LATE PRECAMBRIAN SEDIMENTATION IN THE CENTRAL SPARAGMITE BASIN OF SOUTH NORWAY

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Late Precambrian (Eocambrian) sediments of south Norway, the so-called 'sparagmite sequence', consist of 1500–3000 m of graywackes, shales, arkoses, carbonates, and conglomerates. A system of coarse clastic wedges deposited as fan deltas are found along the margin of the basin, suggesting that sedimentation took place in well-defined grabens as a response to faulting. The Moelv Tillite is developed both as a massive diamictite and as a drop-stone facies in a laminated matrix, indicating local glaciations around the basin.

The sparagmite basin is interpreted as a rift valley structure related to the opening of the protoatlantic ocean. The late Precambrian rifting phase was followed by the Lower Cambrian and Middle Cambrian transgression.

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The late Precambrian sediments referred to as 'sparagmites' by Scandinavian geologists since the days of Esmark (1829) occupy a large region in south Norway and adjacent parts of Sweden. It was soon recognized that these sediments were younger than the surrounding Precambrian basement, and that they had a lower degree of metamorphism and more moderate tectonic deformation in the southern and eastern parts. The sparagmite sequence, Table 1, displays a wide range of lithologies and facies – graywackes, arkoses, orthoquartzites, conglomerates including tillites, shales, and carbonates.

Although the southern and eastern parts of the basin have generally escaped severe deformation, large-scale folding and tectonic imbrication complicate the task of stratigraphical correlation and sedimentological reconstruction. In many areas, notably in Østerdalen, exposures are scarce because of a thick cover of Quaternary deposits, notably glacifluvial deltas.

The sparagmite stratigraphy was first established in the southern parts of the basin around the northern end of the lake Mjøsa (Holtedahl 1953, 1960, Skjeseth 1963). For a historical review, see Skjeseth 1963. More recent mapping in other parts of the sparagmite basin (Bjørlykke 1965, 1969, Englund 1966, 1972, 1973 a,b) has demonstrated that although the stratigraphy of the Moelv – Rena district can be extended northwards with some

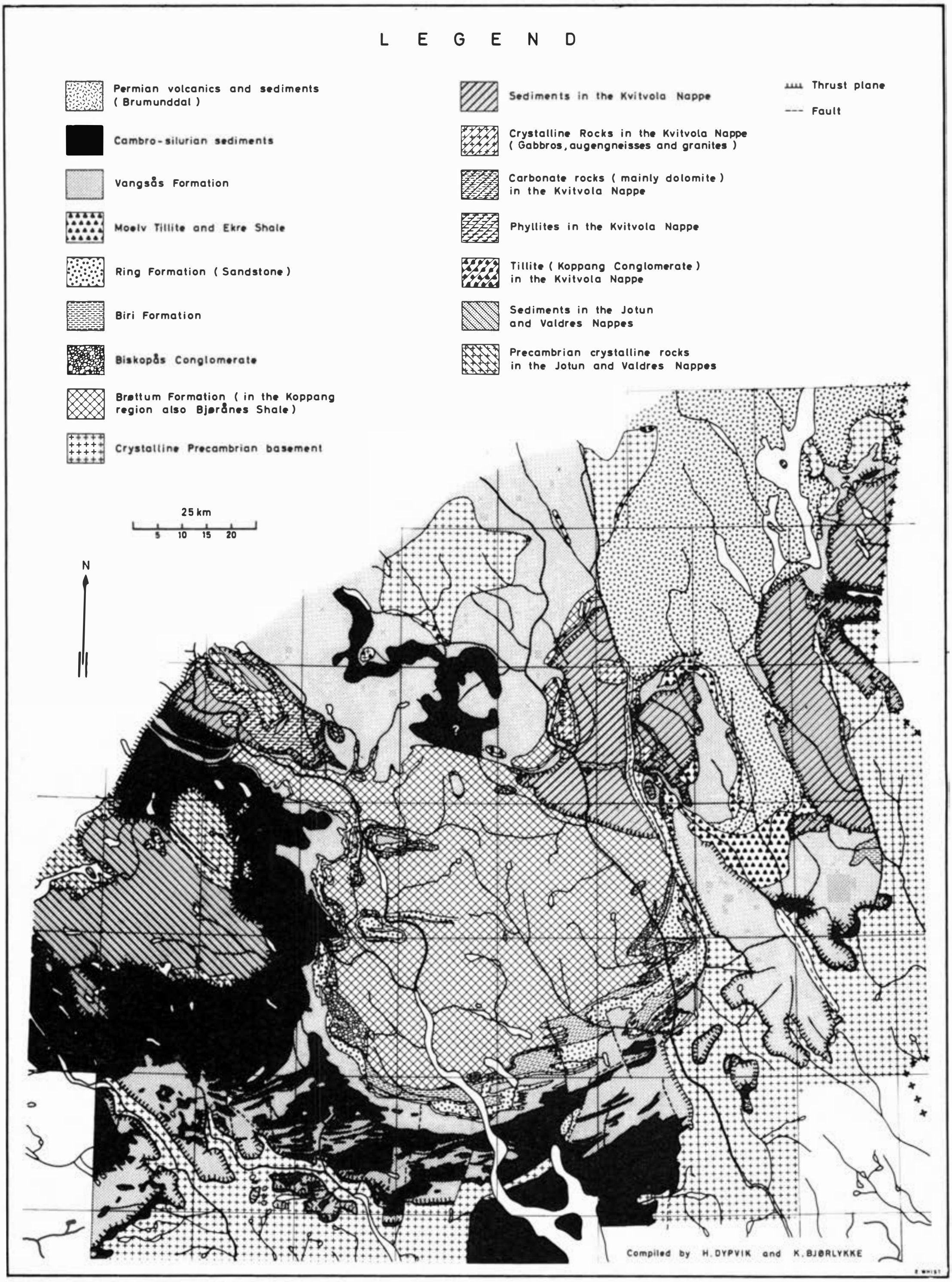

Fig. 1. Geological map of the sparagmite region, south Norway.

modifications, there are considerable variations in lithology and facies development.

The geological map of the region (Fig. 1) shows that the younger formations are mainly confined to the basin margin while the central parts are chiefly occupied by the lower part of the sequence (Brøttum Formation). This is attributable to the pronounced plunging of the fold axes towards the basin margins; they are generally aligned between northeast – southwest and east – west and dip eastwards towards the eastern margin of the main basin along Rendalen and westwards towards the western margin along Gudbrandsdalen. This pattern is probably the result of bulging contemporaneous with the folding of the sparagmites within the rigid surroundings of a Precambrian basin.

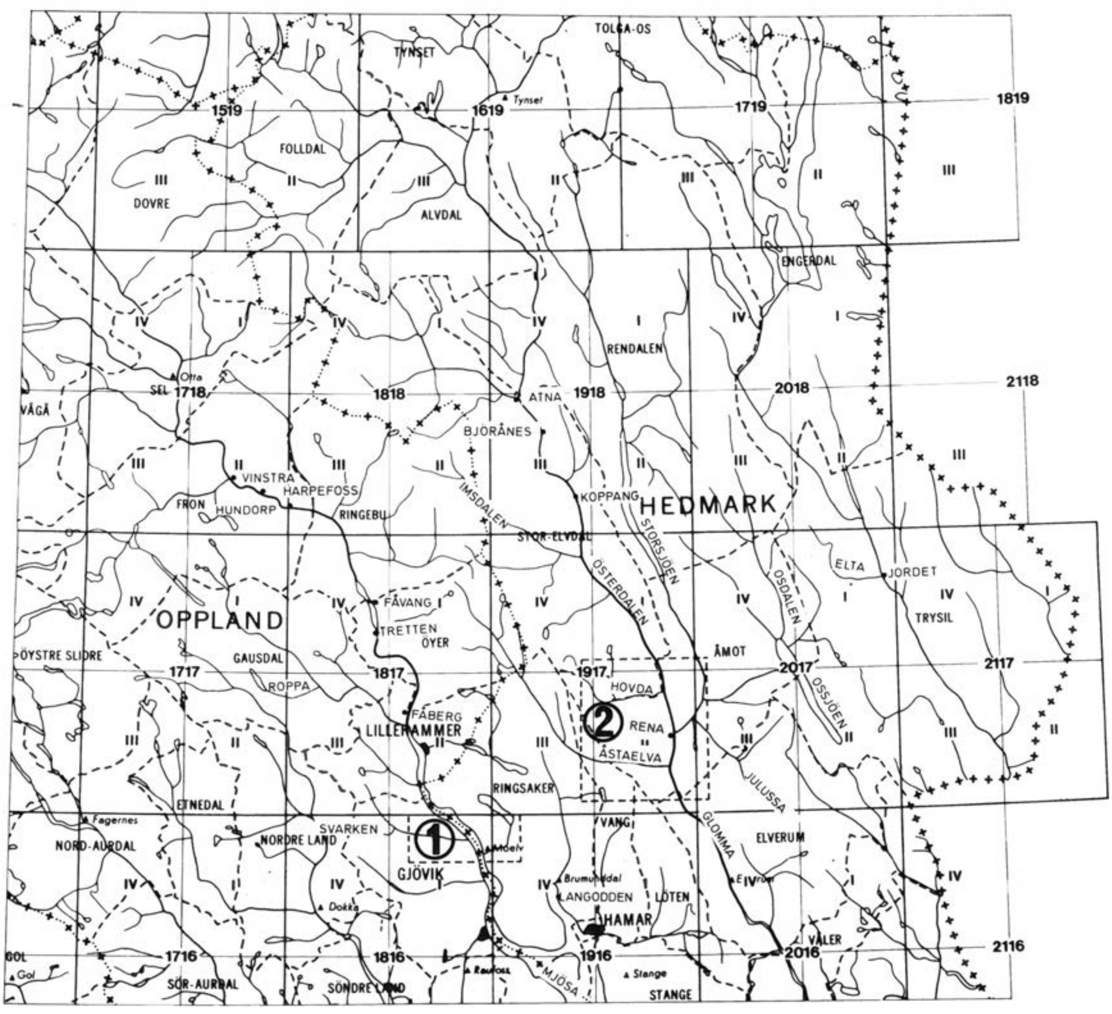

Fig. 2. Map showing the names mentioned in the text.

Most of the sediments in the southern part of the sparagmite basin can be regarded as para-autochthonous. The relationships between basements structures and sediments suggest that the tectonic transport of the latter is

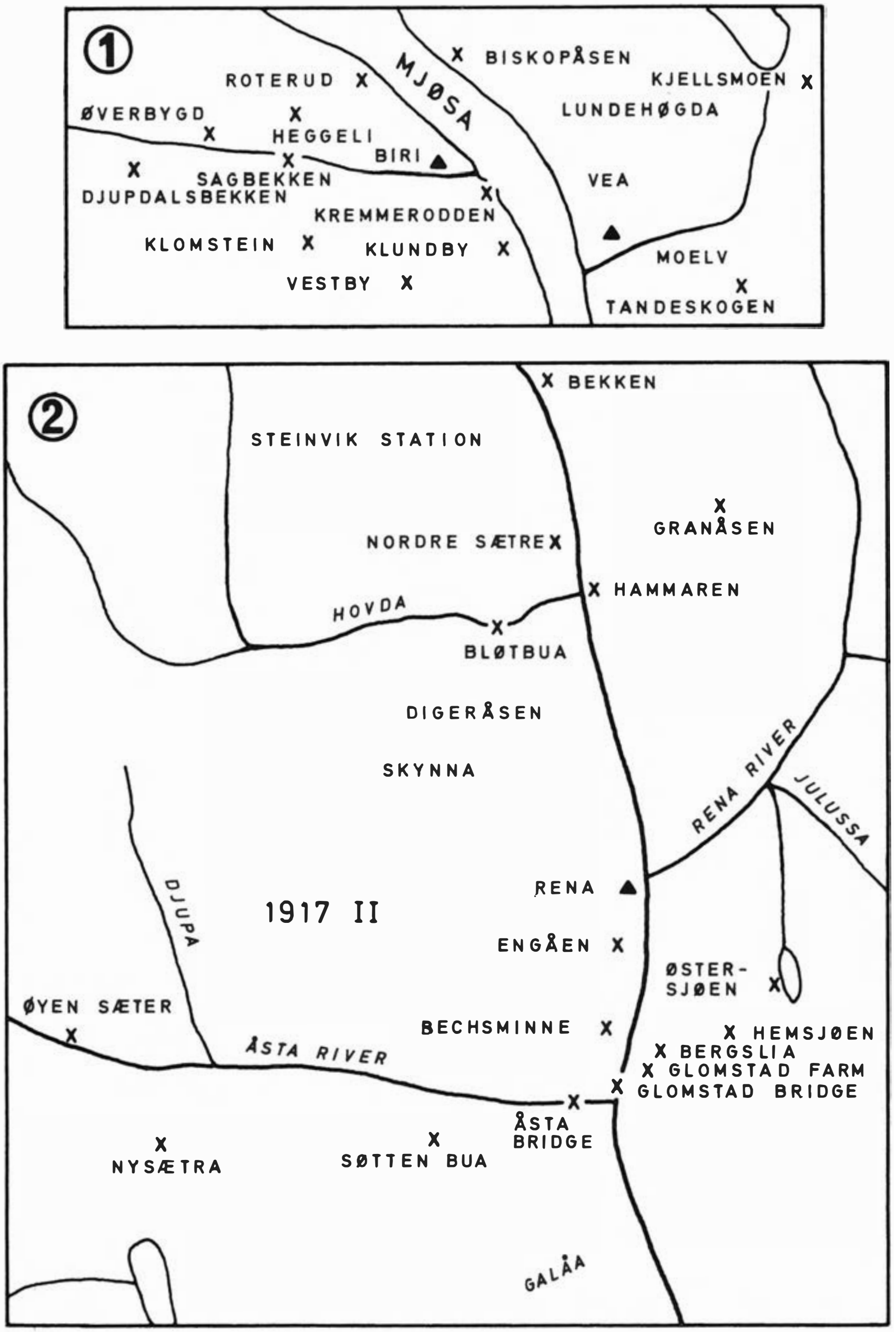

Fig. 3. Detailed map of areas 1 and 2. See Fig. 2.

Table 1. Stratigraphy of late Precambrian (Eocambrian) sediments in the sparagmite basin of south Norway.

	Hedmark Group	
Vangsås Formation	Ringsaker Quartzite Member Vardal Sandstone Member	Rena Subgroup
Ekre Shale		
Moelv Tillite		
Ring Formation		
Biri Formation		Lillehammer Subgroup
Biskopås Conglomerate		
(Biri Formation, mostly eroded)		
Brøttum Formation		

limited to 20–30 km in the southern part. The youngest part of the sequence, the Vangsås Formation, has been moved towards the southeast, sliding on the underlying Ekre Shale, and thrust onto the Cambrian alum shale 30–50 km southeast of the basement proper (Schiøtz 1902).

The Ring Formation may also have been relatively thrusted, the underlying sequence sliding on shales of the Biri Formation. Examples of this can be found on Map sheet Gjøvik, 1816 I (A. Bjørlykke 1973a) and Map sheet Rena, 1917 II (K. Bjørlykke 1976). Isopach maps of depth to basement based on magnetic anomalies (Åm 1976) indicate that the deeper part of the basin (3–4 km depth) was limited to the south by a contemporaneous fault running from Lillehammer to Rena. This suggests that the sequence underlying the Ekre Shale has not been thrusted more than about 20 km to the southeast. The fact that the Moelv Tillite predominantly contains clasts occurring in the locally surrounding basement is also evidence against long-distance thrusting of the main part of the Hedmark Group in the sparagmite basin.

The Brøttum Formation

The Brøttum Formation is the oldest formation in the Hedmark Group and its base is nowhere exposed. A drillhole sunk by Norges Geologiske Undersøkelse into the presumed deepest part of the succession at Fåberg (north of Lillehammer) failed to reach the basement after 500 m. Since the dip is always gentle and the top of the borehole was estimated at over 500 m below the top of the formation, a minimum thickness of 1000 m was deduced (Englund 1966). Geophysical data has since indicated 1500 m as the minimum thickness (Åm 1976). The Eldstad Formation previously considered older than the Brøttum Formation may now be regarded as younger than, or possibly equivalent to, the Vangsås Formation (Englund 1973a: 5).

Fig. 4. Graded sandstone beds in the Brøttum Formation. Locality: Maihaugen, Lillehammer.

In the Brøttum Formation sandstones and black or grey shale facies rapidly alternate, both vertically and horizontally. In the Rena district farther north in the valley of Østerdalen and as far as in the valley of Imsdalen, the formation is characterized by a coarse-grained conglomeratic arkose (Bjørlykke 1966) and shales are rare or absent. This facies may be interpreted as the middle zone of a fan delta (McGowen & Groat 1971).

Westwards towards Gudbrandsdalen graywackes and black shales prevail (Englund 1972, 1973a), indicative of a deeper water facies. Good examples of turbidites have been described from this region by Englund (1966, 1972). A depth of a few hundred metres would have been sufficient for such deposition.

A major obstacle to the reconstruction of the sedimentary evolution of this formation is the absence of marker horizons for correlation between the different areas. The most reliable comparisons are therefore obtained from the uppermost part of the Brøttum Formation directly underlying the Biskopås Conglomerate. The transition from this conglomerate down into the Brøttum Formation is accompanied by a distinctly larger grain size and lesser shale and matrix content in the area north of Rena than in the southern area around Mjøsa. The lack of good continuous sections makes it difficult to quantify this relationship.

In the Rena district to the north, southwest of Storsjøen, and in the area south of Imsdalen, the Brøttum Formation is a conglomeratic arkose. Bedding is recognizable by grain size variations rather than by shaly layers. Sandstone beds are massive or showing some grading (fining upwards) and may have thin (< 10 cm) dark shale on top. The uniform thickness of the sandstone beds and the scarcity of structures that could be interpreted as fluvial suggest that these sediments were deposited in the submarine slope facies of a fan delta rather than in the subaerial fluvial part. This facies usually shows grey colours. But red, cross-bedded conglomeratic arkoses are more prominent in the section along the river Rena south of Storsjøen and in several other locations on Map sheet Evenstad (1971 I) in beds underlying the Biskopås Conglomerate.

In Gudbrandsdalen from Lillehammer to Fåberg – Vinstra, the Brøttum Sparagmite has graded beds separated by shale (Englund 1966, 1972, 1973a) (Fig. 4). Petrographically these are muddy sandstones corresponding to graywackes deposited by turbidity currents (Englund 1966, 1972) in a submarine environment.

In the Fåberg – Øyer district, however, the facies changes stratigraphically into a shallower water facies towards the top of the formation near the Biskopås Conglomerate.

While the Brøttum Formation in Gudbrandsdalen and surrounding areas is dark or greyish, frequently containing considerable amounts of sulphides, the Rena district arkoses are red or grey with low sulphide and carbon contents, as would be expected from shallower more oxidized sandstones. The graywacke facies of the Brøttum Formation is in many areas in Gudbrandsdalen fingering into sandy shale representing distal turbidites and other fine-grained suspended sediments deposited in the deeper parts of the basin. The Bjørånes Shale (Bjørlykke 1965) in Østerdalen is most probably partly contemporaneous with the Brøttum Formation. It is difficult to assess how much of the Brøttum Formation is equivalent to the Bjørånes Shale. The fine grain size and the relatively high carbon content of the Bjørånes Shale suggest, however, a rather slow sedimentation rate.

The Biskopås Conglomerate

Regional mapping of the central sparagmite basin has identified several distinct fans spreading out into the basin from its inferred margin. As Skjeseth (1963) pointed out, these were separate deltaic structures related to the drainage pattern in the surrounding land areas; they may be listed as follows:

Fig. 5. The Biskopås Conglomerate. Locality: Simenstua north of Rena (94.90–24.20). Maximum cobble size 12–15 cm.

Moelv – Biri (Skjeseth 1963, Kirkhusmo 1968).
Rena (Bjørlykke 1966).
Gausdal (Løberg 1970).
Tretten – Øyer (Englund 1972).
Fåvang (Oftedal 1945, Englund 1966).

The Biskopås Conglomerate consists of generally well-rounded pebbles and cobbles, with occasional boulders (Fig. 5). These clasts commonly form a grain-supported structure, often with solution pits at their contacts. The matrix consists of coarse-grained, slightly clayey arkose. Interbedded between the conglomeratic layers are massive sandstones. The cobbles are often imbricated, as can be seen in the road section on both sides of Mjøsa, indicating transport from the south. Obvious fluvial characters such as cross-bedding or fluvial channels are generally absent. The transport direction of the Biskopås Conglomerate can also be inferred from the basin configuration and the pattern of fanning out into the basin (Fig. 6). The conglomerate is up to 200 m thick near the basin margin, and wedges out about 15 km from the basin margin.

The best section through the conglomerate is beside the road E6 at Biskopåsen (62.70 – 87.35) north of Moelv. Here the basal part is well exposed, the lowermost 10–20 m containing carbonate and phosphorite pebbles which have yielded microfossils (Spjeldnæs 1967, Manum 1967). These pebbles are angular, indicating a local source (Spjeldnæs 1967) as opposed to the well-rounded quartzite and gneiss pebbles that are also present. Pebbles of fresh, unaltered diabase are common too in the basal sequence. Spjeldnæs (1967) suggested that the carbonate and phosphorite pebbles were derived from transgressive carbonate beds deposited prior to the conglomerate formation. Thus the Biskopås Conglomerate is believed to have originated during a regional regression (Bjørlykke 1966, Spjeldnæs 1967), eroding first the shelf sediments and then underlying basement rocks (Fig. 6).

At Roterud on the western side of Mjøsa, there are good exposures of the conglomerate beside the lake (62.30 – 85.80) and in a quarry (61.80 – 85.40).

Following the strike westwards one finds a sequence of carbonate beds

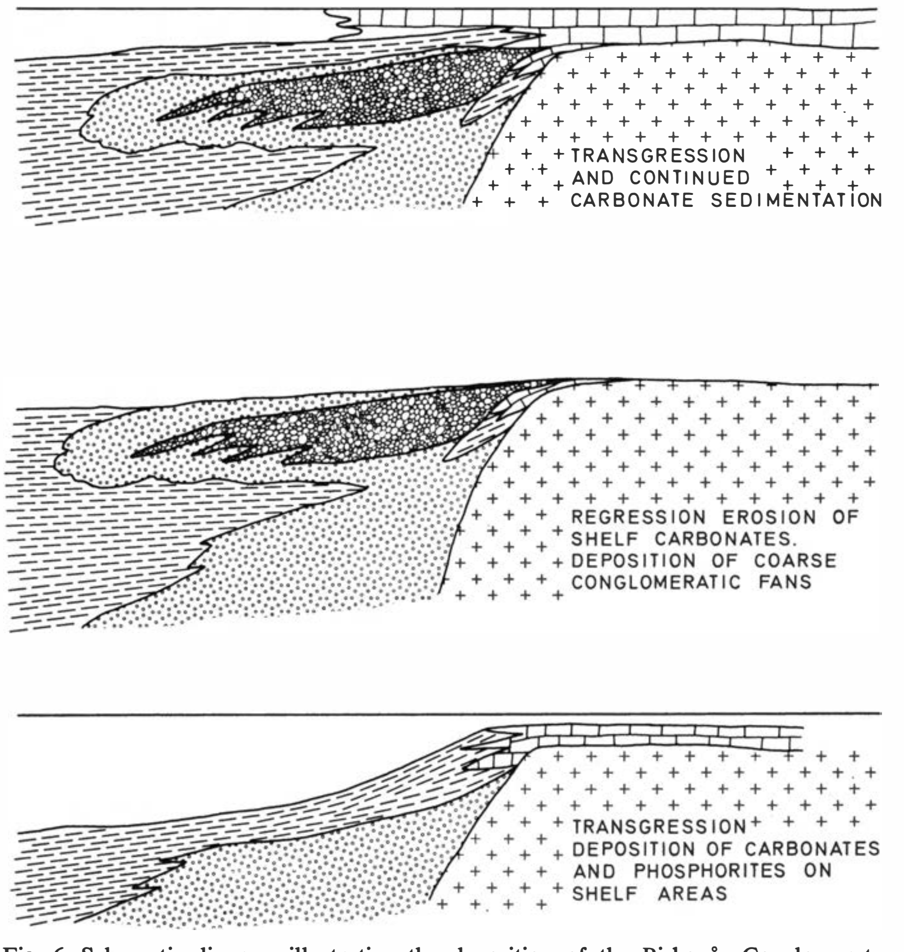

Fig. 6. Schematic diagram illustrating the deposition of the Biskopås Conglomerate. Partly after Spjeldnæs (1967).

between the Biskopås Conglomerate and the Brøttum Formation. These beds below the conglomerate were first noticed by Münster (1900) who named them the Brøttum Shale and Limestone. This unit was also recognized by Kirkhusmo (1968) and mapped in the Moelv and Biri districts by A. Bjørlykke (1973a). It provides additional evidence for a period of carbonate sedimentation preceding deposition of the conglomerate. The lithology of the carbonates underlying the conglomerate resembles that of the overlying carbonates, so it seems logical to include both in the Biri Formation (Bjørlykke, Englund & Kirkhusmo 1967). The Biskopås Conglomerate can therefore be regarded as a clastic wedge in the Biri Formation, though in much of the basin the lower carbonate unit was either eroded or never deposited.

Good exposures of the Biskopås Conglomerate are also found north of Rena, in Digeråsen, Hovda, and around Steinvik station. See also Holmsen & Oftedahl (1956). The conglomerate is also found in Imsdalen and boulders exceeding 1 m can be observed here.

The Hovda section (88.30 – 77.90) reveals a 12 m thick sandstone sequence immediately above the conglomeratic facies, with numerous trough cross-beds indicating fluvial transport to the NNW.

Clast lithology of the Biskopås Conglomerate varies strikingly within the basin. At Fåvang the pebbles and boulders are almost exclusively derived from meta-anorthosites (Oftedahl 1945, Englund 1966). In the Tretten–Øyer district grey gneiss-granite boulders predominate (Englund 1972). However, in Gausdal the basal clasts are mainly composed of granitic material changing to quartzite towards the top (Løberg 1970).

Pebble counts from eight localities in the Rena, Moelv, and Biri districts are shown in Table 2.

The absence of sedimentary structures which may indicate a fluvial environment suggests that the Biskopås Conglomerate was deposited in a submarine environment (R. Otter pers. comm.). These coarse clastic prograding fans, have, however, most probably been fed by fluvial drainage systems. The conglomerate may have been deposited during floods as course sheet flows directly into the basin or as a fluvial delta which became reworked into the submarine environment. We are, however, not dealing here with a deep submarine coarse clastic fan. Depositional depth is not likely to have been much more than 2–300 m as the conglomerate is often found to be conformably overlain by shallow water carbonates as at Rappa (Høy 1976) or by obvious fluvial sediments as in Hovda.

The fact that pebbles from carbonates deposited during a transgressive period (see p. 241) which were eroded first are concentrated towards the base of the conglomerate (i.e. at Biskopåsen) shows that the conglomerate has not been deposited as one unit first and then later been reworked. The data show granitic rocks to be rare or absent at six of these localities, while values around 10 % were recorded at the other two. Carbonate pebbles are also exposed in the basal part of the Biskopåsen section. Quartzite, and to

Table 2. Composition of pebbles in the Biskopås Conglomerate.

			Rena district		Moelv (Biskopåsen)		Biri district	
	1	2	3	4	5	6	7	8
No. of pebbles counted	64	80	207	117	109	92	227	76
Rock types	%	%	%	%	%	%	%	%
Light quartzite	15.5	42	68	41	72	60	69	66
Red quartzite	29	22.5	0.5	4	2	12	1	4
Quartz porphyry	18.5	30	9	6	3		10	
Feldspar phorphyry	23		19	41	9		2.5	
Vein quartz	14.5	4.5	2	4	12	9		12
Granite and gneiss			1.5		2	10	1.5	11
Other rock types		1		4		9	16	7
	100	100	100	100	100	100	100	100

Locality
1. Hammaren (89.40–25.80) about 8 km north of Rena.
2. Nordre Sætre (90.5–25.40), railway section, about 10 km north of Rena.
3. Road section by Bekken (94.90–24.20) about 15 km north of Rena.
4. Granåsen (91.50–30.20), about 10 km north of Rena.
5. River section in the river Hovda (88.30–22.90), Rena district.
6. Section through the upper part of the conglomerate in Biskopåsen (63.70–87.70) north of Moelv.
7. Roterud (61.80–85.40), Biri district.
8. Øverbygd (60.80–79.50), Biri district.

Pebble counts by R. Otter

a varying degree porphyries, are the dominant clast types in these districts. Vein quartz is also common, so the total clast composition does not correspond to the local basement. Although quartzites and porphyries may have had a wider distribution in late Precambrian times (H. Skålvoll, pers. comm.), the abundance of granite and gneiss pebbles in the Moelv Tillite and of potash feldspar in the Brøttum and Ring Formations in the same areas, suggests that much of the local basement consisted of granite and gneiss-granite at that time. The Biskopås Conglomerate is therefore assumed to consist of more long-transported material. The pebbles are also well rounded.

At the Hovda section, the massive Biskopås Conglomerate is overlain by a trough-fill cross-bedded unit corresponding to the distal zone of a fan delta (McGowen & Groat 1971).

The conglomerate resembles Quaternary glaciofluvial conglomerates. However, there is no evidence of ice-contact and the sorting is generally better than in glacial outwash deposits. Thus there is no direct evidence of a glaciation contemporary with this conglomerate. If glaciation was restricted to upland areas some distance from the basin, however, powerful proglacial streams could have transported outwash debris into the basin. The pebbles are often facetted but rounded and are very similar to Quaternary glaciofluvial pebbles found in the same area.

The Biri Formation

The Biri Shale and Limestone (= Biri Formation) (Bjørlykke et al. 1967) has been briefly described by several authors in their regional studies, viz. Skjeseth (1963), Bjørlykke (1966), Englund (1966), and Løberg (1970). However, no attempt has been made to interpret depositional environment of the Biri Formation in a broader context.

Skjeseth (1963) has observed that the massive, pure limestone with ooliths occur near the assumed margin of the basin, while calcareous shales have been formed in the more central, deeper parts. This graduation from shelf to basin facies has been clearly demonstrated in the Rena district (Bjørlykke 1966).

The present account is based mostly on sections studied in the southern part of the sparagmite basin in the Rena, Moelv, Biri, and Gausdal districts (Høy 1976).

The Moelv – Biri district

The outcrops of the Biri Formation occur over lage areas, supporting most of the fertile farm land (A. Bjørlykke 1973a).

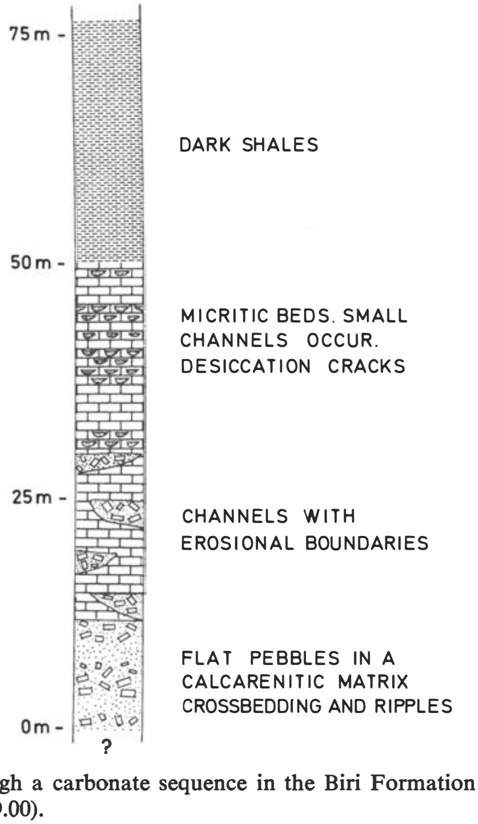

Fig. 7. Section through a carbonate sequence in the Biri Formation at Kremmerodden, Biri district (58.40–89.00).

Fig. 8. Intraformational breccia (tidal channel) in the Biri Formation at Kremmerodden. Longest clast about 1 m.

One of the best exposures is at Kremmerodden, about 2 km south of Biri (Fig. 7). Here an intraformational breccia can be studied in a road section (58.40 – 89.00) and also in a shore section on the peninsula near Vea (58.50 – 90.60) on the east side of Mjøsa.

This breccia consists exclusively of tabular carbonate clasts which may exceed 80 cm in length (Fig. 8) and are lithologically similar to the surrounding limestone. Some of the fragments seem to have been plastically deformed prior to their deposition. The breccia can in some places be shown to be incised into undisturbed carbonate beds.

This type of breccia can only have been formed by tidal channels undercutting carbonate bank sediments, which subsequently fell into the channel. The absence of foreign clasts is consistent with the formation of tidal channels which lack a supply of terrestrial sediments. Despite folding and tectonic deformation, a 50 m sequence of carbonates – mostly with beds of breccia – can be reconstructed.

Fig. 9. Peel of ooides from the Biri Formation. The ooides in the centre of the picture are partly dissolved by stylolites. Heggeli, Biri district (61.20–82.30). Width of section 1 cm.

In the basal part the breccias are comprised of large clasts in a calcarenite matrix, and indicate a channel depth of 2–3 m. The clast size decreases towards the top of the sequence to less than 10 cm, with the breccia occurring as shallow lens-shaped channels or locally eroded and broken up carbonate beds. In these smaller channels the clasts are often imbricated. Adjacent to the channels the carbonate beds display tension cracks possibly caused by sagging of the banks (Høy 1976). While the basal part of the sequence is exclusively calcitic, dolomite occurs frequently further up, alternating with micritic limestone beds. The average size of the clasts and inferred channel depth decreases upwards in the Kremmerodden section. This can be interpreted as a regressive sequence from an intertidal environment to a high-intertidal and supratidal environment. These limestones contain laminations of probable algal origin, though good examples of algal domes have not been observed here. There is no evidence of supratidal conditions in the overlying black shales which probably have been accumulated in deeper water following a regional transgression.

West of Biri, at Heggeli (61.20 – 82.30), about 50–60 m of massive limestones are exposed. They consist predominantly of ooids in a micritic matrix, and algal structures, partly recrystallised (Fig. 9). Well-rounded quartz grains of similar diameter to the ooids are also abundant. This limestone is dark and has most probably been deposited in a reducing environment

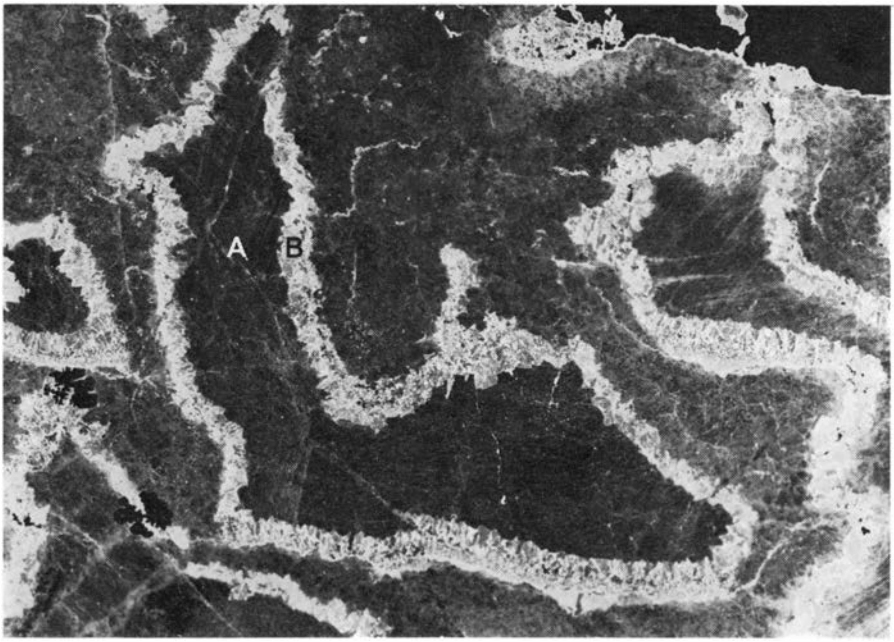

Fig. 10. Fenestral structures filled with sparry calcite (A) and rimmed with dolomite (B). Stained with alizarin red. Width of section 2.2 cm. Locality: Svarken, Biri district (60.30–68.20).

below wave base, next to a carbonate bank responsible for the supply of ooids and quartz grains.

At Svarken (60.30 – 68.20), 15 km west of Biri, massive limestone is exposed below the Moelv Tillite (Fig. 26). This limestone contains fenestral structures (Fig. 10) up to 6 cm wide, filled with sparry calcite and rimmed with dolomite. Abundant fossil karst formations indicate a substantial period of exposure and weathering.

Glomstad – Hemsjøen – Østersjøen area
(Map sheet Rena, 1917 II)
The Biri Formation is well exposed in several localities in this area along the strike from the bridge at Glomstad to Hemsjøen and Østersjøen (Bjørlykke 1966), but the lack of a continuous exposure makes it difficult to correlate the stratigraphical positions from section to section.

Beside the bridge at Glomstad (74.10 – 27.80), dolomite beds and micritic limestones are visible, deformed by an isoclinal fold. The micritic limestones display well-developed parallel laminations and are interbedded with green and red shale bands. Intraformational breccias of the Kremmerodden type (see p. 245) are present, varrying fragments of micritic limestone and dolomite; the thickness of each breccia is only about 50 cm with individual clasts smaller than at Kremmerodden (Fig. 8), rarely exceeding 10 cm (Fig. 11).

The Biri Formation is also exposed as pure limestones, around the farm

Glomstad (74.50 – 28.30), while about 1 km northeast, massive limestones with ooids and quartz grains have been observed. Well-rounded quartz and feldspar sand grains are floating in a carbonate matrix and are tectonically deformed, displaying a lineation parallel to the direction of tectonic transport (370°–170°) (Bjørlykke 1966). Ooids and quartz grains of the same size are also found mixed together. No dolomite has been found in this facies.

At Hemsjøen (76.10 – 30.80) the Biri Formation occurs as an almost pure, pink limestone containing up to 98 % $CaCO_3$ and very little magnesium (< 0.5 % MgO) (Bjørlykke 1966, Høy 1976). During low water dome-shaped stromatolites are visible at Hemsjøen. This limestone has a speckled appearance due to patches of white sparry calcite which probably represent fenestral structures (Bathurst 1971).

To the northeast at Østersjøen (77.75 – 32.20), micritic limestone and dolomites are found with intraformational breccias similar to the ones at the bridge at Glomstad (Fig. 11).

Road sections between the farms Glomstad and Bergslia (75.30 – 28.60) reveal grey, partly calcareous shales, occupying a transitional position between the massive limestones and the overlying Ring Formation arkose. The latter basal sandstones in this section have a carbonate cement and the feldspars are partly replaced by calcite (Bjørlykke 1966: 23).

In the Glomstad sections the dolomite beds seem to occur at the lowest stratigraphical position while the micritic limestone with ooids represents a higher level, passing into shales and the overlying Ring Formation. This is interpreted as a transgressive sequence up to the basal shales of the Ring Formation. The stromatolite facies at Hemsjøen most likely represents an horizon above the dolomite which is either absent or not exposed at Glomstad. The dolomitic beds at Østersjøen are correlated with the same facies at the bridge of Glomstad.

These dolomites and micritic limestones with their intraformational breccias have been formed in a supratidal to intertidal environment. Laminations within the micritic beds are believed to be due to algal mats. The dolomites are early diagenetic products representing depositional periods in the supratidal

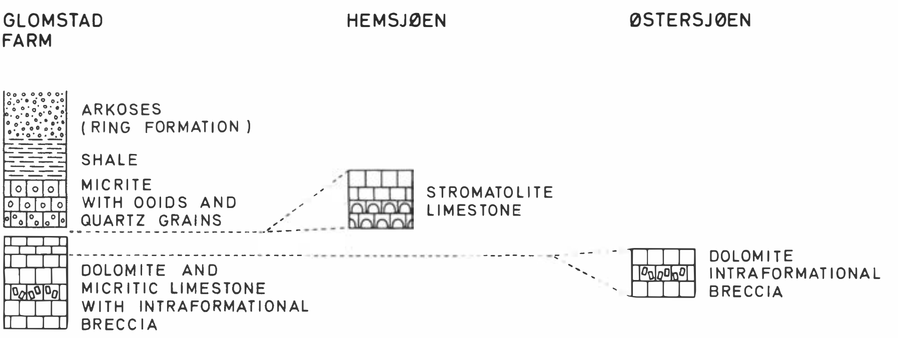

Fig. 11. Lithofacies in the Biri Formation at Glomstad, Rena district.

zone whereas the breccias were formed in tidal channels in the intertidal zone.

At Hemsjøen, the absence of dolomite and mud cracks and the presence of domed stromatolites indicate an intertidal position. Fenestral (birdseye) structures can form in both supra- and intertidal environments (Shinn 1968).

The Hovda river section

(Map sheet Rena, 1917 II)

A section through the Biri Formation is exposed by the river Hovda near Bløtbua (87.60 – 23.50) in the core of a synclinal structure, with the Biskopås Conglomerate exposed on either side. The top of the conglomeratic beds is overlain by a sandstone (12–15 m) interpreted as a more distal facies of the fan delta. This sandstone is succeeded by a 10 m thick unit of green shale. The sandstone/shale ratio decreases upwards towards a 15 m unit of greenish calcareous shales containing occasional carbonate nodules (up to 30–40 cm in diameter). The central part of the syncline is comprised of grey and green laminated calcareous shales with carbonate concretions.

One thin (0.5 m) conglomerate bed is observed within this unit, consisting of clastic quartz granules (1 cm) and larger carbonate clasts.

In the Hovda section the transitional beds between the Biskopås Conglomerate and the Biri Formation are clearly displayed, demonstrating the transgressional sequence from the fluvial environment of the conglomerate to the marine environment of the Biri Formation. No massive pure carbonate beds are found in this section, but this may be due to the fact that the upper part of the section is not exposed. It seems more probable that the Hovda section is located outside the carbonate shelf zone and represents a somewhat deeper water environment.

Øyen sæter – Nysætra

(Map sheet Åsmarka, 1917 II)

In an area with few clean continuous exposures the sections at Øyen sæter – Nysætra in the valley of Åsta provide vital information on the facies distribution within the southern part of the sparagmite basin.

The greatest exposed sequence is along a small tributary which joins the river Åsta south of Øyen sæter (75.80 – 12.00). About 200 m of shale and carbonate beds are exposed (Fig. 12). The basal 45 m of this section consist of red and green calcareous shales, with the colours often cutting across primary bedding, indicating a secondary, late diagenetic alteration.

These shales are overlain by a 100 m thick unit of alternating shale and thin (0.5 – 0.6 cm) micritic carbonate layers. Towards the top of this unit these carbonate beds are often found to be broken up, and occur as intraformational conglomerates. Mud cracks filled with coarser clastic material are also common.

The next unit is a 6 m thick massive carbonate which also contains intraformational conglomerates and resembles the Kremmerodden facies at Biri.

The top unit of this section is an essentially shaly, 35 m thick sequence.

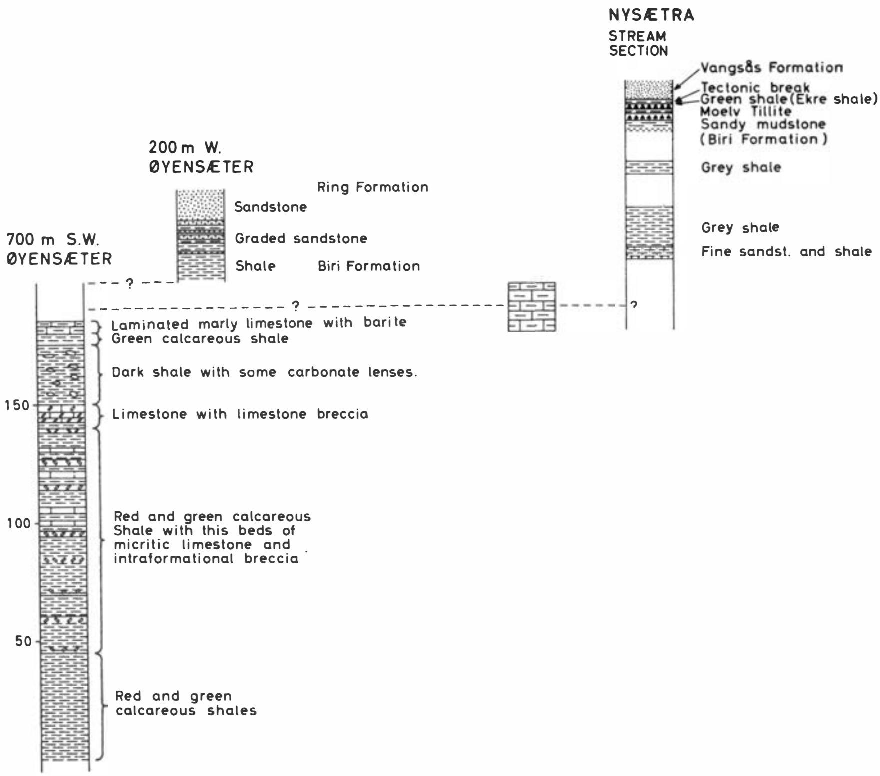

Fig. 12. Section through the Biri Formation and overlying sediments at Øyen sæter (75.80–12.00) and Nysætra (73.70–15.15) in Åstadalen.

It consists mainly of black shale with calcareous lenses. The uppermost 10 m are green calcareous shale passing upwards into finely laminated calcareous shale with small barite lenses (Høy 1976).

At Nysætra (map sheet Rena, 1917 II) about 4 km to the southeast of the above-described section (73.70 – 15.15), a shorter and discontinuous sequence is exposed through the upper part of the Biri Formation beneath the Moelv Tillite (Fig. 12).

The Moelv Tillite is here developed as a conglomerate of pebbles scattered in a laminated, silty shale facies, representing a typical ice-drop facies (see p. 274). Shale and sandy mudstone underlie the Moelv Tillite with a few scattered granule clasts in their upper layers. Some sandy beds contain carbonate cement (up to 16 % carbonate) of both calcite and dolomite. Further downstream 50–60 m of grey to dark grey shale containing some sandy beds (graywackes) in the lowermost part are discontinuously exposed. An isolated exposure at Nysætra consists of finely laminated (0.1 – 1.0 cm) dark limestone, which lacks evidence of subaerial exposure such as mud cracks or tidal channel conglomerates. Thin sections show that the laminae is made up by recrystallized spherical structures. In similarly laminated shale from Sagbekken at Biri these spherical structures can be identified as small ooids (30–80 μ) which may have been transported from the shallower part of the basin (Høy 1976).

The environmental interpretation is that the Biri Formation in the stream section south of Øyen sæter indicates a shallow marine environment with sedimentation of both carbonates and fine-grained clastic material. Intraformational carbonate pebble conglomerates and the occurrence of mud cracks confirm deposition in the intertidal to supratidal zone. The black shale present in the upper part of this sequence probably accumulated in somewhat deeper water, suggesting a regional transgression. The laminated limestones in the top of the section may be equivalent to those at Nysætra. These laminated carbonates with barite lenses may correspond to a subtidal facies in which ooids were blown in or by suspension in water during storm. The barite could have formed as a precipitate from the oxidation of sulphides in black mud (Antun 1967, Bjørlykke & Griffin 1973).

The great significance of the Nysætra section lies in the lack of a coarse clastic facies located beneath the Moelv Tillite and corresponding to the Ring Formation. The Moelv Tillite grades conformably downwards into the underlying fine-grained muddy sediment, suggesting that the Nysætra section represents a somewhat deeper water facies beyond or between coarse clastic deposits (fans). This accords well with the tillite being formed as an ice-drop facies.

Along Åsta, however, the shales of the Biri Formation grade upwards into a sandstone which most probably corresponds to the Ring Formation. This sandstone is comparatively fine grained compared to the typical development of the Ring Formation and is interpreted as a distal facies of the coarse clastic fans (see p. 266). The rapid facies change observed here may be partly due to tectonic compression in a N-S direction.

The river Roppa section

(Map sheet Follebu, 1817 III)

The Roppa section (88.00 – 57.50) provides an almost continuous succession from the Biskopås Conglomerate to Cambrian shales. The sequence is displayed and partly repeated by one fault (Fig. 13).

The transition from the Biskopås Conglomerate to the Biri Formation (see p. 243) is also marked here by the development of a sandstone representing a distal fan facies. Here the Biri Formation totals approximately 70–100 m of carbonate beds together with conglomerates composed of clastic carbonates and siliceous basement material. A calcareous shale with dolomite lenses and small breccias form the basal 10 m at the section and were probably deposited during an early transgressive phase of the Biri Formation.

Disconformably overlying this unit is a conglomerate containing both carbonate clasts and basement boulders; the former are angular and must have been eroded locally while the latter are rounded due to longer trans-

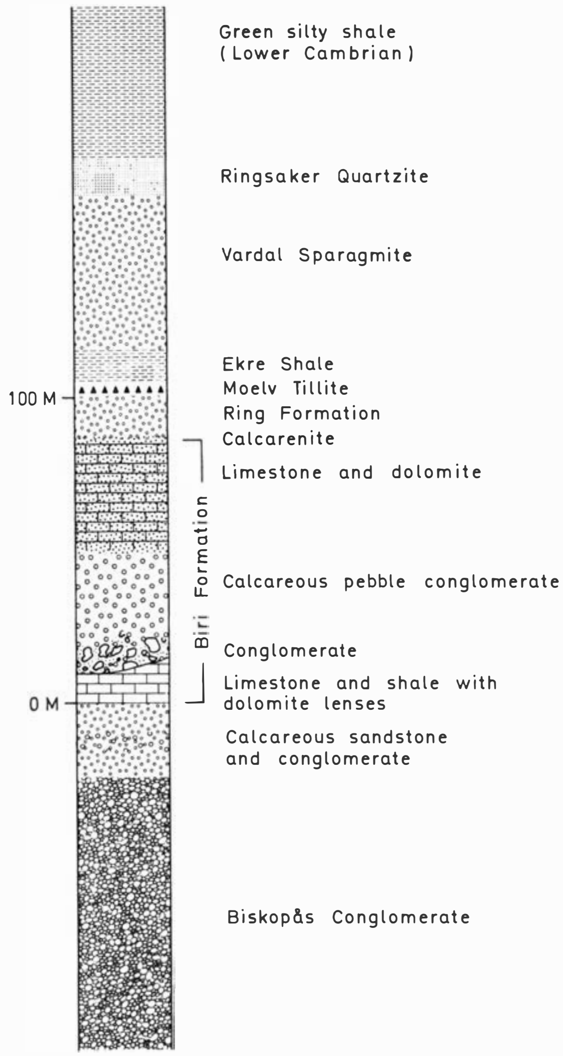

Fig. 13. Section based on almost continuous exposures along the river Roppa, Gausdal district (88.00–57.50).

port or reworking from the Biskopås Conglomerate farther west. This conglomerate in the Roppa section has been interpreted as a tillite by Løberg (1970) on account of its poor sorting and angular boulders. It comprises, however, the basal part of a sequence with an erosional base and more probably originated in a fluvial channel eroding the underlying carbonates. This also agrees better with its stratigraphical position – the conglomerate grades upwards into coarse cross-bedded calcareous sandstones and gravel conglomerates.

The upper 30 m of the Biri Formation consists of limestone and dolomite, partly as calcarenites representing shallow marine conditions. The clastic facies equivalent to the Ring Formation below the Moelv Tillite is only 12 m thick.

Summary of the Biri Formation

The Biri Formation comprises a great variety of lithofacies from dark shales to massive carbonate rocks. Rapid lateral changes of facies from a carbonate shelf environment to basin mud sedimentation (shale) occur over short distances. The scarcity of continuous sections through the Biri Formation and the absence of marker beds present a major problem in the detailed reconstruction of the palaeoenvironment. The accessible data do allow some conclusions to be drawn regarding facies distribution within the basin, and from the descriptions presented above the following main facies types are recognized:

Micritic limestone with algal structures deposited in a shallow subtidal environment on a carbonate platform.

Algal structures are mostly parallel lamination with laterally linked hemispheroids (LLH) which are characteristic of protected environments (Logan 1964, Bathurst 1971).

Carbonates and shales with algal lamination, often containing mud cracks and intraformational conglomerates, deposited in an intratidal to supratidal environment with tidal channels. Dolomite is commonly associated with these sediments:

Oolitic limestones, often mixed with quartz grains, deposited along the carbonate platform margin.

Calcareous shales and sandstones consisting of carbonate sand and mud mixed with siliceous sediment, deposited on the platform slope.

Dark shale deposited as intra-basin mudstone. The carbonate content varies with the supply of carbonate mud from the shelf.

Depositional history. – Carbonate sedimentation in the sparagmite basin commenced prior to the deposition of the Biskopås Conglomerate. This lowermost carbonate unit was recognized by Münster (1900) and Vogt (1924) and referred to as the 'Brøttum Limestone and Shale'. It was later described by Kirkhusmo (1968) from the Moelv district, and mapped at Biri by A. Bjørlykke (1973a).

There is very little observable difference between the facies development of the carbonates underlying the Biskopås Conglomerate and those overlying it. Since this conglomerate represents a clastic wedge that was probably deposited rapidly in discrete fans or deltas, Bjørlykke, Englund & Kirkhusmo (1967) proposed that the 'Brøttum Limestone and Shale' should be regarded as a lower member of the Biri Shale and Limestone Formation, now Biri Formation. In other districts (e.g. Rena) this lower member is absent, due probably to erosion preceding the Biskopås Conglomerate deposition. Although carbonate pebbles are rare in this conglomerate in the Rena district, such pebbles of intraformational conglomerates have been found, indicating carbonate deposition prior to the Biskopås Conglomerate.

The basal part of the Biskopås Conglomerate in the Moelv – Biri district is rich in carbonate pebbles. Phosphoritic pebbles (Spjeldnæs 1967) also point to shallow marine shelf deposition preceding the Biskopås Conglomerate. These phosphoritic pebbles have yielded microfossils – acritarchs (Manum 1967) and *Papillomembrana,* a more 'advanced' fossil of uncertain classification (Spjeldnæs 1967).

The upper member of the Biri Formation lies between the top of the Biskopås Conglomerate and the base of the Ring Formation. Both boundaries are obviously diachronous, and where the Ring Formation is missing (Nysætra section) the top of the Biri Formation is defined by the base of the Moelv Tillite.

Where the Biskopås Conglomerate is absent (i.e. at Djupa, Bjørlykke 1966) the base of the Biri Formation will coincide with the top of the Brøttum Formation. In the Bjørånes district (Bjørlykke 1965) the base of the Biri Formation cannot be defined since the time-equivalent sediments of the Brøttum Formation are also developed as shales. Here the stratigraphical unit 'Bjørånes Shale' includes beds that are probably chronostratigraphically equivalent to the Ring Formation, Biri Formation, Biskopås Conglomerate and probably also parts of the Brøttum Formation. In Gudbrandsdalen in the Tretten – Øyer district (Englund 1972) the Biri Formation is developed as calcitic limestone mixed and interbedded with shale. A similar facies prevails to the north at Fåvang, while to the north-west at Hundrop and Harpefoss only dolomite is found. Here the Biri Formation is directly overlain by the Vangsås Formation (Englund 1973).

The transgression which followed the regression associated with the Biskopås Conglomerate deposition could be eustatic if the regression was contemporary with the first glaciation in Finnmark. The transgression would then be attributable to the ice melting during the deglaciation. Theoretically such a transgression would probably have extended far beyond the basin margins, possibly covering large areas of the surrounding Precambrian shield. There is little evidence left to confirm this, though at Jordet in Trysil the Elta Limestone – equivalent to the Biri Limestone – forms the basal part of the exposed sequence in the eastern basin (Skjeseth 1963), suggesting a continuous carbonate shelf between the Rena and Trysil districts. There is no evidence of hypersaline conditions in the Biri Limestone except in the supratidal environment, which indicates that the sparagmite basin and surrounding shelf areas possessed free circulation with the open sea to the west, or alternatively that precipitation and runoff were sufficient to dilute the basin water.

Dolomite is rare or absent in the subtidal carbonate shelf environment but is commonly formed in the supratidal environment (see p. 16). This suggests that the main water body in the basin had a Mg/Ca ratio comparable with that of normal sea water.

The Ring Formation

The name 'Ring Formation' was introduced (Bjørlykke et al. 1967) to replace 'Moelv Sparagmite' in order to conform with international rules of stratigraphical nomenclature, since the geographical name Moelv is also used in the stratigraphical name 'Moelv Tillite'.

The Ring Formation comprises a series of sandstones – mainly arkoses and conglomerates – underlying the Moelv Tillite. Because earlier descriptions of the Ring Formation were mainly based on the type section at Moelv and other sections located around the sparagmite basin margins, the high degree of lateral facies variation was not appreciated until systematic mapping was carried out both to the north and within the more central parts of the basin.

In the Bjørånes district north of Koppang, Bjørlykke (1965) described sections where the Moelv Tillite developed as a thin ice-drop facies underlain by black shale (Bjørånes Shale). This was an example of a central basin facies in which no coarse clastic sediments were deposited during Ring Formation time and where this formation was therefore missing. Later, Englund (1966, 1972, 1973a) also described developments of the Ring Formation that differed from the development of this formation at Moelv (Kirkhusmo 1968) and Rena (Bjørlykke 1966) localities.

Great thicknesses of the Ring Formation sandstones also occur along the Engerdalen fault zone (Holtedahl 1921). Skjeseth (1963) and Englund (1973a) observed that the distribution of the conglomeratic facies of the Ring Formation coincided with that of the Biskopås Conglomerate.

Description of sections

The present study aimed at interpreting the depositional environment of the Ring Formation from more detailed observations of the vertical and lateral facies variations. Unfortunately continuous exposures are scarce in this region, preventing detailed palaeo-facies mapping, while folding and local fold thrusts compound these difficulties. The following description is therefore of necessity largely based upon some critical sections that show good stratigraphical control.

The Bechsminne section. – This long section is exposed along the main road (riksvei 3) 3.5 km south of Rena (75.90–27.20), through the basal part of the Ring Formation, and was first described by Bjørlykke (1966). See also map sheet Rena 1917 II (K. Bjørlykke 1976). The section can be divided into two parts.

The basal part (26 m thick) overlying the Biri Formation consists of alternating conglomeratic sandstone (maximum grain size 4 cm) and grey, finely laminated, silty or sandy shale (Fig. 14). Loadcasts are common at the base of the conglomeratic sandstone beds while ripple marks are observed in some of the sandy shales.

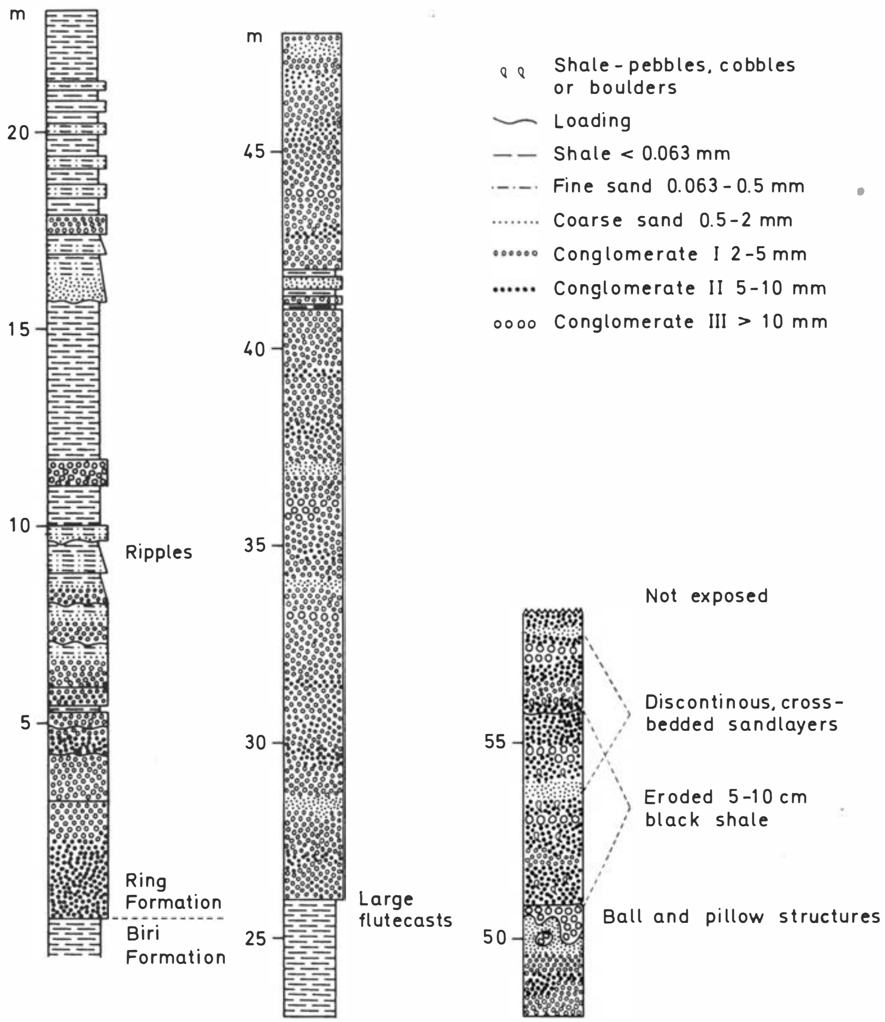

Fig. 14. Section through the lower part of the Ring Formation at Bechsminne, Rena district (75.90–27.20).

The upper part (32 m thick) is composed of a continuous succession of conglomeratic arkose without any shale interbeds thicker than 10 cm. The basal part of this upper unit has an erosional contact towards the underlying shale bed.

Large scale flute casts (3–4 cm) overprinted by smaller ones (10–20 cm) are well exposed at the sole of the conglomeratic sandstone unit, and indicate transport from east to west, allowing for tectonic rotation (Bjørlykke 1966). This conglomerate can be seen incised into the underlying shales, which suggests deposition in a relatively flat-based eroding channel. Pebble conglomerates with 3–4 cm sized grains consist of well-sorted and well-rounded quartz pebbles. Load casts and ball and pillow structures are present

Fig. 15. Conglomeratic arkose in the Ring Formation resting with erosional base on the underlying sandstone. The contact is deformed by loading. From the Bechsminne section, Rena district.

in the basal part of the upper unit where these conglomeratic beds have sunk into the underlying sand (Fig. 15). Most of the conglomerate and sandstone beds display good sorting in relation to maximum grain size and have a sandy, almost mud-free matrix. The conglomeratic beds have a distinct base and usually a diffuse top, generally lacking internal structures, and wedge out laterally over a few metres.

Cross-bedding is usually present only in the finer-grained sandy beds wedging out between the conglomeratic beds. Shale flake conglomerates are common both as scattered flakes in conglomeratic beds and as concentrations in certain horizons due to local rip up of thin shaly beds. These shale flakes, approaching 50 cm length, are a further evidence of the local erosion of thin muddy layers which must have been cohesive and partly lithified, probably through exposure to the air.

The lower part of the Bechsminne section must have been deposited in a shallow marine or lacustrine environment. Episodes of coarse conglomeratic sediment supply have alternated with longer periods of more fine-grained clastic sedimentation. The shale beds represent a quite distinct facies from the conglomeratic beds and could not have been part of a fluvial cycle. The

Fig. 16. Load cast structures on the sole of one of the basal conglomeratic sandstone beds in the Ring Formation. Locality: Bruberg, 500 m east of Rena (80.35–28.60). The section is 3 × 4 m.

upper unit, however, appears to have been deposited in the central or proximal zone of an alluvial fan. Massive conglomerate beds without internal structures suggest flow in the higher flow regime, while cross-bedding in the sandstone beds is evidence of lower flow regime conditions during periods of reduced runoff or channel divergence. The development of regular upward-fining fluvial cycles was prevented by the high current energy and low content of finer particles. The direction of sediment transport from east to west corresponds predictably to flow perpendicular to the inferred basin margin, itself aligned north-south (Bjørlykke 1966).

There is no evidence in the lower part of the section of marine reworking of the sandstone beds, which are therefore believed to have been deposited directly by fluvial outwash streams into a shallow protected environment with little wave energy. The well-developed lamination in the shale points to a freshwater or brackish, rather than marine, environment.

Other sections in the Rena district. – Facies similar to those in the Bechsminne section are present in several shorter sections within the Rena district (Map sheet 1:50000, 1917 II).

The road sections southeast of the river Rena between Rena and Julussa expose the basal part of the Ring Formation, inverted and displaying abundant flute and load casts (Fig. 16) and shale flake conglomerates. Oscillation ripplemarks exist in the sandy shales underlying these conglomeratic sandstones, and they lack any evidence of subaerial exposure so they were most probably deposited in shallow water and shortly afterwards covered by alluvial fan debris.

In the river Skynna 2–3 km NW of Rena station the coarse (5–6 cm) conglomeratic facies of the Ring Formation is well exposed. These discontinuous beds can only be traced for a few tens of metres at the most. However, thin red sandy siltstone interbeds less than 1 m thick can be followed for about 1 km in the E–W direction and these are interpreted as sheet-flood deposits. These layers are at several places ripped up into angular fragments. Shale flakes (up to 100 × 30 × 25 cm) are incorporated in the overlying 5–10 m thick bed of conglomeratic arkose. Some shale flakes are also rounded and ball shaped.

The river Åsta sections. – The Ring Formation is exposed along several parts of the river Åsta and despite considerable facies variation the forma-

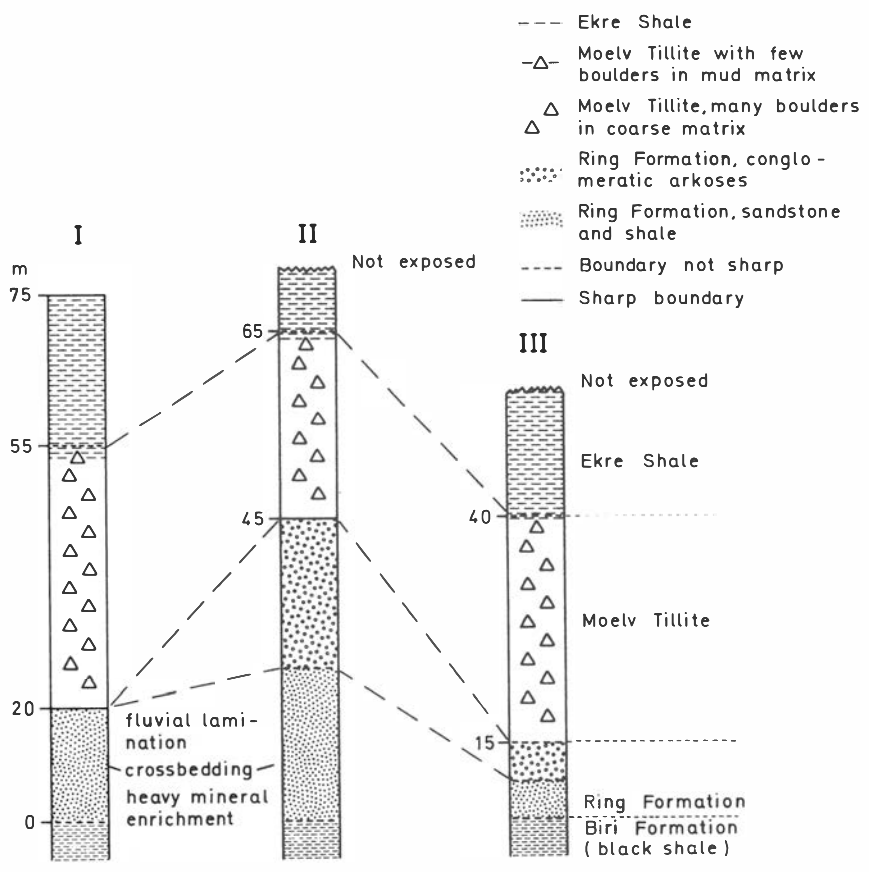

Fig. 17. Three schematic sections through the Ring Formation and the Moelv Tillite from the lower part of the river Åsta. Localities: I (74.85–23.40), II (74.65–24.80), III (74.40–26.10).

tion here is less coarse-grained than further east and north. The massive conglomerate beds are often absent and one has a cross-bedded, medium-grained sandstone with silty shale beds in between. In the lowermost 3 km of the river Åsta above its confluence with the river Glomma, there are several sections through parts of the Ring Formation (Fig. 17) showing trough cross-bedding, planar cross-bedding, and linguoid ripple marks. These sandstones are often dark due to heavy mineral enrichment. Other, discontinuous, sandstone beds of about 50 cm thickness cut red laminated shales.

The described deposits here are thought to comprise the more distal part of a fan, where the grain size is generally smaller and current energy lies within the lower flow regime. Trough cross-bedding, cut and fill structures characterise such distal zones, originating from braided stream activity across the toe of the fan, as described by McGowen & Groat (1971) from the Late Precambrian van Horn sandstone of western Texas. Heavy mineral concentration was also observed in the distal part of the van Horn fan.

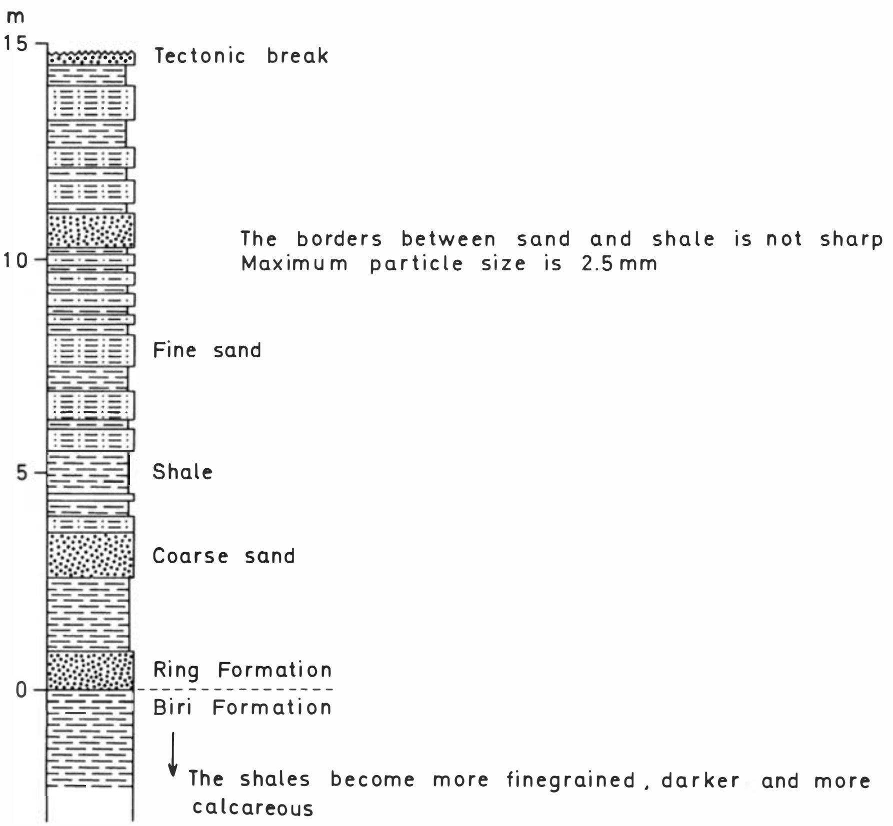

Fig. 18. Section through the basal part of the Ring Formation at Øyen sæter, Rena district (75.90–11.90).

Farther west at Øyen sæter (75.90–11.90) less than 1 km W of the map sheet boundary (sheets Åstmarka 1917 III/Rena 1917 II), a sequence of 0.1–1.0 m thick fine- to medium-grained sandstones alternating with shales is exposed (Fig. 18). They comprise a graded basal unit, a laminated unit, and in some beds an upper unit with micro-ripples, corresponding to a, b, c in a Bouma sequence (Bouma 1962) and probably produced by turbidity currents. Here one is beyond the alluvial fans, in a slope facies leading into the deeper part of the basin. It represents a transitional submarine (or sublacustrine) facies where only clayey sediments were deposited (Fig. 27).

Farther south at Nysætra (73.70–15.15) only fine-grained muddy sediments have been found below the Moelv Tillite, which suggest that the Ring Formation is absent here (Fig. 12). The sections are not continuously exposed and a tectonic squeezing out of some sandstone beds cannot be excluded, but the Moelv Tillite here is developed as an offshore drop stone facies conformably underlain by a few metres of shales (Fig. 27). This strongly suggests accumulation in a deeper part of the basin outside the fan deltas of the Ring Formation.

The Moelv and Biri district. – The Ring Formation in the Biri and Moelv districts is in many respects lithologically similar to that of the Rena district. However, palaeofacies reconstructions are more difficult because the fold axes lie parallel to the depositional strike (i.e. basin margin) in these two districts, and most of the distal facies of the Ring Formation would have lain above the present land surface (Elvsborg 1975).

The Moelv district. – In the Mjøsa section north of the pier at Moelv (56.15–91.35) the transition from the Ring Formation to the Moelv Tillite is tectonically repeated three times. A diffuse conformable boundary can be observed in two of these sequences while the third shows an erosional contact at the base of the tillite. A gradual basal transition to the underlying conglomeratic arkoses is also shown by the Moelv Tillite in the railway section south of Rena station (Vogt 1924) and in the road section at Tandeskogen 4 km east of Moelv.

The upper 50 metres of the Ring Formation exposed in the Mjøsa section reveal an unusually coarse facies with pebbles and cobbles up to 15 cm in diameter. These conglomeratic beds have diffuse boundaries with the adjacent coarse sandstones and are wedging laterally over a few tens of metres. Cut and fill structures are common and pebble imbrication indicates transport from the south. Well-defined cross-bedding is, however, very rare. In this section about 2 km north of the pier at Moelv (57.35–91.20), large shale flakes (200 × 100 × 25 cm) can be observed, interbedded in a poorly sorted conglomerate. Despite a very low mud content this conglomerate is poorly sorted and possibly originated as debris flow. The remaining poorly sorted, coarse-grained conglomerates and arkoses which lack cross-bedding were probably deposited by rapid, braided streams during floods.

The base of the Ring Formation is exposed in a road (E 6) section about 5 km north of Moelv (60.25–40.00) overlying black shales and carbonates of the Biri Formation. The lower part of the Ring Formation here consists of parallel laminated and graded silt- and sandstone beds of presumedly marine or lacustrine origin, deposited conformably above black Biri Shales in a regressive sequence. These sediments are overlain with a sharp boundary by massive 5–10 metre thick conglomeratic beds probably deposited by braided streams in a prograding alluvial fan.

The coarse-grained facies of the Ring Formation is exposed at many localities in the Moelv district (Vogt, 1954, Kirkhusmo 1968) with Lundehøgda, about 5 km north of Moelv, providing the best sections. Here the grain size of the conglomerates rarely exceeds 5 cm and crossbedded sandstones are more common and the trend towards smaller grain size continues as far north as Aursmoberget, 15 km north-east of Moelv (03.50–66.65). Although the absence of continuous sections makes detailed facies studies difficult, a northward decrease in grain size is observed, indicating a reduction in current energy from the higher to the lower regime.

The Biri district. – Close to the western shore of Mjøsa the Ring Formation is tectonically squeezed out (Vogt 1952, A. Bjørlykke 1973). As apparent from the map sheets Gjøvik (1916 I) and Dokka (1816 IV), A. Bjørlykke 1973a, 1973b and from Fig. 1, the Ring Formation can be seen to wedge out westwards, and at Svarken (18 km west of Biri) the Moelv Tillite is resting directly with a primary contact upon Biri Formation limestones (Fig. 26). The upper part of the Biri Formation which underlies the Moelv Tillite shows clear signs of weathering and erosion, including karst forms filled with calcarenite and quartz sand. In the Djupdalsbekken section about 10 km west of Biri (59.00–78.05), at least 30 m of conglomeratic arkose – with feldspar grains exceeding 10 mm – are found above the Biri Formation.

At Sagbekken about 5 km farther east (59.45–82.60), a Ring Formation basal sequence comprising 30 m of graded sandstone beds and shales is overlain by a thick conglomeratic arkose sequence. This transition from a marine/lacustrine facies to a braided river facies is similar to the Mjøsa section.

At Klomstein about 4 km south of Biri (57.40–83.40), the comparable basal unit consists of about 20 m of sand- and siltstones displaying micro cross-lamination and climbing ripples, together with a few thin conglomeratic beds. While the Moelv Tillite is conformably underlain by coarse conglomerates at Vestby (55.95–86.10), finer grain sizes prevail farther east at Klundby (55.60–89.00) near Mjøsa, suggesting that the coarse fan-delta sediments on either side of the lake were not deposited from the same apex but rather represent two coalescent fans (Fig. 21).

Petrology of the sandstones and conglomerates of the Ring Formation

The coarse conglomeratic facies of the Ring Formation is characterised by

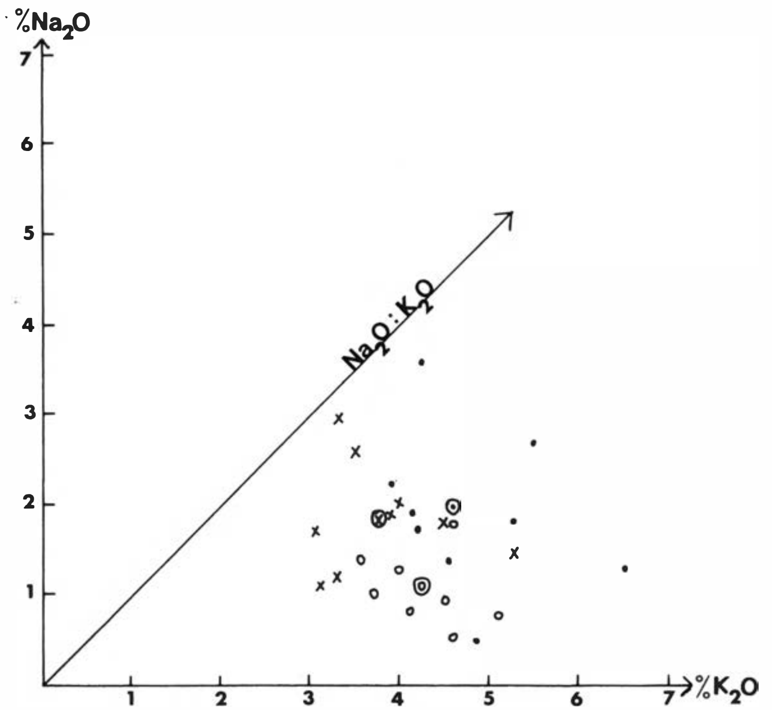

Fig. 19. Atomic absorption analyses of 26 samples of conglomeratic arkoses from the Ring Formation. Rena (x), Moelv (0) and Biri (.) districts. Average values encircled.

its high feldspar content (Fig. 20 A) whereas the fan's more distally deposited sandstones have a higher clay-matrix content (Fig. 20 B). The conglomeratic facies carry mainly quartz and quartzite pebbles while granite fragments are very rare (though single feldspar grains exceeding 1 mm are common). Microcline is the dominant feldspar in the Rena – Moelv district, which is consistent with the high K:Na ratios found in the bulk analyses of the sediments there (Fig. 19) which have also been reported earlier (Bjørlykke 1966, Kirkhusmo 1968). This composition reflects the overall granitic composition of the basement gneisses to the south and east. On the western side of the sparagmite basin, plagioclase gneisses in the basement gave rise to very high plagioclase contents in the arkoses of the Fåvang district (Englund 1966, 1972). Thus the composition of the arkoses of the Ring Formation reflects the local basement composition (Bjørlykke 1966: 37) and is not, as proposed by Barth (1938) and Strand (1951), related to secondary changes during metamorphism.

Ninety thin sections of the Ring Formation rocks from the Rena, Moelv, and Biri districts have recently been examined by one of the authors (Elvsborg 1975).

The matrix of the arkose of the Ring Formation arkose is composed mostly of sericite (illite) and microchlorite, though larger fragments of

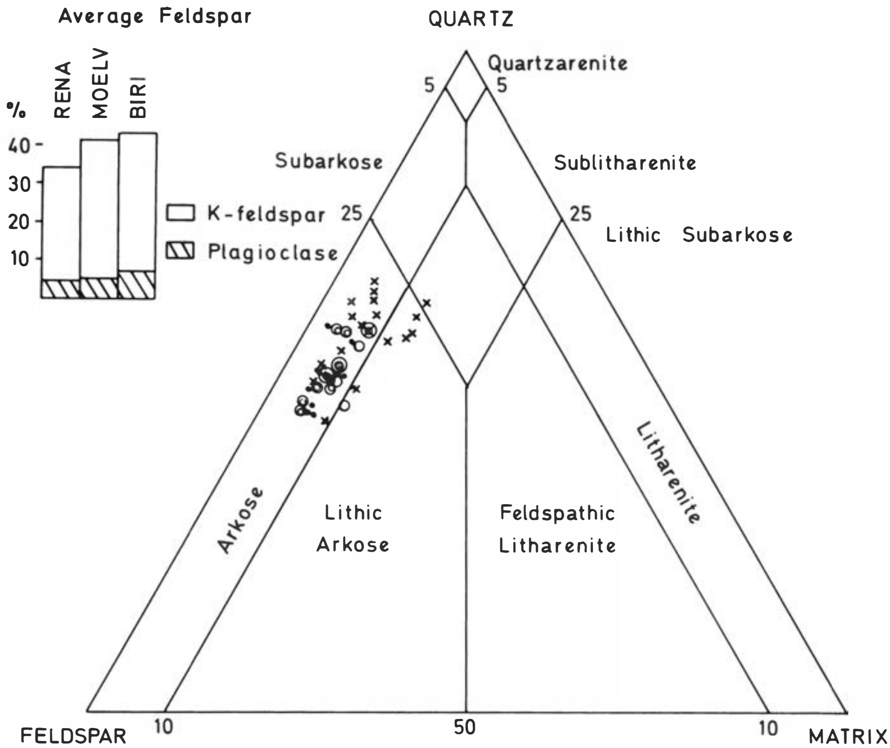

Fig. 20. A. Modal analyses of 36 coarse-grained samples of the Ring Formation from Rena (x), Moelv (0) and Biri districts (.). Average values encircled. Based on 500 points counted on polished and stained slabs.

clastic biotite and muscovite are present. In the lower part of the formation calcite cement is common and quartz, and more particularly feldspars, are frequently corroded or even totally replaced. This was earlier observed by Bjørlykke (1966) in the Rena district and by Englund (1972) in the Tretten – Øyer district.

In the upper part of the Ring Formation, nests of authigenic anatas are the main accessory minerals. Other (clastic) accessory minerals include, zircon, garnet, tourmaline, rutile, amphibole, and epidote. Ore microscopy of the opaque minerals identified limonite, hematite, pyrite, and chalcopyrite. The iron oxides and hydroxides are usually found as coatings on clastic grains, colouring these sandstones red. This effect can be caused by authigenic alteration of iron-bearing clastic grains (Glennie 1970, Ghose et al. 1970). Berner (1969) considers that primary limonite may dehydrate to hematite at or near the ground surface. Sulphides, often rimmed with iron oxides, are most abundant in the basal part of the Ring Formation transitional to the Biri Formation. In the fine-grained submarine slope facies of the Ring Formation at Øyen sæter, pyrites with polyframboidal textures are found. The pyrite is enriched in foreset laminae in the sandstone beds and may have a clastic

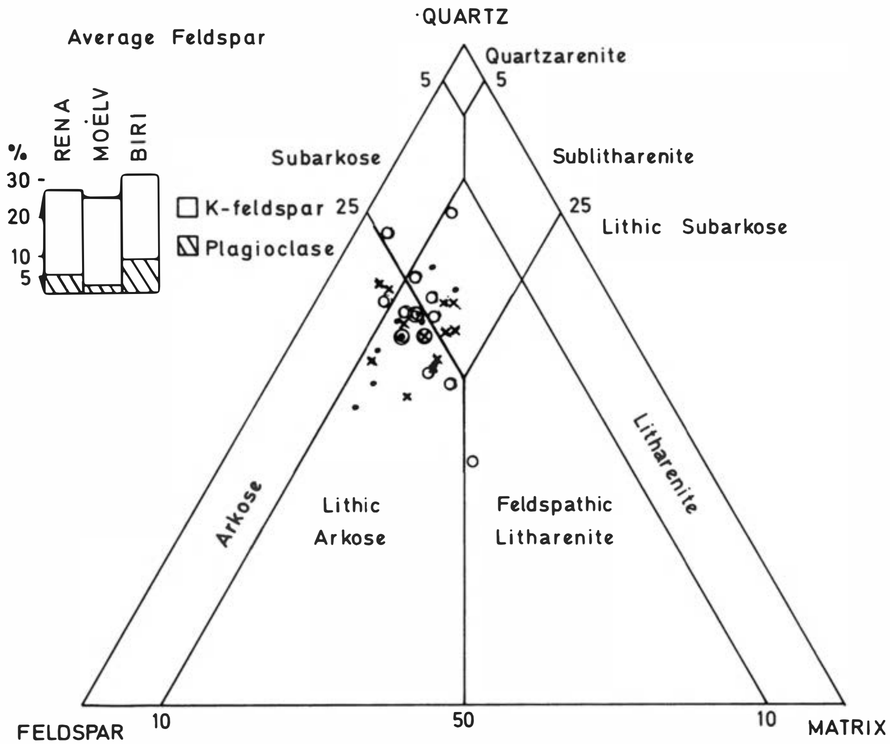

Fig. 20. B. Modal analyses of 28 sandstone samples from the lower part of the Ring Formation. Markings as 21A. Based on 500 points counted on thin sections.

origin, being derived from the underlying black mud. It is similar to what has been described from Silurian graywackes in north Wales (Love 1971).

The feldspars in the Ring Formation generally show very little evidence of pre-depositional weathering.

Depositional environment of the Ring Formation

The Ring Formation represents a regressive sequence from the marine sediments of the Biri Formation through shallow marine or lacustrine sediments in the basal part of the Formation and into the fluvial, braided river deposits of the prograding fan deltas (Fig. 21). A decrease distally in the maximum grain size in these fans can be well documented in the Rena district but can also be inferred from other areas including the Moelv district. Observations of primary sedimentary structures within these districts agree well with alluvial fan descriptions from other parts of the world (Sharp 1948, Blissenbach 1954, Beaty 1963, Drewes 1963, Bull 1964, Ruhe 1964, Bluck 1964, Hooke 1967).

The abrupt variations in grain size, sorting, and bed thickness, which are typical of the upper part of the Ring Formation, indicate transport by currents varying greatly in velocity and channel depth. This is highly character-

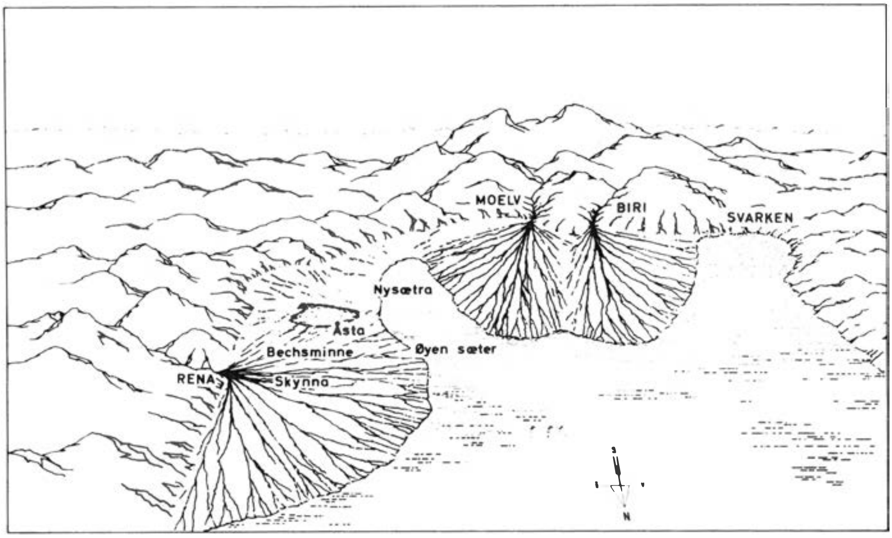

Fig. 21. Schematic reconstruction of the sparagmite basin during the deposition of the Ring Formation. View from north towards south. Rena, Moelv, and Biri are the names for the districts and not the specific places.

istic of modern alluvial fans too (Bull 1972). Cut and fill structures and shale flake conglomerates are typical of ephemeral streams (Picard & High 1973), and many of the rivers feeding the sparagmite basin during the deposition of the Ring Formation may have been seasonal. In particular the Skynna section at Rena displays shaly and silty beds that are continuous over considerable distances parellel to the direction of transport, but quickly wedge out perpendicular to the current direction.

Thin (1–2 cm) clay and siltstone beds which often separate the sandstone beds in the upper part of the Ring Formation are typical deposits of the waning phase of ephemeral flooding of an alluvial fan (Picard & High 1973). Similar patterns have been observed in modern alluvial fans from the Basin and Range Region in North America (Bull 1972). The Ring Formation sandstones are often red, partly attributable to a high percent of red microcline feldspar but mainly to grain coatings of hematite and lepidocrosite formed either during primary weathering or by post-depositional oxidation by circulating ground water (see p. 264).

In several sections the coarse fluvial conglomerates are separated from the underlying Biri Formation by fine-grained ripple-marked sandstones. There is no evidence of strong marine reworking forming cross-bedded sand bars, so these fluvial sediments were deposited in a sheltered environment with low wave and tidal energy. In some areas, as demonstrated by the Bechsminne section at Rena, the coarse conglomeratic facies finger into dark shale. Thus, near the basin margin, coarse fluvial material was discharged directly into the basin and was interbedded with the mudstones that accumulated during periods of fluvial quiescence. Fan progradation towards the deeper parts of the basin was then accompanied by renewed sand and mud transport further down the submarine slopes of the fans by turbidity currents.

Beyond the fan deltas, the Ring Formation does not exist, and the Moelv Tillite rests conformably on the Biri Formation.

The Moelv Tillite

The Moelv Tillite is a poorly sorted polymict conglomerate overlying the Ring Formation. In areas where the Ring Formation is missing the Moelv Tillite rests directly upon shales or limestones of the Biri Formation. It is always overlain by Ekre Shale. The Moelv Tillite is present in the stratigraphical succession of the upper part of the Hedmark Group throughout the central sparagmite basin of south Norway. The conglomerate varies in thickness from as much as 30 m to less than 1 m, but only one tillite horizon has been identified in sections with good stratigraphical and tectonic control.

The interpretation by Holtedahl (1922) of the Moelv Tillite as a glacial conglomerate has been generally accepted by Norwegian geologists. Subsequent investigations by Oftedahl (1945), Holmsen (1954), Spjeldnæs (1964), Bjørlykke (1966, 1974), Englund (1966, 1972, 1973a) and Løberg (1970) all supported or accepted a glacial origin. A summary of the arguments favouring this theory is presented by Bjørlykke (1966). Nelson (1963), however, questioned the glacial hypothesis, suggesting that as turbidity currents

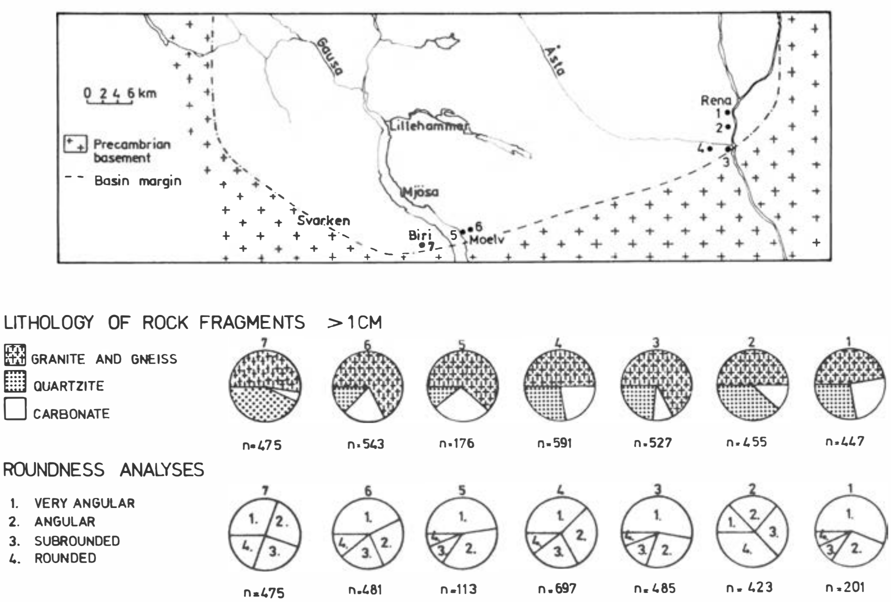

Fig. 22. Lithological composition of clasts in the Moelv Tillite based on counts at localities in the Rena, Moelv, and Biri districts.

were active during deposition of the Brøttum Formation similar processes could have been responsible for the Moelv Tillite.

In single outcrops it is difficult to find satisfactory criteria for distinguishing between mudflow deposits and glacial conglomerates. The strongest evidence for a glacial origin is the lateral and stratigraphical consistency of the Moelv Tillite as a single diamictite horizon (Bjørlykke 1966) which could not be readily accounted for by other processes.

Fine examples of striated pebbles from the Moelv Tillite (Bjørlykke 1974) have added detailed proof of glacial transport to the more circumstancial regional evidence.

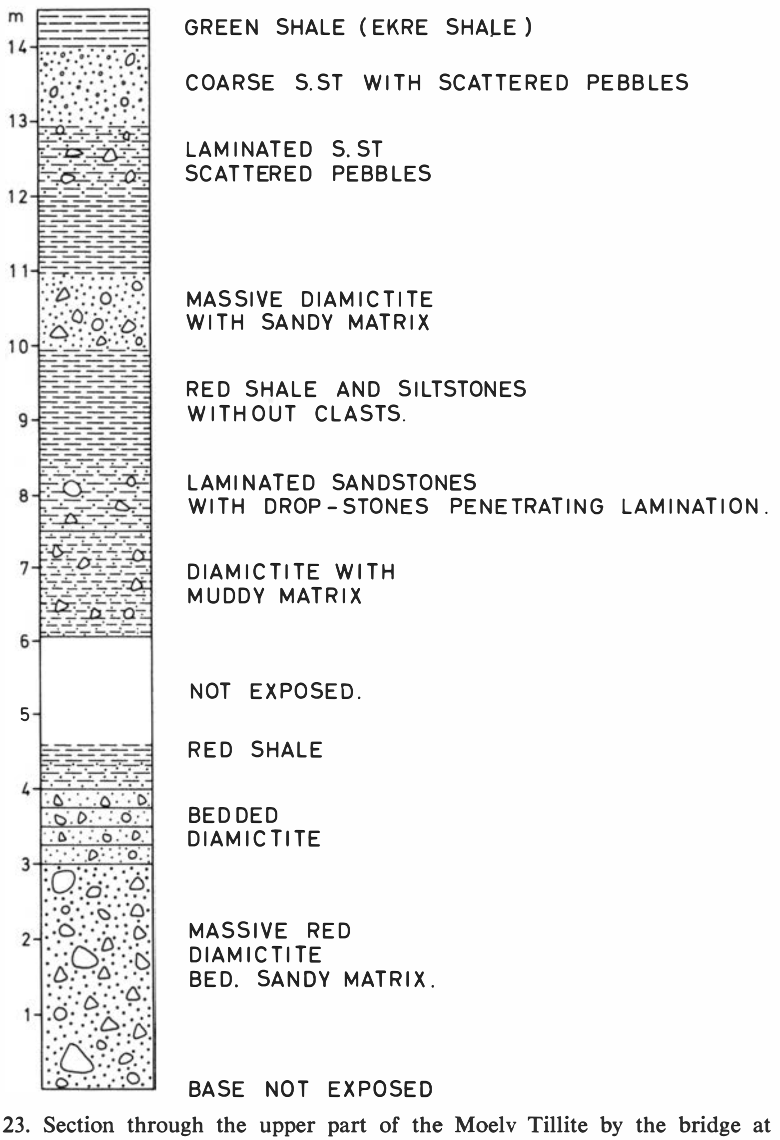

Fig. 23. Section through the upper part of the Moelv Tillite by the bridge at Åsta, Rena district (73.30–26.75).

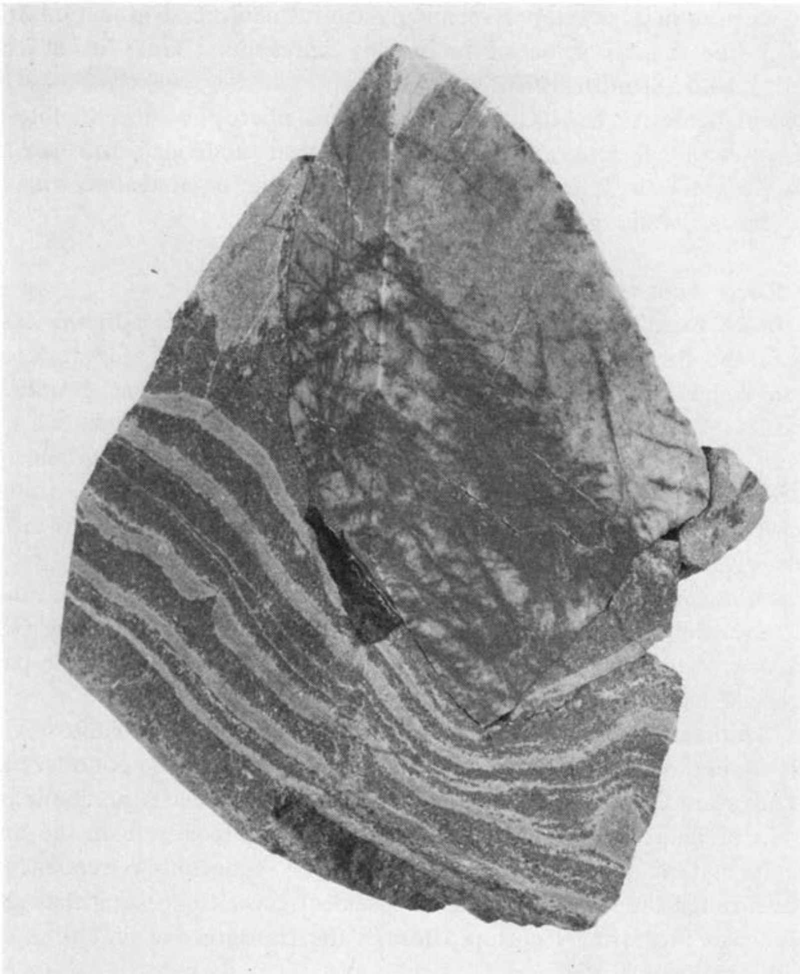

Fig. 24. Ice-dropped boulder deforming primary lamination in the Moelv Tillite from the section by the bridge at Åsta (Fig. 21). Largest diameter of the clast is 14 cm.
is 14 cm.

More recently A. Elvsborg and K. Bjørlykke have undertaken more detailed studies to ascertain the nature of the glacial environment during the tillite deposition. As part of the same project, J. P. Nystuen has examined the Moelv Tillite in the easern sparagmie basin and will discuss in more detail the glacial transport and paleoclimatic aspects of these sediments (J. P. Nystuen pers. comm.).

Since in most areas the Moelv Tillite appears to rest conformably on the Ring Formation without any obvious signs of erosion at the contact, a glaciomarine origin has been proposed (Bjørlykke 1966). However, the massive nature of the underlying Ring Formation would render an erosional unconformity difficult to detect, and it has been demonstrated recently that shelf ice moraines develop a seemingly conformable and gradational base through the incorporation of underlying sediments (Banks et al. 1971, Reading 1965, Spencer 1970).

Recent fieldwork has located some sections where the Moelv Tillite does have an erosional base, and can furthermore be subdivided into two main units – a massive tillite and a laminated sandstone or mudstone with scattered clasts (Elvsborg 1975).

The Moelv Tillite in the Biri district

The facies found in the Moelv – Rena district continues into the eastern part of the Biri district near Mjøsa, where the Moelv Tillite rests upon coarse conglomeratic arkose, but contains chiefly basement granite and gneiss. At Vestby (Fig. 26) the Moelv Tillite can be divided into a lower (10 m) massive diamictite unit with an arkosic matrix, overlain by a siltstone unit containing pebbles and boulders up to 60 cm, some of which clearly penetrate primary lamination. However, lamination is difficult to distinguish in the field.

The clast/matrix ratio in the upper unit is much lower than in the massive tillite and while the matrix of the laminated upper unit is silt or finer (< 0.1 mm) that of the lower unit is coarse arkose (1–4 mm). This lower part is interpreted as moraine, overlain by ice-rafted deposits.

At Svarken, further west in the Biri district, the Moelv Tillite rests directly on carbonates of the Biri Formation (Fig. 24). At the boundary karst structures are developed in the limestone. Limestone clasts predominate in the Moelv Tillite here and are presumedly derived locally from the underlying formation. However, these clasts display a declining frequency upwards through the section relative to basement clasts, suggesting that glacial erosion was increasingly cutting through the transgressive limestone cover into the basement series.

The Moelv Tillite in the Moelv – Rena district

In the Rena district the Moelv Tillite is typically a massive conglomerate about 15–20 m thick. The pebbles consist of Precambrian basement gneisses, quartzites, and limestones.

In some localities quartzite pebbles dominate over the gneisses and granites but usually the latter are most abundant. Carbonate pebbles, when present, rarely attain 30 % of the clast content.

The transition from the coarse-grained arkoses of the underlying Ring Formation is marked by a gradual increase upwards in pebble frequency through the basal 2–3 m, as observed in the Engåen section (77.85–27.45) south of Rena. While gravel-sized clasts in the Ring Formation consist almost exclusively of quartzite, granites and gneisses predominate in the Moelv Tillite (Fig. 22). In thin section an increase in unstable minerals such as biotite, chlorite well-crystallized light mica, clastic carbonates, and epidote can be observed in the matrix of the Moelv Tillite, compared to that of the

Fig. 25. Massive bed in the Moelv Tillite from the section by the bridge at Åsta (Fig. 23).

Ring Formation. At Moelv (Mjøsa section), however, there is an abrupt boundary. The massive, dark tillite with its generally angular granitic clasts is here underlain by a red Ring Formation gravel-pebble conglomerate with well-rounded quartzite pebbles.

Beside the bridge at Åsta (73.30–26.75) the base of the Moelv Tillite is not visible, but 15 m of the upper part of the tillite are exposed, including the contact with the overlying Ekre Shale (Fig. 23). Glacial dropstones penetrating primary lamination are also observed in this section (Fig. 24).

The most plausible interpretation of this section is that it represents a facies near the ice margin. The massive diamictite (Fig. 25) could then be a subglacially deposited tillite. The laminated diamictite represents deposition from a floating ice shelf or ice-bergs, the shale being a marine lacustrine clay accumulated during a period of glacial retreat.

Exposures of the Moelv Tillite along the timber road by Søttenbua (73.60–22.50) show interbedding of a well-sorted ½ m thick conglomerate with well-rounded pebbles and slumping structures.

At Kjellsmoen 7 km northwest of Moelv (61.55–97.40), a sand lens free of clasts has been observed in the Moelv Tillite. This may represent subglacial fluvial sedimentation. Conglomerates which could be interpreted as glaciofluvial equivalents to the Moelv Tillite have not been found in the main sparagmite basin, suggesting that the base of the ice was always submerged.

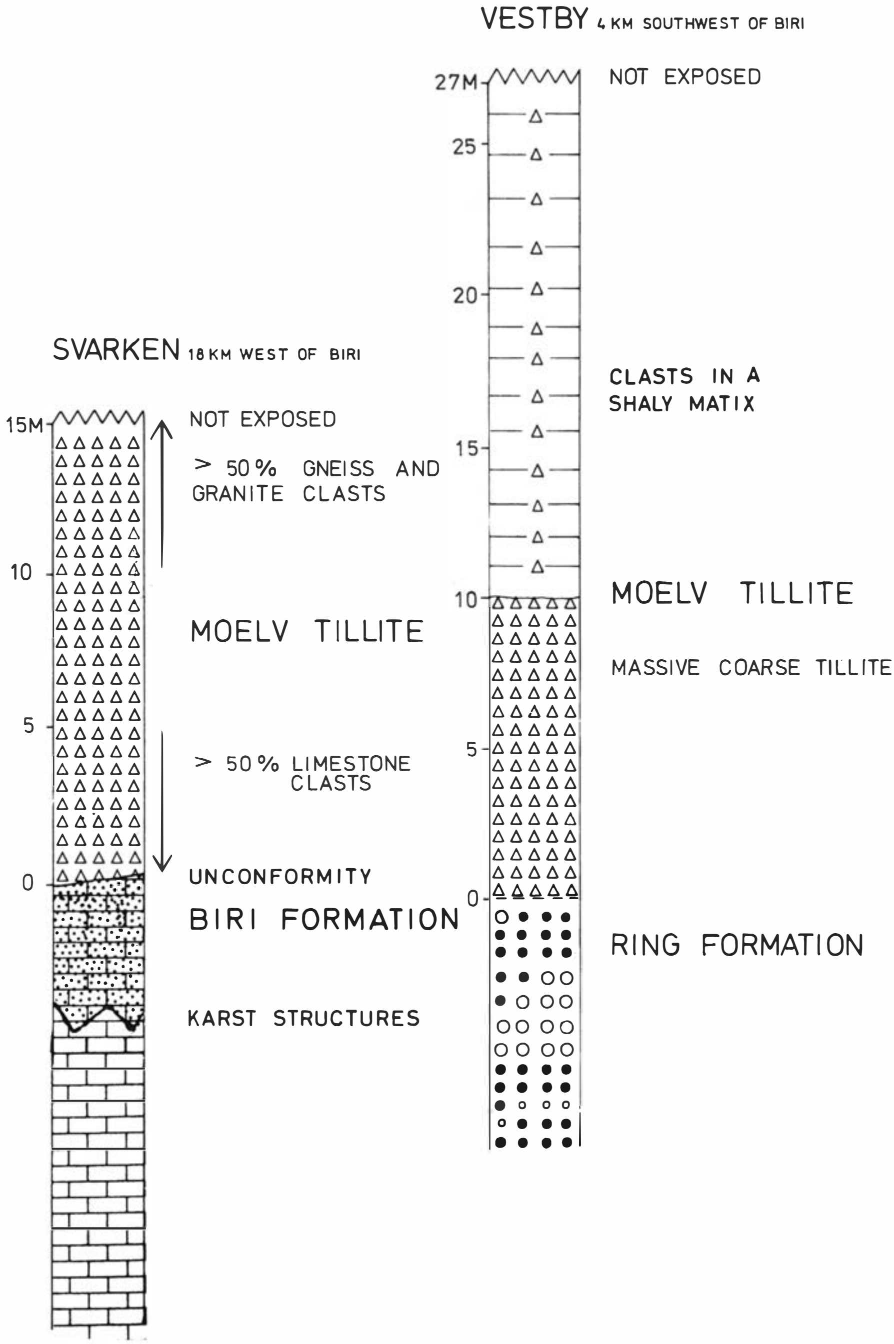

Fig. 26. Section through the basal part of the Moelv Tillite and underlying Biri Formation at Svarken (60.35–68.10) and Vestby (55.95–86.10), Biri District.

In the eastern sparagmite basin in Osdalen the Osdalen conglomerate (Holtedahl 1920) may represent a glaciofluvial conglomerate (Nystuen 1976).

In another section, by the lower reaches of the river Åsta (74.85–23.40), the Moelv Tillite rests apparently unconformably upon a cross-bedded sandstone that corresponds to the more distal zone of a fan delta (Fig. 27).

At Nysætra (73.70–15.15) the Moelv Tillite is only present as a thin (2 m) sequence of scattered pebbles in a laminated shale matrix (Fig. 9). Here one is outside the fan delta deposits of the Ring Formation and since this implies deeper water only a drop stone facies is expected, produced by floating ice (Fig. 27).

The transition from the Moelv Tillite to the Ekre Shale as exposed by the bridge at Åsta (Fig. 21) is conformable showing a gradual decrease in frequency of megaclasts passing gradually into a more fine grained matrix. This sequence must have been deposited during a glacial retreat sequence. There is no evidence of lag deposits showing that no regression took place during the deglaciation.

Depositional environment of the Moelv Tillite

As stated above the glacial nature of the Moelv Tillite has by now been demonstrated beyond reasonable doubt, but it is still important to establish the nature of the glacial environment. The Moelv Tillite can be divided into 'massive diamictite' facies most probably representing basal moraines or glacial ablation sediments, and a laminated part containing scattered pebbles which represents a drop-stone facies associated with floating ice. The clast composition of the Moelv Tillite is essentially derived from the local basement, exotic pebbles being rare. From this it is deduced that ice flowed concentrically into the basin (presumedly from encircling highlands), eroding into the basement and the transgressive limestones of the Biri Formation. At the onset of the glaciation the detailed morphology of the basin margin would have been determined by the distribution of the fan deltas of the Ring Formation.

The front of these fan deltas would approximate to sea (lake) level and here the ice would have been able to erode into the gravel and sand beds deposited immediately prior to the glaciation. Those parts of the basin margin which did not subside during Ring Formation time and thus were not covered by standstones – such as at Svarken (Biri district) – were also eroded; here carbonate clasts predominate among the Moelv Tillite pebbles

Outside the shallow shelf areas and away from the delta complexes, somewhat deeper conditions prevailed, where the ice would have floated and produced an ice-drop facies (Fig. 27). Examples of this facies are found at Nysætra and Bjørånes.

The absence of clasts from underlying sparagmite formations other than the Biri Limestone suggests that these other formations (e.g. Brøttum Formation) were either not deposited along those parts of the basin margin that suffered glacial erosion or were insufficiently indurated to form pebbles.

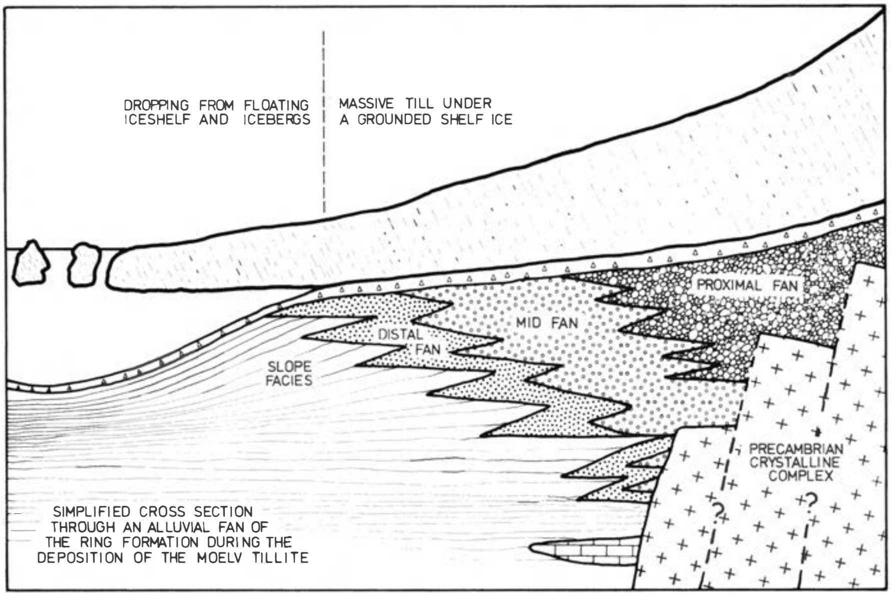

Fig. 27. Diagrammatic section through a fan delta of the Ring Formation during the deposition of the Moelv Tillite.

There is little evidence of a major regression in the sparagmite basin during the Moelv glaciation; indeed in several areas the Moelv Tillite is found in a transgressive position relative to the basement (Holmsen 1954). Since the Moelv Tillite is now considered partly to be ground moraine, particularly in the marginal parts, it may be expected to show a transgressive relationship (Spencer 1975: 232) and may theoretically have been deposited at considerable elevation above sea level, as is the case for Quarternary moraines. The fact that the Moelv Tillite is not eroded and uniformly overlaid by the Ekre Shale and Vangsås Formation, however, suggests very strongly that the Moelv Tillite was deposited below or close to the sea level. The glaciation may, however, have started earlier in other regions and the Ring Formation may have received glaciofluvial material from the higher ground around the basin. The onset of the glaciation in other areas must have led to a regression during the deposition of the Ring Formation.

If the Moelv Tillite was deposited during an extensive world-wide synchronous glaciation it would have been accompanied by a marked eustatic regression. The transgressive position of the Moelv Tillites implies that if the sparagmite basin at that time was marine, ice loading was sufficient to cause an isostatic downwarp, which more than compensated for the eustatic displacement. However, local tectonic subsidence other than ice loading could have contributed to this, and the sparagmite basin may have been a fresh water basin with a water level unrelated to sea level.

The rather limited thickness of the Moelv Tillite in most localities (0.5 – 20.0 m) and the presence of only one tillite horizon indicate that there was only one major glacial advance here, and that of relatively short duration. However, on the basis of a few sections (Fig. 23) this glaciation can be divided into phases of glacial advance and deposition separated by periods of floating ice sedimentation or normal clay sedimentation.

The interbedded, well-sorted conglomeratic sandstones in the Moelv Tillite near Søttenbua (Rena) and at Kjellsmoen (Moelv) may represent subglacial fluvial sedimentation. This is suggestive of deposition under a wetbased glacier following the model described by Carey & Ahmad 1960. But conglomerates that could be interpreted as glaciofluvial equivalents of the Moelv Tillite are very rare in the central sparagmite basin. In the eastern basin the Osdal Conglomerate (Holtedahl 1920) most probably represents a thick glaciofluvial conglomerate (J. P. Nystuen pers. comm.).

The modest thickness of the Moelv Tillite and the absence of a tillite corresponding to the lower (Smalfjord) tillite in Finnmark suggests that south Norway was a marginal area for ice accumulation. Palaeogeographical reconstructions give Norway a near equatorial position for this period (Harland & Bidgood 1959, Harland 1964). Recent palaeomagnetic studies of late Precambrian tillites from Scotland also support an equatorial position for northwest Europe (Tarling 1974).

Late Precambrian tillites have been reported from all the continents except Antarctica (Harland 1964, Schermerhorn 1974). Harland and also Girdler (1964) have interpreted this as implying that the glacial sediments were deposited during synchronous and world-wide rapid changes in world climate. Schermerhorn, however, maintains that many of the reported tillites may not be of glacial origin and regards the evidence inconclusive for providing that the late Precambrian glaciation covered large parts of the world. In the present authors' opinion, Schermerhorn fails to demonstrate convincingly alternative origins of the large number of late Precambrian tillites (or tilloids) discussed in his paper. The Moelv Tillite, however, has many features that can be directly attributed to glacial processes, including glacial striation and drop-stone deformation of primary lamination, and we cannot envisage non-glacial processes being responsible for the deposition of the Moelv Tillite.

It seems natural to assume that the glaciation responsible for the Moelv Tillite emanated from higher ground surrounding the basin. Furthermore it is likely that the fault movements that occurred during the deposition of the Ring Formation produced a relief of horsts and basement escarpments. Even moderate elevations of a few hundred metres would have significantly influenced ice accumulation and flow. If the south of Norway had been covered by a major ice sheet at this time, one would have expected a nearly uniform ice-flow direction across the basin. While the local ice flow deduced for the Moelv glaciation from facies and clast distributions could perhaps be attributed to the closing stages of a more extensive glaciation, there is no evidence to suggest that ice has overridden the entire sparagmite basin at any stage. There is, in the authors' opinion, good evidence that the south of Norway was in a climatically marginal position for ice accumulation and that reconstructions involving large ice sheets with radiating ice flow, as discussed by Spencer (1975) and Chumakow (1968), are not valid for south Norway. Evidence of converging ice flow as found in the sparagmite basin is a good indicator of a restricted mountain glaciation.

The transition from the Moelv Tillite into the Ekre Shale is always conformable within the central part of the sparagmite basin. Neither aggradation nor lag-deposit formation through winnowing by waves or currents are observed in the upper part of the tillite. This points to a transgression accompanying the deglaciation. If the transgression at the base of the tillite (i.e. during the onset of glaciation) was caused by ice-induced crustal downwarping, then a local regression would have been expected at the end of the glaciation. However, erosion around the basin and possible long tectonic crustal subsidence can also account for this transgression, which continues during the deposition of the Ekre Shale and Vangsås Formation. Also if the sparagmite basin was a fresh-water basin a transgression may have been caused by local continued subsidence and/or elevation of the drainage threshold.

The Ekre Shale

The Ekre Shale (Vogt 1924) is a green or red shale, usually about 30–40 m thick. However, in the Tretten – Øyer district its thickness is diminished to only 6–12 m (Englund 1972) while in the eastern sparagmite basin north of Ossjøen it may attain 100 m (J. P. Nystuen, pers. comm.). Varve-like laminations are common (Holtedahl 1953, Skjeseth 1963, Løberg 1970) but are not present in every section.

The Ekre Shale is closely associated with glacial deposits (the Moelv Tillite) and Holtedahl (1953) mentioned the possibility that it could represent a glaciomarine sequence similar to some Quaternary examples. However, the presence of laminated shales, as pointed out by Løberg (1970), indicates brackish or freshwater conditions. This agrees with the low boron values obtained from this formation (J. O. Englund, pers. comm.).

The Vangsås Formation

The Vangsås Formation, which is the uppermost formation of the Hedmark Group (Bjørlykke et al. 1967), has been named the Quartz Sandstone Formation by earlier authors (Schiøtz 1902, Holtedahl 1953, 1960, Vogt 1924). Stratigraphically equivalent formations in Sweden include Strøms Quartzite and the Vemdal Sandstone in the Swedish Caledonides (Kulling 1972), and the name Vemdal Formation has also been used as a stratigraphic name for this formation in Norway (Englund 1966, Bjørlykke 1966). Vogt (1924) divided this formation into two members, the Vardal Sandstone (Sparagmite), which contains considerable amounts of feldspar (10–25 %

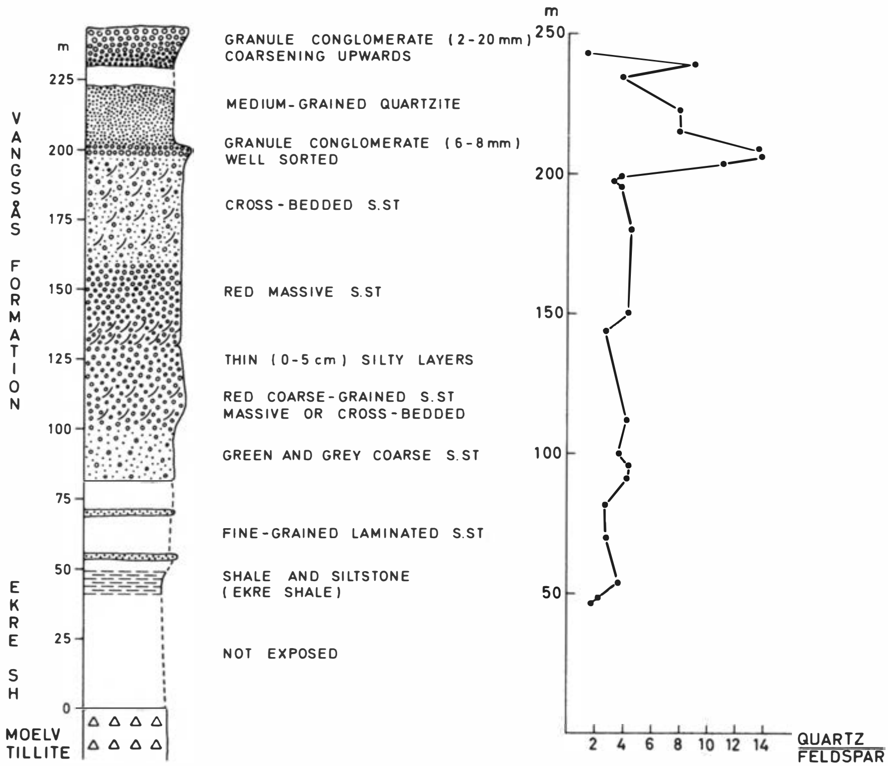

Fig. 28. Section through the Vangsås Formation along the main road (riksveg 3) south of Engåen, 2 km south of Rena station (77.90–27.45). Feldspar content is indicated to the right.

or more) and the Ringsaker Quartzite, which is a quartzite with very low feldspar and matrix content.

The Ringsaker Quartzite is succeeded by a granule conglomerate which varies in thickness from 0.1 to 3.0 m. This conglomerate is in turn overlain by a silty sandstone which grades upwards into a sandy shale yielding Lower Cambrian fossils (Holmia Shale, Kiær 1916). The transition between the Vangsås Formation and the shales of the Holmia series has been described in several sections by Vogt (1924) and Skjeseth (1963).

The Vangsås Formation is found both in a para-autochthonous position on top of the Ekre Shale, and in an allochthonous position (Osen Nappe) thrusted above Cambrian shales. Its thickness at Rena is about 160–130 m (Fig. 28) in the Engåen section (Rena district). Continuous sections through this formation are rare, but along the road section (riksveg 3) at Engåen south of Rena (77.90–77.45) the sequence is almost complete. The transition from the Ekre Shale to the Vangsås Formation is marked by sandstone beds with parallel lamination, interbedded with silty shale. The Ekre Shale here is normally a coarse-grained and silty shale, often containing beds of fine sand. It is therefore difficult to draw a sharp boundary between the Ekre Shale and the Vangsås Formation. The basal 30–40 metres of the Vangsås Formation consist of green to grey sandstone with thin granule beds. Above this unit we find an abrupt facies change. Here cross-bedded red sandstones occur, separated by very thin silty layers (0–5 cm). Beds with indistinct cross-bedding and massive beds are also common. This facies continues upwards to about 130 m above the base of the Vangsås Formation where a well-sorted granule conglomerate (6–8 mm) marks the transition into the overlying medium- to fine-grained reddish quartzite. Towards the top of this formation the quartzite becomes more grey or white. The topmost part of the Vangsås Formation consists of a granule conglomerate (2–20 mm) which is coarsening upwards. This exposure is terminated here by a fault following the stream of Engåen. Grey shales along this disturbed zone may be of lower Cambrian age, but no fossils have been found (Bjørlykke 1966). North of the fault Ring Formation shales and sandstones are exposed.

Interpretation of the Engåen Section

The transition between the Ekre Shale and the overlying Vangsås Formation is developed as an interfingering between the shaly and silty facies of the Ekre Shale and the sandstones of the Vangsås Formation (Fig. 28). The basal sandstones of the Vangsås Formation were described above. These beds have most probably been deposited by deltaic progradation into a marine or a lacustrine environment. There is no evidence of any major development of larger sand bars which might correspond to delta-front sand bars or river-mouth bars. This suggests that the marine (laqustrine) reworking of the deltaic sediments was small. Channels are also absent in these rocks, indicating that due to low tidal energy tidal channels failed to develop. This agrees well with the palaeographical reconstruction of the sparagmite basin as an enclosed basin inhibiting strong wave energy. The low tidal energy would suggest that the basin was in a region of small tidal ranges or that it was wholly or partly a separate basin with fresh or brackish water. Above this transitional zone we find 160 m of continuously exposed sandstone without any shale beds, mostly of presumed fluvial origin. The basal 20 m which is green to grey and lacks cross-bedding may still represent a deltaic environment, while the red cross-bedded sandstones above most probably represent a braided river system. This part of the Vangsås Formation has been deposited by braided rivers with very small suspended loads. Extremely thin silty layers found on top of some sandstone beds may represent fall-out from suspension between periods of flooding (Bull 1972).

The red colour in this sandstone is consistent with a continental environment. The colour is due to hematite staining formed by the circulation of oxidizing ground water. At 45 metres from the top of this formation a well-sorted granule conglomerate occurs which may represent a beach conglomerate indicative of a marine transgression. The overlying quartzite is fine- to medium-grained sandstone with very little feldspar. Low angle cross-bedding is very common in Ringsaker Quartzite (Fig. 29). The upper part of this 45 m thick unit is coarsening upwards into a granule conglo-

Fig. 29. Low-angle cross-bedding in the Ringsaker Quartzite. River section by Glomstad, Rena district (74.10–27.80).

merate, which again probably represents a beach environment. This sandstone with its granule component is practically devoid of clay fraction particles and has a negative skewness – also corresponding to beach environment.

Modal analyses (Fig. 28) of samples from the Engåen section show a sharp increase in quartz/feldspar ratio at the boundary between the Vardal Sparagmite and the Ringsaker Quartzite. This reflects the increased maturity of a shallow marine sandstone compared to the fluvial one. The topmost beds, however, contain granule conglomerates with a high content of coarse feldspar fragments. This suggests that when the top of the Ringsaker Quartzite was deposited, basement sources were exposed nearby, supplying unweathered granitic debris. This conglomerate may be regarded as the transgressional conglomerate at the base of the Lower Cambrian succession.

Regional facies variation in the Vangsås Formation

Compared to most of the other formations in the Hedmark Group the Vangsås Formation shows relatively moderate lateral facies variations within the sparagmite basin. However, despite an apparently uniform lithology, important changes can be observed in the different areas, though no systematic survey has been carried out to map quantitatively facies variations within the formation.

In the Vardal Sparagmite Member of the Vangsås Formation conglomeratic beds (by some authors referred to as the Vardal Conglomerate) are found in several areas, notably at Biri and also in the eastern sparagmite basin (J. P. Nystuen 1967).

Conglomeratic horizons in the Vardal Sparagmite Member can be studied at Biri in the river Bråstad sections (2 km north of Gjørvik) and along highway 4, 3 km north of Redalen (55.15 – 90.15). Here typical cyclic fining upwards can be observed grading from granule conglomerates to siltstones. From the same map sheet (Gjøvik 1816), A. Bjørlykke (pers. comm.) has observed phosphorite bed and phosphorite pebbles up to 50 cm in the Vangsås Formation.

In some areas the Vardal Sparagmite Member is developed as a coarse conglomeratic arkose (5–10 mm) similar to that of the Ring Formation and may be either green or red.

The conglomeratic facies in the Vangsås Formation is limited to the southernmost part of this region and is found in the lowest tectonic units (shortest tectonic transport) in the imbricated tectonic structure of the Osen Nappe (quartz sandstone nappe). This represents a marginal or proximal facies along the southern margin of the sparagmite basin. Further north a facies of uniformly cross-bedded sandstones is found without conglomeratic beds.

A well-exposed continuous section through the Vardal Sparagmite is found by highway 4, north of Redalen (55.15–90.15). Here tabular cross-bedding with angular contact between foreset laminae and the botton set are found (Fig. 30). The sandstone beds which may be 30–60 cm in thickness are separated with very thin (5 cm) layers of silt and fine sands. This suggests transport in a shallow stream with small amounts of suspended sediments.

In the Tretten – Øyer district only parts (about 50 m) of the Vangsås Formation are exposed (Englund 1972) and further to the north in the Ringebu – Vinstra district the thickness of the same formation varies between 70–140 m (Englund 1973). A lower feltspathic sandstone member and an upper quartzite member can be recognised here too.

In the Atna – Bjørånes district the Vardal Sparagmite is developed as a coarse red conglomeratic sandstone with a high feldspar content corresponding to arkose (Bjørlykke 1965). The Ringsaker member here is an orthoquartzite similar to further south. This part of the sparagmite district has received little coarse-grained clastic sediment during the earlier history of the sparagmite basin. The fine-grained facies represented by the Bjørånes shale is probably due to the horst between the eastern and central sparagmite basins trapping sediments before they could reach the central basin. During the deposition of the Vangsås Formation this horst was no longer active and coarse fluvial sediments were fed into the area.

In the topmost part of the Ringsaker Quartzite burrows of the *Scolithus* and *Diplocraterion* type have been found (Skjeseth 1963). This facies is, however, restricted to the southernmost outcrops of the Vangsås Formation. At Langodden, south of Brumunddal, abundant burrows are found (Skjeseth 1963) in a sequence of low-angle cross-bedded quartzites. A phosphorite

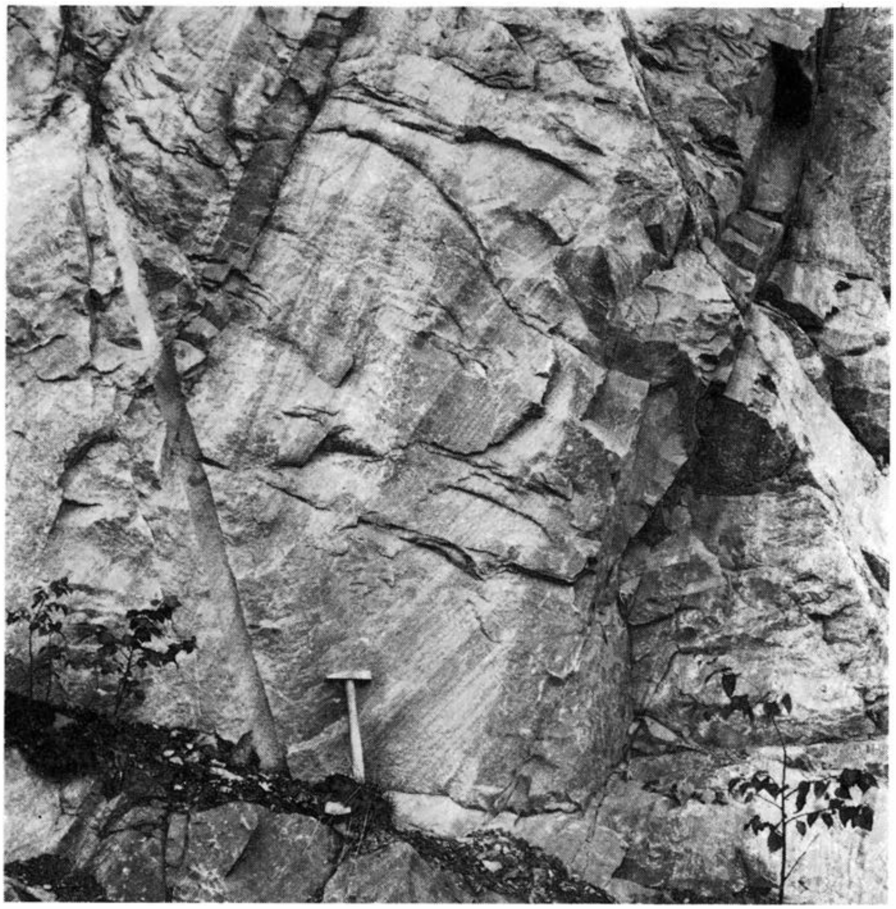

Fig. 30. Tabular cross-bedding in the Vangsås Formation. Road section 3 km north of Redalen (55.15–90.15).

conglomerate occurs at the top of the quartzite marking the transition to the Lower Cambrian Holmia series.

Lead mineralisation in the upper part of the Vangsås Formation

Galena mineralisation has been reported in the Vangsås Formation from several areas in south Norway, in most cases stratigraphically in the uppermost part near the top of the Ringsaker Quartzite. Similar disseminated lead mineralisations are mined in Sweden at Vassbo and Laisvall. In Norway, economically interesting occurrences have been discovered in the Biri district (Snertingdal), at Galå (15 km south of Rena) and at Osen (25 km northeast of Rena). In the latter locality, however, a lower Cambrian sandstone is mineralised. The galena occupies the interstices between the quartz grains in the pure orthoquartzitic facies of the Ringsaker Quartzite; sometimes it has partly replaced the quartz grains and is also found along fracture planes due to later tectonic mobilisation.

The lead mineralisation in the Vangsås Formation has been investigated by Norges Geologiske Undersøkelse, but has so far not proved worth mining (A. Bjørlykke et al. 1973). A. Bjørlykke (pers. comm.) considers this mineralisation to be related to primary weathering of the Precambrian basement. Pb released during the weathering of K-feldspar circulated in the ground water and was trapped in the Ringsaker Quartzite on entering the marine basin.

Depositional environment of the Vangsås Formation
The Vangsås Formation marks the end of a period of subsidence in the sparagmite basin. The increase in maturity of the sediments from the arkoses or feltspathic sandstones in the lower part of the formation to the orthoquartzite (Ringsaker member) at the top is a response to the reduced relief. In the Vardal Sparagmite, red (or green) cross-bedded sandstones most probably represent a fluvial environment. Continuous shale beds are rare or absent within this member, except in the lower part which is transitional to the Ekre Shale. The Ringsaker Quartzite was deposited in a shallow marine beach environment. Thus the Vangsås Formation represents a transgressional sequence. The unconformity between the Vangsås Formation and the grey and green shales of the Holmia series represents a break in sedimentation followed by a transgression with the deposition of a conglomerate.

Summary of late Precambrian (Eocambrian) sedimentation in the sparagmite basin

Sedimentation in the late Precambrian sparagmite basins can be regarded as a response to basement tectonics – mainly faulting – within the area (Fig. 31). These basins covered only relatively small parts of the Fennoscandian shield, so potential detritus source areas were large compared with the areas of deposition. During periods with high relief, caused by block faulting and basin subsidence along marginal faults, sedimentation rates were probably high, though their calculation is prevented by the lack of an absolute time scale for these events.

The source areas surrounding the basins consisted almost exclusively of crystalline metamorphic or igneous rocks. Periods characterised by mainly mechanical weathering combined with fluvial and glacial erosion produced sediments possessing a very high sand:clay ratio. Such deposits are observed in several formations in the Hedmark Group, particularly the Ring Formation but also in the Vangsås and parts of the Brøttum Formation. Large parts of these three formations are attributed to the environment of prograding fan deltas (Fig. 32). On entering the basins these fan deltas apparently escaped significant marine reworking for which their low mud content would have rendered them vulnerable. Indeed the tidal channels found in the Biri Limestone provide the only firm evidence of a marine environment in the whole sequence. No similar indications of tidal activity have been discovered in any of the other formations. Furthermore boron analyses show significantly higher values in the Biri Formation shales than in the other formations (J. O. Englund pers. comm.).

Sedimentological reasoning and geochemical evidence both suggest fresh or brackish water prevailing during the deposition of most of the Hedmark

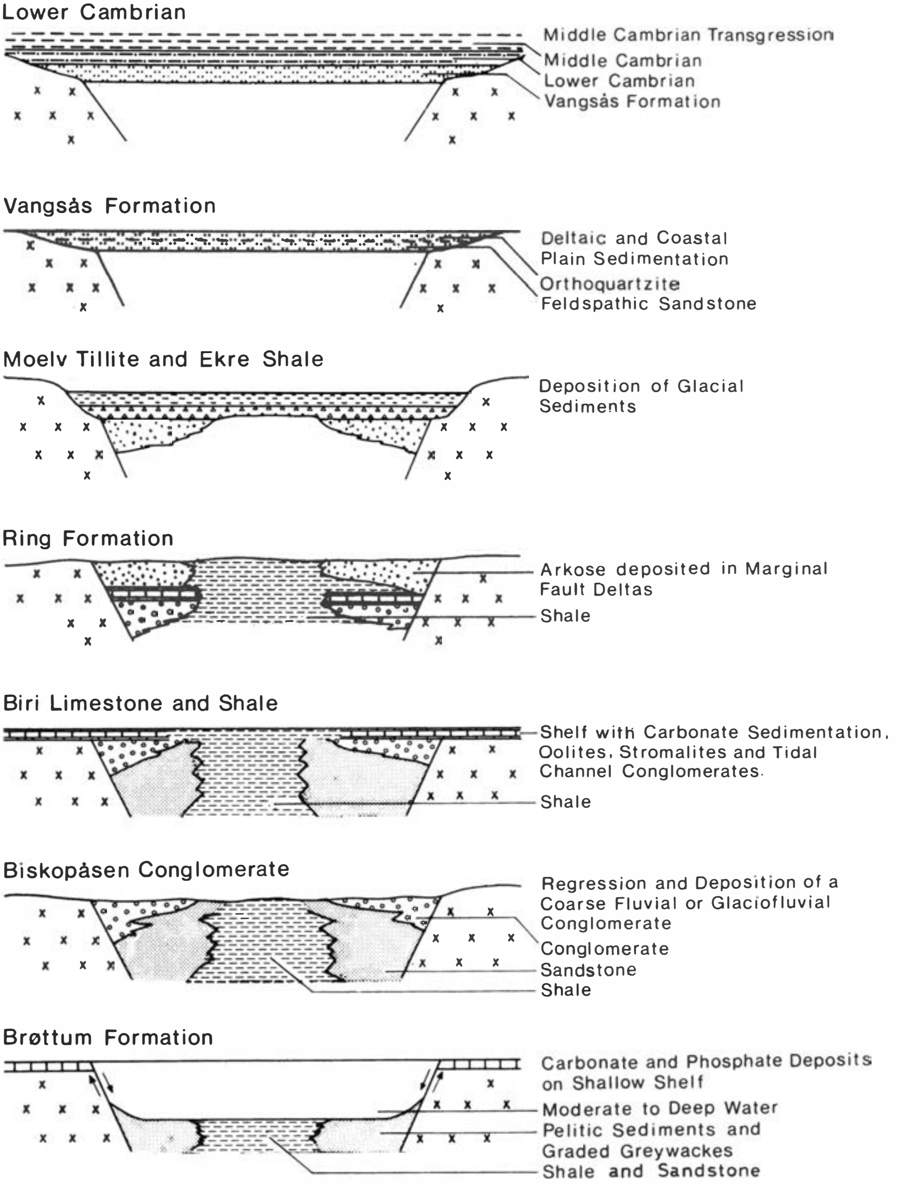

Fig. 31. Simplified diagram which illustrates the history of deposition and distribution of sedimentary facies in the central sparagmite basin in south Norway. The basin is about 60–70 km across (east-west) and 2–4 km deep.

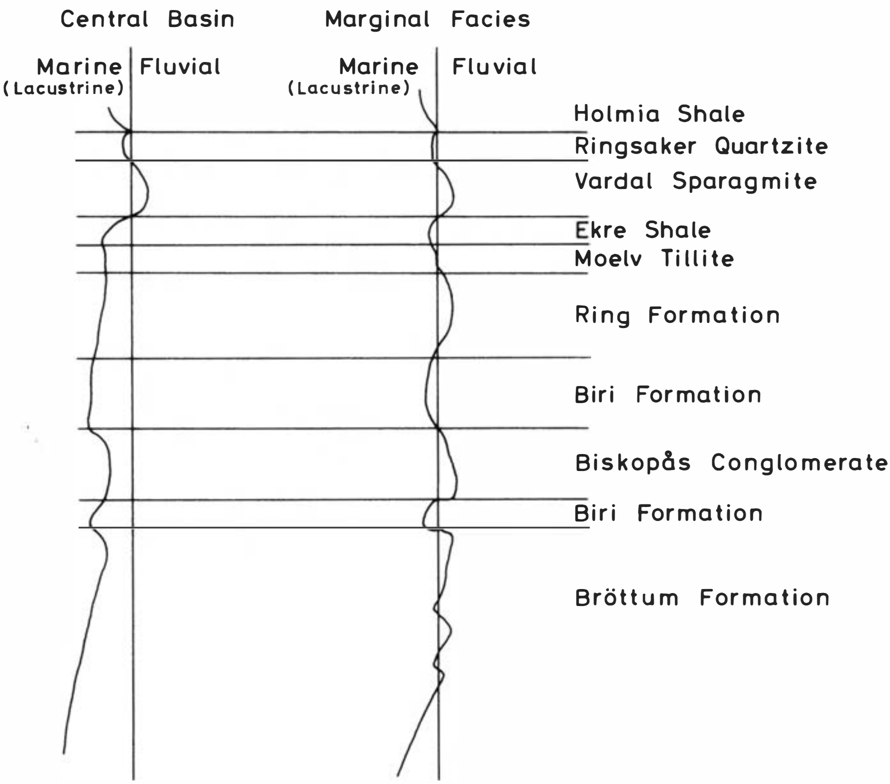

Fig. 32. Interpretation of periods with fluvial and marine (brackish, lacustrine) sedimentation, comparing an area near the basin margin with the central part of the basin

Group (Fig. 32). This agrees well with palaeogeographical reconstructions that suggest a rather narrow basin opening to the north-west (Englund 1972: 24). There is however, no proof that such a passage existed continuously. As a cratonic basin, the sparagmite basin may have been isolated at least intermittently, though the lack of evaporites does suggest uninterrupted drainage to the open sea.

The Biri Formation was deposited during a period of regional transgression and a shallow sea may have extended over large areas of the craton around the basins and thus provided open contact with the ocean.

The Moelv Tillite was probably deposited by local mountain glacies flowing into the basin depositing the conglomerate partly as a moraine, partly as a drop-stone facies. The Moelv Tillite can be correlated with the upper Tillite in Finnmark (north Norway). There is no direct evidence of an earlier glaciation in south Norway as in Finnmark, but the Biskopås Conglomerate may represent reworked glaciofluvial conglomerate formed by glaciers that did not reach the basin.

The Ringsaker Quartzite is interpreted as a transgressive marine deposit that partially reworked the fluvial sediments of the Vardal sparagmite. Low-angle cross-bedding and the presence of burrows, such as *Scolithus* and *Diplocraterion,* are good indicators of shallow marine conditions during the deposition of this thin transgressive sand body (Ringsaker Quartzite). The thin granule conglomerate overlying the quartzite and marking an unconformity with the overlying Holmia series, most probably represents a regression followed by renewed transgression.

Comparison with late Precambrian sediments in other parts of the Baltic shield

The Baltic Shield has probably been mostly emerged in latest Precambrian times. Sedimentation took place in well-defined basins in the craton such as the sparagmite basins (Fig. 33). Fan deltas along the basin margin indicate that the surrounding parts of the shield were land.

Latest Precambrian sediments in Finmark, north Norway (Siedlecka &

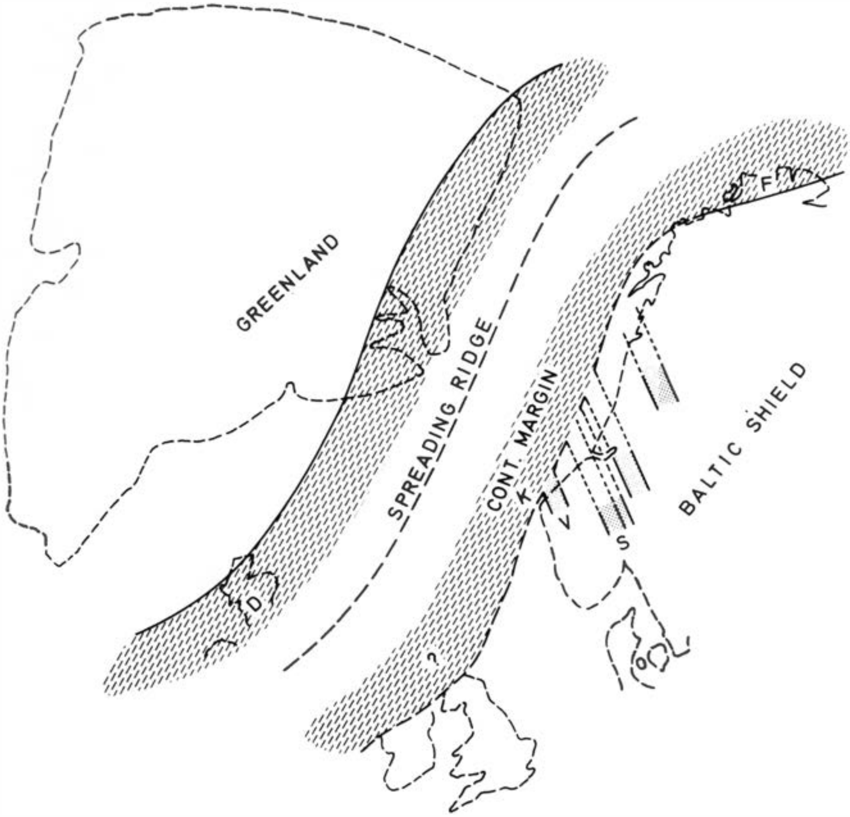

Fig. 33. Hypothetical reconstruction of the north Atlantic Caledonides in late Precambrian times. The sparagmite basin is interpreted as one of several cratonic basins related to the opening of the proto-atlantic ocean. V = Valdres Sparagmite, F = Finnmark, K = Kvitvole Nappe facies, D = Dalradian.

Siedlecki 1972, Siedlecka 1975), represent a very different facies to that of the sparagmite sequence. We here find a thick sequence of sandstones and carbonates, all probably representing a marine-shelf environment. In the northern part of the Varanger halvøya turbidites predominate. Arkoses and coarse conglomerates are rare or absent in the late Precambrian sediments in Finmark and there is little evidence of faulting contemporaneous with sedimentation. Lithological units are very persistent and can be followed for long distances along the depositional strike (Reading 1965). This implies that sedimentation took place in a stable subsiding part of the continental crust near the continental margin.

In south Norway the Kvitvola Nappe is thrusted above the sparagmites and Lower Paleozoic rocks (Fig. 1a). This nappe contains rocks which are more tectonically deformed, usually with a strong lineation, and also contain units of Precambrian basement rocks (Oftedahl & Holmsen 1956, Bjørlykke 1965, Englund 1972). The Kvitvola sediments which are of lower greenshist facies consist of carbonates, and conglomerates are found. The carbonate rocks of the Kvitvola Nappe consist of dolomites and magnesite (Oftedahl & Holmsen 1956, Nystuen 1969) and represent probably a supratidal environment. A polymict conglomerate (the Koppang Conglomerate, Holmsen & Oftedahl 1956) may be interpreted as a tillite equivalent to the Moelv Tillite. Mainly because of the association of carbonates and diamictites the sedimentary rocks of the Kvitvola Nappe have been considered by most workers in the sparagmite region as a metamorphic allochthonous equivalent of the autochthonous Hedmark Group (Holmsen & Oftedahl 1956). However, radiometric data to support or preclude this assumption has so far been lacking. The lithology of the Kvitvola Nappe seems to be regionally uniform and corresponds more to the facies of Finmark, north Norway, than that of the sparagmites. It is probable that the root zone of the Kvitvola Nappe was along the western margin of the Baltic Shield somewhere near the present west coast of Norway (Fig. 33). If the Kvitvola Nappe is contemporaneous with the latest Precambrian sediments in Finmark and south Norway, it may represent a western equivalent to the continental marginfacies of north Norway. This involves thrusting in the order of 200 km to the southeast during the Caledonian orogeny.

The Särv Nappe in Sweden may represent an equivalent to the Kvitvola Nappe and have a root zone in a similar position to dolerite (Strømberg 1969 and Gee 1975). The Särv nappe contains dike swarms of dolerite and dolerites are also found in the Kvitvola Nappe (Holmsen & Oftedahl 1956) and may have intruded during the opening of the proto-atlantic ocean (Gee 1975).

In summary, two distinct facies can be recognised in late Precambrian sedimentary rocks of Scandinavia, depending upon their plate-tectonic position:

(1) In block-faulted grabens (rift valley) in central south Norway and adjacent parts of Sweden (sparagmite facies);

(2) Along the western and northern margin of the Baltic Shield (Finmark and Kvitvola Nappe facies).

The situation along the western margin of the Baltic Shield in late Precambrian times in many ways resembles the sedimentation on both sides of the Atlantic Ocean during the opening of the Atlantic Ocean in Mesozoic and Cenozoic times. In the North Sea (Kent 1975) there is evidence that the rifting phase (Jurassic) largely predates the ocean spreading phase (Cretaceous – Tertiary). The rifting of the sparagmite basin died out at the beginning of the Cambrian period. Although the plate-tectonic significance of this seems uncertain at this stage, it may as an analogy with the pattern found in the modern Atlantic Ocean signify a beginning of the spreading phase during the opening of the proto-atlantic ocean.

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July 1975

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