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Marine Carbonate Facies in Response to Climate and Nutrient Level: The Upper Carboniferous and Permian of Central Spitsbergen (Svalbard)

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Contents

SUMMARY

Carbonate-dominated successions of the Gipsdalen and Tempelfjorden Groups from Svalbard record a significant shift from Photozoan to Heterozoan particle associations in neritic settings during the late Palaeozoic. During the Bashkirian, benthic particle associations which included photoautotrophs such as phylloid algae (Chloroforam Association) characterised shallow subtidal environments. Most depositional settings which endured siliciclastic terrestrial input exhibited poorly diversified associations dominated by brachiopods, bryozoans and siliceous sponges (Bryonoderm Association). During the Moscovian to Asselian, highly diversified associations typified by various calcareous green algae, *Palaeoaplysina*, *Tubiphytes*, fusulinids, smaller and encrusting foraminifers (Chloroforam Association) prevailed in carbonate sediments from supratidal to shallow subtidal environments. During the Sakmarian and Early Artinskian, oolitic carbonate sands (Chloroforam Association) typified intertidal flats, whereas shallow subtidal environments were occupied by moderately diversified associations with fusulinids, smaller foraminifers, echinoderms and bryozoans (Bryonoderm-extended Association) and poorly diversified associations with echinoderms, brachiopods and bryozoans (Bryonoderm Association). During the Late Artinskian to Kazanian, poorly diversified associations characterised by brachiopods, echinoderms and bryozoans (Bryonoderm Association), and sponge-dominated associations (Hyalosponge Association) reigned within siliceous carbonates of intertidal and shallow subtidal environments.

This trend is interpreted as a result of climatic cooling and fluctuations of prevailing levels of trophic resources within shallow-water settings during the studied time period. While raised nutrient levels were restricted to near-shore settings during the Bashkirian, steady mesotrophic conditions arose from the Sakmarian onward and increased to late Permian times.

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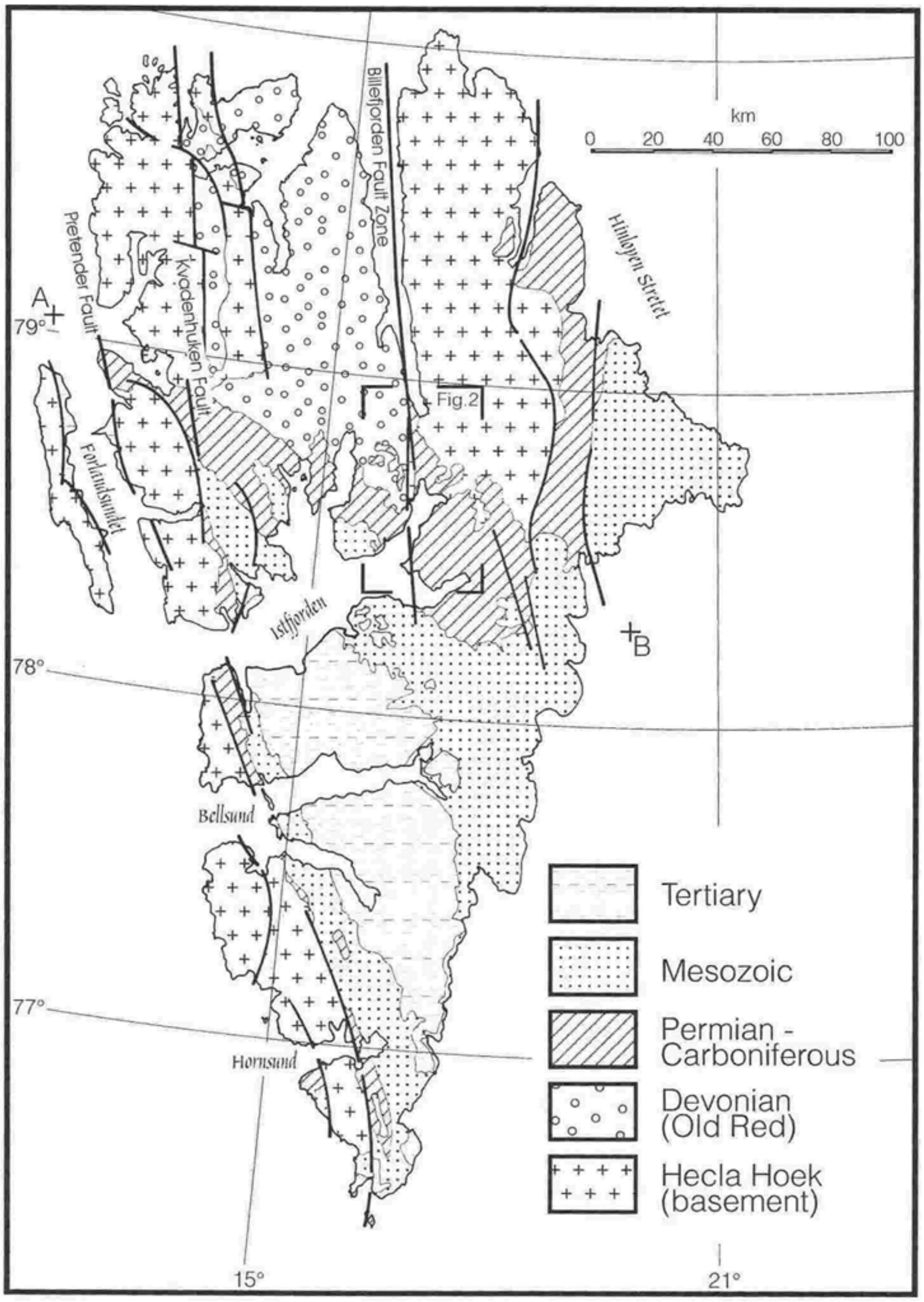


Fig.1. Simplified geologic map of Spitsbergen modified after Dallmann et al. (1999). Letters A and B indicate location of the cross section shown at Fig. 3.

1 INTRODUCTION

The archipelago of Svalbard is the emergent northwestern part of the Barents Sea Shelf, which was uplifted by late-Mesozoic and Cenozoic crustal movements (Fig. 1). The area provides a representative insight into Carboniferous and Permian sediments of the Barents Sea region that were deposited along the northern margin of Pangaea during late Palaeozoic time. These sediments contrast with Palaeotethyan carbonates that were formed around the palaeoequator (e.g. Flügel 1980, Krainer 1992).

The investigated strata of central Spitsbergen have been the focus of a large number of studies (McWhae 1953, Gee et al. 1953, Forbes et al. 1958, Cutbill and Challinor 1965, Winsnes 1966, Malkowski and Hoffmann 1979, Lønøy 1981, 1995, Sundbø 1982, Skaug et al. 1982, Dons 1983, Steel and Worsley 1984, Nakamura et al. 1987, Nilsson 1988, 1993, Dallmann 1993, Ezaki et al. 1994, Johansen 1994, Pickard et al. 1996, Stemmerik 1997, Samuelsberg and Pickard 1999, Buggisch et al. 2001). Changing patterns of sediment thickness, facies configurations, provenance and transport directions clearly reflect the varying tectonic regime during the late Palaeozoic (Steel and Worsley 1984, Johannessen and Steel 1992, Pickard et al. 1996). During the Carboniferous, Spitsbergen developed from a site of fault block tectonism with different subsidence rates and predominant siliciclastic-evaporitic sedimentation to a stable carbonate shelf that experienced uniform subsidence. These

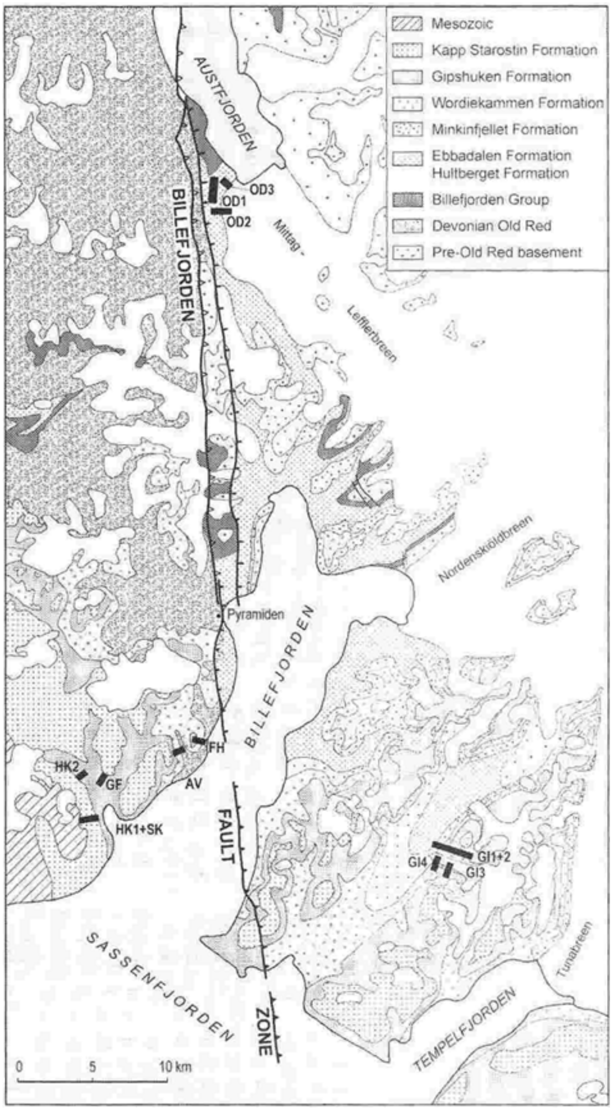


Fig. 2. Simplified geologic map of central Spitsbergen (modified after Dallmann et al. 1999) showing the locations of sample sites. The sections at Skansbukta are located at the S' flank of Gyntfjellet (GF), NE' flank of Högskulefjellet (HK 2) and E' flank of Högskulefjellet (SK, HK 1). The sections at Alvrekdalen are at the SE' flank of Feyling-Hanssen-Fjellet (FH) and NE' flank of Skansen (AV). The sections at Boltonbreen (Gipsdalen) are exposed on the SW' flank of Pyefjellet (GI 1, GI 2). N' flank of Balchinfjellet (GI 3) and NE' flank of Balchinfjellet (GI 4). The sections at Odellfjellet are on the N' flank (OD 1) and E' flank of Odellfjellet (OD 2, OD 3).

conditions lasted until the late Permian when siliciclastic influx revived again. The Upper Carboniferous and Lower Permian strata display a highly cyclic rock succession which is composed of four long-term transgressive-regressive cycles (Samuelsberg and Pickard 1999). These long-term cycles are themselves composed of stacked higher order cycles (Stemmerik et al. 1995, Pickard et al. 1996, Samuelsberg and Pickard 1999).

Climatic cooling, partly attributed to a northward drift of Pangaea, had a great influence on stratigraphical variations and facies development (Steel and Worsley 1984, Worsley et al. 1986, Stemmerik 2000). Palaeolatitude increased from around 25°N in the middle Carboniferous to 50°N during the late Permian (Steel and Worsley 1984, Scotese and McKerrow 1990, Stemmerik 2000), causing a change from early tropical to later temperate conditions. Early Carboniferous grabens were the sites of fluviatile deposition under tropical climatic conditions with coal swamps developed on floodplains and around lakes, whereas red beds, carbonate and sabkha sequences reflect arid climates during the middle Carboniferous and early Permian. A regional transgression during the middle Permian is accompanied by the establishment of a temperate climate in the Svalbard area.

The late Palaeozoic sedimentation in the Svalbard/Barents Sea is quite comparable with that in the Wandel Sea Basin in northeastern Greenland (Håkansson and Stemmerik 1984, Stemmerik and Worsley 1995, Stemmerik 1997, Stemmerik et al. 1995), on the Finnmark Carbonate Platform in the Norwegian Barents Sea (Bruce and Toomey 1993, Ehrenberg et al. 1998, Samuelsberg 1999) and in the Sverdrup Basin in the Canadian Arctic (Beauchamp 1994, Morin et al. 1994, Beauchamp and Desrochers 1997). These areas were originally situated not more than about 1000 km from what is now the west coast of Spitsbergen prior to the break-up of Greenland and Europe and the opening of the Arctic and North Atlantic Oceans.

The present study is part of a research program focused on late Palaeozoic sediments of the St. Johnsfjorden Trough, the Nordfjorden High and the Billefjorden Trough (Fig. 3). Data on conodont stratigraphy have already been published (Buggisch et al. 2001). The main objective of the present paper is to describe and interpret carbonate-dominated facies in the Bashkirian to Kazanian strata of the Nordfjorden High and the Billefjorden Trough. These sediments are part of the Gipsdalen and the Tempelfjorden Group (Cutbill and Challinor 1965, Dallmann et al. 1999). Emphasis is placed on individual facies description and the interpretation of different palaeoenvironments. Time-related depositional facies models are proposed using late Palaeozoic biotic and abiotic benthic particle associations as defined by Beauchamp (1994) (Fig. 8). This paper documents the long-term depositional evolution from Photozoan to Heterozoan carbonates and discusses relevant environmental controls. The cyclic organisation of the investigated successions gives the opportunity to study subtidal, intertidal and supratidal environments at repeated, closely-spaced points in time.

2 REGIONAL GEOLOGY

The basement of Svalbard (Hecla Hock) comprises both late Precambrian and Cambro-Silurian strata. This complex was consolidated during the Caledonian orogeny. Subsequent uplift and extension resulted in the formation of restricted N-S trending grabens during the Devonian and in the formation of the Central Basin of NW Spitsbergen. Devonian siliciclastic deposits (Old Red) were partly deformed during the Svalbardian compression (Vogt 1929).

The sediment accumulation during Carboniferous and Permian times is related to the middle Carboniferous rifting

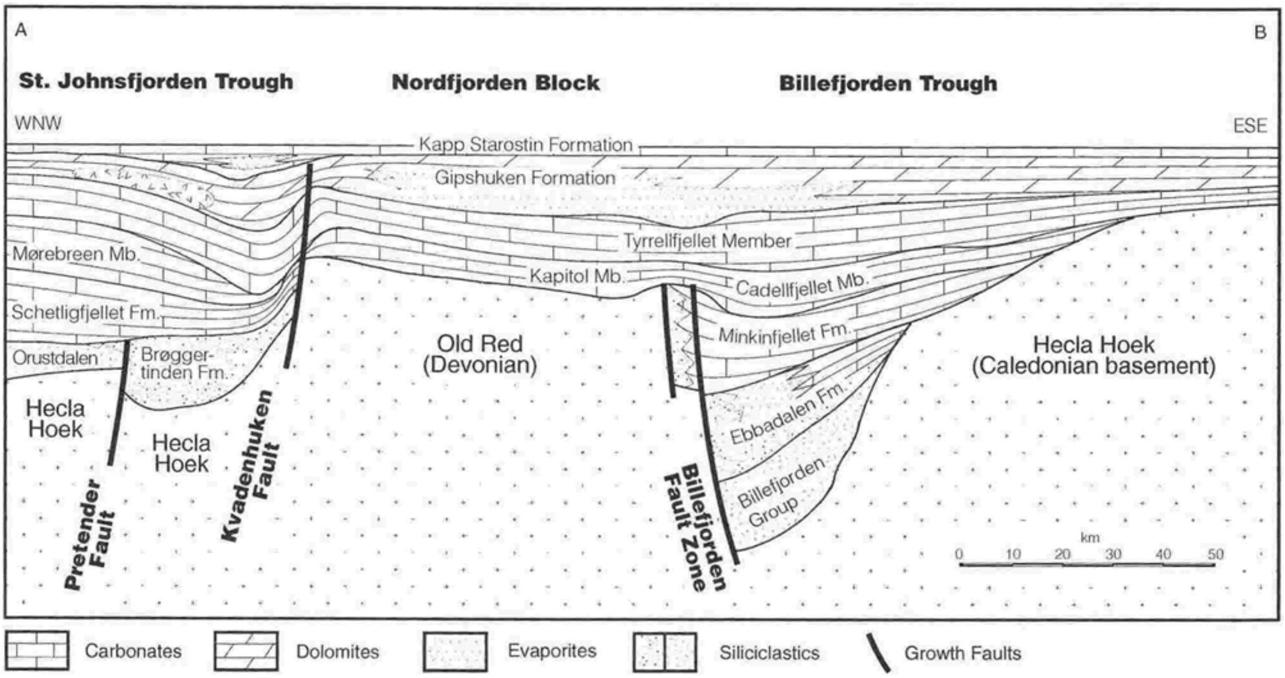


Fig.3. West to east cross section of central Spitsbergen showing the late Palaeozoic horst, graben and halfgraben configuration. Adapted from Wright (in Heafford 1988) and modified after Ludwig (1989) and Johannessen and Steel (1992).

on Svalbard (Fig. 3). Johannessen and Steel (1990, 1992) distinguish three tectonostratigraphic units:

(i) Coal-bearing fluvial and lacustrine sequences of the Billefjorden Group were deposited on flood plains during the "early rift phase".

(ii) The "main rift phase" led to the formation of a horst (Nordfjorden Block), graben, and halfgraben (Billefjorden Trough) topography. Coarse-grained alluvial fan- and sabkha-sediments interfingering with marine sandstones and carbonates of the Ebbadalen Formation were deposited in the subsiding depressions. The narrow and mostly asymmetric geometry of the NNW-SSE trending rift basins is well illustrated in the Billefjorden area.

(iii) During the "late rift and post rift phase" downwrapping was more gentle and subsidence became more and more balanced. Platform carbonates, evaporites and rarely siliciclastics were deposited in extended areas of the Barents Shelf (Minkinfjellet Formation, Wordiekammen Formation, Gipshuken Formation and Kapp Starostin Formation).

The Mesozoic was the time of more or less stable platform conditions (Steel and Worsley 1984). The development of a mobile belt (Western Spitsbergen Alpine Fold Belt) and of sedimentary basins (Central Basin, Forelandsundet Graben) during the Tertiary were probably closely related to the separation of Greenland and Svalbard.

The late Palaeozoic succession deposited across central Spitsbergen normally shows a cyclic organisation of stratification and facies. While local tectonic base-level adjustments mainly controlled the rift-infill sequences in the Billefjorden Trough during the Serpukhovian to Bashkirian (Johannessen and Steel 1992), eustatic sea-level fluctuations more and more controlled the late-rift and post-rift carbonate accumulation (Stemmerik et al. 1995, Pickard et al. 1996, Samuelsberg and Pickard 1999). Upper Carboniferous to Lower Permian strata display four long-term (> 10 Ma) transgressive-regressive cycles (Samuelsberg and Pickard 1999): (1) The oldest transgressive-regressive sequence is of Serpukhovian to Bashkirian age and represents a syn-rift sequence. (2) Also composed of syn-rift sediments is the transgressive-regressive Moscovian-Gzhelian sequence. (3) The late Gzhelian to late Asselian sequence is mainly a post-rift sequence. (3) The youngest sequence is of Sakmarian to possible Artinskian age, and is also composed of post-rift sediments.

The long-term cycles are themselves composed of stacked higher order cycles (Stemmerik et al. 1995, Pickard et al. 1996, Samuelsberg and Pickard 1999). These short-term cycles consists mainly of transgressive-regressive, dominantly shoaling-upward carbonate sequences of widely varying composition and thickness (e.g. Fig. 12). Many cycles are capped by a subaerial exposure surfaces. Local tectonic activity is believed to control to some extent the development of short-term cycles in the syn-rift succession (Samuelsberg and Pickard 1999), and the cyclicity of the in the post-rift succession could have a greater degree of autogenic control (Pickard et al. 1996). However, Late Carboniferous to early Permian cyclicity appears to be a word-wide phenomenon. It has been documented from other areas of the Barents Sea and Greenland (Stemmerik et al. 1995, Stemmerik and Worsley 2000), the Southern Alps (Krainer 1992) and the U.S. Mid-continent (Boardman and Heckel 1989, Goldhammer et al. 1994). The period was characterised by global ice house conditions, with major

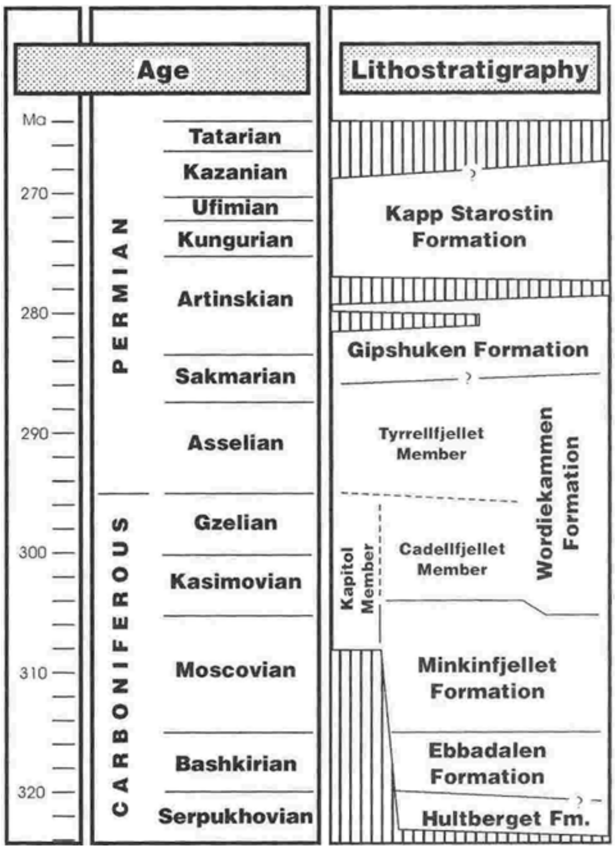


Fig. 4. Biostratigraphy of the Carboniferous and Permian lithostratigraphic units from the Billefjorden Trough and the Nordfjorden Block. Modified from Cutbill and Challinor (1965), Dallmann et al. (1999) and Buggisch et al. (2001). Radiometric ages (Ross et al. 1994) are shown on the left.

glaciations in the southern hemisphere and high-frequency, high-amplitude sea-level fluctuations (Ross and Ross 1985, Veevers and Powell 1987). But the cyclicity of the Artinskian-Kazanian successions, which is separated from the underlying sediments by a subaerial exposure surfaces (Dallmann et al. 1999), is related to low-frequency, low-amplitude sea-level fluctuations (Stemmerik 2000).

3 METHODS

The bulk of information used in this study comes from field observations and description of 11 outcrop localities, representing approximately 2000 m of strata, along the fjord cliffs at Billefjorden and Austfjorden and from the Gipsdalen area (Fig. 2). The authors collected the samples during the joint Erlangen-Jena-Greifswald expedition in 1991 run by W. Buggisch. Buildups were physically traced along the extension of outcrop exposure in order to document the geometry of stratal units. More than 450 samples were collected for lab examination as polished slabs and thin-sections. A total of about 500 thin-sections, stained with alizarin red, were described using terminology established by Folk (1959, 1962), Dunham (1962), Embry and Klovan (1972) and the classification of benthic particle associations by Beauchamp (1994). In the following descriptions, particle concentrations are provided on a semi-quantitative scale including six subdivisions: absent, rare, occasional, common, frequent and abundant, representing 0%, 1-20%, 20-40% 40-60%, 60-80%, >80%, respectively.

4 LITHOSTRATIGRAPHIC FRAMEWORK

4.1 Ebbadalen Formation (Gipsdalen Group)

The Ebbadalen Formation (Cutbill and Challinor 1965) was interpreted as an unconformity-bounded syn-rift succession (Holliday and Cutbill 1972, Johannessen and Steel 1992) that was deposited in an asymmetric, elongated basin on the downthrown side of the Billefjorden Fault Zone (Fig. 3). A more than 700 m thick sedimentary sequence is developed along the western part of the Billefjorden Trough, thinning to only a few tens of metres at the eastern edge of the basin. According to Johannessen and Steel (1990, 1992), the Ebbadalen Formation consists of a transgressive fining-upward sequence at the base (Ebbaelva Member), with a cyclic, fault-bound marginal sequence (Odellfjellet Member) and a basinal sequence on top (Trikolorfjellet Member). Marine carbonates were deposited during repeated rapid submergence. They are overlain by coarsening-upward sequences of alluvial sandstones along the coastal plain, by fanglomerates at the western, fault controlled margin and by regressive evaporites in the basin.

According to conodonts (Buggisch et al. 2001) and fusulinids (Holliday and Cutbill 1972), the onset of accumulation of the Ebbadalen Formation (Fig. 4) started during the early Bashkirian or even earlier (see Johannessen and Steel 1992). In our study, limestone and dolomite interbeds of the Ebbadalen Formation have been sampled at Odellfjellet (NE' flank) near the Billefjorden Fault Zone.

4.2 Minkinfjellet Formation (Gipsdalen Group)

The 150-400 m thick Minkinfjellet Formation is developed only in the Billefjorden Trough (Cutbill and Challinor 1965, Dallmann et al. 1999). According to Johannessen and Steel (1992) it represents the early post-rift platform sediments (Fig. 3). Samuelsberg and Pickard (1999) interpret the succession as the transgressive part of the late syn-rift sediment sequence. The lower boundary is an angular unconformity on a regional scale (Dallmann et al. 1999). The Minkinfjellet Formation consists at the base of reddish, cross-bedded, mica-rich sandstones that are interpreted as basal transgressive shore deposit overlying evaporites of the Ebbadalen Formation (Lauritzen et al. 1989). Proximal parts of the Billefjorden Trough – in respect to the distance from the Billefjorden Fault Zone – are characterised by a variable succession of siliciclastic rocks, anhydritic evaporites and dolomites, whereas limestones dominate in distal parts of the trough towards the east.

The age of the Minkinfjellet Formation (Fig. 4) is Moscovian to early Kasimovian (Pickard et al. 1996). According to conodonts (Buggisch et al. 2001), accumulation in the western part of the trough lasted from Late Morrowian/Early

Atokanian (~ late Bashkirian/early Moscovian) to Late Desmoinesian (~ late Moscovian). In our study an incomplete section (about 120 m) of the Minkinfjellet Formation has been measured at Boltonbreen (SW' flank of Pyefjellet) showing the facies development over the Billefjorden Trough (Fig. 5).

4.3 Wordiekammen Formation (Gipsdalen Group)

The Wordiekammen Formation (Gee et al. 1953) represents post-rift platform sediments which overlie and onlap both successions of mid-Carboniferous age on troughs and older rocks on highs (Johannessen and Steel 1992). Although tectonic activity ceased and a general rise of sea level resulted in a transgressive carbonate-dominated succession, its thickness and sedimentary facies variation are still influenced by the structural horst/halfgraben configuration (Lauritzen et al. 1989). While the Nordfjorden Block persisted as a high with continuous carbonate sedimentation (Kapitol Member), carbonates, dolomites and interstratified siliciclastic carbonates were deposited within the Billefjorden Trough (Cadellfjellet Member). Both Members are overlain by the Tyrrellfjellet Member (Fig. 3).

4.3.1 Kapitol Member

The Kapitol Member (Cutbill and Challinor 1965) is up to 100 m thick and restricted to the Nordfjorden Block (Fig. 3). It rests unconformably on Devonian sandstones (Lauritzen et al. 1989) and correlates with the upper part of the Minkinfjellet Formation and the Cadellfjellet Member of the Billefjorden Trough. In the west the basal part is made up of mica-rich varicoloured sandstones and local conglomerates overlain by a succession predominantly of limestones.

The age of the Kapitol Member (Fig. 4) is late Moscovian through Gzhelian (Cutbill and Challinor 1965, Nilsson and Davydov 1997). In our study, an incomplete section of the Kapitol Member has been measured at Alvrekdalen (SE' flank Feyling-Hanssen-Fjellet) with a thickness of about 50 m, showing the facies development over the Billefjorden Fault Zone (Fig. 6).

4.3.2 Cadellfjellet Member

The up to 250 m thick Cadellfjellet Member (Cutbill and Challinor 1965) forms the lowermost part of the Wordiekammen Formation in the Billefjorden Trough and on eastern Spitsbergen (Fig. 3). A dark, cliff-forming limestone unit of 21 to 50 m thickness, the so-called "Black Crag Beds", marks the base of this member in the western part of the Billefjorden Trough, which is useful for mapping. Further to the east (e.g., at Boltonbreen) marine limestones with algal bioherms form the base of the Cadellfjellet Member (Pyefjellet Beds; Pickard et al. 1996, Dallmann et al. 1999). However, these beds are probably situated at a lower stratigraphic level than the Black Crag Beds. The Cadellfjellet Member represents a single regressive succession (Sundsbø 1982, Lauritzen et al. 1989, Pickard et al. 1996). Open marine carbonate cycles present in the Pyefjellet Beds are succeeded by inner platform, open marine to restricted hypersaline sediments of the Black Crag Beds. The Mathewbreen Beds which form the uppermost part of the member contain evaporites and dolomites and appear to represent intertidal and supratidal settings. According to Samuelsberg and Pickard (1999) the Cadellfjellet Member is part of a late syn-rift sediment sequence (Fig. 3). The facial arrangement within the Billefjorden Trough in sub-parallel facial zones, which was typical during the deposition of the Minkinfjellet Formation, still existed until the end of the Carboniferous, when the Nordfjorden Block was finally flooded (Steel and Worsley 1984, Pickard et al. 1996). Restricted lagoonal environments behind barrier and shoal systems formed by bioclastic banks and algal bioherms are typical for the western part of the trough, whereas open marine platform carbonates accumulated in its eastern parts.

According to fusulinids the age of the member is Kasimovian to Gzhelian (Nilsson and Davydov 1997). This is true for western parts of the Billefjorden Trough, but in its eastern parts the accumulation of the Pyefjellet Beds already started during the Late Moscovian (Pickard et al. 1996). In the present study, sections of the Cadellfjellet Member (about 60 m) have been measured at Boltonbreen (SW' flank of Pyefjellet and NE' flank of Balchinfjellet) showing the facies development over the Billefjorden Trough (Fig. 5).

4.3.3 Tyrrellfjellet Member

The Tyrrellfjellet Member (Cutbill and Challinor 1965) covers the Mörebreen Member of the St. Johnsfjorden Trough, the Kapitol Member of the Nordfjorden Block and the Cadellfjellet Member of the Billefjorden Trough (Fig. 3). Obviously, it represents post-rift sediments (Johannessen and Steel 1992, Samuelsberg and Pickard 1999) and reaches a thickness up to 250 m in Billefjorden Trough (Lauritzen et al. 1989). The base was redefined by Dallmann et al. (1999) to coincide with the base of the distinct Brucebyen Beds, consisting of dark fusulinid limestones with a thickness of up to 10 m. *Palaeoaplysina* bioherms are widespread in the lower part of the member. They form isolated mounds up to 6 m high and tabular mound complexes up to 15 m thick at more than three stratigraphic levels (Skaug et al. 1982). Furthermore, Lauritzen et al. (1989) mentioned bryozoan buildups occurring simultaneously. As suggested by Steel and Worsley (1984), *Palaeoaplysina* bioherms tended to develop along the eastern and western margins of the Nordfjorden Block. The facial interpretation of these bioherms points to a lagoonal shelf environment, whereas the associated dark bituminous limestones suggest restricted lagoonal conditions (Skaug et al. 1982, Lauritzen et al. 1989). The upper part of the Tyrrellfjellet Member is dominated by a succession of well-bedded limestones and dolomites which mainly consist of bioclastic packstones to wackestones and a diverse fossil assemblage. A further distinct, cliff-forming marker horizon, named "Limestone B" by Gee et al. (1953), occurs on the top of the Tyrrellfjellet Member.

An Asselian and early Sakmarian age is assigned to the Tyrrellfjellet Member (Keilen 1991) which is in accordance with conodont data (Buggisch et al. 2001). According to fusulinids (Nilsson and Davydov 1997), the accumulation of the member already started during the late Gzhelian, and the lower boundary between the Tyrrellfjellet Member and the underlying Cadellfjellet and Kapitol Members is regionally diachronous. Within the Billefjorden Trough at Boltonbreen, the accumulation of the member obviously started somewhat later than on the Nordfjorden Block. The development of laterally extensive surfaces of discontinuity and intraformational conglomerates at its base point to a minor uplift of the entire area near the Carboniferous/Permian boundary (Steel and Worsley 1984, Pickard et al. 1996). Nevertheless, renewed transgression during the early Permian led to the re-establishment of open marine environments in most areas.

In our study sections of the Tyrrellfjellet Member were measured at Boltonbreen (SW' flank of Pyefjellet and N' flank of Balchinfjellet) with a thickness of about 190 m and at Alvrekdalen (SE' flank Feyling-Hanssen-Fjellet and NE' flank Skansen) with a thickness of 165 m. These sections represent the facies development over the Billefjorden Trough and over the eastern margin of the Nordfjorden Block along the Billefjorden Fault Zone (Fig. 5, Fig. 6).

4.4 Gipshuken Formation (Gipsdalen Group)

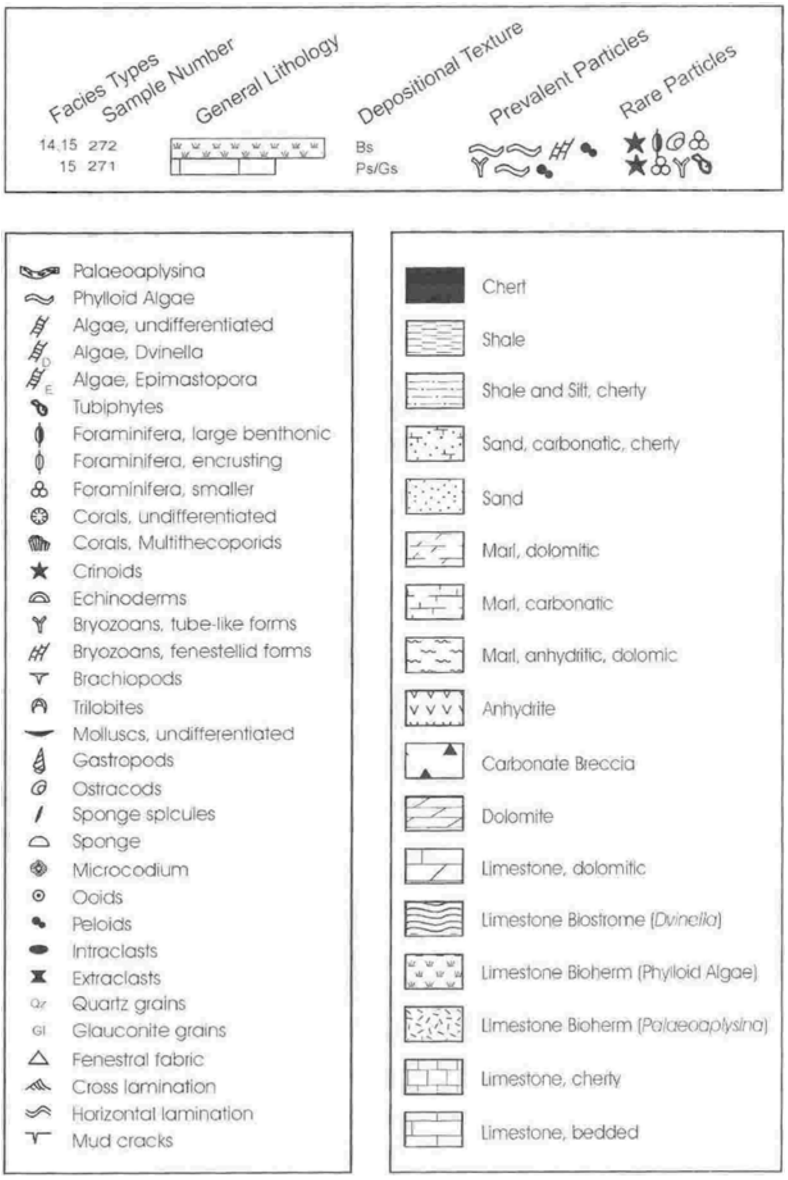
The Gipshuken Formation (Cutbill and Challinor 1965) is characterised by evaporites and associated sediments such as anhydrite, gypsum and dolomite in the lower part, and peritidal dolomites in the upper part. According to Samuelsberg and Pickard (1999) it represents mainly the regressive part of a broad post-rift sequence. Lauritzen (1981) interprets the up to 350 m thick Gipshuken Formation as superimposed sabkha cycles overlain by lagoonal, algal limestones.

According to conodonts derived from the underlying Tyrrellfjellet Member and the overlying basal part of the Kapp Starostin Formation, the age of the Gipshuken is Late Sakmarian to Early Artinskian (Buggisch et al. 2001). This conclusion is in accordance with other data (Nakrem et al. 1992). The base of the Gipshuken Formation, which is commonly defined by the occurrence of the first continuous anhydrite bed (Lauritzen et al. 1989), is possibly diachronous. The eastward progradation of the sabkha sequence across the Nordfjorden Block (Steel and Worsley 1984) exhibits a regressive sequence during the Late Sakmarian that is followed by deposition of a transgressive sequence during the Early Artinskian. The area of the Billefjorden Trough was a site of restricted marine sedimentation. Stemmerik (1997, 2000) suggests a stratigraphic hiatus between the Gipshuken Formation and the Kapp Starostin Formation comprising parts of the Artinskian (Fig. 4).

In the present study, sections have been measured at Skansbukta (NE' flank Högskulefjellet and S' flank Gyntfjellet) and at the basal part of the Formation at Alvrekdalen (NE' flank Skansen) documenting the facies development over the eastern margin of the Nordfjorden Block and the Billefjorden Fault Zone, respectively (Fig. 6, Fig. 7).

4.5 Kapp Starostin Formation (Tempelfjorden Group)

The Kapp Starostin Formation (Cutbill and Challinor 1965) extends all over Svalbard, covering older units of the



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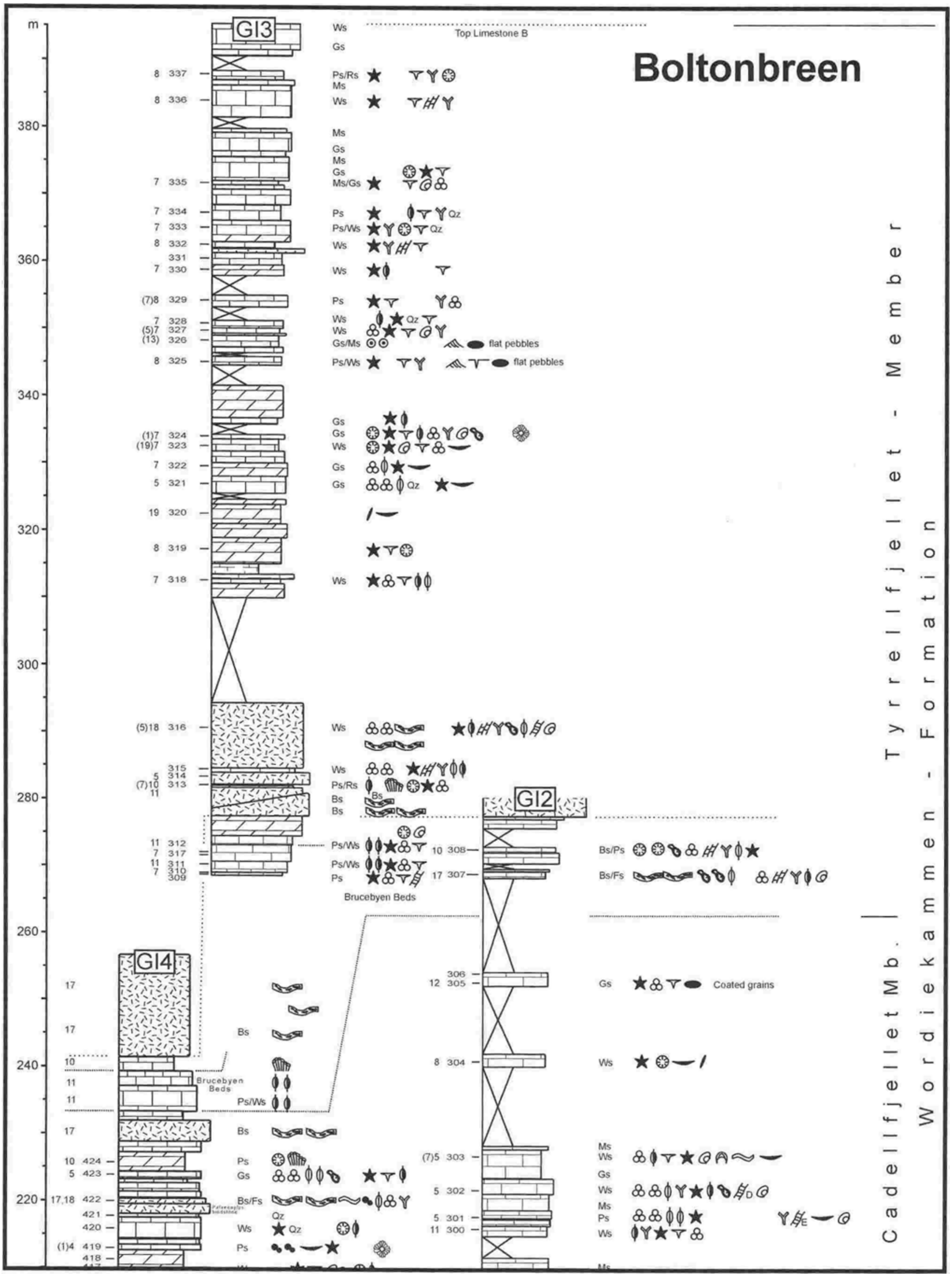


Fig. 5a. Lithology, microfacies and lithostratigraphy of the Moscovian to Sakmarian strata at Boltonbreen. Sections have been measured at the SW' flank of Pyefjellet (GI 1, GI 2), the N' flank of Balchinfjellet (GI 3) and the NE' flank of Balchinfjellet (GI 4).

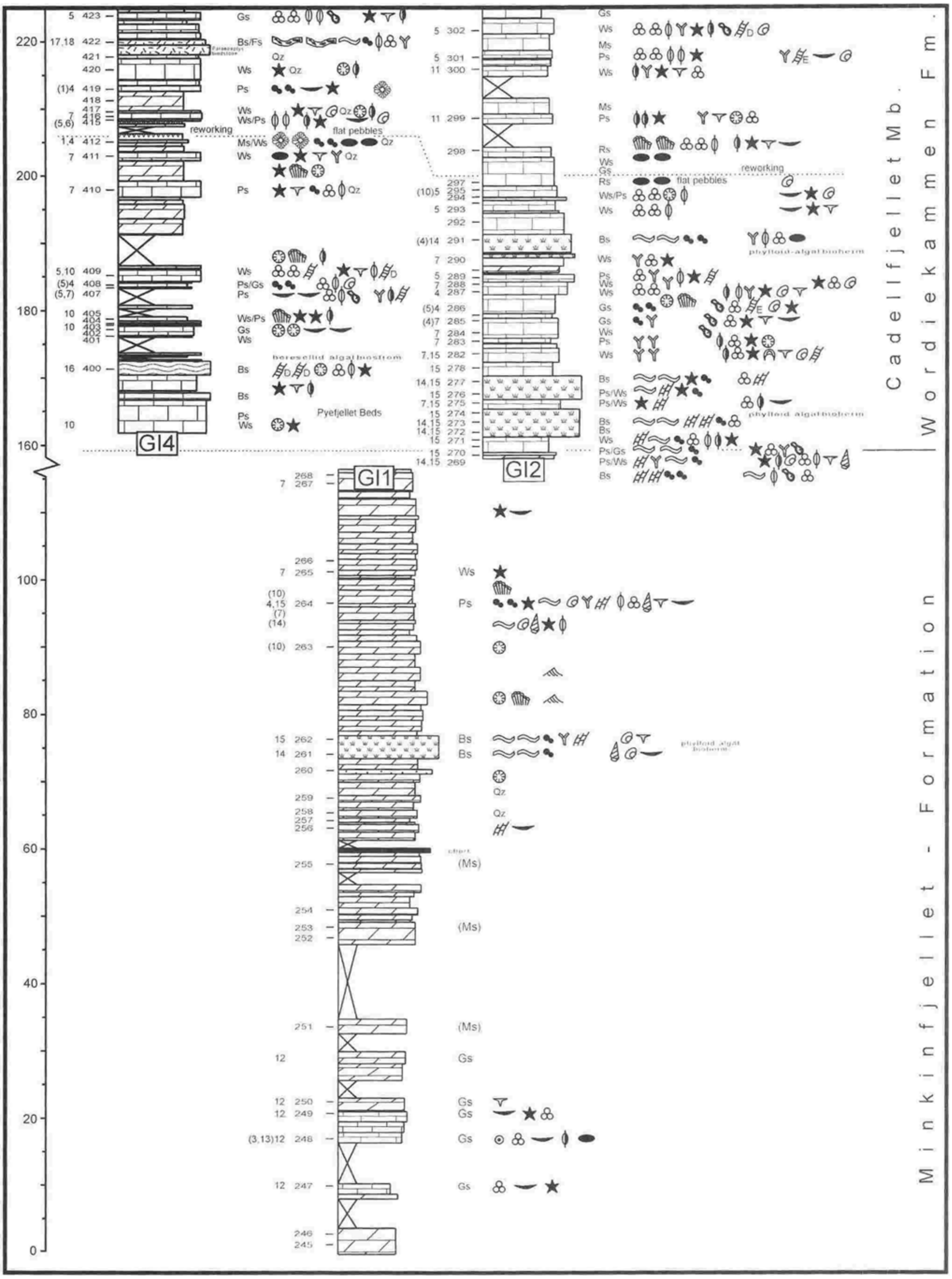


Fig. 5b. Lithology, microfacies and lithostratigraphy of the Moscovian to Sakmarian strata at Boltonbreen. Sections have been measured at the SW' flank of Pyefjellet (GI 1, GI 2), the N' flank of Balchinfjellet (GI 3) and the NE' flank of Balchinfjellet (GI 4).

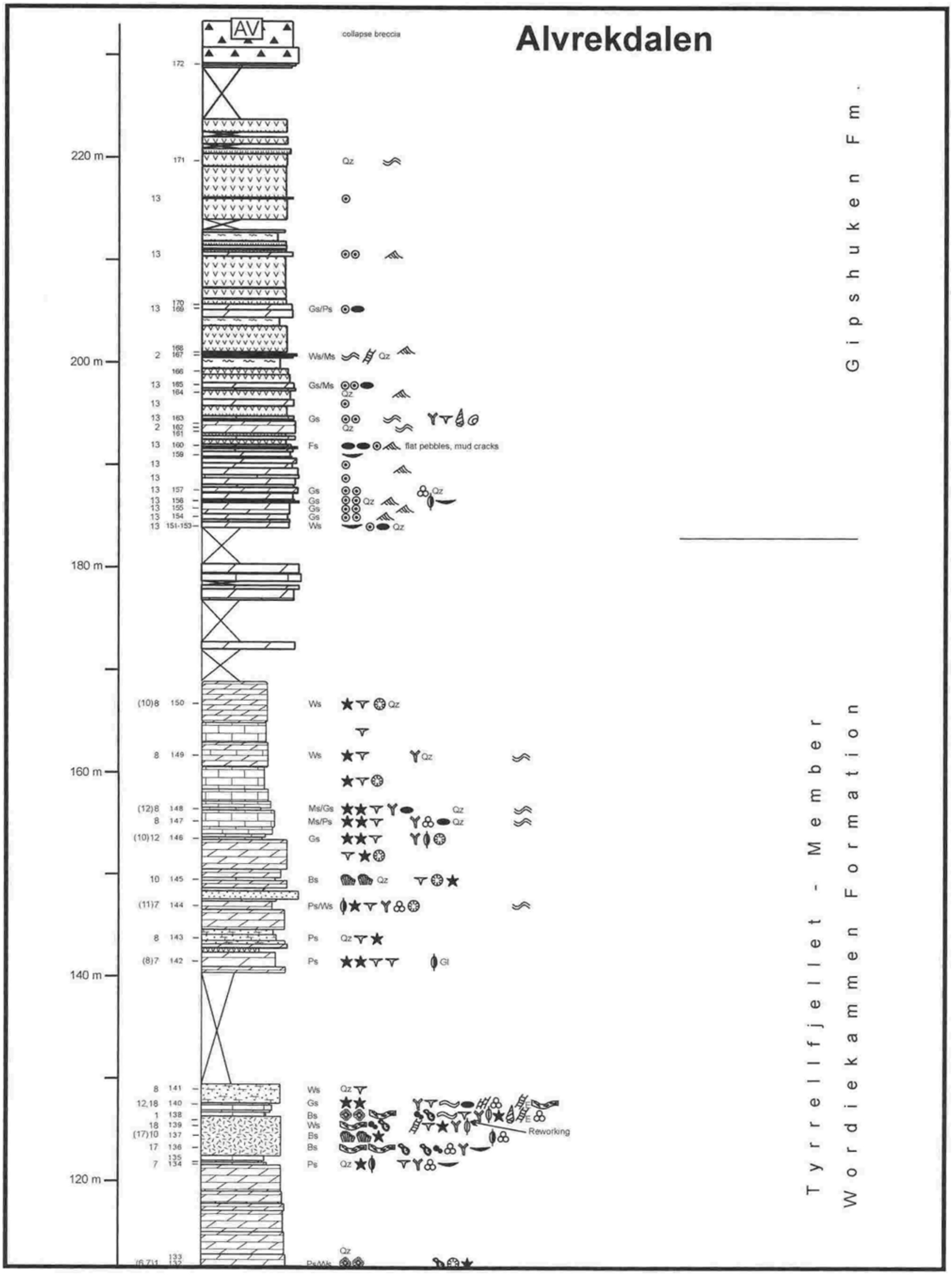


Fig. 6a. Lithology, microfacies and lithostratigraphy of the Moscovian to Early Artinskian strata at Alvrekdalen. Sections have been measured at the SE' flank of Feyling-Hanssen-Fjellet (FH) and the NE' flank of Skansen (AV).

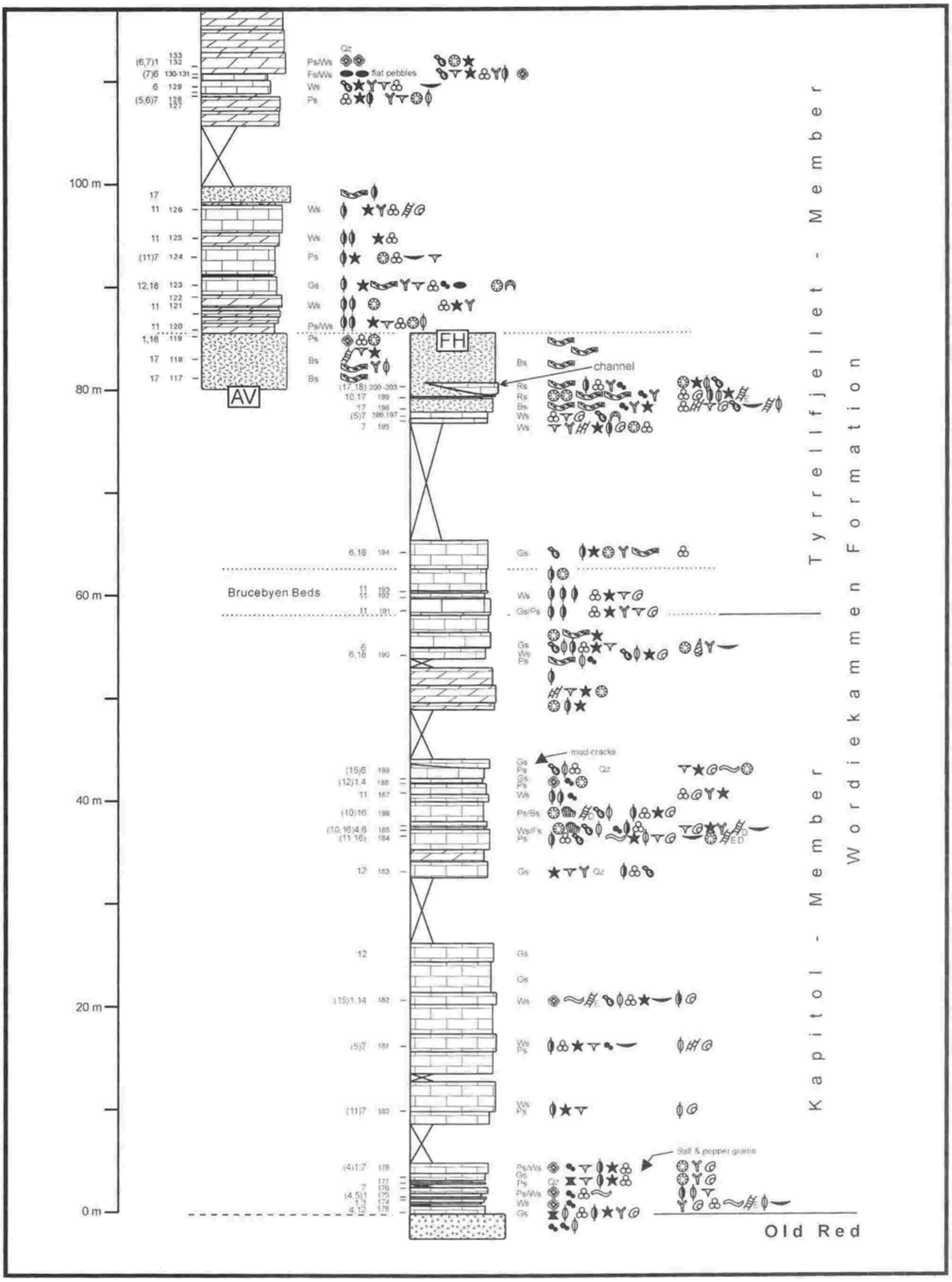


Fig. 6b. Lithology, microfacies and lithostratigraphy of the Moscovian to Early Artinskian strata at Alvrekdalen. Sections have been measured at the SE' flank of Feyling-Hanssen-Fjellet (FH) and the NE' flank of Skansen (AV).

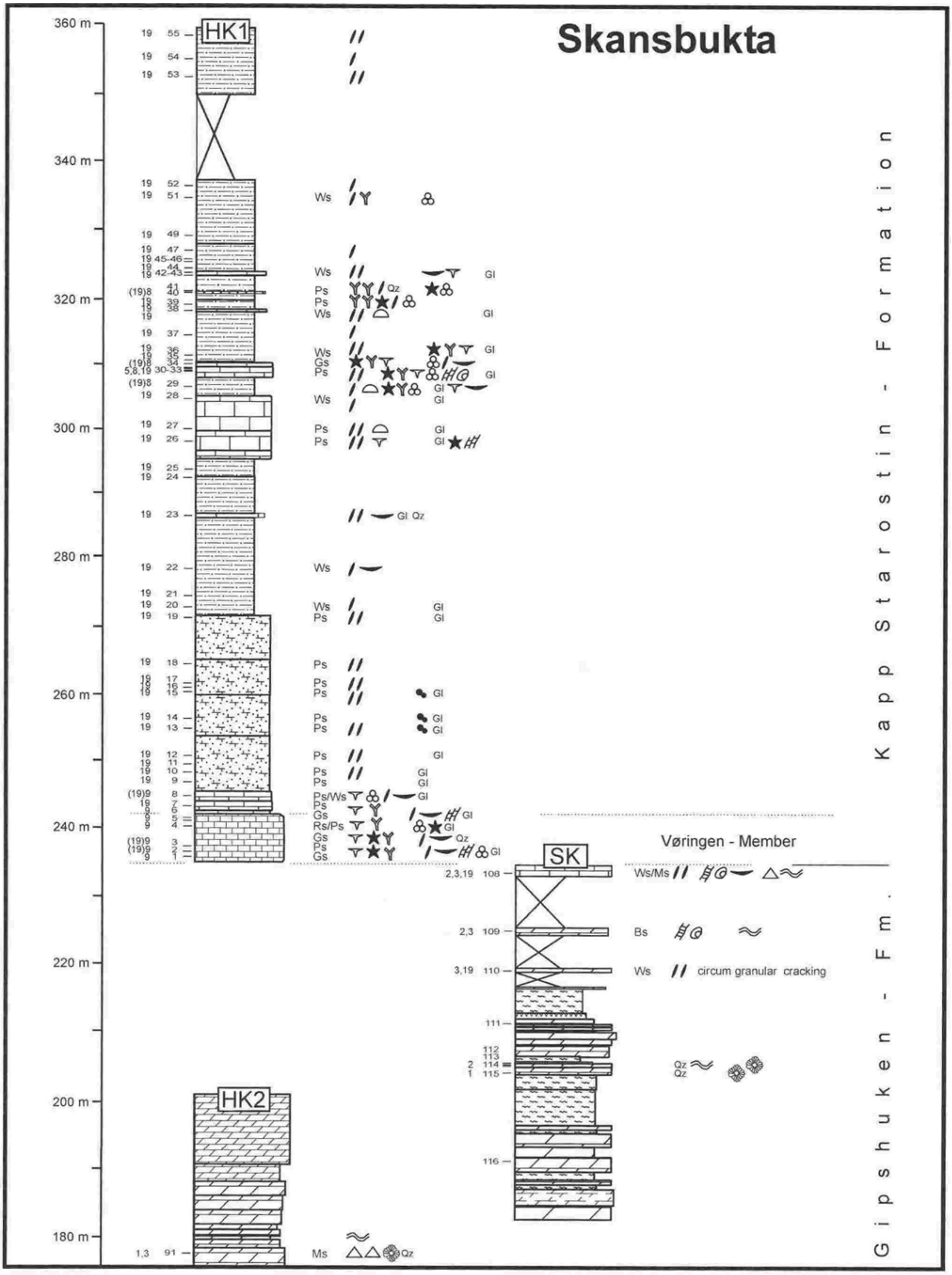


Fig. 7a. Lithology, microfacies and lithostratigraphy of the Sakmarian to Kazanian strata at Skansbukta. Sections have been measured at the S' flank of Gyntfjellet (GF), the NE' flank of Högskulefjellet (HK 2) and the E' flank of Högskulefjellet (SK, HK 1).

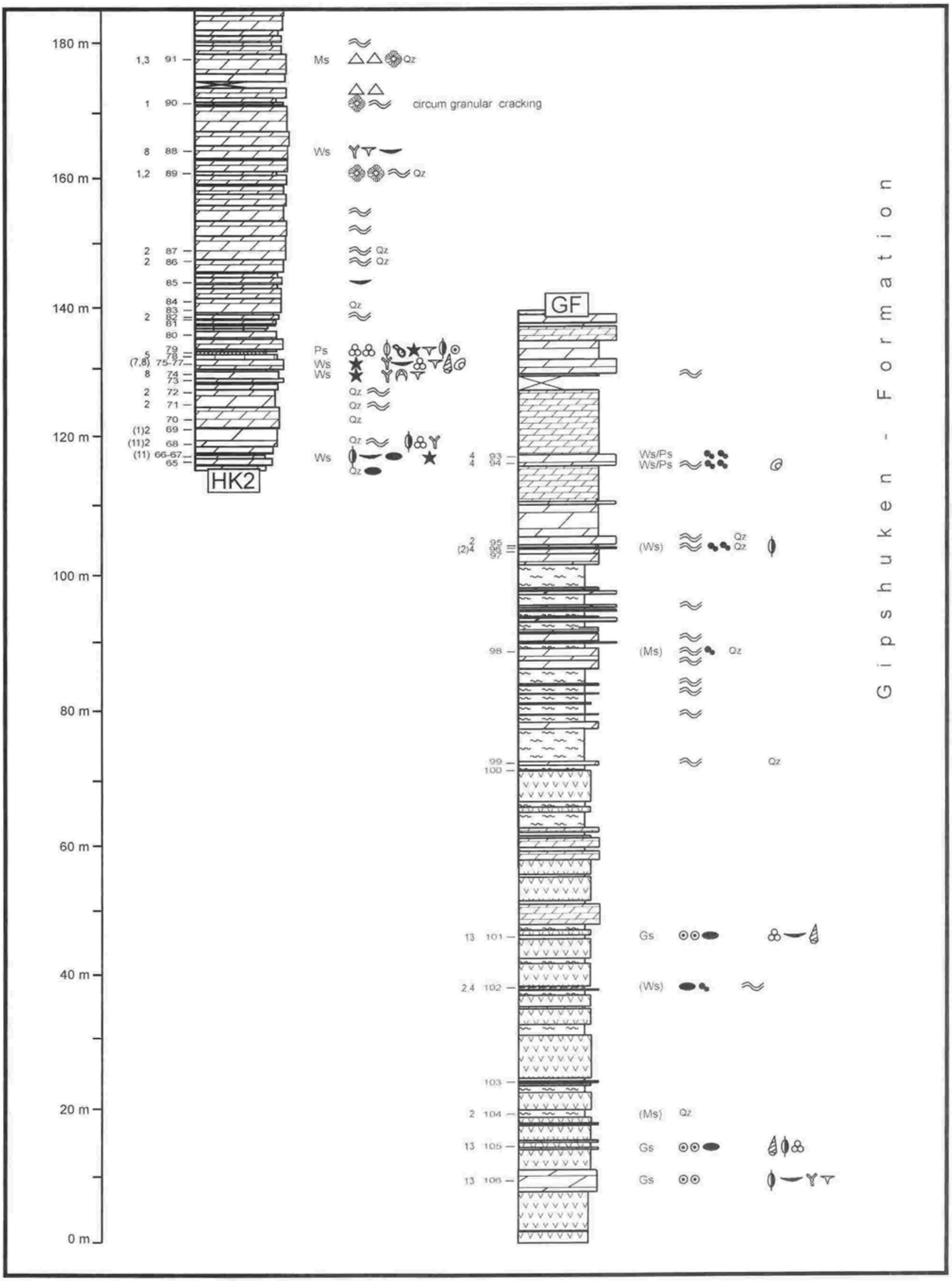


Fig. 7b. Lithology, microfacies and lithostratigraphy of the Sakmarian to Kazanian strata at Skansbukta. Sections have been measured at the S' flank of Gyntfjellet (GF), the NE' flank of Högskulefjellet (HK 2) and the E' flank of Högskulefjellet (SK, HK 1).

St. Johnsfjorden Trough in the west, on the Nordfjorden Block and the Billefjorden Trough in the east (Fig. 3). The up to 460 m thick fossiliferous marine sediments wedge toward the Hornsund High in the south, where they are represented by 10 m of cherty brachiopod limestones (Steel and Worsley 1984).

The basal beds of the formation are built up by bioclastic limestones which represent shoreface deposits, formed by the transgression of barrier sequences across the restricted marine platform and sabkha deposits of the Gipshuken Formation below (Hellem 1981, Steel and Worsley 1984). The succession grades upwards into shales and siltstones interstratified by a few limestone beds and some glauconitic sandstones at the top. Due to a high content of siliceous sponges, all lithologies are more or less silicified and locally pass into bedded cherts. Cherts and spiculitic shales represent deep shelf deposits (Lauritzen et al. 1989). Massive chert bodies are interpreted as sponge buildups by Steel and Worsley (1984), and interbedded units of bioclastic or sandy shoals are most common over the western part of the Billefjorden Trough.

According to conodonts (Szaniawski and Malkowski 1979, Buggisch et al. 2001) the basal boundary of the Kapp Starostin Formation is regionally diachronous. It is attributed to a stratigraphic age of Middle Leonardian and Late Leonardian / Early Rodian, respectively, which is consistent with the diachronous onset of the open marine transgression during the late Artinskian (Fig. 4). The accumulation of the Kapp Starostin Formation lasted until the end of the Permian, although the biostratigraphically dated uppermost beds of the formation are probably not younger than Kazanian (see Steel and Worsley 1984, Nakamura et al. 1987, Stemmerik 1988, Nakrem et al. 1992).

In the present study the basal 120 m of the Kapp Starostin Formation have been measured at Skansbukta (E' flank Högskulefjellet). This section shows the facies development over the Nordfjorden Block (Fig. 7).

5 MICROFACIES TYPES

5.1 *Microcodium* Aggregates

Description: *In situ* aggregates of *Microcodium* grains include spherical clusters of prisms radiating from a central hollow (Pl. 18/4) and sheet-like arrangements of prismatic grains with long axes perpendicular to layers. The size of individual *Microcodium* grains ranges from 100 µm to 400 µm in lenght and from 30 µm to 80 µm in widths. The shape varies from well-defined prisms to vague ellipsoidal or subspherical outlines. Transverse sections show hexagonal or subspherical outlines. Curved faces, mostly convex to outside, tend to be more common than straight faces. The individual calcite grains are yellow to pale brown with dark inclusions concentrated within a peripheral zone. Sweeping extinction within individual grains shows a complete radial-fibrous structure with fibres radiating from a point on the perimeter.

Occurrence: Within the Kapitol Member (late Moscovian) *in situ Microcodium* Aggregates are common at several horizons. Fig. 12/C illustrates a level that is defined by laterally extensive discontinuity surfaces and intraformational conglomerates (see Steel and Worsley 1984). *Microcodium* aggregates also occur at the top of some *Palaeoaplysina* bioherms and in the beds overlying the bioherms within the Tyrrellfjellet Member (Asselian and Sakmarian). Furthermore, *Microcodium* Aggregates are common at several levels within the upper part of the Gipshuken Formation at Skansbukta (Early Artinskian)

Interpretation: *Microcodium* is reinterpreted by Klappa (1978) as being the result of calcification of mycorrhizae,

Plate 16 Moscovian to Gzhelian (late Carboniferous) platform carbonates from Svalbard: Minkinfjellet Formation, Cadellfjellet and Kapitol Members (Wordiekammen Formation) of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light.

Fig. 1. Peloidal grainstone. Some peloids are similar to spherical encrustations around coccoid biotic forms. Sample 286, Boltonbreen (GI 2). x 77

Fig. 2. Foraminiferal packstone. Globular and single-layered apterrinellid foraminifers are the most common skeletal fragments. Sample 415, Boltonbreen (GI 4). x 77

Fig. 3. Colony of multithecoporid corals. Sample 298, Boltonbreen (GI 2). x 2.7

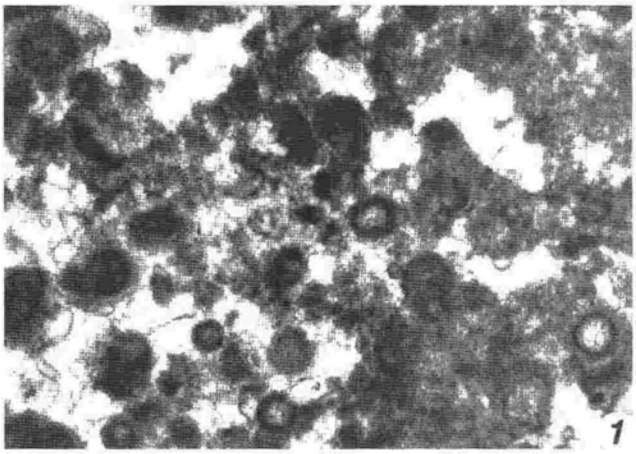
Fig. 4. Bioclastic Packstone/Wackestone Facies. Abundant small foraminifers indicate facial transition to the Small-Foraminifera Facies. Sample 181, Alvrekdalen (FH). x 19

Fig. 5. Beresellid Boundstone Facies. *Tubiphytes* contributes a lot to the sediment stabilisation. Note cross sections of *Tuberitina* and *Bradyina*. Sample 184, Alvrekdalen (FH). x 19

Fig. 6. Beresellid Boundstone Facies. Photomicrograph showing encrusting *Tuberitina* and abundant small foraminifers (such as *Polytaxis*) and *Ammovertella* within the trapped sediment. Sample 185, Alvrekdalen (FH). x 19

Fig. 7. Cylindrical thalli of the beresellid algae *Dvinella* in longitudinal and cross sections. Close-up view of 8. Sample 185, Alvrekdalen (FH). x 77

Fig. 8. Beresellid Boundstone Facies. Abundant beresellid algae are locally encrusted by *Tubiphytes* and *Tuberitina*. Interstices are filled with baffled sediment and cement. Sample 185, Alvrekdalen (FH). x 19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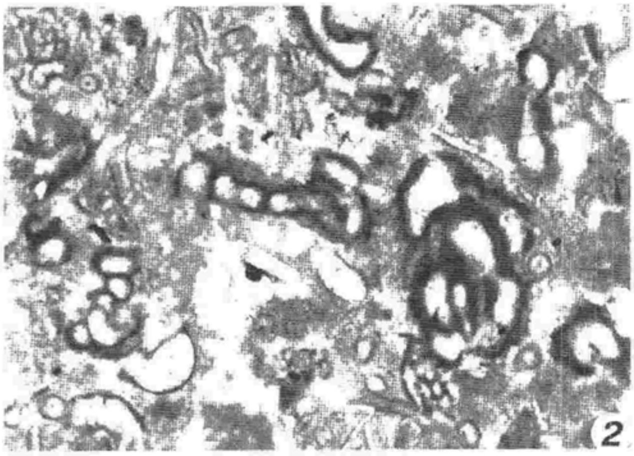
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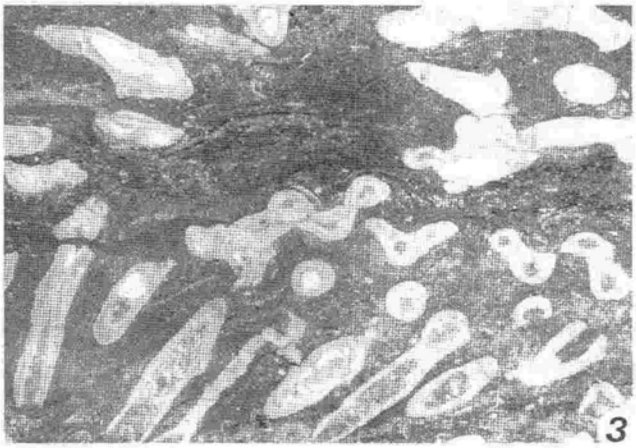
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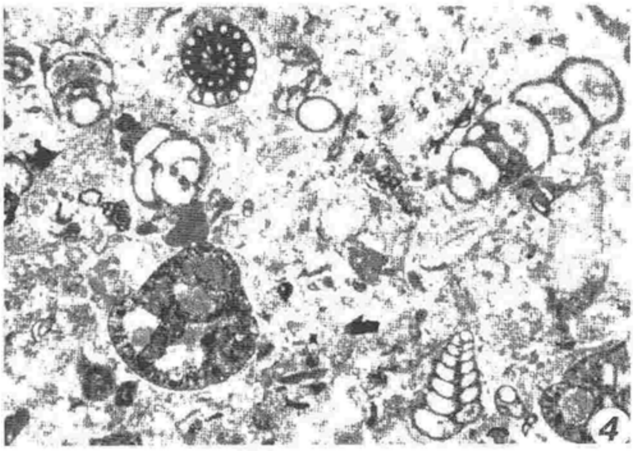
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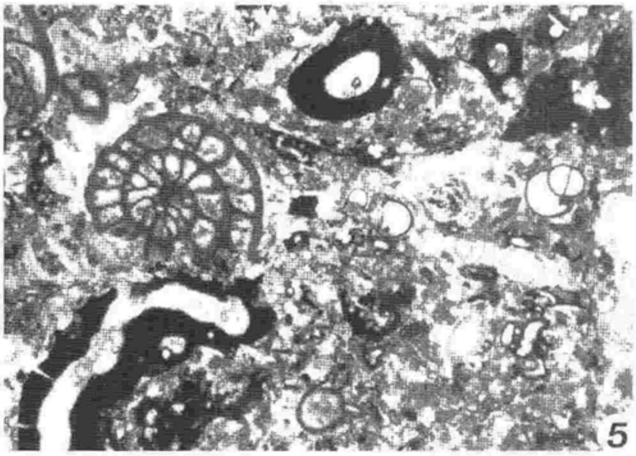
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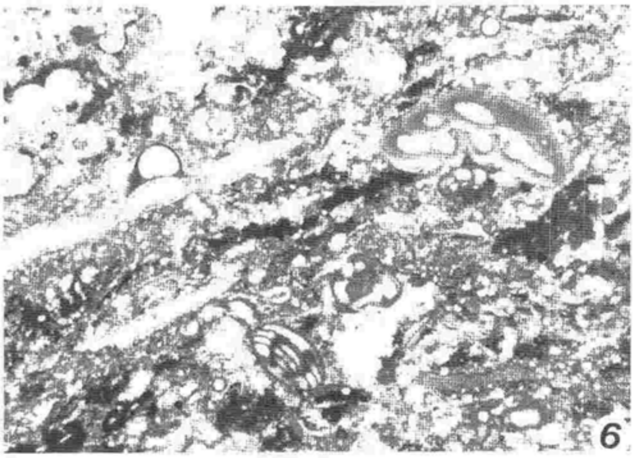
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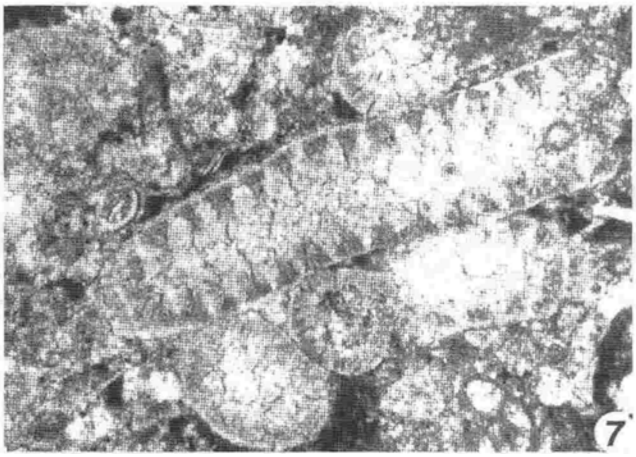
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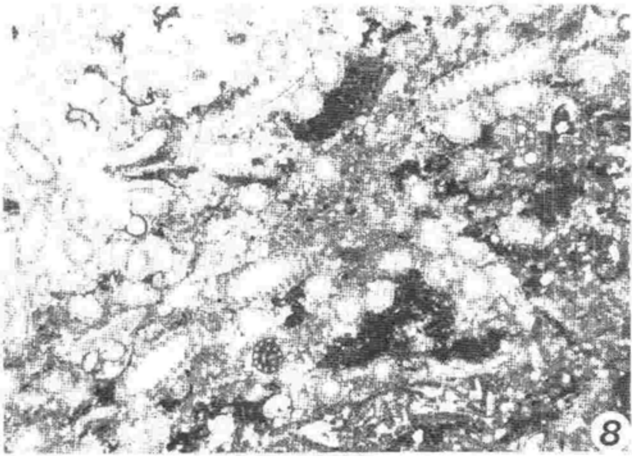
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a symbiotic association between fungi and cortical cells of roots. Thus, *Microcodium* may be used as a criterion for recognition of palaeosols. The presence of *in situ Microcodium* indicates terrestrial conditions, subaerial exposure and cessation of marine sedimentation.

5.2 Laminated Facies (dololaminites and microbial structures)

Description: Horizontal lamination in dolomicrites and dolomicrosparites generally results from 0.1 to 4 mm thick laminae (usually 1 to 2 mm), which are commonly separated from one another by thin bituminous films. Locally the dolomite crystal size decrease towards the top of individual laminae due to primary textural grading of individual laminae. Normal grading is also indicated by a decreasing grain size of silt-sized quartz grains, which may occur in low quantities within some of the units (Pl. 20/1). The presence of quartz grains also enables one to discern the outlines of sand-sized mud pellets, which originally constituted most of the thin laminae, although most of the textural features were destroyed during dolomitization.

While horizontal lamination is most common, layers may also be arranged in laterally-linked hemispherical shapes indicating cryptalgal fabrics. In such cases, open-space structures usually occur and the primary calcite mineralogy is commonly preserved (Pl. 20/4). The orientation of fenestrae can be concordant to lamination or irregularly. These voids may be filled with internal sediments (micrite and peloids) and cement (calcite or anhydrite). Furthermore, the Dololaminite Facies is occasionally associated with *Microcodium* Aggregates and circumgranular cracking.

Occurrence: Dololaminites are typical of the Gipshuken Formation (Late Sakmarian to Early Artinskian) and are most common in its upper parts. They rarely occur within the Ebbadalen Formation (Bashkirian).

Interpretation: Due to the development of extensive infauna or browsing organisms during the late Palaeozoic, the preservation of horizontal lamination is restricted to supratidal and upper intertidal conditions in tidal flat deposits (see Shinn 1983). In sabkhas, plant life is severely restricted and roots are seldom as abundant as to completely destroy sedimentary structures. The distinctly domed and laminated structures, which are clearly microbial in origin, suggest that the more horizontal lamination can also be a result of alternating deposition of marine sediment and algal mat growth. The incorporated detrital material is interpreted to be of eolian origin.

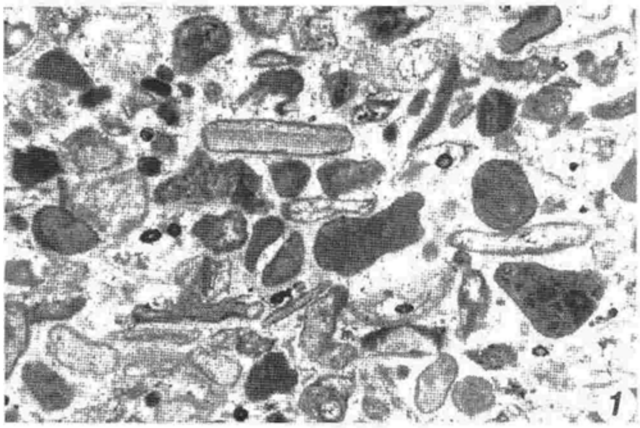
5.3 Dismicrite Facies

Description: Some micritic and pelmicritic limestones and dolomites exhibit open-space structures that in most cases are associated with microbial laminites (Pl. 20/4). The fenestrae are commonly arranged more or less concordantly to micritic and pelmicritic layers, but the outline of fenestrae is mostly irregularly formed. The primary voids may be filled with internal sediments (micrite and peloids) and cement (calcite or anhydrite; Pl. 20/2). Typical stromatactis cavities with a relatively flat base and an irregularly digitated roof are rare.

Occurrence: The Dismicrite Facies is typical of the Gipshuken Formation (Late Sakmarian to Early Artinskian) and is most common in its upper parts.

Interpretation: The preservation and association with the Laminated Facies and more rarely with *Microcodium* Aggregates points to supratidal and upper intertidal conditions in tidal flat environments (see above).

Plate 17	Moscovian to Gzhelian (late Carboniferous) platform carbonates from Svalbard: Minkinfjellet Formation, Cadellfjellet and Kapitol Members (Wordiekammen Formation) of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light (otherwise stated).
Fig. 1.	Bioclastic Grainstone Facies. Sorted and rounded grains showing strong alteration and dolomitization. But their outline is well preserved due to micritic envelopes. Sample 248-1, Boltonbreen (GI 1). x 19 # polars
Fig. 2.	Bioclastic Grainstone Facies. The well-sorted particles include abundant calcitornellid foraminifers, lithoclasts and rare superficial ooids. Sample 305, Boltonbreen (GI 2). x 19
Fig. 3.	Bryozoan/Phylloid Algal Facies. Fenestrate bryozoans overgrown by *Tuberitina*. Sample 269, Boltonbreen (GI 2). x 19
Fig. 4.	Bryozoan/Phylloid Algal Facies. Fecal pellets among poorly preserved phylloid algae. Sample 276, Boltonbreen (GI 2). x 19
Fig. 5.	Bioclastic Grainstone Facies. Close-up view of Fig. 2. Sample 305, Boltonbreen (GI 2). x 77
Fig. 6.	Bryozoan/Phylloid Algal Facies. Interstices of fenestrate bryozoans are filled with peloidal micrite and sparite. Sample 269, Boltonbreen (GI 2). x 19
Fig. 7.	Bryozoan/Phylloid Algal Facies. Photomicrograph illustrating the rare occurrence of abraded and rounded bioclasts of phylloid algal in peloidal packstone texture. Sample 270, Boltonbreen (GI 2). x 19
Fig. 8.	Phylloid algal packstone. The superimposed phylloid algae thalli are poorly preserved. Sample 262, Boltonbreen (GI 1). x 19
Fig. 9.	Phylloid Algal Boundstone Facies. *Eugonophyllum* sp. Note encrusting *Tubiphytes* and *Epimastopora* which occasionally occur within this facies. Sample 182-2, Alvrekdalen (FH). x 19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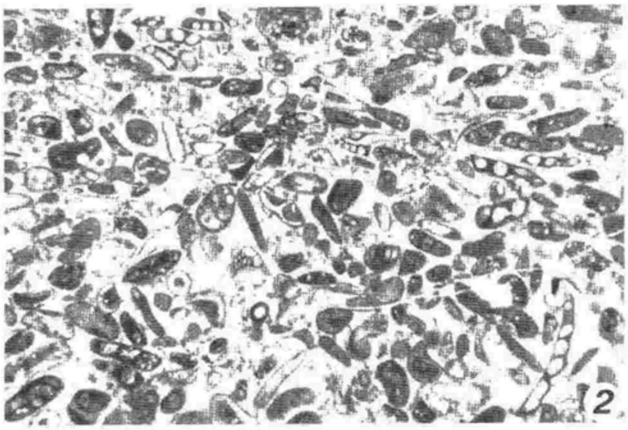
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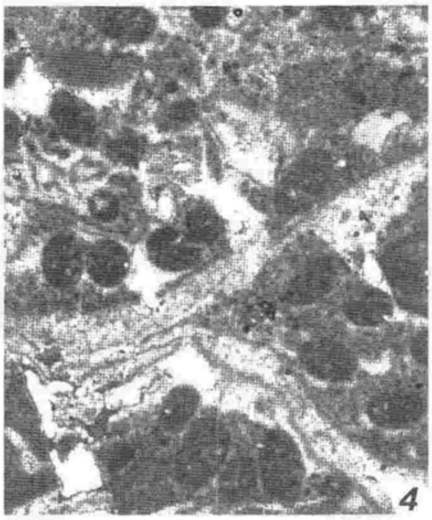
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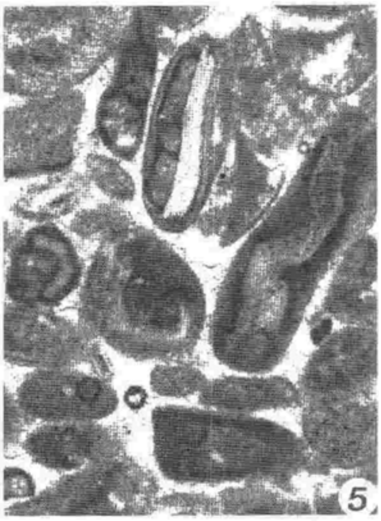
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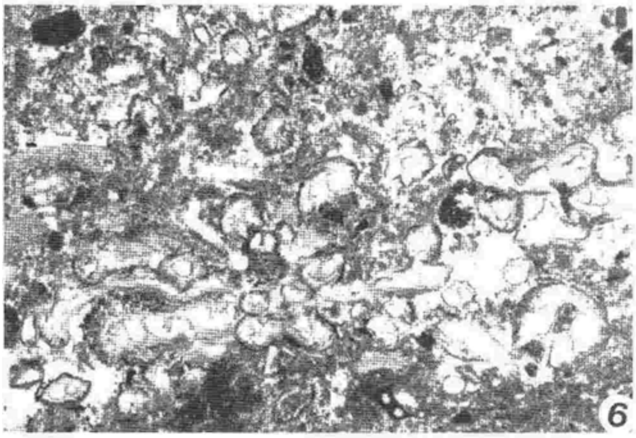
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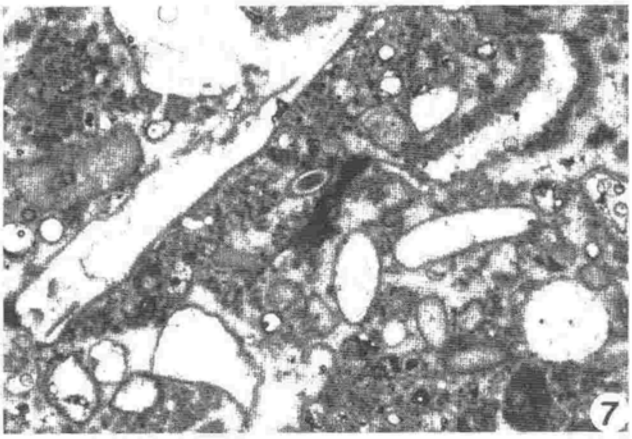
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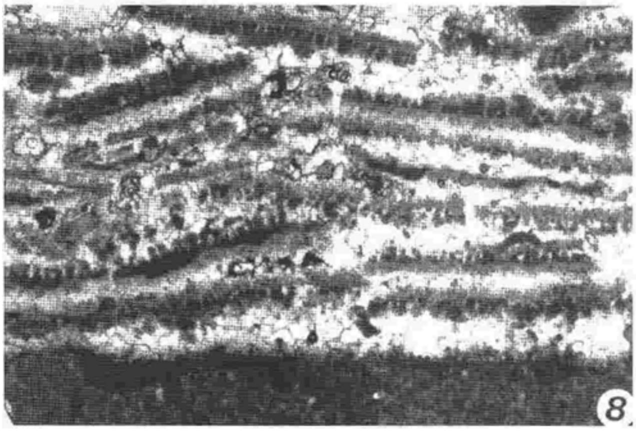
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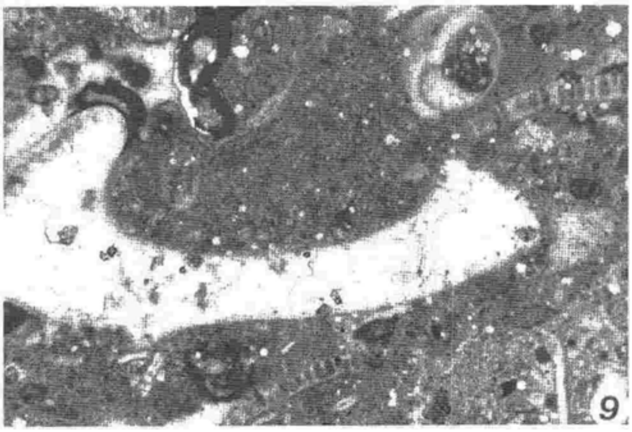
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5.4 Peloidal Facies

Description: This facies is extremely variable in character. Its most regularly occurring components are peloids (more than 50 %) which are accompanied by highly variable amounts of bioclasts. Within bioclastic wackestones and packstones, peloids may constitute a substantial part of the groundmass between bioclastic particles. Peloids are often squashed beyond recognition (0.03-0.2 mm). Therefore, the boundaries of these pellets are difficult to define in the micritic matrix. They usually exhibit subangular and subrounded shapes. Peloids of more irregular shapes may be entirely micritized fragments of bioclasts. A typical biotic association may consist of sessile foraminifers (*Ammovertella*), *Tubiphytes*, other small foraminifers, phylloid algae, ostracods, gastropods, crinoids and bryozoans. Although the typical intergranular cement is calcite, small amounts of the pore space may be filled with gypsum cement.

Peloids in some grain-supported textures usually show irregular shapes, a fluffy structure and diverse sizes (0.05-0.3 mm), suggesting that they represent micritic calcareous microbes or algae (Pl. 16/1). Some of these peloids resemble spherical carbonate encrustations around coccoid biotic forms (blue-green algae ?).

Locally, typical fecal pellets (0.2-0.8 mm) occur as well, but in general make up a small portion among other peloids. They display a spherical outline with a length/width ratio of 2:1, and often thin dark rims. Characteristically, fecal pellets are often concentrated in patches or occur as fills in small burrows.

Peloids that occur in sequences of predominantly mud-supported textures are elongate to irregular in shape with subangular and subrounded outlines. Although some peloids are similar to fecal pellets (see above), most resemble small intraclasts (0.2-2.0 mm Ø). Their micritic structure is identical to that of the surrounding micrite. Depositional textures range from pelmicrites to pelsparites. In the latter case, anhydrite and gypsum frequently occur as intergranular cements or as nodules (1-10 cm). Dololaminites as well as laminated mudstones are closely associated within these sequences.

Occurrence: The Peloidal Facies is associated with various bioherm facies of the Minkinfjellet Formation (Moscovian), Cadellfjellet and Kapitol Members of the Wordiekammen Formation (Moscovian to Gzhelian) and occurs as isolated beds within sequences of anhydrites and dolomites of the Gipshuken Formation (Sakmarian to Early Artinskian).

Interpretation: This facies characterises, but is not limited to, semi-restricted environments. It nearly always occurs in moderate to low energy settings. The transition from Peloidal Facies to open marine facies (in particular to the Palaeoaplysinid Boundstone Facies) is characterised by increasing biotic diversity and decreasing peloid content. Peloidal grainstones are indicative of deposition in very warm water with only moderate water circulation (Flügel 1982).

The Peloidal Facies is typical (1) of shallow lagoonal environments adjacent to isolated shoals in general open-marine platform areas (e.g., formed late during the colonisation stage of *Palaeoaplysina* bioherms; Fig. 11). Restricted environments of inter-buildup lagoonal setting arise during falling stages of sea-level. Furthermore, the Peloidal Facies is typical (2) of tidal flat sequences that include protected coastal lagoons seaward of the intertidal zone (Fig. 9). There, pelleted muds were easily stirred into suspension and transported to the flats during storms.

Plate 18 Latest Gzhelian to Asselian (early Permian) platform carbonates from Svalbard: Lower Tyrrellfjellet Member of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light.

Fig. 1. Bioclastic Grainstone Facies. Intraclasts, crinoids, bryozoans and poorly preserved *Palaeoaplysina* fragments are the most common particles. Sample 194, Alvrekdalen (FH). x 19

Fig. 2. Bioclastic Grainstone Facies. Sorted and rounded particles consist of intraclasts and abraded bioclasts such as crinoids, bryozoans, *Epimastopora* and strongly altered *Palaeoaplysina* fragments. Sample 140, Alvrekdalen (AV). x 19

Fig. 3. Coral Floatstone Facies. The photomicrograph shows bioclastic debris in packstone texture between large rugosan corals. Note *Tubiphytes* encrusting a fenestrate bryozoan and *Tetrataxis*. Sample 308-1, Boltonbreen (GI 2). x 19

Fig. 4. *Microcodium* Aggregates. *Microcodium* prisms usually form spherical clusters radiating from a central hollow. Sample 132, Alvrekdalen (AV). x 19

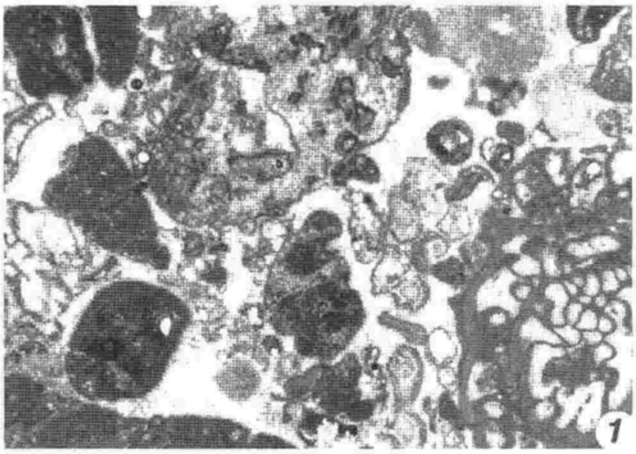
Fig. 5. Bioclastic Packstone/Wackestone Facies. Note the calcareous alga *Ungdarella* and the small foraminifer *Biseriella*. Sample 309, Boltonbreen (GI 3). x 19

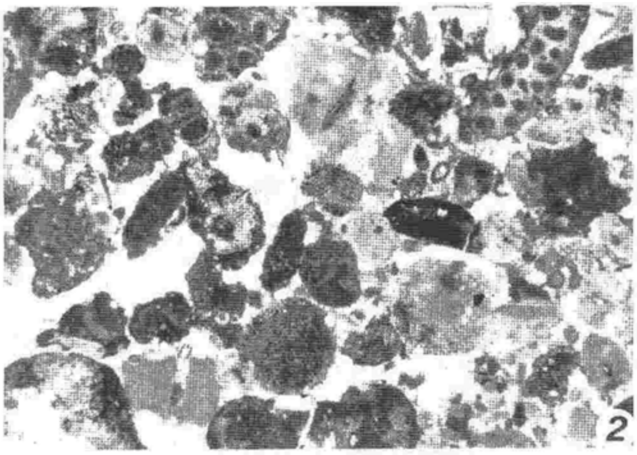
Fig. 6. *Tubiphytes* Facies. *Tubiphytes* commonly overgrows *Ammovertella*, *Calcitornella* and sponge spicules. Sample 129, Alvrekdalen (AV). x 19

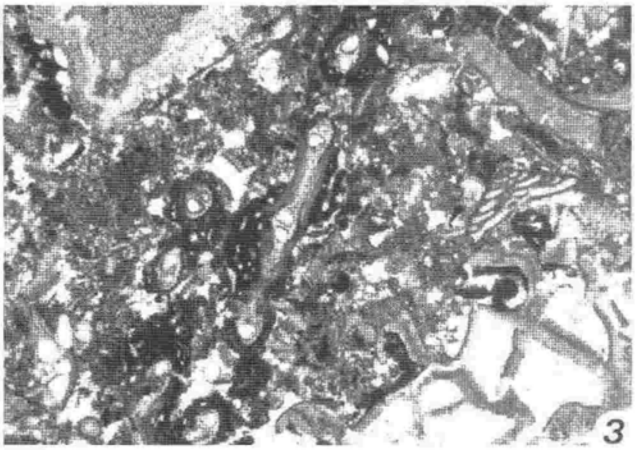
Fig. 7. Bioclastic Packstone/Wackestone Facies. Abundant *Tubiphytes* skeletons show a facial transition to the *Tubiphytes* Facies. Sample 128, Alvrekdalen (AV). x 19

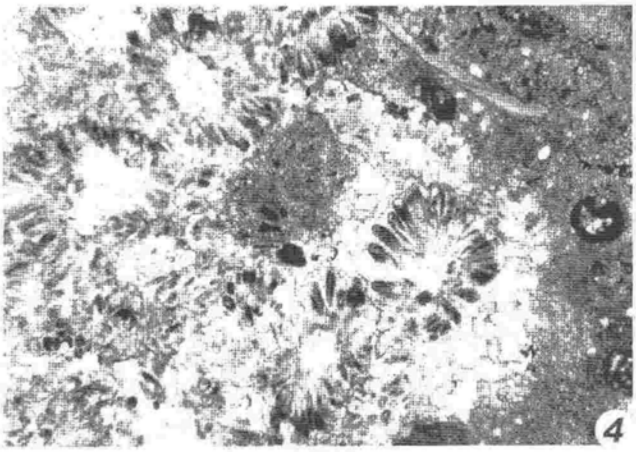
Fig. 8. *Tubiphytes*. Sample 128, Alvrekdalen (AV). x 77

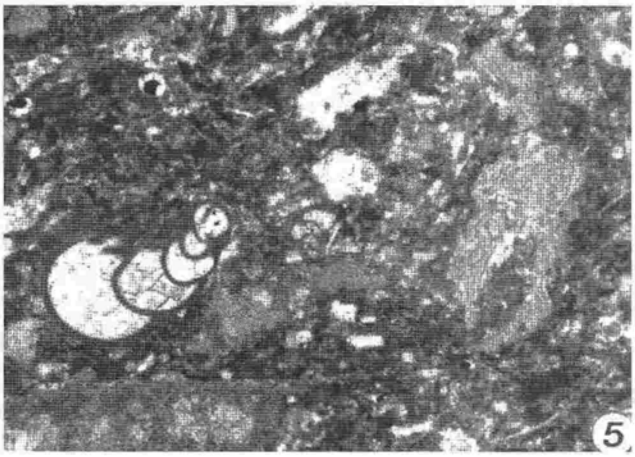
Plate 18

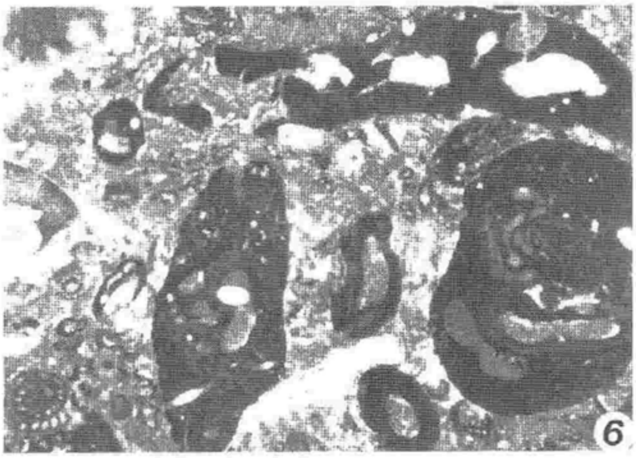


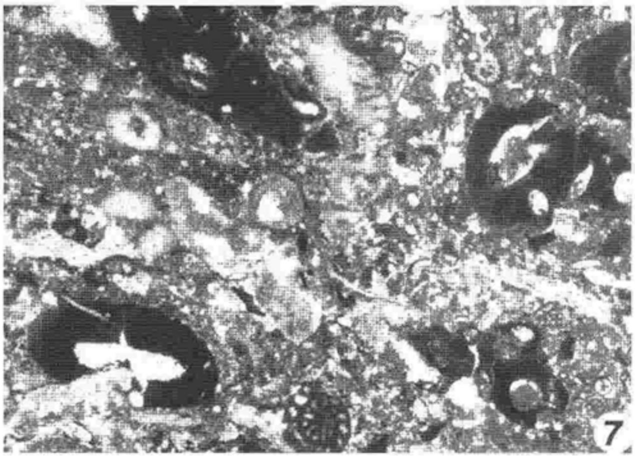


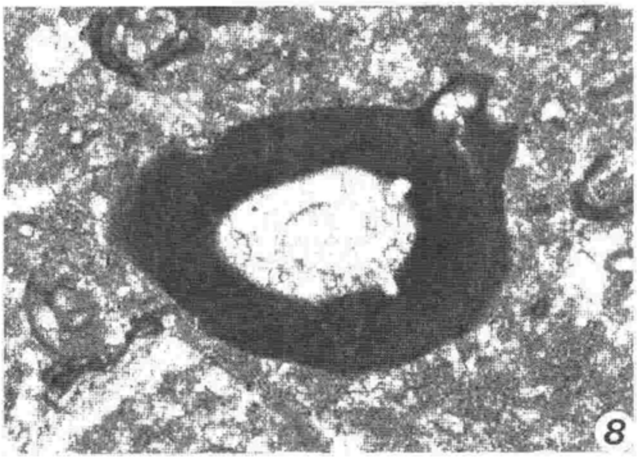












5.5 Small-Foraminifera Facies

Description: Depositional textures of the Small-Foraminifera Facies range from densely packed wackestones to poorly sorted grainstones, but most typical are packstones (Pl. 16/2, 20/6). The characteristic biogenic assemblage is composed of common to abundant small benthic foraminifers that comprise palaeotextulariids (*Climacammina, Cribrogenerina, Deckerella, Palaeotextularia*), tetrataxids, bradyinids, *Biseriella* and *Eonodosaria*. These are usually associated with sessile foraminifers (*Ammovertella, Calcitornella, Tuberitina, Eotuberitina*) in varying amounts (Pl. 19/4). Colonies of large rugosan corals locally occur, while *Ungdarella* and *Girvanella* are rarely present.

Usually fusulinids, bryozoans, crinoids, ostracods, brachiopods, *Tubiphytes* and rare trilobites complete the spectrum of biogenic components. Their occurrence is obviously caused by complete facies transitions between the Small-Foraminifera Facies and the Fusulinacean Facies or the Bioclastic Packstone Facies (Pl. 16/4), respectively.

Occurrence: The Small-Foraminifera Facies repeatedly occurs within the Minkinfjellet and the Wordiekammen Formations (Moscovian to Sakmarian) and is recognised in the Trikolorfjellet Member of the Ebbadalen Formation (Bashkirian).

Interpretation: The Small-Foraminifera Facies was accumulated at the lee-side embayments and protected lagoons of isolated shoals in open marine platform areas which were formed adjacent to *Palaeoaplysina* bioherms (Fig. 11). In these instances the facies is commonly associated with the Bioclastic Packstone/Wackestone Facies and the Peloidal Facies. The Small-Foraminifera Facies also occurs on subtidal platform highs formed by phylloid-algal bioherms. Tidal bars and channels of lagoons are a further suitable setting (Flügel 1982). Wilson (1975) suggested that small foraminifers may be concentrated by gentle winnowing of currents or waves. Perhaps slight restriction or higher water temperature inhibits other biota.

5.6 *Tubiphytes* Facies

Description: The *Tubiphytes* Facies is characterised by wackestone to packstone textures with abundant *Tubiphytes* and sessile foraminifers (*Calcitornella, Ammovertella, Tuberitina*). Additional biogenic elements are echinoderms, small foraminifers and locally bryozoans, brachiopods and ostracods. *Tubiphytes* may occur as isolated particles within the sediment ranging in size up to 3 mm (Pl. 18/6), as smaller epibiont relicts that are attached on broken fossil debris or encrustations on flat pebbles. The

Plate 19 Latest Gzhelian to Asselian (early Permian) platform carbonates from Svalbard: Lower Tyrrellfjellet Member of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light.

Fig. 1. Bioclastic Grainstone Facies. Grains comprise intraclasts, crinoids, fusulinids and bryozoans. Note the crinoid fragment that is overgrown by a bryozoan. Sample 140, Alvrekdalen (AV). x 19

Fig. 2. Bioclastic Packstone/Wackestone Facies. The photomicrograph illustrates the diverse fauna including crinoids, fusulinids, small benthic foraminifers (*Tetrataxis*) and *Tubiphytes*. Sample P13, Boltonbreen (P). x 19

Fig. 3. Fusulinacean Facies. The pressure dissolution is attributed to a raised amount of clay particles. Sample 191, Alvrekdalen (FH). x 19

Fig. 4. Small-Foraminifera Facies. Abundant single-layered calcitornellid foraminifers showing sharp edges and flat base in longitudinal section possibly related to an epibiont mode of life. These biogens are embedded among siliciclastic particles and isolated *Microcodium* prisms indicating redeposition. Sample 321, Boltonbreen (GI 2). x 19

Fig. 5. Palaeoaplysinid Floatstone Facies. *Tubiphytes* encrusting an intensive recrystallized *Palaeoaplysina* fragment and the dasycladacean algae *Epimastopora*, which is a scarcely occurring floral element within this facies. Sample P13, Boltonbreen (P). x 19

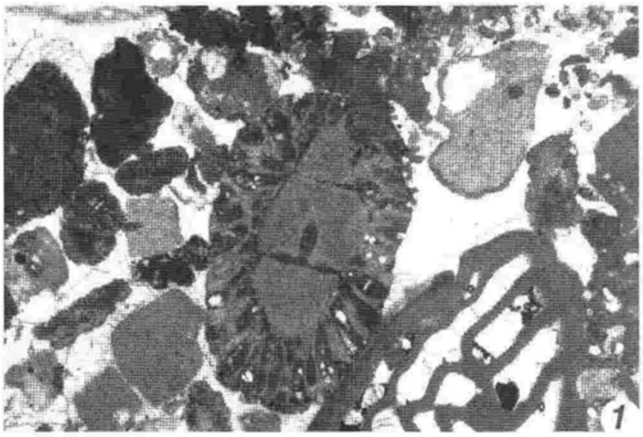
Fig. 6. Cross-section of a *Palaeoaplysina* plate showing remnant preservation of the internal cellular tissue. Most of the internal plate structure is not preserved due to recrystallization to pseudosparite. Channels are blocked up with dark peloidal micrite. Sample 190(2), Alvrekdalen (FH). x 19

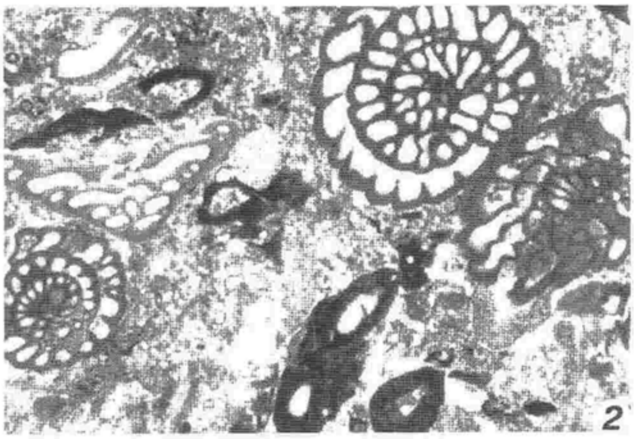
Fig. 7. Palaeoaplysinid Floatstone Facies. *Tubiphytes* micro-reef growing on *Palaeoaplysina*. Sample 190(2), Alvrekdalen (FH). x 19

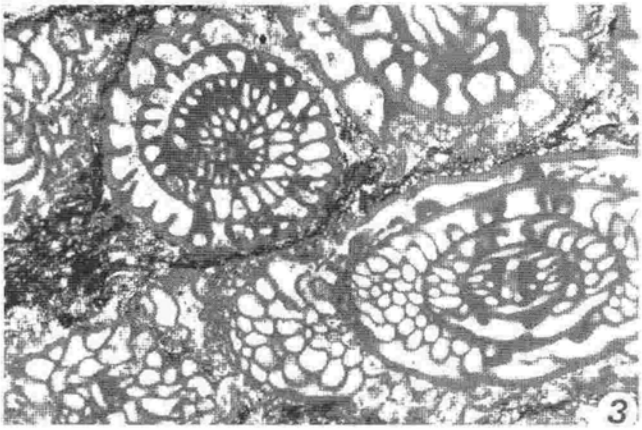
Fig. 8. Palaeoaplysinid Boundstone Facies. Mamelon-like protuberance at the upper-side surface of a *Palaeoaplysina* plate encrusted by *Tubiphytes*. The well preserved outline of the channel system is attributed to an early blocking by peloidal sediment and *Tubiphytes* growing into external parts of these channels. Isolated *Microcodium* prisms and granular cement fill the dissolved parts of the plate. Sample 138-2, Alvrekdalen (AV). x 19

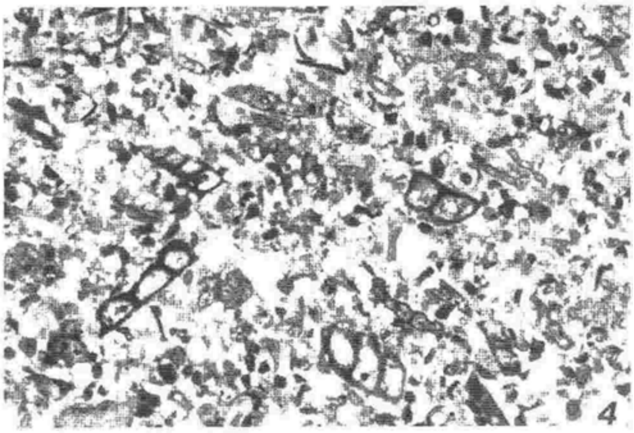
Fig. 9. Palaeoaplysinid Boundstone Facies. Upper part of a *Palaeoaplysina* plate showing branching channels that emerge as pores on its upper-side surface. Note peloidal sediment partly filling these channels. Sample 307, Boltonbreen (GI 2). x 19

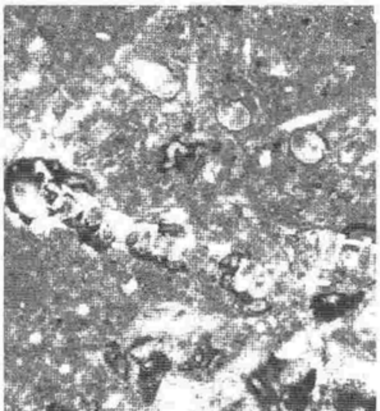
Plate 19

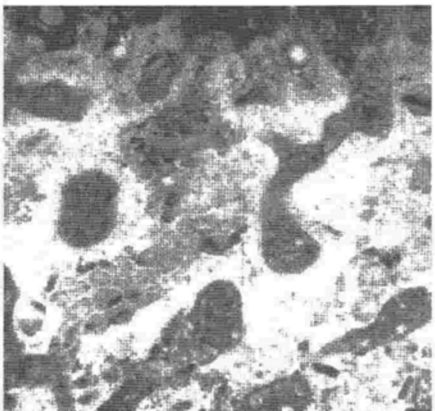


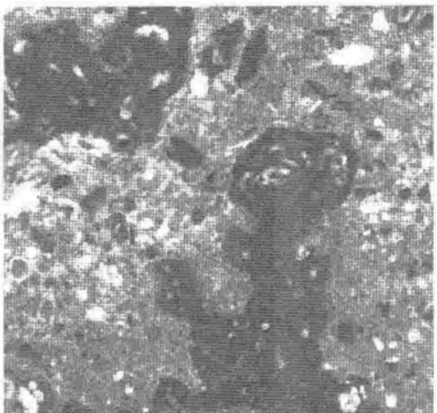


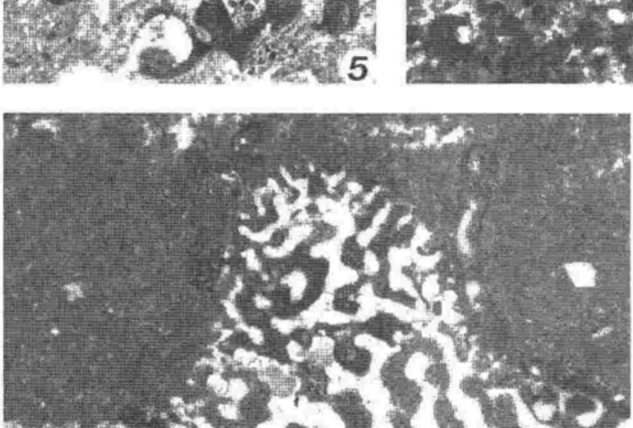


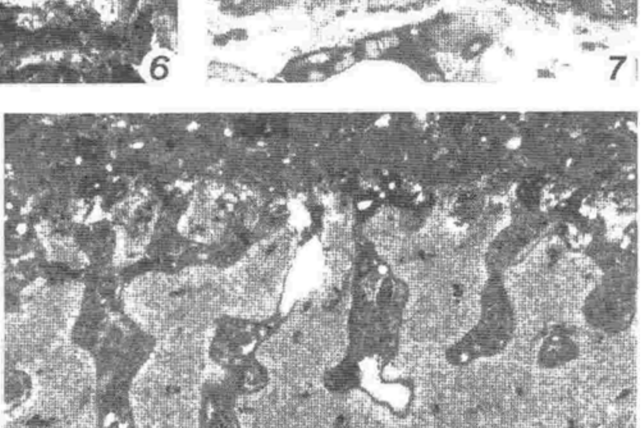












enigmatic taxa is composed of mini-micrite (about 1 micron) and forms a dark, flocculent mass in parts showing concentrically banded internal structures. Internal tubes representing a central canal system are usually present (Pl. 18/8). *Tubiphytes* contributes up to 30 vol.% of the sediment within this facies. Peloids occur as small rounded particles (about 0.05 mm) to larger irregular micritic particles (up to 0.2 mm).

Relics of a primary lamination and sorting are in places preserved. However, various burrow fabrics are abundant and suggest that burrowing organisms were active in the homogenisation of the fine-grained carbonate sediment.

Occurrence: The *Tubiphytes* Facies occurs within successions which are characterised by fast and repeated facial shifts within the Kapitol and the Tyrrellfjellet Members of the Wordiekammen Formation (Moscovian to Sakmarian).

Interpretation: The *Tubiphytes* Facies is an associated bioherm facies. Since *Tubiphytes* and sessile foraminifers did not live in the muddy, low energy sedimentary environment but are typical epibionts growing predominantly on *Palaeoaplysina* plates or bryozoan branches within miscellaneous bioherm facies, the wackestone/packstones of the *Tubiphytes* Facies obviously show a textural inversion in the sense of Folk (1962). This interpretation is borne out by the fact that the *Tubiphytes* Facies occurs in the "vicinity" of both the Palaeoaplysinid Boundstone Facies and the Beresellid Boundstone Facies, and seldom as individual beds within the bioherm successions themselves (Fig. 11). The textural inversion may be either the result of storm events affecting the bioherms and subsequent homogenisation of the abraded sediment by burrowing organisms, or the effect of intensive bioerosion and subsequent redeposition of the rasped and crushed particles.

5.7 Bioclastic Packstone/Wackestone Facies

Description: Typical depositional textures of the Bioclastic Facies are packstones and wackestones whose particles are sometimes broken but usually not abraded (Pl. 19/2). The general lack of lamination and different packing of bioclastic particles is interpreted as a result of prevalent bioturbation. Locally bioturbation is also indicated by the arrangement of bioclasts in a circular pattern. However, the texture is usually homogenised by extensive burrowing. The matrix is micrite or microsparite, generally the former. A further attribute of the facies is a highly diverse fauna including fusulinids, crinoids, brachiopods, rugosan corals, ramose bryozoans, fenestellids, ostracods, tetrataxids, endothyrinids, palaeotextulariids, bradyinids, mollusks, gastropods and sessile foraminifers (*Tuberitina*, *Calcitornella*). The shells of brachiopods and crinoid ossicles are always disarticulated. Pores of the outer tectorium of fusulinid tests are usually filled by micrite indicating redeposition of these bioclasts. The presence of scarce floral elements (e.g. *Tubiphytes*, phylloid algae, *Ungdarella*; Pl. 18/5, 7) and peloids is related to the proximity of bioherm accumulations (Palaeoaplysinid Boundstone Facies, Bryozoan/Phylloid Algal Facies and Beresellid Boundstone Facies). The Bioclastic Packstone Facies is the most variable in composition.

Occurrence: The Bioclastic Packstone/Wackestone Facies occurs occasionally within the Ebbadalen Formation (Bashkirian) and is widespread within the Minkinfjellet and Wordiekammen Formations (Moscovian to Sakmarian).

Interpretation: The Bioclastic Facies points to open marine, low-energy depositional environments occurring in depths below the fair-weather wave base. It constitutes

Plate 20 Sakmarian and Early Artinskian (early Permian) platform carbonates from Svalbard: Uppermost Tyrrellfjellet Member (Wordiekammen Formation) and Gipshuken Formation of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light (otherwise stated).

Fig. 1. Laminated Facies. Siliciclastic dolo-microsparite showing cryptalgal fabrics. Sample 95, Skansbukta (GF). x 2.5

Fig. 2. Dismicrite Facies. Dolomicrite with gypsum-filled birdseyes. Sample 170, Alvrekdalen (AV). x 19

Fig. 3. Spiculite Facies. Photomicrograph illustrating the rare occurrence of spiculitic wackestones within dolomite and anhydrite successions. The open space structure is caused by burrowing. Sample 110-1, Skansbukta (SK). x 19

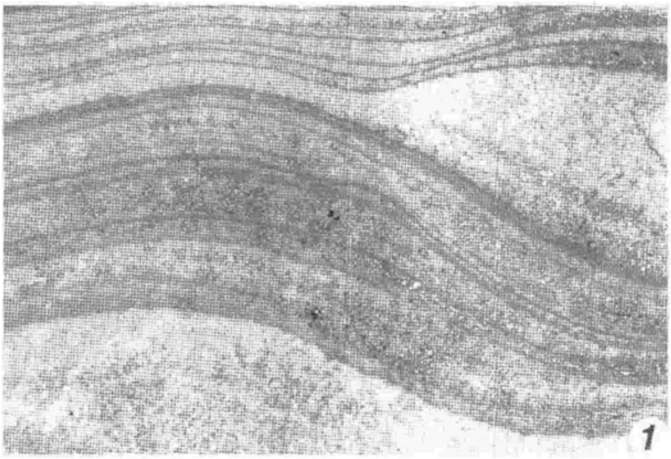
Fig. 4. Dismicrite Facies / Laminated Facies. The dolomite is characterised by fenestral and algal laminated fabrics. Note secondary enlargement of open spaces due to burrowing. Sample 109, Skansbukta (SK). x 19 # polars

Fig. 5. Crinoid-Bryozoan Facies. Sample 337, Boltonbreen (GI 3). x 19

Fig. 6. Small-Foraminifera Facies. Sample 78, Skansbukta (HK2). x 19 # polars

Fig. 7. Oolitic Facies. Note dissolved ooid nuclei. Equant calcite fills these moulds and forms circum-granular and relict meniscus cements within the interparticle pore space. The rest of the pore space is blocked up with gypsum and anhydrite. Sample 105-2, Skansbukta (GF). x 77 # polars

Fig. 8. Oolitic Facies. Dolomitic, poorly preserved ooids in grainstone texture. Interparticle pore space is filled with gypsum. Sample 154, Alvrekdalen (AV). x 77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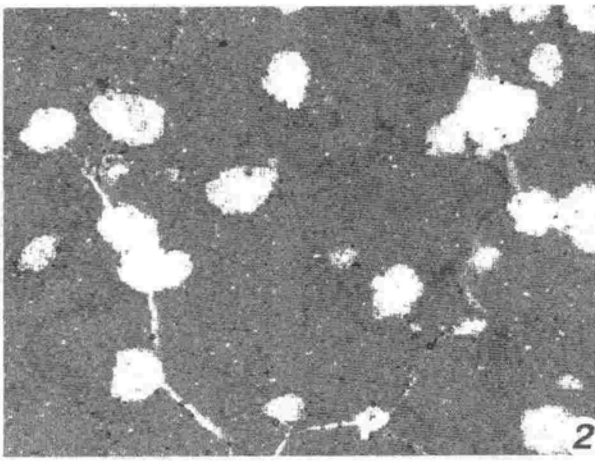
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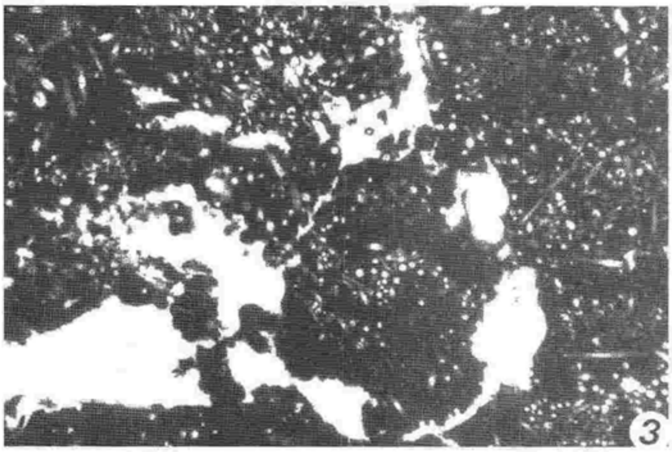
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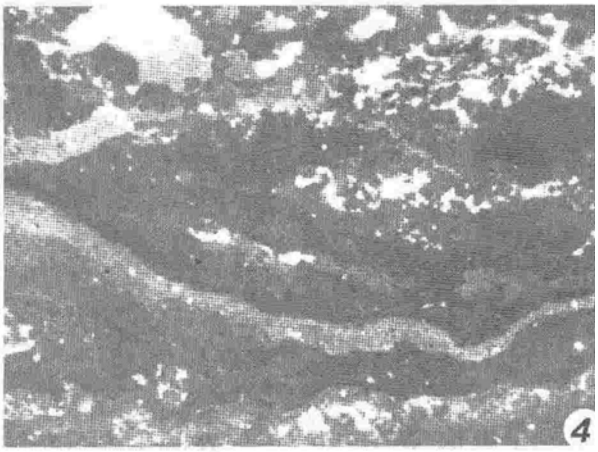
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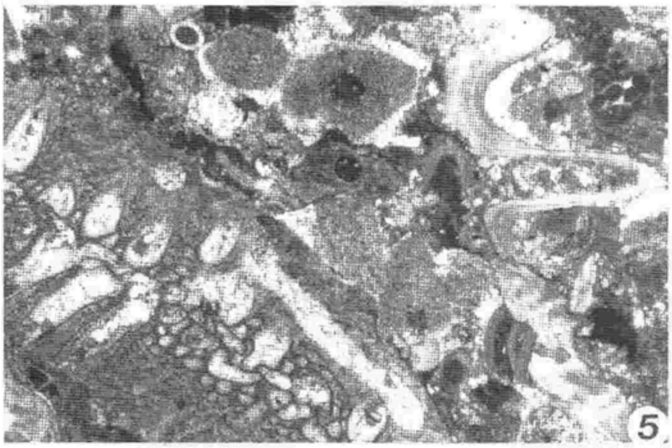
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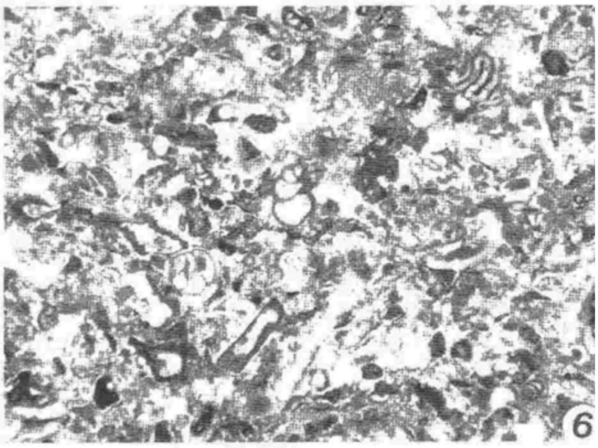
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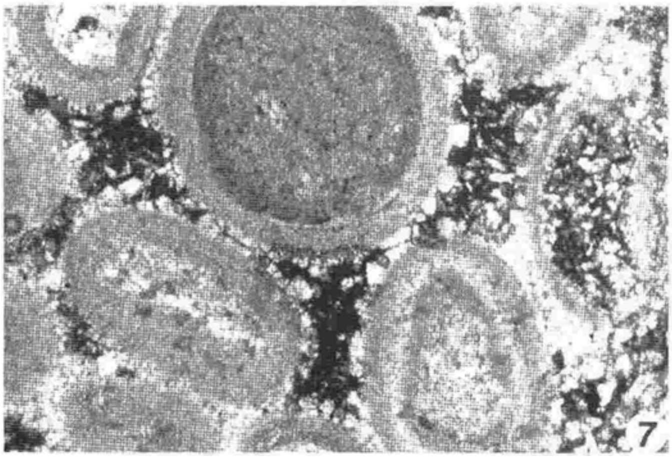
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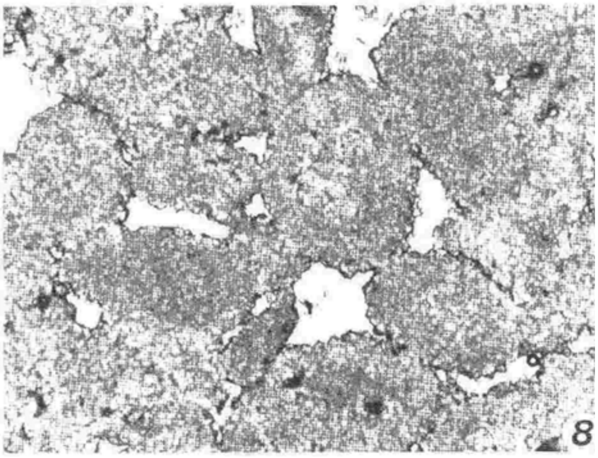
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widespread sheets of "normal marine" limestones between the buildups which grade laterally both into muddy and peloidal sediments (see 5.4. and 5.5) of slightly restricted environments, as well as into the various sediments flanking the *Palaeoaplysina* bioherms and the phylloid-algal bioherms (see 5.12, 5.15, 5.18). The environment was characterised by clear, calm water of open circulation. The diverse stenohaline fauna and flora of this facies and of the associated bioherm facies (compare "chlorozoan" association of Lees and Buller, 1972) is indicative of deposition in warm and tropical shelf waters (see 6.1).

5.8 Crinoid Bryozoan Facies

Description: Depositional textures of the Crinoid Bryozoan Facies range from grain- to mud-supported with particles which are sometimes broken but usually not abraded. Packstones are the most typical, however (Pl. 20/5). The facies exhibits a poorly diverse fossil assemblage comprising crinoids, fenestellid and ramose bryozoans, brachiopods, bivalves and in places silicosponges (Pl. 22/7). Crinoid ossicles may be abundant. Most ossicles are disarticulated and tend to be heavily bored. Prevalent bioturbation is indicated by a general lack of lamination and by an arrangement of bioclasts in a circular pattern. Substantial amounts of mature, silt- and sand-sized siliciclastic particles may be present and cause lamination.

Occurrence: The Crinoid Bryozoan Facies is found sporadically within the Ebbadalen Formation (Bashkirian) and in the upper Tyrrellfjellet Member of the Wordiekammen Formation (Sakmarian), but occurs typically within the Kapp Starostin Formation (Late Artinskian and Kungurian).

Interpretation: The association of skeletal grains corresponds to the "foramol" association of modern shelf carbonates which is characteristic of temperate waters but also extends well into the tropics (see Lees and Buller 1972). The Crinoid Bryozoan Facies occurs in shallow subtidal settings of open-marine, low-energy environments. It forms widespread sediment sheets and patches below the fair-weather wave base and may grade laterally into spiculitic muds (see 5.19). Depositional textures of this facies are comparable with those of the Bioclastic Packstone Facies (See 5.7), although the less diverse fauna points to mesotrophic or more temperate marine waters (see 6.1).

5.9 Brachiopod Facies

Description: The facies consists of grainstones and packstones with abundant brachiopods, various types of bryozoans, crinoids and spicules from siliceous sponges (Pl. 21). Locally brachiopod rudstones occur, in which large and thick-shelled brachiopods are embedded in a packstone matrix of fragmented bryozoans, brachiopods and crinoids. In some cases these particles are well rounded and sorted (Pl. 21/5). Brachiopod shells frequently show punctate and pseudopunctate structures and brachiopod spines are very common. Brachiopods with an atypical prismatic wall structure occur as well (*Composita*).

The sediment is commonly intensively bioturbated with rare primary structures (Pl. 22/2). *Zoophycos* is the only identifiable trace fossil in sandy limestones. Fecal pellets are often concentrated in isolated nests along burrows. Usually the facies is characterised by the presence of chert nodules and glauconite. A complete facies transition with the open-shelf subfacies of the Spiculitic Facies is characterised by a change to mud-supported textures and the disappearance of numerous taxa. Instead, additional biogenic elements such as uniserial nodosariid foraminifers occur.

Occurrence: The Brachiopod Facies is found at the base of the Kapp Starostin Formation (Late Artinskian/ early Kungurian).

Interpretation: We interpret this facies to have been deposited at or above fair-weather wave base in proximal settings. The lack of mud and the abundance of well rounded particles is indicative of intense wave action, although periods of constant reworking are likely to have

Plate 21 Late Artinskian to Kazanian (middle Permian) platform carbonates from Svalbard: Kapp Starostin Formation of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light.

Fig. 1.	Brachiopod Facies. Fenestrate bryozoan among abundant sponge spicules in bioclastic packstone texture. Sample 6, Skansbukta (HK1). x 19
Fig. 2.	Brachiopod Facies. Bryozoan fragment in bioclastic packstone texture. Sample 6, Skansbukta (HK1). x 19
Fig. 3.	Brachiopod Facies. Punctate brachiopod shells in bioclastic packstone texture. Sample 6, Skansbukta (HK1). x 19
Fig. 4.	Brachiopod Facies. Brachiopod shell with spine bases. Sample 6, Skansbukta (HK1). x 19
Fig. 5.	Brachiopod Facies. Rounded and poorly sorted grainstone. Note sections of the pseudopunctate brachiopod shell and spine. Sample 4, Skansbukta (HK1). x 19
Fig. 6.	Brachiopod Facies. Bryozoan fragment. Sample 4, Skansbukta (HK1). x 19
Fig. 7.	Brachiopod Facies. Brachiopod shells and crinoid ossicles in poorly sorted grainstone texture. Sample 4, Skansbukta (HK1). x 19
Fig. 8.	Brachiopod Facies. Strongly bored impunctate brachiopod shell. Sample 4, Skansbukta (HK1). x 19


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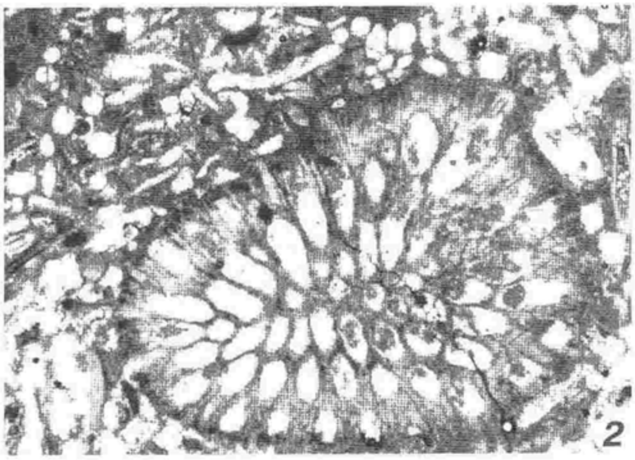
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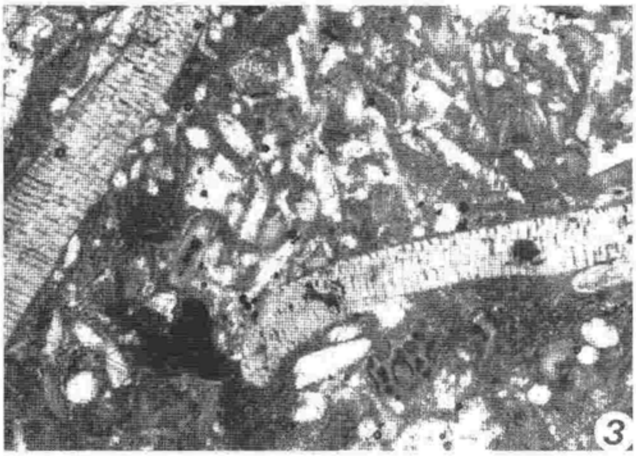
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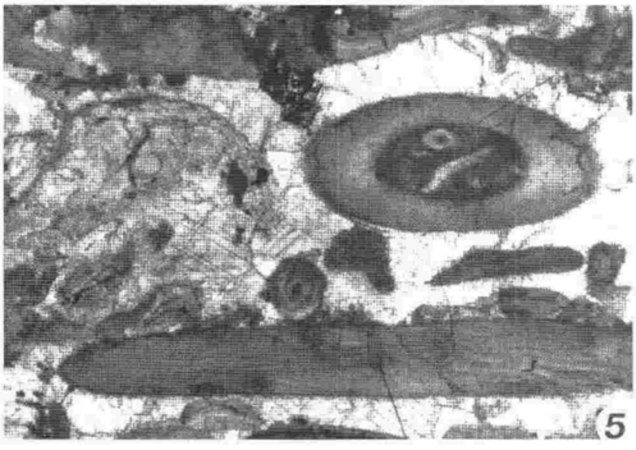
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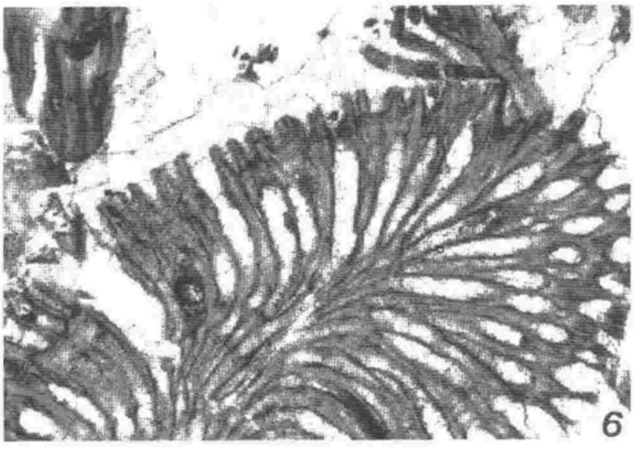
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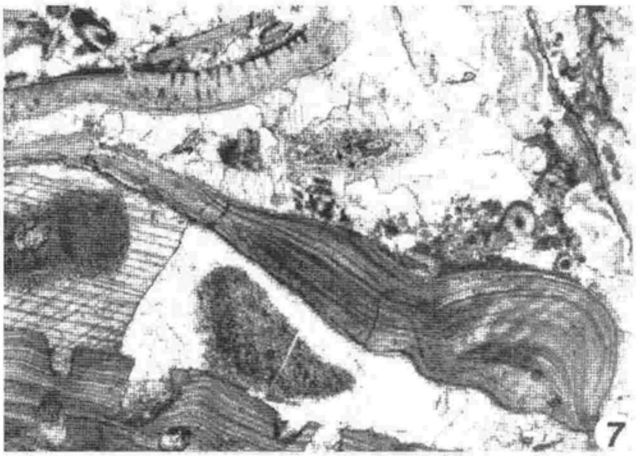
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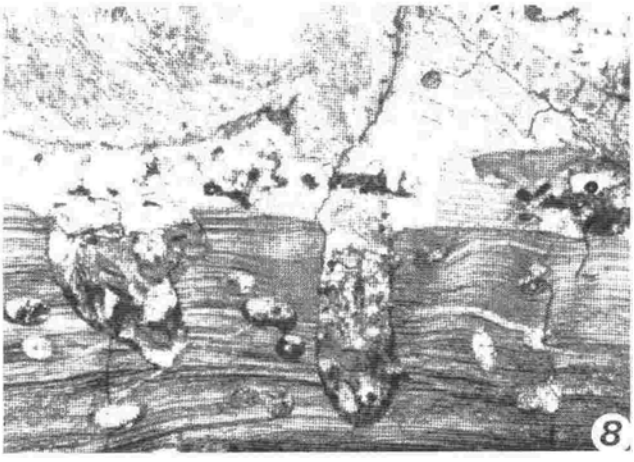
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changed with intervals of rapid growing populations and intense sediment bioturbation. The predominance of brachiopods points to inner platform areas since the ecological zones occupied by bivalves in modern environments were taken by brachiopods during Palaeozoic times (see Stemmerik 1997). The Brachiopod Facies is comparable to the brachiopod-dominated facies described by Malkowski and Hoffman (1979) from Svalbard and by Stemmerik (1997) from North Greenland.

5.10 Coral Floatstone Facies

Description: Isolated colonies of rugosans, multithecoporids, syringoporids and other tabulate corals are found in horizontally well-stratified limestone beds. *Multithecopora* sp. is the most conspicuous element (Pl. 16/3). Individual corals or clusters of delicate skeletons root *in situ* in micrite which contains only a few scattered bioclasts. Sessile foraminifers are typical epibionts. Within a single stratum there may occur a transition to floatstones that are characterised by overturned and sometimes broken coral colonies.

Occurrence: The facies commonly occurs in the Tyrrellfjellet Member of the Wordiekammen Formation (Asselian to Sakmarian).

Interpretation: The Coral Floatstone Facies, which contains preserved infauna and epifauna, corresponds to the whole-fossil wackestone of Flügel (1982) and is interpreted to have been deposited in quiet water below normal wave base. At the base of the Scheteligfjellet Member, multithecoporids and associated solitary corals covered wide areas of the St. Johnsfjorden Trough (Bröggertinden Basin) and built up colonies or extensive biostromes. Ludwig (1991) interpreted these biostromes as deposits of shallow bays or lagoons with open circulation seawards but only reduced wave agitation.

5.11 Fusulinacean Facies

Description: The Fusulinacean Facies is mostly characterised by loosely- as well as densely-packed wackestones and the dominance of fusulinids (Pl. 19/3). Fusulinids occur in rock-forming quantities but are also commonly found in limestone beds with other members of varied normal marine biota. A high organic content is often typical. Some of the more intensively compacted beds, due to higher amounts of siliciclastic mud, give the false appearance of grain supported textures. These are usually dark grey to black bioturbated limestones or marls.

The Fusulinacean Facies includes fusulinid wackestones or packstones with an interstitial matrix of silt to fine sand-sized bioclastic debris that reflects striking bimodal sorting. In many cases the fusulinids are worn and thoroughly micritized, reflecting transport over some distance before deposition. The pores of the outer tectorium are filled and blocked up with micrite, which clearly indicates redeposition of these bioclasts. The mud-supported textures occasionally contain bioclastic particles that show features of textural inversion. Broken fusulinid tests are embedded within the micritic matrix together with locally abundant fragments of large colonial rugosans. More rarely, crushed calcite cements cling to the bioclastic fragments (intraclasts) indicating reworking and redeposition after early diagenetic cementation.

In cases of intensive dolomitization, the fusulinid tests may be well preserved (including the finely porous outer tectorium), while textural features of the groundmass are completely destroyed by euhedral rhombic shaped dolomite crystals.

The presence of crinoids, brachiopods, bryozoans, palaeotextulariids, tetrataxids, bradyinids, *Eolasiodiscus*, sessile foraminifers (*Calcitornella*, *Tuberitina*) and rare ostracods is indicative of gradual transitions between the

Plate 22 Late Artinskian to Kazanian (middle Permian) platform carbonates from Svalbard: Kapp Starostin Formation of the Billefjorden Trough and Nordfjorden Block areas. Photomicrographs of thin sections in plane polarized light (otherwise stated).

Fig. 1. Spiculite Facies. The thick lined circular traces of *Schaubcylindichnus* are dwelling burrows, whereas the dark pellets (crescent-shaped arranged) represent feeding structures of *Scalarituba*. Sponge spicules are indiscernible at this magnification. Sample 19, Skansbukta (HK1). x 2.5

Fig. 2. Brachiopod Facies. Packstone showing circular burrow structure. Sample 6, Skansbukta (HK1). x 19

Fig. 3. Crinoid Bryozoan Facies. Siliceous sponge. Sample 29, Skansbukta (HK1). x 2.9

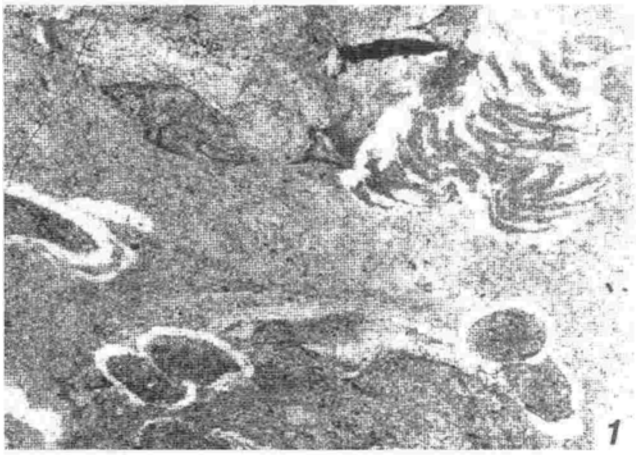
Fig. 4. Crinoid Bryozoan Facies. Abundant spicules within the chambers of a siliceous sponge. Close-up view of Fig. 3. Sample 29, Skansbukta (HK1). x 19

Fig. 5. Spiculite Facies. Section through the periphery of an agglutinated sponge that mainly consists of crinoid ossicles. Sample 38-1, Skansbukta (HK1) x 19 # polars

Fig. 6. Spiculite Facies. Cross sections of siliceous spicules showing well preserved central canal. Sample 23, Skansbukta (HK1). x 77

Fig. 7. Crinoid Bryozoan Facies. Silicified crinoids and sponge spicules in bioclastic wackestone texture. Sample 29, Skansbukta (HK1). x 19

Fig. 8. Crinoid Bryozoan Facies. Silicified crinoid ossicles, brachiopods, bryozoan fragments, spicules and a small foraminifer which are embedded in glauconitic micrite. Sample 29, Skansbukta (HK1). x 19



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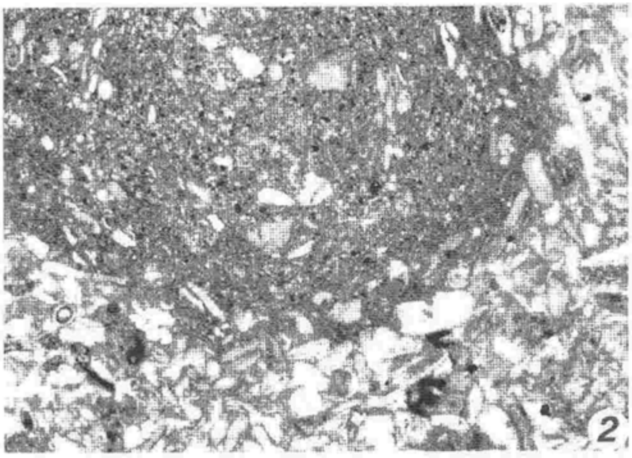

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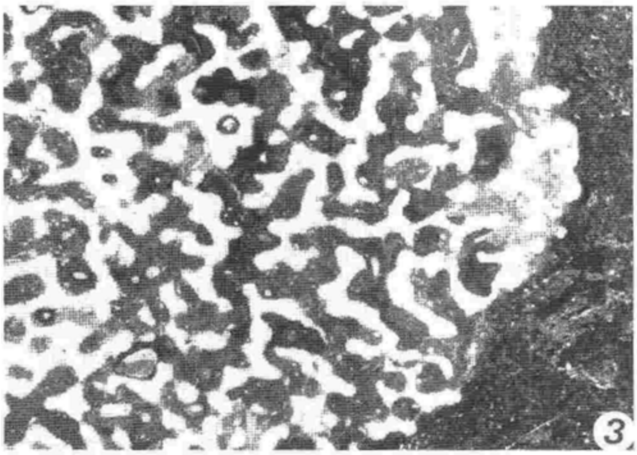

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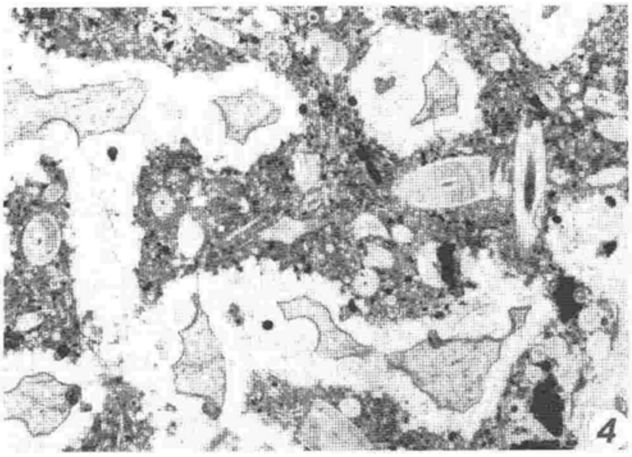

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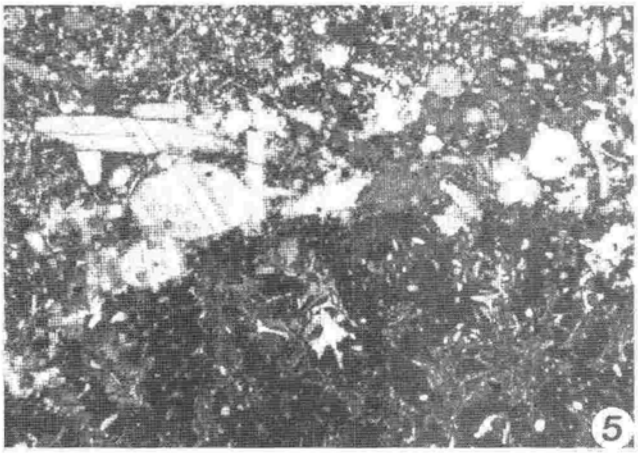

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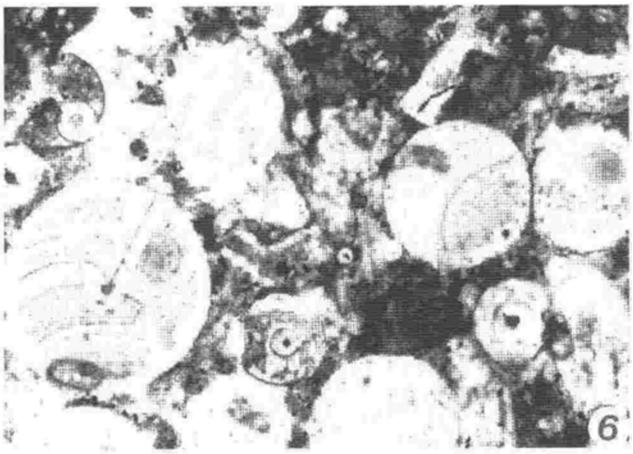

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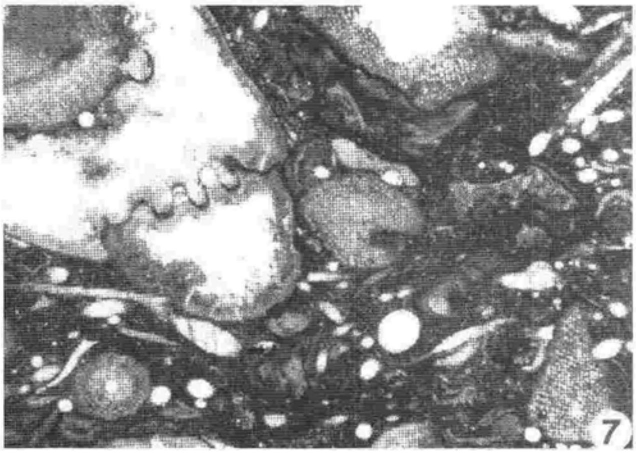

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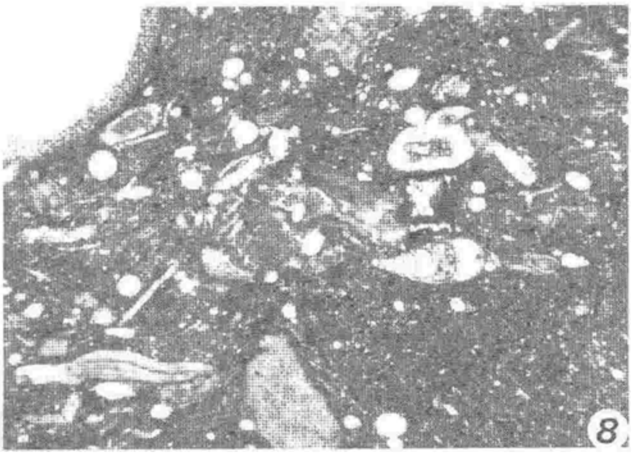

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Fusulinacean Facies and Bioclastic Packstone/Wackestone Facies. In these cases, depositional textures commonly change to primarily grain-supported textures. Fragments of *Palaeoaplysina*, rare phylloid algae and *Epimastopora* among abundant fusulinids point to gradual facial transitions to the Palaeoaplysinid Floatstone Facies that flanks the *Palaeoaplysina* bioherms. Facial transition to the Small-Foraminifera Facies is indicated by beds of small-foraminifera grainstones with intercalated fusulinid horizons.

The most striking variety of this facies are the bituminous "packstones" composed almost exclusively of fusulinids. Clusters of large colonial rugosans are locally abundant. These limestones form a distinct lithologic marker horizon about 10 m thick (Brucebyen Beds). They mainly occur at the base of the Tyrrellfjellet Member, although equivalent lithologies also cap the *Palaeoaplysina* bioherms at Alvrekdalen. The total organic content ranges from about 1% to more than 15% and is mainly derived from terrestrial vegetation (Dons 1983, Dallmann et al. 1999). Deposition is believed to have occurred during a marine transgression (Sundsbø 1982). Such "fusulinites" have been reported from coeval strata on Ellesmere Island (Beauchamp 1989, Morin et al. 1994), possibly indicating a time specific facies (see Walliser 1996).

Occurrence: The Fusulinacean Facies is a peculiar facies of the Tyrrellfjellet Member of the Wordiekammen Formation (Asselian to Sakmarian).

Interpretation: The Fusulinacean Facies represents environments along the inner parts of open shelf lagoons and is closely related to the Palaeoaplysinid bioherms. Sediment accumulation occurred in apparently normal, clear marine water, which ranged in depth from a few meters to a few tens of meters. The Fusulinacean Facies is transitional to the Bioclastic Packstone/Wackestone Facies which represents outer parts of open shelf lagoons. Furthermore, there are gradual transitions between Fusulinacean Facies and sediments flanking the Palaeoaplysinid Boundstone Facies (Fig. 11). According to Wilson (1975), fusulinids must have been able to live on biohermal slopes and hence in quite shallow, rather clear water. Cowen (1983, 1988) interpreted fusulinids to have harboured photosymbionts such as eukaryotic zooxanthellae or prokaryotic cyanobacteria. This is inferred on the basis of test architecture and facies association (Brasier 1995). Photosymbiotic associations are typically found in nutrient-poor habitats or in hosts which depend on a nutrient-poor diet (Muscatine and Porter 1977, Cowen 1983, Hallock and Schlager 1986, Lees and Hallock 1986, Brasier 1995). Therefore, the Fusulinacean Facies should indicate oligotrophic nutrient levels. However, fusulinid tests form resistant particles that were easily transported and are found in many facies types. Consequently, the Fusulinacean Facies include sediments which show textural inversion and are formed in swales close to bioherms and shoals. Dominant particles show features of high energy environments which have moved down local slopes to be deposited in quiet water below the upper photic zone.

Bituminous "packstones" that are composed almost exclusively of fusulinids are probably the result of expanded nutrient levels which usually have a negative effect on photosymbionts, as in recent benthic invertebrates. Temporary higher nutrient-influx levels accompanying the invasion of detrital particles (during regressions) have caused short-lived mesotrophic conditions and widespread dying off of fusulinids. This is due to the high interdependence in a diverse photosymbiotic community structure in which the symbiosis has reached an obligate status (Brasier 1995). The vulnerability of the biota is such that only a relatively minor perturbation may cause the collapse of specialized and interactive ecosystems through the loss of a few key members (Plotnick and McKinney 1993).

5.12 Bioclastic Grainstone Facies

Description: This facies is made of grainstones in which individual particles show evidence of intense reworking. All or almost all of the matrix (micrite) is washed out (Pl. 17/5). Lamination is clearly defined where there are variations in grain size, and in places cross-lamination is a typical structure. Carbonate bioclasts exhibit the broad spectrum from poorly- to well-sorted particles as well as rounded and abraded particles (Pl. 17/1, 18/2). Typical bioclasts are echinoderms, bryozoans (Pl. 19/1), brachiopods, small foraminifers, fusulinids, *Palaeoaplysina* plates (Pl. 18/1), dasycladaceans (*Epimastopora*), *Tubiphytes* and rare phylloid algae. Furthermore, rounded intraclasts, peloids and superficial ooids may occur (Pl. 17/2). Silt to sand-sized quartz particles are locally present.

Occurrence: The Bioclastic Grainstone Facies is widespread within the Minkinfjellet and the Wordiekammen Formations (Moscovian to Sakmarian)

Interpretation: It represents shoal environments of very shallow water with strong wave and current action. A distinct proportion of *Palaeoaplysina* plates, phylloid algae and fusulinids among bioclastic particles occurs particularly in sequences with Palaeoaplysinid bioherms. The Bioclastic Grainstone Facies forms capping beds at the tops of these mounds that indicate temporary reworking during sea-level lowstands (Fig. 11; Pl. 18/2).

Within calcarenites of the lower Kapitol Member more than 50% of the carbonate sand grains are darkened, giving the sediment a speckled appearance. Shinn (1973) described comparable "salt and pepper" sands from subtidal sand accumulations that occur as winnowed lag concentrate in channels of the intertidal zone in the Persian Gulf.

5.13 Oolitic Facies

Description: The Oolitic Facies consists of nearly pure oolitic grainstones (> 90 % ooids within any given sample). Ooids are well-formed and well-sorted with values of grain sizes frequently between 0.2 and 0.5 mm. Usually ooids are composed of several laminae, but locally superficial ooids occur as well. Frequently the distinct concentric laminae coating a nucleus are well preserved, but tangential and/or radial orientation of crystals and other microscopic textural details of ooids are lost during diagenesis where some internal structure has been obliterated by predominant dolomitization (Pl. 20/8). The ooid particles seldom have quartz nuclei. Much of the intergranular pore space between grains is filled by gypsum and anhydrite cement (Pl. 20/7). Besides ooids, intraclasts, rare peloids (micritized ooids ?), abraded bioclasts and sand-sized quartz grains occur. Scattered bioclasts are mollusks, fusulinids, gastropods, small foraminifers, bryozoans and brachiopods.

Thick units (up to 2 m) of primary oolitic grainstones (now dolosparites) exhibit complex cross bedding structures ranging from wedge-planar to trough cross strata. Medium scale cross bedding is characteristic. In these settings, good sorting of ooids and the low content of bioclasts (< 10 %) indicates that the accumulation of ooids occured in the environment of their primary formation (see Loreau and Purser 1973). Small scale ripple cross bedding, wide-span undulating lamination (hummocky cross bedding ?), cross bedding and channel fills occur within oolitic grainstone/mudstone cycles (= dolomicrosparites) which are associated with thin (< 1 m) anhydrite interbeds. Horizontal lamination and wavy horizontal lamination is typical in interlayered beds of oolitic grainstones and in structureless dolomicrosparites within thicker (> 1 m) anhydrite units (Fig. 12/B). Flat pebbles and mud cracks indicate emersion and subsequent reworking. In these settings, polyooids ("Ooidbeutel") and broken or regenerated ooids rarely occur.

Occurrence: The Oolitic Facies is found in the form of thick units at the top of the Tyrrellfjellet Member and as thin interbeds between evaporites at the base of the Gipshuken Formation (Sakmarian to Early Artinskian).

Interpretation: The occurrence of thicker, cross-bedded oolitic units immediately below the base of the prograding sabkha sequence of the Gipshuken Formation points to ooid formation and accumulation in near-coastal environments. The depositional environment is comparable to the wide intertidal sand flats along the Trucial Coast at Sabkha Matti (see Loreau and Purser 1973). There, the intertidal oolitic sands of the open coastal embayment are virtually uncontaminated by non-oolitic elements and extend to a water depth of about 2 m. The ooids were blown by "shamal" winds from the adjacent beach far into the sabkha coastal plain, where they form eolian dunes or thin eolian blankets. The latter may correspond to the thin oolitic grainstone beds that occur within thicker anhydrite units of the lower Gipshuken Formation.

5.14 Phylloid Algal Boundstone Facies

Description: The Facies is predominantly made up of whole and rarely fragmented phylloid algal plates and a peloidal groundmass. The most important genus of the phylloid algae is *Eugonophyllum*. Occasionally, grain types additionally include fenestellid bryozoans in larger quantities (forming Phylloid Algal – Bryozoan Boundstones). In contrast to the Palaeoaplysinid Boundstone Facies it is obvious that the surfaces of algal thalli as well as bryozoans are rarely encrusted or overgrown by *Tubiphytes* and small sessile foraminifers.

Phylloid algal plates may form a relatively open framework in which growth cavities are sediment-filled by peloidal packstone to wackestone (Pl. 17/9). Peloid grains commonly show graded lamination. Locally, bunches of the leaflike algal thalli are still in life position. They may, however, also be tightly packed, closely fitted together and piled up to form grain-supported fabrics (Pl. 17/8). Further additional biogenic elements such as shells of gastropods, ostracods, mollusks, brachiopods, dasycladacean fragments (*Epimastopora*), *Palaeoaplysina* and echinoderms rarely occur (Pl. 17/9).

The thallus of *Eugonophyllum* is composed of large blades which stood erect on the substratum (Mamet 1991). They were attached to the substrate by some type of rhizome and display dense clumps of vertical leaves (Wray 1977a, b). A reconstruction of the phylloid algal community is proposed by Toomey and Babcock (1983). The group is, however, morphological rather than biological and may include both red and green algae (Riding and Guo 1991). Most of the internal structure of the phylloid algae is lost during diagenesis because of the aragonitic primary mineralogy (see James et al. 1988). They occur as broken plates with a length up to several centimetres.

Occurrence: The facies is a typical constituent of the Minkinfjellet Formation and the Cadellfjellet Member of the Wordiekammen Formation, but only occurs locally within the Kapitol Member (Moscovian to Gzhelian).

Interpretation: Phylloid Algal Boundstone Facies is interpreted as a core facies of skeletal mounds. At Boltonbreen these mounds extend across a distance of about 50 m and reach an overall thickness of up to 9 m. They are massive or have a massive core surrounded by inclined flank beds with a relief of up to 4 m. Individual bioherms split laterally into stacked subunits. The skeletal mound facies is generally associated with subtidal platform packstone/wackestone facies which may also form interbeds at marginal parts of the skeletal mounds.

Within the studied material, the uniform and low diverse biota assemblage of this core facies is obvious. As pointed out by Toomey (1976), dense bushes of living phylloid algae induced community restriction on the seafloor, and the only organisms that could survive in such an environment were filter feeders that were attached to the phylloid leaves (e.g., encrusting foraminifers) and grazing animals that could browse and feed on them (e.g., gastropods). Further, organisms may have been washed into these bushes from outside situated environments (by storms). Fenestellid and ramose bryozoans may have colonised in marginal parts (see 5.15). As suggested by many authors, the phylloid algal thalli have the capability to baffle and trap suspension sediments (bafflestones) and thus create mound-like structures on shallow shelves (see Wilson 1975, Flügel 1980, Beauchamp et al. 1989, Morin et al. 1994, Stemmerik and Worsley 2000). Within the Cadellfjellet Member, the rare encrusters present on phylloid thalli played a negligible role in mound development. Stemmerik and Worsley (2000) described dolomitic phylloid algal mounds from the late Moscovian succession of Bjørnøya.

The predominance of phylloid algal bioherms in the central Billefjorden Trough sections, the transition with subtidal platform packstone/wackestone lithofacies, the prevalence of red and green algae, the lack of abraded grains, and the occurrence of peloidal matrix all indicate calm open marine environments with well-oxygenated, detritus-free water for phylloid skeletal mounds. One would expect the depositional setting to be below the storm wave base. Since the phylloid algal bioherms are integrated within shallowing-upward cycles (Pickard et al. 1996, Stemmerik and Worsley 2000), storms have affected mound growth and forced seaward progradation. The abundance of phylloid and dasycladacean algae suggest bioherm growth within the photic zone, not deeper than about 30 m (see Konishi and Wray 1961, Toomey 1976, Davydov and Krainer 1999).

5.15 Bryozoan / Phylloid Algal Facies

Description: This facies associated with the Phylloid Algal Boundstone Facies consists of bioclastic packstones and wackestones or more rarely of grainstones. Biota include common fenestellid and ramose bryozoans and phylloid algae. In general, reworked skeletal fragments of phylloid algae are broken but only slightly abraded and rounded (Pl. 17/7). Palaeotextulariids *(Climacammina, Cribrogenerina)*, tetrataxids, sessile foraminifers (*Ammovertella, Calcitornella, Tuberitina*), fusulinids, crinoids, *Tubiphytes*, *Ungdarella*, ostracods, gastropods and rare brachiopods may be minor constituents in changing proportions (Pl. 17/3). More than 50% of the rock volume may be taken up by the peloidal matrix. Conspicuous elements are fecal pellets which are concentrated in isolated nests (Pl. 17/4).
Occurrence: This facies occurs locally within the Ebbadalen Formation (Bashkirian) and is a typical constituent of the Minkinfjellet Formation and the Cadellfjellet Member of the Wordiekammen Formation at Gipsdalen (Moscovian to Gzhelian).
Interpretation: The Bryozoan / Phylloid Algal Facies mainly occurs as deposits associated with phylloid skeletal mounds. Obviously, fenestellid and ramose bryozoans may have settled in marginal parts and supported mound-building by baffling and trapping lime mud and bioclastic carbonate particles. The higher content of bryozoans – in comparison with the boundstones facies – may also be caused by the poor preservation potential of the reworked phylloid algal plates. The lithofacies of the buildup flanks is transitional to those of the surrounding bedded platform deposits and the massive buildup core. Beauchamp et al. (1991) described bryozoan reefs from shelf margin settings of the Sverdrup Basin that evolved upward into algal reefs.

5.16 Beresellid Boundstone Facies

Description: The small tabular green algae *Dvinella* (Pl. 16/7) forms bafflestones, sometimes in association with multithecoporid framestones. The interskeletal pores of the erect stick-like thalli are filled by cement (Pl. 16/8), a micritic matrix or more typically by encrusting *Tubiphytes* and sessile foraminifers (mainly *Tuberitina*). In the studied area the Beresellid Boundstone Facies forms massive biostromes. The open framework of multithecoporids and colonial rugosans frequently make up the base and grow close to the *Dvinella* bafflestones. Encrusting *Tubiphytes* has a secondary stabilising role during the baffling sediment accumulation. *Ungdarella* is locally common. Associated crinoids are sometimes preserved in the form of still articulated colonies.

Among trapped additional biogenic elements, small benthic foraminifers (especially *Biseriella* and *Bradyina*) dominate (Pl. 16/5). More rarely echinoderm plates, *Calcitornella*, fusulinids, tetrataxids and ostracods occur.
Occurrence: The Beresellid Boundstone Facies is a peculiar type of boundstone present locally within the Kapitol and Cadellfjellet Members of the Wordiekammen Formation (Moscovian to Gzhelian).
Interpretation: The Beresellid Boundstone Facies forms up to 2 m thick massive biostromes. At Balchinfjellet it occurs partly in place of the phylloid algal bioherms. According to Morin et al. (1994), beresellid algae grew in normal marine environments ranging from relative shallow depth to the lower limit of the photic zone. The palaeobathymetric position of the Beresellid Boundstone Facies (deduced from the vertical facies successions) corresponds with the situation observed in the Sverdrup Basin, which suggests a shelf setting at or near the fair-weather wave base. Stemmerik and Worsley (2000) suggest a similar environment for Beresellid Boundstones from Bjørnøya during the Moscovian.

5.17 Palaeoaplysinid Boundstone Facies

Description: *Palaeoaplysina* is made up of sheet-like plates, about 5 mm thick but up to 1 m long (Davies and Nassichuk 1973). The organism has been variously assigned to sponges and algae but is more probably related to hydrozoans (Breuninger 1976). Upper Palaeozoic buildups dominated by *Palaeoaplysina* are apparently restricted to the northern rim of the Laurentian subcontinent (e.g. Stemmerik et al. 1994).

In our material of the Palaeoaplysinid Boundstone Facies, sections perpendicular to the bedding exhibit abundant *Palaeoaplysina* plates in life-position forming a relatively open framework. In other cases, several *Palaeoaplysina* plates are closely superimposed and interparticle pores are partially filled with bioclastic wackestone to packstone material. In both cases the primary boundstone depositional texture is clearly indicated by the flat-lying growth form showing a conspicuously greater portion of epibionts (mostly

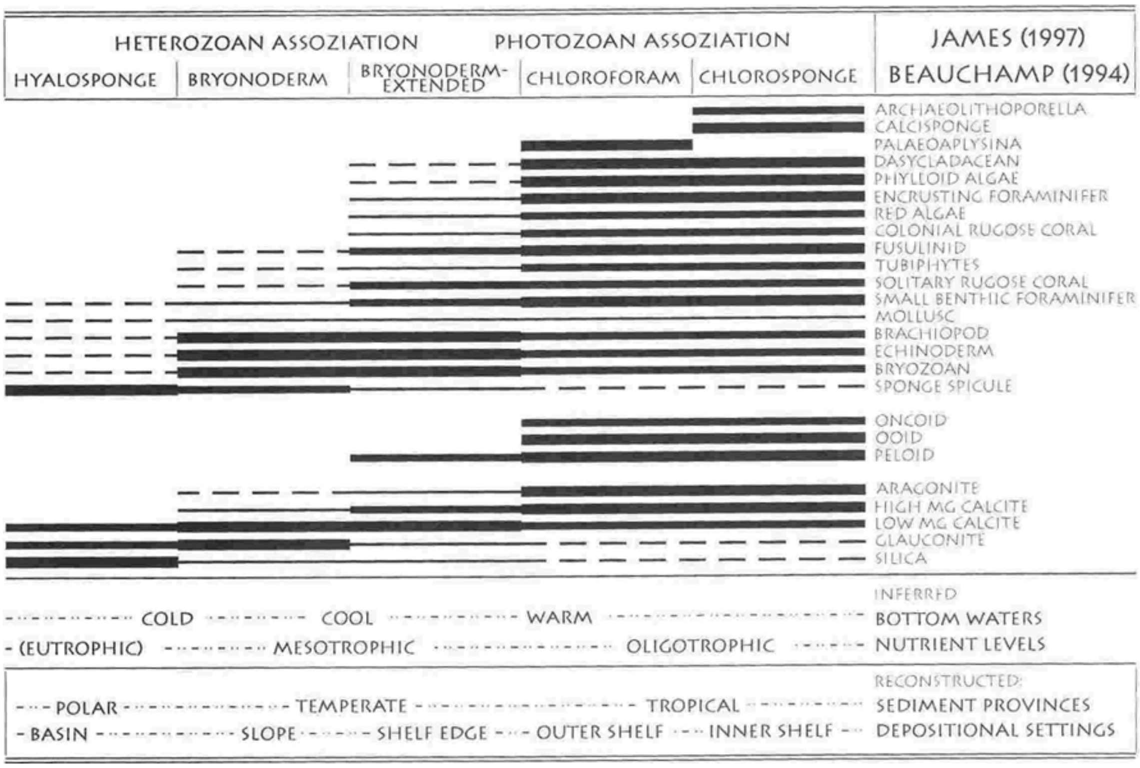


Fig. 8. Composition of benthic particle associations (biota, non-skeletal elements, and mineralogy) of late Palaeozoic marine carbonate sediments according to Beauchamp (1994) and tentative environmental indications (compiled after Lees and Buller (1972), Lees (1975), Hallock and Schlager (1986), Birkeland (1987), Brookfield (1988), Hallock et al. (1988), Föllmi et al. (1994), Beauchamp (1994), James (1997).

encrusting *Tubiphytes*, sessile foraminifers and rare *Ungdarella*) on the mamelon-bearing surface (Pl. 19/7). The latter was thus viewed as the upper surface of the organism by Breuninger (1976) and is generally oriented towards the top within the boundstone facies. Further indications of the primary growth-position of the sheet-like *Palaeoaplysina* plates are provided by the internal network of the channels. The complex branching channels emerge as pores on its outer surface, mostly on the upper mamelon side (Pl. 19/8). Only a few pores pierce the lower flat side of the plates. Therefore, peloidal and bioclastic groundmass predominately fills the pores of the upper sides (Pl. 19/9).

Encrusting *Tubiphytes* frequently forms micro-reefs on the prominent mamelon-like protuberances reaching a maximum height of 0.5 cm (Pl. 19/7). However, epibiont *Tuberitina* locally occur on the more or less flat lower side of the *Palaeoaplysina* plates. Therefore it is also plausible that some *Palaeoaplysina* grew as erect or semi-erect, semi-rigid organisms attached somehow to the substrate, oriented and tilted relative to the main currents (Beauchamp et al. 1989).

A characteristic feature of the Palaeoaplysinid Boundstone Facies is the accumulation of peloidal packstones and wackestones as groundmass between *Palaeoaplysina* plates, interlaminated with bioclastic material. Peloids exhibit geopetal fabrics within the growth cavities or within the locally infiltrated pores of the internal channels. Although some peloids are similar to fecal pellets, most of them show irregular shapes and diverse sizes (0.02 to 1 mm), suggesting that they are of a different origin (algal peloids or micritized particles).

Rugosans, multithecoporid corals and fenestellid bryozoans may occasionally settle among *Palaeoaplysina*. Phylloid algae (*Eugonophyllum*) and dasycladacean (*Epimastopora*, *Anthracoporella?*) occur as rare constituents (Pl. 19/5). Locally small benthic foraminifers are associated in larger quantities. Bryozoans, brachiopods, gastropods, crinoids, ostracods, mollusks and fusulinids represent the biogenic assemblages of the bioclastic wackestones-packstones that are baffled and trapped by *Palaeoaplysina* and fill the growth-cavities.

Remnants of the internal cellular structure are rarely preserved because the originally aragonitic *Palaeoaplysina* plates are affected by extensive recrystallization (Pl. 19/6). The outline of the internal network of channels is most clearly visible when they are either filled by peloidal or micritic material, or when the canal mouths are encrusted by diverse epibionts.

Occurrence: *Palaeoaplysina* bioherms are a typical feature of the Tyrrellfjellet Member of the Wordiekammen Formation (Asselian to Sakmarian). The bioherms are usually found at three stratigraphic levels (Fig. 5). Interbedded rocks mainly consist of dark bituminous fusulinid-rich limestones similar to the Brucebyen Beds. On Bjørnøya a tabular *Palaeoaplysina* complex formed during the late Gzhelian (Stemmerik et al. 1994).

Interpretation: The bioherms are integrated into small-scale shallowing upward cycles (Fig. 12/A). Furthermore, the facial successions of the *Palaeoaplysina* bioherms themselves reflect growth stages that typically arise during mound growth (compare Walker and Alberstadt 1975 and James

1983). Complete cycles ideally exhibit four stages (Fig. 11):
(1) Preparation stage: The beds immediately underlying the mounds are commonly made of dolomitic mudstones, bioclastic dolomicrosparites or dolosparites showing sparse bioclastic content and a distinct (algal ?) lamination in some places. The top surface of these mound under-beds have been interpreted as hardgrounds by Skaug et al. (1982). These authors also point to the sharp lithological change to the overlying mound facies.
(2) Stabilisation stage: At the base of the mounds the core facies is characterised by bioclastic packstones with common fusulinids, echinoderms and bryozoans. Individual bedding planes may be colonised by huge rugosan corals and rare multithecoporids which contribute to the stabilisation of the substrate. Initiation of mound accumulation is, however, indicated by the settlement of encrusting palaeoaplysinids with a clear, flat-lying orientation of their plates. Rudstones, including abraded colonies of rugosans, which are primarily interbedded within lower parts of the mounds, point to temporary reworking during storms.
(3) Colonisation stage: Massive to thick bedded units of *Palaeoaplysina* boundstones are characteristic of comparable high accumulation rates during active mound growth. The low, mono-specific diversity is highly indicative of the colonisation phase of the ecological succession. Only *Tubiphytes* and encrusting small foraminifers populate successfully in sub-environments – attached to the *Palaeoaplysina* plates. The formation of a "crestal" boundstone facies of encrusting foraminifers and/or *Tubiphytes* is interpreted by Wilson (1975) as an indication of growth into the active wave zone.
(4) Cessation stage: The beds immediately overlaying the mounds are composed of different types of facies including fusulinid packstones with high organic content (see 5.11), small-foraminifera wackestones (see 5.5), peloidal bioclastic wackestones and dolomicrosparites, which all more or less point to shallowing or emergence of slightly restricted environments as reasons for the cessation of mound growth. Grainstones indicating shoal sand caps are rare (see 5.12). Furthermore, spherical grouped clusters of *Microcodium* are detectable at the top of some mounds and in the beds overlying the mounds.

In concurrence with Skaug et al. (1982), we interpret the *Palaeoaplysina* bioherms as mound accumulations of a shelf lagoon environment. Initiation of mound growth occurred within open marine settings below the fair-weather wave base following transgression. The presence of dasycladacean algae point to a euphotic setting. During active mound growth, sediment accumulation (by baffling and trapping) continued close or within the fair-weather wave base (see below). The bioherm structures were frequently high enough above the seafloor to affect water circulation. Cessation of mound growth is interpreted as a result of increasing restriction within the shelf lagoon that was followed in some cases by subaerial exposure during a sea-level fall.

5.18 Palaeoaplysinid Floatstone Facies

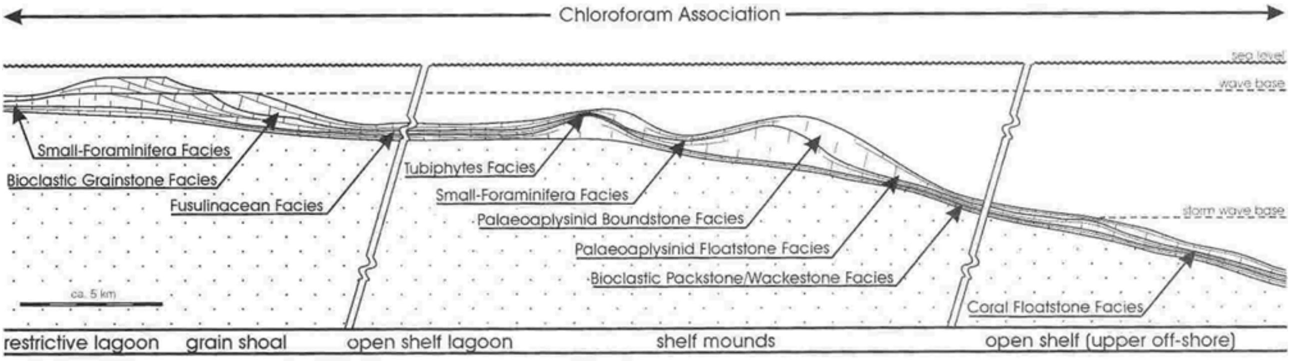
Description: The sheet-like plates of *Palaeoaplysina* are preserved as skeletal fragments up to some decimetres in length. This facies contains broken as well as abraded fragments of *Palaeoaplysina* which are embedded in bioclastic pack- to wackestones or more rarely in grainstones. Peloidal packstone fabrics are common as well. Redeposition of *Palaeoaplysina* is indicated by the loss of any preferred orientation of the skeletal fragments. Consequently, the upper surface of the organism may be turned upwards or downwards. Rudstone fabrics that comprise rounded, but never sorted, *Palaeoaplysina* fragments seldom occur.

The typical epibionts (mostly encrusting *Tubiphytes* and sessile foraminifers) of the bioherm facies are rarely preserved. Phylloid algae (*Eugonophyllum*), *Epimastopora* and other dasycladacean algae are infrequent yet distinctive biogenic elements. The Palaeoaplysinid Floatstone Facies is characterised by highly variable quantities of additional biogenic elements, commonly small benthic foraminifers (*Tetrataxis*, *Cribrogenerina*), fusulinids, bryozoans, brachiopods, crinoids, *Ungdarella* and occasionally ostracods.
Occurrence: The Palaeoaplysinid Floatstone Facies occurs in the vicinity of *Palaeoaplysina* bioherms of the Tyrrellfjellet Member of the Wordiekammen Formation (Asselian to Sakmarian).
Interpretation: The Palaeoaplysinid Floatstone Facies occurs as interstratified beds within the *Palaeoaplysina* bioherms as well as in form of sediments flanking the bioherms (Fig. 11). The presence of abraded or sometimes rounded grains is indicative of turbulent conditions within to the fair-weather wave base. This supports the shallowing upward trend indicated by the whole facial succession. However, the energy-level was usually too low in order to produce well winnowed sediments which are typical of the Oolitic and Bioclastic Grainstones Facies.

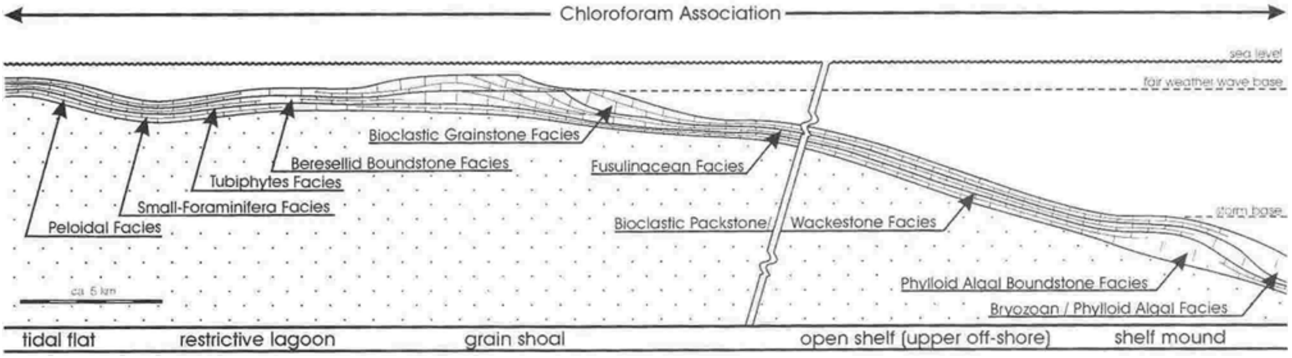
Environmental studies of *Palaeoaplysina* bioherms in Idaho (Breuninger et al. 1989), Yukon (Davies 1989) and Ellesmere Island (Morin et al. 1994) suggest that turbulent conditions favoured the growth of *Palaeoaplysina* plates over fast-growing but fragile phylloid plates. In the study area, phylloid algal mounds do not exhibit indications of growth into fair-weather wave base confirming a deeper depositional environment for phylloid algal mounds (see 5.14).

5.19 Spiculite Facies

Description: The Spiculite Facies is generally composed of abundant siliceous sponge spicules (Pl. 22/6) and is sometimes comprised of fragments of brachiopods, bryozoans and echinoderms that are deposited within bioturbated wackestones or packstones. Further additional biogenic elements such as nodosarid and frondicularid foraminifers and ostracods are seldom found. Isolated specimens of siliceous sponges and agglutinated sponges may be present (Pl. 22/5). Among the siliceous sponge spicules, different



Latest Gzhelian to Asselian (lower and middle Tyrrellfjellet Member)



Moskovian to Gzhelian (Minkinfjellet Formation, Kapitol and Cadellfjellt Members)

Fig. 9. Facies models of the Moscovian to Gzhelian (Minkinfjellet Formation to Cadellfjellet and Kapitol Members of the Wordiekammen Formation) and the Latest Gzhelian to Asselian (lower and middle Tyrrellfjellet Member of the Wordiekammen Formation) carbonate succession of Central Spitsbergen.

sizes can be discerned (up to 1 mm in length) reflecting some kind of bedding. Complex bioturbate textures range from obscure background mottling to well-preserved burrows, easily recognisable by spicules and other bioclasts arranged in circular patterns. Characteristic ichnotaxa seem to be *Schaubcylindichnus* and *Scalarituba* (Pl. 22/1). Conversely, the dominant trace fossils of pure spiculitic shales are *Zoophycos*, *Thalassinoides*, *Chondrites* and *Phycosiphon* (Stemmerik and Worsley 1995).

The Spiculitic Facies is usually rich in glauconite and occasionally contains small amounts of quartz silt. The glauconite usually occurs as rounded pellets (0.3 – 0.1 mm) which are evenly dissipated among other particles. Marly sediment types are characterised by peloidal fabrics. The carbonates contain silicified echinoderm plates and brachiopod shells that are made up of microquartz in varying orientations. Radial-fibrous quartz, known as chalcedonic quartz, is often a pore-filling cement.

The Spiculite Facies is characterised by gradual facies transitions exclusive to the Brachiopod Facies and the Crinoid Bryozoan Facies.

Occurrence: This Facies is the dominant facies of the Kapp Starostin Formation (Kungurian to Tatarian) and occurs in the upper part of the Ebbadalen Formation (Bashkirian).

Interpretation: The lack of high energy sedimentary structures, a low diverse open marine biotic assemblage, the abundance of micrite or siliciclastic mud and intensive bioturbation are clearly indicative of deeper shore-face to shelf-basin depositional settings below wave base. The Spiculitic Facies is subdivided into informal proximal (deeper shore-face to open-shelf) and distal (epicontinental basin) subfacies as evidenced by trace fossils, facial transitions and the discriminating features outlined by Beauchamp (1994): The proximal subfacies comprise medium- to light-grey spiculitic cherts and calcareous shales with common to abundant glauconite. Uniserial nodosarid foraminifers are indicative of additional biotic elements, but monaxon spicules in parts constitute the only biotic remnant. These sediments show intensive bioturbation and/or, more rarely, lamination. Characteristic ichnotaxa are *Schaubcylindichnus*, *Scalarituba*, *Thalassinoides* and *Chondrites*. They pass vertically (and laterally) into well-bedded packstones and grainstones of the Brachiopod Facies (see 5.9) and the Crinoid Bryozoan Facies (see 5.8). Calcareous and agglutinated sponges and bryozoans are found near the transition to the distal subfacies. The distal subfacies comprise black to dark-grey spiculitic cherts and siliceous shales. Prevailing small siliceous spicules are embedded in dark debris that is rich in pyrite and organic matter. *Zoophycos* and *Phycosiphon* are characteristic trace fossils. Additional biotic elements and glauconite rarely occur. Bioturbation is uncommon.

6 ENVIRONMENTAL RECONSTRUCTION

6.1 Modes of carbonate production

Starting from the basic work of Lees and Buller (1972) and Lees (1975), the analysis of particle association has proven to be useful for interpreting depositional settings of modern and ancient carbonate deposits (e.g. Henrich et al. 1992, 1995, James 1997, Betzler et al. 1997, Lukasik et al. 2000). Beauchamp (1994) summarised and defined the biotic and abiotic benthic associations found in late Palaeozoic carbonate sediments (see Fig. 8). While the Bryonoderm and Hyalosponge associations are exclusively produced by organisms that are light-independent, the Chlorosponge and Chloroforam associations additionally include skeletons of light-dependent organisms and/or non-skeletal particles such as oncoids, ooids and peloids. According to James (1997), the Bryonoderm and Hyalosponge associations are **Heterozoan associations**, whereas the Chlorosponge and Chloroforam associations are **Photozoan associations.** We use these benthic particle associations – defined by Beauchamp (1994) – in order to analyse the different environments and the developing sediment formation during late Carboniferous to early Permian time. Facies described in this paper are, however, only part of the larger late Palaeozoic basin fill (Fig. 1). Additional information comes from the St. Johnsfjorden Trough area (Fig. 3, Buggisch and Joachimski, unpubl. data).

6.2 Depositional facies models and depositional history

6.2.1 Bashkirian

The Bashkirian syn-rift succession (Ebbadalen Formation) records the long-term subsidence pattern related to the Mid-Carboniferous extension on Svalbard. While the basal unit is part of a transgressive mega-sequence (Ebbaelva Member), the upper unit represents a part of a regressive to transgressive mega-sequence (combined Trikolorfjellet and Odellfjellet Members) (Johannessen and Steel 1990, 1992).

The organisation into repeated, smaller scale sequences (up to 40 m thick) betrays short-term (~ 1 Ma) episodes of asymmetric subsidence, with initially rapid and then slower rates of base level rise during extension (Johannessen and Steel 1992). Individual sequences record the retreat and submergence of alluvial fans and the subsequent development of beach spit sandstones and marine carbonates, followed by renewed progradation and emergence, which in turn gives rise to the expansion of sabkhas, eolian flats and alluvial fans again. Marine carbonates which occur within the studied successions represent periods of maximum marine flooding, approximately corresponding to the times of the maximum rate of subsidence. Therefore, our knowledge of marine biota is limited. The predominance of siliciclastics limits the insights on snapshots.

The **Chloroforam association** is only found within calcareous interbeds of the Ebbaelven Member representing maximum marine transgression during the Bashkirian. The Bioclastic Packstone/Wackestone Facies (see 5.7) from open lagoonal settings include bioclastic remains of photoautotrophs such as green calcareous algae (*Eugonophyllum, Epimastopora*) and incertae sedis (*Tubiphytes, Palaeoaplysina*) among abundant foraminifers, bryozoans and brachiopods. Most of the bioclastic sediments are characterised by the **Bryonoderm association.** Depositional textures range from grain- to mud-supported, comprising the Bioclastic Grainstone Facies (see 5.12) and the Bioclastic Packstone/ Wackestone Facies (see 5.7). These are limestones and dolomites which commonly contain substantial amounts of siliciclastic particles. The **Hyalosponge association** is represented by the Spiculite Facies (see 5.19) which is most typically found in upper parts of the Odellfjellet Member. These are mostly dolomites containing substantial amounts of siliciclastic particles and authigenic glauconite.

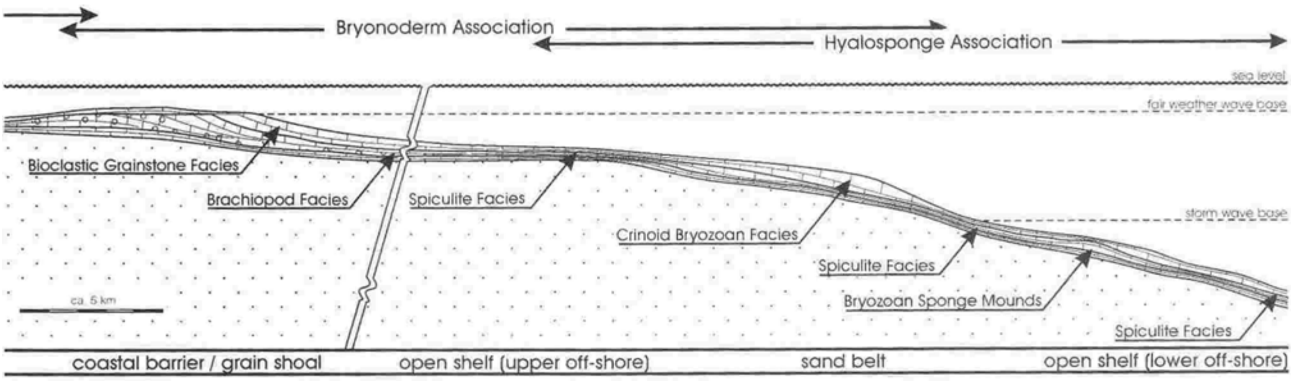
6.2.2 Moscovian – Asselian

The Moscovian to Gzhelian succession (Fig. 9) comprises a broad transgressive-regressive sequence (Minkinfjellet Formation to Cadellfjellet and Kapitol Members of the Wordiekammen Formation) (Samuelsberg and Pickard 1999). Sediments include marginal siliciclastics and evaporites, but mostly marine platform carbonates. The Latest Gzhelian to Asselian succession (Fig. 9) is also characterised by a broad transgressive-regressive sequence (lower and middle Tyrrellfjellet Member), bounded at its base by laterally extensive discontinuity surfaces (Steel and Worsley 1984, Samuelsberg and Pickard 1999). Platform carbonates of mostly open marine environments accumulated in the Billefjorden Trough and on the Nordfjorden Block.

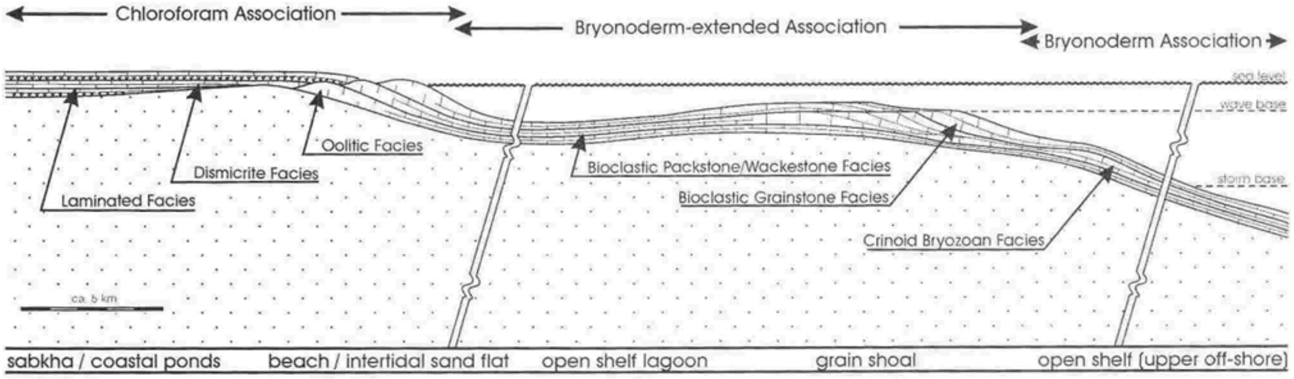
The Moscovian to Asselian carbonate succession is organized in stacked higher order cycles (5-20 m thick; see Fig. 12) and exhibits subaerial unconformities at numerous levels in shallow-water settings, which are indicated by mud cracks, flat pebbles, *Microcodium* aggregates and meteoric cementation (Skaug et al. 1982, Stemmerik et al. 1995, Pickard et al. 1996, Samuelsberg and Pickard 1999). These subaerial exposure horizons are the product of high-frequency, high-amplitude sea-level changes of glacio-eustatic origin (Stemmerik 2000).

The **Chloroforam association** is widespread in the neritic environments from areas with constant wave action reaching down into quiet water below the storm wave base. Consequently, the Bioclastic Grainstone Facies (see 5.12) produced at barriers and shoals as well as the Bioclastic Packstone/Wackestone Facies (see 5.7) of the open shelf include ubiquitous bioclastic remains of photoautotrophs such as green calcareous algae (*Eugonophyllum, Epimastopora, Dvinella*) and incertae sedis (*Tubiphytes, Palaeoaplysina*), among a variety of other organisms. The same is true for the Fusulinacean Facies (see 5.11) which occupies shallow subtidal environments, although vast numbers of fusulinids are characteristic. This facies probably indicates euphotic environments, since fusulinids are interpreted to have harboured photosymbionts (Cowen 1988, Brasier 1995).

Skeletal mounds formed by phylloid algae were widespread during Moscovian to Gzhelian time. These buildups



Late Artinskian - Kazanian (Kapp Starostin Formation)



Sakmarian - early Artinskian (uppermost Tyrrellfjellet Member - Gipshuken Formation)

Fig. 10. Facies models of the Sakmarian to Early Artinskian (uppermost Tyrrellfjellet Member of the Wordiekammen Formation and Gipshuken Formation) and the Late Artinskian to Kazanian (Kapp Starostin Formation) carbonate succession of Central Spitsbergen.

generally grew in subtidal environments with open marine circulation below fair-weather wave base. However, phylloid algae also formed skeletal mounds in association with bryozoans (see 5.14 and 5.15) within deeper photic environments below storm wave base. During Asselian time *Palaeoaplysina* bioherms were common (see 5.17and 5.18). Usually, they initially developed below the fair-weather wave base and became increasingly affected by waves during mound growth. Therefore, facial transitions are more complex in the vicinity of the *Palaeoaplysina* bioherms and shallowing upward successions are recorded during mound growth as proposed by Skaug et al. (1982). The latter are caused by high-frequency glacio-eustatic sea-level changes. Furthermore, isolated colonies of multithecoporids and rugosan corals are widespread in subtidal muddy sediments (see 5.10). Sometimes in association with beresellid algae, these corals formed biostromes (see 5.16) indicating euphotic conditions.

The dominance of photoautotrophs such as green calcareous algae probably indicates a high primary productivity on the shallow platforms. According to James (1997), these organisms are not as affected by increased nutrient levels (transitional oligotrophic to mesotrophic conditions) as certain hermatypic corals that harbour photosymbionts and live mixotroph. When nutrient levels rise, symbiotic animals cannot compete with faster-growing, fleshy algae and asymbiotic animals (Birkeland 1987).

6.2.3 Sakmarian – Early Artinskian

The Sakmarian to Early Artinskian succession (Fig. 10) is characterised by a general transgressive-regressive sequence bounded at its top by a major subaerial surface of exposure (Lauritzen 1983, Samuelsberg and Pickard 1999). The basal package comprises bioclastic shelf carbonates and, more rarely, mixed siliciclastic carbonates that culminate in intertidal oolitic grainstones (uppermost Tyrrellfjellet Member). The overlying part includes supratidal sabkha cycles and eolian deposits (Gipshuken Formation).

Generally, bioclastic shelf sediments with mud-supported textures display a typical **Bryonoderm association**. Therefore, mud-rich types of the Crinoid Bryozoan Facies (see 5.8) exhibit a poorly diversified fossil assemblage comprising crinoids, fenestellid and ramose bryozoans, brachiopods, bivalves and rare silicosponges. It is a typical accumulation of subtidal settings below the wave-base within open-shelf environments.

Bioclastic sediments showing grain-supported textures are characterised by the **Bryonoderm-extended association**, provided that the bioclastic particles are produced within shallow subtidal environments. Particles of the Bioclastic Grainstone Facies (\n 5.12) from shoals and the Bioclastic Packstone/Wackestones Facies (\n 5.7) common to open lagoonal environments include fusulinids, smaller benthic foraminifers and rugosan corals.

The **Chloroforam association** is solely represented by oolitic grainstones (see 5.13) which were formed in intertidal sand flats and probably redeposited in beach and eolian environments. Bioclastic interbeds of the supratidal deposits also include biotic remains of fusulinids, smaller benthic foraminifers and gastropods and/or non-biotic particles such as peloids. In most cases, however, it is doubtful whether the formation of these sediments is autochthonous.

6.2.4 Late Artinskian – Kazanian

The studied Late Artinskian to Kazanian succession (Fig. 10) is characterised by a general transgressive trend. Renewed tectonic activity and inversion along the palaeo-Hornsund lineament led to regression and erosion in southwestern Spitsbergen (Stemmerik et al. 1995). Early Artinskian peritidal dolomites of restricted marine environments (upper Gipshuken Formation) are unconformably overlain by open marine limestones, cherts and spiculitic shales (Kapp Starostin Formation).

Transgressive-regressive cycles (40 – 60 m thick) might be regarded as third-order depositional sequences (Stemmerik 1997). Late Artinskian to Kazanian strata include bioclastic sandy shoal sediments and subtidal limestone deposits with a poorly diversified fossil assemblage comprising crinoids, bryozoans, brachiopods, bivalves and silicosponges. Oolitic sediments are absent. Massive chert bodies are interpreted as sponge buildups by Steel and Worsley (1984). The carbonates are interbedded with sandstones and spiculitic siltstones.

The **Bryonoderm association** is found within the Bioclastic Grainstone Facies (see 5.12) from depositional areas above wave base and within the Crinoid Bryozoan Facies (see 5.8) from settings below wave base. Furthermore, the Bryonoderm association characterises the Brachiopod Facies (see 5.9) that is typical of inner platform areas around wave base and of shoreface deposits. Usually, these limestones are rich in glauconite and lack carbonate mud. The Bryonoderm association is also widespread in siliciclastic inner shelf deposits.

The ecological zones occupied by brachiopods in inner shelf deposits during late Palaeozoic times are taken up by bivalves in modern environments. This suggestion is documented by facial investigations on Svalbard and North Greenland (Malkowski and Hoffmann 1979, Ezaki et al. 1994, Stemmerik 1997). Therefore, the Brachiopod Facies is possibly an Upper Palaeozoic equivalent to the poorly sorted mollusk sand of Boreen et al. (1993) and the coarse-grained bivalve sand of James et al. (1997). Furthermore, the Crinoid Bryozoan Facies resembles outer-shelf bryozoan-dominated sands, deposited between the fair-weather and storm wave base on modern cool-water shelves (see James et al. 1992, 1997, Boreen et al. 1993). This interpretation is in accordance with the conclusions inferred by Stemmerik (1997). The Late Artinskian to Kazanian sediments most likely were deposited at much shallower depth than comparable modern facies off southern Australia (Stemmerik 1997). The critical interface of wave abrasion is expected to be at much shallower depth in intracratonic basins like those discussed here.

The **Hyalosponge association** is found within the Spiculite Facies (see 5.19) which represents subtidal environments below wave base that range from shallow shelf to deeper shelf-basin settings. Kungurian spiculitic limestones and cherts are characterised by intensive bioturbation, high glauconite content, the presence of foraminifers and interstratification with Bryonoderm carbonates. These features may indicate deposition at shallow subtidal areas (see Beauchamp 1994). A deposition within deeper subtidal shelf-basin areas is inferred for cherts and spiculitic shales with high organic matter content, limited bioturbation and a lack of glauconite.

Even within restricted lagoonal and intertidal environments, documented by the uppermost strata of the Gipshuken Formation, sponge spicules are locally abundant. These spicules occur within dololaminites showing open-space structures and cryptalgal fabrics. Although these spicules are most probably allochthonous particles redeposited during storms on tidal flats, their occurrence indicates that the spread of the Heterozoan association reached up into the shallowest shelf areas.

The interstratification of the Spiculite Facies with Bryonoderm carbonates gives rise to significant facial transitions. Common bryozoans and crinoids represent transition to the Crinoid Bryozoan Facies (see 5.8). Brachiopods are more abundant towards the transition with the near-shore Brachiopod Facies, which represents the shallowest water facies dominated by Heterozoan associations. Calcareous and agglutinated sponges and bryozoans are found near the transition to deeper shelf spiculitic shales.

6.3 Discussion of environmental conditions

According to Lees (1975), temperature, and to a lesser extent salinity, control the world-wide distribution of the Photozoan and Heterozoan associations. In modern oceans, a compositional change in bioclastic sediments can be observed both from the equator to the poles and with increasing water depth (Fig. 8). This is caused by decreasing water temperature and salinity in both directions (Brookfield 1988, James 1997). In consequence, carbonates of the Photozoan associations are restricted to shallow-water environments at low latitudes, where carbonates of the Heterozoan associations concurrently occur in deeper-water environments. With increasing latitude, we observe a progressive replacement of Photozoan associations by Heterozoan associations. Therefore, low-latitude deep-water associations are similar to high-latitude shallow-water associations (Beauchamp 1994).

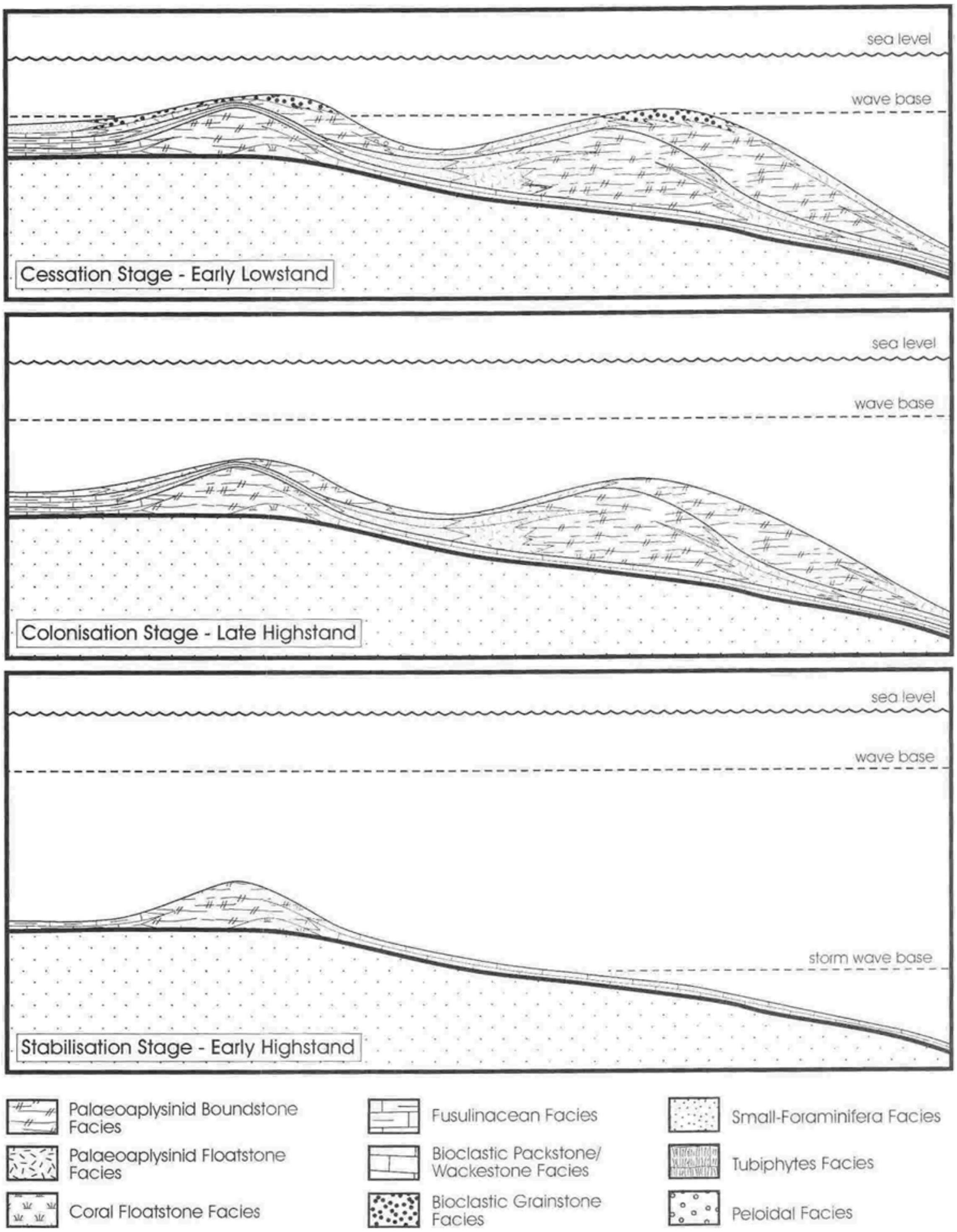


Fig. 11. Facies relationship and stratigraphic succession of the *Palaeoaplysina* bioherms that typically arise during and after skeletal mound growth. The stabilisation stage, the colonisation stage and the cessation stage are shown (for explanation see 5.17).

However, a similar progressive replacement of the Photozoan associations already occurs if nutrient supplies increase (Fig. 8). When nutrient levels rise, photo-symbiotic animals cannot compete with faster-growing, fleshy algae and asymbiotic animals (Birkeland 1987). Increased levels of nutrients also increase planktic biomass and reduce light penetration. In such cases, Heterozoan associations will predominate even in shallow-water environments, and occasionally in photic environments at low latitudes (James 1997). Details of these theories were discussed by James (1997), Beauchamp (1994), Brookfield (1988) and Birkeland (1987).

Furthermore, an interpretative enigma arises because nutrient availability is often intimately tied to temperature and salinity (Hallock and Schlager 1986). By analogy with the present, Föllmi et al. (1994) distinguish between two substantially different carbonate production modes which drive the carbonate factory. The coral-oolite mode of carbonate production documented by Photozoan associations reflects "healthy" platform growth conditions with high average water-surface temperatures, high transparency of the water column, and correspondingly low nutrient levels. The crinoid-bryozoan mode of carbonate production documented by Heterozoan associations reflects mesotrophic environmental conditions with benthic communities of fast-growing, less-specialised, and overall versatile suspension feeders such as stalked crinoids, bryozoans, silicosponges, brachiopods and others. Finally, eutrophic environments are characterised by minimal carbonate accumulation, phosphogenesis and glauconite formation (Föllmi et al. 1994).

A compositional shift in benthic particle associations of shallow-water environments has been mostly interpreted in terms of changing water temperatures. Consequently Beauchamp (1994) and Stemmerik (1997, 2000) have proposed a dramatic climatic cooling in northern Pangaea during the Permian – possibly in response to changes in oceanic circulation patterns – inferred from platform carbonates of the Sverdrup Basin and the Barents Sea. On the other hand, there is an increasing awareness of a relationship between nutrient supply and carbonate production (Hallock and Schlager 1986, Hallock et al. 1988, Föllmi et al. 1994). As a result, it becomes enormously difficult to draw absolute conclusions because nutrient availability is often intimately tied to temperature and salinity.

The biotic change from a (foraminifera-algae-dominated) Chloroforam association to a (bryozoan-echinoderm-dominated) Bryonoderm-extended association is apparently abrupt on the Finnmark carbonate platform off northern Norway and takes place during the late Sakmarian across a 3-cm shale horizon (Ehrenberg et al. 1998). No lithological evidence is seen for a hiatus at this contact. According to the authors, this rapid change resulted from a reorganisation of the oceanic circulation, that introduced progressively cooler waters over the previously warm-water shelves, triggered by a major rise in sea level. The most important change in palaeogeography from the Kungurian to the Kazanian is the opening of a sea-way along the axis of the Greenland-Norway rift from the Barents Sea southwards to the Zechstein basin of northern Europe (Stemmerik 2000).

The platform morphology also changes with the carbonates production mode (James 1997). This has been shown for Permian carbonate platforms in Greenland by Stemmerik (1997). The carbonate factory is located in shallower high-energy settings on Photozoan carbonate platforms in East Greenland, whereas on Heterozoan carbonate platforms in North Greenland the cool-water-factory is located in protected settings below wave base (Stemmerik 1997). The lack of protective rims means that oceanic currents can bring cool and nutrient-rich waters on the platforms.

In the case of the late Permian, many authors have interpreted the entire deep ocean as being devoid of oxygen. The analysis of isotopic data and pelagic cherts of the Panthalassa suggests a stepwise change from a ventilated ocean during early Permian times to a stratified ocean and finally to a superanoxic ocean at the Permian/Triassic boundary, which was accompanied by an upward propagation of anoxic water almost to the ocean surface and into shelf areas (e.g. Margaritz and Schulze 1980, Holser 1984, Isozaki 1997). Furthermore, substantial amounts of silt- and sand-sized siliciclastic particles are commonly used as an approximation of higher nutrient-influx levels derived from the land accompanying the import of detrital particles (Föllmi et al. 1994).

6.4 Conclusions

The depositional evolution of late Palaeozoic epicontinental carbonates from Spitsbergen was controlled by regional and local factors such as tectonic activity, subsidence patterns, relative sea-level, climate and distances to siliciclastic sources (Steel and Worsley 1984, Johannessen and Steel 1992, Pickard et al. 1996, Stemmerik 1997, 2000, Stemmerik et al. 1995, Samuelsberg and Pickard 1999). The primary control on the cyclic depositional pattern is believed to be glacio-eustacy (Stemmerik 1997, 2000). The carbonate production mode has been mainly changed (1) by a significant cooling within shallow-water marine environments of the Barents Sea from the Bashkirian to Kazanian and (2) by increased nutrient levels on the shelf during the Bashkirian and during the Sakmarian to Kazanian. Both extrinsic quantities of influence superimpose each other and cause a dramatic facial change in shallow marine biota associations during the Sakmarian to Late Artinskian time. Chlorosponge associations with abundant hydrozoans and calcisponges (*Inozoan*, *Sphinctozoan*), red algae (*Archaeolithophyllum*) and *Archaeolithoporella*, which characterize shallow-water carbonates of the Palaeotethyan region, were not found.

(1) Cooling: The cooling trend is indicated by a stepwise disappearance of shallow marine biota and their replacement by organisms that were previously thriving in deeper and necessarily colder environments of the water column. From Moscovian to Asselian time Chloroforam associations were widespread within neritic environments from areas with constant wave action reaching down into quiet water below the storm wave base. During the Sakmarian, Chloroforam associations disappeared from most shallow marine environments but remained at intertidal depositional settings bordering a supratidal coastal sabkha environment. Subsequently, euphotic and high-energy environments above the fair weather wave base were occupied by Bryonoderm-extend associations, whereas subtidal environments were taken up by the Bryonoderm associations. During the Artinskian-Kungurian, Bryonoderm-extend associations were taken over from Bryonoderm associations within water-agitated environments above the fair weather wave base. Simultaneously, Hyalosponge associations began to occupy increasingly greater portions of the wide subtidal shelf.

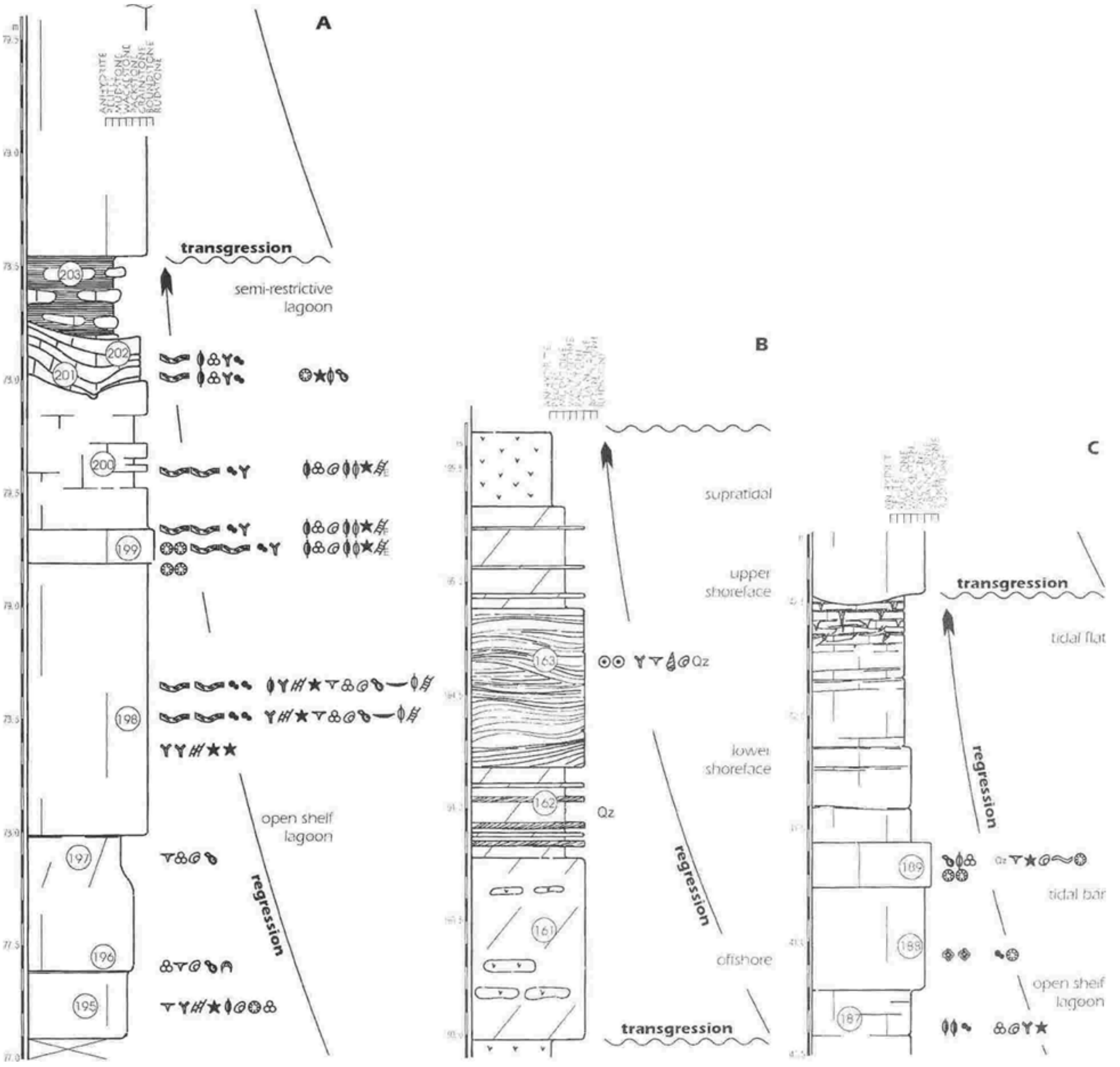


Fig. 12. Details of shoaling-upward carbonate sequences representing short-term cycles of (A) the Tyrrellfjellet Member at Feyling-Hanssen-Fjellet, (B) the lowermost Gipshuken Formation at Skansen (Alvrekdalen) and (C) the upper Kapitol Member at Feyling-Hanssen-Fjellet (for location compare Fig. 6).

(2) Mesotrophic condition: The disappearance of Chloroforam associations and Bryonoderm-extended associations from euphotic and agitated shallow-marine environments and their replacement by Bryonoderm associations and Hyalosponge associations may point to enhanced nutrient levels. Such mesotrophic environmental conditions are inferred when substantial amounts of glauconite and/or ripe, silt- and sand-sized siliciclastic particles also occur. The latter are used here as an approximation of higher nutrient-influx levels accompanying the import of detrital particles (see Hallock et al. 1988, Föllmi et al. 1994, Lukasik et al. 2000).

Mesotrophic conditions, restricted to near-shore settings, already occurred during the Bashkirian. This is indicated by significant quantities of silt- to sand-sized siliciclastics and rare authigenic glauconite within bioclastic carbonates from shoals and subtidal lagoons characterised by Bryonoderm associations. Given the semi-protected inland setting of the Bashkirian basin and the syn-tectonic nature of the basin fill (see Steel and Worsley 1984), nutrients are more likely to have come from the western edge of Billefjorden Trough than from a zone of coastal upwelling. Erosion of older coal-bearing deposits and clastics (Billefjorden Group and Old Red) on the Nordfjorden Block and its basin-ward redeposition by alluvial fans and marine fan deltas caused the raised amount of siliciclastic sediments and nutrients along the basin margin (Fig. 3). Fluctuation in the amount of siliciclastics and nutrients being delivered to the shallow waters affected prevailing levels of trophic resources and gave temporary way to the formation of pure limestones in central and eastern parts of the basin which are characterised by Chloroforam associations.

On the contrary, oligotrophic environmental conditions controlled the neritic accumulation from Moscovian to Asselian times. This can be seen in the widespread Chloroforam associations within shallow marine environments (Fig. 9). However, the carbonate factory is in fact not dominated by photo-symbiotic living corals. Instead, reef mounds and biostromes that occur within the photic zone were formed by phylloid algae, beresellid algae, *Tubiphytes* and *Palaeoaplysina*. This may point to slightly raised nutrient levels (oligotrophic to mesotrophic conditions) having been already present during this time.

The Asselian to Kazanian period is marked by a larger-scale trend from the oligotrophic conditions of the Chloroforam association (Fig. 9) to mesotrophic conditions such as those represented by the Bryonoderm- and Hyalosponge associations (Fig. 10). Increasing amounts of silt- and sand-sized terrigenous clastics, authigenic glauconite formation and condensation accompany this trend from the Kungurian onwards. In deeper, basinal settings, sedimentation took place in protected areas with dysoxic bottom water (Stemmerik and Worsley 1995). Unfortunately, we are not able to determine the nature of nutrient delivery. One obvious candidate for a nutrient source is oceanic upwelling. There is, however, no information available concerning an upwelling of nutrient-rich waters from coeval strata representing outer open-shelf settings of the Barents Sea – Sverdrup Basin shelf area of the Pangaea. Interestingly enough, thought, extensive upwelling led to the deposition of the Phosphoria Formation along the western margin of North America from the Artinskian onward (Sheldon 1989, Herring 1995). Furthermore, chamositic(?) marl/siltstones and a 50 cm thick phosphorite were cored at the inner shelf on the Finnmark Platform in late Permian sediments (Ehrenberg et al. 1998). A possible indication of increased fluvial run-off, which may also be important for delivering more nutrients, is the distribution of heavily bioturbated sandstones in the northwestern part of the basin, particularly during the Kazanian (Steel and Worsley 1984, Stemmerik and Worsley 1995, Stemmerik 1997). This transition from biogenic to siliciclastic sedimentation becomes younger westwards, from Early Permian adjacent to Novaya Zemlya to Late Permian in the Norwegian Barents Sea (Doré 1991) and is related to the Ural Orogeny farther to the east (Johansen et al. 1993). Coeval strata of the Sverdrup Basin received clastic sediments from a northern landmass called Crockerland (Embry 1992). A further explanation could be the rapid transgression of a nutrient-rich, oxygen-depleted water mass onto an oligotrophic platform (e.g. Schlager 1981, 1989, Schlanger et al. 1987). The gradual change from a ventilated ocean during early Permian times to a stratified ocean and finally to a superanoxic ocean across the Permian/Triassic boundary is connected with such an upward propagation of anoxic water almost to the ocean surface and into shelf areas as proposed by Margaritz and Schulze (1980), Holser (1984) and Isozaki (1997).

In general, we confirm a late Palaeozoic evolution from warm-water shelf conditions (arid palaeoclimate) during the Moscovian-Sakmarian to cool and cold conditions (temperate palaeoclimate) during the Artinskian-Kazanian within shallow-water environments. Such a palaeoclimatic shift, in part related to the northward drift of Pangaea and possibly in response to changes in oceanic circulation patterns, was observed in the Barents Sea area (Steel and Worsley 1984, Stemmerik and Worsley 1989, Stemmerik 1997, 2000) and in the Sverdrup Basin (Beauchamp 1994, Beauchamp and Desrochers 1997). The rapid environmental change has, however, been accelerated by a shift from nutrient-poor to nutrient-rich shelf waters during the Sakmarian-Artinskian.

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