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Paleozoic Coral-Sponge Bearing Successions in Austria

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Contents

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1. The Paleozoic of Austria - An Overview

During Variscan and Alpine orogeneses several Paleozoic remnants were dismembered and are now incorporated into the complicated Alpine nappe system. The primary geographic positions and mutual bio(geo)graphic relations of these isolated developments are only poorly understood. A possible arrangement of Paleozoic areas south of the Alpine front, including high grade metamorphosed Paleozoic parts within crystalline complexes, results in a picture shown below (fig. 1).

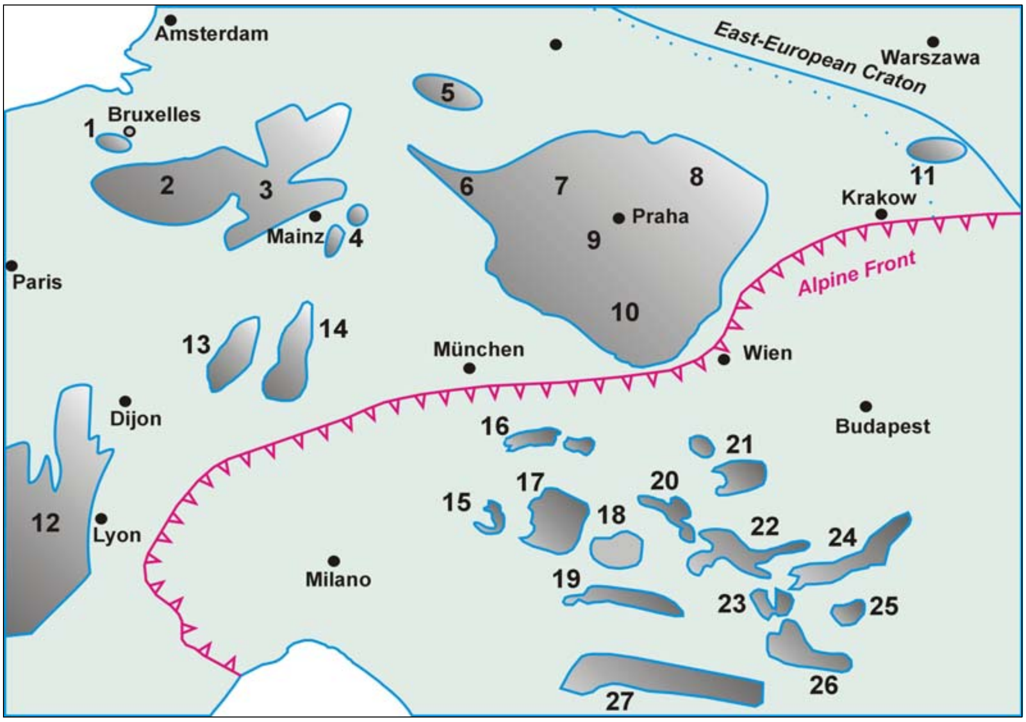

Fig. 1: Variscan regions in Europe. Geographic positions of Paleozoic areas of the Eastern and Southern Alps (15-27) are reconstructed after palinspastic subtraction of alpidic tectonic movements. Redrawn and modified after FAUPL, 2000, and RATSCHBACHER & FRISCH, 1993.

1: Brabant Massif, **2**: Ardennes, **3**: Rhenish Slate Mountains, **4**: Spessart, Odenwald, **5**: Harz, **6**: Thüringerwald, Frankenwald, **7**: Erzgebirge, **8**: Sudetes, **9**: Barrandian, **10**: Bohemian Massif, **11**: Holy Cross Mountains, **12**: Massif Central, **13**: Vosges, **14**: Schwarzwald, **15**: Err-Bernina, **16**: Hohe Tauern, **17**: Sivretta, **18**: Ötztal, **19**: Cristalline Complexes south of the Hohe Tauern, **20**: Quartzphyllites of Innbruck, Radstadt, Ennstal, **21**: Wechsel, **22**: Seckau and Wölzer Alps, **23**: Koralpe, Saualpe, **24**: Greywacke Zone, **25**: Graz Paleozoic, **26**: Gurktal Nappe System, **27**: Carnic Alps, Karawanken Mountains.

Austria's anchizonal to lower greenshist metamorphosed Paleozoic successions are irregularly distributed (fig. 2). Two major regions of Paleozoic developments are distinguished which are separated by the most prominent Alpine fault system, the Periadriatic Line (P.L.). Variscan sequences north of the P.L. form parts of the "Upper Austroalpine Nappe System" whereas sequences south of the P.L. belong to the Southern Alpine System.

Austroalpine Paleozoic areas are the Greywacke Zone of Tyrol, Salzburg, Styria and Lower Austria, the Nötsch Carboniferous, the Gurktal Nappe System, the Graz Paleozoic and some isolated outcrops in southern Styria and Burgenland.

Within Austria's border Paleozoic sequences of the Southern Alpine System are developed in the Carnic Alps and the Karawanken Alps (Southern Carinthia).

Fig. 2: Main regions of anchizonal to lower greenshist metamorphosed Paleozoic strata in Austria. Note the Periadriatic Line (P.L.) separating the Carnic Alps and the Karawanken Mountains (Southern Alps) from other Alpine Paleozoic remnants belonging to the Eastern Alps.

Developmental differences of Austroalpine versus Southalpine areas are visible in different facial and organismic characters as results of independent histories of subsidence rates, amounts of volcanic activities and climatic impacts (SCHÖNLAUB, 1992, 1993; SCHÖNLAUB & HEINISCH, 1993).

2. Review of the Main Weakly Metamorphosed Paleozoic Units in Austria

A. The Greywacke Zone

The Greywacke Zone is a unit of Ordovician to Carboniferous rocks, with fossils either being badly preserved or completely lacking, which represents the base of the Northern Calcareous Alps.
This zone is approximately 23 km in width and has a length of about 450 km.

The Lower Austrian and Styrian parts are named Eastern Greywacke Zone (EGZ), whereas the series in Salzburg and Tyrol belong to the Western Greywacke Zone (WGZ).

The Greywacke Zone represents a thrust complex:
In the EGZ the Noric Nappe together with the lower Kaintaleck- and Silbersberg Nappe is dominated by Lower Paleozoic rocks. They are connected transgressively with the Permo-Mesozoic sequences of the Northern Calcareous Alps.

The lowermost nappe in the EGZ is the Veitsch Nappe which is Carboniferous in age. The different tectonic units of the WGZ have been summarized by SCHÖNLAUB & HEINISCH (1993).

A simplified scheme of the lithostratigraphic units (HUBMANN, 2003, in press.) is shown in fig. 3.

A significant member of the Greywacke Zone are acidic volcanites of Upper Ordovician age. Characteristic for the EGZ is the more than 1500 m thick Blasseneck Quartzporphyry (HEINISCH, 1981) comprising different types of massive ignimbrites, unwelded tuffs and other pyroclasts.

The volcanites are underlain by a more than 1000 m thick sequence of green schists, slates, marls, pyroclastic and basal fragmentites (Wildschönau Slates, Grauwacken Slates). In the Präbichl area (EGZ) the uppermost parts of these sequences include up to 30 m thick limestone lenses containing a rich Condont fauna of Late Caradocian or Early Ashgillian age (FLAJS & SCHÖNLAUB, 1976).

In the WGZ the Wildschönau Slates demonstrate the persistance of the volcanic event up until the Lower Carboniferous. The slates represent turbiditic deepwater sediments interfingering with Upper Silurian to Devonian pelagic limestones, Devonian basaltic lavas and tuffs, as well as Lower Carboniferous basalts. A higher wedge is dominated by carbonates of Silurian to Devonian age. Dolomites prevail in this part.

In the EGZ, chiefly around Eisenerz, the quartzporphyry is overlain by the Polster Quartzite (60 m) and the 13 m thick cystoid limestone belonging to the Ashgillian. The Silurian and Devonian is characterized by up to 350 m thick sequences of different types of limestones. Parts of these limestones - the "eisenführende Kalke" - are metasomatically replaced by siderite and form the iron mine at Erzberg/Eisenerz.

The Devonian sequence is disconformably overlain by a limestone breccia with conodonts spanning the time from Middle Devonian to Lower Carboniferous, and the 100-150 m thick clastic Eisenerz formation probably ranging from the Visean to the lowermost Upper Carboniferous (SCHÖNLAUB & HEINISCH, 1993).

The Veitsch Nappe represents the lowermost unit of the Upper Austro-Alpine thrust sheet. The Lower Carboniferous is characterized by shales, crinoidal limestones and dolomites of Visean age (Steilbach and Triebenstein Formations). Primary to early diagenetic Magnesites are developed. The Upper Carboniferous consists of sequences with conglomerates, sandstones and slates containing plant fossils of Westfalian A-C age (Sunk Formation). Coal measures (altered to graphite) are sometimes interposed.

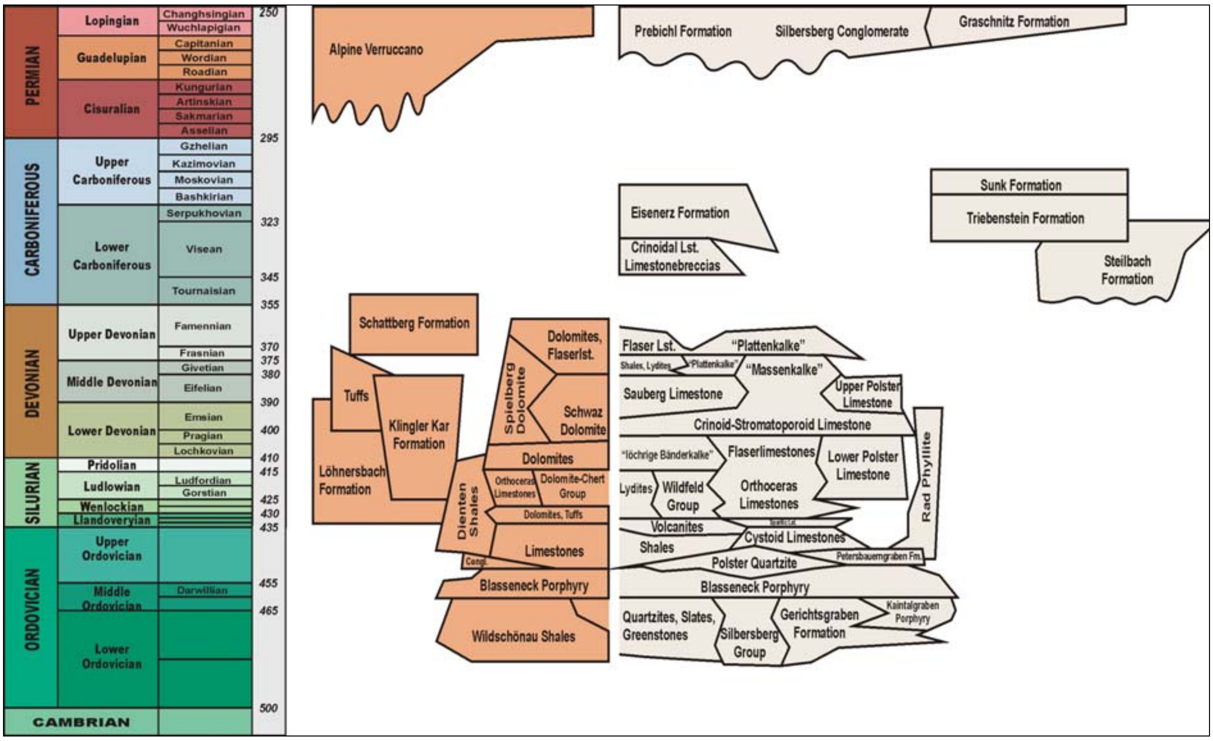

Fig. 3: Lithostratigraphic scheme of the Greywacke Zone (HUBMANN, 2003, in press).

B. The Nötsch Carboniferous

The famous fossiliferous outcrops of Carboniferous age are located in the Gail Valley between Windische Höhe and Mount Dobratsch (In German: Villacher Alpe). It culminates in the peak called Badstube (1369 m) and is crossed by the Nötsch River. The name-bearing village of Nötsch, however, is situated in the Gailtal Crystalline Complex following to the south of the Carboniferous deposits (figs. 4, 5).

Since the beginning of the 19th century the Carboniferous of Nötsch has been famous for its abundance of fossils and thus has attracted many geologists and paleontologists. The east-west directed exposures extend as a narrow fault-bounded wedge over a distance of 8 km, the maximum width of which is 2 km in the east. Further to the west the Carboniferous rocks are squeezed out between the above-mentioned rocks and are also covered by Quarternary deposits, respectively.

The tectonic significance of these Carboniferous rocks has raised many controversial statements in the past. In fact, the true relationship between the Carboniferous sediments and the surrounding units of the Gailtal Crystalline Complex and the Permo-Triassic sequence of the so-called Drauzug has long been a matter of debate and has yet not been solved satisfactorily. One of the main problems concerns the northern boundary of the Carboniferous deposits (see enclosed map). Some authors consider it as a distinct fault zone separating the Carboniferous from the Permian and Triassic, while others assume an originally transgressive relationship between Upper Carboniferous rocks and the overlying Permian clastics. A conclusive decision about one of the two options has significant implications for the tectonic framework of the greater part of the Eastern Alps.

Review of Stratigraphy

Based on a revised map and additional paleontological work carried out in the last few years (SCHÖNLAUB, 1985b; HAHN & HAHN, 1987; FLÜGEL & SCHÖNLAUB, 1990; SCHRAUT, 1990-2000; KABON, 1997; VAN AMEROM & KABON, 1999, 2000), knowledge of most rocks and fossils considerably increased. In the south-dipping sequence which was affected by several NNE-SSW trending distinct faults the oldest part occurs in the north and is named Erlachgraben Formation. Towards the south it is followed by the Badstub Breccia and the Nötsch Formation, respectively.

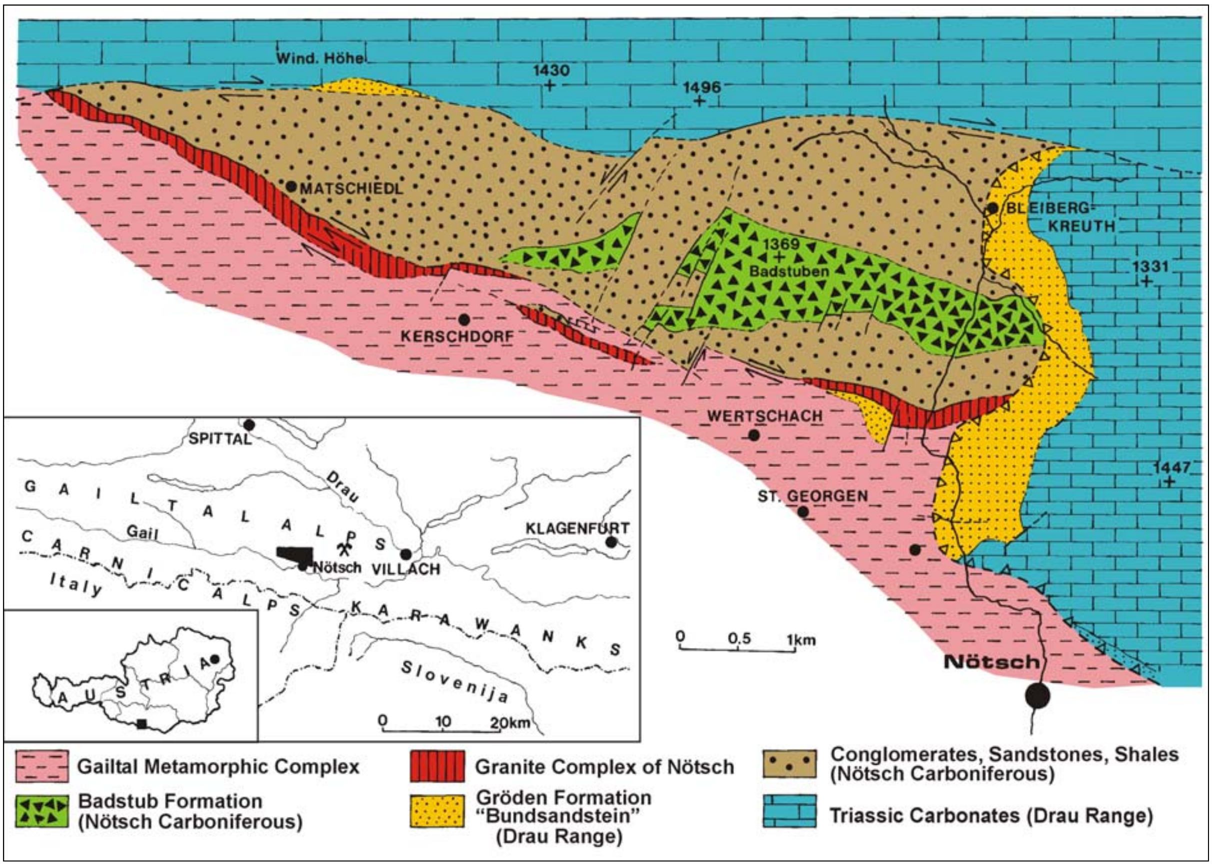

Fig. 4: Simplified geological map of the Nötsch Carboniferous.
Based on a map of SCHÖNLAUB,1985; slightly modified after KRAINER & MOGESSIE, 1991.

Erlachgraben und Nötsch Formations display similar lithologies such as greyish blackish shales, micaceous siltstones, sandstones and quartz rich conglomerates. Locally, both marine faunas and paleofloras occur very abundantly. The disorganized Badstub Breccia is composed of mainly subrounded and rounded crystalline clasts such as amphibolites, ortho- and paragneisses, schists, micaschists, quartz, quartzites, marbles and few limestone clasts embedded in a dense green matrix of tholeiitic composition. From sedimentological evidence SCHÖNLAUB (1985b) and subsequently KRAINER & MOGESSIE (1991) inferred a sedimentary origin for the breccia. Previously a volcanic source was favoured for the origin of this rock. Conodonts recovered from limestone clasts indicate a formation after the *Paragnathodus nodosus* Zone. In terms of the presently used chronostratigraphical subdivision of the Carboniferous, this time corresponds to the latest Visean or more probably, to the early Serpukhovian.

New and revised fossil data suggest an overall Serpukhovian age for the molasse-type Carboniferous sediments. In terms of the western European terrestrial succession a Lower Namurian age is most probable for the whole sequence, deducing from rich occurrences of plants in both the Erlachgraben and Nötsch Formations indicating Namurian A and less probably B, conodonts and the index foraminifera *Howchinia bradyana* (HOWCHIN) in exotic limestone clasts of the Badstub Breccia.

According to FLÜGEL & SCHÖNLAUB (1990) such clasts represent
1. bindstone with fenestral fabric;
2. grainstone/packstone with echinoderms, bryozoans, coated grains and porostromate algae;
3. packstone-grainstone with micritic clasts, peloids, echinoderms and foraminifera, and finally
4. packstone-grainstone with current-layered biogens like algae and bryozoans.

Based on these exotic limestone clasts occurring both in the Carboniferous of Nötsch and the flysch-like deposits of the Hochwipfel Formation of the adjacent Carnic Alps FLÜGEL & SCHÖNLAUB (1990) concluded that during the Visean and Serpukhovian an extensive shallow water carbonate setting of an open marine or a restricted shelf environment existed north of the present Southern Alps and adjacent to a land area. In the Eastern Alps, however, no relics of this Variscan platform sequence have been preserved. The whole carbonate sequence has been completely reworked in an accretionary wedge or underwent subduction due to their low preservation potential at active plate margins.

The dominating fossil groups of the Carboniferous of Nötsch are brachiopods, followed by bivalves, trilobites, gastropods, corals, crinoids, bryozoans, very few cephalopods and plants; microfossils include foraminifera, ostracods and few conodonts. In addition in the clastics trace fossils are fairly common. For more details concerning fossil groups, number of publications and specific reference to taxonomy the reader is referred to the comprehensive summary report of SCHRAUT (1999) in which fossil data and the history of scientific publications are compiled (table 1).

Group (no. of publications)	**Publications on taxonomy**
Fishes (6)	-
Conodonts (10)	-
Ophiocistioids (9)	SCHRAUT, 1992a, 1993a, 1995a
Echinoids (10)	-
Crinoids (30)	-
Brachiopods (52)	DE KONINCK, 1873, HERITSCH, 1918, AIGNER, 1930, 1931, AIGNER & HERITSCH, 1931
Bryozoans (26)	DE KONINCK, 1873
Phyllocarids (5)	SCHRAUT, 1996c
Ostracods (13)	SCHRAUT, 1996b
Trilobites (34)	DE KONINCK, 1873; HERITSCH, 1930; HAHN & HAHN, 1973, 1975, 1987; SCHRAUT, 1990, 1996b
Annelids (1)	SCHMIDT, 1955
Tentaculites (4)	-
Goniatites (15)	AIGNER & HERITSCH, 1930
Nautiloids (13)	DE KONINCK, 1873; AIGNER & HERITSCH, 1930
Bivalves (32)	DE KONINCK, 1873; HERITSCH, 1918
Gastropods (24)	DE KONINCK, 1873; YOCHELSON & SCHÖNLAUB, 1993
Monoplacophores (10)	DE KONINCK, 1873; HERITSCH, 1918; YOCHELSON & SCHÖNLAUB, 1993
Rugose Corals (36)	DE KONINCK, 1873; HERITSCH, 1918; KUNTSCHNIG, 1926; FLÜGEL, 1965; FLÜGEL 1972a
Foraminifera (19)	-
Plants (40)	DE KONINCK, 1873; PIA, 1924; VAN AMERON & SCHÖNLAUB, 1992; KABON, 1997; VAN AMEROM & KABON, 1999, 2000
Trace fossils (15)	-

Table 1: Main fossil groups with number of publications (in brackets) with reference to publications describing relevant species (from SCHRAUT, 1999).

With regard to corals the following taxa were recognized at different localities of the Carboniferous of Nötsch (HERITSCH, 1934; FLÜGEL, 1972):

Pseudozaphrentoides juddi juddi (THOMSON)
Pseudozaphrentoides sp.
"Palaeosmilia" isae HERITSCH
Arachnolasma cylindrica YÜ
Clisiophyllum sp.
Allorisma sp.
Hexaphyllia mirabilis (DUNCAN)

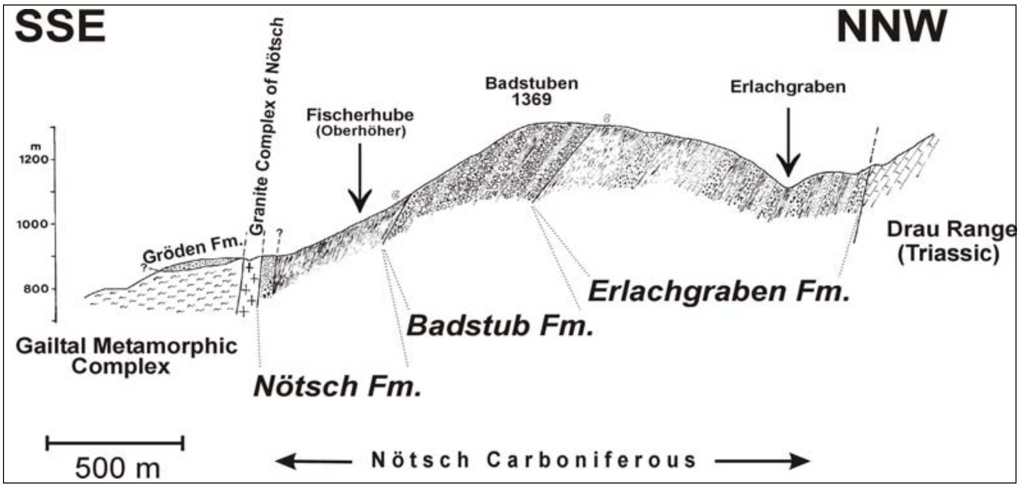

Fig. 5: Simplified stratigraphic section of the Nötsch Carboniferous.
Redrawn from KRAINER & MOGESSIE, 1991.

Discussion

Although the long lasting discussion on the age of the Carboniferous sequence has been settled in recent years three major problems have still not been solved:

1. The basement of the transgressive Carboniferous sequence has yet not been found. It may either be formed by an amphibolite-grade crystalline complex or less probably, by the Gailtal quartzphyllite. Interestingly, at several places north of the village of Nötsch there is clear evidence of a transgressive relationship between the latter and the overlying Permian-Triassic cover of the Drauzug (SCHÖNLAUB, 1985b). It may, thus, be concluded that the present outline of the Carboniferous basin was formed during the Alpine orogeny which affected and rejuvenated older faults and created new ones parallel to the Periadriatic Line. Extensive N-S shortening was mainly responsible for the closely neighbouring different tectonic units observed today; in addition, vertical movements promoted the preservation of Carboniferous deposits distributed today in an apparently distinct and almost exotic setting.
2. Also, the relationship between the Carboniferous deposits and the surrounding Permo-Triassic sequence to the north is yet unclear.
3. The formation of the Badstub Breccia still remains one of the most interesting scientific challenges. As mentioned above this disorganized breccia is composed of subrounded and rounded matrix-supported clasts of amphibolites, ortho- and paragneisses, schists, micaschists, quartz, quartzites, marbles and few limestone clasts embedded in a dense green matrix of tholeiitic composition. From sedimentological evidence, SCHÖNLAUB (1985b) and subsequently KRAINER & MOGESSIE (1991) inferred a sedimentary origin for the breccia. In previous times, however, a volcanic source was favoured for the origin of this rock. In particular, the matrix of the breccia poses problems as it consists of extremely fine-grained material the origin of which suggests either sedimentary or volcanic sources. Hence, further research is needed to decide whether or not the breccia truly represents a sedimentary rock or a volcanic contribution, e.g., a phreato-magmatic component sensu V. LORENZ (1985 ff) must be considered.

C. The Gurktal Nappe System

The Gurktal Nappe System contains Ordovician to Early Carboniferous basement sequences and Late Carboniferous to Triassic cover sequences (fig. 6). In general the nappe complex is subdivided into two major tectonic units, the lower Murau Nappe and the higher Stolzalpe Nappe. Both nappes contain Lower Paleozoic successions showing similar stratigraphic trends but striking differences in detail. The first consists of black shales and phyllites of unknown age overlain by Upper Silurian to Lower Devonian carbonates.

1. The Murau Nappe

The basal sequence of the Murau Nappe consists of phyllites with prasinites and greenshists derived from lava flows, sills and tuffs (NEUBAUER, 1979) which are overlain by a phyllite-rich unit. Carbonatic phyllites, black phyllites, and quartzites with minor greenstones and orthoquartzites build up the next higher stratigraphic unit; at the southern border of the Gurktal Nappe Complex widespread acidic volcanoclastics occur (LOESCHKE, 1989). The overlying sequence is characterized by laterally differentiated carbonates of Late Silurian to Early Devonian age.

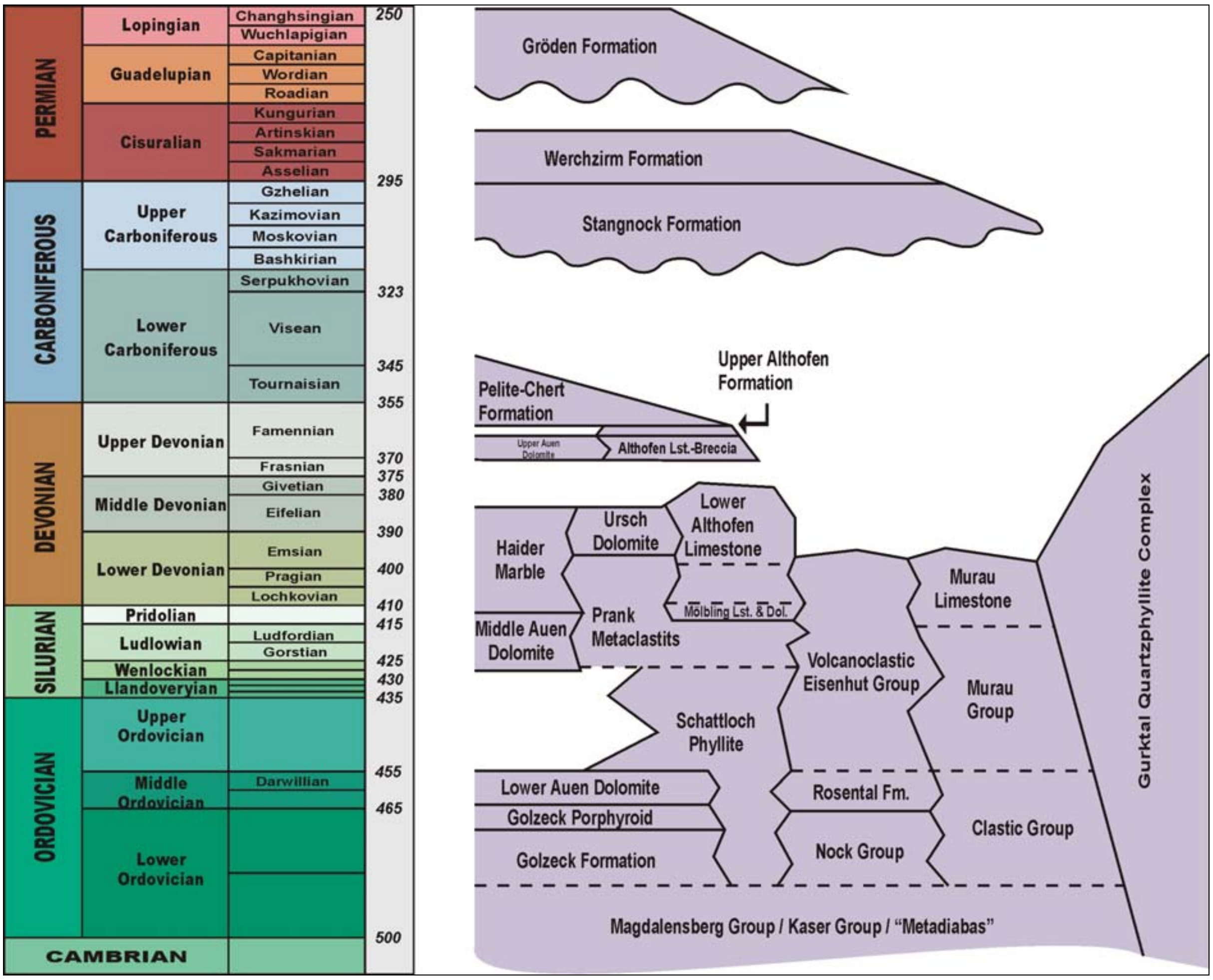

Fig. 6: Stratigraphic column of the Gurktal Nappe System of middle Carinthia and the surroundings of Murau, NW Styria. After SCHÖNLAUB, 1993; HUBMANN, 2003, in press.

2. The Stolzalpe Nappe

Basal parts of the Stolzalpe Nappe are almost similar to those of the Murau Nappe consisting of mafic volcanic sequences. These sequences are divided into the Middle to Late Ordovician Magdalensberg Group and the Nock Group which represents the Late Ordovician followed by the volcanic Early to Middle Silurian Eisenhut Group at the northern edge of the Gurktal Nappe System. These volcanic successions are overlain by sequences dominated by pelitic-psammitic rocks passing into pelagic limestones at the top.

Further reading
NEUBAUER (1980, 1987, 1992), NEUBAUER & PISTOTNIK (1984), SCHÖNLAUB (1993).

D. The Graz Paleozoic

The Graz Paleozoic comprises an outcropping area of approximately 1250 km^2 resting on metamorphic basement. In the northern and western part it overthrusted the Middle Austroalpine (Gleinalm Crystalline), in the eastern part the Lower Austroalpine unit (Raabalpen complex). In its western sector the Paleozoic succession is unconformly overlain by the Upper Cretaceous Kainach Gosau and in the south it is onlaped by Neogene sediments of the "Styrian Basin" (fig. 7).

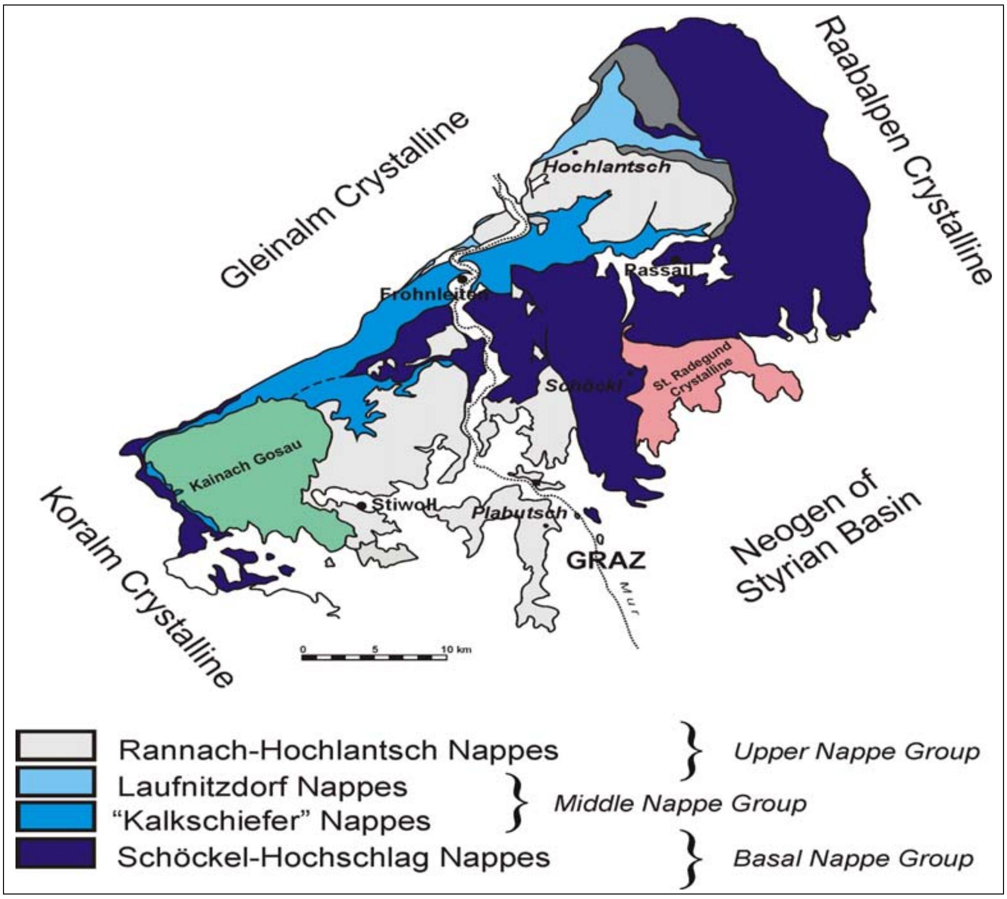

Fig. 7: The Graz Paleozoic is framed by and internally organized in systems of nappes. The Schöckel-Hochschlag-Nappe-group is generally considered to form the "Base Nappe Group"; the "Kalkschiefer Nappe", together with the Laufnitzdorf Nappe forms the "Middle Nappe Group", and the Rannach-Hochlantsch-Nappe-Group forms the "Upper Nappe Group".
The Laufnitzdorf Nappes are characterized by a lower degree of metamorphism than the "Kalkschiefer" Nappes. Therefore, they have a special position within the "Middle Nappe Group".

The Graz Paleozoic itself represents a pile of nappes. The nappes consist of different facial developments. Considering lithological similarities, the tectonic position, and metamorphic superimposition, a basal, an intermediate, and an upper nappe group are discernible:

1) The Basal Nappe System (Upper Silurian - Lower Devonian) comprises the Schöckel Nappe and Anger Crystalline Complex. Besides the common Alpine (Early to Late Cretaceous) deformation of the Graz Paleozoic in this basal nappe system minor Variscan deformation under upper greenschist facies condition (with exceptionally occurring amphipolite facies) is detected. The Schöckel Nappe is made up of pre-Devonian rocks (Passail Group, Taschen Formation) and the Devonian Peggau Group. Generally, volcanoclastics dominate the Late Silurian to Early Devonian, and carbonates the Middle Devonian time span. Part of the Peggau-Group is the Schönberg Formation with eggen-type lead/zinc-barite Sedex mineralizations.

2) The Intermediate Nappe System (Early Silurian to Upper Devonian) includes the "Laufnitzdorf Nappe" and the "Kalkschiefer Nappe" (Early to Upper Devonian). Both nappe groups occur in different structural levels. The former development contains pelagic limestones, shales and volcanoclastics, the latter limestones and siliciclastics.
3) The Upper Nappe System (Upper Silurian to Upper Carboniferous) comprises the Rannach- and Hochlantsch Nappes. Both have a similar facial development in common, especially in the Emsian to Givetian. Successions of the Rannach Nappe are composed of volcanoclastic rocks (Silurian to Early Devonian; Reinerspitz Group), siliciclastics and carbonates rich in fossils (Early to Middle Devonian; Rannach Group) of a littoral environment followed by the pelagic Forstkogel-Group (Late Givetian to Namurian B) and the shallow marine Dult-Group (Namurian B/?Westphalian).

According to a paleogeographical interpretation of the entire Paleozoic succession, the formations of the Rannach- and Hochlantsch Nappes are interpreted as having developed nearest to shore, while the "Laufnitzdorf Facies" represents the furthest from shore. Successions of the Schöckel Nappe occupy an intermediate position in this conception (HUBMANN, 1993) (fig. 8).

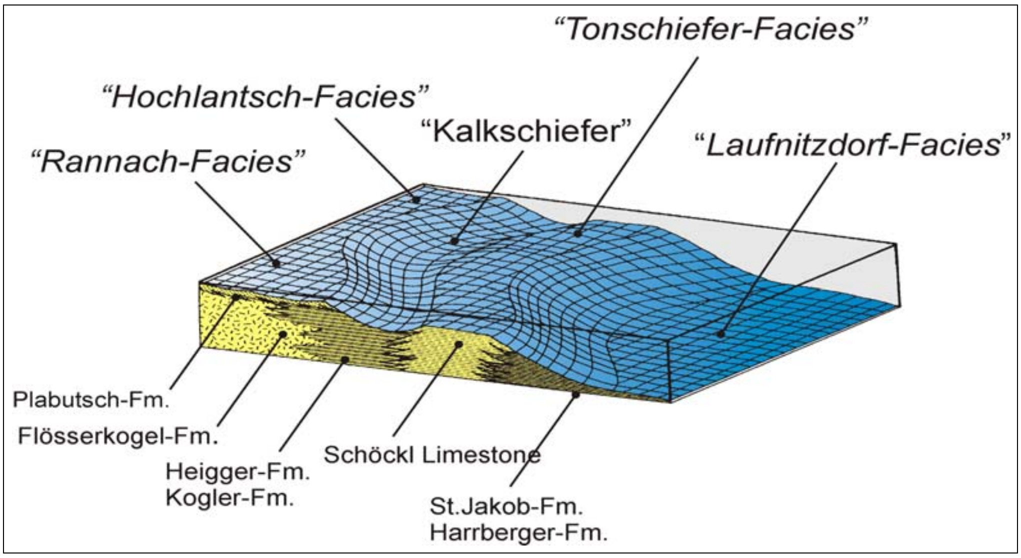

Fig. 8: Paleogeographic reconstruction of the Middle Devonian in the Graz Paleozoic. Modified from HUBMANN, 1993. "Rannach Facies" and "Hochlantsch Facies" are part of the "Upper Nappe Group". The widely distributed "Kalkschiefer" sequences, which are as yet little understood concerning their internal relationships and boundaries ("Kalkschiefer" Nappes), and the "Laufnitzdorf Facies" are subsumed as the "Middle Nappe Group" (FRITZ & NEUBAUER, 1990). In this figure the "Schöckel Facies" contains only parts of the sequences grouped by FLÜGEL (2000) as "Peggau Group".

The stratigraphic sequence indicates a sedimentation area changing from a passive continental margin with the continental breakup (alkaline volcanism) to shelf and platform geometries during the Silurian to Devonian time span (FRITZ et al., 1992). Sea-level changes and probably synsedimentary tectonics had affected both, the lithologic development (i.e. alternations of dolostones and limestones [HUBMANN, 1993]) and the formation of stratigraphic gaps and mixed conodont faunas (EBNER, 1978).
An overview of the stratigraphic development is shown in fig. 9.

Efforts to point out faunal relationships between the Paleozoic of Graz and other remnants of the Paleozoic, especially the Rhenohercynian Zone date back to the early beginning of the investigation history. Some calcareous green algae and tabulate corals (HUBMANN, 1990, 1991, 2000) show biogeographic connections with the Rhenohercynian Zone, the Moravian Karst and the Cantabrian Mountains (HUBMANN, 1991, 1995; HERRMANN & HUBMANN, 1994).

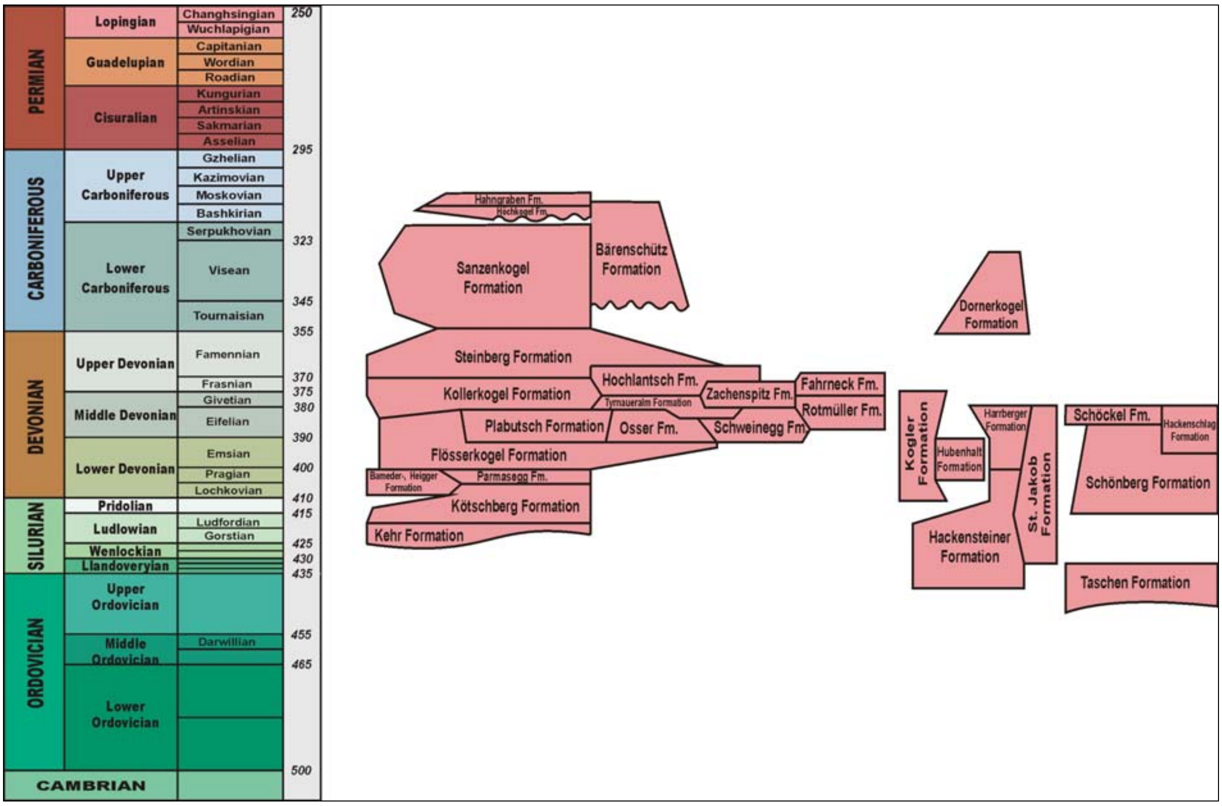

Fig. 9: Stratigraphic overview of the Graz Paleozoic (HUBMANN, 2003, in press).

Summary of the Bioarchitectural History (Upper Nappe Group)

Following the basal volcanoclastics (Kehr and Kötschberg Fms.) (fig. 10a) and marly crinoid-rich sequences (Parmasegg Fm.), the peritidal Flösserkogel Fm. starts perhaps at Lower Pragian. The formation comprises variegated dolostones, silt- to sandstones and subordinated dolomitic limestones which are interpreted as depositions of a supra- to shallow subtidal, barrier-surrounded lagoon, or tidal flats respectively (FENNINGER & HOLZER, 1978). In the vicinity of Graz the lower parts of the succession are interpreted as sand bars whereas the upper parts which are separated by volcanic tuffs contain meadows of *Amphipora ramosa desquamata*. Very rare conodont findings indicate a (lower?) Emsian age (cf. EBNER et al., 2000). Looking over the slightly hump-shaped bodies of the Amphipora Beds, the huge number of individuals and the lack of disarticulated coenostea, they are interpreted as mound structures. In contrast to other lithotypes of the Flösserkogel Formation the Amphipora mounds show a black matrix due to dispersed pyrite and high organic carbon content.

Overlying or interfingering the Flösserkogel Fm. the Plabutsch Fm. is developed. Predominance of typical "reefbuilding organisms" (FLÜGEL, 1975) is conspicuous in all sectional sites. But even so, there is no outcropping evidence of a "true reef" in the field rather coral-stromatoporoid-carpets are the dominant features. Environmental investigations indicate deposition on a differentiated and gently inclined carbonate platform (HUBMANN, 1993). Considering the rarity of *in situ* organisms, the intermittent high supply of clayey sediments (marl-limestone intercalations) and high supply of lime mud (fig. 10b), temporary influx of high amounts of continental phytoclasts and storm impacts (several tempestite sequences within the profiles) and, especially, the effects they had on the biocoenosis, the substrate produced was hardly suitable for the creation of reef structures (HUBMANN, 1995b) (figs. 10c,d).

This phase is terminated by a repetition of tidal flat deposits similar to the Flösserkogel Fm. obviously caused by an eustatic sea level fall.

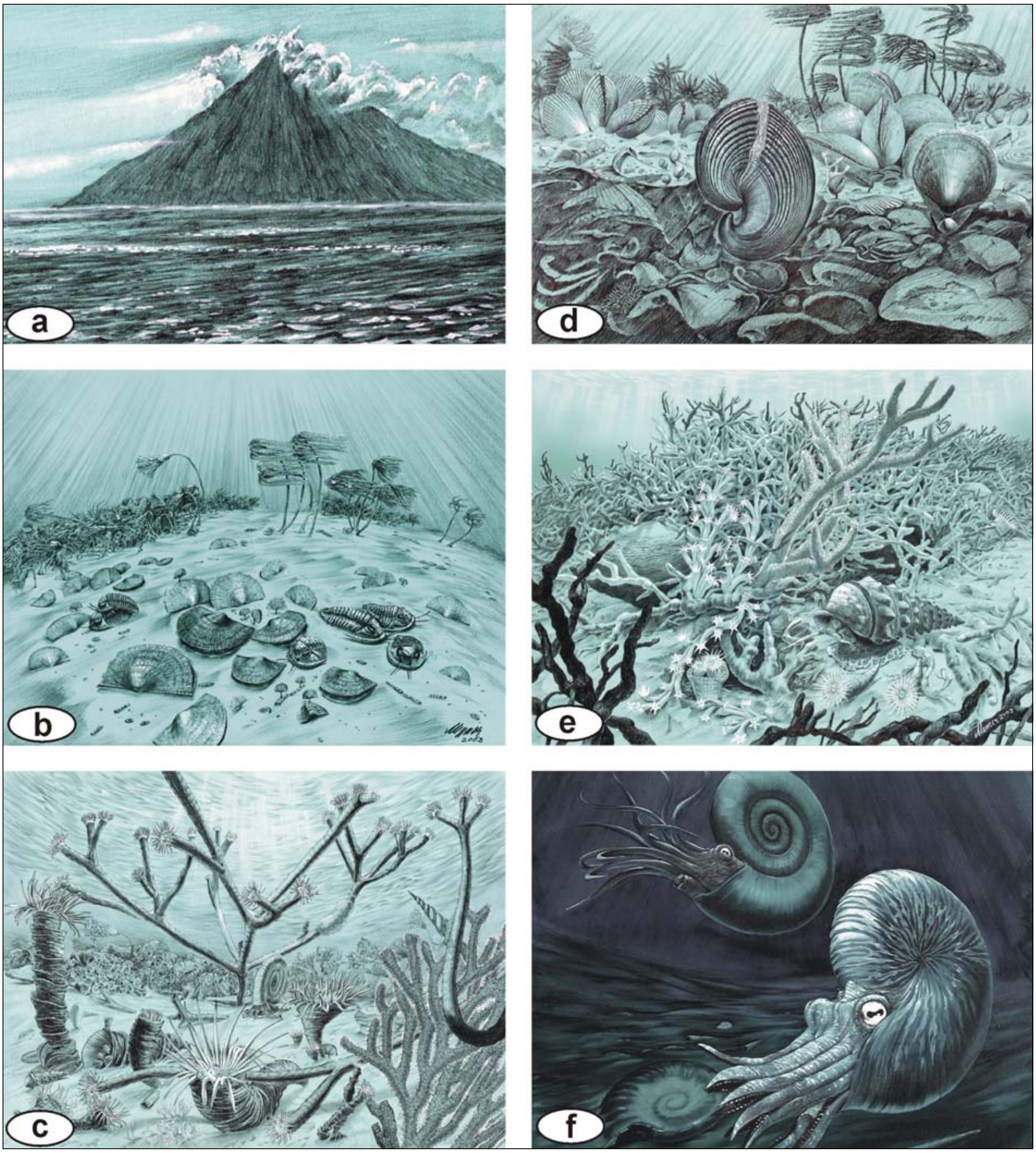

Fig. 10: Artistic reconstruction of the developmental history of the Upper Nappe Group. Biocoenoses are based on specimens collected in the Graz Paleozoic (Fritz MESSNER, unpubl.).

a: Intraplate volcanism, **b**: Gaisberg bed with chonetids, *Maladaia* sp. and crinoids, **c**: Biostrome within the Plabutsch Fm. with characteristic corals (e.g. *Thamnophyllum stachei*, *Thamnopora reticulata*, *"Cyathophyllum" graecense*), **d**: Brachiopod coquina Beds with *Zdimir* cf. *hercynicus* and *"Penta-merus" clari*. **e**: Stachyodes Beds (Kollerkogel Fm.), **f**: Clymenids and goniatids of the Steinberg Fm.

Transgression resulted in a sequence with sharp (bio)facial contrasts between patch-reefs and monotonous mudstones of Givetian age. In both upper nappes, the Rannach Nappe and the Hochlantsch Nappe contemporaneous mudstones as well as small patch reefs or biostromal deposits coexist. The reefal developments are variable due to local environmental constraints. Within the Kollerkogel Fm. small-sized *Stachyodes* thickets (fig. 10e) pass into beds with scattered chaetetids, *Favosites*, *Thamnopora*, *Thamnophyllum*, *Sociophyllum* etc. (Weiße Wand, northern slope of the Rannach). In a transitional zone between Rannach and Hochlantsch Groups a succession consisting of Amphipora Beds, microbolitic lense-shaped bodies and cnidarian patch reefs with *Stachyodes*, *Heliolites*, *Favosites*, *Alveolites*, disk-shaped stromatoporoids and solitary rugose corals are developed at Grabenwarterkogel and Höllererkogel near St. Pankrazen, 30 km NW of Graz (EBNER et al., 2000, 2001). Biostromal bodies constructed by the organisms mentioned above also occur in the Tyrnaueralm Fm. in the Hochlantsch Nappe (KRAMMER, 2001) at the Zechnerhube locality. There, similar to the situation near St. Pankrazen alveolitid corals supply great amounts of the "binder guild".

Restricted only to the Hochlantsch area some 30 km north of Graz the final "bioconstructions" of the Graz Paleozoic are developed within the Zachenspitz Fm. This Upper Givetian formation contains within its succession several basal Amphipora Beds grading into biohermal *Argutastrea*-Alveolitid-Stromatoporoid baffle- to boundstones exposed at the northern slopes of the Hochlantsch mountain. (Micro)Facial investigations indicate a shallow offshore depositional environment with a pelagic fauna dominated by tentaculites in the inter-bioherm facies (GOLLNER & ZIER, 1985). During the uppermost Givetian to lower Frasnian the sedimentation of shallow platform carbonates were replaced by micritic cephalopod limestones (Steinberg Fm., fig. 10f).

Further reading
FLÜGEL (1975), FLÜGEL & NEUBAUER (1984), HUBMANN & HASENHÜTTL (1995), KREUTZER et al. (1997), EBNER et al. (2000), FLÜGEL (2000), EBNER et al. (2001).

E. The Karawanken Alps

The Periadriatic Line divides the Karawanken Alps into a northern part (Northern Karawanken) which belongs to the Eastern Alps and a southern part (Southern Karawanken) belonging to the Southern Alps.

In the Eastern Karawanken Alps, north of the Periadriatic Line, rocks of Paleozoic age have long been known. They belong to the so-called "Diabaszug von Eisenkappel". This narrow belt extends in a W-E direction from Zell Pfarre via Schaidasattel to east of Eisenkappel and continues further east to Slovenia. In Austria this zone has a length of more than 25 km and a maximum width of 3,5 km. The 650 m thick Paleozoic sequence comprises up to 350 m of volcanic and volcaniclastic rocks and sediments. According to LOESCHKE (1970-1977, 1983) and LOESCHKE et al. (1996) the first group is dominated by basic tuffs and tuffitic rocks, massive pillow lavas and basic sills of hawaiitic composition with ultrabasic layers. Sills and pillow lavas represent spilites which differentiated from alkali olivine basalts, the original geotectonic setting of which is yet not known. Subsequent low-temperature metamorphism associated with devitrification and metasomatic replacement processes caused the spilitic mineral composition in these rocks. The sedimentary rocks are monotonous grey shales and slates with intercalations of conglomeratic greywackes, quartzitic and graphitic sandstones and thin limestone beds. The definite age of this succession is yet not exactly known although some poorly preserved single cone conodonts recovered from the limestone intercalations are rather in favour of an Ordovician than of any younger age.

In the Southern Karawanken Alps Paleozoic rocks are exposed in the Seeberg region (fig. 11). Here the sequence starts with acidic to intermediate pyroclastics and shallow marine "flaser" limestones of Upper Caradocian age. The Lower Silurian strata are dominated by siliciclastics passing into Middle to Upper Silurian carbonatic sequences. During the Devonian a carbonate platform is developed with reefal structures resembling present-days atolls (RANTITSCH, 1990). Depending on adequate subsidence the location of the reef core shifted spatially and temporarily during the Devonian. Differing from the Carnic Alps with its 150 m thick reefs of Givetian age, in the Karawanken Alps there are no good records from the Middle Devonian. In both areas, however, the reef development ended in the Frasnian when the former shallow sea subsided being followed by a drowning and erosion of the reefs. Similar to the Carnic Alps in the Karawanken Alps these shallow water deposits were also replaced by uniform pelagic goniatite and clymeniid limestones.

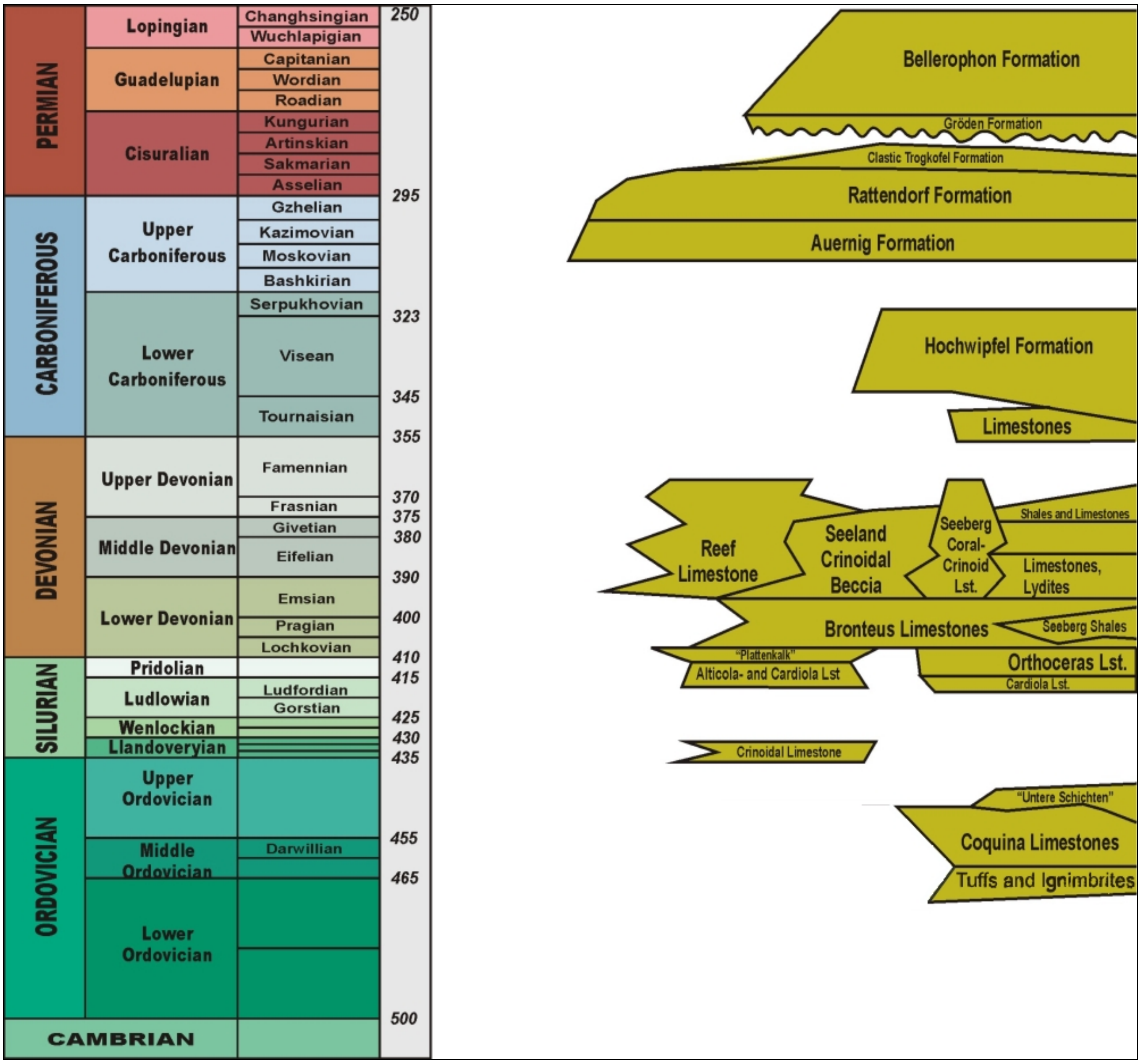

Fig. 11: Biostratigraphic scheme of the Paleozoic sequence of the Karawanken Alps.
After SCHÖNLAUB, 1985, modified.

Upper Carboniferous and Permian molasse-type sediments also occur in the Seeberg area of the Eastern Karawanken Alps (TESSENSOHN, 1983; BAUER, 1983). Although strongly affected by faults the general lithology and the fossil content resemble that of the Auernig Group of the Carnic Alps being dominated by interbedded fusulinid and algal bearing limestones, arenaceous shales, sandstones and massive beds of quartz-rich deltaic conglomerates. Equivalents of the Permian are represented by the Trogkofel Limestone, its coeval detritic Trogkofel Formation and the Gröden Formation. The Bellerophon Formation is only locally preserved.

Further reading
BAUER et al. (1983), TESSENSOHN (1974, 1983).

F. The Carnic Alps
(Fig. 12)

Ordovician

In the Austrian part of the Southern Alps the Ordovician succession comprises weakly metamorphosed fine and coarse clastic rocks named the Val Visdende Group. This more than 1000 m thick sequence is well exposed in the westernmost part of the Carnic Alps on both sides of the Austrian-Italian border on the topographic sheets Obertilliach and Sillian. The lithology ranges from shales and slates to laminated siltstones, sandstones, arkoses, quartzites and greywackes. They are overlain by more than 300 m thick acidic volcanites and volcanoclastic rocks named the "Comelico Porphyroid" and "Fleons Formation" respectively, and their lateral equivalents comprising the Himmelberg Sandstone and the Uggwa Shale. Locally, the latter contain rich fossils such as bryozoans, trilobites, hyoliths, gastropods and cystoids indicating a Caradocian age (HAVLICEK et al., 1987; SCHÖNLAUB, 2000). According to DALLMEYER & NEUBAUER (1994) detrital muscovites from the sandstones are characterized by apparent ages ($^{40}Ar/^{39}Ar$) of circa 600 to 620 Ma and may thus be derived from a source area affected by late Precambrian (Cadomian) metamorphism.

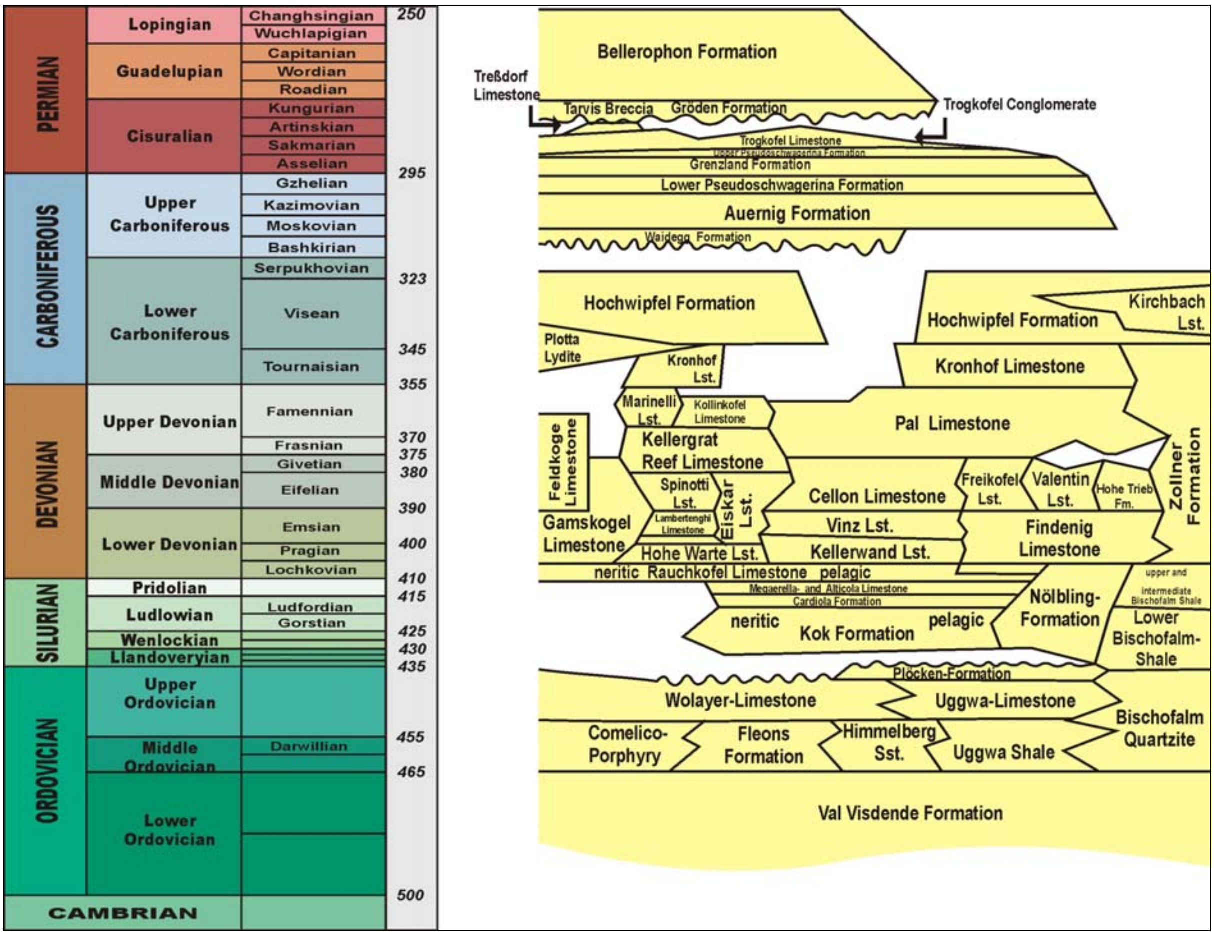

Fig. 12: Biostratigraphic scheme of the Paleozoic sequence of the Carnic Alps.
After SCHÖNLAUB, 1985, modified.

This basal clastic sequence is capped by an up to 20 m thick fossiliferous limestone horizon of early Ashgillian age. It displays two lithologies, namely the massive "Wolayer Limestone" composed of parautochthonous bioclasts (cystoids and bryozoans) which laterally grades into the bedded wackestones of the "Uggwa Limestone" representing a more basinal setting with reduced thicknesses.

In the Carnic Alps the global glacially induced regression during the Late Ashgillian Hirnantian Stage is documented by marly intercalations and arenaceous bioclastic limestones of the Plöcken Formation which presumably corresponds to the graptolite Zone of *Gl. persculptus* (SCHÖNLAUB, 1996). If so it may have lasted during the early and middle Hirnantian Stage for not more than 0.5 to 1 million years. It resulted in channeling, erosion and local non-deposition. In fact, the succeeding basal Silurian strata generally disconformably rest upon the late Ordovician sequence.

Initiation of the fore-mentioned rifting and subsequent movements from higher to lower latitudes may be marked by basic volcanism occurring at various places in the Eastern Alps in pre-Llandeillian strata (for references see SCHÖNLAUB [1992]). In the Southern Alps such rocks have not yet been recognized. The Upper Ordovician faunal affinities, e.g. brachiopods, nautiloids, cystoids, ostracods, conodonts and vertebrate remains indicate links with Bohemia, Thuringia, Baltoscandia, Sardinia and the British Isles (SCHÖNLAUB, 1992; FERRETTI & BARNES, 1998; FERRETTI, 1997; BAGNOLI et al., 1998; BOGOLEPOVA & SCHÖNLAUB, 1998). Moreover, the appearance of carbonate rocks in the Upper Ordovician suggests a position within the broader carbonate belt for this time. However, also a temporary cold-water influx from northern Gondwana may have existed as can be concluded by certain elements of the Hirnantia fauna. Based on the available evidence from the Ordovician of the Southern Alps SCHÖNLAUB (1992) inferred a paleolatitudinal position at roughly 50°S.

Silurian

The Silurian of the Carnic Alps is subdivided into four lithological facies representing different depths of deposition and hydraulic conditions suggestive of a steadily subsiding basin and an overall transgressional regime from the Llandovery to Ludlow (fig. 13). Uniform limestone sedimentation during the Pridoli suggests that more stable conditions were developed at this time (SCHÖNLAUB, 1997). Silurian deposits range from shallow water bioclastic limestones to nautiloid-bearing limestones, interbedded shales and limestones to black graptolite-bearing shales and cherts with overall thicknesses not exceeding 60 m. The available data for the Carnic and Karawanken Alps suggest a complete but considerably condensed succession in the carbonate-dominated facies and a continuous record in the graptolite-bearing sequences.

In the Carnic Alps the Silurian transgression started at the very base of the Llandovery, i.e. in the graptolite zone of *Akidograptus acuminatus*. Due to the disconformity separating the Ordovician and the Silurian at many places a varying pile of sediments is locally missing, which corresponds to several conodont zones of Llandoverian to Ludlovian age. Even uppermost Pridolian strata may disconformably rest upon Upper Ordovician limestones.

The Rauchkofel Boden section is one of the best known and most fossiliferous Upper Silurian sections of the Carnic Alps corresponding to the "Wolayer Facies", an apparently shallower marine environment. The contact with the underlying massive cystoid Wolayer Limestone (Upper Ordovician) and the Mid Wenlock bioclastic limestones with a rich fauna of nautiloids, bivalves, brachiopods and trilobites representing the neritic Kok Formation is marked by an iron-oolitic concentration. Development of microstromatolites is also evident in the lower levels of the sequence. In the Wenlock/Ludlow transition thinly developed cyclic micritic limestone beds of bioclastic accumulations are separated by stylolites and sometimes iron-oolitic concentrations which may mark the end of depositional regimes. Concentrations of apparently juvenile and equidimensional articulate brachiopods, nautiloids and gastropods alternate with the dominantly nautiloid beds (the classic Orthoceras Limestone) in the lower Ludlow demonstrating the changing energy and oxygen levels of the formation while the preservation and orientation of the fauna indicate many accumulated levels with intermittent changes in sea level particularly towards the top of the sequence. The overlying Cardiola Fm., Ludlow in age, comparable with the well-known cephalopod limestone deposited in Bohemia and along the North Gondwana margin is represented by a thinly developed dark limestone showing lateral variation in its outcrop. Nautiloids and bivalves are the dominant fauna in this micritic limestone which represents more current-ventilated conditions. The Alticola Limestone, Pridoli in age, is a fine grey micritic limestone with abundant micritised bioclasts, frequent stylolites and an abundant nautiloid fauna throughout the formation. The associated shallow water fauna is similiar to the Kok Formation except for the presence of rugose corals. A Scyphocrinites Bed bearing complete specimens caps the formation and marks the Silurian/Devonian boundary and the shallowest level of the sequence (FERRETTI et al., 1999).

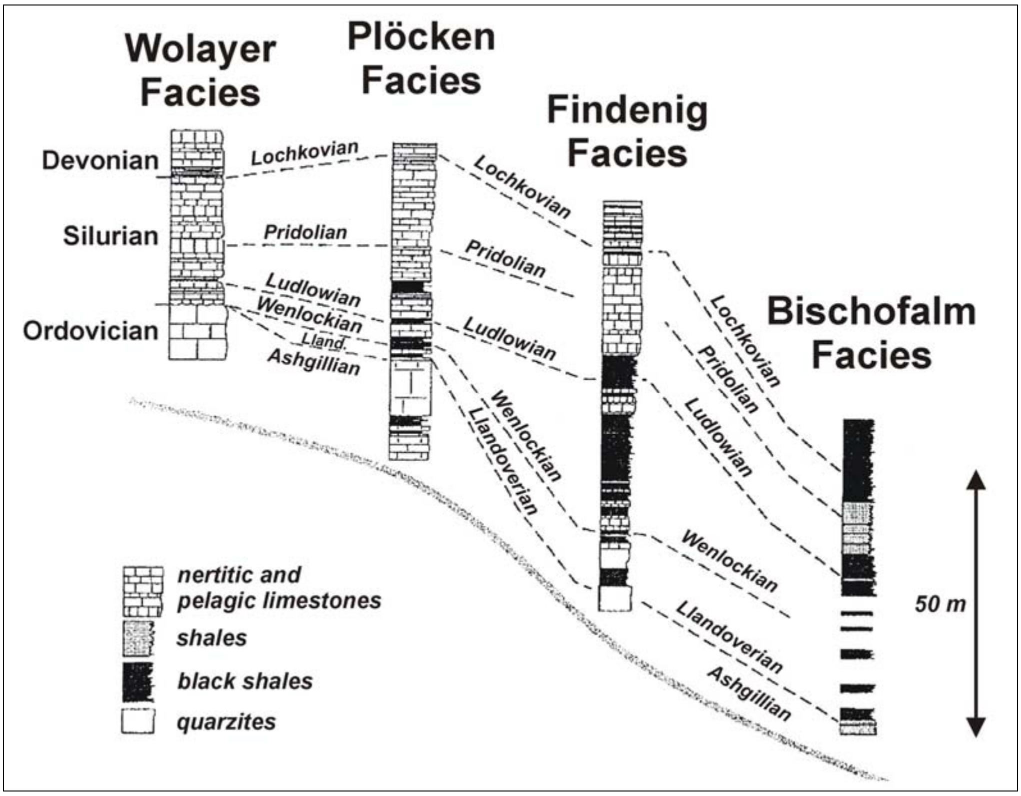

Fig. 13: Lithology of Silurian sediments of the four different lithofacies of the Carnic Alps.

Brickstone: carbonates; black: C_{org} rich graptolite-bearing shales and cherts and C_{org} rich carbon-ates of the Wolayer Facies; light grey: C_{org} poor shales.

Columns from left to right show the sections Rauchkofel Boden, Cellon, Oberbuchach 1-2 and Nölblinggraben-Graptolithengraben. In the latter composite section Lower Silurian sediments are not continuously exposed. After WENZEL, 1997.

The Cellon section represents the stratotype for the Silurian of the Eastern and Southern Alps (WALLISER, 1964) and the "Plöcken Facies" is developed here as a shallow to moderately deep marine carbonate series (FLÜGEL et al., 1977). The condensed nature of the sequence of the Cellon section is clearly demonstrated when correlated with the thicknesses of the same intervals of the more basinal facies of mainly graptolitic shales of the Oberbuchach section and the even more condensed Rauch-kofel Boden section. Underlain by the Uggwa Limestone and the clastic Plöcken Fm. the carbonate sequence of the Plöcken Facies was deposited in a relatively shallow environment, periodically effected by storm currents, with intervals of reduced depositional rates and non-sedimentation in an overall transgressive sequence. The pelagic Kok Formation consists of a transgressive carbonate series with alternating black shales and dark grey to slightly red micritic lenticular limestones occurring at the base of the formation in the upper Llandovery and brown-red ferruginous limestones with abundant nautiloids and frequent stylolites in the Wenlock - lower Ludlow. Two deepening events are documented within the formation: at the transition between the Llandovery and Wenlock and between the Wenlock and Ludlow (SCHÖNLAUB, 1997).

The alternating rapid deposition of black shales and laminated micrites with more time-rich light grey nodular micrites with an abundant nautiloid fauna of the Cardiola Beds (Ludlow) indicates a slightly deeper offshore environment with probable contemporary non-deposition taking place.

A more stable pelagic environment is developed in the Alticola and Megaerella Limestones from the upper Ludlow continuing into the Pridoli (SCHÖNLAUB, 1997) represented by a transgressive carbon-ate series of grey to dark pink micritic limestones with a variety of bed thickness and frequent stylo-lites The beds decrease in thickness in the Pridoli and alternate with interbedded laminated micrites with a dominant nautiloid and brachiopod fauna. Several deepening events marked by the develop-ment of black shales have been documented within the uppermost levels of the Pridoli. An offshore setting frequently ventilated by currents of varying energy is envisaged for the upper Ludlow and Pridoli sequences of the Alticola Limestone. The Megaerella Limestone (Pridoli in age) comprises the upper Pridoli and Silurian/Devonian boundary transgressive sequences of biodetritus-rich carbonates, lenticular micrites and black shales. The boundary between the Silurian and Devonian is drawn based on conodonts with the first occurrence of *Icriodus woschmidti* (WALLISER, 1964). However, the first evidence from graptolites of Lochkovian age is found in bed 50 with the occurrence of *M. uniformis* (JÄGER, 1975). PRIEWALDER (1997, 2000) indicates a rich chitinozoan fauna from the Pridoli-Lochkovian interval, therefore the depositional environment was of a low hydrodynamic regime, favorable for their preservation.

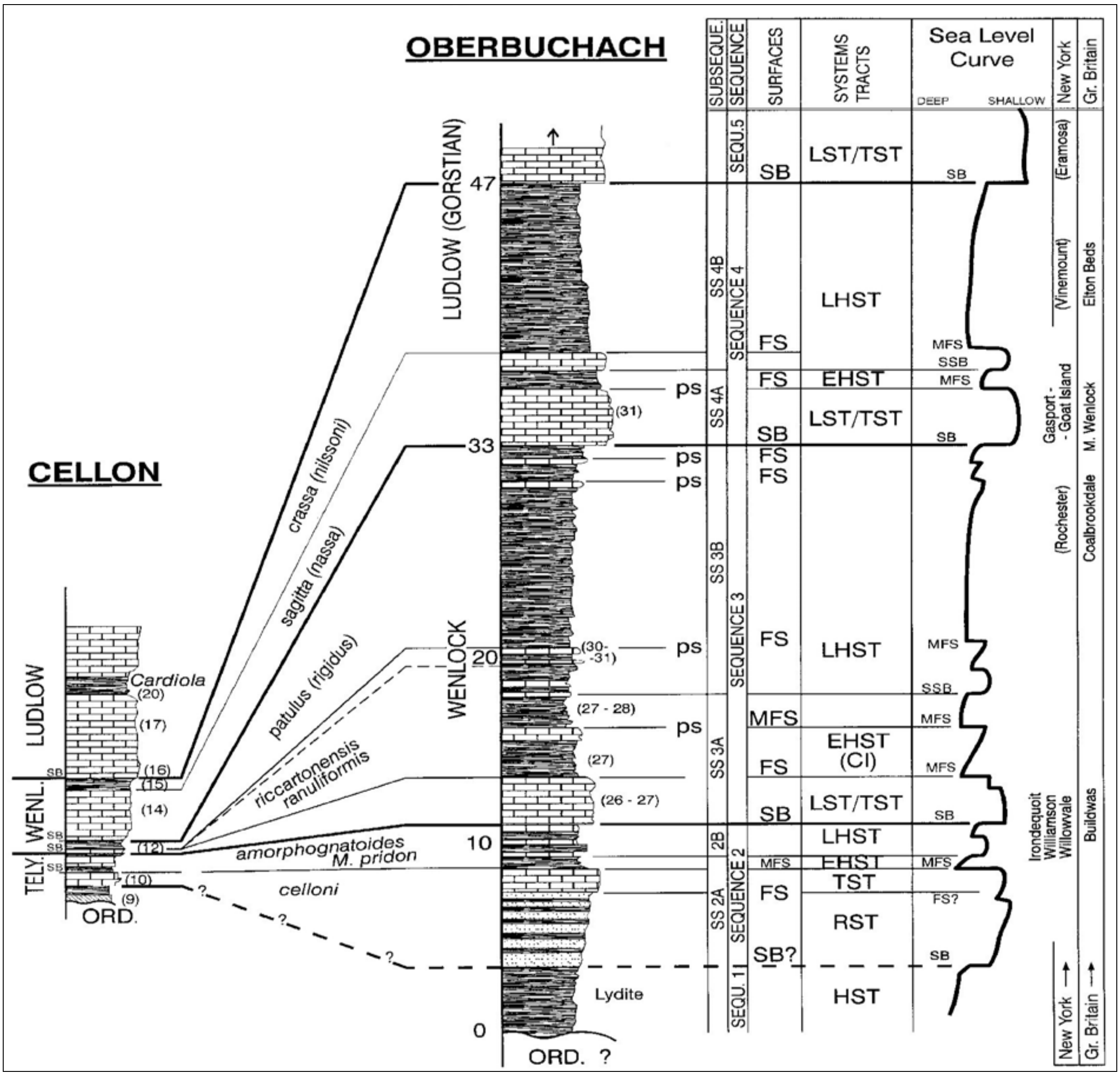

Fig. 14: Correlation and sequence interpretation Llandovery - Lower Ludlow, Carnic Alps. (BRETT & SCHÖNLAUB, 1998).

There appears to be a distinct gradation of beds upwards towards the Silurian/Devonian boundary indicating that the hydrodynamic regime is constantly changing with the shallowest point being reached at the base of the Rauchkofel Limestone (Lochkovian) with the occurrence of a bryozoan fauna (HISTON et al., 1999). A recent taphonomic study of the Silurian of the Cellon section has highlighted in more detail the faunal and environmental changes during this time interval (HISTON & SCHÖNLAUB, 1999).

The large oxygen isotope ratio excursion shown by WENZEL (1997) at the boundary may be supported by the more ventilated setting implied by the bryozoan fauna.

The intermediate "Findenig Facies" occurs between the shallow water condensed sequences outlined above and the starving basinal facies. It consists of the interbedded black graptolitic shales, marls and blackish carbonates which is locally underlain by a quartzose sandstone.
The stagnant water graptolitic "Bischofalm Facies" is represented by black siliceous shales, lydites and clayish alum shales.

The evidence from the Silurian indicates faunal affinities, e.g. conodonts, trilobites, brachiopods, molluscs, chitinozoa and acritarchs with Baltica and Avalonia as opposed to loose relationships with Africa and southern Europe. In addition, first occurrences of rugose and tabulate corals, ooids and stromatolites indicate a moderate climate. An overall island setting may be inferred by a generally condensed and reduced sedimentary pattern without significant clastic imput. These data suggest an ongoing drift towards lower latitudes and consequently a paleolatitudinal position between 30 and 40°S. In the central Alps rifting-related basic volcanism underpins these inferred plate movements (SCHÖNLAUB & HISTON, 1999).

A sea-level curve for the Llandovery/lower Ludlow interval of the Cellon (Plöcken Facies) and Oberbuchach (Findenig Facies) sections of the Carnic Alps has been elaborated by BRETT & SCHÖNLAUB (1998) based on a sequence stratigraphy study of the sections (fig. 14). The variations in sea-level compare quite well with those inferred by JOHNSON (1996) and LOYDELL (1998) for the global sea-level changes during the Lower Silurian. For correlation and sequence interpretation see fig. 14.

Sequence Stratigraphy, Platform Evolution and Paleoecology of Devonian Carbonates in the Central Carnic Alps

The Mid Paleozoic limestones exposed in the Central Carnic Alps preserve the whole range of carbonates encountered on a shelf to basin transect, a scenario rarely encountered in the geologic record. This provided an opportunity to investigate the consequences of sea level changes, shelf sedimentation and margin architecture on a Devonian carbonate system covering a time period close to 50 million years.

Devonian carbonates were investigated in an area extending from Giramondo Paß in the west to Findenigkofel in the east and from Pizzo di Timau in the south to the Gamskofel-Mooskofel Massif in the north. This area encompasses the majority of well-preserved Devonian carbonates in the Carnic Alps. A NNW-SSE oriented differentiation of facies can be recognized with backreef sediments in the south, separated by reef complexes from slope (or ramp) and basin sediments in the north. Tectonic shortening brought the different facies into close proximity and the various depositional environments of the Devonian carbonates are now located in different structural units.

In the Central Carnic Alps numerous sections were measured through reef- and backreef facies (Kellerwand-Hohe Warte Nappe), forereef-, ramp- and/or slope facies (Cellon Nappe) and through pelagic and hemipelagic facies with common gravity flow deposits and interbedded fine-grained siliciclastic units (Findenig Nappe). A pelagic facies with few or no gravity flow deposits occurs in the vicinity of Mount Rauchkofel, and at Zollner Lake cherts and siliceous shales of deep water aspect are exposed (Rauchkofel and Bischofalm Imbricate Nappe Complexes respectively).

The successions reflect the development of a carbonate ramp which was slowly drifting into low-latitudinal warm waters to a tropical carbonate shelf platform with shelfbreak and segmented slope. Masswasting is extensive on the slope and characterizes slope sedimentation. Upper Devonian strata are characterized by overall deepening of the water and backstepping of the shelf edge assembly. The Famennian carbonates of deepwater aspect dominate in all depositional environments and platform drowning is implied.

3. Depositional Environments of the Devonian Carbonates in the Central Carnic Alps

Fig. 15: View from Valentin Törl to the mountainous area in the east showing the proximity of the different depositional environments preserved in the Feldkogel, Cellon, and Rauchkofel Nappes.

Introduction

The Carnic Alps are an east-west striking mountain chain at the border between Southern Austria (Carinthia) and Northern Italy. They represent the Paleozoic basement of the Southern Alps with sequences ranging from Caradoc to Late Carboniferous. The late Paleozoic series were first affected by late Variscan tectonism and later by intense Alpine deformation, which resulted in formation of several thrust sheets, imbricate nappe systems, and dislocations in both, Variscan and post-Variscan Series (SCHÖNLAUB, 1979). Paleogeographically, sediments of the Carnic Alps were deposited in the vicinity of the northern margin of the ancient Gondwana continent. A position removed from a continental or volcanic source area enabled the formation of an almost pure carbonate system.

The area extending from the Giramondo Paß in the west to the Findenigkofel in the east and from the Gamsspitz in the south to the Gamskofel-Mooskofel Massif in the north (fig. 16) encompasses the majority of well-preserved Devonian carbonates in the Carnic Alps. KREUTZER (1990, 1992) recognized a NNW-SSE oriented differentiation of facies, and proposed a paleogeographic model with backreef sediments to the south, separated by reef complexes from slope (or ramp) and basin sediments to the north. Tectonic shortening brought the different facies into close proximity and the various depositional environments of the Devonian carbonates are now located in different structural units (see fig. 15). In the Central Carnic Alps reef- and backreef facies of a carbonate platform complex are confined to the Kellerwand Nappe encompassing Gamskofel Massif, Biegengebirge and Kellerwand-Hohe Warte Complex.

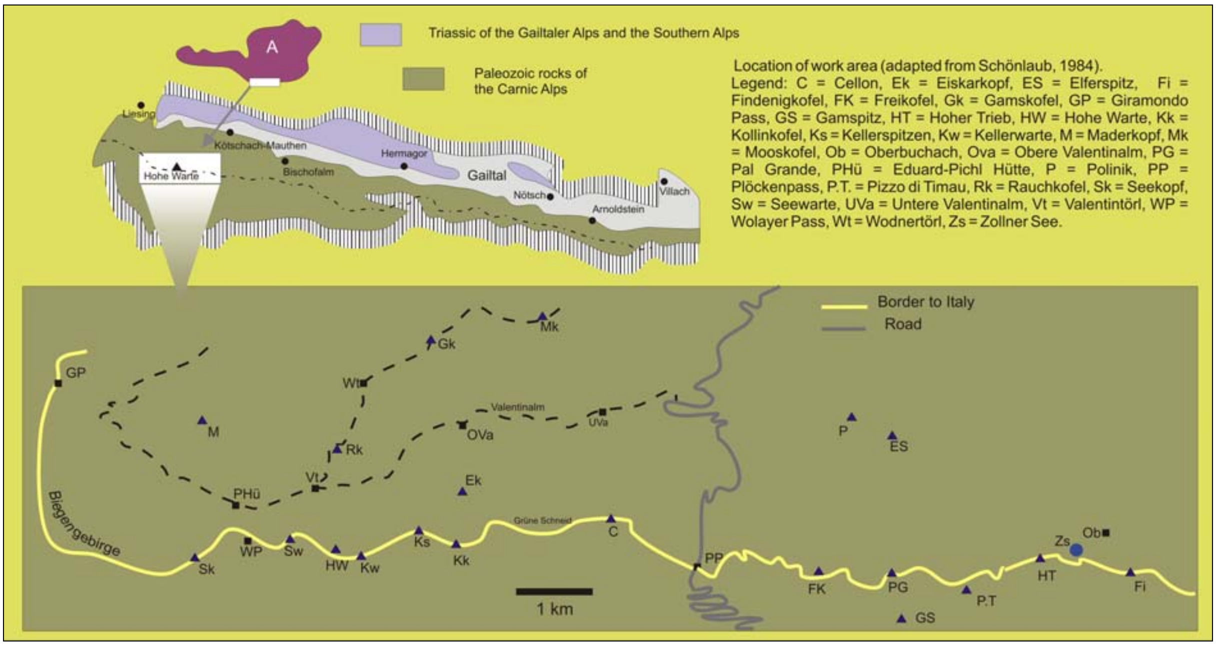

Fig. 16: Location of sections and localities discussed in the text.

The tectonically lower Cellon Nappe contains Silurian to Lower Carboniferous carbonates of fore-reef-, ramp- and/or slope facies. To the northeast along the Cellon Nappe pelagic and hemipelagic limestones occur with common gravity flow deposits and interbedded fine-grained siliciclastic units. A pelagic facies with few or no gravity flow deposits occurs in the vicinity of Mount Rauchkofel and is referred to as Rauchkofel Facies. In the region of the Zollner Lake cherts and siliceous shales occur with graded beds of the Bischofalm Facies. These are interpreted as basin deposits. Sediments of Rauchkofel and Bischofalm Facies display complex imbricated structures and are referred to as Rauchkofel- and Bischofalm Imbricate Nappe Complexes respectively. According to KREUTZER (1992) the intertidal and pelagic zones were spaced about 8-9 km apart with the intervening reef belt about halfway between both zones. Consequently at a few degrees inclination of the slope, the basin floor would have been at about 300 m, at 15°°inclination at over 1000 m water depth (fig. 17).

Although most strata belong to various imbricate thrust slices and nappes that characterize the tectonic style of the Carnic Alps, the internal structure of the allochthonous units is coherent and sections can be correlated based on the biostratigraphy established particularly for slope and pelagic deposits (e.g. BANDEL, 1972, 1974; GÖDDERTZ, 1982; PÖLSLER, 1969; SCHÖNLAUB, 1982). The correlation with the shelf sequences poses more of a problem. The stage boundaries are only loosely defined due to sparse conodont and other biostratigraphically useful faunas and the difficult access to some sections (KREUTZER, 1990, 1992; VAI, 1973).

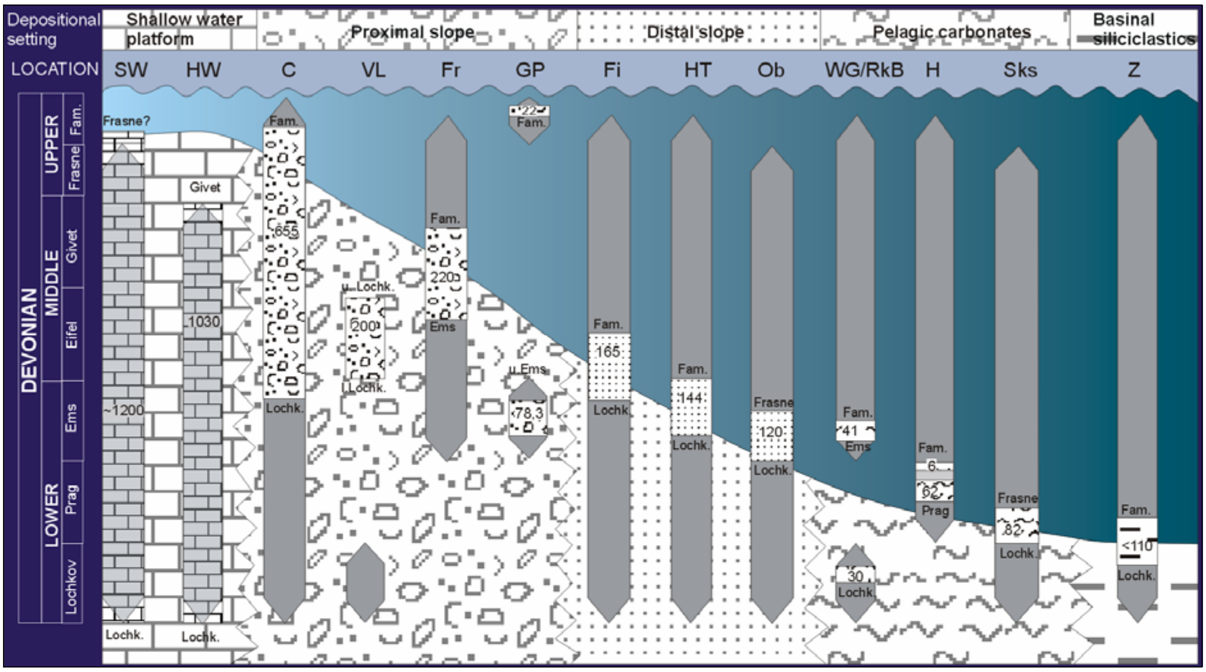

Fig. 17: Thicknesses and ranges of measured sections through the various sedimentary realms.
SW = Seewarte, HW = Hohe Warte, C = Cellon, VL = Valentintal, Fr = Freikofel, GP = Großer Pal, Fi = Findenigkofel, HT = Hoher Trieb, Ob = Oberbuchach, WG/RkB = Wolayer Glacier/ Rauchkofel Boden, H = Hütte, Sks = Seekopfsockel, Z = Zollnersee. For locations see fig. 16.

Conodont Biostratigraphy

Conodont biostratigraphy of sections of the Rauchkofel Facies are well documented from Oberbuchach II, Wolayer Glacier, base of Seekopfsockel and Rauchkofelboden (fig. 18; SCHÖNLAUB, 1981; GÖDDERTZ, 1982; SCHÖNLAUB, 1982). The sections at Findenigkofel were studied by PÖLSLER (1969) and numerous samples collected by BANDEL from various sections were dated by SCHÖNLAUB (in BANDEL, 1972). The latter are kept at the Geological Survey in Vienna and faunas need to be reviewed because much progress has been made in conodont taxonomy and stratigraphy. This is particularly true for the samples from sections of the Cellon Nappe which are not well constrained by conodonts.

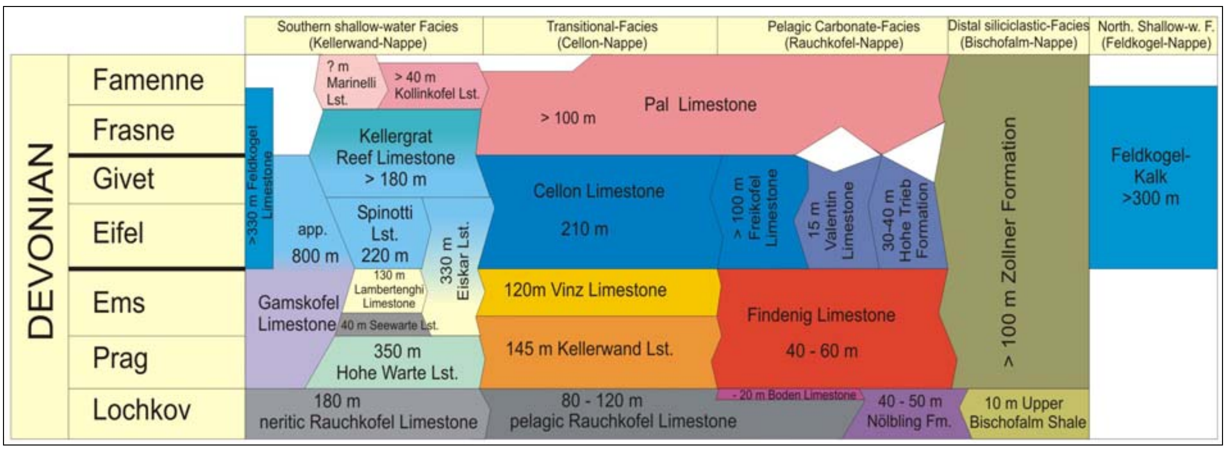

Fig. 18: Stratigraphy of the different Devonian lithofacies on a proximal (left) to distal (right) transect. To the right the northern shallow-water facies of the Feldkogel Nappe is indicated. Adapted from SCHÖNLAUB, 1992.

Southern Shallow Water Facies (Kellerwand Nappe)

The Devonian carbonates of shallow water aspect are preserved in the Kellerwand Nappe Complex and are exposed in the Gamskofel-Mooskofel Massif, Biegengebirge (with Giramondo Paß), and Seewarte-Hohe Warte Massif. Best access and preservation are found at Seewarte, Hohe Warte, and at the base of the Seekopf (BANDEL, 1969, 1972; POHLER, 1982; KREUTZER, 1990, 1992). These sections also show the highest degree of facies differentiation in the region. A section through the southern shallow water facies is accessible at Mount Seewarte (fig. 19).

Lochkovian limestones of the Rauchkofel Limestone are 152 m thick here and can be subdivided into two distinctive units: the lower 96 m consist of dark, thin-bedded finegrained limestones and shales interbedded with three dolomitized conglomerate and mega-conglomerate horizons. The mega-conglomerates contain boulders measuring up to 10 m in diameter. The upper 56 m of the Lochkovian limestone consist of crinoidal limestone with dolomitized groundmass. Graded beds with aligned crinoid debris are interbedded with disorganized massive crinoidal limestones.

The Pragian is represented by 350 m of Hohe Warte Limestone with coarse crinoidal limestone and well developed patch reefs particulary in the upper part (VAI, 1967; JHAVERI, 1967; BANDEL, 1969). It was measured and sampled in detail by BANDEL (1969) at the base of Mount Seewarte. Both Rauchkofel and Hohe Warte Limestone grade laterally into periplatform deposits composed of interbedded pelagic and detrital carbonates (KREUTZER, 1990). This facies is characteristic of the Lower to Middle Devonian sections in the Cellon Nappe and their presence in the shallow water Kellerwand Nappe shows that both sedimentary realms were closely related.

The succeeding Seewarte Limestone is up to 40 m thick and probably early Emsian in age (ERBEN et al., 1962; KREUTZER, 1990; SCHÖNLAUB, 1985). It is characterized by dark-grey colour, large molluscs (*Hercynella*), and abundant algae (PALLA, 1967; JHAVERI, 1969). The limestones are only locally developed and are interpreted as backreef or lagoonal facies. The following, up to 130 m of Emsian Lambertenghi Limestone comprises numerous shoaling upward sequences of 0.5-3 m thick grey limestone beds capped with yellow laminated dolomite (10-30 cm thick layers). Characteristic components are oncoids and other coated grains, algal lumps, bored and enveloped skeletal grains, and algae. Fibrous calcite crusts, algal laminites, open space structures (birdseyes), flat pebble limestone conglomerates and grading are conspicuous elements of the Lambertenghi Limestone. Dolomitization was probably early diagenetic. The sediments are interpreted as peritidal carbonates deposited on a shallow open to semi-restricted marine platform with a water depth ranging from shallow subtidal to supratidal (POHLER, 1982).

The nature of the Lambertenghi Limestone (Emsian) with shallowing upward carbonate-dolomite cycles indicates deposition in arid climate.

The overlying Spinotti Limestone is composed of basal crinoidal and bioclastic limestone (90 m thick) and upper "birdseye limestone" with *Amphipora* (approximately 130 m thick, fig. 20). The lower unit is probably already Eifelian in age (VAI, 1967; KREUTZER, 1992).

The Spinotti Limestone Formation begins at the metal ladder at the base of the Sentiero Spinotti (Track # 145 to Rifugio Marinelli).

Above the massive stromatoporoid debris limestones of the lower Spinotti Limestone follow thickly-bedded unfossiliferous peloidal limestones. They represent 2-3 m thick beds with thin (25-30 cm thick) dolomitic interbeds. This succession is about 60 m thick and is succeeded by about 30 m thick vaguely bedded limestones (0.5-1 m thick beds) followed by 25 m of more distinctively bedded limestones. Characteristic are the dark veining and the laminitic interbeds. Unfortunately thin sections yield little information of this upper part of the succession because of tectonic overprinting. This limestone sequence forms the initial steep part of the Sentiero Spinotti which ends at the ridge at an elevation of 2020 m.

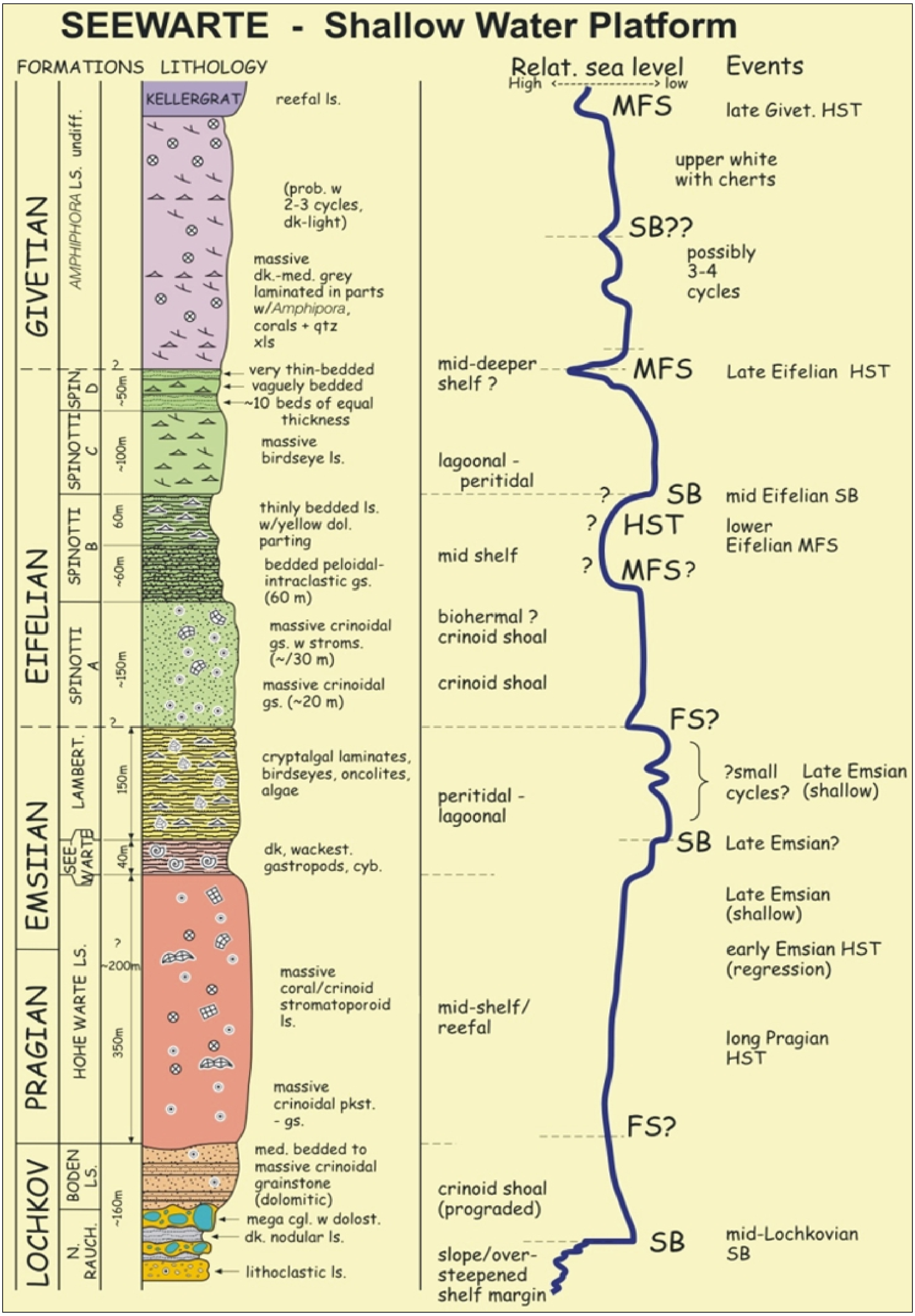

Fig. 19: Section through the southern shallow water facies (Kellerwand Nappe) measured at Mount Seewarte. Sequence stratigraphic interpretation by C. BRETT.

Fig. 20: The succession with the basal Spinotti Limestone at the Sentiero Spinotti.

The track crosses a wide valley that opens to the SW where birdseye limestones are exposed between the 2020 m and the 2200 m ridge (Costone Stella). According to BANDEL (1972) and VAI (1963) they still fall into the Eifelian. In our opinion, however, these strata are equivalent to the basal portion of the Givetian.

A yellow limestone bed is exposed above the trail (fig. 21) at elevation 2120 m, yielding abundant stringocephalid brachiopods.

The trail passes through birdseye limestones with limonitic crusts and intraclasts interbedded with fossiliferous dark Amphipora Limestones containing large gastropods, amphipores, stromatoporoids and stringocephalid brachiopods.

The beds dip with 36º to the south and are overlain by bedded limestones with dolomitic layers (fig. 22). The determination of the brachiopods awaits confirmation, however, it is possible that these beds are already Givetian in age. On fig. 19 this lithological change is indicated between the Spinotti D unit and the Amphipora Limestone.

The track to Costone Stella (2200 m) crosses poorly preserved birdseye limestones with few *Amphipora*-rich horizons. In the following karst terrain dark Amphipora Limestones are exposed in places associated with solitary rugose corals.

They appear to be interfingering with light coloured birdseye limestones. Their thickness is difficult to estimate due to tectonic complications.

The hitherto undescribed birdseye and overlying *Amphipora* limestones are informally referred to as Costone Stella Limestone.

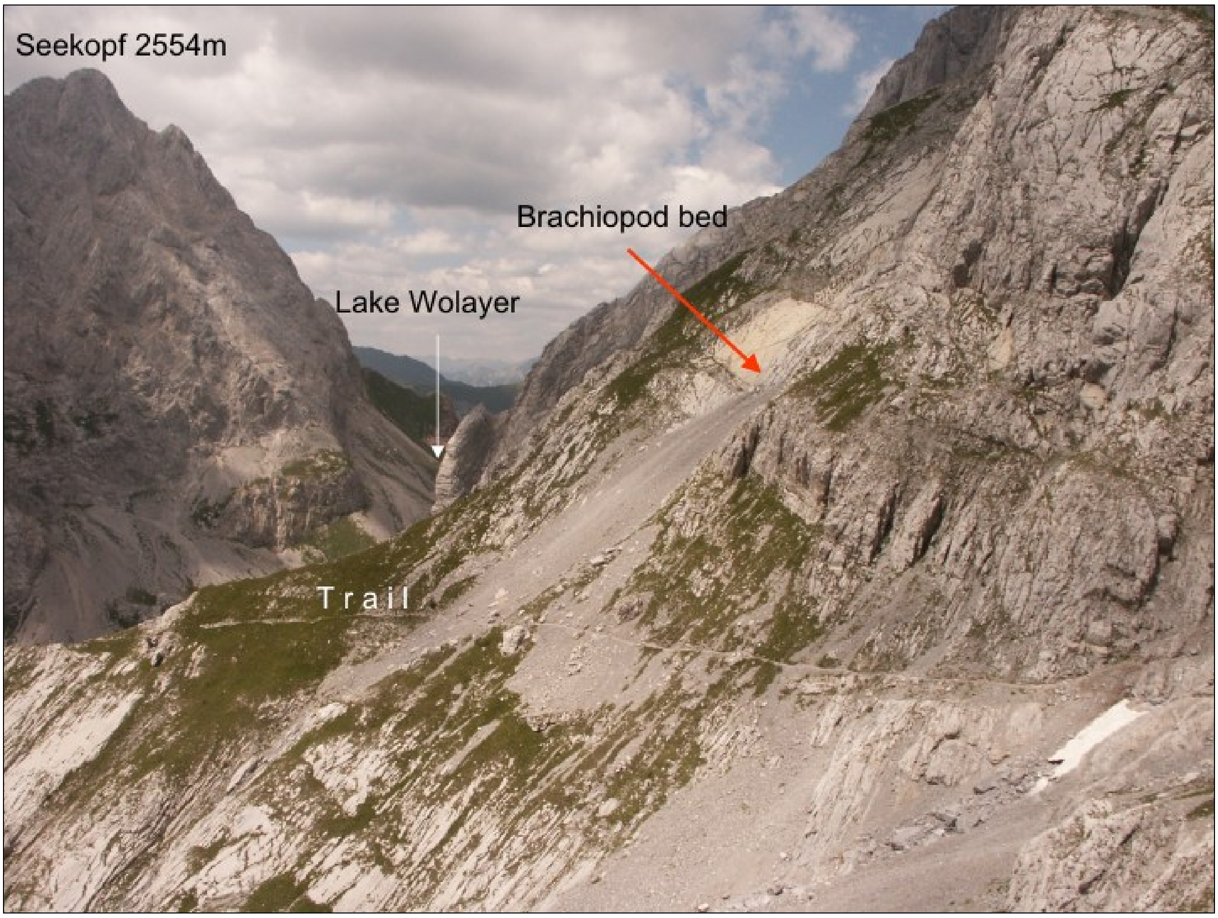

Fig. 21: The yellow bed above the Sentiero at elevation 2120 m.

The karst terrain ends at the track to the south side of Mount Hohe Warte (track # 143a) and here the first reefal limestones of the Kellergrat Limestone Fm. occur. *Amphipora rudis* was determined from this succession and indicates a Givetian to Frasnian age (E. FLÜGEL, pers. comm., 1981). The corals recovered from this area include *Scruttonia julli* (PEDDER) which also suggests a Frasnian age (ÖKENTORP-KÜSTER & ÖKENTORP, 1992). However, both authors caution that the total coral fauna contains elements characteristic of Givetian as well as Frasnian associations.

Along the trail to Hohe Warte *Amphipora* limestones are exposed to the west of the trail and coral limestones to the east. It is likely that a facies transition is present here; however, the rugged terrain and tectonic complications make this relationship difficult to assess.

The succession ends at an unconformity which separates birdseye limestones of unknown age from lower Carboniferous (anchoralis Zone) deep water limestone with goniatites.

To the north, in the upper Kellerwand Nappe both Lambertenghi and Spinotti Limestones grade into Eiskar Limestone, composed of algal-rich grainstones with interbedded "birdseye limestones". This facies ranges from Emsian into middle Givetian and is about 320 m thick. KREUTZER (1990) regarded it as backreef facies (crinoid-cortoid facies).

The Kellerwand Nappe was probably thrust over a segment of the Devonian shelfbreak and upper slope, whose nature is therefore not known. Hints of this facies are reflected in the composition of calciturbidites and other gravity flows which originated at the (now buried) shelfbreak and/or foreslope.

Fig. 22: Birdseye and Amphipora Limestones exposed along the upper part of the trail "Sentiero Spinotti".

Discussion

The term carbonate platform is used herein as a general term for a thick sequence of shallow water carbonates (TUCKER & WRIGHT, 1990).

Prerequisites for the development of a carbonate platform are
1. Presence of plants and animals which produce carbonate minerals rapidly.
2. A shallow illuminated seafloor in tropical to subtropical seas.
3. Warm water (T >18ºC).

Indicators for shallow warm water in Devonian time include
1. Abundance of massive and branching stromatoporoids (*Amphipora*, *Stachyodes*).
2. Colonial rugose corals and tabulozoans (chaetetids and tabulate corals).
3. Calcareous green and bluegreen algae (e.g. Dasycladales and Udoteacean algae).
4. Ooids, oncoids, aggregate grains and common pellets.

Most of these organisms and components occur in the carbonates of the Kellerwand Nappe and the presence of these climate sensitive lithologies in the Carnic Alps indicates deposition in a tropical marine environment (30º or less). In recent oceans cool water carbonates accumulate at depths down to 350 m or more, from carbonates produced by non-phototrophic organisms such as benthic forams, molluscs, bryozoa and red algae (foramol or bryomol assemblages). In the Devonian, crinoids feature prominently in cool water (as well as warm water) assemblages. The condensed Silurian and Ordovician sediments underlying the Devonian carbonate platform show all the hallmarks of cool water carbonates and indicate drifting of the Carnic Alps depositional system from high to low latitudes in the Paleozoic.

Geotectonic settings of shallow marine environments can be
1. Passive continental margins
2. Intracratonic basins
3. Failed rift basins
4. Arc-related basins
5. Oceanic islands
6. Foreland basins.

The lack of volcaniclastic and siliciclastic sediments excludes arc-related and foreland basins as environment for the Carnic Alps carbonates. Deposition in a rift-related basin was suggested by SPALLETTA et al. (1983) and also KREUTZER et al. (1997) proposed an extensional regime of enhanced mobility for the CA depositional system.

Several types of carbonate platform are known, including rimmed shelf, ramp, epeiric platform, isolated platform and drowned platform.

A rimmed shelf is a shallow water platform with a pronounced shelf break and slope into deeper water. Along the shelf margin reefs or shoals may develop, which restrict circulation on the shelf. Some rimmed shelves have deep intrashelf basins behind the shelf rim. Widths of rimmed shelves can vary from a few to 100 km. Accretionary, bypass and erosional types of rimmed shelves can be distinguished.

A carbonate ramp is a sloping surface with a low gradient (a few metres per kilometre) where shallow water carbonates pass gradually into deeper water and then basinal deposits. In contrast to a rimmed shelf there is no distinct break of slope in shallow depth. Two types of ramp can be distinguished. Homoclinal ramps have relatively uniform slopes whereas distally-steepened ramps have an increase in slope gradient in the outer deep ramp region (READ, 1985). The latter are characterized by gravity flow deposits and slumps similar to accretionary shelf margins but differ in the location of the slope break, which is in deeper water. As a consequence the resedimented deposits on the lower slope (or in the basin) consist of outer ramp and upper slope deposits.

Ramp facies are controlled by ocean currents and waves and distinctive sediments are carbonate sands and tempestites. In the shallow ramp region patch reefs, beach barriers and sandy shoals can form and provide sheltered back-ramp areas where lagoonal, shallow subtidal to supratidal flats occur, frequently associated with evaporites and/or paleokarsts and paleosols.

Epeiric carbonate platforms are extensive areas of negligible topography and gradient which covered extensive areas of cratons. Water depth rarely exceeded 10 m and vast areas covered by shallow subtidal to supratidal carbonates are characteristic. Deep intraplatform basins surrounded by ramps or slopes were sometimes present. The influence of tides on these platforms is under debate and a tidal island model is contrasted with a model proposing dampened or no tides and storm domination (JAMES & PRATT, 1986; IRWIN, 1965).

Epeiric carbonate platforms have no recent counterparts but were present particularly in the Paleozoic and in the Triassic-Jurassic in times of lengthy drift phases after plate separation.

Isolated carbonate platforms are shallow water platforms surrounded by deep water. Their size is variable but most are of small size and characterized by steep slopes. Frequently they develop on structural blocks in regions where renting and rifting occurred or on submerged volcanic seamounts. Sedimention on and around isolated platforms is controlled by prevailing wind and storm directions. Reefs are particularly developed at the windward side of isolated platforms and adjacent platform slopes receive little fine grained sediment from the platform interior. Off-platform transport is concentrated on the lee-ward side of the platform where much sediment is redeposited on the platform slope. Drowned platforms typically have deep-water carbonates overlying the shallow-water facies.

The areal extent of a carbonate platform is governed by the size of the platform and amount of siliciclastic sediment. In the CA siliciclastic influx is virtually absent in relation to the large amount of carbonate sediments. The existence of an extensive carbonate platform is indicated by the wide areal extent of the shallow water facies and KREUTZER et al. (1997) calculated a ratio of 12:1 for thicknesses of shallow water versus pelagic carbonates. However, occurrences of shallow water facies are disjunct and it is to date not known whether one continuous or several smaller platforms were present.

The stratigraphic succession investigated at Mt. Seewarte is composed of shallow water carbonates except for the lower Lochkovian interval which consists of allodapic and pelagic limestones and shales. The lateral change to the west is to date not known. To the east the Lower Devonian succession was documented at Hohe Warte (SCHÖNLAUB & FLAJS, 1982; KREUTZER, 1992) and deepening in this direction is indicated. The lithological changes seen in the basal Seewarte section imply that the shallow water facies prograded over a carbonate ramp or slope and suggests that either an accretionary shelf margin or a distally steepened ramp existed in this time interval. The clasts in the lower Lochkovian debris flow deposits are largely slope lithologies.

The overlying mid- to upper Lochkovian limestones are composed of crinoidal grain- and packstones with frequent graded beds interbedded with massive disorganized beds. The lithology suggests deposition at or near crinoid shoals with frequent remobilisation of skeletal debris presumably through storms or gravity flows. The overlying Pragian sediments also contain abundant well preserved crinoids but in contrast to the Lochkovian they are massive carbonates without any graded or other indication of hydraulic sorting. The good preservation of the crinoidal debris indicates deposition close to their original habitat. In addition, stabilization of seafloor sediments permitted local development of carbonate buildups (mounds). The sediments are well washed with little mud and much fibrous calcite. The high diversity of accessory skeletal components indicates good living conditions for organisms in well aerated shallow subtidal marine environment for most of the Pragian. The observed shallowing upward trend implies progradation of shallow platform sediments over the crinoidal storm beds deposited in the Lochkovian.

During late Pragian and early Emsian stromatolites and stromatoporoids became prolific and small patch reefs developed. The succeeding Seewarte Lst. is dark grey with numerous molluscs and stromatolitic bindstone. Deposition in a lagoonal setting is inferred and suggests that for a short time interval restricted circulation occurred at least locally. The reasons could be the formation of lagoonal environments behind substantial Emsian buildups or in an intra shelf basin which became restricted due to a sea level fall.

The succeeding Lambertenghi Limestone (Emsian) was deposited in shallow subtidal to supratidal environments, and hence in shallower water than the previous sediments. Further progradation of the platform is indicated. This continuous trend of shallowing ends with deposition of the Spinotti Limestone (Eifelian), beginning with muddy reefal limestone succeeded by crinoidal packstone.

Initial sediment composition on a platform is largely determined by its carbonate-secreting biota; resultant lithofacies, however, are determined by energy spectrum and sediment binding. In general terms lithofacies on rimmed platforms are dominantly muddy while those on open unrimmed platforms are grainy. On the CA platform grainy facies are dominantly found in the upper Lochkovian to Pragian, whereas lagoonal and muddy facies are found in the dark Seewarte Limestone and (to a lesser degree) in the Lambertenghi Lst. where an oscillation from muddy to grainy to dolomitic facies occurs. This pattern suggests, that a rimmed platform began to form in the mid-Emsian.

The balance between sediment production and sediment transport determines the growth potential of the platform; sea level fluctuations and subsidence cause changes in environmental conditions reflected in vertical accumulation of platform sediments.

Foreslope/Slope Facies (Cellon Nappe)

Several peaks and massifs belong to the Cellon Nappe including Cellon, Freikofel, Großer Pal, Gamsspitz, Pizzo di Timau (fig. 16). BANDEL (1972) and KREUTZER (1990) both interpreted the Devonian sediments of the Cellon Nappe as remains of forereef, foreslope and slope. KREUTZER (1992) subdivided the transitional Devonian facies into several formations based largely on sections at Cellon and Kellerwand. For this study sections located at the Cellonetta avalanche cut (Cac) and along the Steinberger path at the eastern side of Mount Cellon (Pragian to Emsian) are considered (fig. 23).

Fig. 23: View of Mount Cellon with location of Cellonetta avalanche cut and Steinberger path from Mount Großer Pal.

The Lochkovian Rauchkofel Limestone is well exposed at Cac and begins above sample number 47 of WALLISER (1964) just above the Silurian/Devonian boundary. The Lochkovian is 65 m thick here and two units can be distinguished: the lower unit (35 m thick) consists of alternating dark grey to black bituminous platy limestones with subordinate intercalations of black calcareous shales or shaly limestones. The limestones are commonly calcisiltites with coarser bioclastic material at the base of thin laminae. Grading and convolute bedding is common, fine laminations are characteristic. Limestone beds grade upward into shales. In some layers intraformational breccias occur. Dark chert nodules and concretions are common; dolostone beds and patchy dolomitization is characteristic. Bioturbation and ichnofossils are rare in this lower part of the succession. The fauna consists of nektonic, planktonic and benthic organisms. Trilobites and thin-shelled bivalves are abundant and crinoid debris features prominently in most samples. Graptolites are rather sparse in the lower Lochkovian.

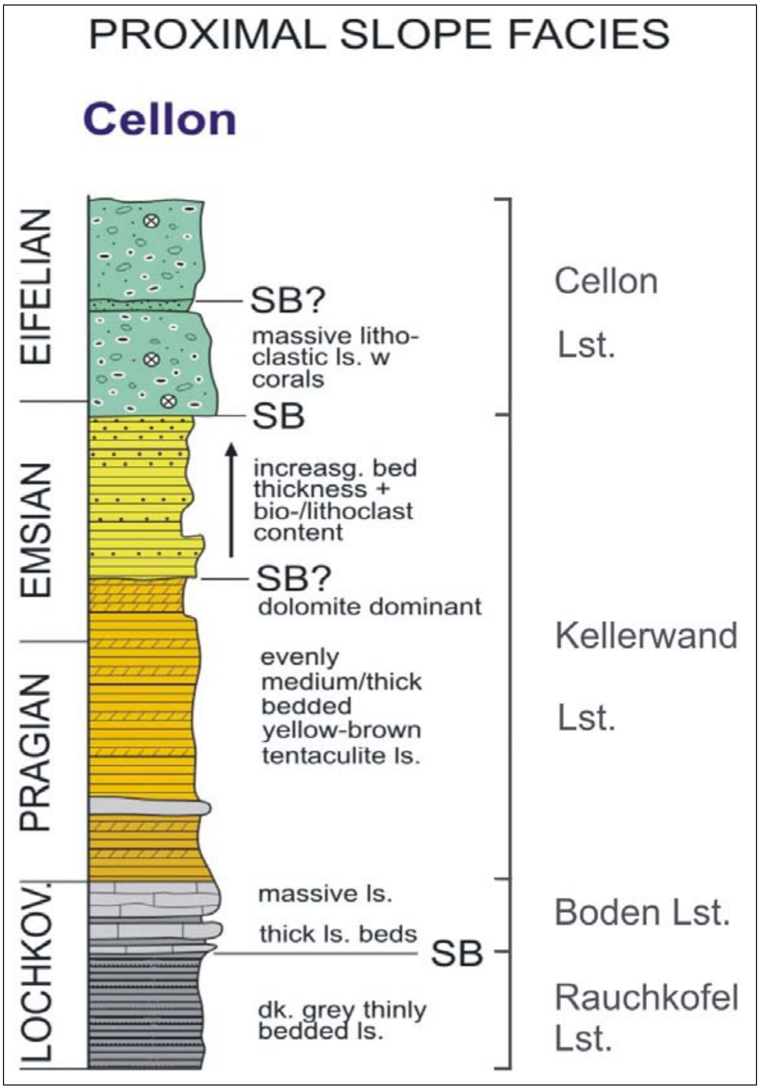

Fig. 24: Generalized section through the proximal slope facies (Cellon Nappe) measured at Mount Cellon.

The upper Lochkovian unit (30 m thick) consists of massive, grey nodular limestone units interbedded with thin-bedded grey limestones (Figs. 23, 24). The nodular limestone typically is bioturbated calcisiltite with numerous peloids and some crinoid debris in a matrix of lime mud. The main difference to the lower Lochkovian is the presence of bioturbation. The interbedded platy and vaguely laminated limestones are fine-grained calcisiltite with dark-brown lamination. Trilobite and other shell debris is oriented parallel to the bedding planes. Bioturbation is reduced. The lithological change from dark platy limestones and shales to dominantly nodular and lumpy limestones can be seen from afar because the latter form prominent steep ribs in the succession, not only at Mount Cellon but also at the inaccessible Kellerwand section to the east.

The Lochkovian/Pragian boundary may coincide at Mount Cellon with the lithological change from grey nodular and platy limestone to yellow dolomitic tentaculite-bearing limestone referred to as Kellerwand Lst. (~145 m thick). At the Steinberger Path this unit is poorly exposed and recessively weathering. The Kellerwand Limestone is largely composed of fine-grained limestone intercalated with muddy calcarenite and calcisiltite beds. The fine-grained lithology is a microskeletal peloidal wackestone interbedded with very abundant broken and complete tentaculites. Skeletal debris of trilobites, crinoids and brachiopod shells is also common. Dolomite crystals are dispersed throughout in variable amounts ranging from numerous xenomorph and idiomorph crystals to pervasive dolomitization. Stylolites and their surroundings are particularly affected by dolomitization. Most of the fine-grained sediment is thoroughly bioturbated.

The coarser calcisiltites and calcarenites are composed of medium sand-sized crinoid debris in a matrix of peloidal grainstone to packstone. In some beds grading can be seen with medium sand-sized crinoid debris grading upward into peloidal grainstone to packstone with few crinoid fragments. Accessory skeletal material is derived from brachiopods and trilobites. In some cases the silty matrix is completely replaced by dolomite. The contacts between grainy beds and muddy lithologies are not sharp but vague and uneven due to the activities of burrowing organisms.

The Kellerwand Lst. is succeeded by the 120 m thick Emsian Vinz Limestone which is characterized by decreasing dolomite content, and increasing bed thickness and lithoclastic content. The Pragian/ Emsian boundary is not clearly defined based on litho- or biostratigraphy. The succession consists of thick bedded bioclastic wackestone intercalated with peloidal/bioclastic pack- and grainstones. The Lower Devonian ends at the transition to grey massive bioclastic wackestones, pack- and grainstones of the Cellon Lst. Fm., averaging 210 m in thickness and forming the peak massif of Mount Cellon.

Discussion

BANDEL (1972) measured and dated several sections through sediments of this "transitional facies" of the Cellon Nappe and also considered the "pelagic limestones with common redeposited beds" (his sections at Woderner Törl, Valentin Alm, Cellon, Cresta di Collinetta, Freikofel, Gamsspitz, Pal Grande, Pizzo di Timau, Elferspitz) transitional between the basin floor and the shallow water platform (BANDEL, 1974). The Lower Devonian interval of this "transitional facies" is characterized by bioturbated thinly bedded to lumpy wackestones and packstones interbedded with thin, fine-grained graded horizons. BANDEL (1972) interpreted the graded beds with abundant shallow-water derived skeletal material as turbidites and the intervening fine-grained beds as pelagic background sedimentation.

BANDEL (1972) noticed, that the composition of resedimented beds reflects the environment of shallow water deposition, where echinoderm fragments were most abundant in the Lower Devonian. KREUTZER (1990: 308) pointed out that debris derived from the backreef travels further downslope than the relatively coarse reefal debris. This may be reflected in the proximal to distal trend from Cellon (proximal) to Gamsspitz (intermediate) to Woderner Törl (distal) postulated by BANDEL (1972).

HLADIL et al. (1996) proposed a turbidite origin for Pragian lime mudstones of the Prague Basin on the basis of graded bedding, abundance of calcisiltite components and imbrication of tentaculite shells. The indistinct outline of the turbidite beds is here ascribed to dewatering after deposition.

To understand the geometry of the carbonate depositional system in the Carnic Alps it is necessary to take into consideration mechanisms of carbonate accumulation.

The surface slope that is maintained by carbonates is determined by the combined effects of (1) rate of *in situ* carbonate accumulation and (2) the depositional angle of the sediment shed from the bank or reef crest as talus and turbidite. Commonly the slope will be steep when most of the carbonate accumulates on the shelf. If carbonate accumulation does not vary much with water depth, then carbonate accumulation will maintain a uniform slope which parallels the underlying surface. This slope may be steepened with time by sediment shed as talus and turbidite from the reef or bank crest.

Distal Slope Facies at Mt. Findenig, Hoher Trieb and Oberbuchach

Sections through the distal slope facies were measured at Hoher Trieb, Oberbuchach and Findenig-kofel. The sections at Hoher Trieb and Oberbuchach were previously documented in terms of litho-facies and biostratigraphy by SCHÖNLAUB (1970, 1985). The succession at Findenigkofel was mapped and studied in detail by PÖLSLER (1969). Limestones of the distal slope environment were measured at Oberbuchach, Findenigkofel and Hoher Trieb (fig. 25). Between 13 and 31 m of section belong to the Lochkovian.

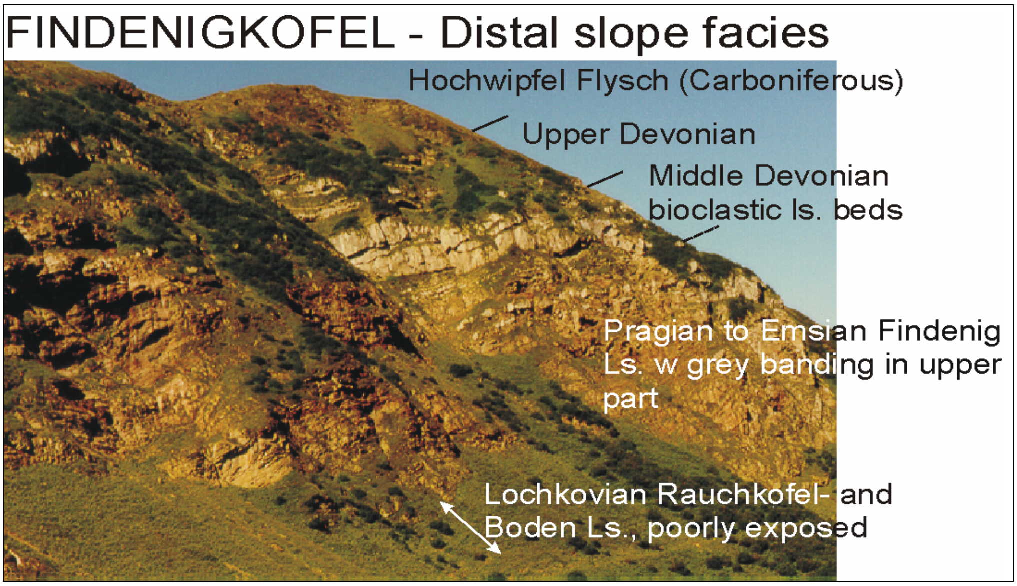

Fig. 25: View of distal slope sediments exposed at Findenigkofel from the Waidegger Alm.

At Oberbuchach and Findenigkofel, the characteristic lithological change from dark platy dolomitic and cherty limestone with graded beds to lighter grey nodular and "flaser" limestones can be observed. Pragian and Emsian limestones are red "flaser" and nodular limestones, both belonging to the Findenig Limestone. The Pragian is well constrained based on conodonts at Oberbuchach II (OB II) and about 30 m thick (figs. 26a, b).

The Emsian segment at OB II is about 32 m thick and characterized by higher limestone content and thin light-grey calcilutite and calcarenite beds intercalated with red "flaser" limestones. Calciclastic beds increase in thickness and coarseness up-section. At Findenigkofel a 100 cm thick grey bed is exposed consisting of 50 cm thick grey lumpy limestone composed of lime-mud with numerous tentaculites with smooth walls, trilobite fragments and few ostracods. Parting material between lumps consists of mm-thick brown material (unresolvable by light microscopy) with silt-sized xenomorph dolomite crystals. This bed is overlain by five centimetres of graded and laminated limestone, composed of fine sand- to silt-sized peloids, crinoid debris (with syntaxial overgrowth) and thick-shelled dacryoconarids (?), followed by seven centimetres of wavy laminated calcisiltite. A six centi-metre thick laminated shale unit concludes the succession. It is overlain by three beds of wavy lami-nated limestone (20 cm thick together) and finally 20 cm of grey lumpy limestone follows, similar to that at the base.

Discussion

The succession observed at Findenigkofel is characteristic of turbidites deposited from low-density flows. Calciturbidites of such small grain sizes show structures similar to siliciclastic turbidites and the succession described above shows Bouma sequences T_a (graded calcarenite), T_b (lower horizontally-laminated division), T_c (cross-laminated division) and T_d (upper horizontally-laminated division). The pelite interval (T_e) is missing. According to STOW (1986) fine-grained turbidites are characteristic of distal slopes or ramps and with increasing distality the T_b and T_c divisions may be missing.

VAI (1980) discussed the sedimentary environment of Devonian pelagic limestones from the Stua Ramaz section north of Paularo in the vicinity of Monte Zermula. VAI (1980: 80) noted the abundance of "grey allodapic limestone beds intercalated with red, partly nodular pelagic beds" which he described in detail. His description suggests that these limestones belong to the Findenig Facies which occurs also to the west at Findenigkofel and Oberbuchach. The colour change from grey to red, associated with increasing siliciclastic/carbonate ratio, is interpreted as lowered sedimentation rate by VAI (1980).

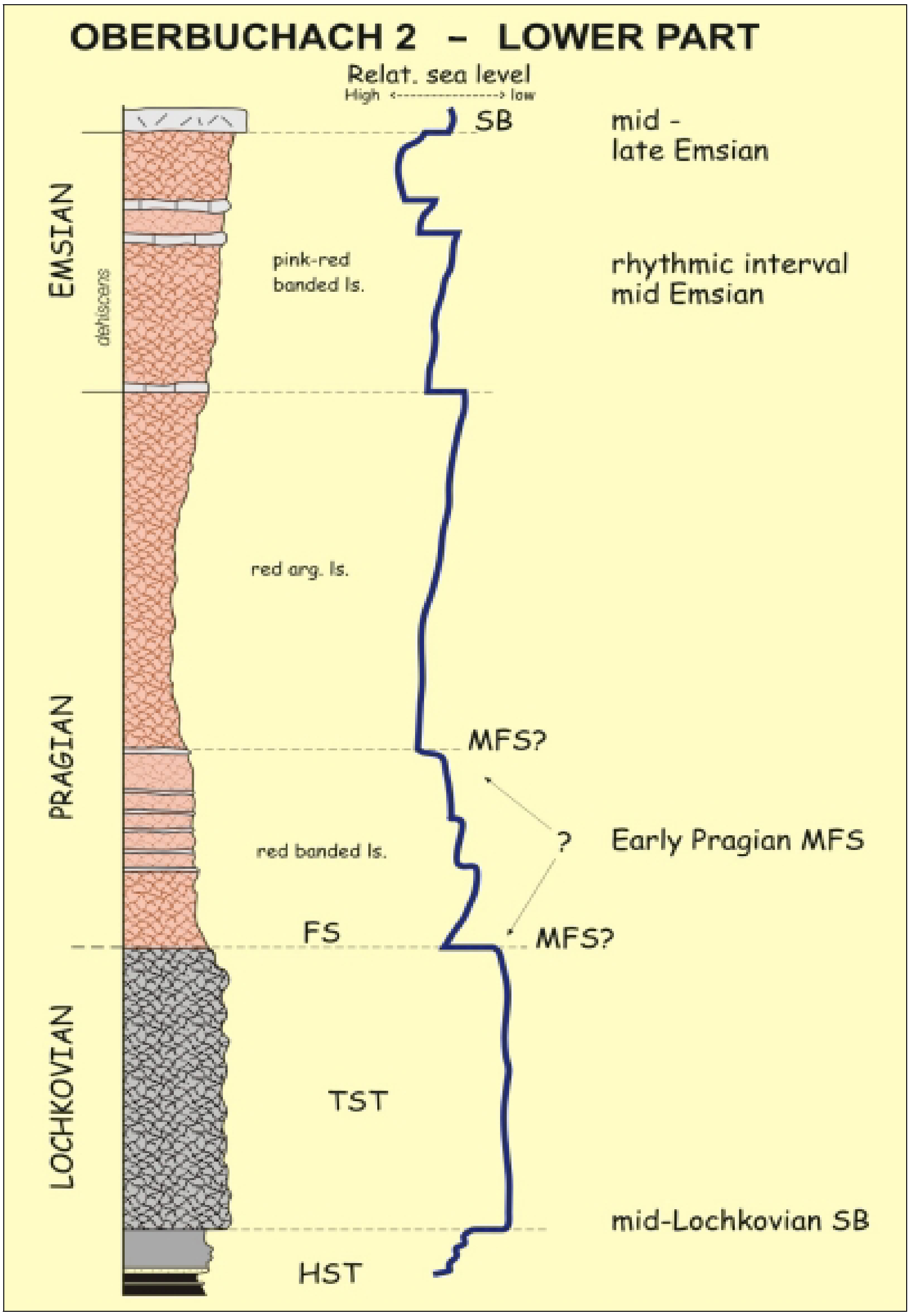

Fig. 26a: Section through the distal slope facies measured at Oberbuchach II, lower part.
Adapted from SCHÖNLAUB, 1985. Sequence stratigraphic interpretation by C. BRETT.

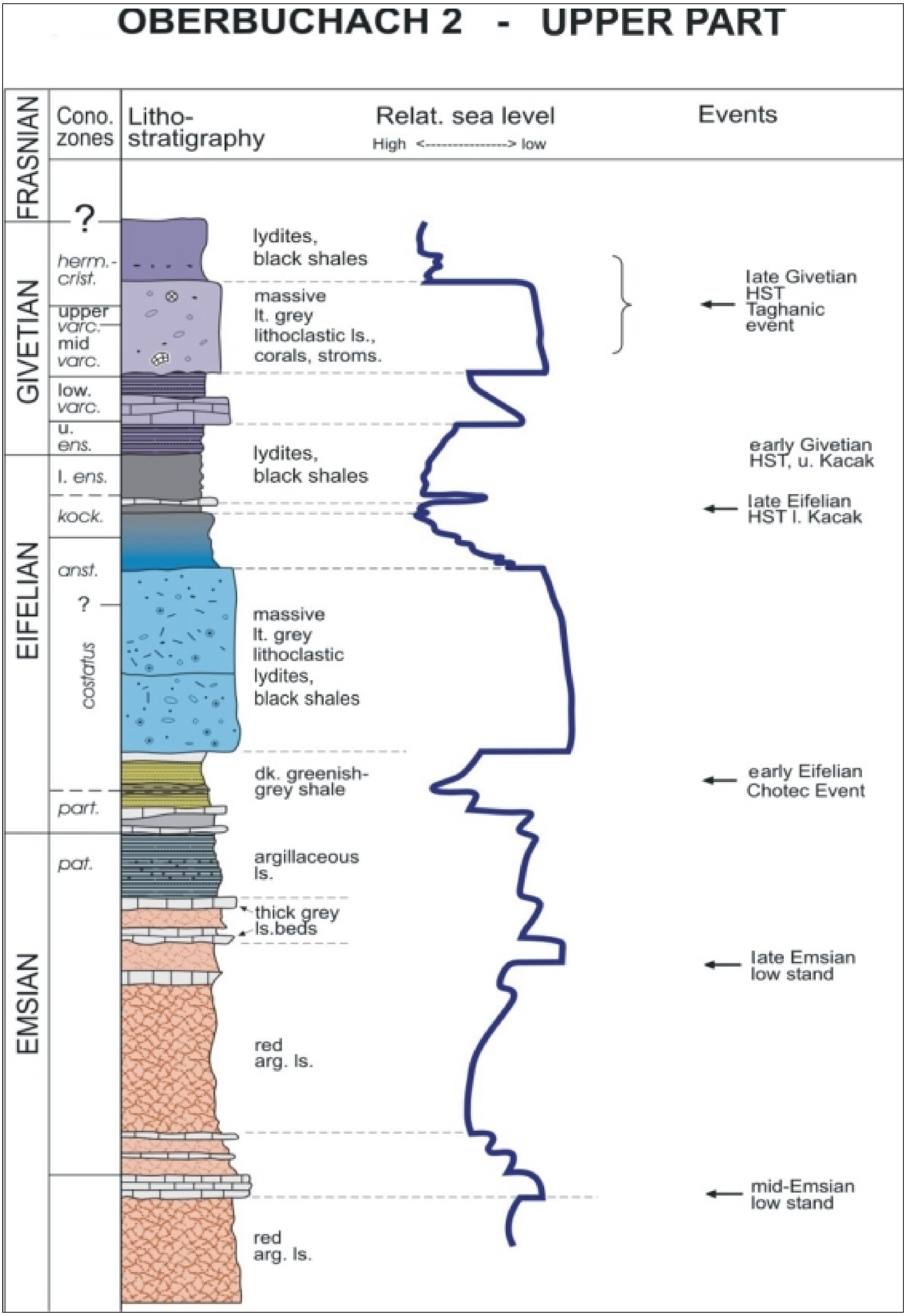

Fig. 26b: Section through the distal slope facies measured at Oberbuchach II, upper part. Adapted from SCHÖNLAUB, 1985. Sequence stratigraphic interpretation by C. BRETT.

He discussed two different types of events that could account for the grey allodapic beds with redeposited shallow water material: (1) storms affecting the carbonate platform could stir up turbid clouds which drifted seaward and settled out of the water column over slope and basin. (2) Turbidity currents resulting from sediment overloading at the platform margin or on the upper slope. The grey limestone bed described above shows all indications of turbidite deposition, however, many of the grey limestones are very fine-grained and deposition from turbid clouds cannot be discounted. It is a well-known mechanism for deposition of fine-grained carbonates on recent carbonate slopes.

Condensed Pelagic Limestone Facies (Rauchkofel Facies)

Lower Devonian carbonates of the Rauchkofel-, Boden-, and Findenig Limestones were assigned to the Rauchkofel Facies (SCHÖNLAUB, 1979, 1985). Outcrops are confined to the Rauchkofel Imbricate Nappe Complex. The limestones of the Rauchkofel Facies are largely devoid of gravity flow and other coarse redeposited units and differ in this respect from the Findenig Facies.

Lower Devonian sections through the Rauchkofel Facies were measured at Seekopfsockel (Sks) and Rauchkofelboden (Rkb); sections through the Pragian/Emsian interval and the Emsian only, at Frauenhügel (H) and Wolayer Glacier (W.G.), respectively.

At Seekopfsockel 83 m of Devonian limestone are exposed of which 77.8 m are Lower Devonian. The Lochkovian interval encompasses 16.1 m of thin-bedded dark limestone, lighter grey "flaser" limestone and pink crinoidal limestones. Undifferentiated Pragian and Emsian "flaser" limestones (Findenig Limestone) with distinctive red colour comprise the remainder of the Lower Devonian succession.

Measuring of the section started at sample number 350 which marks the beginning of the Lochkovian (SCHÖNLAUB, 1980). The Lochkovian succession begins with 2.3 m of dark grey fine-grained bedded limestones with crinoidal debris. Particularly at the base white calcite veining is developed. Above, 1.1 m of thin-bedded dark limestones and shales follow overlain by 2.8 m of grey fine-grained stylo-bedded limestone. Thin sections show microskeletal wackestones with shell debris from trilobites, nautiloids and tentaculites in addition to relatively coarse crinoid debris. In one sample algae, small brachiopods and gastropods were found. Neither grading nor lamination was seen, and most units show signs of bioturbation in contrast to the laminated lower Lochkovian at Mount Cellon. The lithologies do not readily indicate deposition from turbidites but could also be deposits of a deep subtidal environment.

The thin-bedded limestones and shales are succeeded by 6,8 m of grey and pink, hackly weathering, crinoid limestone. All samples from this interval are composed of peloidal grainstone to packstone with varying amounts of coarse crinoid debris. Vague grading is seen in some samples. The different beds vary in the size of the skeletal debris which ranges from medium to coarse sand-size, whereas the peloids are of fine sand-size (rarely silt). The limestones in this interval of section show signs of resedimentation and the coarse crinoid debris was probably transported from a source further up-slope. Crinoidal calcarenites of similar age were reported from the Poludnig-Oisternig region in the eastern Carnic Alps by HERZOG (1988). He interpreted them as debris derived from a shallow water source and deposited down slope among grey lumpy and nodular limestones. The successions of Findenig Facies in this region contain numerous slump horizons, and an interpretation of the crinoidal units as slumps is also possible.

Fine-grained grey limestones with stylo-flaser fabric (3.1 m) form the top of the Lochkovian interval. They are characterized by yellow-brown stains, parting material and stylo-cumulate. Thin sections show skeletal wackestones with debris from tentaculites, ostracods, and trilobites. Crinoid debris is rare. The yellow-brown tinge stems from ferroan dolomite and brown flocculent matter in partings between lensoid limestone lumps.

The Silurian/Devonian boundary interval was investigated in detail in sections Sks. and Rkb. (SCHÖNLAUB, 1980, 1981). In both sections the lower Lochkovian is highly condensed (2 m at Rkb. compared to 9 m thickness at Oberbuchach).

Comparison of the Sks section with that at Rauchkofelboden shows similarities on a large scale but differences in details. The earliest Lochkovian is generally represented by dark, thinly bedded limestones interbedded with shales. This lower condensed unit is succeeded in all sections by the Boden Limestone, a grey, bedded "flaser" limestone with orthoconic nautiloids at section RkB with a central unit of grey to pink echinoderm packstone occurring only at section Sks.

An abrupt colour change from grey to red (between sample # 35 and 36) marks the beginning of the Pragian (SCHÖNLAUB, 1980). The lower 23.5 m of the succession consist of interbedded red shale-rich and red and green mottled limestones, both with stylo-flaser fabric. Commonly units consist of 0.4 to 0.9 m of red recessively weathering shale-rich "flaser" limestone alternating with 1 m to 4 m thick massive red-pink "flaser" limestone. The Pragian/Emsian boundary is not clearly defined to date. SCHÖNLAUB (in BANDEL, 1972) placed the Siegen/Ems boundary about 18 m above the onset of red "flaser" limestone deposition (just above sample # 48 [SCHÖNLAUB, 1980]).

The middle part of the red "flaser" limestone interval consists of 20 m of relatively massive "flaser" limestone with a peculiar pattern of patches of grey arenaceous wackestone that probably represent burrow fills. The fractures run parallel and at low to moderate angles to bedding planes. Their number increases upsection. The cross cutting relationship with calcite filled fractures indicates that they are products of a later stage of deformation rather than synsedimentary fractures. The remaining 18.2 m of red "flaser" limestone are characterized by increasing limestone content and diminishing of the red colour. In the lower part of this interval are up to 15 cm thick grey limestone beds with no or little stylolites developed, spaced about 1 m apart. In the upper part occur 5-20 cm thick shale-rich beds alternating with 10-30 cm thick limestone-rich beds. Another abrupt colour change from red to grey indicates a position close to the top of the Lower Devonian. The exact location of the Emsian/Eifelian boundary is not known; it may coincide with the colour change from red to grey "flaser" limestone or may lie slightly lower as in the Wolayer Glacier section (GÖDDERTZ, 1982).

The Pragian/Emsian interval is characterized in all sections by the distinctive red "flaser" limestone. It appears quite uniform, but three vaguely confined units can be distinguished: (1) a lower unit with pink and red banding and locally, with pink and green mottling (at Sks only), (2) a central shale-rich unit and (3) an upper red and grey banded unit.

Discussion

The deep water sections of the lower Lochkovian Rauchkofel Limestone Formation must be regarded as extremely condensed (SCHÖNLAUB, 1980), the remainder of the Lower Devonian is condensed compared to the distal slope sequences of the Findenig Facies. The additional amount of sediment derived from redeposition could probably explain the different thickness of the Findenig Facies.

BANDEL´s (1974) descriptions of "pelagic limestones with rare redeposited beds" from the region around Mount Rauchkofel include the limestones from sections measured at E. Pichl Hut, Seekopfsockel, Rauchkofelboden and Wolayer Glacier. He distinguished 5 different lithofacies and, based on his analyses, suggested deposition of the pelagic limestones in a basinal environment, ranging in depth between 300 m and 3000 m. SCHÖNLAUB (1980) suggested that the Lower Devonian (Lochkovian) cephalopod and tentaculite limestones of the Rauchkofel Facies were deposited on basinal swells and ridges, which formed as a result of increased bottom mobility at the end of the Lochkovian. KREUTZER et al. (1997) suggested that an extensional tectonic regime was responsible for the increasing bottom topography. A similar interpretation was invoked for the distribution of pelagic limestones and shales in the Frankenwald (H. TRAGELEHN, pers. com., 1999). The condensed pelagic limestones could also be deposits of the proximal basin floor or slope rise (the distal basin floor is preserved in the shales of the Bischofalm Facies) which was not reached by most turbidites. However, inspection of the Zollner Formation shows, that these rocks consist of interbedded cherts and siltstone units with the latter showing sedimentary structures indicative of turbidite deposition (flame structures, graded bedding, cross lamination, convolution). Consequently turbidites did reach the basin floor and their lack in the condensed pelagic limestone facies suggests that turbidites either by-passed this depositional environment or that it represented a separate depositional area.

The study of modern carbonate slopes shows that many different settings can occur along strike, depending on the varying oceanographic and geographic parameters. For example several different settings were documented from the Bahaman carbonate margin and slope, including a diagenetic ramp (MULLINS & NEUMANN, 1979). This model describes slopes which are quite gentle with little mass sediment transport. It is based on the situation on the northern side of the Great Bahama Bank where periplatform facies show downslope transition from hardgrounds to nodular ooze to unlithified ooze. Because of the windward position of this margin, redeposition involves mainly periplatform sediments with platform-derived material being sparse.

In contrast, the leeward margin at the western side of the Little Bahama Bank is characterized by a large percentage of redeposited platform-derived sediment.

The situation at the Tongue of the Ocean gave rise to the concentric facies belt model (SCHLAGER & CHERMAK, 1979). In this setting is sediment being supplied from wind-ward, leeward and tide-dominated platform margins. Facies belts down slope are narrow and slopes are steep. A basinal pelagic facies is not developed because of the closed and narrow nature of the seaway. The Great Bahama Bank is a carbonate platform that was isolated from the American continent.

These models could account for the differences between the condensed pelagic limestone facies at Mount Rauchkofel and the expanded pelagic/redeposited limestone facies at Findenig and Oberbuchach. It could also explain the different pattern of sedimentation at Hoher Trieb where resedimentation is much reduced.

The change between reduced or zero sedimentation and full supply of oozes in pelagic limestones has been explained in terms of third-order sea-level fluctuations in Jurassic sequences of Spain (FELS & SEYFRIED, 1992). These authors found that lithification and erosion took place in the LST (low stand systems tract), ferromanganese crusts formed in the early TST (transgressive systems tract) and red limestones were characteristic for the late TST and HST (high stand systems tract).

VALENZUELA-RIOS & GARCIA-LOPEZ (1997) observed a diachronous event in pelagic sediments of northeastern Spain. In sections measured in the Catalonian Coastal Ranges and in the Spanish Central Pyrenees a change occurs from black shales with minor dark limestones to more massive light-coloured orange/reddish limestones with marl intercalations. This local event near the beginning of the Middle Lochkovian is marked by the disappearance of more endemic conodont faunas including *Icriodus* and the appearance of more cosmopolitan faunas with species of *Ancyrodelloides*. The conodont genus *Flajsella* is also common in this Middle Lochkovian interval (VALENZUELA-RIOS & MURPHY, 1997).

The Lochkovian stage in the Barrandian sections is coincident with the Lochkov Formation and includes two principal lithostratigraphic units (members): the Radotin and the Kotys Limestones. The Radotin Limestone comprises dark bituminous platy limestones with variable amounts of dark shale intercalations and common cherts. Graded bedding and lamination are common sedimentary structures.

The Kotys Limestone is characterized by light-grey thick-bedded bioclastic limestones with debris from crinoids and brachiopods. A transitional facies is presented by the Kosor Limestone, a grey well-bedded bioclastic limestone with minor shale intercalations. SCHÖNLAUB (pers. com., 2001) noted the occurrence of *Ancyrodelloides transitans* in the upper part of the Lochkovian of the Barrandian.

CHLUPAC (1998) summarized facies trends in the Lower and Middle Devonian of central Bohemia. He recognized several stratigraphic events distinguishable in litho- and biofacies.

In the upper Lochkovian a trend of increasing energy and shallowing occurs followed by abrupt deepening at the Lochkovian/Pragian boundary. This is said to be an event of global significance (CHLUPAC & KUKAL, 1986) which can also be recognized in the Carnic Alps where in the basal Pragian of the Rauchkofel Nappes a significant change in lithology occurs (i.e. from grey to red "flaser" limestone).

The Pragian interval in central Bohemia is, according to CHLUPAC (1998), characterized by a trend of increasing water depth interrupted by a shallowing event at the base of the Zlichovian manifested in the increased transport of coarse biodetritus in the northeastern part. This interval approximates the base of the dehiscens Zone. In the sections at Oberbuchach this is the level where the first grey banded units begin to appear, some of these grey beds are bioclastic calcarenites whereas others are light grey micritic beds which may or may not be fine-grained turbidite deposits. It is also the level, where the limestone content in the red "flaser" limestones increases. This increase could either be due to increased lime production/offshore transport or decreased transport of terrigenous material. In view of the connection with increased turbidite flows, it seems more likely, that increased offshore transport and/or production of lime on the platform is the cause for this higher lime content. This would also imply, that the shale rich Pragian succession is a starved sequence where muds were slowly deposited and spent long periods of time exposed to oxygenated bottom waters which caused them to oxidize. The thicker grey beds which were the result of quasi-instantaneous events (i.e. turbidites) with thicknesses of several centimeters to decimeters were only superficially exposed to the bottom water and remained grey.

During the Pragian reefs and crinoidal sands accumulated on the shelf. During this time of presumably high sea level, transport of coarse material onto the slope was reduced (no high stand shedding!) and on the upper slope dominantly hemipelagic sediment was deposited. On the lower slope supply of carbonate was reduced and deposition rate slow. For the offshore deep basinal sequences (Zollner Formation) deposition of cherts is predicted. The pelagic carbonates of the Rauchkofel Facies display red and pink banding attaining a rhythmic character.

The Pragian/Emsian boundary coincides at Oberbuchach with the beginning of the grey banded interval with more calcareous red "flaser" limestones. At Mount Seewarte the boundary was drawn tentatively at the first appearance of *Polygnathus* sp. between sample numbers 16 and 18 (BANDEL, 1969; VAI, 1973), approximately 50 m below the onset of the lagoonal Seewarte Limestone Formation which is also Zlichovian in age (ERBEN et al., 1962; KREUTZER, 1990; SCHÖNLAUB, 1985). The succeeding Lambertenghi Limestone is composed of shallowing upward cycles of shallow intertidal to supratidal limestones and dolomites. Obviously, during the Emsian the platform margin prograded far seaward, a process that must have led to steepening of the slope. This steepening is reflected in the increasing occurrence and number of gravity flow deposits on the proximal slope. The grey limestone beds of the distal slope nappe represent the distal turbidites associated with this progradation.

The succeeding fossiliferous wackestones with favositids clearly indicate deepening on the shelf and backstepping of the shelf margin. There is little biostratigraphic control available for this interval. It may coincide with the early Eifelian Chotec event (transgressive event) observed at Oberbuchach (cf. WALLISER, 1990). Dark-greenish to grey shales were deposited in this time interval. At Mount Cellon and Mount Freikofel or other successions of the proximal slope facies this interval has to date not been identified. This is partly due to the lack of detailed biostratigraphic control and partly to the uniform style of sedimentation.

The overlying crinoidal calcarenites and bioclastic calcirudites probably belonging to the crinoid-cortoid facies of KREUTZER (1990), who interpreted them as back reef or subtidal shelf deposits, suggest progradation again. They are succeeded by peloidal calcarenites and finally birdseye limestones and *Amphipora* limestones suggesting restriction probably associated with the buildup of a rimmed platform margin. The abundance of reefal debris in the upper slope succession supports this interpretation.

The sections reviewed reveal similarities in pattern that suggest widespread allocyclic controls. Moreover, event and sequence stratigraphy of CA sections, particularly those representing medial to distal slope facies (e.g. Oberbuchach road cut), show striking similaries of pattern to coeval Devonian sections of the northern Appalachian Basin (NAB) in eastern Laurentia (especially New York State and Pennsylvania) correlated with conodont biostratigraphy (C. BRETT, unpubl.).

All sections reveal evidence for a period of shallowing in the late Emsian to earliest Eifelian patulus-partitus Zones. In the distal-medial slope sections this event is marked by the appearance of grey crinoid-bearing carbonates that overlie red nodular deeper water carbonates of the earlier Emsian. In the medial to distal slope facies in the Carnic Alps these beds are followed by dark, argillaceous limestones and dark grey shales in the early Eifelian partitus-costatus Zones. The presence of dark organic rich bands near the base of the costatus Zone may be a local representation of the Chotec event, which has been recognized in the Pragian Basin and elsewhere.

The different sections show consistent changes that reflect the development of the Carnic carbonate platform in the Lower Devonian. Several sequence boundaries can be identified. Allocyclic patterns reflecting eustatic sea level changes and other global events are best documented in distal slope sections whereas margin architecture can be best deduced from the proximal slope and carbonate platform settings.

Carboniferous

According to SCHÖNLAUB et al. (1991) in the Carnic and Karawanken Alps the vertical range of the Variscan limestone successions varies considerably. Some end close to the Frasnian/Famennian boundary, others in the middle or upper Famennian, and others range within different levels of the Lower Carboniferous. Yet, at some localities the uppermost beds have yielded diagnostic conodonts and ammonoids of the *anchoralis latus* conodont Zone, thus indicating an age at the Tournaisian/Visean boundary. Recently, a slightly younger age has been inferred from additional sections from the Italian side of the Carnic Alps, west of Plöckenpaß, which provided a "post-*Scaliognathus*" conodont fauna corresponding to the Pericyclus IIγ Stage of the uppermost Tournaisian or lowermost Visean Stage of the Lower Carboniferous (SCHÖNLAUB & KREUTZER, 1993; PERRI & SPALLETTA, 1998a,b; SPALLETTA & PERRI, 1998).

The nature of the transition from the above mentioned limestones to the overlying siliciclastics of the Hochwipfel Formation raised a long lasting controversy about the significance of tectonic events in the Lower Carboniferous.

Apparently, this has been settled after recognition of a wide variety of distinct paleokarst features in the Karawanken and the Carnic Alps (TESSENSOHN, 1974; SCHÖNLAUB et al., 1991). The paleokarst was caused by a drop in sea-level during the Tournaisian. Rise of sea-level and/or collapse of the basin promoted the transgression of the Hochwipfel Formation which presumably started in the Lower Visean.

Based on its characteristic lithology and sedimentology TESSENSOHN (1971, 1983), SPALLETTA et al. (1980), AMEROM et al. (1984), SPALLETTA & VENTURINI (1988) and others interpreted the 600 to more than 1000 m thick Hochwipfel Formation as a Variscan flysch sequence. In modern terminology the Kulm deposits indicate a Variscan active plate margin in a collisional regime following the extensional tectonics during the Devonian and Lower Carboniferous Periods. The main lithology comprises arenaceous to pelitic turbidites and other types of mass flow sediments. In addition to these lithologies, along the northern margin of the region up to 10 m thick plant-bearing sandstone beds (Middle Visean to Namurian age [AMEROM et al., 1984; AMEROM & SCHÖNLAUB, 1992]) consitute a prominent member of the Hochwipfel Formation. Except for trace fossils the paleontological evidence of the flysch sediment is very poor. Other stratigraphic data are derived from the fore-mentioned underlying limestone beds and locally occurring intercalations of limestone clasts with stratigraphically important fossils such as the coral *Hexaphyllia mirabilis*, the algae *Pseudodonezella tenuissima*, the foraminfera *Howchinia bradyana* and early fusulinids. These clasts were supplied from a shelf-like source area located originally to the north of the present Southern Alps but which was completely destroyed by later tectonic events.

According to LÄUFER et al. (1993) the volcaniclastites and basic volcanics of the Dimon Formation occur at the base of the Hochwipfel Formation and not as its lateral equivalents or as a succeeding event. They represent intraplate alkalibasalts indicating the climax of the rifting immediately before the onset of the deposition of the Hochwipfel Formation.

In the Southern Alps the Variscan orogeny reached the climax between the Late Namurian and the Late Westphalian Stages. This time corresponds to the interval from the Early Bashkirian to the Middle or Late Moscovian Stages. According to KAHLER (1983) the oldest post-Variscan transgressive sediments are Late Middle Carboniferous in age and, more precisely, correspond to the *Fusulinella bocki* Zone of the Upper Miatchkovo Substage of the Moscovian Stage of the Moscow Basin (for more details see KRAINER, 1992). In particular between Stranig Alm and Lake Zollner they rest with a spectacular angular unconformity upon strongly deformed basement rocks including the Hochwipfel Formation the Silurian-Devonian Bischofalm Formation and different Devonian limestones. This basal part named the Waidegg Formation consists mainly of basal conglomerates, disorganised pebbly siltstones and arenaceous and silty shales with thin limestone intercalations. Even meter-sized limestone boulders reworked from the basement were recognized at the base of the transgressive sequence (FENNINGER et al., 1976) and which was named Malinfier Horizon by Italian geologists (VENTURINI, 1990).

The lower part of the Bombaso Formation south of Naßfeld, i.e., the Pramollo Member, has also long been regarded as the base of the Auernig Group in this area (VENTURINI et al., 1982; VENTURINI, 1990). Based on new field evidence, however, for this member a clear relationship with the Variscan Hochwipfel Formation is suggested.

In the Naßfeld region the transgressive molasse-type cover comprises the 600 to 800 m thick fossiliferous Auernig Group. Although the oldest part may well correspond to the late Moscovian Stage (PASINI, 1963) the majority of sediments belong to the Kasimovian and Ghzelian Stages. Based on rich fusulinid evidence from the Schulterkofel section west of Rattendorf Alm the Carboniferous/Permian Boundary has recently been drawn by the first appearance of the genera *Pseudoschwagerina* and *Occidentoschwagerina* in the upper part of the Lower Pseudoschwagerina Limestone and not at its base as previously suggested (KAHLER & KRAINER, 1993).

Permian

In the Lower Permian the Auernig Group is succeeded by a series of more than 1000 m thick shelf and shelf edge limestones and clastics (KRAINER, 1992, 1993; FORKE, 1995). They characterize a differentially subsiding carbonate platform and outer shelf setting which were affected by transgressive-regressive cycles from the Westphalian to the Artinskian Stages. This cyclicity may be explained as the response to the continental glaciation in the Southern Hemisphere (KRAINER, 1991; SAMANKASSOU, 1997).

Upper Permian sediments rest disconformably upon the marine Lower Permian or its equivalents, and farther west, on the Ordovician Val Visdende Formation and quartzphyllites of the Variscan basement. They indicate a transgressive sequence starting with the Gröden Formation and followed by the Bellerophon Formation of Late Permian age (BOECKELMANN, 1991, HOLSER et al., 1991, KRAINER, 1993).

4. Austria's Paleozoic Corals: a Brief Review

According to the classification of two major Paleozoic regions (i.e. the Upper Austroalpine Variscan sequences and the Southern Alpine sequences; cf. chapter 1), we also may distinguish an "Austroalpine Coral Fauna" (ACF) and the "Southalpine Coral Fauna" (SCF). Occurrences of both, the ACF and the SCF are restricted to certain locations within these regions. Since all Alpine Paleozoic units were affected by the Variscan or Alpine orogenies - or even by both - major sections of the successions suffered deformation and alteration (tectonic fracturing, dolomitisation, recrystallisation, etc.) thereby destroying the fossil content. The recent distribution of corals obviously does not reflect the original biofacial pattern of dispersion. Especially in the Greywacke Zone and the Gurktal Nappe System corals are rare.

Within Austria's territory Lower Paleozoic corals are frequent in the Carnic Alps and in the Graz Paleozoic. Corals of Upper Paleozoic age are restricted to occurrences in the eastern Greywacke Zone (Lower Carboniferous), Nötsch (Lower Carboniferous), Carnic Alps and Karawanken Mountains (Upper Carboniferous and Lower Permian) (fig. 27).

A review of more than 200 articles, that taxonomically deal with or cite Paleozoic corals in Austria (including coral sites near the border in Italy and Slovenia), lists 220 rugose and 113 tabulate taxa known (FLÜGEL & HUBMANN, 1994, HUBMANN, 1995, 2002). Amomg them 125 taxa (81 Rugosa, 33 Tabulates and 11 Heliolitids) on species level and 16 taxa on subspecies level (12 Rugosa, 4 Tabulates) were described for the first time from Austrian outcrops.

This data base is, however, limited by the need of modern revisional work for some locations, the loss of certain typoids during World War II (e.g. the CHARLESWORTH collection), etc. Nevertheless, the database allows an insight into the diversity of Paleozoic corals of the Alpine region.

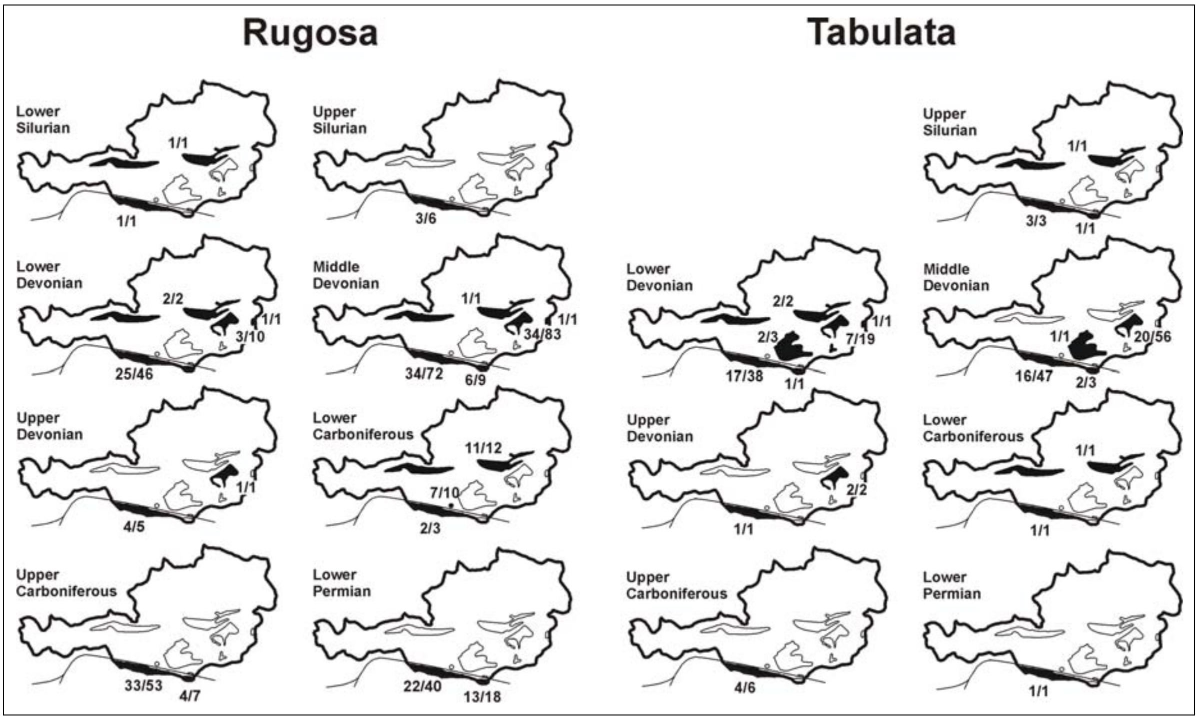

Fig. 27: Variation of numbers of coral genera and species (including taxa with nomenclatura aperta) recorded from the main fossil-bearing Paleozoic regions in the Alps (from HUBMANN, 2002).

A list of all Austrian coral taxa cited in the literature is given in the appendix.

5. Field Trip Stops

Graz Paleozoic: Stops 1-6

Stop 1: Look-out tower "Fürstenwarte" at Plabutsch hill

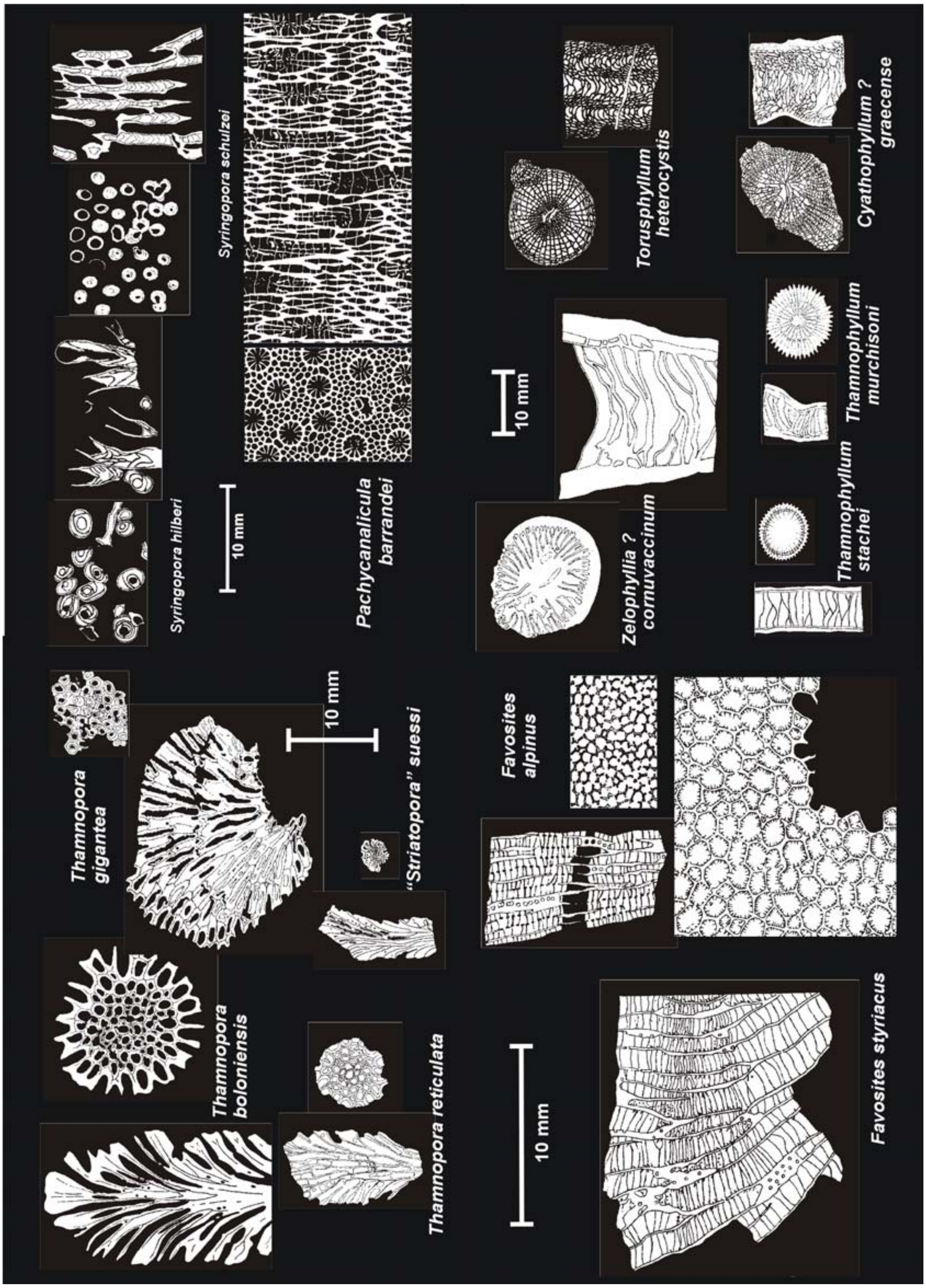

Fig. 28: "Sections" of the most important coral taxa of the Barrandei limestone (Middle Devonian, Eifelian) as they can be seen in the rocks at the Fürstenwarte.
Modified after PENECKE, 1894, and HUBMANN, 1991, 1997.

The Plabutsch (alt. 754 m) is the highest hill within the city limits of Graz. Its coral-bearing strata have been known since 1843 when Franz UNGER - a paleobotanist! - described *Gorgonia infundibuliformis* GOLDF., *Stromatopora concentrica* GOLDF., *Heliopora interstincta* BRONN (*Astraea porosa* GOLDF.), *Cyathophyllum explanatum* GOLDF., *Cyathophyllum turbinatum* GOLDF., *Cyathophyllum hexagonum* GOLDF., *Cyathophyllum caespitosum* GOLDF., *Calamopora polymorpha* a. var. *tuberosa* GOLDF., *Calamopora polymorpha* b. var. *ramoso - divaricata* GOLDF., *Calamopora spongites* a. var. *tuberosa* GOLDF. and *Calamopora spongites* b. var. *ramosa* GOLDF. from here. Although these determinations are only of historic interest, UNGER was the first who recognized the Devonian age of the strata.

The wall of the tower is made up by stones containing most of the fossils known from the Plabutsch Fm. (formerly "Barrandei Limestones" [HUBMANN, 2003]) (see fig. 28).

Stop 2: Abandoned quarry "Marmorbruch"

During the 19th century huge amounts of brickstones were exploited in the Plabutsch area. More than 30 quarries provided the material which was needed for house-building activities during the great expansion of the city at that time. Especially the dark bluish to black limestones of the "Plabutsch Formation" which are very rich in fossils were used for socles of many buildings of the Graz city (e.g. Mausoleum, Palais Prokesch).

These quarries were abandoned at the end of the 19th or at the beginning of the last century and are now covered by vegetation.

Most of the middle Devonian corals of the Graz Paleozic described by PENECKE and HERITSCH were collected in the "Marmorbruch" quarry.

For an interpretation of the depositional environment of the Plabutsch Fm. confer to the figure and text below (fig. 29).

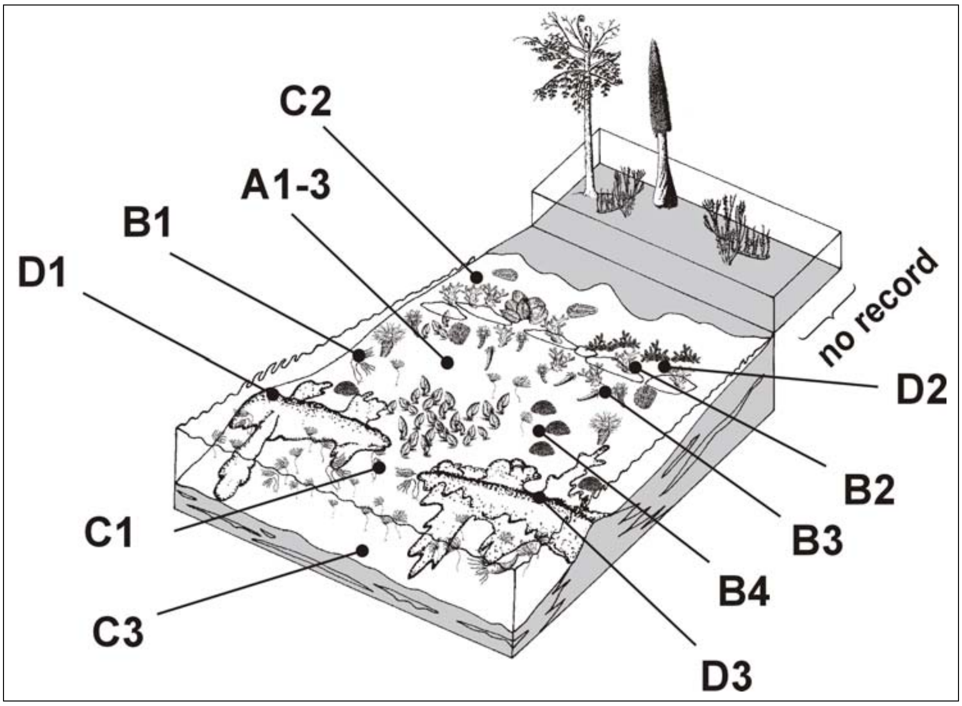

Fig. 29: Facies model of the Plabutsch Formation. After HUBMANN, 1993, 1995; EBNER et al., 2000).

A1-3 Low-energy mud facies: Micritic (to microsparitic) rocks with locally dominant fecal pellets and bioturbation structures. High mud content, as well as totally preserved, easily disarticulated skeletons (i.e.: articulated crinoid stems) suggest a low turbulence hydrodynamic regime.

A1 Mudstone-subfacies: Light grey to blue, yellowish, usually dark grey to black (finely distributed Pyrite and/or organic substance), few fossils.

A2 Calcisphere-wackestone-subfacies: Dark blue to black, micritic limestones with concentrations of calcispheres (and ?Spicula) and biogens/biomorpha of small size. Calcispheres and the rarity of macrofossils indicate deviations from a normal marine environment.

A3 Gastropod-pellet-wacke/grainstone-subfacies: Usually small (size up to 3 mm), trochospiral gastropods with apex upwardly oriented in pelmicritic to pel[pseudo]sparitic matrix. "Fecal pellets" frequently elongate due to pressure in the still unconsolidated sediment. Frequent bioturbation.

B Higher energetic mud facies: Bigger (allochthonous) biogens indicate higher hydrodynamic energy setting during deposition. General lack of rounding indicates short transport (parautochthonous to autochthonous). "Typical reef-building" organisms are characteristic.

B1 Crinoid-Brachiopod-wacke/floatstone-subfacies: Layers with isolated Crinoid stems and thin-shelled brachiopods, frequently with micritic envelops are widely distributed. Brachiopods (usually thin-shelled; Chonetids?) usually double-valved.

B2 "Amphiporid"/"Thamnoporid"-floatstone-subfazies: Accumulations of branching Tabulata of *Thamno-pora* und *Striatopora*-type with dendroid stromatoporoids of *Amphipora*-type with calcispheres and crinoids.

B3 Coral-stromatoporoid-floatstone-subfazies: Rugose, *Thamnophyllum*-dominated and tabulate corals with branching growth-form, as well as lamellar or tabular stromatoporoids in general parallel to layers. Also echinoderms, brachiopods, gastropods und broken shells. Frequent epoecic stromatoporoids on rugose (rarely on tabulate) corals. Orientation indicates directed currents, lack of abrasion indicates short transport.

B4 Brachiopod-coral-floatstone-subfacies: Characterized by thick-valved brachiopods and massive (*Favo-sites*, *Alveolites*, *Heliolites*) as well as dendroid corals (*Thamnophyllum*, *Thamnopora*, *Striatopora*). May be developed as "*Zdimir*-Schill". Lack of imbrication and frequent double-valved brachiopods indicate that at least brachiopods are autochthonous. The corals have no indication for live position.

C High-energy debris facies: Rounded, oriented components, graded fossil debris characteristic.

C1 Crinoid debris-subfacies: Echinoderms-, peloids- and gastropods in sparitic limestones frequently at the bases. Well rounded and sorted biogens with frequent micritic envelopes.

C2 Coarse silt-pellet-subfacies: Besides a high contribution by coarse silt (grain size from 60-125 μ, up to 78%) in micritic to microsparitic or pseudosparitic matrix, also pellets, in particular quartz-silt-grains-rhythmites. Bioturbate structures in planar to wavy laminated hanging-wall.

C3 Eventstone(tempestite)-subfacies: Erosional base, which is not always evident, with following shell debris and biogendebris layers, usually normal gradation. In two-valved organisms the ratio stable to unstable position is about 1:1, geopetal fillings characteristic. "Muddying-upward"-sequences; above the biogen-debris-layers frequently mudstones without fossils.

D "Reef" facies: This facies unites genetically different organismal carpets (algae, stromatopores, corals; autoparabiostromes, cf. HUBMANN [1995: 111]), as well as "coverstones".

D1 "Coverstone"-subfazies: The "coverstone" facies according to TSIEN [1984] characterizes initial reef growth, but is also similar to the tempestite facies. Macroscopic allochthonous components are characteristic, they are covered by autochthonous lamellar organisms. Crinoids, dendroid tabulata, rugosa, heliolitida, brachiopoda and gastropoda are found as detritic (allochthonous) components. Lamellar and tabular stromatoporoids (type *Actinostroma*), as well as favositides with lamellar Corallum act as stabilizers.

D2 Algae wacke/float to bafflestone subfacies: Halimedacean lawns are classified according to MAMET et al. (1984) as "algae-baffle/boundstones". This facies is found in alternation with red marly shales (HUBMANN, 1990, 2000).
This subfacies has residue values which are far elevated above values given in the literature for algal limestones.

D3 Coral-baffle(frame)stone-subfacies: Only in level 7 of the Attems outcrop, where massive, wave-resistant Favositid patches of 0.5 m diameter are found.

Stop 3: Forest road "Attems"

The Plabutsch Formation (see Stops 1, 2) represents a highly fossiliferous sequence whose boundaries are not clearly identifiable till now. Locally the sequence may range from Upper Emsian to Lower Givetian. Generally, four types of microfacies and thirteen types of submicrofacies have been recognized. The comparison of WILSON's types of microfacies with the Plabutsch Fm. suggests that the limestones were deposited in restricted (lagoonal), semirestricted and open platform environments and on the platform margin and foreslope.

The whole section may be roughly subdivided into 5 biofacial parts (fig. 30):

a) Siliciclastic Brachiopod-Trilobite-Biofacies ("Chonetenschiefer") with: *Chonetes* sp., *Maladaia* sp., and Crinoids;
b) Coral-Stromatoporoid-Biofacies with: *Actinostroma sp., Thamnophyllum stachei, Thamnophyllum murchisoni, Favosites styriacus, Thamnopora* sp., *Striatopora* sp., *Pachycanalicula barrandei, Heliolites* cf. *peneckei*, Crinoids;
c) Coral-Brachiopod-Biofacies with *Thamnophyllum stachei, Thamnophyllum murchisoni, Thamnopora reticulata* ?, *Thamnopora* sp., *Striatopora* (?) *suessi*, *Favosites* sp., *Chonetes* sp., "Spiriferids", Crinoids;
d) Algae-Biofacies with *Pseudopalaeoporella lummatonensis*, *Pseudolitanaia graecensis*;
e) Brachiopod-Coral-Biofacies with: *Zdimir* cf. *hercynicus, Thamnopora* cf. *reticulata, Striatopora* (?) *suessi.*

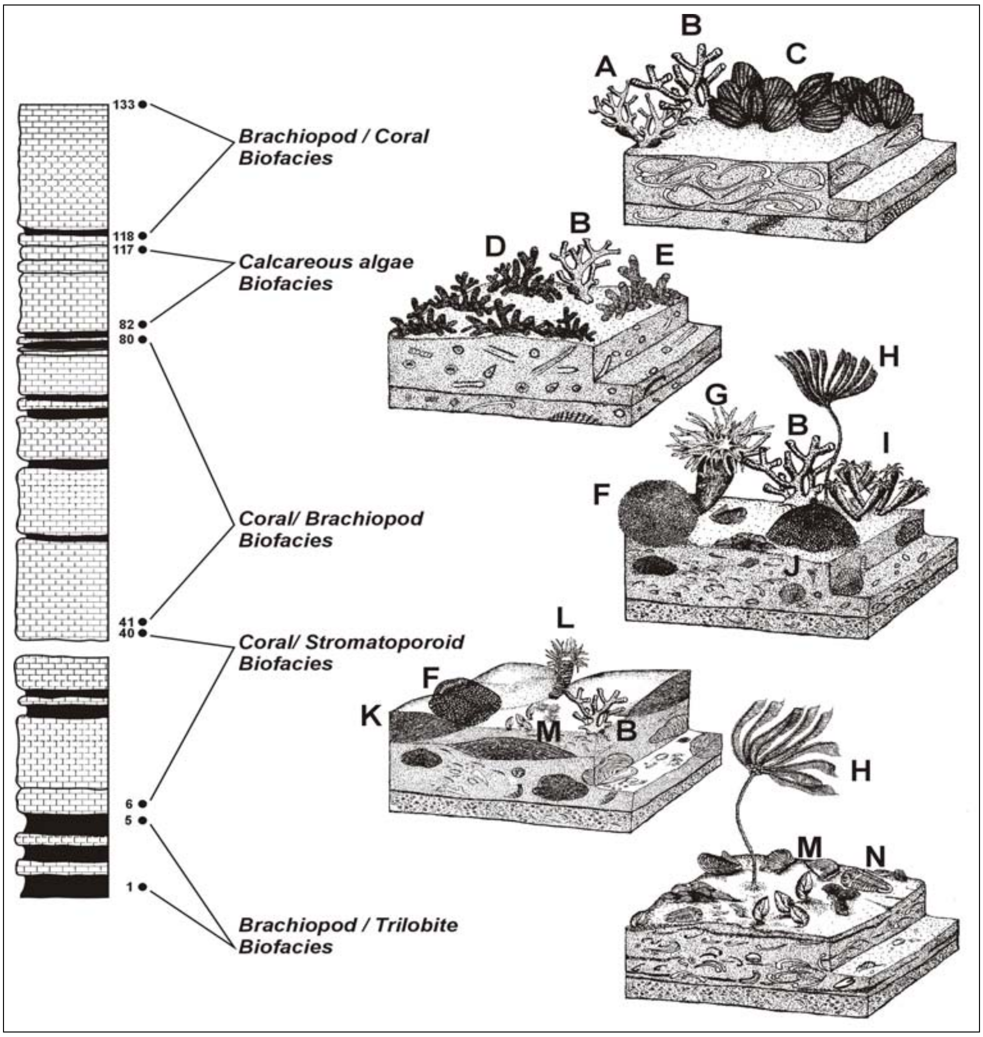

Fig. 30: Simplified distribution of the five biofacies types in the Attems outcrop. Numbers indicate bank sequence.

Stop 4: "Weiße Wand"

Overlying the Plabutsch Fm. a sequence with dolomites and micritic limestones which locally may contain biostromes (Rannach area) and patch-reefs (St. Pankrazen area) is developed. In the Rannach Nappe this sequence is called Kollerkogel Fm., whereas time-equivalent strata in the Hochlantsch Nappe are comprised under Tyrnaueralm Fm. and Hochlantsch Fm. (FLÜGEL, 2000).

The basal parts of the Kollerkogel Fm. consist of biolaminated dolostones (Gaisbergsattel Mb.). Overlying limestones (Kanzel Mb., "Kanzel Limestone") of the varcus Zone locally start with *Amphipora* or *Stachyodes* meadows (fig. 31).

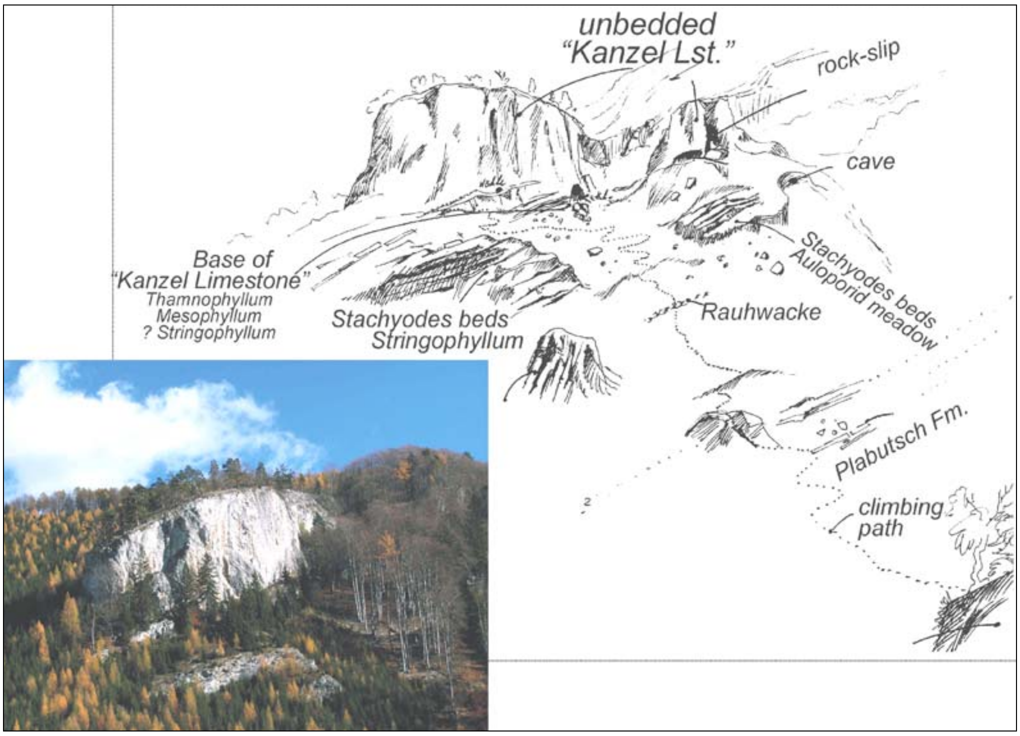

Fig. 31: "Weiße Wand" outcrop.

Stop 5: Teichalm area: "Zechnerhube"

In the Hochlantsch Nappe the Givetian is characterized by a sequence of silty sandstones, laminated dolomites, limestones and volcanites (Tyrnaueralm Fm.). At the southern slope of the Tyrnaueralm a section through the Tyrnaueralm Fm. exhibits in the lower part of the profile alternating peritidal dolomites and sandstones which pass into subtidal limestones (GOLLNER & ZIER, 1985). The latter may contain tempestite layers with coral debris. The source area for this fossil debris probably is represented in the "Zechnerhube" area on the northern slope of the Tyrnaueralm where a coral-biostrome [with *Sociophyllum longiseptatum*, *Cyathophyllum* (*Cyathophyllum*) *dianthus*, *Mesophyllum (Cystiphylloides) caespitosum*, *Zelophyllia cornuvaccinum*, *Alveolites* sp., *Coenites* sp., *Favosites* cf. *styriacus*, *F. alpinus*, *Thamnopora* sp., auloporids, *Heliolites* (very rare) and *Stachoydes* and *Clathrocoilona* (KRAMMER 2001)] is developed.

Stop 6: Hochlantsch: "Zirbisegger"

In the northern part of the Hochlantsch appr. 30 km north of Graz the final "bioconstructions" of the Graz Paleozoic are developed within the Zachenspitz Formation. This Upper Givetian (probably continuing to the Lower Frasnian) formation locally contains in its lower parts several Amphipora Beds grading into biohermal Argutastrea-Alveolitid-Stromatoporoid baffle-to-boundstones exposed at the northern slopes of the Hochlantsch mountain. Due to the pelagic fauna of the interherm facies (dominated by tentaculites) GOLLNER & ZIER (1985) proposed a shallow offshore depositional setting for the formation.

Nötsch Carboniferous: Stops 7-9

Stop 7: Quarry "Jakomini" in the Nötsch-Gorge (fig. 32)

Visit of the Badstub Breccia interbedded by a fossiliferous shaly member ("Zwischenschiefer").

The Badstub Breccia comprises a succession of breccias and conglomerates with amphibolite, granite, gneiss quartzite and carbonate pebbles and intercalated sand- and siltstones. The latter are rather immature and mainly consist of detrital hornblende and fine grained clasts of amphibolites which derive from metamorphic tholeiitic ocean floor (KRAINER & MOGESSIE, 1991). The same material builds up the matrix of the breccias and conglomerates and gives them a "basaltic" appearance. The massive mafic character of the black to green coloured rocks entailed some striking differences in the interpretation of their genesis, i.e. diabase (FRECH [1894]; cf. name of the quarry enterprise!), tectonic breccia (ANGEL, 1932), sedimentary breccia (FELSER, 1936), volcanic breccia (KIESLINGER, 1956), tholeiitic basalt (TEICH, 1982). The very early known fossiliferous sandstone intercalations, especially the horizons with huge brachiopods (*Gigantoproductus*) supported a sedimentary origin of the "Badstub breccia" (cf. SCHÖNLAUB, 1985; KRAINER, 1992).

Fig. 32: Detail of the Quarry "Jakomini" in the Nötsch Gorge exposing the Badstub Breccia and the interbedded Shale Member.

The fossil content (brachiopods, crinoids, plant remains) as well as sedimentary structures (debris flows, subordinate turbidites) indicate a marine depositional environment. KRAINER (1992, 1997) interprets the poor textural and compositional maturity of the conglomerates and the presence of non-disarticulated fossils as being far less transported sediments on small fans or fan-like sediment bodies along an active fault zone at the slope.

Limestone pebbles containing conodonts point to an Upper Visean age (SCHÖNLAUB, 1985) and therefore to an age of the Badstub Breccia not younger than this date, perhaps uppermost Visean.

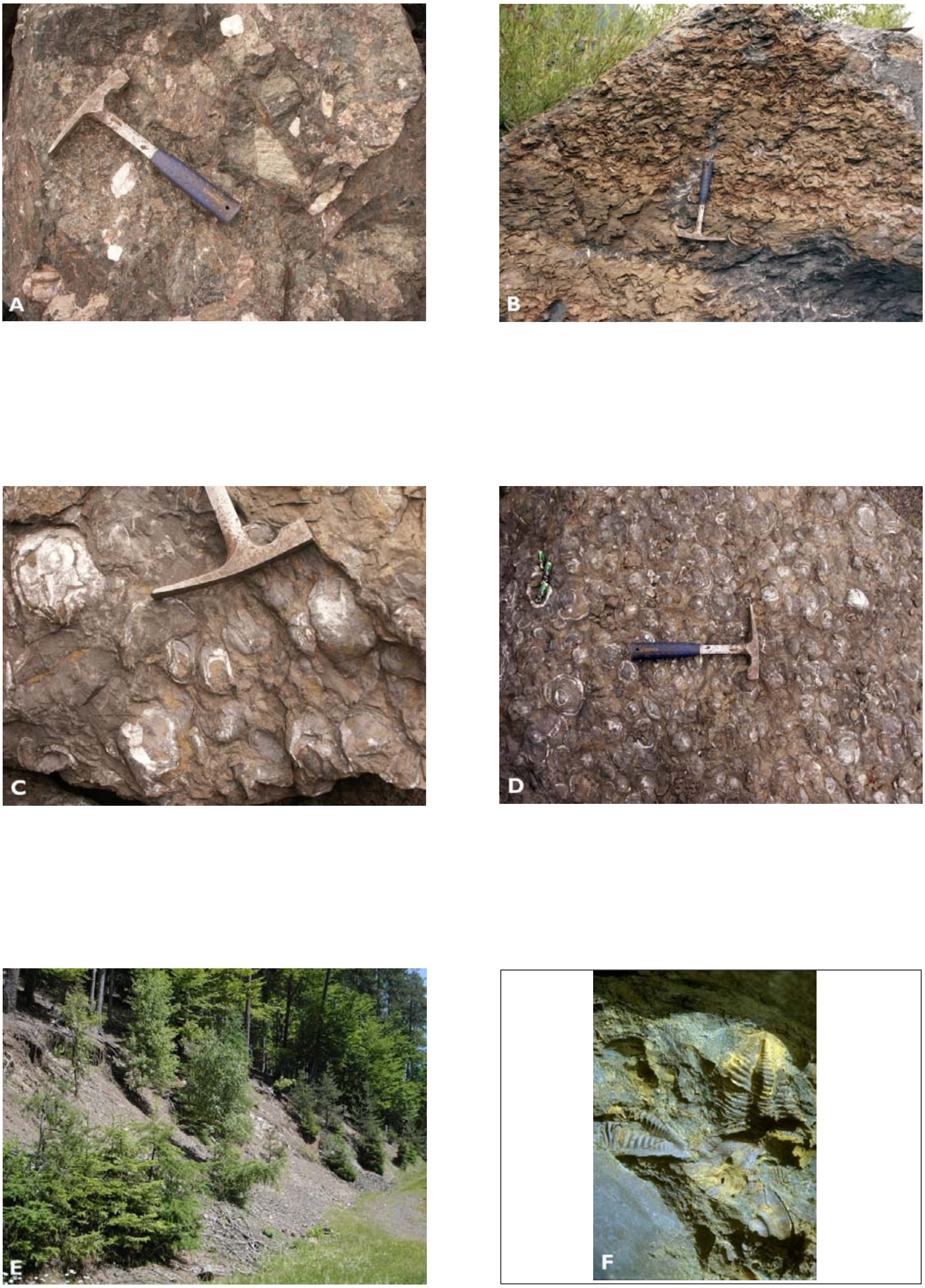

Plate 1: Carboniferous of Nötsch.

A: Badstub Breccia; B, C, D: Brachiopod bearing Shale Member within the Badstub Breccia (*Giganto-productus* sp.); E: Outcrop Oberhöher; F: Trilobite remains from outcrop Oberhöher.

Stop 8: Roadcut to village Hermsberg (optional)

Section through the upper part of the Erlachgraben Formation with locally rich occurrences of brachiopods (among others *Gigantoproductus giganteus*).

Stop 9: Forest road Wertschach - Badstube (loc. O3 at alt. 1080 m) (fig. 33)

Lowermost part of Nötsch Formation charaterized by a highly abundant and diverse fauna and partly also flora preserved as shelly fossils, molds or as "Steinkern".

The Nötsch Formation comprises a sequence of shales and siltstones alternating with sandstones and conglomerates showing typical features of sediments caused by turbidity currents. Especially the fine-grained clastics may show on some places a rich and highly diversified fauna and mega-flora (e.g. AIGNER, 1929; AMEROM & SCHÖNLAUB, 1992; FLÜGEL, 1972; SCHÖNLAUB, 1985; SCHRAUT, 1996a,b, 1999; YOCHELSON & SCHÖNLAUB, 1993 and others). The main constituent of the formation are grey sandy shales supplying the most prominent fossil localities, e.g. bridge at crossing the Nötsch brook at Point 721 and the vicinity of the Fischerhube. Most of the "classical" outcrops mentioned in the literature are lost because of the easy weathering of the sediments and/or urbanisation.

Some hundred meters NW of the farm "Fischerhube" (formerly Oberhöher or Oberhecher) on the southern slope of Badstuben hill a recently built forest road exhibits in a sharp turn rocks with a high fossil content.

Fig. 33: Locality O3 at forest road at an altitude of 1080 m southeast of mountain Badstube.

The dark brown shales and siltstones belong to the lower part of the Nötsch Formation. Especially the dark slightly carbonatic shales contain brachiopods, bivalves, gastropods, crinoidal stems, bryozoans, trilobites, corals and sometimes plant remains.

Due to the foliation no primary sedimentary structures are visible thus precluding an interpretation of the depositional environment.

Carnic Alps: Stops 10-19

Stop 10: Cellon Section

The famous section is located between 1480 and 1560 m on the eastern side of the Cellon mountain, SSW of Kötschach-Mauthen and close to the Austrian/Italian border. lt can be reached within a 15 minutes walk from Plöckenpaß.

The Ordovician to Lower Devonian part of the Cellon section is best exposed in a narrow gorge cut by avalanches (fig. 34). Thus, the German name for the section is "Cellonetta Lawinenrinne".

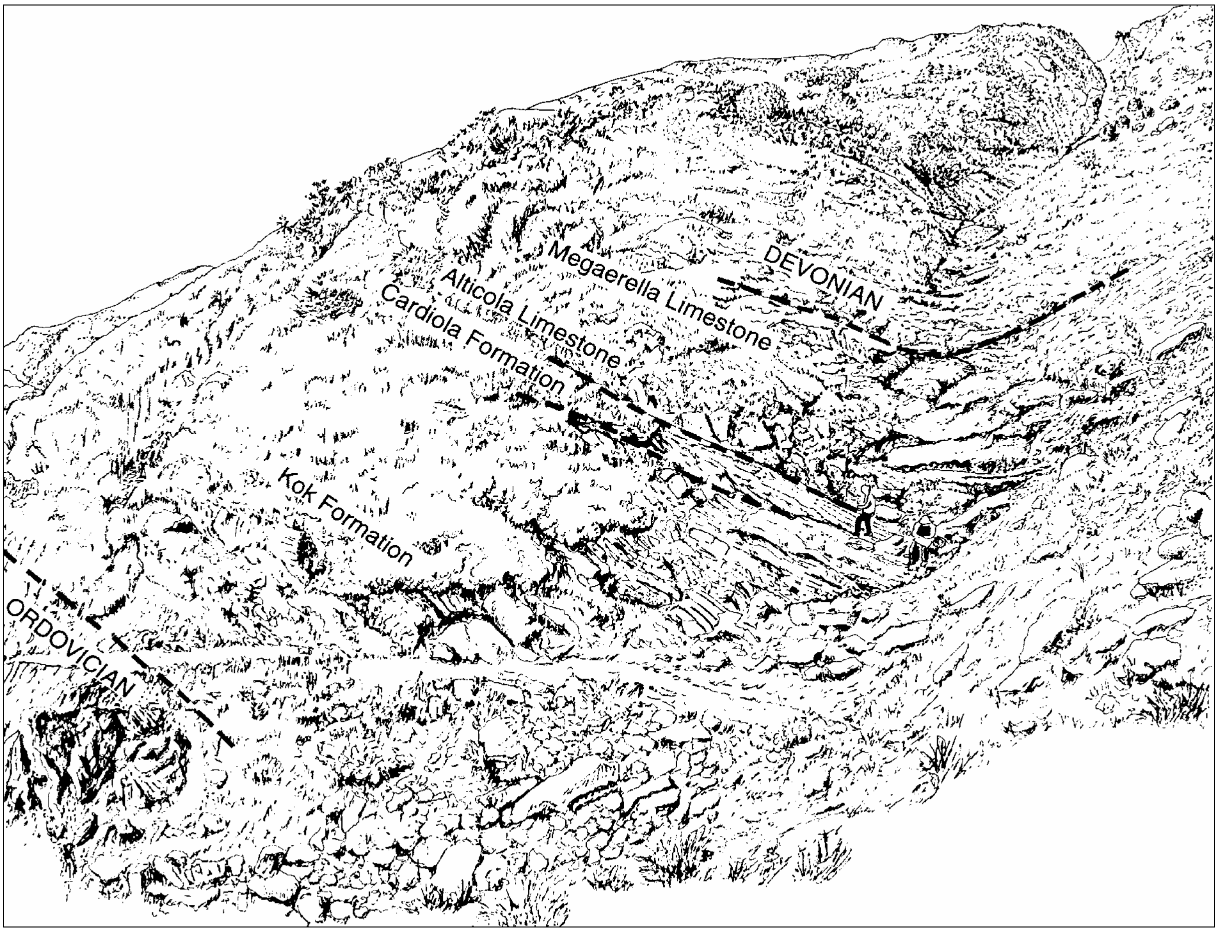

Fig. 34: View of the Cellon section (from HISTON et al., 1999).

Stratigraphy

The Cellon section represents the stratotype for the Silurian of the Eastern and Southern Alps. Nowhere else in the Alps has a comparably good section been found. lt has been famous since 1894 when GEYER first described the rock sequence. In 1903 it was presented to the 9th IGC which was held in Vienna. According to v. GAERTNER (1931) who studied the fossils and rocks in great detail, the 60 m thick continuously exposed Upper Ordovician to Lower Devonian section could be subdivided into several formations. Since WALLISER's pioneering study on conodonts in 1964 it still serves as a standard for the worldwide applicable conodont zonation which, however, has been further detailed and partly revised in other areas during the last two decades. Although the conformable sequence, corresponding to the Plöcken Facies, suggests continuity from the Ordovician to the Devonian, in recent years several small gaps in sedimentation have been recognized which reflect eustatic sea-level changes in an overall shallow-water environment. From top to base the following formations can be recognized (fig. 35):

Lochkovian	80.0 m	Rauchkofel Limestone (dark, platy limestone)
Silurian	8.0 m	Megaerella Limestone (greyish and in part fossiliferous limestone; equivalent to the Pridoli Series)
	20.0 m	Alticola Limestone (grey and pink nautiloid bearing limestone; Ludlow to Pridoli)
	3.5 m	Cardiola Formation (alternating black limestone, marl and shale; Ludlow)
	13.0 m	Kok Formation (brownish ferruginous nautiloid limestone, at the base alternating with shales; Late Llandovery to Wenlock)
Ordovician	5.4 m	Plöcken Formation (calcareous sandstone; Ashgill, Hirnantian Stage)
	6.5 m	Uggwa Limestone (argillaceous limestone grading into greenish siltstone above; Ashgill)
	>50 m	greenish and greyish shales and siltstones (Caradoc to Ashgill)

According to SCHÖNLAUB (1985) the Ordovician/Silurian boundary is drawn between the Plöcken and the Kok Formations, i.e. between sample nos. 8 and 9. In the Plöcken Fm. index fossils of Hirnantian age (brachiopods, trilobites, conodonts) clearly indicate a latest Ordovician age (JAEGER et al., 1975; FERRETTI & SCHÖNLAUB, 2001; SCHÖNLAUB & SHEEHAN, 2003). These strata represent the onset of the end-Ordovician - Lower Silurian transgressive cycle known from many places in the world (SCHÖNLAUB, 1988).

According to conodonts and graptolites from the basal part of the overlying Kok Fm. the equivalence of at least six graptolite and two conodont zones are missing in the Lower Silurian. Renewed sedimentation started in the late Llandovery within the range of the index conodont *P. celloni.*

At present the precise level of the Llandovery/Wenlock boundary can not be drawn. Graptolites and conodonts, however, indicate that this boundary should be placed between levels nos. 11 and 12. Consequently, the rock thickness corresponding to the Llandovery Series does not exceed some three meters.

According to SCHÖNLAUB in KRIZ et al. (1993) the boundary between the Wenlock and the Ludlow Series can be drawn in the shales between sample nos. 15 B1 and 15 B2. Apparently, this level most closely corresponds to the stratotype at quarry Pitch Coppice near Ludlow, England. We thus can assume an overall thickness of some 5 m for Wenlockian sedimentation. By comparison with the Bohemian sections the strata equivalent to the range of *Ozarkodina bohemica* are at Cellon extremely condensed suggesting that during the Homerian Stage sedimentation occurred mainly during the lower part. With regard to the foregoing Sheinwoodian Stage it may be concluded that at its base the corresponding strata are also missing or represented as the thin shaly interval between sample nos. 12 A and 12 C. At this horizon the *M. rigidus* Zone clearly indicates a late Sheinwoodian age.

By correlation with Bohemian sequences and the occurrence of index graptolites for the base of the Pridoli, the Ludlow/Pridoli boundary is drawn a few cm above sample No. 32 (SCHÖNLAUB in KRIZ et al., 1986). This horizon lies some 8 m above the base of the Alticola Lst. The corresponding sediments of the Ludlow have thus a thickness of 16.45 m.

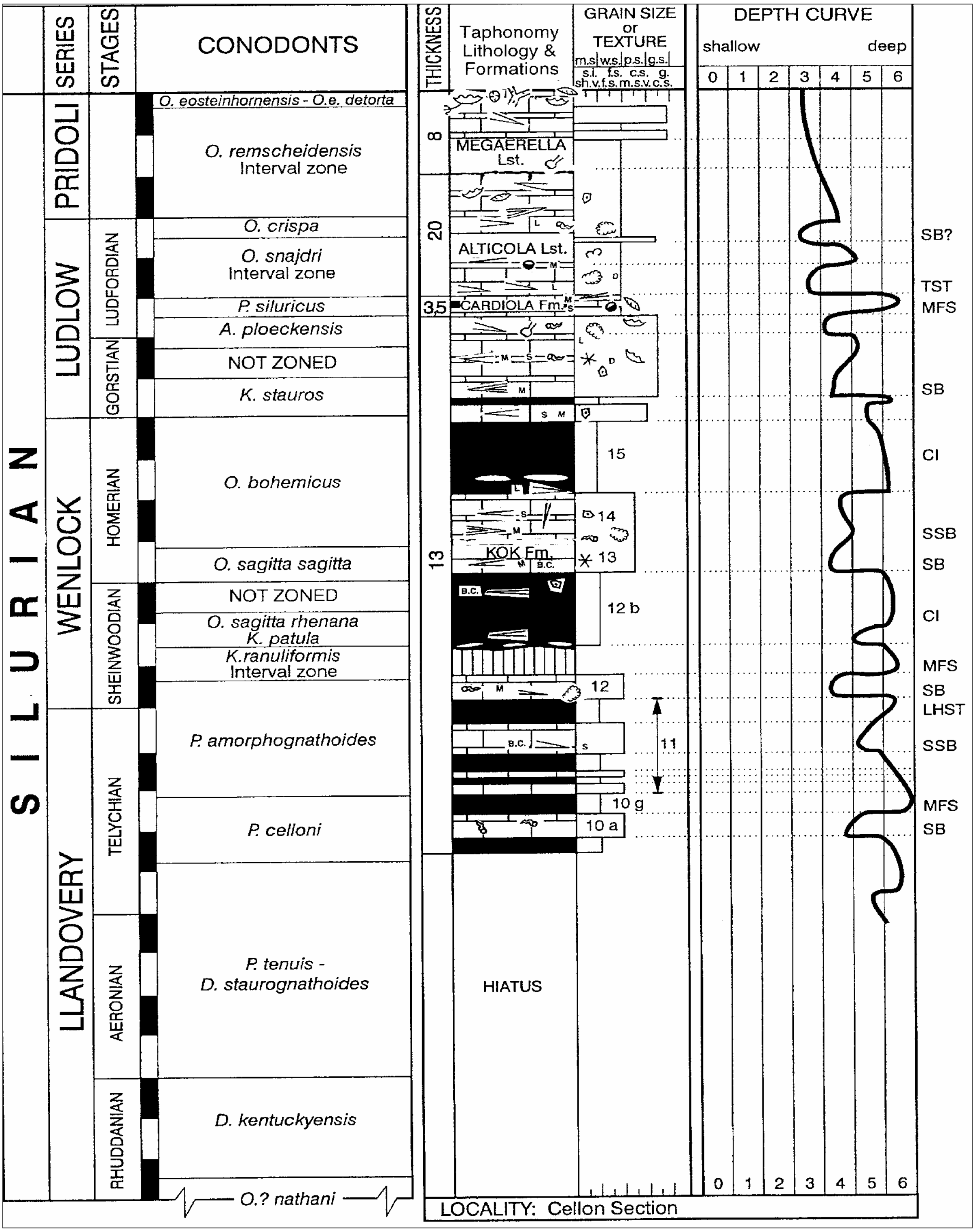

Fig. 35: Conodont stratigraphy, lithology, grain size, significant taphonomic features and depth curve of the Silurian of the Cellon section. Modified from SCHÖNLAUB, 1997.

At Cellon the Silurian/Devonian boundary is placed at the bedding plane between conodont sample nos. 47 A and 47 B at which the first representatives of the index conodont *Icriodus woschmidti* occur. It must be emphasized, however, that the first occurrences of diagnostic graptolites of the Lochkovian is approx. 1.5 m higher in the sequence. JAEGER (1975) recorded the lowermost occurrences of *M. uniformis, M.* cf. *microdon* and *Linograptus posthumus* in sample no. 50. The Pridoli may thus represent a total thickness of some 20 m.

Lithology and Microbiofacies

The first facial investigation at the Cellon section was carried out by FLÜGEL (1965). BANDEL (1972) studied the facies development of the Lower and Middle Devonian in the central part of the Carnic Alps. Middle and Upper Devonian and Lower Carboniferous strata (exposed as steep cliffs and on top of Cellon) were investigated by KREUTZER (1990). Photomicrographs from the Ordovician to Lower Carboniferous sequences comprising the whole Cellon section were published by KREUTZER (1992b) and a preliminary study of the Silurian was given by KREUTZER (KREUTZER & SCHÖNLAUB, 1994). Current work on the cephalopod limestone biofacies in the Carnic Alps with regard to the paleogeographical setting during the Silurian has highlighted many interesting microfacial aspects of the predominantly calcareous sequence.

Uggwa Limestone (Beds 1-4)

This up to 6 m thick limestone horizon comprises indistinctly bedded grey to coloured pelagic "flaser" limestones with ostracod, cystoid and bryozoan debris layers. Sceletal grains consist of brachiopods, ostracods, bryozoans, agglutinate forams, rare cephalopods, trilobites, conodonts and acritarchs.

Plöcken Formation (Beds 6-8)

The 5.40 m thick Plöcken Fm. comprises impure coarse-grained limestones in the lower part consisting of skeletal grains of echinoderms, brachiopods, ostracods, trilobites and conodonts. These grainstones grade into calcareous sandstones and siltstones and shales.

Kok Formation (Beds 9-20) (fig. 36)

Bed 9: Base of the Silurian sequence: At the level of the Ordovician/Silurian boundary the transition from the greenish silts - shales of the Plöcken Fm. to the carbonate sequence of the Upper Llandovery is marked by the occurrence of flattened nodules approximately 3-5 cm in diameter which appear to be micritic, dark grey-black in colour, quite dense and showing iron weathering: The overlying shales and carbonate layers are badly deteriorated: Fossil content not apparent.

Bed 10: Again a series of shales and thin carbonate beds: level E is the best preserved and shows trace fossil features at the base and the first development of "crust" like shales otherwise fossil content not apparent although a trilobite fauna has been described from this level.

Bed 11: The base is marked by micritic lenses or nodules with "crusts". The overlying shales have a crinoid, trilobite and brachiopod fauna towards the top of the sequence: The first occurrence of nautiloids is at the base of the Wenlock with levels of alternating shales and of reddish-grey micritic carbonate levels which have upper and lower crusts. There is a nautiloid fauna both in the shales and limestones. The shales show flow features around the lenses and the nautiloids are enclosed within the shales. They are small to medium in dimension with an abundance of medium nautiloids towards the top of the sequence. They are parallel to bedding with both body chamber and apexes preserved and have an outer oxidised coating only in the carbonates. A change may be noted up the sequence in that the nautiloids become relatively more abundant in the carbonate levels whereas previously they were more abundant in the shales.

General remarks: Thin beds of ferruginous limestone, sometimes bioturbated, are intercalated at the base of the Kok Formation in dark shales locally rich in small brachiopods. At the top of bed 12, a thin and lenticular calcareous horizon (12b) in shales has provided an important cardiolid fauna (KRIZ, 1999). This is a cephalopod wackestone the matrix of which bears many ostracods, echino-derms, rare small bivalves and gastropods. Many muellerisphaerida are present in darker bituminous micritic areas.

Starting from bed 13, the limestone becomes thicker and more massive. The reddish colour and the intensive bioturbation are the most typical features of the upper part of the Kok Formation up to level 17. This cephalopod wackestone is locally rich in brachiopods, echinoderm debris, trilobites, gastropods and ostracods. Some organisms, mostly cephalopods, reveal peculiar iron-banded coatings. Dolomitization is frequent. Around level 15 B1 a singular grainstone of well sorted equidimensional bioclasts occurs which strongly resembles the coeval horizon of the Rauchofel Boden Section. Abundant small thin-shelled bivalves, preserving the two valves still connected, gastropods, trilobites and isolated echinoderm ossicles have iron-stained shells. Shell in shell structures are common there. Starting from around bed 18 the limestone becomes greyer. Pyrite aggregates in the matrix may be occasionally found.

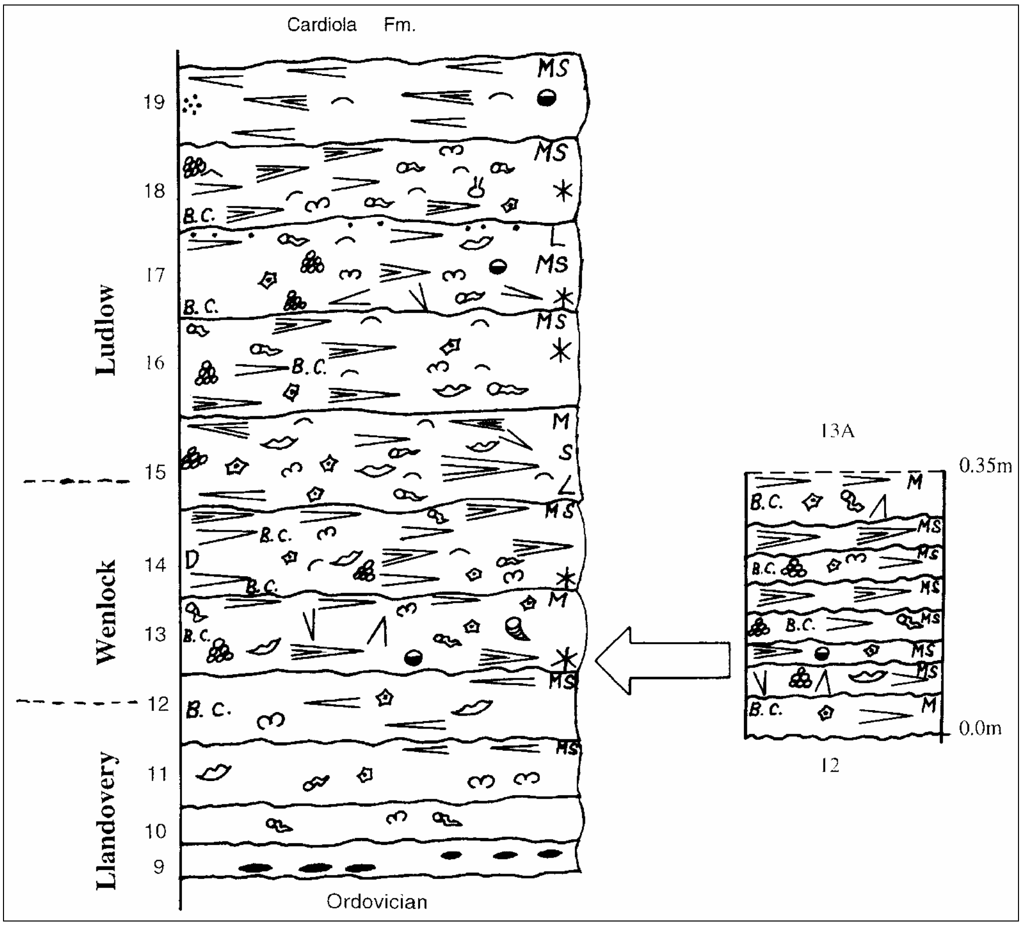

Fig. 36: Taphonomy of the Kok Formation. Note detail of small scale cyclic repetition of beds indicating changes in the hydrodynamic regime.

Cardiola Formation (Beds 21-24)

It is represented by bioclastic shelly layers a few centimeters thick (wackestone-packstone) with a sharp base interbedded with dark shales. At the base of the Cardiola Formation (level 21) bioclastic wackestones rich in cephalopods, trilobites, crinoids and ostracods are intercalated in soft micritic sediments. Scouring traces at the top of the soft sediments, debris grainstone at the base of the overlying horizon with enrichment in iron and manganese oxides would exhibit, according to KREUTZER (1992b), the existence of a Fe-Mn hard-ground. Millimetric pavements of small brachiopods are present in bed 22. When seen in thin-section, they reveal a cephalopod-ostracod bioclastic packstone with abundant brachiopods, but also associated with graptolites, thin-shelled bivalves and micritized grains. Shelter porosity, common orientation of geopetal structures and telescoping of cephalopods have been observed. Sorting is moderate. These shelly laminae decrease in thickness towards the top of the formation and alternate with thin dark bands rich in organic matter and muellerisphaerida, possible ostracods and recrystallized cephalopods.

Alticola Limestone (Beds 25-39)

The Alticola Limestone (Ludlow-Pridoli in age) is distinctive in that it forms the base of the steep slope of the section. The erosive base of the grey dolomitised massive beds contrast sharply with the black shales of the underlying Cardiola Fm. and this reflects an easily recognizable greyish to reddish limestone formation. It has an overall thickness of 20 m and represents a transgressive carbonate series within more stable pelagic conditions (SCHÖNLAUB, 1997). Grey to dark pink limestones represented mainly by a bioclastic packstone with fine-grained micritic matrix with a variety of bed thickness and frequent stylolites are common in the Ludlow with a dominant nautiloid fauna. The beds decrease in thickness in the Pridoli and alternate with interbedded laminated micrites with a dominant nautiloid and brachiopod fauna. Several deepening events marked by the development of black shales have been documented within the uppermost levels of the Pridoli. Cephalopods are abundant, together with crinoids, trilobites, large gastropods and ostracods. Iron-coatings, mostly around trilobites, are again present. Bioturbation is common.

Megaerella Limestone (Beds 40-47 A)

The Megaerella Limestone (Pridoli in age) comprises the upper Pridoli and Silurian/Devonian boundary transgressive sequences of carbonates rich in biodetritus, lenticular micrites and black shales. It has a thickness of 8 m and forms the steep step at the top of the section. Light grey limestones (wackestone to packstones) with cephalopods, ostracods, echinoderm debris and trilobites are dominant. A particular level of juvenile nautiloids (RISTEDT, 1968) occur in bed 40. Bryozoans (*Fenestella* s.l. sp. and a small indeterminate cryptostome [Wyse JACKSON, pers. comm.]) occur on a distinct bedding plane above the Silurian/Devonian boundary together with bivalves. Complete specimens of *Scyphocrinites* (HAUDE, pers. comm.), solitary corals and articulated cridoid stems are common in the lower beds of the Lochkov.

According to KREUTZER (1994) the bathymetric environment for the Upper Ordovician to Devonian sequence can be described as follows:

As early as in the Ordovician a facial differentiation can be recognized for the carbonates. The Cellon section with its Uggwa Limestone development (sample 1-5) represents the late Ordovician Uggwa Facies which is time-equivalent to the Wolayer Limestone of the Himmelberg Facies exposed, e.g., at the Rauchkofel-Boden section. Based on conodonts the Uggwa Limestone is well dated as being Ashgillian in age. According to DULLO (1992), the two formations represent the near-shore parautochthonous cystoid facies (Wolayer Limestone) and an off-shore basinal debris facies (Uggwa Limestone), respectively.

At the end of the Ordovician in the Carnic Alps a regression occurred. The Uggwa Limestone bed nos. 1-4 characterized by pelagic faunal elements, are followed by limestones composed of subtidal components of the Plöcken Formation (bed nos. 5-8). A significant unconformity separates the Plöcken Fm. from the overlying Kok Fm.

Transgression of the Kok Formation started in the Cellon section in the Upper Llandovery (bed no. 9). In contrast to the Cellon section the Rauchkofel section located some 8 km to the northwest exhibits a considerably reduced sequence. At Cellon the basal Silurian succession represents a moderately shallow environment which may have lasted until the Llandovery/Wenlock boundary or until the very beginning of the Wenlock. Sample 11 exhibits a very shallow to intertidal environment. During the remaining part of the Wenlock a transgressive trend can be recognized. However, at the Wenlock/Ludlow boundary (bed nos. 15A-F) some strata may be missing reflecting either submersion or reduced sedimentation.

During deposition of the Cardiola Formation (bed nos. 21-24) contemporary non-deposition may have occurred. Black limestone and shale beds with radiolarians alternate with pelagic limestone beds indicating an offshore environment. The following Alticola Limestone (bed nos. 25-39) reflects stable conditions in a pelagic environment which terminated in a regressive pulse (bed no. 40). With the beginning of the Megaerella Limestone (nos. 41-47A) a further transgressive trend can be inferred.

Starting in the Lochkovian Stage (bed 47B and >; Rauchkofel Limestone) and ranging to the Upper gigas Zone of Frasnian age (top region of the Cellon cliff) the Devonian transitional facies represents a fore-reef facies. While this slope facies accumulated at Cellon, only a few kilometers to the palinspastic SSW (today seen at the Kellerwand region) more than 1000 meters of Devonian shallow-water limestones were deposited. Moreover, coeval carbonates of pelagic origin, i.e. pelagic limestone facies of the Rauchkofel Nappe, with a markedly reduced thickness of not more than 100 meters were deposited within short distances to the NNE (SCHÖNLAUB, 1979, 1985; KREUTZER, 1990, 1992a, b).

During the crepida Zone of the Famennian a short-lasting regression occurred. In the Upper Famennian and Lower Carboniferous uniform cephalopod limestones were deposited (Pal and Kronhof Limestone, respectively). At the beginning of the Visean the flysch of the Hochwipfel Formation transgressed upon the Kronhof Limestone and limestone deposition ended.

In more detail the Devonian to Lower Carboniferous succession is subdivided into the following formations (KREUTZER, 1992). It represents the transitional facies between the southwestern shallow-water realm and the eastern to northeastern deep-water settting:

80 m	well-bedded pelagic Rauchkofel Lst.: dark grey and black plate limestones with occasional organodetritic interbeds (Lochkov);
120-150 m	Kellerwand Lst.: well-bedded yellowish tentaculite limestones alternating with skeletal debris layers (Pragian to Lower Emsian);
120 m	Vinz Lst.: well-bedded dark grey platy limestone interbedded with detritic layers (Emsian);
150-200 m	Cellon Lst.: grey massive limestone beds composed of pelagic biogenes, bioclasts and debris layers (Eifelian-Givetian);
50-100 m	Pal Lst.: greyish to reddish and also pinkish cephalopod limestone (Frasnian to Famennian);
1-3 m	Kronhof Lst.: greyish to reddish cephalopod limestone (Tournaisian).

A short distance to the west of the peak of Cellon at the famous Grüne Schneid section the Devonian/Carboniferous boundary beds are excellently exposed. The detailed distribution of conodonts, goniatites and trilobites as well as the lithology and major and trace element content was recently studied by an international working group (see SCHÖNLAUB et al., 1992).

The Ordovician-Silurian Boundary Event (from SCHÖNLAUB & SHEEHAN, 2003)

The mass extinction at the end of the Ordovician led to the disappearance of about 100 families, which represented about 22% of all marine families (SEPKOSKI, 1982, 1993). This demise affected mainly trilobites, brachiopods, echinoderms, stromatoporoids, corals, bryozoans, ostracods, bivalves, cephalopods, graptolites, conodonts, chitinozoa and acritarchs. In comparison with the great dying at the P/T boundary, this was the second largest catastrophe in the Phanerozoic.

During the last years a number of explanations have been suggested for this extinction, such as the species/area-effect as a negative consequence of the global ice age with an associated regression (BRENCHLEY, 1995; HARPER & JIA-YU, 1995; OWEN & ROBERTSON, 1995), increased sedimentation with related pollution due to a significant sea-level decrease (WYATT, 1995), or changes in the composition of sea water and precipitates (ORTH et al., 1986; WILDE et al., 1986; MELCHIN et al., 1991; GOODFELLOW et al., 1992; WANG et al., 1992, 1993; LONG, 1993). Furthermore, in a few areas, such as South China, the Canadian Arctic, on Anticosti Island, and in South Scotland (Dob's Linn stratotype area), elevated Ir contents were found in boundary layers. However, they have been interpreted to be of terrestrial origin (WANG et al., 1995).

The global ice age at the end of the Ordovician is unusual because it occurred at a time when the atmosphere had increased CO_2 contents and, thus, the Earth should have had a relatively stable greenhouse climate (BRENCHLEY et al., 1994). At that time the atmospheric CO_2 content was supposedly 14 to 16 times higher than today (BERNER, 1990, 1992, 1994; CROWLEY & BAUM, 1991; GRAHAM et al., 1995; MORA et al., 1996).

New studies seem to have clarified the long-standing question regarding the exact timing of the end-Ordovician mass extinction. According to SUTCLIFFE et al. (2001) and SHEEHAN (2001) this event occurred at the base of the Hirnantian Stage of the Upper Ordovician with the beginning of the graptolite Zone of *Normalograptus extraordinarius*. At that time the ratios of the stable isotopes of C, O and S changed, which was explained as a result of biomass reduction, temperature increase, and a short-term flooding of the continental shelf with anoxic water (GOODFELLOW et al., 1992). The latter supposedly was the ultimate cause of the mass extinction. However, an opposite trend was observed for the Hirnantian Stage in middle Sweden by MARSHALL & MIDDLETON (1990).

Based on recent stratigraphic and geochemical studies, the following scenario has been proposed for the Upper Ordovician: Beginning with the so-called gracilis-transgression of the older Caradoc Series, the global sea level increased until the end of the Ordovician (ROSS & ROSS, 1992). This second-order cycle was, however, overprinted by a regressive-transgressive trend in the late Ashgill Series lasting for about 0.5 to 1 million years, i.e. during the early to middle Hirnantian Stage. This trend was caused by glacio-eustatic changes and led to a sea level dropping of about 100 meters (BRENCHLEY & NEWELL, 1980; SHEEHAN, 1988, 2001; BRENCHLEY & MARSHALL, 1999; BRENCHLEY et al., 1991, 1994). The glaciation extended over an area of approx. 30 million square kilometers of the southern hemisphere. It resulted in a significant decrease in the average temperature, as well as in an increased productivity due to increased oceanic circulation. The latter is evident from anomalies in the oxygen and carbon isotope ratios of marine carbonates and in the organic carbon reservoir. This deterioration of the climate and the overall changes of the nutrient supply was held responsible for the first pulse of the twofold mass extinction. After the faunal demise the opportunistic Hirnantia Fauna started to invade middle and higher latitudes. This fauna reached its highest diversity within the *N. extraordinarius* Zone.

At the base of the graptolite Zone of *N. persculptus*, both isotope curves show an opposite trend (MARSHALL et al., 1994; BRENCHLEY et al., 1994; WANG et al., 1994). This signal is assumed to indicate the end of the glacial period, as well as a decreased production of biomass. This condition seems to have been the reason for a second mass extinction shortly before the end of the Ordovician.

Facies sequence data from Upper Ordovician deposits in the Carnic Alps also reflect a regressive-transgressive sedimentary pattern during the Hirnantian Stage (SCHÖNLAUB, 1988; SCHÖNLAUB & SHEEHAN, 2003) (fig. 37). The regressive trend is well documented in the upper part of the Uggwa Limestone and culminates in the bioclastic Plöcken Formation of the late Hirnantian Stage. Newly recovered conodonts and graptolites from this formation seem to indicate a level immediately below the Ordovician/Silurian boundary (FERRETTI & SCHÖNLAUB, 2000; SCHÖNLAUB & SHEEHAN, 2003).

In the Carnic Alps clastic sediments are the dominating lithologies during the Caradocian and early Ashgillian Series. They are succeeded by fossiliferous limestones known as the up to 20 m thick bryozoan and cystoid bearing Wolayer Lst. and the coeval 4-6 m thick Uggwa Lst., respectively. The latter represents a slightly deeper open marine setting being formed during the Rawtheyum Stage at the beginning of the upper Ashgillian.

The Uggwa Lst. is overlain by greyish and greenish siltsones with thin interbedded limestone layers. The transition from calcareous to pelitic sedimentation occurs approx. 0.40 m below the first appearance of the Hirnantia fauna characterizing the base of the succeeding Plöcken Formation. This fauna is associated with trilobites, e.g., *Mucronaspis m. mucronata* and graptolites such as *Normalograptus persculptus*.

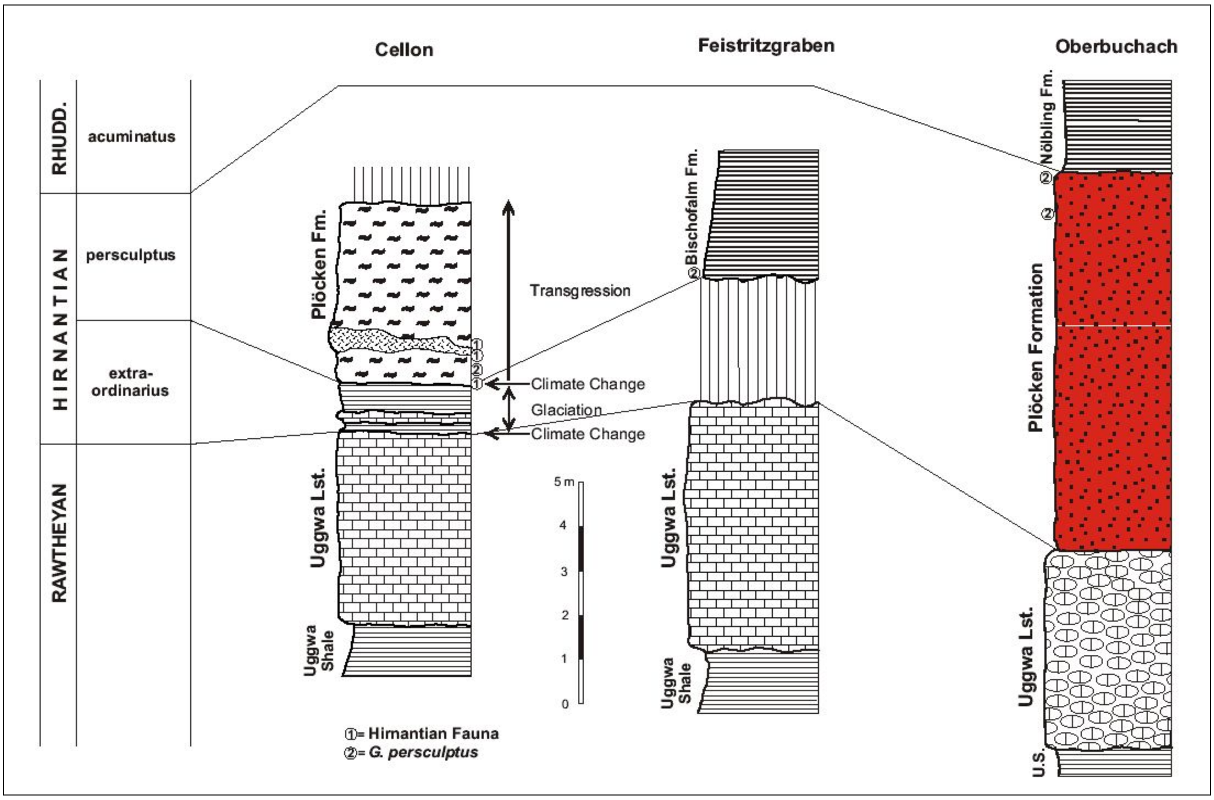

Fig. 37: The Upper Ordovician succession of the sections Cellon, Feistritz Gorge and Oberbuchach (Carnic Alps and Western Karawanken Alps) with inferred glacio-eustatic events.

The change from the Uggwa Lst. to the siltstones reflects significant environmental changes at the beginning of the Hirnantian Stage (fig. 38). Hence, the greenish pelitic shales atop the Uggwa Lst. may represent glacio-marine sediments in distal regions of the icesheet covering large parts of southern Gondwana. They are separated from the overlying transgressive Plöcken Fm. by a distinct uncon-formity.

At the Cellon section the Plöcken Fm. attains a thickness of 5.40 m. The lower 0.80 m thick portion is composed of arenaceous siltstones followed by impure limestones and calcareous sandstones with layers of bio- and lithoclasts. The fossil debris mainly represents disarticulated brachiopod shells but also bryozoans, trilobites, ostracods and conodonts are quite abundant. The whole package is strongly bioturbated, partly graded and convolute bedding and channeling occurs. This lithology suggests a storm-dominated shallow water environment which formed during the retreat of the ice in the *N. persculptus* Biozone.

Representatives of *N. persculptus* occur approx. 0.25 m above the base of the Plöcken Fm. Hence, this index graptolite testifies the upper Hirnantian Stage during which the transgression started on a global scale. Due to local tectonic uplifts, however, in the Carnic Alps a gap in sedimentation occurred at the base of the Silurian. Thus, at the Cellon section the equivalences of the lower and middle Llandovery are missing. Continued sedimentation across the passage from the Ordovician to the Silurian seems to have only occurred in the basinal black shale environment of the Bischofalm Facies.

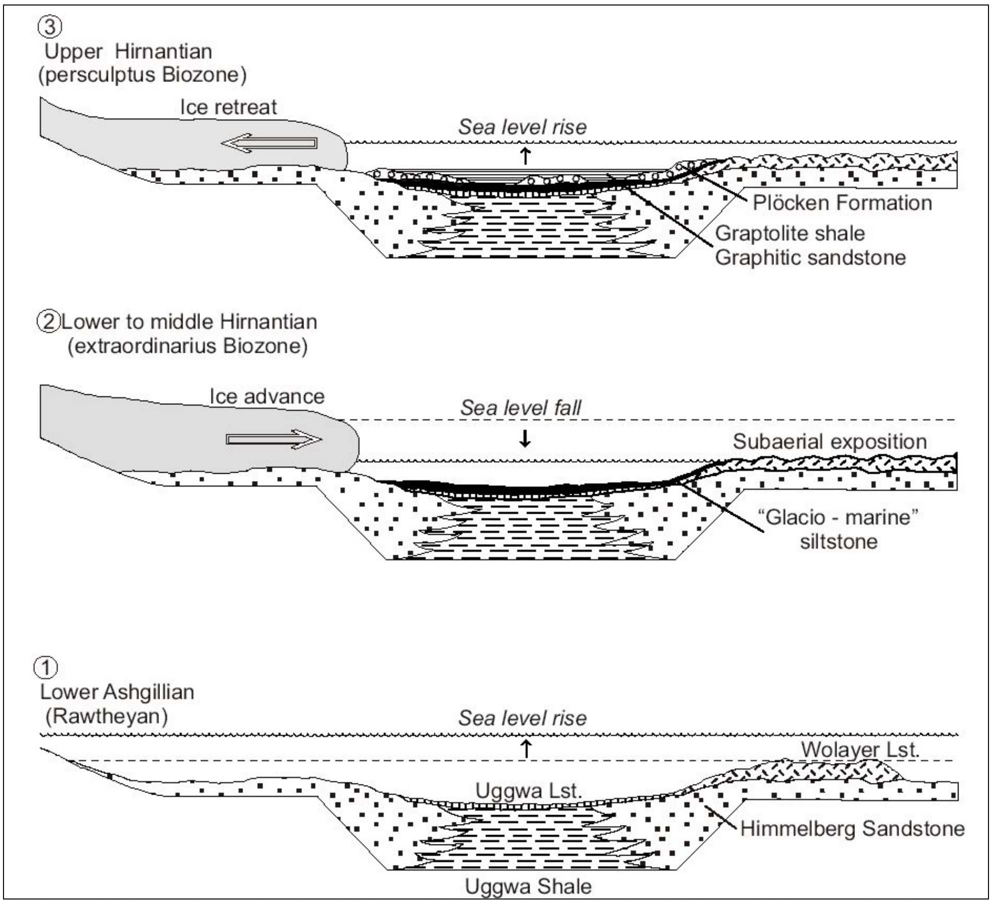

Fig. 38: Model showing the governing sedimentation pattern during the Ashgillian Series in the southern Alps.

Stop 11: Devonian succession at Mount Freikofel

Mt. Freikofel is located to the east of the Plöckenpaß (Passo di Monte Croce Carnico) and can be reached by following Trail # 403 from the Plöcken Haus (1215 m) to the trail head of trail # 401 which climbs to the top of Mt. Freikofel (Cuelat, 1757 m). The trail on the Austrian side shows good exposures of the Frasnian/Famennian succession whereas the branch on the Italian side follows an old army track and shows best exposures of the Lochkovian to Middle Devonian succession.

Note: ***The military track is not difficult to walk (it was made for mules) but it is not secured and drops off steeply to the sides. It is not recommended for those afraid of heights and great care must be taken not to dislodge stones. Good foot wear (boots) is essential.***

Mount Freikofel exhibits a spectacular section which spans almost the entire Devonian, it is easily accessible, well preserved and well exposed.

Based on lithological criteria the succession can be subdivided into five units (fig. 39):

Unit 1 Basal dark grey platy and lumpy limestones (~77 m thick).
Unit 2 Yellow-grey lumpy to nodular bedded limestones with intercalated calcarenite beds (~74 m thick).
Unit 3 Massive lithoclastic limestone with reefal debris and lithoclasts (~68 m thick).
Unit 4 Bedded lithoclastic limestone with intercalated clacarenite units, increasing bed thickness up section (~56 m thick).
Unit 5 Grey stylo-bedded fine-grained limestone and grey to pink burrow-mottled limestone with intercalated calcarenite and calcirudite beds (~36 m thick).

Fig. 39: The Devonian section at Mount Freikofel viewed from southwest.

The basal **unit 1** is exposed at the southern branch of Trail # 401 towards Rossboden Törl (Passo Cavallo).

The succession begins with dark grey fine grained wackestones and mudstones with chert nodules which are probably Lochkovian in age (Unit 1). Up section hackly weathering calcisiltites appear, intercalated with interbedded lumpy and nodular limestones. About 65 m up section the nodular limestones disappear and calcisiltites and thinly bedded wackestones and mudstones dominate the top part of the section up to 77 m. A Conodont sample from the top indicates upper Lochkovian age with *Oz. stygia*. The fauna is similar to that found at Rauchkofelboden (Bodenkalk) and section Oberbuchach (at the transition between black Rauchkofel Limestones to red Findenig Limestone).

About 50 m to the east and separated by a fault (fig. 39) follows the base of the Mt. Freikofel section with unit 2.

Unit 2 begins at the lowest accessible limestones below the military track to the top of Mount Freikofel with yellow-grey lumpy limestones intercalated with stylo-bedded and nodular limestones. The yellow stain comes from high dolomite content and associated iron. About 21 m up section the first substantial lithoclastic limestone bed occurs (1.9 m thick) with up to 8 cm long lithoclasts, some Heliotithes and rugose corals. These lithoclastic beds are spaced in 1-10 m intervals with decreasing distance and increasing thickness up section. The intercalated fine grained limestone beds become more calcareous and massive up section. Unit 2 ends at 74 m.

Three Conodont samples taken from the base of the Freikofel succession indicate *P. dehiscens* Zone, Ems. Obviously the Pragian limestone (Vinz Limestone) is cut out at the fault between lower and upper sections (fig. 39).

Unit 3 begins with 7 to 12 m thick massive lithoclastic limestone beds with a few calcarenite units separating them. The conglomeratic units are dominated by large rafts and flat pebbles of fine-grained limestone lithoclasts with subordinate numbers of bioclastic debris, namely crinoids, rugose and tabulate corals, stromatoporoids, and locally abundant *Stachyodes*. The amount of these reefal bioclasts increases up section, whereas crinoids are abundant throughout. Frequently the coarse lithoclastic units are capped with graded calcarenites suggesting deposition from waning flows. At 112 m large stromatoporoid fragments occur (up to 50 cm diameter) and very fossiliferous limestone clasts can be found in the adjacent loose debris. At 128 m there is a horizon with dark-stained (phosphoritic?) lithoclasts which could indicate the late Eifelian age (Kacak event) or the Eifelian/Givetian boundary (cf. BANDEL, 1972). At 137.8 m Unit 3 ends.

Unit 4 begins with bedded lithoclastic limestones with bed thicknesses ranging from 1 to 2 m. Flat pebble lithoclasts and reefal bioclasts are still common here but smaller in size. Conodont samples from this interval indicate basal Frasnian age with *Ancyrognathus triangularis.*
The dominantly clastic sedimentation ends at 166.9 m with finer-grained bedded calcarenites, calci-siltites, mudstones and wackestones which are, however, still intercalated with calcirudites. Unit 4 ends abruptly with a facies change to fine-grained stylo-bedded limestone at 193.6 m.

Unit 5 is characterized by 0.5-1.0 m thick beds of lithoclastic calcirudite with conspicuous absence of reefal debris. A conodont sample taken from the base of this interval indicates basal Famenne. The thickness of the calcirudite beds decreases up section and stylo-bedded mudstones and bioturbated grey and pink mottled wackestones become dominant. However, at 210.2 m coarser (up to 1.2 m thick) lithoclastic beds re-appear. The succession ends with a 1.0 m thick lithoclastic limestone unit.

Summary

The section at Mount Freikofel is clearly dominated by gravity flow deposits. Their coarseness and abundance indicates proximity to the source. The varying clast sizes and massiveness of the litho-clastic beds reflects changes in the marginal slope or ramp region. Finer-grained units presumably are more distal and indicate back-stepping of the marginal source region and/or lack of transport and/or lack of marginal buildups. The composition of the clasts with reefal debris and rafts of fine-grained limestone lithoclasts reflects a source area with reef growth and (presumably) a fore slope region with fine-grained, early lithified lime mud. The largest clast sizes of reefal and lithoclastic debris are found in the most massive beds and are Middle Devonian in age. This time interval also yields the highest amounts of reefal debris whereas the Famennian interval (predictably) yields hardly any.
The slope succession as a whole reflects the buildup of a massive carbonate platform which reached its acme in the Middle Devonian and began to regress in the Frasnian to finally collapse in the late Devonian and early Carboniferous with onset of the Variscan orogeny.
Compared to the much thicker section at Mt. Cellon the Freikofel was probably located further away from the shelf platform and represents a more distal nappe in a series of imbricated thrust slices of slope sediments.

Stop 12: Abandoned quarry near base of trail # 149 to Rifugio Marinelli

Trail # 149 branches off from the track # 148 to Rifugio Marinelli. At the base of this trail a large ridge of fossiliferous limestone is exposed and higher up the trail are large angular blocks of limestone cut from the rock walls. They provide an excellent and easily accessible exposure of Middle or upper Devonian reef limestone.

The blocks are composed of bioclastic limestone with large colonies of overturned or in situ stro-matoporoids along with various accessory reef builders such as solitary rugose corals, ramose, laminar and massive tabulate corals and much crinoidal debris. The presence of *Heliolites* suggests Middle Devonian age. However, this needs to be confirmed through thin section study.

Most stromatoporoids are massive and reach up to 70 cm in diameter but laminar, encrusting, nodular and ragged shapes are also present along with *Stachyodes* and *Amphipora*. The matrix is fine-grained micrite with a large percentage of coarse bioclastic debris.
Centimetre to decimetre sized cavities are lined with fibrous calcite.

The lithology most likely represents a southern equivalent of the Kellergrat Limestone described by KREUTZER (1992) from the Kellergrat and also present at the southern side of Hohe Warte.

Plate 2: Devonian reef limestone slabs at the abandoned quarry at trail no. 145 to Rifugio Marinelli at alt. 1520 m.

Stop 13: Outcrops in the Wolayerlake area (see fig. 19)

Stop 14-19: Post-Hercynian Succession at Naßfeld region

Principally the Post-Variscan of the Carnic Alps can be divided into three sequences (VENTURINI, 1990):

1) Permian-Carboniferous
2) Permian-Triassic
3) Middle-Triassic

Each of these sequences shows an internal transgressive trend which is interrupted by a regressional erosional surface at the top.

The main deposition area of the Permian/Carboniferous sequences is the Naßfeld basin (Pramollo basin) which extends from Forni Avoltri to Tarvisio.

The excursion deals with the central part of this basin, the Naßfeld region at the border between Austria and Italy (figs. 40, 42, 44).

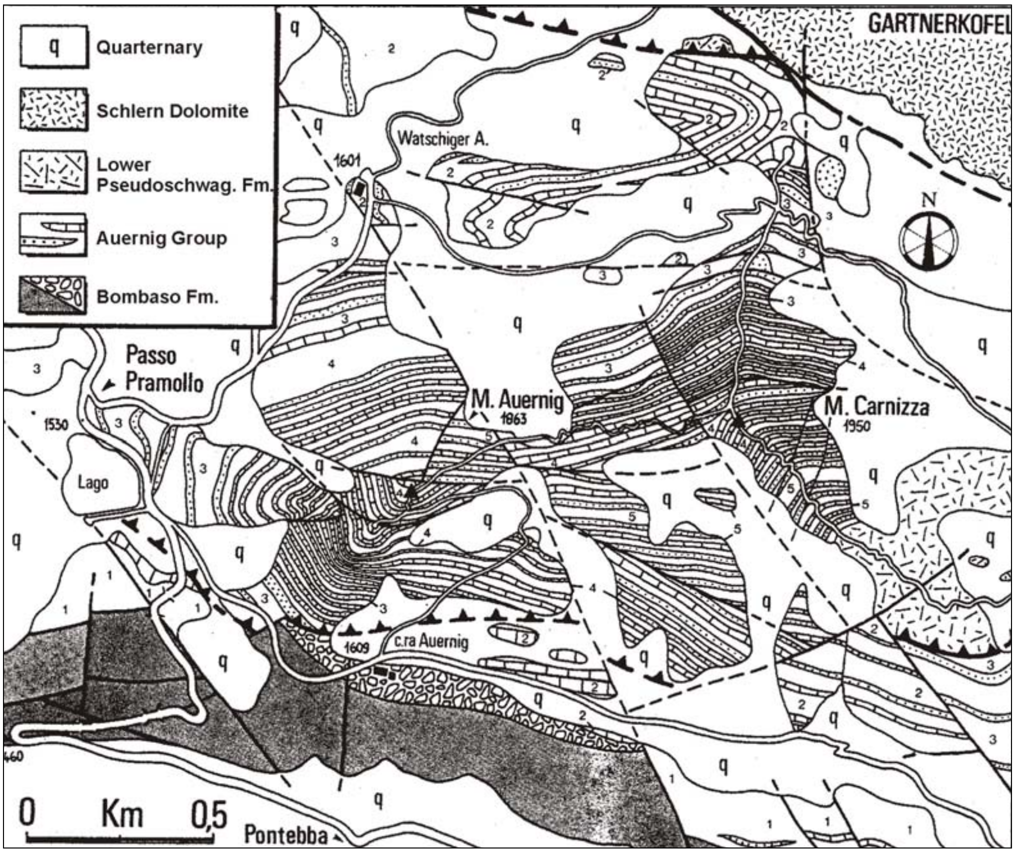

Fig. 40: Geological map of the Naßfeld region (from VENTURINI, 1990).

The Post-Variscan sequence starts in the Westfalian (Moscovian) with the Bombaso Formation and ranges to Artinskian. It is organized into a wide transgressive-regressive 1st-order sequence; mainly the Upper Carboniferous strata can be splitted into several cyclothemes, their thicknesses range from 10-40 m (VENTURINI, 1990). The Bombaso Formation is characterized by series of immature clastic sediments representing alluvial to marine environments.

The Auernig Group, especially the Corona, Auernig and Carnizza Formations are outcropping in the Naßfeld area.

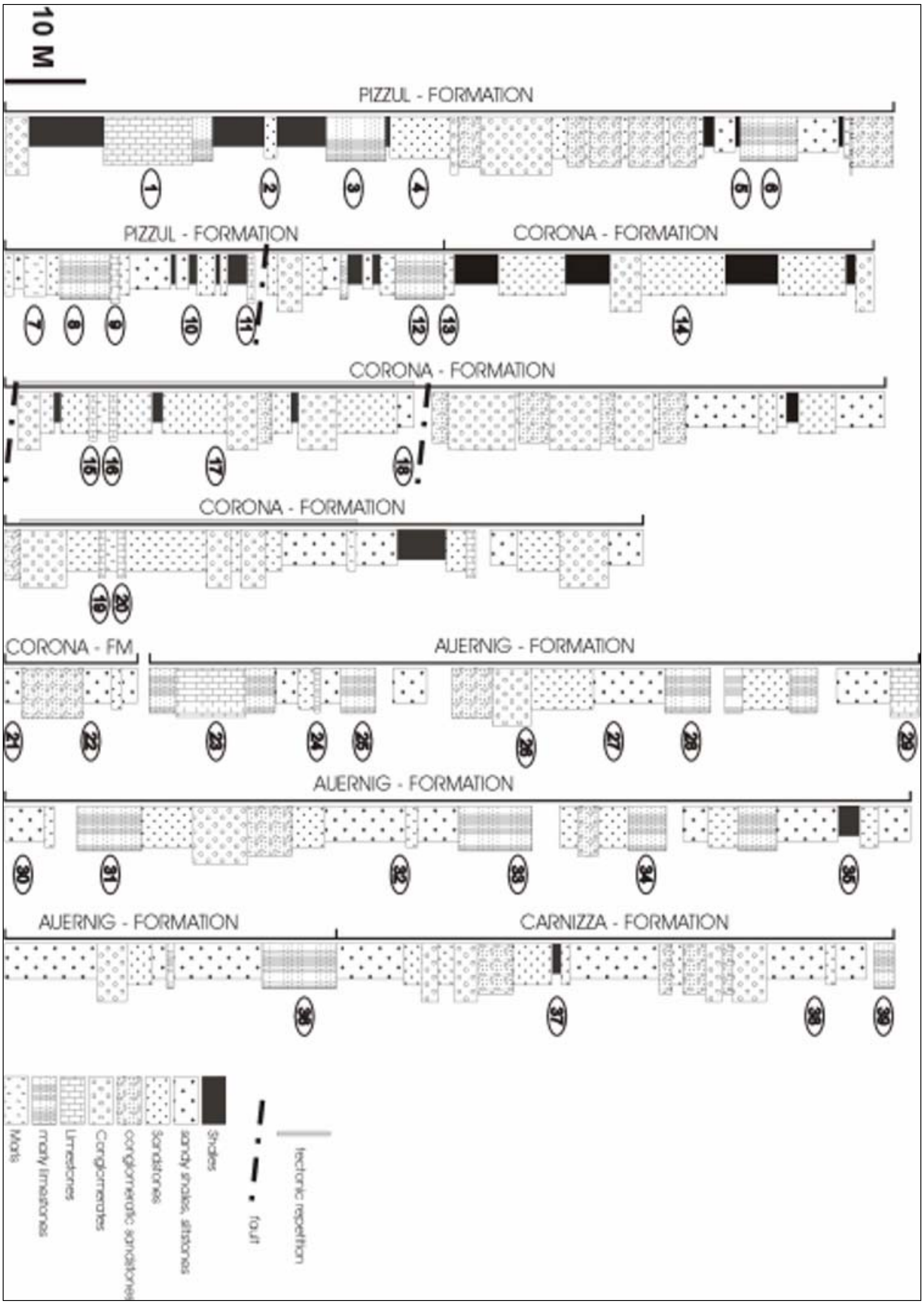

Fig. 41: Section through the Auernig Group in the Naßfeld area.
Compiled and modified after FENNINGER, unpubl., and KRAINER, 1992).

1: rugose corals (partly silicified), **2**: phytoclasts, **3**: fusulinids, **4**: partly bioturbated, **5**: algal marl, **7**: algae, fusulinids, corals (*Sinophyllum carnicum*), brachiopods, **8**: fossil debris, **9**: crinoidal debris, **10**: *Isogramma paotechowensis*, **11**: algae, **12**: fusulinids, brachiopods, corals, **13**: plants, bioturbation, **14**: plants, **15**: algae, **16**: bryozoans, fusulinids, **17**: crinoids, conularids, bivalves, **18**: plants, **19**: algae, **20**: bryozoans, fusulinids, **21**: productids, **22**: plants ("Flora Gugga 1,2"), **23**: fusulinids, algae, echinoderms, brachiopods, bryozoans, **24**: fusulinids, algae, **25**: algae, fusulinids, gastropods, brachiopods, **26**: plant fossils, **27**: plants ("Flora Gugga 3"), **28**: fusulinids, **29**: fusulinids, **30**: plant remains, **31**: algae, fusulinids, echinoderms, bryozoans, archiopods, **32**: plants ("Flora Garnitzenberg 1"), **33**: echino-derms, algae, bryozoans, fusulinids, brachiopods, **34**: algae, fusulinids, bryozoans, **35**: plant remains, bioturbation, **36**: echinoderms, algae, fusulinids, corals, **37**: plants ("Flora Garnitzenberg 2"), **38**: plants ("Flora Garnitzenberg 3"), **39**: algae, fusulinids.

The variegated lithology of the section, containing conglomerates, pebbly sandstones, sandstones, mudstones, marls and limestones is of Gzehlian age and famous for its richness in fossils (calcareous algae, plant fossils, foraminifera, especially fusulinids, ostracods, sphinctozoans, solitary rugose corals, bivalves, gastropods, conulariids, brachiopods and ichnofossils) (figs. 41, 45).

Following rugose corals were mentioned (especially by HERITSCH, 1936) from the Naßfeld area:

Pizzul Fm.:	*Amplexocarinia smithi alpha* HERITSCH
	Lopholasma carbonaria GRABAU
	Lophocarinophyllum acanthiseptatum GRABAU
	Sinophyllum carnicum HERITSCH
	Sinophyllum sp.
	Caninia ? pannonica FRECH
	Caninia ? sp.
	Fomichevella nikitini (STUCKENBERG)
	Caruthersella carnica HERITSCH
	Dibunophyllum ? carnicum HERITSCH
	Lophophylloides carnicum (HERITSCH)
	Palaeosmilia demaneti HERITSCH
	Thysanophyllum vinassai HERITSCH
	Thysanophyllum sp.
	Amandophyllum heritschi MINATO et KATO
	Lonsdaleia sp.
	Geyerophyllum broilii (HERITSCH)
Corona Fm.:	*Zaphrentis omaliusi* MILNE-EDWARDS et HAIME ?
	Lopholasma carbonaria GRABAU
	Amplexus coronae FRECH ? in SCHELLWIEN
Auernig Fm.:	*Lopholasma carbonaria* GRABAU
	Duplophyllum sp.
	Amandophyllum carnicum (HERITSCH)
	Amygdalophylloides sp.
	Carinthiaphyllum carnicum HERITSCH
	Carinthiaphyllum kahleri HERITSCH
Carnizza Fm.:	*Amplexocarinia smithi alpha* HERITSCH
	Lopholasma carbonaria GRABAU
	Bradyphyllum angeli HERITSCH
	Allotropiophyllum sp.
	Hapsiphyllum boswelli HERITSCH
	Lophocarinophyllum acanthiseptatum GRABAU
	Sinophyllum minimum HERITSCH
	Lophophyllidium minimum (HERITSCH)
	Lophophyllidium profundum (MILNE-EDWARDS et HAIME)
	Amygdalophylloides sp.
	Geyerophyllum carnicum (HERITSCH)
	Lonsdaloides boswelli HERITSCH

According to KRAINER (1992) the upper part of the Corona formation as well as the Auernig and Carnizza Formation represent cyclic sequences (Auernig cyclothem) with a thickness of 10-40 m. This mixed clastic-carbonatic sequences are storm dominated shelf sediments with quartz-rich conglomerates of the beach and upper shoreface through crossbedded sandstones (upper shoreface) hummocky crossbedded sandstones (lower shoreface) to bioturbated mudstones and fossiliferous marls and limestones (offshore, below storm wave base).
KRAINER (1992) expected the duration of the cycles in the order of 100.000 years and explained the cycle formation by glacio-eustatic sea level fluctuations caused by the Permo-Carboniferous Gondwana glaciation.

The Permian to Triassic sequence of the Naßfeld-Gartnerkofel area gave rise to a study of the mass extinction event near the Permian/Triassic boundary (HOLSER & SCHÖNLAUB, 1991). A core (GK-1) to a depth of 331 m was accomplished and the core material was studied with respect to biostratigraphy, sedimentology, paleontology, mineralogy-petrology elemental and isotope geochemistry, and paleomagnetics. The conclusions of these studies (HOLSER et al., 1991) were, that the "P/Tr-event may have been spread out over several million years during which extinctions were paralleled by the deterioration of organic carbon deposition consequent on sea level regression".

During the entire late Paleozoic the Naßfeld region formed a bay-like extension at the western margin of the Tethys. Almost until the end of the Permian it remained a coastal region affected by more or less significant sea-level changes which resulted in cyclic sediments such as transgressive highstand marine limestones and regressive lowstand plant-bearing sandstones. The latter have been famous for their rich occurrence of different groups of paleofloras.

Fig. 42: Aerial view of the Naßfeld region with Gartnerkofel, Garnitzenberg and Mount Auernig (courtesy of tourist board Hermagor).

Stop 14: End of Gartnerkofel Cable Road

The limestone horizon occurring at this place represents a rather thin sedimentary wedge of the so-called Auernig Formation which has been preserved in the surroundings of the Naßfeld from the originally thick pile of the post-Variscan molasse type cover (fig. 43). In central Europe similar deposits are missing.

The Upper Carboniferous Auernig Fm. exceeds a thickness of 800 m and comprises an interchange of limestones, marls, shales, sandstones and quartz conglomerates. Its fossil content and lithology reflects near-shore terrestrial deposits with repeated shallow-marine incursions. The general dipping is southward directed with younger sediments following towards south. The cable station was constructed atop fossiliferous limestones with clayish interbeds yielding brachiopods, corals, sponges, gastropods, bivalves, crinoids, algae und foraminifera. Immediately below the station mica-rich calcareous sandstones and brownish shales occur which are locally rich in fossils. Among others the brachiopod *Isogramma paotechowensis* was reported from one distinct layer.

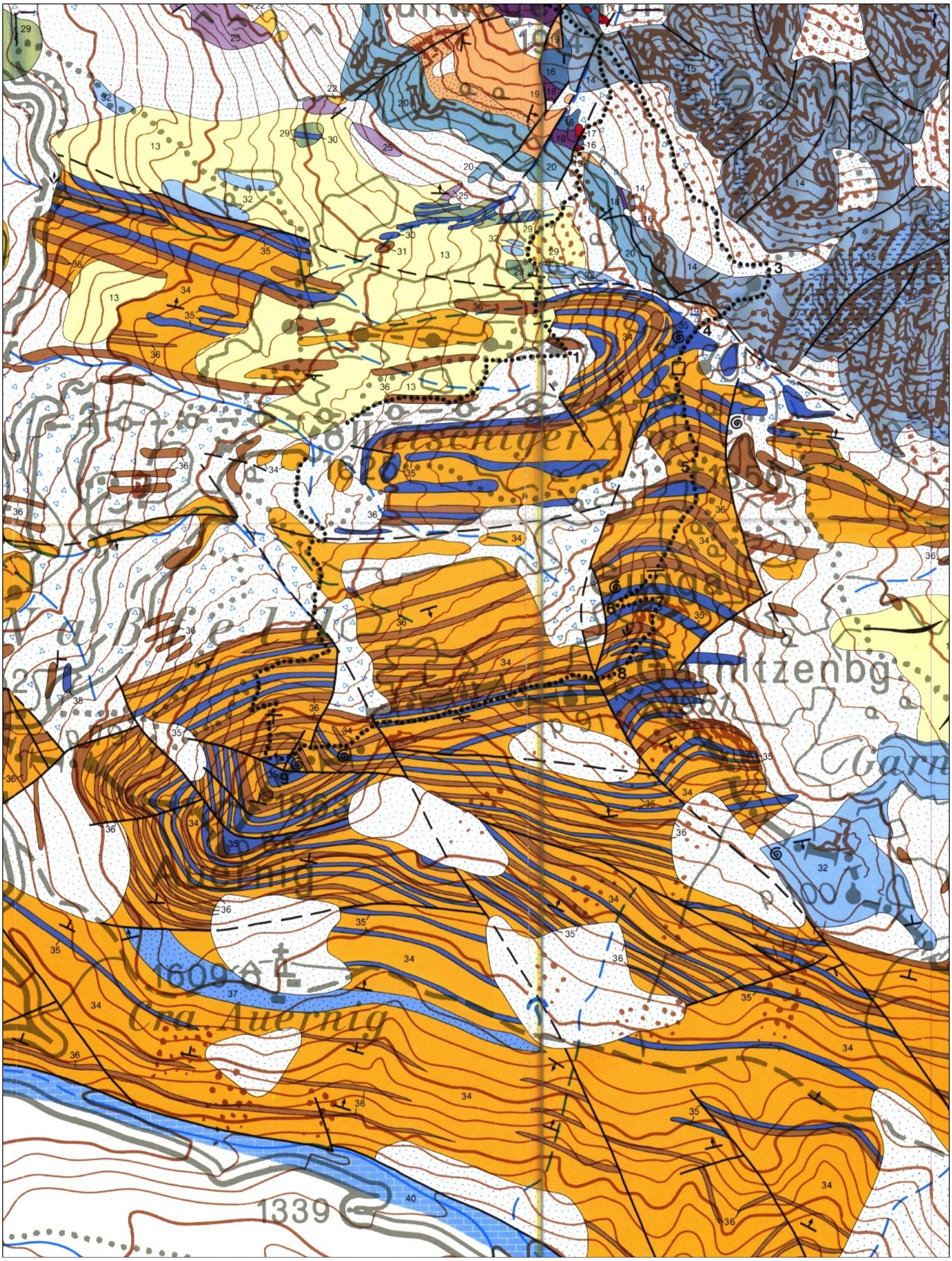

Fig. 43: Geological map of the Naßfeld area (cut from the Geological Map, Sheet 198, Weißbriach, Edit. Geol. Survey of Austria, 1987) showing mainly Auernig Fm. (orange colour) with interbedded limestones (blue) and conglomerates (brown). Bluish part on upper right part comprises Middle Triassic Schlern Dolomite.

Fig. 44: View from Naßfeld to the west with the Trogkofel composed of Lower Permian limestones (courtesy of the tourist board Hermagor).

Fig. 45: Brachiopod coquina from the famous SCHELLWIEN locality east of Naßfeld.

Stop 15: Depression between the cable station and Garnitzenberg, 1856 m

The distinct and more than 10 m thick bed of quartz pebble bearing conglomerates occurs near the middle part of the Auernig Formation. This portion is dominated by sandstones, claystones and quartz gravels. Less abundant are limestones and marls. The very compact conglomerates represent fluvial deposits which were transported from the Variscan hinterland north of the Gail Valley to the Tethyan coast. These gravels are generally well rounded and have diameters between 3 and 4 cm. They consist of more than 90% of quartz. In addition some pebbles comprise slates, quartzite, micaschist, gneisses and even black platy cherts. The latter are most probably derived from Silurian or Devonian rocks formed in deep oceanic water.

Stop 16: Gugga, north of Garnitzenberg, 1928 m

The prominent 12 m thick limestone horizon called "Gugga" represents at its base (1) a transgressive sequence composed of shallow marine siliciclastic sedimentary rocks overlain by fossiliferous limestones and (2) a 15 m thick limestone interval with a massive algal mound facies in the middle part.

The exposed sequence begins with a quartz-rich conglomerate, which erosively overlies fine-grained, hummocky-crossbedded sandstones containing brachiopods and crinoid fragments. The conglomerate is well sorted and rounded, and interpreted to have been deposited in a nearshore environment. The conglomerate facies is overlain by a 10-20 cm thick siltstone-shale interval containing abundant plant fossils. FRITZ & BOERSMA (1990) described from this locality 16 different taxa. The occurrence of *Alethopteris bohemica*, *Odontopteris brardii* and *Pecopteris feminaeformis* refers to the Stefanian Series of the Upper Carboniferous. The fossil-bearing layers are overlain by several meters of siltstones with intercalated fine-grained sandstones.

The clastic sequence is overlain by a 16 m thick limestone succession ("Gugga Limestone") consisting of bedded limestones at the base and on top and massive to indistinctly bedded limestones in the middle part. The latter are composed of *Anthracoporella* wackestones and bafflestones (mound facies), whereas the bedded limestones consist of different types of bioclastic wackestones and packstones (intermound facies). Also, the bedded limestones at the base and particularly on top contain abundand fusulinids. Based on new identifications of fusulinids an Upper Gzhelian age is inferred for this part of the Auernig Formation (KAHLER, 1983, 1985, 1986; KRAINER & DAVYDOV, 1998).

Stop 17: Bedded fossiliferous limestones of the Auernig Fm. with intercalated "micromounds" north of Garnitzenberg

The following stop is located about 100 m northwest of Garnitzenberg (1951 m) at an altitude of 1925 m. The outcrop represents a 7 m thick sequence of well bedded fossiliferous limestones containing *Archaeolithophyllum missouriense*- and *Anthracoporella* micromounds.

The thickness of individual limestone beds ranges from a few cm up to 25 cm. Bedding is wavy to nodular, particularly in the lower part of the sequence. Individual limestone beds are not laterally persistent, but pinch out over short distances of one to several meters. Most interbeds comprise dark, fossiliferous marls ranging in thickness between a few mm and 10 cm and contain algal fragments and/or fusulinids. The dark-grey limestone is locally rich in fossils. In some beds even algae occur in growth position forming small mounds which have been termed "micromounds" by KRAINER (1995). He recognized (1) bioclastic wackestones/packstones, (2) fusulinid wackestones/packstones, (3) *Anthracoporella* wackestones/packstones, (4) *Anthracoporella* packstones/rudstones, (5) bioclastic mudstones, (8) bindstones and (9) bioclastic siltstones/fine-grained sandstones. The micromounds are composed of *Archaeolithophyllum* boundstones/bafflestones and *Anthracoporella* baffle-stones (KRAINER, 1995).

The *Archaeolithophyllum* bafflestones/bindstones consist of a micritic to pelmicritic matrix in which well-preserved, sometimes branching thalli of several cm long *Archaeolithophyllum missouriense* occur in growth position. Growth position is either upright to oblique or parallel to the bedding plane indicating that *A. missouriense* acted as a sediment baffler or as a sediment binder. Many thalli of *A. missouriense* are encrusted by "micritic algae", sessile foraminifers, *Tubiphytes* and rarely by thin crusts of *Archaeolithophyllum lamellosum*. In *Anthracoporella* bafflestones almost all thalli occur in growth position. It formed upright to heights of several centimetres. The space between the algal thalli is filled with micrite and pelmicrite.

The algae *Anthracoporella* seems to have lived in loosely packed associations. Locally, where the algae grew closer, some open pore space remained, which was filled with spar cement. In both micromound types only a few small bioclasts are present in the matrix.

Stop 18: High-frequency cycles in the upper part of the Auernig Fm. and lower part of the Carnizza Fm.

The following section from the peak of Garnitzenberg (1951 m) to the south covers more than 160 m of mainly clastic rocks. Based on current practice along the trail the uppermost 90 meters of the Auernig Formation and the basal 70 meters of the Carnizza Formation are exposed which represent several high-frequency cycles ("Auernig cyclothems"). In the Auernig Formation mixed clastic-carbonate cycles are developed, consisting of thin conglomerates, trough- and hummocky crossbedded sandstones, siltstones with abundant trace fossils, and thin bedded, fossiliferous limestones. The lower part of the Carnizza Fm. is composed of clastic sediments deposited in a shallow marine environment (conglomerates, sandstones and siltstones), which are also vertically arranged to form prominent cycles. Two thin shale intervals within siltstones (localities Garnitzenberg 1 and 2) yielded well preserved plant fossils indicating an Upper Stephanian age (FRITZ & BOERSMA, 1990). The top of the exposed section is formed by a 3 m thick succession of bedded fossiliferous limestones.

Stop 19: Top of the Auernig Mountain, type locality of the Auernig Formation

The Auernig section is one of the most famous outcrops in the Pramollo/Naßfeld area (fig. 46). Beginning at the end of the 19th century, Austrian and Italian geologists investigated the characteristic sequence at the western and southern flanks of this mountain, consisting of an alternation of carbonates, conglomerates, sandstones and marls. The beds were designated by letters (SCHELLWIEN, 1892) or numbers, respectively (GEYER, 1896). The uppermost carbonate bed of the Auernig section named bed "s" by SCHELLWIEN contains selectively silicified organisms such as ostracods, smaller foraminifera, fusulinids, bryozoans, brachiopods and fragments of calcareous algae. This kind of preservation enables the study of silicified microfossils in hydrochloric or acetic acid residues and the comparison between the silicified whole-body fossils and fossils from thin-sections.

Fig. 46: Mount Auernig from the east showing alternating sandstones, conglomerates and limestones.

Because of the extremely finegrained silicification of the shells, ostracods exhibit an excellent preservation (fig. 47). Nearly all details of the ornamentation are preserved. The fauna consists of 62 ostracod species including two new species. About 75% belong to the more or less smooth-shelled Podocopida with a curved hinge, and 25% to the highly ornamented Paleocopida with a straight hinge. The assemblage pattern of the ostracods corresponds to that of shallow marine, near-shore and low-energy environments (BLESS, 1983; BANDEL & BECKER, 1975; BECKER, 1982; FOHRER, 1991). Microfacies analysis supports the environmental interpretation based on the ostracod assemblages. Six microfacies types represented by (1) autochthonous algal wackestones (dominated by *Archaeolithophyllum* and/or *Anthracoporella)* and (2) bioclastic wackestones, packstones and grainstones can be recognized. The bioclastic microfacies types show a high diversity (fusulinids, smaller foraminifera, bryozoans, brachiopods, echinoderms, gastropods and some small trilobites), while the diversity of the algal wackestones is conspicuously low.

Fig. 47: Silicified microfauna (bryozoans, fusulinids, wormtubes, gastropods, bivalves, echinoderms, corals) from bed no. s on the top of Mount Auernig. Foto: B. FOHRER.

The section at this outcrop is composed of a fining-upward sequence with conglomerates, hummocky crossbedded sandstones and siltstones grading upward into bedded and massive limestones. The top consists of a coarsening-upward sequence of bedded limestones grading upward into sandstones with conglomeratic layers. The fining-up sequence is interpreted as a transgressive trend whereas the following coarsening-upward sequence is regressive.

The main features to be observed at this locality are:

- transgressive-regressive sequences
- rapid vertical changes in facies, biotic association and biodiversity
- small algal mounds *(Anthracoporella* and *Archaeolithophyllum missouriense* in growth position) within the "massive limestone"
- high biodiversity within the dark limestones (bryozoans, brachiopods, ostracods, foraminifers and algal fragments) as well as selected silicification of biota.

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Appendix

Distribution of Palaeozoic corals in Austria. For locational areas confer to fig. 2 in the main text.
1: Greywacke Zone, 2: Palaeozoic window of Burgenland, 3: Graz Palaeozoic, 4: Gurktal Nappe (including small Lower Devonian outcrops in southern Styria), 5: Nötsch Carboniferous, 6: Carnic Alps, 7: Karawanken Mountains. Corals of 1 to 5 belong to the ACF, 6 and 7 to the SCF.

Rugosa

1	**2**	**3**	**4**	**5**	**6**	**7**	
○	○	●	○	○	●	○	*Acanthophyllum concavum* (WALTHER, 1928)
○	○	●	○	○	○	○	*Acanthophyllum delicatum* (KETTNEROVA, 1932)
○	○	●	○	○	●	○	*Acanthophyllum heterophyllum heterophyllum* (MILNE-EDWARDS & HAIME, 1851)
○	○	●	○	○	○	○	*Acanthophyllum heterophyllum torquatum* (SCHLÜTER 1884)
○	○	○	○	○	●	○	*Acanthophyllum* cf. *heterophyllum* (MILNE-EDWARDS & HAIME, 1851)
○	○	●	○	○	○	○	*Acanthophyllum* cf. *moravicum* (KETTNEROVA, 1932)
○	○	●	○	○	○	○	*Acanthophyllum smyckai* (KETTNEROVA, 1932)
○	○	●	○	○	○	○	*Acanthophyllum* cf. *smyckai* (KETTNEROVA, 1932)
○	○	●	○	○	●	○	*Acanthophyllum vermiculare* (GOLDFUSS, 1826)
○	○	●	○	○	●	○	*Acanthophyllum* sp.
○	○	○	○	○	●	○	*Acervularia* sp.?
○	○	○	○	○	●	○	*Alaiophyllum jarushevskyi* GORYANOV, 1961
○	○	○	○	○	●	○	*Alaiophyllum wirbelauense* (PICKETT, 1967)
○	○	○	○	○	●	○	*Allotropiophyllum carnicum* HERITSCH, 1936
○	○	○	○	○	●	○	*Allotropiophyllum* sp.
○	○	○	○	○	●	●	*Amandophyllum carnicum* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Amandophyllum heritschi* MINATO & KATO, 1965
○	○	○	○	○	●	○	*Amandophyllum zeliae* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Amandophyllum* sp.
○	○	○	○	○	○	●	*Amplexocarinia geyeri* HERITSCH, 1933
○	○	○	○	○	●	●	*Amplexocarinia heimoi* HERITSCH, 1936
○	○	○	○	○	●	●	*Amplexocarinia muralis biseptata* SOSHKINA, 1932
○	○	○	○	○	●	●	*Amplexocarinia muralis irginae* SOSHKINA, 1932
○	○	○	○	○	●	●	*Amplexocarinia smithi alpha* HERITSCH, 1936
○	○	○	○	○	●	●	*Amplexocarinia smithi smithi* HERITSCH, 1936
○	○	○	○	○	○	●	*Amplexus* ? *carinthiacus* PENECKE, 1887
○	○	○	○	○	●	○	*Amplexus coronae* FRECH ? in SCHELLWIEN, 1892
○	○	○	○	○	●	○	*Amplexus frechi frechi* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Amplexus frechi major* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Amplexus gortanii* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Amplexus hercynicus* ROEMER, 1855?
○	○	○	○	○	●	○	*Amplexus irregularis* KAYSER, 1873?
○	○	●	○	○	●	○	*Amplexus mutabilis* MAURER, 1885
○	○	●	○	○	●	○	*Amplexus tortuosus* PHILLIPS, 1841?
○	○	●	○	○	○	○	*Amplexus ungeri* (PENECKE, 1894)
○	○	●	○	○	●	○	*Amplexus* sp.
○	○	●	○	○	○	○	*Amplexus* sp. aff. *helminthoides* FRECH, 1885
○	○	○	○	○	●	○	*Amplexus* sp. indet. ex aff. *irregularis* KAYSER, 1873
○	○	○	○	○	●	●	*Amygdalophylloides* sp.
○	○	○	○	●	○	○	*Arachnolasma cylindricum* YU, 1933
○	○	○	○	○	●	○	*Arachnophyllum diffluens* (MILNE-EDWARDS & HAIME, 1851)?
○	○	○	○	○	●	○	*Aspasmophyllum* sp.
○	○	○	○	○	●	○	*Aulophyllum fungites* (FLEMING, 1828)
●	○	○	○	○	○	○	*Axophyllum lonsdaleiforme* (SALEE, 1913)
○	○	○	○	○	●	○	*Barrandeophyllum carnicum* SCHOUPPE, 1954

○	○	●	○	○	○	○	*Barrandeophyllum* n. sp. aff. *perplexum* FLÜGEL, 1956
○	○	○	○	○	●	○	*Battersbyia devonica* (BULVANKER, 1958)
○	○	○	○	○	●	○	*Battersbyia petshorense* (SOSHKINA, 1949)
○	○	○	○	○	●	○	*Battersbyia symbiotica* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Battersbyia syringoporoides* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Battersbyia* sp.
○	○	○	○	○	●	○	*Bothrophyllum* ? *densiseptatum* HERITSCH, 1936
○	○	○	○	○	●	○	*Brachyelasma* ? *alpina* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Bradyphyllum angeli* HERITSCH, 1936
○	○	○	○	○	●	○	*Bradyphyllum* sp.
○	○	●	○	○	○	○	*Calceola sandalina* LAMARCK, 1799
○	○	○	○	○	●	○	*Caninia* ? *fredericksi* HERITSCH, 1936
○	○	○	○	○	●	●	*Caninia* ? *koksharowi* STUCKENBERG, 1895
○	○	○	○	○	○	●	*Caninia* ? aff. *koksharowi* STUCKENBERG, 1895
○	○	○	○	○	●	○	*Caninia* ? *pannonica* FRECH, 1906
○	○	○	○	○	●	○	*Caninia* ? *sophiae* HERITSCH, 1936
○	○	○	○	●	●	○	*Caninia* sp.
●	○	○	○	○	○	○	*Caninophyllum archiaci* (MILNE-EDWARDS & HAIME, 1852)
○	○	○	○	○	●	○	*Caninophyllum gortanii* HERITSCH, 1933
○	○	○	○	○	○	●	*Caninophyllum* cf. *oribos* (SALTER, 1855)?
○	○	○	○	○	●	●	*Carinthiaphyllum carnicum* HERITSCH, 1936
○	○	○	○	○	●	●	*Carinthiaphyllum kahleri* HERITSCH, 1936
○	○	○	○	○	○	●	*Carinthiaphyllum suessi* HERITSCH, 1936
○	○	○	○	○	○	●	*Carinthiaphyllum* cf. *suessi* HERITSCH, 1936
○	○	○	○	○	●	○	*Carniaphyllum gortanii* HERITSCH, 1933
○	○	○	○	○	●	○	*Carruthersella carnica* HERITSCH, 1936
○	○	●	○	○	●	○	*Ceratophyllum ceratites* (GOLDFUSS, 1824)
○	○	○	○	○	●	○	*Chonophyllum* sp.
○	○	○	○	○	●	○	*Clisiophyllum* ? *pironai* ANGELIS D'OSSAT, 1895
○	○	○	○	○	●	○	*Clisiophyllum* cf. *praecursor* FRECH, 1885
○	○	○	○	○	●	○	*Clisiophyllum* ? *taramelli* VINASSA DE REGNY, 1918
●	○	○	○	●	○	○	*Clisiophyllum* sp.
○	○	●	○	○	○	○	*Columnaria* cf. *inaequalis* HALL, 1852?
○	○	●	○	○	○	○	*Columnaria* n. sp. A
○	○	○	○	○	●	○	*Columnaria* sp.
○	○	○	○	○	○	●	*Cyathaxonella* sp.
○	○	○	○	○	●	○	*Cyathaxonia cornu* MICHELIN, 1847
●	○	○	○	○	○	●	*Cyathaxonia* aff. *krotowi* STUCKENBERG, 1895
○	○	○	○	○	○	○	*Cyathaxonia rhusiana* VAUGHAN, 1906
○	○	○	○	○	○	○	*Cyathaxonia* aff. *rhusiana* VAUGHAN, 1906
○	○	○	○	○	●	○	*Cyathaxonia* sp.
○	○	●	○	○	●	○	*Cyathophyllum (Cyathophyllum) dianthus* GOLDFUSS, 1826
○	○	○	○	○	●	○	*Cyathophyllum (Cyathophyllum) spongiosum* (SCHULZ, 1883)
○	○	●	○	○	●	○	*Cyathophyllum (Peripaedium) planum* (LUDWIG, 1865)?
○	○	●	○	○	○	○	*Cyathophyllum (Peripaedium) turbinatum* GOLDFUSS, 1826?
○	○	○	○	○	●	○	*Cyathophyllum* ? *alpinum* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Cyathophyllum* ? *angustum* LONSDALE, 1839?
○	○	○	○	○	●	○	*Cyathophyllum* ? *arietinum* FISCHER VON WALDHEIM, 1830?
○	○	○	○	○	●	○	*Cyathophyllum* ? *bathycalyx* FRECH, 1886
○	○	○	○	○	●	○	*Cyathophyllum* ? *canavarii* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Cyathophyllum* ? *collinense* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Cyathophyllum conglomeratum pauciseptatum* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Cyathophyllum* ? *explanatum* GOLDFUSS, 1826
○	○	●	○	○	○	○	*Cyathophyllum* ? *flexuosum* GOLDFUSS, 1826
○	○	●	○	○	○	●	*Cyathophyllum* ? cf. *flexuosum* GOLDFUSS, 1826
○	○	●	○	○	●	○	*Cyathophyllum frechi* PENECKE, 1887?

○	○	○	○	○	●	○	*Cyathophyllum* ? *gortanii* VINASSA DE REGNY, 1918
○	○	●	○	○	●	○	*Cyathophyllum* ? *graecense* PENECKE, 1894
○	○	●	○	○	○	○	*Cyathophyllum* ? cf. *graecense* PENECKE, 1894
○	○	●	○	○	●	○	*Cyathophyllum* ? *hallioides* FRECH, 1886
○	○	○	○	○	●	○	*Cyathophyllum* ? cf. *hallioides* FRECH, 1886
○	○	●	○	○	○	○	*Cyathophyllum* ? *hoernesi* PENECKE, 1894
○	○	○	○	○	●	○	*Cyathophyllum* ? *lindströmi* FRECH, 1885
○	○	○	○	○	●	○	*Cyathophyllum* ? *macrocystis* (FRECH, 1886)?
○	○	○	○	○	●	○	*Cyathophyllum* ? n. sp. ex aff. *dianthus* GOLDFUSS, 1826
○	○	●	○	○	○	○	"*Cyathophyllum* n. sp. aff. *frechi*" PENECKE, 1889
○	○	○	○	○	●	○	*Cyathophyllum* ? *plicatum* GOLDFUSS, 1826?
○	○	○	○	○	●	○	*Cyathophyllum* ? *taramelli* ANGELIS D'OSSAT, 1901
○	○	○	○	○	●	○	*Cyathophyllum* ? *tinocystis carnicum* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Cyathophyllum* ? *vermiculare carnicum* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Cyathophyllum* ? *volaicum* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Cyathophyllum* ? cf. *volaicum* CHARLESWORTH, 1914
●	●	●	○	○	○	●	*Cyathophyllum* sp.
○	○	○	○	○	●	○	*Cystiphyllum* aff. *cristatum* FRECH, 1886
○	○	○	○	○	●	○	*Cystiphyllum* ? *geyeri* ANGELIS D'OSSAT, 1901
○	○	○	○	○	●	○	*Cystiphyllum* ? *intermedium densum* CHARLESWORTH, 1914
○	○	●	○	○	○	○	*Cystiphyllum* sp.
●	○	○	○	○	○	○	*Dibunophyllum* cf. *bipartitum* THOMSON & NICHOLSON, 1876
○	○	○	○	○	●	○	*Dibunophyllum* ? *carnicum* HERITSCH, 1936
○	○	○	○	○	●	○	*Dibunophyllum* sp.
●	○	○	○	○	○	○	*Diphyphyllum lateseptatum* MCCOY, 1849
○	○	○	○	○	●	○	*Diphyphyllum* sp.
○	○	●	○	○	○	○	*Disphyllum aequiseptatum* (MILNE-EDWARDS & HAIME, 1851)
○	○	●	○	○	○	○	*Disphyllum* cf. *aequiseptatum* (MILNE-EDWARDS & HAIME, 1851)
○	○	●	○	○	●	○	*Disphyllum caespitosum caespitosum* (GOLDFUSS, 1826)
○	○	●	○	○	○	○	*Disphyllum caespitosum pashiense* (SOSHKINA, 1939)
○	○	●	○	○	●	○	*Disphyllum goldfussi* (GEINITZ, 1846)
○	○	●	○	○	○	○	*Disphyllum hsianghsienense kostetskae* (SOSHKINA, 1949)
○	○	○	○	○	●	○	*Disphyllum* ? *recressum* (HILL, 1941)
○	○	●	○	○	●	○	*Dohmophyllum helianthoides* (GOLDFUSS, 1828)
○	○	○	○	○	●	○	*Dohmophyllum* cf. *involutum* WEDEKIND, 1923
○	○	○	○	○	●	○	*Dohmophyllum philocrinum* FRECH, 1886)
○	○	●	○	○	○	○	*Dokophyllum* cf. *murchisoni* (MILNE-EDWARDS & HAIME, 1851)?
○	○	●	○	○	○	○	*Dokophyllum* cf. *subturbinatum* (D'ORBIGNY, 1849)?
○	○	○	○	○	●	○	*Dokophyllum* sp.
○	○	○	○	○	●	○	*Duplophyllum mikron* SCHOUPPÉ & STACUL, 1959
○	○	○	○	○	●	○	*Duplophyllum* sp.
○	○	○	○	○	●	●	*Durhamina ampfereri* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Endophyllum acanthicum* FRECH, 1885
○	○	○	○	○	●	○	*Endophyllum carnicum* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Endophyllum priscum* (MÜNSTER, 1840)
○	○	○	○	○	●	○	*Endophyllum* cf. *priscum* (MÜNSTER, 1840)
○	○	○	○	○	●	○	*Endophyllum* sp. ex aff. *acanthicum* FRECH, 1885
○	○	○	○	○	●	○	*Entelophyllum* ? *alpinum* SCHOUPPÉ, 1951
○	○	○	○	○	●	●	*Entelophyllum articulatum* (WAHLENBERG, 1821)?
○	○	●	○	○	○	○	*Entelophyllum* sp.
○	○	○	○	○	●	○	*Favistella (Dendrostella) fluegeli* KODSI, 1971
○	○	○	○	○	●	○	*Favistella (Dendrostella)* cf. *praerhenana* GLINSKI, 1957
○	○	○	○	○	●	○	*Favistella (Dendrostella) trigemme* (QUENSTEDT, 1881)
○	○	○	○	○	●	○	*Favistella (Dendrostella) vulgaris* (SHOSHKINA, 1936)
○	○	●	○	○	○	○	*Favistella (Dendrostella)* sp.
●	○	○	○	○	●	○	*Fomichevella nikitini* (STUCKENBERG, 1895)

○	○	○	○	○	●	○	*GEYERohyllum broilii* (HERITSCH, 1936)
○	○	○	○	○	●	○	*GEYERohyllum carnicum* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Grewingkia* ? *carnica* (SCHOUPPÉ, 1954)
○	○	●	○	○	○	○	*Grypophyllum denckmanni* WEDEKIND, 1922
○	○	●	○	○	○	○	*Grypophyllum* cf. *denckmanni* WEDEKIND, 1922
○	○	●	○	○	○	○	*Grypophyllum frechi* BIRENHEIDE, 1974
○	○	●	○	○	●	○	*Grypophyllum* sp.
○	○	●	○	○	○	○	*Hallia* ? *sophiae* SCHLÜTER, 1937
○	○	●	○	○	●	○	*Hallia* sp.
○	○	○	○	○	●	○	*Hapsiphyllum boswelli* HERITSCH, 1936
○	○	○	○	○	○	●	*Hapsiphyllum elegantulum* GRABAU, 1928
○	○	○	○	○	●	○	*Heliophyllum* sp.
○	○	●	○	○	○	○	*"Hexagonaria" darwini* (FRECH, 1885)
○	○	●	○	○	○	○	*Hexagonaria hexagona* (GOLDFUSS, 1826)
○	○	●	○	○	○	○	*"Hexagonaria"* cf. *santacrucensis* MOENKE, 1954
○	○	○	○	●	○	○	*Hexaphyllia mirabilis* (DUNCAN, 1867)
●	○	○	○	○	○	○	*Hexaphyllia* sp.
○	○	○	○	○	●	○	*Ipciphyllum* sp.
○	○	○	○	○	●	●	*Kabakovitchiella ruedemanni* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Lindstroemia* aff. *laevis* NICHOLSON & ETHERIDGE, 1878
●	○	○	○	○	○	○	*Lindstroemia subduplicata* (MCCOY, 1850)?
○	○	○	○	○	●	○	*Lithostrotion irregulare* (PHILLIPS, 1836)
○	○	○	○	○	●	○	*Lithostrotion junceum* (FLEMING, 1828)
●	○	○	○	○	○	○	*Lonsdaleia duplicata* (MARTIN, 1809)?
○	○	○	○	○	●	○	*Lonsdaleia* sp.
○	○	○	○	○	●	○	*Lonsdaleoides boswelli* HERITSCH, 1936
○	○	○	○	○	●	○	*Lonsdaleoides* cf. *boswelli* HERITSCH, 1936
○	○	○	○	○	●	○	*Lophocarinophyllum acanthiseptatum* GRABAU, 1922
○	○	○	○	○	●	○	*Lophocarinophyllum major* HERITSCH, 1936
○	○	○	○	○	●	○	*Lopholasma carbonaria* GRABAU, 1922
○	○	○	○	○	●	●	*Lopholasma ilitschense* SOSHKINA, 1928
○	○	○	○	○	●	○	*Lophophyllidium kahleri* FELSER, 1937
○	○	○	○	○	●	○	*Lophophyllidium minimum* (HERITSCH, 1936)
○	○	○	○	○	●	●	*Lophophyllidium profundum* (MILNE-EDWARDS & HAIME, 1851)
○	○	○	○	○	●	○	*Lophophyllidium* sp.
○	○	○	○	○	●	○	*Lophophylloides carnicum* (HERITSCH, 1936)
○	○	○	○	○	●	○	*Lophophylloides* sp.
○	○	○	○	○	●	○	*Lophophyllum breve* de KONINCK, 1872
○	○	○	○	○	●	○	*Lophophyllum dumonti* MILNE-EDWARDS & HAIME, 1851?
○	○	○	○	○	●	●	*Lophophyllum* ? *proliferum* (MCCHESNEY, 1860)
○	○	○	○	○	●	○	*Lophophyllum tortuosum* (MICHELIN, 1846)
○	○	○	○	○	●	○	*Lyrielasma subcaespitosa carnica* FLÜGEL, 1962 (nomen nudum)
○	○	○	○	○	●	○	*Lyrielasma* sp.
○	○	●	○	○	○	○	*Mesophyllum (Cystiphylloides) caespitosum* (SCHLÜTER, 1882)
○	○	●	○	○	○	○	*Mesophyllum (Cystiphylloides) macrocystis* (SCHLÜTER, 1889)
○	○	●	○	○	○	○	*Mesophyllum (Cystiphylloides) pseudoseptatum* (SCHULZ, 1883)
○	○	●	○	○	●	○	*Mesophyllum (Mesophyllum) cristatum* (FRECH, 1886)
○	○	○	○	○	●	○	*Mesophyllum (Mesophyllum) originale* (BIRENHEIDE, 1964)
○	○	●	○	○	●	○	*Mesophyllum (Mesophyllum) vesiculosum vesiculosum* (GOLDFUSS, 1824)
○	○	●	○	○	○	○	*Mesophyllum* (?) sp.
○	○	●	○	○	○	○	*Metriophyllum gracile* SCHLÜTER, 1884
○	○	●	○	○	○	○	*Moravophyllum tenuiseptatum* KETTNEROVA, 1932
○	○	●	○	○	○	○	*Neaxon symmetricus* (FRECH, 1886)
○	○	●	○	○	○	○	*Neaxon* sp.
○	○	●	○	○	○	○	*Orthophyllum* sp.
○	○	○	○	○	●	○	*Palaeosmilia demaneti* HERITSCH, 1936

○	○	○	○	○	●	○	*Palaeosmilia hammeri* HERITSCH, 1933
○	○	○	○	●	○	○	*"Palaeosmilia" isae* HERITSCH, 1934
●	○	○	○	●	○	○	*Palaeosmilia murchisoni murchisoni* MILNE-EDWARDS & HAIME, 1848
●	○	○	○	○	○	○	*Palaeosmilia murchisoni pendlensis* PARKINSON, 1926
○	○	○	○	○	●	○	*Palaeosmilia* sp.
○	○	●	○	○	○	○	*Peneckiella achanyensis* SOSHKINA, 1939
●	○	○	○	○	○	○	*Petraia belatula* POCTA, 1902
○	○	○	○	○	●	○	*Petraia benedeniana* DE KONINCK, 1872?
○	○	○	○	○	●	○	*Petraia* ? *confinensis* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Petraia laevis* POCTA, 1902?
○	○	○	○	○	●	○	*Petraia* aff. *laevis* POCTA, 1902?
○	○	○	○	○	●	○	*Petraia* aff. *semistriata* MÜNSTER, 1839?
○	○	○	○	○	●	○	*Petraia* sp.
○	○	○	○	○	●	○	*Pexiphyllum heterophylloides* (FRECH, 1885)
○	○	○	○	○	●	○	*Pexiphyllum* sp.
○	○	●	○	○	○	○	*Phacellophyllum conglomeratum* (SCHLÜTER, 1880)?
○	○	○	○	○	●	○	*Phacellophyllum* cf. *conglomeratum* (SCHLÜTER, 1880)?
○	○	○	○	○	●	○	*Phillipsastrea ananas* (GOLDFUSS, 1828)
○	○	○	○	○	●	●	*Phillipsastrea hennahi* (LONSDALE, 1840)
○	○	○	○	○	●	○	*Polythecalis* cf. *rosiformis* HUANG, 1932
○	○	○	○	○	●	○	*Pseudamplexus bohemicus* POCTA, 1902
○	○	○	○	○	●	○	*Pseudamplexus frechi* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Pseudamplexus* cf. aff. *frechi* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Pseudamplexus* cf. *frechi* (CHARLESWORTH, 1914)
○	○	●	○	○	○	○	*Pseudohexagonaria amanshauseri* (GLINSKI, 1955)
○	○	○	○	○	●	○	*Pseudohuangia* sp.
○	○	○	○	●	○	○	*Pseudozaphrentoides juddi juddi* (THOMPSON, 1893)
○	○	○	○	●	○	○	*Pseudozaphrentoides* sp.
○	○	○	○	○	●	○	*Pterorrhiza dubia* (DE BLAINVILLE, 1830)
○	○	○	○	○	●	○	*Pycnactis mitratum* (SCHLOTHEIM, 1820)?
○	○	○	○	○	●	○	*Scruttonia julli* (PEDDER, 1986)
○	○	○	○	○	●	○	*Sinophyllum carnicum* HERITSCH, 1936
○	○	○	○	○	●	○	*Sinophyllum multiseptatum irregulare* FELSER, 1937
○	○	○	○	○	●	●	*Sinophyllum pendulum carinthiacum* FELSER, 1937
○	○	○	○	○	○	●	*Sinophyllum pendulum pendulum* GRABAU, 1922
○	○	○	○	○	●	○	*Sinophyllum pendulum simplex* (HUANG, 1932)
○	○	○	○	○	●	○	*Sinophyllum* sp.
○	○	○	○	○	●	○	*Siphonophyllia* ? *lonsdalei* KEYSERLING, 1854
○	○	●	○	○	○	○	*Sociophyllum elongatum* (SCHLÜTER, 1881)
○	○	●	○	○	○	○	*Sociophyllum longiseptatum* (BULVANKER, 1958)
○	○	○	○	○	●	○	*Sociophyllum torosum* (SCHLÜTER, 1881)
○	○	●	○	○	○	○	*Sociophyllum* sp.
○	○	●	○	○	○	○	*Sparganophyllum* sp.
○	○	○	○	○	●	○	*Spongophyllum halisitoides* KODSI, 1971
○	○	●	○	○	○	○	*Stringophyllum buechelense* (SCHLÜTER, 1889)
○	○	●	○	○	○	○	*Stringophyllum isactis* (FRECH, 1886)
○	○	●	○	○	○	○	*Stringophyllum* cf. *isactis* (FRECH, 1886)
○	○	●	○	○	●	○	*Stringophyllum praecursor* (FRECH, 1886)
○	○	○	○	○	●	○	*Stringophyllum primordiale* WEDEKIND, 1922
○	○	●	○	○	○	○	*Stringophyllum* ? *schlüteri* (SCHLÜTER, 1894)
○	○	○	○	○	●	○	*Stringophyllum schwelmense* (WEDEKIND, 1925)
○	○	○	○	○	●	○	*Stringophyllum* sp.
○	○	●	○	○	○	○	*Synaptophyllum* ? *heritschi* SCHOUPPÉ, 1949
○	○	●	○	○	○	○	*Syringaxon* cf. *curtum* (POCTA, 1902)
○	○	●	○	○	○	○	*Syringaxon graecense* SCHOUPPÉ, 1954
●	○	●	○	○	●	○	*Syringaxon* ? *zimmermanni* WEIßERMEL, 1941

○ ○ ● ○ ○ ○ ○	*Syringaxon* sp.
○ ○ ● ○ ○ ○ ○	*"Tabulophyllum" cherychevi* BULVANKER, 1958
○ ○ ○ ○ ○ ● ○	*Tabulophyllum delicatum* SOSHKINA, 1952
○ ○ ○ ○ ○ ● ○	*Tabulophyllum heckeri giveticum* FERRARI, 1968
○ ○ ○ ○ ○ ● ○	*Tabulophyllum* sp.
○ ○ ○ ○ ○ ● ○	*Temnophyllum* cf. *latum* WALTHER, 1928
○ ○ ● ○ ○ ● ●	*Thamnophyllum caespitosum* (GOLDFUSS, 1826)
○ ○ ○ ○ ○ ● ○	*Thamnophyllum* cf. *caespitosum* (GOLDFUSS, 1826)
○ ○ ○ ○ ○ ● ○	*Thamnophyllum carnicum* VINASSA DE REGNY, 1918
○ ○ ● ○ ○ ○ ○	*Thamnophyllum cylindricum* (YOH, 1937)
○ ○ ● ○ ○ ○ ●	*Thamnophyllum germanicum germanicum* SCRUTTON, 1968
○ ○ ● ○ ○ ○ ○	*Thamnophyllum germanicum schouppei* SCRUTTON, 1968
○ ○ ● ○ ○ ● ○	*Thamnophyllum hoernesi* PENECKE, 1894
○ ○ ● ○ ○ ○ ○	*Thamnophyllum murchisoni* PENECKE, 1894
○ ○ ● ○ ○ ○ ○	*Thamnophyllum peneckei* SCHOUPPÉ, 1949
○ ○ ● ○ ○ ○ ○	*Thamnophyllum stachei* PENECKE, 1894
○ ○ ● ○ ○ ● ○	*Thamnophyllum* sp.
○ ○ ○ ○ ○ ● ○	*Thysanophyllum vinassai* HERITSCH, 1936
○ ○ ● ○ ○ ○ ○	*Torusphyllum heterocystis* (PENECKE, 1894)
○ ○ ○ ○ ○ ● ○	*Tryplasma devoniana* (SOSHKINA, 1937)
○ ○ ● ○ ○ ● ○	*Tryplasma devonica* (PENECKE, 1894)
○ ○ ● ○ ○ ○ ○	*Tryplasma cf. fasciculare* (SOSHKINA, 1937)
○ ○ ● ○ ○ ● ●	*Tryplasma hercynica* (ROEMER, 1855)?
○ ○ ○ ○ ○ ● ○	*Tryplasma loveni* (MILNE-EDWARDS & HAIME, 1851)?
○ ○ ○ ○ ○ ● ○	*Tryplasma vermiculare* (WEDEKIND, 1927)?
○ ○ ○ ○ ○ ● ○	*Tryplasma cf. vermiculare* (WEDEKIND, 1927)
○ ○ ● ○ ○ ○ ○	*Tryplasma* sp.
○ ○ ○ ○ ○ ● ○	*Tysanophyllum* sp.
○ ○ ○ ○ ○ ● ●	*Ufimia aster aster* (GRABAU, 1922)
○ ○ ○ ○ ○ ● ●	*Ufimia aster cylindroconica* (SOSHKINA, 1928)
○ ○ ○ ○ ○ ● ○	*Ufimia* cf. *aster* (GRABAU, 1922)
○ ○ ○ ○ ○ ● ●	*Ufimia exceptata* (SOSHKINA, 1928)
○ ○ ○ ○ ○ ○ ●	*Ufimia* ? *gracilis* (HERITSCH, 1934) (nomen nudum)
○ ○ ○ ○ ○ ● ○	*Wentzelophyllum arminae* (FELSER, 1937)
○ ○ ○ ○ ○ ● ○	*Wentzelophyllum felseri* MINATO & KATO, 1963
○ ○ ○ ○ ○ ● ○	*Wentzelophyllum volzi alpha* (HUANG, 1932)
○ ○ ○ ○ ○ ● ○	*Wentzelophyllum volzi volzi* (HUANG, 1932)
○ ○ ○ ○ ○ ● ○	*Yokoyamella (Yokoyamella) carinthica* (HERITSCH, 1936)
○ ○ ○ ○ ○ ● ●	*Yokoyamella (Yokoyamella) stillei* (HERITSCH, 1936)
○ ○ ○ ○ ○ ● ○	*Yokoyamella (Yokoyamella) yokoyamai* (OZAWA, 1925)
○ ○ ○ ○ ○ ● ○	*Zaphrentis omaliusi* MILNE-EDWARDS & HAIME, 1851?
● ○ ● ○ ○ ● ○	*Zaphrentis* sp.
○ ○ ○ ○ ○ ● ○	*Zaphrentoides* sp.
○ ○ ○ ○ ○ ● ○	*Zeliaphyllum suessi* HERITSCH, 1936
○ ○ ○ ○ ○ ● ○	*Zeliaphyllum* cf. *suessi* HERITSCH, 1936
○ ○ ● ○ ○ ○ ○	*Zelophyllia* ? *cornuvaccinum* (PENECKE, 1894)
○ ○ ○ ○ ○ ● ○	*Zelophyllia tabulata* (SOSHKINA, 1937)
○ ○ ○ ○ ○ ● ○	*Zonophyllum* sp.

Tabulates (including Heliolitids)

1	2	3	4	5	6	7	
○	○	○	○	○	●	○	*"Actinopora" carnica* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*"Actinopora" proasteriscus* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Alveolites* cf. *reticulatus* STEININGER, 1834
○	○	○	○	○	●	○	*Alveolites collinensis* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Alveolites* cf. *collinensis* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Alveolites crinalis* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Alveolites irregularis* GORTANI, 1913
○	○	○	○	○	●	○	*Alveolites* cf. *irregularis* GORTANI, 1913
○	○	○	○	○	●	○	*Alveolites minor* GORTANI, 1913
○	○	●	○	○	○	○	*Alveolites minutus* LECOMPTE, 1939
○	○	○	○	○	●	○	*Alveolites* n. sp. aff. *reticulatus* STEININGER, 1834
○	○	○	○	○	●	○	*Alveolites reticulatus* STEININGER, 1834
○	○	●	○	○	●	●	*Alveolites suborbicularis* LAMARCK, 1801
○	○	○	○	○	●	○	*Alveolites volaicus* CHARLESWORTH, 1914
○	○	●	○	○	●	○	*Alveolites* sp.
○	○	●	○	○	○	○	*Antholites* sp.
○	○	●	○	○	○	○	*Aulopora conglobata* GOLDFUSS, 1829
○	○	●	○	○	●	○	*Aulopora minor* GOLDFUSS, 1829
○	○	●	○	○	●	○	*Aulopora serpens minor* GOLDFUSS, 1829
○	○	●	○	○	●	○	*Aulopora serpens serpens* GOLDFUSS, 1829
○	○	●	○	○	●	○	*Aulopora* sp.
○	○	○	○	○	●	○	*Beaumontia guerangeri* MILNE-EDWARDS & HAIME, 1851
○	○	○	○	○	●	○	*Caliapora battersbyi* (MILNE-EDWARDS & HAIME, 1851)
○	○	○	○	○	●	○	*Caliapora* ? *carnica* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*Caliapora frechi* (CHARLESWORTH, 1914)
○	○	○	○	○	●	○	*"Caliapora" heritschi* SCHOUPPÉ, 1954
○	○	○	○	○	●	○	*Caliapora* ? *julica* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*"Caliapora labechii"* MILNE-EDWARDS & HAIME, 1851
○	○	○	○	○	●	○	*Caliapora* cf. *labechii* MILNE-EDWARDS & HAIME, 1851?
○	○	○	○	○	●	○	*Caliapora* sp.
○	○	○	○	○	●	○	*Cladochonus macrostomus* ROEMER
●	○	○	○	○	○	○	*Cladochonus michelini* MILNE-EDWARDS & HAIME, 1851
○	○	●	○	○	○	○	*Cladochonus* sp.
○	○	○	○	○	●	○	*Cladopora vermicularis* (MCCOY, 1850)
○	○	●	○	○	○	○	*Cladopora* sp.
○	○	○	○	○	●	○	*Coenites carnicus* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Coenites lonsdalei* (D'ORBIGNY, 1850)
○	○	○	○	○	●	○	*Coenites mariae* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Coenites polonica* GUERICH, 1896
○	○	○	○	○	●	○	*Coenites volaicus* (CHARLESWORTH, 1914)
○	○	●	○	○	○	○	*Coenites* sp.
○	○	●	○	○	●	○	*Favosites alpinus* PENECKE, 1894
○	○	●	○	○	○	○	*Favosites bohemicus* (BARRANDE, 1865)
○	○	●	○	○	○	○	*Favosites eifelensis* NICHOLSON, 1879
○	○	●	○	○	○	○	*Favosites* aff. *eifelensis* NICHOLSON, 1879
○	○	○	○	○	●	○	*Favosites fidelis clavatus* POCTA, 1902
○	○	○	○	○	●	○	*Favosites fidelis fidelis* BARRANDE, 1902
○	○	●	○	○	●	○	*Favosites forbesi forbesi* MILNE-EDWARDS & HAIME, 1851
○	○	○	●	○	●	○	*Favosites forbesi nitidulus* POCTA in BARRANDE, 1902
○	○	○	○	○	●	○	*Favosites* cf. *forbesi* MILNE-EDWARDS & HAIME, 1851
○	○	○	○	○	●	○	*Favosites forojuliensis forojuliensis* VINASSA DE REGNY, 1918
○	○	○	○	○	●	○	*Favosites forojuliensis pinnatus* VINASSA DE REGNY, 1918
○	○	○	○	○	●	●	*Favosites goldfussi* D'ORBIGNY, 1850?
○	○	○	○	○	●	○	*Favosites* aff. *goldfussi* D'ORBIGNY, 1850

○	○	●	○	○	●	○	*Favosites gothlandicus aberrans* REGNELL, 1941
○	○	●	○	○	○	○	*Favosites* cf. *gothlandicus aberrans* REGNELL, 1941
○	○	●	○	○	○	○	*Favosites gothlandicus* cf. *aberrans* REGNELL, 1941
○	○	○	○	○	●	●	*Favosites gothlandicus gothlandicus* LAMARCK, 1816
○	○	●	○	○	○	○	*Favosites* cf. *gothlandicus* LAMARCK, 1816
○	○	●	○	○	○	○	*Favosites graffi* PENECKE, 1894
○	○	●	○	○	○	○	*Favosites* cf. *graffi* PENECKE, 1894
○	○	●	○	○	○	○	*Favosites grandis* HERITSCH, 1937
○	○	●	○	○	○	○	*Favosites grandis* n.ssp. [sensu] BOROVICENY & FLÜGEL, 1962
○	○	○	○	○	●	○	*Favosites* cf. *gregalis* PORFIRIEV, 1937
○	○	●	○	○	○	○	*Favosites hisingeri* MILNE-EDWARDS & HAIME, 1851
○	○	○	○	○	●	○	*Favosites italicus* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Favosites mailieuxi* (LECOMPTE, 1939)?
○	○	●	○	○	●	○	*Favosites* n. sp. aff. *styriacus styriacus* BOROVICZENY & FLÜGEL, 1962
○	○	●	○	○	○	○	*Favosites* ? *radiciformis* (QUENSTEDT, 1881?)
○	○	●	○	○	○	○	*Favosites* ?cf. *radiciformis* (QUENSTEDT, 1881?)
○	○	●	○	○	○	○	*Favosites robiniaefolius* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Favosites styriacus perforatus* SCHOUPPE, 1954
○	○	●	●	○	●	○	*Favosites styriacus styriacus* PENECKE, 1894
○	○	●	○	○	●	○	*Favosites* cf. *styriacus* PENECKE, 1894
○	○	○	○	○	●	○	*Favosites tachlowitzensis* BARRANDE, 1902
○	○	○	○	○	●	○	*Favosites thildae* ANGELIS D'OSSAT, 1901
○	○	○	○	○	●	○	*Favosites volaicus* VINASSA DE REGNY, 1918
●	○	●	●	○	●	●	*Favosites* sp.
○	○	○	○	○	●	○	*Fossopora devonica* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*Heliolites lindströmi* ANGELIS D'OSSAT, 1899
●	○	○	○	○	●	○	*Heliolites* ? *multiporus* CERRI, 1931
○	○	○	○	○	●	○	*Heliolites porosus* (GOLDFUSS, 1826)
○	○	●	○	○	○	○	*Heliolites* ? *praeporosus* KETTNEROVA, 1933
○	○	○	○	○	●	○	*"Heliolites" vesiculosus* PENECKE, 1887
○	○	●	○	○	●	○	*Heliolites* sp.
○	○	○	○	○	●	○	*Helioplasma kolihai* KETTNEROVA, 1933
○	○	●	○	○	○	○	*Hemiplasmopora paucitabulata* (FLÜGEL, 1956b)
○	○	●	○	○	○	○	*Hemiplasmopora peneckei* (FLÜGEL, 1963)
○	○	●	○	○	●	○	*Hemiplasmopora* cf. *peneckei* (FLÜGEL, 1963)
○	○	●	○	○	●	○	*Incedilites spongodes spongodes* (LINDSTRÖM, 1899)
○	○	●	○	○	○	○	*Incedilites spongodes* ssp.
○	○	○	○	○	●	○	*Mcleodea conferta* (Milne-Edwards & Haime, 1851)
○	○	○	○	○	●	○	*Mcleodea minima* (LINDSTRÖM, 1899)
○	○	○	○	○	●	○	*Multithecopora syrinx* (ETHERIDGE, 1900)
○	○	○	○	○	●	○	*Okopites devonicus* (ANGELIS D'OSSAT, 1899)
○	○	●	○	○	●	●	*Pachycanalicula barrandei* (PENECKE, 1887)
○	○	○	○	○	●	○	*Pachycanalicula* aff. *barrandei* (PENECKE, 1887)
○	○	●	○	○	○	○	*Pachycanalicula* cf. *barrandei* (PENECKE, 1887)
○	○	○	○	○	●	○	*Pachyfavosites cronigerus* (D'ORBIGNY)
○	○	○	○	○	●	○	*Pachyfavosites polymorphus* (GOLDFUSS, 1826)
○	○	○	○	○	●	○	*Pachyfavosites* sp.
○	○	○	○	○	●	○	*"Pachypora" coralloides* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Pachypora* sp. aff. *dilacerata"* POCTA in BARRANDE, 1902
○	○	●	○	○	●	○	*Paraheliolites hanusi* (KETTNEROVA, 1933) ?
○	○	●	○	○	●	○	*Paraheliolites minimus* (CERRI, 1931)
○	○	●	○	○	●	○	*Paraheliolites turcicus* (WEIßERMEL, 1939)
○	○	●	○	○	○	○	*Petridictyum petrii* (MAURER, 1874)
○	○	○	○	○	●	○	*Plasmopora carnica* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Platyaxum (Roseoporella) taenioforme gracile* HUBMANN, 1991

○	○	●	○	○	○	○	*Pleurodictyum* sp.
○	○	○	○	○	●	○	*Protomichelinia abichi* (WAAGEN & WENTZEL, 1886)
○	○	○	○	○	●	○	*Pseudoplasmopora confinensis* (CHARLESWORTH, 1914)
○	○	●	○	○	○	○	*Remesia tubaeformis* (GOLDFUSS, 1829)
○	○	●	○	○	○	○	*Roemeria infundibulifera* (GOLDFUSS, 1829)
○	○	○	○	○	●	○	*Scoliopora denticulata* MILNE-EDWARDS & HAIME, 1851
○	○	○	○	○	●	○	*Scoliopora* sp.
○	○	●	○	○	○	○	*Sideriolites repletus* (LINDSTRÖM, 1899)
○	○	●	○	○	○	○	*Squameoalveolites robustus* (PRADACOVA, 1938)
○	○	○	○	○	●	○	*Squameoalveolites* sp.
○	○	○	○	○	●	○	*"Striatopora" angustior* GUERICH, 1896
○	○	○	○	○	●	○	*"Striatopora" gortanii* VINASSA DE REGNY, 1910
○	○	○	○	○	●	○	*"Striatopora" major* VINASSA DE REGNY, 1918
○	○	●	○	○	●	○	*Striatopora* ? *subaequalis* (MILNE-EDWARDS & HAIME, 1851)
○	○	○	○	○	●	○	*Striatopora* ? aff. *subaequalis* (MILNE-EDWARDS & HAIME, 1851)
○	○	○	○	○	●	○	*Striatopora* ? cf. *subaequalis* (MILNE-EDWARDS & HAIME, 1851)
○	○	●	○	○	○	○	*Striatopora* ? *suessi* PENECKE, 1894
○	○	●	○	○	○	○	*Striatopora* ? cf. *suessi* PENECKE, 1894
○	○	●	●	○	●	○	*Striatopora* sp.
●	○	○	○	○	○	○	*Syringocystis eifelensis* (SCHLÜTER, 1889)?
○	○	○	○	○	●	○	*Syringopora alpina* CHARLESWORTH, 1914
○	○	○	○	○	●	○	*"Syringopora" carnica* VINASSA DE REGNY, 1918
○	○	●	○	○	○	○	*Syringopora expansa* MAURER, 1885
○	○	○	○	○	●	○	*Syringopora fascicularis* (LINNÉ, 1767)
○	○	●	○	○	○	○	*Syringopora hilberi* PENECKE, 1894
○	○	○	○	○	●	○	*Syringopora reticulata* GOLDFUSS, 1826
○	○	○	○	○	●	○	*Syringopora samarensis* STUCKENBERG, 1905
○	○	●	○	○	○	○	*Syringopora schulzei* PENECKE, 1894
○	○	●	○	○	○	○	*Syringopora* sp. aff. *schulzei* PENECKE, 1894
○	○	●	○	○	●	○	*Syringopora* sp.
○	○	○	○	○	●	○	*Sytovaelites crassiseptatus* (FLÜGEL, 1956)
○	○	●	○	○	○	○	*Thamnopora boloniensis* (GOSSELET, 1877)
○	○	●	○	○	●	○	*Thamnopora cervicornis* (DE BLAINVILLE, 1830)
○	○	●	○	○	○	○	*Thamnopora dubia* (DE BLAINVILLE, 1830)
○	○	●	○	○	○	○	*Thamnopora* ? *gigantea* (PENECKE, 1894)
○	●	●	○	○	○	○	*Thamnopora reticulata* (DE BLAINVILLE, 1830)
○	●	●	○	○	○	○	*Thamnopora* cf. *reticulata* (DE BLAINVILLE, 1830)
○	○	●	○	○	●	○	*Thamnopora vermicularis* (MCCOY, 1850)
○	●	●	○	○	●	●	*Thamnopora* sp.
○	○	●	○	○	○	○	*Trachypora* sp.