The Ulriken Gneiss Complex and the Rundemanen Formation: a basement-cover relationship in the Bergen Arcs, West Norway

HAAKON FOSSEN

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The Ulriken Gneiss Complex (UGC) and its sedimentary cover, the Rundemanen Formation, have been investigated. The UGC is a Precambrian migmatite complex which suffered strong but heterogeneous reworking during the Caledonian orogeny. Rocks of the Rundemanen Formation occur along three major shear zones in the UGC, and although strongly deformed, primary stratigraphic relations are locally preserved. Based on a few key localities a general stratigraphy has been established. A tentative lithostratigraphic correlation with rocks of the Hedmark Supergroup and the Valdres sparagmites of southern Norway may indicate a Late Precambrian age for rocks of the Rundemanen Formation.

H. Fossen, Statoil, Postboks 1212, N-5001 Bergen, Norway.

Introduction

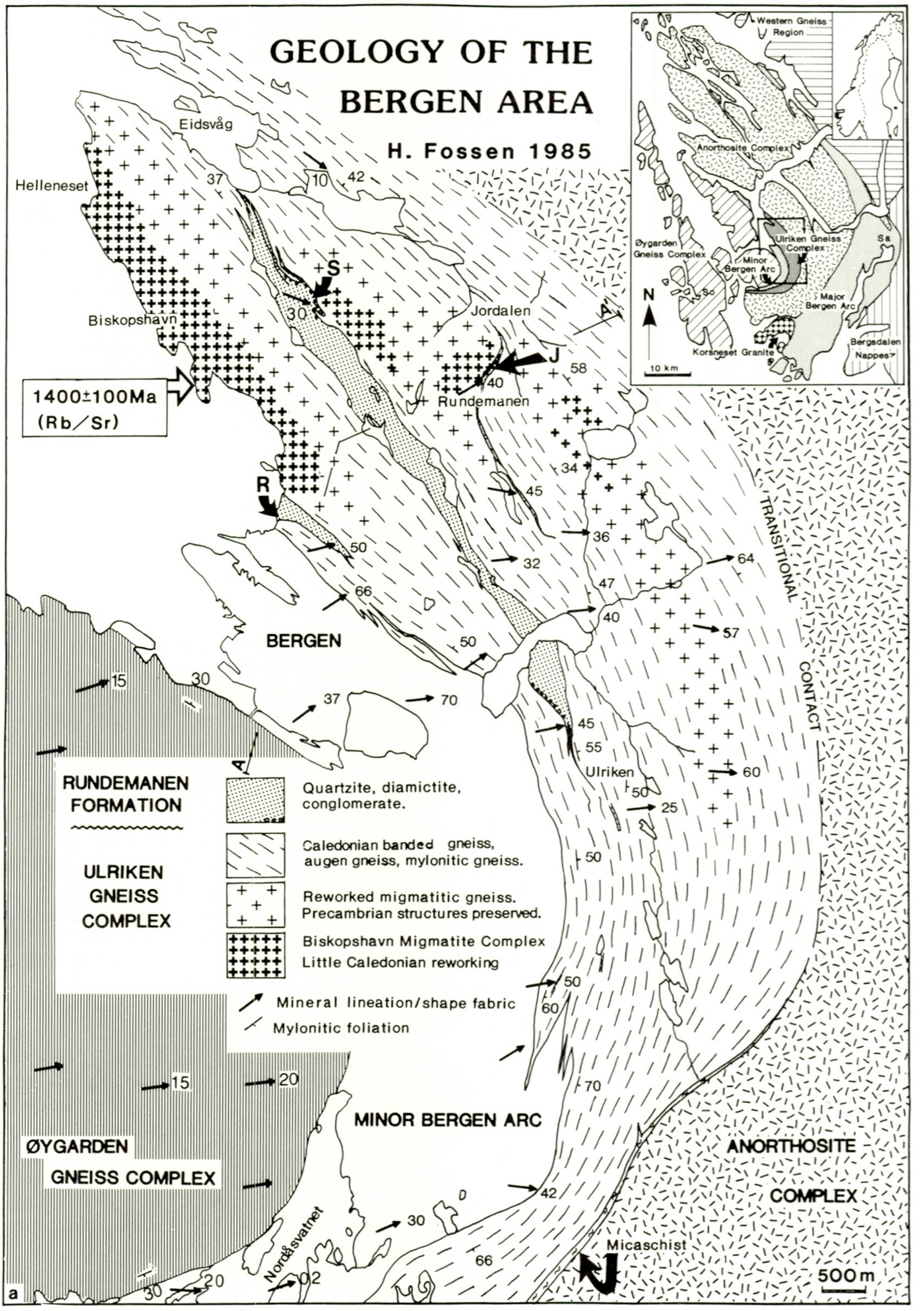
The Rundemanen Formation unconformably overlies the Ulriken Gneiss Complex (UGC), and together they form a basement-cover relationship within the nappe system of the Bergen Arcs (Fig.1). To the west they are separated from strongly deformed, presumed Lower Palaeozoic and older rocks of the Minor Bergen Arc (Kolderup & Kolderup 1940, Fossen 1986a). The contact is clearly tectonic with pronounced mylonitic fabrics, and rocks of the UGC are in contact with different units of the Minor Bergen Arc. Slivers of gneiss and quartz-rich metasediments may represent detached slices of the UGC and the Rundemanen Formation (Fossen 1986a).

To the east the UGC is in contact with the Anorthosite Complex which consists of Precambrian granulite facies gneisses reworked during the Caledonian orogeny, locally to eclogite facies (Austrheim & Griffin 1985). The contact to this complex is not very distinct (see Sturt & Thon 1978), but is marked by a transitional zone of retrogressed gneisses. A thin zone of mica schists to the south and east of Ulriken (Fig. 1) may represent the contact in the south. The marked difference in Caledonian metamorphic grade, the absence of anorthositic rocks and also of Precambrian granulite facies metamorphism in the UGC, and the absence of the Rundemanen Formation or similar metasediments in the Anorthosite Complex suggest that the two complexes were separated until the time of the main Caledonian (Scandian) orogeny. Lower Palaeozoic rocks of the Major Bergen Arc are located to the east of the Anorthosite Complex.

The tectonostratigraphic relations of these units are currently under revision, and the tectonostratigraphy suggested by Sturt & Thon (1978) has recently been modified (Fossen & Ingdahl 1987). Two different tectonostratigraphical interpretations are shown in Fig. 2.

The Ulriken Gneiss Complex

The gneisses of the UGC were generally strongly reworked during the Caledonian shearing, but heterogeneous strain has preserved areas from the otherwise penetrative deformation. Zones of low Caledonian strain are found in the Biskopshavn area and some smaller areas are present near Rundemanen (Fig. 1). In these areas a Precambrian complex of migmatitic gneisses is preserved, named the Biskopshavn Migmatite Complex (Fossen 1986a).


GEOLOGY OF THE BERGEN AREA
H. Fossen 1985
Eidsvåg
Helleneset
Biskopshavn
1400±100Ma (Rb/Sr)
Jordalen
Rundemanen
BERGEN
Ulriken
TRANSITIONAL CONTACT
MINOR BERGEN ARC
ØYGARDEN GNEISS COMPLEX
ANORTHOSITE COMPLEX
Micaschist
Nordåsvatnet
500m
RUNDEMANEN FORMATION
Quartzite, diamictite, conglomerate.
ULRIKEN GNEISS COMPLEX
Caledonian banded gneiss, augen gneiss, mylonitic gneiss.
Reworked migmatitic gneiss. Precambrian structures preserved.
Biskopshavn Migmatite Complex Little Caledonian reworking
Mineral lineation/shape fabric
Mylonitic foliation
Western Gneiss Region
Anorthosite Complex
Øygarden Gneiss Complex
Ulriken Gneiss Complex
Minor Bergen Arc
Major Bergen Arc
Korsneset Granite
Bergsdalen Nappes
10 km
a

The Biskopshavn Migmatite Complex

The oldest part of the Biskopshavn Migmatite Complex (BMC) is represented by granitoid, migmatitic rocks, commonly with a gneissic foliation (PS1) where thin (1-10 mm) schlieren of reddish quartzo-feldspathic leucosome (Mehnert 1968) alternate with narrow biotite-rich bands (the melanosome). Several generations of pegmatite postdate this foliation, but the oldest pegmatites are themselves deformed and transposed into the foliation. Locally, the orientation of PS1 varies considerably, but is commonly subparallel with the Caledonian foliation (NE-SW near Biskopshavn).

The old, commonly stromatic foliation is locally folded and cut by small shear zones. Pegmatites have intruded along some of these shear zones, and cut the PS1 foliation at high angles.

The above-mentioned features are cut by a series of metamorphosed basic dykes, sills and minor mafic bodies. These are now amphibolites or biotite schists. The amphibolites are themselves intruded by pegmatites, commonly net-veining the amphibolites (Fig. 3). At several localities the amphibolites are deformed. Foliated or unfoliated dykes are commonly boudinaged, and pegmatitic material has filled extensional cracks.

Grey, granodioritic dykes which locally transect the amphibolites are cut by a number of very coarse-grained, red pegmatites. The latter, which are up to 10-20 cm in width, are characterized by large crystals of magnetite. These pegmatites were formed both prior to and after the intrusion of pink granitic to granodioritic dykes. The granitoids have partly assimilated xenoliths of the country rocks. Weak

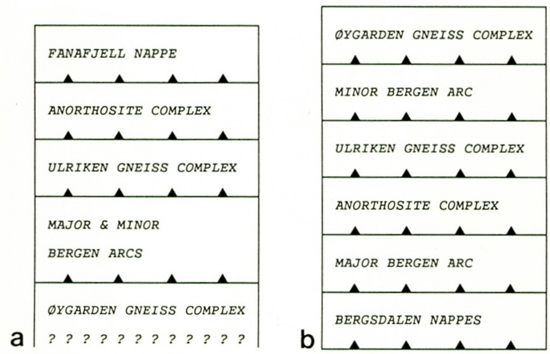


Fig. 2. Two different interpretations of the nappe stratigraphy of the Bergen Arcs; (a) from Sturt & Thon (1978); (b) interpretation based on large-scale asymmetrical folding of the Bergen Arcs (see text and Fig. 22). Note that the Fanafjell Nappe in Fig. 2a is part of the Major Bergen Arc in Fig. 2b.

(probably Precambrian) deformation has locally affected these latest components of the BMC.

A summary of the development of the Biskopshavn Migmatite Complex is given in Fig. 4.

Discussion of the Biskopshavn Migmatite Complex

The rocks of the BMC are clearly migmatitic with a stromatic foliation showing raft, vein, schlieren and diktyonitic structures (Mehnert 1968). An important question is whether the migmatite formed by partial melting of a paleosome or by hydrothermal segregation. Some of the pegmatites have formed in-situ, seen

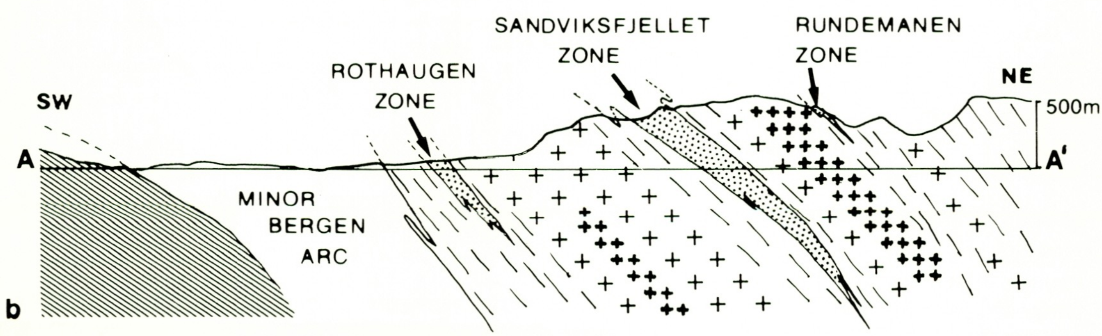


Fig. 1. (a) Geological map of the Ulriken Gneiss Complex and the Rundemanen Formation. The Caledonian mylonitic foliation (S1) and lineation (L1) is indicated. (b) Profile A-A'; the location is shown on themap.

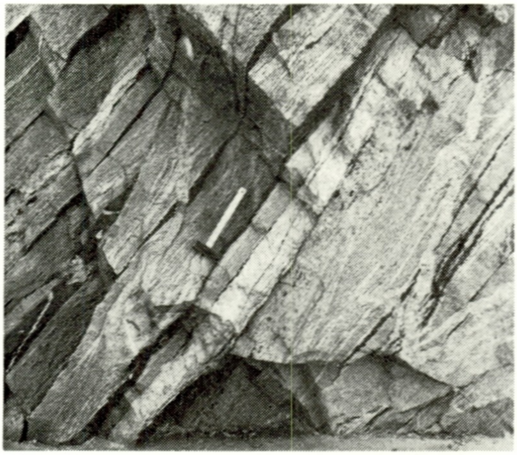


Fig. 3. Pegmatites cross-cutting an earlier Precambrian foliation. Some Caledonian flattening strain may have been superimposed on the rock. Biskopshavn (966037).

by the occurrence of rootless pegmatites (Fig. 5), while others show intrusive relationships (Fig. 3). However, these observations may be explained by partial melting and/or hydrothermal segregation of the gneisses.

Separation of the different migmatite-forming processes has been a matter of discussion for a long time. Yardley (1978) presented criteria for separation, and suggested that different mechanisms may be involved during the formation of a migmatite complex. According to these criteria, the very coarse-grained late pegmatites probably formed by hydrothermal segregation, but the earlier pegmatites (leucosomes) may have formed by partial melting. Similar chemical compositions of the pegmatites and the host rock (Elsoccary & Elbuusei 1983 and unpublished data) support the partial melting hypothesis.

Basic dykes and sills (amphibolites) were boudinaged during the migmatization, but were ductilely deformed and retrogressed to biotite-rich schists during the Caledonian deformation. This may indicate that the temperature was higher during the migmatization than during the Caledonian greenschist facies metamorphism, since the difference in competense between the basic sills and the granitoid migmatitic host rock increases with increasing temperature. However, no mineral parageneses are found which would prove that partial melting had occurred in Precambrian time. Only very detailed geochemical investigations may provide a more definitive answer to the migmatite-forming process in the Biskopshavn Migmatite Complex.

PRECAMBRIAN (Amphibolite facies)

Foliated migmatitic gneiss including 1st generation leucosomes and melanosomes	PS1 gneissic fabric
2nd generation pegmatites (leucosomes).	Rotated into the PS1 foliation, folded by PF2 isoclinal folds.
3rd generation pegmatites.	Cross-cutting PS 1
Intrusion of basic dykes and small bodies.	Uralitization (pyroxene-amphibole). Boudinage of the amphibolites. Local shearing/boundinage of some of the net-veined amphibolites.
4th generation pegmatites net-veining the amphibolites.	
5th generation pegmatites, magnetite-bearing, coarse-grained	
Intrusion of massive reddish granite. 1400 ± 100 Ma.	Decreasing metamorphic grade?
6th generation coarse-grained pegmatites.	

CALEDONIAN (Greenschist facies)

Shear zones cross-cutting all earlier structures

- amphibole biotite
- local development of mylonites and associated isoclinal folding
- rotation of Precambrian elements into the newly formed Caledonian fabric
- late crenulation and overgrowing white mica

Fig. 4. Schematic illustration of the development of the Biskopshavn Migmatite Complex.

The age of the Biskopshavn Migmatite Complex

Only one radiometric date is available from the BMC (Sturt et al. 1975). Late granitoid dykes were dated, yielding a Rb-Sr whole-rock isochron age of 1400±100 Ma ($^{87}Rb=1.42 \times 10^{-11}$). The date represents a minimum age for the migmatization and confirms a Precambrian age for the Biskopshavn Migmatite Complex.

Caledonian reworking

Caledonian reworking of the BMC occurred heterogeneously on all scales. In the Biskopshavn-Helleneset area the Caledonian reworking occurred in two ways:

(a) Heterogeneously, along distinct shear zones which clearly transgress the protolith gneisses.

(b) Less (but still) heterogeneously by transposition (rotation) of Precambrian structures, resulting in the formation of banded gneiss and augen gneiss. This is the dominant Caledonian deformation process which affected the BMC.

Distinct Caledonian shear zones cross-cutting the Precambrian structures are common in the Biskopshavn Migmatite Complex. They are generally recognized as dark, thin streaks or bands, abruptly transecting Precambrian rocks and structures such as pegmatites, basic sheets and Precambrian foliation (Fig. 6). All stages in the development can be studied. Generally, smaller-scale, narrow, Caledonian shear zones involved more cataclastic deformation than the wider shear zones which developed high strains. This may be due to strain softening, whereby the shear zones were initiated in the brittle, cataclastic regime and then passed over into conditions of more ductile flow.

Broader zones (some cm or more) show textures typical for tectonites and mylonites, such as S-C structures, asymmetric augen, and considerable recrystallization and reduction in grain size (Simpson & Schmid 1983, Lister & Snoke 1984). Locally, the deformation produced ambiguous or symmetric S-C structures, but generally the S-C relationship indicates overthrusting to the east. Quartz underwent ductile deformation whilst plagioclase was affected by a more brittle deformation.

Micas are represented by biotite and some white mica. Much of the biotite is an alteration product of amphibole, while the white mica locally grows at the expense of K-feldspar. The white mica has grown at different stages during deformation and occurs both as porphyroclasts and porphyroblasts.

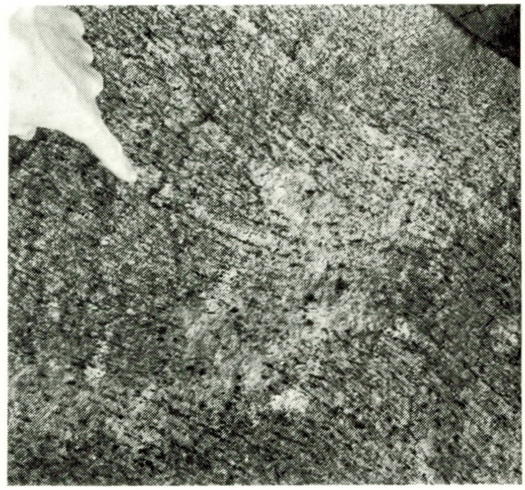


Fig. 5 Rootless pegmatite which may have formed by either anatexis or hydrothermal segregation. Biskopshavn (961046).

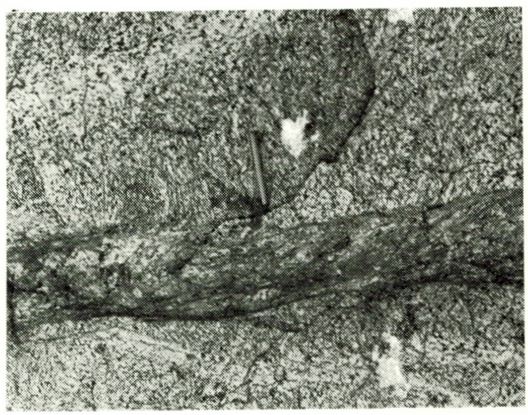


Fig. 6. Narrow Caledonian shear zone in the UGC, called type (a) in the text. Note pen for scale. Near Biskopshavn (975035).

The more important Caledonian reworking of the Precambrian migmatite complex is that of ductile simple-shear rotation and transposition of Precambrian elements into Caledonian banded gneisses. Hence, there is a transition between the BMC and the surrounding gneisses. A progressive increase in Caledonian strain is recognized from the Biskopshavn area towards the east (Fig. 1). This is shown by rotation and apparent flattening of Precambrian elements (i.e. foliations, cross-cutting pegmatites, granites and amphibolites) into the Caledonian schistosity.

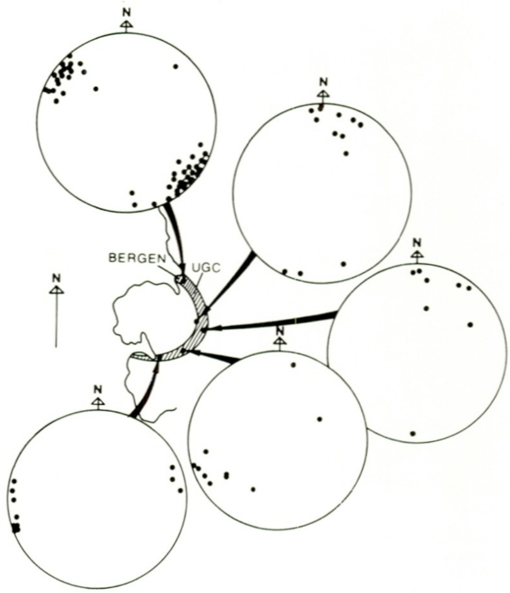


Fig. 7. F2 fold axes at different localities in the UGC and the Rundemanen Formation. Note that the axes follow the arcuate structure which is interpreted as an F3 structure.

Where Caledonian deformation was high, pegmatites are converted to augen gneiss, and basic dykes to stripes of biotite rock. Tightly banded gneisses, revealing isoclinal non-cylindrical folds and refolds represent more advanced stages. The size of the augen (porphyroclasts) decreases from a few cm to less than one mm with increasing strain.

On a larger scale, (proto-) mylonites predominate along the contact with the Nordåsvatnet Complex of the Minor Bergen Arc (Fossen 1986a) (Fig. 1) forming a broad zone which widens southwards. South of Ulriken, the deformation increases so that more or less all of the UGC forms a broad mylonite-zone. This is associated with a marked thinning of the UGC.

To the north, several wide (tens to hundreds of metres), mylonitic high-strain zones can be recognized. These divide the UGC into zones or sheets which are less intensely deformed. These mylonitic zones are concordant with the other Caledonian structural elements. It can be seen from Fig. 1b that the meta sediments of the Rundemanen Formation are localized along three, major, high-strain zones in the UGC.

A consistent, Caledonian, mineral lineation and shape fabric is common in the gneisses. It is best marked by elongated quartz-rods which plunge eastwards at various angles. Axes of Caledonian isoclinal folds are parallel or oblique to this lineation, and the axial planes of the folds are subparallel to the Caledonian mylonitic foliation. In high-strain zones, the mylonitic foliation is folded into typical sheath folds. The relatively consistent trend of the Caledonian lineation is found in all parts of the Øygarden Gneiss Complex; at Sotra (Johns 1981, Bering 1985), Løvstakken (Weiss 1977), Askøy (Askvik 1971) and in the presumed Lower Palaeozoic rocks of the Minor Bergen Arc (Fossen 1986a). This lineation is therefore easily recognized as Caledonian in the UGC. The metamorphism associated with the Caledonian reworking of the UGC is generally indexed by growth of white mica, biotite, chlorite and albite. This mineral assemblage indicates middle green schist facies conditions (biotite zone). However, mylonitic gneisses along the contacts with rocks of the Minor Bergen Arc locally contain metamorphic garnet and amphibole (pargasitic hornblende) in addition to biotite and white mica, testifying to a somewhat higher grade of metamorphism here (uppermost greenschist facies to lower most amphibolite facies).

A series of tight to open eastward-verging folds deform the proto-mylonitic Caledonian banding. The folds are best developed in the mylonitic rocks of the UGC, especially along the contact with the Minor Bergen Arc. Subhorizontal or gently east-dipping axial surfaces are typical. The fold axes are sub-horizontal and are parallel with the strike of the arc (Fig. 7).

The Rundemanen Formation

Three separate zones of predominantly psammitic rocks, named the Rundemanen Formation, are found in contact with the Ulriken Gneiss Complex (Fig. 1). Each zone represents an infolded and sheared part of a cover sequence to the UGC, preserved along Caledonian shear zones in the latter, as indicated by Sturt & Thon (1978). The zones are termed the Rundemanen, Sandviksfjellet and Rothaugen zone (Fig.1b). Based on the regional outcrop pattern (Fig. 1), each zone is interpreted as a sheared-out, non-cylindrical syncline or large-scale sheath fold.

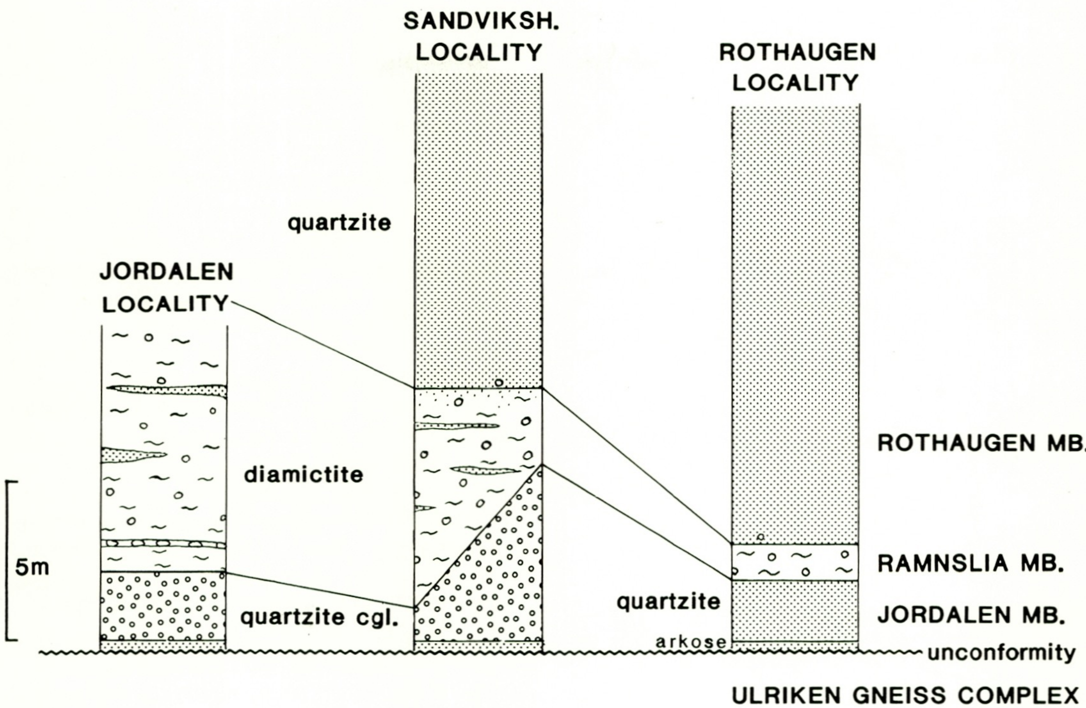


Fig. 8. Lithostratigraphic correlation between rocks of the Rundemanen Formation based on the three localities shown in Fig. 1a.

Stratigraphy

High strains in the Rundemanen Formation have obscured the primary stratigraphy, but from a few localities (Jordalen, Sandvikshytten and Rothaugen, Fig. 1a) it has been possible to establish a lithostratigraphy (Fig. 8). In descending stratigraphic order the Rundemanen Formation is divided into the Jordalen, Ramnslia and Rothaugen Members. Thickness estimates refer to present-day structural thicknesses.

The Jordalen Member

The type locality for the Jordalen Member is in Jordalen (Fig. 1a) (UTM 000039) where rocks of this member locally are practical undeformed. At the base of the Jordalen Member there is a partly reworked weathering arkose with a gradational contact against the underlying gneiss (UGC). The arkose is generally difficult to separate from the gneiss, but can be recognized by the following characteristic features:

- increase in amount of heavy minerals (mainly ilmenite and magnetite);
- somewhat finer grain size than the gneiss;
- increase in the quartz/feldspar ratio (Table 1);
- heavy-mineral bands which may show cross-stratification;
- scattered pebbles of quartzite locally in the arkose.

The presence of cross-stratification and a few scattered quartzite pebbles in the upper part of the arkose indicate that this part of the weathering debris was reworked, probably by rivers. The thickness of the arkose is usually difficult to estimate due to the nature of the basal transitional contact and the later deformation, but may locally have been several metres thick.

The arkose is stratigraphically overlain by a quartzite and/or a quartzite conglomerate (Fig.8). The conglomerate is best developed at the Jordalen locality (Figs. 9 and 10), the Sandvikshytten locality (Fig. 11) and on the northern side of Ulriken (Fig. 1a) (004999). Beds of conglomerate are normally 0.5-1.0 m thick and coexist with usually somewhat thinner beds of gritty arkose which show rare cross-lamination (Fig. 12). The conglomerate

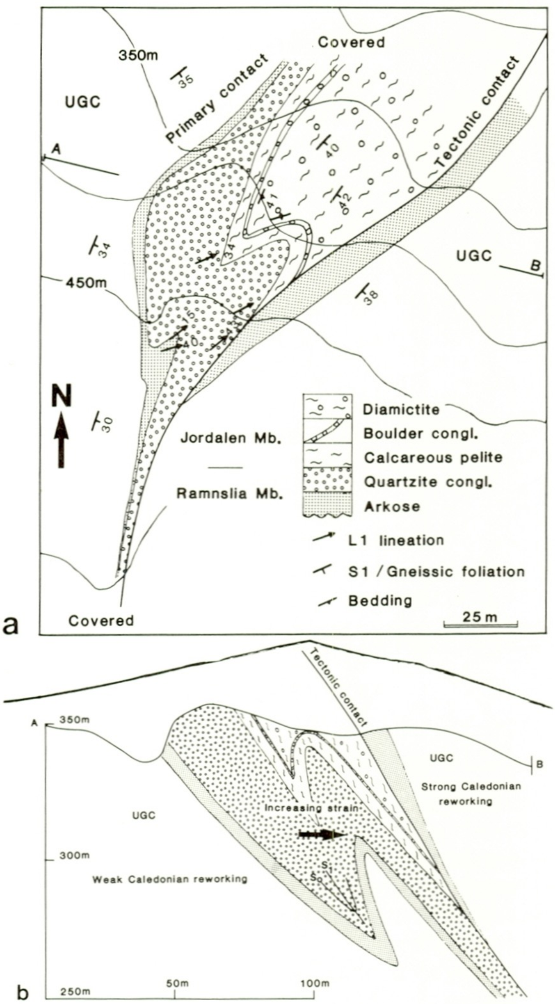


Fig. 9. Geological map (a) and profile (b) from the Jordalen locality. Note the steep relief.

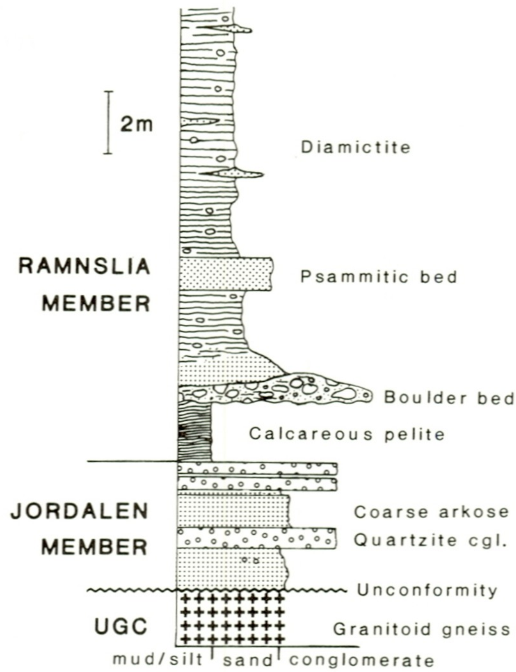


Fig. 10. Lithostratigraphical column from the Jordalen locality.

contains well rounded cobbles and pebbles of quartzite and vein-quartz, and a few pebbles of granitic gneiss. Many of the quartzite pebbles have pre-depositional (internal) tectonic fabrics, and some have pre- or syn-depositional weathering rims. Together with rounding, this may indicate redeposition, though the provenance of the clasts is not known. The original shapes of the pebbles and cobbles are strongly modified by deformation except at the Jordalen locality where undeformed or just slightly deformed conglomerates are locally found along the western contact to the UGC (Fig. 9b).

The thickness of the quartzite conglomerate seems to vary somewhat, being thickest in the fold at the Sandvikshytten locality (Fig. 8). Thinning or wedging out of the conglomerate along strike may be due to both tectonic pinching-out and primary thickness variations. However, the conglomerate is absent in the Rothaugen zone where the Jordalen Member is defined by a quartzite overlying the weathering arkose on the UGC and underlying the Ramnslia Member (Fig. 8). This indicates primary variations in the thickness and distribution of the conglomerate.

Interpretation of the Jordalen Member

While the lowermost arkose of the Jordalen Member seems to be a product of in-situ subaerial weathering and local current reworking, the overlying arkose, quartzite and quartzite conglomerate may be interpreted as braided stream deposits. Trough infills and alternating grit and conglomerate beds support this view, as does the relatively high-angle (35°) cross-stratification in the little-deformed arkose at the Jordalen locality. An alternative view may be that the sediments represent a shallow-marine (delta) deposit, but because of the strong deformation it is difficult to make more precise interpretations.

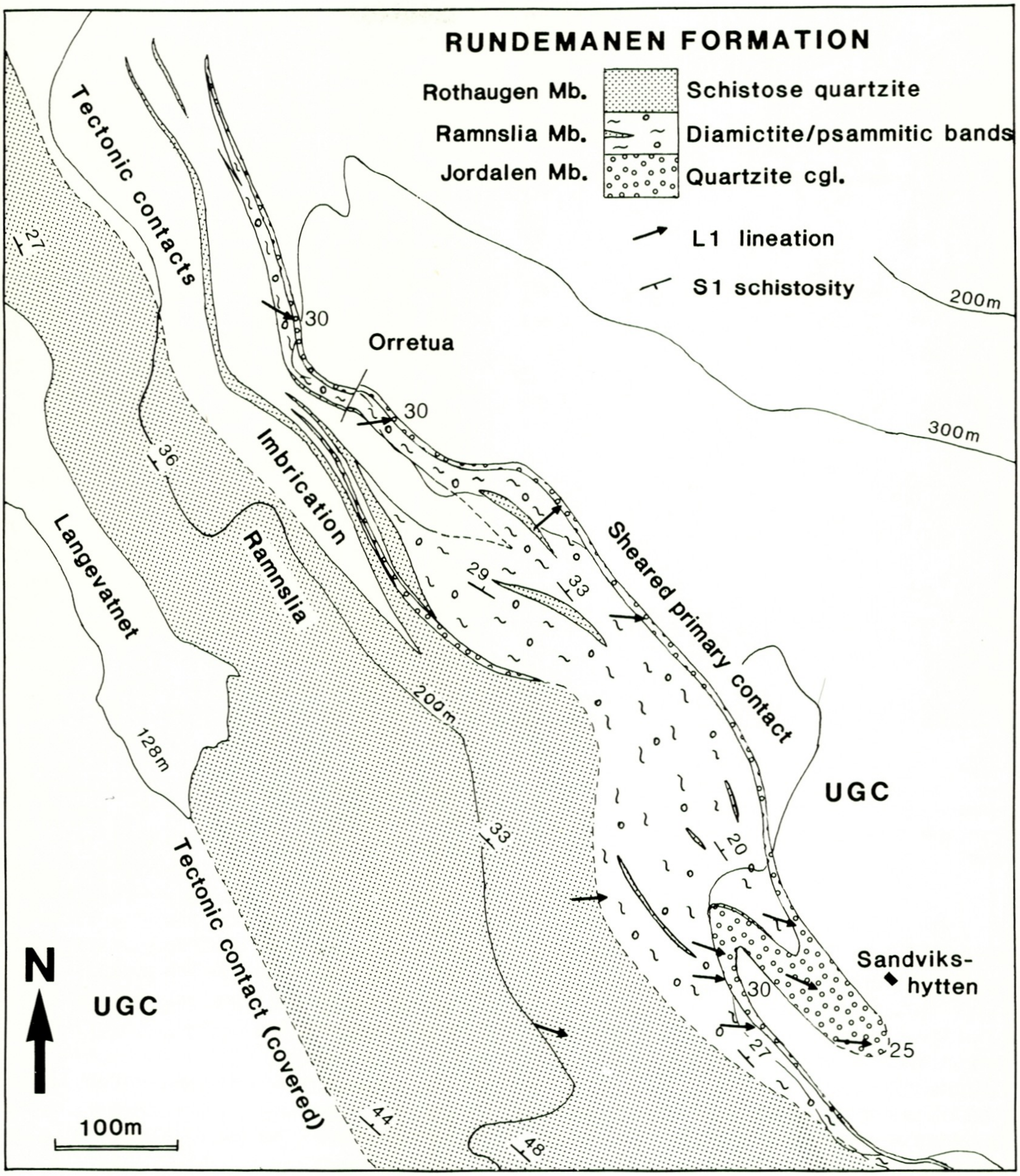


Fig. 11. Geological map of the area around the Sandvikshytten locality.

The Ramnslia Member

Above the quartzite conglomerate or basal quartzite there occur 1-10 m of psammites, semi-pelites and calcareous pelites, assigned to the Ramnslia Member. Oversized pebbles, cobbles and even boulders of granitic gneiss and quartzite (lonestones, Figs.13 and 14) are scattered throughout. Coarse sand-grains of quartz and feldspar are common. The oversized clasts are characteristic for this member.

Generally, the sediments are bedded and psammitic layers (usually 10-100cm) occur interbedded with more pelitic rocks. Locally, sub-angular boulders and pebbles are concen-

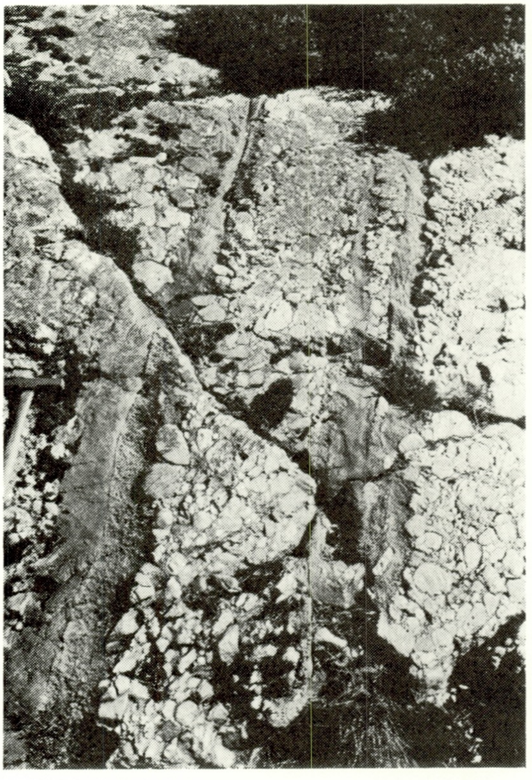


Fig. 12. Beds of quartzite conglomerate alternating with beds of coarse arkose. Section normal to the stretching direction (X-axis). Note cross-lamination in the upper part of one of the arkosic layers. Jordalen Member, Sandvikshytten locality (980045).

trated in bands, but usually they are randomly scattered, forming a diamictite (Flint et al. 1960). Deformation has obscured most primary features in the metasediments and separation of primary and tectonic structures is therefore difficult.

At the *Jordalen locality* (Fig. 10) the Ramnslia Member starts with a 2 m thick, plane-laminated, calcareous pelite with few or no oversized clasts. A boulder bed overlies this pelite, commonly with large granitoid boulders up to 0.5 m^3 and smaller quartzite pebbles scattered in an immature matrix rich in micas and poorly sorted sand-sized grains of quartz and feldspars. A very low-angle disconformity along the base of the boulder bed is probably of depositional origin. Above the boulder bed there are more than 10 m of pelite with lonestones (diamictite) and local psammitic beds and lenses.

The lowermost calcareous pelite is also seen at the *Sandvikshytten locality,* also here overlying a bedded quartzite conglomerate. The boulder bed is missing at this locality and the calcareous pelite grades into the diamictite which also here contains beds and lenses (up to 10m thick) of psammite (Fig.11). Poorly-exposed boulder beds appear, however, in the Sandviksfjellet zone near the western contact to the UGC, 3 to 4 km to the south of the Sandvikshytten locality. The stratigraphic level of these beds is not known.

In the *Rothaugen zone* the Ramnslia Member is represented by 1-3 m of diamictite with an immature, mica-rich matrix and granite and quartzite lonestones. Boulder beds are not present, but calcareous layers (about 10 cm) occur locally near the base of the member, which overlies the basal quartzite.

Because of deformation and the low competence of the rocks of the Ramnslia Member, primary structures are scarce. The best place to look for primary features is at the Jordalen locality where the conglomerate pebbles of the underlying Jordalen Member are virtually unstrained. Some of the lower beds here show minor low-angle disconformities which probably are of depositional origin. Within the boulder bed there is a cobble which has deformed the underlying lamination while the overlying laminae pass straight across (Fig. 14). As strain variations are large over small areas it is difficult to know whether this actually represents a primary feature or is merely a product of deformation.

Interpretation of the Ramnslia Member

The Ramnslia Member contrasts with the other members in that it is generally more pelitic and carries scattered pebbles, cobbles and boulders (lonestones), usually of granitic rocks. Matrix-supported diamictites have normally been interpreted in one of two ways:

(a) as mudflow (debris-flow) deposits,
(b) as a tilloid formed by glaciogenic processes.

It is commonly very difficult to distinguish between these two types of sediment even when undeformed. A discussion of criteria for separation is given by Spencer (1971). In the Ramnslia Member there are some indications in support of a glaciogenic origin. The strongest indication is the common presence of bedding and sand lenses within the metasediment. The bedding may be caused by winnowing of unsorted diamictites by weak bottom currents which were able to move part icles up to medium sandsize (Link & Gostin 1981). Bedding and sand lenses are commonly observed in shallow-marine glacial deposits (Marmo & Ojakangas 1985, Magee & Culver 1986), but debris-flow deposits tend to have a chaotic appearance with few or no signs of bedding or sorting (e.g. Blatt et al. 1980). Another glaciogenic indicator may be the gradual transition into the overlying rocks of the Rothaugen Member. This is a common feature in shallow-marine glaciogenic sediments, as e.g. at the upper contact of the Moelven Tillite in the Hedmark Supergroup (Bjørlykke et al. 1976). If it could be shown that the observed deflection of the lamination beneath the cobble in Fig. 14 was a primary feature, this would be a good indication of the presence of ice-rafted clasts; however, as tectonic deformation may be responsible for the structure, its significance is uncertain. Another point is that the thick weathering arkose of the Jordalen Member may have been formed during a relatively long period of denudation (weathering) of the UGC. This may indicate a relatively flat palaeotopography, whereas gravitational, debris flows are more likely to have formed where the landscape is more undulating.

It may be argued that the relativly thick weathering arkose in the Jordalen Member indicates a comparatively hot and dry climate. Such a climate would contrast with the assumed glaciogenic conditions during deposition of the Ramnslia Member if it represents a tilloid. However, in the Dalradian of Scotland the Vendian tilloids are directly overlain and underlain by stromatolite-bearing dolomites indicative of warm waters (Anderton et al. 1979). This is consistent with palaeomagnetic studies which indicate palaeolatitudes of 10-15° (Tarling 1974). Rapid climatic changes are recorded in other Late Precambrian tilloids, e.g. in the Hedmark Supergroup (Biri Formation, Nystuen 1976, Bjørlykke 1978) and in the Adelaidean Sturtian glacial sequence in Australia (Link & Gostin 1981), testifying to very rapid climatic changes over large parts of the world. Hence, a hot climate prior to the deposition of the Ramnslia Member by no means precludes a glaciogenic origin.

Strong indications of a glaciogenic origin, such as faceted and striated clasts, striated pavement, and definite evidence for ice-rafted dropstones or sand-filled ice-wedges, have

Fig. 13. Granitoid lonestone in the diamictite of the Ramnslia Member, Ramnslia (978046).

not been found. If such structures are or have been present, the strong deformation has made their identification impossible. Thus, no definite conclusion can be drawn about the origin of the Ramnslia Member; but a glaciogenic origin, perhaps as a dropstone facies, is a viable alternative.

The Rothaugen Member

There is a gradual transition from the Ramnslia Member to the overlying Rothaugen Member at most localities, defined by the disappearance of lonestones and commonly by a more psammitic composition. The Rothaugen Member is dominated by variably schistose and massive quartzite, locally with thin micaceous bands. The massive quartzite dominates the Rothaugen zone, and contains local bands enriched in heavy minerals (mainly magnetite and ilmenite). In addition, a few thin (‹20cm) beds of arkose, grit and conglomerate are observed. Structures similar to cross-stratification can be demonstrated to be sheared folds (Fig. 15), and actual sedimentary cross-stratification

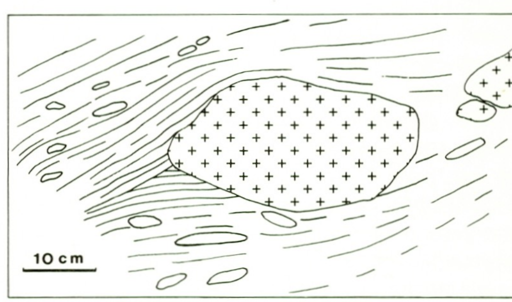


Fig. 14. Boulder in the boulder bed, Jordalen, where the underlying lamination is gently deformed, and cut by the lamination above. Stratigraphically right way up. Redrawn from photograph (000039).

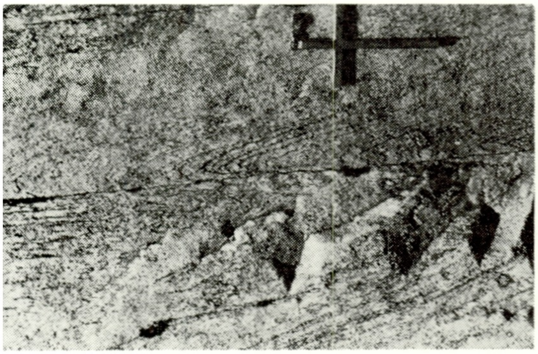


Fig. 15. Sheared F1 fold in the Rothaugen Member. Sandviksfjellet zone (980046).

is impossible to distinguish from the sheared folds. However, structures very similar to those formed by soft sediment deformation are seen along the road sections at the Rothaugen locality (Nye Sandviksvei). Although the primary thickness has been strongly modified by isoclinal folding and shearing (Fig. 16), the Rothaugen Member is the thickest member in the Rundemanen Formation, being several tens of metres thick. The apparent absence of the Rothaugen Member at the Jordalen locality is probably due to tectonic pinching-out.

Because of the tectonic deformation, little can be said about the origin of the psammitic Rothaugen Member. Since it is a more mature sediment than that of the other members it may represent a shallow-marine sand, possibly related to delta sedimentation.

Deformation and metamorphism

The rocks of the Rundemanen Formation are generally pervasively deformed, but at a few localities deformation is relatively weak. These areas have been useful in determining the structural development which will be summarized below.

The first episode of deformation recognized (D1) is the most pronounced. F1 folds are tight to isoclinal, and an axial-planar S1 foliation (Fig. 17) has developed along with a stretching lineation. Intense shear accompanied this deformation, and the S1 fabric is generally pervasive throughout the Rundemanen Formation. The D1 deformation can be traced into the UGC where it appears to be the first Caledonian deformation affecting the migmatitic gneisses (Fossen 1986a). The three zones of metasediments are preserved in Caledonian shear zones which developed during the D1 deformation in the Rundemanen Formation. Where the D1 shearing has been intense in the Rundemanen Formation, the Caledonian reworking has been correspondingly intense in the nearby gneisses. Where deformation is weak or almost absent in the Rundemanen Formation, the gneiss adjacent to the contact shows only Precambrian structures without obvious signs of Caledonian reworking.

Both mesoscopic (Figs. 15, 16 and 17) and macroscopic (Figs. 1a, 9 and 11) F1 folds are present. Mesoscopic folds commonly show that limbs have been removed by shear (Fig. 15), and the Rundemanen Formation is probably preserved in large-scale examples of these sheared F1 folds (Fig. 18).

Mappable F1 folds seem to have fold axes parallel to those of the mesoscopic F1 folds (Fig. 19). From the regional outcrop pattern (Figs. 1a and 11) it can be seen that the axes of the mesoscopic folds must be curved, forming non-cylindrical folds of sheath-like geometry. The fold axes are generally sub-parallel to the stretching lineation even though the latter may have somewhat different orientations at different localities (Figs. 19 and 20). If the stretching lineation is taken to be parallel to the movement (shear) direction, this means that differential movements during D1 may account for the variation. Though fold axes generally are oblique to the lineation where sheath folds curve (e.g. Bell & Hammond 1984), complex strain histories may probably result in sub-parallel lineations also where the fold axes curve. It is not known if this is the case in the Rundemanen Formation, since the fold axes are rarely exposed or have been destroyed by subsequent shearing, but the various orientations of the fold axes (and lineations) may represent different positions on macroscopic sheath-like folds as well as indicating differential movement directions. The variation in orientation cannot be caused by a D2 refolding alone, since the variations in dip where the lineations and fold axes have been measured are small. Fold axes formed during D3 are sub-parallel to L1 and caused no important rotation of L1 and F1.

Fossen (1986a) and Holst & Fossen (1987) reported large variations in finite strain in the quartzite conglomerate of the Jordalen Member. Extreme flattening strains seem to domina-

Fig. 16. Isoclinally folded and transposed bedding where folds have been sheared out. From the Rothaugen Member in the Sandviksfjellet zone (980043).

te, especially on the long limb of F1 folds. Locally the strains were plane or constrictional, as in the short limb and hinge zones of the fold at the Sandvikshytten locality (Holst & Fossen 1987). This fold is interpreted to have developed passively as the result of pertubation of the D1 flow, and the generally flattening finite strain in the conglomerate was locally transformed into constrictional strain. In this case there has been a local dextral shear component (Fig. 21) which is also indicated from kinematic indicators in the UGC at the same locality.

The dextral shear in the Sandvikshytten area is, however, not the dominating bulk flow. S-C structures in D1 shear zones the UGC signify a dominant over thrusting to the east, although movements in the section normal to the thrusting direction also occurred. Asymmetric structures indicate that a component of non-coaxial strain is involved (Berthé et al. 1979, Platt & Vissers 1980, Platt 1984, Chroukroune et al. 1987). However, the bulk flattening finite strain shows that the bulk deformation was not a simple shear, but rather a strain which may be factorized into a component of pure shear and simple shear as illustrated by the strain matrix pre-multiplication

$$\begin{bmatrix} 1+e1 & 0 & \gamma(1+e_3) \\ 0 & 1+e_2 & 0 \\ 0 & 0 & 1+e_3 \end{bmatrix} = \underbrace{\begin{bmatrix} 1 & 0 & \gamma \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{bmatrix}}_{\text{simple shear}} \underbrace{\begin{bmatrix} 1+e_1 & 0 & 0 \\ 0 & 1+e_2 & 0 \\ 0 & 0 & 1+e_3 \end{bmatrix}}_{\text{pure shear}}$$

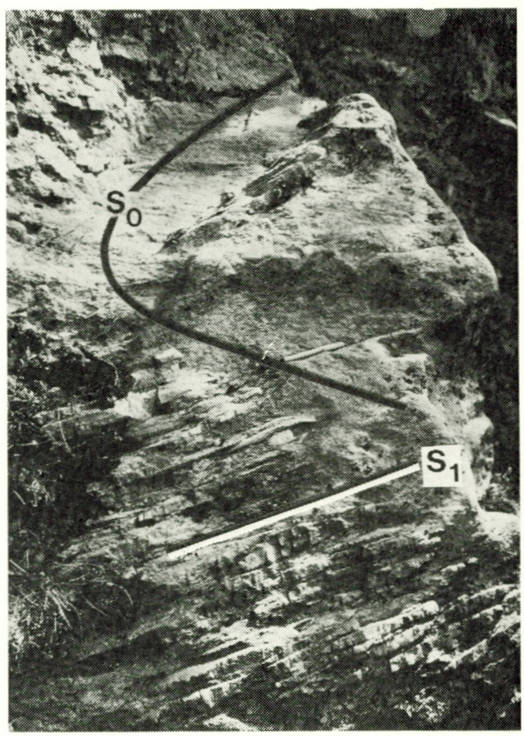


Fig. 17. F1 fold in the Jordalen locality where bedding is folded, and S1 is defined by flattened quartzite pebbles in the Jordalen Member (999037). Looking south.

where γ is the shear strain and $1+e_1>1+e_2>1+e_3$ are the principal extensions (dilation is not considered). In the Rundemanen Formation (and the UGC) it is difficult to quantify the stra-

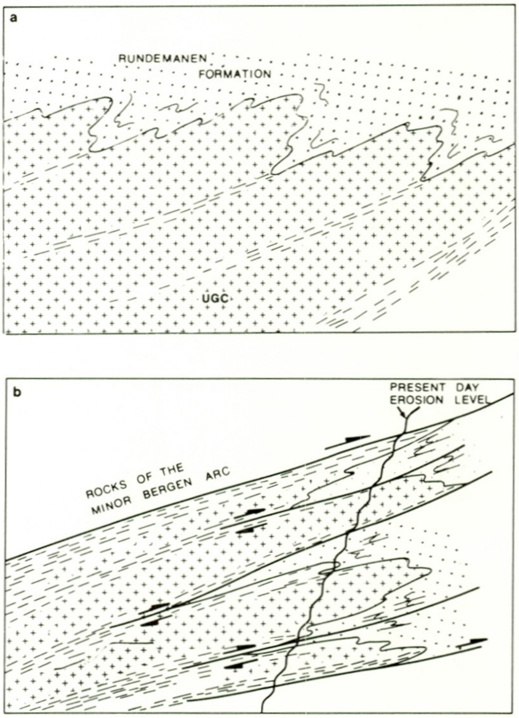


Fig. 18. Cartoon showing how the primary basement-cover relationship may have been modified by shearing in the UGC, and folding and shearing in the Rundemanen Formation.

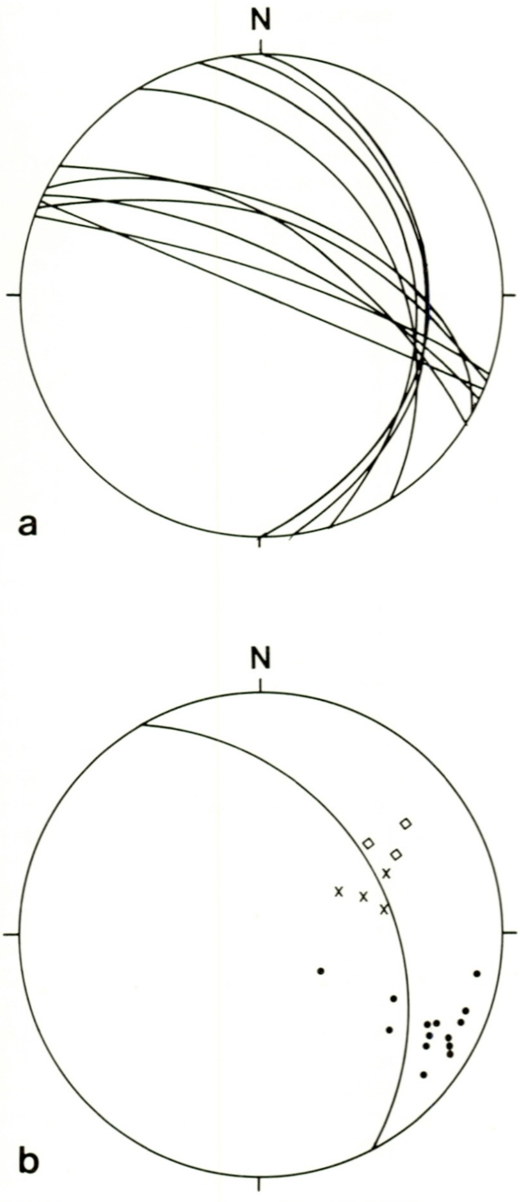


Fig. 19. (a) Measurements of bedding around the lower hinge zone of the F1 fold at the Sandvikshytten locality indicates a theoretica l fold axis plunging to the E or ESE. (b) Orientation of mesoscopic fold axes in the Sandvikshytten locality (dots), Jordalen locality (crosses) and north of Ulriken (squares). The great circle indicates the average plane of foliation in the area (330/35NE).

ins, but $1+e_1 \gg 1+e_2 > 1+e_3$ and the bulk simple-shear factor seems to be less than the pure-shear factor.

Sheath folds formed during this shear movement would soon be modified by the strong flattening component, and local lateral flow (Fig. 21) may involve lateral simple shear of the form

$$\begin{bmatrix} 1 & 0 & 0 \\ 0 & 1 & \gamma \\ 0 & 0 & 1 \end{bmatrix}$$

since bulk pure shear may involve simple shear (or any other non-coaxial shear) on a relatively smaller scale (e.g. Choukroune et al. 1987, Gapais et al. 1987), like the dextral shear at the Sandvikshytten locality.

Open to tight F2 folds deform the S1 foliation and the F1 folds. The F2 folds are asymmetrical with easterly vergence and with axial planes dipping gently to the north-east and east. A crenulation cleavage and associated lineation are commonly developed in the hinge zones. The sub-horizontal F2 fold axes are subparallel to the strike of the S1 foliation, i.e. parallel to the arc structure (Fig.7). Most observed F2 folds have short limbs of 1-20 m and long limbs more than 5 m. The same generation of folds is seen in the UGC and the Minor Bergen Arc (Fossen 1986a). If the folds are parasitic to a larger (km-sized) fold structure, their consistent easterly vergence may indicate that rocks of the Rundemanen Formation, UGC and surrounding units are situated on the inverted short limb of a major monoclinal fold (Fig. 22).The tectonostratigraphy shown in Fig. 2b is a consequence of such a monoclinal fold.

Late-stage shearing is indicated in the Ramnslia area by the development of ultracataclastic to ultramylonitic fault rocks (978048). The importance of this feature is not known, but it seems that the late shearing is related to D2. Local shear-deformation of the D2 crenulation lineation at the Rothaugen locality shows, however, that ductile or semi-ductile deformation occurred locally also after D2.

D3 structures are open, kink-like folds which refold the F2 folds (Fossen 1986a). The D3 folds have sub-vertical axial planes with easterly-plunging axes (Fig. 23), and are related to the formation of the arcuate structure of the Bergen Arcs (Fossen 1986a). They occur sporadically in the Caledonian-deformed rocks of the Rundemanen Formation and the UGC, and their amplitude varies from the microscopic scale to tens of metres or more. Preliminary work indicates a difference in asymmetry in the northern and southern parts of the Bergen Arc structure. Fossen (1986a) interpreted this phase of deformation as having produced the large-scale, open folding of the Bergen Arc System, and the pattern of asymmetry may support the interpretation that the F3 folds are parasitic to the arcuate structure.

The Caledonian metamorphic grade does not appear to have exceeded biotite grade of greenschist facies. The absence of garnet and the composition of the white micas (Fig. 24) lend support to this interpretation. This is consistent with the Caledonian metamorphic grade seen in rocks of the UGC (above), where the grade increases close to the Minor Bergen Arc. The metamorphism of the rocks of the UGC and the Rundemanen Formation is lower than the corresponding Caledonian metamorphism seen in the rocks of both the Minor Bergen Arc (Fossen 1986a, b and in prep.) (Fig. 24), the Major Bergen Arc (Ingdahl 1985, Fossen 1986b and in prep.) (Fig. 24) and the Anorthosite Complex (Austrheim & Griffin 1985).

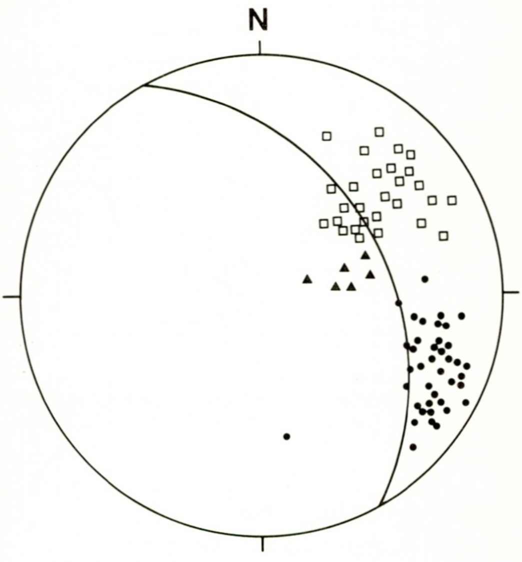


Fig. 20. Stretching lineations in the quartzite conglomerate from the Sandvikshytten area (dots), the Jordalen locality (triangles) and the northern side of Ulriken (squares). The great circle indicates the average plane of foliation in the area (330/3 5NE).

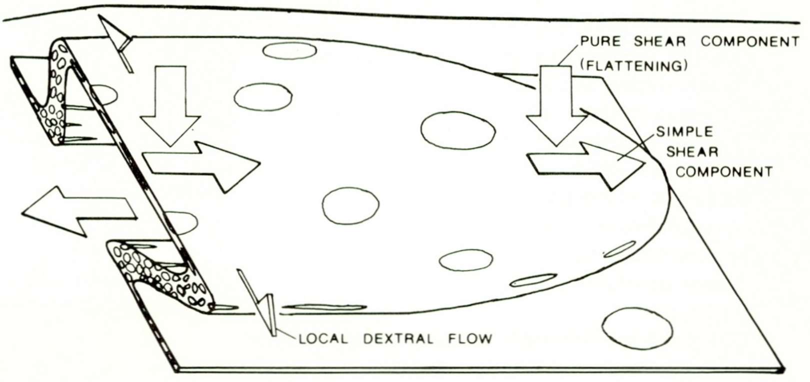


Fig. 21. Schematic illustration of a sheath-fold formed during thrusting to the east. Flattening strain during the development of the fold causes local lateral flow, either dextral or sinistral.

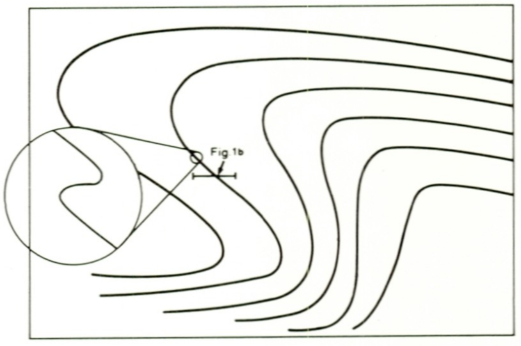


Fig. 22. Illustration of the hypothetical large-scale asymmetric fold to which the mesoscopic F2 folds may be related. The profile in Fig. 1b is indicated.

Age and correlation of the Rundemanen Formation

The deposition of the Rundemanen Formation clearly postdates the latest granitoids of the UGC which are Precambrian in age (Sturt et al. 1975). The main tectonic event of the Caledonian orogeny is expressed by the D1 and D2 structures in the Rundemanen Formation. The D1 (and later) deformation structures can be traced into presumed Upper Ordovician rocks of the Minor Bergen Arc (Fossen 1986a), indicating a probable Silurian (Scandian) age for the deformation and metamorphism. No earlier Caledonian structures have been found in the Ulriken Gneiss Complex or its cover. Hence, the Rundemanen Formation was deposited sometime in the interval from 1400 ± 100 Ma to the time of the main Caledonian orogeny which affected rocks of Llandovery age in the Major Bergen Arc in the Ulven area (Reusch 1882, Ryan & Skevington 1976).

As no fossils are found and no isotopic dates are available for the Rundemanen Formation, a lithostratigraphic correlation appears to be the only means of assigning an age to these rocks. Metasedimentary sequences with similar lithologies in South Norway are:
(a) the Precambrian Telemark Supergroup;
(b) quartzite-dominated rocks in the Bergsdalen Nappes which may be correlated with the Telemark Supergroup;
(c) the Late Proterozoic 'Sparagmites';
(d) the Cambrian basal quartzite;
(e) Silurian quartzite conglomerates and quartzites.

The Precambrian Telemark Supergroup is best known from the Telemark district (Dons 1960), but is also described from the Hardangervidda area (Naterstad et al. 1973). Supracrustal rocks of the Ullensvang Group (Torske 1983) in the Hardangerfjord area are also interpreted as Telemarkian (Reusch & Kolderup 1902, Torske 1983), and allochthonous rocks in the Bergsdalen Nappes (Fig. 1, inset map) east of the Bergen Arcs have also been suggested to be equivalent to the Telemark Supergroup (Kvale 1979).

Like the rocks of the Rundemanen Formation, the Telemark Supergroup comprises quartzites and quartzite conglomerates. However, the Telemark Supergroup shows some important differences from the Rundemanen Formation. The former contains considerable amounts of metavolcanic rocks of both basic and acidic composition, and layers of quartzite and quartzite conglomerate are intercalated with the metavolcanites. The Telemark Supergroup was also deformed, metamorphosed and intruded by granites during the Sveconorwegian orogeny (Dons 1960, Torske 1977, 1983). The Rundemanen Formation shows no signs of pre-Caledonian intrusive activity, deformation or metamorphism, nor is any volcanism recorded. Correlation of the Telemark Supergroup and the Rundemanen Formation therefore seems unlikely.

Quartzites, pelites and quartzite conglomerates of the Ullensvang Group also include basic and acidic volcanic rocks, and are intruded by

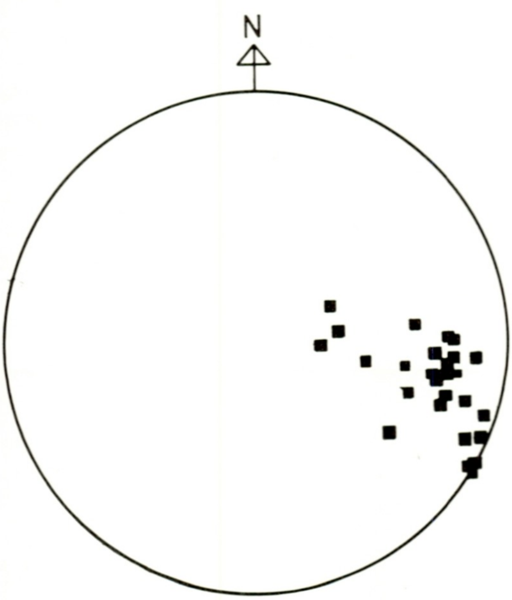


Fig. 23. F3 fold axes from the Rundemanen Formation.

basic and acidic intrusions (Torske 1983 and unpublished data). Several of the intrusions were synkinematic with respect to the polyphasal Proterozoic deformation which affected the metasediments (Torske 1983).

Similarly, quartzites, quartzite conglomerates and volcanites in the Bergsdalen Nappes (Kvale 1947, 1979) were deformed and intruded by basic and acidic intrusive rocks in Precambrian times. Age determinations indicate a Sveconorwegian age for granite intrusions (Brueckner 1972, Pringle et al. 1975, Gray 1978).

Rocks of the Telemark Supergroup, the Ullensvang Group and metasediments in the Bergsdalen Nappes include metasediments similar to those of the Rundemanen Formation. Since there is no sign of intrusive or extrusive rocks, nor any Precambrian deformation structures in the Rundemanen Formation, a correlation seems unlikely. However, the possibility that rocks of the Rundemanen Formation are 'Telemarkian' without volcanite sand which escaped the Sveconorwegian deformation, metamorphism and intrusive activity, cannot be totally excluded.

The Late Precambrian sediments (the 'Sparagmites') of South Norway are best known from the Mjøsa area (Bjørlykke et al.1976) where the deformation and metamorphic grade has been very low (Bjørlykke 1976, 1978). Allochthonous sparagmites occupy a large area to the northwest of Mjøsa, forming the Osen-Røa Nappe Complex (Nystuen 1981, 1982 and 1983) and the Kvitvola and Valdres (Lower Jotun) Nappe Complexes (Loeschke & Nickelsen 1974, Hossack 1976, Nystuen 1983). Metasediments correlated with the Valdres

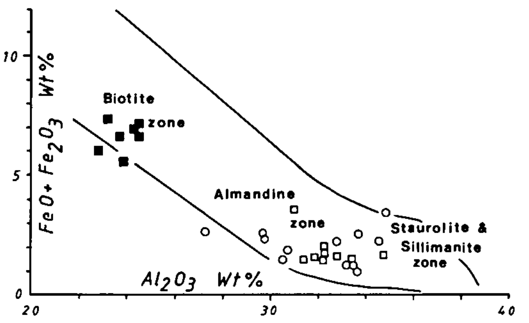


Fig. 24. White micas from the Rundemanen Formation (filled squares) compared with white micas from the Minor Bergen Arc (open squares: Nordåsvatn Complex) and the Major Bergen Arc (open circles: Ulven Group and Samnanger Complex). The formula proportion variation diagram is modified from Miyashiro (1973).

sparagmites are also found in the Sognefjell area near Sognefjorden (O. Lutro in prep.).

At a few localities, these metasediments have preserved primary depositional contacts to Precambrian gneiss slices (e.g. Loeschke & Nickelsen 1968, Hossack et al. 1981, Lutro in prep.). The following similarities exist between the sparagmites and the Rundemanen Formation:

(a) A basal arkose commonly occurs where the primary unconformity is preserved within the thrust-sheets (Hossack et al. 1981, Lutro in prep.).

(b) Fluvial arkoses, quartzites and quartzite conglomerates similar to the Jordalen Member overlie the unconformity, e.g. the Bygdin conglomerate (Hossack 1976), Grønsenknipa (Loeschke & Nickelsen 1968) and Rendal Formations (Nystuen 1981).

(c) Stratigraphically up-sequence, psammites,

Thin-section	Quartz	Feld-spar	White mica	Biotite	Carbonate	Opaques	others	Qtz/Fsp ratio	UTM
16680	58.4	5.4	34.0	–	–	1.8	0.4	10.80	977044
S1F	69.0	11.0	18.1	1.3	0.6	–	–	6.28	975023
S5SH	42.7	16.4	10.9	0.3	27.0	2.1	0.6	2.60	980045
G1SH	12.9	5.4	70.0	0.8	–	10.6	0.2	2.39	980045
18757	46.0	35.3	9.6	0.1	6.0	2.7	0.3	1.30	999039
G4R	16.5	45.1	26.2	9.7	–	1.4	1.1	0.37	999039

Table 1. Modal analyses from the Rundemanen Formation. 16680: quartz-schist, Rothaugen Mb., Munkebotsvatnet. S1F: quartzite, Rothaugen Mb., Rothaugen S5SH: quartzite, top of the Jordalen Mb., Sandvikshytten locality. G1SH: Arkose, lower part of the Jordalen Mb., Sandvikshytten locality. 18757: arkose, lower part of the Jordalen Mb., Jordalen locality. G4R: granitic gneiss (UGC), 1-2 m below the unconformity, Jordalen locality. 800-1000 counts.

in places rich in feldspar, are common (although pelitic facies also occur). A decrease in feldspar content is usually seen away from the unconformity (Hossack 1976, Hossack et al. 1981).

(d) Diamictitic horizons of possible or certain glaciogenic origin correlated with the Varangian tillites in Finnmark are found in the Osen-Røa Nappe Complex (Moelven Tillite; Bjørlykke et al. 1976, Nystuen 1976), Kvitvola (Nystuen 1980), Särv and Valdres (Lower Jotun) Nappes (Loeschke & Nickelsen 1974, Nickelsen 1976). The diamictites are similar to the sediments of the Ramnslia Member. Thus, a correlation between the Rundemanen Formation and the Late Proterozoic sparagmites of southern Norway seems reasonable.

The Cambrian unconformity, known from several areas in southern Norway as well as in northern Scotland, also locally exhibits a weathering arkose which passes up into a thin basal quartzite, or less commonly, a thin basal conglomerate (Hossack et al. 1981). However, the Cambrian sections are dominated by pelitic rocks and some limestones, and diamictites have not been described. The basal quartzite also contains trace fossils (Scolithos). A correlation with the Rundemanen Formation is possible but, based on lithological criteria, not likely.

Late Ordovician and/or Silurian quartzite conglomerates are described from western Norway, e.g. in the Skarfjell Formation, Major Bergen Arc (Ryan & Skevington 1976), in the Hersvik area, Solund (Furnes 1974) and in the Herland Group, Atløy (Brekke & Solberg 1987). These conglomerates occur at the top of cover sequences to Caledonian ophiolite fragments, or contain detritus from ophiolites (Herland Group). In the Major Bergen Arc and the Hersvik area they overlie pelites or green sediments. Therefore, a correlation with the Rundemanen Formation which unconformably overlies Precambrian gneisses and is not associated with ophiolitic (eugeoclinal-type) rocks is considered unlikely.

Conclusions

The Ulriken Gneiss Complex records a Precambrian history, including the development of a migmatite complex (the Biskopshavn Migmatite Complex). These migmatites form the parent rocks to most of the gneisses of the UGC, which were strongly reworked during the Caledonian orogeny. The overlying Rundemanen Formation, which is now preserved in high-strain zones in the gneiss, shows both inverted and right-way-up primary depositional contacts with the UGC. The Rundemanen Formation has taken part in the Caledonian deformation and metamorphism, but not in the Precambrian migmatization.

The Caledonian deformation has been intense in the Rundemanen Formation, but as the strain was heterogeneous it has been possible to reconstruct the primary stratigraphy. Lithostratigraphic correlations across wide areas are uncertain, because of changes in sedimentary facies as well as in the degree of deformation and metamorphism. The allochthonous nature of the UGC and the Rundemanen Formation, and the uncertainty of their palinspastic relations to more well-known, similar, basement-cover units also make correlation uncertain. However, when comparisons are made between the rocks of the Rundemanen Formation and other clastic sequences in southern Norway, a correlation with the Late Proterozoic sparagmites seems the most likely. In particular, the presence of a diamictite, which may be of glaciogenic origin, makes this correlation attractive. If the correlation is correct, the Ramnslia Member could be linked to the Varangianice-age dated at c. 650 Ma ago (Pringle 1973), and the Bergen area added to the list of places where traces of this ice-age are recorded (Spencer 1975). Further, the possible presence of the sparagmites in the Bergen area supports the idea of a wide area of miogeoclinal sparagmite basin sedimentation as suggested by Nystuen(1982). A correlation with the sparagmites was first indicated by Reusch & Kolderup (1902) and later by Kolderup & Kolderup (1940), based on lithological similarities. Correlation with rocks of the Telemark Suite seems less likely, as Precambrian deformation, metamorphism, and intrusive and extrusive rocks are absent in the Rundemanen Formation. However, an Early Palaeozoic age, as indicated by Sturt & Thon (1978), or an age similar to rocks of the Ullensvang Group and comparable rocks in the Bergsdalen Nappes cannot be discounted.

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