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Editors:

R.L. Hall and P.L. Smith

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Advances in Jurassic Research 2000

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Proceedings of the Fifth International Symposium on the Jurassic System, held in Vancouver, Canada, August 12-25, 1998

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R.L. Hall, P.L. Smith

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PUBLICATIONS OF THE 5TH INTERNATIONAL SYMPOSIUM ON THE JURASSIC SYSTEM

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INTRODUCTION

The Symposium

Following earlier meetings in Erlangen (1984), Lisbon (1987), Poitiers (1991) and Mendoza (1994), the 5th International Symposium on the Jurassic System, under the auspices of the Subcommission on Jurassic Stratigraphy (International Union of Geological Sciences), was convened on the campus of the University of British Columbia in Vancouver, Canada, from August 17 to 20, 1998. With 182 attendees (Subcommission members, accompanying members and volunteers) from 27 countries this was the largest Symposium to date.

Delegates to the Symposium were welcomed by the Subcommission Chairman, Giulio Pavia (Italy), and the Chairman of the Organizing Committee, Paul Smith (Canada). The ensuing Special Session was dedicated to Howard Tipper (Geological Survey of Canada) in recognition of his pioneering work on the Jurassic of western Canada; four invited speakers discussed the historical development and current status of Jurassic studies in western Canada:

- "A temporal framework for the Jurassic of western Canada" -- Paul Smith (University of British Columbia)
- "Jurassic tectonics initiate Canadian Cordilleran mountain building" -- Jim Monger (Geological Survey of Canada)
- "A chronology of Jurassic magmatism, tectonic regimes, and mineral deposition in north and central British Columbia: examples from the Iskut River, Queen Charlotte Islands, and Nechako River areas" -- Bob Anderson (Geological Survey of Canada)
- "Applying biogeography to paleogeography: the Early Jurassic tectonic evolution of western Canadian terranes" -- Martin Aberhan (Humboldt University).

Over the following three and a half days, 118 papers were read and 34 posters were displayed in the lobby of the conference venue. Special Sessions, listed below with their organizers, were devoted to specific topics:

- "Extinction and Recovery" -- Beth Carter and József Pálfy
- "Terrestrial Ecosystems" -- Christine Turner and Fred Peterson
- "Tethyan-Pacific Connections: the Hispanic Corridor" -- Giselle Jakobs and Paul Smith
- "Time-Scale Calibration" -- József Pálfy
- "Sequence Stratigraphy" -- Nicol Morton and Terry Poulton

The last afternoon of the Symposium was set aside for meetings of the Working Groups and a plenary session of the Subcommission, during which it was decided that the next Symposium would be hosted by our Italian colleagues.

Public Outreach

Because of the considerable interest in the Jurassic System, both in the Vancouver geological community and the public at large, the Organizing Committee tried to promote the Symposium to a wider audience. The Symposium had its own webpage, created and maintained by József Pálfy. In addition, two Public Sessions were presented during the meeting. One was a talk given by Jim Haggart (Geological Survey of Canada) on the geology of the Queen Charlotte Islands. The other consisted of a two-part, evening presentation by Bob Bakker (Tate Museum, Casper, Wyoming) on dinosaurs. The first part was aimed at young children, the later part at older children and adults. During these presentations Bakker demonstrated his exceptional skills as an artist, scientific communicator and entertainer, and these presentations attracted local TV, radio and newspaper interests. We thank Pamela Galloway for her article on the Symposium published in the Vancouver Sun newspaper, and Mark Forsythe of the Canadian Broadcasting Corporation for his help in promoting the Symposium on the radio.

Field Trips

Five fieldtrips, run before and after the Symposium in Vancouver, were intended to give delegates the opportunity to visit some key Jurassic exposures in western North America while experiencing beautiful mountain, desert or coastal scenery as well as indigenous culture. These trips, the leaders and numbers of participants were as follows:

	Leaders		Nº Participants
A1: Calgary to Vancouver	Russell Hall Jim Monger	Terry Poulton	40
A2: Nevada	David Taylor		10
B1: Haida Gwaii (Queen Charlotte Islands)	Howard Tipper James Haggart Elizabeth Carter	Russell Hall Giselle Jacobs József Pálfy	37
B2: Coast Mountains	Paul Smith Jim Monger Terry Poulton	Andrew Arthur Steve Irwin	14
B3: Harrison Lake	Paul Smith Jim Monger Andrew Arthur	Terry Poulton Steve Irwin	30

Publications

Three volumes were published as a result of the Jurassic Symposium. The editors acknowledge the able assistance and technical expertise of Melanie Butcher during all phases of production.

The Abstracts and Program volume (101 p.) is now out of print. Abstracts were compiled by József Pálfy. A single Guidebook (307 p.), dealing with all of the five field trips mentioned above, is now available through the bookstore of the Vancouver office of the Geological Survey of Canada (101-605 Robson St. Vancouver, B.C. V6B 5J3. Internet address: http://www.nrcan.gc.ca/gsc).

During the Symposium, 61 written papers were submitted for publication in the proceedings; each was sent to two reviewers for critical evaluation. Fifty-one papers appear in this Proceedings volume. Because of the great variety of formats used by authors for writing zonal/subzonal/biohorizon names, the Editors have chosen, for the most part, to print such names as originally used by the author(s).

Acknowledgments

The Organizing Committee particularly wishes to express its appreciation for the enthusiastic and efficient volunteer efforts of members of the Vancouver Paleontological Society. Kate Gordanier-Smith and Esther Tipper organized the program for Accompanying Members, and Kate Gordanier-Smith and Steve Irwin made available photographs of delegates and Symposium activities. A salmon barbeque in the grounds of the Anthropological Museum followed by a tour of its collections, provided a congenial social meeting place for delegates and their guests; we thank John Quackenbush of UBC Food Services for catering. Most importantly, thanks to Brenda Kiernan of the UBC Conference Centre for her organizational skills, patience and constantly cheerful demeanour.

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Timescales and Correlation

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Upper Sinemurian Ammonite Successions Based on 41 Faunal Horizons: an Attempt at Worldwide Correlation

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Keywords: Ammonites, Lias, Sinemurian, Biostratigraphy, Correlation, Faunal Horizons

Abstract: Based on Euroboreal and Alpine data we propose a new composite set of 41 ammonite faunal horizons for the Upper Sinemurian. These are considered the most complete succession potentially correlatable. The sequence of horizons includes published information that is extended and improved for the Raricostatum, Macdonnelli, and Aplanatum Subzones.

Introduction

The present paper is part of a series dealing with correlations of ammonite faunas in terms of faunal horizons for the Upper Sinemurian (Jura, Lias). Following the proposals of Page (1992; 1994) for Great Britain, Dommergues et al. (1994) for Burgundy and Great Britain and Meister (1995) for Europe including the Alps (Blau, 1998), we propose a composite series of horizons which represent the most complete and correlatable subdivision of the Upper Sinemurian to date.

Methods

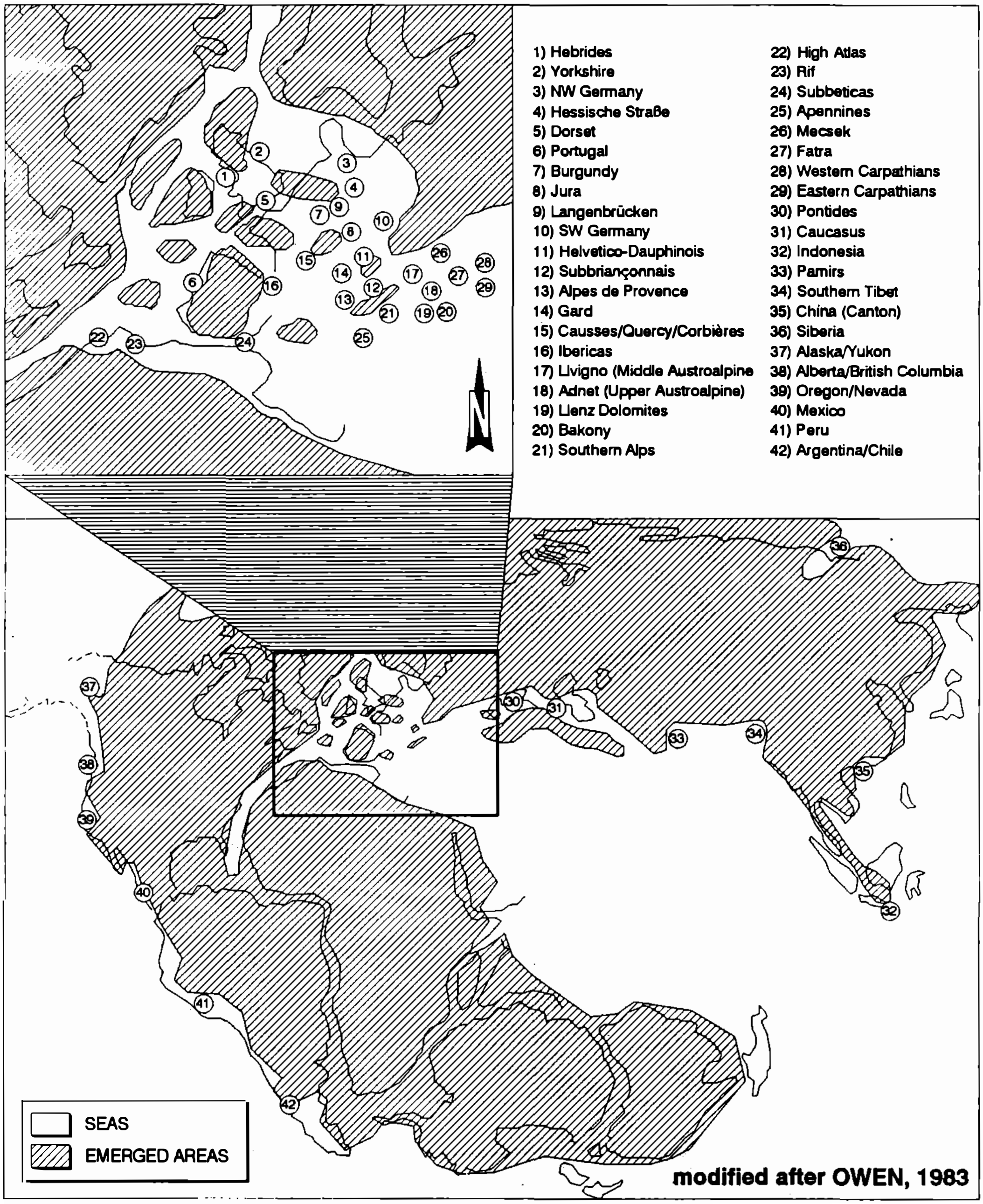
Callomon (1985, p. 624) defined the term "faunal horizon" as follows: "A faunal horizon is a bed or series of beds, characterized by a fossil assemblage, within which no further stratigraphical differentiation of the fauna or flora can be distinguished."

This concept is still applied by some authors while others use the term faunal horizon in the sense of reproducible and correlatable units, without giving formal definitions (for the Lower Jurassic see e.g. Phelps (1985), Meister (1986; 1987; 1995), Dommergues (1987), Dommergues and Meister (1991), Meister et al. (1994), Corna et al. (1997)). The basic idea of these authors is to find guide fossils or fossil assemblages which characterize the horizons. The application of this concept produces correlatable results when a consistent taxonomy is applied by the above mentioned authors.

Probably aware of the mentioned problems, Callomon (1995, p.136) modified his definition from 1985 and introduced the term "characteristic faunal horizon" which matches the latter concept: "A characteristic faunal horizon is a bed or a series of beds, characterized by a specified taxon or assemblage of time-diagnostic guide-fossils, within which no further stratigraphical differentiation of the fauna can be made."

In the present paper we use the term "faunal horizon" in the sense of this definition. It is now necessary to test this concept on a wider scale. For this purpose we have compiled a composite set of horizons which are mainly based on four representative areas (Fig. 1) in which the most complete and continuous faunal successions occur. These are the Dorset Coast for the Obtusum Subzone and part of the Stellare Subzone; Yorkshire (Robin Hoods Bay) and Burgundy for the Stellare to Densinodulum Subzones; the Dorset Coast for Densinodulum and part of the Raricostatum Subzone; and Burgundy, the Alps (Lienz Dolomites) and Yorkshire for the Raricostatum to Aplanatum Subzones (Dommergues et al., 1994; Page, 1992; 1994; Getty, 1972; Dommergues, 1993; Blau, 1998). The Standard NW European zonal and subzonal scheme (Dean et al., 1961) has been used as a reference framework because it is the best established and we feel it necessary to use only one worldwide zonation.

The data are based on numerous field observations (co)authored by one of us (C.M.) but it was also necessary to reinterpret and critically compile data of other authors. All results are presented graphically in a composite range chart (Fig. 2). In this paper we will present an overview of our results restricted to data for the "standard" scale and its key regions (Fig. 2). Additionally, examples from different paleogeographic settings (Fig. 3) are shown to illustrate the problems of correlation.



modified after OWEN, 1983

Figure 1. Liassic paleogeography and locations of the areas dealt within the text.

Comments on the Composite Range Chart (Fig. 2) and the Key Area Table (Fig. 3)

Four key regions have been chosen because together they cover the whole time interval and provide the highest possible resolution in terms of correlatable horizons. Figure 2 shows the faunal associations based on the succession of the composite set of the 41 faunal horizons primarily from the four key regions. Three periods of higher diversity can be observed. The first corresponds partly to the Obtusum Subzone (cf. *confusum* horizon) until Stellare Subzone (*blakei/arnouldi* horizon). The second period spans from the Simpsoni Subzone (*exortum* horizon) to the Oxynotum Subzone (*doris* horizon). The third period of higher diversity ranges from the middle part of the Raricostatum Subzone (*boehmi/intermedium* horizon) to the Aplanatum Subzone (*tardecrescens/romanicum* horizon).

Each of these three peaks of diversity corresponds totally or in part to one of the three major subdivisions of the Upper Sinemurian: the Obtusum, Oxynotum and Raricostatum Oppelian Zones. The first peak is characterized by the genus *Asteroceras*, the second mainly by Oxynoticeratidae and the third peak by *Paltechioceras*. These taxonomic units are distributed worldwide (Figs. 3 and 4).

The two periods of lesser diversity correspond to the upper part of Stellare Subzone and the complete Denotatus Subzone for the first period and to Densinodulum until lower part of Raricostatum Subzone for the second period. *Eparietites* and *Plesechioceras* are key taxa for worldwide correlation of faunal horizons in these intervals.

Obtusum Zone

Obtusum subzone: The highest possible resolution for the Obtusum Subzone in terms of faunal horizons is known from the Dorset Coast of southern England where four units can be distinguished (Page, 1992). By comparison with the other key regions, it is evident that the base of the Obtusum Zone can be defined by the presence of *Asteroceras confusum* (Spath).

The three basal horizons proposed by Page (1992) for the Dorset Coast are based on *Asteroceras confusum* Spath s. l. which appears in all three horizons. Moreover, there is no real difference in the faunal association. In our conception of "characteristic faunal horizons", the guide fossil of the mentioned horizons is *Asteroceras confusum* s. l.; therefore, we suppose the three units are of local significance only and may represent one correlatable faunal horizon.

The upper part of the Obtusum Subzone and the lower part of the Stellare Subzone do not seem to be represented by ammonites in the other key regions. In the Lienz Dolomites and at Adnet, the base of the zone cannot be proved because *Arnioceras ceratitoides* (Quenstedt) and *Arnioceras* sp. can indicate either the Semicostatum or Obtusum Zones. In the Apennines, an association of *Asteroceras* sp. and *Arnioceras* spp. may represent the Subzone. The presence of the three first horizons in Yorkshire (Robin Hoods Bay) is confirmed by *Asteroceras* gr. *confusum* (reinterpretation of *Asteroceras obtusum* from the Bairstow collection). In Chile, the Obtusum Subzone is represented at least by *Asteroceras* cf. *obtusum* (Hillebrandt, 1981).

Stellare subzone: The highest resolution in terms of faunal horizons for the Stellare Subzone is again known from the Dorset Coast where seven local units are distinguished (Page 1992; 1994). The base of the Stellare Subzone is only found in Dorset with the first *Asteroceras* gr. *stellare* (Sowerby) in the *Galaticeras* (now *Bouhamidoceras*, Page, 1995) horizon. The upper part of the Stellare Subzone is known from Yorkshire and Burgundy, with correlation between these areas based on the presence of *Aegasteroceras blakei* (Spath) and *Aegasteroceras sagittarium* (Blake). At the Dorset Coast, a big faunal gap occurs between the *stellare* and *lymense* horizons in the Raricostatum Zone. At Robin Hoods Bay, two horizons (*blakei* and *sagittarium*) are present.

At Adnet (Meister and Böhm, 1993; Dommergues et al., 1995), the Stellare Subzone is represented by two local horizons: *retusum* and aff. *saltriense* horizons, corresponding to an interval comprising *Galaticeras* to *arnouldi* horizons for the first, and the *blakei*/aff. *arnouldi* for the second horizon. In the Lienz Dolomites, the first local horizon which can be attributed to the Obtusum Zone with certainty is the *suevicum* horizon, corresponding to *blakei/arnouldi* horizon. In the Apennines, the presence of *Asteroceras varians* and *Asteroceras margarita* (Cecca et al., 1987) is interpreted by us to indicate the Stellare Subzone because these ammonites show more evolved morphologies in the *Asteroceras* lineage. The exact biostratigraphic position of these ammonites in the framework of faunal horizons is not yet known. We interpret the data given by Palfy and Schmidt (1994) and Pálfy et al. (1994) for western Canada as representing three faunal horizons belonging to the Stellare Subzone. In Argentina and Chile, the Stellare Subzone seems to be represented by at least two faunal horizons: the first characterised by *Epophioceras* and

"*Microderoceras*", and the second by *Asteroceras* cf. *stellare* (Quinzio-Sinn, 1987; Riccardi *in* Westermann, 1992).

Denotatus subzone: In the Denotatus Subzone we distinguish four horizons based on data from England (Yorkshire) (Page, 1992) and France (Jura, Burgundy) (Corna, 1987; Dommergues, 1993). A complete succession of these horizons has not been observed at any other locality until now. The proposed succession of *fowleri* and *denotatus* horizons is due to our interpretation of the data presented by Dommergues, Page and Meister (1994).

Species of *Eparietites* are guide fossils for the Denotatus Subzone. There is no obvious correlation between the successions of Yorkshire and Burgundy. In the Alps, *Eparietites* aff. *glaber* (Guérin-Franiatte) is known from the Subbriançonnais (Dommergues et al., 1990).

Eparietites glaber is known from the Lienz Dolomites and Adnet (Dommergues et al., 1995; Blau, 1998). Hillebrandt (1981) figures "*Eparietites* cf. *undaries*" (Simpson) and Quinzio-Sinn (1987) *Eparietites denotatus* (Simpson). Their whorl sections suggest that these specimens do not belong to either *E. undaries* or to *E. denotatus*. If they represent true *Eparietites*, they would prove the Denotatus Subzone in South America.

Oxynotum Zone

The highest known resolution for the Oyxnotum Zone in terms of faunal horizons is from England (Page, 1992) and Burgundy (Dommergues et al., 1994). The Oxynotum Zone is characterized in South America (Argentina, Chile) by *Palaeoechioceras* sp. and *Oxynoticeras* sp.

Simpsoni subzone: Three horizons can be recognized in the Simpsoni Subzone. On the Yorkshire coast the first and second horizons can be distinguished; the second and third horizons occur in Burgundy.

Oxynotum subzone: We propose a sequence of three horizons for the subzone, which is a combination of the horizons present in Yorkshire and Burgundy. *Oxynoticeras oxynotum* (Quenstedt) ranges through the whole subzone. The *oxynotum* horizon is characterized by *Oxynoticeras oxynotum* without *Bifericeras* and can be observed in Yorkshire. The association of *Bifericeras* (gr.) *bifer* (Quenstedt) and *Oxynoticeras oxynotum* characterizes the *bifer* horizon and is known from Burgundy and Yorkshire. The third association in the *doris* horizon can be observed in sequence only in France.

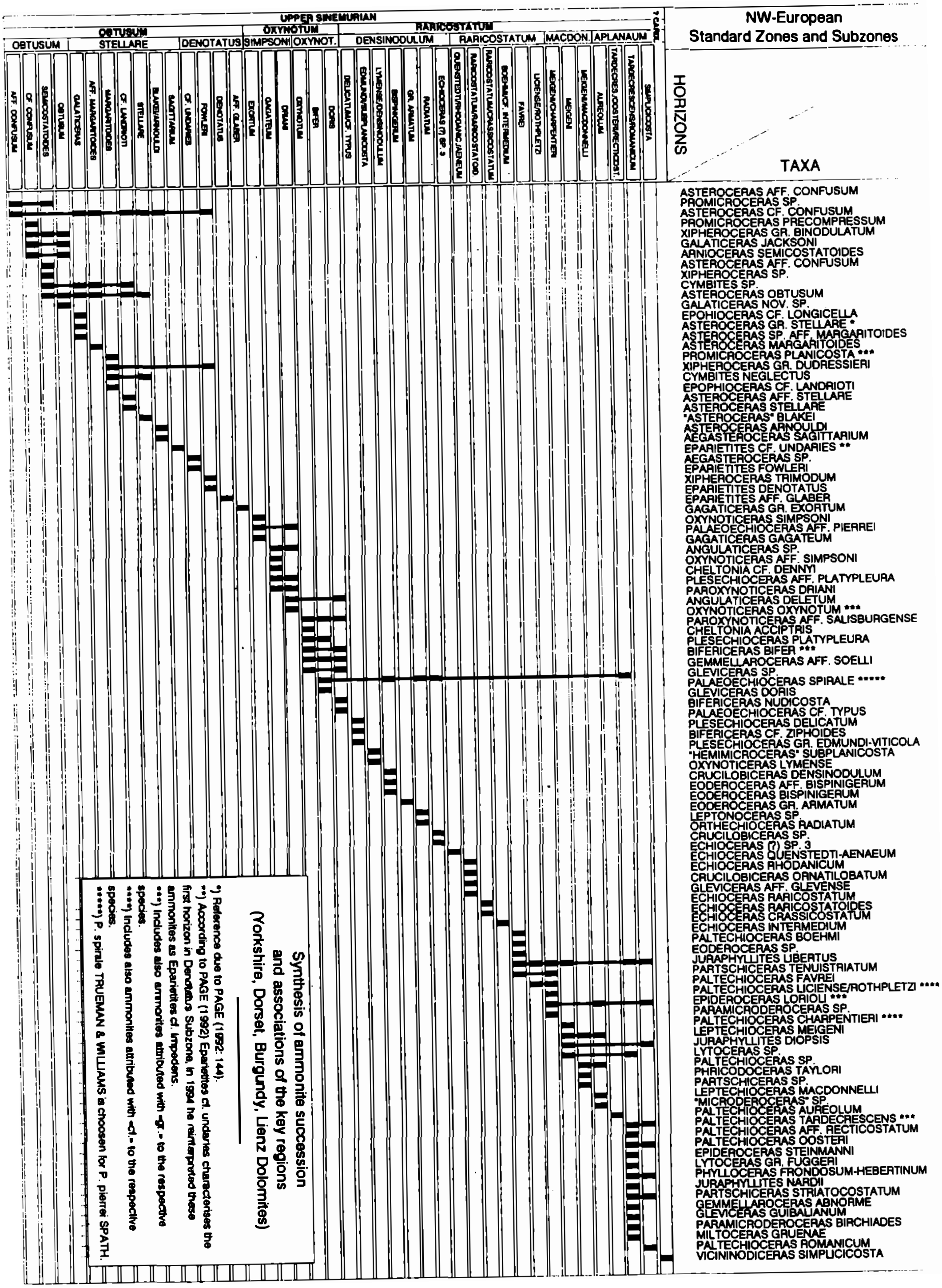
In Yorkshire the occurrence of *G. subguibalianum* is constrained at its base by the *B. bifer* horizon (Oxynotum Subzone) and at the top by the *C. densinodulum* horizon (Raricostatum Zone). In the Lienz Dolomites, the occurrence of *Paroxynoticeras salisburgense* (Hauer), and partly *Gleviceras rigidum* (Pia), are correlated with the last two horizons of the Oxynotum Subzone (Blau, 1998).

Raricostatum Zone

Densinodulum subzone: Only the base and top of the Densinodulum Subzone are well expressed in Burgundy where the two basal and the terminal horizons can be observed. The middle and upper part of the subzone can be recognized at the Dorset coast, the latter again not correlatable with other areas. As already mentioned, the Yorkshire occurrence of *G. subguibalianum* is constrained at its base by the *B. bifer* horizon and at the top by *C. densinodulum* horizon. In the Lienz Dolomites (Blau, 1998), the presence of *Gleviceras rigidum* (Pia) either indicates the topmost part of the Oxynotum Subzone and/or part of the Densinodulum Subzone. At Adnet, *Parasteroceras* sp. and *Microderoceras gigas* (Quenstedt) can be referred, with considerable uncertainty, to the Densinodulum and/or partly to the Raricostatum Subzone. *Plesechioceras arcticum* (Frebold) from Argentina and Chile is also correlated with the lower part of this subzone. The presence of *Plesechioceras* is known from British Columbia and indicates the lower part of the Densinodulum Subzone.

Raricostatum subzone: The finest resolution for the Raricostatum Subzone can be found in Burgundy (Dommergues, 1993) for the base and in the Lienz Dolomites for the upper part. The *subplicatum* horizon of Burgundy corresponds to the two topmost horizons of the Lienz Dolomites, because we interpret *Leptechioceras subplicatum* (Trueman and Williams) to represent an intermediate taxon between the *Paltechioceras favrei-boehmi* (Hug) group and *L. meigeni* (Hug).

Figure 2. Faunal associations based on the succession of the composite set of faunal horizons.



Synthesis of ammonite succession
and associations of the key regions
(Yorkshire, Dorset, Burgundy, Lienz Dolomites)

*) Reference due to PAGE (1992: 144).

**) According to PAGE (1992) Eparietites cf. undaries characterises the first horizon in Denotatus Subzone, in 1994 he reinterpreted these ammonites as Eparietites cf. impedens.

***) Includes also ammonites attributed with «gr.» to the respective species.

****) Includes also ammonites attributed with «cf.» to the respective species.

*****) P. spirale TRUEMAN & WILLIAMS is choosen for P. pierrei SPATH.

In Burgundy, two successive assemblages are observed. The basal assemblage comprises *E. raricostatum* (Zieten) and *E. raricostatoides* (Vadasz), the second only *E. raricostatum* (Zieten) including the variant *crassicostatum* (Trueman and Williams). From Robin Hoods Bay, three horizons are known (Dommergues and Meister, 1992b; Getty, 1972).

In Dorset, only the two basal horizons are present. In the Lienz Dolomites, all horizons except the *raricostatum/crassicostatum* horizon are present and, for the upper part of the subzone, are recognized for the first time. In the Apennines, only the upper part of the Raricostatum Zone is recorded with *Paltechioceras boehmi* (Hug) (here reinterpreted for the *Paltechioceras* ?aff. *rothpletzi* of Cecca et al., 1987), partly *Parasteroceras pulchellum* (Fucini) and *Paltechioceras* aff. *rothpletzi* (Böse) (*sensu* Dommergues, Ferretti, and Meister, 1994). Faunal successions in Canada (British Columbia, Pálfy et al., 1994) can be only doubtfully correlated with ? *Paltechioceras* aff. *boehmi* and *Paltechioceras* cf. *rothpletzi* in the top of the Raricostatum Subzone.

Macdonnelli subzone: The highest local subdivision in terms of faunal horizons can be found in the Lienz Dolomites where three horizons can be recognized. At Robin Hoods Bay, the presence of *Leptechioceras meigeni* (Hug) alone may represent the *charpentieri* and *meigeni* horizons; the topmost *macdonnelli* horizon is indicated by the index. In the Upper Austroalpine of the Adnet area, *Leptechioceras meigeni* represents the Subzone. In the Apennines some *Parasteroceras* (*Parasteroceras* sp. A., *P. pulchellum* (Fucini)) could indicate the presence of the Macdonnelli Subzone. For North America (British Columbia) the occurrence of *Paltechioceras* sp. could indicate an interval from the Macdonnelli to Aplanatum Subzone.

Aplanatum subzone: At Robin Hoods Bay, the *aureolum* and *tardecrescens* horizons are present (Dommergues and Meister, 1992b) and in Burgundy the *recticostatum* and *tardecrescens* horizons represent the subzone (Dommergues, 1993). In the Lienz Dolomites, the *oosteri* horizon is represented by three local units: *tardecrescens*, *gruenae* and *oosteri* horizons. The *romanicum* horizon is the uppermost Sinemurian horizon. It is also present in the Apennines. *P. insigne* (Trueman and Williams) from the Adnet area is correlated with the last two units of the Subzone. The last horizon of the Subzone is indicated by *P.* aff. *romanicum* (Uhlig) associated with *P.* aff. *tardecrescens* (Hauer). The occurrence of *Paltechioceras* cf. *aureolum* (Simpson) from Chile (Hillebrandt, 1981) may indicate the Aplanatum Subzone. *Paltechioceras* sp. from British Columbia is correlated to the two last Subzones without precision.

The *tardecrescens/romanicum* horizon probably represents the last faunal horizon of the Sinemurian above which no echioceratids have been proven. The superjacent *simplicicosta* horizon yields no accompanying fauna to determine whether it belongs to the latest Sinemurian or to the earliest Carixian. The only argument to place the *simplicicosta* horizon in the uppermost Sinemurian is the occurrence of *Vicinodiceras simplicicosta* (Trueman) and *Vicinodiceras tomfryi* (Donovan) 10 m below the first *Apoderoceras* at the Allt Fearns section on the Isle of Raasay, Scotland. The occurrence of *Apoderoceras* for us indicates the first unit of the Pliensbachian. Therefore we consider that the *donovani* horizon of Yorkshire (Dommergues and Meister, 1992b) belongs to the Carixian.

Conclusions

The succession of horizons proposed for the Dorset Coast is, in great part, not correlatable with horizonal successions in the other key regions. This is mainly due to the horizon and taxonomic concepts applied by Page (1992; 1994). In our concept of correlatable horizons, we regrouped some of the horizons proposed by this author which are often based on the acme of very long ranging taxa. This first attempt at transcontinental correlation stresses (a) the homogeneity in taxonomy mainly at the species level and in spite of quite strong faunal provincialism and (b) the application of similar concepts in biostratigraphic analysis. Worldwide correlations at the genus level are possible but the resolution is restricted more or less to zonal or subzonal rank.

The approach of North American stratigraphers is based on zones and first and last appearances of taxa. We cannot use this concept if we want to correlate in terms of faunal horizons, therefore we tried to reinterpret the American data. We have to work with species associations bed-by-bed (it is possible to do this in the field as we saw during the Vancouver meeting excursions). Unfortunately only few papers present bed-by-bed data.

Figure 3. Ammonite horizons of the selected key areas.

NW-EUROPEAN ZONATION			HORIZONS	DORSET	YORKSHIRE	BURGUNDY	LIENZ DOLOMITES (ALPS)
? CARIXIAN			SIMPLICICOSTA		B. DONOVANI + APODEROCERAS SP.		
UPPER SINEMURIAN	Raricostatum Zone	Aplanatum Subzone	TARDECRESCENS/ROMANICUM		P. TARDECRESCENS	P. TARDECRESCENS	P. TARDECRESCENS/P. ROMANICUM
			TARDECRES./OOSTERI/RECTICOSTATUM			P. RECTICOSTATUM	P. TARDECRESCENS/P. OOSTERI/M. GRUENAE
			AUREOLUM		P. AUREOLUM		
		Macdonnelli Subzone	MEIGENI/MACDONNELLI		L. MEIGENI + L. MACDONNELLI	L. MEIGENI + L. MACDONNELLI	L. MEIGENI/L. MACDONNELLI
			MEIGENI		L. MEIGENI	L. SUBPLICATUM + L. MEIGENI	L. MEIGENI
			MEIGENI/CHARPENTIERI				L. MEIGENI/P. CHARPENTIERI
		Raricostatum Subzone	LICIENSE/ROTHPLETZI				P. LICIENSE
			FAVREI				P. FAVREI
			BOEHMI/CF. INTERMEDIUM		E. INTERMEDIUM	P. BOEHMI	P. BOEHMI
			RARICOSTATUM/CRASSICOSTATUM	E. GR. RARICOSTATUM	E. CF. RARICOSTATUM	E. CRASSICOSTATUM	
			RARICOSTATUM/RARICOSTATOIDES			E. RARICOSTATUM	E. RARICOSTATOIDES
			QUENSTEDTI/RHODANICUM/AENEUM	E. GR. QUENSTEDTI	E. QUENSTEDTI	E. RHODANICUM	E. QUENSTEDTI
		Densinodulum Subzone	ECHIOCERAS (?) SP. 3	ECHIOCERAS (?) SP. 3		ECHIOCERAS SP. 3	
			RADIATUM	"O." RADIATUM			
			GR. ARMATUM	E. GR. ARMATUM			
			BISPINIGERUM	E. BISPINIGERUM			
			LYMENSE/DENSINODULUM	O. LYMENSE	C. DENSINODULUM		
			EDMUNDI/SUBPLANICOSTA		G. SUBGUIBALIANUM	P. EDMUNDI+"B." SUBPLANICOSTA	G. RIGIDUM
			DELICATUM/CF. TYPUS			P. DELICATUM	
	Oxynotum Zone	Oxynotum Subzone	DORIS			O. DORIS + B. NUDICOSTA	P. SALISBURGENSE
			BIFER		B. BIFER	O. OXYNOTUM + B. BIFER	
			OXYNOTUM		O. OXYNOTUM		
		Simpsoni Subzone	DRIANI			P. DRIANI	
			GAGATEUM		G. GAGATEUM	G. GAGATEUM	
			EXORTUM		G. EXORTUM		
	Obtusum Zone	Denotatus Subzone	AFF. GLABER			E. AFF. GLABER	E. GLABER
			DENOTATUS		E. DENOTATUS		EPARIETITES SP.
			FOWLERI			E. FOWLERI	
			CF. UNDARIES		E. CF. UNDARIES		
		Stellare Subzone	SAGITTARIUM		A. SAGITTARIUM	A. SAGITTARIUM	
			BLAKEI/ARNOULDI		A. BLAKEI	A. BLAKEI	A. SUEVICUM ? ?
			STELLARE	A. STELLARE			
			CF. LANDRIOTI	E. CF. LANDRIOTI			
			MARGARITOIDES	A. MARGARITOIDES			
			AFF. MARGARITOIDES	A. AFF. MARGARITOIDES			
			GALATICERAS	GALATICERAS			
		Obtusum Subzone	OBTUSUM	A. OBTUSUM			? ? A. CERATITOIDES ↓
			SEMICOSTATOIDES	A. SEMICOSTATOIDES	A. GR. CONFUSUM		
			CF. CONFUSUM	A. CF. CONFUSUM			
			AFF. CONFUSUM	A. AFF. CONFUSUM		A. AFF. CONFUSUM	

The Upper Sinemurian represents a time interval of about 3.5 Ma (Gradstein and Ogg, 1996), which means an average of about 85,000 years for the duration of one horizon. This value is a rough approximation because in the sense of Callomon (1985; 1995), faunal horizons are not contiguous biochronologic units; they might represent either very short or even longer time slices.

References

BLAU, J. 1998. Monographie der Ammoniten des Obersinemuriums (Lotharingium, Lias) der Lienzer Dolomiten (Österreich): Biostratigraphie, Systematik und Paläobiogeographie. Revue de Paléobiologie, 17/1:177-285.

CALLOMON, J. H. 1985. Biostratigraphy, chronostratigraphy and all that-again! p. 612-624. *In* O. Michelsen and A. Zeiss, (eds.): Symposium Volume III. International symposium on Jurassic stratigraphy, Erlangen, September 1-8, 1984.

____. 1995. Time from fossils: S.S. Buckman and Jurassic high-resolution geochronology, p. 127-150. *In* M. J. LeBas (ed.): Milestones in Geology. Geological Society of London, Memoir, 16.

CECCA, F., J.-L. DOMMERGUES, R. MOUTERDE, AND G. PALLINI. 1987. Ammonites méditerranéennes du Lotharingien de Gorgo a Cerbara (M. Nerone, Apennin des Marches, Italie). Les Cahiers de l'Université Catholique de Lyon, sér. Sciences, 1:67-82.

CORNA, M. 1987. Elements de phylogenie des Ariétitidés d'apres les données du Jura méridional. Les Cahiers de l'Université Catholique de Lyon, sér. Sciences, 1:93-104.

____, J.-L. DOMMERGUES, C. MEISTER, AND K. PAGE. 1997. Les faunes d'ammonites du Jurassique inférieur (Hettangien, Sinémurien et Pliensbachien) au nord du massif des Écrins (Oisans, Alpes occidentales françaises). Revue de Paléobiologie, 16/2:321-409.

DEAN, W. T., D. T. DONOVAN, AND M. K. HOWARTH. 1961. The liassic ammonite zones and subzones of the North-West European Province. Bulletin of the British Museum (Natural History) Geology, 4:435-505.

DOMMERGUES, J.-L. 1987. L'évolution chez les Ammonitina du Lias moyen (Carixien, Domérien basal) en Europe occidentale. Documents des Laboratoires de géologie de la Faculté des Sciences de Lyon, 98:1-297.

____. 1993. Les ammonites du Sinémurien supérieur de Bourgogne (France): Biostratigraphie et remarques paléontologiques. Revue de Paléobiologie, 12/1:67-171.

____, A. FERRETTI, AND C. MEISTER. 1994. Les faunes d'ammonites du Sinémurien de l'Apennin Central (Marches et Toscane, Italie). Bollettino della Società Paleontologica Italiana, 33/1:13-42.

____. AND C. MEISTER. 1991. Fréquence de présence des horizons et mise en évidence de fluctuations de l'environnement; exemple dans le Lias nord-ouest européen. Comptes rendus hebdomadaires des séances de l'Académie des Sciences, t. 313, sér. II:977- 982.

____. AND ____. 1992a. *Leptechioceras* (*Neomicroceras*) Donovan, 1966 (Ammonitina, Echioceratidae, Sinémurien, Europe du Nord-Ouest): taxon progénétique ou microconque. Géobios, 25/4:477-493.

____. AND ____. 1992b. Late Sinemurian and Early Carixian ammonites in Europe with cladistic analysis of sutural characters. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 185/2:211-237.

____, ____, AND F. BÖHM. 1995. New data on Austroalpine Liassic ammonites from the Adnet quarries and adjacent areas (Oberösterreich, Northern Calcareous Alps). Jahrbuch der Geologischen Bundesanstalt, 138/2:161-205.

____, ____, AND M. METTRAUX. 1990. Succession des faunes d'ammonites du Sinémurien et du Pliensbachien dans les Préalpes médianes de Suisse romande (Vaud et Fribourg): implications biochronologiques et paléobiogéographiques. Géobios, 23/3:307-341.

____, K. N. PAGE, AND C. MEISTER. 1994. A detailed correlation of Upper Sinemurian (Lower Jurassic) ammonite biohorizons between Burgundy (France) and Britain. Newsletters on Stratigraphy, 30/2:61-73.

GETTY, T. A. 1972. Revision of the Jurassic ammonite family Echioceratidae. Dissertation University of London, p.1-319.

GRADSTEIN, F. M. AND J. OGG. 1996. A Phanerozoic time scale. Episodes 19/1-2:3-5.

Figure 4. Comparison areas.

NW-EUROPEAN ZONATION			HORIZONS	ALBERTA/BRITISH COLUMBIA	ARGENTINA/CHILE	ADNET	APENNINES
? CARIXIAN							
UPPER SINEMURIAN	Raricostatum Zone	Aplanatum Subzone	SIMPLICICOSTA				
			TARDECRESCENS/ROMANICUM	PALTECHIOCERAS SP.	PALTECHIOCERAS SP.	P. INSIGNE	P. AFF. ROMANICUM/P. TARDECRESCENS
			TARDECRES./OOSTERI/RECTICOSTATUM				
			AUREOLUM				
		Macdonnelli Subzone	MEIGENI/MACDONNELLI			L. MEIGENI	PARASTEROCERAS SP. A
			MEIGENI				
			MEIGENI/CHARPENTIERI				
		Raricostatum Subzone	LICIENSE/ROTHPLETZI	? P. CF. ROTHPLETZI + ? P. AFF. BOEHMI		PARASTEROCERAS SP.	P. AFF. ROTHPLETZI P. PULCHELLUM
			FAVREI				
			BOEHMI/CF. INTERMEDIUM				P. BOEHMI
			RARICOSTATUM/CRASSICOSTATUM				
			RARICOSTATUM/RARICOSTATOIDES				
			QUENSTEDTI/RHODANICUM/AENEUM				
		Densinodulum Subzone	ECHIOCERAS (?) SP. 3				
			RADIATUM			M. GIGAS	
			GR. ARMATUM				
			BISPINIGERUM				
			LYMENSE/DENSINODULUM				
			EDMUNDI/SUBPLANICOSTA	P. (?) HARBLEDOWNENSE	P. ARCTICUM		
			DELICATUM/CF. TYPUS	P. AKLAVIKENSE/OXYNOTICERATIDAE/EODEROCERATIDAE			
	Oxynotum Zone	Oxynotum Subzone	DORIS		PALAEOECHIOCERAS SP. + OXYNOTICERAS SP.		
			BIFER				
			OXYNOTUM			O. GR. OXYNOTUM	
		Simpsoni Subzone	DRIANI				
			GAGATEUM				
			EXORTUM				
	Obtusum Zone	Denotatus Subzone	AFF. GLABER		? EPARIETITES SP.	E. GLABER	
			DENOTATUS				
			FOWLERI				
			CF. UNDARIES				
		Stellare Subzone	SAGITTARIUM				
			BLAKEI/ARNOULDI	A. AFF. MARGARITA + A. CF. VARIANS		A. AFF. SALTRIENSE	A. VARIANS + A. MARGARITA + ARNIOCERAS SP.
			STELLARE		A. CF. STELLARE		
			CF. LANDRIOTI	A. AFF. MARGARITA + EPOPHIOCERAS SP. + A. AFF. JELETZKYI + ? A. CF. SALTRIENSE + ARNIOCERAS SP.	EPOPHIOCERAS SP. +MICRODEROCERAS SP.	A. RETUSUM	
			MARGARITOIDES				
			AFF. MARGARITOIDES	A. AFF. MARGARITA + ARNIOCERAS SP.			
			GALATICERAS				
		Obtusum Subzone	OBTUSUM		ASTEROCERAS SP. sensu HILLEBRANDT 1987	ARNIOCERAS SP. ↓	ASTEROCERAS ? SP. + A. REJECTUM + A. PAUCICOSTA ↓
			SEMICOSTATOIDES				
			CF. CONFUSUM				
			AFF. CONFUSUM				

HILLEBRANDT, A. v. 1981. Faunas de amonites del Liásico Inferior y Medio (Hettangiano hasta Pliensbachiano) de América del Sur (excluyendo Argentina), p. 499-537. *In* Comité Sudamericano del Jurásico: Cuncas sedimentarias del Jurásico y Cretácico de América del Sur, 2.

MEISTER, C. 1986. Les ammonites du Carixien des Causses, France. Schweizerische Paläontologische Abhandlungen, 109:1-209.

____. 1987. Comparaison des faunes d'ammonites au Domérien (Jurassique inférieur) entre le bassin des Causses et les Cordillères Bétiques; composition faunique et éléments de corrélation. Comptes rendus hebdomadaires des séances de l'Académie des Sciences, t. 305, sér. II:425-428.

____. 1995. Essai de corrélations au Lias moyen (Sinémurien supérieur et Carixién) entre les Pontides et les principales régions adjacentes de la Téthys occidentale et de l'Europe du nord-ouest. Hantkeniana, 1 (GÉCZY Jubilee Volume), p.75-82.

____, J. BLAU, AND F. BÖHM. 1994. Ammonite biostratigraphy of the Pliensbachian stage in the Upper Austroalpine Jurassic. Eclogae Geologicae Helvetiae, 87/1:139-155.

____, AND F. BÖHM. 1993. Austroalpine liassic ammonites from the Adneth Formation (Northern Calcareous Alps). Jahrbuch der Geologischen Bundesanstalt, 1993, p.163-211.

OWEN, H. G. 1983. Atlas of continental displacement., 159 p., Cambridge University Press.

PAGE, K. 1992. The sequence of ammonite correlated horizons in the British Sinemurian (Lower Jurassic). Newsletters on Stratigraphy, 27/3:129-156.

____. 1994. On the sequence of ammonite chronostratigraphical horizons in the British Sinemurian (Lower Jurassic). Géobios, Mémoire spéciale, 17:369-379.

____. 1995. Biohorizons and zonules: intra-subzonal units in Jurassic ammonite stratigraphy. Paleontology, 38/4:801-814.

PÁLFY, J. AND K. L. SCHMIDT. 1994. Biostratigraphic and facies studies of the Telkwa Formation (Lower Jurassic), Smithers map area, British Columbia. Current Research, Geological Survey of Canada, 1994-E, p.29-38.

____, P. L. SMITH, AND H. W. TIPPER. 1994. Sinemurian (Lower Jurassic) ammonoid biostratigraphy of the Queen Charlotte Islands, western Canada. Géobios, 17:385-393.

PHELPS, M.-C. 1985. A refined ammonite biostratigraphie for the Middle and Upper Carixian (Ibex and Davoei Zones, Lower Jurassic) in North-West Europe and stratigraphical details of the Carixian-Domerian boundary. Géobios, 18/3:321-362.

QUINZIO-SINN, L. A. 1987. Stratigraphische Untersuchungen im Unterjura des Südteils der Provinz Antofagasta in Nord-Chile. Berliner geowissenschaftliche Abhandlungen, Reihe A, 87:1-100.

WESTERMANN, G. E. G. (ed.). 1992. The Jurassic of the circum-Pacific. Cambridge University Press, x + 676 p.

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The Proposed GSSP for the Base of the Sinemurian Stage Near East Quantoxhead/West Somerset (SW England) - the Ammonite Sequence

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Abstract: The ammonite succession given in this paper reveals the sequence across the Hettangian/Sinemurian Stage boundary at the Somerset coast to be the most complete in the Northwest European Province. In the Conybeari Subzone, 13m thick, there can now be distinguished eight well defined ammonite faunal horizons; further potential horizons need additional study. Only a few of these horizons are also known in other parts of the Province. The Stage boundary is situated within an homogeneous bituminous shale. There is a correlation potential with other faunal provinces by means of Hettangian and Sinemurian Arietitinae.

Three new species are introduced: *Vermiceras quantoxense*, *Vermiceras palmeri* and *Vermiceras elegans*.

Introduction

The present ammonite study was carried out where the sequence across the Hettangian/Sinemurian Stage boundary is best exposed and least fractured along the west Somerset coast: close to the beach access north of East Quantoxhead ("Limekiln Steps" in Palmer, 1972); locality map see Page et al., this volume, Fig. 1. The boundary itself is exposed immediately beside these steps in the lower part of the cliff and in the adjoining foreshore and is thus optimally accessible. The sequence passes from the cliff into the ledges on the foreshore so that the identification of individual beds on the foreshore is easy. The facies is typical "Blue Lias", a sequence of marls and bituminous shales and intercalated marly limestone beds. Detailed sections on this part of the coast were first provided by Palmer (1972, p.10, 13-17) and later by Whittaker and Green (1983, p.70-71, 73). The bed numbers of Whittaker and Green are used here, with Palmer's notation in brackets.

Ivimey-Cook and Donovan (1983, p.128-129) gave a brief account of the ammonite fauna which did not permit a subdivision of the Angulata Zone, and details of the ammonite succession in the Conybeari Subzone remained uncertain. After a restudy of this section by the authors in 1994, a more detailed sequence of faunas and biohorizons was established for the Conybeari Subzone (Page, 1995), whereas detailed information on the Hettangian remained lacking. The crushed preservation of the ammonites made determinations difficult, and the present study has been made possible only by comparison with well-preserved material from other parts of the Northwest European Province, especially Germany.

We have studied only the upper part of the Angulata Zone (upper Complanata and Depressa Subzones) through a thickness of about 17m at Limekiln Steps, as faulting and gravel deposits confuse the lower part of the exposure (Extranodosa and Lower Complanata Subzones). The thickness of the complete Angulata Zone can be estimated at around 35-40m (Whittaker and Green, 1983). The Conybeari Subzone could be sampled throughout its full thickness of 13m, up to the first appearance of *Coroniceras deffneri* (Oppel) marking the base of the Rotiforme Subzone.

Material

We have sampled the sequence intermittently over several years (1994 to 1998), mainly around the locality of Limekiln Steps, with additional material from exposures up to 1.5km westwards along the foreshore to Blue Ben. The ammonites were collected mainly by Page. Most of the material is deposited in the Sedgwick Museum in Cambridge; additional specimens are in the Bristol City Museum, and more specimens will be transferred to both collections in due course. The ammonites are generally flattened, especially in the shales and marls. However, characters can be sufficiently recognized in most cases to identify species. In the marly limestones body chambers are occasionally less crushed. Only in scarce, lenticular crinoidal accumulations, which formed a self-supporting texture, are some ammonites preserved uncrushed. In such specimens the phragmocones are filled with coarse calcite. The preparation of specimens from the limestones is often difficult and time-consuming and not all material collected has, therefore, been fully extracted.

Upper Hettangian (Upper Angulata Zone)

Subdivision and correlation of the upper part of the Angulata Zone is of special interest at East Quantoxhead as in the NW European Province uncertainties still exist in this interval. The Angulata Zone had formerly been divided into two subzones, the lower named after *Schlotheimia extranodosa* (Wähner), the upper after *Schlotheimia complanata* Koenen (Donovan in Dean et al., 1961). In southern Germany, however, the fauna of the Complanata Subzone is overlain by a fauna characterized by *Schlotheimia depressa* (Quenstedt), lacking any elements of the Complanata Subzone. A subzone of *Schlotheimia depressa* was therefore proposed by Bloos (1979a, p.37). The relationship between this new subzone and the fauna of the upper Angulata Zone in Britain, characterized by *Schlotheimia pseudomoreana* Spath, remained unclear. Detailed sampling of the East Quantoxhead section has now yielded relevant new data.

The last specimens of the Liasicus Zone are recorded from Bed 76 (Palmer Bed C8). No ammonites are known from Beds 77-79 (C9-C27), about 4.5m thick. The first specimens of *Schlotheimia* (*S. amblygonia* Lange) appear in Bed 80 (C28); they indicate the Extranodosa Subzone (Ivimey-Cook and Donovan, 1983, p.128). Little is known on the fauna of the overlying Beds 81-94 (C29-C50), an interval of about 11m.

Complanata Subzone (Ammonites see Pl. 1.4-8)

Beds 95-129 (C52-C86) contain a typical fauna of the Complanata Subzone including most of the species described from NW Germany by Lange (1951): *Schlotheimia complanata* and closely related forms (Pl. 1.7, 8), *S. oxygonia* (Pl. 1.5), *S. tenuis*, *S. macilenta*, *S. polyeides*, and the *S. angulosa* group (Pl. 1.6). More intensive collecting would most probably yield still more of the species described by Lange, 1951.

In Bed 133 (C90), a specimen of *Schlotheimia* ex gr. *striatissima* (Quenstedt) has been recovered (Pl. 1.4). This rather variable species is characterized by dense, sigmoidal ribbing and rather narrow umbilicus. It occurs in the "Oolithenbank" of SW Germany and in eastern France (Guérin-Franiatte, 1990, pl. 20, Fig. 2).

The Complanata Subzone at Limekiln Steps is at least 16m thick; the true thickness may be greater because the lower boundary is not yet precisely located. The upper portion of the subzone, here indicated by *S.* ex gr. *striatissima*, is extremely thin compared with SW Germany.

Depressa Subzone (Ammonites see Pl. 1.1-3)

Two specimens of *Schlotheimia* cf. *princeps* (S. Buckman) were extracted from Bed 134 (C91) (Pl. 1.3). The specimens differ from the type specimen in having a somewhat wider umbilicus. This species is typical of the horizon with *S. depressa* in southwest Germany and differs from *S. depressa* only in having a narrow umbilicus (see Lange, 1925, pl. 18, Fig. 1). On the upper surface of Bed 134, large specimens of *Schlotheimia*, badly worn by the sea and occurring in isolated groups, are usually not extractable. It is not yet certain, therefore, if *S. depressa* is present in Somerset.

Immediately above Bed 134, *Schlotheimia pseudomoreana* Spath appears and persists to the top of the Hettangian in Bed 145 (C100, 60cm above limestone Bed 144) (Pl. 1.2). *S. pseudomoreana* is very similar to *S. princeps*, but generally has weaker and denser ribbing with less well developed bifurcation (transitional forms do occur, however). The cross section of the whorls is essentially the same in both species and remains ovoid throughout, which is the main difference compared to the large forms of the Complanata Subzone (*Schlotheimia similis* Spath, *S. stenorhyncha* Lange) in which the whorl cross section becomes trigonal during ontogeny and, moreover, the ribbing is considerably denser on the inner whorls.

A smaller form of *Schlotheimia* occurs together with *S. pseudomoreana* (Pl. 1.1). It has a relatively narrow umbilicus and wide-spaced, sharp ribs and probably represents a new species. Similar specimens occur at the top of the Angulata Zone in SW Germany and may, therefore, indicate the base of the *pseudomoreana* horizon in that region.

Significantly, species of the Complanata Subzone do not occur together with *S. pseudomoreana*. This fact, combined with the close relationship between the *S. depressa/princeps* group and *S. pseudomoreana*, suggests that both horizons are part of one subzone, the Depressa Subzone. The thickness of that subzone at East Quantoxhead is around 4.5m. In contrast to the diversity in the Complanata Subzone, the ammonite fauna of the *pseudomoreana* horizon is rather poor.

The Hettangian/Sinemurian Boundary

The boundary between the Hettangian and the Sinemurian is situated between limestone beds 144 (C99) and 147 (C 101) in an argillaceous unit corresponding to Palmer's Bed C100. Whittaker and Green (1983), however, subdivided C100 into two beds, 145 and 146, the lower of which is a bituminous shale in its upper part. Re-measurement yielded the following subdivision:

C101/Bed 147 (70 cm): Divided limestone bed; *Metophioceras* sp. indet B, *Charmasseiceras* sp.
C100/Bed 146 (52 cm): Marl, non-bituminous, transitional to the bed below; *Vermiceras* spp. as below
C100/Bed 145, upper part (58 cm): Bituminous shale; in detail:
  23 cm: Bituminous shale with numerous *Vermiceras quantoxense* n.sp., *V. palmeri* n.sp., rare *Metophioceras* sp. indet. A
  20 cm: No ammonites found.
  15 cm: Bituminous shale with numerous *Schlotheimia* sp., rare fragments of *S. pseudomoreana*
C100 /Bed 145, lower part (45 cm): Marl with *Schlotheimia* sp. and *S. pseudomoreana* as above
C99/Beds 142-144 (43 cm): Limestone bed, partly divided by a thin marl, no ammonites

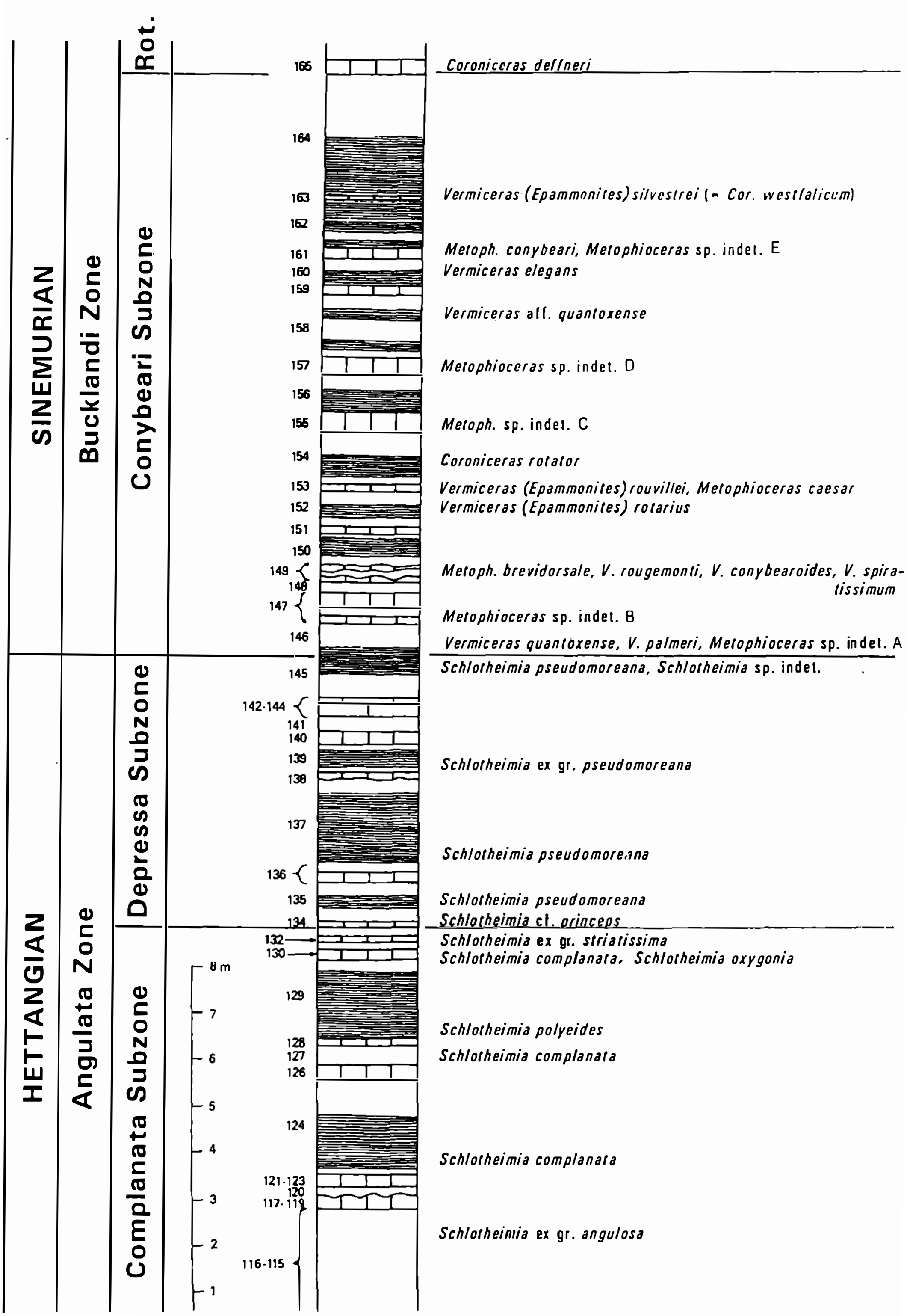
The Hettangian/Sinemurian boundary therefore lies in the bituminous shale at about the middle of unit C100. According to Ivimey-Cook and Donovan (1983, p.129) the first specimens of *Vermiceras* appear in bed 146, above the lithological boundary between bituminous shale and marl. The new records indicate that the genus occurs already in the upper 23cm of the bituminous shale. The highest specimens of *Schlotheimia* (*S. pseudomoreana* and *S.* sp.) have been recovered in the lower 15cm of this bituminous shale. A 20cm interval of the shale horizon has not yet yielded any ammonites; it is here formally attributed to the Angulata Zone. The highest *Schlotheimia* fauna was assigned by Page (1995) to a "*Schlotheimia.* sp. 2 Biohorizon" and the lowest *Vermiceras* to an "aff. *rougemonti* Biohorizon" (numbered '1').

According to Ivimey-Cook and Donovan (1983, p. 129), *Schlotheimia* persists into levels with *Vermiceras*. We could not observe such an overlap; the Hettangian fauna is replaced here as abruptly as elsewhere throughout the NW European Province. The presence of the Stage boundary within a lithologically homogeneous unit meets an essential requirement of the Guidelines of the International Commission on Stratigraphy concerning candidate GSSPs (Remane, 1996).

Lower Sinemurian (Lower Bucklandi Zone)

Conybeari Subzone (Ammonites see Pl. 2)

Lower part (up to Bed 149): The base of the Sinemurian in the NW European Province normally is characterized by an assemblage of *Vermiceras spiratissimum* (Quenstedt), *V. rougemonti* (Reynès), *V. conybearoides* (Reynès), *V. longidomus* (Quenstedt), *Vermiceras (Epammonites) latisulcatus* (Quenstedt), *Metophioceras brevidorsale* (Quenstedt) and *Metophioceras* sp. indet. (misinterpreted as *M. longidomus* thus far). This fauna is often recorded also in Britain as the lowermost of the Sinemurian (e.g., from around Bristol, as indicated by a collection of well-preserved specimens in the Bristol City Museum; see also Donovan, 1956). The horizon can also be identified in the present section, in Bed 149 (C103), which typically is represented by two horizons of lenticular limestone separated by a thin marl layer. The fauna occurs in the upper limestone band and in the intercalated marl below and includes most of the species mentioned above (Fig. 1), *M. brevidorsale* (Pl. 2.1) being the most frequent (="*Metophioceras* sp. 2 Biohorizon" (2b) of Page, 1995).


SINEMURIAN
Bucklandi Zone
Rot.
Conybeari Subzone
HETTANGIAN
Angulata Zone
Depressa Subzone
Complanata Subzone
8 m
7
6
5
4
3
2
1
165
164
163
162
161
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154
153
152
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148
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146
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142-144
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139
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128
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126
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121-123
120
117-119
116-115
Coroniceras deffneri
Vermiceras (Epammonites) silvestrei (= Cor. westfalicum)
Metoph. conybeari, Metophioceras sp. indet. E
Vermiceras elegans
Vermiceras aff. quantoxense
Metophioceras sp. indet. D
Metoph. sp. indet. C
Coroniceras rotator
Vermiceras (Epammonites) rouvillei, Metophioceras caesar
Vermiceras (Epammonites) rotarius
Metoph. brevidorsale, V. rougemonti, V. conybearoides, V. spiratissimum
Metophioceras sp. indet. B
Vermiceras quantoxense, V. palmeri, Metophioceras sp. indet. A
Schlotheimia pseudomoreana, Schlotheimia sp. indet.
Schlotheimia ex gr. pseudomoreana
Schlotheimia pseudomoreana
Schlotheimia pseudomoreana
Schlotheimia cf. princeps
Schlotheimia ex gr. striatissima
Schlotheimia complanata, Schlotheimia oxygonia
Schlotheimia polyeides
Schlotheimia complanata
Schlotheimia complanata
Schlotheimia ex gr. angulosa

Between the level of earliest *Vermiceras* in Bed 146/145 (*quantoxense* Biohorizon) and the above-mentioned widespread level in Bed 149 (*conybearoides* Biohorizon), we found in Bed 147 (C101) rare, large *Metophioceras* with dense ribbing. Specimens are virtually inextractable. Their ribbing is stronger, and the cross section broader than in *M. brevidorsale*. This form is here listed as *Metophioceras* sp. indet. B. Associated small species (*Vermiceras*) have not yet been found. Page (1995) provisionally assigned the fauna of Bed 147 to a "*Metophioceras* sp. 1 Biohorizon" (2a).

Middle and upper part (Beds 150-164): In the NW European Province so far only two additional ammonite horizons were known above the *conybearoides* Biohorizon, a lower one with *Vermiceras (Epammonites) rotarius* (Spath) and an upper one with *Vermiceras (Epammonites) silvestrei* (Reynès) (= *Coroniceras westfalicum* Lange) (Bloos, 1988a; see also Lange, 1925). Both are now known to be present also in west Somerset. Three additional biohorizons were recognized by Page (1995) on the coast of southwest England (Dorset and west Somerset) with, respectively, *Vermiceras (Epammonites) rouvillei* (Reynès), *Coroniceras rotator* (Reynès), and *Metophioceras conybeari* (Sowerby).

We can now recognize further horizons in the East Quantoxhead area. Bed 158 (C112) yielded a form broadly similar to *V. quantoxense*. Only one specimen is presently available, and adequate comparisons with other species are not yet possible.

An additional horizon, characterized by a new form of *Vermiceras* (*Vermiceras elegans* n. sp.), was found in Bed 160 (C114) (see below).

Biohorizons in the Conybeari subzone: The following list gives an overview of the biohorizons actually known in the Conybeari Subzone; additional, unnamed horizons are mentioned which may become biohorizons. The ammonites of the biohorizons are given in Fig. 1.

Rotiforme Subzone:
Bed 165 (C121): *deffneri* Biohorizon (Page, 1992)
Conybeari Subzone:
Bed 163 (C118): *silvestrei* Biohorizon (Bloos 1988a)
(= "Zone des *Arietites westfalicus*" Lange, 1921)
Bed 161 (C115): *conybeari* Biohorizon (Page, 1992)
Bed 160 (C114): *elegans* Biohorizon nov.
Bed 158 (C112): *Vermiceras* aff. *quantoxense*
Bed 157 (C111): *Metophioceras* sp. indet. D
Bed 155 (C109): *Vermiceras* or *Metophioceras* sp.
Bed 154 (C108): *rotator* Biohorizon (Page, 1992)
Bed 153 (C107): *rouvillei* Biohorizon (Page, 1995)
Bed 152 (C106): *rotarius* Biohorizon (Bloos, 1988a)
Bed 151 (C105): *Vermiceras* or *Metophioceras* sp.
Bed 149 (C103): *conybearoides* Biohorizon (Bloos, 1988a)
(= "*Metophioceras* sp. 2 Biohorizon" Page, 1995 and "*longidomus* Biohorizon" Page, 1992)
Bed 147 (C101): "*Metophioceras* sp. Biohorizon" (Page, 1995)
Bed 146/145 (C100): *quantoxense* Biohorizon nov.
(= "cf./aff. *rougemonti* Biohorizon": Page, 1992; 1995)

New Ammonites in the Conybeari Subzone

The basal ammonite assemblage of the Sinemurian near East Quantoxhead consists of two new species of *Vermiceras* and an incompletely known *Metophioceras*.

Vermiceras quantoxense New Species

Holotype.- Original of Pl. 2.4 (Sedgwick Museum Cambridge SMC X29332)

Type locality.- Limekiln Steps, north of East Quantoxhead, near the coastal cliff "Quantock's Head", west Somerset.

Figure 1. Section of the proposed GSSP at "Limekiln Steps" north of East Quantoxhead. After Whittaker and Green 1983:73. The thicknesses are adjusted to the recent measurements; the argillaceous beds are now differentiated into bituminous shale (dark) and non-bituminous marl (white); white in Bed 115/116 and in the upper part of Bed 164 means "undifferentiated" because of unsufficient exposure.

Type horizon. - Upper 75cm of Bed C100 of Palmer, 1972.

Derivation of name. - From the village of East Quantoxhead, at the seaward end of the Quantock Hills, within which parish the type locality lies.

Material. - 12 topotypes, all from the type locality and horizon (Sedgwick Museum Cambridge collections).

Diagnosis. - *Vermiceras* with a characteristic type of ribbing, dense on the inner whorls and becoming widely spaced on the outer, and therefore resembling some *Paltechioceras*. Generally the ribs are slightly curved throughout ontogeny. Size of mature specimens, whose ribbing becomes weak and irregular, around 8cm.

Remarks. - Because of the crushed state, the species can be defined and identified only by its ribbing, the density of which is plotted in Fig. 3. Other characters, such as sculpture of the venter, cross section of the whorls, and relatively simple suture line, vary little within the genus. Therefore, it is likely that these characters will be similar in the present species. The new form was previously identified as *Vermiceras solaroides* (da Costa) as figured in Fucini (1902, pl. 13, Figs. 1, 2), and as *Vermiceras rougemonti* (Reynès) by Ivimey-Cook & Donovan (1983, p.129). Both of these species are rather different from the new species, however. *V. solaroides* most probably is an echioceratid and *V. rougemonti* shows a continuous, rapid increase of rib numbers similar to *V. spiratissimum* (Fig. 2). *V. rougemonti* and *V. spiratissimum* have, moreover, essentially straight ribs.

Vermiceras palmeri New Species

Holotype. - Original of Pl. 2.5 (Sedgwick Museum Cambridge SMC X29333)

Type locality. - Limekiln Steps, north of East Quantoxhead, west Somerset.

Type horizon. - Upper 75cm of Palmer's Bed C100.

Derivation of name. - after C. P. Palmer who gave the first detailed account of the Lower Lias of the coast of west Somerset and recognized the significance of that sequence.

Material. - 8 topotypes, all from the type locality and horizon (Sedgwick Museum Cambridge collections).

Diagnosis. - *Vermiceras* with continuously, but slowly, increasing rib numbers on successive whorls. The ribs are curved on the flanks throughout. The mature size is around 8cm.

Remarks. - In comparison with the co-occurring *V. quantoxense* n. sp., the present species has less dense ribbing on the inner whorls and denser ribbing on the outer whorls (see Fig. 5). At around half the adult diameter (about 30mm umbilical width), there is a complete overlap of rib numbers between both species. In *Vermiceras spiratissimum* and *V. rougemonti*, rib numbers increase more rapidly, especially towards the end of the adult body chamber (Fig. 2), so the present species is therefore distinct. *V. spiratissimum* and *V. rougemonti*, moreover, have essentially straight ribs whereas in *V. palmeri* they are always curved.

For comparison with the two new species of *Vermiceras* from the base of the Sinemurian, we add information on two similar species which occur higher in the succession, *V. spiratissimum* (in Fig. 2) and *V. elegans* n. sp. (in Fig. 4 and below).

Vermiceras elegans New Species

Holotype. - Original of Pl 2.7 (Sedgwick Museum Cambridge SMC X29335)

Type locality. - Foreshore exposure west of Limekiln Steps, between Quantock's Head and Blue Ben.

Type horizon. - Bed 160 (C114), upper part.

Derivation of name. - The name reflects the dense, delicate ribbing.

Material. - 6 topotypes, all from the type locality and horizon, including SMC X29336 (plate 2, figure 8) (Sedgwick Museum Cambridge collections).

Diagnosis. - *Vermiceras* with extremely dense ribbing throughout ontogeny.

Remarks. - The development of ribbing (Fig. 4) resembles that of *V. palmeri* (Fig. 5), but the rib numbers are far higher. In the density of ribbing the species resembles certain echioceratids from the Upper Sinemurian.

Remark on *Metophioceras* sp. indet. A

Scarce fragments (Pl. 2.3) co-occurring with *V. quantoxense* and *V. palmeri* indicate specimens of about 8-10cm diameter. The whorl height exceeds that of *Vermiceras* and, therefore, indicates *Metophioceras*. The dense, weak ribbing resembles that of *M. brevidorsale*, but the size is much smaller. Probably these specimens are not adult.

General Note on the Difference Between *Vermiceras* and *Metophioceras*

Vermiceras: Small to medium-sized, evolute and slowly growing ("worm-like") forms; cross section of inner-most whorls circular (outer whorls variable and less diagnostic); keel appears early in ontogeny, before the ribs; suture line simple (saddles not deeply indented, indentations generally without further subdivision).

Fig. 2

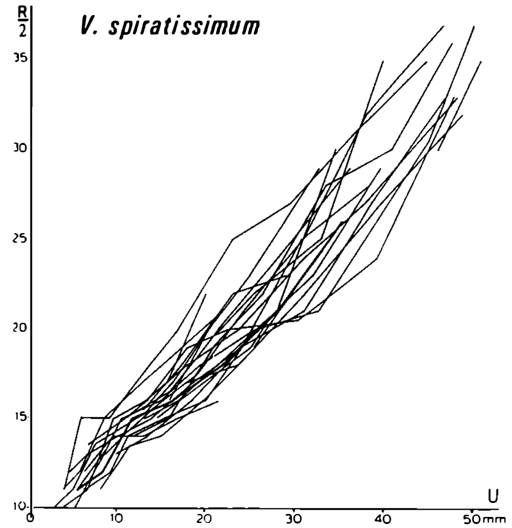


Fig. 3

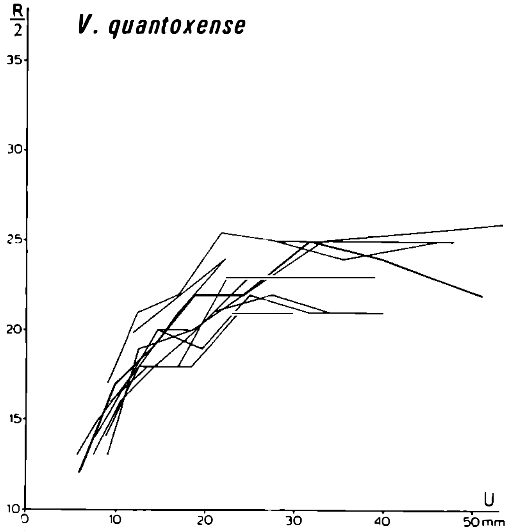


Fig. 4

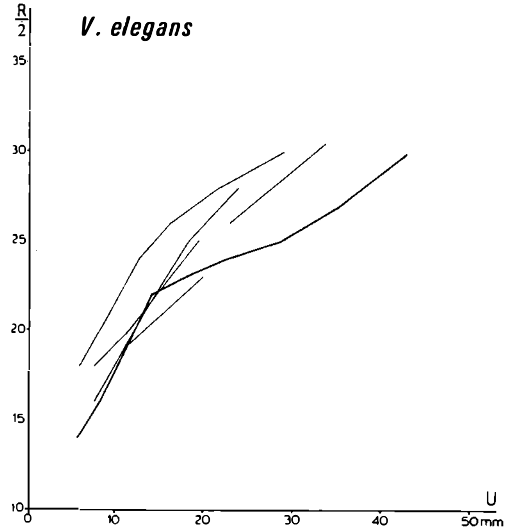
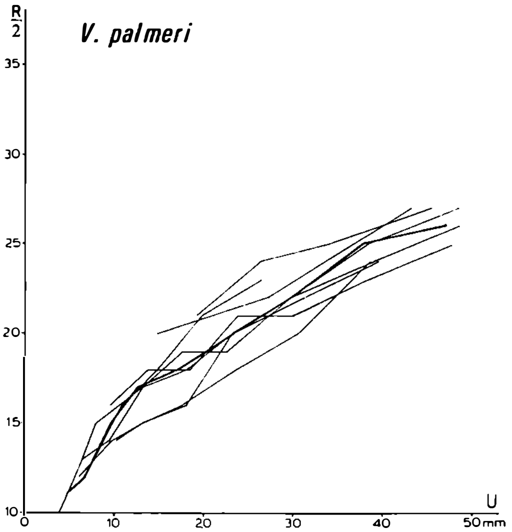


Fig. 5



Figures 2–5. Ontogeny of ribbing in four species of *Vermiceras*. Rib numbers on the half whorl (R/2) in relation to the corresponding width of umbilicus (U) (instead of diameter which cannot be measured on the inner whorls).

2. *V. spiratissimum* (SW Germany) for comparison; the development in *V. rougemonti* is essentially the same. Density of ribbing increases during ontogeny.
3. *V. quantoxense*; rib numbers increase rapidly in early ontogeny but become stagnant or even reduced in late ontogeny.
4. *V. elegans*; highest rib density of all known species of *Vermiceras*. Unlike in *V. spiratissimum*, the increase of rib numbers becomes retarded in later ontogeny.
5. *V. palmeri*; trends similar to *V. elegans*, but the rib numbers are considerably lower.

***Metophioceras*:** Medium-sized to large forms, whorls more rapidly growing in height and width than in *Vermiceras*; cross section of innermost whorls subrectangular, broader than high; ribs appear before the keel; suture line elaborated (saddles deeply indented, indentations subdivided, already early in ontogeny). In all characters *Metophioceras* resembles *Coroniceras*, with only one exception: nodes are lacking. (See also Bloos, 1994.)

Correlation Potential

Correlation of the Somerset sequence with other parts of the NW European Province is not difficult because of the completeness of the ammonite sequence in Somerset.

Detailed correlation with other faunal provinces is still difficult as ammonite successions elsewhere are not so well known. The lower Conybeari Subzone is demonstrably lacking in the northeastern Alps (Bloos, 1988a), even in the most complete sections (the stratigraphic gap between Triassic and Jurassic normally reaches considerably higher, sometimes up to the Pliensbachian).

In the Late Hettangian, provincialism is pronounced over much of the world, whereas in the Rotiforme Subzone ammonite faunas are rather similar. The latter may already be true in the Conybeari Subzone. In the NW European Province, Angulata Zone faunas are uniquely developed, being almost exclusively composed of the genus *Schlotheimia,* and nowhere else is this genus so diverse. In other faunal provinces in the Late Hettangian additional ammonite genera are frequent, most of which are totally lacking in the Northwest European Province. *Schlotheimia* is often so rare in other faunal provinces that it is not possible to distinguish Hettangian from Sinemurian.

Other schlotheimiids besides *Schlotheimia*, known as "*Charmasseiceras*" and "*Sulciferites*" (comprised under *Angulaticeras* in Bloos, 1979b), are characteristic of the Sinemurian but already occur in the Late Hettangian. The characteristic species *"Schlotheimia" marmorea* (Wähner) occurs almost world-wide. It is known in the northeastern Alps only in Wähner's Marmorea Zone. In the NW European Province it has been found exclusively in the Depressa Subzone in SW Germany (Bloos, 1988a; 1988b), eastern France (Guérin-Franiatte, 1990) and, recently by one of us (K.N.P.) together with *Schlotheimia pseudomoreana* in Dorset.

Virtually the only ammonite group to occur world-wide in both Stages is the Arietitinae. As shown by Bloos (1994), the Arietitinae in Wähner's Marmorea Zone of the northeastern Alps are strikingly different from those of the Lower Sinemurian. They are represented by the genera *Paracaloceras, Vermiceras* (subgenus *Gyrophioceras*) and *Schreinbachites. Vermiceras sensu stricto* and *Metophioceras* are lacking.

In the Late Hettangian, Complanata Subzone, *Schreinbachites* is relatively frequent in SW Germany (Bloos, 1994) and also occurs in eastern France (Guérin-Franiatte, 1990: "*Gyrophioceras*", partim) and Luxembourg (Maubeuge, 1987: "*Gyrophioceras*" *praecursor*). Specimens are also known from England near Bristol (Donovan, 1952, pl. 29, Fig. 2) and from Dorset, together with *Schlotheimia pseudomoreana* (Page, 1995). The latter occurrence, near the top of the Hettangian, indicates a major change also in the Arietitinae at the Hettangian/Sinemurian boundary.

The earliest ammonite species of the Sinemurian in the northeastern Alps is *Vermiceras (Epammonites) rouvillei,* indicating the higher part of the Conybeari Subzone. It occurs in a condensed assemblage associated with typical forms of the Rotiforme Subzone, including *Coroniceras rotiforme* and *C. deffneri.*

For global correlation, the Arietitinae will be a group of special significance. The basal Sinemurian may be recognized using species of *Vermiceras* and *Metophioceras.* The latter is a genus which is present, though not yet adequately described, in North and South America (Taylor, 1990; Palfy et al., 1994; Hillebrandt, 1990; 1994; this volume). If it proves possible to identify the Hettangian/Sinemurian Stage boundary in these regions, it will be possible to calibrate other ammonite groups against a standard, thus facilitating correlation with further regions.

Acknowledgments

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References

BLOOS, G. 1979a. Über den Jura am Grossen Hassberg (Unterfranken, N-Bayern) mit Bemerkungen zum Rät. Stuttgarter Beiträge zur Naturkunde, Series B, 44:1–53.

____. 1979b. Über *Ammonites ventricosus* Sow. und ähnliche Schlotheimiiden im tieferen Sinemurium (Unterer Lias). Paläontologische Zeitschrift, 53:142-162.

____. 1988a. On the stage boundary Hettangian/Sinemurian in north-west Europe and in the north-eastern Alps. 2nd International Symposium on Jurassic Stratigraphy, 1:71–83.

____. 1988b. *Ammonites marmoreus* Oppel (Schlotheimiidae) im unteren Lias (*angulata*-Zone, *depressa*-Subzone) von Württemberg (Südwestdeutschland). Stuttgarter Beiträge zur Naturkunde, B, 141:1-47.

____. 1994. Frühe Arietitidae (Ammonoidea) aus dem Hettangium (Angulata-Zone, Unt. Lias) von Württemberg (SW-Deutschland). Stuttgarter Beiträge zur Naturkunde, Series B, 219:1–67.

DEAN, W. T., D. T. DONOVAN AND M. K. HOWARTH. 1961. The Jurassic ammonite zones and subzones of the North West European Province. Bulletin of the British Museum of Natural History, 4, p. 435–505.

DONOVAN, D. T. 1952. The ammonites of the Blue Lias of the Bristol District. Part 1. Psiloceratidae and Schlotheimiidae. Part 2. Arietitidae. The Annals and Magazine of Natural History, Series 12, 5:629–655, 717–752.

____. 1956. The zonal stratigraphy of the Blue Lias around Keynsham, Somerset. Proceedings of the Geologist's Association, 66 (1955):182–212.

FUCINI, A. 1902. Cefalopodi Liassici del Monte di Cetona. Part 2. Palaeontographia Italica, 8:131–217.

GUÉRIN-FRANIATTE, S. 1990. Ammonites du Lias inférieur de France. Part 2. Psiloceratidae, Schlotheimiidae et premiers Arietitidae. Mémoire du Centre d'Etudes et de Recherches géologiques et hydrologiques, 29:1–206.

HILLEBRANDT, A. v. 1990. Der Untere Jura im Gebiet des Rio Atuel (Provinz Mendoza, Argentinien). Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 181:143–157.

____. 1994. The Triassic/Jurassic boundary and Hettangian biostratigraphy in the area of the Utcubamba Valley (Northern Peru). Geobios, Mémoire spécial, 17:297–307.

____. [this vol.]. Ammonite biostratigraphy of the Hettangian/Sinemurian boundary in South America.

IVIMEY-COOK, H. C. AND D. T. DONOVAN. 1983. Appendix 3. The fauna of the Lower Jurassic, p. 126–130. *In* A. Whittaker and G. W. Green (eds.), Geology of the country around Weston-super-Mare, Memoir for 1:50000 geological sheet 279, New Series, with parts of sheets 263 and 295.

LANGE, W. 1925. Zur Paläogeographie und Ammonitenfauna des Lias alpha nebst einer Revision der Nürtinger Psilonotenfauna. Zeitschrift der Deutschen Geologischen Gesellschaft, 77:439–528.

____. 1951. Die Schlotheimiinae aus dem Lias alpha Norddeutschlands. Palaeontographica, A, 100:1–128.

MAUBEUGE, P. L. 1987. Nouvelles données stratigraphiques sur le Grès de Luxembourg au N.O. du méridien de Luxembourg. Bulletin de l'Académie et de la Société Lorraines des Sciences, 26, p.5–19.

PAGE, K. N. 1992. The sequence of ammonite-correlated horizons in the British Sinemurian (Lower Jurassic). Newsletters on Stratigraphy, 27:129–156.

____. 1995. East Quantoxhead, Somerset, England – a potential Global Stratotype Section and Point (GSSP) for the base of the Sinemurian Stage (Lower Jurassic). Proceedings of the Ussher Society, 9:446–450.

____, BLOOS, G., J. L. BESSA, M. FITZPATRICK, S. HESSELBO, M. HYLTON, A. MORRIS, AND D. E. RANDALL. [this vol.]. East Quantoxhead, Somerset: a candidate Global Stratotype Section and Point for the base of the Sinemurian Stage (Lower Jurassic).

PÁLFY, J., P. L. SMITH AND H. W. TIPPER. 1994. Sinemurian (Lower Jurassic) ammonoid biostratigraphy of the Queen Charlotte Islands, Western Canada. Geobios, Mémoir spécial, 17:385–393.

PALMER, C. P. 1972. The Lower Lias (Lower Jurassic) between Watchet and Lilstock in North Somerset (United Kingdom). Newsletters on Stratigraphy, 2:1–30.

REMANE, J. 1996. The revised Guidelines of ICS and their bearing on Jurassic chronostratigraphy. GeoResearch Forum, 1-2:19-22.

SPATH, L. F. 1924. The Ammonites of the Blue Lias. Proceedings of the Geologists' Association, London, 35:186–211.

TAYLOR, D. G. 1990. Two species of Paracaloceras from the Canadense Zone (Hettangian-Sinemurian) in Nevada (USA). Bulletin Société vaudoise Sciences naturelles, 80.2:211-217.

WHITTAKER, A. AND G. W. GREEN. 1983. Geology of the country around Weston-super-Mare. Memoir for 1:50000 geological sheet 279, New Series, with parts of sheets 263 and 295. Geological Survey of Britain; England and Wales.

Plate 1. Characteristic species of *Schlotheimia* in the uppermost Hettangian at East Quantoxhead. Depressa Subzone. **1.** *Schlotheimia* sp. indet., co-occurring with *S. pseudomoreana* Spath; base of Bed 145 (top of Hettangian) (SMC X29321); **2.** *Schlotheimia pseudomoreana* Spath, fragment of body chamber; base of Bed 145 (top of Hettangian) (SMC X29322); **3.** *Schlotheimia* ex gr. *princeps* S. Buckman, probably indicating the horizon of *S. depressa*; surface of Bed 134 (SMC X29323). Complanata Subzone (upper part). **4.** *Schlotheimia* ex gr. *striatissima* (Quenstedt); Bed 133 (top of the Complanata Subzone) (SMC X29324); **5.** *Schlotheimia oxygonia* Lange; Bed 99 (SMC X29325); **6.** *Schlotheimia angulosa/phobetica* group; Bed 99 (SMC X29326); **7.** *Schlotheimia complanata* Koenen, Bed 95 (SMC 29327); **8.** *Schlotheimia complanata polita* Lange; Bed 95 (SMC 29328). Scale: 5 cm.


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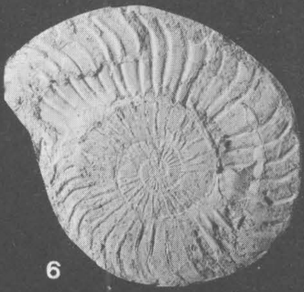
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Plate 1

Plate 2. Characteristic species of the lowermost Sinemurian at East Quantoxhead. Conybeari Subzone. **1.** *Metophioceras brevidorsale* (Quenstedt); Bed 149 (SMC X29329). Scale: 10 cm; **2.** *Vermiceras rougemonti* (Reynès), Bed 149 (SMC X29330); **3.** *Metophioceras* sp. indet A; Bed 145, top of bituminous shale (basal assemblage of the Sinemurian) (SMC X29331). **4.** *Vermiceras quantoxense* n. sp., holotype; Bed 145, top of bituminous shale (basal assemblage of the Sinemurian) (SMC X29332); **5.** *Vermiceras palmeri* n. sp., holotype; Bed 145, top of bituminous shale (basal assemblage of the Sinemurian) (SMC X29333); **6.** *Vermiceras* sp., fragment of a venter (basal assemblage of the Sinemurian) (SMC X29334); **7.** *Vermiceras elegans* n. sp., holotype; Bed 160 (upper Conybeari Subzone) (SMC X29335); **8.** *Vermiceras elegans* n. sp.; Bed 160 (upper Conybeari Subzone) (SMC X29336). Scale of **2-8** same as Plate 1.

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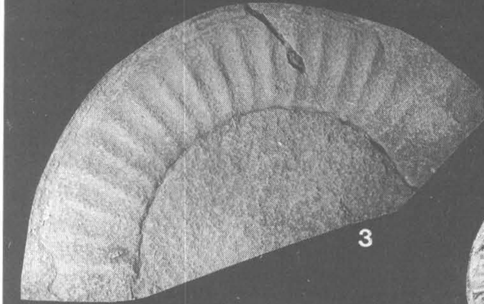

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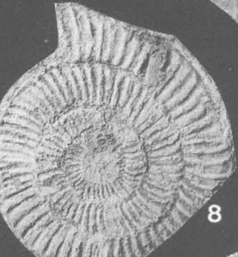

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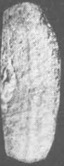

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Plate 2

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The Basal Jurassic Ammonite Succession in the North-West European Province - Review and New Results

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Keywords: Stratigraphy, Lias, Hettangian, Type Section, Faunal Provinces

Abstract: Re-examination of existing collections as well as new, bed by bed sampling revealed the existence of a succession of smooth-shelled psiloceratids in the Planorbis Subzone of the NW European Province. Six different ammonite faunal horizons (biohorizons *sensu* Page) are distinguished. These are best represented in the Wilkesley Borehole (Cheshire, NW England). The recognized sequence is restricted to the NW European Province, indicating a marked provincialism at the beginning of the Jurassic and inhibiting world-wide correlation.

Introduction

Since Sowerby (1822) introduced his *Ammonites planorbis*, further smooth-shelled psiloceratids have been described and named from the earliest Hettangian: *Ammonites sampsoni* Portlock, 1843; *Ammonites psilonotus* Quenstedt, 1849; *Psiloceras erugatum* S. Buckman 1921; *Psiloceras aequabile* S. Buckman, 1923; *Psilophyllites antecedens* Lange, 1931 (1941 genus changed into *Neophyllites*), and *Neophyllites imitans* and further species of the same genus in Lange, 1941b. These forms have generally been found in isolated occurrences, not in stratigraphic sequence and rarely in association with other species. Consequently, their systematic and stratigraphic relationships remained unknown. Most authors believed them to appear at the base of the Jurassic, and there was always a tendency to regard them as varieties of one species: *Psiloceras planorbis*. Thus, *Psiloceras planorbis* indicated the base of the Jurassic ammonite sequence, and it was natural to regard the type region of that species, Watchet in west Somerset, SW England, also as the type region for the base of the Jurassic (Warrington et al., 1994).

There was, however, one indication that these psiloceratids could not all be stratigraphically equivalent: their occurrence as monotypic accumulations. Examples will be given below.

The Ammonites

Psiloceras sampsoni (Portlock) and *Psiloceras psilonotum* (Quenstedt)

In most regions of the NW European Province the Jurassic ammonite succession begins with an evolute, smooth-shelled form (Pl. 1.2) displaying a suture line typical of *Psiloceras* (Fig. 1 a, b). Two names are in use for this form: *Psiloceras sampsoni* (Portlock) in Britain and *Psiloceras psilonotum* (Quenstedt) on the European continent. The synonymy of both names (Donovan, 1952) is not strictly demonstrated but probable. The only uncertainty concerns the type specimen of *sampsoni* from Northern Ireland; there can be no doubt that the mass of British specimens is not different from Quenstedt's species in Germany. In a large number of papers, however, this evolute form is referred to as *Psiloceras planorbis*. The systematic relationship between *Psiloceras sampsoni/psilonotum* and *P. planorbis* has never previously been resolved unequivocally (see below).

Concentrations of *Psiloceras sampsoni/psilonotum* are known to occur at many localities in Britain as well as France (Guérin-Franiatte, 1990) and Germany. Perhaps the most famous accumulation is that of Nellingen near Stuttgart (SW Germany). Specimens from there occur in many collections all over the world.

Psiloceras sampsoni/psilonotum frequently co-occurs with the earliest psiloceratid which is ribbed throughout ontogeny, *Psiloceras plicatulum* (Quenstedt, 1883). In contrast, *Psiloceras planorbis* s. s. is not found together with *Psiloceras plicatulum*, thereby suggesting that both species are not coeval and that this is also true for *Psiloceras planorbis* and *P psilonotum/sampsoni*.

***Psiloceras planorbis* (Sowerby)**

This species (in a strict sense) is historically only recorded from the coastal sections near Watchet (Pl. 1.3). It is generally less evolute than *Psiloceras sampsoni/psilonotum*, but intermediate forms exist (see below). Comparisons were difficult because *Psiloceras planorbis* from the type area is preserved flattened by compaction; it was previously uncertain how much this crushing had influenced the proportions of the shell (e.g., by increasing whorl height). In contrast, specimens of *Psiloceras sampsoni/psilonotum* are preserved uncrushed (Pl. 1.2) in most regions.

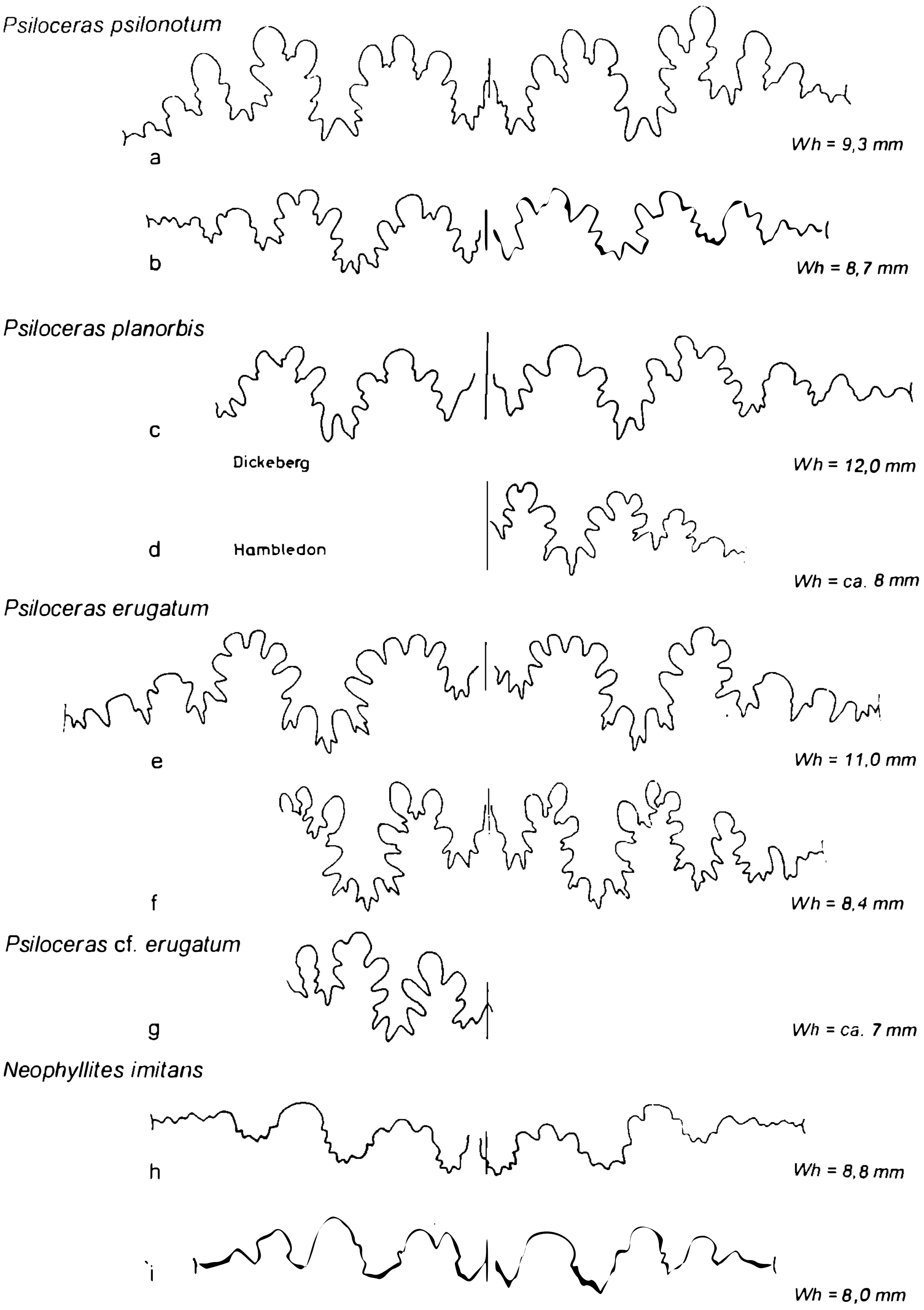
In some regions, however, *Psiloceras sampsoni/psilonotum* is also typically crushed, e.g., in most parts of NW Germany. Nevertheless, even in a totally flattened state, these ammonites remain evolute and can be identified. This is true also for crushed specimens in Great Britain (Pl. 1.1). The effect of crushing is not, as might be expected, a radial movement of the venter which would inevitably increase the circumference of the ammonite. As the shell is rigid, radial fractures would arise and widen to wedge-shaped gaps which separate the fragments. This has not been observed, however. The orientation of the fractures is irregular; long fractures tend to follow the spiral of the whorls (Pl. 1.1, 3, 4, 5, 7, 13).

The changes of shape caused by crushing are compensated partly by some crumbling of the shell, partly by imbricating of shell fragments (see Pl. 1).. The dimensions of the crushed psiloceratids are not, therefore, significantly changed by crushing and meaningful measurements can be made. We have compared the increase of umbilical size in relation to the increase of diameter during ontogeny of both groups; results are given in Fig. 2. The early development of both groups is essentially the same, but in *Psiloceras sampsoni/psilonotum* the umbilicus increases more rapidly than in *P. planorbis*. Thus, in early to middle stages of ontogeny, there is a strong overlap between both groups, making separation difficult. But in late stages of ontogeny, there is essentially no overlap and both groups can well be distinguished. The dissimilar ontogenies indicate that *Psiloceras sampsoni/psilonotum* is not a variety of *Psiloceras planorbis* and should, therefore, be regarded as an independent species. The observation that both forms do not co-occur in accumulations also indicates separation. In the Watchet area the stratigraphic separation is obvious; there *Psiloceras sampsoni/psilonotum* occurs above *Psiloceras planorbis* (see Fig. 5).

This stratigraphic separation is also an argument against a dimorphic relationship between *Psiloceras sampsoni/psilonotum* and *Psiloceras planorbis*, especially as the microconch of *Psiloceras psilonotum* is known (*Psiloceras brevicellatum* Pompeckj). The latter form always co-occurs with *Psiloceras psilonotum* and shows exactly the same variability of proportions. The only differences are: the smaller size, the shorter body chamber and the special, trumpet-like aperture.

Psiloceras planorbis sensu stricto was previously only confirmed in the Watchet area, but a study of unpublished material in British collections has now revealed further localities, at some of which also partly uncrushed specimens were found in marls and limestones (Pl. 1.4, 1.5). In one of the latter specimens the suture line is preserved (Fig. 1d) which was previously unknown. In deep boreholes in NW Germany the species also occurs in a crushed state, but in SW Germany the species is generally lacking. Only one locality, Dickeberg (about 35km SW of Tübingen), has yielded a few – uncrushed – specimens (Pl. 1.6; leg. S. Gräbenstein, Bodelshausen). These specimens are associated with a mass occurrence of *Psiloceras psilonotum*. But as their proportions lie outside the range of variability of *Psiloceras sampsoni/psilonotum* (Fig. 3, top) they are not simply varieties of the latter. *Psiloceras plicatulum* is not present in this accumulation, indicating that it is an earlier assemblage dating from the transition of *planorbis* to *psilonotum*. A separate horizon of *Psiloceras planorbis* has not yet been recognized in SW Germany.

Figure 1. Suture lines of early psiloceratids in the NW European Province. **a**, **b**: *Psiloceras psilonotum*, SW Germany; **a:** Rüdern, SMNS no. 63482, **b:** Nellingen, IGPT no. 1831/51; **c**, **d**: *Psiloceras planorbis*; **c**: Dickeberg, SW Germany (SMNS no. 63686/1); **d**: Hambledon (BGS no.KA 1582); **e**, **f**: *Psiloceras erugatum*, Yorkshire (GSM nos. 23903 and Z 3878); **g**: *P*. cf. *erugatum*, Doniford Bay, bed no. 8 (SMC no. X29338). – **h**, **i**: *Neophyllites imitans*, SW Germany (coll. Wetzel. nos. 943 and 1899, IGPT nos. 1831/36 and 1831/43).


Psiloceras psilonotum
a
Wh = 9,3 mm
b
Wh = 8,7 mm
Psiloceras planorbis
c
Dickeberg
Wh = 12,0 mm
d
Hambledon
Wh = ca. 8 mm
Psiloceras erugatum
e
Wh = 11,0 mm
f
Wh = 8,4 mm
Psiloceras cf. erugatum
g
Wh = ca. 7 mm
Neophyllites imitans
h
Wh = 8,8 mm
i
Wh = 8,0 mm

The study of the section at Doniford Bay near Watchet suggests that there is not a continuous development from *planorbis* to *sampsoni/psilonotum*. The biometric parameters examined for *Psiloceras planorbis* remain essentially the same up to the top of its range, and *Psiloceras sampsoni/psilonotum* is evolute from its first appearance.

***Neophyllites* spp.**

A group of smooth-shelled psiloceratids in the NW European Province differs from *Psiloceras* in the same province by a conspicuously simplified suture line. Already the first observations by Rolle (1858) were rather adequate. The similarity between Rolle's form and *Ammonites* (later: *Psilophyllites*) *hagenowi* Dunker (1847) caused systematic and stratigraphic confusion. Lange (1941b) recognized Rolle's form to be a valid genus different from *Psilophyllites*; he named it *Neophyllites* (see also Bloos, 1999).

Until 1962, *Neophyllites* was known only from loose blocks. Lange (1941a; 1941b) described a mass occurrence from Drove, a locality between Aachen and Köln. He discovered that a further accumulation, mentioned by Thompson (1913) from blocks at the Yorkshire Coast and figured by Spath (1924, textfig. 12 a-c, pl. 18, fig. 1b) belonged to the same group. In both occurrences *Psiloceras* was lacking. Lange therefore created a zone of *Neophyllites antecedens* and supposed that it was earlier than *Psiloceras psilonotum*. In contrast, in Württemberg (SW Germany) *Neophyllites* seemed to occur together with *Psiloceras psilonotum* (Wetzel, 1929; 1932; Lange, 1941; Altmann, 1965).

Deep boreholes in NW Germany (near Bislich and Westerwanna; Bloos, 1999, Fig. 1) revealed *Neophyllites* in sequence with *Psiloceras* for the first time (Hoffmann, 1962); there, as Lange had assumed, *Neophyllites* occurs below *Psiloceras psilonotum* and is the earliest known Jurassic ammonite group in that region. Recently, in SW Germany, a level of *Neophyllites* below *Psiloceras psilonotum* has also been identified (Bloos, 1999); the main species here are the same as at Drove and in the Bislich borehole: *Neophyllites antecedens* and *N. imitans*. *Psiloceras psilonotum* is occasionally associated with *Neophyllites*, but with different species (*Neophyllites becki* and *N. trossingensis*).

Neophyllites is distinguished from *Psiloceras* (and from *Psilophyllites*) mainly by its suture (Fig. 1h; 1i) and it is often difficult, therefore, to identify crushed specimens of *Neophyllites* lacking sutures. Most species of *Neophyllites* differ from *Psiloceras* by having less inflated flanks and a steeper umbilical wall. In partially crushed specimens (e.g., in marls) the steep umbilical wall can be still recognized (Pl. 1.7). In some species, sharp spiral rills occur which are preserved even in totally flattened specimens and are, therefore, a good feature to aid recognition of *Neophyllites* (Pl. 1.7, 8, 9). This character is not, however, developed on all specimens.

***Psiloceras erugatum* S. Buckman**

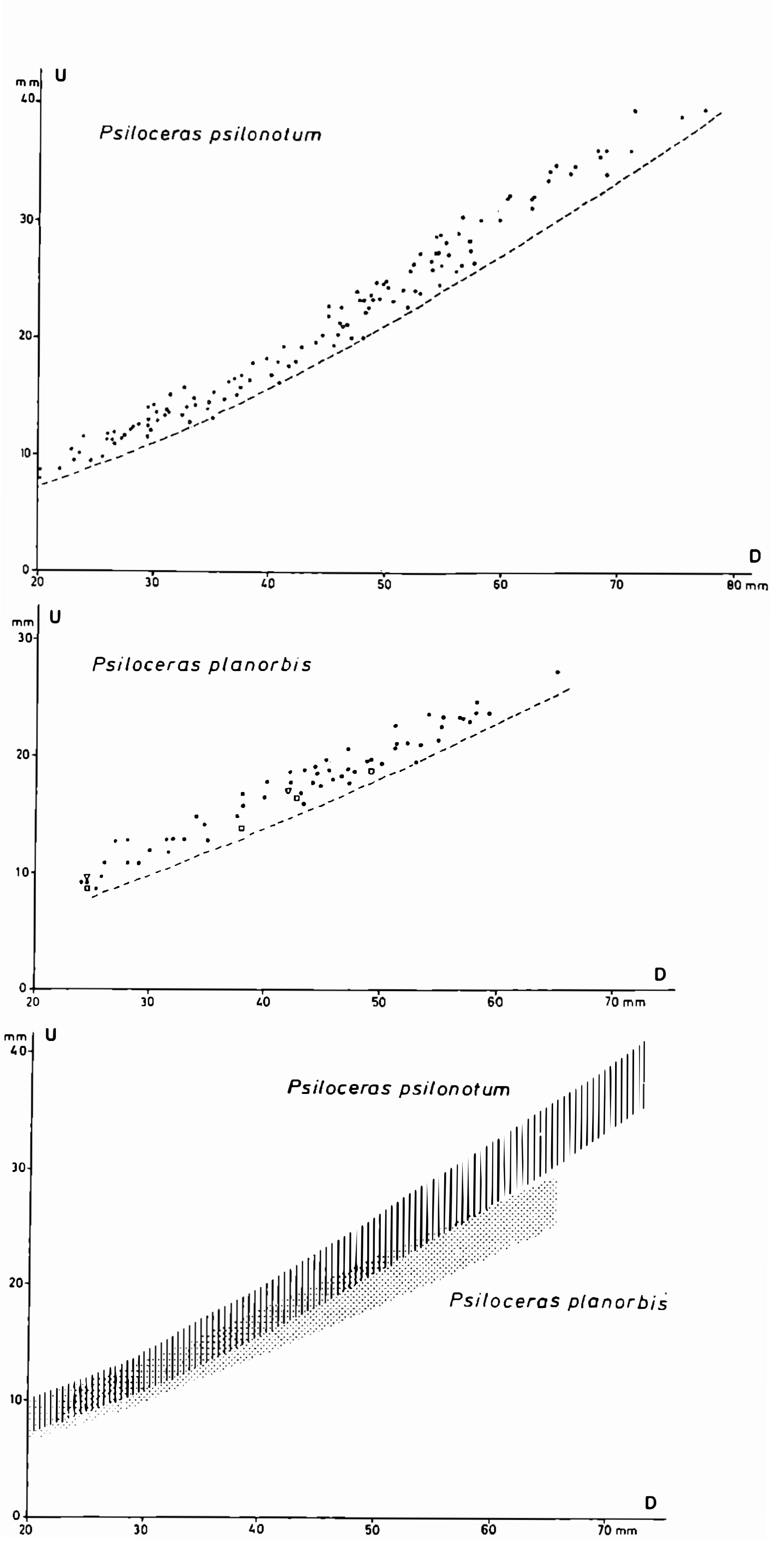
Previously, this species of NW European psiloceratids has been overlooked almost totally. It has the same coiling, cross section, and suture line as *Psiloceras sampsoni/psilonotum* (Fig. 1e; 1f). The only difference is the presence of well-developed tubercles on the innermost whorls, followed by irregular ribbing which ends more or less early in ontogeny (Buckman, 1921, pl. 223).

This species was known virtually only from loose blocks on the Yorkshire Coast (Thompson, 1913); other occurrences were not known. In 1966, however, Donovan identified the species in the Wilkesley Borehole, but regarded it as a subspecies of *Psiloceras planorbis*. This record has been subsequently overlooked, and the stratigraphic significance of that form remained unappreciated.

***Psiloceras aequabile* S. Buckman**

This form (Buckman, 1923, pl. 390), represented by the crushed type specimen, is poorly defined; its horizon is unknown and it is rarely used in the literature. It is therefore not included in the present study. It may be a synonym of *Psiloceras planorbis*.

Figure 2. Relation of width of umbilicus (U) to diameter (D) during ontogeny in *Psiloceras psilonotum* (SW Germany) and *P. planorbis* (England; dots: Watchet, squares: Wilkesley Borehole). – Hatched lines: lower boundaries of variability for comparison with Figure 3, top.


mm
U
40
30
20
10
0
Psiloceras psilonotum
D
20
30
40
50
60
70
80 mm
mm
U
30
20
10
0
Psiloceras planorbis
D
20
30
40
50
60
70 mm
mm
U
40
30
20
10
0
Psiloceras psilonotum
Psiloceras planorbis
D
20
30
40
50
60
70 mm

The Psiloceratid Sequence

Hodges' (1994) was the first to discover ammonites below the traditional "first appearance" of *Psiloceras planorbis* near Watchet in "bed 13" of Whittaker and Green (1983). The preservation of Hodges specimens was poor, however, and specific identification was not possible. In the same year (1994) one of us (K. N. P.) found additional specimens, also in beds 8 and 9 of Whittaker and Green (1983), some of which were better preserved.

Three small specimens have been extracted from bed 8, a marly limestone. Two of them show weak plications (Pl. 1.12), one is smooth. In one specimen a partial suture line is preserved (Fig. 1g), which is of *Psiloceras* type. These specimens are partially crushed and preparation is almost impossible, making illustration difficult. As inner whorls cannot be seen, it is uncertain whether the tubercles characteristic of *Psiloceras erugatum* are present.

The lower part of bed 9, a marl, has also yielded small psiloceratids, especially near the cliffs named Blue Anchor, to the west of Watchet. Some specimens are less crushed and show characters of *Neophyllites*, most probably *N. antecedens* (Pl. 1.10). In the upper part of bed 9 relatively involute psiloceratids appear, probably the first specimens of *Psiloceras planorbis.*

The known sequence in the Watchet area (Fig. 5) is now as follows (based mainly on observations in Doniford Bay); bed numbers follow Whittaker and Green (1983), and thicknesses are given in metres:

Bed 25	First Caloceras (base of the Johnstoni Subzone)	
Bed 24	1.5m	*Psiloceras sampsoni + Psiloceras plicatulum*
	0.4m	*Psiloceras sampsoni*
	0.1m	*Psiloceras plicatulum*
	0.3m	*Psiloceras sampsoni*
	0.6m	*Psiloceras planorbis*
Bed 23-13	1.6m	*Psiloceras planorbis*
Bed 12-10	0.3m	No ammonites
Bed 9	0.25m	*Psiloceras* cf. *planorbis*
	0.10m	Psiloceratidae with indication of *Neophyllites antecedens*
Bed 8	0.15m	*Psiloceras* cf. *erugatum*

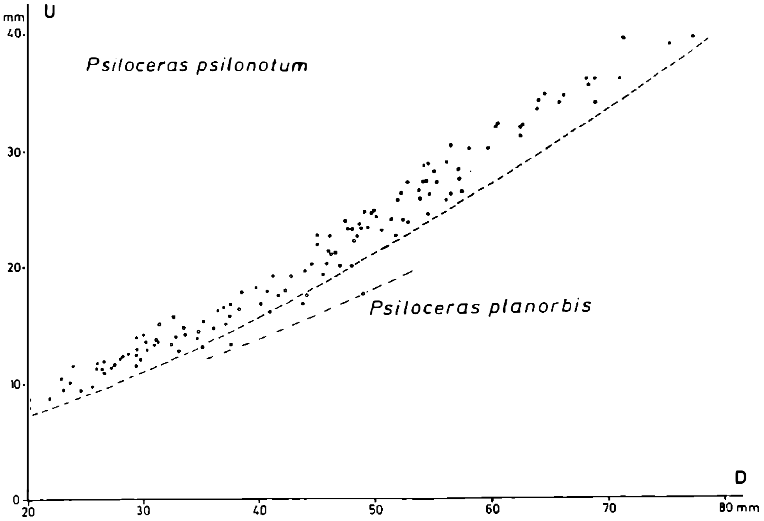
[Formal subdivision of this sequence into biohorizons is discussed by Page *in* Page and Bloos, 1998.]

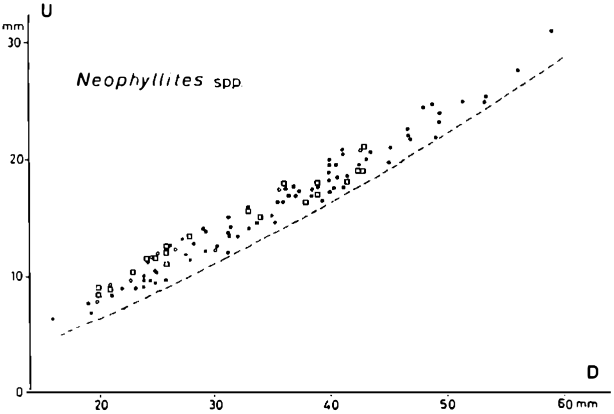
The first section in the NW European Province which yielded the known psiloceratids in stratigraphic succession is the Wilkesley Borehole in Cheshire (NW England; Poole and Whiteman, 1966). From this borehole (made 1959/60) a great number of smooth-shelled, crushed ammonites were obtained over a range of 13.8m (68 specimens from 52 levels, see Fig. 4). Donovan (*in* Poole and Whiteman, 1966) gave a short account of these ammonites, most significantly including the recognition of *Psiloceras planorbis erugatum* (Pl. 1.13) at the base of the sequence. All other material was allocated to *Psiloceras planorbis.*

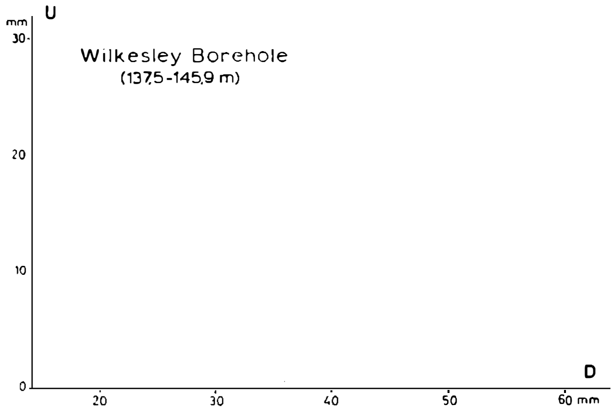
A restudy of the Wilkesley psiloceratids, kindly arranged by S. Tunnicliff and G. Warrington (BGS, Keyworth), has now revealed the following sequence (Fig. 4):

Depth [m]	Thickness [m]	Ammonites present
134.6		First *Caloceras* (base of the Johnstoni Subzone)
136.8	2.2	*Psiloceras sampsoni + Psiloceras plicatulum*
137.5	0.7	*Psiloceras sampsoni*
138.2	0.7	*Psiloceras planorbis*
143.7	5.5	Psiloceratids, *Neophyllites antecedens* predominant
145.9	2.2	Psiloceratids, *Neophyllites imitans* predominant
147.9	2.0	No ammonites
148.4	0.5	*Psiloceras erugatum* (base of Planorbis Subzone)
157.9	9.5	"Pre-Planorbis Beds"

Figure 3. Relation of width of umbilicus (U) to diameter (D) in: top: *Psiloceras psilonotum* with *P. planorbis* from Dickeberg, SW Germany (hatched lines in same position as in Figure 2); middle: *Neophyllites* spp., SW (dots) and NW (squares) Germany; lower: *Neophyllites* in the Wilkesley Borehole.


mm
U
40
30
20
10
0
Psiloceras psilonotum
Psiloceras planorbis
D
20
30
40
50
60
70
80 mm


mm
U
30
20
10
0
Neophyllites spp.
D
20
30
40
50
60 mm


mm
U
30
20
10
0
Wilkesley Borehole
(137.5-145.9 m)
D
20
30
40
50
60 mm

In this section the beds with *Neophyllites* are far thicker than anywhere else in Europe. There is no strict separation of species of *Neophyllites*, but in the upper part the evolute *N. antecedens* predominates whereas in the lower part the less evolute *N. imitans* is most frequent. A plot displaying the relation of the width of umbilicus to the diameter during ontogeny of the specimens from the *Neophyllites* interval in the borehole shows the same variability as *Neophyllites* from Germany (Fig. 3, middle and lower) and is different from the plot of *Psiloceras planorbis*.

A similar sequence has recently been found by one of us (K. N. P.) in the Staithes Borehole in Yorkshire. There, the preservation of *Psiloceras erugatum* is slightly better than in the Wilkesley Borehole and confirms that this species occurs below *Neophyllites*. An account of the succession of ammonites of the Staithes Borehole has not yet been published. There may be other occurrences of *Psiloceras erugatum* in sections which have not been studied in detail (e.g., in the Carlisle Basin, Ivimey-Cook et al., 1995).

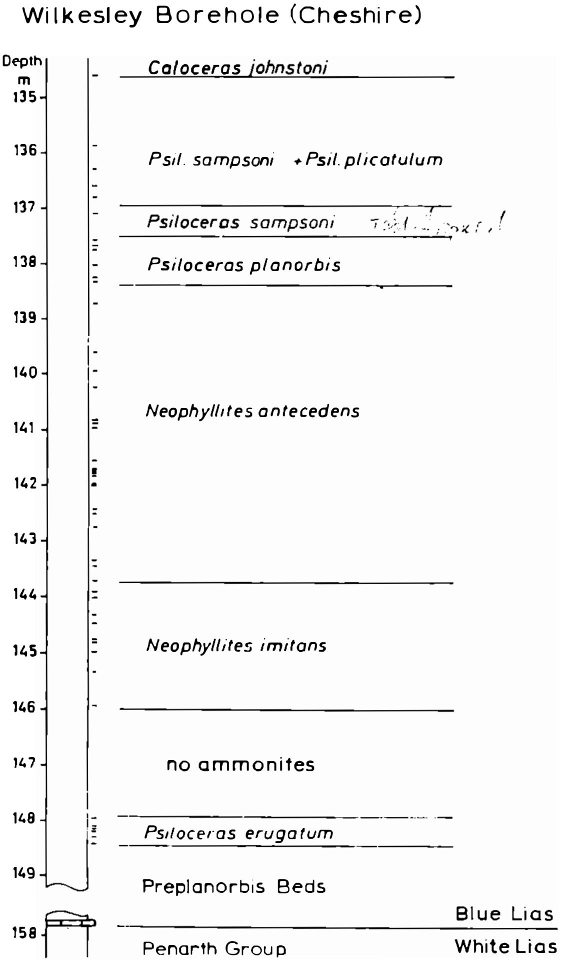


Figure 4. Section of the Planorbis Subzone in the Wilkesley Borehole. In the monotonous marl there are no horizons of limestone or bituminous shale. The levels which yielded ammonites are indicated on the left.

Proposed Type Section for the Base of the Jurassic in Britain

In the Watchet area a candidate GSSP for the base of the Jurassic has been proposed. Because of the better preservation of the earliest psiloceratids, Yorkshire would be more suitable but there no surface exposures are known. In Cheshire, outcrops along the river Weaver (Poole and Whiteman,

1966) may show similar levels to parts of the Wilkesley Borehole, but they have not been studied in detail and it is not even known if they still exist. Watchet will probably remain the area with the most important surface exposures of the basal Jurassic in Britain

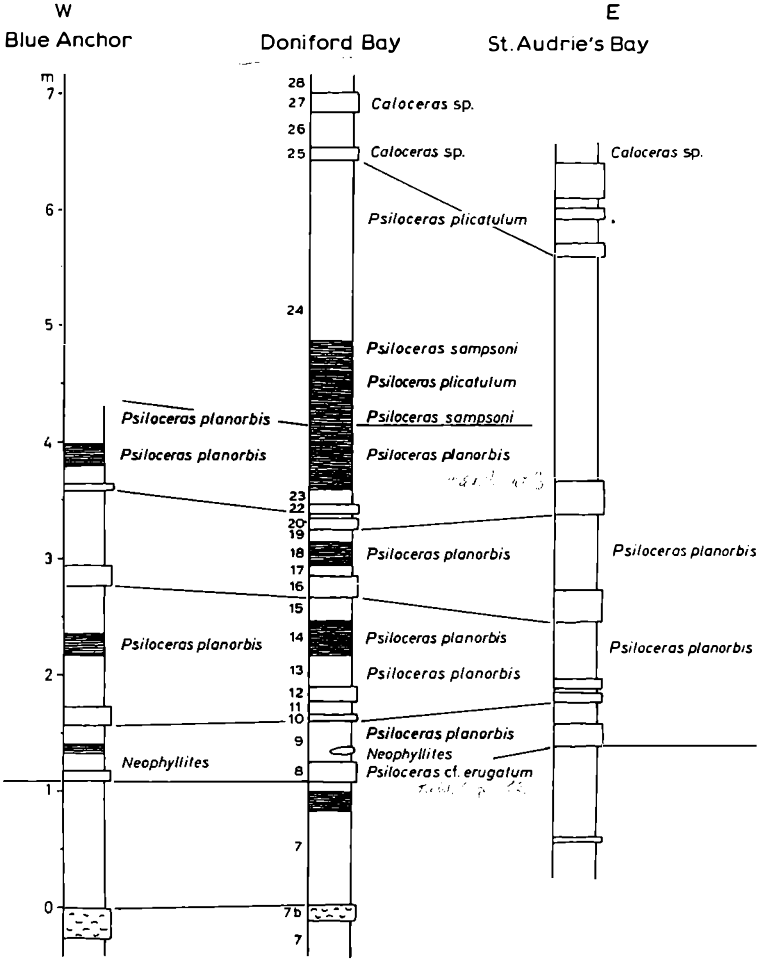


Figure 5. Sections near Watchet/West Somerset. White: marl, dark: bituminous shale, hard layers: marly limestone. In the section 'St. Audrie's Bay' the bituminous layers are not indicated.

Conclusions

There is a stratigraphic sequence of smooth-shelled psiloceratids in the earliest Hettangian (Planorbis Subzone) of the NW European Province. Only in very few places is the sequence complete with *Psiloceras erugatum* at the base. *Neophyllites* is commonly the earliest ammonite group in the subsurface records of NW Germany and in restricted areas in SW Germany. No region is currently known where this sequence begins with *Psiloceras planorbis*. In most regions, *Psiloceras sampsoni/psilonotum* remains the first Jurassic ammonite recorded. The beginning of the Jurassic ammonite succession is obviously diachronous across the Province.

The earliest records of ammonites in different regions do not, of course, necessarily reflect the first arrival of ammonites. The lack of earlier forms may also result from stratigraphic gaps, paleoecological conditions, or collecting failure. Where the first psiloceratids appear immediately above Triassic beds of strikingly different facies, a gap obviously exists. The gap might be caused either by later transgression of the sea or a reworking of older marine beds. Where there are sediments of "Pre-Planorbis" aspect below the first Jurassic ammonites, earlier ammonites may be absent either because of unfavourable conditions (especially in nearshore sediments around uplifts) or because they have been overlooked as was the case in the Watchet area. In part, this can be due to a discontinuous distribution, with ammonites concentrated in local accumulations, perhaps current-generated. This should be kept in mind in discussions of "first appearance".

The new results show that correlation with other faunal provinces is even more difficult than previously assumed. Correlations were formerly based simply on records of "*Psiloceras planorbis*"; many of these records are now believed to be problematic. If, for instance, *Psiloceras tilmanni* in South America is really an equivalent of *Psiloceras planorbis*, as has been assumed, it could hardly be the earliest Jurassic ammonite there. The similarity of these two forms is, however, only superficial. There are certainly more differences than similarities. Hillebrandt has shown in several contributions (1988; 1990; 1994; 1997), an independent sequence of psiloceratids is developed in South America which is similar to that in North America (Guex, 1982; 1995; Guex et al., 1997), but very different from that in the NW European Province. The early psiloceratids of the Alps are also conspicuously different from those of South and North America and of NW Europe. Provincialism is obviously stronger in the earliest Jurassic than previously assumed, and world-wide correlation by ammonites remains difficult.

Acknowledgments

P. Hodges (then National Museum of Wales) kindly showed his first records of pre-*planorbis* ammonites to the authors prior to publication. H. Prudden (Montacute, Somerset) assisted in the field at Doniford Bay and provided information on inland sites. H. C. Ivimey-Cook (formerly of the British Geological Survey) also assisted at Doniford Bay and indicated to the authors the existence of the important Wilkesley Borehole material; G. Warrington and S. Tunnicliff facilitated access to this and other materials in BGS collections. J. Radley assisted both at Doniford Bay and then with access to collections in Bristol City Museum. S. Baker (Natural History Museum London), P. Powell (Oxford University Museum) and D. Norman (Sedgwick Museum Cambridge) similarly assisted with access to historical collections.

References

ALTMANN, H. J. 1965. Beiträge zur Kenntnis des Rhät-Lias-Grenzbereichs in Südwest-Deutschland. Thesis University of Tübingen, 117 pp.

BLOOS, G. 1998. *Neophyllites* (Ammonoidea, Psiloceratidae) in the earliest Jurassic of South Germany. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 207 [in press].

BUCKMAN, S. S. 1919-1930. Type Ammonites. Vols 3-7. London & Wesley.

DONOVAN, D. T. 1952. The ammonites of the Blue Lias of the Bristol District. Part 1. Psiloceratidae and Schlotheimiidae. The Annals and Magazine of Natural History, series 12, 5: 629-655.

____. 1966. The Hettangian ammonites from the Wilkesley Borehole, p. 50-51, 139-140. *In* E. G. Poole and A. J. Whiteman (eds), Geology of the country around Nantwich and Witchurch (Explanation of the one-inch geological sheet 122). Memoir of the Geological Survey of Great Britain, England and Wales.

GUÉRIN-FRANIATTE, S. 1990. Ammonites du Lias inférieur de France. Part 2. Psiloceratidae, Schlotheimiidae et premiers Arietitidae. Mémoire du Centre d'Etudes et de Recherches géologiques et hydrologiques, 29:206.

GUEX, J. 1982. Relation entre le genre Psiloceras et les Phylloceratida au voisinage de la limite Trias/Jurassique. Bulletin de Géologie Lausanne, 250:127-140.

____. 1995. Ammonites hettangiennes de la Gabbs Valley Range (Nevada, USA). Mémoire de Géologie Lausanne, 27:131.

____, M. RAKUS, D. TAYLOR AND H. BUCHER. 1997. Selection of a GSSP candidate for the base of the Jurassic System. 2.2.2. Proposal for the New York Canyon area, Gabbs Valley Range (Nevada) USA. International Subcommission on Jurassic Stratigraphy, Newsletter Number 24:26-30.

HILLEBRANDT, A. v. 1988. Ammonite biostratigraphy of the South American Hettangian. Description of two new species of *Psiloceras*, p. 55-70. *In* R. B. Rocha and A. S. Soares (eds), 2nd International Symposium on Jurassic Stratigraphy, 1.

____. 1990. The Triassic/Jurassic boundary in Northern Chile.- Cahiers de l'Université Catholique Lyon, 3:27-53.

____. 1994. The Triassic/Jurassic boundary and Hettangian biostratigraphy in the area of the Utcubamba Valley (Northern Peru). Geobios, Mémoire spécial, 17:297-307.

____. 1997. Selection of a GSSP candidate for the base of the Jurassic System. 2.2.1. Proposal for the Utcubamba Valley sections in Northern Peru. International Subcommission on Jurassic Stratigraphy, Newsletter Number 24:21-25.

HODGES, P. 1994. The base of the Jurassic System: new data on the first appearance of *Psiloceras planorbis* in southwest Britain. Geological Magazine, 131:841-844.

HOFFMANN, K. 1962. Lias und Dogger im Untergrund der Niederrheinischen Bucht.- Fortschritte der Geologie in Rheinland und Westfalen, 6:105-184.

IVIMEY-COOK, H. C., G. WARRINGTON, N. E. WORLEY, S. HOLLOWAY AND B. YOUNG. 1995. Rocks of Late Triassic and Early Jurassic age in the Carlisle Basin, Cumbria (north-west England). Proceedings of the Yorkshire Geological Society, 50:305-316.

LANGE, W. 1931. Die biostratigraphischen Zonen des Lias alpha und Vollraths petrographische Leithorizonte. Centralblatt für Mineralogie, Geologie und Paläontologie, B, 1931:349-372.

____. 1941a. Über das das Liasvorkommen von Drove bei Düren, Rheinland. Zeitschrift der Deutschen Geologischen Gesellschaft, 93:134-141.

____. 1941b. Die Ammonitenfauna der Psiloceras-Stufe Norddeutschlands. Palaeontographica, A, 93:1-192.

PAGE, K. N. 1994. Appendix to account of Field Excursions: Preliminary observations on the ammonite faunas of the basal Jurassic in Somerset and their global context. Proceedings of the Ussher Society, 8:341-344.

____. AND G. BLOOS. 1998. The base of the Jurassic System in West Somerset, south west England – new observations on the succession of ammonite faunas of the lowest Hettangian stage. Proceedings of the Ussher Society, 12:1-5.

POOLE, E. G. AND A. J. WHITEMAN. 1966. Geology of the country around Nantwich and Whitchurch (Explanation of one-inch geological sheet 122). Memoir of the Geological Survey of Great Britain, England and Wales, p. 42-50.

ROLLE, F. 1858. Über einige an der Grenze von Keuper und Lias in Schwaben auftretende Versteinerungen. Sitzungsberichte der kaiserlichen Akademie der Wissenschaften, mathematisch-naturwissenschaftliche Classe, 26 (1857):13-32.

SPATH, L. F. 1924. The Ammonites of the Blue Lias. Proceedings of the Geologists' Association, 35:186-211.

THOMPSON, C. 1913. The derived Cephalopoda from the Holderness Drift. Quarterly Journal of the Geological Society of London, 69:169-183.

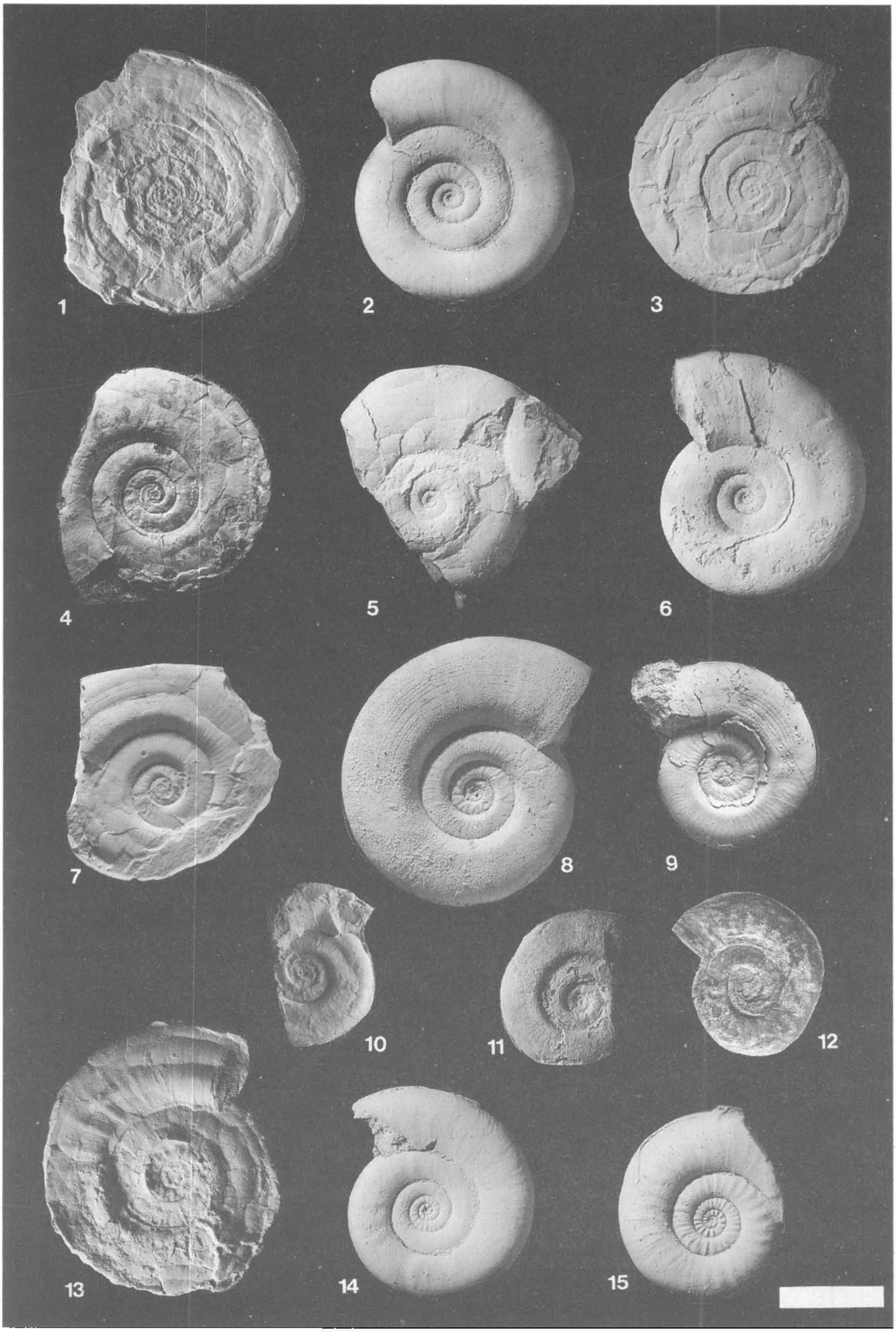
WARRINGTON, G., J. C. COPE AND H. C. IVIMEY-COOK. 1994. St. Audrie‘s Bay, Somerset, England: a candidate Global Stratotype Section and Point for the base of the Jurassic System. Geological Magazine, 131:191-200.

WETZEL, R. 1929. Grenzprobleme zwischen Geologie und Paläontologie. Verhandlungen der Physikalisch-Medizinischen Gesellschaft Würzburg, new series 54:178-187.

____. 1932. "Leitfossil gegen Leithorizont" und die mittelwürttembergische Psilonotenbank. Neues Jahrbuch für Mineralogie, Geologie, Paläontologie, B, 67:455-468.

WHITTAKER , A. AND G. W. GREEN. 1983. Geology of the country around Weston-super-Mare. Memoir for 1: 50000 geological sheet 297, New Series, with parts of sheets 263 and 295. Geological Survey of Great Britain, England and Wales, p. 59-78.

Plate 1. Early psiloceratids from Britain with some specimens from Germany for comparison.- **1.** *Psiloceras sampsoni*, Wilkesley Borehole, depth 137,08 m (GSM no. AL 727); **2.** *Psiloceras psilonotum*, Nellingen, Psilonotenbank, SMNS no. 63697; **3.** *Psiloceras planorbis* from bituminous shale, totally flattened, Doniford Bay, SMNS no.63676; **4.** *Psiloceras planorbis* from marl, less crushed, Blue Anchor (GSM no. Zk 4110); **5.** *Psiloceras planorbis*, from top of a limestone layer, inner whorls uncrushed, Cotham Station, Worcestershire (GSM no. 119451); **6.** *Psiloceras planorbis*, Dickeberg, SW Germany (SMNS no. 63686/1); **7.** *Neophyllites* sp. with spiral rills and steep umbilical wall, Wilkesley Borehole, depth 139,95 m (GSM no. 748); **8.** *Neophyllites becki* with spiral rills on the shell, "Im Höfle" near Rotenzimmern, SW Germany (SMNS no. 63822); **9.** *Neophyllites sulcifer* with spiral rills on the internal mould, Bebenhausen near Tübingen (coll. Wetzel no. 558, IGPT no. 1831/35); **10.** *Neophyllites antecedens*, Wilkesley Borehole, depth 141,96 m (GSM no. 769b); **11.** *Neophyllites antecedens*, Tübingen, SW Germany (leg. H. Hölder, IGPT no. 1831/47); **12.** *Psiloceras* cf. *erugatum*, Doniford Bay, bed 8 (SMC no. X29337); **13.** *Psiloceras erugatum*, Wilkesley Borehole, depth 145,19 m (GSM no. 794a); **14.** *Psiloceras erugatum*, Robin Hood's Bay, Yorkshire (GSM no. 23915); **15.** *Psiloceras erugatum*, ribbing longer persistent, Robin Hood's Bay, Yorkshire (GSM no. Z 3878). Scale: 2cm. Abbreviations: GSM Geological Survey Museum, Keyworth; IGPT Institut und Museum für Geologie und Paläontologie der Universtät Tübingen; SMC Sedgwick Museum Cambridge; SMNS Staatliches Museum für Naturkunde Stuttgart.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On the Proposed Basal Boundary Stratotype (GSSP) of the Middle Jurassic Callovian Stage

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Abstract: Sections through sedimentary successions across the Bathonian–Callovian stage-boundary are numerous and well-known in five of the seven continents. Yet, while the requirements specified by the International Commission on Stratigraphy for the ideal boundary stratotype can all be met individually in one section or another, frequently in many, no section known so far can simultaneously fulfill all of them.

In deciding which requirements should be given most weight it is important to recall the primary purpose of a GSSP. This is perhaps particularly important in the Jurassic, in which chronostratigraphy has always gone beyond the level of Stages in the standard hierarchy. The basic principles are therefore reviewed. They involve the vital distinction between definition and recognition of a boundary, hence the precision of time-correlations attainable by means, here, of ammonites as guide-fossils and hence of the time-scales and time-resolution of their biostratigraphy. The finest time-resolvable ammonite biozones are the so-called faunal horizons.

The proposed Bathonian–Callovian boundary is taken to lie between two faunal horizons that can be most widely recognised and correlated, those of *Clydoniceras discus* below and *Kepplerites keppleri* above. The *keppleri* horizon, or levels very close to it, can be recognised as far afield as Japan, Alaska, British Columbia, East Greenland, most of northern Europe, Russia and the Caucasus. The stratotype section has been chosen close to where the time-diagnostic ammonite faunas have been most extensively characterized, in the Macrocephalen-Oolith Formation at Albstadt-Pfeffingen, 65km south of Stuttgart in Germany. Some results based on non-biostratigraphical methods of correlation, such as geomagnetism and strontium isotope stratigraphy, are reported. None so far approaches the precision of time-correlation of ammonite biostratigraphy.

Introduction

A concrete proposal to designate a basal boundary stratotype of the Callovian Stage was essentially complete in 1990. A meeting of the Callovian Working Group, set up in Erlangen in 1984, was held at Stuttgart and in Albstadt (Swabian Alb, Germany) in September 1990, at which 18 members were present. Detailed arguments were laid out in a document of 20 pages (Callomon and Dietl, 1990, unpublished, CandD for short), supported by an oral presentation by the chairman, and extensive demonstrations of the very rich, characteristic ammonite faunas at the museum in Stuttgart and a visit to the proposed, freshly excavated, type-section near Albstadt-Pfeffingen. At the conclusion, the proposals were adopted as a whole and unanimously. Formal submission to the International Commission on Stratigraphy (ICS) via its Subcommission on Jurassic Stratigraphy (ISJS) has been delayed for practical reasons. The proposals were however reported in the Newsletter of the ISJS (no. 20, 1991) and published in the volume of the Jurassic Symposium at Poitiers in 1991 (Callomon, 1994b). Nothing has happened since to make it necessary to change the proposals in any way. They will, therefore, be repeated here only in outline. A fully documented submission for ratification by the ISJS and ICS will be published elsewhere. The purpose of this note is, however, to resolve some misconceptions contained in, or implied by, the Guidelines of the

ICS and hence to rebut objections to the proposals that have been voiced; and to indicate some new results, particularly of non-biostratigraphical measurements of geochronology, that have been obtained in recent years.

Principles

Historical

First attempts to codify stratigraphic procedure and terminology go back at least as far as the 2nd International Congress (IGC) of 1881. They were renewed in strength after the second world war at the 19th Congress in Algiers, largely under the stimulus of the American Commission on Stratigraphical Nomenclature (Hedberg, 1954). Subsequent developments, through the foundation of the International Union of Geological Sciences (IUGS), have given us the ICS of today, with its numerous Subcommissions (ISJS for the Jurassic) and their Working Groups, charged with the task of completing the codification.

The pronouncements of many of these bodies must, however, strike Jurassic workers in particular as somewhat variable in realism and logic of method. The two documents of immediate interest are the Guidelines of the ICS (Cowie et al., 1986; "G-1" for short) and the Revised Guidelines (Remane et al., 1996; "G-2" for short). These lengthy documents setting out the tasks to be carried out by the world of working stratigraphers seem particularly daunting, especially as the goals to be achieved appear to change with time.

First, the lowest "units for world-wide correlation" (*sic*) were to be Stages. These were to be defined in terms of International Stage Stratotypes. The first attempts to fulfil this objective in the Jurassic were made at the first Jurassic Colloquium held in Luxembourg in 1962 (Maubeuge, 1964). Those of us old enough to have been present remember the confusion resulting from the absence of any clearly formulated, never mind agreed, set of principles of procedure. The idea of Stage Stratotypes was quickly seen to be futile and abandoned, and in the event, Jurassic Stages were defined in terms of their contained Zones, exactly as had been done by Oppel a century previously.

Based on this experience, some of us therefore made a start on deriving such a set of principles (Ager, 1963; Callomon, 1964; 1965) and building on the pioneering work of Arkell (1946); and the Callovian Stage was probably the first to be defined in their terms, down to the typological level of a boundary stratotype. The importance of boundary stratotypes rather than whole-stage stratotypes was particularly stressed, seven years before its adoption by the ICS. There have been subsequent refinements and amplifications (Callomon, 1985b; 1995) and, in the Callovian, much new stratigraphical information has been obtained in the years 1980-94; but the definitions now being proposed differ little in practical effect from those of 1964. The Bathonian–Callovian boundary is today drawn at precisely the same level as it was in 1964, and even since Oppel's time.

What has changed is the meaning of "precisely" and this goes to the heart of the scientific, as opposed to nomenclatural, problem: the precision of distant – as opposed to local – time-correlations in stratified rocks. This may now have reached the attainable limits at the Bathonian–Callovian boundary and is largely independent of any individual section. It also vastly exceeds what is needed for what were meant to be the "lowest units for world-wide correlation", the Stages. The precise choice of any one section for the definition of a Stage boundary stratotype has, perhaps seemingly somewhat paradoxically, become therefore a matter of secondary scientific importance (but, in other Stages, a matter not without geopolitical overtones) – which may be the cause of the puzzlement in the minds of some commentators (Remane, 1996, p. 22).

Principles of Chronostratigraphical Classification

- **P1:** Stages are standard chronstratigraphical units – members of standard chronostratigraphical scales, the rock-equivalents of successive continuous durations of geological time termed Ages.
- **P2:** Like all chronostratigraphic units, standard or general, Stages are defined by bounding time-planes in stratified rocks, corresponding to instants of time marking the beginnings and ends of their reciprocal Ages. They are marked typologically each by a Golden Spike in a type section, the Global Stratotype Section and Point (GSSP). These time-planes extend over the whole of the surface of the Earth ("Global") and occur in strata – or stratigraphical gaps – of corresponding ages everywhere.

But these time-planes cannot be explicitly recognised anywhere other than at the places at which they are defined. The best that can be done is to locate them approximately through time-correlation with their standard stratotypes, by means of some chronometric physical attribute of the rocks. The precision of such correlations depends on the time-resolution attainable in the stratigraphic succession by means of the chronometer, and the universality of such correlations, or chains of correlations, depends on the geographical distribution of the chronometer. Stratified rocks are therefore dated not in terms of the bounding time-planes defining the chronostratigraphic units to which they are assigned, but in terms of what lies between them, in terms of the contents of the units.

These considerations are so vital to the discussion of choice of stratotype section that follow that they have been specially stressed. Most of them are to be found, implicitly or explicitly, somewhere or other in the Revised Guidelines (G-2); but failure to realize fully their implications has led to confusion both in the document itself and in the minds of some of its readers (see below).

- **P3:** Stages are members of a hierarchy of standard chronostratigraphical scales (Callomon, 1965), at level V in a taxonomy recently reviewed by one of us (Callomon, 1995). This principle has now been acknowledged in G-2, *en passant* (section 2.5: "Boundary stratotypes instead of unit-stratotypes" [!], para. 5) but appears there to apply only to Stages and upwards. As is rather well known, however, in the Mesozoic, and in the Jurassic *a fortiori*, this hierarchy has been extended downwards to two further levels - Zones (level VI) and Subzones (level VII) - since Oppel's time.

The taxonomic consequences were also discussed previously (Callomon, 1965), for the principles of zoological taxonomy do in fact provide a powerful analogy (not "Zoological nomenclature", as G-2 would have it in section 2.5, para. 6). A Standard Stage is therefore defined by its contents of Zones and Subzones, and its basal boundary stratotype by that of its lowest Subzone (or Zone, if undivided). And an "Oppel Zone" is just such a standard chronozone in the hierarchy (Callomon, 1995), and not a floating biozone, or even floating general chronozone (see Callomon, 1994a for an explanation of the difference) as Remane (1996, p. 22) would have us believe Guex (1991) showed it to be.

- **P4:** The ability to recognise chronostratigraphical units other than in their type-sections - their practical "globality" - may be limited by the characteristics of the chronometer used, e.g., in biostratigraphy, by the bioprovincialism of guide-fossils, or the constraints of sampling in boreholes. It is, therefore, possible and legitimate to set up as many parallel standard chronostratigraphical scales defined by time-planes as are useful or necessary - all of them "global". The Silurian and Devonian can in parts be zoned almost equally well by graptolites, conodonts or ammonoids, depending on facies; around the Jurassic–Cretaceous boundary, rocks are dated by Russian ammonites (Volgian–Ryazanian Stages), Tethyan ammonites (Tithonian–Berriasian Stages) or calpionellids (Zones). In the Callovian–Oxfordian–Kimmeridgian Stages we have parallel standard ammonite zonations that are of equal merit but that are mutually exclusive. Their classifications and nomenclatures are matters of convention. Their characterization and intercorrelations are matters of practical science. The final step is, therefore, the selection of one of the standards as the <u>primary</u> standard (whose boundary stratotypes might have been more appropriately termed Primary Standard Stratotype-sections and Points – PSSP?).

- **P5:** The selection of a stratotype section should therefore be governed by two considerations: (a) correlatability of levels immediately above and below the proposed boundary level; and (b) adherence as far as possible to historical priority and existing convention, on grounds of stability.

Of these, the first is overriding - on which at least we and the *Guidelines* can agree (G-2, section 2.3). But note: the correlations are <u>*time*</u>-correlations, and there is no reference to thicknesses or other lithological criteria of strata. The second consideration should present few problems in the Jurassic as far as selecting levels is concerned, for Jurassic chronostratigraphy has attained a stable maturity (within the qualifications of P4 above). If these levels also fulfil the first consideration, the choice of locality becomes of secondary importance, as has already been mentioned above.

Definition of the Callovian Stage

The Zones of the Callovian

The first to refine d'Orbigny's Stage was Oppel (1857), who redefined it in terms of three successive constituent Zones (p. 504): Macrocephalus, Anceps and Athleta Zones. These have become more or less the Lower, Middle and Upper Callovian of today. For present purposes, we are interested in the lowest of these, the Macrocephalus Zone. The chronometer here, as elsewhere in the Jurassic, is the biostratigraphy of the ammonites as guide-fossils for correlation and the standard chronostratigraphical units are also all named after ammonites as index-species. (For the important distinction between guide- and index-species, see Callomon 1995).

The history of subsequent refinement by subdivision based on advances in biostatigraphy is summarized in Figure 1 and the current status (right-hand column) is discussed by Callomon, Dietl and Page (1989) and Page (1989). The only point perhaps worth repeating is the reason for the change in name of the Macrocephalus Zone. This had become more and more restricted by subdivision of its original scope, and whereas *M. macrocephalus* (Schlotheim) is a species that undoubtedly occurs within its nominal Zone as originally defined (see P2 above), its precise position within that Zone was not known until recently. It transpired that its type-horizon does not now lie within the restricted Macrocephalus Zone. The latter has therefore to be renamed. *M. herveyi* (Sowerby) was available as alternative index since its proposal by Spath (1932) and it certainly does occur in its renamed nominal Zone. The change is purely nomenclatural: Herveyi Zone, *nom. nov. pro* Macrocephalus Zone, Oppel *partim, emend.* Callomon 1955. (Note here again the distinction between taxonomy and nomenclature, now applied to stratigraphical entities.)

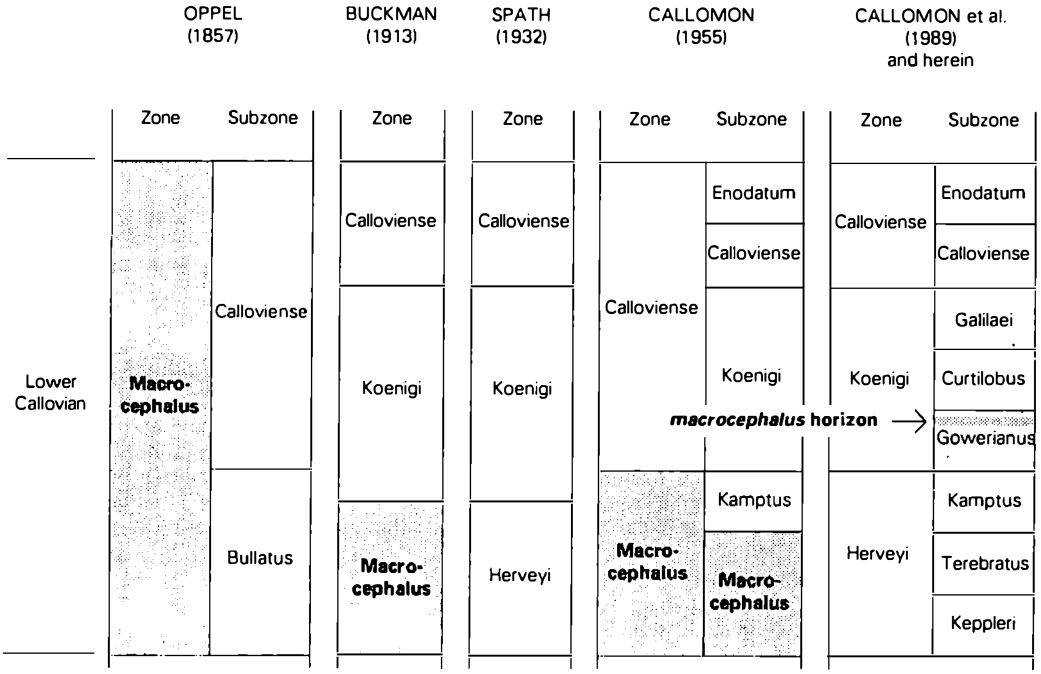

Figure 1. The evolution of the standard chronozonation of the Lower Callovian. (From Callomon et al., 1992).

The Base of the Callovian

The lowest Zone is the Herveyi Zone, whose basal Subzone is the Keppleri Subzone. What factors determine the definition of its base for geochronology, which is the fundamental purpose of chronostratigraphy? They have to be practical: to enable a stratigrapher to answer as precisely as possible (P2) the question: is this piece of rock already Callovian, or is it still Bathonian? The precision of the answer therefore depends on the minimum time-interval between two geological "events" that can be distinguished by the chronometer over as great a distance as possible. This determines the characteristic time-scale of the time-resolution of the chronometer. In biostratigraphical geochronology this depends on the characteristics of the guide-fossils used, which can differ widely. Ammonites are not the same as conodonts, or radiolarians, or calpionellids. Their characteristics have recently been exhaustively reviewed yet again (Callomon, 1995) and compared with those of some other widely-used micropalaeontological guide-fossils (Callomon, 1994a). Dominant among their characteristics is the discontinuity of their record. Most beds in a succession do not yield ammonites and when they do, the succession hardly, if ever, changes continuously. The record is quantized. The most closely-resolvable events in their biostratigraphy are represented by characteristic faunal horizons: faunally distinguishable beds, or series of beds, within which no further biochronological distinction (in the sense of morphological evolution) can be discerned. This is a wholly positivist, operational approach to stratigraphical time-measurement and has three important corollaries:

(a) The definition is biostratigraphical and independent of lithostratigraphy: thicknesses of beds do not enter into it. An ammonite faunal horizon can be 5cm thick (Inferior Oolite, Dorset) or 50m thick (Opalinus-Ton, Germany);

(b) The guide-fossils allow us only to identify time-intervals between the "events" represented by faunal horizons; and,

(c) they can tell us nothing about the durations of these "events": as far as can be determined by means of the fossils alone – the chronometer – the whole of a faunal horizon is effectively of the same age, equivalent to a single "instant", moment of time (Buckman: hemera).

The beds of a faunal horizon may well be "condensed", implying diverse processes within the duration of its formation (Callomon 1995, Tab. 3), but these must all be on shorter time-scales not resolvable by the chronometer. A faunal horizon may be condensed in the sense that it contains an ammonite assemblage at one locality that is resolvable into succession elsewhere (heterogeneous condensation) – an ever-present danger that can only be allayed by further work. Two faunal horizons at two different localities may appear to be identical as far as one can tell from their fossils - be "the same" - but may represent quite different durations of time. And lastly, a succession of faunal horizons is also rarely, if ever, complete. It will contain biostratigraphical gaps that can also be filled only by further work. But at any one time, a succession of faunal horizons represents the best that can be done. (And the point about ammonites is that, in periods and areas in which the study of the biostratigraphy is mature, such as the Callovian, this "best" is spectacular).

The Ammonite Faunal Horizons at the Bathonian–Callovian Boundary

The faunal horizons around the Bathonian–Callovian boundary in NW Europe are shown in Figure 2. The boundary between the standard Bathonian and Callovian Stages lies between the Discus and Herveyi Zones, where it always has been, and between the Discus and Keppleri Subzones of today. The historical criterion, P5(b), is therefore fully met. It remains to define the base of the Keppleri Subzone. This has two components: the choice of chronostratigraphical horizon; and the choice of a typological fixation of that horizon in a boundary stratotype - the Golden Spike.

When choosing the basal faunal horizon of the Callovian, attention should also be paid to the terminal faunal horizon of the underlying Bathonian, for it is the difference between them, and hence the time-interval between their formations, that contributes to the precision with which, and the geographical extent over which, the Bathonian–Callovian boundary can be recognized. The accumulated experience of the last 70 years points unambiguously to the choice of the *keppleri* horizon (above) and the *hochstetteri* horizon (below) between which to draw the boundary. They are by far the most widely recognizable horizons, thus satisfying the overriding criterion P5(a) above.

The ammonite faunas around the Bathonian–Callovian boundary have been listed (Callomon, Dietl and Niederhöfer, 1989). That of the *hochstetteri* horizon has been described in detail by Dietl (1994) and the description of that of the *keppleri* horizon is in progress (Dietl, in prep.). These descriptions are based on new collections of dozens or even hundreds of well-preserved and stratigraphically precisely located specimens made in southern Germany in the last 20 years – a wealth of material unmatched anywhere. The faunas consist of five principal elements, in order of decreasing abundance: (1) Perisphinctinae (*Homoeoplanulites, Parachoffatia*); (2) Macrocephalitinae (*Macrocephalites*); (3) Kosmoceratidae (*Kepplerites*); (4) Bullatimorphitinae (*Kheraiceras*); and (5) Clydoniceratidae (*Clydoniceras*).

Of these, the first are poorly time-diagnostic, although successive assemblages become distinguishable when material is well-preserved and plentiful. The second, the Macrocephalitinae, used to be regarded as an infallible guide to the base of the Callovian, marking it everywhere by its abrupt, albeit cryptogenic, First Appearance. This is now known to have been due only to a combination of stratigraphical gaps and collection-failure in the often only sparsely ammonitiferous beds below. *Macrocephalites* is now known to range upwards in Europe from the Orbis Zone of the

Upper Bathonian (Dietl, 1981), through the Lower and into the low Middle Callovian. In the Pacific, it occurs even earlier. It is in fact the second most common element in the *hochstetteri* horizon of Swabia. It also changes only slowly with time and the genus has turned out to be a relatively poor guide-fossil.

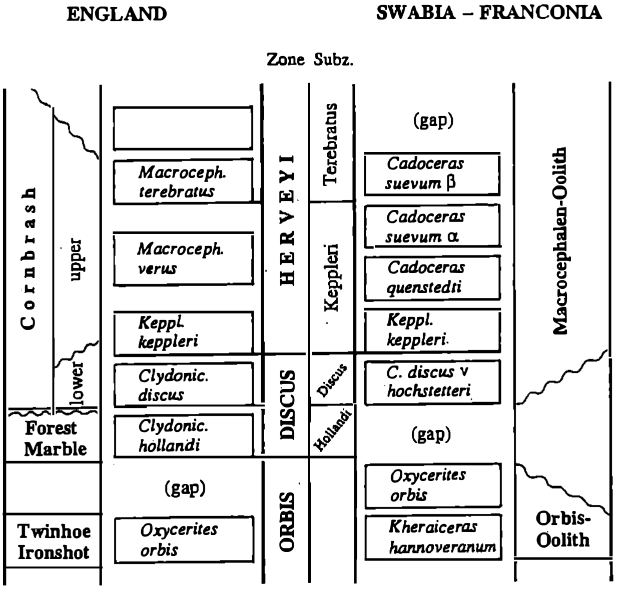

Figure 2. The ammonite faunal horizons around the Bathonian–Callovian boundary in England and Germany.

The species *Kepplerites keppleri*, in contrast, is excellent. It makes a sudden and brief appearance over much of NW Europe as a short-ranging immigrant before retreating again to its eudemic home in the Arctic. But we know also at what point in the evolutionary succession of the Kosmoceratidae as a whole it occurs there. The forms of *Kepplerites* are, like all the Kosmoceratidae, strongly ornamented so that very small evolutionary changes with time are discernible to the eye. Conversely, the transient chronospecies *K. keppleri* is restricted to a very narrow interval even in its eudemic centre of evolution. The time-range of the *keppleri* horizon in Europe must therefore also be very short. Moreover, in Europe *Kepplerites* disappears again above the *keppleri* horizon not to reappear until much later, in the basal *toricelli* horizon of the Koenigi Zone. It is therefore not easily misidentified. The *keppleri* horizon has now been recognized by means of its eponymous guide-fossil in much of Subboreal Europe: England (Dorset, Somerset, Oxfordshire); southern Germany; northern Swiss Jura (Dietl and Gygi, 1998); Caucasus. It also occurs further afield: East Greenland (Arctic Boreal), southern Alaska and the Queen Charlotte Islands (NE Pacific Province, Wrangellia); and even Japan. It therefore spans several biogeographical ammonite provinces at a level at which correlation by means of other groups would be very tenuous.

The Bullatimorphitinae are of little help, those of the latest Bathonian and earliest Callovian being barely distinguishable. They are moreover found only in the more southerly latitudes of the Tethyan Realm. In Europe this means predominantly the shelf-sea sediments of the Submediterranean Province: they have never been found in Britain. Conversely, however, they provide a correlative connection with other faunal provinces world-wide, e.g., with Madagascar and India.

Clydoniceras is valuable because it appears to make its Last Appearance in the Discus Subzone. In Britain, where its stratigraphical distribution has long been best known, its strictly non-overlapping range below *Macrocephalites* was largely the basis of the earliest attempts to draw a systematic boundary between the Bathonian and Callovian Stages (Douglas and Arkell, 1928; 1932; summaries in Arkell, 1933; 1947; 1951). The genus has a very wide distribution in Europe, with excursions as far as the easternmost-preserved northern margin of the Tethys in Tadzhikistan (Krymholz and Zhakarov, 1971); but the records are scattered and material is never abundant. The morphology is also rather nondescript and changes only slowly with time. With some exceptions, therefore, its value as a guide-fossil is rarely greater than as indicator of a definitely pre-Callovian age. *C. discus* itself has some range in the Lower Cornbrash in England, but Arkell was unable to resolve it further. Chronostratigraphically, therefore, the Discus Subzone of England has still to be treated as a single faunal horizon (see below). It is not possible to determine whether *C. hochstetteri* (Oppel, 1857) is a transient of *C. discus* (Sowerby, 1821) in the Discus Zone, or merely a variant. In order not to prejudge the issue, the name *hochstetteri* has been retained for the Swabian faunal horizon.

An additional group of value as guide-fossils is that of the genus *Cadoceras*. But it appears in quantity only a little higher, in the *quenstedti* horizon (Fig. 2). Yet slightly higher still, the *suevicum* horizon provides good correlation with the lowest Callovian ammonite horizon on the Russian Platform, that of *Cadoceras elatmae*.

In summary, by far the most widely biostratigraphically recognizable level at which to draw the Bathonian–Callovian boundary lies between the *hochstetteri* and *keppleri* horizons.

The Choice of Boundary Stratotype

The *hochstetteri* horizon has been most clearly identified at Albstadt-Pfeffingen (Fig. 3a) and Blumberg (Fig. 3b), where it is in each case proven by *Clydoniceras*. One of Oppel's syntypes of *C. hochstetteri* came from Lochenbach, between Balingen and Pfeffingen, as did another specimen (Hahn, 1971); another from Albstadt-Lautlingen (Dietl, 1982); and it has been recorded from Müllheim, between Freiburg and Basel (*id.*). The richest faunas have come from Blumberg and nearby Epfenhofen (Dietl, 1994), made possible by quite exceptionally extensive temporary exposures. Earlier records from the northern Swiss Jura are reviewed by Arkell (1956). The horizon is progressively cut out eastwards by the overlying *keppleri* horizon, which comes to rest on the Orbis Zone already at the Ipf and in Franconia at Sengenthal.

The *keppleri* horizon has been clearly identified at Sengenthal (Callomon et al., 1987, bed C1); the type of *K. keppleri* came from Eningen, near Reutlingen; rich collections have come from the area around Albstadt-Pfeffingen itself and in the Klingenbachtal west of Albstadt-Onstmettingen (Dietl, 1981); at Esslingen, NE of Geisingen (Dietl and Herold, 1986); and at Liesberg (Dietl and Gygi, 1998) - 350km from Sengenthal. It is probably present at the Ipf but not so far proven because of collection-failure. It is definitely missing at Blumberg.

The next higher horizon, of *Cadoceras quenstedti*, can be followed almost uninterruptedly from the Swiss border to the Ipf, including the area around Pfeffingen and Hausen. It has yielded particularly abundant collections at Anwil, SE of Basel (Gygi coll., Basel and Rieber coll., Zürich). (Thierry's claim - 1978, p.339 - that it is accompanied there by *Gowericeras gowerianus* and *Sigaloceras calloviense* is bizarre).

The only area in which both the *hochstetteri* and *keppleri* horizons are well-developed and in contact is that around Albstadt. The choice of this area for the selection of a basal boundary stratotype section of the Callovian Stage seems, therefore, almost automatic.

The Boundary Stratotype Section

There are no natural exposures of the Macrocephalen-Oolith in southern Germany. Even if there were, they would not last long at the hands of the numerous fossil-collectors in search of specimens from a formation famous for the richness of its ammonites. A section was therefore excavated by the staff of the Stuttgart Museum at a site in a Nature Reserve in which geological collecting is generally not permitted. It lies in the upper reaches of the Roschbachtal, 1km west of the centre of Pfeffingen. The section is shown in Figure 3a.

The amount of material that could be obtained from such a small exposure is limited – a limitation that probably applies to most boundary stratotypes. But the section is essentially identical with that of the temporary exposures for deep drainage in the centre of Pfeffingen (Dietl, 1990) from which several thousand ammonites could be collected *in situ*. These are now in Stuttgart and available for study, together with a plentiful supply of matrix for students of micropalaeontology, lithology and geochemistry. The section itself is normally covered again for protection but can be easily reopened when required. It was reopened recently to be sampled for geomagnetic measurements, and a succession of belemnites has been analyzed for strontium isotope geochemistry.

A model of the section made with rock excavated from it has been reconstructed in the "Museum am Kräuterkasten" in Albstadt-Ebingen. It is a branch-museum of the Staatliches Museum für Naturkunde in Stuttgart and is the local museum of Albstadt. The precise level of the Bathonian–Callovian boundary is marked in the section by a suitable Golden Spike and its significance explained in the accompanying text.

Discussion

Conformity with the Requirements of the ICS

To what extent do the facts presented above in support of the case for the proposed GSSP meet the requirements of the ICS as laid out in the Revised Guidelines?

Sections 2.3: correlation and 2.4: priority. Both are fully met and require no further comment.

Section 4: "The stratotype-section should contain the best possible record of the relevant marker events". The record of the relevant marker events [*sic*] is embodied in the *hochstetteri/keppleri* faunal horizons. "Contain" is ambiguous. "Illustrate", yes. "Represent": any boundary that depends on a single section for the "best possible record" is merely in need of further work.

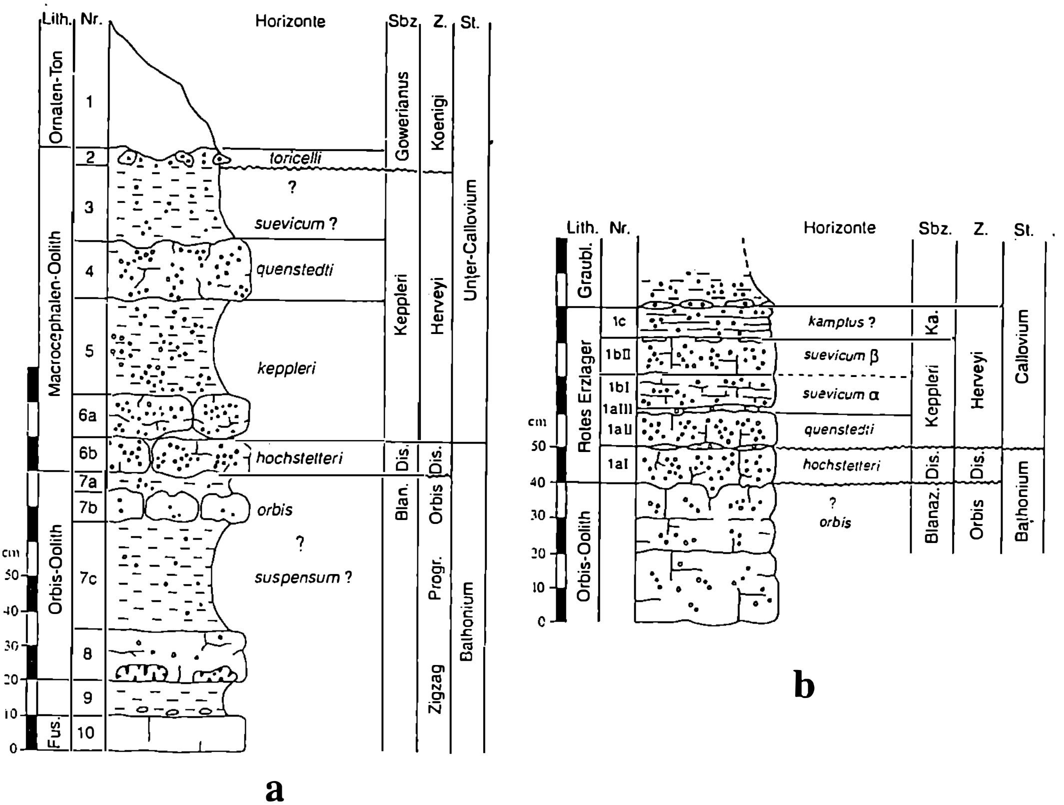

Figure 3. Schematic representation in weathering profile of two sections across the Bathonian–Callovian boundary. 3a: the proposed stratotype section at Albstadt-Pfeffingen. 3b: the important section at Blumberg, southernmost Swabian Alb. (After Dietl, 1994).

4.1: Geological requirements. "Adequate thickness of exposures" in the present case means adequate to characterize the faunal horizons (which are beds) that represent the record in the rocks of the "relevant marker events". These horizons automatically span a sufficient time-interval. We cannot understand what might be meant by reference to the boundary being "also ... determined by interpolation, using auxiliary markers close to the boundary". The faunal horizons are the markers at the boundary.

"Continuous sedimentation: no gaps, no condensation in proximity of the boundary level." This is more serious and at the heart of an objection to the present proposals, as put forward in 1990, by Dr. Rainer Jordan (not a member of the Callovian Working-Group). They were submitted to the then Chairman of the ISJS and to the Chairman of ICS in manuscript form, in German, in 1991. They made no reference to the detailed arguments presented in the document prepared by us (C and D, 1990) in support and explanation of the present case, copies of which were readily available. It is clear from the objections that either Dr. Jordan had not seen the document. or had chosen to ignore it, or just did not understand it. He certainly did not understand the concept of faunal horizon. A rebuttal was therefore prepared in 1992 and circulated. An English translation of the objection was published in Newsletter No. 21 of the ISJS in 1993; the rebuttal was not. The only subsequent reference to this debate so far was in Newsletter No. 24 (1997, p.7), in a brief retrospective review of the activities of the ISJS over the previous seven years, to the effect that "1.2.3. Callovian GSSP ... does not satisfy the requirements defined in the Guidelines of the ICS. The procedure must be reopened." No explanations were given as to why and how.

The crux of Dr. Jordan's misunderstanding is two-fold and simple: there is no such thing as "continuous sedimentation"; and a call for "no condensation" is empty hand-waving without quite detailed explanations of what is meant by "condensation".

Sedimentation can only appear to have been continuous and what matters is: how big must the gaps be before we detect them as sedimentary discontinuities, before sedimentation appears to have been not continuous? Hence, how complete is the sedimentary record, its geochronology? This non-trivial problem has received considerable attention by serious stratigraphers (Schindel, 1980; 1982; Sadler and Strauss, 1990). The answer depends wholly on the time-scale on which the completeness is being calculated. The kind of time-scales on which the Guidelines call for "[apparently] continuous sedimentation: no gaps ..." are those of sedimentary lithochronology: a week (cross-bedding)? – a year (varves, laminated shales)? – a millennium (bioclastic calcarenites)? They are irrelevant to the time-scale of the time-intervals we have been discussing (see above!). The time-scales characteristic of ammonite evolution as resolved in faunal horizons have been recently discussed again by one of us (Callomon, 1995), with some graphic examples going back over a century (Buckman). They are typically of the order of 50,000 - 100,000 years. That is the precision with which a correlation by means of ammonites over greater than purely local distances can be achieved at best, or a time-plane be recognized. One does not buy a chronometer with millisecond time-resolution to check a phone-bill charged by the minute, or even a watch with a second-hand to time a hired car rented out by the day.

Similar arguments apply to "condensation". If a faunal horizon proposed as the basis for a boundary stratotype at one locality is condensed in the sense that it contains mixed faunal assemblages that can be stratigraphically resolved "elsewhere", then "elsewhere" is where the stratotype would be chosen instead. In this sense, the *hochstetteri* and *keppleri* horizons of Swabia-Franconia are not condensed. Their beds are, however, bioturbated, so that on the time-scale of their last bioturbational turnover - a year?, a decade? - they are condensed, even if the time taken to accumulate their sediments - the duration of their formation - was much longer. But this time-scale would also again be irrelevant. (The only sediments not bioturbated or wave-mixed are laminated shales or sandstones. How many GSSPs have been defined in such facies?).

So, gaps in the record there certainly are. It remains only to ask whether that between the *hochstetteri* and *keppleri* horizons is minimal. That can only be answered by regional biostratigraphy and correlation. Where, "elsewhere", have additional distinguishable faunal horizons been found that are identifiably of ages intermediate between those of the *hochstetteri* and *keppleri* horizons? The answer is, that after 140 years of intensive work, nowhere. And the close similarity of the faunas of these horizons suggests that the future chances are small.

In summary, the ammonite faunal horizons at the Bathonian–Callovian boundary stratotype are satisfactory from a sedimentological point of view. Regional evidence shows that at the stratotype, "condensation" is not a problem and that the time-interval resolved across the boundary is minimal.

The definition of the Bathonian–Callovian boundary being proposed is the best that can be done and likely to remain so for the foreseeable future.

4.2: Biostratigraphical requirements. There are no problems here with ammonites. And yes: "An obvious boundary [lithofacial change] should be suspect." But as explained above, so should the absence of obvious boundaries.

Other Methods

(1) Strontium isotope stratigraphy: The stratotype has been successfully sampled by means of belemnites. The analyses were performed by Dr. J. M. McArthur (University College London) at the Radiogenic Isotope Laboratory, Royal Holloway College (University of London), by courtesy of its Director, Dr. M. Thirlwall. The results will be fully published elsewhere, but Dr. McArthur has kindly permitted us to quote them here (Fig. 4). The curve of the ratio of marine Sr(87):Sr(86) changes only slowly in the region of the Bathonian–Callovian boundary. It has been well established below and above and the new determinations agree with the values obtained from the statistically "best" curve (Howarth et al., 1997) almost exactly. While the strontium isotope-stratigraphy therefore contributes little to the definition of the Bathonian–Callovian boundary, its potential value in identifying the boundary in very distant correlations, e.g., in the SW Pacific, New Zealand, the Andes, Mexico and even the Western Interior of the US, can hardly be overestimated.

(2) Magnetostratigraphy: The stratotype was sampled in collaboration with Dr. J. G. Ogg (Purdue University) and the results will also be described more fully elsewhere. Thirty cores in the interval *orbis–toricelli* could be successfully analyzed but the results were not unambiguous. The *toricelli* horizon appears to be N, but the *quenstedti, keppleri* and *hochstetteri* horizons gave results that were mixed, partly N and partly R. The Orbis Zone appears to be predominantly N. The rapid oscillations may be in part genuine, but the question arises as to what extent the signals were affected by bioturbation. The rocks are all ferruginous, fine-grained mudrocks with ironshot ooliths scattered more or less densely in clouds. Magnetostratigraphy is probably a technique that does require more expanded successions, and in the Bathonian–Callovian such successions, precisely correlatable with the stratotype, should be available.

(3) Palynology: The succession at the stratotype has yielded a rich and diverse flora of dinoflagellate cysts. It was described in the Poitiers volume (Smelror and Dietl, 1994), where more details can be found. The assemblages accord with those elsewhere and nothing of special note occurs at the Stage boundary.

Alternatives and Comparisons

Bathonian–Callovian successions have been studied in sections too numerous to list and correlations have been limited only by the quality and quantity of the fossils available. In northern Europe, this means normally down to Zonal or Subzonal level. Further afield, ammonite correlations can become less direct because of their bioprovincialism and chains of correlation have to be used. The subject was reviewed at some length at Erlangen (Callomon, 1985a) and Lisbon (Callomon, 1989) and could now be extended. But for reasons of space the discussion will have to be continued elsewhere.

We will, however, in the present context end with a brief comparison of the proposed Bathonian– Callovian GSSP with the GSSP of a Jurassic Stage that has already been ratified: that of the Aalenian–Bajocian boundary now fixed at Cap Mondego, Portugal (Pavia and Enay, 1997). After long debates it was defined on the basis of a group of ammonites at a point at which the genus *Graphoceras* changes to *Hyperlioceras*. This has long been taken everywhere in Europe to mark the boundary between the Concavum Zone (highest of the Aalenian) and the Discus Zone (lowest of the Bajocian), hence also automatically the boundary between the Aalenian and Bajocian Stages. And as in other cases, this boundary could be everywhere recognized with sufficient precision without the need for a more precise definition. The question now arose: where, exactly, is the generic transition-point? At first it was taken at the lowest level at Cap Mondego alleged to have yielded a *Hyperlioceras*, a "First Appearance Event". But the biostratigraphy of the ammonites at Cap Mondego is limited in detail, both qualitatively and quantitatively. Much richer and better-preserved collections elsewhere then showed that the morphogenera *Graphoceras* and *Hyperlioceras* in fact overlap, but that highly detailed and widely correlatable chronostratigraphy could be done in terms of successive assemblages – and their faunal horizons (Callomon and Chandler, 1990). These faunal horizons could in fact be recognized even at Cap Mondego, and the stratotype boundary was finally placed there so as to coincide with that between horizons Aa-16 and Bj-1 in Dorset!

Conclusions

Since its first definition in 1964, by means of a basal boundary stratotype in England of its lowest standard Subzone, the Macrocephalus Subzone, the accepted position of the boundary between the Callovian and Bathonian Stages has changed little. What has changed is the attainable finesse of the chronostratigraphic subdivision of these Stages around their boundary. The time-resolution achievable is that of the time-intervals between the moments represented by faunal horizons (as nowadays formally defined) of the chronometric guide-fossils, the ammonites. Conversely, this determines the chronometric precision with which a time-plane lying between two faunal horizons can be specified. The choice of such a time-plane defining the base of the Callovian Stage has to be governed above all by the ability to recognise it geographically as far as possible. This consideration picks out the horizon of *Kepplerites keppleri* above all others, to be chosen as the basal ammonite faunal horizon of the Callovian Stage. The choice is based on a combination of over a century's work of correlation over a large part of the world, and very large new collections of the chronometric ammonites from precisely known levels, made in the Swabian Alb.

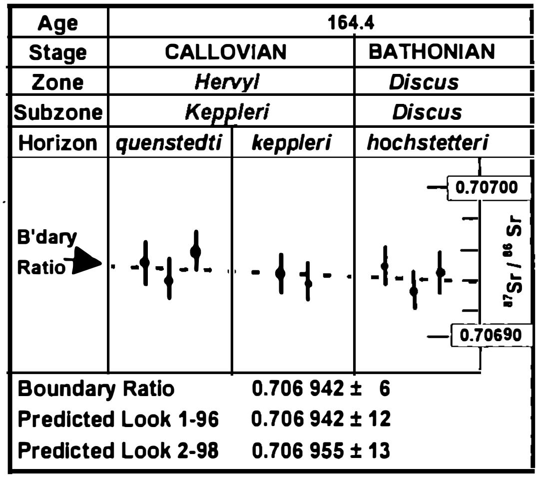

Figure 4. Measured values of the strontium isotope-ratio in belemnites around the Bathonian–Callovian boundary. (Courtesy of Dr. J. M. McArthur, London).

The choice of a stratotype section, therefore, should be guided above all by a search for a section that best demonstrates the succession of these ammonites. No single section could be expected to best characterize the chosen faunal horizons, for these were identified only after correlations of many sections. Such a representative section has been chosen and described at Albstadt-Pfeffingen. It so happens that during Bathonian–Callovian times the region lay in a somewhat sediment-starved basin. Sediments are thin, but on the time-scale of the resolution of ammonite faunal horizons, remarkably persistent and uniform. The stratotype section is therefore thin – "condensed" – but this is of little consequence for the purpose it is meant to achieve. To some degree, this is merely monumental. We confidently predict, however, that ratification of the present proposals will provide a secure and satisfactory definition of the Bathonian–Callovian boundary for the foreseeable future.

References

AGER, D. V. 1963. Jurassic stages. Nature, 198:1045-1046.

ARKELL, W. J. 1933. The Jurassic System in Great Britain. Oxford University Press, Oxford, 681 p.

____. 1946. Standard of the European Jurassic. Bulletin of the Geological Society of America, 57:1-34.

____. 1947. The Geology of Oxford. Oxford University Press, Oxford, 267 p.

____. 1951. Monograph of the English Bathonian Ammonites. Palaeontographical Society, London, part i, 1-46.

____. 1956. Jurassic Geology of the World. Oliver and Boyd, Edinburgh, 821 p.

CALLOMON, J. H. 1955. The ammonite succession in the Lower Oxford Clay and Kellaways Beds at Kidlington, Oxfordshire, and the Zones of the Callovian Stage. Philosophical Transactions of the Royal Society, London, B239:215-264.

____. 1964. Notes on the Callovian and Oxfordian Stages, p. 269-291. *In* P. L. Mauberge (ed.). Colloque du Jurassique à Luxembourg. Publications de l'Institut grand-ducal, Séction Sciences naturelles physiques, mathematiques, Luxembourg. 948 p.

____. 1965. Notes on Jurassic stratigraphical nomenclature. I. Principles of stratigraphical nomenclature. Reports of the VIIth Congress of the Carpatho-Balkan Geological Association, Sofia, part ii, 1:81-85.

____. 1985a. Report of the Callovian Working-Group, p. 77-86. *In* O. Michelsen and A. Zeiss (eds), International Symposium on Jurassic Stratigraphy, Erlangen, 1. Geological Survey of Denmark, Copenhagen.

____. 1985b. Biostratigraphy, chronostratigraphy and all that – again, p. 611-624. *In* O. Michelsen and A. Zeiss (eds), International Symposium on Jurassic Stratigraphy, Erlangen, 3. Geological Survey of Denmark, Copenhagen.

____. 1989. Report of the Stage Working-Groups of the International Subcommission on Jurassic Stratigraphy: Middle Jurassic, Callovian Stage, p. 19-26. *In* R.B. Rocha and A. Zeiss (eds), 2nd International Symposium on Jurassic Stratigraphy, Lisboa 1987, 1. I.N.I.C., Lisboa.

____. 1994a. Palaeontological methods of stratigraphy and biochronology: some introductory remarks. Geobios, Mémoire spécial, 17:16-30.

____. 1994b. Reports of the Stage Working-Groups: Callovian. Geobios, Mémoire spécial, 17:757-760.

____. 1995. Time from fossils: S.S. Buckman and Jurassic high-resolution geochronology, p. 127-150. *In* M. Le Bas (ed.), Milestones in Geology. Geological Society, London, Memoir 16.

____, AND R. B. CHANDLER. 1990. A review of the ammonite horizons of the Aalenian – Lower Bajocian Stages in the Middle Jurassic of southern England. Memorie descrittive della Carta Geologica d'Italia, 40:85-111.

____, AND G. DIETL. 1990. Proposed definition of the Callovian Stage. Field Symposium in Swabia, Stuttgart/Albstadt (16-21 September, 1990) ISJS, Callovian Working Group, MS (unpublished), xi + 20 p.

____, ____, A. GALACZ, H. GRADL, H.-J. NIEDERHÖFER, AND A. ZEISS. 1987. Zur Stratigraphie des Mittel- und unteren Oberjuras in Sengenthal bei Neumarkt/Ofr. (Fränkische Alb). Stuttgarter Beiträge zur Naturkunde, B132:1-53.

____, ____, AND H.-J. NIEDERHÖFER. 1989. Die Ammonitenfaunen-Horizonte im Grenzbereich Bathonium/Callovium des Schwäbischen Juras und deren Korrelation mit W-Frankreich und England. Stuttgarter Beiträge zur Naturkunde, B148:1-13.

____, ____, ____. 1992. On the true stratigraphic position of *Macrocephalites macrocephalus* (Schlotheim, 1813) and the nomenclature of the standard Middle Jurassic "Macrocephalus Zone". Stuttgarter Beiträge zur Naturkunde, B185:1-65.

____, ____, AND K. N. PAGE. 1989. On the ammonite faunal horizons and standard zonation of the Lower Callovian Stage in Europe, p. 359-376. *In* R. B. Rocha and A. Zeiss (eds), 2nd International Symposium on Jurassic Stratigraphy, Lisboa 1987, 1. Universidade Nova de Lisboa.

COWIE, J. W., W. ZIEGLER, A. J. BOUCOT, M. G. BASSETT, AND J. REMANE. 1986. Guidelines and Statutes of the International Commission on Stratigraphy (ICS). Courier des Forschungs-Institut Senckenberg, 83:1-14.

DIETL, G. 1981. Über *Macrocephalites* (Ammonoidea) aus dem Aspidoides-Oolith und die Bathonium/Callovium-Grenzschichten der Zollernalb (SW-Deutschland). Stuttgarter Beiträge zur Naturkunde, B68:1-15.

___. 1982. Das wirkliche Fundniveau von *Ammonites aspidoides* OPPEL (Ammonoidea, Mittl. Jura) aus dem locus typicus. Stuttgarter Beiträge zur Naturkunde, B87:1-21.

___. 1990. *Procerites progracilis* Cox and Arkell und andere Ammoniten aus dem basalen Mittel-Bathonium (Mittl. Jura) der Zollernalb, Schwäb. Alb, SW-Deutschland. Jahresberichte und Mitteilungen des oberrheinischen geologischen Vereins, N.F. 72:329-340.

___. 1994. Der *hochstetteri*-Horizont - ein Ammonitenfaunen-Horizont (Discus Zone, Ober-Bathonian, Dogger) aus dem Schwäbischen Jura. Stuttgarter Beiträge zur Naturkunde, B202:1-39.

___, AND R. GYGI. 1998. Die Basis des Callovian (Mittlerer Jura) bei Liesberg BL, Nord-Schweiz. Eclogae geologicae Helvetiae, 91 (in press).

___, AND G. HEROLD. 1986. Erstfund von *Cadomites* (Ammonoidea) im Unter-Callovium (Mittl. Jura) von Südwest-Deutschland. Stuttgarter Beiträge zur Naturkunde, B120:1-9.

DOUGLAS, J. A. AND W. J. ARKELL. 1928. The stratigraphical distribution of the Cornbrash. I. The south-western area. Quarterly Journal of the Geological Society, London, 84:117-178.

___, ___. 1932. The stratigraphical distribution of the Cornbrash. II. The south-western area. Quarterly Journal of the Geological Society, London, 88:112-170.

GUEX, J. 1991. Biochronological correlations. Springer Verlag, 252 p.

HAHN, W. 1971. Die Tulitidae S. Buckman (Ammonoidea) des Bathoniums (Brauner Jura ε) im südwestdeutschen Jura. Jahreshefte des geologischen Landesamt Baden-Württemberg, 13:55-122.

HEDBERG, H. D. 1954. Procedure and terminology in stratigraphic classification. Congrès Géologique International: Comptes rendus de la XIX Session, Alger 1952. Séction XIII, 1ère partie: Fasc. xiii, 205-233.

HOWARTH, R. J., AND J. M. McARTHUR. 1997. Statistics for strontium isotope stratigraphy: a robust LOWESS Fit to the marine Sr-isotope curve for 0 – 206 Ma, with look-up table for derivation of numeric age. Journal of Geology, 105:441-456.

KRYMHOLZ, G. Ya., AND E. F. ZAKHAROV. 1971. Batskie ammonity Kugitanga, p. 4-40. *In* T. Sh. Shayakuboe (ed.), Paleontologicheskoe obosnovanie opornyk razresov Yurskoi sistemy Uzbekistana i sopredelen'ykh raionov. KGSPE, Tashkent, Sbornik 10. (In Russian).

MAUBEUGE, P. L. (ed.). 1964. Colloque du Jurassique à Luxembourg. Publications de l'Institut grand-ducal, Séction Sciences naturelles physiques, mathematiques, Luxembourg, 948 p.

OPPEL, A. 1857. Die Juraformation Englands, Frankreichs und des südwestlichen Deutschlands. Ebert and Seubert, Stuttgart, 439-694.

PAGE, K. N. 1989. A stratigraphical revision for the English Lower Callovian. Proceedings of the Geologists' Association, 100:363-382.

PAVIA, G., and R. ENAY. 1997. Definition of the Aalenian-Bajocian Stage. Episodes, 20:16-22.

REMANE, J. 1996. The revised Guidelines of ICS and their bearing on Jurassic chronostratigraphy. GeoResearch Forum, Transtec Publications, Switzerland, 1-2:19-22.

___, M. G. BASSETT, J. W. COWIE, K. H. GOHRBANDT, H. R. LANE, O. MICHELSEN, AND WANG NAIWEN. 1996. Revised guidelines for the establishment of global chronostratigraphic standards by the International Commission on Stratigraphy (ICS). Episodes, 19:77-81.

SADLER, P. M., AND D. J. STRAUSS. 1990. Estimation of completeness of stratigraphical sections using empirical data and theoretical models. Journal of the Geological Society, London, 147:471-485.

SCHINDEL, D. E. 1980. Microstratigraphic sampling and limits of paleontologic resolution. Paleobiology, 6:407-426.

___. 1982. Resolution analysis: a new approach to the gap in the fossil record. Paleobiology, 8:340-353.

SMELROR, M., AND G. DIETL. 1994. Dinoflagellate cysts of the Bathonian/Callovian boundary beds in southern Germany. Geobios, Mémoire spécial, 17:453-459.

SPATH, L. F. 1932. The invertebrate faunas of the Bathonian-Callovian deposits of Jameson Land (East Greenland). Meddelelser om Grønland, 87, no.7, 158 p.

THIERRY, J. 1978. Le genre *Macrocephalites* au Callovien inférieur (ammonites, Jurassique moyen). Mémoires géologiques de l'Université de Dijon, 4. 492 p.

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Oxfordian Biostratigraphy from the Lugar Section (External Subbetic, Southern Spain)

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Keywords: Biostratigraphy, Ammonites, Oxfordian, Betic Cordillera, Eastern Subbetic, Southern Spain

Abstract: The section analyzed in the eastern part of Sierra de Lugar (Prov. Murcia) is one of the better outcrops showing Oxfordian deposits with ammonites in epioceanic ammonitico rosso facies in the Subbetic Zone of the Betic Cordillera (southern Spain). The Lower Oxfordian Renggeri Zone, Middle Oxfordian upper Plicatilis (Antecedens Subzone) and Riazi (= Transversarium) Zones, Upper Oxfordian Bifurcatus and Bimammatum Zones, and uppermost Oxfordian or lowermost Kimmeridgian Planula Zone have been identified. Subzonal divisions are recognized in the lower Bifurcatus Zone (Stenocycloides Subzone), Bimammatum Zone (Hypselum and Bimammatum Subzones), and Planula Zone (Planula and Galar Subzones). Intra-Subzone subdivision is envisaged for the Hypselum Subzone (*Fontannesi-Semicostatum* horizon, lower, and *Berrense* horizon, upper). As usually recognized in ammonitico rosso, condensation and/or hiatuses, together with assumed differences in morphologic evolution between epioceanic and epicontinental ammonites, hamper precise correlation of ammonite ranges on the basis of present data.

Introduction

In the Mediterranean Tethys, Oxfordian deposits with ammonites are preserved mainly as more or less condensed, red, nodular limestones showing a wide range between marly and calcareous ammonitico rosso facies. Time-averaging and discontinuous deposition characterize the epioceanic ammonitico rosso, making ammonite biostratigraphy comparatively less accurate (Sequeiros, 1974; Fülop, 1976; Sapunov, 1979; Sarti, 1988; Fözy, 1993a; 1993b) than with more continuous sedimentation corresponding to deposition on epicontinental areas (Meléndez, 1989; Cariou et al., 1991; Schweigert, 1995, among others). In southern and southeastern Spain, outcrops in the Betic Cordillera and Mallorca Island (Balearic Archipelago), respectively, are among the most favourable in Europe for developing precise ammonite biochronostratigraphy in Oxfordian epioceanic deposits (Barthel et al., 1966; Behmel, 1970; Sequeiros, 1974; Checa and Sequeiros, 1990; Olóriz et al., 1998).

Red, nodular, condensed wackestones, known as the ammonitico rosso and related facies, dominated sedimentation during the Late Jurassic in epioceanic swells of the Betic Cordillera (Subbetic Zone or domain) in southern Spain (Fig. 1). Southern swells in the central sector of the Internal Subbetic show Oxfordian deposits characterized by condensed and cephalopod-rich calcareous ammonitico rosso (2-4 m thick). In northern swells of the same sector, the External Subbetic shows marly and cephalopod-poor ammonitico rosso, locally, with a comparatively reduced thickness (less than 2.5 m). The eastern External Subbetic shows comparatively proximal (landward) swells with thicker Oxfordian sections in marly but ammonite-rich ammonitico rosso facies. In the Province of Murcia, the Lugar section in the Sierra de Lugar has been analyzed on the basis of a continuous sampling of 10-20 cm thick samples, and can be proposed as a reference section for the Oxfordian in the External Subbetic.

The Lugar Section

The Oxfordian section investigated at the eastern part of Sierra de Lugar (UTM 30SXH592309 in the topographic sheet of Fortuna: 892) can be reached from the village of Fortuna by the road between Sierra de Lugar and Sierra de La Pila. The Oxfordian at the Lugar section is 9.65 m thick and consists of varying marly/calcareous ammonitico rosso (Figs. 1, 2). Bedding appearance in the outcrop is mainly diagenetic in origin, but minor discontinuities can be identified related to narrow horizons: bioclastic, marly, or intensively burrowed. The Oxfordian succession overlies a ferruginized hardground with Mn nodules ("snuff-boxes") and abundant reworked ammonites and belemnites (Upper Callovian). Overlying the Oxfordian, the identification of the Kimmeridgian-Tithonian section proved possible through ammonite biostratigraphy.

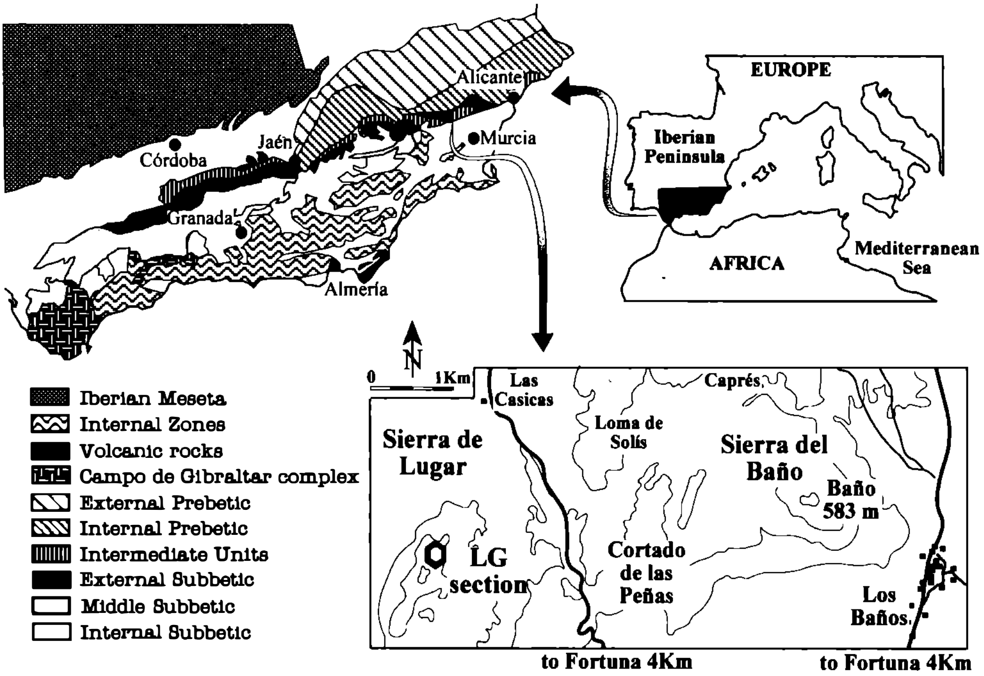


Figure 1. Location of the section studied.

The Oxfordian section begins with well bedded and nodular horizons 45 cm thick (horizons 0-30 and 30-45), which show *Thalassinoides* burrowing, a comparatively enriched benthos, and ferruginized macrofossil remains. Following are 80 cm (horizon 55-105) of comparatively marly and less fossiliferous ammonitico rosso. These lowermost 1.2m are wackestones/packstones rich in *Protoglobigerina* and thin shelled bivalves ("filaments"), together with secondary crinoids (*Saccocoma*), radiolaria, *Globochaetes*, *Stomiosphaera* and indeterminate foraminifera. Cephalopods (mainly ammonites) are the dominant macrofossils; bivalves and occasional gastropods were also recorded.

Overlying are two decimeter-thick beds of calcareous ammonitico rosso (horizons 115-130 and 130-145) with conspicuous slumping, scarce and poorly preserved ammonites, and intercalated horizons with sponges in life position. Above is a 2m thick interval of more or less marly ammonitico rosso rich in fossil remains (mainly ammonites), and then calcareous ammonitico rosso (horizons 0-10b and 15-30b) similar to that at the base of the marly interval. On the whole, microfacies from the slumped beds to the upper calcareous ammonitico rosso horizons are dominated by wackestones with *Protoglobigerina* and *Globochaetes*.

Higher in the section are 2.1m of marly ammonitico rosso (horizons 80-90b to 250-255 to 265-285b) showing upwardly increasing bed thickness in more calcareous subfacies. Microfacies are similar to, but slightly more diversified than, those in lower horizons. Ammonites are more frequent than previously. In the upper part of this interval reworked ammonites are common and the larger ones (Perisphinctidae more than 35cm in size) show eroded upper sides.

Marly ammonitico rosso comprises the upper 1.45m (horizons 325-345b to 605-615). Microfacies are mudstones and wackestones comparatively enriched in *Globochaetes* but impoverished in other bioclasts. The more calcareous horizons show intensive burrowing by *Chondrites* and *Planolites*. Fragmentation and oblique orientation of macrofossils is common.

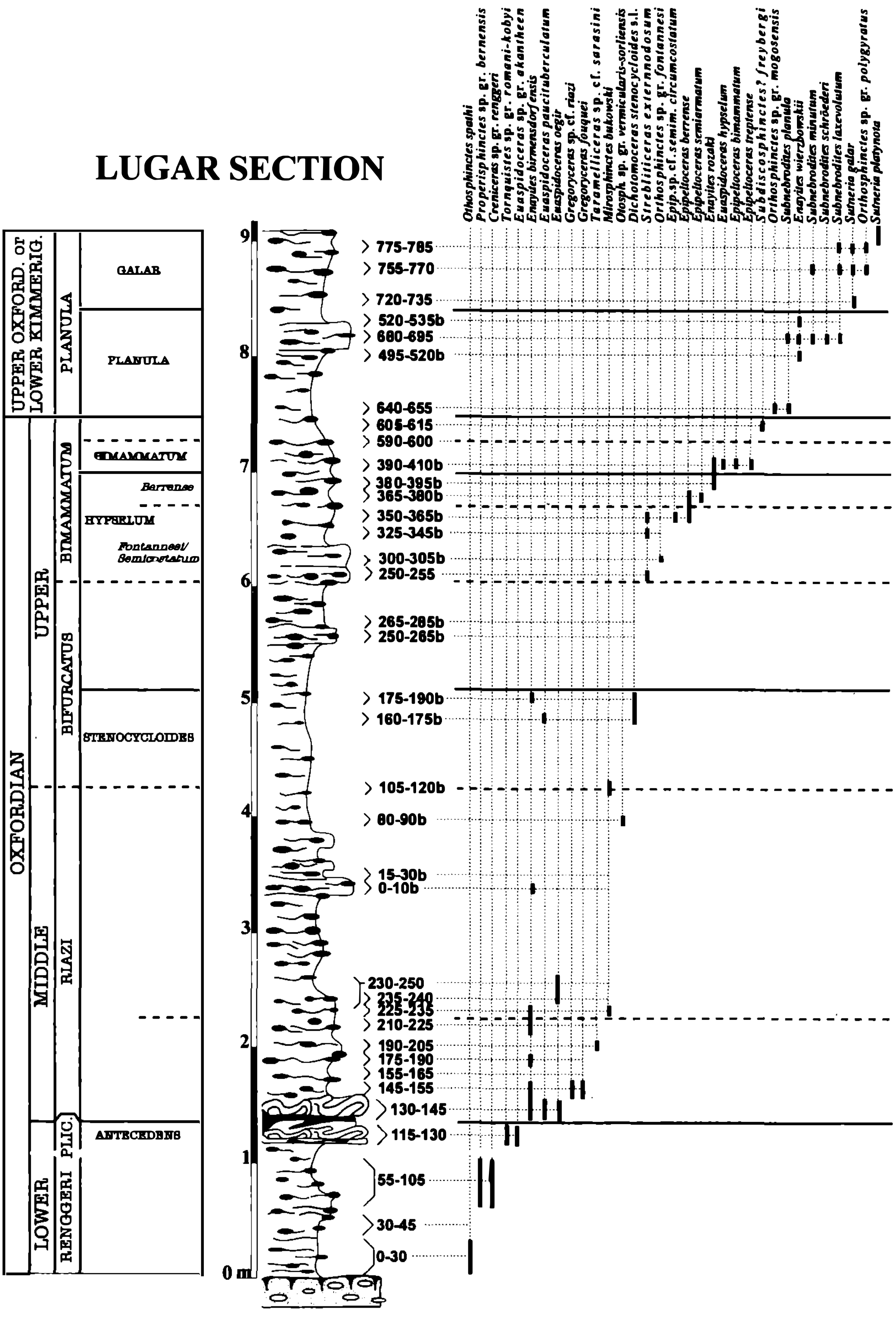
The Oxfordian section ends with 2.1m of marly ammonitico rosso (horizons 640-655 to 777-785) similar to the underlying interval, but showing three decimetre-thick intercalations of calcareous ammonitico rosso, which were intensively burrowed by *Thalassinoides*. With the exception of the composition in the ammonite assemblages recovered, no other significant difference was recognized in microfacies and preservation of macrofossils.

Ammonite Assemblages

A total of slightly more than 2,000 fossils of macro- and mega-invertebrates was collected from continuous samples, each 10-20 cm thick. The preservation of fossils is moderately good and incomplete specimens and fragments overwhelmingly dominate this material, more than 90% of which belongs to ammonites. The 25 ammonite genera identified are distributed at the family/subfamily level between Phylloceratidae (Phylloceratinae, Calliphylloceratinae), Lytoceratidae (Lytoceratinae), Haploceratidae (Haploceratinae, Taramelliceratinae), Glochiceratidae (Glochiceratinae), Oppeliidae (Ochetoceratiane), Pachyceratidae, Aspidoceratidae (Peltoceratinae, Euaspidoceratinae, Epipeltoceratinae, Aspidoceratinae, Physodoceratinae), Perisphinctidae (Pseudoperisphinctinae, Perisphinctinae, Passendorferiinae), and Ataxioceratidae. A range-chart of selected ammonite species is shown in Fig. 2.

- Horizon 0-30 - *Perisphinctes* (*Otosphinctes*) *spathi* Meléndez.
- Horizon 30-45 - *Sowerbyceras* sp., *P.* (*Otosphinctes*) sp.
- Horizon 55-105 - *Lytoceras adelae* (D'Orbigny), *Taramelliceras* (*Taramelliceras*) *callicerum* (Oppel), *Creniceras* sp. gr. *renggeri* (Oppel *in* Sapunov, 1976), *P.* (*Properisphinctes*) sp. gr. *bernensis* (De Loriol).
- Horizon 115-130 - *Pachyceras* (*Tornquistes*) sp. gr. *romani* Douville-*kobyi* Loriol, *Euaspidoceras* (*Euaspidoceras*) sp. gr. *akantheen* (Buckman), *P.* (*Arisphinctes*) sp.
- Horizon 130-145 - *Sowerbyceras tortisulcatum* (D'Orbigny), *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *Taramelliceras* (*Proscaphites*) *anar* (Oppel), *Euaspidoceras* (*Euaspidoceras*) *paucituberculatum* (Arkell), *E.* (*Euaspidoceras*) *oegir* (Oppel), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 145-155 - *Sowerbyceras tortisulcatum* (D'Orbigny), *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *Taramelliceras* (*Proscaphites*) *anar* (Oppel), *Gregoryceras* (*Gregoryceras*) sp. cf. *riazi* (De Grossouvre), *G.* (*G.*) *fouquei* (Kilian), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 155-165 - *Taramelliceras* (*Taramelliceras*) sp. cf. *costatum* (Quenstedt)-*pseudotrachinotum* (Hölder).
- Horizon 175-190 - *Sowerbyceras tortisulcatum* (D'Orbigny), *Paraspidoceras* (*Struebinia*) sp. A cf. *edwardsianum* (D'Orbigny *in* Gemmellaro, 1877), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 190-205 - *Sowerbyceras tortisulcatum* (D'Orbigny), *Taramelliceras* (*Taramelliceras*) sp. cf. *sarasini* (De Loriol), *Glochiceras* (*Glochiceras*) sp. gr. *tectum* Ziegler-*bruckneri* (Oppel) or *G.* (*G.*) sp. gr. *nimbatum* (Oppel).
- Horizon 210-225 - *Calliphylloceras manfredi* (Oppel), *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 225-235 - *Sowerbyceras tortisulcatum* (D'Orbigny), *G.* (*Gregoryceras*) *transversarium* (Quenstedt), *Mirosphinctes bukowskii* (Choffat), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 235-240 - *Calliphylloceras manfredi* (Oppel), *Taramelliceras* (*Taramelliceras*) sp. cf. *pseudotrachinotum* Hölder, *G.* (*Glochiceras*) sp. gr. *tectum* Ziegler-*bruckneri* (Oppel) or G. (*G.*) sp. gr. *nimbatum* (Oppel).

LUGAR SECTION



- Horizon 230-250 - *Sowerbyceras tortisulcatum* (D'Orbigny), *Taramelliceras* (*Proscaphites*) *anar* (Oppel), *E.* (*Euaspidoceras*) *oegir* (Oppel).
- Horizon 0-10b - *Calliphylloceras manfredi* (Oppel), *Holcophylloceras mediterraneum* (Neumayr), *Sowerbyceras tortisulcatum* (D'Orbigny), *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *Taramelliceras* (*Taramelliceras*) *callicerum* (Oppel), *T.* (*Proscaphites*) sp., *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch).
- Horizon 15-30b - *Calliphylloceras manfredi* (Oppel), *Sowerbyceras tortisulcatum* (D'Orbigny).
- Horizon 80-90b - *Holcophylloceras mediterraneum* (Neumayr), *Sowerbyceras tortisulcatum* (D'Orbigny), *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *P.* (*Otosphinctes*) sp. gr. *vermicularis* Lee-*sorliensis* De Loriol.
- Horizon 105-120b - *T.* (*Proscaphites*) *gessneri* (Oppel), *Mirosphinctes bukowskii* (Choffat).
- Horizon 160-175b - *E.* (*Euaspidoceras*) *paucituberculatum* (Arkell), *P.* (*Dichotomoceras*) *stenocycloides* s.l. (Ronchadzé-Siemiradzki).
- Horizon 175-190b - *Sowerbyceras tortisulcatum* (D'Orbigny), *P.* (*Dichotomoceras*) *stenocycloides* s. l. (Ronchadzé-Siemiradzki), *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch), *P.* (*Enayites*) sp. A.
- Horizon 250-265b - *Calliphylloceras manfredi* (Oppel), *Sowerbyceras tortisulcatum* (D'orbigny), *Lissoceras* (*Lissoceratoides*) *erato* (D'Orbigny), *Mirosphinctes* n. sp. A aff. *niedzwiedzkii* (Siemiradzki), *P.* (*Enayites*) sp. C.
- Horizon 250-255 - *Sowerbyceras* sp. C, *Lytoceras orsinii* (Gemmellaro *in* Favre, 1875), *Lissoceras* (*Lissoceratoides*) *erato* (D'orbigny), *T.* (*Strebliticeras*) *externodosum* (Dorn).
- Horizon 265-285b - *Sowerbyceras* sp., *Mirosphinctes* sp.
- Horizon 300-305b - *Orthosphinctes* (*Orthosphinctes*) sp. gr. *fontannesi* (Choffat).
- Horizon 325-345b - *T.* (*Strebliticeras*) *externodosum* (Dorn), *E.* (*Euaspidoceras*) *eucyphum* (Oppel).
- Horizon 350-365b - *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *T.* (*Strebliticeras*) *externodosum* (Dorn), *Epipeltoceras* sp. cf. *semimammatum circumcostatum* (Dorn), *E. berrense* (Favre).
- Horizon 365-380b - *Calliphylloceras manfredi* (Oppel), *Holcophylloceras mediterraneum* (Neumayr), *Holcophylloceras zygnodianum* (D'Orbigny), *Sowerbyceras tortisulcatum* (D'orbigny), *E. semiarmatum* (Quenstedt), *E. berrense* (Favre).
- Horizon 380-395b - *Holcophylloceras mediterraneum* (Neumayr), *T.* (*Taramelliceras*) *sarasini* (Loriol *in* Dorn, 1930), *G.* (*Coryceras*) *microdomum* (Oppel), *Clambites* (*Clambites*) *clambus* (Oppel), *P.* (*Enayites*) *rozaki* Meléndez.
- Horizon 395-410b - *Sowerbyceras tortisulcatum* (D'Orbigny), *E.* (*Euaspidoceras*) *hypselum* (Oppel), *Clambites* (*Clambites*) sp. cf. *schwabi* (Oppel), *E. bimammatum* (Quenstedt), *E.* sp. gr. *berrense* (Favre), *E. treptense* Enay, *P.* (*Enayites*) *rozaki* Meléndez, *Sutneria* sp. gr. *galar* (Oppel).
- Horizon 590-600 - *Subdiscosphinctes* (*Subdiscosphinctes*) n. sp. A. aff. *aeneas* (Gemmellaro *in* Choffat, 1893).
- Horizon 605-615 - *E.* sp. gr. *berrense* (Favre), *Physodoceras wolfy* (Neumayr), *Benetticeras benettii* Checa, *Subdiscosphinctes* (*Subdiscosphinctes*)? *freybergi* (Geyer), *P.* (*Enayites*) sp. gr. *rozaki* Meléndez, *Sutneria* sp. gr. *galar* (Oppel), *Orthosphinctes* (*Praeataxioceras*) sp.
- Horizon 640-655 - *Holcophylloceras mediterraneum* (Neumayr), *Sowerbyceras* sp. A, *S.* sp. C, *T.* (*Taramelliceras*) sp. gr. *costatum* (Quenstedt), *T.* (*Metahaploceras*) *wenzeli* (Oppel), *P.* (*Enayites*) sp. gr. *rozaki* Meléndez, *Subnebrodites planula* (Hehl *in* Zieten, 1830), *Orthosphinctes* (*Praeataxioceras*) sp., *O.* (*Orthosphinctes*) sp. gr. *mogosensis* (Choffat).
- Horizon 495-520b - *T.* (*Metahaploceras*) sp. cf. *pseudowenzeli* (Wegele), *P.* (*Enayites*) *wierzbowskii* Meléndez.
- Horizon 520-535b - *T.* (*Metahaploceras*) sp. cf. *kobyi quenstedti* (Hölder)-*wegelei* Schairer, *Clambites* (*Clambites*) *clambus* (Oppel), *Pseudowaagenia micropla* (Oppel), *P.* (*Enayites*) *wierzbowskii* Meléndez, *Sutneria* sp. gr. *galar* (Oppel).

Figure 2. Ranges of selected ammonites at the Lugar section.

- Horizon 680-695 - *Sowerbyceras* sp. C, *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *T.* (*Metahaploceras*) sp. gr. *falcula* (Quenstedt), *Benetticeras benettii* Checa, *P.* (*Enayites*) *wierzbowskii* Meléndez, *Subnebrodites planula* (Hehl *in* Zieten, 1830), *S. laxevolutum* (Fontannes), *S. minutum* (Dieterich), *S. schröederi* (Wegele), *S.* n. sp. A.
- Horizon 720-735 - *Calliphylloceras manfredi* (Oppel), *Sowerbyceras* sp. C, *T.* (*Strebliticeras*) sp. gr. *externodosum* (Dorn), *T.* (*Metahaploceras*) *wenzeli* (Oppel), *Sutneria galar* (Oppel).
- Horizon 755-770 - *S. tortisulcatum* (D'Orbigny), *S.* sp. C, *Lytoceras orsinii* (Gemmellaro *in* Favre, 1875), *Taramelliceras* (*Strebliticeras*) sp. gr. *externodosum* (Dorn), *Glochiceras* (*Coryceras*) *microdomum* (Oppel), *Sutneria galar* (Oppel), *Subnebrodites laxevolutum* (Fontannes), *S. minutum* (Dieterich), *S.* sp. gr. *proteron* (Nitzopoulos), *S.* n. sp. A, *O.* (*Orthosphinctes*) *polygyratus* (Reinecke).
- Horizon 775-785 - *Calliphylloceras manfredi* (Oppel), *Holcophylloceras mediterraneum* (Neumayr), *Sowerbyceras tortisulcatum* (D'Orbigny), *S.* sp. C, *Lytoceras polyanchomenum* (Gemmellaro *in* Favre, 1876), *Lytoceras orsinii* (Gemmellaro *in* Favre, 1875), *T.* (*Metahaploceras*) *kobyi quenstedti* (Hölder), *T.* (*Metahaploceras*) sp. gr. *kobyi* (Choffat), *T.* (*Metahaploceras*) sp. aff. *subnereus* (Wegele), *Aspidoceras binodum* (Oppel), *A. sesquinodosum* (Fontannes *in* Dumortier and Fontannes, 1876), *Physodoceras altenense* (D'Orbigny), *Benetticeras benettii* Checa, *Sutneria galar* (Oppel), *Subnebrodites laxevolutum* (Fontannes), *O.* (*Orthosphinctes*) sp. cf. *polygyratus* morph. *colubrinus* (Reinecke *in* Olóriz, 1978).

Biochronostratigraphy

The biochronostratigraphic interpretation of the Oxfordian section investigated at the Lugar section is based on the composition of ammonite assemblages recovered from the horizons quoted above. Present data and interpretations improve preliminary ones provided by Caracuel (1996) and the authors in the Abstract Volume of this Symposium:

Lower Oxfordian. Renggeri Zone

Horizons 0-30 to 55-105. The Renggeri Zone is identified by the record of *Perisphinctes* (*Otosphinctes*) *spathi* Meléndez, *P.* (*Properisphinctes*) sp. gr. *bernensis* (De Loriol) and *Creniceras* sp. gr. *renggeri* (Oppel *in* Sapunov, 1976). The presence of *Taramelliceras* (*Taramelliceras*) *callicerum* (Oppel) in the upper part is interpreted as resulting from condensation during late Renggeri and early Antecedens Chrons. A stratigraphically equivalent interval in the section studied at the nearby Sierra de Corque provided *P.* (*Prososphinctes*) sp. gr. *claromontanus-mazuricus* (Bukowski) and rare *Parawedekindia* sp. and *Peltoceratoides* (subgen. indet.) sp.

The Renggeri Zone was identified in the area by Checa and Sequeiros (1990), and correlated with the same interval proposed by Sapunov (1976) in Bulgaria, as well as with the Claromontanus Zone in the Iberian Chain (Meléndez, 1989). In addition, we correlate the interpreted Renggeri Zone with the Cordatum Zone proposed by Enay (1966), Cariou et al. (1971) and Cariou and Meléndez (1990) for epicontinental regions in southern Europe. In more condensed sections, as usually occur in Mediterranean areas, the identification of a stratigraphic interval equivalent to the Paturattensis Zone *sensu* Tarkowski (1990) could apply.

Middle Oxfordian. Upper Plicatilis Zone (Antecedens Subzone)

Horizons 115-130. Among recovered ammonites, *Pachyceras* (*Tornquinstes*) belonging to the *romani* Douville and *kobyi* De Loriol groups, together with *Euaspidoceras* of the *akantheen* (Buckman) group have been considered to be significant for identifying the Antecedens Subzone.

In the Betic Cordillera, the Antecedens Biozone was identified by Sequeiros (1974) as the upper part of the Plicatilis Zone in the central part of the Subbetic Zone, and recognized by Checa and Sequeiros (1990) in the eastern Subbetic at section Corque-5, in the neighbouring Sierra de Corque. Olóriz et al. (1998) identified the Antecedens Subzone, without subdivision, in the Sierra Norte of Mallorca Island. In Bulgaria, Sapunov (1976) interpreted the Antecedens stratigraphic interval at the Zone level and subdivided it into two subzones, the Rotoides (lower) and Dobogrensis (upper) Subzones, which are the only sudivisions proposed for this stratigraphic interval in Mediterranean areas. However, the Antecedens Zone or Subzone is usually condensed and difficult to subdivide in the ammonitico rosso facies characterizing the Mediterranean Tethys.

Middle Oxfordian. Riazi (=Transversarium) Zone

Horizons 130-145 to 105-120b. *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch) is frequent throughout this zone and was also collected from the lower part of the Bifurcatus Zone. Single specimens of *Gregoryceras* (*Gregoryceras*) sp. cf. *riazi* (De Grossouvre), *G.* (*G.*) *fouquei* (Killian), *G.* (*G.*) *transversarium* (Quenstedt) and *Taramelliceras* (*Taramelliceras*) sp. cf. *sarasini* (De Loriol) were recovered from the lower part of the Riazi (=Transversarium Zone), together with *T.* (*Proscaphites*) *anar* (Oppel), the range of which was first specified in the area extending slightly above this interval. Also registered in the lower part were *Euaspidoceras* (*Euaspidoceras*) *paucituberculatum* (Arkell), *E.* (*E.*) *oegir* (Oppel), which extends to the lower upper part of the Riazi (=Transversarium) Zone, *T.* (*Taramelliceras*) of the *pseudotrachinotum* (Hölder) group, and *Paraspidoceras* (*Struebinia*) sp. A cf. *edwardsianum* (D'Orbigny). The range recorded of *Mirosphinctes bukowskii* (Choffat), from horizons 225-235 to 105-120b, is interpreted to indicate roughly the upper part of the Riazi (= Transversarium) Zone, but its younger records could belong to the Bifurcatus Zone, although they are below the first *Perisphinctes* (*Dichotomoceras*) in the Lugar section. Also present in the upper part of the Riazi (=Transversarim) Zone are *T.* (*Taramelliceras*) *callicerum* (Oppel), *T.* (*Proscaphites*) sp., and *P.* (*Otosphinctes*) sp. gr. *vermicularis* Lee-*sorliensis* De Loriol. Less diagnostic species found in the Riazi (= Transversarim) Zone are *G.* (*Glochiceras*) sp. gr. *tectum* Ziegler-*bruckneri* (Oppel) and *T.* (*Proscaphites*) *gessneri* (Oppel).

In comparison with the usual record from the central part of the Subbetic Zone in the Betic Cordillera and the Sierra Norte on Mallorca Island (Sequeiros, 1974; Olóriz et al., 1998), attention should be drawn to the shorter range and the lower abundance of *Gregoryceras* in the Lugar section. The correlation of the Riazi (=Transversarium Zone) in the Betic Cordillera with the Transversarium Zone in southern Europe (Cariou et al, 1971; Cariou and Meléndez, 1990; Cariou et al., 1991) was long ago established by Sequeiros (1974). At present, no subdivision other than the informal one alluded to above can be proposed for the Riazi (= Transversarium) Zone in the Lugar section.

Upper Oxfordian. Bifurcatus Zone

Horizons 105-120b?, and from 160-175b to 250-255. Easy to identify through the appearance of *Perisphinctes* (*Dichotomoceras*) *stenocycloides* s.l. (Ronchadzé-Siemiradzki), the range of which characterizes the lower part of this zone. Also registered in the lower part was *Passendorferia* (*Enayites*) *birmensdorfensis* (Moesch). Scarce specimens of *P.* (*Enayites*) sp. A and sp. C, and a single *Mirosphinctes* nov. sp. A aff. *niedzwiedzkii* (Siemiradzki) were found in the Bifurcatus Zone, the upper part of which is interpreted, indirectly, to be the remainder deposits above those with *Perisphinctes* (*Dichotomoceras*) *stenocycloides* and below the record of species indicating the overlying Bimammatum Zone. *Lissoceras* (*Lissoceratoides*) *erato* (D'Orbigny), a long-ranging species in the central Subbetic Zone (Sequeiros, 1974), was found in the upper part of the Bifurcatus Zone and the base of the Bimammatum Zone in the Lugar section.

In the Subbetic Zone of the Betic Cordillera, the range of the genus *Perisphinctes* (*Dichotomoceras*) characterizes the Bifurcatus Zone (Sequeiros, 1974), and has been traditionally used to correlate with epicontinental and epioceanic deposits in southern Europe. The absence of joint records of *Gregoryceras* and *P.* (*Dichotomoceras*) in the Lugar section is noteworthy, but cannot be conclusively interpreted, since this is well known both westwards, in the central Subbetic Zone (Sequeiros, 1974), and eastwards, at Sierra Norte on Mallorca Island (Olóriz et al., 1998). At present, the Bifurcatus Zone in the Lugar section can be subdivided into a lower Stenocycloides Subzone, which is tentatively correlated with the Stenocycloides Subzone in the Iberian Range (Meléndez, 1989; Cariou and Meléndez, 1990), and an upper part imprecisely known.

Upper Oxfordian. Bimammatum Zone

Horizons 250-255 to 605-615. The base of this zone is recognised by the appearance of *Taramelliceras* (*Strebliticeras*) *externodosum* (Dorn), which was registered in the lowermost part of this zone (as well as forms belonging to the same species group in the lower part of the younger Galar Subzone). Assuming that the record of this species also includes the uppermost Bifurcatus Zone (Sequeiros, 1974), it is difficult to rule out the possibility of condensation of latest Bifurcatus and earliest Bimammatum deposits in the Lugar section due to the rarity of *Perisphinctes* (*Dichotomoceras*) in the upper Bifurcatus Zone. Significant biostratigraphic data in the interpreted lowermost Bimammatum Zone below the first record of *Epipeltoceras* are the last record of

Mirosphinctes sp. and the appearance of single specimens of *Orthosphinctes* (*Orthosphinctes*) sp. gr. *fontannesi* (Choffat) and *Euaspidoceras eucyphum* (Oppel). Less significant ammonites in the Bimammatum Zone are haploceratids, glochiceratids and oppeliids.

The record of *Epipeltoceras* permits the possibility of subdividing of younger horizons within the Bimammatum Zone in the Lugar section. *E.* sp. cf. *semimammatum circumcostatum* (Dorn) and then *E. semiarmatum* (Quenstedt) characterize the lower record of the genus, together with *E. berrense* (Favre). In these horizons *Passendorferia* (*Enayites*) *rozaki* Meléndez is frequent, and *Clambites* (*Clambites*) *clambus* (Oppel) rare. Higher horizons yielded *E. bimammatum* (Quenstedt), *E. treptense* (Enay) and *E.* sp. gr. *berrense* (Favre) associated with *Passendorferia* (*Enayites*) *rozaki* Meléndez, single specimens of *Euaspidoceras hypselum* (Oppel) and *C.* (*Clambites*) sp. cf. *schwabi* (Oppel), and to a new species of *Sutneria* belonging to the group of *galar* (Oppel). The latter species, *S.* sp. gr. *galar* (Oppel), was recorded together with the youngest *E.* sp. gr. *berrense* (Favre), from overlying horizons above the last records of other species of *Epipeltoceras* and below the first record of *Subnebrodites*. Collected from this uppermost part of the Bimammatum Zone were rare *Subdiscosphinctes* (*Subdiscosphinctes*) n. sp. A aff. *aeneas* (Gemmellaro in Choffat, 1893), *S.* (*Subdiscosphinctes*)? *freybergi* (Geyer) and *Orthosphinctes* (*Praeataxioceras*) sp., as well as *Passendorferia* (*Enayites*) sp. gr. *rozaki* Meléndez and the first *Physodoceras wolfy* (Neumayr) and *Benetticeras benettii* Checa.

Since the lower part of the uppermost Bimammatum Zone yielded *Epipeltoceras* sp. gr. *berrense* (Favre), but neither any other species of this genus nor *Taramelliceras hauffianum*, it is difficult at present to say whether the Hauffianum Subzone is missing due to a hiatus or whether its recognition does not apply in Mediterranean ammonitico rosso, as seems to be deduced from the literature. However, Olóriz et al. (1998) identified condensed deposits (ammonitico rosso) with *T. hauffianum* in central Sierra Norte (Mallorca).

According to the above, three subdivisions can be identified within the Bimammatum Zone in the Lugar section, the two lower ones are correlated with the Hypselum and Bimammatum Subzones. However, future research is necessary to characterize the uppermost Bimammatum Zone, correlation of which with the Hauffianum Subzone must be tested. Within the lowermost Hyselum Subzone, a lower *Fontannesi-Semicostatum* horizon and an upper *Berrense* horizon can be envisaged, which probably are equivalent to those identified by Meléndez et al. (1995) in the northeastern Iberian Range. Moreover, on the basis of the biostratigraphic record known from epicontinental areas in southern Europe, variable condensation is assumed to affect the lower boundary of the Bimammatum Zone (the lowermost part without *Epipeltoceras*), the Hypselum Subzone (assembled record of *E. semimammatum* and *E. berrense*) and the uppermost Hypselum and lowermost Bimammatum Subzones (joint records of *Euaspidoceras hypselum* and *E. bimammatum* and *treptense*). In addition, hiatuses probably affected the uppermost Bimammatum Zone in the Lugar section.

Uppermost Oxfordian or Lowermost Kimmeridgian, Planula Zone

Horizons 640-655 to 775-785. The lower boundary is identified by the appearance of *Subnebrodites planula* (Hehl). The upper boundary corresponds to the last record of *Sutneria galar* (Oppel) just below the appearance of *S. platynota* (Reinecke). Ammonite ranges show some significant traits which could indicate condensation, such as the record of *Subnebrodites* together with *S. galar* (Oppel) and no differentiation of ranges in *Subnebrodites* equivalent to that known from epicontinental areas. However, even assuming condensation, some notable biostratigraphic data are the turnover between the ranges of *S.* sp. gr. *galar* (Oppel) and *S. galar* (Oppel), the combined record of *S.* sp. gr. *galar* (Oppel) with *Passendorferia* (*Enayites*), the combined record of the latter with *Subnebrodites*, the record of *Subnebrodites* above the younger *P.* (*Enayites*) *wierzbowskii* Meléndez, and the clearly differentiated record of primitive *Orthosphinctes* (*mogosensis* group) in the older horizons and specimens of the *polygyratus* group above. Moreover, the record of *S. minutum* within the upper range of the genus *Subnebrodites* is noteworthy since corroborates recent reports from epicontinental areas within the Betic Cordillera (Olóriz et al., 1992), as well as in Poland (Matyja and Wierzbowski, 1997).

Future research is necessary to reach a conclusive interpretation of the Planula Zone in epioceanic ammonitico rosso, but the subdivision of the Planula Zone, including the recognition of the Galar Subzone, was proposed by Behmel (1970) and Sequeiros and Olóriz (1979) in the eastern and central Subbetic, respectively. There is no doubt condensation occurred in the Lugar section, but the subdivision of the Planula Zone seems to be possible even assuming peculiarity in some ranges of the ammonite species and/or genera registered. Hence, at present, the identification of the lower Planula Subzone below and an upper Galar Subzone characterized by the record of the index species, together with *Orthosphinctes* of Kimmeridgian affinity and without known species of *Passendorferia* (*Enayites*), seems to be appropriate in absence of significant data provided by the haploceratids recovered. In general terms, the correlation of these two subzones with the standard subdivison of the Planula Zone can be assumed.

Conclusions

The Lugar section is one of the better outcrops for analyzing Oxfordian biostratigraphy in epioceanic ammonitico rosso facies in the Subbetic Zone of the Betic Cordillera in southern Spain. The Lower Oxfordian Renggeri Zone, the Middle Oxfordian upper Plicatilis (Antecedens Subzone) and Riazi (=Transversarium) Zones, Upper Oxfordian Bifurcatus and Bimammatum Zones, and uppermost Oxfordian or lowermost Kimmeridgian Planula Zone have been identified. No ammonite subzones can be recognized in the Riazi (= Transversarium) Zone, and the genus *Gregoryceras* is less abundant and has a stratigraphic range shorter than usually recognized in the Subbetic Zone of the Betic Cordillera. The Stenocycloides Subzone is proposed for the lower part of the Bifurcatus Zone, in which the identification of the upper part has yet to be made. The Hypselum and the Bimammatum Subzones have been identified in the Bimammatum Zone, but no conclusive evidence exists of the Hauffianum Subzone. The subdivision of the Hypselum Subzone into the *Fontannesi-Semicostatum* horizon (lower) and the *Berrense* horizon (upper) is proposed. The Planula Zone can be subdivided in the Planula and Galar Subzones, but ammonite ranges show significant differences with those traditionally recognised. As usually recognized in ammonitico rosso, condensation and/or hiatuses are common, and, together with assumed differences in morphologic evolution between epioceanic and epicontinental ammonites, impede the conclusive interpretation and correlation of the ammonite ranges obtained.

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References

BARTHEL, K.W., F. CEDIEL, O. F. GEYER, AND J. REMANE. 1966. Der Jura von Cehegín (Provinz of Murcia, Spanien). Mitteilungen der Bayerischen Staatssammlung für Paläontologie und historische Geologie, 6:167-211.

BEHMEL, H. 1970. Beiträge Sur Stratigraphie und Paläontologie des Juras von Ostspanien, p. 1-102. *In* Stratigraphie und Fazies im präbetischen Jura von Albacete und Nord-Murcia. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 137(1).

CARACUEL, J. E. 1996. Asociaciones de megainvertebrados, evolución ecosedimentaria e interpretaciones ecoestratigráficas en umbrales epioceánicos del Tethys occidental (Jurásico superior). Ph.D. Thesis, University of Granada, 474 p.

CARIOU, E., R. ENAY, AND H. TINTANT. 1971. Oxfordien. In: Mouterde et al. "Les zones du Jurassique en France". Comptes Rendus Sommaire des Séances de la Société géologique de France, 6:18-21.

___, F. ATROPS, P. HANTZPERGUE, R. ENAY, AND M. RIOULT. 1991. Oxfordien. *In* Réactualisation des échelles d'Ammonites - Groupe Français d'Etude du Jurassique. Abstracts Volume 3rd International Symposium on Jurassic Stratigraphy, Poitiers, 132.

___, AND G. MELÉNDEZ. 1990. A modified Perisphinctid zonation for the Middle Oxfordian of southern Europe Submediterranean Province. Publicaciones del Seminario de Paleontologia de Zaragoza (SEPAZ), 2:129-152.

CHECA, A., AND L. SEQUEIROS. 1990. New data on the Lower Oxfordian from the Subbetic Zone (Betic range, S.E. Spain). Publicaciones del Seminario de Paleontologia de Zaragoza (SEPAZ), 2:153-1160.

CHOFFAT, P. 1893. Description de la faune jurassique de Portugal, Classe des Céphalopodes, 1^{a} série: Ammonites de Lusitanien de la contreé de Torres-Vedras. Direction des Travaux Géologiques du Portugal, Lisbonne, 82 p.

DORN, P. 1930-31. Die Ammoniten des untersten Malm der Frankenalb. Palaeontographica, Bd. 73-74:107-172 (1930); 1-92 (1931).

DUMORTIER, E., AND F. FONTANNES. 1876. Descriptions des Ammonites de la Zone à *Ammonites tenuilobatus* de Crussol (Ardèche) et de quelques autres fossiles jurassiques nouveaux ou peu connus. Mémoires de l' Académie de Lyon, 21:1-161.

ENAY, R. 1966. L'Oxfordien dans la moitié sud du Jura français. I Etude Stratigraphique. II Etude Paléontologique. Nouvelles Archives du Muséum d'Histoire naturelle de Lyon, 624 p.

FAVRE, E. 1875. Description des fossiles du terrain jurassique de la montagne des Voirons (Savoie). Mémoires de la Société Paléontologique Suisse, 2:1-79.

____. 1876. Description des Fossiles du terrain Oxfordien des Alpes fribourgeoises. Mémoires de la Société Paléontologique Suisse, 1- 3:1-75.

FÖZY, I. 1993a. Upper Jurassic ammonite biostratigraphy in the Meesek Mts., southern Hungary. Földtani Közlöny, 123 (2):195-205.

____. 1993b. Upper Jurassic ammonite biostratigraphy in the Gerecse and Pilis Mts. (Transdanubian Central range, Hungary). Földtani Közlöny, 123 (4):441-464.

FÜLOP, J. 1976. The mesozoic basement horst blocks of Tata. Geologica Hungarica, Series Geologica, 16:5-228.

GEMMELLARO, G. G. 1877. Sopra alcuni fossili della zona con *Peltoceras transversarium* Quenstedt sp. del monte Erice, or San Giuliano, nella Provinzia de Trapani. Giornali di Scienze Naturali ed Economiche, Palermo, 12:156-172.

MATYJA, B. A., AND A. WIERZBOWSKI. 1997. The quest for a unified Oxfordian/Kimmeridgian boundary: implications of the ammonite succession at the turn of the Bimammatum and Planula Zones in the Wielun Upland. Central Poland. Acta Geologica Polonica, 47 (1-2):77-105.

MELÉNDEZ, G. 1989. El Oxfordiense en el sector central de la Cordillera Ibérica (Provincias de Zaragoza y Teruel). Ph.D. Thesis. Institución Fernando el Católico e Instituto de Estudios Turolenses, 418 p.

____, I. PÉREZ-URRESTI, AND J. BELLO. 1995. Las subdivisiones bioestratigráficas de la Biozona Bimammatum (Oxfordiense Superior) en la Cordillera Ibérica nororiental: facies y amplitud de las discontinuidades. Geogaceta, 18:117-120.

OLÓRIZ, F. 1978. Kimmeridgiense-Tithonico inferior en el Sector Central de las Cordilleras Beticas (Zona Subbética). Paleontología. Bioestratigrafía. Tesis Doctorales de la Universidad de Granada, 184 (1-2), 758 p.

____, F. J. RODRÍGUEZ-TOVAR, AND G. SCHAIRER. 1992. New record of *Barthelia subbetica* Olóriz and Schairer (Jurassic Ammonitina) from the South Iberian paleomargin (Prebetic Zone, Spain). Neues Jahrbuch für Geologie und Paläontologie Monatshefte, 6: 343-350.

____, B. MARQUES, AND J. E. CARACUEL. 1998. The Middle-Upper Oxfordian of Central Sierra Norte (Mallorca, Spain), and progressing ecostratigraphic approach in western Tethys. Geobios, 31 (3):319-336.

SAPUNOV, J. G. 1976. Ammonite stratigraphy of the Upper Jurassic in Bulgaria. IV Tithonian: Substages, Zones and Subzones. Geologica Balcanica, 7 (2):43-64.

____. 1979. Les fossiles de Bulgarie III/3 Jurassique Supérieur, Ammonoidea. Academie Bulgare des Sciences, 263 p. (In Bulgarian with abridged version in French)

SARTI, C. 1988. Biostratigraphic subdivision for the Upper Jurassic of the Venetian Alps (Northern Italy) on the base of ammonites, p. 459-476. *In* R. B. Rocha and F. Soares (eds.), 2nd Symposium on Jurassic Stratigraphy, Lisboa.

SCHWEIGERT, G. 1995. Zum Auftreten der Ammonitenarten *Amoeboceras bauhini* (Oppel) und *Amoeboceras shulginae* Mesezhnikov im Oberjura der Schwäbischen Alb. Jahreshefte der Gesellschaft für Naturkunde Württemberg, 151:171-184.

SEQUEIROS, L. 1974. Paleogeografía del Calloviense y Oxfordense en el Sector Central de la Zona Subbética. Ph.D. Thesis, University of Granada, 636 p.

____, AND F. OLÓRIZ. 1979. El Oxfordense de la Zona Subbética. Cuad. Geol. Universidad de Granada, 10:463-474.

TARKOWSKI, R. 1990. Les *Taramelliceras* (Ammonitina) de l'Oxfordien Inferieur du Jurassique Cracovien: Valeur stratigraphique. Publicaciones del Seminario de Paleontologia de Zaragoza (SEPAZ), 2:205-216.

ZIETEN, C-H. 1830-34. Die Versteinerungen Württembergs. Stuttgart, 102 p.

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Early to Middle Jurassic Magmatism: New Zircon U-Pb Ages and Comparative Geochemistry from a Crustal Transect through the Bonanza Arc on Vancouver Island, Canada

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Abstract: The Westcoast Crystalline Complex (WCC), Island Intrusions, and Bonanza Group volcanic rocks of Vancouver Island, Canada, comprise different crustal levels of the Early to Middle Jurassic Bonanza island arc. New geochronology and geochemistry confirms ideas proposed by earlier workers that these three igneous suites represent a single age-equivalent magmatic arc. Our study shows that these units represent the deep, middle, and shallow crustal levels of the arc, respectively. Differential uplift has exposed the arc for a strike length of ~500km on Vancouver Island and for another 250km on the Queen Charlotte Islands.

New U-Pb crystallization ages for the WCC (190.3 ±1.0 Ma, 177.3 ±1.8 Ma, 174.2±0.4 Ma) and the Island Intrusions (175.7 ±2.6 Ma, 170.3 ±0.9 Ma, 168.6 ±5.3 Ma) show that these two plutonic units are coeval, and overlap with fossil and isotopic ages previously reported for the Bonanza Group volcanic rocks. Younger Middle Jurassic ages (170-168 Ma) for the eastern Island Intrusions overlap with those for plutonic rocks in the southern Coast Belt and the Queen Charlotte Islands.

All plutonic and volcanic rocks within the arc have geochemical signatures that support their comagmatic origin. The most mafic rocks are typical of mafic island arc basalts worldwide. More felsic compositions (andesite/diorite to rhyodacite/tonalite) display a well-defined calcalkaline trend with REE patterns indicative of both mantle and crustal input.

Introduction

Wrangellia was one of the first accreted terranes recognized in the North American Cordillera. This paper explores the petrologic relationships among three suites of Jurassic crystalline rocks along a transect through Wrangellia on west-central Vancouver Island (Fig. 1) and presents evidence for their cogenesis. These suites include deep crustal metamorphic, migmatitic, and plutonic rocks of the Westcoast Crystalline Complex (WCC), middle- to upper-crustal granitic rocks of the Island Intrusions, and volcanic rocks of the Bonanza Group. The geologic relationships and similar K-Ar and Rb-Sr ages led previous workers to suggest that the WCC, the Island Intrusions and the volcanic rocks of the Bonanza Group might be part of a single magmatic arc system informally called the Bonanza arc (Isachsen, 1987; Muller et al., 1981). However, paucity of comprehensive geochemical data and high-precision U-Pb geochronologic data led to various interpretations of magma sources and space-time relationships among units. For example, Muller et al. (1981) suggested that the WCC was derived entirely by metamorphism of pre-existing crust at deep levels. In their interpretation, melts produced during this metamorphic event migrated to shallower levels in the arc to form the Island Intrusion plutons and Bonanza Group volcanic rocks, devoid of mantle input. However, isotopic data of Andrew et al. (1991) preclude this interpretation and suggest that a significant mantle-derived component must be involved.

The relationship of the Bonanza Group volcanic rocks to the Island Intrusions and the WCC, their potential correlations along strike, and their relationship to the younger Coast Plutonic Complex to the east, are critical petrologic and tectonic issues. Our new U-Pb geochronology and geochemical data confirm earlier proposals that the units represent part of a single magmatic arc.

The data also clearly show that the WCC represents the deep igneous and metamorphic roots to the arc, the Island Intrusions more intermediate crustal levels, and the Bonanza Group volcanic rocks the upper crustal carapace. The new U-Pb geochronology also permits comparisons between these rocks with those to the east in the Coast Mountains and to more northerly parts of Wrangellia in the Yakoun arc of the Queen Charlotte Islands and in the Talkeetna arc of south-central Alaska.

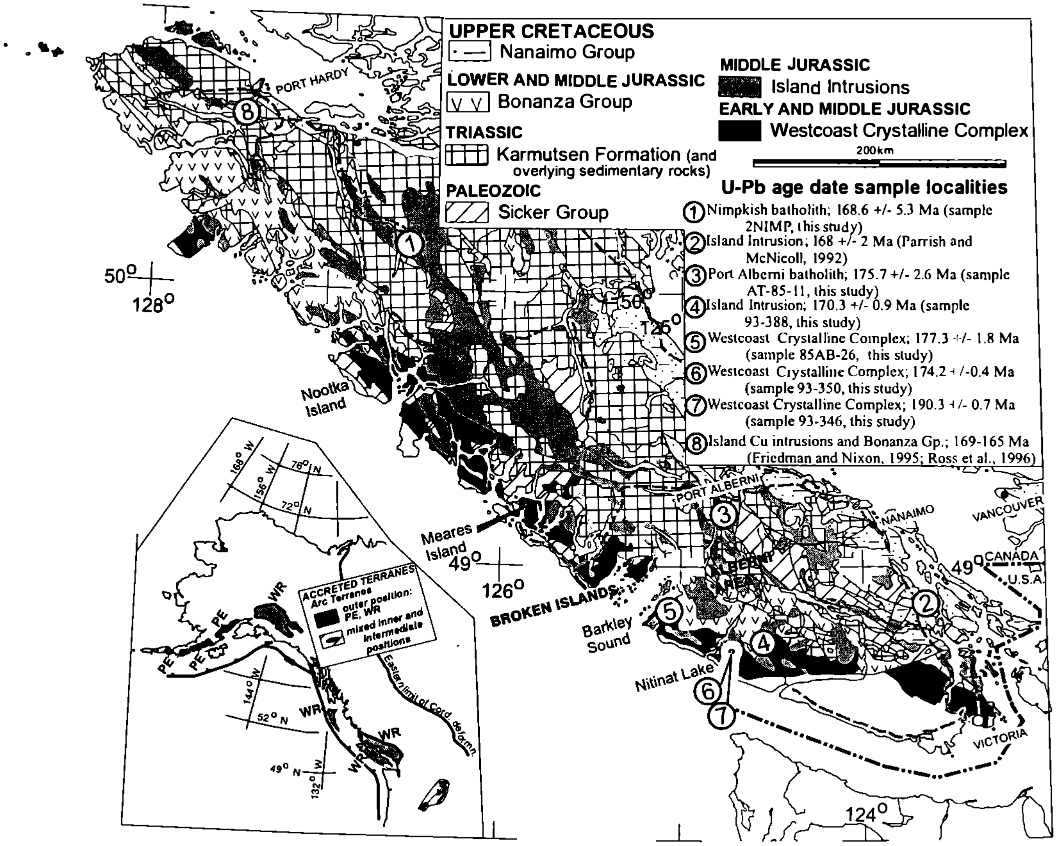

Figure 1. Geologic map of Vancouver Island. Modified after Journeay and Williams (1995). Zircon localities from this and earlier studies are noted by circled numbers. Inset is terrane map of western North America showing the extent of the Wrangellia terrane (dotted pattern) from the Wrangell Mountains in Alaska to Vancouver Island. Wrangellia (WR) is distinguished from the Peninsular Terrane (PE); the latter is the site of the Early Jurassic Talkeetna arc which may be correlative with the Bonanza arc.

Geologic Setting and Description of the Bonanza Arc

The Bonanza arc was built on distinctive Wrangellian basement comprising mid-Paleozoic arc volcanic rocks (Sicker Group), Pennsylvanian-Permian clastic rocks and limestone, and a remarkably uniform, up to 6000-meter-thick, sequence of Late Triassic tholeiitic basalts of the Karmutsen Formation (Fig. 1). These basalts are in turn overlain by Triassic-Jurassic shallow to deep-water carbonate rocks.

The Bonanza arc is thought to have developed on this Wrangellian basement in Early to Middle Jurassic times in response to eastward-directed subduction (Armstrong, 1988). After accretion of Wrangellia to the North American continental margin, subduction continued under an Andean-type continental margin, and the locus of magmatism swept eastward to form the Late Jurassic to Cretaceous Coast Plutonic Complex. Subsequent contraction and possible dextral transcurrent faulting along the continental margin further dismembered the arc (Plafker et al., 1989).

Coeval volcanic, plutonic, and sedimentary rocks of the Bonanza arc define a belt extending for 500 km in a NW-SE direction on western Vancouver Island (Fig. 1). Plutonic rocks of the arc are divided into two separate units, the Westcoast Crystalline Complex (WCC) and the Island Intrusions. The WCC is a heterogeneous assemblage of metamorphic and plutonic rocks that are variably deformed. The Island Intrusions are relatively undeformed, extensive granitoid batholiths and plutons that crop out in northwesterly-trending arrays. Muller et al. (1974a) noted that plutonic rocks in the WCC are commonly indistinguishable from, and grade into, the Island Intrusions. They separated the plutonic rocks of northern Vancouver Island into six parallel, northwest-striking zones. They ascribed relative depths of emplacement to these zones, with the WCC in the southwest displaying the deepest level of exposure (migmatitic catazone) and the Island Intrusions in the northeast displaying mesozonal and epizonal characteristics.

The Westcoast Crystalline Complex (WCC)

Gabbroic to dioritic plutonic rocks, mafic amphibolite, migmatite, and metasedimentary rocks of the WCC occur along the western edge of Vancouver Island (Fig. 1). Contacts of plutons with country rock are usually gradational, and are concordant to foliation in country rock. Foliation of the plutons is very strongly developed adjacent to these contacts, and changes inwards to undeformed pluton interiors. Agmatite is widespread and represents mingled mafic and felsic magmas as well as bulk mixtures of magma and amphibolitic country rock.

Country rocks to WCC plutons were historically included as amphibolite or metasedimentary subunits within the WCC because their metamorphic grade hindered comparisons to pre-Jurassic basement units. However, in some localities, metavolcanic and metasedimentary rocks within the WCC were mapped as part of the Sicker Group (Muller et al., 1981; Muller and Carson, 1969). In the present study, extensive petrographic analysis reveals that in many other areas, amphibolitized and migmatized country rock to (and inclusions within) the WCC are dominantly Sicker Group rocks. Features that identified these rocks with a Sicker Group protolith rather than Karmutsen Formation are the presence of intermediate volcanic rocks, meta-pyroclastic textures, and argillaceous sediments, as well as whole-rock compositions that match unmetamorphosed Sicker Group rocks. The predominance of Sicker Group as country rock is significant in that WCC rocks appear to have intruded only the deepest part of the pre-Jurassic stratigraphic section.

The WCC contains at least five mappable units. The *Gabbro-Peridotite unit* comprises mafic and ultramafic rocks including two-pyroxene hornblende gabbro, pyroxenite, and sheared serpentinite. The gabbro and pyroxenite are cumulates; their modal mineralogy, bulk composition, and textures suggest accumulation of magnesian pyroxene. The *Diorite unit* consists of massive to strongly foliated hornblendite, hornblende gabbro, hornblende diorite to quartz diorite, tonalite, and rare granodiorite. Hornblende diorite and hornblende quartz diorite are the most abundant rock types. Subsolidus deformation is distinguished from magmatic foliation by the presence of deformed and elongate quartz. Mafic enclaves are abundant and represent both migmatized amphibolite xenoliths, mingled mafic magmas, and disrupted mafic dikes. The *Amphibolite unit* consists dominantly of metavolcanic rocks (typically metamorphosed Sicker Group) with granoblastic to schistose and fine-grained gneissic texture. Amphibolite lithologies range in size from cm- to dm-scale xenoliths in diorite to km-scale pendants separating plutonic bodies. The *Migmatite unit* consists of amphibolite melanosomes intermingled with tonalitic to trondhjemitic leucosomes displaying gneissic to agmatitic textures. The leucosomes vary from veins in amphibolite to gneissic tonalitic bodies that enclose sharply-delineated to ghost-like melanosome inclusions. The *Metasedimentary rock unit* comprises rare exposures of amphibolite-grade calcsilicate, marble, metagreywacke, and argillite. They are strongly foliated and display complex isoclinal folding and gneissic to mylonitic texture. Because of their common association with metavolcanic rocks that are clearly Sicker Group, these metasedimentary rocks are most likely the metamorphic equivalents of Sicker Group siliciclastic and limestone units.

The Island Intrusions

The Island Intrusions consist of a belt of northwesterly-aligned batholiths and stocks of hornblende diorite to quartz diorite and granodiorite, with lesser gabbro, tonalite, and granite (Fig. 1). The most southwesterly dioritic members of the Island Intrusions grade into the migmatite and plutonic rocks of the WCC. Plutons of the Island Intrusions more commonly intrude volcanic rocks of the Bonanza Group and Karmutsen Formation than Sicker Group rocks. Contacts between the Island Intrusions and their country rock are much sharper and more discordant than those involving the WCC. Metamorphic grade of the country rock is generally not higher than greenschist facies.

Discrete Island Intrusion plutons east of the WCC display a sequential intrusive chronology of mafic, intermediate, and felsic phases, associated porphyritic dikes, widespread prismatic hornblende, an absence of titanite, and widespread, intense alteration of plagioclase and mafic minerals. In one such pluton, mafic, intermediate, and felsic phases are arranged in an elongate manner from northeast to southwest, parallel to the axis of the pluton. These phases exhibit petrographic features in common with deeper-level, coeval plutons to the west. As in the WCC, Island Intrusion plutons display complex interphase intrusive relationships, demonstrating that they existed as contemporaneous, partly consolidated magmas during intrusion. Some Island Intrusion plutons contain an ultramafic phase (feldspar-hornblende clinopyroxenite) that resembles ultramafic rocks in the mafic subunits of the WCC.

Bonanza Group Volcanic Rocks

The Lower to Middle Jurassic Bonanza Group lithology is varied and heterogeneous. The unit is up to 2500 m thick (Muller et al., 1974a) and consists of interbedded lava, breccia and tuff of basaltic, andesitic, dacitic, and rhyolitic composition.

Basalts and andesites of the study area are dark reddish to greenish-gray in color, and less massive and more coarsely vesicular than Karmutsen Formation basalts. They are generally fine grained to aphanitic, but distinctive coarse-grained porphyries containing densely-packed plagioclase phenocrysts also exist. Dacite contains more sodic plagioclase and less pyroxene than the andesites and basalts and is commonly aphanitic or vitrophyric. Rhyodacite contains phenocrysts of K-feldspar and plagioclase. The Bonanza Group rocks are very similar to parts of the Sicker Group but display a distinctly lower metamorphic grade. The Bonanza volcanic rocks are zeolite to lower greenschist grade, except near pluton margins where the volcanic rocks can reach upper greenschist facies.

Age Relationships

Possible cogenetic relationships among the WCC, Island Intrusions, and Bonanza Group volcanic rocks inferred from field observations, composition, and texture are best tested by determining their ages. If ages are similar and corroborate field relationships, geochemistry can be utilized to test whether they are comagmatic. Muller and coworkers (Muller et al., 1974a; 1981) considered these units to be comagmatic, but the mismatch of middle Early Jurassic Bonanza Group fossil ages and late Early to Middle Jurassic plutonic K-Ar and U-Pb ages was problematic. More recent fossil and geochronologic age determinations have shown the contemporaneity of the Island Intrusions and Bonanza Group (see below), and our new geochronologic data (DeBari et al., 1999) summarized here shows the contemporaneity of the Island Intrusions and WCC.

Previous Work

One of the original dates on the WCC was from a biotite gneiss near Tofino, which yielded a mid-Permian U-Pb age of 264 ±7 Ma (Muller et al., 1974b). This age sparked the controversy over the nature of the relationship between the WCC and the Island Intrusions. Subsequent workers were unable to replicate this analysis, and a sample from the same locality yielded an Early Jurassic U-Pb zircon age of 190 ±12 Ma (Isachsen, 1987). Isachsen (1987) also reports only Jurassic zircon U-Pb zircon ages from other WCC samples near Meares Island (176-190 Ma).

Recent, high quality U-Pb age dates for the Island Intrusions yield younger ages than those previously reported for the WCC. Parrish and McNicoll (1992) reported a U-Pb zircon age of 168 ±2 Ma for a biotite-quartz diorite from an Island Intrusion stock in southeast Vancouver Island (Fig. 1). Friedman and Nixon (1995) reported U-Pb ages for Island Intrusions near the Island Copper mine on northern Vancouver Island. Their results yielded ages of 167 ±0.5 Ma for the composite Wanokana batholith that hosts the Hushamu porphyry Cu deposit and 168 +8/-2 Ma for the stock on Rupert Inlet associated with the Cu-Au mineralized porphyry dike at Island Copper.

Bonanza Group volcanic rocks have been dated both by paleontologic and isotopic techniques. Ammonites from clastic sedimentary rocks interbedded with the volcanic flows of the Bonanza Group generally indicate Early Jurassic (Sinemurian to Pliensbachian) ages (Muller et al., 1981; Muller and Carson, 1969; Muller et al., 1974a), particularly in the study area. However, Bonanza

Group near the Island Copper deposit in northern Vancouver Island may include rocks as young as Aalenian (fossil ages from Poulton, and Tipper, 1991) to Bajocian. Recent zircon U-Pb analyses from near the Island Copper region also indicate younger ages (Friedman and Nixon, 1995). A rhyolite flow near Lemare Lake gave a zircon $^{207}Pb/^{206}Pb$ age of 202 ±3 Ma (earliest Sinemurian) whereas zircon from a rhyolite flow and a welded tuff in the Pemberton Hills area yielded Middle Jurassic (Bathonian-Bajocian) U-Pb ages of ca. 165 Ma and 169 Ma, respectively. Thus, paleontological and isotopic dates bracket the age of the Bonanza Group volcanic rocks between Sinemurian and Bajocian and 202-165 Ma, respectively.

New U-Pb Results

Twenty-one zircon separates from six plutonic rocks of the WCC and Island Intrusions were analyzed (DeBari et al., 1999). Locations of these samples are shown in Figure 1. A foliated WCC granodiorite body just southeast of Nitinat Lake yields a date of 190.3 ±0.7 Ma and is crosscut by an unfoliated WCC diorite dated at 174.2 ±0.4 Ma (sites 6 and 7). Another WCC sample, a mafic diorite from Sugsaw Lake just east of Barkley Sound (site 5) yields an age of 177.3 ±1.8 Ma. An Island Intrusion monzodiorite sample from the Port Alberni pluton (site 3) is essentially the same, with an age of 175.7 ±2.6 Ma. Another Island Intrusion quartz diorite sample east Nitinat Lake (site 4) yields a significantly younger age of 170.3 ±0.9 Ma. An Island Intrusion quartz diorite sample from the Nimpkish batholith (site 1) is younger still (168.6 ±5.3 Ma) and occurs along a northwesterly trend (in present-day coordinates) with the other two younger <171 Ma plutons.

Geochemical Relationships

Geochemical data from the WCC, Island Intrusions, and Bonanza Group volcanic rocks were collected from detailed sampling and mapping traverses in the Alberni Inlet area and the Broken Islands (DeBari et al., 1999). The majority of samples from each of the units contain between 50 and 60 wt.% SiO_2. They define a calcalkaline trend on an AFM diagram, and they belong to the medium-K island arc series (Fig. 2). They show the characteristic enrichment in large-ion lithophile (LIL) elements and depletion in high field strength (HFS) elements typical of arc rocks.

Plots of the major and trace elements against SiO_2 of rocks for all units display highly correlated colinear trends, strongly supportive of their comagmatic nature. Trace element trends show much more scatter, but Cr and Ni clearly decrease with increasing SiO_2, whereas the incompatible elements Ba, Zr, Rb, and Nb increase with increasing SiO_2. Contemporaneous mafic dikes within all units have <50 wt.% SiO_2 and 6.5-9.5 wt.% MgO and may be the closest representative of 'parental' mantle-derived mafic compositions still present within the arc.

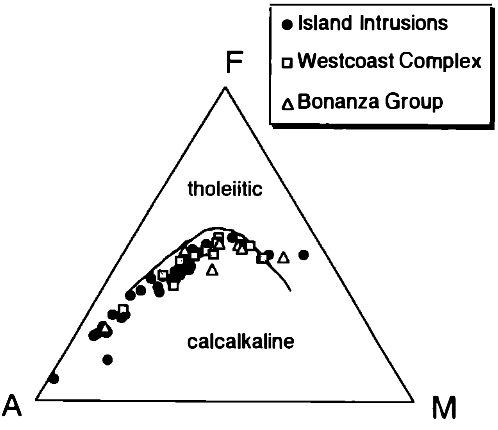

Figure 2. AFM plot of the Jurassic igneous rocks defines a distinctly calcalkaline trend.

Westcoast Crystalline Complex (WCC)

Chemical compositions of WCC plutonic rocks vary widely due to the wide array of lithologies. Most of the complex comprises hornblende gabbro and diorite with 49-60 wt.% SiO_2. Samples with less than 55 wt.% SiO_2 (hornblende gabbro) show significant scatter on variation diagrams, reflecting a probable cumulate origin. This idea is supported by convex REE patterns in these rocks that are suggestive of hornblende accumulation. Dioritic rocks uniformly display light-REE enriched patterns with abundances 30 to 40X chondrite and chondrite-normalized La_N/Yb_N of 2.8-3.5 (Fig. 3).

Tonalitic to trondhjemitic leucosomes associated with amphibolite melanosomes within the WCC display concave-upwards, light-REE enriched and heavy-REE depleted patterns indicative of melting of a hornblende-bearing source (i.e., amphibolite melansomes).

Very limited isotopic data exist for the rocks of the WCC. Initial $^{87}Sr/^{86}Sr$ isotopic ratios reported by Isachsen (1987) range from 0.7034 to 0.7037. These data preclude a source for the WCC that is derived entirely by melting of the Paleozoic Sicker Group rocks. The range in initial

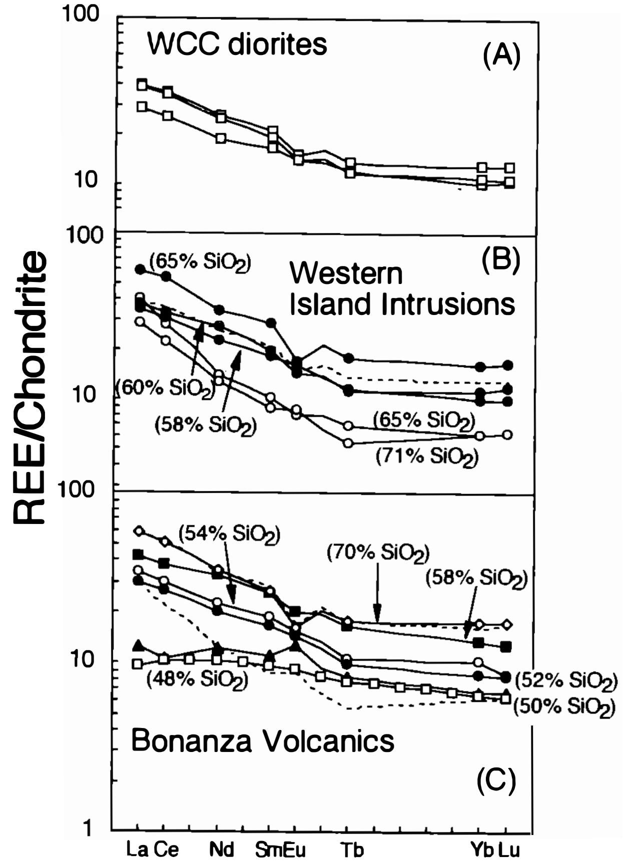

Figure 3. Plots of REE patterns for the plutonic and volcanic rocks of the Bonanza arc: (A) dioritic rocks of the Westcoast Crystalline Complex (WCC); (B) Western Island Intrusions – stippled line denotes REE patterns with the highest abundances in the WCC; (C) Bonanza Group volcanic rocks – stippled line represents REE patterns with the highest and lowest abundances in the Island Intrusions. Note that the REE pattern for the most silicic Bonanza volcanic sample is virtually identical to the pattern of the most silicic Island Intrusion sample.

$^{87}Sr/^{86}Sr$ for Sicker Group rocks during the Jurassic would have been from 0.7043 to 0.7052, which would have been too radiogenic to produce the initial $^{87}Sr/^{86}Sr$ compositions of the WCC rocks.

Island Intrusions

Island Intrusion plutons are dominantly metaluminous, medium-K, calcalkaline granitoids. They display a range of compositions from 53 to 65 wt.% SiO_2, but most commonly contain <60%. There is moderate enrichment of alkalis, with Na_2O ranging from 3 to 4 wt.% and K_2O from 1 to 3 wt.%. These plutonic rocks define a more colinear array on SiO_2 variation diagrams than do equivalent rocks from the WCC, especially for the trace elements. Some elements define distinctive trends for individual plutons but there is no consistent compositional change from west to east in samples with similar SiO_2 contents. The compositions are typical of volcanic-arc related granites worldwide.

Two types of REE patterns typify plutonic rock samples from the Island Intrusions (Fig. 3). Diorite and quartz diorite, regardless of silica content, are characterized by light REE abundances 10-60X chondrite values and by chondrite-normalized La_N/Yb_N of 3.3-3.7. These patterns are indistinguishable from WCC diorite and suggest an origin by fractionation of mantle-derived mafic magma. Tonalite and trondhjemite are characterized by concave-upward patterns, moderate enrichment of light REE (10-40X chondrite values), relative depletion of the heavy REE (5-7X chondrite values), and La_N/Yb_N ratios of 4.7-6.5. This latter pattern suggests derivation by melting of a hornblende-rich source (i.e., the amphibolitic basement to the WCC). Samples from the Port Alberni pluton have the most upwardly concave REE patterns and strongest heavy REE fractionation. These REE characteristics are closely similar to those exhibited by Late Jurassic (153-145 Ma) plutons in the southern Coast Belt (Cui and Russell, 1995a).

Strontium isotopic ratios for the Island Intrusions overlap those of the WCC. Initial $^{87}Sr/^{86}Sr$ values range from 0.7033 to 0.7039 for Vancouver Island and the Queen Charlotte Islands (QCI) (Samson et al., 1990; Andrew et al., 1991). The $_{Nd}$ values range from +4.7 to +5.6 (Samson et al., 1990) and +6.0 to +6.9 (Andrew et al., 1991) for Island Intrusions on Vancouver Island and +6.1 to +6.8 for 171-164 Ma plutons on QCI (Samson et al., 1990). The more primitive ranges are similar with those determined for Late Jurassic intrusions in the southern Coast Belt ($_{Nd}$=+5 to +6.6 and initial $^{87}Sr/^{86}Sr$=0.7033-0.7040) (Cui and Russell, 1995b).

Bonanza Group Volcanic Rocks

The Bonanza Group volcanic rocks show a wide range of compositions from basalt to rhyolite. The mafic endmembers are typically high-alumina basalts with 4-8 wt.% MgO. Their major and trace-element trends on SiO_2 variation diagrams overlap those of the Island Intrusions and WCC, except for higher Sr content at low wt.% SiO_2.

The REE patterns for the Bonanza Group volcanic rocks are slightly to moderately light-REE enriched, with abundances 8-60X chondrite and La_N/Yb_N of 1.9-3.4 (Fig. 2). Patterns for andesite and rhyolite are indistinguishable from diorite and quartz diorite of the WCC and Island Intrusions. However, the volcanic rocks do not have REE patterns that are clearly equivalent to the concave-upward patterns of the tonalite-trondhjemite samples from the Island Intrusions.

Isotopic ratios for the Bonanza Group volcanic rocks are indistinguishable from the Island Intrusions. They have a range of initial $^{87}Sr/^{86}Sr$ values from 0.7035 to 0.7041 and a range of $_{Nd}$ values from +3.7 to +6.9, with most samples between +5.8 and +6.6 (Samson et al., 1990; Andrew et al., 1991).

Discussion

Relationships Among the Jurassic Units

As discussed previously, the temporal and magmatic relationships between the WCC, the Island Intrusions, and the Bonanza Group volcanic rocks have not been clearly delineated. Our new field, petrologic, geochronologic, and geochemical data clarify many of the outstanding questions and expand the evidence for a comagmatic relationship between all of the units as follows:

Field evidence and phase relations: Contacts between Island Intrusions and WCC are entirely gradational. The more concordant and gradational contacts between WCC plutons and country rock, as well as higher metamorphic grade of country rock (comprising only Sicker Group rather than stratigraphically higher Karmutsen Formation), suggest that the WCC represents deeper crustal levels than the Island Intrusions. This observation is corroborated by Al-in-hornblende geobarometry along a NE-SW transect across the arc which yielded pressures of 2-3 kbar for hornblende in Island Intrusion plutonic rocks in the northeast and 3.5-5 kbar for hornblende in WCC plutonic rocks in the southwest (Parent and DeBari, 1995). The more mafic, deep-level plutons of the Island Intrusions are indistinguishable from the plutonic rocks of the WCC.

In all plutonic units there are shared similarities in intrusive relations and in petrographic textures involving the earliest formed minerals. There is an evolution in composition from dominantly mafic in the west (deeper-level) to heterogeneous intermediate and felsic compositions to the east (higher level) as well as a predictable petrographic evolution in the content, alteration, and textures of the later-formed minerals of these intrusions.

Geochronological constraints: Isotopic dating of the WCC and Island Intrusions, along with biochronological and geochronological data for the Bonanza Group volcanic rocks, provide a critical test of contemporaneity. Our new age dates show that the WCC had at least a bipartite history encompassing pre- to syn-deformational magmatism at 190 Ma and post-deformational magmatism at 174 Ma. The younger episode is coeval with dates from Island Intrusions proximal and distal to the WCC (177 Ma and 176 Ma, respectively). The 170-167 Ma dates for the Nimpkish batholith, Island Copper plutons, and an Island Intrusion in the study area represent an even younger episode of magmatism coeval with extensive 172-164 Ma plutonism in the Queen Charlotte Islands (Anderson and McNicoll, 1995; Anderson and Reichenbach, 1991) and the oldest plutons of the SW Coast Belt (Friedman and Armstrong, 1995) (Fig. 4).

Concerns that the Bonanza Group is older than the WCC and Island Intrusions have been addressed in previous studies (e.g., Archibald and Nixon, 1995). Recent discoveries of Aalenian fossils in the Bonanza Group (Poulton and Tipper, 1991) and U-Pb zircon ages as young as 169-165 Ma suggest that the age of the Bonanza Group extends from the Early Jurassic well into the Middle Jurassic. This age range clearly overlaps with that of the WCC and Island Intrusions (Fig. 4).

Geochemical and isotopic constraints: The chemical characteristics of the WCC, Island Intrusions, and Bonanza Group volcanic rocks completely overlap in major element, trace element, and REE geochemistry. The WCC is the most heterogeneous of the three units, which is expected considering the range of rock types that make up the unit, and the incorporation of pre-existing crust into its plutonic section. However, the compositions of the dioritic rocks of the WCC and the Island Intrusions are indistinguishable, which in turn are indistinguishable in composition from the intermediate volcanic rocks of the Bonanza Group. This is shown most clearly in the REE patterns. They completely overlap in abundances and shape. In general, the Island Intrusions display far less geochemical scatter than the WCC, and given their shallower position in the crust, probably represent WCC magmas that have become more uniformly mixed as they migrated upwards in the crust. The WCC, Island Intrusions, and Bonanza Group have overlapping Sr and Nd isotopic compositions, and these compositions are distinct from the Karmutsen Formation or Sicker Group rocks.

Comparison with Middle Jurassic Plutonism in Queen Charlotte Islands and SW Coast Belt

The Middle Jurassic magmatic interval identified in this study from Vancouver Island (177-168 Ma) is comparable to that from the Queen Charlotte Islands (Anderson and Reichenbach, 1991; Lewis et al., 1991) (Fig. 4). The San Christoval suite (172-171 Ma) resembles the Port Alberni pluton in modal composition, homogeneity, abundance of mafic inclusions, and prismatic amphibole habit. A younger (ca. 168-164 Ma), heterogeneous, and highly altered suite (Burnaby Island suite), occurs to the east of the San Christoval suite and has characteristics similar to the younger, more easterly, Island Intrusion plutons (171-168 Ma). There is no unit equivalent to the WCC on the Queen Charlotte Islands.

Some of the oldest plutonic rocks in the dominantly Cretaceous Coast Belt farther east are Middle to Late Jurassic in age (167-145 Ma; Friedman and Armstrong, 1995). The 171-168 Ma Vancouver Island plutons may represent the initiation of a northeastward sweep of arc magmatism from Vancouver Island to the Coast Belt.

Comparison with the Talkeetna Arc of Southern Alaska

One of the intriguing aspects of the Bonanza arc is its similarity in age, field setting, and lithology to the Talkeetna island arc in southern Alaska (Fig. 4). The Talkeetna arc is also located in the Insular Belt, it is Early to Middle Jurassic in age, and it exposes a disrupted arc crustal section (DeBari and Sleep, 1991). The volcanic rocks of the Talkeetna arc (Talkeetna Formation) are calcalkaline and contain interbedded sediments with Sinemurian to Toarcian fossils (Barker and Grantz, 1982) . The plutonic rocks are lithologically similar to the WCC and Island Intrusions (DeBari and Sleep, 1991) and have ^{40}Ar-^{39}Ar cooling ages of 175 to 180 Ma (Onstott et al., 1989).

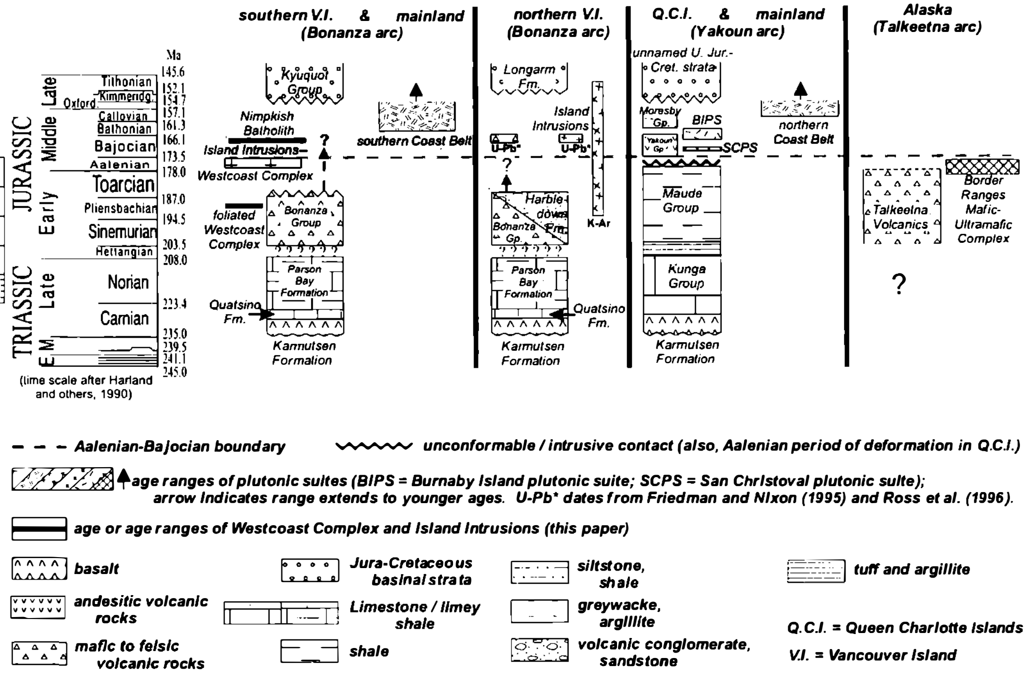

Figure 4. Time-stratigraphic sections which emphasize plutonic-volcanic assemblages for: southern Vancouver Island (Bonanza arc) and mainland (45°-50°) modified from Gabrielse and Brookfield (1992), Monger and McNicoll (1993), and Friedman and Armstrong (Friedman and Armstrong, 1995); northern Vancouver Island (Bonanza arc; 50°-51°) from Gabrielse and Brookfield (1992), Lyatsky and Haggart (1992), Friedman and Nixon (1995), and Ross et al. (1996); Queen Charlotte Islands (Yakoun arc) and mainland (52°-54°) modified from Woodsworth (1991), Lewis et al. (1991), Lyatsky and Haggart (1992), van der Heyden (1989), Anderson and Reichenbach (1991), and Anderson and McNicoll (1995); and the Talkeetna arc (north of 60°) as studied in the Tonsina district (DeBari and Sleep, 1991).

Both suites have abundant hornblende, especially in mafic compositions, and lack titanite in intermediate compositions. Both the Talkeetna and Bonanza arc sequences are bounded on their structurally deeper sides by Mesozoic accretionary complexes (Chugach Terrane and Pacific Rim Complex, respectively). Plafker et al. (1989) suggested that the Talkeetna arc was correlative with the Jurassic arc rocks on Vancouver Island and proposed a reconstruction that requires post-accretion displacement along strike-slip faults.

Conclusions

The Westcoast Crystalline Complex (WCC), the Island Intrusions, and the Bonanza Group volcanic rocks represent different crustal levels of a Jurassic island arc on Wrangellia. The arc extends the entire length of Vancouver Island and into the Queen Charlotte islands, covering an area of approximately 25,000 km^2 with an unknown thickness. The arc is calcalkaline and represents the high-alumina, intermediate-K island arc series.

The WCC represents the deepest exposed levels of the arc and displays a wide variety of lithologic types, including metamorphosed and migmatized segments of pre-existing crust (Sicker Group) and mafic to intermediate plutonic rocks representative of mantle-derived magmas. Stratigraphy of the pre-existing crust suggests that the deepest levels of the arc exposed represent depths of approximately 15 km, which correlates with hornblende barometry. Deformed, 190 Ma WCC plutons were crosscut by undeformed, 175 Ma plutons suggesting that at least two episodes of arc magmatism are represented in the Early to Middle Jurassic.

The higher levels of the Jurassic crust are represented by the plutons of the Island Intrusions. These plutonic rocks are less texturally heterogeneous than the WCC and also include granodiorite compositions. This homogeneity, and lack of migmatized country rock associated with the Island Intrusions, indicate that the crustal melting, mixing with mantle-derived magmas, and homogenization processes took place at deeper crustal levels (represented by the WCC) and were more or less complete before migration of magmas to shallower crustal levels.

The Bonanza Group volcanic rocks record effusive magmatism and mark the highest crustal level. They conformably overlie Triassic rocks and are intruded at their deepest levels by plutons of the Island Intrusions. They are chemically indistinguishable from the Island Intrusions.

The lithology, age, and field setting of the Bonanza arc are strikingly similar to the Yakoun arc in the Queen Charlotte Islands and the Talkeetna arc of southern Alaska. The Talkeetna arc may once have been continuous with the Bonanza arc but was subsequently offset by transcurrent faulting.

Acknowledgments

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References

ANDERSON, R. G., AND V. J. MCNICOLL. 1995. A note on U-Pb dating of Middle Jurassic plutonic suites: Cumshewa Head pluton, southeastern Moresby Island, Queen Charlotte Islands, British Columbia. Current Research 95-A, Geological Survey of Canada, p. 91-96.

____, AND I. REICHENBACH. 1991. U-Pb and K-Ar framework for Middle to Late Jurassic (172-158 Ma) and Tertiary (46-27 Ma) plutons in Queen Charlotte Islands, British Columbia, p. 59-87. *In* G. J. Woodsworth (ed.), Evolution and Hydrocarbon Potential of the Queen Charlotte Basin, British Columbia. Geological Survey of Canada, Paper 90-10.

ANDREW, A., R. L. ARMSTRONG, AND D. RUNKLE. 1991. Neodymium-strontium-lead isotope study of Vancouver Island igneous rocks. Canadian Journal of Earth Sciences, 28:1744-1752.

ARCHIBALD, D. A., AND G. T. NIXON. 1995. 40Ar/39Ar geochronometry of igneous rocks in the Quatsino-Port McNeill map area, northern Vancouver Island (92L/12, 11). B. C. Geological Survey Geological Fieldwork 1994, Paper 1995-1, 49-59.

ARMSTRONG, R. L. A. 1988. Mesozoic and early Cenozoic magmatic evolution of the Canadian Cordillera. Geological Society of America Special Paper, 218:55-91.

BARKER, F., AND A. GRANTZ. 1982. The Talkeetna Formation in the southeastern Talkeetna Mountains, southern Alaska, an Early Jurassic andesitic island arc. Geological Society of America Abstracts with Programs, 14:147.

CUI, Y., AND J. K. RUSSELL. 1995a. Magmatic origins of calc-alkaline intrusions from the Coast Plutonic Complex, southwestern British Columbia. Canadian Journal of Earth Sciences, 32:1643-1667.

____, AND J. K. RUSSELL. 1995b. Nd, Sr and Pb isotopic systematics of the southern Coast Plutonic Complex, southwestern British Columbia. Geological Society of America Bulletin, 107(2):127-138.

DEBARI, S. M.,R. G. ANDERSON, AND J. K. MORTENSEN. 1999. Correlation Amongst Lower to Upper Crustal Components in an Island Arc: the Jurassic Bonanza arc, Vancouver Island, Canada. Canadian Journal of Earth Sciences, in press.

____ AND N. H. SLEEP. 1991. High-Mg, low-Al bulk composition of the Talkeetna island arc, Alaska: Implications for primary magmas and the nature of arc crust. Geological Society of America Bulletin, 103:37-47.

FRIEDMAN, R. M., AND R. L. ARMSTRONG. 1995. Jurassic and Cretaceous geochronology of the southern Coast Belt, B. C., 49°-51°N, p. 95-139. *In* D. M. Miller and C. Busby (eds.), Jurassic magmatism and tectonics of the North American Cordillera. Geological Society of America Special Paper, 299.

___, AND G. T. NIXON. 1995. U-Pb zircon dating of Jurassic porphyry Cu(-Au) and associated acid sulphate systems, northern Vancouver Island, British Columbia, Geological Association of Canada - Mineralogical Association of Canada Annual Meeting, Victoria, B. C., pp. A-34.

GABRIELSE, H., AND A. J. BROOKFIELD. 1992. Figure 1. Correlation chart for the Canadian Cordillera. *In* H. Gabrielse and C. J. Yorath (eds.), Geology of the Cordilleran Orogen in Canada, v. 4. Geology of Canada. Geological Survey of Canada.

ISACHSEN, C. E. 1987. Geology, geochemistry, and cooling history of the Westcoast Crystalline Complex and related rocks, Meares Island and vicinity, Vancouver Island, British Columbia. Canadian Journal of Earth Sciences, 24:2047-2064.

LEWIS, P. D., J. W. HAGGART, R. G. ANDERSON, C. J. HICKSON, R. I. THOMPSON, J. R. DIETRICH, AND K. M. M. ROHR. 1991. Triassic to Neogene geological evolution of the Queen Charlotte region. Canadian Journal of Earth Sciences, 28(6):854-869.

LYATSKY, H. V., AND J. W. HAGGART. 1992. Queen Charlotte basin off western Canada may hold oil, gas potential. Oil and Gas Journal(August):53-56.

MONGER, J. W. H., AND V. J. MCNICOLL. 1993. New U-Pb dates from southwestern Coast Belt, British Columbia. Radiogenic Age and Isotopic Studies, Report 7, Geological Survey of Canada, Paper 93-2, 119-126.

MULLER, J. E., B. E. B. CAMERON, AND K. E. NORTHCOTE. 1981. Geology and mineral deposits of Nootka Sound Map-Area, Vancouver Island, British Columbia. Geological Survey of Canada, Paper 80-16.

___ AND D. J. T. CARSON. 1969. Geology and mineral deposits of the Alberni area, British Columbia. Geological Survey of Canada, Paper 68-50, 35pp.

___, K. E. NORTHCOTE, AND D. CARLISLE. 1974a. Geology and mineral deposits of Alert Bay - Cape Scott map area. Vancouver Island, British Columbia. Geological Survey of Canada, Paper 74-8, 77.

___, R. K. WANLESS, AND W. D. LOVERIDGE. 1974b. A Paleozoic zircon age of the Westcoast Crystalline Complex of Vancouver Island, British Columbia. Canadian Journal of Earth Sciences, 11:1717-1722.

ONSTOTT, T. C., V. B. SISSON, AND D. L. TURNER. 1989. Initial argon in amphiboles from the Chugach Mountains, southern Alaska. Journal of Geophysical Research, 94(B4):4361-4372.

PARENT, L. A., AND S. M. DEBARI. 1995. Aluminum-in-hornblende barometry applied to a proposed island arc crustal section from the Jurassic Bonanza arc, Vancouver Island, Canada. Geological Society of America Abstracts with Programs, 27:A-46.

PARRISH, R. R., AND V. J. MCNICOLL. 1992. U-Pb age determinations from the southern Vancouver Island area, British Columbia, p. 79-86 (ed.), Radiogenic Age and Isotopic Studies: Report 5; Geological Survey of Canada, Paper 91-2.

PLAFKER, G., W. J. NOKLEBERG, AND J. S. LULL. 1989. Bedrock geology and tectonic evolution of the Wrangellia, Peninsular, and Chugach terranes along the trans-Alaska crustal transect in the Chugach Mountains and southern Copper River basin, Alaska. Journal of Geophysical Research, 94(B4):4255-4295.

POULTON, T. P., AND H. W. TIPPER. 1991. Aalenian ammonites and strata of western Canada. Geological Survey of Canada, Bulletin 411, p. 1-71.

ROSS, K. V., R. M. FRIEDMAN, K. M. DAWSON, AND C. H. B. LEITCH. 1996. U-Pb Zircon ages of the Island Copper deposit intrusions, northern Vancouver Island, British Columbia, p. 111-117. *In* Current Research 1996-A, Geological Survey of Canada.

ROSS, G. M., AND R. R. PARRISH. 1991. Detrital zircon geochronology of metasedimentary rocks in the southern Omineca Belt, Canadian Cordillera. Canadian Journal of Earth Sciences, 28:125-1270.

SAMSON, S. D., P. J. PATCHETT, G. E. GEHRELS, AND R. G. ANDERSON. 1990. Nd and Sr isotopic characterization of the Wrangellia terrane and implications for crustal growth of the Canadian cordillera. Journal of Geology, 98:749-762.

VAN DER HEYDEN, P. 1989. Jurassic magmatism and deformation: Implications for the evolution of the Coast Plutonic Complex. Geological Society of America Abstracts with Programs, 21:153.

WOODSWORTH, G. J. (Editor). 1991. Evolution and Hydrocarbon Potential of the Queen Charlotte Basin, British Columbia. Geological Survey of Canada, Paper 90-10.

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Zone Boundaries and Subzones of the Transversarium Ammonite Zone (Oxfordian, Late Jurassic) in the Reference Section of the Zone, Northern Switzerland

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Keywords: Biozonation, Ammonites, Oxfordian, Switzerland

Abstract: The revised Transversarium ammonite Zone by Oppel and Waagen, 1866, is the oldest ammonite biozone that has ever been defined in detail both litho- and biostratigraphically. The zone is redefined in this paper based on more than 3000 ammonites that have been collected *in situ* from several systematic excavations in northern Switzerland since 1962. A reference section is given, and the zone is subdivided into three subzones. The zone as it is redefined here is proposed to be given the rank of an international standard zone. The redefined Transversarium Zone can be adopted in low paleolatitudes almost worldwide.

Introduction

The Transversarium Zone is the oldest well-documented ammonite zone. A Transversarium Zone *sensu lato* was first proposed by Oppel (1863, p. 165). But Oppel stated already at that time that "the proper region of *Amm. transversarius*" was the Birmensdorf Member that had been defined the same year by Moesch (1863, p. 160) in Canton Aargau, northern Switzerland. In a subsequent manuscript, Oppel gave the zone its final definition, but he died before he could publish his paper. Waagen took care of the final editing of the detailed manuscript, and it appeared under the joint authorship of Oppel and Waagen in 1866. In this paper (p. 244), the authors designated the Birmensdorf Member of Canton Aargau with its ammonite fauna as the type unit of the Transversarium Zone. This zone can be recognized in a very wide area. It has been documented in many papers not only in Central and Southern Europe, but also in Central Asia (Besnosov and Mitta, 1995), North Africa (Sapunov, 1973), Madagascar (Collignon, 1959) and Chile (Gygi and von Hillebrandt, 1991). The zone has recently been dated radiometrically by Fischer and Gygi (1989) in northern Switzerland. A controversy about the interpretation of the Transversarium Zone started with a paper by Callomon (1964) and continues to the present. The Transversarium Zone will probably one day be recognized world-wide in low paleolatitudes. Therefore it is necessary to reconsider it with the aim of defining it as a standard zone and achieving stability in stratigraphic nomenclature.

Remarks on the Birmenstorf Member in Canton Aargau

Oppel and Waagen (1866, p. 244) wrote that the type locality of the Transversarium Zone was at Birmensdorf in Canton Aargau. This village is now spelled Birmenstorf and should not be confused with Birmensdorf in the adjacent Canton Zürich (Fig. 1). Gygi (1977, Fig. 2) mapped the vineyards around the locality called Nettel northeast of Birmenstorf, Canton Aargau, at the scale of 1:5000, and found no indication that an outcrop of the Birmenstorf Member ever existed near this village. Moesch (1863) probably named the unit because of the abundant ammonites he found lying on the weathered soil above the Birmenstorf Member in the vineyards of Birmenstorf. For this reason, Gygi (1969, p. 64) proposed the highly fossiliferous section RG 60 of the Birmenstorf Member in the Eisengraben cleft near Gansingen in Canton Aargau as the type section of the member. The fossil excavations RG 210 and RG 225 in the Eisengraben and the excavation RG 230 north of Eisengraben (Fig. 2), as published by Gygi (1977, pl. 11), provide for a composite reference section of the Transversarium Zone (Gygi, 1977, p. 517) in the sense of Hedberg (1976, p. 58). Oppel and Waagen (1866, p. 245) stated that the Transversarium Zone begins in Canton Aargau above the "iron ore with *Amm. Lamberti* and *cordatus*". This is bed 7 of section RG 210 in Figure 2. Gygi (1977, p. 454) named this characteristic unit the Schellenbrücke Bed. The lowermost bed of the Birmenstorf Member and therefore of the Transversarium Zone in section RG 210 (Fig. 2) is bed 8. This is a regional marker bed. The difficult delineation of the top of the Birmenstorf Member beneath the Effingen Member was discussed by Gygi (1969, p. 66).

The Ammonite Fauna of the Birmenstorf Member

Since 1962, Sylvia Gygi and the author have collected more than 3000 ammonites from the Birmenstorf Member of Canton Aargau and its time equivalents in Canton Schaffhausen. Specimens were collected *in situ* from several excavations sampled bed by bed in northern Switzerland. 1179 ammonites have been prepared and are available for taxonomic and biostratigraphic purposes. The principal ammonite taxa and their vertical ranges were published by Gygi (1977). Almost all of the ammonites cited in this paper are kept in the Museum of Natural History Basel, Switzerland (MNHB). Each specimen from this museum has an individual number with the prefix J.

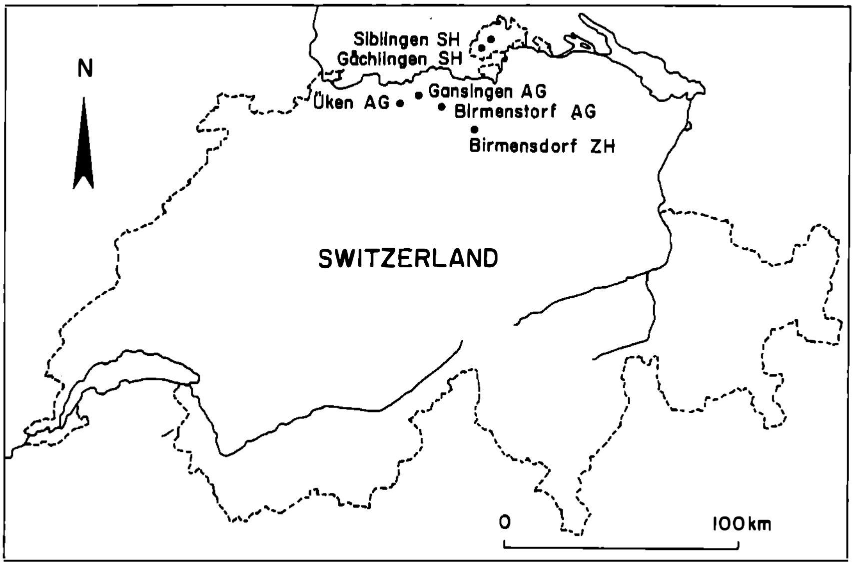

Figure 1. Location of the townships mentioned in the text. The following abbreviations are used in the figure: AG for Canton Aargau, SH for Canton Schaffhausen, and ZH for Canton Zürich.

The lowermost bed of the Birmenstorf Member, which is the oldest bed of the Transversarium Zone as defined by Oppel and Waagen (1866), contains *Cardioceras (Plasmatoceras) tenuistriatum* Borissjak, as figured by Gygi and Marchand (1982, pl. 12, Fig. 2) from excavation RG 208 at Üken near Herznach in Canton Aargau. This taxon is indicative of the Densiplicatum Subzone as proposed by Gygi (1991, Fig. 8). *Perisphinctes (Dichotomosphinctes) antecedens* Salfeld, the index of the overlying subzone, has been found in the same excavation and in the same bed with a thickness of less than 10cm, that is consequently condensed. This regional marker bed in Canton Aargau contains few ammonites. Ammonites of the Densiplicatum and Antecedens Subzones are abundant in two regional marker beds called, respectively, the Mumienmergel Bed and the Mumienkalk Bed (Gygi, 1977, p. 455), which have been excavated for ammonites at several sites in Canton Schaffhausen to the northeast of Canton Aargau. *Cardioceras (Subvertebriceras) densiplicatum* Boden, the index of the Densiplicatum Subzone, occurs in the Mumienmergel Bed, which is in bed 7 of excavation RG 212 (Fig. 2) near Siblingen, Canton Schaffhausen (Gygi and

Marchand, 1982, pl. 11, Figs. 5-6). The first *Gregoryceras (Gregoryceras) transversarium* (Quenstedt) also appear in the condensed Mumienmergel Bed (Gygi 1977, p. 491, pl. 6, Fig. 3), together with *Perisphinctes (Dichotomosphinctes) antecedens* Salfeld, from bed 7 of section RG 212 in Figure 2.

A well-preserved specimen of *Perisphinctes (Dichotomosphinctes) antecedens* Salfeld, from Siblingen SH was found in the lower part of the Mumienkalk Bed, no. 16a, in excavation RG 207 (Fischer and Gygi, 1989, Fig. 5D). The vertical range of this taxon extends from the Mumienmergel Bed to the top of the Mumienkalk Bed. The vertical range of the next subzonal index, *Perisphinctes (Dichotomosphinctes) luciaeformis* Enay, overlaps with that of *P. (D.) antecedens*. The oldest representatives of *P. (D.) luciaeformis* were found in the lower Mumienkalk Bed (e. g., the complete specimen from bed 14a of excavation RG 81b near Gächlingen SH). A well-preserved *Perisphinctes (Dichotomosphinctes) luciaeformis* Enay, was figured by Fischer and Gygi (1989, Fig. 6D) from the lower, glauconitic part of bed 10 in excavation RG 212 at Siblingen (Fig. 2 in this

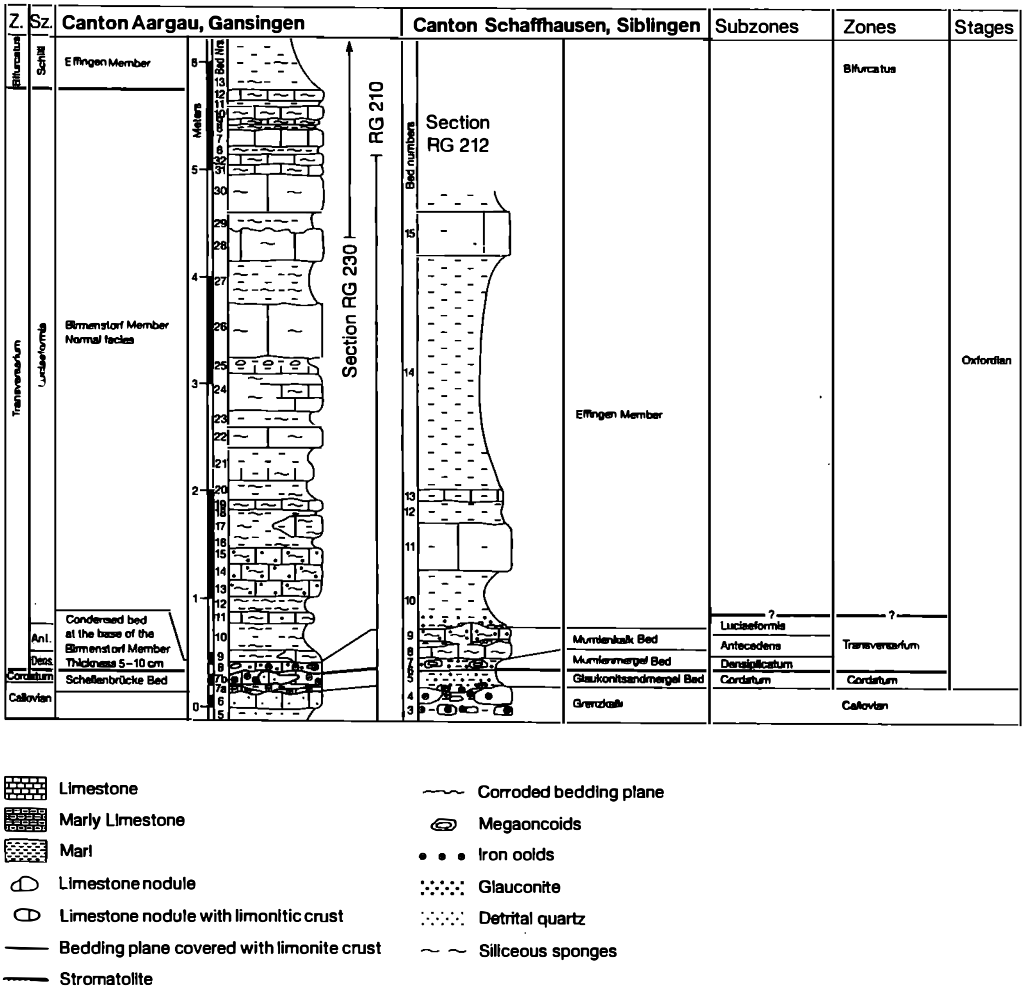

Figure 2. The type section of the Birmenstorf Member which is also the reference section of the Transversarium Zone near Gansingen, Canton Aargau, and a section of the Mumienmergel Bed and the Mumienkalk Bed near Siblingen, Canton Schaffhausen.

paper). This taxon is also represented throughout the normal facies of the Birmenstorf Member in Canton Aargau. The type of the taxon is from southeastern France (Enay, 1966). Further representatives from Spain have been figured by Melendez (1989).

The Subzones of the Transversarium Zone

Densiplicatum Subzone

Cardioceras (Subvertebriceras) densiplicatum Boden is among the oldest ammonites of the Transversarium Zone. As far as we know, it lived during the time interval between the Cordatum and Antecedens Subchrons. Mainly the following names have been proposed for strata equivalent to this time interval:

- *tenuicostatum* beds, Salfeld (1913, tab. 1)
- Vertebrale Subzone, Arkell (1947, p. 98)
- Densiplicatum Subzone, Gygi (1991, p. 12 and Fig. 8)

The Tenuicostatum Subzone has never gained wide acceptance. The Vertebrale Subzone is commonly used (e.g. Cariou et al., 1991). However, its zonal index occurs in Europe probably only in England and in the Baltic, and is thus unsuitable for Switzerland. A Densiplicatum Zone was introduced by Sykes and Callomon (1979, p. 845). The index, *Cardioceras (Subvertebriceras) densiplicatum* Boden, is recorded from the Baltic (Boden, 1911, pl. 1, Fig. 14), England (Arkell, 1942, pl. 52, Fig. 4), southern France (Bourseau, 1977, pl. 8, Figs. 5 and 10) and Switzerland (Gygi and Marchand, 1982, pl. 11, Figs. 5 and 6). The Densiplicatum Subzone is the only one of the three that can be recognized in Central Europe. It is here defined to be equivalent to the vertical range of the index.

Antecedens Subzone

The Antecedens Subzone as proposed by Arkell (1947, p. 98) is now used by most authors and is adopted here in its original sense.

Luciaeformis Subzone

The interval between the Antecedens and Schilli Subzones has been called the Parandieri Subzone by Tintant (1958, p. 2505). It is named after *Perisphinctes (Perisphinctes) parandieri* De Loriol. Only a few complete specimens of this very large taxon are known (see Enay, 1966). No complete adult has ever been found in Switzerland. A substitute must therefore be found. *Perisphinctes (Dichotomosphinctes) luciaeformis* Enay is the best choice, because its vertical range in northern Switzerland fits almost exactly into the interval between the Antecedens and the Schilli Subzones. The taxon has already been suggested as a subzonal index by Melendez (1989, Fig. 22). It is of medium size with a maximum diameter around 150mm. Near-complete, readily identified specimens are therefore easier to find than representatives of *P. (P.) parandieri*. The oldest *P. (D.) luciaeformis* in the collection of the Museum of Natural History Basel is the complete adult, J 24532, from the lower part of the Mumienkalk Bed (no. 14a in excavation RG 81b near Gächlingen, Canton Schaffhausen; see above). Among the youngest representatives is specimen J 24157, complete with peristome and lappets, from the upper Birmenstorf Member in excavation RG 230 north of Eisengraben near Gansingen, Canton Aargau. Another specimen, J 23652, from the condensed glauconitic marl at the base of the Effingen Member in Canton Schaffhausen, bed no. 10 in excavation RG 212 near Siblingen (Fig. 2), has been figured by Fischer and Gygi (1989, Fig. 6D).

Remarks on the Schilli Subzone

Oppel and Waagen (1866, p. 245) stated that the upper boundary of the Transversarium Zone in the type region was at the boundary between the Birmenstorf and the Effingen Members. Gygi (1969, p. 66) showed that the boundary between the Birmenstorf and Effingen Members is not easy to define. Gygi *et al.* (1979, Fig. 3) certainly drew the boundary too high and assigned bed 42 of section RG 276 with *Larcheria schilli* (Oppel) to their Birmenstorf Member. Oppel and Waagen (1866, p. 245) included *Larcheria schilli* (Oppel) in their list of ammonites of the Transversarium Zone. It is important to note that *Gregoryceras (Gregoryceras) transversarium* (Quenstedt) has never been found together with *Larcheria schilli* (Oppel) in the type region of the Transversarium Zone. *Larcheria schilli* is younger than the last *Gregoryceras transversarium.* This is at variance with the range chart of Figure 68 by Enay (1966) that indicates that the two taxa overlap. It is well-known that the vertical ranges of taxa may vary from region to region. The vertical ranges of the taxa in the type region of the Transversarium Zone are to be regarded as critical. It is proposed here to place the upper boundary of the Transversarium Zone at the top of the last occurrence of *Gregoryceras transversarium* just as the top of the Cordatum Zone is defined by the last occurrence of *Cardioceras cordatum.* The Schilli Subzone is therefore assigned here to the Bifurcatus Zone (Fig. 3).

Canton Aargau	**Canton Schaffhausen**	**Subzones**	**Zones**
Effingen Member	Effingen Member	Schilli	Bifurcatus
Birmenstorf Member Normal facies	Basal glauconitic marl	Luciaeformis	Transversarium
Birmenstorf Member condensed bed at the base	Mumienkalk Bed	Antecedens	
	Mumienmergel Bed	Densiplicatum	
Schellenbrücke Bed	Glaukonitsandmergel Bed	Cordatum	Cordatum

Figure 3. Boundaries and subzones of the Transversarium Standard Zone in northern Switzerland.

Conclusions

Systematic excavations for ammonites in the type region of the Transversarium Zone in northern Switzerland lead to the following conclusions (Fig. 3):

- The base of the Transversarium Zone is by the original definition at the base of the Densiplicatum Subzone which is time-equivalent with the Vertebrale Subzone in the Plicatilis Zone of Southern England. The Transversarium Zone therefore begins at the top of the Cordatum Zone like the Plicatilis Zone.

- The vertical range of the zonal index *Gregoryceras transversarium* is from the base of the Antecedens Subzone to the top of the Luciaeformis Subzone.

- The upper boundary of the Transversarium Zone should be drawn in the type region at the top of the vertical range of *Gregoryceras transversarium* which there coincides with the base of the Schilli Subzone.

- The Transversarium Zone in the sense of Oppel and Waagen (1866) should now be given the rank of an international standard zone. This will probably one day be recognized world-wide in low paleolatitudes.

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References

ARKELL, W. J. 1935-48. A monograph on the ammonites of the English Corallian Beds. Palaeontographical Society, London. 420 p.

____. 1947. The geology of Oxford. Clarendon Press, Oxford. 267 p.

BESNOSOV, N. V. AND V. V. MITTA. 1995. Upper Jurassic ammonitids and black shales in Central Asia. Moskva (in Russian).

BODEN, K. 1911. Die Fauna des Unteren Oxford von Popilany in Litauen. Geologische und Palaeontologische Abhandlungen, Neue Folge 10(2):125-199.

BOURSEAU, J.-P. 1977. L'Oxfordien moyen à nodules des "Terres Noires" de Beauvoisin (Drôme). Nouvelles Archives du Muséum d'Histoire naturelle de Lyon 15:1-116.

CALLOMON, J. H. 1964. Notes on the Callovian and Oxfordian Stages, p. 269-291. *In* P. L. Maubeuge (ed.), Colloque du Jurassique, Luxembourg 1962, Volume des Comptes Rendus et Mémoires, Institut Grand-Ducal, Section des Sciences naturelles, physiques et mathématiques. St-Paul, Luxembourg.

CARIOU, E., F. ATROPS, P. HANTZPERGUE, R. ENAY, AND M. RIOULT. 1991. Oxfordien. Abstracts 3rd International Symposium on Jurassic Stratigraphy, Poitiers, p. 132.

COLLIGNON, M. 1959. Atlas des fossiles caractéristiques de Madagascar, Part 4 (Argovien-Raracien). Service Géologique, Tananarive.

ENAY, R. 1966. L'Oxfordien dans la moitié sud du Jura Français. Nouvelles Archives du Muséum d'Histoire naturelle de Lyon 8:1- 624.

FISCHER, H. AND R. GYGI. 1989. Numerical and biochronological time scales correlated at the ammonite subzone level; K-Ar, Rb-Sr ages, and Sr, Nd, and Pb sea-water isotopes in an Oxfordian (Late Jurassic) succession of northern Switzerland.Geological Society of America Bulletin, 101:1584-1597.

GYGI, R. A. 1969. Zur Stratigraphie der Oxford-Stufe (Oberes Jura-System) der Nordschweiz und des süddeutschen Grenzgebietes. Beiträge zur Geologischen Karte der Schweiz, Neue Folge 136:1-123.

____. 1977. Revision der Ammonitengattung *Gregoryceras* (Aspidoceratidae) aus dem Oxfordian (Oberer Jura) der Nordschweiz und von Süddeutschland. Taxonomie, Phylogenie, Stratigraphie. Eclogae Geologicae Helvetiae, 70:435-542.

____. 1991. Die vertikale Verbreitung der Ammonitengattungen *Glochiceras*, *Creniceras* und *Bukowskites* im Späten Jura der Nordschweiz und im angrenzenden Süddeutschland. Stuttgarter Beiträge zur Naturkunde, Serie B, 179:1-41.

____, AND A. VON HILLEBRANDT. 1991. Ammonites (mainly *Gregoryceras*) of the Oxfordian (Late Jurassic) in northern Chile and time-correlation with Europe. Schweizerische Paläontologische Abhandlungen 113:135-185.

____, AND D. MARCHAND. 1982. Les faunes de Cardioceratinae (Ammonoidea) du Callovien terminal et de l'Oxfordien inférieur et moyen (Jurassique) de la Suisse septentrionale: Stratigraphie, paléoécologie, taxonomie préliminaire. Geobios 15:517-571.

____, S. M. SADATI, AND A. ZEISS. 1979. Neue Funde von *Paraspidoceras* (Ammonoidea) aus dem oberen Jura von Mitteleuropa - Taxonomie, Ökologie, Stratigraphie. Eclogae Geologicae Helvetiae 72:897-952.

HEDBERG, H. D. (ed.) 1976. International stratigraphic guide. Wiley, New York, 200 p.

MELENDEZ, G. 1989. El Oxfordiense en el sector central de la Cordillera Iberica (Provincias de Zaragoza y Teruel). Institucion Fernando el catolico, Zaragoza, 418 p.

MOESCH, C. 1863. Vorläufiger Bericht über die Ergebnisse der im Sommer 1862 ausgeführten Untersuchungen im Weissen Jura der Kantone Solothurn und Bern. Verhandlungen der Schweizerischen Natfurforschenden Gesellschaft Luzern 1862:156-168.

OPPEL, A. 1863. Ueber jurassische Cephalopoden. Fortsetzung. Palaeontologische Mittheilungen aus dem Museum des koeniglichen Bayerischen Staates 3:163-266.

____, AND W. WAAGEN. 1866. Über die Zone des Ammonites transversarius. Geognostisch-paläontologische Beiträge 1:207-316.

SALFELD, H. 1913. Die Gliederung des oberen Jura in Nordwesteuropa. Neues Jahrbuch für Mineralogie, Geologie und Paläontologie, Beilageband 37:125-246.

SAPUNOV, I. G. 1973. Ammonites de l'Oxfordien de la partie occidentale des Hauts Plateaux (Algérie). Publications du Service Géologique de l'Algérie, Nouvelle Série, Bulletin 44: 101-137.

SYKES, R. M. AND J. H. CALLOMON. 1979. The *Amoeboceras* zonation of the Boreal Upper Oxfordian. Palaeontology 22:839-903.

TINTANT, H. 1958. Sur la stratigraphie de l'Oxfordien supérieur aux environs de Dijon (Côte-d'Or). Comptes Rendus de l'Académie des Sciences (Paris) 246:2504-2507.

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Aalenian of the Zambujal de Alcaria Section (Central Lusitanian Basin; Portugal)

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Keywords: Aalenian, Central Lusitanian Basin, Ammonites, Biostratigraphy, Paleobiogeography

Abstract: A biostratigraphic zonal scheme for the Aalenian of the Zambujal de Alcaria section (central Lusitanian Basin) based on ammonite distribution is presented. The section is located at the south of Fátima plateau (central Portugal) and corresponds to a reference section for the Aalenian of the central Lusitanian Basin.

Aalenian sediments outcrop in a wide N-S cliff near Zambujal de Alcaria village. They are mainly represented by marl and marly limestone rhythmites, becoming increasingly calcareous upwards. The section may attain up to 70m in thickness and contain an abundant and diverse ammonite fauna.

In this section it was possible to recognize the following biostratigraphic units: Opalinum Zone (Opalinum and Comptum subzones), Bradfordensis Zone (Bradfordensis and Gigantea subzones), and Concavum Zone (Concavum and Limitatum subzones). As in other areas of the Basin, the Murchisonae Zone is absent due to a regional lacuna. The lower boundary of the Aalenian (with the Aalensis Zone) and the upper boundary of the Aalenian (with the Discites Zone) have also been established. A total of 300 well preserved ammonites have been identified.

The ammonite assemblages include North European taxa (Leioceratinae, Graphoceratinae) as well as typical Mediterranean ones (Hammatoceratinae, Grammoceratinae, Erycitidae), whose occurrences can be easily referred to the standard zonal scheme. These ammonite assemblages allow accurate correlation both with other areas of the Lusitanian Basin and the Iberian basins.

Introduction

The Zambujal de Alcaria section is located in central Portugal, 5km to SSW of Porto de Mós village (coordinates: X=516.800; Y=4389.275; h=320m, Serv. Cart. Exército, 1968) (Fig. 1).

Previous work on this section includes that of Choffat (1880), Perrot (1965), Perrot and Mouterde (1957), Ruget-Perrot (1961), Zbyszewski et al. (1974), Azeredo (1993), and Duarte (1995).

Ammonite Assemblages and Biostratigraphy

The section is about 70m thick and includes fossiliferous gray limestones interbedded with gray marls, in shallowing upwards sequences, increasingly calcareous towards the top. Fossil content is mainly ammonites, bivalves, gastropods and brachiopods associated with intense bioturbation and trace fossils (*Zoophycos* and *Chondrites*) (Fig. 2).

The ammonite assemblages recognized in the section allow recognition of the Toarcian-Aalenian (bed ZB 31) and Aalenian-Bajocian boundaries (bed ZA 195), and also the following biostratigraphic units based on occurrences and ranges of ammonoid taxa:

Aalensis Zone; Aalensis Subzone (ZB 12 to ZB 29; 4.5m)

Defined by first occurrence of *Pleydellia aalensis* (Zieten); *P. lotharingica* (Branco), *Pleydellia* sp., *Cotteswoldia limatula* Buckman, *C. egena* Buckman and *Cotteswoldia* sp. may also be found.

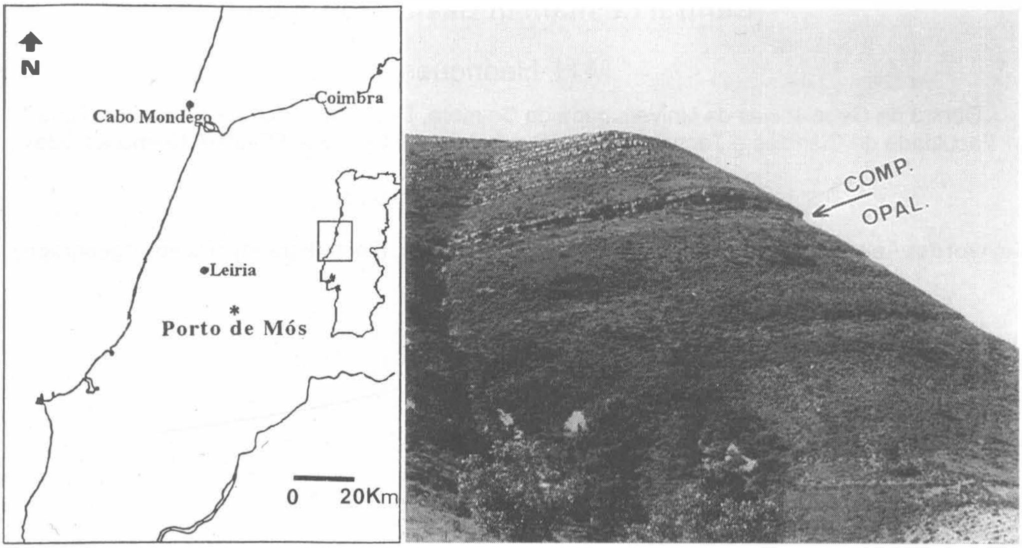

Figure 1. Geographic location and general view of the Zambujal de Alcaria section (Central Lusitanian Basin), including the boundary between the Opalinum and Comptum Subzones.

Opalinum Zone; Opalinum Subzone (ZB 31 to ZA 4; 15m)

Based on first occurrence of *Leioceras opalinum* (Reinecke), which appears associated with *L. costosum* (Quenstedt), *Cypholioceras* sp., *Planammatoceras* sp., *Bredyia* sp. and *Vacekia* sp.

Opalinum Zone; Comptum Subzone (ZA 5 to ZA 93; 17.5m; Pl. 1.1-1.3)

Recognised by first occurrence of *Cypholioceras comptum* (Reinecke); it is also possible to recognize *C. lineatum* (Buckman), *Cypholioceras* sp., *Leioceras* sp., *Tmetoceras scissum* (Benecke), *Tmetoceras* sp., *Vacekia* sp., *Bredyia* sp., *Erycites fallifax* Arkell, *Planammatoceras* sp. and *Spinammatoceras (?)* sp.

Bed 94 is the base of a marl to limestone, shallowing upwards sequence which includes at its top (bed 95) the first representatives of *Brasilia*. The last *Leioceras* occurs in the sequence just beneath. So, the surface of bed 93 represents a stratigraphic gap corresponding to the Murchisonae Zone (Haugi and Murchisonae Subzones), which is also absent in other parts of the Lusitanian Basin (Henriques, 1992; 1995; Henriques et al., 1996).

Bradfordensis Zone; Bradfordensis Subzone (ZA 94 to ZA 155; 13.5m; Pl.1.4-1.5)

Recognised by the occurrence of *Brasilia tutcheri* (Buckman), *B. umbilicata* (Buckman), *Brasilia* sp., *Ludwigella* ? *rudis* (Buckman), *Ludwigella* sp., *Vacekia stephensi* (Buckman), *Fontannesia* sp., *Abbasites* sp., *Bredyia diadematoides* (Mayer), *Bredyia* sp., *Erycites* sp., *Planammatoceras* sp. and *Tmetoceras* sp.

Bradfordensis Zone; Gigantea Subzone (ZA 156 to ZA 175; 3.9m; Pl. 1.6-1.7)

Based on first occurrence of *Brasilia gigantea* (Buckman) associated with *B. similis* (Buckman), *B. tutcheri* (Buckman), *B. umbilicata* (Buckman), *Brasilia* sp., *Ludwigella rudis* (Buckman), *L. cornu* (Buckman), *Ludwigella* sp. and *Malladaites* sp.

Concavum Zone; Limitatum Subzone (ZA 176 to ZA 193; 4.5m; Pl. 2.1-2.3)

Recognised by first appearance of *Graphoceras limitatum* (Buckman), which occurs in bed 176 immediately below the last representatives of *Ludwigella cornu* (Buckman) and *Brasilia gigantea* (Buckman) (bed ZA 175). The Concavum Subzone is absent due to a fault located between beds 175 and 176. The Limitatum Subzone is also characterized by the occurrence of *Graphoceras* *formosum* (Buckman), *G. decorum* (Buckman), *Braunsina compacta* (Buckman), *B. contorta* (Buckman), *B. aspera* (Buckman), *B. projecta* (Buckman), *B.* aff. *rotabilis* (Buckman), *Braunsina* sp., *Haplopleuroceras mundum* (Buckman), *H. subspinatum* (Buckman), *Haplopleuroceras* sp., *Fontannesia explanata* Buckman, *Asthenoceras* sp., *Bradfordia* aff. *platyomphala* (Buckman) and *Euhoploceras* sp.

Discites Zone (ZA 194 to ZA 199; 3.5m; Pl. 2.4-2.6)

The lower boundary of this unit is defined by the first occurrence of representatives of *Toxolioceras* (*T.* aff. *curvum* (Buckman) and *Toxolioceras* sp.). *Braunsina aspera* (Buckman), *Reynesella juncta* (Buckman), *Reynesella* sp. and *Euhoploceras dominans* (Buckman) are also present.

Faunal Composition and Paleobiogeography

The ammonite assemblages collected in the section include 312 specimens. They have been deposited at the Departamento de Ciências da Terra (Universidade de Coimbra), and registered with the prefix letters ZA (Aalenian-Bajocian) and ZB (Toarcian-Aalenian).

The faunal composition of the ammonite assemblages recognized in the Aalenian of Zambujal de Alcaria is characterized by the occurrence of Leioceratinae (Opalinum Zone) and Graphoceratinae (Bradfordensis and Concavum Zones) as major elements for correlation with other basins, and by the absence of Phylloceratina and Lytoceratina (Fig. 3).

The Opalinum Zone is characterized by the relative abundance of Tmetoceratinae (47%) as compared to Leioceratinae (12%). Grammoceratinae is very well represented (20%) in this assemblage, as in other areas of the Lusitanian Basin (Henriques, 1992; Henriques, 1995; Henriques et al., 1996). Ammonite assemblages recognised in the Opalinum and Comptum Subzones can be correlated with the Aa-1 and Aa-2 horizons defined by Callomon and Chandler (1990) in southern England.

The Bradfordensis Zone is characterized by an ammonite assemblage mainly composed of Graphoceratinae (67%). Unlike other parts of the Lusitanian Basin, Tmetoceratinae is scarcely represented (1%), while Grammoceratinae (16%) and Erycitidae (10%) are comparatively less frequent. Again, ammonite assemblages recognised in the Bradfordensis and Gigantea Subzones can be correlated with the Aa-10 and Aa-11 horizons defined by Callomon and Chandler (1990) in southern England.

Ammonite assemblages in the Concavum Zone show a decrease in diversity when compared with the data from the northern Lusitanian Basin. They consist mainly of Graphoceratinae (50%) associated with Sonniniidae (29%) and Erycitidae (12%); Bradfordiinae (8%) and Grammoceratinae (1%) are also present but no Otoitidae, Hammatoceratidae or Strigoceratidae have been found. The ammonite assemblages can be correlated with the Aa-15 horizon defined by Callomon and Chandler (1990) in southern England.

Conclusions

The Aalenian of the Zambujal de Alcaria section corresponds to a marly limestone series of sediments displaying abundant and well preserved ammonite assemblages. This paleontological record includes representatives of Leioceratinae and Graphoceratinae, which allow the recognition of the standard biostratigraphic units (i.e. Opalinum, Comptum, Bradfordensis, Gigantea and Limitatum Subzones) and correlation with other basins. The ranges of representatives of the Grammoceratinae and Erycitidae are similar to those recognised in the Betic Basin which stresses the correlation potential and the biogeographic affinities between both basins already pointed out (Henriques, 1992; 1995; Henriques et al., 1996).

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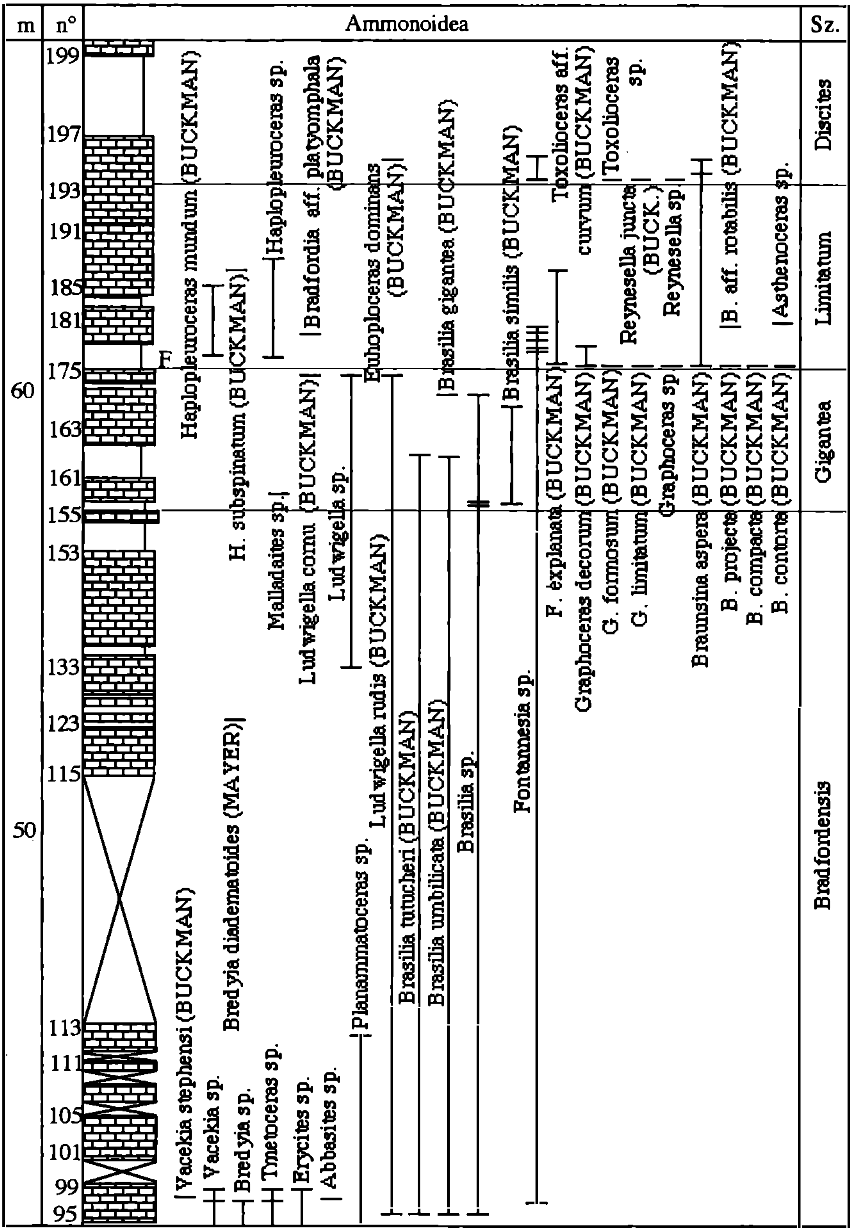
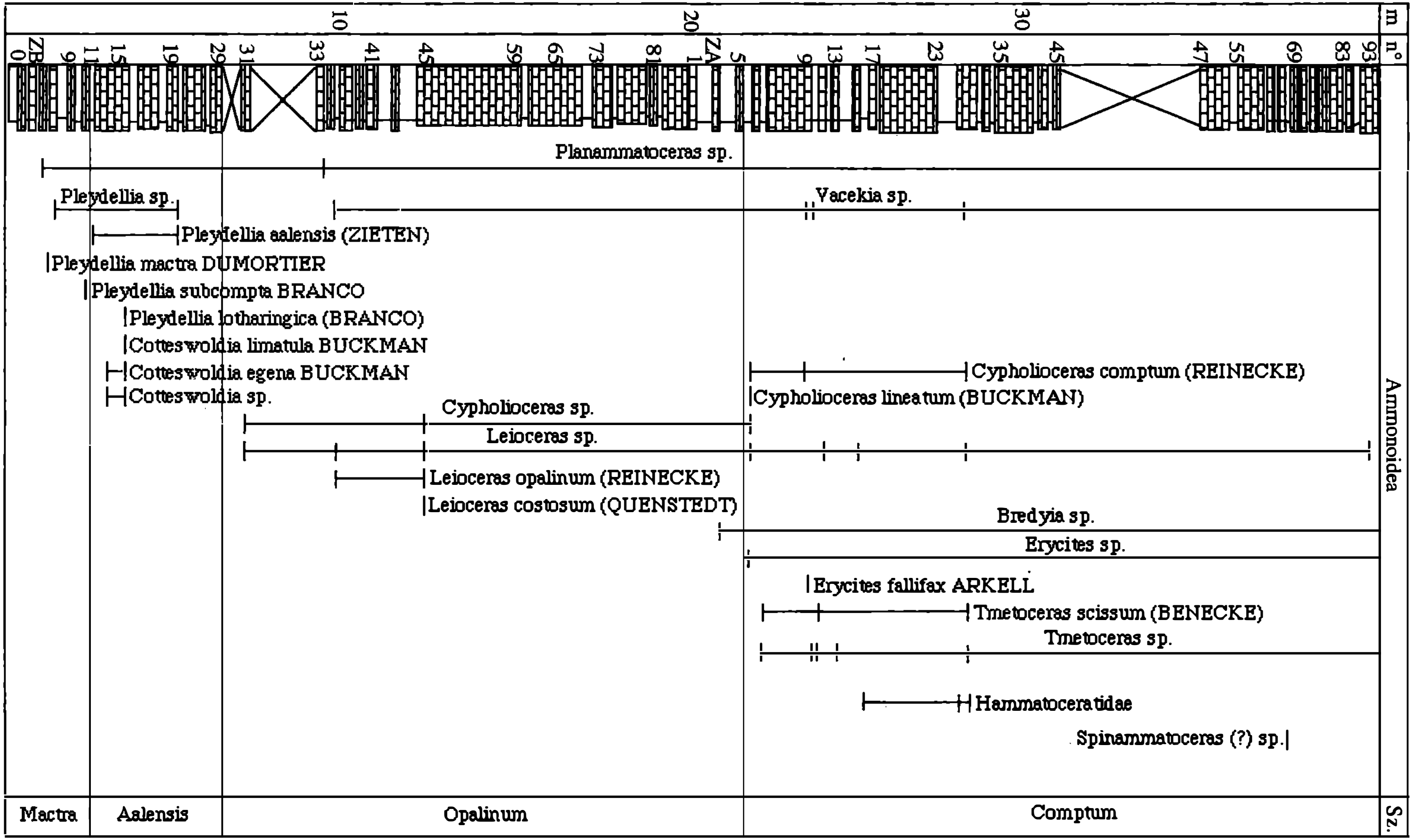


Figure 2. Stratigraphic occurrence and ranges of the ammonoid taxa in the Aalenian of Zambujal de Alcaria section.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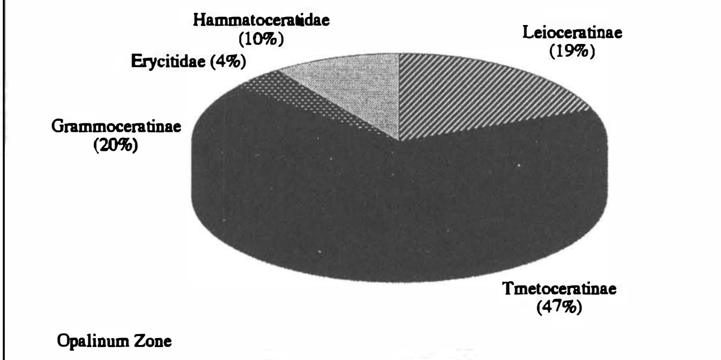
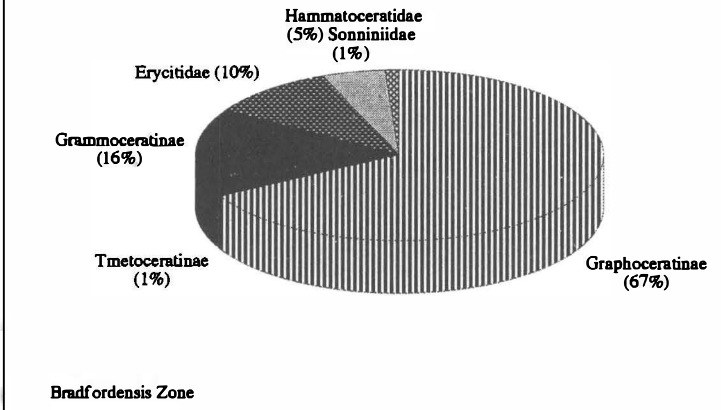
83
93
Planammatoceras sp.
Pleydellia sp.
Vacekia sp.
Pleydellia aalensis (ZIETEN)
Pleydellia mactra DUMORTIER
Pleydellia subcompta BRANCO
Pleydellia lotharingica (BRANCO)
Cotteswoldia limatula BUCKMAN
Cotteswoldia egena BUCKMAN
Cotteswoldia sp.
Cypholioceras comptum (REINECKE)
Cypholioceras lineatum (BUCKMAN)
Cypholioceras sp.
Leioceras sp.
Leioceras opalinum (REINECKE)
Leioceras costosum (QUENSTEDT)
Bredyia sp.
Erycites sp.
Erycites fallifax ARKELL
Tmetoceras scissum (BENECKE)
Tmetoceras sp.
Hammatoceratidae
Spinammatoceras (?) sp.
Ammonoidea
Mactra
Aalensis
Opalinum
Comptum
Sz.


Hammatoceratidae
(10%)
Erycitidae (4%)
Leioceratinae
(19%)
Grammoceratinae
(20%)
Tmetoceratinae
(47%)
Opalinum Zone


Hammatoceratidae
(5%) Sonniniidae
(1%)
Erycitidae (10%)
Grammoceratinae
(16%)
Tmetoceratinae
(1%)
Graphoceratinae
(67%)
Bradfordensis Zone

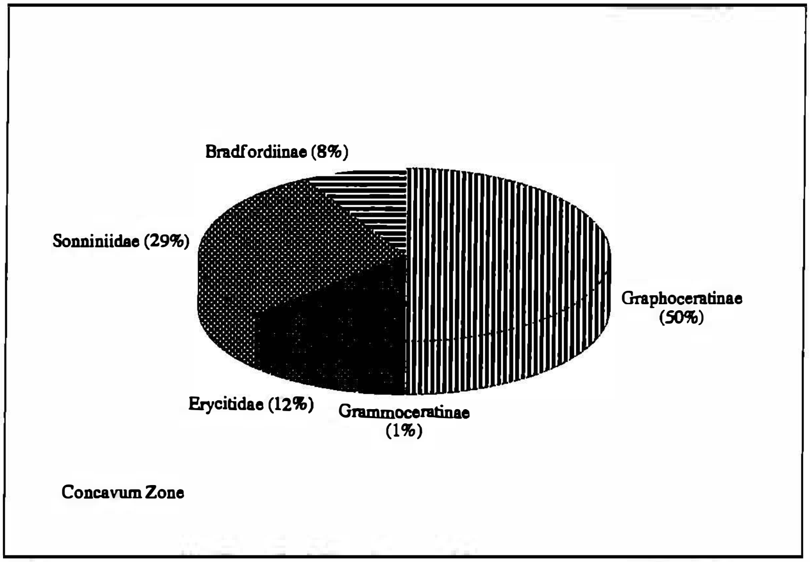


Figure 3. Relative frequencies of the main ammonoid taxa present in the Zambujal de Alcaria section.

References

AZEREDO, A. 1993. Jurássico Médio do Maciço Calcário Estremenho (Bacia Lusitânica): Análise de Fácies, Micropaleontologia, Paleogeografia. Unpublished Ph.D. dissertation, University of Lisboa, 366p.

CALLOMON, J. H. AND R. CHANDLER. 1990. A review of the ammonite horizons of the Aalenian-Lower Bajocian Stages in the Middle Jurassic of southern England, p. 85-112. *In* S. Cresta and G. Pavia (ed.), Proceedings of the Meeting on Bajocian Stratigraphy. Memorie Descrittive della Carta Geologica d'Italia, XL.

CHOFFAT, P. 1880. Étude stratigraphique et paléontologique des terrains jurassiques du Portugal - Première Livraison - Le Lias et le Dogger au Nord du Tage. Memória da Secção de Trabalhos Geológicos, Portugal, 22:1-72.

DUARTE, L. 1995. O Toarciano da Bacia Lusitaniana. Estratigrafia e Evolução Sedimentogénica. Unpublished Ph.D. dissertation, University of Coimbra, 342p.

HENRIQUES, M. H. 1992. Biostratigrafia e Paleontologia (Ammonoidea) do Aaleniano em Portugal (Sector Setentrional da Bacia Lusitaniana). Unpublished Ph.D. dissertation, University of Coimbra, 301p.

____. 1995. Les faunes d'ammonites de l'Aalénien portugais: composition et implications paléobiogéographiques. Geobios M. S., 18:229-235.

____, A. LINARES, J. SANDOVAL, AND M. S. URETA. 1995. The Aalenian of the Iberia (Betic, Lusitanian and Iberian Basins), p.139-150. *In* A. C. Riccardi (ed.), Advances in Jurassic Research. GeoResearch Forum 1-2, Zurich.

PERROT, C. 1965. Études sur l'Aalénien inférieur au Nord du Tage. Comunicações dos Serviços Geológicos de Portugal, XXXVI:129-151.

____. AND R. MOUTERDE. 1957. Études sur l'Aalénien supérieur au Nord du Tage. Comunicações dos Serviços Geológicos de Portugal, XXXVIII:349-371.

RUGET-PERROT, C. 1961. Études stratigraphiques sur le Dogger et le Malm inférieur du Portugal au Nord du Tage, Memória dos Serviços Geológicos de Portugal N.S., 7:1-197.

SERVIÇOS CARTOGRÁFICOS DO EXÉRCITO 1968. Carta Militar de Portugal na escala 1:25,000; folha 318 - Porto de Mós.

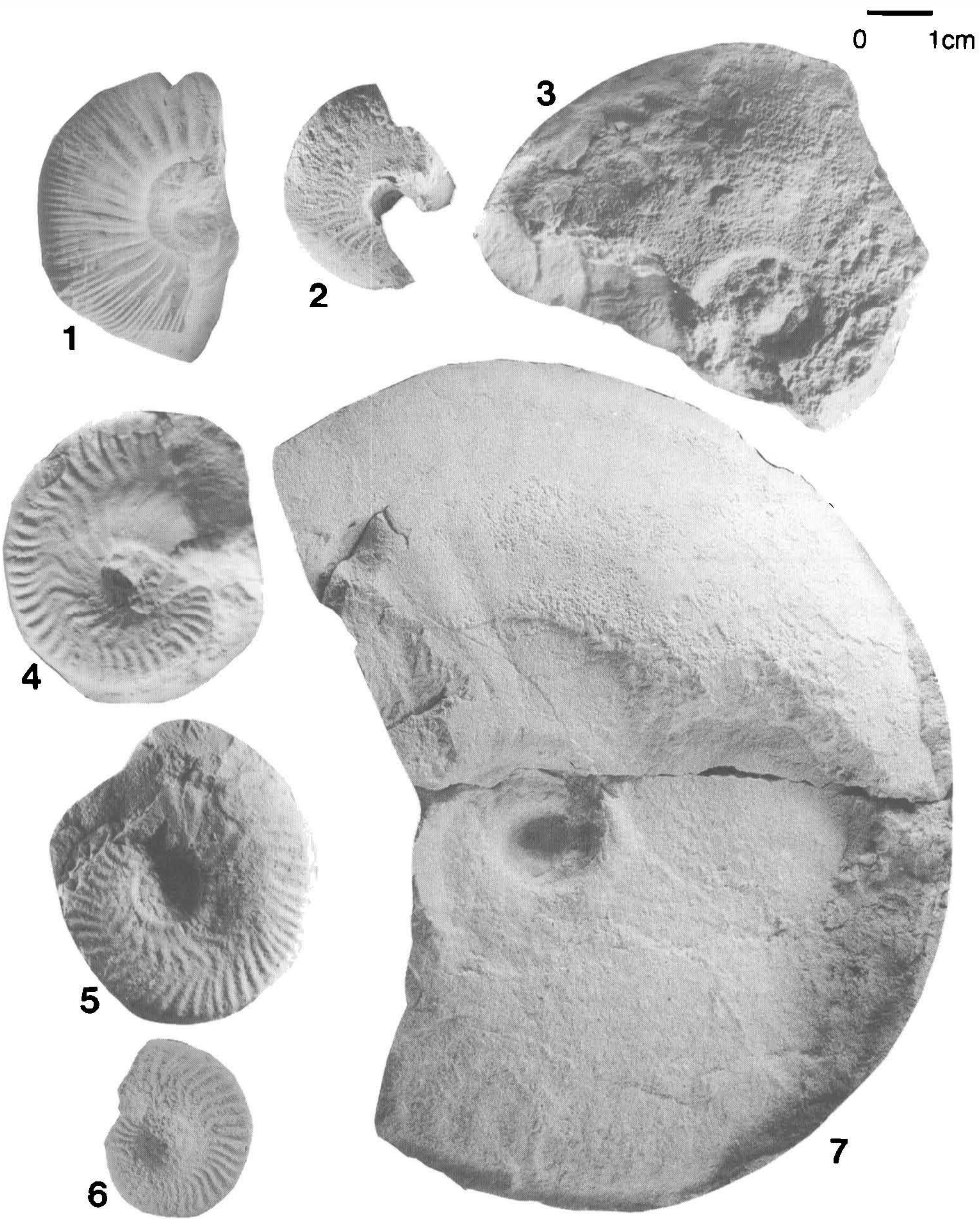


Plate 1. **1.** *Erycites fallifax* Arkell (ZA 9.17; Comptum Subzone; D=50mm; O=16.5; H=16.5; E=10;); **2.** *Cypholioceras comptum* (Reinecke) (ZA 27.1; Comptum Subzone; D=35; O=6.5; H=17; E=7); **3.** *Cypholioceras lineatum* (Buckman) (ZA 5.1; Comptum Subzone; O=19; H=44; E=13.5); **4.** *Ludwigella ? rudis* (Buckman) (ZA 95.1; Bradfordensis Subzone; D=43; O=10.5; H=21.5); **5.** *Brasilia tutcheri* (Buckman) (ZA 95.2; Bradfordensis Subzone; D=44.5; O=9); **6.** *Brasilia umbilicata* (Buckman) (ZA 162.2; Gigantea Subzone; D=29.5; O=8.5; H=12.5; E=7); **7.** *Brasilia gigantea* (Buckman) (ZA 171.1; Gigantea Subzone; D=143.5).

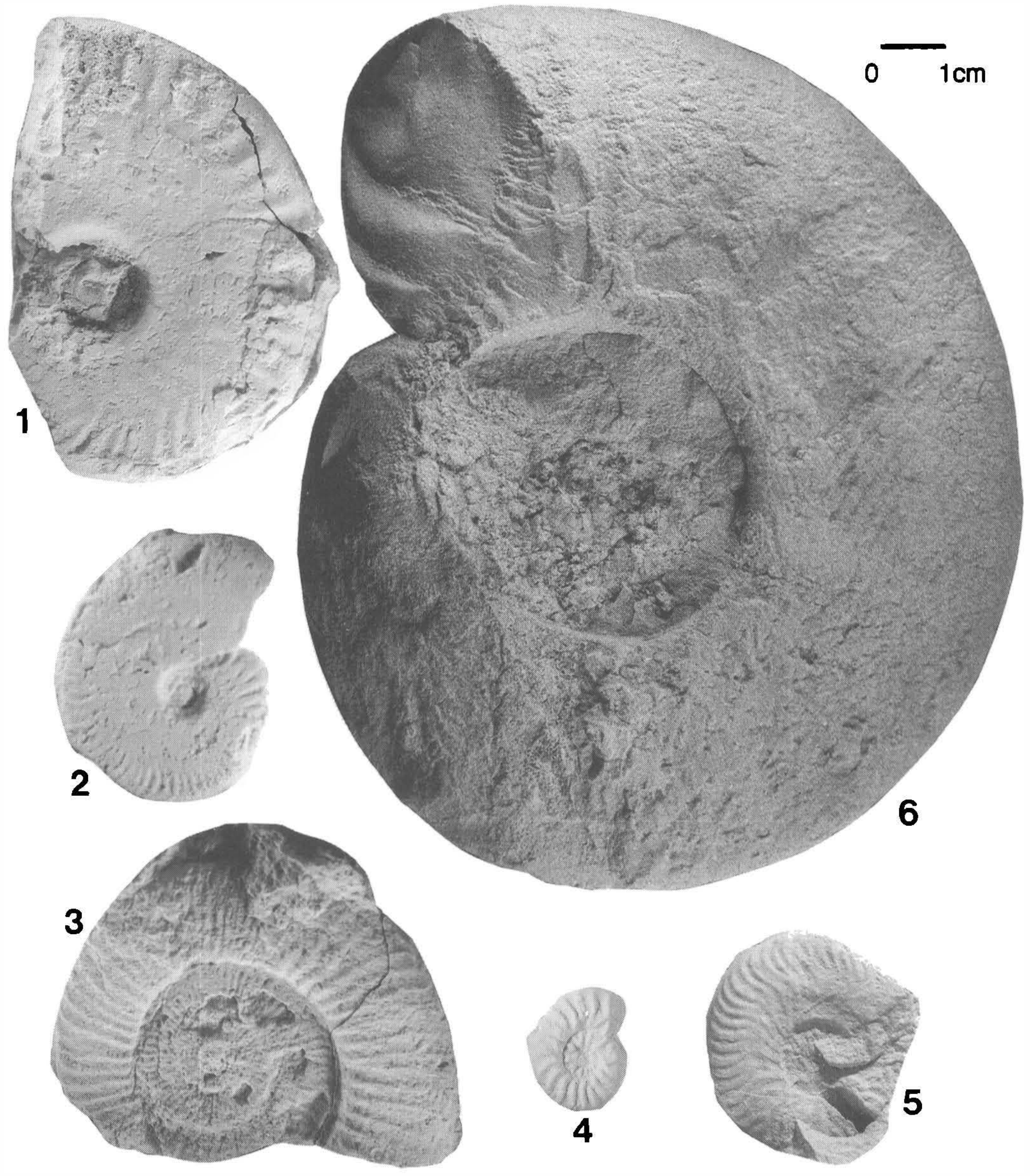


Plate 2. **1.** *Graphoceras limitatum* (Buckman) (ZA 176.8; Limitatum Subzone; D=77mm; O=15.5; H=39; E=16); **2.** *Graphoceras decorum* (Buckman) (ZA 179.1; Limitatum Subzone; D=43.5; O=7; H=19.5; E=5.5); **3.** *Haplopleuroceras mundum* (Buckman) (ZA 185.3; Limitatum Subzone; D=74; O=36; H=26; E=7); **4.** *Braunsina aspera* (Buckman) (ZA 195.3; Discites Zone; D=23; O=7; H=8.8); **5.** *Toxolioceras* aff. *curvum* (Buckman) (ZA 195.1; Discites Zone; D=36.5; O=11); **6.** *Euhoploceras dominans* (Buckman) (ZA 198.1; Discites Zone; D=148.5; O=50; H=60.7; E=35).

ZBYSZEWSKI, G., G. MANUPPELLA, AND O. VEIGA FERREIRA. 1974. Carta Geológica de Portugal na escala de 1/50 000. Notícia Explicativa da Folha 27-A, Vila Nova de Ourém, Serviços Geológicos de Portugal, Lisboa, 82 p.

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Foraminiferal Biostratigraphy of the Lower - Middle Jurassic Sequences in Eastern Arabia

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Abstract: The foraminiferal content of the Lower-Middle Jurassic exposures around Riyadh City in central Saudi Arabia and in ten wells drilled in Qatar Peninsula were studied. The Lower Jurassic Toarcian Marrat Formation in central Saudi Arabia and the Hamlah Formation in Qatar were found barren of foraminifera. In the Middle Jurassic Dhruma Formation in central Saudi Arabia, 95 five foraminiferal species are identified. In the Middle Jurassic Izhara and Araej formations in Qatar, 33 foraminiferal species are identified. Biostratigraphic schemes are proposed for each of the two areas studied. In central Saudi Arabia the Middle Jurassic is divided into eight biostratigraphic zones and in Qatar into seven zones. A biostratigraphic correlation between these two proposed schemes and those proposed by previous authors in Arabia and neighboring areas is attempted.

Introduction

The eastern part of the Arabian Peninsula is one of the main oil and gas reservoirs in the world and most is produced from Jurassic rocks. The general lithologic characters, stratigraphic contacts and faunal contents of the Jurassic units in central Saudi Arabia have been studied by Steineke and Bramkamp (1952), Arkell (1952), Steineke, Bramkamp and Sander (1958), Powers et al. (1966) and Powers (1968). The main work on Jurassic faunas has been done by Redmond (1964a; 1964b; 1965), Imlay (1970), Enay et al. (1986) and Banner et al. (1991). Studies of Jurassic faunas and stratigraphy of Qatar are scarce, the most important being those by Smout and Sugden (1961), Sugden and Standring (1975), and Hewaidy and Al-Saad (1992).

The material on which this study was based consists of samples collected from a composite surface section representing the Lower - Middle Jurassic Marrat and Dhruma formations in the area near Riyadh City in central Saudi Arabia, and core and ditch samples of the Lower - Middle Jurassic from ten deep wells distributed throughout different structural parts of the Qatar Peninsula; of these, seven wells were in the Dukhan oil field, one in the Simsima Arch at the northeast corner of Qatar Peninsula, and two in the central part of Qatar on the central Qatar Arch (Fig. 1).

Biostratigraphy

In this study, 95 foraminiferal species were identified from central Saudi Arabia and 33 from Qatar. Due to the paleoecological and environmental variations between the areas studied in central Saudi Arabia and Qatar, two different biostratigraphic zonation schemes are proposed and correlation between them has been attempted. The Lower Jurassic Toarcian strata in central Saudi Arabia (Marrat Formation) and Qatar (Hamlah Formation) are barren of foraminifera. Based on foraminiferal evidence, the Middle Jurassic Dhruma Formation in central Saudi Arabia is divided into eight biozones, while the Middle Jurassic Izhara and Araej Formations in Qatar are divided into seven biozones (Figs. 2-4). In this scheme we used the best known and most widely distributed foraminifera in the Middle East as markers for the zones, such as *Pfenderina* and *Kurnubia*. Distribution charts for the species identified are shown in Figs. 2, 3, and 4.

The following is an outline of the two biostratigraphic schemes proposed here:

Central Saudi Arabia

Lower Jurassic: The Lower Jurassic in central Saudi Arabia is represented by the Marrat Formation, which did not yield any foraminifera. Nevertheless, based on ammonite assemblages, Arkell (1952), Powers et al. (1966), Powers (1968), and Imlay (1970) assigned a Toarcian age to the Marrat Formation, based on the occurrence of species of *Bouleiceras* in the lower member of the Marrat Formation and of species of *Nejdia* in the upper member. The middle member lacks ammonites. The lower Marrat Member is characterized by the presence of the *Bouleiceras nitescens* Thevenin, *B. elegans* Arkell, *B. arabicum* Arkell, *Protogrammoceras madagascariense* Thevenin and *Protogrammoceras* sp. The upper member yields abundant *Nejdia bramkampi* Arkell and *Hildaites sanderi* Arkell.

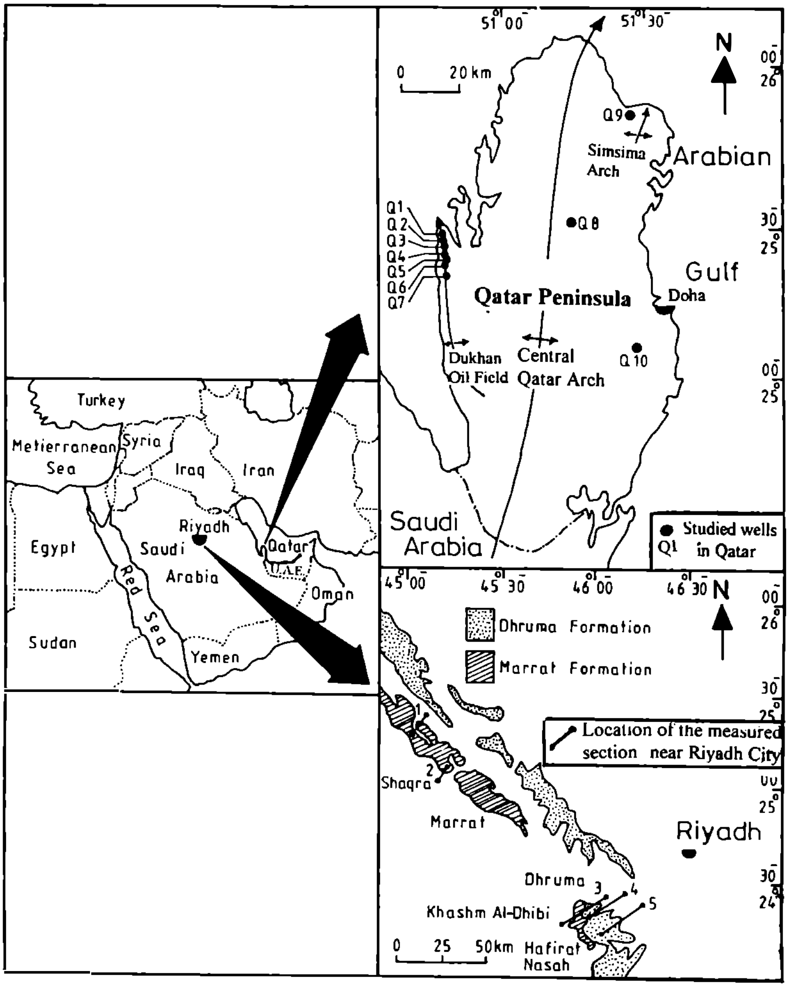


Figure 1. Locality map.

Middle Jurassic: The Middle Jurassic in Central Saudi Arabia is represented by the Dhruma Formation. It is rich in foraminifera in addition to other faunal groups, especially ammonites. The Dhruma Formation is here divided into seven foraminiferal biozones, in addition to a barren interval in the lower part of the lower member of the Dhruma Formation. Descriptions of these zones, beginning at the base, are given below:

1. Barren Zone: This zone occurs in the lower part of the lower member of the Dhruma Formation and is 45m thick. It occurs in yellowish to tan, hard, very fine limestone. It is considered to be of Early Bajocian age (Fig. 5), as it is equivalent to the lower half of the *Dorsetensia* ammonite Zone of Imlay (1970). It is also equivalent to the middle part of the *Haurania* foraminiferal Zone of Powers (1968) which includes the whole of the lower member of the Dhruma Formation.
2. *Haurania deserta* Zone: This zone is represented by the total range of the marker species *Haurania deserta* Henson. It also includes *Amijiella amiji* (Henson) and *Lenticulina muensteri* (Roemer). It occurs in the middle part of the lower member of the Dhruma Formation (Figs. 2, 3), in yellowish, hard limestones. The thickness of this zone is 32m. The genus *Haurania* is an index element in the Bajocian of the Middle East particularly in Palestine (Henson, 1948) and in Iran (Sampo, 1969). This zone is considered to be of Middle Bajocian age (Fig. 5) as it is equivalent to the upper half of the *Dorsetensia* ammonite Zone of Imlay (1970). It is also equivalent to the middle part of the *Haurania* foraminiferal Zone of Powers (1968).
3. *Trocholina minuta* Zone: This Zone is defined by the total range of the zonal marker, *Trocholina minuta*. It includes the upper half of the lower member of the Dhruma Formation (Dhibi Limestone member) and overlies the *Haurania deserta / Amijiella amiji* Zone (Figs. 2, 3). It occurs in gray to light gray limestones and is about 30m thick. In this interval, six foraminiferal species are recorded (Figs. 2, 3). *Trocholina minuta* Derin and Reiss is well known in Bajocian sediments of the Middle East, particularly in Palestine (Henson, 1948) and Iran (Sampo, 1969). The *Trocholina minuta* Zone is considered to be of Late Bajocian age (Fig. 5) as it is equivalent to the *Ermoceras* ammonite Zone of Arkell (1952) and Imlay (1970). It is also equivalent to the upper part of the *Haurania* foraminiferal Zone of Powers (1968).
4. *Pseudomarssonella bipartita* Zone: The total range of the zonal marker *Pseudomarssonella bipartita* is used to delineate this Zone which occupies the lower part of the middle member of the Dhruma Formation (Figs. 2, 3), composed of gray to light yellowish limestones, and about 33m thick. The Zone contains eleven foraminiferal species (Figs. 2, 3). The genus *Pseudomarssonella* is widely distributed in the Bathonian of Saudi Arabia. Powers (1968) considered the *Pseudomarssonella mcclueri* Zone to be of Early Bathonian age. In this study, the *Pseudomarssonella bipartita* Zone is considered as Early Bathonian as it is equivalent to the *Thambites* ammonite Zone of Arkell (1952) and the *Thambites* and *Dhrumaites* Zone of Imlay (1970).
5. *Dhrumella evoluta* Zone: This zone is based on the total range of the marker species, *Dhrumella evoluta* Redmond. This zone occupies the middle part of the middle member of the Dhruma Formation which consists of gray to light gray limestones (Figs. 2, 3). The thickness of this zone is about 37m. This zone yields the richest foraminiferal fauna in the Dhruma Formation, producing 45 species (Figs. 2, 3). It is considered to be of Middle Bathonian age (Fig. 5) as it is equivalent to the *Tulites* ammonite Zone of Arkell (1952) and Imlay (1970). It is also equivalent to the *Dhrumella evoluta* Zone of Powers (1968) in Saudi Arabia.
6. *Riyadhella arabica* Zone: The total range of the marker species *Riyadhella arabica* Redmond is used to define this Zone, which occurs in the upper part of the middle member of the Dhruma Formation (Figs. 2, 3). It is composed in its lower and middle parts of yellowish to tan, thin bedded, laminated limestones while its upper part is composed of massive, yellowish limestones. The thickness of this zone is about 49m. In this zone 24 species of foraminifera are recorded (Figs. 2, 3). The *Riyadhella arabica* Zone is considered to be of Middle Bathonian age (Fig. 5) as it is equivalent to the *Micromphalites* and *Dhrumaites* ammonite Zones of Arkell (1952) and the *Micromphalites*, *Thambites* and *Dhrumaites* Zones of Imlay (1970). It is also equivalent to the *Flabellammina* foraminiferal Zone of Powers (1968).
7. *Pfenderina gracilis / Pfenderella arabica* Zone: This zone is defined between the first appearance of *Pfenderina gracilis* Redmond and last occurrence of *Pfenderella arabica* Redmond. This zone occupies the lower part of the upper member of the Dhruma Formation (Atash Member), which is composed of light to yellowish, massive, chalky limestones. The thickness of this zone is about 24m. In this zone 18 species are identified (Figs. 2, 3). According to Arkell (1952), this part of the upper Dhruma Member is barren of ammonites, while Imlay (1970) recorded a *Grossouvria* assemblage. The genus *Pfenderina* is a very characteristic genus in the Bathonian of the Middle East, where there are many records (Henson, 1948; Smout and Sugden, 1961; Powers et al., 1966; Powers, 1968; Sampo, 1969; Hewaidy and Al-Saad, 1992). The *Pfenderina gracilis / Pfenderella arabica* Zone is considered to be Late Bathonian in age (Fig. 5), as it is equivalent to the foraminiferal Zones of *Pfenderina trochoidea* and *Pseudocyclammina* of Powers (1968).

8. *Praekurnubia crusei* / *Kurnubia palestiniensis* Zone: This zone is defined between the first appearance of the *Praekurnubia crusei* Redmond and last occurrence of *Kurnubia palestiniensis* Henson. This zone occurs in the middle part of the upper member of the Dhruma Formation (lower half of the Hisyan Member), which is composed of yellowish to gray, thick bedded limestones. The thickness of this zone is about 42m. Twenty species of foraminifera are recorded in this zone (Figs. 2, 3). This part of the upper Dhruma is barren of ammonites (Arkell, 1952; Imlay, 1970). The genus *Kurnubia* is a marker for the Callovian in the Middle East (Henson, 1948; Powers et al., 1966; Powers, 1968; Sampo, 1969). The *Praekurnubia crusei/Kurnubia palestiniensis* Zone is considered of Early Callovian age (Fig. 5) as it is equivalent to the *Praekurnubia crusei* foraminiferal Zone of Powers (1968).

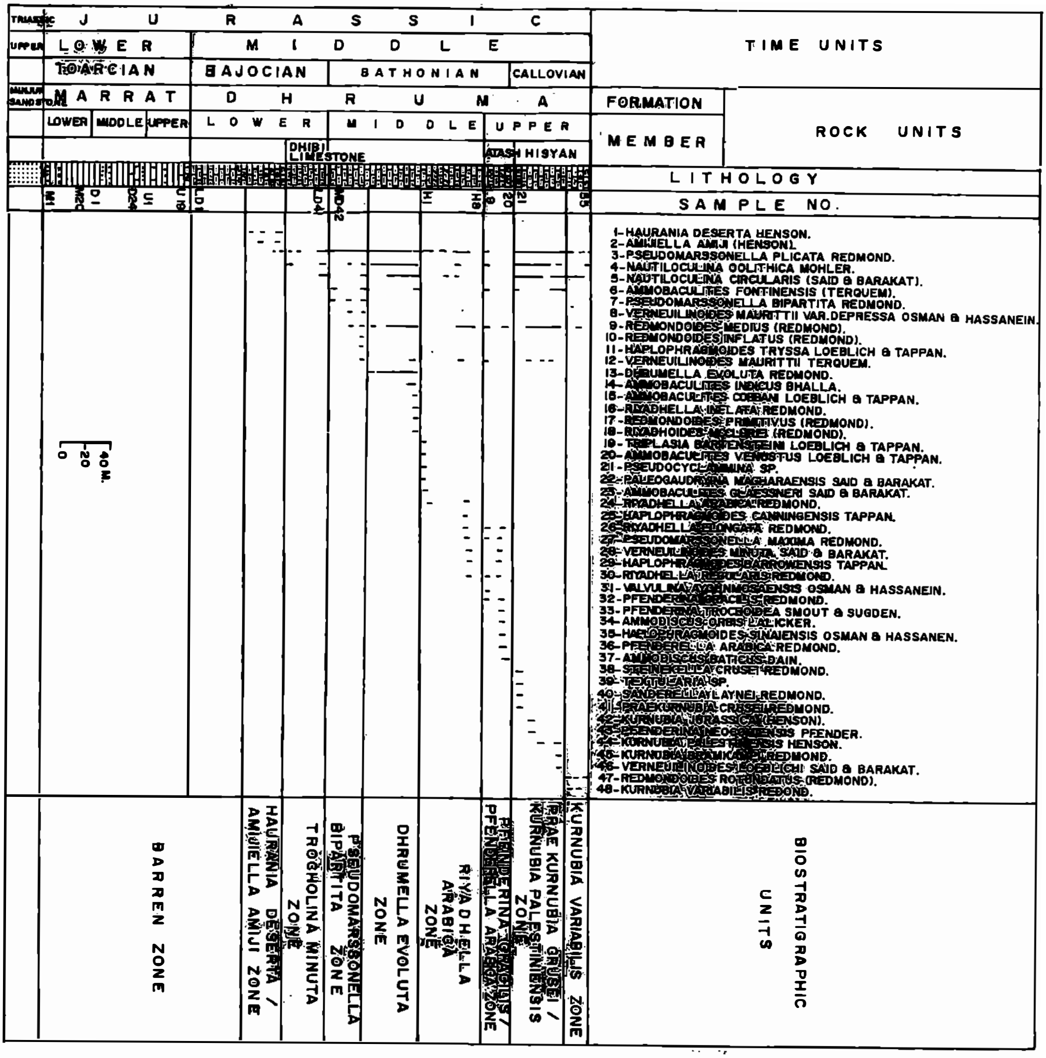


Figure 2. Distribution chart of the identified arenaceous foraminifera in the Dhruma Formation.

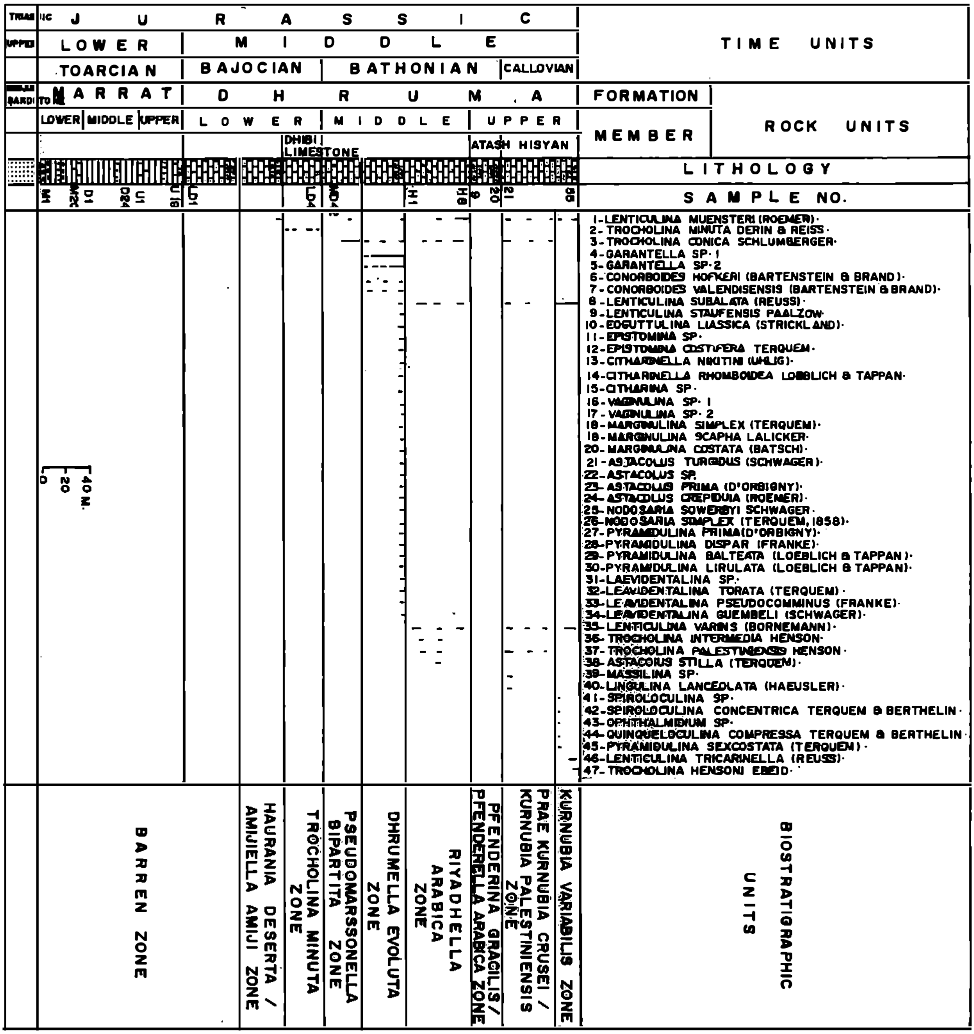


Figure 3. Distribution of the identified calcareous foraminifera in the Dhruma Formation.

9. *Kurnubia variabilis* Zone: The upper limit of the zone is taken at the disappearance of the zonal species *Kurnubia variabilis* Redmond in addition to *Trocholina hensoni* Ebeid, *Lenticulinatricarinella* (Reuss), *Lenticulina subalata* (Reuss) and *Lenticulina muensteri* (Roemer). Its base is defined by the disappearance of the marker species of the underlying zone in addition to *Kurnubia bramkampi* Redmond and *Verneuilinoides loeblichi* Said and Barakat. This zone is represented in the topmost part of the upper member of the Dhruma Formation (upper part of the Hisyan Member), where it occurs in massive, white and creamy, chalky limestones (Figs. 2, 3). The thickness of this zone is about 20m. In this interval 16 species of foraminifera are recorded (Figs. 2, 3). According to Arkell (1952), this part of the upper Dhruma lacks ammonites, while Imlay (1970) recorded *Erymnoceras* and *Pachyceras* assemblages in the topmost part of the upper Member. The *Kurnubia variabilis* Zone is considered as being Early Callovian (Fig. 5) as it is equivalent to the *Kurnubia bramkampi* foraminiferal Zone of Powers (1968).

Qatar

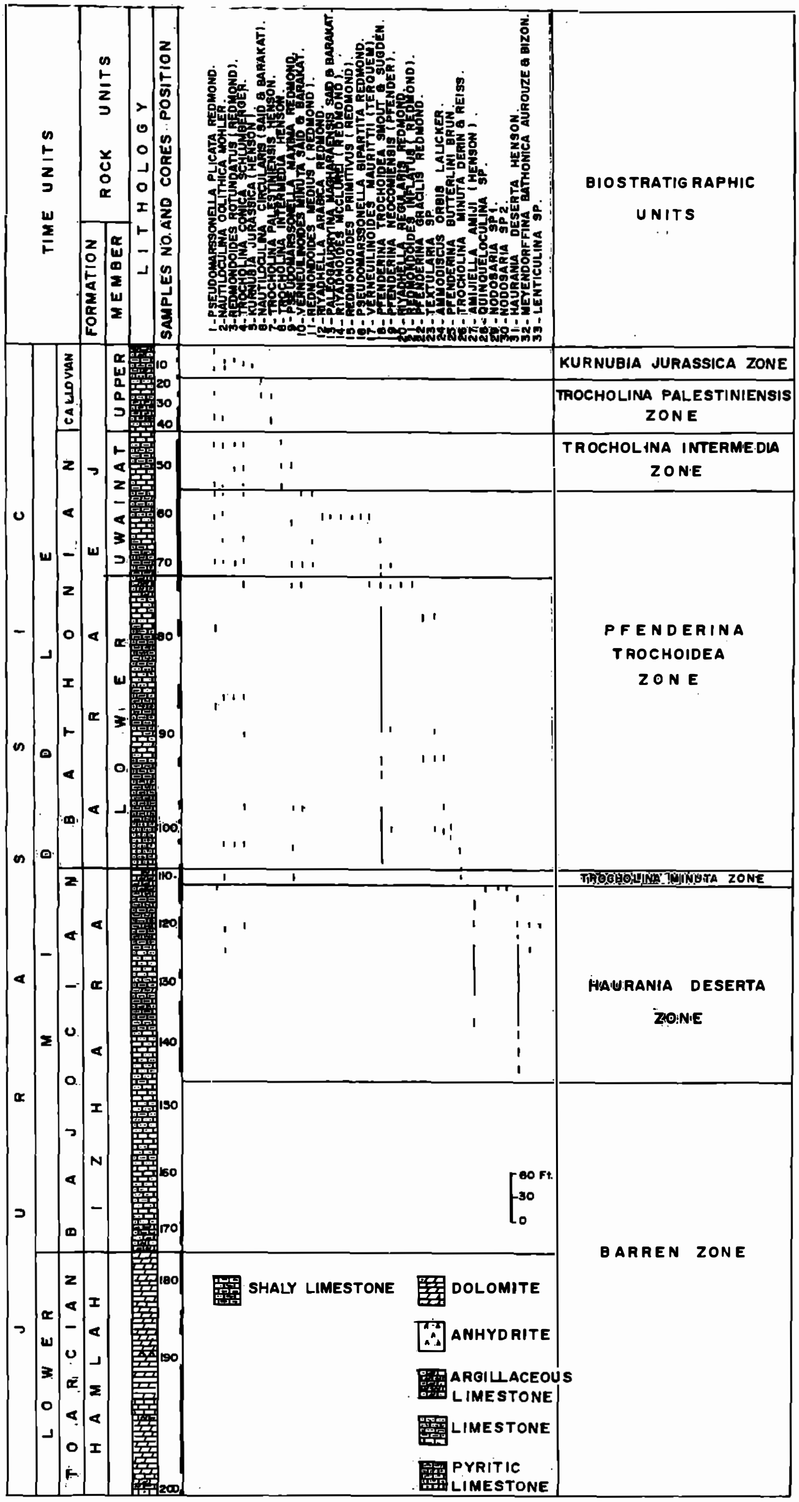
The Lower - Middle Jurassic in Qatar is represented by the Hamlah, Izhara and Araej Formations (from the bottom upwards). In this interval, 33 foraminiferal species are identified in the ten wells studied in the Qatar Peninsula. The distribution of foraminifera in well no. Q1 is used as representative for the distribution of foraminifera in the Lower-Middle Jurassic sequence in Qatar.

Lower Jurassic: The Lower Jurassic is represented in Qatar by the Hamlah Formation. The age of the Hamlah Formation is controversial as it has not yielded any faunas and is similar in lithology to the underlying Gulailah Formation. Sugden and Standring (1975) assigned the Hamlah Formation in Qatar to the Lower Jurassic while the geologists of Qatar General Petroleum Corporation (1983) placed it in the Middle or Upper Triassic with a major unconformity between the Hamlah and overlying Izhara formations. Hamam (1985) assigned a Norian age to Hamlah Formation. Here we consider the Hamlah Formation as a part of the Jurassic because it occurs as a shallow-water, nonclastic facies representing the first stage in the Jurassic marine transgression; in contrast, the Triassic is represented by clastic rocks. In Saudi Arabia, Toarcian strata are clayey, shaly limestones of the Marrat Formation which directly underlie the certainly marine Dhruma Formation. The Hamlah Formation in Qatar may be the lithologic equivalent of the Marrat Formation in Saudi Arabia. Thus, the Hamlah Formation may be of Toarcian age.

Middle Jurassic: The Middle Jurassic in Qatar is represented by the Izhara and Araej Formations. Study of the foraminiferal content of the Araej Formation was previously done by Hewaidy and Al-Saad (1992). In the present study, the Middle Jurassic Izhara and Araej formations have been divided into seven biozones. The lower three zones are Bajocian, the middle two zones Bathonian, while the upper two zones are Early Callovian. Descriptions of these zones, beginning from the top, are given below:

1. *Kurnubia jurassica* Zone: This zone spans the occurrence of the marker species. It occurs in the upper member of the Araej Formation in dark gray, pyritic, shaly limestones. This zone ranges between 7.5m and 12m in thickness in the wells studied. Six foraminiferal species are rarely recorded in this zone (Fig. 4). *Kurnubia jurassica* (Henson) is widely distributed in the Callovian of the Middle East, having been recorded in the Early Callovian in Qatar (Hewaidy and Al-Saad, 1992) and in Egypt (Osman and Hassanein, 1967; Hassanein et al., 1978). The *Kurnubia jurassica* Zone is considered to be of Early Callovian age (Fig. 5) as it is equivalent to the *Kurnubia variabilis* Zone in Central Saudi Arabia in this study.
2. *Trocholina palestiniensis* Zone: This zone is defined by the total range of the marker species, and occurs in the lower part of the upper member of the Araej Formation (Fig. 4) which consists of gray to dark olive, shaly limestones. The thickness of this zone ranges between 23.5m and 75m. Six foraminiferal species are recorded in this zone (Fig. 4). *Trocholina palestiniensis* is well known in the Early Callovian of Palestine and Egypt (Henson, 1948; Derin and Reiss, 1966; Osman and Hassanein, 1967). The *Trocholina palestiniensis* Zone is attributed to the Early Callovian due to its stratigraphic position below the well known Callovian *Kurnubia jurassica* Zone and the absence of any diagnostic Bathonian forms throughout the zone (Fig. 5). It is equivalent to the *Praekurnubia crusei / Kurnubia palestiniensis* Zone in Central Saudi Arabia in this study, and to the Barren Zone of Hewaidy and Al-Saad (1992) in Qatar.
3. *Trocholina intermedia* Zone: The upper limit of this zone is defined by the first appearance of *Trocholina intermedia* Henson, and its lower limit is marked by the first appearance of *Pfenderina trochoidea* Smout and Sugden. This zone is represented in the upper part of the Uwainat Member (Fig. 4), composed of light gray limestones. The thickness of this zone ranges between 18m and 23m. Nine foraminiferal species are recorded in this zone (Fig. 4). The *Trocholina intermedia* Zone is considered to be of Late Bathonian age because of its stratigraphic position conformably overlying the well known Early to Middle Bathonian *Pfenderina trochoidea* Zone. This zone is also equivalent to the *Pfenderina gracilis / Pfenderella arabica* Zone of this study in central Saudi Arabia. It was previously recorded in Qatar by Hewaidy and Al-Saad (1992, Figs. 3, 4).

Figure 4. Distribution chart of the identified foraminiferal species in well Q1, Dukhan Field, Qatar.


TIME UNITS
ROCK UNITS
FORMATION
MEMBER
LITHOLOGY
SAMPLES NO. AND CORES POSITION
1-PSEUDOMARSSONELLA PLICATA REDMOND.
2-NAUTILOCULINA OOLITHICA MOHLER.
3-REDMONDOIDES ROTUNDATUS (REDMOND).
4-TROCHOLINA CONICA SCHLUMBERGER.
5-KURNUBIA JURASSICA (HENSON).
6-NAUTILOCULINA CIRCULARIS (SAID & BARAKAT).
7-TROCHOLINA PALESTINIENSIS HENSON.
8-TROCHOLINA INTERMEDIA HENSON.
9-PSEUDOMARSSONELLA MAXIMA REDMOND.
10-VERNEUILINOIDES MINUTA SAID & BARAKAT.
11-REDMONDOIDES MEDIUS (REDMOND).
12-RIYADHELLA ARABICA REDMOND.
13-PALEOGAUDRYINA MAGHARAENSIS SAID & BARAKAT.
14-RIYADHOIDES MCCLUREI (REDMOND).
15-REDMONDOIDES PRIMITIVUS (REDMOND).
16-PSEUDOMARSSONELLA BIPARTITA REDMOND.
17-VERNEUILINOIDES MAURITII (TERQUEM).
18-PFENDERINA TROCHOIDEA SMOUT & SUGDEN.
19-PFENDERINA NEOCOMIENSIS (PFENDER).
20-RIYADHELLA REGULARIS REDMOND.
21-REDMONDOIDES INFLATUS (REDMOND).
22-PFENDERINA GRACILIS REDMOND.
23-TEXTULARIA SP.
24-AMMODISCUS ORBIS LALICKER.
25-PFENDERINA BUTTERLINI BRUN.
26-TROCHOLINA MINUTA DERIN & REISS.
27-AMIJIELLA AMIJI (HENSON).
28-QUINQUELOCULINA SP.
29-NODOSARIA SP1.
30-NODOSARIA SP2.
31-HAURANIA DESERTA HENSON.
32-MEYENDORFFINA BATHONICA AUROUZE & BIZON.
33-LENTICULINA SP.
BIOSTRATIGRAPHIC UNITS
JURASSIC
MIDDLE
LOWER
CALLOVIAN
BATHONIAN
BAJOCIAN
TOARCIAN
ARAEJ
IZHARA
HAMLAH
UPPER
UWAINAT
LOWER
10
20
30
40
50
60
70
80
90
100
110
120
130
140
150
160
170
180
190
200
KURNUBIA JURASSICA ZONE
TROCHOLINA PALESTINIENSIS ZONE
TROCHOLINA INTERMEDIA ZONE
PFENDERINA TROCHOIDEA ZONE
TROCHOLINA MINUTA ZONE
HAURANIA DESERTA ZONE
BARREN ZONE
60 Ft.
30
0
SHALY LIMESTONE
DOLOMITE
ANHYDRITE
ARGILLACEOUS LIMESTONE
LIMESTONE
PYRITIC LIMESTONE

4. *Pfenderina trochoidea* Zone: The upper boundary of this zone is determined by the first appearance of the zonal marker in addition to other five foraminiferal species. The lower boundary of this zone is determined by the first appearance of *Ammodiscus orbis* Lalicker. This zone encompasses the whole of the lower member of the Araej Formation and the basal part of the Uwainat Member (Fig. 4). Rocks in its lower part are gray, laminated and somewhat argillaceous limestones which gradually change upward to stylolitic limestones. The thickness of this zone ranges between 127m and 152m. Twenty foraminiferal species are recorded in this zone (Fig. 4). *Pfenderina trochoidea* Smout and Sugden is widely distributed in the Bathonian lower member and Uwainat Member of the Araej Formation in Qatar and United Arab Emirates (Smout and Sugden, 1961). This zone is considered to be of Early to Middle Bathonian age (Fig. 5). It is also equivalent to the *Riyadhella arabica*, *Dhrumella evoluta* and *Pseudomarssonella bipartita* Zones of this study in central Saudi Arabia, and to the *Pfenderina trochoidea* and *Ammodiscus orbis* Zones of Hewaidy and Al-Saad (1992) in Qatar.

Time Units			Biostratigraphic Units				
			Suadi Arabia Present study	Qatar Present study	Saudi Arabia Powers, 1968	Qatar Hewaidy & Al-Saad 1992	Western Sinai Osman & Hassanein 1967
Middle Jurassic	Callovian	Early	Kuranubia variabilis	Kurnubia jurassica	Kurnubia bramkampi	Kurnubia jurassica	Valvulinella jurassica
			Praekurnubia crusei / Kuranubia palestiniensis	Trocholina palestiniensis	Praekurnubia crusei	Barren Zone	Haplophregmoides sinaiensis
	Bathonian		Pfenderina gracilis/ Pfenderella arabica	Trocholina intermedia	Pseudocyclammina	Trocholina intermedia	Ammodiscus orbis
			Riyadhella arabica	Pfenderina trochoidea	Pfenderina trochoidea	Pfenderina trochoidea	
					Flabellammina		
			Dhrumella evoluta		Dhrumella evoluta	Ammodiscus orbis	Lingulina longiscata
			Pseudomarssonella bipartita		Pseudomarssonella mcclueri		
	Bajocian		Trocholina minuta	Trocholina minuta	Haurania sp		Barren interval
			Haurania deserta / Amijiella amiji	Haurania deserta / Amijiella amiji			
			Barren interval	Barren interval			

Figure 5. Biostratigraphic correlation between the results of the present study and some previous studies.

5. *Trocholina minuta* Zone: The upper limit of this zone is marked by the first appearance of the zonal marker and its lower limit by the first appearance of *Haurania deserta* Henson. This zone occurs in the upper part of the Izhara Formation (Fig. 4) which consists of gray to olive, marly limestones. The thickness of this zone ranges from 3m and 6m. Four foraminiferal species are recorded in this zone (Fig. 4). As mentioned above, *Trocholina minuta* is widely distributed in the Bajocian of the Middle East. *Trocholina minuta* Zone is considered as Late Bajocian in age (Fig. 5) due to its stratigraphic position conformably overlying the Early Bajocian *Haurania deserta* Zone. This zone is also equivalent to the *Trocholina minuta* Zone in central Saudi Arabia.
6. *Haurania deserta* / *Amijiella amiji* Zone: This zone is represented by the total range of the marker species and occurs in the upper part of the Izhara Formation is composed of gray to olive

pyritic limestones. It conformably overlies the unfossiliferous horizon of the Izhara Formation (Fig. 4). The thickness of this zone in the area studied ranges between 31m and 73m. Ten foraminiferal species were recorded from this zone (Fig. 4) which is considered to be Late Bajocian in age (Fig. 5), as it is equivalent to the *Haurania deserta / Amijiella amiji* Zone of the present study in central Saudi Arabia.

7. Barren zone: This zone occurs in the gray to dark olive, pyritic, shaly limestones of the lower half of the Izhara Formation (Fig. 4). The thickness of this zone in the area studied ranges between 22m and 64m. It is considered to be of Early Bajocian age (Fig. 5) as it conformably underlies the Middle Bajocian *Haurania deserta / Amijiella amiji* Zone. It is also equivalent to the barren zone of this study in central Saudi Arabia.

Conclusions

The Lower- Middle Jurassic foraminiferal faunas from central Saudi Arabia and in ten wells in Qatar have been studied. The Lower Jurassic Toarcian rocks did not yield any foraminifera. The foraminifera of the Middle Jurassic Dhruma, Izhara and Araej formations were examined. 95 foraminiferal species are identified in the Middle Jurassic Dhruma Formation in central Saudi Arabia and 33 species of foraminifera are identified in the Middle Jurassic Izhara and Araej formations in Qatar. The Middle Jurassic Dhruma Formation in Saudi Arabia is divided into eight biozones: two Bajocian, four Bathonian, and two Callovian. In Qatar, the Middle Jurassic is represented by the Izhara and Araej formations. The Izhara Formation is divided into two Bathonian zones, while the Araej Formation is divided into five biozones: three Bathonian, and two Callovian. A regional correlation is attempted between the established zones and their equivalents in the Arabian Peninsula and Egypt.

Acknowledgments

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References

ARKELL, W. J. 1952. Jurassic Ammonites from Jebel Tuwaiq, Central Arabia. Avec une Introduction stratigphique de R A. Bramkamp and M. Steineke. Transactions of the Royal Society (London), Ser. B, 236:241-313.

BANNER, F. T., M. D. SIMMONS AND J. E. WHITTAKER. 1991. The Mesozoic Chrysalidinidae (Foraminifera, Textulariacea) of the Middle East: the Redmond (Aramco) taxa and their relatives. Bulletin of British Museum Natural History (Geology),47:101-152.

ENAY, R., Y.-M. LE NINDRE, C. MANGOLD, J. MANIVIT AND D. VASLET. 1986. The Jurassic of Central Saudi Arabia: new data on lithostratigraphic units, paleoenvironments, ammonite faunas, ages and correlations. Deputy Ministry of Mineral Research, Jiddah, Technical Record BRGM-TR-06-3, 65p.

HAMAM, K. A. 1985. Notes on the Uwainat oil development in Maydan Mahzam and Bul Hanine fields, offshore Qatar. SPE paper, Bahrein.

HASSANEIN, A. M., Z. M. SHARAWY, AND M. Y. HASSAN. 1978. Jurassic stratigraphy of Mamura well, Western Desert, Egypt. Proceedings of Egyptian Academy of Science, 31:1-7.

HENSON, F. R. S. 1948. Stratigraphical correlation by small foraminifera in Palestine and adjoining countries. Geological Magazine, 75.

HEWAIDY, A. AND H. AL-SAAD. 1992. Foraminifera and age of the Araej Formation in the Dukhan oil field, western Qatar, Arabian Gulf. Qatar University Science Journal, 12:200-211.

IMLAY, R. W. 1970. Some Jurassic Ammonites from Central Saudi Arabia. United States Geological Survey Professional Paper 643-D.

OSMAN, A. AND A. M. HASSANEIN. 1967. New Jurassic, Lower Cretaceous and Miocene foraminifera from west central Sinai. Bulletin Faculty of Engineering, Cairo Univesity, Egypt, 315-331.

POWERS, R. W. 1968. Saudi Arabia, In: Lexique Stratigraphique international, Vol. 111, fasc. 10 b1 Asie; Centre National de la Recherche Scientifique, Paris, 177 p.

____, L. F. RAMIREZ, C. D. REDMOND, AND E. L. ELBERG. 1966. Sedimentary geology of Saudi Arabia. In geology of the Arabian Peninsula, United States Geological Survey Professional Paper 560D, 147 p.

REDMOND, C. D. 1964a. The Foraminiferal family Pfenderinidae in the Jurassic of Saudi Arabia. Micropaleontology, 10 (2):251-263.

____. 1964b. Lituolid Foraminifera from the Jurassic and Cretaceous of Saudi Arabia. Micropaleontology, 10 (4):405-414.

____. 1965. Three new genera of Foraminifera from the Jurassic of Saudi Arabia. Micropaleontology, 11 (2):133-140.

SAMPO, M. 1969. Microfacies and Microfossils of the Zagros area, southwestern Iran. International Sedimentary Petrographical Series, V. XII. J. Cuvillier and H. Schurman, (eds.). M. E. Brill, Leiden, Netherlands, 102 p., 105 plates.

SMOUT, A. H. AND W. SUGDEN. 1961. New information on the foraminiferal genus *Pfenderina*. Palaeontology, 4 (4):581-591.

STEINEKE, M., AND R. A. BRAMKAMP. 1952. Mesozoic rocks of eastern Saudi Arabia (abstract). AAPG, 36(5): 909 p.

STEINEKE, M., R. A. BRAMKAMP AND N. J. SANDER. 1958. Stratigraphic relations of Arabian Jurassic oil, p. 1294-1329. *In* L. G. Weeks (ed.), Habitat of Oil, American Association of Petroleum Geologists Symposium.

SUGDEN, W. AND A. J. STANDRING. 1975. Qatar Peninsula. *In* Lexique Stratigraphique International, Vol. III, Asie, Fasc. 10a, 120 p.

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Ammonite Biostratigraphy of the Hettangian/Sinemurian Boundary in South America

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Keywords: Hettangian/Sinemurian Boundary, Ammonite Horizons Peru, Chile, Argentina, Biostratigraphic Correlation (Europe, North America)

Abstract: In northern Peru, Chile and northwestern Argentina many sections document ammonite horizons embracing the Hettangian/Sinemurian boundary. Only a few sections have yielded well-preserved specimens. Different ammonite horizons may be distinguished. The Lower Hettangian is characterized by species of the genera *Schlotheimia*, *Angulaticeras*, *Badouxia* and *Paracaloceras*. An horizon including a new species of *Pseudaetomoceras* is of latest Hettangian or earliest Sinemurian age. The next higher horizons contain different species of the genus *Coroniceras*. Other genera like *Angulaticeras* and ?*Vermiceras* are more rarely found. Correlations are possible with North American and European Hettangian and Sinemurian subzones of ammonites.

Introduction

Sections containing Upper Hettangian and Lower Sinemurian ammonites are known from northern Peru, Chile and northwestern Argentina. At many sections only laterally compressed specimens are found whose specific identification is difficult. Only a few sections are known with continuous sedimentation across the Hettangian/Sinemurian boundary. Most of them do not contain sufficient ammonite horizons for a detailed biozonation.

Peru

Prinz (1985), Hillebrandt (1994) and Prinz and Hillebrandt (1994) described Hettangian and Sinemurian sections from the Utcubamba Valley in northern Peru but transitional sections from Hettangian to Sinemurian strata have not yet been studied in detail. The Upper Hettangian is proved by *Badouxia canadensis* (Frebold) and a new species of *Schlotheimia* (Hillebrandt, 1994, Pl. 1, Fig. 2). Hyatt (1889) figured "*Caloceras*" (=*Alpinoceras* or *Paracaloceras*) *ortoni* from this region which probably proves the latest Hettangian. Prinz (1985, Fig. 12) decribed a section which was said to be of Early Sinemurian age but at least the first horizon with *Coroniceras brevidorsale* (=*Paracaloceras* with a suture line similar to that of *Vermiceras*) and the second horizon with *Vermiceras spiratissimum* (= *Gyrophioceras* with tubercles on the innermost whorls) are of latest Hettangian age (information kindly provided by Dr. G. Bloos, Stuttgart). The next horizon (approx. 1m higher) yielded *Arietites* (*Paracoroniceras*) cf. *nudaries*. Similar whorl fragments are found in northern Chile belonging to *Pseudaetomoceras* of latest Hettangian or earliest Sinemurian age. Several meters higher, different species of *Arnioceras* were found (Semicostatum Zone). The Bucklandi Zone was not in evidence in this section. Evolute arietitids (*Vermiceras* or *Metophioceras*) with densely ribbed inner whorls and more widely spaced ribs on the outer whorls were found at other sections of this region and probably represent the Conybeari Subzone.

Northern Chile

Chong and Hillebrandt (1985), Quinzio (1987) and Hillebrandt (1990) described Hettangian and Sinemurian sections from northern Chile. Most of these sections yielded mainly laterally compressed ammonites and a detailed biozonation was difficult.

Cerros de Cuevitas: The best sections with relief-preserved ammonites and embracing the Hettangian/Sinemurian boundary are found in this area (Fig. 1). Hettangian and Lower Sinemurian sediments cover an area which is 10km wide (E - W). An Upper Triassic to Middle Hettangian transgression from east to west was connected with a change of facies in the same direction. Middle to Upper Hettangian ammonites are mainly found in the middle part (Hillebrandt, 1990, Fig. 2) and the westernmost sections yielded uppermost Hettangian and Lower Sinemurian ammonites (Fig. 2). These sections allowed a detailed biozonation.

Quillagua: In the Coastal Cordillera west of Quillagua (fossil locality LF 2 in Maksaev and Marinovic, 1980) limestones yielded silicified ammonites of latest Hettangian and earliest Sinemurian age.

Aguada de Varas: Upper Hettangian ammonites of this locality were figured by Hillebrandt (1981) and the section was described by Quinzio (1987, Fig. 20, 21). A rich Late Hettangian ammonite fauna is found in the basal limestone bed. The sequence of Early Sinemurian ammonite horizons is incomplete. Quebrada San Juan: The section was described by Chong and Hillebrandt (1985, Fig. 2) and Hillebrandt (1990, Fig. 6). A thick series is exposed but ammonite bearing horizons are rare (Early Hettangian to Early Sinemurian age).

Central Chile

Cecioni and Westermann (1968) investigated an Upper Triassic to Lower Jurassic sequence at Los Molles (160km NW of Santiago de Chile). No well-preserved Sinemurian ammonites were found and the exact stratigraphic position of the Hettangian/Sinemurian boundary is unknown.

Escobar (1980) described sections with Hettangian and Sinemurian ammonites from the area of Curepto (210km SW of Santiago de Chile). Hettangian to Sinemurian strata are very thick but only a few horizons with mostly laterally compressed ammonites were found and a detailed biozonation is not possible. Between the last Hettangian ammonite assemblage (probably Middle Hettangian) and the first Sinemurian assemblage (Semicostatum Zone) the series is 400m thick and no ammonites were found.

Argentina

In the area of the westernmost outcrops of the Rio Atuel Jurassic (Mendoza province) an Hettangian to Lower Sinemurian sequence is found which was described by Riccardi et al. (1988; Riccardi, 1991) and Hillebrandt (1990). The Upper Hettangian to Lower Sinemurian sediments are

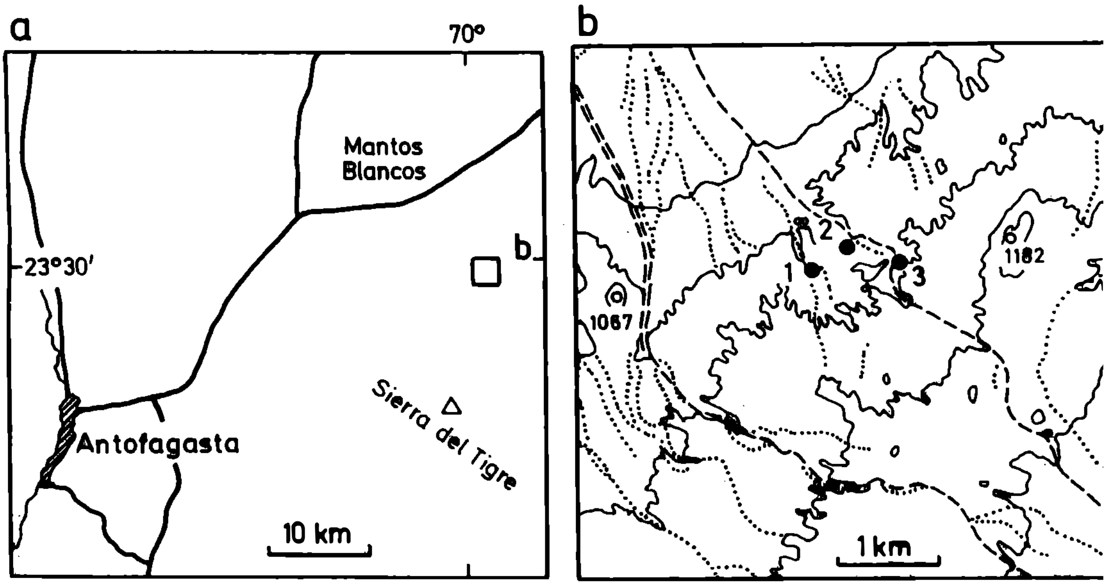

Figure 1. Index (a) and locality map (b) of the westernmost Cerros de Cuevitas sections.

500m thick and are exposed on the eastern bank of Arroyo Malo. Different ammonite horizons may be distinguished which can be correlated well with those found in northern Chile. The boundary between Hettangian and Sinemurian was drawn by Riccardi et al. (1988; Riccardi, 1991) much lower (more or less 100m) than by Hillebrandt (1990).

Biostratigraphy

Different faunal horizons may be distinguished in South America around the Hettangian/Sinemurian boundary. Faunal horizons are used in the sense of Page (1992). The defined horizons are based on the faunal succession within the Upper Hettangian and Lower Sinemurian rocks of South America (Fig. 3).

The South American Hettangian will be subdivided into different horizons, subzones and zones (Hillebrandt in press). South American Sinemurian subzones and zones cannot be established without a detailed description of the ammonites of this age and region. Some species are more or less closely related to species known from other regions and therefore enable a time correlation.

Upper Hettangian

The complete Hettangian biozonation will be described in Hillebrandt (in press). An horizon with *Sunrisites* and *Laqueoceras* is comparable with the upper part of the Middle Hettangian and probably ranges up to the lower part of the Upper Hettangian.

Horizon with *Schlotheimia* n. sp.: This horizon is found in Argentina, Chile and Peru.

Reference section: Cerros de Cuevitas (Hillebrandt, 1990, Fig. 2).

Fauna: The horizon is characterized by a new species of Schlotheimia (figured in Hillebrandt, 1990, Pl. 3, Fig. 7, 8 as Schlotheimia cf. montana). Badouxia canadensis (Hillebrandt, 1990, Pl. 3, Fig. 6) also appears.

Comment: Schlotheimia n. sp. was found at different sections of northern Chile (e.g., Quinzio, 1987, Pl. 2, Fig. 3-6) and northern Peru (Hillebrandt, 1994, Pl. 1, Fig. 2).

Age: The horizon may be correlated with the lower part of the European Angulata Zone.

Horizon with *Paracaloceras* n. sp.: The horizon was proved in Peru, Chile and Argentina.

Reference section: Aguada de Varas (Quinzio, 1987, Fig. 20, 21), first limestone bed of the section. The lowest part of this bed contains *Schlotheimia* n. sp.

Fauna: *Phylloceras* n. sp., *Angulaticeras* cf. *marmoreum* (Oppel), *Badouxia canadensis* (Hillebrandt, 1981, Pl. 1, Fig. 5 and Quinzio, 1987, Pl. 1, Fig. 11), *Paracaloceras* cf. *subsalinarium* (Wähner), *Paracaloceras* n. sp. (Hillebrandt, 1981, Pl. 1, Fig. 7 and Quinzio, 1987, Pl. 1, Fig. 10), *Schreinbachites* (?) sp., *Ectocentrites* cf. *petersi* (Hauer) (Hillebrandt, 1981, Pl. 1, Fig. 6), *Eolytoceras* n. sp.

Comment: This horizon was also demonstrated in the Cerros de Cuevitas region (Fig. 2) and other sections of northern Chile. The *Badouxia canadensis* Zone and the *Vermiceras* Assemblage Zone in Riccardi et al. (1988; 1991) are of the same age. *Vermiceras* cf. *gracile* (Spath) in Riccardi et al. (1988, Pl. 2, Fig. 9; 1991, Fig. 5.1) corresponds to *Paracaloceras* n. sp. and was found by the author in the same section together with *Badouxia canadensis*.

Age: The horizon may be correlated with the upper part of the European Angulata Zone and the lower part of the Canadensis Zone in North America.

Horizon with *Pseudaetomoceras* n. sp.: This horizon was demonstrated in different sections of northern Chile and Argentina.

Reference section: Cerros de Cuevitas (Fig. 2, section 3).

Fauna: Pseudaetomoceras n.sp. (= Agassiceras sp. in Hillebrandt, 1981, Pl. 2, Fig. 1) was found together with Eolytoceras sp.

The same lithologic horizons were observed at sections 1 and 2. The horizon was also found between section 1 and 2. Section 1 yielded a large specimen of Paracaloceras(?) sp. similar to P. coregonense (Wähner).

Comment: The horizon with Pseudaetomoceras n. sp. appears at different sections of northern Chile (e.g., Aguada de Varas). Pseudaetomoceras n. sp. was also figured from Argentina (Riccardi et al., 1988, Pl. 3, Fig. 1; 1991, Fig. 5.6). The author found Pseudaetomoceras(?) sp. together with Badouxia cf. canadensis (transitional form to B. columbiae (Frebold)) in the same section.

Age: Species of Pseudaetomoceras and Paracaloceras coregonense were described by Wähner (1886; 1888) from the Megastoma and Marmoreum Zones (Upper Hettangian). In North America (Guex, 1995) Pseudaetomoceras appears in the upper part of the Middle and the lower part of the Upper Hettangian. The horizon with Pseudaetomoceras n. sp. probably still is of Hettangian age but an earliest Sinemurian age may not be excluded.

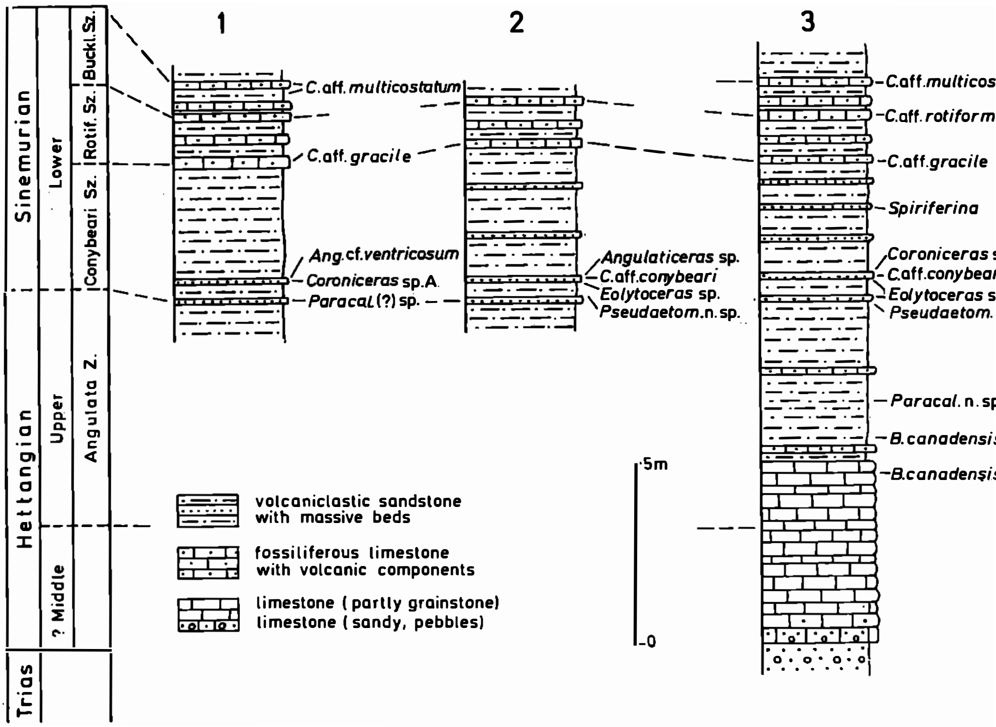

Figure 2. Hettangian to Sinemurian sections in the western region of Cerros de Cuevitas (section numbers correspond to Fig. 1b).

Lower Sinemurian

Different horizons of Early Sinemurian age may be distinguished. Only those of the lower part (Bucklandi Zone) are described.

Horizon with *Coroniceras* aff. *conybeari*: This horizon was shown to exist in Chile, Argentina and ?Peru.

Reference section: Cerros de Cuevitas (Fig. 2, section 3).

Fauna: *C.* (*Metophioceras*) aff. *conybeari* (Sowerby) (Pl. 1.1a, b) is accompanied by a species of *Coroniceras* (?*Coroniceras*) sp. A (Pl. 1.4a, b, Fig. 4.3) with widely spaced ribs, a whorl width much broader than the whorl height and a suture line with an external lobe much deeper than the lateral lobe (Fig. 4.3a). The species is similar to specimens described and figured by Wähner (1894, Pl. 5.3, Figs. 7, 8) under "*Arietites*" *bucklandi* and "*Arietites*" *altesulcatus*. The horizon of the reference section yielded also *Eolytoceras* sp. (Pl. 2.2a, b).

The horizon may be followed to the west (Figs. 1, 2). *C.* (*M.*) aff. *conybeari* (Pl. 1.7a, b) and *C.* (?*C.*) sp. *A* (Pl. 1.1, 2) are accompanied by *Angulaticeras* cf. *ventricosum* (Sowerby) (Pl. 2.8a, b), *Badouxia* cf. *canadensis* (Pl. 2.5,6a, b) and *Eolytoceras* sp. (Pl. 2.3a, b). A fragment of *C.* (*Coroniceras*) cf. *deffneri* (Oppel) (Pl. 2.4a, b) was found at section 2 (Fig. 2) but not *in situ*.

Comment: At some sections appears *Vermiceras*(?) sp. (Pl. 3.3a-c) which is similar to *V.*(?) cf. *coregonense* and *V.*(?) *multicostatum* in Frebold (1967). These specimens are characterized by densely ribbed inner whorls and probably also belong to this horizon. West of Quillagua this horizon also yielded a relatively involute *Coroniceras* (?) sp. (Pl. 3.6a, b) and *Phylloceras* sp. (Pl. 3.7a, b).

The horizon was also shown to exist at the Arroyo Malo section in Argentina. *Coroniceras* (*M.*) aff. *conybeari* appears together with *Coroniceras* (*?Coroniceras*) sp. A (Pl. 1.3a, b, Pl. 3.2.a, b, Fig. 4.2).

Age: The horizon is coeval at least with part of the Conybeari Subzone (earliest Sinemurian) in Europe and comparable with the uppermost part of the Canadensis Zone in North America.

Horizon with *Coroniceras* aff. *gracile*: The horizon was found in northern Chile, Argentina and probably also northern Peru.

Reference section: Arroyo Malo (Mendoza province) (section described by Riccardi et al., 1988, Fig. 2, probably levels 1470 to 1472).

Fauna: At least three beds yielded *C.* (*Metophioceras*) aff. *gracile* Spath (Pl. 2.1a, b, Pl. 3.1a, b, Fig. 4.4) (probably = *Coroniceras* (*Paracoroniceras*) cf. *charlesi* Donovan *in* Riccardi et al., 1988; 1991). No other ammonites were observed.

Comments: This horizon was also found at the westernmost outcrops of the Cerros Cuevitas (Fig. 2). The first one or two beds of a series (2 to 3m thick) of well bedded fossiliferous limestones (beds 10 to 20cm thick) yielded *C.* (*M.*) aff. *gracile*. Furthermore, the species was also found in other sections of northern Chile (e.g., Quebrada San Juan). A specimen similar to *C.* (*M.*) aff. *gracile* was observed in the uppermost part of the Chilingote section (Hillebrandt, 1994, p. 304) of northern Peru.

Age: The horizon is comparable with the upper part of the Conybeari Subzone or the lower part of the Rotiforme Subzone in Europe. In North America the uppermost part of the Canadensis Zone or the lowermost part of the "*Coroniceras*" Assemblage (Palfy et al., 1994) may be correlated with this horizon.

	Europe DEANetal.(1961) BLOOS (1988) Zone/ Subzone		South America Horizon	North America PALFYetal. (1994) Zone/ Assemblage
Lower Sinemurian	Semicostatum			Arnouldi
	Bucklandi	Bucklandi	*C.aff. multicostatum*	"Coroniceras"
		Rotiforme	*C.aff. rotiforme*	
			C.aff..gracile	
		Conybeari	*C. aff. conybeari*	Canadensis
			Pseudaetomoceras n.sp.	
Hettangian	Angulata	Depressa		
		Complanata	*Paracaloceras* n.sp.	
		Extranodosa	*Schlotheimia* n.sp.	
	Liasicus		*S. peruvianus*	

Figure 3. South American Late Hettangian and Early Sinemurian ammonite horizons and correlation with Europe and North America.

Horizon with *Coroniceras* aff. *rotiforme*: The horizon was mainly found in northern Chile but probably also exists in Argentina. No reference section is given because the results of the field observations are not yet sufficient.

Comments: Whorl fragments of *C.* (*Coroniceras*) aff. *rotiforme* (Sowerby) (Fig. 4.1a-c) are found at Cerros de Cuevitas (Fig. 2, section 3) above beds with *C.* (*M.*) aff. *gracile*. A small incomplete specimen (Ø 23mm) comes from approx. 350m SE of section 3. Coiling, ribbing and cross section are very similar to a small specimen figured by Guérin-Franiatte (1966, Pl.

11, Fig. 1). A fragment (Pl. 3.5a-c) with a whorl section broader than high but more closely coiled than *C.* (*C.*) *rotiforme* was found near section 2 (Fig. 2). A whorl fragment (phragmocone) of a giant *C.* (*C.*) aff. *rotiforme* (whorl width 128mm, whorl height 107mm) was collected from the middle part of the Cerros de Cuevitas outcrops (section 2 in Hillebrandt, in press). The distant ribs are nearly rectiradiate. The uppermost beds containing *Coroniceras* (= *C.* (*C.*) *alcinoe* in Riccardi et al., 1988, 1991) in Argentina (Arroyo Malo) probably belong to this horizon.

Age: At least part of the horizon belongs to the Rotiforme Subzone in Europe but it can not be ruled out that the horizon extends up to the Bucklandi Subzone. The horizon corresponds to part of the "*Coroniceras*" Assemblage in North America (Fig. 3).

Horizon with *Coroniceras* aff. *multicostatum*: Up to now the horizon was observed only in northern Chile. No reference section is given because field observations are not yet sufficient.

Comments: More or less densely ribbed arietitids are closely related to *C.* (*Coroniceras*) *multicostatum* (Sowerby). Large specimens (Ø up to 35cm) were found in a layer with large concretions (section exposed in Quebrada El Griton 19km SSE Taltal). An isolated specimen (Ø 21cm, phragmocone) was presented to the author by Dr. T. Bogdanic (formerly Universidad del Norte, Antofagasta). The specimen comes from the Quebrada Pan de Azucar (Quinzio, 1987, Fig. 12).

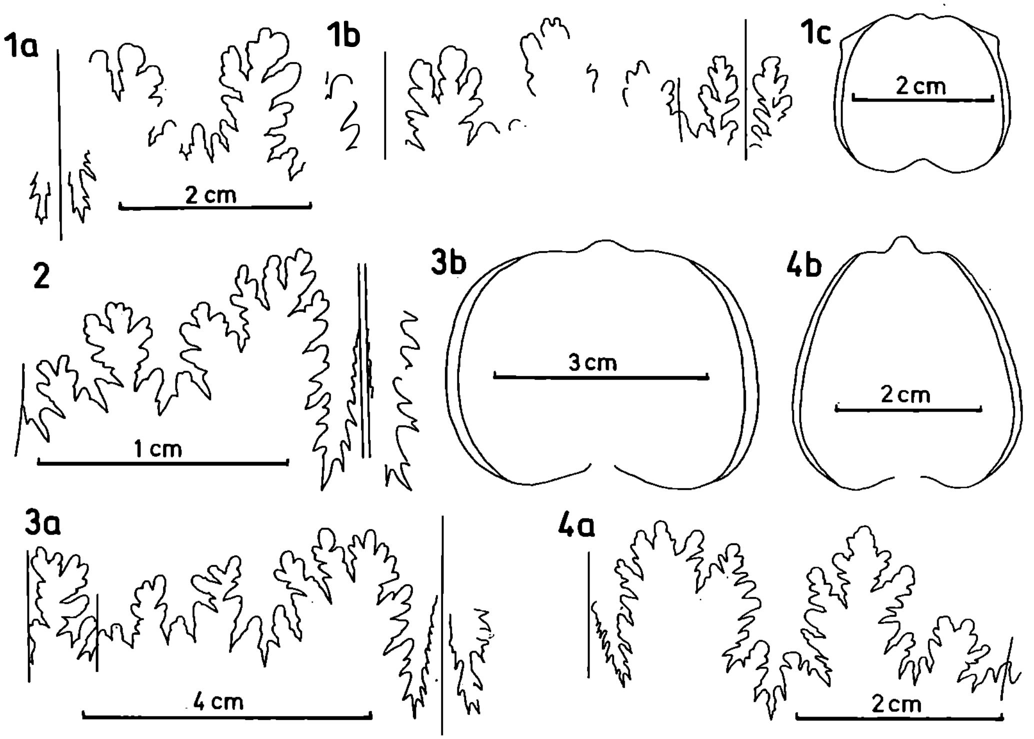

Figure 4. Suture lines and cross sections.
1 a,b,c. *C.* (*Coroniceras*) aff. *rotiforme* (Sowerby), Cerros de Cuevitas (Fig. 2, section 3), *C.* aff. *rotiforme* horizon, WH = 24mm (TUB 790323/2/1) (specimen not figured). **2.** *C.* (?*Coroniceras*) sp. A, Arroyo Malo (Mendoza province), *C.* aff. *rotiforme* horizon, WH=9mm (TUB 880104/5/1) (=Pl. 1, Fig. 3). **3 a, b.** *C.* (?*Coroniceras*) sp. A, Cerros de Cuevitas (Fig. 2, section 3), *C.* aff. *conybeari* horizon, WH = 34mm (TUB 790323/3/1) (=Pl. 1, Fig. 4). **4 a, b.** *C.* (*Metophioceras*) aff. *gracile* Spath, Arroyo Malo (Mendoza province), *C.* aff. *gracile* horizon, WH = 35mm (TUB 880103/1/1) (=Pl. 3, Fig. 1).

Cerros des Cuevitas (Fig. 2, sections 1-3): *C.* aff. *multicostatum* (mostly fragments) was found in the upper part of the limestone beds with *C.* aff. *gracile* and *C.* aff. *rotiforme* and is accompanied by an arietitid broadly resembling *Coroniceras* (*C.*) *lyra* (Hyatt).

Age: The horizon probably belongs to the Bucklandi Subzone (?upper part) but a slightly younger age (Lyra Subzone) cannot be ruled out.

Above the horizon with *C.* aff. *multicostatum* the following beds contain mainly *Arnioceras* spp. (Semicostatum Zone) and their occurrence is proved at many sections in northern Chile (Quinzio 1987).

References

BLOOS, G. 1988. On the stage boundary Hettangian/Sinemurian in North-West Europe and in the North-Eastern Alps. 2nd International Symposium on Jurassic Stratigraphy. 1:71-83.

CECIONI, G., AND G. E. G. WESTERMANN. 1968. The Triassic/Jurassic Marine Transition of Coastal Central Chile. Pacific Geology. 1:41-75.

CHONG, D., AND A. v. HILLEBRANDT. 1985. El Triásico Preandino de Chile entre los 23°30' y 26°00' de Lat. Sur. Actas IV Congreso Geologico Chileno. 1:1-162 - 1-210.

DEAN, W. T., D. T. DONOVAN, AND M. K. HOWARTH. 1961. The Liassic ammonite zones and subzones of the North-West European province. Bulletin of the British Museum, Geology. 4:435-505.

ESCOBAR, T. F. 1980. Paleontología y bioestratigrafía del Triásico superior y Jurásico inferior en el Area de Curepto. Provincia de Talca. Instituto de Investigaciones Geologicas, Boletin. 35, 1-78.

FREBOLD, H. 1967. Hettangian Ammonite Faunas of the Taseko Lakes area, British Columbia. Geological Survey of Canada Bulletin, 158.

GUÉRIN-FRANIATTE, S. 1966. Ammonites du Lias inférieur de France.Centre National de la Recherche Scientifique, Paris, 455 p.

GUEX, J. 1995. Ammonites hettangiennes de la Gabbs Valley Range, Nevada, USA. Mémories de Geologie, Lausanne. 27, 130 p.

HILLEBRANDT, A. v. 1981. Faunas de amonites del liasico inferior y medio (Hettangiano hasta Pliensbachiano) de America del Sur (excluyendo Argentina), p. 499-538. *In* W. Volkheimer and E. A. Musacchio (eds.), Cuencas Sedimentarias del Jurásico y Cretácico de América del Sur.2.

___. 1990. The Triassic/Jurassic boundary in northern Chile. Les Cahiers de l'Université Catholique de Lyon, Série Sciences. 3:27-53.

___. 1994. The Triassic/Jurassic Boundary and Hettangian Biostratigraphy in the Area of the Utcubamba Valley (northern Peru). Geobios, M.S. 17:297-307.

___. (in press). Die Ammoniten-Fauna des südamerikanischen Hettangium (basaler Jura). Palaeontographica Abt.

HYATT, A. 1889. Genesis of the Arietidae. Smithonian Contributions to Knowledge. 673, 238 p.

MAKSAEV J. V. AND N. MARINOVIC S. 1980. Cuadrangulos Cerro de La Mica, Quillagua, Cerro Posada y Oficina Prosperidad. Instituto de Investigaciones Geologicas. Carta Geologica de Chile. 45-48, 63 p.

PAGE, K. N. 1992. The sequence of ammonite correlated horizons in the British Sinemurian. Newsletter on Stratigraphy. 27:129-156.

PÁLFY, J., P. L. SMITH, AND H. W. TIPPER. 1994. Sinemurian (Lower Jurassic) Ammonoid biostratigraphy of the Queen Charlotte Islands, Western Canada. Geobios, M.S. 17: 385-393.

PRINZ, P. 1985. Stratigraphie und Ammonitenfauna der Pucara-Gruppe (Obertrias-Unterjura) von Nord-Peru. Palaeontographica, A, 188:153-195.

___, AND A. v. HILLEBRANDT. 1994. Stratigraphy and Ammonites of the North Peruvian Pucará Group. Palaeontographica, A, 233:33-62.

QUINZIO SINN, L. A. 1987. Stratigraphische Untersuchungen im Unterjura des Südteils der Provinz Antofagasta in Nord-Chile. Berliner geowissenschaftliche Abhandlungen, A 87:100 p.

RICCARDI, A. C. 1991. Hettangian and Sinemurian (Lower Jurassic) biostratigraphy of Argentina. Journal of South American Earth Sciences. 4 (3):159-170.

___, S. E. DAMBORENEA, M. O. MANCEÑIDO, AND S. C. BALLENT. 1988. Hettangiano y Sinemuriano marinos en Argentina. V Congreso Geologico Chileno. 2:359-373.

WÄHNER, F. 1882-1898. Beiträge zur Kenntnis der tieferen Zonen des unteren Lias der nordöstlichen Alpen. Beiträge zur Palaeontologie von Oesterreich-Ungarn und des Orients. 2-11, 291 p.

Plate 1. Bar scale at top left for 1a,b and bar scale at top right for **2** to **7**. **1-4,6.** *Coroniceras* (?*Coroniceras*) sp. A, lateral and external views, *C.* aff. *conybeari* horizon; **1,2,6.** Cerros de Cuevitas (Fig. 2, section 1) (TUB 970110/7/1,2,3); **3.** Arroyo Malo (Mendoza province) (TUB 880104/5/1), (=Fig. 4.2); **4.** Cerros de Cuevitas (Fig. 2, section 3) (TUB 790323/3/1) (=Fig. 4.3); **5,7.** *Coroniceras* (*Metophioceras*) aff. *conybeari* (Sowerby), lateral, external and sagittal views, *C.* aff. *conybeari* horizon; **5.** Cerros de Cuevitas (Fig. 2, section 3) (TUB 970112/7/1); **7.** Cerros de Cuevitas (Fig. 2, section 2) (TUB 90110/10/1).

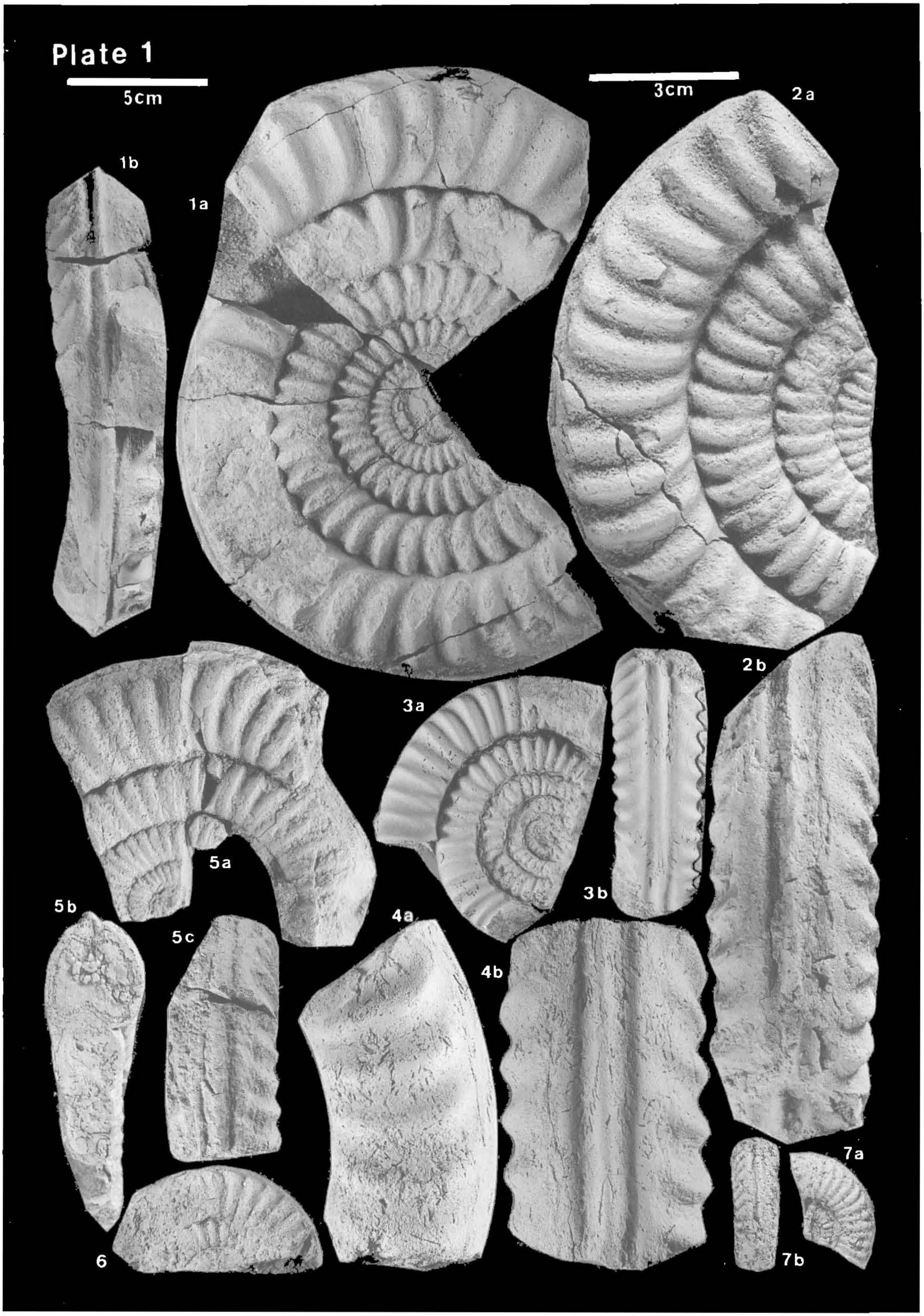
Plate 1
5cm
3cm
1a
1b
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7a
7b

Plate 2. Bar scale at top left for 1a,b and bar scale at top right for 2-8. **1a,b.** *Coroniceras* (*Metophioceras*) aff. *gracile* Spath, lateral and external views, *C.* aff. *gracile* horizon, Arroyo Malo (Mendoza province) (TUB 880103/2/1); **2,3.** *Eolytoceras* sp., lateral and external views, *C.* aff. *conybeari* horizon; **2.** Cerros de Cuevitas (Fig. 2, section 3) (TUB 970112/7/2); **3** Cerros de Cuevitas (Fig. 2, section 2) TUB 970110/10/2); **4a,b.** *Coroniceras* (*Coroniceras*) cf. *deffneri* (Oppel), lateral and ventral views, *C.* aff. *conybeari* horizon, Cerros de Cuevitas (Fig. 2, section 2) (TUB 970110/11/1); **5,6.** *Badouxia* cf. *canadensis* (Frebold), lateral and external views, *C.* aff. *conybeari* horizon; **5.** Cerros de Cuevitas (Fig. 2, between section 1 and 2) (TUB 970112/5/1); **6a,b.** Cerros de Cuevitas (Fig. 2, between section 2 and 3) (TUB 970112/1/1); **7,8.** *Angulaticeras* cf. *ventricosum* (Sowerby), lateral and ventral views; **7.** ? *C.* aff. *conybeari* horizon, W Quillagua (TUB 28-22); **8.** *C.* aff. *conybeari* horizon, Cerros de Cuevitas (Fig. 2, section 1) (TUB 970110/7/4).

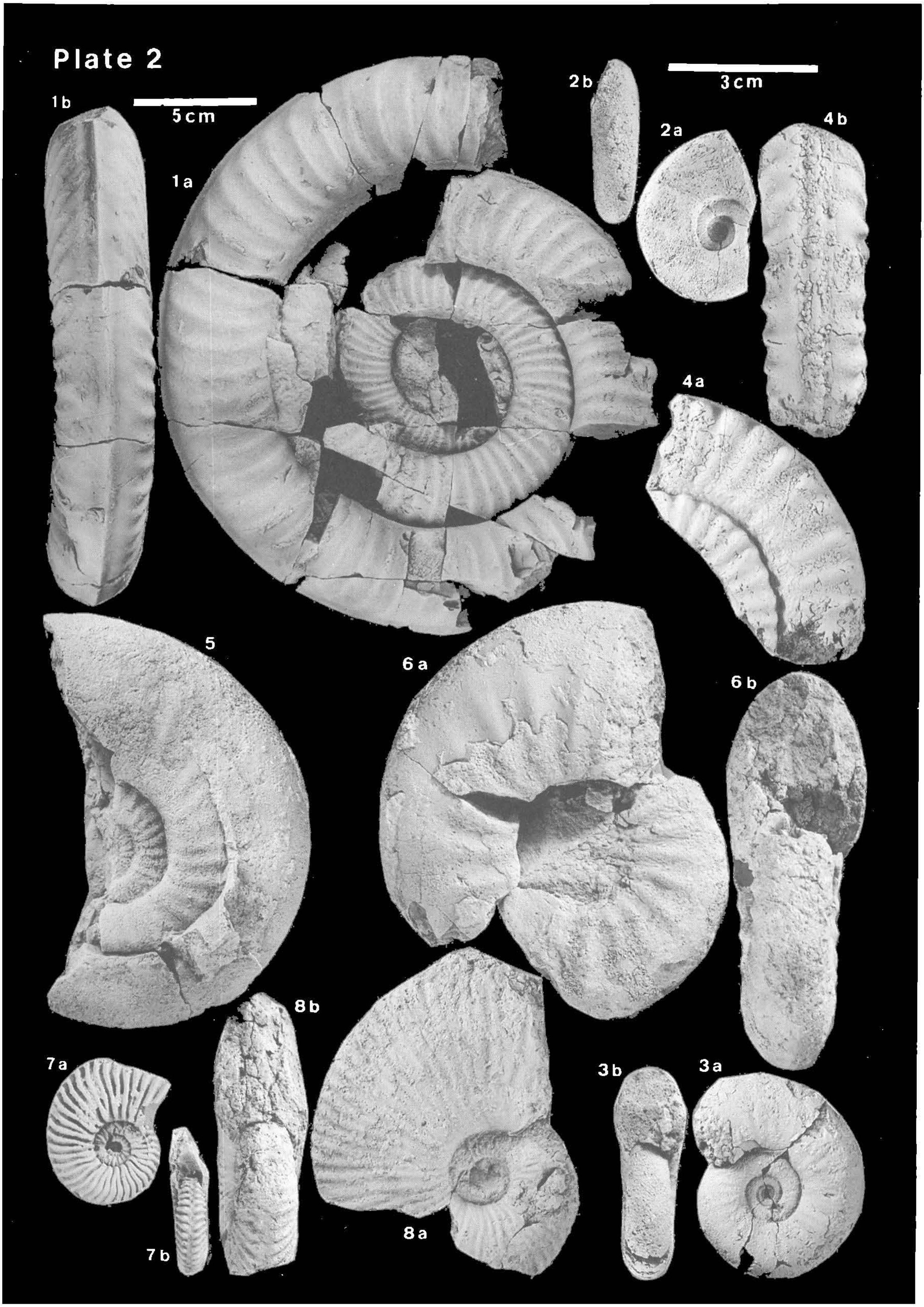

Plate 2

Plate 3. **1a,b.** *Coroniceras* (*Metophioceras*) aff. *gracile* Spath, lateral and external views, *C.* aff. *gracile* horizon, Arroyo Malo (Mendoza province) (TUB 880103/1/1) (=Fig. 4.4); **2a,b.** *Coroniceras* (*?Coroniceras*) sp. A, lateral and external views, *C.* aff. *conybeari* horizon, Arroyo Malo (Mendoza province) (TUB 880105/5/2); **3a-c.** *Vermiceras*(?) sp., lateral and external views, ? *C.* aff. *conybeari* horizon, W Quillagua (TUB 860303/2/1); **4a,b.** *Coroniceras* (*Metophioceras*) aff. *conybeari*, lateral and sagittal views, ? *C.* aff. *conybeari* horizon, W Quillagua (TUB 860303/2/1); **5a-c.** *Coroniceras* (*Coroniceras*) sp. ex gr. *C. rotiforme* (Sowerby), lateral and external views, *C.* aff. *rotiforme* horizon, Cerros de Cuevitas (Fig. 2, near section 2) (TUB 970108/5/1); **6a,b.** *Coroniceras*(?) sp., lateral and external views, ? *C.* aff. *conybeari* horizon, W Quillagua (TUB 860303/2/2); **7a,b.** *Phylloceras* sp., lateral and sagittal views, ? *C.* aff. *conybeari* horizon, W Quillagua (TUB 860303/2/3).

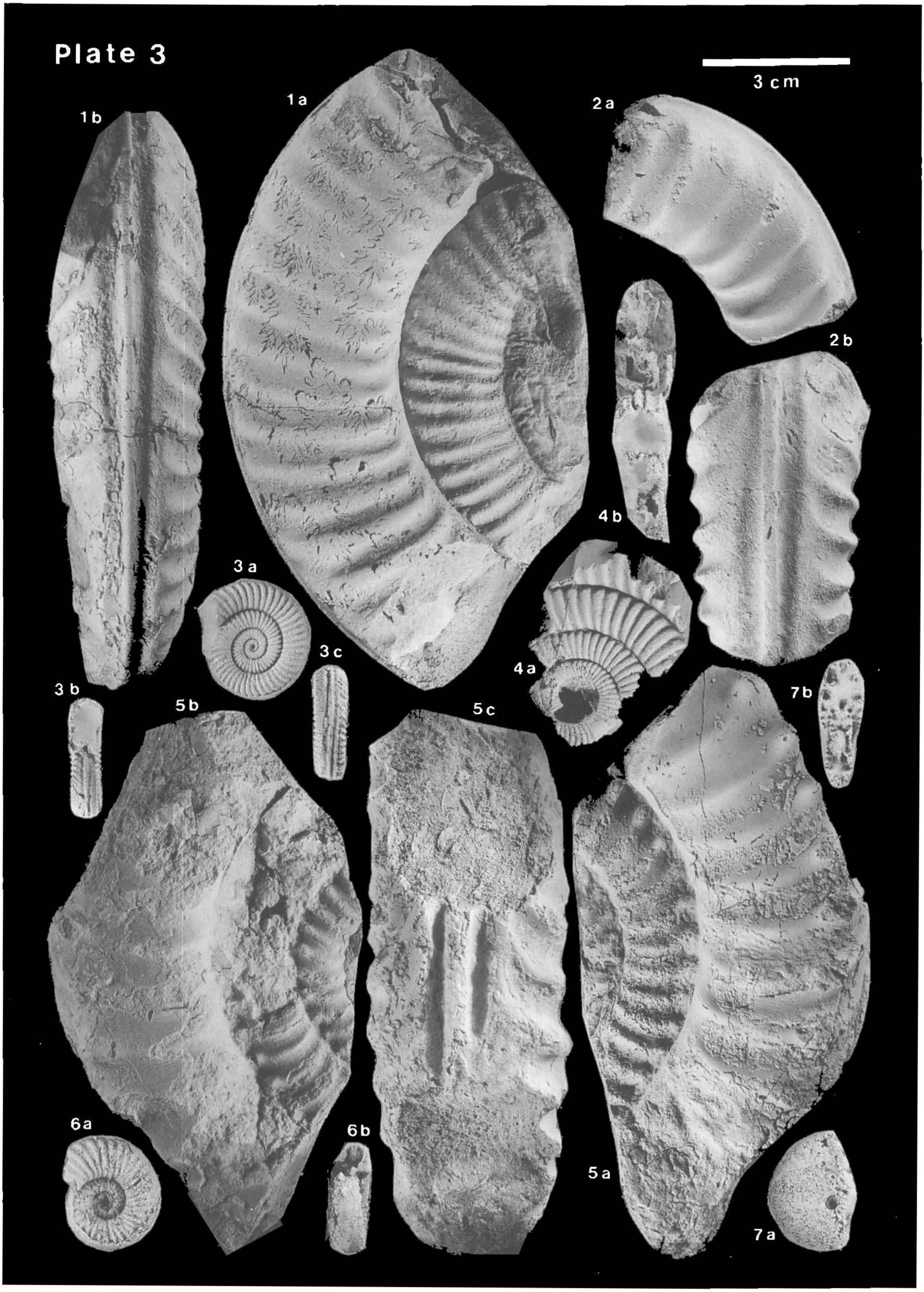

Plate 3

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A Direct Correlation between North American and Japan-Pacific Radiolarian Zonal Schemes for the Upper Jurassic

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Keywords: Upper Jurassic, Radiolaria, Biostratigraphy, Zonal Definition, Correlation, Japan, West Pacific, North America, ODP Site 801

Abstract: Ocean Drilling Program (ODP) Site 801B section, located in the west Pacific, is one of the type sections for the Upper Jurassic radiolarian zonation of the Japan-Pacific region. The *Hsuum maxwelli* Zone (JR 7), *Loopus primitivus* Zone (JR 8), *Pseudodictyomitra carpatica* Zone (KR 1), and *Cecrops septemporatus* Zone (KR 2) are successively recognized in a continuous sequence of brown radiolarites in this section. Our research indicates that both the North American and Japan-Pacific zonations are equally applicable to the west Pacific section, with certain marker taxa providing direct ties between the two zonal schemes. Seven zone-diagnostic bio-events, such as the first occurrence biohorizon (FOB), last occurrence biohorizon (LOB), and evolutionary first appearance biohorizon (EFAB) are identified in the ODP Site 801B section. These events include: (1) LOB of *Mirifusus guadalupensis* Pessagno, (2) LOB of the *Hsuum maxwelli* Pessagno group, (3) FOB of *Vallupus hopsoni* Pessagno and Blome, (4) LOB of Genus *Perispyridium* Dumitrica, (5) EFAB of *Pseudodictyomatra carpatica* (Lozyniak), (6) FOB of *Acanthocircus dicranacanthos* (Squinabol), and (7) LOB of *Ristola altissima* (Rüst). The North American and Japan-Pacific radiolarian zones can be correlated using these bio-events as follows: North American Zone 3/Zone 4 and Zone 4/Zone 5 boundaries are situated within the Japan-Pacific *Loopus primitivus* Zone (JR 8) and *Pseudodictyomitra carpatica* Zone (KR 1), respectively; and conversely, the Japan-Pacific *Hsuum maxwelli* (JR 7)/*Loopus primitivus* (JR 8) and *Loopus primitivus* (JR 8)/*Pseudodictyomitra carpatica* (KR 1) zonal boundaries are situated within North American Zone 3 (Subzone 3α) and Zone 4 (Subzone 4α), respectively.

Introduction

In biostratigraphic studies of Jurassic Radiolaria, parallel zonations have been proposed based on data from North America (Pessagno, 1977; Pessagno et al., 1984, 1987, 1989, 1993), Europe (Baumgartner, 1984, 1987), and Japan and the west Pacific (Matsuoka and Yao, 1986; Matsuoka, 1992, 1995a), which are quite different in their definition and are often considered difficult to correlate with each other. Although challenges to develop a globally acceptable zonation have been initiated (Baumgartner et al., 1995), the results are still being refined.

Middle Jurassic-Lower Cretaceous radiolarians were recovered from a deep ocean basin in the western Pacific by Leg 129 of the Ocean Drilling Program (ODP). The Pacific ODP Site 801B section (Fig. 1) is continuous, well-sampled for biostratigraphic purposes, and contains abundant, well-preserved radiolarians. This section is one of the best among sections available for examining the Upper Jurassic-Lower Cretaceous radiolarian succession. Radiolarian biostratigraphic research of the section was performed by Matsuoka (1992), basically applying a zonal scheme from Japan. Later, Matsuoka (1995a) recognized biostratigraphically significant bio-events in the section and defined radiolarian zones using these bio-events. The Site 801B section is regarded as the type for the Upper Jurassic-Lower Cretaceous interval in the Japan-Pacific region.

Although correlation of Jurassic radiolarian zones established on both sides of the Pacific has been attempted (Pessagno and Mizutani, 1992; Matsuoka, 1995b; Pessagno and Hull, 1996), the results are still controversial. In this study we attempt to apply criteria of the North American zonation (Pessagno et al., 1993) to the Upper Jurassic-Lower Cretaceous section at ODP Site 801. This approach can lead to better understanding of zonal correlation, and eventually to the creation of a more reliable biostratigraphic framework for the Upper Jurassic and Lower Cretaceous. Preliminary results were introduced by Yang and Matsuoka (1997), but lacked the raw data set which is essential for obtaining the results; the data set is included in this paper.

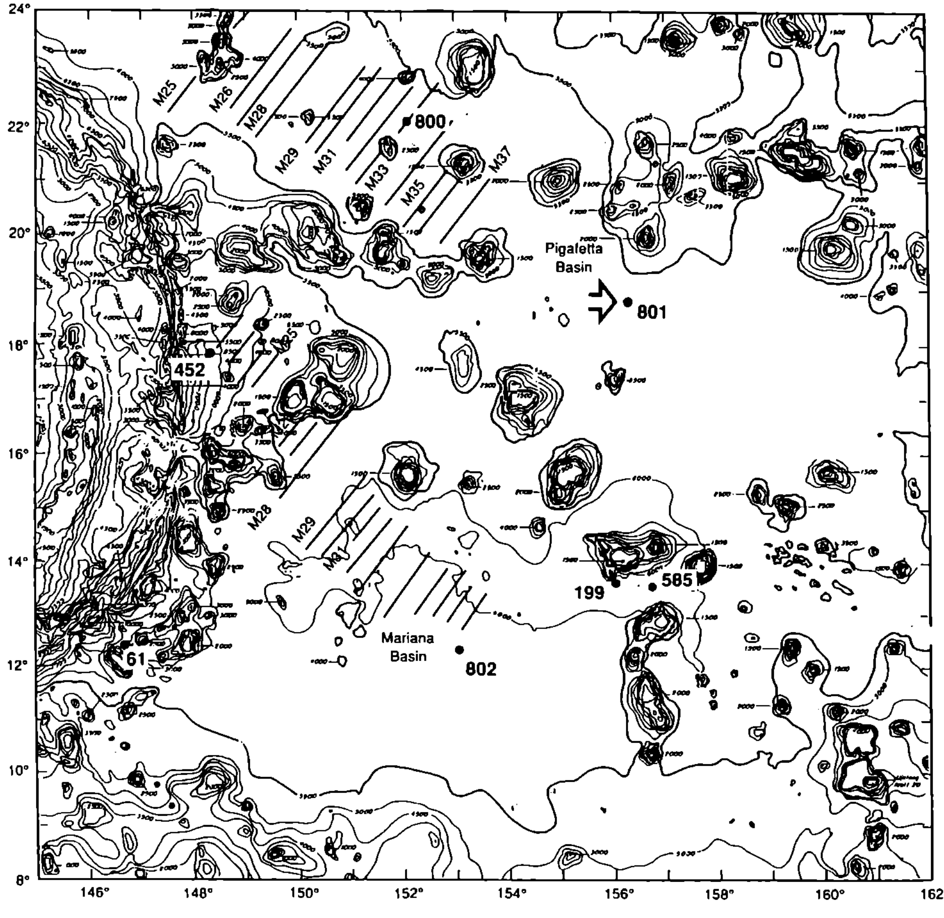

Figure 1. Index map of ODP Site 801.

Stratigraphic Summary and Materials Studied

ODP Site 801 (Fig. 1) is located in the central Pigafetta Basin (18°38.54'N, 156°21.58'E) at a water depth of 5,682m. It is situated on a magnetically quiet zone southwest of the M25-M37 magnetic lineation sequence. According to Lancelot et al. (1990), strata at this site are divided into six lithostratigraphic units. The radiolarian samples studied come from Unit IV (Core 14R to 32R). The base of Unit IV is an unconformable contact with alternating beds of red radiolarite and claystone of Unit V, while the top is overlain by Albian volcano-clastic turbidites of Unit III. Unit IV represents a continuous sedimentary sequence consisting of 124.6m of brown radiolarites and clayey radiolarites with scattered dark brown porcellanite bands and nodules. The porcellanite bands and nodules are more abundant in the upper 82m interval than in lower part of the unit. Radiolarians are abundant and fairly well-preserved in this interval. With the exception of rare nannofossils of Tithonian age in Cores 801B-25R and -26R (Erba and Covington, 1992), the studied section is barren of age-diagnostic fossils other than radiolarians. Paleomagnetic data indicate that ODP Site 801 was located in the vicinity of the paleo-equator during latest Jurassic (Tithonian) time (Steiner and Wallick, 1992).

Radiolarian Biostratigraphy

The radiolarian samples were examined using transmitted-light microscopy, reflected-light microscopy, and scanning electron microscopy. Zonal definitions for the Upper Jurassic to Lower Cretaceous in Japan-Pacific (Matsuoka, 1995a) and for the Upper Jurassic of North America (Pessagno et al., 1993) are presented in Figure 2 and Figure 3, respectively. All zone-diagnostic taxa for the Japan-Pacific zonation were recovered from the ODP materials. However, marker taxa for the North American zonation are not always found in the ODP materials. Occurrence data of selected, biostratigraphically important radiolarian taxa are presented in Fig. 4; other occurrence data are included in Matsuoka (1992, 1995c). The results of this dual zonal scheme approach are outlined in Figure 5.

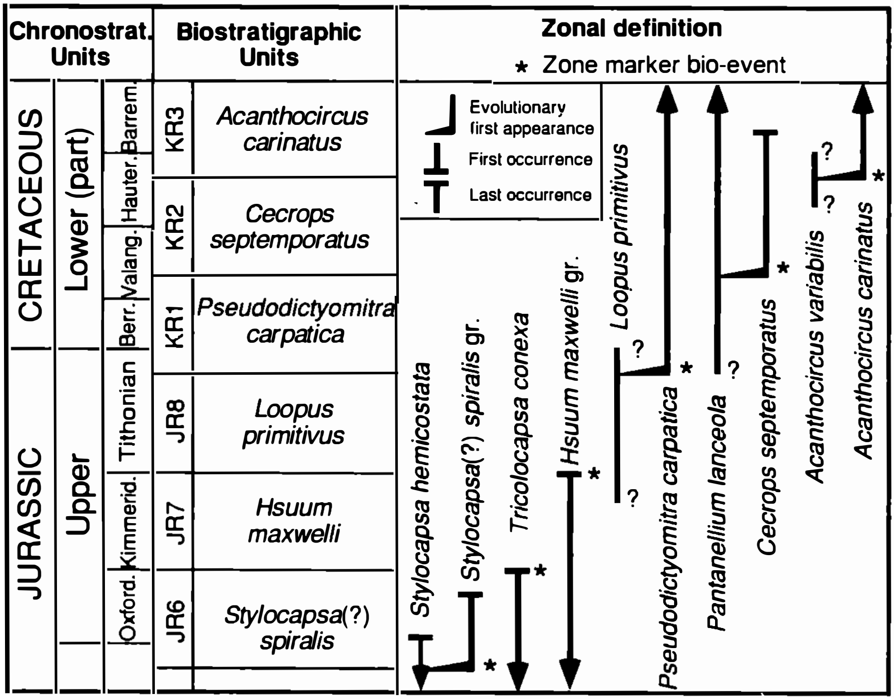

Figure 2. Radiolarian zonal scheme for the Upper Jurassic-Lower Cretaceous in Japan-Pacific. After Matsuoka (1995a).

Biostratigraphic Analysis Using Japan-Pacific Zonal Criteria

The studied interval (Cores 14 to 32) in the section includes the *Hsuum maxwelli* (JR 7), *Loopus primitivus* (JR 8), *Pseudodictyomitra carpatica* (KR 1), and *Cecrops septemporatus* (KR 2) zones (Fig. 5).

The *Hsuum maxwelli* Zone is present in the lowest part of the studied interval. The lower part of this zone is not present due to an unconformity between Cores 32R and 33R. The upper limit is defined by the last occurrence biohorizon (LOB) of the *Hsuum maxwelli* Pessagno group and is drawn between Samples 29R-CC and 29R-1, 16-17cm.

The *Loopus primitivus* Zone is represented by the interval between the LOB of the *H. maxwelli* Pessagno group and the evolutionary first appearance biohorizon (EFAB) of *Pseudodictyomitra carpatica* Lozyniak. The lower and upper limits are drawn between Samples 29R-CC and 29R-1, 16-17cm and between Samples 21R-1, 1-3cm and 20R-CC, respectively.

The *Pseudodictyomitra carpatica* Zone is represented by the interval between the EFAB of *P. carpatica* and the EFAB of *Cecrops septemporatus* (Parona). The lower and upper limits are drawn between Samples 21R-1, 1-3cm and 20R-CC and between Samples 15R-1, 23-25cm and 14R-CC, respectively.

The *Cecrops septemporatus* Zone is identified in the uppermost part of the studied interval. The lower limit of this zone is defined by the EFAB of *C. septemporatus* and is drawn between Samples 15R-1, 23-25cm and 14R-CC. The lowermost part only is presented.

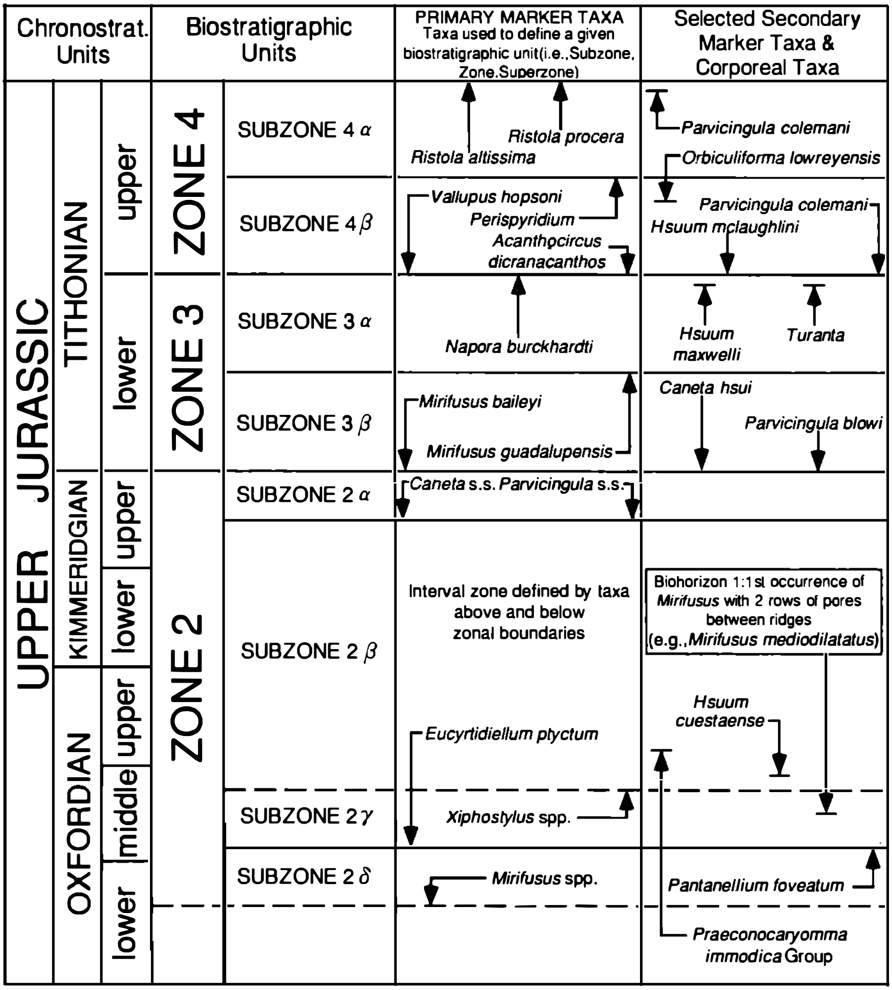

Figure 3. Radiolarian zonal scheme for the Upper Jurassic in North America. After Pessagno et al. (1993).

Biostratigraphic Analysis Using North American Zonal Criteria

When North American zonal criteria (Fig. 3) are applied to the core from Site 801B, Zone 3, Zone 4, and Zone 5 are successively recognized in the studied interval (Fig. 5). The boundary between Zone 3 and Zone 4 is defined by three bio-events involving primary marker taxa; (1) the LOB of *Napora burckhardti* Pessagno, Whalen, and Yeh, (2) the FOB of *Vallupus hopsoni* Pessagno and Blome, and (3) the FOB of *Acanthocircus dicranacanthos* (Squinabol).

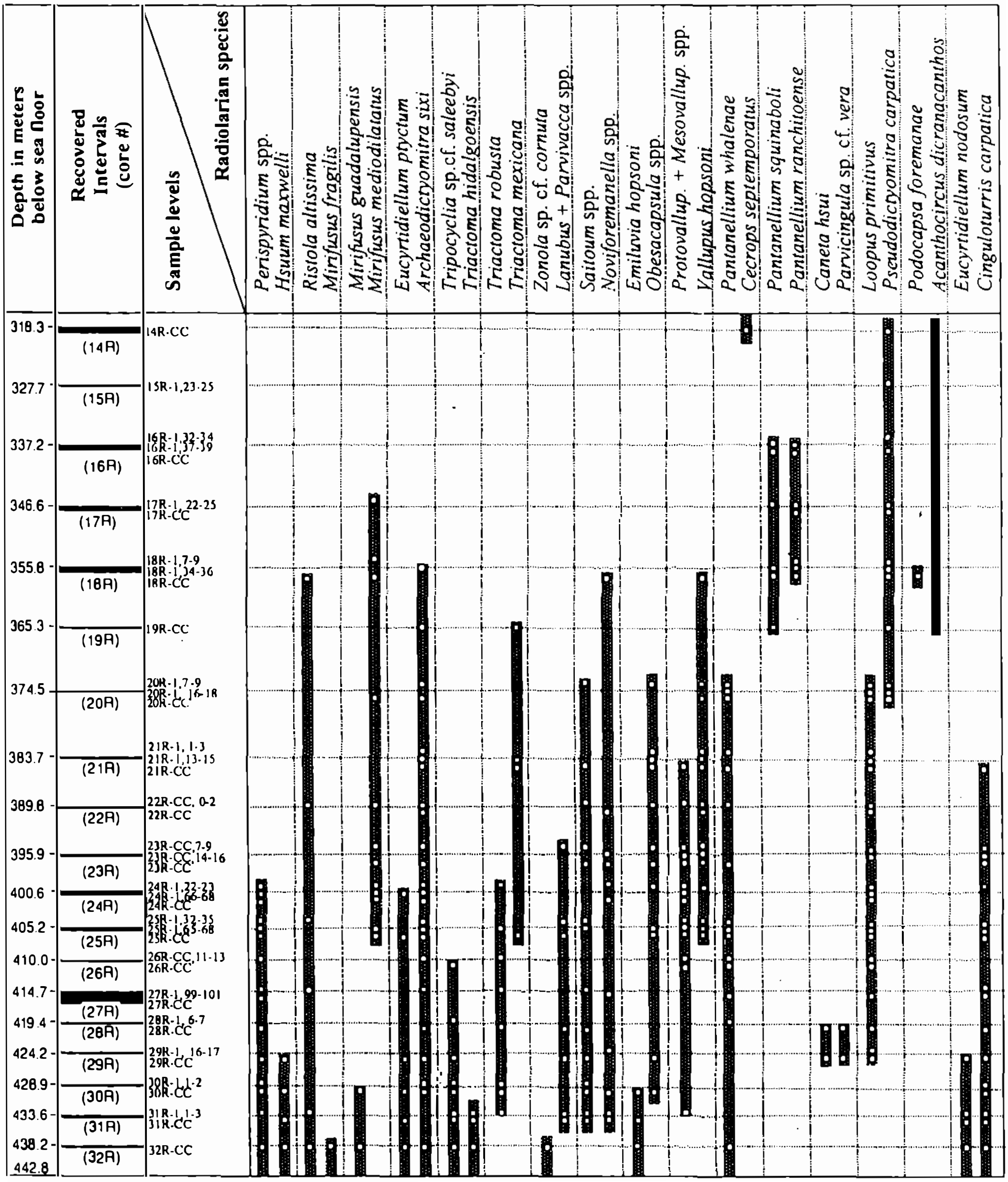

Figure 4. Occurrence data of selected radiolarians in the ODP Site 801B section.

The boundary between Zone 4 and Zone 5 is defined by two bio-events of primary marker taxa: (1) the LOB of *Ristola altissima* (Rüst) and (2) the LOB of *Ristola procera* (Pessagno). The latter species has not been found in the studied section. Based on the LOB of *R. altissima*, the Zone 4/Zone 5 boundary is drawn between Samples 18R-CC and 18R-1, 34-36cm. The top of Subzone 4β is placed above Sample 24R-1, 22-23cm, i.e., immediately above the LOB of *Perispyridium* spp.

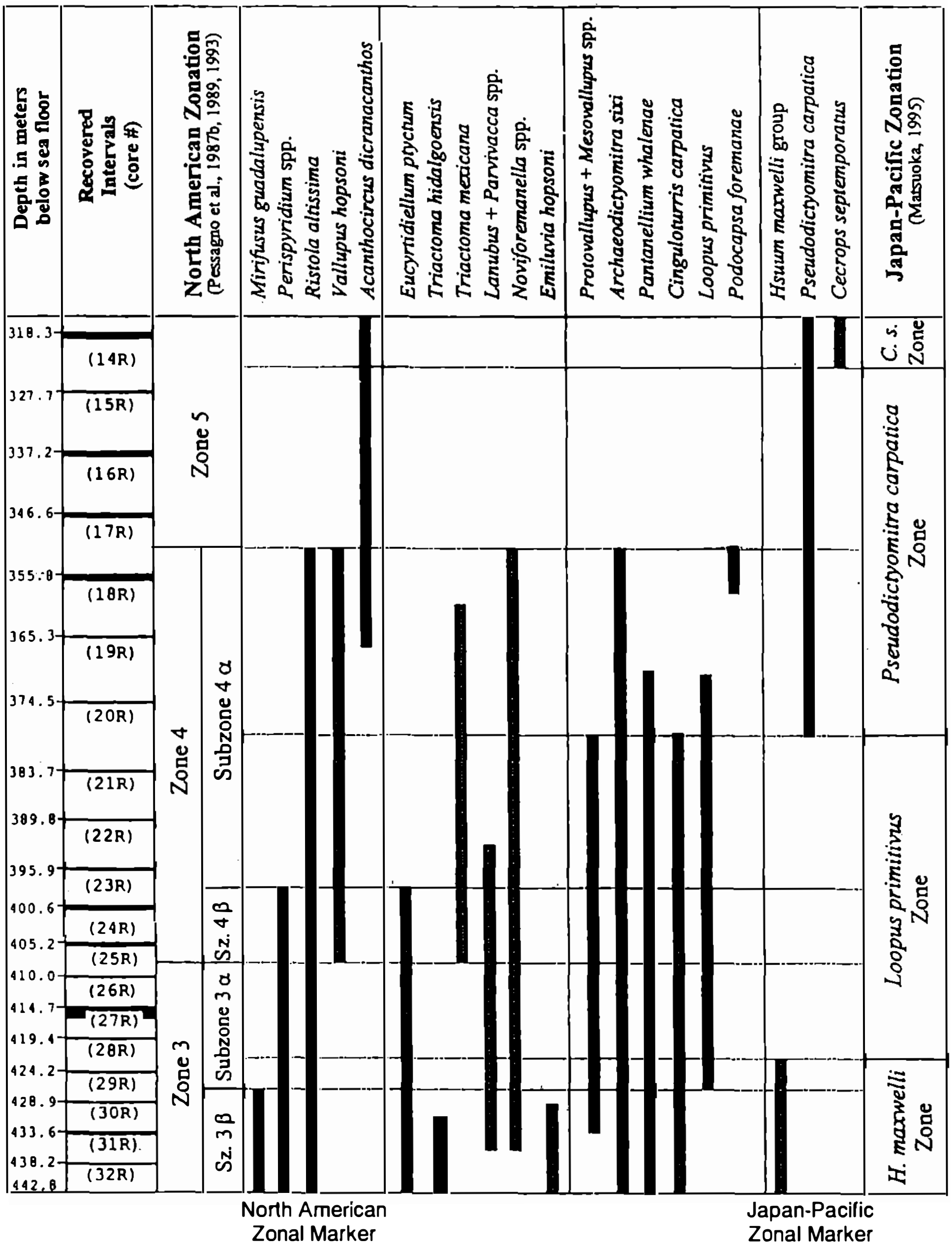

Figure 5. Stratigraphic distribution of zone-diagnostic and other biostratigraphically important radiolarian taxa in the ODP Site 801B section, and biostratigraphic divisions applying to both the North American and Japan-Pacific radiolarian zonal schemes.

Among these taxa, only *N. burckhardti* has not been found in the studied section. The FOBs of *V. hopsoni* and *A. dicranacanthos* are recognized at different horizons. Based on the FOB of *V. hopsoni*, the Zone 3/Zone 4 boundary is drawn between Samples 26R-CC, 11-13cm and 25R-CC. Zone 3 is further divided into lower Subzone 3β and upper Subzone 3α by the LOB of *Mirifusus guadalupensis* Pessagno. The subzone boundary is drawn between Samples 30R-CC and 30R-1, 1-2cm.

Zonal Correlation

Our research indicates that both the North American and Japan-Pacific zonations are equally applicable to the west Pacific section, with certain index markers providing direct ties between the two zonal schemes. The following seven zone-diagnostic bio-events are identified in ascending order: (1) LOB of *Mirifusus guadalupensis* Pessagno, (2) LOB of the *Hsuum maxwelli* Pessagno group, (3) FOB of *Vallupus hopsoni* Pessagno and Blome, (4) LOB of the genus *Perispyridium* Dumitrica, (5) EFAB of *Pseudodictyomatra carpatica* (Lozyniak), (6) FOB of *Acanthocircus dicranacanthos* (Squinabol), and (7) LOB of *Ristola altissima* (Rüst). By using these bio-events the North American and Japan-Pacific radiolarian zones can be correlated (Fig. 6). The North American Zone 3/Zone 4 and Zone 4/Zone 5 boundaries are situated within the Japan-Pacific *Loopus primitivus* Zone (JR 8) and *Pseudodictyomitra carpatica* Zone (KR 1), respectively. Conversely, the Japan-Pacific *Hsuum maxwelli* (JR 7)/*Loopus primitivus* (JR 8) and *Loopus primitivus* (JR 8)/*Pseudodictyomitra carpatica* (KR 1) zonal boundaries are situated within North American Zone 3 (Subzone 3α) and Zone 4 (Subzone 4α), respectively.

Concluding Remarks

This research indicates that both the North American and Japan-Pacific zonations for the Upper Jurassic can be applied equally to the radiolarian-bearing sequence at ODP Site 801B in the west Pacific. This suggests that further international collaboration across the Pacific could potentially lead to the unification of different zonal schemes. We also notice that the stratigraphic ranges of some zone-diagnostic taxa are different among regions. For example, the FOBs of *V. hopsoni* and *A. dicranacanthos* are regarded as the same in North America, while *V. hopsoni* occurs earlier than *A. dicranacanthos* in the west Pacific section. This may be due partly to preservation or to differences in the paleobiogeographic distribution of these two species. To better understand zonal correlation, these factors must be considered.

<table>
<tr><th rowspan="2">CHRONO-STRATIGRAPHY</th><th colspan="3">RADIOLARIAN ZONATION</th></tr>
<tr><th colspan="2">NORTH AMERICA
Pessagno et al., 1993</th><th>JAPAN-PACIFIC
Matsuoka, 1995 a</th></tr>
<tr><td rowspan="2">LOWER CRETA.</td><td rowspan="2">Zone 5</td><td rowspan="2"></td><td rowspan="3"><i>Pseudodictyomitra carpatica</i> Zone</td></tr>
<tr></tr>
<tr><td rowspan="8">UPPER JURASSIC</td><td rowspan="4">Zone 4</td><td rowspan="2">Subzone 4 α</td></tr>
<tr><td rowspan="4"><i>Loopus primitivus</i> Zone</td></tr>
<tr><td rowspan="2">Subzone 4 β</td></tr>
<tr></tr>
<tr><td rowspan="4">Zone 3</td><td rowspan="2">Subzone 3 α</td></tr>
<tr><td rowspan="3"><i>Hsuum maxwelli</i> Zone</td></tr>
<tr><td rowspan="2">Subzone 3 β</td></tr>
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</table>

Figure 6. Correlation of North American and Japan-Pacific radiolarian biostratigraphy in the ODP Site 801B section.

Acknowledgments

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References

BAUMGARTNER, P. O. 1984. A Middle Jurassic-Early Cretaceous low latitude radiolarian zonation based on unitary associations and age of Tethyan radiolarites. Eclogae Geologicae Helvetiae, 77:729-837.

____. 1987. Age and genesis of Tethyan Jurassic Radiolarites. Eclogae Geologicae Helvetiae, 80:831-879.

____, A. BOLTOLINI, E. S. CARTER, M. CONTI, G. CORTESE, T. DANELIAN, P. DE WEVER, P. DUMITRICA, R. DUMITRICA-JUD, S. GORICAN, J. GUEX, D. M. HULL, N. KITO, M. MARCUCCI, A. MATSUOKA, B. MURCHEY, L. O'DOGHERTY, J. SAVARY, V. VISHNEVSKAYA, D. WIDZ, AND A. YAO. 1995. Middle Jurassic to Early Cretaceous radiolarian biochronology of Tethys based on Unitary Associations, p. 1013-1048. *In* P. O. Baumgartner, L. O'Dogherty, S. Gorican, E. Urquhart, A. Pillevuit, and P. De Wever (eds.), Middle Jurassic to Lower Cretaceous Radiolaria of Tethys: Occurrences, Systematics, Biochronology. Mémoires de Géologie (Lausanne), no. 23.

ERBA, E., AND J. M. COVINGTON. 1992. Calcareous nannofossil biostratigraphy of Mesozoic sediments recovered from the western Pacific, Leg .129, P. 431-470. *In* R. L. Larson, Y. Lancelot, A. Fisher, L. Abrams, R. Behl, W. H. Busch, G. Cameron, P. R. Castillo, J. M. Covington, G. Dürr, E. Erba, P. A. Floyd, C. France-Lanord, E. H. Hauser, S. M. Karl, A.-M. Karpoff, A. Matsuoka, A. Molinie, J. G. Ogg, A. R. M. Salimullah, M. Steiner, B. P. Wallick, and W. Wightman (eds.), Proceedings of the Ocean Drilling Program, Scientific Results, vol. 129. College Station, TX (Ocean Drilling Program).

LANCELOT, Y. P., R. LARSON, A. FISHER, L. ABRAMS, R. BEHL, W. H. BUSCH, G. CAMERON, P. R. CASTILLO, J. M. COVINGTON, G. DÜRR, E. ERBA, P. A. FLOYD, C. FRANCE-LANORD, E. H. HAUSER, S. M. KARL, A.-M. KARPOFF, A. MATSUOKA, A. MOLINIE, J. G. OGG, A. R. M. SALIMULLAH, M. STEINER, B. P. WALLICK, AND W. WIGHTMAN. 1990. Proceedings of the Ocean Drilling Program Initial Reports, vol. 129. College Station, Texas, 488 p.

MATSUOKA, A. 1992. Jurassic and Early Cretaceous radiolarians from Leg 129, Sites 800 and 801, western Pacific Ocean, p. 203-220. *In* R. L. Larson, Y. Lancelot, A. Fisher, L. Abrams, R. Behl, W. H. Busch, G. Cameron, P. R. Castillo, J. M. Covington, G. Dürr, E. Erba, P. A. Floyd, C. France-Lanord, E. H. Hauser, S. M. Karl, A.-M. Karpoff, A. Matsuoka, A. Molinie, J. G. Ogg, A. R. M. Salimullah, M. Steiner, B. P. Wallick, and W. Wightman (eds.), Proceedings of the Ocean Drilling Program, Scientific Results, vol. 129. College Station, TX (Ocean Drilling Program).

____. 1995a. Jurassic and Lower Cretaceous radiolarian zonation in Japan and in the western Pacific. The Island Arc, 4:140-153.

____. 1995b. Late Jurassic tropical Radiolaria: *Vallupus* and its related forms. Palaeogeography Palaeoclimatology Palaeoecology, 119:359-369.

____. 1995c. Middle Jurassic to Early Cretaceous Radiolarian Occurrences in Japan and the Western Pacific (ODP Sites 800-801), p. 937-966. *In* P. O. Baumgartner, L. O'Dogherty, S. Gorican, E. Urquhart, A. Pillevuit, and P. De Wever (Eds.), Middle Jurassic to Lower Cretaceous Radiolaria of Tethys: Occurrences, Systematics, Biochronology. Mémoires de Géologie (Lausanne), no. 23.

____, AND YAO, A. 1986. A newly proposed radiolarian zonation for the Jurassic of Japan. Marine Micropaleontology, 11:91-105.

PESSAGNO, E. A. JR. 1977. Upper Jurassic Radiolaria and radiolarian biostratigraphy of the California Coast Ranges. Micropaleontology, 23:56-113.

____, C. D. BLOME, E. S. CARTER, N. MACLEOD, P. A. WHALEN, AND K-Y. YEH. 1987. Studies of North American Jurassic Radiolaria. Part II, Preliminary Radiolarian Zonation for the Jurassic of North America. Cushman Foundation for Foraminiferal Research Special Publication, 23 (Part II), 1-18.

___, AND D. M. HULL. 1996. "Once upon a time in the Pacific": chronostratigraphic misinterpretation of basal strata at ODP site 801 (central Pacific) and its impact on geochronology and plate tectonics models. GeoResearch Forum, 1-2:79-92.

___, ___, D. M. HULL, AND W. M. SIX. 1993. Jurassic Radiolaria from the Josephine ophiolite and overlying strata, Smith River subterrane (Klamath Mountains), northwestern California and southwestern Oregon. Micropaleontology, 39:93-166.

___, ___, AND J. F. LONGORIA. 1984. A revised radiolarian zonation for the Upper Jurassic of western North America. Bulletins of American Paleontology, 87, 51 p.

___, AND S. MIZUTANI. 1992. Radiolarian biozones of North America and Japan, p. 293-295. *In* G. E. G. Westermann (ed.), The Jurassic of the circum-Pacific, Cambridge University Press.

___, W. M. SIX, AND Q. YANG. 1989. Xiphostylidae Haeckel and Parvivaccidae, n. fam., (Radiolaria) from the Jurassic of North America. Micropaleontology, 35:193-255.

STEINER, M. B., AND B. P. WALLICK. 1992. Jurassic to Paleocene paleolatitudes of the Pacific plate derived from the paleomagnetism of the sedimentary sequences at Sites 800, 801, and 802, p. 431-470. *In* R. L. Larson, Y. Lancelot, A. Fisher, L. Abrams, R. Behl, W. H. Busch, G. Cameron, P. R. Castillo, J. M. Covington, G. Dürr, E. Erba, P. A. Floyd, C. France-Lanord, E. H. Hauser, S. M. Karl, A.-M. Karpoff, A. Matsuoka, A. Molinie, J. G. Ogg, A. R. M. Salimullah, M. Steiner, B. P. Wallick, and W. Wightman (eds.), Proceedings of the Ocean Drilling Program, Scientific Results, vol. 129. College Station, TX (Ocean Drilling Program).

YANG, Q., AND A. MATSUOKA. 1997. A comparative study on Upper Jurassic radiolarian biostratigraphy of the Taman Formation, east-central Mexico and the ODP Site 801 B Section, west Pacific. Science Reports of Niigata University Series E, Geology, No. 12, 29-49.

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Biostratigraphical Correlations between the Subboreal Mutabilis Zone and the Submediterranean Upper Hypselocyclum - Divisum Zones of the Kimmeridgian: New Data from Northern Poland

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Keywords: Biostratigraphy, Ammonites, Kimmeridgian, Submediterranean Province, Subboreal Province, Zonal Schemes, Correlations

Abstract: The core sections from the Kuyavy area in northern Poland yielded ammonites indicative of the Hypselocyclum Zone of the Submediterranean subdivision, and the Mutabilis Zone of the Subboreal subdivision. The succession recognised here can be correlated with those of southern England, western France (Aquitaine), and central Poland (Holy Cross Mountains). The occurrence of numerous *Rasenioides* and its allies, together with *Orthosphinctes* (*Ardescia*) *perayensis* Atrops shows that the lowermost part of the Mutabilis Zone *sensu* Birkelund et al. (1983) can be correlated with the *perayensis* horizon representing the topmost part of the Hypselocyclum Zone *sensu* Atrops (1982). During the early Mutabilis Chron, the territory of northern Poland became part of the Subboreal Province, whereas the Submediterranean Province had shifted southwards, occupying central Poland.

Introduction

The correlation of the Kimmeridgian of the Subboreal Province with the Submediterranean Province has presented long standing problems arising from the marked differences in coeval ammonite faunas. Aside from the currently widely discussed position of the lower boundary of the Kimmeridgian Stage, there still exists the problem of the boundary of the Lower and Upper Kimmeridgian Substages (as used in France), which is based mostly on the definition of the Subboreal Mutabilis Zone. It should be remembered that of the Subboreal Kimmeridgian "the part least well known is the Mutabilis Zone" (Birkelund et al., 1983, p.289), and that the lower boundary of this zone is interpreted in different senses in England and in France (mostly Aquitaine, emphasising the difficulties in reaching an accord on definition of the Lower/Upper Kimmeridgian boundary. The purpose of this paper is to discuss the correlation of the Mutabilis Zone (different interpretations of this zone have been given by Birkelund et al., 1983; Hantzpergue, 1989; Hantzpergue et al., 1997) with the typical Submediterranean ammonite zonal scheme of Europe (Atrops, 1982).

New data have been gathered by us during a study of ammonite faunas in cores from the Kuyavy area in northern Poland (Fig. 1), specifically the Kcynia IG-IV Borehole and shallow boreholes drilled in the Zalesie Anticline in the vicinity of Barcin. Ammonites of Subboreal affinity occur in the studied cores together with the Submediterranean forms, enabling correlation of Subboreal and Submediterranean zonal schemes in this stratigraphic interval.

Ammonite Succession in Northern Poland

The general ammonite succession of the Oxfordian and a large part of the Kimmeridgian is well developed in the extremely fossiliferous Kcynia IG-IV Borehole (Matyja and Wierzbowski, 1998). Crucial for correlations discussed here is the core interval from 219.2m to 179.0m of the lime - marly - lumachelle formation (V) of Dembowska (1979), lying directly below Cenozoic deposits, which yields abundant, stratigraphically significant ammonites (Fig. 2). Micritic limestones representing the lower part of the core (219.2 - 211.05 m) contain almost exclusively representatives of the Submediterranean subfamily Ataxioceratinae (Matyja and Wierzbowski, 1998, pl. 3, fig. 5) such as *Ataxioceras* (*Ataxioceras*) *hypselocyclum semistriatum* Schneid, and *A.*

(*A.*) cf. *hypselocyclum* (Fontannes). They are indicative of the *semistriatum* horizon of the Lothari Subzone of the Hypselocyclum Zone (Atrops, 1982) in the Submediterranean zonal scheme.

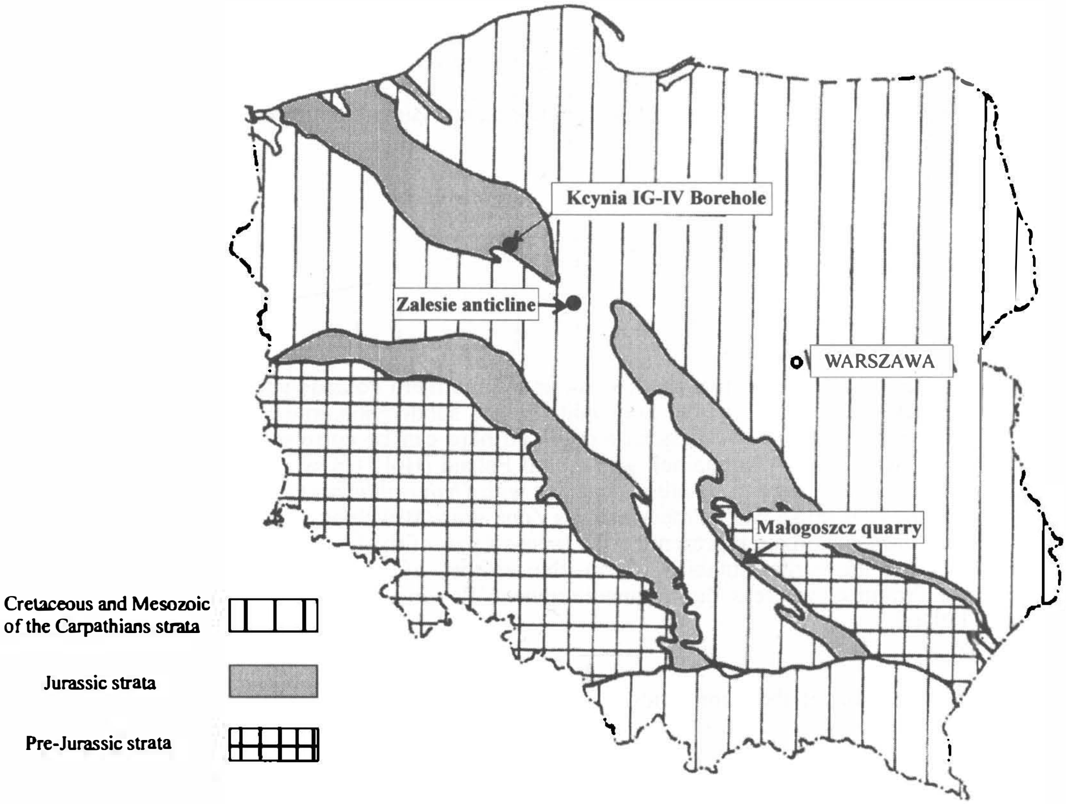

Figure 1. Geological map of Poland showing location of sections studied.

Still higher, the core consists of marly limestones, limestones and dolomites with intercalations of siltstones. In these deposits, well above a marked omission surface at 207.0 - 207.1m, the character of the ammonite fauna changes. Ammonites of the subfamily Ataxioceratinae are extremely scarce, represented by small *Orthosphinctes* (*Ardescia*) *perayensis* Atrops. A single specimen has been found at 179.0 - 180.9m, at the top of the interval studied (Matyja and Wierzbowski, 1998, pl. 4, Figs. 6a-c). This minute species is the last member of the Ataxioceratinae lineage, indicating already the *perayensis* horizon at the top of the Hypselocyclum Zone (Atrops, 1982). Other ammonites of Submediterranean/Mediterranean affinity, which occur in the discussed interval from 207 - 179m, include only a few forms of stratigraphic value, e.g., *Streblites* cf. *levipictus* (Fontannes) from 180.9 - 183.15m (Matyja and Wierzbowski, 1998, pl. 4, figs. 4-5), distinguishable among fairly common, but generally poorly preserved, specimens of the genus *Streblites,* and a single specimen of *Orthaspidoceras* sp. from 190.75 - 191.75m. Both these forms indicate an age not older than the latest Hypselocyclum - earliest Divisum Chrons.

Ammonites of the family Aulacostephanidae are undoubtedly the most common forms, occurring from immediately above the omission surface at about 207m and up to 179m, in the Kcynia IG-IV Borehole. Two groups can be distinguished (Matyja and Wierzbowski, 1998, pl. 3, figs. 6-10, and pl. 4, figs. 1-2): (1) more heavily ornamented forms (at least on inner whorls), including both macroconchs such as *Involuticeras involutum* (Quenstedt) and microconchs of the

Prorasenia type (*Prorasenia* cf. *quenstedti* Schindewolf, *Prorasenia* sp.), and (2) finely ribbed microconchs of the genus *Rasenioides*, such as *Rasenioides thermarum* (Oppel), and *R. lepidulus* (Oppel), replaced in the highest levels of the core (from 191.75 to 180.9m) by a form referred to as ?*Aulacostephanoides* aff. *eulepidus* (Oppel). This latter form shows an incipient ventral smooth band, and thus can be interpreted as transitional between the genera *Rasenioides* and *Aulacostephanoides* (Birkelund et al., 1983).

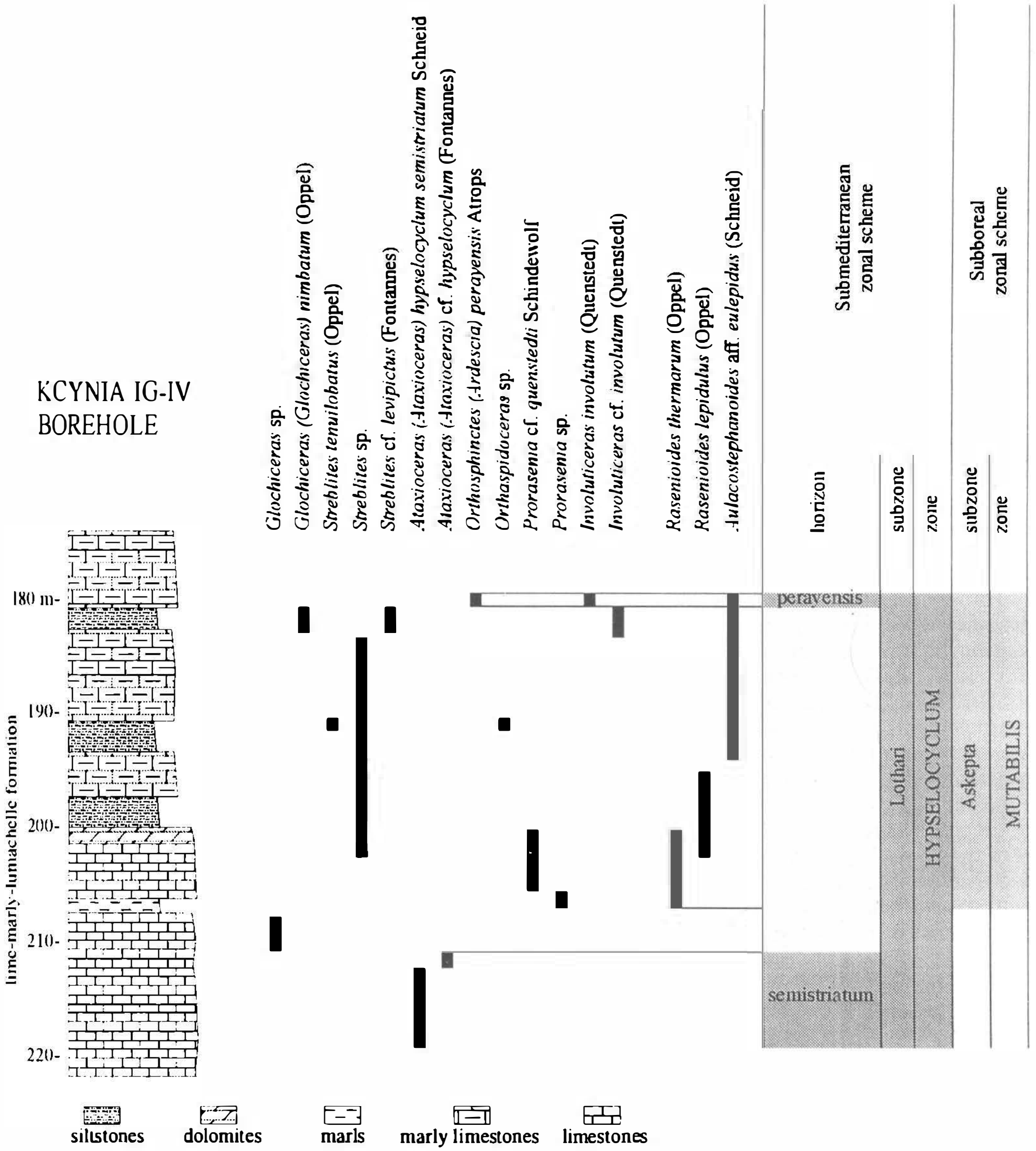

Figure 2. Detailed distribution of ammonites in the upper Hypselocyclum Zone - lower Mutabilis Zone interval in the Kcynia IG-IV Borehole.

The discussed ammonites of the family Aulacostephanidae from the Kcynia IG-IV Borehole enable recognition of biostratigraphical units established in the typical Subboreal Succession in England (Birkelund et al., 1983), as well as in Aquitaine, where Submediterranean and Subboreal ammonites dominate in particular horizons of the Kimmeridgian succession (Hantzpergue, 1989; 1995; Hantzpergue et al., 1997). The stratigraphic range of the genus *Rasenioides*, along with that of ?*Aulacostephanoides* aff. *eulepidus* (Oppel) from our core, corresponds generally to the lower part of the Mutabilis Zone in England, distinguished there as the Askepta Subzone (Birkelund et al., 1983). On the other hand, the same interval in the core when compared with the succession of Aquitaine, corresponds to those levels where ammonites of the genus *Rasenioides* occur commonly, i.e., the upper part of the Chatelaillonensis Subzone (from *askeptus* to *discoidus* horizons) of the Cymodoce Zone (Hantzpergue, 1989; Hantzpergue et al., 1997). However, the studied interval in the Kcynia IG-IV Borehole differs from that of the Aquitaine succession in the common occurrence of aulacostephanids other than *Rasenioides*, such as *Involuticeras* and *Prorasenia*, which have not been encountered in corresponding beds in Aquitaine. In Aquitaine the ammonite fauna consisting of *Eurasenia* and *Rasenia* (as well as the corresponding microconchs of the *Prorasenia* type) precedes the appearance of the genus *Rasenioides*, being known from the *aulnisa* and *manicata* horizons representing the lower part of the Chatelaillonesis Subzone of the Cymodoce Zone (Hantzpergue, 1989; Hantzpergue et al., 1997). Occurrence of *Parataxioceras* gr. *lothari* (Oppel) in the *aulnisa* horizon indicates that it must correspond to an older part of the Lothari Subzone of the Hypselocyclum Zone of the Submediterranean zonal scheme, well below the *perayensis* horizon (Hantzpergue, 1989; Atrops, 1982). It shows that the studied interval in the Kcynia IG-IV Borehole with the *Rasenioides-Involuticeras-Prorasenia* fauna is higher than the lower part of the Chatelaillonensis Subzone of the Aquitaine.

Nine shallow boreholes drilled in the Zalesie Anticline have provided continuous cores, rich in ammonites, at the Lower/Upper Kimmeridgian boundary. The lithological succession consists of limestones with marly intercalations, and marly limestones of the lime - marly - lumachelle formation overlain by siltstones and silty limestones of the Paluki formation (Dembowska, 1979). These two lithostratigraphic units in the studied boreholes in the Zalesie Anticline were distinguished previously (Matyja et al., 1985) as the units "C-3" and "D-1". The lithological character of the deposits, supported by CaO content analysis at one meter intervals, enabled detailed correlation of the cores and construction of a composite section for the area (Fig. 3).

The oldest deposits of the lime - marly - lumachelle formation, developed as micritic limestones and marly limestones, yielded several poorly preserved ammonites of the genus *Ataxioceras*. It is an assemblage typical of the Hypselocyclum Zone (Atrops, 1982). The cores have also yielded small specimens showing typical simplified ornamentation of the *Orthosphinctes* type, identified here (Fig. 4a) as *Orthosphinctes* (*Ardescia*) *perayensis* Atrops. They occur from the uppermost part of the lime - marly - lumachelle formation, where they co-occur with the last representatives of the genus *Ataxioceras*, to the lowermost part of the siltstones of the Paluki formation (Fig. 3). Besides *O.* (*A.*) *perayensis*, a single specimen of *Garnierisphinctes* cf. *semigarnieri* (Geyer) has been found in the lowermost part of the Paluki formation. The recognised stratigraphic range of *O.* (*A.*) *perayensis* corresponds to the *perayensis* horizon of the uppermost part of the Hypselocyclum Zone in the Submediterranean zonal scheme (Atrops, 1982). It should be noted that the occurrence of the last ammonites of the genus *Ataxioceras* in the lower part of the *perayensis* horizon, as seen in these cores, does not seem to be in contradiction with the original characteristics of this horizon (Atrops, 1982; 1994, p. 56).

The deposits lying below those of the *perayensis* horizon in the composite section of the Zalesie Anticline, which yield exclusively ammonites of the genus *Ataxioceras*, should correspond to the *semistriatum* horizon of the Hypselocyclum Zone (Fig. 3).

Besides the discussed ammonites of the family Ataxioceratidae, those studied from the Zalesie Anticline have yielded fairly numerous representatives of the family Aulacostephanidae. Some of them are poorly preserved specimens of the genera *Prorasenia*, as well as *Eurasenia* and/or *Involuticeras*, occurring in the interval of the Hypselocyclum Zone. Others recognised in the lowermost part of the Paluki formation include *Rasenioides* cf. *lepidulus* (Oppel) and a form transitional between *Rasenioides* and *Aulacostephanoides* and characterised by the presence of an incipient ventral smooth band, here referred to as ?*Aulacostephanoides* sp. These ammonites are indicative of the Askepta Subzone of the Mutabilis Zone in the Subboreal subdivision (Birkelund et al., 1983). The occurrence in still younger deposits of the Paluki formation of fairly common

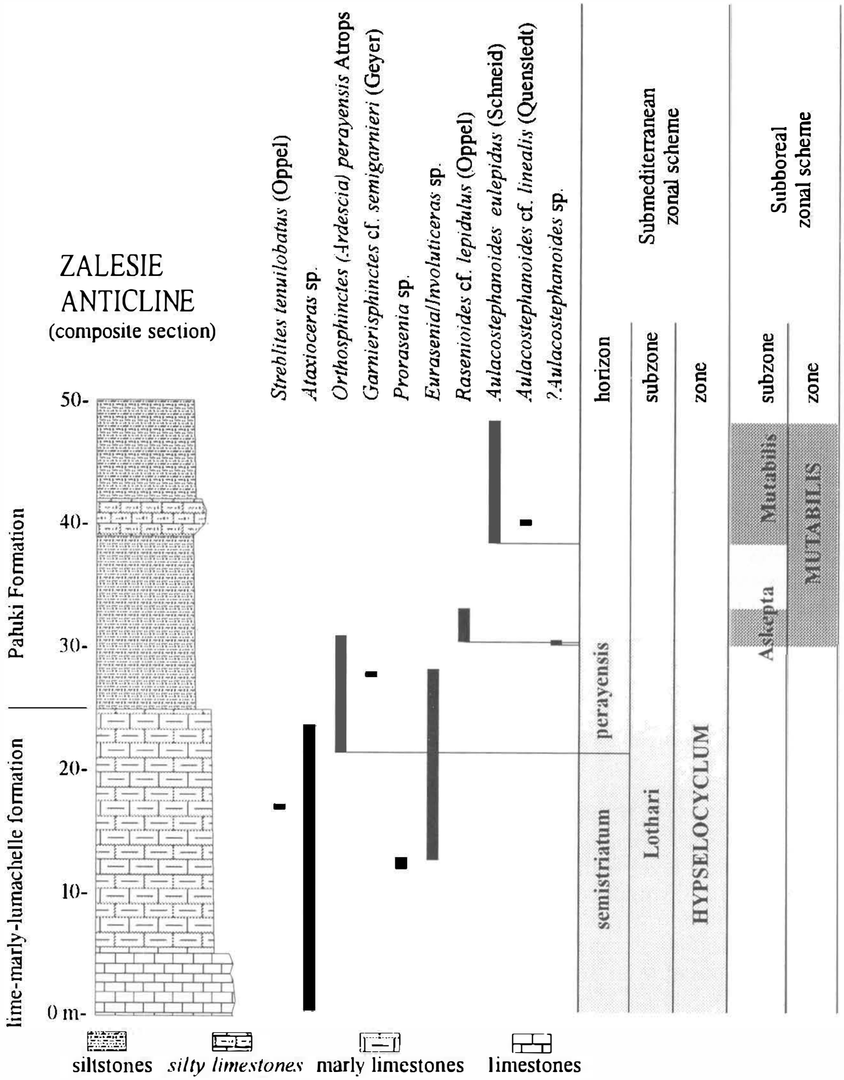

Figure 3. Detailed distribution of ammonites in the upper Hypselocyclum Zone - Mutabilis Zone interval in the composite section of the Zalesie Anticline.

specimens (Fig. 4b) of *Aulacostephanoides eulepidus* (Schneid), associated with *Aulacostephanoides* cf. *linealis* (Quenstedt), and the absence of *Rasenioides*, marks already the presence of the Mutabilis Subzone of the Mutabilis Zone of the Subboreal subdivision (Birkelund et al., 1983).

When compared with the ammonite subdivision of Aquitaine (Hantzpergue, 1989; Hantzpergue et al., 1997) in the Zalesie Anticline succession, the stratigraphic interval characterised by the occurrence of *Rasenioides*, and *Rasenioides* - *Aulacostephanoides* transitional forms should be correlated with the upper part of the Chatelaillonensis Subzone of the Cymodoce Zone, and that denoted by the occurrence of *Aulacostephanoides*, with the Mutabilis Subzone of the Mutabilis Zone.

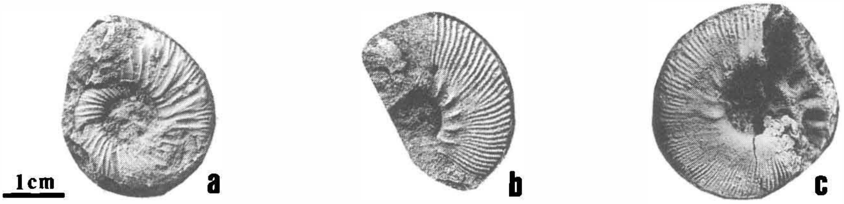

Figure 4. Selected Kimmeridgian ammonites from northern and Central Poland.
a. *Orthosphinctes* (*Ardescia*) *perayensis* Atrops borehole, Zalesie Anticline.
b. *Aulacostephanoides eulepidus* (Schneid), borehole, Zalesie Anticline.
c. *Rasenioides* (*Semirasenia*) *moeschi* (Oppel), Malogoszcz Quarry, Holy Cross Mountains

Correlation Between Subboreal and Submediterranean Ammonite Successions

Kimmeridgian ammonite faunas in the sections studied in northern Poland are mostly of Submediterranean character until the *semistriatum* horizon of the upper Hypselocyclum Zone, enabling differentiation of the Submediterranean zones, subzones, and horizons (Matyja and Wierzbowski, 1998; Atrops, 1982; see also Figs. 2-3 here). The ammonites of the family Aulacostephanidae occurring here, such as *Eurasenia*, *Involuticeras*, and *Prorasenia*, represent the lineage which developed in the Submediterranean Province, German - Polish Subprovince (Matyja and Wierzbowski, 1995), to some extent independent of such forms as *Pictonia* and *Rasenia*, which evolved in the Subboreal Province and have been used to construct the Subboreal subdivision of the Kimmeridgian. The former are thus of little help in recognition of the Subboreal zones and subzones. The only place in Europe where Subboreal ammonites of the genus *Rasenia* occur together with Submediterranean ammonites of the genus *Eurasenia* is western France, especially Aquitaine ("Biome franco - germanique"). This area occupied an intermediate biogeographical position between the Subboreal and Submediterranean Provinces during the Kimmeridgian (Hantzpergue, 1989; 1995). Of special interest for correlation is the *manicata* horizon of the Aquitaine succession with *Eurasenia manicata* (Schneid) and *Rasenia* gr. *evoluta* Spath. The latter is indicative of the highest level with abundant *Rasenia* in the Subboreal succession of England, i.e., the topmost part of the Cymodoce Zone (Birkelund et al., 1983; Hantzpergue, 1989). As the Submediterranean ammonite *Ataxioceras* gr. *lothari* (Oppel) has been found directly below the *manicata* horizon in Aquitaine, detailed correlation between the Submediterranean and Subboreal successions becomes possible at this level (Hantzpergue, 1989; Hantzpergue et al., 1997): the uppermost part of the Subboreal Cymodoce Zone of England thus correlates with part of the upper Submediterranean Hypselocyclum Zone, but not higher than the *semistriatum* horizon of the Lothari Subzone (see also Figs. 2-3, and Table 1).

The appearance of the genus *Rasenioides*, replacing in the Subboreal Province the genus *Rasenia*, marks the base of the Askepta Subzone, corresponding to the lower boundary of the Mutabilis Zone in England (Birkelund et al., 1983). This boundary correlates with the lower boundary of the *askeptus* horizon, distinguished as the lowest horizon with *Rasenioides* in the upper part of the Chatelaillonensis Subzone of the Cymodoce Zone in Aquitaine (Hantzpergue, 1989; Hantzpergue et al., 1997: see also Table 1 here). In the sections studied in northern Poland (Figs. 2, 3), ammonites of the genus *Rasenioides* (*R. lepidulus* and *R. thermarum*) occur commonly in a narrow stratigraphic interval characterised by ammonites of the *perayensis* horizon, well above the *semistriatum* horizon as distinguished here (uppermost Hypselocyclum Zone of the Submediterranean zonal scheme) (Atrops, 1982). It thus may be assumed that the lower boundary of the *perayensis* horizon of the Submediterranean zonal scheme corresponds to the base of the Mutabilis Zone, and more precisely to the base of the Askepta Subzone, in the Subboreal zonal scheme (Birkelund et al., 1983; see also Table 1 here). In the same stratigraphic interval in northern Poland, but mostly in its upper part (Figs. 2, 3), there appear also ammonites transitional in character between *Rasenioides* and *Aulacostephanoides* which are known to occur in the Askepta Subzone of the Subboreal Succession in England, except in the lowermost part of this subzone (Birkelund et al., 1983).

		Submediterranean Province (SE France, S. Germany, C. Poland)			"Biome Franco-Germanique" (Aquitaine)			Subboreal Province (England)	
		ZONE	Subzone	horizon	horizon	Subzone	ZONE	Subzone	ZONE
KIMMERIDGIAN	UPPER	ACANTHICUM (pars)			mutabilis	Mutabilis	MUTABILIS (pars)	Mutabilis	MUTABILIS (pars)
					attenuatus				
					desmonotus				
					linealis				
	LOWER	DIVISUM	Uhlandi		discoides	Chatelaillonensis	CYMODOCE (pars)	Askepta	
			Tenuicostata		chatelaillonensis				
		HYPSELOCYCLUM (pars)	Lothari	perayensis	askeptus				
				semistriatum	maniçata aulnisa				CYMODOCE (pars)
				hypselocyclum		Achilles Cymodoce			
				discoidale					

Table 1. Correlation of the Submediterranean, Subboreal and Aquitaine ("Biome Franco-Germanique") zonal schemes adjacent to the Lower/Upper Kimmeridgian boundary

The sudden appearance of *Rasenioides* in the sections of northern Poland reflects an invasion from the north which partly displaced faunas of Submediterranean affinity existing here before. The same phenomenon is observed in Aquitaine representing a part of the transitional "Biome franco - germanique" where the Submediterranean Ataxioceratidae and Aspidoceratidae are successively replaced by the Submediterranean aulacostephanid genus *Eurasenia*, and later on by the Subboreal aulacostephanid genera *Rasenia* and *Rasenioides*. The level with abundant *Rasenioides* is interpreted there as directly succeeding a maximum sea level rise (Hantzpergue, 1995).

This invasion by Subboreal *Rasenioides* did not extend farther south into central Poland, which remained a part of the Submediterranean Province until the Eudoxus Chron. However, the end of the Hypselocyclum Chron was also here a time of marked change in sedimentation, with a corresponding change in ammonite faunas. In the Kimmeridgian succession of the western border of the Holy Cross Mountains, shallow water carbonates topped by laminated micrites, corresponding to the Lower Kimmeridgian (Platynota-Hypselocyclum Zones), are overlain by various oyster lumachelles. The lower part of these lumachelles, distinguished as the Skórków Lumachelle, contains successive assemblages of ammonites representing the uppermost part of the Hypselocyclum Zone and of Divisum Zone (Kutek, 1968; 1994). The most abrupt change at the end of Hypselocyclum Zone corresponds to replacement of ammonites of the Ataxioceratinae lineage, such as *Ataxioceras* and *Ardescia*, by other groups of Ataxioceratidae such as *Garnierisphinctes* and *Crussoliceras*. The strongest contrast in ammonite faunas between surviving end members of the Ataxioceratinae lineage (*Ardescia*), and the appearance of abundant new representatives of Ataxioceratidae, occurs in the *perayensis* horizon of the Hypselocyclum Zone; the same phenomenon is observed at the same level everywhere in the Submediterranean Province (Atrops, 1982). This turnover of ammonite faunas is interpreted in terms of a marked sea level rise which brought in new ammonites of the family Ataxioceratidae of Mesogean affinity, and resulted in extinction of older Ataxioceratinae (Atrops and Ferry, 1989).

Comparison of the Subboreal ammonite succession at the boundary of the Cymodoce and Mutabilis Zones (*sensu* Birkelund et al., 1983) with that of the Submediterranean succession at the boundary of the Hypselocyclum and Divisum Zones thus indicates approximately synchronous changes in ammonite faunas belonging to quite different lineages in the different ammonite provinces. These changes possibly correlate with transgressive impulses which accelerated evolutionary development and resulted in marked faunal invasions (Hantzpergue, 1995).

The Submediterranean succession of the Divisum Zone yields only a few ammonites in common with the Subboreal succession. One of them is a macroconch of *Rasenioides* type found by us in the middle part of the Uhlandi Subzone of the Divisum Zone at the cement work quarry at Malogoszcz, on the western border of the Holy Cross Mountains (Figs. 1 and 4c). The ammonite is closely comparable with the holotype of *Rasenioides* (*Semirasenia*) *moeschi* (Oppel). Although the precise position of *R. moeschi* in the Submediterranean succession is unknown, it may be regarded as being similar to that of a very close species, *Rasenioides discoidus* Hantzpergue (Hantzpergue, 1989). The species *R. discoidus* occurs in the uppermost part of the Chatelaillonensis Subzone of Aquitaine, directly below the base of the Mutabilis Zone (Table 1).

Still younger strata in the sections studied in northern Poland containing specimens of *Aulacostephanoides* have not yielded any other identifiable ammonites which would allow a closer correlation between the Subboreal and Submediterranean zonal schemes. Although the general correlation of the base of Mutabilis Zone (*sensu* Hantzpergue, 1989; Hantzpergue et al., 1997) with the base of the Mutabilis Subzone (*sensu* Birkelund et al., 1983), as defined by the common occurrence of the *Aulacostephanoides eulepidus* and *A. linealis* groups, is widely accepted, the precise correlation of this level with the base of the Submediterranean Acanthicum Zone does not yet seem established in detail.

References

ATROPS, F. 1982. La sous-familie des Ataxioceratinae (Ammonitina) dans le Kimméridgien inférieur du sud-est de la France; systématique, évolution, chronostratigraphie des genres *Orthosphinctes* et *Ataxioceras*. Documents des Laboratoires de Géologie Lyon, 83:1-463.

____. 1994. The Upper Jurassic in the Dauphinois Basin, general introduction and field guide, p. 32-111. *In* 4th Oxfordian & Kimmeridgian Working Groups Meeting – Guide Book and Abstracts. Lyon and South-eastern France basin.

____, AND S. FERRY. 1989. Sequence stratigraphy and changes in the ammonite fauna (Upper Jurassic SE France), p. 7-9. *In* 2eme Congres Français de Sédimentologie, Mesozoic Eustasy Record on Western Tethyan Margin, Lyon.

BIRKELUND, T., J. H. CALLOMON, C. K. CLAUSEN, H. NOHR HANSEN, AND I. SALINAS. 1983. The Lower Kimmeridge Clay at Westbury, Wiltshire, England. Proceedings of the Geologists' Association, 94(4):289-309.

DEMBOWSKA, J. 1979. Systematyzowanie litostratygrafii jury górnej w Polsce pó³nocnej i œrodkowej. Kwartalnik Geologiczny, 23(3):617-630.

HANTZPERGUE, P. 1989. Les ammonites kimméridgiennes du haut-fond d'Europe occidentale: biochronologie, systématique, évolution, paléobiogéographie. Cahiers de Paléontologie Centre National de la Recherche Scientifique, 428 pp.

____. 1995. Faunal trends and sea-level changes: biogeographic patterns of Kimmeridgian ammonites of the Western European Shelf. Geologische Rundschau, 84:245-254.

____, F. ATROPS, AND R. ENAY. 1997. Kimméridgien, p. 87-96. *In* E. Cariou and P. Hantzpergue (coord.), Biostratigraphie du Jurassique ouest-européen et méditerranéen. Bulletin Centre de Recherches Elf Exploration-Production, Mémoires, 17.

KUTEK, J. 1968. The Kimmeridgian and uppermost Oxfordian in the SW margin of the Holy Cross Mts (Central Poland), I. Stratigraphy. Acta Geologica Polonica, 18(3):493-586.

____. 1994. Jurassic tectonic events in south-eastern cratonic Poland. Acta Geologica Polonica, 44(3-4):167-221.

MATYJA, B. A., T. MERTA, AND A. WIERZBOWSKI. 1985. Stratygrafia i litologia utworów jurajskich struktury Zalesia, p. 19-29. *In* Utwory jurajskie struktury Zalesia na Kujawach i ich znaczenie surowcowe. Warszawa - Wydawnictwa Geologiczne.

____, AND A. WIERZBOWSKI. 1995. Biogeographic differentiation of the Oxfordian and early Kimmeridgian ammonite faunas of Europe, and its stratigraphic consequences. Acta Geologica Polonica, 45(1-2):1-8.

____, AND ____. 1998. The stratigraphical and paleogeographical importance of the Oxfordian and Lower Kimmeridgian succession in the Kcynia IG-IV Borehole (in Polish, English summary). Biuletyn Pañstwowego Instytutu Geologicznego, 382: 35-70.

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Strontium Isotope Stratigraphy of the Aalenian/Bajocian Auxiliary Stratotype Point at Bearreraig, Isle of Skye, NW Scotland

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Keywords: Strontium, Isotope, Stratigraphy, Bearreraig, Skye, Aalenian, Bajocian

Abstract: The $^{87}Sr/^{86}Sr$ value of the Aalenian/Bajocian boundary is 0.707 288 ± 0.000 014 (2 s.e.), based on analysis of belemnites from the Auxiliary Stratotype Point (ASP) at Bearreraig, Isle of Skye. The previous value for this boundary, based on a statistical analysis of data for belemnites from the Inferior Oolite in Dorset, UK, is 0.707 287±0.000 020 (95% C.I.). During the Aalenian, the $^{87}Sr/^{86}Sr$ of marine strontium changed by 0.000 017; correlation and dating of Aalenian strata using strontium isotope stratigraphy is of little value, given the present practical limit of measurement of about ±0.000 004 on multiple repeat analysis and the present uncertainty of ±0.000 020 on the Aalenian Sr isotope calibration curve. Correlation within the Aalenian will be possible when the uncertainty on the Sr isotope calibration curve for the Aalenian approaches the uncertainty of ±0.000 003 currently attained for the best-defined parts of the Neogene isotope calibration curve.

Introduction

Strontium isotope stratigraphy (SIS) is developing rapidly as a valuable tool for correlating and dating marine sediments (for recent reviews see Veizer, 1989; McArthur, 1994). The correlative precision, and the temporal resolution, of SIS depends upon, amongst other things, the rate at which marine $^{87}Sr/^{86}Sr$ was changing when the strata of interest were deposited. Correlation of Maastrichtian rocks using Sr isotopes has been shown to be possible to zonal level, i.e. as good as that achievable from ammonites and belemnites (Crame et al., in press), with associated dates being resolved to ± 0.2 myr or one belemnite biozone. For much of the Jurassic, the rate at which the marine $^{87}Sr/^{86}Sr$ was changing was greater than it was in the Late Cretaceous and correlation with $^{87}Sr/^{86}Sr$ can better that available with ammonites (McArthur et al., in press).

Successful correlation using SIS requires that $^{87}Sr/^{86}Sr$ be defined for well known sequences and especially for stratotypes from which correlations ultimately derive. The use of Sr isotopes to assign numerical ages to rocks requires, in addition, that an accurate age calibration be assigned to such sequences. These combined approaches should lead eventually to a standard global Sr isotope curve showing how $^{87}Sr/^{86}Sr$ of marine Sr has varied through time. As a contribution to the database underpinning any such global standard curve, we present an $^{87}Sr/^{86}Sr$ profile through the Aalenian/Bajocian boundary as it is exposed in outcrops of the Bearreraig Sandstone Formation at Bearreraig, Isle of Skye, NW Scotland, which is the Auxiliary Stratotype Point for this boundary.

Aalenian - Bajocian Boundary

The Aalenian - Bajocian boundary (the base of the Bajocian Stage) has traditionally been defined at the base of the Discites Zone. More recently, under the auspices of the Bajocian Working Group of the ISJS, the ammonite biostratigraphy has been refined and the base of the Discites Zone and of the Bajocian Stage is now recognised and correlated by the first occurrence of an ammonite assemblage including *Hyperlioceras mundum* and related species (Pavia and Enay, 1997).

To establish the chronostratigraphical definition of the base of the Bajocian Stage, possible boundary stratotypes were investigated by the Bajocian Working Group. Two sections were proposed and accepted by the ISJS and the ISC. Murtinheira at Cabo Mondego in Portugal was selected as the Global boundary stratotype (GSSP) and Bearreraig, Isle of Skye, Scotland as the Auxiliary Stratotype Point. This paper gives new stratigraphic data on the Auxiliary Stratotype.

The Bearreraig Section

The Aalenian and Lower Bajocian succession at Bearreraig comprises over 200m of sediments in the Bearreraig Sandstone Formation (Morton, 1990; Morton and Hudson, 1995). Three major coarsening-up siltstone-sandstone cycles occur, within which there are smaller-scale cycles (Fig. 1). Marine fossils, especially ammonites, belemnites and bivalves occur throughout, although the great thickness of sediment means that the distributions of the fossils are discontinuous. All the standard ammonite zones, and most of the subzones, of the Aalenian and Lower Bajocian of the NW European Province can be identified, as can the more detailed ammonite horizons in the uppermost Aalenian to lowermost Bajocian part of the sequence.

The uppermost subzone of the Aalenian Stage (Limitatum Subzone, Concavum Zone) and the lowermost subzone of the Bajocian Stage (Walkeri Subzone, Discites Zone) occur in a thick succession (ca. 25m) of silty shales and siltstones of the Udairn Shale Member. The succession summarised in Figure 1 is compiled from three outcrop sections within 400m of each other. There is a small fault, and a stratigraphic gap calculated to be less than 1m of strata, between the middle section and the upper section (Fig. 1).

Samples

We analysed samples of belemnites occurring between the base of the Murchisonae Zone (Middle Aalenian) to the Ovalis Zone (Lower Bajocian). Many fossils in the Bearreraig section are very well preserved, especially those found in early-diagenetic carbonate nodules. Whilst aragonitic shells have inverted to calcite, and several phases of calcite cementation have been identified (Marshall, 1981), belemnites are well preserved. Except in the immediate vicinity of minor intrusions, such good preservation is not unexpected because fission-track data (Morton, 1990) suggest that the rocks have not been heated to temperatures above 100°C, whilst organic geochemical data (Thrasher, 1992) suggest that they have not been heated above 50°C.

The most detailed sampling was concentrated on strata immediately above and below the Aalenian - Bajocian boundary. The Bearreraig section has been accepted as the Auxiliary Stratotype Point for the base of the Bajocian Stage; therefore, the data presented here are located stratigraphically with respect to their height above or below this boundary. They enable a stratigraphically precise evaluation of the $^{87}Sr/^{86}Sr$ value at the Aalenian - Bajocian boundary which confirms the accuracy of the $^{87}Sr/^{86}Sr$ value of 0.707 287±0.000 020 (95% C.I.) derived for this boundary by Howarth and McArthur (1997) from the data of Jones et al. (1994) and Engkilde (pers. Comm., 1997) on samples from the Inferior Oolite of Dorset, UK.

Analytical Methods and Results

Sample Preparation

We analysed 14 belemnites. Prior to analysis, portions most susceptible to alteration (alveolus, apex, exteriors, apical line) were removed using diamond cutting tools and discarded. The remaining portions were cleaned by brief immersion in 1.2 molar hydrochloric acid, washed with ultra-pure water and then dried. Finally, the remaining pieces were individually fragmented into sub-mm-sized pieces using an agate pestle and mortar. The fragments were again cleaned by brief immersion in 1.2 molar hydrochloric acid, washed with ultra-pure water, and finally dried in a clean environment. Fragments for analysis were selected under the binocular microscope.

Chemical Data and Isotopic Analysis

For $^{87}Sr/^{86}Sr$ analysis, picked fragments were dissolved in 6 M nitric acid and Sr was separated by standard ion-exchange chemistry. Values of $^{87}Sr/^{86}Sr$ were measured with a VG354 five-collector mass spectrometer using a multidynamic routine on three collectors (SrSLL) that include corrections for isobaric interference from ^{87}Rb (Thirlwall, 1991). Data have been normalised to a value of 0.1194 for $^{87}Sr/^{86}Sr$. During data collection the measured value for NIST 987 was within 0.000 025 of the value 0.710248. Data reported in Table 1 have been adjusted to a value of 0.7102 48 for NIST 987 which, in our laboratory, corresponds to a value of 0.709175 for modern seawater Sr. Based upon replicated analysis of standards, the precision of our measurements (2 s.e.) was better than ±0.000 015 for single determinations and ±0.000 011 for duplicates. Total blanks were

<2 mg of Sr; sample contained >5 mg of Sr. For chemical analysis, fragments picked under the microscope were analysed by standard ICP-AES methods for their concentrations of Ca, Sr, Mg, Fe, and Mn, after dissolution in 10% acetic acid. Data are presented in Table 1.

Sample		Zone	Subzone	Age Ma	Level +/- Boundary	$^{87}Sr/^{86}Sr$	% Ca	ppm Element Mg	Na	Sr	Fe	Mn
B16	B	*Ovalis*	-	175.6	53.2	0.707292	38.9	3330	3990	1290	210	28
	a			175.6	53.2	0.707280						
B 14	j	*Discites*	*walkeri*	176.3	11.7	0.707286						
86/5	o	*Discites*	*walkeri*	176.3	10.0	0.707282	38.9	2140	2680	1015	770	19
	c	*Discites*	*walkeri*	176.3	10.0	0.707289						
80/20	i	*Discites*	*walkeri*	176.4	7.5	0.707274	39.7	2750	2970	970	1330	30
	a	*Discites*	*walkeri*	176.4	7.5	0.707282						
80/21	n	*Discites*	*walkeri*	176.4	7.0	0.707296	39.3	2530	2560	985	840	81
80/18		*Discites*	*walkeri*	176.5	0.1	0.707284						
90/5		*Concavum*	*limitatum*	176.5	-0.3	0.707283	39.4	2070	2850	1120	115	3
		Concavum	*limitatum*	176.5	-0.3	0.707287						
75/8	A	*Concavum*	*limitatum*	176.6	-5.3	0.707294	39.6	1980	2860	1010	57	7
	a	*Concavum*	*limitatum*	176.6	-5.3	0.707288						
75/5	l	*Concavum*	*limitatum*	176.7	-6.0	0.707295	39.8	2360	3930	1110	41	<1
	e	*Concavum*	*limitatum*	176.7	-6.0	0.707293						
75/3	n	*Concavum*	*limitatum*	176.7	-9.1	0.707298	40.1	3360	3790	1360	40	3
	i	*Concavum*	*limitatum*	176.7	-9.1	0.707298						
80/22	a	*Concavum*	*limitatum*	176.9	-13.5	0.707281						
No sample	n	*Concavum*	*concavum*									
No sample		*Bradfordensis*	-									
65M		*Murchisonae*	*murchisonae*	178.3	-27.7	0.707280						
No sample		*Murchisonae*	*obtusiformis*									
74/25		*Murchisonae*	*haugi*	179.1	-29.9	0.707297						
74/24		*Murchisonae*	*haugi*	179.1	-30.4	0.707290						

Table 1. Isotopic and elemental data for belemnites from Bearreraig, Skye.

Results and Discussion

Sample Preservation

All samples have concentrations of Ca, Mg, Na and Sr that are within the ranges reported for other belemnites (Jones et al., 1994; McArthur et al., in press) and the $^{87}Sr/^{86}Sr$ ratios for the samples are concordant and close to those found for this time interval by others (Jones et al., 1994; Engkilde, pers. comm., 1997). On these criteria, all samples would be regarded as well preserved.

Diagenetic alteration of skeletal carbonates usually increases the concentrations of Fe and Mn. Uncertainty exists as to the threshold concentration above which iron and manganese may be presumed to show that skeletal carbonate is altered (Veizer, 1983). In recent palaeo-climatic studies, Anderson et al. (1994) and Ditchfield (1997) assumed that the thresholds are 100 ppm Mn and 300 ppm Fe; Jones et al. (1994) set the limit for belemnites at 150 ppm Fe, whilst McArthur (1994), dealing with SIS criteria for brachiopods, bivalves and ammonoids, set the level at 100 ppm for both elements. Our Bajocian samples contain concentrations of Fe that suggest alteration has occurred (samples 80/20, 1330 ppm; sample 80/21, 840 ppm). Our samples are well preserved on other criteria (see above; Table 1) so we speculate that either the high Fe concentrations are not giving a true view of alteration in these two belemnites and may represent trace disseminated pyrite or iron oxide, or that alteration has been too slight to alter the $^{87}Sr/^{86}Sr$ values. We accept the ratios for these belemnites as being representative of Aalenian marine strontium.

Isotopic Ratios

In Fig. 2, Sr isotopic ratios of samples are plotted against their stratigraphic level within the Bearreraig section. The data show a trend to lower values at higher stratigraphic levels. A linear regression of $^{87}Sr/^{86}Sr$ on stratigraphic level has been fitted to the data and is also shown on Figure

2; this regression predicts a boundary value for $^{87}Sr/^{86}Sr$ of 0.707 288±0.000 014 (2 s.e.). The ratio at this boundary has previously been predicted by Howarth and McArthur (1997) to be 0.707 287±0.000 020 (2 s.e.), based on a LOWESS statistical fit to the $^{87}Sr/^{86}Sr$ data of Jones et al. (1994) and M. Engkilde (unpublished belemnite analysis; pers. comm, 1997) for belemnites from the Inferior Oolite of Dorset. The agreement between these values is remarkable; our linear regression predicts a boundary ratio that is analytically indistinguishable from the ratio predicted by Howarth and McArthur (1997), whilst reducing from ±0.000 020 to ±0.000014 the uncertainty associated

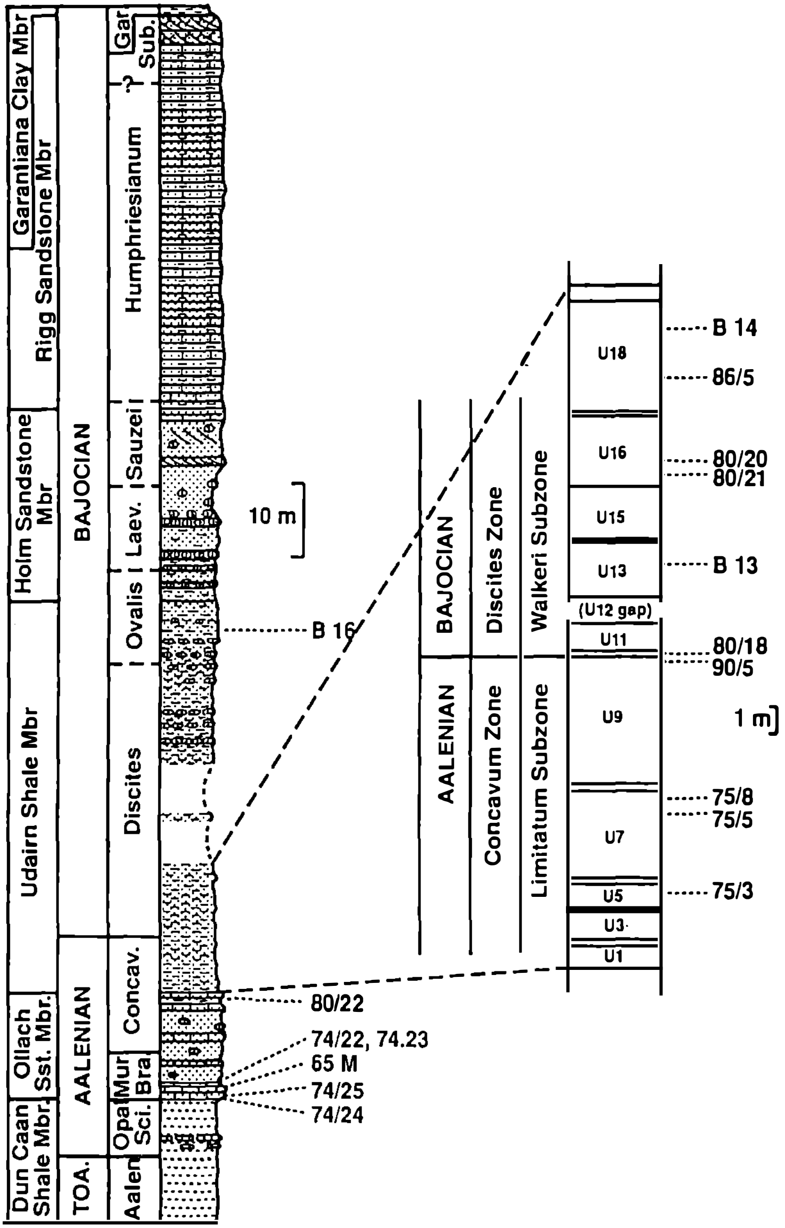

Figure 1. The Aalenian/Bajocian boundary succession at Bearreraig, Isle of Skye, UK.

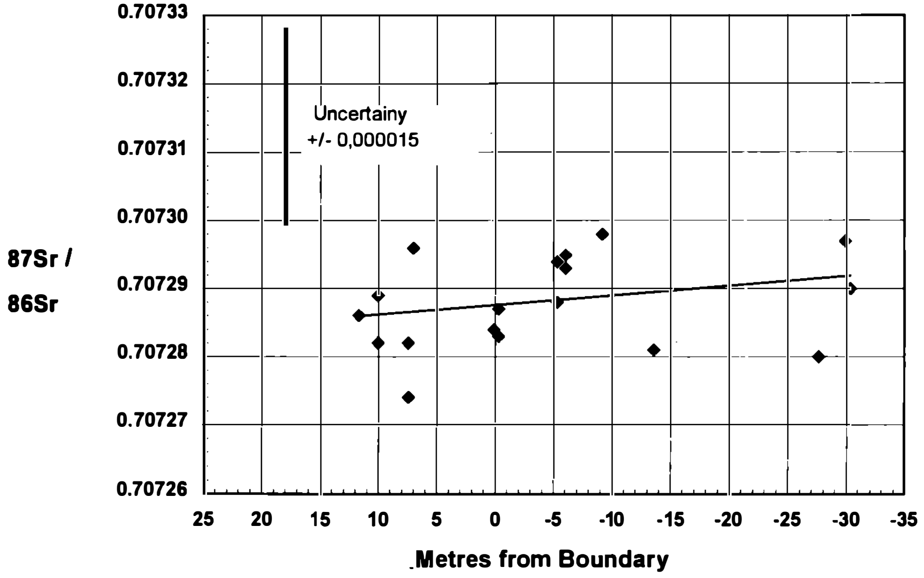

Figure 2. Isotopic ratio for Sr plotted against stratigraphic level, Bearreraig, Isle of Skye.

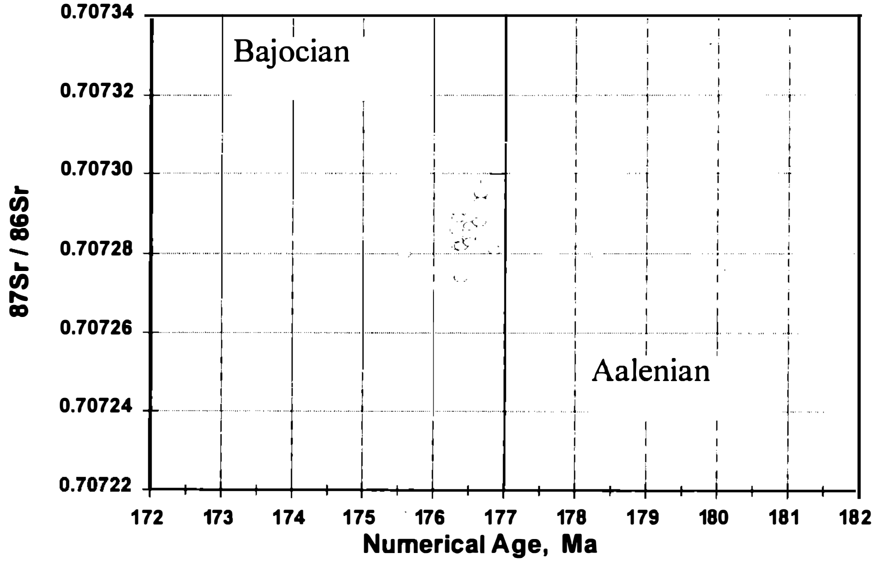

Figure 3. Comparison of our $^{87}Sr/^{86}Sr$ data (ovals) with the data of Jones et al. (1994; triangles).

with the boundary ratio. A comparison of our data with that of Jones et al. (1994) is given in Figure 3; in order to make the comparison possible, a numerical scale has been assigned to our samples on the assumption that subzones are of equal duration and the numerical ages of stage boundaries are those given by Gradstein et al. (1994); 180.1 Myr for the base of the Aalenian, 176.5 Myr for the base of the Bajocian, and 169.2 Myr for the base of the Bathonian stage.

Through the Aalenian, $^{87}Sr/^{86}Sr$ changed by 0.000 017 (Jones et al., 1994; M. Engkilde, pers. com., 1997; Howarth and McArthur, 1997), mostly as a result of a decline in $^{87}Sr/^{86}Sr$ through the upper two thirds of the Stage. Coupled with an analytical uncertainty for $^{87}Sr/^{86}Sr$ of as little as ±0.000 004, a figure achievable by replicate analysis (Crame et al., in press), it seems that strontium isotope stratigraphy might be useful in the Aalenian. Unfortunately, the present strontium isotope calibration curve for the Aalenian carries a statistical uncertainty of ±0.000 020 owing to the small number of data for the interval; until this uncertainty decreases, even replicate analysis will not enable Sr isotope stratigraphy to resolve stratigraphic problems involving Aalenian strata. Clearly, more data are required for this interval. Correlation within the Aalenian will be possible when the uncertainty on its Sr isotope calibration curve approaches the uncertainty of ±0.000 003 currently attained for the best-defined parts of the Neogene isotope calibration curve.

References

ANDERSON, T. F., B. N. POPP, A. C. WILLIAMS, L. Z. HO, AND J. D. HUDSON. 1994. The stable isotopic records of fossils from the Peterborough Member, Oxford Clay Formation (Jurassic), UK: palaeoenvironmental implications. Journal of the Geological Society, London, 151:125-138.

DITCHFIELD, P. W. 1997. High northern palaeolatitude Jurassic-Cretaceous palaeotemperature variation: new data from Kong Karls Land, Svalbard. Palaeogeography, Palaeoclimatology, Palaeoecology, 130:163-175.

CRAME J. A., J. M. MCARTHUR, D. PIRRIE AND J. B. RIDING. (In press). Strontium isotope correlation of the basal Maastrichtian stage in Antarctica to the European and US standard biostratigraphic schemes. Journal of the Geological Society, London.

GRADSTEIN, F. M., AGTERBERG, F. P., OGG, J. G., HARDENBOL, J., VAN VEEN, P., AND HUANG, Z. 1994. A Mesozoic Time Scale: Journal of Geophysical Research, 99:24051-24074.

HOWARTH, M. K., AND J. M. MCARTHUR. 1997. Statistics for strontium isotope stratigraphy: a robust LOWESS fit to the marine strontium isotope curve for the period 0 to 206 Ma, with look-up table for the derivation of numerical age. Journal of Geology, 105:441-456.

JONES, C. E., H. C. JENKYNS, A. L. COE, AND S. P. HESSELBO. 1994. Strontium isotope variations in Jurassic and Cretaceous seawater: Geochimica et Cosmochimica Acta, 58:3061-3074.

MARSHALL, J. D. 1981. Zoned calcites in Jurassic ammonite chambers: trace elements, isotopes and neomorphic origin. Sedimentology, 28:867-887.

McARTHUR, J. M. D. T. DONOVAN, M. F. THIRLWALL, B. FOUKE AND M. BROWNLESS. (in press). Strontium isotope stratigraphy in the early Toarcian – late Pliensbachian (Jurassic): timescale revision and its implications. Earth and Planetary Science Letters.

____. 1994. Recent trends in strontium isotope stratigraphy. Terra Nova, 6:331-358.

MORTON, N. 1990. Bearreraig (Isle of Skye, NW Scotland) as boundary stratotype for the base of the Bajocian Stage, p. 23-48. *In* Cresta and G. Pavia, (eds.) Atti del meetingsulla stratigrafia del Baiociano. Memorie Descrittive Carta Geologica d'Italia, 40.

____ AND J. D. HUDSON. 1995. Field guide to the Jurassic of the Isles of Raasay and Skye, Inner Hebrides, NW Scotland, p. 209-280. *In* P. D. Taylor, (ed.), Field Geology of the British Jurassic. The Geological Society, London.

PAVIA, G. AND R. ENAY. 1997. Definition of the Aalenian-Bajocian Stage boundary. Episodes 20:16-22.

THIRLWALL, M. F. 1991. Long-term reproducibility of multicollector Sr and Nd isotope ratio analysis. Chemical Geology (Isotope Geosciences Section), 94:85-104.

THRASHER, J. 1992. Thermal effect of the Tertiary Cuillins Intrusive Complex in the Jurassic of the Hebrides: an organic geochemical study, p. 35-49. *In* J. Parnell (ed.), 1992. Basins on the Atlantic Seaboard: Petroleum Geology, Sedimentology and Basin Evolution. Geological Society Special Publication 62.

VEIZER, J. 1983. Trace elements and isotopes in sedimentary carbonates, p. 265-299. *In* R.J. Reeder (ed.), Carbonates: mineralogy and chemistry. Reviews in Mineralogy, 11.

____. 1989. Strontium isotopes in seawater through time. Annual Reviews of Earth and Planetary Science, 17:141-167.

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The Nature of Biostratigraphic Boundaries in the Early Middle Jurassic of South West Germany

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Abstract: Several boundaries - of Stages, Zones, Subzones down to faunal horizons - from the South west German Aalenian - Bajocian have been studied using different methods (including geochemistry, macro-, meso- and microfaunal quantitative analysis) to find out the character and details of such boundaries. Two sections without obvious lithological discontinuities were selected. Most of the boundaries studied turned out to be distinctly reflected by changing microfaunal assemblages, some also by shifting geochemical trends. Microfaunal changes can be used to find the exact position of a boundary, sometimes better than with the sporadically distributed ammonites, particularly in boreholes. The observed changes suggest general biological events at the boundaries, possibly caused by sea level changes.

Introduction

Many biostratigraphic boundaries in the South west German Middle Jurassic are marked by iron oolitic beds, such as the so-called 'Murchisonae-Oolith', the 'Sowerbyi-Oolith', 'Humphriesi-Oolith', 'Subfurcaten-Oolith' and so on. Those oolitic boundary beds are usually very fossiliferous (with ammonites, belemnites, bivalves) and easily identifiable in outcrop as well as in bore-holes. In marginal areas, such as the eastern Swabian Alb, the oolites can comprise major parts of the Middle Jurassic. But in central parts of the basin, thick sequences of dark marls and clays are predominant and the oolitic layers can even disappear completely. Such basinal sequences are generally poor in ammonites which are often known from only a few layers. This study attempts to understand better the processes across the boundaries and to find additional criteria for recognising the position of the boundaries in such sequences. For this purpose some boreholes were recently drilled by the Geological Survey Baden-Württemberg to get complete, unweathered cores near natural outcrops of some general importance and with an already more or less established ammonite stratigraphy.

The Lower Aalenian in KB Wittnau

One of these sections is the Lower Aalenian at Wittnau (Oberrhein area, see Fig. 1) which includes also the upper part of the Upper Toarcian. It has been proposed as an Aalenian GSSP and the borehole KB Wittnau was drilled in 1994 to support this project.

Lithology

The Wittnau section consists throughout of uniform grey clays and marls. Its Toarcian part (Jurensismergel Formation) is characterized by accumulated shells of *Bositra buchi* (Roemer). In the Lower Aalenian part (Opalinuston Formation) the shell content diminishes, whereas thin (1-2cm) silt layers are common. No lithological boundary marker beds exist.

Geochemistry

The distribution of geochemical parameters in KB Wittnau was studied by Martin (1996) by x-ray-fluorescence-spectrometry. Most useful are element ratios recently plotted by Martin. For instance the SiO_2/Al_2O_3 ratio very well reflects the Stage boundary with a sudden rise at 57.5m, at

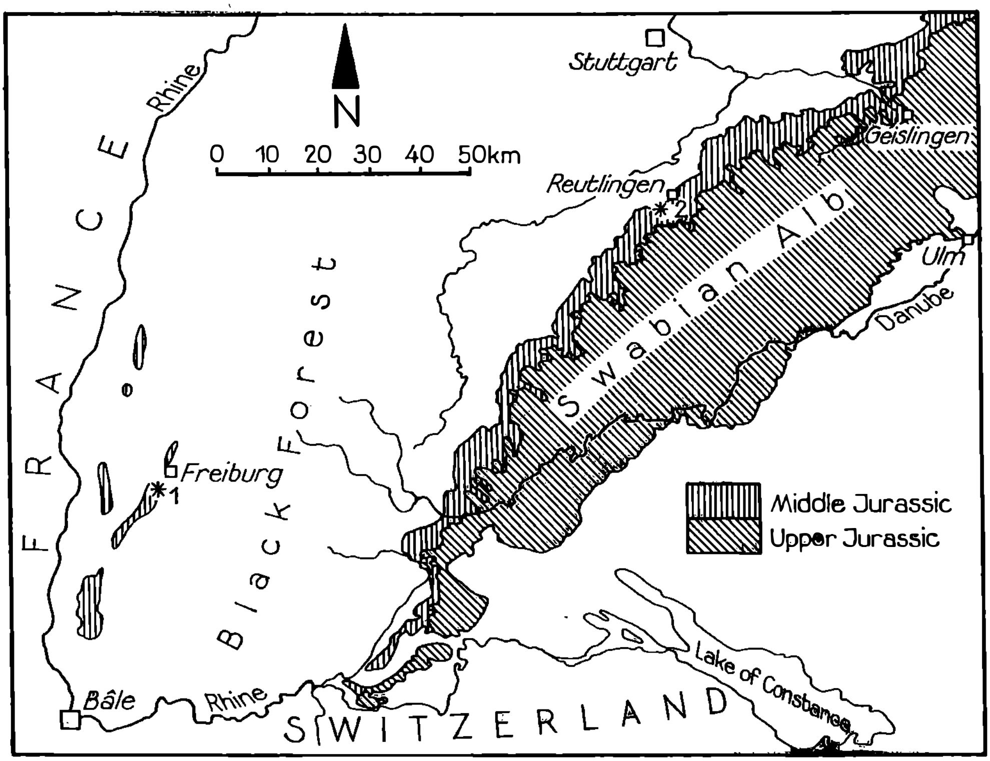


Figure 1. Location of the studied boreholes. 1: KB Wittnau near Freiburg (upper Toarcian-Lower Aalenian), 2: KB Pfullingen-Breitenbach (Lower-Upper Bajocian).

the base of the Aalenian as now agreed upon by a vote of the Aalenian Working Group (Fig. 2). The base of the *Leioceras comptum* Subzone is marked by a maximum value of this ratio. Moreover, the short decline at 37.5m is reflected by nanno-, micro-, and mesofossils. Similarly, the CaO/MgO ratio reflects the base of the Aalenian with a distinct minimum and the basal *L. comptum* Subzone by temporary peaks (Fig. 3), while the horizon at 37.5m is marked by a decline in this ratio.

Calcareous Nannofossils

Baldanza and Mattioli (1996) observed a considerable decrease in the species diversity of calcareous nannofossils across the Toarcian/Aalenian boundary in KB Wittnau. Other minima in diversity in the *Leioceras opalinum* Zone (at about 37m, 18m and 10m) obviously correspond with geochemical and biological events, as for instance declines of the radiolarians and certain arenaceous foraminifera.

Radiolarians

The semiquantitative distribution of radiolarians in KB Wittnau (Fig. 2) (after first appearing in the Upper Toarcian *Hammatoceras insigne* and *Dumortieria levesquei* Zones) exhibits a remarkable increase up to 9 % of the total microfauna beginning with the basal *L. opalinum* Zone, and a distinct decline in the basal *L. comptum* Subzone.

Ostracods

The abundance of ostracods appears to depend generally on the CaO/MgO ratio (Fig. 3), so they exhibit a distinct decline at the basal boundary of the Aalenian and a peak at the base of the *L. comptum* Subzone when the CaO/MgO ratio is rising again. But as with other microfossil groups, their percentages also decrease across the basal boundary of the *Pachylytoceras torulosum* Subzone independent of the CaO content. Certain taxa such as *Paracypris* sp. 1 (generally preferring a deeper marine environment) or *Cytheropterina cribra* (Fischer) have corresponding minima at the boundary of each of the *P. torulosum*, *L. opalinum* and *L. comptum* Subzones. Others, such as *Praeschuleridea, Kinckelinella* and *Cytherelloidea*, increase at these boundaries.

Foraminifera

The most striking correlations with boundaries of established ammonite stratigraphy are exhibited by trends in benthic foraminifera (Fig. 2). The Epistominidae reach a peak (57 %) at the base of the *P. torulosum* Subzone (Upper Toarcian) and a subsequent minimum in the horizon of *Cotteswoldia burtonensis* (Buckman). Another peak follows in the horizon of *Pleydellia comata* Buckman. In the horizon of *Pleydellia misera* (Buckman) at the base of the Aalenian they disappear almost completely, recovering only temporarily to about 7 % at the base of the *L. comptum* Subzone. The Ophthalmidiidae have a similar distribution but attain only low percentages in KB Wittnau.

On the contrary, most arenaceous foraminifera reach a peak just when the Epistominidae are at their minima. As a whole, rising percentages of arenaceous foraminifera correspond with rising SiO_2 content through the Lower Aalenian. But in detail their distribution also appears to be influenced by biological events: for instance, they have a peak in the *C. burtonensis* horizon without any change of the SiO_2 content, and a distinct decline at the base of the *L. comptum* Subzone just when the SiO_2 content reaches its maximum. In particular, *Trochammina topagorukensis* Tappan reacts sensitively at every boundary; for instance, it reflects even the boundary horizon at 37.5m with a short decline, where radiolarians also decrease.

Mesofauna: The samples also yielded nuclei of ammonites, serpulids, and crustacean remains, which may contribute to understanding the character of several boundaries. Thus, the appearance of Lytoceratid nuclei (Fig. 2) at the base of the Aalenian corresponds to peaks of radiolarians and of *T. topagorukensis*. The disappearance of the Lytoceratids at 37.5m coincides with declines of both cited fossil groups. The presence of numerous serpulids at the base of the *L. comptum* Subzone coincides with a minimum of *T. topagorukensis* and a marked decline of radiolarians.

Discussion

The strong increase of radiolarians and Lytoceratid nuclei at the base of the Aalenian suggests an opening of the basin to the Tethys. This is substantiated by earlier observations in the field, where a transgression in the *P. misera* horizon was demonstrated by Schirardin (1961), 40km to the NW of Wittnau. Obviously this transgression is also reflected by the strong rise of certain arenaceous foraminifera (also proved at the base of the Aalenian in northern Switzerland by Tröster, 1987) and the simultaneous decline of Epistominidae. Similarly, the *C. burtonensis* horizon (at 70m) may be interpreted as a short transgressive event within the *P. torulosum* subzone which is generally dominated by the Epistominidae. Conversely, declines of Lytoceratid nuclei and radiolarians at the nodular horizon at 37.5m and again at the base of the *L. comptum* Subzone suggest regressive events.

Upper Aalenian - Basal Bajocian

The Upper Aalenian of South west Germany is less suitable for such investigations: in marginal areas (eastern Swabian Alb, Oberrhein area) it is developed as iron oolites. In the basin there exist only few ammonite bearing horizons and the microfaunas are generally very poor. In the section at Breitenbach near Reutlingen (central Swabian Alb), the uppermost 10m of the Aalenian are barren. The appearance of ammonites and rich microfaunas (Ohmert, 1988, Fig. 3) there coincides with the Bajocian transgression in the 'Sowerbyi-Oolith'. A similar event was reported by Gregory (1990) from Scotland, there slightly preceding the base of the Bajocian. The Epistominidae appear in the Breitenbach section but in the *Witchellia laeviuscula* zone (microfaunal horizon h 5). The upper part of this zone becomes almost barren again.

The Lower Bajocian in KB Pfullingen - Breitenbach

Recently (1997) at Breitenbach a core (KB Pfullingen - Breitenbach) was recovered from the basal Upper Bajocian down to the *Emileia sauzei* Zone (Fig. 4). The succession and its ammonite fauna were partially known from neighbouring brooklets (Ohmert, 1990) but the core now supplied a complete section with reliable thicknesses and unweathered samples. Following the experience at Wittnau the microfauna was studied semiquantitatively.

Lithology

The section begins at the base in the sandy limestones of the Blaukalk Member, followed by 19.5m of clays of the Giganteuston Member. The latter is subdivided by a limestone marker bed, the 'Dorsetensienbank', accompanied by a subjacent pyritic layer. The upper part of this member is characterized by very fossiliferous limestone nodules. The superjacent Blagdenischichten Member consists of alternating marls and limestone beds, overlain by iron oolitic nodules of the Subfurcaten-Oolith and clays of the Hamitenton Formation.

Ostracods

These show remarkable renewals in this section (Ohmert, 1990, Fig. 6) which coincide with the zonal boundaries (*Emileia sauzei/Stephanoceras humphriesianum* and *Stephanoceras humphriesianum/Strenoceras niortense)*. Another distinct change in the ostracod fauna is seen within the *Dorsetensia romani* Subzone, without any recognizable corresponding change of the ammonites. A maximum of the total abundance of ostracods at the base of the horizon of *Stemmatoceras frechi* (Renz) is caused by an increase in specimens of larval stages of only two species [*Systenocythere* cf. *concentrica* (Permjakova) and *Cytheropterina* sp.]. Another peak at the base of the *romani* subzone is caused by a single species of *Bythocypris*.

Foraminifera

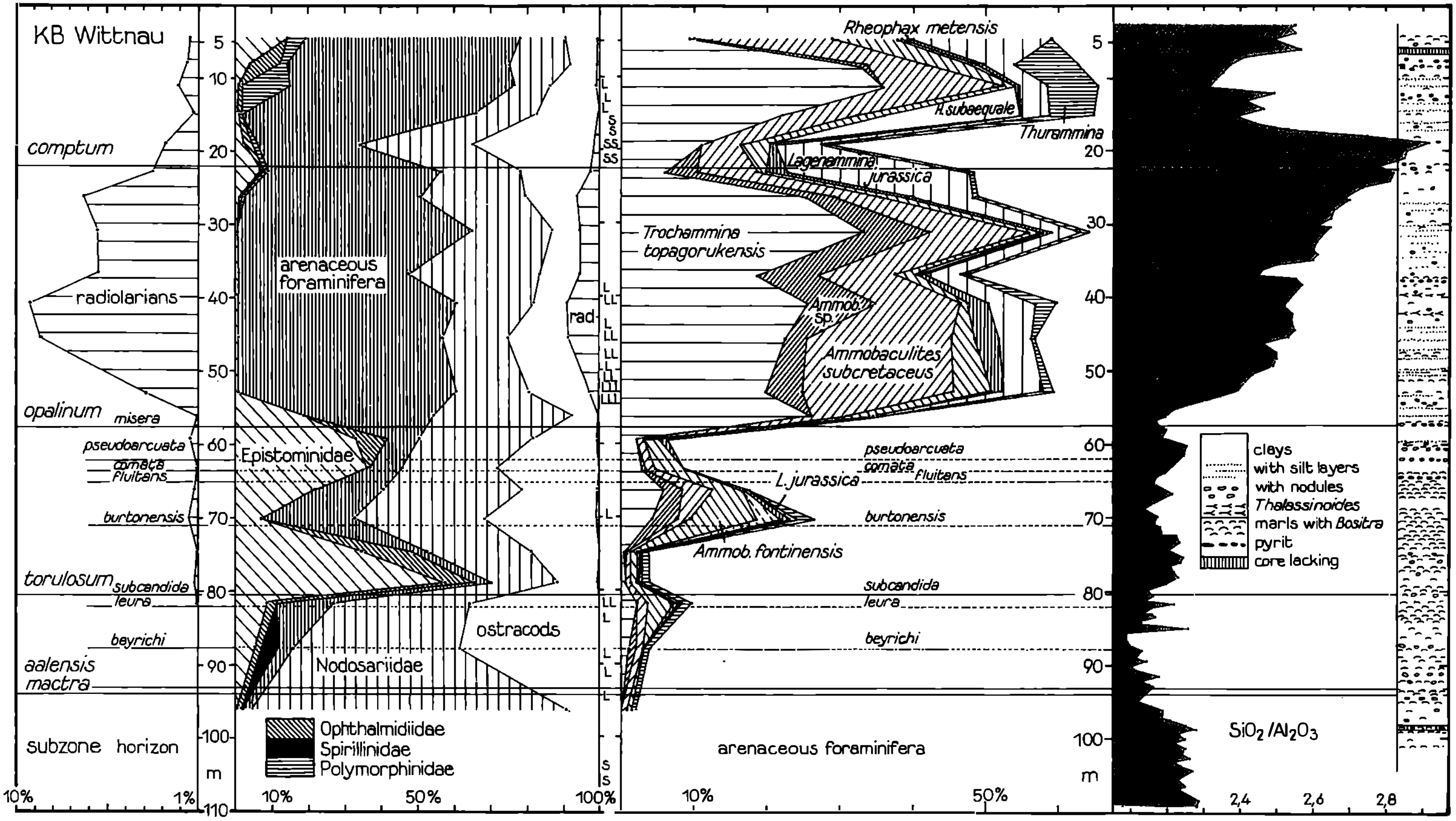
These are dominated in this section by the Epistominidae and Ophthalmidiidae, which together can make up as much as 90 % of the microfauna. The Epistominidae have their maximum in the lower part of the *S. humphriesianum* Zone (*Dorsetensia pinguis* Subzone), decreasing gradually during the *D. romani* and *P. cycloides* Subzones and particularly in the *Teloceras blagdeni* Subzone. In the Upper Bajocian they increase again to more than 80 % of the microfauna (in the *Garantiana baculata* Subzone). Within the *S. humphriesianum* Zone, however, abrupt declines down to less than 5 % are observed to corresponded with almost every boundary of the ammonite horizons. At the base of the *Stemmatoceras frechi* horizon and of the *Stephanoceras umbilicus* horizon they are reduced to less than 1 %, as well as at the base of the Upper Bajocian. The latter boundary in this region was shown by Dietl et al. (1979) to precede the ironoolitic facies of the 'Subfurcaten-Oolith'. As indicated by the minimum of the Epistominidae the boundary may even lie half a meter below the lowermost occurrence of *Teloceras banksi* (Sowerby) so far recorded. At this same level the genus *Ophthalmidium* also passes through a minimum, whereas the encrusting Ophthalmidiidae become dominant. The same was observed at the bases of the *T. blagdeni* Subzone, *P. cycloides* Subzone, and *D. romani* Subzone. In the *D. pinguis* Subzone, however, *Ophthalmidium* increases inversely to the Epistominidae which fall to a minimum at the base of the subzone and particularly at the base of the *S. frechi* horizon. It may be that two close events are included in the same sample (each sample spans about 10cm of the core).

Quantitative distribution of the arenaceous foraminifera also reflects the ammonite boundaries in the lower part of the *S. humphriesianum* Zone. They disappear completely at the base of the zone and again at the base of the *S. frechi* horizon, but soon recover to 20 %. From only 1 % at the base of the *D. romani* Subzone they increase to more than 14 % through this subzone. In the upper part of the zone they remain at constantly low percentages.

Discussion

Within the *S. humphriesianum* Zone there is a repeated alternation between Sonniniid and Stephanoceratid assemblages, as shown by Westermann (1954, fig. 5, tab. 1) 400km away in northwest Germany. The same succession of faunal horizons was observed again in the ironoolitic facies at the Oberrhein (Ohmert et al., 1995, Fig. 56), obviously being independent of lithological facies. The bases of the Stephanoceratid assemblages (*S. frechi* and *S. umbilicus* horizons) correspond in KB Pfullingen-Breitenbach to the most striking minima of the Epistominidae (under 1%).

Figure 2. Relative percentage distribution of the microfauna (on the left) in the uppermost Toarcian-Lower Aalenian of KB Wittnau. L means nuclei of Lytoceratids, S serpulids in the mesofauna. Percentage distribution of the arenaceous (agglutinated) foraminifera (in the middle) compared with the SiO_2/Al_2O_3 ratio on the right.



KB Wittnau
comptum
opalinum
misera
pseudoarcuata
comata
fluitans
burtonensis
torulosum
subcandida
leura
beyrichi
aalensis
mactra
subzone
horizon
radiolarians
10%
1%
5
10
20
30
40
50
60
70
80
90
100
m
110
arenaceous foraminifera
Epistominidae
Nodosariidae
ostracods
rad
Ophthalmidiidae
Spirillinidae
Polymorphinidae
10%
50%
100%
L
S
SS
LL
Rheophax metensis
H. subaequale
Thurammina
Lagenammina jurassica
Trochammina topagorukensis
Ammob. sp.
Ammobaculites subcretaceus
L. jurassica
Ammob. fontinensis
arenaceous foraminifera
clays
with silt layers
with nodules
Thalassinoides
marls with Bositra
pyrit
core lacking
SiO2 /Al2O3
2,4
2,6
2,8

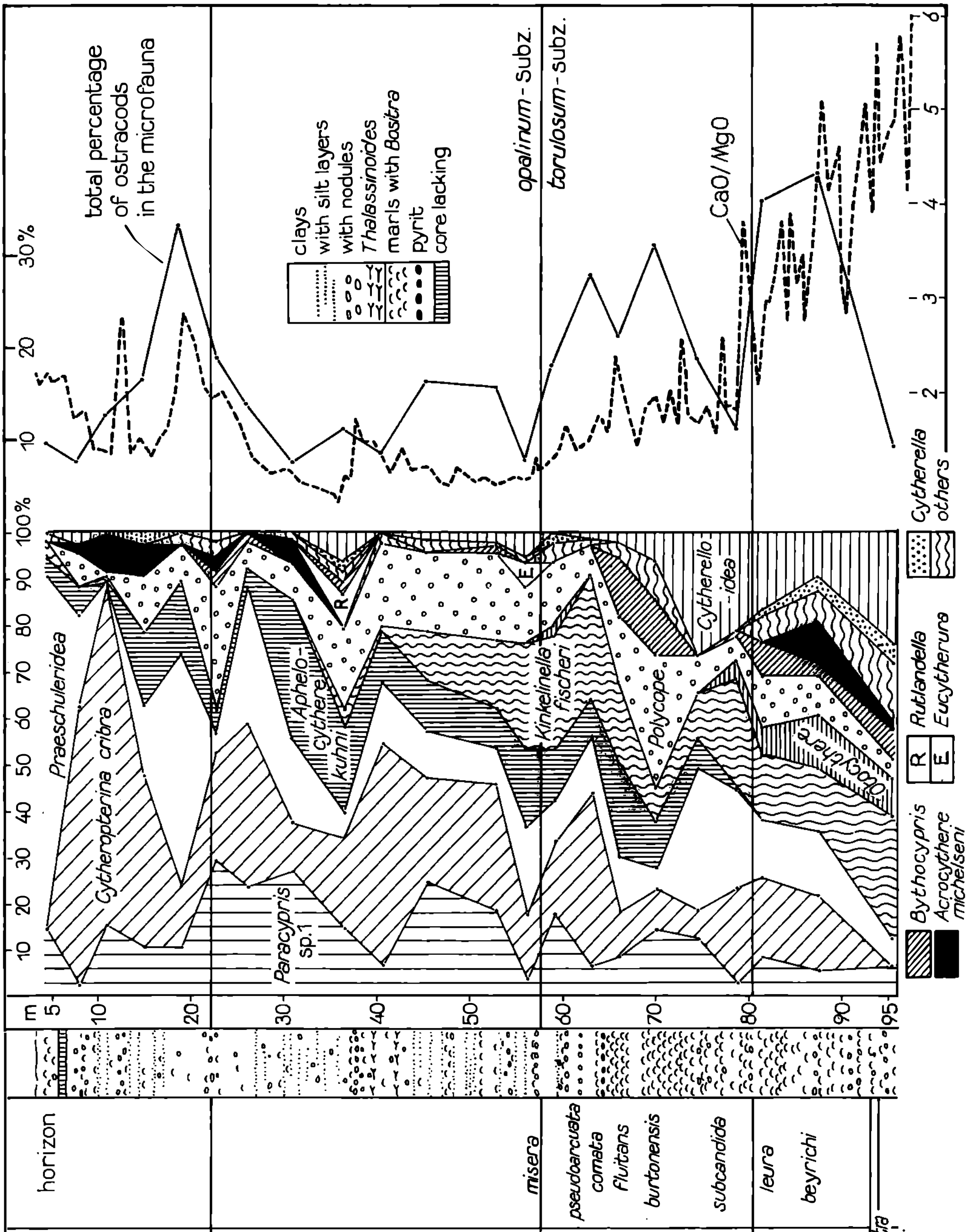


Figure 3. Relative percentage distribution of the ostracods in the uppermost Toarcian-Lower Aalenian of KB Wittnau compared with the CaO/MgO ratio.

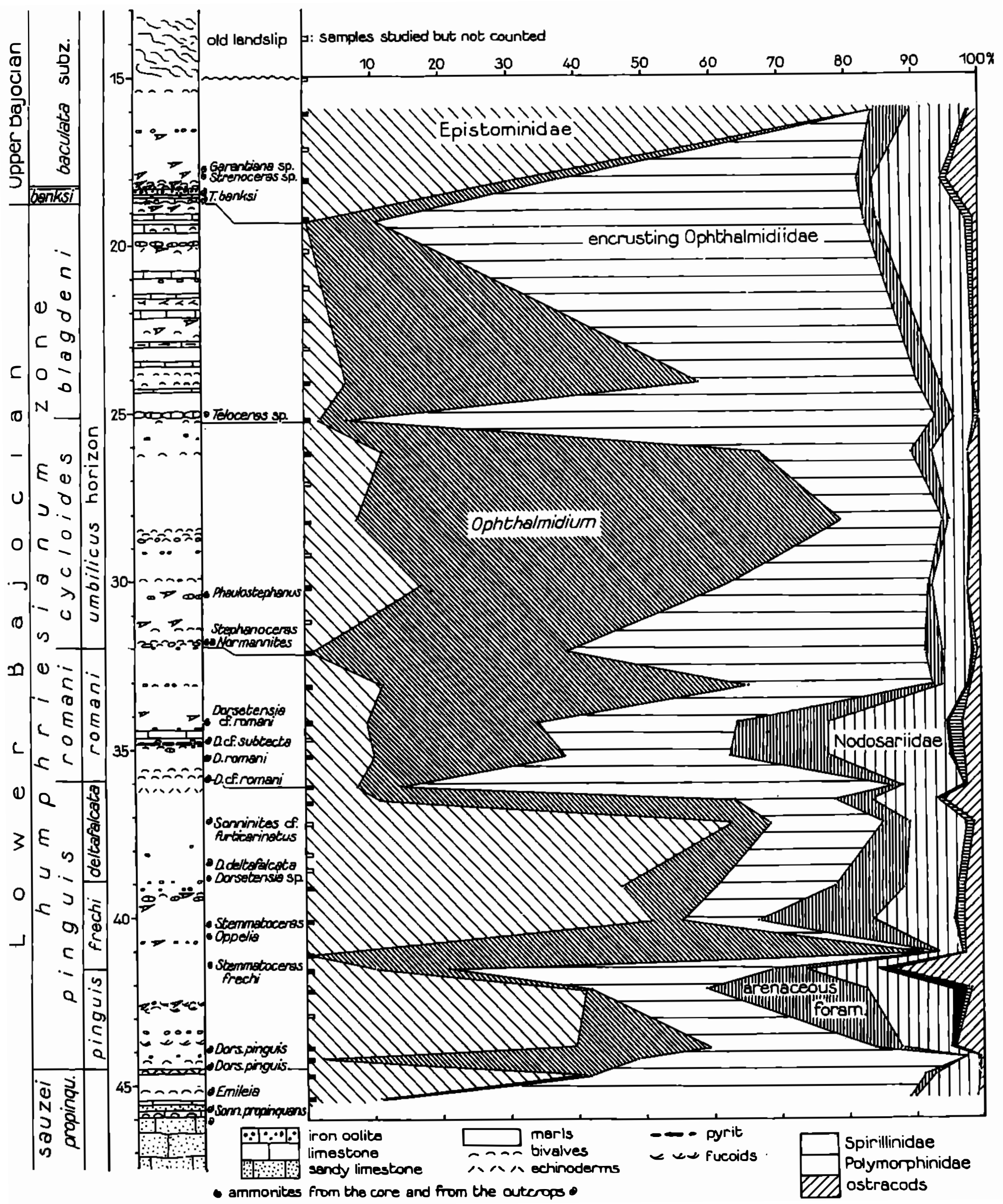


Figure 4. Relative percentage distribution of the microfauna in the Lower-Upper Bajocian of KB Pfullingen-Breitenbach.

An equally strong minimum of Epistominidae at the base of the Upper Bajocian (*T. banksi* Subzone) is also connected with a Stephanoceratid (*Teloceras*) fauna (see Dietl et al., 1979, fig. 3). It corresponds with a transgression as does the (almost) disappearance of Epistominidae at Wittnau at the base of the Aalenian. Accordingly the *S. frechi* and *S. umbilicus* horizons (the latter with a 1cm layer of jet at its base at 32m depth) may be interpreted as transgressive horizons recognizable at least everywhere in Central Europe.

Conclusions

Many fossil groups exhibit distinct changes across the boundaries defined by ammonites in the Aalenian-Bajocian of South west Germany. In particular, the relative percentage distribution of the benthic foraminifera reflects most ammonite boundaries (down to the ammonite horizons), suggesting general biological events. Such microfaunal changes may then be useful for identifying the exact position of a boundary when ammonites are not present throughout. The most striking biological events are related to transgressions (particularly at the stage boundaries), while others may be interpreted as depending on regressive sea level changes. Comparable microfaunal changes in the Toarcian of France were related to eustatic cycles (Bonnet et al., 1992a; 1992b).

References

BALDANZA, A., AND M. MATTIOLI. 1996. The calcareous nannofossils in the Wittnau KB borehole. Aalenews, 6:21-27.

BONNET, L., A. M. BODERGAT, R. CUBAYNES, J. REY, AND C. RUGET. 1992a. Fluctuation des biocoenoses d'Ostracodes dans le Toarcien du Quercy. Relations avec les cycles eustatiques. Comptes rendus des séances de l'Académie des Sciences, Série II, 314:953-960.

____, R. CUBAYNES, J. REY, AND C. RUGET. 1992b. Cycles et rythmes dans les biocénoses de Foraminifères du Toarcien du Quercy. Relations avec les cycles eustatiques et les changements climatiques. Comptes rendus des séances de l'Academie des Sciences, Série II, 314:87-83.

DIETL, G., F. BEMMERER, AND G. NETH. 1979. Zur Stratigraphie des Subfurcaten Ooliths (Ober-Bajocium, Dogger) entlang der Mittleren Schwäbischen Alb, Baden-Württemberg. Jahreshefte der Gesellschaft für Naturkunde in Württemberg, 134:85-95.

GREGORY, F. J. 1990. Foraminiferal biostratigraphy and paleoecology across the Aalenian/Bajocian of the proposed basal Bajocian stratotype at Bearreraig Bay, NE Skye (Scotland). Memorie descrittive della Carta Geologica d'Italia, 40:257-268.

MARTIN, M. 1996. Geochemische Untersuchung der Forschungsbohrung KB Wittnau. Informationen, 8:17-19.

OHMERT, W. 1988. The *ovalis* Zone (Lower Bajocian) in the type area, southwestern Germany. 2nd International Symposium on Jurassic Stratigraphy, 1:255-268.

____. 1990. The *Humphriesianum* Zone in the type area. Memorie descrittive della Carta Geologica d'Italia, 40:117-140.

____, G. GASSMANN, R. H. SCHATZ, AND L. STETTER. 1995. Das höhere Unter-Bajocium (ehemals Mittel-Bajocium) von Lörrach. II. Die *humphriesianum*-Zone. Jahreshefte des Geologischen Landesamts Baden-Württemberg, 35:25-264.

SCHIRARDIN, J. 1961. Sur la limite du Toarcien et de l'Aalenien en Alsace. Bulletin du Service de la Carte Géologique d'Alsace et de Lorraine, 13:95-125.

TRÖSTER, J. 1987. Biostratigraphie des Obertoarcium und der Toarcium/Aalenium-Grenze der Bohrungen Weiach, Beznau, Riniken und Schafisheim (Nordschweiz). Eclogae geologicae Helvetiae, 80:431-447.

WESTERMANN, G. 1954. Monographie der Otoitidae (Ammonoidea), *Otoites, Trilobiticeras, Itinsaites, Epalxites, Germanites, Masckeites (Pseudotoites, Polyplectites), Normannites.* Beihefte zum Geologischen Jahrbuch, 15, 364 p.

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Correlation of Late Bathonian Ammonite Faunas between England and North East Spain and a Proposed Standard Zonation for the Upper Bathonian of Northern and Western Europe

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Abstract: The recent discovery of a relatively complete, ammonite-rich sequence in the Upper Bathonian near Aguilón, Iberian Chain, Spain, combined with a review of all known ammonite faunas of equivalent age in England, has facilitated a reappraisal of existing standard zonations for the substage in Europe. This work shows that the same basic zonal framework used in southern areas ("Submediterranean Province") also works well in northern areas, historically referred to a "Subboreal Province", where a separate zonation has been in use. Following the former scheme, the Upper Bathonian is here divided into two chronozones, respectively Retrocostatum and Discus, the former being the better developed at Aguilón. Three component subchronozones are here recognised in the Retrocostatum Chronozone - Quercinus, Hodsoni and Histricoides - identified on the basis of species of the widespread perisphinctids *Procerites* and *Homoeoplanulites*. Previous Submediterranean correlation schemes relied on genera rare in northern areas and they consequently emphasised differences rather than similarities. The apparent absence of *Clydoniceras* at Aguilón makes recognition of the terminal Bathonian Discus Chronozone problematic. Perisphinctids characteristic of the chronozone do appear to be present, however, and may provide an alternative means of correlating the zone. The relative expansion of the sequence at Aguilón and the association of almost all known European Upper Bathonian ammonite genera, indicate that the section has great potential as a stratotype for the substage in Europe.

Introduction

Correlation between Britain and more southerly areas of Europe (e.g., central and southern France, Iberia) has always been considered to be sufficiently good in the Lower Bathonian that essentially the same zonal scheme can be used in both regions (Mangold and Rioult, 1997). However, for the Middle and Upper Bathonian, separate schemes have been developed, in part due to an emphasis on faunal differences rather than similarities.

The recent discovery of a remarkably expanded, but ammonite-rich, sequence of alternating marls and limestones in the Upper Bathonian near Aguilón, northern Iberian Chain, Spain (Figure 1; Page and Meléndez, 1998), has suggested that virtually all the taxa recorded in Britain either do occur, or can be expected to occur, in so-called "Sub-Mediterranean" areas. Local absences in Britain can, therefore, be explained as being largely due to overall impoverishment of the fauna, as a consequence of unfavourable facies or non-preservation (Page, 1996).

Page (1996) was consequently able to demonstrate that a basic Submediterranean zonation for the Upper Bathonian (e.g. that of Mangold and Rioult,1997 following Mangold, 1970a; 1970b; 1984; 1991) works well in northern areas, historically referred to a "Subboreal" Province (e.g., Britain) where a separate zonation has been in use (Torrens, 1980; Westermann and Callomon, 1988; Mangold, 1991; Mangold and Rioult, 1997). A single zonation for the whole of Europe is therefore practical and preferable.

Following Mangold and other authors (including Torrens, 1974), the Upper Bathonian is here divided into two chronozones, respectively Retrocostatum and Discus, the former being the better developed at Aguilón. Page (1996) recognised three component subchronozones in the Retrocostatum Chronozone, namely Quercinus, Blanazense and Retrocostatum. Correlation of the

Quercinus Subchronozone is straightforward, as the index species, although often mis-named, is a readily identified giant *Procerites*. Recognition of the Blanazense and Retrocostatum subchronozones was, however, more difficult as the index species (and their genus *Prohecticoceras*) are not yet recorded in Britain. These subchronozones were consequently identified on the basis of the more ubiquitous species of the perisphinctids *Procerites* and *Homoeoplanulites*, which had also been found at Aguilón in association with the prohecticoceratid indices.

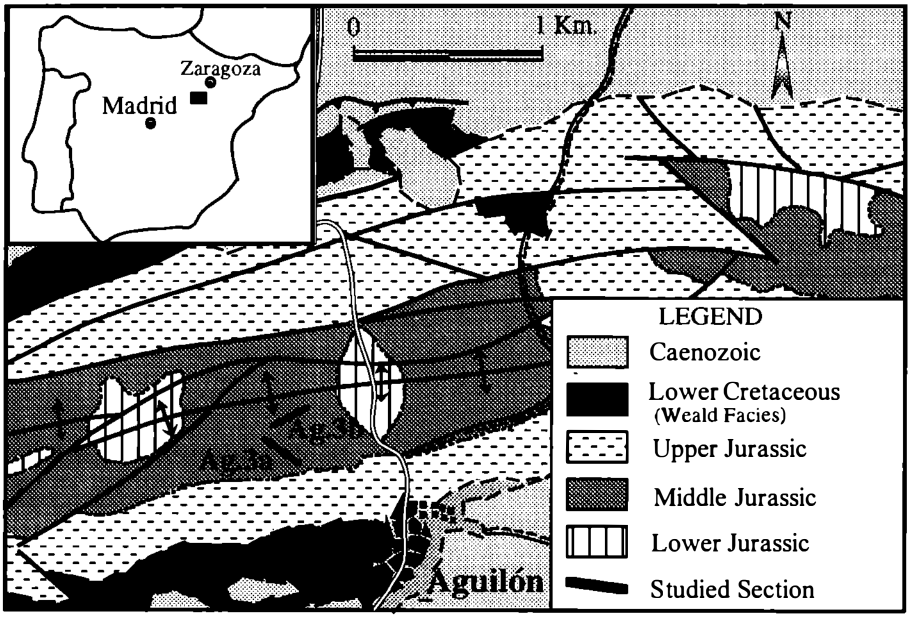


Figure 1. Map showing location of the Aguilón sections.

At Aguilón, as in Britain and elsewhere in Europe (southern Germany: Dietl and Callomon, 1988; France: Mangold 1970a; 1970b) it is these perisphinctids which show the most clear cut stratigraphic changes. Page and Meléndez (1998) therefore proposed that the two higher subchronozones of the Retrocostatum Chronozone be re-defined on the basis of their perisphinctid faunas as, respectively, a Hodsoni Subchronozone nov. (index: *Procerites hodsoni* Arkell) and a Histricoides Subchronozone (index: *Epistrenoceras histricoides* (Rollier)), the latter retaining the historical index but now also correlatable using perisphinctids. This latter scheme is utilised here.

The Sequence of Ammonite Faunas in the Upper Bathonian of Southern England

The Bathonian Stage in Britain is dominated by shallow marine and non-marine facies and ammonites are typically rare or absent. SW England, from the historical "type area" of the stage in the Bath district, to the Dorset coast, has yielded the majority of the known faunas and these were reviewed by Page (1996) as summarised below:

Retrocostatum Chronozone, Quercinus Subchronozone

In SW England the Quercinus Subzone is well developed in the Wattonensis Beds at the base of the Frome Clay in Dorset and South Somerset and their lateral equivalents in north Somerset (the Rugitela Beds, part, of Torrens, 1980). The fauna includes typical *Procerites* ex gr. *quercinus* Arkell ?not Terquem and Jourdy, *Homoeoplanulites* sp., *Eohecticoceras costatum* (Roemer), *Oxycerites* sp. and very rare *Bullatimorphites* characterising a *quercinus* Biohorizon (Page, 1996, fig. 2). Very rare elements of the same fauna, including *P. quercinus*, are known as far north as the East Midlands (Arkell, 1951-1958) in Blisworth Limestone facies.

Hodsoni Subchronozone

The type specimen of *Procerites hodsoni* Arkell came from a stratigraphically mixed fauna around the Fullers Earth Rock / Frome Clay junction in north Somerset but evidence from north east Spain has confirmed its stratigraphic position (Page, 1996; Page and Meléndez, 1998). In the Bath area, *P. hodsoni* and ?*Oxycerites* sp. are recorded in the Combe Down Oolite. These records can be taken together to identify a *hodsoni* Biohorizon.

Histricoides Subchronozone

The fauna of the Twinhoe Ironshot facies of Twinhoe Quarry near Bath includes frequent large *Oxycerites* ex gr. *orbis* Gumbel, *Procerites twinhoensis* Arkell (including the holotype) and *Eohecticoceras* ex gr. *costatum* Roemer. This fauna is comparable to those of the Histricoides Subchronozone in Spain and characterises a *twinhoensis* Biohorizon. Higher faunas are present in the same area and appear to include *Homoeoplanulites* sp. comparable to specimens from high in the Histricoides Subzone at Aguilón (*Homoeoplanulites* sp. Biohorizon). Records of *Clydoniceras*, including *C.* cf. *schlippei* S. Buckman above the Twinhoe Ironshot near Bath, and at equivalent levels farther south high in the Frome Clay (with ?*Oxycerites*) and basal Forest Marble (Boueti Bed), are grouped together within a cf. *schlippei* Biohorizon. A possible record of *Epistrenoceras* was noted by Torrens (1974) and if correctly identified would indicate the Histricoides Subchronozone. Unfortunately, the stratigraphic occurrence of the specimen is poorly known as it occurred in the same mixed assemblage as the type of *P. hodsoni.*

Discus Chronozone, Hollandi Subchronozone

The presence of the zonal index, *Clydoniceras hollandi* (J. Buckman), is virtually the only means of identifying the subzone, but it is very rare in Britain largely due to unfavourable facies. The species is known from around the level of the Digona Bed (Forest Marble) on the Dorset coast and in Wiltshire and Gloucestershire in Bradford Clay facies. A single *Homoeoplanulites* was apparently also recorded from a similar level in Gloucestershire (*hollandi* Biohorizon).

Discus Subchronozone

The Discus Subzone is well developed from the south coast of England (Dorset) as far north as Lincolnshire. At least two horizons are recognisable in the subzone, as first demonstrated by Pocock (1926) in the Oxford district. The lower horizon yields typical *C. discus* (J. Sowerby) (= *discus* Biohorizon) and the upper includes *C. hochstetteri* (Oppel), often with *Homoeoplanulites homeomorphus* Buckman (= *hochstetteri* Biohorizon) (Dietl, 1994). The close similarity, however, of the two *Clydoniceras* species and Arkell's (1951-1958) grouping together of sutural variations, means that most published records are of *Clydoniceras discus* and only rarely is *hochstetteri* separated (Page, 1996).

Herveyi Chronozone, Keppleri Subchronozone [Callovian]

The earliest Callovian faunas in south west England typically occur in the lowest part of the Fleet Member (= "Upper Cornbrash" of others) and correspond to the *keppleri* Biohorizon of the Herveyi Zone of Callomon et al. (1989). The fauna includes *Macrocephalites* cf. *jacquoti* Douvillé, *K. keppleri* and *Homoeoplanulites* ex gr. *homoeomorphus* (Page, 1988; 1996). The presence of the latter two taxa in Berry Member facies near Oxford was noted by Page (1989) but *K. keppleri* is now known from near the top of the unit at Berry Knap on the Dorset coast. Although recovered loose below the outcrop, the specimen came from close to the level yielding the specimen of *Clydoniceras* cited by Page (1995) and it can only, therefore, be assumed that the two taxa do not overlap in stratigraphic range.

The Sequence of Ammonite Faunas in the Upper Bathonian of North East Spain

Natural exposures and road cuttings close to the town of Aguilón (Iberian Cordillera, Zaragoza Province, Aragón) have long been known for their important Upper Jurassic successions. Little work, however, had been carried out on the Middle Jurassic sequence below and was virtually limited to the brief review of the Callovian contained in Sequeiros and Meléndez (1979). Page and Meléndez (1998), however, discovered a relatively expanded and ammonite-rich Upper Bathonian sequence below the Callovian. The faunas collected from the thick sequence of marl- limestone alternations include representatives of virtually all known European ammonite genera (with the exception of Phylloceratina, Lytoceratina and *Clydoniceras*), and have enabled, for the first time, the correct stratigraphic sequencing of many taxa described from elsewhere in Europe. Two separate sections, here designated Ag3A and Ag3B, and around 500m apart, have been sampled (Figure 2).

Retrocostatum Chronozone, Quercinus Subchronozone

A fauna including *Wagnericeras* spp. in Bed Ag3B-70 to Ag3B-78 indicates the latest Middle Bathonian Fortecostatum Subzone. The incoming of giant *Procerites* ex gr. *quercinus* (up to 400mm in diameter) in Bed Ag3B-94 marks the base of the succeeding Quercinus Subchronozone and hence the Retrocostatum Chronozone and the Upper Bathonian. The species persists to as high as Ag3B-124 but through much of this range other ammonites are relatively few and poorly characterised (*e.g. Bullatimorphites, Prohecticoceras*, *Oxycerites*). In Ag3B-118, however, a more diverse fauna, in addition to *P. quercinus*, includes *Bullatimorphites* (*B.*), *P.* ex gr. *blanazense* (Elmi) (including *P. dominyoni* (Elmi)), *Oxycerites* and a very finely ribbed small *Cadomites* (*=Cadomites* sp. 1 Biohorizon of Page and Meléndez, 1998).

Hodsoni Subchronozone

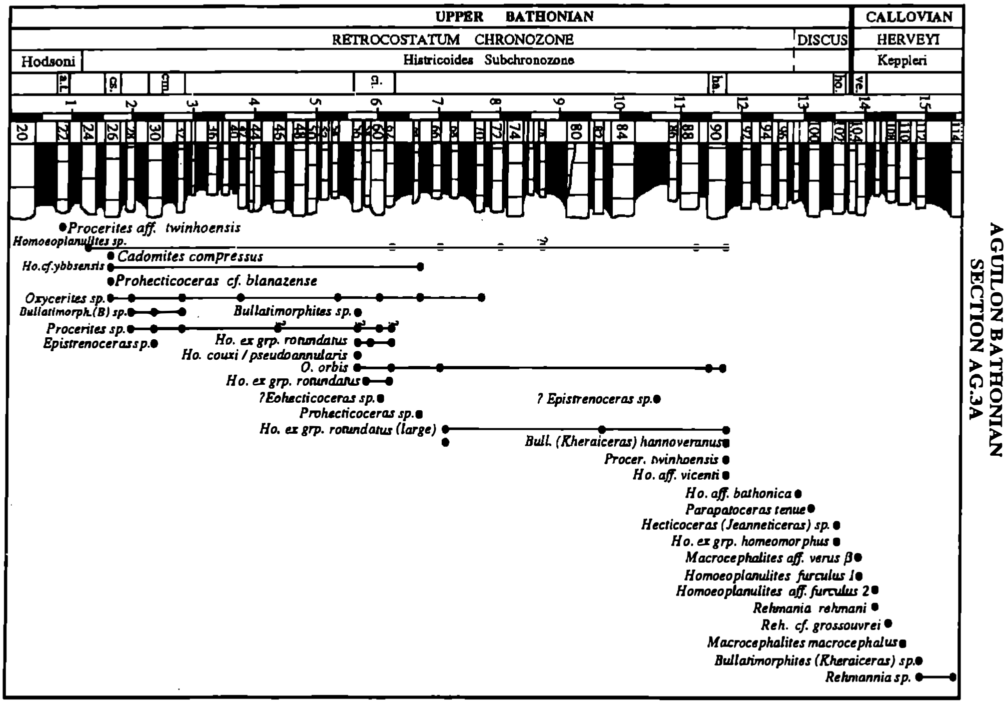
The appearance in Bed Ag3B-126 of relatively evolute *Procerites* with ribbing typically persisting to 230-285mm marks the base of the Hodsoni Subchronozone. Typical specimens have a squarer whorl section than *P. quercinus*. The only other ammonite recovered from Ag3B-126 is a fine ribbed nucleus of *Bullatimorphites* (= *hodsoni* Biohorizon). *P. hodsoni* persists to Ag3B-130 but is replaced by a smaller and more involute *Procerites* in Bed Ag3B-132, with relatively round whorled nuclei (cf. "*Proc. clausiprocerus*"; Arkell, 1951-1958, pl 23, fig 5 only), compressed middle and outer whorls, and macroconchs maturing at around 175-235 mm (= aff. *twinhoensis* Biohorizon). Similar *Procerites* persist at least as high as Ag3B-138, associated with a few *Oxycerites* sp.and a possible *Prohecticoceras*.

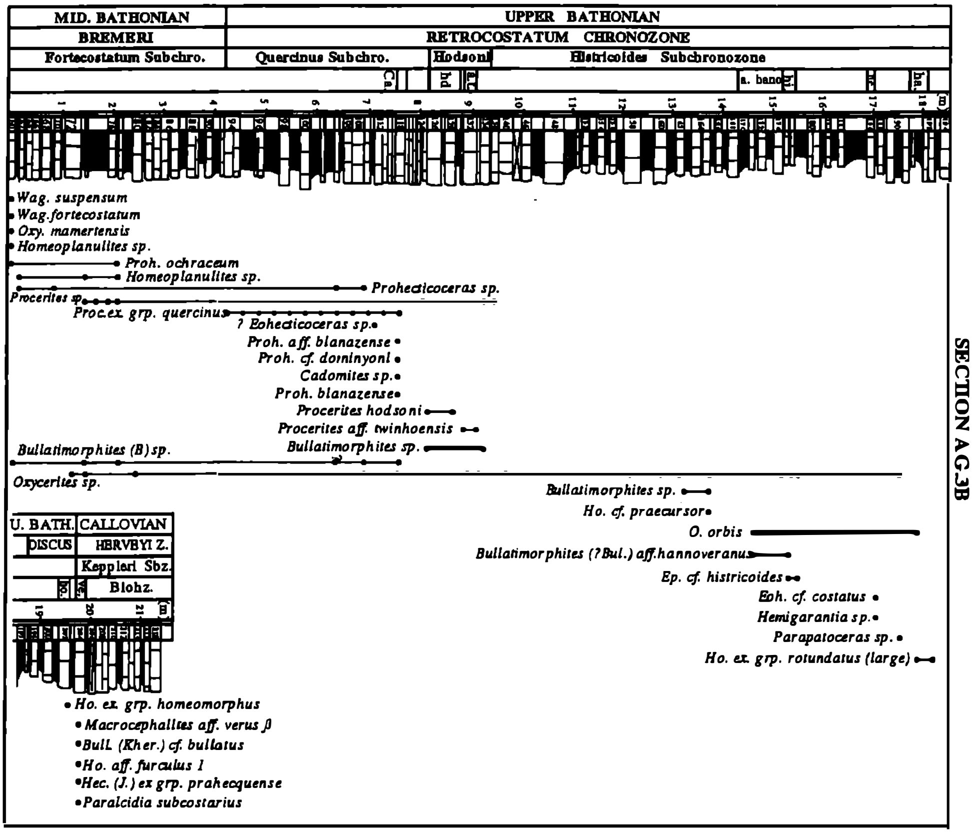
Histricoides Subchronozone

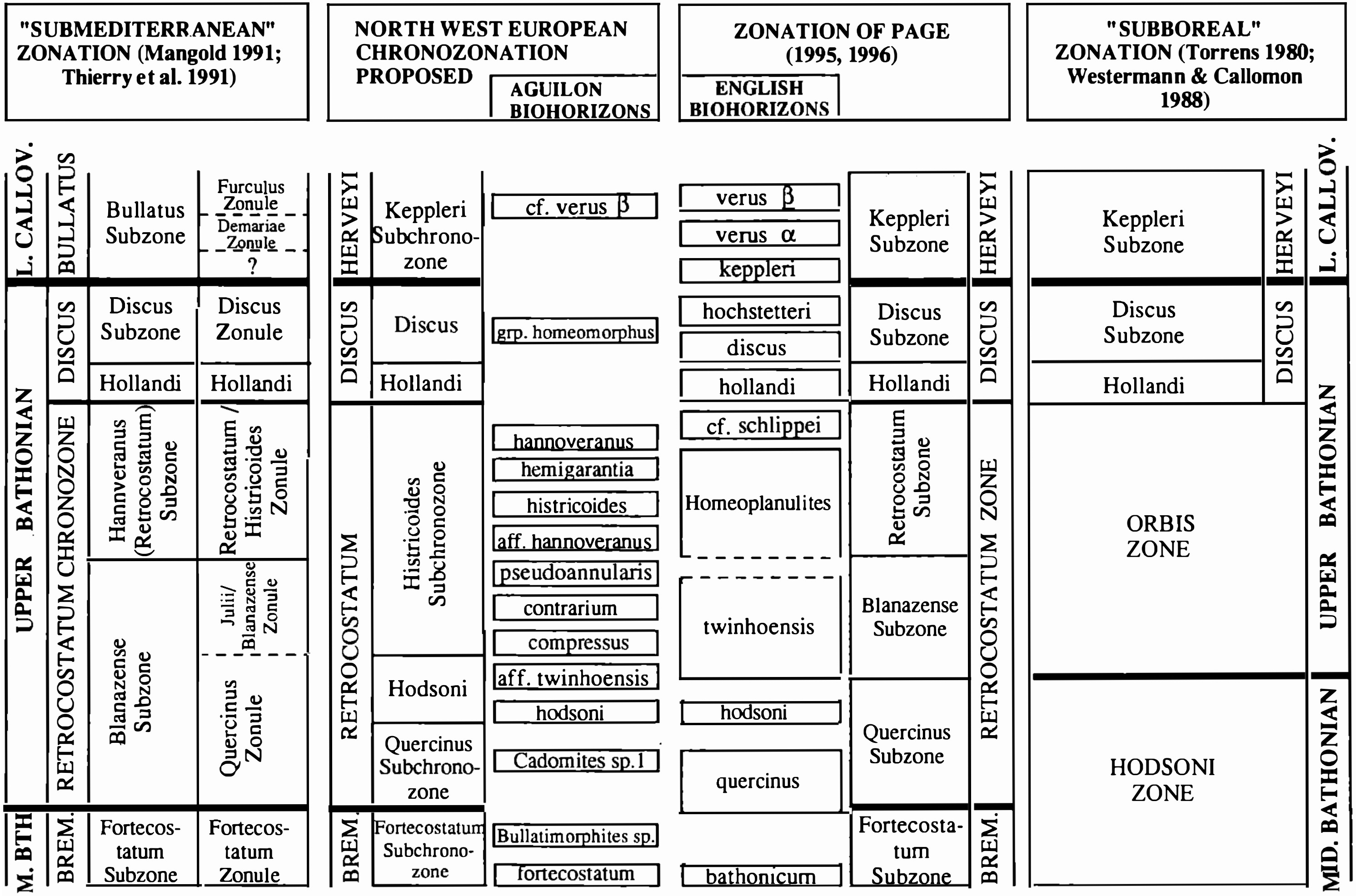
A sudden change from *Procerites* dominated faunas to faunas in which *Homoeoplanulites* is prominent or dominant marks the base of the subzone. In Ag3A the lowest of these is present in Ag3A-24, immediately above *P.* aff. *twinhoensis*. Ag3A-26 yields a distinctive fauna with abundant "*Cadomites*" *compressus* de Groussouvre, small forms with a *Berbericeras*-like body chamber. *Homoeoplanulites* is also present with *Proh.* cf. *blanazense* and *Oxycerites* (= *compressus* Biohorizon). This fauna has not yet been definitely characterised in Ag3B, and correlation of the interval Ag3B-140 to Ag3B-189 with Ag 3A remains poor.

Ag3A-30 to Ag3A-33 yields a limited fauna most notable for the occurrence of *Epistrenoceras* sp. aff. *contrarium* (Behr.), with *Oxycerites*, *Bullatimorphites* (*B.*) sp., coarsely ribbed *Homoeoplanulites* and a few *Procerites* (= *contrarium* Biohorizon). Common *Homoeoplanulites* in Ag3A-56 to 58 include medium-sized forms, *H.* ex gr. *rotundatus* (Roemer), and small, very evolute, *H.* cf. *pseudoannularis* (Lissajous) with *Oxycerites*, *Bullatimorphites* and *Proh.* cf. *retrocostatum* (= *pseudoannularis* Biohorizon nov.). Slightly higher levels in Ag3B yield *H.* cf. *praecursor* (Mangold), possible *Procerites*, *Oxycerites* and *Bullatimorphites* (*?B.*) aff. *hannoveranus* (the early coarse ribbed [M] form of Mangold, 1970a, text figs 96, 97) (= aff. *hannoveranus* Biohorizon). *Epistrenoceras* cf. *histricoides* in Ag3A-85 indicates a further distinctive fauna best characterised in Ag3B-176 to 178, with associated *O. orbis* and *B.* ex gr. *hannoveranus* (= *histricoides* Biohorizon). Ag 3B-186 has yielded *Eohecticoceras* ["*Alcidellus*"] *costatus*, *Ox. orbis*, *Homoeoplanulites* and *Hemigarantia*, the last collected in a loose block but in typical brachiopod-rich accumulations (= *Hemigarantia* Biohorizon).

Figure 2. Stratigraphic distribution of ammonite faunas in the Upper Bathonian at Aguilón. Abbreviations: ve.= aff. *verus* β, ho.= grp. *homoeomorphus*, ha.= *hannoveranus*, He.= *Hemigarantia,* hi= *histricoides,* a.hano.= aff. *hannoveranus,* ps.= *pseudoannularis,* cm.= *contrarium,* cs.= *compressus,* a.t.= aff. *twinhoensis,* hd.= *hodsoni,* Ca.= *Cadomites* sp.1.


UPPER BATHONIAN
CALLOVIAN
RETROCOSTATUM CHRONOZONE
DISCUS
HERVEYI
Hodsoni
Histricoides Subchronozone
Keppleri
•Procerites aff. twinhoensis
Homoeoplanulites sp.
• Cadomites compressus
Ho.cf.ybbsensis
•Prohecticoceras cf. blanazense
Oxycerites sp.
Bullatimorph.(B) sp.
Bullatimorphites sp.
Procerites sp.
Epistrenocerassp.
Ho. ex grp. rotundatus
Ho. couxi / pseudoannularis
O. orbis
Ho. ex grp. rotundatus
?Eohecticoceras sp.
? Epistrenoceras sp.
Prohecticoceras sp.
Ho. ex grp. rotundatus (large)
Bull. (Kheraiceras) hannoveranus
Procer. twinhoensis
Ho. aff. vicenti
Ho. aff. bathonica
Parapatoceras tenue
Hecticoceras (Jeanneticeras) sp.
Ho. ex grp. homeomorphus
Macrocephalites aff. verus β
Homoeoplanulites furculus 1
Homoeoplanulites aff. furculus 2
Rehmania rehmani
Reh. cf. grossouvrei
Macrocephalites macrocephalus
Bullatimorphites (Kheraiceras) sp.
Rehmannia sp.
AGUILON BATHONIAN SECTION AG.3A


MID. BATHONIAN
UPPER BATHONIAN
BREMERI
RETROCOSTATUM CHRONOZONE
Fortecostatum Subchro.
Quercinus Subchro.
Hodsoni
Histricoides Subchronozone
Wag. suspensum
Wag.fortecostatum
Oxy. mamertensis
Homeoplanulites sp.
Proh. ochraceum
Homeoplanulites sp.
Prohecticoceras sp.
Procerites sp.
Proc.ex. grp. quercinus
? Eohecticoceras sp.
Proh. aff. blanazense
Proh. cf. dominyoni
Cadomites sp.
Proh. blanazense
Procerites hodsoni
Procerites aff. twinhoensis
Bullatimorphites (B) sp.
Bullatimorphites sp.
Oxycerites sp.
Bullatimorphites sp.
Ho. cf. praecursor
O. orbis
Bullatimorphites (?Bul.) aff.hannoveranus
Ep. cf. histricoides
Eoh. cf. costatus
Hemigarantia sp.
Parapatoceras sp.
Ho. ex. grp. rotundatus (large)
U. BATH.
CALLOVIAN
DISCUS
HERVEYI Z.
Keppleri Sbz.
Blohz.
• Ho. ex. grp. homeomorphus
• Macrocephalites aff. verus β
•Bull. (Kher.) cf. bullatus
•Ho. aff. furculus 1
•Hec. (J.) ex grp. prahecquense
•Paralcidia subcostarius
AGUILON BATHONIAN SECTION AG.3B



"SUBMEDITERRANEAN" ZONATION (Mangold 1991; Thierry et al. 1991)				NORTH WEST EUROPEAN CHRONOZONATION PROPOSED		AGUILON BIOHORIZONS	ZONATION OF PAGE (1995, 1996): ENGLISH BIOHORIZONS	ZONATION OF PAGE (1995, 1996)		"SUBBOREAL" ZONATION (Torrens 1980; Westermann & Callomon 1988)		
L. CALLOV.	BULLATUS	Bullatus Subzone	Furculus Zonule	HERVEYI	Keppleri Subchronozone	cf. verus β	verus β	Keppleri Subzone	HERVEYI	Keppleri Subzone	HERVEYI	L. CALLOV.
			Demariae Zonule				verus α					
			?				keppleri					
UPPER BATHONIAN	DISCUS	Discus Subzone	Discus Zonule	DISCUS	Discus	grp. homeomorphus	hochstetteri	Discus Subzone	DISCUS	Discus Subzone	DISCUS	UPPER BATHONIAN
							discus					
		Hollandi	Hollandi		Hollandi		hollandi	Hollandi		Hollandi		
	RETROCOSTATUM CHRONOZONE	Hannveranus (Retrocostatum) Subzone	Retrocostatum / Histricoides Zonule	RETROCOSTATUM	Histricoides Subchronozone	hannoveranus	cf. schlippei	Retrocostatum Subzone	RETROCOSTATUM ZONE	ORBIS ZONE		
						hemigarantia						
						histricoides	Homeoplanulites					
						aff. hannoveranus						
		Blanazense Subzone	Julii/ Blanazense Zonule			pseudoannularis	twinhoensis	Blanazense Subzone				
						contrarium						
			Quercinus Zonule			compressus						
					Hodsoni	aff. twinhoensis						
						hodsoni	hodsoni	Quercinus Subzone		HODSONI ZONE		MID. BATHONIAN
					Quercinus Subchronozone	Cadomites sp.1	quercinus					
M. BTH	BREM.	Fortecostatum Subzone	Fortecostatum Zonule	BREM.	Fortecostatum Subchronozone	Bullatimorphites sp.		Fortecostatum Subzone	BREM.			
						fortecostatum	bathonicum					

Ag 3A-90 and its correlatives in Ag3B-191 to 192, yield an abundant fauna dominated by large *Homoeoplanulites* ex gr. *rotundatus* (including *H.* aff. *vicenti* Mangold) with rare large *P.* cf. *twinhoensis* Arkell, occasional *B.* (*Kheraiceras*) *hannoveranus* s.s., typical large *O.* cf. *orbis* and a possible *Prohecticoceras* (= *hannoveranus* Biohorizon). Similar giant *Homoeoplanulites* ex gr. *rotundatus* (Roemer) (=*H. arisphinctoides* Mangold *non* Arkell) up to 320mm in diameter are first recorded in Ag3A-66.

Discus Chronozone

Homoeoplanulites (*H.*) ex gr. *homoeomorphus* in Ag3B-202 and Ag3A-102 appear to indicate the Discus Subzone as they include specimens very close to forms from the English "Lower Combrash" described by Arkell (1951-1958; Page, 1996). A single *Hecticoceras* (*Jeanneticeras*) sp.[m], with strong clavi-like secondary ribs, projected ventro-laterally on the body chamber, is also recorded. In the absence of *Clydoniceras* the base of the Discus Zone is impossible to draw, the only faunal records between Ag3A-90 and 102 and Ag3B-192 and Ag3B-202 being rare *Homoeoplanulites*, *Parapatoceras and ?Oxycerites*.

Herveyi/Bullatus Chronozone, Keppleri Subchronozone [Callovian]

The top of the Bathonian sequence is marked by a marly, iron-rich bed full of Lower Callovian ammonites (Bed Ag 3A/3B-104), suggesting taphonomic condensation. A rich early Callovian fauna with abundant *Macrocephalites* aff. *verus* Buckman (= *M. verus* β in Page, 1995) is present, indicating the upper part of the Keppleri Subzone and therefore confirming the presence of a small non sequence.

Biostratigraphic Correlation Between England and Spain and Other Regions in Europe

The Quercinus Subzone is clearly recognisable in England, Spain and France (as Mangold=s "Horizon à *mirabilis*; 1970a; 1984) and therefore forms a valuable means of correlating the base of the Upper Bathonian (Figure 3). The subzone is broadly equivalent to a single biohorizon in England, but can be subdivided in the Iberian Basin (eastern Spain).

In both Spain and England a *hodsoni* Biohorizon is recognisable and they correlate closely. The biohorizon is also clearly recognisable in Germany (Hahn, 1969, text. fig. 7) and a base for a Hodsoni Subchronozone is therefore potentially a widely recognisable level. The higher part of the subchronozone with "*P.* aff. *Twinhoensis*" in Spain is not presently recognisable in England but due to the great rarity of ammonites at this level in the latter area, this is not surprising. The presence of frequent large *Oxycerites* associated with large *Procerites twinhoensis* and occasional *Homoeoplanulites* (closely resembling those of Bed Ag3B-90) in the Twinhoe Beds in England suggests a correlation with the Histricoides Subchronozone in eastern Spain. *Procerites* is relatively common in England but rare in Spain; *Homeoplanulites* shows a reciprocal distribution, probably due to palaeobiogeographic effects. As noted above, the absence of *Clydoniceras* in Spain makes it impossible to correlate precisely higher levels of the Upper Bathonian. *H.* ex gr. *homoeomorphus* in Spain suggests a correlation with the Discus Subchronozone, but detailed information on the succession of Discus Chronozone perisphinctids is not currently available. *Clydoniceras* is, however, present in northern Germany and France at equivalent levels to Britain (Westermann, 1958; Mangold1970a; 1970b), and a re-examination of the associated perisphinctid faunas may ultimately aid correlations.

Chronostratigraphic Conclusions and Proposed Definitions

Retrocostatum Chronozone

Index: Prohecticoceras retrocostatum (de Grossouvre).
Author: Lissajous (1923).
Historical type area: Maçon, France (Mangold et al., 1974, p. 192; Torrens 1974, p. 590).

Figure 3. Correlation of ammonite biohorizons between England and NE Spain (Aguilón) and previous standard zonations for the Upper Bathonian.

(i) Quercinus Subchronozone:

Index: *Procerites quercinus* Arkell (1951-1958, ?non Terqueum and Jourdy).

Author: Callomon and Cope (1995), after the "Horizon à *mirabilis*" of Mangold (1970a); see Page (1995).

Proposed definition: Base of Bed Ag3B-94, section 3B, Aguilón, Iberian Cordillera, Zaragoza Province, Aragón, Spain.

Characteristic ammonite fauna: Giant species of *Procerites* up to 300mm or more in diameter, including *P. quercinus* sensu Arkell, and "*P. mirabilis*" Mangold non Arkell.

Discussion.- Th e *quercinus* fauna lies at the base of the Retrocostatum Zone of Mangold (1970a; 1970b; 1984) although, as a subzone, it was assigned to the upper part of the Hodsoni Zone by Callomon and Cope (1995). The widespread occurrence of the fauna makes it potentially very useful internationally for correlating the base of the Upper Bathonian.

The use of *P. quercinus* for species at this level follows Arkell (1951-1958). The Subchronozone is used here in a restricted sense, to include only those faunas with *P. quercinus* of others and excludes higher levels with *P. hodsoni* Arkell, included in the Subchronozone by Callomon and Cope (1995) and Page (1996).

(ii) Hodsoni Subchronozone:

Index: *Procerites hodsoni* Arkell.

Authors: Page and Meléndez (1998).

Proposed definition: Base of Bed Ag B-126, section 3B, Aguilón, Iberian Cordillera, Zaragoza Province, Aragón, Spain.

Characteristic ammonite fauna: *Procerites*, including *P. hodsoni* and *P.* sp. nov. aff. *twinhoensis*, with strong, sharp ribbing, persisting at least to middle whorlst.

Discussion: Torrens (1965) proposed an "Unnamed Zone" for faunas dominated by *Procerites*, below those with common *Oxycerites* ex gr. *orbis* (= "*O. aspidoides* Zone" of others) and above those with *Morrisiceras* ex gr. *morrisi* (Morrisi Chronozone). The new zone was eventually consolidated as a Hodsoni Zone by Torrens (1980) and included faunas with both *P. quercinus* and *P. hodsoni* and also earlier *Wagnericeras* of the Bremeri Zone. The first are here separated in a restricted Quercinus Subzone and the sudden appearance of sharply ribbed *Procerites* ex gr. *hodsoni* provides a valuable datum in Spain and Britain (Page, 1996) suitable for correlating the base of a subchronozone. Previous use of a restricted Blanazense Subzone above a broader Quercinus Subzone by Page (1996), is here discontinued as the change from *Proh. ochraceum* to *Proh.* ex gr. *blanazense* takes place already within the now restricted Quercinus Subchronozone.

(iii) Histricoides Subchronozone (emend):

Index: *Epistrenoceras histricoides* (Rollier).

Author: Collignon (1958), Elmi (1967).

Proposed definition: Base of Bed Ag3A-24, section 3A, Aguilón, Iberian Cordillera, Zaragoza Province, Aragón, Spain.

Characteristic ammonite fauna: Early species of *Homoeoplanulites*, including *H. praecursor* and *H. rotundatus*, are typical. *Bullatimorphites* ex gr. *hannoveranus* and large *Oxycerites* ex gr. *orbis* (= *O. oppeli* Elmi) are characteristic, as are *Epistrenoceras* spp., the latter at restricted levels, however. The lowest part of the subzone yields *Proh.* cf. *blanazense* and *Proh.* aff. *dominjoni* but at higher levels *Proh. retrocostatum* is present and appears to be restricted to the subzone.

Discussion: The use of *E. histricoides* as a subzonal index follows Elmi (1967) after Collignon's proposal of a *histricoides* 'Zone' in Madagascar (1958) and is retained here as it is one of the few taxa proposed as index for equivalent levels that genuinely appears to be restricted to the higher part of a conventional Retrocostatum Zone. The name has priority over the Retrocostatum Subzone of Mangold (1970a; 1970b), the Aspidoides Zone of Torrens (1974; later replaced by an Orbis Zone as *Oxycerites aspidoides* (Oppel) is now known to be a species from the Upper Bajocian; Dietl, 1982) and the Hannoveranus Subzone of Dietl and Callomon (1988). The change from *Procerites* to *Homoeoplanulites* dominated faunas is, however, a more reliable indicator of stratigraphic level than any of the taxa conventionally used and provides a better correlation for the base of the subzone and this is therefore the emended use proposed here.

(iv), (v) Discus Zone, Discus Chronozone:

Index: *Clydoniceras discus* (J. Sowerby).

Author: S. S. Buckman (1898).

Type area: southern and eastern England (Torrens, 1974, p. 596). The apparent absence of *Clydoniceras* at Aguilón and throughout the Iberian basin makes recognition of the terminal Bathonian Discus Zone problematic. Perisphinctids characteristic of the zone are present however (e.g., *H.* ex gr. *homoeomorphus*), suggesting that paleobiogeographic factors may be important (*Clydoniceras* appears to be more common in northern regions and infrequent farther south; Page and Meléndez, 1998). As for the higher subzones of the Retrocostatum Zone, an alternative means of correlating the base of the Discus Zone may be necessary, perhaps using *Homoeoplanulites,* although information presently available from Aguilón is insufficient to enable identification of a suitable species. The interpretation of the Discus Zone is discussed further by Torrens (1974) and Page (1996).

Conclusions

The relative expansion of the sequence at Aguilón has facilitated the correct stratigraphic sequencing of many taxa described from elsewhere in Europe whose source levels have remained ambiguous. In particular, the association of almost all known European Upper Bathonian ammonite genera (with the notable exception of *Clydoniceras*), indicates that the section has great potential as a reference for the substage in Europe. This also demonstrates that the historical use of two standard zonations for European faunas of this age ("Subboreal" and "Submediterraenean") is unecessary.

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References

ARKELL, W. J. 1951-1958. A monograph of English Bathonian Ammonites. Monograph of the Palaeontogaphical Society, London, 264 p.

BUCKMAN, S. S. 1898. On the grouping of some divisions of so-called Jurassic time. Quarterly Journal of the Geological Society, 54:442-462.

CALLOMON, J. H. AND J. C. W. COPE. 1995. The Jurassic Geology of Dorset, p. 51-104. *In* P. D.Taylor (ed.), Field geology of the British Jurassic. Geological Society, London.

____ , G. DIETL, AND K. N. PAGE. 1989. On the Ammonite faunal horizons and Standard Zonation of the Lower Callovian Stage in Europe, p. 17-18. *In* R. B. Rocha and A. F. Soares (eds.), 2nd International Symposium on Jurassic Stratigraphy Lisbon, 1.

____ , AND H.-J. NIEDERHÖFER 1989. Die Ammonitenfaunen - Horizonte im Grenzbereich Bathonium/Callovium des Schwäbischen Juras und deren Korrelation mit W-Frankreich und England. Stuttgarter Beitrage zür Naturkunde, Series B. (Geologie und Palaeologie), 148:13p.

COLLIGNON, M. 1958. Atlas des fossiles caracteristiques de Madagascar. Service Géologique, Tanarive.

DIETL, G. 1982. Das wirkliche Funduiveau von *Ammonites aspidoides* OPPEL (Ammonoidea, Mittl. Jura) am *locus typicus*. Stuttgarter Beitrage zür Naturkunde, Series B. (Geologie und Palaeontologie), 87.

____ ,1994. Der *hochstetteri*-Horizont-ein Ammoniten faunen-Horizont (Discus-Zone, Ober-Bathonium, Dogger) ans dem Schwäbischen Jura. Stuttgarter Beitrage zür Naturkunde, Series B. (Geologie und Palaeontologie.), 202, 39 p.

____ , AND J. H. CALLOMON. 1988. Der Orbis-Oolith (Ober-Bathonium, Mittl. Jura). von Sengenthal/Opt., Fränk. Alb, und seine Bedeutung für die Korrelation und Gliederung der Orbis-Zone. Stuttgarter Beitrage sür Naturkunde, Series B. (Geologie und Palaeontologie), 142:31.

ELMI, S. 1967. Le Lias et le Jurassique Moyen 1=Ardeche. Documents du Laboratoire de Géologie, Faculté des Sciences, Lyon, 19, 845 p.

HAHN, W. 1969. Die Perisphinctidae STEINMANN (Ammonoidea) des Bathoniums (Brauner Jura E) im Südwestdeutschen Jura. Jahrbuch geologisch. Landesamt Baden-Württemberg, 11:29-86.

LISSAJOUS, M. 1923. Itude sur la faune du Bathonien des environs de Mâcon. Travails de Laboratoire Géologie de l'Université de Lyon, 5, Memoir 3, 286 p.

MANGOLD, C. 1970a. Stratigraphie des étages Bathonien et Callovien du Jura Meridional. Documents du Laboratoire de Géologie, Faculté des Sciences, Lyon, 41 (1):376p.

____ ,1970b. Les Perisphinctidae (Ammonitinia) du Jura Meridional au Bathonien et au Callovien. Documents du Laboratoire de Géologie, Faculté des Sciences, Lyon, 41 (2), 246 p.

____ ,1984. Report of the Bathonian Working Group, p. 67-75. *In* O. Michelsen and A. Zeiss. International Symposium on Jurassic Stratigraphy, Erlangen I. Copenhagen.

____ ,1991. Bathonien, p. 130. *In* Resumés - Abstracts. 3rd International Symposium on Jurassic Stratigraphy, Poitiers.

____ , ELMI, S. AND J. GABILLY. 1974. Les faunes du Bathonien dans la moitie sud de la France. Essai de zonation et correlations. *In* Colloque sur la Jurassique, Luxembourg, 1971, Memoire du Bureau des Recherches géologiques et minières, 75:103-132.

____ , AND M. RIOULT. 1997. Bathonien, p. 55-62. *In* E. Cariou and P. Hantzpergue (eds), Biostratigraphique ouest-européen et méditerranéen. Bulletin du Centre de Recherche, Elf Exploration et Production, Memoire 17.

PAGE, K. N. 1988.The stratigraphy and ammonites of the British Lower Callovian. Unpublished PhD thesis, University College, London.

____ , 1989. A stratigraphical revision for the English Lower Callovian. Proceedings of the Geologists Association, 100:363-382.

____ ,1995. Biohorizons and Zonules: Intra-Subzonal units in Jurassic Ammonite Stratigraphy. Palaeontology, 38:801-814.

____ ,1996. Observations on the succession of ammonite faunas in the Bathonian (Middle Jurassic) of south west England, and their correlation with a Sub-Mediterranean Standard Zonation. Proceedings of the Ussher Society, 9:45-53.

____ , AND G. MELÉNDEZ. 1998. The Upper Bathonian at Aguilón, Northern Iberian Chain, Spain - a potential reference section for Europe, p. 121-123. *In* G. Meléndez and I. Pérez-Urresti, (eds.), IV Congreso de Jurásico de España; el Jurásico de Iberia y de los Cuencas Peritetisianas: Comunicaciones. Institutión 'Fernando el Católico', publication, 1853 (supplement).

POCOCK, J .I. 1926.The Geology of the Country around Oxford. Memoir of the Geological Survey of Great Britain.

SEQUEIROS, L. AND G. MELÉNDEZ. 1979. Nuevos datos bioestratigráficos del Callovienes y Oxfordiense de Aguilón (Cord. Ibérica, Zaragoza). Cuadernos Geológica, 10:167-177.

THIERRY, J., E. CARIOU, D. CONTINI, D. MARCHAND, AND M. RIOULT. 1991. *In* Resumés - Abstracts. 3rd International Symposium on Jurassic Stratigraphy, Poitiers:131.

TORRENS, H. S. 1965. Revised zonal scheme for the Bathonian staged of Europe. Carpatho-Balkan Geological Association, VII congress, Sofia, Reports, Part II, 1:47-55.

____ , 1974. Standard zones of the Bathonian. *In* Colloque sur la Jurassique, Luxembourg, 1971, Memoire du Bureau des Recherches géologiques et minières, 75:581-604

____ , 1980. Bathonian Correlation chart, p. 21-45. *In* J. C. W. Cope, K. L. Duff, C. F. Parsons, H. S. Torrens, W. A. Wimbledon and J. K. Wright. A correlation of the Jurassic rocks in the British Isles. Part two: Middle and Upper Jurassic. Special Report of the Geological Society, London, 15.

WESTERMANN, G. E. G. 1958. Ammoniten-fauna und Stratigraphie des Bathonien NW-Deutschlands. Beihefte zum Geologischen Jahrbuch, 32, 103

____ , AND J. H. CALLOMON. 1988. The Macrocephalintinae and associated early Callovian (Jurassic) ammonoidea of the Sula Islands and New Guinea. Palaeontographica, A203:1-90.

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East Quantoxhead, Somerset: a Candidate Global Stratotype Section and Point for the Base of the Sinemurian Stage (Lower Jurassic)

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Abstract: The base of the Sinemurian Stage is drawn at the base of the Conybeari Subzone of the Bucklandi Chronozone, the type area of which includes Dorset and Somerset, south-west England. Recent study of extensive coastal sections in West Somerset has revealed one of the most expanded fossiliferous sequences across this boundary known in Europe - the Conybeari Subchronozone alone reaching nearly 13m in thickness. Excellent cliff and foreshore exposures in 'Blue Lias' facies Lower Lias north of East Quantoxhead are eminently suitable as a reference section, and here the Hettangian/Sinemurian boundary is drawn at a level 0.75m below the top of British Geological Survey Bed 146. The terminal Hettangian (Depressa Subchronozone, Angulata Chronozone) is indicated by *Schlotheimia* ex gr. *princeps* S. Buckman and *S. pseudomoreana* Spath and the base of the Sinemurian (Conybeari Subchronozone, Bucklandi chronozone) is characterised by abundant crushed *Vermiceras quantoxense* n. sp. Higher levels in the Conybeari Subchronozone yield species of *Metophioceras, Vermiceras, V. (Epammonites),* early *Coroniceras* and schlotheimiids of the *charmassei* group. These exposures have now been subjected to multidisciplinary stratigraphic analysis in order to establish a range of correlation methods, as is recommended by ICS guidelines. The boundary is now also correlatable using calcareous foraminifera and Gamma-ray logs (at least regionally) and potentially using ostracods. Palynological results are presently poor, however, and remagnetisation makes the section unsuitable for magnetostratigraphical correlation. Although primarily established on the basis of the change in ammonite faunas, the combined multidisciplinary results indicate that the Hettangian/Sinemurian boundary at East Quantoxhead is a viable candidate GSSP and the site is here so proposed.

Introduction

The Sinemurian Stage was proposed by Alcide d'Orbigny in 1842, as the lowest division of the Jurassic System and was named after the town of Semur-en-Auxois, in northern Burgundy (eastern France), where canal construction and other excavations had yielded many characteristic fossils, especially ammonites. The establishment of the Hettangian Stage by Renevier (1864) for the basal part of d'Orbigny's Sinemurian, restricted the latter as the second stage of a conventional Jurassic System (Arkell, 1933).

The basic framework of correlative zones for the Sinemurian Stage was originally established by Oppel (1856) and subsequent refinement led to the standard scheme of Donovan in Dean et al. (1961). The base of the stage is drawn at the base of the lowest subchronozone of the lowest zone, namely the Conybeari Subchronozone of the Bucklandi Chronozone, the type area of which is south-west England, including Somerset (Tutcher, 1918). Donovan in Morton (1971) proposed that sections near Lyme Regis, on the Dorset coast (Page, 1995, fig. 1), described by Lang (1924) should be established as a stratotype for the Sinemurian Stage (although Sinemurian-type faunas have now been recorded slightly lower: Palmer, 1972a; 1972b; Page, 1992, p. 136).

Recent study of coastal sections in West Somerset (Fig. 1), described by Palmer (1972a) and Whittaker and Green (1983- including a brief review of the ammonite faunas by Ivimey-Cook and Donovan), has revealed a considerably expanded Hettangian/Sinemurian boundary sequence nearly five times thicker than at Lyme Regis. The sequence is exposed near East Quantoxhead, east of

Watchet, at a site referred to by Palmer as “Limekiln Steps”. In addition it has been possible to demonstrate that the earliest Sinemurian-type fauna at Lyme Regis is preceded by another, earlier fauna near East Quantoxhead (Page, 1992, p. 136). The Dorset section is therefore not considered to be a suitable stratotype for the base of the Sinemurian, whereas the Somerset succession has great potential.

The combination of a good faunal sucession and expanded sequence (ca.13m for the Conybeari Subchronozone alone) makes the locality unusual in Europe. Elsewhere, for instance near Bristol (e.g., Saltford Cutting; Donovan, 1952a; 1952b; 1956; Donovan and Kellaway, 1984), southern Germany (Bloos, 1985 a; 1985b) and south-east France (Corna, 1985; 1987; Elmi and Mouterde, 1965), sequences are usually much thinner and much less complete even at the historical "type" locality for the Stage near Semur-en-Auxois itself (Corna and Mouterde, 1988). Only in Glamorgan (S. Wales) is the subchronozone known to be thicker (Trueman 1922; 1930; Wilson et al., 1990; pers. obs. 1994), but the faunal sequence seems to be poorer, especially in the region of the Hettangian/Sinemurian boundary.

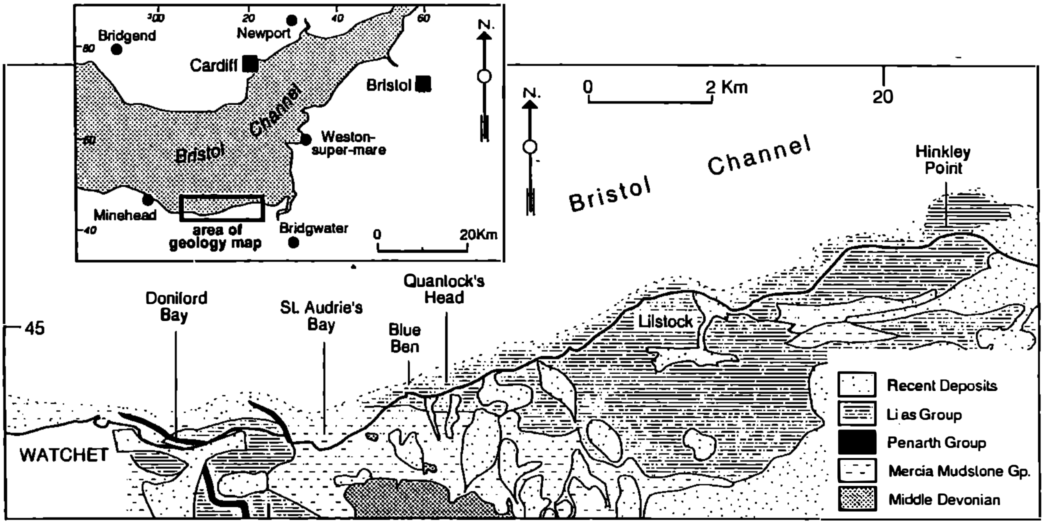

Figure 1: Location map and local geology of the Lower Jurassic (Hettangian-Sinemurian), coastal section at East Quantoxhead (after Warrington and Ivimey-Cook, 1995)

Stratigraphic Characteristics of the Boundary Interval

Ammonite Faunas and Standard Zonations (K.N.P., G.B.)

The faunal sequence in West Somerset is the most complete known in the North West European Province and can form a standard with which to correlate sequences elsewhere (Fig. 2). The sequence of faunas in the Early Sinemurian was characterised as a series of potentially correlatable infra-subzonal units or *biohorizons* (*sensu* Callomon, 1985; Page, 1992) by Page (1995). Further work by Bloos and Page (this volume) has recognised additional faunas in the Early Sinemurian and better characterised the Hettangian sequence below. The change from typically "Hettangian" to typically "Sinemurian" faunas is conventionally indicated by the virtual replacement of faunas dominated by *Schlotheimia* spp. (Schlotheimiidae) of the Angulata Chronozone (Upper Hettangian), by faunas dominated by Arietitinae (including *Vermiceras*, *V. (Epammonites)*, and *Metophioceras*) of the early Bucklandi Chronozone (Lower Sinemurian). The following is a summary of the succession across the boundary (for more details see Bloos and Page, this volume):

Hettangian: Angulata Chronozone, Complanata Subchronozone (>16m): Beds 95-133 (C52-C90) contain a typical fauna of the Complanata Subchronozone of Germany, including

Schlotheimia complanata Koenen, *S. oxygonia, S. tenuis, S. macilenta, S. polyeides, S. angulosa, S. oligoptycha* (all Lange, 1951) and *S.* ex gr. *striatissima* (Quenstedt).

Depressa Subchronozone (4.5m): *S. princeps* (Bayle) is a typical species of the subchronozone in south-west Germany and a similar form is present in Bed 134 (C 91). *S. pseudomoreana* (Spath) appears in Bed 135 and persists up to the top of the Hettangian in Bed 145 (C100, 60cm above limestone Bed 144) (includes the *pseudomoreana* and *Schlotheimia* sp. 2 biohorizons of Page, 1995).

Sinemurian: Bucklandi Chronozone, Conybeari Subchronozone (13m): The base of the Sinemurian is marked by the sudden appearance of *Vermiceras quantoxense* n. sp. (Bloos and Page, this volume), *V. palmeri* n. sp., *Metophioceras* sp. indet. A in Bed 146/145 (C100) with base at a level 0.75m below Bed 147 (*quantoxense* Biohorizon, = cf./aff. *rougemonti* Biohorizon of Page 1992; 1995).

Poorly characterised large *Metophioceras* sp. indet. B and schlotheimiids of the *charmassei* group are present in Bed 147 (C101) (*Metophioceras* sp. 1 Biohorizon of Page 1995). Bed 149 (C103) yields a good fauna including *M. brevidorsale* (Quenstedt), *V. spiratissimum* (Quenstedt), *V. rougemonti* (Reynès), *V. conybearoides* (Reynès), and schlotheimiids of the *charmassei* group (*conybearoides* Biohorizon, = *Metophioceras* sp. 2 Biohorizon of Page, 1995 and *longidomus* Biohorizon of Page, 1992). This fauna is often the lowermost recorded elsewhere in Britain (e.g., from around Bristol) and in Germany and France (Bloos and Page, this volume). Higher faunas in the subchronozone are broadly as characterised by Page (1995), but with the addition of an aff. *quantoxense* Biohorizon (Bed 158/C112), an *elegans* Biohorizon nov. (Bed 160/C114) and the *silvestrei* Biohorizon – the latter already known from Germany (Bed 163/C118) (Bloos and Page, this volume).

Foraminifera (M.H., M.B.H.)

Although planktonic foraminifera probably originated in the Late Triassic, benthic forms predominate during the Lower Jurassic and show very consistent stratigraphic ranges throughout NW Europe (Copestake, 1985; Copestake and Johnson, 1989). Benthic foraminifera are of great value in subdividing and correlating the Lower Jurassic with an obvious importance in situations where ammonites are not recovered.

Sampling: A total of 35 samples of shale and mudstone were taken from the section at East Quantoxhead with an emphasis placed on the boundary shale. Samples labelled EQH were provided by K.N.P. and those labelled QTX were collected by M.B.H. Around 200 grams of material was processed using standard disaggregation techniques and at least 300 individuals from each sample were picked, where possible, to obtain a representative assemblage of the species present (Dennison and Hay, 1967).

Results: The results of a preliminary study of the section at East Quantoxhead are summarized by Hylton (1998) and have now been augmented by additional results providing valuable data on the nature of the foraminiferal fauna across the Hettangian/Sinemurian boundary (Fig. 2). The predominance of calcareous foraminifera and especially the *Lingulina tenera* plexus group is typical of Lower Jurassic sequences in Northern Europe and the United Kingdom. The assemblages also comprise useful index species and other forms diagnostic of the Lower Jurassic and the Hettangian - Sinemurian in particular. The first and last appearances of *Lingulina tenera* plex. *substriata* are restricted to the top of the Hettangian Angulata Chronozone and are therefore useful for correlating the base of the Sinemurian (Copestake and Johnson, 1989). The consistent appearance of *Planularia inaequistriata* and the *Frondicularia terquemi* plexus group above the boundary at East Quantoxhead is a further characteristic to identify the base of the Sinemurian. Copestake and Johnson (1984) also include the important index species *Lingulina tenera* plex. *substriata*, *Planularia inaequistriata* and *Dentalina matutina* in their benthic foraminiferal zonation scheme. This scheme, of international applicability for both southern and northern hemispheres, covers Britain and seventeen other countries and regions throughout the world. The foraminiferal species discussed here are those considered to be the most important for zonation and correlation purposes in the British and Northwest European Lower Jurassic (Copestake and Johnson, 1989). Accordingly, this study has shown that the foraminifera from East Quantoxhead could provide a useful tool for correlation of the proposed GSSP with other sections.

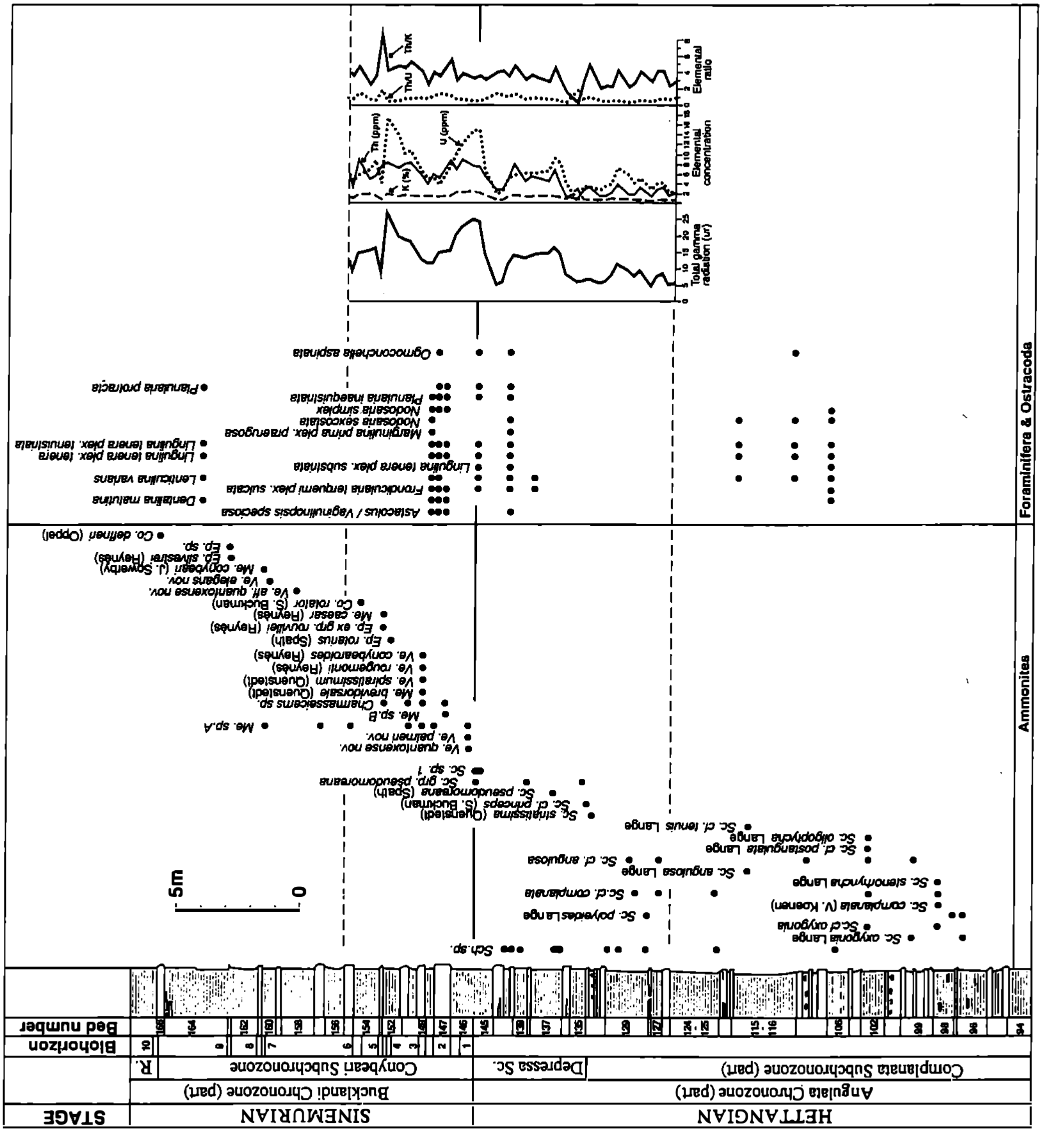

Figure 2. The Hettangian/Sinemurian boundary sequence near Limekiln Steps, East Quantoxhead: Stratigraphical synthesis (bed numbers are those of Whittaker and Green, 1983). Biohorizons numbered as follows: 1- *quantoxense*, 2- *Metophioceras* sp. B, 3- *conybearoides*, 4- *rotarius*, 5- *rouvillei*, 6- *rotator*, 7- *elegans*, 8- *conybeari*, 9- *silvestrei*, 10- *defneri*. (R.= Rotiforme Sunchronozone, *Sc.*= *Schlotheimia*, *Ve.*= *Vermiceras*, *Me.*= *Metophioceras*, *Ep.*= *Epammonites*, *Co.*= *Coroniceras*).

Ostracoda (M.H., M.B.H.)

Sampling: Processing and sample details were as for the foraminiferal study.

Results: Occurrences of ostracod carapaces or valves in the samples were sporadic. Five samples, however, yielded low numbers of specimens (from beds 148, 145, 141 and 114; Fig. 2), dominated by the healdiid genus *Ogmoconchella,* almost entirely as *O. aspinata* (Drexler) which is widespread in the European Hettangian (Lord, 1978). This form ranges from the Hettangian to Lower Sinemurian and is a useful index species for this part of the Lower Jurassic.

Palynology (M.F., M.B.H.)

Samples obtained to date indicate only poorly preserved assemblages (probably largely due to surface weathering), but with continued sampling it is believed that better assemblages will be recovered. No stratigraphic conclusions can therefore be made at this stage.

Magnetostratigraphy (D.E.R., A.M.)

Six beds across the boundary sequence were cored (four in limestone and two in shale). Samples were subjected to progressive alternating field (AF) demagnetisation. Remanences were measured using a JR4 Spinner Magnetometer. IRM acquisition experiments were conducted to determine the magnetic mineralogy.

Results: (i) AF demagnetisation easily removed all the natural remanent magnetism (NRM) of these samples (median destructive field = 10 mT). Clear, single components of magnetisation were isolated using principal component analysis. (ii) IRM experiments showed a rapid increase in isothermal moment with applied fields of up to 150-300 mT. This suggests that the NRM in these samples is carried by magnetite. The low median destructive fields suggest the presence of multi-domain magnetite. (iii) The mean remanence direction is:

In situ:	Dec=003°	Inc=71°	95=11.0°	K=70.4
Tilt corrected:	Dec=010°	Inc=65.6°	95=11.7°	K=62.5

Discussion: The results indicate that the *in situ* direction is in close alignment with the present day axial geocentric magnetic field at the site (Dec=0°, Inc=68°). Primary (Jurassic) remanences are thus expected to have inclinations of around 50°. These data therefore suggest that the section has been remagnetised in the present day direction. This is supported by the slight increase in dispersion after tilt correction (although this is not statistically significant).

Conclusion: The section has been remagnetised, possibly during recent weathering. Primary remanences are not recoverable, even with detailed progressive demagnetisation.

Gamma Ray (J.L.B., S.P.H., H.J.)

A spectral gamma-ray survey of the 'Blue Lias' sequence on the West Somerset coast and other Hettangian-Sinemurian sections in SW England, has recently been published in outline (Bessa and Hesselbo, 1997). Figure 2 shows a detailed survey across the Hettangian/Sinemurian boundary in West Somerset. Sample stations are at 0.3m intervals. Pattern matching of gamma-ray characteristics can be used to correlate outcrop to borehole, to a resolution that varies from about 0.5 to 3m (cf. discussion in Whittaker et al., 1985; Smith, 1989; Bessa and Hesselbo, 1997).

A particular characteristic of the boundary section is the occurrence of two marked peaks in U concentration, one just above the boundary and one some four metres above it. These are linked to the occurrence of dark organic-rich shales, which may themselves be of value when correlating lithostratigraphically within the basin. Correlation with other basins has not yet been attempted but the observations in West Somerset indicate that gamma-ray logs may have some potential for correlation of the Hettangian/Sinemurian boundary, at least regionally.

Satisfaction of ICS GSSP Criteria

The effective communication of the relationships in time and space between different rock bodies and events and processes during the evolution of the Earth requires an unambiguously defined chronostratigraphic nomenclature. *Global Stratotype Sections and Points* or GSSPs unambiguously and objectively define the boundaries between successive divisions of a chronostratigraphic scale at a chosen stratotype locality (Salvador, 1994). Remane et al. (1996) review the ideal requirements of a candidate GSSP, but accept that not all sections are likely to fulfil every requirement. These requirements are listed below with an assessment of the suitability of the Quantoxhead section:

"4.1 Geological Requirements"

(a) "Exposure over an adequate thickness": The large scale of the exposures at East Quantoxhead and the relative stratigraphic completeness (including the great thickness of the trans-boundary sequence) make the section ideal as a candidate GSSP. The boundary interval is exposed and accessible for over 200m along the cliff and foreshore. The section forms part of a continuous stratigraphic sequence from Norian (Upper Triassic) to Semicostatum Chronozone, Lyra Subchronozone (middle Lower Sinemurian).

(b) "Continuous sedimentation": The relative expansion of the boundary sequence when compared to other localities in the region (e.g. 13m for the Conybeari Subzchronozone alone, as opposed to about 5.4m near Lyme Regis on the Devon/Dorset coast), the presence of faunas not known from elsewhere and the lack of any sedimentological indication of condensation or non-sequence near the boundary level, suggests that virtually continuous sedimentation characterises the Quantoxhead succession.

(c) "The rate of sedimentation": The relative thickness of the boundary sequence and the position of the boundary within a major mudrock unit indicate a high rate of sedimentation. The presence of ammonite faunas not known from elsewhere indicates that successive events can be adequately distinguished.

(d) "Absence of synsedimentary and tectonic disturbances": There are no significant synsedimentary disturbances in the boundary sequence. Although the Liassic sequence of the region is incorporated onto a tectonic inversion (Dart et al., 1995), faulting in the area of the proposed GSSP does not significantly disrupt the boundary sequence which lies entirely within a single faulted wedge.

(e) "Absence of metamorphism and strong diagenetic alteration": The area is free from metamorphism and significant diagenetic alteration.

"4.2 Biostratigraphic Requirements"

(f) "Abundance and diversity of well-preserved fossils": Ammonite faunas are abundant at many levels across the boundary sequence. Preservation is often good, with solid specimens in limestones and crushed (but determinable) shells in mudrocks. No other stratigraphically diagnostic macrofossils are present. Microfaunas include foraminifera and ostracods, the former in particular yielding results of taxonomic importance. Palynological information is presently incomplete but may benefit from future sampling of additional levels.

(g) "Absence of vertical facies changes": The boundary interval lies within a thick sequence of alternating mudrocks (bituminous shale, non-bituminous marl) and bedded or nodular limestone. The boundary itself is placed *within* a single bituminous mudrock unit. There is therefore no significant change of litho- or biofacies across the boundary.

(h) "Favourable facies for long-range biostratigraphic correlation": The succession was laid down in an open-marine environment and most of the taxa present are known to have a wide geographical range (for the ammonites, many species are recorded throughout the Northwest European Province and some also in the Tethyan Province).

"4.3 Other Methods"

(i) "Radioisotopic dating": The technique has not been attempted.

(j) "Magnetostratigraphy": The section has been remagnetised and primary remanences have not been recovered, even with detailed progressive demagnetisation. The section is not therefore suitable for magnetostratigraphical correlation.

(k) "Chemostratigraphy": The section offers a good potential for such studies, including strontium isotope analysis.

(l) "Regional paleogeographical context" and "facies relationships": The area lies within the Northwest European Province with free marine connection during the Hettangian and Sinemurian to the rest of Europe and Asia. The widespread occurrence of comparable limestone-mudrock alternations throughout the region suggests that facies variations are unlikely to affect the distribution of faunas and floras of correlative importance.

"4.4 Other Requirements"

(m) "Permanently fixed marker": A marker could be placed, with landowner's permission, adjacent to the exposure at the top of the cliff. Any marking of the cliff itself is likely to be temporary due to active coastal erosion (unless attached directly to Limekiln Steps).

(n) "Accessibility": The site is freely accessible along public rights of way and on open coastal foreshore. Steps down to the beach at grid ref: ST137443 ("Limekiln Steps") descend over the boundary interval - these steps are linked to car parking at Kilve and East Quantoxhead, both around 900m from the proposed GSSP. The site is also accessible along the beach from Kilve when tides are low. Visiting on a falling tide is recommended.

(o) "Free access": Public access along rights of way (footpaths) and along the foreshore is guaranteed under English law.

(p) "Guarantees from the respective authority concerning free access for research and permanent protection of the site": The site is protected under national conservation law (Wildlife and Countryside Act 1981) and lies within the legally designated Blue Anchor to Lilstock Coast Site of Special Scientific Interest (Page et al., 1994). Advice concerning protocols for sampling programs and other enquiries should be directed to English Nature at Roughmoor, Bishops Hull, Taunton, Somerset, TA1 5AA, U.K.

Conclusions

Results to date indicate that the section at East Quantoxhead demonstrates the most complete known Hettangian/Sinemurian boundary succession. The sequence of ammonite faunas is unrivalled and therefore has great potential for detailed international correlations. Foraminifera also facilitate correlation of the boundary and ostracods, although presently incompletely known, may ultimately yield useful results. Gamma ray log correlation is possible, at least regionally, whereas for magnetostratigraphic results, an additional supporting section may be required. When combined with the excellent quality of the exposures and their high level of permanence and protection, the cliffs and foreshore become an ideal candidate Global Stratotype Section and Point for the base of Sinemurian Stage of the Jurassic System.

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References

ARKELL, W. J. 1933. The Jurassic System in Great Britain. Clarendon Press, Oxford, 681 p.

BESSA, J. L. AND S. P. HESSELBO. 1997. Gamma-ray character and correlation of the Lower Lias, SW Britain. Proceedings of the Geologists' Association, London, 108:113-129.

BLOOS, G. 1983. The zone of *Schlotheimia marmorea* (Lower Lias) - Hettangian or Sinemurian? Newsletters on Stratigraphy, 12:123-131.

____. 1985a. On Lower Lias ammonite stratigraphy - present state and possibilities of revision. *In* O. Michelsen and A. Zeiss (eds.), International Symposium on Jurassic Stratigraphy, Erlangen 1984, 1:146-157.

____. 1985b. Les couches basales du Sinémurien - une révision stratigraphique. Cahiers de l'Institute Catholique de Lyon, 14:59-68.

____ AND K. N. PAGE. [this vol.]: The proposed GSSP for the base of the Sinemurian Stage near East Quantoxhead/West Somerset (SW England) – the ammonite sequence.

CALLOMON J. H. 1985. Biostratigraphy, chronostratigraphy and all that – again! p. 611-624. *In* O. Michelsen and A. Zeiss (eds.), International Symposium on Jurassic Stratigraphy, Erlangen 1 1984, 3.

COPESTAKE, P. 1985. Foraminiferal biostratigraphy in the Lower Jurassic, p. 191-206. *In* O. Michelsen and A. Zeiss (eds.), International Symposium on Jurassic Stratigraphy, Erlangen 1984, 1.

____ AND B. JOHNSON. 1984. Lower Jurassic (Hettangian - Toarcian) Foraminifera from the Mochras Borehole, North Wales (UK) and their Application to a world-wide biozonation, p. 183-184. *In* Benthos '83; 2nd International Symposium on Benthic Foraminifera, Pau, April 1983.

____ AND ____. 1989. The Hettangian to Toarcian (Lower Jurassic), p. 129-272. *In* D. G. Jenkins and J. W. Murray (eds.), Stratigraphical Atlas of Fossil Foraminifera, British Micropalaeontological Society Series.

CORNA, M. 1985. Le lias du Jura méridional, paléontologie biostratigraphique du Sinémurien: approche paléoécologique. Thèse 3ème Cycle, Université Claude Bernard, Lyon, 268 pp. (unpublished).

____. 1987. Les horizons Sinémuriens du Calcaire à Gryphées du Jura Méridional Français (Zone à Conybeari - Zone à Obtusum). Geobios, 120:531-536.

____ AND R. MOUTERDE. 1988. Le Sinémurien de Semur (France): essai de biozonation pour le Sinémurien inférieur et le Lotharingien inférieur du stratotype et du Jura méridional, p. 101-107. *In* R. B. Rocha and A. F. Soares (eds.), 2nd International Symposium on Jurassic Stratigraphy, Lisboa 1987, 1.

DART, C. J., K. McCLAY, AND P. N. HOLLINGS. 1995. 3 D analysis of inverted extensional fault systems, southern Bristol Channel basin, UK, p. 393-413. *In* J. G. Buchanan and P. Buchanan (eds.), Basin Inversion, Geological Society Special Publication 88.

DEAN, W.T., D. T. DONOVAN, AND M. K. HOWARTH. 1961. The liassic ammonite zones and subzones of the North-West European Province. Bulletin of the British Museum of Natural History, Geology Series, 4:435-505.

DENNISON, J. M. AND W. W. HAY. 1967. Estimating the needed sampling area for subaquatic ecologic studies. Journal of Palaeontology, 41:706-708.

DONOVAN, D. T. 1952a. The Ammonites of the Blue Lias of the Bristol district. I. Psiloceratidae. Annals and Magazine of Natural History, London, 5:629-655.

____. 1952b. The Ammonites of the Blue Lias of the Bristol district. II. Arietitidae. Annals and Magazine of Natural History, London, 5:717-752.

____. 1956. The Zonal Stratigraphy of the Blue Lias around Keynsham, Somerset. Proceedings of the Geologists' Association, London, 66:182-212.

____ AND G. KELLAWAY. 1984. Geology of the Bristol district: the Lower Jurassic rocks. Memoirs of the Geological Survey, G.B., 69 p.

ELMI, S. AND R. MOUTERDE. 1965. Le Lias inférieur et moyen entre Aubenas et Privas (Ardèche). Travaux du Laboratoire de Géologie du Faculté des Sciences, Lyon. Nouveau Série, 12:143-246.

HYLTON, M. 1998. A preliminary analysis of the foraminifera from the Hettangian-Sinemurian boundary of East Quantoxhead, West Somerset. Geoscience in south-west England, 9:203-208.

IVIMEY-COOK, H. C. AND D. T. DONOVAN. 1983. Appendix 3: The fauna of the Lower Jurassic, p. 126-130. [*In*: A. WHITTAKER G. AND W. GREEN. 1983]

LANG, W. D. 1924. The Blue Lias of the Devon and Dorset coasts. Proceedings of the Geologists' Association, London, 35:169-185.

LANGE, W. 1951. Die Schlotheimiinae aus dem Lias alpha Norddeutschlands. Palaeontographica A, 100:1-128.

LORD, A. R. 1978. The Jurassic Part 1 (Hettangian - Toarcian). *In* R. H. Bate and E. Robinson (eds.), A Stratigraphical Index of British Ostracoda. British Micropalaeontological Society, Seal House Press, Liverpool.

MORTON, N. 1971. The definition of standard Jurassic Stages. Mémoire du Bureau de Recherches géologiques et minières, 75:83-93.

OPPEL, A. 1856-1858. Die Juraformation Englands, Frankreichs and des südwestlichen Deutschlands. Jahreshefte des Vereins für vaterländische Naturkunde in Württemberg, 12:121-556; 13:141-396; 14:129-291.

ORBIGNY, A. D'. 1842-1849. Paléontologie française; Terrains jurassiques. Paris.

PAGE, K. N. 1992. The sequence of ammonite-correlated horizons in the British Sinemurian (Lower Jurassic). Newsletters on Stratigraphy, 27:129-156.

____. 1995. East Quantoxhead, Somerset, England; a potential Global Stratotype Section and Point (GSSP) for the base of the Sinemurian Stage (Lower Jurassic). Proceedings of the Ussher Society, 9:446-450.

___, A. H. KING, AND D. D. GILBERTSON. 1994. Field excursion to examine the Triassic-Jurassic transition in West Somerset and the Quarternary deposits of Doniford Bay, Watchet. Proceedings of the Ussher Society, 8:338-344.

PALMER, C. P. 1972a. The Lower Lias (Lower Jurassic) between Watchet and Lilstock in north Somerset (United Kingdom). Newsletters on Stratigraphy, 2:1-30.

___. 1972b. A revision of the zonal classification of the Lower Lias of the Dorset coast of South-West England. Newsletters on Stratigraphy, 2:45-54.

REMANE, J., M. G. BASSETT, J. W. COWIE, K. H. GOHRBANDT, R. H. LANE, O. MICHELSEN, AND W. NAIWEN. 1996. Revised guidelines for the establishment of global chronostratigraphic standards by the International Commission on Stratigraphy (ICS). Episodes 19/3.

RENEVIER, E. 1864. Notices géologiques et paléontologiques sur les Alpes Vaudoises, et les régions environ Nantes. I. Infralias et zone à *Avicula contorta* (Et. Rhétien) des Alpes Vaudoises. Bulletin de la Société Vaudoise des Sciences Naturelles, 13:218-252.

SALVADOR, A. L. (ed.). 1994. International Stratigraphic Guide (2nd edition). Geological Society of America.

SMITH, D. G. 1989. Stratigraphic correlation of presumed Milankovitch cycles in the Blue Lias (Hettangian to earliest Sinemurian), England. Terra Nova, 1:457-460.

TRUEMAN, A. E. 1922. The Liassic rocks of Glamorgan. Proceedings of the Geologists' Association London, 33:245-284.

___. 1930. The Lower Lias (Bucklandi Zone) of Nash Point, Glamorgan. Proceedings of the Geologists' Association, London, 41:148-159.

TUTCHER, J. W. 1918. The zonal sequence in the Lower Lias (lower part). Appendix I to S.S. Buckman, Jurassic Chronology: I. Lias. Quarterly Journal of the Geological Society, London, 73:278-281.

WARRINGTON, G. AND H. C. IVIMEY-COOK. 1995. The Late Triassic and Early Jurassic of coastal sections in West Somerset and South and Mid-Glamorgan, p. 9-30. *In* P. D. Taylor (ed.), Field Geology of the British Jurassic, Geological Society, London.

WHITTAKER, A. AND G. W. GREEN. 1983. Geology of the country around Weston-super-Mare. Memoirs of the Geological Survey of Great Britain, 147 pp.

___, D. W. HOLLIDAY AND I. E. PENN. 1985. Geophysical Logs in British Stratigraphy. Special Report, Geological Society, London, 18, 72 pp.

WILSON, D., J. R. DAVIES, C. J. N. FLETCHER AND M. SMITH. 1990. Geology of the South Wales Coalfield, Part VI. The country around Bridgend (2nd edition). Memoirs of the British Geological Survey.

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A New Marine Triassic-Jurassic Boundary Section in Hungary

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Keywords: Triassic, Rhaetian, Hettangian, System Boundary, Ammonoids, Radiolarians, Biostratigraphy

Abstract: Reasonably continuous and fossiliferous marine sections across the Triassic–Jurassic boundary are rare; less than ten are known world-wide, and this scarce record has hampered the correlation, selection of a GSSP, and our understanding of biological and geological events at the system boundary. We report preliminary results from a newly recognized boundary section near Csővár, Hungary. There, the Csővár Limestone Formation, a carbonate unit several hundred metres thick, was considered entirely Upper Triassic. However, discoveries of Rhaetian ammonoids (*Choristoceras*) and conodonts (e.g., *Misikella*) in a quarry suggested the presence of uppermost Triassic beds well below the top of the formation, and Hettangian radiolarians from the upper part of the succession indicated that it extends into the Lower Jurassic. We located a nearly 60m thick section in intermittent natural outcrops that contains the Triassic–Jurassic System boundary. Excavation produced a continuous exposure of the boundary interval and permitted macro- and microfossil collection. Latest Triassic ammonoids found in place include *Choristoceras* spp.; these are separated by nearly 20m of strata which have not yet yielded macrofossils, from the lowest Jurassic ammonoids which comprise phylloceratids and psiloceratids with, higher up, *Waehneroceras* indicating the presence of Lower and Middle Hettangian, respectively. Radiolarians are poorly preserved in this part of the section but stratigraphically important Hettangian assemblages were recovered from above the measured section, where chert lenses become more common.

Introduction

Much stratigraphic research effort has been focussed recently on the Triassic–Jurassic System boundary. A primary goal is the designation of a Global Stratotype Section and Point (GSSP) and four GSSP proposals have been made (Kunga Island, Canada; Utcubamba Valley, Peru; St. Audrey's Bay, England; and New York Canyon, USA). Worldwide, there are few other sections which preserve a continuous, fossiliferous marine record of the transition (e.g., Kendelbachgraben, Austria; Sierra Aspera, Chile and others in the proximity to the candidate GSSPs listed above). We report the discovery of an easily accessible section at Csővár (Hungary), which offers opportunities for correlation and a multifaceted study of the system boundary.

Mesozoic rocks were first reported from the vicinity of Csővár by Szabó (1860) who tentatively suggested a Liassic age for "brown shales" in the area. A Jurassic age was confirmed in several reports in the late 1800's but Vadász (1910) assigned the entire Mesozoic package to the Upper Triassic. A Carnian age for the well-bedded brown limestone, based on its alleged correlation with the "Raibl beds" in the Alps, became firmly entrenched in the Hungarian geological literature for several decades. A revision of the age of this unit was first suggested by Kozur and Mostler (1973) who found Late Norian microfossils in the Pokol Valley quarry (Fig. 1). From the same locality, Detre et al. (1988) reported the first finding of an ammonoid (*Choristoceras*) confirming the Late Norian age. Other conodonts (e.g., *Misikella* spp.) and ammonoids (e.g., *Choristoceras* cf. *marshi*) from the highest beds exposed in the quarry indicated a Rhaetian age (Kozur and Mock, 1991; Haas et al., 1997). On Várhegy (Castle Hill), less than 500m from the quarry, Kozur (1993) found Early Jurassic radiolarians which suggested that the formation ranges up into the Hettangian or Sinemurian. It therefore seemed worthwhile to search for the Triassic–Jurassic boundary on Várhegy.

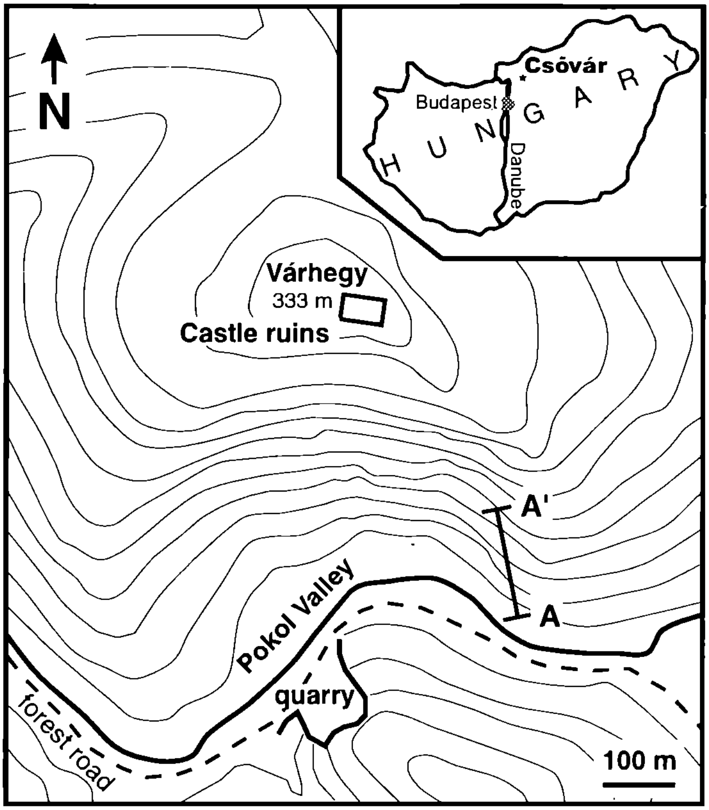

Figure 1. Location of the section at Csővár (A-A': excavated Triassic-Jurassic boundary section). Contour interval is 10m.

Description of the Section

The village of Csővár is located some 50km northeast of Budapest. The south-facing slope of Várhegy (Castle Hill; Fig. 1) is underlain by the Csővár Limestone Formation. Intermittent natural outcrops on the north side of the Pokol Valley have been connected by a newly excavated trench to produce a continuous exposure 0.8km west of Csővár. At present more than 57m of strata are exposed and 124 beds distinguished; some of these are discrete beds, but others are groupings of thinly bedded or laminated strata. The uppermost part of the Csővár Limestone Formation, between the top of the studied interval and the top of the hill (at the the ruins of a medieval castle), is at least 60m thick. The lower, entirely Triassic, part of the formation is best known from the Pokol Valley quarry, located on the other side of the valley, some 150m from the base of our excavated section (Fig. 1). There is strong similarity between the lithologies exposed at the top of the quarry and at the base of the studied section but a precise correlation has not been made.

The Triassic part of the Csővár Limestone Formation, exposed in the Pokol Valley quarry, was described in detail by Haas et al. (1997), who recognized channelized calciturbidites, debris flows and slump structures as well as intervals of laminated, organic-rich mudstone. A toe-of-slope facies, representing the transition from a carbonate platform to a restricted basin was inferred.

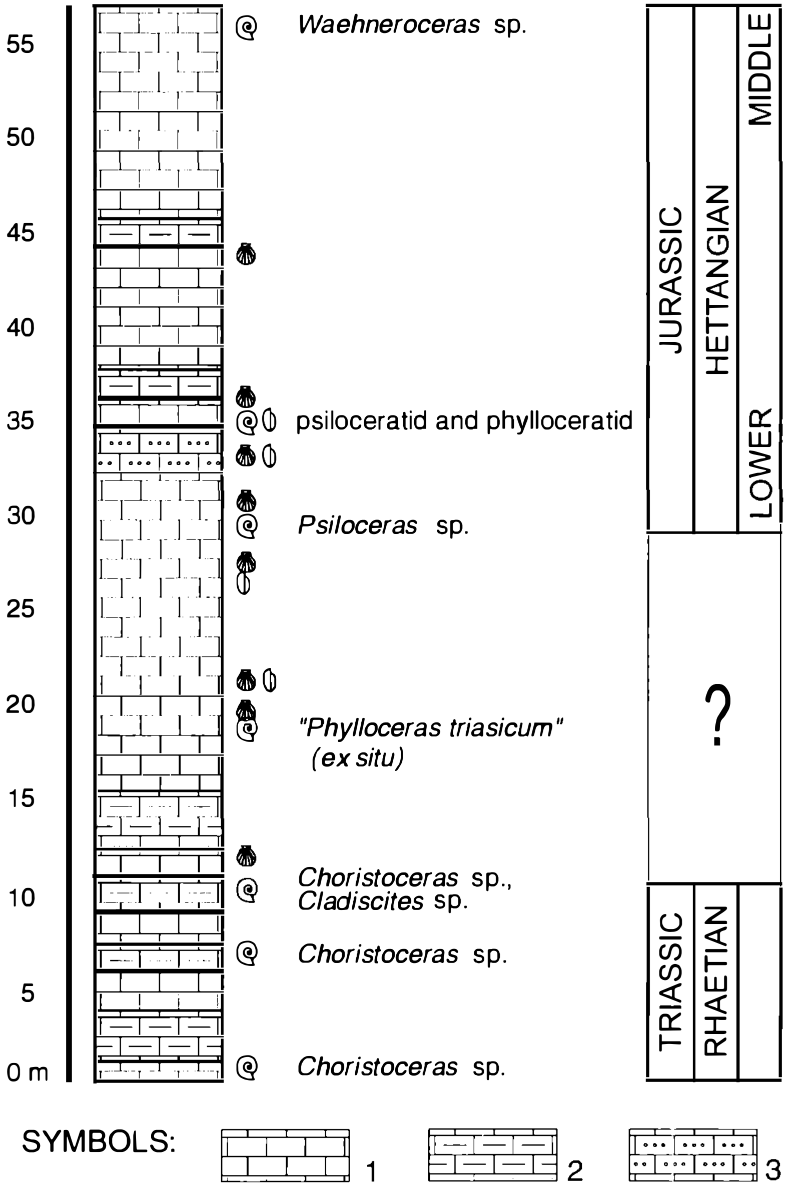

Figure 2. Simplified lithology and biostratigraphy of the Csővár Limestone Formation in the measured stratigraphic section on the south slope of Várhegy at Csővár. Key to lithologic symbols: **1-** Well-bedded limestone with slumps; **2-** Laminated calcareous marl alternating with thin-bedded limestone; **3-** Ooidal limestone.

The lithological features of the section reported here (Fig. 2) are similar to those observed in the Pokol Valley quarry. The lower 11m comprise alternations of laminated calcareous mudstones and thin- to medium-bedded limestones (Fig. 3). Above a nearly one-metre-thick bioclastic limestone bed, laminated calcareous mudstone predominates in the next 3m. Conspicuous slumps are preserved higher upsection for nearly 20m (Fig. 4). Between 32 and 34m, ooidal limestone with abundant brachiopods and bivalves was likely derived from platform environments. The upper half of the measured section is characterized by alternations of laminated calcareous mudstones, bedded limestones and limestones with slump structures. The sedimentological features indicate that deposition was controlled by the interplay of turbidity currents carrying platform-derived carbonate sediments, and a background sedimentation of calcareous mud with a high organic content, perhaps indicating oxygen-depleted bottom conditions. During the time represented by the studied section, the site remained close to the interface of slope, toe-of-slope and basinal depositional environments.

Above the measured section, chert nodules, lenses and layers become increasingly common. The upper part of the formation consists of well-bedded cherty limestone. This resistant, cliff-forming unit is exposed near the top of the hill and is interpreted as a more basinal facies.

Figure 3. Rhaetian (Upper Triassic) laminated limestone, approximately 10m above the base of the Várhegy section.

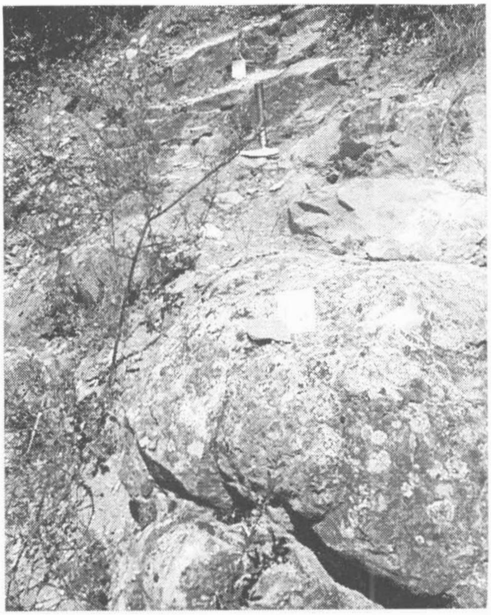

Figure 4. Photograph of a prominent sedimentary slump, interbedded with regularly layered and locally laminated strata in the Triassic-Jurassic transition, approximately 25m above the base of the Várhegy section.

Ammonoid and Radiolarian Faunas

Triassic ammonoids were previously known from the Pokol Valley quarry. Detre et al. (1988) reported *Choristoceras nobile*, and Haas et al. (1997) listed and figured *Choristoceras* cf. *marshi* and *Vandaites stuerzenbaumi*. The first was interpreted as Upper Norian (Sevatian), whereas the others suggested the presence of Rhaetian, indicating the presence of both ammonite zones of the highest Triassic Stage. Although no direct, bed-to-bed correlation has been established between the Pokol Valley quarry and the southern slope of Várhegy, the lithological similarity and the age constraints suggest a broad correlation.

Ammonites occur only sporadically throughout the measured section (Fig. 2). Preservation is commonly poor with the marly, laminated beds containing only flattened specimens; three-dimensional, but often fragmentary, internal moulds are found in the less thinly-bedded strata. Several levels in the lowest 10m yielded specimens of *Choristoceras* spp. (Fig. 5.1, 5.4). A whorl fragment found together with this fauna is tentatively identified as *Cladiscites* sp. (Fig. 5.2). This interval is assigned to the Rhaetian; a more precise age determination awaits the specific identification of the *Choristoceras*.

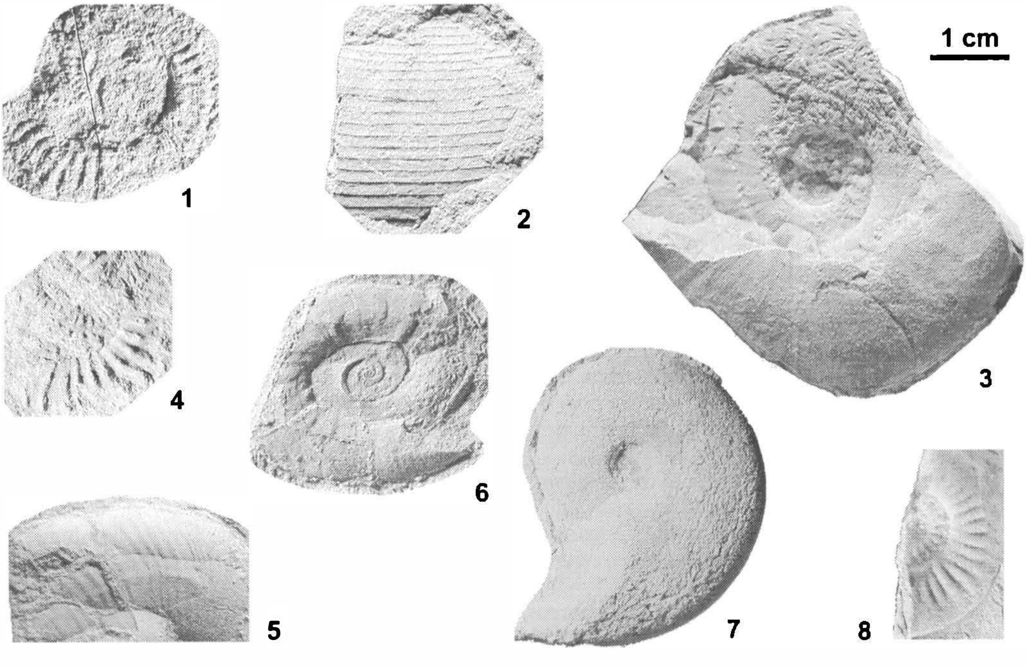

Figure 5. Diagnostic ammonites from the Upper Triassic and Lower Jurassic of the Várhegy section. (Collection levels are given in brackets, in metres above base of section. Specimens are housed in the Department of Paleontology, Hungarian Natural History Museum, under the inventory numbers with prefix M.99.) **1.** *Choristoceras* sp., (6.5 m), Rhaetian, Upper Triassic, M.99.31; **2.** *Cladiscites* sp., (10 m), Rhaetian, Upper Triassic, M.99.32; **3.** *Euphyllites*? sp., *ex situ*, Hettangian, Lower Jurassic, M.99.33; **4.** *Choristoceras* sp., (6.5 m), Rhaetian, Upper Triassic, M.99.34; **5.** *Psiloceras* sp., fragment of body chamber, (29.5 m), Hettangian, Lower Jurassic, M.99.35; **6.** *Pleuroacanthites*? sp., *ex situ*, Hettangian, Lower Jurassic, M.99.36; **7.** "*Phylloceras triasicum*" Vadász, *ex situ*, Hettangian?, Lower Jurassic?, M.99.37; **8.** *Waehneroceras* sp., (56.5 m), Hettangian, Lower Jurassic, M.99.38.

The first ammonoid of Jurassic affinity was found in the scree, 19m from the base of the measured section (Fig. 5.7). The local topography and the lithology of the adhering rock matrix suggest an origin from only a few metres above the level where the specimen was found. The specimen is a phylloceratid closely resembling *Phylloceras triasicum* vadász, although about three times larger in diameter than the holotype, which was described from the vicinity of Csővár by Vadász (1910). We provisionally apply this rarely used species name, although the taxon is probably a junior synonym of another phylloceratid; a comparison can be made to *Phylloceras psilomorphum* Neumayr but further taxonomic studies are needed to establish whether synonymy exists. The lowest Jurassic ammonoid found in place is a finely ribbed body chamber fragment of an evolute form tentatively identified as *Psiloceras* sp. (Fig. 5.5). A comparison to *Psiloceras pacificum* (see Guex 1995), based on the style of ribbing, would imply correlation with the standard

Planorbis Zone. A loose specimen found some 5m higher possesses parabolic nodes and is compared to *Pleuroacanthites* (Fig. 5.6), which ranges from the Planorbis Zone to the Liasicus Zone. Ammonoids found in place at the same level include poorly preserved psiloceratids and phylloceratids that cannot be identified more specifically. A bioclastic limestone bed at the top of the measured section yielded *Waehneroceras* sp. (Fig. 5.8), which indicates the presence of the middle Hettangian. Several intriguing ammonoid specimens, including *Euphyllites*? sp. (Fig. 5.3), *Caloceras*? sp. and *Fergusonites* sp., were found in the talus. These provide further evidence for the presence of the Lower and Middle Hettangian and indicate the potential for further systematic collecting.

Radiolarians were observed in thin sections from the studied section. In the 30 samples processed so far, the tests were secondarily replaced by calcite and could not be separated by acid dissolution. A more abundant and better preserved fauna was reported by Kozur (1993) from the cherty limestone on the upper slope of Várhegy. Two of our spot collections, made several tens of metres above the measured section, yielded identifiable radiolarians. Several of the genera and some species are common to the fauna documented by Carter et al. (1998) from the Lower Jurassic of the Queen Charlotte Islands. A comparison with the integrated radiolarian and ammonoid biostratigraphy from western Canada, which is currently the best studied region in this regard, allows a more accurate radiolarian dating of the Csővár section. Based on the overlap of taxon ranges from the upper slopes of Várhegy, the age of the cherty limestone could range from the early Late Hettangian to the Sinemurian.

Other macrofauna includes, in order of decreasing abundance, bivalves, brachiopods and gastropods. The major faunal turnover caused by the end-Triassic mass extinction allows a gross distinction of Triassic and Jurassic strata based on their bivalve and brachiopod assemblages. Rhaetian beds near the base of the section contain coleoid hooks. Algal filaments, commonly several millimetres in length, are abundant on some bedding planes, and plant macrofossils also occur.

The Triassic–Jurassic Boundary

The Triassic–Jurassic boundary has long had a working definition based on ammonoid biostratigraphy; the first appearance of *Psiloceras* is taken to mark the beginning of the Jurassic. *Psiloceras* is rarely found in succession with the zonal index ammonite of the topmost Triassic, *Choristoceras* (Hallam, 1990) but, if our tentative identifications are correct, this occurs in the Csővár section. Undoubted *Choristoceras* occur at three levels in the basal 10m of the section, and *Psiloceras* cf. *pacificum*? was recovered nearly 30m above the base of the section; this is the lowest Hettangian ammonoid found in place so far. Therefore the Triassic–Jurassic boundary is constrained to a 19m interval that has not yet yielded ammonoids, except for an *ex situ* phylloceratid of probable Jurassic affinity. Further collecting may refine the position of the Triassic–Jurassic boundary in the section.

Discussion

The Csővár section is a significant addition to the small number of marine Triassic–Jurassic boundary sections. Its major advantages include easy access and good exposure, which is partially man-made but can be easily maintained. No obvious hiatus or significant change in lithology has been observed and the section apparently provides a continuous record of sedimentation. The inferred slope-to-basin transitional depositional environment offers potential for correlating bioevents on the platform and in the basin. However, the paucity of ammonoids prevents the section from being proposed as a candidate GSSP for the base of the Jurassic.

This report is based on the first phase of an integrated stratigraphic study. Apart from the work reported here, a detailed sedimentological study by J. Haas is in progress, and magnetostratigraphic investigations have been started by E. Szalay-Márton. Further biostratigraphic work is planned in the near future, including more detailed macrofossil collecting (ammonoids, bivalves and brachiopods), additional sampling for radiolarians, and sampling for conodonts in the lower part of the section. Further work is expected to lead to a refinement of the position of the Triassic–Jurassic boundary in the section. Establishing a stable isotope profile across the boundary will also be attempted.

Apart from its biostratigraphic significance, the section is expected to provide important information bearing on the paleogeographic and tectonic problems of the Mesozoic in Hungary. The study of resedimented, platform-derived components of the biodetrital limestone may shed light on the demise and re-establishment of carbonate platforms and platform-dwelling organisms at the

Triassic–Jurassic boundary. The faunal data will also be interpreted in the context of the end-Triassic mass extinction and subsequent Early Jurassic recovery.

Acknowledgments

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References

CARTER, E. S., P. A. WHALEN, AND J. GUEX. 1998. Biochronology and paleontology of Lower Jurassic (Hettangian and Sinemurian) radiolarians, Queen Charlotte Islands, British Columbia. Geological Survey of Canada Bulletin, 496:1-162.

DETRE, Cs., L. DOSZTÁLY, AND V. HERMANN. 1988. A csővári felső-nóri, sevati fauna (The Upper Norian (Sevatian) fauna of Csővár). A Magyar Állami Földtani Intézet Évi Jelentése az 1986. Évről (Annual Report of the Hungarian Geological Institute, 1986): 53-67.

GUEX, J. 1995. Ammonites hettangiennes de la Gabbs Valley Range (Nevada, USA), Mémoires de Géologie (Lausanne), 27, 131 p.

HAAS, J., E. TARDI-FILÁCZ, A. ORAVECZ-SCHEFFER, F. GÓCZÁN, AND L. DOSZTÁLY. 1997. Stratigraphy and sedimentology of an Upper Triassic toe-of-slope and basin succession at Csővár, Hungary. Acta Geologica Hungarica, 40(2):111-177.

HALLAM, A. 1990. The end-Triassic mass extinction event. *In* V.L. Sharpton and P.D. Ward (eds.), Global catastrophes in Earth history; An interdisciplinary conference on impacts, volcanism, and mass mortality. Geological Society of America Special Paper, 247:577-583.

KOZUR, H. 1993. First evidence of Liassic in the vicinity of Csővár (Hungary), and its paleogeographic and paleotectonic significance. Jahrbuch der Geologischen Bundesanstalt, 136(1):89-98.

____, AND R. MOCK. 1991. New Middle Carnian and Rhaetian conodonts from Hungary and the Alps. Stratigraphic importance and tectonic implications for the Buda Mountains and adjacent areas. Jahrbuch der Geologischen Bundesanstalt, 134(2):271-297.

____, AND H. MOSTLER. 1973. Mikrofaunistische Unterschuchungen der Triasschollen im Raume Csővár, Ungarn. Verhandlungen der Geologischen Bundesanstalt, 1973(2):291-325.

SZABÓ, J. 1860. Geologische Detailkonte des Grenzgebietes des Nograder und Pesther Comitates. Jahrbuch des kaiserlich-königlichen geologischen Reichanstalt, Sitzungberichte (Sitzung am 10. Jänner 1860): 41-44.

VADÁSZ, E. 1910. A Duna-balparti idősebb rögök őslénytani és földtani viszonyai [Paleontology and geology of older blocks on the left side of Danube]. Magyar Királyi Földtani Intézet Évkönyve [Annals of the Royal Hungarian Geological Institute], 18(2):101-171.

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A Revised Numeric Time Scale for the Jurassic

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Keywords: Geochronology, Time Scale

Introduction

Published time scales provide discrepant age estimates for Jurassic Stage boundaries and carry large uncertainties (Pálfy, 1995). The U–Pb or 40Ar–39Ar dating of volcanic and volcaniclastic rocks of precisely known biochronologic age is the preferred method for improving calibration. Here we summarize the results of a Jurassic time scale calibration project which is described in detail elsewhere (Pálfy et al., in press, and references therein).

Calibration of the Jurassic Time Scale

After a systematic effort to obtain new calibration points for the Jurassic numeric time scale, a radiometric age database consisting of more than fifty U–Pb and ^{40}Ar–^{39}Ar ages was compiled. Accepted ages have a precision of ±5 Ma (2σ) or better and every effort was made to constrain each isotopic date at the ammonite zonal level. In some instances, however, only stage level constraints were possible. Many of these calibration points result from integrated bio- and geochronologic dating in the western North American Cordillera and have not previously been used in time scales. Isotopic dates from immediately below or above a stratigraphic boundary, which are considered reliable direct estimates of the boundary age in question, are available only for the Triassic-Jurassic boundary, and the initial boundary of the Crassicosta Zone (Middle Toarcian) and the Callovian stage. The chronogram method (Harland et al., 1990) was used to estimate all other Early and early Middle Jurassic zone boundaries, late Middle Jurassic substage boundaries, and Late Jurassic Stage boundaries. Six consecutive zone boundaries were determined for the Pliensbachian and Toarcian. The derived zonal durations are disparate, suggesting that the assumption of equal zonal duration used for interpolation in several previous time scales, is unfounded. The latest Jurassic isotopic database remains too sparse; therefore, chronogram estimates are improved using interpolation based on magnetochronology (Channel et al., 1995).

Results

The initial boundaries of Jurassic stages are proposed as follows (Fig. 1): Hettangian (Triassic–Jurassic): 199.6 ± 0.3 Ma; Sinemurian: 197.0 + 1.2/– 4.2 Ma; Pliensbachian: 192.0 + 3.8/– 5.2 Ma; Toarcian: 183.6+ 1.6/– 1.1 Ma; Aalenian: 178.0 + 1.0/– 1.5 Ma; Bajocian: 174.0 + 1.0/– 7.3 Ma; Bathonian: 166.0 + 0.6/– 5.4 Ma; Callovian: 159.7 ± 1.1 Ma; Oxfordian: 156.6 + 2.0/– 2.7 Ma; Kimmeridgian: 154.7 + 1.0/– 0.9 Ma; Tithonian: 151.5 + 1.0/– 1.4 Ma; Berriasian (Jurassic–Cretaceous): 144.8 + 2.6/– 3.7 Ma.

References

CHANNELL, J. E. T., E. ERBA, M. NAKANISHI, AND K. TAMAKI. 1995. Late Jurassic-Early Cretaceous time scales and oceanic magnetic anomaly block models. SEPM Special Publication, 54:51-63.

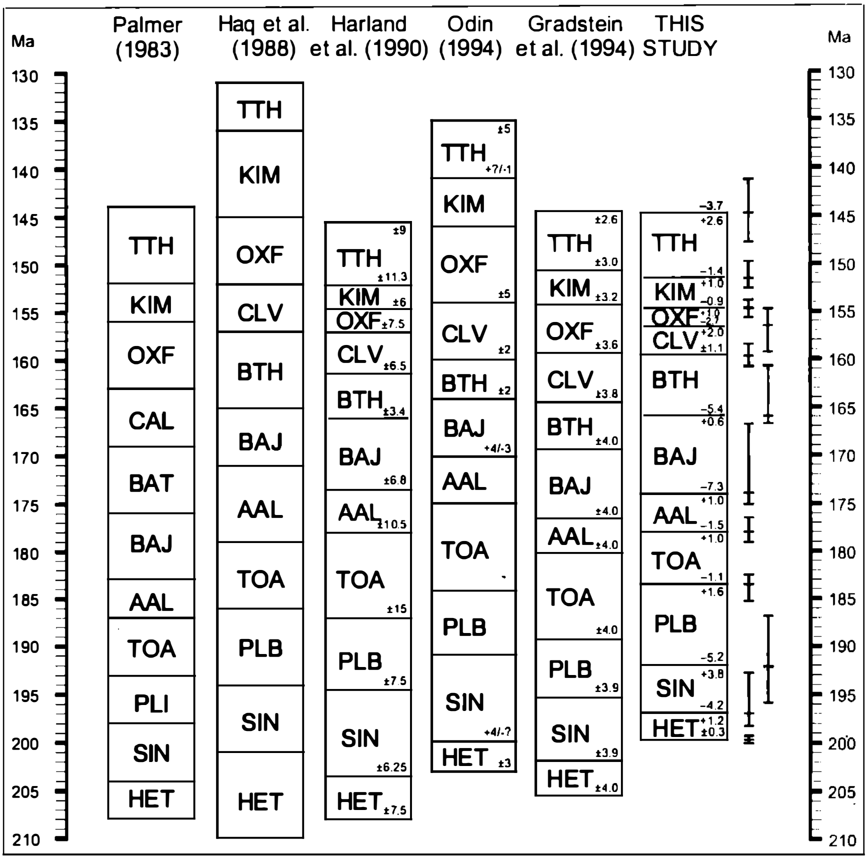

Figure 1. Comparison of the revised Jurassic time scale with other widely used time scales. Stage abbreviations follow those in Harland et al. (1990). Error bars for stage boundaries in the new scale are shown on the right.

References

CHANNELL, J. E. T., E. ERBA, M. NAKANISHI, AND K. TAMAKI. 1995. Late Jurassic-Early Cretaceous time scales and oceanic magnetic anomaly block models. SEPM Special Publication, 54:51-63.

GRADSTEIN, F. M., F. P. AGTERBERG, J. G. OGG, J. HARDENBOL, P. VAN VEEN, J. THIERRY, AND Z. HUANG. 1994. A Mesozoic time scale. Journal of Geophysical Research, B 99:24051-24074.

HAQ, B. U., J. HARDENBOL, AND P. R. VAIL. 1988. Mesozoic and Cenozoic chronostratigraphy and cycles of sea level change. Society of Economic Paleontologists and Mineralogists, Special Publication, 42:71-108.

HARLAND, W. B., R. L. ARMSTRONG, A. V. COX, L. E. CRAIG, A. G. SMITH, AND D. G. SMITH. 1990. A Geologic Time Scale 1989. Cambridge University Press, Cambridge, 263 p.

ODIN, G. S. 1994. Geological time scale. Comptes rendus de l'Académie des sciences, Série II, 318:59-71.

PÁLFY, J. 1995. Development of the Jurassic geochronologic scale. Hantkeniana, 1:13-25.

——, P. L. SMITH, AND J. K. MORTENSEN (in press). A U-Pb and ^{40}Ar-^{39}Ar time scale for the Jurassic: Canadian Journal of Earth Sciences.

PALMER, A. R. 1983. The Decade of the North American Geology 1983 geologic time scale. Geology, 11:503-504.

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Upper Jurassic Foraminiferal Zones and Position of the Jurassic-Cretaceous Boundary in Western Siberia

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Keywords: Biostratigraphy, Foraminifera, Zone, Boundary, Callovian, Oxfordian, Kimmeridgian, Volgian, Berriasian, Western Siberia

Abstract: The biostratigraphy of Upper Jurassic deposits of Western Siberia is based on zones of foraminifera (hereafter abbreviated F-zones) with widespread foraminiferal assemblages. The description of F-zones corresponding to Oxfordian, Kimmeridgian and Volgian stages is given. The Upper Jurassic-Cretaceous boundary coincides with the top of the Upper Volgian F-zone *Ammodiscus veteranus- Evolutinella volossatovi.* The Volgian foraminiferal assemblages cannot be further subdivided; they differ systematically at the species level from the Early Berriasian faunas of the Early Cretaceous.

Introduction

In Western Siberia, Upper Jurassic deposits occur in borehole sections over more than a million square kilometer area. Their upper layers occur at the depths from 750 to 3400m and their thickness is about 100-200m. They were formed during the Late Jurassic transgressive stage with increasing clay content upward in the section. The oil and gas content of these beds was proved 40 years ago, leading to more intensive study of the Upper Jurassic than of other Mesozoic deposits. A great number of wells and different core studies make it possible to concentrate on problems of stratigraphy for areas lacking outcrop. In this respect the Upper Jurassic of Western Siberia can be classified as a typical (standard) region to test the methods and principles of zonal stratigraphy (Msezhnikov, 1984; Anonymous (a), 1991) (Fig. 1).

Over several decades information on the distribution of Jurassic strata in Western Siberia was integrated and complete stratigraphic schemes were given by Rostovtsev (1954), Gurari (1959), Zakharov and Mesezhnikov (1974) and others. Investigators also reconstructed paleogeographical conditions and facies of the Late Jurassic epoch. In 1991 more than 100 regional stratons (the term for all stratigraphic units such as suites, beds, layers and so on) were united into three horizons: Vasyuganski (Callovian to Oxfordian), Georgievski (Kimmeridgian) and Badjenovski (Volgian through Berriasian); these were included in the regional stratigraphical scheme for the Upper Jurassic section of the West Siberian plain.

From the beginning, biostratigraphical investigations of Western Siberia were aimed at microfaunal analysis, mainly foraminifera (Bulynnikova et al., 1972; Anonymous (a), 1991; Anonymous (b), 1991). Foraminiferal zonal assemblages form a definite vertical sequence through the Upper Jurassic section. Very often their areal distribution does not coincide with the lithofacies, nor does their vertical distribution coincide with lithological boundaries. Finding and tracing of detailed biostratigraphical units which are united by the term "zone" is the foundation for the biostratigraphical correlation. We interpret the foraminiferal zone (F-zone) as deposits containing a widespread assemblage of foraminifers with one or two zonal species (index-species) and from 5 to 15 accompanying characteristic species possessing the most narrow vertical ranges. The composition of a zonal assemblage (ratio of agglutinated and calcareous foraminifera) is also taken into account. In the latest (fifth) version of the regional stratigraphical scheme (Anonymous (b), 1991) for the Upper Jurassic deposits of the West Siberian plain two types of division based on foraminifers were retained:

a) The foraminiferal zone (F-zone) means deposits containing zonal species over the whole region; it corresponds to time of their maximum abundance and, as a rule, of their maximum distribution. The F-zone boundaries are taken as isochronous.
b) The foraminiferal layers (f- layers) are those biostratigraphic divisions where index-species are locally developed depending upon particular lithofacies.

The authors give priority to the F-zones for the central portions of the basin lacking outcrops and parallel biostratigraphic schemes based on other groups of organisms. Ammonites, as an example, are so rare that they may be used only to establish correlation with single outcrops of the Upper Jurassic: the North of Siberia, the Subpolar Urals, Mangishlak. Finding a single sample of a zonal ammonite in a borehole core defines neither the zonal distribution nor the position of its lower and upper boundaries. They are often drawn coincident with boundaries of lithological units. In many cases a recognizable sequence of widely distributed foraminiferal assemblages facilitates definition of zone boundaries in monotonous clay-siltstone deposits.

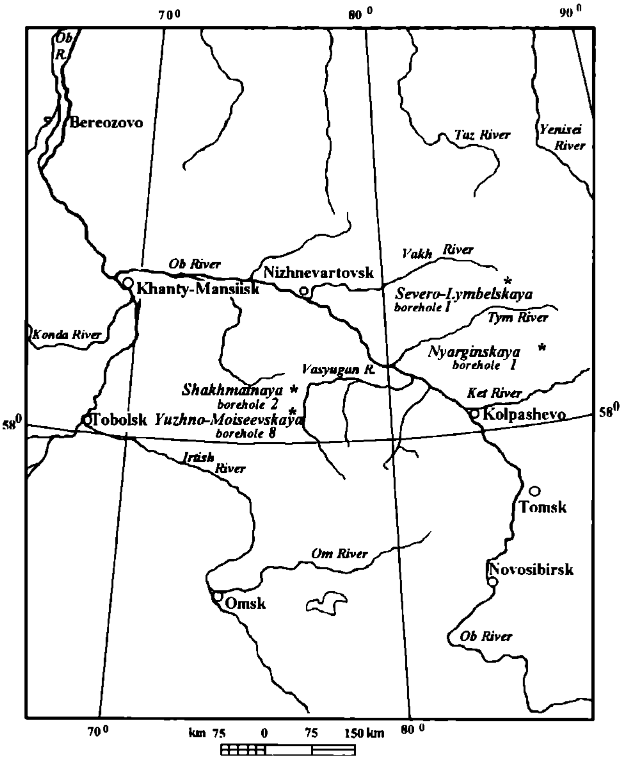

Figure 1. Locality map of boreholes in the west Siberian Plain.

As already noted, Upper Jurassic deposits are an integral part of a very thick oil- and gas-bearing succession in Western Siberia. They have no well-defined lower and upper boundaries. The Callovian- Oxfordian boundary (Middle - Upper Jurassic) lies within monotonous argillites of the Vasyuganski horizon, and is drawn at the contact of the F-zones *Dorothia insperata-Trochammina rostovzevi* (Upper Callovian) and *Ammobaculites tobolskensis - Trochammina oxfordiana* (Lower Oxfordian) (Table 1).

Here we concentrate on description of the Upper Jurassic F-zones which occur mainly in the West Siberian region.

Oxfordian Stage ($J_3 o$). Lower Oxfordian Substage ($J_3 o_1$)

***Ammobaculites tobolskensis - Trochammina oxfordiana* F-Zone**

This zone occurs with considerable variations in all of Western Siberia. In the western sections the layers with *Ammobaculites tobolskensis - Eomarssonella paraconica* correspond to this F-zone (Mesozoic Foraminifera, 1991). In the Polar and Subpolar Urals (eastern slope) the f-layers with *Recurvoides eotrochus* correspond in part with the range of this zone (Bulynnikova et al., 1972). In the Yenisei-Khatangskian depression clayey deposits with an assemblage of numerous foraminiferal species were found. These were distinguished as the f-layers with *Trochammina oxfordiana - Lenticulina memorabilissima* (Sharovsakaya, 1968). Agglutinated quartz-siliceous species in this zone are dominant; calcareous species usually make up 10 %. In the southernmost part of the region the assemblage is represented solely by the calcareous species *Planularia colligatiformis* Kosyreva, *Hoeglundina tjumensis* Kosyreva, is locally distributed. The lower boundary is determined by abundant accumulations of the zonal index-species as well as by appearance of other characteristic species, and the upper boundary by a sharp decrease in their numbers (Fig. 2).

Upper Oxfordian Substage ($J_3 o_3$)

***Recurvoides disputabilis* F-Zone**

This zone also has a wide distribution within this region and occurs in clay and silty facies of the upper part of the Vasyuganski horizon. Agglutinated benthics make up the entire assemblage, representatives of *Recurvoides* and *Spiroplectammina* sometimes making up 100%. In the Anabarski Bay section, the f-layers are .established with the *Trochammina oxfordiana - Ceratobulimina poljarica* assemblage including a great number of Nodosariidae similar to those of the F-zone of *R. disputabilis*. In the eastern and southeastern regions (area along the Yenisei River in Western Siberia) the f- layers with *Glomospirella galinae* were traced into the marine shoreline facies of the Vasyuganskaya and Naunakskaya suites, whose age range corresponds to that of *R. disputabilis* (Sharovskaya, 1968). The lower boundary is established by appearance of the zonal species and other characteristic species, the upper boundary with the substitution of *Recurvoides disputabilis* by representatives of *Verneuilinoides* and *Reophax* (Fig. 3).

Kimmeridgian Stage ($J_3 k$). Lower Kimmeridgian Substage ($J_3 k_1$)

***Haplophragmoides canuiformis - Lenticulina michailovi* F-Zone**

This zone is of general distribution with considerable variations of the second zonal and accompanying species. It corresponds to the Georgievsi horizon. In the northwestern region (the Ob River profile) it is the f-layers with *Cancristiella ambitiosa* Dain (Bulynnikova et al., 1972; Mesozoic Foraminifera, 1991). In the northern regions representatives of *Recurvoides disputabilis planus* Dain and *Lenticulina* occupy an essential place. There are numerous *Trochammina omskensis* Kosyreva in the southeast (Nyarginski typical section). The calcareous benthos, mainly as Lenticulinidae, is common for the shallow water facies of the Marianovskaya, Lopsinskaya, and Sigovskaya suites. The proportion of calcareous forms decreases toward deeper water facies (Georgievskaya, Abalakskaya suites), where members of *Haplophragmoidea* and *Trochammina* are dominant (Fig. 4).

Upper Kimmeridgian Substage ($J_3 k_2$)

***Pseudolamarckina lopsiensis* F-Zone**

This zone was established originally by Dain (*in* Bulynnikova et al., 1972) along the eastern slope of the Subpolar Urals, and later was traced to nearly all the peripheral parts of Western Siberia and the central regions where the Ob River flows from east to west. This zone is traced in clays of the Georgievski horizon. Ceratobuliminidae, sometimes consisting entirely of tests of *Pseudolamarckina,* prevail in the foraminiferal assemblages along the basin margins. In deeper water parts they are associated with unrolled tests of Lenticulinidae (*Marginulina, Vaginulina, Planularia).* In the central part of the region this F- zone was also found, but with other species represented mainly by the calcareous *Lenticulinidae* and rare agglutinated forms. The zone boundaries are determined by appearance and disappearance of the zonal index-species (Fig. 5).

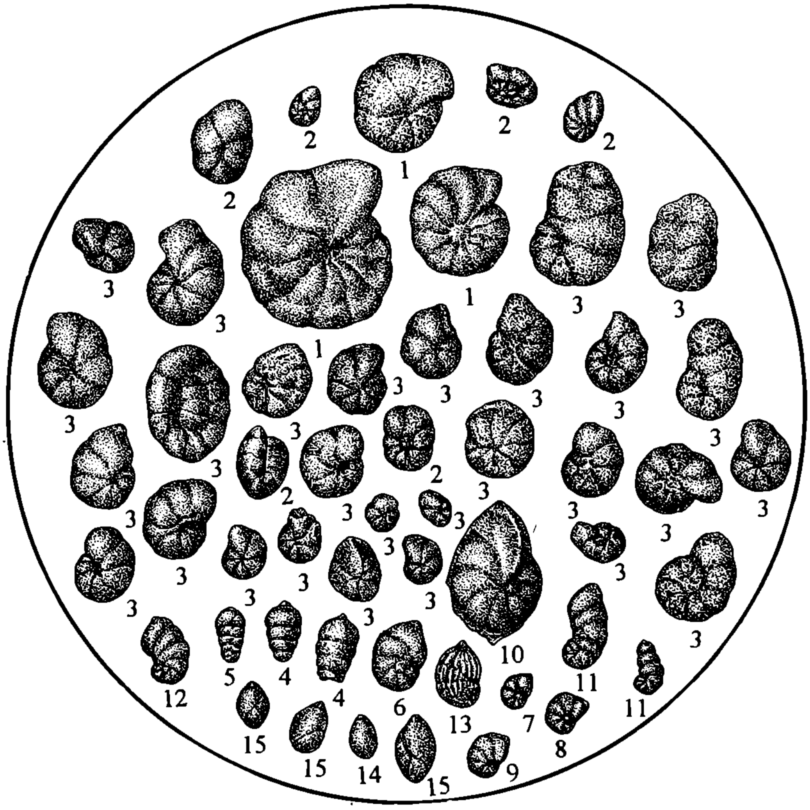

Figure 2. Foraminifera assemblage of *Ammobaculites tobolskensis*, *Trochammina oxfordiana* F-zone, Lower Oxfordian – J_3 ox_1. Western Siberia, Tomsk region, Yuzhno-Moiseevskaya area, borehole 8, depth 2619, 5-2623, 0m, core 6883. **1.** *Haplophragmoides* (*Labrospira*) sp.; **2.** *Recurvoides scherkalyensis* Levina; **3.** *Ammobaculites tobolskensis* Levina and Komissarenko; **4.** *Pseudonodosaria brady* Tappan; **5.** *Geinitzinita praenodulosa* Dain; **6.** *Lenticulina memorabilissima* Gerke and Scharovskaja; **7.** *L.* ex gr. *greisli* Dubrowskaja; **8.** *L. expressa* Putrja; **9.** *L. eurytopica* Putrja; **10.** *L. juganensis* Putrja; **11.** *Astacolus hubrida* Terquem; **12.** *A. pungaicus* Putrja; **13.** *Marginulinopsis praecomptulaformis suprajurensis* Gerke and Scharovskaja; **14.** *Globulina oolithica* Terquem; **15.** *G. paalzovi* Mjatliuk.

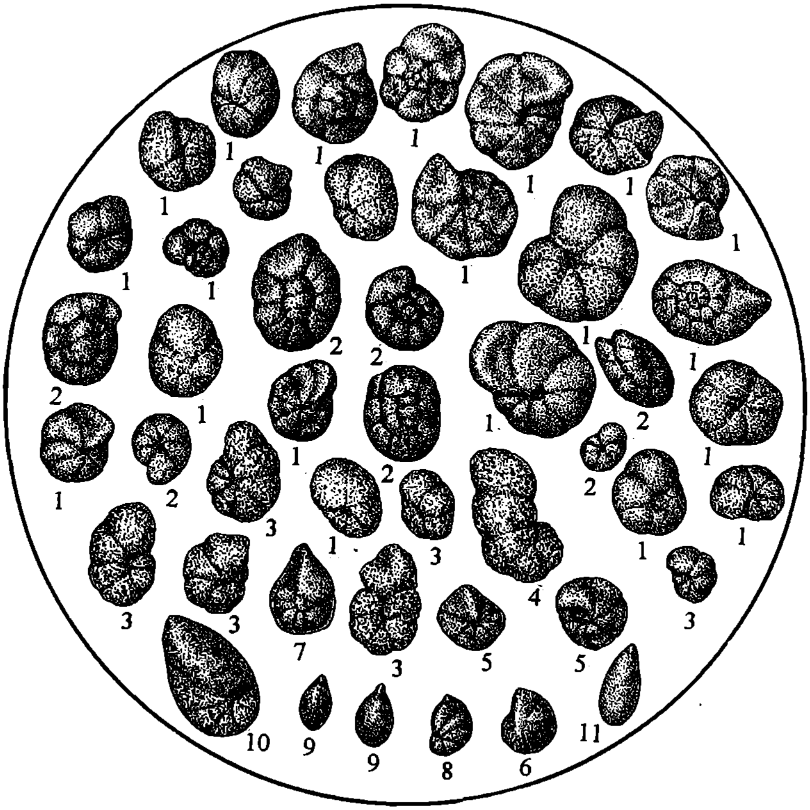

Figure 3. Foraminifera assemblage of *Recurvoides disputabilis* F-zone, Upper Oxfordian – J_3 ox_3. Western Siberia, Tomsk region, Shakhmatnaya area, borehole 2, depth 2481-482m, core 3607. **1.** *Recurvoides disputabilis* Dain; **2.** *R. sublustris* Dain; **3.** *Ammobaculites sundascoensis* Scharovskaja; **4.** *Am. multiformis* Dain; **5.** *Trochammina* ex gr. *minutissima* Dain; **6.** *Lenticulina* ex gr. *orientalis* Kaptarenko; **7.** *Lenticulina* sp.; **8.** *Astacolus inflatiformis* Dain; **9.** *Globulina vulgaris* Dain; **10.** *G.* ex gr. *alexandrae* Dain; **11.** *Globulina* sp.

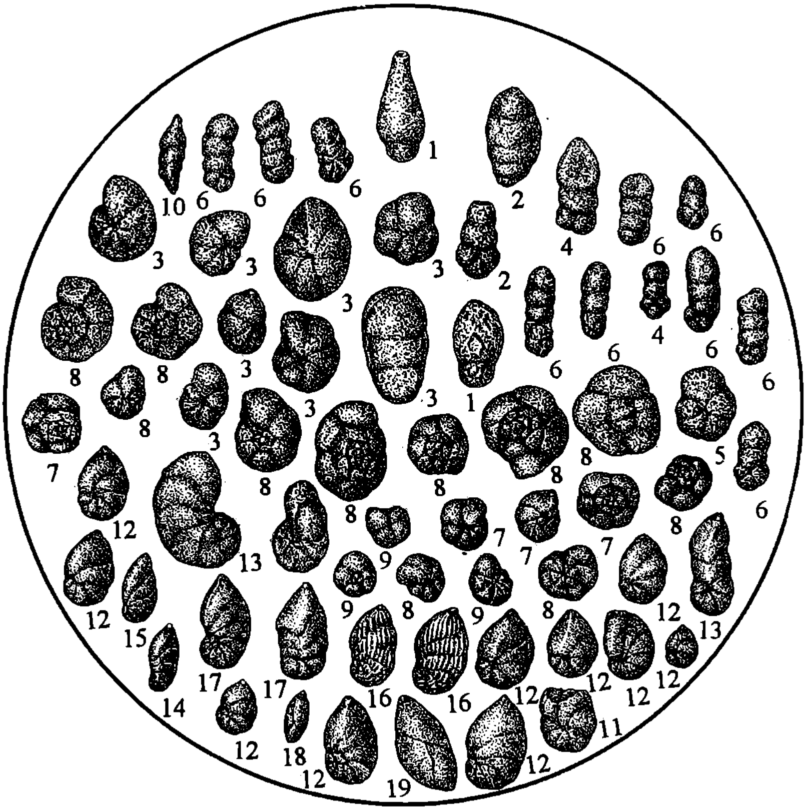

Figure 4. Foraminifera assemblage of *Haplophragmoides canuiformis*, *Lenticulina michailovi* F-zone, Lower Kimmeridgian – J_3km_1. Western Siberia, Tomsk region, Nyarginskaya area, borehole 1, depth 2330, 5-2332, 5m, core 5859. **1.** *Reophax adaptatus* Dain; **2.** *Psammolingulina jurassica* Dain; **3.** *Haplophragmoides canuiformis* Dain; **4.** *Ammobaculites* sp.; **5.** *Am. validus* Beljaevskaja; **6.** *Haplophragmium pocrovkaensis* Kosyreva; **7.** *Trochammina omskensis* Kosyreva; **8.** *Tr.* ex gr. *kumaensis* Levina; **9.** *Tr.* sp.; **10.** *Verneuilinoides graciosus* Kosyreva; **11.** *Lenticulina michailovi* Dain; **12.** *Astacolus* ex gr. *inflatiformis* Dain; **13.** *A. subrusticus* Dain; **14.** *Vaginulina angusta* Putrja; **15.** *V. infida* Putrja; **16.** *Vaginulinopsis* ex gr. *rjavkinoensis* Kosyreva; **17.** *V. dorsoconvexus* Putrja; **18.** *Globulina* ex gr. *oolithica* Terquem; **19.** *G.* ex gr. *alexandrae* Dain.

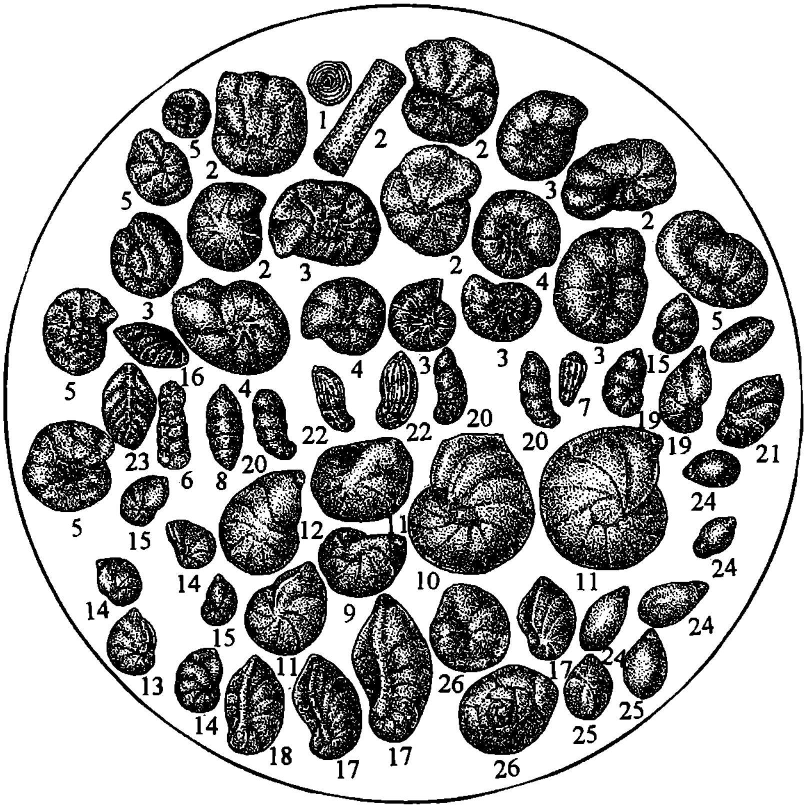

Figure 5. Foraminifera assemblage of *Pseudolamarckina lopsiensis* F-zone, Upper Kimmeridgian – J_3 km_2. Western Siberia, Tomsk region, Severo-Limbelskaya area, borehole 1, depth 2005, 4-2006, 8m, core 5432. **1.** *Spirillina* sp.; **2.** *Haplophragmoides canuiformis* Dain; **3.** *Recurvoides sublustris* Dain; **4.** *R. stschokuriensis* Dain; **5.** *Recurvoides* sp.; **6.** *Spiroplectammina* cf. *vicinalis* Dain; **7.** *Nodosaria* sp.; **8.** *N. tenuitheca* Dain; **9.** *Lenticulina* ex gr. *gerkei* Dain; **10.** *L.* cf. *oppeliformis* Dain; **11.** *L. michailovi* Dain; **12.** *L. limpida* Dain; **13.** *L.* cf. *parahybrida* Dain; 14. *L.* ex gr. *turaensis* Putrja; **15.** *Astacolus rarus* Putrja; **19.** *Vaginulinopsis dorsoconvexus* Putrja; **20.** *V. romanovae* Putrja; **21.** *V. rostovzevi* Putrja; **22.** *V. rjavkinoensis* Kosyreva; **23.** *Citharinella* sp.; **24.** *Globulina prisca* Reuss; **25.** *G. praelacrime* Mjatliuk; **26.** *Pseudolamarckina lopsiensis* Dain.

Volgian Stage (J_3v). Middle Volgian Substage (J_3v_2)

***Dorothia tortuosa - Spiroplectammina vicinalis* F-Zone**

Numerous agglutinated species, mainly of *Spiroplectammina* and *Ammobaculites*, were found. Lenticulinidae are especially prominent among the calcareous forms with sculptured tests of *Marginulina* and others. The systematic composition of foraminifera varies in different regions of Western Siberia. The united F- zone *Dorothia tortuosa - Spiroplectammina vicinalis* consists of two parts along the Urals: the lower part including f-layers with *Spiroplectammina vicinalis - Saracenaria pravoslavlevi* and the upper with *Dorothia tortuosa*. These two f-layers correspond to the lower part of the Badjenovski horizon. On Taimyr the f-layers with *Lenticulina djubapaensis - Citharina nabliuma* containing abundant Nodosariidae are distinguished in the shallow water facies (Grigelis, 1982).

The lower zone boundary is determined by appearance of a considerable number of tests of the zonal species together with ornamented tests of *Marginulina*. Most of characteristic species disappear at the upper boundary. The foraminiferal assemblages of the *Dorothia tortuosa-Spiroplectammina vicinalis* zone were traced from the Baltic in Europe through Western Siberia and the Northern Kamchatka to Alaska and British Columbia of North America. This could be the base for establishing a reliable correlation within the limits of the whole Boreal realm (Fig. 6).

Upper Volgian Substage (J_3v_3)

***Ammodiscus veteranus - Evolutinella volossatovi* F-Zone**

The assemblage with *Ammodiscus veteranus* and *Evolutinella volossatovi* originally established by Komissarenko and Levina (1968) in Western Siberia was traced in all the Mesozoic basins of Northern Siberia and the Barents Sea shelf (Anonymous (a), 1991). The thickness of deposits of the Badjenovski horizon containing this assemblage varies from 5 to 170 m. It contains more than 40 foraminiferal species representing nine families. Great accumulations of agglutinated foraminiferal tests dominated by *Evolutinella* are frequent along margins of the basin. Calcareous shells of Lenticulinidae and Polymorphinidae are prominent in the upper part of this zone. More than half of the species are in common with those of the Middle Volgian and a considerable number of them extend into the Berriasian. In the different f-layers of the *A. veteranus - E. volossatovi* F-zone of Western Siberia agglutinated foraminifera are dominant. Where this occurs, the zonal species make up more than 70 % of the fauna. In most Upper Jurassic beds the systematic composition of this assemblage is less diversified. As a rule the foraminifera are known from shoreline facies, and are notable for the prevalence of genera with more unrolled shells of *Ammodiscus (A. veteranus* Kosyreva*)*, *Evolutinella (E. volossatovi* Scharovskaja*)* and others (Sharovskaya, 1968) (Fig. 7).

At the Jurassic-Cretaceous boundary, benthic foraminifera have essentially changed in systematic composition. The Early Cretaceous assemblages occur in more diverse facies of the Upper part of Badjenovski horizon throughout the whole basin and differ by appearance of more tightly coiled (involute) shells in contrast to Late Volgian faunas. The assemblage with *Trochammina rosaceaformis* was found in the lower part of the Berriasian stage of western regions. Representatives of *Haplophragmoides infracretaceous* Mjatliuk, *Gaudryinopsis gerkei* Sharovskaja are the most abundant in the foraminiferal assemblages of this level. No less than 25 species assigned to 13 genera can be counted in the f-layers with *Trochammina rosaceaformis* in the west and synchronous levels with *Gaudryinopsis gerkei* to the east. There the most characteristic forms are *Haplophragmoides fimbriatus* Scharovskaja, *Ammobaculites* ex gr. *fontinensis*. Terquem, *Orientalia baccula* (Schleifer), *Marginulina zaspelovae* Romanova and *Globulina chetaensis berriassica* Basov. The first appearance of younger species, typical of the Valanginian, is notable at this level (Tatyanin, 1985; Podobina and Tatyanin, 1996).

Table 1. Comparison of foraminiferal and ammonite biozonation for the Upper Jurassic of western Siberia.

System	Section	Stage	Substage	Standard scale		Boreal standard (Zakharov et al., 1997)			F-zones and f-layers (Data of authors)
CRETACEOUS	Lower	Berriasian	Lower	Tirnovella occitanica	Dalmasiceras dalmasi	Hectoroceras kochi			Layers with Trochammina rosaceaformis
					Berriasella privasensis				
					Tirnovella subalpina				
				Berriasella jacobi	Pseudosubpl. grandis	Chetaites sibiricus			
					Berriasella jacobi				
JURASSIC	Upper	Tithonian	Upper	"Durangites"		Volgian	Upper	Craspedites nodiger	Ammodiscus veteranus, Evolutinella volossatovi
				Paraulacosphinctes transitorius				Craspedites subditus	
								Kashpurites fulgens	
			Middle	Microcantoceras pont			Middle	Paracrasp. opressus	Spiroplectammina vicinalis, Dorothia tortuosa
								Epivirgatites nikitini	
				Semiformiceras falluxi				Virgatites virgatus	
				Semiformiceras semiforme				Dorsoplanites panderi	
			Lower	Danubisphinctes palatinum			Lower	Ilovaiskya pseudoscythica	Layers with Pseudolamarckina voliaensis
				Franconites vimineus					
				Usseliceras parvinodosum				Ilovaiskya sokolovi	
				Dorsoplanitoides triplicatus					
				Usseliceras tagmersheimense					
				Hybonoticeras hybonotum				Ilovaiskya klimovi	
		Kimmeridgian	Upper	Aulacostephanus autissiodorensis		Aulacostephanus autissiodorensis			Pseudolamarckina lopsiensis
				Aulacostephanus eudoxus		Aulacostephanus eudoxus			
				Aulacostephanus mutabilis		Aulacostephanus acanthicus			
			Lower	Rasenia cymodoce		Amoeboceras kitchini		Rasenia borealis	Haplophragmoides canuiformis, Lenticulina michailovi
				Pictonia baylei				Pictonia involuta	Layers with Verneuilinoides graciosus, Trochammina omskensis
		Oxfordian	Upper	Ringsteadia pseudocordata		Amoeboceras rosenkratzi			Recurvoides disputabilis
						Amoeboceras regulare			
				Decipia decipiens		Amoeboceras serratum			
				Perisphinctes cautisnigrae		Amoeboceras glosense			
			Middle	Gregoriceras transversarium		Cardioceras tenuiserratum			Layers with Ammodiscus pseudoinfimus, Tolypammina svetlanae
				Perisphinctes plicatilis		Cardioceras densiplicatum			
			Lower	cordatum	Cardioceras cordatum	Cardioceras cordatum			Ammobaculites tobolskensis, Trochammina oxfordiana
					Cardioceras percaelatum	Cardioceras percaelatum			
					Cardioceras bukowskii	Cardioceras gloriosum			
				Vertumniceras mariae		Cardioceras obliteratum			
	Middle	Callovian	Upper	Quenstedtoceras lamberti		Quenstedtoceras lamberti			Dorothia insperata, Trochammina rostovzevi
				Peltoceras athleta		Peltoceras athleta			

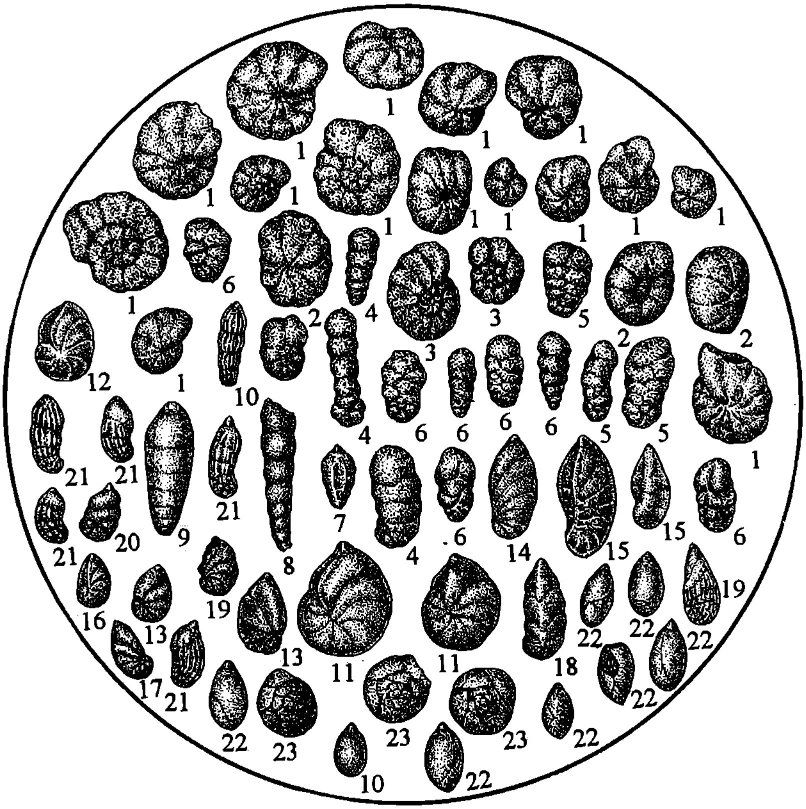

Figure 6. Foraminifera assemblage of Dorotinia tortuosa, Spiroplectammina vicinalis F-zone, Middle Volgian – J_3v_2. Western Siberia, Tomsk region, Nyarginskaya area, borehole 1, depth 2275, 102278, 1m, core 5803. **1.** *Haplophragmoides schleiferi* Scharovskaja; **2.** *Recurvoides praeobskensis* Dain; **3.** *Ammobaculites* cf. *haplophragmioides* Furssenko and Polenova; **4.** *A. gerkei* Scharovskaja; **5.** *Spiroplectammina vicinalis* Dain; **6.** *Dorothia tortuosa* Dain and Komissarenko; **7.** *Geinitzinita praenodulosa* Dain; **8.** *Dentalina* sp.; **9.** *Pseudonodosaria tutkowskii* Mjatliuk; **10.** *Nodosaria* sp.; **11.** *Lenticulina kasanzevi* Furssenko and Polenova; **12.** *L. vladimirskensis* Putrja; **13.** *Astacolus* sp.; **14.** *Planularia mulymiensis* Putrja; **15.** *Saracenaria prolata* Kusnezova; **16.** *S. pravoslavlevi* Furssenko and Polenova; **17.** *Vaginulina infida* Putrja; **18.** *Vaginulinopsis dorsoconvexus* Putrja; **19.** *Vaginulinopsis microcostatus* Putrja; **20.** *Marginulina formosa* Mjatliuk; **21.** *M. complacida* Putrja; **22.** *Globulina lacrima* Reuss; **23.** *Pseudolamarckina voliaensis* Dain.

Figure 7. Foraminifera assemblage of *Ammodiscus veteranus*, *Evolutinella volossatovi* F-zone, Upper Volgian – J_3v_3. Western Siberia, Tomsk region, Severo-Limbelskaya area, borehole 1, depth 1971, 9-1983, 9m, core 5428. **1.** *Ammodiscus veteranus* Kosyreva; **2.** *Evolutinella volossatovi* Scharovskaja; **3.** *Trochammina rosacea* Zaspelova; **4.** *Dentalina* sp.; **5.** *Astacolus* ex gr. *tigjanensis* Schleifer; **6.** *Marginulina complacida* Putrja.

The Jurassic - Cretaceous boundary was drawn in clayey rocks of the Badjenovski horizon. It is difficult to establish the position of this boundary on lithological features and log survey diagrams, but it can be traced by the disappearance of Late Volgian foraminifera of the F-zone *Ammodiscus veteranus - Evolutinella volossatovi* as well as by the appearance of Berriasian foraminifera (f-layers with *Trochammina rosaceaformis*), which are significantly different in systematic composition and reflect the beginning of a new stage in their development (Podobina and Tatyanin, 1996).

We believe the Jurassic-Cretaceous boundary coincides with the top of deposits of the Upper Volgian F- zone *Ammodiscus veteranus- Evolutinella volossatovi*. This zone is closely connected with other underlying Volgian ones. The Volgian foraminiferal assemblages cannot be subdivided and represent related developmental steps. Thus, deposits containing the *Ammodiscus veteranus - Evolutinella volossatovi* assemblage cannot be dated as Early Berriasian (Table 1).

References

ANONYMOUS (a). 1991 .Mesozoic Foraminifera. Practical guidance on microfauna of USSR. Leningrad, Nedra, 5:84-97 (in Russian).

ANONYMOUS (b). 1991 . Regional stratigraphical schemes of Mesozoic deposits of the West Siberian Plain. Proceedings of the West Siberian Research Gas and Oil Institute, Tyumen (A collective of authors, in Russian).

BULYNNIKOVA, S. P., et al. 1972. Foraminifera of Upper Jurassic deposits of Western Siberia. Leningrad, Nedra, p. 1-272 (in Russian).

GRIGELIS A. A. (ed.). 1982. Biostratigraphy of Upper Jurassic deposits of USSR based on foraminifera. Vilnius, Mokslas, p. 120-146 (in Russian).

GURARI, F. G. 1959. Geology and perspectives of oil-and-gas content of the area between Ob and Irtish rivers. Proceedings of Siberian Research Institute of Geology, Geophysics and mineral Resources, 3:1-178 (A collective of authors, in Russian).

KOMISSARENKO, V. K., AND V. I. LEVINA. 1968. To substantiation of age of the *Ammodiscus veteranus* zone in Jurassic. Cretaceous boundary deposits in the West Siberian lowland. Proceedings on stratigraphy of Mesozoic and Cenozoic deposits of Western Siberia, Moscow, Nedra, p. 114-118 (in Russian).

MESEZHNIKOV, M. S. 1984. Kimmeridgian and Volgian stages of northern regions of USSR. Leningrad, Nedra, p. 1-224 (in Russian).

PODOBINA, V. M., AND G. M. TATYANIN. 1996. Stratigraphy of the Upper Jurassic of Western Siberia based on foraminifers, p. 215-220. *In* Advances in Jurassic research. A .C. Riccardi (ed.). Transtec Publications. Switzerland.

ROSTOVTSEV, N. N. 1954. Geological structure and perspectives of oil-and-gas content of the south region of the West Siberian lowland. Data on Geology, Hydrogeology and oil-and-gas content of Western Siberia. Leningrad, p. 5-60 (in Russian).

SHAROVSKAYA, N. V. 1968. Assemblages of foraminifera from Jurassic and Lower Cretaceous deposits of Ust-Yenisei and Turukhan-Ermakovskian regions. Scientific notes of the Research Institute of Arctic Geology. Paleontology and Biostratigraphy, 23:106-117 (in Russian).

TATYANIN, G. M. 1985. Stratigraphy and Upper Jurassic foraminifera of southeast of Western Siberia, Tomsk, 20 (in Russian).

WALL, J. H. 1983. Jurassic and Cretaceous foraminiferal biostratigraphy in the eastern Sverdrup basin, Canadian Arctic Archipelago. Bulletin of Canadian Petroleum Geology, 31:246-281.

ZAKHAROV, V. A., et al. 1997. Boreal zonal standard and Mesozoic biostratigraphy of Siberia. Geology and Geophysics, 38:927-956 (in Russian).

____, AND M. S. MESEZHNIKOV. 1974. Volgian stage of northern regions of the eastern slope of the Urals. Novosibirsk, Nauka, p. 1-216 (in Russian).

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New Biostratigraphic Data from the Kimmeridgian/Tithonian Boundary Beds of SW Germany

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Abstract: An overview is given on the state of biostratigraphic refinement of the Upper Kimmeridgian and lowest Tithonian in SW Germany. The ammonite faunal horizons of the Upper Kimmeridgian Ulmense Zone and the succeeding Lower Tithonian Hybonotum Zone are briefly characterised. The *eigeltingense* horizon is proposed to represent the oldest faunal horizon of the Hybonotum Zone (Riedense Subzone). It is possible to make long-distance correlations of this local/regional biohorizon by the co-occurrence of *Euvirgalithacoceras eigeltingense*, *Subplanites postrueppellianus* and *Hybonoticeras pseudohybonotum.* The existing outcrops, however, do not provide a possible GSSP candidate section; such an exposure must be sought elsewhere.

The Search for a Reference Section of the Kimmeridgian/Tithonian Boundary

Recently, several members of the Tithonian working group have been looking for a possible GSSP for the Kimmeridgian/Tithonian boundary which lies, in biochronostratigraphic terms, at the lower boundary of the lowest standard zone of the Tithonian, the Hybonotum Zone. However, most of the sections in Europe are incomplete, or condensed, or still lacking a modern and detailed biostratigraphic investigation. One of the areas which could provide possible reference sections or auxiliary stratotype sections is the Upper Jurassic of SW Germany (Fig. 1).

Kimmeridgian/Tithonian Lithostratigraphy in SW Germany

The Upper Jurassic in SW Germany has been one of the classical areas of ammonite research and biostratigraphy since the times of Quenstedt and Oppel. This area is part of the Submediterranean Faunal Province, which is characterised by a dominating stock of Tethyan faunal elements along with minor faunal influx from several other areas, especially France (Paris Basin). The lithostratigraphic sequence around the Kimmeridgian/Tithonian boundary consists mainly of micritic limestones ("Liegende Bankkalke Formation", "Hangende Bankkalke Formation"), marly limestones ("Zementmergel Formation") and interfingering massive spongiolithic limestones (Fig. 2). In the middle and eastern part of the Swabian Jurassic small coral reefs occur. The famous locality of Nattheim, known since the last century with its fine silicified preservation of the coral fauna, lies within this interval. In the western part of Swabia, lithographic limestones occur in small outcrops near the village of Nusplingen ("Nusplingen Lithographic Limestone Formation").

Some Remarks on Oppel's *Ammonites ulmensis*

Since the time of Oppel, *Lithacoceras ulmense* (Oppel) has often erroneously been thought to represent a typical Lower Tithonian ammonite species - even by Oppel himself, who changed his original specific concept of *Ammonites ulmensis* (Oppel, 1858) in his later paper of 1863 by including specimens from Franconia (Solnhofen area) in his type series. Unfortunately, Schneid (1914) interpreted *Lithacoceras ulmense* in a way which was not according to the IRZN, and was followed by Arkell (1937, 1956), by taking a specimen of Schneid as "lectotype" of the species. Later, Arkell (1946) proposed an "Ulmense Zone" within the Tithonian succession of southern Franconia following an erroneous correlation of Roll (1933) between the Swabian and Franconian

Upper Jurassic. *Lithacoceras ulmense* and allied lithacoceratids are very common in the interval between the Setatum Subzone and the base of the Tithonian, so that an Ulmense Subzone was re-established and re-defined as the youngest subzone of the Late Kimmeridgian Beckeri Zone (Schweigert and Zeiss, 1994; Zeiss, 1994). Its index species, *Lithacoceras ulmense* (Oppel), was often misinterpreted because of the different species concepts mentioned above, so that a revision of this important species became urgently necessary. The revision was mainly based on newly discovered complete specimens from the Nusplingen Lithographic Limestone (Schweigert and Zeiss, 1998b).

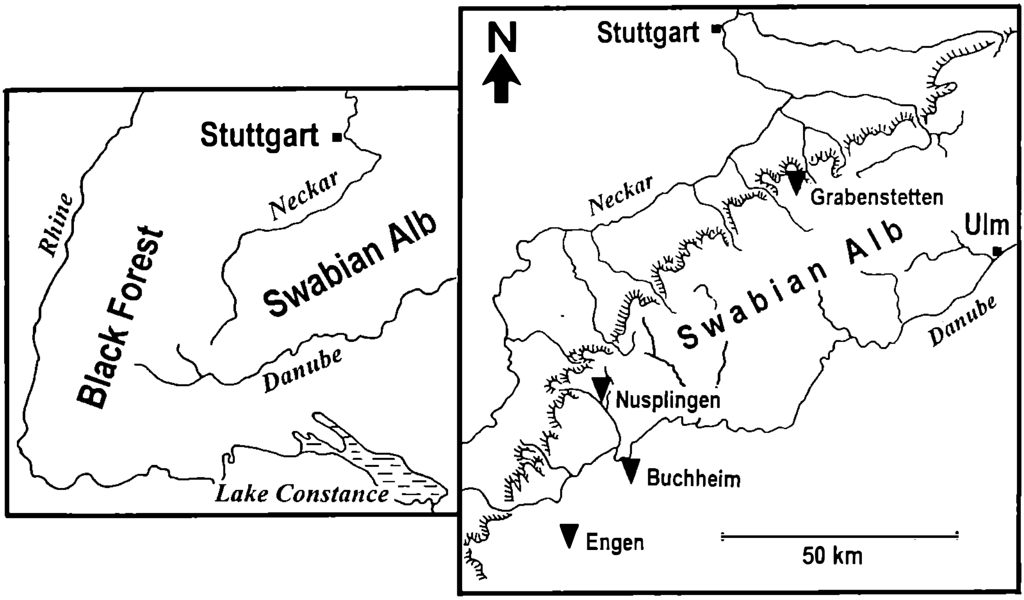

Figure 1. Location of important sections of Kimmeridgian/Tithonian boundary beds in SW Germany.

The Base of the Tithonian

The base of the Tithonian is not yet defined by a GSSP, but there is agreement that it should be at the base of the Hybonotum Zone. Hence, we have to investigate the succession of ammonite faunas. The assumed boundary is arbitrary, at a level selected by consensus of the Jurassic Subcommission and the members of the boundary working group. A criterion for defining the boundary might be, for example, the first occurrence of *Hybonoticeras* of the *H. hybonotum* group or of some other taxon characteristic of the true Tithonian. Additionally, the ammonite faunal horizon which is defined as representing the base of the Tithonian should be characteristic enough to allow long-distance correlations. Hence, selection of an ammonite horizon with high endemism for the base of the Tithonian stage should be avoided.

A high resolution ammonite biostratigraphy based on the concept of ammonite faunal horizons has been recently established (Schweigert, 1993; 1996; 1998; Schweigert and Zeiss, 1994; 1998; 1999; Schweigert et al., 1996, Zeiss et al., 1996). In Swabia, the ammonite faunal horizon which yields one of the oldest *Hybonoticeras* species of the *H. hybonotum* group is represented by the *eigeltingense* horizon. This ammonite horizon is always found in the lowermost part of the "Hangende Bankkalke Formation", just above a prominent disconformity, where the lithofacies changes from marls or heavily bioturbated detrital carbonates to pure micritic limestones. The *eigeltingense* horizon is part of the Riedense Subzone, the oldest Subzone of the Hybonotum Zone.

Whether or not to accept the base of the *eigeltingense* horizon, as proposed herein, for the base of the Tithonian, will have to be agreed upon by the Tithonian working group and the Subcommission.

Besides the index, *Euvirgalithacoceras eigeltingense*, and its antidimorph, *Subplanites postrueppellianus*, the *eigeltingense* horizon yields strange, simple ribbed perisphinctids (*Berckhemeria scherzingeri* [m] pl. 2.6 – *Presimoceras heteromorphum* [M]) which may represent the root stock of the Tithonian heteromorphs (Schweigert and Zeiss, 1998). The occurrence of *Euvirgalithacoceras/Subplanites* together with *Hybonoticeras pseudohybonotum* enables correlations to be made with other Submediterranean areas (SE France, Atrops, pers. comm.; and E Spain, Atrops and Meléndez, 1994) as well as with the Tethyan realm.

The faunal horizons just below the *eigeltingense* horizon should be placed in the Upper Kimmeridgian Beckeri Zone (Ulmense Subzone, respectively), because the *Hybonoticeras* species characteristic of this unit is markedly different from the younger *H. hybonotum* group *(H. pseudohybonotum*, *H. hybonotum*, *H. robustum*) having characteristic keels lacking dorsal spines (Pl. 2.4; for more details on the *Hybonoticeras* lineage see Schweigert et al., 1996). Ammonites known from higher ammonite faunal horizons of the Lower Tithonian in Swabia are listed in Schweigert (1996).

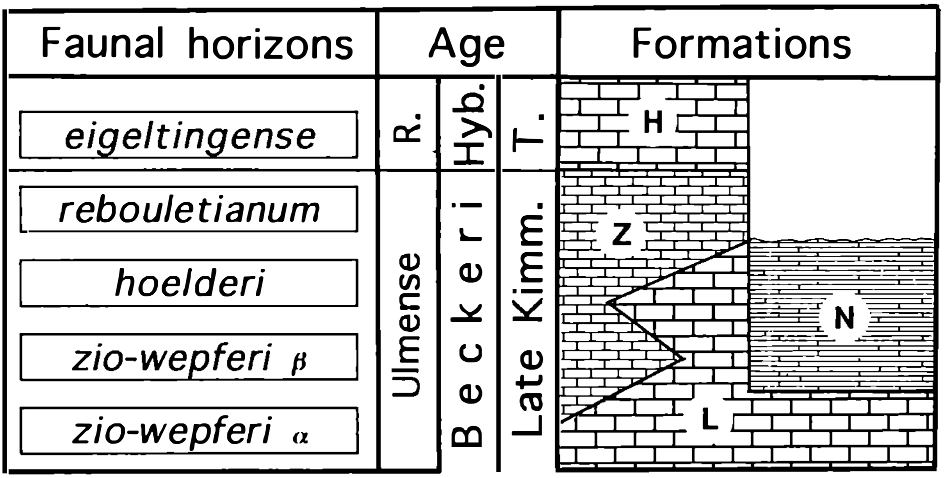

Figure 2. Biostratigraphic subdivision of the Kimmeridgian/Tithonian boundary beds in the Upper Jurassic of Swabia by ammonite faunal horizons. L = Liegende Bankkalke Formation; Z = Zementmergel Formation; N = Nusplingen Lithographic Limestone; H = Hangende Bankkalke Formation; R = Riedense Subzone; T = Tithonian.

The Ammonite Faunal Horizons Around the Kimmeridgian/Tithonian Boundary

In Swabia, the Ulmense Subzone can be subdivided into 4 ammonite faunal horizons. They are recorded in numerous outcrops all over Swabia. Two well exposed outcrops, located along a road near the village of Grabenstetten in the middle part of the Swabian Alb (see Berckhemer and Hölder, 1959: fig. 1) can serve as reference sections for this succession. The ammonite fauna of the *hoelderi* horizon was exhaustively described from the Nusplingen Lithographic Limestone sections (Schweigert, 1998).

zio-wepferi Horizon α

Taramelliceras wepferi Berckhemer [M] (Pl. 1.1) – very common, *Ochetoceras zio* (Oppel) [M] (Pl. 1.3), *Lithacoceras hillebrandti* Schweigert and Zeiss [M], *Silicisphinctes paraboliferus* (Berckhemer and Hölder) [m] (Pl. 1.4), *Lithacoceras* aff. *ulmense* (Oppel) [M], *Silicisphinctes oxypleurus* (Herbich) [m], *Glochiceras* cf. *Lens* Berckhemer [m], *Lingulaticeras nodosum* (Ziegler) [m], *Euvirgalithacoceras albulum* (Berckhemer and Hölder *non* Quenstedt) [M], *Aspidoceras hystricosum* (Quenstedt) [M], *Hybonoticeras extraspinatum* Berckhemer and Hölder [M], *Hybonoticeras harpephorum* (Neumayr) [M]; extremely rare: *Tolvericeras murogense murogense* Hantzpergue. [M].

***zio-wepferi* Horizon β**

Ochetoceras zio (Oppel) [M] – very common, *Taramelliceras wepferi* Berckhemer [M], *Lithacoceras* aff. *onukii* Takahashi [M], *Lithacoceras* aff. *ulmense* (Oppel) (Pl. 2.1), *Silicisphinctes keratinitiformis* Schweigert [m] (Pl. 1.6), *Silicisphinctes oxypleurus* (Herbich), *Glochiceras politulum* (Quenstedt) [m], *Sutneria rebholzi* Berckhemer [m] – very common, *Physodoceras nattheimense* Schweigert [M], *Hybonoticeras harpephorum* (Neumayr) [M]; extremely rare: *Aulacostephanus jasonoides* (Pavlow) [M], *Sarmatisphinctes* cf. *fallax* (Ilovaisky and Florensky) [m+M] (Pl. 2.2), *Tolvericeras murogense katroliforme* Hantzpergue [M].

***hoelderi* Horizon**

Lithacoceras ulmense (Oppel) [M], *Silicisphinctes hoelderi* (Sapunov) [m] (Pl. 1.5), *Lithacoceras fasciferum* (Neumayr) [M], *Silicisphinctes russi* Schweigert [m] (Pl. 1.7), *Ochetoceras* aff. *zio* (Oppel) [M], *Neochetoceras subnudatum* (Fontannes) [M], *Lingulaticeras pseudopercevali* Schweigert [m] (Pl. 1.2), *Physodoceras nattheimense* Schweigert [M], *Sutneria* cf. *rebholzi* Berckhemer [m], *Aspidoceras catalaunicum* (Loriol) [M]; extremely rare: *Gravesia lafauriana* Hantzpergue [M], *Streblites* cf. *zlatarskii* (Sapunov) [M], *Hybonoticeras* cf. *extraspinatum* Berckhemer and Hölder [M] (Pl. 2.3), *Hybonoticeras harpephorum crassicostatum* Olóriz.

***rebouletianum* Horizon**

"Lithacoceras" pseudoulmensis (Furlani) [M], *Neochetoceras rebouletianum* (Fontannes) [M] (Pl. 2.5), *Lingulaticeras planulatum* Ziegler [m], *Sutneria bracheri* Berckhemer [m], *Physodoceras eligmoptychum* (Fontannes) [M], *Hybonoticeras "interlaevigatum"* Berckhemer mscr. [M] (Pl. 2.4), *Aspidoceras catalaunicum* (Loriol) [M]; extremely rare: *Aulacostephanus autissiodorensis* (Cotteau), ?*Gravesia irius* (d'Orbigny) [M].

In Swabia, the following association indicates the basal Tithonian, in accordance with the proposal above to take the *eigeltingense* horizon as the lowermost faunal horizon of the Hybonotum Zone:

***eigeltingense* Horizon**

Euvirgalithacoceras eigeltingense (Ohmert and Zeiss) [M], *Subplanites postrueppellianus* Ohmert and Zeiss [m], *Lithacoceras riedense* (Ohmert and Zeiss) [M], *Silicisphinctes siliceus* Schneid non Quenstedt, *Neochetoceras praecursor* Zeiss [M], *Lingulaticeras* cf. *solenoides* (Quenstedt) [m], *Physodoceras episum* (Oppel) [M], *Sutneria* cf. *apora* (Oppel) [m], *Berckhemeria scherzingeri* Schweigert and Zeiss [m] (Pl. 2.6), *Presimoceras heteromorphum* (Quenstedt) [M], *Hybonoticeras pseudohybonotum* Vigh [M].

Note that all ammonite faunal horizons yield both hybonoticeratids and subboreal ammonites. These faunal elements are important for long-distance correlations between SW Germany, the Tethyan realm (Betic chains of Spain, Italy, Hungary) and adjacent areas (SE France, W France, E Spain, N Germany, Dorset, Poland, Russia). For example, the upper part of the Subboreal Autissiodorensis Zone coincides well with the Tethyan Ulmense Subzone (Kutek and Zeiss, 1997). Moreover, it is possible to recognise some of the Swabian faunal horizons of the Ulmense Subzone, especially the *zio-wepferi* horizon at the base and the *rebouletianum* horizon at the top, in SE France. In Franconia, however, the stratigraphic record of the Ulmense Subzone seems to be very incomplete, and the Torleite section in the classical Solnhofen area does not cross the Kimmeridgian-Tithonian boundary as previously thought, but only the boundary between the Setatum Subzone and a very reduced Ulmense Subzone.

A Possible GSSP for the Base of the Tithonian in SW Germany?

The succession of faunal horizons in Swabia listed above was mainly compiled from a number of isolated outcrops, but unfortunately there is no section which provides a continuous profile and could be proposed as a GSSP candidate section for the Lower Tithonian boundary. The best conditions in this area are around the village of Grabenstetten (middle part of the Swabian Alb) or near the town of Engen NW of the Lake Constance (Fig. 1). Concerning the first section, the ammonite fauna of the *eigeltingense* horizon has been found in an abandoned quarry which lies about 10 meters above the youngest limestones exposed along the road. In the section near Engen, the *eigeltingense* horizon is well developed, but the underlying strata are extremely poor in ammonites. Even the Grabenstetten sections have to be studied in more detail, but there is no doubt as to their value as reference sections for the succession of ammonite faunal horizons of the Ulmense Subzone in SW Germany. Another section in which both the *rebouletianum* horizon and the *eigeltingense* horizon are exposed is located in a large quarry near Buchheim (Danube valley area). There, the boundary lies within a succession of massive spongiolithic limestones.

The higher parts of the Lower Tithonian Hybonotum Zone can be studied in the western part of Swabia, but most fossiliferous strata are only temporarily exposed. Magnetostratigraphic investigations have still not been done. An alternative biostratigraphic zonation across the Kimmeridgian/Tithonian boundary based on dinoflagellate cysts was not very successful for preservational reasons. Nevertheless, the biostratigraphic data obtained from Swabia should be kept in mind when looking for a GSSP candidate section in SE France or elsewhere.

Acknowledgments

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References

ARKELL, W. J. 1937. A monograph on the ammonites of the English Corallian Beds. Part III. Palaeontological Society, 90/406:XLVII–LIV, 43–67.

____. 1946. Standard of the European Jurassic. Geological Society of America Bulletin, 57:1–34.

____. 1956. Jurassic Geology of the World. 757 p.

ATROPS, F., AND G. MELENDEZ. 1984. Kimmeridgian and Lower Tithonian from the Calanda-Bergé area (Iberian Chain, Spain). International Symposium on Jurassic Stratigraphy, Erlangen 1994. 377–392.

KUTEK, J., AND A. ZEISS 1997. The highest Kimmeridgian and Lower Volgian in Central Poland; their ammonites and biostratigraphy. Acta geologica Polonica, 47:107–198.

ROLL, A. 1933. Über den Oberen Malm der südwestlichen Frankenalb. Centralblatt für Mineralogie, Geologie und Paläontologie, Beilagenband, 1933:553–557.

SCHNEID, T. 1914. Geologie der Fränkischen Alb zwischen Eichstätt und Neuburg a. D. Geognostische Jahreshefte, 27:59–170.

SCHWEIGERT, G. 1993. Subboreale Faunenelemente (Ammonoidea) im Oberen Weißjura (Oberkimmeridgium) der Schwäbischen Alb. Profil, 5:141–155.

____. 1996. Die Hangende Bankkalke-Formation im schwäbischen Oberjura. Jahresberichte und Mitteilungen des Oberrheinischen geologischen Vereins, Neue Folge, 78:281–308.

____. 1998. Die Ammonitenfauna des Nusplinger Plattenkalks (Ober-Kimmeridgium, Beckeri-Zone, Ulmense-Subzone, Baden-Württemberg). Stuttgarter Beiträge zur Naturkunde, Serie B, 267:1–61.

____, J. KRISHNA, B. PANDEY, AND D. B. PATHAK. 1996. A new approach to the correlation of the Upper Kimmeridgian Beckeri Zone across the Tethyan Sea. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 202:345–373.

____, AND A. ZEISS 1994. Ammonite biostratigraphy of the Upper Kimmeridgian to Tithonian of southern Germany, p. 27–28. *In* R. R. Leinfelder (ed.), The Origin of Jurassic Reefs: Current Research Developments and Results. Facies, 31.

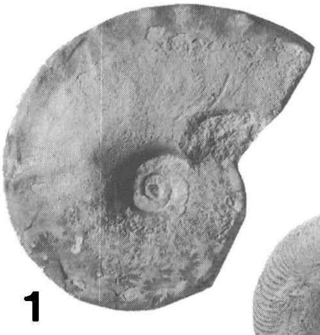
____, AND ____. 1998. *Berckhemeria* n. gen. (Passendorferiinae), eine neue Ammonitengattung aus dem Unter-Tithonium (Hybonotum-Zone) von Süddeutschland. Neues Jahrbuch für Geologie und Paläontologie, Monatshefte, 1998:559–576.

___, AND ____. 1999. *Lithacoceras ulmense* (OPPEL) (Ammonitina) – eine wichtige Leitart des Ober-Kimmeridgiums. Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen, 211:49–73.

ZEISS, A. 1994. Neue Ammonitenfunde aus dem oberen Malm Süddeutschlands. Abhandlungen der Geologischen Bundes-Anstalt, 50:509–528.

____, G. SCHWEIGERT, AND A. SCHERZINGER. 1996. *Hegovisphinctes* n. gen., eine neue Ammonitengattung aus dem Unter-Tithonium des nördlichen Hegaus und einige Bemerkungen zur Taxonomie der Lithacoceratinae. Geologische Blätter für Nordost-Bayern, 46:127–144.

Plate 1. **1**. *Taramelliceras wepferi* Berckhemer. Beckeri Zone, Ulmense Subzone, *zio-wepferi* horizon α. Nusplingen, SW Germany. SMNS No. 63646; **2**. *Lingulaticeras pseudopercevali* Schweigert. Beckeri Zone, Ulmense Subzone, *zio-wepferi* horizon β. Tuttlingen, abandoned Mattsteige quarry, SW Germany. SMNS No. 63079; **3**. *Ochetoceras zio* (Oppel). Beckeri Zone, Ulmense Subzone, *zio-wepferi* horizon α. Tuttlingen, abandoned Mattsteige quarry, SW Germany. SMNS No. 63631/2; **4**. *Silicisphinctes paraboliferus* Berckhemer *in* Berckhemer and Hölder. Beckeri Zone, Ulmense Subzone, *zio-wepferi* horizon α, Gerhausen, SW Germany. SMNS No. 63345; **5**. *Silicisphinctes hoelderi* (Sapunov). Beckeri Zone, Ulmense Subzone, *hoelderi* horizon. Neresheim, SW Germany. GPIT No. Ce 1101/43; **6**. *Silicisphinctes keratinitiformis* Schweigert. Ulmense Subzone, *zio-wepferi* horizon β. Möhringen, Danube valley, SW Germany. SMNS No. 62646. 7. *Silicisphinctes russi* Schweigert. Beckeri Zone, Ulmense Subzone, *hoelderi* horizon. Neresheim, SW Germany. GPIT No. Ce 1101/49. Scale bar 1 cm.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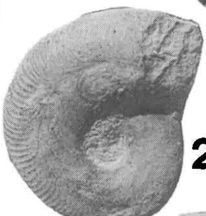
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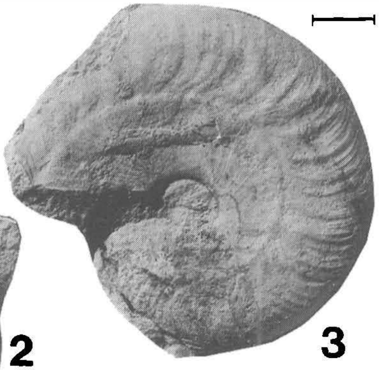
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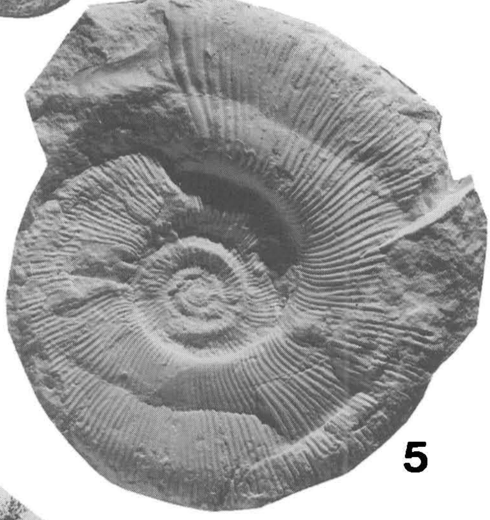
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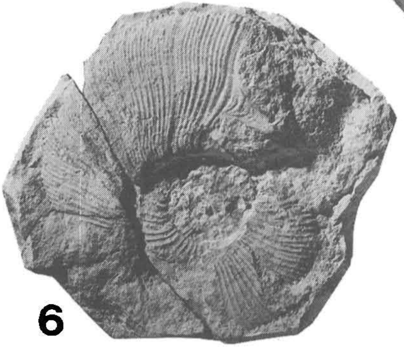
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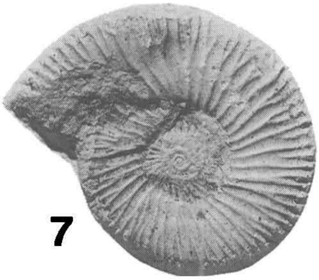
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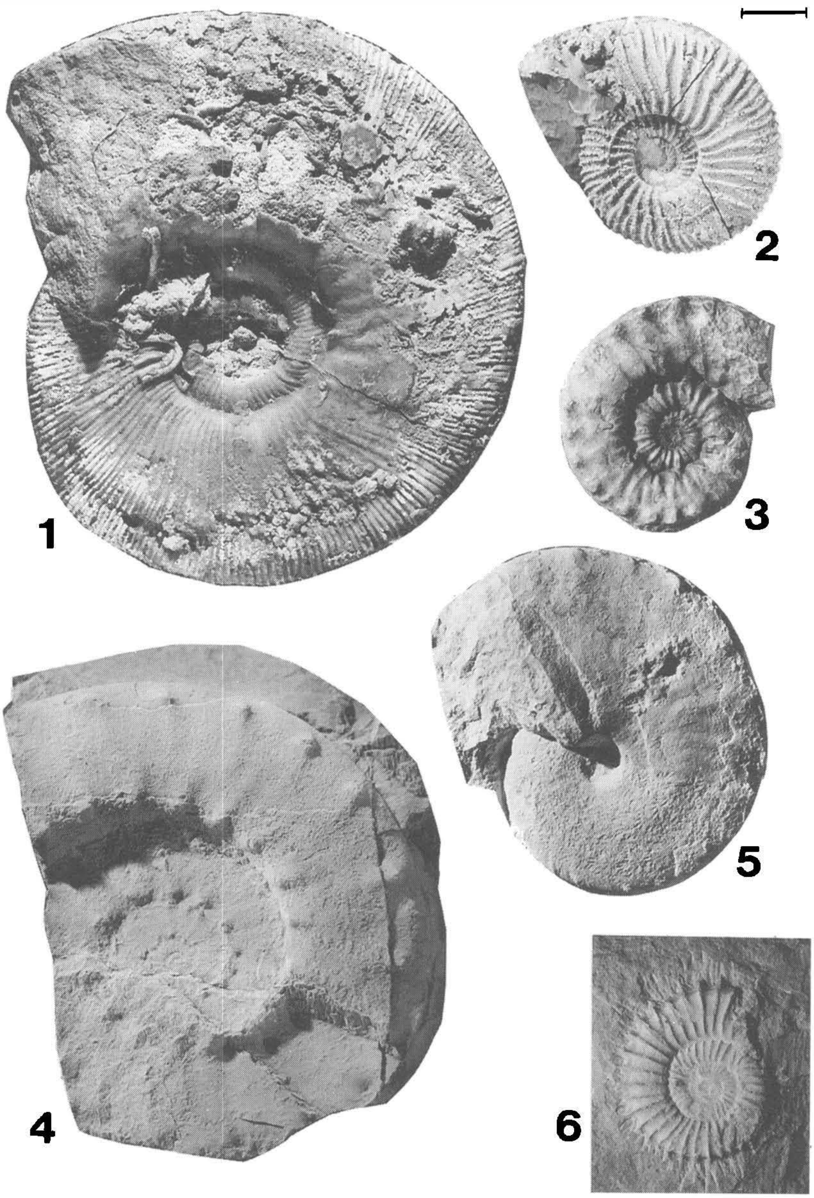

Plate 2. **1**. *Lithacoceras* aff. *ulmense* (Oppel). Ulmense Subzone, *zio-wepferi* horizon β. Nattheim, SW Germany. SMNS No. 62520; **2**. *Sarmatisphinctes* aff. *fallax* (Ilowaisky and Florensky).Ulmense Subzone, *zio-wepferi* horizon β. Gussenstadt, SW Germany. SMNS No. 63647; **3**. *Hybonoticeras* cf. *Extraspinatum* Berckhemer and Hölder. Beckeri Zone, Ulmense Subzone, *hoelderi* horizon, Fridingen, Danube valley, SW Germany. SMNS No. 19427; **4**. *Hybonoticeras interlaevigatum* Berckhemer mskr. Ulmense Subzone, *rebouletianum* horizon. Münsingen Auingen, SW Germany. SMNS No. 14929, x 2/3; **5**. *Neochetoceras rebouletianum* (Fontannes). Beckeri Zone, Ulmense Subzone, *rebouletianum* horizon. Buchheim, SW Germany. SMNS No. 63632; **6**. *Berckhemeria scherzingeri* Schweigert and Zeiss. Hybonotum Zone, Riedense Subzone, *eigeltingense* horizon, Grabenstetten, SW Germany. SMNS No. 19466. Scale bar (except for Fig. 4) 1 cm.

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Immigration of Amoeboceratids into the Submediterranean Upper Jurassic of SW Germany

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Keywords: Ammonites, *Amoeboceras*, Biochronospecies, Paleobiogeography, Late Jurassic, Southern Germany

Abstract: During the Late Jurassic, Southern Germany was part of the Submediterranean Faunal Province. Several ammonite faunal horizons within the rock sequence, however, yield remarkable numbers of amoeboceratids which are of great importance for correlations between the Tethyan Standard Zones and zonal schemes of Subboreal and Boreal areas. Based on the succession of amoeboceratid biochronospecies in Southern Germany, such correlation is possible for the stratigraphic interval between the Middle Oxfordian Transversarium Zone and the Upper Kimmeridgian Acanthicum Zone of the Tethyan standard.

Introduction

Different Upper Jurassic zonal schemes exist for the Tethyan Realm and the more northerly areas of Europe in order to deal with strong bioprovincialism. During the last two decades several efforts have been made to correlate these zonal schemes (Sykes and Callomon, 1979; Matyja and Wierzbowski, 1988). In principle, correlation is possible using Boreal faunal elements which also occur in adjacent parts of the Tethyan Realm. One of the most important ammonite groups for such correlations are the amoeboceratids. While some efforts have been made to analyse the occurrence of amoeboceratids in the Submediterranean Upper Jurassic of E France, N Switzerland and central Poland (Atrops et al., 1993), the exact succession of amoeboceratid biochronospecies in SW Germany remained unknown, although several important species names are based on specimens from there. It is the purpose of this paper to give a brief overview on the state of knowledge in SW Germany (Fig. 1) based on new bed-by-bed collections and additional material from older collections.

Amoeboceratids of the Middle and Late Oxfordian

In the westernmost part of Swabia (Wutach area) and in adjacent parts of Switzerland, the Upper Jurassic begins with a succession of three thin lithostratigraphic members, the "Glaukonitsandmergel" (Cordatum Zone) at the base, followed by the very condensed "Mumienmergel" and an approximately 20 cm thick glauconitic limestone bed in which siliceous sponges accompanied by thrombolithic microbial crusts occur for the first time (Gygi, 1977). The latter, known as "Mumienkalk", is very rich in ammonites which have not yet been described as a whole. In the "Mumienkalk", Submediterranean perisphinctids such as *Subdiscosphinctes elisabethae* (De Riaz) are very common, together with rare macroconchs such as the subzonal index species *Perisphinctes parandieri* (Loriol) (Transversarium Zone, Parandieri Subzone). In contrast to the very condensed and partly reworked glauconitic "Mumienmergel" of the Plicatilis Zone (Antecedens Subzone) below, where cardioceratids are abundant, in the overlying Parandieri Subzone the large majority of ammonites consists of Mediterranean faunal elements: *Ochetoceras canaliculatum* (Buch), *Glochiceras subclausum* (Oppel), *Trimarginites trimarginatus* (Oppel), *Trimarginites arolicus* (Oppel), *Neomorphoceras collinii* (Oppel), *Taramelliceras bachianum* (Oppel), *Taramelliceras semiplanum* (Oppel). Even such typical Tethyan elements as *Gregoryceras riazi* (De Grossouvre), *Sowerbyceras* sp., *Holcophylloceras manfredi* (Oppel) and *Lissoceratoides erato* (d'Orbigny) are recorded from this level; *Gregoryceras riazi* is relatively common at this level (Gygi, 1977). Its intraspecific variability suggests a possible synonymy both with *G. transversarium* (Quenstedt) and with *G. toucasianum* (d'Orbigny). Besides these Tethyan elements, very rarely the first macroconch amoeboceratids occur. They have been identified as *Amoeboceras ilovaiskii* (Sokolov) (Pl. 1.1-2) and *Amoeboceras transitorium* Spath (Pl. 1.3). Due to their rarity, it is uncertain whether these forms are really two different biospecies occurring together, or whether they represent different morphological varieties within a single biochronospecies. Microconch amoeboceratids have not yet been recorded, a striking difference compared to higher levels, where microconchs are predominant. However, correlation is possible with the Glosense Zone of Great Britain, in which identical amoeboceratids occur (Wright, 1996).

Submediterranean Zonation		*Amoeboceras* chronospecies	Subboreal / Boreal Zonations	
Acanthicum		*E. kochi* *E. modestum*	Mutabilis	Kochi
Divisum			Cymodoce	Kitchini
Hypselocyclum				
Platynota	Polygyratus	*A. subtilicaelatum*		
Planula	Galar			
	Planula	*A. haizmanni*	Baylei	Bauhini
Bimammatum	Hauffianum	*A. bauhini*	Pseudocordata	Rosenkrantzi
	Bimammatum	*A. praebauhini*		
	Hypselum	*A. ovale*		Regulare
Bifurcatus	Grossouvrei	*A. alternans*	Cautisnigrae	Serratum
	Stenocycloides			
Transversarium	Rotoides		Pumilus	Glosense
	Schilli			
	Luciaeformis			
	Parandieri	*A. transitorium* *A. ilovaiskii*		

Figure 1. Submediterranean ammonite zonation compared to Subboreal/Boreal zonations. This correlation is mainly based on amoeboceratid biochronospecies from the Upper Jurassic of SW Germany.

In the following beds of the Transversarium Zone, amoeboceratids and other Boreal or Subboreal faunal elements are rare. They reappear in the upper part of the Bifurcatus Zone (not investigated in detail so far) and also at the base of the Bimammatum Zone (Hypselum Subzone, *semimammatum* faunal horizon). At these levels, microconch amoeboceratids are very common, especially adjacent to small sponge bioherms, whereas adult macroconch amoeboceratids are nearly absent. A fragmentary macroconch *Amoeboceras* of the *regulare* group, from the *semimammatum* horizon of Franconia, was named "*Cardioceras*" *neischli* (see Dorn, 1930, Pl. 18, Fig. 28). The microconchs may be identified as *Amoeboceras alternans* (Buch) (Pl. 1-4) or *Amoeboceras ovale* (Quenstedt) (Pl. 1.5). The neotype and lectotype specimens of both morphospecies may come from the same faunal horizon at the foot of the "Lochen" hill near Balingen (Western Swabian Alb), thus representing morphological varieties of the same biochronospecies. It can be noted, however, that specimens with a more coarsely ribbed *alternans* sculpture are more typical in the Bifurcatus Zone. The immigration of amoeboceratids coincides exactly with the initial growth of siliceous sponge mud mounds in the Swabian Jurassic ("Lochen Spongiolithic Limestones"). Another Subboreal influx is demonstrated by the occurrence of the ammonite genus *Ringsteadia* (*R. salfeldi* Dorn) together with its microconch counterpart *Microbiplices/Prorasenia* and the byssate bivalve *Buchia concentrica* (Sowerby). Levels with amoeboceratids are often glauconitic. The glauconitic *semimammatum* horizon with exactly the same ammonite faunal composition is also known from Franconia (Gräfenberg quarry). The following *berrense* faunal horizon of the Hypselum Subzone, mainly characterised by its index *Epipeltoceras berrense* (Favre), contains few ammonites. Only a few records of amoeboceratids are indicated from this level: an unnamed *Amoeboceras* species with bi- and trifurcating secondaries (Pl. 1.6).

In the middle part of the Bimammatum Zone (*bimammatum* faunal horizon) the ammonite fauna of SW Germany reached its greatest diversity during the whole of Oxfordian time. Tethyan faunal elements predominate in this horizon, but amoeboceratids also occur. Again, glauconitic layers are common. Most of the amoeboceratids belong to the microconch species *Amoeboceras praebauhini* (Salfeld) (Pl. 1.8). Just as in Central Poland (Matyja and Wierzbowski, 1988), the variability of this biochronospecies is remarkably high and includes specimens which are transitional to the younger biochronospecies *A. bauhini* (Oppel). This does not mean, however, that *A. bauhini* first appears already in the *bimammatum* horizon, so that its correlation value is well established. The corresponding macroconch of *A. praebauhini* is very close to, or identical with, *Amoeboceras rosenkrantzi* Spath, a species known from Greenland and other Boreal areas. The strange *Amoeboceras transversum* (Quenstedt) (Pl. 1.7, Fig. 2a) is interpreted as being a very coarse ribbed variant of *A. rosenkrantzi* (see Sykes and Callomon, 1979: Pl. 120.1). Adult macroconch amoeboceratids, however, are extremely rare. Other Subboreal elements at this level are *Ringsteadia flexuoides* (Quenstedt) which is close to *R. pseudocordata* Salfeld, and its microconch counterpart which belongs of the genus *Prorasenia*.

Amoeboceratids from the Oxfordian/Kimmeridgian Boundary Beds

In the western part of Swabia the top of the Bimammatum Zone is marked by a further glauconitic level, where ammonites are locally very abundant. In adjacent Switzerland, this bed partly coincides with the "Knollenschicht" of Moesch (see Gygi, 1969). In Swabia it yields the *bauhini* faunal horizon which is mainly characterised by its index species *Amoeboceras bauhini* (Oppel) (Pl. 1.9) together with the Tethyan index *Taramelliceras hauffianum* (Oppel). In the Subboreal succession in Great Britain, *A. bauhini* is indicative for the base of the Lower Kimmeridgian Baylei Zone (Schweigert, 1995, Schweigert and Callomon, 1997) associated with *Pictonia densicostata*. Similar results as in SW Germany have been obtained from central Poland (Matyja and Wierzbowski, 1997). Hence, the overlying Submediterranean Planula Zone falls completely into the Early Kimmeridgian according to the Subboreal zonal scheme. If we accept the traditional definition of the base of the Kimmeridgian in Great Britain as binding on the chronostratigraphic standard time scale, then the Planula Zone of the Submediterranean (Crussolian) standard zonation has to be included within the Kimmeridgian Stage (see Atrops *et al.*, 1998). *Amoeboceras bauhini* is accompanied by its extremely rare macroconch counterpart *Amoeboceras schulginae* Mesezhnikov (Pl. 1.10), but surprisingly another dimorphic pair also occurs, with the macroconch resembling *Euprionoceras* (Pl. 1.11). The best exposure which yields the fauna of the *bauhini* horizon and which is extremely rich in fossils, is the Plettenberg quarry near the town of Balingen.

The Planula Zone is characterised by a Submediterranean fauna comprising mainly perisphinctids (*Subnebrodites*) and oppeliids (*Metahaploceras*). Other ammonite groups occur only sporadically. Amoeboceratids are very rare in the Planula Subzone and at the beginning of the Galar Subzone. Most specimens look somewhat pathologic, with loss both of the keel and sculpture on the body chamber. Fischer (1913) described this strange *Amoeboceras* species as "*Cardioceras haizmanni*" (Pl. 1.12).

Amoeboceratids from Kimmeridgian Beds of SW Germany

In the upper part of the Galar Subzone and in the lowermost faunal horizon of the subsequent Platynota Zone, a last remarkable wave of microconch amoeboceratids swept into the Submediterranean Province, reaching even SE France (Atrops et al., 1993). Some coarser ribbed morphotypes have been misidentified as *A. bauhini* (Oppel) in the past (e. g. specimens like Pl. 1.13), but the retroradiate ribbing style of the latter is a good feature for easy distinction of *A. bauhini* from this younger form. A very fine-ribbed variety coming from this level gives the chronospecific name: *Amoeboceras subtilicaelatum* (Fontannes) (Pl. 1.14). Macroconchs are close to *Amoeboceras bayi* Birkelund and Callomon (Fig. 2b). At higher levels, amoeboceratids are extremely rare in the Swabian Upper Jurassic, and most faunal horizons lack these boreal elements. A few specimens probably belonging to *Euprionoceras* cf. *modestum* Mesezhnikov and Romm (Pl. 1.15) are recorded from a level within the Late Kimmeridgian Acanthicum Zone. The youngest known *Amoeboceras* from Swabia is an immature macroconch of *Euprionoceras kochi* Spath (Pl. 1.16), coming from near the base of the Eudoxus Zone. The ammonite faunas of the Divisum and Acanthicum Zones in SW Germany are far from well-known. Hence, correlations at the base of ammonite faunal horizons are hardly possible, since a high resolution stratigraphy of this interval is still missing.

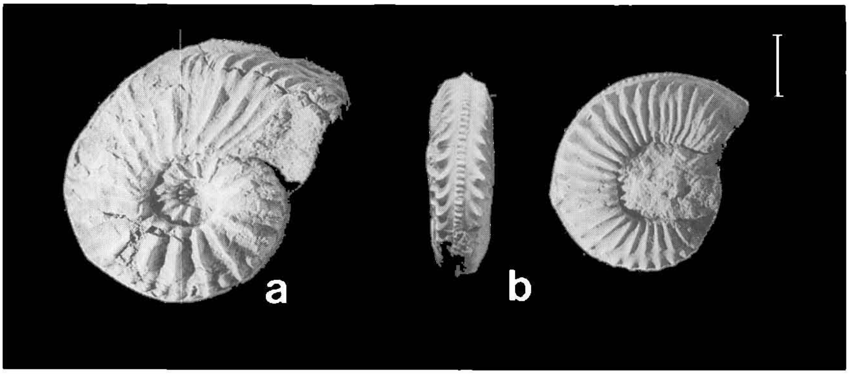

Figure 2. **a.** *Amoeboceras* cf. *rosenkrantzi* (= *Ammonites alternans* in Quenstedt, 1888, Pl. 91.25). Bimammatum Zone, Bimammatum Subzone, *bimammatum* horizon, Lochen area. **b.** *Amoeboceras bayi* Birkelund and Callomon. Platynota Zone, Polygyratus Subzone *sensu* Atrops, *subtilicaelatum* horizon. Aalen-Westhausen, SW Germany. SMNS (Stuttgart) No. 63645. – Scale bar 1 cm.

In Poland and in other Subboreal areas, amoeboceratids occur up to the Late Kimmeridgian Autissiodorensis Zone (Kutek and Zeiss, 1997). During this time, the Subboreal influx in southern Germany diminished, probably because of a change in patterns of marine currents and paleogeographic configurations in the area between Poland and Southern Germany. Thus, no amoeboceratid ammonites have been recorded so far from the late Eudoxus and Autissiodorensis Zones in southern Germany.

Acknowledgments

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References

ATROPS, F., R. GYGI, B. A. MATYJA, AND A. WIERZBOWSKI. 1993. The *Amoeboceras* faunas of the Middle Oxfordian – lowermost Kimmeridgian, Submediterranean succession, and their correlation value. Acta geologica Polonica, 43:213–227.

____, G. MELÉNDEZ, J. OGG, I. PEREZ-URRESTI, K. PAGE, G. SCHWEIGERT, AND J. WRIGHT 1998. The search for the Oxfordian-Kimmeridgian boundary stratotype: Current state-of-progress and the choice for a possible candidate section. 5th International Symposium on the Jurassic System, Vancouver 1998, Abstracts and Programme, 4-5.

DORN, P. 1930. Die Ammoniten des untersten Malm der Frankenalb. Palaeontographica, 73:107-172.

FISCHER, E. 1913. Über einige neue oder in Schwaben bisher unbekannte Versteinerungen des Braunen und Weißen Jura. Jahreshefte des Vereins für vaterländische Naturkunde in Württemberg, 69:31–59.

GYGI, R. 1969. Zur Stratigraphie der Oxford-Stufe (oberes Jura-System) der Nordschweiz und des süddeutschen Grenzgebietes. Beitäge zur Geologischen Karte der Schweiz, Neue Folge, 136:1–123.

____. 1977. Revision der Ammonitengattung *Gregoryceras* (Aspidoceratidae) aus dem Oxfordian (Oberer Jura) der Nordschweiz und von Süddeutschland; Taxonomie, Phylogenie, Stratigraphie. Eclogae geologicae Helvetiae, 70:435–542.

KUTEK, J., AND A. ZEISS 1997. The highest Kimmeridgian and Lower Volgian in Central Poland; their ammonites and biostratigraphy. Acta geologica Polonica, 47:107–198.

MATYJA, B. A., AND A. WIERZBOWSKI. 1988. The two *Amoeboceras* invasions in Submediterranean Late Oxfordian of Central Poland, p. 421–432. *In* R. B. Rocha and A. F. Soares (eds.), 2nd. International Symposium on Jurassic Stratigraphy, Lisboa 1988, 1.

____, AND ____. 1997. The quest for a unified Oxfordian/Kimmeridgian boundary: implications of the ammonite succession at the turn of the Bimammatum and Planula Zones in the Wielu Upland, Central Poland. Acta geologica Polonica, 47:77-105.

QUENSTEDT, F. A. 1888. Die Ammoniten des Schwäbischen Jura. 3. Der Weiße Jura. 817–1140.

SCHWEIGERT, G. 1995. Zum Auftreten der Ammonitenarten *Amoeboceras bauhini* (Oppel) und *Amoeboceras schulginae* Mesezhnikov im Oberjura der Schwäbischen Alb. Jahreshefte der Gesellschaft für Naturkunde in Württemberg, 151:171-184.

____, AND J. H. CALLOMON. 1997. Der *bauhini*-Faunenhorizont und seine Bedeutung für die Korrelation zwischen tethyalem und subborealem Oberjura. Stuttgarter Beiträge zur Naturkunde, Serie B, 247:1-69.

SYKES, R. M., AND J. H. CALLOMON. 1979. The *Amoeboceras* zonation of the Boreal Upper Oxfordian. Palaeontology, 22:839-903.

WRIGHT, J. K. 1996. The *Amoeboceras* faunas of the Upper Calcareous Grit Formation (Jurassic, Upper Oxfordian) of North Yorkshire. Proceedings of the Yorkshire Geological Society, 51:33-43.

Plate 1. **1-2**. *Amoeboceras ilovaiskii* (Sokolov). Transversarium Zone, Parandieri Subzone, *elisabethae* horizon, "Mumienkalk" Bed, Blumberg, SW Germany. **1**. BSPM (Munich) No. 1950 XXX 111; **2**. SMNS (Stuttgart) No. 63643; **3**. *Amoeboceras transitorium* Spath. Transversarium Zone, Parandieri Subzone, *elisabethae* horizon, "Mumienkalk" Bed, Blumberg, SW Germany. SMNS (Stuttgart) No. 63644; **4**. *Amoeboceras alternans* (Buch). Bifurcatus Zone, Lochen Sponge Beds, Lochengründle near Balingen, SW Germany. SMNS (Stuttgart) No. 63640/1; **5**. *Amoeboceras ovale* (Quenstedt). Bimammatum Zone, Hypselum Subzone, *semimammatum* horizon, Lochen Sponge Beds, Lochengründle near Balingen, SW Germany. SMNS (Stuttgart) No. 63641/1; **6**. *Amoeboceras* sp. Bimammatum Zone, Bimammatum Subzone, *berrense* horizon, Impressamergel Formation, former Lochen quarry near Balingen. SMNS (Stuttgart) No. 63638; **7**. *Amoeboceras transversum* (Quenstedt), probably a very coarse ribbed variant of *A*. cf. *rosenkrantzi* (Spath). Bimammatum Zone, Bimammatum Subzone, *bimammatum* horizon. Lochen Sponge Beds, Lochengründle near Balingen, SW Germany. SMNS (Stuttgart) No. 14353; **8**. *Amoeboceras praebauhini* (Salfeld). Bimammatum Zone, Bimammatum Subzone, *bimammatum* horizon, Impressamergel Formation, Mühlheim/Donau, SW Germany. SMNS (Stuttgart) No. 23276; **9**. *Amoeboceras bauhini* (Oppel). Bimammatum Zone, Hauffianum Subzone, *bauhini* horizon, transitional beds between Impressamergel Formation and Wohlgeschichtete Kalk Formation, Plettenberg quarry, SW Germany. SMNS (Stuttgart) No. 62868/5; **10**. *Amoeboceras schulginae* Mesezhnikov. Hauffianum Subzone, *bauhini* horizon, transitional beds between Impressamergel Formation and Wohlgeschichtete Kalk Formation, Lochen quarry, SW Germany. SMNS (Stuttgart) No. 9716; **11**. ? *Euprionoceras* sp. Bimammatum Zone, Hauffianum Subzone, *bauhini* horizon, transitional beds between Impressamergel Formation and Wohlgeschichtete Kalk Formation, Plettenberg quarry, SW Germany. SMNS (Stuttgart) No. 62471; **12**. *Amoeboceras haizmanni* (Fischer). Planula Zone, Planula Subzone, probably *schroederi* horizon, Wohlgeschichtete Kalke Formation, Tuttlingen, SW Germany. SMNS (Stuttgart) No. 63636; **13**. *Amoeboceras subtilicaelatum* (Fontannes), coarsely ribbed variety. Planula Zone, Galar Subzone, *falcula* horizon, Wohlgeschichtete Kalke Formation, road section Mühlheim/Donau – Kolbingen, SW Germany. SMNS (Stuttgart) No. 63642; **14**. *Amoeboceras subtilicaelatum* (Fontannes), finely ribbed variety. Platynota Zone, Polygyratus Subzone, *subtilicaelatum* horizon, Lacunosamergel Formation, SW Germany. SMNS (Stuttgart) No. 63639; **15**. *Euprionoceras* cf. *modestum* Mesezhnikov and Romm. Untere Felsenkalke Formation, Acanthicum Zone, Stettener Tal near Tuttlingen, SW Germany. SMNS (Stuttgart) No. 63634; **16**. *Euprionoceras kochi* Spath. Eudoxus Zone, Untere Felsenkalke Formation, "Paradise cave", Lippachtal near Mühlheim/Donau, SW Germany. SMNS (Stuttgart) No. 63635. Scale bar 1 cm.


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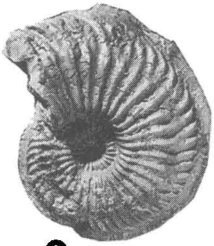
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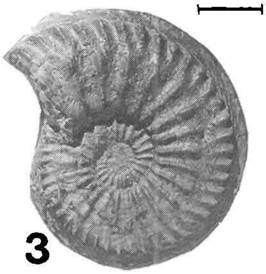
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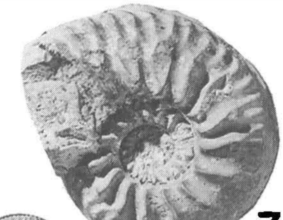
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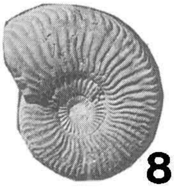
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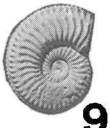
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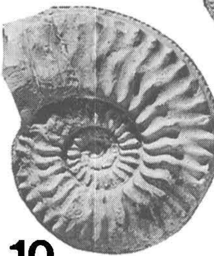
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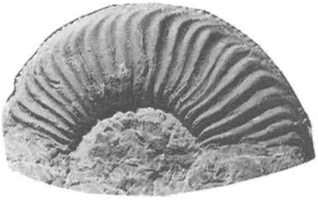
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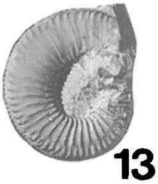
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The Canadensis Zone (Early Jurassic) in the Shoshone Mountains, Nevada

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Keywords: Nevada, Hettangian, Sinemurian, Ammonites, Biochronology

Abstract: The Canadensis Zone in the Shoshone Mountains spans the Hettangian/Sinemurian Stage boundary. *Paracaloceras* cf. *laqueoides* (Hyatt) occurs several meters beneath the Canadensis Zone and offers correlation with the Complanata Subzone, while *Schlotheimia* ex. gr. *stenorhyncha* Lange from the *Metophioceras rursicostatum* beds might support a correlation with the Complanata Subzone. The lower parts of the *Badouxia columbiae* beds are of uncertain stage allocation, while the upper parts of the beds, yielding *Metophioceras* aff. *rotarium* (Buckman) are assigned to the Sinemurian.

Introduction

During his investigation of the Pre-Tertiary stratigraphy of a part of the Shoshone Mountains, Silberling (1959) mapped the Sunrise and Gabbs Formations and provided identifications of ammonoids and bivalves from the section there. He recognized faunas of Hettangian and Sinemurian ages from his stratigraphic units B and C which are now known to be referable in part to the Canadensis Zone (Guex and Taylor, 1976). Taylor (1986; 1990; 1998a; 1998b) has provided additional information on the ammonoids from a section in First Canyon. This section is of special interest because it preserves the most complete record of the Canadensis Zone in Nevada. Ammonoids from the section are illustrated and discussed in the interest of furnishing a means for firmer comparison with Europe.

Stratigraphy

The ammonoid fauna occurs in the Ferguson Hill Member (Taylor et al., 1983), which is the lowest member of the Sunrise Formation. The member corresponds to lithologic units B and C, and perhaps the upper part of Unit A of Silberling (1959). Most of the Sunrise Formation is exposed in the Shoshone Mountains (Fig. 1), although the basal beds in contact with the underlying Gabbs Formation are mostly covered. The upper part of Unit A consisting of dark gray calcareous siltstone might be assignable to the basal part of the Ferguson Hill Member. The lower part of the Ferguson Hill Member is exposed in Milton Canyon, where it is a calcareous siltstone that preserves indeterminate Hettangian ammonoids. Just beneath this is poorly exposed reddish siltstone referable to the Muller Canyon Member. Unit B of Silberling (1959) is about 45m thick and consists of medium-bedded alternating black limestone and mudstone. Unit C is 16m thick and is composed of medium-bedded brown and medium gray bioclastic limestone and siltstone. The contact between the two units is sharp and can be pinpointed by the first gray bioclastic limestone bed. Units B and C were termed the lower and upper divisions of the Ferguson Hill Member respectively (Taylor, 1998a; 1998b) and that terminology is followed here.

The Ammonoid Fauna

The stratigraphic profile (Fig. 2) yields ammonoids referable to the Canadensis Zone and Trigonatum Assemblage (Taylor, 1998b). The Canadensis Zone is divisible into two parts (Fig. 2), the lower *Metophioceras rursicostatum* beds and the upper *Badouxia columbiae* beds (Taylor, 1998b). The Trigonatum Assemblage begins a little over a meter above the highest ammonites referred to the Canadensis Zone.

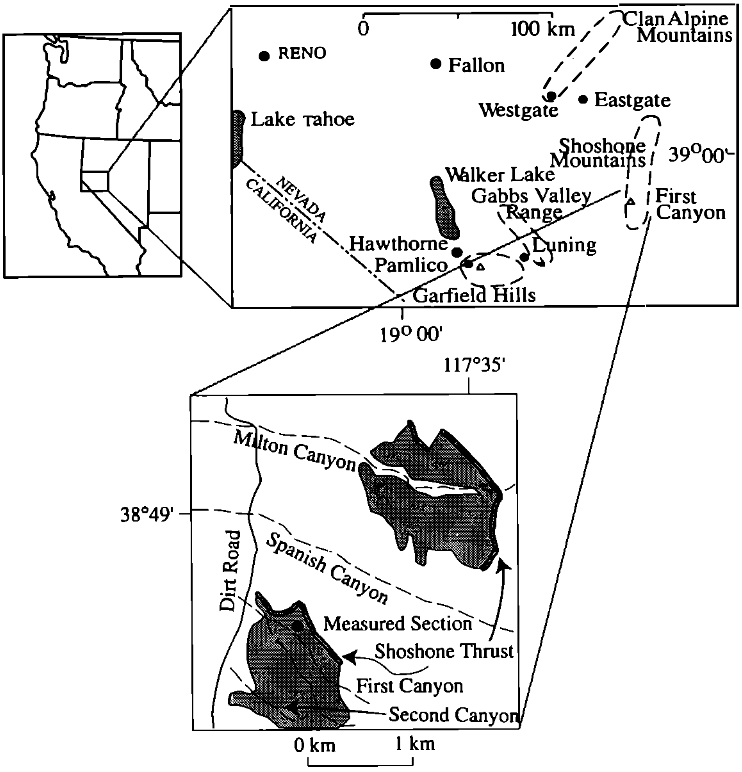

Figure 1. Map giving the location for the stratigraphic section in Late Hettangian and earliest Sinemurian strata in First Canyon, Shoshone Mountains. The shaded areas in the vicinity of Milton and First canyons indicate exposures of the Gabbs and Sunrise Formations.

Plates 1-4 illustrate several of the ammonite species from First Canyon. *Badouxia canadensis* (Frebold) is the most common ammonite in the *Metophioceras rursicostatum* beds, and while this species is not recorded in the stratigraphic section (Fig. 2) in the *Badouxia columbiae* beds (upper division of the Ferguson Hill Member), a float specimen from bioclastic limestone (indicating the upper division) a few hundred meters east of the section suggests that *Badouxia canadensis* ranges that high. A single example of *B.* cf. *occidentalis* (Frebold) (Pl. 2.20-21) occurs above the local range of *Badouxia columbiae* (Frebold).

Metophioceras is common in the Canadensis Zone as well as the overlying Trigonatum Assemblage. Well preserved examples of *Metophioceras rursicostatum* (Pl. 4.1, 4.2, 4.9) occur in the *Metophioceras rursicostatum* beds, while whorl fragments have been found in the *Badouxia columbiae* beds. The upper part of the *B. columbiae* beds yields common but fragmentary specimens of *Metophioceras* aff. *rotarium* (Buckman). Another species of *Metophioceras* (*M.* sp. B) not figured here is much like *Metophioceras* aff. *rotarium* but has closer ribbing and sometimes more evolute coiling. It appears to be a precursor of the abundant *M. trigonatum* (Taylor, 1998b), which is characterized by more rapidly expanding whorls. *Metophioceras* sp. A occurs in the *M. trigonatum* Assemblage. A specimen of *M.* cf. *rougemonti* (Reynès) from the float was recovered from the upper division.

Paracaloceras is also well represented in the Canadensis Zone. *P. retroversicostatus* (Canavari) is common in the *Metophioceras rursicostatum* beds, while a single specimen of *P.* aff. *laqueoides* (Hyatt) was collected from a bed a few meters below the Canadensis fauna. Several forms of *Paracaloceras* also occur in the *B. columbiae* beds, including *P.* cf. *grunowi* (Hauer) (Taylor, 1990), *P. concavum* (Taylor, 1998b), and *P.* n. sp. (Pl. 3.1, 3.2). The latter is distinguished by its coarse ribbing. It is a large species and one example is estimated to have been more than 35cm in diameter. It originally was thought to have been a *Metophioceras* (Taylor, 1990). While outer whorls of *P.* n. sp. may be compressed and subtrigonal, intermediate whorls reveal a broad venter with shallow sulcae typical of *Paracaloceras*. Certain *Paracaloceras* derivatives from the upper

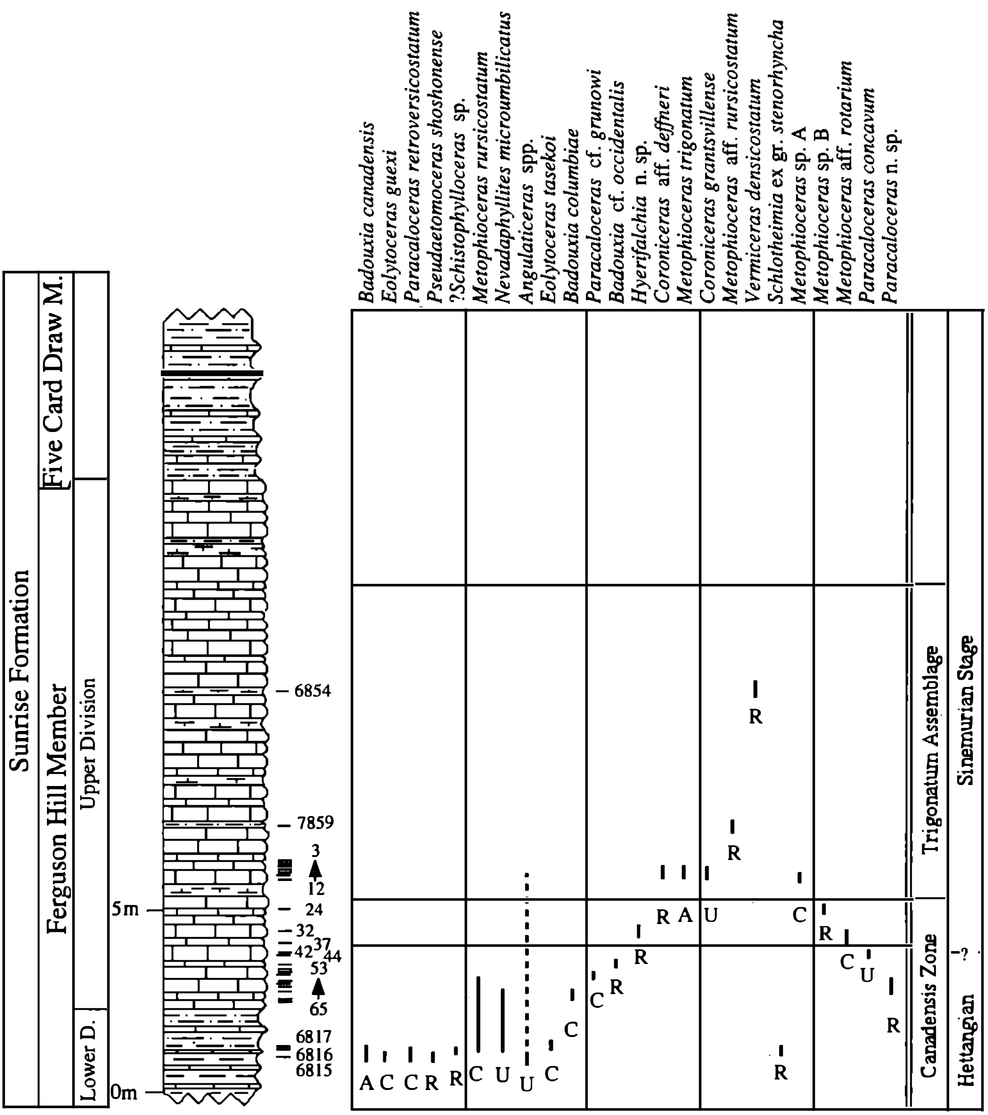

Figure 2. Stratigraphic section showing Late Hettangian and earliest Sinemurian ammonite ranges at First Canyon in the Shoshone Mountains. Ammonite abundances are given as follows: R=1-2 specimens; U=2-5 specimens; C=5-20 specimens; A= greater than 20 specimens.

division include *Hyerifalchia* n. sp. (Pl. 4.6, 4.7) and *Pseudotropites ultratriasicus* (Canavari) (Pl. 3.5-3.8). The latter was found only as float a few hundred meters southeast of the section. *Hyerifalchia* n. sp. (Pl. 4.6, 4.7) differs from the European *H. solitaria* Fucini in having a wider umbilicus and a more compressed whorl section.

Eolytoceras is common in the section where it is restricted to the *Metophioceras rursicostatum* beds. *Eolytoceras tasekoi* is represented in the fauna. While some examples (Pl. 1.1, 1.2, 1.5) are identical to material figured in Frebold (1967), other specimens differ by bearing ventro-lateral nodes (Pl. 1.3, 1.4, 1.6-8). These nodes, however, are very fine and evidently are not present on the few specimens available to Frebold (1967).

Schlotheimiids are rare, and most of the few examples recovered to date are float. *Angulaticeras* ex. gr. *posttaurinus* (Waehner) would suggest a Sinemurian age, although this material (Pl. 1.16-19) was not found *in situ*. One *in situ* example of *Schlotheimia* ex gr. *stenorhyncha* (Lange) occurs in the *Metophioceras rursicostatum* beds (Pl. 1.13-15). The specimen differs from typical *S. stenorhyncha* in having less flexuous ribbing and slightly less forwardly projected ribbing on the venter. Also, ribbing crosses the venter without interruption at the adoral end of the last whorl. The specimen is identical to a Chilean form compared by Hillebrandt (1990) with the alpine species *S. montana* (Waehner). The Nevadan and Chilean specimens do not have the strongly projected ribbing on the venter of *S. montana* and belong to a new species close to *S. stenorhyncha*.

In Nevada the *Metophioceras rursicostatum* beds most likely correlate with the Depressa Subzone (Bloos, 1983) or perhaps the slightly older Complanata Subzone of Late Hettangian age in Europe. Bloos (1994) provided an allocation of *Paracaloceras laqueoides* (Hyatt) to the Complanata Subzone and a correlation with that subzone is, therefore, suggested by the single specimen of *P.* cf. *laqueoides* found several meters below the Canadensis Zone in Nevada. The example of *Schlotheimia* ex. gr. *stenorhyncha* (Lange) suggests a correlation with the uppermost Hettangian, and possibly with the Complanata Subzone.

There is a sharp lithologic contact and possibly a depositional hiatus separating the *Metophioceras rursicostatum* and *B. columbiae* beds in the Shoshone Mountains. Taylor (1990; 1998a; 1998b) suggested that the lower part of the *B. columbiae* beds may be Sinemurian although there is little basis for correlation. However, the occurrence of *Metophioceras* aff. *rotarium* (Buckman) suggests strongly that at least the upper part of the beds is Sinemurian in age. The overlying *Metophioceras trigonatum* Assemblage, yielding *C. grantsvillense* Taylor (1998a; 1998b) (closely allied with *C. hyatti* Donovan) and *C.* aff. *deffneri* (Oppel) is correlated with the lower Sinemurian Rotiforme Subzone.

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References

BLOOS, G. 1983. The Zone of *Schlotheimia marmorea* (Lower Lias) - Hettangian or Sinemurian ? Newsletters on Stratigraphy, 12:123-131.

____. 1994. Frühe Arietitidae (Ammonoidea) aus dem Hettangium (Angulata-Zone, Unt. Lias) von Württemberg (SW-Deutschland). Stuttgarter Beiträge zur Naturkunde, Series B (219) 67 p.

FREBOLD, H. 1967. Hettangian ammonite faunas of the Taseko Lakes area, British Columbia. Geological Survey of Canada Bulletin, 158, 1-35.

GUEX, J. AND D. G. TAYLOR. 1976. La limite Hettangien-Sinémurien, des Préalpes romandes au Nevada. Eclogae Geologicae Helvetiae, 69: 521-526.

HILLEBRANDT, A. v. 1990. The Triassic/Jurassic boundary in northern Chile, p. 27-53. *In* C Ruget and J. Guex (eds.), Rapport du Working Group, "Limite Trias/Jurassique". Comm. Intern. Stratigr. Le Cahiers de L' Université Catholique de Lyon, sér. Sci., 3.

SILBERLING, N. J. 1959. Pre-Tertiary stratigraphy and Upper Triassic paleontology of the Union District, Shoshone Mountains, Nevada. U. S. Geological Survey Professional Paper, 322, 67 p.

TAYLOR, D. G. 1986. The Hettangian-Sinemurian boundary (Early Jurassic): Reply to Bloos 1983. Newsletters in Stratigratigraphy, 16:57-67.

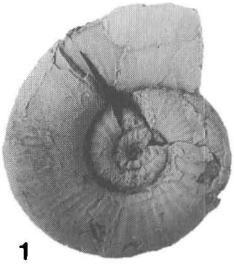
____. 1990. Two species of *Paracaloceras* from the Canadensis Zone (Hettangian-Sinemurian stages) in Nevada (USA). Bulletin de géologie, Lausanne, 309:211-219.

____. 1998a. Nevada field trip guide for the Gabbs and Sunrise Formations in west-central Nevada. Fifth International Jurassic Symposium, University of British Columbia, Canada, 20 p.

____. 1998b. Late Hettangian - Early Sinemurian (Jurassic) ammonite biochronology of the western Cordillera, United Sates. Geobios, 31 (4):467-497.

____, SMITH, P. L., A. LAWS, and J. GUEX. 1983. The stratigraphy and biofacies trends of the Lower Mesozoic Gabbs and Sunrise Formations, west-central Nevada. Canadian Journal of Earth Sciences, 20:1598-1608.

Plate 1. **1-10**, *Eolytoceras tasekoi* Frebold; **1,2**. NWMNH no. 25441, loc. 4045; **3,4**. NWMNH No. 25442, loc. 7045, note ventro-lateral nodes; **5**, NWMNH No. 25443, loc. 6861; **6**. NWMNH No. 25444, loc. 7045, note ventro-lateral nodes; **7,8**. NWMNH NO. 25445, loc. 6817, note ventro-lateral nodes; **9,10**. NWMNH No. 25446, loc. 7046, at adapical end of shell constriction on internal mold only; **11,12**. *Angulaticeras* sp., NWMNH No. 25447, loc. 7045, float from upper division; **13-15**. *Schlotheimia* ex gr. *stenorhyncha* Lange, NWMNH No. 25448, loc. 6815; **16-19**. *Angulaticeras* ex. gr. *posttaurinus* (Waehner); **16,17**. NWMNH No. 25449, 7045, float from lower division; **18,19**. NWMNH no. 25450, float from upper division. Bar scales = 1cm; Scale at upper right is for figures 1-8; 11-19; the other scale is for figures 9, 10.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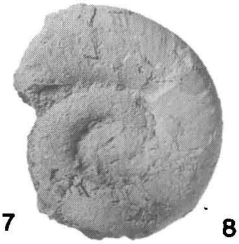
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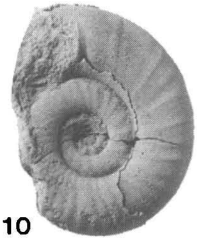
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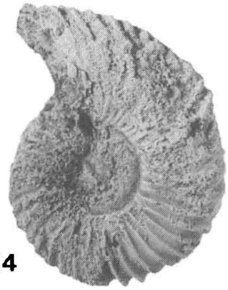
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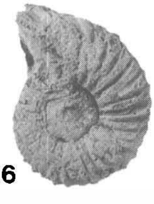
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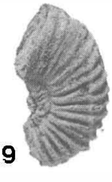
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Plate 2. **1-6**. *Badouxia columbiae* (Frebold). **1,2**. NWMNH No. 25451, loc. 7046; **3,4**. NWMNH No. 25452, upper part of bed 60, involute specimen; **5,6**. NWMNH No. 25453, loc. 7045; **7-19**. *Badouxia canadensis* (Frebold); **7,8**. NWMNH No. 25454, loc. 6816; **9**. NWMNH No. 25455, loc. 6815; **10,11**. NWMNH No. 25456, loc. 6816; **12,13**. NWMNH No. 25457, loc. 7045; **14,15**. NWMNH No. 25458, loc. 7046; **16,17**. NWMNH No. 25459, loc. 6815; **18,19**. NWMNH No. 25460, loc. 6815; **20,21**. *Badouxia* cf. *occidentalis* (Frebold), NWMNH No. 25461, bed 51. Bar scale for all figures = 1cm.

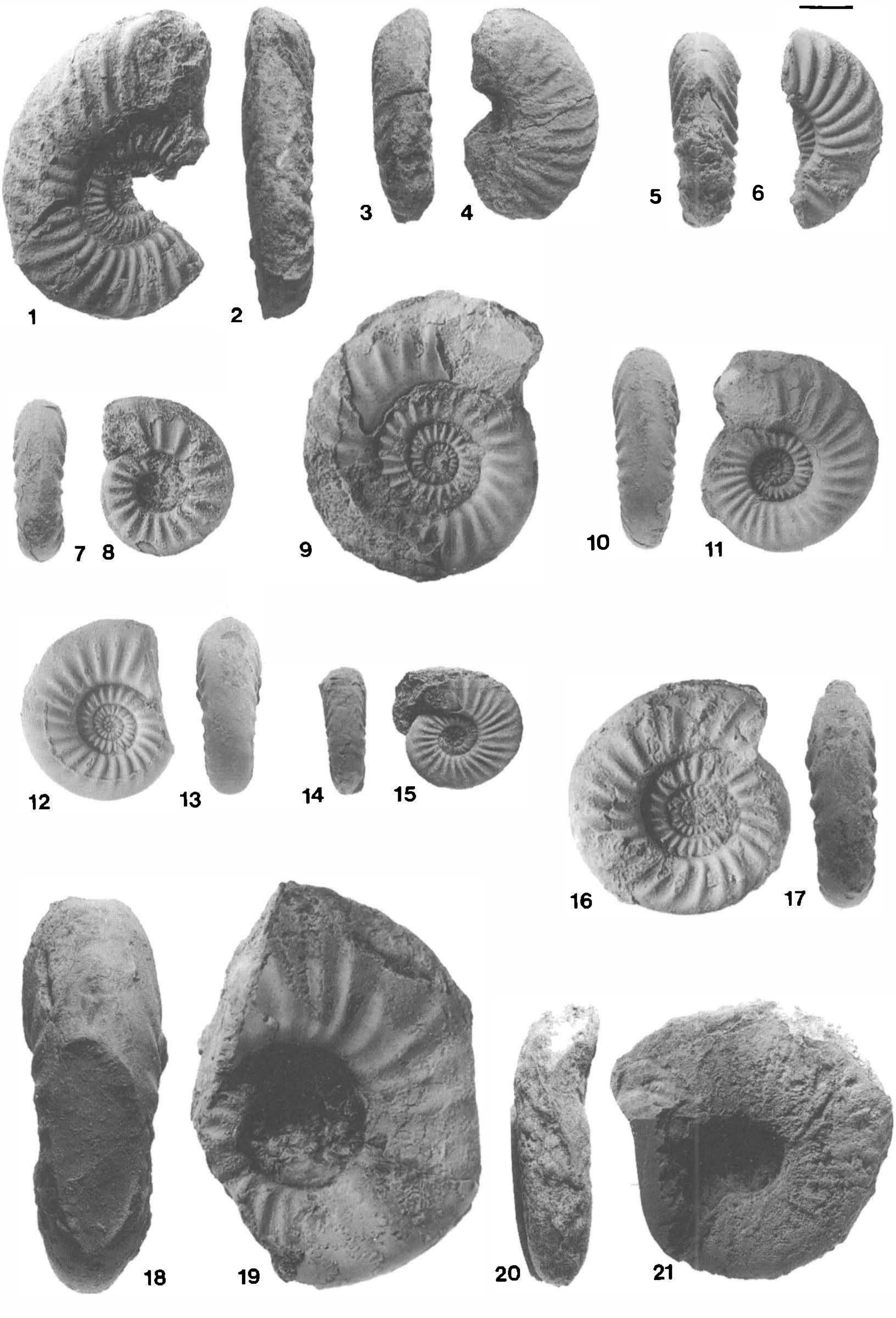
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Plate 3. **1,2**. *Paracaloceras* n. sp. NWMNH No. 25462, bed. 53; **3,4**. *Paracaloceras* aff. *laqueoides* (Hyatt), NWMNH No. 25463, loc. 6/90; **5-8**. *Pseudotropites ultratriasicus* (Canavari); **5,6**. NWMNH No. 25464, float from upper division, body chamber, note chorded keel and strongly depressed whorl section; **7,8**. NWMNH No. 25465, float from upper division; **9,10**. *Metophioceras* sp. A, NWMNH No. 25466, loc. 7446; **11,12**. *Paracaloceras* n. sp. NWMNH No. 25467, bed 54, note slightly subtrigonal outer whorl. Bar scale for all figures = 1cm.


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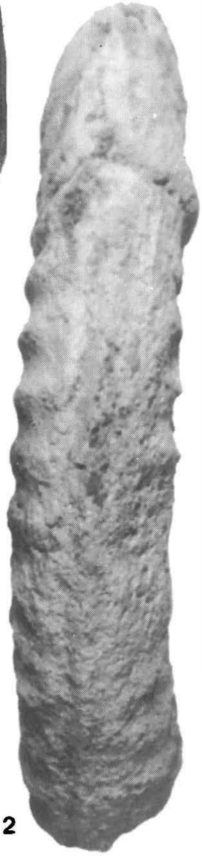
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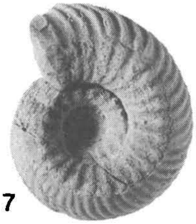
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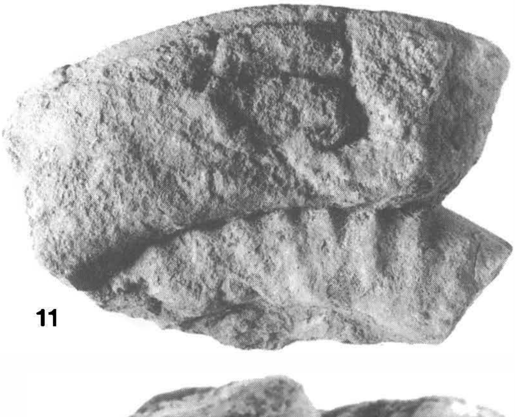
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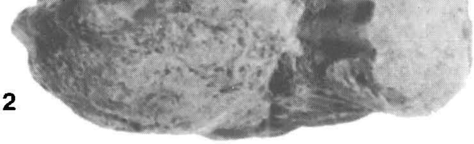
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Plate 4. **1**,**2**,**9**. *Metophioceras rursicostatum* (Frebold); **1**,**2**. NWMNH No. 25468, loc. 6816, evolute specimen; **9**. MWMNH No. 25469, loc. 6816, specimen with rapidly enlarging whorls; **3-5**. *Metophioceras* aff. *rotarium* (Buckman); **3**,**4**. NWMNH No. 25470, bed 37a; **5**. NWMNH No. 25471, bed 37b; **6**,**7**. *Hyerifalchia* n. sp., NWMNH No. 25472, bed 32; **8**. *Pseudaetomoceras shoshonense* Taylor, NWMNH No. 25473, loc. 6815. Bar scales = 1cm; Scale in upper right is for figures 1-5, 8, 9; scale at lower left of plate is for figures 6, 7.

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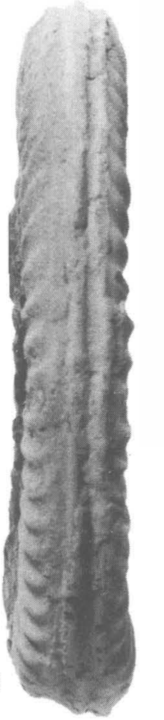

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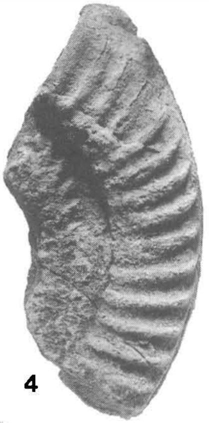

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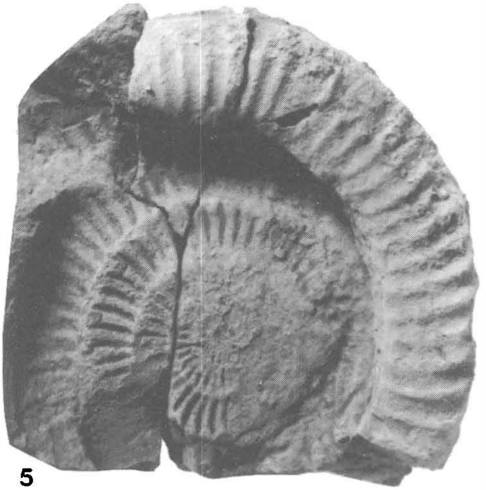

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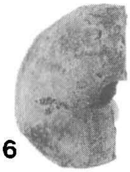

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The Triassic/Jurassic System Boundary in the Gabbs Formation, Nevada

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Keywords: Triassic, Rhaetian, Hettangian, Nevada, Ammonites, Bivalves, Geochemistry, Extinction

Abstract: The Gabbs Formation reveals a comparatively complete record of marine sedimentation across the Triassic/Jurassic boundary. The ammonoid succession is more complete than elsewhere because it yields *Choristoceras crickmayi* stratigraphically above beds with *C. marshi.* In addition, a Jurassic sequence beginning with the *Psiloceras tilmanni* zone followed by the *C. minutum* zone occurs in the same stratigraphic profile and is below the first unquestionably Jurassic *P. pacificum* zone fauna (a correlative of the basal Jurassic Planorbis Zone). The terminal Triassic extinction may have had at least 2 phases, one marked by low diversity Rhaetian ammonoid faunas, followed by an abrupt ammonite and bivalve extinction at the top of the Crickmayi Zone. The latter might have been induced by an environmental disturbance superimposed on faunas already under duress.

Introduction

Muller and Ferguson (1936; 1939) recognized that the Gabbs and Sunrise Formations in west central Nevada yield rich Late Triassic and Early Jurassic faunas and that the beds near the contact between the formations reveal an exceptional record of marine sedimentation across the boundary between the two Systems. Because of its excellent representation of molluscan faunas, and its relatively complete faunal record, the section in the Gabbs Valley Range at New York Canyon (Fig. 1) was proposed by Taylor et al. (1983) to serve as the System boundary stratotype. They suggested positioning the System boundary at the top of the Gabbs Formation, above the highest occurrence of *Choristoceras* and below beds most likely correlative with the Planorbis Zone. Guex et al. (1997; 1998) again proposed that the section serve as the stratotype and provided new range data.

This paper examines macroinvertebrates, lithofacies, and geochemistry of the Gabbs Formation to facilitate interpretation of the state of the faunas leading to the final extinction event. It also provides new data to assist discussion of the Nevada section as a candidate for the System boundary stratotype. Emphasis is placed on the Muller Canyon Member at the top of the Gabbs Formation, where the boundary occurs.

Stratigraphy

The Gabbs Formation has three members (Fig. 2) described by Taylor et al. (1983) and Taylor (1998). The Nun Mine, composed of siltstone and limestone, is the lowest member while the Mount Hyatt is the middle member. The latter has a lower division composed of bioclastic limestone and an upper division with finer-grained siltstone and limestone. The upper member of the Muller Canyon Member furnishes the boundary beds. It is a thin- to medium-bedded, weakly calcareous siltstone ranging up to about 15m in thickness. The basal contact is above a calcareous, sandy siltstone bed while the top is just below the first limestone bed of the overlying Ferguson Hill Member of the Sunrise Formation (Fig. 3, 4).

The member in New York Canyon has two divisions. The lower division at the type locality for the member is 8.2m thick (Fig. 3), weathers to a reddish hue, and is composed of thin- to medium-bedded, soft siltstone and weakly resistant siltstone. This division preserves common small-scale cross lamination and some parallel lamination. The upper division weathers reddish gray, is 6m thick, and is a thin- to medium-bedded, calcareous siltstone and soft siltstone unit. The upper division has more frequent calcareous beds and common fine-scale cross lamination. The boundary between the two divisions is sharp and apparently conformable.

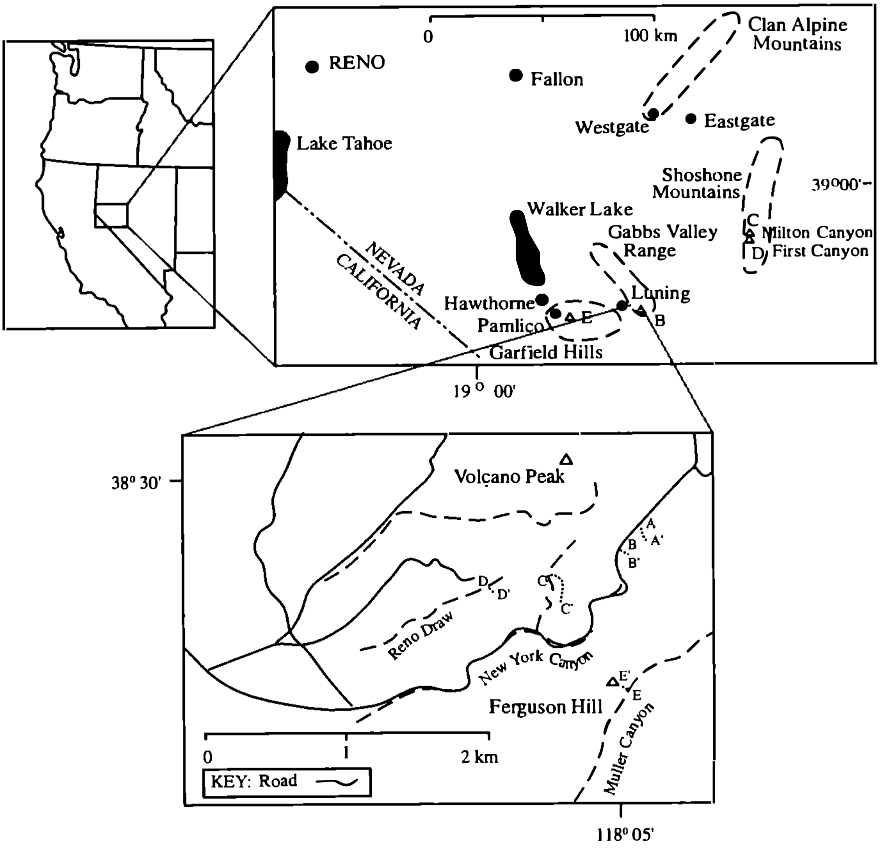

Figure 1. Showing locations of measured sections of Muller Canyon Member, Gabbs Formation.

Biochronology

The ammonoid succession in the Gabbs Formation includes the uppermost Triassic Amoenum and Crickmayi zones. The Amoenum Zone occurs in the Nun Mine Member, while the Crickmayi Zone occurs in the Mount Hyatt to basal part of the Muller Canyon Member (Fig. 2).

The Crickmayi Zone has five horizons which include: (1.) *Vandaites* sp. in the basal beds of the Mt. Hyatt Member, (2.) *Cycloceltites* sp. stratigraphically higher but in the lower part of the member, (3.) *Choristoceras rhaeticum* near the middle part of the member, (4.) *Choristoceras marshi* in the upper few meters of the Mount Hyatt (but not in the uppermost bed), and (5.) *Choristoceras crickmayi* and *Choristoceras* sp. A around the boundary between the Mt. Hyatt and Muller Canyon Members.

The *Psiloceras tilmanni* zone known previously from Peru and Chile (Hillebrandt 1988, 1994) is now recognized from the top of the lower division of the Muller Canyon Member where the name-bearer for the zone is found along with *P. spelae* and "*Juraphyllites*" sp. (Fig. 3, bed 25). This fauna was described recently in Guex et al. (1998).

The *C. minutum* zone, yielding *Psiloceras* and *Choristoceras* (Guex, 1995), occurs in the upper meter of the Muller Canyon Member. Immediately superjacent to the *C. minutum* fauna are typically Hettangian faunas referable to the *Psiloceras pacificum* zone of Guex (1995). Although the System boundary had been indicated between the *C. minutum* and *P. pacificum* zones (Guex 1995; Taylor et al., 1983), a viable alternative is its placement in the beds between the Crickmayi and *P. tilmanni* zones.

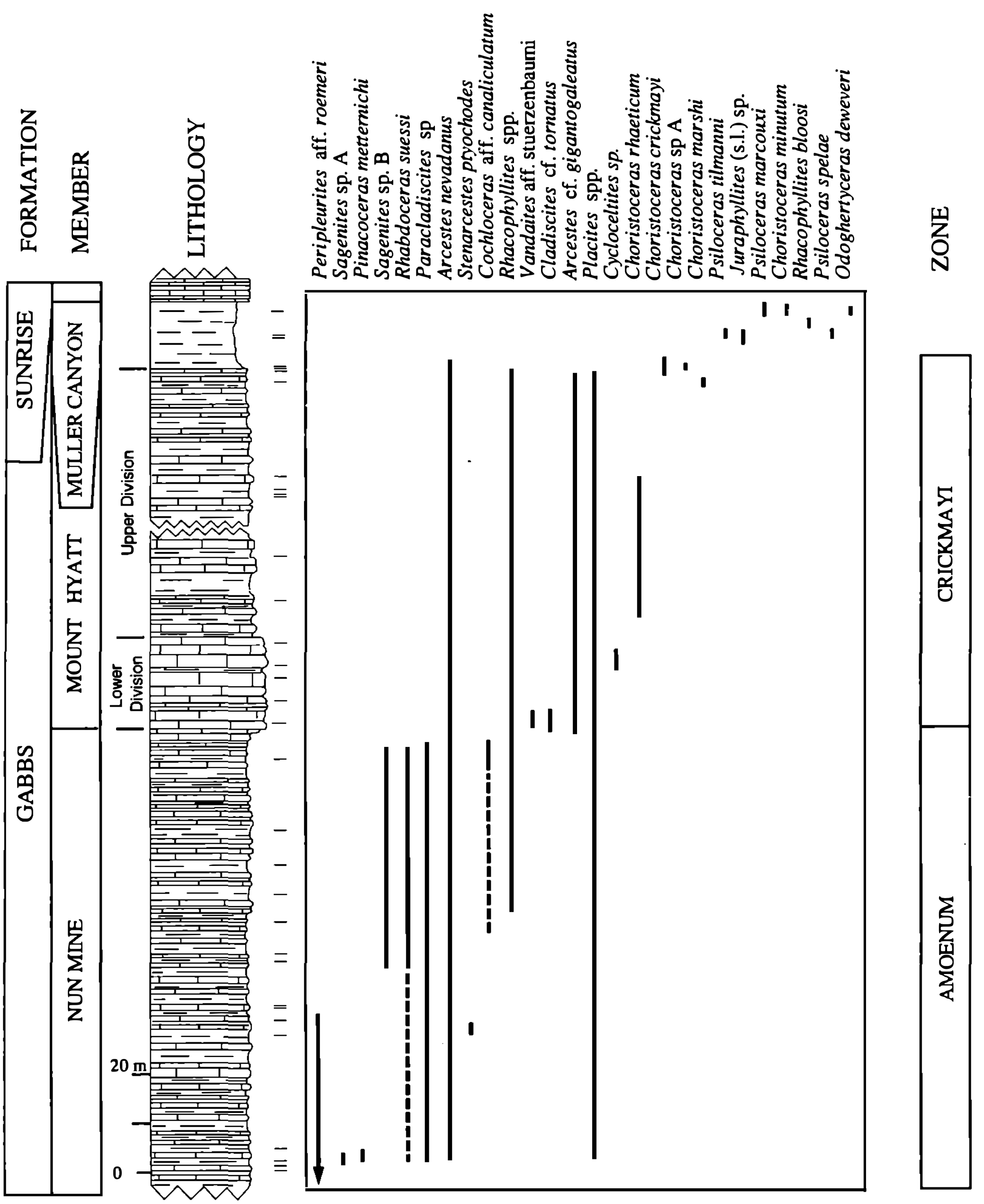

Figure 2. Composite stratigraphic section and ammonite ranges in the Gabbs Formation, New York Canyon. The section is a composite of sections A-E for which locations are shown in Figure 1.

The Fauna from the Boundary Beds

The Gabbs Formation yields abundant bivalves and common ammonites and crinoidal debris. Other macroinvertebrates include coleoids, which are common in the basal beds of the Mount Hyatt Member, uncommon nautilids that occur throughout the section, occasional gastropods, and terebratulid brachiopods. Microinvertebrates include conodonts, known as high as the base of the Mount Hyatt Member (M. Orchard, pers. comm.) and ostracods (A. Lord, pers. comm.). Ichthyosaur material occurs principally near the top of the Nun Mine Member and basal part of the Mount Hyatt Member. A single tooth was recovered from the top of the Mount Hyatt Member. Plate 1 figures several species from the boundary beds. Molluscan faunal associations from the Gabbs Formation at New York Canyon (Laws, 1982) include the *Cochloceras* Association in the Nun Mine Member, the *Plicatula* Association in the basal beds of the Mount Hyatt Member, the *Tutcheria* Association in the lower division of that member just above the latter association, and the *Nuculoma* Association in the upper division of the Mount Hyatt Member and the basal beds of the Muller Canyon Member. The "*Curionia*" Association occurs at the base of the upper division of the Muller Canyon Member.

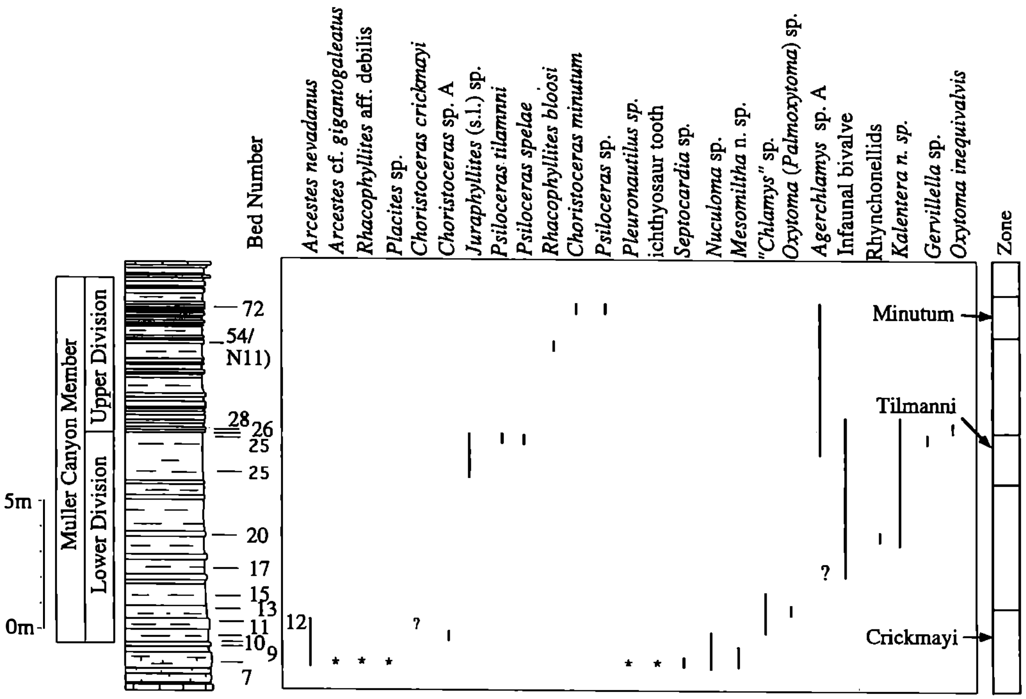

Figure 3. Stratigraphic section for the Muller Canyon Member at its type locality, Ferguson Hill (section E), New York Canyon. Asterisk denotes float material.

The uppermost Mount Hyatt and basal Muller Canyon members yield the *Nuculoma* Association, where the non-siphonate, suspension-feeding bivalve *Septocardia* and ? *Mesomiltha* (Pl. 1.1-5) are common and the deposit-feeding "*Nuculoma*" is abundant. The association persists only into the basal beds of the Muller Canyon Member, and *Arcestes*, *Choristoceras*, and *Rhacophyllites* are the only "Triassic" ammonoids ranging as high as the member (Figs. 3, 4). There are sporadic occurrences of macroinvertebrates from beds higher in the lower division. In the uppermost 15-20cm the pelecypod *Gervillella* (Pl. 1.6) was found along with a possible pectinacean in the *Psiloceras tilmanni* zone. The "*Curionia*" Association occurs near the base of the upper division of the Muller Canyon Member. It yields *Kalentera* (=*Curionia* in Laws, 1982), *Agerchlamys*, *Oxytoma* (Pl. 1.7), and ? *Mesomiltha* (Pl. 1.7). Specimens commonly have valves associated or articulated. Pectinaceans and poorly preserved shallow infaunal bivalves are the dominant benthic elements higher in the upper division and the ammonoid *Rhacophyllites bloosi* was found 2.3m below the *C. minutum* Zone (Guex et al., 1998).

In support of an anoxic event in the basal Hettangian, Hallam (1990) reported that the base of the Ferguson Hill Member lacked benthic invertebrates. These beds, in fact, reveal an abundant bivalve fauna dominated by shallow, infaunal forms. Most of these bivalves are rather small (but not dwarfed), and most commonly are less than a centimeter in length. Bivalves generally increase in size upwards through the Hettangian section so that in the comparatively shallow-water, basal Sinemurian beds large, thick-shelled forms including *Cardinia* and *Weyla* are prominent.

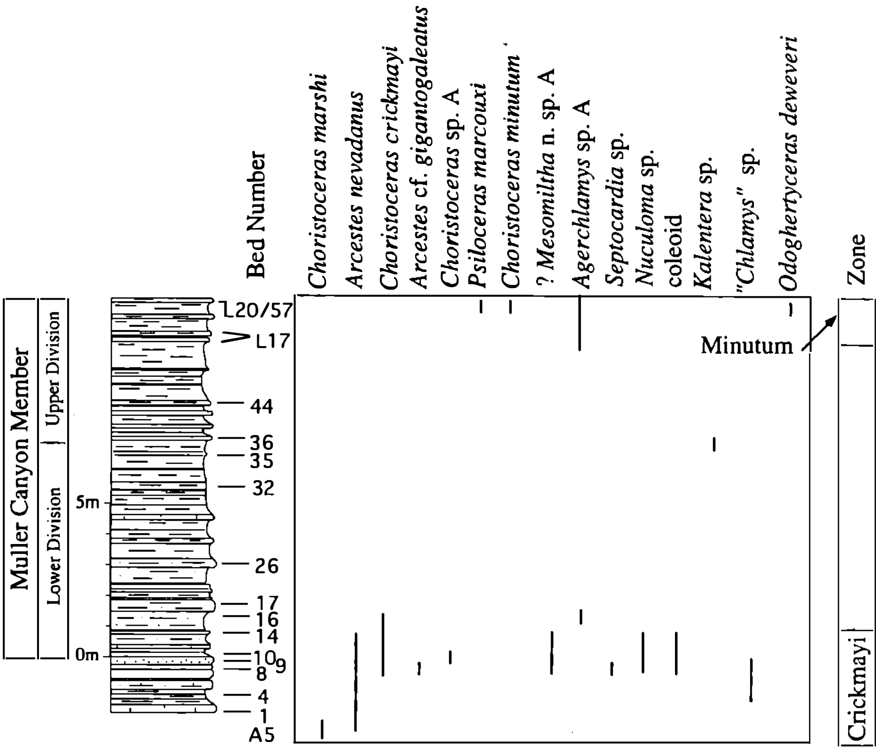

Figure 4. Stratigraphic section, Muller Canyon Member, Reno Draw (section C), New York Canyon.

Discussion

The ammonite fauna from the Gabbs Formation is a low-diversity one, having only 14 genera over a span of two zones. The maximum number of species that co-occur at a given locality is seven (in the upper part of the Nun Mine Member), while localities yield on the average 1-3 species. The formation should yield highly diverse associations, being carbonate-rich and including excellent representation of Composite Assemblages C through D (the optimum range for ammonoids in the Western Cordillera; Taylor, 1982). Tozer (1979) also documented the low global generic diversity of latest Triassic ammonoids.

The ammonite fauna is also remarkable for its few morphotypes. Smooth, involute, compressed shells are represented by *Placites* and *Pinacoceras*; weakly ornate, involute and weakly compressed to globose shells are represented by *Arcestes*, *Stenarcestes*, *Cladiscites*, *Paracladiscites*, and *Sagenites*; midvolute, smooth, compressed shells are represented by *Rhacophyllites*; and serpenticones and heteromorphs are represented by *Peripleurites*, *Choristoceras*, *Vandaites*, *Rhabdoceras*, *Cycloceltites*, and *Cochloceras*. The heteromorphs stand out as the group with the most conspicuous ribbing. Remarkable is the lack of midvolute, ornate ammonoids which normally dominate faunas both in numbers of individuals and generic diversity. Only the heteromorphs show rapid turnover rates, and were still diversifying at the species level until the top of the Rhaetian. There may have been a mutually exclusive relationship between heteromorph and ornate midvolute forms during this time period. The heteromorphs seem to have flourished in terms of numbers of individuals and species during the Rhaetian (Fig. 2). It is noteworthy that the last *Choristoceras* (in the *C. minutum* zone) persists shortly above the first common *Psiloceras*. Thereafter the heteromorphs disappeared, not to be seen again until Middle Jurassic time.

We propose that the low ammonoid diversity and lack of ornate, midvolute ammonoids may indicate communities that were under ecological stress. If this is the case, then an environmental stress leading to the terminal Triassic extinction event may have influenced faunas beginning in the Cordilleranus Zone. The ammonoids hold a comparatively high trophic position in Mesozoic marine invertebrate communities, and as a result may be one of the first groups to show a response to environmental deterioration. Nevertheless, stratigraphic distribution of the ammonoids reveals that several of the Rhaetian ammonoid genera persist to the top of the Stage, where they drop out suddenly. Only *Rhacophyllites* and *Choristoceras* persist a short distance up-section in beds regarded herein to be earliest Jurassic in age. The overall ammonoid pattern suggests, therefore, that the ammonoid communities were deteriorating over a period of time in the Late Triassic, and that this deterioration culminated in, or was punctuated by, an abrupt extinction event at the end of the Rhaetian.

Other taxa reveal trends indicating a sudden extinction at the top of the Rhaetian Stage. There are about 8 nautilid genera in the Gabbs Formation. Most are rare or uncommon, but 2 genera have been observed in the uppermost Mt. Hyatt to basal Muller Canyon members. Other taxa representing lower trophic levels apparently are not markedly affected until the top of the Rhaetian. Laws (1982) commented that the diversity of the bivalve-dominated associations within the Gabbs Formation is not demonstrably low for Mesozoic communities. Also, Carter (1993) documented that Late Triassic radiolaria remained diverse until the end of the Rhaetian.

Litho- and bio-facies suggest that the Muller Canyon Member was transgressive. The rich benthic bivalve-dominated fauna from the top of the Mt. Hyatt Member and basal few beds of the Muller Canyon Member, along with locally common ammonoids, indicate an assignment of that interval to Composite Assemblage C (Taylor, 1982; Taylor et al., 1983). Presence of diverse cephalopods (ammonoids, coleoids, and nautilids) suggests normal marine salinities. Macrofauna becomes sparse in the fine-textured and calcite-poor lower part of the Muller Canyon Member above its basal meter or two, and it probably represents the most transgressive part of the section, assignable to the outer part of Composite Assemblage C to inner part of D. The presence of fairly common small bivalves and ammonoids in bed 25 at the top of the lower division signifies the beginning of a minor regression. The regression is marked notably by the influx of an abundant macrofauna at the base of the upper division. The upper division, along with the lower several meters of the Ferguson Hill Member, indicate the outer part of Composite Assemblage C.

Abundant secondary gypsum, a low diversity ichnofauna, abundant fine-scale cross lamination, some parallel lamination, comparatively fine sediment texture, and paucity of macrofauna might suggest that parts of the lower division of the Muller Canyon Member, in particular, were deposited under dysaerobic conditions. To investigate this problem, we examined thorium and uranium concentrations using Instrumental Neutron Activation Analysis on samples from the Gabbs

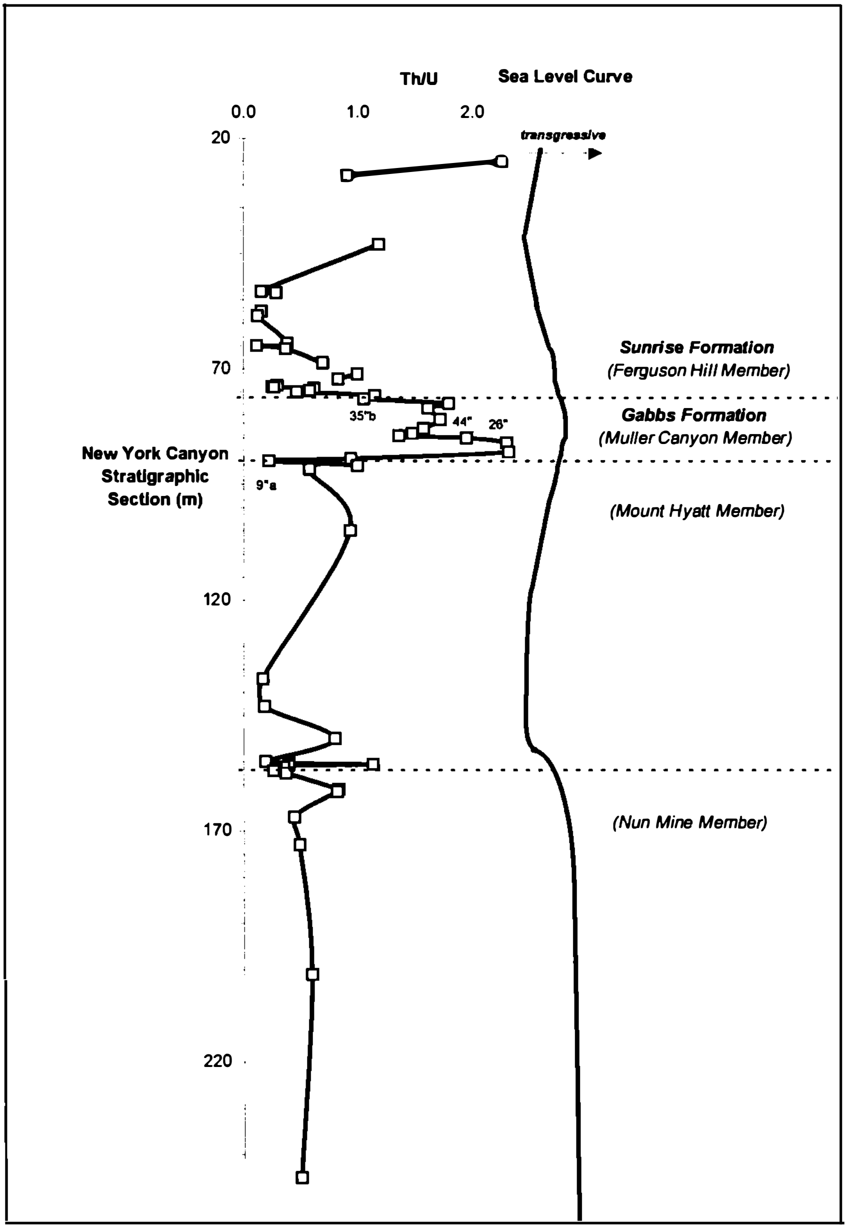

Figure 5. Th/U geochemical signature for the Gabbs and lower part of the Sunrise Formations.

Formation and lower part of the Sunrise Formation (Fig. 5). Th/U ratios are sometimes used as an indication of water oxygen content, with values corresponding to the degree of oxygenation. Thorium is found primarily in the lithogenic portion of the sediment (Pliler and Adams, 1962), while U is removed from sea water by diffusion across the sediment-water interface of organic-rich sediments (Klinkhammer and Palmer, 1991). There is a significant increase in U enrichment in suboxic/anoxic sediments during times of low oxygen bottom water levels and/or high surface water productivity (Klinkhammer and Palmer, 1991). The lower Th/U ratios may indicate poorly oxygenated environments, while the higher values would suggest more oxygenated conditions (Fig. 5). The offshore Nun Mine Member generally has low Th/U ratios, and there are some anomalously low values in the lower part of the Mount Hyatt which preserves a rich benthic invertebrate fauna indicating oxygenated conditions. In general, Th/U values are low in the lower part of the Ferguson Hill Member, where the bioturbated facies preserves a modest diversity bivalve fauna including common infauna. There are comparatively high values in the upper part of the Ferguson Hill Member, which has shallow water facies and a diverse invertebrate fauna characterized by large and ornate bivalves. This fauna indicates well oxygenated conditions.

The highest Th/U ratios occur in the Muller Canyon Member, where facies and fauna might suggest reducing conditions. The higher thorium concentrations in the Muller Canyon Member compared to the rest of the section may result largely from the relatively greater proportion of siliciclastic material in these rocks. In addition, uranium concentrations in the Muller Canyon are on average lower than in the rest of the section, perhaps indicating a change in available organic material. This leaves the question of the oxygen-levels during deposition of the Muller Canyon Member equivocal.

In summary, faunal distributions in the Gabbs Formation indicate an abundant and diverse assemblage of typically Triassic faunas that persist up to the base of the Muller Canyon Member, and suggests that the ensuing extinction event was an abrupt one. The major faunal turnover, based on ammonoids, is between the Crickmayi and *P. tilmanni* zones. Taking into consideration all of the available macrofauna, the major episode of turnover in the type section at Ferguson Hill is narrowed down to a 2.5m interval (between beds 15 and 20). The evidence regarding the extinction suggests a major perturbation that affected numerous invertebrate groups more or less synchronously, such as the ammonoids, nautilids, and bivalves and by extrapolation from the data provided by Carter, microfauna including the radiolaria. The decimation of these taxa indicates a catastrophic global perturbation having a major impact on biotic productivity.

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References

CARTER, E. S. 1993. Biochronology and Paleontology of uppermost Triassic (Rhaetian) radiolarians, Queen Charlotte Islands, British Columbia, Canada. Mémoires de Géologie (Lausanne), Switzerland. No. 11, 175 p.

GUEX, J. 1982. Relations entre le genre *Psiloceras* et les Phylloceratida au voisinage de la limite Trias-Jurassique. Bulletin de Géologie Lausanne, 260:47-51.

____. 1995. Ammonites hettangiennes de la Gabbs Valley Range (Nevada, USA). Mémoires de Géologie (Lausanne).

____, M. RAKUS, D. TAYLOR, AND H. BUCHER. 1997. The Triassic-Jurassic boundary: Proposal for a stratotype defined in the Gabbs Valley Range (Nevada). International Subcommission on Jurassic Stratigraphy, Newsletter 24:26-30.

____, D.TAYLOR, M. RAKUS, AND H. BUCHER. 1998. Deux nouveaux genres et quatre nouvelle espèces d'ammonites (Cephalopoda) du Lias inférieur. Bulletin de Géologie Lausanne, 339:73-85.

HALLAM, A. 1990. The end-Triassic mass extinction event, p. 577-583. *In* V. L. Sharpton, and P. Ward. (eds.), Global Catastrophes in Earth History; An interdisciplinary conference on impacts, volcanism, and mass mortality. Geological Society of America Special Paper 247.

HILLEBRANDT, A. VON. 1988. Ammonite biostratigraphy of the South American Hettangian, Description of two new species of *Psiloceras*. 2nd International Symposium on Jurassic Stratigraphy, Lisbonne, 1:55-70

____. 1994. The Triassic/Jurassic boundary and Hettangian biostratigraphy in the area of the Utcubamba Valley (Northern Peru). Geobios, 17:297-307.

KLINKHAMMER, G. P. and M. R. PALMER. 1991. Uranium in the oceans: Where it goes and why. Geochimica et Cosmochimica Acta, 55:1799-1806.

LAWS, R. A. 1982. Late Triassic depositional environments and molluscan associations from west central Nevada. Palaeogeography, Palaeoclimatology, Palaeoecology, 37:131-148.

MULLER, S.W. and H. G. FERGUSON. 1936. Triassic and Lower Jurassic formations of west central Nevada. Geological Society of America Bulletin, 47:241-252.

_____ and _____. 1939. Mesozoic stratigraphy of the Hawthorne and Tonopah quadrangles, Nevada. Geological Society of America Bulletin, 50:1573-1624.

PLILER, R., and J. A. S. ADAMS. 1962. The distribution of thorium, uranium, and potassium in the Mancos shale. Geochimica et Cosmochimica Acta, 26:1115-1135.

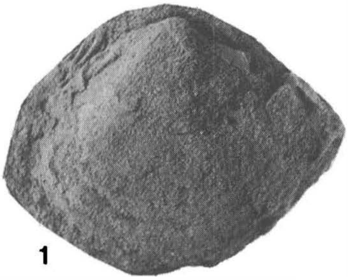
TAYLOR, D. G. 1982. Jurassic shallow marine invertebrate depth zones, with exemplification from the Snowshoe Formation, Oregon. Oregon Geology, 44:51-56.

____. 1998. Nevada field trip guide for the Gabbs and Sunrise Formations in west-central Nevada. 5th International Symposium on the Jurassic System, Vancouver, British Columbia.

____, P. L. SMITH, R. A. LAWS, and J. GUEX. 1983. The stratigraphy and biofacies trends of the Lower Mesozoic Gabbs and Sunrise Formations, west-central Nevada. Canadian Journal of Earth Sciences, 20:1598-1608.

TOZER, E. T. 1979. Latest Triassic (Upper Norian) ammonoid and *Monotis* faunas and correlations. Rivista Italica Paleont., 85 (3-4):843-876.

Plate 1. **1-5**, ? *Mesomiltha* n. sp. A, **1**, (NWMNH 25490), New York Canyon (=NYC), Muller Canyon Member, Section C (Fig. 1), bed 14", right valve; **2-3**, (NWMNH 25491 and 25492), NYC, Mount Hyatt Member, Section C, bed 8", right and left valves; **4-5**, (NWMNH 25493 and 25494), Milton Canyon, Mount Hyatt Member, left and right valves respectively; **6**, *Gervillella* sp. (NWMNH 25495), NYC, Muller Canyon Member, Section E (Fig. 1), top of bed 25; **7**, ? *Mesomiltha* n. sp. B, (NWMNH 25496), NYC, Muller Canyon Member, Section E, bed 28; **8**, *Chlamys* sp. NWMNH 25497, NYC, Muller Canyon Member; **9-11**, *Oxytoma inequivalvis* (J. Sowerby), NYC, Muller Canyon Member, SectionE, bed 28; **9**, (NWMNH x25498), right valve; **10-11**, (NWMNH 25499 and 25500), left valves; **12-13, 16**, *Choristoceras crickmayi* Tozer, NYC, Muller Canyon Member, Section C; **12**, (NWMNH 25374), bed. 14"; **13**, (NWMNH 25377), bed. 14"; **16**, (NWMNH 25373, Muller Canyon Member, section C, bed 14". *Choristoceras marshi* (Hauer), **14-15**, (NWMNH 25379), NYC, about 2.5 m below top of Mount Hyatt Member, Section C (Fig. 1), bed A5; **17**, (NWMNH 25381, NYC, near top of Mount Hyatt Member, Loc. 603. *Choristoceras* sp. A, **18-22**, NYC, Muller Canyon Member, Section C; **18** (NWMNH 25403, bed. 10"; **19** (NWMNH 25503), bed 10"; **20**, (NWMNH 25404), bed 10"; **21**, (NWMNH 25406, bed 10"; **22**, (NWMNH 25407), bed 10"; **23**, ichthyosaur tooth, (NWMNH 25504), NYC, Mount Hyatt Member, float specimen from Section E (Fig.1), bed 9. Bar scale for all figures = 1 cm.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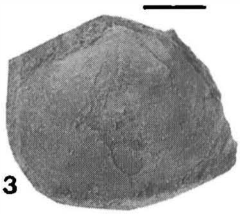
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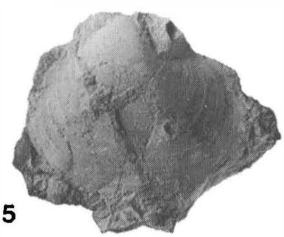
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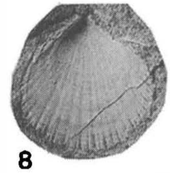
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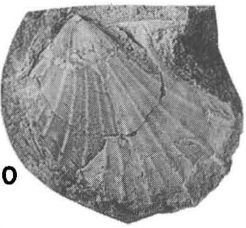
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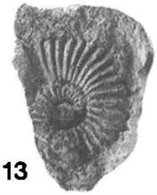
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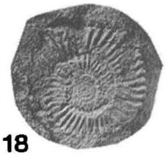
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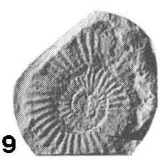
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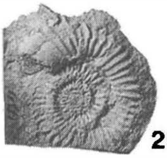
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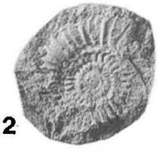
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Recognition of Potential Palynoevents in the Jurassic Sequence of India and their Correlation in Australia

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Keywords: Spore-Pollen Species, Palynoevents, Zonation, Correlation, India, Australia

Abstract: An integrated spore-pollen zonation scheme is presented for Jurassic sequences on the Indian peninsula. This analysis results in the identification of the following major events in the Jurassic palynoflora, i.e., the first occurrences of *Callialasporites turbatus / dampieri, Classopollis* - complex, *Contignisporites cooksoniae, Murospora florida* and *Cicatricosisporites australiensis* in sequential order. Based on FADs (First Appearance Datums) of these species, five palynoevents have been identified in the span of Pliensbachian-Tithonian time, which are correlated with those in Australia of coeval time. These species are cosmopolitan and have synchronous first appearances in other parts of the world, for example *Callialasporites* spp. in the Pliensbachian, *Murospora florida* in the Callovian, and *Cicatricosisporites australiensis* at the Kimmeridgian – Tithonian boundary. The proposed scheme of palynozonation provides a correlation tool in the Jurassic of the Indian peninsula. This review provides evidence of Upper Jurassic deposits in almost all the major basins but with a hiatus from the underlying sediments, which comprise Upper Permian or Triassic sequences.

Introduction

The aim of this contribution is to present an outline of spore-pollen zonation for the Jurassic sequence on the Indian peninsula. The Lower and Middle parts of the Jurassic deposits are not well represented; only the Upper Jurassic has a widespread occurrence recognised in most of the Upper Gondwana basins. From Middle Callovian onwards, the palynozonation is based on stratigraphic resolution as detailed as that for Australian Mesozoic biostratigraphy. The work on terrestrial Jurassic palynoassemblages is mostly from the subsurface, although data from surface material are also known (Fig. 1). An extensive study in the areas of Panagarh and Birbhum, West Bengal, and a critical review of published palynological studies (Singh and Venkatachala, 1988; Ramanujam, 1993; Tiwari and Tripathi, 1992; Vijaya, 1997; 1999a; 1999b) have provided the basis for the present palynozonation of the Jurassic palynoflora. From this analysis, the First Appearances Datums (FADs) of spore-pollen species have been used as biostratigraphic tools for dating and correlation.

Stratigraphy

Lithostratigraphy of the Upper Gondwana sediments on the Indian peninsula has been dealt in broader frame of time. Hence, the lithological units represent relative time span. The formational division of Upper Mesozoic Sequence in each basin has been named differently, as given in Table 1 (Vijaya and Tiwari, 1996).

Potential Events

The basic criterion in palynozonation is the recognition of Palynoevents, and the sequential record of such events provides the frame for palynozones. The first appearance level of a spore-pollen species (FAD) is the key-point in palynoevent stratigraphy.

Event stratigraphy is based on the recognition of short-term events which are synchronous/isochronous over a wider geographic range. The first appearance of a morphological species (Pl. 1, 1-24) of stratigraphic value is a palynoevent, which may or may not be short lived in time. The first step is the identification of such events in continuous sequences, followed by a search for their occurrences in other sections of similar age. To develop palynoevent stratigraphy for the Indian Jurassic sequence, the recognition of such potential events has been attempted here.

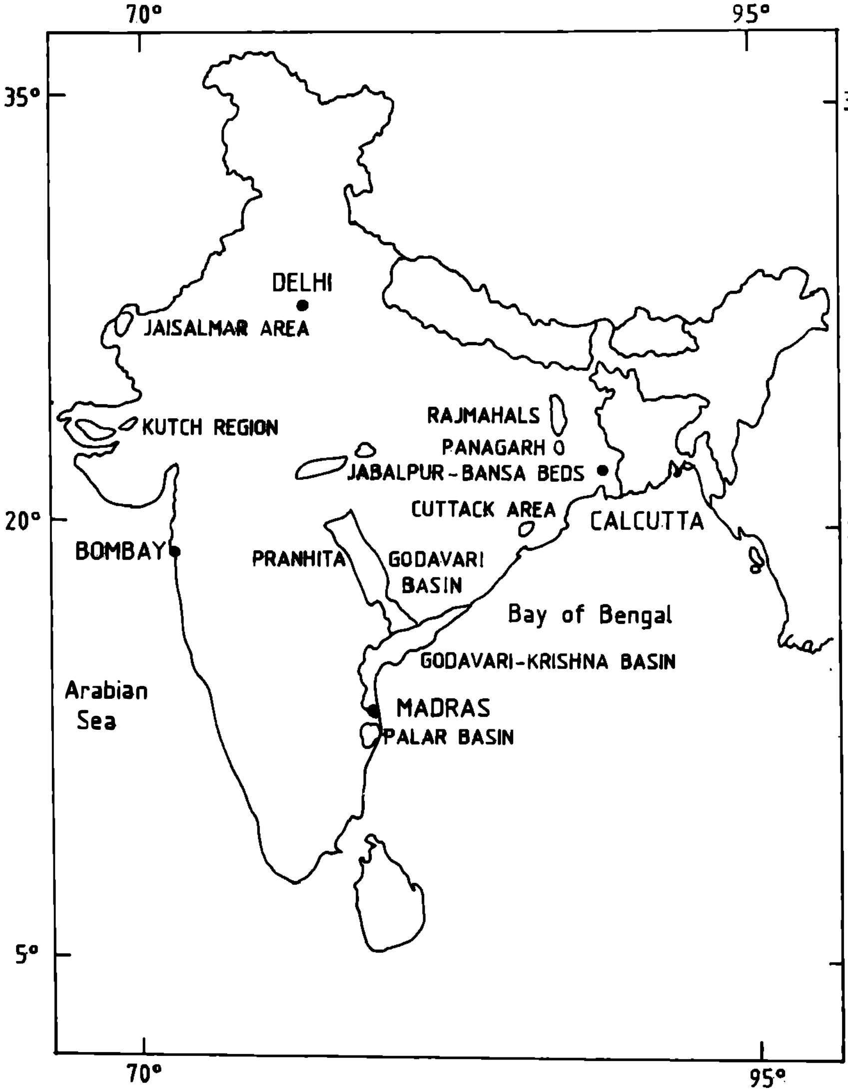


Figure 1. Mesozoic sedimentary basins of the Indian peninsula.

The Jurassic palynoassemblages have not been worked out extensively in the Indian Mesozoic succession. The recently available palynological data from subsurface Mesozoic sediments in the Panagarh area, West Bengal, have provided significant information (Vijaya 1997; 1999a; 1999b). A steady and regular change-over is seen in species appearances (Fig. 2). This fact strongly suggests that further patterns of occurrences may also be traceable, although continuous sequences are not available. Similar sequences of FADs can be traced, which are sufficiently significant to draw the basic palynoevent stratigraphy for this time span.

The first appearance of *Callialasporites turbatus/dampieri* defines the base of Jurassic deposits (Helby et al., 1987). However, the first record of this taxon has been noted in the ? Liassic on the

Indian peninsula (Vijaya and Tiwari, 1992). For the FAD of *Callialasporites* species a Pliensbachian age was assigned by Burger (1990a). The next important level is recognised by the abundance of genus *Callialasporites*. Increased diversity of *Callialasporites* species is attained suddenly in the UP-sections (Vijaya, 1997) and also well represented above the *Classopollis classoides / simplex* zone (Filatoff, 1975).

Period/Areas	Rajmahal	Panagarh	Kachchh	Satpura	S. Rewa	P-G	Mahanadi	K-G	Palar	Cauvery
LOWER CRETACEOUS	Rajamahal FM	Rajmahal FM	Bhuj FM	Jabalpur FM	Bansa FM	Gangapur FM	Athgarh FM	Raghvapuram /Vemavaram	Sriperumbudur FM	Not defined
JURASSIC	Dubrajpur FM	?	Jhuran	?	?	Kota	?	Golapalli	?	?

Table 1. Relative positioning of lithological units in Mesozoic Basins on Indian peninsula.

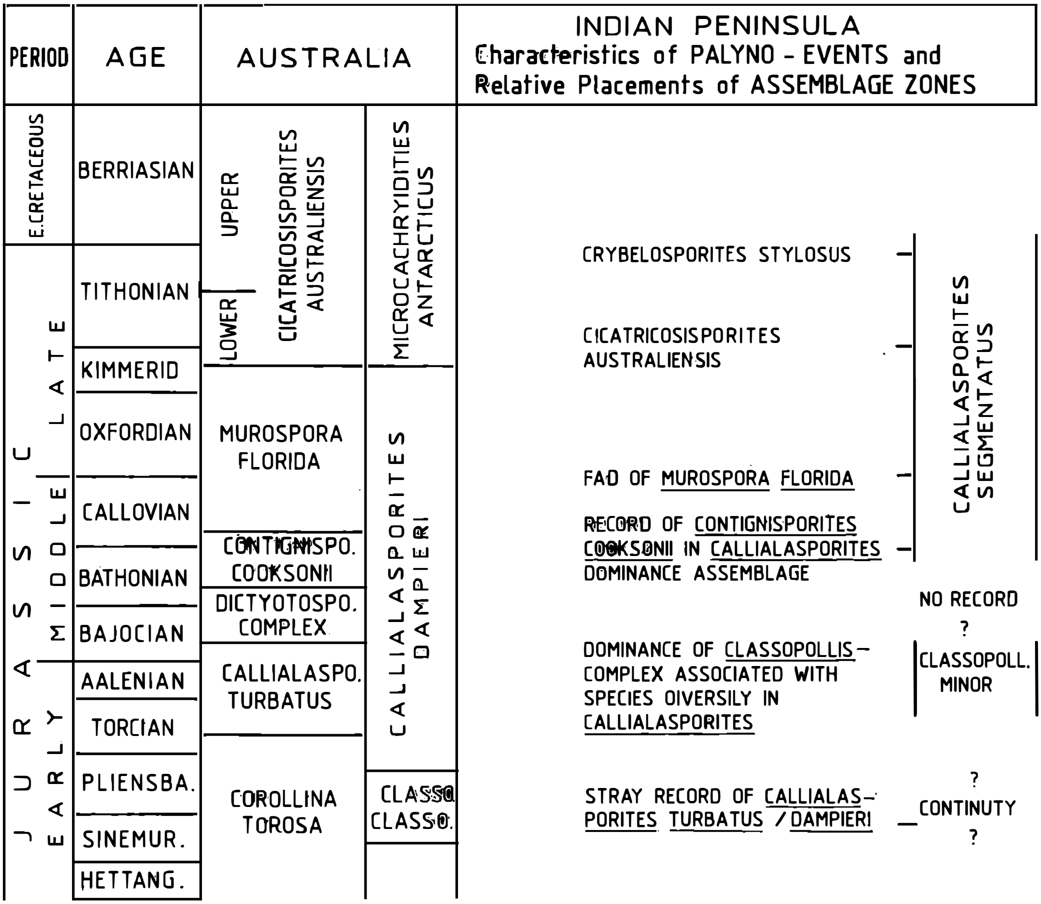


Figure 2. Pattern of occurrences of Jurassic spore-pollen species in the Infratrappean Panchet sediments of Mesozoic succession in Panagarh area, West Bengal. Modes of FADs in sequential order display basic palynoevent stratigraphy for Upper Jurassic deposits.

The Lower Triassic Panchet Formation, where species of *Callialasporites* occur, is represented by about 20-175m of stratigraphic thickness in Panagarh area. The sudden appearance of this taxon in Lower Triassic strata can be regarded as potential event suggesting the initiation of Jurassic sedimentation. This level can further be traced in the areas of -- Panagarh and Birbhum, West Bengal (Vijaya 1997; 1999a; 1999b) but is not physically located in the Mesozic lithosequence (Fig. 2).

Furthermore, in the Upper Jurassic sequential FADs of *Contignisporites cooksoniae, Murospora florida* and *Cicatricosisporites australiensis* in a *Callialasporites - Araucariacites* dominated palynoassemblage mark episodic events in the palynozonation. These palynoevents range from the Bathonian to Tithonian. Although these species have long vertical ranges, they still have their usefulness as event-markers (Burger, 1990a; 1990b). The possible age ranges for each such palynozone containing the marker species has been tied to those in Australia (Fig.3).

Palynozonation

In re-assessing palynological data from Jurassic deposits, it is concluded that the precise ages of various palynoassemblages has not been dealt with precision for their age correlation, except for the Lower and Middle-Upper Jurassic. Presently, two main assemblages in the Jurassic sequence are identified, based on generic dominance: at the lowest level, the *Classopollis-* complex attained prominence, while the upper major part is represented by an abundance of *Callialasporites*. These two characteristic segments of the Jurassic palynoflora were recently defined as the *Classopollis minor* and *Callialasporites segmentatus* Assemblage zones by Tiwari and Tripathi (1992). Here, these assemblages are assessed for palynoevent stratigraphy, on the basis of FADs of those spore species which are important for palynozonation (Helby et al; 1987; Burger 1990a; 1990b).

Classopollis minor Assemblage Zone

***Classopollis minor* palynozone:** The palynoassemblge is characterized by the prominence of the *Classopollis*-complex and rarity of *Callialasporites* species (*C. dampieri, C. triletes, C. segmentatus*). Other associated species are *Gleicheniidites* sp., *Staplinisporites, Verrucosisporites* and microphytoplankton. This zone is recorded only in the western part of India (Table 2), and its placement in the Jurassic palynozonation of Australia (Burger, 1990a) is shown in Figure 3. Stratigraphic information indicates that the abundance of *Classopollis* seems to have been environmentally controlled (Koshal, 1975). Perhaps depositional environment restricted the diversity of the spore-pollen composition of this palynozone. The age assignment of Pliensbachian to Toarcian is probably due to an insufficient record. This is also the case in certain areas of Kachchh and Jaisalmer basins. In India, the record of the *Classopollis*-complex known so far is a short-lived event (Ramanujam, 1993).

Callialasporites segmentatus Assemblage Zone

Contignisporites cooksoniae palynozone: The first record of Contignisporites cooksoniae in the spore-pollen population containing a high abundance of Callialasporites spp. (C. dampieri, C. turbatus, C. segmentatus) defines this particular zone. Recognition of this palynozone has been made in varied facies, but it lacks continuity with the preceding palynozone. Hence, the first unambiguous occurrence (FAD) of this species could not be established here (Table 2, Fig. 3). The appearance of Contignisporites cooksoniae does not seem to be abrupt. The lowest level for this event has been suggested as mid-Bathonian (Burger, 1990a). This species has been observed in many sections of different areas of India (see Table 2).

***Murospora florida* palynozone:** The next important event in the *Callialasporites segmentatus* Assemblage Zone is the first appearance of *Murospora florida*, which has been noted to be independent of facies. The FAD of this particular species defines the base of the *Murospora florida* Zone. Other significant species which make up this assemblage are *Callispora potoniei, Lametatriletes indicus, Boseisporites indicus, Contignisporites* spp., *Crassimonoletes surangei*, and *Podosporites* sp. The highest levels of this zone are characterised by the FAD of *Cicatricosisporites australiensis* (Burger, 1990b). Dominance of *Callialasporites-Araucariacites* is simultaneous worldwide in Upper Jurassic-Lower Cretaceous palynoassemblages. A notable feature at this transition is the presence of *Microcachryidites antarcticus* and many species of *Contignisporites, Lycopodiacidites,* and *Osmundacidites* (Table 2).

Table 2. Age correlation of palynoassemblages on record from India, in the Callialasporites dampieri Superzone of Australia (after Burger, 1990a).

PERIOD	FORMATION	BASIN / AREA OF STUDY	CHARACTERISTICS OF ASSEMPLAGE	PLACEMENT IN JURASSIC SEQUENCE (Burger, 1990a)
Lower Jurassic	Lathi Fm, Subsurface samples	JAISALMER Jaisalmer; Srivastava, 1966	I	I
	Lithounit not defined, Subsurface samples	KACHCHH Between Patcham island & Kachchh Mainland; Koshal, 1975	Prominence of *Classopollis* - complex with *Callialasporites* spp evidence Jurassic., absence of *Murospora, Contignisporites* indicate an older aspect in Jurassic sequence	Equated with *Classopollis classoides* Palynozone, Torcian-Aalenian
	Kota Fm, Surface samples	PRANHITA - GODAVARI Kota Village; Prabhakar, 1989	I	I
Upper Jurassic	Middle Member, Jhuran Formation, Surface samples	KACHCHH Kachchh Mainland Jhuran River Section; Maheshwari & Jana, 1988	*Bhujiasporites* sp., *Impardecispora*, *Lametatriletes, Contignsporites,* absence of *Murospora florida*	I I I I
	Bansa Formation; Surface samples	SATPURA Machrar Nala,; Bharadwaj & Kumar, 1974a	Presence of *Callispora, Boseisporites* with fair occurrence of *Podosporites*	Species composition suggest younger part of *Murospora florida* palynozone, Kimmeridgian - Tithonian in age
	Lithounit not defined, Surface samples	Parsapani coals; Bharadwaj & Kumar, 1974b	*Callispora potoniei, Boseisporites sehoraensis, Lametatriletes indicus, Contignisporites cooksoniae*	I I

Upper Jurassic	Bansa Formation, Surface samples	SOUTH REWA Bansa, Maheshwari, 1975	*Callispora foveolata, Lametatriletes indicus, Boseisporites* spp., *Contignisporites cooksoniae, Crassimonoletes surangei* & *Podosporites sp*	Much species diversity in spores place in *Murospora florida* palynozone; Kimmeridgian-Tithonian in age
	Athgarh sandstone, Surface sample	MAHANADI Sidheswar Hill; Maheshwari, 1975, Jana & Tiwari, 1986	*Contignisporites* spp, *Impardecispora apiverrucata, Foveotriletes, Klukisporites* spp. *Lametatriletes indicus* and *Boseisporites* sp.	
	Lithounit not defined, Subsurface samples	PALAR Kandigai village, Tripathi & Vijaya, 1997	*Contignisporites, Neoraistrickia, Undulatisporites, Murospora florida*	
	Lithounit not defined Subsurface samples	CAUVERY Karaikal well (3055-3060 m); Venkatachala, Sharma & Jain, 1972	*Contignisporites glebulentus, Microcachryidites antarcticus* along with other fern spores	
Upper Jurassic / Lower Cretaceous	Lithounit not defined, Subsurface samples	KACHCHH Between Patcham Island and Kachchh Mainland; Koshal, 1975	*Callialasporites* dominance, *Contignisporites, Klukisporites, Lycopodiacidites, australiensis*	
	Upper Member, Jhuran Formation, Surface samples	Maheshwari & Jana, 1988	Presence of *Cicatricosisporites australiensis, Aequitriradites* sp.	

Upper Jurassic/ Lower Cretaceous		SATPURA		I
	Lithounit not defined, Surface samples	Kotri; Kumar & Kulshreshtha, 1979	*Boseisporites* spp., *Lametatriletes* spp., *Contiginisporites cooksoniae*, *Murospora* sp., *Cicatricosisporites* sp.	I
	Jabalpur Group, Surface samples	Morghat; Maheshwari & Kumar ,1979	*Alsophilidites psilatus, Contigni-sporites glebulentus, Callispora potoniei, Murospora florida & Cicatricosisporites australiensis*	I I
		MAHANADI		
	Athgarh Sandstone, Surface samples	Orissa, Talbasta; Jana, 1990	*Impardecispora spp. Klukisporites spp, Lycopodiacidites* in *Murospora* prominence, & *Acquitrithadites*	Much species diversity suggest lower part of *C. australiennsis* palynozone, Tithonian in age
		PALAR		I
	Equated with Sriperumbudu r Formation	Canjeevaram; Ramanujam & Srisailam, 1974	*Impardecispora apiverucatus, Echinatisporis vembanii, Contignisporites* spp. & *Cicatricosisporites* sp.	I
		RAJMAHAL		I
	Lithounit not defined, Surface samples	Sah & Jain, 1965	*Alsophilidites* spp.,*Callispora-baculoexinus, Neoraistrickia truncatus, Trilobosporites purverulentus & Cicatricosisporites australiensis*	I
Jurassic		SATPURA		
	Chaugaon Beds, Surface samples	Khatama caves; Kumar, 1992	Prominence of *Podosporites*	I I
	Jabalpur Formation, Surface samples	Patbaba Ridge; Kumar, 1986	Recovery and presentation of age marker species not satisfactory	Precise placement not possible
	Nimar Formation, Surface samples	Jhabua-Umali village; Kumar, 1994	*Cingutriletes clavus* present	I

This sudden appearance of *M. florida* in the *Callialasporites-Araucariacites* dominated population, followed by increasing species diversity, is interpreted as a potential palynoevent. The wide geographical range of this species allows recognition of the uppermost part in the Jurassic sequence, as seen in the Australian palynozonation scheme; a Callovian age for this species was originally proposed by Burger (1990a) (Fig. 3).

Discussion

Present knowledge of the first appearances of spore and pollen species in the Mesozoic basins of peninsular India has been reviewed in order to develop palynoevent stratigraphy. Due to the limited palynological data available for the Jurassic sequence, only the upper part of Jurassic is discussed here. The analyses include sections (Table 2) that have been sufficiently well sampled that changes in palynoevents can easily be recognized. These analyses are further supplemented by a palynozonation scheme for Australia covering coeval time (Burger, 1990a; 1990b; Dettmann et al., 1992).

The basic criteria for palynozonation include substantial decrease in the *Classopollis-* complex and corresponding relative increase in *Callialasporites* species diversity. More refined zonation requires the use of first appearances (FADs) of particular marker species.

In Panagarh area, West Bengal, the sediments corresponding to the *Murospora florida* zone are fluvio-lacustrine and the Panchet Formation (Lower Triassic) is capped by lava flows. Concerning the exact age of the Panchet Formtion and oldest Intertrappean of the Rajmahal Formation in the Panagarh sub-basin, the FADs of two spore species (*Murospora florida* and *Cicatricosisporites australiensis*) provide evidence for Upper Jurassic deposits continuing into the Lower Cretaceous (Fig. 2; Vijaya 1997; 1999a; 1999b). Correlation of the.palynozones identified here with those of the Jurassic palynosequence of Australia (Fig. 3) indicates certain compositional differences probably due to climatic differences. This kind of disparity probably reflects a group of spore-pollen species having endemic distribution (Helby et al., 1987).

The Upper Jurassic palynoassemblages in some parts of India exhibit a somewhat complex association of many endemic taxa, e.g., *Callispora, Alsophilidites, Boseisporites,* and *Lametatriletes.* Such varied composition of spore-pollen assemblages from east-coast and central India demonstrates the existence of special palaeoenvironmental adaptations for the survival of vegetation.

The apparent similarity with the Jurassic palynozones on eastern Gondwana suggests that the vegetation comprising the Jurassic floras of Australia and India were of similar composition. Such close relations between the two areas is indicated by the presence of selected spore-pollen species of biostratigraphical significance (Fig. 3).

As indicated above, it does not seem possible to identify the exact first appearance level of *Callialasporites* species, although they are the only criterion providing evidence for Jurassic deposits.

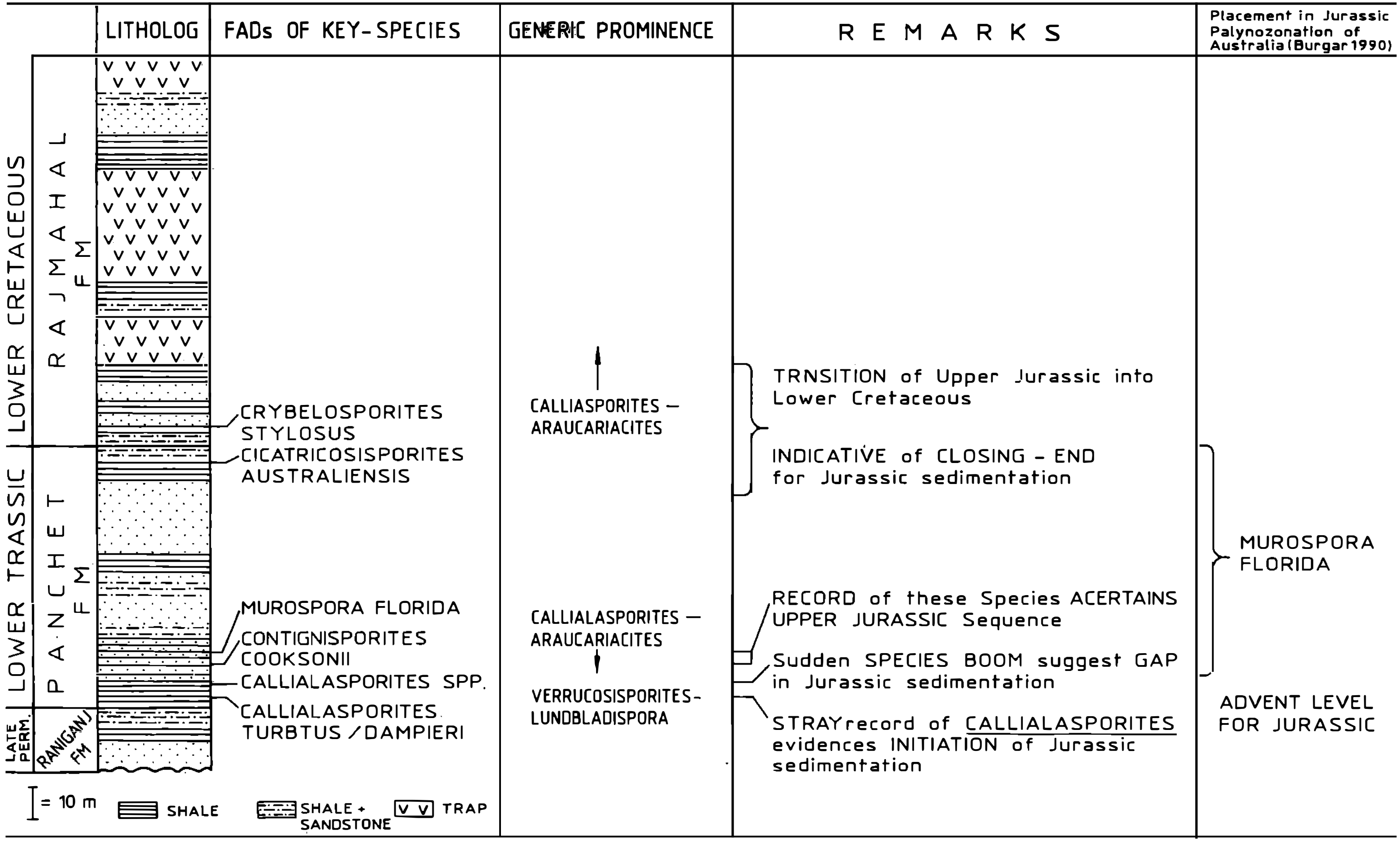
Knowledge of Lower and Upper Jurassic palynoassemblages is still fragmentary, and no definite palynoevents can yet be identified. More research is needed to test whether these occurrences represent regional palynoevents. The first observation of a few specimens of *Callialasporites,* and the subsequently sudden burst in population of the same taxon must be defined with reference to the gap in deposition because, the two happenings cannot be referred to the same time plane, i.e., the FAD and dominance of the same taxon requires a time lapse.

Conclusions

The Jurassic palynoassemblages in the Mesozoic sequence of India are not defined as precisely as those recognised in the Australian Mesozoic (Burger, 1990a; Dettmann et al., 1992). Based on stratigraphical occurrences of key-species in the *Callialasporites segementatus* Assemblage zone, it is proposed that the periodic occurrences of these species may allow a better Palynoevent stratigraphical analysis, as follows:

- The sudden first appearance (FAD) of *Callialasporites turbatus/dampieri* in otherwise Triassic palynoassemblages is interpreted as a sudden evolutionary event corresponding to the advent of Jurassic sedimentation;

Figure 3. Proposed palynozonation scheme for the Jurassic sequence on Indian peninsula, based on relative FADs of spore-pollen species of stratigraphic value.



TRNSITION of Upper Jurassic into Lower Cretaceous

INDICATIVE of CLOSING - END for Jurassic sedimentation

RECORD of these Species ACERTAINS UPPER JURASSIC Sequence

Sudden SPECIES BOOM suggest GAP in Jurassic sedimentation

STRAY record of CALLIALASPORITES evidences INITIATION of Jurassic sedimentation

- Occurrence of the *Classopollis*-complex has the aspect of a palynoevent horizon, because it lacks continuity with older strata, i.e. Triassic;
- Subsequent sudden prominence attained by *Callialasporites* spp. in the *Araucariacites* assemblage indicates a later stage in the sequence;
- Sequential FADs of *Contignisporites cooksoniae* and *Murospora florida* give evidence of Upper Jurassic deposits; and,
- Appearance of *Cicatricosisporites australiensis* is indicative of a transitional phase from the Jurassic into the basal Cretaceous.

References

BHARADWAJ, D. C., AND P. KUMAR. 1974a. Palynostratigraphy of Mesozoic sediments from Machrar Nala, Basin, M.P. India. Geophytology, 4:147-152.

____. 1974b. Palynostratigraphy of Parsapani coals, M.P., India. Geophytology, 4:184-188.

BURGER, D. 1990a. Australian Phanerozoic Timescales, 8. Jurassic biostratigraphic charts and explanatory notes. Bureau of Mineral Resources of Australia Record, 1989/38. Australia. 38 p.

____. 1990b. Australian Phanerozoic Timescales , 9. Cretaceous biostratigraphic charts and explanatory notes. Bureau of Mineral Resources of Australia Record, 1989/38. Australia. 36 p.

DETTMANN, M. E., R. E. MOLNAR, J. S. DOUGLAS, D. BURGER. C. FIELDING, H. T. CLIFFORD, J. FRANCIS, P. JELL, T. RICH, M. WADE, V. RICH, N. PLEDGE, A. KEMP, AND A. ROZEFELDS. 1992. Australian Cretaceous terrestrial faunas and floras: biostratigraphic and biogeographic implications. Cretaceous Research, 3:207-262.

FILATOFF, J. 1975. Jurassic palynology of the Perth Basin, Western Australia. Palaeontographica, 154:1-130.

HELBY, R. J., R. MORGAN, AND A. D. PARTRIDGE. 1987. A palynological zonation of the Australian Mesozoic. Memoirs of the Association of Australian Palaeontologists, 4:1-94, Australia.

JANA, B. N. 1990. Palynology of Mesozoic outcrops of Athgarh Formation exposed near Talbasl, Orissa. Palaeobotanist, 38:155-162.

____, AND R. S. TIWARI. 1986. Further observation on the palynological assemblage from the Athgarh Formation, Sidhes-war Hills, Orissa. Quarterly Journal geol. Min. metall. Soc. India, 58:201-209.

KOSHAL, V. N. 1975. Palynozonation of Mesozoic subsurface sediments of Banni, Kutch, Gujarat. Quarterly Journal of Geology. Min. metall. Soc. India, 47:79-82.

KUMAR, P. 1986. Palynology of Jabalpur Formation at Patbaba Ridge, Madhya Pradesh, India. Indian Society of Geologists Bulletin, 2:37-39.

____. 1992. Palynodating of the Carbonaceous sediments exposed below Khatma caves in Hoshangabad district, Madhya Pradesh. Biological Memoirs, 18:61-73.

____. 1994. Palynology of Carbonaceous clays of Nimar Formation, Jhabua District, Madhya Pradesh, India. Journal Geological Society of India, 44:671-674.

____, AND KULSHRESHTHA. 1979. Palynostratigraphical studies of Carbonaceous shales from Kotri, Narsinghpur District, M.P. India. Geophytology, 9:52-61.

MAHESHWARI, H. K. 1974. Lower Cretaceous palynomorphs from the Bansa Formation, South Rewa Basin, India. Palaeontographica, 146:21-55.

____. 1975. Palynology of the Athgarh Formation near Cuttack, Orissa. Palaeobotanist, 22:23-28.

____, AND B. N. JANA. 1988. Palynozonation of Jhuran and Bhuj formations in Kutch Basin. Palaeobotanist, 36:177-182.

____, AND P. KUMAR. 1979. A Jurassic mioflora from the Jabalpur Groups exposed in Morand River near Morghat, Hoshangabad District, Madhya Pradesh. Geophytology, 9:47- 51.

PRABHAKAR, M. 1989. Palynological evidence and its significance for the Kota Formation in the Pranchita-Godavari Basin. Proceedings 12th Indian Colloquium on Micropalaeontology & Stratigraphy. p. 59-65.

RAMANUJAM, C. G. K. 1993. Spore-pollen suits of some Upper Gondwana deposits of South India - a critical appraisal. Gond. geol. Mag. 1993; Spl. Vol. Birbal Sahni Cente. Nat. Symp. Gondwana India. p. 462-478.

____, AND K. SRISAILAM. 1974. Palynology of carbonaceous shales from a borehole at Kattavakkam near Conjeevaram, Tamil Nadu, India. Pollen spores, 16:67-102.

SAH, S. C. D, AND K. P. JAIN. 1965. Jurassic spores and pollen grains from the rajmahal Hills, Bihar, India: with a discussion on the age of the Rajmahal Intertrappean beds. Palaeobotanist, 13:264-290.

SINGH, H. P., AND B. S.VENKATACHALA. 1988. Upper Jurassic-Lower Cretaceous spore-pollen assemblages in the Peninsular India. Palaeobotanist, 36:168-176.

SRIVASTAVA, S. K. 1966. Jurassic microflora from Rajasthan, India. Micropalaeontology, 12: 87-103.

TIWARI, R. S., AND A. TRIPATHI. 1992. Marker Assembalge-Zones of spores and pollen species through Gondwana Palaeozoic and Mesozoic sequence in India. Palaeobotanist, 40:194-236.

TRIPATHI, A. AND VIJAYA. 1997. Biostratigraphic age determination of Sriperumbiudur Formation in bore hole PBSD-1, Palar Basin, Tamil Nadu: a re-assessment. Journal Palaeontological Society of India, 42:101-116.

VENKATACHALA, B. S., K. D. SHARMA, AND A. K. JAIN. 1972. Palynological zonation of Jurassic-Lower Cretaceous sediments in the sub-surface of Cauvery Basin, p. 172-187. *In* A. K. Ghosh, et al. (eds.)- Proc. Sem. Paleopalynol. Indian Stratigr., Calcutta, 1971, Botany Department, Calcutta University.

VIJAYA. 1997. Palynological dating of infra- and intertrappean sediments in the Mesozoic succession of Borehole PGD-6, Domra sub-basin, West bengal, India. Cretaceous research, 18:833-848.

____. 1999a. Palynological evidence for Jurassic-Cretaceous transition in Rajmahal Formation, Panagarh sub-basin, West Bengal India. Alcheringa.

____. 1999b. Search of Jurassic in subsurface Mesozoic sediments, Birbhum District, West Bengal, India. Proc. XVI Indian Colloquium, Micropalaeontology and Stratigraphy, Goa.

____, AND R. S. TIWARI. 1992. Morpho-evolutionary biohorizon stratigraphy and cladistics in saccate pollen through Gondwana sequences of India. Palaeobotanist, 40:157-193.

____, AND R. S. TIWARI. 1996. Status of continental Jurassic on the Indian peninsula, p. 379-393. *In* M. Morales (ed.). The continental Jurassic Symposium Museum of Northern Arizona Bulletin 60.

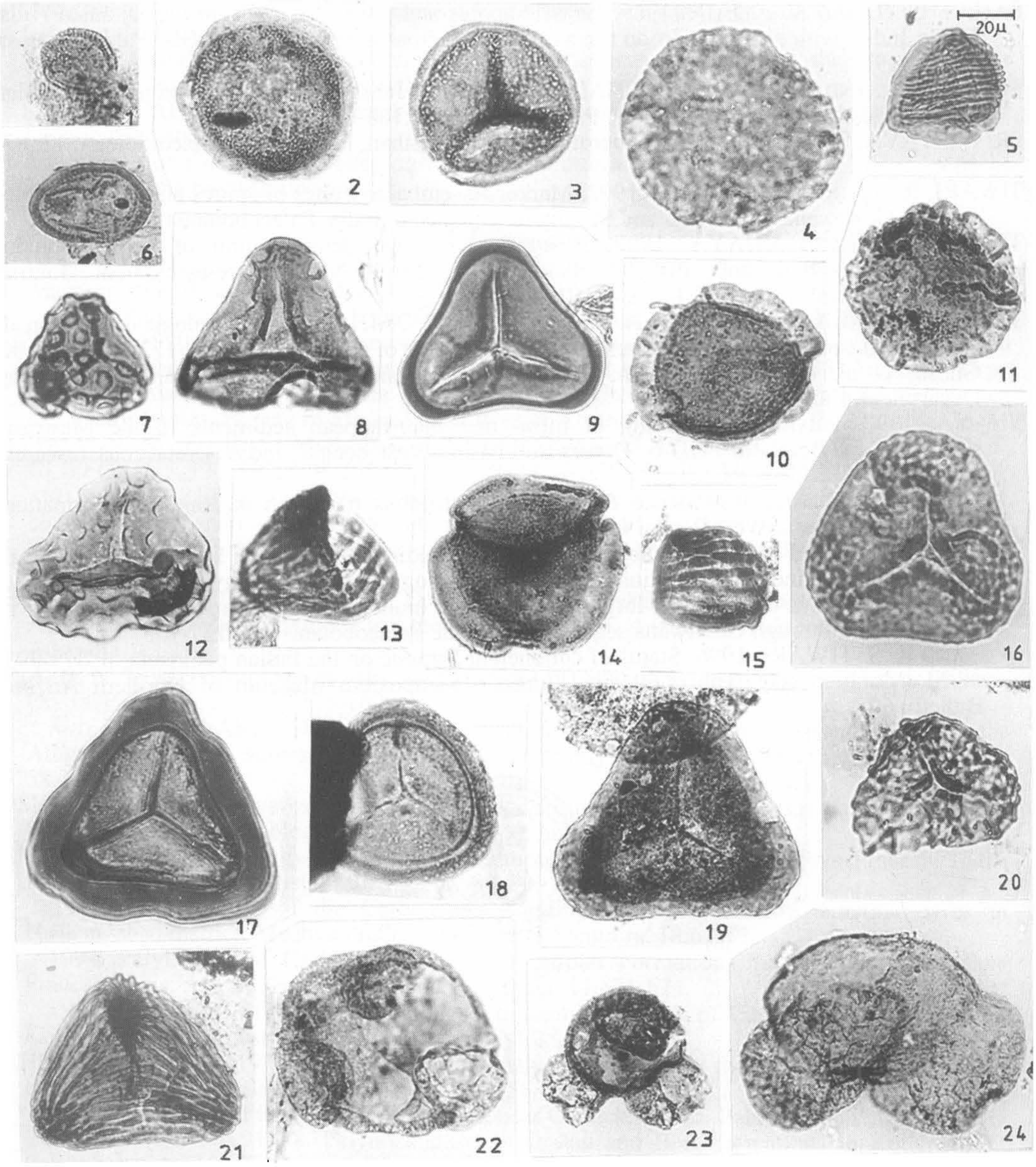


Plate 1. (Bar =µm) Characteristic spore-pollen species in the Jurassic palynoflora. **1.** *Classopollis classoides*. **2.** *Callialasporites velatus*. **3.** *Callialasporites triletus*. **4.** *Callialasporites turbatus*. **5.** *Contignisporites multimuratus*. **6.** *Classopollis chateaunovi*. **7.** *Klukisporites scaberis*. **8.** *Dictyophyllidites venkatachaii*. **9.** *Matonisporites cooksonae*. **10.** *Callialasporites dampieri*. **11.** *Callialasporites segmentatus*. **12.** *Ischyosporites punctatus*. **13.** *Contignisporites cooksonae*. **14.** *Callialasporites trilobatus*. **15.** *Duplexisporites problematicus*. **16.** *Impardecispora purverulentus*. **17.** *Murospora florida*. **18.** *Densoisporites velatus*. **19.** *Boseisporites insignitus* **20.** *Triletes tuberculiformis*. **21.** *Cicatricosisporites australiensis*. **22.** *Podosporites tripakshi*. **23, 24.** *Microcachyidites antarcticus*.

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Recent Advances in Upper Jurassic (Kimmeridgian - Tithonian) Ammonite Biostratigraphy of North-Central Mexico Based on Recently Collected Ammonite Assemblages

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Abstract: The analysis of recently collected ammonite assemblages from type and new sections in central and eastern Mexico allows improvement of Upper Jurassic (Kimmeridgian-Tithonian) biostratigraphy and biochronostratigraphic correlations. The *Schneidia*, lower *Idoceras* without *Sutneria*, upper *Idoceras* with *Sutneria*, *Taramelliceras-Aulacostephanus*, *Coryceras-Procraspedites* assemblages, and the interval zone with *Lingulaticeras*, are all useful in subdividing Kimmeridgian deposits. The precise identification of the Kimmeridgian-Tithonian boundary is difficult, but the upper Beckeri Zone and part of the Hybonotum Zone can be recognized on the basis of *Hybonoticeras*. The *Usseliceras-Franconites*, *Parastreblites-Torquatisphinctes*, and *Andiceras-Kossmatia* assemblages characterize stratigraphic intervals in overlying Lower Tithonian deposits. More research is necessary to determine precisely Upper Tithonian ammonite assemblages in central and eastern Mexico. A revision of traditional and recent biostratigraphic interpretations is provided on the basis of the new material collected in situ bed-by-bed.

Introduction

Biostratigraphic data and biochronostratigraphic interpretations of Kimmeridgian-Tithonian sections studied bed-by-bed in Mexico are rare. Published information showing the influence of valuable papers by pioneers such as Burckhardt (1906 to 1930) and Imlay (1939 to 1943) overwhelmingly dominates, and later contributions usually lack precise ammonite biostratigraphy (Cantú-Chapa, 1967 to 1989). However, recent bio- and biochronostratigraphic reinterpretations of Kimmeridgian-Tithonian ammonites in Mexico have become available both by revision of the literature (Callomon, 1992) and paleontological collections (Linares et al., 1997; Olóriz, Villaseñor and González-Arreola, 1998), as well as by field work on type and new sections providing ammonites collected bed-by-bed (Olóriz et al., 1988; 1990a; 1992; 1993; Olóriz, Villaseñor, González-Arreola, and Westermann, 1998; Villaseñor, 1991; Olóriz and Villaseñor, in press).

The present paper presents recent advances in Kimmeridgian-Tithonian ammonite biostratigraphy, and biochronostratigraphic correlation, obtained by the authors in 14 sections investigated in the Mexican Altiplano, northern-central Mexican States of San Luis Potosí, Zacatecas, and Durango (Fig. 1), improving our previous biochronostratigraphic interpretations (Villaseñor, 1991; Olóriz, 1992; Villaseñor et al. 1993, 1994).

Kimmeridgian

No ammonites older than Kimmeridgian have been found in the La Casita Formation and lateral equivalents (Olóriz et al., 1990a; 1990b; 1992; Villaseñor, 1991; Olóriz, 1992), and ammonites from lateral equivalents of the top of the Zuloaga Limestone have always been interpreted as Late Oxfordian in age (Olóriz et al., 1990b; Myczynski et al., 1998).

Schneidia Assemblage

This assemblage is characterized by the identified range of endemic ataxioceratinae which clearly dominate over secondary haploceratids. No raseniids were found. Microconchiate ataxioceratids are morphologically close to *Schneidia* belonging to the groups of *Schneidia* n. sp. A, *S. collignoni* and *S. lussasense* (Atrops). At present, macroconchs seem to be less frequent and only known from incomplete specimens of *Ataxioceras* and probable *Lithacosphinctes*. Among haploceratids, *Glochiceras* (*Lingulaticeras*) *lingulatum* (Quenstedt), *G.* (*L.*) *nudatum* (Oppel), *Taramelliceras* (*Metahaploceras*) sp. aff. *subnereus* (Wegele), and *T.* (*M.*) *rigidum* (Wegele), as well as specimens belonging to the *kobyi* (Choffat) - *rigidum* (Wegele) groups, were identified in the upper part of this stratigraphic interval.

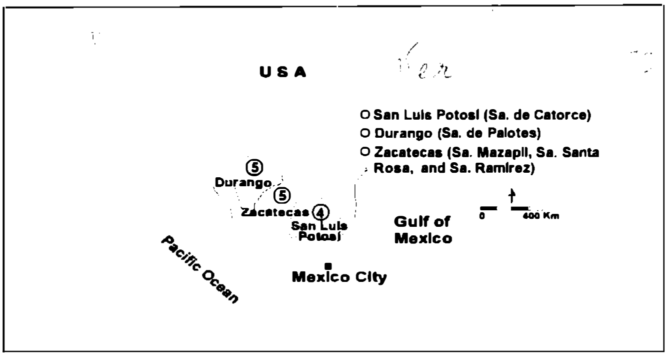
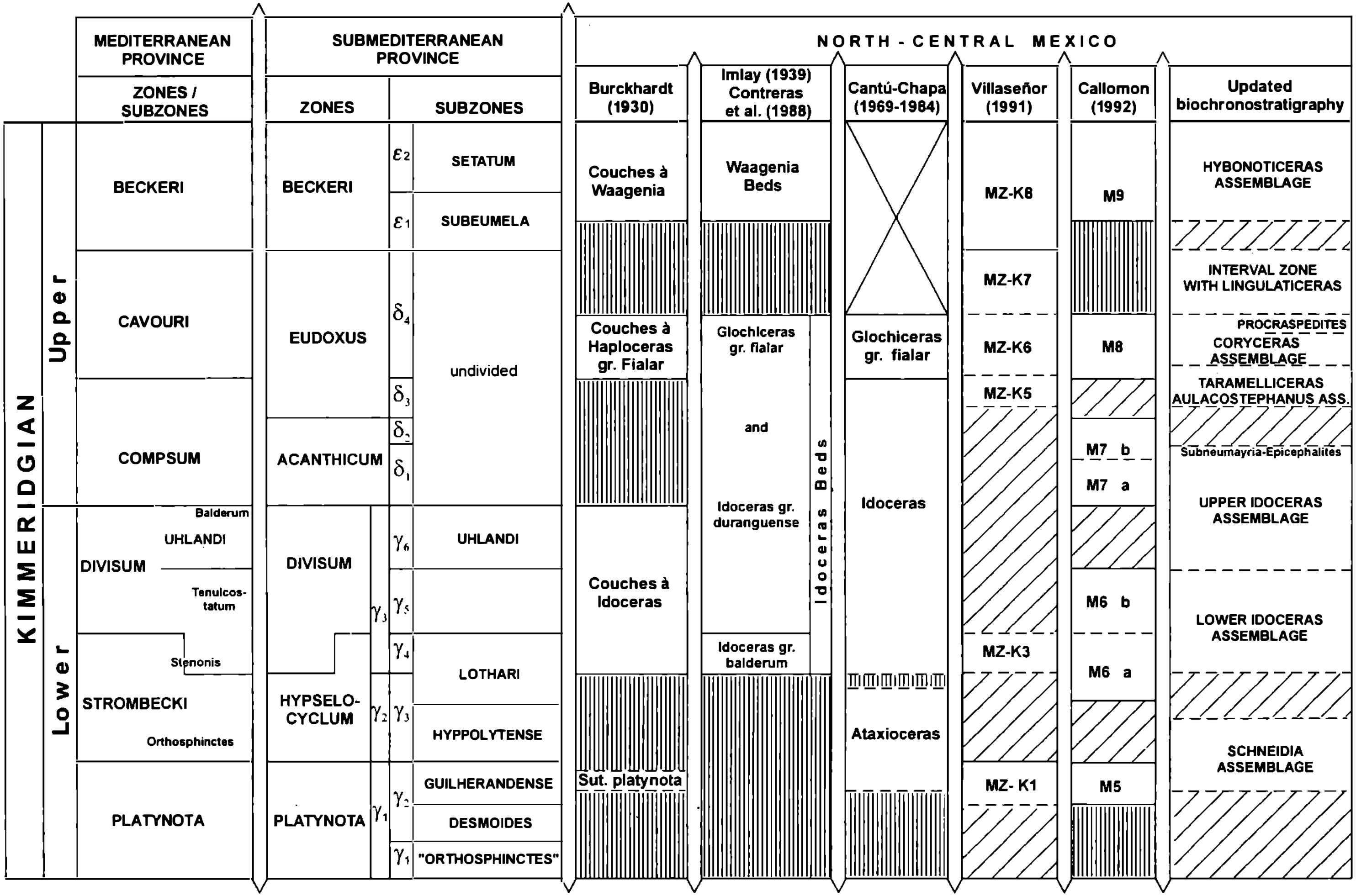


Figure 1. Localization of the studied areas. Numbers relate to the number of sections measured in each area

The *Schneidia* assemblage overlies a ca. 2m thick section rich in bivalves, which is identified directly over the Zuloaga Limestone. This assemblage was gathered from ca. 5m thick section of pinkish siltstones showing the stratigraphic interval which includes the first appearance datum (FAD) and the last appearance datum (LAD) of *Schneidia* in the sections analyzed. We interpret this assemblage to characterize the uppermost Platynota - lowermost Hypselocyclum Zones (Fig. 2). Since we found outcrop limitations (soil cover) persistently overlying the younger records of ataxioceratins, we cannot be conclusive about certain identification of the total range of this assemblage. This interpretation modifies those made by Villaseñor (1991; MZ-K1 stratigraphic interval), Olóriz et al. (1992), and Callomon (1992; M5 faunal horizon), extending the youngest records to the early Hypselocyclum Chron.

The oldest ammonite assemblage identified in La Casita deposits (the *Schneidia* assemblage) is roughly comparable to the so-called "Unidad con *Ataxioceras*" first reported by Cantú-Chapa (1969) from eastern Mexico, but less diversified, since no raseniids were found among the ammonites collected by us. The precise interpretation of differences in the composition of these ammonite assemblages is difficult due to the imprecise biostratigraphic data provided by Cantú-Chapa (1969). According to paleontologic interpretations provided by that author, his assemblage with *Ataxioceras* could expand throughout a slightly wider stratigraphic interval than that interpreted here for the *Schneidia* assemblage, including higher horizons within the Hypselocyclum Zone. Unfortunately, comments by Cantú-Chapa (1969) concerning raseniids, and their assumed presence in ammonite assemblages with *Sutneria platynota* (Reinecke), are too vague to be evaluated. In addition, the record of *Sutneria platynota* (Reinecke) mentioned by Burckhardt (1930) from eastern Mexico was not recognized in the Mexican Altiplano.

Figure 2. Correlation chart of recently collected ammonite assemblages in north-central Mexico. Proposals by Cantú-Chapa, Imlay, Contreras et al., and Burckhardt are reinterpreted. Those by Villaseñor and Callomon are as originally proposed, except of minor adaptations in the latter. Discontinuity in ammonite assemblages: reinterpreted (vertical ruling), originals (oblique ruling). European standards slightly adapted from Hantzpergue et al. (1991).


KIMMERIDGIAN
Upper
Lower
MEDITERRANEAN PROVINCE
ZONES / SUBZONES
BECKERI
CAVOURI
COMPSUM
Balderum
UHLANDI
DIVISUM
Tenuicos-tatum
Stenonis
STROMBECKI
Orthosphinctes
PLATYNOTA
SUBMEDITERRANEAN PROVINCE
ZONES
SUBZONES
BECKERI
ε2
SETATUM
ε1
SUBEUMELA
EUDOXUS
δ4
undivided
δ3
ACANTHICUM
δ2
δ1
DIVISUM
γ6
UHLANDI
γ3
γ5
γ4
LOTHARI
HYPSELO-CYCLUM
γ2
γ3
HYPPOLYTENSE
PLATYNOTA
γ1
γ2
GUILHERANDENSE
DESMOIDES
γ1
"ORTHOSPHINCTES"
NORTH - CENTRAL MEXICO
Burckhardt (1930)
Couches à Waagenia
Couches à Haploceras gr. Fialar
Couches à Idoceras
Sut. platynota
Imlay (1939) Contreras et al. (1988)
Waagenia Beds
Glochiceras gr. fiafar
and
Idoceras gr. duranguense
Idoceras gr. balderum
Idoceras Beds
Cantú-Chapa (1969-1984)
Glochiceras gr. fialar
Idoceras
Ataxioceras
Villaseñor (1991)
MZ-K8
MZ-K7
MZ-K6
MZ-K5
MZ-K3
MZ- K1
Callomon (1992)
M9
M8
M7 b
M7 a
M6 b
M6 a
M5
Updated biochronostratigraphy
HYBONOTICERAS ASSEMBLAGE
INTERVAL ZONE WITH LINGULATICERAS
PROCRASPEDITES
CORYCERAS ASSEMBLAGE
TARAMELLICERAS AULACOSTEPHANUS ASS.
Subneumayria-Epicephalites
UPPER IDOCERAS ASSEMBLAGE
LOWER IDOCERAS ASSEMBLAGE
SCHNEIDIA ASSEMBLAGE

Lower *Idoceras* Assemblage without *Sutneria*

Soil cover and/or marly-shaly deposition has resulted in the lowest record of this ammonite assemblage not being from strata directly overlying those with the earlier *Schneidia* assemblage, from which it is separated by a metre-thick interval lacking ammonites. *Idoceras* shows higher abundance and diversity than in upper horizons (Lara, 1994). *Idoceras johnsoni* (Burckhardt), *I.* sp. cf. *complanatum* (Burckhardt), *I. striatum* (Imlay), and *I. figueroae* (Burckhardt) were found to be characteristic, and *I. zacatecanum* (Burckhardt) and *I. neogaeum* (Burckhardt) attain their maximum abundance. Other "species" of this genus were also collected from upper horizons together with *Sutneria* of the *cyclodorsata* (Moesch) group. No other ammonites collected (incomplete *Glochiceras*, *Lingulaticeras*, and *Taramelliceras*, together with rare and small *Physodoceras*, *Aspidoceras* and perisphinctids) provided relevant data for biochronostratigraphy and correlation, except for the record of very scarce fragments of *Ataxioceras* in the lowermost range of the genus *Idoceras*. At present, data supporting biostratigraphic subdivisions of this ammonite assemblage are inconclusive.

Upper *Idoceras* Assemblage with *Sutneria*

Sampling conditions for this upper assemblage are more favorable and therefore its boundaries are clearly established. The FAD of the *Sutneria* group of *cyclodorsata* (Moesch) identifies the lower boundary while the LAD of *Idoceras* marks the upper boundary. In this assemblage, both abundance and diversity of *Idoceras* decrease progressively upwards. No "species" of *Idoceras* exclusive of this assemblage have been found, hence recorded *Idoceras* (*zacatecanum*, *subdedalum*, *tamaulipanum*, *neogaeum*, sp. cf. *mexicanum*, and *lorioli*) were considered to be long ranging "species". On the whole, the ammonite assemblage is more diversified than in the underlying horizons with *Idoceras*, but contains taxa with a low diagnostic value, among which were identified incomplete bituberculate aspidoceratids and badly preserved glochiceratids and *Taramelliceras*. Potentially more valuable are *Nebrodites*, *Subneumayria* and *Epicephalites*, all of which were collected from the upper part of the *Idoceras* range where the existence of a biohorizon with *Subneumayria* and *Epicephalites* is preliminarily envisaged.

The bio- and biochronstratigraphic interpretations of the "*Idoceras* beds" first reported by Burckhardt (1906) varied among subsequent authors due to inaccurate biostratigraphic data, outcrop limitations and selective sampling of calcareous concretions with well preserved ammonites. Together, these problems resulted in a sudden revival of the earliest hypotheses (Contreras et al., 1988). Recent revisions by Callomon (1992) of Upper Jurassic and lowermost Cretaceous ammonite assemblages reported from Mexico included the tentative reversal of the stratigraphic order as suggested by Imlay (1939) for ammonite assemblages characterizing his "*Idoceras* beds". Callomon (1992) assumed stratigraphic differences between two major ammonite assemblages dominated by *Idoceras*, as reported by Imlay. In fact, the ammonite assemblages involved were gathered from separate areas, such as San Pedro del Gallo in Durango (Burckhardt, 1912), several outcrops in Mazapil, Zacatecas (Burckhardt, 1906; Imlay, 1939), and Sª de la Ventura, San Lázaro Anticline and La Escondida-Soledad in Nuevo León (Imlay, 1939). Ammonite assemblages with *Idoceras* collected from San Pedro del Gallo and Mazapil mainly show the alternative dominance of macroconchiate (together with great size microconchiate?) and microconchiate *Idoceras*, respectively. However, new species described by Imlay (1939) and collected from Nuevo León show *Idoceras* assemblages with varying degrees of morphologic affinity to "species" from both San Pedro del Gallo and Mazapil. Cantú-Chapa (1971, p. 26) found "all varieties of *Idoceras*" (literal translation) to be identified in Huasteca (central-eastern Mexico). Therefore, before interpreting the composition of ammonite assemblages in stratigraphic terms, it would be appropriate to consider taphonomy, as recently done by Olóriz et al. (1997). *Idoceras* assemblages collected from San Pedro del Gallo are richer in bituberculate aspidoceratids (globose shells), while these ammonites are clearly less frequent in the Cañón de San Matias where the majority of microconchiate *Idoceras* were collected by Burckhardt (1906). Our revision of Burckhardt's sections indicates that geographic separation of ammonite assemblages composed mainly of macroconchiate and microconchiate *Idoceras* is only apparent and probably local. Hence we suggest that the noted discrepancies could result mainly from a biostratinomic imprint (i.e. selective post-mortem transportation).

The record of rare *Ataxioceras* within the range of the oldest *Idoceras*, and the first appearance of *Sutneria* of the *cyclodorsata* (Moesch) group exhibiting a range of morphologic diversity within a more diversified ammonite assemblage, suggest the possible subdivision of the "*Idoceras* beds" within the upper Lower Kimmeridgian (intra Divisum Zone). We found *Nebrodites* (*Mesosimoceras*) of *aguilerae-quenstedti* (Burckhardt) affinity, *Subneumayria* and *Epicephalites* within the uppermost range of *Idoceras*, below a single record of *Nebrodites rota* (Burckhardt) and together with *Sutneria* of the group *cyclodorsata* (Moesch). We interpret these assemblages to characterize the upper part of the "Upper *Idoceras* assemblage with *Sutneria*", to which we assign a latest Early to early Late Kimmeridgian age (two-fold division). Failure to collect the assemblage with *Streblites* spp. (Burckhardt, 1912; Imlay, 1939) could result from difficulties in identifying this biohorizon in the San Pedro del Gallo area. According to our data, two alternative stratigraphic intervals could correspond to the record of *Streblites* spp.: in either the lowermost or uppermost range of *Idoceras*. The lower one is usually covered, while the upper could be slightly higher than that identified with *Subneumayria* and *Epicephalites*, partially overlapping the uppermost range of *Idoceras* in the area investigated. A precise biochronostratigraphic interpretation of the single record of *Nebrodites rota* (Burckhardt) cannot be made.

The "Lower *Idoceras* assemblage without *Sutneria*" encompasses a stratigraphic interval slightly higher and wider than the MZ-K3 stratigraphic interval of Villaseñor (1991), while the "Upper *Idoceras* assemblage with *Sutneria*" is correlated with the upper Divisum Zone and an indeterminate part of the lower Upper Kimmeridgian (two-fold division) (Fig. 2). The correlation of these two assemblages with *Idoceras*, with and without *Sutneria*, with the biostratigraphic subdivisions of the "*Idoceras* beds" proposed by Imlay (1939) and the faunal horizons M6 and M7 of Callomon (1992) is unclear (see above).

"*Taramelliceras - Aulacostephanus* Assemblage"

In horizons above the LAD of *Idoceras*, this assemblage was locally identified in a calcareous concretion 1.5m below the widespread black, silty limestone bank with abundant *Glochiceras* (*Coryceras*) *carinatum* (Aguilera) and, therefore, is identified as a reference level in northern and central Mexico. Haploceratids [*G.* (*Lingulaticeras*) sp. and secondary *Taramelliceras* ("*Metahaploceras*") sp.] overwhelmingly dominate. *Taramelliceras* sp. gr. *pseudoflexuosum* (Favre) and a single specimen of *Aulacostephanus* sp. are significant for bio- and biochronostratigraphy, indicating an Upper Kimmeridgian horizon within the Eudoxus Zone (Fig. 2).

The biohorizon with *Taramelliceras* sp. gr. *pseudoflexuosum* (Favre) and *Aulacostephanus* sp. represents the MZ-K5 stratigraphic interval of Villaseñor (1991), being slightly younger than previously interpreted by this author, and correlates with the lower part of faunal horizon M8 of Callomon (1992).

"*Coryceras - Procraspedites* Assemblage"

This ammonite assemblage typically characterizes the black silty limestone bank which is a clear stratigraphic datum in the areas investigated. *Glochiceras* (*Coryceras*) *carinatum* (Aguilera) dominates and *Procraspedites praecursor* (Burckhardt s. Olóriz, Villaseñor and González-Arreola, 1998) is a rare but typical component. Other ammonites collected were *Haploceras transatlanticum* Burckhardt, *Glochiceras* (*Lingulaticeras*) *semicostatum* Berckhemer, *Taramelliceras* ("*Metahaploceras*") *costatum* (Burckhardt *non* Quenstedt) and rare, coarse-ribbed perisphinctids similar to *Pachysphinctes* (under study).

According to Olóriz, Villaseñor and González-Arreola (1998), the biohorizon with *Glochiceras* (*Coryceras*) *carinatum* (Aguilera) and *Procraspedites praecursor* (Burckhardt) belongs to the Eudoxus Zone (Fig. 2). The stratigraphic interval involved represents the upper part of the stratigraphic interval MZ-K6 of Villaseñor (1991), the complete identification of which can be difficult due to poor preservation of fossils in siltstones. MZ-K6 of Villaseñor (1991) correlates with the chronostratigraphic interpretation made by Burckhardt (1930) of his "Couches à Haploceras Fialar" and the M8 faunal horizon of Callomon (1992).

"Interval Zone with *Lingulaticeras*"

Directly overlying the "*Coryceras - Procraspedites* assemblage" and below the registered FAD of *Hybonoticeras*, an approximately 1m thick section of siltstones with *Glochiceras* (*Lingulaticeras*) *semicostatum* (Berckhemer) has been identified. The ammonite assemblage is dominated by haploceratids, but unfavorable preservation impedes the precise identification of the major part of the ammonite fauna. Rare and incomplete *Pseudowaagenia* sp. gr. *haynaldi* (Herbich), as well as more frequent fragments of Lithacoceratinae (*Procraspedites*-like ammonites), were recognized.

The "Interval Zone with *Lingulaticeras*" belongs to a higher part of the Eudoxus Zone (Fig. 2) and represents the stratigraphic interval MZ-K7 of Villaseñor (1991), filling part of the "stratigraphic gap" between faunal horizons M8 and M9 of Callomon (1992).

The Kimmeridgian-Tithonian Boundary

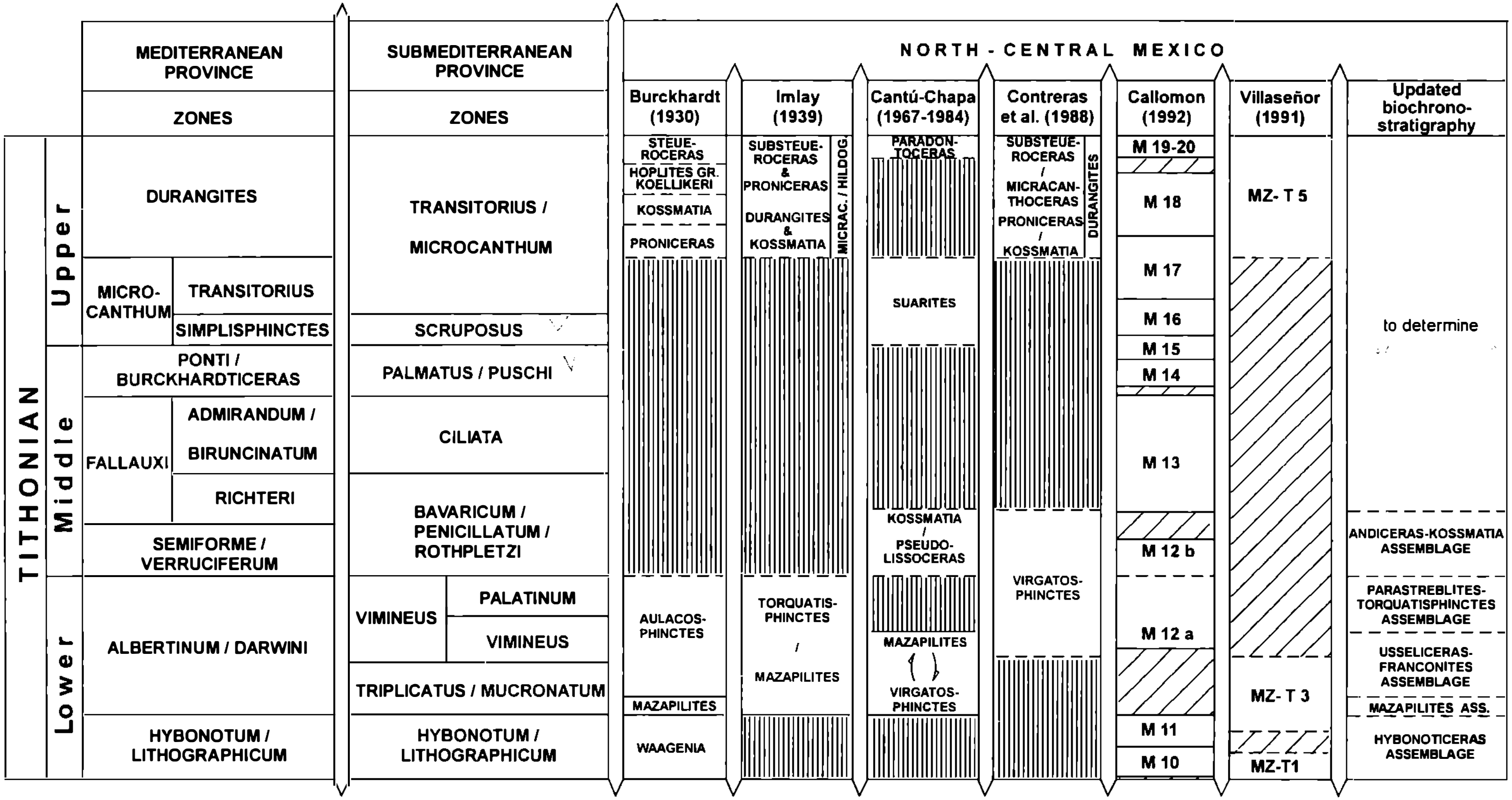
In northern and central Mexico the precise identification of the Kimmeridgian-Tithonian boundary, as it is traditionally accepted in Europe (Geyssant and Enay, 1991; Hantzpergue et al., 1991), i.e. the Beckeri/Hybonotum Zone boundary, is difficult to establish due to poor preservation and scarcity of ammonites accompanying *Hybonoticeras*. The stratigraphic interval containing *Hybonoticeras* in Mexico has been assigned to the uppermost Kimmeridgian and lowermost Tithonian (Imlay, 1984; Olóriz et al., 1990a, 1993; Villaseñor, 1991; Callomon,1992; Olóriz, 1992; Villaseñor et al., 1994; 1995; Olóriz and Villaseñor, in press) including a variety of interpretations of these stages (Burckhardt, 1906; 1930; Imlay, 1939; Cantú-Chapa, 1971; Contreras et al., 1988). The recognition of Mexican specimens of *Hybonoticeras* belonging to the *beckeri* group started with the pioneer paper by Burckhardt (1906), but relation of Mexican specimens to the *hybonotum* group remains problematic. At present, the most precise approach to the Kimmeridgian/Tithonian boundary in Mexico was made by Olóriz et al. (1993) who used the FAD of *Subplanitoides siliceus* (Quenstedt) (presently reconsidered as *Pseudodiscosphinctes*) in Durango to interpret the base of the Tithonian in the area. The assumed biochronostratigraphic significance proposed by Schweigert (1993) for the reinterpretation of that species made by Sapunov (1979) is dubious. The precise recognition of new microconch species described by Olóriz and Villaseñor (in press) is the only way available for recognition of the Kimmeridgian/Tithonian boundary in the Mexican Altiplano.

On the basis of the new material collected, the assumed assemblage with *Hybonoticeras* and *Mazapilites*, proposed by Burckhardt (1919-21), recognized by Imlay (1939; at the top of Unit 7, just underlying beds with *Mazapilites* of Unit 8) and accepted by Cantú-Chapa (1967) and Callomon (1992), resulted from condensation in phosphatic limestones. Hence, the stratigraphic interval with *Mazapilites* is higher within the Lower Tithonian, as proved by our sampling of extended sections exposing pinkish siltstones.

"*Hybonoticeras* Assemblage"

This assemblage is characterized by *Hybonoticeras* sp. gr. *beckeri* (Neumayr) and *Hybonoticeras mundulum* (Oppel). New microconch species described by Olóriz and Villaseñor (in press) permit identification within the Hybonotum Zone of a lower interval with *H.* n. sp. aff. *striatulum* Berckhemer and Hölder, *H. geminatum* n. sp., *H. gonzalezi* n. sp., and *H. cuencamensis* n. sp., all belonging to the Beckeri Zone, and an upper one with *H. evanidum* n. sp. Accompanying ammonites in the lower assemblage are *H. harpephorum* (Neumayr), *Glochiceras* (*Lingulaticeras*) *semicostatum* (Berckhemer), and *G.* (*Lingulaticeras*) sp. gr. *contractum* (Quenstedt). In the upper assemblage *Pseudodiscosphinctes siliceus* (Quenstedt) and close forms, as well as *Taramelliceras rebouletianum* (Fontannes), are significant for correlations; other ammonites collected are *T.* sp. cf. *subnudatum* (Fontannes) and small incomplete aspidoceratids likely close to *Schaireria neoburgensis* (Oppel). *G.* (*Lingulaticeras*) sp., *Haploceras* sp., *Lithacoceras* sp. and indeterminate fragments of Ataxioceratidae become increasingly abundant throughout the range of *Hybonoticeras*.

Figure 3. Correlation of chart of recently collected ammonite assemblages in north-central Mexico. Proposals by Cantú-Chapa, Imlay, Contreras et al., and Burckhardt are reinterpreted. Those by Villaseñor and Callomon as originally proposed. Discontinuity in ammonite assemblages: reinterpreted (vertical ruling), originals (oblique ruling). Gray color for ammonite assemblages of Berriasian age. European standards according to Geyssant and Enay (1991).


TITHONIAN
Upper
Middle
Lower
MEDITERRANEAN PROVINCE
ZONES
DURANGITES
MICRO-CANTHUM
TRANSITORIUS
SIMPLISPHINCTES
PONTI / BURCKHARDTICERAS
FALLAUXI
ADMIRANDUM /
BIRUNCINATUM
RICHTERI
SEMIFORME / VERRUCIFERUM
ALBERTINUM / DARWINI
HYBONOTUM / LITHOGRAPHICUM
SUBMEDITERRANEAN PROVINCE
ZONES
TRANSITORIUS /
MICROCANTHUM
SCRUPOSUS
PALMATUS / PUSCHI
CILIATA
BAVARICUM / PENICILLATUM / ROTHPLETZI
VIMINEUS
PALATINUM
VIMINEUS
TRIPLICATUS / MUCRONATUM
HYBONOTUM / LITHOGRAPHICUM
NORTH - CENTRAL MEXICO
Burckhardt (1930)
STEUE-ROCERAS
HOPLITES GR. KOELLIKERI
KOSSMATIA
PRONICERAS
AULACOS-PHINCTES
MAZAPILITES
WAAGENIA
Imlay (1939)
SUBSTEUE-ROCERAS & PRONICERAS
DURANGITES & KOSSMATIA
MICRAC. / HILDOG.
TORQUATIS-PHINCTES
/
MAZAPILITES
Cantú-Chapa (1967-1984)
PARADON-TOCERAS
SUARITES
KOSSMATIA / PSEUDO-LISSOCERAS
MAZAPILITES
( )
VIRGATOS-PHINCTES
Contreras et al. (1988)
SUBSTEUE-ROCERAS / MICRACAN-THOCERAS
PRONICERAS / KOSSMATIA
DURANGITES
VIRGATOS-PHINCTES
Callomon (1992)
M 19-20
M 18
M 17
M 16
M 15
M 14
M 13
M 12 b
M 12 a
M 11
M 10
Villaseñor (1991)
MZ- T 5
MZ- T 3
MZ-T1
Updated biochrono-stratigraphy
to determine
ANDICERAS-KOSSMATIA ASSEMBLAGE
PARASTREBLITES-TORQUATISPHINCTES ASSEMBLAGE
USSELICERAS-FRANCONITES ASSEMBLAGE
MAZAPILITES ASS.
HYBONOTICERAS ASSEMBLAGE

The lower and upper assemblages with *Hybonoticeras* encompass, respectively, part of the Beckeri and Hybonotum Zones (Figs. 2-3). The absence of known records of *Hybonoticeras* of the *pressulum-verestoicum* groups in the Mexican Altiplano indicates missing evidence for the lower Beckeri Zone (Olóriz and Villaseñor, in press). Further research is necessary to establish the precise identification of the reported Hybonotum Zone. The lower subdivision of the "*Hybonoticeras* assemblage" correlates with both the MZ-K8 stratigraphic interval of Villaseñor (1991), except for its lower part, and the faunal horizon M9 proposed by Callomon (1992) who suggested correlation with the Beckeri Zone without further precision. The upper subdivision of the "*Hybonoticeras* assemblage" correlates with the stratigraphic interval MZ-T1 of Villaseñor (1991), who originally assumed it correlated with a shorter interval within the Hybonotum Zone. The stratigraphic interval that we interpret for the upper assemblage with *Hybonoticeras* encompasses the stratigraphic interval corresponding to faunal horizon M10 in Callomon (1992), the location of which was incorrectly envisaged by him. According to previous comments on the combined record of *Hybonoticeras* and *Mazapilites*, the faunal horizon M11 as envisaged by Callomon could result from condensation, and therefore the stratigraphic range interpreted by him is shorter than indicated by our field data.

The Lower Tithonian above *Hybonoticeras*

"*Mazapilites* Assemblage"

Diverse "species" of *Mazapilites* overwhelmingly dominate this assemblage. Secondary components are taramelliceratinae resembling the small and older *Taramelliceras multinodum* Berckhemer and Hölder, and small *Schaireria* of the *neoburgensis* (Oppel) group. Fragments of small size perisphinctids are difficult to identify, some of them resembling *Subplanitoides*. This composition of the "*Mazapilites* assemblage" is typical of non-condensed siltstone sections.

Overlying the "*Hybonoticeras* assemblage" and below horizons with *Torquatisphinctes*, the "*Mazapilites* assemblage" is correlated with a lower part of the Albertinum/Darwini Zone which is difficult to identify (Fig. 3). The "Beds with *Mazapilites*" were defined as faunal horizon M12 by Callomon (1992) and subdivided into M12(a) [combining species from diverse horizons in Durango following Burckhardt (1919-21; from "Capas superiores con *Mazapilites*" and the "Calizas grises con *Aulacosphinctes*") and Imlay (1939; Unit 8 colls. 9, 10?, 11, 12, 13, 14, and 15)], and M12(b) [yielding ammonite assemblages from condensed deposits in Zacatecas ("Calcaires phosphoritiques rougeâtres" of Burckhardt, 1906)]. Callomon (1992) suggested the stratigraphic interval of M12(a) and M12(b), assuming it to be hypothetical, encompassed a large part of the Albertinum/Darwini and Semiforme/Verruciferum Chrons. The stratigraphic interval MZ-T3 of Villaseñor (1991) is condensed, including the range that we interpret for the "*Mazapilites* assemblage" and that corresponding to the next ammonite assemblage described here.

"*Usseliceras-Franconites* Assemblage"

This ammonite assemblage is typical of condensed deposits labelled by Burckhardt (1906) as "Calcaires phosphoritiques rougeâtres" in the Cañón de San Matias, and includes the so-called Mexican "*Virgatites*". Collected above *Mazapilites* in the areas investigated, these forms are interpreted as endemic macro- and microconch Lithacoceratinae. "*Virgatites*" *mexicanus* Burckhardt (see Cantú-Chapa, 1984 and Burckhardt, 1906, pl. 21, figs. 7-8) is morphologically closer to the *Franconites* group of *vimineus* (Schneid) than *tenuiplicatus* Zeiss. Other "species" (see *Virgatites* sp. in Burckhardt, 1906, pl. 32, fig. 2) could resemble the *Usseliceras* group of *franconicum-fascipartitum* Zeiss and/or *Sublithacoceras subdiffusum* Zeiss, when coarse ribs are developed. Specimens with coarser ribbing are morphologically close to the groups of *Franconites sparsicostatus* Zeiss and *Danubisphinctes multipartitum-subviperinum* Zeiss. Among microconchs we identified *Subplanitoides*, rather than *Subdiscosphinctes*. These ammonites show only bifurcating ribbing as in the Submediterranean group of *Usseliceras altegyratum* Zeiss and other strictly Mediterranean species; others are coarse ribbed specimens which resemble the *Parapallasiceras lateumbilicatum* Zeiss species group.

The "*Usseliceras-Franconites* assemblage" is interpreted to belong to the Albertinum/Darwini Chron (Fig. 3). The morphologic resemblance of these ammonites with older species from the uppermost Kimmeridgian (Beckeri Zone) is clear, but results from convergence as shown by their stratigraphic position. The "*Usseliceras-Franconites* assemblage" represents the M10 faunal horizon identified by Callomon (1992) who misinterpreted its stratigraphic level, suggesting it occurred below *Hybonoticeras* and *Mazapilites* (his faunal horizons M11 and M12). The condensed stratigraphic interval MZ-T3 of Villaseñor (1991) includes the "*Usseliceras-Franconites* assemblage", but was originally proposed to have a shorter range within the Albertinum/Darwini Zone.

"*Parastreblites-Torquatisphinctes* Assemblage"

This assemblage was identified in condensed phosphatic limestones by Burckhardt (1906; 1912; 1919-21) and confirmed by our field work. Significant for biostratigraphy are *Parastreblites mazapilensis* (Burckhardt), *P. circumnodosum* (Fontannes) and *Pseudolissoceras* of the *rasile-planiusculum* (Zittel) groups. Among microconch perisphinctids, *Torquatisphinctes* is typical together with less frequent *Subdichotomoceras* and probable *Biplisphinctes* (many "species" of "*Aulacosphinctes*" in Burckhardt, 1919-21); *Aulacosphinctoides* (*aguilerae* group) is subordinate. Macroconch perisphinctids are poorly known ("*Virgatites*" sp. in Burckhardt, 1919-21, pl. 14, fig.1-3), including rare and very local forms ("*Perisphinctes*" *theodosii* Burckhardt) difficult to assign to known taxa (research in progress). Other ammonites are *Schaireria neoburgensis* (Oppel) and other close "species" (*Aspidoceras fallax*, *A. cajense*, *A. zacatecanum* in Burckhardt, 1906), haploceratids (*Neochetoceras* sp. and a probable new genus), and rare phylloceratids.

The "*Parastreblites-Torquatisphinctes* assemblage" contains *Parastreblites* and *Pseudolissoceras* typical of the Albertinum/Darwini (to lowermost Semiforme/Verruciferum) Zone, as well as perisphinctids with dominant bifurcate and single ribs (Torquatisphinctinae) showing varying similarity to those known from the same age in the Mediterranean Tethys and surrounding areas. The proliferation in globose-smooth aspidoceratids is a well known fact among Tethyan (Mediterranean) ammonite assemblages from the. Albertinum/Darwini (to earliest Semiforme/ Verruciferum) Chron.The "*Parastreblites-Torquatisphinctes* assemblage" represents the majority of the ammonite species included in the stratigraphic interval MZ-T3 of Villaseñor (1991), which was considered to be older, as well as the *Torquatisphinctes* assemblage included in faunal horizon M12 by Callomon (1992) and interpreted by that author as extending upward to include a slightly longer stratigraphic interval than we allow, based on the material collected (Fig. 3).

"*Andiceras-Kossmatia* Assemblage"

This assemblage is typical from Sierra de Catorce (San Luis Potosí) where it characterizes the so-called "Virgatosphinctinae beds" of Verma and Westermann (1973), but some of their more typical components were previously identified by Burckhardt (1906, pls. 35, 36, 37) in his "Calcaires phosphoritiques grisâtres" at the Cañón de San Matias section (Zacatecas). Therefore, the "*Andiceras-Kossmatia* assemblage" is only known from condensed deposits in the Mexican Altiplano, and the composition of ammonite assemblages in separate areas should be expected to vary.

The most complete assemblage is that studied by Verma and Westermann (1973) and later revised through field work by Olóriz et al. (1996; Olóriz, Villaseñor, González-Arreola, and Westermann, 1998). According to the latter authors this ammonite assemblage is composed of *Andiceras* spp., *Kossmatia* spp., haploceratids (*H. transatlanticum* Burckhardt, *H.* sp.), globose aspidoceratids, smooth [*Schaireria neoburgensis* (Oppel)] and bituberculate (*Aspidoceras rogoznicense* [(Zeuschner), *A. rafaeli* (Oppel)], *Pseudolissoceras zitteli* (Burckhardt), *Simoceras* of the group of *aesinense* (Meneghini), and rare *Lemencia*, accompanied by indeterminate phylloceratids and fragmented perisphinctids (remains of macroconch specimens included).

On the basis of the ammonites identified, Olóriz et al. (1996; Olóriz, Villaseñor, González-Arreola, and Westermann, 1998) interpreted the "*Andiceras-Kossmatia* assemblage" to belong to the Semiforme/Verruciferum Chron, assuming the possibility of including the earliest Richteri Chron (Fig. 3). Therefore, the stratigraphic level suggested by Callomon (1992) for the"Virgatosphinctinae beds" of Verma and Westermann, 1973, and its equivalent faunal horizon M13, is younger than here admitted (see discussion in Olóriz, Villaseñor, González-Arreola, and Westermann, 1998). Callomon (1992) considered *Kossmatia* as occurring separately from this ammonite assemblage, and in association with *Substeueroceras*, as characterising his M19 faunal horizon, interpreting a latest Tithonian and probably Berriasian age for his M19 and M20 faunal horizons. We propose that these two horizons are clearly Berriasian in age.

Upper Tithonian

The record of Upper Tithonian ammonites in Mexico is controversial, especially in the Mexican Altiplano. The usual combination of defective sampling and imprecise stratigraphy, the use of the Portlandian by pioneers (Burckhardt, 1906; 1912; 1919-21; 1930; Imlay, 1939), and later the application of a more or less variable Tithonian stage (Imlay, 1943; 1980; 1984; Cantú-Chapa, 1976a; 1976b; 1979; 1980; 1982; 1984; 1989; Verma and Westermann, 1973; Jeletzky, 1984; Contreras et al., 1988) have been involved. Scarcity of well preserved ammonite assemblages with calpionellids (taphonomic or ecologic factors ?), together with particular reinterpretations of their ranges (Salinas in Cantú-Chapa, 1967; 1980; Cantú-Chapa, 1989; Longoria, 1984), resulted in uncertainties when compared with data and interpretations provided by Bonet (1956), Trejo (in Borza, 1969; 1975, in Remane, 1982). Traditionally, Mexican ammonite assemblages interpreted as belonging to the Upper Tithonian contain genera and species that are typically collected from Berriasian deposits in Tethyan areas, as discussed by Olóriz and Tavera (1989). Recent investigations on calpionellids by Adatte et al. (1992; 1994a; 1994b; Adatte, Stinnesbeck and Remane, 1994) and Stinnesbeck et al. (1993), clearly demonstrate this fact. In addition, ammonite assemblages recorded by the latter authors just below horizons interpreted as indicative of boundary beds between calpionellid Zones A and B could be Berriasian according to biostratigraphy in Tavera et al. (1994), who provided the most precise correlation of ammonites and calpionellids in the Mediterranean Tethys. This affects the chronostratigraphic interpretation of the *Salinites-Corongoceras-Durangites* assemblage identified by Adatte et al. (1992; 1994a). Moreover, in the absence of calpionellids, the record of ammonite assemblages containing *Pseudosubplanites* should be interpreted as Berriasian rather than Upper Tithonian, as proposed by Adatte at al. (1994a; Adatte, Stinnesbeck and Remane, 1994). As recognized by Adatte, Stinnesbeck and Remane, (1994) and Olóriz, Villaseñor, González-Arreola, and Westermann (1998), the precise identification of the Tithonian/Berriasian boundary, as accepted in Europe (Remane et al., 1986; Tavera et al., 1994), seems to be difficult in Mexico at present. There is a scanty record of poorly known ammonite assemblages which could be Upper Tithonian, at least in part, such as the ammonite assemblage with *Suarites bituberculatum* but lacking calpionellids (Cantú-Chapa, 1967; 1989), together with a well known absence of Upper Tithonian ammonites typical of Tethyan areas. The significance of these data needs to be investigated, and the identification of hiatuses should be crucial, especially where sharp changes in lithofacies are recorded. Olóriz, Villaseñor, González-Arreola and Westermann (1998) inferred the existence of hiatuses above condensed deposits in Sierra de Catorce, between the El Pastor and El Verde Members of the La Caja Formation in the Alamitos area. Therefore, more research is needed for the precise identification and correlation of Upper Tithonian fossil assemblages in northern and central Mexico.

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References

ADATTE, T., W. STINNESBECK, H. HUBBERTEN, AND J. REMANE. 1994a. Correlaciones multiestratigráficas en el límite Jurásico-Cretácico en el noreste de México. Boletín de la Sociedad Geológica Mexicana, 51(1-2):23-51.

____, ____, H. HUBBERTEN, AND J. REMANE. 1994b. Nuevos datos sobre el límite Jurásico-Cretácico en el noreste y en el centro de México. Boletín de la Sociedad Geológica Mexicana, 52(1-2):11-14.

____, ____, AND J. REMANE. 1994. The Jurassic-Cretaceous boundary in northeastern Mexico. Confrotation and correlations by microfacies, clay minerals mineralogy, calpionellids and ammonites. Geobios Mémoire special 17:37-56.

____, ____, ____, AND H. HUBBERTEN. 1992. The Jurassic-Cretaceous boundary in northeastern and central Mexico. A multistratigraphical approach. Simposios III Congreso Geológico de España y VIII Congreso Latinoamericano de Geología, 4:23-29.

BONET, F. 1956. Zonificación microfaunística de las calizas Cretácicas del Este de México. Boletín de la Asociación Méxicana de Geólogos Petroleros, 8:389-488.

BORZA, K. 1969. Die Mikrofazies und Mikrofossilien des Oberjuras und der Unterkreide der Klippenzone der Westkarpaten. Verlag der Slowakischen Akademie der Wissenschaften, Bratislava, 296 p.

BURCKHARDT, C. 1906. La faune Jurassique de Mazapil avec un appendice sur les fossiles de Crétacique inférieur. Boletín del Instituto Geológico de México, 23:1-217.

___. 1912. Faunes Jurassiques et Cretaciques de San Pedro del Gallo (Durango, Mexico). Boletín del Instituto Geológico de México, 29:1-264.

___. 1919-21. Faunas jurásicas de Symón y faunas cretácicas de Zumpango del Río. Boletín del Instituto Geológico de México, 33:1-135.

___. 1930. Etude synthétique sur le Mésozoïque mexicain. Mémoires de la Société Paléontologique Suisse, 49-50:1-280.

CALLOMON, J. H. 1992. Upper Jurassic, especially Mexico, p. 261-272. *In* G. E. G. Westermann (ed.), The Jurassic of the Circum Pacific. Cambridge University Press, New York.

CANTU-CHAPA, A. 1967. El límite Jurásico-Cretácico en Mazatepec, Puebla. Estratigrafía del Jurásico de Mazatepec, Puebla (México). Instituto Mexicano de Petróleo, Sección Geología, Monografía 1:3-24.

___. 1969. Estratigrafía del Jurásico Medio-Superior del subsuelo de Poza Rica, Veracruz (Área de Soledad-Miquetla). Revista del Instituto Mexicano del Petróleo 1(1):3-9.

___. 1970. El Kimeridgiano inferior de Samalayuca, Chihuahua. Revista del Instituto Mexicano del Petróleo, 2(3):40-44.

___. 1971. La Serie Huasteca (Jurásico Medio-Superior) del centro este de México. Revista del Instituto Mexicano del Petróleo, 3(2):17-40.

___. 1976a. Nuevas localidades del Kimeridgiano y Titoniano en Chihuahua (Norte de México). Revista del Instituto Mexicano del Petróleo, 8:38-45.

___. 1976b. El contacto Jurásico-Cretácico, la estratigrafía del Neocomiano, el Hiato Hauteriviano superior-Eoceno inferior y las amonitas del Pozo Bejuco 6 (centro-este de México). Boletín de la Sociedad Geológica Mexicana 37:60-83.

___. 1979. Bioestratigrafía de la Serie Huasteca (Jurásico Medio-Superior) en el subsuelo de Poza Rica, Veracruz. Revista del Instituto Mexicano del Petróleo, 9(2):14-24.

___. 1980. El límite Jurásico-Cretácico en México. Memoria del I Congreso Latinoamericano de Paleontología, 5:177-184. Argentina

___. 1982. The Jurassic-Cretaceous boundary in the subsurface of Eastern Mexico. Journal of Petroleum Geology, 4(2):311-318.

___. 1984. El Jurásico Superior de Tamán, San Luis Potosí, Este de México. Memoria del III Congreso Latinoamericano de Paleontología, 207- 212. México.

___. 1989. Precisiones sobre el límite Jurásico-Cretácico en el subsuelo del Este de México. Revista de la Sociedad Mexicana de Paleontología, 2(1):26-69.

CONTRERAS, B., A. MARTÍNEZ, AND M. E. GÓMEZ. 1988. Bioestratigrafía y sedimentología del Jurásico Superior en San Pedro del Gallo, Durango, México. Revista del Instituto Mexicano del Petróleo, 20(3):5-49.

GEYSSANT J., AND R. ENAY. 1991. Tithonique. 3rd International Symposium on Jurassic Stratigraphy Abstracts, 134.

HANTZPERGUE, P., F. ATROPS, AND R. ENAY. 1991. Kimmeridgien. 3rd International Symposium on Jurassic Stratigraphy Abstracts, 133.

IMLAY, R. W. 1939. Upper Jurassic ammonites from Mexico. Geological Society of America Bulletin, 50:1-78.

___. 1943. Upper Jurassic ammonites from the Placer de Guadalupe district, Chihuahua, Mexico. Journal of Paleontology, 17(5):527-543.

___. 1980. Jurassic paleobiogeography of the conterminous United States in its continental setting. Geological Survey Professional Paper 1062, 134 p.

___. 1984. Jurassic ammonite successions in North America and biogeographic implication, p. 1-12. *In* G. E. G. Westermann (ed.). Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper 27.

JELETZKY, J. A. 1984. Jurassic-Cretaceous boundary beds of Western and Artic Canada and the problem of the Tithonian-Berriasian Stages in the Boreal Realm, p. 175-256. *In* G. E. G. Westermann (ed.). Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper 27.

LARA, L. 1994. Revisión del género *Idoceras* Burckhardt, 1906 (Ammonitina). Análisis númericos e interpretaciones paleobiológicas. Unpublished M.D. dissertation, Facultad de Ciencias, Universidad Nacional Autónoma de México, 122p.

LINARES, A., F. OLÓRIZ, AND A. B. VILLASEÑOR. 1997. Presencia de *Tropidoceras flandrini* (Dumortier) en Pozo Serna, Sonora (México). Revista Española de Paleontología, 12(2):257-264.

LONGORIA, J. 1984. Stratigraphic studies in the Jurassic of northeastern Mexico: evidence for the origin of the Sabinas Basin, p. 171-194. *In* W. P. S. Ventres, D. G. Bebout, B. F. Perkins, and C. H. Moore (eds.). The Jurassic of the Gulf Rim. Proceedings of the Third Annual Research Conference. Golf Coast Section SEPM.

MYCZYNSKI, R., F. OLÓRIZ, AND A. B. VILLASEÑOR. 1998. Biostratigraphy and correlations of the Middle-Upper Oxfordian in the Americas (southern USA, Mexico, Cuba, and northern Chile). Neues Jahrbuch fur Geologie und Paläontologie Abhandlungen, 207(2):185-206.

OLÓRIZ, F. 1992. North-central and eastern Mexico, p. 100-107. *In* G.E.G. Westermann (ed.), The Jurassic of the Circum-Pacific, part 3, Regional Geology and Stratigraphy, Meso-America. Cambridge University Press, New York.

____, C. GONZÁLEZ-ARREOLA, L. LARA, A. B. VILLASEÑOR AND G. E. G. WESTERMANN. 1992. Significant Ammonites and Calpionellids for correlations within the Upper Jurassic-Lowermost Cretaceous in the southern margin of the North American Plate. Table Ronde Européene, Lyon, Francia, Abstracts, 38.

____, L. LARA, A. DE LA MORA, A. B. VILLASEÑOR, AND C. GONZÁLEZ-ARREOLA. 1993. The Kimmeridgian/Tithonian boundary in the "Barranquito del Alacrán" section at Cuencamé (Durango, Mexico); its biostratigraphy and ecostratigraphic interpretation. Acta Geologica Polonica, 43 (3-4):273-288.

____, A. LÓPEZ, A. B. VILLASEÑOR, AND C. GONZÁLEZ-ARREOLA. 1988. Análisis isotópicos y consideraciones paleoecológicas en el Jurásico Superior de México (Fm. La Casita, Cuencamé, Durango). Datos preliminares. II Congreso Geológico de España, 1:144-148.

____, J. M TAVERA. 1989. The significance of Mediterranean ammonites with regard to the traditional Jurassic-Cretaceous boundary. Cretaceous Research, 10: 221-237.

____, AND A. B. VILLASEÑOR. In press. New microconchiate *Hybonoticeras* from Mexico. Geobios.

____, A. B. VILLASEÑOR, AND C. GONZÁLEZ-ARREOLA. 1997. Factors controlling Upper Jurassic ammonite assemblages in North-central Mexico. Lethaia, 30(4):337-351.

____, A. B. VILLASEÑOR, AND C. GONZÁLEZ-ARREOLA. 1998. Re-evaluation of *Procraspedites* Spath, 1930 (Ammonitina) from the upper Kimmeridgian of Mexico. Bulletin de la Societé géologique de France, 169(2):243-254.

____, A. B. VILLASEÑOR, C. GONZÁLEZ-ARREOLA, AND G. E. G. WESTERMANN. 1990a. Las plataformas mexicanas durante el Jurásico Superior. Un ejemplo de áreas de recepción en biogeografía. Actas de Paleontología, IV Jornadas de Paleontología, Salamanca, 68:277-287.

____, A. B. VILLASEÑOR, C. GONZÁLEZ-ARREOLA, AND G. E. G. WESTERMANN. 1990b. Problems of litho-correlation in the Mexico-Caribbean area and the significance of Upper Oxfordian "*Discosphinctes*". I Oxfordian Working Group. UNESCO, España. Publicaciones del Seminario de Paleontología de Zaragoza, 2: 191-204.

____, A. B. VILLASEÑOR, C. GONZÁLEZ-ARREOLA, AND G. E. G. WESTERMANN. 1996. Interpreting updated ammonite biostratigraphy. The Upper Jurassic-lowermost Cretaceous in the Alamitos area, Sierra de Catorce, San Luis Potosí (North-Central Mexico). Abstract volume IV International Symposium Cephalopods- Present and Past, Granada, España, 134-136.

____, A. B. VILLASEÑOR, C. GONZÁLEZ-ARREOLA, AND G.E.G. WESTERMANN. 1998. Ammonite biostratigraphy and correlations in the late Jurassic - earliest Cretaceous La Caja Formation of north-central Mexico. (Sierra de Catorce, San Luis Potosí), p. 463-492. *In* F. Olóriz and F. J. Rodríguez-Tovar (eds.), Advancing Research in Living and Fossil Cephalopods, Plenum Press.

REMANE, J. 1982. Calpionellids and the Jurassic-Cretaceous boundary. Unedited document of the International Subcommission on Jurassic Stratigraphy (Working Group on the Jurassic-Cretaceous Boundary), 16 p.

____, D. BAKALOVA-IVANOVA, K. BORZA, J. KNAUER, I. NAGY, G. POP, AND E. TARDI-FILACZ. 1986. Agreement on the subdivision of the standard calpionellids zones defined at the 2nd. Planktonic Conference, Roma 1970. Acta Geologica Hungarica, 29:5-13.

SAPUNOV, J. G. 1979. Les fossiles de Bulgarie III.3 Jurassique Superieur. Ammonoidea. Academie Bulgare des Sciences, 263 p.

SCHWEIGERT, G. 1993. Die Ammonitengattungen *Gravesia* Salfeld und *Tolvericeras* Hantzpergue in ihre Bedeutung für den Grenzbereich Oberkimmeridgium/Untertithonium im Schwäbischen Jura. Geologie Bl. NO-Bayern, 43:167-186.

STINNESBECK, W., T. ADATTE, AND J. REMANE. 1993. Mazatepec (Estado de Puebla, Mexico) - Reevaluación de su valor como estratotipo del límite Jurásico-Cretácico. Revista Española de Micropaleontología, 25(2):63-79.

TAVERA, J. M., R. AGUADO, M. COMPANY, AND F. OLÓRIZ. 1994. Integrated biostratigraphy of the Durangites and Jacobi Zones (J/K Boundary) at the Puerto Escaño section in Southern Spain (Province of Cordoba). Geobios Mémoire special, 17:469-476.

TREJO, M. 1975. Los Tintínidos mesozoicos de México. Mémoires du Bureau de Recherches Géoligiques et Minières, 86:95-104.

VERMA, H. M., AND G. E. G. WESTERMANN. 1973. The Tithonian (Jurassic) ammonite fauna and stratigraphy of Sierra de Catorce, San Luis Potosí, Mexico. Bulletins of the American Paleontology, 63 (277):107-320.

VILLASEÑOR, A. B. 1991. Aportaciones a la bioestratigrafia, basada en fauna de ammonites, de la sucesión del Jurásico superior (Kimmeridgiano-Tithoniano) del área de Mazapil, Zacatecas, México. Unpublished Ph. D. dissertation, Facultad de Ciencias, Universidad Nacional Autónoma de México, 154 p.

____, L. LARA, F. OLÓRIZ, AND C. GONZÁLEZ-ARREOLA. 1993. Precisiones bioestratigráficas sobre los géneros *Subneumayria* (Spath, 1924) y *Epicephalites* (Spath, 1928) en el centro-norte de México. Memoria de resúmenes IV Congreso Nacional de Paleontología, México, 108-111.

____, F. OLÓRIZ, AND A. DE LA MORA. 1995. Consideraciones sobre el registro del género Hybonoticeras (Ammonitina) en México. V Congreso Nacional de Paleontología, Memoria de Resúmenes, México, D. F., p. 46-47.

____, F. OLÓRIZ, L. LARA, C. GONZÁLEZ-ARREOLA, AND A. DE LA MORA. 1994. Homotaxy and the Kimmeridgian/Tithonian boundary in north-central Mexico. Abstracts 4th International Congress on Jurassic Geology and Stratigraphy, Mendoza, Argentina, 46

ZEISS, A. 1968. Untersuchungen zur Paläontologie der Cephalopoden des Unter-Tithon der Südlichen Franfenalb. Bayerische Akademie der Wissenschaften. Mathematisch-Naturwissenschaftliche Klasse Abhandlunge, 12:7-191.

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New Results of the Jurassic Stratigraphic Study in the Nyalam Area of Southern Tibet

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Keywords: Nyalam, Southern Tibet, Stratigraphy

Abstract: Jurassic strata of Tethys-Himalayas are well exposed in the Nyalam area of southern Tibet. Abundant fossils such as ammonites, bivalves, brachiopods, corals, belemnites, and foraminifers have been found in this area and provide a biostratigraphic framework for Jurassic stratigraphy of this region. New stratigraphic evidence indicates miscorrelation of some of the previous formations, and led to stratigraphic subdivision of the Jurassic into four formations. In ascending order they are the Pupuga, Nieniexiongla, Chamositic Oolite and Menbu formations. The occurrence of an *Orbitopsella-Lithiotis* foraminifera-bivalve assemblage indicates a Liassic age for the Papuga Formation. A *Witchellia-Dorsetensia* ammonite assemblage indicates an Aalenian-Bathonian age of the Nieniexiongla Formation, while a *Macrocephalites-Indocephalites* ammonite assemblage indicates an Early Callovian age for the Chamositic Oolite Formation. The overlying Menbu Formation contains abundant ammonites such as *Blanfordiceras, Aulacosphinctes,* and *Haplophylloceras*, indicating a Middle Callovian to Tithonian age.

Introduction

The Tibetan Plateau resulted from collision between the Indian and Asian continental plates, which produced complex tectonics near the Indus-Yarlung Zangbo suture zone. The relief of the plateau, which exceeds 4000m elevation, is a result of Neogene subduction of the Indian continental lithosphere under Tibet. Mesozoic sedimentary strata outcrop south of the suture zone, where they belong to the passive continental margin of the Indian plate.

In 1991 and 1995 joint geological expeditions of Chinese, Italian and Canadian scientists focused field studies on the Jurassic stratigraphy and sedimentology of northwestern Nepal and southern Tibet. The study area comprises a 1000km wide region of the central Himalayas. The immense size of the area to be covered and working at altitudes of more than 4000m proved a strenuous task. The purpose of this article, therefore, is not to present a definitive study of the Jurassic of southern Tibet, but rather new results which augment previous stratigraphic studies in this region, and new information about the stratigraphy of the southern margin of Gondwana, as represented by the Jurassic of southern Tibet. Jurassic strata in central Nepal studied by Jansa and Sarti have many similarities to the southern Tibetan Jurassic, which allows us to draw broader conclusions about development of the Jurassic of southern Tibet. Comparisons we make between Nepal and southern Tibet, along with our knowledge of the Jurassic of western Tethys (western Europe, North Atlantic), allow us to fit together many jigsaw pieces which previously could not be put together to form an internally consistent picture of the Jurassic. This article deals only with the stratigraphy; the sedimentary development will be published in a separate paper. The proposed stratigraphic framework is not final, as we are aware that future, more detailed studies will result in more detailed bio-stratigraphic subdivisions. This study was conducted in the region of Nyalam, where five sections were measured (Fig. 1).

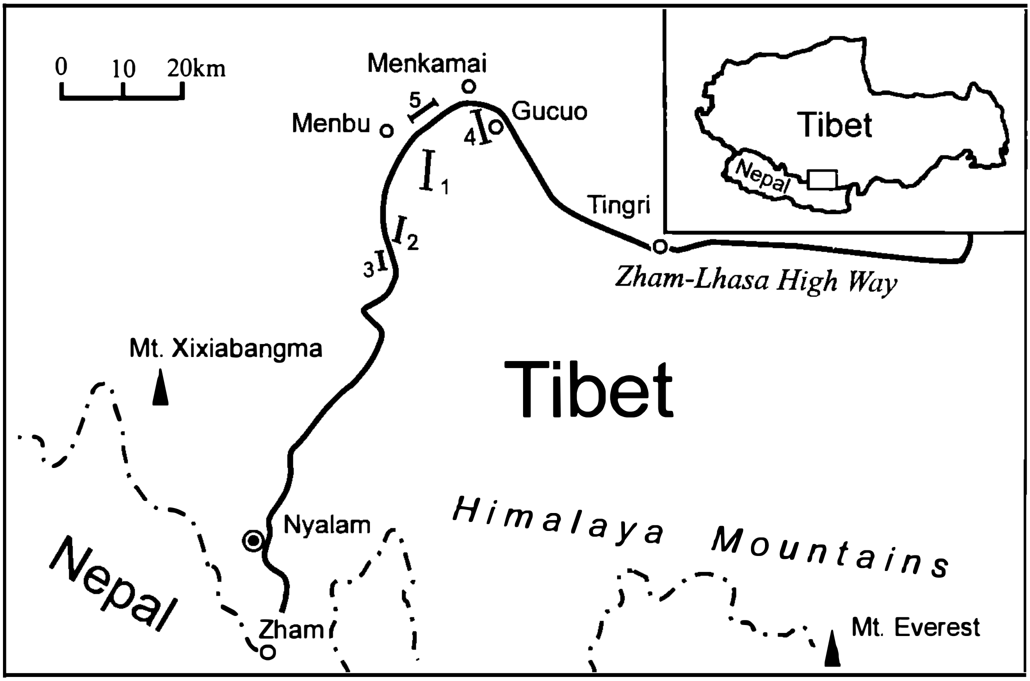

Figure 1. Sketch map showing localities of Jurassic sections in the Nyalam area.
1. Lalongla Section; 2. Nieniexiongla Section; 3. Pupuga Section; 4. Gucuo J/K boundary Section; 5. Xiumo J/K boundary Section.

Stratigraphy

Several biostratigraphic studies have previously been undertaken in the Nyalam region (Wang Yigang and Zhang Mingliang, 1975; Wang Yigang, 1980; Yu Guangming et al., 1983; Westermann and Wang, 1988; Wan Xiaoqiao, 1989; Xu Yuling et al., 1990). They led to stratigraphic subdivision of the Jurassic into five formations in ascending order: the Pupuga, Nieniexiongla, Lalongla, Menbu and Xiumo Formations (Yu Guangming et al., 1983; Xu Yuling et al.; 1990). New stratigraphic evidence indicates miscorrelation of some of the above formations, requiring revision of the previously established stratigraphic framework for the Jurassic of southern Tibet. The new data demonstrate that different names were used at different localities for the same strata.

The Pupuga Formation was named by Yu Guangming et al. (1983), based only on the Pupuga section. The present study confirmed that it occurs also in the Lalongla section; this is the first time the strata in the Lalongla section have been assigned to the Pupuga Formation. Above the Pupuga Formation in the Lalongla Section is a sequence of micritic and oolitic limestones. In previous work, the name Lalongla Formation was applied to these strata and they were dated as Middle Jurassic, and so regarded as being older than the Nieniexiongla Formation. The present study shows that both the Nieniexiongla and Lalongla Formations are lithologically similar and yield similar fossil assemblages. We suggest that the Lalongla Formation in the Lalongla Section is equivalent to the Nieniexiongla Formation in the Nieniexiongla Section; outcrops at these two localities represent the same formation. We retain the name Nieniexiongla Formation for these strata. A chamositic oolite unit is situated above the Nieniexiongla Formation. It was placed at the base of the Menbu Formation. This unit is certainly of sufficient thickness to be a mappable unit, which is the primary characteristics of a formation. In this case, we name this unit the Chamositic Oolite Formation. The so-called Xiumo Formation was illustrated near Xiumo and Gucuo villages and, as well, above the

Menbu Formation in the Lalongla Section (Yu Guangming et al., 1983). This unit seen near Xiumo and Gucuo Villages is lithologically correlative with the upper part of the Menbu Formation, and both contain Tithonian fossils. The strata in the Lalongla Section are wackestones and oolitic limestones. Yu Guangming et al. (1983) referred them to the Xiumo Formation based on their stratigraphic position. Many Lower Jurassic fossils like *Orbitopsella* and *Lithiotis* were found during the present work. This group of strata, therefore, is of Early Jurassic age, and rests on the Menbu Formation above a faulted contact. The Xiumo Formation actually does not exist in the stratigraphic sequence, and it is not retained as a lithostratigraphic unit.

New biostratigraphic evidence presented below allows revision of the Jurassic stratigraphy of southern Tibet into four formations (Table 1). The revised succession consists of the Pupuga, Nieniexiongla, Chamositic Oolite and Menbu Formations. The Menbu Formation is overlain disconformably by the Lower Cretaceous Gucuo Formation in the region. In the proposed stratigraphic succession we separated the Chamositic Oolite Formation because of its wide use as a prominent marker bed during regional mapping.

Wang 1980		Xu et al., 1990		This work	
J_3	Menkadun Formation	K_1	Gucuo Formation	K_1	Gucuo Formation
		J_3	Xiumo Formation	J_3	Menbu Formation
J_2	Nieniexiongla Formation	J_2	Menbu Formation	J_2	Chamositic Oolite Formation
			Lalongla Formation		Nieniexiongla Formation
			Nieniexiongla Formation		
J_1	Pupuga Formation	J_1	Pupuga Formation	J_1	Pupuga Formation

Table 1. Correlation of Jurassic strata in the Nyalam area, southern Tibet

Pupuga Formation (Lower Jurassic)

The name Pupuga Formation was initially applied to strata that outcrop near the Pupuga Bridge, where the top of the sequence is cut by a fault. A better outcrop has recently been found north of Lalongla Pass, where these strata overlie Upper Triassic sandstone. The Pupuga Formation here is 130m thick (Fig. 2), and consists of wackestone, shale and sandstone, with the sandstone and wackestone beds alternating in the lower half of the formation. Fossils are abundant and dominated by foraminifera: *Orbitopsella dubari*, *O. tibetica*, *Vidalina asymmetrica*, *V. martana*, *V. zujovici*, *Labyrinthina recoarensis*, *Rhapydionina urensis*, *R. ellptica*, *Pseudocyclammina liasica*, *Glomospira articulosa*, *G. gordialis*, *G. pattoni*, *G. perplexa*, *G. jurassica*, *G. regularis*, *G. simplex*, *G. tetragona*, *Glomospirella irregularis*, *G. pavida*, *G. spirillinoidea*, *Ammodiscella virgilensis*, and *Textularia vulgaaris*. Rare corals include *Enallhelia venusta*. Characteristic bivalves include *Lithiotis* sp., *Astarte elegans*, *Neocrasina obliqua*, *Anisocardia minima* and *Weyla ambongoensis*. The boundary between the Pupuga Formation and overlying Nieniexiongla Formation is placed at the boundary between sandstone and the first occurrence of oolitic limestones.

Nieniexiongla Formation (Middle Jurassic)

This formation is composed of oolitic limestone and wackestone intercalated with sandstone and shale. The thickness of the formation is 200-550m. The type section was selected to the south of Nieniexiongla Pass. Strata of similar lithology and containing the same fossil assemblage outcrop at Lalongla Pass, where they were named the Lalongla Formation (Yu Guangming et al., 1983).

Bajocian ammonites such as *Dorsetensia romani* occur in both sections, demonstrating age equivalence. They contain abundant fossils, such as foraminifera, bivalves, brachiopods, corals, belemnites and ammonites. Foraminifera are represented by *Labyrinthina recoorensis, Vidalina asymmetrica, V. zujovici, Glomospira articulosa, G. biforma, G. gordialis, G. jurassica, G. regularis, G. tetragona, Glomospirella reata, G. pavida, Milliammina* sp., *Textularia dolffusi*. The bivalve fauna is dominated by *Trigonia brevicostata, T. moorei, Nuculana thompsoni, Tancredia simili, Astarte subminima,* and *A. extensa*. Ammonites include *Wicthellia tibetica, W. laeviusculus, Dorsetensia romani, D. haydeni,* and *Hebetoxyites* sp. Brachiopoda are dominated by *Russirhynchia* sp., *Septaliphora* sp., and *Burmirhynchia* sp.; terebratellids also occur.

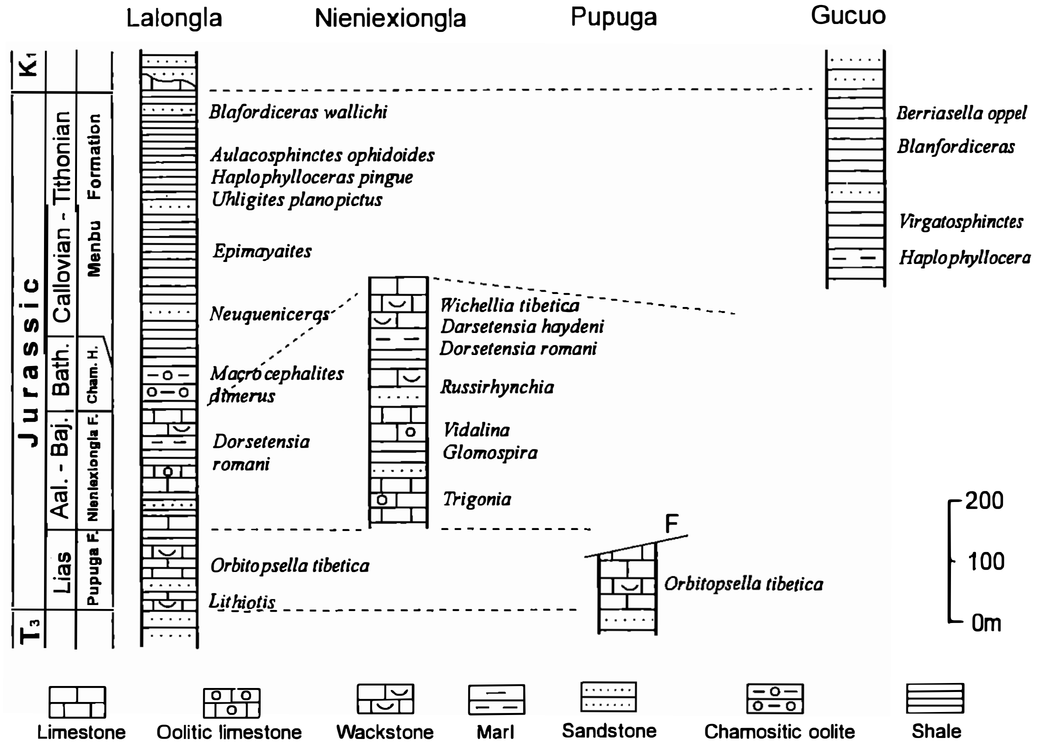

Figure 2. Jurassic stratigraphy in the Nyalam area of southern Tibet.

Chamositic Oolite Formation (Middle Jurassic).

We recognize this bed as a separate stratigraphic unit as it is used as a marker bed during regional mapping. It is 94m thick in the Lalongla Section. This formation separates the Nieniexiongla and overlying Menbu Formation, and was studied in detail by Westermann and Wang (1988). It consists of mainly chamositic oolites, several centimeters to several tens of centimeters thick, underlain by a shelly coquina, an apple green shale layer and gray marl. The Chamositic Oolite Formation is overlain by black micaceous shale with small calcareous concretions, which in turn is overlain by greenish-gray, silty shale. The Chamositic Oolite Formation yields the ammonites *Macrocephalites cosmopolitum, M. compressus, Indocephalites sphaeroidalis, I. diadematus, Parapatoceras calloviese, Helicoceratoides adela, Kheraiceras cosmopolitum, Hecticoceras virgatoum,* the belemnites *Hibolites catlinensis,* and *Belemnopsis lalonglensis,* and the bivalves *Grammatodon virfatus, Protocardia bipi, Posidonia ornati,* and *Astarte kenti,* among others.

Menbu Formation (Upper Middle-Upper Jurassic)

Greenish to dark gray shales intercalated with siltstones overlie the Chamositic Oolite Formation with a sharp basal contact, suggesting a diastem between the Chamositic Oolite Formation and the overlying Menbu Formation. The upper part of the Menbu Formation consists of interbedded, yellowish green siltstone and sandy shale. Calcareous concretions are frequently found within this formation, which at its type section is 416m thick. Fossils are quite abundant and include the ammonites *Blanfordiceras vallichi, Aulacosphinctoides subtorgnatus, A. infundibulus, Aulacosphinctes ophidoides, Haplophylloceras pingue, Uhligites planopictus, Metagravesia decioiens, Macrocephalites cosmopolitum, and Indocephalites sphaeroidalis.* The bivalves *Buchia concentria* and *B. spitiensis* also occur. The top of the Menbu Formation is placed at an erosional contact with overlying quarzitic sandstone of the Lower Cretaceous Gucuo Formation near Gucuo and Xiumo villages. In the Lalongla Pass region, a deeply incised submarine canyon cuts into the top of the Menbu Formation. The sediments of the submarine canyon are overlain by the Lower Jurassic beds yielding *Lithiotis* sp.

The Gucuo Formation yields abundant ammonites and bivalves, the former dominated by *Berriasella grandis, B. privasensis, B. subcallisto, Kilianella costricta, Enthymiceras michaelis,* and *Acanthodiscus subradiatus.*

Conclusions

Biostratigraphic study of the Jurassic strata in southern Tibet has enabled the establishment of several distinct fossil assemblages. The *Orbitopsella-Lithiotis* foraminifera-bivalve assemblage indicates an Early Jurassic age for the Pupuga Formation, a *Witchellia-Dorsetensia* ammonite assemblage provides an Aalenian-Bathonian age for the Nieniexiongla Formation, and a *Macrocephalites-Indocephalites* ammonite assemblage indicates an Early Callovian age for the Chamositic Oolite Formation. Abundant ammonites such as *Blanfordiceras, Aulacosphictes,* and *Haplophylloceras* in the Menbu Formation indicate its age as ranging from Middle Callovian to Tithonian (Fig. 2).

The Early Jurassic *Lithiotis* fauna, found for the first time in southern Tibet during our field work, demonstrates widespread occurrence of this facies over a large area of the Tethys, where it rims the northern margin of Gondwanaland (Bosellini and Loriga, 1971; Loriga and Neri, 1976; Nauss and Smith, 1988).

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References

BOSELLINI, A. AND C B. LORIGA. 1971. I "Calcari grigi" di Rotzo (Giurassico inferiore, Altopiano di Asiago). Annali Universita di Ferrara, Nuova Serie, Sezione 9-Scienze Geologiche e Paleontologiche, 5(1):1-61.

HAQ, B. U., J. HARDENBOL AND P. R. VAIL. 1987. Chronology of fluctuating sea levels since the Triassic. Science, 235:1156-1167.

LORIGA, C. B. AND C. NERI. 1976. Aspetti paleobiologici e paleogeografici della facies a "*Lithiotis*" (Giurese inf.). Rivista Italiana di Paleontologia e Stratigrafia, 82(4):651-706.

NAUSS, A. L. AND P. L. SMITH. 1988. *Lithiotis* (Bivalvia) bioherms in the Lower Jurassic of east-central Oregon, U.S.A. Paleogeography, Paleoclimatology, Paleoecology, 65:253-268.

WAN XIAOQIAO. 1989. Jurassic foraminifera from Nyalam, Tibet. Acta Micropalaeontologica Sinica. 6(2):139-152.

WANG YIGANG. 1980. New data about the stratigraphy in the Himalayan Region (part in China). Stratigraphy Magazine, 4(1):55-59.

____ AND ZHANG MINGLIANG. 1975. Stratigraphy in the Everest Region -Jurassic. *In* Report of the scientific expedition to the Everest Region. Scientific Publishing House, Beijing. p. 124-147.

WESTERMANN, G. E. G. AND WANG YIGANG. 1988. Middle Jurassic ammonites of Tibet and the age of the Lower Spiti shales. Palaeontology, 31:295-339.

XU YULING, WAN XIAOQIAO, GOU ZHONGHAI, AND ZHANG QIHUA. 1990. Biostratigraphy of Tibet in the Jurassic, Cretaceous and Tertiary periods. China University of Geosciences Press. 147 p.

YU GUANGMING, XU YULIN, ZHANG QIHUA, GOU ZHONGHAI, LI XIAOCI AND WAN XIAOQIAO. 1983. Division and correlation of the Jurassic strata in the Nyalam area, Tibet. Contribution to the Geology of the Qinghai-Xizang (Tibet) Plateau, 11: 165-178.

Sequence Stratigraphy

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Sequence Stratigraphic Analysis of the Early and Middle Jurassic Los Patillos Formation, La Ramada Basin (31°30'-32°30' SL), Argentina

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Keywords: Sequence Stratigraphy, La Ramada Rift Basin, Ammonite Zones

Abstract: Stratigraphic studies were carried out in the High Andes of central - western Argentina, San Juan province (31°30' - 32°30' SL). Eight depositional sequences, ranging from Lower to Middle Jurassic, have been recognized in the Ramada Rift Basin for the Los Patillos Formation, allowing a better understanding of depositional mechanisms. Abundant ammonite, bivalve and brachiopod fauna permit precise age control.

Introduction

The Jurassic sequences from the southwestern Main Andes Range of San Juan have attracted the attention of geologists and naturalists since the end of last century. Most have emphasized research in the classic Espinacito Pass Jurassic locality and regional reconnaissance work. First observations on the Jurassic were made by Gay (1838), Philippi (1860), Stelzner (1873), Göttsche (1878), and Bodenbender (in Tornquist, 1898), among others, to whom we owe the first paleontological contributions at these latitudes. Observations on the Mesozoic sequences were later extended by the regional and structural studies of Schiller (1912) and Kühn (1914). Using their biostratigraphic and regional contributions, Groeber (1918) presented the first paleogeographic reconstructions.

The first Lower Jurassic rocks were found in Rancho de Lata creek by Rigal (1930) and recognition of the southward extension of the outcrops, towards the rio de Los Patos valley, was achieved by Lambert (1943; 1944) and Volkheimer et al. (1978a; 1978b). These studies were complemented by Stipanicic (1966), who measured sections at Espinacito Pass, arroyo de las Flores Yeguas Heladas and Los Erizos.

Fossils from the classic localities around Espinacito Pass were studied and illustrated in detailed monographs by Riccardi and co-workers (Westermann and Riccardi 1972; 1979; Riccardi and Westermann, 1991a; 1991b; Riccardi, Damborenea, et al., 1991; Riccardi, Westerman and Elmi, 1991; Riccardi et al., 1984; Damborenea, 1987; and Manceñido, 1990, among others).

Recent work in the area has unravelled the processes acting during basin infill and is partially presented in papers by Ramos et al. (1993; 1996), Alvarez (1996a; 1996b; 1996c), Alvarez et al. (1995) and Ramos and Alvarez (1996) among others. The present paper is intended to provide an overview of the sequence stratigraphy recorded in the Jurassic succession of the La Ramada Basin.

Location

The study area is located in the Andes Main Range at 70° WL between 31°40' and 32°30' SL, in the southwestern corner of San Juan Province, Argentina (Fig. 1). Three belts of Jurassic outcrops are recognized:

1. An eastern belt, which includes the western divide of the La Cerrada Range, the Espinacito Range and its northward extension along the Ramada Range, and the Cordillera de Ansilta. This belt includes the classic Jurassic localities such as Paso del Espinacito (32°09'), Rancho de Lata (32°13'), Arroyo de las Vegas (32°18') and Cordón de la Ramada (32°08'). Outcrops extend from the río Colorado (32°00') to the southern slopes of the quebrada de la Cerrada (32° 30').

2. A central belt occupies the western margin of the Ramada Massif in the southern half and the Cordillera de Santa Cruz towards the north. Outcrops extend from the mouth of the Arroyo de los Bizcochos (31°38') to the Arroyo de las Flores (32°09'). Well known localities such as Paso de los Erizos (31°42'),) and Arroyo de las Flores (32°10') are present in this belt.
3. A western belt lies between the Río de la Pantanosa to the north (31°30') and the Río Alitre beds to the south. These marine Jurassic outcrops have a limited extension, restricted to the headwaters of the Río Pachón (31°44').

La Ramada Basin

La Ramada Basin is a Late Triassic to Early Jurassic rift depocenter isolated from the main Andes basins of Argentina and Chile. An important topographic high, which was present from the Triassic to the mid-Jurassic, separated the La Ramada Basin from the Neuquén Basin to the south. This high, known since the early studies of Groeber (1918), was located between the Las Vacas and Diamante rivers in the northern part of the Mendoza province and separated two different depocenters with similar geologic history (Alvarez, 1996a).

The sedimentary infill of the basin began during the Middle to Late Triassic with deposition of the Rancho de Lata Formation continental, synrift epiclastics and pyroclastics (Alvarez et al., 1995). They are intruded by three north-south aligned rhyolitic domes and both units are cut by porphyritic basaltic dykes. Because neither intrusive is emplaced in the Los Patillos Formation or younger formations, they are interpreted as synrift bimodal volcanics of probable early Early Jurassic age.

Marine limestones and calcarenites of the Los Patillos Formation (Alvarez et al., 1995) unconformably overlie both the Rancho de Lata Formation and the acidic volcanics. They are overlapped by the La Manga Formation, made up of calcareous megabreccias. Deposition of both formations is related to the thermal subsidence stage of the rift system.

Evaporites from the Auquilco Formation overlie the La Manga Formation. A new cycle starts with deposition of the Tordillo Formation, which overlies the latter two formations with a strong erosional unconformity. It is characterised by upward fining red sequences, ranging from coarse conglomerates to medium and fine sandstones, which represent an alluvial fan and braided river environment.

The following units correspond to the Cretaceous Mendoza Group and Huitrín, Diamante, Cristo Redentor and Juncal formations. The Miocene La Ramada Volcanic Complex unconformably overlies all the previous formations.

Los Patillos Formation Depositional Sequences

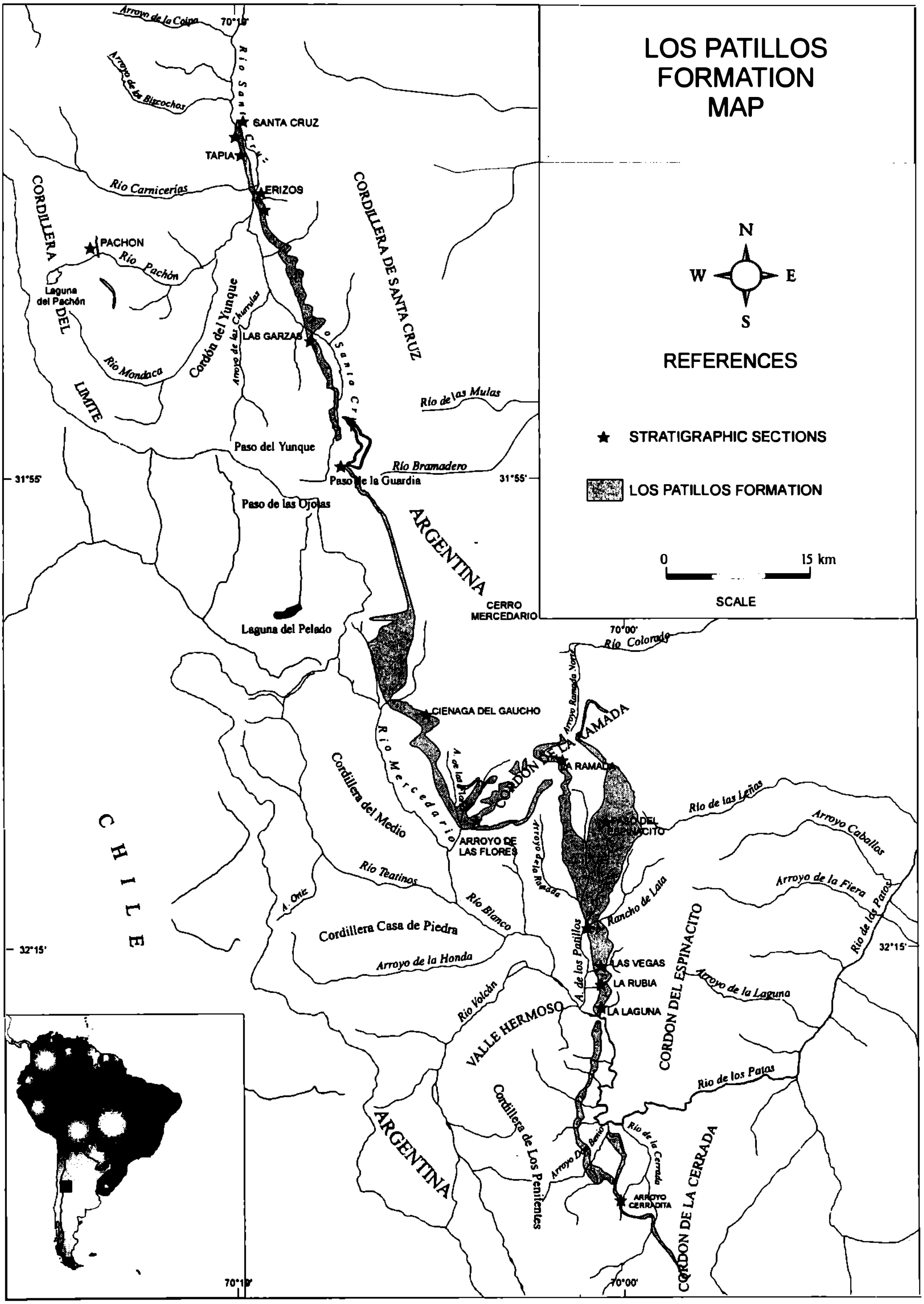
Based on the measured stratigraphic sections, 8 depositional sequences have been recognized within the Los Patillos Formation. They have been informally named as $LP0_1$ (Late Pliensbachian - Early Toarcian), $LP0_2$ (Early Toarcian - Late Toarcian), LP1 (Early Aalenian - Late Aalenian), LP2 (Late Aalenian - Early Bajocian), LP3 (Early Bajocian - Late Bajocian), LP4 (Early Bathonian - Middle Bathonian), LP5 (Late Bathonian - Early Callovian), LP6 (Early Callovian) (Fig. 2). All these sequences have temporal equivalents in the Neuquén Basin where they have been considered as part of the Cuyo subsystem by Riccardi and Gulisano (1992).

The distribution of the different sequences was recognized only through systematic and careful mapping in the eastern belt outcrops, the first of which was done at a 1:10,000 scale in the Espinacito Pass-Ramada Range area. Based on this mapping several unconformities and discontinuities were recognized, which defined the following sequences.

Depositional Sequence LP0

This sequence encloses two depositional subsequences. Its thickness varies from 60 to 110m and has been recognized in La Cerradita, Arroyo de las Vegas and Rancho de Lata (eastern belt), in Ciénaga del Gaucho and Arroyo las Flores (central belt) and in Pachón (western belt). Each subsequence reflects a transgressive marine environment, with a systems tract dominated by storm events and representing a clastic platform facies, and a high stand systems tract genetically related to the first.

Figure 1. Outcrop distribution of Los Patillos Formation.


LOS PATILLOS FORMATION MAP
N
W
E
S
REFERENCES
★ STRATIGRAPHIC SECTIONS
LOS PATILLOS FORMATION
0
15 km
SCALE
Arroyo de la Coipa
70°10'
Arroyo de los Bizcochos
Río Santa Cruz
SANTA CRUZ
TAPIA
ERIZOS
Río Carnicerías
CORDILLERA DEL LIMITE
PACHON
Río Pachón
Laguna del Pachón
Cordón del Yunque
Arroyo de las Chinvilas
Río Mondaca
LAS GARZAS
Río Santa Cruz
CORDILLERA DE SANTA CRUZ
Río de las Mulas
Paso del Yunque
Río Bramadero
Paso de la Guardia
31°55'
31°55'
Paso de las Ojotas
ARGENTINA
CERRO MERCEDARIO
Laguna del Pelado
70°00'
Río Colorado
CIENAGA DEL GAUCHO
Arroyo Ramada Norte
CORDON DE LA RAMADA
RAMADA
Río Mercedario
A. de las Flores
Cordillera del Medio
ARROYO DE LAS FLORES
Arroyo Ramada
Río de las Leñas
ARROYO DEL ESPINACITO
Arroyo Caballos
CHILE
Río Teatinos
Río Blanco
A. Ortiz
Rancho de Lata
Arroyo de la Fiera
Cordillera Casa de Piedra
A. de los Patillos
Río de los Patos
32°15'
32°15'
Arroyo de la Honda
LAS VEGAS
LA RUBIA
Arroyo de la Laguna
CORDON DEL ESPINACITO
Río Volcán
VALLE HERMOSO
LA LAGUNA
Río de los Patos
ARGENTINA
Cordillera de Los Penitentes
Río de la Cerrada
Arroyo Del Bonito
ARROYO CERRADITA
CORDON DE LA CERRADA
70°10'
70°00'

The marine deposits at the base of sequence LP0 onlap the Rancho de Lata Formation as a Type II sequence boundary. This onlap relation is clearly shown on the southern slopes of Rancho de Lata creek. Sequence thickness is variable in each belt; however, they all quickly thin and are replaced by continental beds northward.

Depositional Sequence LP01

The base of this sequence is characterized by a transgressive conglomerate that grades up into fine grained sandstones, with high flow regime structures interpreted as transgressive systems tract tempestites (Fig. 2). Above this systems tract, shales, siltstones and lithoclastic wackestones interbedded with very fine calcareous sandstones represent the late Early Toarcian highest stand successions. Its fauna corresponds to the *Fanninoceras* Association Zone and the base of the *Peronoceras largaense, P. pacificum* and *Collina chilensis* Association Zones (Riccardi et al., 1990). The age of the sequence is thus bracketed between Late Pliensbachian and Early Toarcian.

Depositional Sequence LP02

A discontinuity in the sedimentary record reflects a lowering of sea level. This second depositional sequence is characterized at its base by a new transgressive systems tract, similar to the previous ones. Already close to latest Toarcian time a high stand systems tract starts to develop. It is represented by offshore pelites with hummocky bedded, fine sandstone intercalations, interpreted as distal tempestites. The age of this sequence is based on the occurrence of *Phymatoceras toroense* (Hillebrandt, 1987), *Phlyseogrammoceras tenuicostatum* (Riccardi, 1984), *Pleydellia lotharingica* and *P. fluitans* (Hillebrandt, 1987) Association Zones. Sequence LP02 thus spans the whole Late Toarcian. The systems tracts recognized in these sequences correlate well with the onshore onlap curve of Legarreta et al. (1993) and with the curve of the Andean transgressive events of Hallam (1991).

Depositional Sequence LP1

This sequence has been recognized in outcrops from La Cerradita, Las Vegas and Paso del Espinacito, all eastern belt localities. The sequence thickness ranges from 30 to 50 m, wedging out northward.

A new transgressive cycle occurs in the basin after the Early Aalenian relative sea - level drop. Offshore pelites and siltstones from the previous sequence are overlain by fluvial conglomerate and coarse sandstones along the margin of the basin, followed by shoreface sandstones. The highest flooding successions are generated towards the top of the sequence. They are represented by very fine sandstone, siltstone and bioclastic limestone lithofacies related to decreasing energy of the system in a stormy, offshore environment. The top of the sequence is marked by a discontinuity generated towards the Late Aalenian during a drop in relative sea level. Its age is constrained by the occurrence of ammonites from the Manflasensis Standard Zone (Lower Aalenian).

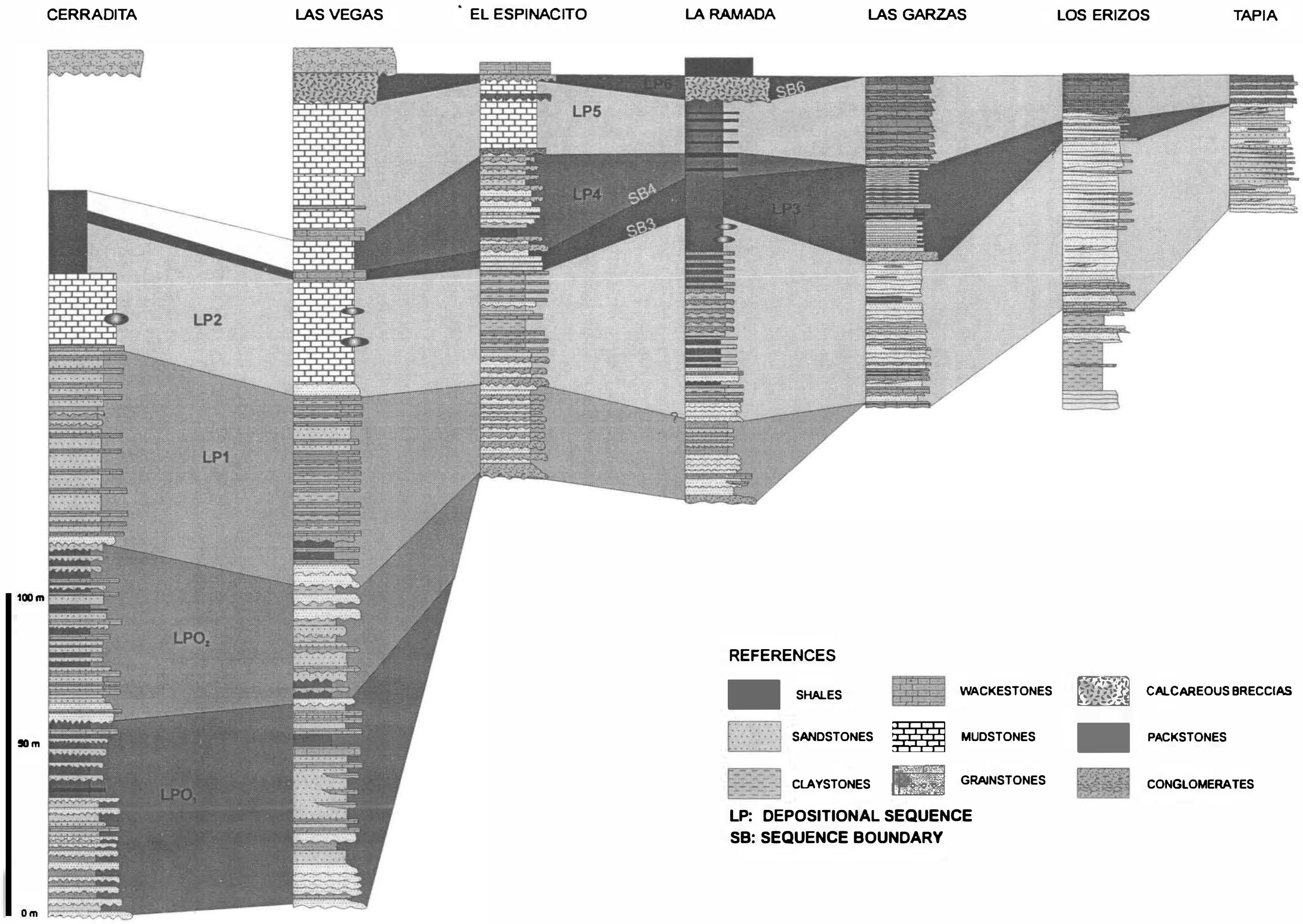
During this period the area covered by marine and transitional marine sediments expands approximately 15 kilometers northwards, covering places such as Espinacito Pass that were previously emergent.

Depositional Sequence LP2

Towards the Late Aalenian - Early Bajocian the Ramada Basin maintained a high subsidence rate, related to its maximum thermal subsidence. The base of sequence LP2 is characterized by a systems tract poorly developed in the eastern belt. Its transgressive character is, however, well established in the central belt, where the Bajocian deposits unconformably overlie the Rancho de Lata Formation volcaniclastics. The age of the sequence spans the Late Aalenian-Early Bajocian, based on ammonites from the Malarguensis, Singularis and Giebeli Zones.

A high stand unconformity marks its top. It is recognized by the onlap between the previous sequences and the offshore pelite and mudstone lithofacies from the easternmost outcrops.

Figure 2. Correlation of stratigraphic sections, showing the development of the different depositional sequences


CERRADITA
LAS VEGAS
EL ESPINACITO
LA RAMADA
LAS GARZAS
LOS ERIZOS
TAPIA
LP6
SB6
LP5
LP4
SB4
SB3
LP3
LP2
LP1
?
LPO_2
LPO_1
100 m
50 m
0 m
REFERENCES
SHALES
WACKESTONES
CALCAREOUS BRECCIAS
SANDSTONES
MUDSTONES
PACKSTONES
CLAYSTONES
GRAINSTONES
CONGLOMERATES
LP: DEPOSITIONAL SEQUENCE
SB: SEQUENCE BOUNDARY

Towards the north of the central belt a prograding shoreface sandstone system replaces the offshore sequences. In the eastern belt, on the other hand, a 25 kilometer northward displacement of the shoreline from its previous location at Espinacito Pass to the headwaters of the Colorado river, has been observed (Fig. 2).

Basin expansion thus may not only be controlled by relative sea level rise but also by regional thermal subsidence.

Depositional Sequence LP3

The base of this sequence is marked by a very important discontinuity in the biostratigraphic record near the late Early Bajocian. This is recognized at the Espinacito Pass and Las Vegas localities by the lack of the *Duashnoceras paucicostatum chilensis* Association Subzone, which defines the top of the Humphriesianum Standard Zone. Towards the south of the eastern belt this zone is completely absent. The basal boundary of this sequence is of type 1 and has been labelled SB3 (Fig. 2).

The base of this lowstand systems tract corresponds to the intra-Bajocian unconformity, bracketed in the Neuquén Basin to the middle part of the Humphriesianum Standard Zone. In the northern La Garzas locality a carbonate breccia marks the beginning of the transgressive systems tract. The top is here defined by ammonites from the late Early Bajocian *Duashnoceras paucicostatum* Association Subzone.

Directly above the previous zone, offshore argillaceous facies reflect the onset of the highstand systems tract. The top of this systems tract corresponds to an erosional unconformity developed on top of the *Megasphaeroceras magnum* bearing beds of the late Early Bajocian Rotundum Zone.

Depositional Sequence LP4

A new depositional sequence, related to lowstand systems tract deposits, is developed on top of the *Megasphaeroceras magnum* beds. The platform is now exposed and its incised channels are filled by very fine-grained, red conglomerates, interpreted as part of a braided fluvial system. This event is interpreted as a forced regression, *sensu* Posamentier et al. (1992), where the coastline migrates inside the basin and the sequence boundary is of Type I (here labelled SB4). The same pattern is described by Legarreta and Uliana (1996) in the Neuquén Basin.

Fluvial sediments are overlain by sandstones, reworked tuffs, siltstones and some air-fall tuffs, interpreted as a shallow marine environment with tidal channels representing the transgressive systems tract of this sequence. Towards the center of the basin these sediments grade laterally into argillaceous and lithoclastic mudstones.

No Lower to Middle Bathonian faunas have been found and therefore the *Cadomites* - Tulitidae mixed Association Zone is missing. This fact is linked to extreme shallowing associated with a relative sea-level drop, recognized in other Argentinian Jurassic basins, and the onset of subduction-linked volcanism west of the basin. Because no ammonites were found in this succession, the age of the sequence is only bracketed by that of the underlying and overlying rocks.

Depositional Sequence LP5

A new transgressive cycle starts with calcareous conglomerates in the Late Bathonian. These are overlain by nodular mudstones that define the highstand systems tract of the sequence. Favourable life conditions were re-established, so that the Late Bathonian - Early Callovian associations are well developed and diverse in all the studied localities, comprising abundant ammonites, bivalves, echinoids and brachiopods. Both the echinoids and bivalves greatly outnumber ammonites in the northern part of the area.

The sequence is dated by ammonites from the Steinmanni, Vergarensis, Rotundum, Bodenbenderi and Proximum Zones of the Early Callovian. An open platform environment that grades into pelites and black, organic-rich mudstones towards the basin interior is represented.

An assemblage of calcareous lithofacies was recognized in the Early Callovian of the central belt. Packstones, oolithic-skeletal grainstones, echinoid and oyster wackestones, rudstones and framestones characterize these facies; they correspond to high-energy, shallow platform deposits.

Depositional Sequence LP6

A new depositional sequence starts at the end of the Late Callovian. The rocks at its base indicate an important sea level drop, dated at 155.5 Ma in the Neuquén-Mendoza Basin (Legarreta et al., 1993), and associated with the Tábanos Formation evaporites.

In the middle part of the La Ramada Basin the sequence is represented by talus deposits related to a lowstand systems tract or to a submarine apron. In the platform facies this sea-level drop is represented by fine conglomerates from the Los Patillos Formation that overlie the Proximum Zone, representing a Type I sequence boundary, here labelled SB6. Its thickness is less than a meter at Espinacito Pass, while in other localities the sequence is absent and is replaced by a discontinuity.

Correlations with the Neuquén Basin

Although several authors have studied the Jurassic sequences within the Neuquén Basin, it was only with Gulisano et al. (1984) that they were treated with a sequential approach. They recognised six depositional sequences, named with the prefix C, within the subsystem Cuyo, which ranges from Early Hettangian to Middle Callovian in age and is partially equivalent to the Los Patillos Formation. Thus, the depositional sequence LP0 may be correlated with their sequence C2, which covers the Late Pliensbachian - Early to Middle Toarcian range and is characterised by clastic platform and turbidite facies. Sequences LP1 and LP2 correspond to their sequence C3, which comprises the Aalenian - Bajocian interval and represents coastal marine, platform and off-shore environments. Sequences LP3 and LP4 correlate with their sequence C4, which developed from the latest Bajocian to the Bathonian. It is made up of red fluvial and deep basinal facies that include abundant pyroclastic material. Sequence LP5 is equivalent to their sequence C5 (Early to Middle Callovian) which represents a carbonate platform environment and sandy turbiditic facies. Sequence LP6 correlates with the calcareous-evaporitic C6 sequence.

Conclusions

Eight depositional sequences have been recognized for the lower to Middle Jurassic Los Patillos Formation in the La Ramada basin. They allow a better understanding of the mechanisms of basin infill. Three major unconformities, the last two showing both eustatic and tectonic control, have been recognized.

The first, dated at 174 Ma (Middle Humphriesianum Standard Zone), corresponds to the SB3 sequence boundary and is equivalent to the intra-Bajocian unconformity in the Neuquén Basin (Zabala, 1996).

The second main discontinuity matches SB4, marking an important facies change and a sea withdrawal of the ocean beyond the platform break. Dated as Early Bathonian (170 Ma), the unconformity relates to the onset of sustained subduction in this Andean segment (Ramos and Alvarez, 1996).

The third major discontinuity corresponds to SB6 and to a forced regressive event controlled by basin encroachment at the end of the Late Callovian, a major paleogeographic change. The basin was isolated from the open sea by a western volcanic arc high.

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References

ALVAREZ, P. P. 1996a. Bioestratigrafía del Jurásico inferior de la cuenca de la Ramada, Alta Cordillera de San Juan (Argentina). Revista Española de Paleontología 11(1):35-47.

____. 1996b. Jurassic ammonite assemblages of the Principal Cordillera of San Juan, Argentina, p. 45-54. *In* A. C. Riccardi (ed.), Advances in Jurassic Research (1-2), GeoResearch Forum, Transtec Publications Ltd, Zurich.

____. 1996c. Los depósitos triásicos y jurásicos de la Alta Cordillera de San Juan, p. 59-138. *In* Ramos et al. Geología de la región del Aconcagua, provincias de San Juan y Mendoza. Dirección Nacional del Servicio Geológico, Anales 24.

____, S. V. BENOIT AND E. G. OTTONE. 1995. Las Formaciones Rancho de Lata, Los Patillos y otras unidades mesozoicas de la Cordillera Principal de San Juan. Revista de la Asociación Geológica Argentina, 49(1-2)(1994):123-142.

DAMBORENEA, S. E. 1987. Early Jurassic bivalvia of Argentina. Part 1 and 2. Palaeontographica, A 199 (1-6):113-231.

GAY, C. 1838. Lettre sur les recherches géologiques dans les Cordillieres d' Elqui, d'Illapel et de Santiago. Comptes Rendus de la Academie des Sciences Paris, 6(26):916-918.

GÖTTSCHE, C. 1878. Über jurassische Versteinerungen aus der Argentinischen Cordillere. Palaeontographica, Suplement 3, 2(3):1-50.

GROEBER, P. 1918. Estratigrafía del Dogger. Boletín de la Dirección de Minas, Geología e Higrogeología 18:1-81.

GULISANO, C. A., A. R. GUTIERREZ PLEIMLING AND R. E. DIGREGORIO. 1984. Esquema estratigráfico de la secuencia jurásica al oeste de la provincia del Nuequén. 9° Congreso Geológico Argentino, Actas 1:236-259.

HALLAM, A. 1991. Relative importance of regional tectonics and eustasy for the Mesozoic of the Andes, p. 189-200. *In* D.I.M. McDonald (ed.), Sedimentation, Tectonics and Eustasy, Special Publication International Association Sedimentology 12.

HILLEBRANDT A. VON. 1987. Liassic ammonite zones of South America and correlations whit other provinces. Bioestratigrafía de los Sistemas Regionales del Jurásico y Cretácico de América del Sur. Mendoza.

KÜHN, F. 1914. Estudio geográfico de las Altas Cordilleras de San Juan. Ministerio de Agricultura, Dirección General de Minas, Geología e Hidrología, Boletín B (Serie Geología) 8:7-44.

LAMBERT, L. R. 1943. Perfil geológico en el valle superior del río de Los Patos Sur (Provincia de San Juan). Museo de La Plata, Revista, Serie Geología 2(11):1-10.

____. 1944. Informe sobre un proyecto de dique de embalse del Río de los Patos sur. Dirección de Mineralogía, Geología e Hidrogeología de la Nación, Boletín 57:1-48.

LEGARRETA, L., C. A. GULISANO AND M. A. ULIANA. 1993. Las secuencias sedimentarias jurásico-cretácicas, p. 87-114. En V.A. Ramos (ed.) Geología y Recursos Naturales de Mendoza. 12° Congreso Geológico Argentino y 2° Congreso de Exploración de Hidrocarburos, Relatorio 1(9).

____, AND M. A. ULIANA. 1996. The Jurassic succession in west-central Argentina: stratal patterns, sequences and paleogeographic evolution. Palaeogeography, Palaeoclimatology, Palaeoecology 120:303-330.

MANCEÑIDO, M. O. 1990. The succession of Early Jurassic Brachiopoda faunas from Argentina: Correlations and affinities, p. 397-404. *In* D. MacKinnon, et al. (eds.), Brachiopods through time.

PHILIPPI, R. A. 1860. Viage al Desierto de Atacama, Halle, Sajonia, 236 p.

POSAMENTIER, H. W., G. P. ALLEN, D. P. JAMES AND M. TESSON. 1992. Forced regressions in a sequence stratigraphic framework: concepts, examples, and exploration significance. AAPG Bulletin, 76 (1):1687 - 1709.

RAMOS, V. A., M. B. AGUIRRE-URRETA, P. P. ALVAREZ, M. I. CEGARRA, E. O. CRISTALLINI, S. M. KAY, G. L. LO FORTE, F. X. PEREYRA AND D. J. PÉREZ. 1996. Geología de la región del Aconcagua, provincias de San Juan y Mendoza. Dirección Nacional del Servicio Geológico, Anales 24:1-510.

____, M. B. AGUIRRE URRETA, AND A. LENCINAS. 1993. El Toarciano fosilífero de Pachón y sus relaciones con el Jurásico marino de la Cordillera Principal de San Juan. XII Congreso Geológico Argentino y II Congreso de Exploración de Hidrocarburos, Actas 2:94-103.

____, AND P. P. ALVAREZ. 1996. Early Mesozoic tectonics of the High Andes of Argentina and Chile, p. 387-396. *In* A. C. Riccardi (ed.), Advances in Jurassic Research, GeoResearch Forum (1-2), Transtec Publications Ltd, Zurich.

RICCARDI, A. C. 1984. Las asociaciones de amonitas del Jurásico y Cretácico de la Argentina. IX° Congreso Geológico Argentino, Actas 4:559-595.

____, S.E. DAMBORENEA AND M. O.MANCEÑIDO. 1990. 3 South America and Antartic Peninsula. 3.1 Lower Jurassic of South America and Antarctic Peninsula. Newsletters on Stratigraphy, 21:75 –103.

____, S. E. DAMBORENEA, M. O. MANCEÑIDO AND S. C. BALLENT. 1991. Hettangian and Sinemurian (Lower Jurassic) biostratigraphy of Argentina. Journal of South American Earth Sciencies 4(3): 159-170.

____, AND C. A. GULISANO. 1992. Unidades limitadas por discontinuidades. Su aplicación al Jurásico andino. Asociación Geológica Argentina, Revista 45(3-4)(1990):346-364.

____, AND G. E. G. WESTERMANN. 1991a. Middle Jurassic Ammonoid Fauna and Biochronology of the Argentine-Chilean Andes. Part 3: Bathonian-Callovian Reineckeiidae. Palaeontographica A 216:111-145.

____ AND ____. 1991b. Middle Jurassic Ammonoid Fauna and Biochronology of the Argentine-Chilean Andes. Part 3: Bajocian-Callovian Eurycephalitinae, Stephanocerataceae. Palaeontographica A 216:1-110.

____, ____, AND S. BALLENT. 1984. Amonitas y estratigrafía del Aaleniano -Bajociano de la Argentina. Con un apéndice micropaleontológico. IX° Congreso Geológico Argentino, Actas 4:559-595.

____, ____, AND S. ELMI. 1991. Biostratigraphy of the Upper Bajocian-Middle Callovian (Middle Jurassic), South America. Journal of South American Earth Sciences 4:149-157.

RIGAL, R. 1930. El Liásico de la Cordillera del Espinacito (Provincia de San Juan). Dirección Nacional de Geología y Minería, Publicación 74:5-9.

SCHILLER, W. 1912. La Alta Cordillera de San Juan y Mendoza y parte de la provincia de San Juan. Ministerio de Agricultura de la Nación, Sección Geología, Mineralogía y Minería, Anales VII(5):1-68.

STELZNER, A. 1873. Mitteilungen an Professor H.B. Geinitz. Uber seine Reise durch die argentinischen Provinzen San Juan und Mendoza und die Cordillere zwischen dem 31° und 33°S. Neues Jahrbuch für Mineralogie, Geologie und Paläontologie, p. 726-744.

STIPANICIC, P. N. 1966. El Jurásico en Vega de la Veranada (Neuquén), el Oxfordense y el diastrofismo divesiano (Agasiz-Yaila) en Argentina. Asociación Geológica Argentina, Revista 20(4)(1965):403-478.

TORNQUIST, A. 1898. Der Dogger am Espinacito-Pass. Paläontologische Abhandlungen, Neue Folge IV(2):3-72.

VOLKHEIMER, W., M. O. MANCEÑIDO AND S. E. DAMBORENEA. 1978a. Zur Biostratigraphie des Lias in der Hochkordillere von San Juan, Argentina. Münstersche Forschungen zur Geologie Paläontologie, 44/45:205-235.

____, ____, AND ____. 1978b. La formación Los Patos (nov.form), Jurásico inferior de la alta cordillera de la provincia de San Juan (República Argentina), en su localidad tipo (Río de Los Patos Sur). Revista de la Asociación Geológica Argentina 32(4):300-311.

WESTERMANN, G. E. G. AND A. C. RICCARDI. 1972. Middle Jurassic ammonoid fauna and biochronology of the Argentine-Chilean Andes. Parte I: Hildocerataceae. Palaeontographica A 140:1-116.

____ AND ____. 1979. Middle Jurassic ammonoid fauna and biochronology of the Argentine-Chilean Andes, Part II: Bajocian Stephanocerataceae. Palaeontographica A 164:85-188.

ZABALA, C. A. 1996. Sequence stratigraphy in continental marine transitions. An example from the Middle Jurassic Cuyo Group, South Neuquén Basin, Argentina, p. 285-294. *In* A. C. Riccardi (ed.), Advances in Jurassic Research, GeoResearch Forum (1-2), Transtec Publications Ltd, Zurich.

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Sequence Stratigraphy of the Callovian-Berriasian (Middle Jurassic-Lower Cretaceous) of the Iberian Basin (NE Spain)

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Keywords: Depositional Sequence, Systems Tracts, Transgressive Events, Discontinuities, Callovian, Oxfordian, Kimmeridgian, Tithonian, Berriasian, Iberian Basin, Spain

Abstract: A sequence stratigraphic study of the Middle Jurassic (Callovian) to lowermost Cretaceous (Berriasian) interval has been carried out in the north-eastern part of the Iberian Cordillera (eastern Spain). Nine depositional sequences have been distinguished. The lower boundary of the lowermost depositional sequence (Callovian Sequence) corresponds to a widespread unconformity involving a regional stratigraphic gap at which the uppermost Bathonian (Discus Zone) and the lowermost Callovian (lower Bullatus Zone) are absent. The upper boundary corresponds to a discontinuity at the end of the Berriasian (lowermost Cretaceous). Sequence boundaries and maximum flooding surfaces for most of the defined sequences are recognised across the basin. It is suggested that the sedimentary evolution and facies distribution in the lower seven sequences, J 2.4 to J 3.6, were partly controlled by external, allocyclic factors (third order sea level cycles). However, significant thickness variation in sequences J2.4, J3.4 and J3.5 indicates local influence of extensional synsedimentary faulting. The upper two sequences (J3.7 and K1.1) were deposited during a relative sea level fall caused by tectonic uplift of the edge of the basin along with a regional sea level fall.

Introduction

The Iberian Cordillera, a mountainous system which extends in a northwest-southeast direction in the northeastern part of the Iberian Peninsula, contains well-preserved outcrops of Jurassic sedimentary rocks. Laterally continuous outcrop allows several cross-sections to be studied across the basin, some hundred kilometres in length, linking the more proximal zones (towards the west) with the more distal ones (towards the east). This lithostratigraphic control, along with the precise biostratigraphic framework based on analysis of the ammonite record, provides the essential constraints to the sequence stratigraphy analysis of the Middle-Upper Jurassic of the Iberian Cordillera (Aurell, 1991).The examined cross-section is located south of Zaragoza, from Veruela to Calanda localities (Fig. 1). The studied rocks have been the site of numerous stratigraphical and sedimentological analyses, that have provided a solid database for our work (Bulard, 1972; Fernández-López, 1985; Meléndez, 1989; Aurell, 1990).

During the Mesozoic, sedimentation in the eastern part of Iberia took place in extensional intracontinental basins, which were episodically flooded by shallow epicontinental seas. Marine sedimentation in these basins generally occurred in shallow and extensive carbonate ramp settings, several hundreds of kilometres across. These ramps were open to the Tethys Sea towards the east. However, during major highstands connections with the boreal realm were made across the Soria Seaway (Fig. 2) to the north-west (Alonso and Mas, 1990; Aurell and Meléndez, 1993). During the Middle-Late Jurassic, a series of palaeogeographic highs detected in the Iberian basin (Bulard, 1972; Aurell, 1990; Aurell and Meléndez, 1993). The Ejulve high, which is located in the east of the study area, had a major effect on sedimentation during Middle- Late Jurassic. This palaeogeographic high, a horst which developed during the Middle Jurassic due to extensive fault activity, was progressively covered by marine units during the successive Late Jurassic floodings.

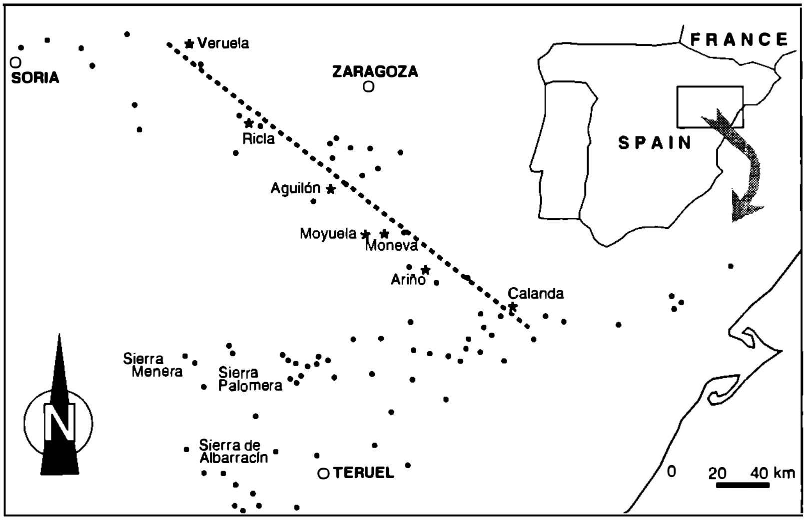


Figure 1. Location of the studied cross-section in the north-eastern part of Spain. Some key sections have been indicated; the black dots correspond to sections measured by the authors in the northern Iberian basin.

Sequence Stratigraphy

Methods

Sedimentary basins are commonly filled by successive unconformity-bounded units. The history of the infill of the sedimentary basin and the presence of breaks in the sedimentary successions, must be understood as the interplay of several factors, such as subsidence, local synsedimentary tectonics, cyclic eustatic fluctuations of different order or magnitude, and the sedimentary input and/or carbonate production variation. The application of sequence stratigraphy concepts has been proved useful in documenting the dynamic evolution of the sedimentary fill of a basin. The basic unit considered in sequence stratigraphy is the depositional sequence. Depositional sequences result from third order cycles of relative sea level variation (from 0.5 to 3 My duration). According to recent refinements (Hunt and Tucker, 1995) of the original postulates of sequence stratigraphy, a sequence boundary should be placed in the inflexion point between sea level fall and sea level rise.

Sequences can be subdivided into systems tracts, according to their relative position on the sea level cycle. In the studied rocks, as is usual in shallow epicontinental platforms, lowstand deposits (LST) are absent or scarcely represented, since there is no space to accommodate them in the interior areas of the platform. Therefore, the sequence boundary is coincident with the transgressive surface. In some cases, facies analysis reveals that the studied sequences consist of deepening-shallowing upward successions, bounded by transgressive surfaces. Accordingly, it has been possible to identify the transgressive (lower, TST) and the highstand (upper, HST) systems tracts. Systems tract assignments are unlikely in some sequences that consist of a single shallowing upward succession or in those that have a very scarce and reworked sedimentary record, related to condensation processes.

Two key surfaces are documented in the carbonate rocks studied: sequence boundaries and maximum flooding surfaces. Identification of the sequence boundary was mainly based on: (1) the presence of significant discontinuities associated with widespread stratigraphic gaps (lacunae in the ammonite record), which have been related to a sea level fall; and, (2) the sharp transition from shallowing upward to deepening upward trends in a succession, which can be traced across the basin. Identification of the maximum flooding surface was based on: (1) the presence of sedimentary condensed events, that may show traces of taphonomic reworking, but usually does not involve a lacuna in the ammonite record. Such boundary events are usually related to starvation processes due to a sea level rise; and, (2) a change in trend from deepening-upward to shallowing-upward in a succession, traceable across the basin.

The Callovian-Berriasian of the northern Iberian basin consists of nine depositional sequences. Facies and thickness distributions of these depositional sequences between the six key localities, are shown in Fig. 3 (see Fig. 1 for location of the studied cross-section). The ages of the sequences and systems tracts, along with other features such as the amplitude of the stratigraphic gaps, can be seen in Fig. 4.

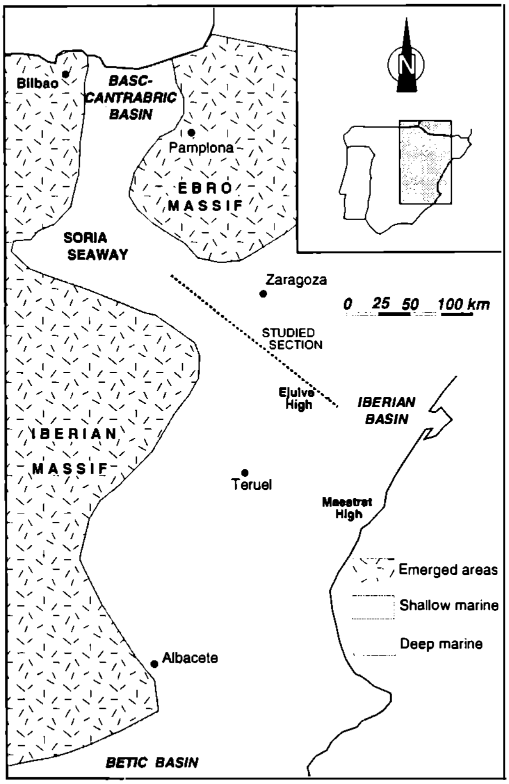


Figure 2. Palaeogeography of the eastern part of Iberia during the early Kimmeridgian. The studied cross-section is located in the northern Iberian basin.

Sequence J2.4 (Callovian Sequence)

According to Fernández-López (1997), the Middle Jurassic of the Iberian basin consists of four deepening-shallowing upward cycles. The upper one is Callovian in age, and in our work is referred to as Sequence J2.4. Its lower boundary corresponds to a widespread unconformity involving a regional stratigraphic gap which affects the uppermost Bathonian (Discus Zone) and the lowermost Callovian (lower Bullatus Zone). The upper boundary appears markedly diachronous, from upper Gracilis to lower Athleta Zones.

This sequence shows marked differences in thickness and facies across the basin. This provides evidence for the strong influence of local tectonics on the general sequence architecture. In the areas influenced by the Ejulve high, between Moyuela and Calanda, the whole sequence is very condensed and incomplete, partly represented by iron-oolite facies. In contrast, a thick succession of skeletal silty limestones and marls occurs in more western localities (Veruela, Ricla, Aguilón). In this succession, it is possible to recognise the Callovian zones, from Bullatus to Athleta Zones (Bulard, 1972; Lardiés, 1990). Facies and paleontological analysis of these expanded sections shows that the maximum flooding surface probably lies at the upper Gracilis Zone, or at the lower-middle Callovian boundary (Gracilis-Anceps Zone). Yet, on the paleogeographic highs at the distal area of the platform, this point is locally marked by a strong regressive event (Fernández-López and Meléndez, 1994, 1996) that could be related to tectonic uplift of the elevated areas due to extensional synsedimentary fault activity.

Sequence J3.1 (Lower Oxfordian Sequence)

The lower boundary of the sequence has an associated stratigraphic gap which comprises at least the uppermost Callovian zone and the lowermost Oxfordian ammonite zone. The maximum hiatus occurred in areas influenced by the Ejulve high, where middle Oxfordian bioclastic limestones (J3.3 Sequence) may rest directly on Bathonian ooid grainstones, which are covered by a subaerial, karsted surface or a planar erosive surface.

In the areas surrounding the Ejulve High (Ariño, Moneva), the J3.1 Sequence is represented by an irregular iron-oolite horizon, up to 0.2 m thick. This horizon is interpreted to have formed under shallow marine environmental conditions, reflecting the flooding of the platform that took place during the Early Oxfordian, after the Late Callovian sea level fall. Iron-ooids and pisoids included in this level are thought to have been formed by weathering processes on the emerged area of the sedimentary high. The resedimentation of the iron-ooids from these areas to the shallow marine environments, along with the arrival of the ammonites as drifted shells, took place during the Early Oxfordian transgressive event (Aurell and Meléndez, 1993; Aurell et al., 1994a). Eventually, the iron ooids and pisoids were mixed and reworked with benthic and open marine fossils on the shallow platform, under extremely low rates of carbonate sedimentation. In these eastern localities, the iron-oolitic horizon contains ammonites: *P.(Parawdekindia) sp., Prososphinctes claromontanus* Bukowski and *Neocampylites delmontanus* Oppel, partly characterising the lower Oxfordian (lower Cordatum Zone or Claromontanus Subzone: see Meléndez et al., 1985).

The iron-ooids are absent in north-western, more proximal areas (Ricla, Veruela), that were located away from the influence of the elevated areas of the Ejulve high. In these localities, the sequence consists of an irregular horizon, up to 0.3 m thick, that includes ammonites of the lower Cordatum Zone along with reworked ammonites (fragmented phosphatic and/or carbonate ammonite internal moulds). Over extensive areas, though, this sedimentary episode has been reworked and the ammonite fossils are incorporated into Sequence J3.2 as reworked fossils (Bulard, 1972; Meléndez et al., 1983; Sequeiros and Cariou, 1984; Ramajo and Aurell, 1997).

Sequence J3.2 (Lower Middle Oxfordian Sequence)

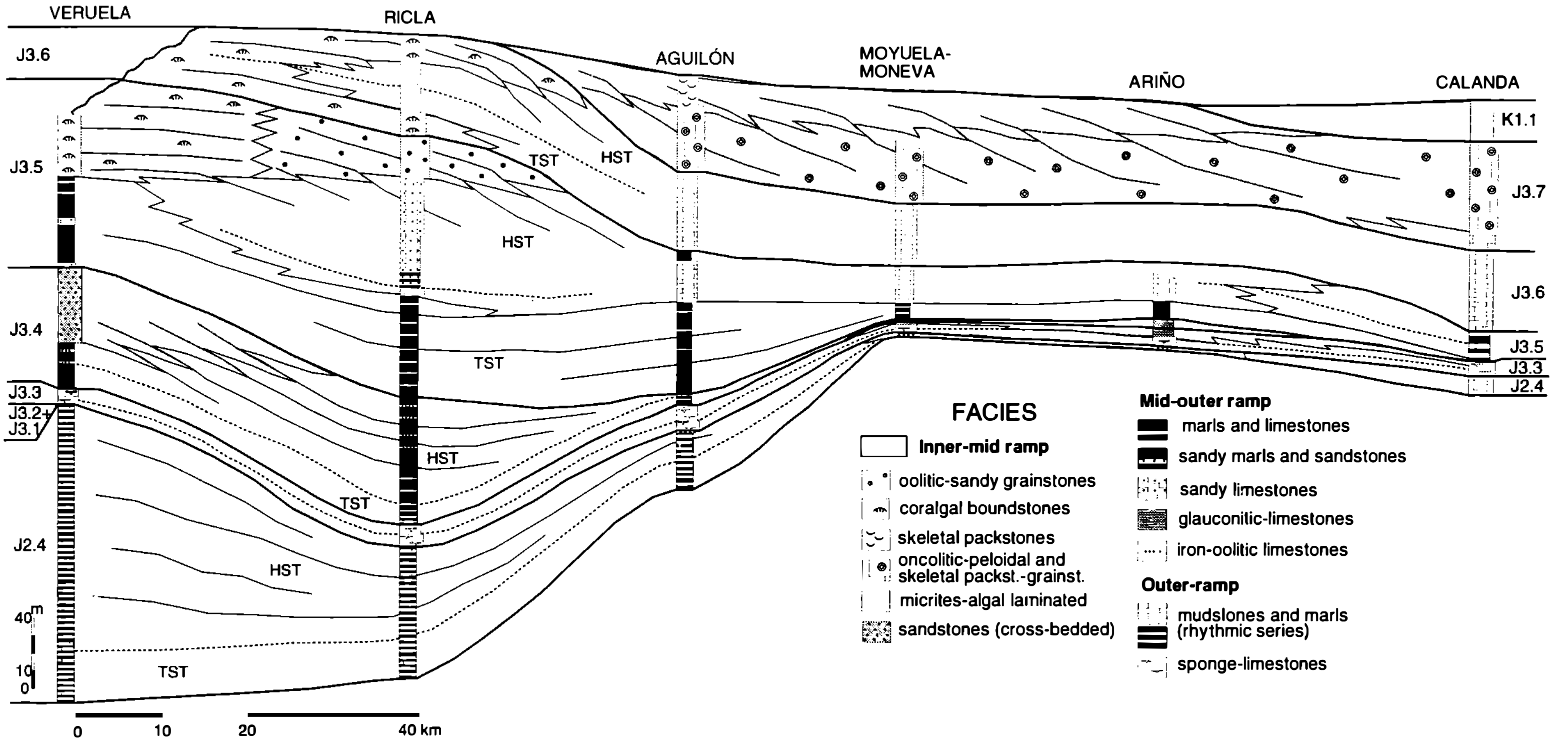
The sequence had a short duration, and it was developed during a transgressive event that took place at the lower-middle Oxfordian transition. The existence of the sequence is indicated by the presence of two widespread stratigraphic gaps, associated with both the lower and the upper boundaries, indicative of sea level falls. The lower one comprises the upper Cordatum Zone, whereas the gap associated with the upper unconformity involves the upper Plicatilis Zone (Ricla) and even the lower Transversarium Zone (Ariño-Calanda).

The sequence is represented by a second irregular, reworked and condensed horizon, also containing iron ooids in areas located near the Ejulve high (Ariño, Moneva), and including reworked fossils from middle to upper Callovian and lower Oxfordian. Non-reworked ammonites found in this bed include some typical forms indicating Middle Oxfordian, lower Plicatilis Zone, Paturattensis Subzone. This sequence is best developed in Ariño, where the characteristic oolitic bed can reach 0.2 to 0.3 m in thickness, and a rich assemblage of lower Plicatilis Zone ammonites has been described (Meléndez et al., 1983; Meléndez, 1989; Meléndez et al., 1997).

Sequence J3.3 (Middle-Upper Oxfordian Sequence)

The lower boundary of the sequence corresponds to the boundary between the upper oolitic bed and an homogeneous lithological unit of sponge limestones. The onset of the sequence involves an important transgressive event in the Iberian basin, which resulted in the establishment of sponge and other benthic and planktic communities across the basin.

Figure 3. Correlation of the key-sections, showing the thickness and main facies distribution in the differentiated sequences and systems tracts. See Fig. 1 for locations.


VERUELA
RICLA
AGUILÓN
MOYUELA-MONEVA
ARIÑO
CALANDA
K1.1
J3.7
J3.6
J3.5
J3.4
J3.3
J3.2+J3.1
J2.4
TST
HST
40 m
10
0
0
10
20
40 km
FACIES
Inner-mid ramp
oolitic-sandy grainstones
coralgal boundstones
skeletal packstones
oncolitic-peloidal and skeletal packst.-grainst.
micrites-algal laminated
sandstones (cross-bedded)
Mid-outer ramp
marls and limestones
sandy marls and sandstones
sandy limestones
glauconitic-limestones
iron-oolitic limestones
Outer-ramp
mudstones and marls (rhythmic series)
sponge-limestones

The lower part of the sequence (Transversarium Zone) or TST appears well represented across the basin, the ammonite record allowing a detailed sequencing of the subzones and horizons defined so far (Meléndez, 1989; Meléndez and Fontana, 1993). Yet the lowermost Parandieri Subzone has only been proven in the north-western part of the platform (Ricla). The open marine, moderately deep, subtidal conditions appear to be present from the very beginning of the unit. This might suggest that the transgressive phase was initiated before deposition of the sponge unit, probably during Antecedens and Parandieri Subzones, leaving no, or a very scarce, sedimentary record. Existence of moderately deep subtidal conditions in the Transversarium Zone are underlined by thinning, deepening upwards sequences, by the development of sponge bioherms, and by the unique reported evidence during this precise stratigraphic interval of ammonite populations colonising the platform (Aurell and Meléndez, 1993; Meléndez et al., 1990, 1997). A hardground surface, resting on a highly condensed interval affecting the top of the Transversarium Zone (Schilli and Rotoides Subzones) and found in distal areas of the platform, is thought to represent the maximum flooding surface.

In western localities (Ricla-Aguilón), the upper part of the sequence (Bifurcatus to Hypselum Zones) or HST, consists of a regular alternation of sponge limestone and marls, showing a typical thickening, shallowing upward trend. In the distal areas (Ariño, Calanda) the HST is represented by strongly condensed sequences (Bello, 1995; Bello et al., 1996; Aurell et al., 1997; Meléndez et al., 1997). The upper sequence boundary is located at the Hypselum-Bimammatum Zone boundary. This upper boundary is represented by an irregular surface involving a sharp facies change from micritic, sponge limestones to more siliciclastic and argillaceous siltstone.

Sequence J3.4 (Upper Oxfordian Sequence)

Above the Hypselum-Bimammatum Zone boundary a sharp facies change demostrates a net increase of the clastic supply from source areas into the basin. This new sedimentary episode extends throughout the Bimammatum and lower Planula Zone (Planula Subzone). It shows a variable development in both facies and thickness, probably reflecting existence of relief on the sea floor and the beginning of a new differentiation into palaeogeographic highs and lows in the sedimentary basin at the onset of the Bimammatum Zone (Pérez-Urresti, 1996; Meléndez et al., 1995).

Maximum thickness of the sequence is reached in the proximal areas (Veruela, Ricla) where the TST consists of an alternation of marls and skeletal wackestones, rich in ammonite fossils, whereas the HST is formed by a coarsening and shallowing upward marly and siliciclastic succession. The age of the cross-bedded sandstones found on top of the HST in the Veruela section is well-constrained by ammonite fossils, having developed in the upper part of the Planula Subzone (Bádenas et al., 1998). In the distal areas (Moyuela, Ariño) the entire sequence is represented by strongly condensed and reworked glauconitic limestones bearing open marine fossils (Pérez-Urresti, 1996; Pérez-Urresti et al., 1996). The upper sequence boundary is located at the Planula-Galar Subzone boundary. This upper boundary is represented by an irregular, slightly diachronous, surface which in some areas may involve a stratigraphic gap ranging from upper Bimammatum to lower Planula Zones.

Sequence J3.5 (Kimmeridgian Sequence)

The sequence spans from the onset of the Kimmeridgian (i.e. Planula Zone, Galar Subzone) to the early Late Kimmeridgian. Ammonites of the Acanthicum Zone are found below and above the upper bounding unconformity in distal basin areas (Calanda), and therefore the upper boundary of the sequence must be placed in the upper Acanthicum Zone.

The onset of the sequence corresponds to a flooding event, indicated by the sharp transition from shallow siliciclastics to open marine marls in proximal areas of the basin (Veruela, Ricla). Shallow and marginal areas of the basin were dominated by reefal and oolitic facies. Most of the mud accumulated in the central part of the basin (i.e., outer ramp rhythmic alternation of mudstones and marls), and must have been derived from shallow ramp productivity areas (Aurell, 1990; Aurell et al., 1995, 1998). In the more distal areas of the basin, the lower Kimmeridgian consists of condensed facies, with abundant ammonites (Atrops and Meléndez, 1985). The maximum flooding surface has been placed in a prominent hard ground found on top of the Divisum Zone. A marly succession some metres thick (i.e., Acanthicum marls), located above this surface is thought to represent the lower HST. The differentiation between the TST and the HST was also based on the change in trend from deepening upward to shallowing upward sequences observed around the lower-upper Kimmeridgian transition (Aurell et al., 1998; Bádenas et al., 1998).

Sequence J3.6 (Upper Kimmeridgian Sequence)

The lower boundary of the sequence is indicated by a flooding event, recognised in the marginal areas of the basin, where coral-algal reef facies (developed in mid ramp environments) sharply overlie inner ramp oolitic and siliciclastic facies of the previous sequence (Bádenas et al., 1993; Aurell and Bádenas, 1997; Bádenas and Aurell, 1997). Facies distribution within this sequence allows distinction of a basal, deepening upward TST, and an upper, shallowing upward or progradational HST. The boundary between these systems tracts, or maximum flooding surface, has been tentatively placed in the last Kimmeridgian Zone (i.e., Beckeri Zone). However, no conclusive paleontological data have been found to locate this surface precisely. In the upper part of the sequence, ammonites of the early Tithonian (Hybonotum Zone) have been recorded in more distal localities (Calanda: Atrops and Meléndez, 1985). As in the previous sequence, the depositional setting consists of an extensive and homogeneous carbonate ramp, with the shallow areas dominated by coral reefs, the distal areas dominated by the deposition of carbonate mud.

Sequence J3.7 (Tithonian Sequence)

The lower boundary of the sequence is indicated by a significant unconformity, produced by a relative sea level fall that involved a considerable basinwards coastal shift at the onset of the sequence. This coastal shift is considered to have been caused by the tectonic uplift of the edges of the basin along with a regional sea level fall. As a consequence, the shallow skeletal, oncolitic and reefal facies developed in the lower part of the sequence sharply overlie the outer ramp micritic facies of the previous sequence in the marginal areas of the basin (Aurell and Meléndez, 1993; Aurell et al., 1994b). Basinwards, there is a gradual transition (i.e., correlative conformity) between the micritic facies of the previous sequence and grain-supported facies.

The sequence was probably developed during the Tithonian, above the Hybonotum Zone. Unlike the preceding sequences, the study area does not reach the most open marine facies belt, which might enable us to be precise about the age of the sequence. Moreover, the sequence shows a single shallowing upward facies arrangement, with no evidence supporting the presence of a lower TST. Therefore, only the HST seems to be represented in the proximal and shallow part of the basin studied in our work (Aurell and Meléndez, 1993).

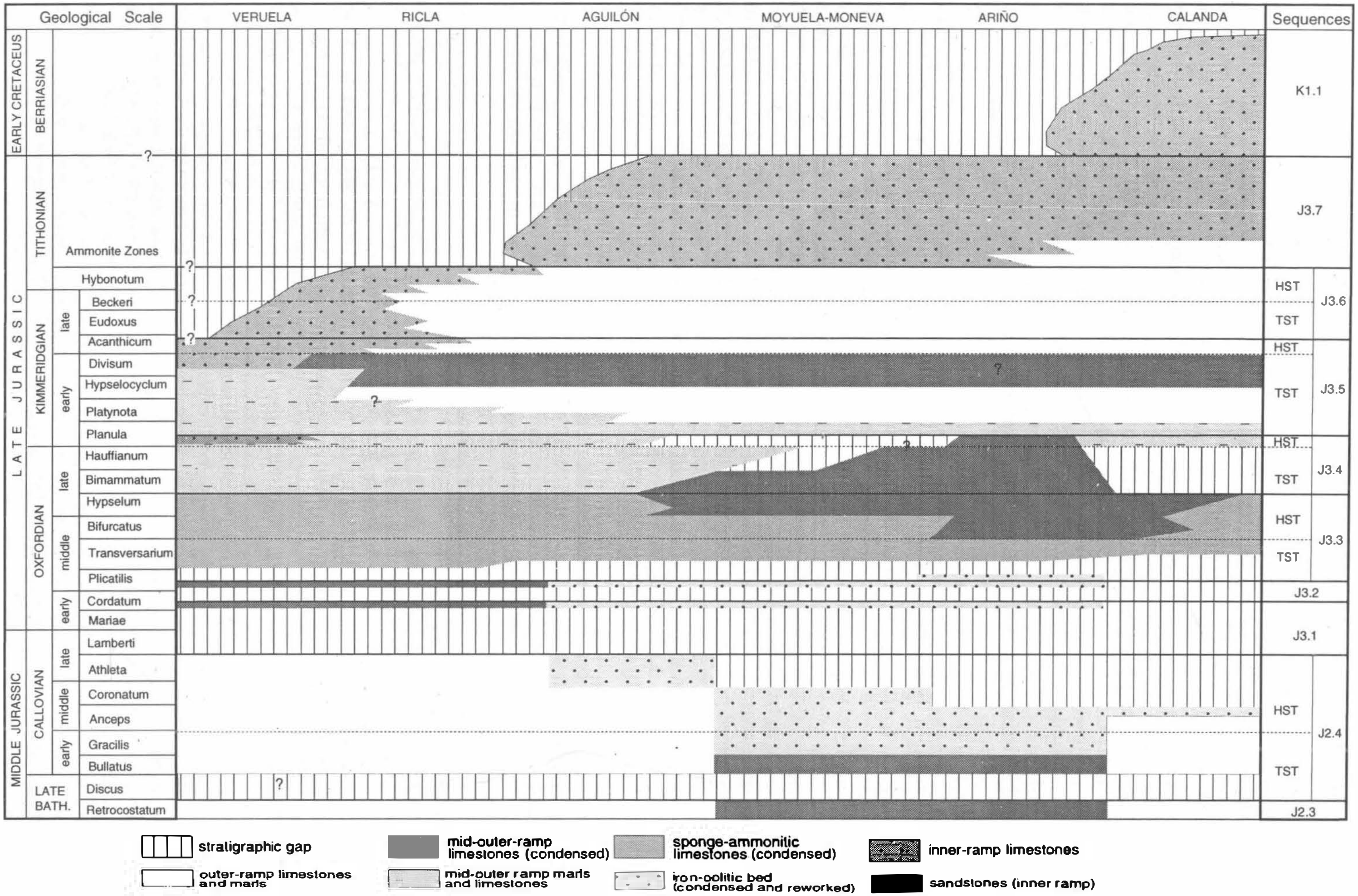
Sequence K1.1 (Lowermost Cretaceous Sequence)

The lower sequence boundary was related to tectonic activity, which involved further basinward coastal shift, and the creation of deeper subbasins in the southeastern part of the Iberian Basin, which were the site of siliciclastic sedimentation (Aurell et al., 1994b). According to the scarce paleontological data, the sequence probably spans from uppermost Tithonian to lower-middle Berriasian. Carbonate facies, deposited on tidal flats and in restricted coastal environments, are found in more stable areas located to the east.

The upper boundary of the studied interval corresponds to a general discontinuity recognised at the end of the Berriasian (lowermost Cretaceous). This major unconformity in the Iberian Basin involves the withdrawal of the sea basinwards (to the east of the studied area) and the generation of a widespread unconformity of variable duration. In proximal areas of the basin, this upper boundary is usually an angular unconformity, with a stratigraphic gap which involves at least all the Valanginian and is also marked by the overlapping of continental units. According to Salas and Casas (1993), this unconformity represents the upper boundary of the so-called Jurassic Supersequence.

Conclusions

In this work we have presented a synthesis of the actual knowledge of the sequence stratigraphy of the Upper Jurassic of the northern Iberian Basin. In previous works (Aurell, 1991; Aurell and Meléndez, 1993), we considered three Upper Jurassic depositional sequences (i.e., Oxfordian, Kimmeridgian and Tithonian-Berriasian sequences). These may actually correspond to second order cycles. Further analysis and new data allow differentiation of four higher order cycles (i.e, third order or depositional sequences) in the Oxfordian Sequence (Sequences J3.1 to J3.4). Both the Kimmeridgian and the Tithonian-Berriasian sequences have been divided into two new depositional sequences (Sequences J3.5-J3.6 and Sequences J3.7-K1.1, respectively).


Geological Scale
VERUELA
RICLA
AGUILÓN
MOYUELA-MONEVA
ARIÑO
CALANDA
Sequences
EARLY CRETACEUS
BERRIASIAN
LATE JURASSIC
TITHONIAN
KIMMERIDGIAN
OXFORDIAN
MIDDLE JURASSIC
CALLOVIAN
LATE BATH.
late
early
middle
Ammonite Zones
Hybonotum
Beckeri
Eudoxus
Acanthicum
Divisum
Hypselocyclum
Platynota
Planula
Hauffianum
Bimammatum
Hypselum
Bifurcatus
Transversarium
Plicatilis
Cordatum
Mariae
Lamberti
Athleta
Coronatum
Anceps
Gracilis
Bullatus
Discus
Retrocostatum
?
K1.1
J3.7
HST
TST
J3.6
J3.5
J3.4
J3.3
J3.2
J3.1
J2.4
J2.3
straligraphic gap
outer-ramp limestones and marls
mid-outer-ramp limestones (condensed)
mid-outer ramp marls and limestones
sponge-ammonitic limestones (condensed)
iron-oolitic bed (condensed and reworked)
inner-ramp limestones
sandstones (inner ramp)

As a whole, the relative sea level variation in the Iberian Basin shows that the Late Callovian was a time of sea level fall. The maximum sea level fall at the end of the Callovian was followed by a continuous sea level rise during the Oxfordian and Kimmeridgian. The Tithonian and Berriasian sequences were deposited during an episode of relative sea level fall, caused by the tectonic uplift of the edge of the basin, along with a regional sea level fall. Significant thickness variations in sequences J2.4, J3.4 and J3.5 also indicate local influence of extensional synsedimentary faulting.

Sequence boundaries and maximum flooding surfaces for most of the defined sequences in the Callovian-Kimmeridgian interval are recognised across the basin. They reveal the occurrence of regional third order cycles of sea level change. Major transgressive events in the Iberian Basin correspond to the lower part of the middle Oxfordian (TST, J3.3 Sequence) and to the early Kimmeridgian (TST, J3.5 Sequence). Transgressive events of minor intensity took place during the Early Callovian, at the lower Cordatum and Plicatilis Zones interval, during middle Late Oxfordian and during middle Late Kimmeridgian. The causal mechanism of these transgressive events remains enigmatic, but it is suggested that they could have resulted from widespread, probably eustatic, events. The geographical extent of the cycles of sea level variation defined in the Iberian Basin may by tested by comparing the age of the sequences and systems tracts with those defined in other sedimentary basins.

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References

ALONSO, A. AND J. R. MAS. 1990. El Jurásico superior en el sector Demanda-Cameros (La Rioja-Soria). Cuadernos Geología Ibérica, 14:173-198.

ATROPS, F. AND G. MELÉNDEZ. 1985. Kimmeridgian and lower Tithonian from the Calanda-Berge area (Iberian Chain, Spain): Some bioestratigraphic remarks. Proceedings of the First International Symposium on Jurassic Stratigraphy, Erlangen. p. 377-392.

AURELL, M. 1990. El Jurásico Superior de la Cordillera Ibérica Central (provincias de Zaragoza y Teruel). Análisis de Cuenca. Tesis Doctoral, Dpto. Geologia, Universidad de Zaragoza, 384 p.

____. 1991. Identification of systems tracts in low angle carbonate ramps: examples from the Upper Jurassic of the Iberian Chain (Spain). Sediment. Geology, 73:101-115.

____, AND B. BÁDENAS. 1997. The pinnacle reefs of Jabaloyas (Late Kimmeridgian, NE Spain): vertical zonation and associated facies related to sea level changes. Cuadernos de Geología Ibérica, 22:37-64.

____, AND A. MELÉNDEZ. 1993. Sedimentary evolution and sequence stratigraphy of the Upper Jurassic in the central Iberian Chain, northeast Spain, p. 343-368. *In* H. Possamentier, C. Summerhayes, B.U. Haq and G.P. Allen (eds.), Sequence Stratigraphy and facies associations. International Association of Sedimentologists, Special Publication, 18.

____, S. FERNÁNDEZ-LÓPEZ, AND G. MELÉNDEZ. 1994a. The Middle-Upper Jurassic oolitic ironstone level in the Iberian Range (Spain): Eustatic implications. Geobios, M.S. 17:547-561.

____, R. MAS, A. MELÉNDEZ, AND R. SALAS. 1994b. El transito Jurásico-Cretácico en la Cordillera Ibérica: relación tectónica-sedimentación y evolución paleogeografica. Cuadernos Geología Ibérica, 18:369-396.

____, D. BOSENCE, AND D. WALTHAM. 1995. Carbonate ramp depositional systems from a late Jurassic epeiric platform (Iberian basin, Spain): a combined computer modelling and outcrop analysis. Sedimentology, 42:75-94.

____, B. BÁDENAS, J. BELLO, G. DELVENE, G. MELÉNDEZ, I. PÉREZ-URRESTI, AND J. RAMAJO. 1997. El Jurásico Superior de Calanda-Mas de las Matas. Libro Guía Excursiones, IV Congreso Jurásico de España. Publicaciones del Seminario de Paleontología, Universidad de Zaragoza, 2:119-144.

Figure 4. The age of the sequences and systems tracts of the late Bathonian-Berriasian in the northern Iberian Basin. See Fig. 1 for locations.

____, B. BÁDENAS, D. BOSENCE, AND D. WALTHAM. 1998. Carbonate production and offshore transport in a Late Jurassic carbonate ramp (Kimmeridgian, Iberian basin, Spain), p. 137-161. *In* V.P. Wright and T.P. Burchette (eds.), Carbonate ramps. Geological Society of London, Special Publication, 149.

BÁDENAS, B. AND M. AURELL. 1997. El Kimmeridgiense de los Montes Universales (Teruel): distribución de facies y variaciones del nivel del mar. Cuadernos Geología Ibérica, 22:15-36

____, M. AURELL, AND A. MELÉNDEZ. 1993. Características sedimentológicas, zonación y evolución de una rampa carbonatada dominada por tormentas (Kimmeridgiense, Cordillera lberica Septentrional). Revista de la Sociedad Geología de España, 6:57-75.

____, M. AURELL, I. PÉREZ-URRESTI, AND G. DELVENE. 1998. Estratigrafía y evolución sedimentaria del Oxfordiense superior-Titónico inferior en Ricla. Geogaceta, 24:35-38.

BELLO, J. 1995. La Biozona Bifurcatus en el sector nororiental de la Cordillera lberica: asociaciones de ammonites y bioestratigrafia. Comunicaciones, X Jornadas Paleontologia, Madrid: 29-32.

____, G. MELÉNDEZ, AND I. PÉREZ-URRESTI. 1996. El límite de las biozonas Transversarium y Bifurcatus (Oxfordiense medio) en la Cordillera Ibérica nororiental: discontinuidades y lagunas estratigráficas asociadas. Geogaceta, 18:14-17.

BULARD, P. F. 1972. Le Jurassique moyen et supérieur de la Chaine Ibérique sur la bordure du bassin de l'Ebre (Espagne). These de Doctorat, Université de Nice, 702 p.

FERNÁNDEZ-LÓPEZ, S. 1985. El Bajociense en la Cordillera Ibérica. Tesis Doctoral, Universidad Complutense de Madrid, 848 p.

____. 1997. Ammonites, clinos tafonómicos y ambientes sedimentarios. Revista Española de Paleontología, 12:102-128.

____, AND G. MELÉNDEZ. 1994. Dispersión biogeográfica y tafonómica de ammonoideos filoceratinos hacia la Cuenca Iberica durante el Jurásico Medio. Coloquios de Paleontologia, Madrid, 46:129-149.

____, AND G. MELÉNDEZ. 1996. Phyllocratina ammonoids in the Iberian basin during the Middle Jurassic: a model of biogeographical and taphonomic dispersal related to sea-level changes. Palaeogeography, Palaeoclimatology, Palaeoecology, 120:291-302.

HUNT, D. AND M. E. TUCKER. 1995. Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall-replay. Sedimentary Geology, 95:307-345.

LARDIÉS, M. D. 1990. Observaciones biostratigráficas y sedimentológicas sobre el Calloviense en la Provincia de Zaragoza. Cuadernos Geología Ibérica, 14:157-172.

MELÉNDEZ, G. 1989. El Oxfordiense en el Sector Central de la Cordillera Ibérica (Provincias de Zaragoza y Teruel). Tesis Doct. Institucion Fernando el Católico, Instituto de Estudios Turolenses, Zaragoza-Teruel, 418 p.

____, AND B. FONTANA. 1993. Biostratigraphic correlation of the Middle Oxfordian sediments in the Iberian Chain, Eastern Spain. Acta Geologica Polonica, 43:193-211.

____, L. SEQUEIROS, AND W. BROCHWICZ-LEWINSKI. 1983. Lower Oxfordian in the Iberian Chain, Spain; Part I. Biostratigraphy and nature of gaps. Bulletin Academia Polonaise Sciences. Serie des Sciences de la Terre, XXX (3-4):158-172.

____, ____, AND ____. 1985. Tentative biostratigraphic subdivision for the Oxfordian of the submediterranean province on the base of Perisphinctids. First Symposium on Jurassic Stratigraphy, Erlangen, Symposium volume II:481-502.

____, M. AURELL, B. FONTANA, AND M. D. LARDIÉS. 1990. El transito Dogger-Malm en el sector nororiental de la Cordillera lbérica: análisis tafonómico y reconstrucción paleogeográfica. Comunicaciones de la reunión de tafonomía y fosilización, Madrid. p. 221-229.

____, I. PÉREZ-URRESTI, AND J. BELLO. 1995. Las subdivisiones bioestratigraficas de la Biozona Bimammatum (Oxfordiense superior) en la Cordillera Iberica nororiental: facies y amplitud de las discontinuidades. Geogaceta, 18:113-1 16.

____, J. BELLO, G. DELVENE, AND I. PÉREZ-URRESTI. 1997. El Jurásico Medio y Superior (Calloviense Kimmeridgiense) en el Sector de la Llanura de Arcos (Ariño-Oliete, Teruel): Análisis tafonómico y bioestratigrafia. Cuadernos Geología Ibérica, 23:269-300.

PÉREZ-URRESTI, I. 1996. Las sucesiones de ammonoideos del Oxfordiense superior en la Cordillera Iberica nororiental: nuevos datos bioestratigraficos. Coloquios de Paleontología, 48: 125-145.

____, J. BELLO, AND G. MELÉNDEZ. 1996. La subfamilia Ataxioceratinae en el Oxfordiense superior de la provincia submediterranea: origen y evolución. Geogaceta, 20 (1):246-249.

RAMAJO, J. AND M. AURELL. 1997. Analisis de las superficies de discontinuidad y depósitos asociados del Calloviense Superior-Oxfordiense Medio en la Cordillera Iberica Noroccidental. Cuadernos Geología Ibérica, 22:213-236.

SALAS, R. AND A. CASAS. 1993. Mesozoic extensional tectonics, stratigraphy and crustal evolution during the Alpine cycle of the eastern Iberian basin. Tectonophysics, 228:33-55.

SEQUEIROS, L. AND E. CARIOU. 1984. Síntesis bioestrátigrafica del Calloviense de Ricla (Zaragoza, Cordillera Ibérica). Estudios Geológicos, 40:411-419.

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Ammonite Taphocycles in Carbonate Epicontinental Platforms

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Keywords: Applied Taphonomy, Sequence Stratigraphy, Sea-Level Changes, Environmental Cycles, Iberian Basin

Abstract: Variations of preservational features of the successive recorded associations of ammonites in carbonate epicontinental platforms enable distinction of taphonomic cycles induced by relative sea-level changes. A taphocycle comprises two or more successive recorded-associations showing cyclical variations in their taphonomic characters, resulting from an environmental cycle. Shallowing-upwards sequences in carbonate outer platforms, and positive taphosequences, were formed during phases of increasing water turbulence and decreasing rate of sedimentation. Positive and negative taphosequences, (or taphosequences of increasing and decreasing turbulence), enable identification of shallowing-upwards sequences and infilling sequences of fifth-order respectively. Condensed sections show different characters in shallow and proximal environments in relation to deep and distal environments of the carbonate epicontinental platforms. The degree of taphonomic condensation in preserved ammonite associations reaches the highest values in shallow and proximal environments of the platform, not in deep and distal environments, though the degree of taphonomic heritage (i.e., the ratio of reelaborated (reworked) elements to total recorded elements) can, in both cases, reach 100%. These taphonomic data are of stratigraphic interest since they provide an independent test of the cycles distinguished in sequence stratigraphy.

Introduction

Preservational characters and the distribution of the ammonites are a junction of, and enable the interpretation of, different sedimentary environments on Mesozoic epicontinental platforms (Fig. 1; Femández-López, 1997a). The ammonite associations preserved in distal and deep environments of the platform show distinctive characters which contrast with those formed in proximal and shallow environments. Some significant features of sedimentary basins, such as variation in the degree of communication between sedimentary environments, as well as water turbulence, rate of sedimentation, and rate of sediment accumulation, can be estimated on the basis of changes in the state of preservation of ammonites. A clear distinction should be made between rate of sedimentation and rate of sediment accumulation (Gómez & Fernández-López, 1994). The rate of sedimentation of a stratigraphic interval is calculated by dividing sediment thickness by the total time interval, including the gaps. In contrast, the rate of accumulation of a stratigraphic interval can be estimated by dividing sediment thickness by the time interval of positive net sedimentation. The distinction between these concepts allows one to predict that the degree of sedimentary and stratigraphic condensation will be higher towards the distal portions of the platforms, whereas the stratigraphic condensation processes without sedimentary condensation will show the maximum intensity and frequency in the shallowest portions of the platforms. Variations in water turbulence and rate of sedimentation are two major factors in these sedimentary environments. From deep to shallow areas, when increases in the water turbulence are associated with decreases in rate of sedimentation, several processes of taphonomic alteration are intensified (Fig. 1). Consequently, more than forty taphonomic gradients can be used to recognize changes in the palaeoenvironmental conditions.

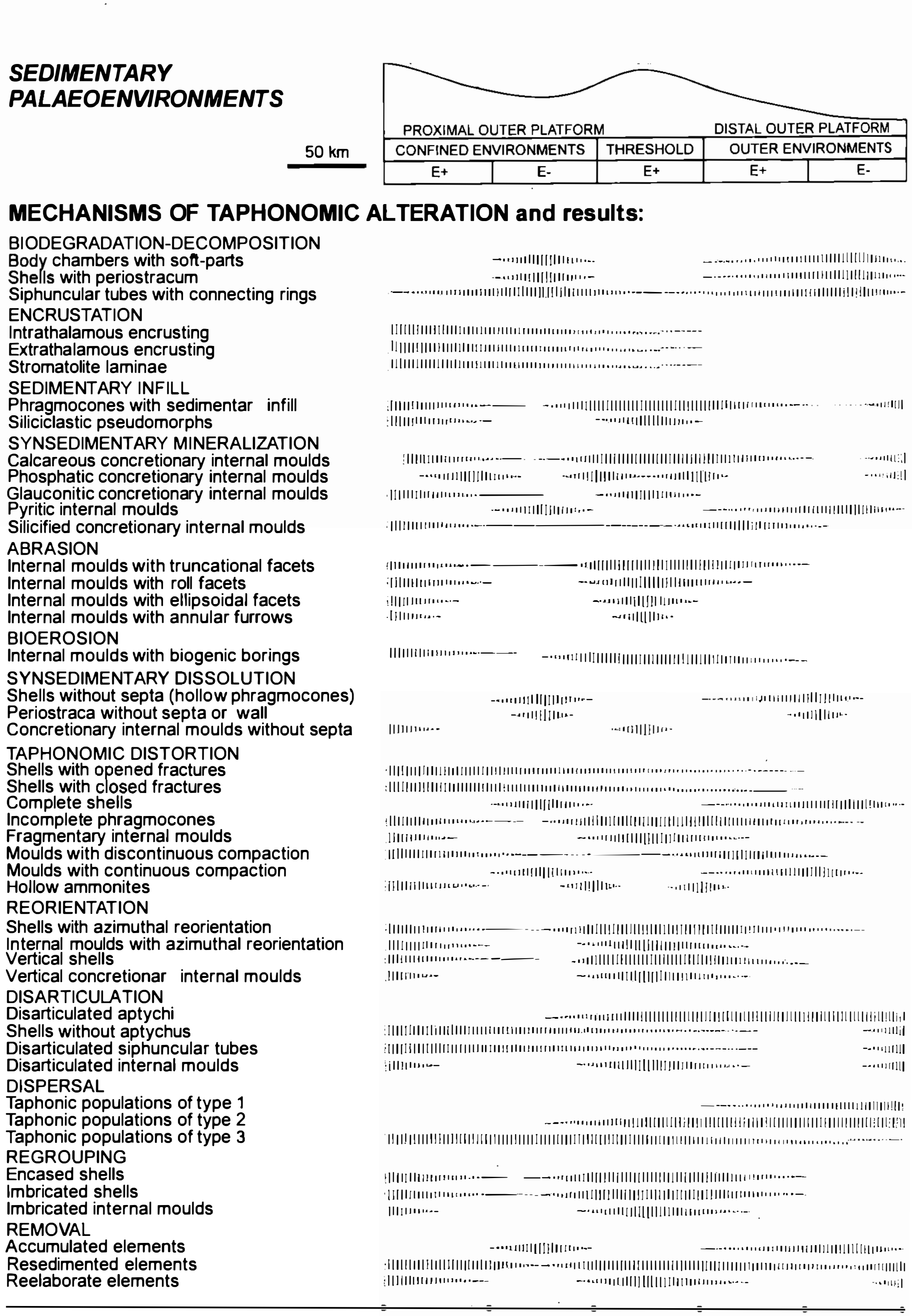

Figure 1. Taphonomic gradients developed in outer platform environments and observed in Jurassic ammonites of the Iberian Range (Fernández-López, 1997a).

Stratigraphic Cycles and Taphonomic Cycles

Stratigraphic cycles result from cyclical environmental modifications (e.g., eustatic, climatic and/or tectonic modifications). In the stratigraphic record it is possible to distinguish stratigraphic cycles and sequences of different order which result from relative sea-level changes. In a similar way, some variations of the preservational features of the successive recorded associations of ammonites in carbonate epicontinental platforms enable distinction of taphonomic cycles and sequences, of different order, induced by relative sea-level changes. A taphonomic cycle or a taphocycle comprises two or more successive recorded-associations showing cyclical variations in their taphonomic characters, as the result of an environmental cycle. Relationships between cyclical processes that have conditioned the continuity/discontinuity of the stratigraphical record or of the fossil record in carbonate epicontinental platforms can be tested on the basis of the relative duration of such processes (Fig. 2). Calibration between stratigraphic cycles and taphocycles was achieved using geochronological and geochronometric data published by Gradstein et al. (1994), Callomon (1995), Odin et al. (1995) and Gradstein and Ogg (1996).

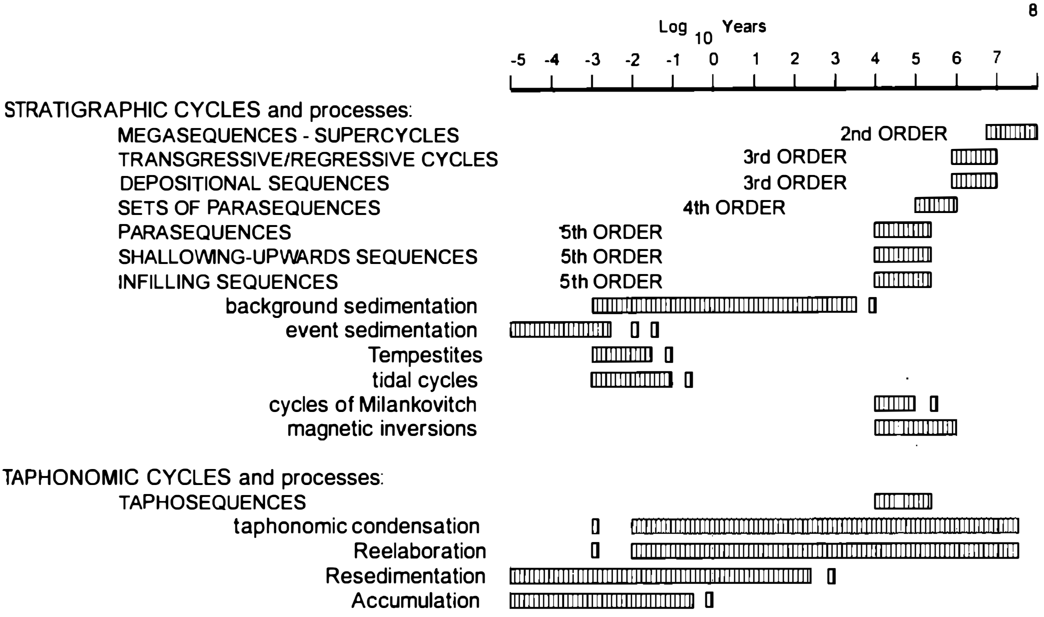

Figure 2. Resolution of stratigraphic and taphonomic cycles of different order, as well as of some processes mentioned in text.

Stratigraphic cycles, stratigraphic sequences, parasequences and taphofacies are genetic terms employed in sequence stratigraphy and applied taphonomy, comprising rock bodies of the stratigraphical record. Taphocycles, taphosequences and taphorecords are genetic terms of taphonomy comprising preserved elements, taphonic populations or preserved associations of the fossil record. The identification of taphonomic sequences and cycles enables testing of genetic differences between the fossil record and the stratigraphical record.

The stratigraphic record and the fossil record are different in nature, and they can be dissociated and studied separately (Fernández-López, 1991). The fossil record may supply relevant data on sedimentary environments and processes which themselves have left no traces in the stratigraphic record. A single stratigraphic level can enclose a set of successive recorded-associations showing distinctive preservational features and composing discrete taphorecords. This set of preserved associations forming a condensed association can even correspond to a time interval without a corresponding stratigraphic record. Interpreting these processes of taphonomic condensation in carbonate epicontinental platforms requires taking into account the relative duration of such taphonomic processes as accumulation, resedimentation and reelaboration (Fig. 2). Accumulation of ammonite shells on the floors of deep sea basins may be practically instantaneous after the death of the organisms. However, in carbonate epicontinental platforms, accumulation of ammonite shells can take place at least several months after death (Fernández-López, 1997a). Resedimented shells may stay on the depositional surface for several tens of years before being buried. Reelaborated internal moulds of ammonites may stay on the depositional surface after being exhumed, and before final burial, for more than ten million years, as indicated by many ammonite moulds subjected to reelaboration processes in the Castilian, Aragonese and Tortosa platforms during the Middle Jurassic (Fernández-López et al., 1996).

Stratigraphic Sequences and Ammonite Taphosequences

Carbonate sediments of shallow epicontinental platforms are organized in shallowing-upwards sequences and infilling sequences, of metre to decimetre thickness, which represent changes in relative depth between subtidal or intertidal and supratidal environments. These sequences represent cyclical variations of fifth-order, for which a time interval of 20,000 to 100,000 years has been estimated, according to Einsele (1992) and Vera Torres (1994); also, 10,000 to 200,000 years according to Miall (1995, 1997).

During the development of shallowing-upwards sequences, variations in the degree of removal (i.e., proportion of resedimented elements plus reelaborated elements) and taphonomic heritage (i.e., proportion of reelaborated elements) of the ammonite associations will depend on variations in rates of sedimentation and in rates of sediment accumulation, rather than exclusively on variations of water turbulence. The degree of removal and the degree of taphonomic heritage of the ammonite associations are inversely proportional to both the rate of sedimentation and rate of sediment accumulation. A decrease in either rate will produce an increase in the degree of taphonomic removal and taphonomic heritage, leading to the development of a positive taphosequence. Yet, an increase in either the rate of sedimentation or rate of sediment accumulation will produce a decrease in the degree of removal and taphonomic heritage of the preserved association, leading to the development of a negative taphosequence (Fernández-López, 1997b). These positive and negative taphonomic sequences (or taphosequences of increasing or decreasing turbulence), enable identification of shallowing-upwards sequences and infilling sequences of fifth-order, respectively.

Shallowing-upwards sequences in carbonate outer platform, and positive taphosequences, were formed during phases of increasing water turbulence and decreasing rate of sedimentation. In contrast, infilling sequences of fifth-order and negative taphosequences were formed during phases of decreasing water turbulence and increasing rate of sedimentation (Fig. 3). In outer platform environments, when decreases in rates of sedimentation are associated with strong turbulence, preserved associations of ammonites show gradual increase in concentration and taphonomic heritage. In such conditions, some taphonomic processes such as biodegradation-decomposition, encrustation, sedimentary infill, synsedimentary mineralization, abrasion, bioerosion, synsedimentary dissolution, fragmentation, reorientation, disarticulation, regrouping and removal of ammonite remains, are intensified. In contrast, when increases in the rate of sedimentation are associated with lowering of turbulence, decreased influence of the same taphonomic processes leads to the formation of ammonite associations with low values of concentration and taphonomic heritage.

Preserved associations of ammonites generated in different environmental conditions compose separate taphorecords. Each taphorecord comprises one or more preserved associations showing distinctive preservational features. Taphorecords and taphofacies have different meanings. Taphorecords are units comprising fossils. Taphofacies comprise rock bodies of the stratigraphical record. Preserved associations of ammonites generated in different phases of these environmental cycles show distinctive preservational characters and compose separate taphorecords. As indicated in Figure 3, shallowing-upwards sequences in carbonate outer platforms, preserving a positive taphosequence and different taphorecords, were formed during a phase of increasing turbulence and decreasing rate of sedimentation. Preserved associations found in positive taphosequences can be grouped in three successive taphorecords: a low turbulence taphorecord (LTT), a moderate turbulence taphorecord (MTT) and a high turbulence taphorecord (HTT). High turbulence taphorecords (HTT) are predominant in shallowing-upwards sequences developed in shallow environments of proximal platforms. In contrast, low turbulence taphorecords (LTT) are commonly formed in shallowing-upwards sequences developed in deep environments of distal platforms.

In the lower portion of a complete shallowing-upwards sequence, where accumulated elements and pyritic ammonites may be found, complete shells are most common. Hollow ammonites (i.e., showing no sedimentary infill in the phragmocone) and hollow phragmocones (i.e., without septa) are the dominant fossils, but are usually compressed by diagenetic compaction. The occurrence of taphonic populations of type 1 (i.e., composed of monospecific shells showing unimodal and asymmetric distribution of size-frequencies, with positive skew) is indicative of autochthonous biogenic production, showing no signs of sorting by necroplanktic drift. Resedimented and reelaborated ammonites become more common in the upper portions of these sequences, as shells are completely infilled with sediment and tend to acquire an encased pattern of grouping.

Towards the top of the sequence, processes of early mineralization are more intense and taphonic populations of type 3 (i.e., composed of polyspecific shells showing uni- or polymodal and asymmetric distribution of size-frequencies, with negative skew) are dominant. Reelaborated concretionary internal moulds become dominant. They may display several distinctive features such as abrasion, fragmentation, disarticulation, reorientation and regrouping. Such reelaborated elements show no traces of deformation by gravitational diagenetic compaction during early burial. However, they may develop abrasion facets formed before final burial. Ammonite shells and concretionary internal moulds tend to produce imbricate patterns of grouping and show azimuthal reorientation. They may be covered by encrusting organisms and biogenic borings. Siphuncular tubes are usually disarticulated as a consequence of intense and lasting biostratinomic processes of biodegradation-decomposition and dissolution. Reelaborated concretionary internal moulds are also preferentially disarticulated along septal surfaces. In the later stage, concretionary internal moulds with ellipsoidal abrasion facets and annular abrasion furrows are formed. Long episodes of emersion and erosion favour the formation of concretionary internal moulds without septa, resulting from dissolution of the septa and filling of interior cavities with sediments (Fig. 4).

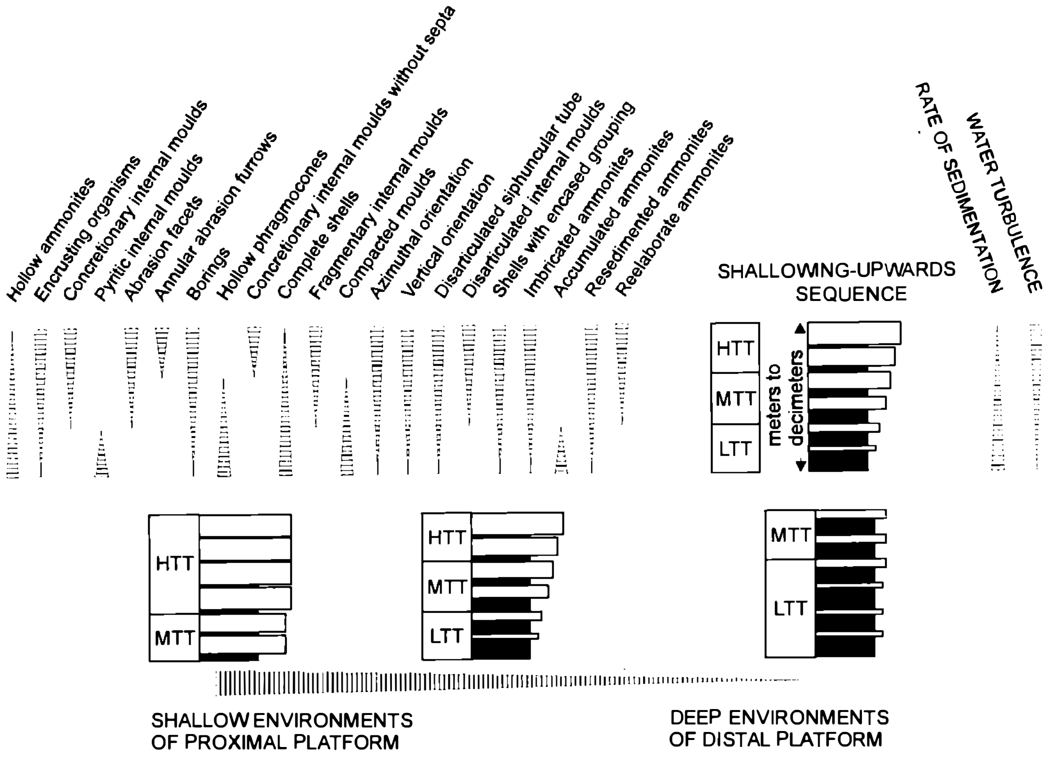

Figure 3. Frequency of different taphonomic characters displayed by ammonites in shallowing-upwards sequences in carbonate outer platform, forming a positive taphosequence and different taphorecords (Fernández-López, 1997b). LTT = low turbulence taphorecord. MTT = moderate turbulence taphorecord. HTT = high turbulence taphorecord.

TAPHONOMIC PROCESSES and results:

BIODEGRADATION-DECOMPOSITION
Body chamber without soft-parts
Shell without periostracum

DISARTICULATION
Shell without aptychus
Disarticulated siphuncular tube

RESEDIMENTATION
Fragmented wall

SEDIMENTARY INFILL
Body chamber with sedimentary infill (F1)
Phragmocone with partial sedimentary infill
(=hollow ammonite)

INITIAL BURIAL (A)
Umbilical cavities with sedimentary infill (F1)

SYNSEDIMENTARY MINERALIZATION
Concretionary internal mould of the shell
Mineralized umbilical plugs of the shell

REELABORATION (B)
Exhumed and moved internal mould
Mineralized umbilical plugs with roll facets

SYNSEDIMENTARY DISSOLUTION (C)
Shell without septa (hollow phragmocone)
Dissolved wall

TAPHONOMIC DISTORTION
Collapsed umbilical plugs
Collapsed inner whorls
Disarticulated internal mould

SEDIMENTARY INFILL (D)
Internal mould without septa (F2)

SYNSEDIMENTARY MINERALIZATION
Concretionary internal mould without septa

REELABORATION
Exhumed and moved, concretionary internal mould
Concretionary internal mould with roll facets

FINAL BURIAL AND COMPACTION (E)
Compacted concretionary internal mould,
without septa

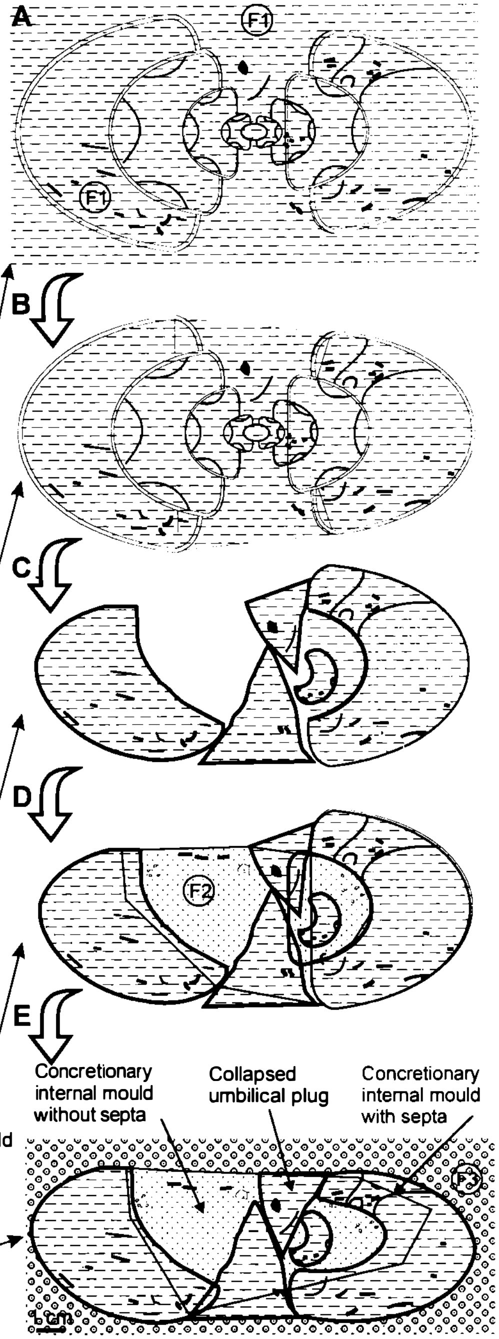

Figure 4. Model for formation of concretionary internal moulds without septa, results of dissolution of septa and infilling of cavities with sediment.

Condensed sections show different characters in shallow and proximal environments as compared to deep and distal environments (Table 1). The highest values of taphonomic condensation in ammonite associations are displayed in shallow epicontinental platforms rather than in deep sea environments. However, the degree of taphonomic heritage (estimated by the ratio of reelaborated elements in the whole assemblage) can reach 100% in both cases. The degree of packing of ammonites (estimated by the difference between the number of specimens and the number of fossiliferous levels subdivided by the number of fossiliferous levels) and the stratigraphical persistence (proportion of fossiliferous levels) display smaller values in proximal than in distal areas. Condensed sections in deep and distal areas usually contain taphonic population of type 1 (Fernández-López, 1997a). In such areas, phragmocones are normally filled by sediment, and concretionary internal moulds display disarticulation surfaces and fractures with acute margins. Pyritic ammonites are common in certain distal areas. On the other hand, in proximal areas, taphonic populations are usually of type 2 or 3, those of type 1 not being represented. Hollow ammonites (i.e., shells showing no sedimentary infill in the phragmocone) are abundant, reelaborated internal moulds show high values of roundness and sphericity as well as frequent biogenic borings, and pyritic ammonites are scarce (Fernández-López, 1997b). Stratigraphic successions in shallow epicontinental platforms are usually more incomplete than those formed in deep basins (cf. Schindel, 1982; McKinney, 1985; Kowalewski, 1996). However, despite the abundance and wide range of biostratigraphic gaps in such sequences, registratic gaps are usually not so important in condensed sections of shallow platforms, the registratic succession being usually more complete than the corresponding biostratigraphic succession.

Condensed sections in shallow and proximal areas:	Condensed sections in deep and distal areas:
Expanded sediments (e.g., tempestites)	Condensed sediments
High taphonomic condensation	Low taphonomic condensation
High taphonomic heritage	Moderate taphonomic heritage
Low degree of packing	High degree of packing
Low stratigraphic persistence	High stratigraphic persistence
Taphonic populations of type 1 are absent	Taphonic populations of type 1 are present
Abundant hollow ammonites	Scarce hollow ammonites
Rounded reelaborate ammonites	Angular reelaborate ammonites
Abundant biogenic borings	Scarce biogenic borings
Scarce pyritic internal moulds	Common pyritic internal moulds
Common stratigraphic gaps	Scarce stratigraphic gaps

Table 1. Differential characters of the condensed sections formed in shallow and proximal areas in relation to deep and distal areas of carbonate epicontinental platforms (Fernández-López, 1997b).

Conclusions

On the basis of changes in the preservational state of successive recorded associations of ammonites, it is possible to distinguish taphocycles and taphorecords resulting from relative changes of sea level in carbonate epicontinental platforms. Preserved associations of ammonites formed in these platforms can be grouped on the basis of taphonomic criteria in taphorecords of different categories: low turbulence taphorecords (LTT), moderate turbulence taphorecords (MTT) and high turbulence taphorecords (HTT). Positive and negative taphosequences, resulting from increasing or decreasing turbulence respectively, enable identification of shallowing-upwards sequences and infilling sequences of fifth-order, respectively.

Condensed sections show different characters in shallow and proximal areas as compared to deep and distal areas. The degree of taphonomic condensation in ammonite preserved associations reaches the highest values in shallow epicontinental platforms, not in deep basins, though the degree of taphonomic heritage (i.e., the ratio of reelaborated elements to the total recorded elements) can, in both cases, reach 100%.

Using paleontological data in sequence stratigraphy analysis of shallow marine platforms requires the prior identification of taphonomic sequences and cycles. These taphonomic data are of stratigraphic interest since they provide an independent test of the cycles distinguished in sequence stratigraphy and genetic stratigraphy. The identification of such taphonomic cycles is of utmost importance in interpreting the stratigraphic cycles of Mesozoic epicontinental platforms when no evidence of coastal onlap is preserved but fossiliferous sediments of the outer platform are widely developed.

Acknowledgments

The author wishes to express his acknowledgement to Dr. G. Meléndez (Faculty of Sciences, University of Zaragoza) and two anonymous referees for critically reading the manuscript. This work is a contribution to the projects PB91-0383 (DGICYT) and PB96-0386 (DGES - CSIC).

References

CALLOMON, J. H. 1995. Time from fossils: S. S. Buckman and Jurassic high-resolution geochronology, p. 127-150. *In* M. J. Le Bas (ed.), Milestones in Geology. Geological Society, London, Memoir 16.

EINSELE, G. 1992. Sedimentary Basins. Springer, Berlin, 628 pp.

FERNÁNDEZ-LÓPEZ, S. 1991. Taphonomic concepts for a theoretical biochronology. Revista Española de Paleontología, 6:37-49

____. 1997a. Ammonites, clinos tafonómicos y ambientes sedimentarios. Revista Española de Paleontología, 12:102-128.

____. 1997b. Ammonites, taphonomic cycles and stratigraphic cycles in carbonate epicontinental platforms. Cuadernos de Geología Ibérica, 23:95-136.

____, M. AURELL, F. GARCÍA JORAL, J. J. GÓMEZ, M. H. P. HENRIQUES, G. MARTÍNEZ, G. MELÉNDEZ, AND L. C. SUÁREZ VEGA. 1996. El Jurásico Medio de la Cuenca Catalana: unidades litoestratigráficas y elementos paleogeográficos. Revista Española de Paleontología: 122-139.

GÓMEZ, J. J. AND S. FERNÁNDEZ-LÓPEZ. 1994. Condensation processes in shallow platforms. Sedimentary Geology, 92:147-159.

GRADSTEIN, F. M., F. P. AGTERBERG, J. G. OGG, J. HARDENBOL, P. VAN VEEN, J. THIERRY, AND Z. HUANG. 1994. A Mesozoic time scale. Journal of Geophysical Research, 99:24,051-24,074.

____, AND J. OGG. 1996. A Phanerozoic time scale. Episodes, 19:3-5.

KOWALEWSKI, M. 1996. Time-averaging, overcompleteness, and the geological record. The Journal of Geology, 104:317-326.

MCKINNEY, M. L. 1985. Distinguishing patterns of evolution from patterns of deposition. Journal of Paleontology, 59:561-567.

MIALL, A. D. 1995. Whither stratigraphy? Sedimentary Geology, 100:5-20.

____. 1997. The geology of stratigraphic sequences. Springer, Berlin, 433 p.

ODIN, G. S., B. GALBRUN, AND M. RENARD. 1995. Physico-chemical tools in Jurassic stratigraphy, p. 507-518. *In* E. Cariou and P. Hantzpergue (eds.), 3rd International Symposium on Jurassic stratigraphy, Poitiers, 1991. Geobios, M.S. 17 (1994)

SCHINDEL, D. L. 1982. Resolution analysis: a new approach to the gaps in the fossil record. Paleobiology, 8:340-353.

VERA TORRES, J. A. 1994. Estratigrafía. Principios y métodos. Editorial Rueda, Madrid, 806 p.

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Definition and Organization of Limestone-Marl Cycles in the Toarcian of the Northern and East-Central Part of the Iberian Subplate (Spain)

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Keywords: Lower Jurassic, Cycles, Sequences, Isochronism, Eustasy, Biostratigraphy

Abstract: An analysis of the organization of the limestone-marl cycles of the Toarcian sediments located north and east of the Iberian Massif has been carried out. Sections representative of different paleogeographical environments located in the Asturian, Basque-Cantabrian and Iberian basins were selected. Four deepening and four shallowing sequences, with marked isochronism, have been recognized in the whole area. The Toarcian transgression included several deepening and shallowing episodes: the Lower Tenuicostatum deepening and shallowing sequences; the Tenuicostatum–Serpentinus deepening sequence; and the Upper Tenuicostatum shallowing sequence. Maximum bathymetric values were reached during the Serpentinus-Bifrons deepening sequence which is followed by the Bifrons-Variabilis shallowing sequence, bounded by a regional unconformity. The Thouarsense-Insigne deepening sequence, which represents a new flooding of the platforms, is followed by the Pseudoradiosa-Aalensis shallowing sequence which continues into the Aalenian. Isochronism is more evident in the tops of the deepening upward sequences, located in the upper part of the Tenuicostatum Zone, within the Bifrons Zone and in the upper part of the Insigne Zone. Results are homogeneous in sections having different evolutionary histories, suggesting that eustatic sea level change is the dominant factor in the generation of the sequences. In the area of research, tectonic activity does not impede recognition of the higher rank sequences.

Introduction

In the Iberian subplate, located between the African and the Euro-Asiatic Plates, Jurassic sediments crop out around the Iberian Massif. In this work, the results of a study of the organization of the limestone-marl cycles in sections of Toarcian sediments are presented. Studied sections are located in the Asturian, Basque-Cantabrian and Iberian basins, located to the north and east of the Iberian Massif (Fig. 1).

Four sections characterized by an alternation of limestones and marls have been studied bed-by-bed. The sections were selected on the basis of their relative sedimentary continuity. However, at a regional scale a total of five units bounded by unconformities have been recognized (Comas-Rengifo et al., 1988; Goy et al., 1994). In these sections ammonite biostratigraphic data allow high-resolution stratigraphy.

In the north, sediments of the Rodiles-Santa Mera section (Fig. 1), representative of the Asturian Basin, correspond to an outer, quite stable and unrestricted platform. The San Andrés section represents the Basque-Cantabrian Basin. The La Almunia-Ricla section is located in a relatively stable area of the northeastern part of the Iberian Basin. The Rambla del Salto (Sierra Palomera) section is located in the central sector of the Iberian Basin, in an area where sedimentation was affected by the movement of active faults.

The rhythmic alternation of marls and limestones (normally mudstones) that constitutes these sections, is organized in thickening-upward sequences characterized by increase or decrease in thickness of the limestone or marls toward the upper part of the sequence. In the sequences where thickness of the limestone beds increases upward, the marly beds generally decrease or maintain their thickness. As a consequence, the limestone/marl ratio increases toward the upper part of the sequence. These sequences have been interpreted by many authors as shallowing upward sequences

(Hallam, 1988; Gómez and Fernández-López, 1994) on the basis of sedimentological, paleoecological and taphonomical criteria (Matyja and Seilacher, 1985; Fernández-López and Suárez-Vega, 1979). On the other hand, in the sequences where the thickness of marls increases toward the upper part of the sequence, limestone beds generally decrease in thickness upwards or maintain a similar thickness. As a consequence, the limestone/marl ratio decreases upwards. In contrast to the previously mentioned sequences, these marly thickening upward sequences can be interpreted as deepening upward sequences (Gómez and Goy, 1996). In turn, several sets of elementary sequences composed of shallowing upward sequences and sets composed of deepening upward sequences can be distinguished. The identification of these cycles has been supported by gamma ray measurements of the outcrops.

The objectives of the present work are to establish the cyclic organization of the sediments of the studied sections, to determine the relationship between the sets of sequences identified in the different basins, to check their position regarding the previously defined biostratigraphic units, and to correlate the sets of sequences found in platforms located in different paleogeographical environments or different structural positions.

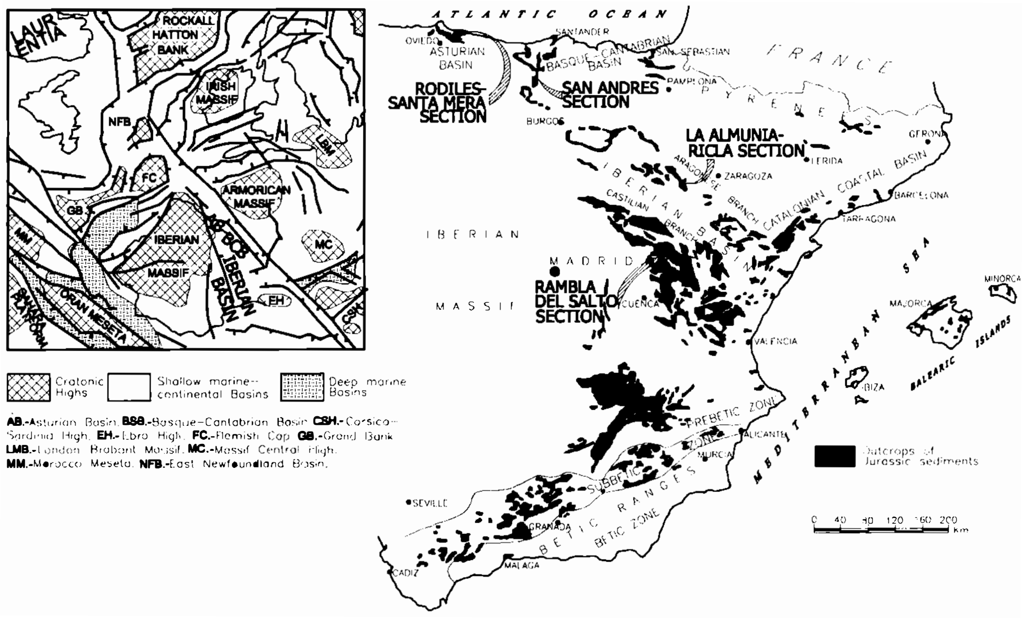
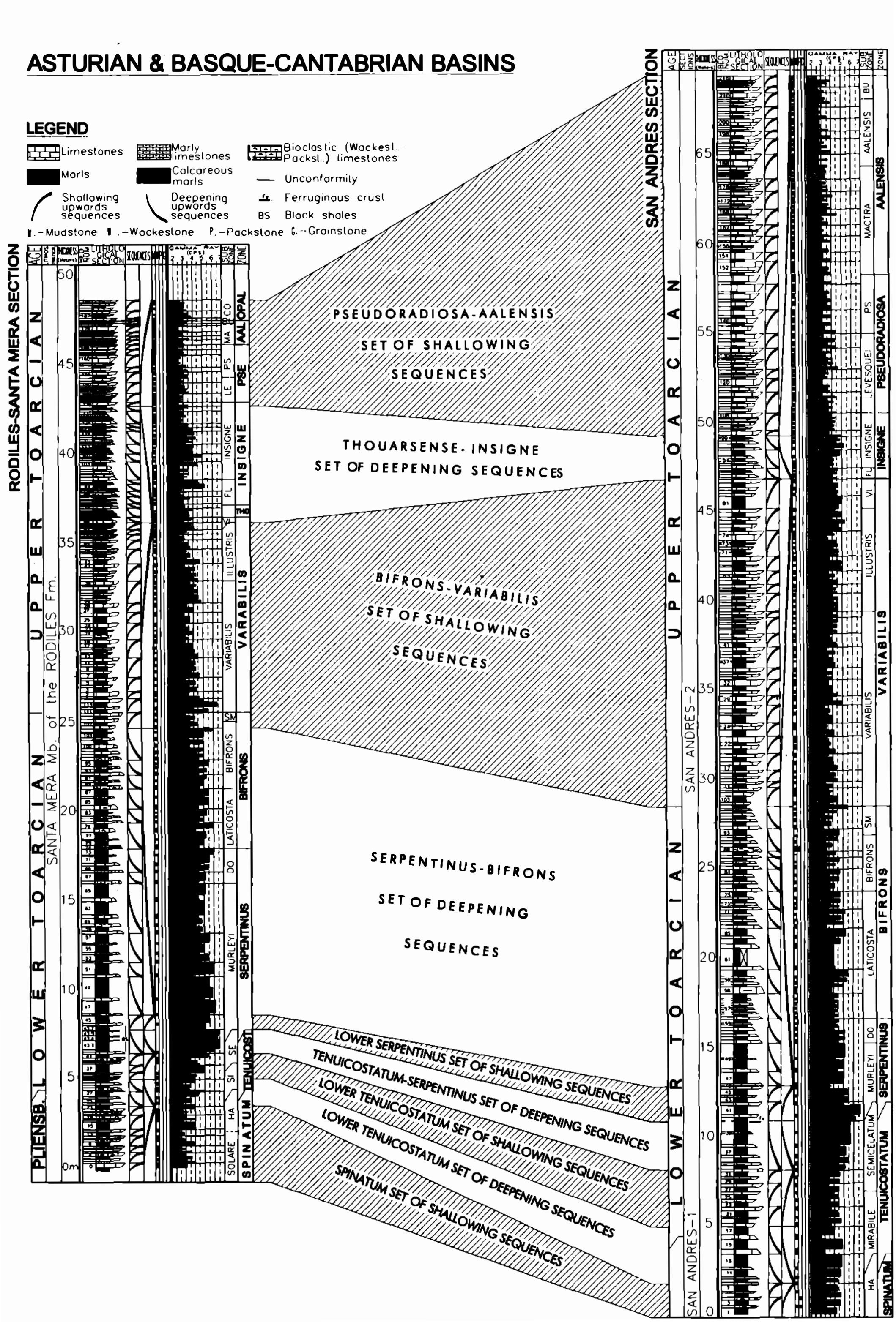


Figure 1. Paleogeography of the Upper Liassic (modified from Ziegler, 1990), showing the northern and eastern Toarcian basins of Spain. Right figure shows the outcrops of Jurassic sediments in Spain as well as the location of the sections in this paper.

Figure 2. Correlation between the Rodiles-Santa Mera Section and the San Andres Section showing Toarcian sequences. Legend for the Zones and Subzones: SPI.-Spinatum. THO.-Thouarsense. PSE.-Pseudoradiosa. AAL.-Aalensis. OPA.-Opalinum. HA.-Hawskerense. SI.-Simplex. SE.-Semicelatum. FA.-Falciferum. DO.-Douvillei. SM.-Semipolitum. VI.-Vitiosa. BI.-Bingmanni. TH.-Thouarsense. FS.-Fascigerum. FL.-Fallaciosum. LE.-Levesquei. PS.-Pseudoradiosa. MA.-Mactra. BU.-Buckmani. CO.-Comptum.


ASTURIAN & BASQUE-CANTABRIAN BASINS
LEGEND
Limestones
Marly limestones
Bioclastic (Wackest.–Packst.) limestones
Marls
Calcareous marls
Unconformity
Shallowing upwards sequences
Deepening upwards sequences
Ferruginous crust
BS Black shales
I.-Mudstone
I.-Wackestone
P.-Packstone
G.-Grainstone
RODILES-SANTA MERA SECTION
SAN ANDRES SECTION
AGE
THICKNESS
LITHOLOGICAL SECTION
SEQUENCES
GAMMA RAY (CPS)
SUBZONE
ZONE
PLIENSB.
LOWER TOARCIAN
UPPER TOARCIAN
SANTA MERA Mb. of the RODILES Fm.
SAN ANDRES-1
SAN ANDRES-2
0m
SPINATUM
SOLARE
HA
SI
SE
TENUICOSTATUM
MURLEYI
SERPENTINUS
DO
LATICOSTA
BIFRONS
SM
VARIABILIS
ILLUSTRIS
VI
THO
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MIRABILE
SEMICELATUM
LEVESQUEI
PSEUDORADIOSA
MACTRA
AALENSIS
BU
PSEUDORADIOSA-AALENSIS SET OF SHALLOWING SEQUENCES
THOUARSENSE-INSIGNE SET OF DEEPENING SEQUENCES
BIFRONS-VARIABILIS SET OF SHALLOWING SEQUENCES
SERPENTINUS-BIFRONS SET OF DEEPENING SEQUENCES
LOWER SERPENTINUS SET OF SHALLOWING SEQUENCES
TENUICOSTATUM-SERPENTINUS SET OF DEEPENING SEQUENCES
LOWER TENUICOSTATUM SET OF SHALLOWING SEQUENCES
LOWER TENUICOSTATUM SET OF DEEPENING SEQUENCES
SPINATUM SET OF SHALLOWING SEQUENCES

Asturian and Basque-Cantabrian Basins

The Toarcian sediments in the Asturian Basin consist of an irregular alternation of mudstone and marly limestones or marls. In the lower part of this succession occurs an interval of black shales locally enriched in organic matter to values of up to 2.5% TOC. The sediments of the Rodiles-Santa Mera section (Suárez-Vega, 1974; Goy et al., 1997) correspond to the Santa Mera Member of the Rodiles Formation (Valenzuela et al., 1986). No important unconformities near the Pliensbachian-Toarcian boundary were recognized. The Lower Toarcian Tenuicostatum, Serpentinus and Bifrons Zones are well developed. In the Upper Toarcian, the Variabilis, Insigne and Pseudoradiosa Zones show a notable development while the Thouarsense and Aalensis Zones are thinner and contain unconformities (Fig. 2).

At San Andrés (Goy et al., 1994), Toarcian sediments constitute an expanded section of limestones and marls deposited in a relatively subsiding area. All the zones are present except for the Thouarsense Zone. The Pseudoradiosa and Aalensis Zones are the thickest in the sections analyzed.

Iberian Basin

In the Almunia-Ricla section (Goy and Martínez, 1990), the Toarcian is represented by alternating marls and lime mudstone of the Turmiel Formation (Goy et al., 1976) and, in the upper portion, by the mudstone-wackestone carbonates of the Chelva Formation (Gómez and Goy, 1979). The sedimentary record in this section is very complete with all the Toarcian ammonite zones and subzones of the Iberian Basin represented. Ammonites from 168 levels were collected and 35 biohorizons defined (Fig. 3).

In the Rambla del Salto section (Comas-Rengifo et al., 1996), Toarcian sediments are represented by the upper part of the Barahona Formation, consisting of bioclastic wackestone-packstone and an alternation of marls and lime mudstone to wackestone corresponding to the Turmiel Formation. The upper part of the section is represented by carbonates of the Chelva Formation (Fig. 3). Toarcian sedimentation occurred in an unstable area affected by extensional tectonic activity. During the Early Toarcian, this area occupied a relative high, although in the Variabilis Zone it corresponded to a subsiding block. The Thouarsense Zone is partially absent. In the uppermost Toarcian, a considerable reduction of thickness is recorded caused by irregularities in sedimentation and the presence of a Toarcian-Aalenian hiatus.

Sequence Analysis

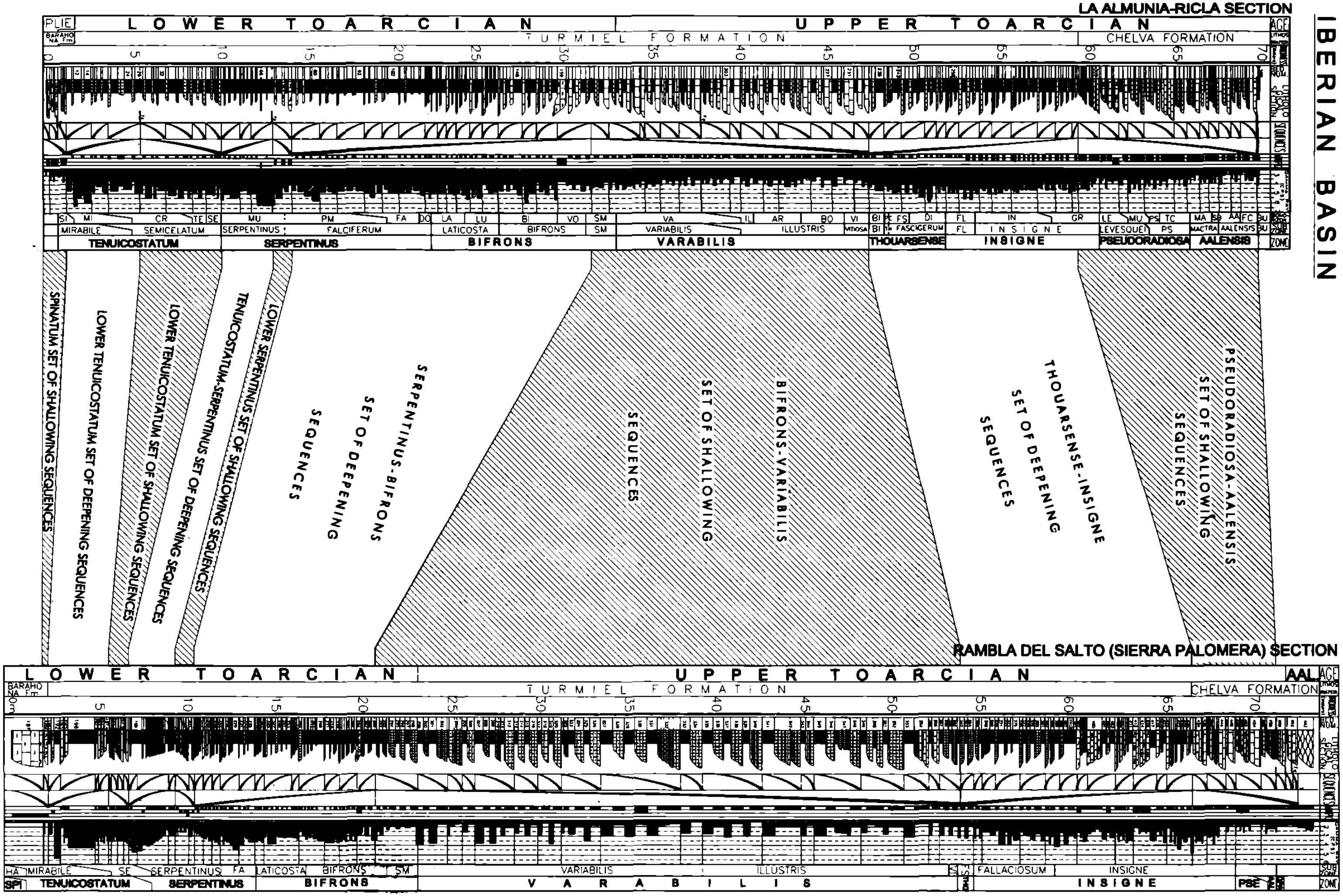
To carry out the sequence analysis of the Toarcian sediments, the La Almunia-Ricla section was chosen as the main reference because sedimentation is more regular and the stratigraphic record more complete. This section is representative of a low energy outer platform although it was occasionally affected by storm action. The La Almunia-Ricla section has been compared with the Rambla del Salto section located in a relatively unstable area of the same basin, where the water was generally shallower and tempestite sedimentation more prevalent.

Single outcrops have also been studied in other platforms with similar and different paleogeographical settings and evolutionary development. The San Andrés section in the Basque-Cantabrian Basin represents similar environments to those found in the La Almunia-Ricla section. In the Asturian Basin, the Rodiles-Santa Mera section represents an outer, relatively deeper platform, in good communication with the platforms of central and northern Europe.

The Pliensbachian, independent of facies, ends with a shallowing sequence (Spinatum set of shallowing sequences) in all the studied areas (Figs. 2 and 3).

In the Toarcian there are four episodes or sets in which the sediments are organized in deepening sequences alternating with four episodes of shallowing sequences (Figs. 2 and 3). The first set of deepening sequences occurs in the lower part of the Tenuicostatum Zone (Lower Tenuicostatum set of deepening sequences). A significant input of marly sediments with abundant nektonic organisms also took place on all the platforms. This is specially well marked in the Iberian Basin where this episode, coinciding with the expansion of the Tethys domain to the north and west of Europe, resulted in the transition from a shallow platform environment to a low energy outer platform environment. This change is marked in the studied stratigraphic columns by a significant increase in the values of radioactivity (Fig. 3).

Figure 3. Correlation between the La Almunia and the Rambla del Salto (Sierra Palomera) Section (Iberian Range) showing Toarcian sequences. Legend in Fig. 2.


IBERIAN BASIN
LA ALMUNIA-RICLA SECTION
PLIE
LOWER TOARCIAN
UPPER TOARCIAN
AGE
BARAHONA Fm
TURMIEL FORMATION
CHELVA FORMATION
0
5
10
15
20
25
30
35
40
45
50
55
60
65
70
LITHOLOGIC SECTION
SEQUENCES
TENUICOSTATUM
SERPENTINUS
BIFRONS
VARABILIS
THOUARSENSE
INSIGNE
PSEUDORADIOSA
AALENSIS
ZONE
SUBZONE
MIRABILE
SEMICELATUM
FALCIFERUM
LATICOSTA
SM
VARIABILIS
ILLUSTRIS
FASCIGERUM
FL
LEVESQUEI
PS
MACTRA
SPINATUM SET OF SHALLOWING SEQUENCES
LOWER TENUICOSTATUM SET OF DEEPENING SEQUENCES
LOWER TENUICOSTATUM SET OF SHALLOWING SEQUENCES
TENUICOSTATUM-SERPENTINUS SET OF DEEPENING SEQUENCES
LOWER SERPENTINUS SET OF SHALLOWING SEQUENCES
SERPENTINUS-BIFRONS SET OF DEEPENING SEQUENCES
BIFRONS-VARIABILIS SET OF SHALLOWING SEQUENCES
THOUARSENSE-INSIGNE SET OF DEEPENING SEQUENCES
PSEUDORADIOSA-AALENSIS SET OF SHALLOWING SEQUENCES
RAMBLA DEL SALTO (SIERRA PALOMERA) SECTION
AAL
0m
MIRABILE
SERPENTINUS
LATICOSTA
FALLACIOSUM
SPI
HA
SE
FA
SM
PSE

A set of shallowing sequences developed in the lower part of the Tenuicostatum Zone (Lower Tenuicostatum set of shallowing sequences). During the deposition of these sequences important changes in the distribution of some benthic organisms occurred (Arias et al., 1992). The overlying Tenuicostatum-Serpentinus set of deepening sequences developed during the upper part of the Tenuicostatum Zone which is characterized by anoxic black-shale facies in the Asturian and Basque-Cantabrian Basins as well as suboxic facies in the northeastern part of the Iberian Basin. This is probably the cause of the extinction of numerous taxonomic groups of benthic and nektonic organisms. This episode, in which a remarkable increase in the radioactive background of the sediments is recorded, can be correlated with the anoxic episode described by Jenkyns (1988) in central Europe.

In the lower part of the Serpentinus Zone, a brief and marked shallowing interval (Lower Serpentinus set of shallowing sequences) is developed in all the studied basins. It presumably represents a relative fall of the sea level that is specially marked in the Rambla del Salto high. Coinciding with this shallowing episode in the Northern Iberian Basin are frequent tempestite levels with resedimented fossils and colonization by the brachiopod *Soaresirhynchia bouchardi* (Dav.). A similar phenomenon has been reported in other areas of western Tethys (Almeras and Elmi, 1993). In the Asturian and Basque-Cantabrian Basins the tempestites show more distal facies and the mentioned brachiopod is not present, probably due to greater depth of the ramp.

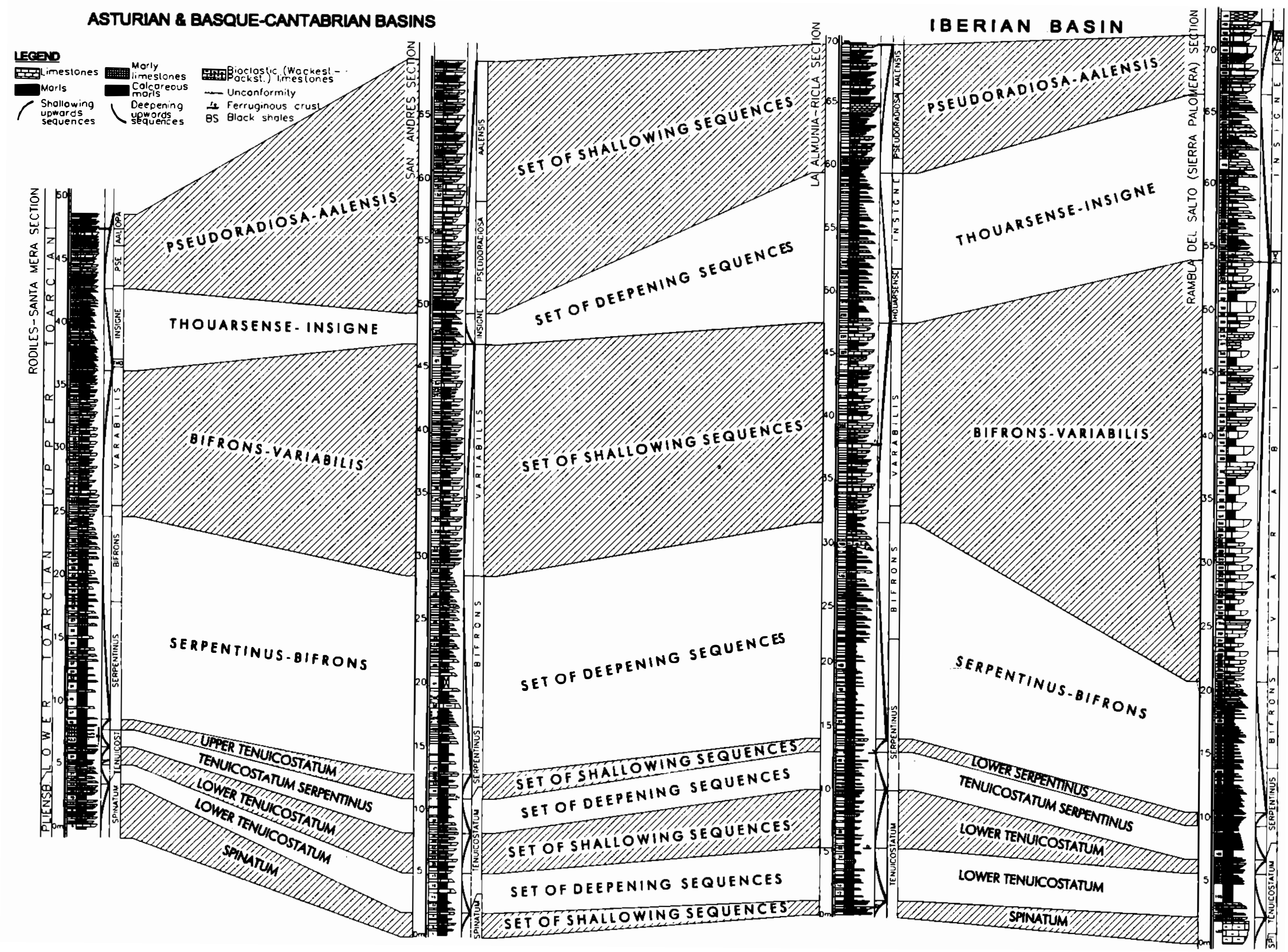
Numerous deepening sequences comprise most of the Serpentinus and Bifrons Zones (Serpentinus-Bifrons set of deepening sequences). It always has a conspicuous and similar thickness in the different sections except in the Rambla del Salto which continues to act as a relative high. It represents the stage of maximum deepening of the Toarcian in the different platforms and almost the only period of time in which the basin was deep enough as to be colonized by ammonites. This occurrence can be extended to other basins of Europe and northern Africa (Elmi, 1986; Soares et al., 1993; Elmi et al., 1994). Nevertheless, the sea bottom was affected by the action of bad weather waves, generating tempestite layers that can be followed for hundreds of kilometers (bed 284 of the Rambla del Salto, bed 159 of La Almunia-Ricla and bed 88 of San Andrés). This continuity indicates the high degree of uniformity shown by the platforms at that time.

In the uppermost part of the Bifrons Zone and the entire Variabilis Zone, a group of shallowing sequences (Bifrons-Variabilis set of shallowing sequences) of similar thickness in the different sections is recorded. In the Rambla del Salto section, the relative high found in previous episodes became a subsiding area and the site of maximum sedimentation rates for the Variabilis Zone. The low radioactivity background values recorded in this sequence are particularly evident in the sediments of the Iberian Basin.

Lastly, a new group of deepening sequences was deposited during the Thouarsense and Insigne Zones (Thouarsense-Insigne set of deepening sequences). Exceptionally, in San Andrés, the first deepening sequence begins in the upper part of the Variabilis Zone where an abrupt influx of fine terrigenous sediments took place, probably due to local tectonic activity. Except for the La Almunia-Ricla section, the Thouarsense Zone is partially absent. The Variabilis-Thouarsense unconformity (or locally Insigne, e.g., the San Andrés area) has been recognized in all the studied basins and normally coincides with the base of the Thouarsense-Insigne set of deepening sequences. In the Insigne Zone and coinciding with the relative maximum deepening, some groups of ammonites locally colonized the platform.

The upper part of the Toarcian (Pseudoradiosa and Aalensis Zones) is dominated by the presence of shallowing sequences (Pseudoradiosa-Aalensis set of shallowing sequences). In the La Almunia-Ricla and in San Andrés the sediments of these zones show a remarkable development and, significant sedimentary gaps are not common. In the Rambla del Salto section by contrast, these Zones are condensed and a relatively important hiatus in the transition between the Aalensis and Opalinum Zones exists. In the Rodiles-Santa Mera section, several remobilization levels have been reported by Fernández-López and Suárez-Vega (1979).

Figure 4. Panel showing the correlation of the Toarcian sequences. Notice that the boundaries of most of these sequences are nearly synchronous.


ASTURIAN & BASQUE-CANTABRIAN BASINS
IBERIAN BASIN
LEGEND
Limestones
Marly limestones
Bioclastic (Wackest - Packst.) limestones
Marls
Calcareous marls
Unconformity
Shallowing upwards sequences
Deepening upwards sequences
Ferruginous crust
BS Black shales
RODILES-SANTA MERA SECTION
SAN ANDRES SECTION
LA ALMUNIA-RICLA SECTION
RAMBLA DEL SALTO (SIERRA PALOMERA) SECTION
PLIENSB.
LOWER TOARCIAN
UPPER TOARCIAN
SPINATUM
TENUICOSTATUM
SERPENTINUS
BIFRONS
VARIABILIS
INSIGNE
PSEUDORADIOSA
AALENSIS
PSE
AAL
OPA
PSEUDORADIOSA-AALENSIS
THOUARSENSE-INSIGNE
BIFRONS-VARIABILIS
SERPENTINUS-BIFRONS
UPPER TENUICOSTATUM
TENUICOSTATUM SERPENTINUS
LOWER TENUICOSTATUM
LOWER TENUICOSTATUM
SPINATUM
SET OF SHALLOWING SEQUENCES
SET OF DEEPENING SEQUENCES
SET OF SHALLOWING SEQUENCES
SET OF DEEPENING SEQUENCES
SET OF SHALLOWING SEQUENCES
SET OF DEEPENING SEQUENCES
SET OF SHALLOWING SEQUENCES
SET OF DEEPENING SEQUENCES
SET OF SHALLOWING SEQUENCES
PSEUDORADIOSA-AALENSIS
THOUARSENSE-INSIGNE
BIFRONS-VARIABILIS
SERPENTINUS-BIFRONS
LOWER SERPENTINUS
TENUICOSTATUM SERPENTINUS
LOWER TENUICOSTATUM
LOWER TENUICOSTATUM
SPINATUM
0m
5
10
15
20
25
30
35
40
45
50
55
60
65
70

Correlation of the Toarcian sets of Sequences

Figure 4 shows the correlation between the studied sections. The remarkable continuity and isochronism of the distinguished sets of sequences across the different basins can be highlighted. Even in cases where the depositional areas have been subjected to different tectonic conditions, the diachronism of the boundaries of the sequences does not surpass the range of a subzone.

The Lower Tenuicostatum set of deepening sequences begins in the uppermost part of the Spinatum Zone in the Asturian and Basque-Cantabrian basins and virtually coincides with the Pliensbachian-Toarcian boundary in the sections studied in the Iberian Basin. This difference can be attributed to the "Toarcian transgression" which probably advanced from north to south. However, that limit is also slightly heterochronous within the Iberian Basin, controlled by the paleogeographical position of the outcrops relative to fault-bounded blocks that were active during sedimentation (Goy et al., 1997).

The anoxic or suboxic episodes of the uppermost part of the Tenuicostatum Zone in the different basins, as well as the top of the Lower Serpentinus set of shallowing sequences, can also be correlated accurately. Other levels of great isochronism along the different basins are the top of the Serpentinus-Bifrons deepening set of sequences and the top of the Thouarsense-Insigne set of deepening sequences, which coincide with episodes of maximum flooding of the basin.

By contrast, the top of the Bifrons-Variabilis set of shallowing sequences, as well as the top of the Pseudoradiosa-Aalensis set of shallowing sequences, are usually related to unconformities and comparison between the sections is difficult due to the local absence of stratigraphic record.

The uniformity of the sequences is also represented by the similar number of smaller scale sequences that have been recorded in the different sections. For example, in the Lower Tenuicostatum set of deepening sequences there are 2 in Asturias and Basque-Cantabrian, and 3 in the Northern Sector of the Iberian Basin. In the case of the Serpentinus-Bifrons set of deepening sequences the number of smaller sequences preserved varies between 12 (Rodiles-Santa Mera and Rambla del Salto) and 16 (La Almunia-Ricla). Also, in the case of the shallowing sequences, the number of recorded smaller sequences is very similar. In the expanded sections at Rambla del Salto and San Andrés, the number of sequences of the Bifrons-Variabilis shallowing set of sequences is 18 and 19 respectively.

Conclusions

Four sets of deepening sequences and four sets of shallowing sequences have been recognized in the Toarcian of the northern and eastern side of the Iberian subplate. These sets of sequences can easily be correlated in the whole studied area.

In general, these sets are markedly isochronous. It can be observed that the tops of the sets of deepening sequences are more isochronous than those of the shallowing sequences. In the cases in which slight heterochronism has been perceived, it is due to causes derived from local tectonic activity, to the presence of stratigraphic unconformities, or to the brief retardation of the transgressive events along the platforms. The homogeneity of the data obtained regarding the distribution of sets of sequences in representative sections of different evolutionary histories suggests that eustatic fluctuations were dominant factors in the generation of these sequences in the northern and central-eastern boundary of the Iberian subplate. In the studied basins, the effect of tectonic activity does not impede recognition of higher rank sequences.

Acknowledgments

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References

ALMERAS, Y. AND S. ELMI. 1993. Palaeogeography, physiography, palaeoenvironment and brachiopod communities. Example of the Liassic brachiopods in the Western Tethys. Palaeogeography, Palaeoclimatology, Palaeoecology. 100:95-108.

ARIAS, C. F., M. J. COMAS-RENGIFO, A. GOY, C. HERRERO, AND C. RUGET. 1992. Variations dans les associations de brachiopodes, foraminifères et ostracodes du Toarcien basal dans le secteur central de le Cordillère Ibérique: un exemple dans La Rambla del Salto (Teruel, Espagne). Cahiers de l'Université Catholique de Lyon, Sér. Sci., 5:5-25.

COMAS-RENGIFO, M. J., J. J. GÓMEZ, A. GOY, C. F. ARIAS, J. BERNAD, F. GARCÍA JORAL, C. HERRERO, G. MARTÍNEZ, AND N. PERILLI. 1996. The Toarcian in the Rambla del Salto (Sierra Palomera) Section. 1st Toarcian and 4th Aalenian Working Group Meeting. Field Trip Iberian Range Guide-Book. 27-48. Department of Paleontology. Universidad Complutense de Madrid.

____, A. GOY, AND A. YEBENES. 1988. El Lias del Sector Suoccidental de la Sierra de la Demanda (Castrovido, Burgos). Ciencias de la Tierra. Geologia, 11:119-141.

ELMI, S. 1986. Corrélations biostratigraphiques et mégaséquentielles dans le Jurassique inférieur et moyen d'Orania. Comapraisons avec les région voisines. Revue de la Faculté des Sciences de Marrakech. M.S. 2, 225-247.

____, J. GABILLY, R. MOUTERDE, L. RULLEAU, AND R. B. ROCHA. 1994. L'étage Toarcian de l'Europe et de la Téthys; divisions et corrélations. Geobios, M.S. 17:149-159.

FERNÁNDEZ-LÓPEZ, S. AND L. C. SUÁREZ-VEGA. 1979. Estudio bioestratigráfico (Ammonoidea) del Aaleniense y Bajociense en Asturias. Estudios geológicos, 35:231-239.

GÓMEZ, J. J. AND S. FERNÁNDEZ-LÓPEZ. 1994. Condensation processes in shallow platforms. Sedimentary geology, 92:147-159.

____. AND A. GOY. 1979. Las unidades litoestratigráficas del Jurásico medio y superior en facies carbonatadas del Sector Levantino de la Cordillera Ibérica. Estudios geológicos, 35:569-598.

____ AND A. GOY. 1996. *In* A. Goy, G. Martínez, C. Arias, J. Bernad, F. García Joral, J. J. Gómez, C. Herrero, N. Perilli, AND S. Ureta. The Toarcian in the sector located between La Almunia de Doña Godina and Ricla: Lithostratigraphy and Sequential Stratigraphy. 1st Toarcian and 4th Aalenian Working Group Meeting. Field Trip Iberian Range Guide-Book. 3-23. Department of Paleontology. Unversidad Complutense de Madrid.

GOY, A., J. J. GÓMEZ, C. HERRERO, AND L. C. SUÁREZ-VEGA. 1997. El Toarcianse inferior en el sector comprendido entre Rodiles y Lastres (Asturias). Comunicaciones IV Congreso de Jurásico de España, p.81-82.

____, ____, AND A. YÉBENES. 1976. El Jurásico de la Rama Castellana de la Cordillera Ibérica (Mitad Norte), I. Unidades litoestratigráficas. Estudios geológicos, 32:391-423.

____ AND G. MARTÍNEZ. 1990. Biozonación del Toarciense en el área de La Almunia de Doña Godina-Ricla (Sector Central de la Cordillera Ibérica). Cuadernos de Geología Ibérica, 14:11-53.

____, ____,. AND S. URETA. 1994. El Toarciense de la región de Pozazal -Reinosa (Cordillera Cantábrica, España). Coloquios de Paleontología, 46:93-127.

HALLAM, A. 1988. A reevaluation of Jurassic eustasy in the light of new data and the revised Exxon curve, p. 71-108. *In* C. K. Wilgus, B. S. Hastings, C. G. S. C. Kendall, H. Posamentier, C. A. Ross and J. C. Van Wagoner (Editors), Sea-Level Changes: An Integrated Approach. SEMP Special Publication, 42.

JENKYNS, H. C. 1988. The Early Toarcian (Jurassic) anoxic event: stratigraphy, sedimentary and geochemical evidence. American Journal of Science, 288:101-151.

MATYJA, B. A. AND A. SEILACHER. 1985. Oyster beds: morphologic response to changing substrate conditions, p. 421-435. *In* U. Bayer and A. Seilacher (Editors), Sedimentary and Evolutionary Cycles. Springer-Verlag, Berlin.

SOARES, A. F., R. B. ROCHA, S. ELMI, M. H. HENRIQUES, R. MOUTERDE, Y. ALMERAS, C. ROUGET, J. MARQUES, L. DUARTE, M. C. CARAPITO, AND J. C. KULLBERG. 1993. Le sous-bassin nord-lusitanien (Portugal) du Trias au Jurassique moyen: histoire d'un "rift aborté". Comptes Rendus de l'Académie des Sciences (Paris). 317:1659-1666.

SUÁREZ-VEGA, L. C. 1974. Estratigrafía del Jurásico en Asturias. Cuadernos de Geología Ibérica, 3 (I&II):1-368.

VALENZUELA, M., J. C. GARCÍA-RAMOS, AND C. SUÁREZ DE CENTI. 1986. The Jurassic sedimentation in Asturias (N Spain). Trabajos de Geología, 16:121-132.

ZIEGLER, P. A. 1990. Geological Atlas of Western and Central Europe. Shell International Petroleum Maastschappij B.V., p. 91-122.

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The Distribution of Foraminiferida in the Oxfordian Sequences of North Dorset, U.K.

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Abstract: The distribution of Foraminiferida from 4 boreholes drilled in north Dorset, U.K., is described and the relationship to sequences documented. These boreholes (one drilled by the British Geological Survey (BGS) and three drilled by the University of Plymouth), together with others in the same area, were designed to intersect the complete succession from the upper part of the Oxford Clay Formation to the lowermost Kimmeridge Clay Formation. The majority of the Jurassic benthonic Foraminiferida are long-ranging and a viable zonation is difficult to establish. Using a variety of statistical and analytical techniques, all of the borehole successions have been successfully correlated and a number of palaeo-environmental changes identified that can be related to the sequence stratigraphy. These changes can be traced from north Dorset southwards across the Wessex Basin to the Dorset Coast and, possibly, the Normandy coastline.

Introduction

Sequence stratigraphy has rapidly gained a vociferous following amongst sedimentologists and hydrocarbon explorationists. Although the application of micropalaeontological data to sequence stratigraphic interpretation is commonly employed within the oil industry (see Jones, 1996) there are relatively few published accounts (Olsson, 1988; Simmons et al., 1991; Powell, 1992; Partington et al., 1993; Emery and Myers, 1996; Hart, 1997) compared with those studies employing more "traditional" (i.e. non sequence stratigraphic) micropalaeontological interpretations.

The concept of "accommodation space" (introduced by Wilgus et al., 1988) provides palaeontologists with a variation in water depth which, in most cases, is the primary control on the distribution of the preserved fauna/flora. The maximum value of "accommodation space" normally equates with the maximum depth (= Maximum Flooding Surface or MFS) at any given locality and tends to reduce the sedimentation rate in offshore marine successions. At the same time the numbers of Foraminiferida in any assemblage reportedly increase both in absolute counts and diversity (Armentrout, 1987, 1991; Armentrout and Clement, 1990; Armentrout et al., 1990). Work to date has tended to use direct counts of genera and species without any allowance being made for the size fractions being investigated or variations in sample size. In this study we have used a variety of statistical indices that negate variations in population size and sample quality.

The simple counting of foraminiferal abundances within sieved sediment samples may provide a guide as to the placement of maximum flooding surfaces, but the aim of this study is to use a variety of other measures to identify a number of ecological changes throughout the succession. These key biohorizons may then be equated with either sequence boundaries or maximum flooding surfaces. This reduces the reliance on suspect palaeoecological comparisons.

Regional Geology

Many of the type-sections for UK Jurassic stratigraphic units are exposed on the Dorset coast. Some of the early pioneers in geology studied these rocks and many of the research papers published on the area have shaped modern geological thinking (Arkell, 1933; 1956; Wilson, 1968a;

1968b; Talbot, 1973; 1974; Brookfield, 1973a; 1973b; 1978; Allen and Underhill, 1989; Sun, 1989; 1990a; 1990b; 1992; Callomon and Cope, 1995).

In contrast, the Upper Jurassic succession of north Dorset (Fig. 1) has, until recently, remained one of the least studied areas in the British Jurassic. Prior to the work of Wright (1981) the only published geological maps of the area were the original "1-inch" BGS maps for Shaftesbury and Wincanton. The mid-Upper Jurassic succession in this area is very poorly exposed but has become better known as a result of recent BGS mapping (Bristow, 1989; 1990; Freshney, 1989; Bristow et al., 1995). As part of their investigations five boreholes were drilled at locations within the mapping area: at East Stour, Cannings Court, Combe Throop, Hazelbury Bryan and Knackers Hole. While all of these provide useful stratigraphic data, the cores made available to the authors (from the East Stour and Cannings Court boreholes) penetrated the same part of the succession (Clavellata Beds Formation to Hazelbury Bryan Formation).

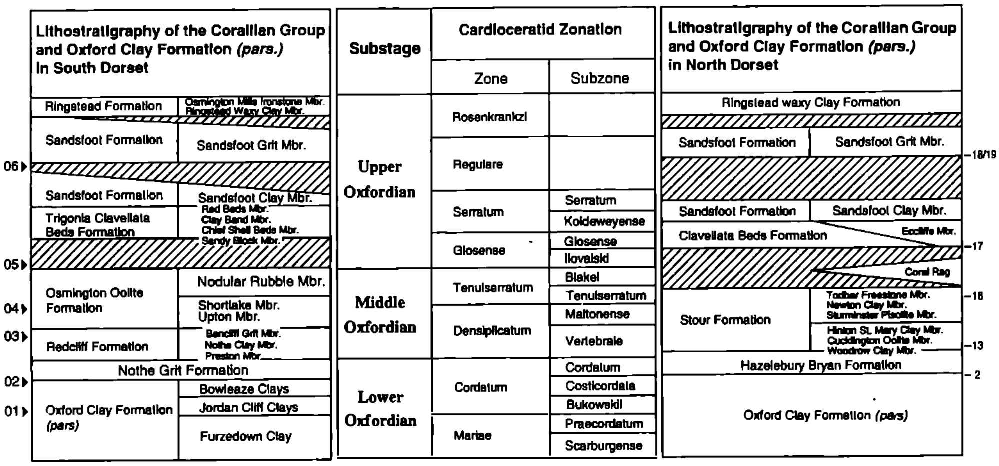

Figure 1. Chronostratigraphic subdivision of the Oxfordian Stage and a comparison of the lithostratigraphy of the Corallian Group and Oxford Clay Formation (pars.) of north and south Dorset (after Sykes and Callomon, 1979, emend. Birkelund and Callomon, 1985; Bristow et al., 1995; Coe, 1995). The unconformities (01 - 06) of Coe (1995) are shown on the left, the "bioevents" of this study are shown on the right.

In order to recover the complete succession from the upper part of the Oxford Clay Formation to the base of the Kimmeridge Clay Formation, three boreholes were drilled by the University of Plymouth in summer 1993 (Fig. 2; inset). Using the lithostratigraphy adopted by the BGS (Bristow et al., 1995 and references therein) almost all of the lithostratigraphical units can be identified from the cores (see Henderson et al., 1994; Henderson, 1997, and Fig. 1 here) and have been sampled for their microfossil content.

Microfaunal Analysis

The fresh samples recovered from all of the boreholes have yielded well preserved, diverse assemblages. Some 148 species have been identified and a monographic description of the fauna is in preparation. The fauna recorded from the north Dorset Upper Jurassic succession is fairly typical of Upper Jurassic microfaunas found north of Tethys, as described and discussed by a number of authors for the Jurassic of Britain (Barnard, 1950-63; Cordey, 1962; Lloyd, 1959; 1962; Barnard et al. 1981; Shipp, 1989; Copestake 1993).

As well as simple presence/absence information, the fauna has been subjected to a number of statistical analyses including: species richness, Margalef's Richness Index, Fisher's Alpha Index, Shannon-Wiener Information Function, Simpson's Dominance Index, Pielou's Equitability Index, cluster analysis and Non-metric Multi-Dimensional Scaling [MDS] (Henderson, 1997). Using an analysis of all these data from every borehole, a number of "events" (or biohorizons) have been identified, many of which can be correlated between the boreholes. The "events" recognised include simple taxon abundance peaks, dramatic shifts in species/generic dominance and abrupt changes in certain statistical indices. These correlations also confirm those based on both the foraminiferal zones and the known ammonite distribution. Some of the "events" mark major changes in many of the statistical measures and are, therefore, regarded as being of greater significance. Some are almost certainly coincident with sequence boundaries and MFS (see Fig. 3).

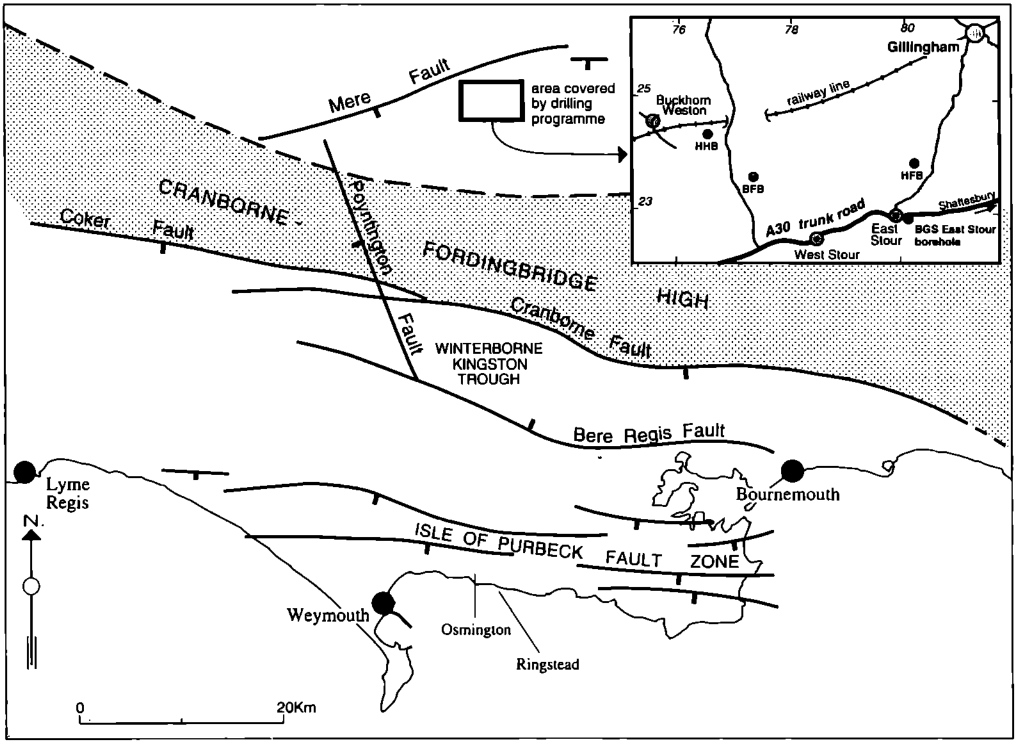

Figure 2. Principal structural features of Dorset and the location of successions discussed in the text. The borehole sites indicated are those of Hartmoor Hill (HHB), Bowden Farm (BFB), Hallett's Farm (HFB) and the BGS East Stour borehole.

The nodosariids and the vaginulinids are the most diverse families represented and also include the most abundant species. The most striking aspect of the fauna as a whole is that in many samples two species are predominant. These are the vaginulinid *Lenticulina muensteri* (Roemer) and the lituolid species *Ammobaculites coprolithiformis* (Schwager). *L. muensteri* comprises up to 70% of the fauna in many samples and is usually greater than 40%. The large variable species *A. coprolithiformis* can dominate samples in the same way, comprising up to 60% of the fauna. Other species and families are very abundant at certain horizons and tend to appear in floods. This is particularly true for some species of the epistominids, where they become the dominant form (e.g. *Epistomina parastelligera* (Hofker) can comprise 80% of the total fauna). Some of the more important epistominid floods are recorded on Figure 3. The epistominids tend to have an inverse relationship with the abundance of the lenticulinids and are most common when *L. muensteri* is least abundant. The Textulariina have a negative correlation with the Robertinina.

Other species often occur in floods and at some horizons also become the dominant taxa within the fauna. *Ammobaculites godmani* (Barnard), *Trochammina globigeriniformis* (Parker and Jones), *Trochammina squamata* Parker and Jones, *Spirillina tenuissima* (Guembel) and *Ophthalmidium strumosum* (Guembel) are examples. Each of these species can be represented by 70% of the fauna in specific samples. In one sample *T. squamata* comprises 93% of the total fauna. How these individual abundances relate to systems tracts is being investigated. Some effects may be preservational. For example, the aragonitic epistominids tend to be preserved in clay-rich sediments and may, therefore, appear to be concentrated adjacent to maximum flooding surfaces. However despite these problems certain "events" recorded within the north Dorset Borehole sequences, are correlatable with MFSs recorded by Partington et al. (1993) for the North Sea Basin. In particular, "events" 13 and 20 are probably equivalent to the J52 MFS and the J56 MFS, respectively, of Partington et al. (1993) despite the distance between basins.

The majority of Jurassic taxa recovered in this study are long-ranging and/or have variable distributions caused by facies changes. The stratigraphic distribution of the fauna has allowed identification of a number of interval zones (following the definition used in the International Stratigraphic Guide; Salvador, 1994) based on last appearance datums (LADs) and first appearance datums (FADs) of stratigraphically reliable taxa (see Fig. 3). This scheme, in general, compares favourably with the ranges of marker species given by Shipp (1989) for the Oxfordian in general, but when compared with the broader Mesozoic biozonation schemes for the Portland-Wight and North Sea Basins (Ainsworth et al., 1998; Partington et al., 1993, respectively) the present scheme is too localised. It is not intended that this be proposed as more than a local zonation, but work is in hand to test its application across the Wessex Basin. Use of the zonation has confirmed the correlation of the boreholes.

Sequence Stratigraphy

As indicated above, it is already established that changes in the abundance of taxa or the increase in diversity of a fauna can be integrated into a sequence stratigraphical framework. (Armentrout, 1987; 1991; Armentrout and Clement, 1990; Armentrout et al., 1990; Shaffer, 1987; 1990; Pacht et al., 1990; Vail and Wornardt, 1990). Much of this work is, however, based on the abundant, well-preserved, faunas of the Plio-Pleistocene succession of the Gulf Coast region of the USA.

The inclusion of FADs and LADs of sponge spicules and bivalve spat has been used in the biostratigraphy of Jurassic sediments of the North Sea (Partington et al., 1993) and Cubaynes et al. (1990) have indicated that other palaeoecological data (e.g., the types and proportions of bioclasts, the numbers of microgastropods and ammonite nuclei, the percentages of *Citharina* and uncoiling *Lenticulina* genera, the type and composition of ostracod assemblages, the amount of phytoplankton in palynofacies analyses) can be used in the recognition of different systems tracts. The importance of the latter work is that it was based on Toarcian (Jurassic) taxa from Quercy in southwest France. The work reported here takes that methodology a little further using a range of statistical analyses and then correlating them across the various successions using a form of "graphic correlation" (see Henderson, 1997, for a full account).

On the Dorset coast (the area between Weymouth, Osmington and Ringstead) even the earliest descriptions of the succession refer to the interpretation of sea-level change and/or cycles, and many authors had developed a "sequence stratigraphy" even before the terminology was developed (Wilson, 1968a; 1968b; Talbot, 1973; 1974; Sun, 1989; 1990a; 1990b; 1992; Wright, 1986). More recently Wilson (1991), Rioult et al. (1991), Coe (1992; 1995) and Oliver (1998) have identified sequence boundaries, and the various systems tracts, using the coastal sections to the east of Weymouth. While there is some agreement in the placing of many of these sequence boundaries (Oliver, 1998, fig. 2.iii), some of the interpretations do not coincide.

In our "event" stratigraphy, based solely on the microfaunal analysis of the north Dorset boreholes, we can identify (Figs 1 and 3) a number of important, or less important, biohorizons.

North Dorset	**Coe, 1995**	**Oliver, 1998**	**Partington et al., 1993**
"Event" 20	-	-	MFS J56
"Event" 18/19	06	5	
"Event" 17	?05	?4	
"Event" 16	04	3	
"Event" 13	?03	?1	MFS J52
"Event" 2	02	-	

Sequence boundary 03 of Coe (1995) is probably the equivalent of 2 (Oliver, 1998) and '149.8' of Rioult et al. (1991) and may be located between ìeventsî 13 and 14, although the placing of this sequence boundary in north Dorset is problematic. The "events" 13 and 20 are possibly the equivalents of MFS J52 and MFS J56, respectively, of Partington et al. (1993).

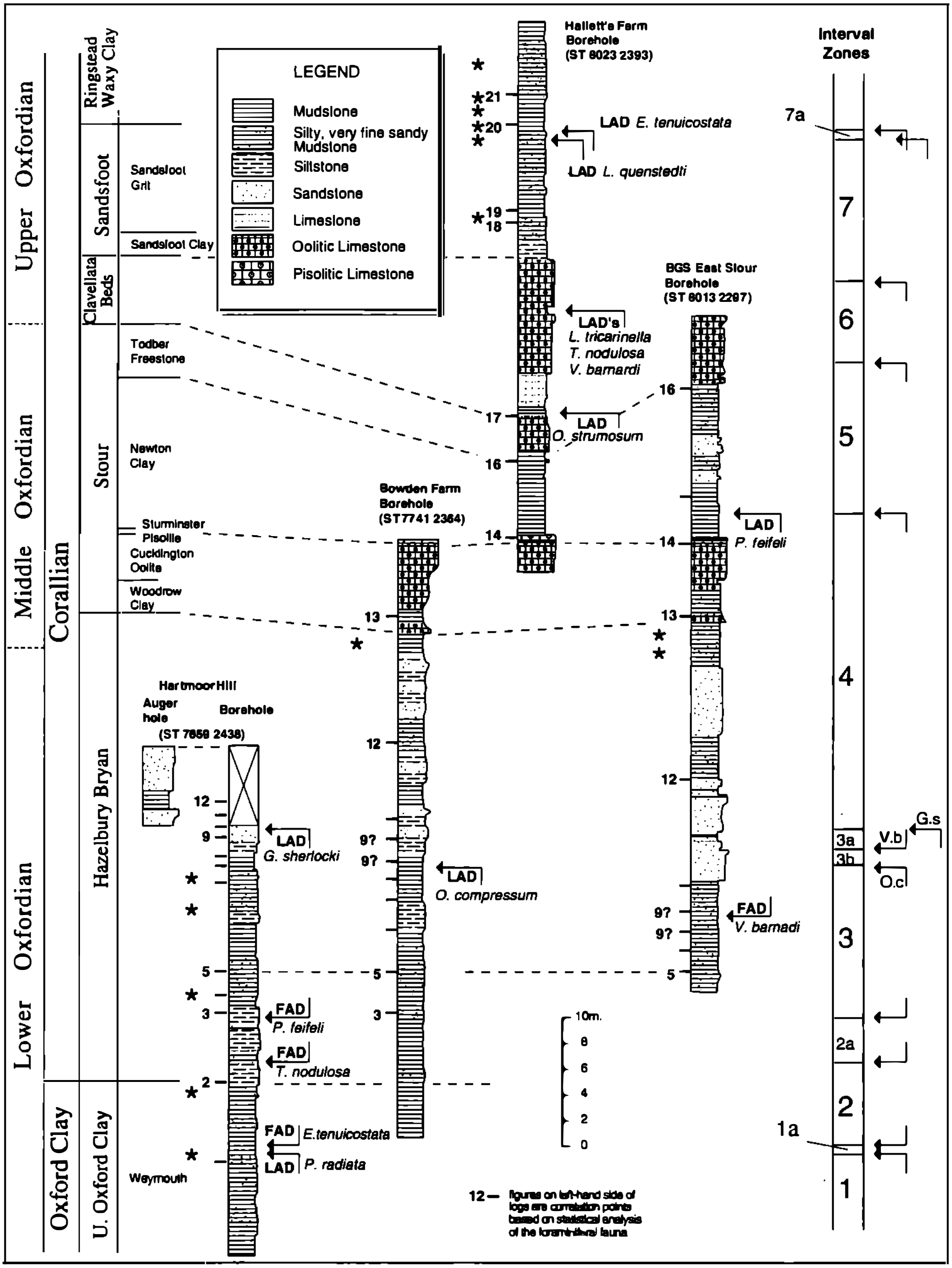

Figure 3. Correlation of the East Stour, Hallett's Farm, Bowden farm and Hartmoor Hill boreholes and the identification of locally applicable interval zones. The "events", or biohorizons, used in the sequence stratigraphical analysis are shown on the left-hand side of the schematic lithological logs and important epistominid floods are represented by asterisks.

Within the individual sequences the distribution of individual taxa (e.g., fluctuations in the percentage of *L. muensteri*) is now being investigated. This taxon is recorded in the East Stour Borehole (see Fig. 3) in some abundance around "event" 5 and again near "event" 9. If "event" 2 is a sequence boundary, as suggested by Coe (1992, 1995), then these levels of *L. muensteri* might equate with the MFS. The distribution of the taxon in the other boreholes indicates that abundance may equate with zones of maximum flooding and absence indicate a proximity to sequence boundaries.

Summary

Foraminiferal biostratigraphy, together with lithostratigraphy, has been used to provide an effective correlation of a number of boreholes from north Dorset.

A range of statistical techniques has also been applied to the samples and an "event" stratigraphy generated. This provides a potential tool for the identification of sequence boundaries and MFSs within the succession, as well as - in many cases - an indication of other palaeoecological changes and cyclical events (some of which may prove to be located within the Milankovitch waveband). Certain "events" established in north Dorset may be equivalent to unconformities previously recognized in Dorset Coast sediments, and possibly equivalent to MFS recognised in the North Sea Basin. The methodology employed (fully outlined in Henderson, 1997) clearly has potential and research on the subject is continuing.

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References

AINSWORTH, N. R., W. BRAHAM, F. J. GREGORY, B. JOHNSON, AND C. KING. 1998. A proposed latest Triassic to earliest Cretaceous microfossil biozonation for the English Channel and its adjacent areas, p. 87-1022. *In* J. R. Underhill (ed.), Development, Evolution and Petroleum Geology of the Wessex Basin. Geological Society, London, Special Publications, 133.

ALLEN, P. A. AND J. R. UNDERHILL. 1989. Swaley cross-stratification produced by unidirectional flows, Bencliff Grit (Upper Jurassic), Dorset, UK. Journal of the Geological Society, 146:241-252.

ARKELL, W. J. (ed.). 1933. The Jurassic System in Britain. Oxford University Press, Oxford.

____. 1956. Jurassic geology of the World. Oliver and Boyd, London.

ARMENTROUT, J. M. 1987. Integration of biostratigraphy and siesmic stratigraphy: Pliocene-Pleistocene, Gulf of Mexico. p. 6-14. *In* Gulf Coast Section, Society of Economic Paleontologists and Mineralogists, Eighth Annual Research Conference.

____. 1991. Paleontologic constraints on depositional modelling; examples of integration of biostratigraphy and seismic stratigraphy, Gulf of Mexico, p. 137-170. *In* P. Weimer and M. H. Link (eds.), Seismic facies and sedimentary processes of submarine fans and turbidite systems. Springer-Verlag.

____, AND J. F. CLEMENT. 1990. Biostratigraphic correlation of depositional cycles: a case study in High Island-Galveston-East Breaks areas, offshore Texas. p. 21-51. *In* Gulf Coast Section, Society of Economic Paleontologists and Mineralogists, Eleventh Annual Research Conference.

____, R. C. ECHOLS, AND T. D. LEE. 1990. Patterns of foraminiferal abundance and diversity: implications for sequence stratigraphic analysis, p. 53-58. *In* Gulf Coast Section, Society of Economic Paleontologists and Mineralogists, Eleventh Annual Research Conference.

BARNARD, T. 1950a. The uses of foraminifera in Lower Jurassic stratigraphy, p. 34-41. *In* The International Geological Congress, 18th session, London (1948), Report., pt. 15 (Proceedings of the International Palaeontological Union).

____. 1950b. Foraminifera from the Lower Lias of the Dorset Coast. Quarterly Journal of the Geological Society of London, 105(3):347-391.

___. 1950c. Foraminifera from the Upper Lias of Byfield, Northamptonshire. Quarterly Journal of the Geological Society of London, 106(1):1-36.

___. 1952. Foraminifera from the Upper Oxford Clay (Jurassic) of Warboys, Huntingdon shire. Proceedings of the Geologists Association, 63(4):336-350.

___. 1953. Foraminifera from the Upper Oxford Clay (Jurassic) of Redcliff Point, near Weymouth. Proceedings of the Geologists Association, 64:183-197.

___. 1956. Some Lingulinae from the Lias of England. Micropaleontology, 2(3):271-282.

___. 1957. Frondicularia from the Lower Lias of England. Micropaleontology, 3(2):171-181.

___. 1958. Some Mesozoic adherent Foraminifera. Palaeontology, 1(2):116-124.

___. 1963. Evolution in certain biocharacters of selected Jurassic Lagenidae, p. 79-92. *In* Koenigswald, G. H. R. von, et al., (eds), Evolutionary Trends in Foraminifera. Elsevier, Amsterdam.

___, W. G. CORDEY, AND D. J. SHIPP. 1981. Foraminifera from the Oxford Clay (Callovian-Oxfordian of England). Revista Espanola de Micropaleontologia, 13(3):383-462.

BIRKELUND, T. AND J. H. CALLOMON. 1985. The Kimmeridgian ammonite fauna of Milne Land, Central East Greenland. Gronlands Geologiske Undersogelse, Bulletin 153.

BRISTOW, C. R. 1989. Geology of the East Stour - Shaftesbury District (Dorset). British Geological Survey Technical Report, WA/89/58.

___. 1990. Geology of the West Stour - Cucklington district. Geological Survey Technical Report, WA/90/48

___, C. M. BARTON, E. C. FRESHNEY, C. J. WOOD, D. J. EVANS, B. M. COX, H. C. IVIMEY-COOK, AND R. T. TAYLOR. 1995. Geology of the Country around Shaftesbury. Memoir of the British Geological Survey, Sheet 313 (England and Wales), London, HMSO, 182p.

BROOKFIELD, M. E. 1973a. Palaeogeography of the Upper Oxfordian and Lower Kimmeridgian (Jurassic) in Britain. Palaeogeography, Palaeoclimatology, Palaeoecology, 14:137-167.

___. 1973b. The palaeoenvironment of the Abbotsbury Ironstone (Upper Jurassic) of Dorset. Palaeontology, 16:261-274.

___. 1978. The lithostratigraphy of the upper Oxfordian and lower Kimmeridgian Beds of South Dorset. Proceedings of the Geologists' Association, 89:1-32.

CALLOMON, J. H. AND J. C. W. COPE. 1995. The Jurassic Geology of Dorset, p. 51-103. *In* P. D. TAYLOR (ed.), Field Geology of the British Jurassic, Geological Society of London.

COE, A. L. 1992. The Oxfordian and Portlandian, p. 37-81. *In* A. L. Coe, S. P. Hesselbo, D. N. Parkinson, and H. C. Jenkyns (eds.), Sequence Stratigraphy of the Dorset Jurassic, a field guide. Sequence Stratigraphy of European Basins, CNRS-IFP, Dijon, France.

___. 1995. A comparison of the Oxfordian successions of Dorset, Oxfordshire and Yorkshire. p. 151-172. *In* P. D. Taylor (ed.), Field Geology of the British Jurassic. Geological Society of London.

COPESTAKE, P. 1993. Application of micropalaeontology to hydrocarbon exploration in the North Sea Basin, p. 93-152. *In* D. G. Jenkins (ed.), Applied Micropalaeontology, Kluwer Academic Publishers, Netherlands.

CORDEY, W. G. 1962. Foraminifera from the Oxford Clay of Staffin Bay, Isle of Skye, Scotland. Senck. leth., 43(5):375-409.

CUBAYNES, R., J. REY, C. RUGET, B. COURTINA, AND A. M. BODEGAT. 1990. Relations between systems tracts and micropalaeontological assemblages on a Toarcian carbonate shelf (Quercy, southwest France). Bulletin, Societe Geologique de France, 8: 989-993.

EMERY, D. and K. J. MYERS. 1996. Sequence Stratigraphy, Blackwell Science, Oxford, 297p.

FRESHNEY, E. C. 1989. Geology of Sheets ST 70NW and NE (Hazelbury Bryan - Mappowder district, Dorset). Technical Report of the British Geological Survey Onshore geology series, WA/90/51, 1-58.

HART, M. B. 1997. The application of micropalaeontology to sequence stratigraphy; an example from the chalk of south-west England. Proceedings of the Ussher Society, 9:158-163.

HENDERSON, A. S. 1997. The palaeoecology and biostratigraphy of the Foraminifera from the Oxfordian of north Dorset. Unpublished Ph.D. Thesis, University of Plymouth.

___, G. M. OLIVER, AND M. B. HART. 1994. The lithostratigraphy of the mid-Upper Jurassic of north Dorset; preliminary results from three new boreholes. Proceedings of the Ussher Society, 8:325-327.

JONES, R. W. 1996. Micropalaeontology in Petroleum Exploration, Oxford University Press Inc., New York, 432p.

LLOYD, A. J. 1959. Arenaceous foraminifera from the type Kimmeridgian (Upper Jurassic). Palaeontology, 1(4):298-320.

____. 1962. Polymorphinid, miliolid and rotaliform foraminifera from the type Kimmeridgian. Micropaleontology, 8(3):369-383.

OLIVER, G. M. 1998. High-resolution sequence stratigraphy and diagenesis of mixed carbonate/siliciclastic successions. Unpublished Ph.D. Thesis, University of Plymouth.

OLSSON, R. K. 1988. Foraminiferal modelling of sea-level change in the Late Cretaceous of New Jersey, p. 289-298. *In* C. K. Wilgus, et al. (eds.), Sea-level changes: an integrated approach. Special Publication of the Society of Economic Paleontologists and Mineralogists, 42.

PACHT, J. A., B. E. BOWEN, J. H. BEARD, AND B. L. SHAFFER. 1990. Sequence stratigraphy of Plio-Pleistocene depositional facies in the offshore Louisiana south additions. Gulf Coast Association of Geological Societies Transactions, 40:1-18.

PARTINGTON, M. A., P. COPESTAKE, B. C. MITCHENER, AND J. R. UNDERHILL. 1993. Biostratigraphic calibration of genetic sequences in the Jurassic-lowermost Cretaceous (Hettangian to Ryazanian) of the North Sea and adjacent areas, p. 371-386. *In* J. R. Parker (ed.) Petroleum Geology of Northwest Europe. Proceedings of the 4th Conference, Geological Society, London, 1.

POWELL, A. J. 1992. Making the most of microfossils. Geoscientist, 2:12-16.

RIOULT, M., O. DUGUE, R. JAN DU CHENE, C. PONSOT, G. FILY, J. M. MORON, AND P. R. VAIL. 1991. Outcrop sequence stratigraphy of the Anglo-Paris Basin, Middle to Upper Jurassic (Normandy, Maine, Dorset).Bull. Centres Rech. Explor. -Prod. Elf Aquitaine, 15: 101-194.

SALVADOR, A. 1994. International Stratigraphic Guide: a guide to stratigraphic classfication, terminology and procedure. IUGS/Geological Society of America Inc., USA.

SHAFFER, B. L. 1987. The potential of calcareous nannofossils for recognising Plio-Pleistocene climatic cycles and sequence boundaries on the shelf. Gulf Coast Section of the Society of Economic Paleontologists and Mineralogists, Eighth Annual Research Conference, 142-145.

____. 1990. The nature and significance of condensed sections in Gulf Coast late Neogene stratigraphy. Gulf Coast Association of Geological Societies Transactions, 40:186-195.

SHIPP, 1989. The Oxfordian to Portlandian. *In* D. G. Jenkins, and J. W. Murray (eds.), Stratigraphical Atlas of Fossil Foraminifera. Ellis Horwood, Ltd. Chichester.

SIMMONS, M. D., C. L. WILLIAMS, AND M. B. HART. 1991. Sea level changes across the Albian-Cenomanian boundary in south-west England. Proceedings of the Ussher Society, 7:408-412.

SUN, S. Q. 1989. A new interpretation of the Corallian (Upper Jurassic) cycles of the Dorset coast, Southern England. Geological Journal, 24:139-158.

____. 1990a. Discussion of swaley cross-stratification produced by unidirectional flows, Bencliff Grit (Upper Jurassic), Dorset, UK. Journal of the Geological Society, 147:396-400.

____. 1990b. Facies-related diagensis in a cyclic shallow marine sequence, the Corallian group (Upper Jurassic) of the Dorset Coast, Southern England. Journal of Sedimentary Petrology, 60:42-52.

____. 1992. A storm-dominated offshore sandstone interval from the Corallian group (Upper Jurassic), Weald Basin, southern England. Marine and Petroleum Geology, 9:274-286.

SYKES, R. M. AND J. H. CALLOMON. 1979. The Amoeboceras zonation of the Boreal Upper Oxfordian. Palaeontology, 22:839-903.

TALBOT, M. R. 1973. Major sedimentary cycles in the Corallian beds (Oxfordian) of southern England. Palaeogeography, Palaeoclimatology, Palaeoecology, 14:293-317.

____. 1974. Ironstones in the Upper Oxfordian of southern England. Sedimentology, 21:433-450.

VAIL, P. R. AND W. W. WORNADT. 1990. Well-log seismic stratigraphy: an integrated tool for the 1990's. p. 379-397. *In* Sequence Stratigraphy as an Exploration Tool: concepts and practices in the Gulf Coast, Gulf Coast Section of the Society of Economic Paleontologists and Mineralogists, 11th annual research conference program and abstracts.

WILGUS, C. K., B. S. HASTINGS, G. St. C. KENDALL, H. W. POSAMENTIER, C. A. ROSS. AND J. C. VAN WAGONER (eds.). 1988. Sea-level changes: an integrated approach. Special Publication of the Society of Economic Paleontologists and Mineralogists, 42: 407pp.

WILSON, R. C. L. 1991. Sequence stratigraphy, an Introduction. Geoscientist, 1:13-23.

____. 1968a. Carbonate facies variation within the Osmington Oolite Series in Southern England. Palaeogeography, Palaeoclimatology, Palaeecology, 4:89-123.

____. 1968b. Upper Oxfordian palaeogeography of Southern England. Palaeogeography, Palaeoclimatology, Palaeoecology, 4:5-28.

WRIGHT, J. K. 1981. The Corallian rocks of north Dorset. Proceedings of the Geologists' Association, 92:17-32.

____. 1986. A new look at the stratigraphy, sedimentology and ammonite fauna of the Corallian Group (Oxfordian) of south Dorset. Proceedings of the Geologists' Association, 97:1-21.

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Transgressive Sediment Intervals in the Late Jurassic of Kachchh, India

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Keywords: Ammonoid, Late Jurassic, Eustatics, Transgression, Kachchh

Abstract: The splendidly exposed ammonoid rich ca 400m Late Jurassic sedimentary succession of Kachchh is examined for 3rd order transgressive intervals and associated maximum flooding signatures in several sections. The Oxfordian is made up of highly condensed to starved, ammonoid-rich nodular conglomeratic oolitic limestones forming hard grounds which, towards the margin, change to over 200m thick sand and subordinate shale. This proximal succession is invariably ammonoid devoid/scarce except for a 10m ammonoid abundant interval. In Kimmeridgian – Tithonian there are several cycles of thin ammonoid-bearing calcareous sands separated by much thicker ammonoid devoid shale/silt and sand. The thin, ammonoid-bearing beds make 3rd to 5th order transgressive intervals under low energy, slow sedimentation, below wave base conditions.

The Kachchh Jurassic is part of the 2nd order Late Bajocian – Late Albian sequence. The 2nd order maximum flooding is interpreted in the ammonoid rich Middle Oxfordian Schilli Subzone since it maximizes 1. margin-ward spatial extent of the ammonoid bearing beds, 2. upward increase in ammonoid density, 3. upward increase in the proportion of North Tethyan ammonoids, 4. upward increase of lyto-phyllo-sphaeroceratid ammonoids, 5. margin-ward increase of fine clastics, 6. gradual decrease of sedimentation rate upward. Most importantly, the Schilli Subzone is the only ammonoid rich interval in the proximal Wagad succession. The significant 3rd order transgressive intervals are: 1. condensed ammonoid rich oolitic limestone in the Kachchh Mainland with *Mayaites*, *Peltoceratoides* etc. of Cordatum Zone which changes margin-ward to normal ammonoid-devoid/scarce sandstone, 2. ammonoid rich, slow sedimented yet paleontologically uncondensed ironstones at Kantkote, proximal to the margin of Schilli Subzone which correspond to part of the submarine non-depositional hiatus in the Kachchh Mainland, 3. thin sand beds with multiple levels of scarce ammonoids (mostly *Orthosphinctes* and *Lithacosphinctes*) in between the intervening ammonoid devoid/scarce thicker sands of late Late Oxfordian, which correspond to the still continuing submarine non-depositional hiatus in Kachchh Mainland, 4. numerous thin ammonoid rich beds with *Aspidoceras*, *Hybonoticeras*, *Taramelliceras* etc. of early Beckeri Zone in Kachchh Mainland sections, 5. ammonoid-bearing thin sand beds with exclusive Indo - East-African *Katroliceras*, *Indodichotomoceras* and *Aulacosphinctoides* so far known only in the Ler – Walakhawas belt of the Kachchh Mainland of Early Tithonian and 6. glauconitic-oolitic marls with *Virgatosphinctes*, *Microcanthoceras* and *Aulacosphinctes* of Microcanthum Zone.

There is clear disagreement between the Kachchh 2nd order sequences and the European Vail models, both with regard to maximum transgression and sequence boundaries. This is here attributed to the overriding influence of vertical tectonics in the India – Africa divergent region. Yet near concordance in the ages of the 3rd order transgressive intervals between the two regions is inferred, presumably indicating eustatic influence.

Introduction

For over a century and half the Kachchh basin of India has been famous for its ammonoid-rich, late Middle to late Late Jurassic sedimentary succession. It is located in the extreme west of India in the prime hydrocarbon belt of the Indian subcontinent (Fig. 1). We have in recent years developed high resolution ammonoid zonation schemes in the Callovian, Kimmeridgian and Tithonian stages (Krishna and Ojha, 1996; Krishna et al., 1996a; 1996b). Studies have also progressed in the Oxfordian (Krishna et al., 1995; 1998) however, these are still long way from producing a formal ammonoid zonation. The Jurassic of Kachchh is also endowed with near continuous sedimentation except for the one large duration, post-Dhosa Oolite, non-depositional gap (particularly evident in the relatively distal West Mainland Kachchh) (Krishna et al., 1998). In context of the hydrocarbon related exploratory geological studies in the basin, we have endeavoured to initiate sequence stratigraphic studies based only on outcrops. The present work discusses the significant 3rd order Late Jurassic transgressive sediment intervals in this famous basin.

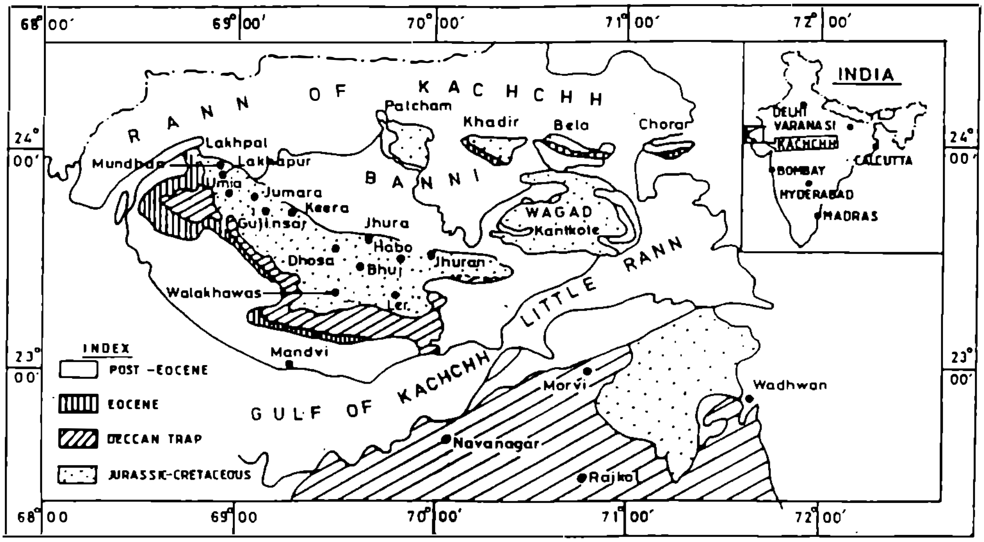

Figure 1. Simplified geological map of Kachchh with principal Jurassic localities.

Basin Geometry

The east-west trending Kachchh basin originated near the Triassic/Jurassic boundary (Krishna, 1987) as part of the rifting related to the dismemberment history of the Gondwana continent. The origin of the basin is influenced by the major east-west trending Narmada-Sone-Damodar lineament and similarly striking Delhi fold belt to the south. The exposed part of the basin from Wagad in the east (proximal most exposed Jurassic) to Mundhan in the west is about 160km while the maximum width is about 150km. The basin is bounded by the Nagar Parker ridge in the north and the Kathiawar uplift in the south. In the east it is bordered by the Radhanpur arch with a west-southwesterly paleoslope and paleocurrent framework. Geophysical studies have also revealed a NNE-SSW striking 'Median High' having originated sometime in the Oxfordian. The Upper Jurassic sediments are presently exposed in six major uplifts which are separated by the Banni plains (Biswas, 1980).

Parameters Considered

Apart from the usual criteria of temporal and spatial variation in lithological characteristics, we have based our recognition of relative bathymetry and transgressive sediment intervals on the quantitative, qualitative and compositional variation of the ammonoids, both stratigraphically and laterally. We have considered the ammonoid density, the proportion in the assemblage of the north Tethyan ammonoids, and the presence of lyto-phyllo-sphaeroceratid ammonoids. Among the sedimentological parameters we have noted the presence and migration of oolitic and glauconitic sediments; condensed and sediment starved intervals; hard grounds; physical and biogenic structures.

3rd Order Transgressive Intervals

Oxfordian

In the Kachchh Mainland, the Oxfordian is encountered in all examined sections from Jawaharnagar in the east to Lakhapur in the west. It is also present in Wagad and Khadir and in between, north-east and north of the Kachchh Mainland. In the West Mainland Kachchh, the Oxfordian is found highly condensed with a thickness up to a maximum of 8m in Jumara and Lakhapur sections in ammonoid-rich limestones that are nodular, bioclastic, conglomeratic, and oolitic. There is a maximum of 7 thin oolitic limestone bands alternating with thin shales (Krishna et al., 1995; 1998) (Figs. 2, 5). A few meters of non-gypsiferous shales lying in between the oolitic limestones above and gypsiferous shales below are also included in the Oxfordian. There is a significant stratigraphic gap above the oolitic limestones almost all over the Kachchh Mainland ranging in age from mid-Middle Oxfordian (Plicatilis Zone) to early Early Kimmeridgian (Hypselocyclum Zone) (Krishna et al., 1996b). However, this gap is not present in the most proximal exposed part of the basin in Wagad (Krishna et al., 1995; 1998) east of the Kachchh Mainland. The overlying sediments at Wagad are almost devoid of ammonoids but for a few thin intervals, among which two are especially noteworthy; one in the early part and the other in the uppermost part (Fig. 2). The older, ca. 10m thick interval (Figs. 2, 5) includes an ammonoid-rich 13 bed/band ironstone succession of Schilli Subzone age at Jadawas north of Kantkote (Krishna et al., 1995; 1998). Near the top of the Jurassic succession in Wagad at Bharodia there is another ammonoid-bearing sedimentary interval of Planula Zone – Platynota Zone age although relatively poorer in ammonoids (Krishna et al., 1995; 1998).

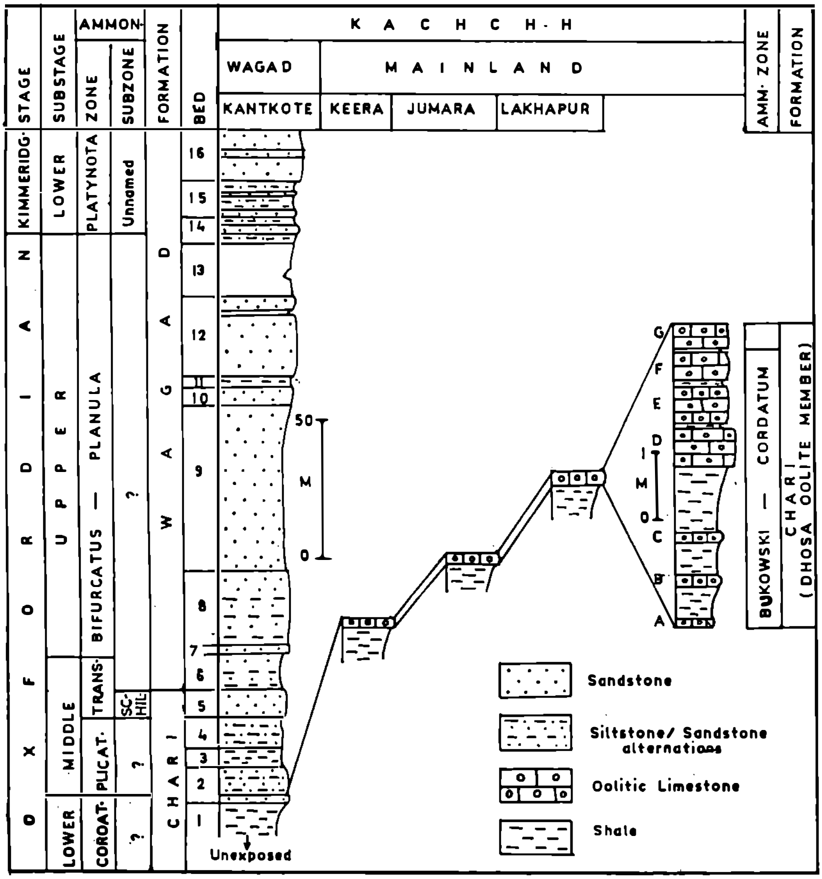

Figure 2. Ammonoid based correlation in the Oxfordian of Kachchh.

Transgressive interval I: The first significant transgressive interval is interpreted in the extremely slowly sedimented, condensed to starved, oolitic limestone with rich *Peltoceratoides*, *Mayaites*, *Otosphinctes*, *Kranaosphinctes*, *Euaspidoceras* etc., correlated to the Cordatum Subzone of the Cordatum Zone (Krishna et al., 1995; 1998; Spath, 1927-33) (Fig. 3). The tentative correlation is also supported by north or southward expansion on the two margins of the Tethys of similar *Peltoceratoides* which is presumably influenced by eustatic rise.

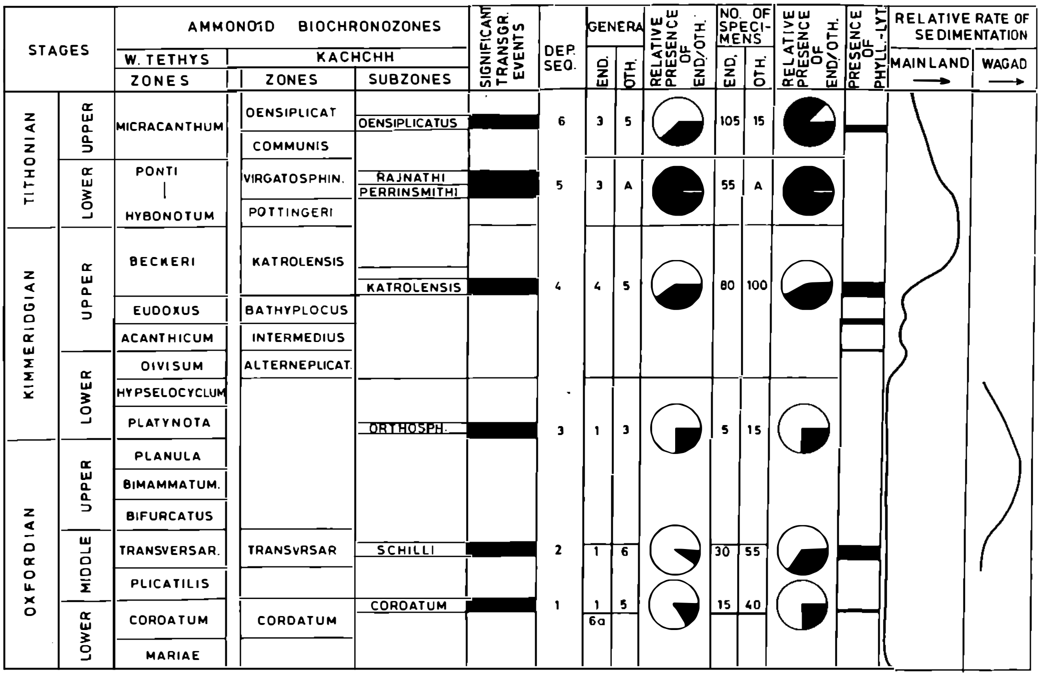

Figure 3. Significant transgressive events through quantitative and compositional variation of ammonoids and relative rate of sedimentation in the Late Jurassic of Kachchh.

The condensation/starvation is interpreted to be caused by sudden significant tectonic-cum eustatic rise in basinal depth leading to eastward shifting of the depositional site towards the proximal side, near Wagad (Krishna et al., 1998). With such a shift, sediments are not able to reach most of the Kachchh Mainland. In terms of generic composition, the Tethyan ammonoid genera (*Peltoceratoides*, *Kranaosphinctes*, *Otosphinctes*, *Campylites*, *Euaspidoceras* etc.) outnumber the Indo – East-African *Mayaites* (inclusive of *Epimayaites*) by almost 5:1 (Fig. 3). Even in terms of abundance, the widely distributed Tethyan ammonoid individuals outnumber the geographically restricted ammonoids by nearly 2:1. In comparison to the preceding Late Callovian depositional sequence, there is significant increase in the number of Tethyan ammonoid individuals in view of increased ammonoid expansion to Kachchh from the Mediterranean and Submediterrnean faunal provinces of Europe obviously having been facilitated by eustatic rise. Extensional tectonics in the vicinity between India and Africa inferred from the presence of intraformational conglomerates (Fürsich et al., 1992) additionally contributed to the net rise in bathymetry. The sediment texture of the oolitic limestones is fine and sedimentation rate is extremely slow. The spatial expansion of these condensed slowly sedimented, ammonoid-rich, oolitic limestones is observed proximally up to Habo and Jhuran with gradually decreasing oolitic content. The density and diversity of ammonoids and the number and extent of the oolitic beds is relatively more than in the proximal, most ammonoid-rich Late Callovian beds of the earlier cycles. For example, ammonoids are scarce to absent below the Middle Oxfordian Schilli Subzone interval in Wagad and Khadir. Ammonoid morphology (Ziegler, 1967; Westermann, 1990; Marchand, 1982) in this transgressive interval suggests a broad estimate of more than 90m basinal depth in the West Mainland localities of Lakhapur and Jumara.

Transgressive interval II: The second significant transgressive interval in the Late Jurassic of the Kachchh basin is in Wagad in the neighbourhood of Jadawas (north of Kantkote) represented by a 10m thick, 13 bed/band ironstone hardground succession (Fig. 5) (Krishna et al., 1995; 1998). Here the individual hardened beds are almost stacked one over the other without any appreciable thickness of alternating shales in between. This interval is also somewhat slow sedimented, however, without any paleontological condensation. The ammonoid abundance/density in this transgressive interval is the maximum of the Kachchh Mesozoic or for that matter of the entire Indian Mesozoic. This interval is located in the proximal-most exposed part of the basin. Further, the rest of the sediment column below and above the ammonoid abundant interval is either ammonoid devoid or ammonoid scarce. Tethyan ammonoids, both number of genera and individuals, heavily outnumber the endemic ones (Fig. 3). The size of the ammonoids in this interval is the maximum of the Kachchh Jurassic. The phylloceratids and lytoceratids are also present in good number which is again the maximum of the Kachchh Jurassic. It may be noted that the *Lytoceras* collected from this interval include an example of an incomplete phragmocone with a diameter approaching 1m. Lithologically the intervening shale thickness in between the successive hardened ammonoid beds is much less than in the previous sequence (Figs. 2, 5). Among the Tethyan genera we have recorded the first *Gregoryceras* and the first *Larcheria* in the Indian subcontinent (Krishna et al., 1995; 1998). This interval also records the farthest proximal presence of ammonoids in the Washtawa locality of Wagad somewhat little more proximal to Jadawas. The above observations suggest maximum basinal depth in the Kachchh basin, the maximum flooding/transgression during the Middle Oxfordian Transversarium Zone Schilli Subzone interval, also the maximum transgression/flooding of the entire 2nd order Kachchh Mesozoic depositional sequence. During the time of this transgressive interval, most or all of the Mainland Kachchh (e.g., Lakhapur, Jumara, Keera etc.) include a significant submarine non-depositional gap. Based on ammonoid morphology in this interval, we suggest basinal depth of 80 to 100m in the neighbourhood of Jadawas (Ziegler, 1967; Westermann, 1990; Marchand, 1982). Obviously the basinal depth at this time at Jumara and Lakhapur localities of West Mainland should be well over 100m.

Oxfordian/Kimmeridgian Transition

Transgressive interval III: There is another ammonoid bearing sediment interval in Wagad at Bharodia (Krishna et al., 1995; 1998). The Wagad Jurassic sedimentary succession in to our opinion is exclusively Oxfordian ranging from early Early Oxfordian to the Oxfordian/Kimmeridgian transition. This ammonoid-bearing interval overlies a nearly ammonoid devoid sand/shale alternation with increase in sand content upward. According to our tentative determination the poorly preserved fragments include *Orthosphinctes* and *Ataxioceras/Lithacoceras/Lithacosphinctes*- a low diversity exclusively Perisphinctinae fauna. The relatively deeper-water, endemic Mayaitinae have almost completely disappeared along with the Tethyan Peltoceratinae, Aspidoceratinae, Taramelliceratinae and also belemnoids which were so rich in the previously discussed transgressive interval. The bivalves (trigonids etc.) are present although different from the previous transgressive intervals. Texturally, the sediments are distinctly much coarser gritty sands in comparison to those of the previous transgessive interval. The Bharodia Perisphinctinae in comparison, are found close to North Tethyan region, yet it is difficult to assign them to particular North Tethyan species. The above indicates that the basinal depth had distinctly decreased due to onset of 2nd order regression after Schilli Subzone maximum flooding.

Some specimens with bifurcating ribs along with frequent single ribs suggest phylogenetic connection to *Torquatisphinctes* which are found abundantly present in the late Early Kimmeridgian Alterneplicatus Zone in the Kachchh Mainland. The meager, exclusively Perisphinctinae assemblage is best assigned to Late Oxfordian Planula Zone with suspected extension to the basal-most Kimmeridgian. It seems that in spite of a distinct shallowing of the proximal part of the basin exposed in Wagad, the Kachchh Mainland remained in deeper water, still unable to receive sediments. The submarine non-depositional gap (Krishna et al., 1995; 1998) in the Kachchh Mainland from Early or mid-Middle Oxfordian continued through the rest of the Oxfordian into Early Kimmeridgian. The 2nd order maximum flooding in the Schilli Subzone differs with that of

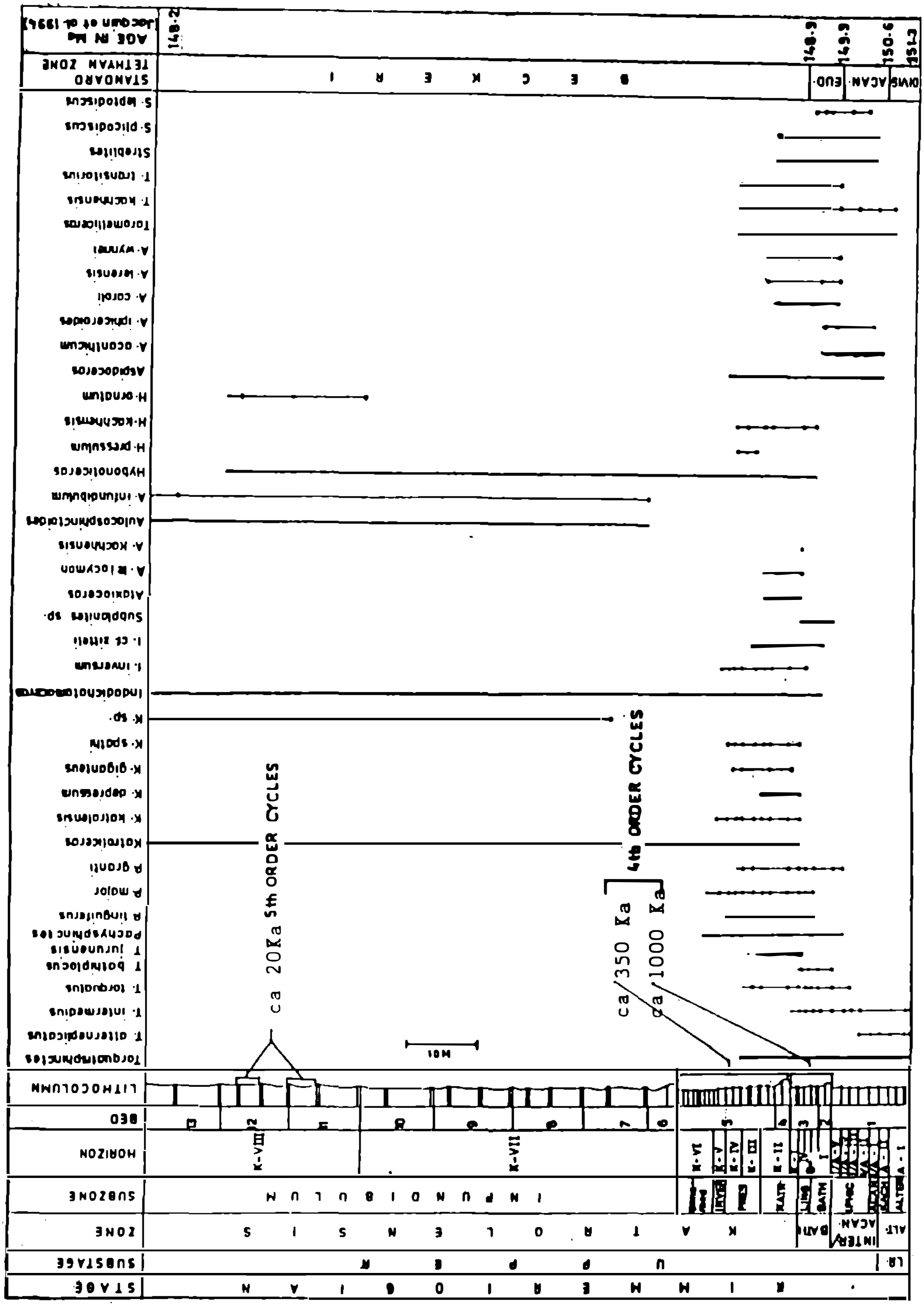

Figure 4. Ammonoid stratigraphic ranges, zonal scheme, and 4th to 5th order depositional sequences with their approximated durations in the Kimmeridgian of south-west Ler, Kachchh.

the Vail European models. In Europe, the 2nd order maximum transgression is farther up either in the Eudoxus - Early Beckeri interval or the Hybonotum - Early Darwini interval (Gygi et al., 1998; Haq et al., 1987; Jacquin et al., 1994). We interpret that the 2nd order regression in Kachchh above the Schilli Subzone is tectonically induced. The tectonic uplift may be related to the origin of the 'Median High' in the basin and seems to have outweighed the eustatic rise. This transgressive interval suggests much reduced basinal depth of the order of 15-30m at Bharodia. The 100% Tethyan character of the ammonoid generic assemblage suggests that ammonoid expansion was maintained from the European Mediterranean and Submediterranean provinces.

Kimmeridgian

The development of the Kimmeridgian Stage in the Kachchh basin is almost restricted to the Kachchh Mainland except for the suspected basal Kimmeridgian in Wagad as discussed above. Even in the Kachchh Mainland, only the post-Hypselocyclum Zone part of the Kimmeridgian is present. In view of the presence of a substantive submarine non-depositional gap almost all over the Kachchh Mainland, the Kimmeridgian disconformably overlies the late Early to early Middle Oxfordian (oolitic limestone) while it is conformably overlain by the Tithonian. The thickness varies between 100 and 150m in different sections. A total of five sections has been examined which from the margin to basin are East Ler and West Ler, Walakhawas, Lakhapur and Jawaharnagar. In general the Kimmeridgian succession comprises alternations of hard, highly fossiliferous, fine to medium grained sandstone and gypsiferous to sandy shale with occasional concretions. The best development is at Ler where ammonoids have been collected from about 40 levels. Ammonoid biostratigraphic studies have resulted in a high resolution ammonoid zonation scheme of 4 zones, 11 subzones and 19 horizons (Krishna et al., 1996b).

Transgressive interval IV: It is clear from the Kimmeridgian ammonoid range chart (Fig. 4) that the maximum ammonoid presence is observed in the Bathyplocus Zone – early Katrolensis Zone interval which has been correlated to the Eudoxus Zone – early Beckeri Zone interval of Europe (Krishna et al., 1996b). Most of the Tethyan elements are restricted to this interval. The geographically restricted Indo-East-African genera (subfamily Torquatisphinctinae) are more abundant than the Tethyan elements. Comparison with the preceding transgressive interval reveals that the proportion of Tethyan genera and individuals has decreased. However, several Tethyan genera are present in view of good ammonoid expansion from the European Tethyan margin. The correlation of the Kachchh Kimmeridgian zones with the European Tethyan scheme has been further strengthened through comparison of the *Hybonoticeras* lineage of Kachchh and Europe (Schweigert et al., 1996).

This transgressive interval has a lesser margin-ward spread than the previous one. It is followed upwards by a relatively rapidly deposited, more sandy, regressive succession which is almost devoid to extremely poor in ammonoids. However, rare presence of *Hybonoticeras ornatum* Spath provides good correlation and dating with the Ornatum Horizon of South-West Germany. In this regressive sediment interval, 4th, 5th and even 6th order depositional sequences are visually differentiable (Fig. 4). The 4th order sequences are tentatively estimated to be of the order of ca. 350,000 to 1,000,000 years. 5th order sequences in the preceding transgressive interval are determined to be of the order of ca 100,000 to 250,000 years. The finest cycles visually identified in the principal transgressive interval of the Kimmeridgian are of 5th order, while in the following regressive interval (late Katrolensis Zone) even the 6th order sequences are differentiable in the sections. It is interesting to note that each 5th order sequence in the regressive sediment interval of ca 20,000 years is found to include a package of 10 to 12 6th order sequences. As such these finest 6th order Late Katrolensis Zone sequences are crudely estimated to be of the order of ca 2000 years. These may probably be among the finest and smallest visually recognizable depositional sequences of the Jurassic anywhere in the world.

Tithonian

The Tithonian is again exclusively in the Kachchh Mainland presumably with still lesser margin-ward extent than the preceding Kimmeridgian transgressive interval. According to our intra-Mainland correlations, the Tithonian is almost completely absent at least in Jawaharnagar except for doubtful basal-most Tithonian. Even in Jhura and Habo sections there is no record as yet of any definite Tithonian. A total of six different sections have been studied (Figs. 6, 7). The Early Tithonian is best expressed at Ler, the Late Tithonian at Lakhapur and Mundhan. In the ca 130m thick Tithonian the number of ammonoid bearing levels in different sections varies between 27 and 48. Ammonoid stratigraphic studies have resulted in an ammonoid zonation scheme of 4 zones, 10 subzones and 13 horizons (Krishna et al., 1996a). The Early Tithonian succession is in general more sandy than the Kimmeridgian with a massive hard and compact dark brown coloured, fine to medium grained sandstone at the base.

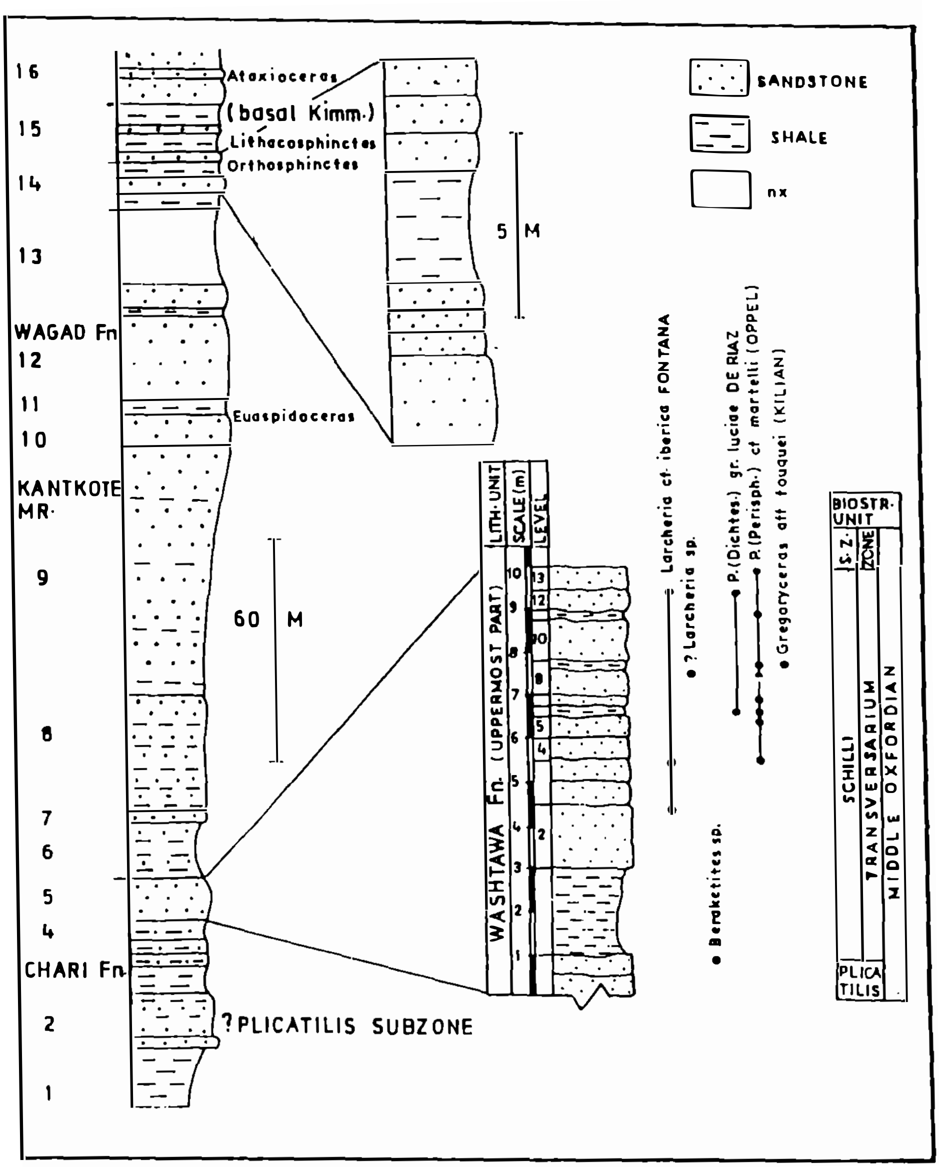

Figure 5. Lithostratigraphic column with position/ranges of ammonoids in the Oxfordian of Wagad, Kachchh.

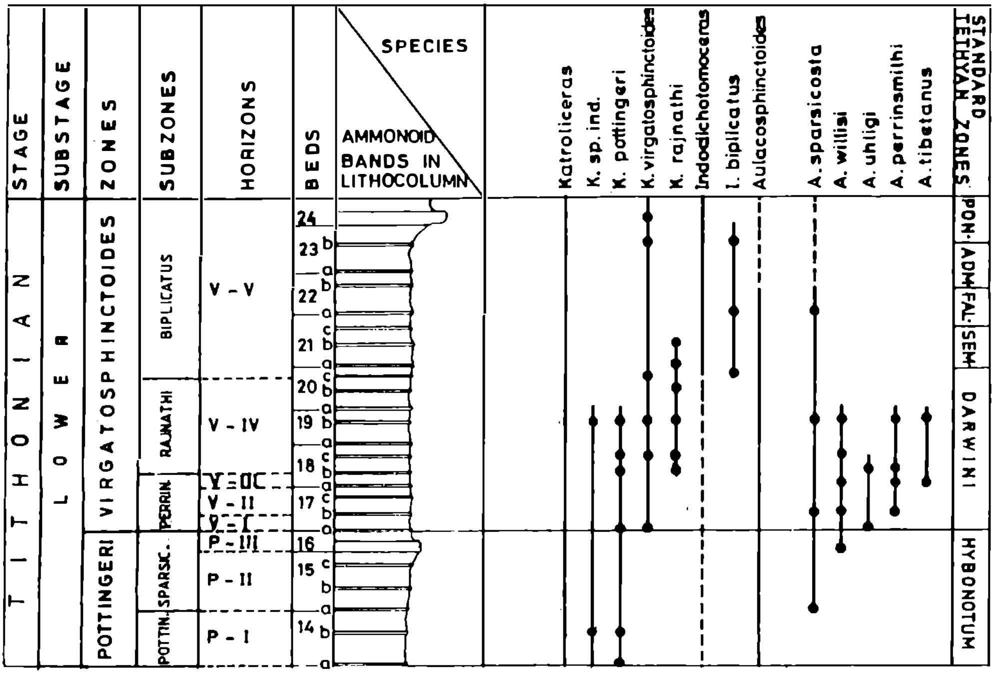

Figure 6. Ammonoid stratigraphic ranges and the zonal scheme in the Early Tithonian of west Ler, Kachchh.

Transgressive interval V: The maximum presence of ammonoids is observed in the early to middle part of the Virgatosphinctoides Zone in Perrinsmithi Subzone – Rajnathi Subzone Early Tithonian interval (Krishna et al., 1996a). The ammonoid assemblage is exclusively restricted to the Indo - East-African province and also consists exclusively of members of the subfamily Torquatisphinctinae. In comparison, the ammonoid density and diversity are distinctly reduced while the Tethyan elements are altogether absent (Fig. 3). Obviously the communication between the two margins of Tethys was reduced and ammonoid expansion prevented. The sediment texture has coarsened. These multiple factors suggest that there has been further reduction in basinal depth. This 3rd order transgressive interval thus suggests continuation of the second order regression. The basinal depth at Ler - Katrol in this interval is again of the order 10-25m. This is also supported by 100% endemism and general reduction in the size of the ammonoids. The interval is broadly dated as Early Tithonian.

Transgressive interval VI: The Late Tithonian is best developed in the West Mainland Lakhapur and Mundhan sections, while a part is also expressed in Ler - Katrol. In West Mainland at Mundhan, the section is ca 45m thick with over ten levels of ammonoids. Lithologically the sedimentary column is made up of medium to coarse sands alternating with silty shales while the topmost 4m thick part includes three 50cm to 1m thick green oolitic glauconitic sandstones which are ammonoid rich. The ammonoid assemblage in this transgressive interval mainly includes dense and finely ribbed species of *Virgatosphinctes*. The transgression is maximum in the 4m thick glauconitic sands assigned to the youngest part of Frequens Subzone of Densiplicatus Zone (Fig. 3). The interval is slow sedimented. A few Tethyan genera are also present (*Microcanthoceras* and *Aulacosphinctes*) (Krishna et al., 1996a). It may be also noted that the section is located in the westernmost exposed part of the Kachchh basin. Ammonoid morphology suggests greater basinal depth than the previous transgressive interval of early Early Tithonian age at Ler – Katrol. Even in Ler – Katrol, the coeval ammonoid-bearing sedimentary interval indicates somewhat greater depth than in the underlying lower Lower Tithonian sediment interval.

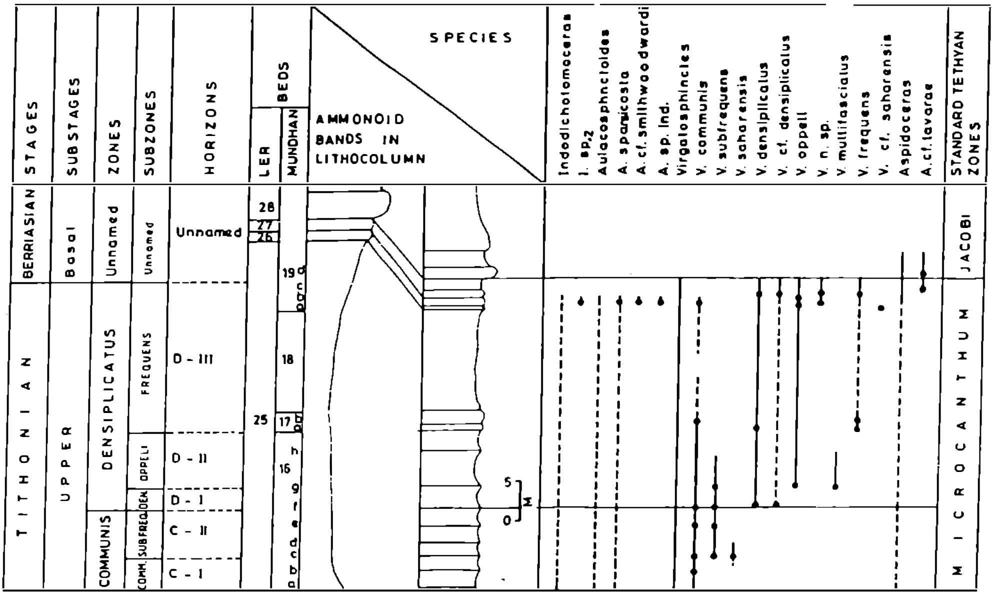

Figure 7. Ammonoid stratigraphic ranges and the zonal scheme in the Late Tithonian of Kachchh.

Comparison with Global Transgressive Interval Framework Formulated in Europe

The comparison with the Kachchh European Late Jurassic sequences strikingly brings out almost total dissimilarity at the 2nd order level. The 2nd order Kachchh maximum flooding is in the Middle Oxfordian Schilli Subzone which differs with those of the European models which are distinctly much later either in the Eudoxus - early Beckeri interval or Hybonotum – early Darwini interval. The 2nd order sequence boundaries also differ. If eustatism is the major control of the Vail sequence in Europe, the disagreement with Kachchh, at the 2nd order level indirectly implies the overriding influence of vertical tectonics in the India – Africa divergent region. In this context, the occurrence of intraformational conglomerates in the uppermost part of Dhosa Oolite of Cordatum Zone age and the origin of the 'Median High' suggestively after the Middle Oxfordian Schilli Subzone 2nd order maximum flooding attest to the dominance of tectonics in the Kachchh Jurassic sequence. It seems that until Schilli Subzone the tectonic subsidence complemented the eustatic rise to strengthen the rapid increase in the basinal depth. However, thereafter the tectonic uplift outweighed the still continuing eustatic rise which resulted in recession of the sea, first from Wagad in Early Kimmeridgian and later from Jhuran, Habo, Jhura belt around Kimmeridgian/Tithonian boundary. Yet, irrespective of the tectonically influenced 2nd order discordance between the Kachchh and European sequences, there is near concordance in the ages of the 3rd order transgressive interval (Figs. 3, 8). We are prompted to postulate that the eustatic signatures in the 3rd order sequences have escaped distortion by the vertical tectonics. It remains to be seen if the same holds good for the other regions of the world, remote from Europe. Similar comparison with other areas, for example Argentina, is in progress.

Conclusions

It is evident from the discussion that the ammonoid studies are not only useful in high order resolution of the geological column in the Mesozoic, but their compositional variation both temporally and spatially in the basin permits identification of 3rd and 4th order transgressive intervals. Further, the ammonoid studies provide tremendous help in sequence stratigraphic integration covering the entire basin irrespective of the presence or absence of ammonoids or their continuation laterally. and vertically. The diverse ammonoid studies have also helped to differentiate individual depositional sequences of different orders and their likely durations. In rare cases the smallest sequences are found to be of the order of 2000 years. The 4th and 5th order sequences suggest correspondence to Milankowich climatic cycles of 20,000, 40,000, 100,000 and 200,000 years. These estimates of the duration of depositional sequences , however crude they may be, in turn should pave the way to a better understanding of sedimentation rates and the tectonic and geological framework of the basin.

Acknowledgments

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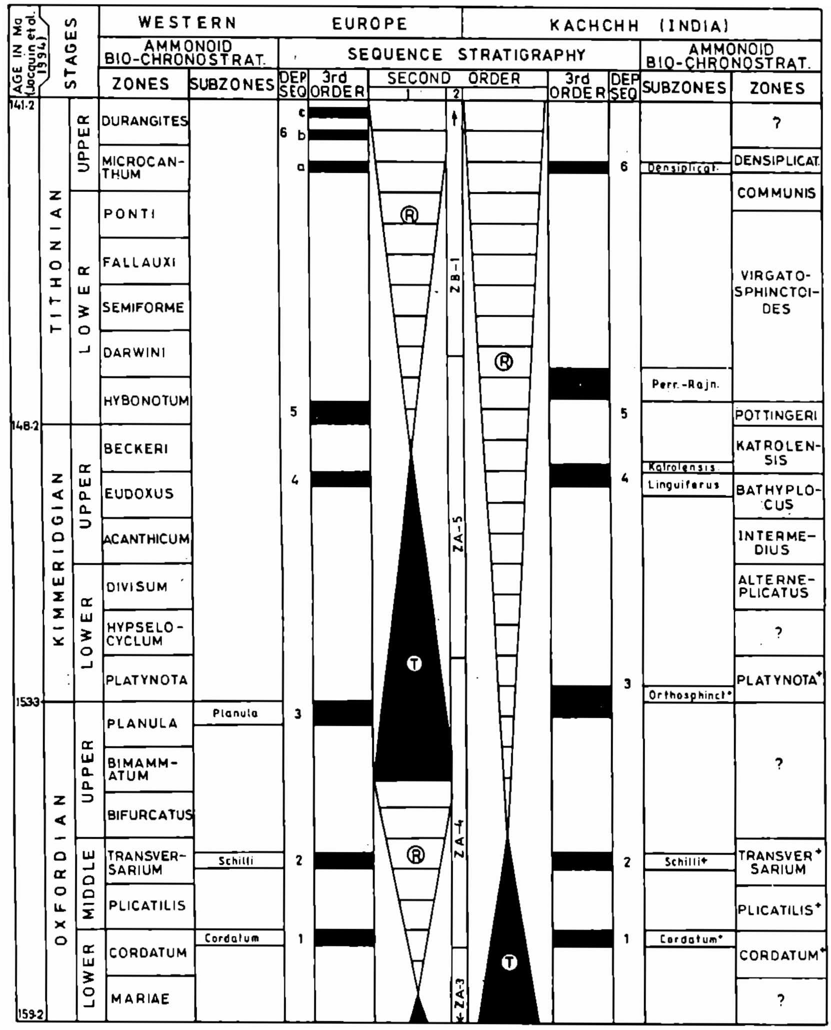

+ Indicate Tethyan Zones identified in Kachchh.

Figure 8. Comparison of the 2nd and 3rd order transgressive intervals and maximum flooding during the Late Jurassic between Western Europe and Kachchh; 1 to 6 indicate the corresponding transgressive intervals, columns 1 and 2 after Jacquin et al., 1994 and Haq et al., 1987, + marks Tethyan zones identified in Kachchh.

References

BISWAS, S. K. 1980. Structure of Kutch-Kathiwar region. Proceedings 3rd Indian Geological Congress, p. 225-272.

FÜRSICH, F. T., W. OSCHMANN, I. B. SINGH, AND A. K. JAITLY. 1992. Hardgrounds, reworked concretion levels and condensed horizons in the Jurassic of Western India: their significance for basin analysis. Journal of the Geological Society of India, 149:313 – 331.

GYGI, R .A., A. L. COE, AND P. R. VAIL. 1998. Sequence stratigraphy of the Oxfordian and Kimmeridgian stages (Late Jurassic) in northern Switzerland. Mesozoic and Cenozoic sequence stratigraphy of European basins, SEPM (Society for Sedimentary Geology), special publication, 60:527-544.

HAQ, B. G., J. HARDENBOL, AND P. R. VAIL. 1987. Chronology of fluctuating sea levels since the Triassic. Science, p. 1150-1166.

JACQUIN, T., P. Ch. de GRACIANSKY, AND P. R. VAIL. 1994. Jurassic sequence cycles chart for western European basins. Project - Mesozoic-Cenozoic sequence stratigraphy of European basins.

KRISHNA, J. 1987. An overview of the Mesozoic stratigraphy of Kachchh and Jaisalmer basin. Journal of the Palaeontological Society of India, 32:136-149.

____, G. MELENDEZ, B. PANDEY, AND D. B. PATHAK. 1995. Characterization of the ammonite genus Larcheria (Middle Oxfordian) in Kachchh (India): biostratigraphy and paleobiogeographic evaluation in the context of Tethyan occurrences. Comptes Rendus de l'Academie des Sciences, 321 (IIa):1187-1193.

____, AND J. R. OJHA. 1996. The Callovian Ammonoid Chronology in Kachchh (India). GeoResearch Forum, Transtec Publications, 1-2:151-165.

____, B. PANDEY, AND D. B. PATHAK. 1996. Ammonoid chronology developed in the Tithonian of Kachchh (India). GeoResearch Forum, Transtec Publications, 1-2:205-214.

____, D. B. PATHAK, AND B. PANDEY. 1996. Quantum refinement in the Kimmerdgian ammonoid chronology in Kachchh (India). GeoResearch Forum, Transtec Publications, 1-2:195-204.

____, ____, AND ____. 1998. Development of Oxfordian (Early Upper Jurassic) in the most proximally exposed part of the Kachchh basin at Wagad outside the Kachchh Mainland. Journal of the Geological Society of India, 52:513-522.

MARCHAND, D. 1982. Role des ammonoides pour les reconstitutions paleogeographiques, paleobathymetriques et paleotectoniques. Examples pris dans le Callovien et l' Oxfordien d'Europe occidentale. Bulletin de la societe geologique de France, Paris, 7, 24,(5–6): 1017–1023.

SCHWEIGERT, G., J. KRISHNA, B. PANDEY, AND D. B. PATHAK. 1996. A new approach to the correlation of the Upper Kimmeridgian Beckeri Zone across the Tethyan sea. Neues Jahrbuch Palaontologie, 202(3):345–373.

SPATH, L. F. 1927-33. Revision of the Jurassic cephalopod fauna of Kachh. Palaeontologia Indica, 9(2):1-945.

WESTERMANN, G. E. G. 1990. New developments in ecology of Jurassic-Cretaceous ammonoids. Atti del secondo convegno interzionale Fossili Evoluzione Ambiente Pergola, 87:459-478.

ZIEGLER, B. 1967. Ammoniten Okologie am Beispiel des oberjura. Geologique Rundschau, 56:439–464.

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Discontinuities and Sequence Stratigraphy of the Antalo Limestone (Upper Jurassic, North Ethiopia)

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Keywords: Mixed Ramp, Discontinuities, Sequence Stratigraphy, Upper Jurassic, Ethiopia

Abstract: The Antalo Limestone of north Ethiopia is an Upper Jurassic marine succession consisting of marls alternating with limestones and subordinate sandstones. Facies range from inner ramp, grain supported sediments to outer ramp marls and micritic limestones.

Different types of discontinuity surfaces (DS) frequently occur: omission surfaces and rock grounds. The former are revealed by firm ground burrows, whereas the latter are characterized by borings, bioencrustations and Fe-oxide coatings. Some rock grounds are related to submarine sediment starving whereas others, restricted to inner ramp facies, are associated with dissolution due to subaerial exposure. Other DS are to be hypothesized at the boundary between marls and the overlying inner ramp calcareous or sandstone bodies, although the surfaces themselves are never observable because of lack of exposure.

Several depositional sequences have been recognized. The most reliable criterion to define sequence boundaries (SB) is a facies contrast: high-energy facies directly overlying marls with open marine fossils may be considered as lowstand deposits and the bounding surface may be taken as a SB. The SB may coincide with the transgressive surface: the previous highstand sediments are sharply overlain by grain-supported sediments in turn either capped by a mineralized, bored and bioencrusted rock ground (maximum flooding surface) or characterized by an enrichment in authigenic grains (condensed section). Marls then follow, marking the renewal of sedimentation after the flooding of the ramp.

The other numerous DS developed within the same facies are interpreted as the product of a higher frequency cyclicity. They are of lesser use in sequence stratigraphy but may be very important for understanding the factors controlling sedimentation in mixed ramp depositional settings.

Introduction

Sequence stratigraphy, in the last decade, has changed the way stratigraphers look at sedimentary rocks because it has highlighted the importance of discontinuities in understanding the stratigraphic record, and directed attention to the genetic relationships between different types of coeval sediments.

Sequence stratigraphic models, developed first for siliciclastic margins and then extended to carbonate ones, have recently also been applied to ramp systems (Tucker et al., 1993). Ramps are depositional environments that, independent of the type of sediment supplied, are characterized by a very gentle and constant angle of slope (Ahr, 1973; Burchette and Wright, 1992). The absence of a major slope break means that: a) relative sea level changes on homoclinal ramps simply result in shifts of facies basinward or landward during regression or transgression, respectively; b) sequence boundaries correspond to subaerial exposure surfaces that pass basinwards into submarine erosional surfaces; and, c) lowstand systems tracts (LST) are characterized by a clastic influx due to increased erosion of the hinterland. Ramps are therefore considered to be simpler than siliciclastic shelves and carbonate rimmed shelves in terms of sequence stratigraphy (Tucker et al., 1993).

Geological Framework

The subject of this study is the Antalo Limestone (Blanford, 1870) which is a package of shallow marine limestone, marl and sandstone alternations about 700 metres thick referable to the Upper Jurassic. It has been studied in the type-area that is the Mekele Outlier, in northeastern Ethiopia (Fig. 1). Here, the Antalo Limestone overlies a thick succession of terrigenous continental sediments. The upper part of this clastic succession consists of red, cross-bedded, fluviatile quartzarenites (Adigrat Sandstone) presumed to be Early-Middle Jurassic in age because of the presumed conformable and transitional boundary with the overlying Antalo Limestone (Merla and Minucci, 1938). The upper part of the Antalo Limestone, as originally defined and described by Blanford (1870) and Merla and Minucci (1938), is poorly exposed because of its marly composition, and was separated as a new lithostratigraphic unit named the Agula Shale by Beyth (1972). The Agula Shale was still considered to be of Jurassic age and, because of the presence of gypsum lenses, attributed to a marginal marine environment. This succession of marine sediments is truncated by an erosional angular unconformity on which lies another unit of continental red sandstones and conglomerates, called the Amba Aradam Formation (Shumburo, 1968 *in* Beyth, 1972) and referred to the Upper Cretaceous, being bracketed between the Upper Jurassic Antalo Limestone and Eocene volcanics. The erosional unconformity cuts increasingly deeper into the Jurassic succession to the west, so that in the western part of the Mekele Outlier (Abi Adi sector) the Antalo Limestone is completely eroded and the Amba Aradam Formation directly overlies the Adigrat Sandstone (Arkin et al., 1971; Bosellini et al., 1997).

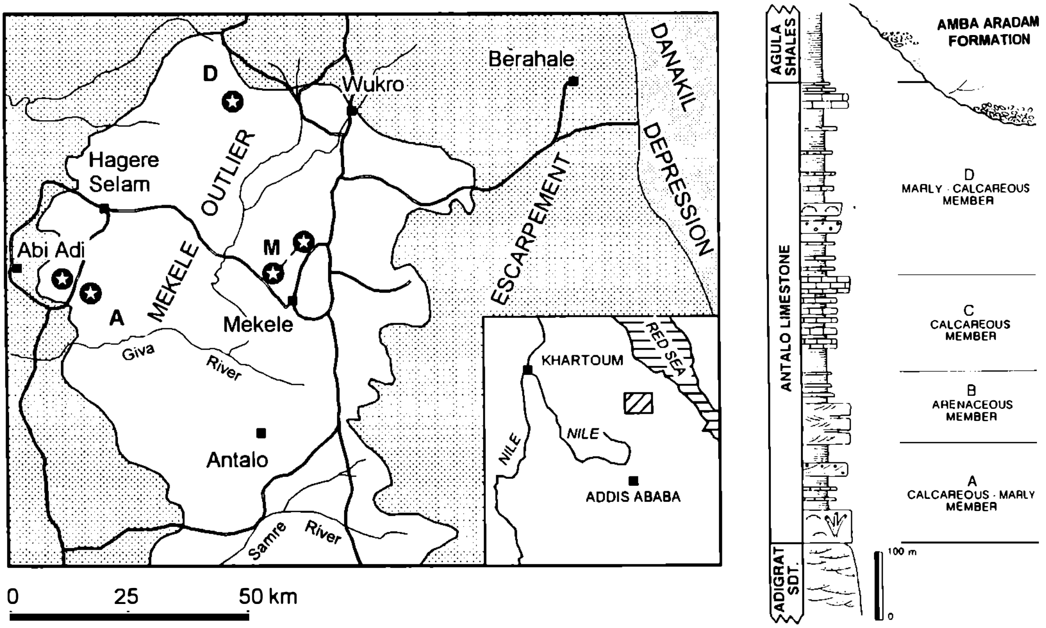


Figure 1. Geographic map of the Mekele Outlier showing locations of the measured stratigraphic sections (A: Agbe; D: Dugum; M: Mekele) and simplified stratigraphic column of the Antalo Limestone and under- and overlying formations.

The Mesozoic stratigraphic succession present in the Mekele Outlier reflects the tectono-sedimentary evolution of the Horn of Africa, summarized by Bosellini (1992), and basically relates to the breakup of the Gondwana supercontinent with the opening of the Indian Ocean.

Transgression in Ethiopia took place only in the Late Jurassic and gave rise to the Antalo and Agula formations. This marine episode, however, was short-lived because an Early Cretaceous tectonic event, interpreted as a distal intraplate effect of the South Atlantic opening, caused an upwarping of the whole Horn of Africa. This resulted in emersion of most of the Ethiopian and

Northern Somalian territory and in the deposition of alluvial conglomerates and sandstones of the Amba Aradam Formation.

The Antalo Limestone

The Antalo Limestone has been the object of several studies in different parts of the Horn of Africa (Turi et al., 1981; Russo et al., 1994; Sagri et al., 1998) and more specifically in Northern Ethiopia (Merla and Minucci, 1938; Beyth, 1972; Merla et al., 1979; Bosellini et al., 1995; 1997). In the most recent papers (Bosellini et al., 1995; 1997) the first sequence stratigraphic interpretation of The Antalo Limestone within the framework of the whole Mesozoic succession was attempted.

The purpose of this paper is to present results of a continuing study of the Antalo Limestone carried out in the north-western part of the Mekele Outlier geographically located between Mekele, Agbe and Wukro (Fig. 1). Several stratigraphic sections were measured in detail during three seasons of field work (January, 1996 to February, 1998), and more than 400 rock and fossil samples were collected which are still under study.

Four main lithostratigraphic units have been distinguished in the Antalo Limestone of the Mekele Outlier (Martire et al., 1998). From bottom to top they are (Fig. 1): Calcareous-Marly Unit A; Arenaceous Unit B; Calcareous Unit C; Marly-Calcareous Unit D. These units are recognizable all over the examined area even though important lateral facies changes may take place within each unit. These four units have been further divided into subunits characterized by homogeneity in lithology and facies. For the sake of clarity, only the lower and middle part of the Antalo Limestone will be considered here (Units A to C) and only three stratigraphic sections, the most suitable for illustrating the main stratigraphic traits within the formation, will be described.

Description of the Stratigraphic Sections

Agbe Section (Figures 2, 3)

This section is actually a composite one resulting from the integration of several tracts of exposures measured on both sides of the valley north of Agbe, along the road Abi Adi-Hagere Selam (Fig. 1). In this sector the boundary with the underlying Adigrat Sandstone is observable and appears to be sharp. The colour changes from red or white to yellow ochre; burrows and the carbonate content, mainly related to skeletal grains (bivalves, brachiopods), become important. The most typical facies of Subunit A1 (about 55m thick), however, is represented by thickly bedded, stromatoporoid- and coral-rich floatstones and rudstones interbedded with wackestones and ooidal-bioclastic grainstones. In the middle part of this subunit, an interval of wackestones to packstones with sponge spicules and rhaxes, is characterized by the presence of chert nodules. The top 5 metres of Subunit A1 are represented by more thinly bedded grainstones with very scarce stromatoporoids and corals (Fig. 4). Subunit A2 (about 55m thick) is mainly made up of marls and hence is poorly exposed. Some limestone beds occasionally crop out and show grain supported textures, basal erosional features and parallel or low angle laminae. Some ammonite moulds have been found in the lower part: *Gregoryceras* cf. *fouquei* (Kilian), *Dichotomosphinctes* cf. *rotoides* (Ronchadze), both typical of the Middle Oxfordian, and *Physodoceras* sp. that points to the Lower Kimmeridgian. The Oxfordian specimens, however, are reworked (reelaborated *sensu* Fernández-López, 1984) as shown by abrasion facets and ostreid encrustation on the moulds. Subunit A3 (over 65m thick) displays a wide variety of facies. The basal contact is very sharp and is marked by a massive body of cross bedded, oolitic grainstones. This is overlain by a marly interval and then another calcareous body consisting of stromatoporoid-coral floatstones to rudstones and oolitic-bioclastic grainstones. Both limestones show local interbedding of dasyclad-rich wackestones. The second is also characterized by closely spaced discontinuities and the presence of Fe-oolites in the upper part. After a poorly exposed marly interval, cross-bedded, quartz-rich bioclastic grainstones with scattered oolites occur and are overlain by pure, bioturbated, sandstones referable to Subunit B1 (21m thick). Above, thin sandstones alternating with marls and ostreid coquinas correspond to Subunit B2 (about 10m thick). After a covered tract, marl and limestone beds are regularly interbedded (Subunit C1, 35m thick). Limestones are represented by wackestones with benthic foraminifera (e.g. *Alveosepta*) and contain *Zoophycos* burrows and ammonite moulds (*Subnebrodites* cf. *hararinus* (Venzo), *Ataxioceras (Parataxioceras)* gr. *polyplocum* (Reinecke)) again referable to the Lower Kimmeridgian. After a covered tract, massively bedded limestones give rise to a clearly distinguishable cliff about 70 m high, as a whole attributed to Subunits C2-C4, that will be better defined in the Mekele section. Among the many facies recognized, noteworthy are cross-bedded, quartz-bearing, oolitic-bioclastic grainstones, nodular marly limestones grading into branching stromatoporoid floatstones to bafflestones, thin bedded mudstones with shrinkage pores, and lithoclastic breccias with collapse structures likely due to dissolution. Above a clear-cut surface, marls with interbedded oolitic and stromatoporoid limestone beds occur, which represent Unit D.

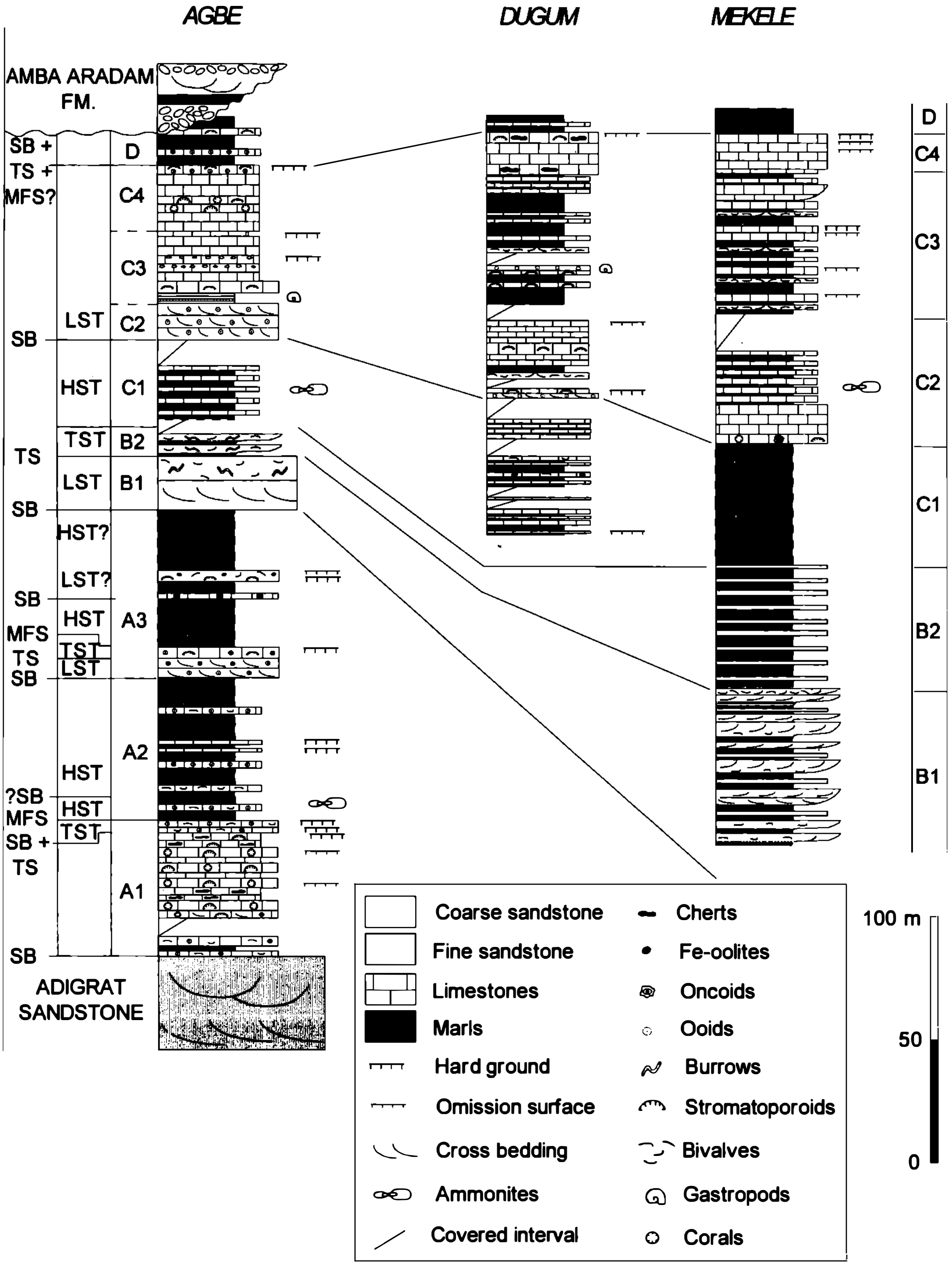
Mekele Section

This was measured a few kilometres north of Mekele, partly in natural outcrops on the escarpment due to an east-west trending normal fault, and partly along the road to Wukro (Fig. 1). In this sector the lowest exposed parts of the Antalo Limestone are the sandstones of Unit B (Fig. 2). Subunit B1 (65m thick) consists of several thickening and coarsening upward cycles, with marly siltstones at the base and cross-bedded coarse brown sandstones at the top. Ostreid-rich coquinoid layers are frequent, especially at the top of the subunit, where a rusty colour shows the enrichment in Fe-oxides. Herringbone cross-bedding is common in the sandstones and documents tidal influence. The overlying subunit B2 (49m thick) is widely covered. It is made up of marls and interbedded thin layers of fine sandstones characterized by parallel or low angle oblique laminae and by erosional bases with burrow casts. After a long tract (about 50m thick) almost completely covered and corresponding to light coloured marls (Subunit C1), Subunit C2 (36m thick) follows, represented by grey to light brown, micritic limestones with marly interlayers. The thickness of the marls increases upward. At the base of this subunit occur some massive beds containing oncoids, benthic foraminifera (*Alveosepta, Everticyclammina*) and fragments of corals and stromatoporoids. The overlying strata are characterized by the presence of *Zoophycos* and by frequent ammonite moulds (e.g. *Pthychophylloceras* sp., *Ataxioceras (Ataxioceras)* gr. *discoidale* Schneid) that allow assignment of these levels to the Lower Kimmeridgian. After a covered tract, Subunit C3 (60m thick) still appears as limestone-marl alternations, but is characterized by the presence of storm layers consisting of grain-supported textures (peloidal grainstones, coquinoid rudstones) and erosional bases with burrow casts. The absence of marls and of storm layers and the thick bedding are the distinctive features of Subunit C4 that crops out as a laterally persistent cliff. A sharp and flat surface characterized by Fe-oxide coatings separates mudstones of Subunit C4 from brachiopod-bearing marls of Unit D.

Dugum Section

In the northern part of the Mekele Outlier (Dugum sector) no complete section of the Antalo Limestone could be measured, despite good outcrop conditions, because of the presence of several faults that break up the succession. Attention was focussed on Unit C which is the best and most continuously exposed unit (Fig. 2). The first 50 metres are closely comparable to Subunit C1 of the Agbe section. Mudstones and wackestones with sponge spicules and benthic foraminifera (e.g. *Alveosepta*), locally also containing small oncoids, are interlayered with marls of variable thickness. After a short covered tract, cross-bedded oolitic grainstones crop out and are followed by a succession of marls and limestones of diverse facies. These range from ostreid coquinas to large gastropod-bearing, oolitic grainstones to stromatoporoid floatstones with a matrix of spiculitic wackestones. These sediments are organized in cycles several metres thick with marls at the base and massive limestones with stromatoporoids at the top. The top of Unit C is remarkable because it is represented by another distinctive cliff where massive mudstones or spiculitic wackestones with chert nodules and silicified stromatoporoids are exposed. Textures and weathering style allow comparison of this package of limestones with Subunit C4 as already seen in both the Agbe and Mekele sections.

Figure 2. Stratigraphic columns for lower and middle parts of Antalo Limestone, showing subunits (A1 to D) and sequence stratigraphy interpretations. SB: sequence boundary; TS: transgressive surface; MFS: maximum flooding surface; HST: highstand systems tract; LST: lowstand systems tract; TST: transgressive systems tract.


AGBE
DUGUM
MEKELE
AMBA ARADAM FM.
SB + TS + MFS?
D
C4
C3
LST
C2
SB
HST
C1
TST
B2
TS
LST
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HST?
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HST
A3
MFS
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A2
HST
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HST
MFS
TST
SB + TS
A1
SB
ADIGRAT SANDSTONE
D
C4
C3
C2
C1
B2
B1
Coarse sandstone
Fine sandstone
Limestones
Marls
Hard ground
Omission surface
Cross bedding
Ammonites
Covered interval
Cherts
Fe-oolites
Oncoids
Ooids
Burrows
Stromatoporoids
Bivalves
Gastropods
Corals
100 m
50
0

Discontinuity Surfaces and Sequence Stratigraphy

Even though a detailed treatment of the sedimentology of the Antalo Limestone is not possible here, it is clear that all ramp environments are represented in the Antalo Limestone, from inner ramp oolitic barrier complexes or arenaceous estuarine systems to outer ramp marls, through a wide range of possibly middle ramp facies belts where corals and stromatoporoids could thrive without building a real margin and where storms regularly interrupted the normal deposition of fine grained sediments.

Discontinuity surfaces (DS) are present in all the units of the Antalo Limestone although they are particularly frequent and evident in limestone intervals (Fig. 2). In the following, we will adopt the classification of DS recently proposed by Clari et al. (1995). We will therefore distinguish omission surfaces when only firm ground burrows occur, and rock grounds where evidence of lithification is present. The latter are further separated into hard grounds and inherited rock grounds if the environments of deposition of the under- and overlying sediments and formation of the surface itself are the same or different, respectively.

Omission surfaces are especially well developed in middle and outer ramp facies within mud supported limestones. Centimetre-sized firm ground burrows, locally displaying U-shaped tubes, occur in the top 10-20cm of the bed and are often filled with coarser and darker sediment such as benthic foraminifer packstones.

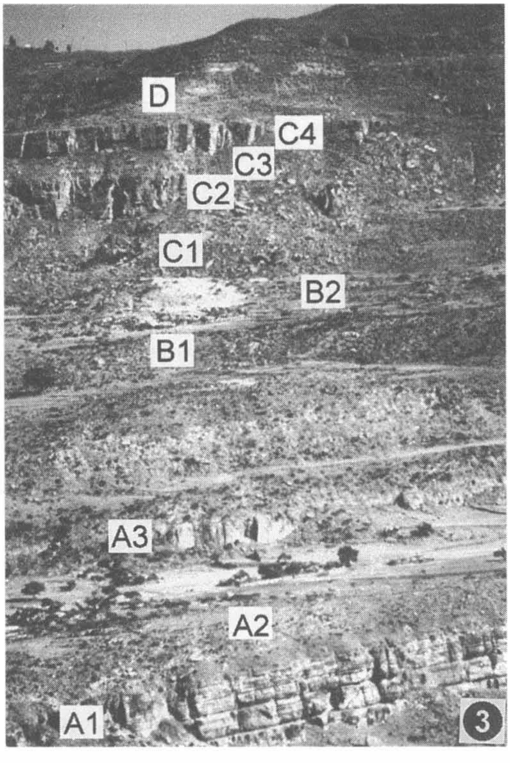

Figure 3. The Antalo Limestone at Agbe and its subdivision in units and subunits.

Rock grounds are clearly recognizable by the presence of one or more of the following criteria: bivalve boring, ostreid encrustation, and Fe-oxide staining and coating of the surface (Fig. 5). Rock grounds are common in all calcareous facies, from desiccated inner ramp mudstones (possible C3-C4 Subunit boundary in Agbe section) to outer ramp wackestones (C3 Subunit of Mekele section). Where these rock grounds occur at the boundary between markedly different facies, e.g. at the top of calcareous bodies overlain by marls such as the top of A1 and C4 Subunits in the Agbe and

Mekele sections, an important environmental change is indicated and the corresponding surface deserves the term inherited rock ground. Other DS are developed within Unit C in the Agbe section and are clearly associated with subaerial exposure. Stromatoporoids and other bioclasts are leached and replaced by a coarse calcite spar or by a fine mosaic of dolomite crystals (Fig. 6); in other instances, less selective dissolution is indicated by the presence of lithoclastic breccias with collapse structures. A fully lithified substrate below these DS has to be inferred although no clear evidence of it is present.

Other important and abrupt facies changes are those at the lower boundary of shallow water rock bodies overlying outer ramp marls and micritic limestones (e.g. the base of Subunits A3, B1 and C2 in the Agbe and Dugum sections). Unfortunately, because of the poor exposure of marls, the bounding DS are not observable.

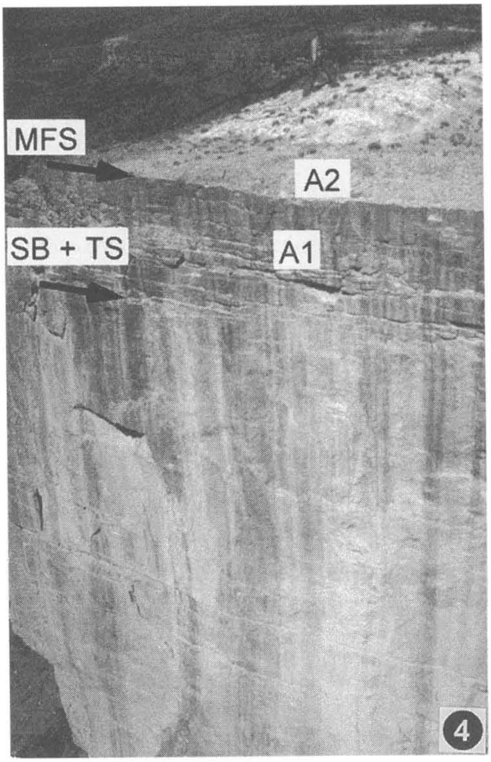

Figure 4. Boundary between subunit A1 and subunit A2 in the Agbe section. The uppermost part of the limestone cliff, characterized by regular bedding, consists of oolitic and bioclastic grainstones interpreted as TST deposits resting direcly on the sequence boundary. The most evident surface, however, separating the limestones of subunit A1 from marls of subunit A2, is the maximum flooding surface.

The DS identified have been used to subdivide the Antalo Limestone into depositional sequences (Fig. 2). Discontinuity surfaces of major interest to sequence stratigraphy are obviously those that separate markedly different facies associations, i.e. systems tracts. In the standard model of a depositional sequence, DS may be found both at the sequence boundary and also in the middle part associated with the condensed section. Most such DS in the Antalo Limestone, e.g. inherited rock grounds, correspond to deepening and may be interpreted as representing transgressive events or maximum flooding surfaces. Sequence boundaries (SB), on the contrary, can be inferred from sharp facies changes corresponding to abrupt downward shifts of the facies belts. However, they cannot be directly observed because they are developed at the top of marl-rich units (highstand systems tracts: HST) which are characterized by very poor exposures. The shallow water limestone (A3, C2) or sandstone (B1) bodies overlying these masked surfaces are consequently referred to lowstand system tracts (LST). In the Mekele section, the SB at the base of C2 is developed within outer ramp facies. It has been put at the transition from marls to micritic limestones that at their base contain oncoids, stromatoporoid and coral fragments. A different situation is present at the top of Subunit A1 where the few metres of grainstones, sharply overlying cherty, stromatoporoid-rich wackestones, are interpreted as transgressive deposits directly overlying the sequence boundary (Fig. 4). In other places, this DS is marked by dissolution of stromatoporoids and is overlain by a veneer of cross-bedded sandstones and hence provides evidence for a period of subaerial exposure.

A comparable situation can be imagined for the top of C4 where, however, transgressive deposits are lacking. In reality a more complicated sequence of events is suggested for the A1/A2 boundary at Agbe. Reelaborated Middle Oxfordian taxa in fact occur a few metres below Early Kimmeridgian ones within the lowest part of the Subunit A2 marl-rich succession. It seems hard to envisage that the Middle Oxfordian ammonite moulds were reelaborated from the underlying stromatoporoid floatstones or transgressive grainstones because of the nature of the moulds, consisting of marly, fine-grained limestones, and the absence of other litho- or bioclasts. A possible explanation is to assume that within the A2 marls a hidden erosional surface occurs, which of course has to be associated with an important chronostratigraphic hiatus. An alternative possibility, difficult to confirm at present, is that the local stratigraphic ranges of the ammonite taxa recognized differ from those of the much better known Mediterranean and European bioprovinces. Only detailed and specific sedimentologic and biostratigraphic studies will help to solve this problem. In this case a new sequence boundary is to be hypothesized a few metres above the A1/A2 boundary as it has been represented in Fig. 2. A peculiar feature of the Antalo Limestone is the marked asymmetry of the depositional sequences that do not show any regressive depositional sequences resulting from progradation. This could be due to strong subsidence rates that outpaced sediment accumulation except for periods of significant sea level fall.

DS developed within the same facies are less important in sequence stratigraphy and can be at most considered as parasequence boundaries. However, they are interesting from a sedimentological viewpoint. In fact, within Unit C, where lateral facies changes are clearly represented, DS due to subaerial exposure and developed at the top of inner ramp, upward shallowing cycles from nodular marls to stromatoporoid floatstones (Agbe section, Fig. 6) can be correlated with omission surfaces or hard grounds occurring at the top of outer ramp cycles displaying an upward thickening of limestone beds (Mekele section, Figures 7, 8). This allows the proposal that micrite originating in shallow water was subsequently redistributed across the ramp by storm currents as recently also proposed by Pittet and Strasser (1998). DS at the top of marl-micritic limestone cycles may, therefore, be due to sea level falls that result in emersion of the inner ramp and deactivation of the "productive factory". Together with this sediment starving effect, an increase in the activity of currents on the outer ramp could have played a role in hindering preservation of the scarce sediment produced during these small scale and high frequency lowstands.

Figure 5. Detail of the bored and encrusted maximum flooding surface developed at the top of the transgressive deposits in the upper part of subunit A1 shown in Fig. 4. Agbe section.

Figure 6. Polished slab of stromatoporoid floatstones of unit C in the Agbe section: most of the fragments of branching stromatoporoids are dissolved and the resulting cavities are filled with a greenish, dolomitized bioclastic packstone.

Figure 7. Limestone-marl alternations of subunit C3 in the Mekele section. The thick limestone bed in the lower part of the photo represents the top of an upward thickening sequence.

Figure 8. Polished slab of the top of the massive bed of Fig. 7. The presence of an omission surface, afterwards evolved into a hard ground, is revealed by the numerous firm ground burrows and borings filled by foraminifer-rich packstones.


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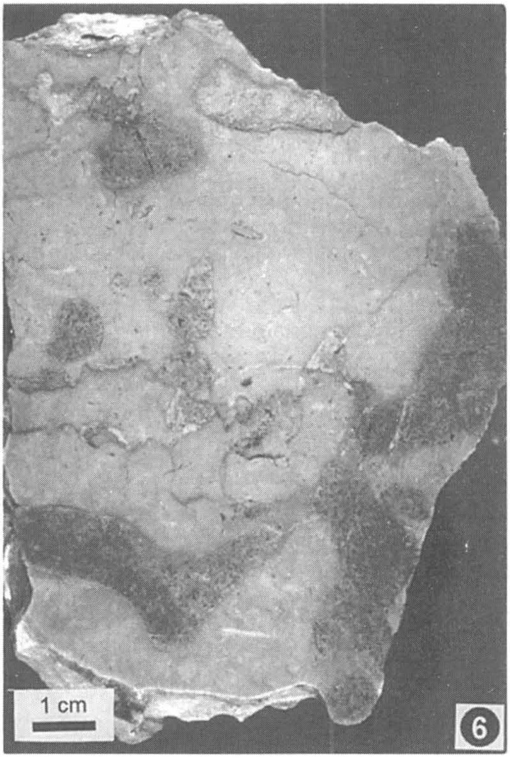
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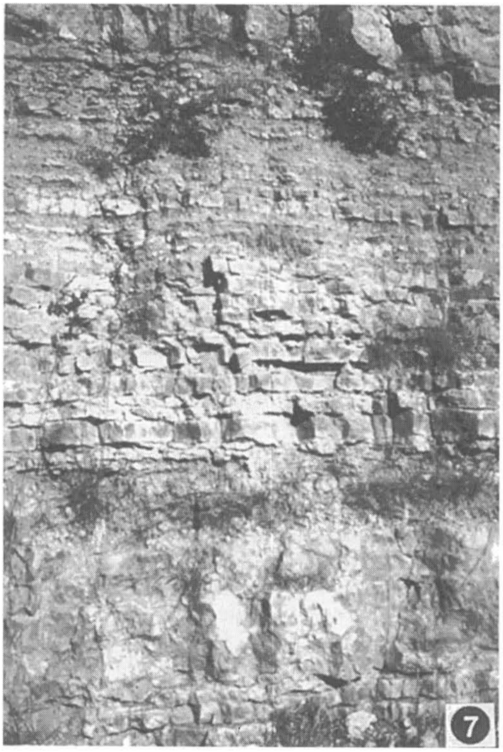
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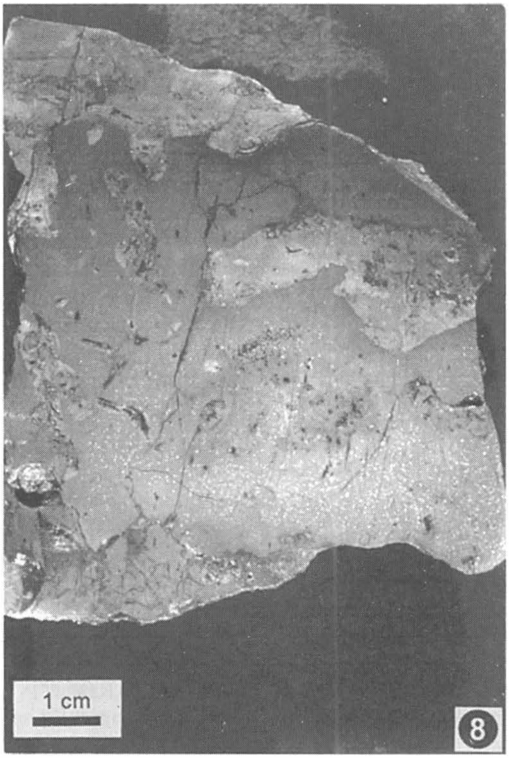
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Conclusions

The present study of the Antalo Limestone in the Mekele Outlier has allowed distinction of six depositional sequences. Sequence boundaries have been put where a sudden shallowing of facies occurs, even if the surfaces and the probable associated erosional features could not be directly observed. Transgressive and maximum flooding surfaces and related condensed deposits, on the contrary, are clearly recognizable and document a rapid deepening.

All the depositional sequences are developed in a mixed, carbonate-siliciclastic ramp over a time span that is not precisely defined but lies within the Early Kimmeridgian, and possibly within only the first two biochronozones, with the exception of the first sequence. This hinders any regional correlation with other sections of the Antalo Limestone in the Horn of Africa, and also with the global sea level curve. Causes for these high frequency, relative sea level cycles are consequently difficult to state. However, as no evidence for synsedimentary tectonics has been recognized in the Antalo Limestone, the mechanism for sea level changes may be found in eustacy or regional intraplate stresses (Cloetingh, 1988).

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References

AHR, W. M. 1973. The carbonate ramp: an alternative to the shelf model. Transactions Gulf Coast Association of Geological Societies, 23:221-225.

ARKIN, Y., M. BEYTH, D. B. DOW, D. LEVITTE, TEMESGEN HAILE AND TSEGAYE HAILU. 1971. Geological map of Mekele sheet Area ND 37-11, Tigre Province, scale 1:250.000. Geological Survey of Ethiopia, Addis Ababa.

BEYTH, M. 1972. Paleozoic-Mesozoic sedimentary basin of Mekele outlier, northern Ethiopia. American Association of Petroleum Geologists, Bulletin 56:2426-2439.

BLANFORD, W. T. 1870. Observations on the geology and the zoology of Abyssinia during the progress of British expedition to that country in 1867-1868, McMillan, London, 487 p.

BOSELLINI, A. 1992. The continental margins of Somalia. Structural evolution and sequence stratigraphy. American Association of Petroleum Geologists, Memoir 53:185-205.

____, A. RUSSO, P. L. FANTOZZI, GETANEH ASSEFA AND TADESSE SOLOMON. 1997. The Mesozoic succession of the Mekele Outlier (Tigre Province, Ethiopia). Memorie di Scienze Geologiche, 49:95-116.

____, A. RUSSO, AND GETANEH ASSEFA. 1995. Il Calcare di Antalo nella regione di Makallè (Tigrai, Etiopia settentrionale). Atti Accademia Nazionale dei Lincei, Rendiconti Scienze Fisiche e Naturali, serie 9, 6:253-267.

BURCHETTE, T. P. AND V. P. WRIGHT. 1992. Carbonate ramp depositional systems. Sedimentary Geology, 79:3-57.

CLARI, P.A., F. DELA PIERRE, AND L. MARTIRE. 1995. Discontinuities in carbonate successions: identification, interpretation and classification of some Italian examples. Sedimentary Geology, 100:97-121.

CLOETINGH, S. 1988. Intraplate stresses: a tectonic cause for third-order cycles in apparent sea-level?, p. 19-29. *In* C. K. Wilgus et al. (eds.), Sea-level changes: an integrated approach. Society of Economic Paleontologists and Mineralogists, Special Publication 42.

FERNÁNDEZ-LÓPEZ, S. 1984. Criterios elementales de reelaboracion tafonomica en ammonites de la Cordillera Iberica. Acta Geològica Hispànica, 19:105-116.

MARTIRE, L., P. CLARI, AND G. PAVIA. 1998. Stratigraphic analysis of the Upper Jurassic (Oxfordian-Kimmeridgian) Antalo Limestone in the Mekele Outlier (Tigrai, Northern Ethiopia): preliminary data, p. 131-144. *In* S. Crasquin-Soleau and D. E. Barrier (eds.), Peri Tethys Memoir 4, Mémoires du Muséum National d'Histoire Naturelle, 179.

MERLA, A., E. ABBATE, A. AZZAROLI, P. BRUNI, P. CANUTI, M. FAZZUOLI, M. SAGRI, AND P. TACCONI. 1979. A geological map of Ethiopia and Somalia. CNR Italy, Firenze, 95 p.

MERLA, G., AND MINUCCI, E. 1938. Missione geologica nel Tigrai. Reale Accademia d'Italia. 336 p., Roma.

PITTET, B. AND A. STRASSER. 1998. Depositional sequences in deep-shelf environments formed through carbonate-mud import from the shallow platform (Late Oxfordian, German Swabian Alb and Eastern Swiss Jura). Eclogae geologicae Helvetiae, 91:149-169.

RUSSO, A., GETANEH ASSEFA AND BALAMWAL ATNAFU. 1994. Sedimentary evolution of the Abbay River (Blue Nile) Basin, Ethiopia. Neues Jahrbuch für Geologie und Paläontologie. Monatshefte, 5:291-308.

SAGRI, M., E. ABBATE, A. AZZAROLI, M. L. BALESTRIERI, M. BENVENUTI, P. BRUNI, M. FAZZUOLI, G. FICCARELLI, M. MARCUCCI, M. PAPINI, G. PAVIA, V. REALE, L. ROOK, AND TEWELDE MEDHIN TECLE. 1998. New data on the Jurassic and Neogene to Quaternary sedimentation in the Danakil horst and northern Afar depression, Eritrea, p. 193-214. *In* S. Crasquin-Soleau and E. Barrier (eds.), Peri Tethys Memoir 3, Mémoires du Muséum National d'Histoire Naturelle, 177.

TUCKER, M. E., F. CALVET, AND D. HUNT. 1993. Sequence stratigraphy of carbonate ramps: systems tracts, models and application to the Muschelkalk carbonate platforms of eastern Spain, p. 397-415. *In* H. V. Posamentier, C. P. Summerhayes, B.U. Haq and G. P. Allen (eds.), Sequence stratigraphy and facies association. International Association of Sedimentologists, Special Publication 18.

TURI, A., L. BIGI, AND C. PIRINI RADDRIZZANI. 1981. Microfacies of the Antalo Limestone (Middle to Upper Jurassic) in some sections of east-central Ethiopia. Bollettino Società Geologica Italiana, 99:437-454.

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Geologic Evolution of the Lusitanian Basin (Portugal) during the Late Jurassic

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Keywords: Upper Jurassic, Sequence Stratigraphy, Rifting, Lusitanian Basin, Portugal

Abstract: During the Late Jurassic, the Lusitanian Basin was located between two Hercynian basement blocks, today represented by the Berlengas islands in the west, and the Iberian block in the east. The graben in between was elongated NNE-SSW with depositional environments becoming less terrestrial and more open marine to the southwest.

A dense network of faults (NNE-SSW, WNW-ESE and NW-SE) acted as an influence on subsidence and paleogeography as proved by major regional variations in facies distribution and thickness.

A detailed analysis of the sedimentary successions across the Lusitanian Basin leads to the definition of eleven basin-wide depositional sequences (A to K) within the Upper Jurassic-Berriasian. In this analysis we consider lateral facies changes and interpreted depositional systems; the identification and classification of depositional surfaces; and the interpretation of sedimentary systems tracts.

The Late Jurassic evolution of the basin involved three major tectonosedimentary stages. Sequences A and B correspond to the onset of rifting, which resulted in widespread drowning of the basin, mainly with carbonate deposition; extensional climax was reached during sequences C, D and E (uppermost Oxfordian to Upper Kimmeridgian). The late stage (uppermost Kimmeridgian, Tithonian and Berriasian) includes sequences F to K and is interpreted as a period of thermal subsidence, sea-level changes and short tectonic pulses.

Introduction

The main goal of the paper is to critically reinterpret the Upper Jurassic sucession of the Lusitanian Basin emphasising a sequence stratigraphic approach. To achieve this goal the large amount of data collected mainly after 1940 were systematically analysed. Sequential analysis at a regional scale was based on a detailed study of every available outcrop, section and well column, as well as on seismic profiles. A significant part of the sediments record continental and transitional depositional systems. One may recognise major variations in accommodation space influenced both by episodes of intense extensional tectonic activity as well as eustatic sea level changes. In these conditions conceptual models risk being oversimplistic. For practical purposes, flooding surfaces and correlative inflection surfaces in terrestrial and deep sea systems (progradation-retrogradation) proved the best markers for bounding cycles (Plint, 1996).

Palaeontological data are integrated for palaeoecological interpretation. The chronostratigraphic framework is based essentially on ammonite biostratigraphy (Atrops and Marques, 1986; 1988a; 1988b; Mouterde et al., 1973; 1979), planktonic and benthic foraminifera (Wightman, 1990), dinoflagellate cysts (Riding and Thomas, 1988) and pollen (Du Chene, 1988; Mohr and Schmidt 1987; Mohr, 1989). For time stratigraphic analysis we used the ammonite-biozone charts of Cariou et al. (1991) for the Oxfordian; Hantzpergue et al. (1991) for the Kimmeridgian; Geyssant and Enay (1991) for the Tithonian and Erba et al. (1995) for the Berriasian. The basin was divided into sectors with different sedimentary records (Figs. 1 and 2). The position of the boundaries between sectors shifted or faded out during the period under consideration.

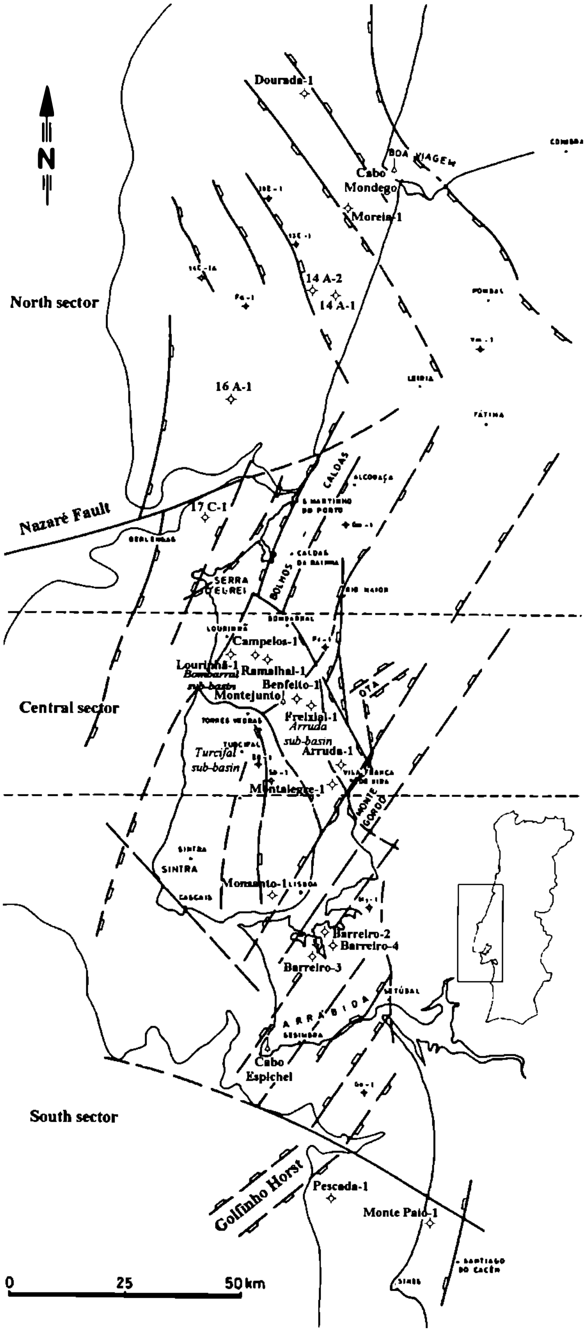
Outline of Upper Jurassic Basin Evolution

During the Late Jurassic, the Lusitanian Basin was located between two Hercynian basement blocks today represented by the Berlengas islands in the west and the Iberian block in the east. The graben in between was elongated NNE-SSW with depositional environments becoming less terrestrial and more open marine to the southwest. The rotation of the Iberian plate with a pole located to the east caused a NNE-SSW or NE-SW crustal extension (Brunet, 1992), resulting in deep fault-controlled sub-basins. A dense network of faults (NNE-SSW, WNW-ESE and NW-SE) controlled subsidence and paleogeography as proved by major regional variations in facies distribution and thickness. A significant tectonic event took place from Callovian-Oxfordian. At the end of the Middle Jurassic the Lusitanian Basin underwent a regional uplift, and erosion affected previous sediments causing karstification in some places (Guéry et al., 1986; Wright and Wilson, 1987).

During tectonic activity, in the northern part of the basin, Lower and Middle Jurassic carbonate sediments were cut into subsident blocks by NW-SE trended normal faults (Fig. 1). To the south of the Nazaré fault, the structure orientation changes to NE-SW. The granitic and metamorphic horst of Berlengas follows the Berlengas fault trend (NNE-SSW). The Caldas structure separated this sector into two subsident areas (Wilson, 1979; Canérot et al., 1995): one in the NW (S. Martinho) with moderate tectonic subsidence and another to SE (Bombarral sub-basin) with intense tectonic subsidence (Fig. 1).

During the Late Oxfordian, extensional activity strongly increased, marking the climax of rifting. According to Mauffret et al. (1988), these changes were associated with the establishment of a rift segment in the Tagus Abyssal Plain, and at the same time a dike network trending NW-SE or E-W was formed (Ferreira and Macedo, 1983; Hill, 1988; Willis, 1988). The main faults in the central area of the basin were very active (Fig. 1), intensifying the definition of blocks, tilting to SE. In the Arruda-1 well (CPP, 1956; Leinfelder and Wilson, 1989), which is next to some main active faults (Pragança, Vila Franca, etc), mixed carbonate and siliciclastic turbidites occur, reaching more than 2000m in thickness (Arruda sub-basin). Strongly subsident sub-basins are defined (Arruda. Bombarral and Turcifal) but where uplifted blocks remain (Ota, Monte Gordo), carbonate platform sediments continue to accumulate. In the southern sector of the Lusitanian Basin the extensional episode led to the establishment of a basically NNE-SSW oriented tectonic complex composed of ESE-tilted blocks in a domino fashion. In northern areas, deltaic systems prograded in highstand sea level conditions (Bernardes and Corrochano, 1992), and on the western border of central areas. alluvial fans and deltas drained to the east. Southward prograding deltaic and clastic slope depositional systems with a high proportion of turbiditic deposits (Abadia Formation), gradually reduced the area of platform slope and pelagic marl deposition (S. Pedro and Ramalhão Formations in the South sector). After the rift climax, coarse-grained siliciclastic sedimentation probably reflects erosion from the borders and subsequent expansion of the drainage basin after cessation of fault movement. The overall progradation of the siliciclastic systems (Abadia Formation) can be interpreted as a highstand-like linkage of depositional systems, created as a response to the tectonic derived relative sea-level rise. In the Late Kimmeridgian (base of Eudoxus Biozone), a short term sea-level rise (within a long term eustatic highstand (Haq et al., 1987; Ponsot and Vail, 1991)) generated an important transgressive surface, followed by widespread deposition of carbonate shelf sediments. The Amaral Formation which includes ooid grainstones and coral bioherms, overlies deeper marine sediments in the south and central sectors of the basin, indicating a decrease in bathymetry. In the north sector, southernmost area (Arrábida), and structural highs (e.g., Ota, Caldas da Rainha and Monte Gordo), the Amaral carbonates and presumed equivalents are a transgressive (deepening) event within shallow marine, marginal and continental clastic deposits. The Tithonian and Berriasian are represented in the centre and north by the Lourinhã Formation. In the axis of the basin, shallow marine marly deposition persisted, over which fluvial and fan delta systems prograded. Up to five short-term, widespread transgressions were recorded within the Lourinhã Formation. The upper beds of this unit are Late Berriasian in age (Rey, 1972; Berthou and Leereveld, 1990). Persistent carbonate deposition was restricted to shallow platform systems in the southern part of the basin (upper part of Mem Martins, Farta Pão and Porto da Calada Formations).

Figure 1. Sketch of Lusitanian basin with main structures active during Upper Jurassic times


N
Dourada-1
BOA VIAGEM
COIMBRA
Cabo Mondego
Moreia-1
14 A-2
14 A-1
North sector
POMBAL
LEIRIA
FÁTIMA
16 A-1
CALDAS
ALCOBAÇA
S. MARTINHO DO PORTO
Nazaré Fault
17 C-1
BERLENGAS
CALDAS DA RAINHA
SERRA D'EL-REI
BOLHOS
RIO MAIOR
BOMBARRAL
LOURINHÃ
Campelos-1
Lourinhã-1
Bombarral sub-basin
Ramalhal-1
Benfeito-1
Montejunto
OTA
Central sector
TORRES VEDRAS
Freixial-1
Arruda sub-basin
TURCIFAL
Turcifal sub-basin
Arruda-1
VILA FRANCA DE XIRA
Montalegre-1
MONTE GORDO
SINTRA
Monsanto-1
LISBOA
CASCAIS
Barreiro-2
Barreiro-4
Barreiro-3
SETÚBAL
ARRÁBIDA
SESIMBRA
Cabo Espichel
South sector
Golfinho Horst
Pescada-1
Monte Paio-1
SANTIAGO DO CACÉM
SINES
0
25
50 km

Sequence Stratigraphic Framework of the Upper Jurassic-Berriasian of the Lusitanian Basin

The integration at basin scale of the data analysed separately by sector made possible the establishment of a sequence stratigraphic framework for the Lusitanian Basin. The proposed chronostratigraphic distribution of facies associations, sequences and the corresponding lithostratigraphic scheme are displayed on Figure 2. Figures 3, 4 and 5 show correlation panels in the three sectors of the basin.

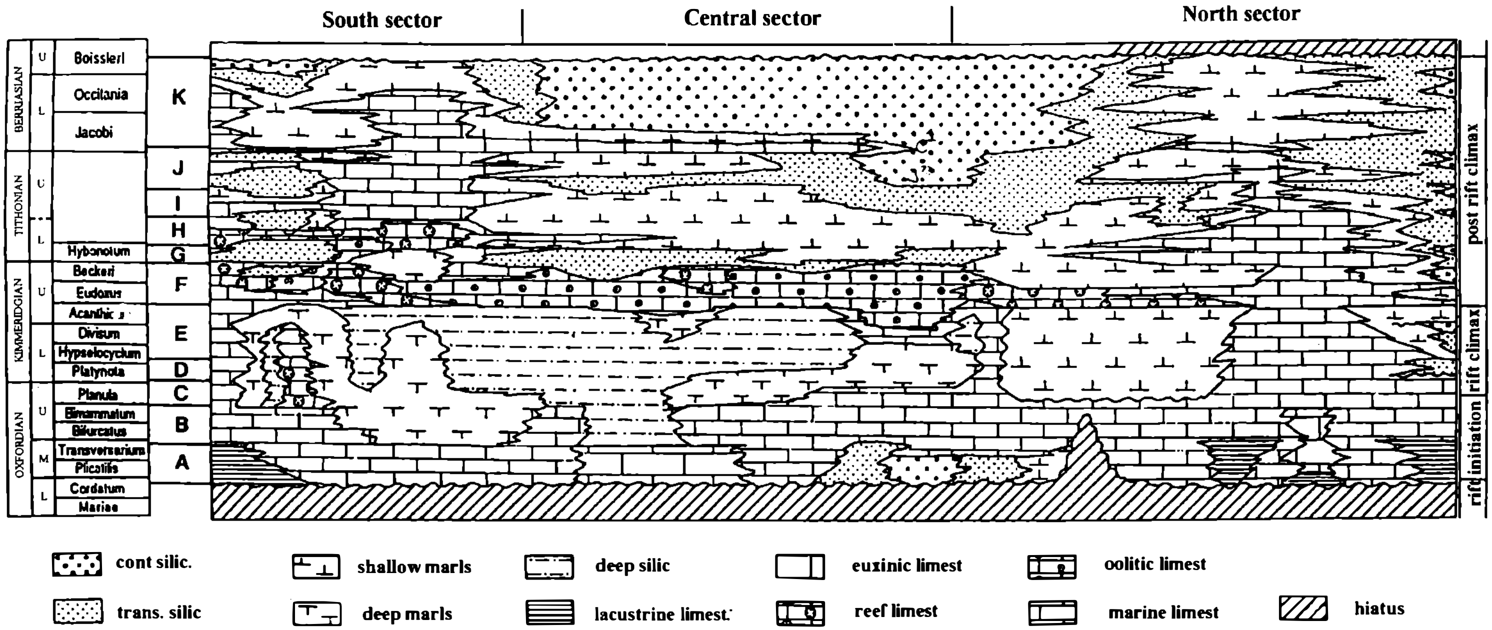
Description and Interpretation of the Depositional Sequences

Sequence A: After a long regressive hiatus that included the Callovian, sedimentation began again during the Middle Oxfordian (late Early Oxfordian?). Lacustrine carbonates (Fig. 2) are well developed in the eastern and north-eastern marginal regions. Some of these carbonates were deposited in hydrologically open shallow lakes, while others (highly bituminous) were formed in closed evaporitic lakes grading to lagoons. A shallow-marine, restricted carbonate platform is represented by a narrow N-S belt but south of Rio Maior, the environment was open marine. (Wright and Wilson, 1985). Near Cabo Mondego area and southwest of Peniche, some siliciclastics fed into the basin. Along the western border there were coastal siliciclastic systems draining to the east. To the south, sedimentation was mainly confined to carbonate systems (including lacustrine deposits) over slower subsiding blocks bounded by faults. Over the Sesimbra fault, very restricted marine environments occurred, probably due to salt flow into the structure. Subsidence of the Alentejo sub-basin appears to have been faster as that area was occupied by an open marine system. Clastic influx from the eastern margin of this sub-basin as well as along the clastic by-passing corridors related with the Arrábida and Golfinho horst was important (Fig. 5).

Sequcnce B (Upper Oxfordian): During deposition of this sequence, marine carbonate sedimentation occupies almost all the northern sector and siliciclastic sedimentation was restricted to a small area north of S. Martinho do Porto (fan delta and lagoon deposits). The outlines of the major sub-basins (Arruda, Bombarral and Turcifal) are defined (Fig. 2). Tectonic subsidence strongly increases within the resulting half-graben blocks. This definition is underlined by the settlement of an "arch" of carbonate platform depositional systems, with reef barriers to the north and east (Ramalhal, Montejunto, Ota and Monte Gordo buildups). Downslope of the major faults, carbonate and siliciclastic fans occur, e.g., Enxara do Bispo (Runa fault), Arruda, Montalegre (Vila Franca fault). In the Bombarral sub-basin, an unusual thickness of oolitic and biohermal limestone occurs (Ramalhal buildup), displaying a prograding geometry towards the southeast, following the Lourinhã-Ramalhal trend (Ellis et al., 1990; Wilson et al., 1990) (Fig. 4, M). Some sedimentation areas of the platform with reef barriers remained isolated e.g., the Ota and Monte Gordo horsts. Except in the eastern Arrábida, the relative sea-level rise allowed a wide distribution of marine carbonate systems and the retreat of clastic systems. The axis of maximum depth and subsidence seems to be located along the Barreiro fault and in the Alentejo sub-basin (Fig. 5). The main control over facies and thickness distribution was probably the activity of NNE-SSW faults.

Sequence C (Lower Oxfordian to lowermost Kimmeridgian): This sequence, which marks the rifting climax at a basinal scale, is difficult to identify in the northern sector. However, it is well represented at Cabo Mondego. In the central sector it is followed by rapidly deposited, thick terrigenous sediments. The activity of the structures (Vila Franca, Pragança, Runa and Caldas da Rainha faults, among others) that controlled the formation of the different sub-basins, lead to the definition of structural steps, bounding the depocenters. These steps allowed the genesis of significant amounts of terrigenous deposits (siliciclastic and carbonate) of submarine fans and canyons (Leinfelder and Wilson, 1989) (Fig. 4, Ca). Sedimentation seems to have proceeded from WNW (Hill, 1989) and ENE (from both margins of the basin) through two by-pass corridors. In general, over the whole southern sector of the Lusitanian Basin, a significant paleogeographical change is recorded, primarily reflecting increased water depth and progradation of clastic systems, mainly linked to Barreiro fault activity.

Figure 2. Biostratigraphic framework of the sequences described, and main facies associations occurring throughout the Upper Jurassic of Lusitanian Basin, (adapted from Pena dos Reis et al. 1996).


South sector
Central sector
North sector
BERRIASIAN
TITHONIAN
KIMMERIDGIAN
OXFORDIAN
U
L
M
Boissieri
Occitania
Jacobi
Hybonotum
Beckeri
Eudoxus
Acanthic.
Divisum
Hypselocyclum
Platynota
Planula
Bimammatum
Bifurcatus
Transversarium
Plicatilis
Cordatum
Mariae
K
J
I
H
G
F
E
D
C
B
A
post rift climax
rift climax
rift initiation
rift
cont silic.
trans. silic
shallow marls
deep marls
deep silic
lacustrine limest.
euxinic limest
reef limest
oolitic limest
marine limest
hiatus

Sequence D (Lower Kimmeridgian): In the area between Cabo Mondego and the Dourada-1C wells, an abrupt change from carbonates to siliciclastic sediments occurs in the top of sequence C, separated by a low angle unconformity (Bernardes, 1992). Other important aspects are the northward progradation of the Berlengas siliciclastic belt and the complete immersion of the Caldas da Rainha block which was, however, still acting as a barrier to the eastward progression of siliciclastic sedimentation. From both borders of the basin (central sector), large amounts of clastic sediments prograded over a decreasing accommodation space. These sediments define deltaic systems (Hill, 1989). In the southwest of this area, hemipelagic sedimentation occurs. In the Alentejo sub-basin, the south-eastward increase in thickness and transition from shallow marine to hemipelagic deposits points to a general tilting, probably controlled by the Santiago do Cacém fault.

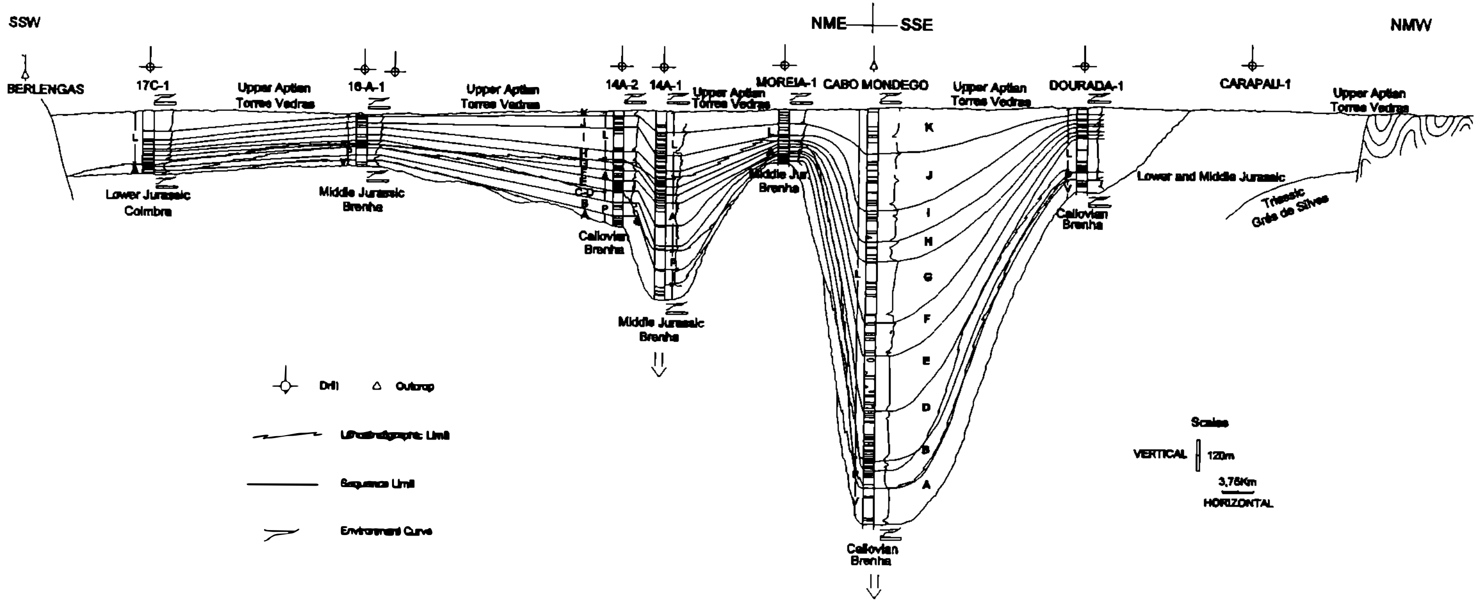
Sequence E (Lower to Upper Kimmeridgian): This sequence is similar to sequence D, displaying a more prograding geometry, and an infilling tendency due to large amounts of clastic sediments arriving into the basin. General progradation of marginal clastic systems is recorded, namely the slope siliciclastics prograding to the SE from the Turcifal and Arruda sub-basins. The axis prograded westward to the south of Barreiro and Arrábida. Compared to sequence D, the pattern of environments and isopach distribution is less complex, which is interpreted as a phase of paleogeographical homogeneity arising from a reduction of tectonic activity.

Sequence F (Upper Kimmeridgian): This sequence corresponds mainly to shallow marine deposits. They are composed of oolitic calcarenites and bioliths representing a widespread transgressive episode which caps the whole underlying succession and nearly infilled the tectonically derived accommodation space. The carbonate sediments display an onlap geometry over a major transgressive surface. They represent the marine invasion of uplifted areas such as the diapiric anticlines and the eastern horsts like Ota or Monte Gordo. Periplatform carbonate systems occupied the NNE-SSW depocenter belt west of Lisbon. Shallow marine carbonates, mainly reefs and oolitic barriers, were widespread in the central sector (Leinfelder, 1987). This facies represents the equivalent of the Amaral Formation (Fig. 4). Retrogradation of the clastic systems and a fairly uniform thickness characterises this sequence, the result of relative sea-level rise coupled with reduced tectonic activity.

Sequence G (Lower Tithonian): Relevant to the deposition of this sequence is the intense progradation of the marginal siliciclastic systems located south of Peniche. These siliciclastic systems replaced the oolitic and reefal facies belt in the northern and central sectors of the basin. At the northern border of the basin, the two marginal siliciclastic belts started to coalesce. To the south, progradation of clastic systems is more prominent on the eastern border. Progradation from the eastern part of Arrábida penetrates deeply into the Alentejo sub-basin, reducing the area occupied by shallow marine carbonates. In the basin depocenter, a shallowing trend is recorded by the onset of a shallow carbonate platform opened to the southwest (including isolated reefal buildups) and by the reduction of the slope of the ramp located south of the Sintra Massif.

Sequence H (Lower to Upper? Tithonian): This sequence is characterized by widespread retrogradation of the siliciclastic systems (except in the Alcobaça area) and the subsidence rate becomes uniform in the north sector. Most of it displays a retrogradational pattern of the clastic systems coupled with a more or less apparent restriction of carbonate systems. The deeper and more subsiding areas probably correspond to gulfs opened to the south with several axes located in the regions of Cascais, Sesimbra and the southern part of the Alentejo sub-basin. At a basin scale, the depocenter corresponds to a belt between Cascais and the Enxara do Bispo-1 well where open marine carbonate systems occur (low angle ramp to platform, including reefal facies). The transitional and continental clastic systems draining to the west still covered most of Arrábida between the Golfinho-1 well and Barreiro.

Figure 3. Correlation of sequences and lithostratigraphic units in the North sector of the Lusitanian Basin (see Fig.1 for location of sections). Strong thickness variations are shown. On the left side of each section lithostratigraphic units are identified according to the following code; Ab, Abadia Fm.; Am, Amaral Fm., C, Cabaços Fm.; Ca, Castanheira Mb.; Cb, Cabrito Mb.; Cr, Casal da Ramada Mb.; Dr, Deixa o Resto Mb.; F, Farta Pão Fm.; Fx, Freixial Mb.; L, Lourinhã Fm.; M, Montejunto Fm.; Mm, Mem Martins Fm.; P, Pholadomya Protei beds; Pc, Porto da Calada Fm.; Sb, Sobral Mb.; V, Vale Verde Fm.


SSW
NME
SSE
NMW
BERLENGAS
17C-1
Upper Aptian
Torres Vedras
16-A-1
Upper Aptian
Torres Vedras
14A-2
14A-1
Upper Aptian
Torres Vedras
MOREIA-1
CABO MONDEGO
Upper Aptian
Torres Vedras
DOURADA-1
CARAPAU-1
Upper Aptian
Torres Vedras
Lower Jurassic
Coimbra
Middle Jurassic
Brenha
Callovian
Brenha
Middle Jurassic
Brenha
Middle Jur.
Brenha
Callovian
Brenha
Callovian
Brenha
Lower and Middle Jurassic
Triassic
Grés de Silves
K
J
I
H
G
F
E
D
B
A
L
Drill
Outcrop
Lithostratigraphic Limit
Sequence Limit
Environment Curve
Scales
VERTICAL
120m
3,75Km
HORIZONTAL

Sequences I and J (Upper Tithonian to lowermost Berriasian?): The sedimentation of marls (mainly prodelta and lagoonal deposits) in a N-S trending, central belt is characteristic of these sequences. In the central sector, sequences G, H, I, J and K. (Fig. 4) (Rey, 1972; Berthou and Leereveld, 1990) record fluvial and deltaic systems and represent an overall trend of progradation and basin infilling (Hill, 1989; Bernardes and Corrochano, 1992). These systems seem to have prograded from both east and west. They gradually invaded a central axial trough, mainly at the end of sequence J and during sequence K, separating two isolated areas of marginal sedimentation. This tendency appears to have undergone a temporary inversion, during sequence H, which records a moderate eastward and westward invasion of coastal and shallow marine systems. Sequence I is relatively thin in the southern sector. The local thinning over the Barreiro (Fig. 5) and Sesimbra structures can be correlated with small vertical movements. Sequence J lacks evidence of significant

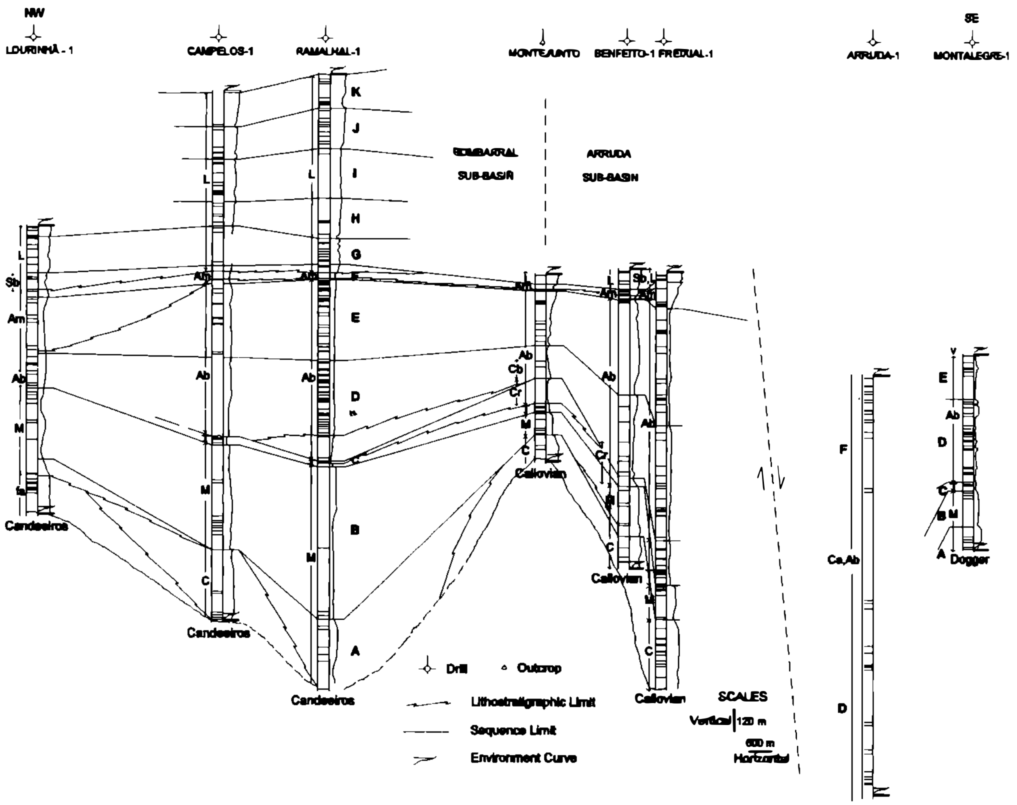
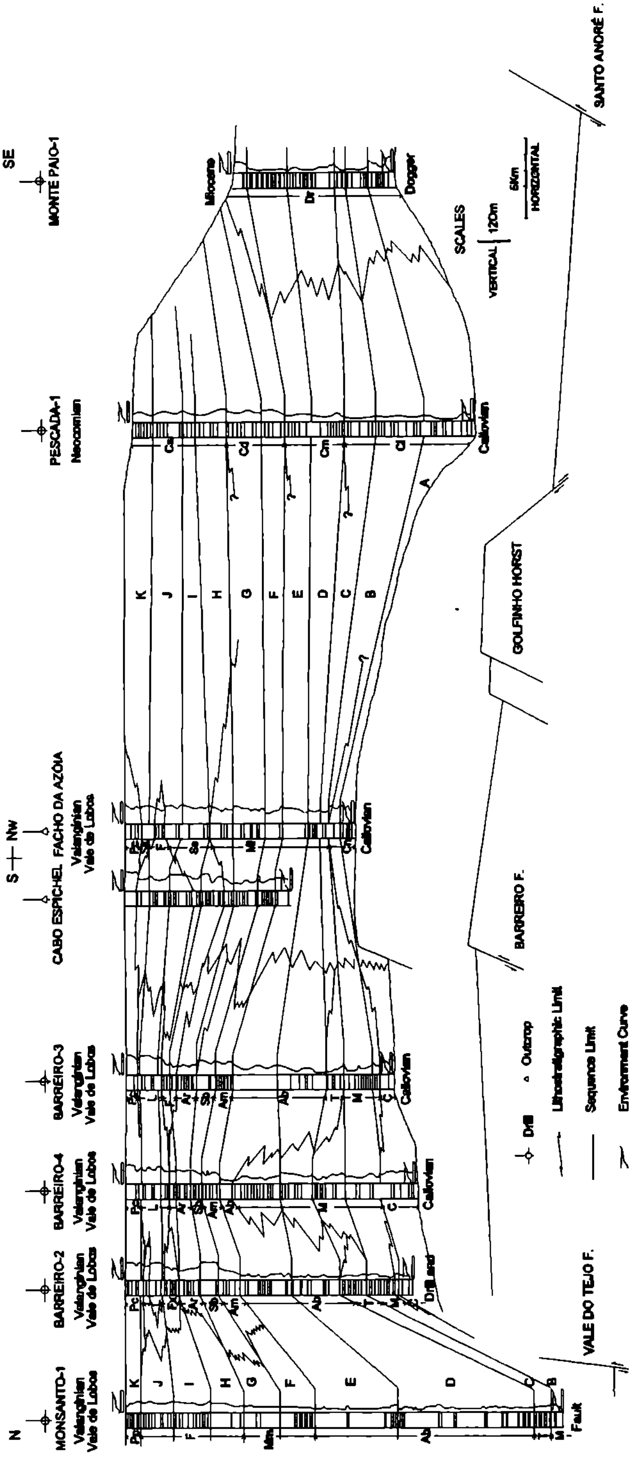


Figure 4. Correlation of sequences and lithostratigraphic units in the central sector of the Lusitanian Basin (see Fig.1 for location of sections and Fig. 3 for legend of units)

Figure 5. Correlation of sequences and lithostratigraphic units in the South sector of the Lusitanian Basin (see Fig.1 for location of sections and Fig. 3 for legend of units).


N
MONSANTO-1
Valanginian
Vale de Lobos
BARREIRO-2
Valanginian
Vale de Lobos
BARREIRO-4
Valanginian
Vale de Lobos
BARREIRO-3
Valanginian
Vale de Lobos
CABO ESPICHEL
S
Nw
FACHO DA AZÓIA
Valanginian
Vale de Lobos
PESCADA-1
Neocomian
MONTE PAIO-1
SE
Miocene
Dogger
Dr
Callovian
Drill and
Cs
Cd
Cm
Cl
Sa
Ml
Pc
F
L
Ar
Sb
Am
Mm
Ab
T
M
C
K
J
I
H
G
F
E
D
C
B
A
Fault
SCALES
VERTICAL
120m
5Km
HORIZONTAL
SANTO ANDRÉ F.
GOLFINHO HORST
BARREIRO F.
VALE DO TEJO F.
Drill
Outcrop
Lithostratigraphic Limit
Sequence Limit
Environment Curve

tectonic movements. Maximum subsidence corresponds to the Cascais-Monsanto gulf and the southwestern Alentejo sub-basin. By this time, the clastic systems prograding from the western border of the central sector (Berlengas Islands region) reached the southern sector. In the transition between the southern and central sectors, a wide WNW-ESE belt of mixed lagoon to protected platform occurs, with a gulf defined towards the Montijo-1 well. Restricted marine carbonate sedimentation continues over a wide area with homogeneous facies distribution in the west and south of the sector.

Sequence K (Lower to Upper Berriasian): The siliciclastic sedimentation covers about two thirds of the total area of the north sector, corresponding to widespread fluvial systems. In the south sector, the highstand systems tract of this sequence corresponds to a rapid regression and is composed of the more continental deposits of the Upper Jurassic-Berriasian cycle. The westward progradation of the clastic systems occupies most of the Arrábida region and an important part of the Alentejo sub-basin. An important progradation also took place towards SW sourced in the Central sector. The thickness of the sequence is fairly uniform.

Conclusions

In the extensional model of the basin, the main features are a deep detachment (Malod, 1987) and an antithetic system of faults that created the graben that constitutes the western border of the basin. The overall sequence allows recognition of the evolution of subsidence. Trends in subsidence rate are similar in all regions of the basin. The rate increased to a maximum at the Oxfordian-Kimmeridgian transition, followed by continuous decrease until the Berriasian.

We interpret this evolution as corresponding to the following tectonic stages: rift initiation, extensional rift climax and gradual transition to post-rift thermal subsidence.

Despite recognition of the tectonic events mentioned above, subsidence rates were very different across the tectonically complex basin. Structures include pull-apart displacement (central sector, Wilson et al., 1990); half-graben (Arruda and Turcifal sub-basins, and Arrábida region) and horst/graben geometries; and halokinetic uprise and withdrawal. This implies strong variations in subsidence. Pena dos Reis et al. (1997) calculated a subsidence of 1647 m/my in the Arruda sub-basin of the Central sector (Sobral well), (Fig. 4).

In detail, a distensive episode seems to have occurred near the beginning of the Late Jurassic (Early to Middle Oxfordian), involving parts of the basin (the Arrábida, the eastern border of the Alentejo sub-basin and the western border of the central sector). On the other hand, fault reactivation under distensive stress since the Early Tithonian and during the Tithonian/Berriasian transition was recorded sedimentologically on a regional scale. The decrease in subsidence rate all over the basin and a long-term eustatic fall resulted in the general progradation of marginal siliciclastic systems. The large amount of sediment accommodation space created during the rift climax was fully infilled. During the late phase of relative tectonic quiescence, eustasy probably became the fundamental control on several recorded short-term sea-level cycles. There is no evidence of significant changes in climate, which is interpreted as generally warm and dry (Hill, 1989). During the late highstand, the progradation of clastic systems increased and the limit between sequences I and J records a peak of continental influence in the Upper Jurassic to Berriasian of the basin.

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In 1992 the Gabinete para a Pesquisa e Exploração de Petróleo (GPEP) through the Instituto Pedro Nunes (IPN) (a structure linked to Coimbra University), contracted the authors to carry out studies on sequence stratigraphy and biostratigraphy of the Upper Jurassic of the Lusitanian Basin. This paper presents a summary of the main results achieved so far. The authors are grateful to GPEP and to the consultants involved in the project.

References

ATROPS, F. AND B. MARQUES. 1986. Mise en evidence de la zone à Platynota (Kimméridgian inférieur) dans le Massif du Montejunto (Portugal): conséquences stratigraphiques et paléontologiques. Geobios, Lyon, 19 (5):537-547.

____ AND ____. 1988a. Précisions stratigraphiques sur les formations à ammonites du Jurassique supérieur dans le massif du Montejunto (Nord du Tage, Portugal). 2nd International Symposium on Jurassic Stratigraphy. Lisboa. 1:505-516.

____ AND ____. 1988b. La faune d'ammonites kimméridgiène des Schistes de Ramalhão (Région de Sintra, Portugal). 2nd International Symposium on Jurassic Stratigraphy. Lisboa. 1:517-524.

BERNARDES, C. 1992. A sedimentação durante o Jurássico Superior entre o Cabo Mondego e o Baleal (Bacia Lusitana): Modelos deposicionais e arquitectura sequencial. Ph.D dissertation. Universidade de Aveiro. Portugal 261 p.

____ AND A. CORROCHANO. 1992. Sequências deposicionais do Jurássico Superior no sector ocidental da Bacia Lusitânica, Portugal. Abstracts of the III Congreso Geologico de España, Salamanca.Actas, p. 81-85.

BERTHOU, P.Y. AND H. LEEREVELD. 1990. Stratigraphic implications of palynological studies on Berriasian to Albian deposits from western and southern Portugal. Review of Palaeobotany and Palynology, Amsterdam, 66:313-344.

BRUNET, M.-F. 1992. Meso-cenozoic subsidence history of the Aquitaine basin. Comparison with neighbouring basins and margins. Abstracts Sequence Stratigraphy of European Basins, Dijon, p. 396-397.

C.P.P. 1956. Relatório geológico da sondagem Arruda-1. GPEP, ref. 859-OFF/CPP, 166p.

CANÉROT, J., J. REY, R. BATISTA, G. MANUPPELA AND B. PEYBERNÈS. 1995. Nouvelle interprétation structurale et géodynamique de la marge atlantique portugaise dans le secteur de Caldas da Rainha (Portugal). Compte Rendus Académie Sciences Paris, t. 320, série II a, 523-530.

CARIOU, E., F. ATROPS, P. HANTZPERGUE, R. ENAY, AND R. MOUTERDE. 1991. Oxfordian. 3rd International Symposium on Jurassic Stratigraphy, Poitiers, p.132.

DU CHENE, J. 1988. Etude palynologique et correlations stratigraphiques entre les sondages "Freixial 1" et "Enxara do Bispo 1", 12p. (GPEP rep. n° 22632).

ELLIS, P. M., R. C. WILSON, AND R. R. LEINFELDER. 1990. Controls on Upper Jurassic carbonate buildup development in the Lusitanian Basin, Portugal. Special. Publications International Association Sedimentologists, 9:169-202.

ERBA, E., F. CECCA, AND P. HOEDEMAEKER. 1995. Report on Early Cretaceous stratigraphy IGCP 262 Pelagic Working Group. TBC Newsletter, Utrecht, 5:7-8.

FERREIRA, M. P AND C. R. MACEDO. 1983. Igneous rocks in the diapiric areas of the western portuguese border: The K-Ar ages and settings of the upper Jurassic suite. Memorias e Noticas Museu e Laboratorio Mineralogico e Geologico Universidade de Coimbra, Portugal 96:159-181.

GEYSSANT, J. AND R. ENAY. 1991. Tithonique. 3rd International Symposium on Jurassic Stratigraphy, Poitiers, 1991, p.134.

GUÉRY, F., C. MONTENANT, AND D. VACHARD. 1986. Evolution tectono-sédimentaire du bassin portugais au Mésozoique suivant la transversale de Peniche (Estremadura). Bulletin Centre de Recherches Exploration-Production Elf-Aquitaine, 10, 1:83-94.

HANTZPERGUE, P., F. ATROPS, AND R. ENAY. 1991. Kimmeridgian. 3rd International Symposium on Jurassic Stratigraphy, Poitiers, 1991, p.133.

HAQ, B. U., J. HARDENBOL, AND P. R. VAIL. 1987. Chronology of fluctuating sea levels since the Triassic. Science, 235:1156-1166.

HILL, G. 1988. The sedimentology and lithostratigraphy of the Upper Jurassic Lourinhã Formation, Lusitanian Basin, Portugal. Ph.D dissertation, Open University, 292p.

____. 1989 Distal alluvial fan sedimentation from the Upper Jurassic of Portugal: controls on their cyclicity and channel formation. Journal Geological Society, London, 146:539-555.

LEINFELDER, R. R. 1987. Multifactorial control of sedimentation patterns in an ocean marginal basin: the Lusitanian Basin (Portugal) during the Kimmeridgian and Tithonian. Geologische Rundschau, 76, 2:599-631.

____ AND R. C. WILSON. 1989. Seismic and sedimentologic features of Oxfordian-Kimmeridgian syn-rift sediments on the eastern margin of the Lusitanian Basin. Geologische Rundschau, 78, 1:81-104.

MALOD, J.A. 1987. Dissymétrie des marges autour de l'Atlantique Nord. Relation avec les modèles de distension. Bulletin Societé geologique de France, 8, 3, 3:465-472.

MAUFFRET, A., D. MOUGENOT, P. R. MILES, AND J. MALOD. 1988. An abandoned spreading centre in the Tagus Abyssal Plain (W Portugal): results of a multichannel seismic survey. Joint Annual Meeting of Can. and Canadian Society of Petroleum Geologists, 13:A80.

MOHR, B. 1989. New palynological information on the age and environment of Late Jurassic and Early Cretaceous vertebrate localities of the Iberian Peninsula (eastern Spain and Portugal). Berliner geowiss. Abh. (A) 106, pp.291-301

____ AND D. SCHMIDT. 1987. The Oxfordian/Kimmeridgian boundary in the region of Porto de Mós (Central Portugal): stratigraphy, facies and palynology. Neues Jahrbuch fur Geologie und Paläontologie Monatshefte, Stuttgart, Abh., 176 (2), pp.245-267

MOUTERDE, R., R. ROCHA, C. RUGET, AND H. TINTANT. 1979. Faciès, biostratigraphie et paléogéographie du Jurassique portugais. Ciências da Terra, vol. 5:29-52, 20 Figs..

____, C. RUGET, AND H. TINTANT. 1973. Le passage Oxfordien-Kimméridgien au Portugal (régions de Torres-Vedras et du Montejunto. Comptes Rendue l'Académie des Sciences, Paris. vol. 277, Séries D. pp.2645-2648.

PENA DOS REIS, R.P.B., P. PROENÇA CUNHA, AND J. DINIS. 1997. Hipersubsident depositional event associated with a rift climax in Late Jurassic of Lusitanian Basin (W Portugal). IV Congreso de Jurásico de España. Alcañiz. p. 101-103.

____, J. DINIS, P. PROENÇA CUNHA, AND P. TRINCÃO. 1996. Upper Jurassic sedimentary infill and tectonics of the Lusitanian Basin (western Portugal), p. 377-386. *In* A.C. Riccardi (ed.), Advances in Jurassic Research. GeoResearch Forum, Zurich, 1-2.

PLINT, A. G. 1996. Marine and nonmarine systems tracts in four-order sequences in the Early-Middle Cenomanian, Dunvegan Alloformation, northeastern British Columbia, Canada, p. 159-191. *In* Howell and Aitken (eds.), High resolution sequence stratigraphy: Innovations and Applications, Geological Society Special Publication No. 104.

PONSOT, C. M. AND P. R. VAIL. 1991. Sequence stratigraphy of the Jurassic: new data from the Paris-London Basin. Terra; Abstract 28/48, EUGIV Congress, Strasbourg, 3,:308.

REY, J. 1972. Recherches géologiques sur le Crétacé inférieur de l'Estremadura (Portugal). Memorias Servicos Geologicos de Portugal, 21(N. S.), 477p.

RIDING, J. B. AND J. E. THOMAS. 1988. Dinoflagellate cyst stratigraphy of the Kimmeridgian Clay (Upper Jurassic) from the Dorset coast, southern England. Palynology, 12, 65-88.

WIGHTMAN, W. 1990. Micropalaeontology of Portugal: correlation of ammonite, ostracod and dinoflagellate zones of Upper Jurassic and Cretaceous age. unpublished UPPL internal report, Utrech, 24p.

WILLIS, M. G. 1988. The tectonic history of the Lusitanian Basin of Portugal. PhD. dissertation. The Open University, Milton Keynes, UK.

WILSON, R. C. 1979. A reconnaissance study of Upper Jurassic sediments of the Lusitanian Basin. Ciências da Terra (UNL), 5:53-84.

____, R. N. HISCOTT, M. G. WILLIS, AND F. M. GRADSTEIN. 1990. The Lusitanian Basin of West - Central Portugal; Mesozoic and Tertiary tectonic, stratigraphic and subsidence history. American Association of Petroleum Geologists Memoir 41:341-361.

WRIGHT, V. P. AND R. C. WILSON. 1985. Lacustrine carbonates and source rocks from the Upper Jurassic of Portugal. IAS, Abs. European Meeting.

____ AND ____. 1987. A terra rossa-like paleosol complex from the Upper Jurassic of Portugal. Sedimentology, 34:257-273.

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Jurassic Sequence Stratigraphy of East Greenland

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Abstract: The Jurassic of east Greenland provides an excellent sedimentary record of the main Mesozoic rift phase in the northern North Atlantic region. The Rhaetian-Lower Bajocian succession of east Greenland forms a pre-rift package deposited in a period of tectonic quiescence, subsidence being governed by thermal contraction following Late Permian-earliest Triassic and Early Triassic rift events. Jurassic rifting was initiated in Late Bajocian time and culminated in the Volgian. The early syn-rift deposits form a succession of backstepping marine units that reflect enhanced rates of extension. In northern east Greenland the Middle Volgian rift climax resulted in the formation of halfgrabens which became filled with coarse, clastic deep-water deposits. In southern east Greenland tilting was less intense and the basin became filled by rapid progradation of shallow marine sands in the Volgian. The Upper Rhaetian-Volgian succession is subdivided into seven tectonostratigraphic sequences that reflect major changes in tectonic style, basin configuration, drainage system, transport patterns, and depositional environments.

Introduction

The Rhaetian-lowermost Bajocian succession of onshore east Greenland can be considered a pre-rift unit and is restricted to the southern part of the 700km long, north-south trending basin (Fig. 1). Subsidence was governed mainly by thermal contraction following rift phases in Late Permian-earliest Triassic and late Early Triassic times. The succession shows an overall layer-cake geometry where both lithostratigraphic and sequence stratigraphic units are relatively sheet-like and stack in a vertically aggradational package (Dam and Surlyk, 1993; 1998). Jurassic rifting was initiated in the Late Bajocian and reached a climax in the Middle Volgian (Fig. 2). The rift phase was associated with major changes in configuration, geometry, extent of the basin, and drainage and sediment transport patterns. Large areas north and west of Jameson Land were transgressed and became part of the depositional basin. The Jurassic basin of east Greenland can conveniently be divided into southern (S), central (C) and northern segments (N). **S** includes Jameson Land and Milne Land, **C** encompasses Traill Ø, Geographical Society Ø and Hold with Hope, while **N** includes Clavering Ø, Wollaston Forland, Kuhn Ø, Hochstetter Forland and Store Koldewey (Fig. 1).

During Late Bajocian-Volgian rifting, **S** behaved as a wide coherent platform which did not break up into halfgrabens, although it underwent some tilting towards W-NW. The syn-rift package shows an overall regressive-transgressive-regressive development. In contrast, rifting in **C** and **N** culminated in the formation of narrow, deep-water halfgrabens. The early syn-rift package shows continued stepwise, transgressive backstepping, and the rift climax unit which is only exposed in **N** comprises mainly coarse-grained deep-water deposits.

The first regressive interlude in **S** and **C** after initial northwards onlap was fairly rapid, marked by strong southward progradation of sandy shelf and shoreface deposits in the Late Bajocian. Onset of transgression with large-scale backstepping was initiated in the Early Bathonian, and the sandy shelf was drowned in the early Early Callovian followed by a short regressive interlude in the late Early Callovian. All sandy depositional systems in **S** and **C** were completely drowned by the Late

Callovian. A similar trend can be recognized in **N** but the sandy shelf areas were first fully drowned by Late Oxfordian-Kimmeridgian times.

The early syn-rift phase was characterized by shallow marine sedimentation and the fill of the basin was axial from north to south. The shelf gradient in **S** became progressively distally steepened through Callovian-Volgian time, so that the physiography changed from almost flat over gently southwards sloping to a distally steepened ramp. The Volgian rift climax documented in **C** and **N** was characterized by formation of deep-water halfgrabens. They were filled mainly by conglomerates, pebbly sandstones, sandstones and subordinate mudstones deposited by a variety of sediment gravity flows (Surlyk, 1978; 1984; 1989).

The aim of the present paper is to describe both long-term tectonostratigraphic sequences (averaging 10 My) and shorter term, unconformity bounded sequences.

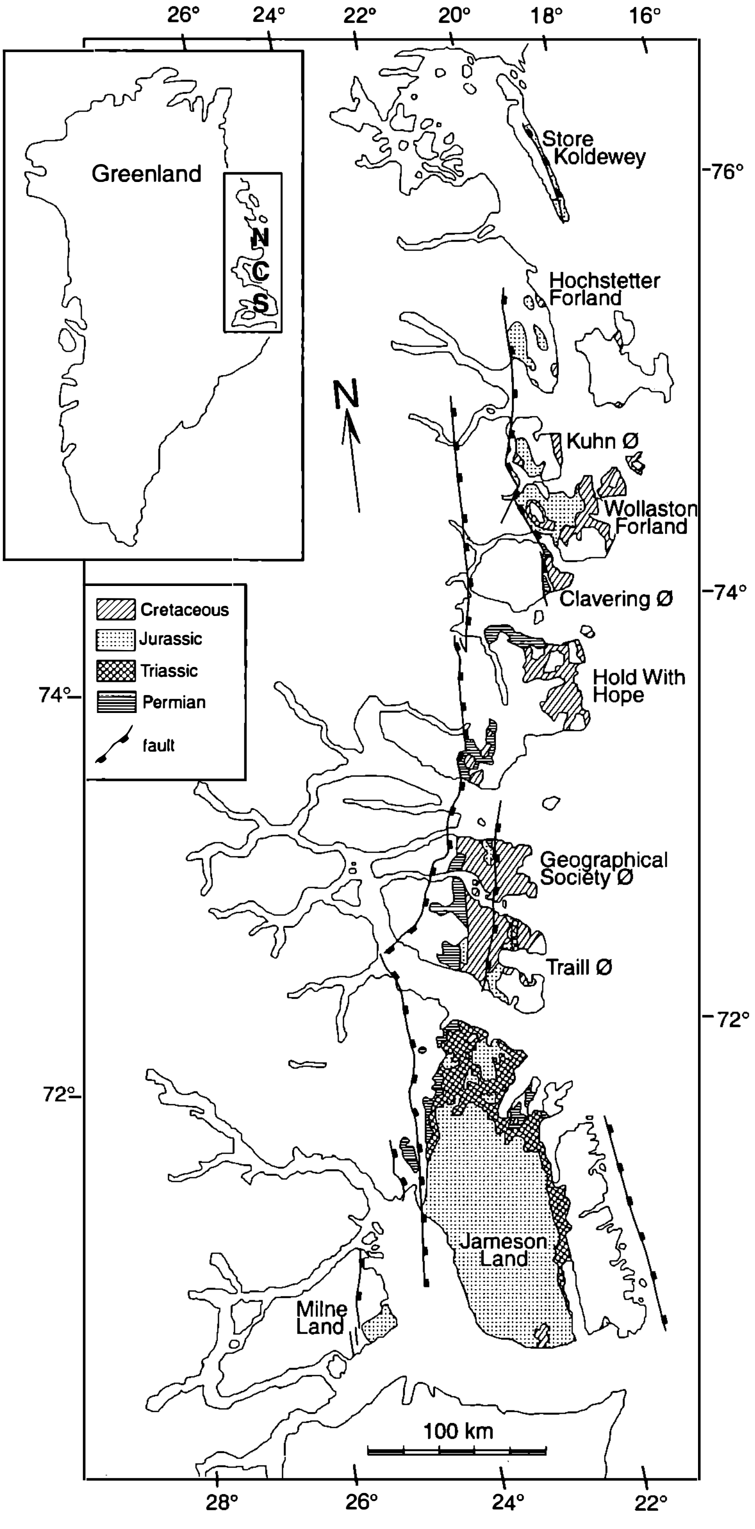
Sequence Stratigraphy

A succession of low order genetic units is recognized in the Rhaetian-Volgian succession of east Greenland (Surlyk, 1991a). They constitute the principal building blocks of the basin fill and their boundaries correspond to major changes in basin configuration, subsidence patterns and depositional environments. They are considered tectonostratigraphic sequences in the sense of Surlyk (1991b) and form the basis for description and interpretation of Jurassic basin evolution. The tectonostratigraphic sequences are comparable in concept and scale to the J0-70 sequences recognized in the North Sea by Rattey and Hayward (1993). A similar numbering system is used for the Jurassic of east Greenland to stress the similarities in stratigraphic development of the two areas and to ease comparison. They are termed JG00-JG70 where J stands for Jurassic and G for Greenland (Fig. 2). The boundaries of the tectonostratigraphic sequences are major unconformities or drowning surfaces commonly associated with erosional hiatuses or offshore condensation. They are subdivided into higher order unconformity bounded sequences (in the sense of Posamentier and Vail, 1988) controlled by changes in eustatic or relative sea-level and sediment influx. Unconformities can easily be recognized in some parts of the succession and represent conventional sequence boundaries, whereas in other parts drowning surfaces or transgressive surfaces of erosion form the most distinct key surfaces. Levels of maximum flooding can commonly be identified but are not used as sequence bounding surfaces as they do not form actual surfaces at the high resolution of field studies, but rather zones. They are situated within relatively uniform mudstone units and thus do not form natural genetic unit boundaries that are useful in field work.

Tectonostratigraphic Sequence JG00: Rhaetian-Sinemurian

The Rhaetian-Sinemurian part of the pre-rift succession was deposited under fluvio-lacustrine conditions and constitutes the lowest tectonostratigraphic sequence, JG00. It contains rich and well preserved floras but the dating is not very precise (Harris, 1937). The lower part belongs to the Lower Rhaetian *Lepidopteris* Zone and the upper part to the Hettangian *Thaumatopteris* Zone. Sinemurian deposits occur in the basin centre as shown by spores and pollen. The marked break between the two floras is accompanied by a change in fluvial facies and may represent a major hiatus. It is possible, however, that the break is real and represents the end-Triassic mass extinction. Lacustrine sedimentation was markedly cyclic showing at least 50 rapid alternations between black, deep water shales and shallow water to terrestrial, delta sheet sands. The cycles are interpreted as reflecting dramatic lake level changes controlled by climatic fluctuations (Dam and Surlyk, 1992; 1993). The higher order lacustrine sequences have no relation to contemporaneous sea-level cycles. However, a long term trend can be recognized which seems to parallel the third order sea-level cycles proposed by Haq et al. (1987) and Hallam (1992). This opens the possibility that sea-level fluctuations exerted some control on base level in the low-lying lake system (Dam and Surlyk, 1993).

Figure 1. Geological sketch map of east Greenland showing distribution of the Permian-Cretaceous deposits and the main faults active during the Jurassic. **S**, **C** and **N** in the inset map indicate the southern, central and northern segments of the Jurassic basin, respectively.


26°
24°
22°
20°
18°
16°
Greenland
N
C
S
Store
Koldewey
-76°
Hochstetter
Forland
N
Kuhn Ø
Wollaston
Forland
Cretaceous
Jurassic
Triassic
Permian
fault
-74°
Clavering Ø
74°
Hold With
Hope
Geographical
Society Ø
Traill Ø
-72°
72°
Jameson
Land
Milne
Land
100 km
28°
26°
24°
22°

Tectonostratigraphic Sequence JG10: Pliensbachian-Lower Bajocian

The lacustrine basin and surrounding low-lying flood plains of JG00 time underwent marine flooding in the Early Pliensbachian. The Pliensbachian-Early Bajocian part of the pre-rift package shows repeated alternations between shallow marine sandstones and deeper marine heteroliths and mudstones. Erosional unconformities, drowning surfaces and marine transgressive surfaces of erosion are well developed, whereas the units bounded by these surfaces show little internal organization in terms of stacked parasequences or overall vertical grain-size trends (Dam and Surlyk, 1998).

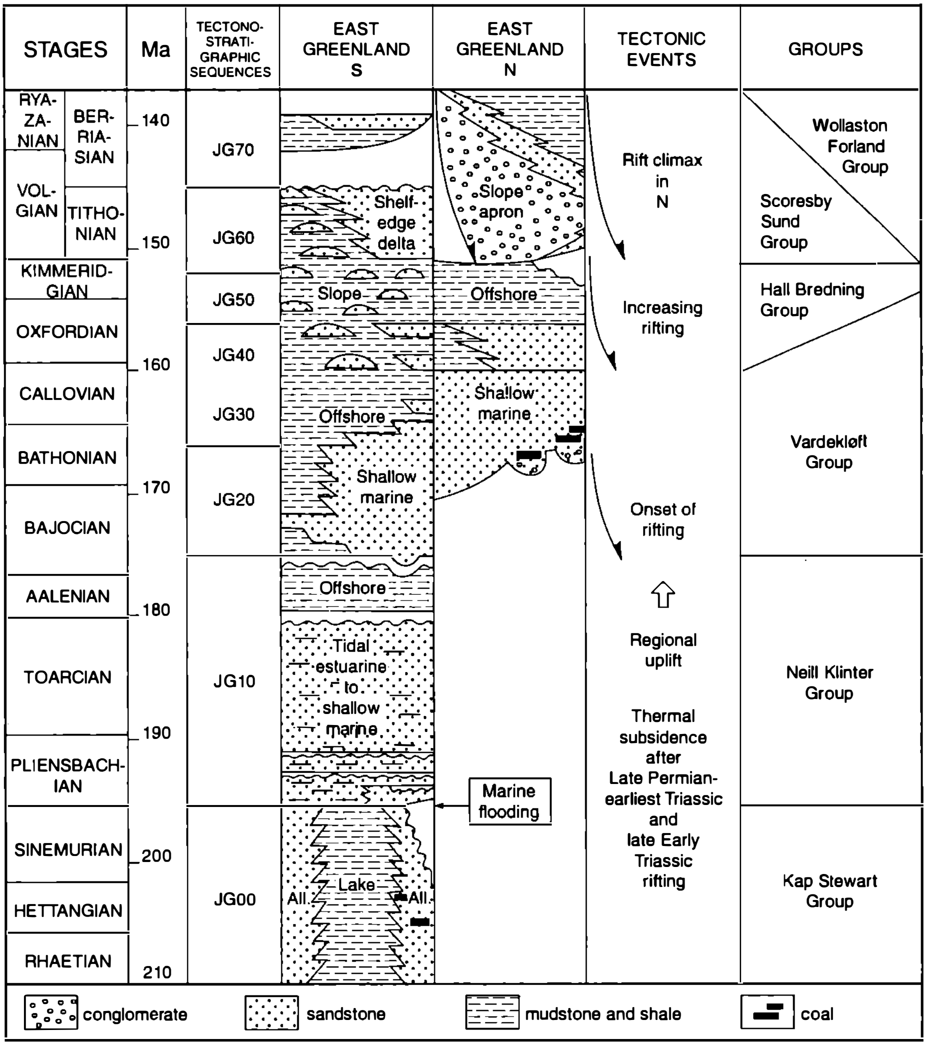


Figure 2. Summary of tectonostratigraphic sequences of the Upper Rhaetian-Jurassic of east Greenland. The lithostratigraphic scheme is based on a recent revision by Surlyk et al. (1998).

Key surfaces are thus in most cases straightforward to identify, whereas systems tracts can only be tentatively interpreted. JG10 consists of seven sequences: three in the Pliensbachian, one straddling the Pliensbachian-Toarcian boundary, one in the uppermost Toarcian-lowermost Aalenian, and one in the Aalenian-Lower Bajocian (Dam and Surlyk, 1998). The age control is poor except for a few levels, and the sequence stratigraphic subdivision was undertaken mainly on the basis of identification of key surfaces, facies analysis and mapping of individual units. The Pliensbachian *U. jamesoni* and lower *P. davoei* and the Lower Toarcian *D. tenuicostatum* Chronozones are well documented. Additional age information is provided by dinoflagellates.

Tectonostratigraphic Sequence JG20: Upper Bajocian-Lower Bathonian

Jurassic rifting was initiated in the Late Bajocian and was associated with major changes in basin configuration, drainage and marine transport patterns. The sequence stratigraphic interpretation of the early syn-rift deposits is based mainly on the work of Vosgerau (1997), Alsgaard et al. (in press), Engkilde and Surlyk (in press), Larsen et al. (in press), and Larsen and Surlyk (in press).

Large areas north of **S** were transgressed, and **C** and **N** became depositional areas for the first time in the Jurassic. The base of JG20 is a major unconformity, and an onlap surface in western **S**, **C** and **N**. JG20 thus rests on progressively older rocks northward, ranging from Triassic in **C** over Upper Permian in southern **N**, to crystalline basement in northern **N**. Deposition in **C** was initiated with fluvial pebbly sandstones probably filling incised valleys. The fluvial succession is topped by coaly shales reflecting a base level rise heralding Late Bajocian marine transgression. The rest of the Jurassic was characterized by marine conditions throughout east Greenland. Terrestrial deposits are thus unknown except for the basal part of the succession in northern **N**, which shows a similar development of fluvial conglomerates topped by a coal-bearing paralic succession. This unit is, however, younger, being of Late Bathonian-Early Callovian age reflecting the northwards onlap.

Late Bajocian marine deposition was initiated by southwards progradation of marine sand and silt forming the bulk of the early syn-rift deposits. The key depositional motif in northern **S**, **C** and **N** is a sand-dominated, coarsening-upward unit topped by a drowning surface, with a lag of pebbles and marine fossils, notably ammonites. The distal southern equivalents comprise coarsening-upward units of mudstone, siltstone and very fine-grained, commonly glauconitic sandstone.

The sequence stratigraphic nature of the units is somewhat ambiguous because direct evidence for sea-level fall is mainly absent. They may, thus, be parasequences topped by marine drowning surfaces, or they may be partial sequences where the pebble lags represent reworked coarse-grained beach and fluvial deposits. The coarsening-upward units are here interpreted as such partial sequences or systems tracts. Stacked systems tracts in any section thus commonly belong to different successive sequences. The southern areas of **S** are volumetrically dominated by lowstand and transgressive systems tracts, whereas the northern areas of **S** and **C** are dominated by highstand systems tracts (Engkilde and Surlyk, in press). This interpretation is based on the character of the lags, lateral thickness changes of the units, and detailed ammonite correlation. The low inclination of the basin axis implies long distance shoreline migration with even minor changes in relative sea level or sediment influx and separation of succeeding systems tracts over tens to several hundred kilometres (Heinberg and Birkelund, 1984; Engkilde and Surlyk, in press). The systems tracts of each sequence are thus commonly laterally offset in a north-south direction over tens of kilometres. The succession in **C** is markedly thinner than in northern **S** and represents a lowstand sediment bypass area. This corroborates the interpretation of the coarsening-upward units as representing partial sequences separated by coalesced sequence boundaries and marine ravinement surfaces.

Four sequences are recognized in JG20. The two oldest represent maximum lowstand progradation to the southernmost part of **S** (*C. borealis, C. indistinctus, C. pompeckji* and lower *A. arcticus* Chronozones). The age determinations of the Middle Jurassic succession follow Callomon (1993). They are topped by drowning surfaces and the associated transgressive and highstand deposits are restricted to more proximal areas in northern **S** and **C**. The succeeding two sequences form an aggradational stack reflecting a change from the initial progradation to long-term backstepping and eventual complete drowning of the sandy depositional system (upper *A. arcticus, A. greenlandicus* and lower *A. ishmae* Chronozones).

Tectonostratigraphic Sequence JG30: Upper Bathonian-Upper Callovian

The succeeding three sequences show large-scale Late Bathonian-Early Callovian northward backstepping over several hundred kilometres (upper *A. ishmae, A. cranocepholoide, C. variabile, C. calyx, C. apertum, C. nordenskioldi, K. koenigi* and *S. calloviense* Chronozones). The sandy depositional system was eventually drowned and draped with offshore muds which show condensed levels and hiatuses in the most distal offshore parts of the basin in southern **S**. A short regressive pulse took place at the end of the Early Callovian (*S. calloviense* Chronozone) marked by southwards progradation of a sandy wedge to central **S**, but renewed Middle Callovian sea-level rise caused drowning of the sandy system (*K. jason* and *E. coronatum* Chronozones). The sequential development can also be recognized in **N** but the preserved sediments are more proximal and sand-dominated. Drowning events are less clearly marked and biostratigraphic age control less precise (Alsgaard et al., in press).

Backstepping was probably governed by increased rate of extension and onset of gentle block tilting acting in concert with a late Middle Jurassic eustatic sea-level rise. In **C** and **N** the main depositional motif is a stepwise backstepping of progradational sandstone-dominated packages separated by drowning surfaces.

Tectonostratigraphic Sequence JG40: Uppermost Callovian-Middle Oxfordian

Renewed regressive phases following the Early and Middle Callovian drowning events took place in latest Callovian and Middle Oxfordian time when sand bodies prograded southwards to reach the shelf edge in central **S**.

Two sequences are recognized in this part of the succession (Larsen and Surlyk, in press). The oldest is of latest Callovian (*P. athleta* Chron) and possibly earliest Oxfordian age, and is marked by incoming of massive base-of-slope sandstones shed from the front of a shelf-edge delta. The rather uniform massive sandstones belong to an undifferentiated late highstand-falling stage-lowstand systems tract, and a sequence boundary cannot be identified. They are topped by a drowning surface which probably passes updip into the transgressive surface (correlative strata removed by modern erosion). The overlying shale unit contains a maximum flooding level of Early Oxfordian age (*Q. mariae* Chron). The shales are overlain by the second unit of massive base-of-slope sandstones, probably of late highstand origin, sharply overlain by shallow marine delta front sandstones of the falling stage and lowstand systems tract (Lower-Middle Oxfordian, *Q. mariae-C. densiplicatum* Chronozones). Key surfaces are difficult to identify and a sequence stratigraphic interpretation is hampered by the relatively deep-water nature of most of the succession.

The Late Callovian and Middle Oxfordian regressive pulses can also be recognized in **C** and **N** where the succession is of shallow marine nature throughout. Progressive northward transgression and onlap first reached the northern part of **N** in Late Bathonian-Early Callovian times, marked by fluvial conglomerates filling an incised valley system overlain by thick coal beds, reflecting base level rise and formation of peat swamps at the top of the almost completely filled valley system. The coal-bearing succession can be subdivided into four sequences, each beginning with a thick coal-seam resting on a rooted paleosol (Petersen et al., 1998). Peat accumulation took place during onset of base level rise and the coal beds represent the lower part of the transgressive systems tract. Dulling-upward cycles in the coal-beds represent parasequences or possibly higher order sequences. The overlying shoreface sandstones form the upper part of the transgressive and highstand systems tracts.

Tectonostratigraphic Sequence JG50: Upper Oxfordian-Upper Kimmeridgian

The shelf-edge deltas of JG40 form the top of the section in central **S**, and younger deposits of JG50 exposed in southern **S** only include base-of-slope and basin plain deposits. They comprise a thick unit of black shales with large bodies of massive sandstones which have been strongly modified by post-burial liquefaction and intrusion into the surrounding shales (Surlyk, 1987; Surlyk and Noe-Nygaard, 1998). The Upper Oxfordian part of the succession has roughly equal amounts of shale and sandstone, whereas the Kimmeridgian part is dominated by massive sandstones. Otherwise, the succession shows no vertical or lateral trends and is completely chaotic. Key surfaces cannot be identified, and a sequence stratigraphic breakdown has not been possible. A correlative offshore shelf succession is, however, exposed in Milne Land at the western basin margin of **S** (Birkelund et al., 1984). Late Oxfordian drowning can also be recognized in this area but was abruptly interrupted in mid Late Oxfordian time by a prograding shallow marine sand body (Fürsich and Heinberg, 1983). The base of the upward-coarsening sand body is a possible regressive surface of erosion formed during falling sea-level, and the sandstone is topped by a sharp transgressive surface of erosion formed during renewed sea-level rise in the Late Oxfordian (*A. regulare* Chron). Maximum flooding was reached in mid Late Kimmeridgian time (*A. eudoxus* Chron) and persisted to the end of the earliest Volgian *P. elegans* Chron, interrupted by mild regression in the latest Kimmeridgian (probably *A. autissiodorensis* Chron). The time interval was characterized by deposition of black, laminated, organic-rich shales and represents the highest relative sea-level in the Jurassic of **S**.

Correlative deposits are poorly preserved in **C** where they subcrop strongly erosional mid-Cretaceous unconformities, but seem to show the same pattern as in **S**. In **N** backstepping continued, and drowning surfaces separate progressively finer-grained and more offshore packages. The maximum flooding interval comprising the *A. eudoxus-P. elegans* Chronozones is very thickly developed and consists of black, laminated shales.

Sequence stratigraphic interpretation is hampered by the offshore mudstone-dominated nature of the succession. The dominant vertical motif is a slightly coarsening-upward unit topped by a drowning surface which is erosional in some cases. The cycles may represent parasequences or distal partial sequences.

Tectonostratigraphic Sequence JG60: Lowermost Volgian-Upper Volgian

Rapid, large-scale progradation characterized the end of the Jurassic in **S**, and the succession along the western basin margin allows precise dating of the onset of relative sea-level fall to the Early Volgian *P. wheatleyensis* Chron (Birkelund et al., 1984). The black, maximum flooding shales of JG60 are sharply overlain by coarsening-upwards, muddy fine-grained sandstones of late Early-Middle Volgian age, and the boundary may represent sea-level fall and associated forced regression. The succession becomes glauconitic at the top, and coarse-grained sandstones occur at several levels. It is topped by a sharp drowning surface of Middle Volgian (*C. anguinus* Chron).

Drowning was followed by a succession of progradational pulses represented by coarse-grained, coarsening-upward sandstones of Middle Volgian age, separated by drowning surfaces which represent major hiatuses. The largest hiatus occurs between sandstones of the Middle Volgian *L. groenlandicus* Chronozone and mid-Valanginian *Tollia*-bearing sandstones (Birkelund et al., 1984). Correlative deposits are extremely well exposed and thickly developed in central **S**, where they form a forestepping stack of coarse-grained, pebbly shelf-edge sandstones (Surlyk and Noe-Nygaard, 1991; 1995). The deposits along the western basin margin are generally finer grained and more marine than those of the basin centre. The former represent highstand deposits, while the latter were deposited during major regressions which reached the shelf-slope break close to the axis of the basin during late highstand, falling stage and lowstand. There is, thus, a spatial separation between early highstand and correlative late highstand-lowstand deposits in the order of 100km . Thin transgressive sandstones and mudstones occur between the regressive shelf-edge sand bodies.

Correlative deposits are not known from **C**, but they are well exposed in **N** and of a totally different type. This area underwent major block faulting in the Middle Volgian with formation of deep-water halfgrabens. Initial deposition was mainly lateral with formation of conglomeratic, scarp-derived taluses, slope aprons and coalesced submarine fans (Surlyk, 1978; 1984; 1989). Sediment transport in the half grabens became mainly axial, with deposition of conglomerates and pebbly sandstones by turbidity currents and related processes. The dominant motif is a fining-upward cycle, tens of metres to some hundreds of metres thick, probably reflecting a major fault episode and associated deepening. The cycles may be correlative to the contemporaneous shallow marine sequences recognized in **S**.

Conclusions

1. The Rhaetian-Volgian succession of east Greenland comprises a pre-rift package of Rhaetian-Early Bajocian age and a syn-rift package of Late Bajocian-Volgian age.
2. Seven tectonostratigraphic sequences are recognized. They correspond to major changes in tectonic style, basin configuration, drainage and transport patterns, and depositional environments. The boundaries are major unconformities or drowning surfaces.
3. The Rhaetian-Sinemurian part of the pre-rift succession is fluvio-lacustrine and the large number of high-order sequences probably result from climatically-controlled lake-level changes. Long-term lake-level changes may on the other hand reflect sea-level controlled base-level changes.
4. The Pliensbachian-Lower Bajocian paralic-shallow marine part of the pre-rift succession shows unconformites and ravinement surfaces, whereas well-developed stacking patterns are mainly poorly developed. Sequences are accordingly well defined, whereas systems tracts are difficult to identify.

5. The early syn-rift package of Late Bajocian-Early Kimmeridgian age reflects major expansion and reorganization of the basin. Deposition was fully marine with exceptions of basal fluvial deposits in a few areas. Stacking patterns are well developed, but differentiation between unconformities, ravinement and drowning surfaces is difficult. It is thus in many cases an open question whether the basic coarsening-upward units are parasequences or partial sequences, or in other words whether their top surfaces reflect drowning only or sea-level fall followed by drowning.
6. The Volgian rift-climax succession is differently developed in the southern and northern parts of the basin. The southern part behaved as a coherent platform and underwent only minor tilting. Maximum Early Kimmeridgian drowning was thus followed by rapid Volgian progradation. Sand was deposited at the basin margin during highstand, at the basin centre during late fall and lowstand, and as a reworked drape during rise. Highstand and lowstand deposits of the same sequence are thus separated by tens of kilometres. In the northern part of the basin deep halfgrabens were formed during Volgian rift climax and filled with thick conglomerates, pebbly sandstones, sandstones and minor mudstones deposited from sediment gravity flows. The main motif is a fining-upward unit many tens of metres thick, probably reflecting major phases of down-faulting and deepening. Sequence stratigraphic key surfaces cannot be identified due to the deep-water nature of the succession. It is, however, possible that the units may be correlative to the contemporaneous shallow marine sequences in the southern parts of the basin and that their boundaries correspond to sequence boundaries.

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References

ALSGAARD, P.C., V.L. FELT, H. VOSGERAU, AND F. SURLYK. In press. The Jurassic of Kuhn Ø, North-East Greenland. *In* F. Surlyk and J.R. Ineson (eds.), The Jurassic of Denmark and Greenland. Geological Survey of Denmark Bulletin.

BIRKELUND, T., J.H. CALLOMON, AND F.T. FÜRSICH. 1984. The stratigraphy of the Upper Jurassic and Lower Cretaceous sediments of Milne Land, central East Greenland. Grønlands geologiske Undersøgelse Bulletin, 153:56 p.

CALLOMON, J.H. 1993. The ammonite succession in the Middle Jurassic of East Greenland. Bulletin of the Geological Society of Denmark, 40:83-11

DAM, G., AND F. SURLYK. 1992. Forced regressions in a large wave- and storm-dominated anoxic lake, Rhaetian-Sinemurian Kap Stewart Formation, East Greenland. Geology, 20:749-752.

____, AND ____. 1993. Cyclic sedimentation in a large wave- and storm-dominated anoxic lake; Kap Stewart Formation (Rhaetian-Sinemurian), Jameson Land, East Greenland, p. 419-448. *In* Special Publications International Association of Sedimentologists, 18.

____, AND ____. 1998. Stratigraphy of the Neill Klinter Group; a Lower-lower Middle Jurassic tidal embayment succession, Jameson Land, East Greenland. Geology of Greenland Survey Bulletin, 175:80 p.

ENGKILDE, M., AND F. SURLYK. In press. Depositional environments and sequence stratigraphy of shallow marine siliciclastics of the Pelion and Fossilbjerget Formations, Middle Jurassic, East Greenland. *In* Surlyk, F. and J.R. Ineson (eds.), The Jurassic of Denmark and Greenland. Geological Survey of Denmark Bulletin.

FÜRSICH, F.T., AND C. HEINBERG. 1983. Sedimentology, biostratinomy, and palaeoecology of an Upper Jurassic offshore sand bar complex. Bulletin of the Geological Society of Denmark, 32:67-95.

HALLAM, A. 1992. Phanerozoic sea-level changes. Columbia University Press. New York, 266 p.

HAQ, B.U., J. HARDENBOL, AND P.F. VAIL. 1987. Chronology of fluctuating sea levels since the Triassic. Science, 235:1156-1167.

HARRIS, T.M. 1937. The fossil flora of Scoresby Sound East Greenland. 5. Stratigraphic relations of the plant beds. Meddelelser om Grønland, 112, 2:1-114.

HEINBERG, C., AND T. BIRKELUND. 1984. Trace fossil assemblages and basin evolution of the Vardekløft Formation (Middle Jurassic, East Greenland). Journal of Paleontology, 58:362-39

____, S. PIASECKI, AND F. SURLYK. In press. Dynamic stratigraphy of a basement-onlapping shallow marine sandstone succession, the Charcot Bugt Formation, Middle to Upper Jurassic, East Greenland. *In* F. Surlyk and J.R. Ineson (eds.), The Jurassic of Denmark and Greenland. Geology of Denmark Survey Bulletin.

LARSEN, M., AND F. SURLYK. In press. A shelf edge delta and slope system of the Upper Callovian-Middle Oxfordian Olympen Formation, East Greenland. *In* F. Surlyk and J.R. Ineson (eds.), The Jurassic of Denmark and Greenland. Geology of Denmark and Greenland Survey Bulletin.

PETERSEN, H.I., J.A. BOJESEN-KOEFOED, H.P. NYTOFT, F. SURLYK, J. THERKELSEN, AND H. VOSGERAU. 1998. Relative sea-level changes recorded by paralic liptinite-enriched coal facies cycles, Middle Jurassic Muslingebjerg Formation, Hochstetter Forland, Northeast Greenland. International Journal of Coal Geology, 36:1-30.

POSAMENTIER, H. W., AND P. F. VAIL. 1988. Eustatic controls on clastic deposition II - sequence and systems tract models. SEPM Special Publication, 42:125-154.

RATTEY, R.P., AND A.B. HAYWARD. 1993. Sequence stratigraphy of a failed rift system: the Middle Jurassic to Early Cretaceous basin evolution of the Central and Northern North Sea, p. 215-250. *In* J.R. Parker (ed.), Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference. Published by the Geological Society London.

SURLYK, F. 1978. Submarine fan sedimentation along fault scarps on tilted fault blocks (Jurassic-Cretaceous boundary, East Greenland). Grønlands Geologiske Undersøgelse Bulletin, 128:108 p.

____. 1984. Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston Forland Group, East Greenland, p. 359-382. *In* E. H. Koster and R. J. Steel (eds.) Sedimentology of Gravels and Conglomerates. Canadian Society of Petroleum Geologists, Memoir 10.

____. 1987. Slope and deep shelf gully sandstones, Upper Jurassic, East Greenland. American Association of Petroleum Geologists Bulletin, 71:464-475.

____. 1989. Mid-Mesozoic syn-rift turbidite systems: controls and predictions, p. 231-241. *In* J. D. Collinson (ed.), Correlation in Hydrocarbon Exploration. Norwegian Petroleum Society, Graham and Trotman, London.

____. 1991a. Sequence Stratigraphy of the Jurassic-Lowermost Cretaceous of East Greenland. American Association of Petroleum Geologists Bulletin, 75:1468-1488.

____. 1991b. Tectonostratigraphy of North Greenland. Bulletin Grønlands geologiske Undersøgelse, 160:27-47.

____, G. DAM, M. ENGKILDE, M. LARSEN, N. NOE-NYGAARD, S. PIASECKI, J. THERKELSEN, AND H. VOSGERAU. In press. Jurassic stratigraphy of East Greenland. *In* F. Surlyk and J. R. Ineson (eds.) The Jurassic of Denmark and Greenland. Geological Survey of Denmark Bulletin.

____, AND N. NOE-NYGAARD. 1991. Sand bank and dune facies architecture of a wide intracratonic seaway: Late Jurassic-Early Cretaceous Raukelv Formation, Jameson Land, East Greenland, p. 261-276. *In* A.D. Miall and N. Tyler (eds.), The three-dimensional facies architecture of terrigenous clastic sediments and its implication for hydrocarbon discovery and recovery. Society of sedimentology and palaeontology 3.

____, AND ____. 1995. High-angle clinoform beds - a recurrent architectural element in Jurassic shallow marine deposits of East Greenland. *In* D. Hunt and M.E. Tucker (eds.), Sedimentary responses to forced regressions: recognition, interpretation and reservoir potential. Abstract volume. The Geological Society of London, 2 p.

____, AND ____. 1998. Massive intrusive sandstones, Upper Jurassic Hareelv Formation, East Greenland. Sediment transport and deposition by particulate gravity currents. Programme & Abstract Volume. University of Leeds 7-9 September 1998, p. 52.

VOSGERAU, H. 1997. Depositional environments and sequence stratigraphy of a Middle-Upper Jurassic early syn-rift succession deposited in a low-gradient epeiric seaway, the Wollaston Forland Basin, Northeast Greenland. Unpublished Ph.D. thesis, 3 parts, University of Copenhagen, 110 p.

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Hispanic Corridor: Its Evolution and the Biogeography of Bivalve Molluscs

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Abstract: An updated set of biogeographic data for bivalve molluscs is analysed to contribute to the understanding of the development of the Hispanic Corridor during Late Triassic to Middle Jurassic times. The succession of Simpson similarity coefficients between the western Tethys (*s.s.*) and several cratonic regions of the eastern Pacific shows "pulses" of increased similarity during that time, prior to the Bajocian. These range from Sinemurian to Toarcian according to the areas considered. Conversely, areas located within the same regions show only minor fluctuations. Since not all ecological types are equally represented in the biotic interchange along the Corridor, a shallow intermittent marine connection which preceded rifting by millions of years is indicated. Benthonic bivalve taxa, and especially epifaunal stocks, some of which can be followed "step by step", show this pattern clearly. Data suggest that this connection acted as a filter, allowing the passage of on-shore species while being an effective barrier for off-shore species.

Introduction

The opening of the North Atlantic, which triggered fragmentation of Pangea and establishment of a marine connection between the western Tethys and eastern central Pacific (Hispanic Corridor), is one of the most important paleogeographic events which occurred during the Jurassic. New biogeographic, geologic, palaeomagnetic and geophysical data are constantly providing rich arguments for the discussion of this event. Among them, biogeography is one of the most powerful tools to understand its nature and timing.

The distribution patterns of bivalves during the Late Triassic–Middle Jurassic interval are now well-known, both along the eastern Pacific (Damborenea, 1996; Aberhan, 1993; 1994; 1998a; 1998b and references therein) and the western Tethys (Hallam, 1976; 1977; Liu, 1995; Liu et al., 1998 and references therein). The distribution of certain bivalves was used to propose the hypothesis of the existence of a shallow marine connection (now called Hispanic Corridor) since Early Jurassic time (Damborenea and Manceñido, 1979) and has been widely discussed since then on the basis of the distribution of bivalves and other marine invertebrates (Hallam, 1983; Smith and Tipper, 1986; Newton, 1988; Smith, 1983; 1989; Smith et al., 1990; Riccardi, 1991; Elmi, 1993).

This hypothesis has been challenged by some authors on the basis of lack of sedimentologic evidence. Along this line, Newton (1988) and Elmi (1993) explained the common occurrence of mollusc species and genera on both sides of the Corridor using alternative hypotheses.

In this paper an updated set of biogeographic data for bivalve molluscs is presented and analysed in order to understand the establishment and development of the Hispanic Corridor during Late Triassic to Middle Jurassic times.

Materials and Methods

The quality of the data base is critical in biogeographic approaches such as this, and the problems arising from the nature of paleobiogeographic data sets have been extensively discussed in the past (Henderson and Heron, 1977). A large amount of the data used here was compiled from papers published over more than a century, and is necessarily uneven. Most published information was re-evaluated, and for this reason the data were treated only at generic/subgeneric level. This admittedly results in loss of information but is common practice in paleobiogeographic papers using bivalve data (Liu, 1995; Liu et al., 1998). The distribution of 237 bivalve genera and subgenera was recorded for 16 geographic areas from the following sources:

South-western Tethys

(1) southern France, (2) Iberian Peninsula (Spain + Portugal) and (3) Morocco and Algeria. References included in Liu, 1995; also Dumortier, 1864-1874; Dareste de la Chavanne, 1920; Dubar, 1925; 1931; Hallam, 1971; Calzada, 1981; 1983; 1986; Mouterde et al., 1981; 1983. Comprehensive revisions of certain bivalve groups have proven especially useful, such as those by Johnson (1984) and Muster (1995).

South America

(4) Colombia, (5) Perú, (6) northern Chile and (7) Argentina + central Chile. References included in Damborenea, 1996; also Körner, 1937; Cox, 1949; 1956; Geyer, 1973; Hayami et al., 1977; Leanza, 1993; Pérez and Reyes, 1994; Romero et al., 1995; Aberhan and Hillebrandt, 1996; Riccardi et al., 1997; Rubilar, 1998; and unpublished data.

"Cratonic" North America

(8) Mexico (mostly "Antimonio terrane"), (9) Sonomia terrane + cratonic USA and (10) North America Boreal craton. References included in Smith et al., 1994; also McLearn, 1924; Smith, 1927; Jaworski, 1929; Burckhardt, 1930; Silberling, 1959; Imlay, 1964; 1967; Laws, 1982; Poulton, 1991; McRoberts, 1992; Tamura *in* Stanley et al., 1994; Aberhan and Muster, 1997; Aberhan, 1998a; 1998b; McRoberts, in press; Damborenea and González León, in press.

Apart from these regions directly related to the question of the Hispanic Corridor, data from Japan were analysed to test the behaviour of bivalve distribution across the Pacific. Although not directly involved in the main subject, data from several suspect terranes of the western North American margin were likewise included.

Time interval	S France	Iberian Peninsula	Morocco	Colombia	Peru	N Chile	W Argentina + central Chile	Mexico	Sonomia + cratonic USA	N America Boreal craton	Wallowa terrane	Wrangellia terrane	Stikinia terrane	Quesnellia terrane	Cadwallader terrane	Japan	Total number of taxa
Bajocian	12	15	73	0	7	9	54	6	37	8	2	2	4	2	0	16	108
Aalenian	21	31	45	0	0	24	35	0	0	12	0	4	2	0	0	3	74
Toarcian	36	57	32	0	2	42	54	0	1	15	0	14	17	2	1	16	110
Pliensbachian	51	48	58	0	0	59	74	16	6	17	3	22	31	6	3	17	137
Sinemurian	19	47	20	13	3	49	39	10	9	30	0	16	25	8	8	23	107
Hettangian	25	11	5	0	2	15	15	0	3	9	0	3	0	0	8	19	62
Norian-Rhaetian	9	6	0	6	22	21	11	13	19	6	28	4	1	0	0	13	68

Table 1. Number of taxa per time slice and area. Data not processed in this study shown in grey.

Table 2. Simpson similarity coefficients between the 16 areas based on the distribution of all bivalve genera / subgenera, for seven time intervals between Norian / Rhaetian to Bajocian. Maximum values for continuous records are shown in bold type.

	S France							Iberian Peninsula							Morocco						Colombia		Peru	N Chile							Argentina + central Chile						
Time slice	N	H	S	P	T	A	B	N	H	S	P	T	A	B	H	S	P	T	A	B	N	S	N	N	H	S	P	T	A	B	N	H	S	P	T	A	B
Iberian Peninsula	33	64	**89**	73	78	57	33																														
Morocco	—	40	53	53	62	76	**83**	—	40	79	62	84	61	**93**																							
Colombia	0	—	46	—	—	—	—	0	—	61	—	—	—	—	—	38	—	—	—	—																	
Peru	22	—	—	—	—	—	—	33	—	—	—	—	—	—	—	—	—	—	—	—	83	—															
N Chile	22	53	**63**	53	50	38	—	0	27	45	60	**64**	54	33	20	55	50	59	58	**78**	66	77	48														
Argentina + central Chile	0	53	58	61	**72**	52	58	0	27	41	**73**	57	55	66	40	40	53	**68**	60	65	33	77	82	82	66	61	86	78	36	89							
Mexico	11	—	40	69	—	—	—	16	—	40	62	—	—	—	—	40	75	—	—	—	50	30	31	31	—	50	81	—	—	—	18	—	70	87	—	—	—
Sonomia+cratonic USA	33	—	22	83	—	—	—	33	—	22	66	—	—	—	—	33	83	—	—	—	50	66	42	42	—	66	**83**	—	—	79	45	—	77	**83**	—	—	70
N American Boreal craton	0	**44**	58	**59**	40	33	37	0	33	43	47	**60**	33	0	40	45	41	47	**75**	**75**	0	61	16	33	**44**	**53**	41	**53**	42	0	16	22	57	65	47	75	**87**
Wallowa terrane	22	—	—	—	—	—	—	0	—	—	—	—	—	—	—	—	—	—	—	—	33	—	45	48	—	—	—	—	—	—	72	—	—	—	—	—	—
Wrangellia terrane	—	0	19	45	57	—	—	—	0	19	36	64	—	—	0	12	27	43	—	—	—	31	—	—	33	56	50	57	—	—	—	0	50	64	64	—	—
Stikinia terrane	—	—	47	61	—	—	—	—	—	56	61	76	—	—	—	40	45	59	—	—	—	69	—	—	—	72	71	76	—	—	—	—	68	87	76	—	—
Quesnellia terrane	—	—	62	50	—	—	—	—	—	50	50	—	—	—	—	25	50	—	—	—	—	25	—	—	—	75	100	—	—	—	—	—	62	83	—	—	—
Cadwallader terrane	—	75	50	—	—	—	—	—	25	50	—	—	—	—	0	37	—	—	—	—	—	50	—	—	62	62	—	—	—	—	—	50	62	—	—	—	—
Japan	11	5	**42**	35	31	—	16	16	0	**56**	29	31	—	27	0	25	29	19	—	**69**	16	46	8	23	0	56	**64**	25	—	33	18	7	39	59	47	—	**62**

	Mexico			Sonomia+cratonic USA				N American Boreal craton						Wallowa	Wrangellia			Stikinia			Quesnellia		Cadwallader	
Time slice	N	S	P	N	S	P	B	N	H	S	P	T	B	N	S	P	T	S	P	T	S	P	H	S
Iberian Peninsula																								
Morocco																								
Colombia																								
Peru																								
N Chile																								
Argentina + central Chile																								
Mexico																								
Sonomia+cratonic USA	23	22	66																					
N American Boreal craton	16	30	12	50	**55**	33	37																	
Wallowa terrane	46	—	—	53	—	—	—	83	—	—	—	—	—											
Wrangellia terrane	—	20	25	—	55	33	—	—	—	56	35	36	—	—										
Stikinia terrane	—	50	43	—	88	66	—	—	—	60	47	47	—	—	69	64	71							
Quesnellia terrane	—	0	50	—	37	33	—	—	—	62	16	—	—	—	62	100	—	62	83	—				
Cadwallader terrane	—	25	—	—	25	—	—	—	12	75	—	—	—	—	50	—	—	75	—	—	37	—		
Japan	15	30	6	23	22	0	**37**	16	0	**39**	23	13	37	31	37	23	14	39	41	12	62	16	12	37

REFERENCES

B	Bajocian
A	Aalenian
T	Toarcian
P	Pliensbachian
S	Sinemurian
H	Hettangian
N	Norian/Rhaetian

North American suspect terranes

(11) Wallowa terrane, (12) Wrangellia terrane, (13) Stikinia terrane, (14) Quesnellia terrane and (15) Cadwallader terrane. References included in Smith et al., 1994; also McLearn, 1924; Smith, 1927; Jeletzky, 1976; Imlay, 1981; Cameron and Tipper, 1985; Newton et al., 1987; Poulton, 1991; McRoberts, 1992; Pálfy et al., 1994; Aberhan and Pálfy, 1996; Aberhan and Muster, 1997; Aberhan, 1998a; 1998b.

West Pacific area

(16) Japan: Hayami, 1975, and references therein.

For each area, data were distributed within seven time slices corresponding to stages, as follows: Norian/Rhaetian, Hettangian, Sinemurian, Pliensbachian, Toarcian, Aalenian and Bajocian. Only records with reliable age assignment to the stage level were included. Data are lacking or very scanty for some areas/time combinations, and those with insufficient data were discarded from the analysis (Table 1). Unfortunately, data are still scarce for some key areas, such as northern South America (Colombia, Peru) and Mexico.

Similarity coefficients were calculated for the seven time intervals considered for every pair of areas having sufficient data (Table 2). The Simpson index (Simpson, 1960) was chosen because it meets six of the nine conditions tested by Shi (1993, see also references therein) when comparing binary similarity coefficients, and thus it is one of those recommended for palaeobiogeographic studies. Relative to other coefficients, it is less sensitive to disparities in sample sizes and sample intensities, and thus reduces the influence of incomplete data on coefficient values (Henderson and Heron, 1977; Newton, 1990), allowing comparison of an heterogeneous data set such as this. For these reasons, it has been widely preferred for large scale paleobiogeographic studies (Hallam, 1983; Liu, 1995).

Results

Table 2 shows the time succession of Simpson similarity coefficients from Norian/Rhaetian to Bajocian times for each possible pair of the 16 areas. In most cases with good continuous record, the maximum similarity occurs in Bajocian times. The similarity maximum occurs during the Sinemurian when certain areas are compared with Japan, when the N American Boreal Craton is compared with the suspect terranes, and for the S France/N Chile relation. On the other hand, for areas on both sides of the Hispanic Corridor, similarity peaks occur either at the Sinemurian, Pliensbachian or Toarcian. These may be only in part influenced by the overall increase in bivalve diversity during the Pliensbachian (see discussion in Aberhan and Fürsich, 1997), as it is also clearly seen when comparing the four major regions in question if only maximum values within each of the four regions are considered (Fig. 1).

The other background pattern which should be taken into account is palaeolatitudinal control of the faunal distributions, which was recognised along the Pacific margins both for the northern (Tozer, 1982) and southern Hemispheres (Damborenea, 1993; 1996) on the basis of mollusc data.

Several further observations can be drawn from the analysis of this set of similarity coefficients, but only a summary of the general patterns observed will be presented here.

Discussion of the relationships among different areas can be grouped as follows (Fig. 2):

a. Within each major region (western Tethys, cratonic North America, South America): as expected, similarity is high through all times considered, and tends to increase slightly with time with only minor fluctuations. The low Aalenian similarity between N Chile and Argentina + central Chile is probably enhanced by uneven knowledge of these faunas (as compared with older and younger ones).

b. Along the eastern Pacific (from North American boreal craton to Argentina + central Chile): again similarity is high (especially for areas located at approximately the same palaeolatitude) and is maintained through time with only local minor fluctuations; see for example the relations between North American Boreal craton and Argentina + N Chile, or Sonomia + cratonic USA and N Chile (Fig. 2). Detailed analysis of these sets of similarity coefficients is especially interesting when compared with those of suspect terranes, but this will not be further discussed here.

c. Across the Pacific: Relationships with Japan show a consistent pattern of low similarity values compared with all other areas analysed here. Regarding changes through time, the observed pattern is also consistent for all regions: there is a decrease in similarity during the Hettangian, followed by a peak during Sinemurian (or Pliensbachian) times, and then relatively low values are maintained with slight fluctuation. It is interesting to note that this pattern is also shown between Japan and the North American Boreal craton, which were located at approximately the same latitude on either side of the Pacific. Nevertheless, it should be pointed out that this is one of the migration routes favoured by Elmi (1993) to explain the common occurrence of ammonites in South America and Tethys during Early and Middle Jurassic times.

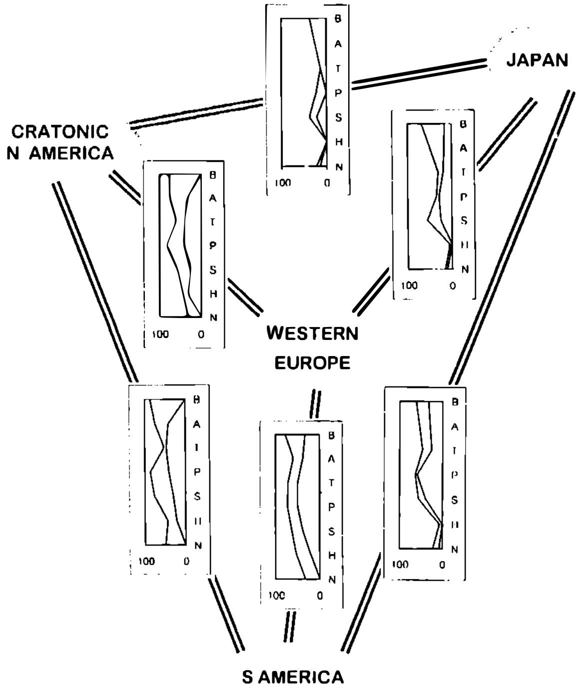


Figure 1. Evolution of the similarity between the four major regions. Maximum and minimum values of Simpson similarity coefficient based on all bivalve genera/subgenera are plotted for each time interval. References for time intervals in Table 2.

d. Across the Hispanic Corridor: There is a sudden increase in similarity values beginning very early in the Early Jurassic (Hettangian-Sinemurian), from low values during Late Triassic time, reaching similarity peaks at about Sinemurian-Pliensbachian time, followed by a decrease by the end of the Early Jurassic and then a new increase from the beginning of the Middle Jurassic to the Bajocian (see examples in Fig. 2). The maximum values observed are similar to those obtained *within* regions and along the eastern Pacific. Maximum similarity peaks across the Corridor fluctuate between the Sinemurian and Toarcian for different regions. Curiously, Norian faunas of western North America are more similar to those from Mediterranean and especially Alpine regions (already pointed out by Muller and Ferguson, 1939) than to those from the westernmost Tethys areas analysed here.

To explain this last pattern, Kristan-Tollmann and Tollmann (1981) envisaged a westerly directed "Tethyan current" for Late Triassic time, and proposed such direction for the migration of the marine fauna. Some bivalve data are indeed in agreement with this hypothesis. For instance, during Triassic time *Palaeolopha* was common in eastern Pacific and Alpine regions and central Europe, but absent from western Europe. This also implied the absence of a direct connection between western Tethys and eastern Pacific at that time.

On the other hand, from Early Jurassic time onwards, comparison of the evolution through time in different areas (Fig. 2) is consistent with the existence of a shallow, intermittent, marine connection between western Tethys and the eastern Pacific. The first marine connection may have been established by about Sinemurian time, and thus preceded rifting by millions of years, but allowed a considerable faunal interchange of benthonic organisms such as bivalves. This connection was certainly well established by Pliensbachian/Toarcian times, as indicated by the high similarity values across it, which equal those recorded between neighbouring localities.

"Off-shore" (sometimes even pseudoplanktonic) bivalve species were analysed separately and, although data are still few and will not be discussed in detail here, the distribution of some of these taxa during the Early Jurassic can be alternately explained using a pantropic model, implying planktotrophic larvae and fast ocean currents. In this case, data are consistent with a western direction of prevailing currents (in agreement with the proposal of Kristan-Tollmann and Tollmann, 1981 for Late Triassic time), since some taxa which are abundant along the eastern Pacific, such as *Posidonotis* and *Otapiria*, do have isolated records in eastern and central (but not western) Tethys. This also suggests that the shallow connection acted as a filter, being an effective barrier for off-shore species, while allowing the passage of on-shore benthonic species.

Dispersal Patterns

Of special interest for this subject are taxa restricted to the western Tethys and the eastern Pacific and their mutual relationships. Several benthonic bivalve taxa, and especially epifaunal stocks, can be followed "step by step" along this migration route, either eastwards or westwards.

Weyla migrated eastwards as discussed by Damborenea and Manceñido (1979; 1988; see also Liu et al., 1998). The *Lithiotis-Opisoma* association (see discussion in Broglio Loriga and Neri, 1976; Hillebrandt, 1981; Nauss and Smith, 1988), followed the same route at approximately the same time but in the opposite direction. Other bivalve taxa which, although not restricted to these regions, probably migrated along this corridor, according to their known distribution with time, are: from Tethys to eastern Pacific: *Palaeolopha*, *Pseudopecten*, *Atreta*, *Terquemia*, *Cardinia*, *Pteromya*, *Goniomya*; from eastern Pacific to Tethys: *Gryphaea*, *Actinostreon*, *Preaexogyra*, *Aguilerella*, *Gervillaria*, *Lycettia*, *Antiquilima* and some trigoniids, such as *Frenguelliella*, *Jaworskiella*, *Vaugonia* and *Psilotrigonia*.

Among ostreaceans, *Gryphaea* was distributed along Arctic and Pacific regions during the Triassic (see McRoberts, 1992) and it reached western Europe already during the Hettangian. But these early faunas are, nevertheless, endemic at specific level. Only from the Sinemurian onwards do Pacific species have strong affinities with western European ones (Rubilar, 1998).

Several *Actinostreon* species known in Europe during the Late Pliensbachian and Toarcian appeared first in South America: *A. costatum* (Sowerby) is known from Late Sinemurian/Early Pliensbachian in Chile, but in western Tethys only appears during the Toarcian; *A. gregareum* (Sowerby) has been mentioned from probably Pliensbachian deposits of Mexico (Jaworski, 1929) but only reached Spain during the Toarcian (de Verneuill and Collomb, 1853, Castell et al., 1955). This genus later spread northwards (Liu et al., 1998).

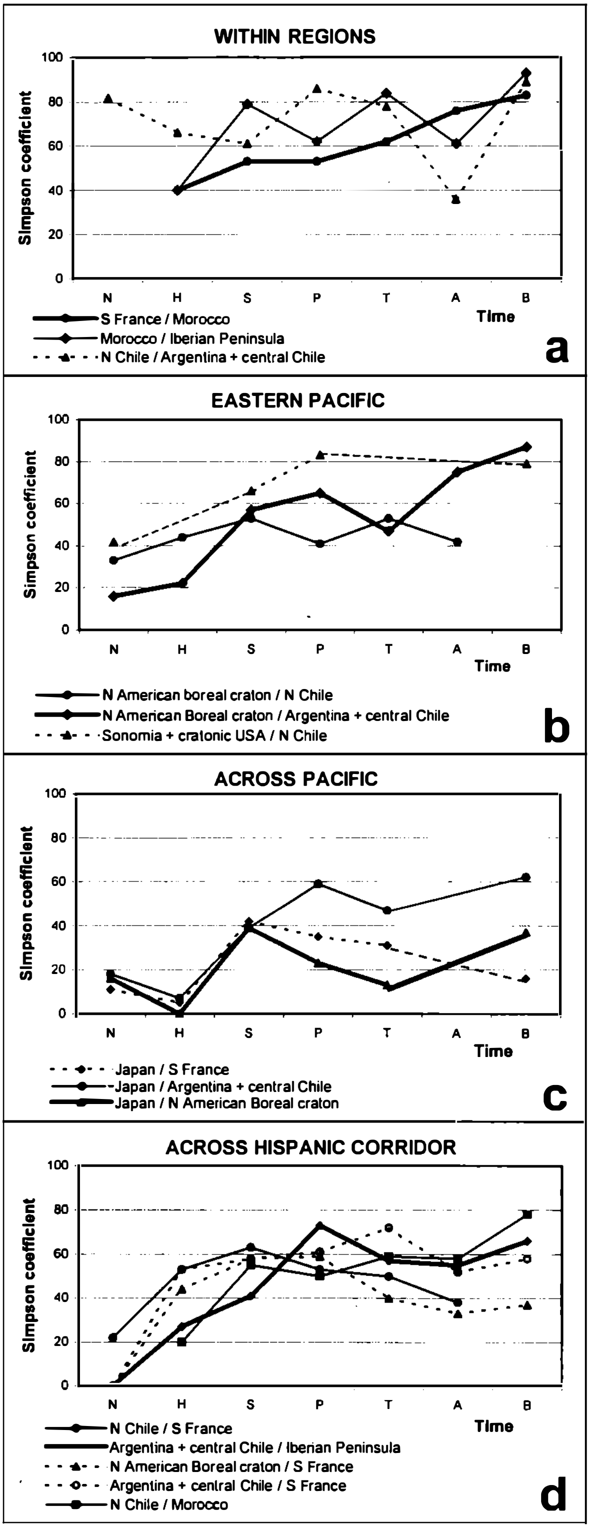
From the author's personal collections and Rubilar (1998), the first representatives of *Praeexogyra* are known in South America (Toarcian) while the genus only becomes common in the Tethys during Middle Jurassic (Bajocian-Bathonian) times.

Conclusions

From analysis of bivalve distribution the following conclusions can be drawn:

1. Within regions faunal similarity is high, and increases with time during the Early and Middle Jurassic. This is also the pattern along the eastern Pacific, especially for areas located at similar palaeolatitudes.
2. Across the Hispanic Corridor: There is a sudden increase in similarity beginning very early during the Early Jurassic (from the Sinemurian at least) from low values during Late Triassic times, followed by similarity peaks at about the Pliensbachian/Toarcian.
3. Japan shows very low values of similarity with all other areas.
4. Most data analysed here are consistent with the existence of a shallow intermittent marine connection between western Tethys and the eastern Pacific. The first marine connection was probably established by the Sinemurian (if not even Hettangian).
5. The shallow connection acted as a filter, being an effective barrier for off-sore species, while allowing the passage of on-shore benthonic species.

Figure 2. Successions of Simpson similarity coefficients for some pairs of areas through time, grouped according to their paleogeographic relationships (discussion in text). References in Table 2.


WITHIN REGIONS
Simpson coefficient
100
80
60
40
20
0
N
H
S
P
T
A
B
Time
S France / Morocco
Morocco / Iberian Peninsula
N Chile / Argentina + central Chile
a
EASTERN PACIFIC
Simpson coefficient
100
80
60
40
20
0
N
H
S
P
T
A
B
Time
N American boreal craton / N Chile
N American Boreal craton / Argentina + central Chile
Sonomia + cratonic USA / N Chile
b
ACROSS PACIFIC
Simpson coefficient
100
80
60
40
20
0
N
H
S
P
T
A
B
Time
Japan / S France
Japan / Argentina + central Chile
Japan / N American Boreal craton
c
ACROSS HISPANIC CORRIDOR
Simpson coefficient
100
80
60
40
20
0
N
H
S
P
T
A
B
Time
N Chile / S France
Argentina + central Chile / Iberian Peninsula
N American Boreal craton / S France
Argentina + central Chile / S France
N Chile / Morocco
d

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References

ABERHAN, M. 1993. Benthic macroinvertebrate associations on a carbonate-clastic ramp in segments of the Early Jurassic back-arc basin of northern Chile (26-29°S). Revista Geológica de Chile, 20(2):105-136. Santiago.

____. 1994. Early Jurassic Bivalvia of northern Chile. Part 1. Subclasses Palaeotaxodonta, Pteriomorphia, and Isofilibranchia. Beringeria, 13:1-115. Würzburg.

____. 1998a. Early Jurassic Bivalvia of western Canada. Part I. Subclasses Palaeotaxodonta, Pteriomorphia, and Isofilibranchia. Beringeria, 21:57-150. Würzburg.

____. 1998b. Paleobiogeographic Patterns of Pectinoid Bivalves and the Early Jurassic Tectonic Evolution of Western Canadian Terranes. Palaios, 13:129-148.

____, AND F. T. FÜRSICH. 1997. Diversity analysis of Lower Jurassic bivalves of the Andean Basin and the Pliensbachian/Toarcian mass extinction. Lethaia, 29:181-195. Oslo.

____, AND A. v. HILLEBRANDT. 1996. Taxonomy, ecology, and palaeobiogeography of *Gervilleioperna (Gervilleiognoma) aurita* n. subgen. n. sp. (Bivalvia) from the Middle Jurassic of northern Chile. Paläontologisches Zeitschrift, 70(1-2):79-96.

____, AND H. MUSTER. 1997. Palaeobiology of Early Jurassic bakevelliid bivalves from western Canada. Palaeontology 40(3):799-815. London.

____, AND J. PÁLFY. 1996. A low oxygen tolerant East Pacific flat clam (*Posidonotis semiplicata*) from the Lower Jurassic of the Canadian Cordillera. Canadian Journal of Earth Sciences, 33:993-1006.

BROGLIO LORIGA, A., AND C. NERI. 1976. Aspetti paleobiologici e paleogeografici della facies a "*Lithiotis*" (Giurese inf.). Rivista Italiana di Paleontologia e Stratigrafia, 82(4):651-705. Milano.

BURCKHARDT, C. 1930. Étude synthétique sur le Mésozoïque mexicain. Première partie. Seconde partie. Mémoires de la Societé Paléontologique Suisse, 49-50:1-280. Bâle.

CALZADA, S. 1981. Sobre tres lamelibranquios del Lías de Camarasa (Lérida) (*Pleuronectites freneixi* n. sp., *Pleuromya rotundata* y *Pleuromya marini*). Cuadernos de Geología, Universidad de Granada, 10:213-218. Granada.

____. 1983. Algunos lamelibranquios liásicos de Alfara (Tarragona). Acta Geologica Hispanica. 17(1-2):121-127.

____. 1986. *Ceratomya bechi* n. sp. del Aaleniense español. Ilerda 46:195-202. Lleida.

CAMERON, B. E. B., AND H. W. TIPPER. 1985. Jurassic stratigraphy of the Queen Charlotte Islands, British Columbia. Geological Survey of Canada, Bulletin 365:1-49. Ottawa.

CASTELL, J., S. DE LA CONCHA, AND J. DE LA REVILLA. 1955. Un yacimiento fosilífero del Lías de Maranchón (Guadalajara). Notas y Comentarios del Instituto Geológico y Minero de España, 40:25-39. Madrid.

COX, L. R. 1949. Moluscos del Triásico superior del Perú (Upper Triassic Mollusca from Perú). Boletín del Instituto Geológico del Perú, 12:1-50. Lima.

____. 1956. Jurassic Mollusca from Peru. Journal of Paleontology, 30(5):1179-1186, pl. 127-128.

DAMBORENEA, S. E. 1993. Early Jurassic South American pectinaceans and circum-Pacific palaeobiogeography. Palaeogeography, Palaeoclimatology, Palaeoecology, 100(1-2):109-123. Amsterdam.

____. 1996. Palaeobiogeography of Early Jurassic bivalves along the southeastern Pacific margin. 13° Congreso Geológico Argentino y 3° Congreso de Exploración de Hidrocarburos, Actas 5:151-167. Buenos Aires. [Includes references]

____, AND C. GONZÁLEZ LEÓN, in press. Late Triassic and Early Jurassic bivalves from Sonora, Mexico. Revista Mexicana de Ciencias Geológicas, 14. México.

____, AND M. O. MANCEÑIDO. 1979. On the palaeogeographical distribution of the pectinid genus *Weyla* (Bivalvia, Lower Jurassic). Palaeogeography, Palaeoclimatology, Palaeoecology, 27(1/2):85-102. Amsterdam.

____, AND ____. 1988. *Weyla*: semblanza de un bivalvo Jurásico andino. Actas 5° Congreso Geológico Chileno, 2:C13-C25. Santiago de Chile.

DARESTE DE LA CHAVANNE, J. 1920. Fossiles liasiques de la région de Guelma. Matériaux pour la Carte Géologique de l'Algerie, 1 ser., Paléontologie 5:1-73. Alger.

DUBAR, G. 1925. Études sur le Lias des Pyrénées Françaises. Mémoires de la Societé Géologique du Nord, 9(1):1-332, pl. 1-7. Lille.

____. 1931. Lias et Jurassique du Bastan (Haute Vallée de la Bidassoa, Espagne). Bulletin de la Societé Géologique de France, ser. 4, 30:589-608, pl. 40-41. Paris.

DUMORTIER, E. 1864-1874. Études paléontologiques sur les dépots Jurassiques du Bassin du Rhone. 1 (1864), 187 pp.; 2 (1867), 252 pp; 3 (1869), 348 pp; 4 (1874), 335 pp. F. Savy (Ed.), Paris.

ELMI, S. 1993. Les voies d'échange faunique entre l'Amérique du Sud et la Téthys alpine pendant le Jurassique inférieur et moyen. Documents des Laboratoires de Géologie, Lyon, 125:139-149. Lyon.

GEYER, O. F. 1973. Das präkretazische Mesozoikum von Kolumbien. Geologisches Jahrbuch B(5):1-155. Hannover.

HALLAM, A. 1971. Facies analysis of the Lias in West Central Portugal. Neues Jahrbuch für Geologie und Paläontologie, Abh., 139(2):226-265. Stuttgart.

____. 1976. Stratigraphic distribution and ecology of European Jurassic bivalves. Lethaia, 9:245-259. Oslo.

____. 1977. Jurassic bivalve biogeography. Paleobiology, 3(1):58-73.

____. 1983. Early and Mid-Jurassic molluscan biogeography and the establishment of the central Atlantic seaway. Palaeogeography, Palaeoclimatology, Palaeoecology, 43:181-193. Amsterdam.

HAYAMI, I. 1975. A systematic survey of the Mesozoic Bivalvia from Japan. Bulletin of the University Museum, University of Tokyo, 10:1-249, pl. 1-10. Tokyo. [Includes references]

____, S. MAEDA, AND C. RUIZ FULLER. 1977. Some Late Triassic Bivalvia and Gastropoda from the Domeyko range of north Chile. Transactions and Proceedings of the Palaeontological Society of Japan, N.S., 108:202-221. Tokyo.

HENDERSON, R. A., AND M. L. HERON. 1977. A probabilistic method of paleobiogeographic analysis. Lethaia, 10(1):1-15. Oslo.

HILLEBRANDT, A. v. 1981. Kontinentalverschiebung und die paläozoogeographischen Beziehungen des südamerikanischen Lias. Geologisches Rundschau, 70(2):570-582. Stuttgart.

IMLAY, R. W. 1964. Marine Jurassic Pelecypods from Central and Southern Utah. United States Geological Survey, Professional Paper, 483-c:1-42. Pl. 1-4. Washington.

____. 1967. Twin Creek Limestone (Jurassic) in the Western Interior of the United States. United States Geological Survey, Professional Paper, 540:1-105. Washington.

____. 1981. Early Jurassic Ammonites from Alaska. United States Geological Survey, Professional Paper 1148:1-49. Washington.

JAWORSKI, E. 1929. Eine Lias-Fauna aus Nordwest-Mexico. Abhandlungen der Schweizerischen Palaeontologischen Gesellschaft, 48(4):1-12. Basel.

JELETZKY, J. A. 1976. Mesozoic and ?Tertiary rocks of Quatsino Sound, Vancouver Island, British Columbia. Geological Survey of Canada, Bulletin 242:1-243. Ottawa.

JOHNSON, A. L. A. 1984. The palaeobiology of the bivalve families Pectinidae and Propeamussiidae in the Jurassic of Europe. Zitteliana, 11:1-235. Taf. 1-11. München.

KÖRNER, K. 1937. Marine (Cassianer-Raibler) Trias am Nevado de Acrotambo (Nord-Peru). Palaeontographica A86(5-6):145-237, pl. 10-14. Stuttgart.

KRISTAN-TOLLMANN, E., AND A. TOLLMANN. 1981. Die Stellung der Tethys in der Trias und die Herkunft ihrer Fauna. Mitteilungen der österreichischen Gesellschaft, 74-75:129-135. Wien.

LAWS, R. A. 1982. Late Triassic depositional environments and molluscan associations from west-central Nevada. Palaeogeography, Palaeoclimatology, Palaeoecology, 37(2-4):131-148.

LEANZA, H. A. 1993. Jurassic and Cretaceous Trigoniid Bivalves from West-Central Argentina. Bulletins of American Paleontology, 105(343):1-95. Ithaca.

LIU CHUNLIAN. 1995. Jurassic bivalve palaeobiogeography of the Proto-Atlantic and the application of multivariate analysis methods in palaeobiogeography. Beringeria 16:3-123. Würzburg. [Includes references]

____, M. HEINZE, AND F. T. FÜRSICH. 1998. Bivalve provinces in the Proto-Atlantic and along the southern margin of the Tethys in the Jurassic. Palaeogeography, Palaeoclimatology, Palaeocology 137:127-151. Amsterdam.

MCLEARN, F. H. 1924. New pelecypods from the Fernie Formation of the Alberta Jurassic. Proceedings and Transactions of the Royal Society of Canada, 18(4):39-61. Ottawa.

MCROBERTS, C. A. 1992. Systematics and paleobiogeography of Late Triassic *Gryphaea* (Bivalvia) from the North America Cordillera. Journal of Paleontology 66(1):28-39. Lawrence.

____. (in press). Late Triassic (Norian-Rhaetian) bivalves from the Antimonio Formation, northwest Sonora, Mexico. Revista Mexicana de Ciencias Geológicas, 14. México.

MOUTERDE, R., J. L. DOMMERGUES AND R.B. ROCHA. 1983. Atlas des fossiles caractéristiques du Lias Portugais. I. Lias inférieur. Ciências da Terra (UNL), 6:49-76. Lisboa.

____, R. B. ROCHA, AND J. DELANCE. 1981. Atlas des fossiles caractéristiques du Lias Portugais. II. Carixien. Ciências da Terra (UNL), 7:187-254. Lisboa.

MULLER, S., AND H. FERGUSON. 1939. Mesozoic Stratigraphy of the Hawthorne and Tonopah Quadrangles, Nevada. Bulletin of the Geological Society of America, 50(10):1573-1624.

MUSTER, H. 1995. Taxonomie und Paläobiogeographie der Bakevelliidae (Bivalvia). Beringeria 14:1-161 (Taf. 1-17). Würzburg.

NAUSS, A. L., AND P. L. SMITH. 1988. *Lithiotis* (Bivalvia) bioherms in the Lower Jurassic of east-central Oregon, U.S.A. Palaeogeography, Palaeoclimatology, Palaeoecology, 65(3-4):253-268. Amsterdam.

NEWTON, C. R. 1988. Significance of "Tethyan" Fossils in the American Cordillera. Science 242:385-391.

____. 1990. Palaeobiogeography. p. 452-460. *In* D. E. Briggs and P. R. Crowther (eds.), Paleobiology – a Synthesis. Blackwell Scientific Publications. Oxford.

____, M. T. WHALEN, J. B. THOMPSON, N. PRINS, AND D. DELALLA. 1987. Systematics and Paleoecology of Norian (Late Triassic) Bivalves from a Tropical Island Arc: Wallowa Terrane, Oregon. Paleontological Society Memoir, 22 (Journal of Paleontology 61,4, suppl.):1-83.

PÁLFY, J., P. L. SMITH, AND H. W. TIPPER. 1994. Sinemurian (Lower Jurassic) ammonoid biostratigraphy of the Queen Charlotte Islands, western Canada. Géobios, Mémoire Spécial, 17:385-393. Lyon.

PÉREZ, E., AND R. REYES. 1994. Presencia de *Maoritrigonia* Fleming, 1962 (Bivalvia; Minetrigoniidae) en el Triásico Superior de Sudamérica: descripción de dos especies nuevas. Revista Geológica de Chile, 21(1):105-117. Santiago.

POULTON, T. P. 1991. Hettangian through Aalenian (Jurassic) guide fossils and biostratigraphy, northern Yukon and Adjacent northwest Territories. Geological Survey of Canada Bulletin 410:1-95. Ottawa.

RICCARDI, A. C. 1991. Jurassic and Cretaceous marine connections between the Southeast Pacific and Tethys. Palaeogeography, Palaeoclimatology, Palaeoecology, 87:155-189. Amsterdam.

____, S. E. DAMBORENEA, M. O. MANCEÑIDO, R. SCASSO, S. LANÉS, AND M. P. IGLESIA LLANOS. 1997. Primer registro de Triásico marino fosilífero de la Argentina. Revista de la Asociación Geológica Argentina, 52(2):228-234. Buenos Aires.

ROMERO P., L., M. ALDANA A., C. RANGEL Z., E. VILLAVICENCIO R., AND J. RAMÍREZ A. 1995. Fauna y flora fósil del Perú. Boletín del Instituto Geológico Minero y Metalúrgico, Serie D: Estudios Especiales, 17:1-332. Lima.

RUBILAR, R. A. 1998. La Superfamilia Ostreacea en Chile y su importancia cronoestratigráfica, paleogeográfica y paleoecológica (Triásico superior-Jurásico). Tesis Doctoral, Facultad de Ciencias Naturales y Museo, Universidad Nacional de La Plata, 363 pp. La Plata. (Unpublished).

SHI, G. R. 1993. Multivariate data analysis in palaeoecology and palaeobiogeography – a review. Palaeogeography, Palaeoclimatology, Palaeoecology, 105(3-4):199-234. Amsterdam.

SILBERLING, N. J. 1959. Pre-Tertiary stratigraphy and Upper Triassic paleontology of the Union District, Shoshone Mountains, Nevada. Geological Society of America Bulletin, 50:1625-1722. Washington.

SIMPSON, G. G. 1960. Notes on the measurement of faunal resemblance. American Journal of Science, 258a:300-311. New Haven.

SMITH, J. P. 1927. Upper Triassic marine invertebrate faunas of North America. United States Geological Survey, Professional Paper, 141:1-135. Washington.

SMITH, P.L. 1983. The ammonite *Dayiceras dayiceroides* and Early Jurassic paleogeography. Canadian Journal of Earth Sciences, 20:86-91.

____. 1989. Paleobiogeography and plate tectonics. Geoscience Canada, 15(4):261-279.

____, J. M. BEYERS, E. S. CARTER, G. K. JAKOBS, J. PÁLFY, E. PESSAGNO, AND H. W. TIPPER. 1994. 5. North America. 5.1. Lower Jurassic. *In* G.E.G. Westermann, and A.C.

Riccardi (eds.) Jurassic taxa ranges and correlation charts for the circum-Pacific. Newsletters on Stratigraphy, 31(1):33-70. Stuttgart. [Includes references]

____, AND H. W. TIPPER. 1986. Plate Tectonics and Paleobiogeography: Early Jurassic (Pliensbachian) Endemism and Diversity. Palaios 1:399-412.

____, G. E. G. WESTERMANN, G. D. STANLEY, AND T. E. YANCEY. 1990. Paleobiogeography of the Ancient Pacific. Science 249:680-683.

STANLEY, G. D. jr., C. GONZÁLEZ-LEÓN, M. R. SANDY, B. SENOWBARI-DARYAN, P. DOYLE, M. TAMURA, AND D. H. ERWIN. 1994. Upper Triassic Invertebrates from the Antimonio Formation, Sonora, Mexico. The Paleontological Society Memoir 36 (Journal of Paleontology 68, 4, suppl.):1-33.

TOZER, E. T. 1982. Marine Triassic faunas of North America: their significance for assessing plate and terrane movements. Geologische Rundschau, 71:1077-1104.

VERNEUIL, E. de, AND E. COLLOMB. 1853. Coup d'oeil sur la constitution géologique de quelques provinces de l'Espagne. Bulletin de la Societé Géologique de France, 2ème. sér., 10(1):61-147, 162-166. Pl. 1-3. Paris.

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Stratigraphic Distribution of Toarcian Brachiopods from the Iberian Range (Spain) and its Relation to Depositional Sequences

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Abstract: The stratigraphic distribution of the Toarcian brachiopods of the Iberian Range is summarized. Several episodes of faunal turnover are distinguished permitting the recognition of 7 successive assemblages. A comparison with the sequence stratigraphy proposed for the basin shows that episodes of faunal change coincide with changes in the trend of the deepening-shallowing sequences. In particular, the most conspicuous changes, and therefore the more useful ones for correlations, take place after the transgressive maxima (uppermost part of the Tenuicostatum Zone, middle part of the Bifrons Zone and upper part of the Insigne Zone).

Introduction

Brachiopods are abundant in the Toarcian of the Iberian Range and have been the subject of specific studies since the middle of last century (Deslongchamps, 1863; Dubar, 1931; Castell et al., 1955). During the last 25 years, the Toarcian of the Iberian Range has been the subject of numerous studies in which the brachiopods have been collected in a systematical way together with ammonites (Goy, 1974; Delance, 1974; Goy and Robles, 1975; Comas-Rengifo and Goy, 1975; García Joral, 1983; 1986; 1993; García Joral and Goy, 1984; 1994; Goy et al., 1984; García Joral et al., 1990; 1997). The aim of this paper is to present an update of the distribution of the different taxa, considering a recent systematic review (García Joral and Goy, in prep.), together with new field data. The distribution data have been related to the sequence stratigraphy recently proposed for the Toarcian of the basin.

Brachiopod Assemblages Recorded in the Toarcian of the Iberian Basin

Study of the stratigraphic distribution of the Toarcian brachiopods in numerous localities of the Iberian Range (Figures 1 and 2) permits us to distinguish 7 assemblages:

1) The first assemblage is present in the Tenuicostatum Zone. It includes taxa that in many cases, particularly at a generic level, range up from the Pliensbachian (*Spiriferina, Liospiriferina, Gibbirhynchia, Quadratirhynchia, Lobothyris, Zeilleria, Aulacothyris*). At a more detailed level, two subsets can be distinguished (García Joral et al., 1997): the first consists of typically Domerian species of wide geographic distribution (*Liospiriferina alpina, Lobothyris subpunctata, L. edwardsi, Aulacothyris resupinata*), while the second includes species close to the previous ones, but of a more endemic nature (*Liospiriferina falloti, L. undulata, Quadratirhynchia attenuata, Gibbirhynchia* nov. sp., *Lobothyris arcta, Aulacothyris iberica*). All these species disappear approximately at the boundary between the Tenuicostatum and Serpentinus Zones. This extinction is followed by a short episode in which no brachiopods are recorded.
2) The second assemblage is composed of generally monospecific associations of *Soaresirhynchia bouchardi* that appear throughout the basin in the lower part of the Serpentinus Zone.
3) The third assemblage extends through the Serpentinus Zone and into the lower part of the Bifrons Zone. It includes taxa which are characteristic of the Basin or of neighboring basins, the so-called Spanish Bioprovince of brachiopods (García Joral and Goy, 1984; Goy et al., 1984), that can also be recognized in Catalonia, Southeastern France, Portugal and the Middle Atlas of

Morocco and Algeria. The associations are dominated by species of the genera *Homoeorhynchia* and *Telothyris* (*H. batalleri*, *H. meridionalis*, *T. jauberti*, *T. pyrenaica*). Other species present in this assemblage are *Pseudogibbirhynchia jurensis*, *Tetrarhynchia? paucicostatae*, Nov. gen. *vasconcellosi*, *Lobothyris hispanica* and *Sphaeroidothyris perfida*.

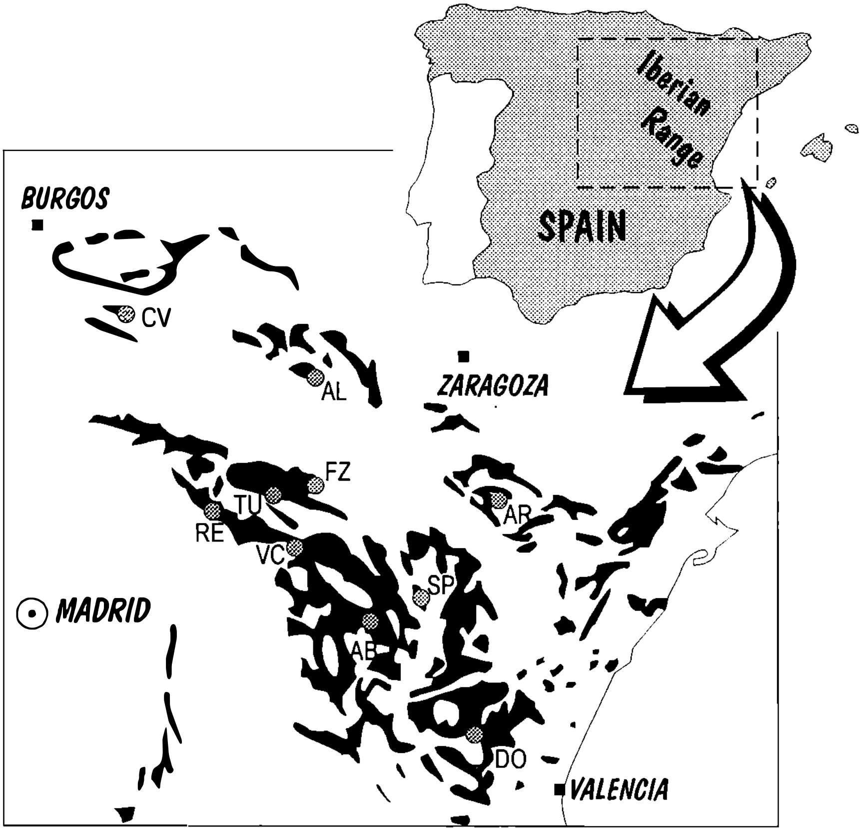

Figure 1. Representative brachiopod localities in the Iberian Range. CV=Castrovido; AL=La Almunia; FZ=Fuentelsaz; TU=Turmiel; RE=Renales; VC=Villar de Cobeta; AR=Ariño; SP=Sierra Palomera; AB=Albarracín; DO=Domeño. Jurassic outcrops in black.

4) The fourth assemblage begins at the middle part of the Bifrons Zone. The majority of taxa included here are endemic to the Basin, such as *Homoeorhynchia* nov. sp., *Praemonticlarella* nov. sp. and Nov. gen. nov. sp. (= *Pseudogibbirhynchia?* nov. sp. *in* García Joral and Goy, 1984). Other taxa are also distributed in neighboring areas, such as *Sphaeroidothyris vari* and *Zeilleria (Z.) lycetti*. Except for *S. vari*, the species of this assemblage are only found at certain localities or particular areas and are, as a rule, scarce and of small size.
5) The fifth assemblage begins at the middle part of the Variabilis Zone and is characterized by a decrease in the diversity. It frequently includes a single species, either *Stroudithyris infraoolithica* or *Homoeorhynchia? marini*, depending on the degree of carbonate content of the enclosing beds. These species are sometimes found associated with *Praemonticlarella* aff. *schuleri*, as well as the last representatives of Nov. gen. nov. sp. or of *Zeilleria (Z.) lycetti*.
6) The sixth assemblage begins with the appearance of *Aulacothyris blakei* associated with *Stroudithyris stephanoides* and extends through the Pseudoradiosa Zone and most of the Aalensis Zone. It includes genera that persist into the Aalenian, such as *Rhynchonelloidea*, *Stroudithyris*, *Loboidothyris* or *Aulacothyris*. The associations are of low diversity (*Rhynchonelloidea goyi*, *S.*

stephanoides and *A. blakei* in the Pseudoradiosa Zone; *R. goyi, S. stephanoides* and *Loboidothyris hypocistis* in the Aalensis Zone), although they can be abundant.

7) The last assemblage includes species normally recorded in the Opalinum Zone of the Aalenian but which, in certain localities such as Fuentelsaz, appear in the upper part of the Aalensis Zone (*Homoeorhynchia cynocephala*, *Praemonticlarella distercica*, *Prionorhynchia rubrisaxensis*, *Stroudithyris pisolithica*, *Zeilleria (Cincta) anglica*).

Relation Between Brachiopod Assemblages and Depositional Sequences

In terms of the sequential organization, two major transgression-regression sequences are recognized in the Toarcian of the Iberian Basin (Goy et al., 1996; Comas-Rengifo et al., 1996; Gómez and Goy, 1997; 1998). The first extends from the base of the Tenuicostatum Zone to the top of the Variabilis Zone and the second from the base of the Thouarsense Zone to the top of the Aalensis Zone. In the first of these megasequences two minor transgression-regression sequences can be distinguished, the boundary between the two being located in the lower part of the Serpentinus Zone (Fig. 2). The distribution of the brachiopod species, the renewals and relative diversity can be related to these sedimentary sequences.

The most important episodes of renewal (those that separate the assemblages described in the previous section) coincide with changes within the sequences. The first of these renewal episodes takes place at the boundary between the Tenuicostatum and Serpentinus Zones coinciding with a maximum depth followed by a sharp and brief fall in sea level. This episode has been related to a notable impoverishment of oxygen at the sea bottom in wide areas of Western Europe (Jenkyns, 1988; Arias et al., 1992; Jiménez et al., 1996; García Joral et al, 1997; Goy et al., 1997) that affects groups of both benthic and nektonic organisms and, in our opinion, provokes the mass extinction of the brachiopods in the Iberian Basin. No brachiopod species survives this event, which is one of the most important in the history of the Mesozoic brachiopods. After this, recovery takes place, beginning with very large populations of *Soaresirhynchia bouchardi*, a species of unspecialized morphology and small size, forming large populations with high level of polymorphism; a set of characters that indicate opportunism or even a pioneer nature.

Another important episode of faunal change takes place in the middle part of the Bifrons Zone and coincides also with a relative depth maximum followed by a progressive fall of the sea level. In several localities, especially those in the north of the Basin, this event leads to the disappearance of brachiopods, that do not reappear until two or three zones later. At other localities, a turnover of species within the same genera is observed (Nov. gen., *Homoeorhynchia*, *Telothyris*, *Sphaeroidothyris*), a change that includes among the rhynchonellids a more or less important reduction in the size of the specimens.

A further episode of faunal change associated with a transgressive maximum takes place between the Insigne and Pseudoradiosa Zones. The changes involved are less evident and the episode is more difficult to observe due to the low diversity of the brachiopods in this interval. *Aulacothyris blakei* appears during this episode, forming large populations preserved in marly facies. Among the rhynchonellids, the sharp substitution of *Homoeorhynchia? marini* by *Rhynchonelloidea goyi* can be observed at different localities whereas, among the terebratulids the substitution of *Stroudithyris infraoolithica* by *S. stephanoides* is gradual within the Insigne Zone.

On the other hand, the relative depth minima (at the upper part of the Variabilis Zone and at the upper part of the Aalensis Zones) can also be related to changes in the brachiopod assemblages. In the first case, the brachiopods are scarce and the changes are less evident and more dependent of local conditions. However, a substitution of rhynchonellid species within the same genus can be observed (*Homoeorhynchia*, *Praemonticlarella*), as well as the gradual change from *Telothyris* to *Stroudithyris* within the terebratulids. Other species can be found both before and after this episode, such as Nov. gen. nov. sp. or *Zeilleria (Z.) lycetti*. In the second sea level minimum, the diversity increases gradually within the Aalensis Zone. At localities in which the rocks are of a more marly facies, *Homoeorhynchia cynocephala, Praemonticlarella distercica* and *Stroudithyris pisolithica* appear in the Aalensis Subzone. The most significant change takes place in the Buckmani Subzone, with the appearance of *Prionorhynchia rubrisaxensis* and *Z. (Cincta) anglica*, two species that are basically distributed in the Opalinum Zone.

As a rule, at the localities in which sedimentation does not show important interruptions or evidence of remobilization (taphonomic mixing), such as at Ariño, Sierra Palomera, or Ricla, it is observed that diversity maxima are related to the beginning of transgressive episodes.

Figure 2. Stratigraphical distribution of Toarcian Brachiopods from the Iberian Range in relation to the Standard Zones, Subzones and Biohorizons proposed for the Basin by Goy et al. (1988). Zonal name abbreviations as follows: S= Spinatum; TEN= Tenuicostatum; SER=Serpentinus; BIF=Bifrons; VAR=Variabilis; THO=Thouarsense; INS=Insigne; PSE=Pseudoradiosa; AAL=Aalensis; O=Opalinum. Subzonal names abbreviations as follows: H=Hawskerense; MI=Mirabile; SE=Semicelatum; ST=Strangewaysi; FA=Falcifer; SU=Sublevisoni; BI=Bifrons; S=Semipolitum;. V=Variabilis; IL=Illustris; VI=Vitiosa; B=Bingmanni; TH=Thouarsense; F=Fascigerum; f=Fallaciosum; IN=Insigne; LE=Levesquei; P=Pseudoradiosa; MA=Mactra; AA=Aalensis; b=Buckmani; O=Opalinum. The subdivisions of the subzones correspond to the 37 biohorizons proposed in Goy et al. (1988).

Conclusions

The stratigraphic distribution of Toarcian brachiopods of the Iberian Range shows changes in the composition of the associations that permit us to distinguish 7 successive assemblages. These changes coincide with changes in the trend of the deepening - shallowing sequences. After the transgressive maxima (uppermost Tenuicostatum Zone, middle part of the Bifrons Zone, upper part of the Insigne Zone) the most evident changes take place, involving especially the rhynchonellids. The most remarkable change takes place at the boundary between the Tenuicostatum and Serpentinus Zones where a mass extinction is observed, even involving a whole order (Spiriferinida), followed by a total turnover of the brachiopod fauna.

During depth minima (upper part of the Variabilis Zone and upper part of the Aalensis Zone), the changes are less obvious, with gradual transitions of some species into others, especially among the terebratulids. In addition, these episodes are more conditioned by local tectonic features so that in certain areas of the basin where relative deepening occurs, the renewal takes place earlier than in the shallower areas.

Concerning the application of the brachiopods in biostratigraphy, the zone boundaries based on renewals related to transgressive maxima are clear and permit good correlations, and thus can be useful for establishing biozonations of local applicability. Garcia et al. (1996) and Garcia and Dromart (1997) consider the times of maximum flooding as those which permit the best biostratigraphic correlations based on brachiopods, although from a marker-beds view instead of a zonal view. In the Toarcian of the Iberian Basin the assemblages that correspond to times of maximum relative depth, which might correspond to these marker-beds, are in fact those that include a higher proportion of species of wide geographical distribution, and those that permit, therefore, wider correlations. Thus, the study of the distribution of the brachiopods in relation to the depositional sequences enhances both approaches, which are not necessarily incompatible.

References

ARIAS, C. F., M. J. COMAS-RENGIFO, A. GOY, C. HERRERO AND C. H. RUGET. 1992. Variations dans les associations de brachiopodes, foraminifères et ostracodes du Toarcien basal dans le Secteur Central de le Cordillère Ibérique: Un exemple dans La Rambla del Salto (Teruel, Espagne). Cahiers Université Catholique Lyon, Série Sciences, 5:5-25.

CASTELL, J., S. DE LA CONCHA AND J. DE LA REVILLA. 1955. Un yacimiento fosilífero del Lías de Maranchón (Guadalajara). Notas y Comunicaciones del Instituto Geológico y Minero de España 40, 23-40

COMAS-RENGIFO, M. J., J. J. GÓMEZ, A. GOY, C. ARIAS, J. BERNAD, F. GARCÍA JORAL, C. HERRERO, G. MARTÍNEZ AND N. PERILLI. 1996. The Toarcian in the Rambla del Salto (Sierra Palomera) Section. 1st Toarcian and 4th Aalenian Working Groups Meeting. Fieldtrip Iberian Range Guide-Book (Department of Paleontology. Universidad Complutense de Madrid), 27-48.

____. AND A. GOY. 1975. Estratigrafía y Paleontología del Jurásico de Ribarredonda. Estudios geológicos 31:297-339

DELANCE, J. H. 1974. Zeilleridés du Lias d'Europe Occidentale. Mémoires Géologiques de l'Université de Dijon 2, 1-408

DESLONGCHAMPS, E. E. 1863. Etudes critiques sur des Brachiopodes nouveaux ou peu connus. Art. 7: Brachiopodes recueillis par M. De Verneuil dans le Lias dEspagne. Bulletin de la Societé Linneane de Normandie, 3:64-76

DUBAR, G. 1931. Brachiopodes liasiques de Catalogne et des régions voisines. Butlletí de la Institució Catalana de Historia Natural, 31:103-180.

GARCIA, J.-P. AND G. DROMART. 1997. The validity of two biostratigraphic approaches in sequence stratigraphic correlations: brachiopod zones and marker-beds in the Jurassic. Sedimentary Geology, 114:55-79

____, B. LAURIN AND G. SAMBET. 1996. Les associations de brachiopodes du Jurassique moyen du bassin de Paris: une échelle biochronologique ponctuée de niveaux-repères pour la contrainte des corrélations séquentielles à haute résolution. Bulletin de la Société géologique de France, 167(3):435-451

GARCÍA JORAL, F. 1983. Caracterización de *Rhynchonelloidea goyi* n. sp. (Brachiopoda, Rhynchonellidae) del Toarciense Superior en la Cordillera Ibérica (NE de España). Boletín de la Real Sociedad Española de Historia Natural (Sección geológica) 80:84-103

____. 1986. Los braquiópodos del Aaleniense de la Cordillera Ibérica. Unpublished doctoral Thesis. Universidad Complutense de Madrid. 326 pp.

____. 1993. Descripción de *Praemonticlarella distercica* nov. gen., nov. sp. (Brachiopoda, Rhynchonellida) del Toarciense superior y el Aaleniense inferior de la Cordillera Ibérica (NE de España). Boletín de la Real Sociedad Española de Historia Natural (Sección geológica) 88, 73-83.

____. AND A. GOY. 1984. Características de la fauna de braquiópodos del Toarciense Superior en el Sector Central de la Cordillera Ibérica (Noreste de España). Estudios geológicos 40:55-59

____. AND A. GOY. 1994. The Associations of Brachiopods from the Toarcian-Aalenian Transition in the Fuentelsaz Section (Iberian Range, Spain). Géobios M.S. 17:223-228

____. AND A. GOY. (in prep.). Los braquiópodos toarcienses de la Cordillera Ibérica. Revista española de Paleontología.

____, A. GOY AND M. DE RENZI. 1997. Las asociaciones de braquiópodos del Toarciense inferior en los bordes este y norte de la Meseta Ibérica IV Congreso de Jurásico de España. Comunicaciones, p. 77-78

____, A. GOY AND S. URETA. 1990. Las sucesiones de braquiópodos en el tránsito Lías-Dogger en la Cordillera Ibérica. Cuadernos de Geología Ibérica 14:55-65

GÓMEZ, J. J. AND A. GOY. 1997. El Jurásico de la Cordillera Ibérica: Estratigrafía Secuencial y Paleogeografía. IV Congreso de Jurásico de España. Comunicaciones, 15-16

____. AND A. GOY. 1998. Sequential Analysis of the Toarcian in the Northern and Central-Eastern Part of the Iberian Subplate (Spain). 5th International Symposium on the Jurassic System. Abstract volume, 32-33

GOY, A. 1974. El Lías de la mitad Norte de la Rama Castellana de la Cordillera Ibérica. Unpublished doctoral Thesis, Universidad Complutense de Madrid, 940 pp. Abstract: Publicaciones del Departamento de Paleontología, Universidad Complutense de Madrid 14, 35 pp.

____, M. J. COMAS-RENGIFO AND F. GARCÍA JORAL. 1984. The Liassic Brachiopods of the Iberian Range (Spain): Stratigraphic Distribution and Biozonation, p. 227-250. *In* O. Michelsen and A. Zeiss, (eds.), International Symposium on Jurassic Stratigraphy. Symposium volume I.

____, M. J. COMAS-RENGIFO, C. F. ARIAS, F. GARCÍA JORAL, J. J. GÓMEZ, C. HERRERO, G. MARTÍNEZ AND A. RODRIGO. 1997. El tránsito Pliensbachiense/Toarciense en el Sector Central de la Rama Aragonesa de la Cordillera ibérica (España). Cahiers de l'Université Catholique de Lyon, Série Science, 10:159-179.

____, A. P. JIMÉNEZ, G. MARTÍNEZ AND P. RIVAS. 1988. Difficulties in correlating the Toarcian ammonite succession of the Iberian and betic Cordilleras. 2nd Symposium on Jurassic Stratigraphy, 20:155-178

____, G. MARTÍNEZ, C. F. ARIAS, J. BERNARD, F. GARCÍA JORAL, J. J. GÓMEZ, C. HERRERO, N. PERILLI AND S. URETA. 1996. The Toarcian in the sector located between La Almunia de Doña Godina and Ricla. 1st Toarcian and 4th Aalenian Working Groups Meeting. Fieldtrip Iberian Range Guide-Book (Department of Paleontology. Universidad Complutense de Madrid), 3-23.

____. AND F. ROBLES. 1975. Resultado de un estudio sobre Braquiópodos liásicos del NE de Guadalajara. Boletín de la Real Sociedad Española de Historia Natural, volumen extraordinario I centenario II, 243-255

JENKYNS, H. C. 1988. The Early Toarcian (Jurassic) anoxic event: stratigraphy, sedimentary, and geochemical evidence. American Journal of Science, 288:101- 151.

JIMÉNEZ, A. P., C. JIMÉNEZ DE CISNEROS, P. RIVAS AND J. A. VERA. 1996. The Early Toarcian Anoxic Event in the Westernmost Tethys (Subbetic): Paleogeographic and Paleobiogeographic Significance. Journal of Geology 104:399-416.

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A Systematic Summary of the Stratigraphic Distribution of Jurassic Rhynchonellide Genera (Brachiopoda)

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Abstract: The number of described Jurassic rhynchonellide genera has nearly doubled over the last three decades. It is thus appropriate to update the present state of knowledge in light of the classificatory framework being adopted for the forthcoming revised version of the Treatise on Invertebrate Paleontology (Part H).

On the basis of the world-wide data-base assembled during such research the known stratigraphical ranges of about 150 valid genera (or subgenera) belonging to 14 families are critically reviewed and their potential usefulness for biostratigraphic application is outlined. Examples of significant genera being increasingly used to develop zonations and correlations of local to regional value are also discussed.

Introduction

It has long been known that brachiopods, like other benthic organisms, while often providing a useful basis for correlation and biostratigraphic zonation, usually attain less resolution than ammonoid biozonations. Nevertheless, brachiopods may be suitable as supplementary guide fossils, and may even become practical tools in facies where cephalopods are extremely scarce or altogether absent. Following seminal work by Ager in the sixties and seventies, this applied interest has led to the development of biostratigraphic schemes based on Mesozoic brachiopods in various regions of the globe. For the Jurassic, illustrative examples of varying scope are available: for a world-wide overview see Ager (1979), updated and with emphasis on the complete British sequences (Ager, 1994); for France (Alméras and Moulan, 1979; Alméras et al., 1991; 1994; Delance et al., 1984; Revert and Tchoumatchenco, 1973); Spain (García-Joral et al., 1990; García-Joral and Goy, 1994; Goy et al., 1984); for Morocco (Alméras, 1993); Algeria (Elmi et al., 1985; Tchoumatchenco, 1984); Hungary (Vörös, 1983; 1984); Bulgaria (Revert and Tchoumatchenco, 1973; Tchoumatchenco, 1971; 1972; 1975; 1977; 1984); Romania (Georgescu, 1990); Central Asia (Prozorovskaya, 1986; 1993; Prozorovskaya and Rostovtsev, 1984); China (Sun, 1986, Sun in Wang, 1988; Shi and Yang, 1990); New Zealand (MacFarlan, 1992); Argentina (Manceñido, 1990; 1994; Riccardi et al., 1991; 1994).

As implicitly recorded and sometimes explicitly acknowledged, in most of these studies, among brachiopod orders the Rhynchonellida stand out as particularly important for biostratigraphic purposes, and are thus the subject of the present synthesis. The number of described Jurassic rhynchonellide genera in the first edition of the Treatise on Invertebrate Paleontology (Ager, 1965) was 74 (plus 10 synonyms) which were grouped in 4 families. Subsequently, Ager et al. (1972) recorded 94 valid genera belonging to 7 families. In both cases only two superfamilies were recognized.

The revised data-base employed in this survey, which includes first and last appearances for each genus as revealed by their referred species, contains significant improvements over that compiled and used by Sulser (1993; 1996). A total of 150 units at generic/subgeneric levels have been classified into 7 superfamilies and 14 families (including 31 subfamilies). Further information on the historical development of the systematics of Jurassic Rhynchonellida may be found in Shi and Grant (1993); for bibliographic references, see also Doescher (1981).

State of Knowledge

Members of the Dimerelloid superfamily (Fig. 1) are often found locally and linked to oceanic cold seep settings. The Rhynchonellininae in particular are typically Mediterranean in distribution, and at least one Pliensbachian biozone has been named after a species of *Sulcirostra* for Morocco (Alméras, 1993). Among Norelloids, *Ochotorhynchia* appears restricted to Sinemurian deposits in Siberia (Dagis, 1968; Manceñido and Dagis, 1992), whereas the norellines *Pisirhynchia* and *Kericserella* provide good stratigraphical markers for the Pliensbachian of Mediterranaean areas (Vörös, 1983; 1984), some species of *Nannirhynchia* are indicative of lowest Toarcian in southwestern Europe, species of *Praemonticlarella* seem confined to the Late Toarcian/Early Aalenian of the same region and species of *Monticlarella* are valuable for the Late Jurassic of central Europe.

Among basiliolid pugnacoids (Fig. 2), species of *Apringia* characterize different segments of the Mediterranean Lower and Middle Jurassic column (Vörös, 1983; 1984), those of *Soaresirhynchia* allow subdivision of the Toarcian of southwest Europe-north Africa, and those of *Fortunella* help characterize the Late Jurassic of southern and eastern Europe-north Africa. The successive generic series from *Stolmorhynchia* to *Lacunosella* provides guide fossils for various levels within the Aalenian-Volgian interval from Britain to the Caucasus. Among erymnariid pugnacoids, biostratigraphic value has been recognized for species of *Calvirhynchia* from the Middle Jurassic of the Caucasus, for species of *Caucasella* from the Late Bajocian-Callovian of Europe and north Africa, for species of *Septocrurella* from the Bathonian-Oxfordian of southern and eastern Europe, whereas other genera are more local in distribution.

Lingering from Triassic faunas, cirpine wellerelloids are mainly typical of Liassic deposits, with *Vincentirhynchia* characterizing the Hettangian of the west Pacific, species of *Calcirhynchia* in the Hettangian-Sinemurian of Europe and north Africa, and species of *Cirpa* widespread in the Pliensbachian of Europe, Asia, Australasia and South America (though the earliest appeared in the Hettangian of the European Alps). Among prionorhynchiid rhynchotetradoids, good index fossils are furnished by species of *Prionorhynchia* from the Sinemurian to Aalenian of Europe, across north Africa and Asia to the Andes, by species of *Lokutella* from the Late Sinemurian to Pliensbachian of Mediterranean Europe, and by species of *Sphenorhynchia* from the Bajocian-Callovian of Europe, north Africa and Asia.

Rhynchonellids (*s. s.*) also yield useful stratigraphic markers (Figs. 3 and 4): species of *Homoeorhynchia* have been successfully used to define various zones of the Sinemurian-Aalenian interval in the Iberian Peninsula, north Africa, Bulgaria and China; successive species of *Herangirhynchia* occur in the Aratauran-Ururoan sequence of New Zealand (MacFarlan, 1992). Similarly, successive species of *Rhynchonelloidea* have allowed subdivision of the Pliensbachian-Aalenian of the Andes. Certain species of *Piarorhynchia* are locally useful within the Pliensbachian-Toarcian, as in western Europe, north Africa and Argentina, whereas a similar role is played in the west Pacific by species of *Sakawairhynchia*, and *Ptilorhynchia* has been used as an index fossil for the Callovian of North America (Crickmay, 1933). Similarly, ivanoviellines become useful biostratigraphic tools for the Bathonian-Oxfordian interval: species of *Rhynchonelloidella* in NW Europe, China and South America, species of *Ivanoviella* over most of Europe to the Himalayas, and those of *Thurmannella* in western Europe, SE Asia and South America. On the other hand, *Bilaminella* is mainly restricted to the Callovian of the Crimea-Caucasus; the incoming of *Furcirhynchia* is typical in the Hettangian of such distant areas as the Austrian Alps, New Zealand and Argentina, with other species persisting into Pliensbachian-Toarcian deposits of Europe and the circum-Pacific; species of *Rimirhynchia* characterize the late Sinemurian-Pliensbachian of boreal areas; those of *Cubanirhynchia* and *Davanirhynchia* occur in the Middle Jurassic of central Asia and most striirhynchines help recognition of particular Middle Jurassic levels over most of Europe, the Caucasus, Transcaucasia and Arabia. Spinose rhynchonellids are widespread in the middle Jurassic, and some such as *Acanthothiris* have been used for biozonations in extra-Tethyan Europe (as in Bulgaria by Tchoumatchenco, 1977; 1984): other genera also seem more restricted areally.

Tetrarhynchiids (Fig. 5) comprise most ordinary-looking Mesozoic brachiopods and are also widely used in biostratigraphy, as testified by species of *Tetrarhynchia, Quadratirhynchia* and *Gibbirhynchia* in the Early Jurassic; or species of *Cymatorhynchia, Daghanirhynchia, Goniorhynchia, Pycnoria, Burmirhynchia, Kallirhynchia* and *Isjuminella* in the Middle Jurassic and by species of *Somalirhynchia, Russirhynchia* and *Mosquella* in the Late Jurassic. Among cyclothiridids (Fig. 6), species of *Fissirhynchia* are distinctive in the Hettangian-Pliensbachian interval of Europe, Asia, Australasia and South America; the incoming of *Rudirhynchia* characterizes the early Pliensbachian of southern France, the Iberian Peninsula and the Andes; in the Aalenian-Bajocian valuable genera (in Europe, both Americas and Asia) are: *Globirhynchia* and *Lacunaerhynchia*, whereas for the Late Jurassic a similar role is played by species of *Torquirhynchia* and *Septaliphoria*. Among cardinirhynchiines species of *Flabellirhynchia* are important for the Aalenian-Bajocian of Europe, both Americas and Antarctica, species of *Parvirhynchia* (such as *P. parvula*) are typical for the Bajocian of Europe, North Africa and South America (although appearing in Spain already in the Aalenian); and species of *Cardinirhynchia* and *Eurysites* help characterize the Bajocian-Callovian of Europe and Asia (the latter reaching South America too). All indorhynchiines seem confined to Bajocian-Callovian of central and southern Asia, whereas all septirhynchiids are conspicuous elements of Callovian faunas from north to east Africa, the Middle East and Italy.

References

AGER, D. V. 1965. Mesozoic and Cenozoic Rhynchonellacea. p. H597-H632. *In* R. C. Moore (ed.), Treatise on Invertebrate Paleontology, Part H, Brachiopoda. Geological Society of America and University of Kansas Press, Lawrence.

____. 1979. The stratigraphical distribution of Jurassic Brachiopoda. Recent Researches in Geology, 4:6-21. Delhi.

____. 1994. Brachiopod Stratigraphy in the Jurassic. Geobios, Mémoire Spécial, 17:57-68. Lyon.

____, A. CHILDS, AND D. A. B. PEARSON. 1972. The evolution of the Mesozoic Rhynchonellida. Geobios, 5:157-235. Lyon.

ALMÉRAS, Y. 1993. The Liassic brachiopod zones of the Middle Atlas, Morocco: comparison with the French North Tethyan zonation, p. 15-21. *In* J. Pálfy and A. Vörös (eds.), Mesozoic Brachiopods of Alpine Europe. Hungarian Geological Society, Budapest.

____, A. BOULLIER, AND B. LAURIN. 1991. Les zones de brachiopodes du Jurassique en France. Annales Scientifiques de l'Université de Franche-Comté, Géologie, (1990), 4:3-30. Besançon.

____, ____, AND ____. 1994. La zonation du Jurassique en Francais par les brachiopodes: limites de

____, AND G. MOULAN. 1979. Biostratigraphie des Terebratulidae et des Zeilleriidae du Lias et du Dogger de la région toulonnaise (Var, France). Geobios, 12:893-897. Lyon.

résolution. Geobios, Mémoire Spécial, 17:69-77. Lyon.

CRICKMAY, C. H. 1933. An attempt to zone the North American Jurassic on the basis of its brachiopods. Bulletin of the Geological Society of America, 44:871-893. Washington.

DAGIS, A. S. 1968. Yurskie i Rannemelovye brakhiopody severa Sibiri. Tr. Inst. Geol. Geofiz., 41:1-170.

DELANCE, J.-H., Y. ALMÉRAS, B. LAURIN, AND G. MOULAN. 1984, p. 127-128. *In* S. Debrand Passard, S. Curboleix and M.J. Lienhardt (eds.), Synthèse Géologique du Sud -Est de la France. Mémoires du B.R.G.M., 125 (I). Orléans.

DOESCHER, R. A. 1981. Living and fossil brachiopod genera 1775-1979. Lists and bibliography. Smithsonian Contributions to Paleobiology, 42:1-238. Washington

ELMI, S., Y. ALMÉRAS, M. AMEUR, AND M. BENHAMOU. 1985. Précisions biostratigraphiques et paléoécologiques sur le Lias des environs de Triffrit (Saïda, Algérie occidentale). Cahiers du Institut Catholique de Lyon, 14:15-41, Lyon.

GARCÍA-JORAL, F., AND A. GOY. 1994. The associations of brachiopods from the Toarcian-Aalenian transition in the Fuentelsaz section (Iberian Range, Spain). Geobios, Mémoire Spécial, 17: 223-228. Lyon

____, A. GOY, AND M. S. URETA. 1990. Las sucesiones de braquiópodos en el tránsito Lias-Dogger en la Cordillera Ibérica. Cuadernos de Geología Ibérica, 14:55-65.

GEORGESCU, M.-D. 1990. An attempt to the correlation of the brachiopod biozones of the Northern Apuseni Mountains (Romania). Crisia, 20:517-541.

GOY, A., M. J. COMAS-RENGIFO, AND F. GARCÍA-JORAL. 1984. The Liassic brachiopods of the Iberian Range (Spain): stratigraphic distribution and zonation, p. 227-250. *In* O. Michelsen and A. Zeiss (eds.), Symposium on Jurassic Stratigraphy, Erlangen, vol. 1. Geological Survey of Denmark, Copenhagen.

MACFARLAN, D. A. B. 1992. Triassic and Jurassic Rhynchonellacea (Brachiopoda) from New Zealand and New Caledonia. Bulletin of the Royal Society of New Zealand, Wellington, 31:1-310.

MANCEÑIDO, M. O. 1990. The succession of Early Jurassic brachiopod faunas from Argentina: correlation and affinities. p. 397-404. *In* D. MacKinnon, D. Lee and J. D. Campbell (eds.), Brachiopods through Time. A.A. Balkema, Rotterdam.

____. 1994. Jurassic brachiopod biostratigraphy of western Argentina: overview and relations. 4th International Congress on Jurassic Stratigraphy and Geology, Abstracts: 28-29. Mendoza, Argentina.

____, AND A. S. DAGIS. 1992. Brachiopods of the circum-Pacific region, p. 328-333, 598-609. *In* G. E. G. Westermann (ed.), The Jurassic of the circum-Pacific. Cambridge University Press, Cambridge (U.K.)

PROZOROVSKAYA, E. L. 1986. Brachiopod biozones in the Jurassic of the South USSR. Biostratigraphie du Paléozoïque, 4:357-361. Brest.

____. 1993. Brachiopod subdivisions in the Jurassic of the southern ex-USSR. Palaeogeography, Palaeoclimatology, Palaeoecology, 100:183-188. Amsterdam.

____, AND K.O. ROSTOVTSEV. 1984. On Jurassic zones of the South of the USSR, p. 795-809. *In* O. Michelsen and A. Zeiss (eds.), Symposium on Jurassic Stratigraphy, Erlangen, vol. 3. Geological Survey of Denmark, Copenhagen.

REVERT, J., AND P. TCHOUMATCHENCO. 1973. Les zones de brachiopodes du Lias en Bulgarie et dans le Sud de la France (Bassin des Causses): essai de corrélation. Documents du Laboratoire de Géologie, Faculté des Sciences de Lyon, 56:181-193. Lyon.

RICCARDI, A. C., S. E. DAMBORENEA, M. O. MANCEÑIDO, AND S. C. BALLENT. 1991. Hettangian and Sinemurian (Lower Jurassic) biostratigraphy of Argentina. Journal of South American Earth Sciences 4:159-170.

____, ____, ____, AND ____. 1994. Middle Jurassic Biostratigraphy of Argentina. Geobios. Mémoire Spécial, 17:423-430. Lyon.

SHI, X.-Y., AND R. E. GRANT. 1993. Jurassic Rhynchonellids: Internal structures and taxonomic revisions. Smithsonian Contributions to Paleobiology, 73:1-190. Washington

SHI, X., AND Z. YANG. 1990. The Jurassic brachiopod sequences of the Qinghai-Tibetan Plateau, p. 405-413. *In* D. MacKinnon, D. Lee and J. D. Campbell (eds.), Brachiopods through Time. A.A. Balkema, Rotterdam.

SULSER, H. 1993. Brachiopoda: Rhynchonellida mesozoica, p.1-281. *In* F. Westphal (ed.), Fossilium Catalogus. I: Animalia. Pars 132. Kugler Publications, Amsterdam.

____. 1996. Notes on the taxonomy of Mesozoic Rhynchonellida, p. 265-268. *In* P. Copper and J. Jin (eds.), Brachiopods. A.A. Balkema, Rotterdam.

SUN, D.-L. 1986. The stratigraphical succession of Jurassic and Cretaceous Brachiopoda of China. Biostratigraphie du Paléozoïque, 4:491-500. Brest.

TCHOUMATCHENCO, P. V. 1971. Notes sur la répartition stratigraphique des brachiopodes du Jurassique inférieur dans les Balkanides centrales et occidentales (Bulgarie). Mémoires du B.R.G.M., 75:183-187. Paris.

____. 1972. Stratigraphical distribution of Lower Jurassic brachiopods in the central and western Balkans and in the Krayshte. Bulgarian Academy of Sciences, Bulletin of the Geological Institute, Series Stratigraphy and Lithology, 21:63-84. Sofia.

____. 1975. On the stratigraphy of Middle Jurassic brachiopods in the Belogradchik region. Bulgarian Academy of Sciences, Geological Institute, p. 314-319. Sofia.

____. 1977. Les biozones de brachiopodes du Jurassique moyen en Bulgarie. Geologica Balkanica, 7:97-108. Sofia.

____. 1984. Les zones de brachiopodes du Jurassique d'Algerie du Nord et leur correlation avec les zones de brachiopodes en Bulgarie, p. 863-882. *In* O. Michelsen and A. Zeiss (eds.), Symposium on Jurassic Stratigraphy, Erlangen, vol. 3. Geological Survey of Denmark, Copenhagen.

VÖRÖS, A. 1983. The Pliensbachian brachiopods of the Bakony Mts. (Hungary): a stratigraphical study. Fragmenta Mineralogica et Palaeontologica, 11:29-39. Budapest.

____. 1984. The Pliensbachian brachiopod zones of the Bakony Mts. /Hungary/ and their correlation with other European areas, p. 295-301. *In* O. Michelsen and A. Zeiss (eds.), Symposium on Jurassic Stratigraphy, Erlangen, vol. 1. Geological Survey of Denmark, Copenhagen.

WANG, Y.-G. 1988. China, p. 95-130. *In* G. E. G. Westermann and A. C. Riccardi (eds.), Jurassic taxa ranges and correlation charts for the Circum-Pacific. 2. Newsletters on Stratigraphy, 19. Stuttgart.

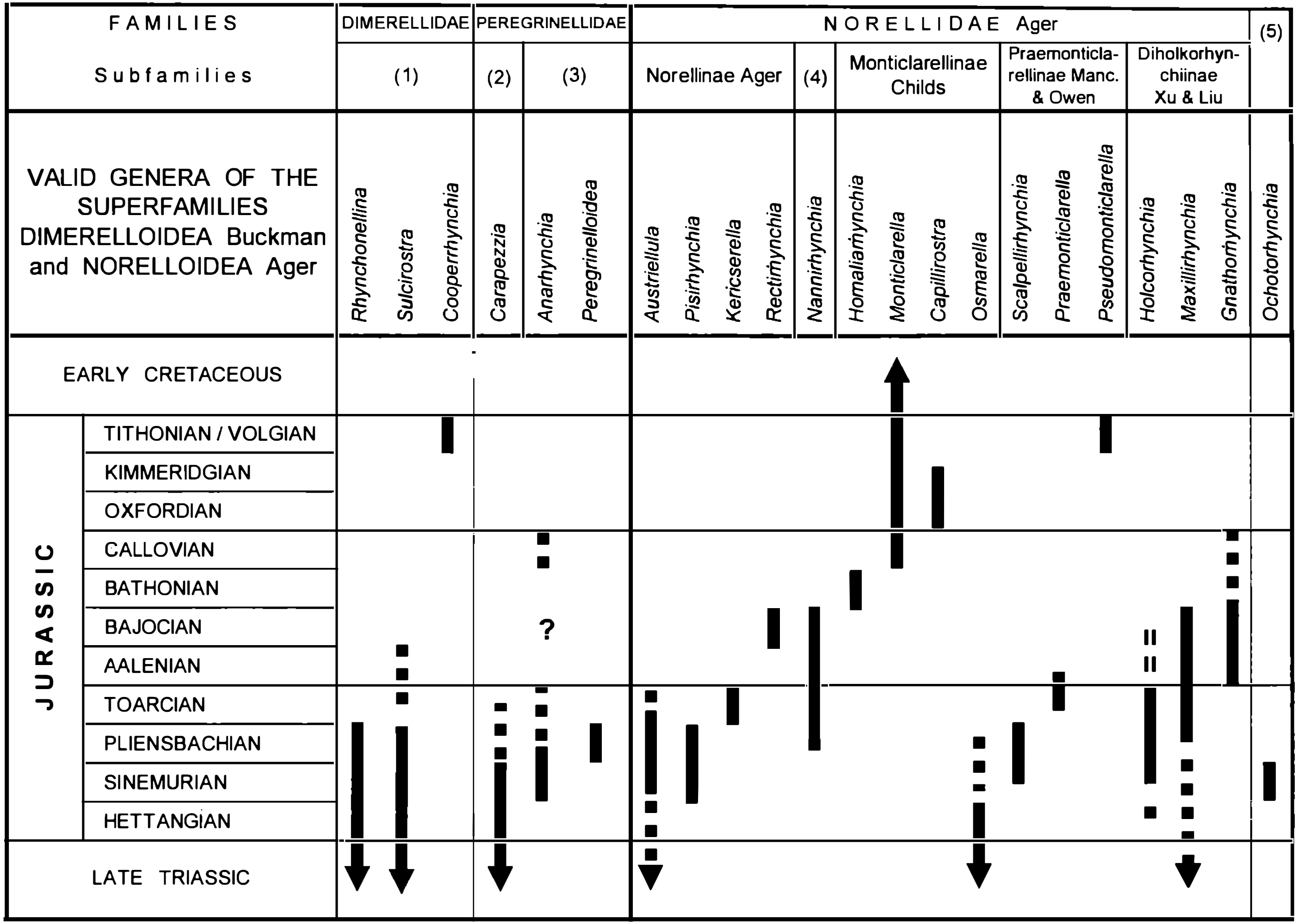
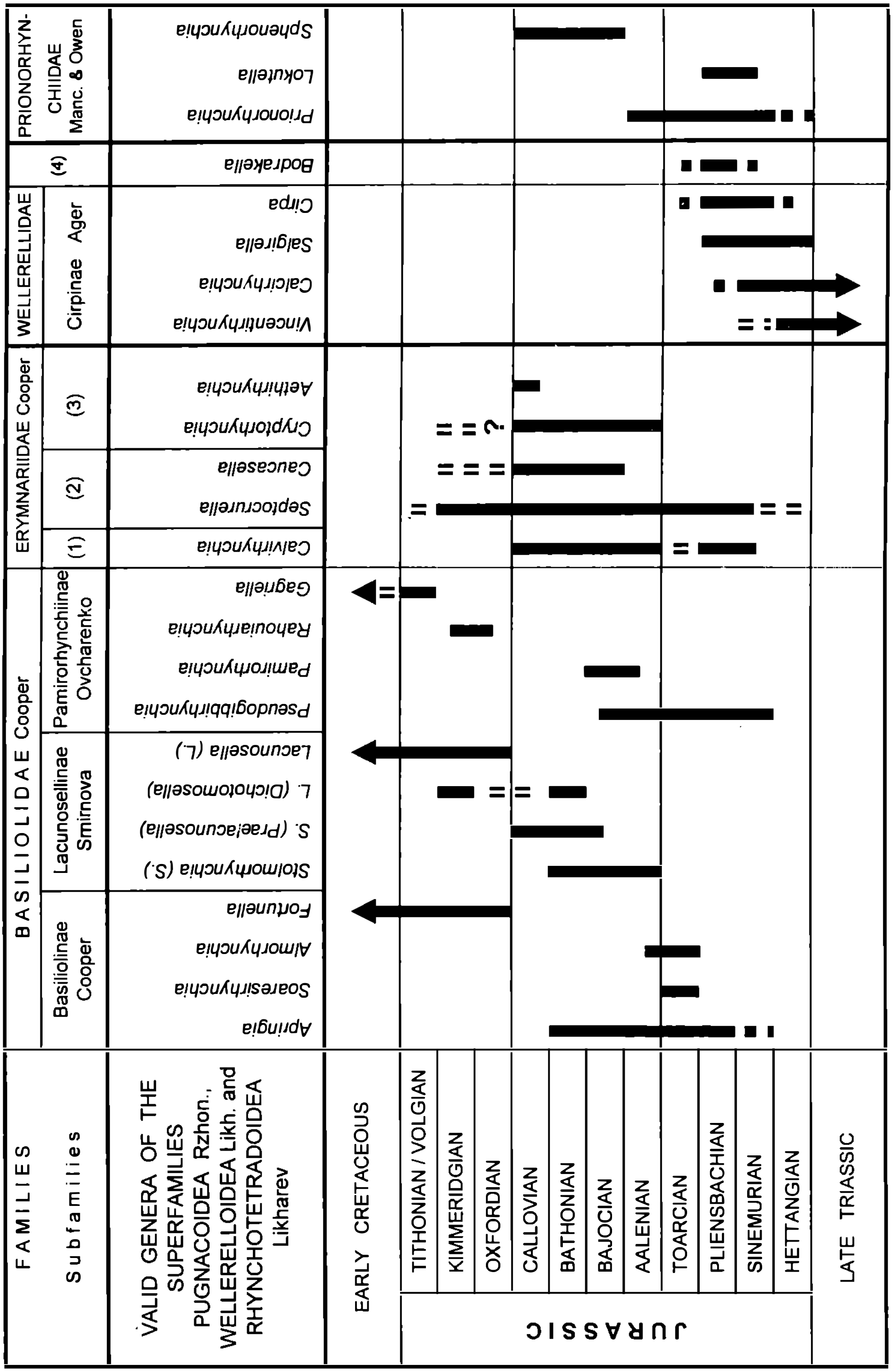


Figure 1.

(1) Rhynchonellininae Ager
(2) Peregrinellinae Ager
(3) Peregrinelloideinae Dagis
(4) Laevirhynchiinae Dagis
(5) OCHOTORHYNCHIIDAE Dagis



(1) Calvirhynchiinae Kamyshan (2) Septocrurellinae Ager, Childs & Pearson (3) Cryptorhynchiinae Shi & Grant (4) PONTISIIDAE Cooper & Grant

Figure 2.

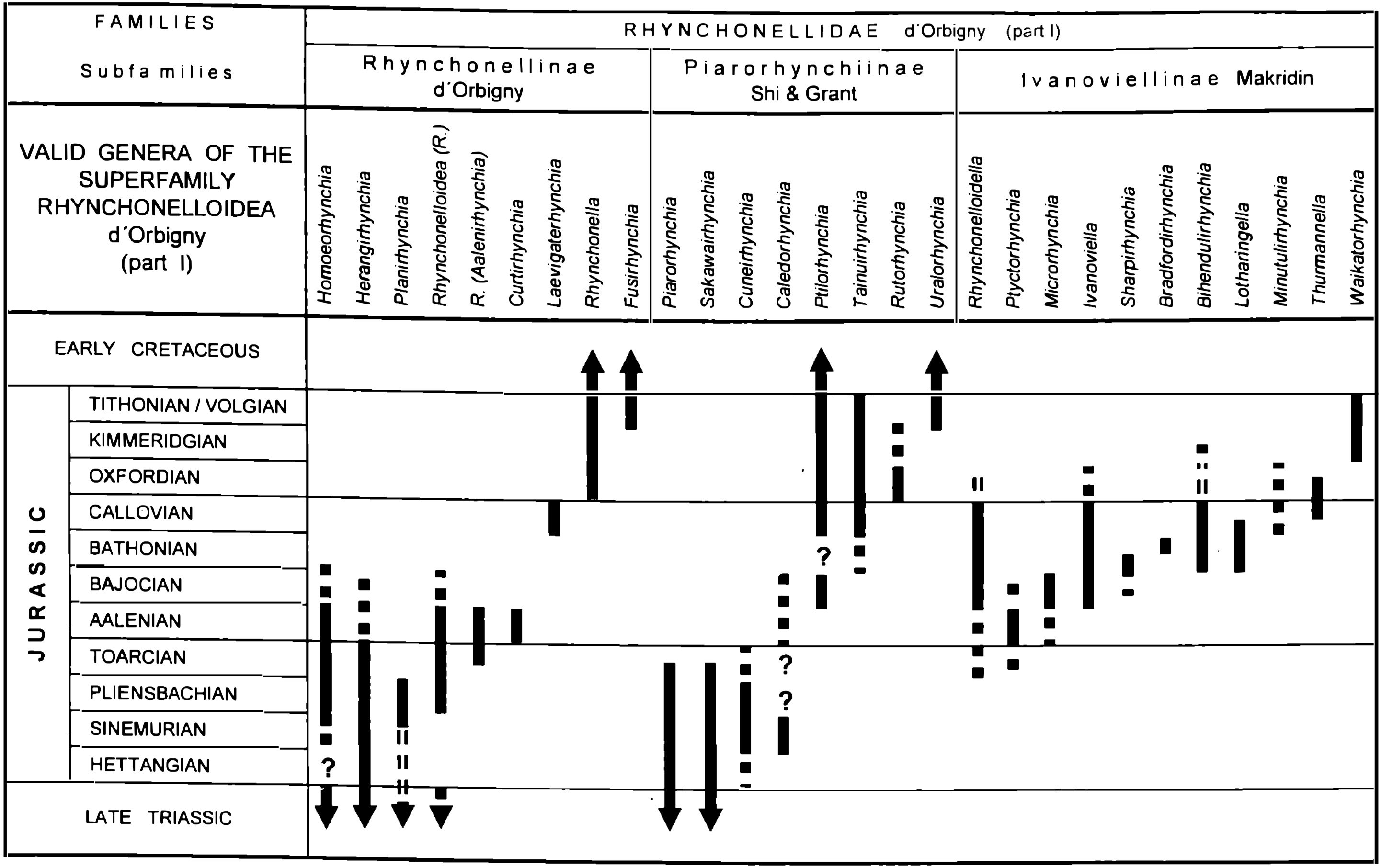

FAMILIES
Subfamilies
RHYNCHONELLIDAE d'Orbigny (part I)
Rhynchonellinae d'Orbigny
Piarorhynchiinae Shi & Grant
Ivanoviellinae Makridin
VALID GENERA OF THE SUPERFAMILY RHYNCHONELLOIDEA d'Orbigny (part I)
Homoeorhynchia
Herangirhynchia
Planirhynchia
Rhynchonelloidea (R.)
R. (Aalenirhynchia)
Curtirhynchia
Laevigaterhynchia
Rhynchonella
Fusirhynchia
Piarorhynchia
Sakawairhynchia
Cuneirhynchia
Caledorhynchia
Ptilorhynchia
Tainuirhynchia
Rutorhynchia
Uralorhynchia
Rhynchonelloidella
Ptyctorhynchia
Microrhynchia
Ivanoviella
Sharpirhynchia
Bradfordirhynchia
Bihendulirhynchia
Lotharingella
Minutulirhynchia
Thurmannella
Waikatorhynchia
EARLY CRETACEOUS
JURASSIC
TITHONIAN / VOLGIAN
KIMMERIDGIAN
OXFORDIAN
CALLOVIAN
BATHONIAN
BAJOCIAN
AALENIAN
TOARCIAN
PLIENSBACHIAN
SINEMURIAN
HETTANGIAN
LATE TRIASSIC
?

Figure 3.

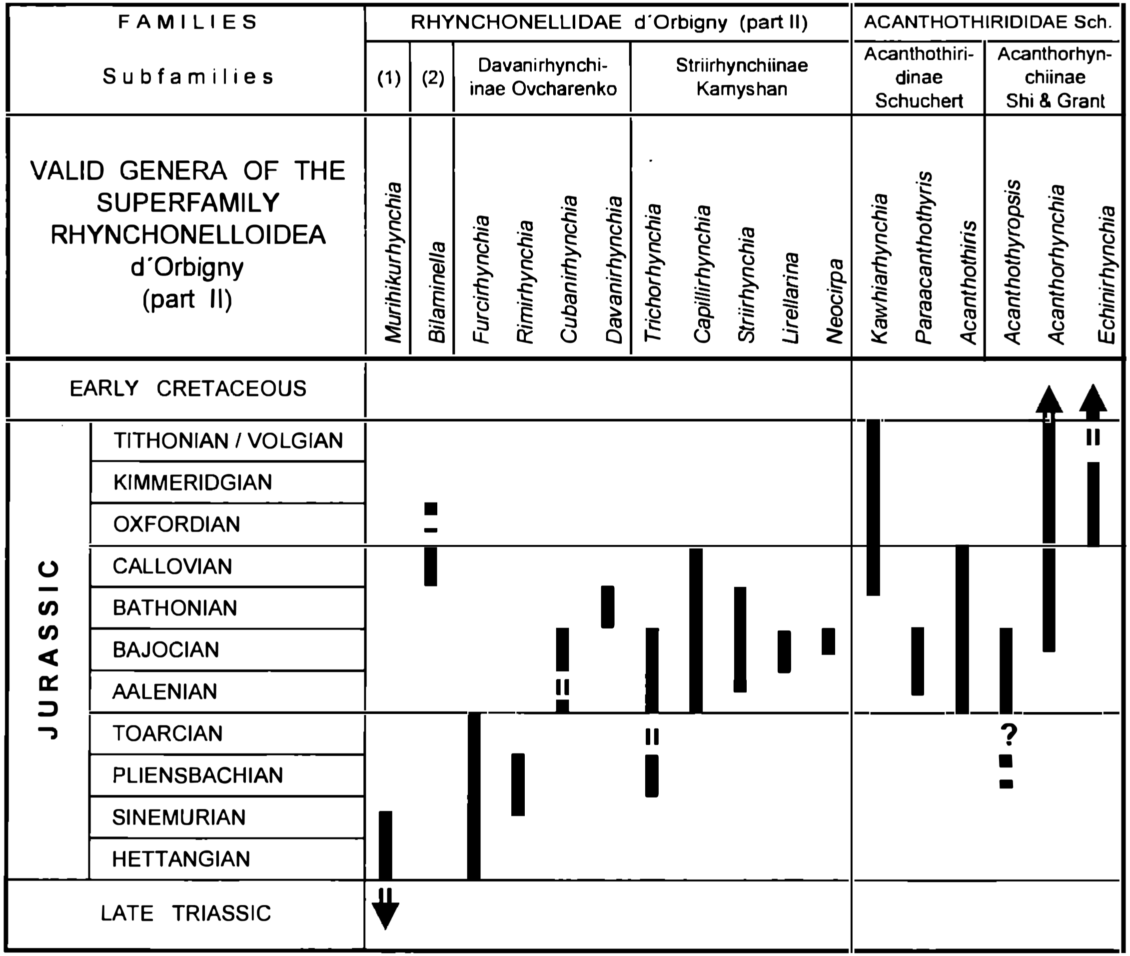


Figure 4.

(1) Nucleusorhynchiinae Xu (2) Bilaminellinae Babanova

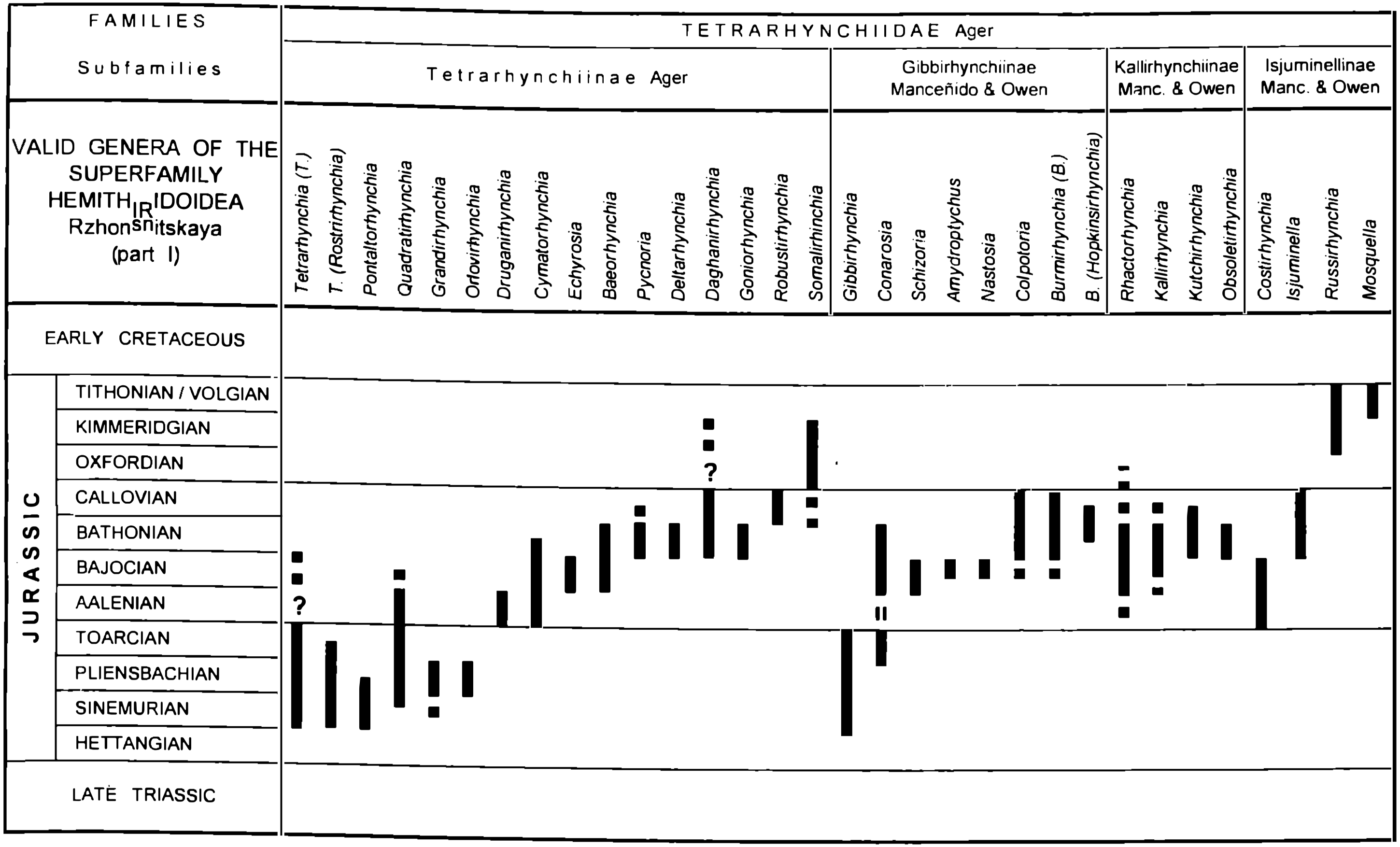

FAMILIES
Subfamilies
VALID GENERA OF THE SUPERFAMILY HEMITHIRIDOIDEA Rzhonsnitskaya (part I)
EARLY CRETACEOUS
JURASSIC
TITHONIAN / VOLGIAN
KIMMERIDGIAN
OXFORDIAN
CALLOVIAN
BATHONIAN
BAJOCIAN
AALENIAN
TOARCIAN
PLIENSBACHIAN
SINEMURIAN
HETTANGIAN
LATE TRIASSIC
TETRARHYNCHIIDAE Ager
Tetrarhynchiinae Ager
Gibbirhynchiinae Manceñido & Owen
Kallirhynchiinae Manc. & Owen
Isjuminellinae Manc. & Owen
Tetrarhynchia (T.)
T. (Rostrirhynchia)
Pontaltorhynchia
Quadratirhynchia
Grandirhynchia
Orlovirhynchia
Druganirhynchia
Cymatorhynchia
Echyrosia
Baeorhynchia
Pycnoria
Deltarhynchia
Daghanirhynchia
Goniorhynchia
Robustirhynchia
Somalirhinchia
Gibbirhynchia
Conarosia
Schizoria
Amydroptychus
Nastosia
Colpotoria
Burmirhynchia (B.)
B. (Hopkinsirhynchia)
Rhactorhynchia
Kallirhynchia
Kutchirhynchia
Obsoletirhynchia
Costirhynchia
Isjuminella
Russirhynchia
Mosquella
?
?

Figure 5.

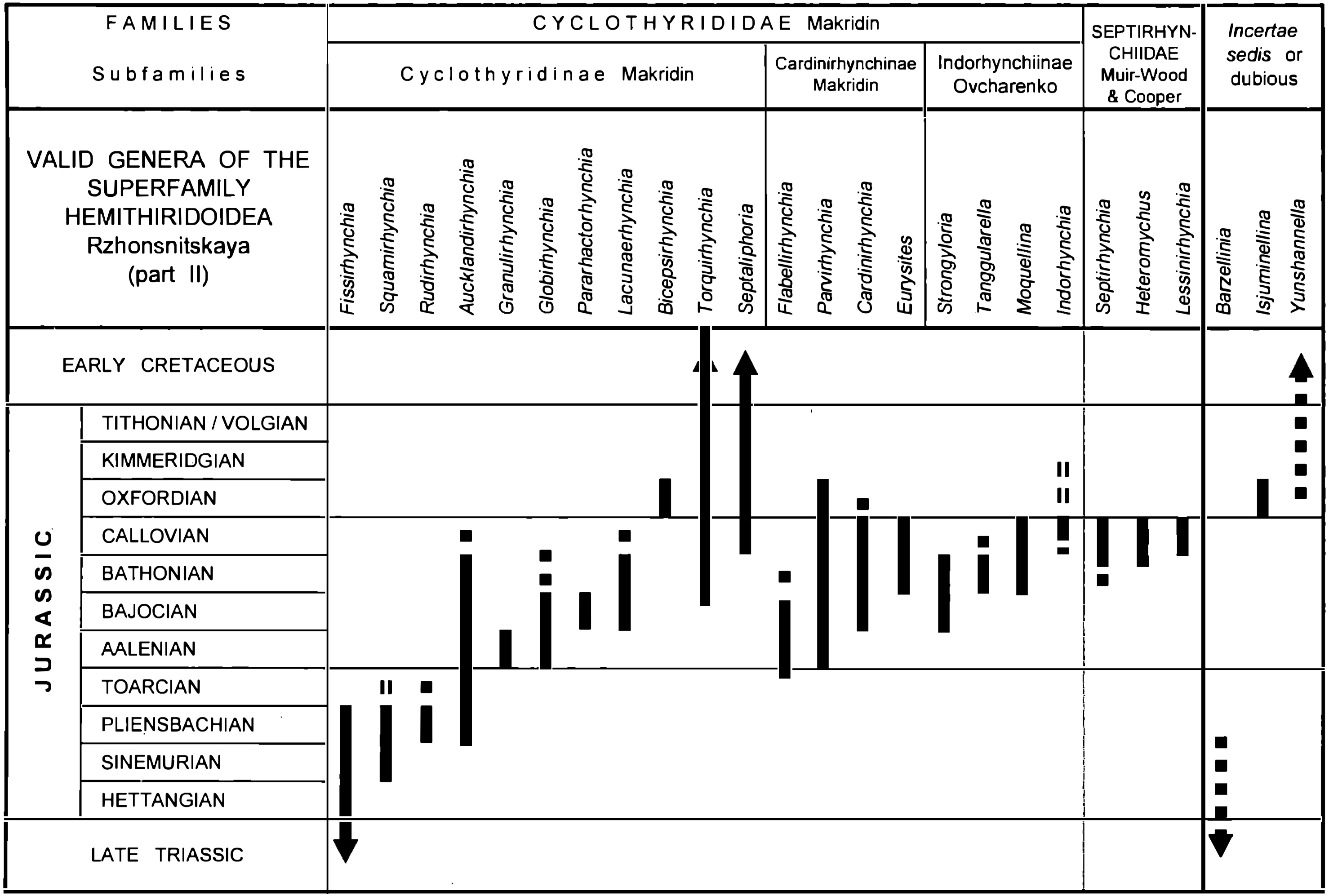

FAMILIES
Subfamilies
CYCLOTHYRIDIDAE Makridin
Cyclothyridinae Makridin
Cardinirhynchinae Makridin
Indorhynchiinae Ovcharenko
SEPTIRHYNCHIIDAE Muir-Wood & Cooper
Incertae sedis or dubious
VALID GENERA OF THE SUPERFAMILY HEMITHIRIDOIDEA Rzhonsnitskaya (part II)
Fissirhynchia
Squamirhynchia
Rudirhynchia
Aucklandirhynchia
Granulirhynchia
Globirhynchia
Pararhactorhynchia
Lacunaerhynchia
Bicepsirhynchia
Torquirhynchia
Septaliphoria
Flabellirhynchia
Parvirhynchia
Cardinirhynchia
Eurysites
Strongyloria
Tanggularella
Moquellina
Indorhynchia
Septirhynchia
Heteromychus
Lessinirhynchia
Barzellinia
Isjuminellina
Yunshannella
EARLY CRETACEOUS
JURASSIC
TITHONIAN / VOLGIAN
KIMMERIDGIAN
OXFORDIAN
CALLOVIAN
BATHONIAN
BAJOCIAN
AALENIAN
TOARCIAN
PLIENSBACHIAN
SINEMURIAN
HETTANGIAN
LATE TRIASSIC

Figure 6.

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New *Subcollina* (Ammonitida) from the Topmost Lower Bajocian: their Phylogenetic and Paleogeographic Significance

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Keywords: *Subcollina*, Ammonitida, Bajocian, Paleobiogeography, Hispanic Corridor

Abstract: The genus *Subcollina* is revised on the basis of material from the Eastern Pacific and Western Mediterranean mid-Bajocian. The Mexican specimens misidentified as *S. lucretii* are renamed *S. sandovali* n. sp. Another species, *S. ochoterenai* n. sp., is described from the uppermost Lower Bajocian of the Digne area, SE France. The latter taxon shows features intermediate between *Subcollina* (subquadratic whorl section, zigzag secondary ribs and feeble furrow on venter) and *Parastrenoceras* (serpenticone, nodes at mid-ventral furrow); it represents the phylogenetic link between these genera, possibly starting from the type-species *S. yeovilensis* to some Late Bajocian *Parastrenoceras*. Distribution of both genera in southwestern Mexico and western Europe is a strong paleobiogeographic argument for two-way faunal dispersal between the Eastern Pacific and Western Tethyan Provinces through the "Hispanic Corridor" during the mid-Bajocian. Strong morphological resemblances between *Subcollina sandovali* and *Phaulostephanus burckhardti* support their relation by cladogenesis in the Mexican Lower Bajocian. *Phaulostephanus* was a "plastic" stephanoceratid giving rise to *Parastrenoceras* via *Subcollina*, *Caumontisphinctes*, and *Leptosphinctes* at different times and more or less directly at the Early/Late Bajocian boundary.

Introduction

Marine faunal exchanges during the Early and Middle Jurassic between the Eastern Pacific and Western Tethyan Provinces through the Hispanic Corridor (Smith, 1983) have been debated in recent years in attempting to explain the close affinity of their fossil assemblages (Smith and Tipper, 1986). Such a marine, bi-directional, albeit intermittent, seaway has been supported with paleobiological and dynamic arguments mainly by North American workers (Westermann, 1993), while European workers suggest migration along the continental masses bordering the Pacific and controlled by sea level changes (Elmi, 1992). Widespread faunal communication is generally accepted as beginning only from Callovian time, when separation between North and South America became effective with the production of oceanic crust in the Central Atlantic (Dercourt et al., 1993).

Any discussion on this subject should take into account both aut-ecology and dispersal potential of organisms, i.e., their migration capacity through epeirogenic seas which are referred to as the Hispanic Corridor. In this respect ammonites can be very useful and, according to Westermann (1992, p. 355), the corridor acted as a restrictive dispersal way for "trachyostracan" ammonites during the Jurassic until the Middle Bajocian, while "leiostracan" taxa were filtered out because of the absence of deep-water or oceanic environments. Meaningful data supporting this interpretation come from the Bajocian successions of Oaxaca, southwestern Mexico (Sandoval and Westermann, 1986), which correspond to the Pacific entrance to the Hispanic Corridor. These authors refer to nearly 30 percent ammonite affinity between the Antofagasta Subprovince and Mediterranean Province based on taxa like *Subcollina* and *Parastrenoceras*, among others. Obviously any new information coming from both the western and the eastern entrances of the seaway can improve such conclusions.

The purpose of this paper is to discuss species referable to the genus *Subcollina*. The starting point is *S. ochoterenai* n. sp. from the middle part of the Calcaires à *Cancellophycus* Formation of SE France (Fig. 1; Pavia, 1983). This new taxon shows a morphology so clearly intermediate between *Subcollina* and *Parastrenoceras* that it can be regarded as the evolutionary link between them, supplying evidence for active exchange through the Hispanic Corridor in the latest Early Bajocian as well.

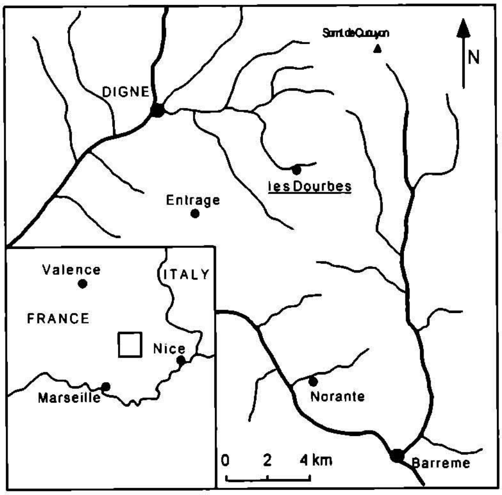

Figure 1. Sketch map of the Digne area, SE France, with the "les Dourbes" type-section of *Subcollina ochoterenai* n. sp.

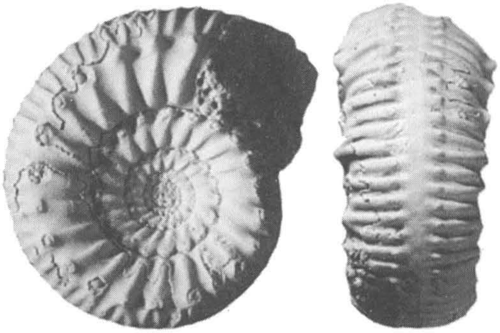

Figure 2. *Parastrenoceras* sp. ind., from the base of the Oolitė de Bayeux Formation, Evrecy, NW France, Upper Bajocian. x 2.

Systematic Paleontology

Suborder AMMONITINA Hyatt, 1889
Superfamily STEPHANOCERATACEAE Neumayr, 1875
?Family STEPHANOCERATIDAE Neumayr, 1875
Genus SUBCOLLINA Spath, 1925

Type species. – *Subcollina yeovilensis* Spath, 1925, for *Aegoceras densinodum* Wright, 1880-1882 (*non* Quenstedt, 1849).

Generalities. – The genus *Subcollina* was established by Spath (1925, p. 171) who renamed as *S. yeovilensis* the ammonite from Yeovil, Dorset, wrongly attributed to *A. densinodum* by Wright (1880-1882, p.350, pl. 38, Figs. 5, 6; see also Sandoval and Westermann, 1986, Figs. 30.1-3). The chronostratigraphy of this specimen changed greatly. Wright suggested Lower Pliensbachian, while Donovan (1954, p. 38) believed it could be a Toarcian dactylioceratid. More recently, Torrens (*in* Sturani, 1971, p. 164) concluded that the "type-horizon must be somewhere in the Inferior Oolite, ... notably the Irony Bed (highly condensed Subfurcatum and Humphriesianum zones)", i.e. the middle part of the Bajocian. Subsequent authors supported such a biostratigraphic statement (Sturani, 1971; Sandoval and Westermann, 1986; Rioult, in Fischer, 1994).

No additional specimens of *S. yeovilensis* having been encountered, and its essential morphology derives from the holotype (London Natural History Museum, inv. C1932): evolute, planulate to serpenticone shell with rectangular whorl section marked by angular ventrolateral edges and almost flat venter; dense primary ribs (22 on the last half-whorl at diameter of 110mm) ending in a row of pointed nodes (spines); two or three evanescent secondaries arranged in zigzag patterns because of alternation of spines on the two whorl sides; mid-ventral weakening of ribs.

A second species, *S. lucretii* (d'Orbigny, 1850), was assigned to *Subcollina* by Sandoval and Westermann (1986). Its features resemble those of the type-species with some important differences: there are only one or two secondary ribs per primary and no median furrow, so that secondaries run with a clear zigzag path (Sturani, 1971). *S. lucretii* was well described by d'Orbigny (1850, p. 262) based on unfigured specimen(s) from the Bajocian of Normandy: "Espèce petite à tours carrés à decouvert, ornée de côtes simples qui ... se terminent en dehors en une pointe d'ou partent deux côtes qui passent sur le dos où elles forment des zigzags réguliers". Unfortunately, the type was already missing at the end of the last century (Haug, 1891, p. 76). *S. lucretii* has been repeatedly misidentified by subsequent authors except Sturani (1971, p.162) who made a detailed description of the taxon from the lower Upper Bajocian of Asiago and stated that the specimen described by Bentz (1928, pl. 15, Fig. 2: see under *S. ochoterenai* n. sp. described below) cannot be d'Orbigny's type as supposed by Ochoterena (1963, pl. 5, Fig. 3); consequently, the species is still in need of a neotype.

S. lucretii was believed to be present also in the mid-Bajocian of southern Mexico (Sandoval and Westermann, 1986), but, due to significant differences, those specimens belong to a different taxon, *S. sandovali* n. sp. A fourth species, *S. ochoterenai* n. sp., is described from southern France. For both the new species, see below. With the broadening of the spectrum of species, the diagnosis of the genus *Subcollina* is here emended after Sandoval and Westermann (1986, p. 1258).

Diagnosis. – Size medium to small. Coiling very evolute to planulate; whorls subquadratic to gently convex; well marked ventrolateral edges; venter flat to slightly arched. Primaries straight, recti- to prorsiradiate, terminating in ventrolateral nodes or spines; two or three, rarely one, secondaries per node crossing the venter in zigzag pattern because of alternation of nodes on both sides; possible feeble median furrow or flattening of secondary ribs at mid-venter.

Remarks. – As far as dimorphism is concerned, the holotype of *S. yeovilensis* (D = 112mm) is a macroconch with trumpet-like peristome. If the Mexican *S. sandovali* n. sp. is indeed dimorphic (see below), then the dimorphism of *Subcollina* is like that of *Phaulostephanus* (Pavia, 1983), but no definitive statement can be made so far.

Mid-Bajocian ammonites similar to *Subcollina* were described by Ochoterena (1963) from southwestern Mexico as *Parastrenoceras* (type-species *P. mixtecum*); the genus was then recorded in northeastern Italy (Sturani, 1971) and Sicily (Wendt, 1971); *P. schlageri* Krystyn (1972) from the Upper Bajocian of eastern Austria needs confirmation, as the holotype shows a morphology similar to the Garantianid *Hlawiceras*. The differences of *Parastrenoceras* from *Subcollina* are: 1) gently convex whorl section, 2) ventral furrow deeply interrupting the secondary ribs, 3) secondaries ending in pointed nodes, 4) undefined alternation of ribs and spines on both sides, i.e. discontinuous or no zigzag pattern on the venter (Fig. 2). Three Mexican species had been originally assigned to *Parastrenoceras* by Ochoterena (1963) who also included the European "*Ammonites Caumonti*" (d'Orbigny) and possibly "*A. lucretius*" (d'Orbigny). The generic position of *P. caumonti* was supported by Sturani (1971) and Rioult (in Fischer, 1994) despite denser ribs and small ventrolateral nodes, which could be regarded as indicative of the genus *Caumontispinctes* (Sandoval and Westermann, 1986, p. 1260); the type of ?*P. caumonti* is missing, so it is impossible to come to any conclusion.

Due to similarities with *Parastrenoceras*, the generic status of *Subcollina* did not obtain general consensus. Sturani and recently Rioult (in Fischer, 1994) suggested the possible synonymy of the two genera, but differences in the whorl section and mainly in the venter suffice to distinguish *Subcollina* from *Parastrenoceras*. Their stratigraphic distributions give further evidence of separation, the latter occurring in the Late Bajocian (Sturani, 1971; Salvador and Westermann in Westermann, 1992, p. 99), while *Subcollina* first occurs in beds below typical assemblages of *Parastrenoceras* (Sandoval and Westermann, 1986), i.e., topmost Lower Bajocian (Sandoval and Westermann, 1986; and *S. ochoterenai* n. sp.).

Occurrence. – Representatives of the genus *Subcollina* are rare but widely distributed. Besides *S. ochoterenai* n. sp. and *S. sandovali* n. sp. pointing to latest Early Bajocian, *S. lucretii* is recorded from early Late Bajocian of Mediterranean (Sturani, 1971) and Submediterranean areas (d'Orbigny, 1850). Just a bit younger are the specimens described by Fernández López (1985, p. 413, as *S.* cf. *lucretii*) and Sturani (1971, p. 127, as ?*Bajocia* n. sp.). Hence, *Subcollina* is recorded through the whole Mediterranean Province to the western opening of the Hispanic Corridor (Mexico).

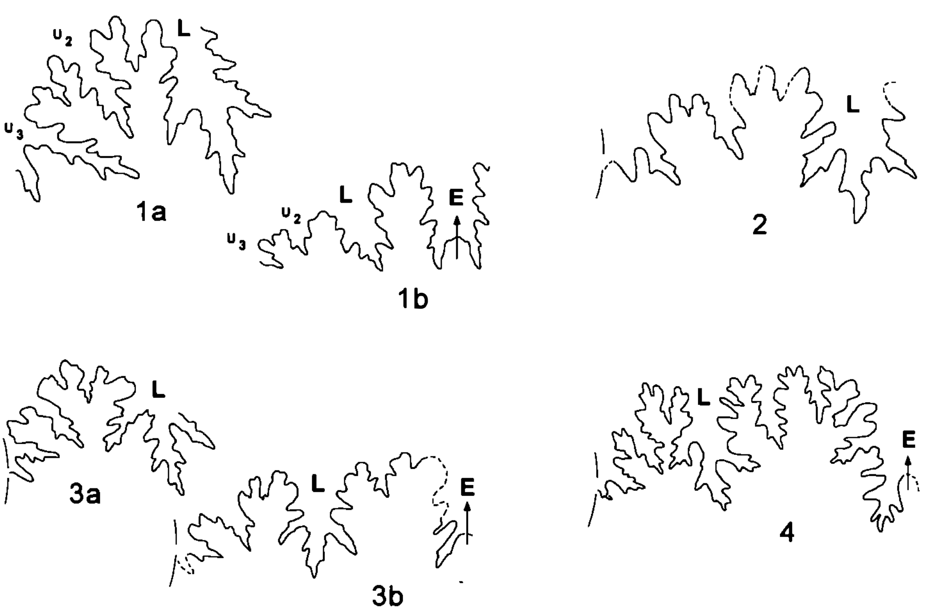

Figure 3. Septal sutures. *1a-b*, ***Phaulostephanus diniensis*** Pavia (1983, Fig. 32) at D = 24 (*a*) and 17 (*b*)mm. *2*, ***Phaulostephanus burckhardti*** Sandoval and Westermann (1986, Fig. 29) at D = 30mm. *3a-b*, ***Subcollina sandovali*** n. sp. (Sandoval and Westermann, 1986, Fig. 31.2) at D = 35mm. *4*, ***Parastrenoceras zapotecum*** ochoterena (Sandoval and Westermann, 1986, Fig. 33) at D = 21mm.

SUBCOLLINA SANDOVALI n. sp.
Fig. 3

Subcollina lucretia (d'Orbigny). Sandoval and Westermann, p. 1258, Figs. 30.5-8, 31.1-2, 32.1-6.

Derivation of name. – In honour of José Sandoval of the Granada Department of Geology.

Holotype. –IGM-3936a illustrated by Sandoval and Westermann (1986) in Figs. 32.1-2: internal mould with body chamber of 40mm diameter, possibly microconch.

Type locality. – Locality Mx3, Barranco los Rebajes, Oaxaca State, Mexico.

Stratotype. – Level 1 of los Rebajes section, Taberna Formation, *Duashnoceras floresi* Biozone, middle part of the Bajocian.

Diagnosis. – Subquadratic whorl section; arched venter; prorsiradiate primary ribs; bifurcation with pointed nodes at 5/6 to 3/4 whorl height; two secondaries per primary rib; irregular zigzag path of secondaries and no median furrow on the venter.

Description. – See Sandoval and Westermann (1986) for morphology and septal suture.

Remarks. – The Mexican specimens show significant differences to *S. lucretii*. In particular, their whorls are more convex and the venter is arched; the division point of primaries with nodes lies on the sides, so that there is no ventrolateral edge, except in Sandoval and Westermann's figures 30.8 and 32.6, while in *S. lucretii* the venter is almost flat (see Sturani, 1971). The secondary ribs of *S. sandovali* n. sp. do not show any regular zigzag pattern but some of them cross the venter and are bounded by opposing spines.

The septal suture of *S. sandovali* n. sp. is similar to that of *Parastrenoceras* (cf. Fig. 3.3-4): the trifid L and retracted umbilical lobe are characteristic. Correspondences are also strong with the suture of *Phaulostephanus diniensis* which, however, differs in the slender saddle E/L (Fig 3.1).

The paratypes of *S. sandovali* n. sp. with low division-point of ribs (Sandoval and Westermann, 1986, Figs. 30.7, 32.3-4) show morphological analogies with *Phaulostephanus burckhardti*, in particular with the specimen of Fig. 30.4. Serpenticone coiling and measurements are the same, except for wider whorls in *Phaulostephanus*; septal sutures are also similar (Fig. 3.2-3). This evidence indicates a phylogenetic link of *P. burckhardti* with *S. sandovali* in the Mexican mid-Bajocian.

Occurrence. – *S. sandovali* is presently known only from the Bajocian of southwestern Mexico.

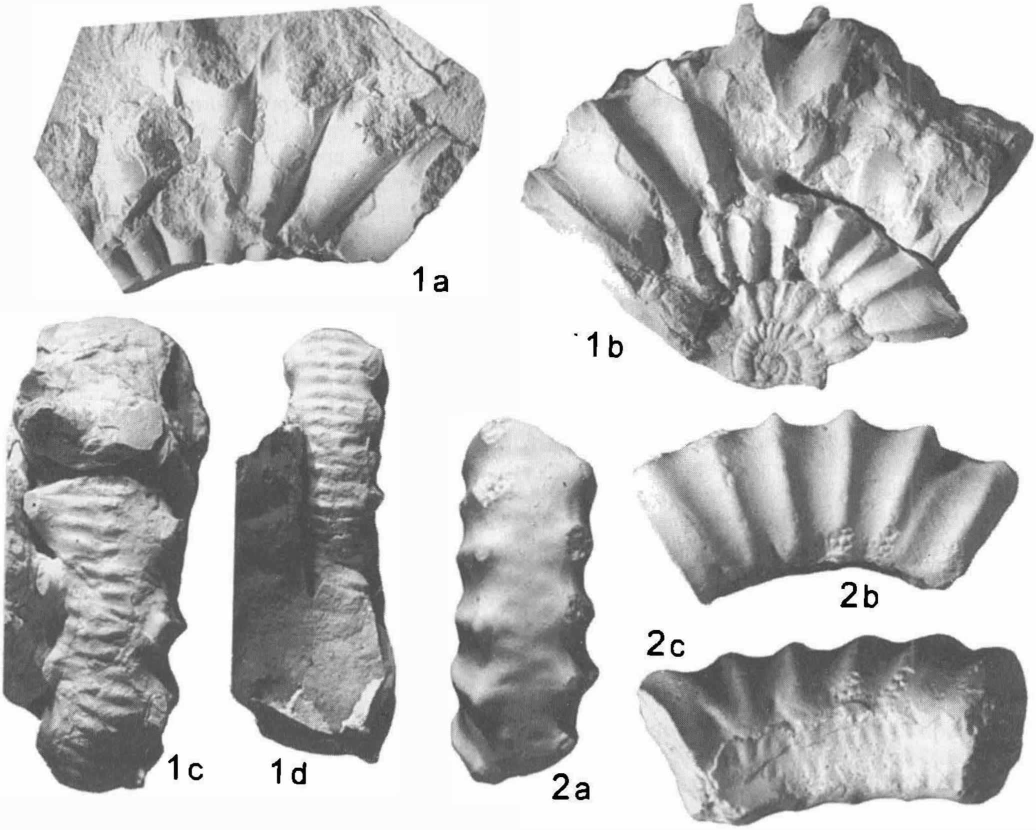

Figure 4. *Subcollina ochoterenai* n. sp. All figures x 1.2.

1a-d: holotype, MGT-BD.17. Les Dourbes section, SE France. *S. humphriesianum* Zone, *T. blagdeni* Subzone, Lower Bajocian. *a*, external cast showing the spines. *b*, lateral view. *c*, ventral view with spines alternating on the two sides. *d*, ventral view of the penultimate whorl with feeble median furrow.

2a-c: gypsum mould of the specimen of Bentz (1928). Oolite de Bayeux Fm., St. Vigor, NW France. *a*, ventral view. *b*, lateral view. *c*, umbilical edge with scar of ventrolateral spines.

SUBCOLLINA OCHOTERENAI n. sp.

Figs. 4.1-2

Strenoceras ? (n.subg.) *lucretius* d'Orbigny. Bentz, 1928, p. 172, pl. 15, Fig. 2.
Parastrenoceras lucretius (d'Orbigny). Ochoterena, 1963, p. 24, pl. 5, Fig.- 3.

Derivation of name. – In honour of Hector Ochoterena, the Mexican paleontologist who first worked in detail on these ammonites.

Holotype. – The fragmentary composite mould illustrated in Fig. 4.1; Geological Museum of the Torino University (MGT-BD.17).

Type locality. – The holotype was collected by M. Kapitzke in 1983 during a joint excursion to the Bajocian succession of Feston river, near Digne (SE France). It was found loose in the middle part of the "les Dourbes" section (Fig. 1; Pavia, 1983).

Stratotype. – From its position on the outcrop and preservation, the holotype has to have come from beds 95.3 of that section, i.e. the upper part of the *Teloceras blagdeni* Subzone (European *Stephanoceras humphriesianum* Standard Zone), topmost Lower Bajocian.

Diagnosis. – Subquadratic whorl section with gently arched sides and flat venter; rectiradiate ribs; well spaced primaries; large and laminar ventrolateral spines alternating on the two sides of outer whorls; three secondaries per primary, slightly prominent by the side of a weak ventral furrow.

Description. – The holotype is represented by 2/5 of a specimen about 60mm in diameter. The preservation as composite mould makes the ornamentation correspond to that of the shell (Pavia, 1983, p. 40-42), whereas the "simple" internal moulds of other specimens of *Subcollina* (cf. Fig. 4.2) only show pointed nodes. The dissolution of skeletal walls prevented the preservation of the septal suture on the last two preserved whorls, so that it is impossible to state whether the holotype is a macroconch or microconch. Diagenetic compaction deformed the fossil and measurements can be only approximated (values as percentages of diameter):

D = 55mm; (whorl height) h = 27; (umbilicus) u = 55; (primaries per whorl) P = ca 20.

The shell is planulate to serpenticone with subquadratic outer whorl section, which is weakly convex but subquadrangular including the ornament. The umbilical wall is rounded and the sides converge to the ventrolateral margin; the venter is flattened on the last two preserved whorls. Coiling is evolute with very feeble whorl overlap; the umbilicus slightly decreases with growth. Prorsiradiate and dense, but prominent, primary ribs are present up to the third or fourth whorl (D = 7mm); on the following whorls the primaries become more widely spaced, rectiradiate and prominent. They are separated by interrib spaces which progressively increase, but the number of primary ribs per whorl remains constant. Primaries become thin and slightly sigmoid on the outer flank. At the ventrolateral edge one spine per primary develops; the primary strings fade on the adoral wall of spines. Spines are high and sharp (Fig. 4.1a), triangular but elongated in shape. On the inner whorls, up to a diameter of 40mm, spines of one side appear to correspond to those of the opposite side (?diagenetic compaction: Fig. 4.1d) while at a diameter of 50mm the spines alternate (Fig. 4.1c). From each spine three secondary ribs arise; because of lateral alternation of primary elements, they form a zigzag pattern on the venter. Secondaries are slightly more prominent beside the feeble furrow that reduces the prominence of ribs at mid-venter.

Remarks. – The fragmentary body chamber described by Bentz (1928) is clearly conspecific. The specimen (Fig. 4.2) came from the base of the Oolite de Bayeux Fm. (Wetzel, 1924, p. 214) which contains condensed ammonite assemblages of the middle and upper parts of the Bajocian (Pavia, 1994). Bentz's fossil shares the same morphological characters as the holotype of *S. ochoterenai* n. sp., as far as coiling, whorl section, primary and secondary ribbing, and ventral furrow are concerned. Ventrolateral spines did exist (Fig. 4.2c) but are reduced to pointed nodes on the internal mould. The sole difference consists of more marked opposition of the ventral spines (Fig. 4.2a) and the clearer zigzag path of secondaries across the venter.

The taxonomic position of Bentz's specimen has been debated many times. It had been referred to *S. lucretii* by Bentz (1928), Ochoterena (1963: even as the holotype of d'Orbigny's species, see general discussion) and Sandoval and Westermann (1986), while Sturani (1971, p. 163) considered it as a possible *Parastrenoceras caumonti*. But the latter species bears slender and denser primaries, reduced and not alternating ventrolateral nodes, and a clearly defined ventral furrow (Rioult *in* Fischer, 1994, pl. 44, Fig. 6). On the other hand, *S. lucretii* has a quadrangular whorl section, denser primaries, and fewer secondaries which cross the venter in characteristic zigzag pattern without significant weakening.

The number of secondaries and the weak ventral furrow make S. *ochoterenai* n. sp. similar to *S. yeovilensis* spath. However, the holotype of the latter species (Sandoval and Westermann, 1986, Fig. 30.1-3) has higher whorl section and denser primary ribbing which results in reduced spines-nodes.

I consider the high number of secondary ribs and the presence of the ventral furrow sufficient to place *ochoterenai* in the genus *Parastrenoceras*. While *Parastrenoceras* bears nodes beside the well marked ventral furrow, its species do not show a clear alternation of the ventrolateral elements, i.e., the zigzag pattern is irregular or absent. Nevertheless, the resemblance to *Parastrenoceras* is so close that *S. ochoterenai* indicates its evolutionary transition with some still unidentified *Parastrenoceras* of the European Late Bajocian. In this respect, *P.* aff. *caumonti* of Sturani (1971, p. 162) assumes prime importance as the possible next phylogenetic step in the derivation of *Parastrenoceras* from *S. ochoterenai*.

Occurrence. –S. ochoterenai is known so far from the holotype which comes from the topmost Lower Bajocian of the Digne succession (SE France). An additional specimen (Bentz, 1928) is from the condensed Bajocian of Normandy (NW France).

Discussion

The phyletic evolution of *S. sandovali* n. sp., i.e. of the earliest known *Subcollina*, from *Phaulostephanus burckhardti* in the upper Lower Bajocian of southwestern Mexico has been discussed in the systematic section. Owing to clear homologies it is not difficult to pursue such a phylogenetic trend, by allopatric speciation, up to the European *S. lucretii* (Fig. 5). In other words it is worth assuming a Mexican, i.e. East Pacific, origin of the genus *Subcollina* from *Phaulostephanus* by cladogenesis, and its further colonisation of West Mediterranean domains around the boundary with the Late Bajocian. In a different way, Sturani (1971) suggested a possible derivation of *S. lucretii* from *Bajocia farcyi*, a poorly known, mid-Bajocian taxon (Brasil, 1895, pl. 2, Fig. 11). This hypothesis was rejected by Sandoval and Westermann (1986). The differences in the ventral morphology and septal suture (narrow and deeper E and non-retracted umbilical lobe) and the biochronological gap between the two taxa militate against their phylogenetic connection. [The Late Bajocian record of *Bajocia* (Donovan et al., 1981, p. 118) refers to the specimens described by Sturani (1971, p. 127) as *Bajocia* ? n. sp., which are best classified as *Subcollina* sp. ind.].

Other European *Subcollina*, i.e., the type-species *S. yeovilensis* and *S. ochoterenai* n. sp., add significant information to the evolutionary pattern of these stephanoceratids. It is here suggested that they constitute the phylogenetic root of *Parastrenoceras*, mainly based on the morphologies of *S. ochoterenai*, that anticipate typical features of *Parastrenoceras* around the boundary with the Late Bajocian. In other words, the genus *Subcollina* can be subdivided in two groups: the first, evolving from Mexican ancestors into the Mediterranean taxa distributed through the *Strenoceras niortense* Zone; the second, giving rise to *Parastrenoceras* at the beginning of Late Bajocian (Fig. 5). Presently, it is impossible to know the link between the Eastern Pacific representatives of *Parastrenoceras* and the Mediterranean ones, because the direct superposition of *Parastrenoceras* assemblages on *Subcollina* beds in the Mexican localities (Sandoval and Westermann, 1986) prevents phylogenetic connection. There are three possibilities: 1) the phyletic evolution in the Oaxaca area is removed at a gap which does not record the transitional forms; 2) the Mexican *Parastrenoceras* constitute a homeomorphic group which evolved somewhere in the central Atlantic from related, but not conspecific, ancestors (Westermann, 1993); or, 3) the presence of *Parastrenoceras* in the Mexican successions is due to migration of taxa from European countries. The third hypothesis appears to be the most likely, but new investigations of Mediterranean material are needed to confirm it.

As to the paleobiogeographical consequence of such a situation, it seems that the "trachyostracan" *Subcollina* and *Parastrenoceras* were able to profit by tectonic events in the "Central-Atlantic Seaway" and to disperse through it. The early origin of *Subcollina* clearly indicates active migration towards the Mediterranean. Westward migration also must have occurred if we assume the development of the Mexican *Parastrenoceras* by migration from the Mediterranean Province. Effective exchanges are supported by the sea-level rise postulated in the late Early Bajocian (Hallam, 1988; Elmi, 1992), thought to be coincident with maximum faunal affinity between both ends of the seaway (Westermann, 1993).

Two-way communication through the Hispanic Corridor may also explain the Western Tethyan and Eastern Pacific distribution of *Phaulostephanus* and *Leptosphinctes* in the late Early Bajocian, the latter supposedly having derived from the former (Pavia, 1983, p. 162). In general, the plastic and simple morphology of *Phaulostephanus* indicates the more or less direct evolution of other European taxa around the Bajocian substage boundary, e.g., *Caumontisphinctes* and

Parastrenoceras, the latter having been repeatedly considered as the possible ancestor of *Strenoceras* and other Garantianinae (cf. Fig. 2 for ventral morphology).

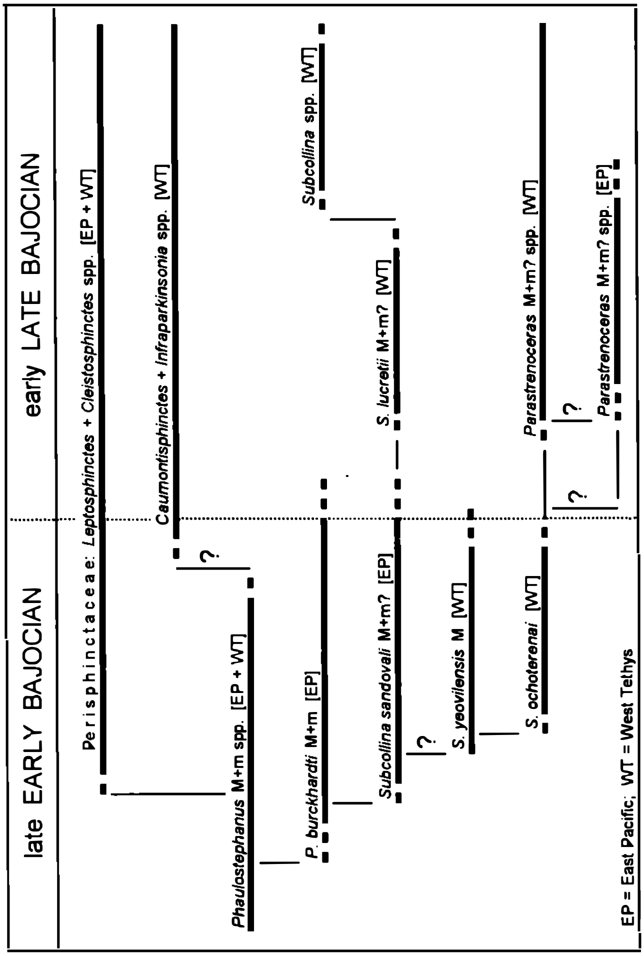

Figure 5. Phylogenetic relationships of selected Ammonitina around the Bajocian substage boundary. Note the central position of the genus *Phaulostephanus* as the possible ancestor of different Late Bajocian taxa. EP (Eastern Pacific) and WT (Western Tethys) figures indicate paleogeographical significance and highlight faunal movements through the Hispanic Corridor. M and m refer, respectively, to macro- and microconchs.

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References

BENTZ, A. 1928. Über Strenoceraten und Garantianen insbesondere aus dem Mittleren Dogger von Bielefeld. Jahrbuch Preussischen Geologischen Landesanstalt, 49:138-206.

BRASIL, L. 1895. Céphalopodes nouveaux ou peu connus des étages jurassiques de Normandie. Bullettin Société Géologique Normandie, 16:27-46.

DERCOURT, J., L. E. RICOU, AND B. VRIELYNCK (eds.). 1993. Atlas Tethys palaeoenvironmental maps. Gauthier-Villars, Paris, 307 p.

DONOVAN, D. T. 1954. Synoptic supplement to Wright's "Lias Ammonites". Palaeontographical Society, 107, 54 p.

____, J. H. CALLOMON, AND M. K. HOWARTH. 1981. Classification of the Jurassic Ammonitina, p. 101-155. *In* M. R. House and J. R. Senior (eds.), The Ammonoidea. Systematic Association, Special Volume 18, London.

ELMI, S. 1992. Les vois d'échange faunique entre l'Amerique du Sud et la Téthys alpine pendant le Jurassique inférieur et moyen. Documents Laboratoire Géologie Lyon, 125:139-149.

FERNÁNDEZ LÓPEZ, S. 1985. El Bajociense de la Cordillera Iberica. Ph.D. dissertation, Universidad Complutense Madrid, 850 p.

FISCHER, J. C. 1994. Révision critique de la Paléontologie Francaise d'Alcide d'Orbigny: Volume 1, Céphalopodes jurassiques. Masson, Paris, 340 p.

HALLAM, A. 1988. A reevaluation of Jurassic eustasy in the light of new data and the revised Exxon curve. Society Economic Paleontologists and Mineralogists, Special Publication, 42:261-273.

HAUG, E. 1891. Les Chaines Subalpines entre Gap et Digne. Bullettin Carte Géologique France, 21, 192 p.

KRYSTYN, L. 1972. Die Oberbajocium- und Bathonium-Ammoniten der Klaus-Schichten des Steinbruches Neumühle bei Wien (Österreich). Annalen Naturhistirisches Musem Wien, 76:195-310.

OCHOTERENA, H. 1963. Amonitas del Jurarisco Medio y Calloviano de Mexico: *Parastrenoceras* gen. nov. Paleontologia Mexicana, 16:1-26.

ORBIGNY, A. D'. 1850. Prodrome de Paléontologie Stratigraphique Universelle des Animaux mollusques et rayonnés: I volume. Masson, Paris, 394 p.

PAVIA, G. 1983. Ammoniti e biostratigrafia del Baiociano inferiore di Digne (Francia SE). Museo Regionale Scienze Naturali Torino, monograph 2, 254 p.

____. 1994.- Taphonomic remarks on d'Orbigny's type-Bajocian (Bayeux, west France), p. 93-111. *In* S. Cresta and G. Pavia (eds.), Proceedings III International Meeting of Aalenian-Bajocian Stratigraphy. Miscellanea Servizio Geologico Nazionale, 5, Roma.

SANDOVAL, J., AND G. E. G. WESTERMANN. 1986. The Bajocian (Jurassic) ammonite fauna of Oaxaca, Mexico. Journal of Paleontology, 60:1220-1271.

SMITH, P. L. 1983. The Pliensbachian ammonite *Dayiceras dayiceroides* and Early Jurassic paleogeography. Canadian Journal of Earth Sciences, 20:86-91.

____, AND H. W. TIPPER. 1986. Plate tectonics and paleobiogeography: Early Jurassic (Pliensbachian) endemism and diversity. Palaios, 1:399-412.

SPATH, L. F.. 1925. Notes on Yorkshire ammonites. The Naturalist, (1925):167-172.

STURANI, C. 1971. Ammonites and stratigraphy of the "Posidonia alpina" beds from Venetian Alps (Middle Jurassic, mainly Bajocian). Memorie Istituti Geologia e Mineralogia Università Padova, 27, 187 p.

WENDT, J. 1971. Genese und Fauna submariner sedimentärer Spaltenfüllungen im Mediterranen Jura. Palaeontographica, A, 136:121-192.

WESTERMANN, G. E. G. (ed.) 1992. The Jurassic of the Circum-Pacific. Cambridge University Press, 676 p.

____. 1993. Global bio-events in mid-Jurassic ammonites controlled by seaways, p. 187-226. *In* M. R. House (ed.), The Ammonoidea: environments, ecology, and evolutionary change. Systematic Association, Special Volume 47, London.

WETZEL, W. 1924. Beiträge zur Stratigraphie und Palaeogeographie des mittleren Doggers von Nordwesteuropa. Palaeontographica, 65:155-247.

WRIGHT, T. 1880-1882. Monograph on the Lias Ammonites of the British Islands: IV and V. Palaeontographical Society, 34:165-264 and 36:329-400.

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Paleoecology and Biostratigraphic Resolution: Review of Jurassic Biostratigraphic Issues in the U.S. Western Interior

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Abstract: While the Jurassic System of Europe has served as the classic example for the development of high-resolution biostratigraphic correlations and concepts, Jurassic biostratigraphy for the United States Western Interior is not well-developed and provides poor stratigraphic resolution. As has been previously pointed out by Jurassic biostratigraphers, the reasons for the unresolved nature of the North American record may be related to the limited study of these strata as compared to European ones and to the high degree of endemism reported in the North American epicontinental seaway. However, the unique paleoceanographic, ecological, biological, and environmental conditions within the Western Interior seaway may also contribute to the lack of biostratigraphic resolution in the Jurassic of western North America.

Middle and Upper Jurassic strata in the U.S. Western Interior represent terrestrial, marginal-marine and marine sedimentation spanning six major sea level changes. The marine benthic invertebrate fauna is generally characterized by low-diversities and includes many long-lived species, genera, and paleocommunities. The low-diversity, generalist nature of the organisms may be related to the rapid sea level changes, harsh environmental conditions, and restricted nature of the seaway. These environmental and ecological characteristics of the seaway may have led to the development of a fauna unlikely to yield good biostratigraphic index fossils. A review of the conditions present in the Western Interior epicontinental seaway may provide insights into the environmental and ecological factors which contributed to the conditions which lead to good—and poor—biostratigraphic resolution in the geological record.

Introduction

Due to the significant tectonic, geochemical, and biological changes which occurred during the Jurassic—such as the fragmentation of Pangaea, changes in ocean chemistry, taxonomic originations, ecological turnover, and the Mesozoic Marine Revolution—it is important to examine this System with a global perspective. However, despite the amount of study which has been conducted on European Jurassic marine strata, comparatively little research has been done on Jurassic marine strata of the United States. One of the reasons for this inattention may be the lack of stratigraphic controls present within the U.S. Western Interior units. In this paper, we present a brief overview of the stratigraphic techniques which have been employed in the U.S. Western Interior Jurassic System and the unique factors within the epicontinental seaway which may have lead to the low level of stratigraphic resolution in the region.

Stratigraphic Techniques within the Marine Jurassic of the U.S. Western Interior

There are many techniques which are used by geologists in order to establish the stratigraphy of regional depositional systems, including radiometric age-dating and sequence stratigraphy. While these techniques have been applied occasionally to the Jurassic System of the U.S. Western Interior, their success has been limited. In the marine Jurassic of the western U.S. there is a limited occurrence of ash beds suitable for numerical dating techniques. There are thin Middle Jurassic bentonite beds in Utah and parts of Idaho within marine and marginal marine facies of the Carmel, Twin Creek, and Temple Camp Formations (summarized by Christiansen et al., 1994). Some of the altered volcanic ash in these beds has been dated using K-Ar, Rb-Sr, Ar-Ar, and apatite fission track techniques (Marvin et al., 1965; Everett et al., 1989; Kowallis et al., 1993; Zhang, 1996). However. working with altered volcanic material involves careful mineralogical and geochemical work in order to eliminate diagenetic problems (Kowallis et al., 1993; Zhang et al., 1996). In addition, no preserved ash layers have yet been found in Montana and most of Wyoming or in any Upper Jurassic marine units. Thus, the use of numerical dating methodologies is limited in the Jurassic of the U.S. Western Interior with regards to geographic, paleoenvironmental, and temporal coverage.

Similarly, the use of sequence stratigraphy in this region has been limited. Although major Jurassic unconformities in the U.S. Western Interior were defined earlier by many workers (Pipiringos and O'Sullivan, 1978), the use of regional unconformities and depositional sequences within sequence stratigraphic frameworks was not well-developed until recently by Peterson (1994) and Brenner and Peterson (1994) for the southern and northern sections of the U.S. Western Interior, respectively. These two papers recognized six major regional unconformities bounding six major marine transgression-regression cycles (more unconformities were recognized in the southern area by Peterson (1994) but may not be recognized in the northern Western Interior). However. while these papers form the basis for a sequence stratigraphic framework, neither employs the terminology associated with sequence stratigraphy, nor do they address smaller sedimentary packages. Peterson (1993) has proposed that a sequence stratigraphic approach to these strata may finally resolve several stratigraphic discrepancies, and Wade and Moore (1995) have suggested that sequence stratigraphy will be very useful in correlating outcrops, cores, and well-logs currently available for many parts of the U.S. Western Interior.

Since both absolute dating and sequence stratigraphy have their limitations and have only been recently developed in the Jurassic Western Interior, biostratigraphy has been the predominant stratigraphic methodology employed in the U.S. Western Interior. Even so, biostratigraphy has not been well-refined and the time-resolution is not particularly good. Overviews of the biostratigraphy of this region have been published by Imlay (1980; 1984), Callomon (1984), Poulton et al. (1992). and Braun (1992). The lack of high-resolution biostratigraphy is attributed by Callomon (1984) to the lack of ammonite collections made in carefully recorded sections, provincialism, few correlatable stage boundaries, and poor representation of ammonites within some time intervals. In the following discussion, we will review the paleoenvironmental and paleoecological mechanisms which may have contributed to these and other factors limiting biostratigraphic and sequence stratigraphic frameworks within the U.S. Western Interior.

Paleoenvironmental and Paleoecological Characteristics of the Jurassic Western Interior Seaway

The paleoenvironmental conditions within the Jurassic North American epicontinental seaway were summarized by Imlay (1980). Generally, it was a shallow seaway with a depth of less than about 100m and there were many topographic highs. Probably due to its large fetch and shallow bottom, there is evidence of much storm activity, especially in fossiliferous units (Brenner and Davies, 1973; Specht and Brenner, 1979). It was lobate in shape, extending thousands of kilometers south from its northward opening (Fig. 1). The presence of echinoids, crinoids, and other stenohaline organisms indicates that normal marine salinities did exist even in the southernmost end of the seaway, but brackish and hypersaline conditions certainly existed in parts of the seaway at different times. The lobate shape may have caused large tidal ranges, especially in the southern part of the seaway (Tang et al., in press; Caputo and Pryor, 1991).

The seaway can be considered a mixed carbonate-siliciclastic system at different scales: regionally, there are both siliciclastic and carbonate units, while in single beds the sediments are often a mixture of siliciclastic grains and carbonate allochems and cements. The mixing of carbonate and siliciclastic elements appears to have resulted from a number of factors, such as those outlined by Mount (1984) (i.e., punctuated events, *in situ* production, and transport of siliciclastics into carbonate-dominated environments). The sources and amount of siliciclastic input can vary greatly both laterally and temporally depending on topography and regional tectonic activity (Blakey, 1994; Brenner and Peterson, 1994; Peterson, 1994).

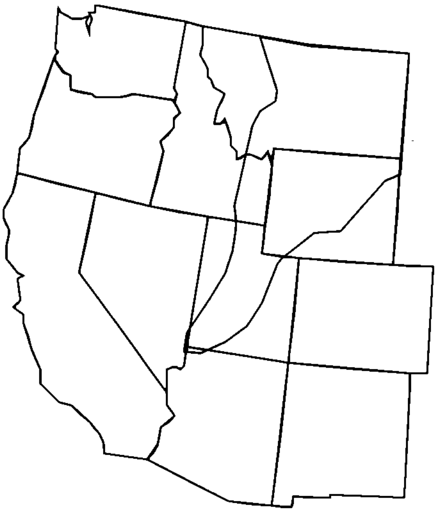

Figure 1. Reconstruction of U.S. Western Interior Jurassic seaway during the Middle Jurassic (modified from Nielson, 1990).

This set of paleoenvironmental and sedimentological characteristics makes both sequence stratigraphic and biostratigraphic analysis difficult. The variability and mixed carbonate-siliciclastic nature of the sediments hinders the establishment of detailed high-resolution sequence stratigraphy at this time. Most sequence stratigraphy models have been developed for pure siliciclastic or pure carbonate systems and only recently have mixed carbonate-siliciclastic models been explored (Handford and Loucks, 1993). In addition, although drastic facies changes and sea level changes allow for the recognition of sequences and parasequences, topographic highs, sea level changes, and the variability of siliciclastic input through time result in a high level of both lateral and temporal facies variations which make regional correlations difficult.

The restricted nature of the seaway may have contributed to the development of a high degree of endemism within the Western Interior starting in the Bathonian (Imlay, 1965; Taylor et al. 1984; Callomon, 1984). This endemic ammonite fauna has been named the Shoshonean ammonite faunal province by Taylor et al. (1984). However, there is some preliminary indication that more detailed systematic work may provide evidence for more non-endemic species than had been previously recognized (Tang et al., in press; M. Wilson, pers. comm.)

Even within the seaway, localized populations of ammonites (Callomon, 1984), microfaunas (Peterson, 1952; Braun, 1992), and bivalves (Wright, 1973) have been documented and are sometimes related to differences in facies. These localized populations may have arisen because of great paleoenvironmental and facies variability due to shallow depths, restricted circulation, and differences in siliciclastic input related to localized uplifts. Braun (1992) suggested that sea level changes also affected the evolution of microfaunal groups by causing isolation and speciation among local populations during major regressions. Tang et al. (1994; in press) and Tang and Bottjer (1997) suggested that the shallow nature of the southernmost end of the seaway during the Middle Jurassic may have excluded large nektonic predators and facilitated the establishment of a faunal association of stalked crinoids rare in other parts of the seaway.

Unique paleoecological conditions may also have arisen due to the shallow and restricted nature of the seaway. The presence of evaporite deposits, freshwater deposits, and many storm beds suggests that paleoenvironmental conditions within the seaway may have been harsh for normal marine organisms. This interpretation of the environment is supported by faunal evidence as well. Based on low diversities of epizoans, some hard substrate faunas have been interpreted as having been deposited under non-normal salinities (Wilson and Palmer, 1994; Tang, 1997). The fauna exhibits fairly low diversities as compared to other parts of the world (Tang, 1996) although preliminary evidence does suggest that diversities may be higher than previously thought (M. Wilson, pers. comm). Low diversity exists within paleocommunities (alpha diversity), between paleocommunities (beta diversity), and regionally (gamma diversity) (Tang, 1996). Tang and Bottjer (1996) and Tang (1996) suggested that the disturbed environments within the Jurassic Western Interior seaway led to the development of a low-diversity generalist fauna.

Whatever the cause of the dominance of a low-diversity, generalist fauna within the seaway, the development of this type of fauna has important biostratigraphic implications. Eurytopic taxa have been theorized to have long species durations (Eldredge and Cracraft, 1980) and among crinoids, it has been recently documented that generalist taxa are longer-lived (Kammer et al., 1997). Possibly compounding this effect, the Jurassic Western Interior benthic fauna is strongly dominated by bivalves which have long species-durations compared to other higher groups (Stanley, 1979). These factors were cited by Tang and Bottjer (1996) as important to the development of long-term stability within the Jurassic Western Interior benthic marine macrofauna. Using a literature database of 325 benthic assemblages, Tang and Bottjer (1996) documented that some species, genera, and paleocommunities persisted through long intervals of time and through several depositional cycles. In fact, there were some bivalve taxa which persisted through the entire 20 million year span of the seaway and through six major regressions and transgressions (Tang and Bottjer, 1996; Tang, 1996). All of these factors may have resulted in the presence of long-lived taxa within the Jurassic Western Interior epicontinental seaway, which would prohibit the development of a high-resolution biostratigraphic framework.

Others studying different geologic time periods have also documented long-term intervals of stability (reviewed in Brett et al., 1996). These patterns of long-term stability of taxa and assemblages through millions of years have been called "coordinated stasis" (Brett and Baird, 1995). Although primarily documented for marine benthic macrofauna, coordinated stasis has also been cited for marine planktonic foraminifera (Lu and Keller, 1995) and terrestrial ecosystems (DiMichele, 1994). The pattern of coordinated stasis is one where species and assemblages remain stable for millions of years before disappearing in a very short interval. During stable intervals, not only does paleocommunity structure remain the same, but species rarely become extinct, new species rarely appear, and morphological characters show no directional change (reviewed in Brett et al., 1996). Although it is unclear how common this pattern of coordinated stasis is and what factors would cause it (Morris et al., 1995; Tang and Bottjer, 1996; Ivany and Schopf, 1996), it is clear that coordinated stasis (or any other pattern of long-term stability) could affect biostratigraphic analyses. For example, within these intervals of long-term stability—which typically persist through several millions of years—it would be difficult to find index taxa capable of subdividing these units. Thus, the factors which control paleoecological and paleobiological stability also affect the level of biostratigraphic resolution possible in different environmental and depositional systems.

Conclusion

A number of tectonic, geographic, bathymetric, eustatic, and ecological factors has contributed to the lack of stratigraphic resolution within the Jurassic of the U.S. Western Interior. An increase in detailed systematic research could improve stratigraphic resolution by identifying new taxa within this region and by splitting species which have previously been lumped together. In addition, there is the possibility that detailed study will reveal a larger number of non-endemic taxa than previously identified. There is already evidence for the identification of non-endemic species within the Carmel Formation of southern Utah based on limited systematic work (Tang et al., in press; M. Wilson, pers. comm). However, even with improved systematics, there may be environmental and ecological characteristics of the epicontinental seaway such as those discussed above which limit the level of biostratigraphic precision attainable. Thus, sequence stratigraphy should be pursued as a possible way to improve regional correlation within the Jurassic Western Interior despite difficulties in correlating facies changes across the basin.

Clearly, the lack of stratigraphic resolution within the Jurassic U.S. Western Interior limits the types of studies which can be conducted. Any study which attempts to look at time slices of less than several million years will encounter difficulties in making regional correlations. Consideration of the low-diversity fauna, patchy distribution of fossils, and poor stratigraphic resolution makes paleontological study of the marine fauna of the Western United States a daunting task. However, given the importance of the Jurassic in global biological and paleoceanographic changes, and the lack of understanding of the epicontinental seaway of North America, there are still compelling reasons to face the challenges of studying the Jurassic System of the Western Interior United States.

References

BLAKEY, R. C. 1994. Paleogeographic and tectonic controls on some Lower and Middle Jurassic erg deposits, Colorado Plateau, p. 273-298. *In* M. V. Caputo, J. A. Peterson and K. J. Franczyk (eds.), Mesozoic Systems of the Rocky Mountain Region, USA. Rocky Mountain Section-SEPM, Denver, Colorado.

BRAUN, W. K. 1992. Ostracods and foraminifers of Western Interior North America, p. 317-321. *In* G. E. G. Westermann (ed.), The Jurassic of the Circum-Pacific. Cambridge University Press, Cambridge.

BRENNER, R. L., AND D. K. DAVIES. 1973. Storm-generated coquinoid sandstone: genesis of high-energy marine sediments from the Upper Jurassic of Wyoming and Montana. GSA Bulletin 84:1685-1698.

____, AND J. A. PETERSON. 1994. Jurassic sedimentary history of the northern portion of the western interior seaway, USA, p. 217-232. *In* M. V. Caputo, J. A. Peterson and K. J. Franczyk (eds.), Mesozoic Systems of the Rocky Mountain Region, USA. Rocky Mountain Section-SEPM, Denver, Colorado.

BRETT, C. E. AND G. C. BAIRD. 1995. Coordinated stasis and evolutionary ecology of Silurian to Middle Devonian faunas in the Appalachian Basin, p. 285-315. *In* D. H. Erwin and R. K. Anstey (eds.), New approaches to speciation in the fossil record. Columbia University Press, New York.

____, L. C. IVANY, AND K. M. SCHOPF. 1996. Coordinated stasis: an overview. Palaeogeograpy, Palaeoclimatology, Palaeoecology 127:1-20.

CALLOMON, J. H. 1984. A review of the biostratigraphy of the post-Lower Bajocian Jurassic ammonites of western and northern North America, p. 143-174. *In* G. E. G. Westermann (ed.), Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper 27.

CAPUTO, M. V. AND W. A. PRYOR. 1991. Middle Jurassic tide- and wave-influenced coastal facies and paleogeography, Upper San Rafael Group, east-central Utah, p. 9-27. *In* T. C. Chidsey, Jr. (ed.), Geology of east-central Utah. Utah Geological Association;

CHRISTIANSEN, E. H., B. J. KOWALLIS AND M. D. BARTON. 1994. Temporal and spatial distribution of volcanic ash in Mesozoic sedimentary rocks of the Western Interior: an alternative record of Mesozoic magmatism, p. 73-94. *In* M. V. Caputo, J. A. Peterson and K. J. Franczyk (eds.), Mesozoic Systems of the Rocky Mountain Region, USA. Rocky Mountain Section-SEPM, Denver, Colorado.

DIMICHELE, W. A. 1994. Ecological patterns in time and space. Paleobiology 20:89-92.

ELDREDGE, N. AND J. CRACRAFT. 1980. Phylogenetic patterns and the evolutionary process. Columbia University Press, New York, 349 p.

EVERETT, B. H., B. J., KOWALLIS, E. H., CHRISTIANSEN, AND A. DEINO. 1989. Correlation of Jurassic sediments of the Carmel and Twin Creek Formations of southern Utah using bentonite characteristics. Utah Geological and Mineralogical Survey Open-File Report 169: 1-60.

HANDFORD, C. R. AND R. G. LOUCKS. 1993. Carbonate depositional sequences and systems tracts—responses of carbonate platforms to relative sea-level changes, p. 3-41. *In* R. G. Loucks and J. F. Sarg (eds.), Carbonate sequence stratigraphy: Recent developments and applications. AAPG, Tulsa, Oklahoma.

IMLAY, R. W. 1965. Jurassic marine faunal differentiation in North America. Journal of Paleontology 39:1023-1038.

____. 1980. Jurassic paleobiogeography of the conterminous United States in its continental setting. U.S. Geological Survey Professional Paper 1062:1-121.

____. 1984. Jurassic ammonite successions in North America and biogeographic implications, p. 1-12. *In* G. E. G. Westermann (ed.), Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper 27.

IVANY, L. C., AND K. M. SCHOPF (eds.). 1996. New Perspectives on faunal stability in the fossil record special issue. Palaeogeography, Palaeoclimatology, and Palaeoecology 127:1-361.

KAMMER, T. W., T. K. BAUMILLER, AND W. I. AUSICH. 1997. Species longevity as a function of niche breadth; evidence from fossil crinoids. Geology 25:219-222.

KOWALLIS, B. J., E. H. CHRISTIANSEN, B. H. EVERETT, K. D. CROWLEY, C. W. NAESER, D. S. MILLER, AND A. L. DEINO. 1993. Possible secondary apatite fission track age standard from altered volcanic ash beds in the Middle Jurassic Carmel Formation, southwestern Utah. Nuclear Tracks Radiation Measurement 21:519-524.

LU, G. AND G. KELLER. 1995. Ecological stasis and saltation: species richness change in planktic foraminifera during the late Paleocene to early Eocene, DSDP Site 577. Palaeogeography. Palaeoclimatology, Palaeoecology 117:211-227.

MARVIN, R. F., J. C. WRIGHT, AND F. G., WALTHALL. 1965. K-Ar and Rb-Sr ages of biotite from the Middle Jurassic part of the Carmel Formation, Utah. U.S. Geological Survey Professional Paper 525B:104-107.

MORRIS, P. J., L. C. IVANY, K. M. SCHOPF AND C. E. BRETT. 1995. The Challenge of ecological stasis: Reassessing sources of evolutionary stability. Proceedings of the National Academy of Sciences 92:11269-11273.

MOUNT, J. 1984. Mixing of siliciclastic and carbonate sediments in shallow shelf environments. Geology 12:432-435.

NIELSON, D. R. 1990. Stratigraphy and sedimentology of the Middle Jurassic Carmel Formation in the Gunlock Area, Washington County, Utah. Brigham Young University Geology Studies 36:153-191.

PETERSON, F. 1994. Sand dunes, sabkhas, streams, and shallow seas: Jurassic paleogeography in the southern part of the western interior basin, p. 233-272. *In* M. V. Caputo, J. A. Peterson and K. J. Franczyk (eds.), Mesozoic Systems of the Rocky Mountain Region, USA. Rocky Mountain Section-SEPM, Denver, Colorado.

PETERSON, J. A. 1952. Stratigraphy and micropaleontology of the Sundance Group, eastern Wyoming. Unpublished Ph.D. dissertation, University of Minnesota, 148 p.

____. 1993. Jurassic sequences of the Western Interior United States regional and global associations. Pangaea Conference Program and Abstracts, Canadian Society of Petroleum Geologists, p. 246.

PIPIRINGOS, G. N. AND R. B. O'SULLIVAN. 1978. Principal unconformities in Triassic and Jurassic rocks, Western Interior, U.S.: a preliminary survey. U.S. Geological Survey Professional Paper 1035A:1-29.

POULTON, T. P., R. L. DETTERMAN, R. L. HALL, D. L. JONES, J. A. PETERSON, P. SMITH, D. G. TAYLOR, H. W. TIPPER, AND G. E. G. WESTERMANN. 1992. Western Canada and United States, p. 29-92. *In* G. E. G. Westermann (ed.), The Jurassic of the Circum-Pacific. Cambridge University Press, Cambridge.

SPECHT, R. W. AND R. L. BRENNER. 1979. Storm-wave genesis of bioclastic carbonates in upper Jurassic epicontinental mudstones, east-central Wyoming. Journal of Sedimentary Petrology 49:1307-1322.

STANLEY, S. M. 1979. Macroevolution: pattern and process. W.H. Freeman and Co., San Francisco, 332.

TANG, C.M. 1996. Evolutionary paleoecology of Jurassic marine invertebrate assemblages of the western interior, U.S.A. Unpublished Ph.D. dissertation, University of Southern California, 235 p.

____. 1997. Evolutionary paleoecology of Jurassic hard-substrate assemblages from the U.S. Western Interior. Geological Society of America Annual Meeting Abstracts with Program 29:A107.

____, AND D. J. BOTTJER. 1996. Long-term faunal stasis without evolutionary coordination: Jurassic benthic marine paleocommunities, western interior, U.S.A. Geology 24:815-818.

____, AND ____. 1997. Low-diversity faunas of the Middle Jurassic Carmel Formation and their paleobiological implications, Brigham Young University Geology Studies 2(2):10-14.

____, ____, AND M. J. SIMMS. 1994. Paleoecology of a Middle Jurassic shallow-water encrinite (Carmel Formation, southern Utah): Implications for the retreat of stalked crinoids to the deep-sea. Geological Society of America Abstracts with Programs 26: A428.

____, ____, AND ____. In press. Stalked crinoids from a Jurassic tidal deposit in western North America. Lethaia.

TAYLOR, D. G., J. H. CALLOMON, R. HALL, P. L. SMITH, H. W. TIPPER AND G. E. G. WESTERMANN. 1984. Jurassic ammonite biogeography of western North America: the tectonic implications, p. 121-141. *In* G. E. G. Westermann (ed.), Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper 27.

WADE, W. J. AND C. H. MOORE. 1995. New correlations in the Wyoming Jurassic: shelf sequence boundaries and mixed carbonate-clastic lowstand wedges. Geological Society of America Annual Meeting Program with Abstracts 27:A83.

WILSON, M. A. AND T. J. PALMER. 1994. A carbonate hardground in the Carmel Formation (Middle Jurassic, SW Utah, USA) and its associated encrusters, borers and nestlers. Ichnos 3:79-87.

WRIGHT, R. P. 1973. Marine Jurassic of Wyoming and South Dakota: Its paleoenvironments and paleobiogeography. University of Michigan Museum of Paleontology Papers on Paleontology 2:1-49.

ZHANG, C., E. H. CHRISTIANSEN, B. J. KOWALLIS, AND A. L. DEINO. 1996. Volcanic ashes in the Middle Jurassic of southern Utah. Geological Society of America Abstracts with Programs 28: 503.

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Siliceous Sedimentation in the Mediterranean Jurassic Caused by Volcanism, Greenhouse Climate and Eutrophication

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Abstract: During the Middle Jurassic in the Mediterranean domain, sedimentation was characterized by the extreme reduction, or a total demise of carbonate production, both in neritic and in pelagic environments, and, in the latter, by the shift to siliceous sedimentation.

Speculation on the causes, and studies in the Umbria-Marche Jurassic basin, lead to the conclusion that this dramatic change had its origin in lithospheric rearrangement and in paleoclimate changes occurring during this time.

The Middle Jurassic represented an unusually fast spreading period in the Pacific leading to high CO_2 release mainly from the circum-Pacific subduction zones. Large aerial volcanism during the Late Aalenian-Early Bajocian has been documented in the Antarctic, Patagonia and Karroo. The release of CO_2 into the atmosphere favored an intensification of greenhouse climate as demonstrated by marked positive shifts in the $\delta\ ^{13}C$ curve. Warm and humid climates intensified water cycling, increased weathering and run-off from the continents and, consequently, the input of nutrients into the sea. Moreover, increased levels of atmospheric pCO_2 may have had a major impact on changing seawater chemistry (higher concentrations of dissolved carbon dioxide), that became less favorable for carbonate producing organisms, and probably aggressive to carbonate shells.

During the Middle Jurassic in the Alpine Tethys and Central Atlantic, important rearrangement of paleogeography occurred with the formation of new spreading ridges which reorganized oceanic currents. The hydrothermal activity along the ridges may also locally promote further release of carbon dioxide and silica into sea-water. In Mediterranean Tethys, the rearrangement of currents allowed easy circulation of nutrients, silica, and aggressive water, resulting in an homogenization of the seawater.

Global and local causes coincided in the Mediterranean Tethys during the Middle Jurassic creating an environment favorable to the productivity and preservation of biosiliceous sedimentation. A partial eutrophication (caused by intensified greenhouse conditions) affected both basins and platforms. Eutrophic conditions caused a crisis in the carbonate producers and at the same time boosted the radiolarian productivity. Furthermore the silica released into seawater favored the development of radiolarians as well as the preservation of the siliceous skeletons, so that siliceous production overcame carbonate production.

Introduction

The Middle Jurassic is characterized by diffused siliceous sedimentation in much of the Tethyan pelagic realm. Siliceous sediments are of two main types: A) radiolarites deposited on newborn oceanic crust above ophiolites in the internal zones, B) cherty limestones or cherts deposited above calcareous sediments resting on continental crust in the external regions. For the first type the environmental situation that caused the siliceous deposition is fairly straightforward (Folk and McBride, 1978) but for the second type, the sedimentary environment and the causes of the transition from carbonate to siliceous deposition have been debated with different and even opposite interpretations. Up to now no clear and unequivocal conclusions have been reached.

Two main models have been proposed: a deep-water model and a shallow water one. For the first model the shift from carbonate to siliceous sedimentation was caused by a rapid deepening of the basin (some thousands of meters) that carried the sea bottom below the CCD. All the carbonate produced in the sea water was dissolved and siliceous sedimentation became dominant (Garrison and Fischer, 1969; Bosellini and Winterer, 1975; Winterer and Bosellini, 1981; Centamore et al., 1971; Bernoulli et al., 1979). This model can explain the siliceous sedimentation but conflicts with the following evidence from field studies of the Calcari Diasprigni unit of the Umbro Marchigiano basin (Central Italy):

1) Rare hermatypic corals present on structural highs close to where cherty beds wedge out (Nicosia and Pallini, 1977; Santantonio, 1993; Santantonio et al., 1996).
2) Calcareous shells (foraminifera and pelecypods) silicified as flint during a diagenetic phase, therefore not dissolved in the sedimentary environment.
3) In the Calcari Diasprigni Formation, nannofossils are generally rare but they are continuously present and become abundant and well preserved in discrete layers, demonstrating that the sedimentary and early burial environment were not particularly aggressive (Mattioli, 1995).
4) Paleomorphological and paleoenvironmental analysis do not agree with abyssal depth, as small structural highs scattered throughout the basins remained in shallow water and do not show evidence of great subsidence (Nicosia and Pallini, 1977; Cecca et al., 1987; Bice and Stewart 1990; Santantonio, 1993; 1994).
5) Formations that precede and follow the cherty beds do not fit with an abyssal environment (Colacicchi et al., 1970; Farinacci et al., 1981; Colacicchi et al., 1988).

The second model proposes sedimentation in shallow water at depths between zero and few hundreds of meters (Colacicchi et al., 1970; 1988; Farinacci et al., 1981; Bice and Stewart, 1990; Santantonio, 1993). Different explanations for the facies change are invoked: A) a change in the sea water chemistry that became unsuitable for carbonates; B) a general cooling of sea water that depressed the carbonate production; C) a period of hyperproduction on the carbonate platforms that sequestrated the carbonate so that the basins were deprived of carbonate. D) Folk (*in* McBride and Folk, 1979) proposed a tide controlled environment as an extreme interpretation; some fractures were considered as shrinkage structures and some quartz and lutecite cement as replacement of sulfates or halite pseudomorphs. This tidal control interpretation is not generally accepted and even McBride, Folks' coauthor, was of a totally different opinion.

In order to address the problem of the siliceous sedimentation, some Middle-Upper Jurassic stratigraphic sections (containing the Calcari Diasprigni Formation) have been studied in the Umbria Marche basin, correlating data coming from different fields: A) stable isotopes of carbon and oxygen; B) abundance and preservation of calcareous nannofossils; C) abundance and preservation of radiolarians; D) development of the adjacent carbonate platform; E) general geodynamic situation. The correlation gives some indications for the causes and the environment of the siliceous sedimentation and gives rise to a new genetic model that is independent of the depth of the basin.

Data and Results

Useful data for environmental studies of flinty chert deposits are difficult to collect. Stratigraphic sections were sought in the Calcari Diasprigni Formation where sedimentation is represented by interbedded calcareous and cherty levels. In the south-eastern corner of the Umbria Marche basin, two sequences (Terminilletto and Colle Bertone) permitted collection of carbon and oxygen stable isotopes and documentation of chert abundance and nannofossil biostratigraphy (Bartolini et al., 1995; 1996). Chronostratigraphic control based on ammonites was also possible. For the same period, sedimentation on the adjacent Latium-Abruzzi carbonate platform can be assessed because development of the platform directly influenced the basinal sedimentation. The following data were collected:

1) The $\delta^{13}C$ curve shows several positive peaks and a direct correlation with chert occurrence (Fig. 1). Increases in the $\delta^{13}C$ correspond to chert increases and, conversely, any decrease of the curve, to a diminution of cherts. Both sections revealed a relative minimum of $\delta^{13}C$ close to 2‰ in the Aalenian. In the Lower Bajocian, the values of the $\delta^{13}C$ increase to a maximum around 3.5‰. This positive excursion may be correlated with analogous and coeval positive excursions found in the Digne region (Corbin, 1994; Bartolini et al., 1996) and in the Subbetic Realm (Bartolini et al., 1997). From Upper Bajocian through Bathonian in the Terminilletto section, the values decrease around 2.3‰. In the Upper Bathonian-Lower Callovian there is another positive

anomaly around 3‰; a similar shift has been found in the Subbetic Realm (Bartolini et al., in prep.). A further positive anomaly of 2.9‰ occurs in the Middle Oxfordian; this one is correlatable with similar positive peaks found in the Southern Alps and southern France (Jenkyns, 1996), in northern and central Poland (Gruszczynski et al., 1990; Hoffman et al., 1991), in eastern Spain (Pisera et al., 1992), Swiss Jura (Bill et al., 1995) and in the Helvetic nappes of eastern Switzerland (Weissert and Mohr, 1995). In the Kimmeridgian-Lower Tithonian of Terminilletto section the values decrease to a minimum of 1.5‰. This variation is analogous to the diminution reported by Weissert and Channel (1989) in the Southern Alps and from Helvetic nappes of eastern Switzerland (Weissert and Mohr, 1995). Given all these analogs, the positive $\delta^{13}C$ events of the Bajocian, Callovian and Oxfordian may represent global events.

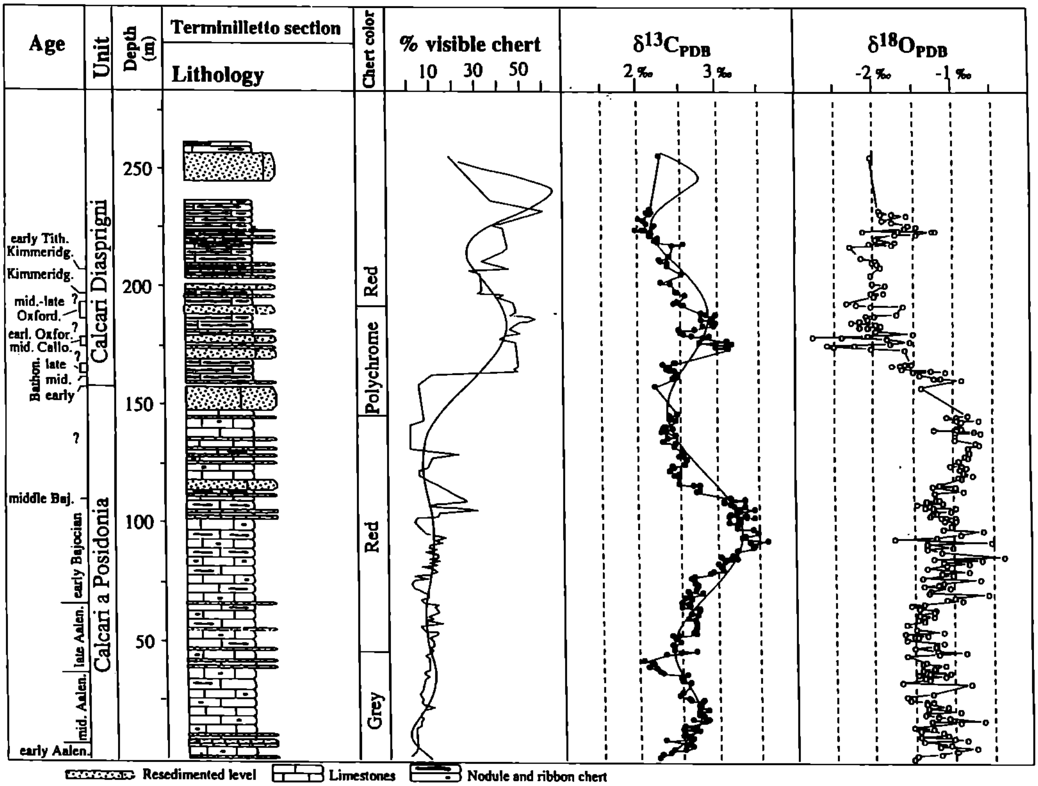

Figure 1. Terminilletto simplified section (Central Italy, Umbria Marche Basin). The low sedimentation rate from Middle Bajocian to Lower Toarcian is evident in the stratigraphic column. Visible chert percent is compared with ^{13}C and ^{18}O. Polynomial regression curves of visible chert and ^{13}C are in phase from Upper Bajocian upward. The ^{18}O curve is highly scattered but it shows clear diminution from Upper Bajocian-Lower Bathonian upward corresponding to an increase in visible chert. Modified from Bartolini (1995).

2) The values of the $\delta^{18}O$ are less accurate, and are scattered between -1‰ and -2‰, and are probably modified by diagenesis. A clear shift towards more negative values starts in the Middle Bathonian; this shift could be due to diagenesis, but it also coincides with an increase of the visible chert percent. Furthermore, the plotted relationships between $\delta^{18}O$ and visible chert abundance reveal a clear negative covariance (Bartolini et al., 1996). A similar $\delta^{18}O$ negative shift, and negative covariance with the chert have been reported and correlated in the site DSDP 367 (Brennecke, 1978) and in the Valdorbia and Fonte Avellana sections (Umbria Marche Basin)

(Haidji, 1991). So beyond any diagenetic overprint, it is reasonable to consider the decrease of the $\delta^{18}O$ as a primary oceanic signal, at least partially.

3) In the micrites of the studied sections, calcareous nannofossils are rare but always present. The abundance and preservation of nannofossils increase from the Late Oxfordian (Mattioli, 1995).
4) Sedimentation rate on the adjacent Latium-Abruzzi carbonate platform shows a dramatic decrease in the Bajocian and Lower Oxfordian, reduced to one fifth of the preceding periods. On the platform during this phase, coral and ooid facies were replaced by crinoidal, bryozoan and silicosponge facies that indicate eutrophication of the environment and a consequent strong crisis of carbonate production (Föllmi et al., 1994). Analogous events have been observed in the Dinarids and other zones of the Tethys (Colacicchi and Bigozzi, 1995). Budyko et al. (1987) outlined the Middle Jurassic as a time of global reduction in carbonate productivity. Periplatform ooze input to the basin was reduced as well; it decreased from Upper Bajocian-Lower Bathonian with a maximum reduction in the Middle Bathonian (Bartolini et al., 1996). The recovery of carbonate production began in the Middle-Late Oxfordian. Coral-chaetetid patch reefs grew in the marginal areas and established a tendency towards an aggrading and then prograding reefal rim (Colacicchi and Bigozzi, 1995). However, little platform ooze reached the basin; reefal rims seem to have retained carbonate mud on the platform. The platform situation is reflected in the proximal Terminilletto section by abundant mass flow deposits mainly containing lime clasts and reefal bioclasts. Distal basinal sedimentation rates reached the lowest values of the entire Jurassic (e.g., Valdorbia and Bosso sections, Baumgartner, 1990; 1995) since when the Middle-Late Oxfordian basinal biosiliceous sediments became gradually diluted by nannofossil ooze. Significant nannoplankton carbonate productivity is revealed in discrete levels of the Terminilletto section by abundant coccolith assemblages found in the Oxfordian and Kimmeridgian (Mattioli, 1995).
5) From a geodynamic point of view, fast spreading in the Pacific during the Middle Jurassic increased volcanism in circum-Pacific subduction zones (Larson and Sager, 1992; Larson et al., 1993). Large aerial volcanism during the Late Aalenian-Early Bajocian has been documented in the Antarctic, Patagonia and Karroo (Courtillot, 1995; Dromart et al., 1996). In the Tethyan region, crustal extension was in progress throughout the Early Jurassic. Crustal separation and the oblique opening of the Liguro-Piemontese Ocean basin occurred in the early Middle Jurassic and must have involved a rearrangement of the prevailing oceanic current pattern (Favre and Stampfli, 1992; Bill et al., 1997).

Discussion and Conclusions

The summarized data reveal an environmental situation more complex than previously hypothesized (Fig. 2).

A) Increase in $\delta^{13}C$ is considered an indicator of high organic productivity and/or high burial rate of organic matter; positive anomalies are often correlated with deposition of black shales, although no black shales are present in our sequences. In fact, the Middle Jurassic carbon isotope shifts seem not clearly related to the deposition of widespread and short-ranging black shales. Transient storage of organic matter in terrestrial and lacustrine environments, in conjunction with high rates of water and organic carbon cycling, may equally remove light carbon form the oceanic reservoir and result in the observed shifts. During the Early Bajocian, organic matter could have been stored over short periods of time in peats and soils formed during equitable and rather wet global climate (Sellwood et al., 1998). At Terminilletto the correlation of positive ^{13}C shifts with increased visible chert is evident. The chert in the Terminilletto section is biogenic in origin, as shown by the abundance of radiolarians and sponge spicules. More generally, the Early-Middle Bajocian, the Callovian and the Oxfordian positive $\delta^{13}C$ shifts correspond to times of onset of radiolarite sedimentation in many areas (Bartolini et al., 1996). Preservation and abundance of biosiliceous sediments are favored by high fertility of surface waters and/or by sluggish deep water circulation and density stratification with disaerobic/anaerobic conditions at the sediment-water interface (Kennett, 1982; Honjo, 1984; Caulet et al., 1992; Milliam and Takahashi, 1995). Furthermore, it is known that high CO_2 is related to greenhouse conditions. At the Terminilletto section, the absence of sedimentological and geochemical evidence for organic matter preservation tends to exclude density stratification with disaerobic/anaerobic conditions at the sediment-water interface (Bartolini et al., 1999). In conclusion, the correlation of positive $\delta^{13}C$ shifts with increased visible chert at the Terminilletto section, agrees well with an interpretation of high primary productivity of surface seawater and greenhouse conditions.

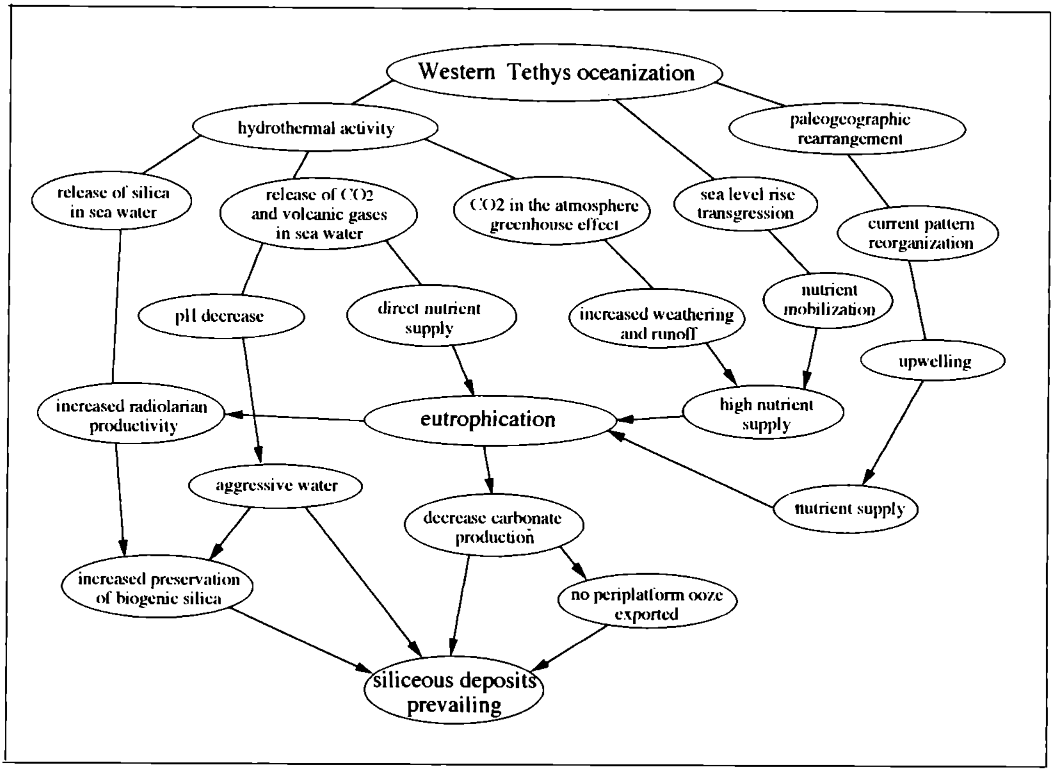

Figure 2. Synthesis of the sedimentary model for siliceous deposition in the Jurassic of Tethys. In the flux diagram the increased spreading activity is considered as the first disturbing cause. All the resulting effects drive an increase in nutrient supply and lead to eutrophication. This depresses carbonate production and increases radiolarian productivity and biogenic silica preservation.

B) The percent of ^{18}O present in a precipitated organic and/or inorganic carbonate is controlled by the isotopic composition of the seawater and, among other things, by the temperature. Thus, according Wetzel (1989), $\delta^{18}O$ may give an idea of the paleo-temperature of the seawater and/or of the burial environment. If the diminution of $\delta^{18}O$ in the Middle Bathonian is a primary oceanic signal, it may indicate an increase of the temperature of around 2 C° (Friedman and O'Neil, 1977), corroborating the greenhouse environment indicated by the $\delta^{13}C$ data (Bartolini et al., 1996).

C) The percent of visible chert is a proxy of the silica/carbonate ratio in the section, which is controlled by biological productivity and sedimentation processes (e.g., the rate of input of periplatform ooze and carbonate platform-derived sediments), as well as by dissolution and diagenesis. Chert formation is favored by the presence of layers in which biogenic silica dominates over carbonates. Dilution of biosiliceous sediment by other sediments (especially clay, but also periplatform and calcareous pelagic ooze) may inhibit chert formation (Baumgartner, 1987; 1990). It is well known that eutrophic conditions of the surface waters are detrimental to carbonate producing organisms, but favorable to biosiliceous organisms (Hallock et al., 1988).

D) During the Early and Middle Jurassic, the calcareous nannoplankton was not yet largely developed. In this situation, most of the micrite present in the basin was not produced in the water column but mainly derived by periplatform ooze produced on the adjacent carbonate platform and exported into the basin (Bombardiere, 1991). The crisis evident on the adjacent carbonate platform resulted in the lack of exported periplatform ooze and was responsible for the decrease in sedimentation rate. In summary, eutrophic seawater conditions caused the carbonate productivity crisis and the reduction of periplatform ooze input into the basin. At the same time, eutrophic seawater conditions favored high radiolarian and siliceous sponge productivity. The combined effect of these two tendencies is a change to biosiliceous sedimentation in the basins (Bartolini et al., 1996).

We may conclude that in the Tethyan region, onset of siliceous sedimentation, positive anomalies of $\delta^{13}C$, crisis in carbonate production and increased volcanism, are interconnected phenomena that suggest a different model for Middle Jurassic siliceous sedimentation.

The fast spreading and intensified volcanism in Pacific regions and elsewhere may be considered the prime controls promoting global climatic change by excess CO_2 in the atmosphere. From a more local point of view, the transition from rifting to drifting that was going on in the western Tethyan and central Atlantic regions may have its impact in sea level variations and in changing ocean current patterns. The hydrothermal activity along ridges may also locally promote further release of carbon dioxide and silica into seawater.

Increased CO2 in the atmosphere resulted in a warm and humid climate, and intensified weathering, run-off, and input of nutrients in the ocean (Weissert and Lini, 1991). Rises in $\delta^{13}C$ are linked to greenhouse conditions. Volcanic gases could directly supply nutrients to the seawater. The changing distribution of continental masses and seaways could modify the general pattern of oceanic currents, resulting in mobilization of nutrients from the sea bottom and a general diffusion of nutrients even on the carbonate platforms. The general transgressive tendency, documented by the curve of Haq et al. (1988) and Hallam (1988), caused seawater to cover and enrich continental platforms with nutrients.

The eutrophic conditions that resulted from the excess nutrients favored the bioproductivity of siliceous organisms and depressed that of calcareous organisms but other phenomena further favored siliceous sedimentation.

In basins surrounded by carbonate platforms, as with the western Tethys, periplatform ooze and carbonate turbidites shed from the platform margins diluted siliceous sedimentation (Baumgartner, 1987; 1990; Bartolini et al., 1996). The crisis of production documented on the Latium-Abruzzi and Dinaric carbonate platforms deprived the basin of this carbonate supply raising the percent of siliceous sediments.

The amount of silica released by increased continental weathering and run-off and by hydrothermal activity made this material available for the radiolarians and at the same time saturated the pore-waters, lowering the solubility of silica and increasing the preservation of the radiolarian shells.

It is known that silica is more soluble in alkaline and less soluble in acid conditions. The reverse is true for carbonates. The chemistry of Middle Jurassic water, given the excess of CO2 and volcanic gases, was reducing the pH causing the dissolution of carbonates and enhancing the preservation of silica.

The proposed model explains the causes of the siliceous sedimentation in the Mediterranean Jurassic taking into account several lines of evidence, all indicating increased production and preservation of siliceous sediments and the depression of carbonates. The model does not need to resort to extreme water depth, nor to a cooling of seawater, nor to sequestration of carbonate on the platforms.

The $\delta^{13}C$ curve of the Upper Oxfordian-Tithonian generally tends to decrease with small positive perturbations whereas percentages of visible chert remained high throughout this time up to the Early Tithonian. On the platform, the carbonate production recovered but the reef facies on the margin prevented the exportation of periplatform ooze. In the basin, micrites intercalating with biosiliceous sediments increase upward and show an abundant and well preserved nanno-flora (Mattioli, 1995). The better preservation of nannofossils in these levels may be due to the absence of periplatform-derived aragonitic muds that, if present, promote significant diagenesis resulting in neomorphism of calcite and the complete destruction of nannofossils. During the Late Oxfordian-Early Tithonian, the coexistence of healthy reefal carbonate platforms and predominantly biosiliceous sedimentation in the basins is difficult to explain in a scenario with continued high trophic levels. We believe that surface water fertility and radiolarian productivity gradually decreased and allowed reefal and nannofossil communities to develop, while radiolarian preservation on the bottom was enhanced due to the presence of silica-saturated pore waters brought up to the sediment surface by compaction.

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References

BARTOLINI, A., P. O. BAUMGARTNER, E. ERBA, J. C. HUNZIKER AND E. MATTIOLI. 1996. Plankton response to CO_2 budget change during the Middle Jurassic (Aalenian-Bajocian). SEPM Wldhouse (Switzerland) June 1996, 19.

____, ____, AND J. GUEX. 1999. Middle and Late Jurassic radiolarian paleoecology versus carbon-isotope stratigraphy. Palaeogeography, Palaeoclimatology, Palaeoecology, 145:43-60.

____, ____, AND J. HUNZIKER. 1996. Middle and Late Jurassic carbon stable-isotope stratigraphy and radiolarite sedimentation of the Umbria-Marche Basin (Central Italy). Eclogae Geologicae Helvetiae, 89 (2):811-844.

____, ____, AND E. MATTIOLI. 1995. Middle and Late Jurassic radiolarian biostratigraphy of the Colle Bertone and Terminilletto sections (Umbria-Marche-Sabina Apennines, Central Italy): an integrated stratigraphical approach, p. 817-832. *In* P. O. Baumgartner et al. (eds.), Middle Jurassic to Lower Cretaceous Radiolaria of Tethys: Occurrences, Systematics, Biochronology. Mémoires de Géologie (Lausanne), 23.

____, E. MORETTINI, L. O'DOGHERTY, J. SANDOVAL, P. O. BAUMGARTNER, AND J. C. HUNZIKER. 1997. Carixian-Bajocian carbon isotope stratigraphy of the Umbria-Marche (central Italy) and Subbetic (Spain) areas. European Union of Geosciences, Strasbourg, France, 23-27 March 1997, 403 p. *In* Abstract supplement N°1, Terra Nova, 9.

BAUMGARTNER, P. O. 1987. Age and genesis of Tethyan Jurassic radiolarites. Eclogae Geologicae Helvetiae, 80:831-879.

____. 1990. Genesis of Jurassic Tethyan radiolarites - The example of Monte Nerone (Umbria-Marche Apennines). Atti II Convegno Internazionale F.E.A. Pergola, p.19-32.

____. 1995. Towards a Mesozoic radiolarian database-Updates of the work 1984-1990, p. 689-700. *In* P. O. Baumgartner et al. (eds.), Middle Jurassic to Lower Cretaceous Radiolaria of Tethys: Occurrences, Systematics, Biochronology. Mémoires de Géologie (Lausanne), 23.

BERNOULLI, D., D. KALIN AND E. PATACCA. 1979. A sunken continental margin of the Mesozoic Tethys: the northern and central Apennine, p. 197-210. *In* Beaudoin and Porser (eds.), La sedimentation jurassique dans le contest europeenne. Ass. Sed. Francaise, Publ. Spec. 1.

BICE, D. M., AND K. G. STEWART. 1990. The formation and drowning of isolated carbonate seamounts: tectonic and ecological controls in the northern Apennine. International Association of Sedimentology Special Publication, 9:145-168.

BILL, M., P. O. BAUMGARTNER, J. C. HUNZIKER, AND Z. D. SHARP. 1995. Carbon isotope stratigraphy of the Liesberg Beds Member (Oxfordian, Swiss Jura) using echinoids and crinoids. Eclogae Geologicae Helvetiae, p.817-832

____, F. BUSSY, M. COSCA, H. MASSON AND J. C. HUNZIKER. 1997. High precision U-Pb and $^{40}Ar/^{39}Ar$ dating of an Alpine ophiolite (Gets nappe, French Alps). Eclogae Geologicae Helvetiae, 90:43-54.

BOMBARDIERE, L. 1991. Analisi sedimentologica ed ultrastrutturale del fango carbonatico della Corniola dei Massicci Perugini (Umbria occidentale) e di alcune sezioni dell' area umbro-marchigiana. Paleopelagos, 3:113-128.

BOSELLINI, A., AND E. L. WINTERER. 1975. Pelagic limestone and radiolarite of Tethyan Mesozoic: A genetic model. Geology, 3 (5):279-282.

BRENNECKE, J. C. 1978. A comparison of the stable oxygen and carbon isotope composition of the Early Cretaceous and Late Jurassic carbonates from the DSDP sites 105 and 367, p. 937-955. *In* Initial Reports of DSDP. 41.

BUDYKO, M. I., A. B. RONOV, AND A. L. YANSHIN. 1987. History of the earth's atmosphere. Springer-Verlag, Heidelberg, p.1-139.

CAULET, J. P., M. T. VÉNEC PEYRÉ, C.VERGNAUD GRAZZINI, AND C. NIGRINI. 1992. Variation of South Somalian upwelling during the last 160 Ka: radiolarian and foraminifera records in core MD85674, p. 379-389. *In* C.P. Summerhayes, et al. (eds.), Upwelling systems: Evolution Since the Early Miocene, Geological Society of London Special Publication, 64.

CECCA, F., S.CRESTA, G.PALLINI, AND M. SANTANTONIO. 1987. Il Giurassico di Monte Nerone (Appennino marchigiano, Italia Centrale): biostratigrafia, litostratigrafia ed evoluzione paleogeografica, p. 63-139. *In* Pallini et al. (eds.), Atti II Convegno Internazionale F.E.A. Pergola 1987.

CENTAMORE, E., M. CHIOCCHINI, G. DEIANA, A. MICARELLI, AND V. PIERUCCINI. 1971. Contributo alla conoscenza del Giurassico dell' Appennino umbro-marchigiano. Memorie Societa Geologica Italiana, 8 (3).

COLACICCHI, R., AND A. BIGOZZI. 1995. Event stratigraphy and carbonate platform-basin interrelations during the Jurassic in the Central Apennines. Paleopelagos 5:111-128.

____, M. NOCCHI, G. PARISI, P. MONACO, A. BALDANZA, S. CRESTA, AND G. PALLINI. 1988. Paleoenvironmental analysis from Lias to Malm (Corniola to Maiolica Formations) in the Umbria-Marche basin, Central Italy (preliminary report). Proceedings 2nd International Symposium on Jurassic Stratigraphy (1987), Lisboa, 2:717-728.

____, L. PASSERI, AND G. PIALLI. 1970. Nuovi dati sul Giurese Umbro-Marchigiano ed ipotesi per un suo inquadramento regionale. Memorie Societa Geologica Italiana, 9:839-874.

CORBIN, J. C. 1994. Evolution géochimique du Jurassique du Sud-Est de la France: influence des variations du niveau marin et de la tectonique. Ph. D. thesis, University Paris VI, p.175.

COURTILLOT, V. 1995. La vie en catastrophes. Fayard, Paris, 278 p.

DROMART, G., P. ALLEMAND, J. P. GARCIA AND C. ROBIN. 1996. Variation cyclique de la production carbonatée au Jurassique le ong d'un transect Bourgogne-Ardèche, Est-France. Bulleitn Societe Géologique de France, 167(3):423-433.

FARINACCI, A., N. MARIOTTI, U. NICOSIA, G. PALLINI, AND F. SCHIAVINOTTO. 1981. Jurassic sediments in the Umbro-Marchean Apennine: an alternative model. Proceedings of Rosso Ammonitico Symposium, ed. Tecnoscienza, Roma.

FAVRE, P. AND G. M. STAMPFLI. 1992: From rifting to passive margin: the examples of the Red Sea, Central Atlantic and Alpine Tethys. Tectonophysics 215:69-97.

FOLK, R. L., AND E. F. MCBRIDE. 1978. Radiolarites and their relation to subjacent "Oceanic crust" in Liguria , Italy. Journal of Sedimentary Petrology , 48 (4):1069-1102.

FÖLLMI, K. B., H. WEISSERT, M. BISPING AND H. P FUNK. 1994. Phosphogenesis, carbon-isotope stratigraphy, and carbonate-platform evolution along the Lower Cretaceous Northern Tethyan margin. Geological Society of America Bulletin, 106:729-746.

FRIEDMAN, I., AND J. R. O'NEIL. 1977. Compilation of stable isotope fractionation factors of geochemical interest, p. 1-12. *In* M. Fleicher (ed.), Data of geochemistry. U.S. Geological Survey Professional Paper.

GARRISON, R. E., AND A. G. FISCHER. 1969. Deep-water limestones and radiolarites of the Alpine Jurassic. p. 20-56. *In* Depositional environments in Carbonate Rocks. SEPM Special Publication, 14.

GRUSZCZYNSKI, M., A. HOFFMAN, K. MALKOWSKI, A. TATUR AND S. HALAS. 1990. Some geochemical aspects of life and burial environments of late Jurassic scleractinian corals from northern Poland. Neues Jahrbuch fur Geologie Paläontologie und Monatshefte. 1990, p. 673-686.

HAIDJI, S. 1991. Stratigraphye isotopique des carbonates pelagiques (Jurassic superieur-Cretace inferieur du bassin d' Ombrie-Marches (Italie). PhD Thesis, Université de Paris, 1_118.

HALLAM, A. 1988. A re-evaluation of Jurassic eustasy in the light of new data and the revised Exxon curve, p. 261-273. *In* Sea-level change: an integrated approach. SEPM Special Publication, 42.

HALLOCK, P., A. C. HINE, G. A. VARGO, J. A. ELROD AND W. C. JAAP. 1988. Platforms of the Nicaraguan Rise: Examples of the sensitivity of carbonate sedimentation to excess trophic resources: Geology, 16:1104-1107.

HAQ, B. U., J. HANDEBOL AND P. VAIL. 1988. Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level changes. SEPM Special Publication, 42:71-108.

HOFFMAN, A., M. GRUSZCZYNSKI, K. MALKOWSKI, S. HALAS, B. A. MATYJA AND A.WIERBOWSKI. 1991. Carbon and oxygen isotope curves for the Oxfordian of central Poland. Acta Geologica Polonica, 43:157-164.

HONJO, S. 1984. Study of ocean fluxes in time and spaces by bottom-tethered sediment trap arrays: A recommendation, p. 305-324. *In* Global Ocean Flux Study: Proceedings of a Workshop, September 10-14 1984, National Academy Press, Washington DC.

JENKYNS, H. C. 1996. Relative sea-level change and carbon isotopes: data from the Upper Jurassic (Oxfordian) of central and Southern Europe. Terra Nova, 8 (1):75-85.

KENNETT, J. 1982. Marine Geology. Prentice-Hall INC. Englewood Cliffs: 813 p.

KOWALLIS, B. J., A. L. DEINO, E. H. CHRISTIANSEN, F. PETERSON AND C. E. TURNER. 1998. Implications of new high precision $^{40}Ar/^{39}Ar$ laser-microbe ages from Utah and Colorado for the Jurassic Time scale, p. 48-49. *In* 5th International Symposium on the Jurassic System, Vancouver.

LANCELOT, Y., R. LARSON, et al. 1990. Proceedings Ocean Drilling Program, Initial Reports, 129: College Station, TX (Ocean Drilling Program), p.1-488.

LARSON, R. L., AND W. W. SAGER. 1992. Skewness of magnetic anomalies M0 to M29 in the northwestern Pacific, p. 471-481. *In* R. L. Larson, , Y. Lancelot et al, Proc. ODP, Sci. Results, 129: College Station, TX (Ocean Drilling Program).

____, A. T. FISHER, R. D. JARRARD, K. BECKER, AND ODP Leg 144 Shipboard Scientific Party. 1993. Highly permeable and layered Jurassic oceanic crust in the Western Pacific, Earth and Planetary Science Letters, 199:71-83.

MATSUOKA, A. 1991. Middle Jurassic radiolarians from the western Pacific. Proceedings of the International Symposium on Shallow Tethys, 3:163-173.

MATTIOLI, E. 1995. Stratigrafia a nannofossili calcarei nel Giurassico del bacino Umbro-Marchiagiano: produttività primaria, preservazione ed effetti della diagenesi. Unpublished Ph.D. Thesis, University of Perugia, 208 pp.

MCBRIDE, E. F. AND R. L. FOLK. 1979. Features and origin of Italian Jurassic radiolarites deposited on a continental crust. Journal of Sedimentary Petrology 49 (3):837-868.

MILLIAM, J. D. AND K. TAKAHASHI. 1995. Carbonate and opal production and accumulation in the ocean. *In* T.M. Usselman et al. (eds.), Global Surficial Geofluxes: Modern to Glacial. National Academy Press.

NICOSIA, U., AND G. PALLINI. 1977. Hermatipic corals in the Tithonian pelagic facies of central Apennines. Evidences of Upper Jurassic sea-level changes. Geologica Romana, 16:243-261.

PISERA, A., M. SATIR, M. GRUSZCZYNSKI, A. HOFFMAN AND K. MALKOWSKI. 1992. Variation in $\delta^{13}C$ and $\delta^{18}O$ in Late Jurassic carbonates, submediterranean province, Europe. Annales Societe Geologique de Poland, 62:141-147.

SANTANTONIO, M. 1993. Facies association and evolution of pelagic carbonate platform/basin system: examples from the Italian Jurassic. Sedimentology, 40:1039-1067.

____. 1994. Pelagic Carbonate Platforms in the Geological Record: Their Classification and Sedimentary and Paleotectonic Evolution. American Association of Petroleum Geologists Bulletin, 78, 122-141.

____, F. GALUZZO AND G. GILL. 1996. Anatomy and palaeobathymetry of a Jurassic pelagic carbonate platform/basin system. Rossa Mts, Central Apennines (Italy), Geological implications. Palaeopelagos 6:123-169.

SELLWOOD, B. W., P. VALDES AND G. D. PRICE. 1998. Geological evaluation of GCM simulations of Late Jurassic palaeoclimate, p. 84-85. *In* 5th International Symposium on the Jurassic System, Vancouver.

WEISSERT, H., AND J. E. T. CHANNELL. 1989. Tethyan carbonate carbon isotope stratigraphy across the Jurassic-Cretaceous boundary: an indicator of decelerated global carbon cycling? Paleoceanography, 4:483-494.

____, AND A. LINI. 1991 Ice age interludes during the time of Cretaceous greenhouse climate, p. 173-191. *In* D.W. Muller, J.A. McKenzie, and H.Weissert (eds.), Controversies in Modern Geology, Academic Press, New York.

____, AND H. MOHR. 1995. Late Jurassic climate and its impact on carbon cycling. Palaeogeography, Palaeoclimatology, Palaeoecology, 122:27-43.

WETZEL, A. 1989. Influence of heat flow on ooze/chalk cementation; quantification from consolidation parameters in DSDP site 504 and 505 sediments. Journal of Sedimentary Petrology, 59:539-547.

WINTERER, E. L., AND A. BOSELLINI. 1981. Subsidence and sedimentation on Jurassic passive continental margins, Southern Alps, Italy. American Association of Petroleum Geologists Bulletin, 65 (3):394-421.

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A Jurassic Rift System in the Canadian Arctic Islands

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Abstract: Geological, geophysical and biostratigraphic data from Prince Patrick Island serve to illuminate rift-related sedimentation and deformation of Jurassic and Lower Cretaceous strata of the western Arctic Islands. Rifting along northerly-striking normal faults began no earlier than the Late Toarcian and continued through as many as eight episodes before the Albian. The parallelism of Jurassic normal faults to the adjacent shelf edge is consistent with an extensional origin for the Canadian portion of the polar margin on the Amerasian Basin.

Introduction and Principal Geological Features

Previous work indicates that sea floor spreading within the Amerasian Basin was preceded by an interval of Middle Jurassic through Early Cretaceous rifting (Moore et al., 1994; Dixon, 1996). An improved understanding of the Canadian portion of this rift system has resulted from recent reconnaissance-scale bedrock geological mapping and related studies by Geological Survey of Canada staff. The current paper identifies the local depocentres and high blocks associated with a portion of the mapped Jurassic fault system, defines the nature and duration of the rift phase, and provides new constraints on viable tectonic models for subsequent plate spreading in the Eglinton Island and Prince Patrick Island area between 75°45' and 77°45' N adjacent to the Canadian polar margin of the Amerasian Basin (Fig. 1).

Four intra-cratonic basins of Jurassic-Cretaceous age, each separated by Devonian basement blocks, lie southwest of the Sverdrup Basin (Fig. 2). These include Banks Basin (max. 3000m thick), Eglinton Basin (to 900m), M'Clure Strait Basin (min. 600m thick) in the western end of M'Clure Strait, and Tullett Basin of west central Prince Patrick Island (max. 2300m thick) most of which underlies younger cover (inset, Fig. 2). The entire region is transected by northerly-striking normal faults, horst blocks, and grabens that contain Triassic and younger sedimentary rocks. However, the fault system responsible for the horst and graben system running through these intra-cratonic basins was inactive prior to the Late Toarcian. Unconsolidated sands and gravel of Pliocene age form an unconformable blanket above deformed Albian and older rocks throughout the western half of Prince Patrick Island (Fig. 2).

Correlation and Biostratigraphy

Measured outcrop sections have been correlated to wells and seismic profiles using standard methods (Figs. 3, 4). These correlations are supported by local and regional studies of ammonites, bivalves, foraminifera, and dinoflagellates. Fossil collections from the report area and throughout the Sverdrup Basin allow recognition of 22 discrete ammonite, and four bivalve (*Buchia*), assemblages representing every stage of the Jurassic in marine facies, possibly including the Hettangian. The oldest Jurassic on Prince Patrick Island, featuring assemblages with *Badouxia* (?), *Coroniceras* sp. and *Arnioceras* (?) sp. of Late Hettangian (?) and Early Sinemurian age (Fig. 3; loc. 16, Fig. 4), also contains diagnostic foraminifera and dinoflagellate assemblages.

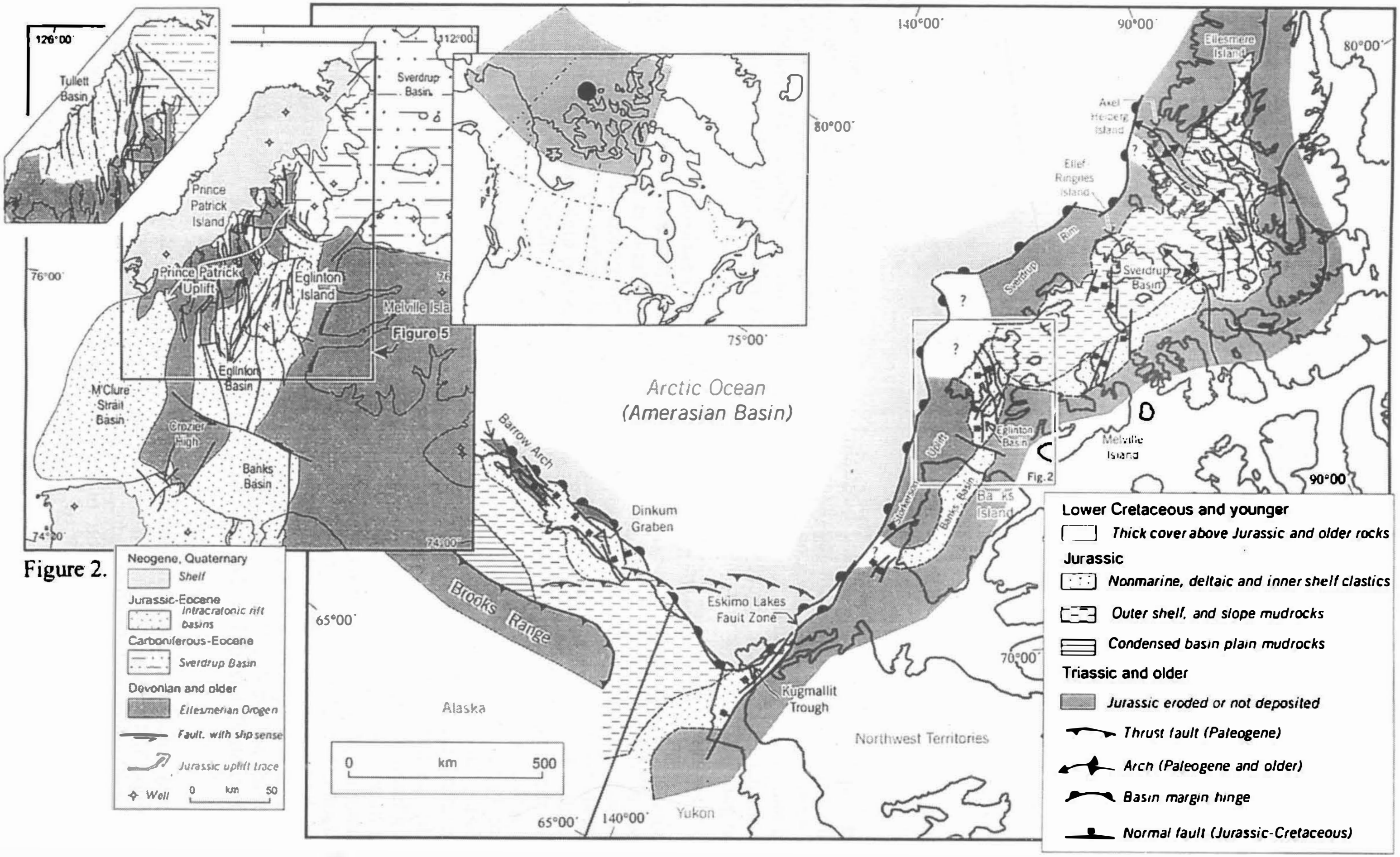


Figure 2.

Figure 1.

The foraminiferal subdivisions are based primarily on collections obtained from outcrop sections on western Axel Heiberg Island and more distal surface and subsurface sections of eastern Axel Heiberg and west-central Ellesmere islands (Fig. 1). The *Glomospira perplexa* assemblage is defined by collections from two outcrop sections on Prince Patrick Island (loc. 16, Fig. 4) and from two well intersections (Jameson Bay C-31 and Satellite F-68; Fig. 5). The dinoflagellate zonation derives from revisions to the original Oppel-Zonation of Davies (1983) and includes two new zones for the Hettangian(?) and Sinemurian.

Depositional History

Pre-rift Jurassic

The pre-rift Jurassic sequence of Prince Patrick Island features the Grosvenor Island Formation (Upper Hettangian? - Sinemurian; Fig. 4) and Intrepid Inlet through Cape Canning members of the Jameson Bay Formation (Upper Pliensbachian - Upper Toarcian).

Rift Phase

Stage 1. Late Toarcian: The oldest structural disturbance, potentially associated with the onset of the Jurassic rift phase, features uplift and erosion of the Jameson Bay Formation (Intrepid and Cape Canning members) prior to deposition of overlapping, nearshore, marine quartz sandstones of the Lower Aalenian Sandy Point Formation on Green Bay Uplift (locs. 14, 15; Figs. 4-6). However, Jameson Bay rocks are differentially preserved in grabens to the southwest. Seismically-defined normal faults may also have been active during this interval, for example, within Hardinge Bay Graben (in unit lmJ on Figs. 6a, 7a).

Stage 2. Late Aalenian and Early Bajocian: Geophysical profiles indicate widespread development of normal faults throughout Tullett Basin during the Late Aalenian to mid-Early Bajocian. This resulted in differential preservation of the Sandy Point-Jameson Bay seismic sequence (unit lmJ) in the downthrown hanging wall of normal faults associated with Hardinge Bay, Richards Point, Tullett Central, and Landing Lake grabens, and local angular unconformities, over and adjacent to, various high blocks (Figs. 6, 7a-c). Some of these faults terminate below the McConnell Island-Hiccles Cove seismic sequence (unit mJ).

Stage 3. Mid-Bajocian through Callovian: Indirect evidence for deformation during this interval includes preservational limits for strata that shift from northwesterly-trending in the Aalenian (i.e., parallel to the Sverdrup Basin margin) to northerly-trending, and parallel to many of the mapped faults (Fig. 5), during the Bajocian and Bathonian. Strata of these ages are found throughout Eglinton Basin. Locally divergent seismic reflectors and growth faults occur within the McConnell Island-Hiccles Cove seismic sequence of Green Bay and Landing Lake grabens (Figs. 7c, d). Additional evidence for erosional unroofing is provided by braidplain sandstones and fluvial channels identified in outcrops of the upper Hiccles Cove Formation and seismically-defined channels that occur within the upper part of the same sequence (unit mJ, Fig. 7b).

Stage 4. Callovian through Early Oxfordian: Constraints on the duration of deformation during this interval are provided by the erosion of Callovian and older sandstones and subsequent onlap by late Early Oxfordian and younger marine shales (Ringnes Formation). Evidence for tectonism prior to deposition of the Ringnes Formation is widespread throughout both Tullett Basin and Prince Patrick Uplift including erosion of the entire Callovian part of the Hiccles Cove Formation on the west edge of Station Creek Graben (locs. 9, 10; Fig. 4), differential preservation of the same Callovian portion of the Hiccles Cove Formation below Ringnes Formation in grabens west and north of Mould Bay (locs. 6 to 8), sub-Ringnes channelling, and a local angular unconformity above the McConnell Island-Hiccles Cove sequence (mJ) on seismic profiles of Hardinge Bay, Richards Point, Tullett Central, and Landing Lake grabens (Figs. 7a-d).

Figure 1. (right). Distribution of Jurassic rocks of Arctic Alaska and Arctic Canada with some key structural elements of Jurassic and younger age. Compilation of data mostly from Embry (1991), Moore et al. (1994), and Wheeler et al. (1996).

Figure 2. (upper left). Principal geological features of the western Canadian Arctic Islands. M'Clure Strait Basin is not featured on Figure 1 as it is believed to contain only Cretaceous and younger strata. The inset at upper left illustrates the major features present beneath Neogene cover.

Stage 5. Early to Mid-Kimmeridgian: Evidence for deformation during this interval is circumstantial and limited. Clast-supported gravel and pebble conglomerates in Landing Lake Graben (loc. 7; Fig. 4), and boulder beds in Giants Causeway Graben (loc. 1), are interbedded with marine shales of the Ringnes Formation. The coarser sediments are interpreted as deltaic channel-fill deposits of potential syntectonic origin. However, distance of transport may be significant as clasts (Devonian sandstones and chert) are mostly subrounded..

Stage 6. Middle and Late Tithonian: Cobble conglomerate and breccia occur with variably well rounded to angular clasts of locally-derived Devonian sandstone in the Awingak Formation in Cape Cam Graben (loc. 2; Fig. 4). Similar cobble conglomerates are interbedded with marginal marine sandstones and minor shales of the Awingak Formation in Comfort Cove Graben on southwestern Melville Island (Fig. 5; Harrison, 1995).

Stage 7. Late Tithonian to Hauterivian: Differential preservation of the mostly Upper Jurassic Ringnes-Awingak-Deer Bay sequence below unconformable cover of Lower Cretaceous Isachsen Formation (Barremian-Aptian), including the correlative seismic units, supports an inference of rift-related fault motion during the Late Tithonian through Hauterivian depositional hiatus. This evidence has been acquired from seismic profiles of northern Eglinton Island within central Eglinton Basin, from surface sections on Prince Patrick Uplift, and from seismic profiles over Houghton Head, Hardinge Bay, and Richards Point grabens in Tullett Basin. Both locally and regionally Lower Cretaceous oversteps the Jurassic and rests directly on Devonian.

Stage 8. Barremian to Aptian: The highest widespread occurrence of rift-related faulting and record of associated deposition occurs in the Lower Cretaceous Isachsen Formation. Features include differential preservation of Isachsen Formation sandstones and the equivalent seismic interval (lK) in the hanging wall of normal faults in Eglinton and M'Clure Strait basins, and various grabens of Prince Patrick Uplift and Tullett Basin (Figs. 6, 7). The age of these strata is provided by Barremian dinoflagellates and other microflora collected from the lower and medial part of the Isachsen Formation in Station Creek Graben. Braidplain gravels of potential syntectonic origin also occur in the uppermost preserved part of the Isachsen Formation in Station Creek Graben where microflora are Barremian to Aptian in age. Syntectonic conglomerates and differentially preserved Lower Cretaceous strata have been documented from northern Banks Basin (Miall, 1979) and on southeastern Melville Island (Harrison, 1995).

Regional Implications and Conclusions

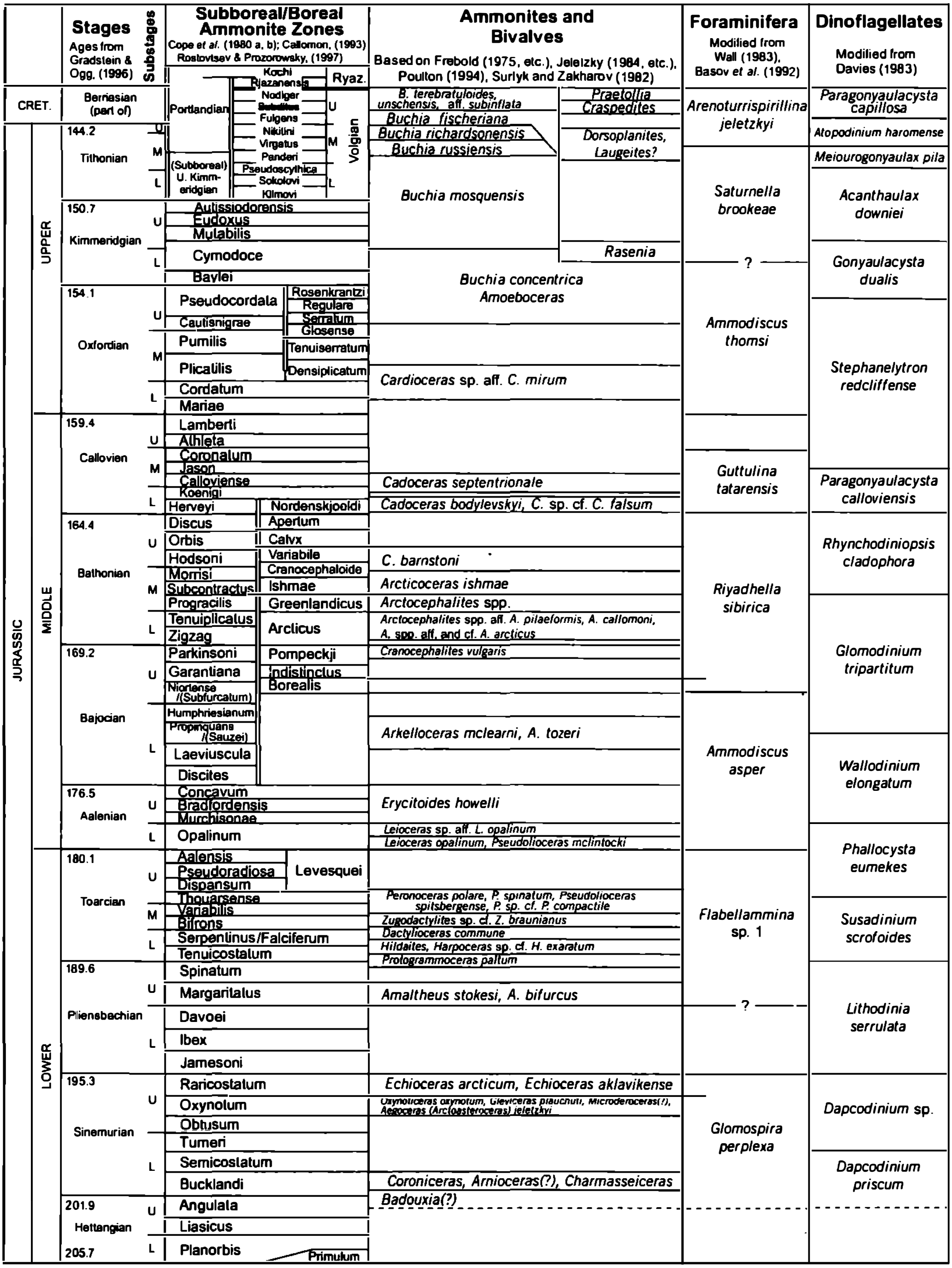
1. Jurassic and Lower Cretaceous rift-related structures in the western Arctic Islands are part of a rift system of comparable age that is parallel and close to the present Arctic continental margin and traceable southwards to the Mackenzie Delta region. Similar structures exist along the same plate margin north of Alaska.
2. Rifting began not earlier than the Late Toarcian. As many as eight pre-Albian rift stages are indicated by field and subsurface data, with local age constraints provided by diagnostic shelly macrofauna, foraminifera and dinoflagellate assemblages.
3. Normal faults in the rift zone are parallel to the adjacent continental margin, and transcurrent faults are more or less perpendicular to this margin. These observations are consistent with plate tectonic models that require extension perpendicular to the Canadian portion of the continental margin during ocean floor spreading in the Amerasian Basin.

References

BASOV, V. A., J. H. WALL, A. R. SOKOLOV, S. P.YAKOVLEV, T. P. POULTON, AND A. F. EMBRY. 1992. The *Riyadhella sibirica* foraminiferal zone in the Middle Jurassic of northern Russia and Canada. Abstracts, First International Conference on Arctic Margins, Anchorage, Alaska, 63.

CALLOMON, J. H. 1993. The ammonite succession in the Middle Jurassic of East Greenland. Geological Society of Denmark Bulletin, 40, 83-113.

Figure 3. Correlation of Boreal Jurassic biozones of Prince Patrick Uplift and the western Sverdrup Basin on Prince Patrick Island. Only the most recent references are cited. Arctic Canada *Buchia* zones have been adjusted partially to reflect ages provided by ammonites in Greenland and Siberia (Surlyk and Zakharov, 1982)



		Stages Ages from Gradstein & Ogg. (1996)		**Substages**	**Subboreal/Boreal Ammonite Zones** Cope *et al.* (1980 a, b); Callomon, (1993) Rostovtsev & Prozorowsky, (1997)				**Ammonites and Bivalves** Based on Frebold (1975, etc.), Jeletzky (1984, etc.), Poulton (1994), Surlyk and Zakharov (1982)		**Foraminifera** Modified from Wall (1983), Basov *et al.* (1992)	**Dinoflagellates** Modified from Davies (1983)
CRET.		Berriasian (part of)			Portlandian	Kochi	Ryaz.					
						Rjazanensis						
						Nodiger	U	Volgian	*B. terebratuloides, unschensis, aff. subinflata*	*Praetollia*	*Arenoturrispirillina jeletzkyi*	*Paragonyaulacysta capillosa*
						Subditus				*Craspedites*		
JURASSIC	UPPER	Tithonian	144.2	U		Fulgens			*Buchia fischeriana*			*Atopodinium haromense*
				M		Nikitini	M		*Buchia richardsonensis*	*Dorsoplanites, Laugeites?*		
						Virgatus			*Buchia russiensis*			
					(Subboreal) U. Kimmeridgian	Panderi			*Buchia mosquensis*		*Saturnella brookeae*	*Meiourogonyaulax pila*
				L		Pseudoscythica	L					*Acanthaulax downiei*
						Sokolovi						
						Klimovi						
		Kimmeridgian	150.7	U	Autissiodorensis							
					Eudoxus							
					Mutabilis							
				L	Cymodoce				*Buchia concentrica Amoeboceras*	*Rasenia*		*Gonyaulacysta dualis*
					Baylei						? *Ammodiscus thomsi*	
		Oxfordian	154.1	U	Pseudocordata	Rosenkrantzi						
						Regulare						
						Serratum						
					Cautisnigrae	Glosense						*Stephanelytron redcliffense*
				M	Pumilis	Tenuiserratum						
					Plicatilis	Densiplicatum			*Cardioceras* sp. aff. *C. mirum*			
				L	Cordatum							
					Mariae							
	MIDDLE	Callovien	159.4	U	Lamberti							
					Athleta							
				M	Coronatum						*Guttulina tatarensis*	
					Jason							
				L	Calloviense				*Cadoceras septentrionale*			*Paragonyaulacysta calloviensis*
					Koenigi							
					Herveyi	Nordenskjooldi			*Cadoceras bodylevskyi, C.* sp. cf. *C. falsum*			
		Bathonian	164.4	U	Discus	Apertum					*Riyadhella sibirica*	*Rhynchodiniopsis cladophora*
					Orbis	Calyx						
					Hodsoni	Variabile			*C. barnstoni*			
				M	Morrisi	Cranocephaloide						
					Subcontractus	Ishmae			*Arcticoceras ishmae*			
					Progracilis	Greenlandicus			*Arctocephalites* spp.			*Glomodinium tripartitum*
				L	Tenuiplicatus	Arcticus			*Arctocephalites* spp. aff. *A. pilaeformis, A. callomoni, A.* spp. aff. and cf. *A. arcticus*			
					Zigzag							
		Bajocian	169.2	U	Parkinsoni	Pompeckji			*Cranocephalites vulgaris*			
					Garantiana	Indistinctus						
					Niortense //(Subfurcatum)	Borealis						
				L	Humphriesianum						*Ammodiscus asper*	
					Propinquans //(Sauzei)				*Arkelloceras mclearni, A. tozeri*			*Wallodinium elongatum*
					Laeviuscula							
					Discites							
		Aalenian	176.5	U	Concavum				*Erycitoides howelli*			
					Bradfordensis							
					Murchisonae							
				L	Opalinum				*Leioceras* sp. aff. *L. opalinum* *Leioceras opalinum, Pseudolioceras mclintocki*			*Phallocysta eumekes*
	LOWER	Toarcian	180.1	U	Aalensis	Levesquei					*Flabellammina* sp. 1	
					Pseudoradiosa							
					Dispansum							
					Thouarsense				*Peronoceras polare, P. spinatum, Pseudolioceras spitsbergense, P.* sp. cf. *P. compactile*			*Susadinium scrofoides*
				M	Variabilis							
					Bifrons				*Zugodactylites* sp. cf. *Z. braunianus*			
				L	Serpentinus/Falciferum				*Dactylioceras commune* *Hildaites, Harpoceras* sp. cf. *H. exaratum*			
					Tenuicostatum				*Protogrammoceras paltum*			
		Pliensbachian	189.6	U	Spinatum							*Lithodinia serrulata*
					Margaritatus				*Amaltheus stokesi, A. bifurcus*			
				L	Davoei						?	
					Ibex							
					Jamesoni							
		Sinemurian	195.3	U	Raricostatum				*Echioceras arcticum, Echioceras aklavikense*		*Glomospira perplexa*	*Dapcodinium* sp.
					Oxynotum				*Oxynoticeras oxynotum, Gleviceras plauchuti, Microderoceras(?), Aegoceras (Arctoasteroceras) jeletzkyi*			
					Obtusum							
				L	Turneri							
					Semicostatum							*Dapcodinium priscum*
					Bucklandi				*Coroniceras, Arnioceras(?), Charmasseiceras*			
		Hettangian	201.9	U	Angulata				*Badouxia(?)*			
					Liasicus							
			205.7	L	Planorbis	Primulum						

COPE, J. C. W., T. A. GETTY, M. K. HOWARTH, N. MORTON AND H. S. TORRENS. 1980a. A correlation of Jurassic rocks in the British Isles, Part 1, Introduction and Lower Jurassic. Geological Society of London Special Report, 14, 73 p.

_____, K. L. DUFF, C. F. PARSONS, H. S. TORRENS, W. A. WIMBLEDON AND J. K. WRIGHT. 1980b. A correlation of Jurassic rocks in the British Isles, Part 2, Middle and Upper Jurassic. Geological Society of London Special Report, 15, 109 p.

DAVIES, E. H. 1983. The dinoflagellate Oppel-zonation of the Jurassic-Lower Cretaceous sequence in the Sverdrup Basin, Arctic Canada. Geological Survey of Canada Bulletin, 359, 59p.

DIXON, J. 1996. Geological Atlas of the Beaufort-Mackenzie Area. Geological Survey of Canada Miscellaneous Report, 59, 173 p.

EMBRY, A. F. 1991. Mesozoic history of the Arctic Islands, p. 369-433. *In* H. P. Trettin (ed.), Chapter 14, Geology of the Innuitian Orogen and Arctic Platform of Canada and Greenland. Geological Survey of Canada, Geology of Canada, no. 3, 369-433.

FREBOLD, H. 1975. The Jurassic faunas of the Canadian Arctic, Lower Jurassic ammonites, biostratigraphy and correlations. Geological Survey of Canada Bulletin, 243, 35 p.

GRADSTEIN, F. M. AND J. OGG. 1996. A Phanerozoic time scale. Episodes, 19:3-5, 26.

HARRISON, J. C. 1995. Melville Island's salt-based fold belt, Arctic Canada. Geological Survey of Canada Bulletin, 472, 331 p.

JELETZKY, J. A. 1984. Jurassic-Cretaceous boundary beds of western and Arctic Canada and the problem of the Tithonian-Berriasian stages in the Boreal Realm, p. 175-255. *In* G.E.G. Westermann (ed.), Jurassic-Cretaceous biochronology and paleogeography of North America. Geological Association of Canada Special Paper, 27.

MIALL, A. D. 1979. Mesozoic and Tertiary geology of Banks Island, Arctic Canada. Geological Survey of Canada Memoir, 387, 235 p.

MOORE, T. E., W. K. WALLACE, K. J. BIRD, S. M. KARL, C. G. MULL AND J. T. DILLON. 1994. Geology of northern Alaska, p. 49-140. *In* G. Plafker and H.C. Berg (eds.), Chapter 3, The Geology of Alaska. Geological Society of America, Boulder, Colorado, The Geology of North America, v. G-1.

POULTON, T. P. 1994. Jurassic stratigraphy and fossil occurrences - Melville, Prince Patrick, and Borden islands, p. 161-193. *In* R.L. Christie and N.J. McMillan (eds.), The Geology of Melville Island, Arctic Canada. Geological Survey of Canada Bulletin, 450.

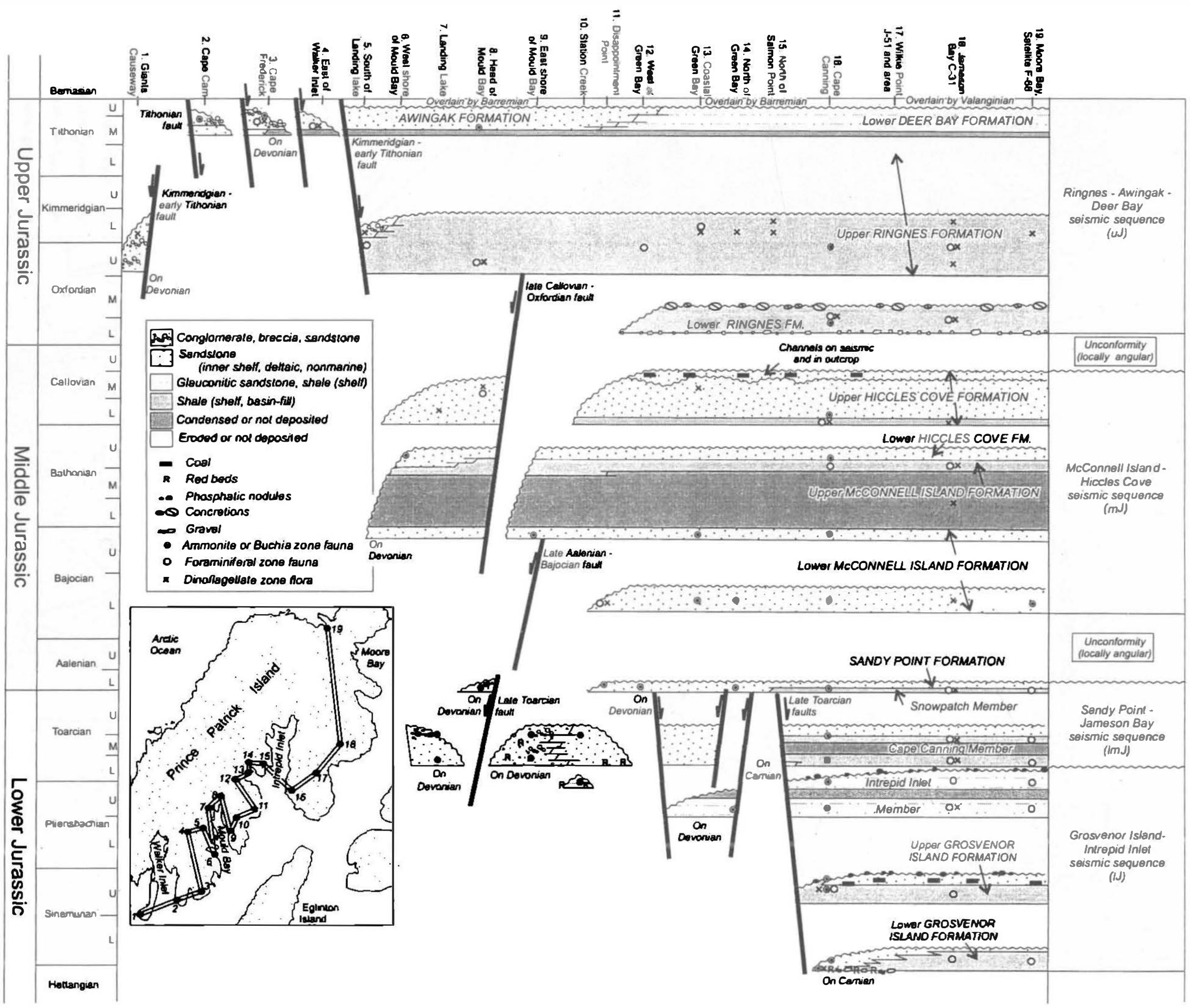
ROSTOVTSEV, K. O. AND V. A. PROZOROWSKY. 1997. Information on resolutions of standing commissions of the Interdepartmental Stratigraphic Committee (ISC) on the Jurassic and Cretaceous systems, International Subcommission on Jurassic Stratigraphy Newsletter, 24:48-49.

SURLYK, F. AND V. A. ZAKHAROV. 1982. Buchiid bivalves from the Upper Jurassic and Lower Cretaceous of East Greenland. Paleontology, 25:727-753.

WALL, J. H. 1983. Jurassic and Cretaceous foraminiferal biostratigraphy in the eastern Sverdrup Basin, Canadian Arctic Archipelago. Bulletin of Canadian Petroleum Geology, 31, 246-281.

WHEELER, J. O., P. F. HOFFMAN, K. D. CARD, A. DAVISON, B. V. SANFORD, A. V. OKULITCH AND W. R. ROEST. 1996. Geological Map of Canada. Geological Survey of Canada Map 1860A, scale 1:5,000,000.

Figure 4. Correlation of Jurassic strata of Prince Patrick Uplift and the western Sverdrup Basin on Prince Patrick Island.


1. Giants Causeway
2. Cape Cam-
3. Cape Frederick
4. East of Walker Inlet
5. South of Landing Lake
6. West shore of Mould Bay
7. Landing Lake
8. Head of Mould Bay
9. East shore of Mould Bay
10. Station Creek
11. Disappointment Point
12. West of Green Bay
13. Coastal Green Bay
14. North of Green Bay
15. North of Salmon Point
16. Cape Canning
17. Wilkie Point J-51 and area
18. Jameson Bay C-31
19. Moore Bay, Satellite F-68
Berriasian
Upper Jurassic
Tithonian
Kimmeridgian
Oxfordian
Middle Jurassic
Callovian
Bathonian
Bajocian
Aalenian
Lower Jurassic
Toarcian
Pliensbachian
Sinemurian
Hettangian
U
M
L
Tithonian fault
On Devonian
Kimmeridgian - early Tithonian fault
Overlain by Barremian
Overlain by Valanginian
AWINGAK FORMATION
Lower DEER BAY FORMATION
Upper RINGNES FORMATION
late Callovian - Oxfordian fault
Lower RINGNES FM.
Channels on seismic and in outcrop
Upper HICCLES COVE FORMATION
Lower HICCLES COVE FM.
Upper McCONNELL ISLAND FORMATION
Lower McCONNELL ISLAND FORMATION
Late Aalenian - Bajocian fault
SANDY POINT FORMATION
Snowpatch Member
Late Toarcian fault
Late Toarcian faults
Cape Canning Member
Intrepid Inlet
Member
On Carnian
Upper GROSVENOR ISLAND FORMATION
Lower GROSVENOR ISLAND FORMATION
Ringnes - Awingak - Deer Bay seismic sequence (uJ)
Unconformity (locally angular)
McConnell Island - Hiccles Cove seismic sequence (mJ)
Sandy Point - Jameson Bay seismic sequence (lmJ)
Grosvenor Island - Intrepid Inlet seismic sequence (lJ)
Conglomerate, breccia, sandstone
Sandstone (inner shelf, deltaic, nonmarine)
Glauconitic sandstone, shale (shelf)
Shale (shelf, basin-fill)
Condensed or not deposited
Eroded or not deposited
Coal
Red beds
Phosphatic nodules
Concretions
Gravel
Ammonite or Buchia zone fauna
Foraminiferal zone fauna
Dinoflagellate zone flora
Arctic Ocean
Prince Patrick Island
Intrepid Inlet
Walker Inlet
Mould Bay
Moore Bay
Eglinton Island

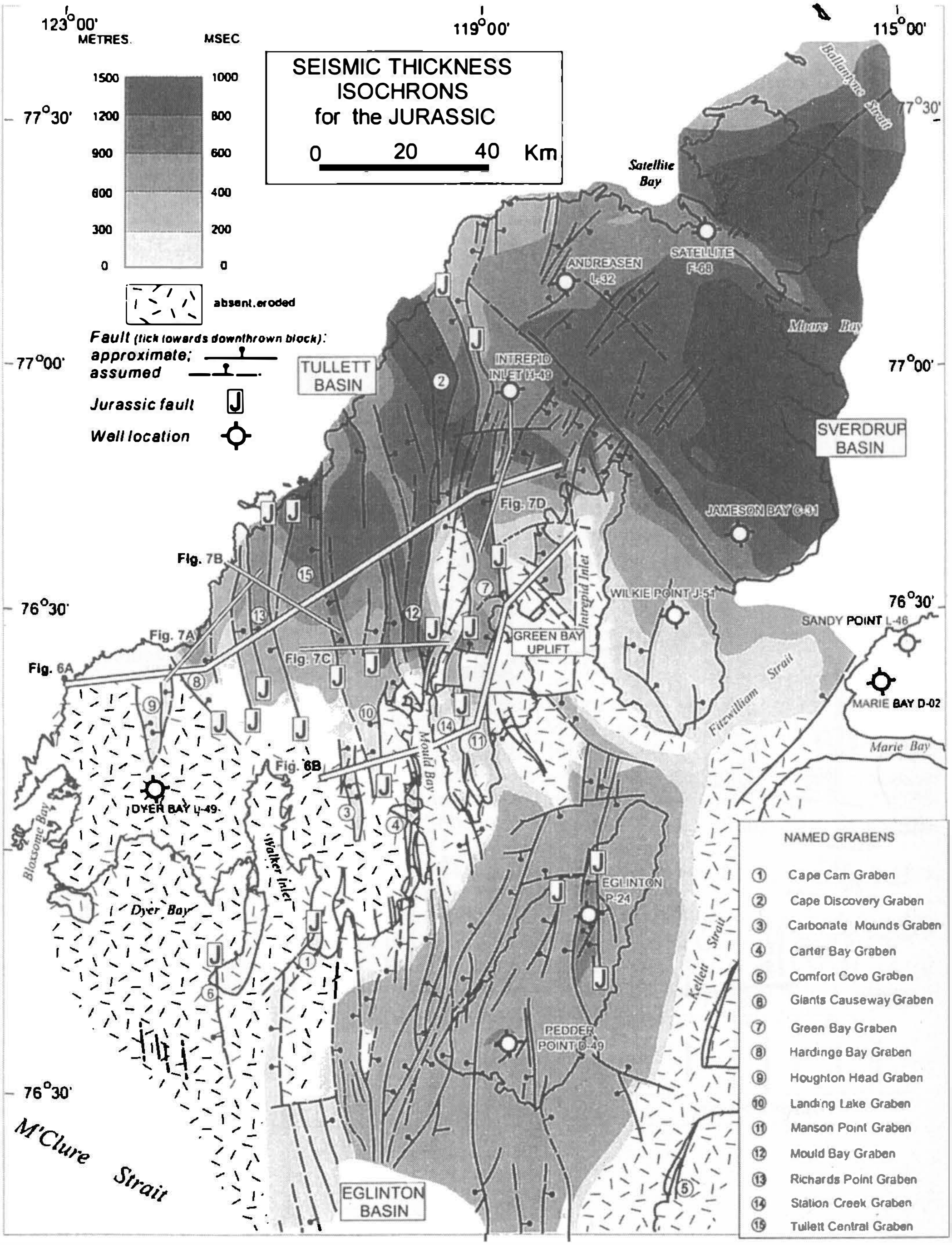


Figure 5. Seismic thickness isochrons for the Jurassic of Prince Patrick and Eglinton islands area. Contour interval is 200m or approximately 300m at 3km/s two-way travel time.

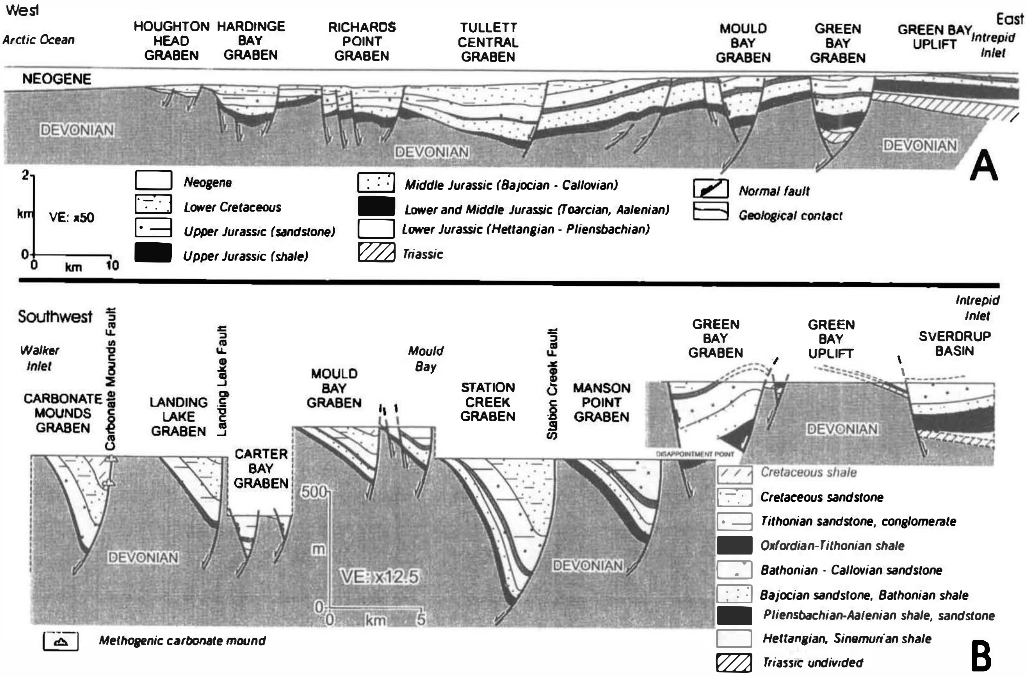
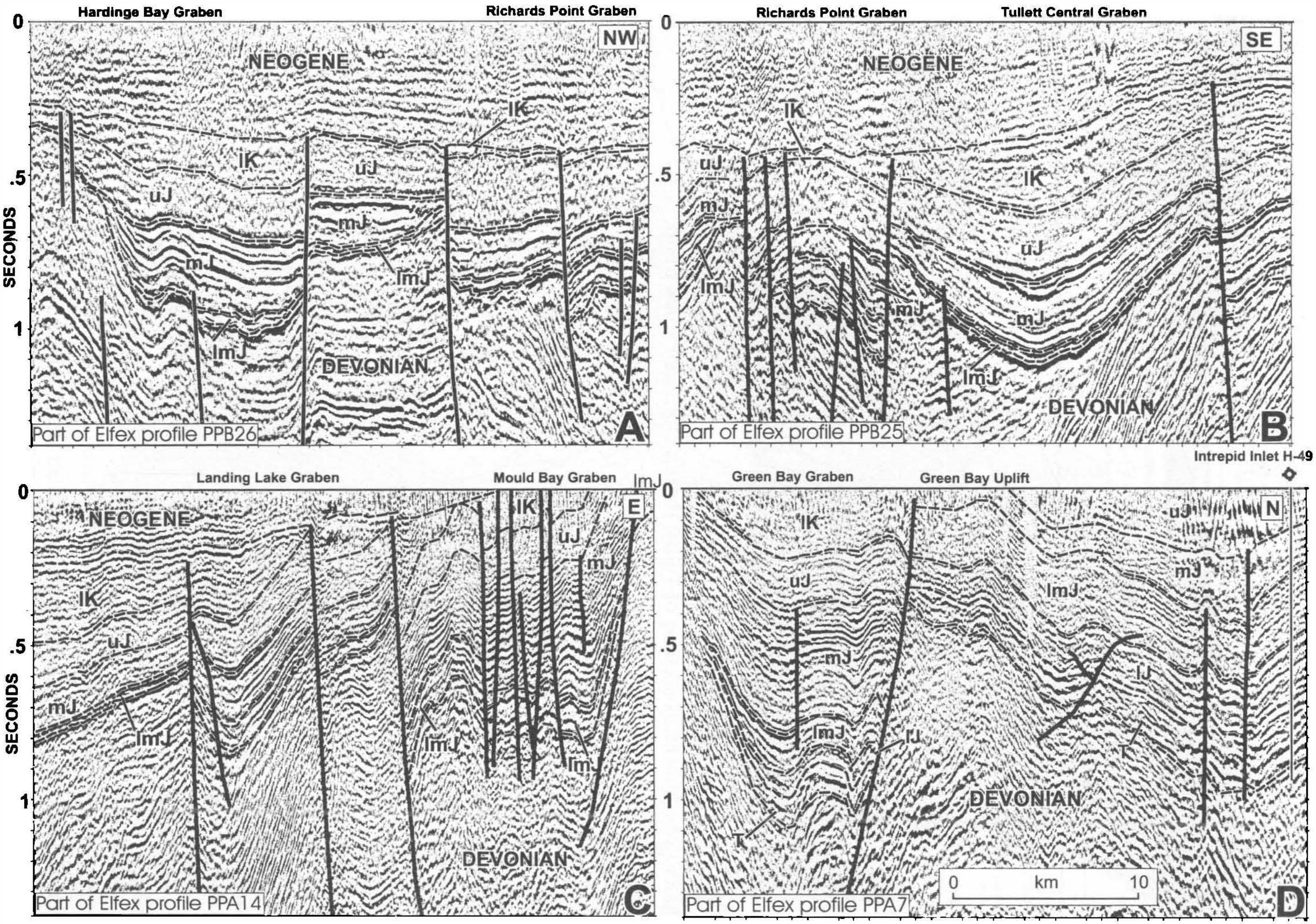


Figure 6. Structural cross-sections of Tullett Basin and Prince Patrick Uplift.

Next Page:

Figure 7. Representative seismic profiles of Tullett Basin. Abbreviations: Tr – Upper Triassic; lJ – Lower Jurassic (Hettangian(?)-Pliensbachian); lmJ – Lower-Middle Jurassic (Toarcian-Aalenian); mJ – Middle Jurassic (Bajocian-Callovian); uJ (Oxfordian-Tithonian); lK – Lower Cretaceous.


Hardinge Bay Graben
Richards Point Graben
NW
NEOGENE
lK
uJ
mJ
lmJ
DEVONIAN
Part of Elfex profile PPB26
A
SECONDS
0
.5
1
Richards Point Graben
Tullett Central Graben
SE
NEOGENE
lK
uJ
mJ
lmJ
DEVONIAN
Part of Elfex profile PPB25
B
Landing Lake Graben
Mould Bay Graben
lmJ
E
NEOGENE
lK
uJ
mJ
lmJ
DEVONIAN
Part of Elfex profile PPA14
C
Green Bay Graben
Green Bay Uplift
Intrepid Inlet H-49
N
lK
uJ
mJ
lmJ
lJ
T
DEVONIAN
0
km
10
Part of Elfex profile PPA7
D

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Lower-Middle Jurassic Rhythmites from the Lombard Basin, Italy: a Record of Orbitally Forced Carbonate Cycles Modulated by Secular Environmental Changes in West Tethys

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Keywords: Cyclostratigraphy, Orbital Forcing, Lombard Basin, Lower Jurassic, Middle Jurassic, Southern Alps, Domaro Limestone, Sogno Formation, Posidonia Alpina Facies, Precession, Obliquity

Abstract: An extensive, 220 meter thick, Upper Pliensbachian-Lower Bajocian basinal carbonate sequence from the Lombardy Prealps, Italy, consisting of hundreds of decimeter scale limestone/marl cycles, shows strong statistical evidence for sustained orbital forcing. The Lower Domerian segment has cyclic components indicative of orbital forcing by the precession-eccentricity syndrome (PES). The physical mechanisms responsible for the cyclicity may be explained as a combination of PES-forced eustatic sea level oscillations acting on neighboring platform carbonate factories supplying material to the basinal depositional site, and direct surface ocean forcing by PES-driven insolation controlling primary pelagic carbonate productivity. The Upper Domerian segment is disrupted by synsedimentary slumping, fracturing and resedimentation, and cannot be used to infer orbital forcing; similarly, the overlying Lower Toarcian interval is marked by a sudden decrease in carbonate accumulation, an influx of siliciclastics, and by a thick (8-10m) black shale representing the well-known Early Toarcian Oceanic Anoxic Event. Subsequently, carbonate sedimentation resumed with thick (meter scale), gradational alternations of limestone and marls, which thin progressively upward to the Toarcian/Aalenian transition, with carbonate cycle frequencies switching from components indicative of PES forcing, to those of obliquity (OB) forcing. This fundamental change was attended by other unusual global events, including a dramatically altered ocean circulation in western Tethys, marine biotic crises, widespread eutrophication of carbonate platforms, and a strong shift in paleoclimatic belts within the Laurasian continent. The sequence ends in the Lower Bajocian with a cessation of carbonate input, and the deposition of thin-bedded, pelagic, residual muds.

Introduction

The Lower and Middle Jurassic series of the South Alpine Province in northern Italy contain extensive sequences of thin-bedded carbonate cycles deposited in passive margin marine environments (Fig. 1). Past work has suggested that the cyclic deposition was forced by the Earth's orbital parameters (Weedon, 1989; Claps, 1993; Hinnov et al., 1994; Hinnov and Park, 1995; Claps et al., 1995; Hinnov and Park, 1998). This work focused principally on Pliensbachian-Toarcian sections where the cycles are particularly well developed, both on the Trento Platform in the east-lying Belluno Trough, and in the western reaches of the Lombard Basin. In the Generoso and Albenza Troughs there are also younger cyclic sequences continuing into the Middle Jurassic, but no studies have yet examined them in a similar context.

To extend our understanding of these high frequency cyclic sequences, we examined the Lower-Middle Jurassic section at Colle di Sogno (Bergamo) (Fig. 2), from the central part of the more than 1km deep Albenza Trough (the Monte Brughetto section of Gaetani and Poliani, 1978), which exhibits nearly continuous cyclic carbonate sedimentation from at least the Sinemurian (Lotharingian) to the Lower Bajocian (*Discites* Zone), with significant interruptions only in the Upper Domerian (tectonics) and in the Lower Toarcian black shale (the Oceanic Anoxic Event; Jenkyns, 1988). The cycles themselves undergo fundamental changes in sedimentation style following Late Domerian tectonics, including a series of frequency modulations after the deposition following Late Domerian tectonics, including a series of frequency modulations after the deposition of the Lower Toarcian black shale. The cycles stabilized during the early Middle Toarcian through to the Toarcian/Aalenian transition; subsequently, carbonate was increasingly replaced by biogenic (radiolarian) chert, and this combined biogenic sedimentation diminished throughout the Aalenian, ceasing altogether in the Early Bajocian (*Discites* Zone), after which a spectacular sequence of centimeter scale radiolarite/clay couplets ("ribbon radiolarites") was deposited (Baumgartner, 1987). The Colle di Sogno section thus spans a time of dramatic change in western Tethys, which began with the opening of the central Atlantic Ocean, altering marine circulation throughout Tethys (Jansa, 1991). Carbonate sedimentation in many South Alpine basinal environments progressively diminished over this time, first with the disappearance of aragonitic fossils in the Early Toarcian, then the remaining carbonate during the Aalenian-Early Bajocian.

Here we discuss the time-frequency evolution of the carbonate cycles of the Colle di Sogno sequence in the context of orbital forcing, sedimentation response, and the evidence for fundamental change in the global system. The frequency of the cycles themselves provides direct clues about global changes that came to influence the local carbonate productivity and depositional processes. Specifically, during the Early Toarcian, there is evidence that carbonate cyclicity underwent a switch from precession to obliquity forcing. A platform source for the bulk of the basinal carbonate during the Early Jurassic becomes crucial in considering the possible causes for the change, since a principal regulator of platform productivity and sedimentation is eustasy, which in turn at high frequencies is a gauge of an orbitally forced continental water (ice?) budget. We discuss how the appearance of a dominating obliquity component in the forcing of locally produced carbonate in a tropical sea is difficult to explain unless the carbonate was produced in platform environments under the control of global sea level oscillations, with obliquity components originating from

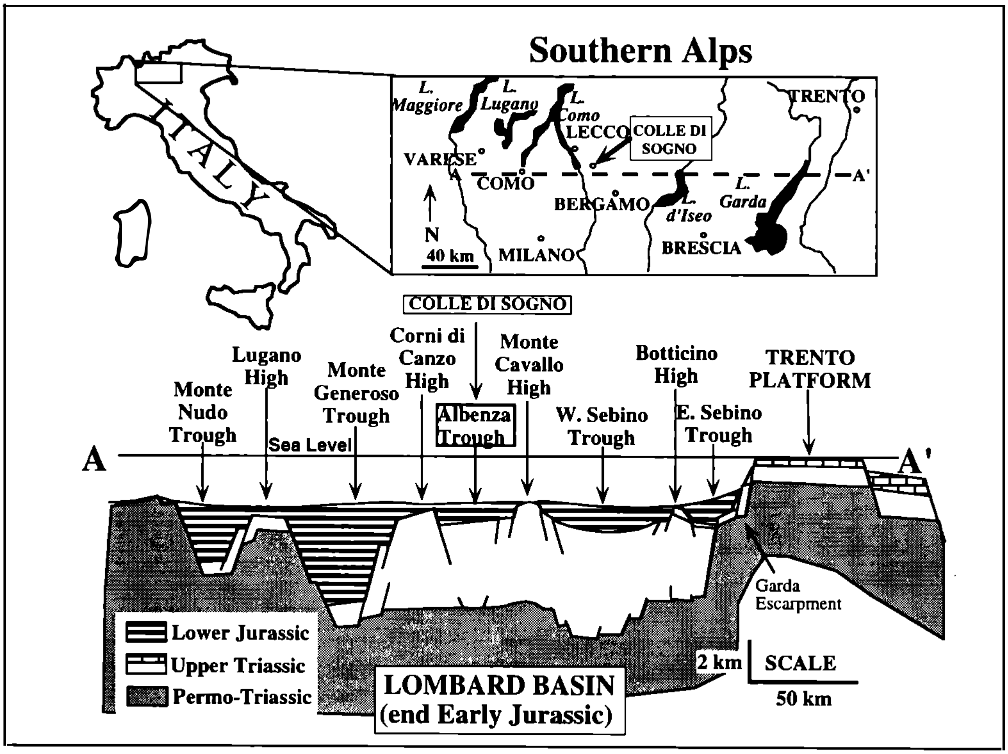


Figure 1. Map of Southern Alpine Province, northern Italy, with location of Colle di Sogno section, and palinspastic west-east cross section through late Early Jurassic paleogeography. Names of basins and structural elements from Gaetani (1975), Gaetani and Poliani (1978), Winterer and Bosellini (1981), Jenkyns (1988) and Sarti et al. (1992). Albenza Trough (boxed) is location of studied Colle di Sogno section.

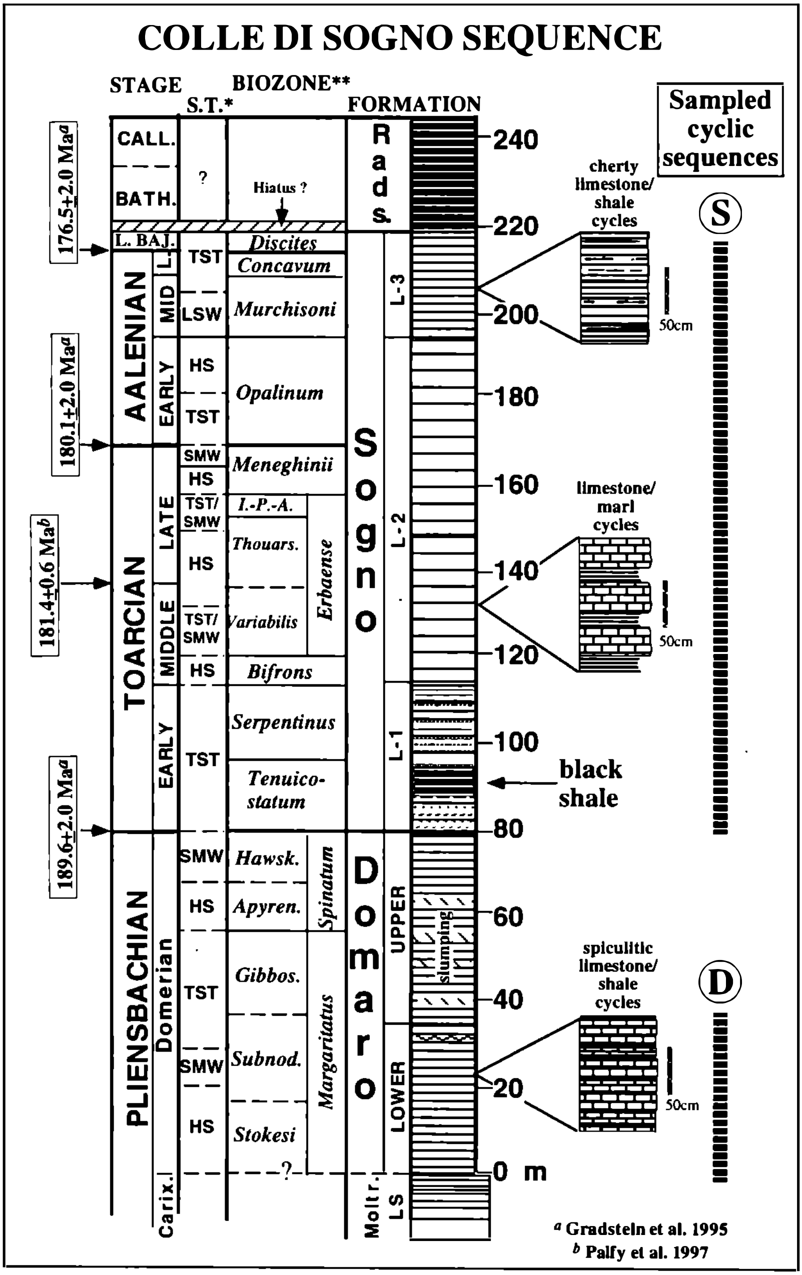


Figure 2. Summary of lithostratigraphy and biostratigraphy of Colle di Sogno section. Biozones estimated from ammonites (Gaetani and Poliani, 1978) and nannofossil stratigraphy developed by Erba and Cobianchi (unpublished study). Range of sampled sequences indicated to the right by D = Domaro sequence, and S = Sogno sequence. S.T. = Sequence Tracts defined by Haq et al. (1988). L-1, L-2 and L-3 are lithozones defined by Gaetani and Poliani (1978). Note that lower boundary of the Domaro is not defined, and may occur stratigraphically lower in the Carixian (Hinnov and Park, 1998).

orbitally forced water sequestration in polar latitudes, or in middle latitudes in both hemispheres simultaneously (Jacobs and Sahagian, 1995; Fischer and Hinnov, 1997; Hinnov and Fischer, 1997). Finally, we review the evidence for glaciation in Siberia during late Early and Middle Jurassic times (Frakes et al., 1992) in the north, and the possibility that an ice sheet could have developed on Antarctica (Valdes et al., 1995).

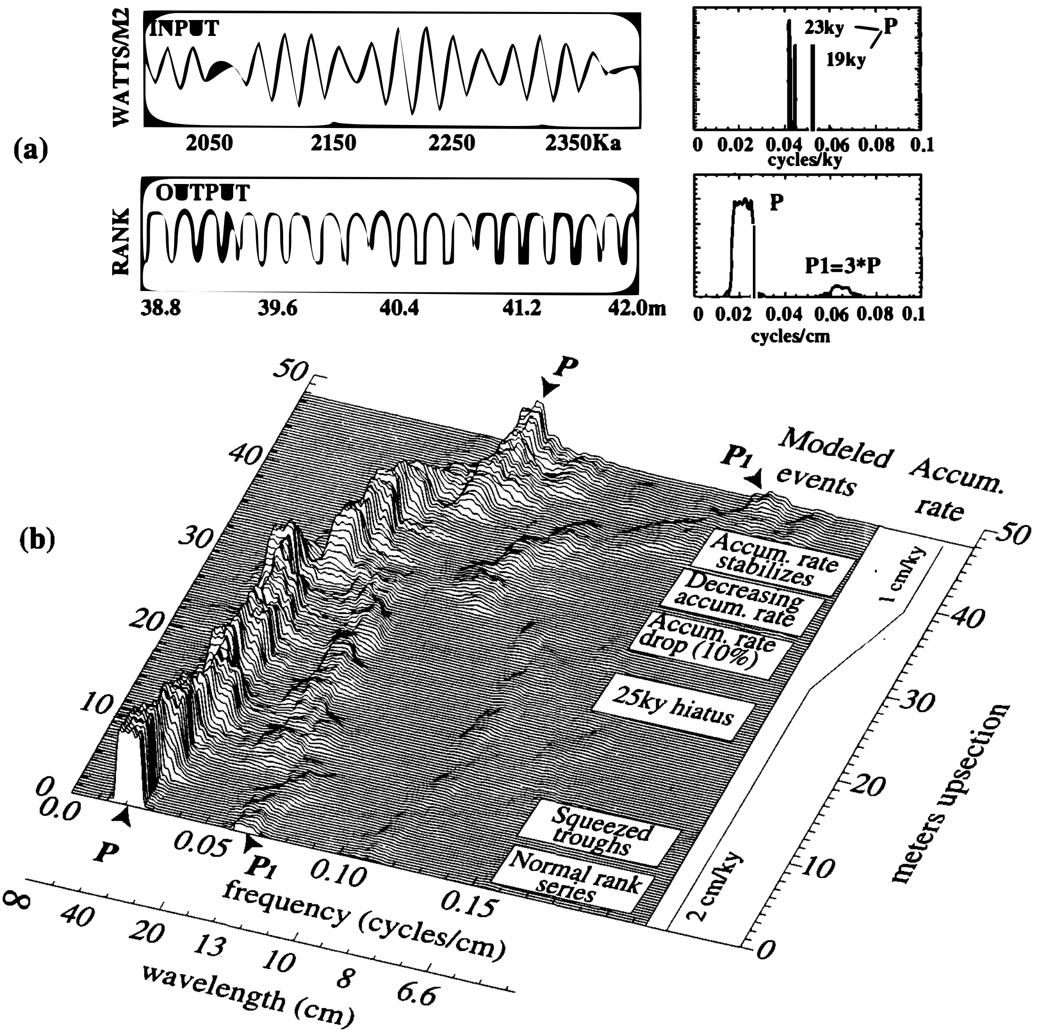
Time Series Modeling

Documentation of cyclic stratigraphic sequences many millions of years long, while desirable from a statistical point of view, is accompanied by practical problems that prevent the application of straight-forward time series analysis and interpretation. These include: (1) significant long term variations in accumulation rate, that can drastically alter the apparent frequencies of an otherwise stable cyclic forcing process; (2) changes in the character of the cycles, e.g., discontinuous lithologic changes versus more gradational cyclic lithologies as in the cyclic formations discussed here; (3) modifications in the relative influence of the orbital forcing parameters, such as the precession-to-obliquity frequency switch that occurs in the Middle Pliocene oxygen isotope curve, and rise of the so-called "Matayuma cycles" of Ruddiman et al. (1986); and, (4) the disruptive effects of multiple hiatuses or rapidly varying sedimentation rates.

To assess the significance of all these problems simultaneously we turned to evolutionary spectral analysis. This use of the evolutionary technique to evaluate the frequency behavior of two modelled, orbitally forced sedimentary sequences consisting of discontinuous values, or "ranks," to simulate the lithologically ranked series that we collected at Colle di Sogno is demonstrated in Figures 3 and 4. In the models we adopted accumulation rates comparable to those estimated at Colle di Sogno to make decimeter-scale lithologic rank cycles, and in the spectral analysis, we used a 4-meter window to obtain a frequency resolution comparable to that of the data analysis. We took into account the "boxcar" shape of a rank series, including simulations for compacted shales ("squeezed troughs"), a simple hiatus, and a variable accumulation rate, including a gradual decline to half the starting rate over 5 My of orbital forcing. The boxcar cycle shape results in spectra with overtones (harmonics); squeezing troughs results in moderate frequency shifts in the overtones. A hiatus has less of an effect on the local spectrum (producing a phase discontinuity only, resulting in a slight drop and local broadening of frequency) than does continued sedimentation through a rapid change in accumulation rate, where spectral distortion is strong over an entire spectral window (here, 4 meters). Of particular note, the models reflect the multicomponent nature of the precession signal with its "lumpy" and generally complicated structure (Fig. 3) versus the unimodal obliquity signal with its stable, 41ky^{-1} spectral peak and smooth vertical sides (Fig 4). The overtones dramatize these characteristics due to their occurrence at higher frequencies where frequency resolution is greatly expanded.

Figure 3. A rank series model of precession forced limestone/shale couplet deposition, with accumulation rates ranging from 1-2cm/ky, recorded as a rank series with limestone=1, shale=0, resolved at Δd = 0.4cm sample spacings. (a) Input precession forced solar radiation (detail, upper left) has over the input 5 million year interval a multicomponent spectrum (upper right) with two main frequencies at 23ky^{-1} and 19ky^{-1}. The output limestone/shale rank series (detail, lower left) was produced by assigning the value of 1 whenever solar radiation was greater than the long term average solar radiation value, and 0 otherwise. The power spectrum (using the algorithm of Thomson (1982), with 2π prolate tapers) for the first 4 meters (lower right) is with a high accumulation rate at 2cm/ky, and does not resolve the multicomponent precession. P = fundamental precession, and P1 = first overtone at 3 times the fundamental period, a reflection of the Fourier expansion of a rectangular function. (b) Evolutionary spectral analysis, using 4 meter windows and a 0.3m increment and the same 2π prolate taper resolution, along the 5 million year model sequence; note the modeled events and evolution of the accumulation rates.

A PRECESSION MODEL



The Domaro Limestone

The 80 meter thick, dm-scale limestone/shale couplet sequence of the Domaro Limestone Lower Member (Fig. 2), represents at least the Lower Domerian and potentially also most of the underlying Carixian (cf. Hinnov and Park, 1998). The dm-scale micritic limestone members are well bioturbated, with pelagic faunas appearing throughout, including juvenile ammonites, protoglobigerinids, nannofossils, radiolaria and pelagic bivalves (*Bositra*). Abundances of sponge spicules collect into semi-continuous cm-scale layers, typically in the lower and upper thirds of the beds, but also occasionally bisecting the beds. The tops and bottoms of the micrite beds typically show increased compaction, with wispy seams of organic material and micro-stylolites. The source of the micrite is principally platform and periplatform ooze (Picotti and Cobianchi, 1996). The intercalated cm-scale shales are strongly compacted and contain illite, micas and small micritic lenses.

The couplet thickness sequence of the Domaro Lower Member has a frequency-modulation history with close affinities to that of the climatic precession (Hinnov and Park, 1998). This result was used to estimate the differential (post-compactional) accumulation rates for the limestone and shale members at 1.57 cm/ky and 0.138 cm/ky, respectively. These rates were used to convert the Domaro rank series from thickness to time. Figure 5 shows the time-frequency evolution of the reconstructed Domaro time series using a 300 ky window. The averaged power spectrum over the first 400ky of this series appears as Fig. 16A in Hinnov and Park (1998), where the two dominant spectral peaks at the 25ky and 18.2ky periods are interpreted as strongly indicative of precession forcing. We have also indicated the possible presence of a minor obliquity component, which occurs at 35ky, close to the predicted 37ky Jurassic obliquity period, although Hinnov and Park (1998) argued that this component is just as likely to have been caused by randomness in the sedimentation process. The spectra also show significant intervals during which the dominant peaks either shift or are replaced by low frequency components (e.g., in the 0.8-1.5 My interval). These shifts can be explained as the result of slowly varying deviations of the true accumulation rates from the modeled rates. The deviations reverse every ~0.5 My and are likely a response to the 400ky component of the eccentricity modulator (of the precession).

The Sogno Formation

The Toarcian-Early Bajocian Sogno Formation is a *Posidonia alpina* facies partly coeval with the Rosso Ammonitico Lombardo elsewhere in Lombardy, and at Colle di Sogno lies conformably over the Pliensbachian Domaro Limestone (Fig. 2). The formation is representative of the sudden and dramatic change in sedimentation that is observed throughout the province at this time. The nearly 140m thick formation is spectacularly cyclic with clear dm-scale limestone/marl alternations persisting over 95% of the section (the other 5% is the Lower Toarcian black shale, which is mostly covered). Unlike the sharp limestone/shale alternations of the underlying Domaro couplets, the Sogno cycles are thicker (averaging 30-50cm), and are characterized by gradational changes between limestone and marl lithologies; radiolarian chert becomes a significant component at the base of the Aalenian. Near the top of the formation (L-3 Zone, see Fig. 2), the cyclicity deteriorates, stratal thicknesses decrease by almost 50%, and carbonate sedimentation diminishes to effectively zero in the Lower Bajocian.

In the field, we recognized 8 consistent lithologies which we have used as "ranks" in a high resolution lithologic series reconstruction of the section: shale (1), marly shale (2), shaly marl (3), marl (4), limey marl (5), marly limestone (6), limestone (7), and cherty limestone (8). It has been shown elsewhere (Elrick and Hinnov, 1996) that these types of field observations of basinal carbonate lithologies closely follow weight percent carbonate and, with the exception of the cherty limestone, we expect a similar result for this series. The highest ranking for the cherty limestone was intended to reflect the insertion of a second major biogenic component into the depositional system. Thus, the Sogno rank series is more precisely a proxy of non-biogenic (low rank) versus biogenic (high rank) sedimentary components, rather than simply non-carbonate vs. carbonate. We measured these 8 lithologies to the nearest millimeter over the entire section and reconstructed a rank series at 1mm spacings over 130m of section. Representative segments of this rank series are shown in Figs. 6a, 7a, and 8a.

The evolutionary spectral analysis of the Sogno rank series appears in Figs. 6b, 7b, and 8b, at a frequency resolution (4 meter windows) identical to the modelled sequence spectra (cf. Figs. 3 and 4). The division of the analysis into the three segments was dictated by the memory limits of our computer software. In Fig. 6b, at the base of the section, Meters 0-3 are "acyclic", dominated by

AN OBLIQUITY MODEL

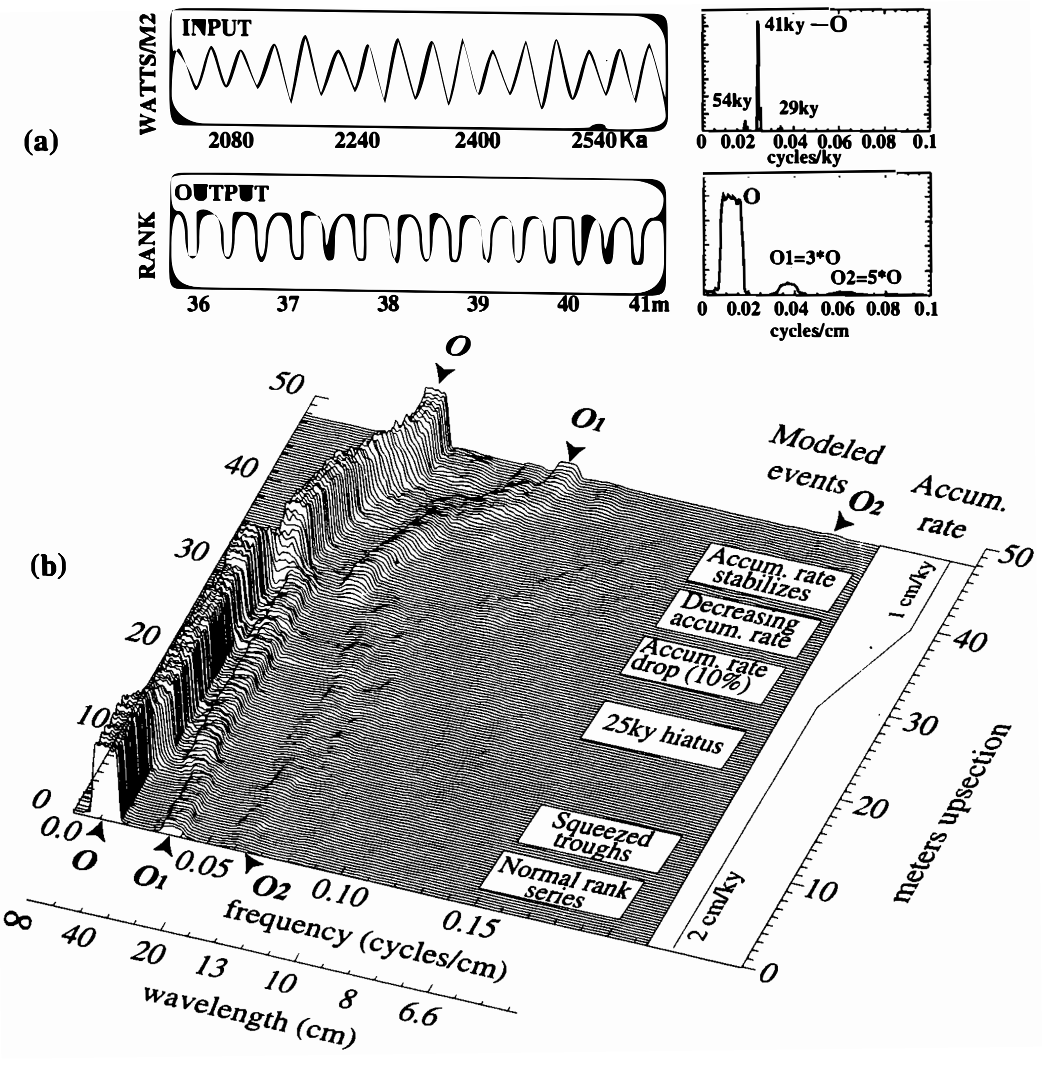
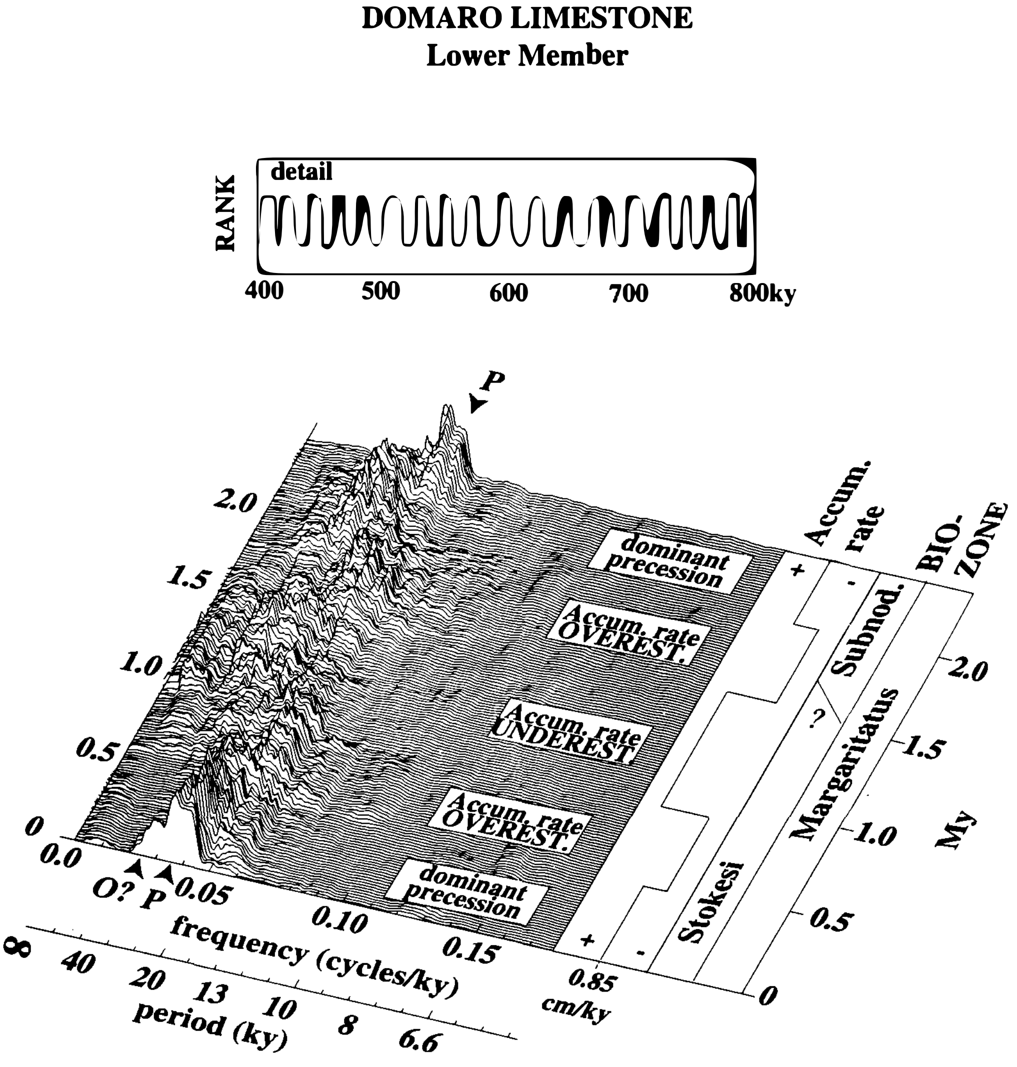


Figure 4. A rank series model of obliquity forced limestone/shale couplet deposition, with the same Δd, accumulation rates and modeled events as in Figure 3. (a) Input obliquity forced solar radiation (detail, upper left) has a major frequency component at 41ky^{-1} in the power spectrum of the input 5 million year interval (upper right); the output rank series (detail, lower left) has over the first 4 meters a spectrum (lower right) with a strong fundamental at O, and two visible overtones at odd multiples of the fundamental period. (b) Evolutionary spectral analysis as in Figure 3.



DOMARO LIMESTONE
Lower Member

Figure 5. Evolutionary spectral analysis of a time reconstruction of the Domaro Limestone sequence. The depth-time transformation as made using an accumulation rate of 1.57cm/ky for limestones, and a rate of 0.138 cm/ky for the (compacted) shales (see Hinnov and Park, 1998 for details), and was sampled at Δt=0.2ky. The window length was 300ky, and the increment was at 10ky, and the power spectra were estimated using the 2π prolate taper resolution as before (see Figure 3).

carbonate beds mixed with coarse siliciclastics. Subsequently, from Meters 3-8, carbonate cycles averaging ~30cm in thickness become established (labelled "P"). These cycles are attended by a ~5:1 bundling indicative of modulation by the eccentricity, which is detected here also as a low frequency, 1.42m cycle (labeled "E" for eccentricity). We have interpreted these together as evidence for PES-driven cycles. Our precession model (Fig. 3) does not simulate PES forcing with a dominant eccentricity response, but such a response has been simulated in 2-D energy balance climate models with Mesozoic paleogeography that respond to orbital forcing in tropical regions of both hemispheres simultaneously, but with different sensitivities (Crowley et al., 1992). Unfortunately, the succeeding black shale (Meters 8-16) is covered at Colle di Sogno by the road; however, a carbonate series sampled in the equivalent biozone in the Belluno Trough (just east of the Trento Platform, Fig. 1), suggests that there is combined precession+obliquity forcing, with PES components dominating the spectrum (Claps et al., 1995).

When the section resumes at Meter 16, the cycles are completely different: an apparent obliquity component briefly dominates the spectrum (Meters 16-18), then the spectral peaks shift dramatically to the left, indicating greatly increased accumulation rates (thicker cycles); the E component reappears and dominates over the next 8 meters (Meters 24-32). Between Meters 32-35, cyclicity gives way to thick carbonate bedding containing small (cm-scale) graded, resedimented turbidites. At Meter 35, frequency components are quickly re-established, with obliquity now the dominant mode. At Meter 52, accumulation rates decrease slightly (peaks move to the right), and "pure" obliquity forcing characterizes the remaining ~85 meters of the formation. The long term frequency modulations of the Sogno cycles have also been examined, and shown to contain "beat" components similar to those of the theoretical obliquity (cf. Hinnov and Park, In press). In Fig. 7b, the spectra exhibit a slow decline in accumulation rate, with the O peak drifting slowly toward higher frequencies (i.e., thinner cycles). From Meters 65-75, the O peak is disrupted due to rapidly shifting accumulation rates; at Meter 80 and again at Meter 90, approaching the Toarcian/Aalenian boundary, there is evidence for sudden reduction in accumulation rate (compare with Meter 28 of the obliquity model in Fig. 4). In Fig. 8b, accumulation rates stabilize from Meters 92-100, and the appearance of chert within the series, along with thinning shales, gives rise to a more prominent overtone at O2 and a new undertone at O-1. Finally, the cycles undergo a series of downstepping accumulation rates, the most severe occurring at Meter 115, coincident with the Type I sequence boundary (LSW) of Haq et al. (1988).

Discussion

The Lower-Middle Jurassic sequence at Colle di Sogno shows evidence for dramatic change, both in sedimentology and fossil content, and in cyclic style and frequency. Spiculitic chert and pelagic faunas occur in the Pliensbachian Domaro Limestone whereas high organic carbon, *Bositra*, and an upward increasing fraction of radiolarian chert characterize the overlying Toarcian-Aalenian Sogno Formation. Multicomponent precession frequencies dominate the ca. 15cm thick, highly etched micrite/shale Domaro couplets whereas the monochromatic obliquity signal dominates the overlying ca. 30cm thick, gradational micrite/marl Sogno cycles. These distinctive sedimentation styles are separated by a disrupted interval in the Upper Domerian (Fig. 2), with synsedimentary slumping and deposition of bioclastic carbonates from platform environments. This disruption is the local expression of well-documented syn-rift tectonics that occurred throughout the province during the latest Pliensbachian - earliest Toarcian (Gaetani, 1975; Winterer and Bosellini, 1981; Sarti et al., 1992; Bertotti et al., 1993).

Any explanation for the fundamental change in cyclic sedimentation response from precession to obliquity forcing (Fig. 9) requires information about the source of the cyclic micritic limestone. This is because the theory of orbital forcing predicts that precession forcing dominates in tropical zones, whereas obliquity forcing is restricted to the polar regions (Berger et al., 1993). Thus, to account for the rise to prominence of an obliquity signal in tropical marine sedimentation requires a global transferral of climate forcing from polar to tropical seas. Below we discuss the possible presence of small polar glaciations, or changes in continental water storage in the middle to high latitudes of both hemispheres, to produce meter-scale sea-level oscillations with a pure obliquity timing.

The problem of the origin of the micrite in the Lower-Middle Jurassic basinal sequences in western Tethys has still not been resolved: is it principally platform-derived carbonate mud or pelagic (nannofossil) ooze? Kälin (1980) argued that nannofossils account for only a fraction of the total volume of these micrites. This was later corroborated for the Lombard Basin by Picotti and

Cobianchi (1996) who, for the Pliensbachian-Toarcian interval, counted only ~2% total abundance of nannofossils in the Generoso Basin, and an order of magnitude less in a coeval section from the eastern Lombard Basin where the nannofossil content was apparently diluted by carbonate mud from the adjacent Venetian Platform. Quantitative assessments of sections from other Early Jurassic basins show similar trends (Lozar, 1992; Bombardiere, 1993; Mattioli, 1993).

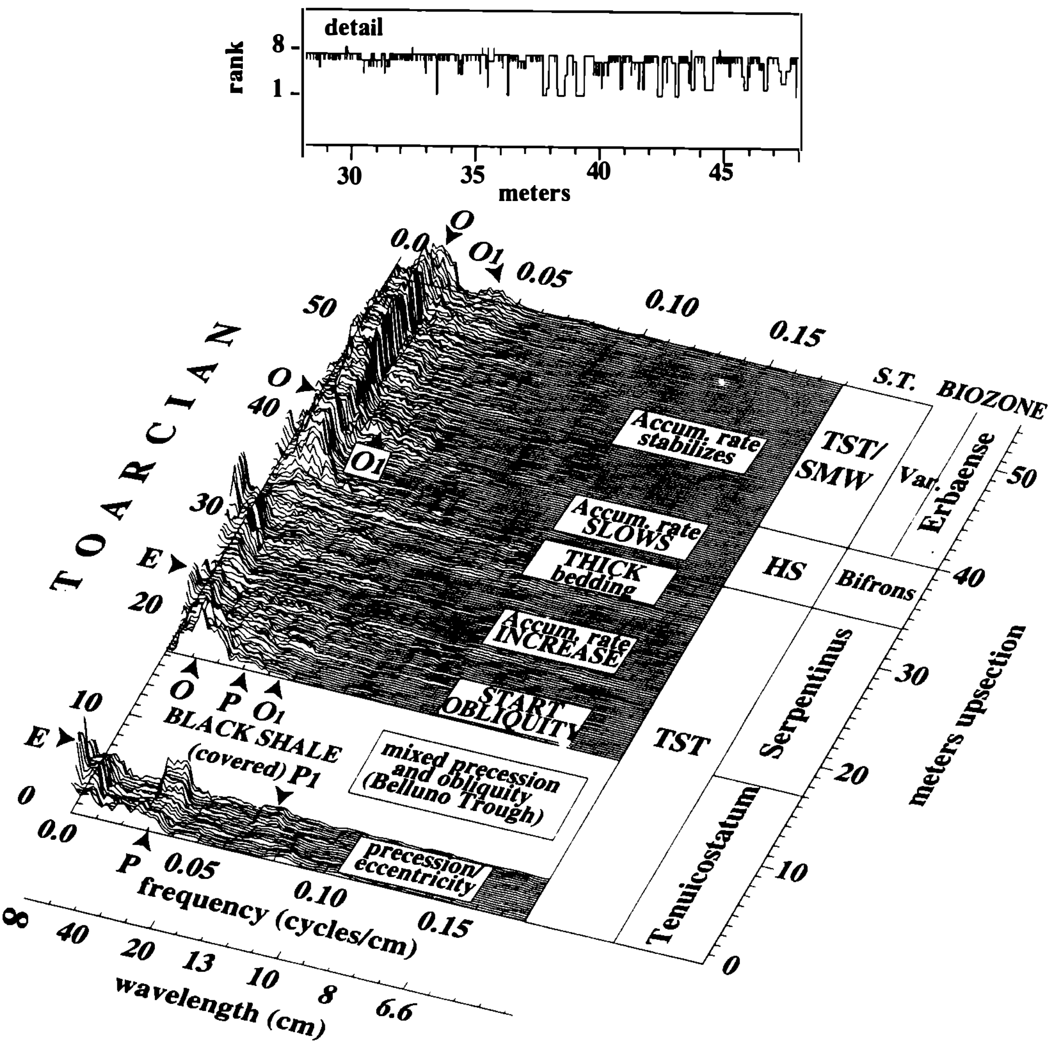

SOGNO FORMATION, 0-55m

Figure 6. Evolutionary spectral analysis of the rank depth series of the Sogno Formation, sampled at Δd = 0.1cm intervals, from 0-55m. The window length was 4 meters, the 2π power spectra were taken at 20cm increments up the section. Note that the depicted detail (top) spans the interval (30-45m) during which the obliquity signal rises to prominence. S.T. = sequence tracts proposed by Haq et al. (1988); biozones are from Figure 2.

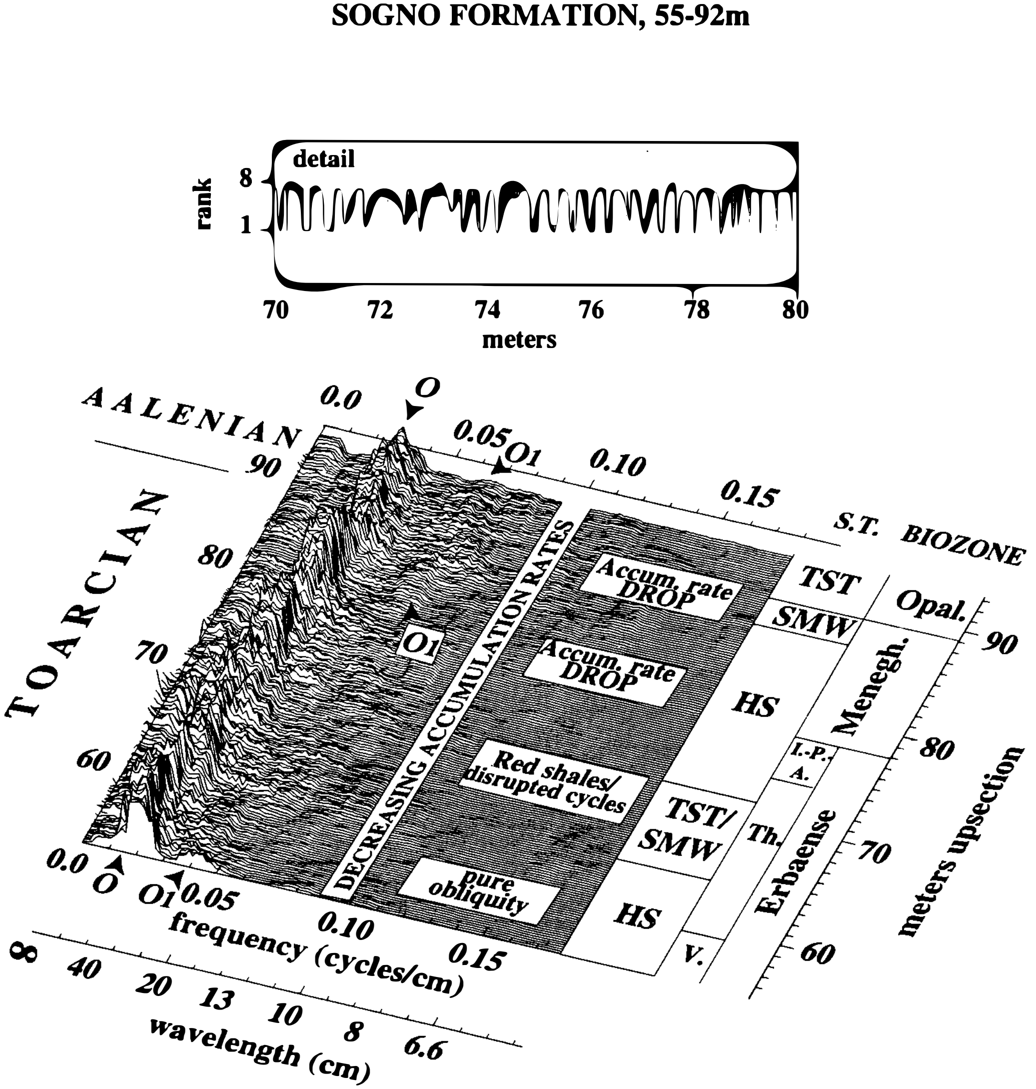


Figure 7. Evolutionary spectral analysis of the Sogno rank depth series from 55-92m. Computation is the same as in Figure 6.

Claps et al. (1995), on the other hand, argued that Sr/Ca data from Lower Toarcian carbonates in the Belluno Trough (Fig. 1) indicate that the primary carbonate phase was calcitic, and not aragonitic as might be expected for allodapic platform mud (although the transition from aragonite to calcite seas appears to have taken place prior to that time (Hardie, 1996)).

The importance of identifying the primary micrite source is that it can help explain the evolution of the strong unimodal obliquity signal in the Sogno cycles. If the Sogno carbonate was indeed derived from a platform, then sea level oscillations caused by small polar glaciations (optimally bipolar, as proposed Fischer and Hinnov, 1997 to explain the pure obliquity forced sea level variations of the Late Pliocene) would provide one possible explanation for the signal. In fact, high frequency sea level oscillations with <5m amplitude are thought to have affected at least some of the cyclic Pliensbachian carbonate platforms in western Tethys (Crevello, 1991; Claps, 1993;

Masetti et al., 1996). Further, Hinnov and Park (1998) called on precession-forced sea level changes to explain the frequency characteristics of the Pliensbachian Domaro Limestone: precession forcing related to transgressing or high sea levels was shown to correspond uniquely with carbonate deposition, i.e., submerged, highly productive platforms with high potential for basinal transport of mud, whereas precession forcing related to receding or low sea levels were shown to correspond uniquely with residual shale deposition, with a reduced platform carbonate factory. The source of allodapic mud to the Colle di Sogno site was most likely the Trento Platform to the east (Fig. 1).

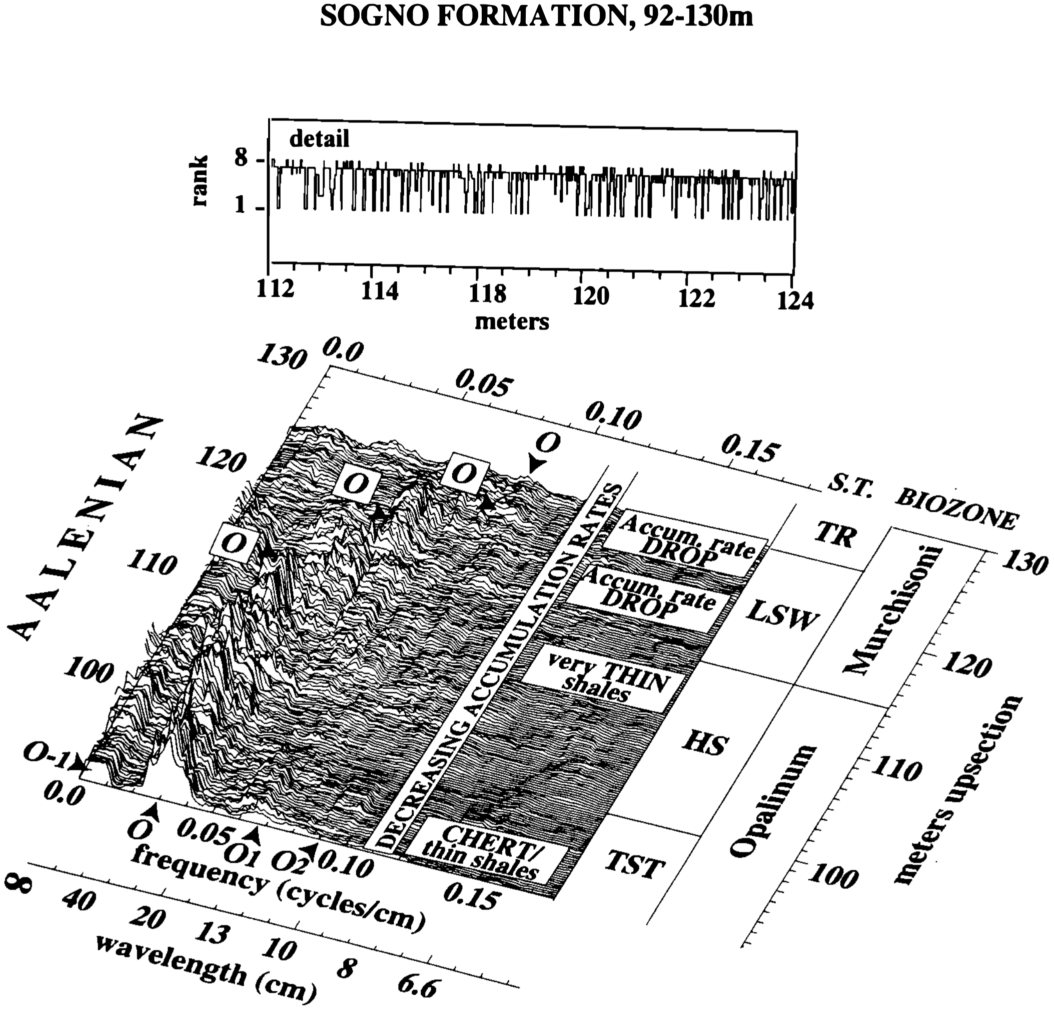

SOGNO FORMATION, 92-130m

Figure 8. Evolutionary spectral analysis of the Sogno rank depth series from 92-130m. Computation is the same as in Figures 6 and 7.

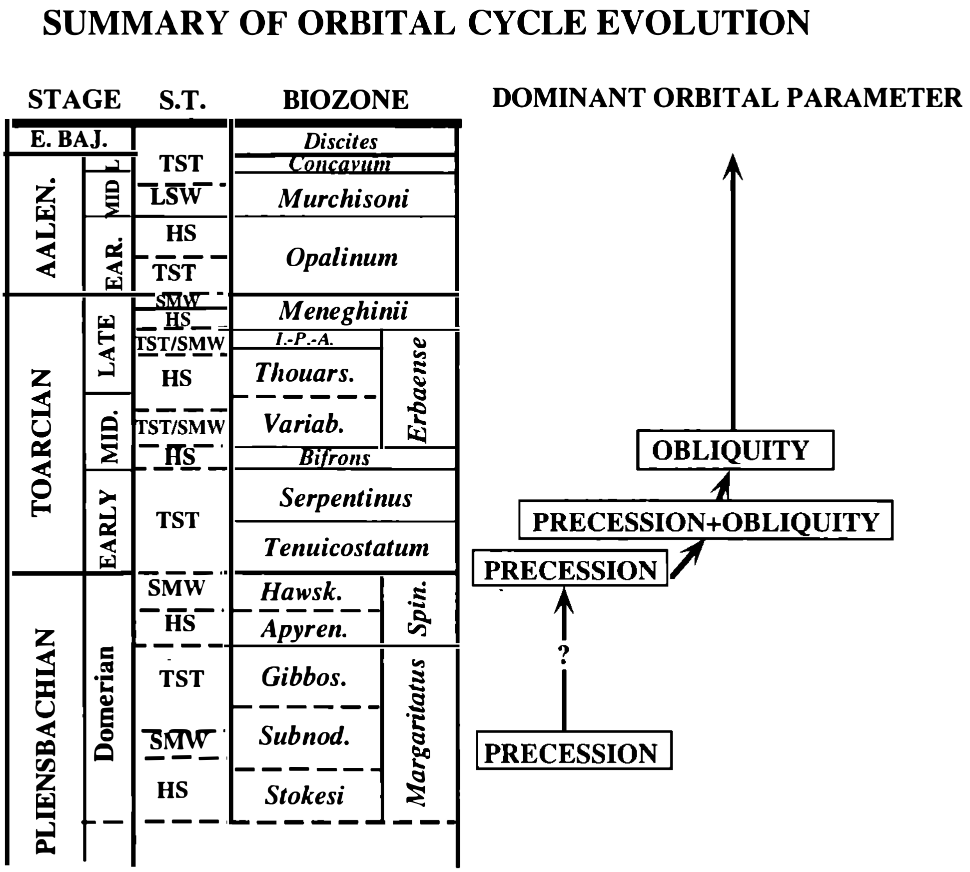


Figure 9. Evolution of orbital forcing of the rhythmites at Colle di Sogno, inferred by the spectral analyses in Figures 5-8.

To explain a switch from precession forced eustasy (indicated in the Pliensbachian Domaro) to a mixed precession and obliquity forced eustasy (the Early Toarcian) to a pure obliquity forced eustasy (Middle Toarcian), we suggest (Fig. 10) that during the Pliensbachian, orbitally controlled water storage occurred in the low to middle latitudes of one hemisphere, either in the form of ice in high mountains or as lakes and groundwater (Jacobs and Sahagian, 1995), varying in tune with the precession. During the Toarcian, either the effective water storage moved poleward (Fig. 10, Path A), or significant water storage began to grow at an opposing latitude in the other hemisphere (Fig. 10, Path B), in either case suppressing the counterphasing precession variation in the summed eustasy, and at the same time reinforcing the interhemispheric phasing of the obliquity variation. In this regard, there is palynological evidence from the southern former U.S.S.R. for widespread development of humid conditions (disappearance of the aridity proxy *Classopolis*) in Laurasia during the Late Toarcian (Vakhrameev, 1991), indicating that significant changes were occurring in water transport into the continental interior of at least one hemisphere. There are also reports of glacial deposits from the Aalenian of Siberia (see summary in Frakes et al., 1992), and while evidence for glaciation in Toarcian Antarctica has not been sought, the potential for an ice sheet there may have existed, as it may have also for the Late Jurassic as modelled by Valdes et al. (1995), when obliquity forcing also appears to have been the dominant mode in marine carbonate cyclicity (Weedon et al., In press). There have also been reports that marine temperatures dropped during the Toarcian-Bajocian interval, and in a review of these studies and a compilation of new marine stable isotope data (Jones et al., 1994), Podlaha et al. (1998) showed that mean $\delta^{18}O$ values underwent a steep increase of ~4‰ following the Early Toarcian anoxic event, coincident with the switch from precession to obliquity forcing observed at Colle di Sogno. This steep increase reflects either a decrease in marine temperatures, progressive removal of O^{16} from the ocean reservoir, or a combination of both. Jenkyns and Clayton (1997) speculated that significant CO_2 drawdown could have occurred as a result of the productivity pulse in the *Serpentinus* Zone, resulting in global cooling (and possible development of polar glaciation).

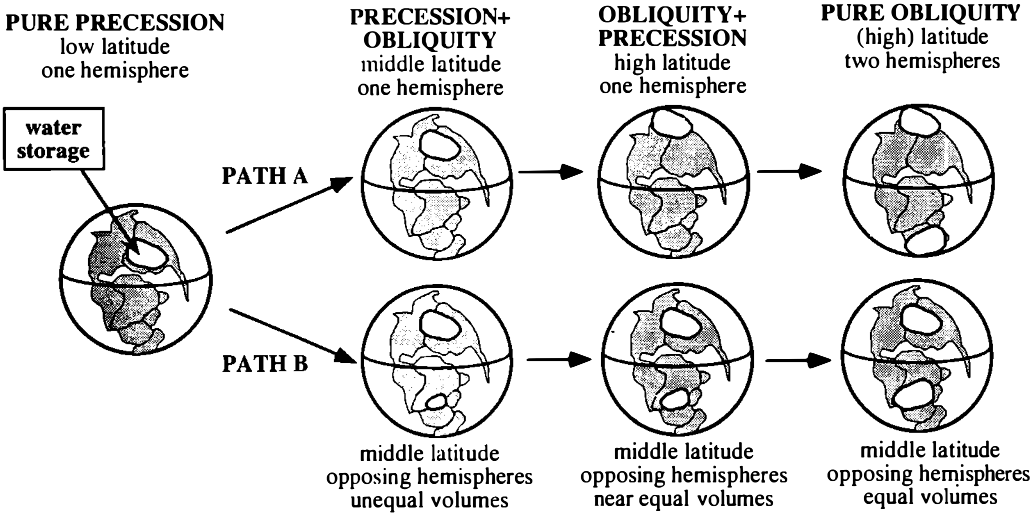


Figure 10. Proposed evolution of water storage on continents that could explain the spectral pattern in Figure 9 (from Hinnov and Park, In press). The idea is that water sequestration on the continents is forced by the local, orbitally forced, solar radiation; ocean volume, viz. high frequency eustasy, is a function of the entire continental water budget. To produce the frequency evolution of the Colle di Sogno cycles, shifts, and in particular, the development of significant reservoirs in both hemispheres, is essential in suppressing the precession, and enhancing the obliquity, as is seen in the Sogno Formation. See text for more discussion on PATH A and PATH B.

Long term modulations in the Sogno cycle spectrum (Figs. 6-8) indicate slowly varying accumulation rates that appear to follow the third order eustatic changes proposed by Haq et al. (1988). The Early Toarcian transgression is attended by cycle components moving rapidly and dramatically to much lower frequencies (i.e., thicker bedding), signalling increasing accumulation rates; this trend intensifies throughout the *Serpentinus* Zone a time also of elevated marine $\delta^{13}C$ (Jenkyns, 1988), then levels off and reverses during the subsequent highstand, and stabilizes with the rise of pure obliquity forcing in the *Bifrons* Zone. The next two third order eustatic cycles (encompassing the *Erbaense* Zone) do not appear to affect accumulation rates, but at the Toarcian/Aalenian boundary there is a rate drop coincident with evidence for a global sea level lowering (SMW). The subsequent transgression (*Opalinum* Zone) is reflected in a slight increase in rates, until at the subsequent late highstand, at Meter 115, the first of a series of rate drops occurs coincident with a major lowstand in the lower *Murchisoni* Zone (associated with ocean drawdown related to continental ice sheet growth?). These long term accumulation rate changes are consistent with reduced platform carbonate productivity (on the Trento Platform) during decreasing sea levels, and higher platform productivity during increasing sea levels, in a fashion analogous to how we envision basinal carbonate accumulation occurred in tune with the superimposed high frequency, orbitally forced, sea level oscillations. The total cessation of carbonate input to the Albenza Trough by Early Bajocian time, however, was probably due to causes other than just third order eustasy. Submergence of the Trento Platform during the Late Aalenian was attended by eutrophication, shutdown of carbonate production, and the ultimate death of the platform, which drowned by the end of the Bajocian (Zempolich, 1993, his Fig. 16).

Conclusions

The Upper Pliensbachian-Lower Bajocian rhythmite sequence at Colle di Sogno in the Lombardian Alps had a remarkable evolution over some 15 million years of basinal cyclic carbonate deposition. The ca. 15cm thick spiculitic limestone/highly compacted shale couplets of the Lower Domerian Domaro Limestone reveal a spectrum consistent with a history of multicomponent precession forcing with low frequency accumulation rate changes in tune with the $400ky^{-1}$ component of eccentricity. Following an interval of syndepositional tectonics and disrupted stratigraphy, the cyclicity resumed in the Lower Toarcian Sogno Formation, this time with a somewhat higher accumulation rate and diluted by siliciclastics, and gaining a prominent $\sim 100ky^{-1}$ eccentricity component; these elements together suggest a significant shift in the global climatic response to orbital forcing. Some time during the deposition of the Lower Toarcian black shale, an obliquity signal appeared in the cyclicity; this intensified in the *Serpentinus* Zone, and eventually replaced the precession signal altogether by the start of the Middle Toarcian (*Bifrons* Zone). This transition occurred at a time when marine $\delta^{18}O$ was rapidly increasing, signifying either a drop in ocean temperature or a progressive reduction in ocean volume, or some combination of the two. Obliquity forcing continued to dominate over the remaining 100 meters of carbonate cycles, with weak modulations occurring in response to the long term eustatic cycles proposed by Haq et al. (1988).

This transition from precession to obliquity orbital forcing is reminiscent of the development of pure obliquity forced, global sea level oscillations during the Middle Pliocene (e.g., Ruddiman et al., 1986), which coincided with the initiation of glaciation in the Northern Hemisphere, counterphased ice volumes at the poles at the precession frequencies, and reinforced ice volumes at the obliquity frequency (Fischer and Hinnov, 1997). Investigation of the provenance of the Domaro and Sogno cycles strongly suggests that platform derived mud was the dominant source of these basinal carbonates, and that the cyclicity was caused by sea level oscillations controlling productivity on the platforms. Changes in the continental water budget, particularly the latitudes of sequestration, are suggested as the cause of changes in frequency content observed in the cycles.

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References

BAUMGARTNER, P. O. 1987. Age and genesis of Tethyan Jurassic Radiolarites. Eclogae Geologicae Helvetiae, 80:831-879.

BERGER, A., M.-F. LOUTRE, AND C. TRICOT. 1993. Insolation and Earth's orbital periods. Journal of Geophysical Research, 98:10341-10362.

BERTOTTI, G., V. PICOTTI, D. BERNOUILLI, AND A. CASTELLARIN. 1993. From rifting to drifting: tectonic evolution of the South-Alpine upper crust from the Triassic to the Early Cretaceous. Sedimentary Geology, 86:53-76.

BOMBARDIERE, L. 1993. Analisi sedimentologica ed ultrastrutturale del fango carbonatico della Corniola dei massicci pergugini (Umbria occidentale) e di alcune sezioni dell'area umbro-marchigiana. Paleopelagos, 3:113-129.

CLAPS, M. 1993. Ciclicità ad alta frequenza nella piattaforma carbonitica dei Calcari Grigi (Giurassico Inferiore, Prealpi Venete): evidence di controllo eustatico, Dottorato di Ricerca in Scienze della Terra, Universita di Ferrara, Italy.

____, E. ERBA, D. MASETTI, AND F. MELCHIORRI. 1995. Milankovitch-type cycles recorded in Toarcian black shales from the Belluno Trough (Southern Alps, Italy). Memorie Scienze Geologiche, 47:179-188.

CREVELLO, P. D. 1991. High-frequency carbonate cycles and stacking patterns: interplay of orbital forcing and subsidence on Lower Jurassic rift platforms, High Atlas, Morocco, p. 207-230. *In* E. V. Franseen, W. L. Watney, C.G. St. C. Kendall, and W. Ross (eds.), Sedimentary Modeling: Computer Simulations and Methods for Improved Parameter Definition, Kansas Geological Survey Bulletin No. 233.

CROWLEY, T. J., K.-Y. KIM, J. G. MENGEL, AND D. A. SHORT. 1992. Modeling 100,000 year climate fluctuations in pre-Pleistocene time series. Science, 255:705-707.

ELRICK, M., AND L. A. HINNOV. 1996. Millennial-scale climate origins for stratification in Cambrian and Devonian deep-water rhythmites, western USA. Palaeogeography, Palaeoclimatology, Palaeoecology, 123:353-372.

FISCHER A. G. AND L. A. HINNOV. 1997. Orbitally forced glaciations during greenhouse and icehouse times, Part 1: a global model. Geological Society of America, Abstracts with Programs, A-211.

FRAKES, L. A., J. E. FRANCIS, AND J. I. SYKTUS. 1992. Climate Modes of the Phanerozoic, Cambridge University Press, 274 p.

GAETANI, M. 1975. Jurassic stratigraphy of the Southern Alps: a review, P. 377-402 *In* C. Squyres (ed.), The Geology of Italy, Libyan Arab Repubic of Earth Sciences.

____ AND G. POLIANI. 1978. Il Toarciano et il Giurassico medio in Albenza (Bergamo), Rivista Italiana Paleontologia e Stratigrafia, 84:349-382.

GRADSTEIN, F. M., F. P. AGTERBERG, J. G. OGG, J. HARDENBOL, P. VAN VEEN, J. THIERRY, AND Z. HUANG. 1995. A Triassic, Jurassic and Cretaceous time scale, p. 95-126 *In* W. A. Berggren, D. V. Kent, M. P. Aubry, and J. Hardenbol (eds.), Geochronology, Time Scales and Global Stratigraphic Correlation, SEPM Special Publication, 54.

HAQ, B. U., J. HARDENBOL, AND P. R. VAIL. 1988. Mesozoic and Cenozoic chronostratigraphy and eustatic cycles, p 71-108 *In* C.K. Wilgus, B.S. Hastings, C.G. St. C. Kendall, H.W. Posamentier, C.A. Ross, and J.C. Van Wagoner. Sea-level Changes: An Integrated Approach, SEPM Special Publication, 42.

HARDIE, L. A. 1996. Secular variation in seawater chemistry: an explanation for the coupled secular varation in the mineralogies of marine limestones and potash evaporates over the past 600 m.y. Geology, 24:279-283.

HINNOV, L. A. AND A. G. FISCHER 1997. Orbitally forced glaciations during greenhouse and icehouse times, Part 1: the stratigraphic evidence. Geological Society of America, Abstracts with Programs, A-211.

____ AND J. PARK. In press. Strategies for assessing Early-Middle (Pliensbachian-Aalenian) Jurassic cyclochronologies. Philosophical Transactions of the Royal Society of London, Series A.

____ AND ____. 1998. Detection of astronomical cycles in the sedimentary record using frequency modulation analysis. Journal of Sedimentary Research, 68:325-338.

____ AND ____. 1995. Modelling Mesozoic precession-forced cyclic stratigraphy using frequency modulation analysis. Eos, 76:186.

____, ____, AND L.A. Hardie. 1994. A precession-to-obliquity climate cycle transition in Middle Jurassic Tethys: carbonate harbingers of a global cool mode? Eos, 75:205.

JACOBS, D. K. AND D. L. SAHAGIAN. 1995. Milankovitch fluctuations in sea level and recent trends in sea-level change: ice may not always be the answer, p. 329-366 *In* B. Haq (ed.), Sequence Stratigraphy and Depositional Responses to Eustatic, Tectonic and Climatic Forcing, Kluwer, Dordrecht.

JANSA, L. 1991. Processes affecting paleogeography, with examples from the Tethys. Palaeogeography, Palaeoclimatology, Palaeoecology, 87:345-371.

JENKYNS, H. C. 1988. The early Toarcian (Jurassic) anoxic event: stratigraphic, sedimentary and geochemical evidence. American Journal of Science, 288:101-151.

____ AND C. J. CLAYTON. 1997. Lower Jurassic epicontinental carbonates and mudstones from England and Wales: chemostratigraphic signals and the early Toarcian anoxic event. Sedimentology, 44:687-706.

JONES, C. E., H. C. JENKYNS, A. L. COE, AND S. P. HESSELBO. 1994. The strontium isotopic composition of Early Jurassic seawater. Geochimica et Cosmochimica Acta, 58:1285-1301.

KÄLIN, O. 1980. *Schizosphaerella punctulata* deflandre and dangeard: wall ultrastructure and preservation in deeper-water carbonate sediments of the Tethyan Jurassic. Eclogae Geologicæ Helvetiae, 73:983-1008.

LOZAR, F. 1992. Biostratigrafia a nannofossili calcarei nel Lias: risultati di alcune sezioni nel Bacino Lombardo e nel Bacino Delfinese. Paleopelagos, 2:89-98.

MASETTI, D., M. CLAPS, M. AVANZINI, A. GIACOMETTI, AND P. PIGNATTI. 1996. I Calcari Grigi della Piattaforma di Trento. 78a Riunione Estiva, Geologia delle Dolomiti, Società Geologica Italiana, San Cassiano, Italy, 48pp.

MATTIOLI, E. 1993. Quantitative analysis of calcareous nannofossils from Pozzale section (Martani Mts., Central Italy): preliminary report. Paleopelagos, 3:261-278.

PÁLFY, J., R. R. PARRISH, AND P. L. SMITH. 1997. A U-Pb age from the Toarcian (Lower Jurassic) and its use for time scale calibration through error analysis of biochronologic dating. Earth and Planetary Science Letters, 146:659-675

PICOTTI V., AND M. COBIANCHI. 1996. Jurassic periplatform sequences of the eastern Lombardian Basin (southern Alps): the deep-sea record of the tectonic evolution, growth and demise history of a carbonate platform. Memorie di Scienze Geologiche, 48:171-219.

PODLAHA, O. G., J. MUTTERLOSE, AND J. VEIZER. 1998. Preservation of $\delta^{18}O$ and $\delta^{13}C$ in belemnite rostra from the Jurassic/Early Cretaceous successions. American Journal of Science, 298:324-347.

RUDDIMAN W. F., RAYMO, M., AND MCINTYRE, A. 1986. Matuyama 41 000-year cycles: North Atlantic Ocean and Northern Hemisphere ice sheets. Earth and Planetary Science Letters, 80:117-129.

SARTI, M., A. BOSELLINI, AND E. WINTERER. 1992. Basin geometry and architecture of a Tethyan passive margin, Southern Alps, Italy: implications for rifting mechanisms, p. 241-258 *In* J. S. Watkins, Z. Feng, and K. J. McMillen-(eds.), Geology and Geophysics of Continental Margins, AAPG Memoir No. 53.

THOMSON, D. 1982. Spectrum estimation and harmonic analyis. IEEE Proceedings, 70:1055-1096.

VAKHRAMEEV, V. A. 1991. Jurassic and Cretaceous Floras and Climates of the Earth, Cambridge University Press, Cambridge, 318 p.

VALDES, P. J., B. W. SELLWOOD, AND G. D. PRICE. 1995. Modelling Late Jurassic Milankovitch climate variations, p. 115-132 *In* M. R. House, and A. S. Gale (eds.), Orbital Forcing Timescales Cyclostratigraphy, Geological Society Special Publication No. 85.

WEEDON, G. P. 1989. The detection and illustration of regular sedimentary cycles using Walsh power spectra and filtering, with examples from the Lias of Switzerland. Journal of the Geological Society, London, 146:133-144.

____, H. JENKYNS, A. COE, AND S. HESSELBO. In press. Progress on an astronomically calibrated time scale for the Jurassic, Philosophical Transactions of the Royal Society of London, Series A.

WINTERER, E. AND A. BOSELLINI. 1981. Subsidence and sedimentation on a Jurassic passive continental margin, Southern Alps, Italy. American Association of Petroleum Geologists Bulletin, 65:394-421.

ZEMPOLICH, W. G. 1993. The drowning succession in Jurassic carbonates of the Venetian Alps, Italy: a record of supercontinent breakup, gradual eustatic rise, and eutrophication of shallow-water environments, p. 63-105. *In* R. G. Loucks and J. F. Sarg (eds.), Carbonate Sequence Stratigraphy: Recent Developments and Applications, American Association of Petroleum Geologists Memoir 57:63-105.

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Benthic Foraminiferal Response to Pliensbachian - Toarcian (Lower Jurassic) Sea-Level Change and Oceanic Anoxia in NW Europe

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Keywords: Mass Extinction, Survival Strategies, Micropalaeontology, Palaeobathymetry

Abstract: The Early Jurassic was a time of rapidly rising sea level associated with the extensive spread of anoxic bottom waters. Sea-level rise across the Pliensbachian - Toarcian boundary culminated in the *falciferum* zone of the Toarcian and is one of the best authenticated eustatic events in the Jurassic. This major transgressive pulse was marked by deeper water marine sequences and was associated with the deposition of organic-rich shales.

The development of anoxia coincides with a notable period of mass extinction of the marine fauna. High resolution sampling and study of the microfaunas through several sequences in the United Kingdom confirm that benthic foraminiferal faunas were similarly affected by an Early Toarcian falciferum zone event. Samples were analysed from Pliensbachian - Toarcian mudstone, clay and shale sequences of the South Dorset Coast, the Midlands and the Yorkshire Coast. The sections studied show distinct changes in assemblages across the Pliensbachian - Toarcian boundary and in the basal zones of the Toarcian.

Evidence for a foraminiferal extinction event in the Pliensbachian - Toarcian includes the elimination of the important Lower Jurassic *Lingulina tenera, Frondicularia terquemi* and *Marginulina prima* plexus groups, initiating a significant period of turnover of the microfauna. A marked change also occurred in the character of associated nodosariid assemblages: the uniserial forms of *Nodosaria, Frondicularia* and *Lingulina*, dominating the Pliensbachian assemblages, were largely replaced by coiled *Lenticulina* in the Early Toarcian. A reduction in test size and a decline in species diversity, compared with Hettangian to Sinemurian foraminiferal assemblages, reflect the development of low oxygen conditions followed by a subsequent renewal of the microfauna in the Middle Toarcian.

The foraminiferal data from the UK sections we have studied correlates with microfaunal responses reported from NW European Pliensbachian - Toarcian black shale sequences in SW Germany.

Introduction

Through their analysis of extinction periodicity, based on family-level data of marine extinctions, Raup and Sepkoski (1984; 1986) and Sepkoski and Raup (1986) claim to have recognised two events in the Jurassic, one of which was placed at the end of the Pliensbachian. This analysis, however, was at the stage level of resolution and the event was later reassessed as probably taking place in the Lower Toarcian, based on the recognition of a clear-cut extinction in Western Europe (Hallam, 1986; 1987).

Sea-level rise across the Pliensbachian - Toarcian boundary culminates in the *falciferum* zone of the Toarcian and is one of the best authenticated eustatic events in the Jurassic (Hallam, 1997). This major transgressive pulse is marked by apparent deepening in marine sequences in extensive parts of the world (Hallam, 1988) and is associated with widespread deposition of organic-rich shales. Jenkyns (1988) recognised that this period corresponds with high rates of carbon burial and a sharp positive excursion in the carbon isotope curve, signifying the Early Toarcian oceanic anoxic event.

This anoxic event coincided with a notable period of mass extinction of marine faunas. Sampling of invertebrate macrofauna through complete expanded sequences in NW Europe has shown that most species extinctions occurred in the Early Toarcian following a regional anoxic event (Little and Benton, 1995). The extinction event first recognised by Hallam (1961), led to an almost complete species turnover among the benthos in the *falciferum* zone.

Background to Early Jurassic Extinction Events

The most complete Lower Jurassic sequence studied for foraminifera in the U.K. was cored in the Llanbedr (Mochras Farm) borehole, north Wales, between 1966 and 1969 (Woodland, 1971). Johnson (1975) studied the foraminifera of the Upper Pliensbachian and Toarcian with the Hettangian to Lower Pliensbachian section being completed by Copestake (1978). These two works, although remaining largely unpublished, contain a considerable amount of information. Data relevant to this study (*margaritatus* to *variabilis* zones) has been re-examined and summarised in Figure 1.

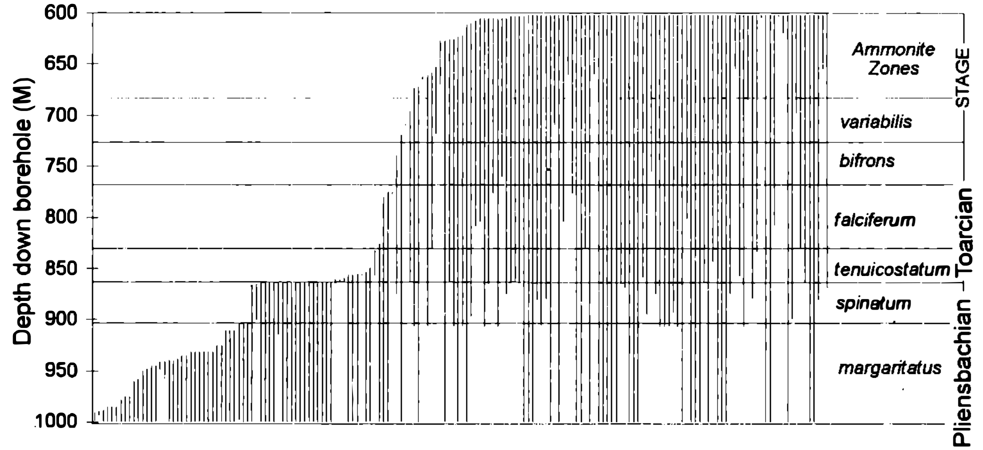

Figure 1. Benthic foraminiferal ranges from the Mochras Borehole, Wales, UK. Ranges compiled from Johnson (1975) and Copestake (1978).

The Mochras borehole foraminiferal fauna in Fig. 1 shows distinct changes in assemblages across the Pliensbachian - Toarcian boundary and during the basal zone of the Toarcian. During the *tenuicostatum* zone foraminifera show a decrease in diversity, as ecological stress developed and continued throughout the zone (Johnson, 1975). Deepening resulted in a reduction in bottom water circulation and this trend reached its acme in the basal *exaratum* subzone of the succeeding *falciferum* zone. The subsequent stratigraphic record of normal marine conditions is here absent due to faulting, although the missing section is entirely within the *exaratum* subzone (Johnson, 1975).

Brasier (1988) noted that while there is little available evidence for foraminiferal mass extinction events in the Lower Jurassic, certain groups were eliminated in boreal regions including the important Lower Jurassic *Lingulina tenera, Frondicularia terquemi* and *Marginulina prima* plexi. Also, there was a marked change in the chamber arrangement of younger nodosariid assemblages: uniserial *Nodosaria, Frondicularia* and *Lingulina* were largely replaced by coiled *Lenticulina* (Copestake and Johnson, 1989).

Sample Details

Figure 2 shows the outcrop distribution of the Lower Jurassic in England, together with the location of the sampled sections in England and SW Germany. The Lower Jurassic has a detailed ammonite zonation scheme (Cope et al., 1980) enabling the sections in this study to be zoned with reasonable accuracy. Figure 3 gives details the stratigraphic intervals sampled in NW Europe for this study.

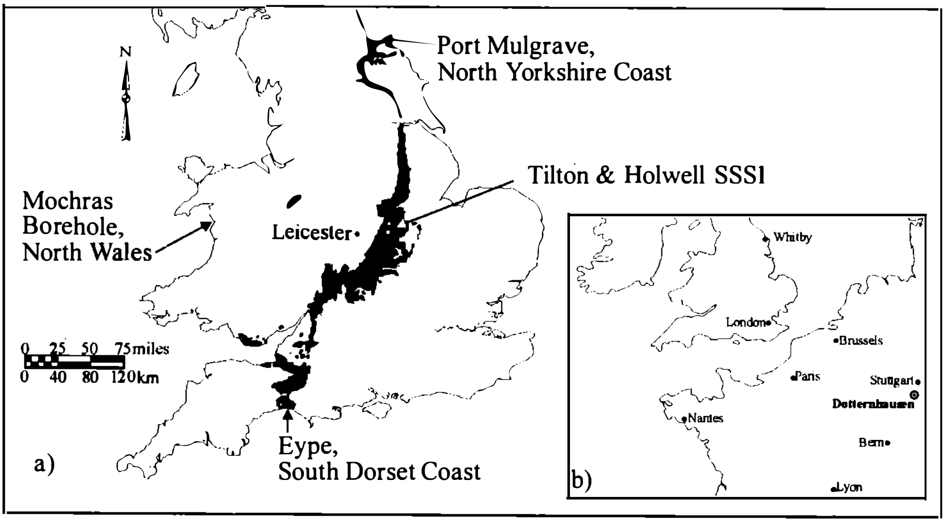

Figure 2. a) Lower Jurassic outcrop in England and location of sections sampled. b) Location of the Dotternhausen Oil Shale Quarry, Baden -Württenburg, S. W. Germany

Along the south Dorset coast, beds of Pliensbachian and Toarcian age are well exposed in accessible cliff sections which have recently been re-measured and described by Hesselbo and Jenkyns (1995). The oldest samples were taken from the Upper Pliensbachian Down Cliff Sands and the Thorncombe Sands. The next six ammonite zones across the Pliensbachian - Toarcian boundary are condensed into the Junction Bed and Marlstone, thin limestone units, which are represented in Yorkshire by 137m of sediment. Above the Junction Bed, Lower Toarcian material was sampled from the Down Cliff Clay.

The section studied at Tilton Railway Cutting in Leicestershire has been described by Hallam (1955). The lowest part of the succession exposes 3m of Upper Pliensbachian silts and clays and the characteristic development of oolitic ironstones of the Marlstone Rock Bed. The base of the Marlstone at Tilton is a non-sequence which represents the whole of the *gibbosus* Subzone, above which the Marlstone is made up of two lithological units: 3m of calcareous and ferruginous sandstones (Sandstone Member) overlain by 2.5m of the Ironstone Member, which is a very oolitic limestone. Howarth (1980) showed that the base of the Toarcian occurs within the Ironstone member, rather than at the 'Transition Bed' level. The Lower Toarcian is represented by 4m of mudstones and clays and lies within the *falciferum* zone (*exaratum* and *falciferum* subzones). A series of samples was taken from the mudstones, clays and shales above and below the Marlstone Rock Bed. North from Tilton a similar exposure is present at the Holwell SSSI. Consisting of two disused quarries conserved as part of a biological and geological reserve, the section exposes nearly 6m of Lower Toarcian laminated shales and mudstones above the Sandstone and Ironstone Members. A series of samples was taken from these sediments comprising the Toarcian *falciferum* zone.

The finely laminated or 'Paper' shales in the *falciferum* zone are analogous to, and roughly coeval with, the Posidonienschiefer of SW Germany and the Jet Rock of the North Yorkshire Coast (Fig. 3). A set of samples from the Whitby Mudstone and Cleveland Ironstone Formations has been provided by C. Little, collected from Port Mulgrave near Whitby. The Lower Toarcian Jet Rock and Bituminous shales of Yorkshire and their equivalents in the North Sea, France (Schistes Cartons), Germany and Switzerland (Posidonienschiefer) are all indicative of periods of deposition of organic-rich shales during the *falciferum* zone (Jenkyns, 1988). The Pliensbachian - Toarcian section sampled in Baden-Württemburg, southwest Germany is thinner and more calcareous than those on the North Yorkshire coast, although it has similar faunal and facies characteristics (Little, 1996).

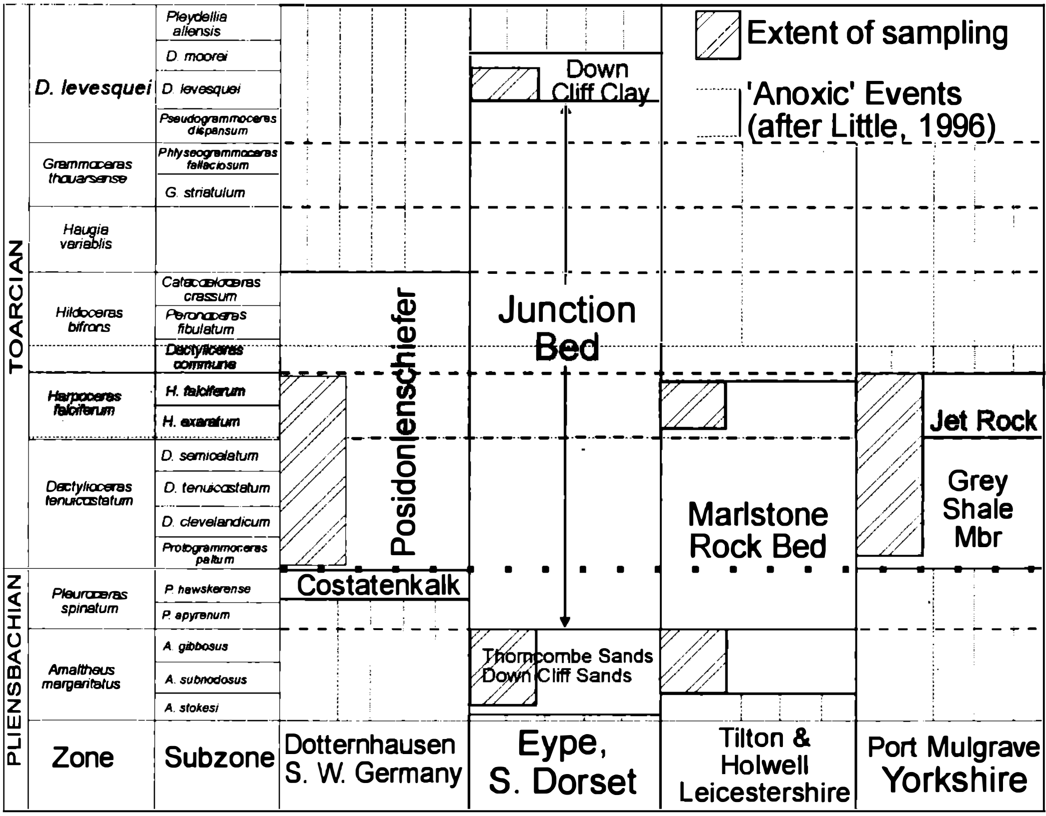

Figure 3. Stratigraphic intervals sampled. Ammonite zonation after Cope et al. (1980).

The main extinction event in south-west Germany occurs towards the top of the *semicelatum* Subzone which is coincident with the most organic-rich part of the Posidonienschiefer and an equivalent facies to the Jet Rock of the Whitby Mudstone Formation. The Posidonienschiefer is well developed in a series of quarries near Dotternhausen which have yielded a rich micro-fauna including foraminifera and ostracods (Riegraf, 1985). For this study, a set of pilot samples, from mid *tenuicostatum* Zone, to the base of the *bifrons* Zone was provided by Dr. Manfred Jäger of Rohrbach Zement.

Overview of Foraminiferal Groups in the Lower Jurassic

The Boreal Realm during the Early Jurassic was characterised by smaller benthic foraminifera. Normal marine shelf assemblages were dominated by nodosariids, with fewer numbers of Miliolina, Robertinidae, Spirillinina, Involutina, Buliminacea, Textulariina, Polymorphinidae and Cassidulinacea (Copestake and Johnson, 1989). Figure 4 illustrates six morphological groups of benthic foraminifera that are considered in this study; these are almost entirely nodosariids except for the Robertiniid genus *Reinholdella*. Within nodosariid-dominated populations, major variances in species and generic abundances are seen in taxa which are long-ranging, suggesting controlling palaeoenvironmental factors. In a statistical analysis of European Lias foraminiferal assemblages, Brouwer (1969) suggested that outer shelf environments were dominated by *Lenticulina* (including *L. muensteri* and *L. varians*), intermediate depths by the *Marginulina prima* plexus and shallower depths by the *Lingulina tenera* plexus.

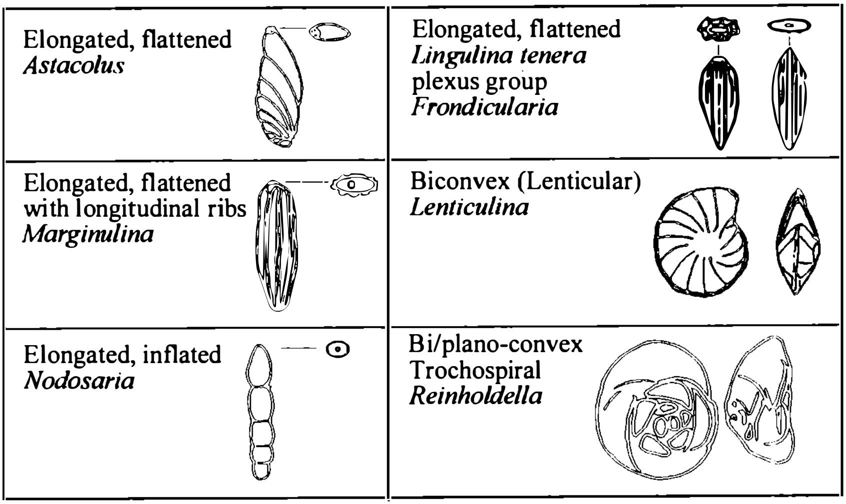

Figure 4. Palaeoenvironmentally significant morphologies in Lower Jurassic assemblages of NW Europe.

However, whilst Brouwer's (1969) analysis clearly indicated that distinctive assemblages exist within the overall nodosariid-dominated shelf areas, he had little direct evidence to support his environmental interpretations. This approach was pursued by Muller (1990) who also, through the use of statistical analysis of Lower Jurassic foraminifera, designated thirteen discrete foraminiferal 'biofacies' groups and proposed a model of Liassic benthic foraminiferal palaeobathymetry ranging from very near shore to outer shelf or shelf basins. Again, the *Lingulina tenera* plexus dominated in the inner to near-shore shelf environments while *Lenticulina* gained importance towards more open marine and deeper water environments. *Reinholdella macfadyeni* dominated deeper water areas.

Foraminiferal Response

The distribution of foraminiferal assemblages in comparison with other palaeoecological and sedimentological indicators allows construction of the model in Fig. 5. Palaeodepths of foraminiferal assemblages appear to concur with the eustatic curve proposed by Haq et al. (1987). In addition to the large scale changes in foraminiferal assemblages in response to changing sea-level and environmental conditions, several small scale occurrences have been observed.

At the base of the *exaratum* subzone, (*falciferum* zone) at Tilton, there appears a high abundance of very small (~100µm diameter), trochospiral foraminifera. These are, as yet, unidentified but are similar to the aragonitic genus *Conorboides* which is also reported from the *falciferum* zone of the Upper Lias at Empingham, Rutland (Horton and Coleman, 1977). Given their extremely low abundances before the Toarcian events and their 'bloom' once the environmental conditions markedly decline, this taxon could be described as a *disaster* or *opportunist* species as described by Harries et al. (1996). The species dominates after a rapid transgression in the *exaratum* subzone before the onset of low oxygen conditions.

Similarly, *Reinholdella macfadyeni* occurs in high abundances in the *tenuicostatum* zone of the North Yorkshire Coast. As an indicator of deep water and transgressive events (Brouwer, 1969), *R. macfadyeni* also disappears with the onset of low-oxygen conditions in the Jet Rock.

Recovery of foraminiferal faunas from the Toarcian anoxic events begins during the *falciferum* subzone of the *falciferum* zone. Members of the coiled genera *Lenticulina* and *Astacolus* appear at Tilton in low diversity upper *falciferum* subzone assemblages, while by the *levesquei* zone at Eype, the assemblages have returned to pre-event diversity levels and are dominated by *Lenticulina*, indicating outer shelf, oxygenated conditions.

Assemblages from the Dotternhausen quarries show a decrease in abundance across the Pliensbachian - Toarcian boundary with few extinctions (this study; Riegraf, 1985). The main extinction level occurs in the *semicelatum* subzone of the *tenuicostatum* zone with some species persisting into the middle of the *falciferum* zone. Early *tenuicostatum* zone assemblages are dominated by *Lenticulina* and *Marginulina* with a minor *Nodosaria* component. The development of anaerobic conditions reflected in the Posidonienschiefer (*falciferum* zone) resulted in an absence of any benthic foraminifera and diversity does not recover until the top of the *bifrons* zone, along with the macrofauna (Little, 1996).

While general in its scope, this study also shows that *Lenticulina,* as the most common genus, must have tolerated a very wide range of environmental conditions. Flattened morphologies predominate in low-oxygen conditions which Bernhard (1986) attributed to an increase in surface area to volume which could either minimise the rate of sinking into low-oxygen sediments, or maximise oxygen intake.

A further noticeable trend in foraminifera from low oxygen conditions is towards a reduction in test size. Comparison with Hettangian - Sinemurian assemblages from the UK comprising large species of *Lenticulina*, *Lingulina tenera*, *Nodosaria* and *Marginulina* (Hylton, 1998) shows a decrease in test size of these genera by the Early Toarcian. Having a small test may also maximise relative surface area (Bernhard, 1986). The foraminiferal assemblages from the Down Cliff Clay (*levesquei* zone) at Eype show a return to greater diversity and a larger test size, indicative of more normal, oxygenated conditions. These assemblages are dominated by the *Lenticulina muensteri* plexus group and *Marginulina.*

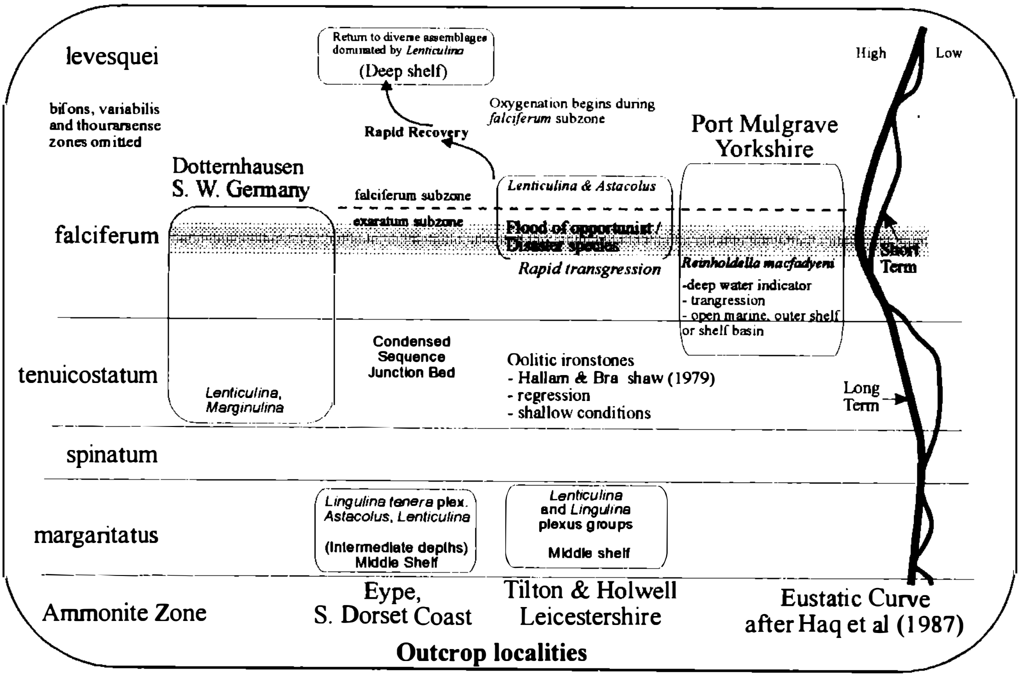

Figure 5. Model of benthic foraminiferal response to environmental conditions in the Late Pliensbachian and Early Toarcian

Conclusions

A fundamental turnover in the foraminiferal assemblage occurs in the *falciferum* zone during which many foraminiferal taxa become extinct. While large scale benthic foraminiferal trends have been previously identified (Brouwer, 1969; Copestake and Johnson, 1989; Brasier, 1988), this high resolution study has revealed the existence of opportunist or disaster foraminiferal species. These species occur during times of environmental stress and exploit available habitat space for short periods of time.

Interpretations of Early Jurassic palaeoenvironments based upon foraminiferal associations generally match sea-level curves and agree with sedimentological evidence. The response of benthic foraminifera during the Late Pliensbachian - Early Toarcian can be linked to changes in sea-level and the resulting development of low oxygen conditions. Uniserial forms such as *Nodosaria*, *Frondicularia* and *Lingulina*, which dominate Pliensbachian assemblages, become less important in Toarcian and subsequent assemblages, whereas the coiled forms like *Lenticulina* dominate after the event. Generalists such as *Lenticulina* most likely had certain adaptations that allowed them to survive low oxygen conditions and contribute to the rapid recovery that started in the *falciferum* subzone.

This study continues with further investigation of Upper Pliensbachian - Lower Toarcian sections in the United Kingdom, South West Germany and southern France.

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References

BERNHARD, J. M. 1986. Characteristic assemblages and morphologies of benthic foraminifera from the anoxic, organic-rich deposits: Jurassic through Holocene. Journal of Foraminiferal Research, 16:207-215.

BRASIER, M. D. 1988. Foraminiferid extinction and ecological collapse during global biological events, p. 37-64. *In* G. P. Larwood, (ed.), 1988. Extinction and Survival in the Fossil Record. Systematics Association Special Volume, 34, Clarendon Press, Oxford.

BROUWER, J. 1969. Foraminiferal assemblages from the Lias of North-western Europe. Verhandlingen der Koninklijke Nederlandge Akademie Van Wetenschappen, Afo., Natuurkunde, Number 25(4), 64 p.

COPE, J. C., T. A. GETTY, M. K. HOWARTH, N. MORTON, AND H. S. TORRENS. 1980. A Correlation of Jurassic Rocks in the British Isles, Part 1: Introduction and Lower Jurassic. Geological Society of London Special Report, 14, 109 p.

COPESTAKE, P. 1978. Foraminifera from the Lower and Middle Lias of the Mochras Borehole. Unpublished PhD thesis, University College of Wales, Aberystwyth.

____, AND B. JOHNSON. 1989. The Hettangian to Toarcian (Lower Jurassic), p. 129-272. *In* D. G. Jenkins, and J. W. Murray, (eds.), Stratigraphical Atlas of Fossil Foraminifera, (second edition), British Micropalaeontological Society Series, London.

HALLAM, A. 1955. The palaeontology and stratigraphy of the Marlstone Rock-bed in Leicestershire. Transactions of the Leicester Literary and Philosophical Society, 49, 17-35.

____. 1961. Cyclothems, Trangressions and Faunal Change in the Lias of Northwest Europe. Transactions of the Edinburgh Geological Society, 18, 124-174.

____. 1986. The Pliensbachian and Tithonian extinction events. Nature, 319:765-768.

____. 1987. Radiations and extinctions in relation to environmental change in the marine Lower Jurassic of north-west Europe. Paleobiology, 13:152-168.

____. 1988. A re-evaluation of Jurassic eustacy in the light of new data and the revised Exxon curve, p. 261-273. *In* C. K. Wilgus, B. S. Hastings, C. A. Ross, H. Posamentier, J. Van

Wagoner, and C. G. St. C. Kendall, (eds.), Sea-level Changes: an integrated approach. Society of Economic Palaeontologists and Mineralogists, Special Publications, 42.

____. 1997. Estimates of the amount and rate of sea-level change across the Rhaetian-Hettangian and Pliensbachian-Toarcian boundaries (latest Triassic to early Jurassic). Journal of the Geological Society, London, 154:773-779.

____, AND M. BRADSHAW. 1979. Bituminous shales and Oolitic Ironstones as indicators of transgressions and regressions. Journal of the Geological Society, London, 136:157-164.

HAQ, B., J. HARDENBOL, AND P. VAIL. 1987. Chronology of fluctuating sea levels since the Triassic (250 million years ago to the present). Science, 235:1156-1167.

HARRIES, P. J., E. G. KAUFFMAN, AND T. A. HANSEN. 1996. Models for biotic survival following mass extinction, p. 41-60. *In* M. B. Hart (ed.), Biotic Recovery from Mass Extinction Events. Geological Society Special Publication, 102.

HESSELBO, S. P. AND H. C. JENKYNS. 1995. A comparison of the Hettangian to Bajocian successions of Dorset and Yorkshire, p. 105-150. *In* P. D. Taylor (ed.), Field Geology of the British Jurassic. Geological Society, London.

HORTON, A. AND B. E. COLEMAN. 1977. The lithostratigraphy and micropalaeontology of the Upper Lias at Empingham, Rutland. Bulletin of the Geological Survey of Great Britain, 62:1-12.

HOWARTH, M. K. 1980. The Toarcian age of the upper part of the Marlstone Rock Bed of England. Palaeontology, 23:637-656.

HYLTON, M. D. 1998. A preliminary analysis of the foraminifera from the Hettangian - Sinemurian boundary of East Quantoxhead, West Somerset. Geoscience in south-west England, 9, 203-208.

JENKYNS, H. C. 1988. The Early Toarcian (Jurassic) anoxic event: Stratigraphic, sedimentary and geochemical evidence. American Journal of Science, 288:101-151.

JOHNSON, B. 1975. Upper Domarian and Toarcian Foraminifera from the Llanbedr (Mochras Farm) Borehole, North Wales. Unpublished PhD thesis, University College of Wales, Aberystwyth.

LITTLE, C. T. S. 1996. The Pliensbachian-Toarcian (Lower Jurassic) Extinction Event. Geological Society of America Special Paper, 307, 505-512.

____, AND M. J. BENTON. 1995. Early Jurassic Mass Extinction: A global long term event. Geology, 23:495-498.

MULLER, F. L., Jr. 1990. The paleoecology of the Liassic benthic foraminifera of Great Britain. Unpublished Ph.D. Thesis, Graduate School - New Brunswick, Rutgers, The State University of New Jersey, 310 p.

RAUP, D. M. AND J. J. Jr. SEPKOSKI. 1984. Periodicity of extinctions in the geologic past. Proceedings of the National Academy of Sciences, USA, 81:801-805.

____, AND ____. 1986. Periodic extinctions of families and genera. Science, 231:833-836.

RIEGRAF, W. 1985. Mikrofauna, Biostratigraphie und Fazies im Unteren Toarcium Südwestdeutschlands und Vergleiche mit Benachbarten Gebieten. Tubinger mikropaläontologische Mitteilungen, 3, 232 p.

SEPKOSKI, J. J., Jr. AND D. M. RAUP. 1986. Periodicity in Marine Extinction Events, p. 3-36. *In* D. K. Elliot (ed.), Dynamics of Extinction, Wiley.

WOODLAND, A. W. (ed.). 1971. The Llanbedr (Mochras Farm) Borehole. Institute of Geological Sciences Report, 71/18, 115 p.

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Jurassic Climatic Events of North China and their Geological Significance

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Keywords: Climate Events, Terrestrial Jurassic, Coal-Bearing Strata, Red Beds, Petrified Forest, Paleolatitude, North China

Abstract: During the Jurassic, greenhouse climatic effects were dominant worldwide, but shorter climatic fluctuations were well-marked on global or vast regional scales. In north China, three distinct climatic events have been recognized: (1) the well-known Early-early Middle Jurassic coal-bearing strata accumulated in the northern humid warm-temperate zone, as shown by abundant biotic and lithologic data. However, a short warm-dry climatic event occurred in the Qaidam Basin and other basins all located west of the Taihang Mountains. Farther east, near the humid Circum-Pacific belt, this event has not yet been so clearly recognized. (2) A more important warm-dry climatic event occurred in the Bathonian and is recognized in both the Eurasian and Circum-Pacific belts. (3) During the Late Jurassic there were clearly different climatic changes in north China and the Circum-Pacific belt. In north China, warm-dry conditions reached an acme, with deposition of mainly red beds, marked by aeolian cross-bedding and volcanics. In the famous Haojiagou section of the Junggar basin, the early Late Jurassic Qigu Formation consists of extensive red beds with a white, tuffaceous horizon near the base. It is interesting to note that this marked horizon represents the maximum flooding surface (MFS) of a sedimentary sequence, and corresponds to the Middle/Late Jurassic biostratigraphic boundary. Obviously, intensive post-Callovian tectonism and volcanism were the main factors causing climate to become warmer and drier both in north China and the Eurasian landmass, reflecting the main process of global climatic change. Proposals for extraterrestrial impacts at the Callovian/Oxfordian and Jurassic/Cretaceous boundaries require investigation of some critical horizons in north China.

Introduction

During the Jurassic, greenhouse climatic conditions dominated throughout the world but, at times, climatic fluctuations occurred on global or vast regional scales (Frakes et al., 1992; Liu and Chen, 1995). The Jurassic System of north China consists predominantly of terrestrial strata, except on the Nadanhada exotic terrane in easternmost Heilongjiang Province. Paleoclimates during deposition of these terrestrial strata are reflected not only by different terrestrial biotas and ecologies, but also by sedimentary and chemical characteristics (Hallam, 1984; Hallam et al., 1991).

The main trend in the Jurassic climate of north China was a general change from humid-warm to dry-hot, but in addition there were three distinct climatic events which occurred in late Early Jurassic (Toarcian), mid Middle Jurassic (Bathonian) and Late Jurassic (Oxfordian-Tithonian) times. Due to the warm-humid air mass and ocean currents in the east, regional variations arose from time to time (Liu and Chen, 1995).

Early-early Middle Jurassic (Hettangian-Bajocian) Climates

Hettangian to Bajocian climates were favorable for the formation of coal beds in north China, including the northern territory of the Paleo-Qinling Mountains (Liu et al., 1987) (Fig. 1). Stratigraphic units include the Beipiao Formation of western Liaoning, the Mentougou (or Yaopu) Formation of Beijing, the Yongdingzhuang and Datong formations of northern Shanxi, the Fuxian, Yan'an and Zhiluo formations of the Ordos basin, the Daxigou and Longfengshan formations of the Gansu Corridor, the Xiaomeigou, Huoshaoshan, Tianshuigou, Yinmagou and Dameigou formations of Qaidam basin, and the Badaowan, Sangonghe and Xishanyao formations of the Junggar and Turpan basins (Table 1).

Age		Junggar & Turpan	Qaidam	Qilian & Gansu Corridor	Ordos	Datong	Beijing & northern Hebei	Beipiao	Southern Da Higgan Mts.
Upper	Tit.	▥	▥	▥	▥	▥	Dabeigou Fm V Zhangjiakou Fm V	▥	Baiyingaolao Fm V
	Kim.	Karaza Fm ●● A	Hongshuigou Fm		Fenfanghe Fm ●●	Baiqi Fm V	Baiqi Fm V		Manitu Fm V Manketouobo Fm V
	Oxf.	Oigu Fm R F .V. 2	Caishiling Fm R	Kushuixia Fm R	Anding Fm R	Tainhechi Fm R	Houcheng Fm R F	Tuchengzi Fm R A	Fujiawazi Fm
Middle	Cal.	Toudunhe Fm 1	Shimengou Fm	Wangjiashan Fm	Zhilou Fm	Yungang Fm	Zhaojishan Fm V Jiulongshan Fm .V.	Lanqi Fm V	Xinming Fm
	Bat.								
	Baj.	Xishanyao Fm ■	Dameigou Fm ■	Longfengshan Fm ■	Yan'an Fm ■	Datong Fm ■	Mentougou Fm ■	Haifanggou Fm ■	Wanbao Fm ■
	Aal.								
Lower	Toa.	Shangonghe Fm ■ R ■	Yinmagou Fm R ■	▥	Fuxian Fm R	Yongdingzhu Fm R		Beipiao Fm ■	Hongqi Fm ■
	Pli.		Tianshuigou Fm ■	Daxigou Fm ■					
	Sin.	Badaowan Fm ■	Huoshaoshan Fm ■		▥	▥	Nandaling Fm V	Xinglonggou Fm V	
	Het.		Xiaomeigou Fm ■						
reference		this paper	Wang 1996	Du 1985	Du et al. 1982	Wang et al. 1982	Xiao et al. 1994	Pu et al. 1985	Wang et al. 1982

■ Coal bed R redbed F petrified forest A aeolian bedding o o conglomarat V volcanic rock .V. tuff

1 Wucaiwan Fm 2 Shishugou Fm

Table 1. Correlation chart for the terrestrial Jurassic in north China.

The dominant rocks of the Hettangian-Bajocian interval are yellowish-brown, yellowish-green, dark grey sandstones and shales, interbedded with important, exploitable coals. The *Neocalamites-Cladophlebis* and *Coniopteris-Phoenicopsis* floras, containing 48-66 genera and 142-178 species, have been found in Beijing and Qaidam areas, respectively.

In the Beijing area, Filicaes and Ginkgopsida make up 59%, Cycadopsida 16.9%, and Coniferopsida 11.2% of the flora (Chen et al., 1984). Dicksoniaceae and Osmundaceae among Filicaes were all present, *Coniopteris* (Dicksoniaceae) reaching its acme and being fairly common; however, typical tropical-subtropical forms, such as Angioteridaceae and Lygodiaceae, were absent, Matoniaceae were rare, and individuals of Dipteridaceae were also uncommon. The flora is regarded as belonging to the subtropical floral province (Wu, 1983) or a transitional position between Vakhrameev's (1965) Indo-European and Siberian flora provinces (Zhang et al., 1987). It is probably closer to a warm-temperate climatic condition because of relatively abundant Cycadopsida and numerous deciduous trees with distinct annual rings (Chen et al., 1984), and is regarded as equivalent to the southern edge of the Siberian province (Vakhrameev, 1965).

Palynologic studies show that gymnospermous pollen dominate the Early Jurassic, making up about 70% of the sporo-pollen assemblage. During the early Middle Jurassic, gymnospermous pollens and pteridophyte spores each represent about 50%. The most abundant pteridophyte spores are *Cycadopites; Classopollis* only makes up 1-3% (Du et al., 1982; Wu et al., 1983). The sporo-pollen assemblages are also similar to those of the Siberian temperate floral province. Certainly, north China lay within the humid warm-temperate climatic zone during Early- early Middle Jurassic times.

The only area with marine Jurassic strata in north China is the Nadanhada terrane, easternmost Heilongjiang province. Early Jurassic radiolarians such as *Parahsuum, Pseudohsuum, Canoptum* and *Canutus* occur in thin bedded, siliceous rocks of this area (Wang et al., 1995). The paleolatitude obtained from the Upper Triassic to Lower Jurassic in this area (Shao et al., 1990) is 21.8°N, suggesting that the Nadanhada terrane lay within tropical to subtropical oceanic/climatic conditions and was located near the Tamba-Mino terrane of southwest Japan in the Early Jurassic (Mizutani et al., 1988).

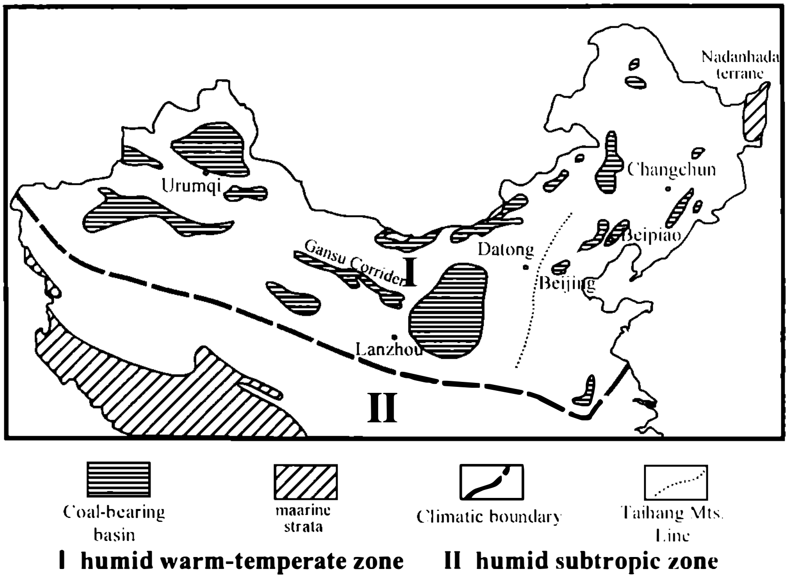

Figure 1. Early-early Middle Jurassic climatic zonation in north China.

Late Middle Jurassic radiolarian faunas with *Stichocapsa, Tricolocapsa, Eucyrtidiellum* and lacking pantanellids, have a different aspect from those of the Tamba-Mino terrane, and show affinities to mid-latitude, cool water faunas (Wang et al., 1995). Obviously the Nadanhada terrane had drifted northward and separated from the Tamba-Mino terrane in Middle Jurassic time (Liu and Lui, 1995). Late Jurassic *Buchia* faunas ranging from Late Oxfordian to Tithonian (Middle Volgian), occur in eastern Heilongjiang both on the Nadanhada terrane and adjacent areas (Sha and Fürsich, 1993). According to the paleolatitude of 43.5°N obtained from the Upper Jurassic in this area (Shao et al., 1990), the Nadanhada terrane accreted to the continental margin of north China and was located in the cool-temperate climatic zone of the Boreal realm during Late Jurassic time (Shao et al., 1992).

A short, warm-dry climatic event has been identified in the late Early Jurassic in north China. Li et al. (1988) and Wang (1996) reported a special flora from the Yinmagou Formation of the Qaidam Basin. This formation consists of red and variegated mudstones and sandstones, and yields *Pagiophyllum, Brachyphyllum, Cladophlebis* and a few *Coniopteris*. *Classopollis* averages 55.9% of the total sporo-pollen assemblage, indicating warm, semi-dry climatic conditions. Similar paleontological and lithological evidence of the same event has also been found in the Junggar (in lower part of Sangonghe Formation), Ordos (in Fuxian Formation) and Datong (in Yongdingzhuang Formation) basins.

It is interesting to note that the evidence for this warm-dry climate occurs only west of the Taihang Mountains. Farther east, near the humid circum-Pacific, this climatic event has not yet been clearly recognized. This climatic event was short-lived but can be traced extensively over Eurasia during Toarcian time (Vakhrameev et al., 1988). It corresponds to the Toarcian anoxic event in European marine strata (Ager, 1981; Jenkyns et al., 1986), possibly related to an eustatic sea level rise as a direct consequence of the initiation of a mid-Atlantic ridge (Hallam, 1972).

Late Middle Jurassic (Bathonian-Callovian) Climate

Global transgression and climatic changes occurred in Late or post-Bajocian time, so that the dry-hot climatic zone expanded from the equator towards higher latitudes (Hallam, 1984; Parrish, 1992). An important adjustment of regional climates took place in north China at the same time (Wu et al., 1994; Liu and Chen, 1995) (Fig. 2).

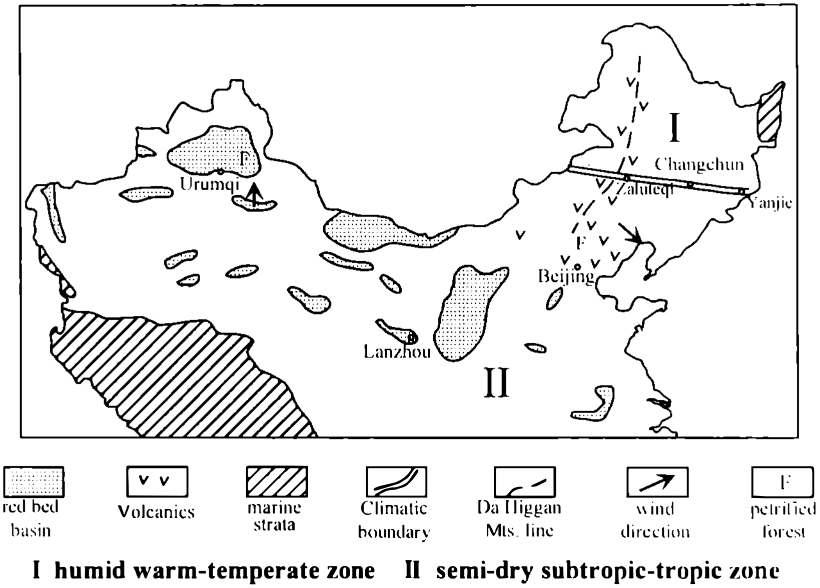

Figure 2. Late Middle-Late Jurassic climatic zonation in north China.

The northern warm-temperate climatic zone is limited to the region north of the Zaluteqi-Changchun-Yanjie Line (climatic boundary in Fig. 2) and is represented by terrestrial-pyroclastic strata. The NNE trending ancient Da Higgan Mountains acted as a barrier and resulted in further regional differentiation on their east and west sides. To the east, strata represent lake-bog deposits with coals and oil-shales. The most abundant Filicinae are members of Dicksoniaceae, Osmundaceae and Polypodiaceae. Coniferopsida ranks as the secondary lush plants. Cycadopsida and Ginkgopsida are not developed (Zhang and Zheng, 1987). This flora reflects warmer and more humid climatic conditions resulting from the effect of the warm paleo-Pacific air mass and ocean currents. To the west no coals have been found. Coniferopsida are dominant with secondary Ginkgopsida and Cycadopsida, but few Filicinae. There is no great difference in palynofloras between east and west sides. Pteridophyte spores and gymnospermous pollen occur in nearly equal amounts. Gymnospermous pollen consists primarily of temperate types of Pinaceae, Cupressaceae and Ginkgopsida (Pu and Wu, 1985).

South of the Zaluteqi-Changchun-Yanjie Line the best climatic indicator is a thick-shelled unionid bivalve fauna of *Lamprotula (Eolamprotula)-Psilunio-Cuneopsis*. *Lamprotula* and *Cuneopsis* are considered to represent warm-temperate to subtropical climates. Some genera of unionids such as *Cuneopsis*, appeared in the Early Jurassic in central Yunnan, but it was most abundant and expanded northward primarily in the Bathonian (Gu, 1982). This thick-shelled, unionid fauna has been dated by the associated marine fossils in paralic facies of western Yunnan and the Tanggula region in northern Tibet.

An important warm-dry climatic event began in the Early Bathonian, and can be traced throughout Eurasia and the circum-Pacific Belt. In the Gansu Corridor the Aalenian-Bajocian Longfengshan Formation contains important coal beds, with abundant *Coniopteris* spp. and the *Cyathidites-Leiotriletes* sporo-pollen assemblage in which the *Classopollis* content reaches only 0.3%. By constrast, in the overlying Bathonian-Callovian Wangjiashan Formation, with yellowish and red sandstones, *Classopollis* suddenly increases from 1.2% to 79.2% upsection (Du, 1985). In Eurasia and the circum-Pacific Belt there are abundant data providing evidence for the expansion of a warm climatic zone and warm oceanic currents during the Late Bajocian - Bathonian interval (Parrish, 1992). This climatic event was possibly of global extent, at least in the Northern Hemisphere (Parrish et al., 1996).

Late Jurassic (Oxfordian-Tithonian) Climate

In north China and adjacent regions of Eurasia, the warm-dry episode began in the Bathonian and strengthened subsequent to the Oxfordian-Tithonian. Red beds and volcanic strata are the main Jurassic rock types (Table 1).

The synchroneity of climatic change with the Middle/Late Jurassic chronostratigraphic boundary in terrestrial strata in north China is still being debated among Chinese geologists (Wang, 1998). In the famous Haojiagou section of the Junggar Basin, the lower Upper Jurassic Qigu Formation consists of a thick sequence of red beds with a white tuffaceous horizon near its base. It is very interesting that this tuffaceous horizon represents the maximum flooding surface (MFS) as well as the transform surface of stratigraphic architecture between Expanding Systems Tract (EST) and HST of a sedimentary sequence (Fig. 3). It should be pointed out that this key surface also corresponds to the Middle/Late Jurassic boundary as recognized in terrestrial biostratigraphy (Fig. 4). This shows that the application of sequence stratigraphy is a method available for solving long-standing arguments, including the establishment of a practicable series boundary. The obvious difference of terrestrial biotic abundances below and above this tuffaceous horizon also reflects an important climatic change between Middle and Late Jurassic times.

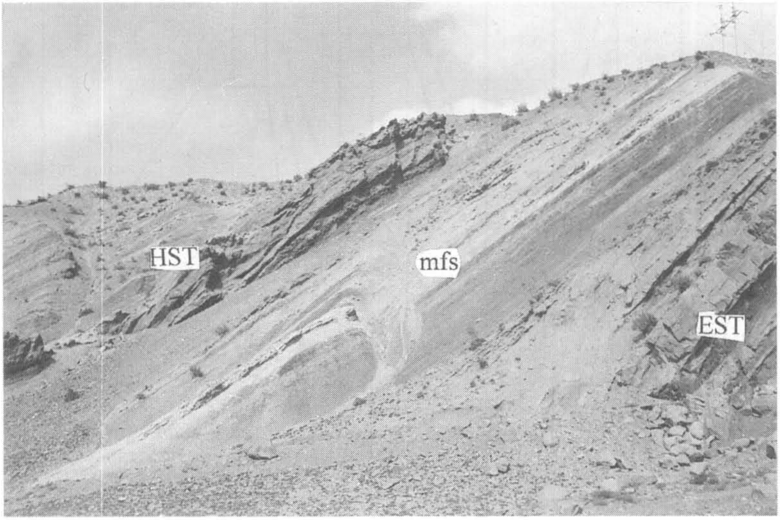

Figure 3. Tuffaceous horizon as a key surface (MFS) near the base of the Qigu Formation in the Haojiagou section, eastern Junggar Basin.

Upper Jurassic eolian deposits and buried petrified forests have been found extensively in north China. The genesis of eolian sands and oases in the lower Upper Jurassic Tuchengzi Formation of the Beipiao area in western Liaoning (Zhang and He, 1983), and the large scale eolian cross-bedding in the uppermost Upper Jurassic Kalaza Formation in the Turpan Basin (Table 1), both indicate a relatively dry climate. Based on measurements of eolian cross-bedding, the main wind direction in the Beipiao area was SE140° (Fig. 5), but in the Turpan Basin was near N360°. These directions conform generally to the northern winter model of atmospheric circulation during the Late Jurassic (Moore et al., 1992; Parrish, 1992).

The magnificent petrified forest of the Shishugou Formation (corresponds to the Qigu Formation) in the eastern Junggar area represents an extensive, flourishing forest that was rapidly buried (Cai, 1996). Similar and smaller petrified forests which occur at the same stratigraphic horizon at Xiadelongwan of northern Beijing (Xiao et al., 1994), probably grew under alternating semi-humid and semi-dry environments.

Uppermost Jurassic strata are more sparsely distributed, and are represented by a volcaniclastic succession including the Baiqi, Zhangjiakou and Dabeigou formations in eastern north China (Table

(Table 1). A few fossils, *Schizolepis* (seed), *Classopollis* and *Cycadopites* (sporo-pollen), have been reported in the Zhangjiakou Formation (Cai, 1986) which contains many zeolite ore beds. Zeolite often forms in alkali lakes (pH≅9.5-10.5), suggesting that the climate was still dry-hot.

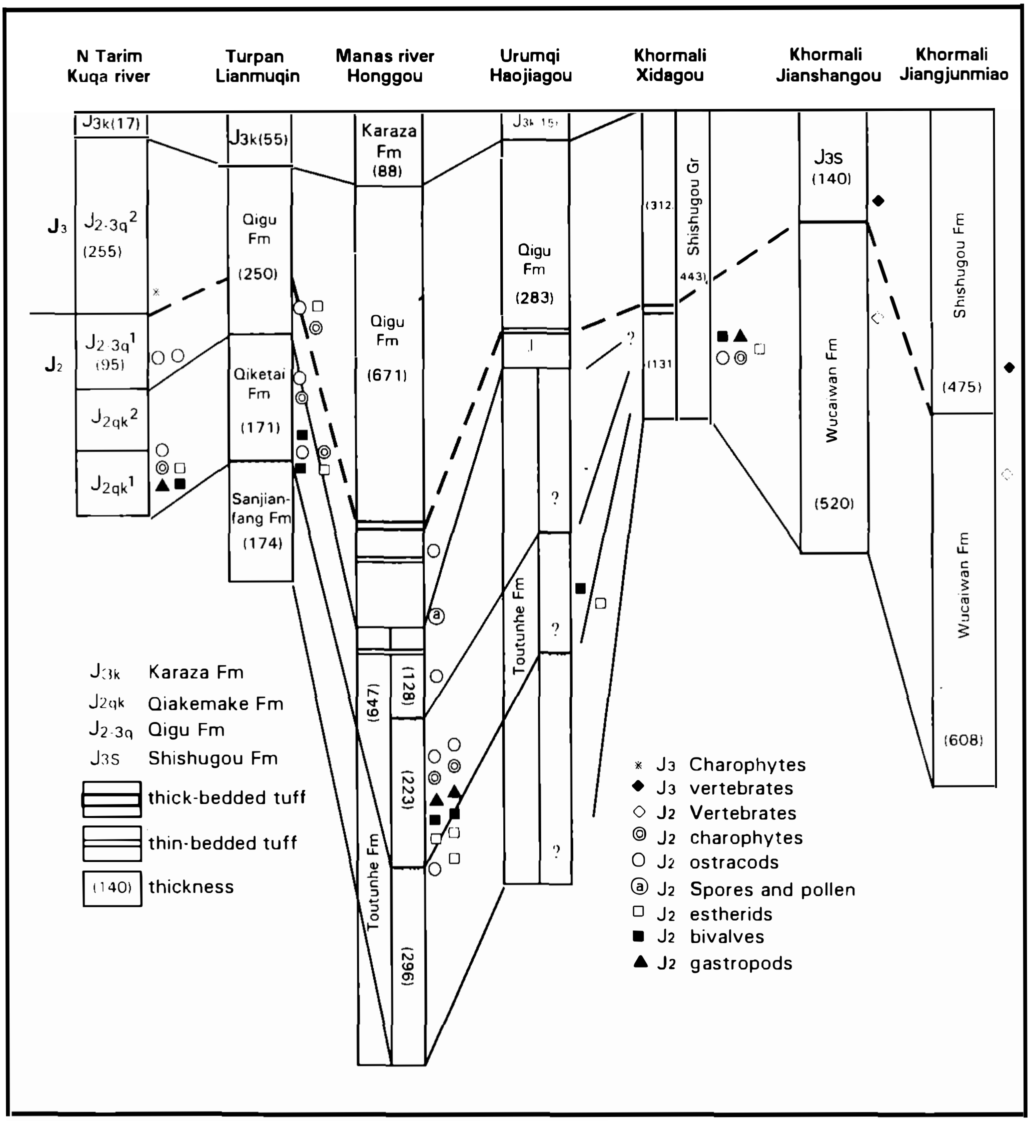

Figure 4. Relationship between chronostratigraphic and lithostratigraphic boundaries of Middle/Upper Jurassic in northern Xinjiang.

In the circum-Pacific Belt, marine fossils belonging to both the Boreal and Tethyan realms indicate the presence of seawater of different temperatures. The southward migration of Boreal faunas during the Late Callovian to Kimmeridgian (Hillebrandt et al., 1992) shows a strengthening of cold currents from high latitudes and falling seawater temperatures. This differs from the tendency torward warm-dry climates during the Late Jurassic in Eurasia. These differences reflect a complex relationship between dynamic changes of the lithosphere, hydrosphere and atmosphere of the Earth during the critical time of accelerated rifting of Pangea with increased rates of sea-floor spreading (Moore et al., 1992; Liu et al., 1997a).

Recent paleomagnetic research in the Tarim plate indicates that the paleolatitude of this region changed from 14.2±3.7°N in the Devonian to 30.2±5.8°N by the Middle Jurassic (Fang, Chen, et al., 1990). This northward movement coincides with the history of plate drift and collision recorded by geology. But other paleolatitude results for the Tarim plate, averaging 30.1°N by the Early-Middle Jurassic and 24.8°N by the Late Jurassic, probably reflect a southward rebounding (Li et al., 1996). Paleomagnetic results from the eastern part of north China (Sino-Korean plate) indicate that the paleolatitude of this region was 32.2±3.8°N in the Middle Jurassic, and that a counterclockwise rotation of 53.0±9.0° occurred between the Middle Triassic and Middle Jurassic (Fang, Tan, et al., 1990). Hence, it is concluded that the northward movement of the warm-dry climate zone in Eurasia during the Late Jurassic was also related to this southward movement of the Tarim plate. It should be pointed out that the break-up of Pangea resulted in intensive crustal movement and volcanism after the Callovian (160Ma±) and is the main explanation for the climate becoming gradually warmer and drier in Eurasia. During the interval, marine biogeographic changes between the Boreal/Subboreal and Tethyan realms in the northern circum-Pacific region may have been brought about by regional changes in paleogeographic conditions rather than by global climatic cooling.

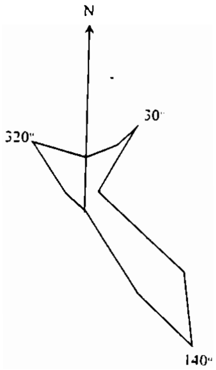

Figure 5. Eolian cross-bedding dip rose diagram of the Tuchengzi Formation (after Zhang and He, 1983)

Possible Cosmic Factors Affecting Jurassic Climatic Change

With respect to reports of a possible extraterrestrial impact event at the Callovian/Oxfordian boundary, suggested for Europe (Brochwicz-Lewinski et al., 1987), the Mjölnir structure (143±20 Ma) in the Barents Sea (Gudlaugsson, 1993), and the Gosses Bluff crater near the Jurassic/Cretaceous boundary (142.5 Ma) in Australia (Glikson, 1997), it is necessary to check some key horizons and suspect circular structures in north China, such as the tuffaceous horizon near the basal boundary of the Qigu Formation in the Junggar Basin and the Duolun circular structure about 250km north of Beijing. More detailed study is necessary to ascertain whether any record of an impact event is preserved in these horizons or structures (Liu et al., 1997b).

Acknowledgments

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References

AGER, D. V. 1981. Major marine cycles in the Mesozoic. Journal of the Geological Society, London, 138:159-166.

BROCHWICZ-LEWINSKI, W., A. GASIEWICZ, W. E. KRUMBEIN, et al. 1987. Iridium anomaly at the Middle/Upper boundary. Abstract of Second Working Session, IGCP Proj.199, 24.

CAI, K. Y. 1986. Discovery of sporo-pollen fossils of Zhangjiakou Formation in Fengning county, northern Hebei. Geological Information of Hebei, 3:72-75 (in Chinese).

____. 1996. Forest of Jurassic Period in General Gobi of Qitai, Xinjiang of China. 30th International Geological Congress Abstract, Vol. 2:69.

CHEN, F., Y. W. DOU, AND Q. S. HUANG. 1984. The Jurassic Flora of West Hills Beijing (Peking. Beijing: Geology Publishing House, 136p., 38pls. (in Chinese with English summary).

DU, B. A. 1985. Stratigraphic and paleogeographic significances of Middle Jurassic sporo-pollen in Wangjiashan area, Jinyuan County, Gansu Province. Geological Review, 2:131-141 (in Chinese with English abstract).

____, X. R. LI, AND W. H. DUAN. 1982. Sporo-pollen assemblages from Yan'an and Zhiluo formations in Chongxin County, Gansu Province. Acta Palaeontologica Sinica, 21 (5):597-606 (in Chinese with English abstract).

FANG, D. J., H. L. CHEN, G. H. JIN, et al. 1990. Late Paleozoic and Mesozoic paleomagnetism and tectonic evolution of the Tarim terrane. *In* T.J. Wiley et al. (eds.), Terrane Analysis of China and the Pacific Rim. Houston, Texas, U.S.A., p.251-255.

____, X. D. TAN, Y. B. GOU, et al. 1990. Paleomagnetism and tectonic evolution in Mesozoic for North China plate. *In* T. J. Wiley et al. (eds.), Terrane Analysis of China and the Pacific Rim. Houston, Texas, U.S.A., p.359-363.

FRAKES, L. A., J. E. FRANCIS, AND J. I. SYKTUS. 1992. Climate Modes of the Phanerozoic. Cambridge: Cambridge University Press, 274pp.

GLIKSON, A. Y. 1997. Asteroid/comet impact connections of Precambrian episodes: Re-emergence of the catastrophic paradigm in earth science. Earth Science Frontiers (China University of Geosciences, Beijing), 4 (3-4):23-46.

GU, Z. W. 1982. Distribution and development of nonmarine Mesozoic bivalves and formations in China. Scientia Sinica, Series B, 25:438-452.

GUDLAUGSSON, S. T. 1993. Large impact crater in the Barents Sea. Geology, 21:291-294.

HALLAM, A. 1972. Diversity and density characteristics of Pliensbachian-Toarcian molluscan and brachiopod fauna of the North Atlantic margins. Lethaia, 5:389-412.

____. 1984. Humid and air zones in the Jurassic and Cretaceous. Palaeogeography, Palaeoclimatology, Palaeoecology, 47:195-223.

____, J. A. GROSE, AND A. H. RUFFELL. 1991. Palaeoclimatic significance of changes in clay mineralogy across the Jurassic-Cretaceous boundary in England and France. Palaeogeography, Palaeoclimatology, Palaeoecology, 81:173-187.

HILLEBRANDT, A. von, G. E. G. WESTERMANN, J. H. CALLOMON, et al. 1992. Ammonites of the Circum-Pacific region. *In* G.E.G. Westermann (ed.), The Jurassic of the Circum-Pacific. Cambridge University Press, p.342-359.

JENKYNS, H. C. AND C. J. CLAYTON. 1986. Black shales and carbon isotopes in pelagic sediments from the Tethyan Lower Jurassic. Sedimentology, 33:87-106.

LI, P. J., HE, Y. L., WU, X. W. et al. 1988. Early and Middle Jurassic Strata and Their Floras from Northeastern Border of Qaidam Basin, Qinghai. Nanjing: Nanjing University Press, 231pp, 140pls (in Chinese with English summary).

LI, Y. A., LI, X. D., SUN, D. J. et al. 1996. Tectonic Evolution of Qiangtang Block and Kangxiwar Structure Zone in Kara-Kunlun Mountains, Southwest of Xinjiang, China. Urumqi: Xinjiang Science & Technology & Hygiene Publishing House, p.73-94 (in Chinese).

LIU, B.P. AND F. CHEN. 1995. Mid-Jurassic bioclimate events and their dynamic significance of geospheres. Laboratory of Lithosphere Tectonics and Its Dynamics, Annual Report 1994. Beijing: Seismological Press, 151-159.

____. AND X. S. CUI. 1997a. An approach to rhythms in Earth's history and their astronomic factors. Laboratory of Lithosphere Tectonics and Its Dynamics, Annual Report 1996. Beijing: Geology Publishing House, p.117-126.

____, Q. S. YU, AND S. R. YANG. 1987. Bivalve fauna from Mentougou Formation in Western Hills, Beijing and the age of Early Mesozoic coal epoch in North China. Proceeding of the Symposium on Mesozoic and Cenozoic Geology, China. Beijing: Geology Publishing House, p.137-148.

____. AND S. H. ZHANG. 1997b. Rhythms of different geospheres and their relations in Middle Jurassic-Early Cretaceous. Earth Science Frontiers, 4(3):65-74 (in Chinese with English abstract).

LIU, S. F. AND B. P. LIU. 1995. Comparative study on the juxtaposition events of Jurassic terranes in the northern Circum-Pacific belt. Laboratory of Lithosphere Tectonics and Its Dynamics, Annual Report 1994. Beijing: Seismological Press, 138-147.

MIZUTANI, S., S. KOJIMA, J. A. SHAO, et al. 1988. Mesozoic radiolarians from the Nadanhada area, Northeast China. Proceedings of the Japanese Academy, Ser. B, 62:337-340.

MOORE, G. T., D. N. HAYASHIDA, C. A. ROSS, AND S. R. JACOBSON. 1992. Paleoclimate of the Kimmeridgian/Tithonian (Late Jurassic) world: I. Results using a general circulation model. Palaeogeography, Palaeoclimatology, Palaeoecology, 93:113-150.

PARRISH, J. T. 1992. Jurassic climate and oceanography of the Pacific region. *In* G.E.G. Westermann, (ed.), The Jurassic of the Circum-Pacific. Cambridge Universiy Press, p.365-379.

____, M. T. BRADSHAW, A. T. BRAKEL, S. M. MULHOLLAND, J. M. TOTTERDELL, AND A. N. YEATES. 1996. Paleoclimatology of Australia during the Pangean interval. Palaeoclimates--Data and Modeling, 1:23-57.

PU, R. G. AND H. Z. WU. 1985. Mesozoic sporo-pollen assemblages in western Liaoning and their stratigraphic significance. *In* Mesozoic Stratigraphy and Palaeontology of Western Liaoning, 2:121-212, 25 pls (in Chinese with English abstract).

SHA, J. G. AND F. T. FÜRSICH. 1993. Biostratigraphy of the Upper Jurassic-Lower Cretaceous bivalves Buchia and Aucellina of eastern Helongjiang, northeast China. Geology Magazine, 130(4):533-542.

SHAO, J. A., C. Y. WANG, K. D. TANG, et .al. 1990. Relationship between stratigraphy and terrane in the Nadanhada Range. Journal of Stratigraphy, 14 (4):286-291 (in Chinese with English abstract).

____, C. Y. WANG, AND K. D. TANG. 1992. A new approach to the tectonics in the Ussuri (Wusuli) region. Geological Review, 38 (1):33-39 (in Chinese with English abstract).

VAKHRAMEEV, V. A. 1965. Jurassic floras of the USSR. Palaeobotanist, 14:118-123.

____, V. I. ILJINA, AND N. I. FOKINA. 1988. Subdivision of the continental Jurassic on plants. *In* G.Y. Krymholts, et al. (eds.), The Jurassic Ammonite Zones of the Soviet Union, p.63-71.

WANG, S.. E. 1998. Correlation of the continental Jurassic in the north of China to the paralic Jurassic in Northwest Scotland, United Kingdom. Acta Geologica Sinica, 72 (1):11-21 (in Chinese with English abstract).

____, Z. W. CHENG, N. W. WANG, et al. 1982. The Jurassic System of China. Stratigraphy of China, 11. Beijing: Geology Publishing House, 95-99 (in Chinese).

WANG, W. L., S. L. ZHENG, L. J. ZHANG, et al. 1995. Tectono-stratigraphy of the Circum-Pacific Belt in Northeast China. Beijing: Geol. Publ. House, 263-267 (in Chinese).

WANG, Y. D. 1996. Vegetation and palaeoclimate change of Early-Middle Jurassic palynoflora from coal-bearing basins in Northwest China. Intern. Geol. Cong., 30th, Beijing, 1996, Abstract, Vol.1: 23.

WU, H. Z. AND X. L. ZHANG. 1983. Palynological assemblage from the upper coal-bearing member of the Beipiao Formation of Liaoning. Acta Palaeontologica Sinica, 22 (5):564-570, 2pls (in Chinese with English abstract).

WU, S. B. AND J. N. TONG. 1994. Jurassic. *In* H.F. Yin (ed.), The Palaeobiogeography of China. Oxford Biogeography Series, No. 8. Oxford: Clarendon Press, 219-245.

WU, S. Q. 1983. On the Late Triassic, Lower and Middle Jurassic floras and phytogeographic provinces of China. *In* Palaeobiogeography Provinces of China. Beijing: Science Press, 121-130 (in Chinese).

XIAO, Z. Z., H. L. YANG, AND Q. S. SHAN. 1994. The Mesozoic Stratigraphy and Biota of Beijing Area. Beijing: Geology Publishing House, 33-35 (in Chinese with English summary).

ZHANG, C. B. AND Y. L. HE. 1983. Late Middle Jurassic aeolian deposits near Beipiao, Liaoning. Acta Sedimentologica Sinica, 1 (4):48-62, 2pls (in Chinese with English abstract).

ZHANG, W. AND S. L. ZHENG. 1987. Early Mesozoic fossil plants in western Liaoning, Northeast China. *In* Mesozoic Stratigraphy and Palaeontology of Western Liaoning, 3:239-338, 30pls (in Chinese with English abstract).

ZHAO, Y., Z. Y. YANG, AND X. H. MA. 1994. Geotectonic transition from Paleoasian system and Paleotethyan system to Paleopacific active continental margin in eastern Asia. Scientia Geologica Sinica, 29 (2):106-119 (in Chinese with English abstract).

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Biological and Physical Agents of Shell Concentrations of *Lithiotis* Facies Enhanced by Microstratigraphy and Taphonomy, Early Jurassic, Trento Area (Northern Italy)

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Keywords: Skeletal Concentrations, Taphonomy, Event Beds, Burrowing, Carbonate Platform, Pliensbachian, Venetian Prealps, Italy

Abstract: During the Early Jurassic the "Calcari Grigi" Formation was deposited on the carbonate Trento Platform, a large structural and paleogeographic unit of the Jurassic continental margin of the Southern Alps (northern Italy). The "Rotzo" member, a carbonate lagoonal sequence mainly in the upper part of the Lower Jurassic, has been studied since last century in the Venetian Prealps due to abundance of land plants and large bivalves of the *Lithiotis* facies. In Valbona Pass, near Folgaria, a composite sequence 67m thick deposited during the Late Pliensbachian prior to the *Lithiotis* bioherms has been studied. This sequence consists largely of bioturbated limestones and marls with abundant skeletal concentrations, 1-45cm thick, containing brachiopods, large and small bivalves, turriculate gastropods, echinoderms, encrusting algae, solitary corals and large benthic foraminifers. Taphonomic analysis of these shell layers (utilizing thin sections, polished rock slabs and image-processing computer analyses), reveals that of 450 studied layers, 345 show taphonomic features. This comprehensive facies study is significant, because nearly 75% of studied beds show taphocoenoses, skeletal fabrics and features referable to physical and biological agents.

Five types of skeletal concentrations have been observed: 1) bimodal gastropod concentrations, oriented by fair-weather waves, in a mud-supported fabric (SkC 1); 2) bioclast-supported packstones deposited by combined-flow agents (SkC 2); 3) convex-up oriented bivalve grainstones with shelter porosity, winnowed by weak and repetitive (tidal?) currents (SkC 3); 4) erosive bioclast-supported rudstones-packstones resulting from winter storms or hurricanes and forming proximal to distal tempestites (SkC 4a and b); 5) peloidal to skeletal packstones as forced infillings of subsurface burrows (tubular tempestites, SkC 5). Burrow systems are complex, generally tiered, boxworks and show straight tunnels, meander mazes, shafts, turning chambers, apertures and conical mounds. In tidal areas of modern shelves analogous burrow systems are actually created by crustacean decapods, especially thalassinidean shrimps such as *Callianassa*, or crustacean predators such as mantis shrimps (stomatopods).

Microstratigraphy of skeletal concentrations reveals a strong vertical redistribution of grains by intense biological activities (such as burrowing) that reworked and destroyed partially or completely the original physical structures. Physical structures of densely-packed beds, for this reason, such as gradation, scoured bases and laminated cross-stratification (including HCS and wave ripples) are discontinuous and only locally preserved. Four taphonomic classes have been recovered and discussed: 1) disarticulation, grouping, dispersion; 2) abrasion, truncation, filling with mud; 3) taphonomic distortion, and 4) crustacean burrow activity. Biological agents consist of predators, scavengers, bioadvectors, surficial and deep burrow-lining infauna, encrusters and borers.

Introduction and Previous Studies

The *Lithiotis* facies developed during the Pliensbachian and Toarcian in several carbonate platform sequences of Tethys and Panthalassa, distributed through the Hispanic Corridor between North and South America and the Tethyan region, central Asia and western Pacific as far as Timor (see paleogeographic reconstructions of Nauss and Smith, 1988), before the extinction of genus *Lithiotis* by the end of the Toarcian. These facies represent a mine of information for palaeontologists and palaeobotanists as far back as the 19th century. In fact, in the Pre-Alps the main scientific interest derived from their well preserved, terrestrial floras of such tropical elements as Coniferales and Bennettitales recovered in certain dark, clayey levels of carbonate platform sequences, the famous Rotzo and Lessini land floras (De Zigno, 1885; Wesley, 1956 among others).

Other authors have studied the biostratigraphy, paleobiology, paleogeography and sedimentology of the *Lithiotis* facies in the Venetian area, focusing on the Rotzo and Lessini sequences (Bosellini and Broglio Loriga, 1971; Broglio Loriga and Neri, 1976; Masetti et al., 1996). Other analyses have been directed to taxonomy and systematics of *Lithiotis* and *Cochlearites*, defining morphofunctions, characters and microstructures of *Lithiotis* shells and analysing mollusk communities such as *Mytiloperna*, *Gervilleioperna*, *Opisoma*, *Lithioperna*, *Isognomon*, *Lithiopedalion* and other large to small bivalves (Berti Cavicchi et al., 1971; Accorsi Benini and Broglio Loriga, 1977, 1979; Accorsi Benini, 1979; 1980; 1981; 1985; Benini and Loriga, 1974; Broglio Loriga and Neri, 1976).

Palaeoenvironmental indicators in the Grey limestones ("Calcari Grigi") such as lumachelles (coquinas) and bioherms suggest that the Trento carbonate platform was colonized during the Pliensbachian by large to small bivalves of reef-subtidal-intertidal environments, associated with other invertebrates such as high-spired gastropods (Nerineacea), brachiopods (terebratulids), isolated corals, sponges, red and green algae (*Paleodasycladus*, *Thaumatoporella*, Dasycladaceae) and benthic foraminifers (mainly *Orbitopsella* and *Lituosepta*: see Bosellini and Broglio Loriga, 1971; Accorsi Benini and Broglio Loriga, 1979; Broglio Loriga and Neri, 1976; Broglio Loriga and Fugagnoli, 1997; Fugagnoli, 1996). These diversified assemblages populated shallow-water lagoons and tidal inlets bordered by marshes, swamps, mud flats and oolitic or sandy barrier island complex (Bosellini and Broglio Loriga, 1971; Broglio Loriga and Neri, 1976 and references therein). Other authors have proposed wide and open lagoons linked to gently inclined ramps showing asymmetric cycles of fossiliferous limestone-marl couplets that become progressively mud-dominated from Asiago towards the Val D'Adige, and are bordered by oolitic facies in the M. Baldo area (Masetti et al., 1996).

Although the general paleontology and sedimentology of the Veneto area is well understood, taphonomy of different shell beds has not been sufficiently studied in order to relate the paleoenvironmental setting with biologic activity and sedimentation. Until now, storm deposits in carbonate sequences of the Jurassic of this area have been considered only marginally (Bosellini and Broglio Loriga, 1971) and trace fossils have not been analyzed in detail, having been regarded as subordinate to event stratigraphy and neglected in this area for precise paleoenvironmental interpretations (Bromley, 1996).

The aim of this paper is to apply to each skeletal concentration of the Pliensbachian Rotzo member, taphonomic categories known in the literature (Fürsich and Oshmann, 1993; Fernández López, 1997a; 1997b). A methodology based on taphonomic recurrences by computer image processing helped to identify the origin of many skeletal concentrations of shallow neritic setting; such as fair-weather waves or combined-flow agents, proximal to distal tempestites due to winter storms or hurricanes, and winnowed beds due to currents (Aigner, 1985; Duke, 1985; Fürsich and Oschmann, 1993; Monaco, 1995). The purpose is to distinguish, by means of microstratigraphy and taphonomy, biological from physical agents, such as burrowing and taphonomic distortion of biogenic origin, in order to define the allocation of skeletal remains after their deposition and to reconstruct the degree of reworking from the action of benthic organisms.

Geological Setting and Stratigraphy

The Grey Limestones, known as the "Calcari Grigi" Formation of the Southern Alps, northern Italy (Bosellini and Broglio Loriga 1971), constituted during the Early Jurassic a wide, shallow-water carbonate platform developed in the Western Venetian Prealps. This large structural and paleogeographic unit, the Trento Platform, formed a horst, 70-80km wide, with deep, marine troughs to both the west and east, the Lombardo and Belluno basins, respectively (Fig. 1). The Trento platform was formed from an epicontinental stage at the end of the Triassic with uniform sedimentation of paralic-epicontinental type, evolving at the beginning of the Jurassic into an oceanic platform, due to disintegration of the original Triassic epicontinental platform, showing a Bahamian oolitic type of sedimentation. During the final stage of the Early Jurassic some calcarenitic facies developed along belts connecting the basin-platform system, as in the M. Baldo area (M. Lessini) located to the west and the M.Grappa-Bassano area to the east, providing evidence of high-energy regimes in a marine setting which bordered in the center a lagoonal domain in the

Trento-Rovereto area (see paleogeography of Bosellini and Broglio Loriga, 1971).

Recently the Grey Limestones Formation of the Asiago area has been divided into three members. The upper member has been formally named "Member of Rotzo" (near Rotzo, Asiago area), and studied since last century near Rotzo village, for the chronostratigraphic value of its terrestrial plants, algae and microfossils of the *Lithiotis* facies and for lithostratigraphic markers (see extensive literature reported in Broglio Loriga and Neri, 1976). This upper part of the Grey Limestones, the Rotzo member, is of mainly Late Pliensbachian age (Domerian, of the *Paleodasycladus mediterraneous* zone, or, from the bottom, *Orbitopsella* and *Lituosepta compressa* subzones, according to Accorsi Benini and Broglio Loriga, 1979; see also Cita, 1959), with the upper part, made up of some bioherms and oolitic deposits, extending probably into the Aalenian (see also Zempolich, 1993 and references therein).

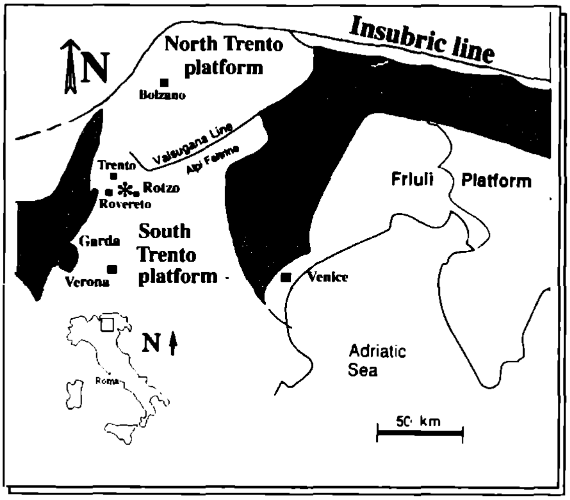


Figure 1. Paleogeographic sketch of the Trento platform and location map of the studied area (see asterisk) (redrawn from Zempolich, 1993).

Methods

Two sections were studied: the SVB section of a total thickness of 27.15m and outcropping in the Valbona Pass (near Folgaria, Rovereto area, Fig. 1), along a path through the southern side of M. Campomolon; and the PVB section of a total thickness of 39.9m located along an old road from the Valbona Pass through the M. Toraro.

Two analytical methods, based on microstratigraphy and taphonomy, were used in the study of shell beds of the Rotzo member: 1) field characteristics and laboratory study, and 2) photocomputer analysis.

Field Characteristics and Laboratory Study

A stratigraphic log was made in Valbona Pass, near Folgaria, analysing 67m of Upper Pliensbachian section immediately below the *Lithiotis* bioherms (the *Lithiotis* bioherms have not been considered in this study). Field analysis reveals a carbonate sequence formed largely of bioturbated (sometimes nodular) white or light-brown limestones (25 to 230cm in thickness), grey marls, locally fossiliferous, and black clayey-marls or clays 5-65cm thick. Textures, granulometry, physical and biological structures have been reported, as well as fossils, accessory minerals, diagenetic features and colour variations. Cyclicity and lithologic alternations of limestone and marls were recorded. The following characteristics were photographed: skeletal beds ≥1cm and burrowed limestones, shelter porosity of large bivalves shells or brachiopods, imbrication and patterns of turriculate gastropods, remains of echinoderms also in pavements of burrows, encrusting and boring, solitary coral remains and large benthic foraminifers, and coarse to fine skeletal debris characters.

Taphonomic features: In the Rotzo member more than 150 shell beds were analyzed (with photographs) considering skeletal concentrations related to physical agents (such as meteorologic phenomena) and biological agents (Table 1). Lateral continuity, sedimentary structures, pathways and taphonomic signatures (see also Brett, 1995; Anderson and McBride, 1996) of such physical agents as waves, storm flows, currents and burial of sediment, were considered, according to nine types of skeletal concentrations classified by Fürsich and Oschmann (1993) for the Jurassic of India. Taphonomic characters of the shell beds (tapho-facies) have been analyzed by means of thin sections and emery-polished rock slabs, with a minimum length of 20cm for each x-y-z axis. The microstratigraphy of skeletal debris and/or horizons of shells was employed to emphasize the vertical redistribution of grains caused by biological agents that partially or completely destroyed the original physical structures (Kidwell, 1991). The taphonomic categories are essentially those reported in Fernández López (1997a) for fossil associations of Mesozoic epicontinental platforms.

Trace fossils: This analysis included the search for and measurement of burrow systems, such as three-dimensional boxworks (tiered, regular, irregular; Frey et al., 1978), mazes (meander, regular and irregular), shafts, tunnels, turning chambers, and mounds of mud. Tiering of the same or different burrows was considered. Wall-pellet ichnofabrics (Frey et al., 1978; Anderson and Droser, 1998) and structure of the burrow lining were also observed in thin-section. If boxworks are preserved on slabbed rock surfaces, orientation and shape have been considered (if present they are generally well preserved from the coarse-grained debris inside the burrow), following the methodology of Anderson and Droser (1998); these authors show a diagram (their Fig. 6) with inclinations and shapes of burrows in cross-section. Architectural style of burrows can provide additional information useful in characterizing depositional environments and can be integrated with taphonomic analysis techniques to refine sedimentary interpretations.

Photocomputer Analysis

All photographs of tapho-layers were imported as images into a PowerPC computer by digital photo camera, scanner or Photo CD format (slides converted mainly in the 1536x1024 format). A PowerPC (PowerMacintosh) image-processing package was used for graphic retouching (brightness/contrast, levels and curves, B/N conversion, filters, crop of undesirable area; rotate and scale). All significant details of such figures were retraced using painting software (Monaco, in press). Text, numeric and graphic fields, histograms and specific fields for taphonomy (for each field: n=absent; r=rare; c=common; a=abundant; d=dominant) were recorded in a database in which, by drag and drop technique, all the figures previously mentioned have been pasted together with retraced details and diagrams (Monaco, in press).

Results

Microstratigraphic and taphonomic analysis of SVB (27.15m thick) and PVB (39.90m) sequences from the Pliensbachian in the Valbona area reveals that of a total of 450 studied layers of fossiliferous limestones and marls from both sections, 345 had a clear taphocharacter (151 for SVB, 194 for PVB). The total number of such layers is significant, because nearly 70% of studied beds show taphocoenosis, skeletal fabrics and features referable to physical and biological activities. Only the lower part of studied PVB section showed dark clays and some limestones and marly-limestones that were unfossiliferous and not bioturbated and, consequently, they could not be utilized in this study.

Skeletal concentrations (SkC) that show biological re-organization are interpreted as deposits formed by physical or biological agents (Kidwell, 1991) and they have been analyzed using methods previously described. Sedimentary fabrics and taphonomic characters are described and then discussed below.

Skeletal Concentrations Due to Physical and Biological Agents

SkC 1: Bimodal-oriented Gastropods in Mud-supported Fabric (Wackestones of Fair-weather Waves).

Skeletal concentrations, 5-8cm thick, of 80-90% high-spired gastropods (Nerineacea) show bimodal orientation of shells in plan-view (Fig. 2). Scattered bivalves are rare (10-20%), valves are disarticulated and poorly fragmented and without orientation in plan-view. No shelter porosity has been observed in side-view. Gastropods generally abraded or broken, although some fresh shells also occur, as indicated for several gastropod concentrations in Jurassic of India (Fürsich and

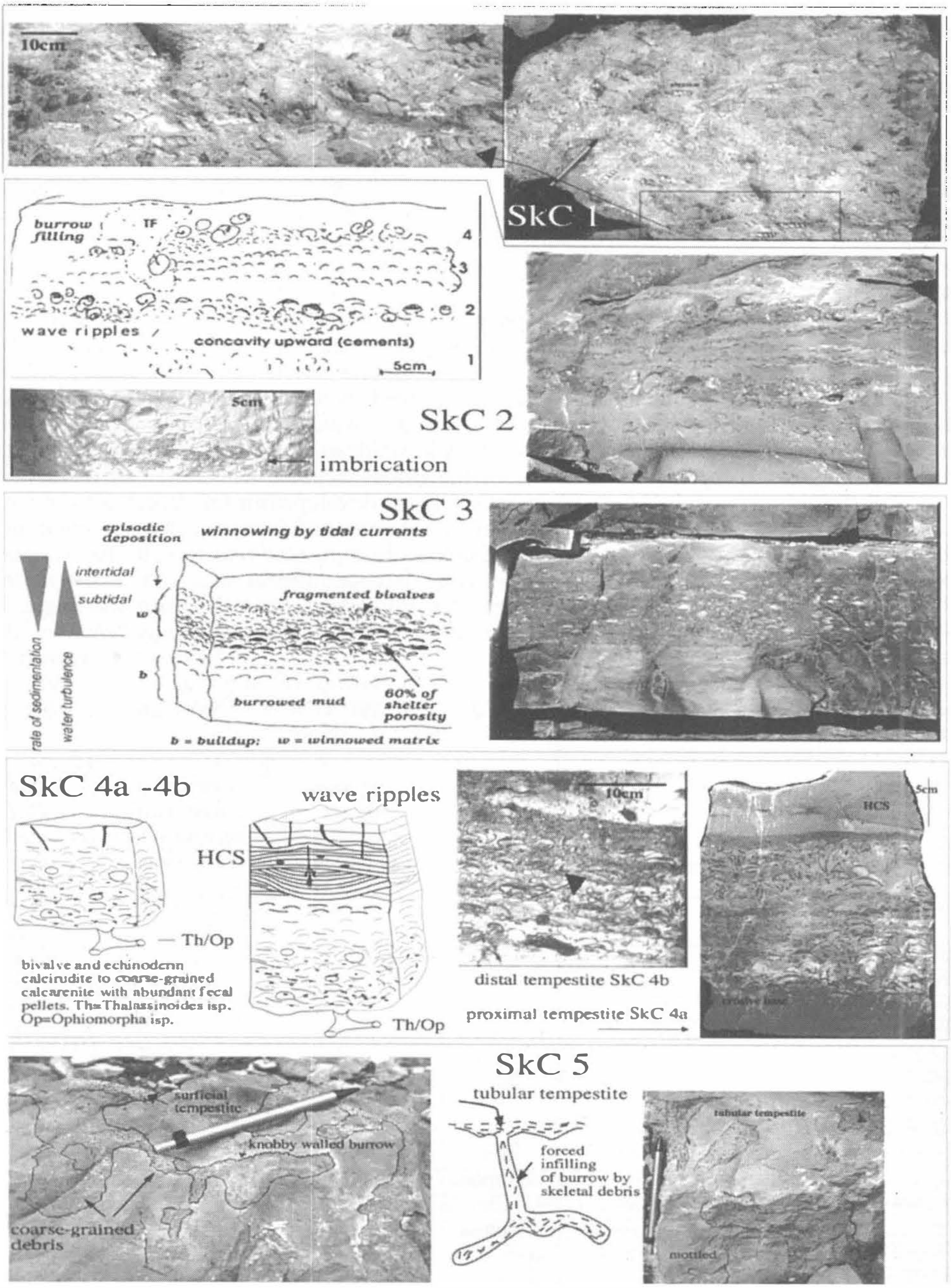
Oschmann, 1993). No encrusting or boring phenomena have been noted. This fabric is well developed in the middle-upper part of PVB section.

Physical agents: Abrasion, breakage and bimodal orientation of shells suggest fair-weather wave concentration on a shallow-water, lagoonal sea-bottom, where mud-supported fabric was formed under normal wave regime inducing abrasions on shell before burial, but without transport of skeletal material. Weak winds, characteristic of fair-weather periods, are incapable of generating shelf currents or waves with sufficient strength to entrain skeletal material or fine sand (Snedden et al., 1988). Locally, abnormal concentrations (grouping) of non-abraded gastropod shells are probably due to rolling action of dead shells in shallow (few cm deep) depressions on the lagoonal floor, e.g., between ripple crests (Fig. 3) or around carbonate mud mounds formed by crustacean decapods.

Biological agents: Burrowing and taphonomic distortion (predation, chewing, crumbling) are largely developed, inducing discontinuous shapes on gastropods concentrations and inducing strong vertical reworking of skeletal fragments by predators and scavengers (fish and/or crustacean activity?). Product of taphonomic distortion as fine-grained, skeletal debris infills small depressions developed around cone-shaped mounds produced by burrowing shrimps. Such mounds, probably related to the excavation-excrement activity of thalassinidean crustacean decapods, constitute a highly irregular topography on the sea-floor of the Rotzo member with relief of 10-20cm. This biological activity, indicated by several burrow entrances developed in the depressions around the mounds, is also well documented in modern carbonate shelves of subtropical, shallow environments during fair-weather conditions (mounds of endobenthic shrimps, *Callianassa* spp.: Braithwaite and Talbot, 1972; Wanless et al., 1988; Curran, 1994 and references therein; *Callianassa subterranea* in the North Sea: Stamhuis et al., 1997; burrows of *Callianassa major*: Weimer and Hoyt, 1964; burrows of thalassinidean shrimps *Callianassa californensis* or *Upogebia pugettensis* of British Columbia: Swinbanks and Luternauer, 1987).

skeletal concentrations:	SkC 1	SkC 2	SkC 3	SkC4a-b	SkC 5
PHYSICAL AGENTS					
abrasion, truncation:	very abundant	common	common	common	rare
dissolution:	rare	absent	absent	common	rare
disarticulation:	rare	dominant	common	very abundant	abundant
reorientation:	common	common	common	abundant	very abundant
crumbling:	absent	rare	common	common	very abundant
grouping:	rare	common	common	very abundant	abundant
winnowing:	rare	common	very abundant	common	rare
concavity-upward	-	common	rare	common	common
concavity-downward:	-	abundant	very abundant	abundant	common
filling with mud:	abundant	abundant	rare	common	common
shelter porosity:	-	rare	very abundant	rare	-
BIOLOGICAL AGENTS					
incrustation:	absent	rare	no	rare	rare
aerobic biodegradation:	rare	common	rare	rare	rare
concretion	rare	rare	rare	rare	rare
tapho-distortion (boring)	absent	common	no	rare	common
biologic reorientation:	common	common	rare	abundant	rare
predation, chewing:	common	common	surficial	abundant	abundant
biologic accumulation:	abundant	abundant	rare	abundant	abundant
burrow casts:	common	common	rare	common	very abundant
crumbling:	common	abundant	surficial	common	rare
bioadvection:	rare	common	rare	common	common

Table 1. Taphonomic characteristics of skeletal concentrations (see text).


10cm
SkC 1
burrow filling
TF
4
3
2
1
wave ripples
concavity upward (cements)
5cm
5cm
imbrication
SkC 2
SkC 3
episodic deposition
winnowing by tidal currents
intertidal
subtidal
rate of sedimentation
water turbulence
w
b
fragmented bivalves
burrowed mud
60% of shelter porosity
b = buildup; w = winnowed matrix
SkC 4a -4b
wave ripples
HCS
Th/Op
bivalve and echinoderm calcirudite to coarse-grained calcarenite with abundant fecal pellets. Th=Thalassinoides isp. Op=Ophiomorpha isp.
Th/Op
10cm
distal tempestite SkC 4b
proximal tempestite SkC 4a
5cm
HCS
erosive base
surficial tempestite
knobby walled burrow
coarse-grained debris
SkC 5
tubular tempestite
forced infilling of burrow by skeletal debris
tubular tempestite
mottled

SkC 2: Bioclast-supported Packstones (Combined Flow Concentrations)

These common deposits (3-5cm thick) are made up of concentrations of disarticulated and broken bivalve, gastropod and brachiopod shells, mixed with abundant peloids. Shells are stacked reciprocally, generally convex-up, but without shelter porosity. Erosive bases, wave ripples and cross-lamination are rare. Nerineacea show unimodal orientation and imbrication occurs frequently (Fig. 2). Locally, concentrations of whole brachiopods have been found (almost floatstones). Bioclast-supported fabrics are well developed in both PVB and SVB sections, particularly in the upper part of PVB section.

Physical agents: The motion associated with normal waves and currents has been called combined flow (Harms, 1969). As indicated by several authors that have studied modern lagoons and shelves (Swift et al., 1983; Snedden et al., 1988), combined flow produces a total stress on the bottom of greater magnitude than the simple sum of wave and current stresses (Grant and Madsen, 1986). This interactive flow induces sediment transport alongshore as a function of both waves and currents, which often interact in a non-linear fashion (Snedden et al., 1988). The concentrations of skeletal remains in the Rotzo member with this characteristic bioclast-supported fabric, showing no erosive base, coarse-grained wave ripples or cross-bedding (as Hummocky cross-stratification (HCS)), are not tempestites, but probably were formed from combined flow agents. Generally, valves are disarticulated but not strongly fragmented (the sorting is moderate). Species diversity is high but of the same lagoonal setting. In fact, microstratigraphy and side-view observations reveal that benthic fauna derives from the same area and no mixing with outside skeletal remains (such as crinoids) has occurred, indicating a short transport due to a long-term but feeble intensity mechanism (alongshore current concentrations).

Biological agents: As for SkC 1, deep burrowing. and taphonomic distortion (predation, chewing, crumbling) are commonly represented (Figs. 1, 2). Taphonomy and microstratigraphic observations suggest that dead shells are crushed by bottom-feeding organisms, or they have been exhumed by bioadvection. Reorientation below the sea-floor by bulldozing organisms or scavengers also occurs (Fig. 3). Intense biogenic reworking produces changes in thickness and lateral continuity of skeletal beds. It was probably caused by the activity of predators and scavengers in shell-gravel pavements (fish, crustaceans and other predators; see live/dead interactions on shell-gravel substrates, in Kidwell, 1991). These biological activities, very intense in a subtidal environment such as the Rotzo lagoon, induced rapid breaking of shells into small pieces, and produced biological transport (vertically and/or horizontally) of particles and whole shells within a soft substrate from the historic to the active layer, before redistribution by such physical agents as combined flow (see the description of biological reworkings in Bromley, 1996; Kidwell, 1991; and paleobiology of subtidal ichnocoenoses in Curran, 1994). Depressions around mounds formed by excavation-excrement activity of thalassinidean crustacean decapods are commonly preserved and they are filled by such skeletal concentrations.

SkC 3: Convex-up Oriented Bivalves Grainstones with Shelter Porosity (Winnowed Beds with Stacking by Tidal Currents)

Monospecific shell beds, 6-30cm thick, are rare but locally very representative of the lower parts of both PVB and SVB sections. Bivalves are normally disarticulated, but never fragmented, and their stacking is orderly and progressive (from floatstone to rudstone upwards, no erosive base, and gradation). Taphonomic observations suggest that 90-100% of stacked shells are convex-up, and shelter porosity is always developed (transparent, sparry calcite cement below the shells, Fig. 2). These deposits have great continuity and maintain the same thickness in the field laterally for many hundreds of metres without facies changes. A thin (2-6cm thick), cross-bedded muddy layer, locally bioturbated, locally overlies shell concentrations.

Figure 2. Skeletal concentrations (SkC) created by physical agents, Rotzo member, *Lithiotis* facies, Pliensbachian, Valbona pass. SkC 1: bimodal-oriented gastropods in a mud supported fabric (fair-weather waves). Note the imbrication. SkC 2: bioclast-supported packstones (combined flow concentrations); SkC 3: convex-up oriented shell grainstones with shelter porosity largely developed due to winnowing by tidal waves; SkC 4a-b: erosive bioclast-supported rudstones to packstones (proximal to distal tempestites); SkC 5: peloidal and skeletal packstones as forced infillings of subsurface burrows (tubular tempestites). See text for explanations.

Physical agents: Onshore-directed waves, induced by tidal cycles, produced on the lagoonal sea-floor of the Rotzo area a progressive reworking of muddy substrates containing a monospecific bivalve infauna taphocoenosis (probably semi-infaunal, R. Posenato, personal commun.). It is possible that during a sea-level lowstand repetitive water motion produced by tidal currents induced a slow but continuous winnowing of the mud and fine-grained particles between dead shells, without sufficient energy to collect shells and transport them (see substrate analysis of Goldring, 1995; Fernández-López, 1997b). Currents promoted a simple re-orientation of dead, autochthonous bivalve fauna, at the same time winnowing the finer fraction, and accumulated whole shells on the sea-floor with slow deposition during abating wave energy (Fig. 2). The final product is a progressive concentration (locally very thick) of unbroken, stacked shells in which shelter porosity and sparry calcite cements may dominate due to winnowing.

Biological agents: These deposits are very poorly burrowed. Pre-erosional events such as large burrow tunnels and side chambers of crustacean decapods (mainly *Thalassinoides*) observed in SkC4 are lacking here (Fig. 3). Post-event colonizers are represented by a few opportunistic taxa that rework only a few centimeters of shell concentrations. In most cases deep burrowing did not occur and significant taphonomic distortion phenomena such as bioerosion and encrusting have not been recorded (Table 1). This was probably due to the absence of suitable conditions for predators and adverse paleoenvironmental conditions for many burrowers, borers and encrusters on the sea-floor after the winnowing event, and before the raising of sea-level.

SkC 4a and b: Erosive, Graded Bioclast-supported Rudstone to Packstones (Proximal to Distal Tempestites)

These deposits, 10-40cm thick, are similar in composition to bioclast-supported packstones (SkC 2), also consisting of skeletal concentrations of bivalves, brachiopods, gastropods and other lagoonal fauna, but differing in many sedimentary and taphonomic characters. Differences are: outside skeletal sand such as crinoids mixed with other fauna, sharp and erosive bases, small flute casts (including gutter casts; see Aigner, 1985), a skeletal lag interval locally overlain by HCS (Monaco, 1994) and coarse-grained wave ripples; moreover, a high rate of taphonomic distortion of shells (*sensu* Fernández-López, 1997a; 1997b) is present, including predation and chewing (Table 1). SkC 4a shows more coarse-grained skeletal fragments (coarse sand) than SkC 4b which is made up mainly of peloidal and fine-grained skeletal (silt to fine sand) remains, often bioturbated on the top. Many skeletal deposits are monospecific concentrations of echinoderms (crinoidal or echinoid packstones). Locally, buried burrow tunnels of crustacean decapods, 5 to 12cm in width, have been exhumed and then filled with bivalve skeletal remains (cut and fill concentrations, Fig. 2).

Physical agents: Fining-upward sequences with a basal skeletal lag, followed by an HCS interval (although very reduced in thickness; see Fig. 2), wave ripples and bioturbated mud blankets, seem to indicate a tempestite or storm-induced sediment transport across the shelf in deeper parts of the lagoons or the ramp (Aigner, 1985; Monaco, 1992; Masetti et al., 1996). Skeletal storm lags are developed locally with characteristic shaped coarse-grained wave ripples, with sharp bases and asymmetry of the crests (Fig. 3); and these deposits commonly occur in close lateral association with hummocky cross-stratification (Leckie, 1988). Wave conditions producing such coarse-grained wave ripples are not the result of particularly large storms (wave heights can range from 1-9m in shallow water depths, generating crest spacings of 0.2-3.0m; see Leckie, 1988). Shoreward- (crinoid concentrations) or seaward-flowing current induced by strong meteorological perturbations such as winter storms or moderate hurricanes promoted near bottom shear-stress and sediment-laden flow that caused skeletal transport for long distances, producing proximal to distal HCS deposits in the shallow carbonate shelf or in the deeper part of the ramp, respectively (see the extensive literature on tempestites and HCS deposits reported in Aigner, 1985; Duke, 1985; Monaco, 1992; 1994; 1995; Myrow and Southard, 1996).

Biological agents: Pre-event burrows are irregular boxworks of crustacean decapods (*Thalassinoides* isp.). Post-event colonizers occur mainly in SkC 4b facies and are such opportunistic taxa as *Skolithos* and *Chondrites*. Deep burrowing as shafts of *Ophiomorpha* involves cross-stratification. Shell breakage, reorientation (by bulldozing organisms?) and crushing of dead shells by bottom-feeding organisms provide evidence for biogenic accumulation or dispersion (Fig. 3). Under conditions of low accumulation rates (inter-storms), bioeroders and encrusters developed, as testified by reduced thicknesses of internal walls of brachiopod and bivalve shells due to intense boring activity (algae, microendolithic bacteria and sponges).

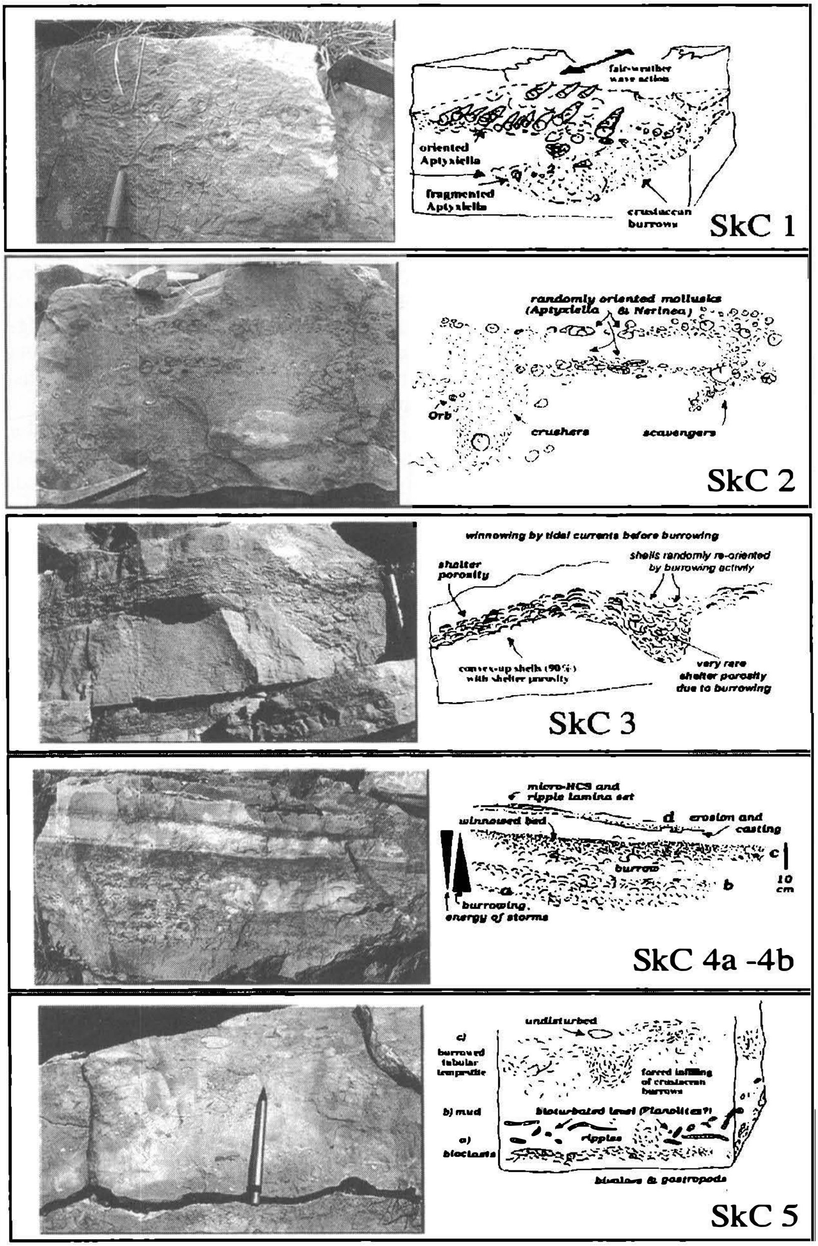


Figure 3. Biological agents involving physical concentrations, Pliensbachian, *Lithiotis* facies of the Rotzo member, Valbona pass. Burrowing and taphonomic distortion (predation, chewing, crumbling, bioadvection) are largely developed in lime muds, causing breakage and reorientation of skeletal remains deposited by physical agents (compare Fig. 1). Biological activity follows the physical and reworks particles inducing anomalous concentrations by predators and scavengers (see text for explanation; for terms of skeletal concentrations see Fig. 1)

SkC 5: Peloidal and Skeletal Packstones as Forced Infillings of Subsurface Burrows (Tubular Tempestites)

Many muddy layers in the Rotzo member show elliptical spots, from 3 to 8cm wide in side-view, made of peloidal, bivalve and/or echinoderm packstones, that are clearly distinguishable from the neighbouring muddy fabric. These spherical, coarse-grained, skeletal concentrations are sections of a three-dimensional boxwork that represents a coarse-grained burrowing system, developed 50 to 120cm beneath the sediment surface. A surficial, thin, tempestite packstone (2-3cm thick), made of the same skeletal debris, blankets the muddy layer and connects the three-dimensional boxwork with burrow entrances (Fig. 3).

Physical agents: Wanless et al., 1988 recently demonstrated that three-dimensional or tubular coarse-grained concentrations of shells in modern, tropical carbonate, storm systems are forced infilling of dwelling burrow tubes (often abandoned) and side chambers of crustacean decapods such as the burrowing shrimp *Callianassa* (on burrows of *Callianassa major* as geologic indicators of littoral and neritic environments, see Weimer and Hoyt, 1964). Infillings occur during repeated storms or hurricanes in a subtidal muddy environment, forming the so-called tubular tempestites (for the effects of the moderate hurricane Kate across the Caicos platform, see Wanless et al., 1988). During the intense wave action and currents of the storm, coarse and heavier shells or shell fragments fall into the burrow by gravity or pumping action due to rapid fluctuations in the hydrostatic pressure through the burrows (Webb and Theodor, 1968). The remaining coarse material that filled open burrow entrances and depressions between cone-shaped excavation-excrement mounds if they are not flattened by the storm action (Wanless et al., 1988).

Biological agents: Locally, these coarse-grained deposits are burrowed after the infilling storm event. Many halo burrows of crustacean decapods have been seen overlapping tubular tempestites. Burrowing after abatement of storm energy mainly involves surficial tunnels, destroying prevalently surficial tiers.

Taphonomic Recurrences and Conclusions

After analysing all taphofacies of the Early Jurassic Rotzo member, it is clear that some taphonomic classes repeatedly occur and common characteristics resulting from physical and biological agents can be grouped in four genetic types as follows:

1. Disarticulation, grouping, dispersion. This category constitutes the dominant taphonomic recurrence of all considered skeletal concentrations (35% of the total). Characteristics may be summarized as follows: • transport with or without selection by combined flow currents, winter storms, and hurricane events; • mixed fossils coming from remote areas (biogeographic dispersion), each showing its own taphonomic history; • turriculate gastropods aligned due to wave action or sea-floor flow; • stacked bivalves produced by winnowing of mud due to repetitive events, and accumulated over longer periods (hiatal concentrations of Kidwell, 1991); • biogenic dispersion by predators, scavengers and surficial burrowing.
2. Abrasion, truncation, filling with mud. The taphofacies analyzed here represents 10% of all categories. Characteristics are: • attached shells on firm substrate truncated by currents; • mechanical fracturing and abrasion (cigar-shape) of shells under high turbulence and wave flow regimes; • cavities (mainly of brachiopods) filled with mud by aspirant currents; • high turbulence + accumulation + instability of shells = scarce filling with mud; • biogenic activity on the sea-floor by predators of living shells (infauna), mud infilling by passive actions of deep burrowers.
3. Crustacean burrowing. This is an important category of studied taphonomic features, involving many of the strata considered (50% of total). Characteristics can be grouped as follows: • surficial and deep burrowing systems (mainly tiered mazes, irregular boxworks) such as *Ophiomorpha* isp. and *Thalassinoides* isp., referable to the feeding activity of crustacean decapods and analogous to modern thalassinidean shrimps (*Callianassa* or *Upogebia*); • simple cylindrical tubes 80cm in length, analogous to those created by modern carnivorous crustacean predators, such as mantis shrimps or stomatopods (Braithwaite and Talbot, 1972; Bromley, 1996; Hof and Briggs, 1997), or other unidentified burrowers.
4. Taphonomic distortion. Finally, several minor features (5%) can be summarized as follows: • borings, fractures, perforation and deformation of shells by predation activity of carnivorous mollusks and necrophagous feeders; • reduced thickness of internal wall of brachiopod and bivalve shells (algae, microendolithic bacteria and sponges) developed between storms.

Biofabric and taphonomic signatures of skeletal concentrations record different histories based on physical and biological agents. They represent, therefore, a useful tool in palaeoenviromental analysis but they are best used in an integrated approach combining taphonomy, ichnology and sedimentology.

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References

ACCORSI BENINI, C. 1979. Lithioperna, un nuovo genere fra i grandi Lamellibranchi della facies a "*Lithiotis*". Morfologia, tassonomia ed analisi morfofunzionale. Bollettino della Società Paleontologica Italiana, 18 (2):221-257.

____. 1980. Microstruttura di Opisoma Stoliczka, 1871 lamellibranco eterodonte della facies a "*Lithiotis*" (Liassico). Annali Universita Ferrara, sez.9, v.6:339-351.

____. 1981. Opisoma Stoliczka, 1871 lamellibranco eterodonte della facies a "*Lithiotis*" (Giurassico inf., Liassico). Revisione morfologica e tassonomica). Bollettino della Società Paleontologica Italiana, 20 (2):197-228.

____. 1985. The large Liassic bivalves: symbiosis and longevity. Palaeogeography, Palaeoclimatology, Palaeoecology, 52:21-33.

____. AND C. BROGLIO LORIGA. 1974. Isognomon (Mytiloperna) v. Ihering 1903 e Gervilleioperna Krumbeck 1923 fra i grossi lamellibranchi della facies a "*Lithiotis*" del Veneto. Acc.Naz. Lincei, Rend. s.8, v.57 (3-4):233-243.

____ AND ____. 1977. *Lithiotis* Gümbel, 1871 e Cochlearites Reis, 1903 I° revisione morfologica e tassonomica. Bollettino della Società Paleontologica Italiana, 16 (1):15-60.

____ AND ____. 1979. Facies a "*Lithiotis*", Calcari Grigi (Giurassico inferiore). Altipiano dei Sette comuni (Vi); M.Lessini occidentali (Vr). Paleobenthos, 1° Conv. Ecologia e paleoecologia delle comunità bentoniche, Ferrara 8-12 Ott. 1979, libr. guida escurs.:18-27.

AIGNER, T. 1985. Storm depositional systems: Dynamic stratigraphy in modern and ancient shallow-marine sequences. *In* G.M. Friedman, H.J. Neugebauer and A. Seilacher (eds.), Lecture notes in Earth Sciences: Springer-Verlag, NewYork, v.3, 174p.

ANDERSON, B.G. AND M.L. DROSER. 1998. Ichnofabrics and geometric configurations of Ophiomorpha within a sequence stratigraphic framework: an example from the Upper Cretaceous US western interior. Sedimentology, 45:379-396.

ANDERSON, L.C. AND MCBRIDE. 1996. Taphonomic and palaeoenvironmental evidence of Holocene shell-bed genesis and history on the northeastern Gulf of Mexico shelf. Palaios, 11:532-549.

BERTI CAVICCHI, A., A. BOSELLINI, AND C. BROGLIO LORIGA. 1971. Calcari a *Lithiotis* problematica Gümbel o calcari a "*Lithiotis*"? Mem. Geopaleont. Univ. Ferrara, 3 (1):41-53.

BOSELLINI, A. AND C. BROGLIO LORIGA. 1971. I "Calcari grigi" di Rotzo (Giurassico inferiore, Altopiano di Asiago) e loro inquadramento nella paleogeografia e nella evoluzione tettono-sedimentaria delle Prealpi Venete. Annales Universita di Ferrara s.9, v.5 (1):1-61.

BRAITHWAITE, C.J.R. AND M.R. TALBOT. 1972. Crustacean burrows in the Seychelles, Indian ocean. Palaeogeography, Palaeoclimatology, Palaeoecology, 11:265-285.

BRETT, C.E. 1995. Sequence stratigraphy, biostratigraphy, and taphonomy in shallow marine environments. Palaios, 10:597-616.

BROGLIO LORIGA, C. AND A. FUGAGNOLI. 1997. Nerineacean dominated communities from Lower Jurassic shallow water carbonates (Calcari Grigi, Trento platform; northeastern Italy). GEOITALIA 1° forum FIST, Bellaria, 5-9 ott. 1997, vol.2, sess.6:67-68.

____ AND NERI C. 1976. Aspetti paleobiologici e paleogeografici della facies a "*Lithiotis*" (Giurese inf.). Rivista Italiana di Paleontologia e Mineralogia, 82 (4):651-706.

BROMLEY, R.G. 1996. Trace fossils. Biology, taphonomy and applications. Chapman & Hall Publ. London (second edition): pp.361.

CITA, M.B. 1959. Osservazioni stratigrafiche sul Lias del M.Baldo. Rivista Italiana di Paleontologia e Stratigrafia, 65, n°4, Milano.

CURRAN, H.A. 1994. The palaeobiology of ichnocoenoses in Quaternary, Bahamian-style carbonate environments: the modern to fossil transition, p. 83-104. *In* Donovan, Stephen K. (eds.) The Palaeobiology of Trace Fossils: New York, John Wiley and Sons, 308.

DE ZIGNO. 1856-1885. Flora fossilis formationis oolithicae. Due volumi, Padova.

DUKE, W.L. 1985. Hummocky cross-stratification, tropical hurricanes, and intense winter storms. Sedimentology, 32:167-194.

FERNÁNDEZ LÓPEZ, S. 1997a. Ammonites, clinos tafonómicos y ambientes sedimentarios. Revista Española de Paleontologia, 12 (1):102-128.

____. 1997b. Ammonites, ciclos tafonómicos y ciclos estratigráficos en plataformas epicontinentales carbonáticas. Revista Española de Paleontologia, 12 (1):151-174.

FREY, R.W., J.D. HOWARD, AND W.A. PRYOR. 1978. Ophiomorpha: its morphologic, taxonomic, and environmental significance. Palaeogeography, Palaeoclimatology, Palaeoecology, 23:199-229.

FUGAGNOLI, A. 1996. On the occurrence of Duotaxis metula Kristan (foraminifera) in the Lower Jurassic (Calcari grigi, Venetian Prealps, Italy). Revue de Paléobiologie, 15 (2):385-392.

FÜRSICH, F.T. AND W. OSCHMANN. 1993. Shell beds as tools in basin analysis: the Jurassic of Kachchh, western India. Journal of the Geological Society of London, 150:169-185.

GOLDRING, R. 1995. Organisms and the substrate: response and effect, p. 151-180. *In* D.W.J. Bosence & P.A. Allison (eds.), Marine Palaeoenvironmental analysis from Fossils, Geological Society special publication n°83.

GRANT, W.D. AND O.S. MADSEN. 1986. The continental shelf bottom-boundary layer. Annual Review Fluid Mechanics, 18:265-305.

HARMS, J.C. (1969). Hydraulic significance of some sand ripples. Geological Society of America Bulletin, 80:363-396.

HOF, C.H.J. AND D.E.G. BRIGGS. 1997. Decay and mineralization of mantis shrimps (Stomatopoda: Crustacea) - A key to their fossil record. Palaios, 12:420-438.

KIDWELL, S.M. 1991. Taphonomic feedback (live/dead interactions) in the genesis of bioclastic beds: keys to reconstructing sedimentary dynamics, p. 268-282. *In* G.Einsele, W.Ricken & A.Seilacher (eds.) Cycles and Events in Stratigraphy, Springer-Verlag, Berlin.

LECKIE, D. 1988. Wave-formed, coarse-grained ripples and their relationship to hummocky cross-stratification. Journal of Sedimentary Petrology, 58 (4):607-622.

MASETTI, D., M. CLAPS, M. AVANZINI, A. GIACOMETTI AND P. PIGNATTI. 1996. I Calcari Grigi della Piattaforma di Trento (Lias inferiore e medio, Prealpi Venete). 78° riun. est. Soc. Geol. It.: la Geologia delle Dolomiti, S. Cassiano (BZ) 19-21 sett. 1996, guida all'escursione: pp. 48.

MYROW, P.M. AND J.B. SOUTHARD. 1996. Tempestite deposition. Journal of Sedimentary Research, 66 (5):875-887.

MONACO, P. 1992. Hummocky cross-stratified deposits and turbidites in some sequences of the Umbria-Marche area (central Italy) during the Toarcian. Sedimentary Geology, 77:123-142.

____. 1994. Hummocky cross-stratifications and trace fossils in the Middle Toarcian of some sequences of Umbria-Marche Apennines. Geobios, M.S.17:679-688.

____. 1995. Relationships between trace-fossil communities and substrate characteristics in some Jurassic pelagic deposits in the Umbria-Marche basin, central Italy. Geobios, M.S.18:299-311.

____. in press. Macintosh computer database as a tool to investigate taphonomy and events in carbonate platform environments. Paleopelagos, sp. vol. (1999), Roma.

NAUSS, A.L. AND P.L. SMITH. 1988. *Lithiotis* (bivalvia) bioherms in the Lower Jurassic of east-central Oregon, U.S.A. Palaeogeography, Palaeoclimatology, Palaeoecology, 65:253-268.

SNEDDEN, J.W., D. NUMMEDAL, AND A.F. AMOS. 1988. Storm- and fair-weather combined flow on the central Texas continental shelf. Journal of Sedimentary Petrology, 58 (4):580-595.

STAMHUIS, E.J., C.E. SCHREURS, AND J.J. VIDELER. 1997. Burrow architecture and turbative activity of the thalassinid shrimp Callianassa subterranea from the central North Sea. Marine Ecology Programs Series, 151:155-163.

SWIFT, D.J.P., A.G. FIGUEIRODA, Jr., J.L. FREEMAN, AND G.F. OERTEL. 1983. Hummocky cross-stratification and mega-ripples. A geologic double standard? Journal of Sedimentary Petrology, 53:1295-1318.

SWINBANKS, D.D. AND J.L. LUTERNAUER. 1987. Burrow distribution of thalassinidean shrimp on a Fraser delta tidal flat, British Columbia. Journal of Paleontology, 61 (2):315-332.

WANLESS, H.R., L.P. TEDESCO, AND K.M. TYRRELL. 1988. Production of subtidal tubular and surficial tempestites by hurricane Kate, Caicos Platform, British West Indies. Journal of

Sedimentary Petrology, 58 (4):739-750.

WEIMER, R.J. AND J.H. HOYT. 1964. Burrows of Callianassa major Say, geologic indicators of littoral and shallow neritic environments. Journal of Paleontology, 38 (4):761-767.

WEBB, J.E. AND J. THEODOR. 1968. Irrigation and submerged marine sands through wave action. Nature, 220:682-683.

WESLEY, A. 1956. Contributions to the knowledge of the flora of the Grey Limestones of Veneto: Part I. Memorie Istituti di Geologia e Mineralogia Universita di Padova, v.19.

ZEMPOLICH, W.G. 1993. The drowning succession in Jurassic carbonates of the Venetian Alps, Italy: a record of supercontinent breakup, gradual eustatic rise, and eutrophication of shallow-water environments, p. 63-105. *In* R.G. Loucks and J.F. Sarg (eds.), Carbonate Sequence Stratigraphy - Recent Developments and Applications, American Association of Petroleum Geologists Memoir n°57, Tulsa.

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Stable Isotopic Signal of Carbon and Oxygen in Jurassic Marlstone-Limestone Rhythms (Italy, Central Apennines)

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Keywords: Early Jurassic, Rhythmic Bedding, Stable Isotopes, Benthic Foraminifers

Abstract: Rhythmic alternating burrowed limestone and marlstone beds were deposited in the Umbria-Marche Basin (Fiuminata, central Apennines, Italy) during the Middle Toarcian-Early Aalenian interval. The processes responsible for the formation of hemipelagic and pelagic marlstones/limestone rhythms may be cyclic environmental changes controlled by orbital variations.

We provide data for interpreting deposition processes of a 30 metres thick succession of pelagic marlstone/limestone rhythmic couplets (Calcari e Marne a Posidonia Unit = Calcari e Marne a Posidonia) as related to a Dilution Cycle and to a Proliferation Cycle defined here. Marl and calcareous beds, rich in thin-shelled bivalves, mainly differ in benthic foraminifers abundance and stable isotopic content: marlstones contain abundant foraminifers and show lighter $\delta^{13}C$ and $\delta^{18}O$ isotopic ratios with respect to limestones, which show paucity of benthic foraminifers and heavier $\delta^{13}C$ and $\delta^{18}O$ isotopic ratios.

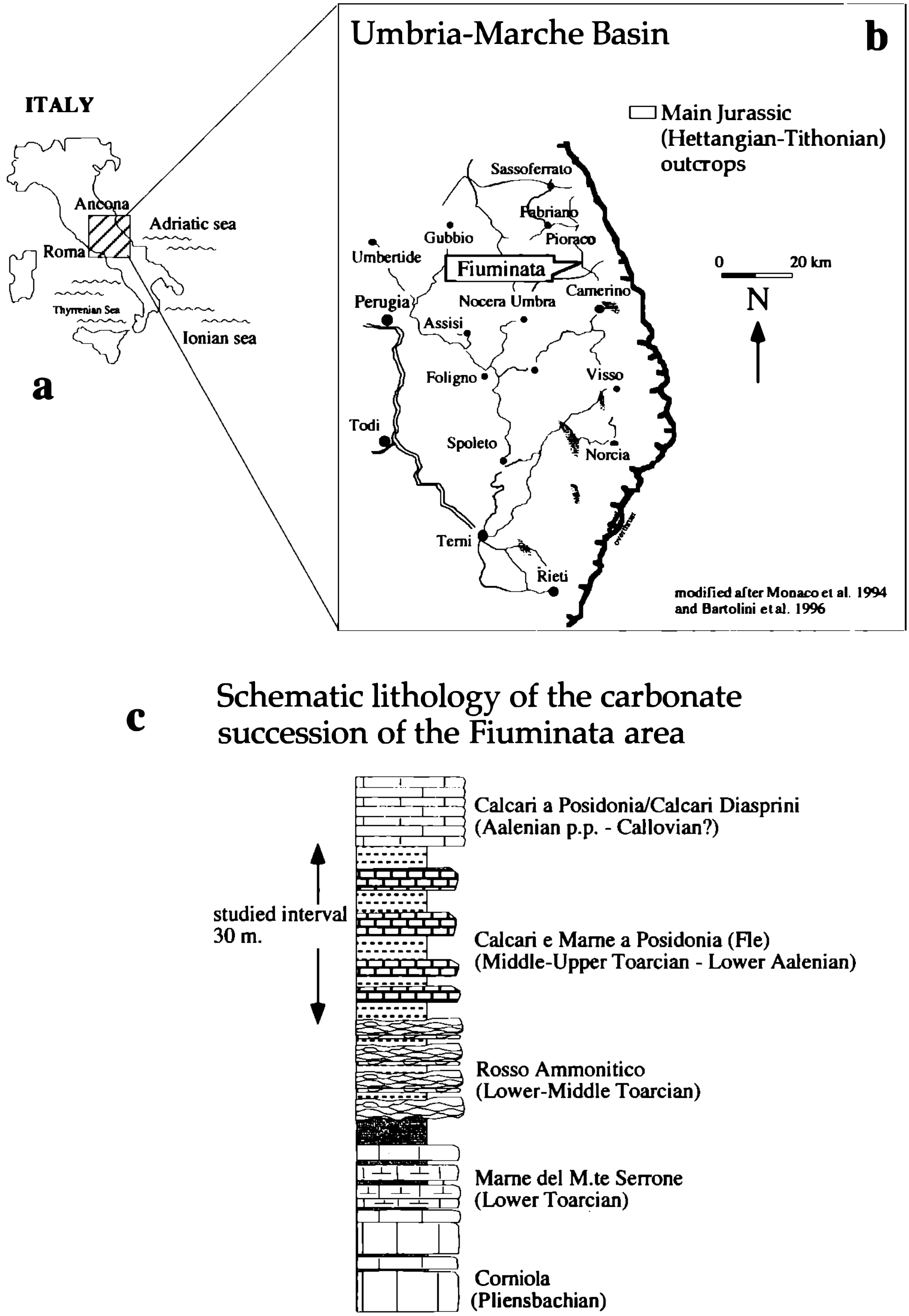
Isotopic curves and benthic faunal abundances covary rhythmically, enhancing climatic and lithological cyclicities. The mechanism which linked climatic changes to deposition processes seem to go from a dilution cycle, triggered by strong runoff on continents, to a proliferation-production cycle which also reflect the chemistry of water column.

Introduction

Lower Cretaceous to Paleogene pelagic formations consisting of rhythmic marlstone-limestone alternations are common in the Northern-central Apennines. Erba, (1992) and Mattioli, (1997), have focused the attention on calcareous nannoplankton as a key factor to understand marlstone-limestone alternations.

Nannofloral contrast between the limestone beds with schizospheres and the marly interbeds with coccoliths have, *a priori*, been linked to preservational conditions. Moreover the two sediment types reflect cyclic environmental changes combined with changes in the faunal content.

This study is focused on Lower Jurassic bivalve-bearing deposits that show dm-scale rhythmic marlstone/limestone couplets (Calcari e Marne a Posidonia unit, Conti and Cresta, 1982; Colacicchi et al., 1988; Cresta et al., 1988; Conti and Monari, 1992; Mattioli, 1997; Monaco and Morettini, in press). The area of study is Fiuminata (Pioraco, Macerata, Northern Apennines, Umbria-Marche Basin, Fig. 1). Marlstone/limestone rhythmic alternations have been approached sedimentologically, micropaleontologically and also through isotopic studies, and interpreted in terms of organic productivity, water salinity, and burial depth (Scholle and Arthur, 1980). Many authors have argued that differences in the isotopic signature of marlstones and limestones are primary (Barron et al., 1985; R.O.C.C. Group, 1986; Eicher and Diner 1989; Ricken, 1991). However, until recently, very few studies have investigated short term variations in carbon and oxygen stable isotope composition of Jurassic and Cretaceous rocks (Arthur et al., 1984). This work documents detailed $\delta^{13}C$ and $\delta^{18}O$ variations performed on 78 limestone and marlstone samples spaced between 5 and 90cm (over 20,000y), from a 30m thick section (Fig. 2).


ITALY
Ancona
Roma
Adriatic sea
Thyrrenian Sea
Ionian sea
a
Umbria-Marche Basin
b
Main Jurassic (Hettangian-Tithonian) outcrops
Sassoferrato
Fabriano
Pioraco
Gubbio
Umbertide
Fiuminata
Camerino
Perugia
Nocera Umbra
Assisi
Foligno
Visso
Todi
Spoleto
Norcia
Terni
Rieti
overthrust
0
20 km
N
modified after Monaco et al. 1994 and Bartolini et al. 1996
c
Schematic lithology of the carbonate succession of the Fiuminata area
studied interval 30 m.
Calcari a Posidonia/Calcari Diasprini (Aalenian p.p. - Callovian?)
Calcari e Marne a Posidonia (Fle) (Middle-Upper Toarcian - Lower Aalenian)
Rosso Ammonitico (Lower-Middle Toarcian)
Marne del M.te Serrone (Lower Toarcian)
Corniola (Pliensbachian)

Geological Setting

The studied unit crops out in the Fiuminata area, in the mountain of Colle Corno, after which the section is named. The area is located in the Internal Umbria-Marche Apennine fold and thrust belt. Mesozoic and Paleogene marine formations, deposited on the African continental margin since the Triassic, were involved in a late Tertiary compressive phase that formed partially thrusted and overturned folds showing NNW to SSE axial trends (Lavecchia and Pialli, 1980; Lavecchia, 1981). The extensional tectonic of Jurassic age, responsible for the thinning of the Apulian continental margin (Bernoulli and Jenkyns, 1974; Channell et al., 1979), was succeeded by this Tertiary compressive phase

The studied section consists of limestones, marly limestones, marlstones and clays, spanning the Early to Middle Jurassic. It is part of a complete sequence in the sense that all the units that make up the Umbria-Marche carbonate sedimentary successions are present in stratigraphic order in the area (see extensive literature on stratigraphic units in Cresta et al., 1988; Monaco et al., 1994). The section is located in an east vergent anticline structure containing Lower Jurassic units in the core, and Cretaceous-Paleogene units on the flanks.

The Colle Corno section (Fle, Fig. 1c) consists of the Calcari e Marne a Posidonia unit made up of marlstone (5-60cm) and limestone (5-80cm) couplets rich in thin-shelled bivalves (Fig. 2). Limestone beds are brown to grey, and consist of bivalve, radiolarian and nannofossil micrites (Mattioli, 1997) and calcarenitic intercalations, mainly shell-lag types (Fig. 2). The Calcari e Marne a Posidonia unit was dated as Middle Toarcian to Early Aalenian on the basis of ammonites and nannofossils (Mattioli, 1994; Mattioli, 1997; Monaco and Morettini, in press). This unit overlies the Rosso Ammonitico Umbro-Marchigiano and is topped by the Calcari Diasprini unit. The Colle Corno section displays a higher clay content than coeval calcareous units of the Umbria-Marche realm (Middle-Late Lias to Dogger in age, Conti and Cresta, 1982; Colacicchi et al., 1988), where significant clay contents are restricted to the Upper Pliensbachian to Middle Toarcian interval (Ortega-Huertas et al., 1993). In the Fiuminata area the interval from the Pliensbachian to Bajocian is over 400m thick (Fig. 1c). The usual thickness of the same interval for the "complete" sections ranges from 300 to 380m (Colacicchi et al., 1988; Parisi et al., 1998). The anomalous thickness seems to be the result of a continuous clay input, not observed in the Upper Toarcian–Aalenian of other Umbria-Marche sections.

The studied marlstone/limestone couplets occur in an overall thickness of ca. 30m, and crop out over a lateral extent of ca. 300m. No regional bed by bed correlation is possible, since the rhythmic bedding is seldom found in Toarcian-Aalenian outcrops of Umbria-Marche.

Analytical Techniques

We performed laboratory analyses of carbon and oxygen isotope composition and $CaCO_3$ content at bed scale, at the Laboratory of Isotope Geochemistry, Insitute of Mineralogy and Petrography of the University of Lausanne (UNIL).

Micropalaeontological analyses were based on 6 gr of washed residue that was separated into a fine (63-150 μm) and coarse (>150 μm) fraction. Residues were observed *in toto*, the foraminiferal assemblage was picked and the total number of specimens was considered. Radiolarians, ostracodes, brachiopods, bivalves, crinoid fragments, holothuroids, ophiuroids, echinoid spines were observed in a semiquantitative way, establishing relative abundance through five classes: rare, (R < 10%), common (C between 10 and 30%) and abundant (A > 30%). Some benthic foraminifer species were chosen according to those selected by Stam, 1986; and by Nocchi, 1992, that should give the best chronological, palaeobathymetric and palaeoenvironmental resolution; foraminifer distribution chart is shown in Fig. 3.

The $\delta^{13}C$ and $\delta^{18}O$ compositions of whole rock (32 marlstones and 46 limestones) samples were measured following the standard procedure of McCrea (1950). Isotopic ratios were obtained from 15-20mg of powdered sample that was digested in 2ml of 100% ortophosphoric acid (H_3PO_4)

Figure 1. a) Schematic location map of the Umbria-Marche basin. b) Umbria-Marche main Jurassic outcrops (modified after Monaco et al., 1994). c) The studied rhythmic sequence of Calcari e Marne a Posidonia Unit (Middle-Upper Toarcian - Lower Aalenian).

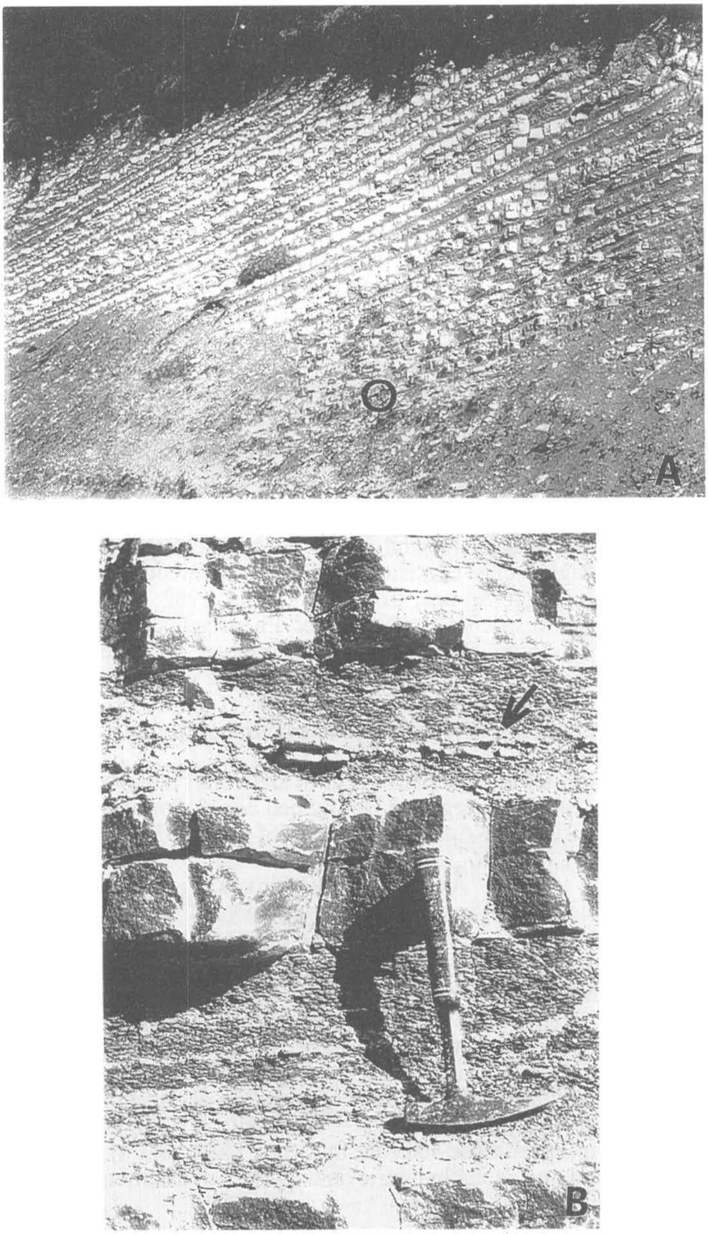


Figure 2.(A) The Colle Corno (Fle) stratigraphic section, Middle-Upper Toarcian - Lower Aalenian. Rhythmic marl/limestone bivalve-rich couplets (Calcari e Marne a Posidonia = Calcari e Marne a Posidonia Unit). Hammer inside circle of 35cm. (B) Detail of a marl-limestone couplet: rhythmic sedimentation is randomly interrupted by bivalve lags (see arrow).

per mill deviation from the PDB (Pee Dee Belemnite) standard. Reproducibility of sample measurements was checked on duplicate analyses, and is better than 0.05 ‰ for $\delta^{13}C$ and better than 0.1‰ for $\delta^{18}O$ ratios. Twenty out of 78 analyses were also performed on samples previously treated with sodium hypochloride (NaOCl), filtered and dried at 80°C (Charef and Sheppard, 1984)

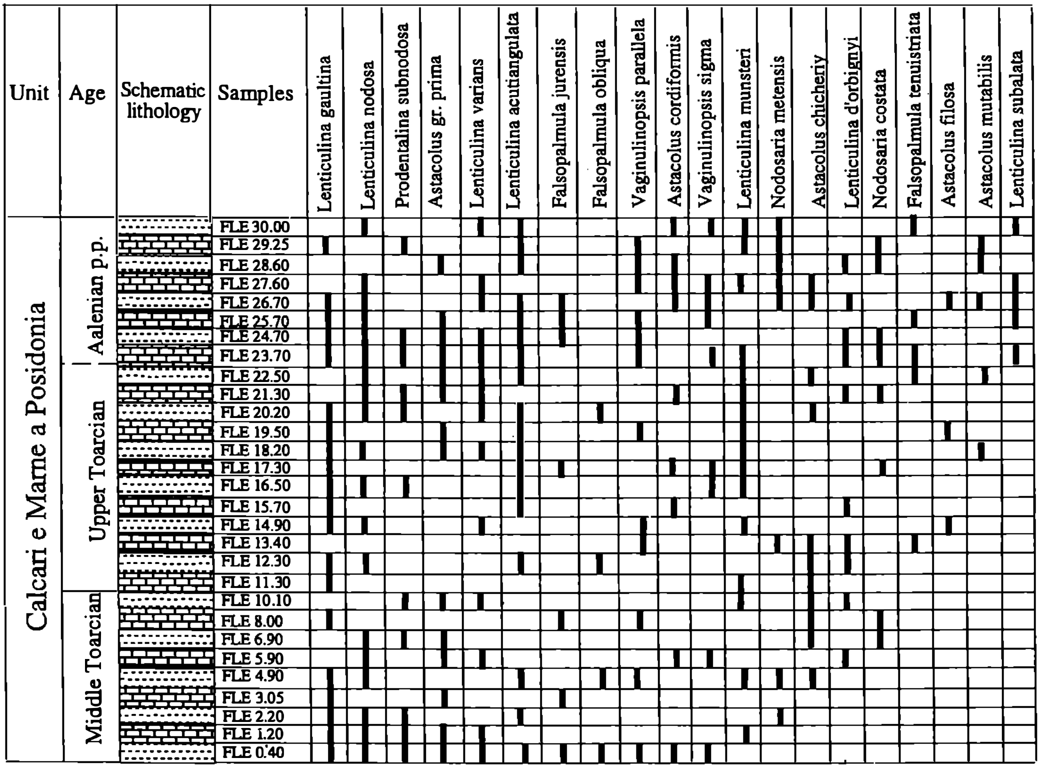


Figure 3. Distribution chart of selected foraminifera species of the Colle Corno (Fle) stratigraphic section (Calcari e Marne a Posidonia Unit), against schematic lithology and age. The species selection comes from the marly intervals, limestone showing extreme paucity of specimens. Note that *Falsopalmula Jurensis* and *Lenticulina d'orbigny* first occurrences mark the onset of the Middle Toarcian (Nocchi, 1992). *Eoguttulinas* and *Prodentalinas* are the most abundant genera and indicate a relatively deep outer-sheld (Stam, 1986; Nocchi, 1992)

in a bath at 25°C for 12 hours. The obtained carbon dioxide (CO_2) was analysed in a Finningan MAT 251 mass spectrometer. Results are reported according to the usual δ notation, expressed as to eliminate organic matter. Isotopic values obtained from the samples with and without treatment, differed by a value less than the analytical error.

The isotopic variations discussed in this paper are defined by 78 data points, with only a few samples deviating from the mean trend (Figs. 4, 5). The difference of carbon isotopic values between individual marlstones and limestones in a couplet (0.3-0.5‰) is 3 to 5 times larger than measurement reproducibility.

Limestone and Marlstone Characteristics

Palaeontology

Thin-shelled bivalve (posidoniids) mudstones, wackestones, packstones and grainstones represent the most common limestone biofacies. Microscopic analysis led to the identification of different textures, cements, carbonate grains, and initial porosity where bivalves and radiolarians are common. An intense infaunal activity in the limestone seems to have modified the original radiolarian-bivalve wackestone-packstone texture. Bivalve lags (Fig. 2 B; see arrow) found in the Fiuminata section, show tight packing of bivalve shells oriented parallel to bedding, with the convex side up, and showing no sign of abrasion. Shelter porosity is common. Since grainstone facies show the same type of fauna as the mudstones and packstones, we interpret the textural differences as resulting from winnowing of muddy sediments on the sea-floor, either by catastrophic events (Kreisa, 1981; Kreisa and Bambach, 1982) or by bottom currents (Kidwell, 1991; Monaco, 1992).

The marly-clay portion shows common (10-30%) to abundant (>30%) benthic foraminifers, ostracodes, bivalves, holothurians, ophiurids, crinoid fragments and spines, while radiolarians, which are very abundant in limestones, appear to be scarce or absent (except for sample Fle 10.10). Textures suggest that the original faunal assemblages were disturbed or even mechanically destroyed by an intense infaunal activity.

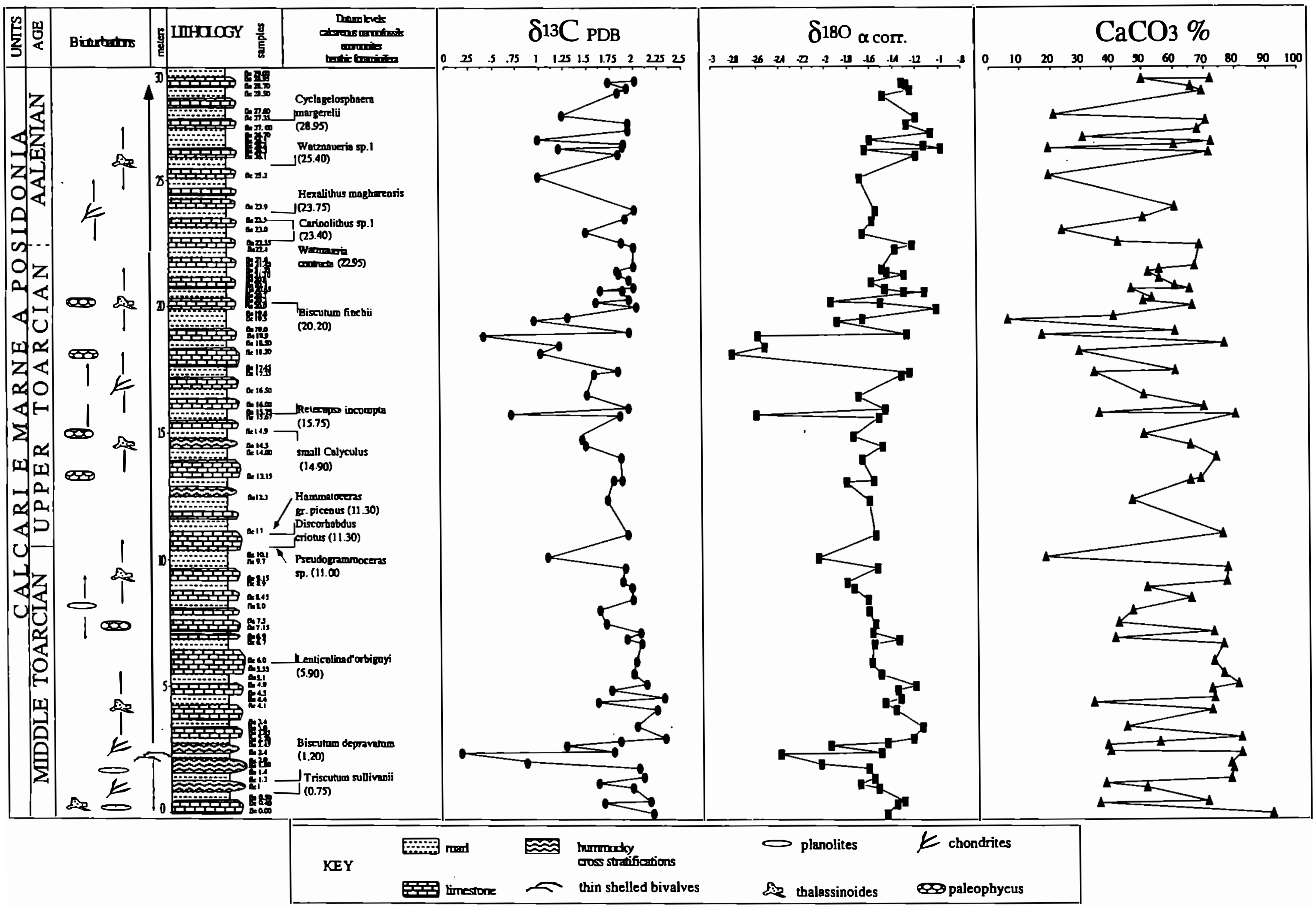
Marly levels show residual faunas dominated by benthic foraminifers (examination of more than 100 limestone by thin sections, conversely, has revealed absence of foraminifers and no more than 3 lenticulinas in samples Fle 1.4; 4.10; 8.45). Marlstone benthic foraminifer assemblages can be characterised as follows (according to the assemblages proposed by Nocchi 1992):

1) The assemblages of section Fle are dominated by *Lenticulina*, *Prodentalina* and *Eoguttulina*, common *Spirillina*, abundant *Conicospirillina*, and large *Astacolus*; *Lenticulina d'orbignyi* and *Falsopalmula tenuistriata* appear (samples Fle 5.90 and 13.40 respectively, Fig. 3, assemblage G according to Nocchi, 1992). This assemblage is typical of the Middle-Upper Toarcian.
2) Benthic foraminifer abundances fluctuate rhythmically between marlstones and limestones showing high abundances in the marly intervals (Fig. 3).
3) Number of species decreases at the Late Toarcian/Early Aalenian boundary (Fig. 3).
4) Benthic foraminifers, bivalves and ostracodes show no evident signs of erosion, abrasion, dissolution, or recrystallization when observed under the microscope.
5) Posidoniid shells are constantly found throughout the section both in marlstones and limestones (abundance >30% in washed residues).
6) Calcified radiolarians are very rare in marlstones but common in limestones.

Interpretation

a) A suitable environment for posidoniid bivalves occurred both during deposition of marlstones and limestones.
b) A favourable environment for benthic foraminifers occurred mainly during deposition of marlstones. In this basinal setting it is likely that increased influx of clays derived from continental settings could also signal the influx of additional organic matter from terrestrial environments. During carbonate influx minima, more organic matter could have possibly made it to the sea-floor, for consumption by benthic communities. The clay input from continental settings diluted the production of carbonates (thin shelled bivalves and calcareous nannofossils) through a Dilution Cycle. This latter, linked to fluctuating environmental conditions (Claps et al., 1997), might have triggered, through an increase of organic matter to the sea-floor, a rhythmicity in benthic foraminiferal abundances, here named a *proliferation cycle*.
c) More abundant radiolarians found in limestones could suggest higher surface fertility conditions during limestone sedimentation. However, we rather interpret the paucity of radiolarians in the marlstones as a result of poor preservation (Baumgartner, 1987).

Figure 4. Diagram showing Fle schematic lithology related to $CaCO_3$, foraminifera abundance, $\delta^{13}C$ and $\delta^{18}O$ variations. Note the opposite trend between $CaCO_3$ variations and foraminifera abundances, and the covariance of $CaCO_3$ percentage and $\delta^{13}C$ oscillations. For the exact correspondences between $CaCO_3$, $\delta^{13}C$ and $\delta^{18}O$ variations see Fig. 5 B.. Datum levels on calcareous nannofossils refer to Mattioli, (1994; 1997). Ammonites occurrences:Venturi, personal communication.


UNITS
AGE
Bioturbations
meters
LITHOLOGY
samples
Datum levels
calcareous nannofossils
ammonites
benthic foraminifera
δ13C PDB
δ18O α corr.
CaCO3 %
0 .25 .5 .75 1 1.25 1.5 1.75 2 2.25 2.5
-3 -2.8 -2.6 -2.4 -2.2 -2 -1.8 -1.6 -1.4 -1.2 -1 -.8
0 10 20 30 40 50 60 70 80 90 100
CALCARI E MARNE A POSIDONIA
MIDDLE TOARCIAN
UPPER TOARCIAN
AALENIAN
0
5
10
15
20
25
30
Cyclagelosphaera margerelii (28.95)
Watznaueria sp.1 (25.40)
Hexalithus magharensis (23.75)
Carinolithus sp.1 (23.40)
Watznaueria contracta (22.95)
Biscutum finchii (20.20)
Retecapsa incompta (15.75)
small Calyculus (14.90)
Hammatoceras gr. picenus (11.30)
Discorhabdus criotus (11.30)
Pseudogrammoceras sp. (11.00)
Lenticulina d'orbignyi (5.90)
Biscutum depravatum (1.20)
Triscutum sullivanii (0.75)
KEY
marl
limestone
hummocky cross stratifications
thin shelled bivalves
planolites
thalassinoides
chondrites
paleophycus

Geochemical Characteristics

Isotopic data: the $\delta^{13}C$ and $\delta^{18}O$ profiles

The absolute range of values for $\delta^{13}C$ varies between 0.14 and 2.3 ‰ (Figs. 4 and 5). The average value of $\delta^{13}C$ at Fiuminata is 1.69 ‰. Values below 0.9‰ were suspected to result from diagenetic calcite veins or voids. Measured values for marlstones differ systematically from those measured in limestones (Fig. 5). The mean $\delta^{13}C$ value for marlstones is 1.52‰, and ranges from 0.68 to 1.93‰. The mean $\delta^{13}C$ value for limestones is 1.85‰, and ranges from 0.14 to 2.29‰. The $\delta^{13}C$ signature for marlstones is on average 0.3‰ lighter than that for limestones. The average $\delta^{13}C$ difference between adjacent marl/limestone in individual couplets shows a value of 0.5‰ with maximum difference of 0.7‰. The absolute range of values for the $\delta^{18}O$ varies between -2.83 and -1.03‰. The average value for $\delta^{18}O$ is -1.53‰. The difference between marlstones and limestones is present also for $\delta^{18}O$, though to a smaller extent. The mean $\delta^{18}O$ value for marlstones is -1.63‰, and the absolute the range of values goes from -2.83 to -1.14‰. The mean $\delta^{18}O$ value for limestones is -1.44‰, and range from -2.45 to -1.21‰. The $\delta^{18}O$ signature for marlstones is on average 0.2 ‰ more negative than that for limestones. The average $\delta^{18}O$ difference between adjacent marl/limestones in individual couplets shows a value of 0.2‰. The mean shift recorded for oxygen is therefore by 0.1 ‰ smaller than the shift of the $\delta^{13}C$. It seems that oxygen values record environmental variations to a lesser extent than carbon values, and because of this, our interpretation of isotope values significance will concentrate on carbon. Scatter diagrams of $\delta^{13}C$ and $\delta^{18}O$ and of $\delta^{13}C$ versus $CaCO_3$ values is shown in Fig. 5.

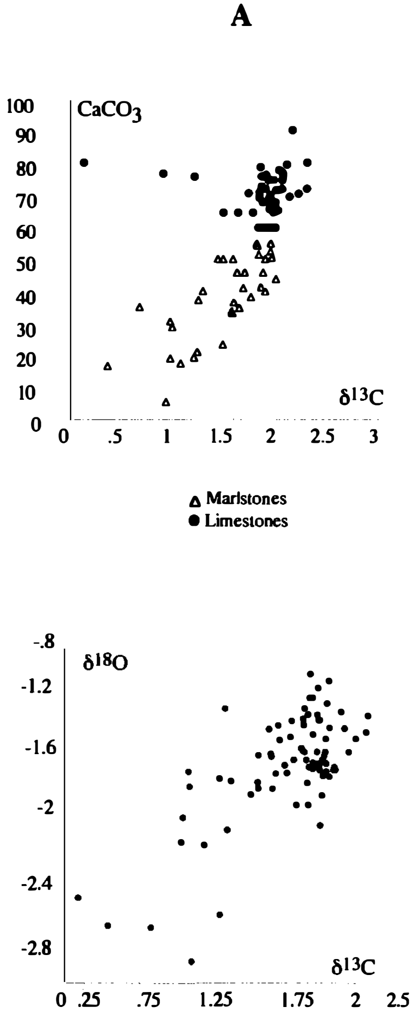
Interpretation

These measurements indicate a systematic difference in isotopic levels for limestones and marlstones in each couplet. No isotopic studies on Jurassic and Cretaceous rocks have been able to determine uniquivocally the primary or diagenetic origin of short term isotopic variations (Foucault and Renard, 1987; Arthur et al, 1994;). Whether our isotope data reflect primary fluctuations in the water column or early diagenetic changes at/or beneath the sediment-water interface cannot be determined. However, the isotopic data correlate well with fluctuations in benthic foraminifers abundance and lithology. We therefore believe that these data record short term environmental fluctuations either in the bottom waters or in the earliest diagenetic environment.

Although very small variations in the isotopic signature are often considered to be of solely diagenetic origin (Thierstein and Roth, 1991), some authors (R.O.C.C. Group, 1986; Barron et al., 1985, among others) have developed palaeoenvironmental and palaeoclimatic models based on different isotopic signatures for marlstones and limestones.

Isotopic values recording a primary signal: if the observed isotopic variations are considered primary signals, the lower $\delta^{13}C$ of marlstones might reflect different conditions at the sea-bottom and in the superficial waters. Phytoplankton (Mattioli, 1997) is less abundant during marly deposition, while benthos proliferates. The $\delta^{13}C$ in marly beds is lighter than in limestones, results that can be explained by lower ^{12}C uptake by phytoplankton. Benthic fauna proliferation is probably encouraged by enhanced organic matter discharge from the continents to the sea-floor, triggered by clay input, and not recycled because of stratified waters (pyrite was found in residues). The positive values of $\delta^{13}C$ in limestone beds stands as an indicator of higher uptake of nutrients from phytoplankton in the upper part of the water column; Mattioli (1997) notices higher productivity during the limestone hemicouplet deposition, concomitant with heavier $\delta^{13}C$ values. This hypothesis leads to the conclusion that short term nutrient/productivity fluctuations of the water column were recorded as small but significant isotopic variations.

Isotopic values recording an earliest diagenetic signal: if the observed isotopic variations are considered as an early diagenetic signal, the heavier isotopic value of limestones could be explained by rapid cementation taking place at the sediment-water interface. The isotopic signature of limestones is closer to equilibrium with sea water than the marlstones which tend to become



B

Meters	$\delta^{13}C_{rom}$	$\delta^{18}O_{corr}$	$CaCO_3$	$\delta^{13}C$ (mean v.)
0.0	2.15	- 1.49	91	
0.4	1.64	- 1.41	35	1.895
0.5	2.12	- 1.34	70	
1.0	1.94	- 1.56	70	2.03
1.2	1.58	- 1.72	36	
1.4	2.05	- 1.60	77	1.815
1.8	2.01	- 1.65	78	
2.0	0.91	- 2.07	77	1.46
2.4	0.14	- 2.43	81	
2.45	1.75	- 1.54	38	0.945
2.7	1.24	- 1.99	37	
2.85	1.82	- 1.48	54	1.53
3.0	2.29	- 1.26	81	
3.43	1.99	- 1.17	44	2.14
4.1	2.21	- 1.40	71	
4.4	1.98	- 1.50	33	1.90
4.55	2.28	- 1.36	72	
4.9	1.72	- 1.39	71	2.00
5.1	2.09	- 1.23	80	
5.55	1.96	- 1.53	75	2.03
6.0	1.98	- 1.61	72	
6.7	2.04	- 1.58	75	2.01
6.9	1.89	- 1.37	40	
7.15	2.04	- 1.60	72	1.97
7.5	1.68	- 1.57	41	
8.0	1.61	- 1.6[illegible]	46	1.65
8.45	1.96	- 1.64	65	
8.9	1.95	- 1.77	51	1.96
9.15	1.85	- 1.83	76	
9.7	1.88	- 1.56	77	1.87
10.1	1.07	- 2.08	17	
11.0	1.91	- 1.57	75	1.49
12.35	1.69	- 1.62	46	
13.1	1.76	- 1.83	65	1.73
13.15	1.85	- 1.58	68	
14.0	1.85	- 1.69	73	1.85
14.5	1.48	- 1.51	65	
14.9	1.43	- 1.76	50	1.46
15.67	1.83	- 1.54	79	
15.75	0.71	- 2.62	35	1.27
16.0	1.92	- 1.49	69	
16.5	1.48	- 1.72	50	1.70
17.35	1.56	- 1.34	34	
17.45	1.82	- 1.28	60	1.69
18.2	0.99	- 2.83	29	
18.5	1.19	- 2.54	76	1.09
18.9	0.36	- 2.61	17	
19.0	1.93	- 1.29	60	1.15
19.5	0.92	- 1.91	6	
19.6	1.28	- 1.68	40	1.10
20.0	2.01	- 1.03	66	
20.2	1.58	- 1.52	50	1.80
20.3	1.93	- 1.96	53	
20.68	1.86	- 1.14	46	1.90
20.7	1.63	- 1.32	65	
20.8	1.98	- 1.49	60	1.81
21.1	1.93	- 1.60	55	
21.35	1.82	- 1.32	51	1.88
21.45	1.81	- 1.47	55	
21.6	1.97	- 1.51	67	1.89
22.4	1.98	- 1.40	68	
22.55	1.85	- 1.25	41	1.92
23.0	1.47	- 1.68	23	
23.5	1.89	- 1.59	50	1.68
23.9	1.98	- 1.56	60	
25.2	0.97	- 1.71	19	1.48
26.1	1.83	- 1.21	71	
26.3	1.19	- 1.66	19	1.51
26.4	1.87	- 0.99	60	
26.5	1.88	- 1.14	72	1.88
26.7	0.97	- 1.61	31	
27.0	1.92	- 1.08	68	1.45
27.35	1.93	- 1.29	70	
27.6	1.23	- 1.21	21	1.58
28.5	1.82	- 1.49	69	
28.7	1.91	- 1.26	66	1.87
28.95	1.73	- 1.29	50	
29.0	2.01	- 1.33	72	1.87

Figure 5.(A) Scatter diagram of carbon versus oxygen isotopes of marlstones and limestones. (B) Geochemical values for the Fiuminata Colle Corno section.

cemented at later diagenetic stage, being influenced by the circulation of isotopically lighter fluids. Burrowers may have influenced the earliest diagenetic environment by provoking earliest cementation; a sufficiently firm substrate, created by early cementation will record burrowing, while the softer marly uncemented substrate will not (as confirmed by ichnofacies, Monaco et al., 1994).

Isotopic values recording burial compaction diagenesis: in this sense, isotope fluctuations simply mirror variations in lithology, therefore entirely depending on diagenesis; once the whole succession is buried and compacted, different diagenetic paths for limestones and marlstones occur, giving rise to different isotopic signatures.

Discussion and Conclusions

Marlstone/limestone cyclicity recorded in the Colle Corno section can be modeled following the four cycle types proposed by Einsele et al. (1991): dissolution cycle, calcareous redox cycle, productivity cycle, dilution cycle. A pure productivity cycle is characterised by a fluctuating supply of pelagic carbonate during a steady contribution of clay (Einsele and Ricken, 1991; Ricken, 1991). Since only extreme oceanic environmental changes, i.e., shifting of upwelling cells and climatic belts, can cause major changes in productivity, cyclic variation in productivity seldom seems the only factor controlling marl/limestone rhythms. In the Colle Corno section, thin-shelled bivalves might provide sufficient carbonate for a "steady" production, while the oscillating parameters are represented by periodic fluctuations of calcareous benthic foraminifers and by intermittent clay input in the basin. A study of calcareous nannofossils seems also to indicate rhythmic variations (Mattioli, 1994, 1997).

The absence of benthic foraminifers in the more calcareous lithologies and the abundance within the marly portion might be related to enhanced influx of organic matter from terrestrial environments during carbonate minima. Once on the sea-floor, the terrestrial organic material might have been available for consumption by benthic communities. Since proliferation of benthic foraminifers follows cyclic variations between carbonate-poor and carbonate-rich beds, we propose to adopt the term *proliferation cycle* for this type of benthic fauna fluctuation. The benthos fluctuation cannot be identified with the productivity cycle of Einsele and Ricken (1991), since the fluctuating parameter (benthic foraminifers) is not a source of pelagic carbonate. The change in benthic life can be interpreted as a trophic cycle at the substrate leading to a starvation during times of limestone sedimentation. Bottom waters might have been enriched in organic matter during periods of marlstone deposition, also due to lower O_2 contents (as proposed by Mattioli, 1997). The isotopically lighter values of marlstones might reflect lower phytoplankton nutrient uptake and higher quantities of organic matter reaching the sea-floor, only partially oxidized in the water column.

The alternation of carbonate-rich and carbonate-poor layers are therefore interpreted as the sedimentary expression of productivity and proliferation cycles induced by intermittent clay discharge in the basin.

The second hypothesis considers the isotopic signal simply as a mirror of lithological cyclicity. Carbon and oxygen isotopes can therefore enhance lithological cyclicity, but do not provide an independent signal. The third hypothesis is concerned with an entirely diagenetic signal.

We favour the first hypothesis because phytoplankton cyclicities, benthic foraminifer fluctuations and intermittent clay discharge, seem to be well correlatable with the systematic isotopic variations.

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References

ARTHUR, M. A., W. E. DEAN, D. BOTTJER, AND P. A. SCHOLLE. 1984. Rhythmic bedding in Meso-Cenozoic pelagic carbonate sequences: the primary and diagenetic origin of Milankovitch-like cycles, p. 191-222. *In* A. Berger, J. Imbrie, J. Hays, G. Kukla, and B. Saltzman, (eds.), Milankovitch and climate, Part 1, Riedel, Hingman.

BARRON, E. J., M. A. ARTHUR, AND E. G. KAUFFMAN. 1985. Cretaceous rhythmic bedding sequences: a plausibile link between orbital variations and climate, Earth and Planetary Science Letters, 72:327-340

BAUMGARTNER, P.O. 1987. Age and genesis of Tethyan Jurassic Radiolarites, Eclogae Geologicae Helvetiae, 80-3:831-879.

BERNOULLI, D. AND H. C. JENKYNS. 1974. Alpine, Mediterranean and central Atlantic Mesozoic facies in relation to the early evolution of the Tethys, p. 129-160. *In* R.H. Dott and R.H. Shaver, (eds.), Modern and Ancient geosynclinal Sedimentation, Society of Economic Paleontologists and Mineralogists, Special Publication, 19.

CHANNELL, J. E. T., B. D'ARGENIO AND F. HORVATH. 1979. Adria, the African Promotory, in Mesozoic Mediterranean paleogeography, Earth Sciences Review 15:213-292.

CHAREF, A. AND S. M. F. SHEPPARD. 1984. Carbon and Oxygen isotope analysis of calcite and dolomite associated with organic matter, Isotope Geoscience, 2:325-333.

CLAPS, M., E. MATTIOLI, E. MORETTINI, B. PITTET, AND M. RIPEPE. 1997. Orbitally driven palaeoclimatic control on plankton productivity and stable isotope record: an example from the Early-Middle Jurassic of Italy. Abstract Supplement N. 1, Terra Nova Vol. 9, 271.

COLACICCHI, R., M. NOCCHI, G. PARISI, P. MONACO, A. BALDANZA, S. CRESTA, AND G. PALLINI. 1988. Paleoenvironmental analysis from Lias to Malm (Corniola to Maiolica Formations) in the Umbria-Marche basin, Central Italy (preliminary report), p. 717-728. *In* R.B. Rocha and A.F. Soares, (eds.), 2nd International Symposium on Jurassic Stratigraphy Sept. 1987, Lisboa, vol. 2.

CONTI, M. A. AND S. CRESTA. 1982. Considerazioni stratigrafiche e paleontologiche sui "livelli a Posidonia" (Auct.) dell'Appennino Umbro-Marchigiano, Paleontologia Stratigrafica ed Evoluzione, Quaderno N°2, 73-80.

____ AND M. MONARI. 1992. Thin shelled bivalves from the Jurassic Rosso Ammonitico and Calcari a Posidonia Formations of the Umbria-Marche Apennine (Central Italy), Paleopelagos 2:192-213.

CRESTA, S., F. CECCA, M. SANTANTONIO, G. PALLINI, P. BRÖNNIMANN, A. BALDANZA, R. COLACICCHI, P. MONACO, M. NOCCHI, G. PARISI, AND F. VENTURI. 1988. Stratigraphic correlations in the Jurassic of the Umbria-Marche Apennines (Central Italy), p. 729-744. *In* R.B. Rocha and A.F. Soares, (eds.), 2nd International Symposium on Jurassic Stratigraphy Sept. 1987, Lisboa, 2.

EICHER, D. L. AND R. DINER. 1989. Origin of the Cretaceous Bridge Creek cycles in the Western Interior, United States, Paleogeography, Paleoclimatology, Paleoecology, 74:127-146.

EINSELE, G. AND W. RICKEN. 1991. Limestone-marl alternations – an overview, p. 23-47. *In* G. Einsele, W. Ricken and A. Seilacher, (eds.), Cycles and Events in Stratigraphy, Springer-Verlag, Berlin, Heidelberg, New York.

____, ____, AND A. SEILACHER, (eds.). 1991. Cycles and Events in Stratigraphy. Springer-Verlag, 956 p., Berlin, Heidelberg, New York.

ERBA, E. 1992. Calcareous nannofossil distribution in pelagic rhythmic sediments (Aptian-Albian Piobbico core, Central Italy), Rivista Italiana di Paleontologia e Stratigrafia, 97:455-484.

FOUCAULT, A. AND M. RENARD. 1987. Contrôle climatique de la sédimentation marno-calcaire dans le Mésozoique d'Espagne (Sierra de Fontcalent, province d'Alicante): argumentes isotopiques, Comptes Rendus de l'Académie des Sciences, Paris, 305, série II, 517-521.

KIDWELL, S. M. 1991. Taphonomic feedback (live/dead interactions) in the genesis of bioclastic beds: keys to reconstructing sedimentary dynamics, p. 268-282. *In* G. Einsele W. Ricken and A. Seilacher, (eds.), Cycles and Events in Stratigraphy, Springer Verlag, Berlin, Heidelberg, New York.

KREISA, R. D. 1981. Storm-generated sedimentary structures in subtidal marine facies with examples from the Middle and Upper Ordovician in southwestern Virginia. Journal of Sedimentary Petrology, 51,3:823-848.

____ AND R. K. BAMBACH. 1982. The role of storm processes in generating shell beds in Paleozoic shelf environments, p. 200-207. *In* A. Seilacher and G. Einsele, (eds.), Cyclic and Event Stratification. Springer Verlag, Berlin, Heildelberg, N.Y.

LAVECCHIA, G. 1981. Appunti per uno schema strutturale dell'Appennino Umbro-Marchigiano: 3) lo stile deformativo, Bollettino della Società Geologico Italiana, 100:271-278.

____ AND G. P. PIALLI. 1980. Appunti per uno schema strutturale dell'Appennino Umbro-Marchigiano: la copertura, Studi Geologici Camerti, 6:23-30.

MATTIOLI, E. 1994. Calcareous nannofossils of the Toarcian - Aalenian of the Fiuminata section (Central Apennines, Italy), Paleopelagos, 4:175-187.

____. 1997. Nannoplankton productivity and diagenesis in the rhythmically bedded Toarcian-Aalenian Fiuminata section (Umbria-Marche Apennine, central Italy): Paleogeography, Paleoclimatology, Paleoecology, 130:113-133.

MCCREA, J.M. 1950. The isotopic chemistry of carbonate at a paleotemperature scale, Journal of Chemical Physics, 18:849-857.

MONACO, P. 1992. Hummocky cross-stratified deposits and turbidites in some sequences of the Umbria-Marche area (central Italy) during the Toarcian, Sedimentary Geology, 77:123-142.

____ AND E. MORETTINI. in press. Marl-limestone rhythmites and event beds in the Toarcian-Aalenian "Calcari e Marne a Posidonia" unit of Fiuminata (Pioraco, central Apennines). Bolletino Servizio Geologico d'Italia Roma, 96 (1997).

___, M. NOCCHI, M. ORTEGA-HUERTAS, I. PALOMO, F. MARTINEZ, AND G. CHIAVINI. 1994. Depositional trends in the Valdorbia section (Central Italy) during the Early Jurassic, as revealed by micropaleontology, sedimentology and geochemistry, Eclogae Geolicae Helvetiae, 87,1:157-223.

NOCCHI, M. 1992. Associazioni a foraminiferi bentonici del Bacino Umbro-Marchigiano (Appennino centrale) durante il Lias, Paleopelagos , 2:37-53.

ORTEGA-HUERTAS, M., P. MONACO, AND I. PALOMO. 1993. First data on clay mineral assemblages and geochemical characteristics of Toarcian sedimentation in the Umbria-Marche basin (central Italy), Clay Minerals, 28:297-310.

PARISI G., A. BALDANZA, L. BENEDETTI, E. MATTIOLI, F. VENTURI, AND S. CRESTA. 1998. Toarcian stratigraphy of the Colle D'Orlando section (Umbria, central Italy, northern Apennine). Bolletino della Società Paleontogica Italiana, 37 (1):3-39.

RICKEN, W. 1991. Variation of sedimentation rates in rhythmically bedded sediments. Distinction between depositional types, p. 1-19. *In* G. Einsele, W. Ricken and A. Seilacher, (eds.), Cycles and Events in Stratigraphy, Springer Verlag, Berlin, Heidelberg, New York.

R.O.C.C. Group. 1986. Rhythmic bedding in Upper Cretaceous pelagic carbonate sequences: varying sedimentary response to climatic forcing, Geology, 14:153-156.

SCHOLLE, P. AND M. A. ARTHUR. 1980. Carbon isotopic fluctuations in pelagic limestones: potential stratigraphic and petroleum exploration tools. American Association of Petroleum Geologists Bulletin, 64:67-87.

STAM, R. 1986. Quantitative analysis of Middle-Late Jurassic foraminifera from Portugal and its implications for the Grand Banks of New Foundland, Utrecht Bulletins, p. 1-166.

THIERSTEIN, H.R. AND P.H. ROTH. 1991. Stable isotopic and carbonate cyclicity in Lower Cretaceous deep-sea sediments: dominance of diagenetic effects, Marine Geology, 97:1-34.

Terrestrial Ecosystems

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Early Middle Jurassic Plant Communities in Northwest Scotland: Paleoecological and Paleoclimatic Significance

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Abstract: Recently discovered plant megafossil assemblages from Bearreraig, Skye, northwest Scotland offer a rare opportunity to study the anatomy of Middle Jurassic land-plants. The 14 plant-bearing horizons occur throughout ca. 120m of strata and have yielded 15 disarticulated organ-species, including unique, calcareously permineralized, cycadophyte leaves and conifer strobili. They represent a minimum of eight whole-plant species: one equisetalean, four filicalean ferns, one cycad, at least one bennettite, and at least one conifer. Calcitic nodules enclosing the plant megafossils also contain diverse assemblages of well preserved miospores and representatives of many major groups of contemporaneous marine biota. Integration of narrow ammonite biozones with standard chronostratigraphies suggests prolonged accretion, from the earliest Aalenian to the early Bajocian. Tectonic, sedimentologic, and paleontologic data all indicate deposition in fully marine conditions, ca. 25km offshore from at least one substantial delta. The river systems drained large and topographically variable catchments, and therefore probably sampled and homogenized components from several biotic communities. Taphonomic depletion least affected organ-species that were (1) resistant to mechanical disaggregation (i.e., robust), (2) resistant to biodegradation (most effectively conferred by charcoalification), and (3) entered the transport system closest to the site of deposition. Plants probably rafted seaward until encountering the turbulent zone of column inversion and mixing, where they settled through the water column and were rapidly buried by dominantly terrigenous sediment. Complex physico-chemical interactions in this unstable depositional environment allowed permineralization of plant tissues by calcite leached from surrounding invertebrate tests. Paleoecological interpretation of these highly allochthonous fossil plant assemblages necessitates distinction between (1) putative whole-plant species of different communities within each assemblage, and (2) evolutionary versus ecological causes of directional changes in species compositions of successive assemblages. At least two distinct associations apparently persisted throughout the notional 10my of accumulation: a coastal/deltaic community dominated by the araucarian conifer *Brachyphyllum* cf. *mamillare–Taxodioxylon*–cf. *'Araucarites'* (admixed *B.* cf. *crucis* may represent an associated cheirolepidiacean conifer), and an inland community represented primarily by burnt foliage of the matoniaceous fern *Phlebopteris*. Although exaggerated by taphonomic depletion, the low species diversity and remarkable robustness of these land-plant communities apparently reflect growth under appreciable extrinsic stresses, notably periodic aridity. Prodelta and marine shelf facies are undoubtedly under-exploited as potential sources of fossil megafloras that indicate contemporaneous terrestrial environments.

Introduction

Latest Triassic and Jurassic strata have yielded some of the best-known of all fossil floras (Wing et al., 1992); notable examples include the Rhaetic flora of Scoresby Sound, east Greenland (Harris, 1937), the largely Bajocian flora of northeast Yorkshire (Harris, 1961–1979) and the largely Kimmeridgian flora of the French Jura (Barale, 1981). The plants typically occur in fine-grained deltaic, estuarine, and shallow marine, clastic sediments, and consist largely of adpressions: essentially two-dimensional fossils that provide information on gross morphology but lack

Stage	Aalenian										Lower Bajocian			
Ammonite zone	*Opalinum*		*Murchisonae*		*Concavum*						*Discites*	*Laeviuscula*		
Lithostratigraphic unit	Dun Caan Sh.		Ollach Sandstone				Udairn Shale				Holm Sandstone			
Plant-bearing horizon	P1	P2	P3	P4*	**P5**	**P6***	P7*	P8*	**P9***	**P10***	P11	P12	P13*	P14*
Sphenopsida: Equisetales														
'Equisetites' cf. *columnaris*	–	–	–	–	–	–	–	□	–	–	–	◆↓r	–	–
Pteropsida: Filicales														
cf. *Cladophlebis denticulatus*	–	–	–	–	□↓	–	–	–	–	–	–	–	–	–
Hausmannia dichotoma	–	–	–	–	–	■	–	–	–	–	–	–	–	–
Hausmannia buchii	–	–	■	–	–	■	–	–	–	□	–	–	–	–
Coniopteris cf. *hymenophylloides*	–	–	–	–	–	■	–	–	–	–	–	–	–	–
Phlebopteris woodwardii	■	–	–	■↓	■ f	■ f	■	–	–	■	■	■↓	■	■ f
Rachises (?mostly *Phlebopteris*)	■●	□	■	–	□↓	■●	–	–	■	■●	–	■↓	–	■ r
Gymnospermopsida: Cycadales														
Nilssonia cf. *tenuinervis*	–	–	–	–	–	–	–	–	–	●	–	–	–	–
Gymnospermopsida: Bennettitales														
cf. *Otozamites penna* (small)	–	●	–	–	–	○	–	–	●	●	○	–	○	–
cf. *Otozamites penna* (large)	●	–	–	–	–	–	–	–	–	–	–	–	–	–
?Female cone (robust)	–	–	–	–	–	–	–	–	–	–	–	–	–	●
?Isolated sporophylls (cf. *Cycadolepis* sp.)	–	–	–	–	–	–	–	–	●	–	–	–	–	–
?Isolated ovules	–	–	–	–	–	–	–	–	●	–	–	–	–	–
Gymnospermopsida: Coniferales														
Large woody axes (mostly cf. *'Taxodioxylon'*)	–	–	–	–	○↓	–	–	–	●	●	–	●↓	●	●
Leafy shoots (mostly *Brachyphyllum* cf. *mamillare*)	●	●	●	●	●	●	●	●	●	●■	●■	●↓	–	●
Female cone (cf. *'Araucarites'*)	●	–	–	–	–	–	○	–	○	–	–	–	–	○
?Isolated ovules	–	–	●	–	–	–	–	–	–	○	–	–	–	–
Unidentified wood fragments (?mostly cf. *'Taxodioxylon'*)	●■	●■	●■	–	–	–	–	■	■●	●■	–	●■↓	–	–

anatomical detail. In the Jurassic of western Europe, the most frequently encountered sources of anatomical data are burnt fern foliage and calcareously petrified fragments of dominantly coniferous driftwood. Elsewhere, siliceous petrifaction has preserved rare examples of ferns, cycadophytes, and conifers. Calcareously petrified floras, frequent elsewhere in the geological column, are especially rare throughout the Jurassic worldwide (for a rare and taxonomically restricted exception see Seward and Bancroft, 1913).

This paper, abstracted from a more detailed manuscript currently in preparation, describes a new sequence of allochthonous, anatomically-preserved Middle Jurassic floral assemblages from coastal sections at Bearreraig, Skye, northwest Scotland (Morton, 1990; Bateman and Morton, 1994; Dower and Bateman, 1998). Although the plants have been severely taphonomically disarticulated into a wide range of isolated organs, quality of preservation by calcareous petrifaction is high. The plant megafossils occur at several stratigraphic levels, together with diverse assemblages of well preserved miospores and contemporaneous marine organisms. These include dinoflagellates, calcareous nannofossils, foraminifera, ostracodes, crinoids, bivalves, gastropods, scaphopods, belemnites, and ammonites, together with fragmentary arthropods, crinoids, and fish; nearby occur tetrapods, including the only known Scottish dinosaur bones. The biostratigraphic value of most of these fossil groups was recently reviewed in detail, leading to Bearreraig's Auxiliary Stratotype status for the Aalenian–Bajocian boundary (Morton, 1990; Morton and Hudson, 1995).

Here we use Bearreraig as a case-study in the interpretation of sequences of strongly allochthonous fossil plant assemblages. The newly-acquired paleobotanical information is integrated with tectonic, sedimentologic, and paleozoologic data to infer the nature of the taphonomic cycle experienced by the plants. These interpretations in turn provide a framework for tentative reconstructions of their source communities and of the prevailing climatic regime.

Location, Stratigraphy, Age, and Materials

Bearreraig Bay is situated on the east coast of the Trotternish Peninsula, 10km north of Portree on the Isle of Skye, northwest Scotland. Our detailed measured section currently represents six exposures extending laterally over 2km. The Jurassic stratigraphy of Bearreraig is based on zonal and subzonal boundaries interpolated between numerous horizons that yielded abundant, well preserved ammonite assemblages, thereby providing an unusually precise biozonation (Morton, 1990) that has since been supplemented with other biostratigraphic, paleomagnetic and Sr isotope data (Morton and Hudson, 1995).

The 14 plant-bearing horizons discovered to date (excluding biologically uninformative wood fragments) span a 120m-thick central portion of the Bearreraig Sandstone Formation (Fig. 1), from the middle of the Dun Caan Member (*opalinum* Zone: earliest Aalenian) through the Ollach Sandstone and Udairn Shale Members to the middle of the Holm Sandstone Member (*laeviuscula* Zone: early Bajocian); thus, they slightly pre-date or overlap the classic Yorkshire floras described by Harris (1961–1979), and represent a nominal 10my of potential evolution (Harland et al., 1989).

Numerous calcareous nodules were collected by N. Morton between 1963 and 1984. They were fractured repeatedly into irregular (typically 1–5cm) fragments in order to expose abundant, well preserved ammonites and bivalves, which yielded valuable biostratigraphic and paleobiologic information (Morton, 1990; Morton and Hudson, 1995). As an unexpected by-product, small but well preserved, land-plant fossils were also revealed, intimately associated with the marine molluscs responsible for their calcitic preservation. Following re-examination of the stratigraphic collections, the most promising plant-bearing horizons (notably P6, P9, and P10; Fig. 1) were sampled more extensively by R. Bateman and O. Farrington in 1986 and 1995, supplemented with loose blocks collected from the foreshore.

Figure 1. Stratigraphic distribution of petrified plant megafossil organ-species at Bearreriag Bay, Skye. Preservation: circle = non-fusainized, square = fusainized, diamond = pyritized (if two states present, dominant is listed first); data reliablity: open symbol = no specimen identified with confidence, arrow = all specimens from loose blocks only; detail of ferns: **f** = fertile pinna, **r** = possible rhizome; horizons yielding the largest samples are in boldface, those yielding palynological samples are asterisked.

Plant megafossils were prepared for photography, cellulose acetate anatomical peeling, petrographic thin-sections, and SEM examination using standard paleobotanical techniques (Jones and Rowe, in press). Plant microfossils and cuticles were macerated from the inorganic matrix; this was analyzed geochemically to investigate modes of preservation (Bateman *in* Jones and Rowe, in press).

Organic-walled Microfossils

Miospores and dinoflagellate cysts dominate the palynological preparations, supplemented with occasional acritarchs, algae (marine prasinophyceans and the presumed brackish/freshwater chlorophyte *Botryococcus*), and foraminiferal test-linings. They occur with variable quantities of palynodebris, dominantly land-plant tracheids (internal) and cuticle (external).

Dinoflagellate assemblages from Bearreraig contain ca. 17 morphospecies (Riding, 1991) and generally resemble those from Jurassic strata elsewhere in northwest Europe. The suites are dominated by *Nannoceratopsis* species (notably *N. ambonis* and *N. gracilis*), together with the pareodinoid *Caddasphaera*. Samples higher in the succession (P13 - P14) are more diverse, including appreciable quantities of the gonyaulacoid *Dissiliodinium* and the pareodinoid *Kallosphaeridium*.

Miospore assemblages are appreciably more diverse. Riding (1991) recorded at least 31 miospore species, even without attempting to differentiate morphospecies of the stratigraphically unhelpful (but paleoecologically highly significant) bisaccate group that dominates most samples. A substantial increase in morphospecies diversity approximating the junction of the Dun Caan Shale and Ollach Sandstone Members occurs ca. 8 m above the first appearance of identifiable plant megafossils in the succession (horizon P1; Fig. 1). Putative gymnosperm pollen includes abundant bisaccates, together with several organ-species of *Callialasporites* and the characteristic Jurassic miospore '*Classopollis*' (strictly, *Corollina*). Also consistently present are *Cerebropollenites macroverrucosus*, *Perinopollenites elatoides*, and *Cycadopites* species.

The most abundant and persistent putative pteridophyte spore-species are *Retitriletes austroclavatidites*, *Ischyosporites variegatus*, and *Cyathidites* species (notably *C. australis* and *C. minor*). Frequently occurring, but less abundant, spores are *Coronatispora valdensis*, *Concavissimisporites verrucosus*, *Todisporites minor*, *Osmundacidites wellmanii*, *Lycopodiacidites semimuris*, and *Leptolepidites* species (Riding, 1991). Sadly, few of these spore species have been conclusively correlated with their source megafossil species.

Plant Megafossils

As systematic study of the plant megafossils is still in progress, no attempt has been made here to revise their taxonomy. Work on the pteridophytes is regarded as complete, whereas further paleobotanical investigation is planned for the gymnosperms. Due to severe disarticulation, taxa listed in Figure 1 are isolated organ-species rather than whole-plant species.

Early hopes of detecting non-vascular plants at Bearreraig were dashed when the specimens in question (non-charcoalified axes from horizons P1, P5, P8, P10, and P12) proved to resemble the organic internal frameworks of sponges!

Turning to the vascular land-plants, the Sphenopsida (Equisetales) is represented by a large, pyritized putative rhizome of *Equisetites* cf. *columnaris*; comparison with related extant horsetails suggests that it reached the size of a small tree. The ferns (Pteropsida: Filicales) are more diverse and uniformly fusainized (charcoalified), thereby revealing intricate histological details under the SEM. Much the most abundant and stratigraphically widespread fern is *Phlebopteris woodwardii* (Matoniaceae), which occurs as both sterile and fertile pinnules derived from the palmate leaves, plus occasional subtending rachises and one possible rhizome fragment. Next most frequent is *Hausmannia* (Dipteridaceae), which occurs as two form-species, *H. dichotoma* and *H. buchii*. Their co-occurrence in horizon P6 at Bearreraig, and their histological similarity under the SEM, support Harris' (1961) speculation that the two organ-species represent a single dimorphic whole-plant species that possessed specialized fertile and sterile fronds respectively (analogous to the extant Stags-horn Fern, *Platycerium*). Much rarer are single pinnule fragments of cf. *Cladophlebis denticulatus* (of the water-loving Osmundaceae) and *Coniopteris* cf. *hymenophylloides* (of the typically shrub/tree-sized, forest-dwelling Cyatheaceae). Thus, the four putative whole-plant ferns represent four different (and phylogenetically relatively primitive) families (Bateman and Morton, 1994); moreover, all four characterize the Yorkshire Jurassic floras (Harris, 1961).

The remainder of the Bearreraig plants are gymnosperms. Although well preserved and rich in morphological and histological characters, the 'cycadophytes' proved a serious challenge to identification. The well known evolutionary convergence in gross leaf morphology between true Cycadales and the more phylogenetically derived, unrelated Bennettitales was revealed primarily by epidermal and stomatal specialisms of the latter mirrored in cuticlar imprints (Harris, 1964; 1969), which are poorly preserved on the otherwise highly informative Bearreraig petrifactions. The rarer of the two cycadophytes is a single central portion of an entire, strap-shaped leaf plus distinctive rachis that appears assignable to the cycad *Nilssonia* cf. *tenuinervis* rather than the bennettite *Nilssoniopteris* (Dower and Bateman, 1998). The larger (and more frequent), pinnate cycadophyte genus is a bennettite, assigned to cf. *Ptilophyllum* by Bateman and Morton (1994). Most specimens compared most closely with *P. pecten*, but a large specimen with elongate pinnae from the lowest horizon in the succession resembled *P. pectinoides*. These two organ-species may represent different growth forms of the same whole-plant species, an interpretation reinforced by the recent discovery of a clearly juvenile frond by Dower and Bateman (1998). These authors tentatively reassigned all the Bearreraig bennettite leaves to the very similar genus *Otozamites* cf. *penna*. Isolated sporophylls and ovules from P9, and a cone fragment from P14, were probably borne by cycadophytes.

The conifers have received least taxonomic attention thus far, despite being the most abundant plant remains at Bearreraig (Fig. 1). Well preserved leafy shoots occur throughout the succession, dominating most horizons. A few are sufficiently large to show two orders of branching. Leaves are thickly xeromorphic and bear distinctive sunken, pentagonal stomata. They vary considerably in morphology, from appressed sheaths through club- to egg-shaped; all fit the broadly delimited form-genus *Brachyphyllum*. Most appear assignable to *B.* cf. *mamillare* (Araucariaceae), though the convergent *B.* cf. *crucis* (Cheirolepidiaceae) may also be present. Frequent woody axes and wood fragments throughout the sequence probably represent these conifers; those in the upper part of the succession are sufficiently well preserved to identify as cf. '*Taxodioxylon*.' Two excellent intact female cones (one flattened) and a few isolated ovules are also consistent with araucarian derivation.

To summarize, the Bearreraig organ-species most likely constitute a minimum of eight and maximum of eleven whole-plant species (Bateman and Morton, 1994). This is a startlingly low species diversity for a sizeable taphonomic catchment (for example, compared with the much richer Late Aalenian–Bajocian floras of Yorkshire; Harris, 1961–1979), but remarkably the assemblage encompasses eight or nine families and five orders/classes (*sensu* Stewart and Rothwell, 1993).

Tectonic Framework and Sedimentary Environment

The severe fragmentation and fully marine depositional setting of the floras conclusively demonstrate their highly allochthonous nature. Nevertheless, many floral assemblages are preserved in the 120m-thick plant-bearing sequence, documenting repeated input of plant debris. Inferences concerning their taphonomy therefore require knowledge of the prevailing tectonic framework and the resulting patterns of sedimentation.

The Lower to Middle Jurassic transition was marked by an episode of lithospheric stretching in the proto-North Atlantic, which tectonically renewed subsidence in the sedimentary basins and uplifted the hinterlands, rejuvenating topography (Morton and Hudson, 1995). Large quantities of terrigenous material (including land-plant debris) were discharged into basins surrounding northern Britain. Marginal and non-marine depositional environments dominated throughout the Aalenian and Bajocian, persisting into the Bathonian in all areas except the northernmost North Sea. Estuarine and deltaic sediments incorporated important adpression floras in the Bajocian of the Yorkshire Basin (Harris, 1961–1979) and the Bathonian of Sutherland (Harris and Rest, 1966; Van Konijnenberg-Van Cittert and Van Der Burgh, 1989).

The Middle Jurassic Hebrides Basin was greatly elongate north to south, connected to a broad shelf sea to the south but closed at the northern end; consequently, it was strongly tidal. Clast petrography and fine-sand mineralogy of the Bearreraig Sandstone indicate that associated siliciclastics were derived from Moine and Dalradian sources eastward in the Scottish Highlands; the Middle Jurassic coastline lay close to the present-day coastline of the Scottish mainland, 25–30km east of Bearreraig. North to south variations in detrital mineralogy indicate that several rivers drained into the Hebrides Basin.

The Bearreraig Sandstone varies considerably in thickness across the Skye–Raasay area, reaching an intermediate 210m at Bearreraig itself. The most striking facies variation is the diachronous spread northwards of cross-stratified sandstones and sandy limestones that were deposited as migrating tidal sand waves. Orientations of cross-strata show that sediment transport was predominantly longitudinal to the Hebrides Basin, parallel to the north - south oriented coastlines and fault-related topographic features within the basin. Terrigenous sediment from the rejuvenated hinterland to the east was transported by rivers to the sea, where it was redistributed by tidal currents and mixed with varying proportions of diverse marine shell debris.

Stenohaline nektic/nekto-benthic faunas occur in most beds at Bearreraig (Morton, 1990; 1992). A change in sea bottom conditions resulted in a paucity of cephalopods and a transition from dominantly calcitic to dominantly sideritic cementation (with a concomitant decline in the frequency and quality of plant preservation) in the upper part of the Udairn Shale Member. Bivalves dominate the benthic megafaunas, and the only common gastropod is a small species of *Actaeonina*. Current studies of microfossils from Bearreraig may eventually provide greater environmental resolution than the megafossils.

Although the presence of land-plant debris and the brackish/freshwater alga *Botryococcus* clearly demonstrate input of onshore biota, the closest evidence (both geographically and temporally) of temporary emergence occurs 9km south of Bearreraig, where root-casts delimit a paleosol. This horizon occurs higher in the succession than the uppermost plant-bearing horizon (P14) at Bearreraig. Thus, the entire Bearreraig Sandstone was deposited in a normal-salinity marine environment. The sediments and faunas are typical of soft, muddy to sandy sea-floors on an offshore, shallow shelf experiencing relatively rapid accumulation of organic-rich sediment.

Taphonomy

The Bearreraig plant assemblages are highly fragmented and were deposited under fully marine conditions, 25 - 30km from the nearest emergent terrain. They are, therefore, highly allochthonous; a series of taphonomic filters (*sensu* Behrensmeyer et al., 1992) has severely depleted the number of organ-species present and greatly modified the relative proportions of those that remain. Even tentative reconstruction of their source communities, therefore, requires consideration of the possible effects of all the biases listed by Bateman (1991) as potentially resulting in differential preservation .

The stratigraphic distribution of the permineralized floral assemblages (Fig. 1) is strongly positively correlated with the occurrence of nodular concretionary horizons that are cemented primarily by calcite rather than siderite. Laminar concretions in the upper Dun Caan Shale and basal portion of the Ollach Sandstone Member, rich in molluscs, yielded floral assemblages P1 -P4; they probably reflect decreased supply of siliciclastics relative to rates of biogenic carbonate precipitation. The most diverse and best-preserved biotic assemblages occur in bands of calcareous nodules and doggers, which vary considerably in size and shape. Nodules from the most productive horizons (P6 - P10) are typically 10 - 30cm in maximum diameter and roughly ellipsoidal in longitudinal section. Geochemical analyses demonstrate that the main distinguishing feature of the nodules is the concentration of both biogenic and authigenic $CaCO_3$ (Bateman *in* Jones and Rowe, in press).

The heterogeneous distribution of biotas within individual horizons is probably largely a secondary phenomenon, reflecting non-preservation of most of the megafossils that accumulated on the sea floor at each point in time. Decay and dissolution of shells released sufficient calcium carbonate to allow local re-precipitation of authigenic calcite, which nucleated around the remaining shells. Cementation filled pore spaces, thus protecting intra-nodular biota from further chemical dissolution and from physical compression. Meanwhile, biotas in the enclosing sediment were severely depleted by further dissolution. Plant fragments (including charcoal) that were intimately associated with the molluscs were thus fortuitously preserved by calcitic permineralization.

Some of the conifer branches reveal fossilized excreta of wood-boring insects, but these probably formed pre-mortem. Evidence for post-mortem modification is confined to the best preserved of the cf. *Otozamites* leaves, which contain sparse degradative fungi within the mesophyll (Dower and Bateman, 1998). Otherwise, the paucity of chemical and/or microbial degradation of the plant tissues at Bearreraig suggests that permineralization (and therefore nodule formation) occurred very rapidly after deposition, as the relatively shallow water column and sandy substrate are unlikely to have allowed the development of anoxic conditions at the water - sediment interface (this inference is supported by evidence of extensive bioturbation). Subsequent recrystallization of calcite, which can severely disrupt anatomical detail in calcareously permineralized plant fossils, has only seriously affected the larger woody conifer branches.

The abundance of plant-bearing fossils in the Bearreraig Sandstone and the dominance of *Brachyphyllum* and *Phlebopteris* throughout the succession indicate repeated input from broadly similar communities during the notional 10my gross accumulation period. Together with the rapid accumulation of terrigenous sediment, this suggests deposition offshore from one or more persistent river discharges. Although disarticulated, the floral assemblages are dominated by delicate, non-charcoalified, leaf-bearing conifer twigs and charcoalified (and therefore brittle) fern pinnules. Neither component could have survived the minimum of 25km of transport from the nearest landmass across an ocean floor that was rapidly accumulating medium - coarse clastic sediment. Thus, the floral assemblages are more likely to have been carried into the shelf environment as rafts, before becoming waterlogged at the zone of saline inversion and backflow, and subsequently sinking through the water column to the ocean floor (Bateman and Morton, 1994).

The long distance from shore and rapid accumulation of terrigenous sediment suggest that discharge from the rivers was substantial. Relatively high gradients (reflecting considerable topographic relief) and a large catchment are postulated. Thus, the Bearreraig floral assemblages potentially contain components of several ecologically distinct communities, sampled by the rivers from a considerable geographical area that encompassed an appreciable range of altitudes. Many authors argue that long-distance aqueous transport results in plant megafossil assemblages that preserve general community patterns (Behrensmeyer et al., 1992), though preservation bias is likely to be severe and temporal acuity poor.

Several pieces of evidence indicate a high degree of taphonomic bias at Bearreraig:

1. Although the organ-species assemblages contain representatives of five orders (*sensu* Stewart and Rothwell, 1993), they can be resolved into a minimum of only eight whole-plant species. Moreover, only the araucarian conifer *Brachyphyllum* cf. *mamillare* and the matoniaceous fern *Phlebopteris woodwardii* are frequent, the former dominating the preserved biomass. This low 2. Most of the foliar organs recovered possessed thick, robust leaves that appear potentially resistant to both mechanical and chemical degradation. Any more vulnerable plant debris that entered the water-course was presumably eliminated earlier in the taphonomic cycle.
3. Dispersed diaspores (notably cycadophyte and, especially, conifer seeds) are surprisingly rare; it seems likely that they lacked the flotation potential of leaves, strobili, and wood fragments, and hence were winnowed out as bedload to form more proximal assemblages.
4. The Bearreraig palynofloras appear considerably more diverse than the associated plant megafossils. This situation characterizes most well preserved fossil floras, as dispersal ranges for miospores generally greatly exceed those of megafossils derived from the same individual plant. Thus, the miofloras may provide a more accurate assessment of overall species diversity in the hinterland, though they are far too admixed to record accurately the composition of individual communities.

Any attempt to differentiate source communities within the Bearreraig floral assemblages should, therefore, take account of their profound taphonomic depletion, and will inevitably be tentative.

Terrestrial Paleocommunities

It is logical to assume that the Bearreraig discharge sampled both riparian/coastal and hinterland communities. The most compelling (albeit circumstantial) evidence for the presence at Bearreraig of plant megafossils from more than one community is provided by variation in preservation.

The single large *Equisetites* axial fragment is pyritized, suggesting that it was a component of a riparian community and was therefore able to enter the water-course directly. The remaining plant megafossils exhibit two distinct modes of preservation. Portions of the fern *Phlebopteris* (together with the single frond fragments of *Cladophlebis*, *Hausmannia*, and *Coniopteris* from horizon P6) are invariably black and brittle, whereas the various organ-species of conifer (together with the cycadophyte leaves *Nilssonia* and cf. *Otozamites*, mostly from horizons P9 and P10) are invariably brown and less prone to fragmentation. The black specimens are regarded as fusain (fossil charcoal). Harris (1961) argued that *Phlebopteris* dominated extensive inland heaths, analogous to those dominated by Bracken (*Pteridium aquilinum*) in present-day northwest Europe. Periodic wildfire swept through the seasonally dry heaths, consuming the more flammable components of the vegetation such as conifers but charcoalifying the thick, heavily cutinized and comparatively resin-deficient *Phlebopteris* fronds. *Cyathidites* spores, which may in part represent *Phlebopteris*, are frequent in the Bearreraig palynofloras, providing additional (albeit equivocal) evidence that this fern was a significant component of the regional vegetation. Modern descendants of all four fern genera from Bearreraig preferentially inhabit nutrient-poor soils.

Coniferous remains occur even more frequently in the Bearreraig succession than *Phlebopteris* (Fig. 1). Most specimens are rich brown in colour, relatively well articulated, and their size often appears constrained only by the dimensions of the preservation envelope conferred by the enclosing nodules. Similar preservation is exhibited by the cycadophyte leaves *Nilssonia* and cf. *Otozamites*. The *Brachyphyllum* cf. *mamillare* twigs, *Taxodioxylon*-type wood and rare cf. *'Araucarites'* cones are all consistent with derivation from araucarian confers (Harris, 1979; Stewart and Rothwell, 1993). However, the very similar foliage-species *B.* cf. *crucis*, which has also been correlated with *Taxodioxylon*-type wood, reputedly possesses pollen-bearing cones assigned to the highly morphologically divergent Cheirolepidiaceae. Interestingly, the palynofloras consistently contain *'Classopollis'*, which is considered to be exclusively cheirolepidiacean in origin. However, it is diluted by an abundance of bisaccate pollen, more consistent with derivation from Pinaceae or Podocarpaceae (Hart, 1987).

The relative abundance and articulation of *Brachyphyllum* species at Bearreraig suggest that they occupied distal (possibly intertidal) zones around the deltas. The araucarian *B.* cf. *mamillare* appears more abundant than the putative cheirolepidiacean *B.* cf. *crucis*, though it may simply have grown closer to the water-courses and thus been preferentially preserved. The poorly represented cycadophytes *Nilssonia* and cf. *Otozamites* may have been less common components of this riparian (possibly mangrove-like) community, or they may have occupied drier (possibly back-swamp) habitats. There are currently no obvious criteria for distinguishing between elements of upland and lowland communities in the area.

Thus, two coarsely resolved putative communities persisted throughout the ca. 10my depositional history represented by the Bearreraig floral assemblages: an inland (presumably basinal) community characterized by *Phlebopteris woodwardii* and a coastal (possibly delta-swamp) community characterized by the *Brachyphyllum–Taxodioxylon* aggregate. Given the probability of severe taphonomic depletion, neither *Phlebopteris* nor *Brachyphyllum s. l.* can be confidently regarded as having dominated their respective communities, though they must have been important components. Interestingly, the Bearreraig fern assemblages are paralleled in the Jurassic adpression floras of eastern Scotland (Bathonian–Kimmeridgian: Harris and Rest, 1966; Van Konijnenberg-Van Cittert and Van Der Burgh, 1989) and Yorkshire (Aalenian–Bajocian: Harris, 1961–1979), but the gymnosperm assemblage is not; in short, gymnosperm communities appear more regional in composition that fern communities. Sadly, the current qualitative data from Bearreraig (Fig. 1) are insufficient to (1) identify quantitative changes in relative contributions from the two putative communities through the succession, or (2) support further ecological subdivision. Nonetheless, they have prompted investigations into the potential calorific value gained from the plants by the associated herbivorous dinosaurs (Dissanayake, 1996).

Paleoclimatic and Evolutionary/Biostratigraphic Implications

During the Aalenian–Bajocian, Scotland was marginal to a large aggregation of continents to the east and an ocean to the south and west (Scotese and McKerrow, 1990). The Bearreraig flora grew at a latitude of ca. 35°N, in a dominantly semi-arid climatic zone that did not ameliorate until the end of the Bajocian, when persistent brackish water lagoons occupied the Hebrides Basin, indicating reliable input of freshwater. Atmospheric oxygen approximated present-day levels (22 ± 2%: Berner and Canfield, 1989), thereby offering similar environmental constraints to plant respiration.

Sedimentation patterns in the region suggest the following tentative scenario: tectonic rejuvenation of the hinterland generated considerable topographic relief. This steepened the river gradients, increasing erosion of the coarse-textured putative source rocks in the upper reaches and also increasing the load-carrying capacity of the lower reaches. Precipitation was low and strongly seasonal; wet periods allowed transport of large volumes of bedload, and dry periods inhibited pedochemical weathering (which decreases overall particle size) and development of continuous vegetation cover (which reduces erosion rates). Once the clastics reached river mouths, strong longshore tides provided optimal conditions for redistribution across the shelf.

This sedimentologically based hypothesis can be evaluated by morphotypic interpretation of the plant megafossils. With the exception of the rare ferns *Cladophlebis* and *Coniopteris*, all of the species present had thick, fleshy leaves that were probably thickly cuticularized. Stomata are confined to the abaxial surfaces of leaves and recessed behind large subsidiary cells (*Phlebopteris*, *Hausmannia*, cf. *Otozamites*, *Nilssonia*) or protected by pentagonal arrays of subsidiary cells (*Brachyphyllum s. l.*). In addition, *Phlebopteris* fronds were hirsute. All of these features characterize xeromorphy in extant plants and have been cited as xeric specializations in other Mesozoic species (Wing et al., 1992). The apparent co-existence of an araucarian (*Brachyphyllum* cf. *mamillare*) and a cheirolepidiacean (*B.* cf. *crucis*) in a similar deltaic paleoenvironment at Bearreraig may reflect iterative evolution, wherein severe extrinsic environmental stress induced similar vegetative morphologies in only distantly related clades of conifers. The classic cheirolepidiacean spore genus *'Classopollis'*, present at low frequencies at Bearreraig, is also routinely associated with semi-arid vegetation.

However, the paleobotanical evidence for at least seasonal tropical aridity is prone to over-interpretation. The same robust characteristics of the Bearreraig organ-species that indicate xeric adaptation also undoubtedly aided their survival through severe and probably prolonged taphonomic 'filtration'; thus, they may provide a distorted impression of the overall morphological 'facies' of their source communities. Independent paleoclimatic data are desirable, perhaps exploiting tree-ring patterns and isotopic signatures.

Only two plant megafossil genera occur sufficiently frequently through the Bearreraig succession to monitor putative evolutionary changes: the fern *Phlebopteris* and the conifer *Brachyphyllum–Taxodioxylon.* No evidence of morphological change was detected in *Phlebopteris*, and evolutionary interpretation of the conifers is compromized by the paucity of reproductive organs and the difficulty of delimiting biologically-meaningful species among the vegetative fragments. However, more detailed scrutiny of the numerous, petrifed, conifer-dominated megafossil assemblages at Bearreraig provides an unusually good opportunity to increase taxonomic resolution, as histological characters of leaves and twigs can be combined with characters representing gross morphology and epidermal features. Male cones are desirable to allow correlation of in situ and dispersed conifer pollen.

Admittedly, further paleobotanical investigation will not negate the problems of distinguishing true evolutionary innovation from local ecological replacement; at a local level, speciation is difficult to distinguish from immigration, and global extinction from regional loss (extirpation).

Hypotheses of subtle environmental change affecting community structure may be better tested using the miospore assemblages rather than the less diverse and apparently taphonomically more modified megafloras. Unfortunately, few of the spore-species have been correlated with their source plants, and species delimitation is especially poor among the abundant bisaccate pollen types that lack obvious correlates among the conifer megafloras. The dispersed miospore assemblages vary little in composition throughout the succession (Riding, 1991), indicating remarkable ecological and evolutionary conservatism in the terrestrial realm. Intriguingly, this pattern contrasts with the unusually high species turnover evident in many associated groups of marine organisms, including ammonites (Morton, 1990), foraminifera (Gregory, 1990), and dinoflagellates (Riding, 1991).

Conclusions

The presence of numerous petrified plant megafossil assemblages within the Bearreraig succession ostensibly provides a rare opportunity to examine time-related phenomena such as evolution, large-scale changes in regional floral composition, and (crudely and on a much shorter time-scale) local ecological succession. It also offers the potential to relate these biotic phenomena to changing environmental parameters. However, in the absence of reliable and constant periodicity of deposition, such interpretations require assumptions regarding the biostratigraphic and lithostratigraphic completeness of the section and time-averaging of depositional rates between supposedly fixed and accurately dated stratigraphic markers (in this instance, the primarily ammonite-delimited Tithonian–Aalenian, Aalenian–Bajocian, and Bajocian–Bathonian boundaries). Such assumptions may have some validity for successions deposited in large, persistent lakes (over relatively short time periods) and abyssal oceans (over longer time periods), but Bearreraig is unlikely to demonstrate an equivalent level of temporal acuity. Variable sedimentation rates, and probable depositional breaks, severely reduce the likelihood of accurately dating speciation events. Assuming that the succession represents a ca. 10my period, megafloras have only been preserved on average once every 0.7 my. This level of time resolution, together with the highly allochthonous nature and presumed severe taphonomic depletion of the megafloras, only permits detection of large-scale and exceptionally long-term ecological changes.

However, this study does indicate the potential value of considering the taphonomy and paleo-ecology of biotas that are unusually diverse, both in terms of the range of higher taxa present and, especially, the modes of life of individual species. At Bearreraig, the highly allochthonous terrestrial floras outlined in this paper provide a stark contrast with the dominantly autochthonous and/or parachthonous marine biotas (Bateman, 1991). Also, the Bearreraig assemblages contradict the commonly expressed opinion that prodelta and shelf sediments are poor sources of paleobotanical material. Indeed, the suggested mechanism of deposition at Bearreraig - rafting followed by waterlogging at the zone of water-column inversion - is probably generally applicable to estuarine and deltaic environments, providing a dynamic, chemically complex system that allows formation of calcareous nodules and thus fortuitously permineralizes plant debris. Thus, paleobotanical surveys of similar depositional settings throughout the Phanerozoic might prove unexpectedly rewarding.

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References

BARALE, G. 1981. La Paléoflore Jurassique du Jura Français. Documents de Laboratoire de Géologie Lyon 81.

BATEMAN, R. M. 1991. Palaeoecology, p. 34–116. *In* C. J. Cleal (ed.), Plant Fossils in Geological Investigation: The Palaeozoic. Horwood, Chichester, U.K.

____, AND N. MORTON. 1994. New petrified Middle Jurassic floras from nearshore marine sediments at Bearreraig, Skye, W. Scotland. American Journal of Botany, 81 (abstracts supplement):88.

BEHRENSMEYER, A. K., R. HOOK, C. E. BADGLEY, J. A. BOY, R. E. CHAPMAN, P. DOBSON, R. A. GASTALDO, R. W. GRAHAM, L. D. MARTIN, P. E. OLSEN, R. A. SPICER, R. E. TAGGART AND M. V. H. WILSON. 1992. Paleoenvironmental contexts and taphonomic modes, p. 14-136. *In* A. K. Behrensmeyer, J. D. Damuth, W. A. DiMichele, R. Potts, H.-D. Sues and S. L. Wing (eds.), Terrestrial Ecosystems Through Time. Chicago University Press, Chicago.

BERNER, R. A., AND D. E. CANFIELD. 1989. A new model for atmospheric oxygen over Phanerozoic time. American Journal of Science, 289:333-361.

DISSANAYAKE, R. 1996. Macrotaxonomic Survey of the Embryobionta, with an Emphasis on Calorific Values with Implications for Vertebrate Herbivory. MSc thesis, Royal Botanic Garden Edinburgh/University of Edinburgh.

DOWER, B. L., AND R. M. BATEMAN. 1998. Morphological and anatomical investigations of exceptionally preserved cycadophyte leaves from the Jurassic of Skye, NW Scotland. American Journal of Botany, 85 (abstracts supplement):72-73.

GREGORY, J. 1990. Foraminiferal biostratigraphy and palaeoecology across the Aalenian/Bajocian of the proposed basal Bajocian stratotype at Bearreraig Bay, NE Skye (Scotland). Memoria Descrittive della Carta Geologica d'Italia, 40:257-268.

HARLAND, W. B., R. V. ARMSTRONG, A. V. COX, L. E. CRAIG, A. G. SMITH, AND D. G. SMITH. 1989. A Geologic Time Scale 1989. Cambridge University Press, Cambridge.

HARRIS, T. M. 1937. The fossil flora of Scoresby Sound, East Greenland. 5. Stratigraphic relations of the plant beds. Meddelelser om Grønland, 112:1-114.

____. 1961–1979. The Yorkshire Jurassic Flora (5 vols.). British Museum, London.

____, AND J. A. REST. 1966. The flora of the Brora coal. Geological Magazine, 103:8-109.

HART, J. A. 1987. A cladistic analysis of conifers: preliminary results. Journal of the Arnold Arboretum, 68:269-307.

JONES, T. P., AND N. P. ROWE (eds.) In press. Fossil Plants and Spores: Modern Techniques. Geological Society of London Special Publication.

MORTON, N. 1990. Bearreraig (Isle of Skye, NW Scotland) as boundary stratotype for the base of the Bajocian Stage. Memoria Descrittive della Carta Geologica d'Italia, 40:23-48.

____. 1992. Late Triassic to Middle Jurassic stratigraphy, palaeogeography and tectonics west of the British Isles, p. 53–68. *In* J. Parnell (ed.), Basins on the Atlantic Seaboard. Geological Society of London Special Publication 62.

____, AND J. D. HUDSON. 1995. Field guide to the Jurassic of the Isles of Skye and Raasay, Inner Hebrides, NW Scotland, p. 209–280. *In* P. D. Taylor (ed.), Field Geology of the British Jurassic. Geological Society of London.

RIDING, J. R. 1991. Toarcian to Bathonian (Jurassic) palynology of the Inner Hebrides, northwest Scotland. Palynology, 15:115-179.

SCOTESE, C. R., AND W. S. McKERROW. 1990. Revised world maps and introduction, p. 1-12. *In* W. S. McKerrow and C. R. Scotese (eds.), Palaeozoic Palaeogeography and Biogeography. Geological Society of London Special Publication, 12.

SEWARD, A. C., AND N. BANCROFT. 1913. Jurassic plants from Cromarty and Sutherland, Scotland. Transactions of the Royal Society of Edinburgh, 48:867-888.

STEWART, W. N., AND G. W. ROTHWELL. 1993. Paleobotany and the Evolution of Plants (2nd edn). Cambridge University Press, Cambridge.

VAN KONIJNENBERG-VAN CITTERT, J. H. A., AND J. VAN DER BURGH. 1989. The flora from the Kimmeridgian (Upper Jurassic) of Culgower, Sutherland, Scotland. Review of Palaeobotany and Palynology, 61:1-51.

WING, S. L., H.-D. SUES, B. H. TIFFNEY, R. K. STUCKY, D. B. WEISHAMPEL, R. A. SPICER, D. JABLONSKI, C. E. BADGLEY, M. V. H. WILSON AND W. KOVACH. 1992. Mesozoic and Early Cenozoic terrestrial ecosystems, p. 326-416. *In* A. K. Behrensmeyer, J. D. Damuth, W. A. DiMichele, R. Potts, H.-D. Sues and S. L. Wing (eds.), Terrestrial Ecosystems Through Time. Chicago University Press, Chicago.

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Jurassic Floras of North China

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Abstract: The Early Jurassic *Neocalamites - Cladophlebis* Flora in the north of China is composed mainly of ferns, Ginkgopsida, conifers and numbers of equisetaceans dominated by *Neocalamites*. It is further divided into the Hettangian-Sinemurian assemblage, the Pliensbachian assemblage, and the Toarcian assemblage.

The Middle Jurassic *Coniopteris-Phoenicopsis* Flora is more extensively distributed than the Early Jurassic one. Ferns are the most developed group, Ginkgopsida ranks second, followed by conifers; Cycadopsida is also common. Abundance of *Neocalamites* greatly decreases, while that of *Equisetites* usually increases substantially. It is also divided into 3 assemblages: the Aalenian-Bajocian *Coniopteris hymenophylloides-Baiera* spp. assemblage, the Bajocian *Coniopteris hymenophylloides-Anomozamites thomasi* assemblage, and the Bathonian *Coniopteris hymenophylloides-Brachyphyllum* assemblage.

A few Late Jurassic plants have been reported, chiefly from Hebei and Liaoning.

Introduction

A broad territory in the north of the Kunlun Mountains, Qinling Mountains and Dabie Mountains is discussed in this paper. The territory includes northwestern, northern and northeastern China. The Jurassic is one of the most widespread systems in this region, mainly characterized by continental deposits, except for a few marine and continental-marine alternating rocks of the Middle-Upper Jurassic which are restricted to the east of Heilongjiang Province, northeastern China. Generally speaking, the Lower Jurassic and the lower part of the Middle Jurassic rocks are coal-bearing units deposited in warm-humid climates and rich in plant and palynomorph fossils. The upper part of the Middle Jurassic and the Upper Jurassic are usually characterized by red beds or multicolored sediments which were deposited in semi-dry or dry climates, with rare plant megafossil and palynomorph assemblages. However, these sediments are relatively rich in Reptilia and freshwater fossils such as Charophyta, Ostracoda, Chonchostraca and Bivalves.

The study of the Jurassic plants of the North of China started in the 19th century, but not until the 1930s did Chinese paleobotanists begin their work (Sze, 1931; 1933). Then, in the 1950s and 1960s, a series of papers was published dealing with the Jurassic plants of Xinjiang (Sze, 1956), Qinghai (Sze, 1959), Gansu (Sze, 1960; Shen, 1961) and Shanxi (Lee, 1955). However, the most systematic research results were completed during the past two decades (Huang and Chow, 1980; Chen et al., 1984; Duan, 1987; Li et al., 1988; Zhang and Zheng, 1987; Mi et al., 1996; Zeng et al., 1995).

The present paper briefly introduces the Jurassic floras in north China based on previous studies by other paleobotanists as well as our current work.

Early Jurassic Floras in Northern China

The Early Jurassic *Neocalamites-Cladophlebis* Flora is best developed in Qaidam Basin, Qinghai and is widely distributed in the North of China (Fig. 1). With more than 100 recorded species, this flora is composed mainly of ferns, Ginkgopsida, conifers and numerous equisetaceans dominated by *Neocalamites*. This flora differs from the Late Triassic one by the disappearance of *Danaeopsis* and *Bernoullia*. Among the ferns, dipteridaceans are quite abundant. *Cladophlebis* appears both in abundance and high diversity and is characterized by large pinnules. The dicksoniaceous fern *Coniopteris* is very rare in the early Early Jurassic but is usually represented by 2 or 3 species with slender fronds and small pinnules in the late Early Jurassic. Cycadopsida is not abundant, while the Ginkgopsida is represented by numbers of *Ginkgoites* and abundant Czekanowskiales. The conifers are characterized by long leaf types including *Elatocladus*, *Pityophyllum*, *Storgaardia* and varied *Podozamites*. *Cycadocarpidium* is common in the early Early Jurassic but decreases in the late Early Jurassic. It can be divided further into 3 assemblages.

Hettangian - Sinemurian Assemblage

Ferns, Ginkgopsida and numbers of *Neocalamites* dominate the assemblage, while Cycadopsida and conifers are usually represented by a few species. The presence of numbers of *Todites*, *Thaumatopteris*, *Phlebopteris*, *Hausmannia*, *Dictyophyllum*, *Clathropteris* and *Cycadocarpidium*, and absence of *Coniopteris*, indicate that this assemblage is very closely comparable with those of the Hettangian-Pliensbachian *Thaumatopteris* flora of eastern Greenland (Harris, 1931-37), floras of the same age in Middle Asia (Vakhrameev, 1991) and the Hettangian-Sinemurian (or Pliensbachian) Guangyintan flora of Southwestern Hunan, China (Zhou, 1984). This assemblage has been discovered from the lower part of the Badaowan Formation of Xinjiang, Xiaomeigou Formation of Qaidam, Qinghai, Daxigou Formation of Gansu and the Hongqi Formation of Jilin Northeast China.

About 15 species have been described from the Xiaomeigou Formation of Qaidam, Qinghai (Li et al., 1988), among which *Neocalamites* sp., *Todites williamsonii*, *Cladophlebis ingens*, *Dictyophyllum* sp., *Zamites* cf. *gigas*, *Czekanowskia pumila*, *C.* cf. *nathorsti*, cf. *Vittifoliolum segregetum*, cf. *Storgaardia spectabilis*, *Podozamites* cf. *mucronatus* and *Elatocladus* sp. occur. Most of them are the same as, or similar to, the *Thaumatopteris* flora of eastern Greenland (Harris, 1931-37).

The flora from the lower part of the Badaowan Formation of Toksun, Turpan, Xinjiang is composed mainly of dipteridacean ferns, such as *Dictyophyllum* cf. *nathorsti*, *Clathropteris elegans*, *Hausmannia* (*Protorhipis*) *nariwaensis*, and matoniaceaen ferns *Phlebopteris microphylla*. *Cycadocarpidium* and, questionably, *Coniopteris* is also present (Wu et al., 1986).

Recently, the authors have collected a large number of plant fossils from the lower part of the Badaowan Formation of Junggar Basin, Turpan Basin and Yanji Basin, Xinjiang. The main species are *Equisetites lateralis*, *Neocalamites hoerensis*, *N. carrerei*, *Todites princeps*, *Clathropteris elegans*, *Cladophlebis suluktensis*, *C. haiburnensis*, *C. hirta*, *C.* cf. *kaoiana*. But no *Coniopteris* has been found. This form now seems to be early Early Jurassic in age. Another possible Hettangian-Sinemurian Flora is known from the Daxigou Formation, Jingyuan, Gansu (Xu, 1986a; Liu, 1982). The more than 30 species are principally ferns, Ginkgopsida and conifers. Cycadopsida is rare. The ferns are represented by *Todites williamsonii*, *T. denticulata*, *Marattiopsis asiatica*, *Thaumatopteris hissarica*, *T.* sp., *Cladophlebis suluktensis*, *C. nebbensis*, *C. jingyuanensis*, *C. tsaidamensis*. *Cycadocarpidium* is rather abundant.

The flora from the Daxigou Formation of Lanzhou (near Jingyuan), is composed of 46 species (Yang and Shen, 1988), with abundant dipteridacean ferns including *Hausmannia*, *Dictyophyllum*, *Clathropteris* and 9 species of *Cladophlebis*, but no *Cycadocarpidium* was recorded. The age is considered to be Early-Middle Lias.

In Jarud Qi of Inner Mongolia, Northeastern China, the coal-bearing unit of the Hongqi Formation has yielded 34 species of plants (Yang and Sun, 1985). *Neocalamites* and ferns are predominant. Ginkgopsida and conifers are common, while Cycadopsida is represented by only one species. The great abundance of *Neocalamites*, *Todites*, *Cladophlebis*, the presence of *Thaumatopteris schenki* and *Cycadocarpidium*, and absence of typical Middle Jurassic or Early-Middle Jurassic elements, such as *Coniopteris* and *Eboracia*, indicate that it belongs to the early Early Jurassic.

Pliensbachian Assemblage

The Pliensbachian assemblage is characterized by numerous species of *Neocalamites*, e.g., *N. carcinoides*, *N. hoerensis*; by the predominance of *Cladophlebis* and reduction of dipteridacean ferns; and by the first appearance of 1-2 species of the dicksoniaceous fern *Coniopteris*. *Cladophlebis* reaches a great diversity, usually comprising 20-30%, even 45%, of the assemblage.

This assemblage is recorded in Xinjiang and Qaidam of Qinghai, but is poorly studied in other areas.

In Qaidam, the Pliensbachian assemblage, with 33 species, is characterized by 45% *Cladophlebis*.

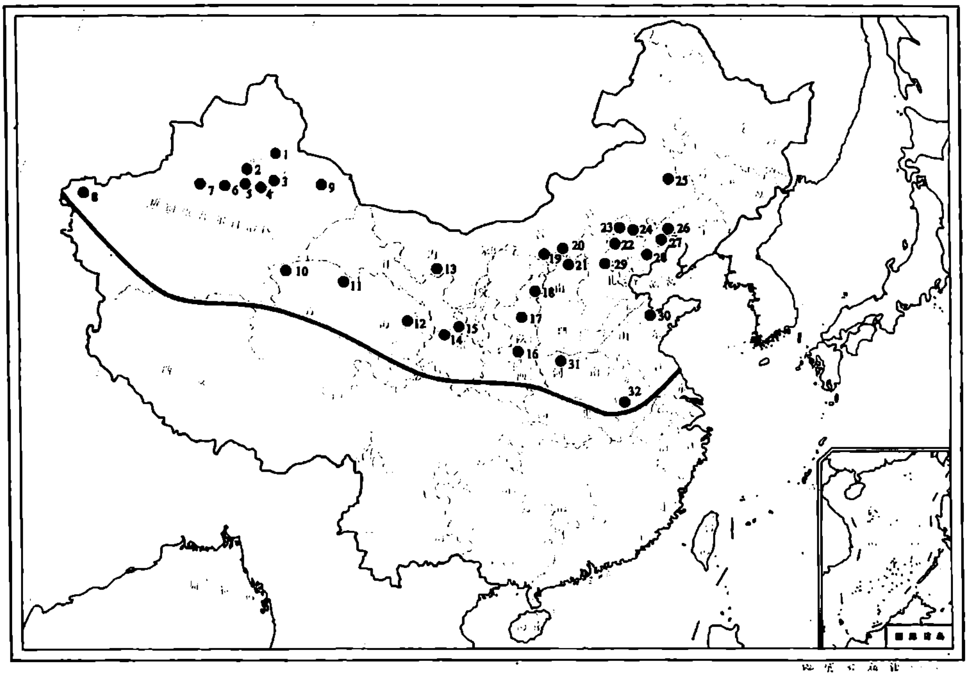

Figure 1. Distribution of the Jurassic plants in the North of China

1. Toutunhe Fm. (J_2^2), Jiangjunmiao, Junggar Basin, Xinjiang; 2. Badaowan Fm. (J_1^1), Sangonghe Fm. (J_1^2) and Xishanyao (J_2^1), Junggar Basin, Xinjiang; 3. Badaowan Fm. (J_1^1), Sangonghe Fm. (J_1^2), Xishanyao (J_2^1), Taoshuyuan, Turpan Basin, Xinjiang; 4. Lower part of the Badaowan Fm. (J_1^1), Toksun, Turpan Basin, Xinjiang; 5. Badaowan Fm. (J_1^1), Awelgou, Turpan Basin, Xinjiang; 6. Badaowan Fm. (J_1^1), Yanji Basin, Xinjiang; 7. Yangxia Fm. (J_1^2) and Kezilnur Fm. (J_2^1), north Tarim Basin, Xinjiang; 8. Saritax Fm. (J_1^1), Kangsu Fm. (J_1^2) and Yangye Fm. (J_2^1), southwest Tarim Basin, Xinjiang; 9. Xishanyao Fm. (J_2^1), Sandaoling, Turpan Basin, Xinjiang; 10. Caishiling Fm. (J_3), Qaidam Basin, Qinghai; 11. Xiaomeigou Fm. (J_1^1) and Dameigou Fm. (J_1^2-J_2), Qaidam Basin, Qinghai; 12. Riyueshan Fm. (J_1), Xining, Qinghai; 13. Qingtujing Group (J_2), Yabulai Basin, Gansu; 14. Daxigou Fm. (J_1^1), Tandonggou Fm. (J_1^2) and Yaojie Fm. (J_2^1), Lanzhou, Gansu; 15. Daxigou Fm. (J_1^1), Longfengshan Fm. (J_2^1) and Wangjiashan Fm. (J_2^2), Jingyuan, Gansu; 16. Fuxian Fm. (J_1^2), Fuxian, Shaanxi; 17. Yan'an Fm. (J_2^1), Yan'an, Shaanxi; 18. Fuxian Fm. (J_1^2) and Yan'an Fm. (J_2^1), Shenmu, Shaanxi; 19. Wudanggou Fm. (J_1^2) and Zhaogou Fm. (J_2^1), Shiguazi Basin, Inner Mongolia; 20. Nansuletu Fm. (J_1^2), Nansuletu, Inner Mongolia; 21. Yongdingzhuang Fm. (J_1^2), Datong Fm. (J_2^1) and Yungang Fm. (J_2^2), Datong, Shanxi; 22. Houcheng Fm. (J_3) and Zhangjiakou Fm. (J_3), Fengning, Hebei; 23. Houcheng Fm. (J_3) and Zhangjiakou Fm. (J_3), Weichang, Hebei; 24. Jiashan Fm. (J_1^2), Chende, Hebei; 25. Hongqi Fm. (J_1^1), Jarud Qi, Inner Mongolia; 26. Beipiao Fm. (J_1^2), Haifanggou Fm. (J_2^1) and Lanqi Fm.(J_2^2), Beipiao, Liaoning; 27. Yixian Fm. (J_3?, K_1?), Chaoyang, Liaoning; 28. Beipiao Fm. (J_1^2), Shimenzhai, Hebei; 29. Yaopo Fm. (J_2), Longmen Fm. (J_2^2) and Yudaishan Group (J_2^2), West Hills, Beijing; 30. Fangzi Fm. (J_2), Weifang, Shandong; 31. Yima Fm. (J_2^1), Henan; 32. Fanghushan Fm. (J_{1-2}), Hefei Basin, Anhui.

Another possible Pliensbachian assemblage of Qinghai is from the coal-bearing unit of the Riyueshan Formation of Xining. It consists of 17 species, of which 8 belong to *Cladophlebis.*

In the upper part of the Badaowan Formation of Junggar, Xinjiang, *Cladophlebis* (with 9 species) accounts for 40% of the assemblage. It is significant that *Coniopteris gaojiatianensis* also appears.

Toarcian Assemblage

The Toarcian assemblage is more extensively distributed in north China than the older assemblages. It is characterized by abundant dipteridacean ferns, such as *Hausmannia* and *Clathropteris*; scale leaf conifers, including *Brachyphyllum* and *Pagiophyllum*; and, even a possible desert plant, *Ephedrites*, in Qaidam. A high proportion (usually 30%-50%) of *Classopollis* pollen from the sediments is frequently recorded. Two or three species of *Coniopteris* commonly appear, including *Coniopteris gansuensis*, *C. simplex* and *C.* cf. *hymenophylloides*. They are characterized by distinctive features, such as small fronds with slender rachis and small pinnules, but relatively larger sori.

This assemblage is somewhat comparable with the Toarcian floras in Southern China and Central Asia. The high proportion of *Classopollis* pollen commonly occurs in plant bearing strata. This feature indicates a climatic change in the Toarcian, similar to that in Siberia and Middle Asia (Vakhrameev, 1991).

In Junggar and Turpan of Xinjiang, the Sangonghe Formation has an assemblage with 20 species. The main elements are *Neocalamites* spp., *Marattiopsis asiatica*, *Todites princeps*, *Phlebopteris polypodioides*, *Coniopteris gansuensis*, *C. simplex*, *C.* cf. *hymenophylloides*, *Hausmannia ussuriensis*, *Clathropteris elegans* and much *Cladophlebis*. Ginkgopsida and conifers are similar to that from the Badaowan Formation.

In the south of Tarim, Xinjiang, there is an abundance of *Nilssonia* and Bennettitales including *Dictyozamites*, *Otozamites*, *Ptilophyllum*, and *Nilssoniopteris*, in the Kangsu Formation. Thus, the climate was hotter than northern areas during the Toarcian (Wu, 1990).

A list of 13 species from the Tandonggou Formation of Gansu was published (Xu, 1986b). *Cladophlebis* and Ginkgopsida, some *Otozamites*, *Zamites*, *Nilssonia* and one *Coniopteris* were reported, besides the commonly observed elements, such as equisetaceans.

In Ordos Basin, Shaanxi, the upper part of the Lower Jurassic Fuxian Formation is composed of red, purple and gray sediments. The 27 species in the Fuxian Formation (Huang et al., 1980) include *Todites williamsonii*, *Phlebopteris digitata*, *Coniopteris tatungensis*, *C. hymenophylloides*, *C.* sp., *Hausmannia papilinacea*, numbers of Ginkgopsida and some conifers. However, the frond of *Coniopteris hymenophylloides* is morphologically different from the Middle Jurassic specimens and possibly belongs to another species.

The Nansuletu Formation of the center of Inner Mongolia is rich in plants, including *Marattiopsis asiatica*, *Asterotheca acumirata*, *Todites daqingshanensis*, *Klukia exilis*, *Phlebopteris brauni*, *P. polypodioides*, *Thinnfeldia rhomboidalis*, *Otozamites hsiangchiensis*, *O.* cf. *mixomorphus*, *Ptilophyllum acutifolium*, *P. contiguum*, *Nilssonia complicatis*, *N. undulata* and *Pseudoctenis* cf. *lanei* (Wang, 1984). It is characterized by ferns and Cycadopsida, but contains no ginkgos or conifers. It is significant that most of the species are usually distributed in the Lower Jurassic of Southern China and Europe.

In the Yongdingzhuang Formation of Datong, Shanxi, more than 30 species have been found (Wang, 1984; Li and Hu, 1984). Ferns are dominant, and Cycadopsida, Ginkgopsida and conifers are also quite common. It is noteworthy that many elements, which commonly appear in the Xianxi Flora Hubei, southern China (Wu et al., 1980), such as *Todites princeps*, *Phlebopteris brauni*, *Coniopteris gaojiatianensis*, *Tyrmia nathorsti*, *Ixostrobus magnificus* and even *Brachyphyllum muensteri*, are present.

Similarly, among the 17 species from the Jiashan Formation of Chende, Hebei (Wang, 1984), *Todites hsiehiana*, *T. princeps*, *Phlebopteris brauni*, *Clathropteris obovata*, *Cladophlebis gigantea*, *Phaphidopteris rugata*, *Ptilophyllum contiguum* and *Swedenborgia cryptomerioides* are Early Jurassic elements usually occurring in South China.

The Beipiao flora from western Liaoning, Northeastern China and Shimenzhai, north Hebei, has been systematically studied in detail in the last decade (Zhang et al., 1987; Mi et al., 1996). The total of about 60 species from western Liaoning consists mainly of ferns, Ginkgopsida and conifers. Four or five species of Cycadopsida are present as well. It is assigned to the late Early Jurassic due to the presence of *Marattiopsis hoerensis*, *M. muensteri*, *Todites denticulata*, *T. williamsonii*, *Thaumatopteris pusilla*, *Dictyophyllum nathorsti*, *Clathropteris meniscioides*, a high percentage of *Cladophlebis* with large pinnules, *Cycadocarpidium* and a few *Coniopteris*.

Middle Jurassic Floras

Middle Jurassic floras are more amply represented than Early Jurassic ones in the northern China due to the more extensive development of Middle Jurassic continental, coal-bearing deposits. The flora of Qaidam, Qinghai and West Hills of Beijing, named the *Coniopteris-Phoenicopsis* Flora, have been studied in detail (Li et al., 1988; Chen et al., 1984).

This flora includes more than 150 species. Ferns are the most developed group, accounting for 30-40% of the flora. Ginkgopsida ranks second, with about 30% of the flora, followed by conifers. Cycadopsida usually makes up of about 10% of the flora, except in a few regions. *Neocalamites* gradually declines in the Middle Jurassic, while *Equisetites* usually increases greatly. Of the ferns, Dicksoniaceae is extremely well developed and reaches its maximum diversity, with *Eboracia*, *Gonatosorus* and more than 10 species *Coniopteris*. Although a few species with small pinnules are represented, i. e. *Coniopteris simplex* (*C. tatungensis*), most have large fronds and pinnules such as *C. hymenophylloides* and *C. burejensis*. *Cladophlebis* persists in abundance, mostly represented by forms with large pinnules. Dipteridaceans and matoniaceans are represented by *Clathropteris* and *Hausmannia*. Cycadopsida is usually represented by great varieties of *Pterophyllum*, *Anomozamites* and *Nilssonia*. However, in the eastern part of this area, such as in western Liaoning, Cycadopsida may even dominate the flora and besides the genera above, *Zamites*, *Ptilophyllum*, *Tyrmia*, *Williamsoniella*, *Zamiophyllum* and *Ctenis* occur as well. Ginkgopsida flourishes both in number of individuals and variety of genera and species. *Ginkgo*, *Ginkgoites*, *Baiera*, *Sphenobaiera*, *Pseudotorellia*, *Eretmophyllum*, *Phoenicopsis*, *Czekanowskia*, *Sphenarion*, *Vittifoliolum* and many reproductive organs are present. Among the Coniferopsida, *Pityophyllum* and *Podozamites*, *Ferganiella* and *Elatocladus* are frequently represented and dominate the coniferous group. The scale leaf conifers, such as *Brachyphyllum* and *Pagiophyllum*, appear in the late Middle Jurassic. *Cycadocarpidium* disappears.

This flora is comparable in composition with the Yorkshire Flora (Harris, 1961-79; Harris and Millington, 1974) and with more than 20 Yorkshire species, particularly those elements of biostratigraphic significance, such as *Coniopteris hymenophylloides*.

Based upon the data of Qaidam Basin, Qinghai (Li et al., 1988) and of West Hills of Beijing, this flora can be separated into 3 assemblages as follows:

1. *Coniopteris hymenophylloides* - *Baiera* spp. assemblage (Aalenian-Bajocian). Dicksoniaceae reaches its maximum diversity, with *Coniopteris hymenophylloides*, *C. tatungensis* and *Eboracia lobifolia* appearing together. Numerous *Cladophlebis* are characterized by large pinnules. Ginkgopsida and conifers are abundant while Cycadopsida makes up a lower percentage.
2. *Coniopteris hymenophylloides* - *Anomozamites thomasi* assemblage (Bajocian). Characterized by the persistent development of dicksoniaceous ferns, by a great reduction of *Cladophlebis* and ginkgos, and by relatively abundant Cycadopsida and long leafed conifers.
3. *Coniopteris hymenophylloides* - *Brachyphyllum* assemblage (Bathonian). This is a depauperate flora and most of the early species have disappeared. Some ferns, including *Coniopteris hymenophylloides*, *C. tatungensis* and *Gleichenites*, occur. The scale leaf conifer *Brachyphyllum* and *Pagiophyllum* appear. Cycadopsida makes up a high percentage of the assemblage and is represented by varied Bennettitales such as *Otozamites* and *Nilssoniopteris*.

These three assemblages succeed one another gradually in time. In most areas, it is difficult to divide them definitely, based on the current level of understanding. Therefore, we will discuss them in geographical order from the west to the east of China.

Northwestern China

Recently, the flora from the coal-bearing Xishanyao Formation of the Turpan and Junggar Basins, and the Kezilnur Formation of north Tarim Basin, Xinjiang, has been systematically studied by the authors. It is composed of more than 100 species dominated by ferns and Ginkgopsida. Cycadopsida, with *Anomozamites*, *Nilssoniopteris*, and *Pterophyllum*, are abundant and more abundant than conifers. This flora is possibly Aalenian to Bajocian in age.

Bathonian plants in Xinjiang are represented by numerous silicified woods yielded in the Toutunhe Formation of Jiangjunmiao, Junggar Basin. However, no leaf fossils have been found.

The Middle Jurassic flora in Gansu has been studied in detail and can be divided into two assemblages. The lower one, from the Yaojie Formation of Lanzhou (Sun, 1986) and the correlative Longfengshan Formation of Jingyuan (Xu, 1986b), is dominated by ferns and Ginkgopsida, and is similar to the Aalenian-Bajocian assemblages of Qaidam. The upper one, from the Honggou Formation of Lanzhou (Sun, 1986) and the correlative Wangjiashan Formation of Jingyuan (Xu, 1986b), is probably of Bajocian age.

A possible Bathonian, and most likely Late Bathonian, flora was recently found by the authors in the upper part of the Qingtujing Group of Yabulai Basin, western Inner Mongolia. It is composed of 20 species and dominated by Cycadopsida, including *Otozamites*, *Taeniopteris*, *Nilssonia* and *Cycadolepis*. Ferns are represented by a few taxa, including *Coniopteris hymenophylloides*, *Cladophlebis delicatula*, and *Gleichenites* with very small pinnules. Only a few specimens of *Neocalamites* sp. and *Ginkgoites* sp. are found.

The Middle Jurassic flora from the coal-bearing unit of the Yan'an Formation of Ordos Basin, with about 50 species, is dominated by Ginkgopsida and ferns (Huang et al., 1980; He, 1987). Cycadopsida is only represented by a few species, while the conifer *Elatocladus* is rather abundant. It is Aalenian-Bajocian in age. The Zhiluo Formation, which lacks coal, lies conformably on the Yan'an Formation. About 30 species from this formation are dominated by ferns and Cycadopsida, while Ginkgopsida is represented by few species.

Northern China

The Mentougou flora of West Hills, Beijing is dominated by ferns and Ginkgopsida; however, Cycadopsida and conifers are also common. The ferns consist mostly of *Cladophlebis* and Dicksoniaceae. More than 20 species of Ginkgopsida include *Ginkgo*, *Ginkgoites*, *Baiera*, *Sphenobaiera*, *Czekanowskia*, *Sphenarion*, *Phoenicopsis* and *Stenorachis*. Cycadopsida is represented of various species of *Nilssonia*, *Pterophyllum*, *Anomozamites* and one or two species of *Otozamites*, *Nilssoniopteris*, *Beania*, *Tyrmia* and *Ctenis*. Among the Coniferopsida, *Pityophyllum* is abundant, *Podozamites*, *Ferganiella* and *Elatocladus* are common, a few *Storgaardia* are represented, but no scale leaf conifer (i.e. *Brachyphyllum* and *Pagiophyllum*) occurs (Chen et al., 1984; Duan, 1987).

The flora is subdivided into two assemblages (Chen et al., 1984). The lower assemblage, with 73 species, in the lower part of the Yaopo Formation, is considered to be Early Aalenian-Middle Bajocian in age. The upper assemblage, with 55 species, in the upper part of the Yaopo Formation and Longmen Formation, has a much lower diversity than the lower assemblage, while the Cycadopsida with 12 species is much increased in number and percentage compared to the lower assemblage. It is Bajocian in age.

The Yudaishan Group of west Beijing and north Hebei yields a possible Bathonian assemblage (Wang, 1984; Chen et al., 1986). Although Dicksoniaceae and *Cladophlebis* are each represented by 5 species, ferns are greatly reduced. Ginkgos are represented by a few elements, while czekanowskias are absent. The presence of *Ptilophyllum*, the scale leaf conifer *Pagiophyllum*, and a high proportion of *Classopollis* indicate a dry and hot climate.

The flora of the Yima Formation, Yima Basin, Henan, consists of more than 100 species and is similar to the Mentougou flora in composition (Zhou, 1995; Zhou and Zhang, 1989; Zeng et al., 1995). It is dominated by ferns, which are represented by 6 species of *Coniopteris* and dozens of *Cladophlebis*. Ginkgopsida, the second most abundant group, has more than 20 species. *Nilssonia*, of Cycadopsida, is abundant and varied. Conifers are represented by *Pityophyllum*, *Elatocladus*, *Sewaediodendron*, *Schizolepis* and *Podozamites*.

The Middle Jurassic floras from the Wudanggou Formation and the Zhaogou Formation of Inner Mongolia (Lee, 1954; Wang, 1984); the Datong Formation (Wang, 1984) and the Yungang Formation (Lee, 1955), Datong, Shanxi; and the Fangzi Formation of Shandong and Fanghushan Formation of Hefei Basin, Anhui are incompletely described. There are 15-30 species recorded at these locations, which are similar to the Mentougou and Yima Floras in their features and ages.

Northeastern China

The flora from the Haifanggou Formation of western Liaoning, northeastern China, is composed of about 130 species (Zhang et al, 1987; Mi et al., 1996). The Cycadopsida, with 15 species of Cycadales and 21 of Bennettitales is the dominant component of the flora. Ferns are of secondary importance. Dicksoniaceae reaches its maximum diversity, with 9 species of *Coniopteris*. Dipteridacean ferns are represented by *Clathropteris* and *Hausmannia*. *Todites* of the Osmundaceae, and *Cladophlebis*, are common. Ginkgopsida is almost the same as the Early

Jurassic except for a few new elements. The coniferous group is dominated by Pinaceae, including *Pityophyllum*, *Pityocladus* and *Pityospermum*. *Equisetites*, rather than *Neocalamites*, dominates the Equisetales. The age is early Middle Jurassic, or Aalenian-Bajocian.

The Lanqi Formation, which overlies the Haifanggou Formation, is composed of volcanic rocks intercalated with sediments in its middle-upper part. The flora from the sediments includes 76 species (Zhang et al., 1987) and is completely dominated by Cycadopsida. The Cycadopsida, with 36 species, is characterized by typical southern China types, such as *Zamites*, *Ptilophyllum*, *Cycadolepis*, *Williamsonia*, *Williamsoniella* and *Bennetticarpus*. Although the dipteridacean ferns have greatly declined, being represented only by *Hausmannia*, it is very abundant. The Dicksoniaceae and other ferns are similar to those from the Haifanggou Formation. Ginkgopsida, with 8 species, occupies the third position and is much reduced in diversity. High diversity and numerous Cycadopsida reflect a hot climate and indicate a Bathonian age.

Late Jurassic Plants

The Jurassic flora in north China began its decline in the late Middle Jurassic due to widespread increased aridity. The Upper Jurassic is largely represented by red beds in the west of the Taihang Mountains and by red beds intercalated with volcanic rocks in the east of the Taihang Mountains. Plant fossils are scarce in these areas.

No undoubted Late Jurassic flora has been reported from northwestern China. Some poorly preserved specimens of *Otozamites* sp. recently have been found by the authors in the base of the Caishiling Formation of Qaidam. This formation is assigned to the Middle Jurassic based on Ostracoda and Conchostraca, but to the Upper Jurassic based on spores and pollen.

Synthesizing the data by Wang (1984) and Xiao et al. (1994), a short list of plants from the Houcheng Formation of Fengning and Weichang, Hebei, includes *Equisetites* cf. *sarrani*, *Neocalamites* sp., *Saginopteris colpodes*, *Coniopteris hymenophylloides*, *C. tatungensis*, *Onychiopsis elongate*, *Pterophyllum angustum*, *Zamites* sp., *Ginkgoites* sp., *Sphenarion lineare*, *Podozamites* sp., *Pityophyllum* sp., *Elatides* sp. and *Xenoxylon latiposum*. Wang and Xiao both considered the age as Late Jurassic, while Zhou (1995) was doubtful about this.

Late Jurassic plants from the Zhangjiakou Formation of Weichang and Fengning, Hebei (Wang, 1984) include *Equisetites rugosus*, *Czekanowskia rigida*, *Leptostrobus sphaericus*, *Sphenarion lineare*, *Pityolepis larixiformis*, *P. pingquanensis*, *Schizolepis fenglingensis*, *S. liaoxiensis*, and *Pagiophyllum* sp. It is characterized by quite abundant conifers and Ginkgopsida, and by a lack of ferns.

It is notable that in recent years many plant fossils have been found from the "Jehol Biota" bed of the Yixian Formation and its correlative units in Hebei, western Liaoning and Jilin. There include angiosperms such as *Archaefructus* (Sun et al., 1998), quite abundant czekanowskias, conifers such as *Pityophyllum* and *Podozamites*, and some unnamed plants. However, it is still a matter of controversy whether the "Jehol Biota" is Late Jurassic or Early Cretaceous in age. The "Jehol Biota" is composed of fish, insects, Conchostraca, Ostracoda, Bivalves, plants, spores and pollen, Amphibia, Reptilia, Aves (including *Confuciusornis*, *Sinosauropteryx*, *Liaoningornis*, *Protarchaeopteryx*) and ancient Mammalia. Most palaeontologists believe that it is Early Cretaceous, while others think it should be Late Jurassic, of similar or slightly younger age than the *Archaeopteryx* bed of Solnhofen, Germany. Some isotopic data from the volcanic rocks of the Yixian Formation support the Early Cretaceous viewpoint, but other data not. Although plant fossils are abundant and very significant, current studies are insufficient to allow a definite age assignment to be made.

References

CHEN FEN, DOU YA-WEI AND HUANG QI-SHENG. 1984. The Jurassic Flora of West Hills, Geology Publishing House, Beijing, 136p.

____, DOU YA-WEI AND MENG XIAN-YING. 1986. Fossil Plants from the Yudaishan Group in Xiahuayuan, Hebei Province. Journal of China College of Mining and Technology, 2:47-61.

DUAN SHU-YING. 1987. The Jurassic Flora of Zhaitang, Western Hills of Beijing, Stockholm, Akademitryck. Edsbruk, 95p.

HARRIS T. M.. 1931-37. The fossil flora of Scoresby Sound, East Greenland, Parts 1-5, Meddelelser om Grønland, 85(2, 3, 5), 112(1, 2).

____. 1961-79. The Yorkshire Jurassic flora, I-III, V. British Museum (Natural History).

____, AND W. MILLINGTON. 1974. The Yorkshire Jurassic flora, IV. 1. Ginkgoales, 2.

Czekanowskiales. British Museum (Natural History). 150p.

HE DE-CHANG. 1987. Yan'an flora from Yuling, Shenmu and Fugu areas, p. 78-85. *In* Qian Lijun et al., (eds.), Jurassic coal-bearing strata and the characteristic of coal accumulation from northern Shaanxi. Northwest University Press, Xi'an.

HUANG ZHI-GAO, AND CHOW HUI-JIN. 1980. Palaeobotany, p. 43-114. *In* Chinese Academy of Geological Sciences, (ed.), Institute of Geology, Mesozoic stratigraphy and Palaeontology of Ordos Basin, Vol. 1. Geological Publishing House, Beijing.

LEE, H. H. 1954. On the stratigraphy of Shihkuaitze Coal Field, Tatsingshan, Inner Mongolia, with special reference to the unconformities. Acta Geol Sinica, 34(4):411-436.

____. 1955. On the age of the Yunkang Series of the Tatung Coal Field in North Shansi. Acta Palaeontologia Sinica, 3(1):25-46.

LI BAO-XIAN, AND HU BIN. 1984. Fossil plant from the Yongdingzhuang Formation of the Datong Coalfield, northern Shanxi. Acta Palaeontologia Sinica, 23(2):135-147.

LI PEI-JUAN, HE YUAN-LIANG, WU XIANG-WU, MEI SHENG-WU, AND LI BING-YOU. 1988. Early and Middle Jurassic strata and floras from northeastern border of Qaidam Basin, Qinghai. Nanjing University Press, Nanjing, 231p.

LIU ZI-JIN. 1982. A Preliminary Study of Early Jurassic Beds and the Flora in Eastern Gansu. Bulletin of Xi'an Institute of Geology and Mineral Resources, Chinese Academy of Geological Sciences, 5:87-100.

MI JIA-RONG, SUN CHUN-LIN SUN YUE-WU, CUI SHANG-SEN, AND AI YONG-LIANG. 1996. Early-Middle Jurassic Phytoecology and Coal-accumulating Environments in Northern Hebei and Western Liaoning. Geological Publishing House, Beijing, 169p.

SHEN KUANG-LUNG. 1961. Jurassic fossil plants from Mienhsien Series in the vicinity of Huicheng Hsien of S. Kansu. Acta Palaeontologia Sinica, 9(2):165-179.

SUN BAI-NIAN. 1986. A Preliminary Study of Middle Jurassic Fossil Plants from Yaojie Coal-Field of Lanzhou. Journal of Lanzhou University (Natural Sciences), 22(1):113-118.

SUN GE, D. L. DILCHER, ZHENG SHAO-LING, AND ZHOU ZHE-KUN. 1998. In Search of the First Flower: A Jurassic Angiosperm: *Archaefructus*, from Northeast China. Science, 282:1692-1695.

SZE, H. C. 1931: Beitrage zur Liassischen Flora von China. Mem. Nat. Res. Inst. Geol., Academy Sinica, *Danaeopsis* and *Bernoullia*. 12:1-85.

____. 1933. Jurassic plants from Shensi. Mem. Nat. Res. Uinst. Geol. Acad. Sinica, 13:65-86.

____. 1956. The fossil flora of the Mesozoic oil-bearing deposits of the Dzungaria-Basin, northwestern Sinkiang. Acta Palaeontologia Sinica, 4(4):461-476.

____. 1959. Jurassic plants from Tsaidam, Chinghai Province. Acta Palaeont. Sinica, 7(1):1-31.

____. 1960. Jurassic plants from Hanxia, Yumen, Kansu, p. 27-31. *In* Institute of Geology and Palaeontology, Institute of Geology, Academia Sinica, Peking College of Geology, eds. Contribution to Geology of the Mt. Chilien, 4(1). Science Press, Beijing.

VAKHRAMEEV V. A. 1991. Jurassic and Cretaceous floras and Climates of the Earth. Cambridge University Press, Cambridge, 318p.

WANG ZI-QIANG. 1984. Plant Kingdom. p. 224-302. *In* Tianjin Institute of Geology and Mineral Resources, Ministry of Geology and Mineral Resources, ed. Palaeontological atlas of North China (2), Mesozoic. Geology Publishing House, Beijing.

WU SHAO-ZU. 1990. Characteristics of Early Jurassic Flora of Xinjiang. Xinjiang Geology, 8(2):119-131.

WU SHUN-QING, YE MEI-NA, AND LI BAO-XIAN. 1980. Upper Triassic and Lower and Middle Jurassic plants from the Hsiangchui Group, western Hubei. Mem Nanjing Institute of Geology and Palaeontology, Academia Sinica, 14:64-131.

WU XIANG-WU, HE YUAN-LIANG, AND MEI SHENG-WU. 1986. Discovery of *Ephedrites* from the Lower Jurassic Xiaomeigou Formation, Qinghai. Acta Palaeobotanica Palynologica Sinica, 1:13-21.

XIAO ZONG-ZHENG, YANG HONG-LIAN, AND SHAN QING-SHENG. 1994. The Mesozoic Stratigraphy and Biota of Beijing Area. Geological Publishing House, Beijing, 135p.

XU FU-XIANG. 1986a. Early Jurassic plants of Jingyuan, Gansu. Acta Palaeontologica Sinica, 25(4):417-425.

____. 1986b. An outline of the Mesozoic plants of Gansu. Gansu Geol., 5:105-116.

YANG SUN, AND SHEN GUANG-LONG. 1988. Early Liassic Fossil Plants from Daxigou Formation of Aganzhen Coal Field, Lanzhou. Journal of Lanzhou University (Natural

Sciences), 24(4):139-144.

YANG XEU-LIN, AND SUN LI-WEN. 1985. Jurassic fossil plants from the southern part of Dahingganling, China. Bulletin of Shengyang Institute of Geology and Mineral Resources, China Academy of Geological Sciences, 12:98-111.

ZENG YONG, SHEN SHU-ZHONG, AND FAN BING-HENG. 1995. Flora from the Coal-bearing Strata of Yima Formation in Western Henan. Sci. Tech. Press. Jiangxi, Nanchan, 92p.

ZHANG WU, AND ZHENG SHAO-LIN. 1987. Early Mesozoic fossil plants in western Liaoning, Northeast China, p. 239-368. *In* YU XI-HAN et al. (eds). Mesozoic stratigraphy and palaeontology of western Liaoning. Geological Publishing House, Beijing.

ZHOU ZHI-YAN. 1984. Early Liassic plants from southwest Hunan, China. Palaeontologia Sinica, Whole No. 165, N. S. A. 7, 85p.

____. 1995. Jurassic Floras, p. 343-410. *In* LI XING-XUE, (ed.), Fossil Floras of China Through the Geological Ages. Guangzhou: Guangdong Science and Techology Press.

____, AND ZHANG BO-LE. 1989. A Middle Jurassic *Ginkgo* with ovule-bearing organs from Henan, China. Palaeontographica, Abt B., 211:113-133.

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Constraining Late Jurassic Paleoclimate within the Morrison Paleoecosystem: Insights from the Continental Carbonate Record of the Morrison Formation (Colorado, USA)

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Abstract: The continental carbonate record of the Upper Jurassic Morrison Formation is well-developed in east-central Colorado and contains sensitive climatic and hydrologic indicators. This record includes 1) a succession of open- and marginal-lacustrine carbonates and 2) pedogenic carbonate within paleosols. Open-lacustrine carbonate facies commonly contain Magadi-type chert, suggesting alkaline-saline lake deposition in semi-arid settings. Marginal-lacustrine carbonates typically contain pseudo-microkarst and microkarst features indicating that lake margins were repeatedly subjected to variations in lake level. The marginal-lacustrine deposits also contain minor pseudomorphs after evaporites and gypsum nodules; bedded evaporite deposits are restricted to the lower Morrison in southeastern Colorado. Pedogenic carbonate displays evidence of multiple cycles of nodule breaking, healing and re-cracking.

Overall, the continental carbonate record suggests that the Morrison paleoecosystem in east-central Colorado was dominated by semi-arid conditions during deposition of the lowermost Morrison, with evaporitic playas in the southeast and freshwater to alkaline-saline lakes to the north. Exposure indicators suggest the majority of the lake complex was deposited under a semi-arid to 'intermediate' (transitional between semi-arid and sub-humid) climate, with distinct wet and dry periods; hydroperiods ranged from ~180-325 days for open-lacustrine areas and ~15-310 days for marginal-lacustrine settings.

Introduction

A semi-arid to arid Late Jurassic climate with strong seasonal contrasts has been suggested for the western United States as a result of the breakdown of the Pangean megamonsoon (Parrish, 1992; 1993a; 1993b) and based on the global sedimentary and fossil records (Hallam, 1992). Global climate models also have suggested a semi-arid to arid climate for most of the Western Interior during the Late Jurassic (Moore et al., 1992; Valdes, 1992). On the Colorado Plateau, sedimentologic evidence from the Morrison Formation corresponds well with a Late Jurassic paleoecosystem dominated by semi-arid to arid conditions. This sedimentologic evidence is derived from two main sources: 1) the siliciclastic and volcaniclastic deposits associated with the extensive alkaline-saline Lake T'oo'dichi' complex of the Brushy Basin Member (Turner and Fishman, 1991); and 2) the eolian deposits associated with the Bluff Sandstone Member (Peterson and Turner-Peterson, 1987).

Controversy over Morrison climatic interpretations has arisen due to apparently conflicting paleontological interpretations from the plant and terrestrial vertebrate records. Paleofloral interpretations of pteridophytes, cycadophytes, and other plants have suggested a humid and subtropical climate for the Colorado Plateau during deposition of the Brushy Basin Member (Jensen, 1966; Tidwell and Medlyn, 1992). Similarly, the metabolic demands of large terrestrial vertebrates, such as dinosaurs, have resulted in the invoking of strong seasonal precipitation to correspond with high seasonal plant productivity to try to explain dinosaur biogeography and taphonomy (Dobson et al., 1980).

One aspect of the sedimentary record that has yet to be fully utilized in the discussion of Late Jurassic paleoclimate are the lacustrine and pedogenic carbonate deposits; this is despite the fact that these deposits are valuable archives of continental paleoclimatology and paleohydrology on a fine temporal scale. The purpose of this paper is to: 1) describe the major climatically- and hydrologically-sensitive sedimentologic indicators associated with the lacustrine and pedogenic carbonate record; and 2) present a paleoclimatic synthesis for east-central Colorado during the Late Jurassic. Although continental carbonates are not volumetrically significant, these deposits are present throughout the Morrison Formation and provide a high resolution record of subtle Late Jurassic climatic and hydrologic changes. This record is particularly well-developed in east-central Colorado, and a detailed examination should provide critical climatic insights and a framework for studies in other parts of the Western Interior where continental carbonates occur.

Geologic Setting

During the Late Jurassic, the Morrison Formation was deposited as part of the thick Middle Jurassic-early Eocene succession in the back-bulge depozone of the Cordilleran foreland basin system (Fig. 1; DeCelles and Currie, 1996) at a paleolatitude of approximately 31-35°N (Peterson, 1988). In east-central Colorado, the Upper Jurassic Morrison Formation contains a well-developed lacustrine carbonate succession within a complex sequence of interbedded floodplain and fluvio-lacustrine deposits with minor evaporitic and marine deposits (Figs. 1 and 2; Peterson and Turner-Peterson, 1987; Peterson, 1994; Dunagan, 1998). This lacustrine complex is different from the alkaline-saline Lake T'oo'dichi' complex, associated with Morrison Formation in the Colorado Plateau region. In east-central Colorado, the lacustrine complex represents numerous shallow, perennial to ephemeral, carbonate lakes and ponds with low-gradient ('ramp'-type) margins surrounded by extensive mudflats and distal alluvial plain deposits (Dunagan, 1998).

Paleoclimatic and Paleohydrologic Indicators

The utilization of lacustrine carbonate deposits as sensitive paleoclimatic and paleohydrologic indicators is not a new approach (Allen and Collinson, 1986; Platt and Wright, 1991). Numerous

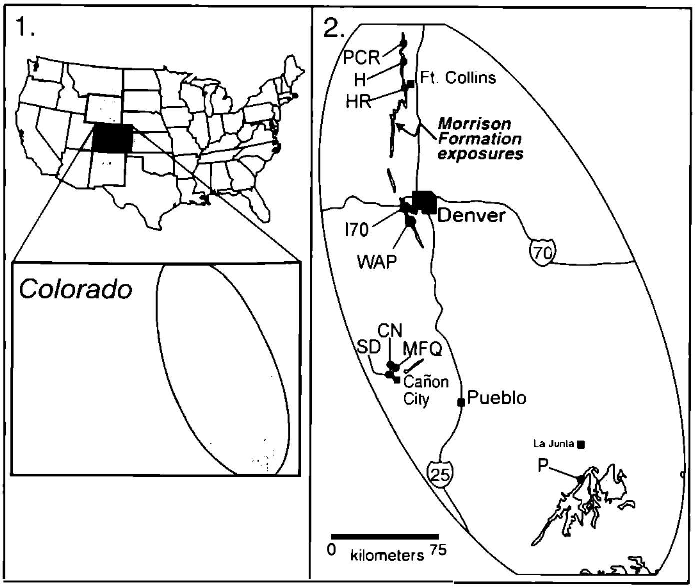

Figure 1. **1**, Paleogeographic distribution (shaded) of the Morrison Formation in the Western Interior basin (from Peterson, 1972) and the primary study area; **2**, location of nine stratigraphic sections in east-central Colorado. Localities: PCR (Park Creek Reservoir); H (Highway 287); HR (Horsetooth Reservoir); I70 (I-70 exposure); WAP (West Alameda Parkway); SD (Skyline Drive); CN (Cope's Nipple); MFQ (Marsh-Felch Quarry); and P (Purgatoire area); see Dunagan (1998) for detailed locality information.

climatically and hydrologically sensitive sedimentologic features are present within the lacustrine complex in east-central Colorado, including palustrine limestones, evaporite deposits, Magadi-type chert, and pedogenic carbonate.

Palustrine Limestones

Marginal-lacustrine carbonate facies containing abundant pseudo-microkarst, pedogenic, and subaerial exposure features are referred to as 'palustrine' (Freytet and Plaziat, 1982; Wright and Platt, 1995). Palustrine features are commonly associated with the open- and marginal-lacustrine facies of the Morrison Formation, which suggests that Morrison lake margins were repeatedly subjected to fluctuations in lake level. Pseudo-microkarst is common and includes: complex voids filled with vadose silt and internal sediment, micritic clasts, and calcite cement (Fig. 3.1); circumgranular, desiccation, horizontal, and septarian cracks (Fig. 3.2); brecciation and grainification (Fig. 3.2); and columnar and stacked rhizoconcretions, and root traces (Fig. 3.3). Microkarstic features also were observed, such as dissolution of metastable skeletal allochems and dissolution voids.

Platt and Wright (1992) noted that climatic control exerts a particularly strong effect on the evolution of palustrine facies. Their study of Carboniferous to Quaternary palustrine examples from Europe and the United States indicates that three characteristic types of palustrine sequences are recognizable, each deposited under different climatic regimes: semi-arid, intermediate, and sub-humid (Table 1; Platt and Wright, 1992). Based on the abundance of pseudo-microkarst features and the relative paucity of evaporites and calcrete horizons, Morrison lacustrine carbonates collectively imply deposition under Platt and Wright's (1992) 'intermediate'-type climatic conditions. The presence of bedded evaporites in the basal Morrison Formation in southeastern Colorado (discussed below) suggests playa-lake deposition under semi-arid climatic conditions.

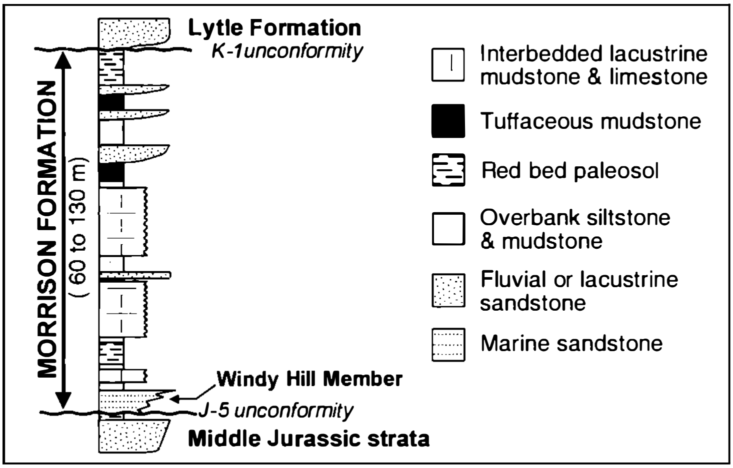

Figure 2. General stratigraphic section of the Morrison Formation in east-central Colorado (Dunagan, 1998). Note that the lacustrine deposits primarily occur within the lower portions of the Morrison Formation.

The progressive effects of subaerial exposure on the lacustrine carbonate substrate produce characteristic sedimentary structures and diagenetic features that can be used to construct a conceptual freshwater carbonate exposure index (Platt and Wright, 1992). Using the exposure features associated with the lacustrine facies associations from the Morrison Formation, an inferred hydroperiod is estimated for both the marginal- (i.e., palustrine) and open-lacustrine facies associations (Fig. 4). The marginal-lacustrine deposits have an estimated hydroperiod of about 15-310 days as compared to approximately 180-325 days for the open-lacustrine facies. The duration

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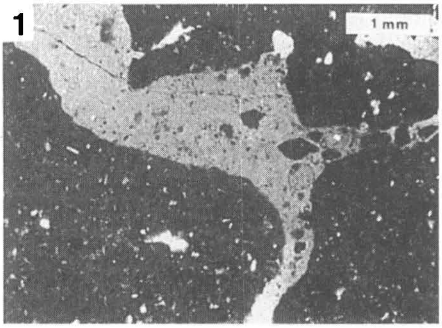

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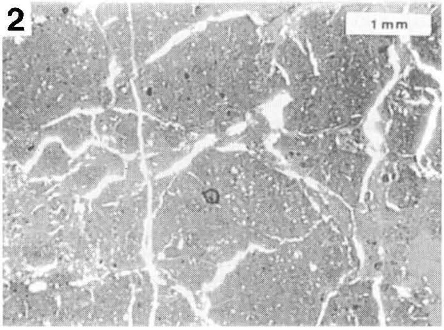

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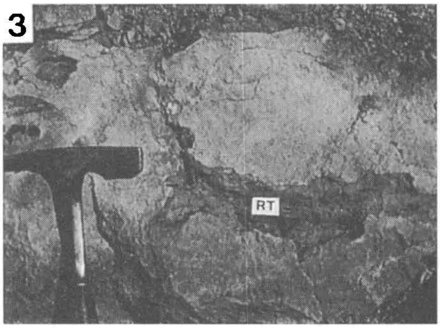

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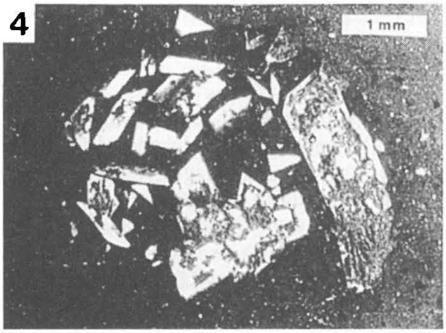

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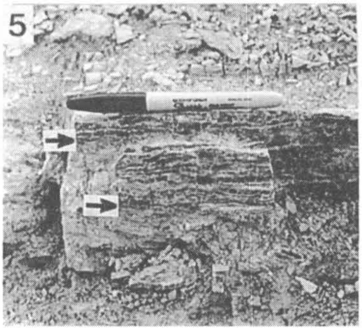

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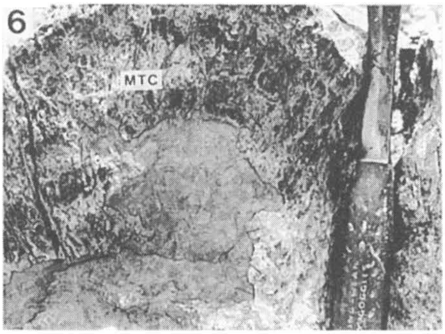

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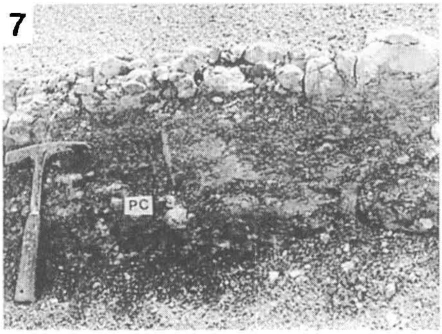

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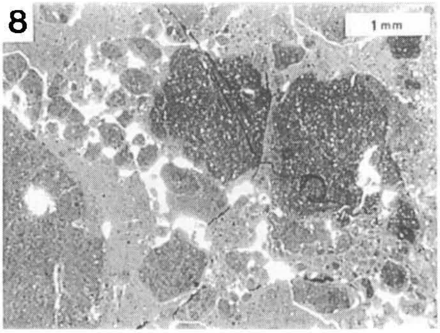

and intensity of exposure experienced by the different lacustrine settings is reflected in the two distinct hydroperiods. This difference in hydroperiods also illustrates the effects of lake-level fluctuations on open- and marginal-lacustrine deposits in low-gradient ('ramp'-type) settings. The exposure index suggests that the Late Jurassic climate was characterized by a distinct wet and dry season during periods associated with lacustrine carbonate deposition.

Climatic Regime	Palustrine Features
Semi-arid	Calcrete development, evaporites, brecciation and nodularization, and laminar coatings
Intermediate	Extensive pedogenic activity (rhizoliths), desiccation and pedogenic cracks, *Microcodium*, pseudo-microkarst and microkarst
Sub-humid	Well-developed organic-rich horizons (coals, lignites), channel deposits, and blackened pebbles ± desiccation cracks and rhizoliths

Table 1. Climatic regimes and associated palustrine features (Platt and Wright, 1992).

Evaporite Deposits

Calcite and chert pseudomorphs after evaporites, evaporite nodules, and evaporite pore-fillings are associated with the lacustrine carbonate deposits. Pseudomorphs after evaporites, such as gypsum and trona?, are present within micritic mudstones and skeletal mudstone to packstones (Fig. 3.4). These pseudomorphs are typically displacive and disrupt the sedimentary fabric. Sweet (1984) noted calcite pseudomorphs after evaporites in the Cañon City area and interpreted them as pseudomorphs after trona, gypsum, shortite, and gaylussite. Other evaporites, such as anhydrite, and pseudomorphs after halite, are present primarily within pore-fillings. The evaporite nodules are rare; they are ellipsoidal (up to 0.7cm diameter) and are composed of calcite pseudomorphs after gypsum and possibly trona. Bedded evaporites are restricted to southeastern Colorado, where a prominent sequence of nodular-bedded gypsum (10-130cm) interbedded with red, gypsiferous mudstone, approximately 25m thick, is present in the lower Morrison Formation.

The presence of evaporites and evaporite nodules in lacustrine carbonates indicates hydrologic periods characterized by a high evaporation rate relative to groundwater, meteoric, and/or fluvial input into Morrison lakes. In addition, the presence of evaporites (as pseudomorphs) and small scale brecciation and grainification associated with the marginal-lacustrine carbonates points toward short-term climatic shifts to dominantly semi-arid conditions. The bedded gypsum-mudstone deposits represent deposition within an evaporitic playa environment and were apparently restricted to the lowermost Morrison Formation in southeastern Colorado. The distribution of these deposits suggests deposition in a hydrologically closed basin under semi-arid to arid conditions. These conditions apparently moderated because a distinct lacustrine carbonate sequence is present above the evaporite deposits.

Magadi-Type Cherts

Magadi-type cherts (MTC) represent important indicators of continental hydrochemistry in lacustrine successions. These cherts have been reported from a variety of ancient and modern lacustrine settings (see references in Surdam et al., 1972; Sheppard and Gude, 1986; Krainer and Spötl, 1998), and more rarely from marginal marine settings (Houser, 1982) in association with sodium silicate minerals (magadiite, $NaSi_7O_{13}(OH)_3{\cdot}3H_2O$) and typically sodium carbonate-bicarbonate minerals (trona, primarily). Chert formation commonly involves a process of evaporite precipitation and subsequent alteration, Na-silicate precipitation, and subsequent alteration to a

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Figure 3. Climatically- and hydrologically-sensitive sedimentologic features of the Morrison Formation; 1, 2, 4, and 8, photomicrographs; 3, 5, 6, and 7, field photographs; 1, 2, pseudo-microkarst features from palustrine carbonates; 3, root trace (RT); 4, calcite pseudomorphs after evaporites; 5, bedded Magadi-type cherts; 6, microbial dome silicified by Magadi-type cherts (MTC); 7, pedogenic carbonate (PC); 8, pedogenic carbonate micro-nodules.

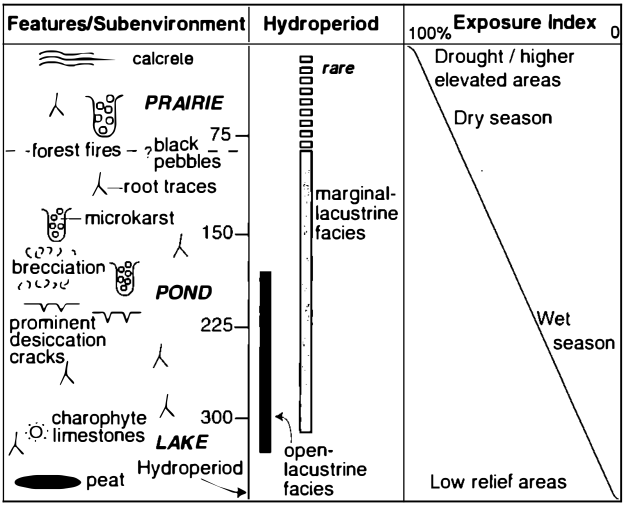

Figure 4. Estimated hydroperiod and exposure index for Morrison lacustrine environments, based upon features observed from the lacustrine facies associations. The hydroperiod represents the number of days the sediment surface is inundated over the course of the year. The exposure index is the percentage of time the sediment surface is exposed. The freshwater carbonate exposure index is based on Platt and Wright (1992, Figure 8, p. 1069).

silica gel, which dehydrates to chert. Although the dynamics of this chert forming process are not completely understood, the restriction of modern MTC to semi-arid environments and limnological settings characterized by high alkalinity (pH > 9.0-9.5; Eugster and Jones, 1968; Hay, 1968; Sheppard and Gude, 1986) make the identification of MTC useful in interpreting paleoclimate and paleohydrogeochemical conditions within ancient lake deposits.

Open-lacustrine facies are typically restricted to the lower portions of the Morrison Formation. Within this lacustrine facies, Magadi-type cherts associated with microbialite- and charophyte-rich carbonates are present (Figs. 3.5 and 3.6; Dunagan et al., 1997). The occurrence of Magadi-type cherts strongly suggests that carbonate sedimentation was common in both freshwater and alkaline lakes in east-central Colorado. Previously, Magadi-type cherts were known only from Wyoming (Surdam et al., 1972); however, the distribution of Magadi-type cherts now includes east-central Colorado (Dunagan et al., 1997) and the Colorado Plateau (Dunagan, unpublished data). Alkaline-saline conditions required for modern MTC occurring in East African rift lakes, such as Lake Natron and Lake Magadi (Hay, 1968), suggest chert formation in alkaline Morrison lakes during climatic periods characterized by semi-arid conditions.

Pedogenic Carbonates

Reddish-gray to drab olive gray paleosols interbedded with lacustrine and fluvial deposits are present in east-central Colorado. The paleosols are characterized by pedogenic slickensides, pedogenic carbonate, and pseudoanticlines. Pedogenic carbonate occurs as sharp to diffuse nodules disseminated through the paleosol matrix (Fig. 3.7), stacked nodules, and columns. Petrographically, the pedogenic carbonate is composed of nodules and micro-nodules displaying circumgranular cracks, craze planes, and evidence for multiple cycles of nodule breaking, healing and re-cracking similar to features observed in marginal-lacustrine carbonates (Fig. 3.8).

Implications for Late Jurassic Climate

The Morrison Formation occupies a unique paleogeographic position that is useful for interpreting climatic evolution in North America during the Mesozoic. Strong monsoonal conditions have been documented for the Triassic and into the Early Jurassic (Dubiel et al., 1987; Parrish, 1993a) followed by the breakdown of the Pangean monsoon during the remainder of the Jurassic (Parrish, 1993b). Eventually, the establishment of zonal circulation and relatively cooler climate occurred, which was characteristic of the Cretaceous (Parrish, 1993b; Schudack, in press). Interpretations of a semi-arid to arid Late Jurassic climate for the western United States are based on sedimentologic evidence of the waning Pangean monsoon (Parrish, 1992, 1993a; 1993b). A semi-arid to arid climate for most of the Western Interior has also been interpreted for the Late Jurassic based on global climate models (Moore et al., 1992; Valdes, 1992). Moore et al. (1992, Figure 6) also suggested that a high-pressure system dominated southwestern North America during the Late Jurassic, which contributed to seasonality with surface temperatures during June/July/August of 30-40°C and 0-20°C during December/January/February. Schudack (in press) recently suggested that taxonomic compositions and oxygen isotope compositions from ostracodes and charophytes point toward a generally arid climate that cooled throughout Morrison time.

Numerous sedimentologic indicators point toward a dry, semi-arid to arid climate, but much of the debate is currently centered around the presence or absence of seasonality in the Morrison paleoecosystem, because there is also evidence that has been interpreted as "wet" indicators. These "wet" indicators include: an abundance of dinosaur remains that must have had large vegetation requirements (Dobson et al., 1980); the necessary plant population would have required large quantities of water; and the "freshwater" lake deposits are abundant and scattered throughout the depositional area (see references in Peterson and Turner-Peterson, 1987; Peterson, 1994).

Sedimentologic evidence from Western Interior Jurassic deposits documents extensive evaporite precipitation and eolian dune fields, and the formation of a large saline-alkaline lake complex toward the latest Jurassic (Peterson, 1994), which appears to agree with the interpretations of Parrish (1993a; 1993b) and the predictions of global climatic models. Peterson (1994) presented two main interpretations of Morrison climate: 1) an overall semi-arid climate that fluctuated between relatively wetter and drier conditions; or 2) an overall dry climate with abundant water resources brought into the region by fluvial systems originating from the western and southern highland regions.

The well-developed continental carbonate record in east-central Colorado points toward a climatic regime for this region characterized by seasonality. In the study area, pedogenic carbonate displays evidence of repeated cracking, micro-nodule formation, healing and re-cracking, which suggests a wet and dry climate as well as a net water deficit in the ancient soils (Cerling, 1991). The type of pseudo-microkarst features associated with Morrison open- and marginal-lacustrine carbonates (see Dunagan et al., 1996; Dunagan, in review) are common from climatic regimes 'intermediate' (*sensu* Platt and Wright, 1992) between semi-arid and sub-humid; the exposure index suggests these palustrine features are associated with a distinct wet and dry season (Fig. 4).

Platt and Wright (1992) demonstrated the close association between the hydroperiod, seasonality, and exposure features in modern freshwater marshes, which serve as an analog for "palustrine" limestone forming environments. Platt and Wright (1992) noted that a freshwater exposure index, similar to the marine exposure index of Ginsburg et al. (1977), could be developed thereby linking the development of characteristic pseudo-microkarstic and exposure features found in marginal lake settings to hydroperiod and seasonality (Fig. 4). Assuming that this model is applicable to the lacustrine carbonate record and that the spectrum of features observed from the marginal-lacustrine carbonates of the Morrison Formation is representative of the average climatic and hydrologic conditions, Morrison lakes and ponds had sum hydroperiod of 15-325 days, corresponding to an exposure index between 15 and 98%. Open-lacustrine deposits have a hydroperiod of ~180-325 days; marginal-lacustrine settings had a greater hydroperiod range of about 15-310 days. This broad hydroperiod and exposure range suggests that Morrison open- and marginal-lake settings were constantly evolving as the lakes shallowed and each basin eventually filled.

Conclusions

1. The Morrison lacustrine carbonate succession in east-central Colorado contains sensitive sedimentologic indicators that record variations in paleoclimate and paleohydrology.
2. The abundance and type of pseudo-microkarst associated with the lacustrine carbonate succession indicate the Late Jurassic climate was characterized by a distinct wet and dry season that was primarily intermediate between semi-arid and sub-humid conditions. Magadi-type cherts, evaporite deposits, and some types of pseudo-microkarst record brief climatic intervals during lacustrine carbonate deposition associated with more semi-arid conditions.
3. Exposure features associated with the lacustrine facies association indicate that the shallow-water, lake settings within the Morrison paleoecosystem were characterized by a wide range of

hydroperiods, approximately 180-325 days for open-lacustrine areas and about 15-310 days for marginal-lacustrine settings.

References

ALLEN, P. A., AND J. D. COLLINSON. 1986. Lakes, p. 63-94. *In* H. G. Reading (ed.), Sedimentary Environments and Facies. Blackwell Scientific Publications, Oxford.

CERLING, T. E. 1991. Carbon dioxide in the atmosphere: evidence from Cenozoic and Mesozoic paleosols. American Journal of Science, 291:377-400.

DECELLES, P. G. AND B. S. CURRIE. 1996. Long-term sediment accumulation in the Middle Jurassic-early Eocene Cordilleran retroarc foreland-basin system. Geology, 24:591-594.

DOBSON, P., A. K. BEHRENSMEYER, R. T. BAKKER, AND J. S. MCINTOSH. 1980. Taphonomy and paleoecology of the dinosaur beds of the Jurassic Morrison Formation. Paleobiology, 6:208-232.

DUBIEL, R. F., R. H. BLODGETT, AND T. M. BOWN. 1987. Lungfish burrows in the Upper Triassic Chinle and Dolores Formations, Colorado Plateau. Journal of Sedimentary Petrology, 57:512-521.

DUNAGAN, S. P. 1998. Lacustrine and palustrine carbonates from the Morrison Formation (Upper Jurassic), east-central Colorado, USA: implications for depositional patterns, paleoecology, paleohydrology, and paleoclimatology. Unpublished Ph.D. dissertation, University of Tennessee at Knoxville, Knoxville, 276 p.

____. In review. Lacustrine and palustrine carbonate deposition within the Upper Jurassic Morrison Formation (east-central Colorado, U. S. A.). Journal of Sedimentary Research, 25 pages, 10 figures, 2 tables.

____, T. M. DEMKO, S. G. DRIESE, AND K. R. WALKER. 1996. Lacustrine and palustrine carbonate facies of the Morrison Formation (Upper Jurassic): implications for paleoenvironmental reconstructions. Geological Society of America Abstracts with Programs, 28(6):336.

____, S. G. DRIESE, AND K. R. WALKER. 1997. Paleolimnological implications of Magadi-type Cherts from lacustrine carbonates in the Morrison Formation (Upper Jurassic), Colorado, U.S.A. Geological Society of America, Abstracts with Program, 29(6):270.

EUGSTER, H. P., AND B. F. JONES. 1968. Gels composed of sodium-aluminum silicate, Lake Magadi, Kenya. Science, 161:160-163.

FREYTET, P., AND J.-C. PLAZIAT. 1982. Continental Carbonate Sedimentation and Pedogenesis - Late Cretaceous and Early Tertiary of Southern France. Contributions to Sedimentology No. 12, 213 p.

GINSBURG, R. N., L. A. HARDIE, O. P. BRICKER, P. GARRETT, AND H. R. WANLESS. 1977. Exposure index: a quantitative approach to defining position within the tidal zone, p. 7-11. *In* L. A. Hardie (ed.), Sedimentation on the Modern Carbonate Tidal Flats of Northwest Andros Island, Bahamas. John Hopkins University Press, Baltimore.

HALLAM, A. 1992. Jurassic climates as inferred from the sedimentary and fossil record, p. 79-88. *In* J. R. L. Allen, B. J. Hoskins, B. W. Sellwood, R. A. Spicer, and P. J. Valdes (eds.), Palaeoclimates and Their Modelling. Chapman and Hall, London.

HAY, R. L. 1968. Chert and its sodium-silicate precursors in sodium-carbonate lakes of East Africa. Contributions to Mineralogy and Petrology, 17:255-274.

HOUSER, B. B. 1982. Chert derived from magadiite in the Middle Eocene McBean Formation, South Carolina. Geological Society of America, Abstracts with Program, 14(1-2):27.

JENSEN, J. A. 1966. Foreword, p. 137-171. *In* M. E. J. Chandler (ed.), Fruiting organs from the Morrison Formation of Utah, USA. Bulletin of the British Museum of Natural History 12.

KRAINER, K., AND C. SPÖTL. 1998. Abiotic silica layers within a fluvio-lacustrine succession, Bolzano Volcanic Complex, northern Italy: a Permian analogue for Magadi-type cherts? Sedimentology, 45:489-505.

MOORE, G. T., D. N. HAYASHIDA, C. A. ROSS, AND S. R. JACOBSON. 1992. Paleoclimate of the Kimmeridgian/Tithonian (Late Jurassic) world: I. Results using a general circulation model. Palaeogeography, Palaeoclimatology, Palaeoecology, 93:113-150.

PARRISH, J. T. 1992. Jurassic climate and oceanography of the Pacific region, p. 79-88. *In* J. R. L. Allen, B. J. Hoskins, B. W. Sellwood, R. A. Spicer, and P. J. Valdes, (eds.), Palaeoclimates and Their Modelling. Chapman and Hall, London.

____. 1993a. Climate of the supercontinent Pangea. Journal of Geology, 101:215-233.

____. 1993b. Mesozoic climates of the Colorado Plateau, p. 1-11. *In* M. Morales (ed.), Aspects of Mesozoic Geology and Paleontology of the Colorado Plateau: Museum of Northern Arizona Bulletin 59.

PETERSON, F. 1988. Pennsylvanian to Jurassic eolian transportation systems in the western United States. Sedimentary Geology, 56:207-260.

____. 1994. Sand dunes, sabkhas, streams, and shallow seas: Jurassic paleogeography in the southern part of the Western Interior basin, p. 233-272. *In* M. V. Caputo, J. A. Peterson, and K. J. Franczyk, (eds.), Mesozoic Systems of the Rocky Mountain Region, USA.

____, AND C. E. TURNER-PETERSON. 1987. The Morrison Formation of the Colorado Plateau: recent advances in sedimentology, stratigraphy, and paleotectonics. Hunteria, 2(1):1-18.

PETERSON, J. A. 1972. Jurassic system, p. 177-189. *In* W. W. Mallory, (ed.), Geologic Atlas of the Rocky Mountain Region. Rocky Mountain Association of Geologists, Denver, Colorado.

PLATT, N. H., AND V. P. WRIGHT. 1991. Lacustrine carbonates: facies models, facies distribution, and hydrocarbon aspects, p. 57-74. *In* P. Anadon, Ll. Cabrera , and K. Kelts, (eds.), Lacustrine Facies Analysis. Special Publication Number 13 of the International Association of Sedimentologists.

____, AND ____. 1992. Palustrine carbonates and the Florida Everglades: towards an exposure index for the fresh-water environment? Journal of Sedimentary Petrology, 62:1058-1071.

SCHUDACK, M. (in press). Ostacoda (marine/nonmarine) and paleoclimate history in the Late Jurassic of Central Europe and North America. Marine Micropaleontology.

SHEPPARD, R. A., AND A. J. GUDE, 3rd. 1986. Magadi-type chert - a distinctive diagenetic variety from lacustrine deposits, p. 335-345. *In* F. A. Mumpton (ed.), Studies in Diagenesis. U.S. Geological Survey Bulletin 1578:

SURDAM, R. C., H. P. EUGSTER, AND R. H. MARINER. 1972. Magadi-type chert in Jurassic and Eocene to Pleistocene rocks, Wyoming. Geological Society of America Bulletin, 83:2261-2266.

SWEET, R. G. 1984. Evidence for playa sedimentation in the Morrison Formation, Canon City, Colorado. Geological Society of America, Abstracts with Programs, 16(6):672.

TIDWELL, W. D., AND D. A. MEDLYN. 1992. Short shoots from the Upper Jurassic Morrison Formation, Utah, Wyoming, and Colorado, USA. Review of Palaeobotany and Palynology, 71:219-238.

TURNER, C. E., AND N. S. FISHMAN. 1991. Jurassic Lake T'oo'dichi': a large alkaline, saline lake, Morrison Formation, eastern Colorado Plateau. Geological Society of America Bulletin, 103:538-558.

VALDES, P. J. 1992. Atmospheric general circulation models of the Jurassic, p. 79-88. *In* J. R. L. Allen, B. J. Hoskins, B. W. Sellwood, R. A. Spicer, and P. J. Valdes, (eds.), Palaeoclimates and Their Modelling. Chapman and Hall, London.

WRIGHT, V. P., AND N. H. PLATT. 1995. Seasonal wetland carbonate sequences and dynamic catenas: a re-appraisal of palustrine limestones. Sedimentary Geology, 99:65-71.

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The Taphonomy and Paleoecology of the Upper Jurassic Morrison Formation Determined from a Field Study of Fossil Localities

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Keywords: Dinosaurs, Vertebrates, Silicified Logs, Fluvio-Lacustrine, U.S. National Parks, Utah, Colorado

Abstract: A paleontological survey of the Upper Jurassic Morrison Formation was conducted in and near several U.S. National Parks in the Rocky Mountain region. Over 500 documented fossil localities provide data on the distribution of diverse fossil material with respect to stratigraphy and lithofacies. The most common fossils are dinosaur bones exhibiting a range of preservation from a few complete skeletons to abundant fragments of bone. This supports the interpretation for the fauna of the Morrison Formation as being an attritional sample. Silicified logs indicate the presence of large trees in the region, although how near the site of preservation cannot be precisely determined. Small vertebrates and invertebrates inhabited near-channel, floodplain environments where water was present. Dinosaurs are distributed throughout the Salt Wash and Brushy Basin members of the Morrison, but other elements of the biota are more restricted stratigraphically.

Introduction

The Upper Jurassic Morrison Formation of the Rocky Mountain region of North America has long been known for its fossil vertebrate fauna. Some of the first great collections of dinosaurs in North America were made from the Morrison Formation (Ostrom and McIntosh, 1966; Breithaupt, 1998), and the mammalian fauna of the Morrison provides one of the most diverse samples of Mesozoic mammals in the world (Engelmann and Callison, 1998). In addition, a less widely known fauna of lower vertebrates and invertebrates and a flora of plant macrofossils and palynomorphs makes the Morrison one of the best sampled biotas in the fossil record (Chure et al., 1998). With so many elements of this ancient ecosystem available, the Morrison is an appealing choice for paleoecological interpretation. This is especially so because of the curiosity excited by the dinosaurs, especially the large sauropods like *Apatosaurus*, so different from any animals alive today, and the resulting wish to interpret them as living animals. The high diversity of these gigantic animals makes the paleoecology of the Morrison unique in the history of North America (Fiorillo, 1998).

There have indeed been many attempts at interpreting the depositional environment of the Morrison Formation from sedimentologic and paleontologic evidence; one of the most recent to consider the dinosaur fauna in this context (Dodson et al., 1980) provides a brief summary of earlier work. Early interpretations were strongly biased by the interpretation of the large sauropod dinosaurs as aquatic, and tended to favor a very wet environment with deep lakes. In contrast, Dodson et al. (1980) suggested that the depositional environment of the Morrison was characterized by strong seasonality, with a wet season and an intense and perhaps prolonged dry season. They cited growing sedimentologic evidence for arid environments in the Morrison and the reinterpretation of the functional morphology of the sauropods suggested they could live as fully terrestrial animals. More recent work on the sedimentology of the Morrison Formation has provided additional and stronger support for arid conditions (Dunagan, et al., 1996; Ekart, 1996; and Demko and Parrish, 1998).

In addition to synthesizing diverse evidence for paleoenvironments of the Morrison, Dodson et al. (1980) focused specifically on the nature of occurrences of the dinosaurs. The distribution of dinosaur localities with respect to the four predominant sedimentary facies they identified, and the degree of disarticulation of specimens, were used to infer the probable habitat preferences of the dinosaurs and the characteristic features of the depositional environment that preserved them. They relied on well-documented quarry samples for their analysis.

Paleontological Survey

As part of a multidisciplinary study of the Morrison Formation, a survey of Morrison fossil localities was undertaken within National Park Service units of the Rocky Mountain region. Parks included in the survey were: Dinosaur National Monument, Arches National Park, Capitol Reef National Park, Colorado National Monument, Black Canyon of the Gunnison National Monument/ Curecanti National Recreation Area, Glen Canyon National Recreation Area, Bighorn Canyon National Recreation Area, and Yellowstone National Park (Engelmann, in press).

The survey sought to locate and document occurrences of fossil material of any type that could be found by examination of surface exposures within the Morrison Formation. Fossil localities were identified wherever fossils exposed at the surface, even if poorly exposed or fragmentary, indicated the presence of fossil material *in situ* in the rock.

The survey documented 574 localities within the Morrison Formation for occurrences of a wide range of fossil material, including dinosaur bone (in varying degrees of completeness of preservation), small vertebrate bone, invertebrates, silicified wood, other plant fossils, and vertebrate and invertebrate traces. In addition to documenting the location and type of fossil material present, each site was also characterized in terms of stratigraphic position within the Morrison Formation and the lithofacies within which it occurred. Locality data gathered by the survey are maintained in the archives of the respective parks surveyed.

Only at Dinosaur National Monument and Black Canyon of the Gunnison National Monument/ Curecanti National Recreation Area was the survey exhaustive, examining all available exposures. In most other parks, stratigraphically and lithologically representative intervals were examined throughout the outcrop area, but not all outcrops were visited. It is largely for this reason that the majority of localities recorded (360) are at Dinosaur National Monument; but in addition, the Morrison at Dinosaur National Monument was more prolific than at some other parks. Other parks at which significant numbers of localities were found are Arches National Park, Capitol Reef National Park, and Colorado National Monument. Many fewer localities were found at the remaining parks probably, in large part, because of more limited exposures. Although each outcrop at Black Canyon of the Gunnison National Monument/Curecanti National Recreation Area was examined thoroughly for fossil material, the total area of Morrison exposures for these parks was small compared with other parks.

By including all fossil occurrences in the survey, rather than just collectable specimens, we can see a more accurate representation of the distribution of fossils with respect to stratigraphy and sedimentary facies. Although the incomplete or fragmentary nature of most of the material, and the methods and intent of the survey greatly limit how specific the taxonomic identifications can be (e.g., dinosaur bone is the most specific identification possible in many cases), it allows us to greatly increase the sample size.

Dinosaur Localities

Because Dodson et al. (1980) focused on the dinosaurs and the interpretation of their habitat and their place in the ecology, it may be informative to consider the dinosaur localities first. It might be argued that the dinosaur localities are one of the most useful taphonomic indicators by virtue of their abundance. Our survey corroborates the casual observation that dinosaur bone is common in the Morrison. With 340 (62%) of the localities, it is the most commonly occurring fossil in the survey, both overall, and at each of the parks.

Why is dinosaur bone the commonest fossil? It seems unlikely that this reflects simple abundance of living dinosaurs in the Morrison ecosystem. Rather, we believe this is the result of the greater durability of the large, massive dinosaur bones compared with the remains of other organisms when subjected to decomposition, weathering, and other pre-burial processes. Preferential preservation of larger animal bones is consistent with patterns observed in modern environments. For example, in their study of modern East African mammal assemblages, Behrensmeyer and Dechant Boaz (1980) concluded that a preservational bias was present for animals greater than 100kg in size. Furthermore, dinosaur bones are generally very robust. More robust bones tend to survive better than less robust bones according to Lyman's (1994) summary of the effects of bone density on their survival. Thus, the sample available for burial and preservation is limited by the time it takes for the remains to disintegrate beyond recognition. For small and/or lightly built organisms, this would be a very short time, while for the dinosaurs this would represent a much longer time. This interpretation demands that, overall, the Morrison represents an attritional sample of the dinosaur populations.

Evidence from the nature of the fossil occurrences in the survey supports the interpretation of an attritional accumulation. If the dinosaur localities are classified according to completeness of specimens, one can see that the number of localities is inversely related to the degree of completeness (Fig. 1). Only five sites represent associated and articulated skeletons (e. g., Fiorillo and May, 1996). There were 27 localities with more than one bone, representing either disarticulated, incomplete specimens, or a concentration of transported elements. Isolated bones were found at 126 localities.

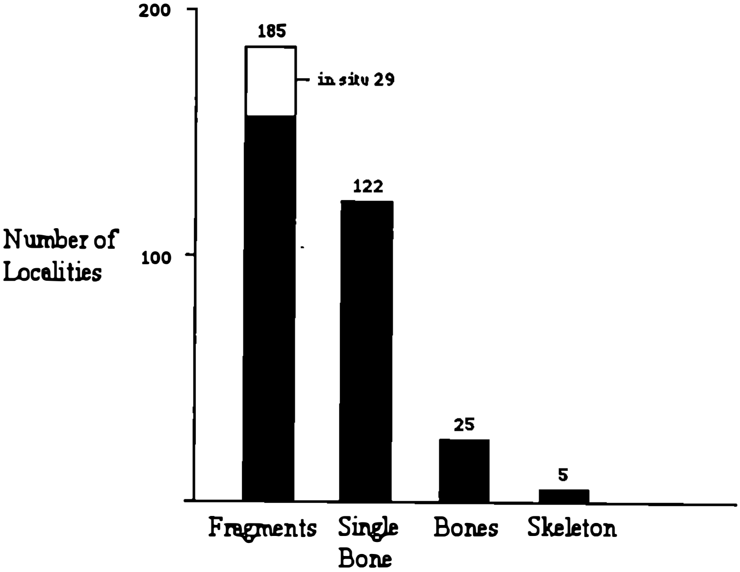

Figure 1. Number of dinosaur localities according to the degree of completeness of the specimen(s) present. Localities at which fragments were observed in place in the rock are represented as a subset of localities for bone fragments.

Only fragments of bone were found at 191 localities. Some of these sites may represent isolated elements that had broken apart on exposure to weathering; others may represent specimens buried as bone fragments. At 53 of these localities, the fragments were observed *in situ* in the rock, indicating the latter explanation. In some cases bone fragments found *in situ* in conglomeratic layers had been rounded by abrasion to form bone pebbles. The range of conditions represented includes every stage of pre-burial modification of a specimen, from the death of the dinosaur to the complete disintegration of its remains. The pattern described above would be expected for a sample that has accumulated and been exposed on a surface over an interval of time sufficient to achieve equilibrium between addition of new material and destruction of old material. It could represent a much longer time.

Fossil Wood

By far the greatest number of plant fossil localities documented by the survey represented silicified wood (120). In fact, after dinosaur bone, silicified wood was the next most common fossil in the Morrison Formation. The wood occurred primarily as segments of logs or small pieces of logs. Although the specimens are not sufficiently complete to estimate the size of the trees that produced them, some are quite large, a meter or more in diameter. Most were relatively short segments, less than 3 meters in length, but one log was more than 20 meters long. With only a few exceptions, virtually all of the fossil wood was found in fluvial channel sandstones and the logs were undoubtedly transported some distance from their growth sites. Broad inferences can be made based on our current appreciation of plant taphonomy. It is reasonable to expect that plant material in lag and point-bar deposits is not far removed from the source (Scheihing and Pfefferkorn, 1984; Spicer, 1989), though channel-margin vegetation has a higher probability of transport downstream (Gastaldo et al., 1987). Waterlogged versus non-waterlogged tree trunks have differing potentials for transport in a fluvial system (Macdonald and Jefferson, 1985), and with no further data available, the distance of transport for these Morrison tree trunks cannot be ascertained. However, further work to determine the depth of channels where the logs occur and, more importantly, the depth of upstream channels within the paleodrainage basin, might be useful in limiting the maximum distance some of the larger specimens could have been transported.

In most cases the silicified wood is rather poorly preserved with the cellular structure largely destroyed and only the gross morphology of the log preserved. This greatly limits the opportunity for taxonomic identification.

Stratigraphic Distribution of Fossils

Within the Colorado Plateau where most survey sites were located, several members within the Morrison can be recognized (Peterson and Turner-Peterson, 1987; Peterson and Turner, 1998). In most areas the Tidwell, Salt Wash, and Brushy Basin members could be readily distinguished. Very few fossils were found in the Tidwell, so our data apply primarily to the Salt Wash and Brushy Basin members. In most cases, it was possible to identify stratigraphic position more specifically as being within the lower, middle or upper part of the Salt Wash, or the lower or upper part of the Brushy Basin, with the upper part divisible again into lower and upper parts. Stratigraphic position of fossil localities is presented here in terms of these 6 subdivisions of the Salt Wash and Brushy Basin members of the Morrison Formation.

Fossils were found throughout both the Salt Wash and Brushy Basin members of the Morrison Formation, although specific types of fossils were restricted to, or decidedly more common within particular stratigraphic intervals. Dinosaur bone was found in all of the stratigraphic intervals (Fig. 2), and it was somewhat surprising to find that it was as common in the Salt Wash as in the Brushy Basin. In fact, more dinosaur localities were found in both the middle (87) and upper (98) parts of the Salt Wash than in the lower part of the upper Brushy Basin (77). Other parts of the Salt Wash and Brushy Basin had still fewer dinosaur localities.

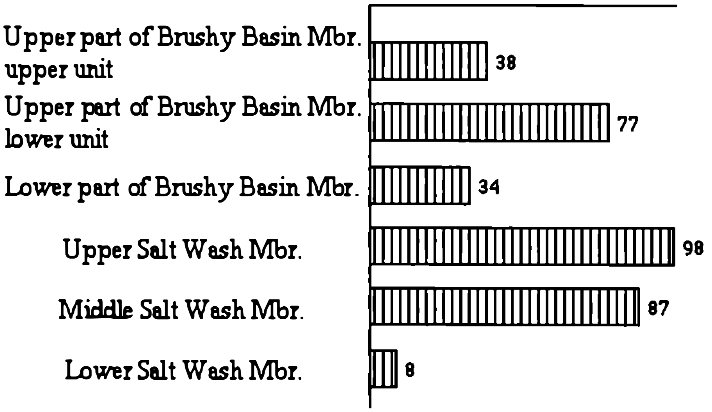

Figure 2. Stratigraphic distribution of dinosaur localities documented by the survey.

Although silicified wood was found throughout both the Salt Wash and Brushy Basin members, it was far more common in the Salt Wash, and most of the large logs were in that member. The greater abundance of fluvial channel sandstones in the Salt Wash than in the Brushy Basin may be responsible, in part, for this distribution because of the close association of the fossil wood with this facies (see below). It is noteworthy that there is a difference along stratigraphic lines in the quality of preservation of the fossil wood. Although abundant, silicified wood found in the Salt Wash was poorly preserved. Silicified plant remains found in the Brushy Basin, on the other hand, were much more likely to preserve fine details of the plant's internal cellular structure. The higher volcanic ash content of the Brushy Basin may have made silica more available in the diagenetic environment, leading to silicification sooner after burial than in the Salt Wash.

Microvertebrates, including mammals, small amphibians and reptiles, have been found only in the lower unit of the upper part of the Brushy Basin. However, they occur only within the mudstones (see below), which are abundant in the Brushy Basin and scarce in the Salt Wash. The number of such sites is small, so it is possible that this lithologic bias is primarily responsible for the stratigraphic restriction.

Another type of fossil that exhibits a marked stratigraphic restriction is the invertebrate traces that occur within a horizon spanning the upper Salt Wash and lower Brushy Basin. This is especially true for deep burrows such as termite nests, which were common within this interval in some areas, but applies to all invertebrate traces. The types of invertebrate traces include termite nests, unionid clam burrows, crayfish burrows, vertical tube burrows, and probable beetle burrows.

Lithofacies

The Morrison Formation was deposited in diverse environments and includes equally diverse lithofacies, but fluvio-lacustrine facies dominate the section in most areas and include virtually all of the fossil localities. For purposes of this summary, fossil localities can be characterized as occurring within one of three broad categories of rock type. Fluvial sandstones, representing channel and near channel environments, are present throughout the section and dominate in some parts of the section, notably the Salt Wash Member in most areas. Mudstones deposited in various overbank environments such as levee, crevasse splay, pond, and other floodplain environments are present throughout the section and dominate in parts of the section, especially the upper part of the Brushy Basin Member. Carbonates occur as thin layers throughout the section, and represent lacustrine, palustrine, and pedogenic environments. These lithofacies equate roughly with the four lithofacies recognized by Dodson et al. (1980), with the exception that, for this study, we have not distinguished between variegated mudstones and drab mudstones as they did. In most cases, the variegated mudstone lithofacies would probably represent the most appropriate equivalent.

Fossil localities are not evenly distributed among the various lithofacies of the Morrison Formation. Most of the localities occurred in sandstone (409); substantially fewer were in mudstones (109), and only a small proportion was found in carbonates (19). To some extent this reflects the dominance of sandstone as the principal lithology of the Salt Wash Member in most areas. However, when localities are broken down according to stratigraphy and lithofacies (Table 1), it is apparent that even in the upper Brushy Basin Member that is dominated by mudstones, localities occur nearly equally in sandstone and mudstone. A similar effect can be seen when considering dinosaur localities alone. Of 340 dinosaur localities in the Morrison, 62 were in mudstone. Almost all of the others were in sandstone. In the upper part of the Brushy Basin, 55 were in mudstone and 54 in sandstone. Dodson et al. (1980) made a similar observation concerning the distribution of the dinosaur localities they studied and that occurred primarily within the Brushy Basin.

Hypotheses that explain the predominance of sandstone as the lithofacies within which fossils are found are not mutually exclusive. One obvious explanation is that organic remains were concentrated by streams by being swept into the channel from the floodplain during floods or as lateral migration of the channel consumed floodplain sediments with whatever had accumulated on or in them. It is also possible that the living organisms were concentrated to some extent along streams. Supporting this interpretation is the observation that most of the mudstones that contain fossil localities are closely associated with fluvial sandstones. In contrast, in parts of Capitol Reef National Park, mudstones of the upper Brushy Basin include extremely little sandstone and are devoid of fossils. Finally, the bias toward sandstone may represent a collection bias to some extent. Exposures of mudstone in the Morrison Formation are usually deeply weathered, and often the mudstones are highly smectitic. Shrinking and swelling of the clays in the weathering zone break up fossils, and the irregular "popcorn" texture of the surface, with numerous cracks, may make small fragments difficult to see. Fossils may have little or no surface expression and be easily overlooked.

Sandstones are more likely to present a relatively unweathered exposure and do not subject fossils to such great stresses in weathering.

The bias toward occurrence in the channel sandstones is especially pronounced for fossil wood. Of 116 fossil wood localities, only 3 were in mudstone. All others were found in sandstone. It seems likely that this is a result of preservational processes, including, and perhaps especially, postdepositional processes. Termites, for which there is good evidence in part of the section, and microorganisms may have consumed woody material even after it was buried, except where the entombing sediment remained saturated for some time after burial.

	Sandstone	Mudstone	Carbonate
Upper Brushy Basin, upper unit	24	21	2
Upper Brushy Basin, lower unit	59	64	4
Lower Brushy Basin	37	6	2
Upper Salt Wash	126	3	0
Middle Salt Wash	128	14	10
Lower Salt Wash	35	1	1

Table 1. Distribution of localities according to stratigraphy and lithology.

A strong bias is also apparent for microvertebrates. The few sites that have produced amphibians, lizards, and mammals (3) have all been found in mudstone. The sites represent overbank sediments interpreted as pond and crevasse splay deposits. Of several other sites with small bone fragments identifiable only as small vertebrate, almost all occurred in mudstone. It is perhaps simplest to explain these occurrences as indicating the habitat of these small vertebrates, living in and around small ponds and other floodplain subenvironments. It is not clear why these organisms are not known from the sandstones. The fact that their remains are very small and delicate and unlikely to survive any significant transport can account for their general absence from the sandstone, but clay pebble lag deposits have preserved equally delicate specimens in Cretaceous and Tertiary sediments. Similar sediments have been observed and examined in the Morrison Formation, but were devoid of fossils.

Although large unionid clams are known from channel sandstones (they are the most common fossils in the Carnegie Quarry at Dinosaur National Monument), most of the unionid clam localities documented by this survey occurred within mudstones of the middle, shaly part of the Salt Wash in the vicinity of Colorado National Monument. These gray mudstones, interspersed with thin carbonate ledges, are very different from most other mudstones that contain fossil localities in the Morrison Formation, and undoubtedly represent a distinct environment. The presence of carbonates with charophytes and gastropods indicates an environment that is more paludal or lacustrine than fluvial.

Conclusion

Although our approach to the taphonomic interpretation of fossils of the Morrison Formation is different from that of Dodson et al. (1980), we arrive at some of the same conclusions. The abundance of dinosaur bone in the Morrison Formation and the nature of occurrences of dinosaur bone are best explained as an attritional sample. Dinosaurs are disproportionately represented because they are the largest elements of the fauna with the most durable remains. Like Dodson et al.

(1980), we found that the dinosaur localities are distributed in both fluvial channel and overbank sediments, indicating that dinosaurs ranged throughout the near-channel and floodplain environments rather than having been restricted to a particular subenvironment. We corroborate their observation that there is a bias toward preservation in the channel sandstones. We agree that this is probably largely the result of concentration of the sample by sedimentary processes, but suggest the possibility that it is also the result to some extent of concentration of the biota along major drainages.

Silicified logs are also abundant in the Morrison and are largely a transported sample, although the distance they were transported cannot be determined. The silicified logs occur almost exclusively in the channel sands, but this may be because of a preservational bias in postdepositional processes.

In contrast, small vertebrates, such as mammals and small herpetofauna, occur preferentially in mudstones. This appears to be the result of habitat preference for small ponds and similar floodplain environments.

The stratigraphic distribution of localities reveals some new information. The dinosaurs appear to be as abundant in the Salt Wash Member of the Morrison as in the Brushy Basin, in spite of the fact that most of the classic dinosaur localities are in the Brushy Basin. Silicified logs, on the other hand, are noticeably more common in the Salt Wash than in the Brushy Basin, probably because of the much greater abundance of sandstone in the former unit. Small vertebrate localities exhibit a complementary distribution, being primarily in the Brushy Basin, possibly because of the close association with mudstone. Invertebrate traces are particularly common and widespread in some areas in the upper part of the Salt Wash and lower part of the Brushy Basin and may provide a useful horizon for interpretation of the stratigraphy and tectonics of the region.

References

BEHRENSMEYER, A. K., AND D. E. DECHANT BOAZ. 1980. The recent bones of Amboseli National Park, Kenya, in relation to East African paleoecology, p. 72-92. *In* A. K. Behrensmeyer and A. P. Hill (eds.), Fossils in the making, vertebrate taphonomy and paleoecology. University of Chicago Press, Chicago, 338 p.

BREITHAUPT, B. H. 1998. Railroads, blizzards and dinosaurs: a history of collecting in the Morrison Formation of Wyoming during the nineteenth century, p. 441-463. *In* K. Carpenter, D. J. Chure, and J. I. Kirkland (eds.), The Upper Jurassic Morrison Formation: An interdisciplinary study, Part 2. Modern Geology, 23.

CHURE, D. J., R. J. LITWIN, S. T. HASIOTIS, K. CARPENTER, AND E. EVANOFF. 1998. The flora and fauna of the Morrison Formation, p. 507-537. *In* K. Carpenter, D. J. Chure, and J. I. Kirkland (eds.), The Upper Jurassic Morrison Formation: An interdisciplinary study, Part 2. Modern Geology, 23.

DEMKO, T. M. AND J. T. PARRISH. 1998. Paleoclimatic setting of the Upper Jurassic Morrison Formation, p. 283-296. *In* K. Carpenter, D. J. Chure, and J. Kirkland (eds.), The Upper Jurassic Morrison Formation: An interdisciplinary study, Part 1. Modern Geology, 22.

DODSON, P., A. K. BEHRENSMEYER, R. T. BAKKER, AND J. S. MCINTOSH. 1980. Taphonomy and paleoecology of the dinosaur beds of the Jurassic Morrison Formation. Paleobiology, 6:208-232.

DUNAGAN, S. P., T. M. DEMKO, S. G. DRIESE, AND K. R. WALKER. 1996. Lacustrine and palustrine carbonate facies of the Morrison Formation (Upper Jurassic): implications for paleoenvironmental reconstructions. Geological Society of America Annual Meeting 1996, Abstracts with Programs, 28:A336.

EKART, D. D. 1996. PCO2 during deposition of the Late Jurassic Morrison Formation and other paleoclimatic/ecologic data as inferred by stable carbon and oxygen. Geological Society of America Annual Meeting 1996 Abstracts with Programs, 28:A252.

ENGELMANN, G. F. (in press). Stratigraphic and geographic distribution of fossils in the upper part of the Upper Jurassic Morrison Formation of the Rocky Mountain region. *In* D. D. Gillette (ed.), Vertebrate fossils of Utah. Utah Geological Survey.

____, AND G. CALLISON. 1998. Mammalian faunas of the Morrison Formation, p. 343-379. *In* K. Carpenter, D. J. Chure, and J. Kirkland (eds.), The Upper Jurassic Morrison Formation: An interdisciplinary study, Part 2. Modern Geology, 23.

FIORILLO, A. R. 1998. Dental microwear patterns from the sauropod dinosaurs Camarasaurus and Diplodocus: evidence for resource partitioning in the Late Jurassic of North America. Historical Biology, 13:1-16.

____, AND C. L. MAY. 1996. Preliminary report on the taphonomy and depositional setting of a new dinosaur locality in the Morrison Formation (Brushy Basin Member) of Curecanti National Recreation Area, Colorado, p. 555-561. *In* The Continental Jurassic, Morales, M. (ed.). Museum of Northern Arizona Bulletin 60.

GASTALDO, R. A., D. P. DOUGLASS, AND S. M. MCCARROLL. 1987. Origin, characteristics, and provenance of plant macrodetritus in a Holocene crevasse splay, Mobile Delta, Alabama. Palaios, 2:229-240.

LYMAN, R. L. 1994. Vertebrate Taphonomy. Cambridge University Press, Cambridge, 524 p.

MACDONALD, D. I. M., AND T. H. JEFFERSON. 1985. Orientation studies of waterlogged wood: a paleocurrent indicator? Journal of Sedimentary Petrology, 55:235-239.

OSTROM, J. H. AND J. MCINTOSH. 1966. Marsh's dinosaurs: the collections from Como Bluff. Yale University Press, New Haven, p. 388.

PETERSON, F. AND C. E. TURNER-PETERSON. 1987. The Morrison Formation of the Colorado Plateau - Recent advances in sedimentology, stratigraphy, and paleotectonics: Hunteria, 2:1-18.

____, AND C. E. TURNER. 1998. Stratigraphy of the Ralston Creek and Morrison formations (Upper Jurassic) near Denver, Colorado, p. 3-38. *In* K. Carpenter, D. J. Chure, and J. Kirkland (eds.), The Upper Jurassic Morrison Formation: An interdisciplinary study, Part 1. Modern Geology, 22.

SCHEIHING, M. H. AND H. W. PFEFFERKORN. 1984. The taphonomy of land plants in the Orinoco Delta: a model for the incorporation of plant parts in clastic sediments of Late Carboniferous age of Euramerica. Review of Palaeobotany and Palynology, 41:2-5-240.

SPICER, R. A. 1989. The formation and interpretation of plant fossil assemblages. Advances in Botanical Research, 16:95-191.

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