MICROPALEONTOLOGY, DEPOSITIONAL ENVIRONMENT, AND BIOSTRATIGRAPHY OF THE UPPER PROTEROZOIC HEDMARK GROUP, SOUTHERN NORWAY

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ABSTRACT. Abundant and relatively well preserved microfossils were discovered in turbidite shales, mudstones, sandstones, and matrix supported conglomerates from several formations of the Upper Proterozoic-Lower Cambrian Hedmark Group in the allochthonous Hedmark Basin in southern Norway. Various microfossil morphotypes are attributed to the probably chroococcalean genera Spheeroconggregus Moorman, 1974 and Palaeogomphosphaeria Palacios Medrano, 1989.

Organic-rich black hemipelagic shales in turbidite sequences of the Brøttum Formation generally yield vast numbers of possibly cyanobacterial coccoids, whereas organically more lean rocks of the Biskopåsen and Biri Formations also yield acritarchs. Acritarchs (encysted and/or motile life stages of algal protists) are rare in the Hedmark Group.

The carefully identified facies associations of the Hedmark Group preceding the Varanger glacial event are largely the result of turbidite deposition in a tectonically labile, largely stratified, and anoxic intracratonic rift basin where primary productivity might have been restricted to the uppermost oxygen-rich layer. Upwellings might have been brought about by various mechanisms, of which seasonal changes in wind direction and air pressure and local factors causing inversions of the stratified, largely anoxic, water body are but a few. We speculate that tectonically controlled sedimentation of high-density turbidity currents and other types of gravity-flows may have impelled phosphate-rich bottom waters to circulate and flow into the shelf areas. Productivity surges could have been followed by plankton blooms, enhanced burial of organic matter, and formation of sedimentary phosphates. A large part of the recorded microbiota could be interpreted alternatively as anoxygenic photosynthetic bacteria occupying the H₂S-rich waters of stratified suboxic basins with a very shallow chemocline. Under the depicted conditions a substantial contribution to productivity could have been supplied by anoxygenic photosynthesis.

Micropaleontological data are in agreement with ages provided by isotopic dating. Acritarchs from units preceding the Vendian Varanger glaciation in southern Norway are considered contemporaneous with the intracratonic, Upper Riphean-Vendian Visingsösö basin in southern Sweden.

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INTRODUCTION

The well-known silicified (chert) microbiotas from Precambrian laminated mat communities and stromatolites consist, with few exceptions (Knoll and Calder, 1983), of autochthonous associations of microorganisms which generally allow taphonomic study and paleoecological interpretation (Hofmann, 1976; Knoll and Golubic, 1979). In general, inferences based on gross morphologic similarity, life cycles, and paleoecological parallels with extant photoautotrophic bacteria indicate that these biotas contain many components present in modern laminated microbial communities. Rare components, interpreted in part as occasional occurrences of allochthonous plankters, are also part of laminated microbial assemblages (Knoll, 1981; Knoll and Calder, 1983; Vidal, 1976a).

Organic-walled remains of cyanobacteria and encysted and/or motile life stages of eukaryotic protists are found in relative abundance in siliciclastic and carbonate rocks. Because of inherent problems in establishing their proper taxonomic affiliation, they are generally treated among the acritarchs (Downie, Evitt, and Sarjeant, 1963; Downie, 1973), although morphologically simple forms are occasionally treated among the cryptarchs (Diver and Peat, 1979). Acritarchs and associated microfossils of presumably cyanobacterial affinity are found in various kinds of sedimentary rocks thought to have formed in a variety of depositional settings. Generally, and with but very few exceptions, facies associations and sedimentological evidence indicate marine depositional conditions (Al-Ameri, 1983; Dorning, 1981). However, in a few instances acritarchs are believed to have accumulated in non-marine environments (Balme and Sergroves, 1966; Churchill and Sarjeant, 1962; Harland and Sarjeant, 1976).

In this paper we report on the occurrence of microfossils (acritarchs and possible planktonic cyanobacteria and/or photosynthetic bacteria) from Upper Proterozoic sedimentary rocks in the Sparagmite Region in southern Norway. Several carefully studied sections through critical parts of the Hedmark Group are described for the first time. The sedimentological data are here used to build up a paleoenvironmental and depositional frame for the investigated basin. We consider that well constrained models of basin development are of paramount importance in paleobiology. Hence, the microfossil assemblages from investigated Upper Proterozoic rock units are here set in relation to models of basin evolution and deposition. Acritarchs from Lower Cambrian strata in the Lake Mjøsa district merit separate treatment and will be the subject of a forthcoming paper.

Most studies of Proterozoic acritarchs have dealt with shallow marine, platform edge depositional settings (Vidal, 1981), although well documented Late Proterozoic (Riphean and Vendian) deep-shelf deposits and turbidites have been the objects of extensive micropaleontological investigation (Mansuy and Vidal, 1983; Palacios Medrano, 1989;
Vidal and Siedlecka, 1983; Vidal, 1990). This is composed of organic-rich, laminated pelagic deposits (Palacios Medrano, 1989) which contain microfossils that, following the contentions by Vidal and Knoll (1983), were interpreted as planktonic chroococcalean cyanobacteria (Mansuy and Vidal, 1983; Palacios Medrano, 1989; Vidal and Siedlecka, 1983). The present study contributes to the knowledge of the Proterozoic microbiotas in turbidite-dominated depositional settings.

**GEOLOGY**

Upper Proterozoic sedimentary rocks in southern Norway occur within a series of stacked nappes and thrust sheets in the eastern thrust belt of the Scandinavian Caledonides (fig. 1). The dominating coarse-grained feldspathic arenites of these allochthonous sequences were termed “sparagmites” by Esmark (1829), and this term has since been used as a trivial name for these rock units and the term “Sparagmite Region” for their outcrop area.

The material dealt with in this paper derives from strata in the lowermost nappe unit in the Sparagmite Region, the Osen-Røa Nappe Complex. The nappe complex consists of thrust sheets of crystalline basement rocks, the minimum 3000 to 4000 m thick Upper Proterozoic-Lower Cambrian Hedmark Group and up to about 2000 m thick epicontinental Cambrian-Silurian strata. The sole thrust is located within Cambrian shales of the eastern foreland succession, and thrusting and folding fade out in the southern part of the Oslo Region south of Lake Mjøsa (Oftedahl, 1943; Strand, 1960; Bockelie and Nystuen, 1985; Morley, 1986).

The Hedmark Group, formally defined by Bjørlykke, Englund, and Kirkhusmo, (1967), consists of eight formations in the Lake Mjøsa type area (figs. 1 and 2). Additional formations, also including a tholeiitic basalt unit, are defined from the eastern and northern parts of the Sparagmite Region (Sæther and Nystuen, 1981: Nystuen and Siedlecka, 1988). Internal tectonic deformations in the Hedmark Group include imbricate fans and duplex structures as well as broad open-folds and small-scale folds and thrusts (Nystuen, 1983; Morley, 1986). Penetrative fracture cleavage and foliation have modified sedimentary structures and textures in the northern part of the Sparagmite Region, but in the southern Mjøsa-Rena area (fig. 1) sedimentary textures are only distorted in local thrust zones. The rocks are altered into the lower part of the greenschist facies (Englund, 1973a; Morad, 1988).

The paleogeographic position of the sedimentary basin of the Hedmark Group has been widely discussed (for summary, see Nystuen, 1981: Bockelie and Nystuen, 1985). Previous (Oftedahl, 1943; Strand, 1960) and recent palinspastic restorations (Morley, 1986) suggest a minimum tectonic displacement of 130 to 150 km for the sequence in the Lake Mjøsa area. Nystuen (1981, 1982), Kumpulainen and Nystuen (1985), and Nystuen (1987) restored the Hedmark Basin about 200 to 250 km to the northwest, formed here as one of several similar basins by
riifting during the late Proterozoic (about 800–650 Ma) break-up of the Baltoscandian-Laurentian supercontinent (Gale and Roberts, 1974; Gee, 1975; Schwab, Nystuen, and Gunderson, 1988). The rift basin, though located in a continental setting, was in the west open to a marine seaway which by later sea-floor spreading in the Cambrian and early Ordovician evolved into the Iapetus Ocean. From geochemical studies on the boron
Fig. 2. Idealized composite section through the Hedmark Group in southern Norway. Bracketed vertical bars indicate the approximate stratigraphic position of sections in figures 3 to 5 and of sampling localities in text and in app. 1.

content in the Hedmark Group rocks, Bjørlykke and Englund (1979) concluded that the basin water was fresh or brackish, except for the depositional time for the Biri Formation. The fossil microbiota from the Hedmark Group reported in this paper indicate that the basin was connected to a marine seaway for most of its history.

Manum (1967) and Spjeldnæs (1963, 1967) first reported microfossils from the Hedmark Group in the Mjøsa type area. Later finds are briefly mentioned by Vidal (1981, 1990). The fossiliferous units in the
Mjøsa-Rena area (fig. 1) were deposited in the southwestern part of the Hedmark Basin. The general lithostratigraphy and sampled intervals of the Hedmark Group from this area and the central and southeastern part of the Sparagmite Region are shown in figure 2.

Brøttum Formation.—The Brøttum Formation is at least 3000 to 4000 m thick and is dominated by coarse- to fine-grained feldspathic sandstones with intercalated black carbonaceous shales. In the Østerdalene district (fig. 1) the formation also includes conglomeratic facies within massive, thick-bedded coarse-grained arkosic sandstones. It is suggested here that the unit represents a system of prograding fan deltas (Bjørlykke, Elvsborg, and Høy, 1976; Sæther and Nystuen, 1981; Nystuen, 1982, 1987). These fan deltas were supplied from the east, and the about 3000 m thick fluvial Rendalen Formation in the eastern and northeastern part of the Sparagmite Region is interpreted as a lateral equivalent to the marine Brøttum Formation in the west (Nystuen, 1982, 1987).

In the Mjøsa-Gudbrandsdalen area (fig. 1) the amount of black shale in the Brøttum Formation is up to 25 percent. The TOC (total organic content) in the shales is up to 2 percent and consists of pyrobitumen. The individual sandstone beds vary in thickness from some few centimeters to 12 to 15 m. Beds ranging in thickness about 0.01 to 2 m are usually normally graded, comprising the Bouma a and b intervals (massive and planar lamination). Current ripples (Bouma interval c) are present infrequently in the uppermost part of the graded beds, but most commonly the sandstones pass into a thin laminated, dark gray very fine-grained sandstone or siltstone interval (Bouma d), or they are directly overlain by black hemipelagic shale. The thick and very thick sandstones can be graded or ungraded and are generally massive and consist of two or more amalgamated subunits. Clasts of black shale are normally present in the sandstones, and sole marks like flute, groove, and gutter casts are present. Load-deformed, contorted, and slumped sandstone beds are common within shale-rich intervals, and in these sequences sand beds may have been mobilized by pore-water expulsions from compacting muds, forming intruding sand veins or thin dikes.

The Brøttum sandstones in the Mjøsa-Gudbrandsdalen area were deposited from turbidity currents (Englund, 1966, 1972, 1973b; Bjørlykke, Elvsborg, and Høy, 1976; Nystuen, 1982, 1987). The beds occur in cycles suggested here to represent sandstone channel-fill sequences, sandstone-shale levee-facies, and inter-channel and inter-lobe facies dominated by hemipelagic organic-rich shale and thin-bedded fine-grained turbidites. The Brøttum sequence in this area is located in a lower stratigraphical level than the fan-delta facies in Østerdalene to the east, and the turbidites are assumed to have been deposited from the west to the east (Englund, 1972, 1973b; Nystuen, 1982, 1987).

Two sections in the City of Lillehammer were sampled. Most fossils have been recorded from the Maihaugen section (fig. 3). All samples are finely laminated hemipelagic black shales. The organic-rich mud likely
settled vertically from suspension plumes and from highly diluted, low-density turbidity currents in tails of major turbidity currents.

_Biskopåsen Formation._—Biskopåsen Formation, also termed Biskopåsen Conglomerate (Bjørlykke, Englund, and Kirkhusmo, 1967), consists of several large wedge-shaped conglomerate bodies that interfinger into Brøttum turbidite sandstones and shales of the Birí Formation in the southern and western part of the Hedmark Basin (Bjørlykke, 1966; Løberg, 1970; Englund, 1966, 1972, 1973b; Bjørlykke, Elvsborg, and Hoy, 1976). In these areas the formation attains a maximum thickness of 200 to 400 m, whereas the stratigraphical equivalent, Imsdalen Conglomer-
Fig. 4. Lithostratigraphical section through part of the Biri Formation and Biskopåsen Formation at Hjelland, western side of Lake Mjøsa (585400/671900). Amalgamated contacts are stippled. Positions of samples shown.

erate in the northern part of the basin, is at least 700 m thick (Sæther and Nystuen, 1981).

The formation consists of about 0.1 to 5 m thick conglomerate beds interbedded with dark gray, massive medium- to coarse-grained sandstone layers. The formation thins and fines upward toward the overlying upper member of the enveloping Biri Formation (figs. 2 and 4). Most conglomerate beds are matrix-supported, but there are all gradations from sandstone beds with some few isolated pebbles or cobbles to clast-supported conglomerate facies. The matrix consists of sandstone with similar grain-size and texture as the adjacent sandstone beds. The clasts are dominated by extrabasinal metaquartzite and gneiss in well-rounded pebbles, cobbles, and boulders. Coarse-tail normal and inverse grading are common in the conglomerate beds which also display clast imbrication. Most contacts between adjacent conglomerate beds or between conglomerate and sandstone beds are gradational or amalgam-
ated (fig. 4). Seen in cross sections, the conglomerate beds form lensoid sheets.

On the eastern side of Lake Mjøsa, the Biskopåsen Formation overlies the Birí shales with a marked erosional unconformity. The basal portion of the conglomerate contains numerous clasts derived from the underlying member of the Birí Formation, including various limestone facies and calcareous sandstones and, less frequently, phosphorite pebbles. The intrabasinal clasts of calcareous sandstone are 4 to 5 m in diameter. On the western side of Lake Mjøsa (at the sampling site of M-77-60 in fig. 4) intrabasinal clasts are also present in the middle part of the formation. East of Lake Mjøsa clasts of tholeiitic basalt, reaching boulder size, are abundant in the boulder-rich basal part. But in the section west of Lake Mjøsa clasts of this rock type are less frequent up to the middle part of the formation. The tholeiitic basalt fragments may have been derived from local basalt flows erupted along fissures in the fault-bounded basin margin (Nystuen, 1982, 1987). All clasts of intrabasinal origin, including basalt fragments, are angular to poorly rounded. Phosphorite clasts from the basal part east of Lake Mjøsa yielded scattered microfossils (Manum, 1967) and specimens of the problematic fossil *Papilomembrana compta* (Spjeldnæs, 1963, 1967; Vidal, 1990).

The wedge-shaped bodies of the Biskopåsen Formation were previously interpreted as formed in fluvio-deltaic settings (Skjeseth, 1963; Bjørlykke 1966; Spjeldnæs, 1967; Bjørlykke, Englund, and Kirkhusmo, 1967). Skjeseth (1963) suggested that the conglomeratic unit was deposited during a phase of fault-induced basin subsidence, whereas Bjørlykke (1966), Spjeldnæs (1967), and Bjørlykke, Elvsbørg, and Høy (1976) ascribed its origin to a regional regression. The regression hypothesis was suggested because of the presence of clasts of limestone, phosphorite, and calcareous sandstone. The clasts derive from eroded beds of the Birí Formation and from the lower member of this unit beneath the Biskopåsen Formation (Bjørlykke, Englund, and Kirkhusmo, 1967: Bjørlykke, Elvsbørg, and Høy, 1976). This part of the Birí Formation formed during an early stage of carbonate deposition. R. Otter (in Nystuen 1982) interpreted the Biskopåsen Conglomerate wedges as subaqueous fans formed in front of deltas by gravity flow processes. The margin origin of the conglomerate is demonstrated by the presence of microfossils in dark shales from a section west of Lake Mjøsa (fig. 4). Furthermore, the subaqueous origin is indicated by the interfingering between the Biskopåsen Formation and the shales and sandstones of the Birí and Brøttum Formations, respectively. The internal organization of conglomerate and sandstone beds, matrix-supported texture, frequent amalgamated contacts, and total lack of sedimentary structures is typical of gravel-sized fluvial deposits. These features are suggestive of transport and deposition from high-density turbidites and subaqueous mass flows. Thus, in our opinion, there is no evidence supporting the existence of a subaerial fluvial deltaic counterpart of the subaqueous facies
association recorded in the Biskopåsen Formation in the Lake Mjøsa area.

The boulder-bearing basal part in the eastern side of Lake Mjøsa is interpreted as a subaqueous channel-fill deposit. The channel may have been cut down during an event of regression, eroding calcareous deposits of the lower member of the Birí Formation which formed in shallow-marine platforms along the basin margin. The high amount of extra-basinal clasts and the architectural style of the Biskopåsen Conglomerate wedges do imply a strong tectonic control. Furthermore, the association of basaltic volcanism indicates that the “Biskopåsen depositional event” reflects a rifting climax in the Hedmark Basin (Nystuen, 1987). However, an increased run-off in drainage area, perhaps due to climatic changes, might have been a contributing factor to the influx of gravel-sized debris to the basin.

The material sampled for this study consists of dark gray, carbonaceous, laminated hemipelagic mudstone, preserved as thin beds in the upper part of the formation (fig. 4). Clasts of similar mudstone lithology, for example, sample M-77-60, are abundant in the conglomerate and sandstone beds. This indicates that hemipelagic deposition of organic-rich mud took place between surges of gravity-flows carrying coarse-clastic debris. One single sample of sandstone was also investigated (app. 1).

Birí Formation.—The Birí Formation has a thickness of 50 to 200 m and crops out in all parts of the Sparagmite Region. The formation is complex and contains a variety of interfingering lithofacies, including pure micritic limestone, oolitic limestone, dolomite limestone, dolomite, arenaceous limestone, intraformational limestone breccia, calcareous sandstone, heterolithic facies with interbedded laminated limestone, calcareous mudstone and fine-grained sandstones, calcareous laminated siltstone, and laminated black shale. The carbonate-dominated facies are interpreted to represent deposition in shallow-marine carbonate platforms in shelf settings, including intertidal and supratidal environments. However, laminated micrites of the Birí Formation were interpreted as deep shelf or basinal deposits (Tucker, 1983). Furthermore, arenaceous facies and black shales are interpreted as basin-slope and basin-floor deposits, respectively (Bjørlykke, Elvsborg, and Høy, 1976; Nystuen, 1982). Calcareous sandstones formed in high-energy coastal environments also occur, and some intraformational limestone breccias may also have originated along edges of carbonate platforms (Nystuen, 1982).

In the Mjøsa type area, shales, siltstones, and calcareous rocks also occur beneath the Biskopåsen Formation (fig. 4), and these beds are regarded as a lower member of the Birí Formation (Bjørlykke, Englund, and Kirhusmo. 1967, Bjørlykke, Elvsborg, and Høy, 1976). The various lithofacies represented in the carbonate clasts and the phosphorite fragments that occur in the lower part of the Biskopåsen Formation (see above) are considered to have been eroded from this lower member of
the Biri Formation (Spjeldnæs, 1967; Bjørlykke, Elvsborg, and Høy, 1976). The lower member is missing in a great many areas, and this may be due to erosion during the formation of the Biskopåsen Conglomerate wedges (Bjørlykke, Elvsborg, and Høy, 1976) or to non-deposition of the early Biri fine-clastic and calcareous facies. In these areas the Biskopåsen Formation rests directly on and interfingers with the Brøttum turbidite sandstones. In areas where the Biskopåsen Formation is absent, basinal shales and calcareous mudstones of the Biri Formation succeed the Brøttum sandstones or their laterally equivalent formations (Nystuen, 1982, 1987).

The deposition of the Biri Formation is interpreted to have taken place during a regional transgression, possibly as the result of a eustatic sealevel rise (Spjeldnæs, 1967; Bjørlykke, Elvsborg, and Høy, 1976). However, the nature of the sealevel change should be considered as relatively doubtful (Bjørlykke, 1982) due to the degree of uncertainty involved in the effect of eustatic sealevel changes in an intracratonic rift basin with high tectonic activity (Nystuen, 1987). The transgressive phase is clearly demonstrated in the northern and eastern part of the basin where the fluvial Rendalen Formation is overlain by limestone, black shale, and dolomite of the Biri Formation (Nystuen, 1982, 1987). The organic-rich black shales (with TOC up to 3 percent) of the Bjørånes Shale Member in the central and northern part of the Sparagmite Region appear to have been deposited in a sub-basin formed by a sudden tectonic subsidence of the basin floor. Black shales with interbedded sandstones and subaqueous gravity-flow conglomerates rest directly on fluvial arkoses (Sæther and Nystuen, 1981; Nystuen, 1987).

The complex interrelationships between the Biri Formation and adjacent formations of the Hedmark Group and the great facies variations in the formation suggest that this unit was deposited during conditions of sealevel variations and tectonically controlled sedimentation. The presence of gypsum and halite pseudomorphs together with magnesite and chert from the stratigraphically equivalent Hylleråsen Formation in the Kvitrøla Nappe Complex also suggests that arid climatic conditions prevailed during at least part of Biri deposition times (Nystuen, 1969: 1980). In an analysis dealing with the origin of organic-rich limestones and mudrocks of the Biri Formation, Tucker (1983) presented a model for the deposition of the Biri Formation that implies accumulation connected with transgressive events leading to increased organic productivity and subsequent development of oxygen-depleted bottom waters. Bjørlykke, Elvsborg, and Høy (1976) suggested a normal-marine salinity and open-marine conditions during the deposition of the Biri Formation.

The Biri Formation was sampled in the upper member in the Lake Mjøsa area, in a basin-slope setting at Rena, in a more central basin position at the River Åsta, and in a sub-basin position at n. Bjøråa (where the Bjørånes Shale Member rests on underlying fluvial sandstone; figs. 1, 4, 5 and app. 1). The samples taken close to and just above the Bisko-
Fig. 5. Upper part of Biri Formation and lower part of the Ring Formation at Becksminne, 4 km south of Rena (627050/6775550). Positions of samples shown.

Påsen Formation (fig. 4) consist of hemipelagic mudstones within thin-bedded, fine-grained sandstones, calcareous siltstones, and laminated limestones of turbidite origin. This lower part of the upper member of the Biri Formation probably represents an upward shallowing sequence formed at the slope of a prograding carbonate platform.

Ring Formation.—The Ring Formation occurs as several coarse-clastic wedges along the basin margin in the western part of the Hedmark Basin. It consists of shales, thin-bedded turbidite sandstones, coarse-grained, conglomeratic arkoses, and massive to graded quartz-pebble conglomerates of channel-fill origin (fig. 5). Clasts of dark gray
mudstones are common in the conglomerates. The contact to the underlying Birí Formation is in some places sharp and erosional (Løberg, 1970; Englund, 1972, 1973b). In the type locality at Mjösa and in the Rena area the boundary is characterized by a transition from black shales to grayish-green mudstones, siltstones, and very fine-grained sandstones beneath the lowermost conglomerate channel-fills of the Ring Formation. Bjørlykke, Elsvborg, and Høy (1976) interpreted rocks of the Ring Formation as fan delta deposits in which subaerial as well as subaqueous facies were present. Nystuen (1987) suggested that the Ring Formation, like the Biskopásen Formation, is primarily preserved as a subaqueous facies deposited by gravity flow processes on fronts of prograding fan deltas. Samples were collected from the fine-grained mudstones in the lowermost part of the Ring Formation (fig. 1, app. 1).

Moelv Tillite and Ekre Formation (shale).—The Moelv Tillite of the Varangerian glaciation (about 650 Ma) has a wide distribution in the Sparagnite Region. It is underlain by a regional erosional unconformity and rests on the Ring, Birí, Rendalen Formations (fig. 2) and the autochthonous crystalline basement (Nystuen, 1976; Bjørlykke, Elsvborg, and Høy, 1976; Bjørlykke and Nystuen, 1981; Bockeleie and Nystuen, 1985). It is composed of a number of facies, including subglacially deposited massive diamicrite and, usually overlying, glaciomarine laminated mudstone with dropstones (Bjørlykke, Elsvborg, and Høy, 1976; Nystuen, 1976). The Moelv Tillite sequence has been interpreted as the result of the advance and retreat of a continental ice sheet across most of the Hedmark Basin and other rift- and shelf-basins at the margin of the Baltoscandian craton (Nystuen, 1985). In the deepest part of the basin the ice-sheet was probably floating, and close to the buoyancy line the ice might have eroded on underlying unconsolidated sediments without imposing any high burden that could destroy incorporated microfossils.

The Ekre Formation (or Ekre Shale) succeeds conformably the postulated glaciomarine upper facies of the Moelv Tillite. The transitional boundary is marked by an upward decrease in ice-dropped clasts (Nystuen, 1976, 1982). The Ekre Formation is dominated by grayish-green or red, laminated siltstones. In the upper part it is interbedded with deltaic sandstone beds of the overlying Vangså Formation (Bjørlykke, Elsvborg, and Høy, 1976; Nystuen, 1982; Dreyer, 1988). A marine origin for the lower part is suggested by the conformity with the underlying Moelv Tillite (Nystuen, 1982, 1985). From geochemical studies on the Ekre Formation, Englund (1973b) concluded that the shale unit was deposited in an oxygen-rich and well-aerated water body, either marine or lacustrine. Although our samples from the Ekre Formation are totally barren, we visualize that the basin had indeed a connection with the sea during Ekre deposition times. The absence of microfossils is explainable in terms of insufficient sampling due to poor exposure of suitable facies. Furthermore, the sampled greenish-gray mudstones
seem virtually devoid of organic matter. This feature may be the result of the destruction of organic matter in oxygen-rich water and oxygenated sediments.

_Vangås Formation._—The Vangås Formation is the uppermost formation in the Hedmark Group and consists of two members, the Vardal Sandstone Member and the Ringsaker Quartzite Member (fig. 2). The lower member is characterized by great lateral facies and thickness variations. In the eastern and northeastern part of the Hedmark Basin, it includes coarse- to fine-grained turbidite sandstones, subaqueous debris-flow deposits, and black organic-rich shales, besides (overlying) deltaic and fluvial, braided-stream sandstones and conglomerates (Nystuen, 1982, 1987). In the southern and western part of the basin the Vardal Member consists of feldspathic arenites of deltaic and braided-stream origin (Bjørlłykke, Elvsborg, and Høy, 1976; Dreyer, 1988).

The Vardal Member passes into the overlying shallow-marine Ringsaker Quartzite Member through a zone of marine reworking (Bjørlłykke, Elvsborg, and Høy, 1976; Nystuen, 1982; Dreyer, 1988). The Ringsaker Member facies is interpreted as representing a beach facies, tidal channel deposits, shallow-marine storm-deposits, and offshore bars (Nystuen, 1982; Dreyer, 1988). Bioturbation structures consisting of _Scolithos, Monococraterion_, and _Diplocraterion_ occur in the uppermost part of the member in the Mjøsa area (Skjeseth, 1963); this member is immediately overlain by fossiliferous shales of the Lower Cambrian Holmia “stage” (Skjeseth, 1963; Vidal, 1981; Ahlberg, Bergström, and Johansson, 1986; Moczydlowska and Vidal, 1986). The Ringsaker Quartzite Member is a major transgressive unit and onlaps the autochthonous Precambrian basement (Bockelie and Nystuen, 1985).

Only the Vardal Sandstone Member is dealt with in this paper: samples were collected from black shales and turbidite sandstones in the Osen-Jordet area (fig. 1; app. 1).

**MATERIAL AND METHODS**

Samples were collected from structurally simple and little tectonized, comprehensive sections and isolated outcrops through units of the Hedmark Group. This is composed of a total of twenty localities (see fig. 1 and app. 1).

Rocks of the Hedmark Group are well indurated. Whenever possible, pieces of rock with an approximate weight of 50 g were cut from the core of each sample. The samples were then pretreated and processed according to the method described by Vidal (1988). Concentrated organic residues from each sample were prepared in permanent mounts using glycerol jelly (M-77 samples) or petroepoxy (M-82 samples; Vidal, 1988). Permanent mounts were examined under an interference contrast transmission light microscope, and micrography was generally carried out using a X 100 oil immersion lens.
RESULTS

A total of eighty one (81) samples from the Hedmark Group were examined for acid-resistant organic walled microfossils. Additionally, petrographic thin sections of forty five samples cut parallel to the bedding were also examined. Thirty two samples yielded microfossils.

The recovered microfossils display a variable state of preservation. Probably because they are small (10–30 μm in diam), microfossils from the Brøttum Formation are generally preserved in three dimensions. This feature is particularly common among microfossils from the Biskopåsen and Ring Formations, where a supporting detrital or carbonate rock matrix has resulted in extremely good preservation.

The color of particulate kerogen (and microfossils) varies from amber, dark-brown to gray, thus reflecting moderate to considerable thermal alteration. The lowest observed thermal alteration index (TAI) is among microfossils from the Biskopåsen, Bir, and Ring Formations which yield amber to dark-brown microfossils. More generally, the small microfossils from turbidite shales are light brown (hue 10YR 5/6), while a dark-brown color (hue 10YR 3/2) dominates the larger and thicker-walled specimens (Pearson, 1984). This corresponds to TAI 3+ to 4 (AMC 5+ to 6; Rovnina, 1981) which may indicate burial temperatures of the order of 175 to <200°C (Rovnina, 1981) and late catagenesis-early metagenesis stages (Hayes, Kaplan, and Wedeking, 1983). However, samples of these units also yielded kerogen suggesting a higher stage of maturation in agreement with burial temperatures around 200°C.

In general, samples from the Brøttum and Vangssås Formations are rich in black to grayish, strongly biodegraded amorphous kerogen. Whenever present, microfossils are generally small, ranging about 6 to <30 μm in diam. The color of particulate organic matter and microfossils is variable. Thin envelopes and sheaths are generally gray and translucent and in agreement with a supermature thermal stage indicating temperatures around 200°C or higher, corresponding to a very low grade of metamorphism (Morad, 1988). But condensed organic accumulations inside discrete cells and thicker cell envelopes are generally opaque black.

Rocks of the Brøttum Formation were sampled at five localities (see app. 1). Seventeen processed samples of hemipelagic/low-density turbidite shales (see above) yielded only abundant carbonaceous material; microfossils were recovered from twelve samples (fig. 6). The samples yielded scattered small (<30 μm across) smooth spheromorphs (fig. 7C), abundant specimens of small spheromorphs displaying inner dense, dark bodies (fig. 7F–F’, fig. 8D), small isolated and clustered specimens of spherical cellular aggregates attributed to Sphaerocongregus sp. (fig. 7A, B, D, E), some of which are contained in thin, hyaline envelopes (fig. 7G, I, J–L). The assemblage also includes scattered specimens of flattened, 60 to 70 μm in diam, opaque unidentifiable spheromorphs.
Seven samples were collected from the Biskopshaven Formation. Microfossil occurrences are shown in figure 9. Samples M-77-61, 62A are of hemipelagic/low-density turbidite shale interbeds in conglomeratic portions of the unit (see above), and M-82-53, 54 are samples of shale at the base of conglomerate beds.

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<td>Clustered smooth spheroids</td>
<td>Smooth, hyalineous spherobidial sheaths with inner bodies</td>
<td>Sphaerocongregus sp.</td>
<td>Sphaerocongregus sp. with loose cells</td>
<td>Sphaerocongregus with enclosing transparent envelope</td>
<td>Smooth, hyalineous spheroids</td>
<td>Double-walled, smooth spheromorphs</td>
<td>Smooth, solitary spheromorphs</td>
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Fig. 6. Distribution of microfossils in fossiliferous samples from the Frejdam Formation.
(fig. 4). Sample M-77-39A is of matrix rich portions in the conglomerate, whereas sample M-77-59 is from a sandy interbed in the conglomerate. In addition, M-77-60 represents a shale clast within the conglomerate (fig. 4). Five samples yielded microfossils (fig. 9). The microfossils generally do not exceed 30 μm in diameter. Microfossils from the matrix sample (M-77-39A) are small (10-22 μm) spherical, smooth spheroids comparable to those in the Brøttum Formation. Abundant microfossils attributed to Sphaerocongregus sp. (fig. 7A, B, D) occur in M-77-60. More irregular-shaped colonial microfossils resembling pleurocapsal or chroococcalean cyanobacteria (fig. 11C-C') were recovered in small numbers.

Scattered three-dimensional preserved microfossils attributed to Vandalosphaeridium varangeri Vidal, 1981 (fig. 12C-C'), Cymatiophosphoridae sp. (fig. 11A-A', B-B', E, F-F'), and Palaeogomphosphaeria caurensis Palacios Medrano, 1989 (fig. 8A-A'', B, D-D') occur in M-77-62A.

A total of twenty samples of the Biri Formation and ten samples of the Bjørnæs Shale Member were processed (M-77-85 to 95). Most samples are of black, laminnated, or massive mudstones (M-77-58, 85, 86, 87, 88, 89, 90, 91, 92, 93, 36), and black, organic rich carbonate within the formation (M-77-36, 36B) and yielded detrital organic matter, and thirteen samples yielded microfossils (fig. 10). Most microfossils are small, spheroidal, and smooth, often displaying intracellular condensed organic matter and are here referred undetermined to cocoid cyanobacteria comparable to microfossils from the Brøttum Formation (fig. 7C, F-L).

As in the Brøttum Formation, colonial spheroids are common and can be attributed to the possibly chroococcalean Sphaerocongregus sp. (fig. 9 A-B, D-E, G-I). Smooth flattened spheromorphs are attributed to Leiosphaeridia spp. (fig. 12F). Additionally, three samples yielded abundant specimens of polygonomorphic acritarchs attributed to Octoedryxium truncatum (Rudavskaya, 1971) Vidal, 1976a (fig. 12A-B'). Acritarchs with a clearly reticular wall surface are rare and represented only by two specimens recovered in one single sample (M-77-85: fig. 8C). A possibly spinous acritarch displaying short processes is present in sample M-77-72 (fig. 12G).

Four samples from the Ring Formation were examined for the present study. They are derived from shales within the conglomerate facies of the formation (M-77-37) and from shale clasts within the conglomerate facies of the unit (M-77-38, M-82-20, 21). Three samples

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Fig. 7. Sphaerocongregus variabilis Moorman, 1974 (A-B, D-E, G-L), (A-B, D) specimen LO 4108, slide M-77-83:01; coordinates Y/37-2, Z/38-3, and Y/37-2, respectively. (E) specimen LO 6150, slide M-77-82:1A:Z/39-40. (G) specimen LO 6103, slide M-77-80:01:C/44. (H-I, L) specimen LO 6101, slide M-77-74:01; coordinates X/46, Z/36 (lower) and Z/36, respectively. (J-K) specimen LO 6104, slide M-77-81:1A; U/45 and X/46 (X). Small spheroidal microfossils displaying dense inner, dark bodies (C, F-F'), specimen LO 6105, slide M-77-82:1A; coordinates Z/26-2 and Z/43 (lower). All specimens from the Brøttum Formation. Length of bar at A 10 μm for all micrographs.
yielded microfossils that consist of abundant small, smooth spheroids with internal coagulated organic matter (compare above; fig. 8E-F) comparable to those recovered from underlying units, specimens attributed to Sphaerocongregus sp., and scattered thick-walled, smooth spheromorphs attributed to Leiosphaeridia sp.

Samples of matrix-rich, dark-gray diamicite from three sections of the Moelv Tillite (samples M-77-35, 101, 102; see app. 1) and of gray-green mudstone from the Ekre Shale (samples M-77-103, 104) were processed. Sample M-77-101 yielded a few specimens of brownish-colored acritarchs attributed to Leiosphaeridia sp. (fig. 12F) and T. timofeevi and one single specimen of V. varangeri (fig. 12C-C'). The remaining samples were totally barren.

Five samples of black, pyrite-rich mudstones of the Vardal Sandstone Member of the Vangås Formation (M-77-105, 108, 109, 110; see app. 1) were processed. All samples yielded abundant, extensively biodegraded dark-gray to black organic matter. A sample (M-77-106) yielded specimens of microfossils considered as colonial clusters of cyanobacteria attributed to Sphaerocongregus sp. and smooth, thick spheromorphs attributed to Leiosphaeridia sp. Sample M-77-105 from the uppermost part of the Vardal Sandstone Member yielded the only identifiable microfossils consisting of two specimens of Fimbrioglomerella minuta (Jankauskas 1979; Moczydlowska and Vidal, 1988).

MICROFOSSILS

The most common microfossils in the Hedmark Group consist of smooth, thin-walled, translucent, solitary, or clustered spheroidal microfossils. Their dimensions range from 6 to 30 μm in diameter. They often display internal, dark, often granular, and irregular-shaped bodies (fig. 7G-L). They are possibly envelopes of colonial cell aggregates (here attributed to Sphaerocongregus, see below) or solitary cells of cyanobacteria. Intracellular granular dark bodies (fig. 7C, E, F-F'; fig. 8A-A', D; fig. 12E) probably represent blebs of degraded protoplasm (Hofmann, 1976; Knoll and Barghoorn, 1975; Knoll and Golubic, 1979). Numerous samples yield spheroidal colonial cell aggregates, here referred to the acritarch species Sphaerocongregus sp.

Filamentous microfossils consisting of translucent, septated, and coiled sheaths 10 to 15 μm in diam (fig. 12D) occur in small numbers in

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Fig. 8. Palaeogomphosphaeria caurenis Palacios Medrano, 1989 (A-A", B). (A-A") three-dimensionally preserved specimen at three different focal levels. Micrographs A-A" and B from specimen LO 6099, slide M-77-62A:02; coordinates M/27 and N/42. Unnamed spheromorph with reticular surface ornamentation (C); specimen LO 6106, slide M-77-85:01; coordinates X/42. Unnamed smooth spheromorph with internal coagulated (?) organic material (D); specimen LO 6102, slide M-77-77:01; coordinates N/43-4. Colonial cell clusters with dark coagulated (?) intracellular organic material (E-F); specimen LO 6097, slide M-77-36B:01; coordinates Y/27-4 and Z/27-4. Specimen at A-B from the Biskopasen Formation. Specimen at D from Brøttum Formation. Specimens at C, E, F from the Birr Formation. Length of bar at E 20 μm for A-A", B, D, E-F; 13 μm for C.
Fig. 9. Distribution of microfossils in samples of the Biskopåsen Formation.

most samples. Their simple morphology adds little information as to their taxonomic attribution. Smooth, as well as annulated and segmented, tubular sheaths are common components in numerous late Proterozoic Riphean microfossil accompanied by abundant acritarchs (Jankauskas, 1989) from shallow marine settings. The Biri filaments could admittedly represent rare occurrences of planktonic oscillatoriacean cyanobacteria. However, the Biri samples are from rocks that formed at considerable depth, and no acritarchs accompany the filaments. We consider that the available morphological, depositional, and paleoecological information equally well could accommodate interpreting the filamentous microfossils as abandoned sheaths from filamentous cyanobacteria or as heterotrophic, sulfide oxidizing beggiatoacean bac-
Fig. 10. Distribution of microfossils in samples of the Biri Formation.

teria. Viewed in the perspective of the present model of deposition of the Hedmark Group (see below), we favor the latter interpretation.

**TAXONOMY**

LO numbers refer to the collections of the Geology Department University of Lund. The position of figured specimens in the permanent mounts is indicated after the colon following the slide number.
Fig. 11. Cymatosphaeroides sp. (A-A', B-B', E-F'). *Palaeogomphosphaeria caurensis* Palacios Medrano, 1989 (D-D'). Pleurocapsaean or chroococcalean cyanobacteria (C-C'). All from specimen LO 6098, slide M-77-62A:01; coordinates C/29 (A-A'), Q/34 (B-B'), P/40 (C-C'), Z/40-3 (D-D'), Z/47 (E) and P/32 (F-F'). All specimens from the Biskopåsen Formation. Length of bar at A 10 μm.
Genus Octoedryxium Rudavskaya, 1973

Octoedryxium truncatum (Rudavskaya, 1973) Vidal, 1976a
Figure 12A–B′

Synonymy: See Vidal (1976a)
Description: See Vidal (1976a)

Dimensions: The dimensions of O. truncatum vary considerably (Vidal, 1976a). Specimens in the present material from units of the Hedmark Group are consistently small (N = 20, $\bar{x} = 14.1$ $\mu$m, $\sigma = 2.23$ $\mu$m). One single specimen recovered from the Biskopåsen Formation (sample M-77-62A) is 30 $\mu$m across, a size found among specimens from the USSR and southern Sweden (compare Vidal, 1976a). On the other hand, dimensions recorded among scattered specimens from the Vendian Tillite Group in East Greenland (Vidal, 1979) compare well with those of the present material.

Remarks: Specimens of O. truncatum are particularly common in samples of shales from the Biri Formation. One single well preserved specimen was found in the Biskopåsen Formation. Specimens from the Biri Formation are dark-brown in color due to thermal maturity and are preserved in a state of slight compressional distortion. The single specimen recovered from the Biskopåsen Formation, however, is almost completely flattened. Specimens from the Biri Formation display sharp pointed apical cusps, although Vidal (1976a) indicated great variability concerning this feature. This is illustrated by the fact that the single specimen deriving from the Biskopåsen Formation displays more rounded vesicle cusps.

Occurrence:—Biskopåsen Formation and Biri Formation (near to junction with Ring Formation). Previous occurrences of O. truncatum were listed by Vidal (1976a, 1979).

Stratigraphic range: Vendian and Lower Cambrian (?). Previous records strongly suggest that the occurrence is restricted to early Vendian deposits (in the sense used by Vidal and Siedlecka, 1983). Occurrences in the Mostsk Beds (Motskaya Svita) in Irkutsk were interpreted as suggesting occurrence in the Lower Cambrian. However, the age of the Mostsk Beds remains problematic (Vidal, 1976a).

In the Northwest Atlantic region, the youngest recorded specimens of O. truncatum are scattered, possibly reworked specimens in the Vendian Tillite Group in East Greenland (Vidal, 1979).

Genus Sphaerocongres Moorman, 1974

Sphaerocongres variabilis Moorman, 1974
Figure 9A–B, D–E, G–L


Description: As for Sphaerocongres Moorman, 1974.
Dimensions: Colonial aggregates; N = 560, $\bar{x} = 23.17$ $\mu$m, $\sigma = 7.03$ $\mu$m. Discrete cells; N = 100, $\bar{x} = 1.68$ $\mu$m, $\sigma = 0.59$ $\mu$m.

Remarks: Microfossils here attributed to Sphaerocongresus sp. and S. variabilis were formerly attributed to Bavlinaella faveolata (Shepeleva) Vidal 1976a. Vidal (1976a) relied on Timofeev's identification of cellular aggregates with Bavlinaella faveolata Shepeleva (Timofeev, 1969). In so doing, he transferred obviously biogenic cell aggregates attributed to Sphaerocongresus variabilis Moorman 1974 to B. faveolata Shepeleva 1962. But subsequent examination by one of us (G. V.) of the type specimen of B. faveolata from the Bavl Formation in the Volga-Ural region in the USSR revealed, as previously indicated by Volkova (1974), that the type specimen illustrated by Shepeleva (1962) is in fact the organic residue after maceration of frambooidal pyrite. Pyrite frambooids can and do resemble certain types of microorganisms (Vidal, 1974; Knoll, Blick, and Awramik, 1981). As with numerous previous occurrences (Knoll, Blick, and Awramik, 1981) the microfossils in question are not necessarily associated to the presence of pyrite, and, connecting to previous sources of evidence (Vidal, 1977; Knoll, Blick, and Awramik, 1981), it deserves mentioning that the microfossils are transparent to transmitted light and that different stages of preservation reveal specimens in various stages of disaggregation which allow viewing the biogenic nature of cell clusters (fig. 9E). Although formerly misidentified with probable pseudomicrofossils, we are certain that the structures under discussion represent a biological organization. The arguments supporting this view were presented by various authors, and a complete discussion in favor of the biogenicity of comparable fossils can be found in Cloud, Moorman, and Pierce (1975). Vidal (1977), Knoll, Blick, and Awramik (1981), Mansuy (1983). A conspecific taxon (Sphaerocongresus Moorman 1974: to which the present microfossils are referred) was considered by Moorman (1974) possibly to be related to Pleurocapsalae cyanobacteria, a view subsequently followed by Knoll, Blick, and Awramik (1981).

Moorman (1974) interpreted various unicells, dyads, and cell aggregates as part of the life cycle of a single taxon. On grounds of spatial

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Fig. 12. Octoedryxium truncatum (Rudavskaya, 1973) Vidal, 1976 (A-B). (A) specimen LO 6099, slide M-77-62A:02; coordinate R/31. (B-B') specimen LO 6107, slide M-77-98:1G; coordinate Y/28-2. Vandalosphaeridium varangeri Vidal, 1981. (C-C'). Specimen LO 6109, slide M-77-101:01; coordinate J/31. Abandoned translucent sheath, probably deriving from filamentous cyanobacteria or heterotrophic, sulfide oxidizing bepgiatocean bacteria (D). Specimen LO 6106, slide M-77-85:01; coordinate Z/42. Probably cyanobacterial envelopes of dividing cells displaying condensed (?) intracellular organic matter (E). Specimen LO 6108; slide M-77-83:01; coordinate Z/43. Leiosphaeria sp. (F). Specimen LO 6107; slide M-77-98:1G; coordinate L/33. Unnamed spinous (?) acritarch (G). Specimen LO 6100; slide M-77-72:01; coordinate W/26. Specimen at A from the Biskopasen Formation. Specimens at B-B', D, F and G from the Bir Formation. Specimen at E from the Brøttum Formation. Specimen at C-C' from the Moelv Tillite. Length of bar at E 10 $\mu$m for A, B-B', E, F-G; 15 $\mu$m for C-C', and 25 $\mu$m for D.
association, patterns of cell division, and cell growth, Knoll and Swett (1985) argued for the formerly proposed (Moorman, 1974) pleurocapsalean affinity. While, several of the above mentioned smooth, thin-walled spheromorphs can be identified with some of Moorman's stages, we refrain from attempting more precise taxonomic attributions. This is because we consider that the very simple morphology of these allochthonous microfossils and their advanced state of degradation make the approach highly speculative. As it is the case with numerous previous occurrences of comparable microfossils (Knoll, Blick, and Awramik, 1981; Mansuy, 1983; Palacios Medrano, 1989), the present material yields a range of morphotypes clearly transitional between the multisphere aggregates of *Sphaerocongregus* and clusters of spheroidal unicells, and this circumstance is indicated in figures 6 to 8. In our opinion, the complete range of variation observed among multisphere aggregates (figs. 9A–L: 12E), encompassing specimens with cells ranging in diameter from <1μm to 4μm, can be accommodated within the limits of morphologic variation of one single taxon. But they may in fact represent several taxa of only remotely related bacteria. The same applies to the more regular cell aggregates (fig. 9A–B), displaying evident geometrically arranged cell limits, which could as well be attributed to Chroococcalean cyanobacteria (the extant marine genus *Cyanocystis*). We thus see that the various morphotypes could possibly be interpreted to belong to different genera (alternatively one single genus) of extant cyanobacteria, although they may as well be interpreted as representing taxa within different orders (the Pleurocapsales; Moorman, 1974).

The microfossils under discussion occur in detrital sedimentary rocks formed in various depositional settings (Knoll, Blick, and Awramik, 1981), this being in agreement with a planktonic mode of life (Mansuy and Vidal, 1983; Palacios Medrano, 1989). This feature, coupled to close association to high productivity and eutrophic conditions, probably in connection with interglacial conditions (Knoll, Blick, and Awramik, 1981) and turbidite deposition (Mansuy and Vidal, 1983), justified their attribution to planktonic Chroococcalean cyanobacteria comparable to *Gomphosphaeria*, *Coelosphaerium*, and *Microcystis* (Mansuy and Vidal, 1983). Taxonomic attribution remains controversial pending more conclusive evidence (Knoll and Swett, 1985). An alternative previously unexplored interpretation of these microfossils as photosynthetic anoxygenic bacteria could be accommodated within the frame of inferred paleoecological conditions and depositional settings (see discussion below under environmental aspects).

**Occurrence:** See Mansuy (1983), Knoll and Swett (1985), and Palacios Medrano (1989).

**Stratigraphic range:** For a complete discussion see Knoll, Blick, and Awramik (1981) and Knoll and Swett (1958).
Palaeogomphosphaeria caurensis Palacios Medrano, 1989
Figure 10A–A'', B; 11D–D'


Description: Tightly packed, spherical clusters of radially arranged ovoidal cells, occasionally displaying condensed cells contents observed as irregular-shaped dark-brown spots.

Remarks: Microfossils attributed to Palaeogomphosphaeria caurensis Palacios Medrano, 1989 are morphologically comparable to Recent species of Gomphosphaeria (the marine plankton G. aponina) and occur in small numbers in samples from the Biskopåsen Formation and the Biri Formation.

Dimensions: Vesicle diam 20 to 25 μm, cell width 2.5 to 7 μm, cell length 4 to 6 μm, N = 4.

Occurrence: Taxa regarded by Palacios Medrano (1989) as conspecific with P. caurensis were reported from northwestern China (Yangjaping and Niutiltang Formations) and from the Cijara Formation in central Spain (Palacios Medrano, 1989).


Genus Vandalosphaeridium Vidal 1981
Vandalosphaeridium varangeri Vidal, 1981
Figure 12C–C'


Material: Nine specimens in generally fair state of preservation.

Description: Organic-walled microfossils consisting of a thick, massive inner body provided with scattered thin, simple, filiform processes which support an outer, transparent membrane.

Dimensions: Inner body; N = 9, $\bar{x} = 21.18$ μm, $\sigma = 3.22$ μm. Membrane; N = 8, $\bar{x} = 13.87$ μm, $\sigma = 2.97$ μm.

Remarks: The dimensions of specimens of V. varangeri recovered in the present material are comparable to those recorded in previous occurrences. The species was originally reported from Upper Riphean-Lower Vendian rocks in Varanger Peninsula, northern Norway (Vidal, 1981). Morphologically comparable taxa (V. reticulatum, V. walcottii and Cymatosphaeroides kulligii) were reported from Upper Riphean to Lower...
Vendian rocks in Sweden (Vidal, 1976a), Arizona (Vidal and Ford, 1985), and Svalbard (Knoll, 1984).


Stratigraphic range: Upper Proterozoic, Upper "Riphean"-Lower Vendian (sensu Vidal and Siedlecka, 1983). The single specimen recovered from a sample of the Moelv Tillite is probably reworked from older strata.

Leiosphaeridia spp.

Acritarchs here attributed to Leiosphaeridia spp. consist of medium size (about 45-50 µm across) compressed, probably spheroidal, smooth, and robust-walled microfossils. These microfossils are rare, being restricted to samples from the Biskopåsen Formation, the Biri Formation, and the Moelv Tillite (see below).

Distinctive operculated excystment openings were reported among certain unornamented leiosphaerids (Vidal and Ford, 1985). On other grounds, the leiosphaerids from the present material are indistinguishable from those reported from previous occurrences, either Proterozoic or early Paleozoic.

Environmental Aspects

In Proterozoic times, photosynthetically generated O₂ became abundant and aerobic conditions prevailed, thus contributing to establishing aerobic environments. In modern environments the accumulation of encysted stages of marine algal protists preferentially occurs in relatively shallow waters. This feature has a parallel in the Proterozoic record of encysted stages of microphytoplankton (Knoll, 1981; Vidal, 1976a; Vidal and Knoll, 1983).

The general existence of distinctive associations of acritarchs in early Paleozoic times has been explained in various ways. Thus, Colbath (1980) suggested successive fluctuations of water masses inhabited by assemblages of planktonic protists differing in taxonomic composition, whereas Dorning (1981) proposed the alternation of near-shore shelf and deep-water environments. and Al-Ameri (1983) suggested successive lagoonal, intermediate, and open marine environments. Previous studies of Proterozoic generally shallow shelf sequences confirm the recurrent nature of previously recognized patterns in the occurrence of assemblages of microfossils of planktonic algal protists and possible cyanobacteria (Knoll, 1985; Vidal, 1976a; Vidal and Knoll, 1983).

The vertical and lateral distribution of microfossils from units of the Hedmark Group parallels those reported in previous studies of various basinal and turbidite Late Proterozoic sequences in Spain (Palacios Medrano, 1989) and the late Proterozoic Brioverian of western France (Mansuy, 1983).
Organic-rich shale of hemipelagic origin in the Brøttum Formation yielded a fossil assemblage dominated by single celled and colonial spheroidal microfossils of presumable chroococcalean cyanobacterial or photosynthetic bacterial affinity (see below). These are considered to have been carried in suspension by oxygen-deficient, nutrient-rich waters and probably also by low-density sediment plumes and fine tails of turbidity currents that accumulated at top of channel fills, in levee environments, and inter-lobe areas (fig. 13). The total absence of presumably eukaryotic algae in the Brøttum Formation is a puzzling feature in accordance with observations from similar sedimentary facies in other Proterozoic sequences (Mansuy, 1983; Mansuy and Vidal, 1983; Palacios Medrano, 1989).

Additional to abundant microfossils of planktonic single celled and colonial spheroidal microfossils, the investigated laminated black shales of the Biri and Biskopåsen Formations contain few acritarchs (figs. 7, 8) identified as Cymatosphaeroides ? sp., Vandalosphaeridium varangeri, and Octoedryxium truncatum, previously recorded from shallow shelf and intertidal shales and carbonates (Vidal, 1976a, 1981; Vidal and Siedlecka, 1983).

Occurrences of acritarchs in the Ring Formation, Moelv Tillite, and Vardal Sandstone (Vangås Formation) are extremely poor and generally restricted to sporadic finds of smooth-walled leiosphaerids. Paralleling occurrences in coeval tillites in Varanger Peninsula (northern Norway; Vidal, 1981), acritarchs from the Moelv Tillite are considered to be reworked from older strata, although matrix-rich counterparts of the
unit in East Greenland may have accumulated during deposition in a glacial-marine environment (Vidal, 1976b, 1979). The occurrence of a well-preserved specimen of *Vandaloosphaeridium* in the diamictite facies of the Moœl Tillite can be due to erosion from underlying marine fan delta facies in the Ring Formation close to the buoyancy line of the glacial ice sheet that flowed into the basin in Moœl time.

Two main questions from the previous discussion demand closer scrutiny: (1) to what extent are reasonably well-identified turbidite sequences devoid of acritarchous microfossils?; (2) what mechanism or combination thereof could exclude presumed eukaryotic plankters in favor of presumably prokaryotic microorganisms? In one single instance, acritarchs attributed to eukaryotic algal protists were reported in supposed association with turbidite deposition. Thus, Zang and Walter (1989) reported complex acritarchs from the late Proterozoic Pertatalaka Formation and interpreted the apparent rarity of late Proterozoic large acanthomorphic acritarchs (but see Knoll and Butterfield, 1989) as depending on insufficient sampling of deep-marine settings. Nonetheless, turbidite deposition in the Pertatalaka shales has not been documented, thus leaving the occurrence of Proterozoic acanthomorphs open to alternative explanations (Knoll and Butterfield 1989; Vidal, 1990). Microfossils from late Proterozoic (Riphean and Vendian) turbidites have been extensively sampled, but results are very different from those reported by Zang and Walter (1989).

Deeper marine settings, for example, continental slope and submarine fans are well documented in the Proterozoic rock record (Bjørlykke, Elvsborg, and Høy, 1976; Pickering, 1979, 1981; Siedlecka and Edwards, 1980; Nystuen, 1981, 1987). Detrital rocks that formed in shelf areas normally yield rich and diverse assemblages of acritarchs (Dorning, 1981; Al-Ameri, 1983; Vidal and Knoll, 1983), whereas rocks that formed from suspension currents and gravity-flows in deeper basin troughs and submarine fans yielded predominantly simple spheroidal (presumably cyanobacterial) microfossils (Palacios Medrano, 1989). It is unlikely that these latter microfossils are part of an *in situ* living biota, and most likely represent either reworked bacterial benthos or, alternatively, rich accumulations of prokaryotic phytoplankton. Where sampled, Proterozoic hemipelagic mudstones are usually rich in sapropel-like organic matter (Palacios Medrano, 1989) and yield microfossils of possible planktonic chroococcacean cyanobacteria (Mansuy and Vidal, 1983; Palacios Medrano, 1989). Notwithstanding, virtual absence of microfossils of inferred algal protists in offshore pelagic deposits does not imply a total absence of eukaryotic plankton in the water column. Putative algal protists may have accumulated in negligible numbers or, alternatively, may not have produced preservable cysts. This may be the case with turbidite deposits of the Late Proterozoic Kongsfjord Formation in Varanger Peninsula investigated by Vidal and Siedlecka (1983). Great uncertainty remains about the preservability of geologically more recent motile stages of the algal protists. Some inferences can be ven-
tured from available information on modern and fossil dinoflagellates (Evitt, 1985). However, this adds additional uncertainty as to whether the organic-walled envelopes of motile stages that might have been produced by extinct Proterozoic and Paleozoic algal protists were preservable in the rock record. However, our study of numerous thin sections of pelagic mudstones from the Hedmark Group did not reveal any traces of hypothetical vegetative envelopes, and this observation parallels available data from recent studies of dinoflagellates (Evitt, 1985). Present data indicate that Proterozoic algal protists were planktonic and that they occurred preferentially in nutrient-rich, well-oxygenated shallow shelf waters bordering epicratonic and intracratonic basins, a feature perhaps reflecting an early stage in evolutionary significant adaptations aimed at the efficient recovery of water-dissolved nutrients (Vidal, 1984).

The second question demands a more elaborate answer intimately interwoven to the previous discussion and to the depositional history of the Hedmark Basin.

Black, organic-rich shales in the Birí Formation were interpreted as being accumulated during transgressive events that led to enhanced carbon burial through increased organic productivity and subsequent development of oxygen-depleted bottom waters (Tucker, 1983). In this context, the above mentioned presence of filamentous microfossils (see chapter on Microfoossils) here regarded as sheath building bennigatoacean organotrophic sulfide oxidizers (fig. 12D) is in agreement with the inferred oxygen-depleted conditions of sedimentation. Early sediments of the Birí Formation (lower member) may have formed in a shelf slope where cold oxygen-depleted and nutrient-rich waters carried by ascensional currents may have developed plankton blooms. In our model of a deep basin (at least some hundred meters) in which organic debris has been accumulated and concentrated in hemipelagic mud over long periods, the most obvious source of phosphate is from ascending, cold, oxygen-depleted and nutrient-rich waters from shelf-slope and basin-floor areas (figs. 13–16). It is suggested that such currents developed plankton blooms in other Late Proterozoic sequences of similar environmental framing (Palacios Medrano, 1989). Upwellings in the Hedmark Basin might have been brought about by seasonal changes in wind direction and air pressure and by more local factors causing inversions of a stratified largely anoxic water body. The tectonically controlled sedimentation of high-density turbidity currents and other types of gravity-flows may have impelled phosphate-rich bottom waters to circulate and flow into the shelf areas.

In the depicted scenario, productivity by plant protists was probably strongly restricted to a thin layer of superficial well-oxygenated water. In such a setting, surges in sediment discharge by riverine waters, either seasonal or triggered tectonically, transported water soluble mineral nutrients (notably phosphates). This could induce productivity peaks followed by increased cyanobacterial abundances (Mansuy and Vidal,
Fig. 14. Cartoon showing model of the depositional environment for the lower member of the Birí Formation, southwestern part of the Hedmark Basin.

Fig. 15. Cartoon showing model of the depositional environment for the Biskopåsen Formation, southwestern part of the Hedmark Basin.
and oxygen depressing mineralization of organic matter. This latter hypothetical model is founded on the basic assumption that primary productivity in the Hedmark Basin was largely carried out by oxygenic cyanobacterial phytoplankton. However, recent studies of the Black Sea demonstrate a previously suspected important contribution of anoxygenic photosynthesis to carbon cycling. This might in fact provide the foundation for an additional alternative interpretation to productivity in the Proterozoic (Repeta and others, 1989). Indeed, the apparently permissible assumption that a number of microfossils identified in the Hedmark Group and elsewhere may be chroococcalean or pleurocapsalean cyanobacteria could be basically wrong. These attributions were established largely on morphological, cell division (Moorman, 1974; Knoll and Swett, 1985), paleoecological and paleoenvironmental (Knoll, Blick, and Awramik, 1981; Mansuy and Vidal, 1983; Palacios Medrano, 1989) parallels to extant cyanobacteria. Indeed, the microfossils from the Hedmark Group and comparable turbidite sequences in Spain (Palacios Medrano, 1989), France (Mansuy, ms; Mansuy and Vidal, 1983), and northern Norway (Vidal and Siedlecka, 1983) could alternatively be interpreted as anoxygenic photosynthetic bacteria occupying the H$_2$S-rich waters of stratified suboxic basins with a very shallow chemocline (Repeta and others, 1989).

Additional to providing a satisfactory framework for comprehensive carbon fixation and enhanced burial of organic carbon (Tucker, 1983), we consider the above hypothesis attractive in that it explains
satisfactorily the apparent absence (or exclusion) of obviously aerobic eukaryotic primary producers from organic-rich, pyrite-rich shale facies in various Proterozoic basins.

The conditions shaping deposition and organic productivity in the Hedmark Basin were propitious to enhanced carbon burial and accumulation of sedimentary phosphates and appear consistent with a formerly proposed shallow shelf origin for minor occurrences of phosphates as penecontemporaneously deformed (and hence intrabasinal) reworked pebbles in the Biskopåsen Formation (Bjørlykke, Elvsborg, and Høy, 1976; Spjældnes, 1967). Moreover, microfossil evidence suggests that the various identified facies associations of the Hedmark Group preceding the Varangerian glacial event could be coherently accommodated in a simple model of turbidite sedimentation, formation of coarse-clastic subaqueous fans in a deep-marine basin with upwelling onto shelf platforms, enhanced organic production, and burial of organic carbon in hemipelagic muds.

**Biostratigraphy**

Age data on rocks of the Hedmark Group are restricted to one single, Rb/Sr whole rock age (Welin, unpublished; Rankama, 1971) of 612 ± 18 Ma on the Ekre Shale. An indirect estimation of the age of the Hedmark Group is offered by Rb/Sr whole rock dates of the Ottafjället dolerite dike swarm in the Särv Nappe (Claesson, 1976, 1977; Claesson and Roddick, 1983) which yielded ages of 720 ± 260 and 665 ± 10 Ma. The dikes penetrate the Toossfjället Group which is a correlative of the Hedmark Group (Kumpulainen and Nystuen, 1985).

Units of the Hedmark Group yielded few microfossil taxa, and this makes age assignments on palynomorph evidence difficult. The only units yielding relatively abundant and age-diagnostic acritarch taxa are the Biskopåsen Formation and the Birí Formation. Thus, *V. varangerii* occurs in the Ekkerøy Formation at the top of the late Proterozoic Vadso Group in Varanger Peninsula in northern Norway. Morphologically comparable forms are *C. kullingii*, *V. reticulatum*, and *V. walcottii*. The former was reported by Knoll and Calder (1983) from the late Proterozoic Hunnberg Formation of the Realldtoppen Group in Svalbard. The age is poorly constrained, but preceding the deposition of the Varangerian tillites in the stratigraphically higher Gothia Group, it was regarded as late Riphean. *C. kullingii* and *V. walcottii* were reported in the late Proterozoic (about 845-796 Ma; Elston and McKee, 1982) Kwagunt Formation of the upper Chuar Group in northern Arizona (Vidal and Ford, 1985). *V. reticulatum* was reported from the upper unit of the Visingsö Group (about 663-703 Ma; Bonhome and Welin, 1983). Although supported by extremely scarce evidence, the micropaleontological data are in harmony with the directly and indirectly available isotopic age data. It thus seems possible to conclude that units preceding the deposition of the most likely Vendian Moelv Tillite (Nystuen, 1985) may be in part contemporaneous with the deposition of the Visingsö Group.
in the adjacent intracratonic Visingsö Basin in southern Sweden (Kumpulainen and Nystuen, 1985; Vidal, 1985).

Only one sample of the Vangsås Formation (Vardal Sandstone Member) yielded two microfossil specimens attributed to the acritarch *Fimbriaglomerella minuta* which was reported from the early Cambrian in various parts of Baltoscandia and the East European Platform (Moczydłowska and Vidal, 1986, 1988; Moczydłowska, 1989). Its occurrence in the Vardal Member of the Vangsås Formation may be consistent with an early Cambrian age. Only unidentifiable acritarchs attributable to *Comassphaeridium* sp. were recovered from a sample of the overlying Ringsaker Quartzite Member of the Vangsås Formation. Indirect evidence suggests that the Ringsaker Quartzite Member (as its possible time equivalent Hardeberga Quartzite; Vidal, 1981; Moczydłowska, 1989) may be of early Cambrian Talsy age.

**SUMMARY**

The study of well-defined facies associations in sections of the late Proterozoic Hedmark Group sheds additional light on the environments of deposition that dominated the formation of the Hedmark Group. This combined study provides further information on the special conditions regulating productivity and organic carbon burial in basinal marine settings during late Proterozoic times. Undoubtedly, future studies of the stable isotope geochemistry of the Hedmark Group will provide further insight on the late Proterozoic and early Cambrian evolution of the Hedmark Basins and the paleoecological framework of the Riphean to early Cambrian biota.

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**APPENDIX I**

All coordinates are in the UTM system, referring to mapsheets in the M711 Series 1:50 000 (see map in fig. 1)

**BRØTTUM FORMATION**

**Lillehammer:** section at Lake Mjøsa. LILLEHAMMER 579000/6775300. Black mudstone samples M-77-62b, M-77-63, M-77-64, and M-77-65. M-77-63 fossiliferous, all other samples yielded black organic detritus.

**Fåberg:** LILLEHAMMER 575150/6782150. Black mudstone in road cut, 2 m above borehole level in core of open anticline, west of Gudbrandsdalslågen River. Sample M-77-66. Black organic detritus.

Maihaugen: LILLEHAMMER 579450/6776150. Vertically dipping turbidite sandstones and black shales (fig. 3). Samples M-77-70 to M-77-83. M-77-70, M-77-72, M-77-73, M-77-74, M-77-77, 78, 79, 80, 81, 82, 83 fossiliferous. All other samples yielded black organic detritus.


**BISKOPÅSEN FORMATION**

Biskopåsen: GJØVIK 587000/6763850. Sandstone bed below conglomerate. Sample M-77-39 is barren while M-77-39A is fossiliferous.

Hjellund: 3 km northwest of Biri. GJØVIK 585400/671900. Southern corner of quarry. Shale interbeds in conglomerate. M-77-60 fossiliferous black mudstone clast in conglomerate, M-77-61, M-77-62a, M-82-53, and 54 fossiliferous shale beds in conglomerate (fig. 7).

**BIRI FORMATION**

Kremmerodden: GJØVIK 588900/6758360. Road cut in Highway E 6 (former highway 4), westernmost and least disturbed part of section. M-77-53, 54, 55 samples of fossiliferous laminated black mudstone.

Hjellund: 3 km northwest of Biri, GJØVIK 585600/6761700. Road cut on west side of local road (former highway 4). Sample M-77-56 massive dark-gray calcareous mudstone 10.6 m above upper boundary of Biskopåsen Formation. M-77-57, as previous 12.0 m above top of conglomerate, and M-77-58, just above the Biskopåsen Formation.

Becksminne: 4 km south of Rena, RENA 627050/6775550. Road cut at Highway 3 through mudstones and sandstones of Biri Formation and conglomerate facies of Ring Formation (fig. 5). Samples of fossiliferous finely laminated shale M-77-96, 98, 99, 100.

Rise: GJØVIK 590150/6760100. Road cut at Highway 213 (former E 6) through shale and limestone of the Biri Formation and conglomerate of the Ring Formation. Fossiliferous samples and samples containing fragmentary microfossils derived from mudstone within the Ring Formation (M-77-37) and mudstone fragments (M-77-38, M-82-20, 21) within the conglomerate facies of the Ring Formation.

S. Bjørnásbru: Ásta River, ÁSMARKA 608100/6776500. Road cut, alternating dark mudstone and laminated calcarous argillite and impure sandstone. Samples of mudstone M-77-85 to M-77-95. M-77-85, 90, 91, 93, 95 fossiliferous. Other samples yield abundant dark organic detritus.

**BJØRÅNES SHALE MEMBER (BIRI FORMATION)**

N. Bjøråa: STOR-ELVDAL 599600/6838400. Road cut at Highway 3 through cleaved black laminated mudstone with sandy layers and abundant pyrite. Mudstones are stratigraphically underlain by grayish, arkosic sandstone of the fluvial Rendalen Formation. Sample M-77-114 consists of black mudstone and yielded black organic detritus.

**MOELV TILLITE**

Moelv: GJØVIK 591450/6755750. Gray, matrix-rich diamicite at the shore of Lake Mjøsa, at the old Moelv quay.

Skjærodden: 2 km south of Rena, RENA 627500/6777800. Road cut at Highway 3. Sample M-77-101 of fossiliferous gray-green matrix-rich diamicite,

**EKRE FORMATION (SHALE)**


**VANGSÅS FORMATION; VARDAL SANDSTONE MEMBER**

Storbekken: 12 km northwest of Osen, NORDRE OSEN 640000/6805750. Cleaved black, pyritous shale in upper part of sequence of gray feldspathic arenite. Few microfossils, abundant black organic detritus.

Grønnsjøbrenna: 10 km west-southwest of Jordet, JORDET 659700/6809500. Road outcrop at Highway 215 in black pyritous mudstone. Fossiliferous, organic rich sample M-77-106.


Smeia River: 3 km northwest of Jordet, JORDET 345800/6815800. Road cut on north side of road through alternating black mudstone coarsening upward into dark-gray sandstone and matrix-rich granular conglomerate. Samples rich in black organic detritus from black argillie (M-77-109), black mudstone (M-77-110), and black conglomerate (M-77-111).

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MOLECULAR FOSSILS AND MICROFOSSILS OF PROKARYOTES AND PROTISTS FROM PROTEROZOIC SEDIMENTS

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ABSTRACT. Organic matter is ubiquitous in Proterozoic sedimentary rocks. Although the bulk of it consists of amorphous and poorly characterized polymeric material, often highly altered from its original state in living organisms, certain components are either microscopically or chemically recognizable or both. This paper reviews and summarizes the results of recent biomarker studies of well-preserved sediments in the light of current micropaleontological and geochemical perceptions. We now know of several occurrences of Proterozoic hydrocarbons, including one from the Early Proterozoic, where there is good reason to believe that the biomarkers represent the remains of microbes living at the time the sediments were deposited. The identities of specific hydrocarbons in these assemblages correlate well with information derived from independent geological and paleontological observations. In particular, bitumens from sediments of the Early Proterozoic McArthur Group, Middle Proterozoic Roper Group (both in northern Australia), the Late Proterozoic Chuar Group (Arizona), Late Proterozoic sediments of southern and central Australia, and oils reservoired in Late Proterozoic sediments in Oman and Siberia all contain sterane, triterpane, and extended acyclic isoprenoid alkanes biomarkers, consistent with inputs from eukaryotes, eubacteria, and archaebacteria respectively. Furthermore, the biomarker distribution patterns are unusual when compared to Phanerozoic sediments, suggesting that Proterozoic microbial ecosystems were generally different in composition to those from younger sedimentary environments.

INTRODUCTION

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"Indeed the range of uninvestigated and insufficiently investigated materials and problems, and the diversity of applicable techniques and interests suggests that we are on the verge of a new era in the investigation of pre-Paleozoic rocks; organic geochemistry and paleomicrobiology will continue to play inter-related and significant roles. . . ." (Cloud, 1969). Although considerable progress has been made in the intervening years since this statement, we remain at an early and superficial stage in the search for reliable and comprehensive chemical and biochemical data concerning the exact nature of preserved Proterozoic microfossils. However, as suggested by Cloud (1969), significant and interesting new developments have resulted from the inter-play of disciplines, the application of new techniques, and the study of materials unknown or unrecognized at the time of earlier endeavors; decoding the record of