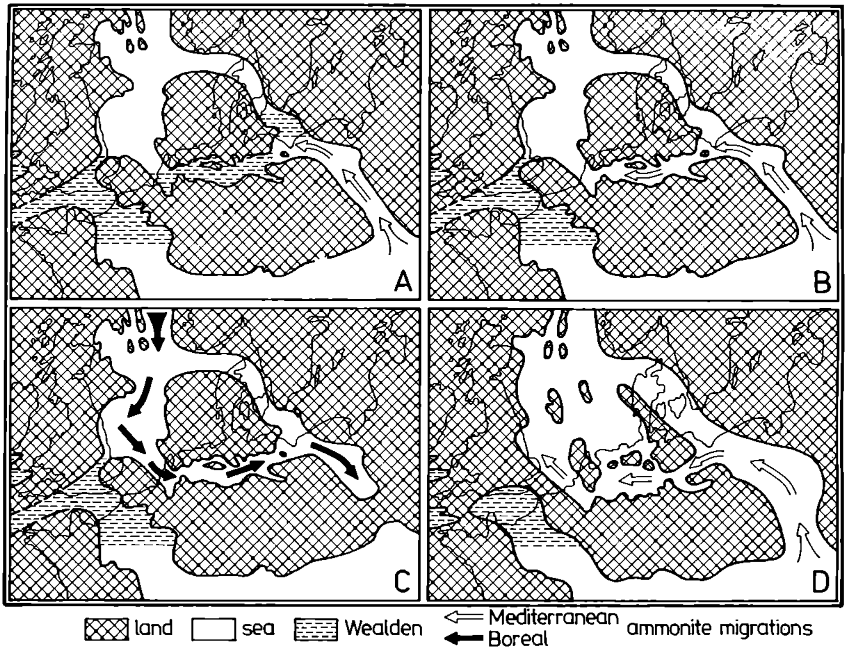
Relics of Neocomian sediments occur in Southern Poland in a position corresponding to the southern slope of the Meta-Carpathian Arch. On the western side of the present Mid-Polish Anticlinorium Neocomian sediments have been encountered in the borehole Stasiówka (Text-Fig. 1A). These sediments occur within the cratonic substrate of the Carpathian Skole Nappe as a layer ca. 10 m thick, intercalated between Upper Cretaceous and Upper Jurassic (probably Kimmeridgian) strata. The ostracod and foraminiferal assemblages in these sediments, most probably of late Valanginian age, contain, in addition to species known from the Polish Lowland, also species exclusively known from Tethyan regions, especially from the Crimea (GEROCH et al. 1972). A fragment of a lytoceratid has also been found at Stasiówka, which is of interest because no lytoceratids have been encountered in the Neocomian sediments of the Polish Lowland.

On the eastern side of the Mid-Polish Anticlinorium, thin and poorly dated Neocomian sediments are preserved in the Lubaczów region (Text-Fig. 1A). The age of basal Neocomian sediments seems to be late Valanginian, at least in some sections (MORYC & WAŚNIOWSKA 1965, MAREK & RACZYŃSKA 1979b).

Southeast of the Lubaczów region, in Western Ukraine, a Neocomian sequence is beginning with marine sediments probably belonging to the Upper Valanginian (PASTERNAK 1971).

11. Ammonite biogeography

It is commonly accepted that the late Berriasian Neocomitidae, represented by *Berriasella, Fauriella, Malbosiceras, Euthymiceras* and *Neocosmoceras* in the Polish Subbasin of the Central-European Basin, reached this region from the southeast, i. e. from the Carpathian Tethys (MAREK 1967, MAREK & RACZYŃSKA 1973, 1979a, MAREK 1983a, JELETZKY 1984). In this subbasin, the late Berriasian transgression gave rise to a marine sequence which probably encompasses, in addition to the Paramimouna and Picteti Subzones, also the Callisto Subzone and the lowest Valanginian Otopeta Zone (Text-Fig. 6A), including not only the sediments hitherto ascribed to the "Upper Berriasian" but also some lower parts of the Beds with *Neocomites* and *Platylenticeras*. This makes *Neocomites premolicus* also a Mediterranean faunal constituent of this sedimentary sequence.



Text-Fig. 6. Ammonite migrations (paleogeographic maps slightly modified after MICHAEL 1979); A: late Berriasian, B: early *Platylenticeras* time, C: late *Platylenticeras* and *Polyptychites* times, D: *verrucosum* time.

The specimen of *Karakaschiceras quadristrangulatum* found in bed "1" at Wąwał (Text-Fig. 2) has been derived from this sequence, probably from some upper part of it. As this is the earliest known occurrence of the genus *Karakaschiceras,* its origin in Poland out of the neocomitid main stock is a possibility. This would mean its origin in an epicratonic basin connected with the Tethys by a rather narrow sea-way centered over the southeastern part of the Mid-Polish Rift (Text-Fig. 6A).

Also an immigration of the genus *Riasanites* from Tethyan regions into the Polish Subbasin has been suggested in papers by MAREK (1967, 1983a) and by MAREK & RACZYŃSKA (1973, 1979a). This interpretation is consistent with the tectonical and paleogeographical pattern outlined in the preceding sections of this paper, which was favourable for faunal exchange with the regions of the Crimea and Caucasus. A direct eastward migration of *Riasanites* from the Central Russian Basin into Poland has been proposed by JELETZKY (1984), partly on the assumption that the genus *Riasanites* originated in that basin in pre-late Berriasian time. This interpretation, however, is rejected in a recent paper by HOEDEMAEKER (1987). In this context it is worth to note that *Riasanites* appears in sections of the Polish Lowland at levels where unquestionably Mediterranean late Berriasian Neocomitidae are also present (MAREK 1967). Thus, there is no evidence for an earlier, independent migration of *Riasanites* into Poland. However, the concept of a direct biogeographical link between the Central Russian Basin and the Polish Subbasin may find some support, provided that Russian species of *Surites* do occur in the Polish Lowland, as suggested e. g. by the identification of *Surites subtzikwinianus* (MAREK et al. 1984b).

The Tethyan ammonites which occupied the Polish Subbasin as a result of the late Berriasian transgression did not reach Northwest Germany where sedimentation of the Wealden persisted from the Berriasian to earliest Valanginian (*otopeta*) time (KEMPER et al. 1981, HOEDEMAEKER 1987).

Following regressive tendencies in some parts of Poland, the transgression at the beginning of *Platylenticeras* (= *pertransiens*) time (Text-Fig. 6B) brought the genus *Platylenticeras*, which is known from the Western Carpathians (VAŠÍČEK 1979), from the Tethys across Poland to Northwest Germany (KEMPER 1961, KEMPER et al. 1981).

The second level of occurrence of the neocomitid genus *Karakaschiceras* in the Wąwał section falls within a rather low interval of the *Platylenticeras* Beds. No *Karakaschiceras* has been found in Northwest Germany in the *Platylenticeras*-Schichten (KEMPER 1961, KEMPER et al. 1981), so that the occurrence of this genus in *pertransiens* time in Central Europe seems to be restricted to the peri-Tethyan region of Poland.

Significantly, there is no evidence for any migration of Tethyan cephalopods into the Polish Subbasin in late *Platylenticeras* (*pertransiens*) time. On the contrary, the presence of *Polyptychites* and *Acroteuthis* in the upper part of the *Platylenticeras* Beds in Poland testify to immigration of Boreal forms (Text-Fig. 6C).

The Polish *Polyptychites* Beds yielded only a few fragments of ammonites representing the Boreal genus *Polyptychites*. This, together with the presence of non-marine sediments in the *Polyptychites* Beds in the Polish Lowland and of a gap corresponding to these beds in the Tomaszów Syncline, suggests that the sea-way towards the Tethys in Southeast Poland was then closed (Text-Fig. 6C) and that region subjected to erosion in *Polyptychites* (*campylotoxus*) time (possibly as early as in late *Platylenticeras* time). Such an interpretation may account for the fact that several Neocomian sections in southeastern Poland start with Upper Valanginian sediments.

The wide-spread transgression at the beginning of the late Valanginian (*verrucosum* time), largely resulting from a global rise of the sea-level (HOEDEMAEKER 1987), brought assemblages of Tethyan ammonites (Text-Fig. 6D), including *Karakaschiceras*, to the Polish Subbasin, and also to Northwest Germany and England (KEMPER et al. 1981). A fall of the sea-level (HOEDEMAEKER 1984, 1987) may account for the abrupt replacement of the diversified Tethyan-related assemblage of the *verrucosum* Horizon by an assemblage dominated by *Dichotomites*, as displayed by the Wąwał section (Text-Fig. 2).

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Sedimentäre und paläotektonische Entwicklung der epikontinentalen Unterkreide Polens

SYLWESTER MAREK, Warszawa

Mit 6 Text-Figuren und 1 Tabelle

MAREK, S. (1989): Sedimentäre und paläotektonische Entwicklung der epikontinentalen Unterkreide Polens. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 755-770. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The sedimentary and paleotectonic development of the Lower Cretaceous platform areas in Poland is reconstructed in facies, tectonic, and isopach maps. The epicontinental Lower Cretaceous Basin in Poland shows a transgressive character with alternating marine influence from Tethys and/or North Atlantic and freshwater input from the surrounding uplands, respectively. Analysis of sediments points to a marine, clastic sand/clay facies in the NW and a sand/clay/carbonate facies in the SE.

A complete section of the Lower Cretaceous is only developed in the Middle Polish Trough with a maximum thickness of 650 m and in synsedimentary graben structures. In the central part of the trough occur several paleostructures, which are generally related to salt tectonics. Stratigraphic gaps and thinning-out of lithostratigraphic units characterize the margin of the Middle Polish Trough.

Kurzfassung: Die sedimentäre und paläotektonische Entwicklung der Unterkreide-Plattformgebiete Polens wurde auf Paläomächtigkeits-, Fazies- und einer paläotektonischen Karte dargestellt. Das polnische epikontinentale Unterkreidebecken hatte einen expansiven Charakter mit alternierenden marinen Einflüssen von Tethys und/oder Atlantik bzw. Süßwasser-Einfluß von den umgebenden Hochgebieten. Die Sedimentanalyse ergibt, daß die Unterkreide eine marine, terrigene Assoziation mit einer Sand/Ton-Subassoziation im NW und einer Sand/Ton/Kalk-Subassoziation im SE darstellt. Das vollständige Profil der Unterkreide ist auf die Mittelpolnische Furche beschränkt und erreicht im Zentrum eine Mächtigkeit von 650 m. In der Zentralzone der Furche kommen zahlreiche, meist salinare Paläostrukturen vor. Außerhalb der Furche sind starke stratigraphische Reduktionen erkennbar. Auskeilende Sedimentkörper bilden hier Monoklinalen und Strukturterrassen. Innerhalb dieser Strukturen kommen synsedimentäre Gräben mit vollständigen, für die Mittelpolnische Furche typischen Profilen vor.

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1. Einleitung

Im vorliegenden Beitrag werden die sedimentäre und paläotektonische Entwicklung der epikontinentalen polnischen Unterkreide auf 5 Paläomächtigkeits- und Fazieskarten (Text-Fig. 1-5) und auf einer paläotektonischen Karte (Text-Fig. 6) dargestellt. Mitautoren der Paläomächtigkeits- und Fazieskarten sind ANNA RACZYŃSKA (NW-Polen) und WŁADYSŁAW MORYC (S-Polen), Mitautor der paläotektonischen Karte ist ANNA FELDMANN. Diese Karten wurden auf der Grundlage von mehreren hundert Bohrungen und zahlreichen Tagesaufschlüssen der Unterkreide im Heiligkreuzgebirge erstellt.

Die Stratigraphie basiert auf der Grundlage von faunistischen, lithologisch-petrographischen und geochemischen Untersuchungen, verbunden mit Bohrlochgeophysik.

Die Paläomächtigkeits- und Fazieskarten stellen die Stufen des Berrias (Ryazan), des *Platylenticeras*-Untervalangin, des *Polyptychites*-Untervalangin, des Obervalangin, des Hauterives und des Barreme-Mittelalb dar. Sie wurden mit der quantitativen Mächtigkeits- und Lithofaziesanalyse mit Faktoren-Kombination konstruiert: Sand-Tonsteinschiefer-Faktor (S/T) und klastischer Faktor ([S + T]/K). Ferner erfolgten der Vergleich rezenter und fossiler Sedimentationsräume, eine Rekonstruktion der Mächtigkeiten und Fazies in epigenetischen Erosionszonen, eine Analyse der Subsidenzzonen-Grenzen und Sedimentationslücken.

Die paläotektonische Karte der Unterkeide (Text-Fig. 6) entstand durch Superposition einzelner Paläomächtigkeits- und Fazieskarten. Sie stellt ein paläostrukturelles Bild des Liegenden der Unterkreide dar, wie es am Ende des Mittelalb bestanden haben dürfte. Diese Karte gibt die Resultate der Mächtigkeitsanalyse, der Subsidenzprozesse, die Sedimentationslücken und Diskordanzen, aber auch Resultate der Fazies-Analyse und der regionalen Paläotektonik wieder.

2. Zur Paläogeographie der polnischen Unterkreide

Die Stratigraphie der Unterkreide der Plattformgebiete Polens ist immer noch nicht eindeutig.

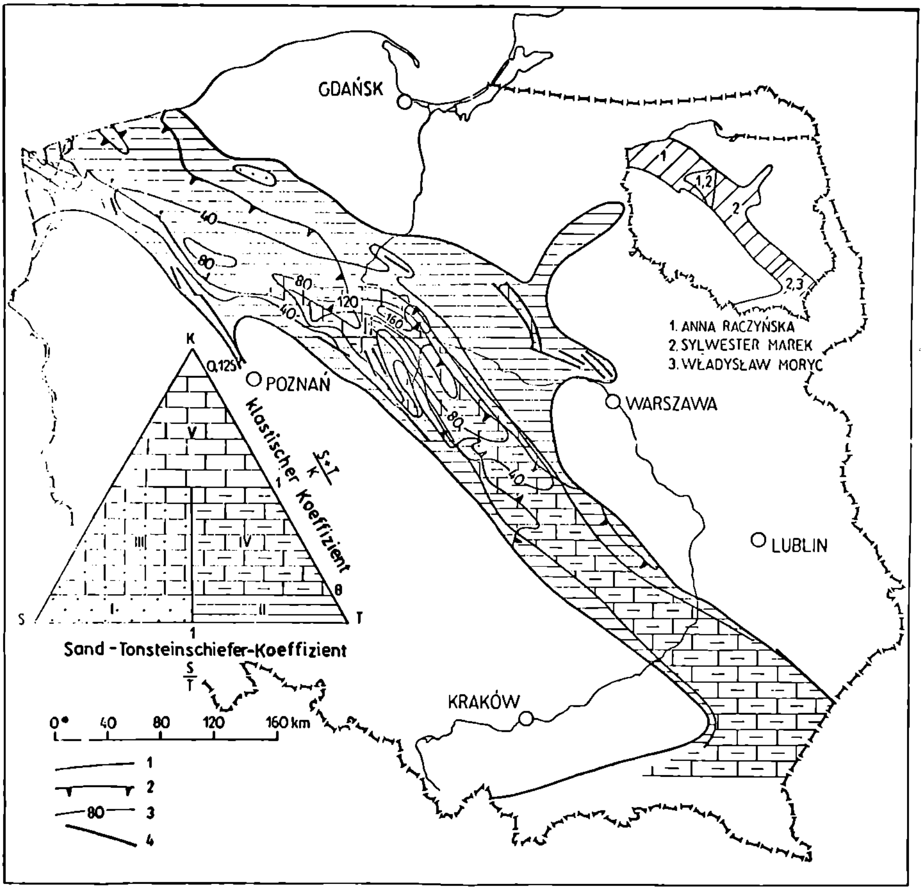
In zentralen Beckenteilen liegt eine verhältnismäßig gute Dokumentation für die Berrias (Ryazan)-Stufe, für das *Platylenticeras*-Untervalangin und für das Obervalangin vor. Sedimente des *Polyptychites*-Untervalangin, des Hauterives und Mittelalbs sind dagegen kaum mit Versteinerungen belegt. Im Barreme, Apt und Mittelalb mangelt es an charakteristischer Fauna (CIEŚLIŃSKI 1959, 1960, MAREK 1967, 1968, 1969, 1983, RACZYŃSKA 1967, 1971, 1979, WITKOWSKI 1969, SZTEJN 1969, 1984, MAREK & RACZYŃSKA 1973, 1979b, MAREK et al. 1984).

Faunistische Daten sind auch für die Randzonen des Unterkreidebeckens nicht vorhanden. Deswegen wurde in den letzten Jahren neben einer chrono- und biostratigraphischen Gliederung die lithostratigraphische Einteilung bevorzugt (RACZYŃSKA 1979, MAREK & RACZYŃSKA 1979a) (Tabelle 1). Wesentliche Bedeutung für die Stratigraphie des obersten Jura und der Unterkreide haben mikrofloristische Untersuchungsergebnisse (MAMCZAR 1986).

Das Unterkreidebecken steht in genetischem Zusammenhang mit der Zechstein-Mesozoischen Mittelpolnischen Furche, die sich am Kontakt von osteuropäischer präkambrischer Plattform und mittel/westeuropäischer paläozoischer Plattform gebildet hat. Diese Furche verband zwei paläogeographische Provinzen, die des Atlantiks im NW und die der Tethys im S und SE.

Das Unterkreidebecken hatte einen expansiven Charakter und ist gekennzeichnet von aufeinanderfolgenden Oszillationen mariner und intrakontinentaler Perioden.

2.1 Berrias und *Platylenticeras*-Untervalangin (Text-Fig. 1)



Text-Fig. 1. Paläomächtigkeits- und Fazieskarte des Berrias/Untervalangin (mit *Platylenticeras*).

1 - primäre Verbreitung der Sedimente (Null-Isopache) des Untervalangin mit *Platylenticeras*, 2 - Verbreitung der vollkommenen epigenetischen Späterosion nach der Oberkreide, 3 - Isopachen, 4 - synsedimentäre Verwerfungen; Klassifikationsdreieck für Text-Fig. 1-4: S - Sandsteine, T - Tonsteine, K - Karbonate; Lithofazien: I - Sandstein - siltig, II - siltig-tonig, III - sandig-karbonatisch, IV - tonig-karbonatisch, V - karbonat-klastisch.

Tabelle 1. Stratigraphische Gliederung der epikontinentalen Kreide in Polen.

Chrono- und Biostratigraphie			Leitfauna und charakteristische Fauna	Lithostratigraphie Glied	Formation	
1			2	3	4	5
Alb	Mittel		*Anahoplites praecox*			
			Hoplites dentatus			
			Dimorphoplites hili	Kruszwica		
			Neohibolites minimus			
			Cymatoceras radiatus			
	Unter		-			
Apt			*Ammonites* sp. indet.	Gopło	Mogilno	Text-Fig. 5
Barreme			-	Pagórki		
Hauterive	Ober-	*Simbirskites*	*Simbirskites (Craspedodiscus) gottschei*	Żychlin		
			Simbirskites (Craspedodiscus) sp.			Text-Fig. 4
	Unter-	*Endemoceras*	*Endemoceras* aff. *enode*			
			Endemoceras sp. (ex gr. *noricum-enode*)	Gniewkowo		
			Endemoceras sp.			
			Endemoceras cf. *amblygonium*		Włocławek	
			Bochianites neocomiensis			
Valangin	Ober-	*Dichotomites*	*Saynoceras verrucosum*			
			Dichotomites bidichotomus	Wierzchosławice		Text-Fig. 3
			Dichotomites cf. *biscissus*			
		und	*Polyptychites* cf. *petschorensis*			
		Saynoceras	*Polyptychites* cf. *ramulicostatus*			
			Polyptychites cf. *michalskii*			
			Polyptychites cf. *ascendens*			
			Polyptychites cf. *scalarinus*			
			Polyptychites cf. *latissimus*			
			Polyptychites sp. ex gr. *gradatus*			
			Polyptychites sp. ex gr. *keysserlingi*			
			Valangites nucleus			
			Neocraspedites complanatus			
			Neocraspedites sp.			
			Astieria cf. *sayni*			
			Astieria bachelardi			
			Leopoldia provincialis			
			Leopoldia cf. *biassalensis*			
			Leopoldia sp.			
			Neocomites biformis			

		Polyptychites		*Polyptychites* sp. *Polyptychites* cf. *gravidus* *Oxyteuthis primus*	-	Bodza- nów	Text- Fig. 2
	Unter-	*Platylenti- ceras* und *Neocomites*		*Platylenticeras (Pl.) heteropleurum posterum* *Platylenticeras (Tolypeceras) marcouisianum inflatum* *Platylenticeras (Pl.) gevrilianum gevrilianum* *Neocomites neocomiensis* *Neocomites neocomiensis* var. *praemolica* *Neocomites* sp.	Opoczki		
Berrias	Ober-	*Surites* und *Euthymiceras*		*Surites* cf. *subtzikwinianus* *Surites* cf. *spasskensis* *Surites* sp. div. *Peregrinoceras* sp. ex gr. *albidum* *Riasanites* sp. *Euthymiceras* cf. *euthymi* *Neocosmoceras* cf. *platycostatus* *Neocosmoceras prebalcanicum* *Neocosmoceras* sp. cf. *curelense* *Berriasella* sp. div.	Zakrzew	Rogoźno	Text- Fig. 1
	Unter-	*Riasanites* und *Malbosiceras*		*Riasanites swistowianus* *Riasanites rjasanensis* *Pseudosubplanites (Hegaratella) jauberti* *Malbosiceras* cf. *malbosi* *Berriasella (Picteticeras)* aff. *chomeracensis* *Berriasella (Picteticeras) picteti* *Berriasella (Picteticeras)* cf. *aurousei* *Berriasella vranensis* *Himalayites* sp.	Kajetanów		
Wolga-St.	Obere	Purbeck	A	*Cypridea posticalis, Pachycytheridea compacta,* *Eoguttulina witoldi* *Ammobaculites kcyniensis,* *Galiocytheridea* cf. *postsinuata*	Skotniki	Kcynia	
			B	*Klieana kujaviana, Nodophthalmocythere kcyniensis*			
			C	*Cypridea binodosa polonica,* *Cypridea tumescens granulosa* *Cypridea praealta, Cypridea tumescens acrobeles* *Cypridea tumescens praecursor*			
			D	*Cypridea dunkeri dunkeri, Cypridea dunkeri inversa* *Cypridea granulosa polonica,* *Cypridea binodosa binodosa*			
	Mittlere		E	*Fabanella ansata, Cypridea dunkeri sowerbyi*			
			F	*Mantelliana purbeckensis, Procytheropteron brodiei*			

Die ersten unterkretazischen Meeresingressionen gelangten auf das polnische Flachland von der Tethys und veränderten die stark ausgesüßten Becken der obersten Wolga-Stufe in brackisch-marine Becken der untersten Berrias-Stufe (KUTEK 1962, BIELECKA & SZTEJN 1966, MAREK 1967, 1984, MAREK et al. 1969, KUTEK & GŁAZEK 1972, DEMBOWSKA 1973, DEMBOWSKA & MAREK 1975, 1976). Während des jüngeren Berrias und im *Platylenticeras*-Untervalangin entwickelte sich daraus ein typisches intrakontinentales Meeresbecken, nun auch mit einer Verbindung zum Atlantik. Dafür spricht eine Vergesellschaftung atlantischer und tethydischer Ammoniten der Gattungen *Riasanites, Surites, Tollia (?), Peregrinoceras, Malbosiceras, Euthymiceras, Neocosmoceras, Pseudosubplanites, Berriasella, Himalayites*, ebenso wie *Platylenticeras, Neocomites* und *Polyptychites*. Die anfangs herrschende kalkarenitische und siltige Sedimentation zeigt später einen Übergang in siltig-tonige Sedimentation mit Sphärosideriten. Die siltig-tonigen Sedimente sind überwiegend dunkel gefärbt. Gegen Süden nimmt der terrigene Sedimentanteil zugunsten des karbonatischen Anteils ab. Allerdings ist die Kenntnis dieser Sedimente im SE-Teil der Mittelpolnischen Furche ungenügend. Diese Sedimente fielen größtenteils früh- und spätepigenetischen Erosionen zum Opfer. Dasselbe trifft auch für die jüngeren Glieder der Unterkreide zu.

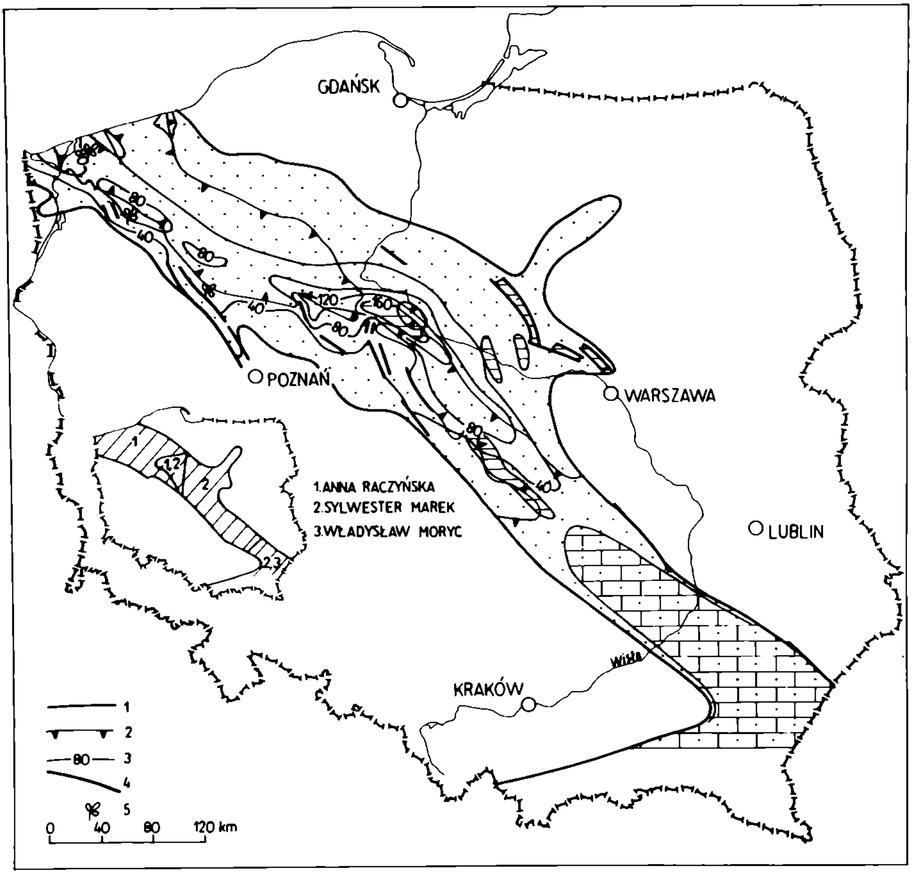
Das intrakontinentale Meer des Berrias und des *Platylenticeras*-Untervalangin war im wesentlichen auf die Mittelpolnische Furche begrenzt und hat nur im mittleren Teil, dem Kujawischen Abschnitt, den E-Rahmen überschritten. Ingressionen des Berrias-Meeres erreichten die Zone der synsedimentären tektonischen Gräben von Żuromin, Płońsk und Nasielsk-Dębe; das *Platylenticeras*-Valangin-Meer dringt als schmale Bucht in das Mazury-Land ein. Im zentralen Teil der Mittelpolnischen Furche erreichen die Sedimente eine Mächtigkeit von 80 bis 160 m und nehmen gegen die Außenzonen ab. Es zeichnen sich deutlich Gräben und synsedimentäre Mulden ab. Außer den obengenannten Gräben sind besonders die Gräben von Grzęzno und Szamotuły-Człopy zu erwähnen.

Mächtige Ablagerungen des Berrias und des *Platylenticeras*-Untervalangin und eine vollständige und kontinuierliche Entwicklung in den erwähnten tektonischen Gräben sprechen dafür, daß diese Gräben ständig mit dem offenen Meer verbunden waren, wahrscheinlich durch eine Bucht in der Gegend von Płońsk. Diese an der Jura-Kreide-Wende angelegten Gräben persistierten in das jüngere Valangin, das Hauterive, ebenso wie ins Barreme und Apt. Salzstrukturen zeigten geringe tektonische Mobilität und übten unbedeutenden Einfluß auf die Mächtigkeitsdifferenzierungen aus.

Mit einem 3-Faktoren-Klassifikationsdiagramm wurden zwei vorherrschende Lithofazien unterschieden: eine siltig-tonige Lithofazies im NW (Pommerschen) und mittleren (Kujawischen) Abschnitt, daneben eine hauptsächlich im SE entwickelte tonig-karbonatische Lithofazies. Außerdem war im NW lokal eine sandig-siltige Fazies ausgebildet. Die Sedimentation hat im seichten, aber offenen Meer im Wirkungsbereich der Gezeitenströmungen und Wellenbewegung stattgefunden. Nur im untersten Berrias existierte vorübergehend ein brackisch-marines Becken mit geringem Salzgehalt (Ende der Purbeck-Sedimentation).

2.2 *Polyptychites*-Untervalangin (Text-Fig. 2)

Auf die transgressive Etappe des Berrias/*Platylenticeras*-Untervalangin-Meeres folgt im *Polyptychites*-Untervalangin eine geringe Beckenverbreiterung bei



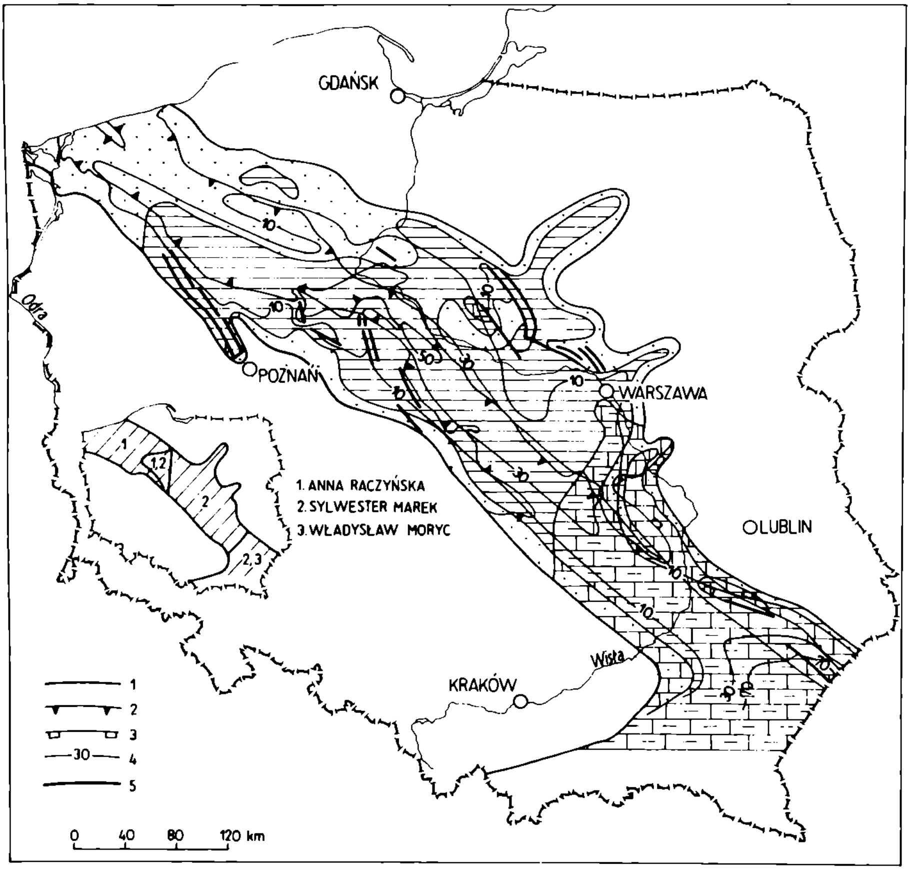
Text-Fig. 2. Paläomächtigkeits- und Fazieskarte des Untervalangin (mit *Polyptychites*).
1 - primäre Verbreitung der Sedimente (Null-Isopache) des höheren Untervalangin mit *Polyptychites*, 2 - Verbreitung der vollkommenen epigenetischen Späterosion nach der Oberkreide, 3 - Isopachen, 4 - synsedimentäre Verwerfungen, 5 - Flora des Moor/Brackwasser-Milieus; Klassifikationsdreieck siehe Text-Fig. 1.

gleichzeitiger Verflachung. Zunächst erfolgte eine Abschwächung der marinen Einflüsse, im Pommerschen Gebiet entwickelte sich sogar Moorsedimentation.

In diesem Zeitabschnitt entstanden hauptsächlich sandige, untergeordnet siltig-sandige Ablagerungen mit Rhizoiden und Holzresten, die für einen Süßwasserzufluß sprechen. Die Moorsedimentation erfolgte in seichten Beckenteilen, teilweise bildeten sich ausgedehnte tidale Ablagerungsräume. Marine graubraungefärbte feingebankte Ton- und Siltsteine mit Eisenkonkretionen wurden nur im unteren und oberen Profilteil des Kujawischen Gebiets festgestellt. Im unteren Teil dieses Profils wurden vereinzelt *Polyptychites*-Bruchstücke gefunden, die für Atlantik-Einflüsse sprechen. Allgemein überwog im NW und mittleren Beckenteil eine sandig-siltige Fazies. Terrigene

Sedimentation wird im SE von einer kalkarenitischen Sedimentation vertreten. In dem am meisten subsidenten Kujawischen Teil der Polnischen Furche erreichen die *Polyptychites*-Schichten eine Mächtigkeit von 120-160 m. Zu betonen ist ein rasches Absinken der Mächtigkeiten in Richtung auf die Ränder der Furche. Zunehmende Mächtigkeiten außerhalb der Furche sind nur in synsedimentären Gräben festzustellen. Nach wie vor zeigen die Salzstrukturen nur geringe tektonische Aktivität.

2.3 **Ober-Valangin** (Text-Fig. 3)



Text-Fig. 3. Paläomächtigkeits- und Fazieskarte des Obervalangin.
1 - primäre Verbreitung der Sedimente (Null-Isopache) des Obervalangin, 2 - Verbreitung der vollkommenen epigenetischen Späterosion nach der Oberkreide, 3 - Verbreitung der epigenetischen Erosion vor dem Alb, 4 - Isopachen, 5 - synsedimentäre Verwerfungen; Klassifikationsdreieck siehe Text-Fig. 1.

Im Ober-Valangin erfolgte die nächste transgressive Etappe, während der die Ränder der Mittelpolnischen Furche weit überschritten wurden. Außerhalb der Furche liegen die Obervalangin-Sedimente diskordant auf verschiedenen Schichtgliedern der Wolga-Stufe und des Kimmeridge.

Im NW und zentralen Beckenteil bilden sich nun hauptsächlich dunkle Ton- und Siltsteine mit sandig-dolomitischen und sideritischen Einlagerungen. Diese siltig-schieferartige Fazies ist bis in die Randzonen des Sedimentationsbeckens verbreitet und wird hier von einer sandig-siltigen Fazies begrenzt. In SE-Richtung ist eine Zunahme von organodetritischen und oolithischen Kalken und Mergeln zu beobachten. Dabei überwiegt hier eine tonig-karbonatische und sandig-karbonatische Fazies, teilweise ist sogar eine karbonatische Fazies entwickelt.

Im zentralen, Kujawischen Teil der Furche erreichen die Obervalangin-Sedimente eine Mächtigkeit von 30-50 m. Lokale Mächtigkeitsdifferenzierung ist mit beginnender Mobilität einiger Salzstrukturen verbunden.

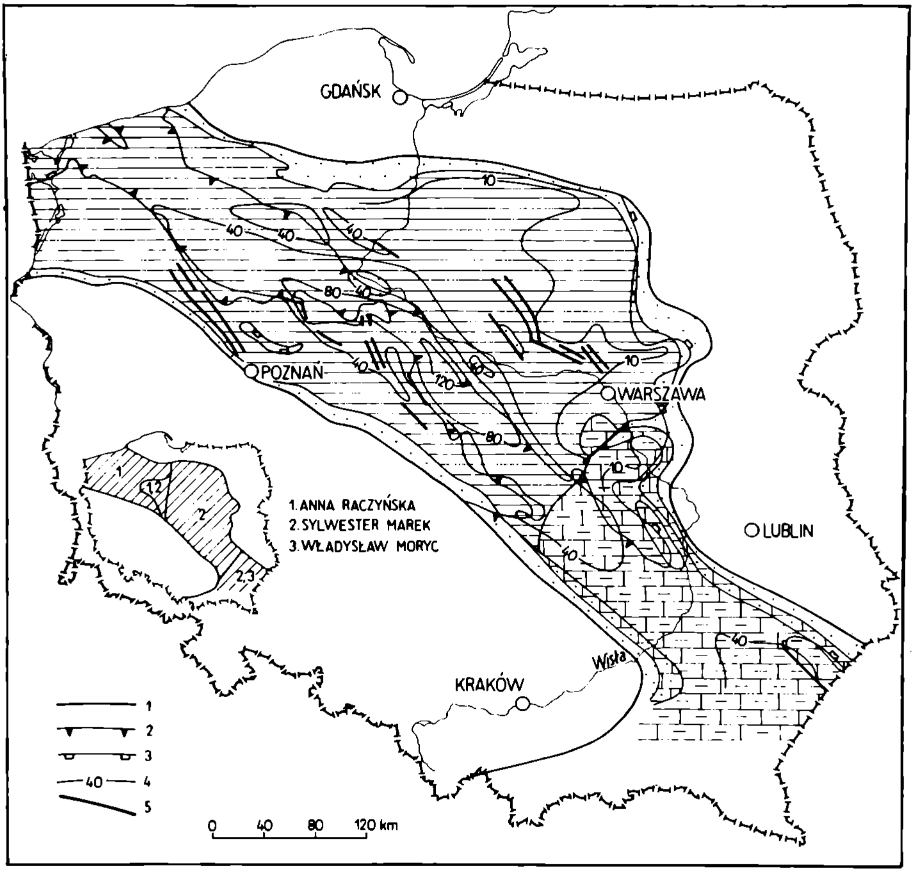
Im SE, in der Vorkarpaten-Senke erreichen die Obervalangin-Oolithkalke eine Mächtigkeit von über 70 m und lokal, an synsedimentären Verwerfungen 80 m (Bohrung Basznia 1).

In der Randzone des Karpatenflyschs (mit Ausnahme der Bohrung Stasiówka 1) wurde unter dem Turon und auf Kimmeridge eine 10 m mächtige Schicht epikontinentaler mergeliger Tonsteine von wahrscheinlichem Obervalangin-Alter durchbohrt (GEROCH et al. 1972). Für eine Tethys-Verbindung im Obervalangin sprechen Ammoniten der Gattungen *Bochianites, Saynoceras, Astieria, Leopoldia, Neocomites* und *Valanginites*. Eine Verbindung mit dem Atlantik bestätigen die Ammonitengattungen *Neocraspedites, Dichotomites* und *Polyptychites*.

2.4 **Hauterive** (Text-Fig. 4)

Auch das Hauterive-Meer war - nach einer kurzen Regression gegen Ende des Valangin - weiterhin transgressiv, belegt durch weit übergreifende Sedimente. Im NW und zentralen Beckenteil repräsentieren das Unterhauterive dunkle Ton- und Siltsteine mit Sideriteinlagerungen, zuweilen mit Eisenoolithen. Den Anfang des Oberhauterives kennzeichnen eine Verflachung des Sedimentationsraumes, erneute Erosion am Beckenrand und starker terrigener Input. Abgelagert wurden zu dieser Zeit teils dolomitische, teils sideritische Sandsteine, die stellenweise Goethit-Chamosit-Ooide enthalten. Im jüngeren Oberhauterive erfolgte eine geringe Verbreiterung des Meeresarmes mit einer überwiegend tonig-siltigen Sedimentation. Gegen Ende des Hauterive erfolgte wieder eine Zunahme sandiger Schüttungen mit örtlicher Konzentration von Oolith-Siderit-Chamositerzen. Im NW und zentralen Bekkenteil überwog im Hauterive eine siltig-tonschieferige Fazies, im SE-Bekkenteil dagegen eine tonschieferig- und sandig-karbonatische Fazies. An den Beckenrändern herrschte sandig-siltige Sedimentation. Die größten Mächtigkeiten erreichten die Hauterive-Sedimente mit 80-120 m in Zentralpolen, in tektonischen Gräben (Szamotuły-Człopa) sogar bis 217,5 m (Bohrung Szamotuły 21).

Zu den Rändern des Hauterive-Beckens nimmt die Mächtigkeit deutlich ab, wobei diese Reduktion sowohl primär als auch sekundär sein kann. Es erfolgte hier früh-epigenetische Erosion an der Wende Hauterive-Barreme, im Barreme, Apt und auch im Unter/Mittelalb. Die Salinartektonik war nicht intensiv und hatte nur geringen Einfluß auf die Sedimentation. Ammo-



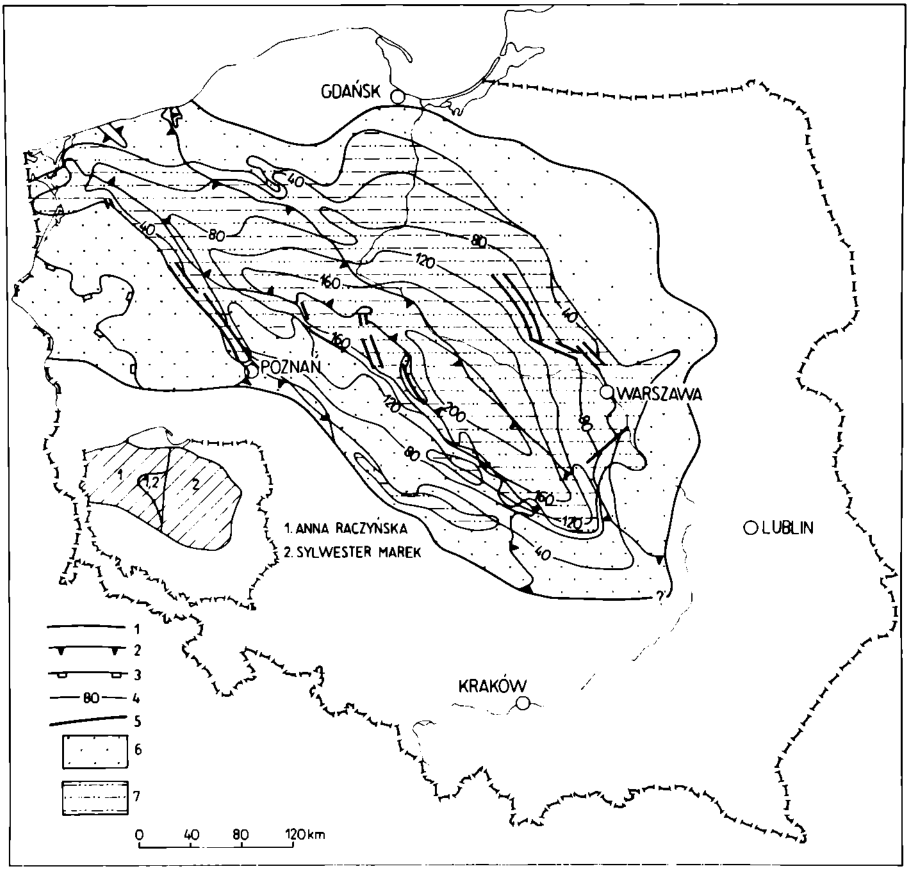
Text-Fig. 4. Paläomächtigkeits- und Fazieskarte des Hauterive.
1 - primäre Verbreitung der Sedimente (Null-Isopache) des Hauterive, 2 - Verbreitung der vollkomenen epigenetischen Späterosion nach der Oberkreide, 3 - Verbreitung der epigenetischen Erosion vor dem Alb, 4 - Isopachen, 5 - synsedimentäre Verwerfungen; Klassifikationsdreieck siehe Text-Fig. 1.

niten der Gattungen *Endemoceras* und *Simbirskites* sprechen für eine Dominanz atlantischer Einflüsse.

2.5 **Barreme-Mittelalb** (Text-Fig. 5)

Barreme-Mittelalb bestehen überwiegend aus sandigen Sedimenten, lithostratigraphisch drei Schichtglieder einer Formation: Pagórki-, Gopło- und Kruszwica-Schichten. Sie entsprechen dem Barreme, Apt und Unter/Oberalb (RACZYŃSKA 1979).

Im Barreme ist eine starke Verflachung und Einengung des Beckens zu beobachten, begleitet von starken Sandschüttungen. Abgelagert wurden vor allem helle Quarzsandsteine, oft Kaolinsandsteine mit inkohltem Pflanzen-



Text-Fig. 5. Paläomächtigkeits- und Fazieskarte des Barreme-Mittelalb. 1 - primäre Verbreitung der Sedimente (Null-Isopache) des Mittelalb, 2 - Verbreitung der vollkommenen epigenetischen Späterosion nach der Unterkreide, 3 - Verbreitung der epigenetischen Erosion vor dem Alb, 4 - Isopachen, 5 - synsedimentäre Verwerfungen; Lithofazien: 6 - Sandstein, 7 - Sandstein-Silt.

detritus. Das Becken hatte einen intrakontinentalen Charakter mit geringen Atlantik-Einflüssen; es war von der Tethys getrennt durch die nach dem Hauterive gebildete "Unter-San-Schwelle" (Małopolska-Land).

Das Apt zeigte eine neue schwache marine Ingression mit der Ablagerung sandig-schluffiger Sedimente mit Glaukonit, zuweilen mit einem Anteil von Eisenoolitherzen. Das Meer war seicht, turbulent und von geringem Salzgehalt. Das Apt-Meer war ebenso wie das des Barreme von der Tethys abgesondert und hatte nur mit dem Atlantik Verbindung.

Im Unter- und Mittelalb erfolgte eine neue, sich kontinuierlich entwickelnde marine Transgression, die von einer erneuten Verflachung des Sedimentationsbeckens begleitet war. Abgelagert wurden vor allem helle Sandsteine mit coarsening-upward- oder fining-upward-Sequenzen; die Sandsteine haben hier die größte Verbreitung. Diese Sandsteine scheinen im SE Polens zu fehlen, wo die Obervalangin- und Hauterive-Sedimente unmittelbar von kalkigen Sandsteinen des Oberalb bedeckt werden (MORYC & WAŚNIOWSKA 1965).

Das Unter- und Mittelalb-Meer war seicht und von normalem Salzgehalt; es war nur im Unteralb sicher mit dem Atlantik verbunden. Im Mittelalb existierte wahrscheinlich auch eine Verbindung mit der Tethys. Die Cephalopoden-Gattungen *Hoplites, Anahoplites, Dimorphoplites, Cymatoceras* und *Neohibolites* belegen eine Dominanz der Atlantik-Einflüsse (CIEŚLIŃSKI 1959, 1960).

Allgemein herrscht im Barreme-Mittelalb im Zentrum der Furche eine sandig-siltige Lithofazies vor (Schluffsteine der Gopło-Schichten). In den Randzonen dominiert die Sandstein-Fazies (transgressive Sedimente des Kruszwica-Glieds).

Die Mächtigkeit der Mogilno-Formation überschreitet im zentralen Bekkenteil 200 m. Größere Mächtigkeiten wurden lokal in tektonischen Gräben festgestellt.

Im Bereich der Salinartektonik sind starke Mächtigkeitsschwankungen der Pagórki- und Kruszwica-Schichten zu beobachten.

3. Paläotektonische Karte der Unterkreide (Text-Fig. 6)

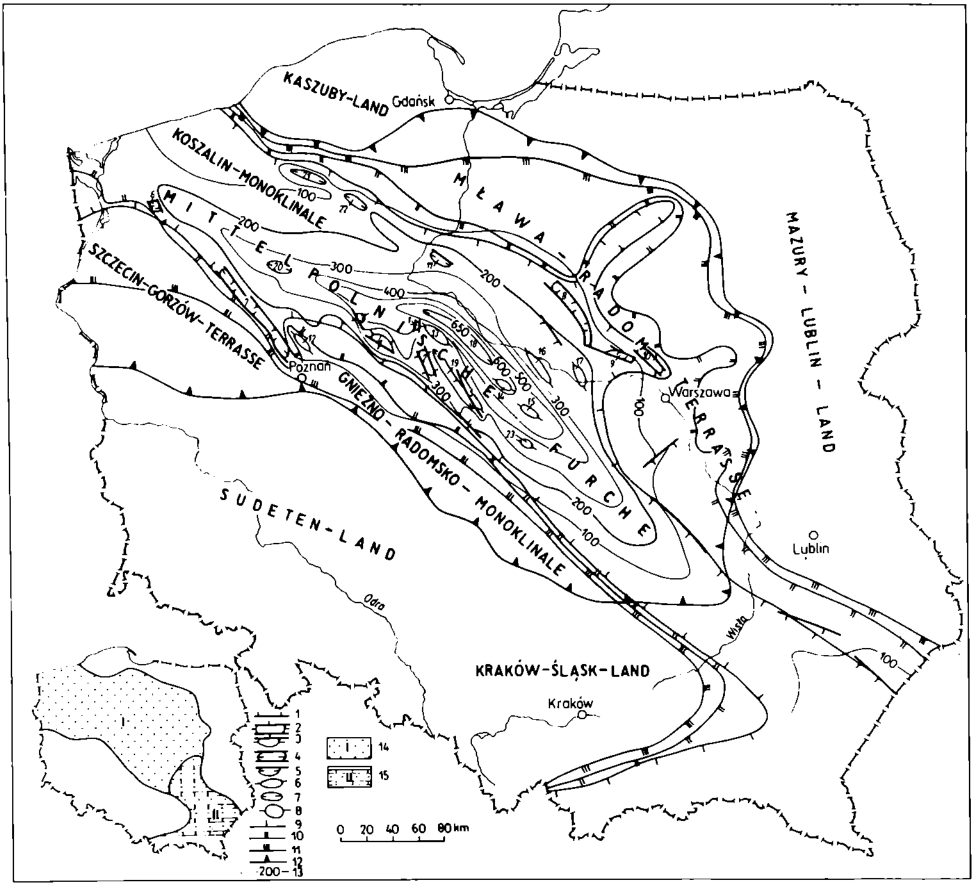
Die epikontinentale Unterkreide Polens hatte einen expansiven Charakter, mit wechselnden marinen Einflüssen von Tethys und/oder Atlantik bzw. Süßwassereinflüsse aus den begrenzenden Hochgebieten.

Die Sedimentation des Berrias und *Platylenticeras*-Untervalangin, des Obervalangin, Hauterive und Apt erfolgte in einem seichten, aber offenen Intrakontinentalmeer. Im NW Beckenteil wurden überwiegend dunkle Ton- und Siltsteine mit bedeutendem Gehalt an Eisensulfiden und -karbonaten, Glaukonit, Chamosit und Muskovit abgelagert. Diese Sedimente entstanden meist bei ungenügender Beckendurchlüftung. Im SE Beckenteil wurden zu dieser Zeit vor allem helle Kalke und Mergel mit großem Anteil oolithischer und organodetritischer Kalke abgelagert, die für eine gute Durchlüftung sprechen.

Im *Polyptychites*-Untervalangin, im Barreme und im Unter/Mittelalb unterlag das ganze Becken starker Verflachung, im *Polyptychites*-Untervalangin und im Barreme sogar bedeutender Aussüßung. Die Sedimentation war zu dieser Zeit überwiegend sandig.

Die Faziesverteilung erlaubt die Folgerung, daß das Becken dem Einfluß von zwei Klimazonen unterlag. Im NW herrschte ein sehr feuchtes Klima; durch mechanische Verwitterung und relativ tiefe Erosion der das Becken umgebenden Länder (Kaszuby-, Mazury- und Sudetenland) erfolgte starker detritischer Eintrag in das Becken. Dagegen stand der SE Beckenteil unter dem Einfluß eines wesentlich trockeneren und wahrscheinlich wärmeren Klimas. Im Bereich der benachbarten Hochs von Kraków, Śląsk und Lublin überwog die chemische Verwitterung, so daß weit geringere Mengen detritischen Materials angeliefert wurden.

Aus der Sedimentanalyse geht hervor, daß die epikontinentale Unterkreide Polens das früh-transgressive Stadium der tektonischen Entwicklung in einer marin-terrigenen Fazies-Assoziation repräsentiert. Im NW Beckenteil ist eine klastische Sand- und Ton-Subassoziation, im SE dagegen eine klastisch-kalkige Sand-Ton-Kalk-Subassoziation ausgebildet.



Text-Fig. 6. Paläotektonische Karte der Unterkreide.
1 - synsedimentäre Verwerfungen, 2 - synsedimentäre Horste (1 - Inowrocław, 2 - Gopło), 3 - Halbhorste (3 - Damasławek, 4 - Mogilno, 5 - Izbica-Kłodawa), 4 - Gräben (6 - Grzęzno, 7 - Szamotuły-Człopa, 8 - Żuromin, 9 - Płońsk, 10 - Nasielsk-Dębe), 5 - Halbgraben (11 - Chełmża), 6 - Schwellen (12 - Oborniki, 13 - Konary, 14 - Lubień-Łanięta, 15 - Wojszyce, 16 - Gostynin, 17 - Wyszków), 7 - Depressionen (18 - Gniewkowo, 19 - Piotrków Kujawski-Grzegorzew, 20 - Piła, 21 - Biały Bór-Rzeczenica, 22 - Chojnice), 8 - Kuppeln (23 - Rogoźno); Beckengrenzen: 9 - *Platylenticeras*-Untervalangin, 10 - Obervalangin, 11 - Hauterive, 12 - Barreme-Mittelalb, 13 - Paläoisohypsen des Liegenden der Unterkreide; marine terrigene Assoziation: 14 - klastische Sand/Ton-Subassoziation, 15 - klastisch-kalkige Sand/Ton/Kalk-Subassoziation.

Ein vollständiges Unterkreideprofil ist auf die Mittelpolnische Furche begrenzt, in der folgende Zonen unterschieden werden: im NW ein Pommerscher, im Zentrum ein Kujawy-Rawa- und im SE ein Gielniów-Heiligkreuzgebirge-Abschnitt. Subsidenz und tektonische Mobilität sind in den einzelnen

Abschnitten der Mittelpolnischen Furche sehr unterschiedlich entwickelt. Die größte Subsidenz ist im zentralen Teil der Furche zu beobachten (Kujawy-Rawa-Furche), wo ununterbrochene Sedimentation in Jura und Kreide erfolgte und die Unterkreide-Sedimente eine Mächtigkeit von 650 m erreichen.

Die Mittelpolnische Furche, konventionell mit einer 200 m-Paläoisohypse gezeichnet, ist ca. 500 km lang, während ihre Breite stellenweise 100 km erreicht. Die Furchenachse verläuft entlang der SW-Kante der Pommerschen Furche und danach kulissenartig entlang der NE-Kante der Kujawy-Rawa- und Gielniów-Heiligkreuzgebirge-Furche. Die Kujawy-Rawa-Furche war von starker Mobilität gekennzeichnet (zahlreiche positive und negative Paläostrukturen). Diese Paläostrukturen sind meist mit der synsedimentären Bildung der Salinarstrukturen verbunden, mit einem hohen Anteil aktiver Verwerfungen. Zu bemerken sind hier die Horste von Inowrocław und Gopło, die Halbhorste von Domasławek und Mogilno (im NE) und Izbica-Kłodawa, die Rogoźno-Kuppel und die Hochs von Konary, Lubień-Łanięta, Wojszyce, Gostynin und Wyszków. Negative Paläostrukturen sind der Halbgraben von Chełmża und die langgestreckten Depressionen von Gniewków, Piotrków, Kujawski-Grzegorzewo und Piła. Paläostrukturen der Salinartektonik wurden hauptsächlich an der Wende Wolga-Stufe/Berrias, im Obervalangin und an der Wende Valangin/Hauterive, aber auch im Barreme und Unteralb gebildet.

In der SE-Verlängerung der Mittelpolnischen Furche entstand nach dem Hauterive eine große positive Paläostruktur, die Unter-San-Schwelle. Diese Schwelle begann wahrscheinlich schon im Hauterive eine Rolle zu spielen, die Verbindung mit der Tethys wurde im Barreme und Apt ebenso wie im Unteralb abgeriegelt.

Außerhalb der Mittelpolnischen Furche wurden zum Teil starke stratigraphische Reduktionen und Auskeilen der Sedimente festgestellt. Die Ränder der Furche werden deutlich von Sedimenten des Obervalangin, Hauterive, teilweise des Barreme und Apt und besonders von denen des Mittelalb überschritten. Diese Sedimente bilden große einseitige Paläostrukturen (Monoklinalen und Strukturterrassen), die mit den Gebieten geringerer Subsidenz zusammenfallen. Sie sind von den Paläoisohypsen mit Werten von 200 bis 0 m abgebildet.

Im NE wird die Mittelpolnische Furche von der Koszalin-Monoklinale und Mława-Radom-Terrasse, im SW von der Szczecin-Gorzów-Terrasse und Gniezno-Radomsko-Monoklinale begrenzt.

Besondere Beachtung benötigen die synsedimentären Gräben, die mit Blocktektonik verbunden sind und sich innerhalb der Terrassen in der Nachbarschaft der Mittelpolnischen Furche entwickelten. SW der Pommerschen Furche entstanden die Gräben von Szamotuły-Człopa und Grzęzno, im SE der Kujawy-Furche die Gräben von Żuromin, Płońsk, Nasielsk und Dębe. Diese Gräben, die bereits im Mittel- und sogar Unterjura angelegt wurden, zeigten während der ganzen Unterkreide Aktivität. Das stratigraphische Profil dieser Gräben ist bedeutend vollständiger und die Mächtigkeiten sind beträchtlich größer als die angrenzender Gebiete. Im Żuromin-Graben überschreitet die Mächtigkeit der Unterkreide-Sedimente stellenweise 300 m und im Człopy-Graben sogar 600 m. Diese Mächtigkeiten entsprechen der stärksten Subsidenz der Mittelpolnischen Furche. NE der Pommerschen Furche wurde ebenfalls eine Zone verstärkter Subsidenz festgestellt (Depressionen von Biały Bór-Rzeczenica und Chojnice).

Im NW-Teil der Gniezno-Radomsko-Monoklinale schließlich ist die kleine Oborniki-Schwelle zu erwähnen, die möglicherweise auch mit einer Salinarstruktur verbunden ist.

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Correlating Boreal and Subtethyan Valanginian with Buchias and Ammonites

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With 2 Tables

ZAKHAROV, V. A. & BOGOMOLOV, J. I. (1989): Correlating Boreal and Subtethyan Valanginian with Buchias and Ammonites. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 771-774. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

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Significant progress has been made in attempts to correlate the Boreal and Subtethyan Valanginian due to the contributions of R. IMLAY, J. JELETZKY, E. KEMPER, V. SACHS, N. SHULGINA, J. P. THIEULOY, J. WIEDMANN, and others. However, precise correlation of the Boreal and Mediterranean Valanginian is still difficult. Agreements have neither been achieved on the subdivision of the Valanginian into an upper and lower substage, nor on the correlation of the lower and upper boundaries of the Valanginian Stage, nor on the correlation of Subtethyan and Boreal ammonite zones.

Work on the subdivision of the Boreal Valanginian, especially of the Arctic Valanginian, into ammonite zones is still in progress. In the USSR only a few specialists are involved in the study of the Boreal Valanginian. They have different points of view as shown by the two different reports on the stratigraphy of the Valanginian presented at this meeting (see SHULGINA, this vol.).

The problem of correlating the Boreal and Subtethyan Valanginian is further complicated by the fact that none of the standard ammonite zones from the stratotype of the Valanginian in Switzerland and from the hypostratotype in southeastern France can be recognized in the Boreal Realm. Furthermore, there are significant differences in ammonite zonations of the Subboreal regions, namely northeastern England, the Lower Saxony Basin of West Germany, the Russian Platform, the High-Boreal or Arctic regions of northern Siberia, Greenland and Arctic Canada. All this complicates a direct correlation of the Arctic, Subboreal, and Tethyan ammonite zonations.

Correlations of the Boreal and Submediterranean Valanginian is, however, possible by means of the bivalve genus *Buchia*. In the Arctic and some Subboreal regions there is a continuous sequence of Buchia zones.

An attempt was made to demonstrate that identical or very similar Buchia zonations can be defined in the northern regions of the USSR and along the Pacific Coast of the United States. It has been demonstrated by V. ZAKHAROV at the 1st Cretaceous Symposium in Münster in 1978 that a high degree of correlation exists between the strata in the two regions. The main disagreement still remains as to the age assignments of the Buchia zones given by IMLAY & JONES (1970) and by ZAKHAROV (1981). Ammonite data of BOGOMOLOV lend support to the former interpretation.

Based on the occurrence of *?Neocomites (Parandiceras)* cf. *rota* and five species of *Tollia* in British Columbia, IMLAY & JONES (1970) assigned a basal Valanginian age to the Buchia tolmatschovi Zone. The illustrated specimen of *Neocomites* is a juvenile one that lacks the specific features necessary for identification at the generic or subgeneric level. Moreover, the genus *Neocomites* has a long stratigraphic range; it is not restricted to the Valanginian but extends from the Upper Tithonian to Valanginian. According to SHULGINA (1972) all five species of *Tollia* listed in the above paper are of late Berriasian age. In the north of the USSR, *Buchia tolmatschowi* is only known from the Bojarkia mesezhnikovi ammonite Zone and from the *Tollia tolli* beds. Therefore, the Buchia tolmatschowi Zone of IMLAY & JONES (1970) is assigned to the Upper Berriasian. Overlying strata with *Buchia pacifica,* which is vicarious of *Buchia inflata,* are placed in the lower Middle Valanginian by IMLAY & JONES (1970); their age assignment is based on the presence of the ammonite genera *Tollia, Sarasinella, Thurmanniceras,* and *Kilianella* occurring further up in the section. The first, but rare *Thurmanniceras* and *Kilianella* appear in the uppermost Berriasian and become very abundant in the basal Valanginian. The ammonite genus *Tollia* was revised and the genus *Neotollia* was introduced by SHULGINA (1972); the latter is characteristic for the lowermost Valanginian of the Boreal Realm. In southeastern France the genus *Sarasinella* occurs exclusively in the Lower Valanginian. In the northern USSR (Pechora River Basin and Khatanga Depression) *Buchia inflata* is quite rare in the *Tollia tolli* beds which belong to the Boreal uppermost Berriasian (ZAKHAROV 1981). Therefore, it is proposed that the oldest beds, in which *Thurmanniceras, Kilianella,* and *Buchia inflata* occur in northern California, are of latest Berriasian age. The Berriasian - Valanginian boundary should be drawn at the level where *Tollia mutabilis* has its first appearance. We feel that this species should be assigned to the genus *Neotollia.* The acme of *Neotollia mutabilis* is used by IMLAY & JONES (1970) to define their Mutabilis Zone. In northern Siberia *Buchia inflata* is most abundant in the Lower Valanginian Neotollia klimovskiensis Zone, co-occurring with abundant *Neotollia* and *Temnoptychites.* The upper boundary of the lowermost Valanginian zone is best placed where *Buchia inflata* and *B. pacifica* disappear, just above the last common occurrence of *Neotollia mutabilis* (see Table 1).

Overlying beds containing *Buchia keyserlingi, Sarasinella* sp., *"Polyptychites" trichotomus, Polyptychites* sp., *Neocraspedites giganteus, Crioceratites, Thurmanniceras jenkinsi,* and *Olcostephanus* sp. are placed to the Middle Valanginian by the North American authors. However, the occurrence of *Sarasinella* in the basal part of the Buchia keyserlingi Zone is indicative of an early Valanginian age. Furthermore, *Polyptychites* sp., which appears 0.4 m above the base of this zone, resembles the subgenus *Euryptychites* which is found in the Euryptychites (Euryptychites) astieriptychus Zone of northern Siberia. Their *"Polyptychites" trichotomus* does not resemble true polyptychitids. Because of the flatness of its shell and the character of its ribs, this species closely resembles Lower Valanginian craspeditids, especially the genus *Thornsteinssonoceras* JELETZKY found in the Euryptychites (Euryptychites) astieriptychus Zone of northern Siberia (Table 2).

The epibole of the genus *Thurmanniceras* is found in the Lower Valanginian according to WIEDMANN (1980). Olcostephanids with coarse ribs are restricted to the Valanginian sensu lato according to IMLAY & JONES (1970). These authors also consider the presence of *Neocraspedites giganteus* as indicative of uppermost Upper Valanginian. However, it should be pointed out that: 1) neocraspeditids already occur in the Lower Valanginian, and the type species of the genus *Neocraspedites, N. semilaevis,* is restricted to the Lower Valanginian; 2) the Californian specimen of *Neocraspedites giganteus* does not display inner whorls. This species could thus belong to another genus almost completely lacking ribs in adult specimens, such as *Dichotomites* KOENEN or *Homolsomites* CRICKMAY. Thus, this specimen is not a reliable age indicator. The range of the genus *Crioceratites* extends over the entire Valanginian and may even extend into the Hauterivian and Barremian, according to WIEDMANN (1980) and others.

In the northern USSR and in Greenland, the *Buchia keyserlingi* ranges throughout the Lower Valanginian and extends into the Upper Valanginian as shown by ZAKHAROV (1981) and by SURLYK & ZAKHAROV (1982). The Buchia keyserlingi Zone includes three ammonite zones; it is correlated with the Lower Valanginian by two interregional ammonite datum levels. Firstly, the base of the Propolyptychites quadrifidus Zone of northern Siberia is correlated with the middle part of the Platylenticeras robustum Zone in the Lower Saxony Basin which in turn corresponds to the base of the Pertransiens Zone of the standard zonation. Secondly, the base of the Polyptychites beani Subzone of the Siberites ramulicosta Zone of northern Siberia correlates with the base of the Polyptychites multicostatus Zone in the Lower Saxony Basin, which coincides with the base of the Campylotoxum standard Zone (Table 2). Based on these facts it is concluded that the Buchia keyserlingi Zone of California and Oregon belongs to the Lower Valanginian (Table 1).

The highest beds in California that include *Buchia crassicollis* are assigned to the Upper Valanginian. Overlying beds (without *Buchia*) contain the ammonite genera *Spitidiscus* and *Wellsia* indicating a basal Hauterivian age. The ammonites *Homolsomites quatsinoensis* and *Olcostephanus pecki* are also found in the strata with *Buchia crassicollis*, but ammonites of the genus *Dichotomites* sensu lato were not encountered, but are common in the Upper Valanginian of the Lower Saxony Basin of West Germany, northern Siberia, Arctic Canada and southeastern France. It is noteworthy that the Buchia sublaevis Zone is missing in California and Oregon, but occurs in the Boreal Valanginian between the B. keyserlingi and the B. crassicollis zones embracing the entire Upper Valanginian in northern Siberia, eastern Greenland, northern Alaska, and in the Arctic Canadian Archipelago. Questions are: Which parts of the Upper Valanginian in northern California and Oregon correspond to the Buchia sublaevis Zone? Is there a gap?

According to KEMPER (1971), the presence of ammonites of the genus *Olcostephanus (Astieria),* without members of other ammonite genera, is indicative of the uppermost part of the Upper Valanginian in England and the Lower Saxony Basin.

In northern Siberia beds with these ammonites are absent. However, numerous subgenera of the genus *Polyptychites,* e. g. *Dichotomites, Paleodichotomites,* and *Polyptychites* s. str., occur together with *Buchia sublaevis* which is the marker species of the Sublaevis Zone. A late Valanginian age for the Buchia sublaevis Zone is indicated because: 1) it correlates with strata containing *Dichotomites bidichotomus* in northern Siberia; 2) it also correlates with beds containing *Dichotomites* spp. in the Lower Saxony Basin which, according to KEMPER, correlate with the Verrucosum Zone of the standard zonation, and 3) because beds with *Dichotomites bidichotomoides* correlate with the basal part of the Trinodosum standard Zone.

Beds with *Buchia crassicollis* and ammonites belonging to the genus *Homolsomites* overlie this Buchia Zone both in northern Siberia and in northern California and Oregon. Beds with ammonites of the genus *Homolsomites* from the Sverdrup Basin of northern Canada correlate with strata containing *Dichotomites* in the Lower Saxony Basin (KEMPER & JELETZKY 1979), that is with the middle part of the Upper Valanginian. Beds in northern Siberia contain *Homolsomites* but not *Dichotomites* and were assigned to the Lower Hauterivian by SHULGINA. At the present state of knowledge the true range of these *Homolsomites* is still debatable.

In any case, our investigation of Buchia zonations has led us to the conclusions that the lower part of the Upper Valanginian is missing in the California and Oregon sections of the USA. The youngest beds with buchias have yielded Lower Hauterivian ammonites from the northern Russian Platform and Lower Saxony Basin of West Germany. Thus, the precise age of beds with *Buchia crassicollis* in the Boreal Realm is still an open question.

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Table 1. Zonal correlation chart of Boreal and Subtethyan Valanginian.

Standard zonation	Stage	Sub-stage	Pacific coast of the Northern America: Ammonites	Pacific coast of the Northern America: Buchias	Arctic Canada: Buchias	Arctic Canada: Ammonites	North of Siberia: Buchias	North of Siberia: Ammonites	NW Germany: Ammonites	S England: Ammonites
Radiatus	Hauterivian	Lower	*Spitidiscus* sp., *Wellsia oregonensis*	No Buchias	No Buchias	Marine beds devoid ammonites	*Crassicollis*	*Bojarkensis*	*Amblygonium*	*Amblygonium*
Callidiscus	Valanginian	Upper	*Olcostephanus pecki*,	*Crassicollis*	*Buchia crassicollis* s. str.	*? Homolsomites ?*; *Dichotomites* spp.; *Tozeri*	*Sublaevis*	*Bidichotomus*: *Neocraspedites kotschetkovi* Beds	"*Astieria*" fauna; *Tuberculata*	"*Astieria*" fauna; *Pitrei*
Trinodosum			*Homolsomites quatsinoensis*			*Ammundense*		*Bidichotomoides*	*Bidichotomoides*	*Dichotomites* spp.
Verrucosum					*Sublaevis*	*Pseudopolyptychum*		*Triplodiptychus*	*Crassus*; *Polytomus*; *Hollwedensis*	
Campylotoxum		Lower	*Olcostephanus* sp.,	*Keyserlingi*	*Keyserlingi*	*Polyptychites* spp.; *Siberites rectangulatus*	*Keyserlingi*	*Ramulicosta*: *Beani*; *Ramulicosta*	*Sphaeroidalis*; *Clarkei*; *Pavlowi*	*Polyptychites* spp.
Pertransiens			*Thurmanniceras jenkinsi*, *Sarasinella angulata*, *Euryptychites astieriptychus*, ? *Propolyptychites trichotomus*,			*Euryptychites astieriptychus*, *Propolyptychites stubendorffi*; *Ellesmerensis*; *Kemperi*		*Astieriptychus*; *Quadrifidus*	*Involutum*; *Heteropleurum*	*Paratollia* spp.
Otopeta			*Neotollia mutabilis*, *Thurmanniceras californicum*	*Inflata*	?	*Troelseni*; Basal beds devoid ammonites	*Inflata*	*Klimovskiensis*	*Robustum*	
Boissieri	Berriasian	Upper	*Thurmanniceras* sp.	*Tolmatschowi*	*Buchia* aff. *volgensis*	*Tollia tolli*	*Tolmatschowi*	*Tollia tolli* Beds; *Mesezhnikowi*	Wealden 6 ? ?; Wealden 5	*Albidum*
Occitanica			*Spiticeras (Negreliceras) stonyensi*, *Neocosmoceras* sp.	*Buchia uncitoides*, *B. okensis*	*Uncitoides*		*Jasikovi*	*Analog.*: *Analogus*; *Subquadratus*	Wealden 4; Wealden 3	*Stenomphala*
		Lower			*Okensis*		*Okensis*	*Kochi*: *Praeanalogus*; *Constans*; *Kochi*		*Icenii*

Table 2. Interregional markers and correlation of zonal schemes for northern Middle Siberia, northwestern Germany and southeastern France.

Stage	Substage	Hypostratotype (SE France)	Inter-regional markers	NW Germany	Inter-regional markers	North of Middle Siberia (BOGOMOLOV 1986)	
Ht	Ht_1	*Radiatus*	*L. leopoldi*	*Amblygonium*		*Bojarkensis*	
VALANGINIAN	Upper	*Callidiscus*		"*Astieria*" fauna		*Bidichotomus*	*N. kotschetkovi* Beds
			Olcostephanus spp.				
			Di. houdardi	*Tuberculata*			
			Di. tuberculata				
		Trinodosum		*Bidichotomoides*			*Bidichotomoides*
			D. ramulosus				
			D. bidichotomoides	*Triptychoides*			
					D. bidichotomoides		
		Verrucosum		*Crassus*			*Triplodiptychus*
			S. verrucosum	*Polytomus*			
			Valanginites spp.				
			Karakaschiceras spp.	*Hollwedensis*			
			Dichotomites spp.		*D. bidichotomus*		
	Lower	*Campylotoxum*	*Th.* cf. *campylotoxum*	*Sphaeroidalis*		*Ramulicosta*	*Beani*
				Clarkei			
				Multicostatus			
			P. polyptychus		*P. polyptychus*		
		Pertransiens		*Pavlowi*			*Ramulicosta*
				Involutum		*Astieriptychus*	
			Pl. heteropleurum latum	*Heteropleurum*		*Quadrifidus*	
			Pl. marcousianum		*Pr. quadrifidus*		
		Otopeta		*Robustum*		*Klimovskiensis*	
Brs	Brs_2	*Boissieri*		"Weald"		*T. tolli* Beds	

Subdivisions of the Marine "Neocomian" in Siberia and Correlation with the Type Sections – A Review

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With 1 Table

SHULGINA, N. (1989): Subdivisions of the Marine "Neocomian" in Siberia and Correlation with the Type Sections - A Review. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 775-778. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

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The zonal scale for the "Neocomian" of northern Siberia varies in some details. Only the subdivision of the Berriasian (Table 1) is not a controversial problem among Siberian geologists, and is largely based on the first draft given by BODYLEVSKY (1939). Definition and extent of the Valanginian, its substages and zonal subdivisions are, however, differently interpreted by Soviet (SAKS 1972, GOLBERT & KLIMOVA 1983, SHULGINA & BURDYKA 1983, BOGOMOLOV 1986, ZAKHAROV & BOGOMOLOV, this vol.) and foreign scientists (e. g., KEMPER & JELETZKY 1979, JELETZKY & KEMPER 1988). For northern Siberia a twofold subdivision in a Lower and Upper Valanginian Substage seems to be adequate. No agreement is, however, achieved on the presence of Lower Hauterivian in northern Siberia which is in part denied.

And, similarly, there is no agreement at all about the correlation of the Siberian "Neocomian" with the southernmore Subboreal, Subtethyan and Tethyan sections. Correlation with the Berriasian type sections in SE France can be made only at a first approximation, and is limited to 2 or 3 zones or individual markers. In the Berriasian type sections 3 zones may be generally accepted and recognized (Table 1), i. e. in ascending order the Pseudosubplanites grandis/jacobi, the Tirnovella occitanica and the Fauriella boissieri zones. The Lower Berriasian boundary can be defined in both areas, the Tethyan and Boreal realms, by the disappearance of the Virgatosphinctinae. In many sections, the upper boundary is marked by the appearance of the Valanginian genera *Platylenticeras* (France, West Germany, England, Poland), *Paratollia* (West Germany, England), *Menjaites* (England, Russian Platform, northern Siberia), *Pseudogarnieria* (England, Russian Platform), *Propolyptychites* (West Germany, England, northern Siberia), and *Neotollia* (northern Russian Platform, northern Siberia, northern Urals, Greenland, northern Norway?, North America). This means that the

Table 1

Stage / Substage	South-East France	North-East England	Northern West Germany	Russian Platform	Northern Urals	Northern Siberia
Lower Hauterivian	Crioceratites loryi	Endemoceras Beds	Endemoceras noricum	Speetoniceras versicolor	Speetoniceras versicolor ?	—
	Acanthodiscus radiatus		Endemoceras amblygonium	Pavlovites polyptychoides; Homolsomites bojarkensis	Homolsomites bojarkensis	Homolsomites bojarkensis
Upper Valanginian	Teschenites callidiscus [Dichotomites]	?	„Astieria" fauna; Dicostella pitrei	Dichotomites, Homolsomites	Dichotomites, Homolsomites	Dichotomites, Homolsomites
	Himantoceras trinodosum [Dichotomites]	Dichotomites Beds	Dichotomites Beds	Dichotomites spp. u. Polyptychites polyptychus	Dichotomites ramulosus	Polyptychites polyptychus, Dichotomites spp.
	Saynoceras verrucosum [Prodichotomites]				„Amundiptychites" spp. u Polyptychites canadensis	Temnoptychites mediatus, Dichotomites, Polyptychites
Lower Valanginian	Thurmanniceras campylotoxum [Polyptychites]	Polyptychites Beds	Polyptychites Beds	Polyptychites michalskii	Polyptychites michalskii	Polyptychites michalskii
				Temnoptychites hoplitoides	Temnoptychites syzranicus	Temnoptychites syzranicus [Menjaites]
	Thurmanniceras pertransiens [Platylenticeras]	Platylenticeras, Paratollia, Pseudogarnieria, Menjaites Beds	Platylenticeras, Paratollia Beds	Pseudogarnieria undulato-plicatilis, Menjaites, Neotollia	Neotollia spp.	Neotollia klimovskiensis [Menjaites]
	Thurmanniceras otopeta [Platylenticeras]					
Berriasian	Fauriella boissieri	Peregrinoceras albidum		Peregrinoceras albidum	Bojarkia payeri	Bojarkia mesezhnikowi
		Bojarkia stenomphala		Surites tzikwinianus		
		Lynnia icenii		Euthymiceras, Riasanites, Surites	Surites analogus	Surites analogus
	Tirnovella occitanica	Hectoroceras kochi		Riasanites Hectoroceras	Hectoroceras kochi	Hectoroceras kochi
	Berriasella grandis/jacobi	Praetollia runctoni		Riasanites, Garnieriiceras ?	Chetaites sibiricus	Chetaites sibiricus
Upper Volgian	Durangites	Volgidiscus lamplughi			Volgidiscus pulcher [Virgatosphinetes spp.]	Chetaites chetae [Virgatosphinctes spp.]
		Subcraspedites preplicomphalus		Craspedites nodiger	Craspedites taimyrensis [Virgatosphinetes spp.]	Craspedites taimyrensis [Virgatosphinctes spp.]
	Paraulacosphinctes transitorius	Subcraspedites primitivus		Craspedites subditus	Craspedites subditus [Virgatosphinetes spp.]	[C. okensis] Subcraspedites originalis [Virgatosphinctes spp.]
				Kachpurites fulgens	Kachpurites fulgens [Virgatosphinetes spp.]	[C. okensis] C. okensis; V. exoticus [Virgatosphinctes spp.]

upper Berriasian boundary is much more reliable than the lower one. The Berriasian ammonite assemblages of northern Siberia, the northern Urals, Russian Platform, Poland and England have a number of genera and species in common; these are *Subcraspedites (Borealites)*, *Surites (Surites) spasskensis*, *S. (S.) subtzikwianus*, *Bojarkia stenomphala*, and others. In the Russian Platform and Poland, the above-mentioned forms are associated with southern ammonite genera such as *Neocomites*, *Rjasanites*, *Berriasella*, and *Euthymiceras*. In the Tethyan and Subtethyan sections the latter is known both from the Boissieri and partly from the Occitanicus Zone. On these lines, correlation of the two or three upper zones of the Siberian Berriasian with the standard zonation is feasible. Problems, however, arise with the lower portion of the Siberian Berriasian, especially with the Sibiricus Zone the equivalence of which needs to be elaborated.

In northern Siberia, *Neotollia klimovskiensis* defines - together with *Menjaites* - lowermost Valanginian. Correlation with the zones of Thurmanniceras otopeta and pertransiens at the standard is based on the following observations: (1) co-occurrence of Tethyan and Boreal ammonite genera like *Thurmanniceras*, *Kilianella*, *Tollia*, and *Neotollia* in the western U. S. A., (2) co-occurrence of *Neotollia* and *Menjaites* on the Russian Platform and in northern Siberia, (3) co-occurrence of *Pseudogarnieria* and *Menjaites* on the Russian Platform and in England where *Platylenticeras* is contemporaneous, and thus (4) final correlation of the Klimovskiensis Zone with the distribution of *Platylenticeras* in NW Germany and SE France.

The following Siberian Zone of Temnoptychites syzranicus is tentatively correlated with the lower Campylotoxum Zone of the standard by positioning. The Zone of Polyptychites michalskii yielded in the Siberian sections ammonites with close affinity to *"Neocraspedites" (= Prodichotomites) fissuratus, flexicosta*, and *undulatus* having been recorded from the lowermost Saynoceras verrucosum Zone of SE France. This means that the Michalskii Zone is preferably placed at the Lower/Upper Valanginian transition.

Correlation of the French late Valanginian Trinodosum and Callidiscus zones is made feasible due to the co-occurrence of *Dichotomites* ex gr. *bidichotomus* in the type sections as well as in the Boreal Zone of Polyptychites polyptychus. In consequence, the Valanginian/Hauterivian boundary cannot be correlated with the necessary certainty.

These problems continue into the Hauterivian. Only tentatively the Boreal Homolsomites bojarkensis Zone can be related to the lower portion of the *Endemoceras* Beds or the Acanthodiscus radiatus Zone, respectively. For KEMPER & JELETZKY (1979), the Homolsomites bojarkensis Zone is still late Valanginian. In western Siberia and the northern Urals, this is followed by the Zone of Speetoniceras versicolor which is considered to be time-equivalent to the French Zone of Crioceratites loryi. In northern Siberia, however, the Bojarkensis Zone is followed by continental to lagoonal sediments.

From this brief discussion it is obvious that the best traceable correlation is that of the Neotollia klimovskiensis Zone and the Berriasian/Valanginian boundary. It is therefore reasonable to draw the Jurassic/Cretaceous boundary at the base of the Valanginian as has been proposed previously by WIEDMANN (1967, 1968, 1975) and DRUSHTCHIC (1968), especially since no reliable marker is known from the base of the Berriasian joining Boreal and Tethyan realms.

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Late Cretaceous Palynoflora from the Western Tarim Basin of South Xinjiang, NW China

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With 3 Plates, 3 Text-Figures and 1 Table

YI-YONG ZHANG (1989): Late Cretaceous Palynoflora from the Western Tarim Basin of South Xinjiang, NW China. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 779-791. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The marine Upper Cretaceous from the Tarim Basin of Southern Xinjiang, NW China, contains a plenty of pollen and spores, and a great variety of marine fossils. On the basis of the features of palynological composition and vertical changes, three palynofloral suites can be distinguished from the late Cretaceous strata, viz. *Archaeotriporopollis-Taurocusporites* Suite (Cenomanian), *Cranwellia-Interulobites* Suite (Turonian) and *Xinjiangpollis-Senegalosporites* Suite (early Senonian). According to the Middle Cretaceous palynogeoprovinces by BRENNER (1976), the present Cenomanian-Turonian palynoflora shows some characteristics of both palynogeoprovinces, viz. the South Laurasian and the North Gondwana palynoflora. The early Senonian palynoflora contains even more North Gondwana elements, especially the diagnostic forms *Senegalosporites* and *Galeacornea,* which occupy a prominent position in the palynoflora.

Kurzfassung: Die marine Oberkreide des Tarim-Beckens im südlichen Xinjiang (NW China) enthält eine Vielzahl von Sporen und Pollen neben zahlreichen marinen Organismen. Auf der Grundlage der palynologischen Daten und vertikalen Veränderungen können drei verschiedene Palynofloren-Folgen ausgeschieden werden: (a) die *Archaeotriporopollis-Taurocusporites*-Folge (Cenoman), (b) die *Cranwellia-Interulobites*-Folge (Turon) und (c) die *Xinjiangpollis-Senegalosporites*-Folge (frühes Senon). Beim Vergleich mit den von BRENNER (1976) ausgeschiedenen Palynogeoprovinzen der Mittelkreide zeigen die Cenoman- und Turon-Floren die Merkmale beider Provinzen, d. h. der südlichen Laurasischen und der nördlichen Gondwana-Provinz. Die frühsenone Flora enthält dagegen mehr nördliche Gondwana-Elemente, nämlich die typischen *Senegalosporites* und *Galeacornea.*

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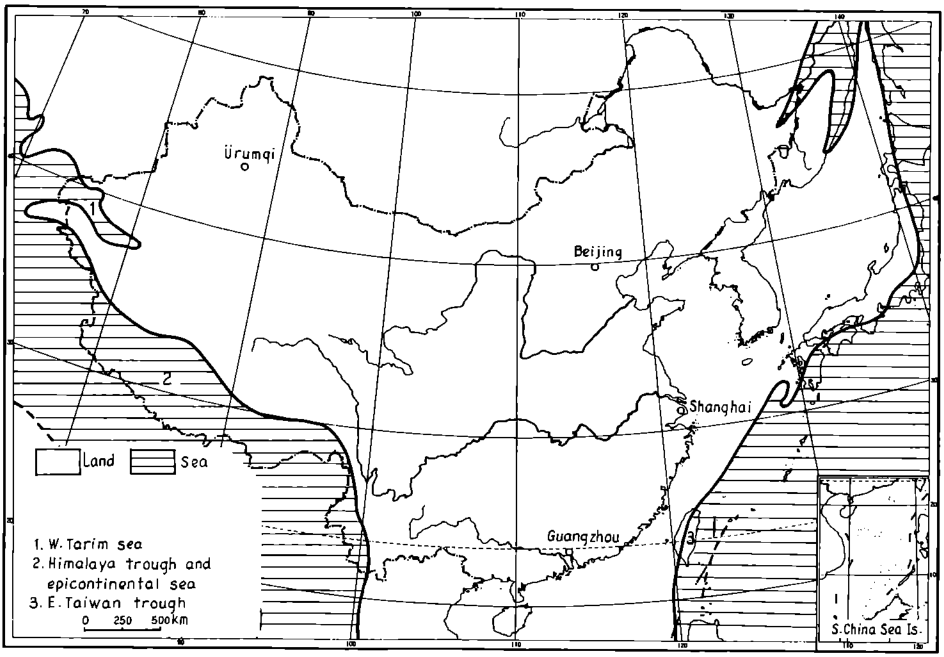
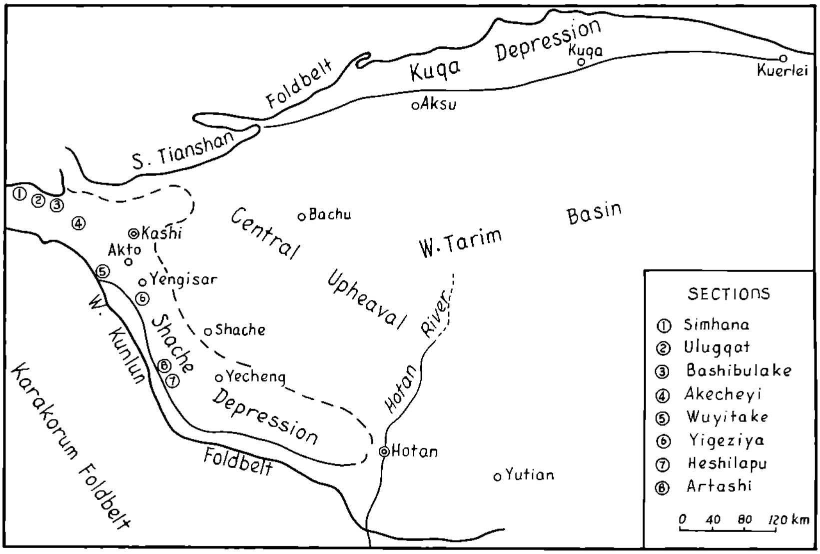
1. Introduction (Text-Figs. 1, 2)

Across the vast land and the continental shelf of China, marine Upper Cretaceous is only distributed at the West Tarim Basin, the Himalaya Folded Region and its marginal zone, and at the East Taiwan Foldbelt. Among these, the Upper Cretaceous from the Western Tarim Basin of South Xinjiang has attracted much attention of Chinese geologists for prospecting oil since the 1950's. Based mainly on the fossil molluscs found therein they have established stratigraphical sequences, named some formations and dated them preliminarily. As a result of the successive finding of various microfossils from those strata, it has been possible to correlate them with Upper Cretaceous stratotypes of the West Mediterranean region over the past decade. Particularly the finding of well-preserved pollen and spores from the same beds plays an important role for the correlation of both marine and non-marine Upper Cretaceous in China.

Pollen and spores mainly occur (1) from the Kukebai Formation to the Dongba Formation at sections of Simuhana, Wulukeqiat and Bashibulake along the foot of the southern South Tianshan Mountains; and (2) from the Kukebai Formation to the Wuyitake Formation at sections of Akecheyi, Wuyitake and Heshilapu along the foot of the northern West Kunlun Mountains. Among these, the Bashibulake section contains much better-preserved and less carbonized pollen and spores than other sections, and the palyno-



Text-Fig. 1. Sketch-map showing late Cretaceous paleogeography of China and adjacent areas.



Text-Fig. 2. Location map of studied sections of marine Upper Cretaceous in the western Tarim Basin.

logical record is thus more continuous. On the basis of associated marine macro- and microfossils, the ages of the palynofloras range from Cenomanian to early Senonian.

The present paper is part of the comprehensive results of the project "Study of Biostratigraphy and Deposited Environment of Marine Upper Cretaceous in the Western Tarim Basin" accomplished by the South Xinjiang Working Team (Nanjing Institute of Geology and Palaeontology) and the Exploration and Development Research Institute of Xinjiang Petroleum Administration (Oil Ministry of P. R. China).

This paper aims to describe briefly the components and the characteristics of each palynofloral suite, and also to discuss the problems related to palynoprovincialism and palaeoclimate.

2. Stratigraphy (Table 1)

The paper entitled "A Preliminary Study of the Upper Cretaceous of the Western Tarim Basin (South Xinjiang, China) with Special Reference to its Transgressions" (by YANG et al. 1983) has described the stratigraphical sequence in the studied area. On the basis of studies of fossils of more than ten groups, some stratigraphical units have been revised. The new classification of the stratigraphical sequences and their correlation between the foot areas of the southern South Tianshan Mountains and those of the northern West Kunlun Mountains is shown in Table 1.

Several revised stratigraphic units should be further mentioned as follows:

(1) The Kukebai Formation, formerly divided into two members, has now been divided into three members; and (2) the former "Wuyitake Formation" named for the strata along the foot of the southern South Tianshan

Table 1. Stratigraphical sequences and correlations between the foot areas of the southern South Mt. Tianshan and the northern West Mt. Kunlun.

<table>
<tr><th rowspan="2">AGE</th><th colspan="4">LITHOLOGICAL STRATIGRAPHY</th></tr>
<tr><th colspan="2">The foot of southern S. Mt. Tianshan</th><th colspan="2">The foot of northern W. Mt. Kunlun</th></tr>
<tr><td rowspan="3">SENONIAN</td><td rowspan="4">DONGBA FORMATION</td><td rowspan="2">U.</td><td colspan="2">TUYILOUKE FORMATION</td></tr>
<tr><td rowspan="2">YIGEZIYA FORMATION</td><td>U.</td></tr>
<tr><td>M.</td><td>L.</td></tr>
<tr><td rowspan="4">CENOMANIAN — TURONIAN</td><td>L.</td><td colspan="2">WUYITAKE FORMATION</td></tr>
<tr><td rowspan="3">KUKEBAI FORMATION</td><td>U.</td><td colspan="2" rowspan="3">KUKEBAI FORMATION</td></tr>
<tr><td>M</td></tr>
<tr><td>L</td></tr>
<tr><td>EARLY CRETACEOUS</td><td colspan="4">KEZILES GROUP</td></tr>
</table>

Mountains does not coincide with the Wuyitake Formation exposed at the stratotype locality; only the lower part corresponds to the Wuyitake stratotype. It has therefore been renamed as the Dongba Formation, in which three members are distinguished, viz. the lower member is composed of red argillaceous siltstones intercalated with thin beds of gypsum; the middle one consists of dark brown argillaceous siltstones and silty mudstones intercalated with several thicker greyish-green argillaceous siltstones and a few thin beds of limestones; and the upper one is largely the same as the lower member in lithology.

3. Palynoflora and background of palynoprovince (Text-Fig. 3)

According to the features of palynological composition and vertical changes, three suites of floral successions can be distinguished in the late Cretaceous strata. They are described as follows in ascending order:

(1) *Schizaeoisporites-Taurocusporites-Psilatricolpites* Suite (from the lower and middle members of the Kukebai Formation).

Common to abundant species in the suite are:

Schizaeoisporites spp. (23-53 %)
Gleicheniidites senonicus (1-5 %)
Deltoidospora spp. (1-2 %)
Cyathidites minor (+)
Hymenophyllumsporites spp. (1-2 %)
Cicatricosisporites spp. (0-1.6 %)
Plicatella spp. (+)
Foraminisporis spp. (1-9 %)
Interulobites spp. (2-12 %)
Taurocusporites spp. (4.2-9 %)
Seductisporites spp. (0.5-18 %)
Trochicola spp. (0-1.7 %)
Trisolissporites spp. (0-1.6 %)
Ephedripites (E.) spp. (0-7 %)
Cycadopites spp. (3.6-8.1 %)
Monosulcites spp. (1-3 %)
Jugella spp. (0-1.5 %)
Saccate pollen (1-2.5 %)
Classopollis spp. (0-2.7 %)
Psilatricolpites spp. (1-4 %)
Retitricolpites spp. (0-3.6 %)
Archaeotriporopollis spp. (+)
Asteropollis sp. (+)

This suite is characterized by plentiful spores of pteridophytes such as *Schizaeoisporites, Taurocusporites, Foraminisporis, Interulobites* and *Seductisporites. Cicatricosisporites* and *Plicatella* often occurring in early Cretaceous are few in number. The gymnospermous pollen are mainly composed of *Cycadopites, Ephedripites* and *Monosulcites*, with rare bisaccate pollen. Angiospermous pollen occur in small amounts, about 2.7-9 %, including *Psilatricolpites, Retitricolpites* and *Liliacidites*, which are considered to be primitive forms. *Archaeotriporopollis* and *Asteropollis* make their first appearance in this suite. The assemblage indicates a tropical climate and a Cenomanian age.

(2) *Schizaeoisporites-Interulobites-Cranwellia* Suite (from the upper member of the Kukebai Formation).

The following species are common or characteristic elements within the suite:

Schizaeoisporites spp. (30-60 %)
Cyathidites minor (0.3-1.8 %)
Hymenophyllumsporites (0-5.7 %)
Divisisporites digitaformis (0-1.7 %)
Plicatella nankingensis (+)
Appendicisporis sp. (+)
Foraminisporis spp. (0-2.8 %)
Interulobites spp. (1-8.5 %)
Taurocusporites spp. (0-2.8 %)
Polycingulatisporites spp. (0.3-5.7 %)
Seductisporites spp. (1.1-5.1 %)
Trisolissporites spp. (0-2.8 %)
Ephedripites (E.) spp. (1-4.5 %)
Regalipollenites spp. (0.3-1.7 %)
Cyadopites spp. (2-4.3 %)
Jugella spp. (0.6-3.6 %)
Saccate pollen (1-1.6 %)
Psilatricolpites spp. (0-1.3 %)
Retitricolporites oblatus (+)
Archaeotriporopollis (+)
Cranwellia spp. (3.8-6.6 %)

This suite is recognized by the first occurrence of *Cranwellia* which may amount to 6.6 % in the assemblage. Other elements are largely the same as those of the first suite. Bisaccate pollen remain in small amounts. Pollen of angiosperms increase in number, being much more abundant and diverse than that of the first assemblage. The brevaxonate-type pollen, *Cranwellia*, is a prominent feature. In addition, there are also *Cupanieidites, Retitricolporites oblatus, Retitricolporoidites*, whereas *Psilatricolpites* begins to decrease gradually. This suite is confirmed to be of Turonian age by associated marine fossils.

(3) *Schizaeoisporites-Senegalosporites-Xinjiangpollis* Suite (from the lower and middle member of the Dongba Formation and Wuyitake Formation).

Main species present in this assemblage are:

Schizaeoisporites spp. (44-67 %)
Hymenophyllumsporites spp. (0-2 %)
Lygodioisporites smithianum (0-2.4 %)

L. spp. (1-6 %)
Foraminisporis spp. (0.6-2 %)
Interulobites spp. (0.3-1.3 %)
Polycingulatisporites spp. (0-2.4 %)
Taurocusporites spp. (0-0.6 %)
Seductisporites spp. (0.3-4.1 %)
Trisolissporites spp. (0-2.4 %)
Gabonisporis spp. (0.6-8.9 %)
Senegalosporites cretaceus (0-4.6 %)
S. wangchengensis (2.4-9.2 %)
Ephedripites (E.) spp. (2.4-9.2 %)
Cyadopites spp. (0-2.4 %)
Jugella spp. (0-1.7 %)
Saccate pollen (+)
Araucariacites australis (0-1.3 %)
Retitricolpites spp. (+)
Retitricolporoidites sp. (+)
Cranwellia spp. (0-5 %)
Xinjiangpollis spp. (1-3 %)
Galeacornea tarimensis (1-6.7 %)

The suite is characterized by the presence of *Senegalosporites* and *Xinjiangpollis*, etc., the majority of them are commonly observed and restricted to this suite. *Foraminisporis, Interulobites, Taurocusporites,* and *Cycadopites* began to decrease.

The age of the suite most likely belongs to late Turonian - early Senonian.

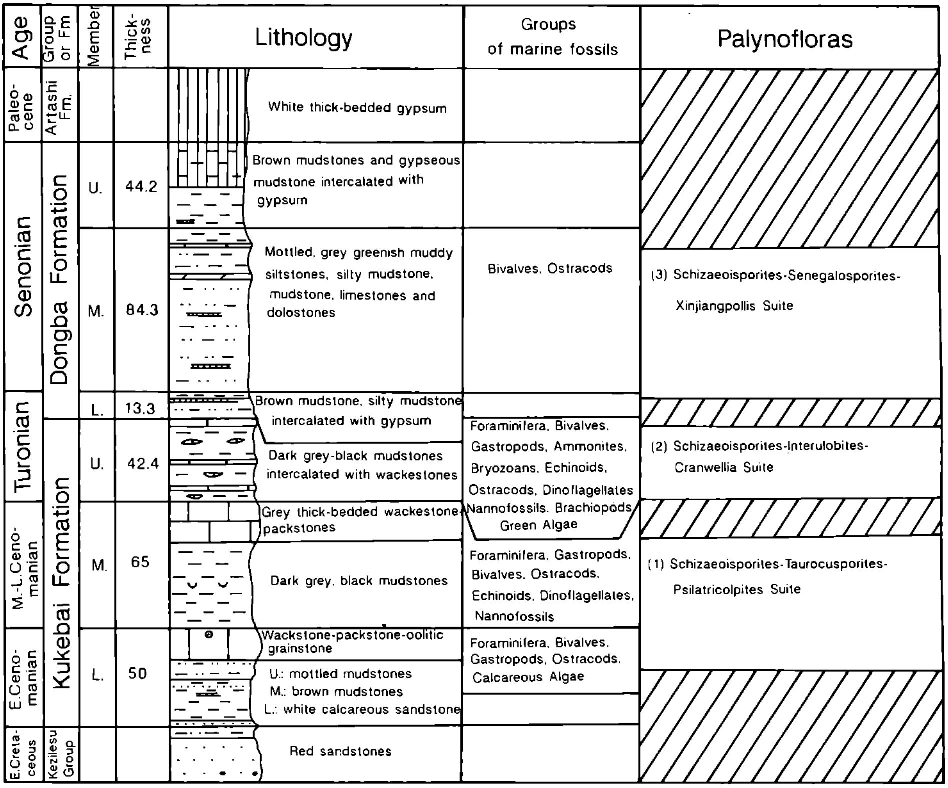
The palynomorphs are only sporadically present in a few stratigraphical units of the Upper Cretaceous along the foot of northern West Kunlun Mountains. The palynological sequence is discontinuous, presumably reflecting a comparatively arid climate.

The present palynofloras basically reflect the expected features of a vegetation surrounding the "Tarim Sea". The available data show that although the studied area was located in the Arid-belt of Central Asia-Southern China, the vegetation seems to have been flourishing and diverse during the early Upper Cretaceous along the foot of southern South Tiashan Mountains owing to the Eastern Tethyan extending into this area.

The affinity of most of the components in these assemblages is unknown at present. By comparing some types with spores of present plants, it seems that the climate was rather warm, corresponding to that of the tropical or southern subtropical zone of the present mainland of China. On the other hand, the frequent occurrence of Bryophyta spores allows us to assume that the northern coast of the "Tarim Sea" seems to have been comparatively moist, but the southern coast appears to have been arid as indicated by the meagre and discontinuous palynological record with the exception of the higher content of *Classopollis* in the assemblage.

Compared with the Middle Cretaceous palynogeoprovinces proposed by BRENNER (1976), the present Cenomanian-Turonian palynoflora appears to be similar to the South Laurasian Palynoprovince on the one hand, due to the occurrence of various Pteridophyta spores especially *Schizaeoisporites, Cicatricosisporites* and *Gleicheniidites*, etc.; and also appears to be similar to the North Gondwana Palynoprovince on the other hand, owing to the presence of great variety of striate pollen types and lack of saccate pollen. Thus, it can be said that the Cenomanian-Turonian palynoflora in the studied area bears possibly some characteristics of both the above mentioned palynoprovinces. Going into detail, the present Cenomanian-Turonian palynoflora is more similar to the Cenomanian-Turonian palynoflora of West Siberia studied by CHLONOVA (1962), due to the common presence of diagnostic and elegant *Trisolissporites* with three "radiata" which may serve as a marker to recognizing the Cenomanian and Turonian stages.

The early Senonian palynoflora contains more North Gondwana elements, particularly the diagnostic forms *Senegalosporites* and *Galeacornea*. These forms occupy a prominent position in the assemblage, but their species are unlike those found in West Africa. For example, we have not found *Galea-*



Text-Fig. 3. Palynostratigraphy of the Upper Cretaceous in Wuqia district, western Tarim Basin.

cornea causea, G. clavis and *Senegalosporites costatus* which are described in Senegal (West Africa) by STOVER (1963) and by JARDINE & MAGLOIRE (1965).

Besides, an element of angiospermous pollen having a brevaxonate outline, i. e. *Xinjiangpollis,* appears to have played an outstanding role in the early Senonian palynoflora, which has not yet been found from other areas of China. There are many species of *Xinjiangpollis* and their sum may reach a total of 5 %. However, they seem to have disappeared in overlying Tertiary strata.

The Upper Senonian consists of red siltstones and mudstones without any fossils. It is therefore difficult to ascertain whether the late Senonian palynoflora belongs to the Normapollis Province or to the Aquilapollenites Province.

Acknowledgements. The author wishes to express sincere thanks to Mr. OUYANG SHU and BAI CHENG-PENG for correcting the English manuscript and to ZANG XIAO-QUAN for drawing figures.

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Plate 1

All figures x 680.

Schizaeoisporites-Taurocusporites-Psilatricolpites Suite

Figs. 1-5. *Schizaeoisporites* spp.
Fig. 6. *Cyathdites minor* COUPER
Fig. 7. *Polycingulatisporites triangularis* ZHANG & ZHAN
Fig. 8. *P. caelatus* ZHANG & ZHAN
Fig. 9. *Cicatricosisporites minor* (BOLCH.) POCOCK

Fig. 10. *Hymenophyllumsporites granulatus* ZHANG & ZHAN
Fig. 11. *Interulobites symmetricus* ZHANG & ZHAN
Fig. 12. *Foraminisporis symmetricus* ZHANG & ZHAN
Fig. 13. *Heliosporites altmarkensis* SCHULZ
Fig. 14. *Seductisporites eminens* TSCHUDY
Fig. 15. *Trochicola xinjiangensis* ZHANG & ZHAN
Figs. 16, 17. *Taurocusporites segmentatus* STOVER
Fig. 18. *Trisolissporites stellatus* (CHLONOVA) TSCHUDY
Fig. 19. *Monosulcites* sp.
Figs. 20-22. *Cycadopites* spp.
Fig. 23. *Ephedripites (E.)* sp.
Fig. 24. *Classopollis annulatus* (VERB.) LI
Fig. 25. *Podocarpidites podocarpoides* (THG.) KRUTZSCH
Fig. 26. *Psilatricolpites parvulus* (GROOT & PENNY) NORRIS
Fig. 27. *Archaeotriporopollis singularis* ZHANG & ZHAN
Fig. 28. *Retitricolpites minutus* PIERCE
Fig. 29. *Retitricolporites oblatus* ZHANG & ZHAN

Plate 2

Schizaeoisporites-Interulobites-Cranwellia Suite

Figs. 1-4. *Schizaeoisporites* spp.
Fig. 5. *Punctatisporites labiosus* ZHANG & ZHAN
Fig. 6. *Cyathidites minor* COUPER
Fig. 7. *Foraminisporis obovatus* ZHANG & ZHAN
Fig. 8. *F. symmetricus* ZHANG & ZHAN
Fig. 9. *Gabonisporis vigourouxii* BOLTENHAGEN
Fig. 10. *Polycingulatisporites verus* ZHANG & ZHAN
Fig. 11. *Zlivisporis unioreticulatus* ZHANG & ZHAN
Fig. 12. *Taurocusporites segmentatus* STOVER
Fig. 13. *Interulobites symmetricus* ZHANG & ZHAN
Fig. 14. *Hymenophyllumsporites granulatus* ZHANG & ZHAN
Fig. 15. *Trisolissporites stellatus* (CHLONOVA) TSCHUDY
Fig. 16. *Regalipollenites granulatus* ZHANG & ZHAN
Fig. 17, 18. *Cycadopites* spp.
Fig. 19. *Jugella claribaculata* MTCH. & SHAK.
Fig. 20. *Ephedripites (E.) viesensis* KRUTZSCH
Fig. 21. *Podocarpidites* sp.
Fig. 22, 23. *Retitricolpites georgensis* BRENNER
Fig. 24. *Extratriporopollenites thiergarti minutus* PFLUG
Figs. 25-27. *Cranwellia* spp.
Fig. 28. *Sindorapollis* cf. *granulatus* TSCHUDY
Fig. 29. *Retitricolporites oblatus* ZHANG & ZHAN

Plate 3

Schizaeoisporites-Senegalosporites-Xinjiangpollis Suite

Figs. 1-6. *Schizaeoisporites* spp.
Figs. 7-9. *Senegalosporites* spp.

Plate 3 (continued)

Fig. 10. *Polycingulatisporites verus* ZHANG & ZHAN
Fig. 11. *Foraminisporis baculus* ZHANG & ZHAN
Fig. 12. *Trisolissporites stellatus* (CHLONOVA) TSCHUDY
Fig. 13. *Interulobites circinatus* ZHANG & ZHAN
Fig. 14. *Gabonisporis reticulatus* ZHANG & ZHAN
Fig. 15. *Hymenophyllumsporites granulatus* ZHANG & ZHAN
Fig. 16. *Lygodioisporites bellus* ZHANG & ZHAN
Fig. 17. *Trilobosporites bernissartensis* (DEL. & SPRUM) POT.
Fig. 18. *Cycadopites minimus* (COOKSON) KRUTZSCH
Fig. 19. *Ephedripites (E.) wolkenbergensis* KRUTZSCH
Figs. 20-21. *Galeacornea tarimensis* ZHANG & ZHAN
Fig. 22. *Cranwellia crassicolpita* YU
Fig. 23. *Retitricolpites georgensis* BRENNER
Figs. 24-27. *Xinjiangpollis* spp.

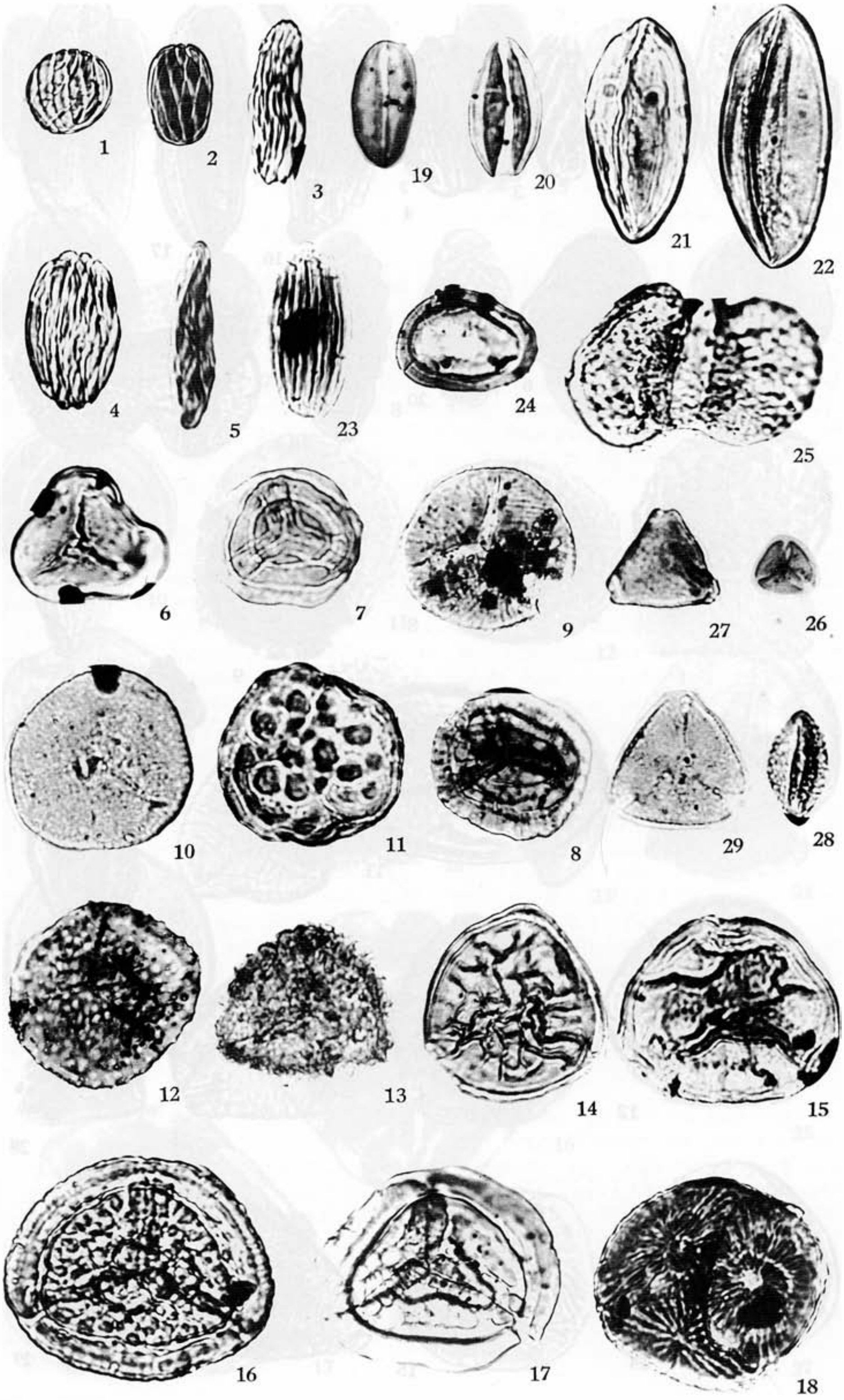

Plate 1

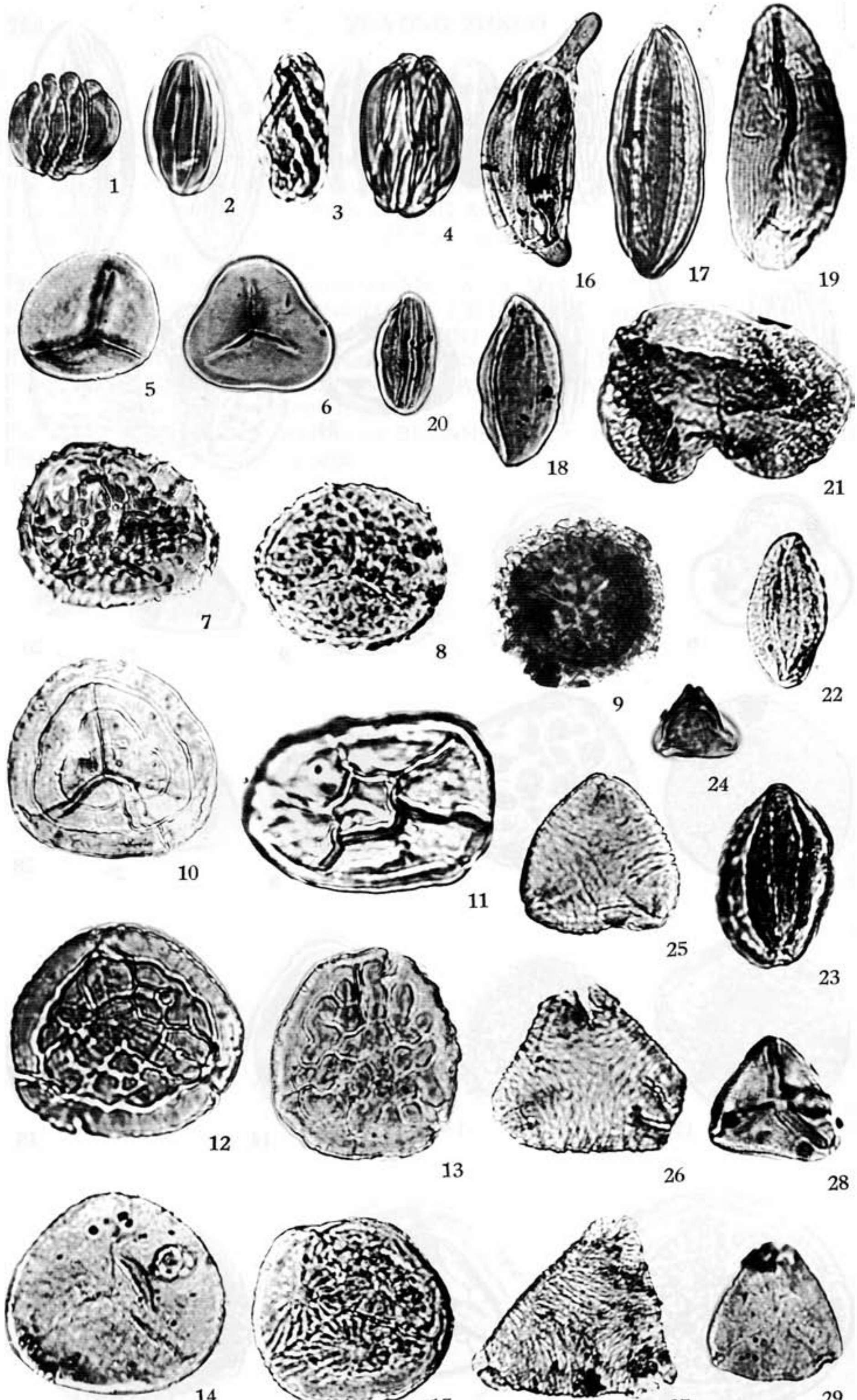

Plate 2

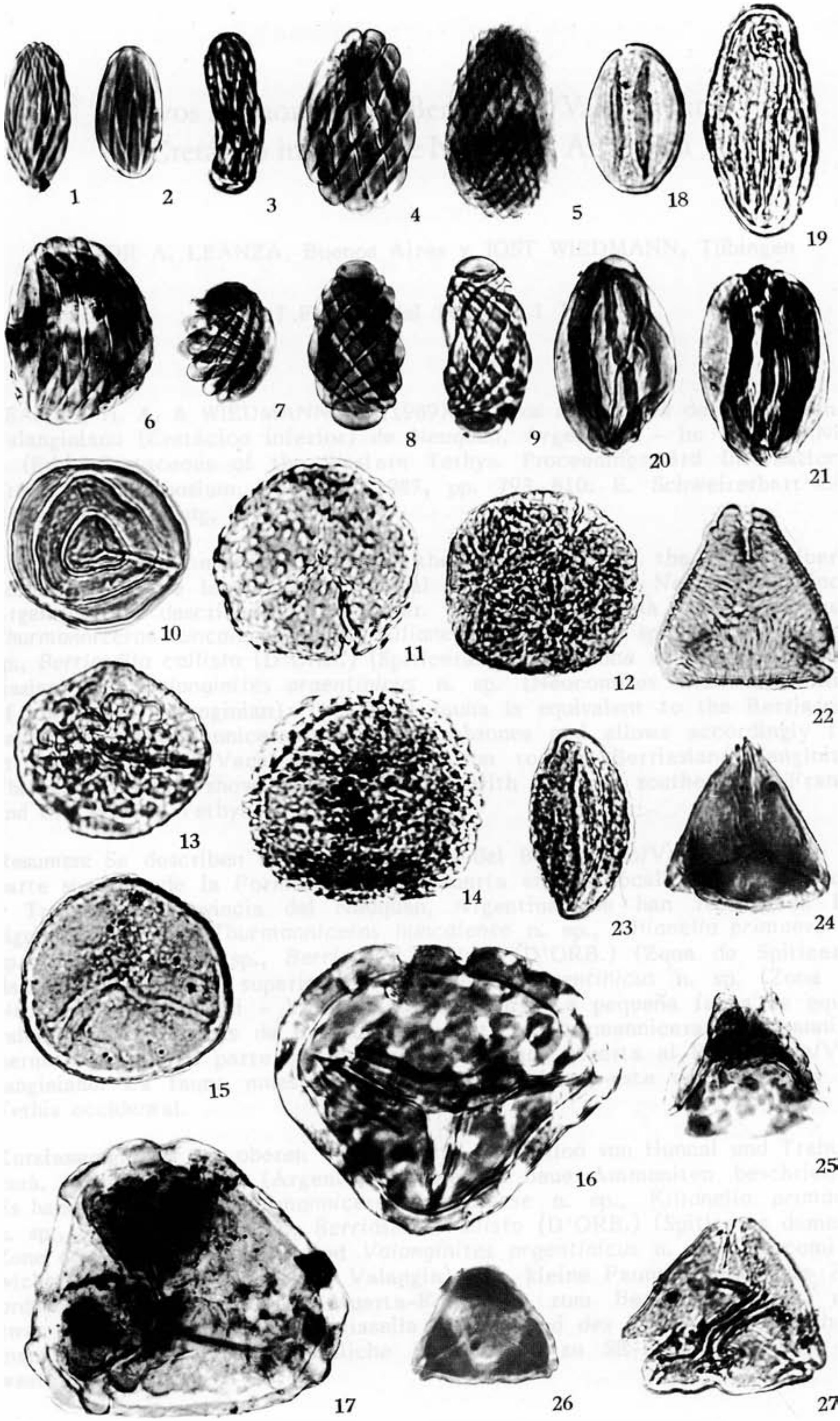

Plate 3

Nuevos ammonites del Berriasiano/Valanginiano (Cretácico inferior) de Neuquén, Argentina

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Con 7 Figuras del Texto y 1 Tabla

LEANZA, H. A. & WIEDMANN, J. (1989): Nuevos ammonites del Berriasiano/Valanginiano (Cretácico inferior) de Neuquén, Argentina. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 793-810. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: New ammonites found in the upper part of the Vaca Muerta Formation in the localities of Huncal and Trahuncurá, Neuquén province, Argentina, are described in this paper. The following taxa were recognized: *Thurmanniceras huncalense* n. sp., *Kilianella primaeva* n. sp., *Protancyloceras* sp., *Berriasella callisto* (D'ORB.) (Spiticeras damesi Zone of uppermost Berriasian) and *Valanginites argentinicus* n. sp. (Neocomites wichmanni Zone of lowermost Valanginian). The small fauna is equivalent to the Berriasella callisto and Thurmanniceras thurmanni subzones and allows accordingly the attribution of the Vaca Muerta Formation to the Berriasian/Valanginian. The studied fauna shows close affinities with that of southeastern France and the western Tethys.

Resumen: Se describen nuevos ammonites del Berriasiano/Valanginiano de la parte superior de la Formación Vaca Muerta en las localidades de Huncal y Trahuncurá, provincia del Neuquén, Argentina. Se han reconocido los siguientes taxones: *Thurmanniceras huncalense* n. sp., *Kilianella primaeva* n. sp., *Protancyloceras* sp., *Berriasella callisto* (D'ORB.) (Zona de Spiticeras damesi - Berriasiano superior) y *Valanginites argentinicus* n. sp. (Zona de Neocomites wichmanni - Valanginiano inferior). La pequeña fauna es equivalente a las subzonas de Berriasella callisto y Thurmanniceras thurmanni y permite atribuir la parte superior de la F. Vaca Muerta al Berriasiano/Valanginiano. La fauna muestra afinidades con el sud-este de Francia y el Tethis occidental.

Kurzfassung: Aus der oberen Vaca Muerta-Formation von Huncal und Trahuncurá, Provinz Neuquén (Argentinien), werden neue Ammoniten beschrieben. Es handelt sich um: *Thurmanniceras huncalense* n. sp., *Kilianella primaeva* n. sp., *Protancyloceras* sp., *Berriasella callisto* (D'ORB.) (Spiticeras damesi-Zone des obersten Berrias) und *Valanginites argentinicus* n. sp. (Neocomites wichmanni-Zone des untersten Valangin). Die kleine Fauna erlaubt die Zuordnung der höheren Vaca Muerta-Formation zum Berrias-Valangin, und zwar zu den Subzonen der Berriasella callisto und des Thurmanniceras thurmanni. Die Fauna zeigt deutliche Beziehungen zu SE-Frankreich und der westlichen Tethys.

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1. Introducción

El presente trabajo tiene por objeto dar a conocer un conjunto de interesantes ammonites hallados en la parte superior de la Formación Vaca Muerta entre los parajes de Huncal y Trahuncurá, en la parte occidental de la provincia del Neuquén, Argentina (Fig. 1). Los mismos permiten efectuar una segura correlación con las subzonas de Berriasella callisto y Thurmanniceras thurmanni del sud-este de Francia, cuyas edades corresponden al Berriasiano superior y al Valanginiano inferior respectivamente.

La mayoria del material fue donado para su estudio por JOSE I. GARATE (Neuquén, Zapala), mientras que el resto fue coleccionado por los autores, quienes en Abril de 1985 tuvieron oportunidad de recorrer nuevamente la región precisando la proveniencia estratigráfica del material.

Ubicación de las localidades fosilíferas

Los taxones *K. primaeva* n. sp., *Th. huncalense* n. sp. y *Protancyloceras* sp. fueron hallados en la región situada 8 km al suroeste de la escuela de Huncal, en dirección a Loncopué. En línea recta, Huncal dista 67 km al noroeste de Bajada del Agrio y 33 km al noreste de Loncopué. *B. callisto* y *V. argentinicus* n. sp. fueron hallados en el faldeo occidental del Cerrito

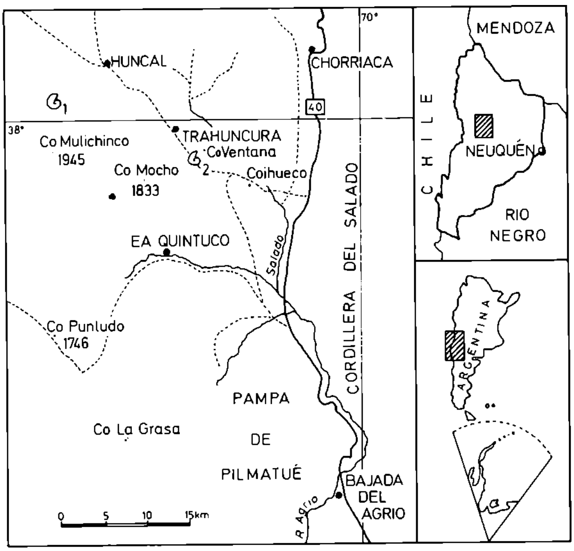

Fig. 1. Mapa de ubicación mostrando las localidades fosilíferas de 1) Huncal, 2) Trahuncurá.

de la Ventana en la localidad de Trahuncurá, situada 48 km al noroeste de Bajada del Agrio (Fig. 1). Ambas localidades fosilíferas, indicadas con 1) y 2) en el mapa de ubicación, están situadas en la Hoja 33 b, El Huecú, de la carta geológico-económica de la República Argentina a escala 1 : 200.000.

Preservación de los fósiles

Los restos de *K. primaeva* n. sp., *Th. huncalense* n. sp. y *Protancyloceras* sp. se hallan piritizados y exhiben una coloración pardo rojiza. Probablemente como resultado de este tipo de fosilización solamente se han conservado las vueltas del fragmocono, sin que en ninguno de los 265 ejemplares examinados esté presente la cámara de habitación. Si bien ello constituye un obstáculo para arribar a una clasificación satisfactoria, el excelente estado de las partes conservadas ha hecho factible su exacta ubicación taxonómica. La característica saliente de este grupo de ammonites es su pequeño tamaño, estando la mayoría de ellos comprendidos en un diámetro de 5 a 10 mm. Solo como excepción, un ejemplar de *Th. huncalense* n. sp. alcanza los 27 mm de diámetro.

V. argentinicus n. sp. y el fragmento de *B. callisto* presentan una conservación normal.

Abreviaturas

Las abreviaturas empleadas en las descripciones sistemáticas son las siguientes:

D = diámetro máximo en un determinado estado de crecimiento
U = diámetro del ombligo medido en la sutura umbilical
H = altura de la última vuelta medida en el plano de arrollamiento
W = ancho máximo de la última vuelta en ángulo recto con el plano de arrollamiento

Solamente se ofrecen las dimensiones de los ejemplares figurados en este trabajo.

2. Estratigrafía

La geología de la región donde proceden los fósiles fue dada a conocer por H. LEANZA (1973: 102). Allí existen extensos afloramientos de pelitas negras de la Formación Vaca Muerta (WEAVER 1931, emend. H. LEANZA 1972) cubiertas por conspicuas cornisas de areniscas de la Formación Mulichinco (WEAVER 1931) que determinan el flanco septentrional del valle de Huncal-Trahuncurá, girando luego en dirección meridional según un arco que pasa al oeste de Coihueco y al norte de la Ea. Quintuco, quedando aislados en el sector central a modo de relictos los cerros Mulichinco y Mocho.

En ocasión de describir el género *Acantholissonia*, H. LEANZA (1972: 64) dió a conocer la sección estratigráfica del Cto. de la Ventana, consignando para la Formación Vaca Muerta, que aflora sin base visible, un espesor de 343 m, distinguiendo en ella las siguientes asociaciones de ammonites: 1) *Thurmanniceras neogaeus* y *Th.* aff. *keideli*, más dos especies indeterminadas de *Thurmanniceras*, y 2) *Lissonia riveroi* y *Acantholissonia* *gerthi*. A esta lista se agrega ahora a la primera asociación *Berriasella callisto* y a la segunda *Valanginites argentinicus* n. sp.

H. LEANZA & C. HUGO (1977: 256) efectuaron una sección estratigráfica a lo largo del camino que une Huncal con Pichaihue, 12 km al noroeste de Trahuncurá, donde la Formación Vaca Muerta acusa un espesor parcial de 826 m, desde el nivel del valle hasta el contacto superior con la Formación Mulichinco. Según los mismos autores, en la Formación Vaca Muerta pueden distinguirse en orden ascendente las zonas de Substeueroceras koeneni, Argentiniceras noduliferum, Spiticeras damesi y Neocomites wichmanni, que indican edades del Tithoniano superior, Berriasiano inferior y superior, y Valanginiano inferior respectivamente.

Los taxones *K. primaeva* n. sp., *Th. huncalense* n. sp. y *Protancyloceras* sp. proceden de la Zona de Spiticeras damesi, es decir de los niveles que portan *Berriasella callisto* en Trahuncurá.

Repositorio

Los ejemplares estudiados se encuentran depositados en el Institut und Museum für Geologie und Paläontologie de la Universidad de Tübingen (GPIT) de la República Federal Alemana (Sigwartstraße 10, D-7400 Tübingen, BRD), y en el Museo Juan Olsacher (MOZ) de la Dirección General de Minería de Neuquén (Olascoaga 421, 8340 Zapala, Neuquén, Arg.). También hay moldes en yeso disponibles en los repositorios del Servicio Geológico Nacional (Av. Santa Fé 1548, 1060 Buenos Aires, Arg.) y del Museo de La Plata (Paseo del Bosque, 1900 La Plata, Arg.).

3. Descripciones sistemáticas

Orden	Ammonoidea ZITTEL 1884
Suborden	Ammonitina HYATT 1889
Superfamilia	Perisphinctaceae STEINMANN 1890
Familia	Neocomitidae SALFELD 1921
Subfamilia	Neocomitinae SALFELD 1921

Género *Kilianella* UHLIG 1905

Especie tipo: *Hoplites pexiptychus* UHLIG 1881, por designación subsecuente de ROMAN 1938.

Kilianella primaeva n. sp.
Fig. 2. 1-5

Holotipo: Ejemplar GPIT 1674/7; Fig. 2. 2.
Locus typicus: Huncal, provincia del Neuquén, Argentina.
Stratum typicum: Parte superior de la Formación Vaca Muerta.
Edad: Berriasiano superior, Zona de Spiticeras damesi.

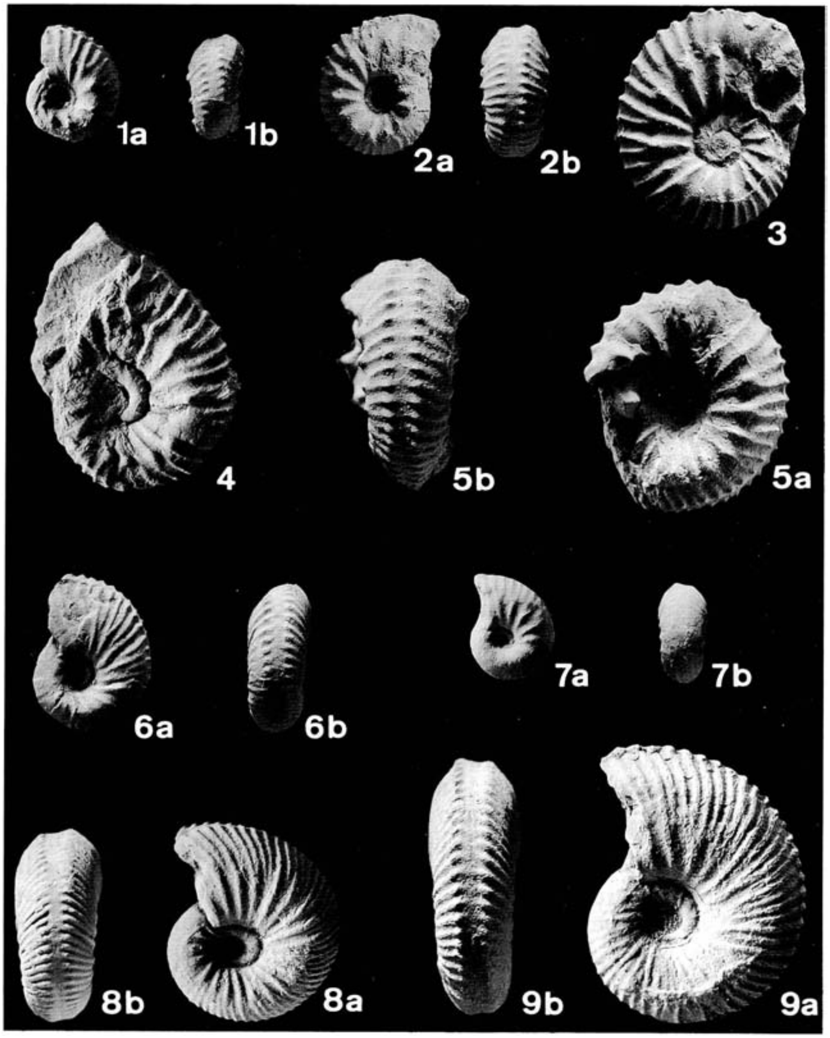

Fig. 2. 1 - 5: *Kilianella primaeva* n. sp. 3/1.
1a-b: vistas lateral y ventral, GPIT 1674/6.
2a-b: vistas lateral y ventral del Holotipo, GPIT 1674/7.
3: vista lateral, GPIT 1674/8.
4: vista lateral, GPIT 1674/9.
5a-b: vistas lateral y ventral, GPIT 1674/10.
6 - 9: *Thurmanniceras huncalense* n. sp. 3/1.
6a-b: vistas lateral y ventral, GPIT 1674/16.
7a-b: vistas lateral y ventral, GPIT 1674/12.
8a-b: vistas lateral y ventral, GPIT 1674/13.
9a-b: vistas lateral y ventral, GPIT 1674/14.
Berriasiano superior. Zona de Spiticeras damesi. Formación Vaca Muerta. Huncal, Neuquén, Argentina.

Derivatio nominis: Alude al carácter primitivo de esta nueva especie de *Kilianella*, cuyo acmé se registra en el Valanginiano.

Diagnosis: Conchilla moderadamente involuta, con sección de las vueltas apenas más alta que ancha, algo inflada. Ornamentación constituida por costillas simples, algo proyectadas. Algunas de ellas se bifurcan en dos ramas, la posterior levemente retroversa, existiendo en el punto de bifurcación un tubérculo pequeño y puntiagudo. Hay también algunas costillas intercaladas que se pierden en la mitad del flanco. Todas las costillas terminan en pequeños nodos lateroventrales, permitiendo la presencia de una banda ventral lisa. Constricciones no muy frecuentes y poco profundas. Línea de sutura simple, de tipo neocomítido. Cámara de habitación desconocida.

Material

27 ejemplares, de los cuales 25 muestran solamente los estadios ontogenéticos iniciales y las primeras vueltas, mientras que los 2 restantes consisten en fragmentos de fragmoconos. La cámara de habitación no está conservada en ningún ejemplar. Los 5 ejemplares figurados en el presente trabajo se identifican bajo la sigla GPIT 1674/6-10.

Descripción

Conchilla moderadamente involuta (rel. O/D = 0,21 a 0,30), con sección de las vueltas apenas más alta que ancha, algo inflada. A partir de temprana edad la ornamentación comienza a manifestarse vigorosa y con peculiares características. Las costillas primarias son levemente proyectadas, bifurcándose a la altura de la mitad del flanco en dos ramas, la posterior de las cuales es algo rursiradiada. En el punto de bifurcación aparecen tubérculos pequeños y puntiagudos, siendo el relieve de las costillas resultantes menos vigoroso que el de las primarias. También existen costillas intercaladas que se pierden en la mitad del flanco. Los diferentes tipos de costillas culminan en el borde lateroventral en un pequeño nodo, interrumpiéndose totalmente sobre el vientre permitiendo la presencia de una banda ventral lisa. También existen de tanto en tanto constricciones no muy marcadas. La línea de sutura es simple, de tipo neocomítido. Como ya se expresó, nada puede agregarse acerca de la morfología de las últimas vueltas y de la cámara de habitación, que permanece desconocida.

Dimensiones (en mm), ejemplares figurados solamente:

Ejemplar nº	D	O	O/D	H	W	H/W
GPIT 1674/6	5,0	1,3	0,26	2,6	2,5	1,04
GPIT 1674/7	6,0	1,6	0,26	3,1	3,0	1,03
GPIT 1674/8	9,5	2,0	0,21	4,5	-	-
GPIT 1674/9	10,5	3,0	0,30	5,0	-	-
GPIT 1674/10	10,0	3,0	0,30	4,5	4,2	1,07

Observaciones

El material descripto exhibe una evidente y estrecha relación con *Kilianella* UHLIG 1905 (especie tipo: *Hoplites pexiptychus* UHLIG 1881, lám. 4, figs.

4-5), género que posee conchillas de pequeño tamaño cuyas vueltas internas, especialmente aquellas de algunas especies figuradas por SAYN (1907) se asemejan sumamente a las de nuestro material, si bien muestran constricciones más numerosas y profundas.

De las 8 especies y 2 variedades de *Kilianella* figurados por SAYN (1907) del Valanginiano del sud-este de Francia, que también se hallan coincidentemente piritizadas, nuestro material recuerda especialmente al ejemplar juvenil de *K. lucensis* SAYN (1907: 50, lám. 6, fig. 13), aunque este último posee constricciones más frecuentes y profundas. *Neocomites? longi* SAYN (1907: 37, lám. 3, fig. 19; lám. 4, figs. 1-2) también resulta muy similar para iguales estadios ontogenéticos, aunque aparenta tener una costulación mucho más densa que nuestros ejemplares. Según nuestra opinión, se duda que esa especie pertenezca a *Neocomites,* apareciendo mucho más vinculada con *Kilianella*. En el museo del Instituto de Paleontología y Geología Histórica de la Universidad de München existen ejemplares del Valanginiano de los Alpes Bajos, cerca de Sisterón, clasificados *in schedis* como "*Thurmannites (Kilianella) asperrimus* D'ORB." que son sumamente parecidos a *K. primaeva* n. sp.

Con respecto a la cita de *Kilianella* en Argentina hecha por GERTH (1925: 98), se sabe que es errónea, dado que este autor incluyó en ese género a *Hoplites burckhardti* MAYER-EYMAR (in BURCKHARDT 1903: 61, lám. 10, figs. 17-20) que constituye la especie tipo de *Lytohoplites* SPATH (1925: 144). Por su parte, GIOVINE (1950) figuró con interrogante a una especie indeterminada de *Kilianella* (op. cit.: 46, lám. 7, fig. 5) consistente en un fragmento de vuelta procedente del Hauteriviano de Neuquén.

Distribución

K. primaeva n. sp. está asociada con *Th. huncalense* n. sp. y *Protancyloceras* sp. en sedimentitas de la parte alta de la Formación Vaca Muerta que corresponden a la Zona de Spiticeras damesi, cuya edad equivale al Berriasiano superior.

Género *Thurmanniceras* COSSMAN 1901

Especie tipo: *Ammonites thurmanni* PICTET & CAMPICHE 1858-60.

Thurmanniceras huncalense n. sp.
Fig. 2. 6-9; Fig. 3; Fig. 4

1931 ? *Thurmanniceras quintucoense* WEAVER, p. 447, lám. 43, fig. 292; lám. 56, figs. 358-359.

Holotipo: Ejemplar GPIT 1674/15, Fig. 3a, b.
Locus typicus: Huncal, provincia del Neuquén, Argentina.
Stratum typicum: Parte superior de la Formación Vaca Muerta.
Edad: Berriasiano superior, Zona de Spiticeras damesi.
Derivatio nominis: Alude al lugar de su hallazgo.

Diagnosis: Conchilla moderadamente involuta, con vueltas más altas que anchas, proporción que se acentúa con la edad. Costillas primarias fuertes, radiales en la región interna y fuertemente proyectadas en la región externa, bifurcándose en la mitad del flanco o bien permaneciendo simples. Costillas intercaladas frecuentes que se pierden en la mitad del flanco. Todas finalizan en el borde lateroventral en un pequeño nodo, permitiendo la presencia de una banda ventral lisa. Línea de sutura simple, de tipo neocomítido. Cámara de habitación desconocida.

Material

208 ejemplares, todos los cuales consisten en vueltas internas. La cámara de habitación no se halla preservada en ningún ejemplar. El tamaño de los ejemplares oscila entre los 3 y 15 mm de diámetro, con excepción del holotipo, que alcanza 27 mm. Los 5 ejemplares figurados se identifican con la sigla GPIT 1674/11-16.

Descripción

Los estadios ontogenéticos más jóvenes presentan conchilla globosa, apenas más alta que ancha. A partir de los 5 mm de diámetro, la altura gana a expensas del ancho. Los flancos son convexos, pasando insensiblemente a la pendiente umbilical, mientras que la periferia es subplanada. La relación O/D varia entre 0,24 y 0,25. La ornamentación de los estadios ontogenéticos iniciales (hasta 4 mm de diámetro) (ver Fig. 2. 7a-b) recién empieza a definirse en el cuarto final de la primera vuelta, mostrando primarias fuertes que se dividen en ramas tenuemente marcadas que se extinguen hacia la periferia. Entre los 5 y los 10 mm de diámetro (Fig. 2. 6-9) existen costillas rectiradiadas en la región interna del flanco y fuertemente prorsiradiadas en la región externa. En el tercio interno del flanco suelen dividirse en dos ramas del mismo grosor, sin que aparezca ningún tipo de tubérculo en el lugar donde se produce la bifurcación. Existen costillas intercaladas que se pierden en la mitad del flanco. Todas culminan en un pequeño nodo, permitiendo la presencia de una banda ventral lisa. En el holotipo (Fig. 3a-

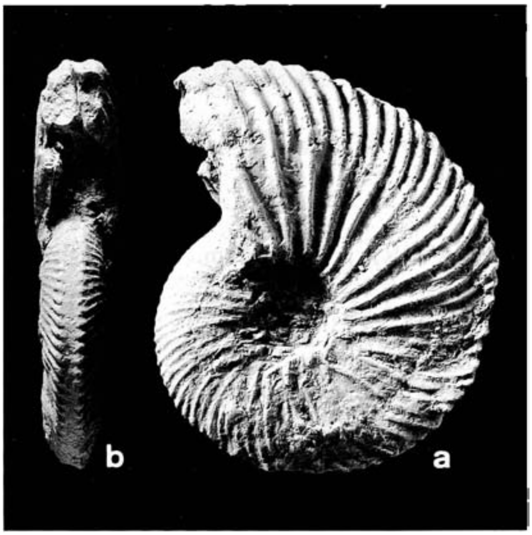

Fig. 3. *Thurmanniceras huncalense* n. sp. Holotipo, 2/1.
a-b: vistas lateral y ventral, GPIT 1674/15. Berriasiano superior. Zona de Spiticeras damesi. Formación Vaca Muerta. Huncal, Neuquén, Argentina.

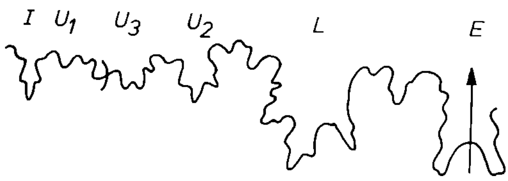

Fig. 4. Línea de sutura de *Thurmanniceras huncalense* n. sp. GPIT 1674/16. Altura de vuelta = 5,5. 10/1.

b), se puede apreciar la costulación resultante en un estadio ontogenético más avanzado, con costillas fuertes y filosas, apenas flexuosas, rectiradiadas en la región interna del flanco y prorsiradiadas en la región externa. A su vez, la altura de la división de las costillas se eleva hasta la mitad del flanco, mientras que algunas permanecen simples y otras se intercalan. Todas las costillas culminan en un pequeño nodo lateroventral permitiendo la presencia de una banda ventral lisa, proporcionalmente más estrecha que en los estadios iniciales. La línea de sutura (Fig. 4) es simple, de tipo neocomítido, con L trífido. La cámara de habitación permanece desconocida.

Dimensiones (en mm), ejemplares figurados solamente:

Ejemplar n°	D	O	O/D	H	W	H/W
GPIT 1674/11	4,1	1,0	0,24	2,0	2,0	1,00
GPIT 1674/12	6,0	1,5	0,25	3,0	2,5	1,20
GPIT 1674/13	8,0	2,0	0,25	5,0	3,5	1,41
GPIT 1674/14	12,0	3,0	0,25	6,0	4,0	1,50
GPIT 1674/15	27,0	6,5	0,24	14,0	6,0	2,30

Observaciones

El tipo de ornamentación del material descripto coincide con aquella del género *Thurmanniceras* COSSMANN 1901 (especie tipo: *A. thurmanni* PICTET & CAMPICHE 1858-60). Si bien en este género las costillas tienden a atravesar el vientre sin interrumpirse, no ocurre lo mismo en las vueltas internas, las cuales están provistas de banda ventral lisa. Si bien posee afinidades, el género *Cuyaniceras* A. LEANZA 1945 (especie tipo: *Odontoceras transgrediens* STEUER 1897) se diferencia por poseer en las vueltas internas costillas que comienzan a dividirse a la altura del borde umbilical.

Especificamente, el material descripto encuentra estrechas afinidades con varias especies de *Thurmanniceras* del Valanginiano del sud-este de Francia descriptas por SAYN (1907), aunque éstas, en similares estadios ontogenéticos, se hallan invariablemente provistas de constricciones más o menos profundas y de distinta frecuencia, las cuales no han sido detectadas en los ejemplares argentinos. Exceptuando estas diferencias, el ejemplar juvenil de *Th. thurmanni* figurado por SAYN (1907, lám. 5, fig. 3) resulta muy similar a nuestros especímenes. En el Museo del Instituto de Paleontología y Geología Histórica de la Universidad de München existen ejemplares del Valanginiano de los Alpes Bajos, cerca de Sisterón, clasificados in schedis como "*Neocomites (Lyticoceras) neocomiensis* D'ORB." que son sumamente parecidos a *Th. huncalense* n. sp.

Es posible que eventualmente *Thurmanniceras quintucoense* WEAVER (1931, p. 447, lám. 43, fig. 292; lám. 56, figs. 358-9) represente el estado adulto de nuestra nueva forma, pero de cualquier modo ello no pudo ser comprobado, pues presenta estadios ontogenéticos diferentes. Las restantes especies argentinas de *Thurmanniceras* figuradas por GERTH (1925, 1926), WEAVER (1931) y A. LEANZA (1945) no presentan analogías.

Atendiendo a las consideraciones expuestas, se propone designar al material descripto como una nueva especie de *Thurmanniceras*, denominada *Th. huncalense* n. sp.

Distribución

Th. huncalense n. sp. está asociado con *K. primaeva* n. sp. y *Protancyloceras* sp. en sedimentitas de la parte alta de la Formación Vaca Muerta que corresponden a la Zona de Spiticeras damesi, cuya edad equivale al Berriasiano superior.

Subfamilia Berriasellinae SPATH 1922

Género *Berriasella* UHLIG 1905

Especie tipo: *Ammonites privasensis* PICTET 1867, por designación subsecuente de ROMAN 1938.

Berriasella callisto (D'ORBIGNY 1847)
Fig. 5

1973 *Berriasella callisto* (D'ORB.). - LE HEGARAT: 53, lám. 4, figs. 1-3; lám. 38, fig. 11. (Con sinonimia).
1987 *B. callisto* (D'ORB.). - COMPANY: 104, lám. 3, figs. 9-11; lám. 18, fig. 4. (Con sinonimia).

Material

Un fragmento de vuelta aplastado, GPIT 1674/1; Fig. 5.

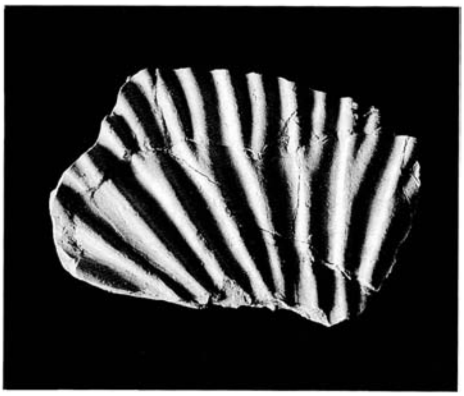

Fig. 5. *Berriasella callisto* (D'ORB.). Fragmento de fragmocono, GPIT 1674/1. 2/1. Berriasiano superior. Zona de Spiticeras damesi. Formación Vaca Muerta. Trahuncurá, Neuquén, Argentina.

Descripción

Se dispone de un fragmento (24 x 15 mm) en el que son visibles costillas de nítido relieve, prosoversas, netamente flexuosas, separadas por interespacios un poco más anchos que ellas. Todas las costillas se bifurcan en la mitad del flanco según ángulos bastante estrechos.

Observaciones

Si bien se trata de un pequeño fragmento, pueden reconocerse en él los típicos rasgos de *Berriasella callisto* (D'ORBIGNY 1847: 551, lám. 213, figs. 1-2) reilustrada por LE HEGARAT (1973, lám. 4, figs. 1-3; lám. 38, fig. 11) al encargarse del estudio del Berriasiano del sud-este de Francia. En tal sentido, *B. callisto* constituye la especie índice de la subzona más alta del Berriasiano francés.

Distribución

B. callisto fue hallada en el Cto. Ventana en sedimentitas de la Formación Vaca Muerta en la Zona de Spiticeras damesi que en consecuencia puede correlacionarse en sentido amplio con la Zona de Fauriella boissieri del sud-este de Francia.

Familia Olcostephanidae HAUG 1910
Subfamilia Olcostephaninae HAUG 1910

Género *Valanginites* KILIAN 1910

Especie tipo: *Ammonites nucleus* (PHILIPPS?) ROEMER 1841.

Valanginites argentinicus n. sp.
Fig. 6. 1a-c

Holotipo: Ejemplar GPIT 1674/2; Fig. 6. 1.
Locus typicus: Trahuncurá, provincia del Neuquén, Argentina.
Stratum typicum: Parte superior de la Formación Vaca Muerta.
Edad: Valanginiano inferior, Zona de Neocomites wichmanni.
Derivatio nominis: Alude a la República Argentina.

Diagnosis

Conchilla globosa, extremadamente involuta (rel. O/D = 0,12), con ombligo profundo y borde umbilical redondeado. Cámara de habitación ornamentada por tubérculos periumbilicales de los que parten dos costillas de escaso relieve que atraviesan el vientre sin interrupción. Hacia la apertura la superficie de la conchilla se torna prácticamente lisa, al tiempo que los tubérculos periumbilicales son más distantes y puntiagudos.

Material

Un ejemplar GPIT 1674/2. Correspondiendo casi el 90 % de la última vuelta a la cámara de habitación.

Descripción

Conchilla globosa, casi esférica, extremadamente involuta (rel. O/D = 0,12). Ombligo profundo con bordes umbilicales redondeados. La cámara de habitación ocupa casi el 90 % de la última vuelta. Dada la gran involución, las vueltas internas no son visibles. Existen tubérculos periumbilicales en los que comienzan un par de costillas radiales muy poco prominentes y de sección redondeada que atraviesan transversalmente el vientre sin interrupción. En la última mitad de la cámara de habitación la superficie de la conchilla se torna prácticamente lisa al desaparecer las costillas, carácterística que distingue a esta especie de las demás conocidas. Al mismo tiempo, los tubérculos periumbilicales se tornan más distantes y puntiagudos.

Dimensiones (en mm):

Ejemplar n°	D	O	O/D	H	W	H/W
GPIT 1674/2	40,5	5	0,12	17	36	0,47

Observaciones

El género *Valanginites* KILIAN caracteriza a formas globosas extremadamente involutas de talla mediana a pequeña (25 a 60 mm) con costillas primarias radiales de escaso relieve que parten generalmente en haces desde el borde umbilical, pudiendo formar allí bullas moderadas ó tubérculos agudos. *Dobrodgeiceras* NIKOLOV 1962 (especie tipo: *D. ventrotuberculatum* NIKOLOV 1962) representado en Perú por *D. broggianum* (LISSON 1937) (cf. RICCARDI & WESTERMANN 1970) tiene afinidades con *Valanginites* pero se diferencia de éste por la presencia de tubérculos ventrolaterales y costillas primarias muy prominentes. Dadas sus características morfológicas, nuestro ejemplar cae naturalmente en *Valanginites* KILIAN.

Específicamente, *V. argentinicus* n. sp. se diferencia de *V. nucleus* (ROEMER) (cf. THIEULOY 1977: 426, lám. 8, figs. 23-24) del Valanginiano superior (Zona de S. verrucosum) del sud-este de Francia, por poseer costulación más débil y tubérculos periumbilicales.

V. psaephoides paludensis THIEULOY (1977: 431, lám. 9, figs. 19-20) de la misma región posee, como la especie argentina, tubérculos periumbilicales, pero las costillas aparecen muy bien marcadas en toda la cámara de habitación. *V. tijerensis* IMLAY (1937: 562, lám. 72, figs. 4-5) del arroyo San Lázaro, Nuevo León, México, posee también en la cámara de habitación tubérculos periumbilicales pero se diferencia de *V. argentinicus* n. sp. por exhibir haces de 3-5 costillas y constricciones bien marcadas. "*Valanginites*" *angusticoronatus* IMLAY (1938: 557, lám. 4, figs. 1-3) de la Formación Taraises en la Sierra de Parras, México, fué transferido por THIEULOY (1977: 426) al género *Olcostephanus*.

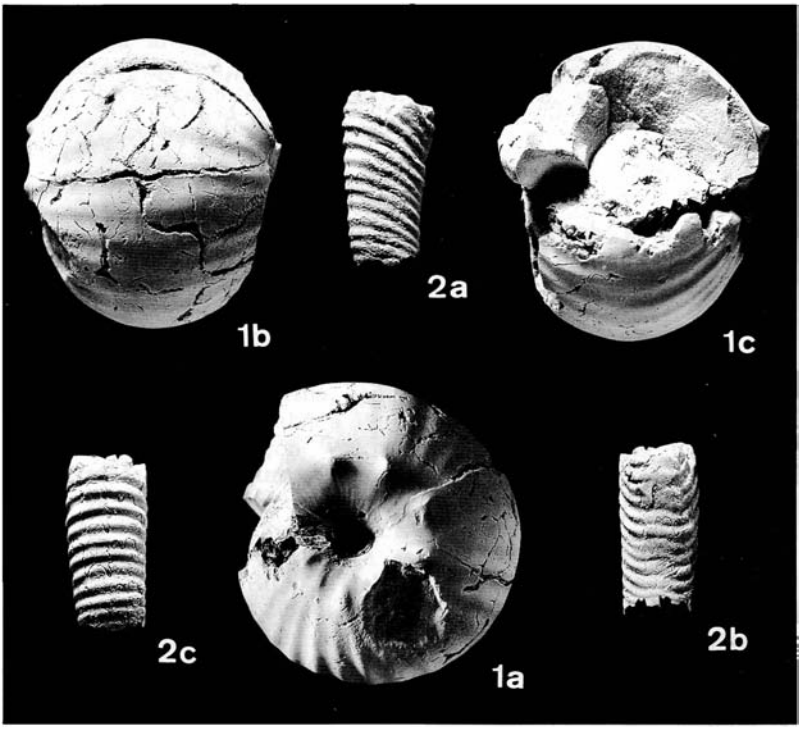

Fig. 6. 1a-c: *Valanginites argentinicus* n. sp. Holotipo, GPIT 1674/2. Vistas lateral, ventral y anterior. Valanginiano inferior. Zona de Neocomites wichmanni. Formación Vaca Muerta. Trahuncurá, Neuquén, Argentina. Tamaño natural.
2a-c: *Protancyloceras* sp. 3/1. Vistas lateral, dorsal y ventral. P.1686 MOZ. Berriasiano superior. Zona de Spiticeras damesi. Formación Vaca Muerta. Huncal, Neuquén, Argentina.

Distribución

V. argentinicus n. sp. procede de la parte superior de la Formación Vaca Muerta en el faldeo occidental del Cerrito de la Ventana, Trahuncurá, Neuquén, en sedimentitas correspondientes a la Zona de N. wichmanni, en asociación con *Lissonia riveroi* (LISSON) y *Acantholissonia gerthi* (WEAVER), cuya edad equivale al Valanginiano inferior.

Suborden	Ancyloceratina WIEDMANN 1966
Superfamilia	Ancylocerataceae GILL 1871
Familia	Bochianitidae SPATH 1922
Subfamilia	Protancyloceratinae BREISTROFFER 1947

Género *Protancyloceras* SPATH 1924

Especie tipo: *Ancyloceras guembeli* OPPEL (in ZITTEL 1868).

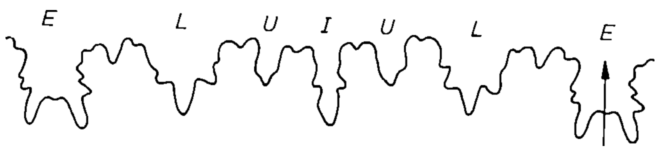

Fig. 7. Línea de sutura de *Protancyloceras* sp. P.1686 MOZ. Altura de vuelta = 3,5 mm, 8/1.

Protancyloceras sp.
Fig. 6. 2a-c; Fig. 7

Material

Un pequeño fragmento de fragmocono muy bien preservado, P.1686 MOZ, Figs. 6. 2 y 7.

Descripción

Pequeño fragmento de fragmocono en el cual se puede apreciar que las vueltas se arrollan según un espiral muy abierto. La sección de las vueltas es suboval, más alta que ancha, algo más ancha en la parte dorsal que en la ventral. La costulación está estrechamente espaciada, curvada y fuertemente inclinada hacia adelante. Al cruzar la región dorsal, las costillas se atenuan y se arquean suavemente hacia adelante. En la región ventral son visibles con dificultad tubérculos ventrolaterales redondeados, cruzando luego las costillas el vientre algo atenuadas según un chevron dirigido hacia adelante. Línea de sutura típica de *Protancyloceras,* según la fórmula E L U I (ver. Fig. 7).

D i m e n s i o n e s (en mm):

Ejemplar n°	H	W	H/W
P.1686 MOZ	4	3	1,33

Observaciones

Si bien no existen dudas para incluir este pequeño fragmento en el género *Protancyloceras* SPATH, el mismo no es suficiente para arribar a una clasificación específica. No obstante, el ejemplar neuquino posee afinidades con *P. kurdistanense* SPATH (1950: 121, lám. 9, figs. 1-5) del Tithoniano de Kurdistán, norte de Iraq, y con *P. eximium* ARNOULD-SAGET (1951: 118, lám. 11, fig. 10a-d) del Tithoniano-Berriasiano de Túnez central. *Protancyloceras* sp. constituye la primera cita de ese género en la Cuenca Neuquina-Mendocina.

Distribución

Protancyloceras sp. ocurre en la parte superior de la Formación Vaca Muerta en la región situada al suroeste de Huncal, Neuquén, en sedimentitas de la Zona de Spiticeras damesi, cuya edad equivale al Berriasiano superior. Está asociado con *Th. huncalense* n. sp. y *K. primaeva* n. sp.

4. Analisis de la fauna

Combinando los resultados obtenidos en este trabajo con la información brindada por H. LEANZA (1972, 1973), LE HEGARAT (1973) y WIEDMANN (1980), puede establecerse la correlación de Tabla 1.

La fauna estudiada comprende los géneros *Thurmanniceras, Kilianella, Berriasella, Valanginites* y *Protancyloceras* en los cuales se han reconocido 3 nuevas especies. *Valanginites argentinicus* n. sp. corresponde a la primera cita de ese género en la Argentina, pues anteriormente su distribución conocida en el dominio andino llegaba a Perú (cf. RAWSON 1981). La especie argentina posee afinidades con *V. psaephoides paludensis* THIEULOY (1977), pero mientras la primera especie se ubica en la parte baja del Valanginiano inferior, la especie francesa corresponde a la parte alta del mismo piso (= Zona de Th. campylotoxum) (cf. THIEULOY, op. cit.). En Argentina *V. argentinicus* n. sp. se encuentra asociado con *L. riveroi* (LISSON) y *Acantholissonia gerthi* (WEAVER) en la Zona de Neocomites wichmanni. Por su parte, *Thurmanniceras huncalense* n. sp. y *Kilianella primaeva* n. sp. consisten en pequeños fragmoconos piritizados con sorprendentes afinidades con aquellos descriptos por SAYN (1907) en el sud-este de Francia. El pequeño fragmento de *Protancyloceras* sp., tambien piritizado, constituye la primera cita de ese género en el oeste central de la Argentina (Cuenca Neuquina/Mendocina). Finalmente, la presencia de un fragmento que indudablemente pertenece a *Berriasella callisto* (D'ORB.) en la Zona de Spiti-

Tabla 1. Esquema de correlación.

	Sud-este de Francia (LE HEGARAT 1973, WIEDMANN 1980)		Argentina (H.LEANZA 1981)	Especies de ammonites citadas entre Huncal y Trahuncurá, Neuquén, Argentina (H.LEANZA 1972, 1973 & este trabajo)
	ZONA	SUBZONA	ZONA (pars)	
VALANGINIANO	Kilianella roubaudiana	Thurmanniceras thurmanni	Neocomites wichmanni	*Lissonia riveroi* (LISSON). *Acantholissonia gerthi* (WEAVER). *Valanginites argentinicus* n.sp.
BERRIASIANO	Fauriella boissieri	Berriasella callisto	Spiticeras damesi	*Berriasella callisto* (D'ORB.), *Thurmanniceras huncalense* n. sp., *Th.* aff. *keideli* GERTH, *Th. neogaeus* A.LEANZA, *Kilianella primaeva* n.sp., *Protancyloceras* sp.

ceras damesi, en la cual están contenidos *Th. huncalense, K. primaeva* y *Protancyloceras* sp., asegura la correlación con la Subzona de B. callisto - correspondiente a la Zona de Fauriella boissieri - del Berriasiano más alto del sud-este de Francia.

Paleogeográficamente los ammonites descriptos son una prueba más de una imigración instantánea de la fauna mediterránea Eurafricana (Mediterranean Leptoceras Province in WIEDMANN 1988) en las cuencas de los Andes y de Neuquén.

Cabe consignar, además, que durante abril de 1985 los autores tuvieron oportunidad de recorrer nuevamente la región de Huncal y Trahuncurá pudiendo coleccionar fragmentos que se esperan estudiar en el futuro pertenecientes a los grupos de *Th. thurmanni* y *F. boissieri,* los cuales confirmarían una vez más la correlación de parte de las capas allí presentes con los estrâtos limîtrofes Berriasiano/Valanginianos del sud-este de Francia.

Agradecimientos. Los autores están profundamente agradecidos al señor JOSE I. GARATE ZUBILLAGA (Zapala, Neuquén) por facilitarnos para su estudio parte del material descripto. El Dr. OSCAR LOSADA (Director General de Minería, Zapala, Neuquén) proveyó un vehículo para una breve visita realizada en Abril de 1985. La Fundación Alexander von Humboldt facilitó al primer autor permanecer unos días en Tübingen (Alemania Occidental) donde se finalizó el estudio. El señor WERNER WETZEL (Tübingen) tuvo a su cargo la obtención de las fotografías. A las personas e instituciones nombradas expresamos nuestro más sincero agradecimiento.

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Biostratigraphy of the Non-Marine Cretaceous of Argentina Based on Calcareous Microfossils

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With 5 Plates, 8 Text-Figures and 3 Tables

MUSACCHIO, E. (1989): Biostratigraphy of the Non-Marine Cretaceous of Argentina Based on Calcareous Microfossils. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 811-850. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The present paper summarizes the available micropalaeontological information from three of the principal oil bearing basins in Argentina: the San Jorge Gulf, the Neuquén and the North-western Argentine basins. The Neuquén basin, in particular, includes different marine sequences with ammonoids giving a complementary framework allowing the gauging of the interbedded or related continental facies.

The stratigraphical distribution of 77 ostracod species and 36 charophyte species, previously described and illustrated, is shown. Seven formal zones are proposed (from which only five are Cretaceous) and their chronological assignation is discussed:

7) Zone of Peckichara cf. varians meridionalis (Upper Paleocene)
6) Zone of Tolypella grambasti (Maastrichtian)
5) Zone of Ilyocypris wichmanni (Campanian - ?early Maastrichtian)
4) Zone of Flabellochara harrisi (Aptian)
3) Zone of Atopochara trivolvis triquetra (late Hauterivian - early Barremian)
2) Zone of "Gomphocythere" dorsoacuminata (early Hauterivian)
1) Zone of Bisulcocypris barrancalensis (Callovian - Oxfordian?)

Emphasis is laid on some cases of taxa showing a large geographical distribution, as well as on the relative degree of endemism as it is suggested in different assemblages. There is evidence that ostracods and charophytes interchanged between Northern and Southern Hemispheres, at least during the Aptian and the Campanian/Maastrichtian ages. Marked differences exist in the structure of species from both groups.

Resumen: Este trabajo resume la información micropaleontológica disponible en tres de las principales cuencas petrolíferas argentinas: la Cuenca del Golfo de San Jorge, la del Neuquén y la del Noroeste argentino. La Cuenca del Neuquén, en particular, incluye diversas secuencias sedimentarias marinas, bien datadas con amonites, las que permiten calibrar las facies continentales interestratificadas.

En el mismo se ilustra la distribución cronológica de 77 especies de ostrácodos y de 36 de carófitos, previamente descriptas. Se proponen siete zonas formales (de las cuales solamente cinco son cretácicas) y se analiza su asignación cronológica. Estas son:

7) Zona de Peckichara cf. varians meridionalis (Paleoceno).
6) Zona de Tolypella gramabasti (Maastrichtiano).
5) Zona de Ilyocypris wichmanni (Campaniano - ?Maastrichtiano temprano).
4) Zona de Flabellochara harrisi (Aptiano).
3) Zona de Atopochara trivolvis triquetra (Hauteriviano tardío - Barremiano temprano).
2) Zona de "Gomphocythere" dorsoacuminata (Hauteriviano temprano).
1) Zona de Bisulcocypris barrancalensis (Calloviano - Oxfordiano?).

Se destacan algunos casos de taxa que muestran amplia distribución geográfica, como así también se analiza el relativo grado de endemismo que exhiben las diferentes asociaciones. Hay evidencias de intercambio de carófitos y ostrácodos con el Hemisferio Norte, al menos durante el Aptiano y el Campaniano/Maastrichtiano.

Se destacan, finalmente, diferencias en la estructura del diseño de las especies en los dos grupos microfósiles.

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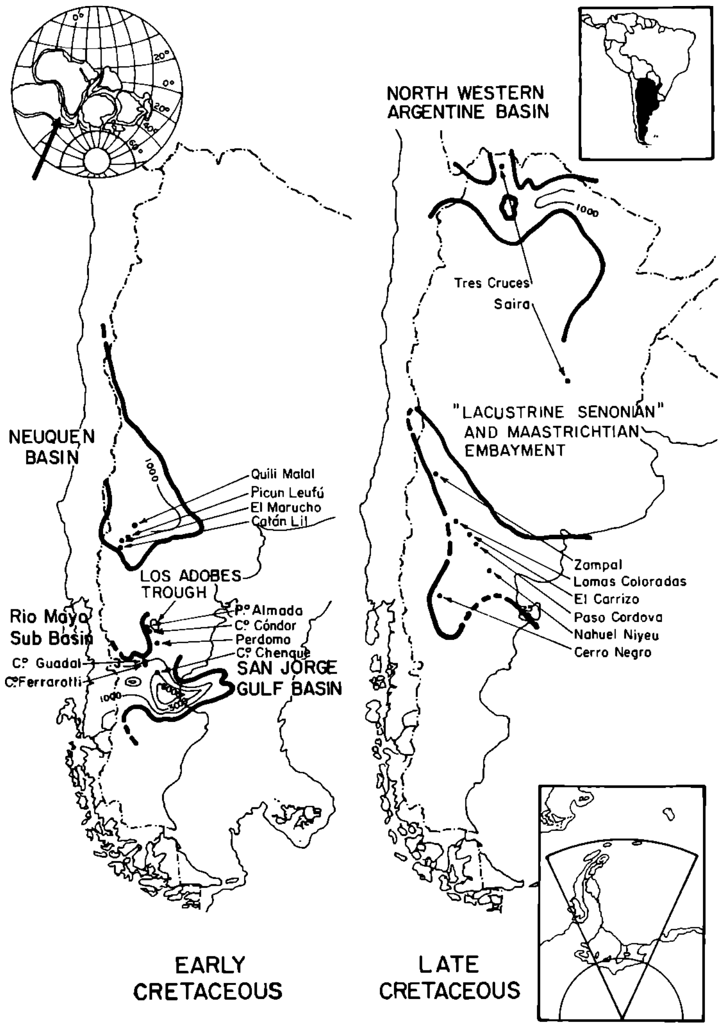
1. Introduction

Cretaceous non-marine ostracod and charophyte remains of Argentina have only received attention in the past 20 years. Now it is possible to design a first outline dealing with the systematics, the stratigraphical distribution and some biogeographical evidences of the known Cretaceous assemblages from the southernmost part of South America, principally from Patagonia. The early Cretaceous assemblages are of the "Wealden-facies" type, including abundant carapaces of the genus *Cypridea* BOSQUET (Ostracoda). Most of the recognized taxa of this genus are provincial or endemic, but a few cases of wide-distributed species were noted. The allied charophyte remains are also abundant, including armored gyrogonites of species with world-wide distribution belonging to the family Clavatoraceae. In the late Cretaceous, however, the ostracods of the genus *Ilyocypris* BRADY & ROBERTSON and other "*Ilyocypris*-like" ostracod-carapaces are dominant, but utricles of clavatoraceans have not been found up to now.

In addition, the present paper contributes to the first attempt towards a zonation based on species of ostracods and charophytes recognized only in outcrops. Many of the proposed zones have proved their capability of comparing different profiles of the same region or between different basins. For the chronological approach, the Neuquén Basin is the most solid reference. In this basin many ammonite-bearing levels allow to gauge the interbedded continental deposits or, at least, give reliable top limits for the beginning or the end of some non-marine sequences. On the basis of the best dated assemblages it was preferred to use the world-wide geologic time-scales rather than other provincial chronological terms. In other cases, for instance in the San Jorge Gulf Basin, the principal support for the chronological approach is the correlation with similar Neuquén assemblages, as well as the proper information displayed by the clavatoraceans. However, the age of some assemblages can be, up to the present, only roughly assigned. In spite of the present advances, some basins, many areas and complete profiles have not been examined for microfossils.

2. Geological Framework



Text-Fig. 1. Sedimentary basins and most important localities in Argentina where Cretaceous non-marine microfossils are found.

Coinciding with the time of rifting and initial Atlantic spreading, an intra-Malmean orogeny of Kimmeridgian age took place in different regions and basins of South America and was followed by a late Jurassic/early Cretaceous sedimentary megacycle. The early Cretaceous sedimentary environments in Argentina are both well diversified and well distributed. Many good exposures show stratigraphical continuity, without a break at the Jurassic/Cretaceous boundary. Some of the most important oil-bearing fields of the country are found in sediments of the mentioned megacycle. More advanced in the early Cretaceous epoch, the regional identity of the same megacycle decreased owing to the local or provincial character of different epeirogenic movements.

A second sedimentary megacycle mainly belonging to the late Cretaceous cannot be, likewise, well defined in its origin at different basins. However, at the end of the megacycle wide areas of the country were conspicuously transgressed during the Maastrichtian up to the Danian by a marine event giving rise to an Atlantic connection. The second megacycle ends with this event which took place prior to the Paleocene "Laramide" phase of the Andean orogeny.

2.1 The San Jorge Gulf Basin

This cratonic basin, filled mainly by continental sediments, is settled on a Jurassic substratum mostly composed of volcanic rocks that overflowed at different times after the Gondwanid orogeny. A precocious Liassic-Dogger continental rifting of tensional origin, related to a presumably aborted essay of Gondwanaland fragmentation, built an initial basin pattern which was later modified. The previous Jurassic volcanic-sedimentary megacycle can be best analysed at the outcrops comprising the northern arm of the primitive basin at the Medium Valley of the Chubut River. Recurrent flows, mainly andesitic, filled a trough with local thickness of more than 2.500 m, including Liassic marine interbedded deposits and Callovian-Oxfordian? lacustrine sediments with ostracods.

The late Jurassic and Neocomian sedimentation that followed filled a West-East directed basin ruled in part by transform faults related to the Atlantic spreading. The involved sediments are mainly epiclastics, lacustrine and fluvial in origin. The western arm (Paso Rio Mayo Subbasin) communicated westward with the Pacific Andean domain where marine early Cretaceous sediments of the back-arc type seem to be interbedded.

During Aptian age, a volcanic-pyroclastic event (Divisadero Group) at the Andean domain spread out widely to the extra-Andean region (involving the Pozo D-129 and Cerro Barcino Formations with non-marine microfossils). This event is, in turn, overlain by different and local late Cretaceous sequences, mainly pyroclastics and volcanic in the origin of the clastic materials, having local and regional correlations not yet fully established.

At the end of the period, isolated Maastrichtian marine sediments, and other Danian more widely distributed ones preceded the first Tertiary Laramide orogeny.

Authors contributing to these problems were STIPANICIC & RODRIGO 1970, NULLO & PROSERPIO 1975, RAMOS 1979, BIANCHI 1981, LESTA et al. 1980, MUSACCHIO 1981, SCIUTTO 1981, BARCAT et al. 1984.

2.2 The Neuquén Basin

An elongated embayment from the Jurassic/early Cretaceous, located subparallel to the west continental margin of the central regions of Argentina and Chile, has been the ruling factor for the Neuquén Basin. At least in part, the basin is of the back-arc type. However, its longitudinal axis point curved to the south-east, where the basin subsided on a cratonic and mobile substratum. This "basement" is not settled at the continental margin properly, but on a belt contiguous to the Colorado Trough, lying to the east.

In this basin, GROEBER (1946) proposed three sedimentary megacycles, classical for the Andean domain: the "Jurassic", the "Andico" and the "Riográndico", which deserve general assent today, only with minor modifications.

The "Andico" follows the intra-Malmean orogeny of Kimmeridgian age. These movements have displayed a continental synorogenic blanket on the basin ("areniscas de relleno") interbedded toward the west in Chile with the andesitic deposits ("ocoitas") of the magmatic arc. The greater subcycle "Mendociano", which is mainly marine, was settled over this substratum from the Tithonian to the early Barremian. The Jurassic/Cretaceous boundary is stratigraphically transitional. This subcycle was affected by epeirogenic block-tilting which together with eustatic fluctuations explains the peculiar facies changes of the basin. Many biostratigraphical zones of ammonoids (classical on the literature) allow the gauging of the allied Eocretaceous interbeddings. The second subcycle belonging to the Andico ("Rayosiano" = Rayoso Group; emend. by ULIANA et al. 1975) follows a main epeirogenic phase belonging to the initial Miranic movements during the Upper Barremian/Aptian. The marine communication remains, but very attenuated during the Aptian; the facies are, however, mostly continental including also thick "red bed" sections.

The "Riográndico" megacycle begins with a continental environment ("Neuqueniano" = "Dinosaurian beds"), but at the end of the Cretaceous and up to the early Paleocene, however, a marine event ("Malalhueyano") connected with the Atlantic embayment of the Colorado Trough, ends conspicuously the "Riográndico". This marine event provided an excellent fossil register for chronological references (see under the Grambasti Zone).

Authors contributing to these problems were: WEAVER 1931, ULIANA et al. 1975, LEANZA 1980, DIGREGORIO & ULIANA 1980, ULIANA & DELLAPE 1981, LEANZA & WIEDMANN 1980, LEANZA 1981, ORCHUELA & PLOSZKIEWICZ 1984, DIGREGORIO et al. 1984.

2.3 The North-West Argentine Basin

A well-spread brackish basin having a variable degree of marine influence is displayed in the north of Argentina, north of Chile and Bolivia. The representative Yacoraite Formation in Argentina, including carbonatic rocks with stromatolites and pelites belongs to the Maastrichtian up to the Paleocene. It overlies a continental late Cretaceous sequence, including red beds (Grupo Pirgua) through a transition area which anticipates the marine communication (Lecho Formation). Fossils and sedimentary rocks of the Yacoraite Formation suggest a surrounding low landscape, stagnant in part, up to anoxic conditions in the basin itself. Even though the biota is not scarce, the major ecosystem is not taxonomically well diversified, including continental and other fossils of mixed-environment with marine contact (FRENGUELLI 1936, GROEBER 1953, LOHMANN & BRANISA 1962, LEANZA 1969, YRIGOYEN 1969, MARQUILLAS 1986, SALFITY & MARQUILLAS 1986, ZAMBRANO 1987).

3. Biostratigraphical Zones

In different late Jurassic to early Tertiary non-marine assemblages from Argentina more than 77 taxa of Ostracoda and more than 36 of Charophyta were recognized. The minimum chronological and stratigraphical distribution of the species is shown in the Tables 1 and 2.

In each one of the locally sampled profiles, the vertical distribution of the different species was examined. Some groups of species, or groups of very similar morphs, which are also found in different places and even in other basins are now selected for correlations as secondary markers. The zonation now proposed attempts to emphasize the chronological interest of these groups of taxa but with independence of their possible local stratigraphical significance. In this sense the units attempt to be oppelizones, but with limitations. In some cases, the first limitation is the paucity of the number of species characterizing the zones.

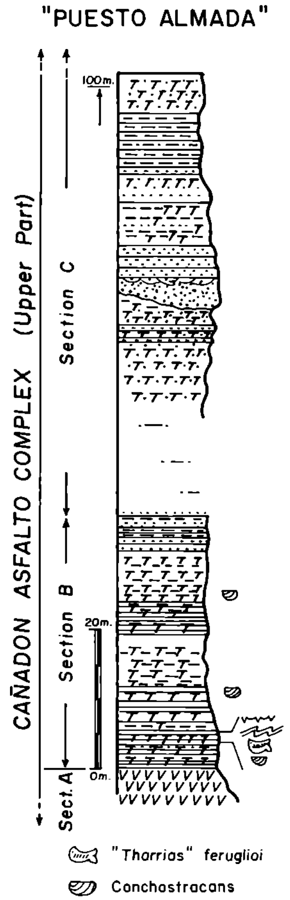
A particular case is that of the Zone of Ilyocypris wichmanni. The first marker of the zone is an ostracod that shows its acme at the lacustrine event which ends the continental Neuquén Group, previous to the Maastrichtian marine transgression.

3.1 Zone of Bisulcocypris barrancalensis (Callovian - Oxfordian?)

The continental Cañadón Asfalto Complex (STIPANICIC et al. 1968) is well developed at the Cerro Cóndor region, in the Chubut Province. It is composed of black shales, cherty limestones, mudstones, basic volcanics (besides other allied Cretaceous hypabyssal traps) and, finally, of an uppermost member (or formation?) which includes the present ostracods, together with conchostracan and fish remains. This member is characterized at its lower part by gray and olive-grayish shales, marls and mudstones, occasionally including a basal violet-purple conglomerate as in the Cerro Cóndor itself, and at its upper part by tufaceous sandy sediments, gray to grayish-pink. A complete and relatively undisturbed profile, having a thickness of more than 400 m outcrops approximately 15 km NE of the Cerro Cóndor at the Cerro Bayo, where it underlies unconformably - owing to the intra-Malmean movements - the Neocomian Los Adobes Formation (STIPANICIC et al. 1968).

For additional information as regards the stratigraphy and the paleontology see: STIPANICIC & RODRIGO 1970, TASH & VOLKHEIMER 1970, NULLO & PROSERPIO 1975, LESTA et al. 1980, BONAPARTE 1981, CIONE & PEREIRA 1987, VALLATI, in press, including the listed literature.

The reference profile for the present zone can be better studied 2 km north of the Almada Outpost, immediately west of the main Provincial Road 12, 18 km north of the Cerro Cóndor hamlet. These outcrops exhibit only a part of the uppermost and unnamed Member of the Canadón Asfalto Complex (sections B and C of Text-Fig. 2). The microfossils were recovered from grayish pelites, which in part include laminated siltstones with the



Text-Fig. 2. Reference profile for the Barrancalensis Zone (from MUSACCHIO et al., in press).

"Tharrias" feruglioi assemblage of fishes, recently reviewed by CIONE & PEREIRA (1987).

The present ostracods (see MUSACCHIO et al., in press) can be better compared to others from the Middle Jurassic and, in particular, from the Upper Jurassic in the Province of Gansu in China (see LI ZU-WANG 1985, and QI HUA 1985). However, the age of the assemblage has been estimated mainly taking into account the stratigraphical relationships rather than the information arising from the microfossils properly.

Bisulcocypris barrancalensis MUSACCHIO s. str. and *Darwinula* sp. (in MUSACCHIO & CHEBLI 1975) were also found at "El Barrancal" (next to Paso de Indios in the Province of Chubut), in sediments that we now consider equivalent to the present reference profile (cf. MUSACCHIO & CHEBLI 1975: 126-128).

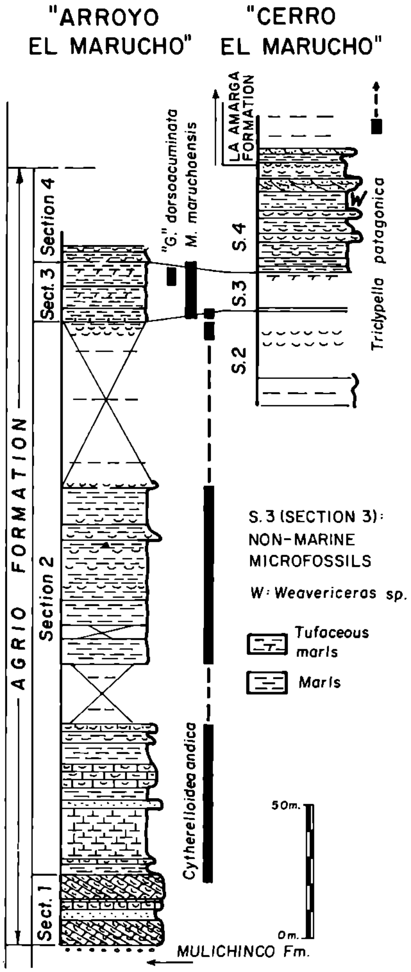
Darwinula sp. (in MUSACCHIO & CHEBLI 1975) is proposed as a secondary marker.

Reference for the assemblage: MUSACCHIO et al. (in press).

3.2 Zone of "Gomphocythere" dorsoacuminata (early Hauterivian)

The reference profile for this zone lies in the Province of Neuquên (Neuquén Basin) approximately 80 km south of the town of Zapala, and next to the National Road N° 40 at El Marucho. Good exposures of the lower part of the Agrio Formation (WEAVER 1931) can be studied here at El Marucho brook; however, the section having the *dorsoacuminata* assemblage is placed at the middle-upper part of the Agrio Formation, outcropping on both sides of the mentioned Road 40, on the northern rise of the El Marucho.

The Agrio Formation forms part of the "Mendociano" sedimentary cycle (Tithonian - early Barremian), mainly of marine origin and belonging to the Andico megacycle. The present profile, however, next to the southern border of the "embayment" (see above: "Neuquén Basin") includes non-marine intertongues. The ostracods and charophytes were found in a section which is approximately 20 m thick, composed of gray and grayish-violet



Text-Fig. 3. Reference profile for the Dorsoacuminata Zone.

marls, siltstones, and very fine sandstones, in part tuffaceous. Macro- and microfossils suggest a brackish or mixed environment for this section. Besides, some samples include foraminifers and ostracods with marine influence which are also well represented in the underlying marls of the section 2 (see Text-Fig. 3), such as the *Cytherelloidea andica* MUSACCHIO (1978), a conspicuous species from the lower Agrio Formation. The stratigraphical relationships relative to the under- and overlying sediments of the non-marine member (or section 3) seem to be concordant.

The underlying marine section 2 includes the ammonoids *Lyticoceras pseudoregale* BURKHARDT, *Acanthodiscus* ex gr. *vaceki* NEUMAYER, *Holcoptychites* aff. *compressus* LEANZA & WIEDMANN belonging to the early Hauterivian Pseudoregale Zone of GERTH (1925).

The overlying marine section 4 (see Text-Fig. 3) includes the ammonoid *Weavericeras* sp. belonging to the early Hauterivian Neuquensis Zone of GERTH (1925) (see below, discussion on Trivolis triquetra Zone).

This is a reliable basis to assign the Dorsoacuminata Zone of non-marine microfossils to an early Hauterivian age.

The provisional generic assignation of *"Gomphocythere" dorsoacuminata* MUSACCHIO (1970) awaits for a clear phyletic relationships approach of this and other Cretaceous new Limnocytheridae from Argentina.

Mesochara maruchoensis SIMEONI & MUSACCHIO (1986) is considered a secondary marker.

Reference for the assemblage: SIMEONI & MUSACCHIO (1986).

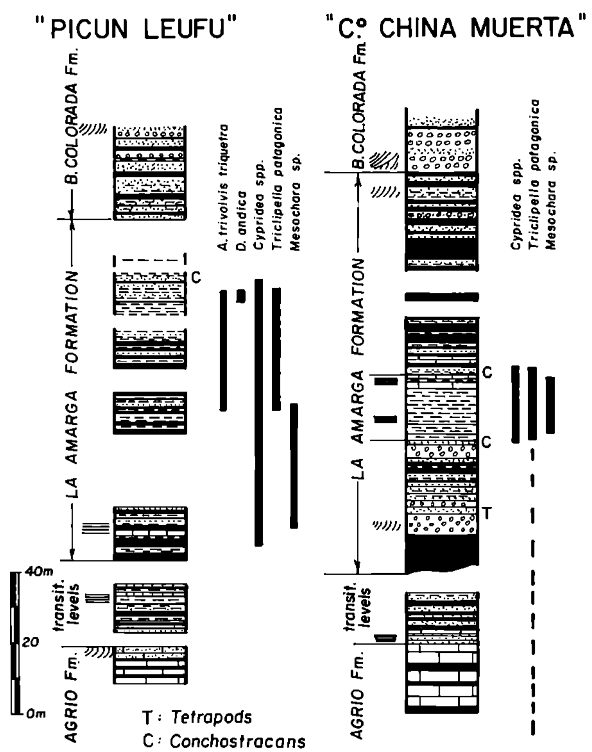
3.3 Zone of Atopochara trivolvis triquetra (late Hauterivian – early Barremian)

The reference profile for this zone belongs to La Amarga Formation (PARKER 1964) at the Picún Leufú River in the Province of Neuquén, south of the town of Zapala, 25 km downstream from the river-crossing of the National Road 40. In this place, the La Amarga Formation is mainly composed of lacustrine gray and grayish-yellow marls including scarce limestones, variegated mudstones and siltstones. It overlies transitionally the Agrio Formation.

The type-locality of the La Amarga Formation outcrops, instead, 6 km to the north-east of the Aguada Florencio (the last one at the National Road 40, south of the Picún Leufú River). It overlies, also transitionally, marine sediments of the Agrio Formation. The lithology shows, however, differences with the correlated column of Picún Leufú. At the lower part of the Cerro China Muerta profile, La Amarga Formation includes continental sandstones and conglomerates, and upwards many limestone levels, also of continental origin. The last ones replace fertile levels of marls and end, in turn, the lacustrine section, which is more reduced in this profile. The microfossils in this area are filled with reddish zeolites.

In both places these sediments are, in turn, transitionally overlain by an epiclastic sequence, including conglomerates, from red to reddish-purple.

Besides the first marker of the zone *Atopochara trivolvis triquetra* GRAMBAST (1965), the charophytes *Triclypella patagonica* nov. sp. and *Mesochara* cf. *stipitata* (WANG, S. 1965) seem to be good secondary markers for correlations. The last species was originally described as *Mesochara* sp. (in MUSACCHIO 1971a), but recently SCHUDACK (1987: 153) rightly emphasized similitudes with the reference-taxon from China. Dealing with



Text-Fig. 4. Reference profile for the Triquetra Zone at Picún Leufú, Neuquén (MUSACCHIO 1971a).

ostracods, good markers also seem to be the species *Cypridea (U.) australis, Cypridea (U.) subcuadrata, Cypridea (U.) modestissima, Huillicythere gambasti,* and *Dryelba picunleufensis.*

MUSACCHIO (1971a: 35) assigned the La Amarga Formation assemblage to the Barremian, taking into account that the available register in 1971 for the *Atopochara trivolvis triquetra* ssp. and the genus *Triclypella* was limited to that age. The same author did not reject, however, a possible Hauterivian age for the same assemblage owing to local stratigraphical relationships. Here, the non-marine levels belonging to the La Amarga Formation are intertongued between marine sediments of the uppermost Agrio Formation at El Marucho/China Muerta (ibid. 1971a: 36, foot-note). These continental replacements increase from the Aguada Florencio to Catán Lil, toward the proximal border of the basin, where many red bed sections are intercalated between marine Hauterivian yellow limestones and green marls.

After MUSACCHIO (1971a), both *A. trivolvis triquetra* ssp. and the genus *Triclypella* were registered in Hauterivian sediments (CANEROT 1979, 1982 in SCHUDACK 1987: 136, WANG ZHEN & LU HUI-NAN 1982). On the other hand, the ammonoids from the El Marucho are now better known thanks to the identifications of Dr. H. LEANZA (CONICET - Argentina) on our collections (see: Dorsoacuminata Zone, discussion in this paper). Dr. LEANZA admits a replacement of the late Hauterivian Crioceratites andinus Zone of ammonoids, absent at El Marucho/China Muerta area, by the non-marine La Amarga Formation.

Summarizing: The levels of the La Amarga Formation with *Triclypella patagonica* nov. sp. overlying marine limestones with *Weavericeras* sp. of the Neuquensis Zone (Text-Fig. 4, at Cerro El Marucho) are, in the present paper, assigned to a late Hauterivian age. There is a good basis to enlarge the age of the present Triquetra Zone, at Picún Leufú, from late Hauterivian to Barremian. In fact, the possible Barremian age for the upper part of the same interval in this place cannot be rejected. The levels bearing *A. trivolvis triquetra* at Picún Leufú are younger than those being included or intertongued at the uppermost part of the Agrio Formation at El Marucho/Aguada Florencio. At Picún Leufú, the Agrio Formation lacks continental interbeddings or replacements such as those more frequently exhibited toward the basin border at the south. The time when the regression of the marine environment reached the Picún Leufú locality is, up to now, unknown.

Reference for the assemblage: MUSACCHIO (1970, 1971a and 1971b).

3.3.1 The *Cypridea (Ulwellia) modestissima* assemblage

In the Province of Chubut at Cerro Guadal and next to the V. Parada outpost, the Albornoz Formation (FERNANDEZ GARRASINO 1977) includes different lacustrine pelitic levels with non-marine ostracods, recently described (MUSACCHIO et al., in press) and characterized by many species of the subgenus *Cypridea (Ulwellia)* ANDERSON 1939. More than a half of the taxa keep similitudes of different degrees with the ostracodal microfauna of the La Amarga association at China Muerta. The age of this assemblage is considered Hauterivian to Barremian.

Reference for the assemblage: MUSACCHIO et al. (in press).

3.3.2 The *Looneyellopsis* cf. *brasiliensis* assemblage

In the Province of Chubut, at the Cerro Ferrarotti profile (few km south of the above mentioned V. Parada profile) the lowermost sediments of the Cretaceous column, just overlying Jurassic green shales with conchostracan, include a well preserved association of ostracods and charophytes. Among the ostracods the presence of *L.* cf. *brasiliensis* KRÖMMELBEIN & WEBER (1971) is emphasized. The reference species was described from the Candeias Unit at the Tucano-Recôncavo basins, Brazil.

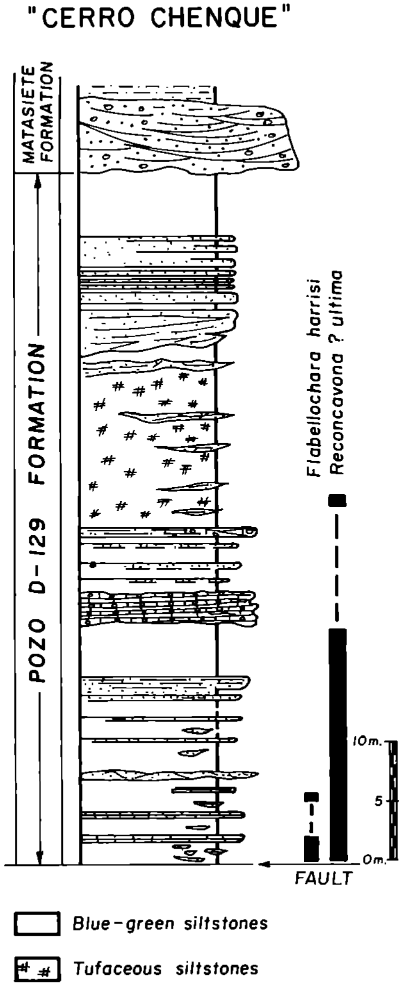
This assemblage certainly belongs to the Lower Cretaceous on account of the composition of the microfossils. The same precedes the *modestissima* assemblage according to its stratigraphical position. However, the relationships between these two geographically neighboring assemblages is not known with exactness. Up to now it can be roughly assigned to the Neocomian.

Reference for the assemblage: MUSACCHIO et al. (in press).

3.4 Zone of Flabellochara harrisi (Aptian)

The Pozo D-129 Formation (LESTA 1968) from the San Jorge Gulf Basin is well known in the subsoil where it can exceed 2.500 m of thickness. The relatively short uppermost section of this formation outcropping at Cerro Chenque, 80 km north of the town of Sarmiento, in the center of the Province of Chubut, seems to be, however, a significant biostratigraphical reference for the basin. These outcrops include a well preserved microfossil assemblage as well as good exposures allowing the study of the upper and lateral stratigraphical relationships of the formation (HECHEM et al. 1987).

The green tufas, in part lacustrine, of the Cerro Barcino Formation (CHEBLI in MUSACCHIO & CHEBLI 1975), which are well and continuously



Text-Fig. 5. Reference profile for the Harrisi Zone (from HECHEM et al. 1987).

exposed at the middle valley of the Chubut River area includes, likewise, a well preserved *F. harrisi* assemblage. Both the uppermost Pozo D-129 and the Cerro Barcino assemblages are considered coeval in this paper. Some complete profiles of the Cerro Barcino Formation which include many fertile levels, could be chosen as type-reference for the zone, instead of the partially outcropping Cerro Chenque profile. The regional significance of the Cerro Barcino beds is, however, much more limited in comparison, owing to the restricted or local distribution of this formation.

The age of the *F. harrisi* assemblage of the Cerro Barcino was already considered Aptian or next to this age, taking into account the presence of five species of ostracods and charophytes very similar or identical to other morphs found in Aptian sediments in USA (see MUSACCHIO in MUSACCHIO & CHEBLI 1975). The recently recovered Cerro Chenque assemblage includes also *F. harrisi* (PECK) and *Porochara mundula* (PECK) and, besides, two species of ostracods which can be well compared to other Brazilian taxa. They are: *Reconcavona? ultima* KRÖMMELBEIN & WEBER 1971 from Candeias beds at the Recôncavo Basin and *Pattersoncypris* cf. *angulata angulata* (KRÖMMELBEIN & WEBER 1971) from the Riachuelo Formation at the Sergipe Basin, both of late Eocretaceous age.

Moreover, the last mentioned assemblage includes the first register up till now in Argentina of *Neuquenocypris* MUSACCHIO and *Ilyocypris* BRADY & NORMAN. Other species interesting for correlations are *"Gomphocythere" herreriensis* MUSACCHIO & CHEBLI (1975) and *Rayosoana* sp. (in MUSACCHIO & CHEBLI 1975), although the last one is scarcely represented.

A formerly described *Flabellochara* aff. *harrisi* assemblage in the Ranquiles Formation WEAVER 1931 (Rayoso Group) from the Neuquén Basin at Quili-Malal was assigned to the Aptian or next to this age (MUSACCHIO & PALAMARCZUK 1975). The ostracods and charophytes are in the same levels allied to foraminifers. This fact suggests a remaining of the marine communication of the basin after the Miranic initial tilting movements ending the "Mendociano" sedimentary cycle.

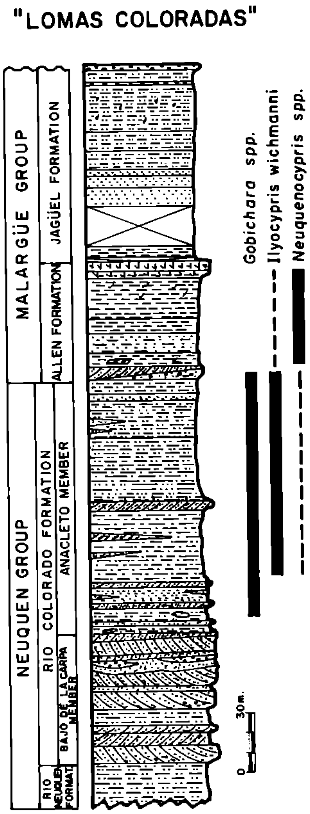
Secondary markers of the zone are: *Cypridea diminuta, Cypridea craigi, Cypridea amerikana, Reconcavona? ultima, Rayosoana* sp. (Ostracoda) and *Porochara mundula* (Charophyta).

References for the assemblage: MUSACCHIO & CHEBLI (1975), HECHEM et al. (1987).

3.5 Zone of Ilyocypris wichmanni (Campanian - ?Lower Maastrichtian)

I. wichmanni MUSACCHIO (1973) is a species of Ostracoda of non-marine? to brackish waters, which can be abundantly recovered from the variegated lacustrine mudstones of the Anacleto Formation (or Member) at the uppermost part of the Neuquén Group (= "Dinosaurian beds"). The same species show a high degree of population variability. This fact is presumably related to the variations of the physical and chemical conditions of the environment. For this reason a broad definition of the species was given.

The reference profile at Lomas Coloradas (see ULIANA 1979) is next to the "Jagüel de Rosauer" (= "Bajo del Jagüel"), east of the town of Neuquén. In this profile it is possible to study minutely the apparition and increasingly abundance of gyrogonites of *Gobichara (Pseudoharrisichara)* spp. and *"Cypridea"* sp. These levels are overlain first by grayish-green, then violet, and finally green mudstones, now displaying abundant carapaces of



Text-Fig. 6. Reference profile for the Wichmanni Zone (from ULIANA 1979).

I. wichmanni. This species is here allied to *Wolburgiopsis neocretacea* (BERTELS) and *Neuquenocypris* sp.

A similar association was recovered from the same variegated mudstones next to Paso Córdoba in the Province of Río Negro (cf. WICHMANN 1927: XIII) and near Vista Alegre, in the Neuquén Province, from the higher Neuquén River cliff next to the Provincial main Road no. 234. Likewise, *I. wichmanni* was found at the Lower Coli-Toro Formation of the Cerro Negro profile (see COIRA 1979: 31) next to the town of Ing. Jacobacci.

At Lomas Coloradas, the Anacleto Formation is overlain by brackish (not normal-marine) sediments with foraminifers and different species of ostracods of the genus *Neuquenocypris* belonging to the Allen Formation. In the Allen Formation at Lomas Coloradas and in the Loncoche Formation at Zampal, Aguada de Pérez and Aguada de Isaac (in the Province of Mendoza) some levels were found with *I. wichmanni*. According to a broad definition of the taxon this is not a peculiar species for the Anacleto Formation and their equivalents, but is is also present in brackish (oligohaline?) uplying sediments (belonging to the Grambasti Zone) of Maastrichtian age. Moreover, some Eotertiary morphs of the El Carrizo Formation keep morphological similitudes with *I. wichmanni*. However, their acme at the Anacleto Formation is considered a good stratigraphical marker. The abundance of this ostracod allows the characterization of the corresponding lacustrine biotope, without any equivalent in other underlying pelitic sections of the Neuquén Group.

Good markers for correlations of the Anacleto Formation and its equivalents are *Gobichara (Pseudoharrisichara) walpurgica* (MUSACCHIO 1973) and *Gobichara (Pseudoharrisichara) tenuis* (MUSACCHIO 1973). These charophytes are in many levels allied to *I. wichmanni*.

The presence of a species belonging to the Talicyprideinae (HOU YOU-TANG 1982), Ostracoda, is emphasized (Plate 5, Figs. 7-10).

References for the assemblage: MUSACCHIO (1973) and MUSACCHIO in ULIANA & MUSACCHIO (1978).

3.6 Zone of Tolypella grambasti (Maastrichtian)

The reference profile for this zone outcrops in the Province of Mendoza, south of Bardas Blancas and next to the National Road No. 40. In this place, the Loncoche Formation is composed of yellow sandstones, green mudstones and marls with gypsum and carbonates (see Text-Fig. 7). These sediments overlie the continental lacustrine deposits of the Anacleto Formation. The Loncoche Formation includes in this profile macro- and microfossils of mixed behaviour, showing a variable degree of restriction relative to open marine environment. Among the foraminifers the presence of *Guembelitria cretacea* (CUSHMAN 1933) and *Pyramidina prolixa* (CUSHMAN & PARKER 1935) is emphasized. The Locoche Formation is correlated, in part, with the normal-marine Jagüel Formation with *Eobaculites ootacodensis* (STOLICZKA) (see RICCARDI 1974) at Río Negro and, in part, with the mixed facies of the Allen Formation in the Provinces of Neuquén and Río Negro (ULIANA 1979, DIGREGORIO & ULIANA 1980, and ULIANA & DELLAPE 1981). The Cretaceous/Tertiary boundary in the present profile cannot be fixed up to now. The uppermost limestones that follow, also with microfossils of brackish origin, were correlated with the Roca Formation of Danian age (see Text-Fig. 7, after ULIANA in ULIANA & MUSACCHIO 1978).

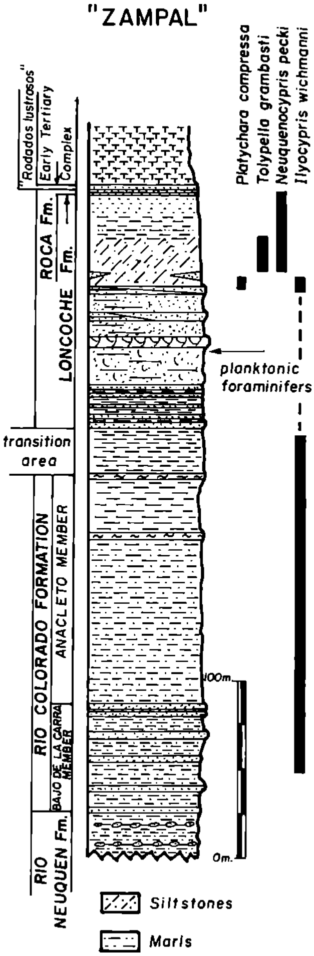
The microfossil assemblage is characterized by different species of the ostracod genus *Neuquenocypris* MUSACCHIO (1983). Some peculiar species of this last taxon were found in sediments of brackish environments, but never in others of normal-marine origin. At present this genus is found in Argentina in continental and mixed facies of Aptian to Paleocene ages.

Some species of *Neuquenocypris* belonging to the Grambasti Zone are allied to the charophytes *Platychara compressa* (KNOWLTON) and *Porochara* cf. *gildemeisteri* KOCH & BLISENBACH in El Zampal.

Other species of non-marine ostracods seem to deserve interest for correlations, although they are scarce and not well preserved in our collections. They are: *Timiriasevia* sp., *Cyprois?* sp. and *Ilyocypris* sp. B (see MUSACCHIO in ULIANA & MUSACCHIO 1978).

Secondary markers for this zone are: *Neuquenocypris pecki, Neuquenocypris alleniensis, Neuquenocypris zampalensis, Neuquenocypris nauel niyeuensis, Timiriasevia* sp., *Cyprois?* sp., *Ilyocypris* sp. B.

Reference for the assemblage: MUSACCHIO in ULIANA & MUSACCHIO (1978).



Text-Fig. 7. Reference profile for the Grambasti Zone (from ULIANA in ULIANA & MUSACCHIO 1978).

3.6.1 The Yacoraite and Mariano Boedo assemblages (Maastrichtian to Danian)

Up to the present, the assemblage of the Loncoche Formation is the most diversified of the terminal Cretaceous. The Grambasti Zone is well represented in a relatively continuous vertical interval of the wide-distributed Maastrichtian basin of the north of Argentina, Chile, Bolivia and equivalent sediments from the south of Perú (see above: "The North-West Argentine Basin"). To this interval belongs part of the assemblage of the Yacoraite Formation (MUSACCHIO 1972), and the Mariano Boedo Formation from the subsoil of Córdoba, Argentina, the El Molino Formation from Bolivia (GRAMBAST 1968) and equivalent sediments of the Moho Group from Perú (PECK & REKER 1947).

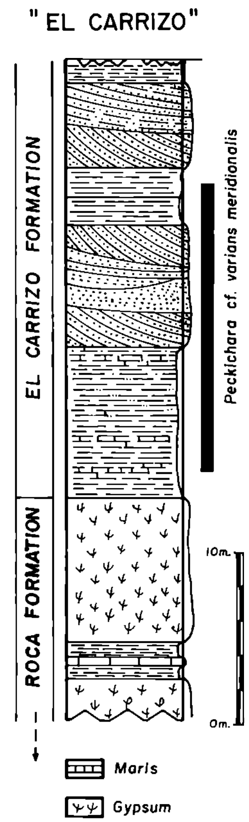
The systematics of these microfossils are relatively monotonous and the presence of the charophytes *"P." compressa* and *"P." gildemeisteri* is characteristic. However, in some profiles charophytes with affinities to others of the Eotertiary age from the north of Perú were found. They are *"Grambastichara" yuntaishaensis* WANG ZHEN (1978) from the Yacoraite Formation and a *Peckichara* sp. from the Mariano Boedo Formation which resembles *P. varians meridionalis* MASSIEUX (1981). These species suggest a Danian age for the uppermost sediments of the two last mentioned units.

3.7 Zone of Peckichara cf. varians meridionalis (Paleocene)

In the Province of Río Negro at "Barda Negra" and "Meseta del Humo" the terminal gypsum of the Roca Formation is overlain by continental fluvial and lacustrine sediments of the El Carrizo Formation.

The underlying marine Roca Formation includes the planktonic foraminifers of Danian age (BERTELS 1970):

Globoconusa daubjergensis (BRÖNNIMANN 1953)
Globorotalia pseudobulloides (PLUMMER 1927)
Subbotinna triloculinoides (PLUMMER 1927)



Text-Fig. 8. Reference profile for the "cf. varians meridionalis" Zone (from ULIANA 1979).

The microfossils of the El Carrizo Formation are abundant and well preserved but limited in the number of species. Some species show similitudes with others of the Grambasti Zone. Such is the case of *Platychara compressa* and morphotypes of *Ilyocypris* sp. (see MUSACCHIO & MORONI 1983). The genus *Peckichara* GRAMBAST itself is known in the late Cretaceous and a species of this genus was found in Maastrichtian levels of the Coli Toro Formation, but is different from the present first marker. Besides, the genus *Limnocythere* of ostracods was recorded in the late Cretaceous. According to these evidences, the present assemblage, which is up till now the only well-dated from the Argentine Eotertiary, does not show a rough taxonomical change but an attenuation of the Maastrichtian diversity.

The reference-section of this zone is just the type-profile proposed by ULIANA (1979) for the El Carrizo Formation.

Limnocythere sp. is considered a secondary marker for the zone.

Reference for the microfossils: MUSACCHIO & MORONI 1983.

4. Palaeobiological Problems

This is the right moment to point out two palaeobiological problems arising from the systematics of the microfossils under consideration. The first one deals with the biogeographical relationships. The second deals with the nature of the species structure of the two groups of the non-marine microfossils discussed.

An important part of the Cretaceous taxa seems to be endemic both in the species and, in cases, in the genera levels. A few other species are common or very similar to other species described from some Brazilian and West African continental margin basins. The best examples of these cases are *Looneyellopsis* cf. *brasiliensis* KRÖMMELBEIN & WEBER (1971), from the Neocomian levels of the Ferrarotti profile, and two ostracods of Aptian age from Pozo D-129 Formation at Cerro Chenque profile included in the taxa *Reconcavona? ultima* KRÖMMELBEIN & WEBER (1971) and *Pattersonocypris* cf. *angulata angulata* (KRÖMMELBEIN & WEBER 1971). These species were found in the San Jorge Gulf Basin. However, in average, the Brazilian similitudes are up to now scarce.

Now it is interesting to focus on the cases of possible interchange of ostracods and charophytes with the Northern Hemisphere:

The scarcely diversified Middle/Upper Jurassic assemblage from the Cañadón Asfalto Complex is the only one found in Argentina from this period. It strongly resembles, at generic level, other coeval assemblages from the Northern Hemisphere.

Species belonging to the Neocomian assemblages were found which resemble other taxa with world-wide distribution. Such are the cases of the charophytes *Atopochara trivolvis triquetra* GRAMBAST and *Triclypella patagonica* nov. sp. Among the allied ostracods the most conspicuous "Wealden-facies" genus *Cypridea* is well represented, but no species having "normal" carapace as in *Cypridea* s. str. was found. All of the recognized taxa and morphotypes show the right valve overlapping the left valve as in *Cypridea (Ulwellia)* ANDERSON. On the other hand, several cases of faunal and floral similitudes with the record of the Northern Hemisphere suggest the possibility of interchange. The most significant evidences of this interchange with the Northern Hemisphere are known in the advanced early Cretaceous and in the terminal late Cretaceous.

The Flabellochara harrisi Zone, of presumably Aptian age and represented by different post-orogenic sequences both in the Neuquén as in the San Jorge Gulf Basins, includes species having different degree of similitude to a counterpart in the Northern Hemisphere, particularly in the USA, belonging to the Comanchean sedimentary cycle and other equivalent strata. MUSACCHIO & CHEBLI (1975) emphasized the cases of the ostracods *Cypridea diminuta* VANDERPOOL (1928), *Cypridea amerikana* MUSACCHIO (1975), *Cypridea craigi* MUSACCHIO (1975), *Flabellochara harrisi* (PECK 1941) and *Porochara mundula* (PECK 1941).

Different Campanian and Maastrichtian assemblages show also evidences of interchange with the Northern Hemisphere. The case of *Platychara compressa* (KNOWLTON) - a charophyte from Maastrichtian/Paleocene levels in Argentina, originally described in USA - is an example. Asian affinities deserve particular interest. Many genera of charophytes as *Gobichara, Amblyochara, Grambastichara* and *"Porochara"* are also well represented in Asia or resemble, in cases, other Asian species (WANG ZHEN 1978, MUSACCHIO 1981). Moreover, "cristid" or "lip-bearing" ostracods belonging to the typically Asian subfamily Talicyprideinae HOU YOU-TANG (1982) were recently found in Northern Patagonia (see this paper, Plate 5).

Different mechanisms of dispersion mainly for eggs of ostracods, such as the action of winds and the transport by insects, have been previously proposed (see mainly HELMDACH and SOHN). However, the palaeogeography during the Cretaceous period, particularly with regard to the relationships among South America and the Northern Hemisphere, is not completely known. Thus, a complete outline for the distribution of the inherent continental biota and their changes is still needed.

Finally, there is the problem of the species structure. The nature of species in continental charophytes and ostracods seems to be mutually quite different. Dealing with charophytes, GRAMBAST (see mainly: 1966, 1968, 1974) has recognized, in a series of classical papers, the existence of trends in clavatoraceans. This is a group with sexual reproduction and low provincialism showing, moreover, a relatively slow evolutionary ratio. Other fossil groups, in spite of differences in the systematic position and in their ecology, exhibit similitudes. It is the case, for instance, for the patterns of evolution of the Lower Cretaceous benthonic foraminifers studied by BETTENSTAEDT (1968) and BETTENSTAEDT & SPIEGLER (1975) from the Hannover Basin including, as examples, the cases of *Vaginulina procera* (transformation) and *Lagena hauteriviana* (cladogenesis).

Non-marine speciation and patterns of evolution in continental ostracods seem, on the other hand, very difficult to be organized in lineages. Such is the case in the genus *Cypridea* which is fruitful in morphotypes but not well known in its biology, particularly in the reproduction. The great number of taxa in different Lower Cretaceous basins of the world suggests speciation mechanisms which, if they are not well known, can be better analysed comparing them with the recent cases of "species-flock" and "explosive evolution" shown, for instance, in the cichlid fishes from the African lakes (GREENWOOD 1974). Our fossil register includes groups of well established morphotypes (? groups of clones) with continuity in space and permanence in time. They can be found together with cases of isolated morphs of "species" of sudden apparition, without continuity in the column.

Table 3. Non-marine microfossil assemblages.

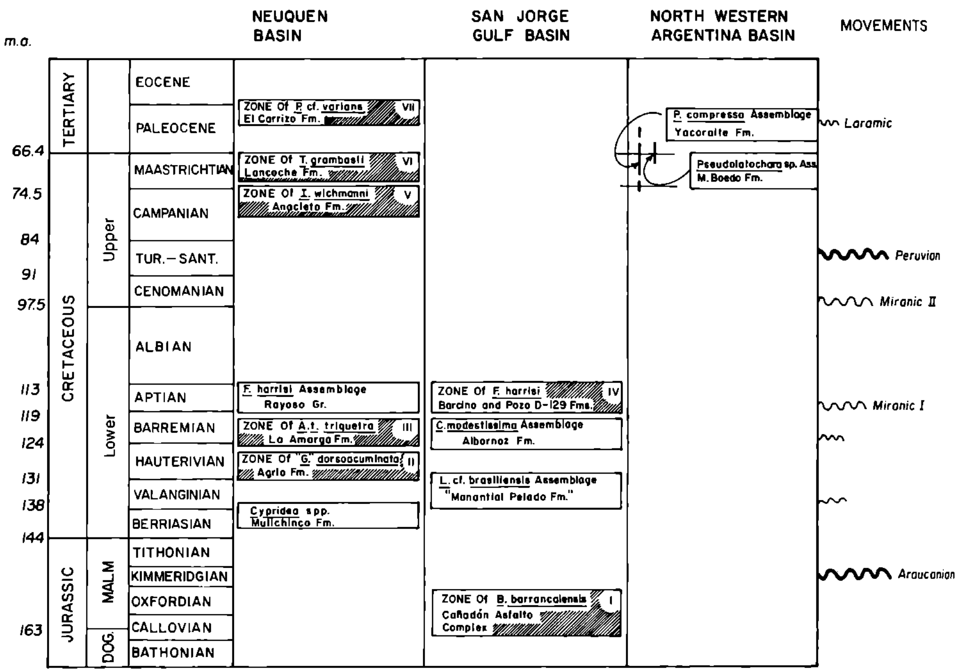


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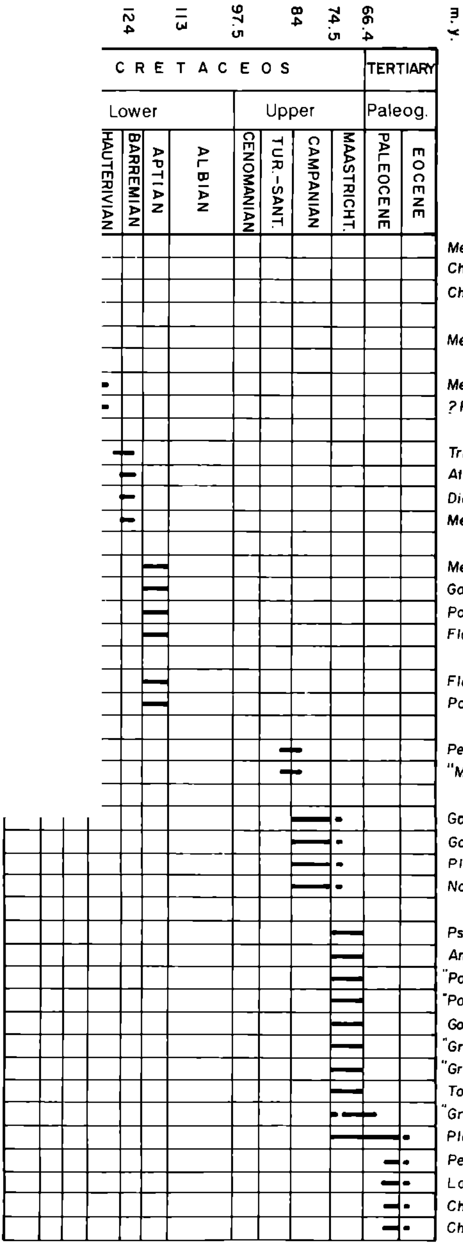
Table 1. Chronological distribution of the Cretaceous charophytes from Argentina.



Mesochara cf. maruchoensis SIM. & MUS. 1986	Lowermost Cretaceous at Ferrarotti Profile
Charoidea gen. et sp. indet. 1 MUS. 1987	
Charoidea gen. et sp. indet. 2 MUS. 1987	
Mesochara sp.	MULICHINCO FM.
Mesochara maruchoensis SIM. & MUS. 1986	AGRIO FORMATION
?Peckisphaera sp. SIM. & MUS. 1986	
Triclypella patagonica nov. sp.	LA AMARGA FORMATION
Atopochara trivolvis triquetra GRAMBAST 1968	
Diectochara andica MUS. 1971	
Mesochara cf. stipitata (WANG S. 1965)	
Mesochara sp. MUS. & PAL. 1975	RANQUILES FORMATION
Gobichara sp. MUS. & PAL. 1975	
Porochara sp. MUS. & PAL. 1975	
Flabellochara cf. harrisi	
Flabellochara harrisi (PECK 1941)	Cº BARCINO and POZO D-129 FMS.
Porochara mundula (PECK 1941)	
Peckisphaera portezueloensis MUS. 1973	PORTEZUELO MEMBER (NEUQUEN GROUP)
"Mesochara" ameghinoi MUS. 1973	
Gobichara walpurgica (MUS. 1973)	ANACLETO FORMATION (NEUQUEN GROUP)
Gobichara tenuis (Mus. 1973)	
Platychara cf. caudata GRAMBAST 1971	
Nothochara apiculata MUS. 1973	
Pseudolatochara sp.	LONCOCHE, MARIANO BOEDO and YACORAITE FORMATIONS
Amblyochara sp. MUS. 1972	
"Porochara" ovalis (FRIETZSCHE 1924)	
"Porochara" gieldemeisteri KOCH & BLIS. 1960 s.l.	
Gobichara groeberi (MUS. 1978)	
"Grambastichara" sp. 1 (MUS. 1978)	
"Grambastichara" sp. 2 (MUS. 1978)	
Tolypella grambasti MUS. 1978	
"Grambastichara" yuntaishaensis W. ZHEN 1978	
Platychara compressa (KNOWLTON 1888)	LONCOCHE, MARIANO BOEDO and YACORAITE FORMATIONS / EL CARRIZO FORMATION
Peckichara cf. varians meridionalis MASS. 1981	EL CARRIZO FORMATION
Lamprothamnium sp. MUS. & MOR. 1983	
Charoidea forma A MUS. & MOR. 1983	
Charoidea forma B MUS. & MOR. 1983	

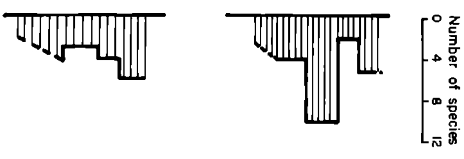
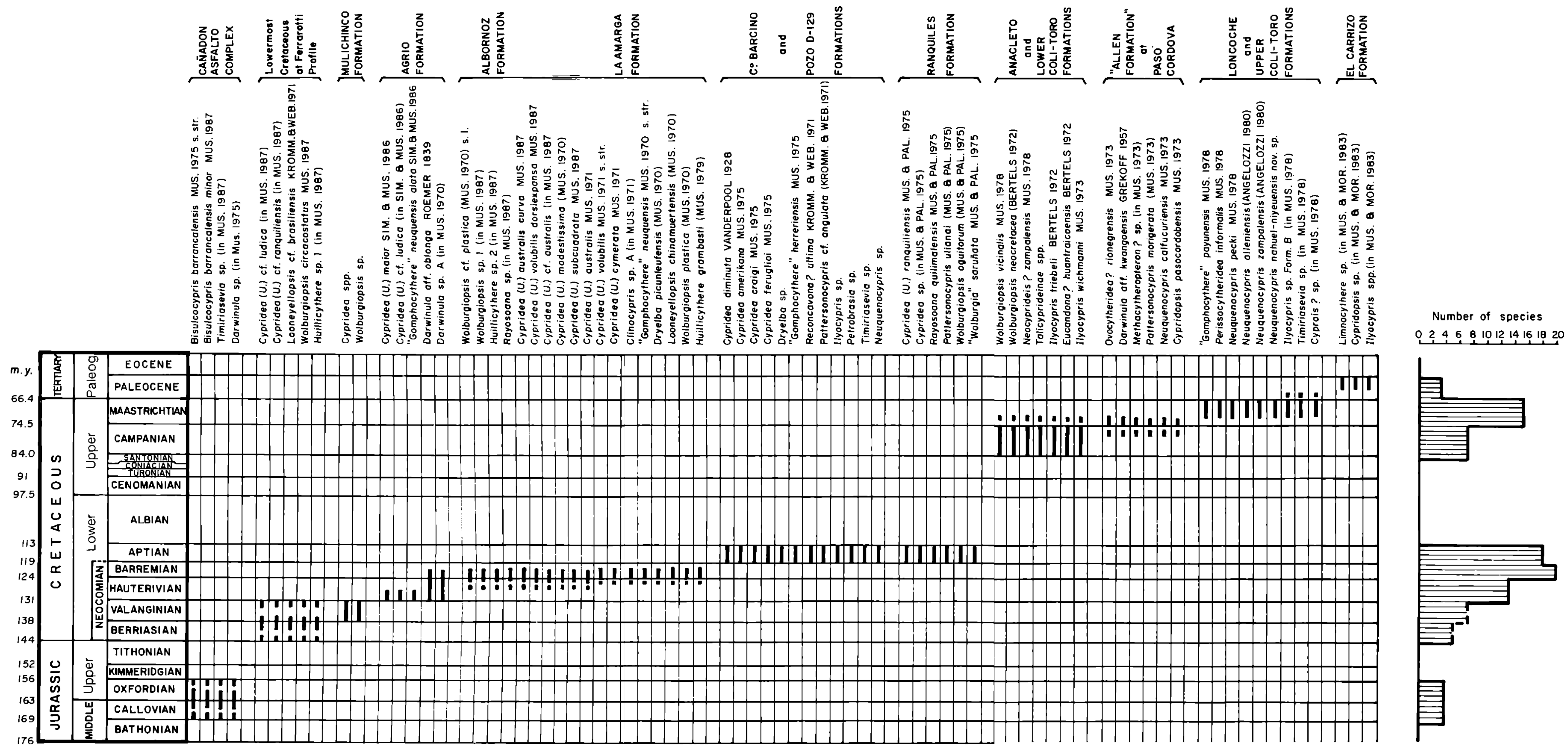


Table 2. Chronological distribution of the Cretaceous non-marine ostracods from Argentina.



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Systematic Appendix

Subclass Ostracoda LATREILLE 1806
Superfamily Cypridacea BAIRD 1845
Genus *Neuquenocypris* nom. transl.

e x: *Ilyocypris (Neuquenocypris)* MUSACCHIO 1973.

T y p e - s p e c i e s: *Ilyocypris (Neuquenocypris) calfucurensis* MUSACCHIO 1973.

D e f i n i t i o n: Subrectangular to subtrapezoidal carapace of medium to great size and well marked cardinal angles. RV greater than the LV. Antero-dorsal depression with two not well marked sulci. Well-ornamented surface including hollow spines, pustules, tubercles, papillae, and/or punctuations; marginal denticulations and marginal alligned papillae. Hinge adont. Inner lamellae lacking lists or smooth. Central muscle scars pattern as in *Ilyocypris*.

R e m a r k s: The present taxon includes the following species:

Cypridacea gen. et sp. indet. (in MUSACCHIO & MORONI 1983: 30), Paleocene.
Neuquenocypris nahuel-niyehuensis nov. sp., Maastrichtian/?Danian.
Neuquenocypris peckii (MUSACCHIO 1978) nov. comb., Maastrichtian.
Neuquenocypris zampalensis (ANGELLOZI 1980) nov. comb., Maastrichtian.
Neuquenocypris alleniensis (ANGELLOZI 1980) nov. comb., Maastrichtian.
Neuquenocypris calfucurensis (MUSACCHIO 1973) nov. comb., Campanian/Maastrichtian.
Neuquenocypris sp. this paper, Plate 3, Figs. 5-6, Campanian.
?*Neuquenocypris* sp. (in HECHEM et al. 1987), Aptian.

This genus can be compared with *Ilyocypris* BRADY & NORMAN, 1889, and *Pelocypris* KLIE, 1939. However, the inner lamellae seems always smooth and the RV overlaps the LV at the extracardinal area. Some species of *Neuquenocypris* resemble *Parailyocypris* HOU 1956, and *Ilyocyprimorpha* MANDELSTAM 1956. In particular *Ilyocyprimorpha palustris* MANDELSTAM 1956 is thick-walled, the single anterior-dorsal depression is not well defined; the spines are massive and the outline from lateral view - which is described as concave - seems, however, more rectilinear than in most of the neuquean taxa.

Neuquenocypris nahuel-niyeuensis nov. sp.
Plate 4, Figs. 7-12

Type-level: Coli Toro Formation (Maastrichtian/?Danian).
Type-locality: Rincón de Treneta (also well represented at Nahuel-Niyeu), Province of Río Negro.
Holotype: BA-G-CM 80/1 (Fig. 7, carapace) (Museo Rivadavia, Colección Musacchio).
Paratypes: BA-G-CM 80/2-6 (Figs. 8-12).
Definition: Carapace of rounded-subrectangular outline from lateral view, moderately tumid; pitted surface, bearing papillae regularly distributed, which are, besides, alligned at the margin in both valves.

Division Charophyta
Orden Charales
Family Clavatoraceae PIA 1927
Genus *Triclypella* GRAMBAST 1969

Triclypella patagonica nov. sp.

References: *Triclypella* aff. *calcitrapa* GRAMBAST 1969, in: MUSACCHIO 1971: 23-26, pl. 1, figs. 1-4, pl. 2, figs. 19-23, text-fig. 3. *Triclypella* sp. MUSACCHIO 1979, pl. 7, figs. 16-17.
Type-level: La Amarga Formation, section of marls and limestones, late Hauterivian.
Type-locality: Cerro China Muerta, Province of Neuquén.
Holotype: LP - PB (La Plata - Paleobotánica) Pm 569 (MUSACCHIO 1971: plate 1, fig. 2).
Paratypes: LP - PB, Pm: 569 (the other specimens).
Definition: Globular utricle formed by three ramules of 10 (9-12) cells. Each ramule radiates at or just over the 1/2 height of the utricle from a tabular subvertical unit. This unit starts from the basis and grows upwards turning tenuously to the left. Outline of the utricle from upper view subcircular to roughly triangular, lacking of "horn-like" form.

INDEX MICROFOSSILS

Plate 1

1) Zone of Bisulcocypris barrancalensis (Callovian-Oxfordian?)

Figs. 1-4: *Bisulcocypris barrancalensis* MUSACCHIO 1975 s. str. (in MUSACCHIO & CHEBLI 1975)

Characteristics: Dimorphic carapace with ellipsoidal-rectangular outline (lateral view). Surface with a delicate polygonal reticulation, papillae and antero-ventral marginal denticulations; ventrally, the reticulation is resolved into several fine longitudinal crests.
Fig. 1: Holotype (♀), dorsal view; length 1.120 µ, height 680 µ, width 695 µ.
Fig. 2: Paratype (♂), dorsal view; length 1.180 µ, height 660 µ, width 575 µ.
Fig. 3: Instar (-1) from left side.
Fig. 4: Carapace (♂) from right side.
All specimens from El Barrancal, Chubut.

Fig. 5: *Bisulcocypris barrancalensis minor* MUSACCHIO 1987 (in MUSACCHIO et al., in press)

Characteristics: A subspecies with carapace more swollen and smaller than the *barrancalensis* s. str. morphotype; the papillae are more reduced.
Fig. 5: Holotype (♀), dorsal view; length 969 µ, height 586 µ, width 590 µ. Canadón Asfalto Complex, upper part, at Puesto Almada, Chubut.

Fig. 6: *Darwinula* sp. (in MUSACCHIO & CHEBLI 1975)

Characteristics: Carapace of lengthened-ovoidal form. Ventral margin, in lateral view, almost rectilinear but tenuously concave at the anterior medium part.
Fig. 6: Carapace from right side; length 1.055 µ, height 520 µ. El Barrancal, Chubut.

2) Zone of "Gomphocythere" dorsoacuminata (early Hauterivian)

Figs. 7-9: *"Gomphocythere" dorsoacuminata* MUSACCHIO 1970

Characteristics: Lanceolate carapace in the male morph and pyriform in the female morph, both in dorsal view. Surface with a ventral rib, which is more expanded in the male morph, several hollow tubercles, a polygonal reticulation and papillae. The large right valve overhangs the left valve at the dorsal part, forming a curved dorsal margin in lateral view.
Fig. 7: Carapace (♀), dorsal view (x 55). Topotype from the Agrio Formation at El Marucho, Neuquén.
Fig. 8: Carapace (♂), dorsal view (x 55). Topotype from the Agrio Formation at El Marucho, Neuquén.

Fig. 10: *Mesochara maruchoensis* SIMEONI & MUSACCHIO 1986

Characteristics: Minute gyrogonite with subprolate to sphaeroidal-prolate form and basal pole projected but truncated. 6 (5-7) cellular turns

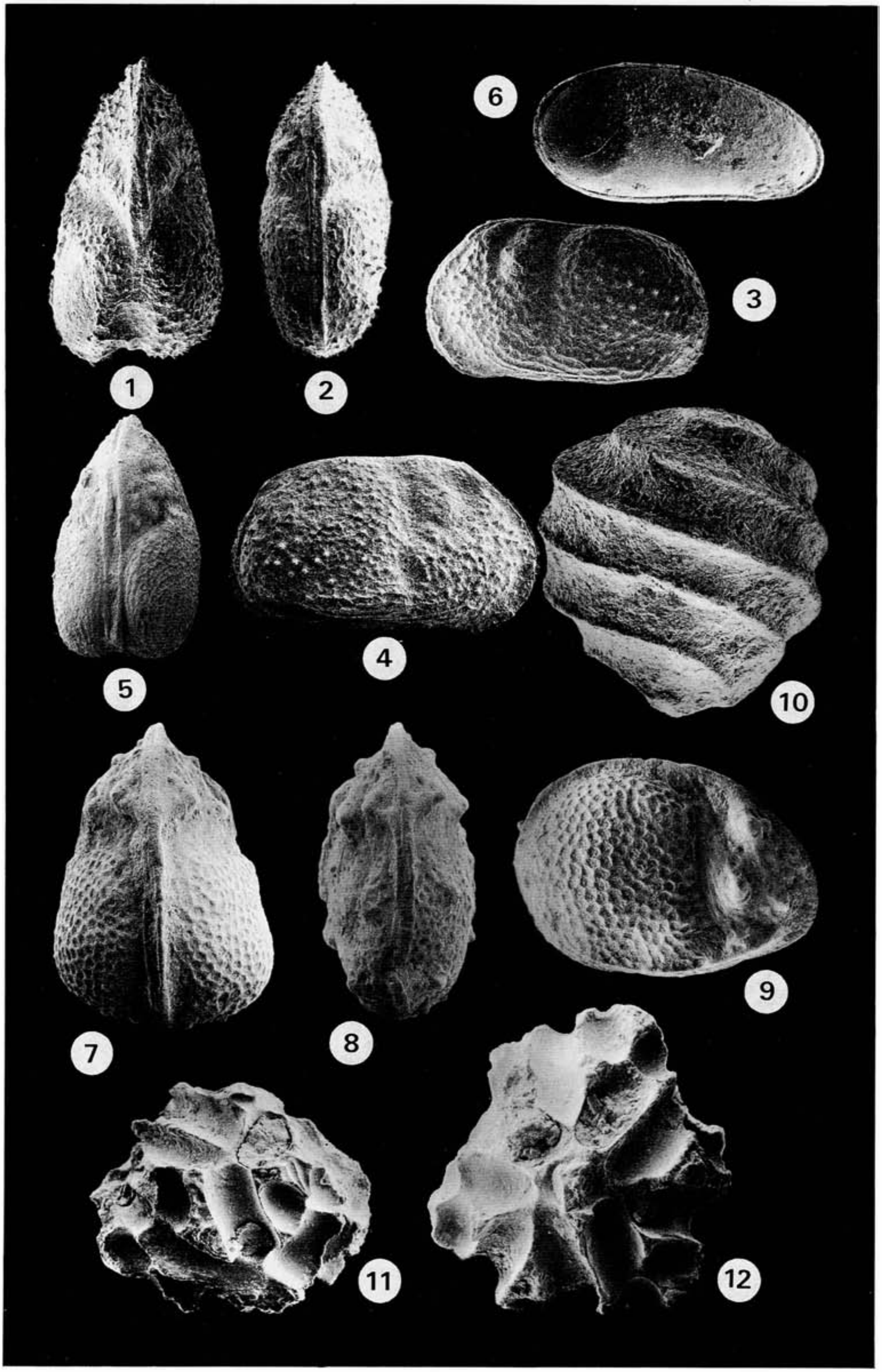

Plate 1

Plate 1, cont.

in lateral view. LPA ranging from 250 to 360 µ and LED from 250 to 300 µ.
Fig. 10: Gyrogonite (paratype) from lateral view. Agrio Formation at El Marucho, Neuquén.

3) Zone of Atopochara trivolvis triquetra (late Hauterivian-early Barremian)

Figs. 11-12: *Atopochara trivolvis triquetra* GRAMBAST 1968

Characteristics: "Utricle with the structure of *Atopochara trivolvis*, showing a more or less triangular and irregular outline when seen from apex or base. Branches of each of the three fundamental utricular groups sometimes not closely adpressed. Cell of antheridial origin sometimes retaining slight shield-cell ornamentation." (GRAMBAST 1968: 9).
Fig. 11: Utricle in lateral view (x 55).
Fig. 12: Utricle in basal view (x 55). Different specimens from La Amarga Formation at Picún Leufú, Neuquén.

Plate 2

(Zone of Atopochara trivolvis triquetra-Secondary markers)

Figs. 1-2: *Triclypella patagonica* nov. sp.

Characteristics: Globular utricle formed by three ramules of 10 (9-12) cells. Each ramule radiates at or just over the 1/2 height of the utricle from a tubular subvertical unit. This unit starts from the basis and grows upwards turning tenuously to the left. Outline of the utricle from upper view subcircular to roughly triangular but without three-horned shape.
Fig. 1: Utricle from lateral view (x 66). La Amarga Formation at Cerro China Muerta, Neuquén (late Hauterivian).
Fig. 2: Utricle in lateral view (x 66). La Amarga Formation at Picún Leufú (early Barremian).
Reference: *Triclypella* aff. *calcitrapa* GRAMBAST (in MUSACCHIO 1971: 23).

Fig. 3: *Cypridea (Ulwellia) australis* MUSACCHIO 1971

Characteristics: Carapace with right valve overlapping the left valve; however, the left valve reaches the maximum height at the anterior cardinal angle zone. Trapezoidal outline in lateral view. Surface strongly punctuated and with numerous and minute papillae. Circular cyatus. Length of the holotype 1.035 µ, height 710 µ.
Fig. 3: Carapace in left view (x 60). Topotype from La Amarga Formation at China Muerta, Neuquén.

Fig. 4: *Cypridea (Ulwellia) subcuadrata* MUSACCHIO 1971

Characteristics: Carapace of subcuadrate outline in lateral view.

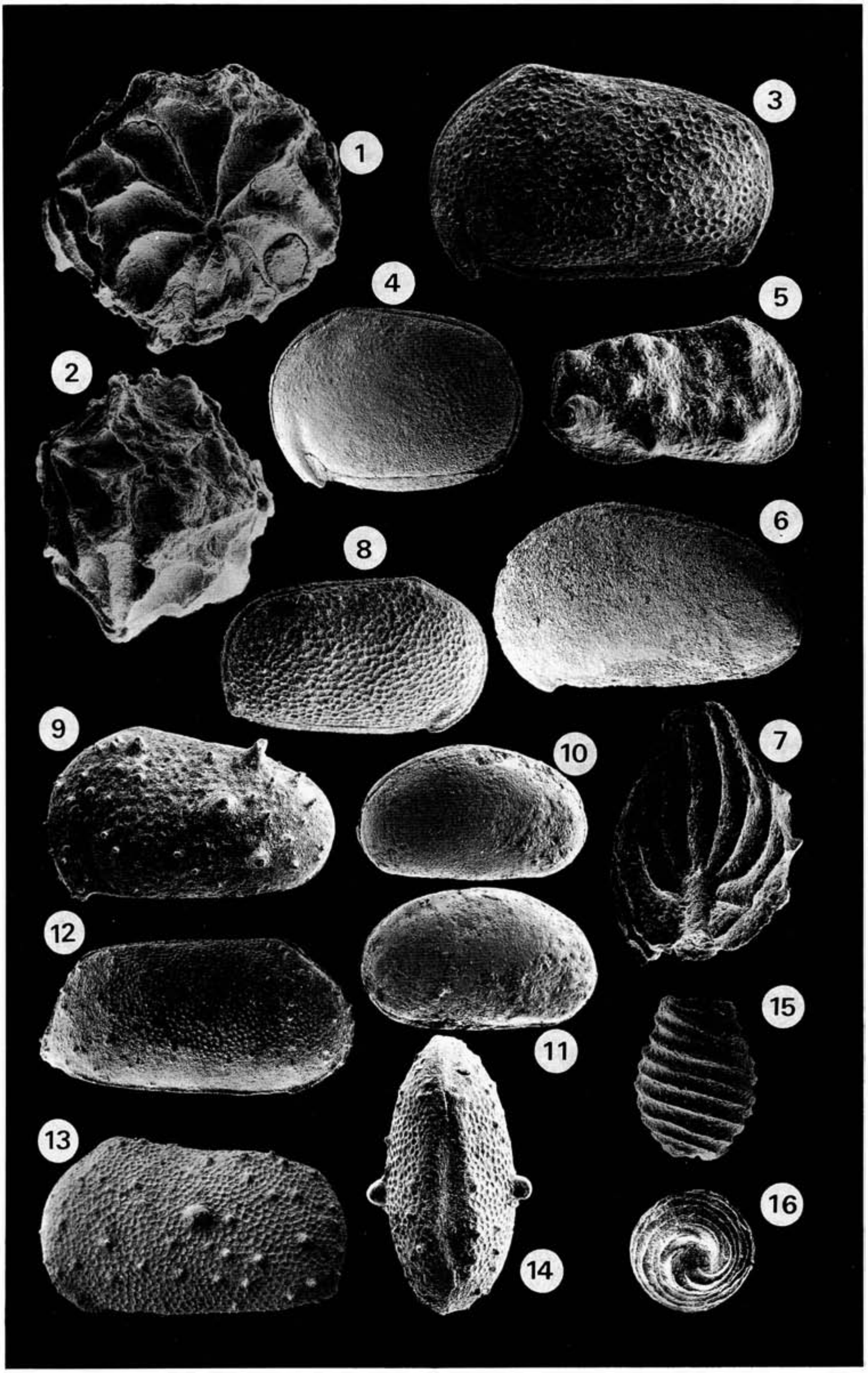
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Plate 2, cont.

Right valve overlapping the left valve. Subcircular cyatus. Surface tenuously punctuated.
Fig. 4: Carapace in lateral view (x 55). Albornoz Formation, Chubut.

Fig. 5: *Huillicythere grambasti* MUSACCHIO 1979

Characteristics: Small carapace with lanceolate form in dorsal view and subrectangular outline in lateral view. Seven prominent tubercles: three of them ventral in position, attached together by an alar ventral expansion; the other four are dorsal. Surface roughly reticulated, particularly at the flattened ventral side, and with papillae.
Fig. 5: Paratype in lateral view; length 495 μ, height 265 μ, width 285 μ.

Fig. 6: *Cypridea (Ulwellia) modestissima* (MUSACCHIO 1971)

Characteristics: Laterally flattened carapace with right valve overlapping the left valve. Rounded trapezoidal outline in lateral view with anterior margin continuously curved. Smooth or tenuously punctuated surface.
Fig. 6: Carapace in left view (x 55). Albornoz Formation, Chubut.

4) Zone of Flabellochara harrisi (late Hauterivian-early Barremian)

Fig. 7: *Flabellochara harrisi* (PECK 1941) GRAMBAST 1969

Characteristics: Utricle of subglobular form with roughly bilateral summetry. It is composed of two groups of radiating cells which fan out from a short vertical lower furrow to cover a considerable part of the gyrogonite.

The specimens of Paso de Indios include 10 (8-11) cells for each fan group and the utricles are more frequently 695 (770) 880 μ of height and 545 (602) 680 μ of width (in 50 specimens).
Fig. 7: Utricle in lateral view (x 55). Cerro Barcino Formation at Paso de Indios, Chubut.

Fig. 8: *Cypridea diminuta* VANDERPOOL 1928

Characteristics: Carapace with left valve overlapping the right valve. Subrectangular outline in lateral view. Small rostrum turned upwards. Reticled surface with conjunctive papillae. In 46 specimens from Paso de Indios, length 720-830 μ, height 435-505 μ.
Fig. 8: Carapace in lateral view (x 55). Cerro Barcino Formation, Paso de Indios, Chubut.

Fig. 9: *Cypridea craigi* MUSACCHIO 1975 (in MUSACCHIO & CHEBLI 1975)

Characteristics: Carapace with left valve overlapping the right valve. Oval-trapezoidal outline in lateral view with anterior cardinal angle well marked and the posterior angle rounded. Punctuated surface with papillae and spines, including a "polar" and a "lumbar" pair of the largest spines, variably developed in different populations. Holotype from Cerro Barcino Formation at Paso de Indios, Chubut: length 1.100 μ, height 635 μ, width 635 μ.

Plate 2, cont.

Fig. 9: Carapace from left side (x 40). Cerro Barcino Formation at Paso de Indios, Chubut.

Figs. 10-11: *Reconcavona? ultima* KRÖMMELBEIN & WEBER 1971

Characteristics: Smooth carapace with the left valve overlapping the right valve. Suboval outline in lateral view. Greatest height approximately at 1/2 length. Anterior margin a little more broadly rounded than the posterior margin.
Figs. 10-11: Carapaces of different specimens from left side (x 50). Pozo D-129 Formation (from subsoil), Chubut.

Fig. 12: *Rayosoana* sp. (in MUSACCHIO & CHEBLI 1975)

Characteristics: A very exiguously represented species of great size from Paso de Indios. Carapace lengthened, with rectangular outline in lateral view. Posterior margin slopping at great angle toward the pointed postero-ventral end, coinciding with the maximum length. Right valve overlapping the left valve; length 1.550 μ, height 725 μ.
Fig. 12: Carapace in lateral right view.

Figs. 13-14: *Cyprideo amerikana* MUSACCHIO 1975 (in MUSACCHIO & CHEBLI 1975)

Characteristics: Carapace with left valve overlapping the right valve. Rectangular outline in lateral view. Reticled surface with papillae and a "polar" tubercle.
Fig. 13: Holotype from right side; length 970 μ, height 560 μ, width 455 μ. Cerro Barcino Formation at Paso de Indios, Chubut.
Fig. 14: Carapace in dorsal view (topotype).

Figs. 15-16: *Porochara mundula* (PECK 1941) GRAMBAST 1968

Remarks: The Argentine morphotype has prolate to subprolate (ISI) and subovoidal to ellipsoidal (ANI) gyrogonites; the average size is smaller and the form more cylindroid than the type-material described by PECK (1941). Measurements of 100 specimens from Paso de Indios, Chubut: LPA 360 (435) 515 μ, LED 230 (296) 335 μ; 11 (9-12) cellular turns in lateral view.
Fig. 15: Gyrogonite in lateral view (x 66). Cerro Barcino Formation at Paso de Indios.
Fig. 16: Gyrogonite in apical view (x 66). Cerro Barcino Formation at Paso de Indios.

Plate 3

5) Zone of Ilyocypris wichmanni
(Campanian-?early Maastrichtian)

Figs. 1-2: *Ilyocypris wichmanni* MUSACCHIO 1973

Characteristics: Carapace with swollen lobes; particularly the posterior (L 3) is notably swollen and arranged continuously lengthwise at the postero-dorsal and posterior margin. Punctuated or smooth surface with papillae having variable development.
Fig. 1: Holotype, dorsal view; length 720 μ, height 430 μ, width 390 μ.
Fig. 2: Right valve (topotype x 50). Both from Anacleto Formation at Vista Alegre, Neuquén.

Fig. 3: *Gobichara walpurgica* (MUSACCHIO 1973)

Characteristics: Gyrogonite more frequently subprolate (ISI) and subovoidal to ellipsoidal (ANI). Summit broadly rounded and tenuously depressed at the periapical periphery. Pyramidal-truncated basal expansion, prominent and variably truncated. LPA in 20 specimens ranging from 445 (545) to 565 μ and LED from 370 (428) to 490 μ; 9-10 cellular turns in lateral view.
Fig. 3: Holotype in lateral view. Anacleto Formation at Paso Cordova, Río Negro.

Fig. 4: *Gobichara tenuis* (MUSACCHIO 1973)

Characteristics: Gyrogonite of prolate to subprolate (ISI) form and ellipsoidal to subovoidal (ANI). Rounded summit, tenuously depressed at the periapical periphery. Prominent and variably truncated basal expansion. LPA in 20 specimens between 495 (544) 615 μ, LED 365 (386) 460 μ; 11 (10-12) cellular turns from lateral view.
Fig. 4: Holotype. Anacleto Formation at Paso Cordova, Río Negro.

Figs. 5-6: *Neuquenocypris* sp.

Fig. 5: Carapace from left side (x 55).
Fig. 6: Same specimen in ventral view (x 55). Anacleto Formation at Cerro Villegas, Neuquén.

Figs. 7-8: *"Cypridea"* sp.

Fig. 7: Carapace from left side (x 50). Anacleto Formation at Vista Alegre, Neuquén.
Fig. 8: Detail (x 150).
Reference: MUSACCHIO 1973: 20.

6) Zone of Tolypella grambasti
(Maastrichtian)

Figs. 9-10: *Tolypella grambasti* MUSACCHIO (in ULIANA & MUSACCHIO 1978)

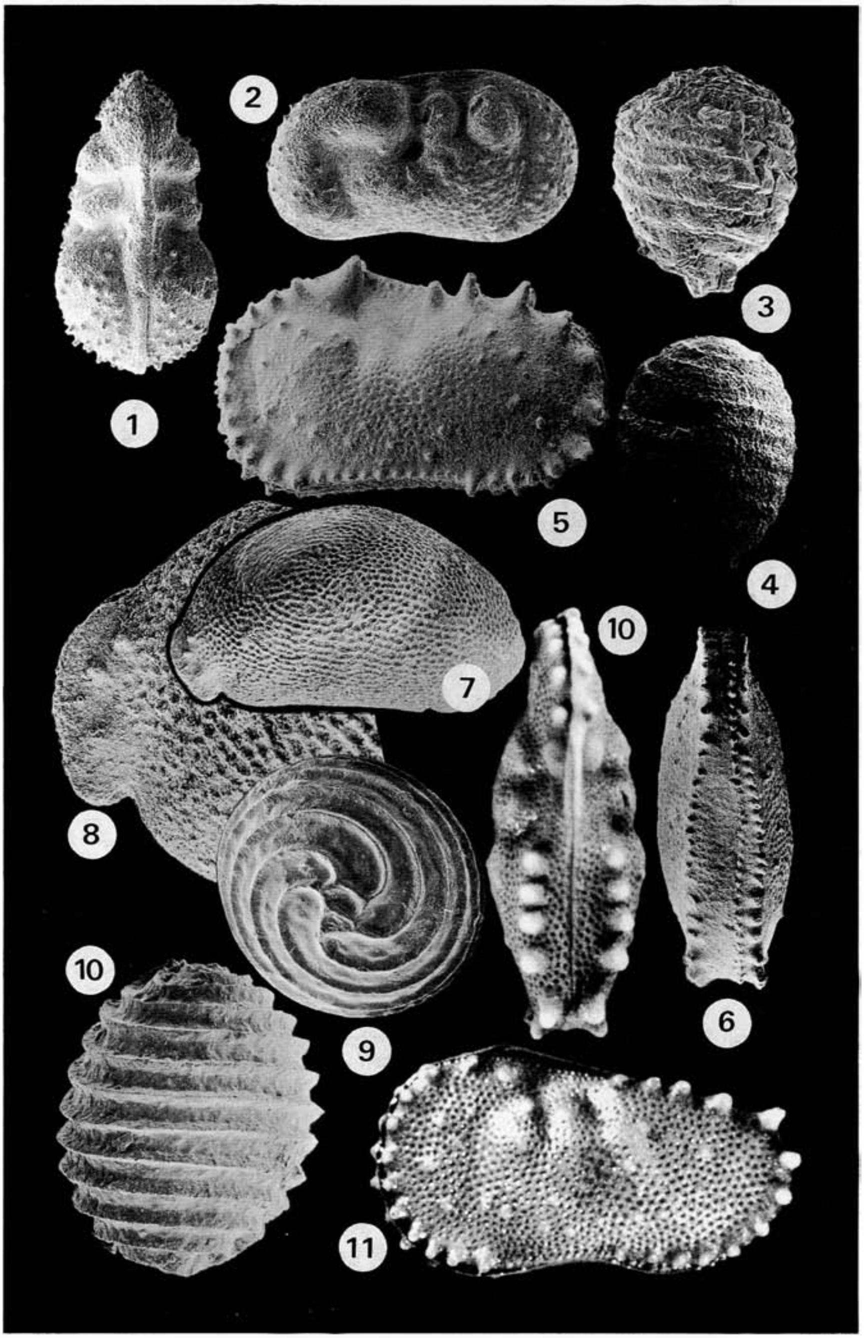

Plate 3

Plate 3, cont.

Characteristics: Minute gyrogonite of ellipsoidal to sphaeroidal-subprolate form. Basal plate duplicate. LPA in 25 specimens 270 (288) 320 µ, LED 230 (249) 285 µ; 10.5 (9-12) cellular turns in lateral view.
Fig. 9: Paratype in basal view. Loncoche Formation at Zampal, Mendoza.
Fig. 10: Holotype in lateral view. Loncoche Formation at Zampal, Mendoza.

Figs. 11-12: *Neuquenocypris pecki* MUSACCHIO (in ULIANA & MUSACCHIO 1978)

Characteristics: Carapace laterally compressed, particularly in the anterior part. Prominent hollow spines forming a peripheral linear "fringe-like" arrangement. Punctuated surface with papillae.
Fig. 11: Paratype, dorsal view (x 60).
Fig. 12: Holotype from left lateral side (x 60). Loncoche Formation at Zampal, Mendoza.

Plate 4

(Zone of Tolypella grambasti-Secondary markers)

Figs. 1-2: *Neuquenocypris alleniensis* (ANGELOZZI 1980) nov. comb.

Characteristics: Carapace of subtrapezoidal outline in lateral view with a prominent antero-ventral border broadly rounded. Subcentral tubercle (T3) and a secondary similar one underlying, both covered with an "influorescence-like" of numerous papillae. Punctuated surface with regularly distributed papillae.
Fig. 1: Carapace in basal view (x 50).
Fig. 2: Carapace from left lateral side (x 50).
Both specimens from Allen Formation at Puesto Rebolledo, Río Negro.

Fig. 3: *Neuquenocypris zampalensis* (ANGELOZZI 1980) nov. comb.

Characteristics: Carapace with subrectangular outline in lateral view. Numerous papillae regularly distributed, except in the sulci at the antero-dorsal depressed area. Marginal denticulations well marked and arranged lengthwise of the extracardinal border of the right valve.
Fig. 3: Carapace from left side (x 50). Topotype from Loncoche Formation at Zampal, Mendoza.

Figs. 7-12: *Neuquenocypris nahuel-niyeuensis* nov. sp.

Characteristics: Carapace of rounded-subrectangular outline in lateral view, moderately tumid; pitted surface, bearing regularly distributed pore-papillae and forming a marginal arrangement in both valves.
Fig. 7: Holotype, carapace from left side (x 57).
Fig. 8: Paratype; right valve from inner side (x 230).
Fig. 9: Paratype; carapace from left side (x 57).
Fig. 10: Paratype; central muscle scars (x 230).
Fig. 11: Paratype; carapace in basal view (x 57).

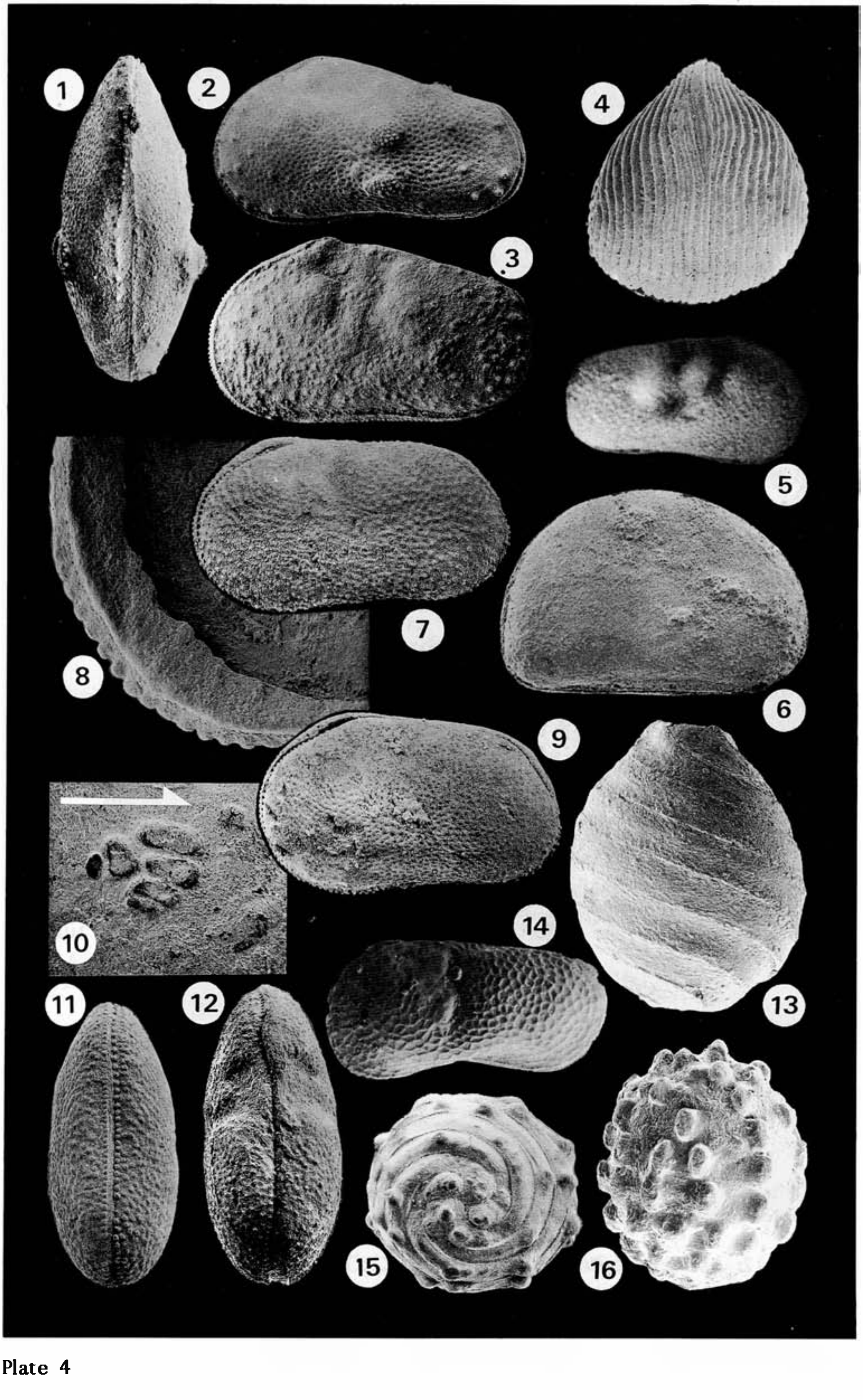

Plate 4

Plate 4, cont.

Fig. 12: Paratype; carapace in dorsal view (x 57).
Type level: Coli Toro Formation at Treneta, Río Negro.

Other species of the Grambasti-Zone exiguously or not well represented, up to the present, but with interest in correlations:

Fig. 5: *Ilyocypris* sp. morphotype B (in ULIANA & MUSACCHIO 1978)

Carapace from right side (x 66); Loncoche Formation at Zampal, Mendoza.

Fig. 4: *Timiriasevia* sp. (in ULIANA & MUSACCHIO 1978)

Carapace in dorsal view (x 120); Loncoche Formation at Zampal, Mendoza.

Fig. 6: *Cyprois?* sp. (in ULIANA & MUSACCHIO 1978)

Carapace from right side (x 150); Loncoche Formation at Zampal, Mendoza.

Fig. 13: *"Porochara" gildemeisteri* KOCH & BLISENBACH 1960

Gyrogonite in lateral view (x 00). Mariano Boedo Formation, Córdoba.

7) Zone of Peckichara cf. varians meridionalis (Paleocene)

Fig. 14: *Limnocythere* sp. (in MUSACCHIO & MORONI 1983)

R e m a r k s : Lengthened carapace with rectilinear dorsal margin and surface ornamented by a polygonal reticle. Length 800 µ, height 400 µ in the figured specimen.
Fig. 14: Left valve; El Carrizo Formation (type-profile), Río Negro.

Figs. 15-16: *Peckichara* cf. *varians meridionalis* MASSIEUX 1981

R e m a r k s : Globular gyrogonites with spiral cells ornamented by tubercles, including a variably projected rosette at the apical tips. Base tenuously prominent in many specimens. LPA (in 100 gyrogonites) 460 (759) 1015 µ; LED 415 (675) 965 µ; 8 (6-9) cellular turns in lateral view.
Fig. 15: Gyrogonite from apical side (x 57); El Carrizo Formation (type-profile), Río Negro.
Fig. 16: Gyrogonite in lateral view (x 57); different specimen of the same precedence.

Plate 5

Other microfossils of palaeobiogeographical interest

Figs. 1-2: *Looneyellopsis* cf. *brasiliensis* KRÖMMELBEIN & WEBER 1971

Fig. 1: Left valve.

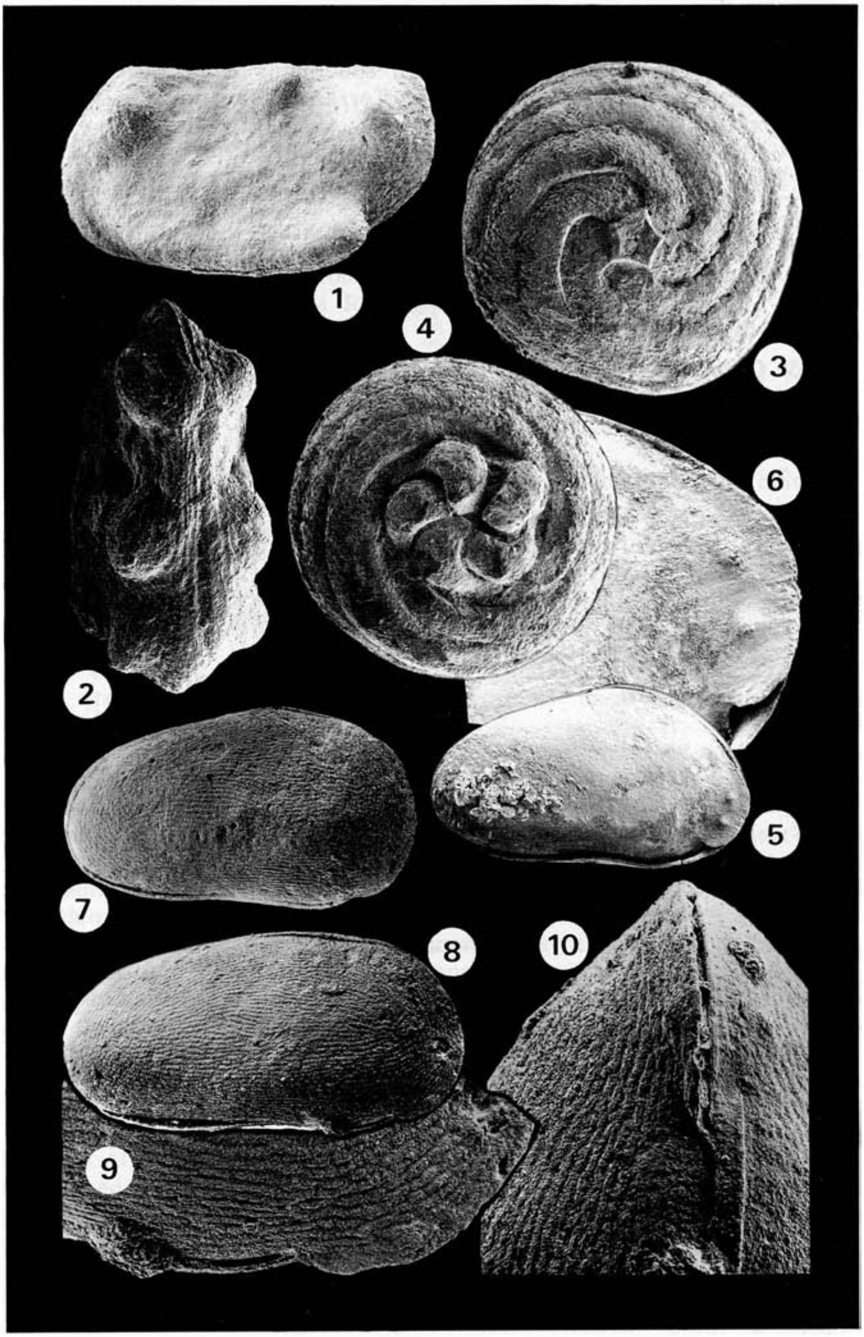

Plate 5

Plate 5, cont.

Fig. 2: Carapace in dorsal view. Both specimens from the lowermost Cretaceous sediments at the Cerro Ferrarotti Profile, Chubut (Neocomian). x 100.

Figs. 3-4: *Platychara compressa* (KNOWLTON 1888)

Fig. 3: Gyrogonite in basal view (x 66).
Fig. 4: Gyrogonite from apical side (x 66). Both specimens from the Loncoche Formation at Zampal, Mendoza (Maastrichtian).

Figs. 5-6: *Altanicypris?* sp. 1

Fig. 5: Carapace from right side (x 66). Coli Toro Formation at Nahuel Nique, Río Negro (Maastrichtian).
Fig. 6: Detail (x 150).

Figs. 7-10: *Altanicypris?* sp. 2

Fig. 7: Carapace from left side (x 66).
Fig. 8: Carapace from left side (x 66).
Fig. 9: Detail in the same specimen of Fig. 8 (x 150).
Fig. 10: Same specimen of Fig. 8 in basal view (anterior part) (x 150).
Both specimens from Anacleto Formation at Cerro Villegas, Neuquén (Campanian - ?Lower Maastrichtian).

Lower and Mid-Cretaceous Coralline Sponge Communities of the Boreal and Tethyan Realms in Comparison with the Modern Ones – Palaeoecological and Palaeogeographic Implications

JOACHIM REITNER, Berlin

With 25 Text-Figures

REITNER, J. (1989): Lower and Mid-Cretaceous Coralline Sponge Communities of the Boreal and Tethyan Realms in Comparison with the Modern Ones - Palaeoecological and Palaeogeographic Implications. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 851-878. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: The coralline sponges (chaetetids, stromatoporoids, sphinctozoans, pharetronids) are sponges with a rigid calcareous basal skeleton. Within modern oceans 3 communities are observed:

1. An Indopacific reef community characterized by calcified demosponges like *Acanthochaetetes, Astrosclera, Stromatospongia, Merlia,* and pharetronid calcarean sponges (Murrayonida, *Minchinellia*).
2. A Mediterranean non-reefal community characterized by pharetronids (e. g. *Petrobiona*) and the calcified demosponge *Merlia normani.*
3. A Caribbean reef community characterized by ceratoporellid/agelasid calcified demosponges.

From the mid-Cretaceous comparable communities are known. Non-reefal subtropical pharetronid communities are known from the Aptian of Faringdon (England) and the Cenomanian of the Essener Grünsand (West Germany). These sponge communities are characterized by sycettid sphinctozoans (*Barroisia, Thalamnopora*), minchinellid pharetronids (*Lymnorella*), and rarely by the calcified demosponge *Neuropora.*

The Barremian/Lower Aptian coralline sponge communities from the Tethyan Realm (e. g. Helvetic "Schrattenkalk") are not comparable with the modern ones. After the Middle Aptian (Gargasian) events new coralline sponge communities occurred within the North Spanish Urgonian reef platforms, e. g. the *Acanthochaetetes* community which has close taxonomic and ecological relationships with the modern Pacific *Acanthochaetetes* community.

The genus *Acanthochaetetes* is first reported from the Albian Mural Limestone in Arizona (USA).

From a late Cenomanian hardground (Liencres, Northern Spain) a deeper water sphinctozoan demosponge community is described (*Vaceletia/Acanthochaetetes* community) which has also a close correspondence to some modern occurrences of coralline sponges within deeper water environments of Indopacific reefs.

The extant Indopacific *Acanthochaetetes/Vaceletia* community and the Mediterranean coralline Calcarea community are relict faunas ("living fossil communities") from the mid-Cretaceous Tethyan Ocean.

Kurzfassung: Coralline Spongien (Chaetetiden, Stromatoporen, Sphinctozoen und Pharetroniden) sind Spongien, die neben ihrem spikulären Skelett ein kalkiges Basalskelett besitzen. Innerhalb der heutigen Ozeane lassen sich drei Gemeinschaften unterscheiden:

1. Eine indopazifische Riffauna, die charakterisiert wird durch die Demospongier-Gattungen, *Acanthochaetetes, Astrosclera, Stromatospongia* und *Merlia,* sowie durch die Ordnungen Murrayonida und Minchinellida der pharetroniden Calcarea.
2. Eine mediterrane Spongiengemeinschaft, charakterisiert durch Pharetroniden (*Petrobiona*) und den Demospongier *Merlia normani.*
3. Eine karibische Fauna, die durch ceratoporellide Demospongier charakterisiert wird und keine engeren Beziehungen zu den mediterranen und pazifischen Vorkommen aufweist.

Aus der Unter- und Mittelkreide von Europa, Nordamerika und Kleinasien sind vergleichbare coralline Spongiengemeinschaften bekannt. Aus dem Apt von Faringdon (England) und dem Cenoman von Essen sind reine Pharetroniden-Gemeinschaften bekannt, die Übereinstimmungen mit den rezenten Mittelmeer- und Pazifik-Faunen besitzen. Die fossilen Faunen unterscheiden sich jedoch durch den hohen Anteil von ausgestorbenen syscettiden Sphinctozoen (z. B. *Barroisia*).
Die Barreme- und Unterapt-"Sklerospongier" des Tethys-Raumes besitzen keine unmittelbaren Übereinstimmungen mit den rezenten Faunen. In den Oberapt- und Alb-Riffen von Nordspanien, Südosteuropa und Asien findet sich eine *Acanthochaetetes*-Gemeinschaft, die enge taxonomische und ökologische Übereinstimmungen mit der indopazifischen *Acanthochaetetes*-Gemeinschaft aufweist. Die Gattung *Acanthochaetetes* wird erstmals im Alb des Mural Limestone (Arizona, USA) beobachtet.
Aus dem Obercenoman von Nordspanien wird eine Hardground-Fauna von corallinen Spongien beschrieben (*Vaceletia/Acanthochaetetes*-Gemeinschaft), die einem tieferen subtidalen Milieu zuzuordnen ist und sehr enge Übereinstimmungen mit rezenten pazifischen Vorkommen aufweist. Die rezente *Acanthochaetetes/Vaceletia*-Gemeinschaft ist eine überlebende Reliktfauna der tethyalen Mittelkreide.

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1. Introduction

Coralline sponges are sponges with a secondary rigid calcareous skeleton in addition to the spicular skeleton. This type of sponge belongs to the siliceous demosponges and the order Calcarea. Normally these sponges are called "chaetetids", "stromatoporoids", "sphinctozoans", "pharetronids", and "sclerosponges". These terms represent only grades of organization and cannot be used as taxonomic units (REITNER 1987b, c). All coralline sponges can be classified within the modern sponge classification (VACELET 1985, REITNER 1987b, c, d, WOOD 1987, WOOD, in press).

The modern coralline sponges are restricted to distinct ecological niches within tropical reefs and subtropical environments and it is possible to distinguish today among three types: a Caribbean community, a Mediterranean community, and a Pacific community (HARTMAN & GOREAU 1970, 1975, VACELET 1967, 1981, 1985, REITNER & ENGESER 1987, a. o.).

There are several coralline sponge communities within Lower Cretaceous reefs and so-called boreal realms which demonstrate a remarkable systematic and ecological similarity with modern communities. The purpose of this paper is to compare selected coralline sponge communities from the Bar-

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Text-Figs. 1-18

All figured specimens without numbers or from private collection are housed in the Institut für Paläontologie der Freien Universität Berlin (IPFU)/88 RE/1.

Text-Fig. 1. *Acanthochaetetes wellsi* HARTMAN & GOREAU 1975.
Text-Fig. 1a. Reef cave specimen with superficial astrorhizal patterns (15 m water depth), Lizard Island Great Barrier Reef, Australia. Scale 0.5 cm.
Text-Fig. 1b. Histological cut of *A. wellsi* (VACELET coll.). a: Choanocyte chambers, b: Archaeocytes, c: Tabula. Scale 1 mm.

Text-Fig. 2. *Vaceletia crypta* (VACELET 1977).
Text-Fig. 2a. Reef cave specimen (15 m water depth), Lizard Island, Great Barrier Reef, Australia. Scale 0.5 cm.
Text-Fig. 2b. Histological cut of *V. crypta* (VACELET coll., org. VACELET 1979); a: Spongocoel, b: Choanocyte chambers, c: newly formed skeletal chamber without any carbonate, d: Aragonitic basal skeleton. Scale 2 mm.

Text-Fig. 3. *Astrosclera willeyana* LISTER 1900. Reef cave specimen with internal astrorhizal patterns (15 m water depth), Lizard Island, Great Barrier Reef, Australia. Scale 0.5 cm.

Text-Fig. 4. *Murrayona phanolepis* KIRKPATRICK 1910. Christmas Island, Pacific Ocean (Scanning Electron Microscope (SEM) photograph).

Text-Fig. 5. *Minchinella lamellosa* KIRKPATRICK 1908.
Text-Fig. 5a. Specimen from the Christmas Island, Pacific Ocean; a: Oscular tubes. Scale 2 mm.
Text-Fig. 5b. Rigid sclere skeleton (acanthose calthrops) of *M. lamellosa;* a: epitactic cements (SEM) (British Museum Nat. Hist., DENDY coll. no. 25 11 1 75).

Text-Fig. 6. *Petrobiona massiliana* VACELET & LEVI 1958.
Text-Fig. 6a. Rocky beach specimen, Isle of Ischia, Mediterranean Sea (SEM).
Text-Fig. 6b. Different sclere types of *P. massiliana* cemented by epitactic cements of diagenetically altered scleres. Reef cave specimen, Marseille, Mediterranean Sea (SEM). Scale 100 µm.

Text-Fig. 7. *Merlia normani* KIRKPATRICK 1908.
Text-Fig. 7a. Rocky shore specimen, Madeira Island, Atlantic Ocean.
Text-Fig. 7b. Histological cut of *M. normani* (VACELET coll.). Classical chaetetid structure. a: Soft tissue with plumose arranged tylostyle megascleres. b: Tabulae with a central opening. Archaeocytes are embedded in older parts of the tubes. Eastern Mediterranean Sea.
Scale 250 µm.

Text-Fig. 8. Aptian pharetronids from the Faringdon "Sponge gravels" (South England); a: *Peronidella* sp., b: *Barroisia anastomans*, c: *Lymnorella* cf. *faringdonensis*. Scale 0.5 cm.

Text-Fig. 9. Cenomanian pharetronids from the Essener Grünsand (Münsterland Basin, West Germany); a: *Thalamnopora cribrosa*, b: *Peronidella* cf. *pistilliformis*, c: *Corynella multiformis*, d: *Lymnorella* sp.

Text-Fig. 10. *Murania lefeldi* SCHOLZ 1984 (holotype).
Text-Fig. 10a. An axinellid calcified demosponge. Late Barremian Schrattenkalk, Allgäu (South Germany). Scale 200 µm.
Text-Fig. 10b. *M. lefeldi* (holotype); detail of the spicular skeleton. Scale 100 µm.

Text-Fig. 11. *Chaetetopsis favrei* DENINGER 1906, a hadromerid calcified demosponge. Late Barremian Schrattenkalk, Gottesacker Wände, Kleines Walsertal (Vorarlberg, Austria). Scale 1 cm.

Text-Fig. 12. *Parastromatopora* sp., an agelasid calcified demosponge. Late Barremian Schrattenkalk, Gottesacker Wände, Kleines Walsertal (Vorarlberg, Austria). Scale 0.5 cm.

Text-Fig. 13. *"Monotrypa"*, an undescribed chaetetid from the late Albian Murguia reef (Northern Spain). Scale 0.5 cm.

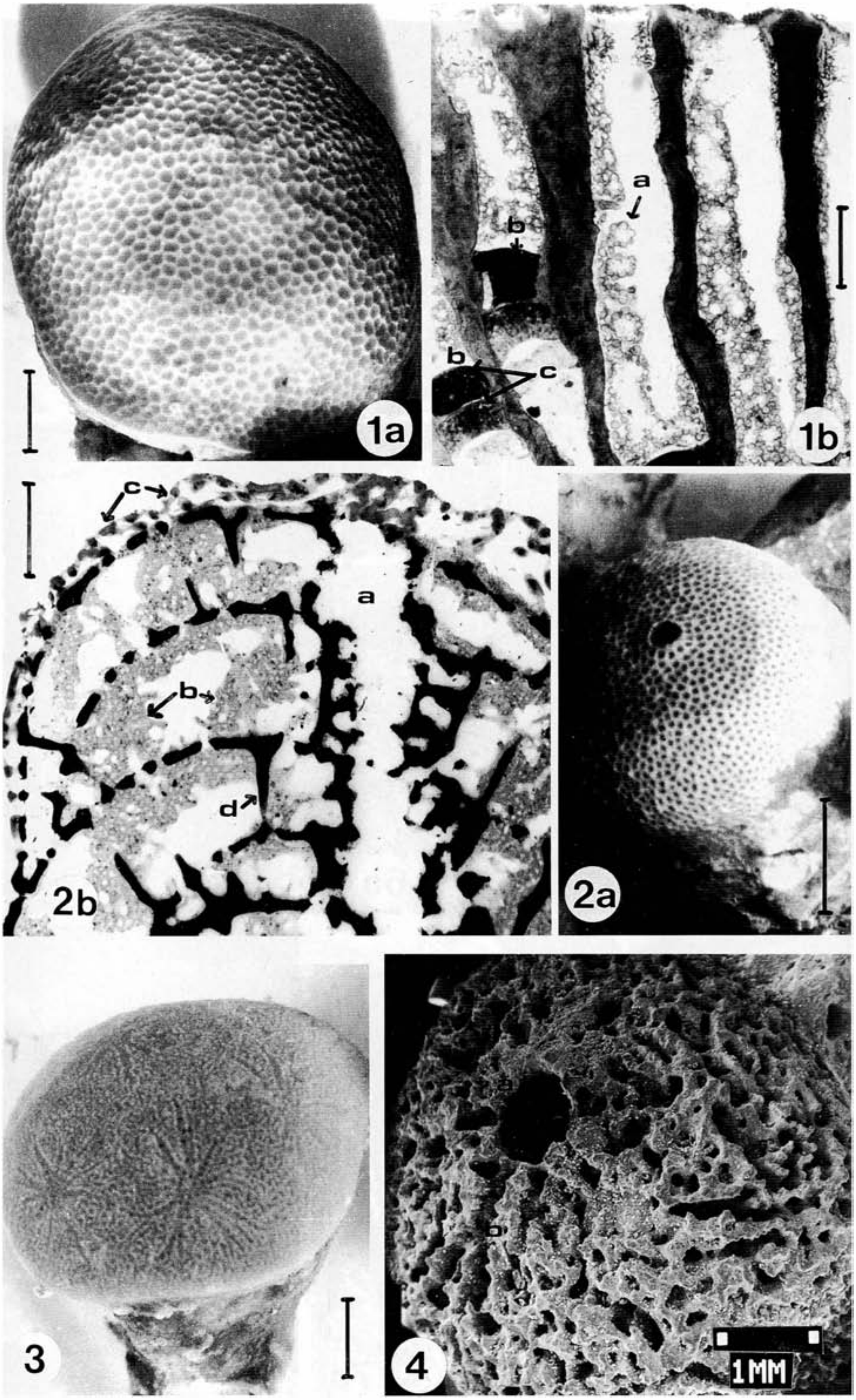
Text-Fig. 14. *Menathalamia caniegoensis* REITNER & ENGESER 1985; *Vaceletia*-type sphinctozoan demosponge from the late Albian Caniego Limestone (Northern Spain). a: Spongocoel. Scale 2 mm.

Text-Fig. 15. *Acanthochaetetes seunesi* FISCHER 1970; a hadromerid chaetetid with close relationship to the extant *A. wellsi*. Late Albian Albeniz/Eguino-Reef (Northern Spain). Scale 0.5 cm.

Text-Fig. 16. *Astrostylopsis* n. sp.; a demosponge stromatoporoid from the late Albian fore-reef of the Albeniz/Eguino Reef (Northern Spain). a: Internal astrorhizal system. Scale 0.5 cm.

Text-Fig. 17. *Actinostromaria cantabrica* SCHNORF-STEINER 1957. A demosponge stromatoporoid from the late Albian platform of Comillas (Northern Spain). Scale 2 mm.

Text-Fig. 18. *Vaceletia*-type demosponge sphinctozoan community of the late Cenomanian hardground of the Liencres-Beach. a: *Vaceletia* n. sp., b: *Astrosclera*-type stromatoporoids. Scale 0.5 cm.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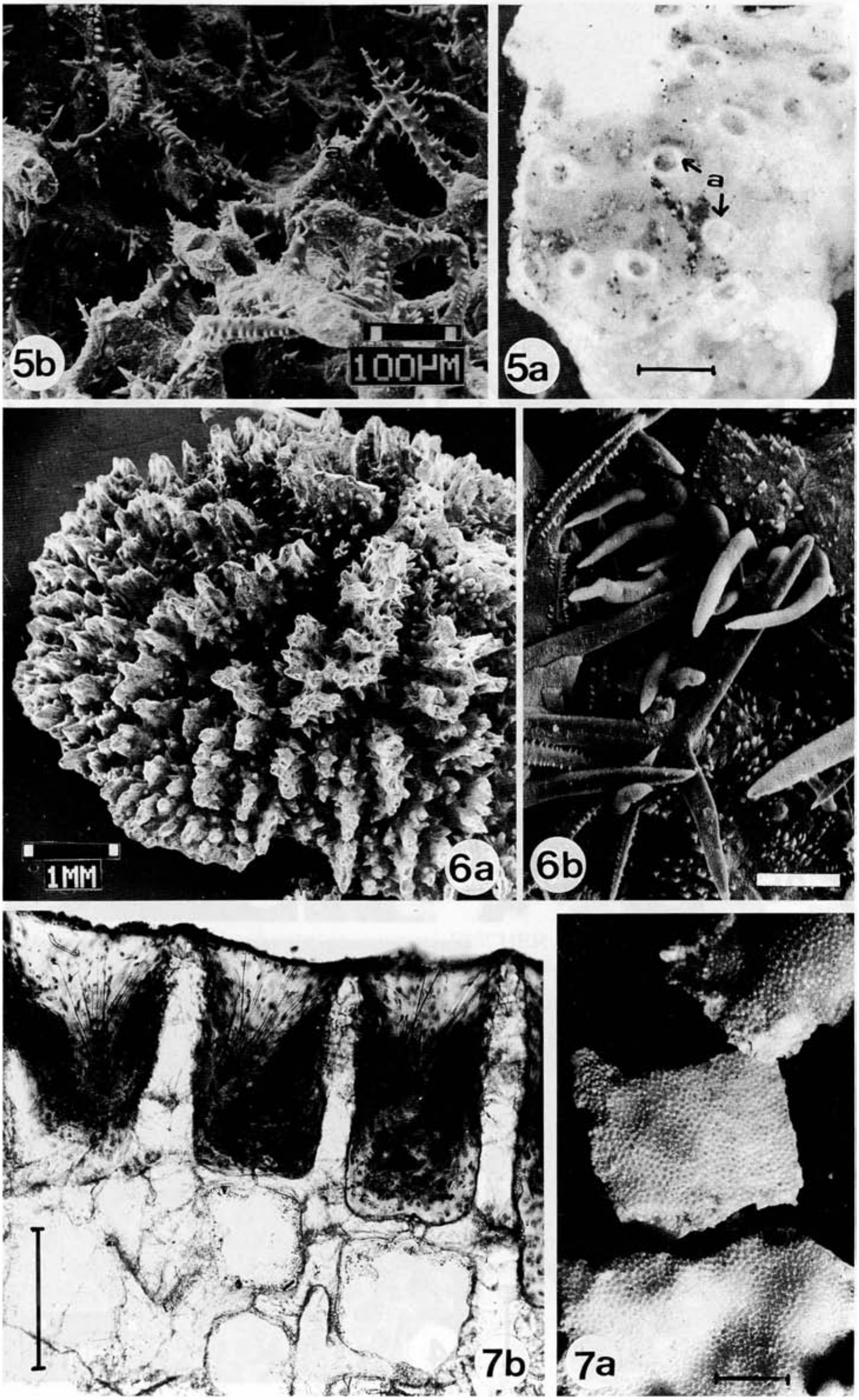
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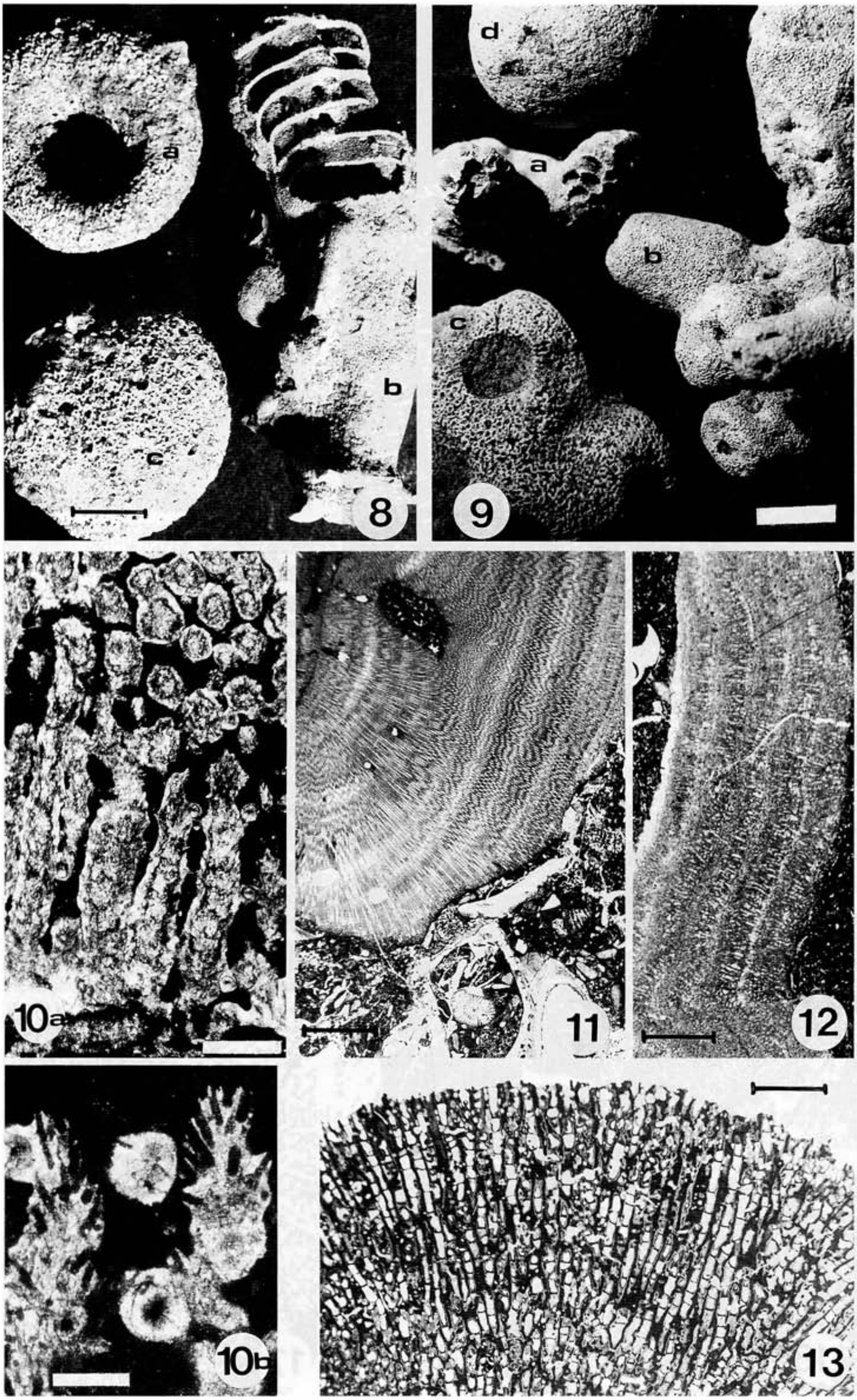
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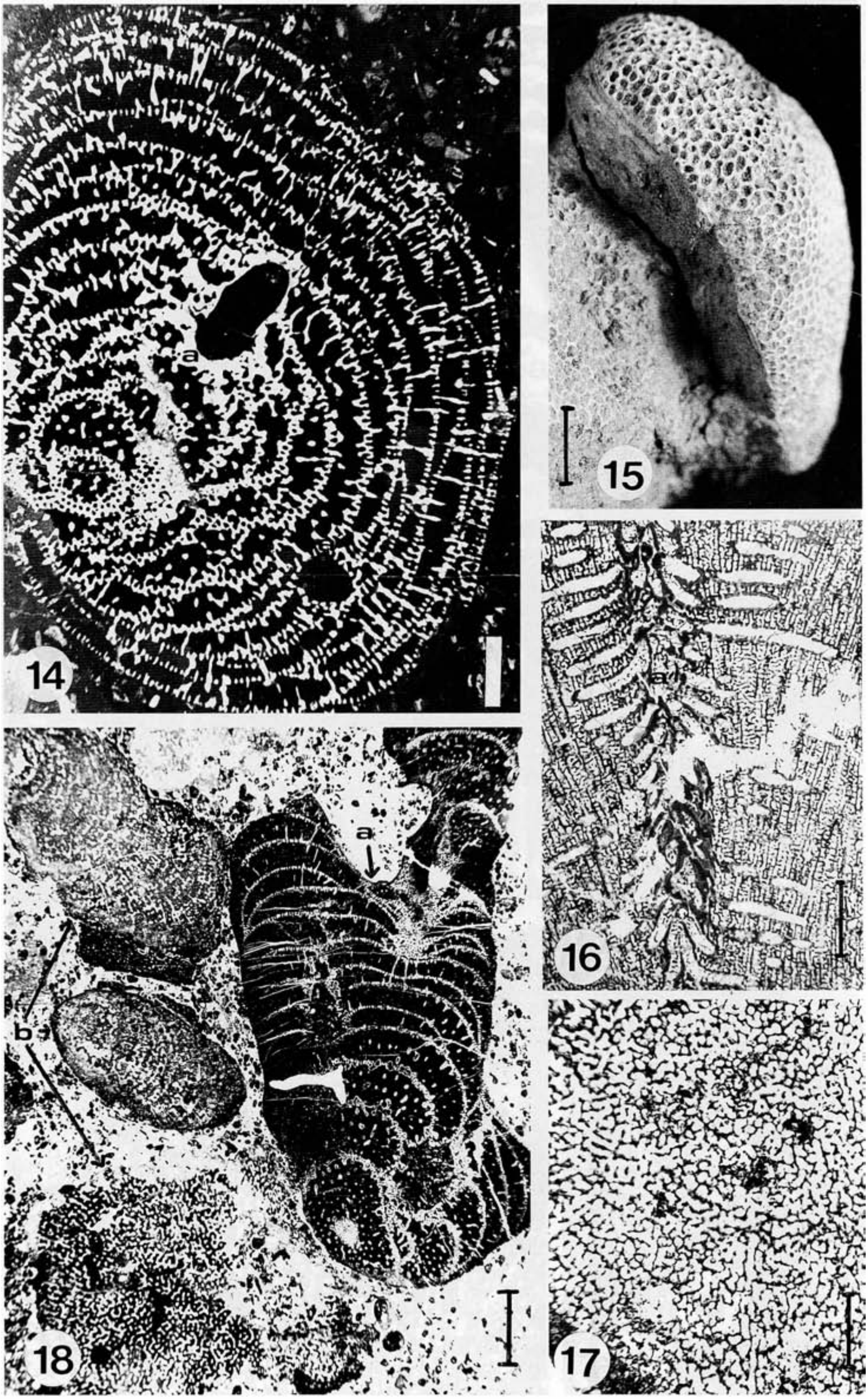

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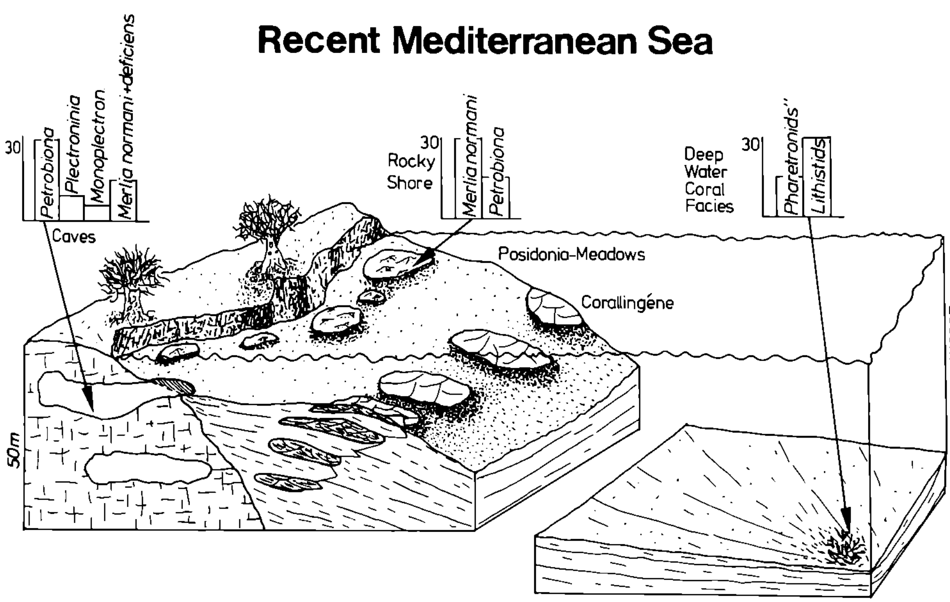
remian, Aptian, Albian, and Cenomanian from the Tethyan/Atlantic Ocean Realm with modern coralline sponge communities from the Mediterranean Sea and the Pacific Ocean.

2. Coralline sponge province of the Mediterranean Sea
(Text-Figs. 6-7, 19)

In the Mediterranean Sea sponges with a rigid calcareous skeleton occur only in dark or dimly light habitats such as in caves or underneath large boulders along a rocky shore. Famous are the caves around Marseille. The fauna and flora of these cryptic habitats were studied by RIEDL (1966), LABOREL & VACELET (1958), PERES (1967). The fixosessile hard-bodied faunas within these caves are dominated by sponges, Hydroidea, Scyphozoa, and octocorals. Minor faunal elements are crustose foraminifera, thecidean brachiopods, and serpulids (RIEDL 1966). The sponges are the most important faunal element; more than 170 species are known (RIEDL 1966).



Text-Fig. 19. Simplified model of the environments of the western Mediterranean Sea. Data based on RIEDL 1966, LABORELL & VACELET 1958, VACELET 1964.
Abbreviations (for all figures): Bryoz. - Bryozoans; Brach. - Brachiopods; Phar., Pharetr., Pharetron. - Pharetronids; Millep., Milleporell. - Milleporellidae; Lithist. - Lithistids; Hex. - Hexactinellids; Strom., Stromatop. - Stromatoporoids; Acanthochaet., Acanthochaetet. - *Acanthochaetetes;* Actinostr. - *Actinostromaria.* Basis of the histograms are individuals collected statistically in outcrops, counted in thinsections, or/and data from literature.

About 80 % of these sponges are demosponges without a rigid skeleton. The rest of this sponge community are Calcarea; however, hexactinellid sponges are not present. The occurrence of coralline Calcarea ("Pharetronids") and some lithistid sponges (Text-Fig. 19) is significant.

2.1 *Petrobiona/Merlia normani* community (Text-Figs. 6-7)

Petrobiona massiliana VACELET & LEVI 1958 (Text-Figs. 6a, b)

The most important and frequent species is *Petrobiona massiliana* (VACELET & LEVI 1958). This species forms thick crusts of a light coloured high Mg-calcite secondary skeleton. The sponge tissue is restricted to the upper surface of the calcareous skeleton (chaetetid grade of organization). Within small fissures of the calcareous body the sponge possesses chambers with archaeocytes (internal gemmulae). This strategy allows the sponge to survive environmental crises (VACELET in press). The archaeocytes regenerate the calcareous basal skeleton. The rigid calcareous body is formed by diagenetically altered calcitic spicules. The genus *Petrobiona* is linked to the order Minchinellida, subclass Calcaronea.

In rare cases, besides *Petrobiona,* the species *Plectroninia hindei* KIRKPATRICK, ssp. *mediterranea* VACELET, and *Monoplectroninia hispida* POULQUEN & VACELET occur. Both coralline Calcarea are minchinellid sponges. In contrast to *Petrobiona* these sponges are extremely small (1-2 mm diameter.

Merlia normani KIRKPATRICK 1908 (Text-Figs. 7a, b)

Coralline demosponges are generally absent from the caves except for one form: the chaetetid poecilosclerid sponge *Merlia normani* KIRKPATRICK. This sponge possesses high Mg-calcite with a crustose chaetetid structure. The sponge has, within the ontogenetically older parts of the skeleton, internal gemmulae comparable with *Petrobiona* (VACELET in press). The genus *Merlia* includes two species: *Merlia normani,* with a calcareous skeleton and *Merlia deficiens* VACELET, without a calcareous skeleton. Both species demonstrate the same type of spicules. Perhaps both species represent different grades of ecological adaptation.

Further types of sponges with a rigid skeleton are known. These are lithistid demosponges *Corallistes masoni* (BOWERBANK), *Discodermia polydiscus* BOCAGE, and *Desmanthus incrustans* (TOPSENT) (POULIQUEN 1971-1972).

Within the Mediterranean Sea these sponges are variously distributed (Text-Fig. 19). *Petrobiona* is restricted to the caves near Marseille and the Isle of Ischia near Naples. *Petrobiona* is an endemic species of the western Mediterranean Sea. *Merlia normani* is mainly found only in the eastern Mediterranean Sea. From the western Mediterranean Sea only the species *M. deficiens* is known.

The coralline sponge community is dominated by coralline Calcarea. Only one coralline demosponge *Merlia normani* is known. Lithistid demosponges are occasionally part of this sponge community. VACELET has observed small pharetronids from deep water occurrences (pers. comm.).

3. Coralline sponge community of the Indopacific Ocean
(Text-Figs. 1-5, 20)

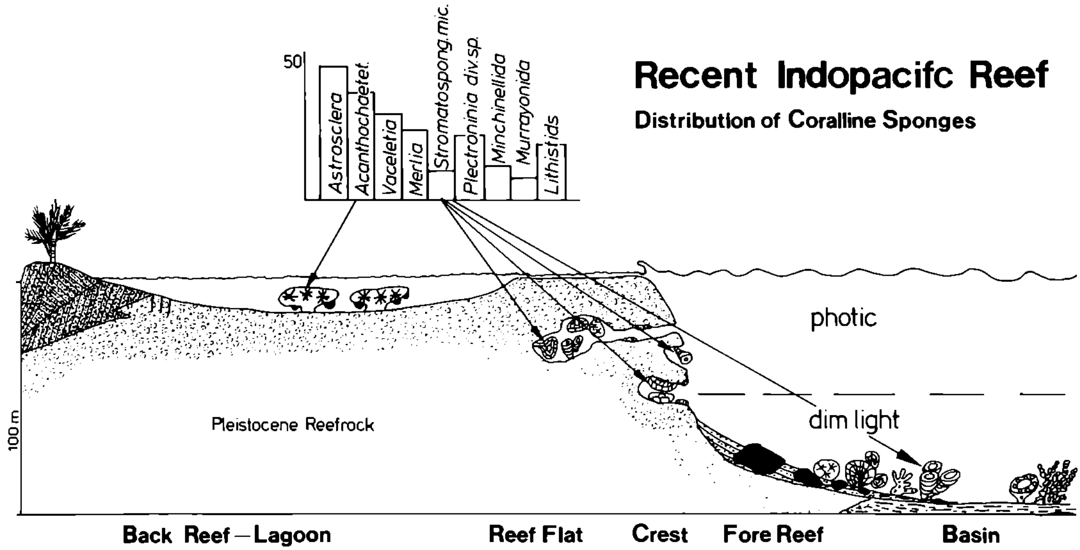
Coralline sponges are an important component of the reef communities of the Indopacific- and Pacific Ocean.

The sponges occur in reef caves, overhangs, or deep fore-reef areas. These aphotic or dimly light areas are comparable with the Mediterranean ones. The coralline sponge community is characterized by a relatively high diversity of coralline demosponges ("Sclerospongia") as well as a high diverse fauna of coralline Calcarea (AYLING 1982, BASILE et al. 1984, DÖDERLEIN 1897, HARTMAN & GOREAU 1975, 1976, KIRKPATRICK 1900, 1908, 1910, 1911, 1912, LISTER 1900, REITNER & ENGESER 1987, VACELET 1967a, b, 1977, 1979, 1981, VASSEUR 1974).

3.1 *Astrosclera/Acanthochaetetes/Vaceletia/Murrayona* community
(Text-Figs. 1-5)

Astrosclera willeyana LISTER 1900 (Text-Fig. 3)

The most common species of this community is *Astrosclera willeyana*. This species possesses a stromatoporoid aragonitic basal skeleton constructed of intracellularly formed spherulites. The aragonitic skeleton shows a clearly developed calcified exhalant system (astrorhize system). The soft tissue covers the ontogenetic younger parts of the calcareous skeleton (stromato-



Text-Fig. 20. Simplified model of a Recent Indopacific reef. Data based on VASSEUR 1974, HARTMAN & GOREAU 1975, 1976, VACELET 1981, AYLING 1982, BASILE et al. 1984, SAUNDERS & THAYER 1987.

poroid grade of soft tissue organization). The rest of the basal skeleton is free of soft tissue. The spicular skeleton is characterized by small acanthostyle megascleres (LISTER 1900, AYLING 1982). Microscleres are not known. *Astrosclera* is linked to the Agelasidae, a common non-calcified demosponge with a heavy spongin skeleton (VACELET 1985). *Astrosclera* is observed in the Red Sea, the Indian Ocean, and the Pacific.

Acanthochaetetes wellsi HARTMAN & GOREAU 1975 (Text-Figs. 1a, b)

Acanthochaetetes wellsi is also a common element of this community. This sponge is characterized by a chaetetid high Mg-calcite skeleton with a unique microlamellar/irregular microstructure (HARTMAN & GOREAU 1975, REITNER & ENGESER 1987). The soft tissue of this sponge covers only the upper surface of the calcitic skeleton (chaetetid grade of soft tissue organization). Within the ontogenetically older part of the chaetetid skeleton, internal gemmulae can be observed (VACELET in press) (Text-Fig. 1b). In rare cases imprints of the exhalant channels are visible (Text-Fig. 1a). The spicular skeleton is constructed of spiraster microscleres and tylostyle megascleres. The spicular skeleton is similar to the genus *Spirastrella,* a non-calcified genus of the demosponge order Hadromerida.

Acanthochaetetes wellsi is restricted to the Pacific Ocean. MORI's species *Tabulospongia japonica* and *T. horiguchii* (MORI 1976, 1977) are synonyms of *Acanthochaetetes wellsi* (REITNER & ENGESER 1987).

Vaceletia crypta (VACELET 1977) (Text-Figs. 2a, b)

Vaceletia crypta is common within some reefs of the Indian Ocean (Madagascar, VACELET 1977) and the Pacific Ocean (New Caledonia, Barrier-Reef o. a., AYLING 1982, BASILE et al. 1984). *Vaceletia crypta* contains a thalamid aragonitic basal skeleton with a prominent spongocoel (Text-Fig. 2a, b). The microstructure is irregular. The soft tissue covers only the ontogenetically younger parts of the calcareous skeleton (sphinctozoid grade of soft tissue organization). Spicules are not known. The anatomy of the soft tissue and the larvae type is similar to ceractinomorph demosponges.

Murrayona phanolepis KIRKPATRICK 1910 (Text-Fig. 4)

BASILE et al. (1984) first reported the coralline Calcarea *Murrayona phanolepis* ("Pharetronida") as part of the *Astrosclera* community. This sponge is a Calcarea with a secondary spherulitic high Mg-calcite skeleton (WENDT 1979, REITNER 1987d). The calcareous spicules are not integrated within the basal skeleton (KIRKPATRICK 1910b); the surface of the sponge is characterized by dermal scales. The sponge possesses a spongocoel with an internal astrorhizal system. The soft tissue shows affinities to the subclass Calcinea (basal choanocyte nuclei).

The ecology of the *Astrosclera* community (Text-Fig. 20) was observed in caves and on overhangs protected from wave action and light in a water depth between 20-35 m. The coralline sponge community is quite often associated with the thecidean brachiopod *Thecidellina congregata*, the terebratulid brachiopod *Frenulina sanguinolenta* (GMELIN), soft body sponges, serpulids, bryozoans, hydrozoans, ahermatypic corals and other fixosessile benthos. Algae are in most cases absent. AYLING (1982) reported a size control by strong water currents; e. g. large specimens of *Astrosclera* occur only in open caves and tunnels with constant, strong water currents.

Acanthochaetetes occurs also in intertidal areas (SAUNDERS & THAYER 1987) in close association with numerous brachiopods.

3.2 Other coralline sponges of Indopacific reefs

Beside the *Astrosclera* community, further types of coralline sponges are known from reef caves, overhangs, and fore-reef belts. The syn- and autecology of these sponges is in most aspects unknown. Most of these sponges are extremely small (1-2 mm diameter) or they live in the distal parts of reef caves which are impossible to reach, even with SCUBA diving.

Stromatospongia micronesica HARTMAN & GOREAU 1976

This coralline demosponge was first reported by HARTMAN & GOREAU (1976) from the western region of the Pacific Ocean. *Stromatospongia micronesica* demonstrates some similarities with the *Ceratoporella/Stromatospongia* community from the Caribbean. *Stromatospongia* is a ceratoporellid sponge with affinities to the non-calcified Agelasidae (VACELET 1985). This species prefers the darkest and most distal parts of reef caves in a water depth of 3-40 m and is not found with other coralline sponges (HARTMAN & GOREAU 1976, VACELET 1981).

Merlia KIRKPATRICK 1908

Both species of the chaetetid sponge, *Merlia normani* and *M. deficiens*, are reported from the underwater caves of the Indian Ocean (VASSEUR 1974) and New Caledonia (VACELET 1981).

Minchinellid Calcarea (Text-Figs. 5a, b)

The minchinellids are calcaronid sponges with the largest diversity within the "Pharetronids". Most of these species are very small (diameter in some millimeters) and they occur on cave walls and within deeper water conditions.

Most of the minchinellids are characterized by a rigid sclere-skeleton formed by tetractines. The tetractines are linked by special cements (VACELET 1981, REITNER 1987d). This rigid sclere-skeleton is comparable with the desma-skeleton of the lithistids. The dermal sclere-skeleton is not rigid and characterized by tuning fork triactines. Most of the species are bearing prominent oscular tubes.

The following species with a rigid skeleton are known from the Indic and Pacific Ocean: *Tulearina stylifera* VACELET, *Monoplectroninia hispida* POULIQUEN & VACELET, *Minchinella lamellosa* KIRKPATRICK (Text-Figs.

5a, b), *Minchinella kirkpatricki* VACELET, *Petrostroma schulzei* DÖDERLEIN, *Plectroninia hindei* KIRKPATRICK, *P. deansii* KIRKP., *P. lepidophora* VACELET, *P. neocaledoniense* VACELET, *P. pulchella* VACELET, *P. tecta* VACELET, *P. radiata* VACELET, *P. tetractinosa* VACELET, *P. microstyla* VACELET, *P. vasseuri* VACELET, *P. minima* VACELET (VACELET 1981). Besides minchinellids with a rigid skeleton there are minchinellids without cemented basal scleres: *Lelapia,* and *Lepidoleucon* (VACELET 1967, 1981).

4. Aptian and Cenomanian "Boreal" coralline Calcarea communities (Text-Figs. 8, 9, 21)

4.1 Geological and sedimentological setting of the Aptian "Sponge-Gravels"

The most famous locality of Aptian "Boreal" pharetronids are the quarries around Faringdon near Oxford (England). The so-called "Sponge-Gravels" of the Lower Greensand are sediments of a transgressive sequence. The Lower Greensand facies is characterized by channels. The shape and direction of these channels was linked to the topography of the pre-Aptian basement (Jurassic) (KRANTZ 1972). The type of channels and the interbedding of interdistributary quiet water mud facies with storm layers and mud pebbles are characteristic of subtidal/tidal-dominated environments. The sponges are parautochthonous and probably grew on hardgrounds of the interdistributary area or on topographic peaks of the Jurassic basement.

4.2 *Barroisia* community (Text-Fig. 8)

The sponge fauna is characterized by Calcarea only. The most typical sponge is the "sphinctozoan" species *Barroisia anastomans.* This species possesses a thalamid skeleton and typical calcaronean scleres (HINDE 1883, RAUF 1913, REITNER 1987d). The most abundant genera are *Rhaphidonema,* a large form with a diameter of more than 3 cm, *Peronidella, Oculospongia,* and *Corynella.* The latter three are pharetronids nearly similar to the thalamid genus *Barroisia* (REITNER 1987d). *Barroisia* and *Peronidella* are probably ancestors of the modern Sycettida (REITNER 1987d). Besides the pharetronids with phylogenetic affinities to the non-rigid Calcaronea there are some species which possess a rigid basal skeleton and oscular system similar to the Recent Order Minchinellida. These types were described firstly by DIGHTON-THOMAS (1971) from the Lower Greensand as *"Actinostromaria" faringdonensis.* The stromatoporoid skeletons and the sclere types are, however, similar to the genus *Lymnorella* (REITNER 1987d).

The whole community is characterized by coralline Calcarea. From the Sponge-Gravels ca. 20 species and 7 genera are known. Calcified demosponges are not present.

4.3 Cenomanian Essener Grünsand *Thalamnopora* community (Text-Fig. 9)

From the Cenomanian "Essener Grünsand" (West Germany), a transgressive sequence similar to that in the Aptian Lower Greensand from Faringdon, a coralline Calcarea community is known (WELTER 1910, DUNIKOWSKI 1883). Recently a diverse sphinctozoan fauna was reported by HILLMER & SENOW-

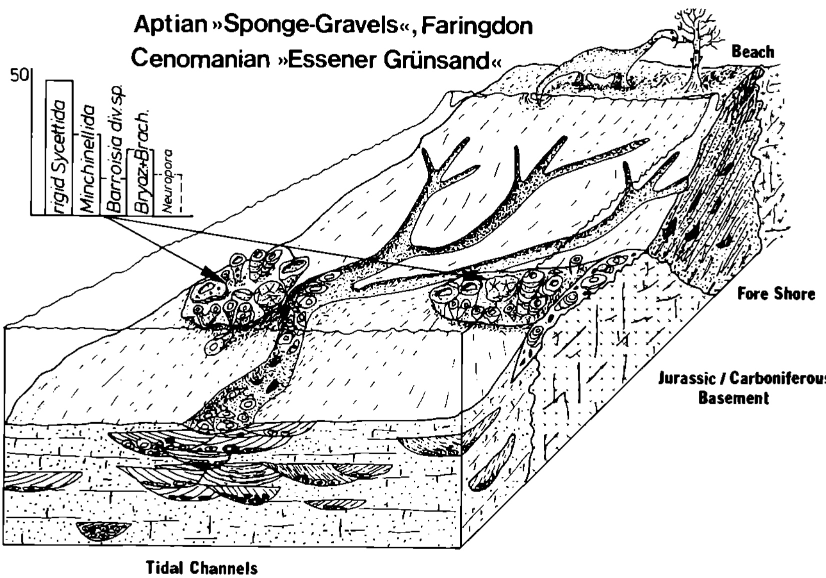
BARI-DARYAN (1986) from the Cenomanian cliffs of Mühlheim-Broich (Germany), including the genera *Thalamnopora, Barroisia, Sphaerocoelia,* and *Celyphia.* Except for *Celyphia,* these sphinctozoans are syscettid Calcarea. *Celyphia* probably does not have a true basal skeleton. This type of "aporate" skeleton is probably a calcified gemmulae cyst.

The main components of the *Thalamnopora* community are minchinellid pharetronids. Important genera are *Porosphaerella, Steinmannella, Sagittularia, Raphidonema, Synopella* etc. and syscettid non-thalamid forms like *Peronidella* and *Corynella.* Calcified demosponges are rare. Only *Neuropora* (*"Crysaora venosa"*) is reported by GOLDFUSS (1826-33) and SIMONOWITSCH (1871). *Neuropora* possesses intramural long style megascleres (KAZMIERCZAK & HILLMER 1974).

A nearly similar community of coralline sponges is reported by REUSS (1845) from the late Cenomanian ("Quader-Sandstein") of Poland.

4.4 Paleoecology and discussion of both Cretaceous communities (Text-Fig. 21)

Elements of the fauna, besides sponges, include cyclostome bryozoans, oysters, rhynchonellid and terebratellid brachiopods, and regular echinoids. Algae are not known. In some cases small pharetronids were observed below big *Raphidonema.*



Text-Fig. 21. Facies model of the Aptian and Cenomanian pharetronid community. Data based on KRANTZ 1972, HILLMER & SENOWBARI-DARYAN 1986 and own observations.

The occurrence of the calcified demosponge *Neuropora* in the Cenomanian outcrops is important. *Neuropora* was first described as a bryozoan (KAZMIERCZAK & HILLMER 1974). *Neuropora* is a common sponge of the so-called "boreal" Lower Cretaceous and is the only known demosponge species of the pharetronid community. This genus is therefore very common within the Hauterivian "Hils-Sandstein", together with pharetronids. This community is comparable with that of the Essener Grünsandstein.

The absence of algae, the sedimentological evidence, and the rest of the fauna indicate a shallow marine subtidal hard substrate lacking in light. The paleoenvironment was probably strongly influenced by tidal currents. The problem is, that the modern pharetronids live only under clear-water conditions. "Reef" and "fore-reef" environments suggested by KRANTZ (1972) and HILLMER & SENOWBARI-DARYAN (1986) do not exist at this time. No reef structures and hermatypic organisms are reported. Some small rare corals, observed within the Essener Grünsand, do not suggest a reefal environment.

Both the Aptian sponge fauna from Faringdon and the Cenomanian Essener Grünsand sponge fauna were probably linked to warm water currents from the Tethys or the proto-Gulf stream which influenced the boreal areas of southern England and northern Germany (MICHAEL 1979). These warm water currents produce temporary subtropical conditions in which some Tethyan benthonics like coralline Calcarea, cyclostomate bryozoa and some corals can survive or have optimal life conditions.

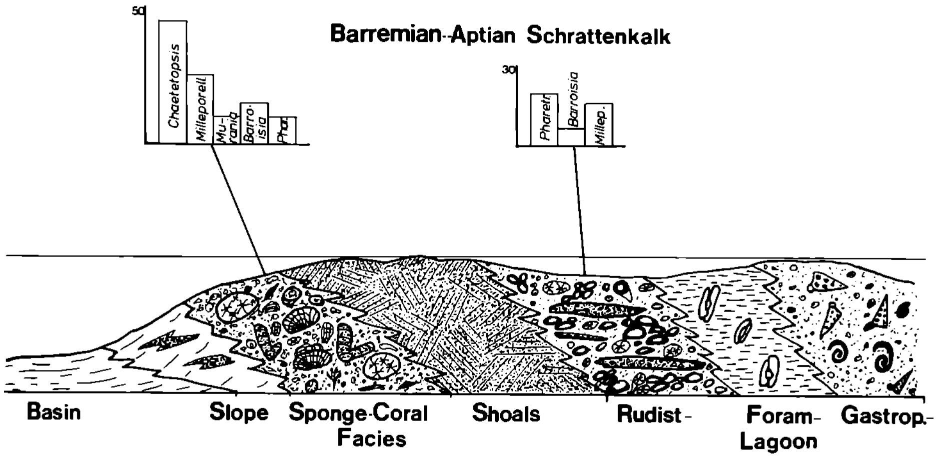
5. Tethyan tropical areas with coralline sponge communities
(Text-Figs. 10-22, 24)

Within the Tethyan-Mediterranean tropical reef belt, different coralline sponge communities from the Lower Cretaceous and Cenomanian are known. Selected were three communities from different times and geographical areas.

5.1 The Barremian/Aptian Helvetic Schrattenkalk *Chaetetopsis* community (Text-Figs. 10-12, 22)

Within the Helvetic Schrattenkalk platform facies small mounds occur with a coralgal-coralline sponge frame. The characteristic sponge is the spicule-bearing chaetetid *Chaetetopsis favrei* (DENINGER 1906, KAZMIERZCAK 1979) (Text-Fig. 11). *Chaetetopsis* possesses intramural simple tylostyle megascleres. The sponge is probably linked with the tetractinomorph demosponges. The *Chaetetopsis* specimens are big and possess an average diameter of 5-8 cm.

Further components of the community are rare milleporellid stromatoporoids (*Parastromatopora* sp.) (Text-Fig. 12), which are also calcified demosponges (WOOD in press). The occurrence of the calcified axinellid sponge *Murania* (Text-Fig. 10a, b) is important. The genus *Murania* was first described by KAZMIERCZAK (1974) (*Murania merbeleri*) from the Barremian Schrattenkalk equivalent in the Tatra Mountains. SCHOLZ (1984) has reported *Murania lefeldi* (Text-Fig. 10a, b) from the late Barremian Schrattenkalk from the Allgäu area. Probably both species are the same. *Murania* is part of a long-ranging calcified demosponge lineage (Carnian-



Text-Fig. 22. Facies model of the Barremian-Aptian Schrattenkalk platform facies. Modified after SALOMON 1987 (compare this volume).

Albian). *Murania* is an encrusting sponge occurring mostly as a thin crust and together with coralline red algae.

Coralline Calcarea are represented mainly by *Barroisia* cf. *cretacea* in this community. This sphinctozoan sycettid sponge is common within the grainstone facies and occurs rarely together with *Chaetetopsis* (compare SALOMON, this volume). Other pharetronids are rare and mostly represented by the genus *Corynella*.

Characteristic for this community is the close relationship with coralline red algae and hermatypic corals (Text-Fig. 22). There is no indication of a cryptic paleo-environment. *Chaetetopsis* is the main framebuilder within the coral mound. TURNSEK & MASSE (1973) describe a comparable community from the Barremian Urgonian of Provence (Southern France), but the diversity of the coralline sponges is much higher. This community also contains a non-cryptic fauna.

5.2 Late Aptian/Albian *Acanthochaetetes* community from the Urgonian reefs of Northern Spain (Text-Figs. 13-17, 23)

The late Aptian/Albian coralline sponge faunas are different from the Barremian/early Aptian ones. During the Middle Aptian (Gargasian), many sedimentological, tectonic, and oceanic events changed different biotas. Important newcomers include several planktonic foraminifera, radiolitid and caprinid rudists, different larger foraminifera, branching coralline red algae, and new coralline sponges (REITNER 1987a).

The most important newcomer of the sponges is the hadromerid chaetetid genus *Acanthochaetetes* (Text-Fig. 15). The Albian species of *Acanthochaetetes* possess tylostyle megascleres and different spiraster microscleres (REITNER & ENGESER 1983, 1987). The Albian genus is very similar to the modern species *Acanthochaetetes wellsi* (see above). It is possible to distinguish three different species: *Acanthochaetetes seunesi* (Text-Fig. 15), *A.*

ramulosus, and *A.* aff. *seunesi*. *A. seunesi* and *A.* aff. *seunesi* are very similar to the extant form (REITNER & ENGESER 1987). *A. ramulosus* is a branching form in contrast to the globular and pyriform *A. seunesi*. *A. ramulosus* preferred photic environments and occurred in "boreal" areas without any reef or carbonate platform influence (REITNER 1987a). All Albian species bear intramural spicules. A common member of this community is the stromatoporoid sponge *Astrostylopsis* which is characterized by prominent inner and outer astrorhizae (Text-Fig. 16) and occasionally mamellons. Spicules are not known. The microstructure is similar to the stromatoporoid genus *Actinostromaria* (Text-Fig. 17), another major member of the community. The most important species is *A. cantabrica* (Text-Fig. 17) (REITNER 1987a, SCHORF-STEINER 1957). Rare stromatoporoids are *Disjectopora* sp. and *Steinerella* sp. (REITNER 1987a). The haplosclerid stromatoporoid *Euzkadiella* is known from lagoonal environments (REITNER 1987c). Chaetetids, beside *Acanthochaetetes*, are rare. They possess a granular micritic microstructure and the shape of the skeleton is similar to the genus *"Monotrypa"* (FISCHER 1970) (Text-Fig. 13).

Sphinctozoan demosponges are rare but the diversity is high. It is possible to distinguish choristid sphinctozoans (*Murguiathalamia, Boikothalamia*) and ceractinomorph sphinctozoans (Verticillitida), which are closely related to the extant genus *Vaceletia* (*Vascothalamia, Menathalamia*) (Text-Fig. 14). *Vascothalamia* bears simple monaxone spicules (REITNER & ENGESER 1985). The axinellid calcified demosponge *Murania* and an undescribed sphaerastermicroscleres bearing coralline sponge are relatively abundant.

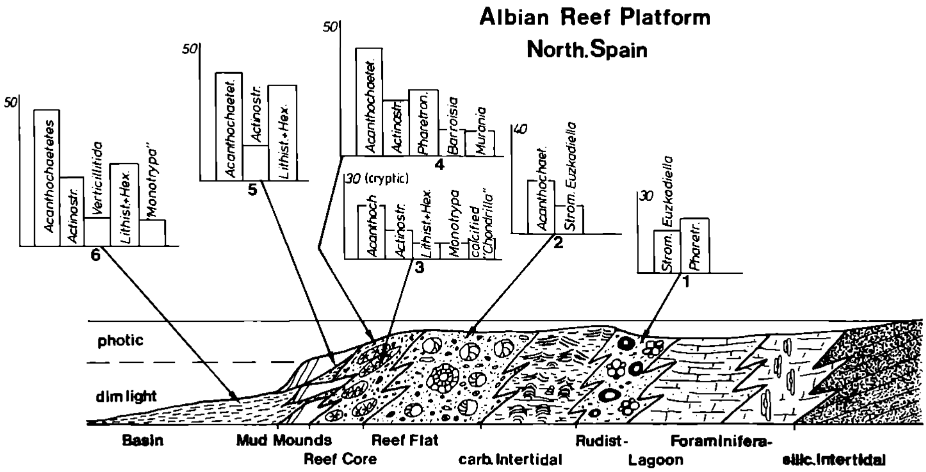
The coralline Calcarea are rare and represented by different species of *Barroisia* (e. g. *B. gandaraensis*) and by several undescribed species of *Corynella* (REITNER 1987a, d).

Different types of lithistid demosponges are common and an important part of this community. Most specimens are linked to the Rhizomorina. Hexactinellids (Lychniscosa and Hexactinosa) are rare and occur only in the deeper water *Acanthochaetetes* community.

5.2.1 Paleoecology (Text-Fig. 23)

The *Acanthochaetetes* community inhabitated six paleoecological niches within the Urgonian reefs:

1. Within the normal marine lagoonal facies belt, the haplosclerid stromatoporoid *Euzkadiella erenoensis* and some pharetronids were sometimes part of the requienid rudist community.
2. *Acanthochaetetes ramulosus* and some stromatoporoids (*Actinostromaria, Steinerella*, disjectoporid stromatoporoids) occurred together with caprinid rudists.
3. Small individuals (ca. 1 cm diameter) occur within small reef caves and fissures. Observed were demospongid sphinctozoans (*Vascothalamia, Murguiathalamia, Menathalamia*), small stromatoporoid sponges of *Actinostromaria* type and disjectoporoids, the chaetetid *"Monotrypa"*, small pharetronid sponges, and lithistid sponges. Beside the sponges crustose foraminifera, bryozoans and serpulids are parts of the hard substrate fauna.
4. Within the photic zone of the reef belt a different community was present, characterized by *Acanthochaetetes ramulosus* and *Barroisia* div. sp. Typical were closely related coralline red algae (*Archaeolithothamnium rude, Agardhielopsis cretacea, Ethelia alba*).



Text-Fig. 23. Facies model of the Albian Albeniz/Eguino reef platform (Northern Spain). Modified after REITNER 1987a.

5. The fifth occurrence was within deeper water mud mounds which are situated at the margins of the reef platforms. Some of the mounds were constructed of crusts of problematic blue-green algae, lithistid demosponges, microsolenid corals and coralline sponges of the *Acanthochaetetes* community (ENGESER et al. 1986, REITNER 1987a, PASCAL 1985).

6. The last occurrence of this community is in the deep fore-reef areas of the Urgonian reef systems. All observed specimens of this facies belt are big (more than 1 cm diameter; average size 5-10 cm). *Acanthochaetetes seunesi* was the dominant sponge. The stromatoporoid sponge *Astrostylopsis* was also a common faunal element. Lithistids were very common and hexactinellids were rare.

5.3 *Acanthochaetetes* community from the Albian Mural Limestone (Arizona)

A unique *Acanthochaetetes* community is known from the early Albian Urgonian Mural Limestone of southern Arizona (USA) (REITNER et al., in press). The Mural Limestone is a vast rudist platform with characteristic patch reef mounds (SCOTT 1979). The *Acanthochaetetes* community is characterized by two species of *Acanthochaetetes* and is quite common within the *Microsolena* biofacies.

Closely related with this chaetetid are pharetronids (*Lymnorella, Corynella*) (RIGBY & SCOTT 1981). Lithistids are rare and no other coralline sponges are observed. This is the first occurrence of *Acanthochaetetes* within the Gulf of Mexico area.

WELLS (1934) has reported a *Vaceletia*-type sphinctozoan from the Cenomanian Buda-Limestone from Central Texas.

5.4 *Acanthochaetetes* from the Peoples Republic of China

DENG (1982) has reported from the Lower Cretaceous (Albian) of Xizang an *Acanthochaetetes* specimen. This is the first occurrence of the genus *Acanthochaetetes* in the Far East and relatively close to the modern Pacific Ocean.

5.5 *Acanthochaetetes ramulosus* of non-reef influenced "boreal" environments

Acanthochaetetes ramulosus is reported from shallow marine non-cryptic Cenomanian strata from northern France (FISCHER 1970) and southern England (HART & JOHNSON 1984). *Acanthochaetetes ramulosus* is also known from the non-cryptic paleoenvironments of the Albian reefs of northern Spain. Also FISCHER & VOIGT (1978) have reported from the Maastrichtian of Maastricht acanthochaetetids associated with warm water organisms, e. g. larger foraminifera (*Orbitoides*).

6. Late Cenomanian Liencres rudist ramp
(Text-Figs. 18, 24)

6.1 Geological and sedimentological setting

An important outcrop containing a Cenomanian coralline sponge community is the rocky beach at Liencres. The Middle and late Cenomanian strata on the northern part of the Toñanes syncline are well-bedded and interfinger with grey marls (REITNER 1987a). The microfacies of the Cenomanian beds are grainstones, packstones, and wackestones. The whole facies is characterized by numerous hardgrounds with limonitic impregnations and oncoliths. An important unit is the terminal hardground sequence in which several reworking events are recognizable, probably caused by storm-events. Some beds demonstrate hummocky cross stratification with symmetric ripples on the top of the beds. Most of the sponges are preserved within a limonitic matrix. In some cases the sphinctozoan sponges demonstrate spicule preservation (REITNER, in press). The autochthonous organisms are dominated by different coralline sponges, brachiopods, bryozoans, and serpulids. This terminal hardground is covered by Turonian hemipelagic marls.

6.2 *Vaceletia*/*Acanthochaetetes* community (Text-Fig. 18)

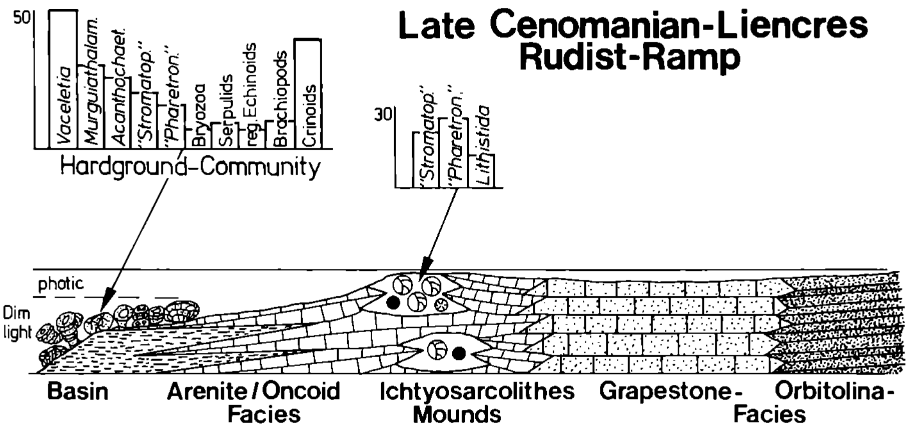
The fauna of the terminal hardground is characterized by different demosponge sphinctozoans with a trabecular basal skeleton. Very important is the abundant occurrence of an undescribed species of *Vaceletia,* which is the oldest known report of this species. Besides the ceractinomorph sphinctozoans, undescribed tetractinomorph sphinctozoans of Murguiathalamida are common.

The hadromerid *Acanthochaetetes* cf. *seunesi* is an abundant member of the community. The Cenomanian Acanthochaetetidae do not have any intramural spicules, just like the extant species *Ac. wellsi*.

Quite common are coralline sponges with a stromatoporoid skeleton. No skeletal microstructures are preserved due to the intense diagenesis. The skeletal arrangement of most of the stromatoporoids is comparable with the modern genus *Astrosclera*. But any relationship is as yet uncertain. Pharetronids are rare and lithistids are not observed.

6.3 **Paleoecology and discussion** (Text-Fig. 24)

This community is unique in the world. It is characterized by typical hard substrate organisms and no highly developed algae (e. g. coralline red algae) or corals are observed. The facies distribution, the microfacies, and the sedimentological analysis indicate a distal deeper water ramp facies, influenced by storms.



Text-Fig. 24. Facies model of the late Cenomanian Liencres rudist ramp. Modified after REITNER 1978a.

This conclusion is supported by the absence of algae. The common observed limonitic coated grains are probably not formed by algae. The hardgrounds indicate permanent bottom currents; the sponges preferred to grow in protected niches of the hardground surface. This community is different from the Albian ones and much more comparable with modern deeper water coralline sponges from the Pacific Ocean.

7. Palaeobiogeography

7.1 Recent distribution of coralline sponges

In the modern oceans, three communities of coralline sponges are known and distributed in the Atlantic Ocean, the Indopacific Ocean, and the Mediterranean Sea.

The pharetronids of the Indopacific and the Mediterranean Sea are more or less similar. Only *Petrobiona* is an endemic species. Pharetronids are not known from the Atlantic Ocean.

Most of the modern calcified demosponges from the Indopacific Ocean are also relics from the Cretaceous Tethyan community. The long-lived genera are *Acanthochaetetes, Vaceletia,* and probably *Astrosclera.* The occurrence of *Astrosclera* from the Cretaceous strata is doubtful, but *Astrosclera*-type coralline sponges are known from late Permian and Triassic outcrops from the Tethyan Realm.

There are some differences in the distribution and occurrences of this community of sponges. *Acanthochaetetes* and *Stromatospongia micronesica* occur only in the Pacific Ocean.

There are no fossil relatives of ceratoporellid/agelasid *Stromatospongia micronesica* and the poecilosclerid *Merlia. Stromatospongia micronesica* is an endemic species of the Indopacific. Both genera occur within the tropical realm of the Atlantic and Indopacific Ocean. *Merlia normani* occurs also in the Mediterranean Realm and the Isle of Madeira in subtropical environments (KIRKPATRICK 1911). From boreal environments no coralline sponges are known.

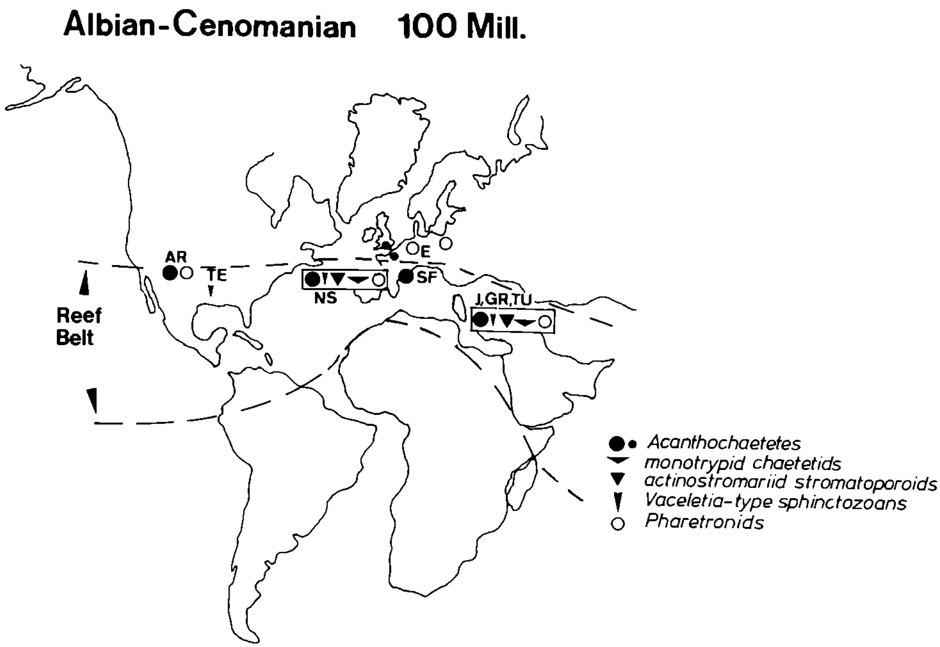
The Caribbean coralline sponge community is different and characterized by ceratoporellid sponges and demonstrates, except for the genera *Merlia* and *Stromatospongia,* no coincidence with the Indopacific/Mediterranean Tethyan relic community.

7.2 The Albian-Cenomanian palaeobiogeography of the pharetronid and the *Acanthochaetetes* communities (Text-Fig. 25)

7.2.1 Regional distribution of the *Acanthochaetetes* communities

During the Aptian and Cenomanian four types of coralline sponge communities are observed. Most of these types of sponges are restricted to the tropical reef belt and linked with coralgal reefs. The European *Acanthochaetetes* reef community appears in the late Aptian (Clanseyian) and is up to now only known from northern Spain (REITNER 1987a, PASCAL 1985). There are some reports of Acanthochaetetidae from the late Jurassic (ZUFFARDI COMERCI 1926), but the stratigraphical evidence is uncertain. The Gargasian events are probably responsible for this newly evolved community. From the early Albian of the western part of the Gulf of Mexico Realm (Mural Limestone) (Arizona) an *Acanthochaetetes* community is known, but the total diversity of the community is much smaller than the European ones.

Within the central European Tethyan Realm, a widespread distribution is known from the mid-Cretaceous. Besides the northern Spain occurrences the *Acanthochaetetes* reef community is found in southern France, Jugoslavia, Greece, Turkey, and China (Xizang). The non-reefal acanthochaetetids are known from the Cenomanian of northern France and southern England.



Text-Fig. 25. Paleogeographic distribution of Albian/Cenomanian coralline sponge communities. Map based on SMITH & BRIDEN (1977).
Abbrev.: AR - Arizona; E - Essener Grünsand; Gr. - Greece; J - Jugoslavia; NS - Northern Spain; TE- Texas; TU - Turkey; CHX- China Xizang.

7.2.2 Distribution of pharetronid communities

Pure pharetronid communities occur only within the northern part of Europe. Famous are the Aptian Sponge Gravels of Faringdon and the Cenomanian Essener Grünsand communities. Less important outcrops are known from the Aptian of Normandy and the Cenomanian of Poland (MICHAEL 1979, REUSS 1845).

8. Conclusions

a. During the Aptian and Cenomanian several regional communities of shallow marine coralline sponges are known. It is possible to distinguish a Tethyan reef community with calcified demosponges and pharetronids and a temperate ("Boreal") community of pharetronids only in the north central parts of Europe.
b. Within the Barremian/Aptian Tethyan reef platforms the chaetetid *Chaetetopsis* and milleporellid stromatoporoids are the dominant coralline sponges. These communities are comparable with the late Jurassic ones (Oxfordian-Kimmeridgian).

c. From the "Boreal" Aptian of South England (Faringdon) only coralline Calcarea are known. The paleoclimate was comparable with the modern Mediterranean.
d. In the late Aptian the Tethyan *Acanthochaetetes* reef community occurs. Most of the Barremian and Aptian coralline sponges disappeared due to the Middle Aptian (Gargasian) Events.
e. The genus *Acanthochaetetes* is reported for the first time from the western Cretaceous Gulf of Mexico Realm (Arizona).
f. *Acanthochaetetes ramulosus* occurs also in non-reefal Temperate ("Boreal") environments.
g. During the Albian and Cenomanian, the *Acanthochaetetes* community is widespread within the central Tethyan Realm.
h. From the Cenomanian a pharetronid (*Thalamnopora*) community is known from Essener Grünsand. Only one calcified demosponge (*Neuropora*) is known, a characteristic "Boreal" coralline sponge.
i. The modern distribution patterns of coralline sponges are comparable with the mid-Cretaceous ones, except for the Caribbean community. It is possible to identify a Subtropical Mediterranean pharetronid community with only one calcified demosponge (*Merlia normani*). This community is comparable with the Cenomanian *Thalamnopora* community.
j. The modern Indopacific coralline sponge communities demonstrate a strong coincidence with the Cretaceous Tethyan *Acanthochaetetes* community. The extant genera and the ecological niches are the same as the fossil ones.
k. The Mediterranean and Indopacific coralline sponges are survivors from the Cretaceous Tethyan Ocean.

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Facies and Composite Biostratigraphy of Late Cretaceous Strata from Northeast Egypt

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With 53 Text-Figures and 1 Table

KUSS, J. & MALCHUS, N. (1989): Facies and Composite Biostratigraphy of Late Cretaceous Strata from Northeast Egypt. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 879-910. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Late Cretaceous sediments of NE Egypt were investigated with respect to lithofacies, microfacies and biostratigraphy. The late Cretaceous marine strata were subdivided into five major cycles: the late Cenomanian/early Turonian cycle with carbonates and siliciclastics - the late Turonian cycle with predominantly sandstones and few marly limestones - the middle/late Coniacian cycle is dominated by limestones, marls and sandstones - the Campanian/Maastrichtian cycle with phosphates, claystones and chalks - a local early/late Maastrichtian cycle is characterized by limestones and sandstones.

The lithofacies of these shelf-deposits show many lateral and vertical changes due to non-sedimentation and synsedimentary structural movements; the interfingering of continental sandstones with marine sandstones, claystones and limestones is common in most strata.

Different microfacies types (including a characteristic microfauna and -flora) are described from limestones of the late Cenomanian/early Turonian sequence. The late Turonian/Campanian strata contain only minor amounts of carbonates, while an isolated Maastrichtian sequence yields shallow-water and deeper shelf limestones and sandstones.

A regional oyster biostratigraphy was established, based on ammonoids, foraminifera and lithostratigraphic correlations, which allows a supplementary subdivision of the nearshore and shallow-marine strata. Differences of lithology and microfacies in the area of Wadi Qena and the Galala Plateaus enabled to distinguish a northern and a southern facies province; they coincide with differences in the macrofaunal communities: a Wadi Qena faunal province in the south and a Galala faunal province in the north is proposed.

Kurzfassung: Es wurden oberkretazische Sedimente Nordostägyptens lithofaziell, mikrofaziell und biostratigraphisch untersucht. Die kretazische Abfolge beginnt mit überwiegend kontinentalen Sandsteinen (Unterkreide/Mittelcenoman). Die oberkretazische marine Abfolge wurde in fünf übergeordnete Zyklen unterteilt: der Obercenoman/Unterturon-Zyklus wird aus Kalken und Klastika aufgebaut, der Oberturon-Zyklus überwiegend aus Sandsteinen und wenigen mergeligen Kalken, der Mittel/Oberconiac-Zyklus besteht aus Sandsteinen und Mergeln, der Campan/Maastricht-Zyklus aus Phosphaten, Tonen und Kreidekalken, und ein lokal bedeutsamer Unter/Obermaastricht-Zyklus setzt sich aus Kalken und Sandsteinen zusammen.

Diese Schelfsedimente zeigen laterale und vertikale Fazieswechsel, die auf Schichtlücken und synsedimentäre strukturelle Bewegungen zurückzuführen sind. In den meisten Abfolgen ist eine Verzahnung von kontinentalen und marinen Sandsteinen, Tonsteinen und Karbonaten ausgebildet.

Verschiedene Mikrofazies-Typen (einschließlich einer charakteristischen Mikrofauna und -flora) werden von den Obercenoman/Unterturon-Ablagerungen beschrieben. In den Oberturon/Campan-Ablagerungen sind nur untergeordnet Kalke enthalten. Durch mikrofazielle Untersuchungen einer isolierten Maastrichtabfolge mit Flachwasserkalken, tieferen Schelfkarbonaten und Sandsteinen wurden synsedimentäre tektonische Bewegungen nachgewiesen.

Eine regionale Austern-Biostratigraphie wird vorgestellt, die mit Ammoniten und Foraminiferen geeicht wurde und eine detaillierte Unterteilung und Korrelation der küstennahen und flachmarinen Ablagerungen gestattet.

Die lithologischen und mikrofaziellen Unterschiede der Profile des Wadi Qena (Süden) und der Galala-Plateaus (Norden) zeigen unterschiedliche Ablagerungsbereiche einer südlichen und nördlichen Faziesprovinz an, die sich auch in den Faunenvergesellschaftungen unterscheiden. Sie werden hier als Wadi Qena- und Galala-Faunenprovinzen bezeichnet.

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1. Introduction

During the last four years, Cretaceous strata of the Eastern Desert of Egypt were reinvestigated concerning lithology, microfacies and fauna (BANDEL et al. 1987, BANDEL & KUSS 1987, HENDRIKS et al. 1987, KUSS 1986a, 1986b, 1987, 1989). The aim of the present paper is to describe the most important paleoenvironments of the Cretaceous sequence based on sedimentologic features and microfacies analysis, and to introduce an oyster-biostratigraphy, calibrated with ammonoids, benthic and planktonic foraminifera. The results are based on field work realized together with K. BANDEL during 1984 in Wadi Qena and during 1986 along the two Galala plateaus.

Cretaceous deposits are widely distributed in NE Egypt and are composed of a great variety of rock types (Text-Figs. 3-5). They overlie Precambrian basement rocks (Red Sea Crystalline) as well as sediments of different Phanerozoic ages (Cambrian to Jurassic). The early Cretaceous cycle is dominated by sandstones. The late Cretaceous strata were subdivided into five major cycles (KUSS 1989):

a. The late Cenomanian/early Turonian cycle with mixed siliciclastic limestones;
b. The late Turonian cycle, dominated by sandstones with subordinate marly limestones;
c. The middle/late Coniacian cycle, composed of sandstones and marls;
d. The Campanian/Maastrichtian cycle including phosphates, shales and chalks;
e. The Maastrichtian cycle, characterized by limestones, dolomites and sandstones, which is of local importance.

In general, deposition took place on a broad shelf system that deepened from south to north with an extension of more than 400 km in N-S direction (KUSS 1987). Its sedimentary environments are characterized by the interfingering of shallow-marine facies with continental facies. Though pelagic organisms found in several subunits suggesting significant depths, the study area was mainly characterized by shallow shelf environments with several intra-shelf basins. Synsedimentary folding (in the north), block-faulting, changes in sea-level and episodic influx of fresh water formed different genetic and diagenetic environments, resulting in a great variability of litho- and biofacies types (chapt. 2).

The microfacies description of the carbonates is generalized - based on all collected data from figured sections (Text-Figs. 3-5) - and is compared with sections of adjacent areas (Sinai). The microfacies concept was used for the limestone-dominated sequences of late Cenomanian/early Turonian, Coniacian and Maastrichtian. No thin section analysis was carried out for the other sequences (chapt. 3).

Together with WIEDMANN (pers. comm.) and LUGER (pers. comm.), a zonation of the late Cenomanian/late Campanian strata was elaborated, based on ammonite assemblage zones A-K (reference sections M/N: middle Wadi Qena, and section O: southern Wadi Qena; both see Text-Fig. 5) and was used to establish a regional oyster biostratigraphy. A comprehensive systematic description of the Cretaceous oysters from Egypt was given by MALCHUS (in press). Maastrichtian sediments were separated and subdivided with the help of benthic and planktonic foraminifera (KUSS 1986b) and oysters (reference section X; Text-Fig. 5).

Macrofaunal assemblages, facies and microfacies in Wadi Qena and the Galala plateaus differ significantly. Therefore, a southern and northern facies province were differentiated, coinciding with the here proposed Wadi Qena faunal province and Galala faunal province.

The sections presented in Text-Figs. 3, 4, and 5 are chosen as representative sections of the described area. Sections 9/2, 13/2, 15/2, and 18/2 (for location see Text-Fig. 1) were used additionally for description. Two of the sections are composite: W/U/V (Text-Fig. 3) and the section "south and north Galala" (Text-Fig. 4). The latter is composed of sections 15/2 (above), 13/2 (middle), and 20/2 (below).

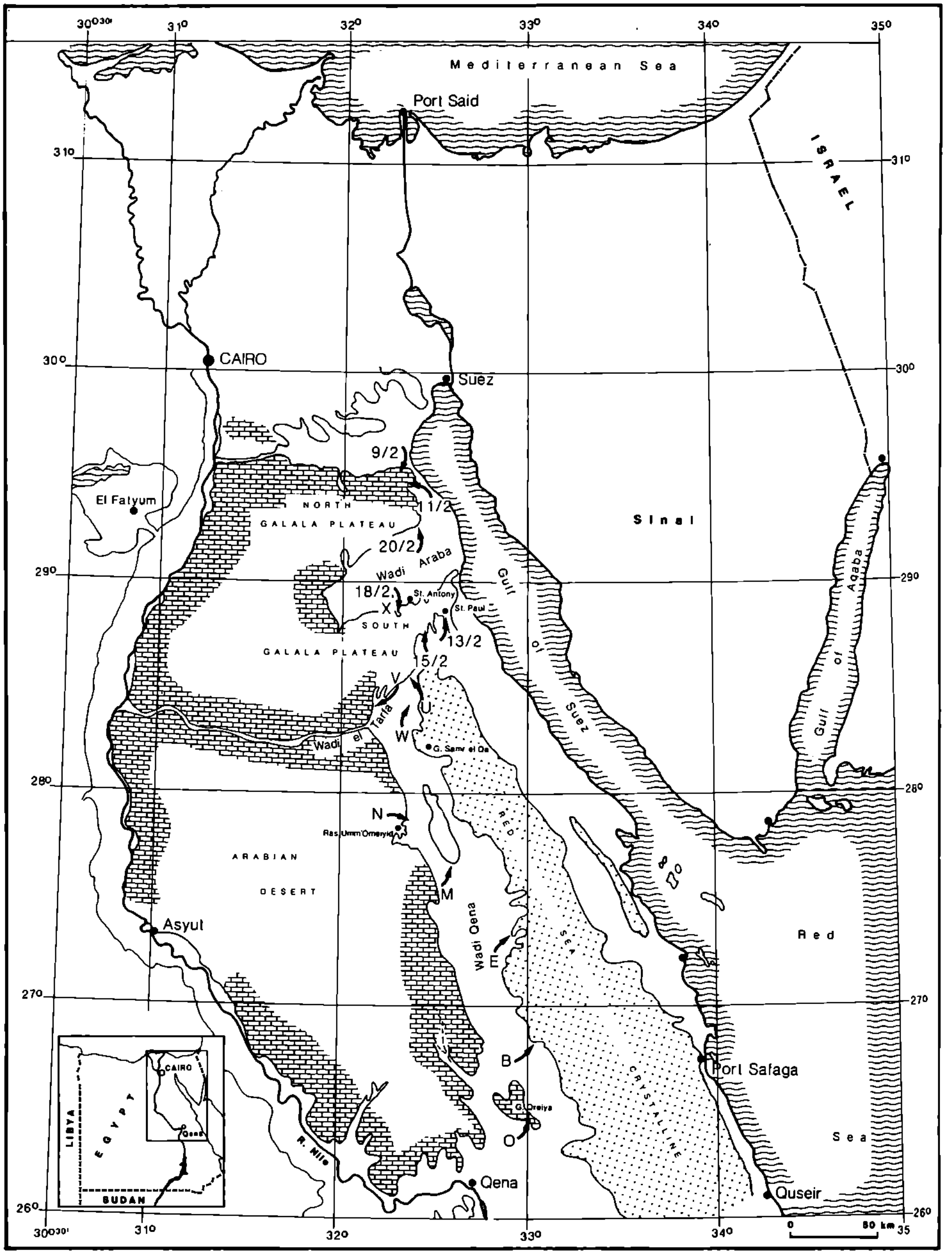
2. Description of the Cretaceous lithotypes (macrofacies)

2.1 Early Cretaceous/Middle Cenomanian cycle

During Neocomian/middle Cenomanian times thick continental sandstone and siltstones accumulated in most areas of northern Egypt (Wadi Qena and the Galala Plateaus included), due to emergence of the shelf areas. Earliest marine sediments are of late Aptian/early Cenomanian age. They are confined to the northern Sinai and to subsurface equivalents in the northern part of the Western Desert (KUSS & SCHLAGINTWEIT 1988). Further to the south Aptian sediments were formed at the transitional zone between marine and fluviatile facies (Abu Ballas Formation of BÖTTCHER 1985), indicating a southward prograding sea. Southward transgressive units were found in all younger Cretaceous strata, as indicated by facies patterns.

2.2 Late Cenomanian/early Turonian cycle (zones A-E)

The late Cenomanian/early Turonian strata rest either on basement rocks or sandstones of Carboniferous to Middle Cenomanian age and are overlain by



Text-Fig. 1. Location map of the investigated area in the Eastern Desert of Northern Egypt. Sections were taken betwen North Galala Plateau (in the north) and southern Wadi Qena (in the south). Brick-signature indicates the Tertiary limestone plateaus.

Coniacian limestones and dolomites as well as by Campanian chalks. The sequence can be subdivided into five biostratigraphic units A-E (chapt. 4), which are characterized by a rapid change of deposition, non-deposition and erosion. The sediments are formed by fluviatile and near-shore sandstones, marls, limestones and dolomites. They are gradually decreasing in thickness from 200 m in the northern Galala Plateau to 55 m in the southern Wadi Qena. The thickness and the content of limestone beds increases from south to north. Variations in thickness and composition of the strata resulted from undulations of the seafloor, terrigenous influx and synsedimentary tectonic movements.

In many sections of the Wadi Qena facies province (e. g. section B; Text-Fig. 1) the late Cenomanian/early Turonian cycle starts with a transgressive sequence. Basal conglomerates are overlain by littoral sandstones and fine-grained, often bioturbated silts with intercalations of sandy/limy oyster beds. These beds are topped by marls and marly limestones with echinoids, ammonoids and gastropods. Quickly deposited channel-sandstones (formed by erosion of the crystalline Red Sea Hills and the older Nubian strata) often interrupt the sequence.

The Wadi Qena facies province is characterized by the dominance of siliciclastic deposits, due to the input of continental erosives. In sections of the Galala facies province (further north) limestones, dolomites and marly/silty shales predominate, indicating an offshore environment.

2.3 Late Turonian cycle (zone F)

Late Turonian strata are dominated by glauconitic sandstones and siltstones, with minor amounts of marly limestones. Most siliciclastic deposits of this cycle are characterized by varying thicknesses, due to nearshore deposition in channels (BANDEL et al. 1987). Their major occurrences were found in the northern Wadi Qena.

2.4 Middle/late Coniacian cycle (zone G)

In northern Wadi Qena, deposition of massive channel-sandstones continues to Coniacian times. Though the Coniacian transgression reached far to the south in Egypt (documented by inocerams found near Aswan; KLITZSCH 1986), in the Eastern Desert deposits of this age are known only from isolated occurrences. They are represented by sandstones, marls and few limestones. In contrast, LEWY (1975) described open marine carbonates from the Sinai.

Due to the north/northwestward drift of the African Plate, tectonic movements intensified in late Turonian/early Santonian times (SCHANDELMEIER et al. 1987). These movements caused folding (in the north) and block-faulting (further south). They resulted in erosion or non-deposition on uplifted areas, whereas deposition took place in down-faulted blocks.

2.5 Campanian (zones H-K)/Maastrichtian cycle

The Campanian/Maastrichtian strata in the Eastern Desert are either formed by siltstones, sandstones, shales and phosphates, or by chalks and marls.

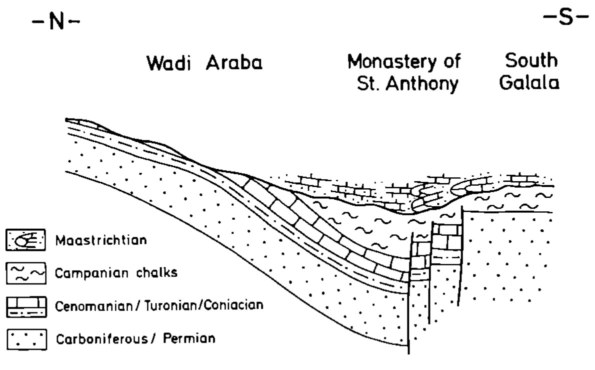
The enrichment of late Campanian/early Maastrichtian phosphates is connected to synsedimentary tectonic movements (BANDEL et al. 1987): Dissolved phosphates (concentrated in the sea) were enriched in phosphorite sands, when reworking and winnowing of fine particles took place. The thick layers of the phosphate formation (Abu Had Formation; BANDEL et al. 1987) from section O and some neighbouring localities of the southern Wadi Qena are parts of a broad E-W running phosphate facies-belt (GERMANN et al. 1984). These strata were not found in the Galala area; thin shales and few cherts are the possible equivalents here.

The well-bedded or massive chalks (St. Paul Formation: BANDEL et al. 1987 in the South Galala; Sudr Chalk in Sinai) overlie truncated beds of late Cenomanian to late Coniacian age as well as phosphates of the Abu Had Formation. Oysters of *Pycnodonte vesiculare* (SOW.) are frequent in layers near the base of the chalk sequence (chapt. 4.2). This chalk sedimentation persisted locally to late Maastrichtian times (ABDELMALIK et al. 1978). At the monastery of St. Anthony the chalks were truncated by Lower Maastrichtian sandstones and limestones.

2.6 Maastrichtian cycle

The occurrence of early to middle Maastrichtian limestones is restricted to a 5-7 km broad, E-W running outcrop at the northern slopes of the Southern Galala Plateau (near monastery of St. Anthony, section X; Text-Fig. 1). This unique sequence of shallow-water limestones, mixed with deeper shelf carbonates, marls and sandstones (KUSS 1986b) could not be compared with any neighbouring outcrop, neither in the western parts of Egypt, nor in Sinai or Israel. BANDEL & KUSS (1987) named this 190 m thick sequence St. Anthony Formation, comprising early to late Maastrichtian strata.

As a consequence of faulted-block tectonics, the shallow-marine sediments of this formation slid northwards. This results in both slumping-



Text-Fig. 2. Model of the formation of Maastrichtian shallow-water strata, which were deposited on top of a post-Campanian truncation-surface (section X near monastery of St. Anthony, Text-Fig. 1). North-dipping slump axes indicate E-W directed synsedimentary transcurrent-faults, creating and lineating a small pull-apart basin.

structures with north-dipping axes (BANDEL & KUSS 1987) and mixing with the autochthonous, pelagic sediments (chapt. 3.4). As no autochthonous age-equivalent strata of the early/middle Maastrichtian shallow water facies could be found, it is assumed that the Maastrichtian carbonate platform, where the shallow-water limestones were originally formed, was exposed further south and eroded in post-Maastrichtian times.

We suppose that the strata of St. Anthony Formation were conserved in a pull-apart basin, formed by synsedimentary E-W-striking faults (Text-Fig. 2). This basin is due to pre-rifting processes of the Gulf of Suez Graben: The Arabian Plate moved quicker northwards than the Nubian Plate (LINKE 1986), enabling the opening of such small E-W-striking basins.

3. Description of the microfacies

Deposition of carbonates on the Cretaceous shelf-platform was controlled by two main factors:

(1) Development of local shelf basins within the shelf-platform.
(2) Episodic input and changing sedimentation rates of silty/sandy terrigenous material.

The following description of the main microfacies-types is concentrated on the major constituents and the paleoecologic conditions, which formed the different environments. As the distance from one section to the next is great, and lateral facies changes are very common, it is not possible to correlate these microfacies types (MF-types) bed by bed. They represent summarized associations, present in carbonates of the ammonite zones A-K, and the overlying Maastrichtian limestones (Text-Fig. 3).

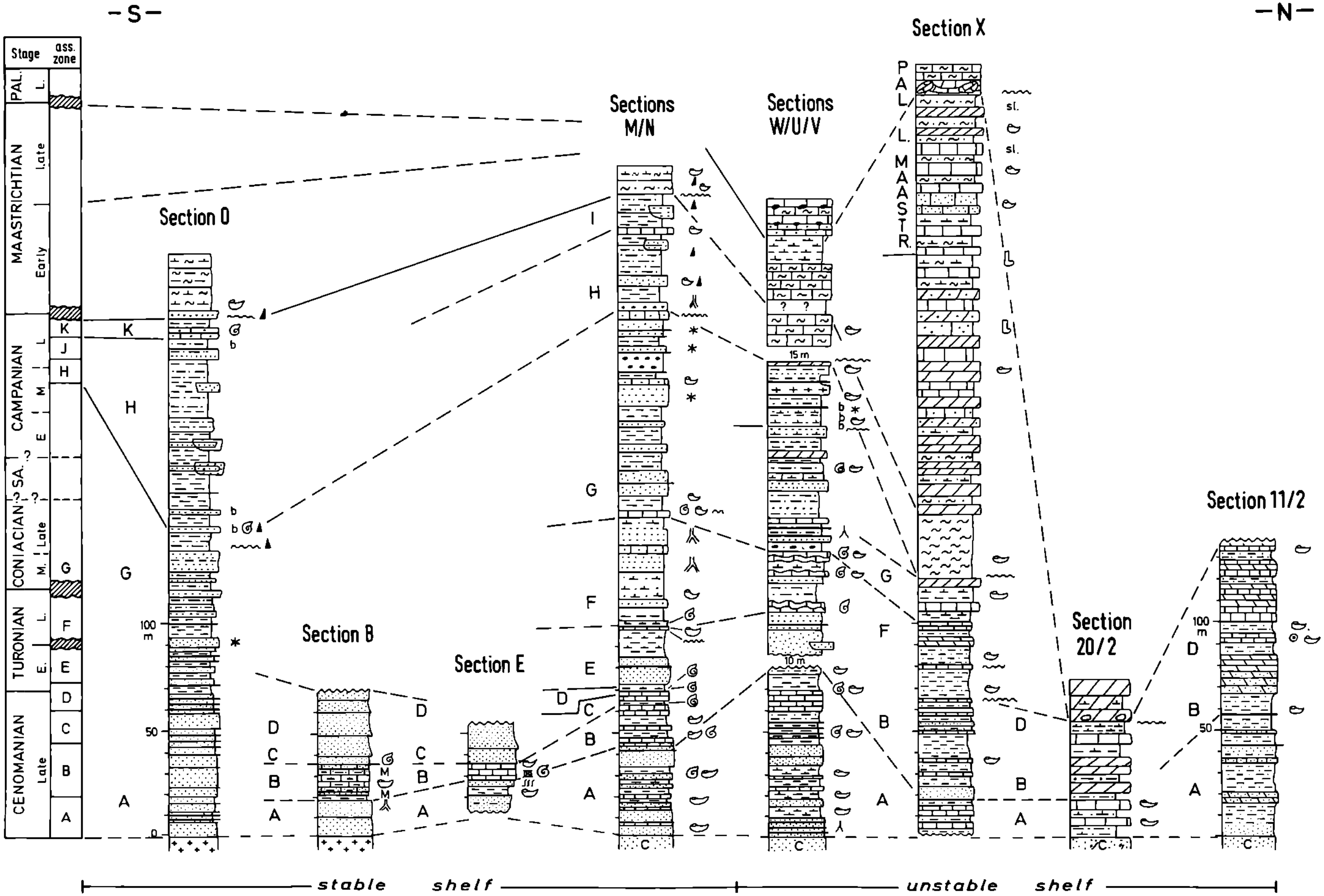
As many of the marly or nodular limestones are dolomitized, it is often impossible to get diagenetically unaltered carbonates. Many steinkerns of large snails, ammonites and echinids provided a better, unaltered composition, because cavities of the shells protected the infilled sediment.

3.1 Late Cenomanian to Turonian limestones

3.1.1 Coral-bearing limestones

Carbonates of this facies-type are restricted to section E (Text-Figs. 1, 3) and few neighbouring outcrops. Several flat coral knobs were found in massive limestones together with gastropods, bivalves (among them oysters) and ammonites. They allow the stratigraphic correlation with zone B.

The sequence starts with rudstones and floatstones, mainly composed of oyster-debris, clastic particles and phosphatic remains. The coral-bearing limestones were classified as wackestones and floatstones. No reefal structures were found here: a zonation in back-reef and fore-reef environments is impossible. Two types of corals were distinguished, according to growth-forms and dimensions: flat coral knobs of *Actinastrea* sp. (colony-diameter up to 43 cm) with thamnasteroid corallites contrast to branching forms, similar to *Thecosmilia* (2.8-3.1 mm diameter of single corallites: Text-Fig. 15). Both coral types are encrusted by serpulids and different algae: *Lithothamnium* sp., *Pseudolithothamnium album* PFENDER, *Archaeolithothamnium* sp., *Pycnopyridium sinuosum* (JOHNSON & KONISHI), and *Pseudochaetetes* sp.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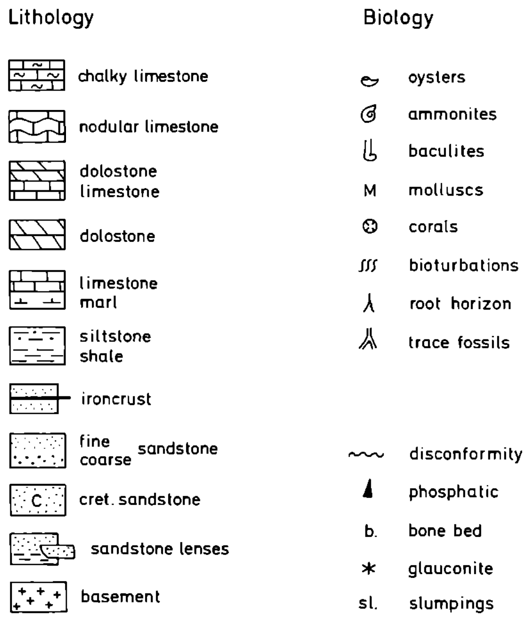
Section 0
Section B
Section E
Sections M/N
Sections W/U/V
Section X
Section 20/2
Section 11/2
PAL.
L. MAASTR.
100 m
50
0
15 m
10 m
sl.
stable shelf
unstable shelf

They occur together with solitary individuals of *Griphoporella* sp. (Text-Fig. 16), *Acicularia* sp., *Neomeris cretacea* STEINMANN, *Trinocladus tripolitanus* RAINERI (Text-Figs. 17, 51), and *Lithocodium* sp. (most of these algae were described in detail by KUSS 1986a). Besides this diverse flora, which is also rich in individuals, single foraminifera of *Charentia cuvillieri* NEUMANN (Text-Fig. 38) and *Gavelinella* sp. occur together with few agglutinated foraminifera. The absence of a more diverse benthic fauna (especially of foraminifera) is surprising in such shallow-water environments. Ostracods, fragments of echinids, molluscs, serpulids and intraclasts are frequent constituents, which often show thin micritic envelopes.

3.1.2 Mixed siliciclastic limestones

This most frequent facies type can be subdivided into four microfacies types (a-d), found in the lower and middle parts of sections B, E, M/N, W/U/V and X, where they often alternate with pure siliciclastic units. The mixing of clastic and carbonate sediments depends on several different factors, the most important are ecologic and tectonic reasons (HAY et al.



Text-Fig. 3. Bio- and lithostratigraphic correlation of the sections indicated in Text-Fig. 1, based on ammonite assemblage zones A-K (late Cenomanian - late Campanian) and on foraminifera (Maastrichtian). Section W/U/V is composite.

1988). It is assumed (compared with other MF-types of the region), that the mixed siliciclastic limestones described herein represent a "facies mixing on rimmed platforms" (MOUNT 1984). The mixing between nearshore tidal flat environments and subtidal carbonate environments is primarily controlled by the terrigenous clastic input, coast-parallel currents, and rates of lateral facies migrations. Periodic breaks in the nearshore clastic sedimentation and lag-deposition are indicated by iron-staining and phosphatic particles. Field observations indicate the interfingering between different nearshore sedimentary environments, as already mentioned by KUSS (1987).

a. Bioclastic wackestones

They are often intercalated within nearshore sand- and siltstones, and were found in the lower parts of sections B, E, M and 20/2. All variations are characterized by a reduced benthic fauna. The thickness of these limestone beds ranges between 15-40 cm; their contacts to the clastic horizons are sharp.

These bioturbated limestones are dominated by changing amounts of terrigenous, clastic components. Most frequent are subrounded quartz grains (which are often iron-stained) together with subordinate fragments of crystalline and metamorphic rocks and some phosphatic remains. The HCl-insoluble matrix consists of clayey, silty material. Fragments of echinids, bryozoans, small serpulids, oysters and other bivalves are frequent biota together with steinkerns of small gastropods and serpulid colonies (Text-Fig. 10); few oncoids occur. A conspicuous micritization affected nearly all components. Sometimes well preserved dasycladacean algae occur: *Boueina* sp. and *Dissocladella undulata* RAINERI. The poor macro-fauna yields some gastropods of *Pterodonta* sp. and oysters.

Many dasycladacean algae and oysters tolerate brackish water and the bioclastic wackestones are characterized by a reduced benthic fauna. Therefore, it is assumed that these mixed siliciclastic limestones were formed in nearshore environments, possibly with freshwater influences.

b. Bioclastic packstones

The hard limestones of this facies type are very frequent in most sections (B, M/N, W/U/V). They are composed of skeletal debris from bivalves, especially oysters, gastropods and echinids. The mechanically reworked shells are concentrated as angular to subangular (ruditic/arenitic) fragments with a moderately good sorting (Text-Figs. 8, 9). Typical prismatic microstructure of *Rhynchostreon siliceum* (LAM.) were identified in several thin-sections (Text-Fig. 9). This form is the only Cretaceous exogyrid oyster with a calcitic prismatic microstructure. Most of the broken shells in Text-Fig. 8 were covered by thin micritic crusts, due to borings by endolithic algae and/or bacteria. Larger cavities (100-300 µm) were mainly found in the outer regions of mollusc shells, caused by different boring organisms, possibly by sponges.

Reworking of the shells was caused by wave action, indicated by cross-bedding, ripple marks and sharp erosional contacts to the underlying strata. The intercalations of these storm deposits with pelmicritic layers (representing low-energy strata) demonstrate rhythmic changes of different shallow-water environments.

c. Peloidal biomicrites

This MF-type consists of pellets (40-60 %) and subrounded fossil-debris (20-30 %) including different agglutinated foraminifera. Coprolites of the genus *Favreina* are frequent (Text-Fig. 11). The quartz and clayey silt content varies between 5-10 %, but may exceed up to 45 %. Pellets are often compressed. Sparitic areas occur within these pelmicrites, which are irregular-shaped, resembling burrows, but may also form thin layers. The sparitic subfacies shows the same composition of biota and skeletal grains as the micritic one. The latter occurs also as reworked intraclasts within the sparitic subfacies.

Three different stages of a conspicuous dolomitization were recognized mainly in samples from sections 20/2 and 11/2, exposed further north: Dolomitization first affected the matrix, later the pellets. Finally the fossil debris was altered to often euhedral dolomitic crystals.

Micritic wackestones with highly fragmented gymnocodiacean and dasycladacean remains and few benthic foraminifera interfinger with the above described facies-type. This "algal debris facies" of the Middle East was first described by ELLIOTT (1958). It includes burrowed micrites with mollusc debris and/or algae. The allochems are often encrusted or micritized. They represent quiet-water deposits in back-shoal environments.

d. Micritic/allochemical sandstones

Besides quartz-grains and rock-fragments, only oyster debris could be identified (Text-Figs. 6, 7). It is not possible to classify this facies-types (most frequent in sections M/N and B) in the well-known terminology of DUNHAM (1962) and FOLK (1962). MOUNT (1985) proposed a classification, which is very useful for such admixtures of siliciclastic and carbonate particles: micritic and allochemical sandstones. They are the most frequent facies types of the mixed siliciclastic limestones.

3.1.3 Miliolid-bearing limestones

Carbonates of this microfacies were only found in section W/U/V and further north in sections X, 20/2, 11/2, and 9/2. Thus they are restricted to the Galala facies province. The miliolid grainstones/packstones contain codiacean fragments (*Boueina pygmaea* PIA, *Boueina* sp.), lumps, cortoids and a diverse fauna of miliolids. Besides *Nezzazata simplex* OMARA, *Cribrostomoides* cf. *paralens* OMARA, *Triloculina* sp., *Bolivinopsis* sp., *Cuneolina* cf. *conica* D'ORBIGNY, *Quinqueloculina* sp., *Cyclogyra* sp., *Haplophragmoides* sp., and *Textularia* sp., which were already described by KUSS (1986b) from zone A-E of section X, the following taxa were determined in section W: *Praealveolina cretacea* D'ARCHIAC (Text-Figs. 18, 24) occurs together with *Textularia* sp., *Dicyclina* sp., small involutinids of *Hensonina lenticularis* HENSON (Text-Figs. 42, 43) and plenty fragments of *Boueina* sp. (Text-Fig. 18). Further to the north, in sections 20/2, 11/2, and 9/2, *Vidalina* sp.

was found with *Biconcava bentori* HAMAOUI & SAINT MARC (Text-Fig. 47), *Pseudorhipidionina casertana* (DE CASTRO) (Text-Fig. 33) and many small miliolids of unknown taxonomic position. Here also fragments of dasycladaceans were found: *Cylindroporella parva* RADOICIC (det. M. CONRAD, Text-Fig. 19) and *Clypeina* sp. (Text-Fig. 50).

Miliolid-bearing wackestones (Text-Fig. 14) usually interfinger with mudstones, which may contain a very diverse fauna of exclusively miliolids. Conspicuous fenestral fabrics (Text-Fig. 13) were found in both facies types. These thick-bedded limestones are often bioturbated, indicating subtidal depositional environments with low clastic input. Miliolid-bearing wackestones are described from Cenomanian strata of several adjacent localities (HAMAOUI & SAINT-MARC 1970, BISMUTH et al. 1981, ARKIN & HAMAOUI 1967). They occur in sections of the Galala facies province and also in equivalent strata of the Sinai. The following foraminifera were determined from both sides of the Gulf of Suez (mainly based on SCHROEDER & NEUMANN 1985): *Biconcava bentori* HAMAOUI & SAINT-MARC (Text-Fig. 47), *Broeckina (Pastrikella) balcanica* CHERCHI, RADOICIC & SCHROEDER (Text-Fig. 36), *Dicyclina* sp. (Text-Fig. 34), *Chrysalidina gradata* D'ORBIGNY (Text-Fig. 39), *Nummofallotia apula* LUPERTO SINNI (Text-Figs. 44, 45), *Merlingina* cf. *cretacea* HAMAOUI & SAINT-MARC (Text-Figs. 48, 49), *Praechrysalidina* sp., *Praealveolina tenuis* REICHEL (Text-Figs. 25-27), *Pseudedomia drorimensis* REISS, HAMAOUI & BECKER (Text-Figs. 28, 29), *Pseudolituonella reicheli* MARIE, *Pseudorhipidionina casertana* (DE CASTRO) (Text-Figs. 32, 33, 40), *Pseudorhapydionina dubia* (DE CASTRO) (Text-Figs. 35, 40), *Spiroloculina* sp. (Text-Fig. 13), *Trochospira avnimelechi* HAMAOUI & SAINT-MARC (Text-Fig. 46). Fragments of *Cylindroporella* sp. (Text-Figs. 52, 53) occur together with other dasycladaceans, which will be described later.

3.1.4 Oncolitic biomicrites

This MF-type is characterized by oncolitic encrustations of coarse arenitic/ruditic shell-fragments, algae, ooids, and intraclasts (Text-Fig. 12). Tiny particles are fixed within the concentric (often asymmetric) crusts. Monospecific algae of *Boueina* sp. and *Boueina pygmaea* PIA are frequent, while echinids, ostracods and few miliolids occur with minor frequencies. As mentioned before, this MF-type interfingers with mudstones or bioclastic packstones of the lumachelle-type, usually found in sections of the Galala facies province.

The described oncoids are morphologic and ecologic adaptions of encrusting organisms to soft-bottom environments (MONTY 1976). Their co-occurrence with mudstones indicates a low-energy depositional environment.

3.1.5 Oolitic grainstones/packstones

Ooid-bearing limestones were found in sections of the Galala facies province and in age-equivalent sections of the northern Sinai. They increase in thickness and frequency from south to north, and are often developed as cross-bedded shoal beds (KUSS 1987), which may be topped by hardgrounds with small boring cavities (section 9/2). Besides ooids and oolithic grapestones (which both reach 40-65 % of the total content), coated grains, algal lumps, ostracods, echinids, and phosphatic particles occur. No benthic foraminifera were found, and only very poorly preserved fragments of few green algae are present. The good sorted ooids are mixed with other bioclasts in both micritic and sparitic cements (Text-Fig. 20).

The single ooids are characterized by thick, radial-fibrous crusts. Compared with investigations on recent ooids, LOREAU (1979) described this ooid type to be formed in marginal marine areas of hypersaline coastal regions, and in terrestrial salt-lakes. It is assumed the here described late Cenomanian to Turonian radial-fibrous ooids were originally formed in lagoonal quiet-water environments, and became later enriched in shoal banks.

The variations of ooid-content and accompanying clasts are due to the palaeogeographic position of ooid-generating environments, which shifted in time. The periodical oscillations of deposition energy resulted in the episodic input of components originated in different depositional environments.

3.1.6 Distribution of late Cenomanian/early Turonian microfacies

Most carbonates of the Wadi Qena facies province are characterized by a high content of clastics, due to the mixing of shallow-marine with clastic-dominated (?freshwater) environments. Though mixed siliciclastic limestones were also found in the Galala facies province, they are more frequent in sections further south.

The offshore environments of the Galala facies province are characterized by thick limestone/dolomite units with minor clastic admixtures. Oolitic, oncolitic and various types of miliolid-bearing limestones prevail here and can be traced to sections in the Sinai.

3.2 Coniacian sequence

Coniacian sediments (zone G) were found at few outcrops only in the Wadi Qena and at section X (chapt. 4). The Coniacian age of the strata from the latter locality were not identified until now (KUSS 1986b), due to the lack of oyster biostratigraphy. These sediments are represented by fossiliferous oolitic packstones, which clearly differ from similar facies types of the Cenomanian/Turonian beds below. The content of phosphatic and glauconitic grains is high; these grains are also acting as nuclei for ooids. Only very small foraminifera of *Praeglobotruncana* sp. and *Gavelinella* sp. are discernable. They occur with reworked algal fragments of *Dissocladella undulata* RAINERI and some debris of corallinacean algae. The intercalations of glauconitic shales, iron-crusts and glauconitic, nodular, sandy *Ophiomorpha*-limestones indicate a nearshore, clastic-dominated regime.

These oolitic bioclastic packstones interfinger with sandy limestones and pure sandstones in age-equivalent sections W, M/N and at the Western Sinai (LEWY 1975).

3.3 Campanian/Maastrichtian phosphates, shales and chalks

As already mentioned before, microfacies analyses were not carried out for both phosphate and chalk facies.

3.4 Maastrichtian limestones

As described before (chapt. 2.6), the occurrence of Maastrichtian limestones is restricted to section X and few neighbouring outcrops (Text-Fig. 1). The first limestones with *Orbitoides media* SCHLUMBERGER (Text-Fig. 21) overlie about 40 m of sandy wackestones at the base of the Maastrichtian section. The wackestones mainly contain fine bioclastic debris without stratigraphic or facial significance. It is assumed that these basal carbonates were formed in deeper shelf areas below the photic zone. Allochthonous *Orbitoides*-packstones are interbedded, containing a diverse shallow-water microfauna and -flora. The before described wackestones occur here as extraclasts, formed by erosion and reworking. Four to five layers of *Orbitoides*-limestones (each 20 to 60 cm thick) can be differentiated, all with a sharp erosive base to the underlying wackestones.

The following foraminifera were determined within the shallow-water *Orbitoides*-limestones: *Hedbergella* sp., *Nodosaria* sp., *Planorbulina* sp., *Pseudoorbitoides* sp., *Rotalipora* sp., *Textularia* sp., *Omphalocyclus* sp., *Siderolithes calcitrapoides* LAMARCK, *Siderolithes vidali* DOUVILLE, and *Sulcoperculina* sp. The last four foraminifera form together with *Orbitoides media* SCHLUMB. a typical early Maastrichtian shallow-water fauna, described from different Tethyan areas (DILLEY 1973). Besides hydrozoan colonies (Text-Fig. 22), several calcareous algae occur, which were figured and described in detail by KUSS (1986a): *Archaeolithothamnium* sp., *Lithothamnium* sp., *Parachaetetes asvapatii* PIA, and *Pseudolithothamnium album* PFENDER.

The marly dolomitic wackestones above (similar to those from the base) yield four species of globotruncanids which prove a middle Maastrichtian age (Gansserina gansseri Zone, KUSS 1986b). Several planktonic forms without stratigraphic importance were also found here. The chalky/sandy limestones on top of the sequence with oysters (chapt. 4.2) again indicate a regressive stage. A late Maastrichtian age is assumed for these strata according to MAZHAR et al. (1979).

4. Biostratigraphy

Biostratigraphic correlations are mainly based on ammonoids (late Cenomanian to late Campanian) and foraminifera (early to late Maastrichtian). They were used to elaborate a regional oyster biostratigraphy (for the following discussion see Table 1).

The faunal assemblages found in Wadi Qena and the Galala Plateaus differ considerably and lead to the proposal of a Wadi Qena faunal province and a Galala faunal province (coinciding with Wadi Qena- and Galala facies provinces: chapts. 2 and 3). The first comprises the area of south and middle Wadi Qena (sections O, B, E, M/N), the second north Wadi Qena and the two Galala Plateaus (sections W/U/V, X, 20/2, and 11/2 - both Text-Fig. 3).

Table 1. Stratigraphical and geographical distribution of ammonites (assemblage zones A-K) and foraminifera used for biostratigraphic subdivision (Gta = *Globotruncanita;* stippled areas indicate hiatus).

Series		Ass. zone	South Wadi Qena (ref. section O)	Middle Wadi Qena (ref. section M/N)	North Wadi Qena	South Galala Plateau	N. Galala Plateau
MAASTRICHTIAN	l.						
	m.					Gansserina gansseri zone	
	e.		upper Gta. falso-stuarti-zone	upper Gta. falso-stuarti-zone		Orbitoides-Siderolithes-Omphalocyclus-assemblage	
CAMPANIAN	l.	K	Nostoceras Solenoceras Libycoceras				
		I		Manambolites/Libycoceras hybrid			
	m.	H	Manambolites Canadoceras				
	e.						
SAN.							
CON.	m/l.	G		Metatissotia Subtissotia			
	e.						
TURONIAN	l.	F		Coilopoceras reqienianum	Coilopoceras requienianum		
	e.	E		Fagesia Mammites		Pseudotissotia Mammites Vascoceras	Pseudotissotia Vascoceras
CENOMANIAN	late	D		Vascoceras (only globose types)		Vascoceras (transitional types between non globose and globose) and Vascoceras (non globose types)	
		C		Vascoceras (non globose types)			
		B		Acanthoceras + Neolobites or Ac.+ Ne.+ Vascoceras or Ne. + Metoicoceras		Neolobites + Metoicoceras or Ne. + Vascoceras	Metoicoceras
		A		Neolobites	Neolobites		

4.1 Ammonoid and foraminifera biostratigraphy

4.1.1 Wadi Qena faunal province

In middle Wadi Qena, section M/N contains eight ammonite-bearing horizons ranging from late Cenomanian to early late Campanian times. A better subdivision of the Campanian strata was possible with two ammonite horizons of section O in southern Wadi Qena. These ammonites were used to define 10 assemblage zones A-K; sections M/N and O were chosen as reference sections. Table 1 shows the stratigraphic range and geographic distribution of the assemblage zones (compare LUGER & GRÖSCHKE 1989).

In section M/N the first three sedimentary cycles (described in chapt. 2) are represented by zones A-E, F and G. They comprise a time span from late Cenomanian to middle/late Coniacian. The upper boundary of zone G can neither be defined biostratigraphically nor lithostratigraphically. It is assumed that the clastic strata overlying the ammonite horizon of zone G also include sediments of Santonian age.

Younger strata with well preserved ammonoids were only found in southern Wadi Qena (section O). Ammonoids of zone H indicate a late middle Campanian time, those of zone K belong to the late Campanian. As both zones were not traceable further to the north, it is difficult to recognize time-equivalent sediments in middle Wadi Qena. In section M/N (middle Wadi Qena) a problematic ammonite (found in strata far above zone G) yields a poor hint for correlation: It combines suture-lines of *Manambolites* and *Libycoceras* (LUGER pers. comm.). As *Manambolites* is the ancestor of *Libycoceras*, this problematic ammonite is interpreted as an intermediate form ("*Manambolites/Libycoceras*-hybrid"). It was used to define zone I restricted to section M/N. This zone lies between zones H (with *Manambolites*) and K (with *Libycoceras*) of section O in southern Wadi Qena (Table 1). Strata between zones G and I in middle Wadi Qena comprise therefore a ?Santonian to middle Campanian time. Strata younger than zone K did not produce any ammonoids.

In section O a horizon with reworked phosphatic particles occurs a few meters above zone K. It is overlain by marls with planktonic foraminifera of the upper Globotruncanita falsostuarti Zone (LUGER, pers. comm.), which is also present in section M/N of middle Wadi Qena. Thus, lithology and biostratigraphy indicate a hiatus ranging from ?uppermost Campanian to early Lower Maastrichtian.

4.1.2 Galala faunal province

In the Galala faunal province sediments of late Cenomanian to late Turonian age (zones A-F) comprise ammonoid assemblages similar to those of the Wadi Qena faunal province (Table 1). Therefore the zonation elaborated for the latter could be applied for the Galala faunal province (composite section W/U/V: Text-Fig. 3, and composite section "south and north Galala": Text-Fig. 4, which summarizes sections 20/2, 13/2, and 15/2; locations see Text-Fig. 1).

The ammonoid record starts with ammonites of zone A (represented by section W of W/U/V: Text-Fig. 3).

Ammonoids of the assemblage zone B were found in sections 13/2, 14/2, and 15/2 (location see Text-Fig. 1) from the southern escarpment of the south Galala Plateau and in section 9/2 of the northern Galala Plateau. Section 13/2 also comprises ammonoids representing vascoceratids transitional between non-globose and globose types together with non-globose types. Therefore no distinction was possible between zones C and D (Table 1).

Ammonoids belonging to zone E were found in sections 15/2 and 9/2.

The characteristic ammonoid of zone F - *Coilopoceras requienianum* - occurs in several horizons of sections U and V (northern Wadi Qena) and could not be traced to sections of the Galala Plateaus.

In northern Wadi Qena and the southern Galala Plateau strata younger than late Turonian lack ammonoids and other index fossils (with the exception of section X). Thus, oysters (chapt. 4.2) and lithostratigraphic correlations were used to recognize the relative position of sediments.

The subdivision of the Maastrichtian strata of section X (at the northern escarpment of the South Galala Plateau) is based on foraminifera: The Globotruncanita falsostuarti Zone described from sections M/N and O of the Wadi Qena faunal province is here represented by an assemblage of benthic foraminifera (chapt. 3). Sediments above contain planktonic foraminifera of the G. gansseri Zone, indicating a middle Maastrichtian age (KUSS 1986b). Late Maastrichtian strata were described by MAZHAR et al. (1979).

In the northern Galala Plateau all Cretaceous strata investigated belong to the late Cenomanian/early Turonian cycle.

4.2 Oyster biostratigraphy

Oysters are widely distributed in both faunal provinces, especially in strata of the late Cenomanian to early Turonian sedimentary cycle. Text-Figs. 4 (late Cenomanian/early Turonian) and 5 (late Turonian/late Maastrichtian) demonstrate the distribution of oyster species in the two faunal provinces. Their stratigraphic ranges are correlated to the ammonoid assemblage zones A-K in both figures, in Text-Fig. 5 also to foraminifera and to relative positions between oysters themselves (sections U, V and X).

4.2.1 Wadi Qena faunal province

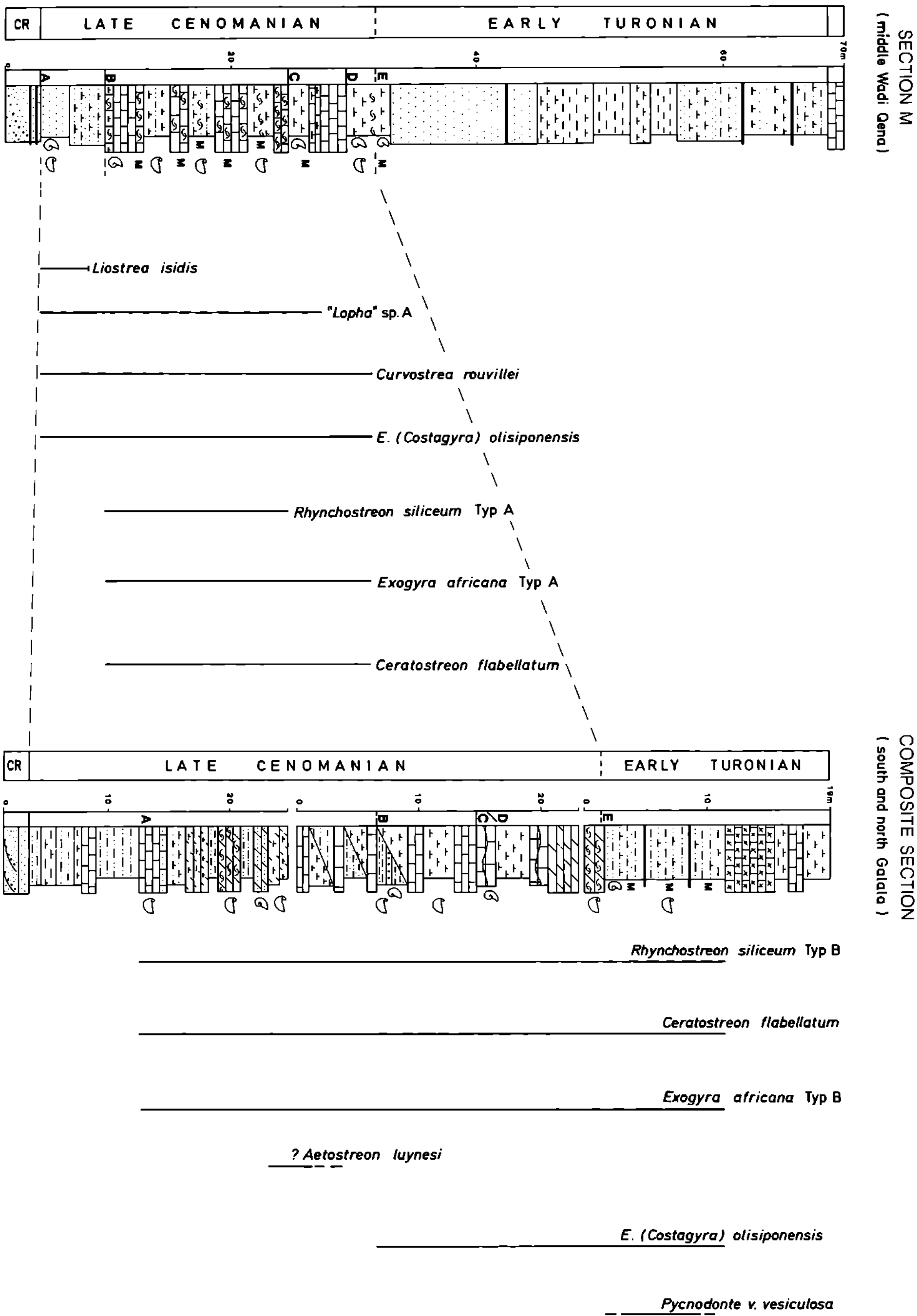
Section M (lower part of section M/N: Text-Fig. 4) comprises only the first sedimentary cycle (zones A-E). Here zone A is characterized by the "index-oyster" *Liostrea isidis* (FOURTAU) co-occurring with "*Lopha*" sp. A, *Curvostrea rouvillei* (COQUAND) and *Exogyra (Costagyra) olisiponensis* SHARP.

The onset of zone B is marked by the absence of *Liostrea isidis* (FOURT.) and by the first appearance of *Rhynchostreon siliceum* (LAMARCK) type A, *Exogyra africana* (LAMARCK) type A and rare specimens of *Ceratostreon flabellatum* (GOLDFUSS).

The oyster composition of zone C is nearly identical to that of zone B, but lacks *R. siliceum* (LAM.) type A.

Just below the top of zone C "*Lopha*" sp. A also disappears, so that zone D is characterized by *C. rouvillei* (COQ.), *E. (C.) olisiponensis* SHARP, *E. africana* (LAM.) type A and *C. flabellatum* (GOLDF.).

The lower part of zone E (early Turonian) lacks oysters. Only *C. rouvillei* (COQ.) reoccurs in the upper part of this zone, which may already belong


SECTION M
(middle Wadi Qena)
CR
LATE CENOMANIAN
EARLY TURONIAN
0
20
40
60
70m
A
B
C
D
E
Liostrea isidis
"Lopha" sp. A
Curvostrea rouvillei
E. (Costagyra) olisiponensis
Rhynchostreon siliceum Typ A
Exogyra africana Typ A
Ceratostreon flabellatum
COMPOSITE SECTION
(south and north Galala)
CR
LATE CENOMANIAN
EARLY TURONIAN
0
10
20
0
10
20
0
10
19m
Rhynchostreon siliceum Typ B
Ceratostreon flabellatum
Exogyra africana Typ B
? Aetostreon luynesi
E. (Costagyra) olisiponensis
Pycnodonte v. vesiculosa

to middle/late Turonian (HENDRIKS et al. 1987: 60; compare section M: Text-Fig. 4, and section M/N: Text-Fig. 5).

Zone F is subdivided in a lower limestone sequence with oysters of *C. rouvillei* (COQ.) and an unfossiliferous upper sandstone sequence.

The characteristic oyster of zone G is *Gyrostrea roachensis* (FOURTAU), which forms wave-resistent build-ups with lateral extensions of 10-20 m and a height of less than 1 m. *G. roachensis* (FOURT.) was still found in the upper part of section M/N; therefore it is not restricted to Coniacian strata. According to the litho- and biostratigraphic correlations between sections O and M/N (Text-Fig. 5), *G. roachensis* (FOURT.) also characterizes ?Santonian to early Campanian strata.

A thin horizon with *"Lopha" bella* (CONRAD) was found above layers with the last occurrence of *G. roachensis* (FOURT.). This small oyster may be of middle to early late Campanian age. In a somewhat higher position of section M/N *Nicaisolopha nicaisei* (COQUAND) and *Pycnodonte (Phygraea) vesiculare* s. str. (LAMARCK) represent the typical oyster assemblage of zone I (early late to late Campanian). *N. nicaisei* (COQ.) does not reach higher in the section, whereas *P. vesiculare* s. str. (LAM.) was still found in the upper G. falsostuarti Zone (late early Maastrichtian of sections M/N and O).

Oyster horizons of the younger strata are monospecific with the exception of the two co-occurring species *N. nicaisei* (COQ.) and *P. vesiculare* s. str. (LAM.). All species but *P. vesiculare* (LAM.) are restricted to the middle Wadi Qena, while oysters found in northern Wadi Qena already belong to the Galala faunal province. In the southern Wadi Qena Cenomanian/Campanian sediments contain no oysters (section O: Text-Fig. 3).

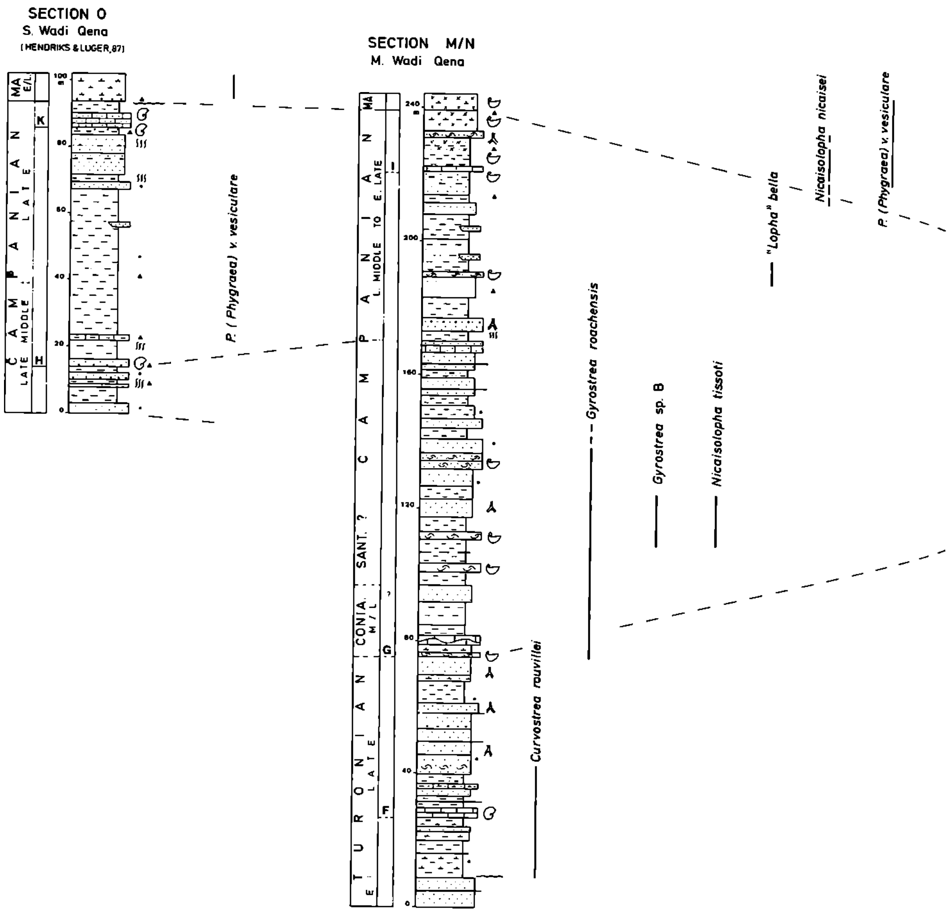
Pycnodonte (P.) vesiculare s. str. (LAM.) also occurs in Campanian chalks (sections V, 13/2, and X) and middle/late Maastrichtian limestones (section X) both of the Galala faunal province, and in early Maastrichtian marls (section O) of the Wadi Qena faunal province (all Text-Fig. 5).

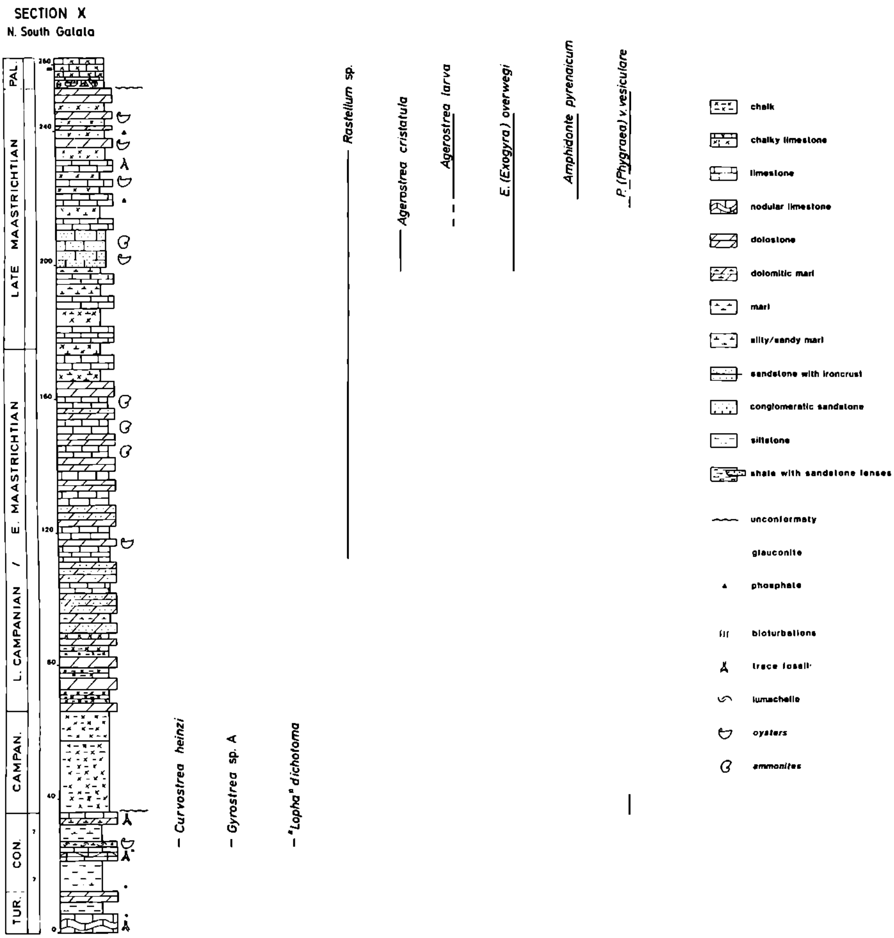
4.2.2 Galala faunal province

The stratigraphic range of oysters found in the late Cenomanian/early Turonian cycle of the Galala faunal province was correlated with the composite section "south and north Galala" (Text-Fig. 4). Stratigraphy of the upper two thirds of this section is based on ammonoids, whereas the lower third was compared with section W of northern Wadi Qena (lower part of composite section W/U/V: Text-Fig. 3).

The late Cretaceous marine sequence in section W starts with shales, containing *Rhynchostreon siliceum* (LAM.) type B, *Exogyra africana* (LAM.) type B, abundant *Ceratostreon flabellatum* (GOLDF.) and *?Aetostreon luynesi*

Text-Fig. 4. Correlation of the late Cenomanian/early Turonian sedimentary cycle (Wadi Qena province and Galala province), including the stratigraphic distribution of oysters; ammonite assemblage zones A-E are indicated. The composite section comprises from base to top sections 20/2, 13/2, and 15/2 (for location see Text-Fig. 1).


SECTION O
S. Wadi Qena
(HENDRIKS & LUGER,87)
CAMPANIAN
LATE MIDDLE
LATE
MA
E/L
H
K
SECTION M/N
M. Wadi Qena
TURONIAN
E
LATE
CONIA.
M/L
SANT. ?
CAMPANIAN
L. MIDDLE TO E. LATE
MA
F
G
I
P. (Phygraea) v. vesiculare
Curvostrea rouvillei
Gyrostrea roachensis
Gyrostrea sp. B
Nicaisolopha tissoti
"Lopha" bella
Nicaisolopha nicaisei
P. (Phygraea) v. vesiculare



Text-Fig. 5. Correlation of late Turonian to early Paleocene sections from Wadi Qena (O, M/N) and southern Galala (X). The geographic provenance and stratigraphic range of oysters are indicated according to the ammonite assemblage zones F-K (for location see Text-Fig. 1, ammonite symbols in section X represent baculitids).

(LARTET). The upper part of the section produced three layers with *Neolobites*, which can be correlated to the assemblage zone A of middle Wadi Qena. The uppermost Cretaceous sediments of section 20/2 (lower part of composite section in Text-Fig. 4) contain the same oyster assemblage with *?Aetostreon luynesi* (LAR.) which was found at the base of section W. The *Neolobites*-horizon that should follow above is missing in section 20/2 due to erosion of the younger strata.

Zone B (proved by ammonoids in sections 13/2, 14/2, 15/2, and 9/2: Text-Fig. 1) and zones C/D (ammonoids in sections 13/2 and 15/2) mostly comprise the same monotonous oyster assemblage consisting of *E. africana* (LAM.) type B, *R. siliceum* (LAM.) type B and *C. flabellatum* (GOLDF.). *E. (Costagyra) olisiponensis* SHARP occurs sporadically.

In addition, section 23/2 contained *Pycnodonte (Phygraea) vesiculare vesiculosum* (SOWERBY) associated with small individuals of *R. siliceum* (LAM.) type B and *C. flabellatum* (GOLDF.). The stratigraphic position of this oyster assemblage is unclear, as *P. v. vesiculosum* (SOW.) does not occur in other sections and no ammonoids were found in this horizon. Lithologic correlations suggest that the strata might belong to zone C/D. The overlying early Turonian strata (zone E) are bare of oysters.

None of the exogyrids mentioned above were found in the late Cenomanian/early Turonian sequence of section X. Only ostreids occur in several thick lumachelles, but were not determinable.

All younger strata of the Galala faunal province are restricted to north Wadi Qena and to the South Galala Plateau. Zone F includes the youngest ammonite horizon that connects the two faunal provinces. This horizon was found in sections U and V, but could not be traced further north. Oysters found in lumachelles of zone F and in directly overlying strata belong to the *Gyrostrea* group, but cannot be determined to the species level.

Text-Fig. 6. Micritic/allochemical sandstone with fragments of oysters, gastropods and dasycladacean algae in a clayey-micritic matrix; M/4, scale 5 mm.

Text-Fig. 7. Micritic/allochemical sandstone with rounded quartz-grains and pressure-solution contacts to oyster shells; M/5, scale 0.5 mm.

Text-Fig. 8. Bioclastic packstone with micritized mollusc-shells; MII/10, scale 5 mm.

Text-Fig. 9. Packstone with oyster-debris of *Rhynchostreon siliceum* (LAM.), identified by its typical prismatic shell-microstructures; M/7, scale 2.5 mm.

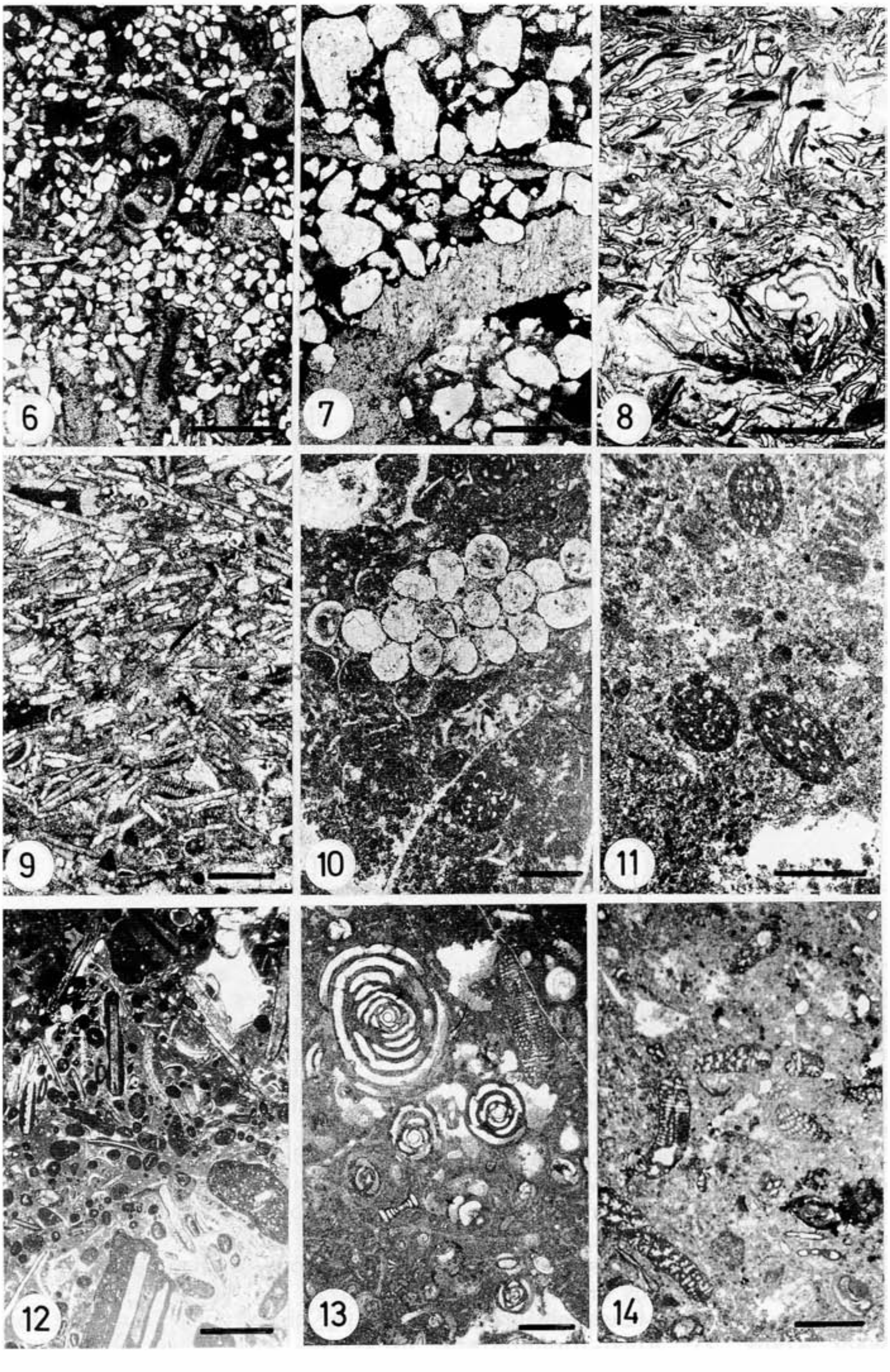
Text-Fig. 10. Peloidal mudstone with a serpulid colony in the centre and few *Favreina* sp.; VI/12b, scale 1 mm.

Text-Fig. 11. Peloidal wackestone with different sections of *Favreina* sp.; 15/2, 1e, scale 1 mm.

Text-Fig. 12. Oncolitic biomicrite, characterized by encrustations of bioclasts and intraclasts; plenty of algal remnants; VI/10, scale 5 mm.

Text-Fig. 13. Miliolid wackestone with sections of *Nummoloculina* sp., *Spiroloculina cretacea* REUSS, *Cuneolina* sp. and fenestral fabrics; VI/12b, scale 0.5 mm.

Text-Fig. 14. Miliolid wackestone with *Cuneolina* sp., *Pseudorhipidionina* sp., *Dicyclina* sp.; XI/3, scale 1 mm.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The overlying sequence in sections V and U mainly consists of channel-sandstones cut into claystones with a low fossil-content. In a higher position of section V two layers of *Gyrostrea* sp. B occur, followed by strata with *Nicaisolopha tissoti* (P. & TH.). The latter was also found in sections 13/2 and 15/2. In contrast section X contains "*Lopha*" *dichotoma* (BAY.) together with *Curvostrea heinzi* (P. & T.) and G. sp. A in a comparable position (see Text-Fig. 5). On Sinai (Ain Sudr) "*Lopha*" *dichotoma* (BAY.) is restricted to the CA 4 zone of LEWY (1975) with early late Coniacian ammonites.

In sections V, 13/2, and X (as well as at Ain Sudr) the sediments with *Nicaisolopha tissoti* (P. & TH.) respectively "*Lopha*" *dichotoma* (BAY.) are discordantly overlain by a conspicuous chalk sequence with an accumulation of *Pycnodonte vesiculare* (LAM.) s. str. in the lower part. This chalk is overlain in section X by late Campanian/early Maastrichtian sediments, supposing that it is of an early to middle Campanian age (although a Santonian age cannot be excluded for the lower part).

Younger oysters than the before mentioned *P. vesiculare* (LAM.) were only found in section X (Galala faunal province): *Rastellum* sp. occurs first in early Maastrichtian sediments and ranges to the top of the sequence. *Exogyra (E.) overwegi* (v. BUCH) and *Agerostrea cristatula* (DOUV.) were found in middle Maastrichtian strata (G. gansseri Zone). In layers above, *A. cristatula* (DOUV.) is substituted by *A. larva* (LAM.), which is accompanied by *Amphidonte pyrenaicum* (LEYMERIE), *E. (E.) overwegi* (BUCH), and *P. (P.) vesiculare* (SOW.) s. str. Only the last oyster was found at the erosive top of the Maastrichtian sequence, which is overlain by middle Paleocene sediments (Text-Fig. 5).

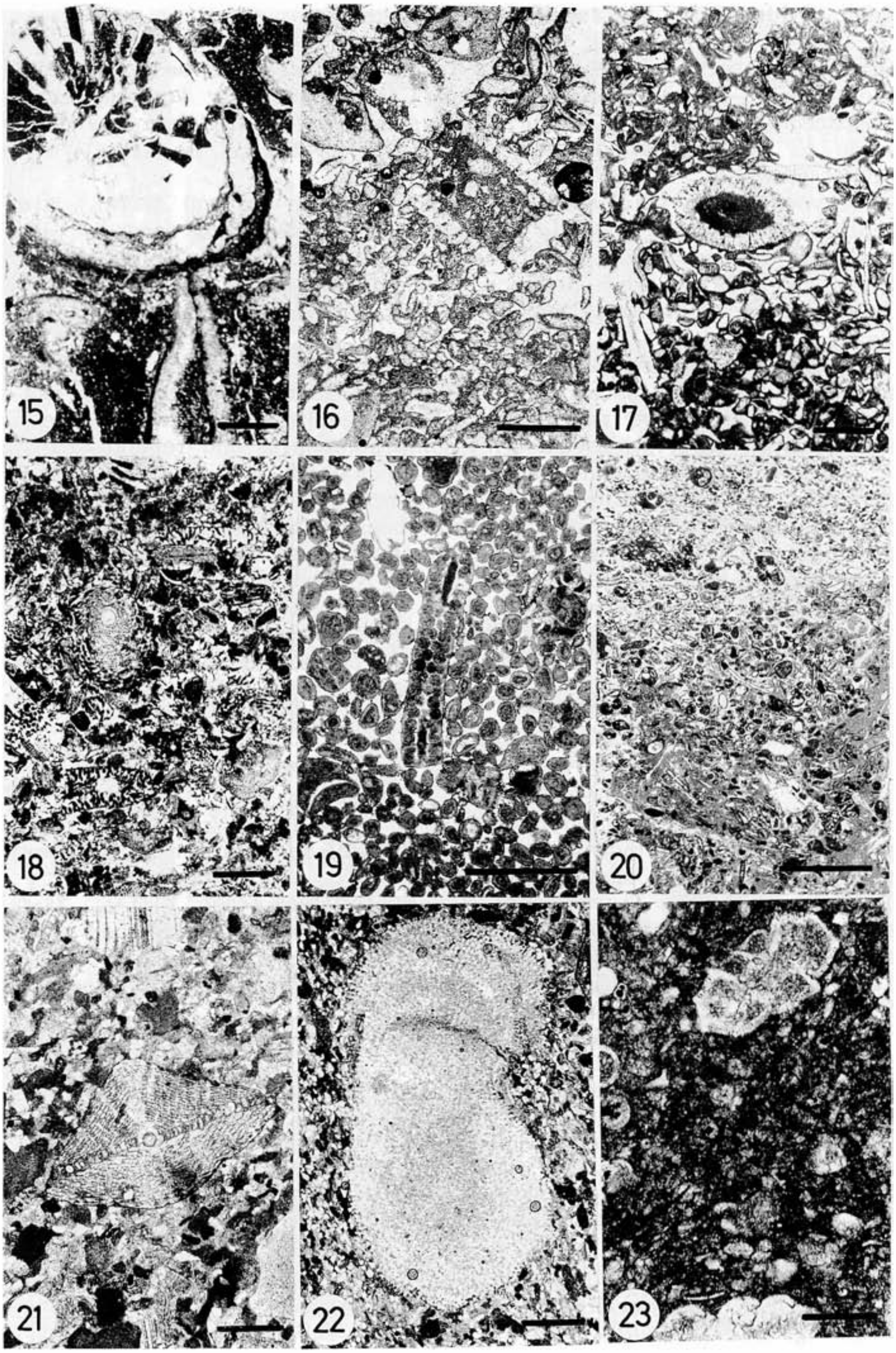
Text-Figs. 15-17. Coral-bearing limestones of the middle Wadi Qena. 15: Solitary coral with encrustations of *Pseudochaetetes* sp. (light colour) and *Lithothamnium* sp. (dark colour); E/5,2, scale 1.4 mm. 16: Grainstone with cortoids and algal remains: a longitudinal section of *Griphoporella* sp. in the centre; L/3,3, scale 1 mm. 17: Grainstone with coated grains, peloids, echinids and algae: *Trinocladus tripolitanus* STEINMANN in the centre; L/3, scale 1 mm.

Text-Fig. 18. Algal packstone with debris of *Boueina* sp. and *Praealveolina cretacea* D'ARCHIAC; W3/2, scale 1.2 mm.

Text-Fig. 19. Miliolid grainstone with *Cylindroporella parva* RADOICIC, stem cut in a longitudinal section; 20/2,10, scale 1.25 mm.

Text-Fig. 20. Partly washed-out oolitic wackestone/grainstone with foraminifera, peloids, cortoids and mollusc-debris; 11/2, 10a, scale 2 mm.

Text-Figs. 21-23. Maastrichtian shallow-water limestones from the southern Wadi Araba. 21: Algal-foraminiferal packstone with a vertical section of *Orbitoides media* SCHLUMBERGER; X/29b, scale 0.85 mm. 22: Algal-foraminiferal packstone with sandy quartzes and a hydrozoan colony; X/29b, scale 2.4 mm. 23: Pelagic wackestone with *Globotruncana* sp.; X/9, scale 150 µm.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4.3 Evidence for the two faunal provinces

Differences between the proposed faunal provinces are best documented in the exogyrid-assemblages of the late Cenomanian/early Turonian cycle: *Ceratostreon flabellatum* (GOLDF.) occurs mainly in the Galala faunal province while *E. (Costagyra) olisiponensis* SHARP is typical for the Wadi Qena faunal province.

Rhynchostreon siliceum (LAM.) and *Exogyra africana* (LAM.) are both represented by two ecologic variants, here called "sp. type A" and "sp. type B"; however "sp. type B" is characteristic of the Galala faunal province. In addition, this province yields *?Aetostreon luynesi* (LART.) and *Pycnodonte vesiculare vesiculosum* (SOW.); both have not been found in Wadi Qena.

Evidence for the two coexisting provinces is also given by other faunal elements: *Plicatula* sp. and *Neithea* sp. are abundant in zones B and C of Wadi Qena and were not found in time-equivalent strata of the Galala area. Similar observations exist for some ammonites and regular sea-urchins.

These differences can neither be explained by differences of water depth nor by salinity fluctuations. A normal marine salinity is indicated by the accompanying fauna in the oyster-bearing horizons of both areas (e. g. pectinids, plicatulids and pinnids, ammonoids, and sea-urchins). The ammonite fauna suggests similar water depths of both faunal provinces (WIEDMANN, pers. comm.).

Instead it is assumed that the discrepancy between the oyster assemblages of the Wadi Qena and the Galala Plateaus is due to differences of

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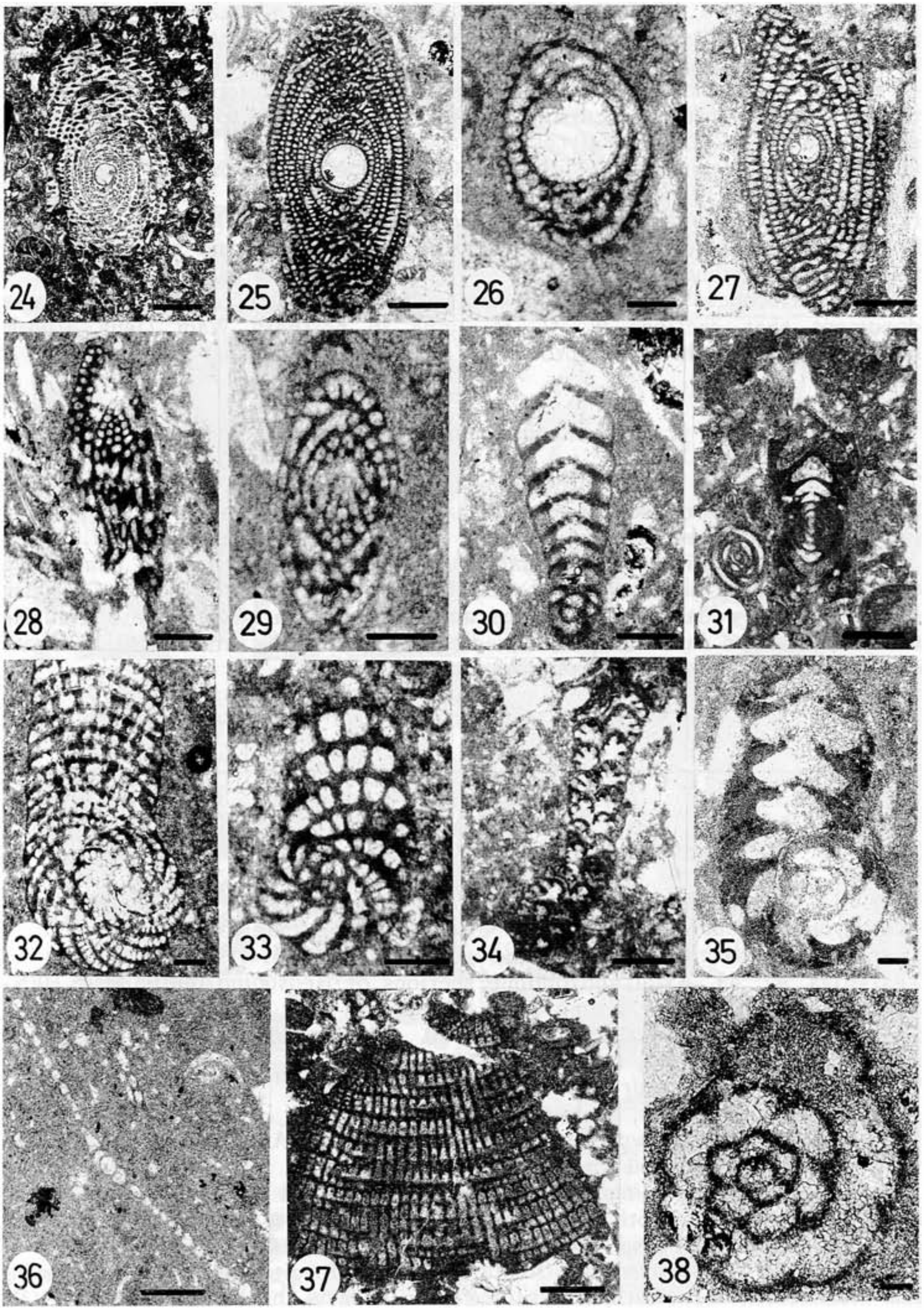
Text-Fig. 24. *Praealveolina* cf. *cretacea* D'ARCHIAC, subaxial section of a partly leached-out individuum (original wall-composition is diagenetically transformed); W/3,2, scale 350 µm.
Text-Figs. 25-27. *Praealveolina tenuis* REICHEL; all VI/13a. 25: Axial section; scale 360 µm. 26: Juvenile form with a slightly flattened proloculus; scale 140 µm. 27: Nearly axial section with gulo; scale 360 µm.
Text-Figs. 28, 29. Two subaxial sections of *Pseudedomia drorimensis* REISS, HAMAOUI & ECKER; both VI/13a, scale 300 µm.
Text-Fig. 30. Axial section of *Pseudorhapydionina laurinensis* (DE CASTRO); VI/13a, scale 200 µm.
Text-Fig. 31. *Nummoloculina* sp., axial section of a microspheric form; VII/14, scale 230 µm.
Text-Figs. 32, 33. *Pseudorhipidionina casertana* (DE CASTRO), both sections are cut oblique-subaxial. 32: VII/12, scale 150 µm. 33: VII/14, scale 110 µm.
Text-Fig. 34. *Dicyclina* sp., oblique section; VII/14,2, scale 180 µm.
Text-Fig. 35. *Pseudorhipidionina dubia* (DE CASTRO), equatorial section; VI/14, scale 70 µm.
Text-Fig. 36. *Broeckina (Pastrikella) balcanica* CHERCHI, RADOICIC & SCHROEDER, axial section; VII/14, scale 380 µm.
Text-Fig. 37. *Cuneolina pavonia* HENSON, subaxial section; 11/2,2, scale 250 µm.
Text-Fig. 38. *Charentia cuvillieri* NEUMANN, equatorial section; 11/2, 13a, scale 60 µm.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the clastic influx: While a high terrigenous input prevailes in the south, shales, silts, and carbonates increase in frequency and thickness further north (chapts. 2 and 3).

Faunal differences were also preserved in late Turonian to Campanian times. Ostreid oysters occurring in Wadi Qena faunal province are: *Curvostrea rouvillei* (COQ.), *Gyrostrea roachensis* (FOURT.), *"Lopha" bella* (CON.), and *Nicaisolopha nicasei* (COQ.). The Galala faunal province is characterized by *Gyrostrea* sp. A and B, *N. tissoti* (P. & TH.), *C. heinzi* (P. & TH.), *"L." dichotoma* (BAY.) and *Pycnodonte vesiculare* (LAM.) s. str. Both facies-interpretation and accompanying fauna gave no hints to explain this oyster distribution.

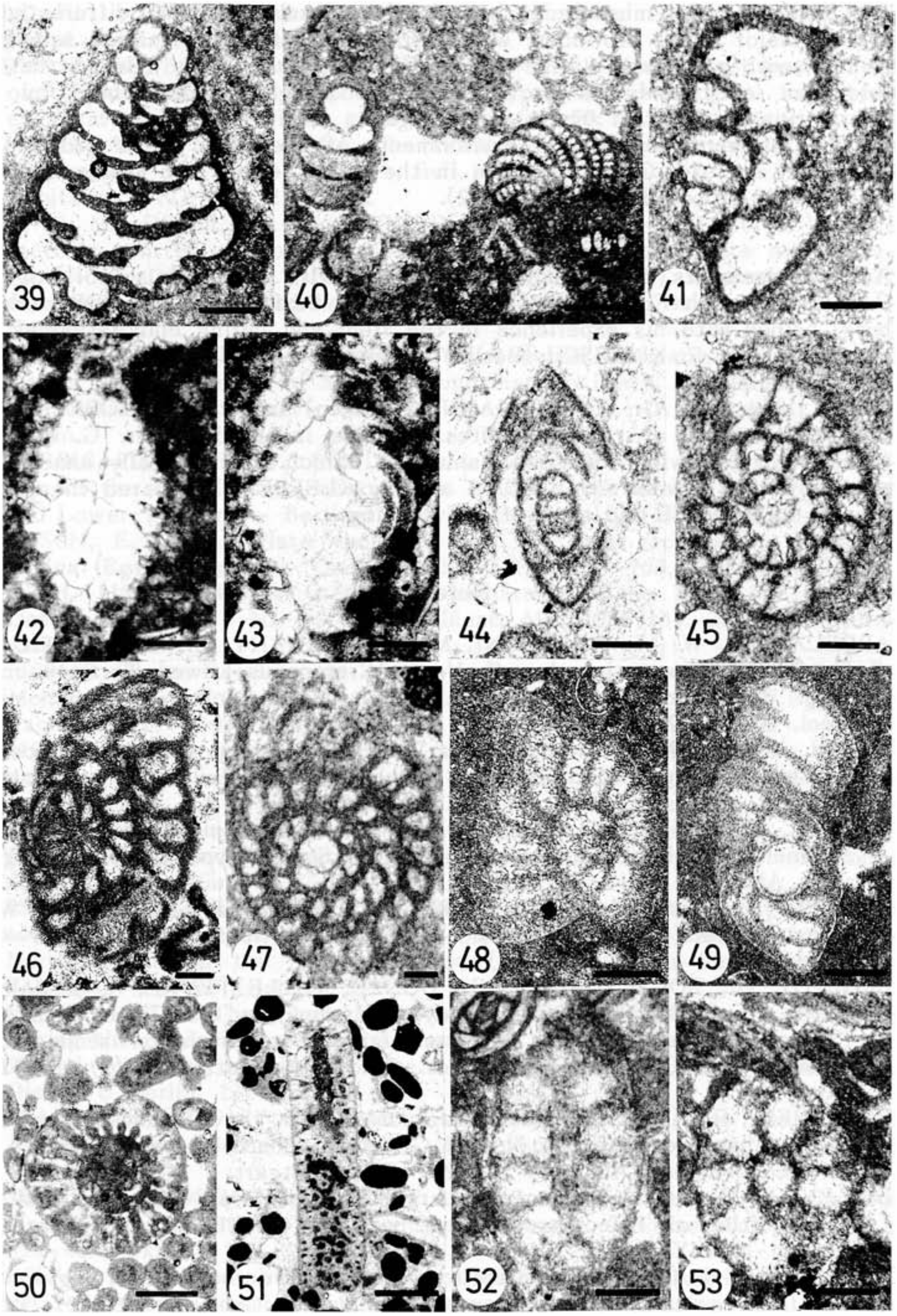
The differences of late Cenomanian to Campanian oyster assemblages document that sections W, U, V, lying in northern Wadi Qena already belong to the Galala faunal province.

5. Results and conclusions

The present study deals with facies interpretation and biostratigraphy of late Cenomanian to late Maastrichtian strata in northeastern Egypt. The investigated area represents a segment of the southern Tethyan shelf.

The new established oyster biostratigraphy produced results comparable to the ammonite biostratigraphy for the late Cenomanian to late Coniacian strata. Only few data of oyster and ammonite biostratigraphy were available for Santonian to late Campanian sediments, while the Maastrichtian strata could be subdivided with foraminifera and oysters.

Text-Fig. 39. *Chrysalidina gradata* D'ORBIGNY, axial section; VII/14, scale 280 µm.
Text-Fig. 40. *Pseudorhipidionina casertana* (DE CASTRO) - right, and *Pseudorhapydionina dubia* DE CASTRO) - left; VI/12b, scale 240 µm.
Text-Fig. 41. *Nezzazata* sp., oblique axial section; VI/14, scale 110 µm.
Text-Figs. 42, 43. Axial and oblique sections of *Hensonina lenticularis* HENSON; W/3, both scales 70 µm.
Text-Figs. 44, 45. Axial and equatorial sections of *Nummofallotia apula* LUPERTO SINNI; VI/14, both scales 75 µm.
Text-Fig. 46. *Trochospira avnimelechi* HAMAOUI & SAINT-MARC, oblique subaxial section; VI/13a, scale 65 µm.
Text-Fig. 47. *Biconcava bentori* HAMAOUI & SAINT-MARC, equatorial section; VI/13a, scale 60 µm.
Text-Figs. 48, 49. *Merlingina* cf. *cretacea* HAMAOUI & SAINT-MARC, axial and oblique sections; both VI/12b, scale 180 µm.
Text-Fig. 50. *Clypeina* sp., horizontal section; 20/2, 10d, scale 480 µm.
Text-Fig. 51. *Trinocladus tripolitanus* RAINERI, longitudinal section, showing the arrangement of whorls and primary, secondary and tertiary pores; E/3b, scale 1.1 mm.
Text-Figs. 52, 53. *Cylindroporella* sp., horizontal and oblique sections; both VI/13, scales 160 µm.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Lithofacies and microfacies investigations resulted in the differentiation of two major facies provinces, with terrigenous influenced marine sediments in the south and open marine strata further north. They induced the development of two faunal provinces, which are best documented by the distribution of different oyster assemblages.

The different depositional environments of the Wadi Qena province in the south and the Galala province in the north reflect the stable and unstable shelf proposed by SAID (1962).

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Problems of Upper Cretaceous Inoceramid Biostratigraphy and Paleobiogeography in Europe and Western Asia

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With 8 Text-Figures

TRÖGER, K.-A. (1989): Problems of Upper Cretaceous Inoceramid Biostratigraphy and Paleobiogeography in Europe and Western Asia. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 911-930. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Besides ammonites, belemnites and echinoids, the inoceramids are the most important key fossils of the Upper Cretaceous. In contrast to ammonites and belemnites, the lateral distribution of inoceramids is, in most cases, independent of lithofacies differentiations. A major part of all described inoceramid species and subspecies have an "intercontinental" or cosmopolitan distribution. The utility of inoceramids for stratigraphical purposes owes to their rapid evolution especially between the Cenomanian and Campanian stages. The Cenomanian to Campanian strata can be divided into 32 assemblage or range zones using the inoceramid faunal successions. The levels of the Cenomanian/Turonian boundary, the Turonian/Coniacian boundary and the Coniacian/Santonian boundary are fixed by the appearance of ammonite species or subgenus and may be confirmed by inoceramids.

The North European Province (E. G. KAUFFMAN) including the Crimea-Caucasus-Mangyschlak-Kopet Dag region, is situated at the northern border of the Mediterranean Province (Tethyan Realm). A widespread dispersal of both inoceramid species and subspecies took place during the Upper Cenomanian-Lower Turonian (*Inoceramus pictus* SOWERBY, *Inoceramus (Mytiloides) labiatus*-lineage), during the Upper Turonian-Lower Coniacian (*Inoceramus (Mytiloides) incertus* JIMBO, *Inoceramus dresdensis* TRÖGER) and during the Middle Coniacian-Lower Santonian (involute inoceramids, *Inoceramus (Magadiceramus) subquadratus* SCHLÜTER). A connection is visible between the time of migration and the pulse of global transgressions and regressions. It seems that the migration pathway between the North European Province and the Japan-East Asia Subprovince followed the northern border of the Tethyan region. The migration pathway from the North Temperate Realm (especially N-Jenissei region) to the North European Province followed the Turgai seaway east of the Ural Mts.

Kurzfassung: Neben Ammoniten, Belemniten und Echiniden besitzen vor allem die Inoceramen in der Oberkreide eine große biostratigraphische Bedeutung. Im Gegensatz zu den Ammoniten und Belemniten sind die Inoceramen, von Ausnahmen abgesehen, in ihrer lateralen Verbreitung nicht faziesabhängig. Ein großer Teil aller beschriebenen Arten besitzt außerdem eine "interkontinentale", in einzelnen Fällen nahezu kosmopolitische Verbreitung. Durch die

schnelle Entwicklung hauptsächlich im Zeitraum Cenomanium - Santonium ist die biostratigraphische Bedeutung der Inoceramen bedingt. Der genannte Abschnitt kann auf der Grundlage der Inoceramen-Entwicklung in 32 Zonen gegliedert werden. Die mit Hilfe der Inoceramen-Biostratigraphie festgelegten Grenzen zwischen dem Cenoman und Turon, dem Turon und Coniac und dem Coniac und Santon stimmen mit den aufgrund der Ammoniten-Entwicklung festgelegten Grenzen überein.

Die Nordeuropäische Provinz im Sinne von E. G. KAUFFMAN, die nördlich der Mediterranen Provinz gelegen ist, schließt die Gebiete der Berg-Krim, des Kaukasus, des Mangyschlak und des Kopet Dag mit ein. Eine weitgehende Migration einzelner Arten und Unterarten ist besonders an die nachfolgenden Zeiträume gebunden:

Ob.-Cenoman/Unt.-Turon	(*Inoceramus pictus* SOWERBY, *Inoceramus (Mytiloides) labiatus*-Entwicklungsreihe)
Ob.-Turon/Unt.-Coniac	(*Inoceramus (Mytiloides) incertus* JIMBO, *Inoceramus dresdensis* TRÖGER)
Mitt.-Coniac/Santon	(involute Inoceramen, *Inoceramus (Magadiceramus) subquadratus* SCHLÜTER)

Es bestehen ursächliche Zusammenhänge mit den oberkretazischen globalen Trans- und Regressionen. Die Migrationen zwischen der Nordeuropäischen Provinz und der Japan-Ostasien-Subprovinz erfolgten augenscheinlich entlang des Nordrandes der Tethys. Vom Boreal (North Temperate Realm) insbesondere vom N-Jenissei-Gebiet ausgehend wurde die Turgai-Straße benutzt, die seit dem Turon geöffnet war.

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1. Introduction

The boundaries of the series and stages including the substages are fixed by the stratigraphical range of ammonite species in the Cretaceous System. It must firstly be mentioned, that in the Upper Cretaceous rocks, especially in some lithofacies, the ammonites are rare. Mostly they are bound to the limy and marly lithofacies. On account of this reason, especially inoceramids, belemnites, echinoids and crinoids were used for stratigraphical divisions, too (see Text-Figs. 4 and 5). In Europe the importance of inoceramids for stratigraphical purposes especially was pointed out by O. SEITZ (1922) and R. HEINZ (1928a, b). R. HEINZ (1928b) investigated the Upper Cretaceous sequence of Lüneburg (Cenomanian-Lower Maastrichtian) and discriminated 15 layers characterized by inoceramid species or inoceramid assemblages consisting of 2-9 inoceramid species. In recent years, the utility of inoceramids in global biostratigraphy and for international correlations has been firmly established by O. SEITZ for Europe, T. MATSUMOTO for Japan, M. A. PERGAMENT for Eastern Asia and E. G. KAUFFMAN for North and Latin America. Three facts made the inoceramids useful for stratigraphical purposes:

1. A rapid evolution of inoceramids with an ample increase of both species and subspecies took place from Cenomanian to Santonian times (see Text-Figs. 1-4). Evolutionary lineages can be observed and are used for stratigraphical divisions (for example *Inoceramus costellatus pietzschi*

TRÖGER - *Inoceramus costellatus costellatus* WOODS: Middle and Upper Turonian, Text-Fig. 2).

2. In the most cases, inoceramids are abundant and not bound to particular lithofacies. They were observed in limestones, marls, claystones and fine to middle grained sandstones. Only in the facies of submarine swells and reefs they are absent or extremely rare.
3. According to E. G. KAUFFMAN (1977), a great amount (75 %) of all described species and subspecies have an "intercontinental" to cosmopolitan distribution. The rapid and widespread dispersal may be caused by means of long-lived, environmental tolerant and planktonic larvae (E. G. KAUFFMAN 1977) or by an epiplanktonic way of living on floating wood or vegetation. Therefore, a great part of all inoceramid species and subspecies does not play an important role in the definition of paleobiogeographic units.

The following remarks are bound to investigations of sequences and collections being sampled in the North European Province including the Caucasian, the Mangyschlak and the Kopet Dag area.

2. Inoceramid zonation of the Cenomanian to Campanian stages in the North European Province

The rapid evolution of Upper Cretaceous inoceramids, especially between the Cenomanian and Santonian, comprises the base for dividing the interval between the Lower Cenomanian and Upper Campanian into 32 zones (on the Text-Figs. 1-5 numbered 1-32). In the most cases the zones are inoceramid assemblage zones. In some cases we may speak of taxon range zones (nos. 3, 4, 28), too. Between the Cenomanian and Lower Santonian the boundaries fixed by means of inoceramids or inoceramid assemblages are corresponding or nearly corresponding with the boundaries fixed by ammonites.

2.1 Cenomanian

The proposals of W. J. KENNEDY (1984) and U. KAPLAN, S. KELLER & J. WIEDMANN (1984) were used to divide the Cenomanian stage into 10 ammonite zones (Text-Fig. 1). It is possible to distinguish 6 inoceramid zones or assemblage zones, the boundaries of which partly agreeing with the boundaries of the ammonite zones.

The basal Lower Cenomanian is characterized by the appearance of large specimens of *Inoceramus crippsi crippsi* MANTELL. Besides, rare specimens of the *Inoceramus anglicus*-group can be observed. The inoceramid assemblage zone 2, corresponding with the upper part of the Lower Cenomanian, consists of *Inoceramus crippsi crippsi* MANTELL, *Inoceramus crippsi hoppenstedtensis* TRÖGER, *Inoceramus virgatus* SCHLÜTER (tall and small subspecies), *Inoceramus tenuis* MANTELL and *Inoceramus etheridgei etheridgei* WOODS. The most important species of Middle Cenomanian age are *Inoceramus schoendorfi* HEINZ (zone 3) and *Inoceramus atlanticus* HEINZ (zone 4). In some Middle Cenomanian sequences (for example: Subhercynian Cretaceous Basin/GDR and Crimea area) the rocks are unfossiliferous. In this case it is impossible to divide the zones 3 and 4. It is not quite clear if an overlapping interval of the range of the 2 species exists.

The Upper Cenomanian is characterized by the *Inoceramus pictus*-lineage with several subspecies all bound to the basal Upper Cenomanian (zone 5, Text-Fig. 1). On the contrary, the uppermost Upper Cenomanian (zone 6) only yields the small inoceramid subspecies *Inoceramus pictus bohemicus* LEONHARD.

2.2 Turonian

Using ammonites for establishing the Cenomanian/Turonian boundary causes some difficulties described especially by W. J. KENNEDY (1984) and T. BIRKELUND, J. M. HANCOCK et al. (1984). In the past, the appearance of the assemblage zone of *Watinoceras coloradoense* mostly was used to fix the boundary level. In some boundary sequences, for example in the Subhercynian Cretaceous Basin, the rocks are almost devoid of fossils. In other profiles ammonites are missing. There are widespread condensations and breaks in western Europe including the type area (W. J. KENNEDY 1984) and in great parts of eastern Europe as well (GERASIMOV et al. 1962, NAIDIN 1969 cit. in K.-A. TRÖGER 1978). E. SEIBERTZ (1979), K.-A. TRÖGER (1981) and T. BIRKELUND, J. M. HANCOCK et al. (1984) therefore proposed to fix the boundary with the entry of first members of the *Inoceramus (Mytiloides) labiatus* lineage. The basal Lower Turonian inoceramid zone (no. 7, Text-Fig. 2) contains *Inoceramus (Mytiloides) submytiloides* SEITZ, *Inoceramus (Mytiloides) mytiloides* MANTELL and *Inoceramus (Mytiloides) labiatus* (v. SCHLOTHEIM). Regarding the range, there is a small overlap with *Inoceramus pictus bohemicus* LEONHARD at the base of the zone 7. A further problem is the level of the Lower/Middle Turonian boundary which was set by O. SEITZ (1922), applying to CL. SCHLÜTER, at the appearance of *Inoceramus lamarcki* PARK. Breaks exist in many sequences of eastern Europe ranging from Lower to Upper Turonian (GERASIMOV et al. 1962, NAIDIN 1969 cit. in K.-A. TRÖGER 1978). At several places, for example in the Saxony Cretaceous Basin, *Inoceramus (Mytiloides) hercynicus* PETRASCHECK is accompanied by the Middle Turonian ammonite species *Collignoniceras woollgari* MANTELL. The base of the Middle Turonian therefore must be placed in the inoceramid assemblage zone 9 or at the base of this zone. The further division of the Middle Turonian is given by the development of the *Inoceramus lamarcki*-lineage (assemblage zones 13-16), the *Inoceramus (Striatoceramus) striaconcentricus* lineage and the *Inoceramus costellatus* lineage (zone 17, Text-Fig. 2). The lower part of the

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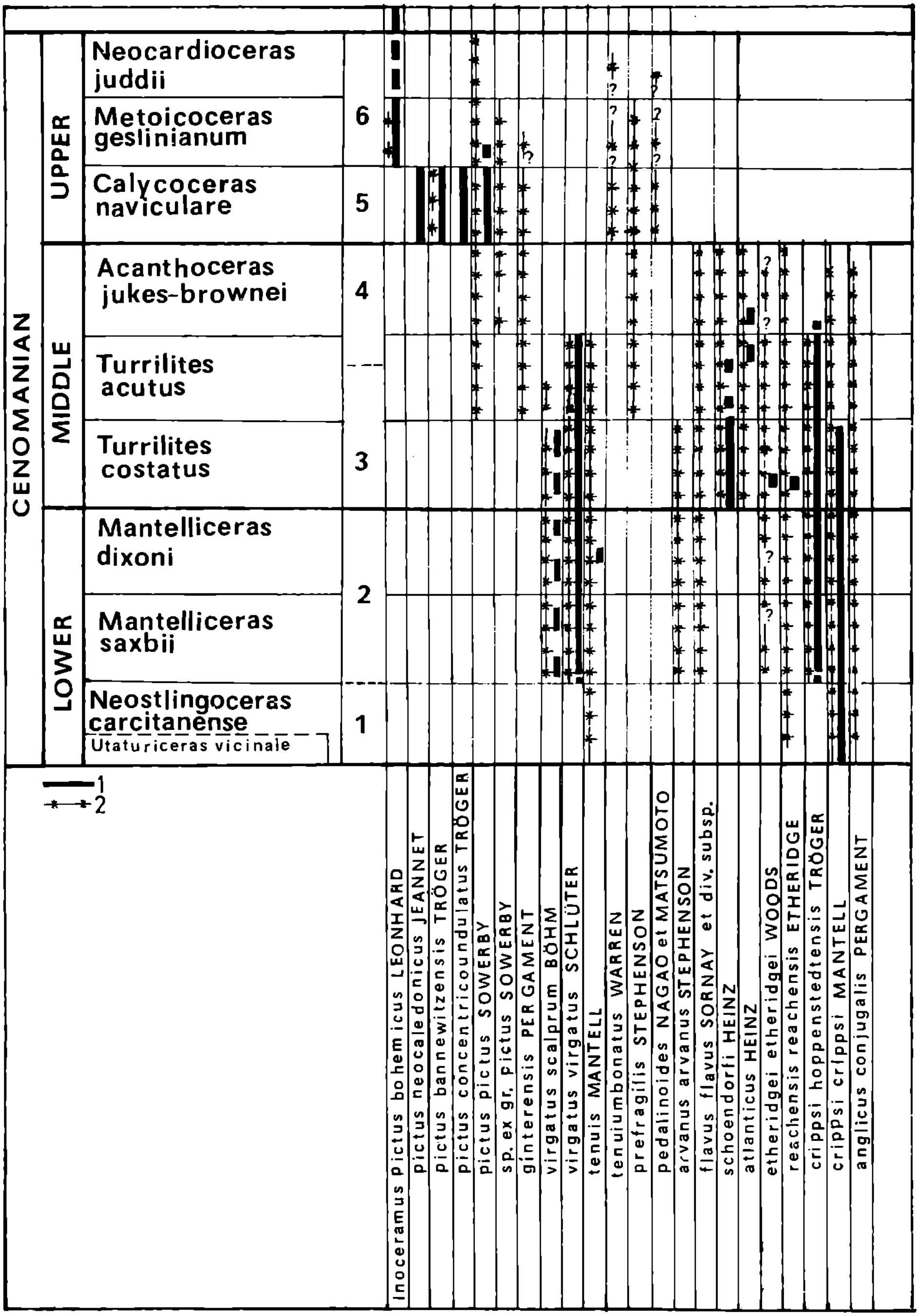
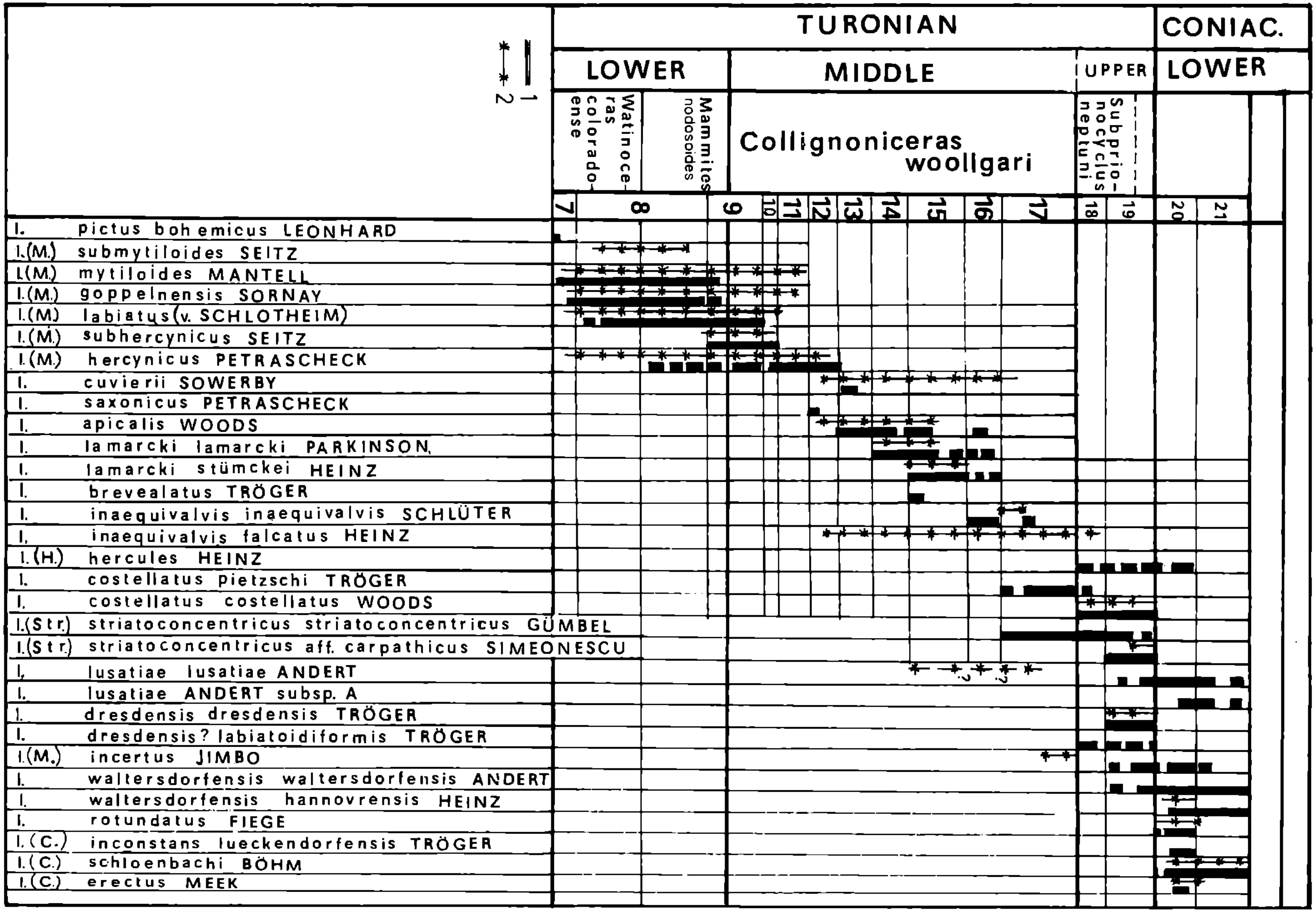
Text-Fig. 1. Diagram showing the stratigraphical distribution of the most important Cenomanian inoceramid species and subspecies and the zonal division of the Cenomanian in the North European Province using inoceramid assemblages according to U. KAPLAN, S. KELLER & J. WIEDMANN (1984), E. G. KAUFFMAN (1976), W. J. KENNEDY (1981, 1984), W. J. KENNEDY & M. J. HANCOCK (1970), R. MARCINOWSKI (1974), D. P. NAIDIN (1981), F. ROBASZYNSKI (1984) and K.-A. TRÖGER (1981).
1 - distribution in Middle and Eastern Europe, 2 - distribution in Western Europe.


CENOMANIAN
UPPER
MIDDLE
LOWER
Neocardioceras juddii
Metoicoceras geslinianum
Calycoceras naviculare
Acanthoceras jukes-brownei
Turrilites acutus
Turrilites costatus
Mantelliceras dixoni
Mantelliceras saxbii
Neostlingoceras carcitanense
Utaturiceras vicinale
6
5
4
3
2
1
1
2
Inoceramus pictus bohemicus LEONHARD
pictus neocaledonicus JEANNET
pictus bannewitzensis TRÖGER
pictus concentricoundulatus TRÖGER
pictus pictus SOWERBY
sp. ex gr. pictus SOWERBY
ginterensis PERGAMENT
virgatus scalprum BÖHM
virgatus virgatus SCHLÜTER
tenuis MANTELL
tenuiumbonatus WARREN
prefragilis STEPHENSON
pedalinoides NAGAO et MATSUMOTO
arvanus arvanus STEPHENSON
flavus flavus SORNAY et div. subsp.
schoendorfi HEINZ
atlanticus HEINZ
etheridgei etheridgei WOODS
reachensis reachensis ETHERIDGE
crippsi hoppenstedtensis TRÖGER
crippsi crippsi MANTELL
anglicus conjugalis PERGAMENT

Text-Fig. 1


1
2
TURONIAN
CONIAC.
LOWER
MIDDLE
UPPER
LOWER
Watinoceras coloradoense
Mammites nodosoides
Collignoniceras woollgari
Subprionocyclus neptuni
7 8 9 10 11 12 13 14 15 16 17 18 19 20 21
I. pictus bohemicus LEONHARD
I.(M.) submytiloides SEITZ
I.(M.) mytiloides MANTELL
I.(M.) goppelnensis SORNAY
I.(M.) labiatus (v. SCHLOTHEIM)
I.(M.) subhercynicus SEITZ
I.(M.) hercynicus PETRASCHECK
I. cuvierii SOWERBY
I. saxonicus PETRASCHECK
I. apicalis WOODS
I. lamarcki lamarcki PARKINSON
I. lamarcki stümckei HEINZ
I. brevealatus TRÖGER
I. inaequivalvis inaequivalvis SCHLÜTER
I. inaequivalvis falcatus HEINZ
I.(H.) hercules HEINZ
I. costellatus pietzschi TRÖGER
I. costellatus costellatus WOODS
I.(Str.) striatoconcentricus striatoconcentricus GÜMBEL
I.(Str.) striatoconcentricus aff. carpathicus SIMEONESCU
I. lusatiae lusatiae ANDERT
I. lusatiae ANDERT subsp. A
I. dresdensis dresdensis TRÖGER
I. dresdensis? labiatoidiformis TRÖGER
I.(M.) incertus JIMBO
I. waltersdorfensis waltersdorfensis ANDERT
I. waltersdorfensis hannovrensis HEINZ
I. rotundatus FIEGE
I.(C.) inconstans lueckendorfensis TRÖGER
I.(C.) schloenbachi BÖHM
I.(C.) erectus MEEK
?

Text-Fig. 2

Upper Turonian yields especially the subspecies *Inoceramus (Striatoceramus) striatoconcentricus* GÜMBEL and *Inoceramus costellatus costellatus* WOODS (zone 18). The inoceramid assemblage of the upper part of the Upper Turonian (partly limestones of Dresden-Strehlen) consists of the two mentioned species together with members of the *Inoceramus (Mytiloides) incertus* lineage and the *Inoceramus dresdensis* lineage.

2.3 Coniacian

According to T. BIRKELUND, J. M. HANCOCK et al. (1984) and having studied the classical references, the Turonian/Coniacian boundary was placed at the base of the zone with *Barroisiceras haberfellneri* (= specimens of the type area at Charente belonging to *Forresteria (Harleites) petrocoriensis*). In most sequences, ammonites belonging to the subgenus *Forresteria (Harleites)* are rare or absent. E. SEIBERTZ (1979) and K.-A. TRÖGER (1981) proposed to establish the boundary level with the entry of *Inoceramus rotundatus* FIEGE (zone no. 20, Text-Fig. 2). Besides this species, the assemblage zone contains *Inoceramus (Mytiloides) incertus* JIMBO; *Inoceramus waltersdorfensis waltersdorfensis* ANDERT, *Inoceramus waltersdorfensis hannovrensis* HEINZ and in the upper part *Inoceramus (Cremnoceramus) inconstans lueckendorfensis* TRÖGER and *Inoceramus (Cremnoceramus) schloenbachi* J. BÖHM. At Zatzschke (Saxony Cretaceous Basin) the zone 20 contains *Placenticeras d'orbygnianum* GEINITZ as well.

The upper part of the Lower Coniacian (zone no. 21) especially yields *Inoceramus (Cremnoceramus) schloenbachi* J. BÖHM and *Inoceramus (Cremnoceramus) ernsti* HEINZ. The distribution of *Inoceramus gradatus* EGOJAN in this zone is restricted to eastern Europe (Crimea area, Caucasus area). The appearance of involute inoceramids, for example *Inoceramus (Volviceramus) koeneni* G. MÜLLER and *Inoceramus (Volviceramus) involutus* SOWERBY marks the Middle Coniacian. In addition to involute inoceramids, the Middle Coniacian inoceramid assemblage (zone no. 22, Text-Fig. 3) consists of members of the *Inoceramus (Platyceramus) mantelli* lineage, of *Inoceramus frechi* FLEGEL, *Inoceramus kleini* G. MÜLLER, *Inoceramus percostatus* G. MÜLLER and inoceramids acquainted with *Inoceramus troitzkii* SCHUL-

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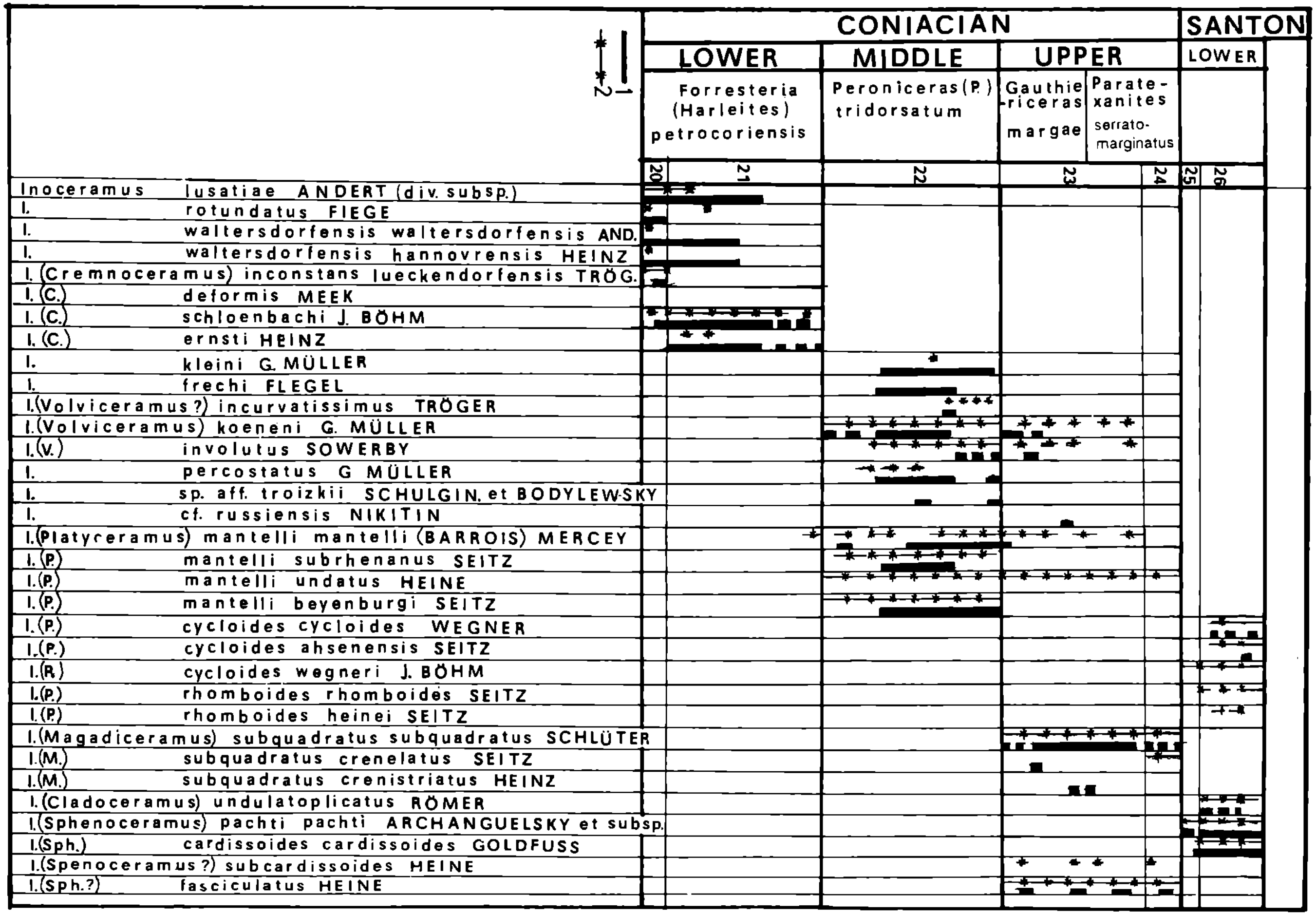
Text-Fig. 2. Diagram showing the stratigraphical distribution of the most important Turonian inoceramid species and subspecies and the zonal division of the Turonian in the North European Province including the Turonian/Coniacian boundary by using inoceramid assemblages according to F. AMEDRO, C. COLLETE et al. (1982), A. DEVRIES, C. MATHIEU et al. (1974), U. KAPLAN (1986), S. KELLER (1982), W. J. KENNEDY (1984), W. J. KENNEDY, C. W. WRIGHT & J. M. HANCOCK (1980), F. ROBASZYNSKI, G. ALCAYDE et al. (1982), J. SORNAY (1982), E. SEIBERTZ (1978, 1979), K.-A. TRÖGER (1981), C. W. WRIGHT (1979), C. W. WRIGHT & W. J. KENNEDY (1981).
1 - distribution in Middle and Eastern Europe, 2 - distribution in Western Europe.
I. - *Inoceramus*, M. - *Mytiloides*, Str. - *Striatoceramus*, H. - *Heroceramus*.


1
2
CONIACIAN
SANTON
LOWER
MIDDLE
UPPER
LOWER
Forresteria (Harleites) petrocoriensis
Peroniceras (P.) tridorsatum
Gauthiericeras margae
Parate-xanites serrato-marginatus
20
21
22
23
24
25
26
Inoceramus lusatiae ANDERT (div. subsp.)
I. rotundatus FIEGE
I. waltersdorfensis waltersdorfensis AND.
I. waltersdorfensis hannovrensis HEINZ
I. (Cremnoceramus) inconstans lueckendorfensis TRÖG.
I. (C.) deformis MEEK
I. (C.) schloenbachi J. BÖHM
I. (C.) ernsti HEINZ
I. kleini G. MÜLLER
I. frechi FLEGEL
I. (Volviceramus?) incurvatissimus TRÖGER
I. (Volviceramus) koeneni G. MÜLLER
I. (V.) involutus SOWERBY
I. percostatus G. MÜLLER
I. sp. aff. troizkii SCHULGIN et BODYLEWSKY
I. cf. russiensis NIKITIN
I. (Platyceramus) mantelli mantelli (BARROIS) MERCEY
I. (P.) mantelli subrhenanus SEITZ
I. (P.) mantelli undatus HEINE
I. (P.) mantelli beyenburgi SEITZ
I. (P.) cycloides cycloides WEGNER
I. (P.) cycloides ahsenensis SEITZ
I. (P.) cycloides wegneri J. BÖHM
I. (P.) rhomboides rhomboides SEITZ
I. (P.) rhomboides heinei SEITZ
I. (Magadiceramus) subquadratus subquadratus SCHLÜTER
I. (M.) subquadratus crenelatus SEITZ
I. (M.) subquadratus crenistriatus HEINZ
I. (Cladoceramus) undulatoplicatus RÖMER
I. (Sphenoceramus) pachti pachti ARCHANGUELSKY et subsp.
I. (Sph.) cardissoides cardissoides GOLDFUSS
I. (Spenoceramus?) subcardissoides HEINE
I. (Sph.?) fasciculatus HEINE

Text-Fig. 3

GINA & BODYLEWSKI. The basal part of the Upper Coniacian is mainly characterized by *Inoceramus (Magadiceramus) subquadratus* SCHLÜTER including subspecies, involute inoceramids and *Inoceramus (Sphenoceramus?) subcardissoides* HEINZ (zone no. 23). In the uppermost parts of the Upper Coniacian (zone no. 24) involute inoceramids are missing.

2.4 Santonian

According to T. BIRKELUND, J. M. HANCOCK et al. (1984) the appearance of the subgenus *Texanites (Texanites)* is a good marker for the Coniacian/Santonian boundary level. Yet, the subgenus is completely absent in stratigraphically important sequences in Europe, for example in the Aquitaine Basin. Other proposals are to use the appearance of *Inoceramus (Cladoceramus) undulatoplicatus* ROEMER. There is a general consensus of opinion that *Cladoceramus* and *Texanites (Texanites)* are nearly coincident (CHR. WOOD, A. SWIECICKI et al. 1984). The third possibility, which was especially remarked by O. SEITZ (1965) is the entry of members of the *Inoceramus (Sphenoceramus) pachti/cardissoides* group (Cardissoides Teilzone according to O. SEITZ 1965 = zone 25 in this paper, Text-Fig. 4). This latter possibility which is not confirmed by the ammonite biostratigraphy was used especially in the NW German - Polish Basin. For the western part of this basin O. SEITZ (1961) proposed the following division of Santonian sequences using the inoceramid faunal successions:

Pinniformis Zone Mülleri Zone (?)	} Upper Santonian
Brancoiformis Zone (?) Haenleini Zone	} Middle and Upper Santonian
Cordiinitialis Zone (?)	Middle Santonian
Cordiformis Zone	Middle and Lower Santonian
Undulatoplicatus Faunenzone	Lower Santonian
Sphenoceramen-Teilzone (= Cardissoides Teilzone)	Lower Santonian

There are some difficulties in using this scheme. First, the species *Inoceramus (Cordiceramus) cordiinitialis* SEITZ, *Inoceramus (Cordiceramus) brancoiformis* SEITZ and *Inoceramus haenleini* G. MÜLLER are rare and can therefore not be used for practical purposes. Second, there are great over-

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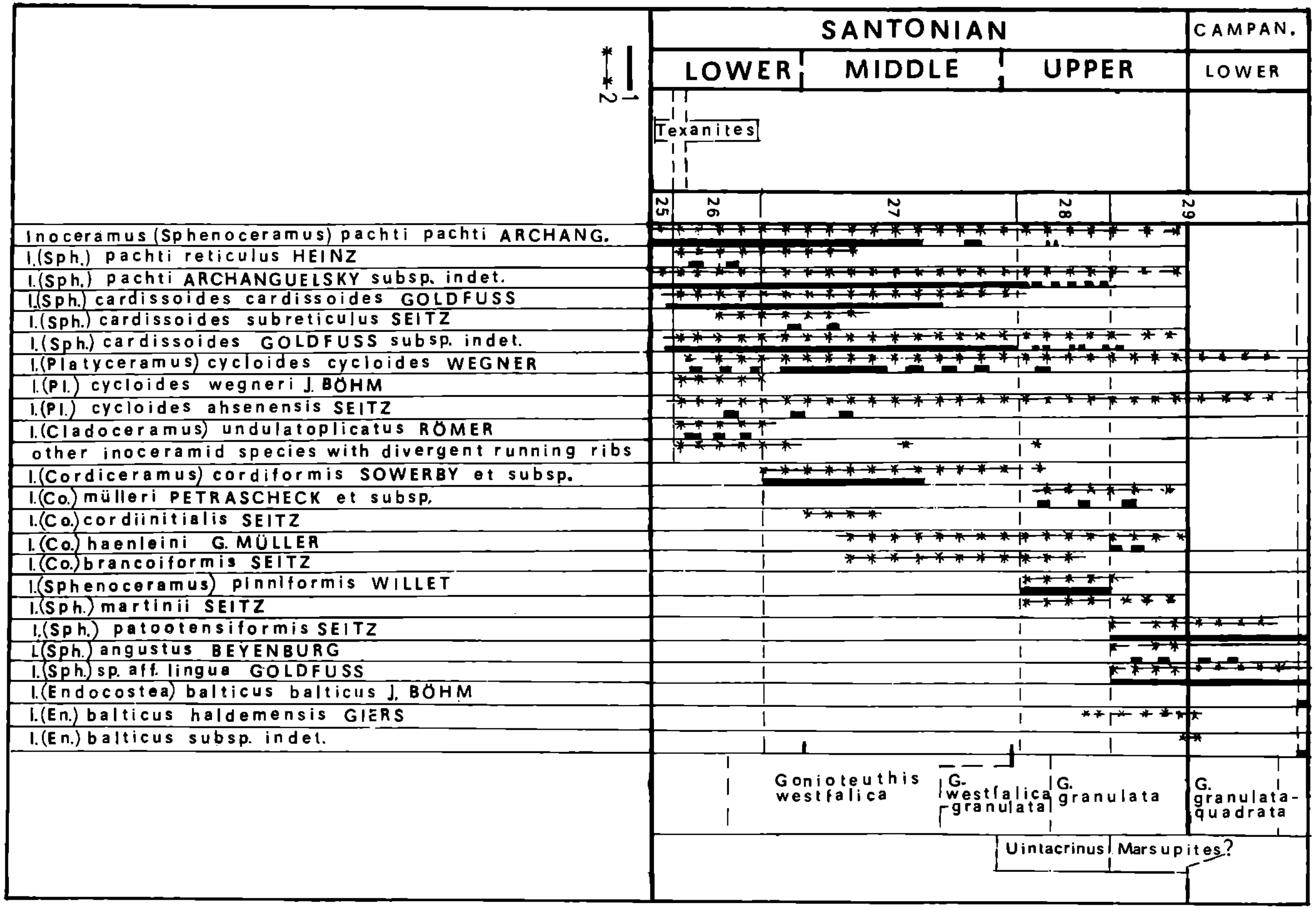
Text-Fig. 3. Diagram showing the stratigraphical distribution of the most important Coniacian inoceramid species and subspecies and the zonal division of the Coniacian in the North European Province including the Coniacian/Santonian boundary by using inoceramid assemblages according to F. AMEDRO & F. ROBASZYNSKI (1978), T. BIRKELUND, M. J. HANCOCK et al. (1984), M. COLLIGNON, E. CREGUT et al. (1979), G. ERNST (1970), S. KELLER (1982), W. J. KENNEDY (1984), O. SEITZ (1961, 1962, 1965, 1970) and K.-A. TRÖGER (1981, 1987).
1 - distribution in Middle and Eastern Europe, 2 - distribution in Western Europe.


SANTONIAN
CAMPAN.
LOWER
MIDDLE
UPPER
LOWER
1
2
Texanites
25
26
27
28
29
Inoceramus (Sphenoceramus) pachti pachti ARCHANG.
I.(Sph.) pachti reticulus HEINZ
I.(Sph.) pachti ARCHANGUELSKY subsp. indet.
I.(Sph.) cardissoides cardissoides GOLDFUSS
I.(Sph.) cardissoides subreticulus SEITZ
I.(Sph.) cardissoides GOLDFUSS subsp. indet.
I.(Platyceramus) cycloides cycloides WEGNER
I.(Pl.) cycloides wegneri J. BÖHM
I.(Pl.) cycloides ahsenensis SEITZ
I.(Cladoceramus) undulatoplicatus RÖMER
other inoceramid species with divergent running ribs
I.(Cordiceramus) cordiformis SOWERBY et subsp.
I.(Co.) mülleri PETRASCHECK et subsp.
I.(Co.) cordiinitialis SEITZ
I.(Co.) haenleini G. MÜLLER
I.(Co.) brancoiformis SEITZ
I.(Sphenoceramus) pinniformis WILLET
I.(Sph.) martinii SEITZ
I.(Sph.) patootensiformis SEITZ
I.(Sph.) angustus BEYENBURG
I.(Sph.) sp. aff. lingua GOLDFUSS
I.(Endocostea) balticus balticus J. BÖHM
I.(En.) balticus haldemensis GIERS
I.(En.) balticus subsp. indet.
Gonioteuthis westfalica
G. westfalica-granulata
G. granulata
G. granulata-quadrata
Uintacrinus
Marsupites?

Text-Fig. 4

lapping intervals between the vertical range of the mentioned species, as O. SEITZ (1961, 1965) has shown, too.

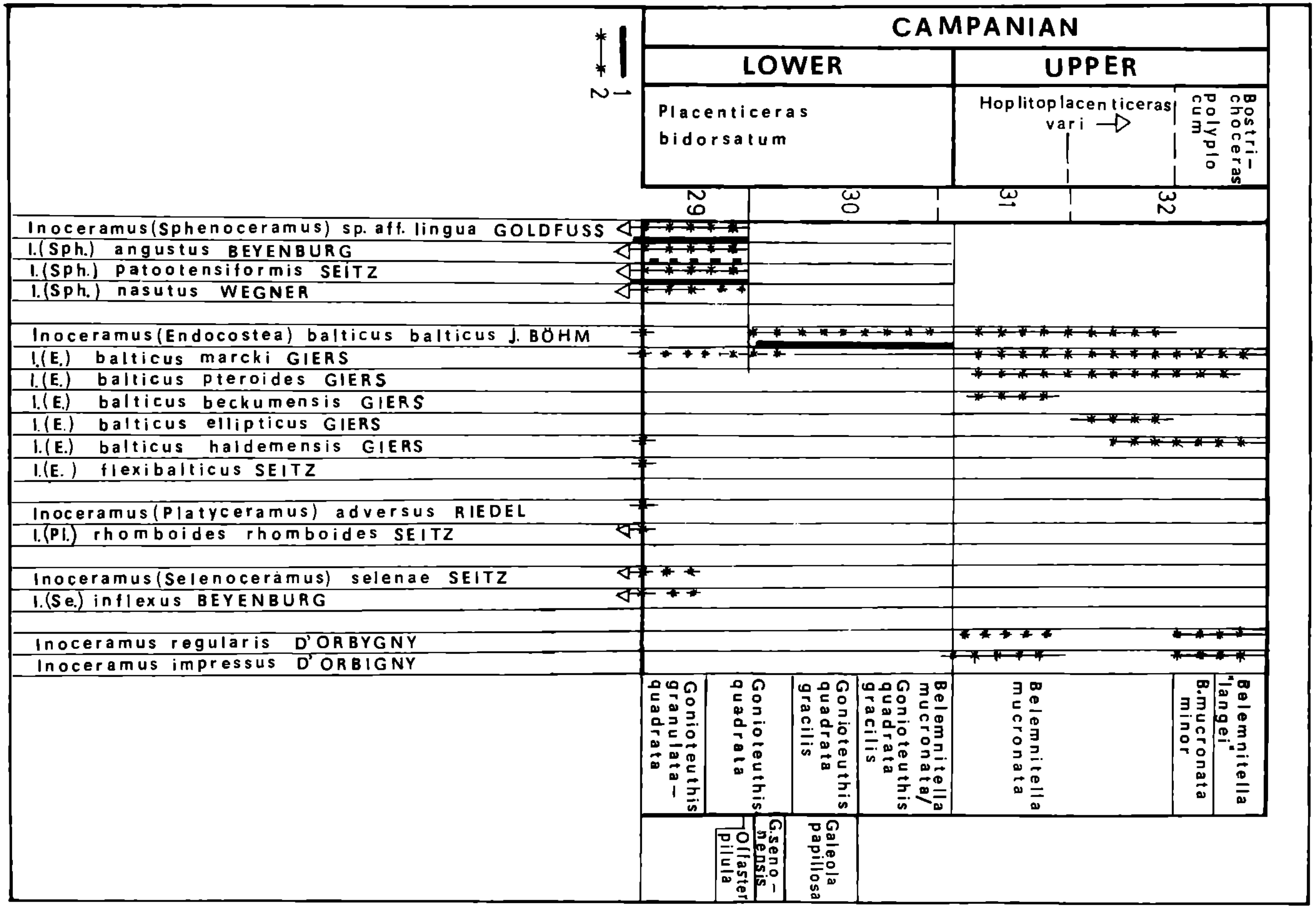
The division of the Santonian into 5 assemblage zones is used in this paper (zones no. 25-29, Text-Fig. 5). The basal zone (zone 25) yields *Inoceramus (Sphenoceramus) pachti* ARCH. and subspecies, *Inoceramus (Sphenoceramus) cardissoides* GOLDFUSS and subspecies, *Inoceramus (Sphenoceramus?) bornholmensis* n. sp. (TRÖGER & CHRISTENSEN, in press) and *Inoceramus (Platyceramus) cycloides ahsenensis* SEITZ. The assemblage zone 26 - Undulatoplicatus Faunenzone sensu O. SEITZ - mainly consists of *Inoceramus (Sphenoceramus) pachti* ARCH. and subspecies, and *Inoceramus (Sphenoceramus) cardissoides* GOLDFUSS and subspecies. These two sphenoceramids are accompanied by *Inoceramus (Cladoceramus) undulatoplicatus* RÖMER (rare) and different members of the *Inoceramus (Platyceramus) cycloides* group. The assemblage zone 27 is especially characterized by *Inoceramus (Cordiceramus) cordiformis* SOWERBY and subspecies. The upper level of this zone can be fixed by the appearance of *Inoceramus (Sphenoceramus) pinniformis* WILLET which is the key fossil for zone 28. The uppermost Santonian (zone 29, partly), especially consisting of *Inoceramus (Sphenoceramus) patootensiformis* SEITZ, *Inoceramus (Sphenoceramus) angustus* BEYENBURG and small inoceramids near to *Inoceramus (Sphenoceramus) lingua* GOLDFUSS, can be parallized with the distribution of *Marsupites* (Text-Fig. 4).

2.5 Campanian

The different definitions of the Santonian/Campanian boundary level, referring to the classical literature, are discussed by T. BIRKELUND, J. M. HANCOCK et al. (1984) and by W. J. KENNEDY (1984). In the classical definition the boundary is given by the appearance of *Placenticeras bidorsatum* (RÖMER). Yet, the species is extremely rare in the type area and the adjacent areas, too. In the NW German Basin the appearance of *Gonioteuthis granulataquadrata* (STOLLEY) was used to establish the level of the Santonian/Campanian boundary (G. ERNST 1964). It is possible that this level is nearly coincident with the extinction of *Marsupites* (see Text-Fig. 4).

By means of inoceramids it is impossible to fix the boundary. The assemblage zone 29 is extending from the Upper Santonian to the Lower Campanian with *Gonioteuthis quadrata* (BLAINVILLE). During the Campanian the development of the *Inoceramus (Endocostea) balticus* group took place.

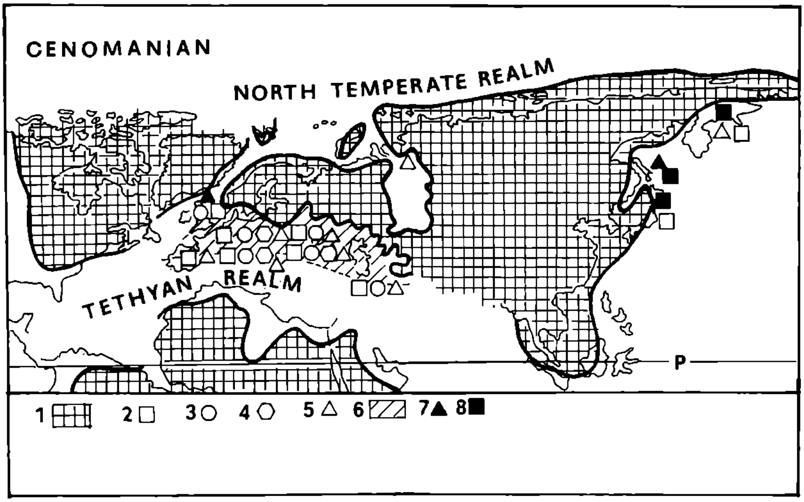
Text-Fig. 4. Diagram showing the stratigraphical distribution of the most important Santonian inoceramid species and subspecies and the zonal division of the Santonian including the Santonian/Campanian boundary. Inoceramid assemblages are used and compared with the distribution of belemnite, echinoid and crinoid species and subspecies in Western Europe, mainly in the Münsterland Basin (O. SEITZ 1961, 1965, F. ROBASZYNSKI, M. J. M. BLESS et al. 1985, G. ERNST 1964, 1970) and Middle Europe, mainly Subhercynian Cretaceous Basin (K.-A. TRÖGER & W. HALLER 1966, H. ULBRICH 1971).
1 - distribution in Middle Europe, 2 - distribution in Western Europe.


1
2
CAMPANIAN
LOWER
UPPER
Placenticeras bidorsatum
Hoplitoplacenticeras vari
Bostrichoceras polyplocum
29
30
31
32
Inoceramus(Sphenoceramus) sp. aff. lingua GOLDFUSS
I.(Sph.) angustus BEYENBURG
I.(Sph.) patootensiformis SEITZ
I.(Sph.) nasutus WEGNER
Inoceramus(Endocostea) balticus balticus J. BÖHM
I.(E.) balticus marcki GIERS
I.(E.) balticus pteroides GIERS
I.(E.) balticus beckumensis GIERS
I.(E.) balticus ellipticus GIERS
I.(E.) balticus haldemensis GIERS
I.(E.) flexibalticus SEITZ
Inoceramus(Platyceramus) adversus RIEDEL
I.(Pl.) rhomboides rhomboides SEITZ
Inoceramus(Selenoceramus) selenae SEITZ
I.(Se.) inflexus BEYENBURG
Inoceramus regularis D'ORBYGNY
Inoceramus impressus D'ORBIGNY
Gonioteuthis granulata–quadrata
Gonioteuthis quadrata
Gonioteuthis quadrata gracilis
Belemnitella mucronata/Gonioteuthis quadrata gracilis
Belemnitella mucronata
B.mucronata minor
Belemnitella "langei"
Offaster pilula
G.senonensis
Galeola papillosa

Text-Fig. 5

A further division is given by the appearance of *Inoceramus regularis* D'ORBIGNY and *Inoceramus impressus* D'ORBIGNY.

3. Paleobiogeographic remarks



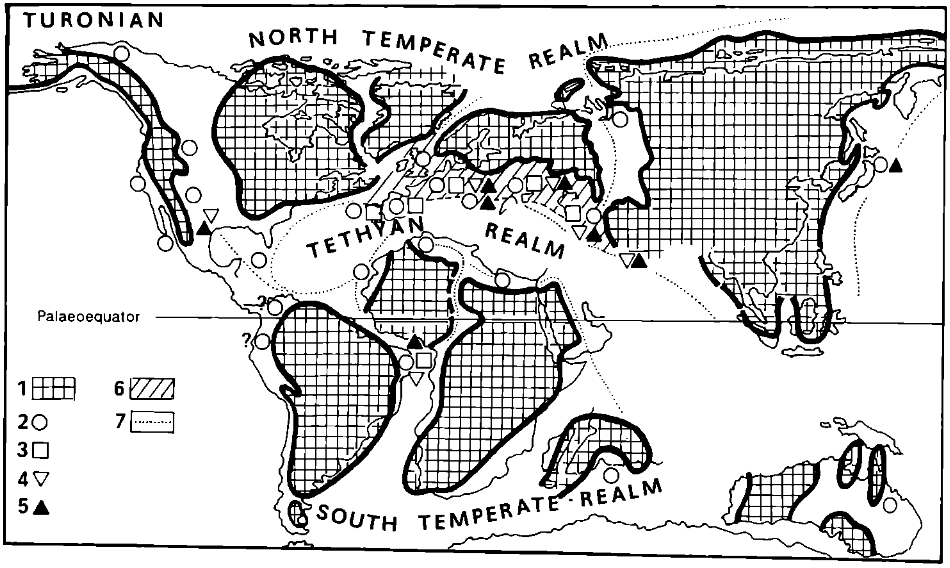
Text-Fig. 6. Distribution of the most important Cenomanian inoceramid species in Europe and Asia according to T. MATSUMOTO (1977), D. P. NAIDIN (1981), M. A. PERGAMENT (1978), J. SORNAY (1978), G. THOMEL, A. BIDAR et al. (1973) and K.-A. TRÖGER (1981). The paleogeographical reconstruction is based on the publication of A. G. SMITH, A. M. HURLEY & J. C. BRIDEN (1982).

1 land area
2 *Inoceramus crippsi crippsi* MANT. and *Inoceramus crippsi hoppenstedtensis* TRÖGER
3 *Inoceramus virgatus* SCHLÜTER and subspecies
4 *Inoceramus schoendorfi* HEINZ
5 *Inoceramus pictus* SOWERBY and subspecies
6 North European Province
7 *Inoceramus ginterensis* PERGAMENT
8 *Inoceramus pennatulus* PERGAMENT
P = Paleoequator

Text-Fig. 5. Diagram showing the stratigraphical distribution of inoceramid species and subspecies in Campanian layers by using the publications of G. ERNST (1964, 1970), O. SEITZ (1967, 1970), F. ROBASZYNSKI, M. J. M. BLESS et al. (1985) and H. ULBRICH (1971).
1 - distribution in Middle and Eastern Europe, 2 - distribution in Western Europe.

The North European Province established by E. G. KAUFFMAN (1973) on the base of bivalve molluscs is situated at the northern border of the Tethyan Realm (Text-Fig. 7). This North European Province is partly separated from the Tethyan region by a string of small islands which are not visible on the Text-Figs. 5 - 7. The greatest part of the mentioned inoceramid species and subspecies is distributed laterally over the whole North European Province. Nevertheless, there are small differences between the inoceramid assemblages of the western and eastern part of the North European Province.

The Lower Coniacian inoceramid species *Inoceramus (Volviceramus) wandereri* ANDERT having been described firstly from the North Bohemian

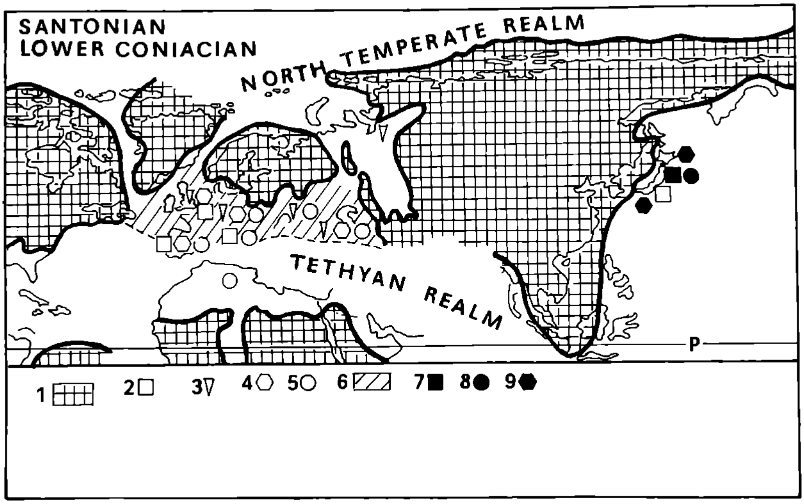


Text-Fig. 7. Global distribution of some Lower and Upper Turonian inoceramid species and subspecies according to E. M. ARSUMANOVA (1965), F. AMEDRO, H. MANIVIT et al. (1978), K. AYYASAMI & RANJIT K. BANERJI (1984), P.-Y. BERTHOU (1984), M. M. RIBEIRO HESSEL (1987), D. LUPU (1976), T. MATSUMOTO (1975, 1977), D. P. NAIDIN (1981), M. A. PERGAMENT (1978), R. A. REYMENT (1980), J. SORNAY (1972, 1981), L. SZASZ (1982) and K.-A. TRÖGER (1981, 1987). The paleogeographical reconstruction is based on the publication of A. G. SMITH, A. M. HURLEY & J. C. BRIDEN (1982).

1 land area
2 *Inoceramus (Mytiloides) submytiloides* SEITZ, *Inoceramus (Mytiloides) mytiloides* MANTELL, *Inoceramus (Mytiloides) labiatus* (v. SCHLOTHEIM)
3 *Inoceramus hercynicus* PETRASCHECK
4 *Inoceramus dresdensis* TRÖGER and subspecies
5 *Inoceramus (Mytiloides) incertus* JIMBO and subspecies
6 North European Province
7 pathway of migration

Cretaceous, is extremely rare in the western part of the North European Province. In the eastern part of the province the species is widespread in the Donbass-Crimea-Caucasus area and is used as a key fossil for the Lower Coniacian. *Inoceramus (Cremnoceramus) gradatus* EGOJAN is only distributed in the eastern part of the North European Province showing the possibilities of small faunal differentiations within this province.

Studying the inoceramid assemblages of the North American Province, of the Mediterranean Province (Tethyan Realm), of the South Atlantic Subprovince and of the Japan-East Asia Subprovince, there are some similarities in the composition of the inoceramid assemblages especially during Upper Cenomanian - Lower Turonian, Upper Turonian - Lower Coniacian and Lower Santonian. This fact has been mentioned, too, by E. G. KAUFFMAN (1973) and by M. A. PERGAMENT (1971, 1974) as is visible on the Text-Figs. 6-8. This corresponds with global eustatic cycles during the late Cretaceous which were considered to be related to pulses of plate movements, especially to Atlantic seafloor spreading and subsidence of ocean ridges. The late



Text-Fig. 8. Distribution of some Upper Coniacian and Santonian inoceramid species according to M. JA. BLANK, G. JA. KRIMGOLTZ et al. (1974), A. V. IVANNIKOV (1979), G. LOPEZ (1986), D. LUPU (1976), M. M. MOSKVIN (1959), M. A. PERGAMENT (1974, 1978), O. SEITZ (1961, 1965), J. SORNAY (1982), G. THOMEL, A. BIDAR et al. (1973).

1 land area
2 *Inoceramus (Magadiceramus) subquadratus* SCHLÜTER and subspecies
3 *Inoceramus (Sphenoceramus) pachti* ARCH. including subspecies, and *Inoceramus (Sphenoceramus) cardissoides* GOLDFUSS including subspecies
4 *Inoceramus (Cladoceramus) undulatoplicatus* RÖMER and subspecies
5 *Inoceramus (Platyceramus) cycloides* WEGNER and subspecies
6 North European Province
7 *Inoceramus mihoensis* MATSUMOTO
8 *Inoceramus amakusensis* NAGAO & MATSUMOTO
9 *Inoceramus japonicus* NAGAO & MATSUMOTO

Albian-Cenomanian-Lower Turonian transgressions reached their maximum in the Upper Cenomanian/Lower Turonian (widespread dispersal of *Inoceramus pictus* SOWERBY (Text-Fig. 6) and *Inoceramus (Mytiloides) labiatus* lineage (Text-Fig. 7)). They were followed by a regression in the Middle Turonian causing widespread breaks mainly in eastern Europe. The Coniacian -Santonian transgressions were initiated in the late Turonian (widespread dispersal of *Inoceramus (Mytiloides) incertus* JIMBO and *Inoceramus dresdensis* TRÖGER in both the northern and southern hemisphere). The North European Province was also influenced by immigration of different species (for example *Inoceramus russiensis* NIKITIN and *Inoceramus troitzkii* SCHULGINA & BODYLEWSKI) from the north Jenissei area belonging to the North Temperate Realm. All possible migration pathways are visible on Text-Fig. 7.

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Biostratigraphy of the Cenomanian of NW Germany

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With 4 Text-Figures and 2 Tables

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Abstract: Intensified collecting in the Cenomanian "Normalfazies" of the Osning Range (northern Westphalia) allows
(1) a more precise subdivision of the North German Cenomanian by means of ammonites and inoceramids,
(2) a better comparison with the existing subdivision based on planktonic foraminifera,
(3) a better delimitation of the North German Cenomanian,
(4) its better comparison with the standard subdivision of the Anglo-Paris Basin, and
(5) a better correlation of North Temperate and Tethyan ammonite succession.
From the southern marginal facies is reported:
(1) the recognition of *Hypoturrilites carcitanensis* in the transgressive early Cenomanian of Kassenberg, Mülheim-Broich, and
(2) the discovery of late Albian ammonites in the "Cenomanian" Essen Grünsand of Bergkamen. This means that - at least locally - the "Cenomanian" transgression in the southern Münster Basin was of late Albian age.

Kurzfassung: Intensivierte Neuaufsammlungen in der "Normalfazies" des Cenoman entlang der Osning-Kette, Nord-Westfalen, ermöglichen
(1) eine präzisere Gliederung des norddeutschen Cenoman mit Ammoniten und Inoceramen,
(2) einen besseren Vergleich mit der bereits vorliegenden Gliederung auf der Grundlage planktonischer Foraminiferen,
(3) eine bessere Abgrenzung des norddeutschen Cenoman,
(4) seinen besseren Vergleich mit der Standardgliederung des Anglo-Pariser Beckens und
(5) eine bessere Korrelation von borealer und tethydischer Ammonitenfolge.
Von der südlichen Randfazies ist zu berichten
(1) der Nachweis von *Hypoturrilites carcitanensis* im transgressiven Cenoman-Rotkalk des Kassenbergs, Mülheim-Broich, und
(2) das Vorhandensein von Alb-Ammoniten im "cenomanen" Essener Grünsand von Bergkamen. Das bedeutet, daß die "Cenoman"-Transgression auf die Rheinische Masse auch im S' Münsterland lokal bereits im höheren Alb einsetzte.

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1. Introduction

Continued collecting efforts in the "Normalfazies" of the Cenomanian along the Osning, northern Westphalia (Text-Fig. 1, localities 4-11) allow a better definition and subdivision of this stage in Northern Germany. These investigations are based on ammonites and inoceramids and compared with planktonic foraminifera (WEISS 1982).

The so-called Normalfazies consists of a marl-limestone sequence (Text-Fig. 2) which was deposited under continuous neritic conditions which can be followed from Rheine in the west to Schwaney in the east, continuing even into the easternmost Lower Saxony Basin (Text-Fig. 1, localities 13-16). It contrasts, however, to the more southern condensed greensand facies ("Essener Gründsand") which is typical for the marginal areas of the Westphalian Basin (Text-Fig. 1, localities 2, 3, 12, 13). Moreover, at the Kassenberg section, Mülheim-Broich, a very thin red limestone layer is deposited on the Carboniferous of the Rhenish Massif at the Cenomanian shoreline.

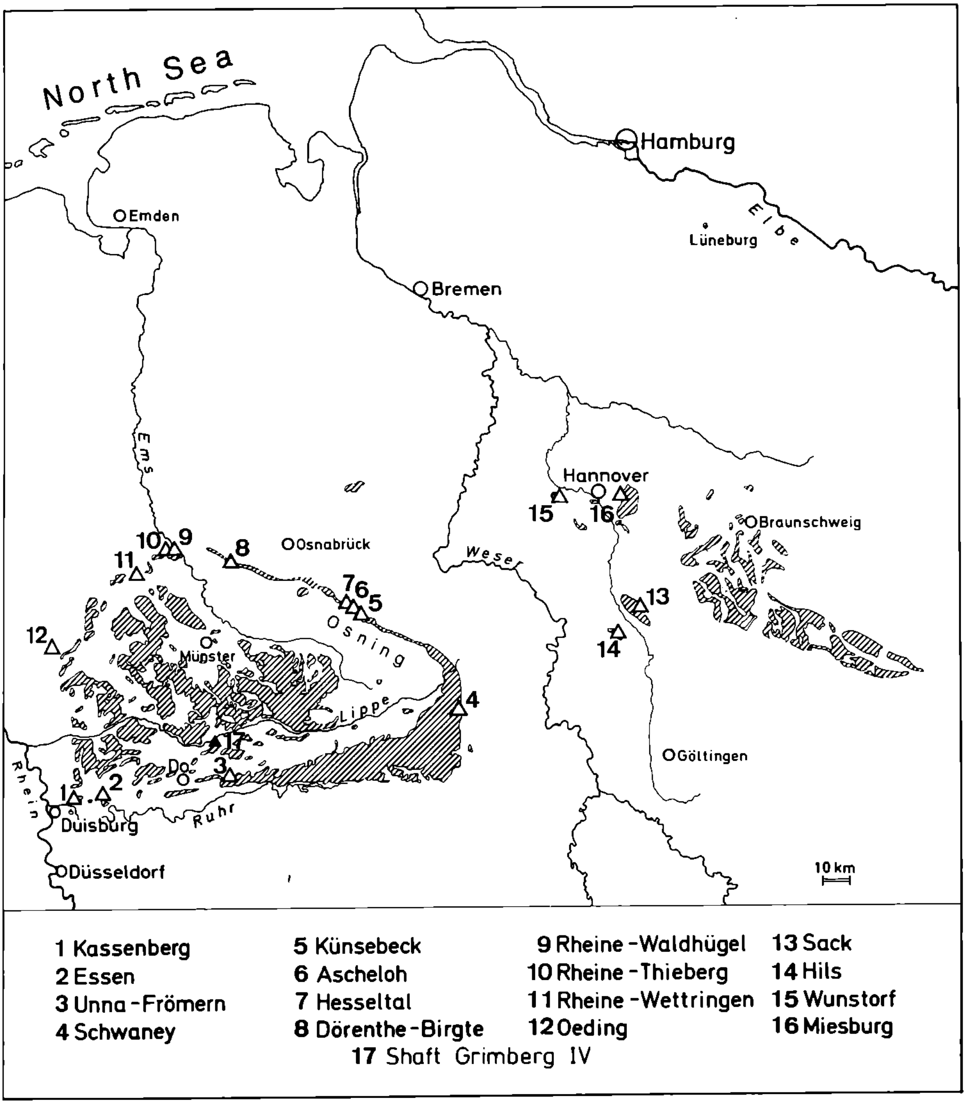
The well-preserved ammonite fauna of the glauconitic Cenomanian of southern Westphalia was described by SCHLÜTER (1871-1876) and HISS (1982); it is, however, stratigraphically meaningless. The likewise well-preserved cephalopod fauna of the Kassenberg was first recognized by HANCOCK et al. (1972) and later described in detail by WIEDMANN & SCHNEIDER (1979). Unanimously it was considered to represent the base of the Cenomanian. Despite the absence of the index species it was attributed to the Zone of Hypoturrilites carcitanensis by HANCOCK et al. (1972: 447). WIEDMANN & SCHNEIDER (1979: 673) proposed to re-establish the Zone of Utaturiceras vicinale which was originally erroneously referred to the late Cenomanian by SPATH (1926) and WRIGHT (1957). Meanwhile, *Hypoturrilites carcitanensis* (MATH.) has been found at Kassenberg (Text-Fig. 3) and can now be added to this interesting fauna.

The Kassenberg fauna is of importance for the present purpose since the equivalent level, i. e. the base of the Cenomanian and the top of the Albian, is actually not exposed in the Normalfazies area. Further disadvantages of the Normalfazies are the comparatively poor ammonite preservation and the scarcity of the ammonite record.

After SCHLÜTER's monograph (1871-1876), a first illustration of ammonite and inoceramid faunas of the Normalfazies was given by KAPLAN et al. (1985). Now, some new information can be added.

2. Cenomanian Normalfazies

In Text-Fig. 2, a simplified and generalized section of the Normalfazies of the Osning Range is reproduced. It is mainly based on the quarries at Rheine-Waldhügel which actually offer the most complete section with some tectonics restricted to its central area. As mentioned above, equi-



Text-Fig. 1. Map of Upper Cretaceous outcrops and most important Cenomanian localities in North Germany.

valents of the Kassenberg red limestone are actually unexposed in the Normalfazies. From mapping and boreholes we can estimate that the unexposed portion of early Cenomanian has a thickness of about 20 m and a facies similar to the marls outcropping above (KEMPER 1984, KAEVER & BECKER 1985). The Kassenberg red limestone and its index ammonites are provisionally placed at the base of the standard section to fill the respective gap.

From base to top the Normalfazies exhibits the following simplified section:

40 to 50 m of grey, thin-layered marls with only subordinate individual limestone beds or nodular layers; this lower portion represents the upper part of the Mantelli Zone.

35 m of grey, flaser-bedded marls and limestones; thickness of beds varies between 0.10 and 0.20 m. Stratigraphically these beds represent the Dixoni and the lower part of Rhotomagense zones. Synsedimentary events are documented in individual layers below and at the Lower to Middle Cenomanian boundary. At this boundary a 1 m thick marly layer seems to be consistent. It is followed by 7 to 8 m of thick-bedded yellowish marly limestones including the "*Actinocamax primus* Event" of various authors (e. g. DAHMER et al. 1986).

40 to 45 m of grey, well-bedded limestones with thicknesses of beds varying between 0.20 and 0.40 m and very thin marly interlayers. Represented are the upper Rhotomagense, Acutus, Jukesbrownei, and Pentagonum zones. Near the Middle to Upper Cenomanian boundary two marly layers are separated by a 7 m thick limestone, the lower one containing *Pycnodonte baylei,* the upper *Amphidonte*.

Finally, the top of the Cenomanian and the early Turonian are developed at Rheine by 10 to 15 m of red clays, marls, and thin-layered limestones alternating with a few greenish-grey marly layers ("Rotpläner"). Only the lower 5 m are equivalent to the late Cenomanian *Plenus* Marl, i. e. the zones of Metoicoceras geslinianum and the following of Neocardioceras juddii.
Towards the north and south-east, the Rotpläner is substituted by a similar alternation of black and grey marls of greater thickness. The Turonian part of both sequences can be identified by mass occurrences of *Inoceramus (Mytiloides) mytiloides* MANTELL and *I. (M.) labiatus* (v. SCHLOTH.), and rare *Watinoceras* sp.

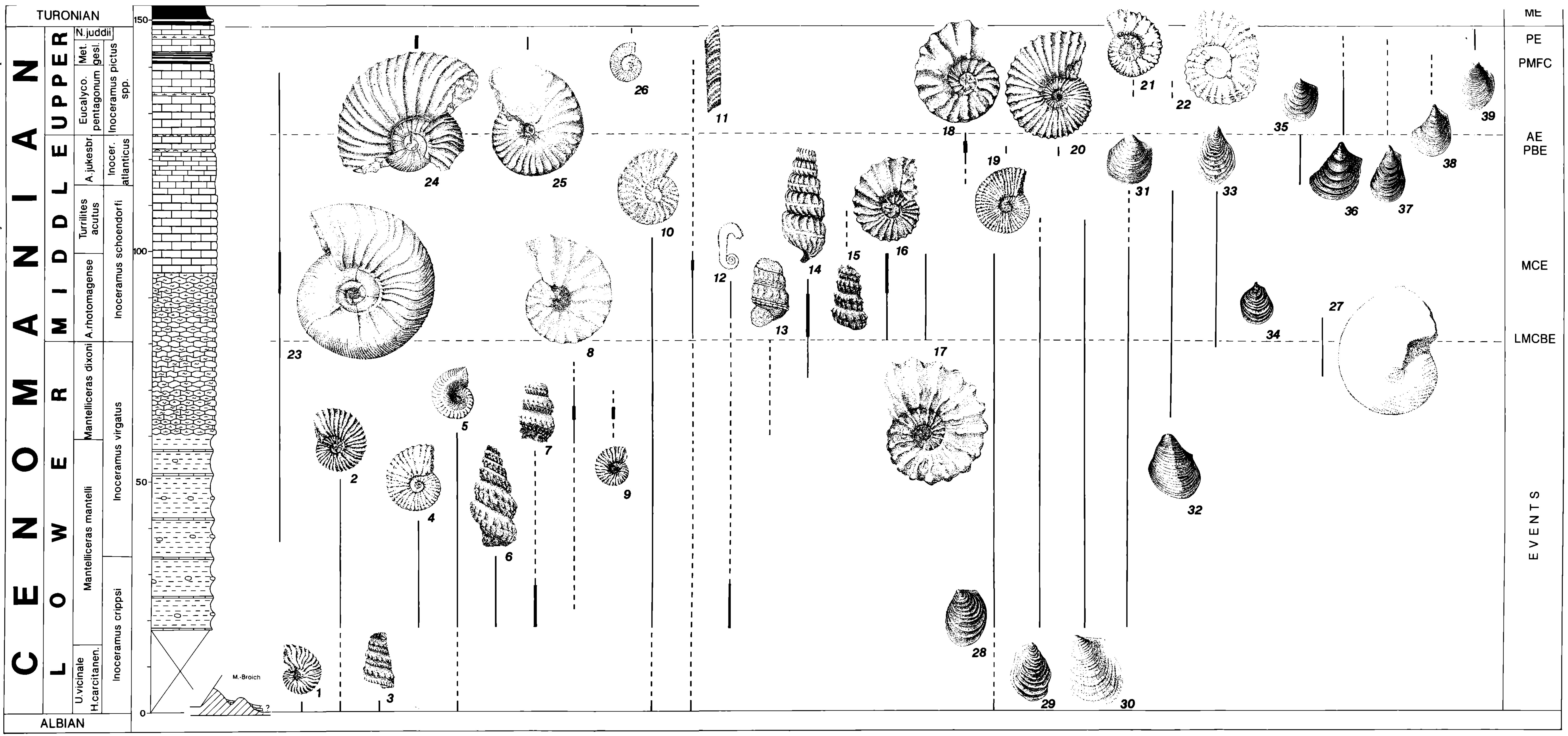
In summary it can be said that there is a continuously increasing substitution of marls by limestones through the Cenomanian except for the *Plenus* Marls. It is correlated with a similar increase in thickness of marly layers or limestone beds, respectively.

3. Biostratigraphy

3.1 Ammonites

During the last years, increasing collecting effort brought to light many of the ammonite species attributed to the North Temperate Realm and the Anglo-Paris Basin in uncondensed succession. The most important species of this succession are reproduced in Text-Fig. 2.

In describing the ammonite faunas of the British Lower and Middle Chalk, WRIGHT & KENNEDY (1981, 1984) proposed the following Cenomanian zonation (Table 1). This zonation can be adopted with minor changes (see Text-Fig. 2).



Text-Fig. 2. Generalized Cenomanian section of Osning "Normalfazies" (northern Westphalia) and biostratigraphic frame based on ammonites, inoceramids, and events.
1. *Utaturiceras vicinale* (STOL.) - 2. *Mantelliceras saxbii* (SHARPE) - 3. *Hypoturrilites carcitanensis* (MATH.) - 4. *Mantelliceras mantelli* (J. SOW.) - 5. *Hyphoplites falcatus* (MANTELL) - 6. *Mariella cenomanensis* (SCHLÜTER) - 7. *Hypoturrilites tuberculatus* (BOSC) - 8. *Acompsoceras renevieri* (SHARPE) - 9. *Mantelliceras dixoni* SPATH - 10. *Schloenbachia varians* (J. SOW.) - 11. *Sciponoceras baculoide* (MANTELL) - 12. *Worthoceras* sp. - 13. *Turrilites boerssumensis* SCHLÜTER - 14. *Turrilites costatus* LAM. - 15. *Turrilites acutus* PASSY - 16. *Acanthoceras rhotomagense* (BRONGN.) - 17. *Cunningtoniceras inerme* (PERV.) - 18. *Acanthoceras jukesbrownei* (SPATH) - 19. *Eucalycoceras rowei* (SPATH) - 20. *Calycoceras newboldi* (KOSSM.) - 21. *Lotzeites aberrans* (KOSSMAT) - 22. *Thomelites* sp. - 23. *Austiniceras austeni* (SHARPE) - 24. *Pachydesmoceras denisonianum* (STOL.) - 25. *Metoicoceras geslinianum* (D'ORB.) - 26. *Neocardioceras juddii* (BARROIS & GUERNE) - 27. *Forbesiceras* sp. - 28. *Inoceramus crippsi crippsi* MANTELL - 29. *I. crippsi hoppenstedtensis* TRÖGER - 30. *I. virgatus virgatus* SCHLÜTER - 31. *I. virgatus scalprum* BÖHM - 32. *I. tenuis* MANTELL - 33. *I. schoendorfi* HEINZ - 34. *I. "tenuistriatus"* sensu KELLER - 35. *I. atlanticus* (HEINZ) - 36. *I. pictus neocaledonicus* JEANNET - 37. *I. pictus pictus* J. SOW. - 38. *I.* sp. aff. *pictus concentricoundulatus* TRÖGER - 39. *I. pictus bohemicus* LEONH.
Events: LMCBE - Lower/Middle Cenomanian boundary event - MCE - Mid-Cenomanian event - PBE *Pycnodonte baylei* event - AE - *Amphidonte* event (this must be placed 4 m higher up in the section) - PMFC - *Plenus* Marl facies change - PE - *Pachydesmoceras* event - ME - *Mytiloides* event.

Table 1. Cenomanian ammonite zones used by WRIGHT & KENNEDY (1981, 1984).

SUBSTAGE	ZONE	SUBZONE
LOWER TURONIAN (PART)	*Watinoceras coloradoense*	
UPPER CENOMANIAN	*Neocardioceras juddii* *Metoicoceras geslinianum* *Calycoceras guerangeri*	
MIDDLE CENOMANIAN	*Acanthoceras jukesbrownei* *Acanthoceras rhotomagense*	 *Turrilites acutus* *Turrilites costatus*
LOWER CENOMANIAN	*Mantelliceras dixoni* *Mantelliceras mantelli*	 *Mantelliceras saxbii* *Neostlingoceras carcitanense*
UPPER ALBIAN (PART)	*Stoliczkaia dispar*	*Mortoniceras (Durnovarites) perinflatum*

a) Albian/Cenomanian boundary, Zone of Utaturiceras vicinale/Hypoturrilites carcitanensis

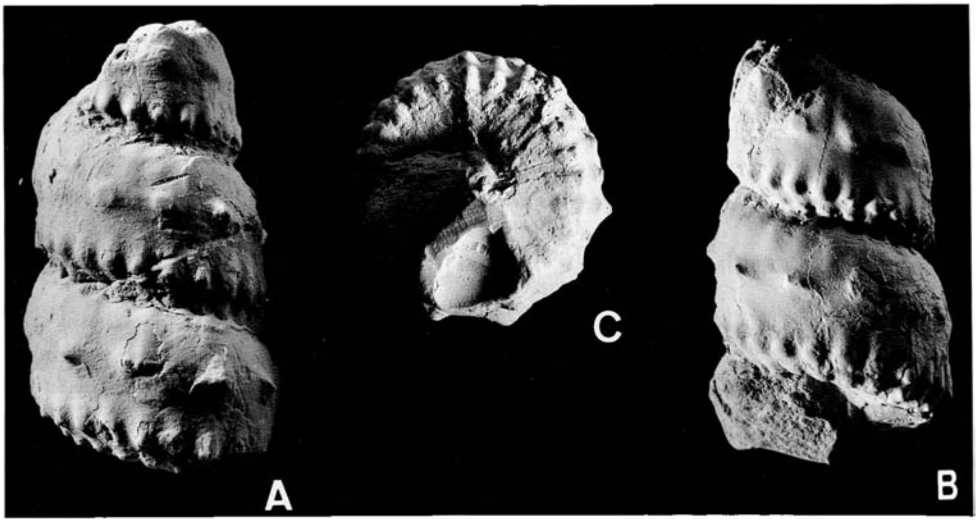
As indicated above, neither the Albian/Cenomanian boundary nor the lowermost Cenomanian are actually exposed along the Osning Range, i. e. in the Normalfazies.

WIEDMANN & SCHNEIDER (1979) recognized - in describing the Kassenberg cephalopods - the existence of *Hypoturrilites* and *Hyphoplites* species transitional between late Albian and early Cenomanian. This was the reason to consider the red limestone from Kassenberg to represent the lowermost Cenomanian. Furthermore the first proof of Indo-Malgache species such as *Utaturiceras vicinale* (STOL.) and *Mantelliceras lateretuberculatum* COLL. in a European section, was not explained by weakening of faunal provincialism but rather with a depositional gap at this boundary in large areas of Europe (SCHOLZ 1973).

Utaturiceras vicinale (STOL.) was therefore proposed as index-species for lowermost Cenomanian, especially since *Hypoturrilites carcitanensis* (MATH.) was not yet known from Kassenberg. This zone was compared with the "*Submantelliceras*" level of NW Africa (PERVINQUIERE 1907, 1910), and with the *Graysonites* beds in Texas (YOUNG 1958, MANCINI 1979) and Spain (WIEDMANN 1960, 1980).

Meanwhile, *Hypoturrilites carcitanensis* was found at Kassenberg (Text-Fig. 3) and both zones might thus be time-equivalent; however, we need more information from this earliest Cenomanian succession. Noteworthy is the presence of *Mantelliceras saxbii* (SHARPE) in the same level. This species is therefore abandoned as a zonal marker in the Lower Cenomanian (Table 1). But it is important to remember that *Mantelliceras mantelli* (J. SOW.) is still absent at Kassenberg[1]. We would like to note the first ap-

[1] WRIGHT & KENNEDY (1984: 99) considered *Mantelliceras tuberculatum* (MANTELL) to be synonymous with *M. mantelli*. In consequence, this would have implications for separating the Carcitanensis and Mantelli zones. *M. tuberculatum* is a common species of the Carcitanensis Zone and is also present in the Kassenberg fauna (WIEDMANN & SCHNEIDER 1979).

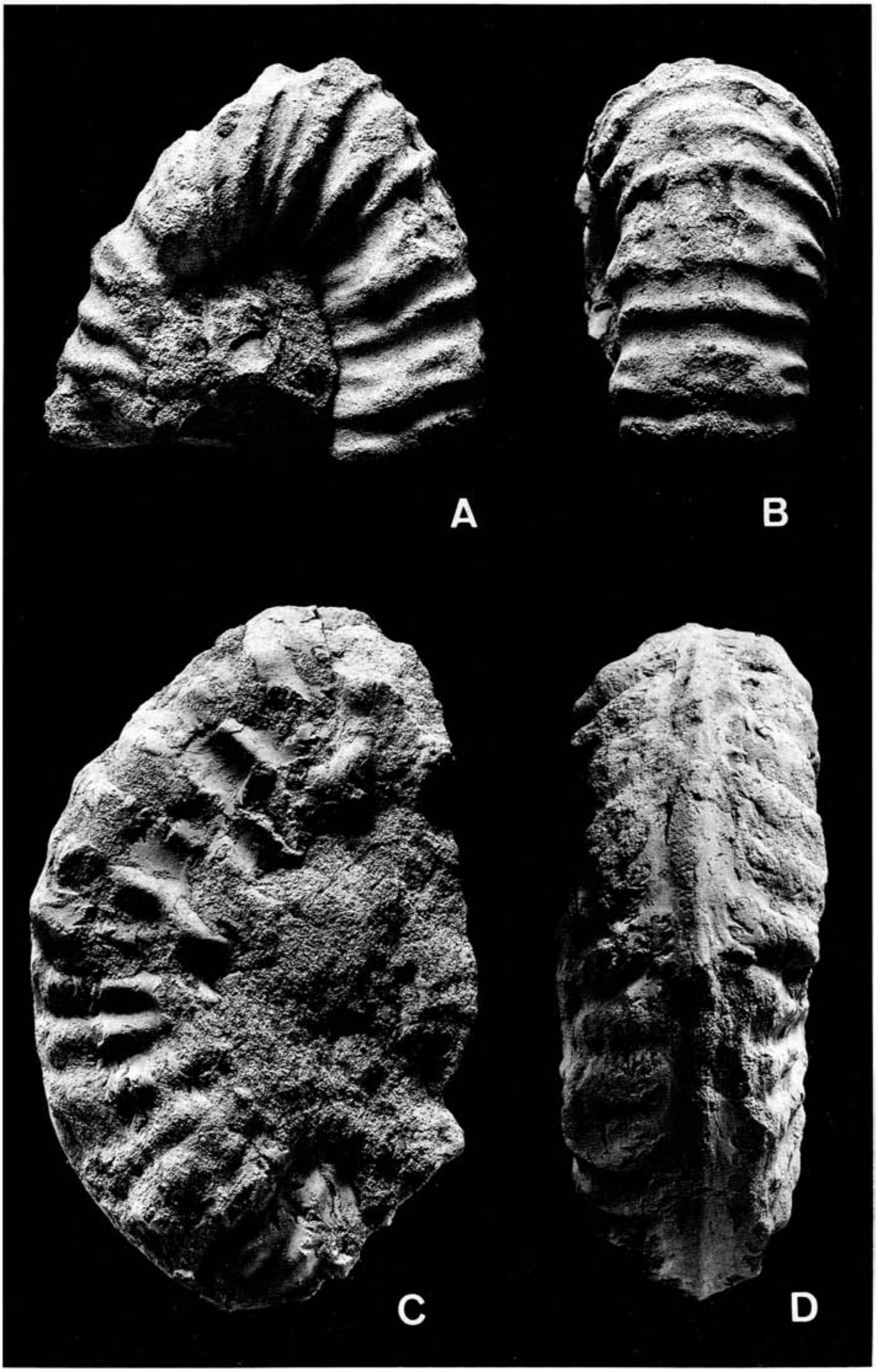


Text-Fig. 3. *Hypoturrilites carcitanensis* (MATHERON).
A, B. Lateral views. C. Basal view. 1.5 times. Zone of Utaturiceras vicinale/ Hypoturrilites carcitanensis. Red limestone, Kassenberg, Mülheim-Broich (Westph.). Coll. Hilpert/Datteln.

pearance of the genus *Lewesiceras* (*L. cenomanense* WIEDM. & SCHN.), the presence of a few more heteromorphs, of *Schloenbachia varians* (J. SOW.) with its broad variability in shell form and sculpture, and the uncommon abundance of nautiloids.

While there is a continuous marine transition from Albian "Flammenmergel" to the Cenomanian marls in the Osning area of Normalfazies as well as in the eastern Münster Basin (KAEVER & JORDAN 1985), at the western and southern margins of the Westphalian Basin the basal Cenomanian rests unconformably on the Carboniferous. As mentioned above, it is formed as a thin layer of red limestone at the Kassenberg paleo-shoreline, but as a calcareous greensand ("Essen Grünsand") at the southern margin. In accordance with the classic monograph of SCHLÜTER (1871-1876), HISS (1982) was able to show that the base of the Essener Grünsand transgression is of Carcitanensis age, that the Grünsand mostly comprises the Lower Cenomanian, but locally the Middle and sometimes the Upper Cenomanian also.

Text-Fig. 4. Late Upper Albian ammonites from the "Cenomanian" Essen Grünsand. Shaft Grimberg IV, near Bergkamen (Westph.).
A, B. *Anisoceras (A.) perarmatum* PICT. & CAMP. Lateral and ventral view. 1/1. Coll. Westf. Berggewerksch.-kasse Bochum (WBKB) no. 1.
C, D. *Mortoniceras (M.) rostratum* (J. SOW.). Lateral and ventral view. 1/1. Coll. WBKB no. 2.


A
B
C
D

In revising older collections of Essen Grünsand ammonites, it was very surprising to note the presence of late Upper Albian species, i. e.

Mortoniceras (M.) rostratum (J. SOW.) - Text-Fig. 4 C, D
and *Anisoceras (A.) perarmatum* PICT. & CAMP. - Text-Fig. 4 A, B

from shaft Grimberg IV near Bergkamen (Text-Fig. 1, locality 13). These forms indicate that the "Cenomanian" transgression on the Rhenish Massif started also in this area - at least locally - with the late Albian.

b) Zone of Mantelliceras mantelli

The basal 40 to 50 m of exposed Normalfazies can be referred to the Mantelli Zone. The oldest known fauna has been collected at Rheine and Ascheloh. It contains - as usual in the marly facies - poorly preserved specimens, mainly *Mantelliceras mantelli* (J. SOW.) and *M. saxbii* (SHARPE). In the Osning sections the latter seems to continue after the disappearance of the index species, but both disappear distinctly before the first appearance of *Mantelliceras dixoni* SPATH. No significant species is actually available to characterize the upper Mantelli Zone. As mentioned above, the Saxbii Zone cannot be maintained, especially since the index species first appears at the base of the Cenomanian.

Acompsoceras species [*A. renevieri* (SHARPE), *A. sarthense* (GUER.), *A. essendiense* (SCHLÜT.)] range through the total zone and have a maximum at the top. The presence of large specimens of *Hypoturrilites tuberculatus* (BOSC) and the common occurrence of *Worthoceras* are typical for the Mantelli Zone, during which the maximum of ammonite diversity is achieved for the Lower Cenomanian.

Further ammonite species are: *Hypophylloceras seresitense* (PERV.), *Hyphoplites falcatus* (MANTELL), *Schloenbachia varians* (J. SOW.), *Mantelliceras cantianum* SPATH, *M. costatum* (MANTELL), *M. tuberculatum* (MANTELL), *Forbesiceras sculptum* CRICK, *Sciponoceras baculoide* (MANTELL), *Hamites simplex* D'ORB., *Anisoceras* cf. *plicatile* (J. SOW.), *Ostlingoceras* sp., *Mariella cenomanensis* (SCHLÜTER), *M. essenensis* (SCHLÜTER), *Turrilites (T.) scheuchzerianus* BOSC, *Scaphites equalis* J. SOW., *Sc. obliquus* J. SOW.

c) Zone of Mantelliceras dixoni

The base of the Dixoni Zone coincides approximately with the facies change from the thin-layered basal marls to flaser-bedded marls and limestones typical for the central portion of the Cenomanian. 20 m of this sequence are considered to represent the Dixoni Zone.

In the lower marly part of the Dixoni Zone the maximum occurrence of *Mantelliceras dixoni* SPATH is found simultaneously with *Acompsoceras renevieri* (SHARPE), *A. sarthense* (GUER.) and *A. essendiense* (SCHLÜT.). *Hyphoplites falcatus* is still present in the lower Dixoni Zone. *Schloenbachia varians* (J. SOW.), *Scaphites equalis* J. SOW., and *Sc. obliquus* J. SOW. were found throughout. *Forbesiceras* spp. have a maximum occurrence near the Lower/Middle Cenomanian boundary. The range of *Turrilites (T.) boerssumensis* SCHLÜTER coincides with that of the Dixoni Zone. *Turrilites (T.)*

costatus LAM. appears in the upper portion of the Dixoni Zone and is therefore inadequate to characterize the lower Middle Cenomanian.

d) Zone of Acanthoceras rhotomagense

KENNEDY (1969) proposed a tripartite subdivision of the Middle Cenomanian Rhotomagense Zone of which the youngest *Acanthoceras jukesbrownei* assemblage was later considered to represent a separate zone (WRIGHT & KENNEDY 1984). The lower two assemblages of *Turrilites costatus* and *Turrilites acutus* were later regarded (WRIGHT & KENNEDY 1984) to be subzones of the Rhotomagense Zone. Due to the first appearance of *Turrilites (T.) costatus* LAM. in the Dixoni Zone, at least this zone cannot be maintained. *Turrilites (T.) acutus* PASSY, however, can be used as a zonal marker for the middle part of the Middle Cenomanian although it is very rare.

The Rhotomagense Zone is here understood to comprise the upper 15 m of flaser-bedded marly limestones at the base and 5 m of the well-bedded limestones at the top (Text-Fig. 2). It starts with a 1 m thick marly layer followed by 7 to 8 m thick-bedded yellowish marly limestones of the "*Primus*-Event". This sequence is again characterized by a peak occurrence of ammonites. The index species becomes, however, frequent towards the facies change and disappears very quickly in all the Normalfazies sections. *Cunningtoniceras inerme* (PERV.) seems to have a comparable range but is rather rare. *Turrilites (T.) costatus* LAM. becomes abundant in the lower Rhotomagense Zone together with *Anisoceras plicatile* (J. SOW.), *Sciponoceras baculoide* (MANTELL), and *Forbesiceras* sp. *Austiniceras austeni* (SHARPE) is attaining a maximum together with that of *Acanthoceras rhotomagense* (BRONGN.). *Turrilites (T.) scheuchzerianus* BOSC and *Schloenbachia varians* (J. SOW.) are decreasing in density. *Worthoceras* also has a last peak occurrence. The scaphitid species persist together with *Puzosia subplanulata* (SCHLÜTER), *Acompsoceras sarthense* (GUER.), and others.

e) Zone of Turrilites acutus

Due to the early disappearance of *Acanthoceras rhotomagense* in the Osning sections, the subsequent 15 m of well-bedded limestones might be considered - in accordance with the English authors - to form a Zone of Turrilites acutus. Near the base of the zone, a sedimentary "non-sequence" was recorded by ERNST et al. (1983).

These limestones are very rare in fossils; therefore the absence of *Acanthoceras rhotomagense* may be a preservational problem, because the species persists in England into the lowermost Upper Cenomanian (WRIGHT & KENNEDY 1987, text-fig. 85).

Nevertheless, rare occurrences of the following ammonites have to be mentioned: *Austiniceras austeni* (SHARPE), *Lewesiceras cenomanense* (WIEDM. & SCHN.), *Sciponoceras baculoide* (MANTELL), and *Turrilites (T.) acutus* PASSY.

f) Zone of Acanthoceras jukesbrownei

The upsection following 10 m of well-bedded limestones are considered to comprise the Jukesbrownei Zone. A 3.30 m thick limestone bed is differentiated at the top and is bounded by two thin marly layers. The lower one is comparatively rich in *Pycnodonte baylei*. Around the *Pycnodonte baylei* layer, large-sized *Acanthoceras jukesbrownei* (SPATH) are common. *Eucalycoceras rowei* (SPATH) and *Calycoceras newboldi* (KOSSM.) share the same distribution. *Austiniceras austeni* and *Sciponoceras baculoide* are still present.

g) Zone of Eucalycoceras pentagonum

No agreement has been achieved how to define the lower part of the Upper Cenomanian. HANCOCK (1960) proposed a Calycoceras naviculare Zone for the Upper Cenomanian excluding the - at that time Turonian - *Plenus* Marls. JUIGNET & KENNEDY (1976) criticized, however, that *Calycoceras naviculare* (MANTELL) characterizes only the upper part of its zone but continues into the *Plenus* Marls. They proposed *Eucalycoceras pentagonum* (JUKES-BR.) instead because it has the necessary range as well as a wide distribution. Continuing into the *Plenus* Marls, it was again substituted by *Calycoceras guerangeri* SPATH (WRIGHT & KENNEDY 1984). This substitution is, however, unnecessary because of the definition of zones by the first appearance of the index species - without any relevance to its disappearance. Moreover, *Calycoceras guerangeri* has a rather restricted distribution. For these reasons, *Eucalycoceras pentagonum* is favoured by the present authors.

In the Westphalian Normalfazies the upper 15 m of well-bedded limestones can be attributed to the Pentagonum Zone. Unfortunately, ammonites become very rare in this part of the sequence, similar to the conditions found in SE England (WRIGHT & KENNEDY 1984: 7). None of the three species discussed above have been found in northern Westphalia. The only ammonites filling the gap between Jukesbrownei and Geslinianum zones are *Calycoceras (Lotzeites) aberrans* (KOSSM.), *C. (L.)* cf. *lotzei* WIEDM., *Thomelites* sp., *Austiniceras austeni* (SHARPE), *Schloenbachia* cf. *lymense* SPATH, and *Sciponoceras* sp.

h) Zone of Metoicoceras geslinianum

The lower 5 m of the Rotpläner marls and limestones are considered to correlate with the Geslinianum Zone despite the fact that also *Metoicoceras geslinianum* (D'ORB.) is a very rare species and up to now restricted to the base of this zone. In the same level large-sized *Pachydesmoceras denisonianum* (STOL.) occur (KAPLAN & SCHMID 1983, KAPLAN et al. 1985).

At the coastline section of the Kassenberg the red limestone with *Utaturiceras vicinale* and *Hypoturrilites carcitanensis* is directly overlain by a grey-green glauconitic marly limestone, 0.10 - 0.50 m thick, and yielding *Actinocamax plenus* (BLAINV.) and *Metoicoceras geslinianum* (D'ORB.) (WIEDMANN & SCHNEIDER 1979).

i) "Zone of Neocardioceras juddii" and Cenomanian/Turonian boundary

In the Wunstorf quarry (Lower Saxony) and the middle part of the Osning, Rotpläner is replaced by the black-grey alternation. Here, two thin (0.50 m) levels with *Neocardioceras juddii* (BARR. & GUERNE) together with *Sciponoceras* sp. and *Allocrioceras annulatum* (SHUM.) exist on top of the Geslinianum Zone. This is not the place to decide whether these levels should be considered an individualized ammonite zone (WRIGHT & KENNEDY 1981, 1984, BIRKELUND et al. 1984, KENNEDY 1984) or much better a *Juddii* Event. The latter idea, however, has many points in its favour.

The problem of the Cenomanian/Turonian boundary was discussed in detail by BIRKELUND et al. (1984). Since there is a majority favouring a boundary between the "Neocardioceras juddii Zone" and the Zone of Watinoceras coloradoense, this solution is also accepted in the present paper. It coincides, moreover, with a pronounced change in inoceramid evolution.

3.2 Inoceramids

The same accuracy of subdividing the Cenomanian cannot be achieved by inoceramids because of the larger ranges of inoceramid species. KELLER (1982) elaborated a subdivision of the Cenomanian of Lower Saxony in five inoceramid zones, which to a great extent are reliable with the ammonite zonation (KAPLAN et al. 1985). It largely agrees with the inoceramid succession studied by TRÖGER (1981, this vol.). The Upper Cretaceous inoceramids have - in contrast to their Lower Cretaceous ancestors - a nearly cosmopolitan distribution and allow correlation between the North Temperate and the Tethyan realms (WIEDMANN & KAUFFMAN 1978).

a) Zone of Inoceramus crippsi

Inoceramus crippsi crippsi MANTELL may be regarded as index species of the early Cenomanian. It is known from the glauconite horizon with *Neohibolites ultimus* (D'ORB.) in the Sack Syncline (KELLER 1982) and from the early Cenomanian marls of the Osning area. At the base of the Cenomanian it is competing with the last descendants of the *Inoceramus anglicus* lineage, i. e. *I. comancheanus* CRAGIN (see MARCINOWSKI 1974, KAUFFMAN 1978). *I. crippsi crippsi* becomes extinct at the end of the Zone of Acanthoceras rhotomagense. The Zone of Inoceramus crippsi may, however, be defined to correspond to the Zone of Utaturiceras vicinale/Hypoturrilites carcitanensis and to the lower Zone of Mantelliceras mantelli.

b) Zone of Inoceramus virgatus

This zone comprises the range of *Inoceramus virgatus* SCHLÜTER up to the appearance of *Inoceramus schoendorfi* HEINZ. This corresponds with the upper Mantelli and Dixoni zones. The index species appears together with *I. crippsi hoppenstedtensis* TRÖGER. Near to the facies change to the flaser-bedded marls of the Dixoni Zone, *Inoceramus tenuis* MANTELL appears. All species and subspecies mentioned range into the Middle Cenomanian.

c) Zone of Inoceramus schoendorfi

Also for the Middle Cenomanian a subdivision into two inoceramid zones seems appropriate. The first zone is defined by the range of *Inoceramus schoendorfi* HEINZ and corresponds with the two zones of Acanthoceras rhotomagense and Turrilites acutus. The central part of the Middle Cenomanian can be characterized by a late form of the *crippsi* lineage, i. e. *I. "tenuistriatus"* sensu KELLER 1982, non NAGAO & MATSUMOTO 1939; but systematic position and stratigraphic range need to be ascertained. During the life-time of this form, all Lower Cenomanian survivors disappear.

d) Zone of Inoceramus atlanticus

The upper part of the Middle Cenomanian can be defined by another late form of the *crippsi* lineage, i. e. *Inoceramus atlanticus* (HEINZ). It first appears somewhat below the range of *Acanthoceras jukesbrownei* and the level with *Pycnodonte baylei*. It decreases in abundance from Westphalia to Lower Saxony, but at the same time - at the Middle/Upper Cenomanian boundary - inoceramid abundance is generally reduced.

e) Zone of Inoceramus pictus

In the Osning Normalfazies and in Lower Saxony, *Inoceramus atlanticus* is immediately followed by representants of the group of *Inoceramus pictus* J. SOW. At the top of the lineage there is, however, a short overlap of *I. pictus bohemicus* LEONH. with the early Turonian *I. (Mytiloides) mytiloides* MANTELL. Despite the fact that there is no facies change at the North German Cenomanian/Turonian boundary, a subgeneric evolutionary change in inoceramids coincides with this boundary as defined above. However, it is sometimes very difficult to distinguish late members of the *pictus* lineage from early representants of the *mytiloides* stock.

There is, however, a problem related with the lower boundary of the Zone of Inoceramus pictus, since this species seems to extend its range into the Middle Cenomanian in Western Europe (see TRÖGER this vol., Text-Fig. 1). In North Germany *I. pictus pictus* and *I. pictus neocaledonicus* JEANNET seem to characterize the Pentagonum Zone, *I. pictus concentrico-undulatus* TRÖGER the transitional beds between well-bedded limestones and Rotpläner, and the small-sized *I. pictus bohemicus* the Zone of Metoicoceras geslinianum (including the *Juddii* Event).

3.3 Planktonic Foraminifera

The next step has to be the correlation of the data obtained from ammonites and inoceramids with those of planktonic foraminifera. Despite the important contribution of WEISS (1982) we are still far from satisfying results. We learned from WEISS that ranges of Cenomanian *Rotalipora* and Turonian globotruncanids differ considerably between the Mediterranean Southern France and the Temperate North Germany. Since his research started from selected localities, it is not yet clear how far the results can be generalized.

Table 2. Comparison of Cenomanian subdivisions of North Germany (WIEDMANN & SCHNEIDER 1979, KELLER 1982, WEISS 1982, KAPLAN et al. 1985, this paper) and Northern Spain (WIEDMANN 1960, 1980, WIEDMANN & KAUFFMAN 1978).

		NORTH GERMANY			MEDITERRANEAN	
		Ammonite Zones	Inoceramid Zones	Plankt.Foraminifera (WEISS 1982)	Ammonite Zones (WIEDMANN & KAUFFMAN 1978)	Inoceramid Zones (WIEDMANN & KAUFFMAN 1978)
TURONIAN		Mammites nodosoides Watinoceras coloradoense	Mytiloides labiatus Mytiloides mytiloides	Whiteinella archaeocretacea	Fallotites subconciliatus Vascoceras gamai	Mytiloides mytiloides Mytiloides opalensis M. submytiloides
CENOMANIAN	UPPER	Neocardioceras juddii	Inoceramus pictus bohemicus	Rotalipora cushmani & R. greenhornensis (R. thomei)	Metoicoceras geslinianum	Inoceramus pictus s. l.
CENOMANIAN	UPPER	Metoicoceras geslinianum			Metoicoceras muelleri	
CENOMANIAN	UPPER	Eucalycoceras pentagonum	Inoceramus pictus pictus		Calycoceras naviculare & Lotzeites lotzei	?
CENOMANIAN	MIDDLE	Acanthoceras jukesbrownei	Inoceramus atlanticus	Rotalipora cushmani	Eucalycoceras spathi	I. aff. prefragilis
CENOMANIAN	MIDDLE	Turrilites acutus	(I. tenuistriatus s. KELLER)	?		Inoceramus etheridgei
CENOMANIAN	MIDDLE	Acanthoceras rhotomagense	Inoceramus schoendorfi	Rotalipora reicheli	Cunningtonic. cunningtoni	Inoceramus reachensis
CENOMANIAN	LOWER	Mantelliceras dixoni	(I. tenuis)	Rotalipora appenninica	Mantelliceras mantelli	
CENOMANIAN	LOWER	Mantelliceras mantelli	Inoceramus virgatus			
CENOMANIAN	LOWER	Hypoturrilites carcitanensis Utaturiceras vicinale	Inoceramus crippsi	?	Hypoturrilites mantelli & Graysonites sp.	?
ALBIAN		Stoliczkaia dispar	(I. gr.anglicus)		Stoliczkaia dispar	?

In any case, *Rotalipora appenninica* RENZ occurs in Lower Saxony and Helgoland much later than in the Mediterranean Realm, i. e. in the late Lower and the Middle Cenomanian. Thus the early Cenomanian cannot be adequately defined yet by foraminifera.

Also *Rotalipora reicheli* MORNOD arrives in North Germany with delay and may characterize the early Middle Cenomanian (Zone of Acanthoceras rhotomagense). In Southern France, however, it follows *Rotalipora brotzeni* (SIGAL) during the late Cenomanian. *Rotalipora cushmani* (MORROW) first occurs in Southern France with *Turrilites acutus,* in North Germany it appears delayed again in the late Middle Cenomanian. A transitional zone between Middle and Upper Cenomanian can generally be defined by *Rotalipora thomei* (HAGN & ZEIL). The Middle Cenomanian Acutus Zone has, however, not yet any specific index species.

The Upper Cenomanian Pentagonum Zone can be considered to correlate with a Zone of Rotalipora cushmani and R. greenhornensis despite the fact that *R. cushmani* first appears in the Middle Cenomanian. *R. greenhornensis* (MORROW) is still present in the first 2 m of the *Plenus* Marls, while *R. cushmani* disappears 1 m above. *Rotalipora deeckei* FRANKE co-occurs in the Mediterranean Realm as well as in North Germany with *R. greenhornensis.* Rotpläner or "black-grey alternation" correspond with the "Zone à grandes Globigérines" (LEHMANN 1963) and are in consequence characterized by the first appearances of *Dicarinella hagni* (SCHEIBN.) and *Whiteinella archaeocretacea* PESSAGNO.

These remarkable differences in stratigraphic ranges do, however, not restrict the use of planktonic foraminifera for correlation, but they have to be kept in mind.

4. Cenomanian Events

There is still much controversy about synchroneity and use of sedimentary/faunistic events for stratigraphic correlation. With the necessary reservation we have to admit that at least some Cenomanian events are found in Westphalia as well as in Lower Saxony (Hannover, Hildesheim) in the same or a similar stratigraphic position. These are:

a) **Lower/Middle Cenomanian boundary event** documented by intraclast layers up to mass flows as the response to synsedimentary tectonics near the boundary;
b) **Mid-Cenomanian event** as documented by an omission surface or "nonsequence" in the area of Hildesheim which is placed at the boundary between Rhotomagense and Acutus zones (DAHMER et al. 1986). It was first recognized in the Anglo-Paris Basin by CARTER & HART (1977);
c) **Pycnodonte baylei event**, placed below and
d) **Amphidonte event**, placed above the boundary between Middle and Upper Cenomanian; both are characterized by an uncommon accumulation of oysters;
e) **Plenus Marl facies change** which is documented by the Rotpläner facies in the southern and by a "black-grey alternation" in the northern, basinal sections; the latter corresponds with the oceanic blackshale event (CTBE) and both drastic changes coincide with the boundary of Pentagonum and Geslinianum zones;
f) **Pachydesmoceras event** as it is documented by the occurrence of large-

sized *Pachydesmoceras denisonianum* (STOL.) in the first thick limestone bed of the Geslinianum Zone; and finally

g) **Mytiloides event** characterizing the base of the Turonian by a mass occurrence of *Mytiloides* shells.

4. Conclusions

The North German Cenomanian was reinvestigated in order to define its boundaries and subdivisions with more accuracy (Table 2). In this paper, especially the "Normalfazies" of northern Westphalia (Osning Range) is studied. The advantage of this Normalfazies is the complete lack of glauconitic condensation, thus continuous sedimentation can be assumed. Ammonites and inoceramids are studied at the same time, and an attempt is made to tie both zonations together, and to compare them with that of planktonic foraminifera (WEISS 1982).

Interesting is above all the presence of the lowermost Cenomanian at the famous Kassenberg section, including *Utaturiceras vicinale* together with *Hypoturrilites carcitanensis*, and with forms intermediate to faunas of the uppermost Albian. Unfortunately, the equivalent beds of Normalfazies are not exposed.

Interesting is the agreement with the results obtained in the largely condensed sections of the Anglo-Paris Basin (KENNEDY 1960, JUIGNET & KENNEDY 1976, WRIGHT & KENNEDY 1981, 1984, 1987). Also with the Cenomanian successions and subdivisions of Saxony (TRÖGER 1981, this vol.) and Poland (CIESLINSKI 1959, MARCINOWSKI 1974) the agreement is perfect as it is with the Cenomanian of the Gulf of Regensburg (FÖRSTER et al. 1983), and even the Northern Calcareous Alps (IMMEL 1979). This has paleogeographic implications.

Very different are, however, the ammonite successions of late Cenomanian-early Turonian on the Iberian Peninsula (CHOFFAT 1898, WIEDMANN 1960, 1980, WIEDMANN & KAUFFMAN 1978), NW Africa (PERVINQUIERE 1907, 1910), Nigeria (REYMENT 1954a, 1954b, 1955, 1956, BARBER 1957, POPOFF et al. 1986, ZABORSKI 1985), and Israel (FREUND & RAAB 1969, LEWY & RAAB 1978). Due to the discovery of inoceramids in the Cenomanian of Northern Spain (WIEDMANN & KAUFFMAN 1978), and to co-occurrences in Israel (LEWY et al. 1984), Tethyan and North Temperate ammonite zones can adequately be correlated (Table 2).

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E. Volcanism, Magneto-Stratigraphy

Vestiges of Volcanic Activity in Cretaceous Sediments of Europe

WINFRIED ZIMMERLE, Celle

With 2 Plates, 3 Text-Figures and 1 Table

ZIMMERLE, W. (1989): Vestiges of Volcanic Activity in Cretaceous Sediments of Europe. - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 951-987. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: Vestiges of volcanic products, as well as other evidence of igneous activity, are widespread in the Cretaceous sediments of Europe, though sometimes they are camouflaged by alteration or weathering. They are classed in order of importance as tangible criteria into four groups: 1) intrusive rocks, volcanic rocks, pyroclastic rocks, and hydrothermal mineralization; 2) volcanic rock fragments and single pyroclastic mineral grains; 3) inorganic geochemical and geophysical evidence; and 4) specific lithological associations, diagnostic minerals, and certain fossil groups. Examples of methods for identifying these vestiges are dealt with briefly; they include macroscopic inspection, thin-section analysis, phase microscopy, scanning electron microscopy, and heavy-mineral examination.

The evaluation of evidence for volcanic activity in Europe during the Cretaceous is substantiated by literature data and own observation. The problem of provenance of the various pyroclastic and volcaniclastic deposits as seen by various authors is discussed. A chart summarizes the stratigraphic distribution of the volcanic vestiges observed in the Cretaceous sediment of Europe, as well as associated igneous intrusions, area by area: North Sea, British Isles, France/Benelux, Central Europe, northern Europe, eastern Europe, Alps (Helveticum), and Iberia. Finally, the importance of tracking the volcanic vestiges is emphasized for any paleogeographic reconstruction and their relevance to the applied geology of the Cretaceous.

Kurzfassung: Spuren vulkanischer Tätigkeit und Hinweise auf magmatische Aktivität sind verbreitet in kretazischen Sedimenten von Europa, jedoch vielfach schwer erkennbar infolge von Umwandlungs- und Verwitterungsvorgängen. Diese Spuren werden in vier Gruppen klassifiziert, die nach der Bedeutung ihrer Beweiskraft angeordnet sind: (1) Intrusivgesteine, vulkanische Gesteine, Pyroklastika und hydrothermale Mineralisation; (2) vulkanische Gesteinsfragmente und einzelne pyroklastische Mineralkörner; (3) anorganisch-geochemische und geophysikalische Hinweise und (4) spezifische lithologische Assoziationen, diagnostische Minerale und bestimmte Fossilgruppen. Methoden zur Erkennung solcher Spuren werden aufgezeigt: megaskopische Untersuchung, Dünnschliff-Analyse, Phasenkontrast-Mikroskopie, Rasterelektronen-Mikroskopie und Schwermineral-Untersuchung. Die Beweiskette für vulkanische Tätigkeit während der Kreidezeit in Europa wird aufgebaut anhand von Literaturdaten und eigenen Beobachtungen. Die geographische Herkunft der verschiedenen Pyroklastika und der vulkanoklastischen Ablagerungen ist in ihrer zeitlichen Abfolge aufgelistet. Eine Übersichtstabelle stellt die stratigraphische Verteilung der vulkanischen Spuren sowie der assoziierten magmatischen Intrusionen für die Kreide in Europa in der folgenden geographischen Gliederung zusammen: Nordsee-Gebiet mit Rockall, Britische Inseln, Frankreich/Benelux, Zentraleuropa (BRD), Nordeuropa, Osteuropa, Alpen (Helvetikum) und Iberische Halbinsel. Schließlich wird auf die Wichtigkeit des Erkennens und Verfolgens vulkanischer Spuren hinsichtlich jeglicher paläogeographischer Rekonstruktion hingewiesen und ihre Bedeutung für die angewandte Geologie in der Kreide hervorgehoben.

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1. Introduction

In the classical geology and petrography of Europe, the conception of paleovolcanism (Permian and older) and neovolcanism (Tertiary and younger) persisted for a long time: That between Permian and Tertiary there were periods with no igneous activity in Europe. HARRISON, JEANS & MERRIMAN (1979: 57) made the pertinent statement to this point: "Prior to isotopic age determination of British rocks, minerals and associated hydrothermal events, the British Isles were generally considered to have been magmatically dormant from the close of the Hercynian orogeny to the widespread volcanism of the Tertiary." Oil exploration in northwestern Europe, especially in the North Sea, furnished supplementary data and observation which no longer allowed to sustain this conception: Mid-Jurassic volcanic rocks were discovered in the Moray Firth (HOWITT et al. 1975, WALMSLEY 1975).

Thus, there is an urgent need to collect data and observation on Mesozoic and Cenozoic volcanism in Europe and to summarize this data. This is done for the Cretaceous period in the present contribution. The data is by no way exhaustive and the author would be grateful for additional observation and information on the subject.

The pertinent data is compiled in Text-Fig. 1, a synoptic graph, entitled "Vestiges of volcanism and related phenomena in the Cretaceous of Europe". This compilation includes observation by all methods available: geological mapping, geophysical surveys, megascopic examination, thin-section petrography, phase microscopy, scanning electron microscopy, heavy-mineral mounts, cathodoluminescence, microprobe analysis, and X-ray diffractometry.

The Plates 1-2 depict photographs with various examples of the methods applied: (1) Megascopic examination of tuffaceous layers (Plate 1, Figs. 1-4), (2) thin-section micrographs of tuffs (Plate 1, Figs. 5-6), (3) phase-microscopic graphs of tuffs (Plate 1, Figs. 7-8), (4) scanning electron micrographs of single components (Plate 2, Figs. 1-3), (5) micrographs of heavy minerals (Plate 2, Figs. 4-11), and (6) scheme of a highly altered volcaniclastic clay (Plate 2, Fig. 12).

2. Vestiges of volcanism and related phenomena

Vestiges of volcanism are of various types and sizes (submicroscopic to kilometer dimensions). They can be either of direct or indirect evidence. For

Table 1. Vestiges of volcanism and related phenomena, ranked in four categories of degrading convictive evidence.

FIRST - ORDER
PLUTONICS - VOLCANICS - PYROCLASTICS - HYDROTHERMAL MINERALIZATION

SECOND - ORDER
VOLCANIC ROCK FRAGMENTS- PYROCLASTIC MINERALS

THIRD - ORDER
GEOCHEMICAL EVIDENCE - GEOPHYSICAL HINTS

FOURTH - ORDER
SILICEOUS ROCKS - FLINTS - SMECTITE - PHOSPHORITES - SIDERITE - GLAUCONITE - CRISTOBALITES - ZEOLITES

the present purpose, they are ranked in four categories, namely first- to fourth-order vestiges of degrading convictive evidence (Table 1):

First-order vestiges are direct indications of all types of igneous activity whose Cretaceous age is known by radiometric dating or clear stratigraphic evidence:
- anorogenic or orogenic plutonic rocks or intrusives (sensu ZIEGLER 1982, Encl. 38) of calc-alkalic and alkalic composition, respectively;
- anorogenic or orogenic hypabyssal and volcanic rocks;
- pyroclastic rocks in a stratigraphic sequence;
- hydrothermal mineralization of magmatic origin, ordinarily postdating plutonic as well as volcanic activities.

Evidence of igneous rocks, hydrothermal mineralization, and volcanogenic sediments deserves collation.

Second-order vestiges are direct indications, but of detrital origin whose geological age is not necessarily known or determined and which are of much smaller dimensions (mm to cm). They are unaltered volcanic rock fragments*, crystallized volcanic rock fragments as well as volcanic glass fragments, and single pyroclastic minerals.

* Altered volcanic rock fragments, however, such as volcanic groundmasses or volcanic glass particles are normally difficult to identify.

Among the pyroclasts range minerals like pyroclastic quartz, sanidine, pyroclastic biotite (sensu GAIDA et al. 1978: 80, 81) or specific ore minerals of volcanic derivation.

Third-order vestiges are indirect evidences only and hint at igneous or hydrothermal activity. These vestiges are essentially obtained by inorganic geochemical analysis and geophysical surveys. Geochemical evidence might be based on major elements (e. g. titanium) and certain element ratios (e. g. Ca-K-Na), on trace elements, e. g. Zr, Nb, Y, La, Cr, Ce, and Th, and trace element ratios, or on radioactive elements such as K, U, and Th. Geophysical evidence comprises the whole spectrum of methods: seismics, gravimetry, magnetics, and wire-line log measurements (SERRA 1985). Magnetostratigraphy and other magnetic measurements will play an increasing role (HAMBACH 1985: 84ff., Anlage 6; 1987). Natural gamma-ray spectrometry furnishes valuable data on mineral and rock composition and derivation (ENGELL-JENSEN, KORSBECH & MADSEN 1984).

Major and trace elements of Cretaceous sediments have been analysed and interpreted by PORTHAULT (1982). Relative ratios of Al, K and Na and the Al-Ti-Ba-Na association were used. Moreover, Si and Sr-bearing smectites are supposed to be indicative of volcanic ashes decomposed by halmyrolysis (PORTHAULT 1982: 132). Unusual high boron contents (PORTHAULT 1982: 127), previously and normally used to determine paleosalinity, have been considered as indicative of volcanic derivation (KEMPER & ZIMMERLE 1982b: 674) because of the association with volcanic tuffs. Furthermore, high amounts of fluorine are associated with the high boron contents in the rock suite analysed by PORTHAULT (1982).

Fluorine is a key element (HARRISON, STUART & STRONG 1982). A high fluorine content points likely to volcanic derivation. In this context it is noteworthy that volcanic quartz is characterized by high amounts of fluorine.

Even the derivation of heavy metals from volcanic sources is postulated by some authors. For instance, in the Oligocene Carpathian Flysch of the USSR synchronous volcanism is said to have led to the concentration of chalcophilic elements like copper, lead, and zinc which are derived from decaying pyroclasts according to AFANASYEVA (1986). This interpretation might even have some bearing on explaining the trace element content of crude oils.

Fourth-order vestiges are considered by the author to be additional hints - together with other criteria or in combination with other vestiges - at volcanic influence. Taken alone they are no valid indicators. The fourth-order vestiges comprise (1) related rock associations, (2) related minerals, and (3) certain fossils.

The rock associations being considered as fourth-order vestiges to volcanism are siliceous rocks s. l., like cherts, radiolarites, gaize, opoka, diatomites or Flammenmergel. Numerous Recent siliceous sediments are related to volcanism, too.

Russian petrographers even minted the term volcanogenic-siliceous sedimentary rock association (BURLIN et al. 1976) or tuffo-siliceous association.

Even flint, a characteristic rock component of the Upper Cretaceous in Europe, is included here as closely associated with the group of biosiliceous rocks. The origin of flint nodules is a thoroughly and controversially discussed subject (FÜCHTBAUER 1974: 329, BROCKAMP 1976: 173, VOIGT 1979, EHRMANN 1986).

The Upper Cretaceous flints are mostly brownish or dark grey nodules with a white crust and consist of chalcedony. Inclusions of bryozoans, echinoids, bivalves, belemnites, and foraminifera which are replaced by silica suggest that the flint is a replacement. Soft microfossils with preserved organic material point to an early diagenetic origin. The silica is supplied mainly by sponge needles in the chalk.

Some authors claim that an interrelation of flint formation and volcanic influence exists. Thus, VALETON (1960, fig. 2, 39) emphasized the close spatial association of alternating tuffite layers and flint horizons in Upper Cretaceous profiles examined, especially in Lüneburg, insinuating a genetic relationship. Also FÜCHTBAUER (1974: 329) commented that a volcanic origin of the silica cannot be excluded "in view of the tuffite layers found by VALETON (1960)." Moreover, BROCKAMP (1976: 173) related lithological peculiarities of the Cretaceous, such as the presence of flint nodules, to volcanism. Further observation and additional analytical work are necessary to decipher the primordial derivation of flint silica; the diagenetic, often complex recycling of flint silica is a fact beyond any doubt.

Other characteristic, but polygenetic minerals, discussed in the context with volcanic activity in the Lower Cretaceous (ZIMMERLE 1982a: 85ff., 1982b, KEMPER & ZIMMERLE 1982a, 1982b) are smectites, phosphates and phosphorites, complex carbonate minerals (kutnahorite) and siderite, glauconite, cristobalite, and zeolites, especially clinoptilolite. Aspects of phosphorite occurrence and volcanic influence have been briefly discussed by ZIMMERLE (1982b: 198), and KEMPER & ZIMMERLE (1982a: 251/252).

The origin of siderite is also a debated subject. For instance, BOSTRÖM (1970) considers submarine volcanism as one source for iron. And in Recent sediments, neoformation of siderite by volcanic exhalations has been observed: e. g. in the Laacher See and on the Santorin Island (PUCHELT et al. 1973). Moreover, during decomposition of volcanic rocks siderite can newly form.

Diatoms, radiolarians, and/or siliceous sponges are either solitary rock constituents or characteristic components of biosiliceous rocks. Their presence indicates firstly abundant supply of silica and only secondly some link to volcanic activity.

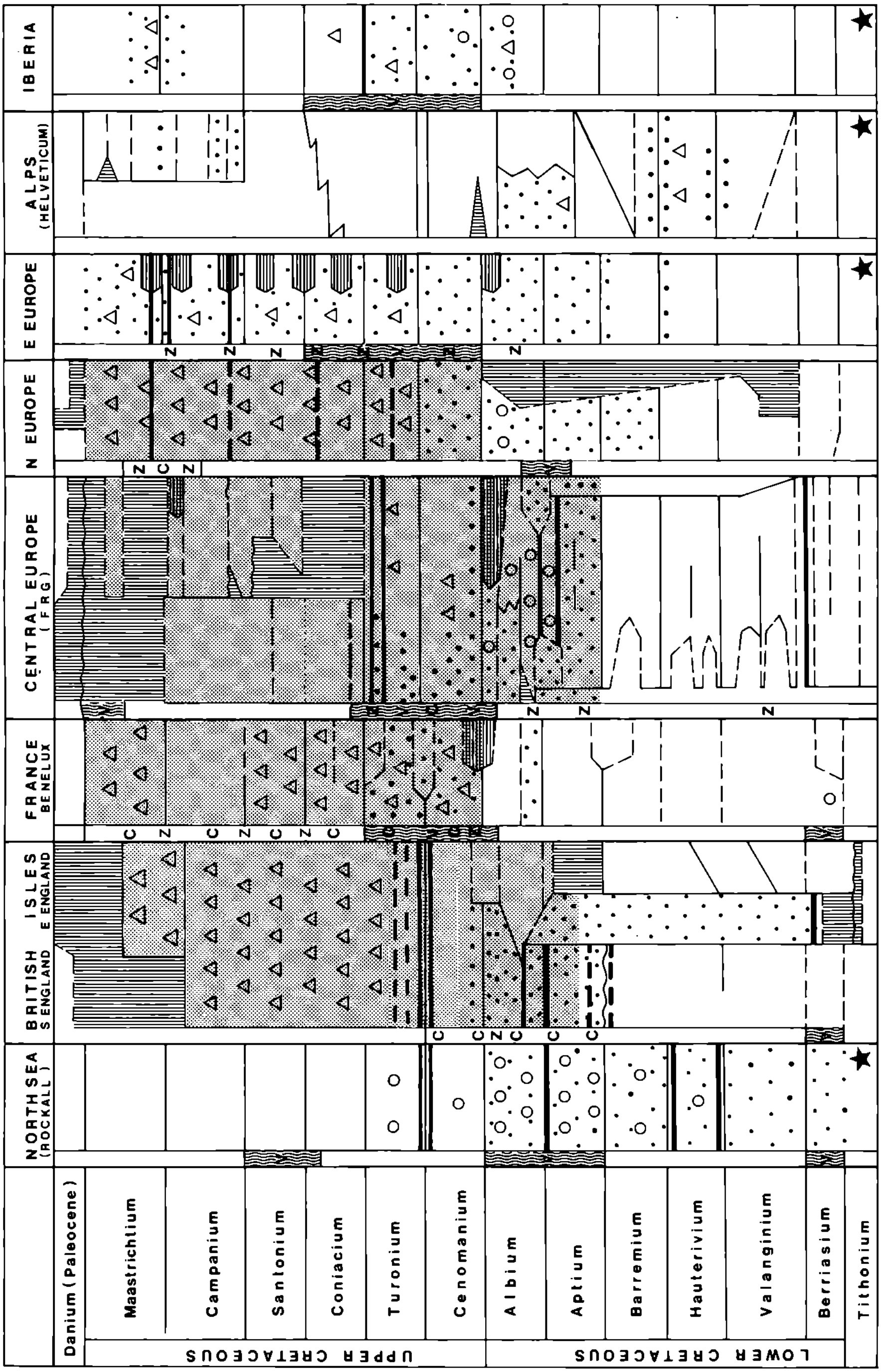
The degree of testimony decreases from the first-order to fourth-order vestiges. Only the sum of all can lead to a reasonable conclusion.

The final stage of alteration is the formation of cryptovolcanic argillaceous sediments or camouflaged pyroclastics as described by Russian petrographers (KHVOROVA 1978: 548/549). Such sediments are the end member of decaying volcaniclastic series; they are an important and voluminous pelitic sediment type in the Cretaceous of Europe. The progressive alteration of the unstable mineral and rock components of volcanogenic rocks makes the recognition of such particles very difficult.

3. Areal and stratigraphic distribution of volcanic vestiges and related phenomena

Text-Fig. 1 summarizes the volcanic vestiges and related phenomena in the Cretaceous of Europe: stratigraphic distribution of igneous rocks, tuffs, smectitic clay minerals, flint nodules, cristobalite, zeolites (clinoptilolite), glauconite, and solitary pyroclasts.

Emphasis is laid mainly upon the non-Tethyan Cretaceous of Europe: North Sea (Rockall), British Isles, France and Benelux, Central Europe,


IBERIA
ALPS (HELVETICUM)
E EUROPE
N EUROPE
CENTRAL EUROPE (F R G)
FRANCE BENELUX
BRITISH ISLES
S ENGLAND
E ENGLAND
NORTH SEA (ROCKALL)
Danium (Paleocene)
Maastrichtium
Campanium
Santonium
Coniacium
Turonium
Cenomanium
Albium
Aptium
Barremium
Hauterivium
Valanginium
Berriasium
Tithonium
UPPER CRETACEOUS
LOWER CRETACEOUS

N Europe, E Europe, Alps, and Iberia. The column North Sea (Rockall) includes all occurrences with the Dutch, U. K., Danish and Norwegian North Sea as well as the shelf W from the British Isles including the Rockall area. The column British Isles is subdivided into S England and E England. France and Benelux are another column. The Central Europe (FRG) column is identical with the western and eastern portion of the Lower Saxony Basin (BRINKMANN & KRÖMMELBEIN 1986: Übersicht 17 Kreide). N Europe includes northernmost Germany, GDR, and Scandinavia. E Europe comprises the epicontinental portion of Poland and the European portion of the USSR. Alps (Helveticum) refers mainly to the Calcareous Alps of southern Germany and Switzerland. Iberia includes data on Spain and Portugal.

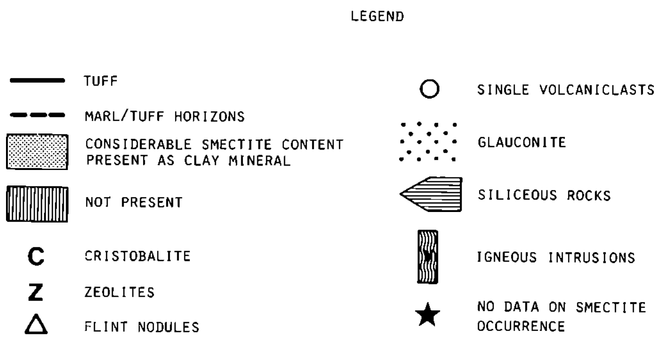
The stratigraphic subdivision and the framework were taken from the chart entitled "Übersicht 17 Kreide" (BRINKMANN & KRÖMMELBEIN) modified and supplemented by some additional columns. The Alpine Flysch zone of Austria, the W Carpathian Flysch zones of the CSSR and Poland, the E Carpathian Flysch zones (Hungaria, Bulgaria) and the Mediterranean (Italy) are not listed.

3.1 North Sea (Rockall)

In the column North Sea (Rockall), data so far known from this area is compiled. The tectonic framework is mainly governed by the spreading of the North Atlantic and associated rifting (ZIEGLER 1982). The Rockall Trough may represent a region of oceanic crust formed in the Cretaceous or in the Permian. Spreading had probably begun in the Bay of Biscay, west of Iberia in the Rockall-Trough-Faeroe-Shetland channel by Aptian time. By late Cretaceous time spreading had ceased in the Bay of Biscay and Rockall Trough. The geological history of the Goban Spur was described by ROBERTS et al. (1981).

First-order vestiges of volcanic activities were found in the past decades mainly thanks to oil exploration (ZIEGLER 1982). According to HARRISON et al. (1979: 59), late Cretaceous igneous activity (Coniacian/Santonian)

Text-Fig. 1. Vestiges of volcanism and related phenomena in the Cretaceous of Europe.



with microgabbros, dolerites, lavas, and tuffs was observed in NW Ireland, in the Blackstones Bank igneous complex in the Sea of the Hebrides, at Helen's Reef (olivine-microgabbro) near Rockall Island, and at seamounts in the Rockall Trough (Rosemary Bank, Anton Dohrn, and Hebrides Terrace). This late Cretaceous volcanism culminated in the Paleocene volcanism of the Thulean Igneous Province.

In the 72/10-1 A well of the Western Approaches BENNET et al. (1985: 259) observed a variegated sequence (36 m) of altered volcanic rocks and intercalated tuffaceous sediments of Aptian age containing evidence of subaerial exposure. These rocks are similar in age and composition to a rift-related olivine-dolerite in the Fastnet Basin (CASTON et al. 1981).

In the **Zuidwal-1** well, onshore the northern Netherlands, alkaline intermediate to basic volcanic rocks (trachytes, phonolites, and leucite-bearing lavas) were found (COTTENCON, PARANT & FLACELIERE 1975). Their age was determined with 144 ± 1 Ma (^{40}AR/^{39}Ar - DIXON, FITTON & FROST 1981: 131) and 145 Ma (K-Ar - HARRISON et al. 1979: Appendix 2).

Age determinations (^{40}Ar/^{39}Ar) of alkaline basic volcanic rocks (olivine nephelinites and basanites) from wells in the **southern Netherlands** (Andel-2, Andel-4, and Loop-op-Zand-1) and a mafic biotite phonolite (biotite monchiquite) from the 29/25-1 well on the northern flank of the Mid-North Sea High at the western edge of the Central Graben revealed crystallization ages near the Jurassic/Cretaceous boundary (DIXON, FITTON & FROST 1981: 132). The volcanic intrusions in the SW Netherland Basin presumably caused the geothermal high shown by CORNELIUS (1975, Fig. 6).

In this context KETTEL (1983) postulated for the heat anomaly caused by the **East Groningen Massif**, previously called Emden Massif (CORNELIUS 1975, Fig. 6), which is associated with a positive aeromagnetic anomaly, a late Kimmerian age (Jurassic/Cretaceous boundary). Further magnetic anomalies, however of unknown age, occur in the subsurface of northwestern Germany.

Mid-Cretaceous (Albian) undersaturated basaltic rocks (olivine nephelinites) and alkali basalts were found in the K/14-FA 103, L/13-3, and Q/7-2 wells of the Dutch North Sea (DIXON, FITTON & FROST 1981, Fig. 6).

All these intrusions are part of the Jurassic-Cretaceous rift magmatism, similar to that of the Eastern Rift of Africa. The Mid-Jurassic to Mid-Cretaceous igneous activity occurred essentially in three episodes, approximately coeval with known tectonic phases: 175-165 Ma - 145-130 Ma - 95-11 Ma (DIXON, FITTON & FROST 1981, ZIEGLER 1982). In the North Sea area the volcanic activity seems to have started in Mid-Jurassic in the north and became confined to the southern North Sea in the early and middle Cretaceous. This trend is also indicated by the presumable intrusion age of the Bramsche and Vlotho Massifs in NW Germany. It likely reflects the progressive breakup of the continent from the Atlantic rim into Central Europe.

Tuffs and pyroclastic deposits (Hauterivian-Turonian with a culmination in the Aptian/Albian) in the North Sea area (e. g. in the UK well 22/1-2A) were briefly described or mentioned by numerous authors (FRODESEN 1979: 12, HARRISON et al. 1979: Table 3, DIXON, FITTON & FROST 1981: 130, ZIEGLER 1982: Encls. 21, 22, 23, 31, 32, 35, 36, 37). For instance, nine thin volcanic tuff bands (bentonites) of late Aptian to early Albian age were found in mudstones of the 81/40 borehole in the western part of the Central North Sea Basin (LOTT, BALL & WILKINSON 1985: 243).

Second-order vestiges such as admixtures of single volcaniclasts are not often recorded. Many of the observations remain still unpublished.

In the southern part of the Porcupine Trough a major volcanic center, recognized on reflection seismics and associated with a positive magnetic anomaly (third-order vestiges), was active during the early Cretaceous (ZIEGLER 1982: 71, Encl. 21).

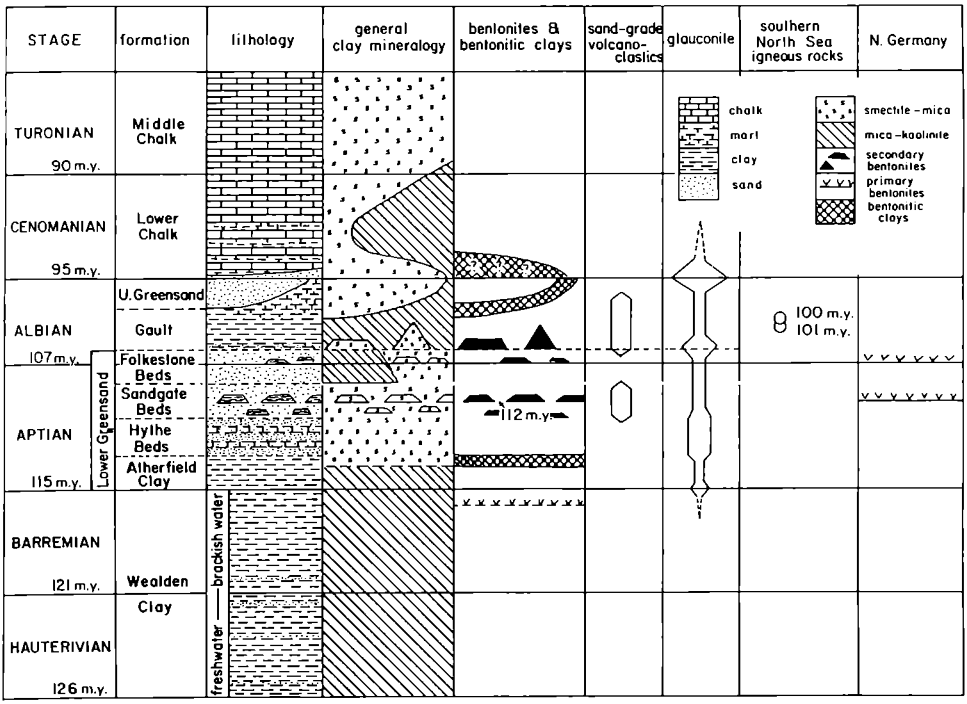
Fluid inclusions in authigenic quartz rims of sandstones from the Brent Formation are considered to be fourth-order vestiges indicating, according to PAGEL et al. (1986), a thermal event (120 ± 10 °C) during the Cretaceous, which is related to a large-volume hot water circulation.

3.2 British Isles

In the column British Isles, first-order vestiges of volcanic activity are tuffs and pyroclastic rocks in the Berriasian (KNOX & FLETCHER 1978), in the Aptian/Albian (JEANS 1978, JEANS et al. 1982), and in the Cenomanian/Turonian (JEFFERIES 1963, JEANS et al. 1982, PACEY 1984).

Intrusive rocks are unknown, except for the Wolf Rock nosean phonolite, Cornwall, with an age of 131 Ma.

Sporadic and scattered hydrothermal events, e. g. uranium mineralization (pitchblende and coffinite), and hydrothermal metasomatism, age-dated



Text-Fig. 2. Lithology, clay mineralogy and distribution of volcanogenic material in the Cretaceous sediments of southern England and their relationship to primary bentonites in the Cretaceous sediments of northwestern Germany (after JEANS et al. 1982).

between 130 and 100 Ma and related to the initial opening of the North Atlantic, are known to occur in Britain (HARRISON et al. 1979: 60, Table 2). Hydrothermal mineralization correlates in general with isotopic ages of magmatic rocks and volcanogenic materials.

Volcaniclastic admixtures in Mid-Cretaceous strata and the volcanogenic origin of some of the glauconite grains were described in great detail by JEANS et al. (1982). The data is depicted in Text-Fig. 2.

The derivation of smectites, especially in the Upper Cretaceous, was and is still thoroughly discussed (JEANS 1978, HARRISON et al. 1979: 61, JEANS et al. 1982).

Crystals of volcanogenic clinoptilolite-heulandite are found in the Lower Cretaceous Fuller's Earth, Upper Greensand and Upper Cretaceous chalk (JEANS et al. 1977: 25, JEANS et al. 1982). Clear golden yellow sphene is a characteristic accessory mineral of the Lower Cretaceous Fuller's Earth (JEANS et al. 1977: 31) as it is in comparable bentonite layers of north-western Germany.

According to HANCOCK (written communication) cristobalite is widespread in small quantities. However, in the Upper Albian Malmstone of Surrey, southern England, and in the Middle Turonian Tuffeau du Bourré of Touraine, France, cristobalite occurs as a rock-forming constituent. The origin of cristobalite, however, is still a subject of debate: biogenic versus volcanogenic.

The thin black clay layer (about 10 cm thick), the so-called Black Band, within the Chalk at the Cenomanian-Turonian boundary of central eastern England is bituminous and composed of smectite. Many geologists who work on the Chalk (e. g. JEANS et al. 1982, PACEY 1984) postulate an ultimately volcanic origin for the smectite. Moreover, within the post-Cenomanian Chalk sequence of the same area about 20 thin persistent marl horizons occur that formed by in-situ alteration of eolian-transported volcanic ashes of pantelleritic composition (PACEY 1984). The morphology of pyroclastic particles suggests that the volcanic eruptions responsible for the ejection of the original ash into the air were of the shallow-water phreatomagmatic (Surtseyan) type.

3.3 France/Benelux

In the column France/Benelux first-order vestiges are the massive intrusions and volcaniclastic rocks of Albian to Turonian age in the orogenic belt of the Pyrenees (THIEBAUT et al. 1979, BEBIEN & GAGNY 1980: 116, SOUQUET et al. 1985).

The following rock types occur in form of lavas and tuffs: melanocratic basalts (ankaramites and spilites) and keratophyres. Theralitic diabases, theralites, picrites, and augite syenites form small intrusions, and lamprophyres such as camptonites, monchiquites, fourchites, micro-episyenites, and albitites occur as dikes. The rock spectrum ranges from ultrabasic to syenitic. Two phases of igneous activity can be distinguished: (1) a pre-orogenic volcanic phase Upper Albian/Lower Cenomanian, and (2) a syn-orogenic to post-orogenic, entirely intrusive phase during Cenomanian, Turonian and post-Turonian. Even the age of mylonitic events in the St. Barthelemy Massif, North Pyrenean Zone, was determined as 100 and 110 Ma. (COSTA & MALUSKI 1987).

Except for the Pyrenees, no Cretaceous tuffs and pyroclastic rocks have been reported so far. The Wadden Zee intrusives (Zuidwal-1 well) were already discussed in the North Sea (Rockall) paragraph.

Tuff particles are found in the "Tuffaceous Series" of basal "Wealden" age in the western Netherlands.

Hydromica and montmorillonite clay minerals as well as fragments of volcanic glass have been reported by TIMOFEEW & BOGOLUBOWA (1978: 3) in the Aptian/Albian black shales of the Bay of Biscay (DSDP Site 402 A and 400).

Furthermore, the smectite content of Upper Cretaceous carbonate rocks in France and its origin was discussed in detail by POMEROL & AUBRY (1977). The authors explained the smectite to be of volcanogenic origin being derived from the Mid-Atlantic Ridge in Cretaceous time. Cristobalite and zeolites (clinoptilolite, and minor wairakite, stilbite, and harmotome) are associated with these sedimentary rocks, mainly in the Turonian (ROBASZYNSKI et al. 1982).

Flint nodules (silex) are abundant throughout the Upper Cretaceous of France, Belgium and the Netherlands. Some French authors (e. g. BRETON 1981: 33/34) prefer to derive the silica by postulating that "Au Crétacé l'Atlantique nord s'ouvrait lors de l'activité de la dorsale, qui fonctionnait par à-coups les éruptions volcaniques sousmarines enrichissaient périodiquement en silice l'eau de mer."

Glauconitic gaize with phosphorite nodules and rich in siliceous sponge fragments, being the lithological equivalent of the Flammenmergel with smectite and minor illite, occurs in Belgium and northern France and was described in great detail (FAUVEL & PETIT 1984).

In Mid-Cretaceous clay cores from the sites 549 and 550 (DSDP Leg 80) on the continental rise W off Brest (Celtic Margin) PASCAL & GILLOT (1984) discriminated between the following clay mineral parageneses:

(1) an early Barremian syn-rift paragenesis composed of kaolinite, illite, chlorite, and mixed-layer minerals,
(2) a late Barremian syn-rift paragenesis composed of mixed-layer minerals as well as of smectite,
(3) an Albian/early Senonian post-rift paragenesis composed of smectites, illite, mixed-layer minerals and often chlorite.

The authors propagating the integration of methods (X-ray mineralogy, geochemistry, and multivariant analysis), explain this rather striking difference in clay mineral composition by change of climatic conditions in the source area (hinterland). Exactly the same pattern of clay mineral composition was observed by BROCKAMP (1976) in clays of the Barremian/Albian interval in northwestern Germany in the surroundings of Hannover. However, based on additional thin-section evidence, the clay mineral composition was interpreted by increasing volcanogenic influence and only indirectly by climate.

3.4 Central Europe

In the column Central Europe, first-order vestiges of volcanic activity are mainly found in NW Germany in form of tuffs and pyroclastic deposits in the Berriasian (ZIMMERLE 1979), in the Aptian/Albian (GAIDA et al. 1978, KEMPER & ZIMMERLE 1978, ZIMMERLE 1979, KEMPER & ZIMMERLE

1982a), in the Turonian (DORN & BRÄUTIGAM 1959, BRÄUTIGAM 1962, SEIBERTZ & VORTISCH 1979) and throughout the Upper Cretaceous (BRÄUTIGAM 1962, VALETON 1959, 1960). However, numerous marl layers, termed tuffs, have not been studied by microscopic means.

Intrusions of volcanic rocks of Mid-Cretaceous age from Trois Epis (Vosges) were age-dated by LIPPOLT et al. (1974: 219). DILLMANN & NEGENDANK (1982) referred to a Middle Cretaceous Lapilli-Tuffschlot, south of Trier. Hercynian basalt dikes in Hessen have a presumable Cretaceous intrusion age (GRUMBT & LÜTZNER 1983, table). The Katzenbuckel volcanics have an age of 66 Ma (HOFMANN 1976).

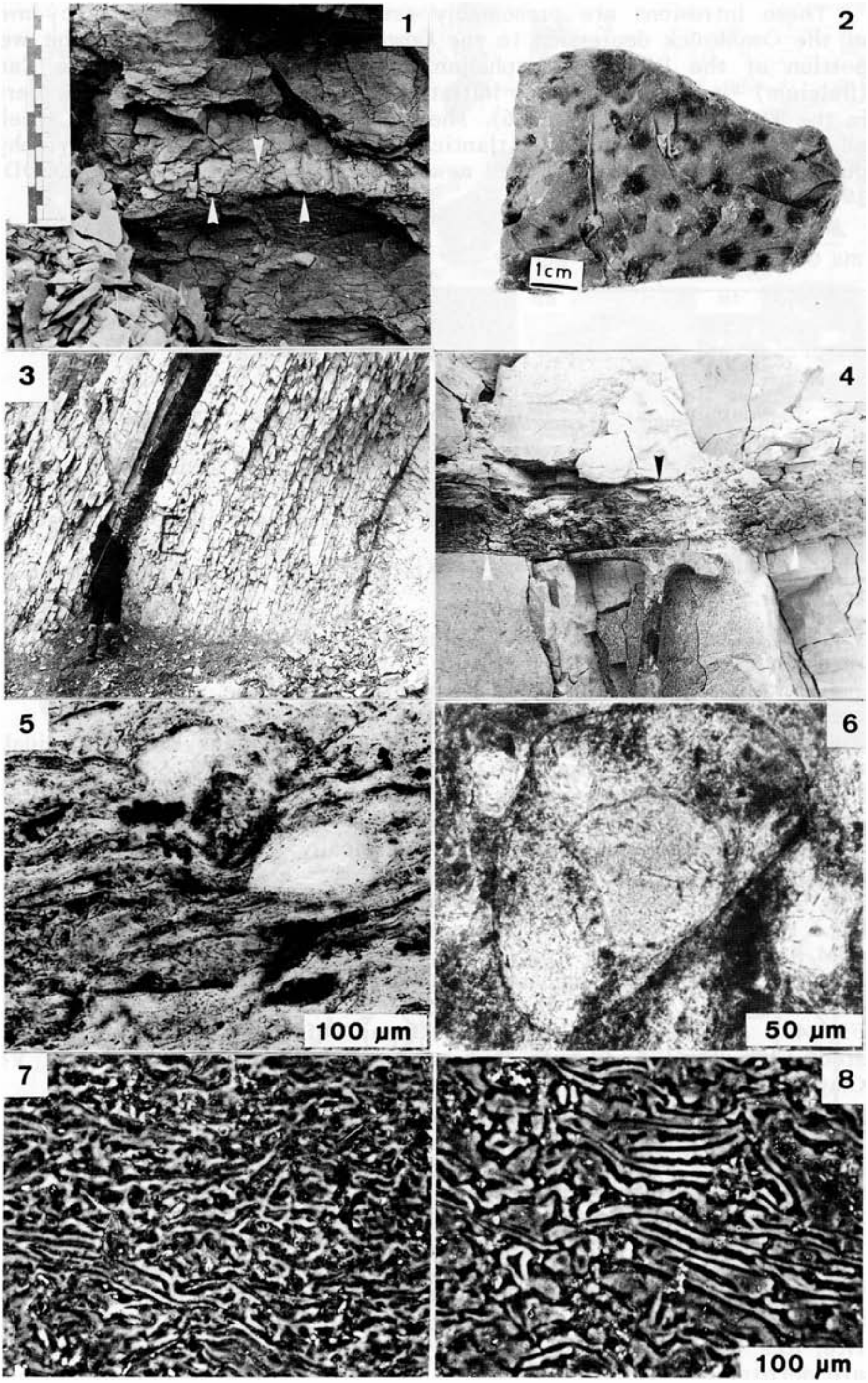
Other first-order vestiges in Central Europe, directly proven by hydrothermal mineralization and indirectly by seismics, vitrinite reflectance, mineralogical studies, and other evidence, are the Bramsche and Vlotho Massifs, presumably of intermediate to basic composition, with minor, more acidic intrusions near Uchte and Ellerburg in a depth of approximately 5 km (CORNELIUS 1975: 58, Figs. 6, 12 + 13, BUNTEBARTH & TEICHMÜLLER 1979, DEUTLOFF et al. 1980).

Coalification measurements (BUNTEBARTH & TEICHMÜLLER 1979, DEUTLOFF et al. 1980, BUNTEBARTH 1985) demonstrated a paleotemperature gradient between 65 and 92 °C/km. The age of these igneous intrusions is most likely Cenomanian-Turonian (Coniacian) (STADLER & TEICHMÜLLER 1971, CORNELIUS 1975: 58/59, BUNTEBARTH & TEICHMÜLLER 1979). ZIEGLER (1982: 71), however, propagated an Aptian age.

Plate 1

Fig. 1. Light-coloured tuff band (3 cm thick) embedded in dry marlstone. Lower Tuff Horizon, Upper Aptian, Sarstedt Clay Pit, 21 km SE of Hannover.
Fig. 2. Fragment of tuff band with dispersed glauconite spots (dark - view upon bedding plane). Upper Tuff Horizon, Upper Aptian, Sarstedt Clay Pit, 21 km SE of Hannover.
Fig. 3. Dark-coloured tuff horizon in limestone sequence. Tuff Horizon E, Turonian, Salder Quarry, Lesse Syncline, 37 km SE of Hannover.
Fig. 4. Thin tuffite horizon (5 cm thick) containing swelling clay minerals in a limestone sequence. Tuff Horizon B, Turonian, Söhlde Quarry (Dammann K. G.), Lesse Syncline, 35 km SE of Hannover.
Fig. 5. Thin section - Vitroclastic tuff with single grains of sanidine, volcaniclasts, and leucoxene (TS 26037, // nicols). Upper Tuff Horizon, Upper Aptian, Sarstedt Clay Pit, 21 km SE of Hannover.
Fig. 6. Thin section - Vesicular volcanic rock fragment (TS 29 271, // nicols). Tuff Horizon B, Turonian, Söhlde Quarry (Dammann K. G.), Lesse Syncline, 35 km SE of Hannover.
Fig. 7. Phase microscopy - Vitroclastic texture in volcanic tuff (TS 25 737). Upper Tuff Horizon, Upper Aptian, Sarstedt Clay Pit, 21 km SE of Hannover.
Fig. 8. Phase microscopy - Vitroclastic texture in volcanic tuff (TS 26 348). Aptian/Albian Boundary Tuff, Vöhrum Clay Pit, 32 km E of Hannover.

Note that tuffs of the same vitroclastic texture (Fig. 7 + 8) show different clay mineral composition (GAIDA et al. 1978: 94)!


1
2
1cm
3
4
5
100 µm
6
50 µm
7
8
100 µm

These intrusions are presumably associated with the tectonic inversion of the Osnabrück depression to the Lower Saxony tectogene. In the western portion of the Rhenish-Westphalian coal mining area and in the Campine (Belgium) "inversion tectonics initiated possibly in the Cenomanian, certainly in the Turonian" (ROSSA 1986). They were also triggered by the accelerated opening of the North Atlantic. A variety of minerals (pyrophyllite, guembelite, and dickite) formed newly in the surrounding rocks (KEDDEINIS 1967: 95, SCHREYER 1969).

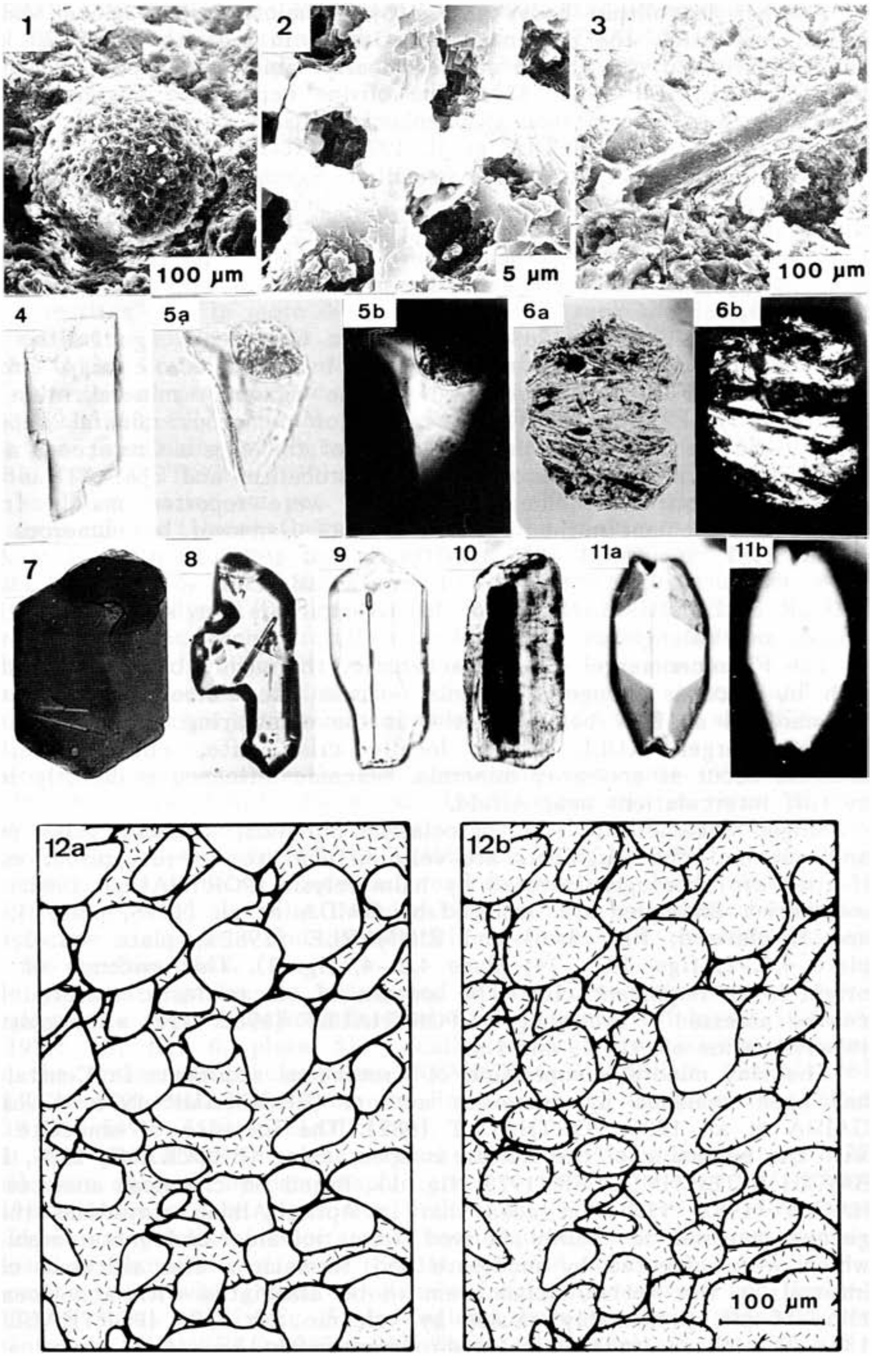
Plate 2

Fig. 1. Scanning electron microscopy - Radiolarian completely replaced by zeolite (clinoptilolite).
Fig. 2. Scanning electron microscopy - Close-up of radiolarian replaced by clinoptilolite.
Fig. 3. Scanning electron microscopy - Imprint of siliceous sponge spicule in marl.
Fig. 1-3. Gott 12 sample, Upper Aptian, Sarstedt Clay Pit, 21 km SE of Hannover.

Fig. 4. Light-mineral mount - Cleavage fragment of sanidine (HM 4067, // nicols).
Fig. 5. Light-mineral mount - Fragment of sanidine with bubble-wall texture (HM 4139). a // nicols, b + nicols.
Fig. 6. Light-mineral mount - Volcanic rock fragment with trachoidal texture (HM 4139). a // nicols, b + nicols.
Fig. 7. Heavy-mineral mount - Pseudohexagonal biotite with prismatic apatite inclusions (HM 4068, // nicols).
Fig. 8. Heavy-mineral mount - Magmatically corroded zircon (HM 4152, // nicols).
Fig. 9. Heavy-mineral mount - Prismatic apatite (HM 4068, // nicols).
Fig. 10. Heavy-mineral mount - Prismatic apatite with pleochroic core (HM 4128, // nicols).
Fig. 11. Heavy-mineral mount - Euhedral sphene twin (HM 4069). a // nicols, b + nicols.
Fig. 4-7 and 9-11 come from the Upper Tuff Horizon, Upper Aptian, Sarstedt Clay Pit; Fig. 8 Tuff Horizon from Aptian/Albian Boundary, Vöhrum Clay Pit. All grains depicted are between 63 and 200 µm in size.

Fig. 12. Scheme of highly altered volcaniclastic clay. Middle Albian, Open Pit Berufsschule Alfeld, 45 km S of Hannover.

a = // nicols: Rounded transparent sand grains (white) are quartz, altered feldspar, and altered volcaniclasts; the rock matrix (randomly dashed) is composed of cryptocrystalline clay, mainly smectite. In the thin section the grain boundaries appear more diffuse than shown in the scheme.
b = + nicols: The major portion of the rounded sand grains, namely feldspars and volcaniclasts are altered into cryptocrystalline clay, mainly into smectite, like the rock matrix (randomly dashed). Two sand grains only (white) are detrital quartz.


1
100 μm
2
5 μm
3
100 μm
4
5a
5b
6a
6b
7
8
9
10
11a
11b
12a
12b
100 μm

Another laccolithic body, proven by seismics, formed since Mid-Cretaceous time below the present Rhine Graben in a depth of 25-30 km; volcanic activity in the graben started nearly simultaneously (ZIEGLER 1975, fig. 18, WALTHER 1983: 323). The olivine nephelinite magmatism on the flanks of the Rhine Graben (Katzenbuckel) dates back to the early Upper Cretaceous (95 Ma) (LIPPOLT et al. 1974, text-fig. 2).

Hydrothermal activity, the so-called Saxonian mineralization, is also known from the Cretaceous. Ba, Pb, Fe mineralization is associated with the Bramsche and Vlotho Massifs (TISCHENDORF 1987). Also the barite-fluorite mineralization in the western Harz Mts., in the Thüringer Wald, and Thüringer Schiefergebirge is of early Saxonian age (GRUMBT & LÜTZNER 1983, table).

Moreover, KETTEL (1983) attributed a Cretaceous age to the chalcocite, bornite, and covellite mineralization in the Paleozoic of a Groothusen well. SCHAEFFER (1984) referred to the Saxonian mineralization in the Sauerland. WALTHER (1983) advocated for numerous mineral deposits of the Alpidic metallogenetic province north of the Alps a Cretaceous age.

Authigenic zeolites (clinoptilolite), cristobalite, and opal CT as well as pyroclastic biotite, sanidine, and titanite were reported mainly from the Aptian to Turonian in the neighbourhood of Hannover by numerous authors (HEIM 1957, SCHÖNER 1960, GAIDA et al. 1978, KEMPER & ZIMMERLE 1978, KULL 1979, PORTHAULT 1982: tables 4.3-1 + 4.3-2, p. 132, ZIMMERLE 1982a, EHRMANN 1986: 40). Analcite is yet known from the Berriasian and Valanginian.

The **Flammenmergel** - a characteristic, thoroughly bioturbated lithofacies rich in siliceous sponge fragments - rims the Mittelgebirge of northern Germany as an E-W belt. Smectite is the dominating clay mineral of the Flammenmergel (KULL 1979); locally cristobalite, clinoptilolite, and/or analcite occur as accessory minerals. Volcanic influence is directly indicated by tuff intercalations near Alfeld.

Single volcaniclasts, i. e. pyroclastic minerals, volcanic glass particles, and volcanic rock fragments are very difficult to be recognized, especially if they are thoroughly altered by halmyrolysis (PORTHAULT 1982: 131) or weathered. Examples are depicted by GAIDA et al. (1978, plate 4, figs. 4 and 5; plate 6, figs. 3-6) and ZIMMERLE (1982a, plate 4.2.-1, fig. 7; plate 4.2.-2, figs. 13 + 14; plate 4.2.-4, fig. 3). This evidence of volcanic origin is an important clue. The amount of volcaniclastic material admixed can be assessed - according to PORTHAULT (1982: 131) - by plotting the relative ratios of Al, K, and Na.

The **clay mineral composition** of Cretaceous sediments in Central Europe has been analysed by numerous authors (BROCKAMP 1976, KULL 1979, GAIDA et al. 1981, PORTHAULT 1982). The increase of smectite content with the beginning of the Aptian is spectacular (BROCKAMP 1976, BUNTEBARTH & TEICHMÜLLER 1979, fig. 1). Based on chemical analyses PORTHAULT (1982: 130) concluded that in Aptian/Albian claystones the terrigenous material is mainly derived from volcanic and metamorphic rocks which were affected by moderate soil formation. The siliceous claystone intervals in the Aptian/Albian seem to be associated with a noticeable decline of pH, presumably caused by volcanic ash falls (PORTHAULT 1982: 133). This change in chemical environment might have also caused a change in microfauna.

The **origin of the smectite** is much discussed (BROCKAMP 1976: 197, KULL 1979, KEMPER & ZIMMERLE 1982b: 675ff., PORTHAULT 1982: 132).

Both, BROCKAMP (1976) and PORTHAULT (1982: 132) advocated a mainly volcanic origin of the Mid-Cretaceous smectites. PORTHAULT (1982: 125ff.) discriminated three types of smectite in the Aptian/Albian claystones around Hannover:

(1) pure smectites rich in silica and with high amounts of B, Zr, and La in a tuff sample,
(2) Sr-bearing siliceous smectites of authigenic origin from tuffaceous claystones highly altered by halmyrolysis,
(3) aluminous smectites with Al, Ti, Ba, and Na.

In most samples analysed all three types occur. It seems important to analyse the smectites even in more detail as to better track their origin.

The marl/tuff horizons, the so-called "Tonbänder" or "Mergellagen", within the Upper Cretaceous (Turonian-Maastrichtian) are difficult to analyse by conventional petrographic means (DORN & BRÄUTIGAM 1959, VALETON 1959, 1960, BRÄUTIGAM 1962, SEIBERTZ & VORTISCH 1979). Except for the painstaking petrographic studies by VALETON (1959, 1960) and SEIBERT & VORTISCH (1979), no thorough examination of the marl/tuff horizons has been carried out yet. This is due to the extremely fine particle size of most of the Upper Cretaceous marls and clays, comparable with that of the Upper Cretaceous bentonites from the Carpathians in southern Poland (LEFELD 1976: 707-710) evidencing cryptovolcanic influence. It is obvious that thoroughly altered dust tuffs can hardly be identified as volcanic products, except when carefully examined by scanning electron microscope and microprobe.

Moreover, the **chemical assimilation** of the pyroclastic layers and the adjacent country rocks is rapid and complete (e. g. GAIDA et al. 1978: 84, table 6).

Glauconite is very common in the Cretaceous sediments (VALETON & ABDUL-RAZAK 1974, VALETON et al. 1982), especially in those of the Mid-Cretaceous (Glaukonitische Randfazies nach KEMPER & ZIMMERLE 1982b: 665, fig. 11-4). KEMPER & ZIMMERLE (1982b, 1983) interpreted this glauconite belt as being related to coastal upwelling. Later, HAMBACH (1985: 59ff.) presented a similar model taking, as a rather modern example, the Pleistocene glauconitic and phosphoritic sediments of the SW African shelf at the Agulhas Bank.

Single glauconite grains are widespread in the Cretaceous strata. Spectacular glauconite spots (Glaukonitflecken) form in some tuff layers (GAIDA et al. 1978: 68, fig. 6, plate 5). Locally even glauconitization horizons occur in the Aptian/Albian claystones (KEMPER & ZIMMERLE 1982a: 248) which have been described also from the Barremian and Albian in the Celtic Sea (ODP Site 549 + 550) (GILLOT 1984).

About 100 glauconite samples of Hauterivian to Maastrichtian age (27 locations) were analysed by petrographic and geochemical means (VALETON et al. 1982). Many of the glauconite pellets are considered to have been formed in-situ as gel-like precipitates. These glauconitic pellets are normally composed of glauconitic mica of the 1M and 1Md mica type.

Glauconite grains with ghosts of volcanic fabric, as observed and depicted by JEANS et al. (1982: fig. 5) in England, have not been observed yet in NW Germany (ZIMMERLE 1982a: 94). There is, however, a definite affinity between the neoformation of glauconite and pyroclastic deposits rich in smectite (ZIMMERLE 1982a: 93/94).

Complex carbonate concretions rich in Fe and Mn are common in the Aptian/Albian of northern Germany. Together with siderite concretions they are considered by some authors to be a direct or indirect product of volcanic exhalations (BROCKAMP 1976, 1978, KEMPER & ZIMMERLE 1982a). Also other siderite occurrences are considered to be of volcanic derivation: e. g. the Devonian siderite concretions in the Rhenish Schiefergebirge (ZACHMANN & JUNGMANN 1984) and the Cenozoic siderite deposit of Deveci, Turkey (ÜNLÜ 1983: 72).

The paragenesis of glauconite and siderite is common and widespread in the Cretaceous sediments of northwestern Europe (ZIMMERLE 1982a: 89ff., KEMPER & ZIMMERLE 1982b: 665, HAMBACH 1985: 23).

Rock-magnetic observation on Cretaceous sediments of NW Germany by HAMBACH (1985: 84ff., 1987) showed that magnetization (NRM) which depends on the absolute content of magnetic minerals within a given sediment, shows a rather sudden, significant increase up to several hundred mA/m at the Cenomanian/Turonian boundary. The maximum intensity was observed in the Plenus Beds which are known to contain volcanogenic components, so far only reported from England (PACEY 1984). HAMBACH (1985: 174) interpreted the present anomaly as caused by a diluted bentonite (verdünnter Bentonit). Higher up in the Turonian section the magnetization decreases gradually. The carrier of magnetization within the marls and marly limestones of the Turonian are minerals of the magnetite group (magnetite and titanomagnetite dissolved at high temperatures), presumably derived from volcanic ash falls. Change in polarity of the paleomagnetic field was also observed in sediments at the Cenomanian/Turonian boundary in Morocco. This rock-magnetic observation may indicate an interrelation between magnetic field, volcanism, climate, and lithofacies (HAMBACH, written communication) supplementing the author's model (ZIMMERLE 1985).

3.5 N Europe

In the column N Europe, the sole first-order vestige is associated with the Tornquist-Teisseyre Zone. The Tornquist-Teisseyre Zone is a lineament "representing the southwestern margin of the East European/Fennoscandian Precambrian basement platform. It is characterized by complex, often rejuvenated, dip-slip and strike-slip faulting and frequently by tectonic inversion" (PEGRUM 1984). The zone extends across Poland beneath the

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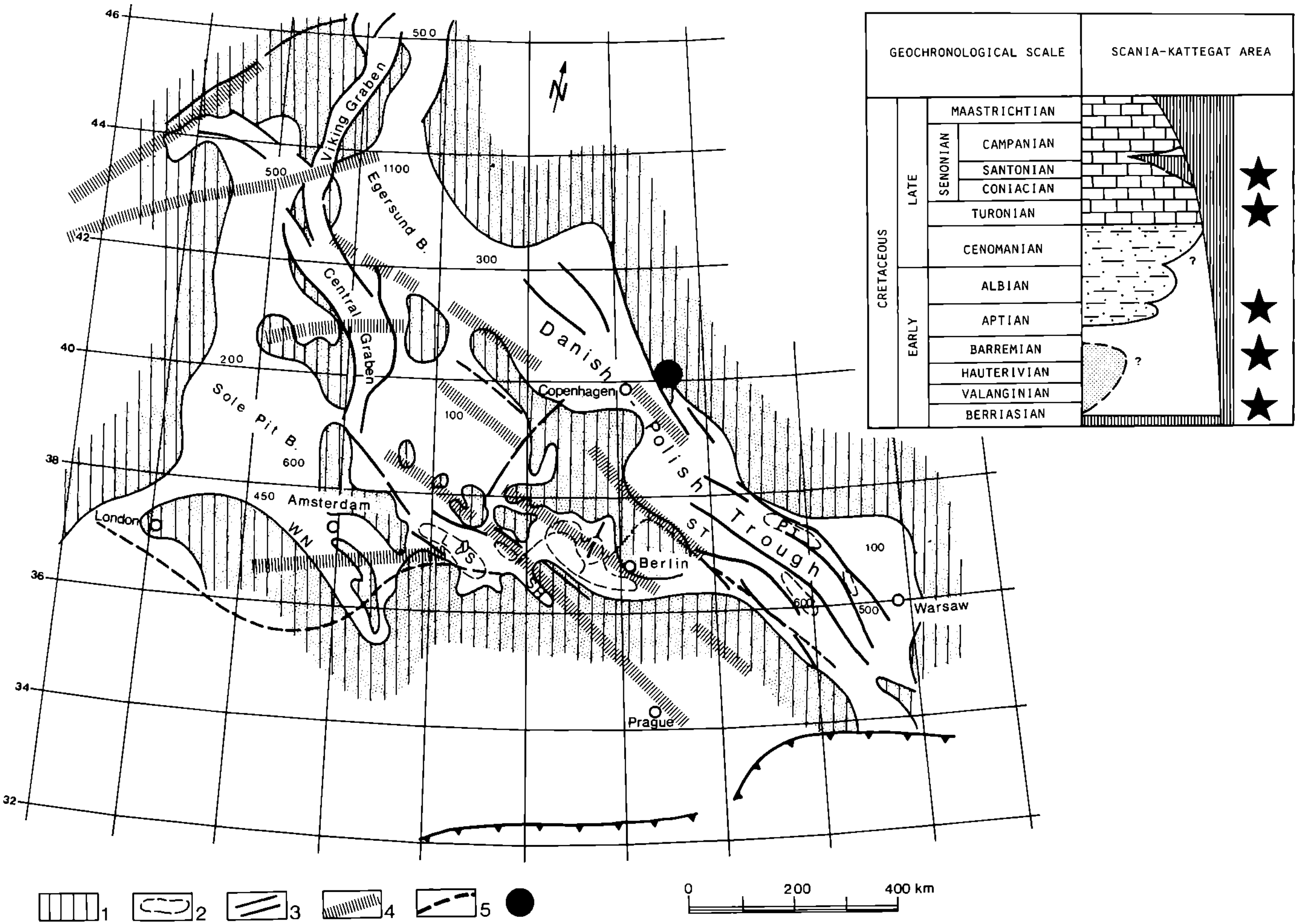
Text-Fig. 3. Paleotectonic scheme of the Lower Cretaceous in northern Europe (after SCHWAB et al. 1980). Large dot marks approximate location of the Scania volcanic eruption centers. Legend: 1 = land, 2 = thickness, 3 = main faults, 4 = lineament zones, and 5 = northern boundary of the folded Variscides.
Abbreviations: LS = Lower Saxony Basin, P = Prignitz-Westbrandenburg Trough, PT = Pomeranian Trough, SH = Sub-Hercynian Basin, ST = Szczecin Trough, and WN = West Netherland Basin.
Insert: Stratigraphic column Scania-Kattegat area (after ZIEGLER 1982: encl. 36).


GEOCHRONOLOGICAL SCALE
SCANIA-KATTEGAT AREA
CRETACEOUS
LATE
EARLY
SENONIAN
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
CENOMANIAN
ALBIAN
APTIAN
BARREMIAN
HAUTERIVIAN
VALANGINIAN
BERRIASIAN
Viking Graben
Egersund B.
Central Graben
Sole Pit B.
Danish - Polish Trough
Copenhagen
Amsterdam
London
Berlin
Warsaw
Prague
WN
LS
ST
PT
46
44
42
40
38
36
34
32
500
1100
300
200
100
600
450
0
200
400 km
1
2
3
4
5

southern Baltic into Scania and northern Denmark and beneath the southern Norwegian North Sea. Repeated volcanic intrusions of Cretaceous age are associated with this zone (ZIEGLER 1982, Encl. 21-23, 36), especially in Scania, southern Sweden.

The Scania eruption center (PRINTZLAU & LARSEN 1972, KLINGSPOR 1976, NORLING et al. 1977: 455) is characterized by the wide age span of the igneous rocks as depicted by ZIEGLER (1982, Encl. 36 - Scania-Kattegat area). The alkali olivine basalt flows and plugs of Scania are evidently petrographically similar to the Forties ankaramites (PRINTZLAU & LARSEN 1972). Conventional K-Ar whole-rock age dating shows groups between 81 and 108 Ma and between 86 and 106 Ma, and coincide broadly with the southern North Sea Mid-Cretaceous episode. PRINTZLAU & LARSEN have a remaining group between 122 and 133 Ma, whereas KLINGSPOR has two samples at 141 ± 2 and 151 ± 2 Ma. In all of the charts published by ZIEGLER (1982) this area is characterized by repeated intrusions during Cretaceous time. These facts stress the importance of the Tornquist-Teisseyre Zone during the Cenozoic. No doubt, more volcanic intrusions of different age will be detected in the future along the Tornquist-Teisseyre Zone.

Bentonites occur in the Turonian and Coniacian in northern Germany (VALETON 1959, 1960, SEIBERTZ & VORTISCH 1979) and in the Lower Maastrichtian of Rügen, GDR (STÖRR 1967).

In the GDR clay mineral analyses of Cretaceous sediments (claystones, marlstones, and sandstones) revealed the occurrence of illite/muscovite, kaolinite, mixed-layer minerals, and alkali and alkaline-earth smectites (LANDGRAF et al. 1980). Chlorite is rare (less than 5 %). 1M and 1Md glauconites are also present. In a Lower Albian claystone sample from the Altmark alkali smectite particles are extremely minute; they occur only in the < 0.63 µm clay fraction. The Lower and Middle Turonian is characterized by an extremely high smectite content. Smectite is being interpreted essentially as a detrital component within a given facies scheme. The interpretation of smectite as being of volcanogenic derivation remains restricted to the Upper Cretaceous marl horizons analysed by STÖRR (1967).

Mineral grains of zeolites (clinoptilolite) and cristobalite are reported from the Upper Cretaceous of the GDR and of the Lägerdorf-Kronsmoor-Hemmoor area (EHRMANN 1986: 34ff.). STÖRR (1967) elaborated on the clay minerals in the Upper Cretaceous of the GDR.

The paragraphs on flints and on marlstones/tuffs in Central Europe apply also to the corresponding rocks in N Europe.

3.6 E Europe

Most information on E Europe is available from Poland. In the epicontinental part of Poland north of the Carpathians, especially in the Miechow Trough, numerous bentonites and tuffites of Senonian age (mainly Lower Campanian) occur (RUTKOWSKI 1976). This bentonite zone, an important key horizon, extends nearly 300 km from Cracow to Warsaw and Lublin. It has been correlated with the Lower Campanian bentonized tuffites of NW Germany (VALETON 1960) and with coeval bentonites farther to the east in the Crimea (RUTKOWSKI 1976: 656). The bentonites are commonly associated with siliceous sediments (opoka) which are locally very glauconitic. Phosphorite nodules are also common associates.

Zeolites (clinoptilolite), normally associated with glauconite, siliceous rocks, and phosphatic rocks or with volcanic-sedimentary rocks, are widespread in the Cretaceous of the Russian Platform, Armenia, and W Kazakhstan (GOTTARDI & OBRADOVIC 1978).

Some of the clinoptilolite occurrences are definitely being interpreted as alteration products of pyroclastic material.

3.7 Alps (Helveticum)

The column Alps (Helveticum) contains only few data. First-order vestiges of igneous intrusions and pyroclastics are not known yet. Limestones and marlstones predominate in the lithological column. Characteristic, however, are the remarkable glauconite content in the Mid-Cretaceous (Aptian/Albian) and phosphorite concentrations. Siliceous sediments such as siliceous limestones (Kieselkalk) occur in the Hauterivian. Camouflaged volcanic interference cannot be excluded for the Helvetic Cretaceous eo ipso.

3.8 Iberia

The column Iberia contains data on Spain and Portugal. Cretaceous outcrops occur in the Pyrenees, Iberides, Betic Cordillera, Estremadura, and Algarve; the Pyrenees and the Betic Cordillera belonging to the Mediterranean orogenic belts. Except for biostratigraphy, data on lithology, petrography, and stratigraphic clay mineral distribution is sparse.

Major igneous events were observed in the orogenic belt of the Pyrenees (SOUQUET & DEBROAS 1980, PEYBERNES 1982: 998), especially in the Pente Milieu Océanique (AMIOT, FLOQUET, MATHEY, PASCAL, RAT & SALOMON 1982: 55, Fig. 10), and - to a minor extent - in the Betic Cordillera (VERA 1982).

Albitic trachytes, together with basic volcanic rocks, occur in the Cretaceous of the Basque Province of Guipuzcoa in northern Spain. BOESS & HOPPE (1986) also mentioned submarine volcanism near the Lower/Upper Cretaceous boundary in the Vasco-Cantabrian strike-slip fault Basin of northern Spain.

OSANN & ROSENBUSCH (1923: 463) referred to teschenites near Lisbon, Portugal, and BERTHOU et al. (1979) mentioned igneous events of Senonian and younger age. These igneous intrusives are to be seen in connection with the Atlantic border tectonics. SCHMIDT (1985) also reported sporadic igneous activity in the Cretaceous of Portugal with a major episode near the Jurassic-Cretaceous boundary (about 135 Ma).

MATHEY (1982: fig. 3.6, p. 130) reported volcanic tuff horizons from the Turonian/Coniacian boundary within the Iciar Formation of the Basque Arch (Arco Vasco), northern Spain.

Repeatedly mentioned in literature (SEGURA & WIEDMANN 1982) are horizons of green or greenish clays, claystones, and marls from the Cenomanian and Turonian of Spain. They are often barren of fossils, but may possibly contain volcaniclastic admixtures. This is concluded from lithological analogies with similar beds in Central Europe and from the extremely exposed tectonic setting of the Iberian Peninsula (FRISCH 1981, figs. 1-5).

According to the literature references available, glauconite grains and silex nodules seem to be less common than in Central Europe.

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OSANN & ROSENBUSCH (1923: 463) referred to teschenites near Lisbon, Portugal, and BERTHOU et al. (1979) mentioned igneous events of Senonian and younger age. These igneous intrusives are to be seen in connection with the Atlantic border tectonics. SCHMIDT (1985) also reported sporadic igneous activity in the Cretaceous of Portugal with a major episode near the Jurassic-Cretaceous boundary (about 135 Ma).

MATHEY (1982: fig. 3.6, p. 130) reported volcanic tuff horizons from the Turonian/Coniacian boundary within the Iciar Formation of the Basque Arch (Arco Vasco), northern Spain.

Repeatedly mentioned in literature (SEGURA & WIEDMANN 1982) are horizons of green or greenish clays, claystones, and marls from the Cenomanian and Turonian of Spain. They are often barren of fossils, but may possibly contain volcaniclastic admixtures. This is concluded from lithological analogies with similar beds in Central Europe and from the extremely exposed tectonic setting of the Iberian Peninsula (FRISCH 1981, figs. 1-5).

According to the literature references available, glauconite grains and silex nodules seem to be less common than in Central Europe.

In the sediments of the so-called Cenomanian/Turonian Boundary Event (CTBE) from the Gibraltar Arch area (THUROW, KUHNT & WIEDMANN 1982, THUROW & KUHNT 1986, THUROW 1987) volcanic interference and admixture of volcaniclastic material is likely. The CTBE rock suite resembles the Lower Carboniferous phtanitic rock suites in the Rhenish Schiefergebirge and Harz Mountains as well as the Upper Jurassic phtanitic rock suites in the Northern Calcareous Alps and in the Apennine, both with cherts, radiolarites, and tuffs. Recently, GIMENO (1987: 191) described so-called exhalites, rocks related to hydrothermal events, which much resemble these rock associations.

Similar to the wrench faulting and transform faulting of the Pyrenees with oceanic crust exposed is the mirror image of wrench faulting in the Gibraltar Arch (FRISCH 1981, Figs. 1-5). Consequently, the geological setting and the petrographic rock composition should be similar.

3.9 Tethyan Realm

In order to supplement the sparse and rather incomplete data on volcanic activity in the orogenic belt of the **Tethys**, such as in the Alps (Helveticum), striking first-order vestiges are briefly mentioned from the Tethyan Realm.

In the lower part of the Upper Cretaceous (Cenomanian and higher) picrites and picritic tuffs occur in the Klippen region of Ober-St.-Veit near Vienna (WESSELY 1974).

Similar basic rocks, namely augitites, are known in Albian sediments from the Carpathian Klippen Zone and teschenite-picrite effusions of Hauterivian-Aptian age in the Silesian and Baska Zone of the Carpathian Flysch (CSSR).

For the Cretaceous of the Carpathian Flysch of Poland, LEFELD (1976: 706-711) summarized all pertinent data: spilitic rocks and sills of basic igneous rocks (teschenites), acidic to intermediate bentonites (Barremian, Aptian, Albian, Cenomanian, Turonian, Senonian, Campanian, and Maastrichtian) frequently associated with dark shales, siliceous shales, gaize, spongolites, and radiolarites.

The pyroclastic deposits within the Carpathian Flysch mark the transition into the epicontinental Cretaceous farther to the north of Poland. The mineral composition of the tuffs resembles very much that of the tuffs in Central Germany, e. g. Sarstedt (GAIDA et al. 1978). Zircon, apatite, and titanite are characteristic heavy minerals.

In the Carpathian Flysch sediments of Poland the following Mid-Cretaceous horizons are rich in organic matter (GUCWA & WIESER 1980): Lgota Beds, Gaize Beds - Spongolites, "Manganese" and Radiolaria Shales, and Siliceous Marls. The high silica content of the Albian Lgota Beds is due to relatively large amounts of siliceous sponges and diatoms as a result of intense volcanic activity. The Cenomanian sediments within the Carpathian Flysch geosyncline formed in the period when great changes in the configuration of sea basins occurred, presumably due to the spreading out of the sea floor of the Tethys. The intense submarine volcanism in the Eastern Carpathians, the pelagic character of the sediments (Radiolaria Shales), and the appearance of clinoptilolite claystones are proof of this great change. During the sedimentation of the Upper Albian gaize strong tectonic changes took place initiated by eruptions of andesitic and andesitic-basaltic lavas. Hydrolysis of volcanic glass supplied anomalously high amounts of manganese (0.45 % in gaize beds and 1.06 % in spongolites). Silica also produced by hydrolysis of volcanic glass went into the biotic cycle, being absorbed by diatoms, siliceous flagellates and sponges (e. g. presence of α-cristobalite). Smectites prevailing in the Cenomanian sediments are the relics of the decomposition of pyroclastic rocks. Volcanic ash falls induced lithofacies changes in the Turonian siliceous marls (GUCWA & WIESER 1980).

Atlantic-type volcanic rocks of Valanginian and Aptian age occur in the Montagne Mecsek, Hungaria (VADASZ & FÜLÖP 1959).

In the eastern Carpathian of Romania, the "Black Shale Flysch" ("Flisul bisturilov negre") of Barremian-Albian age contains thin volcanic ash layers (K.-W. MÜNTZ, pers. comm.). In the Black Shale Nappe (Andia Nappe) of the Eastern Carpathian Flysch zone IANOVICI et al. (1968) refer to black shales of Mid-Cretaceous age (Vraconian-Cenomanian) associated with a thin horizon (20-40 m) of variegated shales with tuffite and arkose intercalations. Moreover, the authors refer to volcanic activity (diabases and melaphyrs) including basic tuffs (cinerites) during Barremian and Albian. FILIPESCO (1959) mentioned pyroclastics in the Santonian to Campanian and POPOV (1981) volcanogenic deposits in the Banat-Srednogorie Rift.

Upper Senonian tuffites occur in SE Bulgaria (KARASZEWSKI 1975).

In the Albian of the western USSR (Carpathians and Crimea) the glauconitic and phosphoritic rock associations of quartz sandstones, claystones, and biosiliceous sediments (spongolites, opoka, gaize) are explained by GRIGJALIS et al. (1978) to have originated by upwelling accompanied by abundant silica supply from submarine volcanoes (vulkanogene Abfuhr in das Osteuropäische Meer). Moreover, GORBUNOVA (1966) postulated after the careful study of Middle to Upper Albian tuffs from Daghestan USSR "that volcanic ash was probably the parent material of the Middle-Upper Albian black calcareous clays of predominantly montmorillonitic composition throughout the whole of the Northeast Caucasus".

According to LEBEDINSKII et al. (1978: 615) intensive volcanism took place during Albian, Cenomanian, and Turonian in the piedmont belt of the northern flanks of the Lesser Caucasus and within the North Crimean graben-shaped downwarp.

4. Provenance of tuffs and volcaniclastics

An important question is where the tuffs and volcaniclastics in the Cretaceous of Europe are derived from. Recent age determinations of Cretaceous intrusive and volcanic rocks by DIXON, FITTON & FROST (1981: 129, 133) supplement previous attempts to locate potential source regions, especially for the Aptian-Albian ash beds (JEANS et al. 1977, GAIDA et al. 1978, KEMPER & ZIMMERLE 1978, ZIMMERLE 1979, KEMPER & ZIMMERLE 1982a). Mid-Cretaceous intrusions have been also found in the Dutch North Sea.

Since three major phases of volcanic activity are known, the question about the provenance is answered in chronological order:

(1) During the **Jurassic/Cretaceous boundary eruption phase** (mainly in the Berriasian) tuffs have been derived so far:

in Speeton (England):
- from Zuidwal 1, Wolf Rock or some unknown source,
- from a yet undiscovered volcanic source in the southern North Sea, different from the Waddenzee and Wolf Rock volcanic centers (KNOX & FLETCHER 1978: 25).

in the Schüttorf 3 well (FRG):
- from deep-seated fractures like the Tornquist Line and the Fair Isle-Elbe Line (ZIMMERLE 1979: 397-399).

(2) During the **Mid-Cretaceous (Aptian/Turonian) eruption phase** tuffs have been derived so far:

in SE England:
- from the Mesozoic province of intermediate to basic alkaline volcanism located in the South-Western Approaches to Britain, e. g. Wolf Rock (COWPERTHWAITE et al. 1972: 325),
- "from the Zuidwal volcano" (JEANS et al. 1977: 41/42),
- from the "undersaturated basic magmatism" close to the axis of the North Sea (DIXON et al. 1981: 129, 133),
- from "an easterly source probably in the southern North Sea area", other areas like the English Channel and the North Channel with submarine basaltic magmatism are not excluded (JEANS et al. 1982: 153),
- "from the Wolf Rock volcano" (ZIEGLER 1982: 73, Encl. 37).

in Lower Saxony (FRG):
- from deep-seated lineaments, probably from the Waddenzee eruption center (NW Netherlands) (GAIDA et al. 1978: 98),
- from the Waddenzee eruption center (ZIMMERLE 1979: 399),
- from the "undersaturated basic magmatism" close to the axis of the North Sea (DIXON et al. 1981: 129, 133),
- from boundary faults of the Lower Saxony Basin (KEMPER & ZIMMERLE 1982a: 250).

(3) During the **Upper Cretaceous eruption phase** tuffs or tuffogene smectites of the smectite-clinoptilolite-cristobalite association have been derived so far:

in SE England:
- from volcanic centers W or NW of mainland Britain, especially from the Anton Dohrn Seamount in the Rockall Trough. "However, ashfalls responsible for the formation of the traces of smectite in the chalk matrix were probably of the general background (stratospheric) type and may have been drawn from more distant centres" (PACEY 1984).

in France:
- from volcanic material whose ultimate source area was the "active Mid-Atlantic Ridge" (POMEROL & AUBRY 1977).

in northern Germany (FRG):
- from precursory extrusions of the Skagerak Eocene volcanism or from extrusive equivalents of the Lingen, Bramsche, and Vlotho Massifs - also similar to the Upper Cretaceous tuffs and bentonites of Poland (VALETON 1959: 198ff.),
- no data as to the provenance (VALETON 1960: 41),

- from the outer zone of the Carpathian geosyncline, analogous to the Upper Cretaceous bentonites from Poland and Romania (SEIBERTZ & VORTISCH 1979: 649).

Summing up all observation and speculations it seems to be most likely that the Cretaceous tuffs and pyroclastic rocks are derived from various source areas with decreasing geological age. The older intrusions are presumably closer to the opening North Atlantic than the younger intrusions. All rocks seem to be fault-controlled, nonorogenic rift volcanics. Evaluating the stratigraphic correlation charts by ZIEGLER (1982: encl. 35-37) it becomes evident that the longest volcanic activity in the Cretaceous of Europe - as to our present knowledge - was observed in the Scania-Kattegat area of the Danish-Polish Trough (Text-Fig. 3, insert). This observation speaks for an almost continuous volcanic activity along the Tornquist-Teisseyre Zone. The paleotectonic map of the Lower Cretaceous after SCHWAB et al. (1980) stresses the importance of the Tornquist-Teisseyre Zone (Text-Fig. 3) and depicts a rather land-locked Lower Cretaceous epicontinental sea in central and northern Europe.

5. Conclusions

Literature survey and own petrographic observation over many years lead to the following conclusions:

(1) As summarized in Text-Fig. 1, **a variety of direct and/or indirect vestiges of volcanic activity** are observed in the Cretaceous of Europe, such as magmatic intrusions and extrusions, pyroclastic and volcaniclastic deposits, pyroclastic minerals and volcanic rock fragments as sedimentary components, geochemical and geophysical hints, and certain sedimentary rock and mineral associations. These vestiges were ranked into first to fourth-order vestiges according to the reliability of the criteria.
The fourth-order vestiges such as siliceous rocks including flint, smectites, phosphates and phosphorites, complex carbonate minerals (kutnahorite) and siderite, glauconite, cristobalite, and zeolites might be remotely related to volcanism and, thus, indirectly hint at volcanic activity.
Three maxima of strong igneous activity can be recognized: (1) at the Jurassic/Cretaceous boundary (early Kimmerian), (2) during Mid-Cretaceous time, and (3) in the late Cretaceous.
Text-Fig. 1 also demonstrates the Europe-wide extent of volcanic vestiges and related phenomena.

(2) A characteristic feature of Cretaceous sediments is, that tuffs and single volcaniclastic particles are in many instances thoroughly altered into crypto- to microcrystalline clay. Moreover, they were deformed and partially replaced by the microcrystalline clay matrix. Also, the chemical assimilation of these unstable components proceeds rapidly. Consequently, pyroclastic and volcaniclastic sediments associated with **camouflaged volcanism** are thought to be a common sedimentary rock type, much more widespread in the Cretaceous of Europe than commonly accepted. This observation was thoroughly corroborated by HARRISON et al. (1979: 61) stating: "the growing evidence of Mesozoic volcanism around the British Isles and in north-west Europe suggests that derived

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volcanogenic materials are probably far more common contemporaneous sediments than has been suspected in the past."
As studies in modern sediments show, ashes occur in deep-sea sediments as discrete layers or in a dispersed state. Ash beds can be obliterated or disrupted by reworking (currents, seafloor fauna, slumping) and diagenesis (post-depositional alteration). Ash beds in older deposits are therefore difficult to be recognized because they blend with the surrounding sediments as the result of these alteration processes (ZIMMERLE & GAIDA 1980).
Thus, a significant aspect of lithogenesis, namely the admixture of volcaniclastic components, is commonly not recognized or simply overlooked. Solely integrated thin-section, X-ray diffraction, scanning electron microscope, and energy-dispersive X-ray fluorescence analyses will furnish the full answer as to the lithogenesis.

(3) The igneous activity (intrusions, extrusions, and pyroclastic deposits) in the epicontinental areas of Europe consists essentially of a **bimodal, strongly alkaline, nonorogenic rift volcanism.**
The discrepancies between classifying fresh volcanic rock samples as strongly alkaline and undersaturated rocks, and highly weathered pyroclastic deposits as acidic to intermediate rocks is caused by the easy weathering and extinction of the diagnostic, but unstable feldspathoids and mafic components in the altered volcanogenic rocks.

(4) The igneous activity during Cretaceous time is associated with the **opening of the North Atlantic** (HARRISON et al. 1979, SCHWAB et al. 1980, fig. 1, ZIEGLER 1982). The highest spreading rate started in the upper portion of the Lower Cretaceous leading to the formation of the Rockall Trough. A progressive breaking-up of the European continent seems to have taken place during the opening of the North Atlantic. It started at the margin of the Proto-Atlantic in Jurassic time and reached the central portion of the European continent during Upper Cretaceous time. This progressive break-up becomes also evident by the different ages of major active faults in Central Europe (SCHWAB et al. 1980: 326, fig. 10) and by the progressive mineralization (WALTHER 1983).
The importance and longevity of the **Tornquist-Teisseyre Zone** are evident in the majority of paleogeographic reconstructions (SCHWAB et al. 1980, ZIEGLER 1982, PEGRUM 1984). Consequently, among the numerous stratigraphic correlation charts shown by ZIEGLER (1982) that of the Scania-Kattegat area shows a consecutive number of volcanic events in the Cretaceous similar to those from the Pyrenees. This tectonic setting is shown in Text-Fig. 3. The tectonic movements along the lineaments, especially those re-activated by Upper Cretaceous inversion tectonics as described by ZIEGLER (1982: 78), make them to be ideal conduits for ascending magmas, plutonic as well as volcanic rocks.
The break-up of the European continent also caused the stepwise transgressions to the south.

(5) In Aptian/Albian claystones and Upper Cretaceous marlstones and chalks **smectite** is the dominant clay mineral; it is of polygenetic origin (see discussion by HARRISON et al. 1979: 70 and KEMPER & ZIMMERLE 1982b). But in the opinion of the author a considerable amount of

smectite is volcanogenic, derived from the thorough alteration of volcanic rock fragments or volcanic glass.
Smectite or argillized volcanic rock fragments are also the precursory substance of glauconite. Glauconite occurrence climaxed in Mid-Cretaceous time when smectite supply was greatest.
For the formation of flint and zeolites a volcanic influence, direct or indirect, cannot be excluded.

(6) Knowing site, depth, and geothermal gradient of Cretaceous igneous intrusions and extrusions has important exploratory aspects, for both **hydrocarbon and mineral exploration.** Igneous bodies of a certain size can increase heat flow for a longer time and accelerate maturation of organic matter. Examples for precocious maturation are:
- the Bramsche and Vlotho Massifs in northwestern Germany (CORNELIUS 1975, VAN WIJHE et al. 1980),
- the East Groningen Massif (KETTEL 1983),
- the northern foreland of the Pyrenees (CLARET, JARDINE & ROBERT 1981: 396).

VAN WIJHE et al. (1980: 16, fig. 13) showed conclusively how the "widespread volcanic activity" in the northern Netherlands in the late Jurassic/early Cretaceous led to "a somewhat increased geothermal gradient" which definitely affected the main CH_4 generation periods.
Moreover, the hyperthermal anomaly in the Aquitanian Basin, southern France, was produced by Mid-Cretaceous igneous activity in the northern Pyrenees (CLARET et al. 1981: 396). "Ces anomalies seraient responsables de la transformation en gaz des materières organiques dans tout le bassin de Lacq au sud du synclinal d'Arzacq."
Cretaceous igneous intrusions, increased geothermal gradients, and maturation of organic matter on the one hand, and Upper Cretaceous inversion tectonics on the other hand are intimately interrelated (CORNELIUS 1975, ZIEGLER 1975: 14ff., fig. 15, VAN WIJHE et al. 1980: 16, ZIEGLER 1982: 78).
Future ore and mineral prospection likewise depends on the knowledge of ore-bringing intrusives.

(7) Recognizing volcanic events and volcanic activities helps to explain chemical anomalies (e. g. isotopes), variations in climate and changes in evolution (ZIMMERLE 1985), development of the atmosphere as well as interference between stagnant oceans, petroleum, stratabound ores and mass extinction. For instance, a carbon isotope anomaly, namely a strong positive ^{13}C excursion, was observed by HILBRECHT et al. (1986) at the Cenomanian/Turonian boundary in northwestern Germany. Likewise SCHOLLE & ARTHUR (1980) demonstrated significant fluctuations of ^{13}C in Cretaceous pelagic limestones in the Circum Atlantic-Western Tethyan region, the Gubbio area included.
In this context DEAN et al. (1986: 146) are inclined to interpret the depletion of ^{13}C in Cretaceous marine organic matter by higher concentrations of CO_2 during Middle Cretaceous. This more available CO_2 might have been responsible for the lighter $\delta^{13}C$ values of Cretaceous marine organic matter. A tectonically very active Mid-Cretaceous with increasing volcanism would make this interpretation more plausible. Also WALKER (1986: 159) postulated that isotopic and concentration data of carbon from carbonate minerals and sulfur analysed in Cretaceous samples, especially of those from the Aptian/Albian, can be

only reconciled "by a model which invokes a significant flux of hydrothermal sulfide to the deep sea, at least during the Cretaceous." According to this author, the Cretaceous was "a time of markedly perturbed geochemical cycles for carbon, sulfur and oxygen." ZIMMERLE (1985) postulated that increased volcanic activity and deposition of very fine-grained volcanic debris was one of the major factors in creating "anoxia" in the Mid-Cretaceous times. Major marine transgressions such as the one during the Mid-Cretaceous are commonly accompanied by increased volcanic activity at the subduction zones (POMEROL & AUBREY 1977). Also HERBIN et al. (1986: 390, fig. 1) advocated that "the increase of sea-floor spreading between the Cenomanian and anomaly 34 suggests intense tecto-magmatic processes on the mid-oceanic ridge during the CTBE." Global tectonics are ultimately the first-order steering mechanism for numerous phenomena in the Cretaceous, among others also for the anoxic events (ZIMMERLE 1985: 406, Fig. 7).

(8) Volcanism is normally connected with tectonic activity. Since tectonic activity, anorogenic rifting as well as orogenic folding, is observed to a rather large extent throughout the Cretaceous of Europe, more petrographic studies of Cretaceous argillaceous and calcareous sediments, especially by means of thin sections, shall reveal further vestiges of volcanism in the rock column.
Such petrographic studies should also include inorganic-geochemical analyses with the microprobe in order to identify submicroscopic minerals of volcanic derivation and vestiges of exhalative/hydrothermal activities such as the specific trace element distribution in fish scales (OUDIN & COCHERIE 1987). In addition, the evaluation of the K, U and Th content of the pelitic rocks, as determined by chemical analysis and/or by natural gamma-ray spectrometry (ENGELL-JENSEN et al. 1984, SERRA 1985), seems to be a promising approach in indirectly tracking volcanic vestiges in argillaceous sediments.

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Upper Cretaceous and Paleogene Magnetic Stratigraphy from the "Atlas Gulf" (Western High Atlas Mountains, SW Morocco)

JOSEF HEMLING, Köln

With 8 Text-Figures

HEMLING, J. (1989): Upper Cretaceous and Paleogene Magnetic Stratigraphy from the "Atlas Gulf" (Western High Atlas Mountains, SW Morocco). - In: WIEDMANN, J. (Ed.), Cretaceous of the Western Tethys. Proceedings 3rd International Cretaceous Symposium, Tübingen 1987, pp. 989-1005. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Abstract: In the western High Atlas, Turonian to Eocene sediments were sampled for paleomagnetic investigations. Alternating field and thermal demagnetization experiments revealed the polarity of the primary component of the natural magnetization. The behaviour of the earth magnetic field is described although significant secondary components were present. A well defined polarity record is observed in the Agadir and Erguita Section. The polarity pattern of these sections correlates very well with the geomagnetic time scale (LOWRIE & ALVAREZ 1981) from anomaly 34 to 22. The Long Cretaceous Magnetic Quiet Zone ranging from the base of Aptian to the top of Santonian (Anomaly 34) is interrupted by a mixed polarity during the Turonian.

The typical Upper Cretaceous reversal frequence suggests that the deposition is relatively continuous. The complete Upper Cretaceous magnetic polarity record gives a hint at the Cretaceous/Tertiary boundary which is up to now in Morocco hardly dated by biostratigraphic data.

Kurzfassung: Im westlichen Hohen Atlas wurde die Sedimentfolge vom Turon bis ins Eozän paläomagnetisch untersucht. Die Polarität der Primärkomponente der natürlichen Magnetisation wurde durch Wechselfeld- und thermische Entmagnetisierungsversuche ermittelt. Das Verhalten des Erdmagnetfeldes wird beschrieben, obgleich beträchtliche Sekundärkomponenten vorhanden waren. In den Profilen Agadir und Erguita lassen sich die magnetischen Polaritäten gut ermitteln. Die magnetischen Umkehrungen dieser Profile stimmen sehr gut mit der geomagnetischen Zeitskala von Anomalie 34 bis Anomalie 22 überein (LOWRIE & ALVAREZ 1981). Die magnetisch ruhige Zone in der Kreide, die von der Basis des Apt bis zum Top des Santon (Anomalie 34) reicht, wird während des Turons durch wechselnde magnetische Umkehrungen unterbrochen.

Die für die Oberkreide typische Frequenz der magnetischen Umkehrungen läßt eine relativ kontinuierliche Ablagerung vermuten. Der vollständige geomagnetische Bericht der Oberkreide gibt Hinweise auf die Kreide-Tertiärgrenze, die bisher in Marokko kaum durch biostratigraphische Daten erfaßt wurde.

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1. Introduction

Magnetic reversals are synchronous global events, independent from facies and environmental changes. The establishment of the magnetic polarity sequence in sections containing biostratigraphic tiepoints and the correlation of the resultant sequence with the magnetic time scale allows the section to be placed within an independent geochronological framework. This chronological framework is independent from local lithostratigraphic limits or stage evolution.

The aim of the present study is to continue the previous paleomagnetic work (KRUMSIEK 1982) for Upper Cretaceous to Lower Tertiary sedimentary series of SW Morocco, and to possibly verify a Turonian mixed polarity zone (KRUMSIEK 1982) interrupting the "Long Cretaceous Magnetic Quiet Zone" (HELSLEY & STEINER 1970). Moreover, magnetostratigraphical informations concerning the C/T boundary are expected.

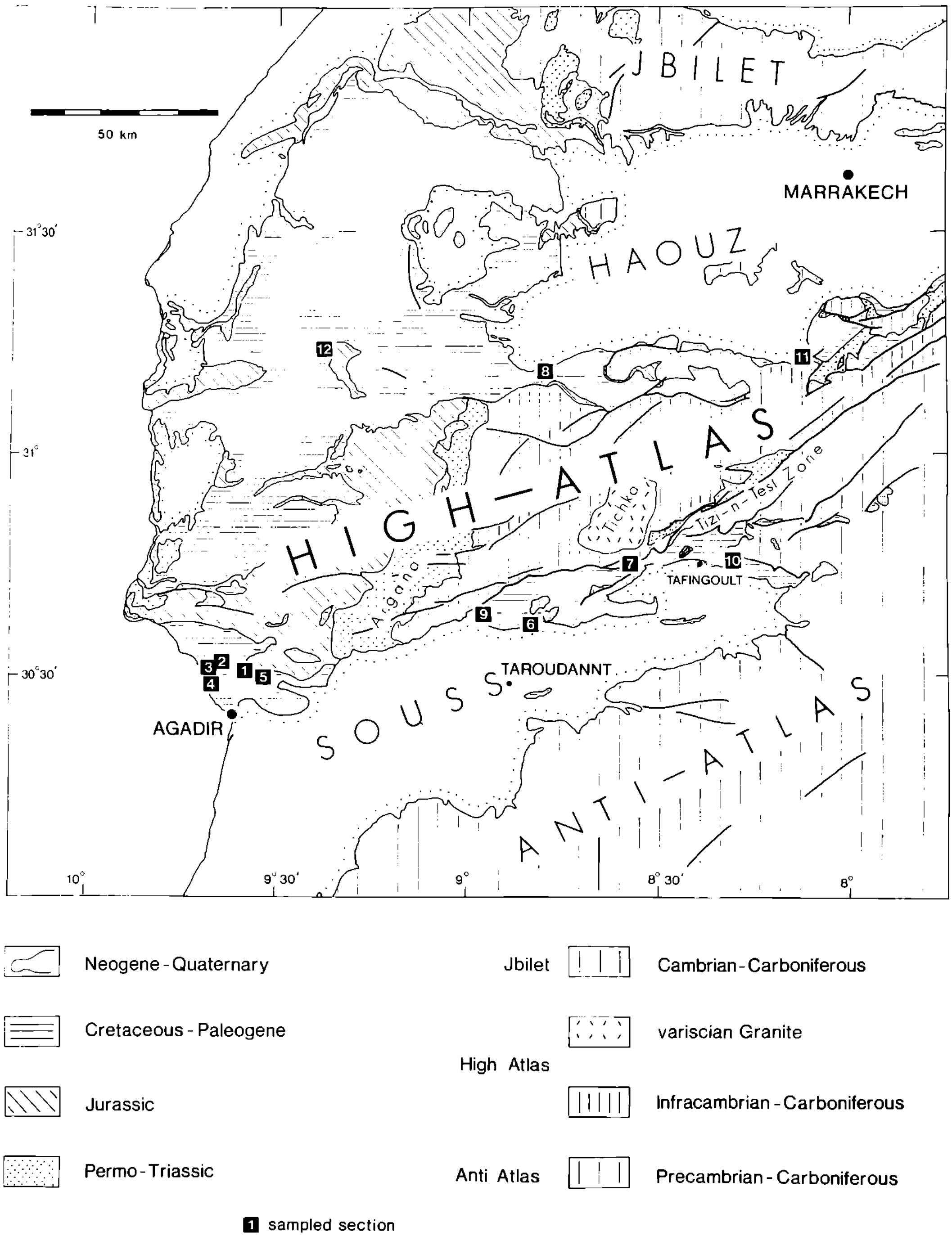
Only two sections are suitable for a biostratigraphic calibration (WIEDMANN et al. 1978, 1982, AMBROGGI 1963): the Agadir region and the southern margin of the High Atlas (Text-Fig. 1). If it is possible to evaluate a paleomagnetic correlation chart for these two sections, a composite magnetic record may be used in biostratigraphically undated sections, while at the moment, only a lithostratigraphic correlation is possible (BELOW 1977, BRIX 1976, 1981, HEMLING 1984, SCHMITZ 1977, WELLPOTT 1977).

2. Geological setting

During the Jurassic a "gulf system" (STETS & WURSTER 1982a, b, WIEDMANN et al. 1982) developed in NW Africa. The High Atlas, bordered to the north by the Moroccan Meseta as a part of the Variscan mobile belt, and to the south by the Anti Atlas as part of the African craton (Text-Fig. 1) was an area of high subsidence after the Hercynian orogenesis. Since Upper Cretaceous this belt was inverted (BRIX 1981, STETS & WURSTER 1981, 1982b).

In the western, northern and southern part of the High Atlas thick mesozoic sediments were deposited (Text-Fig. 1). These basins show typical facies and syntectonic features, and different rates of sedimentation were triggered by block faulting (AMBROGGI 1963, SCHAER 1966, MICHARD 1976, BIX 1981, WURSTER & STETS 1981, 1982a, WIEDMANN et al. 1982, BUTT 1982).

Text-Fig. 1. Schematic geological map of the investigated area (after CHOUBERT 1955).


JBILET
50 km
MARRAKECH
HAOUZ
31°30′
12
8
11
HIGH—ATLAS
31°
Tichka
Tizi-n-Test Zone
Argana
7
10
TAFINGOULT
9
6
3
2
1
5
4
30°30′
TAROUDANNT
AGADIR
SOUSS
ANTI—ATLAS
10°
9°30′
9°
8°30′
8°
Neogene - Quaternary
Cretaceous - Paleogene
Jurassic
Permo - Triassic
Jbilet
High Atlas
Anti Atlas
Cambrian - Carboniferous
variscian Granite
Infracambrian - Carboniferous
Precambrian - Carboniferous
sampled section

2.1 **Agadir section** (Text-Fig. 1, nos. 1, 2, 3; Text-Fig. 2)

The Agadir section is reconstructed from short sections (Text-Fig. 1, nos. 1, 2, 3):

Immouzzer:	Turonian - Coniacian 30°30' N; 9°37' W (Text-Fig. 1, no. 1)
Imi-n-Mekki:	Coniacian - Lower Campanian 30°32' N; 9°39' W (Text-Fig. 1, no. 2)
Tamghart:	Maastrichtian 30°31' N; 9°40' W (Text-Fig. 1, no. 3)

For detailed biostratigraphic and paleogeographic results it is referred to BUTT (1982) and WIEDMANN et al. (1978). In the following, however, only a short description of the sampled sequences is given.

In contrast to the Cenomanian marls, the Turonian is composed of carbonates. This "barre calcaire turonienne" is a distinct marker in the field. It comprises thin bedded bituminous limestones followed by siliceous limestones which interfinger offshore with black shales, i. e. typical Cenomanian/Turonian OAE (oxic anoxic event) sediments (BRUMSACK 1986, BUTT 1982, JENKYNS 1980, SCHLANGER & JENKYNS 1976, THUROW et al. 1982, THEIN 1985, WIEDMANN et al. 1978, 1982). The sandy limestones and the shell beds of the Upper Turonian indicate a progressive shallowing trend during late Turonian and Coniacian (BUTT 1982).

The Coniacian and Santonian sediments are comprised of marls, calcareous sandstones and sometimes also of large-scale crossbedded sandstones marking a regressive trend in the Atlas Basin. The Lower Campanian encompasses dolomitic crossbedded sandstones, while the Upper Campanian facies is passing into siliceous calcareous marls and yellow limestones.

Thick bedded sandstones with calcareous concretions are typical for Maastrichtian sediments in the western High Atlas.

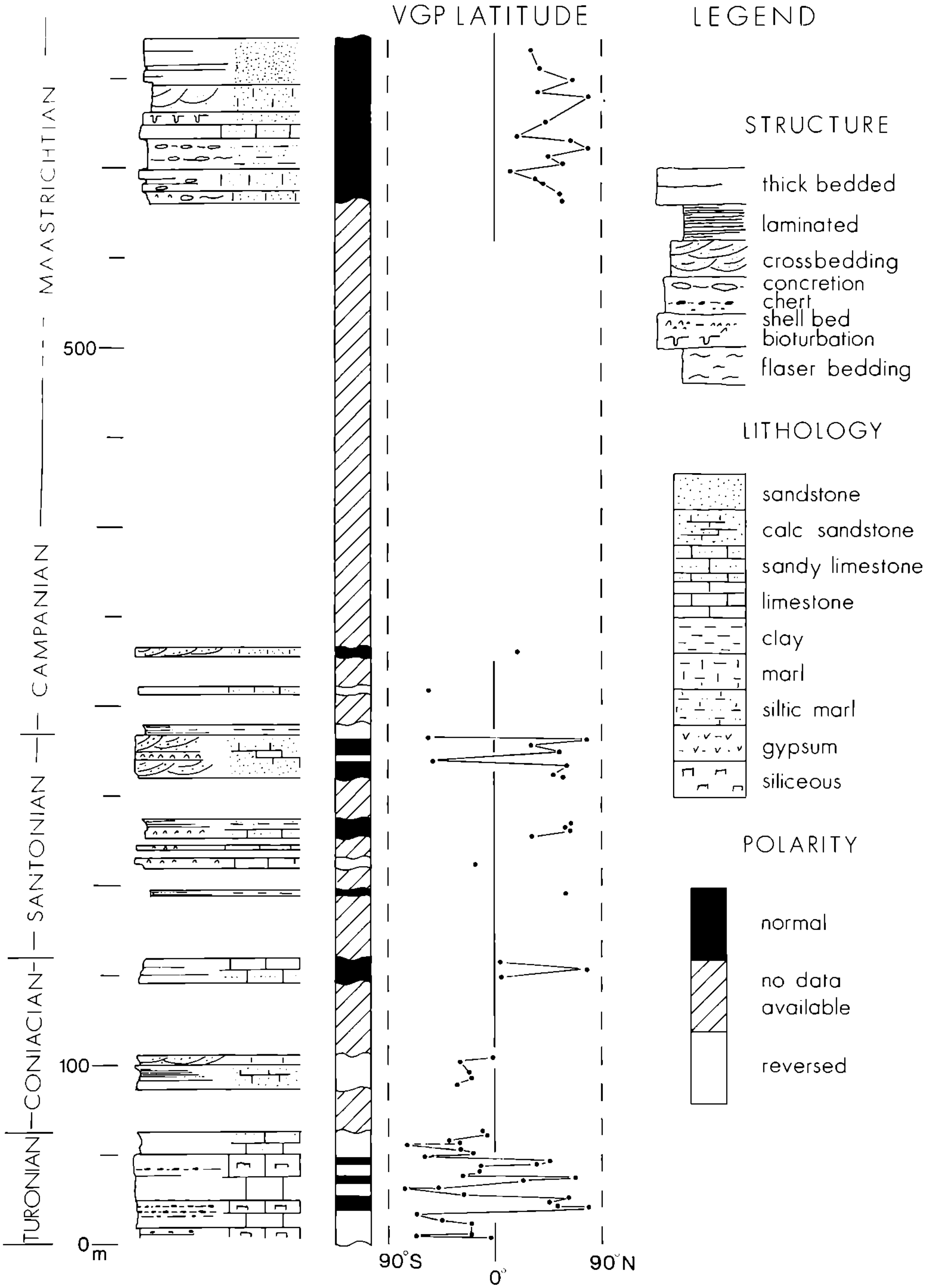
The early Tertiary marls (WIEDMANN et al. 1978) are unconformably overlain by Quarternary conglomeratic beds.

2.2 **Qued Erguita section** (Text-Fig. 1, no. 9; Text-Fig. 3)

This section, 15 km north of Taroudant, was described in detail by AMBROGGI in 1963 (Text-Fig. 1, no. 9).

The sampling started in the lowermost Santonian because the Coniacian lagoonal deposits are unsuitable for paleomagnetic sampling. The Santonian sediments (Text-Fig. 3) are characterized by limestones, which are sometimes dolomitic, by siltic marls and by two red bed layers indicating the climax of the Santonian regression.

Text-Fig. 2. Paleomagnetic data from the Agadir section. Lithologic column, interpreted polarity column and VGP latitude values are plotted against their stratigraphic thickness.


VGP LATITUDE
LEGEND
STRUCTURE
thick bedded
laminated
crossbedding
concretion
chert
shell bed
bioturbation
flaser bedding
LITHOLOGY
sandstone
calc sandstone
sandy limestone
limestone
clay
marl
siltic marl
gypsum
siliceous
POLARITY
normal
no data available
reversed
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
CONIACIAN
TURONIAN
500
100
0 m
90°S
0°
90°N

Marly limestones with siliceous concretions, alternating with sandstones, were deposited during the Campanian.

The Maastrichtian sequence of the Oued Erguita section is an important key for Upper Cretaceous facies and basin reconstruction. In the Oued Erguita Section more than 200 m of phosphatic sandstone were deposited.

This phosphatic sedimentation continued up to the Lower Eocene. The Middle Eocene is built up by a transgressive sequence of limestones. In the Upper Eocene the facies changed to continental red beds.

In general, in the southern High Atlas a facies alteration is observed from shallow marine conditions in the western parts to highly continental influenced conditions in the eastern parts. In view of the Maastrichtian phosphatic sediments, we have to discuss two different basin evolutions:

- the Agadir Basin without deposition of Maastrichtian phosphatic sediments,
- the Souss Basin with the formation of thick phosphatic sediments and higher rates of subsidence than in the Agadir region.

3. Paleomagnetic sampling and measurement

Using the time scale of HARDENBOL et al. (1982) and the sediment thickness of each of the major late Cretaceous stages, appropriate sedimentation rates have been calculated. In order to achieve a resolution equivalent to about $1 \cdot 10^5$ - $2 \cdot 10^5$ years, a paleomagnetic sample spacing of about 0.5 m in sequences with low sedimentation rates like the Turonian is required.

In both sections more than 600 cores were drilled with a diamond-tipped, gasoline powered core drill of fairly standard design. In all but a few cases the core was oriented while still attached to the rock. The remanence was measured on more than 2500 standard sized cylinders, using a UGF-4 spinner magnetometer with a noise level of $5 \cdot 10^{-5}$ A/m. An alternating field demagnetization unit was used for magnetic cleaning along three orthogonal axes, while thermal cleaning was carried out with a Schoenstedt demagnetizer.

4. Rock magnetic properties

4.1 Isothermal remanent magnetization (IRM)

The magnetic mineralogy was investigated by coercivity spectrum analysis (DUNLOP 1972). An electromagnet was used to give 50 samples a succession of isothermal remanent magnetization (IRM) in progressively increasing fields up to 0.3 Tesla. In most of the samples the highest part of the coercivity spectrum is below 0.2 T. Between 0.2 and 0.3 T a flattening of the IRM acquisition curve is observed which can probably be associated with magnetite (Text-Fig. 4). Additional support for the presence of

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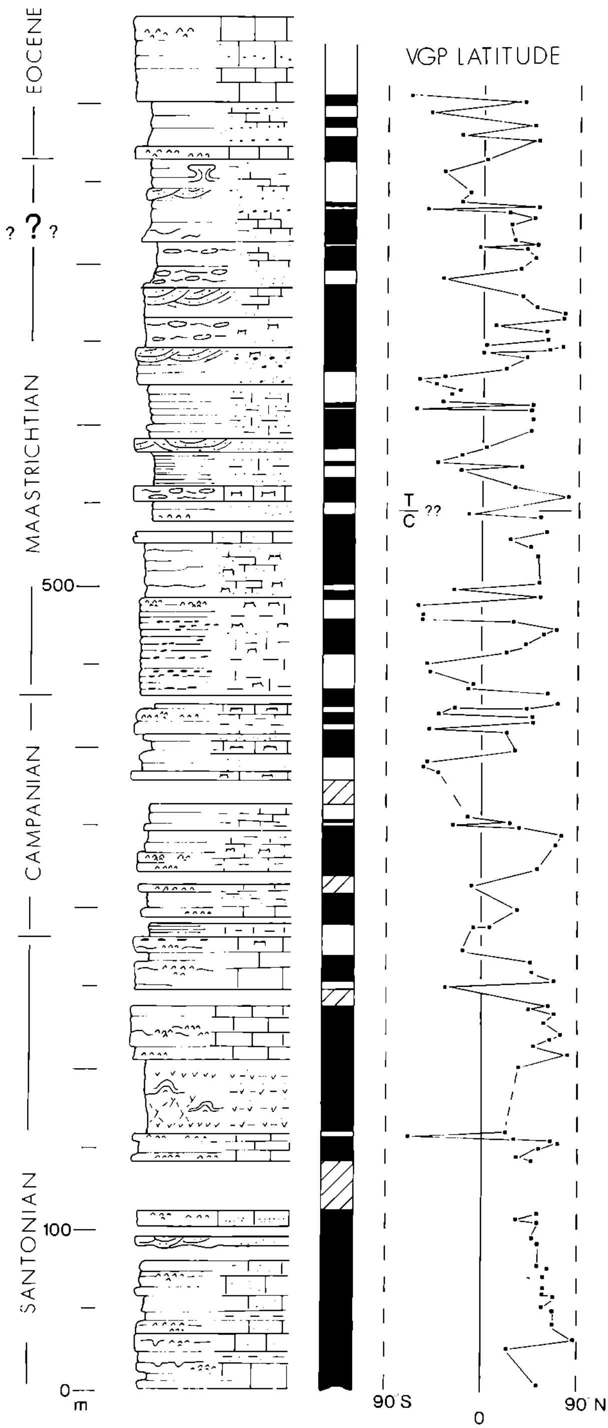
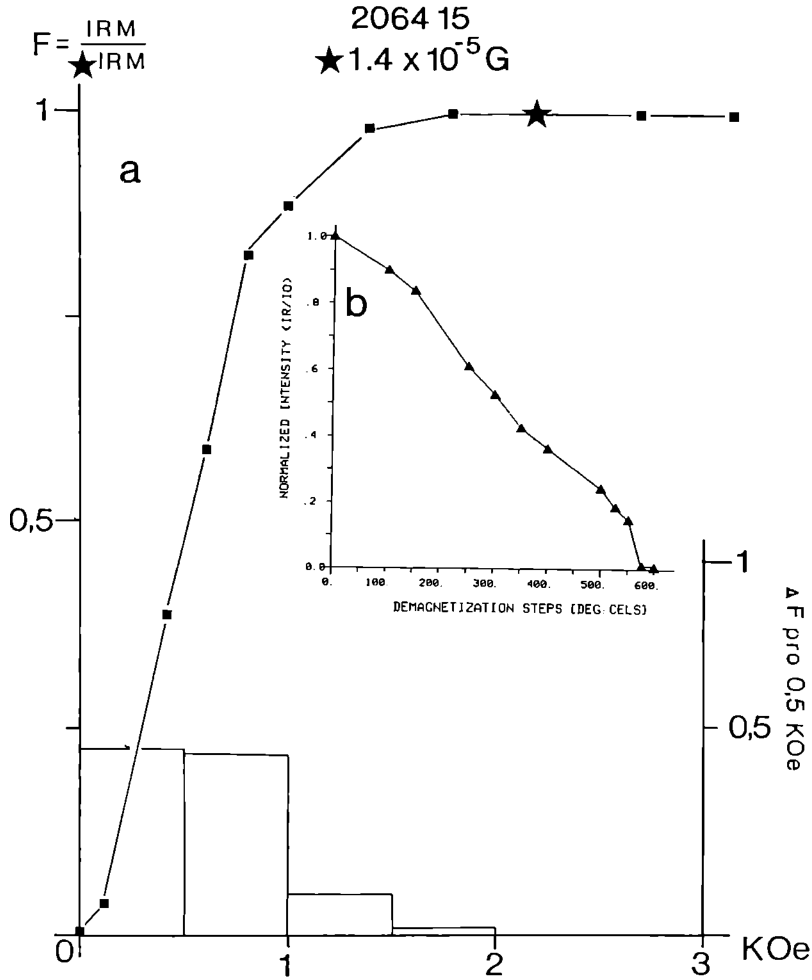
Text-Fig. 3. Paleomagnetic data from the Oued Erguita section (for explanations see Text-Fig. 2).

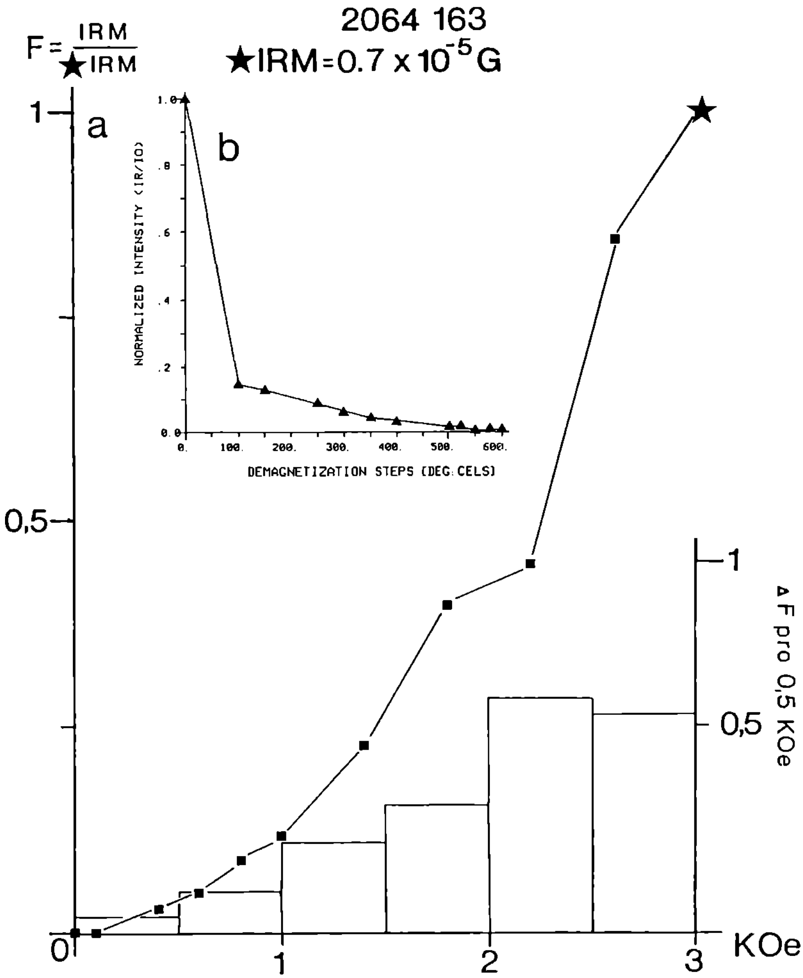

EOCENE
? ? ?
MAASTRICHTIAN
CAMPANIAN
SANTONIAN
500
100
0
m
VGP LATITUDE
T
C
??
90°S
0
90°N



Text-Fig. 4. a) IRM acquisition curve and coercivity spectrum (Histogramm). b) Continuous thermal demagnetization of the IRM. c) Room-temperature monitoring of susceptibility after stepwise heating to high temperatures.

magnetite is due to thermal behaviour of the IRM because the maximum unblocking temperature is generally reached at 550 °C (Text-Fig. 4a). In other samples goethite was found to be the dominant magnetic mineral which is also revealed in the unblocking temperature (Text-Fig. 5). In nearly all IRM experiments it is obvious that more than one carrier of remanence is present.

During the heating at high temperature an alteration of magnetic phases is monitored by susceptibility measurements (Text-Fig. 6c, 7c). It is possible to define two alteration generations. The first generation (Text-Fig. 6) occurs at a temperature above 100 °C, perhaps due to dehydration effects of clay minerals, or to crystallization of poorly crystallized goethite (HEL-



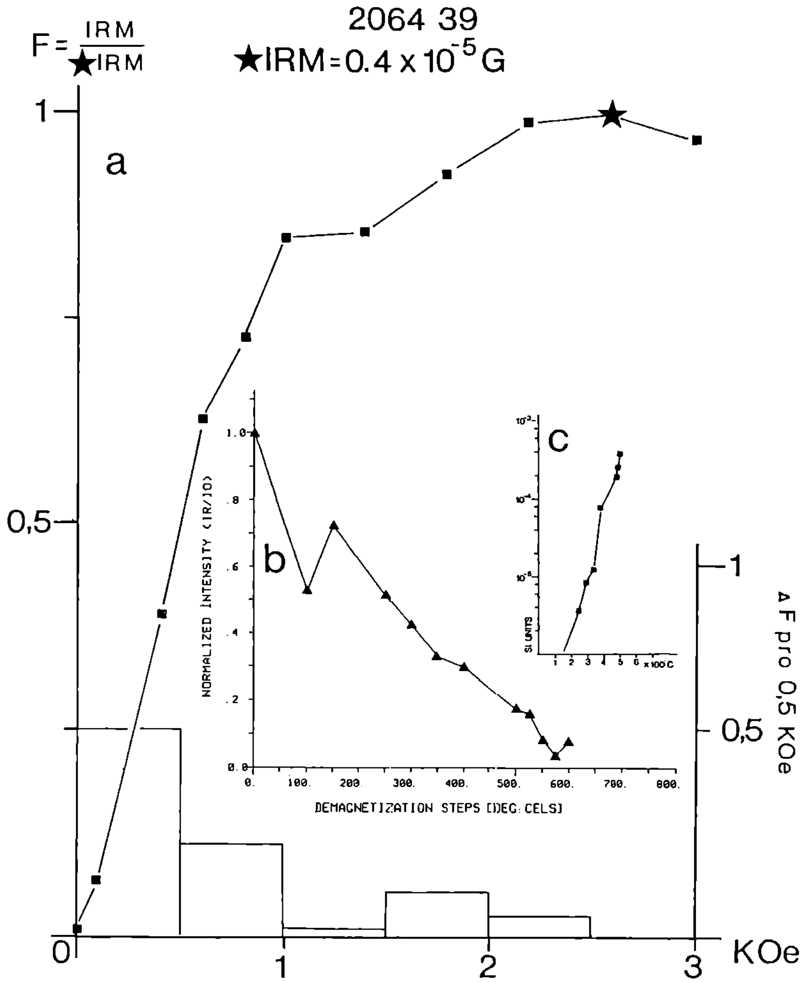
Text-Fig. 5. Explanation, see Text-Fig. 4.

LER 1978). The next alteration (Text-Fig. 7) takes place above 300 °C, the dramatic increase of susceptibility indicates the creation of a new magnetic phase. IRM acquisition characteristic after heating revealed that the new phase has a low coercivity and is probably magnetite.

4.2 Natural remanent magnetization (NRM)

The NRM intensities of rocks in both sections are between $2 \cdot 10^{-5}$ A/m typical for carbonates without any terrestrial input and $2 \cdot 10^{-3}$ A/m for large-scale crossbedded deltaic sandstones.

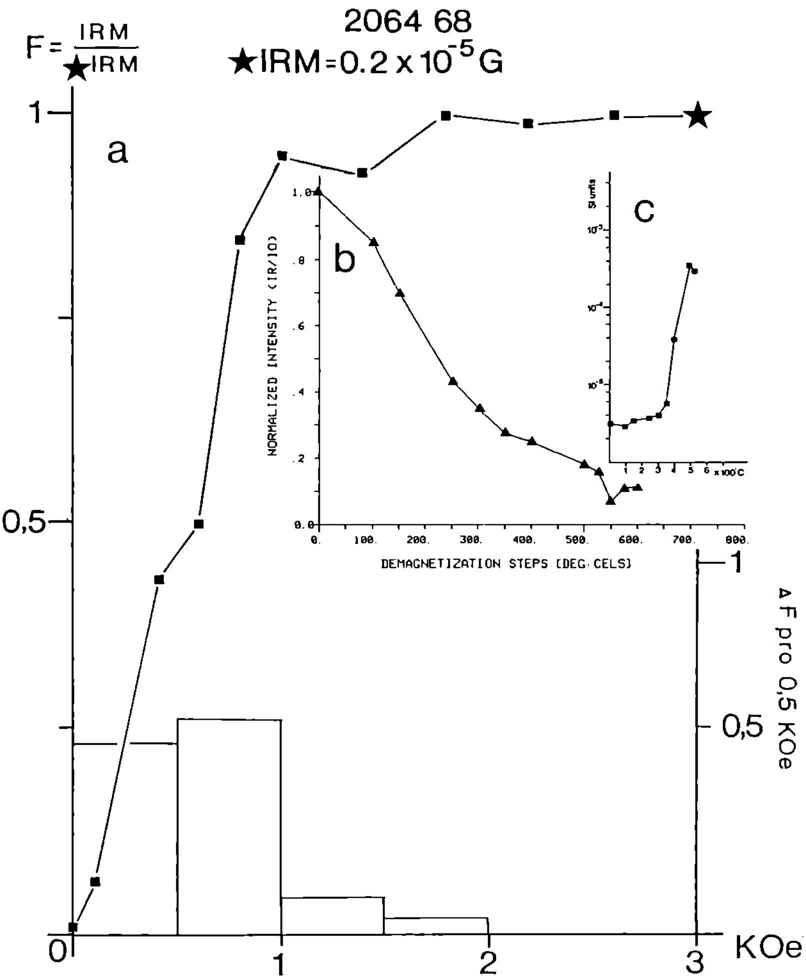
The directions of NRM are scattered around the axial dipol field of Morocco, indicating an overprinting of the paleofield direction by the present day field.



Text-Fig. 6. Explanation, see Text-Fig. 4.

The stability of NRM was analysed by progressive demagnetization of pilot specimens, at least one from each core, at relatively close intervals (usually 50 Oe or 50 °C). Stable end-point curves (ZIJDERVELD 1967) provide a clear illustration of the direction of a hard magnetic component, usually considered the primary component, which represents the normal and reversed polarity produced by the ancient Earth's magnetic field.

The normal and reversed stable directions are not exactly antipodal. These observations suggest that a perfect magnetic cleaning was probably not achieved. This indicates a hard component, which is either secondary or the resultant of more than one stable component different from the CARM (characteristic remanent magnetization).



Text-Fig. 7. Explanation, see Text-Fig. 4.

5. Magnetic stratigraphy

The declination and inclination of the stable primary component, the so-called characteristic remanent magnetization (CARM), together with the present geographic coordinates of the sections is being used to calculate the virtual paleomagnetic pole (VGP). Values of northern VGP latitude represent normal polarity whereas values of southern latitude represent reversed polarity (Text-Figs. 2, 3). The VGP latitude values define discrete zones of uniform polarities when plotted against their stratigraphic positions.

Because of secondary normal polarity overprints not completely erased by demagnetization techniques, some of the sites do not reach high negative VGP latitudes. Nevertheless, the direction during the demagnetization process clearly trends towards reversed polarity. Thus we can confidently assign a reversed polarity to the primary CARM. Text-Figs. 2 and 3 show the calculated VGP values and the polarity sequence determined from these data after optimum demagnetization.

5.1 Agadir section

According to the detailed biostratigraphic studies of the Tübingen University Research Group, the Agadir section must be considered as a standard section for the western High Atlas. The stratigraphic positions of paleomagnetic samples are determined by a sufficient lithostratigraphic correlation with the biostratigraphic section (WIEDMANN et al. 1978, 1982, BUTT 1982).

In this section a clear polarity sequence can be distinguished with reversals at the Turonian/Cenomanian boundary, in the Middle Turonian and in the Upper Turonian. This mixed polarity is in perfect agreement with older data of SW Morocco (KRUMSIEK 1982) but in disagreement with the magnetostratigraphic standard time scale.

Demagnetization experiments distinguished magnetite, haematite, goethite, and pyrrhotite as carriers of remanence.

The depositional remanent magnetization (DRM) is generally due to the alignment of detritic magnetic grains, e. g. magnetite in the direction of the ambient magnetic field. This DRM can be considered a primary magnetization component that records the direction of the geomagnetic field at the time of deposition.

Ironhydroxides or Fe-oxides like goethite or haematite are often produced during diagenesis of sediments, carrying a chemoremanent magnetization (CRM).

The Lower and Middle Turonian reversals are carried by magnetite as detrital input, the Upper Turonian NRM is sometimes carried by haematite. These reversed polarized remanence components are positively correlated to the lithological change from limestones to Mg-rich limestones or even to dolomites, which could be due to a younger CRM acquired after its deposition during dolomitization processes. The direction of the haematite component does not differ significantly from that of the magnetite component, and both were most likely acquired almost contemporaneously with deposition. A scattering of these two directions carried by magnetite or haematite is not observed. These paleomagnetic results, combined with geochemical and microfacies results (STAMM & THEIN 1982, THEIN 1985) permit the assumption that the distribution of the dolomites is controlled by facies and not by late diagenetic processes. This indicates that the NRM was acquired a short time after deposition and is therefore reliable for magnetostratigraphic determinations.

Many parts of the Agadir outcrops were either not accessible or covered by debris plains, or they were otherwise not suitable for paleomagnetic sampling. Nevertheless, within the Santonian/Campanian boundary' two reversals are well marked, with abrupt changes of hemisphere in the VGP latitude (Text-Fig. 2).

The polarization of the sampled Maastrichtian sequence is normal. Only two low latitude peaks possibly indicate a reversed component overprinted by a normal polarity component.

5.2 Oued Erguita section

In this section north of Taroudannt (Text-Fig. 1, no. 9) it was possible to sample an Upper Cretaceous to Middle Tertiary sequence without any big sampling gaps.

There is a clear difference between the lower 200 m of the section, which are characterized by normal polarity only interrupted by a short reversal at meter 155, and the uppermost Santonian where two reversals correspond to the polarity pattern at the Santonian/Campanian boundary in the Agadir section (Text-Fig. 3).

In the following Upper Cretaceous to Lower Tertiary shallow marine sediments a typical reversal frequency with long normal and short reversed polarity intervals is observed. This reversal frequency which is characteristic for the Upper Cretaceous (LOWRIE & KENT 1983) is a promising result for the determination of the Cretaceous/Tertiary boundary.

6. Conclusions

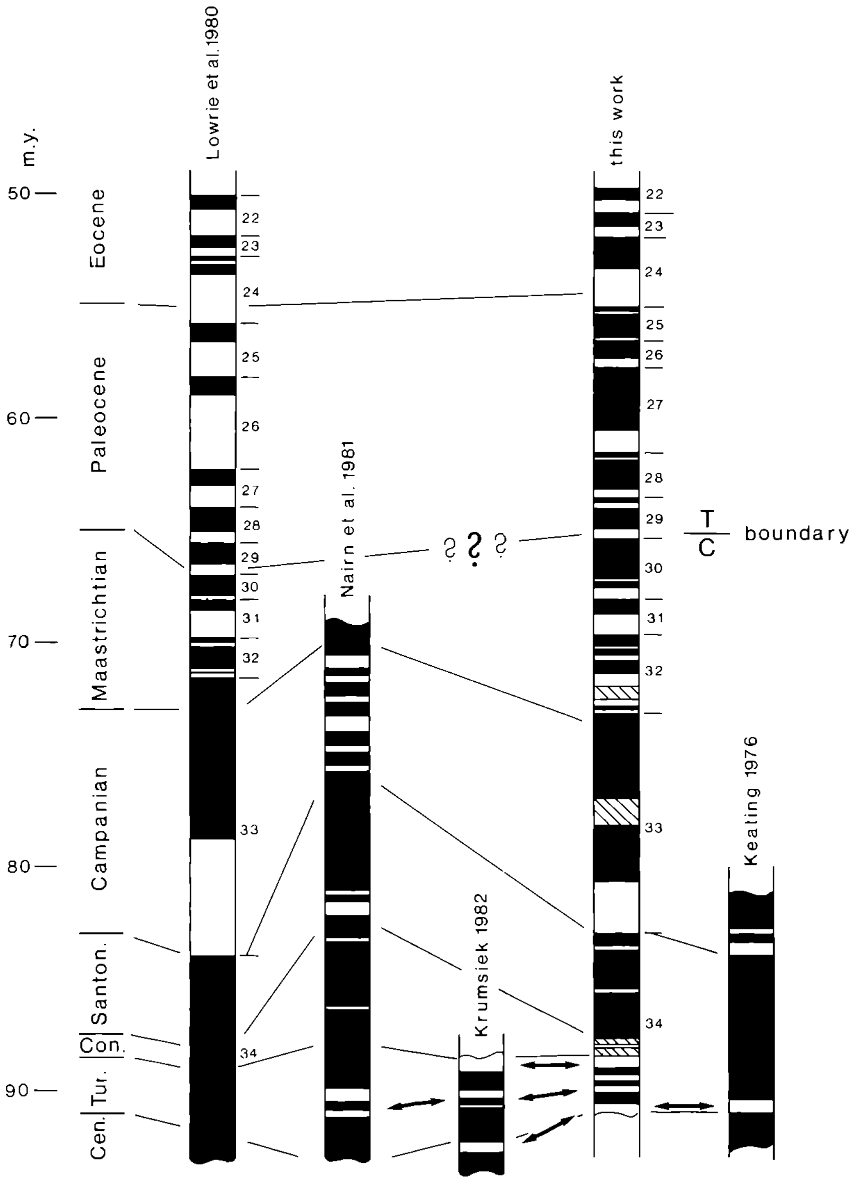
A generalized magnetostratigraphy for the late Cretaceous to the Lower Tertiary of SW Morocco and a correlation between other polarity sequences is given in Text-Fig. 8. The principal correlative lithological marker is the organic-rich anoxic/oxic event at the Cenomanian/Turonian boundary (BRUMSACK 1986, WIEDMANN et al. 1978, 1982, THEIN 1985, THUROW et al. 1982, SCHLANGER et al. 1976). The Santonian/Campanian boundary is considered as a magnetostratigraphic tiepoint age of 84 Ma (KENT & GRADSTEIN 1985), well defined as the end of the "Long Cretaceous Magnetic Quiet Zone".

The comparison of the composite magnetic reversal sequence of Morocco with data (Text-Fig. 8) from literature matches the standard magnetostratigraphy given by LOWRIE & ALVAREZ (1981) which is a compilation of Umbrian sections (e. g. Gubbio) and oceanic magnetic anomalies from Barremian to Oligocene with the only exception of the Turonian mixed polarity zone. Reversed Turonian polarity events have been observed in DSDP Site 361 (KEATING et al. 1975, 1978), especially at the Cenomanian/Turonian boundary, and in several land sections as Mexico (BÖHNEL 1985, CABALLERO et al. 1987), the Dinarian fold belt (MARTON & VELJOVIC 1987), Tunesia and - last but not least - Morocco (BEHRENS et al. 1978, KRUMSIEK 1982).

Recent studies on borehole sections such as Konrad 101 in the German Saxonian basin revealed a typical Middle Turonian reversal (LEITMANN, in prep.). The compilation of all data reveals a typical reversal at the Cenomanian/Turonian boundary which provides an important correlation device for this boundary.

All ambiguities in age determination of these reversals interrupting the Long Magnetic Cretaceous Quiet Zone call for a practical definition of the quiet zone, which will be useful in worldwide paleomagnetic correlations. It may be suggested that the Cretaceous Long Quiet Zone encompasses parts of the late Cretaceous which is characterized by predominantly normal polarity rather than exclusive normal polarity.

Even though magnetostratigraphy is only a binar code of normal or reversed polarity, it could be powerful in sequences like the phosphatic Upper Cretaceous and Tertiary sandstones where no detailed biostratigraphy has been done yet.



Text-Fig. 8. Correlation chart with other magnetostratigraphic records. Numbers are standard magnetic anomaly identification (LABRECQUE et al. 1977).

Compared to the standard magnetostratigraphic time scale, anomalies 34 to 22 are observed for the Upper Cretaceous to Middle Eocene. This complete magnetostratigraphic record suggests a relatively continuous deposition and gives further cause to confirm the Cretaceous/Tertiary boundary. Anomaly 29 enables us to divide the phosphatic sandstones, which are of Maastrichtian age according to AMBROGGI (1963), into a Maastrichtian and a Paleocene part. This magnetostratigraphic division is the first indication of the existence of Paleocene deposits in this area of the High Atlas Mts. (see also WIEDMANN et al. 1978, 1982).

In future, the chronologic framework thus developed will be used as a correlation tool for numerous sections (Text-Fig. 1, nos. 6, 7, 8, 10, 11, 12) in the eastern and northern High Atlas which are still insufficiently dated.

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