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The geology of the Leirpollen area, Tanafjord, Finnmark

By

D. R. V. Beynon[1]), G. R. Chapman[1]), R. O. Ducharme[1]) and J. D. Roberts[1]).

Abstract.

900 metres of sedimentary rocks, including Eocambrian tillites, have been mapped on 1 : 50,000 and 1 : 25,000 scales around Leirpollen and correlated with the Vestertana Group which outcrops on the Digermul Peninsula. The rocks consist almost entirely of clastic sediments; volcanic rocks are absent and carbonates extremely rare. The Vestertana Group begins with two tillite formations which appear to be absent in the east. The upper part of the group consists of cleaved mudstones, siltstones, orthoquartzites and greywacke sandstones. Body fossils have not been found but trace fossils are present and are common in the upper part of the group.

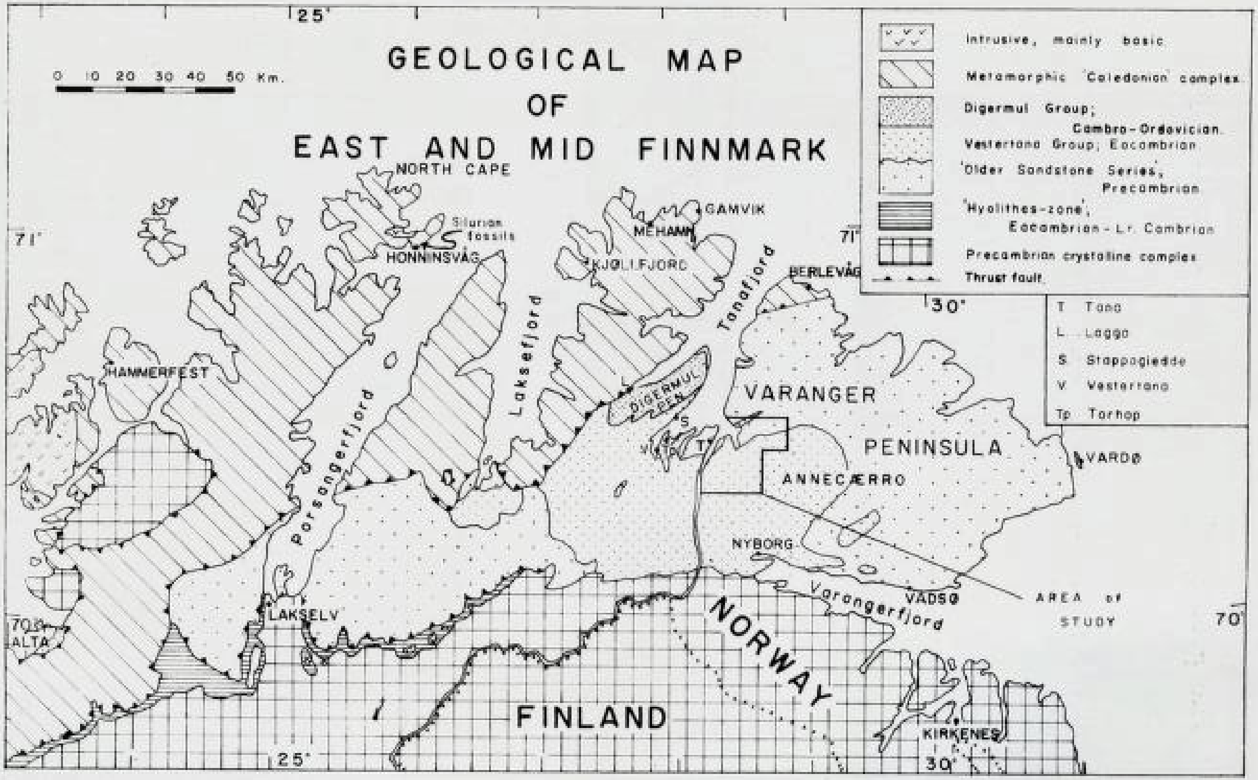
The rocks are folded with axes trending approximately SSW—NNE. The intensity of the folding decreases towards the east. In the extreme west of the area some overturned synclines occur and are associated with steeply-dipping reverse faults.

Introduction and geological background.

The Leirpollen area (Maps, next page and end of paper.) lies immediately the east of the mouth of the Tana River at the head of Tanafjord on the north coast of Finnmark, between longitudes 28°10′ and 29° east and between latitudes 70°15′ and 70°30′ north.

The area was first described by Holtedahl (1918) and the general features of the geology were elucidated. In 1933 and 1934 the western part of the area was visited by Føyn (1937) who established the presence of an uncontormity beneath the Lower Tillite. In 1933 the area had been mapped on the scale of 1 : 100,000 by Rosendahl (1945). Since 1950 five parties from Oxford University, under the direction of Dr. H. G. Reading, have visited East Finnmark, working mainly on the Digermul Peninsula. The stratigraphic sequence for the Digermul Peninsula was worked out in detail and in 1964 J. D. Collinson made a brief visit to the Leirpollen district. The purpose of this paper is to

[1]) Department of Geology and Mineralogy, Oxford, England.


GEOLOGICAL MAP
OF
EAST AND MID FINNMARK
0 10 20 30 40 50 Km.
25°
71°
70°
30°
Intrusive, mainly basic
Metamorphic 'Caledonian' complex
Digermul Group; Cambro-Ordovician
Vestertana Group; Eocambrian
'Older Sandstone Series', Precambrian
'Hyolithes-zone', Eocambrian - Lr. Cambrian
Precambrian crystalline complex
Thrust fault
T Tana
L Lagga
S Stappogiedde
V Vestertana
Tp Torhop
NORTH CAPE
Silurian fossils
HONNINSVÅG
GAMVIK
MEHAMN
KJØLLEFJORD
BERLEVÅG
HAMMERFEST
Porsangerfjord
Laksefjord
Tanafjord
DIGERMUL PEN.
VARANGER
PENINSULA
ANNECÆRRO
VARDØ
NYBORG
VADSØ
Varangerfjord
AREA of STUDY
LAKSELV
ALTA
NORWAY
FINLAND
KIRKENES

present the results of an expedition from Oxford which went to the Leirpollen area in the summer of 1965 with the aim of remapping the area using the detailed stratigraphical section obtained from Digermul.

1 : 50,000 scale A.M.S./M 711 maps were used, supplemented in the western part of the area by 1 : 25,000 scale aerial photographs.

The authors are indebted to Norges Geologiske Undersøkelse for the generous provision of a grant, maps and aerial photographs. to Dr. S. Føyn who suggested visiting the area and gave much practical assistance; and to Dr. H. G. Reading who provided advice and encouragement throughout. The authors would also like to acknowledge the friendly assistance given by the people of Leirpollen, in particular Olaf Henriksen and his son Øystein.

Stratigraphy.

The Older Sandstone Series is overlain with slight unconformity by the Vestertana Group (Eocambrian), which outcrops around the shores of Vestertanafjord and is there about 1450 metres thick. It passes conformably up into the Digermul Group which contains Cambrian and Tremadocian fossils (Reading, 1965). Within the Leirpollen area no rocks above the Lower Breivik Member of the Vestertana Group occur and the local thickness measured, from the base of the Lower Tillite Formation, was 900 metres. No body fossils were found but trace fossils were fairly common, particularly in the higher horizons.

Older Sandstone Series.

The uppermost 250 metres were studied in order to ascertain the nature of the unconformity beneath the Vestertana Group. The following succession is seen in the north-west of the area:

Vagge Quartzite	15 metres
Vagge Shale	90 metres
quartzite	over 200 metres

Towards the south-west the unconformity beneath the Lower Tillite Formation cuts down through the above succession and in the extreme south-west removes 130 metres of strata and rests on the quartzite below the Vagge Shale. An angular unconformity of 1—2 degrees was calculated, agreeing with that found by Føyn (1937).

Vestertana Group.

Lower Tillite Formation.

The thickness of the formation varies from 50 metres in the south-west to 8 metres in the north-west and it is apparently absent in the east. The rock is generally massive but stratification is sometimes present, taking the form of thin lenses of sand. It is unsorted, containing particles ranging from clay-grade to large cobbles. The bulk of the rock is composed of material finer than coarse sand, which is light-grey when fresh and weathers to dark-grey or brown. The coarser fraction consists of angular, sub-angular or rounded blocks of dolomite, metamorphic rocks, grey mudstone and quartzite. Dolomite makes over 50 per cent of the blocks which are frequently dissolved out leaving characteristic cavities. The Lower Tillite Formation is generally homogenous throughout its thickness but in the west of the area the formation contains a band of slates and mixed sediments 3—5 metres thick.

Nyborg Formation.

This formation is well exposed in the west of the area where it separates the two tillite formations. The thickness is commonly 15—20 metres, increasing to 90 metres in the south-west (Hanaelven valley). The formation thins towards the east but there is no accompanying facies change; it retains its character as a quiet-water sediment. This suggests that a pre-Upper Tillite Formation unconformity is responsible for the thinning.

The lower limit of the formation is taken at the abrupt disappearance of tillite blocks at the top of the Lower Tillite Formation. This is generally followed by a grey lithic sandstone 2-3 metres thick which grades quickly up into purple shales, but on the north-western slopes of Lammeskallfjell the Lower Tillite Formation is overlain by a dolomite 1.5 metres thick. The bulk of the formation consists of purple shales and lighter-coloured siltstones, the siltstones being commoner in the higher horizons. Immediately beneath the Upper Tillite Formation the beds are generally sandy laminated siltstones showing cross-lamination, but some massive lithic sandstones occur.

Upper Tillite Formation.

The Upper Tillite outcrops extensively in the western part of the area. It is commonly 9-15 metres thick with a maximum of 50 metres. It was not found in the east and is probably absent. It is an unsorted rock, so similar in most respects to the Lower Tillite Formation that it is difficult to distinguish them on lithology alone. Føyn (1937) states that in general the Upper Tillite Formation contains fewer dolomite and more crystalline blocks than the Lower Tillite Formation. In this area this distinction does not seem to hold. The blocks consist of dolomite, including pisolitic and striated varieties, with subordinate quartzite, gneiss, vein-quartz, mudstone, siltstone, chert, conglomerate, acid igneous rocks, and one block of pegmatite containing galena and pyrite was seen. Faint stratification, represented by the parallel orientation of small pebbles, is sometimes depressed under large blocks, suggesting that they have been dropped in from above. In some places there is evidence of reworking by currents.

Lateral variation of the succession in the east of the area

Along the northern margin of the main outcrop of the Vestertana Group, the Older Sandstone Series rocks form steep cliffs and the boundary between the Older Sandstone Series and the Vestertana Group is often obscured by quartzite screes. The bedrock is seen in only a few places.

West of Hanglefjell the two tillite formations are seen with the Nyborg Formation at the most only 15 metres thick. The succession in this part is thus similar to that seen in the west of the area near the Tana River but the units are essentially thinner and there is evidence of much tectonic crumpling of the beds.

5 kilometres to the east the following succession occurs:

4. Blue-green and red-violet slate member	over 200 m.
3. Purple mudstones (Nyborg Formation?)	15 m.
2. Laminated grey silty sandstone	8 m.
1. White quartzites of Older Sandstone Series	over 100 m.

The two tillite formations and possibly the Nyborg Formation appear to be absent. None of these formations was found at any place further to the east. Their absence could be explained by structural dislocations similar to the reverse faults in the west of the area. However, the present authors favour a stratigraphical explanation and believe the Lower Tillite and Nyborg Formations have been removed by a pre-Upper Tillite unconformity; the evident thinning of the tillites and Nyborg (the latter with no facies change) point to this conclusion. The Upper Tillite Formation itself, which sometimes shows shallow-water features, was eroded prior to the deposition of the Stappogiedde Formation due to the shallowing of the basin or the existence of land in this region immediately after the deposition of the Upper Tillite.

It is thought that the unconformity is only a local feature. In the west of the area the contact between Nyborg and Upper Tillite Formations is gradational, suggesting that there is no unconformity in this region.

Stappogiedde Formation.

Following the terminology of Reading (1965) this is divided into three members:

3. Red quartzitic sandstones with greywacke sandstones and mudstones.
2. Blue-green and red-violet slate.
1. Quartzitic sandstones*).

1. *Quartzitic sandstone member:* This outcrops frequently in the western part of the area but was not seen in the east. Its thickness varies irregularly from 5-55 metres. It consists of massive grey orthoquartzites, 2-8 metres thick, alternating with thin-bedded lithic sandstones and siltstones. Medium-scale cross-bedding is common. The upper part shows increasing amounts of fine material and grades up into the overlying member.

2. *Blue-green and red-violet slate member*: This is at least 250-300 metres thick. A thickness of 550 metres was calculated for the eastern part of the area but this apparent increase is probably due to undetected folding.

The lower 10-100 metres are dark purple mudstones, with, in the north, one band of white orthoquartzite 1.7 metres thick. Bands of light-green occur within the purple and are both parallel and transverse to the bedding, suggesting post-depositional reduction of the ferric to ferrous iron.

The upper part of the member consists of blue-green or grey mudstones containing laminae and thin lenses of siltstone. Ripples and channels occur, the channels being filled with muddy siltstones similar to the sediments they cut. The highest horizons become finer and purple mudstones again occur.

Red quartzitic sandstone member: This consists of three bands of red quartzitic sandstone separated by two bands of greywacke sandstones with mudstone alternations. The local thickness is between 230 and 250 metres.

3rd red quartzitic sandstone	10— 12 metres
Greywacke sandstones and mudstones	30— 40 metres
2nd red quartzitic sandstone	30— 40 metres
Greywacke sandstones and mudstones	45— 55 metres
1st red quartzitic sandstone	80—100 metres

The bases of the bands of greywacke sandstones and mudstones are abrupt. The greywacke sandstones occur as thin (10-14 cm.) graded beds with sharp bases, and are interbedded with silty mudstones. Horizontal and vertical burrows are present. The greywacke sandstones become thicker and the ratio of sandstones to mudstones increases towards the top of the bands. The transi-

*) The sandstone terminology of this paper follows Pettijohn (1957) with the addition that "quartzitic sandstone" is a bulk term including both orthoquartzites and lithic sandstones.

tion into the overlying red quartzitic sandstone is lithologically gradual although the actual colour-change is usually abrupt. However, on the west side of the Hanaelven valley there is at the base of the 1st red quartzitic sandstone a massive pale-grey orthoquartzite, 1.5-3.0 metres thick, overlying siltstones; this is found only in this valley and a fluviatile origin is suggested. Similar massive "white" quartzites occur more widely near the base of the second red quartzitic sandstone and at the top of the third, where a 3-4 metre bed forms a distinctive mapping horizon. In general the red quartzitic sandstone bands consist of red orthoquartzites and lithic sandstones with a small proportion of mica and decomposed felspar grains. Channels, medium-scale cross-bedding and ripple-marked surfaces are present.

The member thus shows a succession of coarsening-upward cycles. Each cycle begins with silty mudstones and greywacke sandstone alternations, the latter becoming thicker and more abundant upward. These pass gradually into the red quartzitic sandstone bands which contain little fine-grained material. There is a sharp base beneath the succeeding cycle.

Breivik Formation.

Rocks of this formation are stratigraphically the highest beds seen in the Leirpollen area. They consist of medium-grained lithic sandstones, orthoquartzites and greywacke sandstones, all interbedded with silty mudstones. The lithic sandstones and orthoquartzites occur as parallel-sided units 5-200 centimetres thick and showing cross-stratification and scoured bases. The greywacke sandstones occur as thin graded units 2-10 centrimetres thick with sharp, occasionally conglomeratic, bases and abundant horizontal burrows.

Reading (1965) has divided the formation into two members but probably only the lower member occurs here since the characteristic Upper Breivik Member lithology of the Digermul of thin mudstones and greywacke sandstones, without lithic sandstones and orthoquartzites, is absent. In addition the maximum observed thickness is only 230 metres, compared with 220-255 metres for the Lower Breivik Member on the Digermul.

Structure

The major structural elements consist of pitching folds with varying axial directions. In the south-west the axes trend NNE-SSW. These change to NE-SW in the east and north, while in the extreme north-east of the area the trend of the steeply dipping rocks adjacent to the Older Sandstone Series appears to be NW-SE. Intensity of the folding increases towards the west where folds are often asymmetrical, with axial planes dipping west. In the region of Annecaerro the beds are practically horizontal. Much minor folding occurs in the Stappogiedde Formation in the west of the area.

Three large reverse faults occur in the area of greatest deformation. They follow the trends of the folds and are thought to be contemporaneous with them. The two western examples, on the river-cliffs of the Tana and at Lavvonjargga may possibly be the same fault; in both these cases the Older Sandstone Series is brought up against overturned younger rocks. In all cases the fault-planes dip in a westerly direction. Small reverse faults are found in the noses of some minor folds, where sandstone beds have deformed by fracturing but adjacent fine-grained beds show plastic deformation.

Normal faults are fairly common throughout the area and generally have an E-W trend. However, in most cases the throws are too small to justify their representation on the map.

Flow cleavage, increasing westwards, was found to be parallel to the major and minor fold axes.

Geomorphology

In the north of the area the Older Sandstone Series rocks form high hills, up to 600 metres high, covered with a surface of bare rock and screes. The Vestertana Group underlies lower ground to the south, where an extensive plateau surface occurs at 340-380 metres, this is deeply dissected by wooded valleys. In the north-flowing Julelven tributaries there is a striking correlation between anticlines and stream valleys.

The whole area shows abundant evidence of the action of land-ice in the recent past; overdeepened valleys, with hanging tributaries, and large erratic blocks of tillite and crystalline rocks, the latter derived from the Precambrian basement in the south, are common. Drumlins are found on Hanglefjell. Many valleys on the plateau are dry, or only carry snow melt-water in the early summer.

Deep incision of valleys, terraces at many levels and raised beaches above the fjord appear to indicate that a large lowering of base-level was the most recent geological event in the area.

Conclusion

The Leirpollen area shows a succession of clastic sediments 1150 metres thick. The lowest 250 metres comprise the upper part of the Older Sandstone Series. The remainder is in the Vestertana Group, placed in the Eocambrian by Reading (1965) who, however discusses, (pp. 186-187) the possibility that the uppermost portion, the Breivik Formation, may be of Lower Cambrian age.

The Vestertana Group can be correlated in detail with the succession described by Reading (1965) from the Digermul Peninsula.

Two tillite formations were found in the area. NE of Leirpollen a thin typical Nyborg lithology was found between two tillite horizons. The tillite by the Leirpollen Bridge is considered to be the Upper Tillite. Two faulted tillite horizons are present in the extreme west of the area and in the extensive but hitherto unrecorded exposure in the upper Hanaelven valley. Reading (1965) accepts a marine glacial origin for the tillites on the evidence of lithology, lateral extent and constancy of the two tillite horizons. The lithological evidence from the Leirpollen area supports this view. However, in the east of the area unconformities beneath both tillites and removal of the Upper Tillite by a possible unconformity beneath the Stappogiedde Formation suggest that the margin of the basin lay in this area, and that erosion alternated with periods of deposition.

The Nyborg Formation (90 metres) is thinner than in the centre of the Tana district where a minimum of 200 metres was recorded, and in the Leirpollen area no turbidites were found in this formation.

The Stappogiedde Formation (500-600 metres) shows a considerable increase in thickness compared to the Digermul area where it totals 330-475 metres. The increase is in the two upper members of the formation. The basal member of the Stappogiedde Formation contains shallow-water quartzites and conglomerates, but quiet water conditions seem to have prevailed during the deposition of the blue-green and red-violet slate member. The topmost member of the formation contains three bands of red quartzitic sandstones with many shallow-water features, alternating with greywacke-sandstones and mudstones in a series of coarsening-upwards cycles.

The Breivik Formation, consisting of greywacke-sandstones, quartzites and mudstones outcrops extensively in shallow synclines underlying most of the high ground in the south.

The greywacke-sandstones in the Stappogiedde and Breivik formations may have been deposited by turbidity currents, but the closely associated orthoquartzites and lithic sandstones were probably deposited by traction currents. The basin of deposition was thus probably essentially shallow, but occasionally became deeper, allowing deposition of fine-grained sediments and turbidites. Current directions in the greywacke-sandstones of the Stappogiedde Formation show a south to north movement, but are more variable in the red quartzitic sandstones.

Trace-fossils are common in the higher formations. They first occur in the blue-green and red-violet slate member and increase in abundance upwards.

No body fossils were found. The oldest fossils found in the Digermul succession are of Lower Cambrian age. They occur in the Duolbasgaissa Formation of the Digermul Group, about 700 metres above the highest horizon mapped in the Leirpollen area.

The structural history is similar to that on the Digermul, but over the greater part of the area deformation is less intense. It comprises folding along NNE-SSW axes, with contemporaneous reverse-faulting or thrusting in the west. There was a later episode of minor normal faulting along E-W lines. Dating of these events is impossible due to the absence of younger rocks.

The investigation has shown that the map of Rosendahl (1945) is substantially correct. Nevertheless, there are important modifications: Rosendahl was of the opinion that only one tillite occurred at Leirpollen, but two were found. Only one red quartzite sandstone was recorded from the area by Rosendahl but three have now been proved as on the Digermul. The Breivik Formation outcrops more extensively than is shown on Rosendahl's map; the white quartzite detritus on Annecaerro is considered to be derived from this formation, and not from the Duolbasgaissa Formation as suggested by Rosendahl.

SAMMENDRAG

De geologiske forhold i Leirpollen-området, Tanafjord, Finnmark.

Berggrunnen i området omkring Leirpollen består av sandsteiner, morenekonglomerater (tillitter) og leirsteiner. Karbonat-bergarter (dolomitt) er svært sjeldne og det finnes ikke vulkanske bergarter i området. Lagrekken er i området i alt 1150 m tykk. Den er en del av de sedimentbergarter som forekommer på strekningen fra Varangerhalvøya over Tana, Laksefjordvidda og videre til SV for Porsangerfjorden.

Den laveste delen av lagrekken i Leirpollen-området, omkring 250 m, hører til den såkalte "eldre sandsteinslagrekke". Den hører i alder til yngste prekambrium. Mellom den eldre sandsteinslagrekke og den øvre del av lagrekken er det et brudd som representerer en erosjonsperiode.

Den øvre delen, som altså er omkring 900 m tykk, svarer til den del av lagrekken som H. G. Reading har kalt Vestertana-gruppen i sin beskrivelse av bergartene på Digermulhalvøya vest for Tanafjorden (1965). Vestertana-gruppen er inndelt i 5 formasjoner, nemlig (nedenfra og oppover): Undre tillitt, Nyborg-formasjonen, Øvre tillitt, Stappogiedde-formasjonen og Breivik-formasjonen. Av den sistnevnte formasjonen er omkring 230 m representert i Leirpollen-området, d.v.s. ca. halvparten av Breivik-formasjonen i Digermulhalvøya. Det geologiske tidsrom som Vestertana-gruppen hører til, kalles eokambrium, som kan oversettes med "kambriums demring".

Spor etter krypende eller gravende organismer i form av horisontale og vertikale rørformede dannelser i bergartslagene, er i Leirpollen-området funnet fra og med Stappogiedde-formasjonen og oppover. Sporene blir mer og mer tallrike oppover i lagrekken. Ingen rester av organismene selv, altså egentlige fossiler, ble funnet der.

De to tillittformasjonene inneholder blokker av grunnfjell og av dolomitt. En regner med at tillittene er dannet ved at blokkene er falt ned fra flytende is. Siden det er to tillittformasjoner, må en regne med to istidsperioder.

Mellom de to tillittformasjonene ligger Nyborg-formasjonen som i Leirpollen-området er opptil 90 m tykk. Den består av skifrig rød leirstein og moleir-stein. Det synes å ha vært et brudd i lagrekken mellom den øverste delen av Nyborg-formasjonen og øvre tillitt.

Stappogiedde-formasjonen begynner med 40 m kvartsittisk sandstein. Derover følger 300 m blågrønn og rødfiolett leirstein og moleir-stein med noen få sandsteinsbenker i. Det øverste ledd i Stappogiedde-formasjonen er omkring 240 m tykt, det består av en veksling av rød kvartsittisk sandstein og mer gråaktige sandsteiner og leirstein.

Av Breivik-formasjonen er som før nevnt bare den undre halvpart representert i området. Den er karakterisert ved en sterk veksling av kvartsitter og andre slags sandsteiner og av leirsteiner.

Bergartene i området er blitt foldet under den kaledonske fjellkj[illegible]e-dannelse. Foldenes retning er fra SV til NØ. Graden av deformasjon øke[illegible]mot V i området, der en kan se skjeve folder som er veltet over, og hvor [illegible] også har skjedd forkastninger med store oppskyvninger. Mindre forkastni[illegible]bevegelser langs linjer som går i retning V-Ø, har også forekommet.

Den endelige utformning av landskapet er skjedd under og etter [illegible]rtærtidens istider, altså i en geologisk sett meget sen periode av jordh[illegible]orien. Virkningen har vært til dels nedsliting av berggrunnen på grunn av ismassenes bevegelse, og dels en avleiring av løse jordlag — morenemateriale og elveavsetninger — som nå dekker berggrunnen i større eller mindre grad.

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Geological investigations in the Snåsa—Lurudal area, Nord-Trøndelag

By
David Roberts.

Abstract.

Metasediments and meta-volcanic rocks of supposed Middle Ordovician age overlying a basement complex consisting predominantly of granite-gneisses, are described from an area near the north-eastern end of the lake Snåsavatn. A generalised succession (in ascending order) of mica schists, phyllite, limestone and greenstones with greenschist and pyroclastics has been recognized. Observations of the basement gneisses and their relation to the mica schists indicate that while a local concordance of banding is common, on a regional scale a very slight discordance would appear to exist, the various evidence favouring this as being a primary feature.

A tripartite division of the main Caledonian deformation is recognisable and the various minor structures are briefly described. Minor folds of the second generation are found to be of considerable value in positioning the axial plane traces of associated major folds.

Introduction.

The area under consideration is situated on the SW flank of the Grong culmination at the northern extremity of the Trondheim region, in the tract of ground between the NE corner of Snåsavatn and the valley of the Luru river, an areal extent of some 180 km². Metasediments and meta-volcanic rocks of the Snåsa Group are preserved in a major synclinal structure, the so-called Snåsa syncline (Carstens 1956). A granite-gneiss basement complex forms the north-western limit of the mapped area while similar rocks occur as a wedge-shaped outcrop on Kolåsfjell widening north-eastwards beyond Lurudal (Plate I).

Although no work dealing specifically with this particular area has been published, information either directly or indirectly relating to the geology immediately NE of Snåsavatn can be found in the papers of Carstens (1955, 1956), Oftedahl (1955, 1964), Birkeland (1958) and Peacey (1964). Carstens'

(1956) paper contains a large-scale map of the iron-ore district from Snåsa to Stjørna, the map terminating in the Snåsa-Kolåsfjell area. The present area lies partly on Foslie's 1 : 100,000 map-sheet "Sanddøla", published by NGU (no description available), but largely on the 1 : 100,000 rectangle sheet Overhalla; only the eastern part of the latter has so far been mapped (S. Foslie and H. Carstens manuscript map, NGU archives).

Fieldwork for the present study was carried out for Norges Geologiske Undersøkelse (The Geological Survey of Norway) in the summer of 1964 at the suggestion of statsgeolog Fr. Chr. Wolff, and was intended as a contribution towards the compilation of the new map-sheet Grong, shortly to be published on a scale of 1 : 250,000 by NGU. Partly on account of bad weather and partly because time was spent in the Sanddøla valley area examining the Limingen-Sanddøla and Eastern Cambro-Silurian Series, field mapping was restricted to less than 8 weeks. Field expenses were very kindly defrayed by Norges Geologiske Undersøkelse. Prior to commencing the work, mapping problems were described to the writer in considerable detail by Dr. Janet Peacey: for this valued advice, and the constant help and support given by statsgeolog Wolff, the writer is extremely grateful.

Geological setting.

The northern limit of the Trondheim region of Cambro-Silurian eugeosynclinal sediments is marked by the east-west ridge of Pre-Cambrian gneissic rocks referred to as the Grong culmination (Oftedahl 1955) — this, in effect, is the ridge connecting Asklund's (1955) 'Olden-Anticline' and 'West Norwegian basement rocks'. Part of this basement ridge has been called the 'Olden nappe'.

To the north of the Grong culmination Lower Palæozoic eugeosynclinal sediments continue as the Nordland facies (Strand 1960, 1961), and are dissected by several thrust planes and nappes of which the Seve nappe is the most extensive. The Cambro-Silurian sediments of the Trondheim region themselves constitute part of this Seve nappe which can here be subdivided into the main Seve nappe and an upper nappe, the latter recognised by Peacey (1964) in the Tømmerås - Hegsjøfjell region. It is more than probable that the bulk of the Trondheim region metasediments belong to this upper nappe (Wolff 1967, Roberts 1967), since its thrust plane is traceable down to the south of this region; in this regard, Wolff has suggested that the name 'Trondheim nappe' be adopted for this allochthonous metasedimentary pile.

Metasediments and volcanic rocks of the Snåsa - Lurudal area belong exclusively to the main Seve nappe. Collectively they are referred to the Snåsa

Group, since it is possible to trace several of the lithologies of this group from the Tømmerås-Snåsavatn area described by Peacey (1964) around the closure of the Snåsa syncline into the present area. Significant facies changes are, however, apparent. Furthermore, as the sequence does not reach up to the basal conglomerate of the Upper Hovin Group (Carstens 1956, 1960), the rocks for the most part almost certainly belong to the regional Lower Hovin Group, largely of Middle Ordovician age. Fragments of gastropods found in the Snåsa limestone (Carstens 1956, 1960) would appear to confirm this view. The age of the mica schists below the limestone is uncertain; these rocks quite possibly extend down into the Lower Ordovician.

While the Snåsa syncline dominates the structural picture in this area, two other less prominent major folds are present, mappable largely but not entirely on minor structural evidence. Both these folds and the Snåsa syncline deform the regional schistosity or foliation. South-east of the present area two major folds of considerable magnitude — the Tømmerås anticline and Verdal synform — have been described by Peacey (1964), both deforming the prevalent foliation and in the latter case deforming a major early isocline in the Hegsjøfjell area.

The lithological succession.

Evidence is more or less lacking in this area for the establishment of a chrono-stratigraphy. Just off the map in the extreme south the Steinkjer conglomerate, the supposed base of the Lower Hovin Group, is at least 15 m thick at Navlus (Peacey, 1964) but loses its identity north-eastwards along the strike; at Agle a granule conglomerate or coarse grit is thought to represent this hori?on.

Difficulty also arises in tabulating an accurate litho-stratigraphical succession, primarily because of the facies variations encountered across the area. It is, nevertheless, possible to establish a generalized lithological succession. No sedimentary structures have been observed and consequently the sequence is based partly on tectonic structural evidence and partly on correlation with the successions recorded by Carstens (1956, 1960) and Peacey (1964). The generalized succession is as follows:

4. Greenschists, greenstones and hornblende schists.
3. The Snåsa and Kjennerås limestones.
2. Phyllite; with granule conglomerate in the south.
1. Mica schists.

1. The mica schists.

These rocks, regarded as the oldest member of the sequence, are exposed on either side of the Kolåsfjell ridge of basement granite-gneisses and leptites and also quite extensively in the north-west of the area striking NE-SW through Kultjern. While the north-western outcrop of mica schists is roughly lenticular in shape due both to folding and original sedimentary variation, those flanking the Kolåsfjell granite-gneiss are thickest in the east or north-east and appear to thin out westwards, eventually wedging out altogether so that greenschists come to lie adjacent to the gneisses.

Typically, this lithology is a grey or brownish grey, medium-grained biotite schist frequently containing garnet (and sometimes biotite) porphyroblasts and displaying a rusty-brown staining along schistosity planes. Garnets are usually small ($\leqslant$ 3 mm) and of variable abundance but examples have been found of rhombdodecahedra up to 1 cm across. Muscovite is infrequently present though certain horizons are muscovite-rich with a corresponding lack of garnet.

Hornblende schist bands are not uncommon particularly towards the top of the schist sequence where the boundary with the greenschist (where phyllite and marble are absent) is often quite graditional. In considering the strike extent of the mica schist from Troldvatn to Lurudal, a generalization is that the amount of amphibole decreases towards the north-east. At the same time garnet is generally more profuse in the east and north-east, but both this variation and that of the amphibole may be merely a reflection of original sedimentary character rather than metamorphic grade.

Thin graphitic phyllite bands sometimes weathering a sulphurous yellow colour are occasionally present in these schists. Bands of iron ore (pyrite and magnetite) occur sporadically but one 15 m thick ore horizon in the extreme north-east (Lurudal) is noteworthy. Other lithological variants include thin limestone and calcareous schist horizons E and NE of Kultjern and tuffitic or quartz-keratophyre bands — together with biotite-amphibole schist — in the vicinity of Kultjern. Tectonic inclusions of limestone are also distinguishable. In the basal part of the mica schists psammitic or quartz schist intercalations are present locally.

A notable feature of the schists, moreso in the north-west of the area, is their gradual induration and changing character towards the basement gneiss. The schist invariably becomes finer grained and massive or flaggy towards its base with micaceous leptite bands appearing, at first sporadically but becoming progressively more common, until the lithology is largely a fine-grained grey or pink-grey leptite or micaceous leptite. But even the more massive leptitic rock-type contains many intercalations of fine-grained biotite schist or amphibole schist, so that the field mapping of this transitional lithology can be somewhat frustrating. Although the precise origin of leptites and micaceous leptites has been the subject of much debate, the field evidence in this area would appear to point to a sedimentary derivation for at least the micaceous leptites. This is in agreeance with Peacey's (1964) suggested origin for the micaceous leptites of the Tømmerås area.

Quartz veinlets and segregations, often boudined, are ubiquitous in these mica schists throughout the northern and north-eastern parts of the area. These are invarably parallel to the dominant schistosity but are deformed by later folds. It is thought probable that much of this quartz is of metamorphic segregatory origin and as such is of fairly local derivation. On the northern border of the Kolåsfjell granite-gneiss, pegmatitic and granitic streaks, segregations and diffuse lenticles are present in the schist, again paralleling the recognisable banding and schistosity but deformed by second generation structures, such that an origin concomitant with the main foliation and metamorphism seems evident. Moreover, the greater part of this granitic material appears to be of replacive origin.

2. *Phyllite.*

This member of the sequence crops out in the south-eastern part of the area both N and NE of Sjysjøen and farther south beneath the Snåsa limestone in the Agle district, on the northern and southern limbs of the Snåsa syncline respectively. On Foslie's "Sanddøla" map the southern outcrop of phyllite extends north-eastwards from Agle for some 13 km until the pattern of outcrop suggests the presence of a major fold closure. The phyllite outcrop is then severed by an apparent tectonic break (trending ca. 060-065°), re-appearing some 4 km to the south-west on the north side of this line. The present writer, agreeing with Peacey's (1964) assumption, regards this line of discontinuity as a fault.

As a distinctive lithology the phyllite is a grey biotite-muscovite phyllite frequently containing conspicuous pyrite with or without magnetite. It is sometimes of greenish grey colour moreso where it grades into the typical greenschist. Quite often, where the grain size is slightly larger, it is best described as a phyllitic schist, but all gradations from phyllite into the underlying mica schist can be found. On the south side of the Snåsa syncline, the phyllite is generally darker grey and graphitic; thin limestone and quartzitic ribs are sometimes present.

At Agle, some 30-50 m below the base of the Snåsa limestone, the phyllite contains a distinctive schistose granule conglomerate. Difficulty arises in choosing a name for this rock-type, moreso as particle size varies across the strike. Where particles, closely packed and constituting the bulk of the rock, are always less than 3 mm across, it can be described as a schistose grit but such a size restriction is uncommon since particles in many bands measure up to 5 or 6 mm and sporadic pebbles of greenstone up to 3 cm are present. In view of this variation a middle course has been chosen and Twenhofel's (1950) term 'granule conglomerate' adopted for the lithology of this horizon.

Particles consist mainly of a blue or bluish feldspar, white quartz and pale grey quartzite with subordinate fragments of greenstone, greenschist and, rarely, jasper. Larger pebbles always appear to be of greenstone or a similar schistose amphibolitic rock-type. This is very interesting since it indicates that the sequence from which the pebbles were derived - the Støren Group - was probably somewhat metamorphosed prior to the deposition of the conglomerate. The matrix is normally a dark phyllite or schist. Although a three-dimensional study of the various particles was not attempted on account of the poverty of exposure, a stretching direction is perceptible which is oblique to the trend of local minor folds. These small folds, varying from microfolds up to structures of 3 m wavelength and amplitude, deform the regional schistosity and also appear to post-date the particle lineation: the fold axes plunge towards 260°-270° whereas the lineation of small fragments is towards 050°-060°. On the other hand, a transverse section though these minor folds often shows ca. 60-70 % of particles flattened parallel to the axial planes — yet the orientation of particle c-axes does not accord with that of the fold axes.

Nearer the boundary with the main limestone in this same small area, a calcareous schist is found to contain intercalated psammitic ribs which have frequently been sheared and dissected into lenticular or discoidal fragments. At times this lithology resembles a highly tectonized conglomerate though it is clearly not of primary origin. Some 4 km further NE along the strike, due south from Sjysjøen, a calc-phyllitic schist contains many thin quartzitic ribs, there only partially disrupted by shearing.

Returning to the granule conglomerate, it is almost certain that this lithology is the strike continuation of the Steinkjer conglomerate which is so well-developed further south-west. A gradual thinning, together with important facies changes accounts for the character of the lithology at Agle (here more of a coarse grit than a true conglomerate).

About 2 km north-east from the Agle occurrence along the same stratigraphical horizon, a rather poor exposure of gravelly phyllite has been observed. The particles here are of quartzite and greenstone. Many pseudo-pebbles of quartz are present, these having originated tectonically by shearing and rotation.

This above-mentioned gravelly phyllite would appear to be the last vestige of the Steinkjer conglomerate when the latter is traced north-eastwards on the southern limb of the Snåsa syncline. No lithology resembling either a true conglomerate or a granule conglomerate as that described from Agle has been found on the northern limb of the syncline. However, two localities situated some 800 m apart just below the limestone in the vicinity of Kolåstjern display interesting features which are suggestive of a possible correlation with the Steinkjer conglomerate horizon. Both these exposures are of greenish grey schist which contains scattered but fairly abundant drawn-out fragments, granules or very small pebbles of either greenstone or a pale grey psammitic rock-type: these are up to 5 mm across. While the phyllitic member of the sequence is absent hereabouts it is significant that the stratigraphical position of these granule-bearing schists corresponds quite favourably with that of the Steinkjer conglomerate. Considering the facies variations inherent in this conglomeratic horizon on the southern limb of the Snåsa syncline, correlation of the Kolåstjern granule-bearing schist with the Steinkjer conglomerate is therefore, not too improbable a suggestion.

3. The limestones.

The Snåsa limestone occurs on the southern border of the mapped area trending ENE towards Agle and becoming progressively thinner. Beyond Agle this limestone thins gradually; on Foslie's "Sanddøla" map it disappears due east of Sjysjøen.

A second major limestone outcrop is that in the vicinity of the farm Kjennerås, south-east of Kultjern. The rapidly varying thickness of this limestone is partly primary but largely ascribed to tectonic causes. Whereas to the south-west of Kjennerås this limestone extends into Aadalen beyond Troldvatn, it thins out quite rapidly north-eastwards. Other thinner bands of limestone (up to 60 m thick) occur prominently within the greenstone-greenschist sequence.

Since the position of the Kjennerås limestone in relation to the mica schist and greenstone-greenschist members of the sequence is more or less identical to that of the Snåsa limestone in the Snåsa-Agle area, correlation of these two limestones seems highly probable and this assumption is supported by lithological similarities. Both are fairly well-banded, blue-grey, recrystallized limestones often containing thin dark grey or dark blue stringers of graphitic phyllite giving the rock a striped appearance. Finely disseminated pyrite may be observed. The boundaries of the limestone are frequently gradational into the adjacent schists, either through progressive increase of pelite content or alternations of limestone and schist which cannot be indicated on the present map. Within relatively short distances along the strike, lithological variations are perceptible, these clearly being of a primary nature. Near Kjennerås farm the limestone is locally coarsely crystalline and poorly banded.

Lithological facies variations are indeed of appreciable importance in this part of Nord-Trøndelag, not least in a consideration of the geology of this Snåsa - Lurudal area. Peacey's (1964) remarks concerning the Snåsa limestone can perhaps be quoted here — "... the limestone itself is certainly diachronous since near Kvam it occupies the whole span of the Lower Hovin Group as delimited by the Steinkjer and the polygenous Middle Ordovician conglomerates, whilst north-east along the strike it thins to nothing and its place is taken by greenschists and amphibolites".

4. The greenschists, greenstones and hornblende schists.

This member of the sequence constitutes the greater part of the Snåsa synclinal basin, also extending north-eastwards to beyond Flåtjern. The lithology is, for the most part, a green or pale greyish-green, poorly schistose rock though with a pronounced linear element always noticeable in the field. Where a schistosity is pervasive the rock-type can be referred to as a greenschist; otherwise, greenstone is the accepted terminology. At times the lithology is massive and essentially a tuffitic greenstone but all transitions to quartz keratophyres, rhyolite tuffs and keratophyre - agglomerate appear to be present, though not common. On the map (Plate I) only the more prominent tuffaceous or keratophyric bands have been indicated.

Towards Troldvatn and in a belt north-eastwards to beyond Flåtjern the lithology is more of a hornblende schist or mixed hornblende schist - greenschist than the typical greenstone - greenschist. Hornblende is certainly the predominant porphyroblastic mineral in this northern area and intercalated bands of hornblende-garbenschiefer are mappable with amphibole up to 16 cm in length though generally $\leqslant$ 10 cm. It is noteworthy that these garbenschiefer horizons, many of which have had to be omitted from the map, invariably occur close to limestone bands. In the area between the two thin but extensive limestone bands in the central part of the area and the wedge of basement granite-gneiss, hornblende schist is very abundant and thin impersistent limestone ribs are not uncommon. Limestone ribs, and occasionally bands up to several metres thickness, are also demonstrable in the Snåsavatn area, just above the main Snåsa limestone.

Pyrite and magnetite are common minerals throughout the greenstone-greenschist sequence, sometimes in segregations or thin ore-bodies but often as disseminated crystals showing excellent cubic (pyrite) and octahedral (magnetite) form. Magnetite is usually present in quartzitic bands. Detailed descriptions of the ore can be found in Carstens' 1956 paper. Epidote and quartz-epidote-calcite segregations are also present in this greenstone-greenschist; epidote veinlets occur locally. West of Kolåstjern the greenschist immediately above the main limestone band contains abundant biotite porphyroblasts, some up to 5 mm across.

Just to the west of Flåtjern a finely banded tuff or keratophyre can be seen to change laterally into a rubbly tuffitic greenstone and then into a conspicuous agglomerate. This latter rock-type, and to a lesser extent the adjacent massive rhyolite tuff, is densely net-veined with quartz and felsitic veinlets (some of only 1 mm thickness) which stand out as ribs on the white-weathered surface. Disoriented schist fragments up to 2·5 cm in length are present within this pyroclastic lithology while yellow-weathering concentrations of pyrite are not infrequent.

Evidence as to the origin of the bulk of the greenstone tends to be masked by the metamorphism, but tuffs, tuffaceous greenstones, rhyolitic tuffs and keratophyre-agglomerates leave no doubt as to their volcanic derivation. The typical greenschist or partially schistose greenstone is less easily accounted for. Occurrences are found of transitions from limestone through calc-schists to greenschists and even tuffaceous or keratophyric greenstone so that a sedimentary origin would here appear incontrovertible. The bulk of the greenschists and some of the poorly schistose greenstones therefore probably represent reworked lava detritus. The iron-ore bands are also almost certainly of sedimentary origin, but the association of pyrite with the pyroclastic material near Flåtjern leads one to consider the possibility that gas exhalations connected with the volcanism may have supplied some of the elements constituting the ore-mineral segregations, as suggested by Oftedahl (1958).

The amphibolite situated within mica schists north of the Kolåsfjell granite gneiss is regarded as a meta-gabbro. In the central parts of the body it is typically a black to dark green, coarse-grained, garnetiferous meta-gabbro: garnets up to 5 mm across are quite prominent. Towards its margins this basic sheet is more of a fine-grained schistose amphibolite, almost a true schist in part, containing abundant small garnets.

South of Kolåsfjell, and again within mica schist, another amphibolite sheet is present. This is broadly similar to that occurring north of the granite-gneiss except that garnets are now quite uncommon. Moreover this amphibolite is, at best, less coarse-grained than the northern one and tends to be more schistose. Despite these differences it would appear that the two amphibolites are related and quite feasibly represent segments of an originally more extensive and possibly continuous gabbroic intrusive sheet. The structural evidence for this point of view is presented later.

The basement complex.

Rocks of this group, which underlie the metasediments and volcanics of the Snåsa Series, are largely of granitic composition and have been regarded by various authors as being of Pre-Cambrian age. As can be seen from the map (Plate I), basement rocks occur in two separate areas. On tracing these occurrences in a general easterly direction onto Foslie's Sanddøla map-sheet, they are found to link up into one extensive area of 'gneissic granite'. This broad E-W belt of granitic rocks forms part of the Grong culmination and while the larger part of this belongs to the so-called Olden nappe, a marginal zone is referable to the Seve nappe.

On the present Snåsa - Lurudal map no distinction is made between rocks of these two tectonic units. In the time available, and since investigations were purposely and basically concerned with the Cambro-Silurian sequence, little work has been attempted on the basement rocks.

Granite-gneiss is the most widespread rock-type occurring in this basement complex. Typically it is a pink or greyish pink, medium- to coarse-grained gneiss of granitic character. Feldspars are often prominent as porphyroblasts with biotite as the chief mafic mineral. Some zones of granite-gneiss are muscovite-rich and are correspondingly whiter in colour.

Grain size is sometimes observed to be extremely variable. Where the lithology is of finer grain, it is often difficult to differentiate between a gneiss and a leptite, moreso where gradations occur.

Augen gneiss is notable only to the north-east of Kolåsfjell in Lurudalen. In one locality feldspars (chiefly microcline) up to 6 cm in length were observed but this is quite exceptional as most porphyroblasts rarely exceed 1·5-2 cm. Growth of secondary feldspar tends to have been quite irregular in distribution and porphyroblastic feldspars also occur sporadically in the schists adjacent to the basement rocks.

Leptite is here taken to mean a pinkish grey, fine-grained (rarely medium-grained), banded or massive rock, consisting mainly of feldspar and quartz with subordinate amounts of micas, epidote, garnet, sphene and zircon. It may be rather flaggy and is generally brittle and closely fractured. Wafer-thin biotitic stringers are sometimes present, particularly nearer the overlying schists. Micaceous leptites are also distinguishable, and gradations both into true leptites and mica schists have already been described. Furthermore transitions into gneissic granite are recognisable.

Nowhere throughout the area could any sharp contact be traced between either granite gneiss and leptite or leptite and mica schist; a transitional boundary is invariably present. In a railway cutting some 2.6 km north of Lurudal Station the contact between basement rocks and mica schists is excellently exposed. As far as dips and strikes are concerned the sequence is conformable, the mica schists being separated from the leptite or leptitic gneiss by a zone of augen gneiss. The full sequence shows a fine- or medium-grained granite-gneiss becoming perceptibly more leptitic towards the north; this is followed by a 30-35 metre thick zone of leptitic augen gneiss with feldspars up to 1 cm across. North of this, for 3-5 metres, feldspar augen become smaller and more scattered, the bulk of the rock now of darker colour and finer grain and more of a true biotite schist: then follows the ordinary biotite schist but this is also found to contain sporadic small feldspar porphyroblasts over a distance of at least 40 m — the exposure is then discontinued. Thin leptite bands and intercalations of hornblende schist are present in the mica schist at this locality.

Without a thorough petrological examination of the leptites it is difficult to comment on their precise origin. Based on field criteria alone — the gradations into micaceous leptites and schists, local rapid alternation with indubitable metasediments and fairly regular 'stratification' parallel to that in the sediments — a sedimentary origin would not appear improbable. Peacey (1964), while putting forward arguments for a possible volcanic interpretation, leaves the question of their origin quite open. At this stage the present writer also feels inclined to remain uncommitted on this question.

The basement contact.

The basement contact — the contact between the basement rocks and the overlying Cambro-Silurian metasediments — is rarely exposed well enough for an appraisal of its precise nature. Commonly the critical few metres of contact are not sufficiently well exposed, or else a gradational boundary appears to exist which quite often may be partially masked by secondary effects.

The earlier mentioned section exposed along the railway line north of Lurudal provides the only really clear-cut profile displaying the boundary relationships in this area. There, it will be recalled, the sequence from schist to granite-gneiss is conformable but an augen gneiss separates these rocks: feldspar porphyroblastesis is also pervasive in the schists hereabouts, so that the original nature of the contact cannot be stated with any certainty. It is noteworthy that augen gneiss occurs only in this Lurudal area on the northern side of the Kolåsfjell basement complex. In the coarsest type of augen gneiss, the groundmass between the feldspar augen has an appearance not unlike some mylonitic lithologies. From this it is possible to suggest that the present concordance between basement and cover in this particular small area is secondary. It is necessary, however, to consider other points before generalizing on this important relationship.

Partially transitional boundaries between schist and leptite and leptite and granite-gneiss have been mentioned previously. These are particularly wide in the W and NW of the area but are observed elsewhere on a lesser scale. Sharp contacts have not been found though the boundary may be located within a distance of two or three metres. Despite this apparent concordance measurements of banding in both schist and gneiss (or leptite-gneiss) in any one locality adjacent to this narrow contact show interesting discrepancies. These are quite systematic and cannot therefore be dismissed as errors of measurement. Seven such localities show that while strikes are broadly comparable in schist and basement, dips of banding in the basement granitic rocks are always steeper than in the stratigraphically overlying metasediments by anything up to 12°. Only in the rail-cut locality north of Lurudal Station is there an absolute concordance of banding. Considering the profile A-B (Fig. 3) across the area, if the major folds responsible for the present pattern of outcrops are unrolled until the metasediment banding is horizontal, the banding in the gneisses is found to dip gently and quite constantly in a south-easterly direction, thus suggesting the presence of a slight, regional, primary unconformity. It must be stressed that this slight discontinuity is a regional feature — it has not been possible to place ones finger on any angular unconformity in the field. On studying the map (Plate I) the wedging out of the mica schist both on the north and the south sides of the Kolåsfjell basement area could also be taken as evidence of a large-scale unconformable relationship between basement and cover.

In recent years a certain amount of controversy has surrounded the question of the nature of the basement contact in the Trondheim region; the papers of Birkeland (1958), Oftedahl (1964 and 1965), Peacey (1964) and Holmsen (1965) provide a fair range of views on this subject. While not wishing to take sides on this matter, at least not at this stage, the present writer can only point to the observations from this comparatively small Snåsa-Lurudal area, and the conclusions reached from the investigation cannot be said to hold for other parts of the Central Norwegian Caledonides. In summary, the salient features of the basement contact in this area are that, firstly, there is little evidence (except possibly in a small area NNW of Lurudal Station) for the existence of a general pseudo-conformity, i.e. a Caledonian tectonized contact.

Secondly, while the basement-cover boundary is concordant on outcrop scale, measurements point to a very slight discordance. This, together with the larger scale Kolåsfjell gneiss-schist relationship as seen on the map, would suggest that on a regional scale a very slight unconformity exists, the implication being that the basement rocks were tilted very slightly (in a general south-easterly direction) but unfolded at the time of deposition of the sediments and volcanics of the Snåsa Group.

These conclusions would appear to confirm, to a large extent, the findings of Peacey (1964) from the Tømmerås area south of Snåsavatn. Moreover they are broadly consistent with Oftedahl's (1964) opinion that "the Pre-cambrian rocks were essentially flat-lying und unfolded" at the commencement of Eocambrian or Cambrian sedimentation, but the present writer does not subscribe to the view of the basement surface being perfectly horizontal. Considering both Peacey's conclusions from Tømmerås and the present investigations, a gently undulating Pre-Cambrian surface is the more likely case. A final point is that in stating these conclusions the writer regards them as applicable only to the extreme northern part of the Trondheim region. To the south and west of this extensive region, it is known that prior to the Eocambrian - Silurian sedimentation the basement gneisses were subjected to a complex sequence of deformation, intrusion and metasomatism (see e.g. Banham and Elliot 1965). With regard to the nature of the basement and its boundary with the cover rocks it is clearly unsafe to generalise over an area as broad as that of the Trondheim region.

Structure.

The major structure, the Snåsa syncline, can be followed far beyond the limits of this small area, particularly in a south-westerly direction. To the north-east, some 12-13 km NE of the lake Sjysjøen to be exact, the closure of this fold can be readily identified on Foslie's "Sanddøla" map. A brief description of the structural features of this Snåsa syncline closure area have been given by Peacey (1964, pp 78-80) and Fig. 34 of that paper illustrates the general situation, and the relationship of the Snåsa syncline to other major folds, quite adequately.

In the Snåsa - Lurudal area the Snåsa syncline is an asymmetrical structure the axis of which plunges in a general WSW direction. This is reflected in an overall convergence of lithologies towards the ENE. The synclinal axial plane, the trace of which is depicted on Fig. 1, dips steeply in a NNW direction — the northern limb of the fold dips steeply, at times near vertically, to the SSE whereas the southern limb is generally inclined at a moderate angle

(30° - 50°) towards the NNW. Minor folds, clearly related and congruous to the main syncline, show variable axial plunges generally WSW but locally horizontal or ENE.

Before considering the structures in the northern part of the area it is important here to note that the Snåsa syncline, and its parasitic minor folds and phyllitic crinkles, deforms the main schistosity which is present in almost all the metasediments. Rarely, in calc-silicate schists, tight or isoclinal folds are found, the pervasive schistosity being axial planar to these folds. In several places, minor folds related to the Snåsa syncline (or other major folds occurring further north) deform an earlier linear element. From this it will be appreciated that the deformation which produced the Snåsa syncline was not the first to affect these metasediments.

The northern half of the area is characterised by steep-dipping metasediments and granite-gneisses, but it is nevertheless possible to demonstrate the existence of a further two major structures with the aid of minor structural evidence. The disposition of the various metasediments and basement rocks to the SW of Kolåsfjell would appear to indicate the presence of an anticlinal structure complementary to the Snåsa syncline. Although no indubitable fold hinges were discernible, a study of the minor folds and their relative vergence reveals the presence of a tight fold closing upwards (Fig. 2), thus confirming the stratigraphical indications. South of the axial surface trace of this fold, minor folds deforming the schistosity are overturned towards the NNW — this is also the northern limb of the Snåsa syncline. Immediately north of the axial trace the direction of fold overturning is reversed, but folds become inconspicuous away from the hinge zone of this anticline.

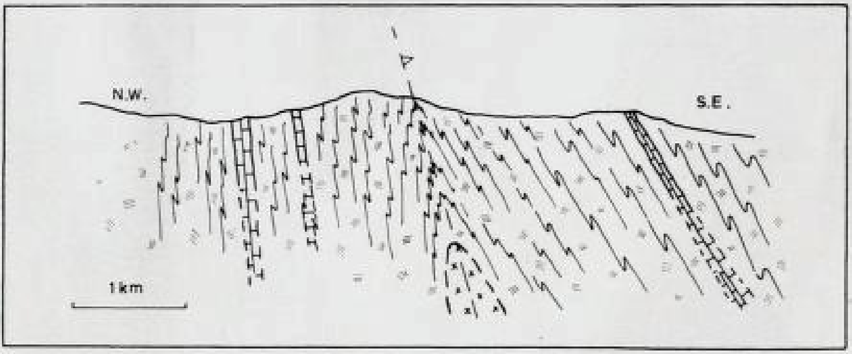

Fig. 2. Diagrammatic profile (SE of Kjennerås) depicting anticline axial plane located by congruous minor folds.

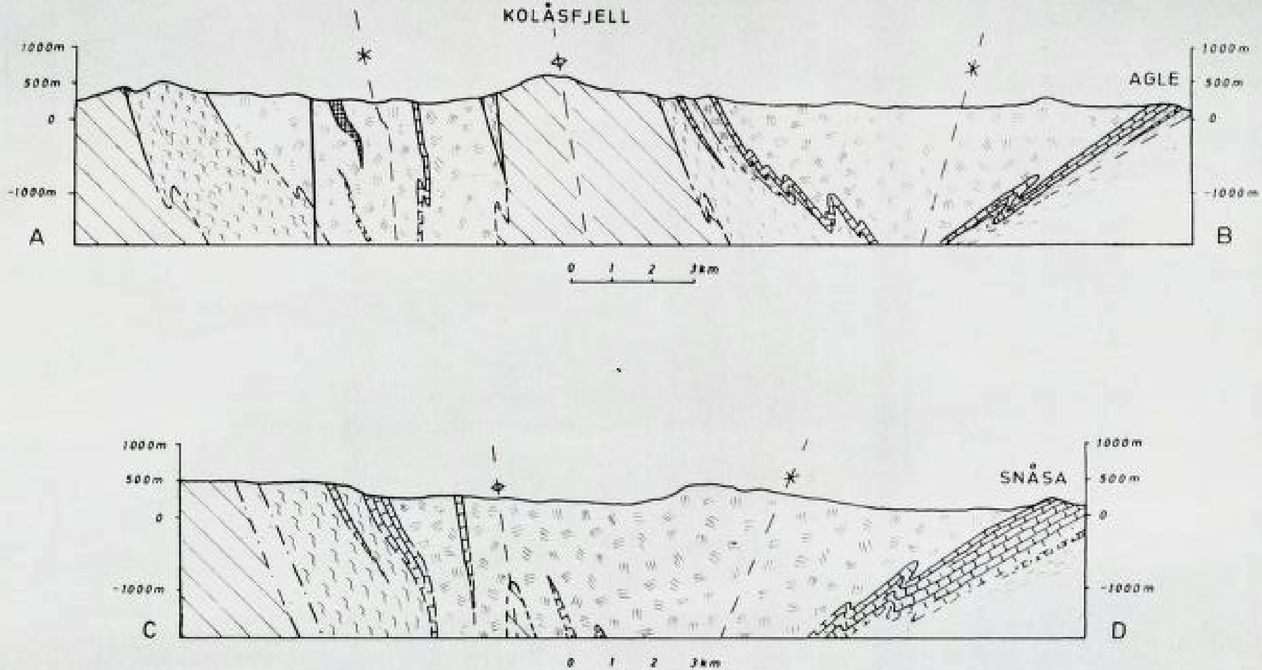

Fig. 3. Geological sections across the area.

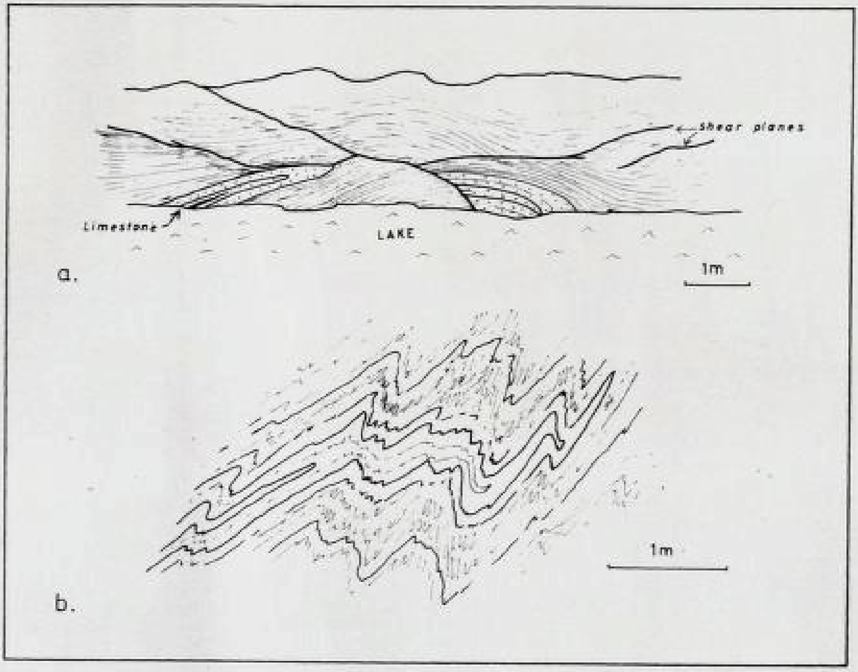

Fig. 4. Early (first episode) structures. (a) Sheared isocline with third phase warps. Calcareous greenschist and limestone, NE shore of Snåsavatn. (b) Isocline deformed by second phase folds. Limestone, Troldvaselven.

Further to the north-west minor structural indications are that another fold, this time synclinal, is present within the greenstone-greenschist sequence (Figs 1 and 3) the axial surface trace extending north-eastwards beyond Flåtjern and separating the two outcrops of mica schist in the extreme NE part of the area. This again supports the stratigraphical evidence. To the south-west, west of the Bruvoldelven valley, it has not been possible to trace this fold.

The changing style of the regional major folds is rather interesting; in the northern part of the area the 'Flåtjern syncline' is a very tight structure while the anticline further south is only slightly less acute. In comparison the Snåsa syncline is a relatively open structure. This trend — of progressively more open style to the S or SE — is continued if the Tømmerås anticline is brought into the picture (Peacey 1964).

Though a systematic analysis of minor structures and lineations was not possible in the time available, their observation has shown that three main episodes of deformation have affected the rocks in this area. Faulting may be counted as a possible fourth phase of the deformation history.

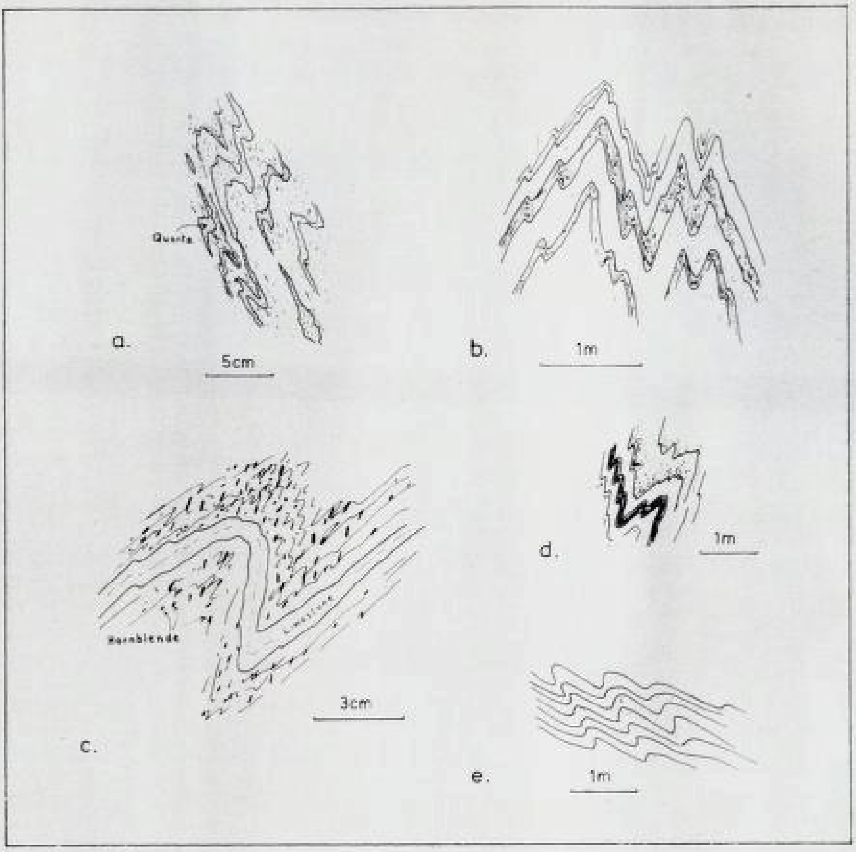

Fig. 5. Second episode structures. (a) Mica schist with vein quartz. (b) Interbanded leptite and schist. (c) Hornblende schist with limestone ribs. (d) Greenschist with psammite bands. (e) Greenschist.

The most abundant structure representative of the first episode of deformation is the schistosity or foliation displayed by metasediments, volcanic rocks and gneisses. Folds to which this folation is axial planar are quite uncommon, tending to be restricted to interbanded limestone - calc-silicate schist and some greenschist lithologies (Fig. 4). Where present they are essentially of isoclinal type. Boudinage and stretching phenomena constitute an associated linear element as does the alignment of small granules and pebbles in the conglomerate within the phyllite member of the succession. Major first episode folds have not been recognised.

Second episode folds and lineations are prominent over most of the area. Minor

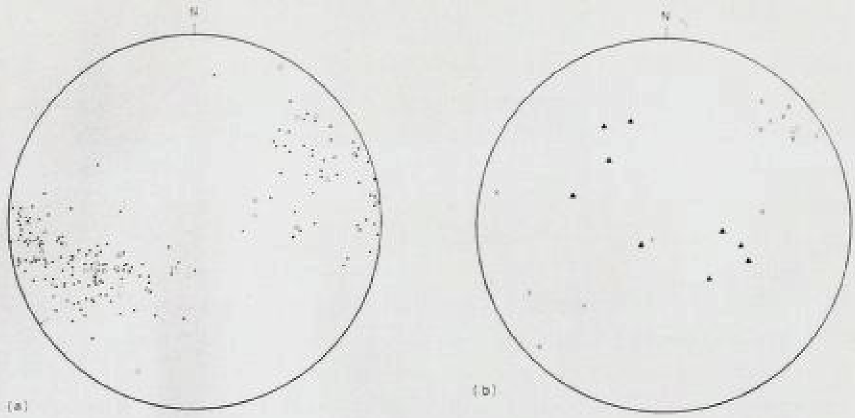

Fig. 6. Stereographic projections of linear elements (Wulff net, lower hemisphere). (a) Second episode structures; dots — folds axes and phyllitic lineation; open triangles — mineral lineation; circles — lineation in basement rocks. (b) First and third episode structures; crosses — first episode fold axes and lineations; ellipse — pebble elongation (first episode); full triangles — third episode fold axes.

folds vary considerably in style, a variation which is only partially dependent on lithology, since in the south these folds tend to be less acute than in similar greenschist farther to the N-NW. In the limestones a maximum style variation is observed, from quite open to near-isoclinal structures.

Minor folds axes and lineations belonging to this generation are depicted on the stereogram, Fig. 6. The variation of trend is largely a reflection of the attitudes of the major structures of this second episode and only to a minimal extent by later deformation. In the south, for example, both in the hinge zone and on the southern limb of the Snåsa syncline, minor fold axes generally plunge at low angles to the W-WSW; locally however, plunges are to the east. Moving N and NW away from the axial trace of the syncline the angles of plunge of these minor folds steepen until, in the closure zone of the adjacent anticline, they are in the range 30° - 50°. At the same time the direction of plunge moves round close to SW.

Within this same anticlinal hinge zone away from the Kolåsfjell gneiss, second episode minor folds and lineations plunge less steeply and towards Troldvatn the direction of plunge is often ENE. An ENE fold plunge is also common in the mica schist and limestone in the north-western part of the map area. It would thus appear that the steeper lineations and fold axes in the hinge zone of the 'Kolåsfjell anticline' just to the SW of the wedge of gneiss are a consequence of granite-gneiss acting as some kind of buffer during the deformation, so resulting in a deviation of linear elements developing in the less competent metasediments.

In the Flåtjern area, lineations of this generation are quite irregular in trend swinging round to N-S and then reverting to the general WSW plunge in the extreme north-east of the area. While this deflection appears to be due to a later episode of deformation, impression of linear elements on a pre-existing irregularity cannot be entirely ruled out.

Boudinage is also found associated with second episode folds. In some places two orthogonal directions of boudinage may be observed, moreso where conjugate shear planes disrupt the picture: this latter case, with boudins aligned in 'a', is not uncommon in the greenstone-greenschist lithology. Fold mullions are demonstrable in the mixed leptite-schist lithology west of Kultjern — these are parallel to the local second phase fold axes.

Acicular hornblende frequently displays a preferred lineation paralleling the second fold axes, but cases of two amphibole lineations in the same rock have been observed in the hornblende schists of the Kolåsfjell anticline hinge zone. Where this occurs, the earlier lineation is only weakly developed and is quite oblique to the prominent later element.

A crude linear element, essentially of quartzo-feldspathic material, is manifest in the granite-gneisses: this appears to be parallel to the second episode lineation developed in neighbouring metasediments.

Third episode structures are relatively uncommon, at least as minor folds. Where recognisable they are quite open folds or warps, although locally in phyllitic lithologies they may be represented by kink folds or strain-slip cleavage.

Distortions of earlier lineations on a large scale may also be attributed to this deformation phase. In general the trend of these late warps or folds is somewhere between NW-SE and WNW-ESE and is, therefore, more or less normal to the dominant second phase lineation.

Faulting appears to be a relatively insignificant feature over this small area. The few faults present show a marked NE-SW trend, although the major fault in the south and south-east varies from ENE to NE. This fault may quite feasibly be an extension of the major strike-fault present just north of Snåsavatn (Peacey 1964), but further investigations are needed before a definite opinion can be voiced.

Since the faults are representative of a notably brittle deformation, they are probably largely of fairly late development. As several of them are strike faults or oblique faults, they could be envisaged as developing simultaneously with the upheaval of the Grong culmination, itself a late structural feature (Oftedahl 1955). Downthrows along these faults are noticeably to the SE or SSE, an observation which would seem to accord well with Oftedahl's postulate of a late upheaval of the basement to the north of this Snåsa area. On the other hand, minor slides in the limestones, associated with the second episode of folding, are often seen to develop into faults along the same dislocation. These faults sometimes exhibit features indicative of horizontal displacement. While most faults over the area appear to be normal, the major fault in the south which displaces the Snåsa syncline axial trace may have an additional tear component, but the evidence for this is indistinct.

The penetrative schistosity or foliation seen in all lithologies in this area has been shown to have developed concomitantly with the first folding. Although this would appear to restrict the main metamorphism and recrystallization to this deformation phase, the picture is not so straightforward. The second generation structures deform the foliation but they also deform quartz veins and segregations and granitic material locally pervading the mica schist which can be shown to post-date the first folds and foliation.

Vein and segregatory quartz, while usually paralleling the banding of schistosity, is not infrequently seen to transsect these S-planes. It is however strongly deformed, often boudined, by the second folds. Similarly, granitic and pegmatitic material occurring in the mica schist near the basement contact has been emplaced ensuing the development of schistosity and is clearly of replacive origin. This too is affected by the second folding. These phenomena suggest, therefore, that metamorphic and metasomatic processes continued into the static interval separating the first and second deformation phases but a precise dating of the acme of the metamorphism cannot be given until a thorough petrographical study has been carried out. While this accounts for the main regional metamorphism, it is quite likely that the pre-Lower Hovin rocks were also affected by an earlier metamorphic event, as noted previously.

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Orienterende undersøkelser vedrørende sprøhet og flisighet av bergarter

Av

Thor L. Sverdrup og *Erling Sørensen.*

Forord.

Undersøkelsen er foretatt for å få rede på om borkjernemateriale kan nyttes for bedømmelse av bergarters anvendbarhet i asfaltslitedekker.

Resultatene er foreløpig basert på en forekomst i Sør-Trøndelag, men resultatene er såvidt interessante at undersøkelsen vil bli fortsatt ved andre forekomster og i andre bergarter.

Feltet som er undersøkt ligger ved Kalvå, Ørlandet, hvor firmaet Dyre Halse A/S har oppført et moderne pukkverk.

De geologiske undersøkelsene er utført av statsgeolog Sverdrup, de mekaniske undersøkelsene av tekn. ass. Erling Sørensen, mens boringene ble ledet av formann Bratli, NGU.

Innledning.

Da man ved Geologisk avdelings seksjon for mineralske råstoffer og bygningssten, NGU, arbeider med bergarters brukbarhet som tilslagsmateriale til vegdekker, og da spesielt til asfaltslitedekker, mottar vi i blant borkjerner til sprøhets- og flisighetsanalyser. Da borkjerner har en krum overflate, vil det være naturlig å tro at et nedknust materiale av en slik kjerne vil gi et gunstigere resultat enn hva tilsvarende utskutt materiale gir. Hvor sterkt begunstiget kjernematerialet blir har man ikke kjennskap til. Da hverken Statens Veglaboratorium eller Geologisk institutt, NTH, som driver tilsvarende undersøkelser har benyttet borkjerner for analyser, har vi ikke hatt noe å bygge på. Da vi videre må regne med at diamantboringer mer og mer blir trukket inn i undersøkelsesarbeidet av slike forekomster, har vi funnet det nødvendig å foreta orienterende undersøkelser for å bringe klarhet i forholdet kjernemateriale — utskutt bergart.

Feltet som er undersøkt ligger ved Kalvå, Ørlandet i Sør-Trøndelag fylke. NGU fikk i oppdrag fra firmaet Dyre Halse A/S, Trondheim, å finne en bergart i kyststrøkene nær Trondheim som ville egne seg for pukk og som tilslagsmateriale til asfaltslitedekker for maksimal belastning.

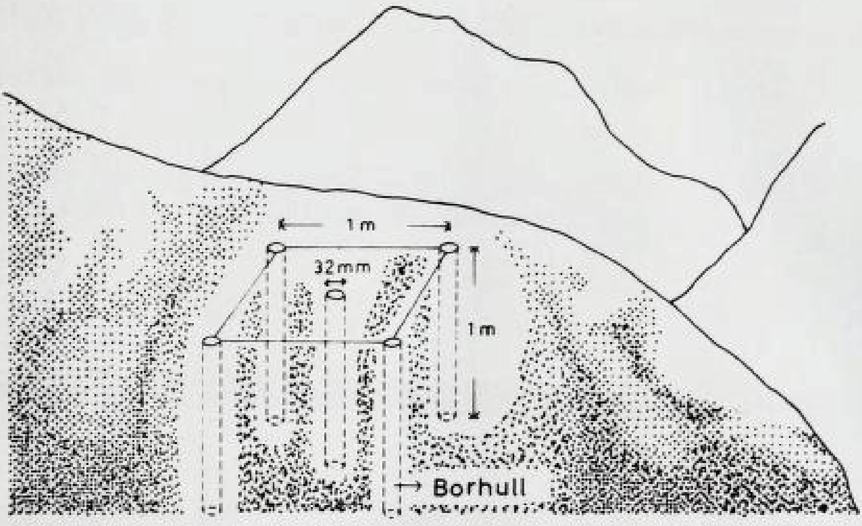


Fig. 1. Skisse som viser hvorledes prøvene bores ut av bergarten. Materialet skytes etterpå ut fra det sentrale borhullet.

Det ble foretatt systematiske undersøkelser av en rekke bergarter i distriktet før en konsentrerte seg om feltet ved Kalvå. (Analysearbeidet ble i begynnelsen utført av konstruktør John Wilhelmsen.)

Geologisk beskrivelse.

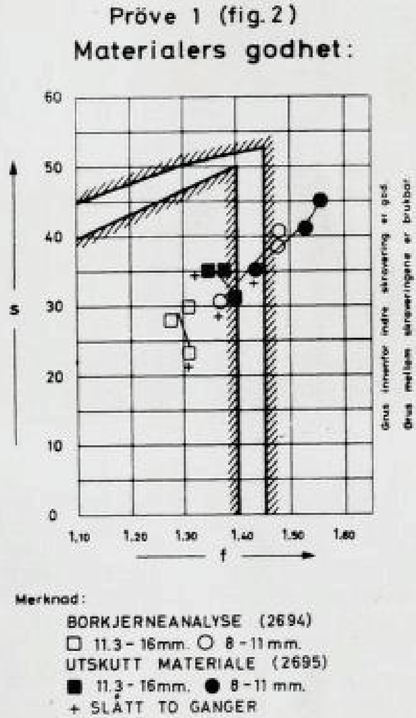
Bergarten er en mylonittisert grønnstein. Hovedmineralene er en noe ufrisk plagioklas og epidot. Plagioklasen er en sur andesin til basisk oligoklas. På slepper opptrer litt kloritt. Svært små gehalter av kvarts sees i slip. Kvartsen er sikkert sekundær og står i forbindelse med mylonittiseringen. Grønnsteinen har fremdeles en ofittisk struktur. Den er svært finkornet, hard og med innfiltrerte mineralkorn, noe som fører til bergartens store holdfasthet. Bergarten er av kambro-silurisk alder.

Boring og materialuttak.

Ved prøveboringen ble det lagt ut fire forskjellige prøvefelt innen forekomsten. På hver borplass ble det boret 5 hull à 1 m. Hullene ble påsatt i en avstand av 1 m i kvadrat med et femte sentralt hull (se fig. 1). For å få tilstrekkelig borkjernemateriale med 32 mm kjernediameter er det nødvendig med ca. 5 løpemeter kjerne.

Straks boringen var avsluttet ble det skutt opp prøver fra det sentrale hullet. Det innsamlede materiale ble uttatt av det minst knuste, og samtlige stykker var av knyttnevestørrelse. Det ble uttatt ca. 30 kg fra hver lokalitet.

Fig. 2—5. Figurene viser sprøhet og flisighet for henholdsvis utskutt materiale og borkjernemateriale.



Resultat av fallprøve.

Resultatene er vist i fig. 2—5. Som det vil fremgå av figurene, begunstiges både bergartens flisighet og sprøhet i borkjernemateriale i prøvene 1, 2 og 4, mens prøve 3 viser liten variasjon.

Gjennomsnittlig sprøhet og flisighet for de fire prøvene er satt opp i fig. 6.

Studerer en fig. 6 finner en således at såvel sprøhet som flisighet begunstiges i de prøver vi foreløpig har hatt til disposisjon. I prosent får en følgende resultat:

Tabell 1.

Begunstigelse av sprøhet og flisighet for borkjernemateriale i %.

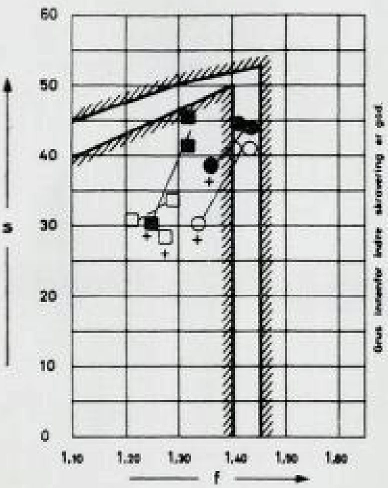
	Fraksjon	Begunstigelse	Begunstigelse slått 2 ganger
Flisighet	8 - 11 mm	2,05 %	2,88 %
Flisighet	11,3 - 16 mm	3,76 %	1,55 %
Sprøhet	8 - 11 mm	4,72 %	10,29 %
Sprøhet	11,3 - 16 mm	11,14 %	15,53 %

Ved å bore ut bergartskjerner istedet for å skyte ut prøver er det trolig to faktorer som har betydning ved bedømmelsen av materialet.

1. Kjernens form med krumme flater i forhold til utskutt materiale.

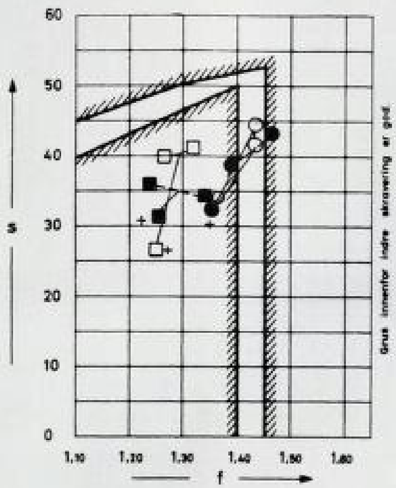
De krumme flatene vil bevirke større motstandskraft mot knusing ved slag. En må vente en avtagende favorisering jo finere (i mm) fraksjonen en arbeider med er, da en ved fin-fraksjonen får friknust flatene i større grad enn ved

Pröve 2 (fig. 3)
Materialers godhet:



Merknad:
BORKJERNEANALYSE (2696)
□ 11.3 - 16 mm. ○ 8 - 11 mm.
UTSKUTT MATERIALE (2697)
■ 11.3 - 16 mm. ● 8 - 11 mm.
+ SLÅTT TO GANGER

Pröve 3 (fig. 4)
Materialers godhet:



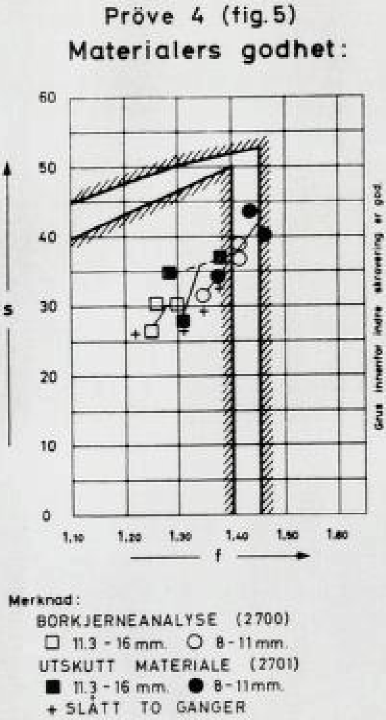
Merknad:
BORKJERNEANALYSE (2698)
□ 11.3 - 16 mm. ○ 8 - 11 mm.
UTSKUTT MATERIALE (2699)
■ 11.3 - 16 mm. ● 8 - 11 mm.
+ SLÅTT TO GANGER

grovere fraksjoner. Ser en på sprøhetstallene for fraksjonen (8 - 11) mm og 11,3 - 16) mm ser en tydelig tendensen 4,72 % mot 11,14 %. Videre må en vente en reduksjon av forholdet ved å øke kjernediameteren, men her har vi foreløpig ikke noen data å støtte oss til.

Vedrørende flisigheten må en naturlig nok vente en bedring i resultatet såfremt en ikke utfører boringer parallelt eller tilnærmet parallelt skifrighet eller lineasjonen. I en såvidt homogen bergart som en her har synes det imidlertid ikke som en kan si noe absolutt om det er fin- eller grovfraksjonen som prefereres spesielt.

2. Sjokk i bergarten p.g.a. skyting.

Den andre faktoren som naturlig vil begunstige borkjernematerialet er at materialet ikke blir utsatt for sjokk og dermed oppknusninger som utskutt materiale blir utsatt for. Dette vil i denne bergarten sannsynligvis gjøre størst utslag i sprøheten. Hvor mye av prosentforbedringen som imidlertid skyldes skyting og hvor mye som skyldes borkjernens form, kan vi ikke avgjøre i dag.



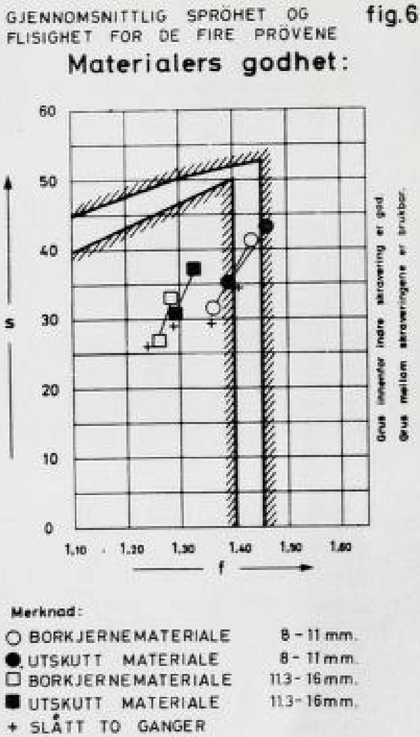


Fig. 6. Figuren viser den gjennomsnittlige sprøhet og flisighet for de fire prøvene.

Konklusjon.

De fremkomne data viser såvidt markert begunstigelse av borkjernemateriale både hva sprøhet og flisighet angår, at en skal være meget varsom med å benytte borkjernemateriale ukritisk for bedømmelse av bergarters anvendbarhet i faste veidekker.

Undersøkelsen er hittil begrenset til et felt, men p.g.a. resultatene har institusjonen funnet det nødvendig å fortsette arbeidet også i andre områder, såvel i massive som skifrige bergarter.

Contributions to the geology of Hardangervidda (West-Norway)[1]

I. An explosion-breccia occurrence in Hjølmodalen.

By *Sverre Svinndal* and *Henri Barkey.*

Abstract.

An occurrence of a breccia, which is thought to have been formed by a fluidization process, is described. The breccia is supposed to be genetically related to similar breccias encountered in other parts of the Precambrian of southern Norway. A short comparison is made with the previously described Gardnos breccia.

The occurrence of the breccia is situated in Hjølmodalen, about 2 km to the south of the village Øvre Eidfjord (see Fig. 1 for the location).

The breccia is outcropping over a distance of about 110 m along a fresh road-cut of a road under construction, leading from the valley up to the mountain plateau of SW Norway (Hardangervidda).

The dimensions of the outcrop are estimated to be about 130 m x 50 m, with the elongated direction along a major fault system. The border contacts of the breccia are very irregular and many minor offshoots radiate into the country rocks, which are mainly gneisses and granite-gneisses of Precambrian age. Some minor amphibolitic bands occur.

A geological section of the breccia along the road is given in Fig. 1. From this section and Figs. 2 and 3, it is clearly seen that blocks and fragments are in all stages of detachment from the walls and from each other. The separating medium consists of veins of "intrusive tuff" which also have perforated and dissected the blocks and fragments themselves.

H. Cloos (1941) proposed the term *tuffisite* for this kind of "intrusive tuff",

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[1]) The authors are engaged in engineering geological work in the NW part of Hardangervidda in connection with a designed hydro-electric power project by the Norwegian state hydroelectric power organisation. (NVE. STATSKRAFTVERKENE). The engineering geological investigations started in 1962 and the technical results are presented in a series of internal reports to NVE. Statskraftverkene, for whom the investigations were carried out. In a series of short papers the authors intend now to present and describe some geological features which might contribute to a better knowledge of the geology of this part of Hardangervidda.

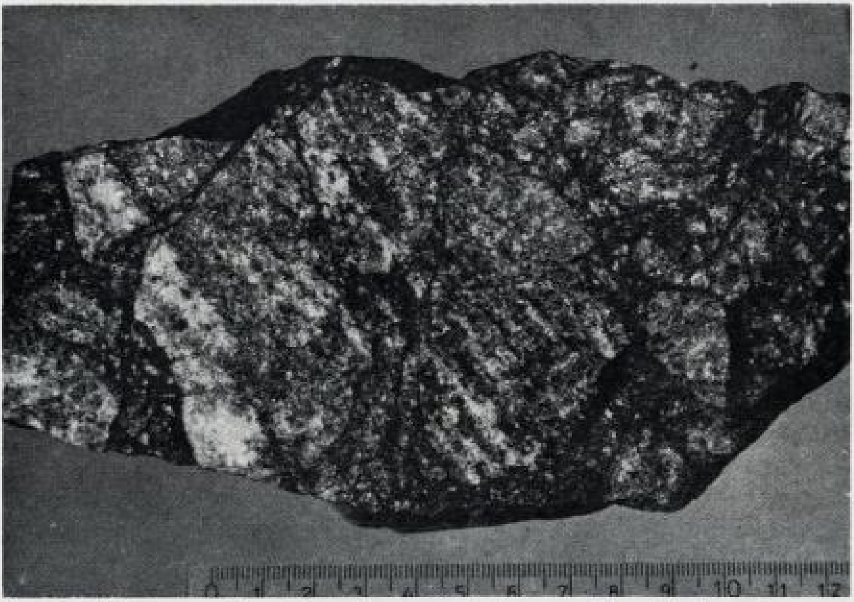

Fig. 2. Rock fragments in an incomplete stage of detacment. Note the larger fragments with preserved orientation of the foliation.

formed from the country rocks to distinguish it from the tuffs normally deposited over the surface as volcanic ash. According to Cloos's extensive description of the Tertiary tuff pipes of Swabia, the tuffisite consists of al lkinds of country rock debris down to dust size and in addition solidified droplets (lapilli) of lava.

No detectable volcanic material has been observed by the authors in the breccia described here.

The most typical feature of the breccia is the cataclastic texture (Fig. 4). Cognate crystals and fragments of the country rock are embedded in a submicroscopic granulated groundmass, in which epidote minerals and a brownish biotite are quite abundant, in part, these minerals are newly crystallized. In addition, patches of opaque ore minerals and some scattered graphitic material occur. Often, these minerals occur even along the finest fissures and cracks.

Quartz, feldspar (both plagioclase and microline), biotite with sagenitic inclusions, muscovite, carbonate, chlorite, sericite, saussurite and accessory minerals (zircon, apatite, ilmenite with leucoxene rims) are minerals which are also characteristic of the gneissose country rock.

Most of the cognate crystals and fragments of the country rock are angular

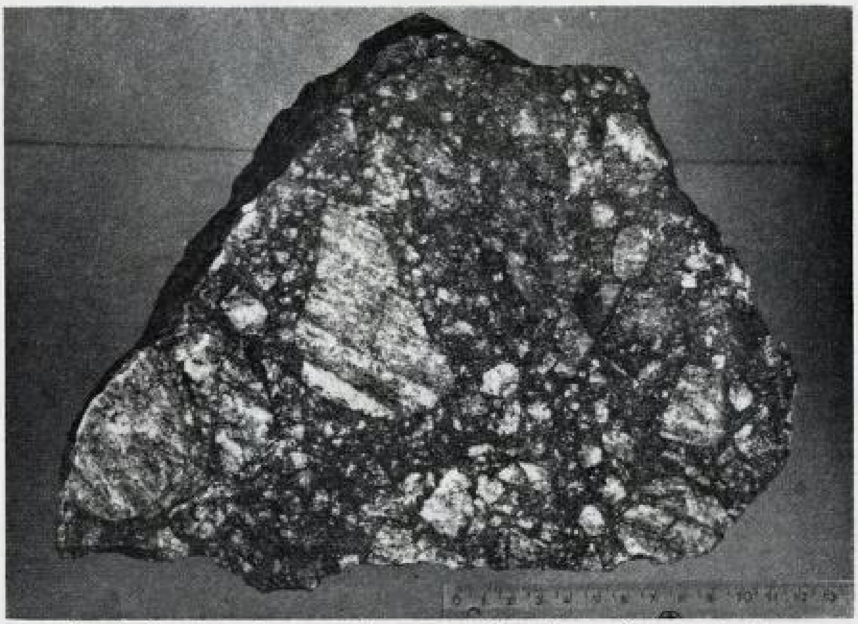

Fig. 3. Fragments in a complete stage of detachment from each other.

to sub-angular, sometimes with pronounced abrasion borders. Rounded to sub-rounded forms occur, but are not abundant.

Sometimes rounded microscopic inclusions of quartzitic rocks (Fig. 5) are encountered, while in the cross-section some larger fragments of mica-schists are present. These rock types are not found in the immediate vicinity of the breccia outcrop, but they are not uncommon in the gneissose rock series of this part of Hardangervidda.

Some foliation measurements on fragments and blocks indicate that transition from parallel (to the foliation of the country rock) to completely random orientations exist. No vertical grading of the fragments is observed. Fragments of country rock appear even in very small fissures.

The rocks adjacent to the breccia are not visibly tectonized, most often the contacts being conspicuously clean and sharp (Fig. 6). Only where the ramifying veins in the surrounding gneissose rocks become abundant does a kind of "transition" exist. The observed foliation in some rock fragments and blocks, being parallel to the foliation of the country rock, is in striking contrast to the chaotic jumble that would have resulted from the explosive process usually envisaged. Volcanic explosions and their effects are familiar enough, but here we are dealing with phenomena that cannot be so easily explained.

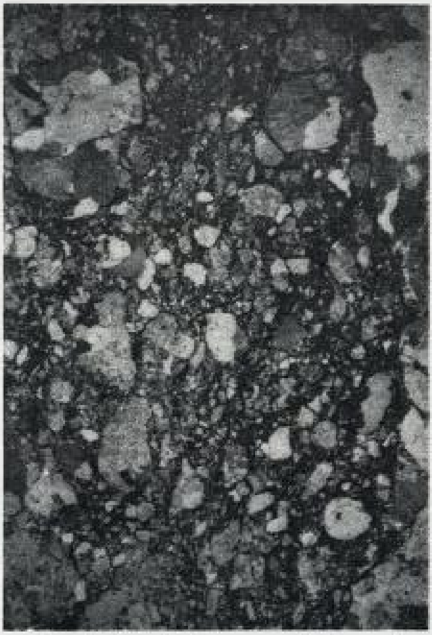

Fig. 4. Cataclastic texture in the tuffisite. Crossed nicols. x 10 magnification.

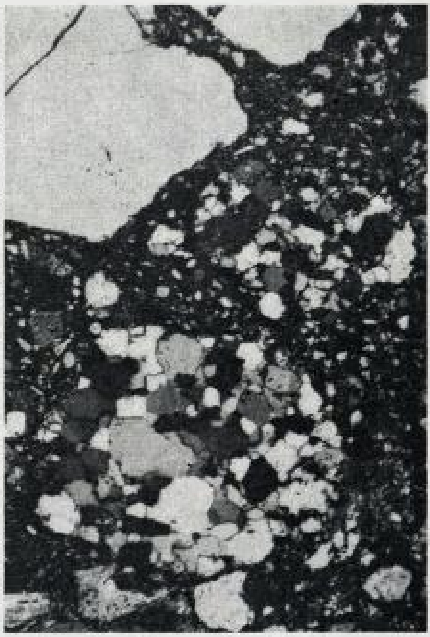

Fig. 5. Microscopic inclusions of rounded quartzitic rocks. Crossed nicols. x 40 magnification.

In the authors' opinion, fluidization or a gas-solid streaming process, such as that envisaged by Cloos (1941) and Reynolds (1954), could adequately explain all the intrusion and comminution phenomena encountered in the breccia outcrop in Hjølmodalen.

In a discussion after a paper presented by Coe (1966), Dr. Doris Reynolds pointed out that in several recent papers the term "explosion-breccia" had been used as though explosion and fluidization were one and the same process.

With the term "explosion-breccia" the authors just want to stress that an explosion must have been the generating force that resulted in "opening" the rocks for the fluidizing process. Any explosive energy that may have been liberated at great depths will be unable to bring about the swift outburst of fragments that one observes in quarry blasting. It will, however, give rise to blast waves, which will shatter the adjoining rocks, and to heat, which will increase both the pressure and the temperature of the gas present. As soon as cracks have been generated (or old fractures sufficiently re-opened), rising streams of high-pressure expanding gas will force their way to the surface

Fig. 6. Sharp contact between the tuffisite and the adjacent gneissose rocks. Roadcut Hjølmodalen.

through passage ways which they can widen by abrasion: the streams of gas are thus arming themselves with dust and fragments that will add to their erosive capabilities. Once a crack is wide enough, the dust-laden gas that streams through it will quickly increase the size of the opening by liberating bigger fragments. Thus all fragments, large and small, will be constantly worn down by abrasion.

Such a process would account for the lack of pyrometamorphism along the margins of the fragments and blocks. The degree of metasomatic alteration of the fragments is strongly dependant upon the gas temperature and chemical composition.

Nowadays, fluidization is a commonly accepted process to explain the peculiar features observed in kimberlites and kimberlitic rocks. This explains perhaps the close genetic relationship between kimberlites and the present type of autoclastic breccia, as observed by so many authors.

For similar breccias encountered in other parts of the Precambrian of southern Norway, Ramberg and Barth (1966) assume a genetic relationship to the Fen volcanism. According to isotope-ratio determinations they postulate an Eocambrian age for this regional volcanic activity. In this connection the authors want to stress Davidson's (1964) conclusion, that it is not necessary to assume that the kimberlitic composition of kimberlites was introduced in a molten state. Consequently, radiometric age determinations on minerals from the kimberlitic fragments do not necessarily represent the real age of the kimberlite emplacement.

The autoclastic breccia in Hjølmodalen is very similar in appearance to some parts of the previously described Gardnos breccia (Broch 1945).

The junior author has had the opportunity to study a detailed section of the Gardnos breccia in a water supply tunnel (Nes hydro-electric power plant), which had been driven through the breccia pipe. A report on these investigations will appear in a separate paper. Peculiar to the Gardnos breccia is the rather high content — as compared to the breccia in Hjølmodalen — of graphitic material which gives the tuffisite its black colour. Some chemical analyses of carbon content for both the Gardnos breccia and the breccia in Hjølmodalen are given below. The sample from Hjølmodalen represents a very fine-grained dark tuffisite, sample G3 from the Gardnos breccia represents a very fine-grained black tuffisite taken in the central part of the breccia pipe and sample G2 is a fine-grained black tuffisite taken closer to the wall contact.

The analyses were carried out by A. Flårønning at the chemical department of NGU.

Sample no.	C %
558 A E30 tuffisite Hjølmodalen, explosition-breccia	0.12 %
J 21 G3 tuffisite Gardnos breccia, (tverrslag T2 venstre)	1.52 %
J 21 G2 tuffisite Gardnos breccia, (tverrslag T2 høyre)	1.02 %

Fairly pronounced metasomatic alteration phenomena along fissures and the margins of fragments were observed in the Gardnos breccia, especially in the central part of the pipe, whereas such features are not perceptible in the breccia of Hjølmodalen. The alteration phenomena and the graphite-content of the tuffisite in the Gardnos breccia seems to diminish towards the contact with the adjacent rocks (mainly gneissose Precambrian rocks). Especially close to the wall contact, the Gardnos breccia is difficult to distinguish from the breccia in Hjølmodalen. A genetic relationship between these two breccia occurrences seems to be very likely. In the authors opinion, a systematic mapping of the Precambrian area in this part of Norway will certainly reveal more of these peculiar breccias.

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Grundfjeldstektoniske studier omkring Moss (SØ-Norge)

En foreløbig meddelelse

av

Asger Berthelsen

Indledning.

Efter aftale med Norges geologiske Undersøkelse, gennem professor Steinar Skjeseth, er detaljerede studier af grundfjeldet i Moss-området påbegyndt af forfatteren i forbindelse med afholdelsen af karteringskurser for geologistuderende fra Aarhus Universitet (somrene 1965 og 1966). Sommeren 1967 er det tanken, at studerende fra Københavns universitet deltager.

Moss-områdets blotningsgrad er relativt god, og ikke mindst de mange nysprængte vejprofiler omkring rigsvejen og dens til- og frakørsler indbyder til detaljstudier. Moss kommune har velvilligst stillet fotogrametriske kort (1 : 5000 med 5 m højdekurver) til vor rådighed.

Arbejdet er blevet udført med støtte fra NGU og de implicerede universiteter. Vi er også Moss kommune og Norges Landbrukshøyskole tak skyldig for megen hjælp.

Af tidligere arbejder foreligger Gleditsch' (1945) oversigtskort visende hovedbjergartstypernes regionale udbredelse og de større strukturer i Østfold, men egentlige detaljarbejder er kun udført længere mod nord (Broch, 1926; Gleditsch, 1952 a og b). For NGU leder professor Skjeseth nu en kartering i 1 : 50.000, til hvilket projekt vort arbejde knytter seg som specialundersøgelse.

Selv om arbejdet må anskues som en "femårsplan" med begrænset indsats hvert år, har vi valgt at præsentere enkelte resultater allerede nu for at orientere kolleger om, hvad vi holder på med.

Moss-områdets regionale placering.

Moss-områdets grundfjeldsbjergarter danner ligesom tilsluttende dele af Østfolds gnejser en nordlig fortsættelse af det sydvest-svenske Stora Le-Marstrand kompleks (Larsson, 1956; Geijer, 1963). Iddefjord-graniten, som begrænser Moss-området i syd, har sin svenske fortsættelse i Bohus-graniten, der gennemsætter Stora Le-Marstrand komplekset og opbygger kysttrakten sydover til Lysekil på nær Kosterøerne i vest (Asklund, 1950). Bohus-graniten tildeles en sen-dalslandisk alder, mens Stora Le-Marstrand komplekset, bl. a. af P. H. Lundegårdh (Lundegårdh og Lunqvist, 1964), tolkes som svekofennisk grundfjeld (med Stora Le-Marstrand serien som svekofennisk suprakrustal), der har gennemgået en delvis gotisk omprægning i forbindelse med foldningen af det østligere strøg med Åmåls formationen.

Dette regionale mønster passer godt sammen med Moss-områdets komplicerede tektonik, hvor liggende folder med vestdykkende akser dominerer, men hvor flere foldefaser kan udskilles. Vest for Oslofjorden viser grundfjeldet en tilsvarende kompleks opbygning (Wegmann og Schaer, 1962; Falkum, 1966.)

Træk af Moss-områdets grundfjeldstektonik.

Moss-områdets yngste grundfjeldsbjergarter udgøres af Iddefjordgranitens pegmatiter, der træffes over hele området, men som især er hyppige i dets sydlige del, hvor de når mægtigheder på op til 10 m. De er ufoldede, stryger i ca. N-S og er østhældende gange med tydelige intrusivkontakter.

I Moss-områdets nordvestlige del, omkring Kambo, gennemsætter Iddefjord pegmatiterne foldede og omkrystalliserede metabasiter, der optræder som delvis konforme, 1—30 m mægtige, gang-intrusiver i gnejs, øjegnejs og migmatitgnejs. Metabasiterne ses tydeligst i vejprofilerne langs rigsvejen, hvor oprindelige intrusivstrukturer (apofyser, afkølingskontakter, xenolither, diskordant afskæring af folder og migmatitårer i gnejsen) også kan ses bevaret, selv om metabasiterne er omkrystalliseret til granat-amfibolit, der lokalt har foliation. Metabasiterne er ellers især karteret omkring overgangszonen eller grænsen mellem to forskellige gnejstyper, en kvartsrig, og en mafitrig type, hvoraf den sidstnævnte ofte er udviklet som øjegnejs eller migmatitgnejs; men også den kvartsrige type kan være migmatiseret.

Metabasiterne, der klart er yngre end denne migmatisering, er foldede med varierende intensitet omkring vestdykkende akser. De ældre strukturer i sidestenen viser også vestdykkende akser. Forekomsterne muliggør derfor en tidsadskillelse mellem to foldeakser, der begge har frembragt vestdykkende akser: V_1-foldningen (ældst) og V_2-foldningen (yngst).

Sydligere, omkring Lauersbakåsen (se fig. 1), kan også udskilles strukturer af forskellig alder. Her er en migmatitisk amfibolit (med net-året kvartsplagioklas neosom) infoldet i sliret migmatitgnejs med et resulterende kompliceret lagforløb på kortet. I grænsen mellem amfibolit og gnejs optræder et tyndt lag af mellemkornet, ofte båndet og noget feldspatholdig kvartsit, der normalt er en halv til 1 m tyk, men som når større mægtigheder i ombøjningszonerne. Selv om kvartsiten kan mangle (være tværet ud?), har den kunnet karteres over det meste af området i fig. 1 og længere mod NØ, hvor-

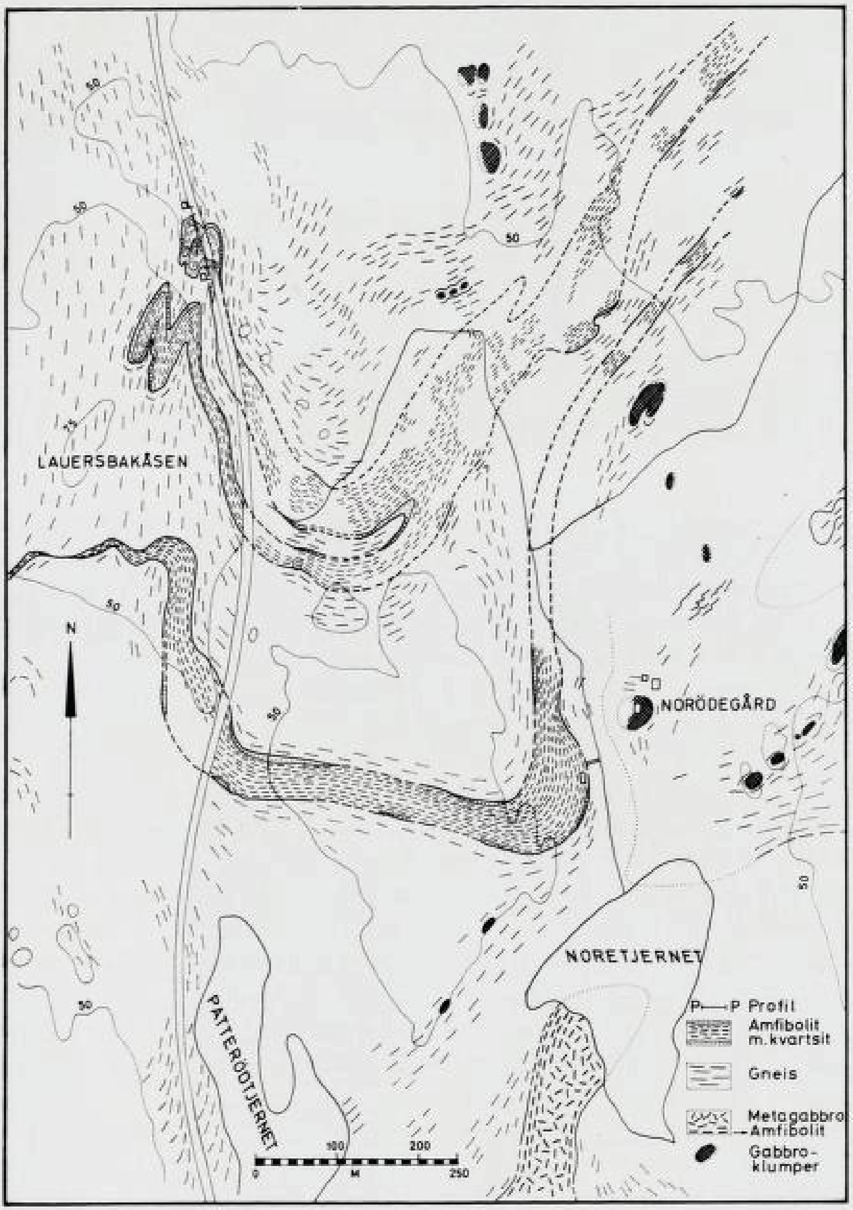

Fig. 1. Geologisk kortskitse af området mellem Lauersbakåsen og Noretjernet, nord for Vålervejens skæring med rigsvejen. Signaturstregerne angiver lag- eller foliationsforløb i blotningerne, og signaturens udbredelse markerer blotningernes fordeling. P——P viser placeringen af profilet i fig. 2, langs vestsiden af rigsvejen.

for det ligger nær at tillægge kvartsiten en suprakrustal oprindelse. Denne tolkning indebærer, at amfiboliten lukker isoklinalt "i sig selv". Det fremgår også af profilet (fig. 2) gennem den nordligste zig-zag foldede "amfibolit gren", at kvartsiten her lukker omkring amfiboliten, der således danner kernen i en isoklinal struktur. Denne isoklinal er senere gennemfoldet med vesthældende aksialplaner, hvorved dens zig-zag-forløb er fremkommet. Både de mindre og større strukturer i denne komplicerede fold adlyder vestdykkende akser. Alle lagflader/foliationer indplottet i et stereogram (se fig. 2) giver en veldefineret π-zone svarende en vestdykkende akse (i 263-25°), omkring hvilken de målte småfolder grupperer sig. Rumvinklen mellem den ældre isoklinals akse og de senere overprægede liggende folders akser må derfor være lille (ca. 10°). Det er fristende at jævnføre disse to foldeakser, der begge har vestdykkende akser, men forskjelligt orienterede aksialplaner, med V_1 og V_2 foldningerne, der kunne adskilles lidt nordligere ved intrusionen af basiske sills og gange.

I profilet i fig. 2 ses i væggens nordlige del et par tynde amfibolitlag indfoldet i migmatitgnejs. Nederst i væggen ses således en liggende fold, der mod syd lukker omkring en vestdykkende akse (V_2 fold at dømme efter aksialplanets orientering). I foldens nedre flanke ses hvert af dens to amfibolitlag imidlertid at lukke "i sig selv" med en afvigende akse (215-7°), parallelt med hvilken der er udviklet en lineation i amfiboliten. Det drejer sig her sandsynligvis om en akse og lineation (ældre end både V_1 og V_2 foldningen), som er foldet omkring vestaksen. Der er også andre steder i Moss-området fundet små og større strukturer, som vidner om en ældre tektonik end V_1 foldningen. Det er imidlertid endnu for tidligt at udtale sig om samhørigheden og betydningen af disse "relikter", da en øget forskel i orienteringen af henholdsvis V_1 og V_2 akserne også fremkalder komplikationer i det tektoniske billede.

Metagabbroen syd for Noretjernet (se fig. 1) danner således kernen i en liggende fold, som er karteret i gnejserne øst herfor med en amplitude på ca. 1 km og med en NV-dykkende akse. Da aksialplanssporet løber fra metagabbroen ombøjning i Noretjernet mod ØNØ til gnejsombøjningen (udenfor fig. 1) repræsenterer den NV-dykkende akse sandsynligvis V_1 foldningen, og V_2 foldningens indvirkning må følgelig her være mindre end omkring Lauersbakåsen.

Metagabbroen, som danner kernen i den liggende gnejsfold, er godt blottet i vejprofilet langs Vålervejen syd for Noretjernet. Metagabbroen overlejres her i vest af migmatitisk amfibolitgnejs, og den underlejres i øst (med tektonisk kontakt) af rødlig migmatitgnejs. Langs sine grænser er gabbroen stærkt opdelt i linser og boudins (med relikte primærstrukturer), hvorom folieret til

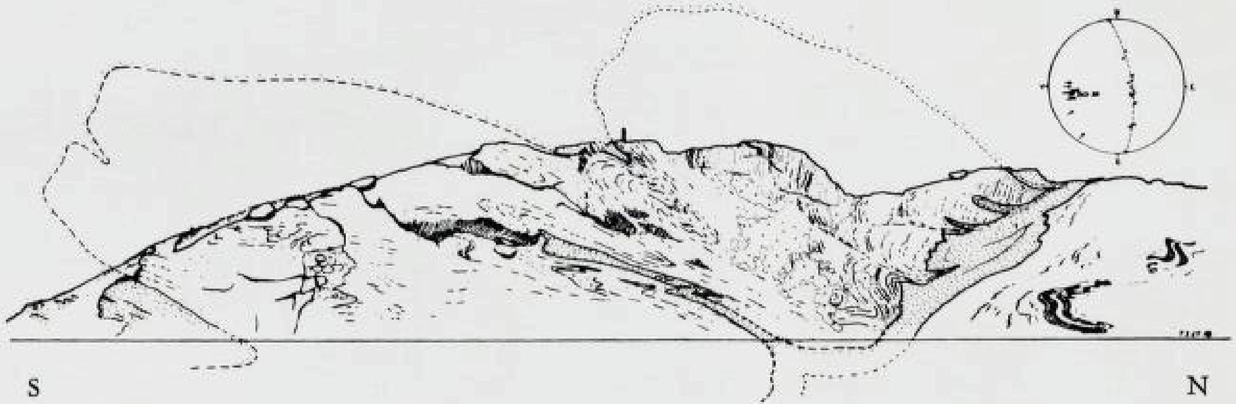

Fig. 2. Profil gennem den nordligste amfibolitgren i Lauersbakåsens struktur. Amfibolit har parallel stregsignatur, hvor de tykke streger viser observeret foliationsforløp. Pegmatitbeklædninger er vist med små kors. Kvartsiten, som omgiver amfiboliten, er prikket. Nordligst i profilet forekommer migmatitgnejs med enkelte tynde, indfoldede amfibolitlag (mørke). Stereogrammet (Wulff net, nedre halvkugle) indeholder normaler (kryds) til lagflader og foliationer og viser orienteringen af målte småfolder (med åbne hoveder) og den konstruerede foldeakse (263-25°). Alle målinger (i gamle 360 grader) hidrører fra profilet og dets nærmeste omgivelser. Profilet er tegnet efter feltskitser og fotos. Dets nordligste del er noget formindsket i fohold til sydenden. Profilet er ca. 50 m langt.

migmatitisk amfibolit smyger sig. Linserne og de forskellige boudins udviser en tydelig strækningsretning parallel med småfolderne i de tektoniserede omgivelser. Disse strækningsretninger og akser har — indplottet i et stereogram — en fordeling, der tenderer mod et lillecirkelmønster. Dette antyder en genfoldning af en ældre liniation (strækningsretninger og småfoldeakser) omkring en NV-orienteret akse. Herfor taler også, at VNV til NV-dykkende strækningsretninger fremhersker i metagabbroens "øvre" (vestlige del), mens N til ØNØ-dykkende strækningsretninger dominerer i metagabbroens "nedre" (østlige) del. Profilet skærer således en synform. Opdelingen og migmatiseringen af metagabbroen skulle derfor kunne henføres til en foldefase, der kan anses for at være ældre end V_1 foldningen.

Tynde, svagt deformede amfibolitgange, der gennemsætter metagabbroboudins i profilet ved Vålervejen, kan muligvis jævnføres med de metabasiske intrusioner omkring Kambro, hvilket ville stemme overens med V_2 folddningens svage indvirkning på metagabbroen langs Vålervejen.

Den fortsatte kartering og de videre petrografiske studier vil selvsagt have som et af sine hovedformål at udrede aldersforholdet mellem de forskelligt orienterede strukturer, migmatitiseringen og dertil knyttede metamorfe processer.

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Geologic and structural studies around two geophysical anomalies in Troms, Northern Norway

by

Asger Berthelsen

Abstract.

The geology and structures of two windows (the Mauken and the Divielva windows, see Fig. 1) are outlined. For the Precambrian supracrustal rocks of the Mauken window, two formation names and one group name are suggested. Hyolithus zone sediments in autochthonous position along the southern margin of the Mauken window show that the Caledonian thrusting exceeded 70 km. In both window structures "Reliefüberschiebung" took place. A magnetic anomaly in the Mauken window is explained by local concentrations of normally accessoric magnetite in an antiformal hinge zone in sericite-chlorite schists. The relations between an E-W directed magnetic anomaly and the structures of the Divielva window are discussed.

Sammenfatning.

De geologiske og strukturelle forhold omkring to vinduer (Mauken og Divielva vinduerne, se Fig. 1) opridses. To formationsnavne og et gruppe-navn bringes i forslag for de prækambriske suprakrustaler i Mauken-vinduet. Fund af Hyolithus zone sedimenter langs sydranden af Mauken-vinduet viser, at den kaledonske overskydning beløb sig til over 70 km. I begge vindue-strukturer fandt reliefoverskydning sted. En magnetisk anomali i Mauken-området forklares ved koncentration af ellers accessorisk forekommende magnetit i en antiform ombøjningszone i sericit-klorit skifre. Relationen mellem en Ø-V rettet magnetisk anomali og strukturerne i og omkring Divielva-vinduet diskuteres.

List of contents:

Introduction.

Since 1963 SYDVARANGER A/S has conducted systematic prospecting in eastern Troms. This work has embraced geophysical surveying, general prospecting such as "block hunting" and geochemical search for ore in addition to photogeologic and field geologic mapping. The coordination of all these activities and the programming of resulting drilling operations were managed by bergingeniør Andreas Eriksen, Sydvaranger A/S. Mr. Eriksen is cordially thanked for many inspiring discussions.

The writer also wishes to thank Mr. Bruno Rothé, geologist (Syvaranger A/S), fil. lic. Nils Marklund, consulting geologist, and his colleagues and students from Aarhus University, who participated in the work, for their collaboration and permission to draw from their unpublished results. Last, but not least, the writer wants to thank *all* who assisted in and contributed to a successful realisation of the field program.

The first part of this paper deals with the Mauken area and in particular its stratigraphy, structures and their relation to some geophysical anomalies. The Mauken area has also been studied and is still being studied by the director of the Natural Museum of Tromsø, Dr. K. Landmark, who is preparing a comprehensive description including petrographic and chemical data to accompany the geological map of the Målselva area. Our work, however, have been carried out independently and differs in methods and scope.

The writer is indebted to Sydvaranger A/S for the permission to publish this report. He also wishes to express his gratitude to Norges Geologiske Undersøkelse for accepting this paper for publication.

Regional Geology.

In Troms the eastern front of the Scandinavian Caledonides (Strand, 1961, Oftedahl, 1966) crosses the Norwegian territory and the extreme south-eastern part of the district is made up of Precambrian crystalline basement rocks comprising: gabbro, syenite, gneisses and granites. Close to the marginal Caledonian thrust, these basement rocks are nonconformably overlain by a fairly thin cover of fossiliferous Cambrian strata, the so-called Hyolithus zone of the eastern foreland. The overthrust, metamorphic units of the Caledonides rest with a marked tectonic contact on the autochthonous cover; Vogt (1918) and Holtedahl (1953 and 1960).

In Fig. 1 the Precambrian terrain of the eastern foreland is shown together with two tectonic windows inside the Caledonides, where the Precambrian basement rocks (wholly or partly fringed by a thin autochthonous cover of the foreland type) are exposed below the allochthonous units, which owe their

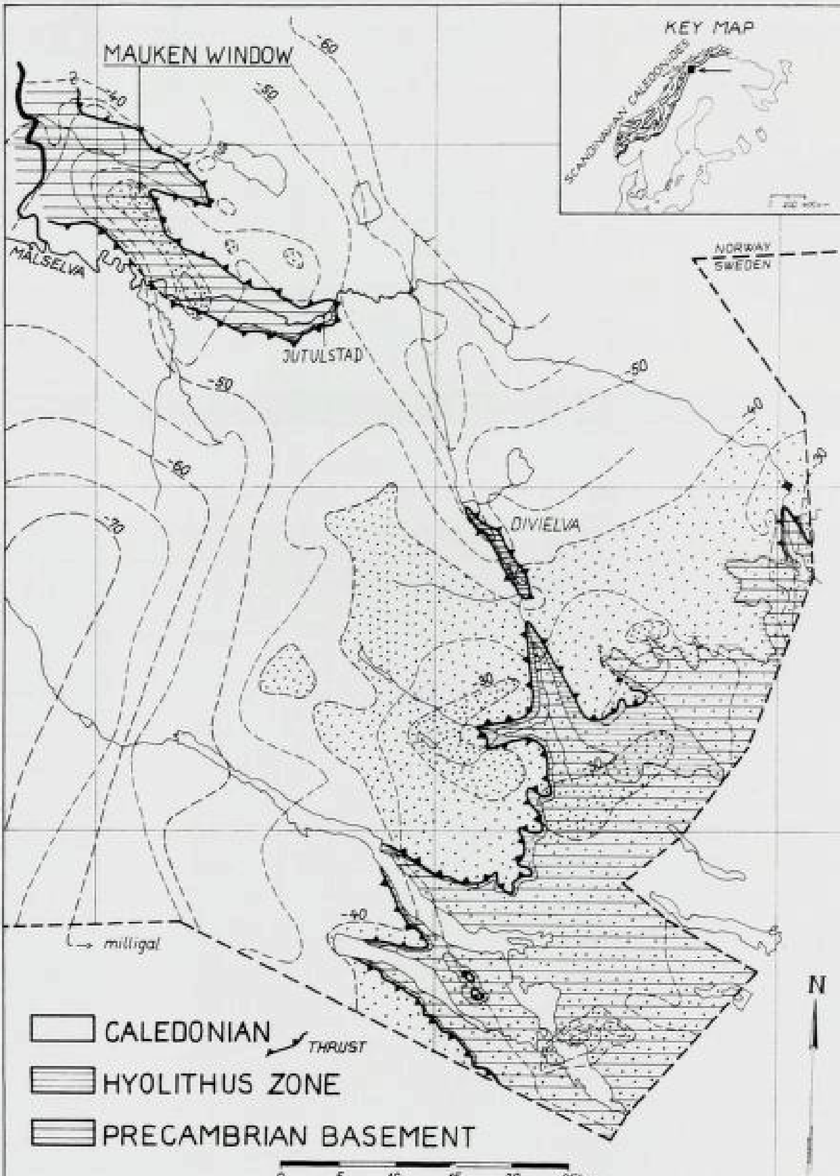

Fig. 1. Sketch map of eastern Troms.

"mise en place" to the Caledonian orogeny. Since these units may include rocks of Precambrian age, they are referred to in the following as the *Caledonian overthrust units.*

The main feature of the gravity field are also shown in Fig. 1. The milligal curves were constructed by bergingeniør Andreas Eriksen, Sydvaranger A/S, from the data supplied by Norges Geografiske Oppmåling. Although no corrections were made, the map clearly shows that the regional gravity field depends on the large scale geologic structures. A pronounced minimum in the south-west corresponds to the depression within the Caledonian thrust masses, and a similar minimum is indicated to the north-east. A gravimetric high extends from the Precambrian foreland area (south-east corner of the map) in a north-western direction, and is clearly accentuated in the Mauken window around Målselva and Takelva, and even the elongated window along Divielva (Gustavson, 1963) influences the anomaly trend.

No doubt, the gravity field betrays a major culmination within the Caledonides caused by a NW-SE directed uplift of Precambrian basement rocks.

The Mauken Window.

The area of the Mauken window is, for its greater part, covered by the geological map (1 : 100.000) published by K. Landmark (1959). Obviously, the idea that Precambrian rocks occur in this region were in Landmark's mind when he drew his map, since in the stratigraphic legend he placed the granodioritic rocks as older than the overthrust units of Stormauken. However, in the geological map accompanying Holtedahl (1960), the granodiorites are shown as Caledonian intrusives.

Photogeological studies and reconnaissance work carried out by the author and his collaborators (1964-65) showed that the Precambrian rocks of the Mauken window are completely surrounded by Caledonian overthrust rocks and that Hyolithus zone rocks occur along the southern and southeastern edge of the window below the basal Caledonian thrust. Gustavson (1966, p. 18 and pp. 47-48) discovered these autochthonous rocks in 1963.

On the spur rising from Alappmo towards Alapp Mt. we found the basal ? Cambrian conglomerate overlying both granodiorite and chlorite schists unconformably. The conglomerate is succeded by dense quartzites grading into grey, partly phyllitic, shales which are in thrust contact with hard-schists and quartzites of the lower "quartzite division" of the Caledonian sequence. Lithologically, the Hyolithus zone rocks of the Mauken area show a close resemblance to the autochthonous rocks found within the southern part of the Divielva window (see p. 70).

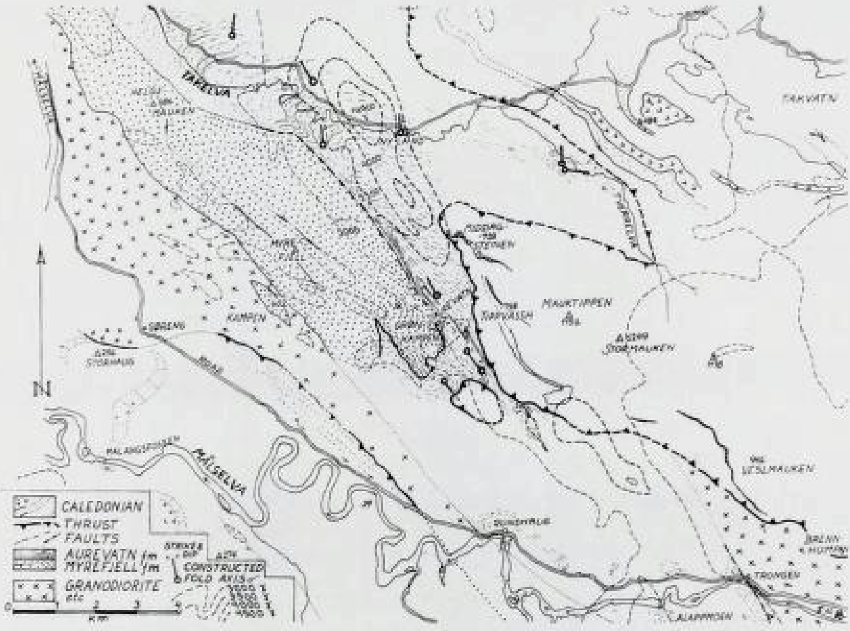

Fig. 2. Geologic sketch map of the Mauken area. Geology from Landmark (1959), Bruno Rothé, J. F. Touborg and the author. Aeromagnetic anomalies from a total intensity survey performed by Compagnie Générale de Géophysique (Paris) on behalf of Sydvaranger A/S. For location of the area see Fig. 1.

East of Jutulstad, the basal Caledonian thrust crosses Målselva and climbs the south-facing slopes of Veslmauken and Stormauken. At Brennhumpan and Veslmauken, the thrust zone, as mapped by Landmark (1959) cuts into Precambrian granodiorite (see also Fig. 2). The two localities offer an illustrative example of a tectonically produced pseudoconformity between basal crystalline and Caledonian overthrust units, the latter including here large masses of reworked Precambrian material.

To the NE and SE of Aurevatn it can be demonstrated that the thrust has sheared off the culmination of a Precambrian antiform. Minor slices of Precambrian supracrustals can be seen in the overlying (transported) units which also contain tectonized and double folded gneisses, e.g. north of Tippvassh (see Fig. 2).

The thrust contact delimiting the Mauken window in the south presumably crosses Målselva east of its cofluence with Kirkeselva, since in the mountain slopes between Rundhaug and Søreng SSW-didpping phyllonites, formed at the expense of the granodiorite, overlie undistrubed granodiorite. Further west, the window border appears on the north side of Storhaugen in a steep, or almost vertical (?faulted), position. The granodiorite reported by Landmark (1959) at Malangsfossen (south of Storhaugen) was not observed by the author. Along the south bank of Målselva, at Malangsfossen, conglomeratic limestones forming part of the Caledonian sequence were seen.

The Hyolithus zone rocks which so far could be traced only along the southern and south-eastern edge of the window, occur, so to say, on the lee side of the window structure, when seen in relation to the Caledonian direction of tectonic transport (NW→SE).

The discovery of the Hyolithus zone rocks in an autochthonous position at the said localities, yields evidence for a Caledonian overthrusting (or underthrusting) totalling more than 70 km (see Fig. 1). The thrusting within this part of the Caledonides can also be described as a kind of "Reliefüberschiebung" since the Precambrian rocks of the window evidently formed a topographic high prior to the thrusting.

Moreover, the Mauken window forms an interesting stepping stone between the Rombak window in the south and the windows of Finmarka in the north.

The Precambrian rocks of the Mauken window.

The Precambrian rocks of the Mauken window form four major units: *The Myrefjell formation* (metavolcanics with intermixed and intercalated metasediments), *the Aurevatn formation* (feldspatic and micaceous quartzites, subordinate calcareous rocks, sericite and chlorite schists), *the Øverbygd crystalline complex* (intrusive granodiorite, gneiss and migmatite, metabasaltic rocks and gabbro), and *the Kampen granodiorite* (massive to foliated granodiorite with a few relics of gneiss).

The supracrustal rocks of the Myrefjell and the Aurevatn formations form a 5-10 km broad SE-NW trending belt flanked on both sides by infracrustal rocks, i.e. the Kampen granodiorite to the south-west and the Øverbygd complex to the east.

Although no direct connection exist between the Kampen granodiorite and the similar rocks of the Øverbygd complex, the assumption can be made that the granodioritic rocks on both sides of the supracrustal belt are of the same age. Landmark (1959) has chosen this point of view. There are, however, some differences between the two regions of infracrustal rocks. While the Kampen granodiorite is mostly made up of fairly homogenous type, the Øverbygd complex, in addition to such types, consists also of grey, banded gneisses and migmatites. The occurrence of amphibolitized and deformed metabasaltic dykes within the gneisses also indicates a more complex history of the Øverbygd rocks. In the new road cuts north-east of Jutulstad and south of Målselva, the Øverbygd granodiorite is seen to intrude and partly assimilate a minor body of biotite-gabbro. Along the same road, the granodiorites, gneisses and migmatites can also be seen forming a SE-plunging antiform, the nose of which is eroded by the stream joining Målselva at Jutulstad.

K. Landmarks correlation of the granodioritic rocks of the Kampen and Øverbygdas area is supported by the fact that both granodiorites show intrusive relations to the separating belt of supracrustals. The Kampen granodiorite intrudes the Myrefjell formation, e.g. on the 602 m hill north-east of Kampen where a more than one kilometer long enclave of Myrefjell metavolcanics is enclosed by granodiorite. In a low road cut leading to Rundhaug from the west, the original intrusive character of the granodiorite (against amphibolitic greenstone) is also preserved. More often, however, the contact runs parallel to the banding, or foliation, shown by the metavolcanic rocks and has in many places been subjected to postintrusive shearing.

The contact between the Øverbygd complex and the rocks of the central supracrustal belt is well exposed along the roads both north of Målselva (at Trongen) and south of this river (the road turnings east of Alappmoen), where a steeply dipping fault separates the two rock units. This fault can be traced southwards along the spur rising towards Alapp Mt. On the south part of this spur supracrustal rocks (chlorite schists) also appear east of the fault and are here intruded by the Øverbygd granodiorite, which contains schist xenoliths. The basal conglomerate of the Hyolithus zone overlies both schists and granodiorite nonconformably.

The Myrefjell formation.

The name, the Myrefjell formation, is suggested for a sequence of generally steeply dipping, mainly metavolcanic rocks (the Mauken amphibolite of Landmark (1959). The metavolcanics are mixed with sedimentary material which occurs also as separate layers of minor thickness. The predominant rock types are epidote-amphibolites and epidotic greenstones. Relic pillow structures (e.g. due north-west of Myrefjell), graded tuff bedding (north-west of Aurevatn) and flow structures (as in the gorge SW of Grønkampen) all suggest an extrusive, or effusive, origin of the metavolcanics. The intercalated metasediments, which usually dip as steeply as the amphibolites and greenstones, represent originally fairly pure carbonate and arenaceous sediments. At one locality, a quartz-banded iron ore was found in a quartzite bed. In thin section, the tuffogenic metavolcanics are seen to contain notable amounts of terrigenic detritus.

Since repetition on a larger scale may well occur in the Myrefjell rocks, their true thickness is difficult to estimate. With a dip of 70°, or steeper, the formation measures more than 3 km.

The rocks of the Myrefjell fm are generally in faulted contact with the rocks of the Aurevatn formation. Only south and west of Grønkampen normal contacts are found (see p. 66).

The Aurevatn formation.

The term the Aurevatn formation is suggested as a collective name for the metasedimentary and, probably, mixed metasedimentary and metavolcanic rocks that make up an overturned (to recumbent) antiform, which culminates just south of Aurevatn and plunges northerly under Takelva. The lowermost and ?oldest) rocks exposed around and just south of Aurevatn are mainly compact feldspatic and schistose micaceous quartzites. Then follows a sequence of sericite-chlorite schists, which in the overturned (western) flank of the antiform are sheared and reduced in thickness, but which form a thicker succession of mainly lens-structured chlorite schists in the upper flank of the structure. On both flanks it often contains considerable amounts (5-10 %) of disseminated magnetite which may be concentrated in cm thick bands in hinges of local folds. NE of Aurevatn a thin layer of dark, rusty graphite schists is found in the chlorite schists.

A little further north-east of Aurevatn these rocks are directly overridden by the Caledonian thrust masses, and due east of Aurevatn, the basal thrust cuts into the quartzites occupying the core of the antiform.

North of Aurevatn, the sheared, magnetite-bearing chlorite schists in the western flank of the antiform are underlain by a thin limestone bed and sericite schists, which downwards (? stratigraphically upwards) become calcareous.

South of Aurevatn calcareous schists overlie the Myrefjell metavolcanics in the east flank of the hill 675 m south-east of Grønkampen. These calcareaus schists may be correlated with the calcareous sericite schists found north of Aurevatn.

The described succession from the lower flank of the antiform is also developed around its closure and in its upper flank in Takelva valley, where the different members attain considerably greater thicknesses. The limestones mapped by Landmark (1959) at Soleng (lower Takelva valley, just outside the area of Fig. 2) thus represent an antiformal hinge zone concentration of the mentioned topmost calcareous strata of the structure.

The chlorite schist and other mafic rocks of the Aurevatn fm are wholly or partly of volcanic origin. In fact, it is often difficult to distinguish hand specimens of the highly sheared Myrefjell metavolcanics from those of the chlorite schists of the Aurevatn fm, just as it is difficult to distinguish between the Aurevatn feldspatic schists and quartzites on one side and the mylonitic rocks of the overthrust unit on the other side. As for both problems, structural mapping affords the only solution (see Fig. 2).

Concerning the age relations between the Myrefjell formation and the Aurevatn formation, the evidence is conflicting. The faulted contact between the two formations (from Aurevatn to lower Takelva) might create the impression that the Aurevatn rocks, due to their lesser dips, originally rested unconformably on top of the Myrefjell rocks. This situation is actually seen south of Grønkampen, where the quartzites of the Aurevatn fm lie with low dips (about 10° to ESE) on top of greenstones belonging to the Myrefjell fm, which dip vertically and strike at almost right angle to the quartzite. Sills of greenstone have, however, been encountered within the Aurevatn quartzites and in these sills traverse cleavage and foliation may be seen. Along the upper course of a stream west of Grønkampen graphitic dark slates and calcareous phyllite apparently underlie the Myrefjell metavolcanics and show very low dips in contrast to the steep structures of the overlying metavolcanics. Poor exposures and insufficient field work render it difficult to interpret the stratigraphic relations here. The steep fold axes found in the Myrefjell rocks could be explained as having been formed along with flexuring which accompanied later faulting, and, therefore, these axes can not be used as safe evidence that that the Myrefjell rocks were deformed before the deposition of the Aurevatn formation.

Because of a lack of evidence to the contrary, and due to the occurrence of metabasic rocks within both formations, it might be advisable to group the Myrefjell fm and the Aurevatn fm together and consider them belonging to one larger unit, for which we suggest the term the *Målselva group*. Thus, the Målselva group includes all the Precambrian supracrustal rocks, forming a belt in a SE-NW direction and extending throughout the Mauken window from Alappmo in SE to the lower Takelva valley in NW.

The supracrustal rocks occurring around Alappmoen could also conveniently be ascribed to the Målselva group, since their correlation with either the Myrefjell or the Aurevatn fms remains somewhat uncertain. They comprise epidotic greenstone, lens- structured chlorite schists and subordinate calcareous layers. At Målselva layers of magnetite ore also occur locally. They are exposed on a small peninsula on the south bank of the river, due NW of Alappmoen.

These occurrences are of no economic value. At Alappmoen and also north of Målselva saussurite-gabbro has been met with. It is not known whether this rock type is of Caledonian or Precambrian age.

Structural control of the Nyland anomaly.

The reason that geological reconnaissance and structural mapping was started within the area of the Mauken window was the discovery of a distinct aeromagnetic anomaly during the systematic geophysical prospecting conducted in the summer of 1963. Comparisons between the NNW-SSE trend of this anomaly and the general NW-SE strike of the formations mapped by Landmark (1959) indicated a "hidden cause" for the anomaly and focussed interest on it, even if its magnitude is not exceptional. Photogeologic studies were made made before the field season of 1964, and soon after the start of the field work it became obvious that the main anomaly (around Nyland, see Fig. 2) could be related directly to a structural element, i.e. the axis of an overturned to recumbent antiform, made up of the rocks now grouped as the Aurevatn fm. The existence of this structure was first proved when its hinge zone around Takelva west of Nyland was mapped. Since the axial plunge was greater than the slopes of the mountains south of Takelva, it could be predicted that the subsurface rocks of the main anomaly region at Nyland would appear on the mountain plateau around Aurevatn. The main emphasis, therefore, was placed on the study of this plateau. The discovery of magnetite-bearing sericite and chlorite schists (which, when crushed, reacted strongly to a hand magnet) both within the lower and upper flank of the antiform left little doubt that the Nyland anomaly should be explained by a slightly higher magnetite content in the schists of the plunging nose of the antiform. Thus, neither the rock association nor the size of the anomaly left much hope for finding ore deposits of economic interest, and, consequently, prospecting was abandoned.

From Fig. 2, where the principal results of the aeromagnetic survey and the geologic mapping are shown, it is also clear that the values over 3500 gamma were obtained only above supracrustal rocks and especially above rocks of the Aurevatn fm.

The Nyland anomaly may serve as an example of a "discordant" anomaly caused by thickening and concentration of magnetite-bearing rock members along the axis of an overturned to recumbent antiform. As shown in Fig. 2, there exists a close parallelism between the anomaly direction and the axial trend. The fold axes shown in Fig. 2 were constructed from foliation poles.

The Divielva window.

During the regional aeromagnetic survey referred to above, a small, but conspicuous anomaly was also found west of the Precambrian window in Dividal. This so-called Frihetsli anomaly was checked in 1965 by means of a geophysical ground survey and geologic field work. Ultimately, a drilling was performed in order to verify the interpretation thus obtained.

The main purpose of the geologic field work was to assist the geophysicists in chosing the right model for further interpretation and calculation, and to arrive at an estimate of the depth to the Precambrian basement below the planned site of a drill hole.

Since the anomaly maximum lies due west of the exposed Precambrian rocks, its cause could be sought either within the overthrust Caledonian rocks or in the underlying Precambrian rocks. A detailed mapping of the overthrust rocks of the anomaly area and of the Precambrian exposures was undertaken, and the autochthonous rocks of the Hyolithus zone were studied as well. This work was greatly facilitated by the grid established for the geophysical ground survey and by air photographs kindly placed at our disposal by the Forestry Department of Troms.

The Precambrian basement rocks.

As mentioned by Vogt (1918) and more recently by Gustavson (1963), Precambrian crystalline rocks and Cambrian sedimentary rocks of the Hyolithus zone outcrop along Divielva in an elongated window. Gustavson (1963) mentions two windows, but strictly speaking there is only one window with two inliers. Since these earlier works contain conflicting results, a redescription of the Divielva window will be given along with the presentation of the structural data. The southernmost and the largest area of Precambrian rocks here called the Frihetsli inlier, is found on the east side of the Divielva around the lower courses of Kleivbekken and Kvernelva, where good sections in the NW-striking basement rocks are seen. Additional exposures are found along a small stream north of Kleivbekken, along the road to Frihetsli and scattered within the forest, which largely grows on morainic deposits.

The predominant rock type of the Frihetsli inlier is a medium-grained, pink gneiss of almost aplitic composition. In the Kvernelva section, it contains layers and lenticles of fine-grained amphibolite, which, like the gneiss, is transsected by slightly discordant pegmatites (usually a few tens of cm thick). The amphibolites may represent folded and metamorphosed basic intrusives (Pettersen, 1874). In the gneisses along Kvernelva, metre-large enclaves and lenticles of talc-tremolite ultramafics also occur. Neither the amphibolites nor

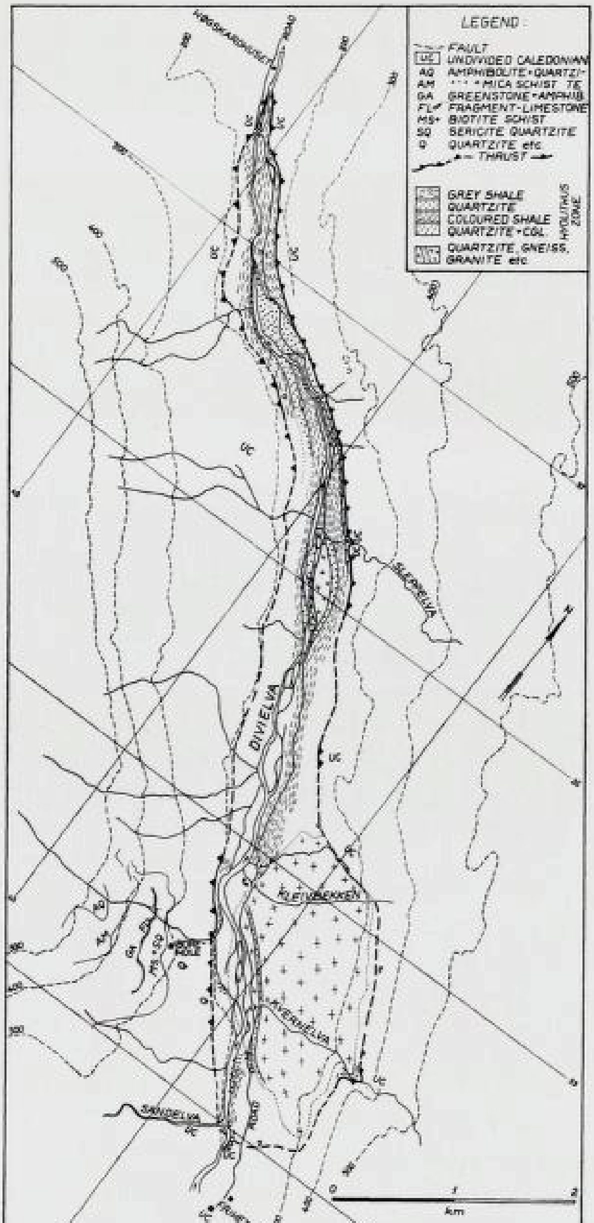

Fig. 3. Geologic sketch map of the Divielva window. Due to scale, the quartzites below and above the coloured shales are not differentiated (cfr. Fig. 4).

the ultramafics contain ferro-magnetic minerals in appreciable amounts and both rock types are quantatively of little importance.

The only mappable member in the basement rocks of the Frihetsli inlier is an, at least 10 metre broad belt of synformally downfolded and lineated, feldspatic quartzites. Thin layers of biotite gneiss also occur in the quartzite belt. The feldspars of the quartzite (up to 10 %) are often whitish weathered, but a similar weathering is also met with in the gneisses and pegmatites in exposures lying close to or representing the sub-Cambrian surface on which the rocks of the Hyolithus zone vere originally laid down. Gustavson (1963) correctly attributed a late Precambrian age to this weathering.

The Precambrian rocks of the northerly Sleppelva inlier are made up of homogeneous pink gneisses and granitic rocks.

The rocks of Frihetsli inlier show a general NW strike and moderate SW dips. Small folds, development of lineation and a slight curvature of strike make possible a construction of the fold axis (see diagram I of Fig. 6). The plunge of 25° at 308° is, however, only representative of the norternmost part of the inlier. Towards the south-east the axis assumes a horizontal position, and it shows a weak SE plunge at Kvernelva.

The means that neither the strike nor the axial trend concur with the general EW trend of the neighbouring anomaly. Thus none of the exposed rocks extend — along their strike or down their axis — into the anomaly area. Therefore, a study of the joint directions in the basement rocks was undertaken. Most joints proved to be cross joints (following the ac plane). They showed sporadic coatings of pyrite. Another well developed system strikes NW and dips at moderate to low angles to NE. Coatings with hematite and locally also epidote are found along this system. No joint system, however, parallels the EW trend of the anomaly.

Supposing that the magnetic anomaly is caused by rocks forming part of the buried basement rocks west of the window, it can be deduced that the anomaly-causing body shows discordant relations to its surroundings.

The Hyolithus zone.

The distribution of the various rock members of the Hyolithus zone is shown in Fig. 3 and the stratigraphic relations are illustrated in the diagrammatic profile of Fig. 4.

The Hyolithus zone generally starts with a basal (?Cambrian) conglomerate and quartzites. The basal conglomerate contains rounded pebbles of quartz and, sometimes, also of quartz intergrown with feldspar as in pegmatite. The feldspar grains of these pebbles are whitish weathered. The basal quartzite

Fig. 4. Diagrammatic profile through the Divielva window.

itself now weathers light brownish but is light greenish to grey when fresh. It is often cut by a system of quartz-filled joints. The basal conglomerate and quartzites are overlain by coloured (red to green) shales except for the southernmost part of the window where the coloured shales are missing. In the bore hole west of the Frihetsli inlier, the coloured shales rest directly on the Precambrian basement. Similar shales are well exposed along the road between Kleivbekken and Sleppelva, showing in places traverse cleavage, which dips steeper to the WSW than the bedding. Folding of the originally vertical joint planes was also noticed in several road cuts.

In the southernmost part of the window (i.e. along the eastern bank of Divielva, north and south of Kvernelva), the basal conglomerate and quartzites are overlain by darkbluish quartzites or sandstones of gritty appearance (due to recent weathering of scattered pyrite grains?). This bluish quartzite is correlated with the more light-coloured and banded quartzites which further north overlie the coloured shales. In the south, the dark bluish quartzite is succeeded by dark shales which upwards become more light-grey. Both the quartzites and the grey shales are folded into a syncline, which is overturned to the SW. Its axis plunges about 10° at 299° (see diagram II of Fig. 6). The grey shales attain a phyllitic appearance and may enclose a few metre large lenticles of chlorite schists in their top parts. The grey shales were not differentiated by Gustavson (1963), according to whom the quartzites (at Sleppelva) are in direct contact with the overthrust Caledonian masses. Gustavson (1963) estimated the thickness of the Hyolithus zone at Sleppelva to 75-80 m. Vogt (1918) mentioned that it varies from a few to hundred metres within the window.

Since in the southern part of the window the present topographic surface lies close to the sub-Cambrian level of erosion, weathering and deposition, it is possible to draw a contour map of this depositional plane. The contours represent the horizon separating the weathered basement and the basal conglomerates. The map (Fig. 5) shows a gentle culmination within a generally WSW dipping surface. In the eastern part of the Frihetsli inlier the basement reaches altitudes of more than 300 metre as mentioned by Vogt (1918), but

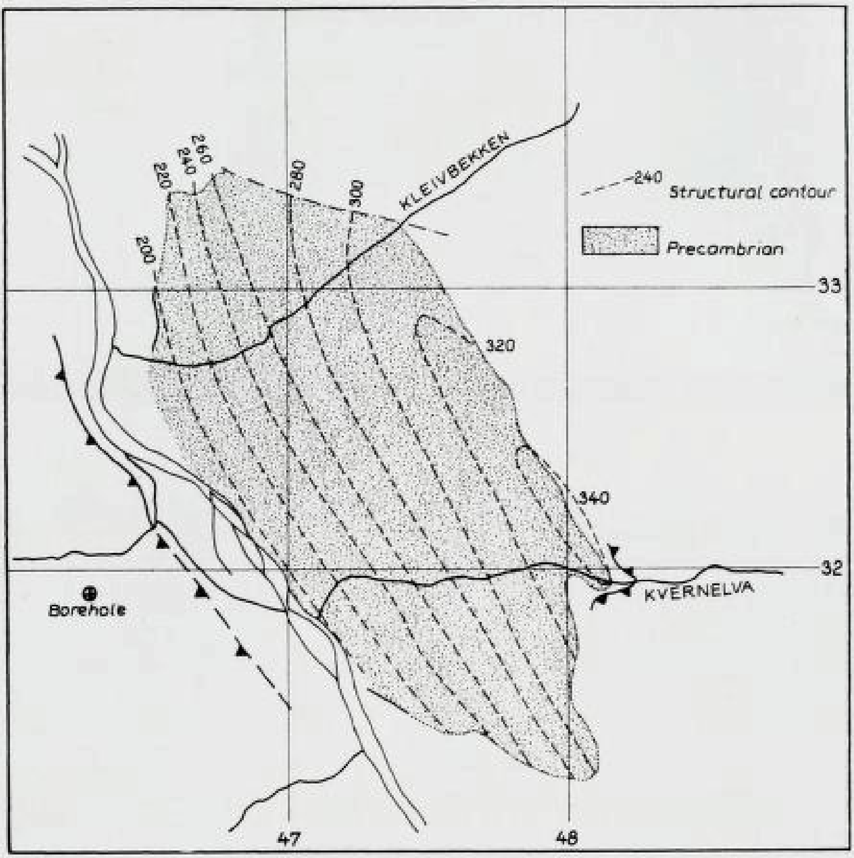

Fig. 5. Contour map of the sub-Cambrian surface (Frihetsli inlier). The borehole is showed by a cross within a circle.

doubted by Gustavson (1963). The depth to the Precambrian rocks at the bore hole site west of the Frihetsli inlier was estimated by assuming that the dip of the sub-Cambrian surface remained constant. This estimate proved surprisingly correct.

The general WSW dip of the sub-Cambrian surface around Divielva is a result of a larger basement uplift, which must be limited to the ENE by a fault or a flexure.

Returning to the diagrammatic section of Fig. 4, it may be pointed out that the slight culminations shown by the Sleppelva and Frihetsli inliers do not show any direct relation to the observed wedging out of the coloured shales of the Hyolithus zone. Rather, it seems that increased uplift corresponds to an originally greater thickness of the autochthonous cover rocks. The way in which the Caledonian thrust cuts off different sedimentary units in various parts of the window, indicates that the uplift took place prior to the thrusting, as advocated by Gustavson (1963) and already mentioned by Vogt (1918). The folding noticed within the rocks of the Hyolithus zone just south of the Frihetsli inlier may have been controled by pre-existing basement structures and may predate the overthrusting because of its opposite directed "Vergenz".

The allochthonous rocks and their structures.

Within the anomaly area, which was mapped in 1 : 10.000, the following rock sequence was found above the basal Caledonian thrust:

Table.

Structurally highest:
(legend in Figs. 3
and 7)

AQ	>400 metres of massive to schistose amphibolite (of intrusive origin) and white quartzites.
AM	about 150 metres of massive to schistose amphibolite (of intrusive origin) with thin layers of biotite schists and a discontinuous layer of graphite-bearing biotite schists at the base.
GA	about 150 metres of banded greenstones grading upwards into amphibolitic schists (of probable effusive origin).
FL	0.5-3 m of limestone with slump (?) structures.
MS +	about 10 metres of coarse-grained biotite schists.
SQ + Q	about 150 metres of the "quartzite division" embracing (from top to bottom): sericite quartzites, dense and banded quartzites or intensively mylonitized rocks, a discontinuous horizon of calcareous, pyritic dark schists, and at the base, dark quartzites.

— — — — — — — — — — basal Caledonian thrust — — — — — — — —

Hyolithus zone.

The only unit of this sequence which will be given closer attention is the limestone layer (FL of Fig. 7). This unit, particularly in its lower part, carries fragments of varying size and shape. Some fragments are rounded to ellipsoidal, whereas others are angular. Usually they occur irregularly scattered in the limestone, but in some cases they are concentrated along bedding planes giving rise to gritty laminae. Several rock types may be identified among the fragments: biotite gneiss and granite, tremolite ultramafics, rock quartz, as well as quartzites and rusty, coarse-grained, biotite schists. Except for the quartzites and biotite schists which are very similar to some rocks in the underlying Caledonian units, the rock fragments were obviously derived from a pre-Caledonian basement, i.e. a Precambrian crystalline complex of a composition astonishingly similar to that of the Frihetsli inlier. The limestone layer is generally light to bluish grey and may be somewhat silicified in its top part. It often shows intricate folding highly reminiscent of slump structures. It might, however, be difficult to distinguish such structures from those caused by Caledonian stresses in an incompetent layer. The writer is inclined to interpret the limestone layer as either a conglomeratic deposit or an agglomeratic calcituffite. The occurrence of metavolcanics (of ?effusive origin) on top of the limestone could be taken as a support for the last-mentioned way of origin.

The lowermost portion of the "quartzite division" (i.e. the portion below the sericite quartzites) shows a constant strike (Fig. 7). The higher units take the shape of a west-plunging syncline accompanied by disharmonic folding on a smaller scale. As shown in Figs. 6 and 7 the plunge of the synclinal axis decreases in the higher units. The stereograms V and VI (Fig. 6) compiled from readings taken within the amphibolites of intrusive origin also show a wider scatter of the foliation poles. The seemingly complex pole distribution of stereogram V is due to the interference between the W-plunging synclinal axis and a SW-plunging axis.

Several of the smaller structures in the major syncline have given rise to local and rather insignificant pyrrhotite mineralisations in the anticlinal hinges. These small concentrations explain fully the local anomalies superimposed on the major anomaly pattern.

The general westward plunge of the major synclinal axis seemingly fits well in the E-W trend of the magnetic anomaly (Fig 7). This parallelism, however, is by chance. Apart from the above mentioned local pyrrhotite concentrations in small anticlinal hinges, no traces of mineralisation were found within the overthrust rocks of the anomaly area, and the major syncline itself appears to have exerted a rather negative structural control on possible mineralisation.

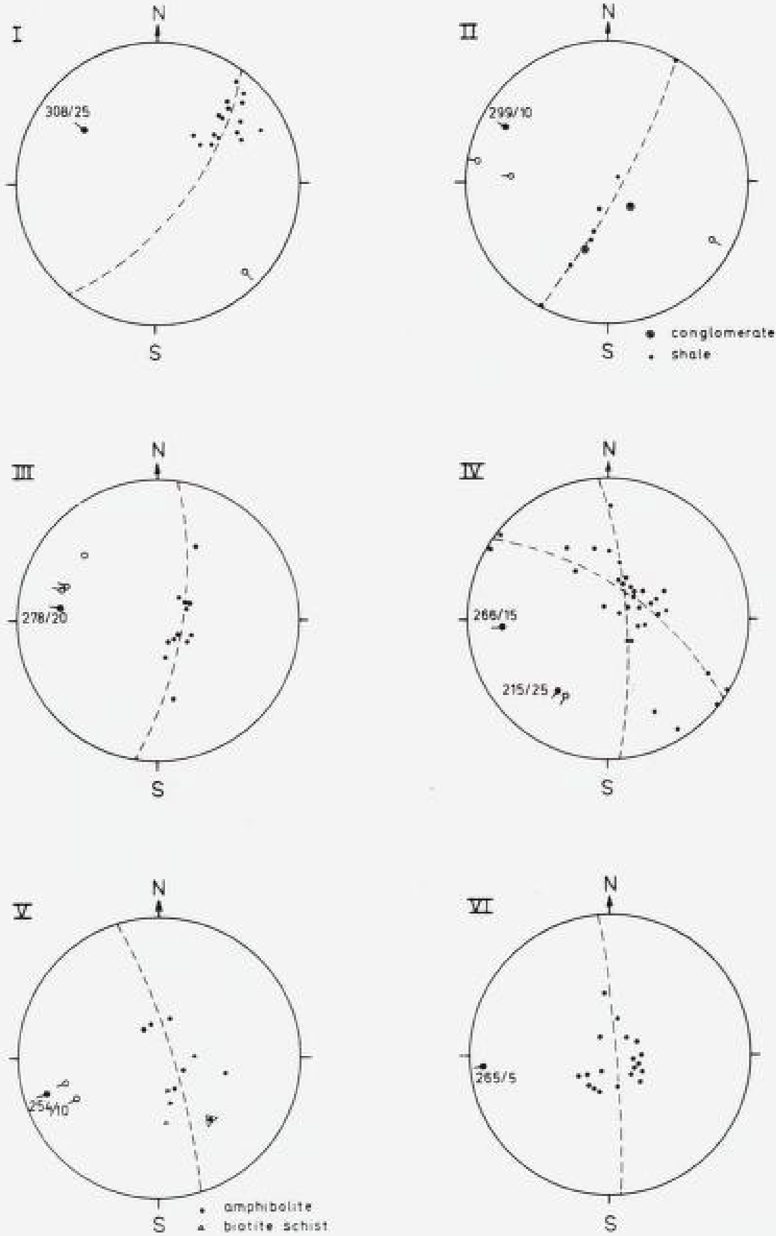

Fig. 6. Structures of the rocks of the Frihetsli area.

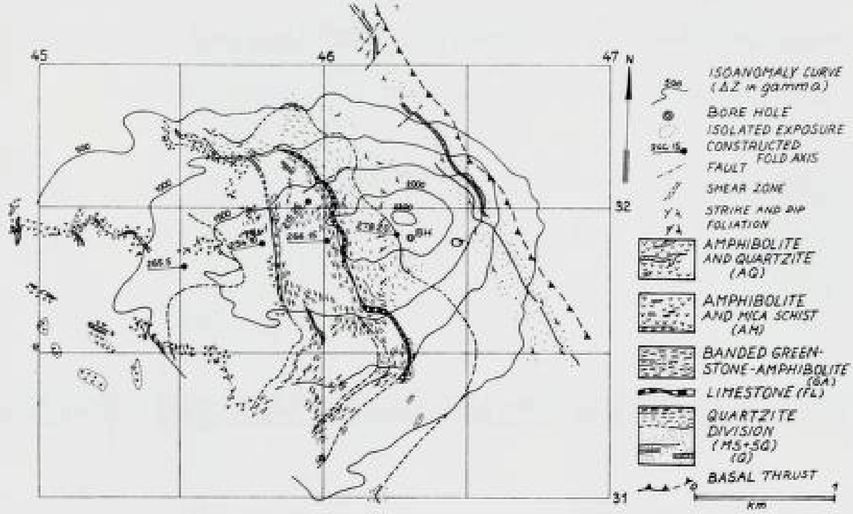

Fig. 7. The Frihetsli magnetic anomaly. Isoanomaly curves (△Z) from a ground survey performed in 1965 by ABEM (Stockholm) on behalf of Sydvaranger A/S. Geology by the author.

Now, after the boring has been drilled, and the cause of the anomaly is known to be a gabbro body forming part of the Precambrian basement, these arguments naturally gain in strength!

The upwards (and westwards) decrease in plunge in the major syncline suggests the presence of an axial flexure. In addition to the attempt to estimate the depth to the basement below the site of the planned bore-hole by means of the known dip of the near-by exposed sub-Cambrian peneplane (see Fig. 5), an attempt was made to use the structures in the overthrust masses. In an E-W profile (including the bore-hole) the basal thrust was thus extrapolated below the mountains west of Divielva, the thrust plane being drawn parallel to the assumed flexured axes in the overthrust masses. The estimate thus made showed to be way too big. The plunge variations, therefore, can now be explained as being due to the formation of imbricate structures within the overthrust masses in response to the obstacle presented by the uplifted basement rocks further to the east.

Concluding remarks.

The two examples described were chosen for publication not only because of their geological interest, but also because they show how geophysical and structural methods can supplement each other. The structural as well as compositional disharmony between 1) the Precambrian basement, 2) its autochthonous cover, and 3) the Caledonian thrust masses resulted in superimposition of internally complex geophysical effects which can only be interpreted satisfactorily once the tectonics of the region are known. In those cases, however, where multiple geophysical methods alone permit a definite conclusion, structural geologists might benefit from the models thus established.

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Structural history of the Bygdin area, Oppland

By

John R. Hossack[1])

Abstract.

The tectonic history of the area, which is part of the marginal thrust zone of the Norwegian Caledonides, is described. Within the area, the Upper Jotun Nappe has been thrust over the Valdres Sparagmite and the Bygdin Conglomerate during the first Caledonian movement phase recognized in the area (F1). This thrusting induced cataclastic textures in the nappe rocks and the sediments, and during a late phase of the F1 deformation, produced northwest trending linear structures, including the pebble elongation of the Bygdin Conglomerate.

During the second movement phase (F2) a large northwest trending antiform, the Bygdin Antiform, was produced which refolded the basal thrust plane of the Jotun Nappe. The axial trend of the minor structures associated with this antiform is exactly parallel to that of the first movement phase.

Structures of the third movement phase (F3) are found only on the minor scale and have an orthorhombic symmetry similar to those described in the Moine Thrust Zone, Scotland.

During the last movement phase in the area (F4), brittle structures which include joints, joint-drag folds, and tension gashes were formed.

1. Introduction.

The area described in this paper lies on the south-east margin of the Jotunheim Mountains, 190 km northwest of Oslo in the County of Oppland. Structurally, the area is part of the marginal thrust zone of the Norwegian Caledonides and within the area, the Upper Jotun Nappe, which is the highest nappe of the marginal thrust zone, has been thrust over the Bygdin Conglomerate and the Valdres Sparagmite (Fig. 1).

The Jotunheim district was described by Goldschmidt (1916) who recognized two large crystalline nappes, the Lower and the Upper Jotun Nappes which were separated by a series of sediments, the Valdres Sparagmite. Gold-

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[1]) Grant Institute of Geology, West Mains Road, Edinburgh 9. Present address: Geology Department, Sir John Cass College, London, E.C. 3.

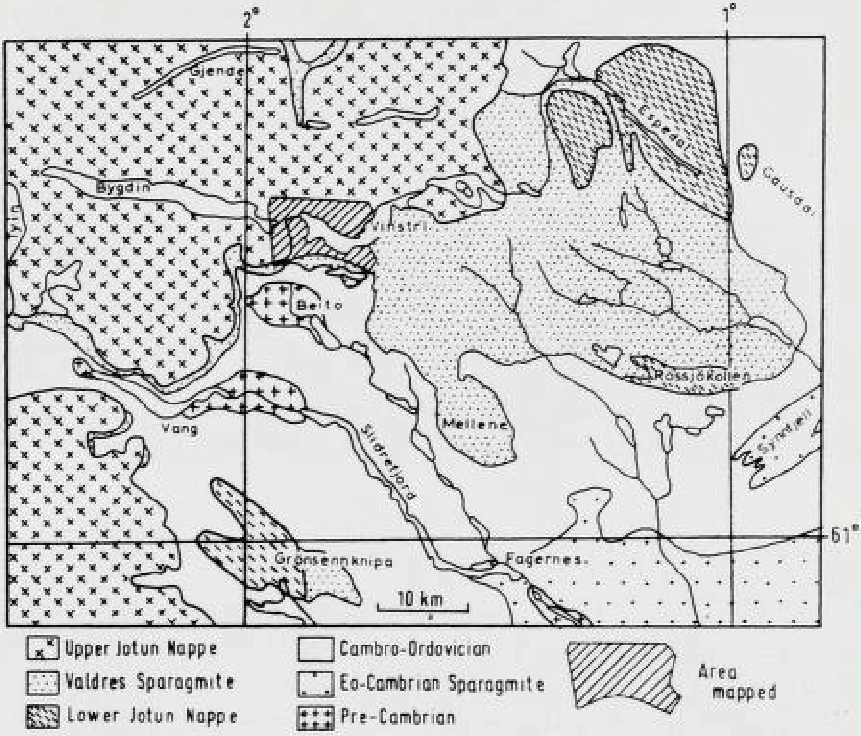

Fig. 1. Geological map of the Valdres district.

schmidt suggested that the Lower Jotun Nappe had been emplaced during an early phase of thrusting and that the Valdres Sparagmite was deposited after or during the emplacement of the Lower Nappe. During a later phase of thrusting, the Valdres Sparagmite was overthrust by the Upper Jotun Nappe. Goldschmidt regarded the Jotun Nappes to be rooted in the large synclinal depression in which they now lie. In addition, Goldschmidt figured the deformed Bygdin Conglomerate and included some measurements of deformed pebbles.

Holtedahl (1936) was the first to postulate a large transport distance for the Jotun Nappes. He suggested that the Jotun Nappes were not rooted in the depression but in fact "floated" on a basal thrust plane above the underlying sediments without showing signs of rooting anywhere. Holtedahl's hypothesis has since been accepted by most Scandinavian geologists.

The Bygdin area was described by Strand (1945) who discussed the structural petrology of the Bygdin Conglomerate and figured microfabrics of the deformed pebbles. He described near-orthorhombic quartz girdles with northwest axes parallel to the pebble elongation, minor fold axes, and the postulated movement direction of the nappe. He also illustrated the exterme variation in the shape of the pebbles from rod shapes to flat pancake shapes. The stratigraphy of the Bygdin area and the surrounding district has been described by Strand (1951, 1958). Strand's work on the Bygdin Conglomerate has been used by other workers to demonstrate the so-called "*a*-lineation" associated with thrusting, which is formed parallel to the postulated movement direction of the thrust (Anderson, 1948, Kvale, 1953).

Flinn (1959) noted the geological similarities between North-east Shetlands and the Jotunheim, and also the comparable pebble deformation in the Funzie Conglomerate (Shetland) and the Bygdin Conglomerate (Flinn, 1961). Both conglomerates show similar types of rod and cake deformation.

In the present study, the Bygdin area was mapped on the scale of 1 : 20.000 and the major structure and the structural history of the area determined by using the minor fold structures and modern structural techniques such as those described by Ramsay (1957 a, 1957 b, 1960) and Turner and Weiss (1963).

2. Stratigraphy and Petrography.

(A) Jotun Nappe Rocks.

The geological map of the Bygdin area is shown in Fig. 14 in which the Upper Jotun Nappe is divided into two separate thrust sheets. To the south of Lake Bygdin towards the Bitihorn, including the peninsula at the east end of Lake Bygdin, the Upper Jotun Nappe is composed almost entirely of gabbroic rocks. The rest of the Jotun Nappe to the north and north-east of Bygdin, along the north side of Lake Vinstra, is composed dominantly of granite and intermediate igneous rocks. Between Lake Bygdin and the Bitihorn, the gabbro and the granite sheets are separated by a thin wedge of Bygdin Conglomerate, with both the granite and the conglomerate wedge dipping westwards underneath the gabbro. The conglomerate may have been a thin wedge which has been caught up by the thrusting and squeezed in between two moving thrust sheets.

The Bitihorn gabbros are coarse grained epigranular rocks with white or pink feldspar and dark ferromagnesian minerals. In contrast to the granite sheet, the gabbros show little sign of cataclasis induced by the Upper Jotun Nappe thrust movements, except in the area above the thrust wedge of conglomerate, where the gabbro has been broken down to a fine grained, schistose phyllonite.

Most of the gabbros clearly show typical igneous textures, including some rhythmic layering of feldspar and ferromagnesian minerals, graded bedding, and load casting of dense ferromagnesian layers into less dense feldspar-rich layers. Also in thin section, hornblende and spinel show well developed intercumulus textures suggesting crystallization of a late liquid in the interstitial spaces between the cumulus phases. The dominant assemblage is labradorite-clinopyroxene-orthopyroxene-hornblende, commonly with accessory mesoperthite, quartz, apatite, and black spinel. The proportion of amphibole may vary considerably and orthopyroxene may be completely replaced in the assemblage by amphibole.

An ultrabasic body with sheared margins outcrops within the gabbro on the west side of the Bitihorn massif. The core of the body consists of an olivine-clinopyroxene-anthophyllite-labradorite-biotite-green spinel assemblage with a clinopyroxene-orthopyroxene margin. This body seems similar to the ultrabasic bodies described by Battey (1960) elsewhere in the Jotunheim. The anthophyllite appears to be replacing olivine, and the stability field of anthophyllite (Greenwood, 1963) suggests recrystallization temperatures between 650°C and 750°C. However, the addition of iron to the anthophyllite system can lower the minimum temperature of the stability to 520°C (Hellner et al, 1965, p. 167).

The Bitihorn gabbros appear to have undergone a post-igneous recrystallization. Many parts of the gabbro are cut by a later series of anastomosing shear planes and parts of the gabbro on the peninsula at the east end of Lake Bygdin are remobilized and contain xenoliths of gabbro with contorted layering, "floating" in a pegmatite matrix. The contrast between the mobility of these structures and the more brittle deformation in the gabbro phyllonite suggests that the remobilization was pre-thrusting.

The granite sheet is composed almost entirely of granitic and intermediate acidic rocks. Most of the rocks have a strong cataclastic texture induced by the thrusting of the Jotun Nappe, which has destroyed much of the original texture of the rock. Hornblende and biotite granites are dominant, and most of the rocks contain microcline perthite and a plagioclase which may range in composition from albite to oligoclase/andesine. The intermediate rocks are characterized by a higher dark mineral content. One specimen of interest showed a quartz-mesoperthite-orthopyroxene-clinopyroxene assemblage with an accessory amount of oligoclase, which suggests that this rock lies in the granulite facies. Just above the thrust plane at Bygdin and at various horizons throughout the granite sheet, the granitic rocks are broken down to fine grained hornblende phyllonites.

Age relationships of the igneous rocks can be determined. On the north side of Lake Vinstra (Map reference 924005, sheet 1617 1), angular xenoliths of intermediate igneous rocks were found "floating" in a granite matrix, suggesting that the intermediate rocks pre-date the granite. North of Synshorn, four xenolith-like lenses of labradorite-hornblende amphibolite were found in the granite. These amphibolites seem to be related to the hornblende gabbros of the Bitihorn, and in fact the gabbro is intruded by granite on the peninsula at the east end of Lake Bygdin.

A metadolerite dyke with a north-northeast trend was discovered intruded into the granite at the east end of Lake Vinstra. Eleven similar dykes have been recorded intruding rocks south of the Tyin-Gjende fault around Tyinholmen and Eidsbugarden by McRitchie (1965, pp. 85-87), and the larger of these trend northeast, though they may have been disturbed by later folding. One metadolerite dyke running north by west is known in the granulites north of the Tyin-Gjende fault on Storegut (Battey, personal communication).

Piecing together these age relationships, the igneous history appears to be

(i) Gabbro crystallization and emplacement of the ultrabasic body.

(ii) ?Intermediate igneous rock crystallized.

(iii) Intrusion of granite.

(iv) Intrusion of dolerite dyke.

All the rocks above, including the dolerite dyke, show an overprinting of the greenschist facies metamorphism associated with the thrusting of the Upper Jotun Nappe, proving that the complete igneous history is pre-thrusting and probably of Pre-Cambrian age.

(B) Valdres Sparagmite and Bygdin Conglomerate.

The Valdres Sparagmite is a typical meta-arkose with a quartz-microcline-albite/oligoclase groundmass which is strongly deformed with the small quartz and feldspar grains forming a partly recrystallized mozaic. Biotite and muscovite form a schistosity and much of the groundmass is flattened parallel to this schistosity. Large round relic clastic grains of perthite and quartz can be found, although these clastic grains are often shattered and broken. Tectonic evidence suggests that the sparagmite is highly deformed and thickened tectonically although some relic sedimentary structures, such as current bedding, are still evident. The petrology of the sparagmite is constant throughout the whole of the Bygdin area showing no distinct alternation of bedding of differing composition.

The petrology of the matrix of the Bygdin Conglomerate is exactly the same as that of the sparagmite. Most of the pebbles are quartzite with under 5 % microcline and chlorite, although a few epidosite and granitoid pebbles are present. The original texture in the pebbles has been completely destroyed by cataclasis, but a few pebbles show some banding which is suggestive of relic bedding. The contact of the Valdres Sparagmite with the conglomerate in the less deformed area at Olefjell (Fig. 14) is very irregular with much interdigitation of conglomerate and sparagmite. The sparagmite and conglomerate are over 360 m thick within the area mapped.

3. Minor structures of the Bygdin area.

(A) Introduction.

At least four sets of minor structures were recognized in the area, and have been separated by using their interference relationships (e.g. refolding with each other). If different sets of structures in an exposure failed to interfere with one another, they were tentatively arranged in the movement sequence by comparing their style with structures of known age elsewhere in the area. The second set of minor structures are synchronous with the only set of major folds recognized in the area. Three of the minor sets include axial plane cleavages, lineations, and minor folds. The fourth set includes folds and tension gashes which are associated with joint formation.

A movement phase which pre-dates the minor structures discussed above, may be present in the Bitihorn gabbro. The poles of the igneous rhythmic layering have a diffuse girdle pattern with a northeast plunging axis. This northeast folding has resulted in steep or inverted dips in the igneous layering of the gabbro (determined from inverted igneous graded bedding and an inverted load cast), but does not fold the thrust plane below the gabbro (synchronous with F1). Structurally the gently dipping thrust plane appears to cut across the steep layering of the gabbro and is thus probably later than the northeast folding. It is suggested that the northeast folds are of Pre-Cambrian age.

(B) First Movement Phase (F_1).

The first Caledonian movement phase at Bygdin consisted of large scale thrusting of the Upper Jotun Nappe over the Valdres Sparagmite and the conglomerate. The thrusting induced cataclastic textures in the sediments and the Jotun rocks, and formed the first schistosity of the area (S1) parallel to the thrust plane. S1 in the granitic rocks of the nappe, is defined by a cataclastic layering in which new biotites and muscovites have crystallized. The general effect of the F1 cataclasis has been to reduce massive ?igneous granites to rocks with a gneissic texture in hand specimen. Regular and irregular phyllonite bands from a few centimeters to over one meter thick, occur throughout the whole of the granite outcrop. In addition to the bands of phyllonite, small patches or isolated "knots" of phyllonite can be found. True mylonites are absent at Bygdin because the strongly sheared rocks are too coarse grained to be defined as mylonites. Because of their phyllitic appearance, they are classified as phyllonites (Knopf, 1931, p. 19). Hornblende phyllonites are especially well developed near the Bygdin Hotel.

Most of the phyllonites have a closely spaced cleavage produced by the preferred orientation of hornblende and biotite crystals, which is parallel to the cataclastic banding, but locally this cleavage is absent or even oblique to the phyllonitic shear bands. In addition. cataclastic and phyllonite layers can be found which are folded in a "similar" style by F1 folds about an S1 axial plane cleavage (Fig. 2, A). This age relationship is taken to indicate that the cataclasis generally was prior to the formation of the F1 folds and S1 schistosity. Only a small part of the gabbro has been altered to phyllonite. A thin sheet of gabbroic phyllonite, 10 m thick, occurs just above the thrust plane 2 km south of Bygdin.

In the non-conglomeratic sediments below the nappe, the biotite-muscovite schistosity (S1) is parallel to the ?cataclastic layering which is defined by alternating layers of slightly differing grain size. Some clastic textures can still be recognized in the sediments in hand specimens and thin sections, e.g. large well-rounded relic clastic grains of perthite and quartz which have been flattened and elongated in a northwest-southeast direction in S1 (Fig. 2,B).

The deformed pebbles in the Bygdin Conglomerate are flattened within S1 (Fig. 3) and in thin section the biotites and muscovites of S1 can be seen to sweep round the flattened pebbles. These relationships are taken to indicate that the main pebble deformation was during the F1 phase.

S1 throughout the area is a biotite-muscovite schistosity, but in many of the igneous rocks, relic igneous hornblende has been reorientated to form planar (S1) and linear (L1) fabrics. The L1 linear structures in the granite are due to the parallelism of elongate feldspar and hornblende crystals. In some localities, L1 is formed by the hinge lines of "similar"-type folds which fold the cataclastic layering and by the intersection of this layering with S1. The L1 lineation was detected at only one locality in the gabbro, indicating its general resistance to the F1 deformation.

The L1 structures in the conglomerate are defined by elongate pebbles and a fine striation lineation on the pebble surface exactly parallel to the longest pebble axis. In some localities, the pebbles have flat pancake-like shapes, and in these pebbles, the longest axis is not always determinable. However the fine striation lineation on the pebble surface is assumed to define the trend and plunge of the longest axis of the pebbles. The assumption that this fine

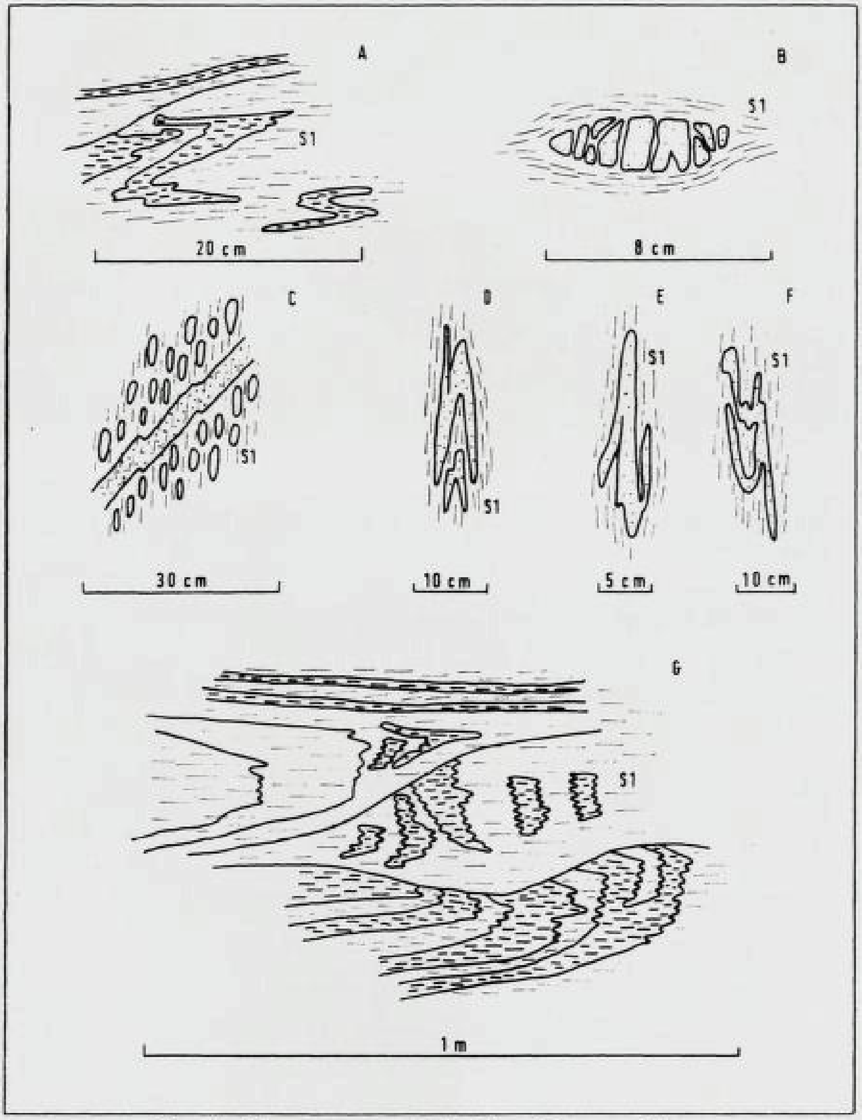

Fig. 2. Minor structures of the First Movement Phase.

striation lineation is of F1 age might be in error because a similar striation on the pebble surfaces is parallel to the axes of the later F2 folds and F2 mica crinkle lineation (Fig. 8, F). Hence much of the striation lineation might be of F2 age. However in two localities the surface striation is folded by F2

Fig. 3. Deformed quartzite pebbles lying flattened in the S_1 schistosity.

minor folds (Fig. 8, D) and must therefore be at least partly of F1 age.

The conglomerate matrix is also lineated, with L1 being picked out by elongated feldspars and micas. Cleavage bedding intersection lineations are present, especially in thin sparagmite bands that indicate original bedding surfaces within the conglomerate (Fig 2, C). Most of these intersections are parallel to the regional L1 trend with gentle plunges to the northwest, but some plunge quite steeply to the northeast.

Two types of F1 folds can be distinguished in the conglomerate. The first type is picked out by folded pebble-free sparagmite bands (Fig. 2, C), and the second type by folded pebbles. The latter type is often difficult to distinguish from the folds in the pebbles developed during the second movement phase. However, the first "pebble folds" have S1 as an axial plane cleavage (Fig. 2, D, E, F) whereas the second "pebble folds" always refold S1, the plane of pebble flattening (Fig. 7). In contrast to the almost homogeneous deformation of the majority of the deformed pebbles, these first "pebble folds" must be the result of inhomogeneous deformation. Like the L1 cleavage bedding intersection lineations, the F1 folds in the pebbles have the regional F1 northwest trend and the gently or steeply plunging northeast directed trends.

The F1 lineations in the sparagmite are defined by elongate feldspars and micas, and cleavage bedding intersections, all having a northwest trend. Because the compositional homogeneity of the Valdres Sparagmite has given rise to a

Fig. 4. Large F_1 bedding folds in the Valdres Sparagmite.

general homogeneous deformation, folds of all ages are rare. However, F1 folds can be found in quartz veins folded about S1. In addition, examples of folded green and pink epidote- and feldspar-rich bands (relic bedding?) are present on a minor scale. These folded structures have an extremely ductile style in which the fold cores are commonly detached from their limbs (Fig. 2, G) and may either be tectonic or sedimentary slump folds. The S1 schistosity is parallel to the axial planes of all these folds. In addition to these minor folds, a few large F1 folds can be found in the sparagmite (Fig. 4). These fold the sparagmite bedding and are recumbent isoclinal "similar" folds.

F1 boudinage structures were observed at three localities in the nappe rocks and the sediments. They are thought to represent elongation strains of the first deformation. At two of the localities, the long and intermediate axes of the boudins lie in the S1 schistosity plane (B.A. 1 and 3, Fig. 5) and were probably formed by an extension axis which lay normal to the long axis of the boudin in S1 (though this is not necessarily true, c.f. Rast, 1956, Flinn, 1962). Application of the experimental results of Ramberg (1959) suggests that the axis of maximum shortening for these two examples lay approximately normal to the long axis of the boudin and S1. At the other locality (B.A. 2) there are rhomboid boudins and the extension and contraction axes have been derived by bisecting the angles between the intersecting rhomboid planes. Maximum extension and contraction axes are plotted for the three localities

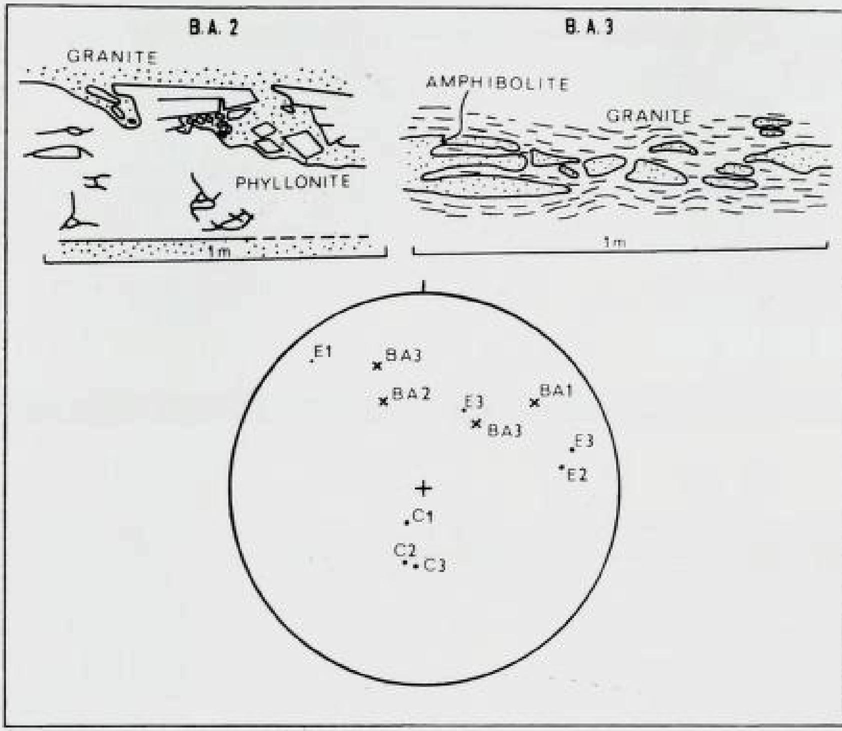

Fig. 5. F_1 boundinage structures and their calculated extension (E) and contraction axes (C). *Crosses* - long axes of the boudins.

(Fig 5). The strain axes derived from the boudins are consistent with the kinematics of pebble deformation which will be described in a future paper.*)

The rhomboid boudins discussed above (locality 2½ km east of Bygdin at 924998, map sheet 1617 1) occur in a phyllonite shear band and indicate that boudinage occurred after cataclasis and phyllonitization, but the age relationships of the F1 folding and F1 boudinage are unknown. A phase of F1 pegmatite injection can be dated with respect to the boudinage. A boudinaged phyllonite band with pegmatite injected into the spaces between the boudins was found at 949982 (map sheet 1617 1) 5½ km east of Bygdin. The pegmatite was emplaced either with or after the boudinage formation, but the pegmatite has undergone greenschist facies metamorphism which can be shown to be associated with the F1 phase. This pegmatite injection may be

*) Hossack, 1968. Pebble deformation and thrusting in the Bygdin area, Southern Norway. Tectonophysics, in press.

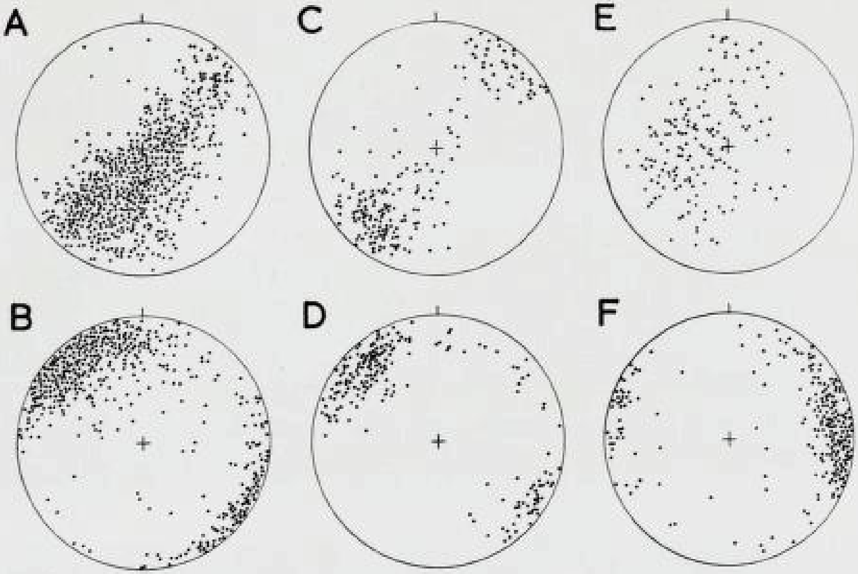

Fig. 6. Stereographic plots of the minor structures of the Bygdin area. A. S_1 poles. B. F_1 fold axes and lineations. C. S_2 poles. D. F_2 fold axes and lineations. E. S_3 poles. F. F_3 fold axes and lineations.

equivalent to the pegmatites injected along thrust planes at Eidsbugarden (Battey, 1965).

The detailed history of the F1 movement phase is as follows,

(i) Thrusting of the Upper Jotun Nappe with phyllonitization and cataclasis.

(ii) Folding and boudinage of the phyllonite bands. S1 formation and pebble deformation. ?Pegmatite injection.

(iii) Greenschist facies metamorphism (synchronous with (ii) ?).

S1-poles throughout the whole of the Bygdin area display a girdle pattern with a gently plunging northwest axis (Fig. 6, A). This girdle distribution is a result of folding by F2 major and minor folds. The F1 linear structures and fold axes have a wide variation in orientation (Fig. 6, B) with a maximum which corresponds to the gently plunging regional trend of F1. However, east and northeast trends are present in the F1 lineations at Barnesodden (Fig. 15). In addition at various localities near Bygdin, F1 folds in the pebbles and cleavage bedding intersections in the sparagmite have northeast trends. The steeply plunging F1 structures of Fig. 6, B are result of later F2 and F3 refolding. The F1 minor structures and the strains indicated by the deformed pebbles give no clue to the direction of movement of the Upper Jotun Nappe.

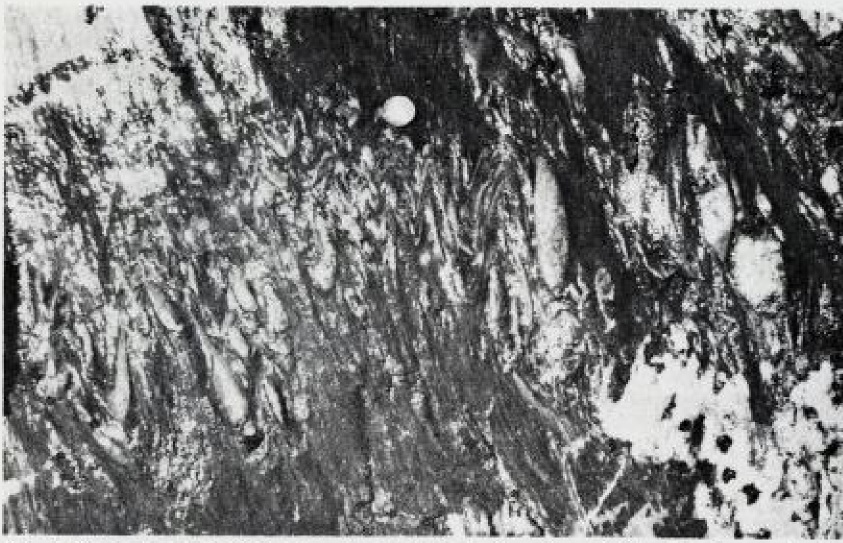

Fig. 7. Plane of pebble flattening (S_1) folded by F_2 minor folds with the production of an S_2 axial plane crenulation cleavage.

(C) Second Movement Phase (F_2).

The linear structures of the second movement phase (L2) are generally difficult to distinguish from those of the first movement phase because the first and second linear structures are parallel (Fig. 6, B and D). However, the second folds can be shown to post-date the first movement phase because they refold S1 (Fig. 7) and F1 folds (Fig. 8, A and C). The shapes of the F2 folds range from gentle to isoclinal (terminology of Fleuty, 1964) and the style from "parallel"- to "similar"-type (Fig. 8, E and B) often with a crenulation axial plane cleavage (Fig. 7). Both asymmetrical and symmetrical F2 folds are present, the various shapes being related to the major structure. The different styles of the F2 folds do not appear to be controlled by rock composition but occur throughout all rock types. The styles appear to be controlled by the grain size of the rock with the "parallel"-folds occurring in coarse grained rocks and the "similar"-type folds in the fine grained schistose rocks (e.g. the phyllonites).

Deformed pebbles can be found lying with their shortest dimension normal to the S_2 crenulation cleavage (Fig. 7). These pebbles have either been rotated as rigid bodies or have redeformed during the second movement phase. In addition, the original plane of pebble flattening appears to have been locally rotated by F_2 folding until S_1 can sometimes lie almost parallel to S_2. The

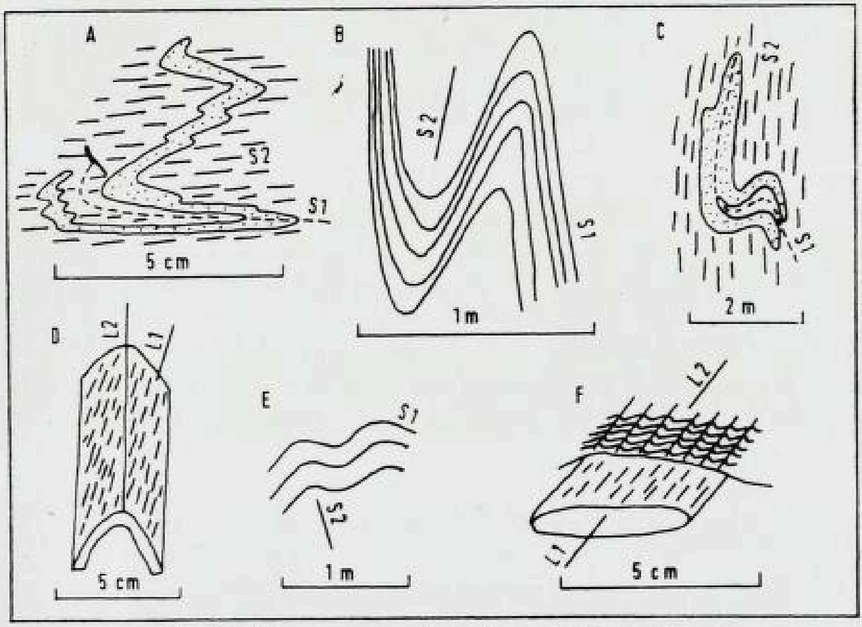

Fig. 8. Minor structures of the Second Movement Phase.

formation of S_2 axial plane cleavage is restricted to the area covered by the inset map of Fig. 15 and to the core of the Bygdin Antiform. The majority of the F_2 minor folds are restricted to the conglomerate and are rare in other rocks of the area.

One noteworthy feature about the F_2 structures is that throughout the whole area, the F_2 linear structures (L_2) are, with two exceptions, always parallel to L_1 (Fig. 8, F). In these two exceptions, a fine L_1 striation lineation is oblique to and folded by "parallel" F_2 minor folds (Fig. 8, D). The fact that much of the fine striation lineation on the pebble surfaces appears to be of F_2 age. has already been mentioned. These examples of refolded L_1 striations prove that some of the striations are of F_1 age. The divergence in trend of L_1 and L_2 in the two exceptions is about 5° to 10°. The parallelism of L_1 and L_2 could be due to the following.

(i) The F2 deformation has oriented L1 into parallelism with L2.

(ii) L_1 has controlled the orientation of L_2.

Hypothesis (i) states that the pre-F_2 trend of L_1 has been rotated into its present regional northwest trend by the F_2 deformation. However, extensive rotation of L_1 seems unlikely, because at Olefjell F_2 minor structures are absent but L_1 still has the regional northwest trend. The absence of F_2 structures suggests that F_2 strains might be absent here (though this is not necessarily true) and hence the northwest trend of L_1 at Olefjell might show the pre-F_2 trend of L_1 for the whole area. In addition, the unrolling of the refolded L_1 striation lineation (Fig. 8, D) which occurs in the area of antiform-A (Fig. 15) gives the pre-F_2 trend of this lineation as northwest-southeast. Both these lines of evidence suggest that prior to F_2 the trend of L_1 was more or less parallel to its present trend.

Hypothesis (ii) states that the pre-F_2 trend of L_1 was northwest-southeast and the formation of L_1 induced an anisotropy in the rocks which controlled the direction of the later F_2 folds. This hypothesis is tentatively accepted by the writer because of the structural relations at Barnesodden. Here L_1 departs from the regional northwest trend into an east or northeast trend. Similarly, L_2 departs from its regional northwest trend into the same east or northeast trend, suggesting some anisotropic control.

The dominant trend of the F_2 fold axes and linear structures is northwest-southeast with low plunges to the northwest or southeast (Fig. 6, D). However, in the area of the Barnesodden peninsula (as discussed above) the F_2 axial directions have a northeast or an east trend. The S_2 axial surfaces have a northwest strike throughout most of the area and dip steeply to the northeast or southwest. The northeast girdle of S_2 poles in Fig. 6, C, appears to be the result of syntectonic refolding of the S_2 planes by the F_2 folds; the evidence for this conclusion will be discussed later. Folding of the S_1 schistosity by F_2 on the major and minor scales accounts for the S_1 pole girdle of Fig. 6, A, with the girdle axis plunging gently to the northwest parallel to the trend of the major and minor F_2 fold axes. With the exception of Barnesodden, F_2 minor folds are generally restricted to the conglomerate southeast of Bygdin and in the granite to the north and southwest of Bygdin.

(D) Third Movement Phase (F_3).

The third minor fold structures vary in their symmetry from near-perfect orthorhombic, through monoclinic, to triclinic, and in their style from "similar"-type folds to "parallel"-folds (Fig. 9). The third structures refold both F_1 and F_2 structures (Fig. 9). The "similar"-type F_3 folds are normally disharmonic (Fig. 9, A) and sometimes have a crenulation axial plane cleavage (S_3). Most F_3 folds have this style and are usually asymmetrical monoclinic folds.

An indication that the third deformation contains a flow component of deformation parallel to the axial surface of the fold is suggested by the patterns of deformed L_1 lineations developed in parts of the conglomerate. The fine

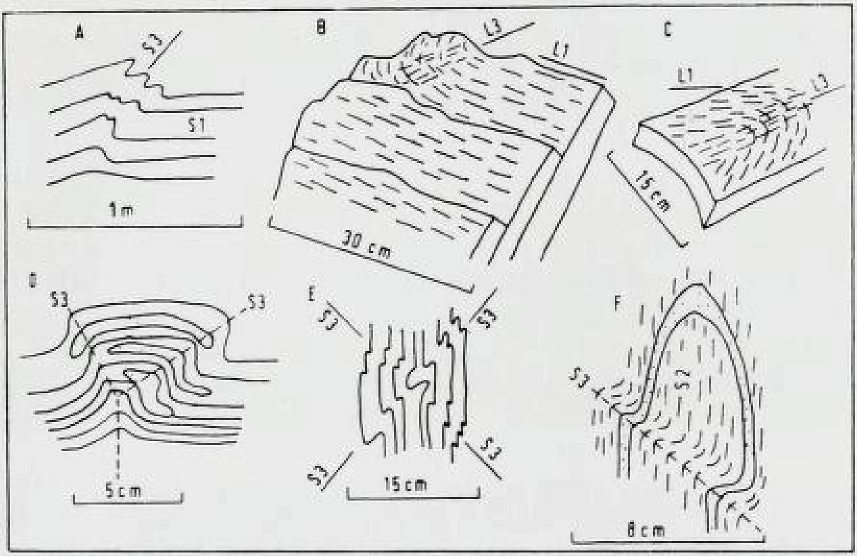

Fig. 9. Minor structures of the Third Movement Phase.

L_1 striation lineation is deformed in a sine-wave pattern on planar surfaces (i.e. the flattened pebble surfaces) (Fig. 9, B). This structure is analogous to the deformed lineations reported by Ramsay (1960, p. 80) in Glenelg. Ramsay (ibid. p. 90) suggested that if the movement direction which deforms a lineation, lies within the plane of the lineation, no folds will develop but the lineation will be deformed into a sine-wave pattern. Any new lineation developed on the surface by the deforming movements will be parallel to the *a*-direction. The L_3 lineation in the above structure is thus probably a true *a*-lineation.

Some of the "similar"-type third folds described above, deform L_1 and L_2 lineations so that they now occupy partial great circles in projection. The *a*-directions of these folds have been calculated (c.f. Ramsay, 1962, a). In some exposures, L_1 and L_2 appear to have been deformed during the production of F_3 folds. The angle between the lineation and the F_3 fold axis now varies across the F_3 fold, signifying that the third deformation included a component other than that of "parallel" buckling. Because of the scale of these folds, individual deformed lineations could not be measured, but the approximate plane in which the lineation lies was measured in the field and the *a*-axis calculated. Although this method is subject to inaccuracies of measurement, it probably gives an indication of the approximate movement directions within the F_3 folds. All the measured *a*-lineations and *a*-directions calculated from deformed lineations are given in Fig. 11, B. Almost all the *a*-directions plunge between 0° and 45° to the east and southeast. Some of these deformed lineations lie in gentle "parallel" folds with the fold axis parallel to the L_3 lineation (Fig. 9, C).

The third folds show an extreme variation in axial plane and fold axis orientation (Fig. 6, E, F). This variation cannot be explained as being wholly due to the control of already folded S_1 surfaces on the orientation of the F_3 axes, or by later refolding. Much of the variation is thought to be the result of the orthorhombic symmetry of the F_3 structures. Some examples of box folds (Fig. 9, D) were found in the conglomerate with axial trends parallel to neighbouring F_3 folds. In addition, in the more schistose parts of the granite, a few conjugate folds with kink bands parallel to the axial plane were found which have a style comparable to those described by Johnson (1956) in the Coulin Forest, Scotland. Most of these folds consist of isolated unpaired kink bands but two folds were found which had the characteristic paired kink bands inclined towards one another, of complete conjugate folds.

Only one F_3 fold with perfect orthorhombic symmetry was found (Fig. 9, E). Unfortunately this was in a fallen block and so its true orientation could not be determined. An L_1 lineation passes over one of the fold hinges so that the angle between the deformed lineation and the F_3 axis varies around the fold. Because of the scale of the fold it was not possible to determine accurately whether the pattern of the deformed lineation lies on a great circle or complex curve (Ramsay, 1963). At least part of the deformation of this fold was accompanied by a "similar" style component (i.e. a component which does not involve flexural slip). This component may have been "continuous simple shear" (Dewey, 1965).

The difference in style between the rounded box folds in the conglomerate and the more angular conjugate folds in the schistose granite is probably a result of the difference in the type of schistosity they fold. True conjugate folds form in closely laminated rocks (Ramsay, 1962 b, p. 517) and hence are resticted at Bygdin to the strongly schistose parts of the granite. Box folds on the other hand formed in the conglomerate where the S_1 schistosity is more widely spaced.

Although most of the F_3 folds are asymmetric monoclinic folds, the poles to S_3 from the whole area have an orthorhombic symmetry defined by a diffuse girdle pattern (Fig. 6, E). This orthorhombic symmetry can be compared directly with the orthorhombic symmetry of the box and conjugate folds and suggests they are all of the same age. In three sub-areas (Fig. 10) more compact

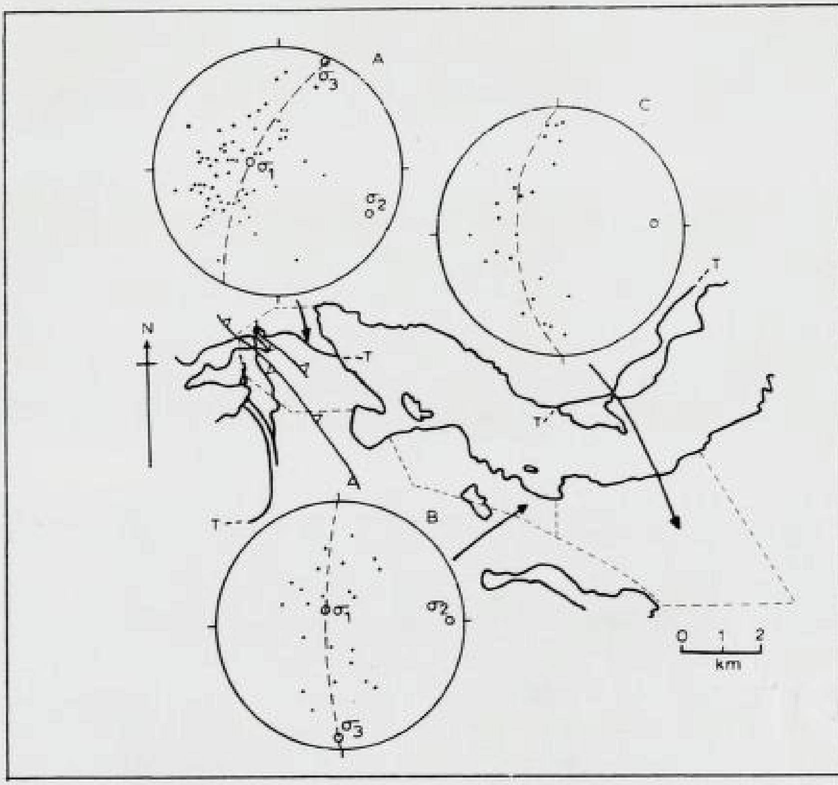

Fig. 10. F_3 pole girdles from three sub-areas at Bygdin with their calculated stress axis.

S_3 pole girdles are displayed and suggests that the diffuseness of the composite S_3 pole diagram for the whole area (Fig. 6, E) is probably a result of the variation in orientation from sub-area to sub-area. The axes of these girdles represent the line of intersection of the S_3 axial planes which fan about these axes. The regional girdle axis plunges at a low angle to the east. In the three sub-areas, the girdle axes plunge through the average orientation of the S_1 schistosity. This asymmetry produces divergent S_1/S_3 intersections and accounts for the complex F_3 axial distribution (Fig. 6, F). The geometrical relationships of this type of structure has already been described by Ramsay (1962 b) and need not be elaborated here.

The fact that some of the F_3 folds are true conjugate folds suggests that the deformation of the other orthorhombic folds may have been similar to that of the conjugate folds. Hence an analysis of the third fold stress axes can be

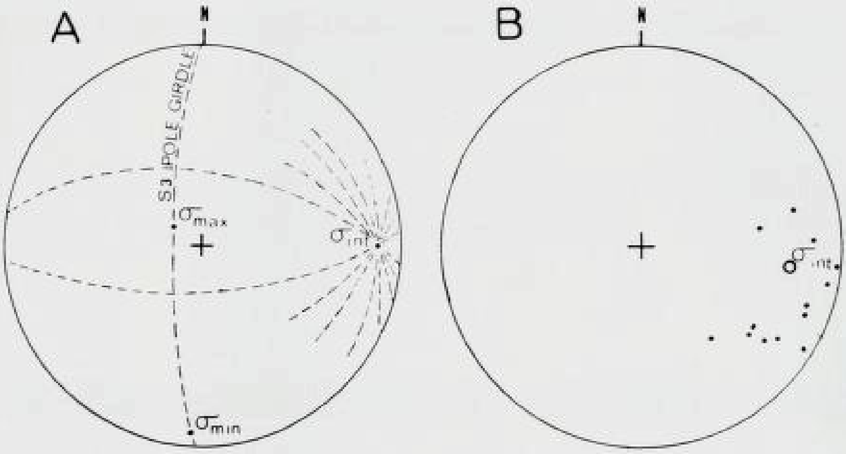

Fig. 11. A) Calculation of the stress axes from the S_3 pole girdle. B) Stereographic plot of the a_3-directions and the regional intermediate principal stress axis (σ_{int}) of the Third movement phase.

attempted using the symmetry of the F_3 structures (Johnson, 1956, Ramsay, 1962 b). The intermediate stress direction (σ_{int}) is assumed to lie along the line of intersection of the orthorhombic axial planes (and hence along the girdle axis). The σ_{int} directions for the three sub-areas of Fig. 10 thus plunge at low angles (10°-30°) to the east and southeast. An approximate girdle axis for the diffuse regional girdle of S_3 poles (Fig. 6, E) plunges at 25° towards 100° east. Now the maximum and minimum stress directions (σ_{max} and σ_{min} respectively) must lie normal to σ_{int} and hence lie somewhere within the S_3 pole girdle.

In a simple conjugate fold with two intersecting axial planes, σ_{max} and σ_{min} will bisect the angles between the two axial planes. In sub-areas A and B of Fig. 10 however, almost complete girdles are present. Distinct gaps in the girdles occur near the horizontal plane (representing gaps in the spread of the axial planes near the vertical). The σ_{max} axis must bisect the angle of this gap (Fig. 11, A) as it is highly unlikely that an axial plane in a conjugate fold would form parallel to the σ_{max} axis. In fact in the conjugate folds produced experimentally by Paterson and Weiss (1966) the axial planes are all at 60° to σ_{max}. The σ_{min} axis will then lie 90° from σ_{max} on the S_3 pole girdle. The σ_{max} axes for sub-areas A and B lie near the vertical and the σ_{min} axes near the horizontal in a north-south direction (Fig. 10). Comparison of the calculated stress axes, the movement sense, and the direction of maximum shortening (e.g. the horizontal direction in the fold of Fig. 9, D and the

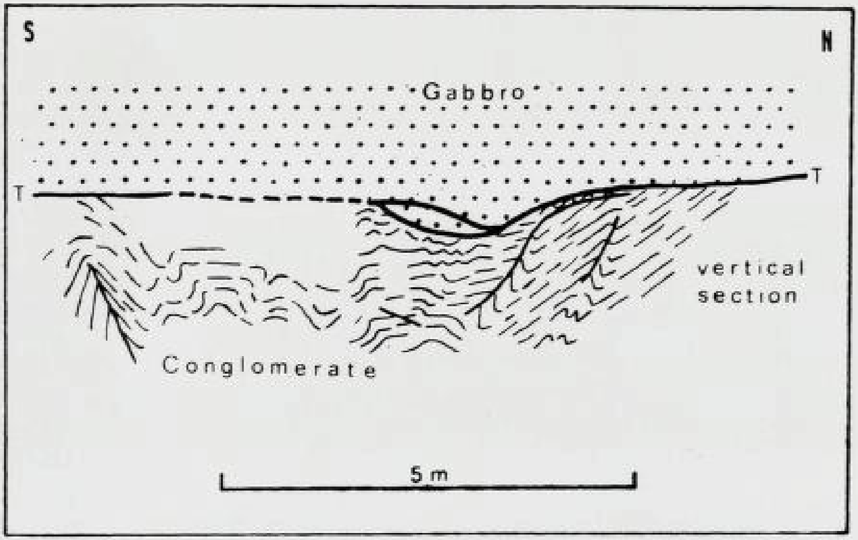

Fig. 12. Basal thrust plane truncating F_3 box folds on the Bitihorn crags.

vertical direction in the fold of Fig. 9, E) displayed by the F_3 folds in sub-areas A and B are essentially compatible. However, the movement sense and shortening directions of some of the F_3 folds of sub-area C (Fig. 10) do not fit with a vertical σ_{max} axis. The geometry of some folds in fact suggest a horizontal σ_{max} axis. Horizontal shortening directions were found elsewhere in the area with the best examples being displayed just below the thrust plane on the Bitihorn crags (Fig. 12). Within the Bygdin Antiform (sub-area A, Fig. 10) the axis of maximum compression appears to have been near vertical but to the west and east of the Antiform the maximum stress direction appears to have rotated into a horizontal direction. However, the intermediate stress direction (σ_{int}) appears to have been approximately constant in direction throughout the area, because the S_3 pole girdles for the three sub-areas of Fig. 10 have a relatively constant axial direction.

The third deformation was probably not a plane strain (Love, 1944, p. 45) but had some extension (either positive or negative) in the direction parallel to the σ_{int} axis, because the *a*-axes of the F_3 deformation group around the regional σ_{int} axis (Fig. 11, B). In spite of the fact that the directions of some of the *a*-axes are only approximate because of the difficulty of accurate measurement, and also that an *a*-axis does not necessarily represent the only line of movement within a rock but one of many, the grouping of *a*-axes about σ_{int} demonstrates that at least some of the movement took place along the intermediate stress direction.

(E) Fourth Movement Phase (F_4).

This is the last movement phase in the area to develop minor structures, all of which have a brittle style and seem to be intimately associated with joint formation. A joint analysis has been presented by Strand (1945) who recognized three joint sets in which the dominant trend was northeast-southwest, approximately normal to the F_1 and F_2 linear structures. The northeast cross joints are often represented by two conjugate joint sets intersecting in a small angle. The other two sets of joints trend north-south and west north west - east south east. Because all the joints cut through the structures of the first three movement phases, they are regarded as being later.

Small joint-drag folds can be found associated with the cross joints and because of the symmetry of the cross joints, the joint-drag folds may have a conjugate symmetry. They are also present where joints are absent. Other structures of the fourth movement phase include quartz or chlorite filled tension gashes which may or may not occur with the joint drag folds. The tension gashes, which have both dextral and sinistral movement senses, cut F_3 folds and thus post-date the third movement phase.

5. Major Structure

The only major structures recognized at Bygdin were formed during the first and second movement phases. That of the first movement phase is the basal thrust zone of the Upper Jotun Nappe, and those of the second movement phase are northwest trending antiforms and synforms.

(A) Basal Thrust Zone.

In the extreme southwest of the Bygdin area (Fig. 14) the thrust zone is defined by a clean cut thrust at the base of the gabbro. Here the thrust has a northeast strike (at 059°) with a dip varying between 14° and 47° to the northwest. The change in trend of the thrust trace as seen on the map from east-west to north-south in the crags below the Bitihorn summit, is purely a result of topography. The rocks only show slight crushing below the Bitihorn and phyllonitic rocks are absent. Most of the original textures of the gabbro are retained to within a few meters of the thrust plane.

The thrust plane at the base of the gabbro transgresses to a higher structural level 1½ km north of the Bitihorn and a second thrust plane continues northwards to define the basal thrust zone of the Jotun Nappe. A basal phyllonitized gabbro outcrops where the thrust plane at the base of the gabbro transgresses upwards. This is a fine grained schistose rock which can be traced upwards from the thrust for 10 m and gradually passes into less cataclastic gabbro. At

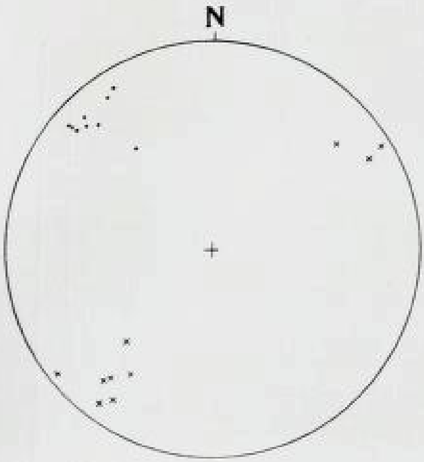

Fig. 13. Stereographic plot of the F_2 major structures. *Dots* - fold axes. *Crosses* - poles to axial planes.

the transgression, the thrust plane is refolded by antiform-A (Fig. 15) and the trace of the thrust swings round to a northwest trend. The thrust zone on the southwest limb of this antiform is composed of a series of thrust wedges which dip westwards below the gabbro. The structural sequence going down from the gabbro is:

(i) Conglomerate wedge closing towards the north.
(ii) Granite sheet wedging out towards the south.
(iii) Conglomerate and sparagmite.

The granite and gabbro probably form two separate thrust sheets separated by the conglomerate wedge. The conglomerate pinches out on the peninsula at the east end of Lake Bygdin. The contact to the north of this is not exposed but is probably a thrust zone.

The thrust at the base of the granite (unit ii above) can be traced northwards towards Bygdin defining the basal thrust zone of the Jotun Nappe. Near the Bygdin Hotel, the thrust is folded about major antiforms of the second movement phase (inset map of Fig. 15). This folding has produced steep or inverted dips in the thrust plane. To the northeast of the antiforms the thrust plane trace continues southeast into Lake Vinstra. The strike here is northwest with steep dips of 60° to 70° to the northeast.

A basal zone of phyllonite is present in the granite just above the thrust plane, and can be traced from the thin granite wedge on the west limb of antiform-A, round the antiforms at the Bygdin Hotel towards Lake Vinstra. The phyllonites vary in thickness from one meter in the core of antiform-A to 100 meters at Bygdin. In addition to the basal phyllonite zone, extensive phyllonite horizons are exposed up to 400 meters above the thrust plane to the north and northeast of Bygdin (Fig. 14).

Five thin slices of deformed conglomerate occur within the nappe to the north and northeast of the Bygdin Hotel. Three of the slices lie between 5 and 20 meters above the thrust plane in the hinge zone of the antiformal structures at Bygdin (inset map, Fig. 15). The other two slices are structurally about 300 meters above the basal thrust plane on the northeast limb of the Bygdin

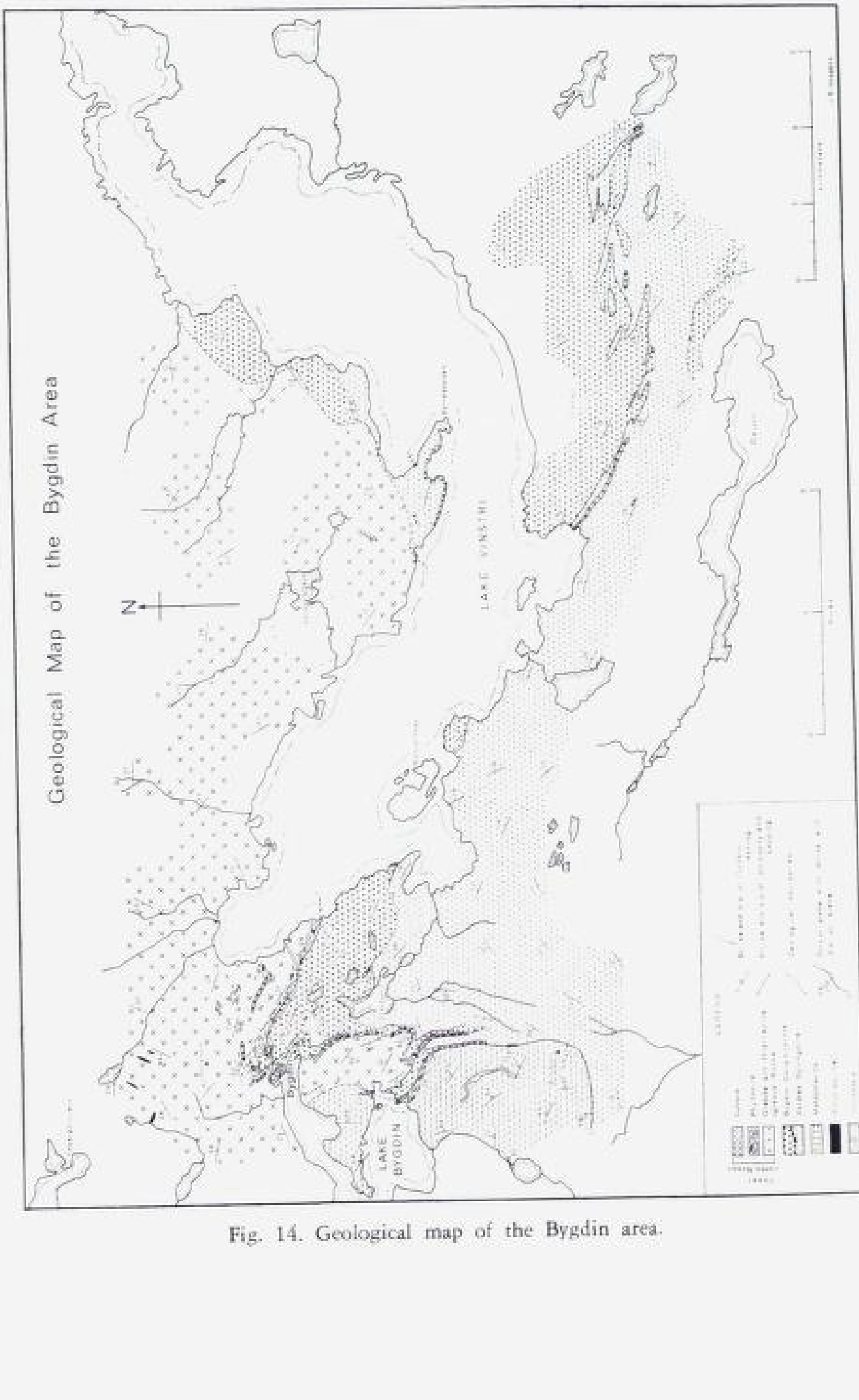

Fig. 14. Geological map of the Bygdin area.

Antiform. Four of the slices are shown on the geological map (Fig. 14). The fifth, which is too small to show on a geological map at the scale of Fig. 14, is depicted as the small folded lens (of F_2 age) on the southwest limb of antiform-H on the inset of map of Fig. 15. The slices are all believed to have been picked up during thrusting. It does not appear that they represent fold cores because the structural data do not indicate closures. The thrust plane trace reappears on the north side of Lake Vinstra on the Barnesodden peninsua, but is only exposed in one locality to the northeast of the peninsula.

The conglomerate and sparagmite just below the thrust plane have not been mylonitized. However, in the thrust zone near the Bygdin Hotel, some intermixing of conglomerate and igneous rocks has taken place. Thin bands of blue hornblende phyllonite (derived from igneous rocks) up to 10 cm thick, can be found within the conglomerate near the thrust plane. Within the phyllonite which lies just above the thrust plane, small fragments of deformed conglomerate pebbles can be found. These pebble fragments were probably picked up during the thrusting and phyllonite formation and were intermixed with the phyllonite. As recognized by Flinn (1961), the thrust plane near Bygdin is not a clean cut fault, but is defined by a lithological break rather than an obvious tectonic break.

In several outcrops on the Bitihorn crags, the thrust plane was found to truncate F_3 folds (Fig. 12). This indicates that the Bitihorn gabbro has undergone a post-F_3 phase of thrusting. This late movement could be a result of readjustment along the thrust plane during a late phase of the third movement period. This readjustment is similar to that reported Christie (1963) in the Moine Thrust Zone. However, kakirites or secondary mylonites are absent on the Bitihorn.

Major F_1 folds were not recognized in the area. The form of the F_1 minor folds (i.e. S- or Z-form, c.f. Fleuty, 1964, p. 475) varies from outcrop to outcrop. Now the form of minor parasitic folds must change as a larger fold of the same generation is crossed from one limb to the other. Hence the variation in the form of the F_1 minor folds at Bygdin suggests that F_1 folds exist on a larger scale than any folds found in the area. However, from the number of F_1 minor folds available, it is not possible to define any of these larger scale folds or determine their true dimensions. The largest F_1 folds visible in the area are those of Fig. 4.

(B) Major Second Folds.

Eight major F_2 folds are recognized in the area. The axial plane traces (A to H, Fig. 15) can be determined by using changes in the strike and dip

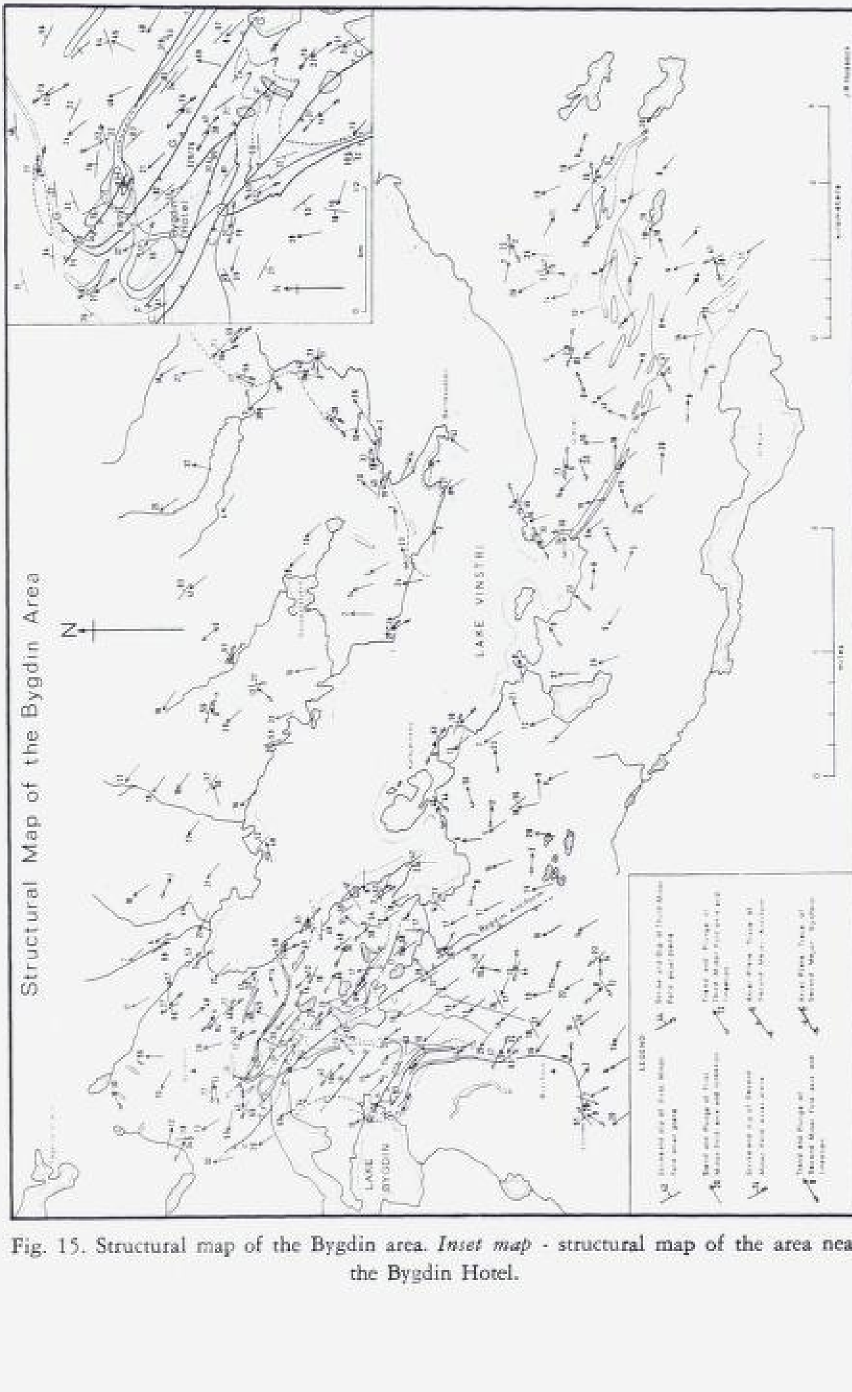

Fig. 15. Structural map of the Bygdin area. *Inset map* - structural map of the area near the Bygdin Hotel.

of S_1. In addition, the form of the F_2 minor folds (i.e. S- or Z-form) can be seen to change across the fold axial plane traces, proving that the major folds are contemporaneous with the minor folds and hence are of F_2 age. The largest fold of the area (Fold-C, Fig. 15) is referred to here as the Bygdin Antiform.

The Bygdin Antiform is disharmonic as can be seen when it is traced from the lower structural levels in the south of the area towards Bygdin. In the south the Antiform is an open fold (cf. Fleuty, 1964) with dips of 15° to 30° on each limb. It becomes tighter to the northwest and north of Stavtjern (inset map, Fig. 15) dips of up to 80° were recorded on the limbs. All other F_2 major folds of the Bygdin area are large scale parasitic folds (de Sitter, 1958) on the limbs of the main Bygdin Antiform. The axial plane traces of the other folds, apart from the Bygdin Antiform, can only be followed for short distances, and the folds themselves are open except for folds G and H.

The axial planes and fold axes of the large parasitic folds and of the Bygdin Antiform (which has been divided into three sub-areas along the axial plane trace because of its disharmonic style) have been calculated from the S_1 pole girdles and F_2 minor folds within each major fold (Fig. 13). The trends of the F_2 major structures are parallel to those of the F_2 minor structures (Fig. 6, C and D) and the whole system is homoaxial.

Evidence of refolding relationships is given by folds G and H (inset map. Fig. 15). The trace of G can be followed from the small tarns to the north of Stavtjern, passing northeast of the Bygdin Hotel. On the northeast limb of G, just north of the Bygdin Hotel, the thrust plane is folded about minor parasitic folds of antiform-G. The shape of these folds is consistent with the geometry of G. The trace of G then swings northeast to cross the trace of antiform-H. This refolding of G by H causes antiform-G to change from an upward closing fold (in the lower structural levels) to a sideways closing fold (as it crosses the core of H).

Antiform-H is excellently exposed in the island in the outlet stream of Lake Bygdin and has a style and orientation similar to the F_2 folds. In addition the minor parasitic folds on the limb of antiform-H are refolded by minor F_3 folds. It is suggested that antiform-H is of F_2 age and that it has refolded antiform-G during a later phase of the second movements. The causes of this refolding may be similar to those discussed by Wynne-Edwards (1963). This refolding of G by H also causes refolding of the axial planes of minor F_2 folds within antiform-G and results in the northeast girdle of S_2 poles in the composite plot of S_2 poles of the Bygdin area (Fig. 6- C). This girdle has an axis with a low plunge to the northwest and is parallel to the trend of the major and minor F_2 structures. Refolding of the second minor folds by a later major second fold is also present in antiform-A. This refolding is probably analogous to the refolding in antiform-H.

No major folds are recognized to the east of Bygdin. The Vinstra area is not folded on a major scale and has a regional dip of 20° to 30° to the northeast. In addition, major third folds are absent in the area. However, a plunge culmination of F_1 and F_2 axial structures exists in the conglomerate one km southeast of Bygdin. To the north of the culmination, F_1 and F_2 axes plunge northwest and to the south the axes plunge southeast. This culmination has a north or northeast trend which suggests that it may be a result of the third movement phase.

6. Metamorphic history

The metamorphic history of the area will only be summarized to indicate the metamorphic effects of the movement phases at Bygdin. Strand (1945) and Flinn (1961) have already discussed the metamorphic effects of the large scale thrusting of the Upper Jotun Nappe.

The effect of thrusting on the igneous rocks of the nappe was largely to reduce the grain size and to produce phyllonitic and schistose derivatives from originally coarse grained ?igneous rocks. The pre-thrusting amphibolite facies of the granitic rocks has been partly retrograded to greenschist facies but the granulite facies of the gabbro is largely untouched. During the greenschist metamorphism induced by the thrusting, muscovite and biotite were formed to produce the S_1 schistosity. In addition, porphyroblasts of epidote and sphene crystallized and the hornblendes in the granite have been partly or completely decolourized.

The thrusting also produced a greenschist facies metamorphism in the sediments below the nappe. Biotite and muscovite crystallized in the S_1 schistosity and porphyroblasts of epidote and sphene developed. The main greenschist metamorphism can be shown to pre-date the second movement phase as epidotes are now seen to be bent and broken in the F_2 minor fold hinges.

Both the F_2 and F_3 movement phases were accompanied by the recrystallization of biotite and muscovite in the S_2 and S_3 schistosities, but this recrystallization is not widespread and seems to be of a lower grade than the main greenschist facies metamorphism associated with the thrusting of the Upper Jotun Nappe.

7. Conclusions

Four movement phases are recognized by the writer in the area around Bygdin. The first movement phase was synchronous with the thrusting of the Upper Jotun Nappe over the conglomerate and the Valdres Sparagmite.

However the minor structures associated with this thrusting give no clue to the direction of thrust movements. This thrusting caused retrogression of the amphibolite and granulite facies of the nappe to greenschist facies and produced cataclastic textures in the nappe rocks. Extensive phyllonite horizons were formed within and at the base of the nappe and folding of these phyllonite bands by F_1 folds about an S_1 axial plane cleavage suggests that the phyllonitization was prior to the formation of the F_1 minor structures. S_1 is the first schistosity of the area and is parallel to the thrust plane throughout most of the area. The deformed pebbles of the Bygdin Conglomerate now have a flattened and elongated form with their long axes lying in the schistosity. Most of the pebble deformation was accomplished during the first movement phase. The F_1 lineation and pebble elongation direction have a regional northwest trend but on the Barnesodden peninsula in Lake Vinstra, the northwest F_1 lineations swing into east and northeast trends. The cause of these changes in the trend of L_1 and the relationship of L_1 to the movement of the nappe will be presented in a future paper describing the deformation of the Bygdin Conglomerate.

The thrust plane and the deformed pebbles are folded by the later F_2 major and minor folds which include the Bygdin Antiform and its associated major and minor parasitic folds. The F_2 axial trend is exactly parallel to the F_1 axial trend both on the major and minor scales.

The third movement structures occur only on a minor scale and have an orthorhombic or conjugate symmetry. The stress axes of the third deformation are tentatively deduced from the symmetry of the third folds. Within the Bygdin Antiform, the maximum principal stress appears to have been near-vertical with the minimum principal stress near-horizontal in a north-south direction. Throughout the area, the intermediate principal stress direction appears to have had a low plunge to the east or southeast. However, to the east and west of the Bygdin Antiform, the maximum and minimum principal stress directions would appear to have rotated through 90° bringing the minimum stress direction to a near-vertical position.

A rejuvenation of movement along the basal thrust plane of the Jotun Nappe took place on the Bitihorn after the formation of the F_3 folds.

Joint formation and associated minor structures were formed during the last movement phases recognized at Bygdin. Movement along the joint surfaces formed joint-drag folds and tension gashes. Because of the conjugate symmetry of the cross-joints, dextral and sinistral folds and tension gashes were formed.

Acknowledgements

The research described in this paper was carried out with the aid of a Department of Scientific and Industrial Research Studentship which is gratefully acknowledged. The author would like to thank Dr. M. R. W. Johnson for his helpful supervision during the research and Prof. G. Wilson, Prof. J. G. Ramsay, and Dr. D. Elliott for their critical reading of the manuscript.

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Geologi og petrografi på Nord-Karmøy

(Geology and petrology on northern Karmøy)

Av Hans-Peter Geis.

Summary

The author gives a description of rocks and tectonics on the northern and central part of the island of Karmøy, West-Norway, and concludes with an interpretation of the observations and a geologic history of the area.

The following rocks are described, proceeding from the youngest to the oldest, that is from the north to the south:

Greenschist, originally tuff, a slaty greenish rock, consisting of oligoclase, quartz, chlorite and biotite.

Lava greenstone. The main type is a more massiv greenish rock with albite-oligoclase and actinolite as the main component and quartz, biotite, chlorite, calcite and ore as accessories. There are typical lava structures to be found such as pillows and other proof of lava flow. The megascopic picture of the rock varies, holocrystalline, epidote-rich and quartz-rich types are to be found. Greenschist, quartzite and quartz-keratophyric rocks occur as concordant layers.

Amphibolite, a coarse rock, consisting mainly of albite-oligoclase and actinolite, appears lowest in the sequence. The coarse rock is always intercalated with rocks of the lava greenstone type, with the coarse constituent as layers, lumps and veins.

These rocks are intruded by

epidiorite, a rock similar to amphibolite but with granitic texture and sharp, disconcordant contacts,

trondhjemite, dikes and lumps of a greyish-white rock consisting mainly of albite and quartz with intersertale texture. Accessories are epidote, actinolite, chlorite, ilmenite, leucoxene, pyrite, limonite, biotite.

Concordant deposits of pyrite, sphalerite, chalcopyrite, pyrrhotite and magnetite are interpreted as submarine-exhalative. One deposit is exhausted, another one is mined, both consisting of pyrite with sphalerite and chalcopyrite. The other deposits are of no commercial value.

Youngest is the quarternary till: boulders, sand and clay.

The tectonic processes have given rise to a syncline that in general follows the eastern shore line of Karmøy, with NNW-plunge. West of the syncline the strike is northwesterly with increasing dip angle westover. East of the syncline the strike is northerly. The lineation of the rocks goes parallel with the above mentioned syncline (B-lineation), another one (B') goes oblique to the B-lineation.

There are ruptures that strike N-S with more or less horizontal movements and others with strike NW-SE that are overthrusts. Both of them have to some extent formed the topography.

All of the rocks are metamorphosed in the "greenschist-facies". The following local facies types have been observed from the north to the south:

1. albite/oligoclase - quartz - chlorite - biotite - epidote
2. albite/oligoclase - actinolite - epidote (quartz - biotite - chlorite)
3. same as 2, but coarse crystalline.

It is not difficult to interpret the greenschist and the lava-greenstone as tuff and lava, respectively. The epidiorites are small intrusives. It is more difficult to give an interpretation of the amphibolite. Other places in Norway these rocks are interpreted as intrusions. The author is of the opinion that there is little proof for such an interpretation on Karmøy. He thinks that they are higher metamorphosed lava greenstones and the trondhjemite might be a mobilisate in connection herewith.

The age of the rocks is Lower Ordovician.

I. Overblikk over Karmøy på grunnlag av tidligere publikasjoner

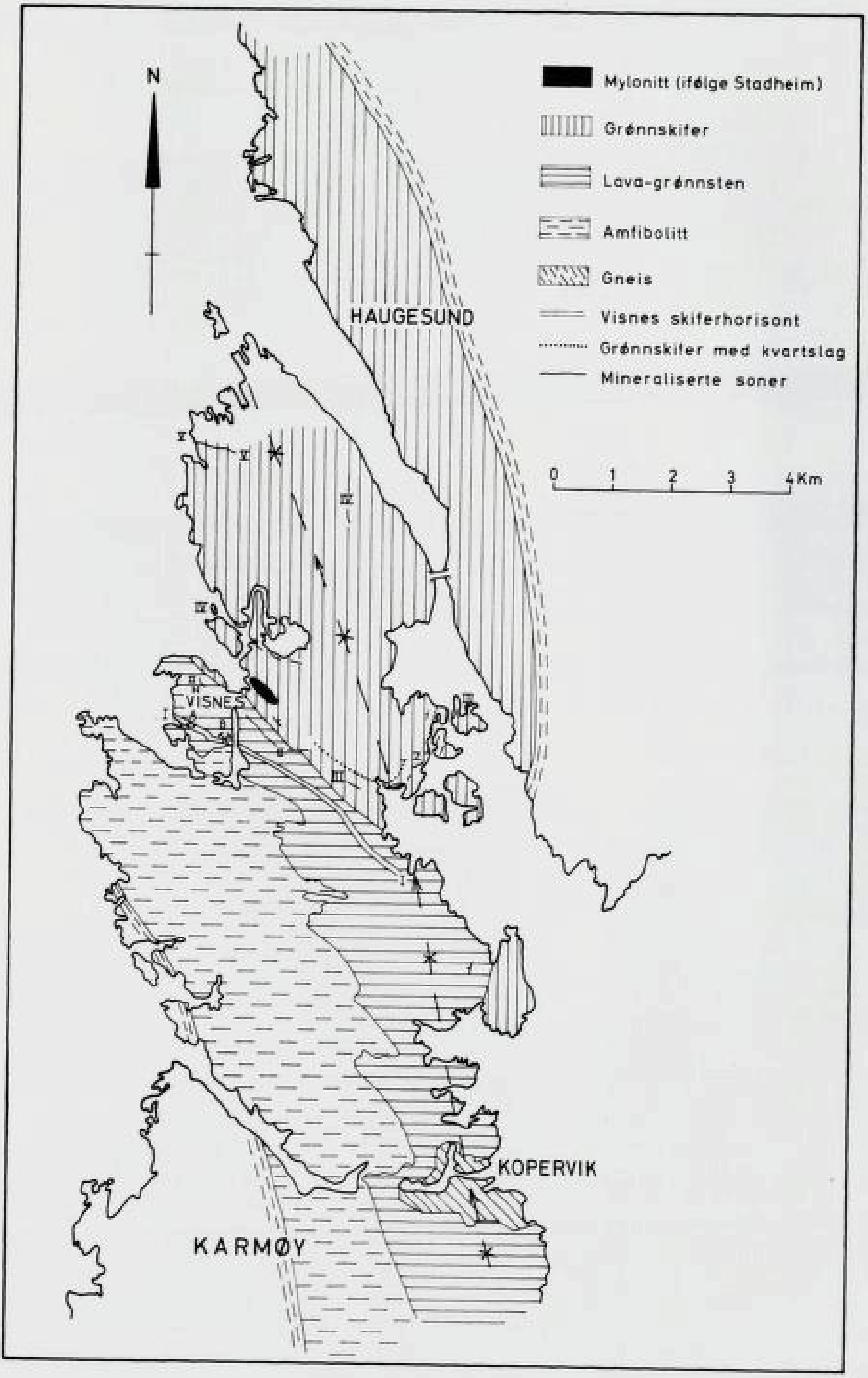
Karmøy i sin helhet er tidligere bare undersøkt en eneste gang. Det var av Reusch, og han skriver om dette i sin bok "Bømmeløen og Karmøen med omgivelser". Reusch påviste denne gang at bergartene på øya stryker i NV-retning. Han inndelte bergartene i

a) en sone med forskjellige gneisbergarter i SV
b) i midten et belte med "dioritiske bergarter"
c) i nordøst på Karmøy og det tilstøtende fastland et område med "grønne, krystallinske skifere" (på fastlandet igjen omgitt av gneis)
d) på nordspissen av Karmøy "sortaktig og grå lerglimmerskifer"
e) i sør ligger et "polygent konglomerat" på gneisene.

Senere laget Goldschmidt (1922) følgende inndeling i Stavanger-området:

Sedimentbergarter (konglomerat) på Karmøy (muligens Downton)
Avdeling av de grønne skifere (Overordovicium til Silur)
Fylittavdeling (Kambrium og Underordovicium).

"Saussuritgabbroen" på Karmøy har etter hans oppfatning tydelig karakter av en dypbergart.


N
Mylonitt (ifølge Stadheim)
Grønnskifer
Lava-grønnsten
Amfibolitt
Gneis
Visnes skiferhorisont
Grønnskifer med kvartslag
Mineraliserte soner
0 1 2 3 4 Km
HAUGESUND
VISNES
KOPERVIK
KARMØY

I konglomeratet som inneholder rullesten av de andre bergarter på Karmøy, fant Isachsen 1939 noen fossiler (Broch, Isachsen, Isberg & Strand 1940). På grunnlag av disse ble konglomeratets alder fastslått til overgrensen for Ordovicium.

Holtedahl (1953) innordner grønnstenene på Vestlandet lavere enn Goldschmidt, nemlig Underordovicium.

Endelig foreligger det et upublisert kart av Stadheim fra NGU's bergarkiv. Han utførte i 1940-årene geologisk kartlegging av et mindre område vest for Visnesvann i målestokk 1 : 5000.

Kolderup (1931) beretter ganske kort om noen observasjoner fra Karmøy, deriblant nevner han en synklinal på østsiden av Karmøy.

II. Kartleggingen

De mest inngående undersøkelser ble av forfatteren utført i området omkring Visnes og omkring den gamle Sørstokke-gruve S for Kopervik, hvor det ble utført kartlegging på flyfoto i målestokk 1 : 5000.

Grensen mellom lava-grønnsten og amfibolitt ble kartlagt på flyfoto over hele Karmøy, det samme gjelder grensen mellom lava-grønnsten og grønnskifer. I amfibolitt-området ble det utført noen oversiktsbefaringer og dens ligg-grense ble kartlagt i 1 : 50 000. En innleiring på Lande ble kartlagt, en annen i Torvastad (Bjørgene) ikke. Nordgrensen av grønnskiferen ble overtatt etter Reusch (1888).

Fig. 1. Geologisk kart over Nord-Karmøy samt tilstøtende fastland (foregående side).
Geological map of northern Karmøy and the nearby mainland.

grønnskifer	*greenschist*
lava-grønnsten	*lava greenstone*
amfibolitt	*amphibolite*
gneis	*gneiss*
Visnes skifer	*Visnes schist*
grønnskifer med kvartsårer	*greenschist with quartz veins*
mineraliserte soner	*mineralised zones*
I Visnes malmsone	*Visnes ore zone*
A. Gamlegruben	*Old mine*
B. Rødklev grube	*Rødklev mine*
II malmsone	*ore zone*
Haugesundsynken — Patmos — Klevensynk	
III malmsone	*ore zone*
Våge — Bukkøy	
IV malmsone	*ore zone*
Vikingstad — Storesund	
V Dale malmsone	*Dale ore zone*
Dale synker	*Dale shafts*

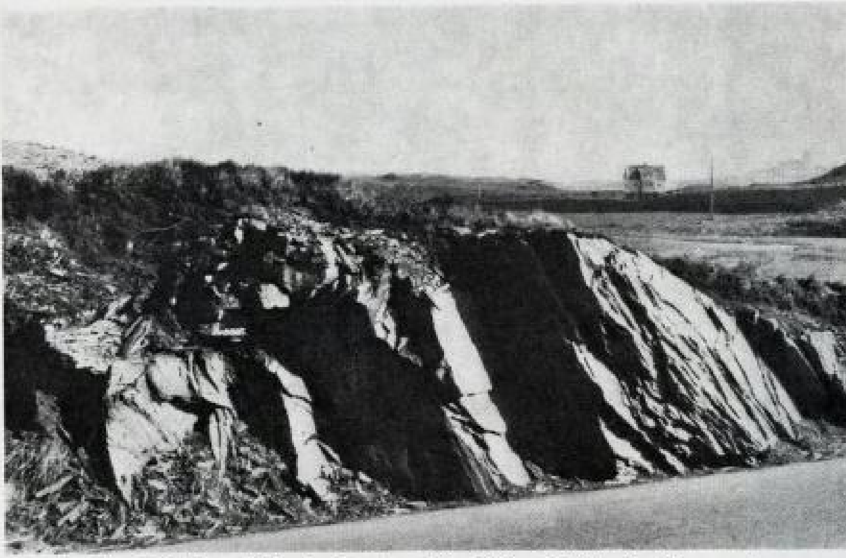


Fig. 2. Frisk veiskjæring i grønnskifer. Riksvei 14 øst for Torvastad.
New road cut in greenschist. Highway 14 east of Torvastad.

Forfatteren vil her takke Metallgesellschaft AG. ved Dipl.Ing. Reddehase samt A/S Vigsnes Kobberverk ved direktør Leiv Løvold for den velvilje de viste under arbeidet og for tillatelsen til å publisere resultatene. Prof Chr. Oftedahl og statsgeolog M. Gustavson takkes for kritisk gjennomsyn av teksten og noen kommentarer.

III. Bergartsbeskrivelser

Den overveiende del av forfatterens undersøkelser ble utført innenfor de bergartene som Reusch kalte for "dioritisk bergart" og bare i mindre utstrekning i de "grønne, krystallinske skiferes" område. På alle de andre bergartene ble det kun utført oversiktsbefaringer. Den av Kolderup (1931) nevnte glimmerskifer i Torvastad ble således ikke skilt ut. Da undersøkelsene måtte utføres meget nøyaktig i forbindelse med malmletingsoppgaver, var det av oppdragsgiveren på forhånd bestilt flyfotografier i målestokk 1 : 5000.

Det vesentligste resultat av undersøkelsene er at det ble utført en todeling av Reusch's "dioritiske bergarter", nemlig i

Grønnsten med lava-strukturer (Lava-grønnsten) og
Amfibolitt.

En bergart med liknende egenskaper som amfibolitten opptrer tydelig diskordant i lava-grønnstenen. Den ble kalt "epidioritt". Med hensyn til den areal-

Fig. 3. Horisont med konkordante kvartsårer i grønnskifer. Gloppe.
Layer with concordant quartz veins in greenschist. Gloppe.

messige fordeling av bergartstypene henvises til fig. 2 i Geis (1965).

I det følgende skal det gis en beskrivelse av de forskjellige bergarter samt deres innleiringer.

a) Grønnskifer (Reusch's "grønne, krystallinske skifere").

Dette er de yngste bergartene i vårt område. De ligger over lava-grønnstenen. Grønnskiferen har en grønngrå farge og er tynnskifrig og tynnplattig (fig. 2). I motsetning til lava-grønnstenen lar den seg kløve og man får glatte og jevne bruddflater. I uforvitrete blotninger er det lett å holde de to grønnstenene fra hverandre. Men dette kan være meget vanskelig hvor bergartene er forvitret eller polert av isen.

Grensen mellom dem er ikke alltid skarp. På Skeie f.e. har man inntrykket av at de veksellagrer slik man venter det i grenseområdet mellom lava og tuff. Grensen ble i dette tilfelle tegnet hvor vedkommende bergartstype overveier.

Det finnes *kvartsittiske* innleiringer av samme type som de som lenger nede vil bli beskrevet fra lava-grønnstenen. Fra Skeisvoll over Våge til Gloppe går en horisont med *konkordante årer av kvarts* (fig. 3). Den viser meget pent lagenes ombøyning og den flatere aksestilling på østsiden av Karmøy. I dens

Amfibol og plagioklas i noen slip

Slip nr.	Bergart	$Z \wedge c$	Amfibol størrelse (mm)	Amfibol % i slip	$X' \wedge (010)$	Plagioklas størrelse (mm)	Plagioklas % i slip
30	lava-grønnsten	17,0	0,04 x 0,5	35			
32	»				— 9,4 (An 12)	—	50
42	»	17,7	0,125	90			
60	»	21,0	0,5 x 0,125	25	— 14,7 (An 5)	—	60
172	»				— 13,8 (An 6)	—	30
91	lava-grønnsten keratofyrisk		ingen		— 14,4 (An 5)	1,5 x 0,5	70
99	epidotisert lava-grønnsten	16,0	< 1	5	— 11,2 (An 10)	0,025 x 0,006	25
29	holokryst. lava-grønnsten	20,3	< 1	30	— 18,4 (An 0)	1,5 x 0,5	60
22	Visnesskifer				— 17,7 (An 0)	0,5 x 0,025	80
48	»	17,5	0,025 x 0,125	15			
112	»	20,2	0,007 x 0,1	25	— 18,6 (An 0)	—	50
83	kvartskeratofyr i lava-grønnsten	16,6	0,05 x 3	5	— 22,6 (An 0)	—	70
71	grønnskifer		ingen		— 10,6 (An 11)	0,025	60
6	epidioritt	20,6	1,5 x 2,5	25	— 12,4 (An 8)	—	50
36	»	16,2	1,5 — 2,5	70	ikke mulig å identifisere		
11	amfibolitt	12,6	5 x 10	50	— 11,0 (An 11)	—	20
14	»	17,8	5 x 10	60	— 15,3 (An 3)	—	30
34	»	19,8	2 — 2,5	50	— 17,0 (An 0)	—	40
95	trondhjemitt	10,2	1 x 0,25	5	— 24,9 (An 0)	—	85

forlengelse mot NV er kvartsen mer kompakt anriket og inneholder mm-tykke pyrittbånd som tidligere ga anledning til en del skjerpevirksomhet.

Bortsett fra de nevnte innleiringer viser selve grønnskiferen nord for gården Visnes visse variasjoner. Her opptrer tildels en sterk *epidotisering* som delvis har ført til dannelsen av epidot-biotitt-gneis.

Tynnslipene av grønnskifer viser en fin parallelltekstur med vekslende mengder kvarts og plagioklas (oligoklas 11 An). Bånd av kloritt og biotitt og årer av granoblastisk kvarts og plagioklas gir det hele også en makroskopisk parallellstruktur.

Fig. 4. Lava-grønnsten med putestrukturer. 200 m nord for Visnes havn.
Lava greenstone with pillow structures. 200 m N of the harbour of Visnes.

Epidot-biotitt-gneisen består makroskopisk av biotittlag i en grunnmasse med skitten-gul farge. Under mikroskopet ser man at grunnmassen består av kvarts med fin granoblastisk tekstur, heri er anriket slirer av epidot og noe fibrig aktinolitt.

Grønnskifer av en liknende type som på Karmøy har Strand (1958) beskrevet fra Helgeland.

b) Lava-grønnsten.

Dette er den bergarten som ble studert mest inngående. Heri ligger Visnes-malmsonen med de mest betydningsfulle malmforekomster på Karmøy. Bergarten dekker den nordøstlige del av Reusch's "dioritisk bergart". Den danner et NV-SO-gående belte av 500-1000 m bredde over størsteparten av Karmøy (se også avsnittet "Tektonikk og metamorfose"). Den har en rettlinjet grense mot grønnskiferen i hengen (NO) og en buktet og meget uklar begrensning mot amfibolitten .

Bergarten er masseformet med ujevne brudd- og avsondringsflater. På forvitret overflate er fargen grålig, på friskt brudd grønngrå. Bergarten er tett, stedvis med anrikninger av små biotitt- eller klorittflak. Også kalkspatårer sees leilighetsvis. Bergarten forvitrer slik at flytestrukturer som i flytende lava blir synlig. N for Visnes havn kan man i veiskjæringen noen steder se putestrukturer (fig. 4). Putene er imidlertid langstrakt etter hovedfoldningsaksen og er ikke så karakteristisk som f.e. i Lahn-Dill-området i Tyskland.

Ved hjelp av parallellstrukturen som samtidig er skifrigheten er det mulig å oppdage foldestrukturer.

Under mikroskopet ser man i det vesentlige en parallelltekstur av plagioklaslister (albitt-oligoklas 5-11 An) og av fibrig aktinolitt. Det hele er strødd over av epidotkrystaller. Aksessorisk finnes: kvarts, kloritt, biotitt, kalsitt, erts. Årer av kalsitt eller epidot og kvarts trekker seg gjennom det hele. Såvel biotitt som kloritt er helt frisk og ofte intenst sammenvokset med aktinolitt.

Bergartens sammensetning og utseende varierer innenfor temmelig vide grenser. Dette gjelder spesielt for gehalten av aktinolitt. Den kan utgjøre hovedmengden, men den kan også forsvinne nesten fullstendig, slik at det oppstår en skifer med *keratofyrisk til kvarts-keratofyrisk* karakter slik den finnes et stykke vest for Rødklevgruven.

Den normale lava-grønnsten kan vise et *mikroskopisk bilde forskjellig* fra det ovenfor beskrevne. En slik variant ble spesielt funnet nord for veien fra Visnesvann til Visnes havn. Det eneste makroskopisk påfallende er små epidotflekker av noen mm diameter. Under mikroskopet viste det seg at bergarten overveiende består av meget finkornet plagioklas (albitt 10 An) og mindre kvarts med granoblastisk texstur. Det hele er gjennomsatt av en masse små epidotkorn. Kloritt og biotitt som tildels er intenst sammenvokset, danner slirer og flekker. Mellom alt skyver seg bittesmå aktinolittfibrer. Disse danner delvis også større slirer. Aksessorisk opptrer: større aktinolittkrystaller, leukoksen, erts, allanitt (etter bestemmelse av Dr. Stam). Plagioklas- og kvartskornene inneholder vanligvis jevnt fordelt bittesmå inneslutninger som ikke har latt seg bestemme. Om aktinolitt og plagioklas se også tabell.

Den bergarten N for gamle Visnes gruve, som av Reusch (1888) ble betegnet som "*konglomerat*" hører også til denne type. Den særpreges av store klumper og slirer av epidot med kvarts. Disse står frem når bergarten forvitrer og gir den et visst konglomeratliknende utseende (fig. 5). Ved en sammenlikning med de ekte konglomeratene på Sør-Karmøy og på Bømlo oppdager man snart at det her må dreie seg om noe annet, nemlig en spesiell type epidotisering. Dessuten er også den muligheten tilstede at det dreier seg om et agglomerat.

I en annen varietet er *nesten hele bergarten erstattet med kvarts* i granoblastisk tekstur. Innimellom sitter det små mengder med stråleformet plagioklas. Magnetittkrystaller er strødd over det hele. Forøvrig er både det opprinnelige makro- og mikrobilde bevart. Mandelsten-hulrom er fylt igjen med litt grovere granoblastisk kvarts.

Fig. 5. Utvitrete epidotrike partier i lava-grønnsten. 400 m NO for Visnes havn.
Epidote concentrations in lava greenstone which are more resistant to weathering. 400 m NE of the harbour of Visnes.

Videre finnes det slireaktig og gangaktig *holokrystallinske partier* som under mikroskopet består av plagioklas (albitt 0 An) og aktinolitt i intersertal tekstur. Det finnes alle overganger fra en fibrig aktinolitt-kloritt- substans — hvor konturene av en porfyroblast såvidt kan skimtes — til tydelig intersertaltekstur. Plagioklaskrystallene er som oftest fylt med små epidotkorn. Det er påfallende at plagioklasene i tydelig holokrystallin-intersertale partier inneholder mindre epidot, og her er også ilmenitt anriket. Delvis er denne omvandlet til leukoksen. Leukoksenen opptrer også uten synlig sammenheng med ilmenitt. Aksessorisk finnes: kloritt, kalsitt, kvarts.

Det finnes også en rekke *konkordante innleiringer*. Først og fremst må nevnes *tynnskifrige grønne bergarter* (Visnesskifer). I disse ligger gamle Visnes- og Rødklev-gruvens malmleier. Under mikroskopet viser disse bergarter plagioklas (albitt 0 An) og kloritt med parallell til granoblastisk tekstur. Jevnt fordelt over det hele opptrer epidot, mens aktinolitt bare foreligger i meget små mengder. Aksessorisk finner man dessuten kalsitt, kvarts, leukoksen. Omkring de to gruvene er innenfor den samme sone også andre bergarter representert som bare ble observert på liggsiden av malmen. Det er hovedsakelig hvite til lysegrå skifere som består av meget finkornet kvarts samt kloritt i parallell anordning, dette er gjennomsatt av noen årer med grovere kvartskorn (aksessorisk pyritt, leukoksen, epidot). Innleiringer som likner på lapillituff ble iakttatt. Under mikroskop viste de lapilli-liknende stedene seg å bestå av plagioklas-aggregater som tildels er sterkt kalsittisert og epidotisert.

En annen gruppe av *innleiringer* er meget *kvartsrike* og disse må vel tydes som sedimentære dannelser, muligens delvis kjemiske utfellinger på havbunnen. Det kan ikke sies noe om utgangsproduktene er tilført hydrotermalt eller om de ble anriket i sjøvannet på annen måte. Bergarten kalles vanligvis for "blåkvarts", mens Reusch (1888) kalte den for "kvartsit". Den har blågrå farge og finnes hovedsakelig langs med kysten omkring Visnes; mot SO er den sjeldnere. Da det opptrer en del slike linser omkring malmforekomstene ble de tidligere brukt som "ledehorisonter" under malmletingen. Som tynnslipene viser, er blåkvartsen omkring malmleiene faktisk rene kvartsbergarter, mens en stor del av de utenforliggende er av kvartskeratofyrisk sammensetning. Under mikroskop kan man se jevne overganger fra den ene til den andre variant. Det er mulig å stille opp en rekke som begynner med en bergart som bare består av granoblastisk kvarts med antydet paralleltekstur, overstrødd med noen bittesmå magnetittkrystaller; aksessorisk sees epidot. Rekken fører videre over bergarter som består av finkornet granoblastisk kvarts og plagioklas (Albitt 0 An) med vekslende kloritt- og epidotgehalt og med noen årer av grovere kvarts. Rekken slutter med en bergart av fint sammenfiltret kvarts-plagioklas-grunnmasse med isprengt aktinolitt, epidot, biotitt og erts (magnetitt og pyritt). Aksessorisk ligger heri titanitt og rutil. Den siste type viser en viss likhet med kvartskeratofyriske ganger sør for Visnes. Disse er imidlertid makroskopisk tette og av hvit farge med noen klorittiske slirer.

I lavagrønnstenens område opptrer grønnsten også *gangaktig*. Både makro- og mikroskopisk er det en stor likhet med de før nevnte grønnstensvarianter. En unntakelse er en lagergang N for Visnes som det var mulig å følge over 600 m lengde. Grunnmassen i denne er fullstendig erstattet med meget finkornet granoblastisk kvarts samt litt epidot og aktinolitt. Som innsprengninger opptrer feltspatkrystaller som er delvis fortrengt av epidot og granoblastisk kvarts. Konturene av disse er både makro- og mikroskopisk godt synlige.

c) Epidioritt.

Noen steder ble det observert holokrystallinske bergarter som støter diskordant mot de omgivende bergarter, man må altså anta at de er av intrusiv opprinnelse. Bergartene ble funnet på følgende steder: V for gamle Visnes-gruve, NV for Rødklev-gruve, 700 m NV for gamle Visnes-gruve og 300 m S for Rødklev-gruve. Sørgrensen av epidioritten NV for Rødklev-gruven ble på-

Fig. 6. Slirig-lagig amfibolitt. Kvalavåg.
Amphibolite, consisting of coarse grained and fine grained layers. Kvalavåg.

truffet på 40 m- og 150 m-etasjene i gruvens Vestkis-parti. Den samme bergart finnes også på etasje 150 og 210 i gruvens Østkis-parti.

Jevn granittisk tekstur, skarpe grenser mot de omgivende bergarter og massiv avsondring er karakteristisk for epidioritten. Dette utgjør også dens forskjell fra amfibolitten (som skal omtales i neste avsnitt). I randsonene foreligger ofte sterk forskifring.

Under mikroskop viser det seg at bergarten har en intersertaltekstur av frisk aktinolitt og fullstendig gjenfylte plagioklaskrystaller (albitt 8 An). Fyllmassen er en uklar blanding av titanitt, epidot, kvarts, kalsitt, aktinolitt-trevler, kloritt, apatitt. Større aktinolittaggregater består i sitt indre delvis av kloritt.

De forskifrete partiene ser under mikroskop prinsipielt like ut. Men det hele er her gjennomsatt av fintrevlet aktinolittsubstans og aktinolittkrystallene er mer trevlete.

d) Amfibolitt.

Lava-grønnstenens ligg-grense er ikke skarp, men mot liggen opptrer det flere og flere grovkrystallinske partier i linse- og slireform. I dem ser man med det blotte øye uregelmessige aktinolitt-individer i en hvit, uklar feltspat-

Fig. 7. Breksjert lava-grønnsten med trondhjemittisk kittsubstans. 2 km vest for Kopervik.
Brecciated lava greenstone, cemented by trondhjemite. 2 km W of Kopervik.

mellommasse. Aktinolittkrystallenes størrelse varierer mellom noen mm og 3 cm. I hele amfibolitt-området forekommer partier av lava-grønnsten-type i like stor mengde som grovkrystallinske partier. Større arealer med bare grovkrystallinsk amfibolitt finnes ikke. Grensen ble på kartet trukket hvor det ikke bare opptrer enkelte, men mange grovkrystallinske partier. De grovkrystallinske og de tette partier danner for det meste tydelig avgrensete lag som alternerer med hverandre (fig. 6). Men det er heller ikke sjeldent at grovkrystallinske partier gjennomvever de tette i form av slirer og uregelmessige klumper som ser ut som lateralsekresjoner. Enkeltkrystallenes størrelse er over hele området meget variabelt, d.v.s. teksturen kan på hver plass variere mellom meget tett og sterkt grovkrystallinsk uten noen generell økning av kornstørrelsen mot liggen.

Det er denne bergart som av Goldschmidt (1922) ble kalt "Saussuritgabbro". Liknende bergarter opptrer også andre steder i Kaledonidene og omtales under samme navn. På Karmøy syntes jeg imidlertid ikke det var riktig å bruke dette navnet da man med gabbro mener en basisk dypbergart. Mineralsammensetningen viser at det er en intermediær bergart, og jeg tolker den heller ikke som en dypbergart. Det henvises i denne forbindelse til det som er anført nedenfor. Navnet "amfibolitt" passer etter min mening bedre, iallfall er det nøytralt med hensyn til bergartens dannelse.

Fig. 8. Amfibolaggregater som svever i trondhjemitt-substans, har omriss av breksjebruddstykker. 2 km vest for Kopervik.

Aggregates of amphibole — with contours similar to the brecciated lava greenstone — cemented by trondhjemite. 2 km W of Kopervik.

Den *grovkrystallinske* amfibolitt i egentlig forstand har en temmelig grov subofittisk tekstur (Tyrrell 1926) i hvilken plagioklas (albitt-oligoklas 0-11 An) og aktinolitt er nesten grafisk sammenvokset. Begge to er tydelig vokset gjennom hverandre og det er ikke mulig å avgjøre aldersforholdet. Aktinolitten viser pleokroisme og er ofte vokset sammen med og fylt med kloritt og kalsitt slik at den ved studium med en nikol ser ut som et enhetlig individ. I aktinolittsubstansen sitter små epidotkorn i mindre antall. Også rester av ilmenitt, som er omgitt av leukoksen, sitter ofte i hornblenden. Feltspatene er vanligvis friske med tydelig tvillinglamellering, men en stor del er også fylt med små mineralkorn. Disse består overveiende av epidot, dernest aktinolitt, fine aktinolitt-trevler, apatitt og rutil. Epidotinneslutningene synes selv å ha inneslutninger som imidlertid ikke kunne bestemmes.

Meget interessante er noen iakttakelser fra *grenseområdet mellom amfibolitt* og lava-grønnsten. For det første finnes her trondhjemittene som skal omtales i neste kapittel. For det andre ble det noen plasser i lava-grønnstenen iakttatt feltspatporfyroblaster i grønngrå grunnmasse. Endelig er — spesielt V for Kopervik — lava-grønnstenen i grenseområdet breksjert (fig. 7) og binde-

Fig. 9. Trondhjemittganger. 350 m sør for Visnes havn.
Trondhjemite veins. 350 m S of the harbour of Visnes.

midlet består av trondhjemittsubstans. Litt lenger borte finner man en bergart med anrikninger av grovkrystallinsk amfibol i trondhjemitt-grunnmasse. Disse anrikninger har formen av slike breksje-bruddstykker (fig. 8).

Mikroskopisk består grunnmassen i *lava-grønnstenen med feltspatporfyroblaster* av aktinolitt og feltspatkrystaller som nesten fullstendig er fylt igjen med epidot og aktinolitt-trevler i intersertal- eller parallell anordning. Dertil kommer større mengder leukoksen og litt fri epidot. Som innsprengninger sitter heri feltspatporfyroblastene som imidlertid er nesten fullstendig fylt med epidot.

Den *breksjerte lava-grønnsten* består av granoblastisk aktinolitt og plagioklas, overstrødd med leukoksen, litt epidot og spor av ?rutil. Bindemiddel er finlamellert, frisk plagioklas med grov granoblastisk tekstur samt noe epidot og litt aktinolitt, ilmenitt og leukoksen.

Til slutt skal det nevnes at *grønnsten* av lavatypen tydelig gjennomsetter amfibolitten *gangaktig*, særlig godt sees dette på nesset V for Visnes. Et slip fra en slik gang viser plagioklas med grove tvillinglameller i intersertal anordning, med trevlete aktinolittbiter av blågrønn farge imellom. Det hele er strødd over med epidotkorn og -aggregater. Aksessorisk kommer leukoksen til.

e) Trondhjemitt.

Som allerede nevnt i det foranstående kapittel, opptrer det i grenseområdet mellom amfibolitt og lava-grønnsten en nesten hvit, sur bergart. Den danner uregelmessige ganger, slirer og klumper fra noen m til 100 m lengde med diskordante grenser og hvis strøkretning er subparallell med de øvrige bergarter (fig. 9). Fargen skyldes forvitringen, på friskt brudd er bergarten av middelsgrå farge.

Allerede med det blotte øye er det mulig å skjelne mellom to typer: en kvartsrik type og en i hvilken man ikke kan se kvarts. Undersøker man bergarten i tynnslip så iakttar man i det vesentlige et intersertalt flettverk av plagioklas (albitt 0 An) i hvilket det sitter kvarts-granoblaster. Ofte har man inntrykk av at de har fortrengt plagioklasen, men på en hel rekke steder foreligger det kvarts og plagioklas med grafisk implikasjonstekstur som altså tyder på samtidig utkrystallisering. Hvor man med det blotte øye ikke kan se kvarts, finner man under mikroskop også vesentlig mindre mengder av dette mineral.

Plagioklasene er alle av lik sammensetning. De ser friske ut, er ofte lamellert og inneholder vekslende mengder med inneslutninger. Disse består hovedsakelig av epidot, dessuten aktinolittnåler og noe kloritt. Også kvartsen er ofte uklar p.g.a. innesluttede epidotkorn. Aksessorisk opptrer følgende mineraler i trondhjemitten: epidot, aktinolitt, kloritt, ilmenitt med leukoksen, pyritt med limonitt, biotitt.

f) Malmsoner.

Hovedformålet med forfatterens arbeide var en undersøkelse av malmforekomstene. Det skal her gis et kort overblikk (sml. også Geis 1961 og 1965).

Hovedforekomstene er, resp. var, gamle Visnes-gruve og Rødklev-gruve. Begge ligger i den ovenfor nevnte Visnesskifer-horisonten. Gamle Visnesgruve like ved Visnes havn var i drift fra 1866—1894 og produserte i dette tidsrom ca. 1.364.500 t. råmalm med i gjennomsnitt 1,66 % Cu (Foslie nevner i "Norges svovelkisforekomsetr" et lavere tall som antakelig er en addisjon av tallene i Norges Bergverksdrift; der er imidlertid for tiden 1866—1883 den produserte mengde *salgsprodukt* oppgitt). Gruven ble drevet på seks steiltstående malmstokker og nådde et dyp på 732 m. All drivverdig malm er utdrevet. Ved undersøkelser i de senere år ble det heller ikke funnet ny malm. Beskrivelser foreligger fra Reusch (1884, 1888), Dittmarsch-Flocon (1875), Helland (1871), Knudsen (1885), Vogt (1894, 1910), Foslie (1926).

Rødklev-gruven som ligger ca. 500 m SO for den gamle gruven, har vært i drift siden 1904 med noen avbrytelser i årene 1920—1924. Her opptrer det malm på to sider av Visnesvann-forkastningen, "Vestkisen" og "Østkisen". Vestkisen kan kalles for en linjal med en rekke fortykkelser og forgreninger. Østkisens oppbygning er mer komplisert. En rekke forkastninger har skjøvet den sammen nesten som et trekkspill. Produksjonen i 1959 var 49.970 t råmalm med 23,48 % S og 0,45 % Cu. Gruven har nådd et dyp på 375 m. Vestkisen er beskrevet av Foslie (1926).

Mindre mineraliseringer opptrer på Visnesskifer-sonen øst og vest for gruven.

Utover den nevnte horisonten opptrer følgende mineraliserte soner som er uten økonomisk betydning:

Linderotsynker - Hopkins gruve - Huelva II - Hutte gruve - Kolstø (svovelkis og magnetkis),

Haugesundsynken - Patmos gruve - Kleven synk (svovelkis og magnetkis),

Landanes - Skeisvoll - Våge - Sørvelde - Gloppe - Bukkøy (svovelkis, magnetkis, magnetitt),

Vikingstad - Hauge (magnetitt, svovelkis),

Dale (magnetkis),

Klostein gruve (svovelkis, kobberkis),

Sørstokke gruve (svovelkis, kobberkis, har produsert tilsammen 7300 t råmalm med 0,5 % Cu),

Gåsvannet - Yttraland (svovelkis, kobberkis).

Forekomstene antas å være dannet submarint-ekshalativt.

IV. Kvartæravsetninger

Sen paleosoiske, mesosoiske og tertiære bergarter mangler i arbeidsområdet. Av yngre avsetninger opptrer bare istids- og muligens alluviale dannelser. Istidsavleiringene består av sand med rullesten av forskjellige størrelser, noen steder også leire som danner råmaterialet til et teglverk på Avaldsnes. Rullestenene er overveiende fremmede bergarter. Det dreier seg vel overveiende om bunnmorene. Dens mektighet øker fra vest til øst på Karmøy. Landskapet gjenspeiler dette: vestkysten er bergfull og meget tynt befolket, østkysten derimot flat, med godt jordsmonn og av den grunn tett befolket. Istiden har også gitt fjellknausene og forhøyningene sin nåværende form, det er runde og ovale koller.

Alluviale dannelser er vel stenurene ved noen bratte kanter, og torvmyrene som man gjerne finner i mangelfullt drenerte områder.

V. Tektonikk og metamorfose

a) Storstilet foldning.

På kartet (fig. 1) er det inntegnet en rekke innleiringer i lava-grønnstenen og i grønnskiferen, som gjenspeiler strøk og fall av lagdelingen. Denne vises meget godt av den allerede av Reusch (1888) konstaterte grense mellom lavagrønnsten og grønnskifer ("dioritisk bergart" og "grønne krystallinske skifere").

På størstedelen av Karmøy er strøkretningen NV-SO med steilt fall (ca. 60° NO) og en steil stupning av foldningsaksene i nordlig retning. Mot østkysten blir foldningsaksenes stupning suksessivt mye flatere. Samtidig bøyer bergartsgrensene seg fra NV-SO over O-V i SV-NO-retning, viser altså det typiske bilde av en synklinal. Synklinalstrukturen blir allerede nevnt av Kolderup (1931). Ombøyningen ble ved hjelp av tektoniske spesialundersøkelser funnet så langt sør som ved Sørstokke-gruve, hvorfra den forløper i NNV-retning mot nordspissen av Karmøy.

Bergartene, spesielt grønnskifrene, er ved østkysten stuet sammen til nesten isoklinale folder. Av den grunn får man her strøkretning ca. N-S. Svakere lokalfoldning forekommer også andre steder. Den er bl. a. antydet i Visnesskiferens forløp, dessuten i mineraliseringssonen Hopkins- - Patmos-gruve - Haugesundsynken og endelig i kvartsittenes forløp vest for den sistnevnte sone.

Den tektoniske oppbygning på fastlandet er likedan som på Øst-Karmøy, d.v.s. strøkretning N-S med en meget svak stupning av foldningsaksene mot N. Liknende strukturforhold som på Vest-Karmøy finnes på Feøy.

b) Småfoldning ("makroskopisches Gefüge")

Foruten foldene med km- og 100 m-utstrekning finnes også folder av mindre dimensjoner. Det ble iakttatt både steiltstående og flattliggende folder. Særlig lett er dette å se i grønnskiferen like NO for grensen til lava-grønnstenen, men også i Rødklev-gruven. Her er både kontakten mellom malmen og sidebergarten, malmens bånding, innleiringer i malmen og i tynne malmbånd småfoldet. Lengdeaksene av tynne aktinolittnåler i malmen og dens omgivelser forløper parallelt med de lokalt fremtrende hovedfoldningsaksene.

Allerede Reusch (1888) skriver at malmlinjalenes lengdeutstrekning (i gamle Visnes-gruve) faller sammen med bergartenes strekningsretning. Den østerrikske geolog Dr. F. Karl viste i en spesialundersøkelse (Karl, 1956) at lineasjonene (fine striper) på planstrukturene (lagdeling og skifrighet) i og omkring Rødklev-gruve går parallelt med hovedfoldningsaksene og at det utover dette består visse relasjoner mellom foldningsaksene og malmlinjalenes lengdeakser.

Dr. Karl viste at de nevnte småstrukturene omkring Visnes til ca. 1 km SO for Rødklev-gruve stuper steilt mot NO. Ved Jordan-gruve allerede blir deres stupning flatere og går i retning N til NNV. Disse målingene ble av forfatteren siden ført videre over hele det her omtalte område på Karmøy. Resultatene stemmer godt overens med de under "Storstilet foldning" nevnte observasjoner.

Omkring den gamle Sørstokke-gruve var det overhodet ikke mulig å se lagdelingen eller noe tilsvarende. Av den grunn ble det her utført en liknende undersøkelse som av Dr. Karl på Visnes. Diagrammene av planstrukturene tyder her på en liknende synklinalstruktur som på Nord-Karmøy. Anordningen av mer eller mindre lagergangaktige innleiringer av amfibolitt tyder på det samme. Lineasjonene stuper overveiende i nordlig retning.

Målingene av lineasjonene i amfibolitt-området viser en ganske stor likhet i deres forløp samt en likhet med lineasjonenes forløp omkring Visnes. I en rekke tilfelle ble det også iakttatt at de makroskopiske lengdeakser av hornblendekrystaller forløper parallelt med den lokalt herskende lineasjon. Av dette kan man slutte at lineasjonene i og utenfor amfibolitten ble preget samtidig. Paralleliteten av hornblendenes lengdeakser med lineasjonen tyder også på at dannelsen av lineasjonen og omkrystallisasjonen skjedde — iallfall delvis — samtidig.

Men det finnes ikke bare de typiske B-akse-lineasjoner parallelt med foldningsaksenes stupning, men også en B'-lineasjon*) som i alminnelighet står loddrett på den første, bare sjeldent står de skjevt til hverandre. B'-lineasjonen er vanligvis uten større betydning, bare i Rødklev-gruven synes den å gjøre forholdene enda litt mer komplisert. ac-sprekker står vanligvis ikke helt loddrett på B-lineasjonen resp. foldningsaksene.

Mange steder opptrer det små kvartsganger av noen cm tykkelse og noen dm lengde langs etter hovedstrøkretningen, som selv er foldet. Jeg antar at det er sprekker som er oppstått i de tidligste faser av foldningen og som i det videre forløp av foldningen selv ble foldet.

c) Bruddtektonikk.

De første antydninger til bruddtektonikk ble allerede nevnt i forbindelse med de foldete små kvartsganger. Den neste fase representeres vel av dm-tykke kvartsganger som imidlertid er svært sjeldne.

N-S-gående bruddsoner oppfattes som de yngste dannelser som — etter kartet å dømme — er av en viss betydning også utenfor Karmøy. Her skal bare nevnes: Karmsund, Grindefjord, Ålfjord. I arbeidsområdet er de representert ved den med sikkerhet påviste Visnesvann-forkastning, Karmsund-forkastningen og den antatte Fiskåvann-forkastning. Det er meget vanskelig å bestemme størrelsen av forskyvningene. For Visnesvann-forkastningens vedkommende var det mulig. Her er østpartiet skjøvet 75-100 m nordover.

Visnesvann-forkastningen er morfologisk meget utpreget. Det er en påfallende forsenkningssone med bratte fjellvegger på begge sider. Heri ligger Visnesvannet og 750 m lenger S enda et vann.

Ved *Fiskåvannet* begynner også en forsenkningssone med noen vann, som fortsetter 4 km i sørlig retning. Av likheten med Visnesvann-sonen kan det

*) Ifølge Sander (1948) opptrer det utenom den vanlige B-akse-lineasjon som går parallell med foldningsaksene, på de samme flatene ofte også en lineasjon mer eller mindre loddrett på B. Denne lineasjon kaller han B'.

antas at det her dreier seg om en liknende forkastningssone. P.g.a. overdekning med morene, vann og myr er det imidlertid ikke mulig å påvise dette med sikkerhet.

Sikkert påvist er *Karmsund-forkastningen* ved Bukkøy fyr. Her ser man en sprekk i 1 m bredde som er fylt igjen med limonittiserte bruddstykker av grønnskifer som er kittet sammen av en kaolinliknende substans. Den forløper 170°/70° V.

Mindre N-S-gående forskyvninger ble observert mellom Rødklev-gruve og gamle Visnes-gruve.

Bortsett fra disse N-S-gående forkastninger kunne det p.g.a. lederhorisontenes og de geologiske grensenes forløp sluttes på O-V-forkastninger. Det ser ut til at deres N-partier er sunket ned. De har dessuten bevirket horisontale forskyvninger på 50-200 m. Mer kan det ikke sies om dem. Det dreier seg antakeligvis ikke om tverrsprekker til de flate foldningsaksene ved vestkysten, da de også opptrer i områder med steile akser.

Til slutt skal jeg komme inn på *forkastninger langs etter strøket.* En rekke av dem ble observert i Rødklev-gruvens Østkis-parti. De faller vanligvis noe slakere (55°) enn lagdelingen (65°). Etter observasjonene å dømme er de eldre enn Visnesvann-forkastningen. I ethvert tilfelle er hengpartiet forskjøvet nedover i forhold til liggpartiet. Hvilken utbredelse disse forkastningene har i terrenget er det vanskelig å avgjøre. Spesielt på vestkysten finnes det et stort antall små daler og forsenkninger i terrenget som går parallelt med bergartenes strøkretning, men ingen steder finner man entydige beviser (f.e. glidespeil, breksje) for at det er forkastninger. Alderen av disse forkastninger er vanskelig å avgjøre. På den ene siden er det mulig at de er paradeformative (i forhold til foldningen) da de følger bergartenes strøkretning. På den andre siden er det like så godt tenkelig at de er av postdeformativ alder og delvis har benyttet seg av planstrukturene som dannet svakhetssoner.

En pekepinn på det innbyrdes aldersforhold mellom de tre forkastningssystemer får vi av iakttakelser i Rødklev-gruven. Små sprekker viste her denne aldersfølge:

1.) NV-SO 2) O-V 3.) N-S.

d) Metamorfose.

Som allerede antydet synes det å være sammenheng mellom tektonikk og metamorfose. Metamorfosen omhandles av den grunn sammen med tektonikken.

Bergartene i det område som ble kartlagt tilhører grønnskiferfasies og kvarts-albitt-epidot-biotitt-subfasies som definert av Turner & Verhoogen (1960). Selv om samtlige bergarter tilhører den samme metamorfe fasies kan man dog merke en økning av den metamorfe påvirkning fra NO til SV. Av de tidligere beskrivelser fremgår det at de følgende lokalfasies følger hverandre i den nevnte retning:

1.) Albitt/oligoklas-kvarts-kloritt-biotitt-epidot.
2.) Albitt/oligoklas-aktinolitt-epidot-(kvarts-biotitt-kloritt). Finkrystallinsk.
3.) Samme mineralsammensetning som 2., men grovkrystallinsk.

Som en variant av 1.) opptrer

1 a.) Albitt/oligoklas-kvarts-epidot-biotitt.

Ifølge Barth (1952) opptrer liknende bergarter i Sulitjelma-området, og der oppfattes disse mineral-kombinasjoner å tilhøre samme metamorfosegrad, med intrusivbergarter som utgangsmateriale for 1.) og sedimentbergarter for 1 a.). Det skal også bemerkes at grensene mellom metamorfosegradene av Vogt (1927) i Sulitjelma ble trukket litt annerledes. Etter ham hører våre lokalfasies 2.) og 3.) til

"epidot-amfibolitt-fasies",

hvilken av Turner & Verhoogen er opptatt i grønnskifer-fasies.

VI. Tydning av de petrografisk-geologiske observasjoner

Vi kan anta at grønnskiferen er en metamorf tuff, muligens med noen mindre mellomliggende strømmer av tynnflytende lava, men dette kan vanskelig avgjøres i dag.

Vi har sett at lava-grønnstenen inneholder delvis ganske gode reliktstrukturer som tyder på en effusiv dannelse. Innenfor lava-grønnstenen iakttok vi små stokkformete partier av en jevn krystallinsk bergart som ble kalt for "epidioritt" og som har skarpe begrensningslinjer. Ved disse to egenskaper — jevnheten i krystalliniteten og skarpe grenser — skiller den seg tydelig ut fra amfibolitten. Amfibolittens karakteristika er som tidligere beskrevet den vanligvis uskarpe grense mot lava-grønnsten og den slireaktige blanding av tette og forskjellige mer grovkrystallinske partier.

I avsnittet "Metamorfose" så vi at metamorfosegraden øker fra grønnskifer over lava-grønnsten til amfibolitt, som danner tre trinn i arbeidsområdets metamorfose. Gneisområdet på SV-Karmøy danner et fjerde trinn.

Amfibolitten og gneisene på SV-Karmøy støter — som Reusch's kartlegging og egne resultater viser — med en rett grense mot hverandre. Denne grense forløper samtidig subparallell med bergartenes alminnelige strøkretning slik den kommer til uttrykk f. eks. i grensen mellom lava-grønnsten og grønnskifer. Vest for Buasdalsvann (V Sørstokke) ble det foretatt noen målinger som viser at gneisen faller under amfibolitten. Lineasjonsmålingene viste meget god overenstemmelse med målingene ved Sørstokke gruve. Det ser altså ut til at bergartsgrensene er primære og at det dreier seg om en i det vesentlige konkordant lagfølge hvis alder og metamorfosegrad avtar fra SV til NO.

Bortsett fra amfibolitten er det ikke vanskelig å gi en tydning av bergartene. Dette er i en viss utstrekning allerede skjedd i kapitlet "Bergartsbeskrivelse". Amfibolitten kan imidlertid være oppstått på 2 forskjellige måter. For det første kan det opprinnelig ha vært en basisk intrusivstokk. For det andre kan det være en omkrystallisert effusivbergart. Av Goldschmidt's benevnelse "Saussurit-gabbro" synes det å fremgå at han oppfattet den som en intrusivbergart. Tilsvarende bergarter på Stord ble av Kvale (1937) kalt for "gabbro". Oftedahl (1961) påpeker at man i de fleste kaledonske gabbromassiver i Norge finner liknende forhold som de som her ble beskrevet fra Nord-Karmøy. Han antar at det ble ekstrudert store lavamasser hvis dypeste partier hadde lengre tid å krystallisere og ble grovkrystallinske.

Etter min mening finnes det en hel rekke argumenter som taler imot en intrusiv opprinnelse og imot Oftedahls forklaring. Vi så for det første forskjellen mellom de små epidioritt-stokkene og amfibolitten. Man savner den skarpe grense og en jevn krystallinitet som man ellers er vant til i det indre av et slikt kompleks. At ligg-grensen går parallelt med bergartenes lagdeling er for en intrusivbergart også høyst ualminnelig. I de sentrale deler finner man dessuten bånding og sliring som har en viss likhet med granittiseringen, spesielt slik den opptrer øst for Haugesund, bare med den forskjell at årene og båndene i arbeidsområdet ikke er så skarpt avgrenset fra resten av bergarten. Men det kan ha sin grunn i at granitt-migmatittene tidligere har vært sedimentbergarter med godt utpreget lagdeling, noe som ikke er tilfelle med pillow-lava. Ofte er man fristet til å snakke om en slags injeksjonsmetamorfose. Under mikroskop ser man heller ikke reliktstrukturer etter basiske dypbergarter, men helt overveiende friske, fylte og jevnt oppbyggete plagioklas- og aktinolittkrystaller som er vokset sammen på en slik måte at man må anta en samtidig dannelse. Eskola (1939) nevner som de viktigste kjennetegn på den metamorf-krystalloblastiske tekstur bl. a.:

samtlige mineraler er dannet omtrent samtidig,
utvikling av krystallflater er sjelden,
parallelltekstur ved at komponentene fortrinnsvis vokser i én retning,
sonarbygning av komponentene mangler,
inneslutningene følger vanligvis ikke sonarbygningen,
bergartene er kompakte, ikke amygdaloide e.l.

Nettopp disse kjennetegn finner vi igjen i amfibolitten. Jeg antar av den grunn at amfibolittene er sterkt metamorfoserte lavabergarter av samme eller liknende type som utgangsproduktet for lava-grønnstenen. En støtte for denne oppfatning finner man hos Harker (1950), som skriver om metamorfose av basiske lava-bergarter: "In most of the examples which have been cited, even in the lowest grades of metamorphism, the original fabric of the igneous rock (gabbro, dolerite or basalt) had been completely broken down".

Hvordan skal vi nå forestille oss gangen i denne metamorfosering-omkrystallisasjon? En pekepinn gir observasjonene som før er meddelt fra Kopervik: Her fantes innenfor lava-grønnsten-området en breksje av lava-grønnsten med trondhjemitt som kittsubstans. I grenseområdet mot amfibolitten finner man på samme sted liknende bilder, bare med den forskjellen at "bruddstykkene" her består av aktinolitt og plagioklas i grov subofittisk anordning. I andre prøver fra samme sted ser man at disse strukturene er mer utvisket og etter hvert går over i rent granittiske teksturer.

Det ser ut at vi her har nøkkelen til forklaringen på de trondhjemittiske gangene ved amfibolittens henggrense: jeg tyder dem som en slags "migmatitt-front". Ifølge Harker blir det frigjort kvarts ved omvandlingen av augitt til hornblende, og Eskola viser at det blir frigjort SiO_2 og Al_2O_3 ved omvandling av anortitt til epidot. Forklaringen på opprinnelsen til den høye Na-gehalten i de nå foreliggende plagioklaser byr på visse vanskeligheter. Men dette gjelder i like stor utstrekning for de andre bergarter i arbeidsområdet. Man står overfor de samme problemer når man skal forklare diabasenes spilittisering. Også i dette tilfelle må man anta Na-tilførsel på en eller annen måte (Eskola 1939, jfr. også Geis 1962 a og b).

Utkrystallisasjonen av de grove amfibolittpartiene skjedde sikkert ikke ut av en smelte, men vi kan med Eskola anta en mer eller mindre metasomatisk, intergranular stoffvandring. Den ofte iakttatte anordning av disse partiene langs planstrukturene tyder på at oppløsningene eventuelt har brukt veier som de tektoniske hendelser har forberedt for dem.

Til slutt kan vi nå sammenfatte resultatene og forsøke å tegne et bilde av hendelsesforløpet.

I begynnelsen ble det avsatt mer eller mindre *klastiske avleiringer*. Derpå begynte temmelig abrupt den basiske *initial-vulkanisme*. Den begynte med og besto i sin første fase av submarine effusjoner med etter hver effusjon følgende tuffavleiring. I sin annen fase besto den vel nesten utelukkende av tufferupsjoner — grønnskiferens plattig-lagige karakter tyder iallfall på dette. I forbindelse med denne vulkanisme ble det også sedimentert annet materiale, f. eks. blåkvarts som muligens kan tydes som ekshalasjoner av kiselsyre. Likedan tydes forekomstene av svovelkis, magnetkis og magnetitt. Tydningen av de keratofyriske "ganger" er ikke så lette p.g.a. de metamorfe forandringer.

Det kan f. eks. være episodiske sure effusjoner eller — hvis man tyder blåkvartsen som et sediment — metamorfe sedimentbergarter, muligens også "welded tuffs". Endelig er det ikke umulig at det er metamorfe mobilisasjonsprodukter.

Som det tredje ledd opptrer igjen sedimentbergarter, denne gang rene leirsedimenter som man finner på nordspissen av Karmøy.

Lagfølgen viser en viss likhet med den i Trondheimsfeltet omkring Berkåk. Her ble denne lagfølge observert:

fyllitter
tynnplattig-skifrig grønnsten
lava-grønnsten

En stund etter at de nevnte bergartene på Karmøy var kommet på plass fulgte vel først metamorfosen; det ble konstatert at metamorfoseringen i det store og hele opptrer etasjeformet, det vil si at metamorfosegraden avtar parallelt med bergartenes alder. Grensen mellom amfibolitt og lava-grønnsten viser dette ganske godt. Men i amfibolitten er hornblende-krystallene lokalt orientert parallelt med hovedfoldningsretningen. Dette tyder på en samtidig metamorf og tektonisk virksomhet. Antakelig begynte den tektoniske deformasjon i en senere fase av metamorfoseringen.

Vi forsøker så å ordne vårt område inn i det geologiske helhetsbilde: Vi må anta at det fantes en del-geosynklinal i hvilken de grønne bergartene og fyllitten ble avsatt. Del-geosynklinalen ser ut til å strekke seg fra området nord for Stavanger over Karmøy - Bømlo - Stord - Tysnes - Ølve - Varaldsøy inn i Hardangerfjorden. Mot dypet er bergartene sterkere metamorfosert og danner således en naturlig overgang til de enda sterkere metamorfoserte partiene øst og vest for del-geosynklinalen. Ifølge Sørbye (1946, 1954) er også dette kaledonske bergarter. Dette bilde virker for meg mer naturlig enn å anta en tynn, plateformet intrusjon av "saussurittgabbro" som strekker seg fra Karmøy og helt til Hardangerfjorden.

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On kink-zone development and metamorphic differentiation in the low-grade schists of Norwegian Sulitjelma

by

Robin Nicholson[1])

Abstract

The common minor scale kink-zones of upper levels of the Furulund schists of eastern Norwegian Sulitjelma generally do not occur in conjugate pairs; usually only the member with northwest dipping axial plane is developed although at lower levels rarely both conjugate pairs and single members occur. The kink zones fit the simple shear mechanism of development proposed by Dewey and Roberts. The upper kink-zones are characterised by a well-developed metamorphic differentiation structure resulting from the juxtaposition of zones alternately rich in mica and quartz. In many of the kink-zones the mica-rich zones are bent by smaller scale folds unknown elsewhere in the region but which are geometrically conjugate to the minor-scale structure in which they lie. Although they clearly are later than the kink-zones it is judged that they are directly related to them.

Introduction

The structures with which this account is concerned are developed in the structurally higher levels of the low-grade quartz-mica schists of the eastern edge of the central metamorphic core of the Scandinavian Caledonides at Norwegian Sulitjelma. The rocks of the area of kink-zone and metamorphic differentiation development (generally east of Lomivann, Fig. 1) are of two main lithologies, black quartz-muscovite phyllites (some homogeneous and some with well-developed thin semi-pelitic layers) and volcanic-chloritic rocks with much marble (Vogt, 1927; Nicholson, 1966). Differentiation is most marked in folds of distinct kink-zone style, the differentiation layering along the kink-zone sometimes being so well-developed that at first sight it resembles bedding.

Development of kink-zones and differentiation structure

1. Most of the kink-zones described here have northwest dipping axial planes (Fig. 7) and axes plunging east of north (regional dips of bedding and

1) Department of Geology, University of Manchester, England.

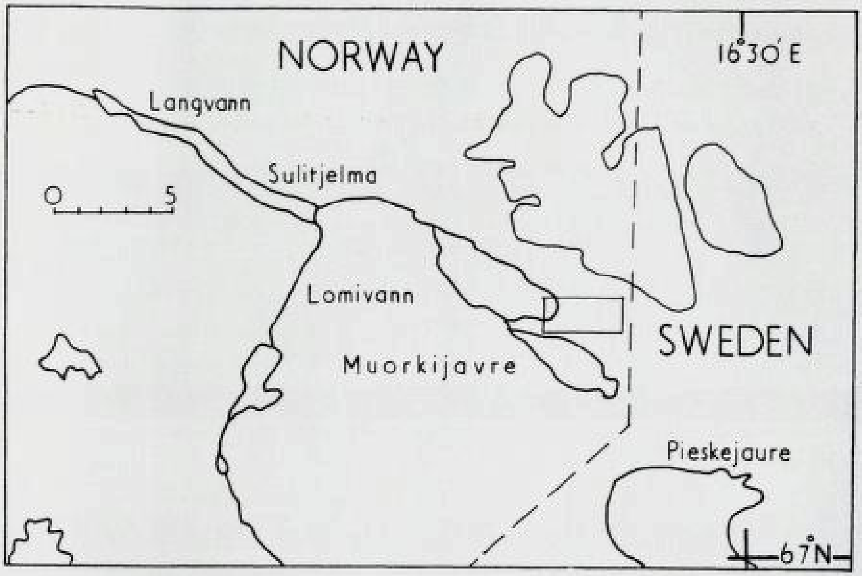

Fig. 1. Outline map of the Sulitjelma region; inset rectangle east of Lomivann limits area in which observations were made. Scale in kms.

schistosity are northerly). At other levels than that at the east end of Lomivann the other member of the conjugate set sometimes is developed on its own and occasionally both members are present. Here we are concerned with a level just below the amphibolites of Sulitjelma (Vogt, 1927, Kautsky, 1953) some 10 km long in which only the northwest dipping minor scale member is present. From west to east at this level there is a gradual decrease in metamorphic grade (Vogt, 1927); the most angular kink-zones are developed in the low-grade rocks to the east while to the west end of Lomivann the apparent equivalents are of somewhat different style but basically the same geometry but without angular hinges. Such folds are associated with one another in what in the field were described as stacked folds in which half a dozen or so folds of the same size and sense of overturn occur together. No conjugate arrangements at all are known in the west.

The sharply angular style of the eastern kink-zones developed in the finely schistose rocks of the east Lomivann area is replaced by a rounder style where more thickly constructed layers are reached and a kink-zone may die out against a thick sequence of such rocks while being able to grow through a bed a few centrimetres thick.

As is common with kink-zones (Paterson and Weiss, 1966, 352) the margin

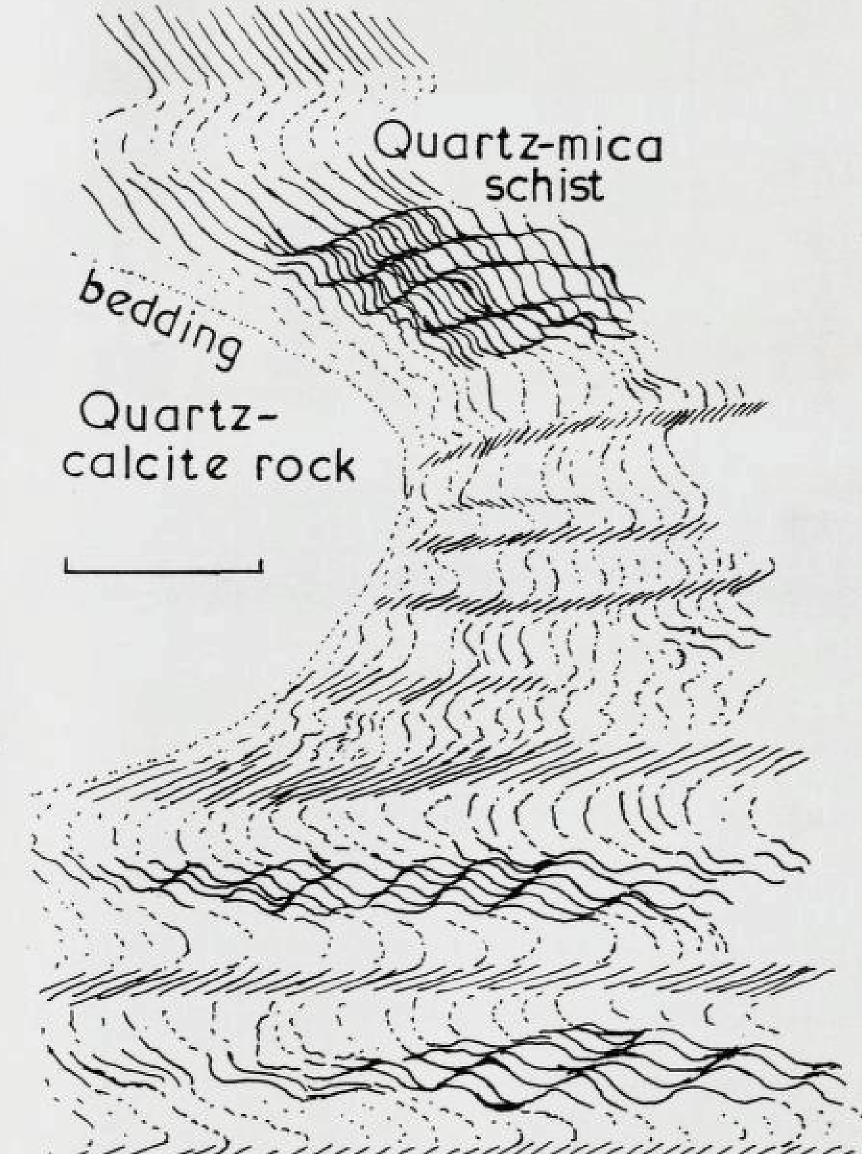

Fig. 2. Sketch of thin section of east Lomivann black phyllite with marked metamorphic differentiation; later folding of suitably oriented micaceous laminae. A part of the area of this sketch is shown in fig. 1. Scale mark 1.0 mm.

Fig. 3. Black phyllite of fig. 2 showing detail of differentiation structure and its folding. Scale mark 0.5 mm.

about bisects the angle between bedding and schistosity inside and outside the kink-zone (bedding and schistosity are about parallel in this region and equivalent to the primary composite banding of Roberts, 1966, 833). The kink-zones often lie at about 30° to 40° to this primary banding and generally are no more than 10 cm. thick. On favourable outcrops the kink-zones can be seen to be little more than 2 to 3 m square, throughout being characterised by continuity of the inherited structure through the kink-zone.

2. As described below the kink-zones often have a finely layered appearance that on close examination is seen to result from the presence of mica-rich and mica-poor layers developed across the primary banding and parallel to the kink-zone margin. Figs. 2, 3, 4 and 5 show examples of the layered structure present in the kink-zones of Norwegian Sulitjelma. The sequence of repeating elements of the structure, quartz-rich and mica-rich layers, may be 20-30 cm thick, the thickness of the kink-zone itself. The layered structure is found only in the finely foliated (or schistose) semi-pelitic rocks, stopping as massively chloritic or psammitic and more broadly bent layers are reached; behaviour which is perfectly understandable when it seen that the two elements of the

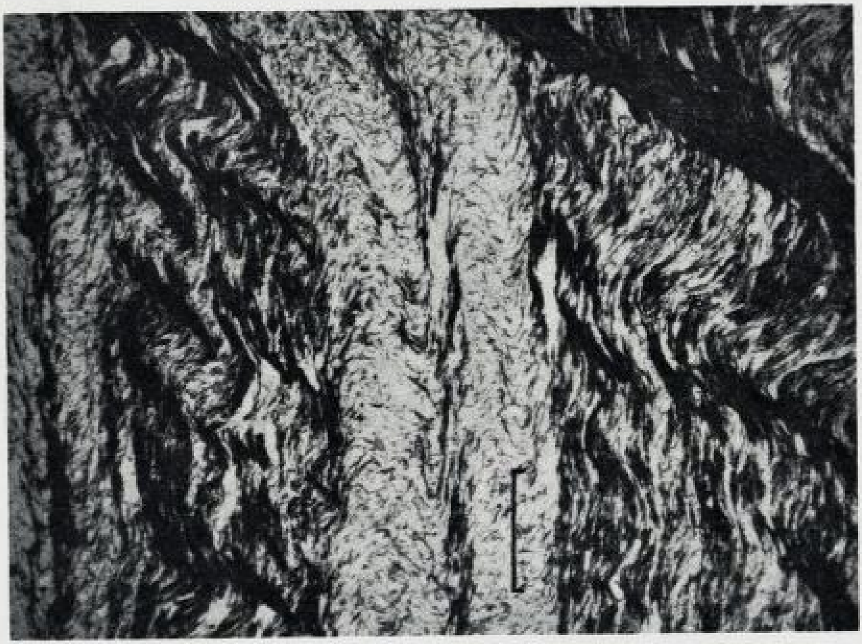

Fig. 4. Quartz-mica differentiation in east Lomivann black phyllite. Folding later than its development; here affecting both quartz-rich as well as mica-rich layers. Scale mark 0.5 mm.

secondary structure are constructed from different parts of the small folds within the kink-zone. Clearly only material which can fold on this small scale can give rise to it. As Figs. 2 to 6 show the Sulitjelma kink-zones are yet more complex for the differentiation structure itself is folded.

Fig. 2 of a section normal to the kink-band axis (and to the axes of the larger of the microscopic-scale folds seen in this figure and the photomicrograph of Fig. 3) shows the schistosity of the black phyllites passing through the differentiation layers. Where mica is preponderant, schistosity lies at an acute angle to the layer while in quartz-rich zones schistosity lies more or less normally across the band. This layered differentiation structure was described first by Clough (1897) and its units were called strain bands by him. The short limbs of the Sulitjelma kink-zones are composite therefore, being made of several almost microscopic sets of folds or strain bands through which continuity of fabric is preserved as it is on the larger scale of the kink-zone. Since the layered structure lies along a kink-zone it follows that the limbs of the small folds making up the structure are strictly parallel to neither the long

Fig. 5. Metamorphic differentiation structure in kink-band of black phyllite (trending gently upwards from left to right) and folds of this structure (trending more steeply upwards from right to left).

limb of the kink-zone nor to the average attitude of the short limb, both limbs have been deformed. Also it is clear that the kink-zone margin is not the bisector of the angle between primary banding attitudes inside and outside of the small folds (Fig. 2). In some ways these small folds of about equal limb length are comparable in style with chevron folds, in this case however, unequal thicknesses of the primary banding in different limbs allows the axial plane to be nearer one set of limbs than the other.

3. Figs. 2, 3 and 4, 5 and 6 show not only the differentiation structure typical of Sulitjelma kink-zones but also folds of this structure while Fig. 3 shows the pattern of Fig. 2 over a wider zone of the same rock. The folds of micaceous elements clearly are later than the differentiation structure. They have the same sense of rotation over the whole of the kink-zone (being on that scale of conjugate motion with it), while on the small scale the relationship is of alternately conjugate and non-conjugate motion. The two fold systems are not contemporary of course; all member of one set are earlier than all of the other. Functionally the second set are kink-zones also although they are not tabular but occur only within thin mica-rich zones. Examples of the geometric relations between the two sets are given in Fig. 7 parts 2, 3 and 4. The angle between the minor and microscopic scale developments varies

Fig. 6. Kink-type fold in Lomivann black phyllites with set of crenulations whose axial traces lie along the trace of the minor-scale fold (along direction of pen) and a subsidiary and crossing set (along direction of pencil), restricted to the axial zone of the fold.

considerably from one example to another although most commonly the angle is about 50°, as in Fig. 7, parts 3 and 4, so that the two "kink-zones" are at an acute angle to one another over the presumed maximum principal stress.

Fig. 2 provides an interesting example of the way in which differentiation might upset the orderly development of kink-fold sets for the new planar structure is less easily kinked than the rock from which it formed. Secondly it illustrates the way in which the quartz-poor regions even when strongly folded cannot develop the type of modal variation possible in the initial quartz-mica schist. The pseudo-conjugate structure is developed only in the kink-bands themselves and on the scale described above rather as if it could form only on the transposed schistosity of the kink-bands. In spite of the obvious age difference the writer regards the two fold sets as being almost as closely related as truely conjugate structures.

Genesis of kink-zones and differentiation structure

1. Historically there have been two sources of interest in kink-like structures, firstly the natural occurrences of such structures in rock on a minor fold scale and secondly the development of geometrically similar arrangements in deformed crystals; both enquiries date back into the last century. The most

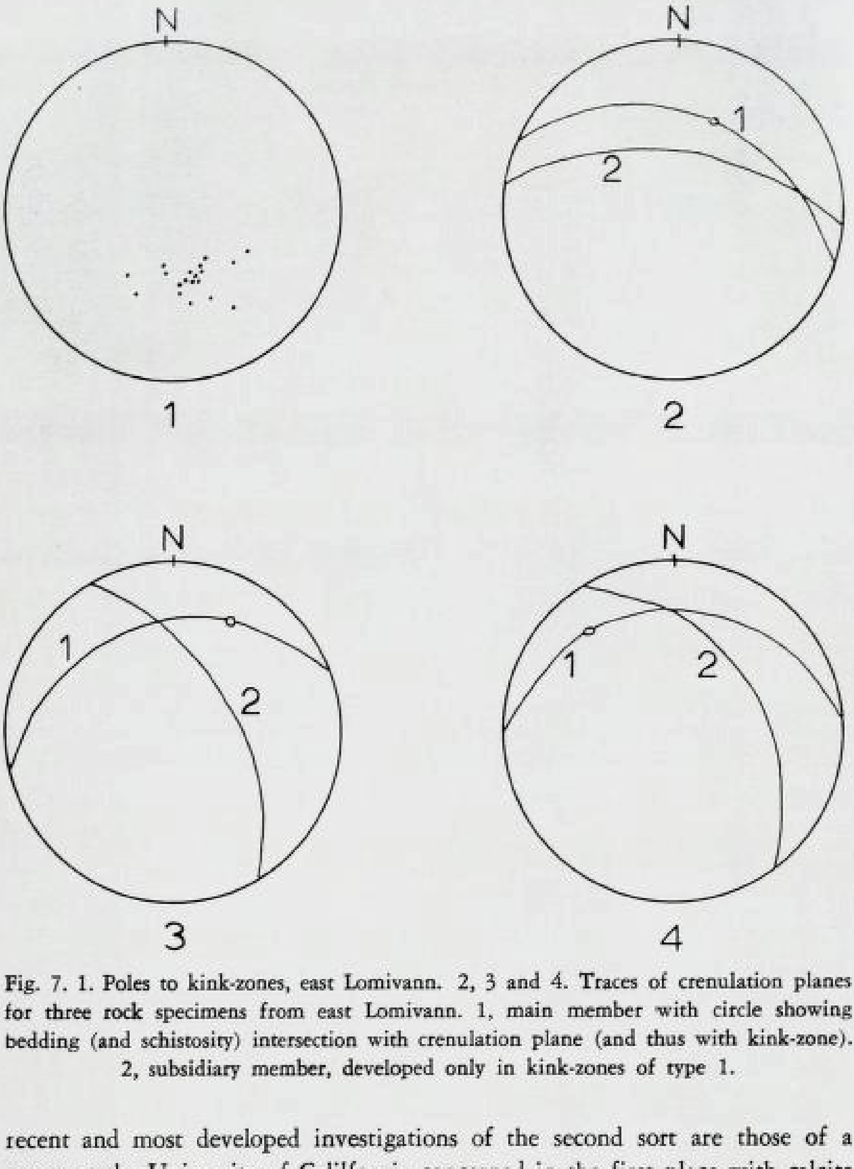

Fig. 7. 1. Poles to kink-zones, east Lomivann. 2, 3 and 4. Traces of crenulation planes for three rock specimens from east Lomivann. 1, main member with circle showing bedding (and schistosity) intersection with crenulation plane (and thus with kink-zone). 2, subsidiary member, developed only in kink-zones of type 1.

recent and most developed investigations of the second sort are those of a group at the University of Calilfornia concerned in the first place with calcite (for example, Turner, Griggs and Heard, 1954); from this study have grown analytic and experimental studies of geometrically similar rock structures (Patterson and Weiss 1962; Heard, Turner and Weiss 1965; Patterson and Weiss 1966). Amongst the latest studies based on the field examination of kink-zones and in both cases using analytical tools developed by Flinn (1962), are accounts by Dewey (1965) and Roberts (1966). As Roberts has very recently pointed out there are substantial differences in the conclusions reached in these two ways for both Dewey and Roberts accept an important class of kink structures produced by simple shear while the Californian school propose that the same class of structures are the products of bend-gliding.

The writer finds the geometric arguments of Dewey and Roberts compelling and thus accepts that some kink-zones are produced by simple shear and that those produced by bend-glide (flexural slip is an alternative name for this mechanism) should have dislocations on the margins; thus continuity of structure through the kink-zone is one piece of evidence in favour of the simple shear mechanism. If this mechanism is accepted then such kink-zones in rock are more comparable to the slip domains of calcite described by Heard, Turner and Weiss (1965), rather than the kink-zones of calcite. Basically the difference of opinion concerns the character of the strain in the marginal planes of kink-zones; all recognise the need for strain there that is compatible with the strain regimes on both sides. Dewey and Roberts point out, however, that it is not enough that such compatibility be present at the end of development of the structure but that it must exist throughout so to preserve the continuity of the earlier developed arrangements to typical of kink-zones. It is on this need that they decide the inadequacy of bend-glide of flexural slip mechanism for structures like those described here from Sulitjelma. At Sulitjelma as for the Dalradian rocks described by Roberts there is an absence of the structure necessary to define the strain in detail. Again like the Dalradian rocks, however, there is an absence also of the contemporary structures which might suggest triaxial strain which therefore is accepted as substantially biaxial.

The simple shear mechanism of kink-zone development does not seem to provide any sufficient reason why the kink-zone margin should commonly bisect the angle between primary banding inside and outside the kink-zone. If this relationship is regarded as common and revealing then this is a deficiency in the theory. Dewey, however, has described folds (for instance 1965, Fig. 17 B) in which the kink-zone margin is far from bisecting the banding angle. Thus it is not absolutely clear that near equality of angle is especially significant although it is reported by Patterson and Weiss as typical of their experimental results (1966, Fig. 6) and certainly occurs in the Lomivann examples. Roberts (1966, 847) clearly supposes that the primary banding of the short limb will not be rotated when it has reached a position normal to the maximum principal stress; on his simple model of kink-zone folds this is achieved when the two angles from the kink-zone margin to the banding on either side are equal. Such an end to rotation seems obvious in the homogeneous strain investigated by Flinn (1962) but not so obvious in zones of simple shear in which the banding supposedly is mechanically passive.

As Roberts suggests (1966, 851) the appearance of only one member of the potential conjugate set might result from a stress field oblique to the composite banding. Kehle (1964, p. 284) has proposed for fractures a slightly different context in which there might be unequal development of the members of a conjugate set, viz. that the local stress field might be symmetric while the far stress field in general might not be so and thus that the work necessary to cause virtual displacement on one member of a conjugate set would be less than that necessary on the other. Both arguments might apply here as the kink structures seen to have formed late in a thrusting episode and the far field stress field then presumably was not symmetrical.

2. It has been suggested that the differentiation structure developed by the diffusion of quartz from highly strained zones followed by precipitation in adjacent less strained ones so that modal variation was produced in a once homogeneous rock (Nicholson, 1966, Roberts, 1966, p. 850). The structures which define the quartz and mica-rich zones clearly are part of the minor-scale kinks and coeval with them. The inner complexity of the kink-zones seems to have been early in development for if the kink-zone long limbs are undeformed during much of kink development and a simple shear mechanism operative (both suggested by Dewey (1965, p. 470-471) and Roberts (1966, p. 841)) then simple shear must be responsible for it too and there is no reason to separate its time of development from that of he kink-zone itself. Since quartz diffusion presumably is a relatively slow process it seems that the kink-zones themselves did not achieve their finished form abruptly. In addition it is easier to suppose that the kink-zone started with its present width rather than grew wider as Paterson and Weiss (1966, p. 367) suggest; in this way all the component zones have the same time for development (clearly they have the same structural character).

There is no evidence at Lomivann for migration of quartz beyond the major kink-zones; quartz veins are not common and those that do occur show no special distribution with respect to them. Since the differentiation structure is always on the same scale and does not occur much magnified in bigger folds it is clear that conditions for its development are suitable only in the small folds, perhaps because both the contrast and degree of strain necessary to produce such quartz migration in schistose quartz - mica fabric only takes place on this scale.

Clough (1897, 22) described the development of differentiation as an example of "... the law that in the thinned limbs quartzose parts of the bands diminish in quantity, while the micaceous remain the same or increase," and Dewey (1956, 475) has already suggested quartz mobility in kink-zones although he did not relate it closely to the development of strain-bands. Although Clough (1897, 20) like Dewey (1965, 486) and Rast (1965, 89), does seem to suppose that some of the mica of the Cowal kink-zones is new (thus supposing mobility of its components) it is clear that the attitude of some of it fits the mica of the pre-strain-band schistosity as in the example of Fig. 2 for Clough writes (1897, 20) that "mica in the slips [strain slips or strain-bands] is not always quite parallel to the sides of the slips, but has a small angle, approximately that at which the earlier banding [and usually schistosity (Clough 1897, 9)] approaches the slips." This description fits the common case described above where there is a continuous mica fabric through moth the mica and quartz-rich zones. Later development as proposed by Roberts (1966, 850) may include the growth of a mica-felt along the kink-zone margin as Dewey (1965, 486) and Rast (1965, 90) have earlier suggested. This stage is not reached at Lomivann.

The usual starting point of discussions of metamorphic differentiation is Eskola (1932) and the development of strain-bands can be conducted in its terms. The formation of alternating mica- and quartz-rich layers in the folded fabric provides examples of Eskola's second and third principles of differentiation within a rock mass (1932, 70), namely the principles of enrichment in the most stable constituent and the solution principle. The "exogenous force" (Eskola 1932, 70) that drives the process is deforming stress. Thus the differentiation phenomenon is a synkinematic one.

It has long been supposed that quartz readily migrates and Eskola (1932, 77) made special reference to such migration. Voll (1960, 556, 557) has described its importance both in general in metamorphic rocks and in particular in the development of mica films on secondary schistosity planes in situations very like that under discussion here. Voll, however, attributes the generation of the fine plications of the schistosity directly to shear planes, the mica films forming on such planes by the solution of quartz and relative enrichment in micas. In the Sulitjelma rocks, however, no clear-cut shears are present for schistosity is continuous through the kink-bands and their internal folds although there is markedly varied strain from layer to layer of the structure.

Conclusions

It is difficult to fit the Lomivann kink-bands into Dewey's classification of kink structures although those parts of kink-bands which pass through less closely foliated rock are close in geometry to this fourth group, that of smoothly curving similar folds. However, most of the kink-bands are present in black phyllites and are slab-sided with foliation continuous through them; thus they are superficially similar to Dewey's third group (1965, 485) although lacking the felt of mica along the kink-boundaries that he described as characteristic of this group. It is the writer's view that Dewey is wrong in putting Clough's strain bands in this third group where indeed he makes them the type examples. Although their geometry is close to that of the fourth group of Dewey the writer regards Clough's examples as of the same origin as those of Lomivann, that is, essentially simple shear kink-folds.

Clough's structures perhaps have more similarity with ordinary crenulation cleavage than with kink-bands and their presence here widens the discussion to include the development of differentiation in crenulation or strain-slip cleavage, a matter given general consideration by Roberts (1966, 848). The writer has already suggested a close similarity (Nicholson, 1966) for as Turner and Weiss (1964, 465 and 487) have said crenulation cleavage is characterised by "periodically spaced thin domains of intense strain" while "the intervening laminae (the microlithons of De Sitter) show evidence of pronounced internal strain in the form of crenulated relic S foliation". It seems that in general these crenulation folds are simple shear structures as Roberts has described (1966, 848) sometimes modified by SiO_2-diffusion and metamorphic differentiation.

The metamorphic differentiation of the Sulitjelma rocks is a synkinematic phenomenon involving the solution of quartz in, and its diffusion from, zones of high strain and its precipitation in adjacent less strained zones, mica remaining passive, the folds controlling its regular development being crenulations of an earlier schistosity. Such crenulations and the differentiation structure may occur within clear kink-band structures or within minor scale folds of other style. In all cases there is a clear relationship between fold size and style and the nature of the lithological sequence. When developed in bigger folds the structure parallels the axial surface of the larger fold.

Since kink-bands are best developed in thinly foliated rocks the bands and the related differentiation structure are characteristic products of late deformation phases in schistose rocks. Except for mica and quartz the mineralogy of the earlier rock is not relevant. For example at Ntungamo in Uganda (Nicholson, 1965) good differentiation occurs in staurolite schists (Fig. 1, Nicholson, 1966) while at Sulitjelma and Cowal it occurs in much lower grade rocks. It is clear that the development of differentiation will inhibit or strongly modify kink-zone evolution so that it may be only in mica-rich rocks that the climax of the conjugate relation, the development of folds affecting all the rock will be reached (Paterson and Weiss, 1966, 367, Fig. 17, Roberts, 1966).

McNamara (1965), has proposed that the development of quartz segregations (not all of kink-band type) in the lower-grade schists of Cowal has a broad P-T control as segregations do not occur in the lowest grade of the region. He also proposes (1965, 374) that the marked differences in amount of quartz segregation between occurrences in massive greywacke and mica schists is attributable "to the increased amount of solution of quartz in the more sheared rock, because abnormally high surface energies on portions of the detrital quartz grains, due to stress, cause the silica to dissolve more readily, leading to supersaturation of the pore fluid; on relaxation of the stress, the quartz is redeposited". To fit the petrographic evidence of strain bands it is necessary to modify McNamara's hypothesis to allow synkinematic precipitation in a less strained zone.

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Note on a molybdenite-dolomite-bearing pegmatite in Velfjord, Nordland, Norway

By

Christoffer Oftedahl

Geologisk Institutt, N.T.H., Trondheim.

Abstract

A small area of basic gabbro occurs within the granodiorite pluton north of Velfjord. In this gabbro was found a peculiar pegmatite dike, rich in molybdenite, and composed of quartz, albite, and dolomite as major minerals.

Velfjord is situated in Søndre Helgeland, the southernmost part of Nordland fylke. North of Velfjord the geologic map of Rekstad (1902, p. 8) shows a large massif of intrusive granite. Not shown on the map is a gabbro massif which may be several kilometers in diameter, situated east of the little fjord Andalsvågen. The massif contains a great variety of gabbroic rocks. Fine-grained to medium-grained gabbros predominate, but very coarse-grained varieties are met with, and a number of peridotite areas are found. About 1.0 km northwest of the top of the mountain Andalshatten (900 m high) a little pegmatite dike occurs in very coarse-grained peridotite.

Petrographically the granitic rocks are for the most part medium to coarse-grained granodiorite, or close to quartz diorite, with around 10 % microline. In places they are porphyritic, with plagioclase crystals of size up to 4 cm.

Some results from a detailed study of the rather homogenous medium-grained gabbro around Andalsvågen may be mentioned. Bright green augite is the most important mineral (25 - 50 %), followed by plagioclase (10 - 30 %), olivine (10 - 20 %), orthorhombic pyroxene (5 - 15 %), light brownish hornblende (5 - 15 %), and similarly light brownish biotite (5 - 15 %). Ore minerals make up less than one per cent. The geologic setting and mineral composition suggest a basic gabbro, crystallized or recrystallized under conditions of strong regional metamorphism.

The pegmatite dike was found by Mr. Olav J. Andal and was subsequently mined for molybdenite in 1936. The outlines of the dike could not be seen during my visit on Aug. 16, 1958, due to snow filling of the mined-out area, but Mr. Andal tells that it was 35 cm wide and 8 m long. The dumps contain

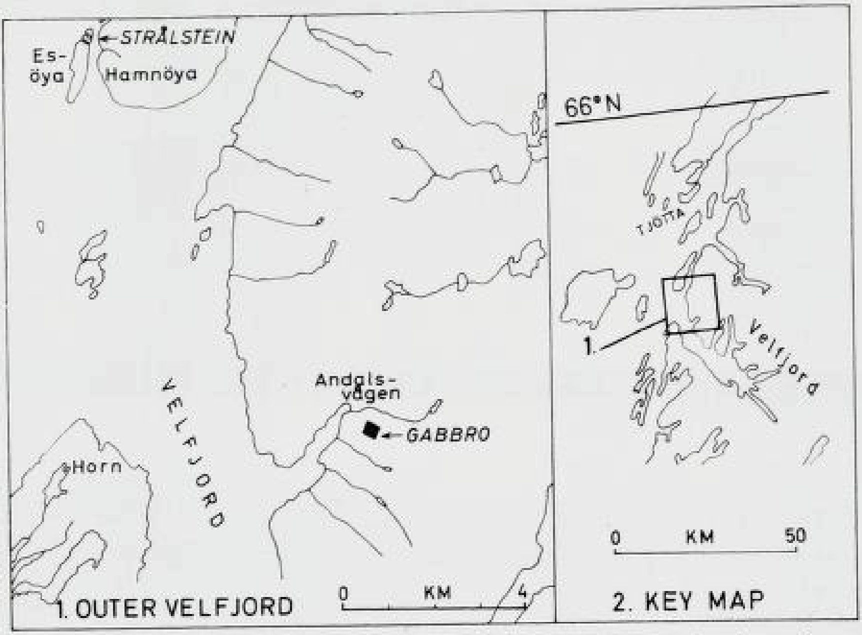

Fig. 1. 1. Sketch map of outer Velfjord with the gabbro area at Andalsvågen schematically shown. Also shown is the outcrop on the small island Esøya of an extraordinary rock type (strålstein). It is composed of fibrous actinolite aggregates of a very bright and fresh green colour, set in a matrix of white calcite.

2. Key map showing the location of Velfjord on the Nordland coast, at 65°30′ N.

big pieces of the pegmatite with the following minerals: molybdenite, quartz, ankeritic dolomite, albite, biotite, muscovite, talc, tourmaline, and traces of chlorite.

The molybdenite occurs in the dike as scattered platy crystals and blades with a diameter up to 3 cm, with a reported maximum size of 8 cm. Most of the molybdenite occurs as blades larger than 1 cm, accordingly much molybdenite could simply be handpicked by blasting the pegmatite.

The main minerals of the pegmatite must have a zonal arrangement, because all inspected fragments are rich in either quartz or in albite and dolomite. The quartz-rich type carries muscovite, tourmaline, and molybdenite as accessory minerals, and the albite plus dolomite type contains talc, biotite, a few clear quartz grains, molybdenite, and a little chlorite. In the latter type the feldspar and dolomite may have all proportions in the various pieces available for inspection. Both quartz, dolomite, and albite may make up large continuous mineral aggregates with single crystals of size up to 2 - 3 cm.

The mica blades are less than 0.5 cm in diameter, and the talc occurs as fine plates, 0.1 cm in size, growing on the dolomite. The tourmaline needles are less than 0.1 cm in diameter and may have a length up to 1.0 cm.

The occurrence of this pegmatite dike rich in feldspar and dolomite is reported here because such a type of pegmatite dike seems to be rather rare and is not reported earlier from the Caledonian zone of Norway. Probably this dike is the same as the one mentioned by A. Bugge (1963, p. 116) as molybdenite-bearing tourmaline vein. If so, this characterization is erroneous due to incomplete or wrong information given to Mr. Bugge. In the vicinity I found a few molybdenite-bearing quartz-tourmaline veins, also reported by Bugge. Grains of molybdenite occur also in hornblende veins cutting the peridotite and in small granite veins. Quartz veins reported to carry gold are also found within the gabbro massif.

Molybdenite mineralization is reported in various places in Velfjord in the area around Andalsvågen. This mineralization must be seen in relation to the adjacent massif of granodiorite, which is clearly an intrusive batholithic body with crosscutting contacts against the metamorphosed sediments and with a slight contact metamorphism.

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The geology of an area north of Gåsbakken, Sør-Trøndelag

By
Paul Carter.

Abstract

This paper constitutes a description of the geology of the Lower Ordovician strata outcropping in a 70 sq.km area, north of the village Gåsbakken in Sør-Trøndelag.

A review of the stratigraphy is given. Three phases of deformation are described.

Introduction

The mapping of this area was undertaken in order to correlate the known stratigrahical succession and structure of adjoining areas, which have been mapped by Th. Vogt (1945), Carstens, C. W. (1952) and Chadwick et. al (1964).

A general map and sections are presented showing the geology of the area. Correlation with adjacent areas is demonstrated by a general geological map of the area, and also a structural trend map.

Stratigraphical succession

The sedimentary, volcanic, and intrusive strata outcropping in the locality have been subdivided by Vogt (1945) into the Støren Series (mainly volcanic) of Arenigian age, and the Hovin Series (mainly sedimentary) of Llanvirnian and Llandeilian ages.

Table 1 shows the local succession in detail.

Carstens, C. W. (1951) showed the Støren Series (which he called the "Bymark Group") to be underlain by the Røros Group, which consists in the main of metamorphosed argillaceous sedimentary rocks.

With reference to Table 1, Blake (1962), from the evidence of graptolites found in the Bogo Shales of the Fjeldheim Beds (of Lower Hovin Series), has shown this horizon to be equivalent to the Phyllographtus Densus Zone (3b) of Middle Arenigian age. This suggests that the age boundaries suggested by Vogt must be revised.

Table 1 (after Th. Vogt, 1945).

		Break
Llandeilian	Lower Hovin Series	5. Hølonda Andesite 4. Hølonda Limestone (fossiliferous)
Llanvirnian		3. Hølonda Shale 2. Gaustadbakk Breccia & Almås Mudstone 1. Venna Conglomerate
		Break
Arenigian (Skiddavian)	Støren Series	3. Upper Støren Greenstone & Houe Slate 2. Jåren Beds 1. Lower Støren Greenstone

The succession established by the present author is given in Table 2, with those established by Vogt (1945) and Chadwick et.al (1964).

Table 2

Present author	Vogt (1945)	Chadwick et.al (1964)
HOVIN SERIES Upper Arenaceous Sequence (sandstones and grits)	HOVIN SERIES	HOVIN SERIES Nyplassen Beds (shales and sandstone)
Porphyrites (intrusive and/or extrusive)	Hølonda Andesites	Intrusive Porphyrites
Shale and limestone Sequence	Limestone Shale	Fjeldheim Beds Shales Limestones Sandstones
Lower Arenaceous Sequence Limestones and sand-stones grits	Gaustad Breccia and Almås Mudstone	
Conglomerates	Venna Conglo-merate	Fjeldheim Conglomerate
BREAK	BREAK	BREAK
TUFFS STØREN SERIES LAVAS (undifferentiated lavas)	STØREN SERIES	STØREN GROUP (sedimentaries, vol-canics, pyroclastics)

A detailed description of the succession tabulated by the present author is given below.

The Støren Series

In the area the Støren Series strata consist of basic lavas, overlain by tuffaceous deposits.

Basic Lavas

The basic lavas include both massive (flow) and pillow lavas, indicating underwater deposition, at least in part. The characteristic sagging or "U"ing of younger semi-molten pillows into those that have already been deposited and solidified, gives good younging data. Secondary alteration of the lavas has resulted in the widespread occurrences of epidote.

Structureless quartz with ferrous impurities - locally known as jasper - occurs in many of the interstices between the pillows. This indicates a silica-rich sea, due to the underwater effusion of silica during the formation of the lavas.

Tuffs

The tuffs show two main varieties. A very distinctive tuff bed immediately overlies the basic lavas. It contains large idiomorphic crystals of feldspar and quartz in a fine-grained matrix predominantly of white mica. The concentration of idiomorphic crystals in the tuff varies from almost totally constituting the deposit, to being sparsely scattered in the fine-grained matrix. The mica matrix is probably the result of secondary alteration after fine-grained feldspars. Secondary alteration has also produced chlorite around the edges of the idiomorphic crystals, and in the fissures.

A very fine-grained tuff bed is occasionally found overlying the distinctive tuff band, and below the Lower Hovin Conglomerate. Its original composition and texture has been greatly changed by secondary alteration.

The Støren Series is overlain by a conglomerate of Lower Hovin age which consists of fragments of Støren lavas and angular fragments of jasper at the base. More rounded cobbles and pebbles occur as the perpendicular distance from the Støren Series strata increases. The deposition of the thick conglomerate layer indicates uplift of the Støren volcanic rocks accompanied by rapid erosion, producing widespread pebble beaches. The Hovin Series strata do not, however, overlie the Støren Series with a very marked degree of angular uncomformity.

The Hovin Series

The Hovin Series is represented in the area mapped by four main sequences of strata which for convenience may be termed the Lower Arenaceous Sequence, the Limestone and Shale Sequence, the Porphyrites, and the Upper Arenaceous Sequence. These names are strictly for local use in the area under consideration and are not being put forward as alternatives for the many names that have already been given for the rocks of Lower Hovin age, outcropping in adjacent areas.

Lower Arenaceous Sequence

This sequence is made up of conglomerates, which occur mainly at the base, grits, sandstones and tuffaceous material, with very local developments of limestone and shale. The sequence thins from circa 600 m by Langkjøsen to circa 300 m in the Jårengrenda.

The conglomerates rarely have a framework of pebbles in contact with each other. More usually pebbles are scattered in greater or lesser quantities in an unsorted matrix. Poor sorting is general in this lowest sequence and in any one outcrop grain size can vary from cobbles to grits or shales. Overall it can be said that the beds become coarser downwards. The pebbles are usually of jasper (in the local sense of the word - amorphus quartz, stained red), and green probably volcanic rocks ascribed by Chadwick et.al. to the Støren "Greenstone" lavas.

The jasper ranges up to blocks almost a metre in diameter in the conglomerate exposed in road cuttings along the north edge of Svorksjøen, and is often less rounded than the well rounded rock pebbles.

The beds sometimes show good sedimentary structures. An outcrop on the southern shore of Morsjøen shows a sharply defined layer of pebbles lying on fine sandstones but grading upwards gradually through grits with scattered pebbles into sandstones over a vertical distance of 2 m. What appear to be turbidite units 30 cm thick with grit bases and shaly upper parts occur in the valley north of Sjømoen and examples of load casting occur nearby. False bedding is shown in an outcrop on the hill just west of Sundet. All these structures provide good younging data for analysing the structure of the area.

Limestones and Shale Sequence

It is convenient to consider together the limestones and shales which lie above the Lower Arenaceous Sequence, since their outcrops suggest them to be local developments within the same general horizon. The sequence varies in thickness from over 300 m. in the Jårengrenda to nil in the far west

Limestones develop locally at or near the top and bottom of the sequence and are fairly continuous along the strike. They can be white, grey or black, are always recrystallized, and are coarse- or fine-grained depending on the amount of recrystallization that has taken place. It is therefore usually difficult to ascertain their original form of deposition, although recognisable reef-breccias sometimes occur, for instance in outcrops between Konstadløkken and Blokkan. The limestones also often contain "ruckled bands" of arenaceous material which are usually about 1-3 mm thick. The bands are sedimentary features and the ruckling illustrates well the plastic deformations of the limestones during folding.

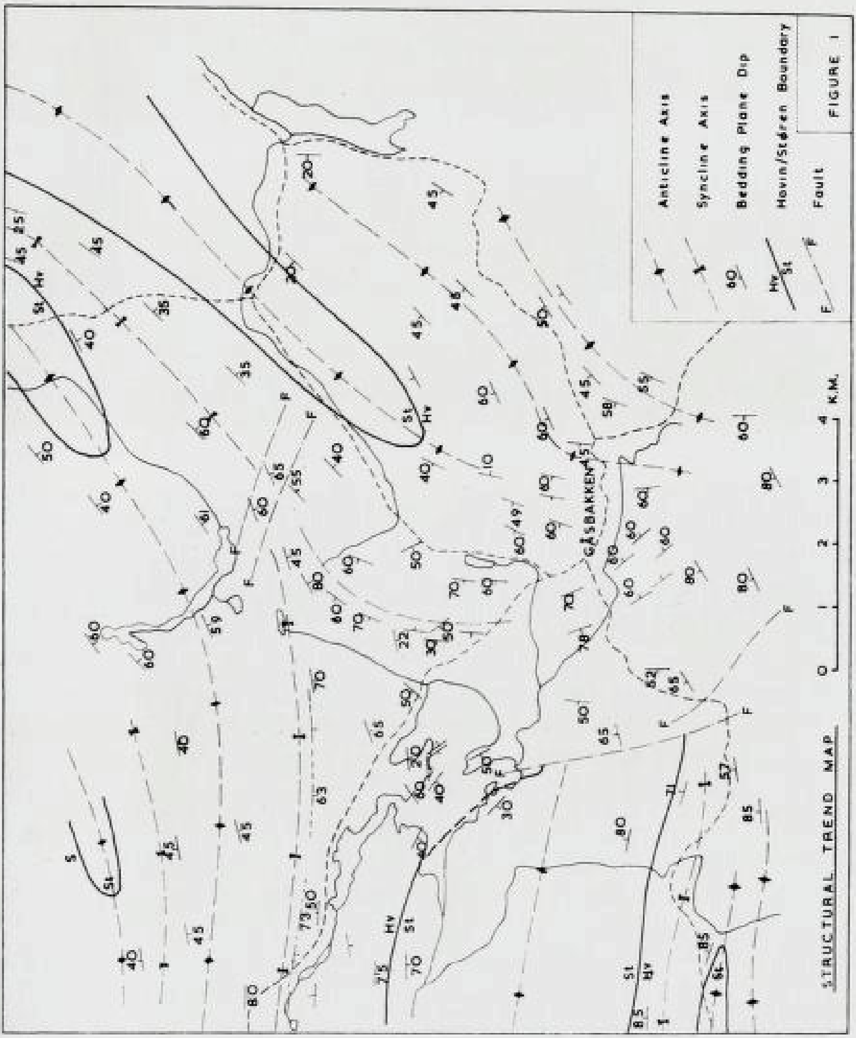
The shales are predominantly grey though sometimes green. Inclusions or flecks of an iron mineral, identified by Vogt (1945) as pyrrhotite, often occur. Occasional sandy bands indicate the bedding, and the shales also sometimes contain lime nodules.

The Porphyrites

It is convenient to deal with the Porphyrites here, as they occur structurally above the Limestone and Shale Sequence, and below the Upper Arenaceous Sequence. The Porphyrite sheet varies from a maximum of circa 300 m around Konstadløkken to nil near Klefstad.

The two pioneer workers in this area, Th. Vogt (1945 and Carstens C. W. (1951), clearly disagreed as to the stratigraphical relationships of the porphyrites. The latter regarded them as intrusive and probably discordant in most cases, whilst the former author regarded them as lavas normally interbedded in the Hovin Group. Vogt even subdivided the porphyrites stratigraphically into Almaas and Berg types, but he acknowledged a discordant intrusive relationship for a minority of the outcrops. Chadwick et.al. considered them to be mainly concordant intrusions.

The present author noted that the Porphyrite sheets do not show any of the typical features normally associated with lava flows such as "blocky" or "ropy" flow surface features, or any evidence of separate flows. However, a prominent Pyroclastic layer is frequently found directly underlying the Porphyrite. This does suggest a possible volcanic origin. Some fragments of the country rock could however, be excepted to be found at the base of a concordant intrusion due to fracturing of the country rock during intrusion. Evidence of strong heating of the country rock in the immediate vicinity of the Porphyrite is given by metamorphosis of the limestone adjacent to the Porphyrite into marble. Porphyrite dykes were found in three outcrops, intruded into the



Støren Series strata and the Lower Hovin conglomerates. These were presumably feeders for the overlying Porhyrite sheets. The author found the field evidence for the precise origin of the Porphyrites to be inconclusive.

Upper Arenaceous Sequence

This sequence forms the youngest strata found in the area. It follows a continuous outcrop from Restad extending south-west along the strike until it is cut off by the major fault south of Morsjøen. The beds are equivalent to the "Restadgrøtås sandstones" of Vogts "Jåren Beds". A green-white weathering surface is displayed by these beds which are better sorted than the beds of the Lower Arenaceous Sequence, they consist of fine-grained grits, sandstones and some shaly bands The sequence is about 150 m thick with the top not seen.

Structural geology

The structural outline plan (Fig. 1) indicates the main directions of the folding in the area. The structure has been determined by three phases of deformation, which took place during the Caledonian orogeny. Chadwick et.al. (1964) have called these the F_1, F_2 and conjugate folding phases respectively.

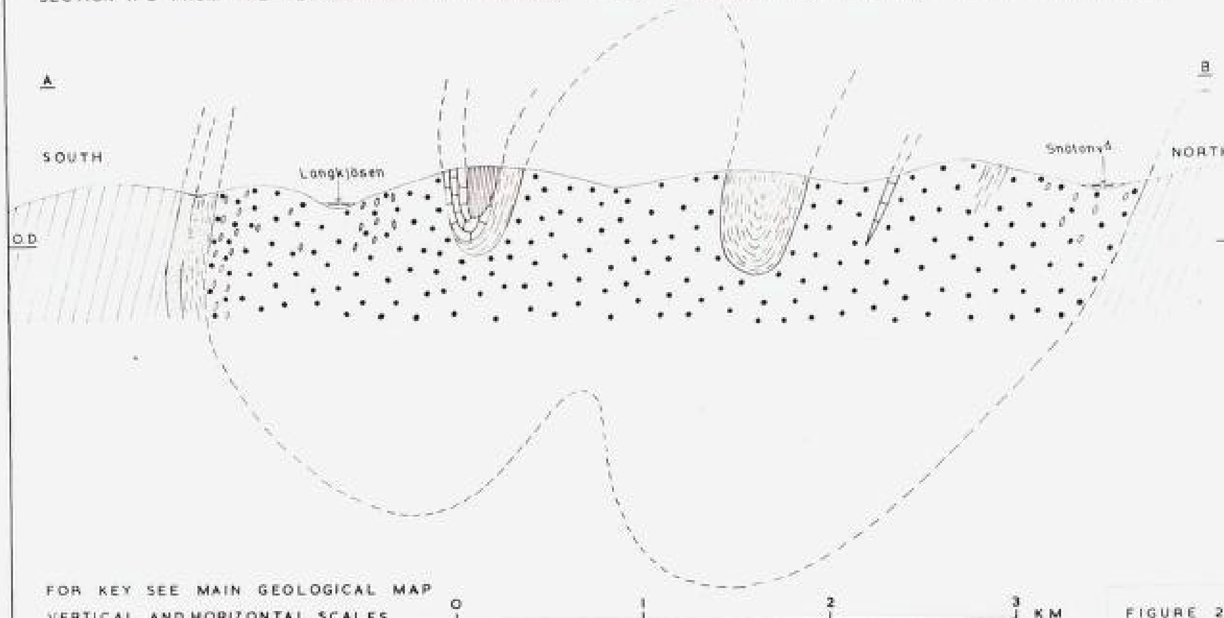
The F_1 phase has produced most of the major folding occurring in Sør-Trøndelag. The strike of the F_1 folding follows a south-westerly direction from the Trondheim area, swinging round more to the west in the area mapped by the author. The F_2 folding phase constituted a second major deformation producing minor conjugate folding. Some minor faulting has occurred. The overall structure of the area constitutes a major F_1 syncline upon which has superimposed the effects of the F_2 folding.

First or F_1 Phase of Folding

The major F_1 syncline strikes south-westerly parallel to the Jårengrenda in the eastern part of the area, swinging round into a westerly strike to the north of Svorksjøen. Hovin Series sediments, with a core of porphyrites and Upper Arenaceous Sequence rocks, have been downfolded by the syncline into Støren Series volcanic rocks, which form the synclinal limbs (Fig. 2). Thus the southern limb of the syncline brings Støren Series rocks to the surface south of Svorksjøen, and south-east of the Jårengrenda. The strata outcropping on the southern limb are overturned between 0° and 20° from the vertical. Evidence for the overturning is given by sedimentary structures in the Hovin Series strata and "V"ing of the pillows in the Støren volcanics. On the northern limb of the syncline, Støren Series volcanics outcrop near Snoton Lake in the extreme north-west of the area, and at Konstad Sæter in the extreme north-east.

The major F_1 folding was accompanied by minor folding, rodding structures parallel to the minor fold axes, shear planes and tension cracks. These structures are best developed in the Hovin Series strata to the north and west of Svorksjøen. The direction of minor fold axes and rodding is east to west, following the strike of the major folding. The majority plunge towards the east, at angles of between 5° and 35° from the horizontal. The tension cracks,

SECTION A-B FROM THE WESTERN ARM OF SVORKSJÖEN IN THE SOUTH-WEST TO SNÖTONV^{d} IN THE NORTH-WEST



FOR KEY SEE MAIN GEOLOGICAL MAP
VERTICAL AND HORIZONTAL SCALES

FIGURE 2

generally quartz filled, occupy planes perpendicular to the fold axes. The poles to the tension cracks therefore plunge parallel to the minor fold axes. Shear planes are found, parallel to the minor fold axial planes.

Second or F_2 Phase of Folding

The major F_2 folding has been shown by Chadwick et.al. (1964) to be superimposed on the F_1 folding, from the evidence of minor fold interference patterns. A north-westerly striking syncline has been produced with its axis running through the village of Gåsbakken. The core of the syncline is occupied by the Porphyrites, and the limbs by the Lower Arenaceous Sequence strata of Hovin age, and by rocks of the Støren Series. This fold phase appears to be responsible for the arcuate strike of the F_1 syncline and could be responsible for the generally easterly plunge of the F_1 minor fold features in the west of the area. Minor folding and associated structures along F_2 axes confirm the general direction of this phase of folding as shown by the bedding plane strike. The folding due to superimposition of F_2 folding on F_1 structure is complex, but difficult to interpret in detail, due to the poor rock exposure.

Third or Conjugate Phase

The third phase of deformation did not affect the major structures of the area. Compressive stresses in the rock produced conjugate folding on a minor scale. This has refolded the minor structures produced by previous fold phases.

Faults

Faulting has occurred in this area only on a minor scale. Since the faults transsect F_1 structures, but have not been affected by the F_2 deformation, they are contemporaneous with, or younger than the latter.

Two parallel faults striking north-west occur immediately south of Morsjøen in the east-central part of the area. A fault block of (probably) Lower Hovin conglomerate has been uplifted into the porphyrite and Upper Arenaceous Sequence core of the syncline. Due to the differential erosion, fault scarps occur leaving the conglomerate block upstanding.

On the Ramsberget Peninsula which juts out into Svorksjøen, a fault striking nearly east-west, has brought the porphyrites down against Lower Hovin conglomerates. A prominent fault scarp has resulted from differential erosion, with the porphyrites forming the scarp-face.

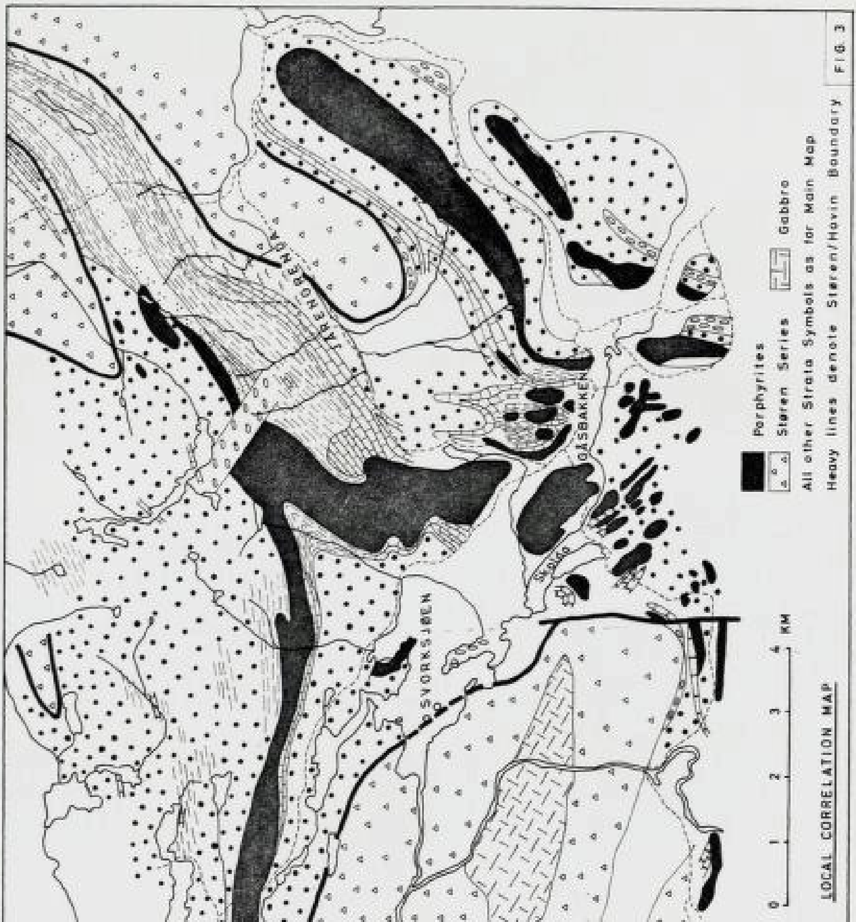


FIG. 3

Local correlation

The stratigraphy and structure of the area correlates well with those established for adjacent areas by Th. Vogt (1945), Carstens C. W. (1951 and 1952) and Chadwick et.al. (1964). The probable correlation of the strata in the area mapped by the present author, with the strata outcropping in the adjacent areas, is shown in table 2. The geological sketch-map (Fig. 3) indicates the continuity of the strata along the strike into the surrounding areas. The structural trend map (Fig. 1) shows the continuity of the structures. Points of disagreement with previous authors are as follows:

Vogt (1945) believed the Hovin Series strata, downfolded by the F_1 syncline

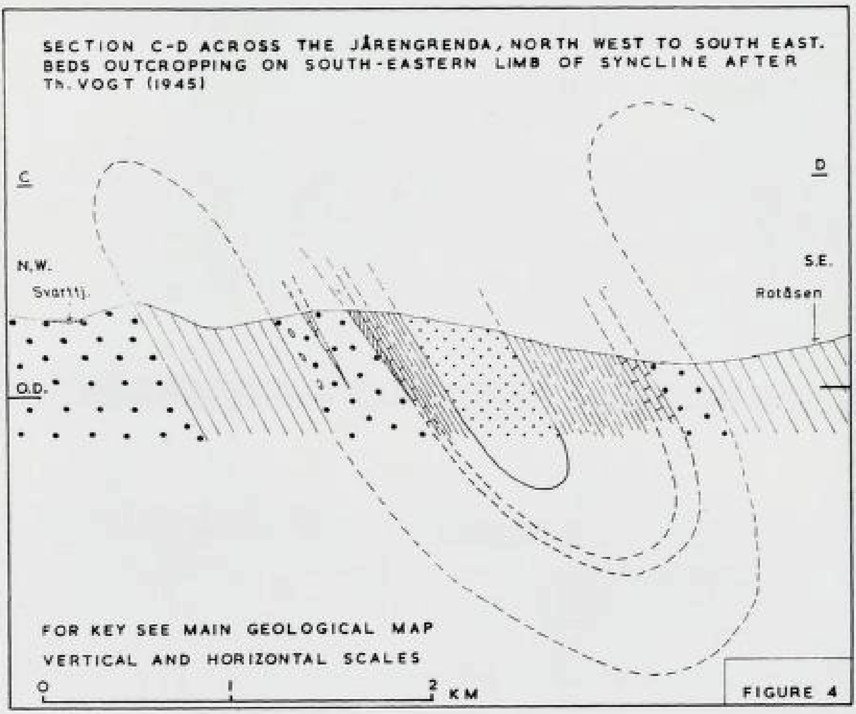


FIGURE 4

in the Jårengrenda, to be a sedimentary series of Støren age. He gave the name "Jåren Beds" to this series, and named the Støren volcanic rocks brought to the surface on the north-western and south-eastern limbs of the syncline, the "Lower Greenstones" and "Upper Greenstones" respectively. Carstens (1951) opposed this interpretation claiming the "Jåren Beds" to be of Hovin age. This sedimentary series has in fact been shown by the present author to comprise synclinally folded strata of Lower Hovin age. A section across the syncline is given in Fig. 4.

Carstens C. W. (1952) believed Støren Series and Røros Group strata to outcrop to the north of Bøverdalshaugen in the extreme west of the area. The strata at this point however belong to the Lower Hovin Series. The outcrop mapped by Carstens as Støren "Greenstone" is in fact porphyrite of Lower Hovin age.

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The author wishes to express his gratitude for the great help, both in the preliminary preparations for the work and in the field, given by Per Sandvik of the Orkla Grube Aktiebolag. A detailed report of the survey lies with

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Geophysical measurements in Jeløya

By

B. L. Ramachandra[1]) G. Lind[1]) S. Saxov[1])

Summary

As a part of geological/geophysical student field courses a series of magnetic and gravimetric observations has been carried out at Jeløya. Geological samples have been collected and the physical properties of the samples have been determined in the laboratory.

The results of the surveys and the laboratory investigations are presented in form of figures.

Introduction

Geological Institute, Aarhus University, arrange every summer a compulsory field course in geological/geophysical mapping. The course is attended by students of geology and geography by preference in the summer following their second year of study. After consultation with Professor, Dr. St. Skjeseth, Geological Institute, NLH, Vollebekk, this course takes place in Østfold fylke, Norway. The courses commenced in 1965 at the road-cuttings in the Moss area along the new highway from Svinesund to Oslo. Reports on the geological results will be published on a later occasion. Geophysical measurements, i.a. gravimetric and magnetic observations, were undertaken in the same area in 1965 and extended in 1966 to include Jeløya. This report, which is considered preliminary, is concerned with the Jeløya survey. It is intended to make additional geophysical measurements later on and also to investigate the geological conditions in detail.

Geology

Jeløya belongs geologically to the Oslo field with young rocks, paleozoic and permian, downfaulted in the precambrian. The geological setting is shown on fig. 1, which is adopted from Brøgger and Shetelig (1926). The oldest

[1]) Institute of Applied Geophysics, Department of Geology, Aarhus University, Aarhus C, Denmark

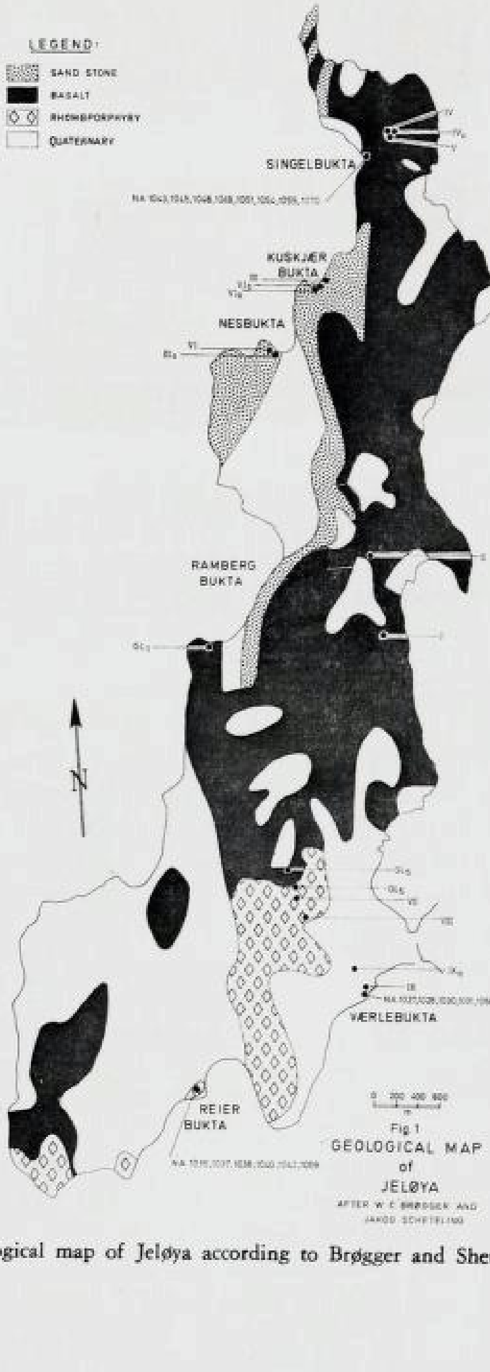

Fig 1. Geological map of Jeløya according to Brøgger and Shetelig (1926).

rocks on Jeløya are downtonian sandstone and conglomerate which occupies the western part of the island, and having a gentle dip towards the east. On top of these lies permian basalt (B_1), which covers the main part of the island. The youngest rocks are the permian rhomb-porphyries (Rp_1) with outcrops in the southern part of the island. The above mentioned faultline runs east of the island in a north-south direction. Where no outcrops are shown on the map, the rocks are covered with quaternary clay, sand, and moraine.

Table 1 shows the rocksamples which were taken and analysed for their physical properties. The localities are shown on both fig. 2 and fig. 3 together with density values on the gravimetric map and susceptibility on the magnetic map. It should be noticed that the samples with numbers beginning with NA are taken for paleomagnetic measurements and that is why many samples are taken at the same locality. Samples GL 6 and II, which both are basalts, seem to differ a little from the rest in that group. Sample GL 6 is a porous type and strongly zeolitised, sample II is a basalt with phenocrysts of plagioclase and it seems to differ from the rest only by having a more reddish groundmass. A more detailed study of the petrographic properties of the rocks compared to their physical properties will be made in the near future.

Gravity

The gravimetric readings have been carried out by means of Worden gravimeters 142 and M 779. The observations are referred to Oslo Fundamental Gravity Stations with G = 981.92815 gals. A density value of 2.67 gr/cm^3 has been employed. The coordinates are taken from the topographical maps in the scale of 1 : 5 000, the reference point being Jeløya radio $\varphi = 59°26'00''$ and $\lambda = 10°35'42''$ E Grw. The elevation values are taken partly from polygon-points, the heights being known in millimetres or centimetres, and partly from dot-points in the topographical maps, the heights being in metres or half-metres. Topographic correction is not applied.

Two stations determined by the Geographical Survey of Norway, station G 36 top 3 Jeløya radio $\varphi = 59°26'00''$, $\lambda = 10°35'70''$ E Grw, h = 0.1 m, G = 981.90938 gals, and Bouguer anomaly + 30.15 mgals, and G 36 top 4 Nestangen $\varphi = 59°28'93''$, $\lambda = 10°38'57''$ E Grw, h = 0.3 m, G = 981.90820 gals, and Bouguer anomaly + 25.14 mgals are included in the map. We have measurements very close to these stations, No. 173 with an anomaly of + 30.03 mgals, and No. 4 with an anomaly of + 26.07 mgals. The agreement between the two sets of observations must be considered satisfactory.

In fig. 2 the Bouguer anomalies are plotted and isogal contours are drawn, the equidistance being 1 mgal. The general trend is decreasing anomalies from

Table 1

Sample No.	Rock-Type	Locality	Density g/cm³	Remanent Magnetiz. cgs	Suscepti-bility cgs	Q
VII	Rhomb-Porphyry	Vardeveg	2.56	0.00016	0.00016	1.95
NA 1035	—	Reierbukta	2.56	0.00006	0.00012	1.00
NA 1068	—	Værlebukta	2.61	0.00156	0.00032	9.64
NA 1069	—	Reierbukta	2.56			
NA 1037	—	—	2.56			
NA 1038	—	—	2.56			
NA 1040	—	—	2.56			
NA 1042	—	—	2.56			
VIII	—	Vannbasseng	2.60	0.00012	0.00004	5.07
IX	—	Værlebukta	2.60	0.00129	0.00006	39.92
GL 5	—	Vardeveg	2,61			
NA 1027	—	Værlebukta	2.61	0.00084	0.00054	3.10
NA 1028	—	Værlebukta	2.61	0.00036	0.00007	10.08
NA 1030	—	—	2.61	0.00113	0.00010	22.87
NA 1031	—	—	2.61	0.00082	0.00016	10.08
IX a	—	200 m NW of Værlebukta	2.64	0.00114	0.00179	1.27
GL 6	Basalt	Orkerød	2.62			
GL 1	—	Rambergbukta	2.73			
NA 1043	—	Singelbukta	2.74	0.00089	0.00236	0.76
NA 1045	—	—	2.74	0.00047	0.00233	0.41
NA 1048	—	—	2.74	0.00040	0.00299	0.27
NA 1049	—	—	2.74	0.00036	0.00278	0.27
NA 1051	—	—	2.74	0.00078	0.00140	1.11
NA 1054	—	—	2.74	0.00041	0.00341	0.24
NA 1055	—	—	2.74	0.00062	0.00330	0.38
NA 1070	—	—	2.74	0.00032	0.00419	0.16
IV	—	300 m NNE of Singelbukta	2.76	0.00233	0.00565	0.83
V	—	—	2.77	0.00202	0.00655	0.63
IV a	—	—	2.78	0.01484	0.00503	6.03
I	—	Nesvegen	2.79	0.00117	0.00008	26.19
II	—	Kjellandsvik	2.87	0.00209	0.00439	0.96
VI	Sandstone	Nesbukta	2.63			
III	—	Kuskjærbukta	2.64			
III a	—	Nesbukta	2.65			
VI b	—	Kuskjær	2.67			
VI a	—	—	2.69			

west to east. The peak value + 31.09 mgals is obtained in the south-western part of the island, while the lowest value of + 21.81 mgals is to be found at the northern end. The general trend is also seen at the gravimetric map Oslo-Feltet (NGO, 1960).

The location of the samples listed in table 1 are plotted on fig. 2 and the corresponding density values are listed. Rhomb-Porphyry samples have values from 2.56 to 2.64 gr/cm^3, the mean being about 2.60 gr/cm^3. Basalt runs from 2.62 to 2.87 gr/cm^3; sample GL 6 with 2.62 gr./cm^3 and sample II with 2.87 gr/cm^3 are not considered representative for basalt; the mean value of the remaining samples is about 2.76 gr/cm^3. Sandstone varies from 2.63 to 2.69 gr/cm^3, the mean value being 2.66 gr/cm^3.

Magnetic Measurements

An Askania (GfZ) torsion magnetometer with a scale constant of 230 γ per scale division, where the vertical component Z can be measured with an accuracy of $\pm$ 10 gamma, was used for the magnetic investigation. The field work was carried out in July 1966 during seven days and it covered most of the area. The points measured lie mostly along roads and around the coast. The distance between the points measured varies from 200 - 500 m. A main base station at Ås (NLH) was established, and five sub-base stations in the Jeløya have been selected, being remeasured two to three times a day.

Fig. 3 shows the magnetic map of $\triangle$ Z. The anomaly contours are marked in gamma units. The location of the samples and the magnetic properties of the analysed samples which have been measured by means of an oerstedmeter are listed in table 1 and plotted on fig. 3.

Rhomb-Porphyry samples have remanent magnetization and susceptibility values from 0.00006 to 0.00156 cgs, the mean being 0.00074 cgs and 0.00032 to 0.01484 cgs, the mean being 0.00105 cgs, respectively.

Basalt samples have RM values from 0.00004 to 0.00179 cgs, the mean is 0.00033 cgs and susceptibility value from 0.00008 to 0.00655, the mean being 0.00339 cgs. From these values it is seen that the basalts are stronger magnetised than Rhomb-Porphyry and that basalts also have higher susceptibility.

Because of the complexity in magnetic anomalies only representative curves are drawn. More variation and non-uniformity in the magnetic intensity can be seen from the values. This is accounted for inhomogenity in the rocks.

It is worth noting that basalt sample II with rather strong magnetic properties is situated close to the local magnetic maximum of 1550 γ while basalt sample I with very weak magnetic properties is placed close to the even stronger magnetic maximum of 3150 γ.

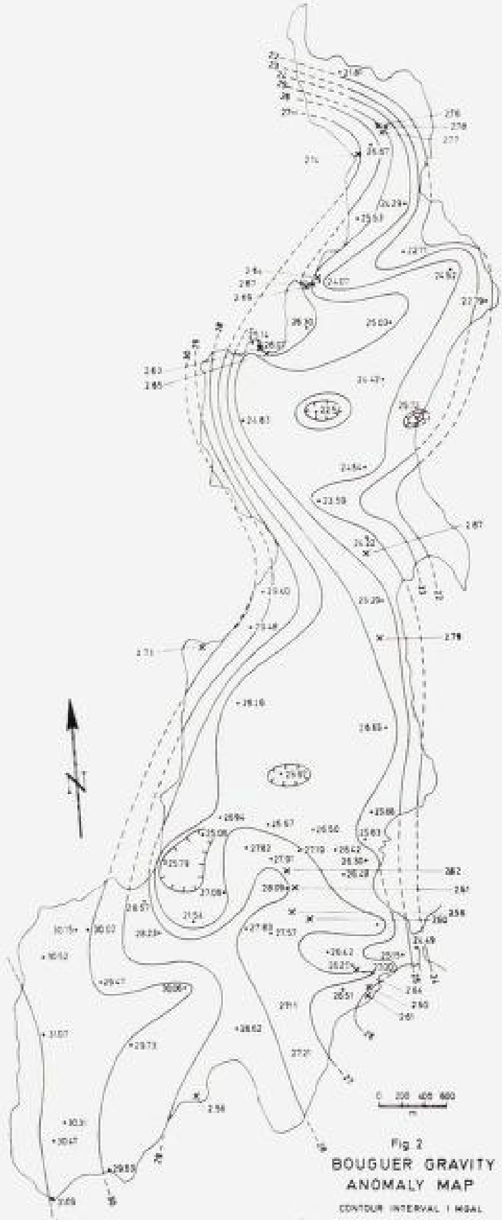

Fig. 2. Gravimetric map of Jeløya.

Discussion

By a comparison of the geological picture (fig. 1) and the isogal contour picture (fig. 2) a certain correlation can be seen; i.g. the minimum of 26 mgals between the two basaltic blocks and the minimum of 26 mgals in the midst of the basaltic block; see also the gravimetric features in the sandstone area. Large parts of the Jeløya island is covered by overburden, and it is the intention by geophysical means to determine this thickness during the coming field season. More discussion concerning this problem is therefore to be taken up on a later occasion.

The magnetic anomaly picture is characterized by rapid changes of several hundred gammas. Some of the anomalies can be correlated to the geological formations. The very strong, local maxima must be due to special conditions in the basalts.

The correlation between the gravimetric and the magnetic anomalies is but small. Certain trends can be seen, however, before either more details of geophysical data or a reworked geological map are present, and at the same time the geological/geophysical conditions in Jeløya can be considered together with similar conditions in the main-land area there is no purpose in studying the present figures.

References

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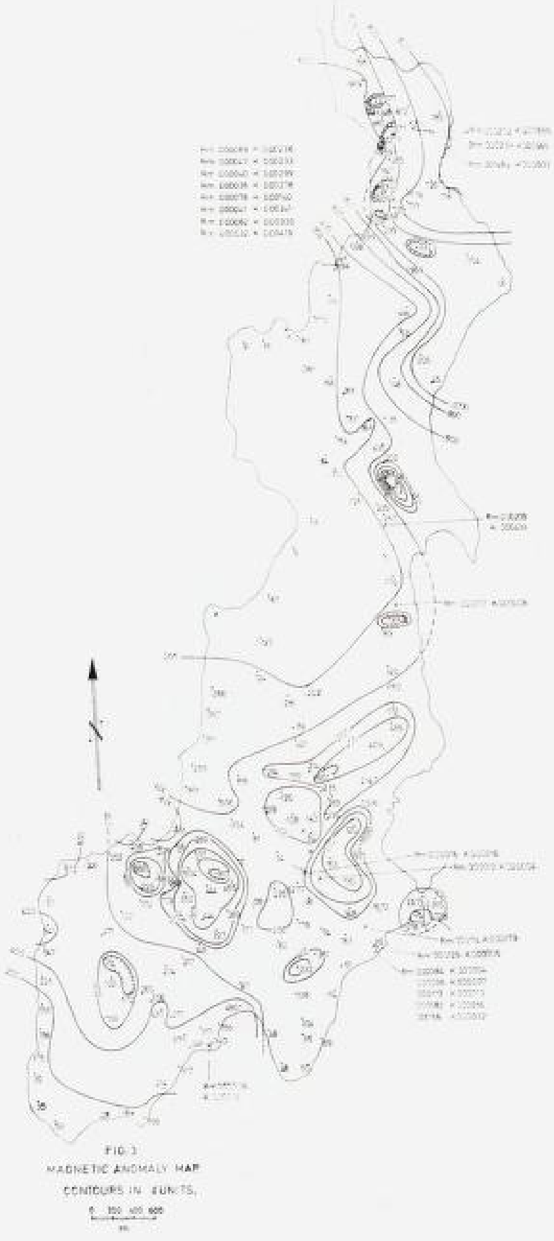


Fig. 3. Magnetic vertical intensity map of Jeløya.

Appendix

Bouguer anomalies computed in the usual manner, being referred to Oslo Gravity Fundamental Station G = 981.92815 gals, the International Formula of 1930 being applied as far as the normal gravity in gals in the surface of the sea is concerned, and a density value of 2.67 gr/cm^3 being employed. Elevations are either levelling points or dot-points from the 1 : 5 000 topographical maps.

No topographic correction is applied.

Station No.	φ	λ	H	G	Bouguer Anomalies
2	59°25′.76	10°39′.20	2.4 m	981.90295 gals	+ 24.49 mgals
3	24.84	36.19	1.0	90714	+ 29.63
4	27.91	38.16	0.5	90774	+ 26.07
5	27.51	39.77	0.5	90542	+ 25.72
31	26.74	37.99	100.22	88647	+ 25.97
32	26.38	38.28	53.129	89579	+ 26.50
33	26.29	38.46	42.486	89768	+ 26.42
34	26.15	38.51	29.120	90021	+ 26.49
35	26.19	38.81	16.603	90253	+ 26.30
36	26.41	38.80	31.053	89951	+ 25.83
37	26.56	38.93	34.204	89914	+ 25.88
38	26.83	39.11	37.669	89957	+ 26.65
39	26.24	38.19	59.83	89485	+ 27.19
40	26.24	37.81	76.293	89233	+ 27.91
41	26.33	37.84	79.197	89178	+ 27.82
42	26.51	37.93	85.971	88966	+ 26.67
43	26.02	37.97	80.339	89153	+ 28.09
44	25.86	37.92	59.491	89487	+ 27.57
45	25.92	37.48	42.352	89863	+ 27.83
46	25.96	37.08	36.381	89958	+ 27.54
47	25.94	36.77	28.773	90173	+ 28.23
48	25.76	38.33	32.804	89889	+ 26.42
49	25.52	38.34	15.321	90211	+ 26.51
50	25.45	37.94	46.033	89657	+ 27.11
51	25.58	38.67	8.626	90396	+ 27.00
52	25.71	38.73	12.932	90252	+ 26.20
53	25.73	39.11	7.512	90256	+ 25.15
54	25.85	38.85	11.250	90312	+ 27.14
93	30.25	39.28	10	90470	+ 21.81
94	29.88	39.42	45	90218	+ 26.67
95	29.58	39.74	25	90336	+ 24.29
96	29.31	39.78	44	89767	+ 22.71
97	29.18	40.23	50	89813	+ 24.52

Station No.	φ	λ	H	G	Bouguer Anomalies
98	29.08	40.58	5	90513	+ 22.79
99	29.01	39.47	90	89053	+ 25.03
100	28.73	39.35	59	89565	+ 24.42
101	29.54	39.24	25	90433	+ 25.33
157	27.42	39.18	25	90149	+ 25.29
158	27.83	39.03	35	89900	+ 24.22
159	28.08	38.54	20	90157	+ 23.59
160	28.54	37.92	10	90546	+ 24.83
161	28.58	38.63	15	90225	+ 22.54
162	29.20	38.78	5	90656	+ 24.07
163	28.98	38.60	25	90456	+ 26.30
164	28.24	39.10	65	89405	+ 24.64
165	27.75	38.12	2.5	90604	+ 29.40
166	27.50	37.85	30	90480	+ 29.48
167	27.07	37.60	84	89068	+ 26.54
168	26.09	37.32	44	89779	+ 27.08
169	26.36	36.83	2	90512	+ 25.79
170	26.43	37.18	20	90096	+ 25.08
171	26.64	37.37	100	88734	+ 26.94
172	26.11	36.58	20	90403	+ 28.57
173	25.92	35.81	19	90542	+ 30.02
174	25.74	36.08	20	90444	+ 29.47
175	25.39	36.31	20	90423	+ 29.73
176	25.65	36.92	20	90491	+ 30.06
177	25.39	37.35	12	90470	+ 28.62
178	25.22	38.08	20	90149	+ 27.21
367	25.82	35.54	7.5	90806	+ 30.52
368	25.56	35.45	10	90776	+ 31.07
369	25.10	35.77	35	90145	+ 30.31
370	25.07	35.60	75	89366	+ 30.47
371	24.78	35.64	0.5	90779	+ 31.09

Description of the geological maps "Tromsø" and "Målselv", Troms

I. The Precambrian window of Mauken-Andsfjell.

By

Kåre Landmark.

Contents

Abstract

This paper is the first part of a description of the bedrock on the two 1 : 100,000 map sheets, "Målselv" and "Tromsø". The geological map, "Målselv", accompanies the paper. This part deals with the "window" of Precambrian rocks exposed beneath the Caledonian nappes in Målselv, just N of latitude 69° N. The basement consists of mountainous areas that rise far above the surrounding Precambrian peneplane. These Precambrian rocks are composed of two main units, the Mauken amphibolite and the Andsfjellet granodiorite. The amphibolite is thought to be a meta-basalt. Chemical analyses indicate a "normal", relatively acid, basaltic composition. The analyses are compared to analyses of meta-basalts from other parts of the North Scandinavian Precam-

Fig. 1. Location of the area described (arrow).

brian. The granodiorite is believed to be a product of a Precambrian granitisation. Chemical analyses of the granodiorite are given. It is shown to have an intrusive contact against the amphibolite. The junction between these two Precambrian units and the overlying schists is a pronounced thrust plane, which is a part of the Caledonian "major thrust". The question whether these basement outcrops formed actual mountains as early as in the Precambrian, or whether they result from a Caledonian uplift is discussed, and the latter hypothesis is thought most likely.

Preface

The geological mapping of the "Tromsø" and "Målselv" sheets began as early as 1947, but because of other duties the ultimate working up of the field material has had to be put off until recently. The geological map of the "Målselv" sheet was printed in 1959, the "Tromsø" sheet is still in manuscript. The author has made various supplementary observations within the relevant areas since the completion of the actual mapping. Since the topographic base of the "Målselv" sheet was printed, several new roads have been built, and attention is drawn to them in the text where they affect the area now under consideration.

These two sheets, stretching from Tromsø in the N to Øverbygd, Målselvdalen, in the S, provide a profile across the Caledonian mountain chain at this latitude - from the highest nappes in the NW to the Precambrian basement in the SE, described in this paper. The relatively highly metamorphosed rocks making up the greater portion of the bedrock within these two sheets, are traditionally thought of as thrust nappes of Caledonian age. No fossils have been found within these maps.

Research in the area treated in this first paper has, prior to the present work, been confined to the mapping carried out by the geologist, Karl Pettersen, who was also attached to Tromsø Museum (Pettersen 1887). He mentions the granodiorite ("granite") of Andsfjell and Mauken, but not the amphibolite. He comments on the discordance between the granodiorite and the overlying schists, and concludes that, because of it, the lower rock is of Precambrian age. It is also designated such on his map of "Tromsø Amt" (Tromsø county) published in 1890, shortly after his death.

The An-content of the plagioclases has been determined by refractive index measurements. Extinction angles for hornblende have been calculated using the universal stage. In order to aid the localisation of place names on the map, an individual co-ordinate system (in blue) is superimposed. The co-ordinates are given in the text as, e.g. (8.3, 2.6).

The chemical analyses have been carried out by Statens Råstofflaboratorium (State Raw Materials Laboratory) and Norges Geologiske Undersøkelse (Norwegian Geological Survey). Norges Geologiske Undersøkelse and Norges almenvitenskapelige forskningsråd (Norwegian Scientific Research Council) have financed the research, and the last-named body has covered the cost of printing the map. The manuscript has been translated by my assistant, Richard E. Binns, B.Sc., who has also given valuable help and good advice during the preparation of the manuscript.

Introduction

The morphology of the S part of the "Målselv" sheet is dominated by the Mauken-Andsfjell ridge. The Målselv river is flanked in its middle stretch by Mauken, but further N has cut through the ridge to leave Andsfjellet as an isolated remnant to the NW. The ridge continues eastwards to Skjold, a few kilometres off the map.

The bedrock of the Mauken-Andsfjell ridge is the structurally lowest unit within the two map-sheets, "Målselv-Tromsø". A marked thrust plane everywhere forms the junction between this unit and the overlying rocks. The rock clearly represents autochthonous Precambrian, and the thrust plane is the Caledonian "major thrust" separating the Precambrian from the overlying allochthonous and par-autochthonous Caledonian metamorphics. This nappe outcrops on the highest summits of the Mauken ridge. The basement consists of an older amphibolitic complex, and a younger, chiefly granodioritic, unit. The central part of the basement has a common foliation which strikes NW-SE, with a very steep, frequently vertical, dip. The contact between the amphibolite and the granodiorite is conformable to this foliation so that the granodiorite flanks the amphibolite in the SW, generally with a very steep dip. The basement foliation differs from this only near the thrust plane where it bends to parallel the latter (see Figs. 9-12). Other rock types occur in the basement, but these are genetically related to the two main types. Thus, in the amphibolite are conformable sheets or sills of very fine grained quartz dioritic rocks. "Transition rocks" of dioritic composition - clearly granitisation products of the amphibolite - outcrop in the border zone between the amphibolite and the granodiorite, especially in Andsfjell and near the SE edge of the map. Finally, at the junction between the amphibolite and the overlying nappe is a layer of green schists, which may be interpreted either as a border facies of the amphibolite or as an autochthonous basal greywacke.

The amphibolite group

The Mauken amphibolite is a steeply dipping layer trending NW-SE. In its northermost part, near Moen, it is about 1200 m thick. Around Myrefjell, some kilometers further S, the thickness reaches some 3000 m. Whether the thickness is primary or due to tectonic effects, is not easy to determine. This

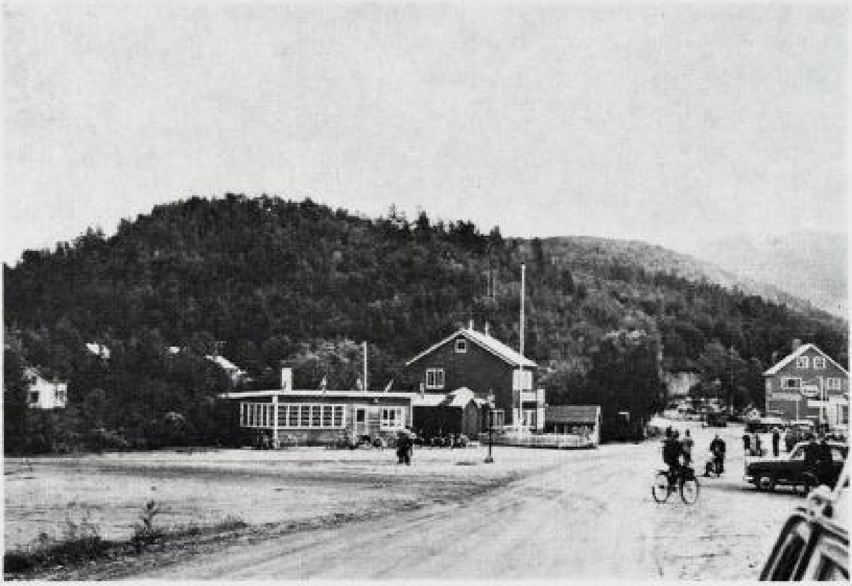


Fig. 2. Mauken seen from Olsborg (11.7, 8.6) towards SE.

matter will be discussed later. The unit has a uniform character, not varying particularly in composition or structure, either transversally, NE-SW, or longitudinally, NW-SE.

The amphibolite is very well exposed over the whole N part of Mauken. A particularly accessible profile is in the extreme N, eastwards from the bridge across Takelva, near Olsborg (11.5, 8.5) (Fig. 2). The amphibolite body shown on the map on Andsfjellet, consists, as we shall see, only to a lesser degree of amphibolite proper.

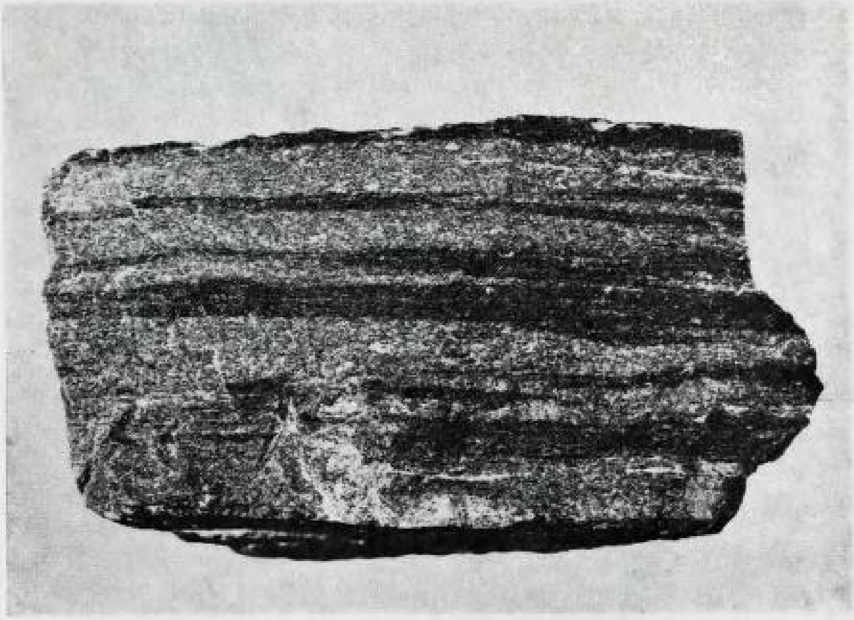
The foliation is primarily due to a parallel orientation of the hornblende (and biotite) crystals. Its strike is mostly 140 - 150°, but can range between 130 and 160°. This scatter is apparently due to local variations developed during the one deformation phase. Throughout the central area the foliation has a steep, often vertical, dip. Along the junction in the NE the amphibolite dips to NE. It dips to SW in the SW, near Maukdal (around 17, 2) in Målselvdalen, where it borders against the overlying schists and is not represented by the granodiorite.

The foliation also represents shear planes along which vertical movement has clearly taken place along the vertical lamination planes, not only during the recrystallisation phase of the hornblende and other minerals, but also continuing after that. Hence, the amphibolite minerals are frequently cataclastic. The shear planes are often coated with thin films of hydrothermally-formed quartz. Vertical shear planes are also frequently seen to trend in other directions - especially around 30°. Relatively minor slickensiding is occasionally developed.

Petrographically the amphibolite is rather heterogeneous. Thus, the amphibolite, itself, frequently has a streaky or banded structure. In addition, interlayered sheets, or sills, of quartz dioritic composition occur within the amphibolite. The quartz diorite bodies vary in thickness from about 20 cm to several metres. Such layers can, for example, be seen in the profile near Olsborg, or on the summit of Andsfjellet, as well as in the SE corner of the massif, outside the area covered by the map. The quartz diorite is a massive, grey rock with a very low content of mafic minerals. On weathered surfaces it appears very similar to the amphibolite, so that it is difficult to determine with accuracy the relative amounts of each. The quartz diorite, anyway, makes up a subordinate part. All the rocks of the amphibolite massif are completely recrystallized and highly cataclastic. The larger hornblende crystals are usually fractured, and often separated into isolated fragments. Feldspar and quartz often show mortar structure. Micro-shears are sometimes visible in thin sections.

The amphibolites are dark grey-green to nearly black. The dominant type is a homogeneous, massive rock, but it alternates with banded types. The banded amphibolites seem to occur subordinately in the massif, but the quantitative relationship between the two types is difficult to determine. Nor is it clear whether the banded types are fairly evenly distributed or mainly confined to certain zones. The homogeneous amphibolites frequently display only a weak NW-SE foliation parallel to the steeply dipping foliation of the massif as a whole, and show little tendency for splitting along that plane. The banded amphibolites are also massive rocks, without real schistosity, but with a greater tendency to split parallel to the banding. The banding results from an alternation of lighter grey, and darker layers. The individual bands have a breadth which varies from a few millimeters up to a couple of centimeters. The bands are often not very persistent, but wedge out, so that the structure becomes more streaky than banded (Fig. 3). In several places the amphibolites are cut by irregular networks of very fine quartz veins. The veins are usually only a few millimeters broad.

a



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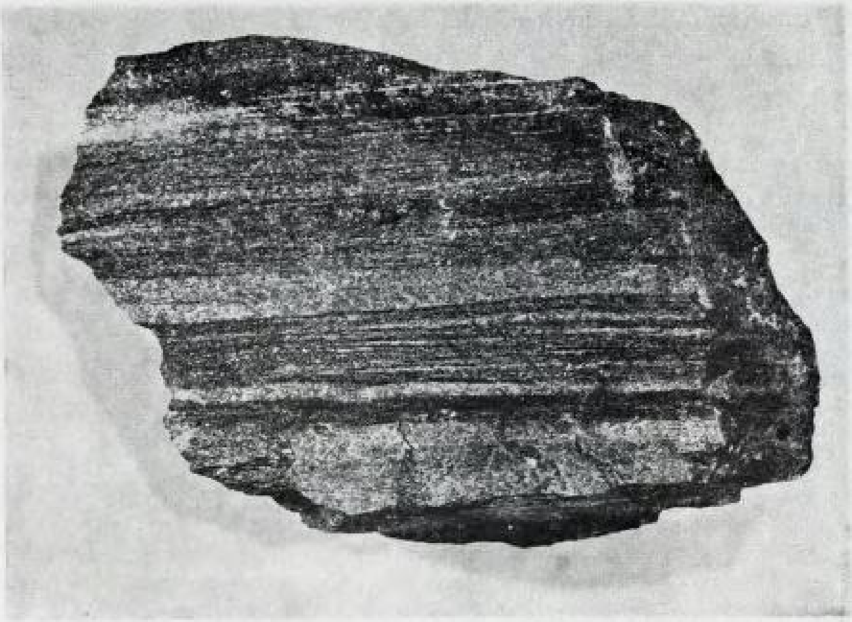

Fig. 3. Banded amphibolites: a) coarser bands, wedging out; b) streaky variety. ⅔ nat. size.

TABLE 1

Mineral assemblages of amphibolites from Mauken.

No.			Plag.	Hornbl.	γ/c	γ	Biot.	Musc.	Zois.
2287	Homogen. section		x	70-80 %	26°	1.670			
2248c	»	»	x	40-50 »	28°	1.68			
2290	»	»	x	30-50 »	24°	1.674			
617	»	»	x	20-25 »	26°	—	20-25 %		
2288	»	»	(x)				(x)	30 %	30 %
2261	Streaky section	Dark band	x	70-80 »	25°	—	(x)		
		Light band	x	15 »	—	—			
2446b	»	Dark band	x				50 %		
		Light band	x	20 »	26°	1.678			

On the basis of the mineral content of these rocks they can be divided into the following 4 types:

a) Amphibolites with hornblende as the sole dark mineral.

b) Amphibolites with hornblende and lesser amounts of biotite.

c) Bands with biotite, but without hornblende.

d) Bands with (clino-)zoisite, muscovite and biotite.

The pure amphibolite without biotite is the major variant. Table 1 summarizes the petrographic data on which this division is based.

These rocks contain phenocrysts of hornblende in an extremely fine grained groundmass of mainly light-coloured minerals. The hornblende forms relatively long-prismatic crystals, mostly 0.3 - 0.5 mm long, but occasionally up to 1 mm. The grain size of the groundmass is usually 0.02 - 0.04 mm. Where the hornblende content is especially large this mineral forms a more or less continuous mesh. The hornblende crystals are occasionally poikilitic and frequently have highly irregular outlines. Most of the hornblende individuals show sub-parallel orientation. Both the hornblende and the minerals of the groundmass display cataclastic effects. The crystals are often fractured, partly into separate pieces, and the quartz usually shows undulating extinction. (See Plate 7.)

The hornblende is an ordinary green hornblende, never actinolite. It is strongly pleochroitic, frequently in olive-green and pale yellow. Birefringence varies somewhat, between 0.020 and 0.025. The refractive index varies from 1.67 to 1.68. The extinction angle γ/c is high (25 - 28°). This is much higher than usually given for green hornblendes in metabasites from the mountain chain of N Norway (e.g. Vogt 1927, Bugge 1948, Vokes 1957). But Randall (1959) reports a similar value for hornblende in the Lyngen gabbro in Troms.

The biotite is brown and strongly pleochroitic. It occurs partly as independent flakes equal in size to the hornblende individuals, partly on the borders of the hornblende crystals where it is clearly recrystallized from that mineral. In the biotite-rich part of specimen 2446 b the biotite is less affected by cataclasis than are the other minerals.

The fine grained groundmass consists predominantly of feldspar and quartz. The feldspar is an acid plagioclase. Potash feldspar is not observed. The feldspar and quartz grains are of about equal size. Narrow twin lamellæ can be seen only in a few of the larger feldspars. The feldspar is often almost clear, and difficult then to distinguish from quartz, but it often contains small inclusions of sericite and rod-shaped zoisite. Large numbers of inclusions are never seen in the plagioclase. All the plagioclase seems to have the same degree of acidity, with an An content of 12 - 15 %.

Quartz occurs in all the specimens, though in varying amounts. The exact proportions of quartz and feldspar are difficult to determine, however. They seem to be about equal in specimen 2290, but otherwise quartz is considerably subordinate to feldspar. In specimen 2288, which chiefly consists of muscovite and zoisite, feldspar is insignificant in amount. The quartz almost always has undulating extinction.

Small grains of (clino-)zoisite and epidote occur in the groundmass of the majority of the specimens. In the zoisite-rich specimen, 2288, zoisite occurs as highly irregularly-shaped aggregates 0.3 - 1 mm in size. The zoisite gives anomalous, bluish, interference colours. The muscovite in this specimen consists chiefly of fine grained sericite (0.01 - 0.02 mm) in contact with the zoisite, but a few larger, scattered flakes are also seen.

Small amounts of sphene and iron-ore occur in all the specimens. A little apatite is found in specimen 617.

No pyroxene relicts have been observed, nor any alteration of hornblende to chlorite. The mineral combination muscovite-hornblende does not occur in any of the specimens. The small amounts of epidote and zoisite normally present, indicate that the original plagioclase has been calcium-poor.

According to the mineral assemblages outlined above the rocks in question must belong to the upper part of the epidote-amphibolite facies, in which a plagioclase with up to about 15 % An is stable.

TABLE 2

A comparison of the Mauken amphibolites with some other basaltic rocks

	Daly I	Mauken Amphibolites II	III	IV	Average V	Green sch. VI	Birtavarre VII	Finland VIII
SiO_2	48.8	50.0	49.0	49.4	49.3	48.8	48.5	49.4
TiO_2	2.2	1.8	1.2	1.8	1.6	1.2	1.2	1.7
Al_2O_3	14.0	14.2	14.4	13.3	14.0	14.6	17.3	14.1
Fe_2O_3	3.6	3.0	2.7	3.4	3.0	2.2	1.3	2.3
FeO	9.8	10.7	9.9	10.3	10.3	10.6	7.6	12.0
MnO	0.2	0.3	0.2	0.3	0.3	0.2	0.1	0.2
MgO	6.7	6.0	7.5	6.8	6.8	7.9	7.9	6.1
CaO	9.4	10.0	10.1	9.9	10.0	6.5	11.9	10.5
Na_2O	2.6	2.7	2.8	2.2	2.6	3.1	2.4	2.3
K_2O	0.7	0.4	0.2	0.3	0.3	0.2	0.3	0.6
H_2O	1.8	1.4	1.6	1.5	1.5	4.7	0.9	0.6
P_2O_5	0.3	0.1	0.1	0.2	0.1	0.1	0.3	tr.

I Daly 1933, p. 17, no. 60: Plateau Basalts, average of 43 analyses.
II Mauken, amphibolite no. 613.
III » » » 2447.
IV » » » 2287.
V Average of II, III, IV.
VI Mauken, green schist no. 850.
VII Birtavarre, Troms, metabasics, average of 5 analyses (Vokes 1957, p. 67).
VIII Sodankylä, Finland, amphibolite. (Mikkola 1941, p. 257).

The chemical composition of the amphibolite

Three chemical analyses of the "pure" (biotite-free) amphibolite have been carried out, and also one of the green schist which forms the junction zone between the amphibolite and the overlying meta-sedimentary schists. Chlorite is the chief mineral in the green schist, green hornblende only occurring as relicts. This rock will be treated in greater detail in another connestion (p. 197), so we shall only consider the three amphibolites here. Their analyses are given in Table 2 columns II, III and IV, the average of these being in column V.

There is a striking similarity between the three analyses. None of the components show any significant spread of values. These data should therefore provide a good foundation for a general consideration of the composition of the amphibolites.

The analyses fall well within the normal variation trend for gabbroid minor intrusives and extrusives. To aid comparison the table also contains "the average of 43 analyses of plateau basalts" given by Daly (1933, p. 17, no. 60). The similarity between this average analysis and the other three is quite striking. Only the potassium content differs noticeably, the Mauken analyses having 0.2 - 0.4 % K_2O against Daly's 0.7 %. The potassium content is, anyway, low compared with that normally given for basic magmatic rocks (e.g. Daly 1933, Johannsen 1932). (This is also a well-known feature of the *Caledonian* green schists of Norway.) The silica content of the new analyses is that of a basaltic magma of moderate acidity.

Without drawing any definite conclusions from the few analyses available, I will refer to some analyses from parts of the N Scandinavian Precambrian that are geographically close to the Mauken area.

From Vest-Finnmark, in Norway, all the analyses of "greenstones" (Holmsen, Padget and Pehkonen 1957) and of "diabase-like rocks" (Gjelsvik 1958) are too sodium-rich to bear comparison with the Mauken amphibolites. However, there are some analyses of non-spilitic meta-basites from Swedish and Finnish Lappland, that provide more suitable comparative material. Geijer (1931, p. 179, no. 1) gives an analysis of an "effusive diabase" from the Kiruna area. This contains the same amount of SiO_2 (49.1 %) as the Mauken specimens, but some more Al_2O_3 (15.7 %). None of the other components differs significantly from the Mauken amphibolites. The potassium values in several of Geijer's Precambrian meta-basites are rather low, similar in value to those of Mauken.

From Finland there is an analysis of an "amphibolite" from Sodankylä (Mikkola 1941, p. 257), which is the nearest of all in composition to the Mauken amphibolites. It is reproduced in Table 2. The only differences it shows from the amphibolites is in the F_2O_3/FeO relationship and in the somewhat higher potassium content.

Finally it should be mentioned that the *Caledonian* meta-basites from Birtavarre in Troms (Padget 1955, Vokes 1957) seem to differ markedly in chemical composition from the Precambrian amphibolites described above. Five analysed specimens from Birtavarre all contain significantly more aluminium and less iron. The average of these analyses is given in Table 2.

The quartz dioritic rock within the amphibolite

As described previously, this quartz diorite occurs as sheets or sills of light grey rock, concordantly interbedded here and there in the amphibolite. The rock is frequently translucent on fracture edges. It has no hornblende, and contains only small or completely insignificant amounts of biotite. Muscovite is absent. Plagioclase is the only feldspar (An 15-17 %). In the types poorest in biotite there is only a very faint foliation, but no schistosity.

Three specimens from different localities can be described in more detail.

Spec. 2503 a. From Andsfjell. Micro.: Plagioclase and quartz are the chief minerals, and have a fairly uniform grain size (0.05 - 0.07 mm). The quartz occurs as relatively equidimensional grains, without significantly undulating extinction. The feldspar forms the matrix between the quartz grains. Twin lamellæ are seldom. An = 15 %. Small slivers of dirty brown biotite and small lathes of zoisite, up to 0.02 mm long, occur in very insignificant amounts. The zoisite is always in the feldspar.

Spec. 2446 a. From Olsborg. Micro.: Mostly identical to the previous specimen. The quartz content is about 2/3 that of the plagioclase. The plagioclase is An 17 %.

Spec. 1674 a. From Skjold, about 2 km SE of the SE corner of the "Målselv" sheet. Micro.: The minerals are mostly the same as in the previous specimens, but a little coarser. Quartz and feldspar measure about 0.15 mm, and the biotite is also correspondingly larger. Frequent twin lamellæ are seen in the feldspar. An = 17 %. Small sphene crystals are scattered throughout the section. A few tiny grains of apatite and calcite are seen.

A chemical analysis of specimen 1674 a is given in Table 3 (p. 186). The norm of the analysis, with Q = 30.6, Or = 4.0, Ab = 51.5 and An = 10.0, can be assumed to correspond very well with the mode of the rock. Calculated from the normative amounts of Ab and An the plagioclase has a composition of $\frac{10.0 \times 100}{61.5} = 16.3$ % anorthite.

The composition of the rock corresponds very closely to some of Goldschmidt's trondhjemites (Goldschmidt 1916) from the Norwegian mountain chain. Strand (1958, p. 126) refers to sheets in the green schists of Helgeland, Nordland, which seem, from their mineral content, to have a similar composition, too.

The rock is especially characterized by its pronounced leucocratic composition and its very low potassium content. It may best be called a very acid (meta-) leucodacite.

The granodiorite

The granodiorite is exposed in several isolated areas, namely in Andsfjellet, along the W side of Mauken from Moen and southwards for about 12 km*), at Storhaug near the confluence of Målselva and Barduelva, and along the W

*) As a revision of the printed map, this outcrop is connected to the tiny one in the N, just E of Moen, by a narrow zone, and does not stop at Fredriksberg, as the map shows. In addition the structure symbol at (11 - 6) on Andsfjell should show a dip of 25° to SE and not to NW.

slope of Mauken near the S edge of the map. The rock is particularly easily studied in Andsfjellet, itself, and beside nearby Andsvatn. Andsfjellet is almost dome-shaped, with the granodiorite standing out in the central part surrounded by the overlying mica schists in the NW, SE and SW. Tectonic movements have, however, also given the junction of the granodiorite an exceptionally irregular form, with correspondingly irregular branches (see map). The exposures in the S along the Målselv valley have more regular shapes dominated by the NW-SE strike of the rock.

It is possible to distinguish the following variations in this unit:

1. Medium grained, massive rock, with only a weakly developed foliation.
2. Pegmatitic type.
3. Aplitic dykes.
4. Migmatitic-like variants.
5. Coarse grained breccia.
6. Foliated mylonitic gneiss.

The massive granodiorite, type 1, is the most extensive of these variants. It occurs primarily in the central part of the unit, inside the narrow border zone which abuts against the amphibolite and the overlying rocks.

Mineralogically there is very little difference between the various structural types. All contain, though in somewhat varying amounts, quartz, acid plagioclase and microcline as the chief minerals, and only small numbers of dark minerals. The massive granodiorite will be described first.

This rock is generally megascopically rather homogeneous. It has a medium grained, white to light grey groundmass of quartz and feldspar, which is interspersed with very irregularly shaped aggregates of dark minerals. These are often drawn out to lengths of ½ - 1 cm and are about 2 mm in breadth. They are arranged more or less parallel and give the rock a mottled appearance, or one which is more or less streaky according to how far the foliation is developed. The total content of the mafic minerals is always low, but varies somewhat. Subordinate variants that are almost lacking in dark minerals also occur. The foliation is only defined by the parallel orientation of the dark aggregates. In the parts lacking these dark minerals no foliation can be discerned. Whilst the overall impression obtained from a study of these types is of a greyish-looking granodiorite, the microcline is sometimes slightly pinkish so that the colour of the rock is influenced by that. Under the microscope (examples are from thin sections of specimens 546, 794, 795, 797, 798, 2262, 2466 and 2480) the quartz, microcline and plagioclase show a granoblastic texture (Plate II, a). The grain size of these minerals is rather similar, varying usually between about ½ and 1 mm. The quartz is completely clear and shows no undulating extinction. The microcline is also wholly clear and displays beautifully developed lamellæ. It usually contains no perthitic inclusions. The plagioclase often has an almost rectangular shape, and then always has its long axis perpendicular to 010. The shape of the microcline is more irregular. It surrounds or includes numerous grains of plagioclase. The plagioclase shows very well developed albite twins with relatively broad lamellæ. It usually contains some inclusions, frequently in insignificant numbers and rarely in considerable amount. The numbers of inclusions vary from specimen to specimen. Plagioclase grains that are almost completely free of inclusions are seen. The inclusions mostly are small flakes of sericite, but also very small, mostly rod-shaped, crystals occur which are clearly mainly zoisite. The anorthite content of the plagioclase is the same in all specimens examined, namely only 3 - 5 %. It is hence a rather acid albite.

The proportions of microcline and albite vary a good deal from one specimen to another (as the analyses also indicate), and are not easy to determine with any exactitude, even under the microscope, as the distribution within any one thin section is rather uneven. The main type, however, contains a good deal more albite than microcline, though specimens occur where the two feldspars are more or less equal in amount. Quartz and feldspar usually lack all evidence of cataclasis.

The dark minerals consist of irregular intergrowths of biotite, hornblende, epidote and iron ore. The strongly birefringent epidote is the most common. Afterwards comes biotite, as dirty greenish-yellow, strongly pleochroic flakes. The hornblende is also deep green and strongly pleochroic. It occurs least, and is an unstable mineral.

Two analyses of the rock, both from Andsfjell, are given in Table 3, p. 186. The analyses indicate granodioritic or granitic rocks with a high percentage of silica and a relatively low aluminium content. The calcium content is low. The sodium/potassium relationship is very different in the two specimens. Specimen 546, with a greater proportion of Na_2O than K_2O, corresponds best to the main variant of the rocks and according to Johannsen (1932) this ought to be defined as a sodaclase-granodiorite.

To try to compare the analyses with some from other Precambrian granodiorites and granites from nearby regions, we may look at Ödman's reference to "7 Karelian migmatite granites" from Norrbotten county, Sweden (Ödman 1957, p. 126). None of these are like the Andsfjell granodiorite. They all

TABLE 3

Analyses of quartz diorite rock and granodiorites from the Mauken area

	Quartz diorite No. 1674a. Wt. %	Cation %	Granodiorites No. 546 Wt. %	Cation %	No. 795 Wt. %	Cation %
SiO_2	72.85	67.7	75.85	71.6	76.23	71.9
TiO_2	0.46	0.3	0.24	0.2	0.08	0.1
Al_2O_3	14.40	15.8	11.23	12.5	12.91	14.3
Fe_2O_3	0.32	0.2	2.73	1.9	0.58	0.4
FeO	1.56	1.1	1.24	1.0	0.75	0.6
MnO	0.01		0.04		0.01	
MgO	0.55	0.8	0.20	0.3	0.13	0.2
CaO	2.71	2.7	1.03	1.0	0.55	0.6
Na_2O	5.68	10.3	4.80	8.8	3.30	6.0
K_2O	0.66	0.8	2.40	2.9	5.00	6.0
H_2O-	0.04		0.04		0.05	
H_2O+	0.17		0.24		0.45	
CO_2	0.18	0.3	—		—	
P_2O_5	0.10		0.03		0.02	
	99.69	100.0	100.07	100.2	100.06	100.1
Q		30.6		35.2		34.8
Or		4.0		13.5		30.0
Ab		51.5		44.0		30.0
An		10.0		2.5		3.0
Σ fem		3.9		4.8		2.2
		100.0		100.0		100.0

contain less SiO_2, and with one exception, more Al_2O_3, and all contain considerably more K_2O than Na_2O. There is an especially large discrepancy between the Andsfjell specimens and the two analyses from the Vassijaure granite from close to the Norwegian border near Torneträsk. However, Strand's (Foslie and Strand 1956, p. 72, no. 1) analysis of a Precambrian "granitic gneiss" from Børgefjell, Nordland, Norway, corresponds fairly closely with specimen 795.

Some of the variants of the granodiorite that differ structurally or texturally from the main type, may now be treated.

Pegmatitic bodies occur at several localities within the basement, though in insignificant amounts. They consist of irregular dykes or lenses. Both pegmatites with pale pink microcline as the dominant feldspar (e.g. no. 798 b. from

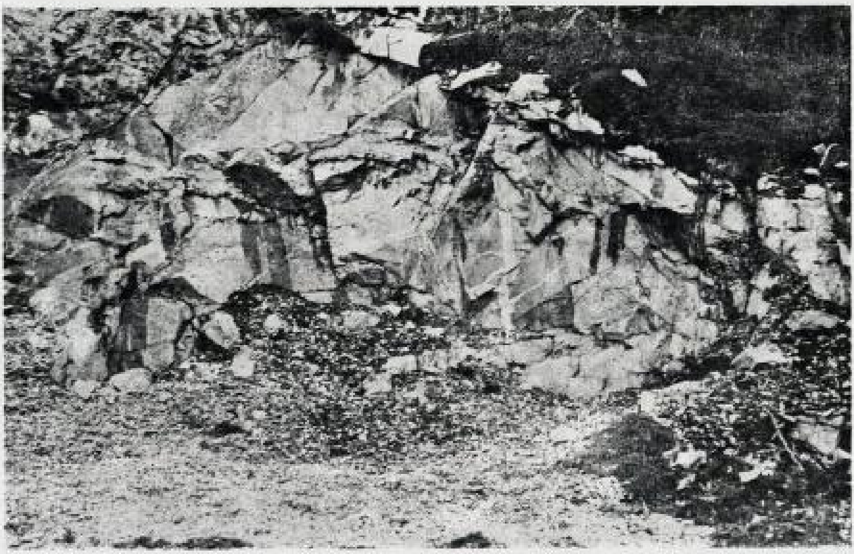

Fig. 4. Migmatitic structure within the granodiorite.

Andsvatn), and types with light grey albite as the main mineral (e.g. no. 2467 from the E side of Andsfjellet), occur. The feldspar crystals have dimensions of ½ - 2 cm. The albite, also in these, has only 3 - 5 % An. The mafic minerals again occur in extremely irregular "clumps", and only in very subordinate amounts. A pegmatite area of relatively large dimension is found on the E side of Andsfjellet. It forms a ridge-shaped area stretching from Bukteholmen (10.5, 8.5) more or less due W up the hillside. The pegmatite is a steeply dipping dyke-net with a strike parallel to the direction of the ridge. In these pegmatites are some beautiful quartz crystals up to several centimetres in size.

About due W of this pegmatite-net the map shows the granodiorite as having a worm-like branch within the otherwise overlying mica schists. This is a quartz segregation along a steep, brecciated fault line. This fault and the pegmatite ridge are probably tectonically associated.

Aplitic dykes, clearly cutting the granodiorite foliation, are found at several localities. Such dykes can, for example, be seen in the quarry near the bend of the main road WSW of the bridge over Målselva, near Buktemoen, Moen. They trend somewhat variously here, but generally with a 40° strike and about 60° dip to N. The dykes are often only about 10 - 20cm across, and are exposed for up to 20 m. The material is light grey to white, or light red, according to which type of feldspar dominates.

Fig. 5. Granodiorite intrusion in amphibolite (seen through water in the floor of a stream).

Migmatitic structures are seen in a few exposures. One is beside the road (13.5, 4.0) on the N side of Storhaugen (Fig. 4). Grey (black on the photograph) flake-shaped agmatitic inclusions are found in the granodioritic material here. These paleosome fragments have the same mineral content as the granodiorite, but have a larger proportion of biotite and hornblende, so that they achieve a definitely granodioritic composition. The agmatites are surrounded partly by light greyish granodioritic material, similar to the main type of the Andsfjell granodiorite, partly by irregular bodies of pegmatitic material, which are themselves cut by aplitic veins and dykes.

The clastic or mylonitic variants of the granodiorite will be described later.

The relationship between the amphibolite and the granodiorite

The contact between the amphibolite and the granodiorite can be observed at many localities. This junction has, in a descriptive sense, a distinct "intrusive" character, with the amphibolite as the older unit and the granodiorite as a younger mobile component. The contact zone, however, has a somewhat different development in the various exposures.

Fig. 5 shows a section of the contact zone in an exposure in a stream about 0.5 km SE of Fredriksberg (12.5, 6.0). Granodioritic dykes are seen cutting the amphibolite. Their dominant trend is nearly perpendicular to the

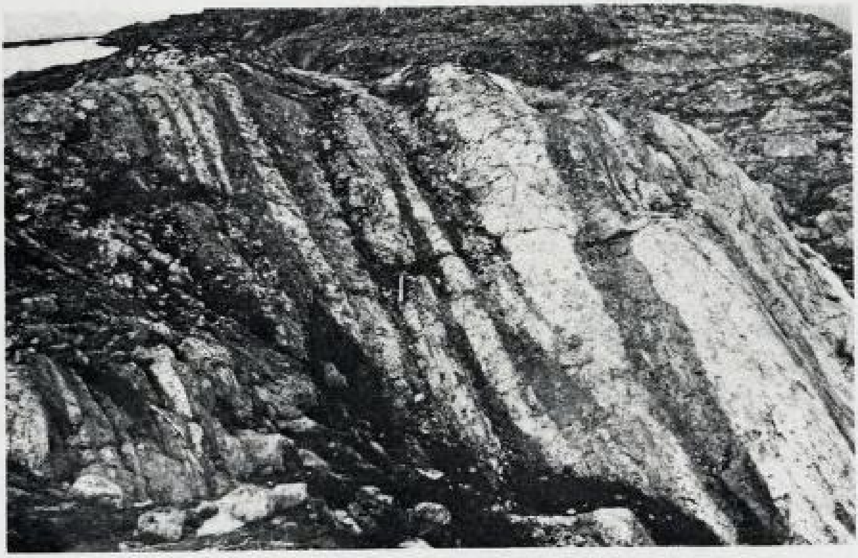

Fig. 6. Granodiorite intruding amphibolite, Andsfjell.

steep junction which itself trends NW-SE (the massive granodiorite, itself, is a few metres below the area shown in the photograph). Near the bottom of the photograph fragments of amphibolite can be seen in the granodioritic material. The surrounding fine grained amphibolite, here, displays no marked effects of metasomatism.

A characteristic granitization zone in the amphibolite is exposed over a large part of the summit plateau of Andsfjellet. The bedrock consists of a continuous layered alternation of the melanocratic host-rock and light granitic material. The layers have a regular strike of 155°, and a very steep dip. The thickness of both the melanocratic parts and the granitic horizons varies from about 10 cm up to several metres. The regular parallel structure can, at first glance, invite some uncertainty as to whether the process has been granitic intrusion of a basic rock, or the opposite. However, granitic apophyses can be observed crossing basic layers in several places, whilst the basic layers are never seen to cut granitic horizons. It can be observed, too, that the basic layers show a diffuse disintegration within the granitic material (Fig. 6, right-hand part).

The basic layers have been somewhat more deeply eroded than the granitic material has, and this tends to strengthen the false impression that there has been intrusion of basic material (Fig. 7).

Fig. 7. Dark bands of amphibolite in granodioritic material, Andsfjell.

The granitic material largely consists of white or pale pink aplite (e.g. specs. 2371 and 2504), but there is some coarser grained material, too. The quartz-diorite-aplite of the amphibolite occurs here, too, and can easily be mistaken for this, but it has a greyer, slightly glassy appearance.

The melanocratic layers are greyish rocks of dioritic appearance, and have only a faint foliation which parallels the strike of the layers. All the specimens are nearly equigranular, with the grain size varying in different layers from 1 - 2 mm and down to about 0.3 mm. Acid plagioclase is the main component, with varying amounts of hornblende and biotite as the dark minerals.

Microscopically, specimen 2503 from these beds, shows a pronounced granoblastic texture. About ⅔ of the thin section reveals plagioclase, with a grain size of around 0.3 mm and extremely irregular borders. It contains many inclusions, mostly of epidote and small muscovite flakes, but also of all the other minerals elsewhere in the thin section. Twin lamellæ cannot be seen, but cleavage cracks are visible in most grains. The refractive index, determined by immersion, indicates 5 % An, so it is a considerably more acid plagioclase than that of the amphibolites. Much quartz is present. Biotite is the major dark mineral, about 20 %, whilst hornblende is relatively rare. Both minerals are seen in thin section as sub-parallely orientated crystals, and are strongly pleochroic. The biotite is brown, with a slight greenish tinge which distinguishes it somewhat from the pure brown biotite of the amphibolites. The biotite and hornblende are often intergrown and the latter also occurs as inclusions in the former. The borders between the two minerals are often diffuse, and the hornblende gives the overall impression of being unstable. Scattered throughout the section are a few flakes of muscovite, up to 0.2 mm in size. Epidote occurs fairly commonly, and ore (magnetite?) in small amounts.

If we then pass on to the granodioritic area further up the Målselv valley, we find a type of granitization that has led chiefly to products with a somewhat different texture. This can easily be studied beside the main road, near Trongen (25.0, 0.5), and on the other side of the river, E of Alappmoen, or by the main road near Skjold, about 15 km further S along the valley (outside the area shown on the map). Coarse grained or pegmatitic intrusions of pinkish granite have, here, transformed the amphibolite to dark, granodioritic types in which the feldspar crystals are of considerably larger dimension than the other minerals, and often form definite porphyroblasts. This applies to the albite as well as the microcline. The microcline, especially, produces well-shaped rectangular cross sections, with dimensions up to ¾ x 1½ cm.

The granite intrusions E of Alappmoen, are primarily made up of sheets parallel to the steep foliation of the amphibolite. The sheets vary in thickness from about 20 cm to several metres. This metasomatically derived granodiorite forms a distinct zone some 10 m or more thick.

Three specimens from this area have been studied under the microscope: spec. 2450 from near the SE corner of the map, spec. 2449 b. from near Alappmoen, and spec. 541, from some kilometres E of the map. The samples are uniform with only slight variations in development. Acid plagioclase is always the chief mineral. Microcline occurs in varying amounts, which are difficult to determine exactly. The most microcline-rich sample is no. 541, in which this mineral seems to approach close to plagioclase in amount. The two other samples contain considerably less. Microcline forms definite phenocrysts only in the first of the specimens, otherwise the grain size of all the minerals varies between about 0.5 and 3 mm. The dark minerals vary between about 10 and about 25 % of the whole. Quartz is present in all the specimens. A few quite clear plagioclases with zonal structure are seen, the core having a somewhat higher extinction than the outer zone. In other cases the zoning is produced by a highly sericitic core surrounded by clearer material, or the sericite has developed along a definite zone with clearer plagioclase both in the core and in the outer zone. The refractive index, measured in comparison to quartz and canada balsam, indicates the plagioclase of the outer zones to be a rather pure albite.

Microcline occurs in various different habits. Plagioclase grains are seen in which the microcline occurs only as patchy antiperthitic inclusions. At other points the microcline forms bays in the plagioclase, and there are often only relicts left from the latter mineral. Furthermore, microcline phenocrysts containing inclusions of plagioclase (and of the dark minerals) are seen, though some microcline crystals show an almost complete lack of inclusions. Myrmekite is developed occasionally, but is not very typical.

The dark minerals of specimen 2450 consist of green hornblende and brownish-green biotite in more or less equal amounts. Both are strongly pleochroic and intergrown. Brownish speckled patches are often seen in the hornblende, clearly indicating an embryonic biotitization. No. 2449 b. has slivers of dirty brown biotite as the chief dark mineral. No hornblende is seen, but there is some epidote with a high interference colour. There are also a few allanite crystals, surrounded by borders of epidote. In No. 541 an iron-rich epidote forms about half of the dark minerals. There is also some dirty pale green hornblende with small inclusions of ore and epidote. Biotite is not present. The dark minerals are gathered into clumpy aggregates, like those in the "pure" granodiorite. Some sphene is seen in all the thin sections.

Whilst the contact between the amphibolite and the granodiorite is a very definite zone of granitization, it is also characterized by the tectonic tensions and movements that have taken place. The result of this shows itself to differing extent in different areas. On the summit plateau of Andsfjellet, with its characteristic "intrusive" structure, no conspicuous tectonization was observed. However, we are not at the actual contact of the granodiorite and the more easterly situated massive amphibolite of Mauken, as the amphibolite in Andsfjellet is an isolated patch "floating" in the granodiorite. At the eastern border of the Andsfjell massif, beside the main road to Finnsnes*), near Buktemoen (10.3, 8.8), a sharp contact between the Mauken amphibolite proper and the granodiorite can be observed. In the immediate few metres parallel to the contact, both the granodiorite and the amphibolite display a definite breccia structure, the rocks having been crushed somewhat, though no significant thrusting has taken place. The corresponding contact for the granodiorite of Mauken, SW of Myrefjell, has a decidedly tectonic character, even though the

*) A new road - not shown on the geological map.

intruded dykes (Fig. 5) show only slight signs of this. The contact is, here, accompanied by a zone where both the granodiorite and the amphibolite are brecciated. This zone has a width on the ground of some 10 m, through which the bedrock surface (mostly concealed) is often eroded into small elongated hollows. The contact has a marked brecciated character on either side of the river near Trongen and Alappmoen. Whilst the cataclastic textures connected with the Caledonian overthrusting follow the junction plane between the basement and the thrust cover, the breccia zone parallels the steep NW-SE foliation of the basement.

Concerning the time relationship between the brecciation and the granitization the following can be noted. As shown in Fig. 5, the granitic dykes reveal no sign of post-intrusive movements despite their location within that area of Mauken where the junction is so clearly tectonic. When we move further S, to the granitic dyke material in the Trongen/Alappmoen area, we see that this has a most distinct brecciose character (spec. 2449 a). On the other hand the tectonic effects are only weakly developed in the feldspathized types of this area, specs. 2450, 2449 b, and 541. Megascopically, these show no sign of brecciation, but the quartz is seen to have very marked undulating extinction, and mortar structure is developed at the borders of the feldspar crystals. On the whole, however, the newly formed phenocrysts of plagioclase and microcline show little evidence of crushing, though a few plagioclase grains have bent twin lamellae and some cracks, filled mostly with epidote. All this suggests that the granitization has continued after the major part of the brecciation took place.

The genesis of the granodiorite

Every aspect of the granodiorite and the granite make it natural to consider their formation by a process of granitization. It seems feasible, for all the areas described, to consider that the older, now granitized, rock, once consisted of parts of the neighbouring amphibolite. Paleosome relicts of obviously different rocks are not observed. It seems reasonable, furthermore, to attribute the granitization to a time after (or contemporaneous with) that when the amphibolite layers assumed their present vertical position. Such an interpretation finds support in the fact that the junction towards the amphibolite always runs parallel with the rather uniform NW-SE strike of the amphibolite, instead of having irregular bodies of massive granodiorite penetrating laterally into the amphibolite. The conditions along the junction can be attributed to a diapiric uplift of the granodiorite. The tendency to brecciation that can be observed along the junction zone can be attributed to such an uplift. This movement has affected the granodioritic rock along the contact, though the granitization has continued beyond the main period of uplift.

Granitic rocks also occur immediately beneath the post-Precambrian sediments in the nearby Dividalen district. It seems that this granite is somewhat different from the granites described here, though I can say nothing definite about this at present.

The relationships at the junction between the basement and the overlying metasediments

The contact between the granodiorite and the overlying metasediments

More or less metamorphosed sediments overlie the Mauken amphibolite complex and the Andsfjell granodiorite. These meta-sediments vary a good deal, including low metamorphic chlorite schists, muscovite-biotite schists, quartz-rich schists and quartzites. These will be dealt with in greater detail in a later paper - including a discussion on how far they represent par-autochthonous Cambrian, Eocambrian or Precambrian deposits or are parts of long-transported thrust nappes.

In the two nearby basement windows, at Bardu in the S and Dividalen in the SE, the relationships near the contact of the Precambrian gneisses are more simple than is the case here, as we there have relatively unmetamorphosed sediments - shales and sandstones - resting on the basement surfaces. In the Mauken-Andsfjell region all the immediately overlying sediments are more strongly tectonized and recrystallized.

In addition, the granodioritic and amphibolitic rocks which abut against the meta-sediments display special tectonic and metamorphic characteristics. The granodiorite is transformed into streaky and banded gneisses. Instead of amphibolites we get green schists in which hornblende is unstable and chlorite is the prevailing mineral. The junction, thus also symbolizes a sharp break in the metamorphic facies.

The following can be said regarding the structural relationships at the junction. Neglecting an area extremely close to the junction plane, there is a distinct angular disconformity between the foliation in the basement and that in the overlying schists. Whilst the former has a NW-SE strike with a steep dip, the schists have an E-W or WSW-ENE strike with a moderate northerly dip. The foliation in the schists coincides with the sedimentary layering. A special feature with this discordance is that the foliation of the schists also abuts against the *junction plane* at an acute angle. A schematic picture of this is shown in Fig. 8.

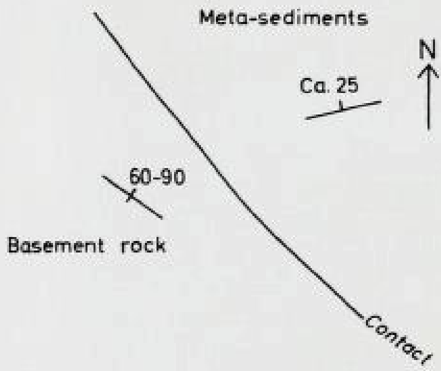

Fig. 8. Schematic sketch of relations at the junction between the basement and the overlying meta-sediments.

This feature is shown in a general way on the 1 : 100,000 map, in the divergent directions of the foliation in the Mauken amphibolite and the meta-sediments along the N side of Takelvdalen (around 15, 19.). In the immediate vicinity of the junction plane, however, the foliations of both the basement and of the schists are parallel to the plane. These relationships will be further explained by reference to some more detailed observations.

The contact between the granodiorite and the overlying meta-sediments

The junction is exposed in a road cutting by the N side of Andsvatn, about midway along the lake (5.5, 4.3). The plane of contact here trends northwards at about 160° with about 40° W dip. Within the immediate 5 m or so of the contact the foliation of both the biotite schists and the granodiorite is in full conformity with the junction plane. West of the contact the mica schist gradually modifies its trend across about 50 m, until it has a stable strike of about 20°, with a dip of about 25° towards the NW. In this transition belt the schist has many small folds with nearly horizontal axes trending at 160°. Eastwards from the contact the trend of the granodiorite foliation changes gradually through some 10 m to a strike of 170°, and a dip of 70° W. The sketch map and profile in Fig. 9 shows this. The granodiorite near the junction is strongly foliated.

The junction near the eastern outlet of Andsvatn (7.0, 4.1) is seen in the road cutting, here, in a 10-15 m high exposure of granodiorite overlain by a couple of metres of biotite schist (Fig. 10).

No discordance can be observed at this point because the exposure only embraces the concordant beds in the vicinity of the contact. The junction plane, near the top of the cutting, has a moderate dip. The granodiorite exposed here reveals a very fine picture of the tectonization suffered by this rock near the contact. Here, the granodiorite as a whole has a banded character conformable to the junction plane, but from the bottom of this gneiss layer to the top, adjacent to the schist, several textural variants reveal the gradual evolution of the mylonitization of the rock. At the foot of the exposure is a type of

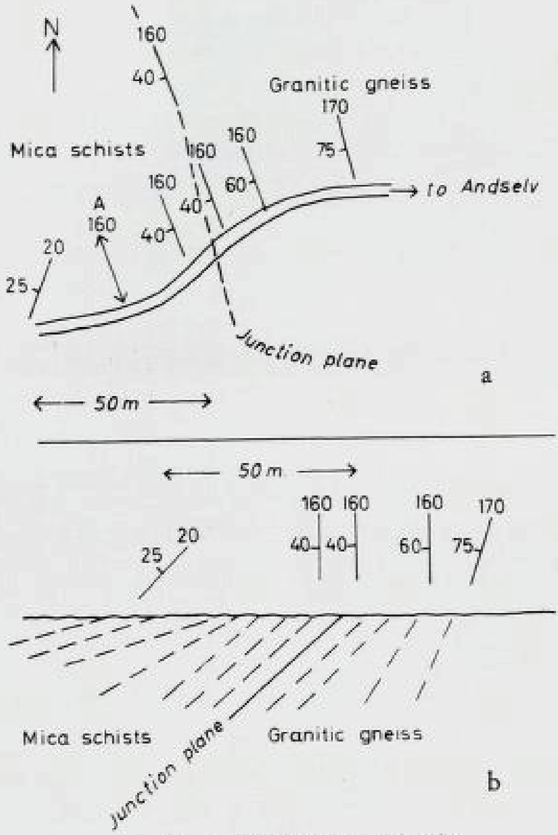

Fig. 9. Plan a) and profile b) of the contact near Andsvatn.

granodiorite not differing particularly from the massive types described previously, in which the dark minerals produce a foliation whilst no real cataclasis is developed. Above this is a distinctly cataclastic, foliated rock, in the lower part with a coarse streaky texture in which the foliation is picked out by mica-covered shear planes that do not occur more tightly than to produce a brecciose texture in the rock. Higher up, the shear planes are closer together and become more parallel to each other. The rock splits easily along these planes. At the top is a megascopically homogeneous, fine grained, light grey, strongly foliated gneiss. Under the microscope this is seen to have the mineral assemblage of the massive granodiorites. Acid plagioclase is the dominant feldspar, with microcline less common, and only very small amounts of dark minerals. The minerals show a very well defined mortar structure. The feldspars, with a grain size of 0.2 - 0.5 mm, are scattered in a fine grained groundmass. The vast majority of these phenocrysts display cataclastic effects. They are broken up, often into separate pieces, and the twin lamellae are often bent (Pl. II b). The biotite schist overlying the gneiss is a highly crushed, compact rock, rich in quartz lenses (specs. 547, 812 and 2263).

Storhaug, near the confluence of Målselva and Barduelva (13.5, 3.5), is an isolated hill, with a smooth slope in the S, but ending towards the N in a steep, terraced cliff going down to the drift covered valley floor about 200 m below. The junction plane lies in a nearly horizontal position in the summit of this hill. Immediately above the plane is a dark, heavily tectonized phyllitic

Fig. 10. Mylonitic gneiss and overlying mica schists near Andsvatn.

schist, cut throughout by numerous glide planes and crumbling easily when hammered. Below this are horizontal layers of mylonitized granodioritic gneisses identical with those at the last locality described. At the foot of the cliff is the exposure of migmatitic granodiorite mentioned earlier (p. 188), apparently showing no trace of the thrusting movements near the summit.

The contact between the amphibolite and the overlying meta-sediments - and their relationships to the green schist

Between the amphibolite complex, as we have termed it previously, and the overlying meta-sediments, we find some specially characteristic features along the border in the NE from Olsborg up through Takelvdalen.

In this stretch a rather extensive belt of green schists forms the junction zone for the amphibolite. Their thickness is only some few metres, but because of the coincidence of the topographical surface and their layering they have a relatively large outcrop. Chlorite is the chief mineral and green hornblende is unstable or is completely destroyed. Selvages of greenish biotite can be seen intergrown with the chlorite in a few specimens. Albite is rather common, and epidotic minerals are ubiquitous. All samples, too, contain significant amounts of sphene, and some ore occurs in a few. Above this schist is a thickness of a few metres of a quartz-bearing, chloritic biotite schist which goes over to a quartz-rich schist, which after a few metres is replaced by a purely quartzitic rock. The lowermost meta-sedimentary layers do not differ greatly in composition from the hornblende-free green schists, and as the colour of the biotite is fairly similar to that of the chlorite and hornblende it is very easy to confuse these green rocks in the field.

We shall consider this feature in more detail at an exposure of the junction that is relatively well suited for that study. This is beside Takelva, about 3 km above Olsborg (13.5, 8.0). There is quite a good exposure beside, and in the river. Along the short NW-SE trending stretch of river the junction zone is crossed over about 15 m. SW of here the amphibolite has a strike of 130 - 140° and a dip to NE of 60 - 70°. To the NE of the junction the quartz schists outcrop with a strike of about 80° and a dip of 25° to N, which is a constant attitude for the Takelva meta-sediments as a whole.

It should be mentioned here that immediately W of this locality there is a strip of limestone within the pure (chlorite-free) amphibolite. Beside the river this layer is only a few metres thick and it wedges out and disappears towards the SE, just beyond the S bank of the river. The limestone is distorted, partly by boudinage, and it must be assumed to represent a tectonic inclusion of a limestone that occurs somewhat higher up in the sedimentary succession.

The circumstances beside the junction are shown in the sketch (Fig. 11). Furthest to the S is a relatively massive, only slightly schistose "greenstone" which reveals under the microscope a tightly intertwined assemblage of green hornblende and green chlorite (spec. 2292). The hornblende is the remains of crystals of 0.2 - 0.3 mm, which are now in process of disintegration and chloritization. The hornblende has a somewhat lighter colour than is normal in the chlorite-free amphibolites. The chlorite is quite deep green, and strongly pleochroic. It has brownish-violet, anomalous interference colours. The crystals often form rosette-shaped aggregates. Feldspar and quartz form an extremely fine grained groundmass of crystals of 0.02 mm size, all of which show strongly undulating extinction.

About 5 m NE of this sample (i.e. 3 m higher in the profile) the green schist has a strike of 100° with a dip of about 50° to N (spec. 2293). This sample contains the same assemblage as the previous one. A few plagioclase

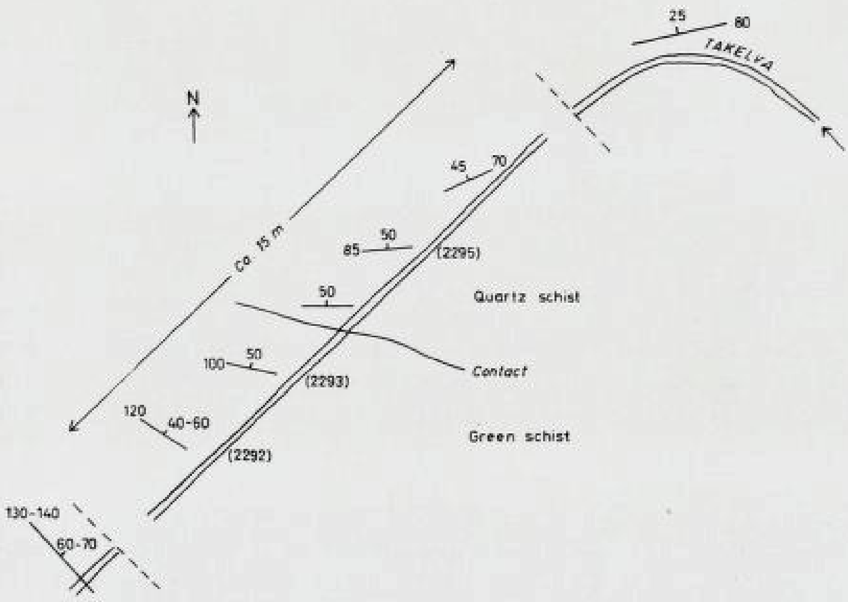

Fig. 11. Plan of the contact at the river, Takelv.

grains of dimensions 0.2 - 0.3 mm are preserved among the fine-grained groundmass, though frequently in a very fractured state.

A couple of metres above this specimen is a small exposure in the river, consisting of a platy, chlorite-bearing schist, which crumbles very easily and it was impossible to obtain a specimen of it. This strikes at 90°, with a dip of about 50° N. The schist resembles the next specimen rather than the one described earlier.

About 2 m above is a greyish biotite schist with a strike of 85 - 90° and a dip of about 60° to N (spec. 2295). This schist has a brownish-green biotite as the chief dark mineral. Intergrown with this biotite, but in much inferior amounts, is a green chlorite similar in appearance to that in the previous specimens. Hornblende is not found. The schist contains considerably more quartz than the previous specimens. Plagioclase is present in considerable quantity, chiefly a crystals of dimensions 0.2 - 0.5 mm. These crystals are often very cataclastic. A few calcite aggregates occur. There is a little ore, but sphene is only present in insignificant amounts. Higher up in the profile this schist clearly goes over gradually, through a few metres, to a feldspathic quartzite, which I named the "lower quartzite zone" on the geological map. Whilst specimen 2293 belongs to the amphibolite complex, specimen 2295, about 4 m higher in the profile, without any doubt belongs to the overlying meta-sediments.

I shall also refer to a couple of other specimens from the green schist of Takelvdalen.

No. 850 is an excellently foliated, phyllonitic schist, chiefly of fine grain (0.05 mm), but with a few larger relict feldspar crystals (0.2 - 0.3 mm). Over half of the thin section is taken up by a light green mesh, mostly of chlorite, but with some (10 - 15 %?) slivers of hornblende. The hornblende has the same pale colour as the chlorite and therefore is not easily distinguishable from it until they are seen under crossed nicols.

No. 623 reveals, under the microscope, highly elongated foliated aggregates of biotite and chlorite. The biotite shows a brownish-green colour parallel to its base, the chlorite is pale grass green. Some rod-like hornblendes occur. In cross section these prove to lie parallel to the micas. Some calcite is seen in the thin section.

No. 614 is also a platy, fine grained green schist. Sphene crystals are for the most part arranged in "stringy" parallel rows about 0.3 - 0.6 mm apart. Green chlorite, as in the previous specimen, occurs in intergrown habit with slivers of brownish-green biotite, and makes up a continuous mesh in the thin section. Hornblende does not seem to be present. One or two grains of calcite are seen.

Some more data on the minerals from these schists may be added.

Plagioclase in specimens 850 and 614 gave, by the immersion method, an anorthite content of 5 % or less. The anorthite content is thus considerably lower than that in the feldspar of the amphibolite which had 12 - 15 % An. The plagioclase in the schists (2295) also has an anorthite content of less than 5 %.

The hornblende rods of specimen 623 gave the following data $\gamma \sim 1.67$ and $\gamma/c = 22 - 23°$. $\gamma - \alpha = 0.02 - 0.025$. It is highly pleochroic. Despite it shape it therefore proves to be a green hornblende and not actinolite.

The refractive index of the chlorite was determined in the green schists (specimens 614, 623 and 850) to 1.625, in specimen 2292 to 1.63. The chlorite is distinctly birefringent, but the optical characters are difficult to determine in the very small flakes. The refractive index places it midway between iron-free and iron-bearing chlorite components.

Concerning the biotite that occurs in the green schists and in the overlying schist (spec. 2295) it may be said that the brownish-green colour distinguishes it from the pure brown biotite which, among elsewhere, occurs in the amphibolite.

However, there is no significant difference in the refractive indexes of the two types of biotite. Both the greenish biotite (spec. 2295) and the brown biotite in the amphibolite specimen 2446b (p. 179) have $\gamma \sim 1.64$.

The meta-sediments (2295) immediately overlying the green schists in this profile, contain significant amounts of albite, as well as biotite and chlorite, of the same types as in the green schists. The quartzites of the "lower quartzite zone", which follow, are also frequently feldspathic and in my opinion represent original arkosic rocks. It is natural to assume that specimen 2295 represents a basal greywacke overlying the amphibolite. This will best explain its transitional character between the green schists and the overlying meta-sediments. It may be thought, too, that the green schists, proper, contain some very little altered debris from the amphibolite surface, but, otherwise it seems very feasible to assume that the green schists represent a part of the amphibolite that has been made schistose by Caledonian tectonism. A chemical analysis of one of the green schists, specimen 850, is given in Table 2, p. 181, no. VI. Its composition differs very little from that of the massive amphibolite, though the content of Al_2O_3 proves to be somewhat higher than in the latter, and K_2O somewhat lower. It is significant that the calcium content is considerably lower, but otherwise the differences from the amphibolites are so small that the analysis indicates that no weathering of the amphibolite has taken place - as would be expected if it referred to a clastic sediment. On the contrary, the analysis supports the theory that the green schists represent a lower metamorphic facies of the amphibolite, perhaps where a certain degree of metasomatism has been effective (the H_2O content of the schist is so much higher than in the amphibolites that this must be taken into account in a more detailed consideration of the values of the analyses).

A profile NE-SW from Takelvdalen up towards Humpen (15.0, 6.4). somewhat further S than the last observations, shows the following features. Beside the farm, Skogvang, furthest in the NE is green schist with a strike of 90° and a dip of 25° N. In the slope above, the green schist has a strike of 110° with 20-25° N dip. At a height over the valley floor of 200 m the green schist strike has swung to 130°, with a dip to N of about 30°. Further up the slope the dip increases gradually through 40 to 60° and finally becomes vertical. Then at about 300 m over the valley floor comes the amphibolite, with a strike of 140° and a vertical dip. At the contact the amphibolite is a dark mylonitic rock some metres thick.

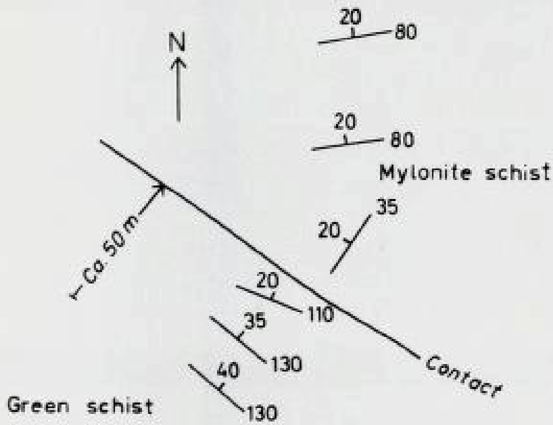

Fig. 12. Plan of the contact at Rundhaug.

A profile from the slope above Rundhaug in Målselvdalen (20.5, 1.2) shows the following features. Up to the junction, at about 150 m over the valley floor, are compact, massive green schists. The schistosity is not always easy to identify, but where it is conspicuous it has a strike of 120 - 130°, and a dip of 20 - 40° to N. The schists are frequently cut by millimetre-broad quartz veins. These mainly are parallel to the foliation but also branch into a network of veins.

The green schists, 5 - 10 m below the contact, have a strike of 110° and a dip of 20° N. Immediately above the contact the meta-sedimentary schists are distorted with strikes varying from 20 to 35 and 40° and with a dip of 20° to N. A few metres above this, the schists are more constant with strike 80° and dip 20° N. Fig. 12 shows all this.

The sedimentary rocks immediately above the green schists are very tectonized sericite-chlorite schists (phyllite). It seems very likely that these could represent a metamorphosed variant of the shales from the Hyolithus zone, as it is seen in Dividalen.

Discussion and Conclusions

The term amphibolite group is used to describe the pure amphibolites, their streaky horizons of varying composition, and the somewhat thicker layers of quartz diorite or leucodacite.

The pure amphibolite consists mostly of green hornblende. It has no pyroxene or chlorite. The plagioclase has about 15 % An. Epidote and zoisite occur only in small amounts. Both the petrography and the chemical composition suggest that the rock is an altered effusive of basaltic character. Its composition is particularly close to certain average analyses of plateau basalts indicated in the literature. Assumed relict variolitic structures are observed, but the rock is otherwise so completely recrystallized that traces of primary structures are mostly destroyed. No examples of pillow structures have been observed.

Even though everything points to the rock being an effusive, the possibility of its representing a highly metamorphosed and tectonised plutonic rock cannot be rejected. The shape of the outcrop, for example, allows this possibility. I shall, however, discuss this rock in the basis of its being an effusive.

The light, streaky horizons in the amphibolite are distinguished from this by having little or no hornblende. In some of them biotite is the chief mineral, in others muscovite and zoisite are most important. Muscovite and hornblende do not occur together.

A similar streaky banding in amphibolites is not an unusual occurrence, and several examples are described from the Caledonian nappes. Foslie (1941, p. 64) refers to rocks from Tysfjord, Nordland, which, according to the mineral assemblages he reports, seem identical with the ones here. Kulling (1955, p. 168) gives a comprehensive account of banded amphibolites with similar mineralogical compositions, from Västerbotten in Sweden, and has a photograph which could easily be from Mauken. I have observed a similar structure in an amphibolite at Mosjøen, Nordland. The three interpretations for this banding that seem particularly appropriate are that they derive from:

1. an alternation of tuffitic material and lava flows,
2. detrial sedimentary material deposited in alternation with lava or tuff,
3. metamorphic differentiation.

Kulling, somewhat tentatively, considered the last to be most likely. Foslie also reached no definite conclusion, allowing the possibilities of tuffitic or detrital sedimentation to remain open. Strand (Foslie and Strand, 1956) describes banded greenstones from the Namsen area, in Trøndelag, in which he believes the banding to be due to metamorphic differentiation, whilst from Helgeland, Nordland (1958), he describes banded green schists which he ascribes to an alternating development of volcanic and sedimentary material.

In the present case I find it most probable that the bands derive from tuffitic deposits alternating with lava flows. The homogeneous composition of the amphibolites, indicated by the chemical analyses, suggests the improbability of any significant metamorphic differentiation. Nor does the composition of the light streaks point directly to a sedimentary origin.

I also find it reasonable to interpret the very fine grained quartz dioritic layers in the amphibolite as tuffites, even though the texture of the rock gives no real cause for this. They bear a certain megascopic similarity to the Andsfjell aplites, but differ mineralogically.

Three analyses of pure amphibolite, taken from a transverse profile, are available. The analyses agree closely with each other. They are used for comparing the amphibolites with metamorphosed basic effusives from the Precambrian of Norwegian, Swedish and Finnish Lappland. The few analyses available from the Lappland Precambrian show a noticeable - in one case striking - similarity to the Mauken amphibolites. The Mauken amphibolites seem to have a clearly different composition from the *Caledonian* metabasalts in Birtavarre, N Norway.

Green schists — Near their junction against the overlying meta-sediments, the amphibolites go over into a relatively thin zone of chlorite-bearing schists containing only a little hornblende, if any. An analysis of these green schists shows that they have a very similar composition to the amphibolites. I believe it most reasonable to consider these green schists as a border facies of the amphibolites, which have undergone retrograde metamorphism during the Caledonian thrust movements. On the other hand the green schists seem to be connected, by a degree of transition, to the overlying meta-sediments of assumed Eocambrian-Silurian age. The reason for this is not fully clear. It is also possible that the green schists partly represent a thin bed of altered debris (a basal greywacke) that has overlain the Precambrian amphibolite surface.

The granodiorite forms the next largest unit in the area. According to its feldspar content it varies from granodioritic to granitic. Most of it seems to have a granodioritic composition, and it is that collective term that is used on the map and in the text. The chief variant of these rocks is medium to fine grained, but aplites and pegmatites also occur. I believe the granodiorite has been formed by a granitization process, for which - in the present area - the Mauken amphibolites apparently represent the host rock. The granitisation seems to have taken place during a diapiric uplift movement.

The granodiorite is distinctly intrusive in its relation to the amphibolites. Breccia structure is, however, also seen at several points along the junctions, thus showing that movements must have taken place at a relatively late stage in the granitisation. However, a few of the bodies, intrusive into the amphibolite, are little affected by cataclasis, thus indicating that these are of somewhat younger date.

The conditions along the junctions between the basement and the overlying meta-sediments — The schists above the basement lie with a distinct angular disconformity on both the granodiorite and the amphibolite. They show, furthermore, not only disconformity towards the foliation of the amphibolite and granodiorite, but also towards the intermediate junction plane. Only beside the contact, itself, have both the basement and the schists a foliation that is expected within the overlying sediments, but these do not seem to occur;

It seems possible that the massif has been uplifted in Caledonian times, after the overlying meta-sediments had assumed their present position, or simultaneous with that. A fairly gentle bulging-up of the surface is a preferable mechanism to an uplift "en bloc". Gustavson (1963) assumes such a possibility for the basement windows in Dividalen. The many vertical minor faults, and not least the very steep shear planes of the amphibolite, make such an interpretation probable. Cataclastic phenomena are a normal occurrence in the otherwise massive amphibolite, and indicate that shearing took place not only during the formation of the hornblende, but also continued afterwards. The existing parallelism between the structural trends of the basement and the overlying schists, just beside the contact, must be interpreted as being derived from thrust movements across an extremely uneven surface. But this conformity may also be due to a pressure from below, brought about by an uplifting massif, after the schists had been emplaced above. However, in this case we should expect to find extensive chloritisation of the hornblende in the central part of the massif, something which does not in fact occur. I will, therefore, refrain from taking up any definite standpoint in this question.

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conformable to the junction plane. All the surrounding rocks near the junction are more or less cataclastic, mylonitised and schistose. The junction must therefore be considered as a tectonic thrust plane. The overlying meta-sediments clearly consist of Caledonian nappes, though remains of par-autochthonous Cambrian (and Eocambrian?) sediments are also present. The rocks in the immediate vicinity of the junction have been imprinted with a coincidental foliation.

It should, however, be noted that, whilst the NW-SE structural trend of the basement is discordant to the WSW-ENE strike of the overlying schists, this first direction is nevertheless a well-known structural feature within the Caledonian nappes (the "cross-folding"). This is especially apparent in the nappes which make up the greater part of the map-sheets "Målselv" and "Tromsø", where there are synform and antiform systems of very similar scale to what we have in the Mauken area. Any possible connection between these will not be discussed in the present paper.

The structure of the basement — The Mauken amphibolite complex has a steep NW-SE foliation in its central part, and has a dip to NE in its eastern limb and one to SW in its western limb. It is therefore natural to consider the structure as an isoclinal antiform with a very steep axial plane. On the other hand, the uniformity in petrography and structure that the rock shows along a transverse profile, gives no direct support for such an assumption. As an alternative the amphibolite complex could perhaps represent one very steep limb of a large fold whose original shape it is impossible to reconstruct. In the latter case the basaltic layer has had a thickness of a least 3000 m, in the former, at least half that. The research carried out has not produced any more information supporting or contradicting either of these alternatives.

The amphibolite complex forms a ridge-shaped area, the granodiorite of Andsfjell, with its rather dome-like appearance, terminating this ridge in the NW. The Precambrian surface, here, reaches at least 500 m higher than it does in the surrounding region. The sides of this Precambrian massif mostly slope quite steeply (30-50°). The question then arises whether these Precambrian outcrops have been monadnocks on the Precambrian peneplane, or whether they have assumed their present position during later movements. Against the monadnock theory may be noted, among others, the following points:

1. the mountain sides have been relatively steep and the mountains rather high to have belonged to a peneplane;

2. with such steep mountain slopes, traces of coarse debris (talus) could be

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Stratigraphical consequences of the discovery of Silurian fossils on Magerøy, the island of North Cape

By

Sven Føyn.

Contents

Abstract

Crinoid stems were discovered east of Nordvågen (east of Honningsvåg) on Magerøy in 1959. In 1960 an additional find of fossils proved a Silurian age for the limestone and conglomerate beds there. Two years later the author found crinoid stems in similar rocks at Sardnes (west of Honningsvåg). In this paper the author elucidates the present state of the age problem as regards the sedimentation of (later metamorphosed) rocks in the central area of Magerøy, especially the tillite-like conglomerate at Duksfjord. A Silurian age for this rock is, with the present state of knowledge, a more reasonable one than the Eocambrian which was formerly presumed. The alteration of the view of the age implies that the glacial origin is now to be controverted.

The existence of the fossils proves that the intrusions of gabbro on Magerøy are not older than Silurian age.

An unpublished report with a map prepared by Mr. J. J. C. Geul in 1958, has been the basis for the planning of the excursions and a guide in the field.

Introduction

By the middle of the last century the geology of Magerøy was fairly well known, taking into consideration the general knowledge of the geology of Norway. *Gæa Norvegica* contains a description and a map (Keilhau 1850, p. 255 and 270, and Pl. V. Fig. 1), compiled on the basis of investigations by L. v. Buch (1810), R. Everest (1829) and Keilhau himself. During the next hundred years, contributions to the geological literature on Magerøy seem to be confined to only two papers, based on short visits by Reusch (1924) and O. Holtedahl (1944).

Holtedahl in particular describes the occurrence of a tillite-like conglomerate at Duksfjord in the north-eastern part of the island. The conglomerate shows no stratification. It is metamorphosed, the fabric of the matrix being crystalloblastic, but almost no schistosity appears. The rock fragments are mostly angular, consisting in the main of fine-grained limestone or dolomitic limestone, but Holtedahl also reports the presence of quartzite among the fragments. The tillite-like conglomerate rests on a series of garnet-bearing mica schists, quartzites, bedded conglomerates and crystalline carbonate rocks (limestones and dolomites).

The thought naturally suggested itself that the tillite-like conglomerate might be a matamorphosed equivalent of one of the Eocambrian tillites of Finnmark. Consequently, the adjoining rocks also had to be looked upon as parts of the latest Precambrian and Eocambrian sedimentary suite of Finnmark. The term Eocambrian is, in this paper, applied in the "restricted" sense, viz. from the time starting with the deposition of the lower tillite in Finnmark. When I visited the locality in 1957, 1958 and 1959, I fully agreed with this view of Holtedahl.

In the south-eastern part of Magerøy the metamorphism is of a lower grade, the argillites being phyllites and not mica schists. On the geological map of Norway of 1953 (Holtedahl and Dons), these rocks were designated as probably Precambrian, based on a lithological similarity to a part of the Raipas suite in the Alta district.

The mapping in 1958 by J. J. C. Geul

In 1958 J. J. C. Geul, a Dutch post-graduate student who wished to work in the Caledonides for a time, was employed by Norges Geologiske Undersøkelse (Geological Survey of Norway) for the summer season. The director of the Survey gave him the task of investigating the interesting, differentiated gabbro complex near to Honningsvåg on Magerøy, and of mapping the surroundings, including the sedimentary rocks of the Duksfjord area and the

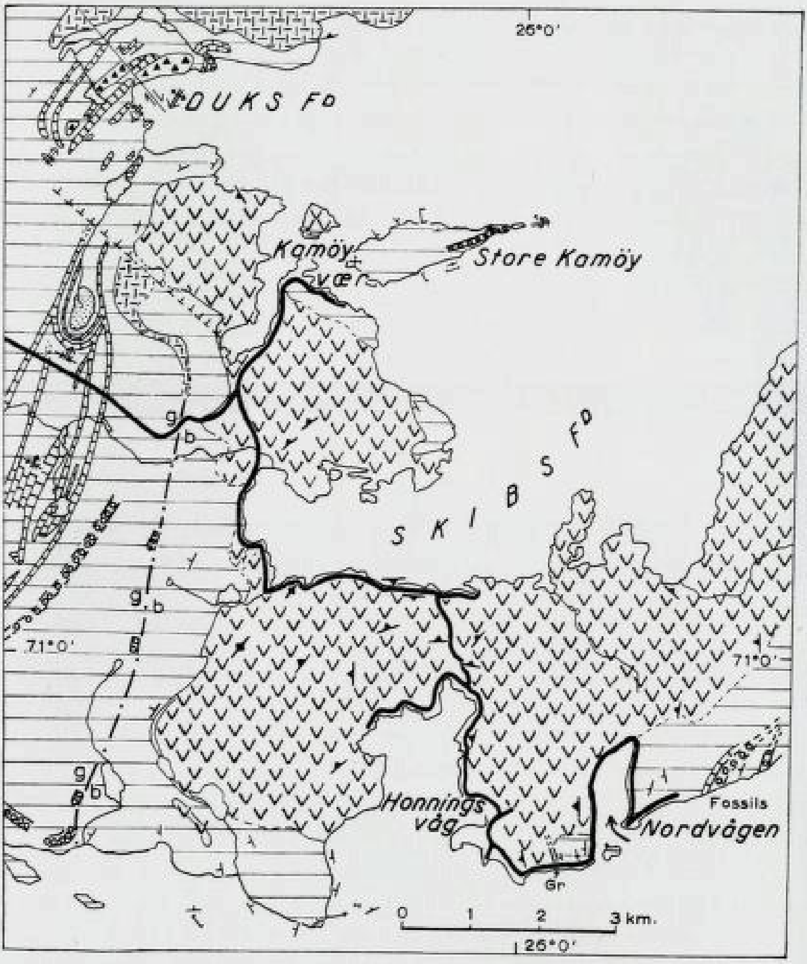
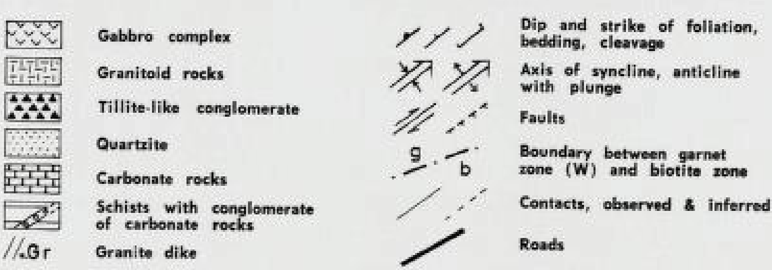


Fig. 1.
Geological map of south-eastern Magerøy by J. J. C. Geul.



North Cape plateau. Geul's report represents a very valuable contribution to the knowledge of the geology of the eastern part of Magerøy. An abstract of the report and a simplified map were published in the Guide to excursion no. A 3, International Geological Congress, Norden 1960 (Reitan 1960, pp. 55-57 and Fig. 12). The abstract mainly gives a review of the petrology of four types of gabbro and of the more important sedimentary rocks. The map, Fig. 1 in this paper, has been reproduced after that in the Guide-book. Note that hornfelses, with more or less sedimentary relics, are included in the "Gabbro complex". To the abstract in the Guide-book some important conclusions from Geul's report concerning the age problems are added here:

1. The sedimentary rocks of the south-eastern part of Magerøy differ only in the degree of metamorphism from those of the central part and the Duksfjord area. Accepting an Eocambrian age for the Duksfjord "tillite", he thus dismissed a Precambrian (Raipas) age for the south-eastern sedimentary rocks.
2. The intrusions of the gabbro effected a contact metamorphism of the adjacent rocks.
3. The gabbro shows no signs of intense shearing. Most probably the intrusions belong to a late phase of the Caledonian orogeny.

Further, Geul's map and report show that intraformational conglomerates and carbonate rocks occur east of Nordvågen and are common in the area between Duksfjord and Sardnes (which is a locality at the south-western corner of the map Fig. 1 - the name Sardnes is omitted).

When, in 1958, I accompanied Mr. Geul in the field for a couple of days, he especially called my attention to this similarity between conglomerates east of Nordvågen and at Sardnes, both containing pebbles mainly of limestone or dolomite, and having a carbonate-rich matrix.

The discovery of the fossils

In 1959, Holtedahl, Reitan and I visited Magerøy, in connection with the planning of the international excursion no. A 3 of 1960. Using Geul's map we went along the shore to the westernmost of the two conglomerate localities east of Nordvågen. In an impure limestone close to the conglomerate we discovered crinoid stems (see Guide to excursion no. A 3, 1960, p. 57, and Strand 1960, p. 165).

In 1960, G. Henningsmoen and F. Nicolaisen collected more crinoid stems from the limestone bed, and also from limestone pebbles in the conglomerate. In addition they found a few poorly preserved, straight monograptids in a shale slab. During the Congress excursion the same year one of the excursion

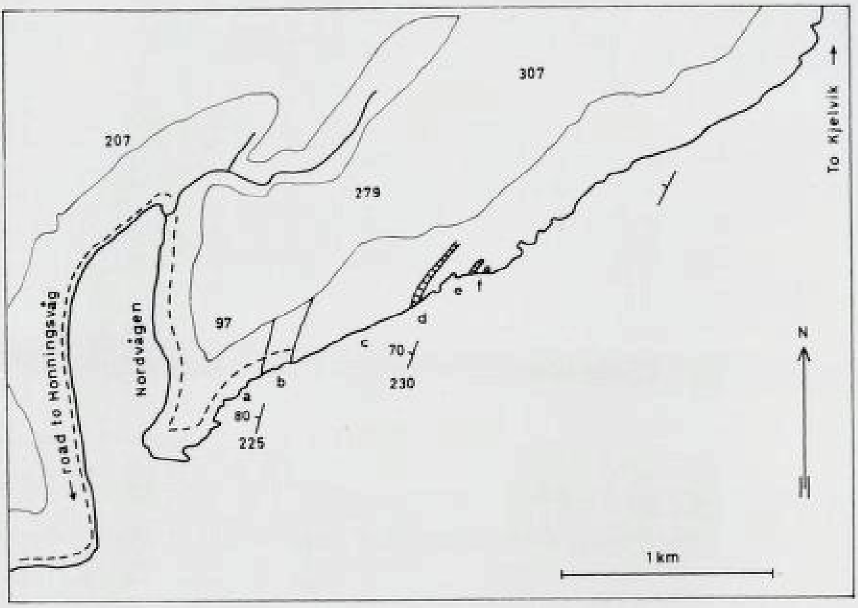


Fig. 2. Sketch map of the coast of Magerøy between Nordvågen and Kjelvik, drawn from air photos. 207, 97, 279, 307 are heights in metres above sea-level. The thin line indicates the edge of the plateau. The figures of strike and dip are based on the 400° scale. a: fine-grained, grey coloured arenaceous beds of 20 - 30 cm thickness and dark banded argillites. b: thin-splitting black rusty phyllite. c: dark grey slates. d: limestone and conglomerate with crinoid stems and sandstone. e: dark grey slates. f: limestone and limy sandstones with fossils. Further to the east: Mainly dark grey slates with intercalations of arenaceous beds.

Kartskisse over kysten av Magerøy mellom Nordvågen og Kjelvik, tegnet etter flyfotografi. 207, 97, 297, 307 angir høyde over havet i meter. Den tynne linjen antyder randen av platået. Strøk og fall er oppgitt i grader basert på kompass med 400° inndeling. a: finkornete, grå, 20 - 30 cm tykke sandsteinslag i veksling med mørke leirsteinsbergarter. b: tynnspaltende mørk fyllitt med rustbelegg. c: mørke grå skifre. d: kalkstein og konglomerat med enkrinitt-stilker (sjøliljestilker) og sandstein. e: mørke grå skifre. f: kalkstein og kalkholdig sandstein med fossiler. Videre østover: hovedsakelig mørke grå skifre med innleiringer av sandige lag.

members, Dr. John Rodgers, found at the eastern locality (se maps Fig. 1 and 2), chain corals, crinoid stems and pentamerids, without doubt of Silurian age (Henningsmoen, 1961). The discovery of this fauna implies that the other sedimentary rocks east of Nordvågen most probably also have to be ascribed to the Silurian period. Moreover, it makes it necessary to reconsider the age problem of the rocks in the central part of Magerøy, among them the Duksfjord tillite. As mentioned above, in Geul's opinion the rocks east of Nordvågen and those of the Duksfjord-Sardnes area are stratigraphically related.

Fig 3. The "first" fossil locality east of Nordvågen (d on the map Fig. 2). Text see p. 214.

Den vestlige av de to fosillokalitetene øst for Nordvågen (d på fig. 2). Fra venstre mot høyre: Mørk grå leirsteinsbergart med tynne lag av lys sandstein, 2 m kalkstein, 3 m sandstein med konglomeratbånd, 10 - 15 m grovt konglomerat med linser av sandstein. Lagstillingen er invertert (overkippet), oppad i lagrekken er mot høyre.

In order to try to contribute to the solving of this problem, I visited Magerøy for a few days in 1962 and 1966. In addition to investigations at the two fossil localities east of Nordvågen (where more fossils were collected) and a reconnaissance trip along the shore to Kjelvik (about 3 kilometres further to the east), I went (by boat) to Sardnes and also traversed from Sardnes to the "North Cape-road".

The fossil localities east of Nordvågen

The main features of the geology along the shore east of Nordvågen appear in Fig. 2. All the beds have apparently about the same strike, N 25° E (based on 400° compass) near to Nordvågen, somewhat more easterly further to the east. The dip is towards the NW, about 80° (vertical = 100°) near to Nordvågen, decreasing a little eastwards along the coast. It is difficult to decide whether or not there is any repetition of beds in the section, although the presence of nearly isoclinal folding must be considered possible or perhaps very probable.

Fig. 4. Conglomerate at the "first" fossil locality east of Nordvågen (d on the map Fig. 2). The beds are inverted.

Konglomerat i den vestlige av de to fossillokalitetene øst for Nordvågen. Lagene er invertert.

The western (in this paper called the "first") fossil locality is situated about 1300 metres east of the populous area of Nordvågen. Details of this locality (d on the map) are seen on the photograhp Fig. 3. From the left (west) the sequence is:

1. Dark grey argillite with thin beds of light coloured sandstone.
2. 2 m limestone.
3. 3 m sandstone with conglomerate bands.
4. 10 - 15 m coarse conglomerate with lenses of sandstone.
5. (Outside the photo) 10 m sandstone.
6. (" " ") Dark grey argillite.

Apparently there is a slight unconformity between 1 and 2. Judging from the tectonic picture of the locality as a whole, I believe that this apparent unconformity is of tectonic, and not of stratigraphical, origin.

The limestone, sandstones and conglomerates are usually light grey in colour, with a yellowish coating. The conglomerate contains pebbles consisting mostly of limestone, but also of sandstone. Sedimentary structures indicate that the sequence youngs to the right (see Fig. 4). This view is supported by the

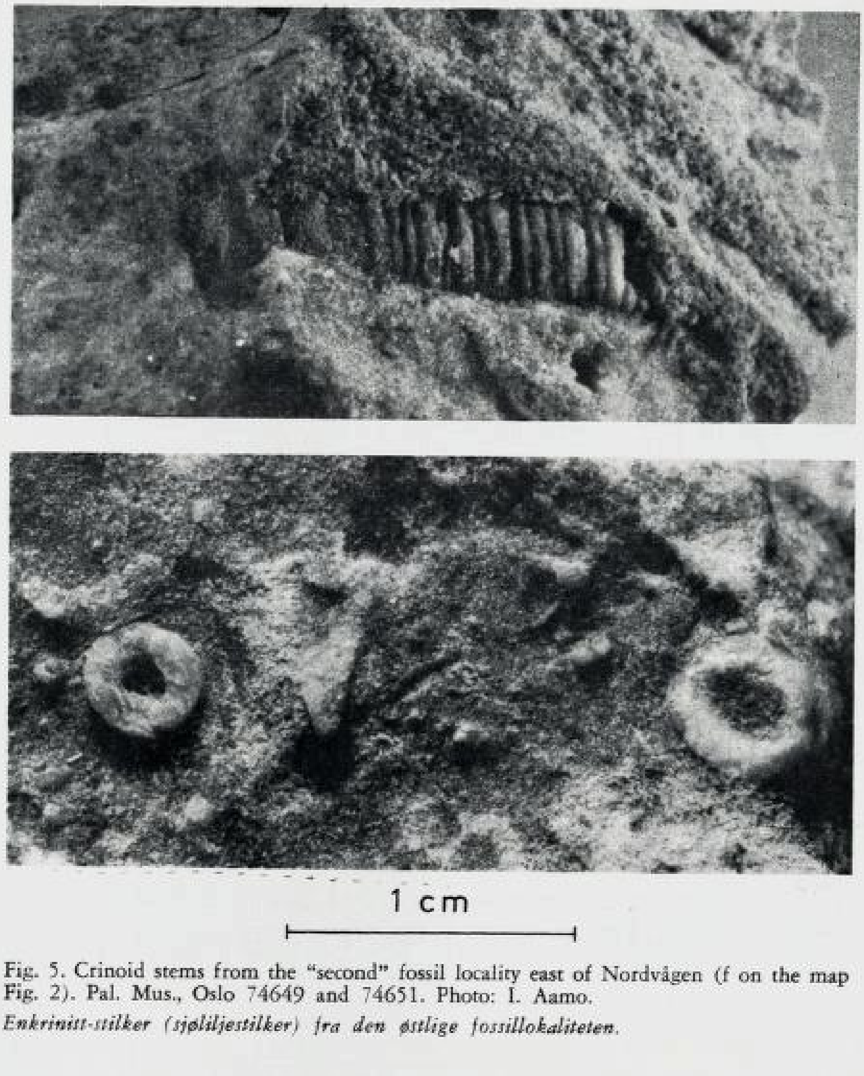


Fig. 5. Crinoid stems from the "second" fossil locality east of Nordvågen (f on the map Fig. 2). Pal. Mus., Oslo 74649 and 74651. Photo: I. Aamo.
Enkrinitt-stilker (sjøliljestilker) fra den østlige fossillokaliteten.

fact that crinoid stems are found not only in the limestone, but also in pebbles of the conglomerate. The beds are thus inverted here. The pebbles have, during the metamorphic processes, generally been somewhat elongated. The bedding planes show a lineation, which plunges 70° to the NNE.

The "second" fossil locality on the map is situated about 350 metres east of the "first". The dark argillites (with sandy beds) grade into a 4 metres thick, bluish-grey, somewhat sheared limestone which contains fragments of

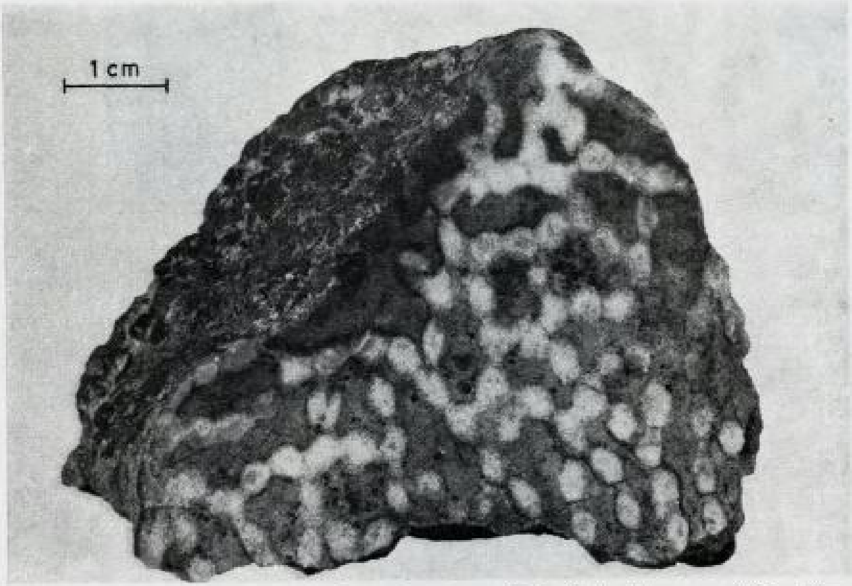


Fig. 6. Chain coral (Catenipora?) from the "second" fossil locality east of Nordvågen (f on the map Fig. 2). Pal. Mus., Oslo 74648. Photos: I. Aamo.

Kjedekorall fra den østlige fossillokaliteten.

crinoid stems and a rugose coral. Then comes (eastwards) a ridge of light yellowish sandstone, 6 metres thick. This rock is partly a limy sandstone or sandy limestone, in which most of the fossil specimens were found (crinoid stems, pentamerids and corals). About 10 metres further to the east, at the shore line, another "body" of fossiliferous limy sandstone or sandy limestone is seen. It is separated from the above-described band by dark argillites.

The fossil-bearing rocks seem to be entirely recrystallized. The fossil fragments consist of calcite grains, while the groundmass contains, in addition to calcite, quartz and a mineral which is probably dolomite.

Figs. 5 - 7 show examples of the collected fossiles.

The existence of a pentamerus species makes an Upper Llandoverian age for the sediment most likely.

The two fossil bearing bodies at the "second" locality are found only near to the shore line, no northward continuation on the mountain slope could be found. Signs of strong tectonic movements are present (joints, quartz lenses and quartz veins). In my opinion, the two bodies are probably parts of one bed. A connection may exist offshore, or the original bed may have been split during the deformation into rods or lenses with a steep orientation, corresponding to a supposed north-eastern plunge of the fold axes.

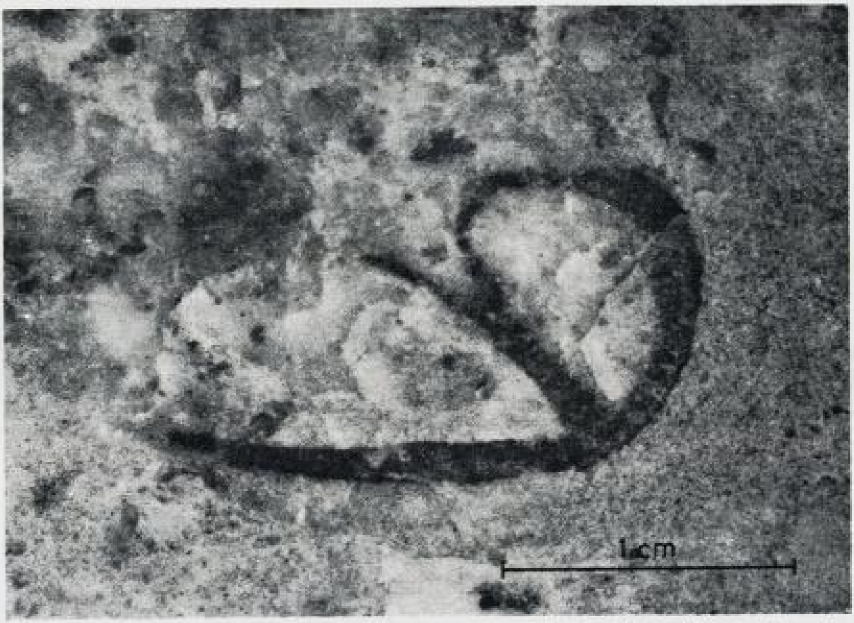


Fig. 7. Pentamerus (?) from the "second" fossil locality east of Nordvågen (f on the map Fig. 2). Pal. Mus., Oslo 74650. Photo: I. Aamo.
Pentamerus (?) fra den østlige fossillokaliteten.

The convergence of the beds of the two fossil localities as drawn by Geul (see map Fig. 1) and indicating an anticline, is not in accordance with my view. The absence of known continuity between the fossil-bearing deposits of the two localities, does not exclude the possibility that they belong to one horizon. Nor does the absence of conglomerate at the second locality, as we know from other parts of the island that beds of conglomerate may vary considerably in thickness, swelling into lenses and then diminishing and even disappearing.

The area between Sardnes and Duksfjord

As already mentioned, Geul in 1958 showed me occurrences of a conglomerate at Sardnes very similar to that at the "first" fossil-bearing locality east of Nordvågen. According to Geul's map, the Sardnes area is the southern part of a large synclinorium which extends towards the NNE to a fault situated south of Duksfjord. The synclinal form west of Duksfjord, of which the tillite-like conglomerate is an integral (upper) part, is apparently a parallel synclinorium. At the inner part of Duksfjord, anticlinal structures seem to dominate.

In the lower parts of the hillsides at Sardnes the rocks are of almost the same metamorphic grade as the rocks east of Nordvågen (e.g. phyllites). On the plateau, west of the line drawn by Geul on the map (Fig. 1), the rocks are of a higher metamorphic grade (garnet- and hornblende-bearing mica schists).

East of the outer part of the bay at Sardnes the beds have a moderate dip towards the NW. Lowermost there are grey sandstones with dark argillaceous bands, then (westwards) black, rusty, fissile phyllites. These two members have a general appearance very like that of the members a and b of the section east of Nordvågen.

Lying above the phyllite horizon, north of the bay and above the scree, these beds are found: 4 m bluish limestone, 0,5 m light coloured sandstone and (uppermost) 15 to 20 m limestone conglomerate (yellowish on the surface). Because of the soil and vegetation on the conglomerate bed, it was not possible to observe its upper boundary.

The limestone appeared to contain small fragments of thin crinoid stems. The preservation of the fossils is poorer than at the localities east of Nordvågen. Because of the crystallinity and schistosity of the rock, they were seen only on weathered rock surfaces. At the foot of the hill several boulders of limestone yield remnants of crinoid stems - which I also found in a pebble in each of two conglomerate boulders.

Outcrops of the limestone band are seen westwards for a distance of about 3 kilometres to the west side of Sardnespollen (the innermost end of the fjord west of Sardnes, off the map), and at several localities I found fragments of crinoid stems. The limestone horizon continues further up the valley onto the plateau, the strike curving gradually through north-west to north. About 2 kilometres from Sardnespollen it thins out and disappears beneath the cover of moraine. From Sardnespollen, the limestone gradually becomes more coarse and granoblastic, and no traces of fossils were found. Also the adjacent argillaceous rock shows traits of an increased degree of metamorphism.

Over the last kilometre the dip of the limestone band is about 35° towards the east. The band thus forms an integral part of the western flank of the large synclinorium on Geul's map (but west of the westernmost bed drawn on the map Fig. 1).

Following the strike further towards the NNE, I observed outcrops of conglomerate in addition to those drawn on Geul's original map. Near to the "North Cape road" the westernmost conglomerates strike towards the west side of Duksfjord, thus suggesting that the Duksfjord syncline may be an integral part of the large synclinorium. Time did not allow me to continue my field trip north of the road, and therefore the significance of the fault drawn by Geul with regard to the general geology between the road and Duksfjord is not known. At the present time it is, therefore, safer to regard the Duksfjord syncline as a fold parallel to the large synclinorium.

Garnet-bearing mica schists are the most common type of rock in the central part of Magerøy. Between Duksfjord and Sardnes, however, bluish or white-coloured, bare rock-faces of limestone constitute a conspicuous feature of the landscape. Several bands seem to be present, separated by mica schists; fossil remains are not to be expected, as the limestones mostly have a coarse granoblastic fabric, and in fact, all my searching in that respect was without success.

Bands of conglomerate are subordinate but not rare. They are bedded and contain rounded or subangular pebbles of sandstones, argillaceous rocks and carbonate rocks, and pebbles of quartz dioritic and granitic rocks have also been observed. Thus, their pebble content (and their appearance in general) to some extent differs from that of the conglomerates associated with the fossil-bearing limestones at Sardnes and east of Nordvågen, wherein the pebbles are mostly of carbonate rocks. (The conglomerate of Store Kamøy, where the metamorphism has been of a low grade, is polymict and thus unlike the latter. Fossiliferous rocks seem to be absent.)

Discussion of the age of the tillite-like conglomerate at Duksfjord

The situation is now as follows:

1. The crinoid-bearing limestone and conglomerate at Sardnes is of about the same age as the fossil-bearing limestones east of Nordvågen, i.e. of Silurian (Upper Llandoverian?) age.
2. The fossiliferous beds at Sardnes belong to the lower part of a succession of phyllites and mica schists with conglomerate horizons, limestones and quartzites, a succession which tectonically forms a large synclinorium.
3. At Duksfjord another smaller synclinal fold, parallel to the large synclinorium (perhaps an integral part of one broader synclinorium), is comprised of mica schists, limestones, quartzites and conglomerates. The tillite-like conglomerate is one of the uppermost (perhaps *the* uppermost) bed of this syncline.
4. Although the existence of recumbent folds must be taken into consideration, there is no evidence of an inversion on a grand scale of the upper part of the Duksfjord syncline and/or the large synclinorium.
5. Locally, thrusts certainly exist. The common features of the sequence as a whole, however, tell against the presence of any large-scale thrusting of an older (Eocambrian?) group of rocks upon a younger (Silurian).

Holtedahl and I have discussed the age problem on the basis of the above-mentioned points. We find that an Eocambrian age for the tillite-like conglomerate is now less probable than it was before the discovery of the Silurian fossils. The available evidence favours the view that the tillite-like conglomerate at Duksfjord is a younger member of the sequence than the fossil bearing horizon at Sardnes, and therefore should be ascribed to the Silurian.

Furthermore, as the presence of a tillite of Silurian age in Norway would be very astonishing, the revised view with regard to the age means that the supposed glacial origin of the conglomerate is now questioned. Most likely, the unsorted and unbedded character of the rock must be looked upon as due to other agencies, e.g. landslides. An interpretation of the rock as a tectonic breccia is hardly conceivable, for it appears as a layer with a sedimentary boundary with the underlying member of the sequence.

The suggested mode of formation of the conglomerate is, at present, only guesswork. A final solution to the problems of its origin, age and metamorphic history would be very welcome. Indeed, a detailed investigation of the stratigraphy, sedimentology and tectonics of the rocks of Magerøy would be an important contribution to the geology of Northern Norway.

Some remarks on relations with rock formations outside Magerøy

Magerøy is the only place in Finnmark where Silurian fossils have been found. Elsewhere the youngest fossils (Tremadocian) are present on the Digermul Peninsula west of Tanafjord, in the top formation of the groups of non-metamorphosed rocks of Finnmark (Reading 1965). In the Guide to excursion A 3 (1960) it was suggested that east of Nordvågen there might occur equivalents of the dark coloured Cambrian and Lower Ordovician sediments on the Digermul Peninsula. This is prehaps drawing too detailed conclusions, especially after the Silurian (and not Ordovician) age of the limestones east of Nordvågen was demonstrated, as it requires a break at the base of the Silurian limestone, with a hiatus corresponding to the greater part of the Ordovician period. Nevertheless, the discovery of the Silurian fossils is of importance in a regional respect, especially because Magerøy is situated in the metamorphic belt of the Caledonides. Earlier, the sedimentation of the (later metamorphosed) rocks of the nappes of eastern Finnmark was presumed to have taken place mainly in latest Precambrian and/or Eocambrian times. The existence of Silurian fossils on Magerøy favours the tendency to regard the larger part of the rocks in question as being of a somewhat younger, viz. Cambro-Silurian age (see Strand, 1960, p. 275).

Acknowledgements

The report of Mr. J. J. C. Geul has been of the greatest value to this work on Magerøy. His geological map formed the basis of the planning of the excursions and was my guide in the field. I thank him too for the interesting time we spent together on the island.

I am also highly indebted to Prof. dr. Olaf Holtedahl for the exchange of views during our visits to Magerøy in 1959 and 1960, as well as during the preparation of the present paper.

Finally, thanks to Dr. David Roberts, who has kindly corrected the English.

Sammendrag

Stratigrafiske konsekvenser av oppdagelsen av siluriske fossiler på Magerøy

I 1959 og 1960 ble det funnet fossiler fra silurtiden i to lokaliteter ved stranden øst for Honningsvåg, henholdsvis 1300 og 1650 m øst for bebyggelsen ved Nordvågen (se kartskissen fig. 2). Fossilene består av sjøliljestilker, kjedekoraller, hornkoraller, og brakiopoder av slekten pentamerus (se fig. 5 - 7). Bergarten som inneholder disse fossilene er en grå kalksandstein med gulaktig forvitringshud. Sjøliljestilker ble også funnet i et kalksteinskonglomerat (fig. 4) ved den vestligste (fig. 3) av de to lokalitetene. I skifer ble det funnet graptolitter. I den vestlige fossilforekomsten ble det funnet bare sjøliljestilker (og graptolitt i skifer i nærheten), de øvrige slags fossiler ble funnet i den østlige lokaliteten. Bergartene langs stranden er ellers mørke skifre i veksling med grå sandsteiner. Bergartene har vært utsatt for metamorfose, som bl. a. har ført til at kalksteinen er blitt krystallinsk.

Funnet er oppsiktvekkende da det ikke tidligere er kjent sikre siluriske fossiler nord for Trøndelag. For Magerøys vedkommende har funnet ført til at spørsmålet om alderen av også de andre bergartene på øya måtte tas opp til ny vurdering. Av spesiell interesse er et usortert konglomerat ved Duksfjord i den nordlige del av øya (se kartet fig. 1). Det ble tidligere ansett for å være tillitt (morenekonglomerat), avleiret under samme istid som tillittene i Alta og Øst-Finnmark, d.v.s. i eokambrisk tid (tidlig kambrium). Resultatet av de undersøkelser som nå foreligger, tyder på at det tillitt-liknende konglomeratet er et av de øverste ledd (kanskje det øverste) i en lagrekke av glimmerskifre, krystallinske kalksteiner og konglomerater, en lagrekke som danner et stort synklinorium (foldningstrau) mellom Duksfjord og Sardnes, og hvorav noen av de laveste lagene ses nord for Sardnes (i det sørvestlige hjørne av kartet fig. 1 - navnet står ikke på kartet). Disse lagene ved Sardnes minner om bergartene øst for Nordvågen. Forfatteren av denne artikkel fant i 1962 (dårlig

bevarte) sjøliljestilker i kalkstein og i konglomerat i flere lokaliteter nord for Sardnes. Det er dermed sannsynlig at også det tillitt-liknende konglomerat ved Duksfjord er av silurisk alder. Forekomst av tillitt fra silurisk tid i Norge ville være svært overraskende, det er derfor ikke lenger rimelig å anse konglomeratet for en istidsavleiring, det er mer naturlig å ty til andre forklaringer på dets usorterte karakter, f. eks. at det er dannet ved et slags undersjøisk ras.

Gabbroen i den sørøstlige del av øya kan ikke være eldre enn fra silurisk tid, idet dens fremtrengning førte til at de sedimentære bergartene omkring ble kontaktmetamorfosert (omdannet til hornfels. Merk at "Gabbro complex" på kartet fig. 1 også omfatter hornfelser, som til dels har tydelige sedimentære strukturer).

En upublisert rapport fra den hollandske geolog J. J. C. Geul, som i sommerhalvåret 1958 utførte geologisk kartlegging på Magerøy etter oppdrag fra Norges geologiske undersøkelse, var grunnlag for de undersøkelser som førte til funnet av fossilene og har vært et verdifullt hjelpemiddel ved det videre feltarbeid på Magerøy. Fig. 1 er et forenklet kart tegnet på grunnlag av det kart som hører til J. J. C. Geuls rapport.

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Big boulders of tillite rock in Porsanger, Northern Norway

By
Sven Føyn.

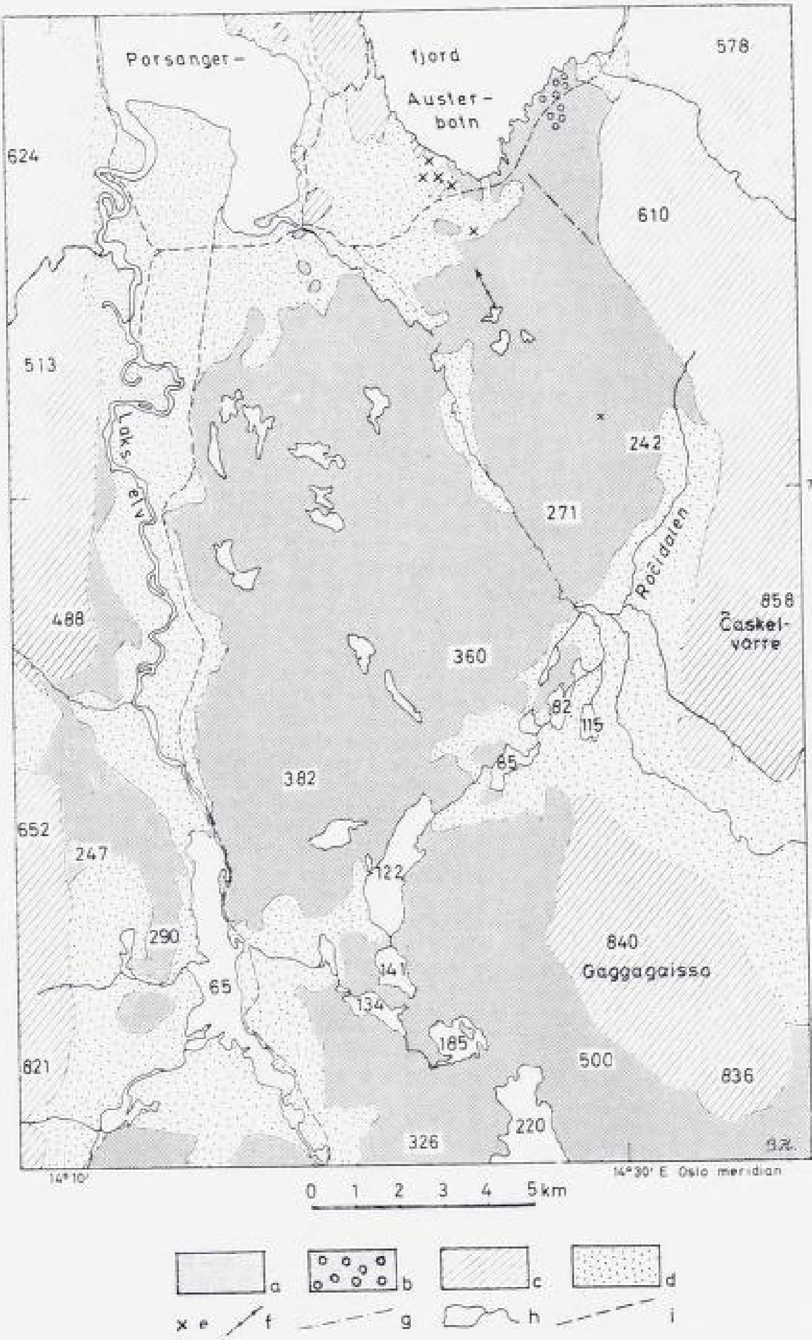
Abstract

In 1959 numerous erratic boulders of tillite rock were discovered at the head of Austerbotn, the eastern arm of the Porsangerfjord. Some of the boulders are very big, having volumes of up to 20 m³. In 1965 another two boulders were found about 7 km to the south-east of the head of the fjord. The presence of the tillite boulders shows that Eocambrian tillite occurs - or at least has existed - in the Porsanger region. No deposits of tillite occurring in situ have, however, been reported from this district. The writer suggests that the source of the boulders is most probably in Ročidalen, a depression in the Precambrian surface of the broad Lakselv valley, about 10 km south of the head of the Porsangerfjord. As there are no rock exposures in the bottom of Ročidalen on account of the thick cover of Quaternary deposits, this theory can hardly be proved. Possible future finds of tillite boulders may bring other parts of the Lakselv valley into focus.

Introduction

Numerous erratic boulders of tillite rock occur west of the head of Austerbotn, the eastern arm of the Porsangerfjord (lat. 70° 4′ N, long. 24° 68′ E). The boulders are found mainly on the slope facing the sea, but also on the small hill north of the main road. No occurrence of tillite in solid rock has been reported from Porsanger. The nearest known in situ deposits of Eocambrian tillite are those south of Laksefjord more than 50 kilometres to the NE, and at Altafjord about 60 km to the west. As the only agent which can be responsible for their transport and present distribution is the glaciers of the Quaternary period, and since the ice movement was from a southerly direction, the boulders at the head of the Porsangerfjord cannot have been derived from either of the above-mentioned deposits.

The boulders were discovered by Prof. O. Holtedahl, Dr. P. Reitan and


Porsanger-
fjord
Auster-
botn
578
624
610
513
242
70°00'
271
Loks-
elv
Roćidalen
858
Časkel-
várre
488
360
82
115
85
382
652
247
122
290
840
Gaggagaissa
65
141
134
185
500
821
836
220
326
14°10'
14°30' E Oslo meridian
0 1 2 3 4 5 km
a
b
c
d
e
f
g
h
i

the writer in 1959, during field preparation for the international geological excursion to Finnmark, 1960. In 1962 the present writer spent a few days in Porsanger in order to try to find the origin of the boulders.

General geology

The geology of the region south of the Porsangerfjord is known from the work of Holtedahl (1931), Crowder (1959) - see the map Fig. 1 - and Skålvoll (the Quaternary deposits) (1960). In the Lakselv valley, which is the continuation of the Porsangerfjord depression, the rocks are Precambrian supracrustals, considered to belong to the Karelides. The steep sides of this broad valley consist mainly of the Porsanger sandstone formation (of late Precambrian age), which forms a nappe lying with thrust contact on beds of the Dividal group (= the "Hyolithus zone").

The Dividal group in Finnmark was formerly, in its entirety, ascribed to the Lower Cambrian. The lower part of it has recently been shown to be of Eocambrian*) age, being equivalent to the formation next to the upper tillite of eastern Finnmark (Føyn 1967).

The Austerbotn area

At least a dozen of the boulders at Austerbotn have a diameter (or length) of more than one metre. Some of them have a volume of about 20 m³. The colour of the matrix is partly red-brown, partly grey. All the pebbles and

*) The term Eocambrian is here applied in the "restricted" sense to the time beginning with the deposition of the lower tillite in Finnmark and ceasing at the base of conventional Lower Cambrian.

Fig. 1. Geological map of the Lakselv valley area (p. 224).

a) Precambrian crystalline rocks.
b) Patches of conglomerate of the Dividal Group on the Precambrian surface east of Austerbotn.
c) Caledonian rocks. (Dividal Group (= "Hyolithus zone") and Porsanger Sandstone Formation).
d) Quaternary moraine, gravel and sand deposits and bogs.
e) Localities with boulders of tillite rock.
f) Main direction of Quaternary glacial striae.
g) Fault.
h) Lake and river, sea-shore.
i) Road.

Geological boundaries largely after Crowder (1959).

boulders in the tillite consist of Precambrian crystalline rocks. The tillite shows practically no signs of tectonic deformation, its unsheared character indicating that it is unlikely to have belonged to the rocks constituting the nappes. It is far more likely that the source has to be searched for close to the old Precambrian surface. Remnants of a tillite cover may perhaps have been preserved in depressions in the Precambrian surface, like the tillite deposits described by P. Holmsen from West-Finnmark (Holmsen 1956 and 1957).

In point of fact, the distance between the present position of the boulders and the old surface of the Precambrian basement is not large - both in the lateral and vertical sense. Outcrops of coarse, feldspathic sandstone and green and reddish shales show that the solid rock in the hill and along a part of the shore consists of beds belonging to the basal part of the Dividal group. Rocks belonging to the Precambrian basement crop out, however, only a few hundred metres east of the hill. Furthermore, in the bog on the southern side of the main road south of the hill, crags of Precambrian rocks are seen, and south of the bog, which is about 700 m broad, the Precambrian rocks are well exposed.

At the southern border of the bog one (rather small) boulder of the same sort of tillite rock as in the accumulation on the hill has been observed. The boulder had appeared from the Quaternary moraine below the peat, during a farmers digging of a drainage ditch.

The peneplain is fairly well preserved about 2 km NE of the head of Austerbotn (Holtedahl 1918, p. 134). The main road crosses an area where patches of the basal conglomerate of the Dividal group are seen on the surface of the Precambrian rocks, which there are mostly hornblende schists. The peneplain dips about 3° (angular measurements based on 100° scale) to the north-west. Between this area and that of the Dividal rocks west of Austerbotn, a block faulting must have taken place at a time later than the deposition of the Dividal beds. One of the fault lines is shown on the map Fig. 1 (see also Crowder's map (1959)). More parallel faults are evident, judging from the marked linear breaks in the landscape. The trend of these lines is N 50° W, each south-western block having been raised slightly relative to its north-eastern counterpart.

In order to find more tillite boulders, if any, the writer made excursions in the area of the Precambrian rocks as far as about 10 km south of the head of the fjord. In this area a rather massive greenstone (hornblende diorite) is seen as solid rock. Boulders of this greenstone are also found west of Austerbotn together with those of the tillite and are of the same order of size as the latter. Their direction of transport corresponds to the main trend of the glacial striae,

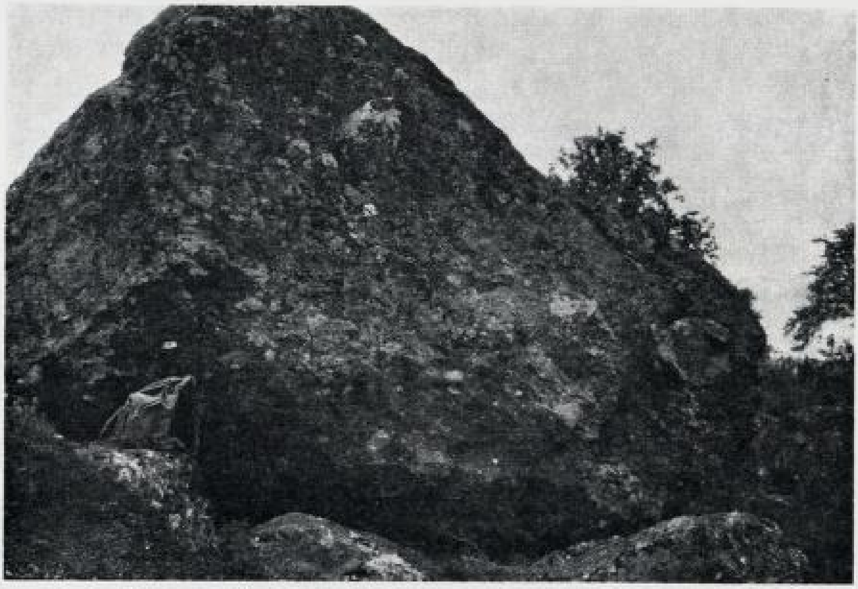


Fig. 2. Boulder of tillite rock west of the head of Austerbotn.

which is about N 30° W. (Very locally, on the Precambrian surface south of the main road, at Austerbotn, a trend N 25° E was observed). In spite of a rather thorough search, not a single block of tillite could be found in the area.

As a result of the investigations in 1962, the present writer concluded that the tillite boulders of the large, but local concentration at Austerbotn could hardly have been transported over any long distance. It was thought that, in all probability, the source must be just south-east of the hill, in the depression now occupied by the bog. This theory would also have accounted for the existence of the tillite boulder at the southern boundary of the bog.

Another boulder locality

It is now known, however, that this theory is not tenable, or at least, it does not accord with the full facts. In 1965 and 1966, geologist cand. real. Gunnar Juve conducted thorough investigations in the Precambrian area south of Porsangerfjord, with the ore mineralization as his main topic. When the writer visited the area in 1966, Juve drew attention to two big boulders of tillite rock, situated about 7 km south-east of the head of Austerbotn. The compass direction from these two boulders to the hill west of Austerbotn roughly coincides with the main direction of the glacial striae in the area.

The consequence of Mr. Juve's find of these two boulders is that the source of the tillite boulders must be farther to the south or SSE.

Ročidalen - the place of origin of the tillite boulders?

As realized by Holtedahl (1931), the Precambrian surface of the Lakselv valley was bowed up in early Caledonian time into a ridge running SW-NE, about 15 km to 10 km south of the head of Porsangerfjord. The ridge has, on the whole, a gentle slope (about 3°) towards the fjord, but is much steeper on its south-eastern side, after which a gentle rise of the Precambrian peneplain towards the south can be demonstrated. While the Caledonian thrust plane at and to the north of the ridge cuts the Dividal beds only ten or a few tens of metres above the basement surface, the thickness of the Dividal group south of the ridge is about 250 m.

Ročidalen is the north-eastern part of the depression south-east of the ridge. In this valley Quaternary deposits, probably of considerable thickness, entirely conceal the solid rock.

South of Ročidalen, along the south-western side of the mountains Časkelvarre and Halkkavarre*) and around the mountains Gaggagaissa and Mellemfjellet*), the contact between the Precambrian rocks and the basal beds of the Dividal group is exposed at several places. This area has been studied in detail by the geologist, mining engineer Bernt Røsholt and assistants, and the writer has also visited the area for a couple of days. No occurrences of tillitic rock have been observed.

Thus, Ročidalen seems to be the most probable place of origin of the tillite boulders, at least in the eastern part of the Lakselv valley. As mentioned above, no solid rock is visible, and because of this, proof of the possible existence of tillite below the Quaternary deposits is hardly expected to be forthcoming. If more tillite boulders were to be found north of the valley, especially on its very slope, this would be a useful and additional indication that the source of the boulders is hidden in the bottom of the valley.

Ročidalen, however, should not be looked upon as the only place likely to be concealing in situ tillite. Future finds of tillite boulders may bring a more south-western part of the Lakselv valley into focus. Concerning the boulders at Austerbotn, floating icebergs as a contributory transportation agent can not be left completely out of account (M. Marthinussen, personal communication).

The question of the source of the tillite boulders in Porsanger is not solved. Important as this kind of rock is as a source of knowledge of palaeoclimatology and palaeogeography, it would be of great interest to find the tillite as solid rock also in this part of Finnmark, or - if the deposits do not exist there any

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*) Halkkavarre and Mellemfjellet are off the bottom right-hand corner of the map Fig. 1.

longer - to find more indications as to where the position of such deposits may have been.

The main purpose of this paper has been to call attention to the fact that Eocambrian tillite may occur - or has existed - in the Porsanger district. Any new finds of tillite boulders may bring the problem closer to its solution. Reports of such finds to Norges Geologiske Undersøkelse will, therefore, be very welcome.

Acknowledgements

I wish to thank Prof. O. Holtedahl, Mr. G. Juve and Mr. B. Røsholt for pleasant collaboration in the field and valuable discussions afterwards. I am also indebted to Mr. M. Marthinussen for critical reading of the manuscript and to Dr. D. Roberts for correcting the English.

Sammendrag

Store flyttblokker av tillitt-bergart i Porsanger, Nord-Norge

I 1959 ble det oppdaget et stort antall flyttblokker av tillitt-bergart (= morenekonglomerat) på den lave høyden og i skråningen mot sjøen like vest for den innerste ende av Austerbotn i Porsanger. Noen av blokkene er svært store, opp til 20 m^3 i volum, og minst et dusin har en diameter (eller lengde) på over en meter. Nok en blokk ble funnet (i 1962) i søndre del av myra på sydsiden av riksveien. I 1965 ble ytterligere to store blokker (1 m^3 og 2 m^3) funnet ca. 7 km sørøst for fjordbotnen. Blokklokalitetene er merket med x på kartet fig. 1.

Flyttblokkene viser at tillitt fra eokambrisk tid forekommer — eller i det minste må ha eksistert — i Porsanger-området. En kjenner imidlertid ikke til noen forekomst av tillitt i fast berg i dette område. De nærmeste forekomster av tillitt i fast berg er sør for Laksefjorden, minst 50 km i luftlinje mot nordøst, og ved Alta, 60 km i luftlinje mot vest. Flyttblokkene kan ikke være kommet fra noen av de nevnte stedene, da transporten må være skjedd med isbreer under den kvartære istid, og i Lakselvdalen var brebevegelsen fra sørsørøst. Forfatteren av denne artikkel mener at Ročidalen må være det mest sannsynlige sted for opprinnelsen til tillittblokkene. Ročidalen er en senkning i grunnfjellsområdet i Lakselvdalen, ca. 10 km sør for Austerbotn. Uheldigvis (i denne forbindelse) er bunnen av Ročidalen fylt av løse avleiringer fra kvartærtiden, disse jordlagene er sannsynligvis svært tykke, og fast berg er ikke å se. En kan derfor ikke vente å få noen endelig avgjørelse på spørsmålet om det finnes tillitt i fast fjell liggende på grunnfjellet under jorddekket. Dersom det kunne

finnes flere blokker av tillitt nord for Ročidalen, særlig i selve dalsiden, ville det være til støtte for teorien om Ročidalen som opphavsområde for tillitt-blokkene.

En bør imidlertid ikke anse Ročidalen som det eneste mulige sted for opprinnelsen til tillitt-blokkene. Skulle det i fremtiden bli funnet blokker i andre deler av Lakselvdalen, kan også mer sørvestlige deler av denne komme på tale som opphavssted for blokkene.

Tillittbergartene er viktige som kilde for kjennskapet til paleoklimatologi og paleogeografi. Det ville derfor være av stor interesse å finne tillitt også i form av fast berg i Porsanger, eller å finne hvor eventuelle forekomster av tillitt en gang må ha ligget. Ethvert nytt funn av tillitt-blokker kan bringe løsningen av problemet nærmere. Norges geologiske undersøkelse er derfor interessert i å få opplysninger om slike eventuelle funn.

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A preliminary account of the geology of the Signaldalen-Upper Skibotndalen area, Inner Troms, N. Norway

By

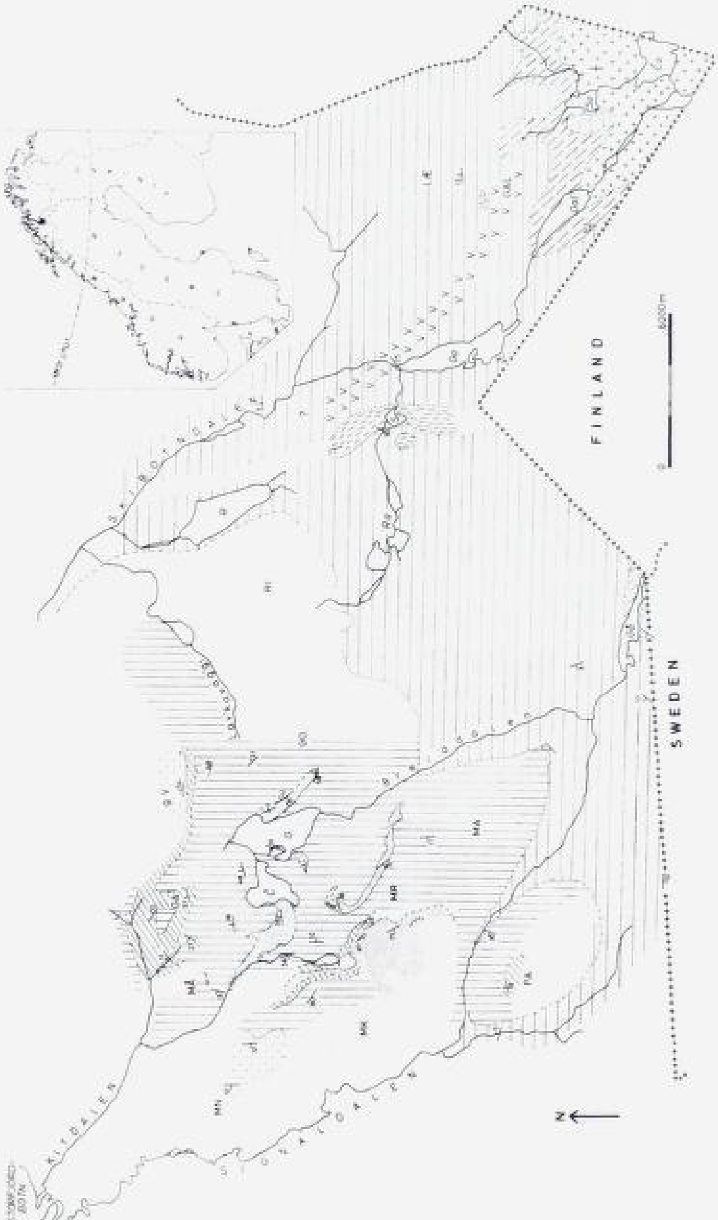
Richard E. Binns[1]

Contents

Abstract

A preliminary account of the petrography, stratigraphy and structure of an area at the Caledonian front in inner Troms, N. Norway, is given. Precambrian autochthonous gneisses are overlain by autochthonous and par-autochthonous Cambrian sediments - the Hyolithus Zone. These are themselves overlain by Caledonian thrust metamorphics, divided into the Lower and Upper Allochthonous units. The lower unit consists mostly of quartzo-feldspathic phyllonites, which may be mainly sedimentary in origin. It is bordered below and above by low angle thrusts. The Upper Allochthonous unit has a very varied lithology of calcareous, quartzitic, pelitic and basic rocks all in a fairly advanced metamorphic state. Some of the rock boundaries (especially quartzites) are tectonic. A provisional stratigraphy is suggested but this is uncertain due to the structural complexities in the area. At least one early period of isoclinal folding is recognised, and the main thrusting is associated with this. These structures were refolded by a system of open NW-SE folds and a possibly later system of gentle NE-SW flexures. The regional significance of this structural history is discussed.

[1]) Department of Geology, Tromsø Museum.


FINLAND
SWEDEN
N

Introduction

The area being mapped lies in the innermost part of Troms county, N Norway, some 400 km N of the Arctic Circle, at about 69° 10′ N, 20° 30′ E. The precise location is shown on the maps (Fig. 1).

This is an area of elevated lake plateaux and high, steep mountains. The area is geologically fairly well exposed, though lakes, and lake and river deposits, together with morainic material, cover extensive stretches of the flatter land, whilst block-fields are extensive on the mountains in the eastern

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Fig. 1. Simplified Geological Map of the Signaldalen - upper Skibotndalen area (p. 232).

Undifferentiated rock		UPPER ALLOCHTHONOUS UNIT
Norddalen Schist		UPPER ALLOCHTHONOUS UNIT
Limestone, schists, basic and acid rocks	Plateau Limestone Group	UPPER ALLOCHTHONOUS UNIT
Quartzite	Plateau Limestone Group	UPPER ALLOCHTHONOUS UNIT
Phyllonite (granitised - V V)		LOWER ALLOCHTHONOUS UNIT
Phyllite and Limestone		LOWER ALLOCHTHONOUS UNIT
Cambrian		
Precambrian basement		
Glacier		

25 Dip of Bedding and Schistosity

Vertical Foliation

MOUNTAINS

gV *gaskamus Viessugas*
GA *Gas'kasuorgigai'si*
GAL *Gal'laoai'vi*
GO *Goatte raš'ša*
LÆ *Læt'tačåk'ka*
MA *Mar'kus Mal'la*
MK *Markusfjellet*
MN *Mannfjellet*
MR *Mar'kusriep'pi*
MÅ *Måskogai'si*
PA *Paras*
RI *Riep'pe gai'si*

LAKES

C *Čaccajav'ri*
CO *Čoap'peluobbal*
G *Gåv'dajav'ri*
Ga *Galgujav'ri*
Gal' *Gal'lajav'ri*
Gå *Gåldajav'ri*
R *Riep'pejav'ri*
Ra *Raš'šajavrit*
S *Sallujav'ri*
SO *Sađ'gejav'ri*

half of the area. Snow cover in the past three summers has been unusually heavy and has, together with the generally poor weather, proved a serious drawback to the mapping programme.

Very little work has been done in the area previously. Pettersen (1868, pp. 91-2, 1870, pp. 4-15) has mentioned some observations he made, mostly on the edges of the area. Padget (1955) has described a profile along Skibotndalen. An account of the Hyolithus Zone and associated rocks in the Finnish-Norwegian border area has been given by Hausen (1942) and the geology of the neighbouring part of N Sweden has recently been outlined by Kulling (1964).

The present mapping was begun in 1961 whilst I was a member of an undergraduate scientific expedition from University College of Wales, Aberystwyth. Some observations were made in 1963, but the bulk of the work has been done in parts of the summers of 1964-66 and it is hoped to complete most of the work in the summer of 1967. The present paper is a preliminary account of the stratigraphy, petrography, and structure of the area. As the mapping so far accomplished has revealed several features which do not fit in well with some of the previously published ideas on the geology of the region as a whole, it seems desirable to present some of these matters, now, so that other workers in the region, now and in the near future, can be aware of the other possibilities in interpretation and correlation that may be applicable.

Mapping is being done mainly on a scale of 1 : 25,000 using base maps photographically enlarged from the new Series M 711 1 : 50,000 topographical maps.

Autochthonous Precambrian

The autochthonous Precambrian is exposed only in the SE corner of the area where it continues across the Finnish-Norwegian border from the extensive granodioritic-quartz dioritic complex of Svecofennian age in N Finland (Eskola 1963). Glacial drift and lakes cover most of the solid rock, but scattered exposures of granodiorite, with some quartz diorite, granite and granite-aplite occur. Intersecting and partly cross-cutting, semi-amphibolitised dark-green basic bodies occur here and there, sometimes sub-parallel to the steep, near N-S, foliation which is generally developed in the acid rocks.

The acid rocks are normally poor in dark minerals, which even in the granodioritic types, are mostly represented by a dirty green biotite. The granodiorite contains considerably more plagioclase than potash feldspar, and the plagioclase seems to be older as it is filled with minute sericitic and other inclusions, whilst microcline is noticeably fresh.

The Precambrian surface is uneven and has been subject to sub-Cambrian weathering, as also observed by Hausen (1942, p. 16) at the same locality, NE of Saðgejav'ri.

The basement rocks show clear evidence of shearing and the constituent minerals are sometimes very tectonised. This has affected rocks both at the Precambrian surface and some way below it.

Autochthonous and par-autochthonous Cambrian

Some details of the Cambrian Hyolithus Zone stratigraphy in the SE corner of this area have been given by Hausen (1942). According to Hausen, and Kulling (1964), the Cambrian also extends across the border from Sweden beside Treriksrøysa. A fairly intensive search in most of the bog and woodland here, has failed to reveal any exposures of anything E of the approximate boundary (shown on the map in Kulling 1964) between the Cambrian and the overlying thrust rocks. The junction on Fig. 1 is therefore based on the maps published by Kulling and Hausen.

The Cambrian sediments have only been studied in a very cursory manner. They consist chiefly of grey, reddish and greenish-brown shales and fine grained sandstones, with a basal conglomerate rich in pegmatitic quartz and other pebbles derived from the underlying weathered basement. The rocks are generally more or less untectonised and should offer good opportunities for finding fossils. However, the uppermost greenish-brown shales are considerably deformed in the exposures S of Gal'lajav'ri, and must be considered par-autochthonous.

Lower Allochthonous Unit

As indicated on the map, this group of rocks is very extensive in the S of the area. It chiefly consists of more or less phyllonitic, banded, quartz-feldspar-mica tectonites which have been thrust over the Cambrian sediments. Small scale shears and thrusts are often seen, and small tight, isoclinal and recumbent folds are frequent in many parts. Joint drags are common. Most of the unit is metamorphosed in the quartz-albite-epidote-biotite subfacies of the Green Schist grade (Turner and Verhoogen 1960, pp. 537-8).

The major part of this unit can quite likely be equated with the so-called sparagmite which outcrops in large areas of outer Finnmark and in Troms, and continues S in Sweden and Norway.

Light- and subordinate dark-grey, fine grained banded phyllonite and grey-green, fine grained phyllonite are the most extensively exposed varieties, and reach a total thickness of several hundred metres. The latter variety, especially, has a distinctly green and glassy appearance when wet.

The term "phyllonite" is used to describe these rocks, because this, as defined and used by Knopf (1931) and Christie (1961) seems to best describe their macroscopic and microscopic appearance.

Feldspar porphyroclasts are frequently present and microscopic examination shows that these are mostly of micro-perthitic microcline, with some albitic or oligoclasic-albitic plagioclase. A few plagioclases are zoned but generally they are more or less clouded with inclusions. These feldspars seem to be partly regenerated relicts incompletely crushed by the tectonisation.

The granulated, mostly recrystallized, groundmass consists chiefly of quartz and mostly untwinned feldspar, with varying amounts of mica mainly segregated in layers. Most of the mica is white - partly large muscovite felts containing biotite inclusions, and partly fine sericitic flakes. Biotite occurs also as tiny flakes of late date. Garnets are present in some horizons, but are inconspicuous. They occur as near euhedral, fresh grains, but often as more or less altered relicts filled or replaced with epidote, zoisite, chlorite, etc. The feldspar content varies somewhat but often approaches 30 % of the total mineral content. Though the frequent independent occurrence of epidote, in coarser quartz-feldspar-mica schist layers, might suggest that these rocks were originally semi-calcareous (Harker 1939, pp. 247-8), some of the epidotes contain a core of allanite, normally an accessory mineral of acid igneous rocks. The near uniformity in composition of the plagioclases may also suggest an acid eruptive origin for these phyllonites. This problem will be mentioned again later (pp. 248-9).

This phyllonite is clearly the same rock as that described by the name "Sparagmite Schist", by Skjerlie and Tan (1961, pp. 185-6) from Reisadalen, E of here.

In the Gåldajav'ri area the lowest part of this Lower Allochthonous unit is a highly tectonised, light coloured or pinkish-brown rock of apparent granitic origin, with minor amounts of darker, partly more basic, rock.

This relatively insignificant portion of the unit is probably derived by imbrication from the Precambrian basement. Another similar body is seen high up in the thrust sheet W of Står'magied'di in Breiddalen.

In thin section it is seen to consist of large and medium-sized porphyroclasts of potash and alkali feldspars, with epidote, ore, and especially quartz and mica. The potash porphyroclasts are micro-perthitic and both microcline and orthoclase are represented, though the former is dominant and outweighs the albite-oligoclase alkaline feldspar. Some of the feldspars are zoned, but generally the plagioclases are cloudy and filled with small inclusions and all the feldspars are extensively cracked and disrupted, the cracks being generally filled with quartz and a lesser amount of untwinned feldspar, whicn cohtinue unbroken from the groundmass.

The groundmass is mostly of very small, sutured quartz, with some feldspar and a little calcite, and muscovite has crystallized along some shears and around porphyroclasts. Green biotite occurs as very small flakes with a semi-parallel alignment.

In Skibotndalen are two other horizons worthy of special mention here. The first is the zone of granitization described by Padget (1955), of which the southerly outcrop is within the area now being studied, and the second is the belt of dark, somewhat phyllitic rocks with pale-coloured limestone bodies that immediately overlies this southerly granitisation belt.

Padget believed that the granitisation zone overlay his so-called sparagmitic schists (banded phyllonite, here), but reconnaissance mapping in the fairly well-exposed ground N of Gal'lajav'ri indicates otherwise. Here the banded phyllonites overlying the Cambrian sediments on the lower S slope of Gal'laoai'vi become gradually more granitized and mixed with basic material in the upper part of the hill. In the summit region pink and white quartzo-feldspathic layers alternate with fairly coarse pale- and grey-green bands similar to the normal phyllonitic schist. Green amphibolitic horizons are important. In the floor of the high E-W trending valley N of Gal'laoai'vi is a small anticlinal core containing a pure, white, crystalline limestone and a highly puckered phyllite. The relatively complex folding of the granitic/basic zone and the speed of the mapping here make it a little uncertain, but this seems to immediately overlie the granitisation zone and is overlain by a little more granitic/basic material and then feldspar-porphyroclastic grey phyllonite of a type very characteristic for this Lower Allochthonous unit and described by Padget from Helligskogen (Padget 1955, pp. 10 and 52-53). This rock appears to outcrop continuously between Skibotndalen and the Finnish border for several kilometres northwards.

Scattered observations in the ground between Galgujav'ri and Gal'laoai'vi indicate that the granitisation belt on the latter is continuous with that by the road at the N end of the lake. W of the road it continues to outcrop in a NNW direction and is overlain in the area around Mul'kejav'ri-Sallujav'ri-Råg'gejav'ri by a broad belt of dark phyllitic schists with irregular pale-coloured limestone horizons. This is the other limb of the open NNW-SSE antiform that occurs in the upper part of Skibotndalen, and this horizon is doubtless equivalent to the similar rock N of Gal'laoai'vi.

The approximately 600 m thick, dark, fine grained phyllitic belt is mostly highly tectonized quartz-biotite-feldspar-clinozoisite rock, but appears to be derived from amphibole-rich material as this still occurs in rare relict porphyroclasts and in more or less untectonized and unrecrystallized bands. Skarn is developed locally in association with one amphibolite layer. The quartz often occurs in lenses and is then coarse grained and has strongly undulating extinction. These lenses sometimes contain relict feldspar fragments. Larger pods of quartz occur at two horizons and they then simulate sheared and folded remnants of once more continuous bands. The feldspar is partly microcline, but mostly plagioclase. The plagioclase is extensively altered and occurs mostly as small porphyroclasts of relict albite. Amphibole, sphene, zircon, chlorite, white mica, allanite and ore occur in varying amounts.

The limestone is white, pale-grey or pale-green and partly severely tectonized. It consists essentially of medium to very fine grained calcite and dolomite, with varying amounts of colourless amphibole and white mica, both of which sometimes occur as very large, bent, porphyroblastic sheets.

The limestone is not confined to one horizon and has often reacted extremely plastically to the deformation forces. A large outcrop in the cliff immediately WSW of Sallujav'ri is enclosed within the fine grained, light grey, banded phyllonite which otherwise overlies this belt, here. The limestone varies from a few to over 60 m in thickness.

Except for the basic nature of the non-calcareous rock and its structural position it would be natural to correlate this belt with the Rautas rocks of N Scandinavia (the par-autochthonous Blue Quartz - Shale - Dolomite of Hausen (1942) and the Lower Thrust Rocks of Kulling (1964)). The rock may also be correlateable with the greenschists described from a similar horizon in Reisadalen (to the E) by Skjerlie and Tan (1961, pp. 186-7). But the question must be left open for the present.

Allanite occurs especially often in the granitic layers of the granitized belt but is sometimes present in the dark phyllitic rock and the normal phyllonite. The allanite nearly always has an outer shell of epidote (?clinozoisite - with blue interference colours - in the phyllites) but is occasionally seen in tiny pleochroic haloes in biotite. Its significance will be subject to a closer study, later, but it is provisionally considered to be a pneumatolytic mineral formed during the granitisation phase and to have migrated from the highly granitized parts by hydrothermal action along the shear planes in the surrounding phyllonites, thus dating the granitisation to a fairly late stage in the recrystallization and deformation of the Lower Allochthonous rocks.

Cataclasis is widespread throughout the Lower Allochthonous rocks, being represented among other ways by tectonically segregated quartzo-feldspathic material, by fracturing and bending of feldspars, granulation, undulating extinction and Böhm lamellae in quartz, bent micas and the stretching of epidotic minerals, garnet, sphene and zircon - though not allanite.

But the most extreme forms of dynamic metamorphism are restricted to certain levels representing shear zones varying from a few millimetres to many metres in thickness. The larger zones have been mostly observed and studied in the upper part of the unit in Parasdalen and Stordalen, the upper branches of Signaldalen.

They take the form partly of very dark rocks consisting of an extremely fine grained mylonitic groundmass containing stringers of medium grained cataclastic quartzo-feldspathic material. Elsewhere they are medium to fine grained, brown weathering, puckered phyllites, or sheared lenses of amphibolitic material. Kyanite and sillimanite are widely developed and garnet occurs. Trains of graphite and the development of Cu/Fe sulphides are common features. In the amphibolitic rocks the coloured amphiboles appear both as squat, often euhedral, prisms, and long blades. The plagioclase is more calcic (oligoclase-andesine) than in the bulk of the phyllonites, though some albite still occurs sometimes, and labradorite is found. Most of the mica is brown biotite.

These mylonitic rocks are associated with the major thrust separating the Lower Allochthonous and Upper Allochthonous units, which is mentioned in more detail later (p. 247).

Upper Allochthonous Unit

This is exposed in the northerly, mainly higher, parts of the area being mapped. The structure, as outlined in a later chapter, is rather complex and the following divisions are only in a provisional stratigraphical order. The rocks are mostly metamorphosed in the lower half of the almandine-amphibolite facies (Turner and Verhoogen 1960, pp. 544-8).

Plateau Limestone Group

This name is used, for the moment, for the most extensive association of rocks within this unit. All the lithological variations outcrop typically on and around the 700-750 m lake plateau, in the centre of the region, around Čaccajav'ri and Gåvdajav'ri.

The group has a very varied lithology. Calcareous rocks, both nearly pure limestone and calc-silicate schist, quartzites, semi-pelitic and pelitic schists, and basic rocks are all important, and the lithological changes are sometimes lateral. Metasomatic processes, particularly, have led to gradations between various rock types and to a frequent development of semi-gneissic and granitic rocks. A detailed description of these rocks will be reserved for a later paper, but some of the most characteristic features may be mentioned here.

The light coloured crystalline limestones are mostly fairly coarse grained, light-grey or pale-brown weathering, and vary greatly in thickness laterally. They are sometimes nearly or wholly replaced by pelitic and semi-pelitic schists. Sometimes only one thick (ca. 60 m) bed occurs in a profile (e.g. at Mar'kusriep'pi, S of Čaccajav'ri), but more often several thinner ones are found. They swell and thin out rather rapidly along the strike, and this and the complex minor folding, seen very clearly in some water-washed exposures, demonstrate the plasticity of the limestone during the metamorphism. The limestone, strictly defined, makes up a relatively small proportion of the group.

Mineralogically the limestone consists predominantly of calcite, with some diopside, clinozoisite, pyroxene, quartz, feldspar and varying amounts of ore, mica, graphite, yellow vesuvianite and fluorite. The last two have not been observed in thin sections containing mica. The vesuvianite occurs either as individual euhedral to anhedral crystals or in clots up to 1½ cm across, and where developed most coarsely fluorite is sometimes associated with it. Until identified by X-ray diffractometer powder patterns the yellow minerals were thought to belong more probably to the humite group, and this may still be true for some of them.

Where silicate minerals are more strongly developed the limestone becomes a brown, calc-silicate schist, or a characteristic green calc-amphibolite in which thin pale-green epidotic bands alternate rapidly with equally thin dark-green amphibolite ones. The calc-silicate schists, especially, reveal small-scale complex folding and shearing, and are particularly strongly developed just NW of upper Breiddalen, but are also seen in upper Norddalen, lower Stordalen and on the SE end of Mar'kus Mal'la around 950 m. The calc-amphibolite is widespread around Čaccajav'ri, and usually grades into a dark green amphibolite.

Quartzitic rocks occur in varying thicknesses (up to ca. 400 m on Goatte raš'ša) at several horizons within the Plateau Limestone Group. The larger bodies, at least, usually have tectonic boundaries with the encompassing rock. They are dominantly pale-brown weathering, more or less white, medium grained rocks containing interbedded, sometimes slightly transgressive, very thin garnetiferous amphibolites, and speckled semi-pelitic schists occur. They are frequently quite feldspathic, containing abundant sodic-plagioclase and some potash feldspar. Muscovite, and especially biotite, are scattered throughout and segregated into layers or laminae which, together with the basic layers and flattening of quartz, help to define the schistosity. The present outcrop pattern of the quartzites is partly due to structural repetition.

Those semi-pelitic rocks that are not associated with the quartzites are also highly feldspathic, but some of this oligoclase-andesine is very fresh and clearly metasomatically introduced. The rocks are otherwise characterized by garnets and by the abundance of biotite and muscovite, the latter occurring both as unorientated poikiloblastic grains and in parallel intergrowth with the better orientated biotite. These semi-pelitic rocks are extremely varied in appearance, at least partly because of differing degrees of metasomatism.

More pelitic schists, varying in thickness from one traverse to another, are often tectonized and display abundant evidence of stress, with strained quartzes and feldspars, bent micas, stretched and partly recrystallized garnets, and the development of kyanite. The most pronounced of these are believed to represent relatively minor thrust zones, partly contributing to the regional imbrication.

Thin, irregular, grey or dark-green garnetiferous amphibolites occur throughout the group, and are often boudined. Some are oblique to the foliation. Garnet-epidote skarn is frequently developed in the contact zone of neighbouring limestones, but vesuvianite is never observed in close proximity to them.

Granitisation is irregularly developed throughout the Plateau Limestone Group, and several zones of granite, granodiorite, and granite-gneiss occur. Much of the granitisation however shows itself by small scale metasomatic introduction and recrystallization of quartzo-feldspathic material, and as porphyroblasts, clots, veins and lenses in the semi-pelitic, quartzitic aid basic parts.

Except for the development of the vesuvianite and fluorite, and some other porphyroblasts, the limestone members are little affected, even where they are in close proximity to migmatised rocks, as on the 1100 m plateau area W of Mar'kusriep'pi and on the upper N slope of Markusfjellet. In some cases the granitisation seems to be more marked in the vicinity of late NW-SE fold cores. Pegmatitic dykes, lenses and pockets are common in some parts, and the dykes often trend around NNE-SSW.

Norddalen Schist

The Norddalen Schist, named after one of its typical occurrences in Norddalen, is apparently tectonically interlayered with the Plateau Limestone Group, in this area.

It is mostly a platy to well-bedded, dark-green to black, hornblende schist containing, in addition to hornblende, variable, but small, amounts of quartz, intermediate to sodic plagioclase, clinozoisite-epidote, biotite, sphene and ore. Garnets are not always present and tend to be concentrated in certain layers.

Interbedded with this are rare, thin, highly pelitic schists and very subordinate, but often characteristic, dolomitic limestones and calc-silicate schists. The limestones, and especially the calc-silicate schists are notable for their often high amphibole content. In thin section this mineral is colourless, lacks pleochroism and is often poikiloblastic. It tends to form large porphyroblasts which are black and sometimes conspicuously ribbed in hand specimen. In addition to the dominant carbonates and amphibole, a little quartz, rare sodic-plagioclase, muscovite and rare biotite, a little clinozoisite, much graphite, and in one section, abundant diopside, occur. Calcareous rock makes up no more than some 5-10 m out of perhaps 80-200 m in any of the profiles examined. Its appearance varies markedly from layer to layer and sometimes between separate occurrences. In one profile, in the N slope W of Mar'kus-riep'pi, a nearly pure limestone occurs, a thin section of which reveals no amphibole, and only a little muscovite, quartz, clinozoisite and ore, whilst immediately below is the more typical, amphibole-rich, calc-silicate schist.

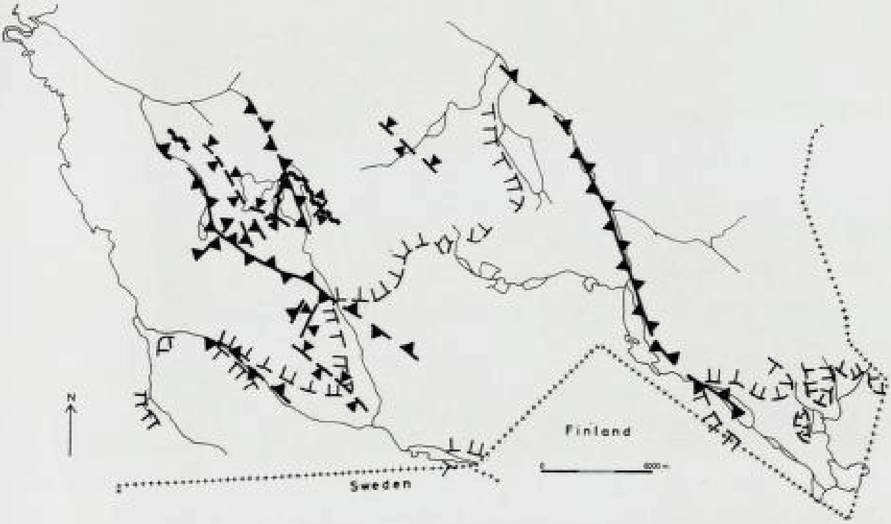
In the upper 500 m of the E shoulder of Mannfjellet, outcrop other, structurally higher, stratigraphical elements. These have been only superficially studied so far, due to their elevated occurrence and consequent heavy snow cover in recent years. They probably have quite an extensive distribution on the Markusfjellet-Mannfjellet range, making up much of the rock on the upper W side of these mountains, where mapping still remains to be done.

On Mannfjellet a 200 m thickness of thinly platy, green hornblende schist containing notable amounts of quartz and intermediate plagioclase, overlies a granitic horizon which is interpreted as belonging to the Plateau Limestone Group. Above that is a grey quartz-garnet-hornblende schist with poorly orientated porphyroblastic hornblende, and then a grey, apparently non-feldspathic, quartzite. Above that come several hundred metres of mixed pelitic and semi-pelitic schists, partly granitized near the top, and always with conspicuous garnets where the rock could actually be observed at close hand.

On the upper part of gaskamus Viessugas are brown schists which cannot at present be correlated definitely with any other unit in the area.

Structure

As stated previously the tectonic history and structure of this area is rather complex. It seems possible to draw a certain parallel between the Skibotndalen-Signaldalen district and that around Glomfjord in Nordland, some 400 km SW in the Caledonian chain, that has been described in a series of detailed papers in recent years (e.g. Rutland and Nicholson 1965, Holmes 1966).



Axial plane trace of late antiformal fold

Axial plane trace of late synformal fold

Axial plane trace of early fold

Major thrust

Fig. 2. Simplified Structural Map of the Signaldalen - upper Skibotndalen area.

However, due at least in part to the much less favourable exposure here, especially during the heavy snow cover of recent years, the tectonic history and structure have still not been fully worked out with certainty. Fig. 2 shows the main structural features. An outline of the main features and problems will be given below.

The most obvious structural trait in the area is a system of fairly open, more or less symmetrical folds, trending about NW-SE. Two major antiforms with a complementary synform cross the area. The westernmost one, which may be called the Kitdalen-Breiddalen antiform, is a complex anticlinorial fold, whilst the easternmost one (and apparently the synform too) is more simple. This easterly antiform, which trends more or less along Skibotndalen, has only

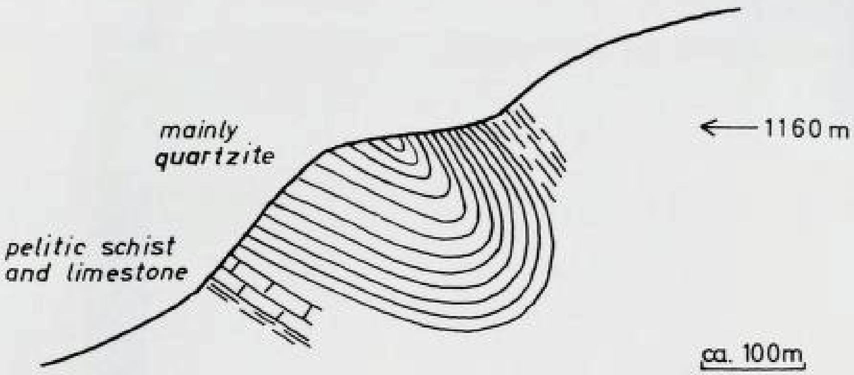


Fig. 3. Sketch-profile of part of the S face of the N part of Goatte rašša to show the shape of the early isoclinal fold here.

been observed where it affects rocks below the Upper Allochthonous unit and this may be the reason for its apparent simplicity as the westerly antiform appears to have a much simpler form in the outcrop of the Lower Allochthonous unit.

These folds are crossed at about right angles by more gentle, open folds. They may have developed simultaneously with the NW-SE folds, but several features point to a somewhat later development.

It is undoubtedly these fold systems that were recognized by Padget in the adjacent Birtavarre area (Padget 1955).

As the mapping has proceeded several features that will not fit in with this relatively simple structural pattern alone have become apparent, and it is clear that in this area, and consequently doubtless in the adjacent region too, the rocks have had a more complex structural history. Intensive deformation has resulted in inclined and recumbent isoclinal folds which pre-date the fold systems mentioned above (as these refold the earlier structures), and early thrusting has also occurred.

A couple of examples will suffice to demonstrate this at present.

In the W side of the N part of Goatte rašša is a large early fold, the closure being seen in the quartzite outcrop at the S end of the W face. As indicated on the map (Fig. 1) and in Fig. 3, the rocks dip for the most part to the E at a moderate angle, but about ¾ of the way across the outcrop (moving from W to E) a belt of vertical and near vertical dips is encountered. Fig. 3 shows the shape of this structure. Pelitic schists with thin limestones and basic bodies dip beneath the quartzite on the lower slopes and are seen on the other side of the quartzite along the N face of the mountain. The top part of the mountain consists of highly tectonized, heavily (reddish-brown) weathered rock

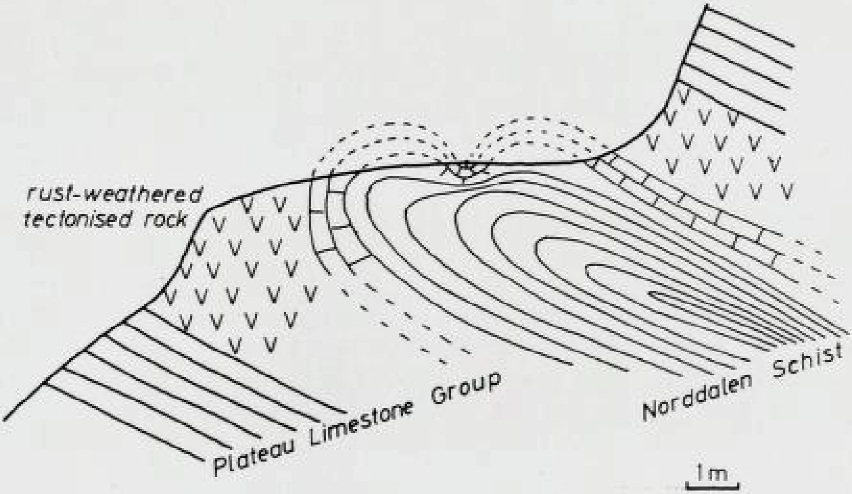


Fig. 4. Field sketch of relationships at the early fold closure on the W slope of Måskogai'si.

which seems to be granodioritic. This weathering colour is extremely characteristic for the whole of this mountain block between here and Riep'pejav'ri, as has been noted already by Pettersen (1870, p. 9).

As Figs. 1 and 2 show, this early fold is refolded by anticlinorial folds of the E limb of the Kitdalen-Breiddalen antiform, which also trend in a general NW-SE direction plunging to the NW at a moderate angle. The early fold is also refolded by a very gentle flexure trending about NE-SW. This occurs about half way along the W face of Goatte raš'ša (see Fig. 2). The fold closure is characterized by intensive shearing and very contorted minor folding, and there is a tendency for the basic material in the quartzite to be gathered in the lower horizons whilst the upper part of the trough consists of more pure quartzite than is normal.

The second example concerns a recumbent fold which outcrops across the mountains of Måskogai'si and Gas'kasuorgigai'si and no doubt continues further E and N, too. A closure consisting of a core of fissile hornblende schist, with pelitic schists and limestones folded round it, is seen in an exposed position high on the W side of Måskogai'si (Fig. 4). The E side of this mountain has suffered a large landslide at the position of the continuation of this hornblende schist, and consists mostly of large unstable boulders and displaced blocks of solid rock and the actual shape of the outcrop has not been determined yet. But on Gas'kasuorgigai'si a thick succession of hornblende schists with very subordinate pelitic and calcareous horizons (Norddalen Schist) overlies schists and limestones of the Plateau Limestone Group and underlies more schists and limestone, in no way distinguishable from the Plateau Limestone succession and very like that high on Måskogai'si. This recumbent fold is refolded by the anticlinorial waves of the Kitdalen-Breiddalen antiform.

The presence of this recumbent fold, here, seems difficult to reconcile with the succession seen in the E side of Markusfjellet. As the map (Fig. 1) shows this consists of two thick horizons of typical Norddalen Schist with an intervening thick quartzite, and underlain by Plateau Limestone rocks. The upper horizon of hornblende schist is overlain by a granitic gneissic zone, which is at present extensively snow covered except in the hollow between Markusfjellet and Mannfjellet. However, sticking out of the snowfield on the upper slope is a large exposure (the structure symbol on the map) of crystalline limestone of a type very characteristic for the Plateau Limestone lithology. The quartzite, moreover, is to all appearances equivalent to the quartzitic horizons in the Plateau Limestone Group. No satisfactory explanation for the nature of this succession has yet been evolved, but it is likely that very low angle thrusting has played some part in its development. There are several slide-like zones in the lower horizon of the Norddalen Schist, and the lower junction of the quartzite is a tectonic boundary - a characteristic of nearly all of the major quartzite junctions seen in the Plateau Limestone Group. The close proximity of this succession to the recumbent fold closure on Måskogai'si may be accounted for by postulating a slide between them, but this will not be discussed further at the present stage.

Unless the main quartzitic horizons of the Upper Allochthonous unit, included now within the Plateau Limestone Group, are really correlateable with part of the lithology of the Lower Allochthonous unit, no part of that unit occurs at a higher structural level in the present area. It nevertheless shows abundant evidence of having taken part in the early severe isoclinal folding, as recumbent and inclined, stacked zig-zag folds are frequently seen on a minor scale. Middle limbs are often sheared out. On the other hand the later deformation is not so marked in the Lower Allochthonous unit, probably because of the dying out of its effects at depth.

Minor structures related to all the major folds are frequently observed. The majority of the minor fold axes and the mineral and other lineations trend around NW-SE. But trends of anything between N-S and about E-W are common too. Further discussion of the minor structures will be postponed until a later occasion.

The thrust between the Lower Allochthonous unit and the underlying Cambrian sediments is well documented in the literature of the Caledonian chain. On the other hand evidence for a similar tectonic discordance above what is termed the Lower Allochthonous unit in this paper, is contradictory. Padget (1955) found no indication of a discordance in the profile he studied E of Helligskogen, but Skjerlie and Tan (1961, p. 189) found evidence for it further E in Reisadalen. Kulling (1964, pp. 86 and 148) says that a tectonic break is not discernable locally in N Sweden.

The present mapping has revealed a marked tectonic discordance at the top of the Lower Allochthonous unit at the head of Signaldalen. It continues in the area to the E too, and was noted by Pettersen (1870, pp. 8-9) at the S foot of Goatte raš'ša, though his interpretation of the relationship of the units is open to question.

This thrust is clearly folded by the late NW-SE deformation phase, which, as it to some extent affects both the thrust units along the same axial lnies, must post-date the main thrusting. It is therefore likely that the thrusting can be correlated with a culminating stage of an earlier isoclinal folding phase.

A summary of the structural history of the area, as far as it has been discerned at this stage, may now be given.

The sub-vertical foliation of the Precambrian basement may be attributed to a Precambrian orogenic phase, but the local shearing and tectonization in the basement is perhaps mostly derived from the Caledonian orogeny. At least one phase of isoclinal and recumbent folding has affected the allochthonous units. This plastic deformation was accompanied by disruption of certain horizons, notably at junctions between more and less competent rocks (e.g. the tectonic boundaries of the quartzitic horizons).

Sliding and low angle thrusting took place during and after the early fold phase. These structures are seen both intraformationally and bounding the lower thrust unit.

These structures were refolded by more or less open folds trending mostly NW-SE and about NE-SW, which may or may not have been formed during one and the same folding phase.

Regional Implications

Considering the paucity of recent published data on the Caledonides of Troms and neighbouring regions, and the considerable differences between the provisional results of the present mapping and those of other workers in much of the region, it may be rather premature to try and draw far-reaching correlations. However, it seems appropriate to make a few suggestions which can be tested and confirmed or disproved by degrees as the mapping of the region becomes more complete, and isolated areas are linked up. Kulling (1964) has also attempted some broad correlations, between N Sweden and N Norway.

There seems little room for doubt that at least one early period of isoclinal folding with accompanying thrusting has taken place on a fairly large scale in the present area. This folding period (if only one) can scarcely be confined to an area of 700 sq. km or so, and must therefore be identifiable in the surrounding region when the right localities are hit upon and when more continuous mapping of individual horizons is accomplished. Ball, Gunn, Hooper and Lewis (1963) have recognized a similar fold episode (F_1) in the Loppa-Øksfjord district, further NE. The Bedford College (London) group working in the Sørøy region, still further NE, have also identified an early recumbent folding phase (see e.g. Dr. B. A. Sturt's comments (pp. 107-8) in the discussion after Rutland and Nicholson 1965).

Thus an implication from the present discoveries and those in Øksfjord and Sørøy is that there has probably been an early isoclinal folding phase in the Birtavarre area also. This implies that stratigraphical repetition, unrecognized during the Birtavarre mapping, may be present there. My own observations, from a short excursion in the Abmelašvag'gi profile, E of Manndalen in the Birtavarre district, reveal that the Big Limestone succession there is indistinguishable from the Plateau Limestone succession described in this paper. Furthermore, the very characteristic calc-silicate schist of the Norddalen Schist is the dominant rock type in the Guolas Limestone in that profile. The hazard a guess at which units may in fact be repetitions of each other in the Birtavarre area, I would suggest that the Schists-with-thin-Limestones may be equivalent to the Big Limestone.

If we look further afield, to the SW, we may cite a calc-silicate schist (etc.) succession of exactly the same appearance in the field to that in the Norddalen Schist, found in considerable thickness in the E side of Kirkesdalen (near Bjørkås).

We may now look at the "sparagmite" problem. Padget (1955) named rocks in Skibotndalen, that I have called banded phyllonite, Sparagmite Schists, and correlated them with the Eocambrian feldspathic quartzites which are supposedly extremely extensive in their metamorphic state, from the N coast of Finnmark through Sweden and Norway for hundreds of kilometres southwards. The lower quartzite zone of the Målselv area (Landmark 1959) is believed to represent the Eocambrian Sparagmite (Landmark, personal communication). As stated earlier (p. 236) Skjerlie and Tan (1961) have also proposed the same correlation for similar rocks in inner Reisadalen. Kulling (1964, pp. 127-8, 161-2) has cast doubt on Padget's and Skjerlie and Tan's correlations, and even on the identification of the original, pre-metamorphic, material, stating that apparently identical rocks in the Pältsa area (among elsewhere) of N Sweden consist of fine grained gneisses, granites and syenites. He states furthermore, that elsewhere in N Sweden there are fine grained overthrust feldspathic rocks which are leptitic gneisses, granites, and sediments of the Precambrian Sjöfall Series. Kulling then suggests that the above sparagmitic schists of inner Troms are older than Eocambrian and probably mostly or wholly non-sedimentary in origin. He partly uses the evidence of the greenschists described within the sparagmitic schist by Skjerlie and Tan to deny a Caledonian age for the unit, and it may be remarked that these greenschists could perhaps be represented in Skibotndalen by the basic phyllitic belt. I have not yet, as stated previously, found any basis for definitely determining the original character of the major varieties of the phyllonitic rock. Though various features suggest that it may be a meta-sedimentary rock, the only slight variation in plagioclase composition and the presence of allanite may point to an acid eruptive parent rock. If this large area of problematic rock is really of pre-Eocambrian age the implication is that the same applies to the rest of the adjoining belt to the NE (e.g. Strand 1960, pp. 270-1) which is generally correlated with sedimentary Eocambrian rocks of eastern N Norway and S Norway.

On the basis of this generally accepted Norwegian theory, Ramsay and Sturt (1963) have accepted an Eocambrian age for the meta-sedimentary succession of Sørøy.

The lowest part of this, termed the Klubben Quartzite, is white and grey, partly feldspathic, quartzites and quartz schists, with micaceous layers, and bears some similarity to rock seen on the E side of Lyngenfjord, NE of Kvesmenes (Landmark, personal communication) and at the head of Ullsfjord. This latter possibility has also been thought likely by Dr. D. Roberts (personal communication) who has made a close study of the Klubben Quartzite. Part of the Ullsfjord succession has been described by Randall (1959) under the name of Quartzite Series and this as he admits is scarcely distinguishable from his Blue Schist Group in this part of his area.

Green, fine grained, tectonized quartzo-feldspathic rock similar to horizons in the Lower Allochthonous unit, is also present at the head of Ullsfjord. At the S end of Nord Fugløy (Binns, in preparation) is a quartzitic succession that bears strong resemblance to these rocks in Lyngenfjord, Ullsfjord and on Sørøy.

Well over 2000 m of the Upper Allochthonous rocks separate the Lower Allochthonous phyllonites from these quartzo-feldspathic schists and quartzites of Lyngenfjord and Ullsfjord. If the Lower Allochthonous unit does not prove to be wholly of Precambrian age the thrusting and overfolding recognized in the area described in this paper, may enable this to be correlated with these quartzitic rocks. Only a more extensive field and laboratory study can confirm this.

Acknowledgements

I am especially indebted to Mr. K. Landmark, the head of this department, for assisting me in many ways, and not least for stimulating and helpful discussions on the geological problems of Troms. I am also grateful to Drs. B. A. O. Randall, and D. Roberts, for valuable discussions. Field expenses have mostly been defrayed by the Norwegian Geological Survey (N.G.U.) which has also carried out two X-ray determinations. I am most grateful for this assistance.

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A preliminary note on the geology of the area between Altevatn and Målselva, Indre Troms, N. Norway

By

Feiko Kalsbeek[1]) and Niels Østerby Olesen[1])

Abstract

An area of approx. 1,600 square kilometres in Indre Troms, N. Norway, has been geologically mapped. The area is built up of 1) granites and migmatites of the basement, overlain by 2) non-metamorphic sediments of the Hyolithus zone, on which, in overthrust position, rest 3) metamorphic Caledonian rocks. The overthrust Caledonian rocks have been divided into three 'sequences', which may represent different thrust sheets. The most common rocks and structures found in the three sequences are given in table I, and illustrated in the profiles of figures 2 and 3. Fig. 1 shows a sketch map of the area.

Introduction

During the summers of 1965 and 1966 an area of approx. 1,600 square kilometres in Indre Troms, N. Norway, was geologically mapped by students of the Geological Institute of Aarhus University. This mapping formed part of an exploration program of A.S. Sydvaranger, mining company[2]). Although the mapping is not yet completed (we reckon to be able to finish the map in 1967, while local detail problems probably will require prolonged fieldwork) it seems worth-while to report some of the results, arrived at during the first two summers, in this note.

The investigated area, which lies in northern Norway, is located between latitudes 68°30′ and 69° and between longitudes 19° and 20°15′, approx. 120 km SSE of Tromsø and 80 km ENE of Narvik. The area mapped until

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[1]) Geologisk Institut, Aarhus Universitet, Aarhus, Denmark.

[2]) Our sincere thanks are due to A.S. Sydvaranger for the opportunity it gave us to investigate this interesting area. Although, of course, the company was mainly interested in the economical aspects of the investigations, we had complete freedom to pursue our own scientific interests. Especially we appreciate the cooperation of Hr. bergingeniør Andreas Eriksen of A.S. Sydvaranger and his kind help in many matters of technical importance.

now is limited to the south by Altevatn, to the east by the Norwegian-Swedish boundary, to the north by Rostaelva and Målselva, and to the west by a line through Kirkesdalen to the northwestern tip of Altevatn.

Geologically speaking, the area mapped is situated in the marginal zone of the Caledonian mountain chain, where the rocks of the Caledonian chain overlie the granites and migmatites of the Precambrian basement.

As a topographical base for the mapping the maps on a 1 : 50,000 scale of Norges Geografiske Oppmåling and aerial photograps on a scale of approx. 1 : 20,000 of Widerøe's Flyveselskab were used.

The mapping was done at a rate of approx. 200 km^2 per man per summer lasting 2½ months. At this rate, a fairly good covering of most of the area is possible because, in general the stratigraphy of the area is relatively simple and because rather large parts of the area are almost completely covered by Quaternary deposits.

Apart from the two authors of this report, the following students of the Geological Department of the Aarhus University took part in the mapping: Holger L. Andersen and Peter B. Sørensen in 1965, Arvid H. Mortensen and Niels F. Schrøder in 1966. All of these contributed essentially to the results reported here.

Mainlines of tectonics and stratigraphy in the investigated area

Fig. 1 gives a very simplified geological sketch of the area mapped. The most southeastern part of the area is mainly built up of Precambrian granites and migmatites (in the following shortly called the 'basement'). On top of these granites etc., non-metamorphic sediments (arkoses, conglomerates, quartzites and shales) of the Hyolithus zone, up to approx. 150 m thick, occur. On top of these non-metamorphic sediments and separated from them by a clearly tectonic contact, a thick sequence of metamorphic rocks (a.o. mica-schists, amphibolites and metagabbros, marbles, quartzites) is found. This sequence attains an estimated thickness of approx. 2,500 m in the central part of the area.

The fact that these schists (often containing kyanite and staurolite), and other fairly high-grade metamorphic rocks, lie on top of non-metamorphic sediments, proves that the former have been thrusted on top of the latter.

In the southern part of the area the top of the basement, with its cover of Hyolithus zone sediments, dips gently (estimated average dip approx. 2°) towards NW, and disappears under the surface. In Dividalen, the basement and Hyolithus zone sediments appear on the surface again in an approx. 9 km

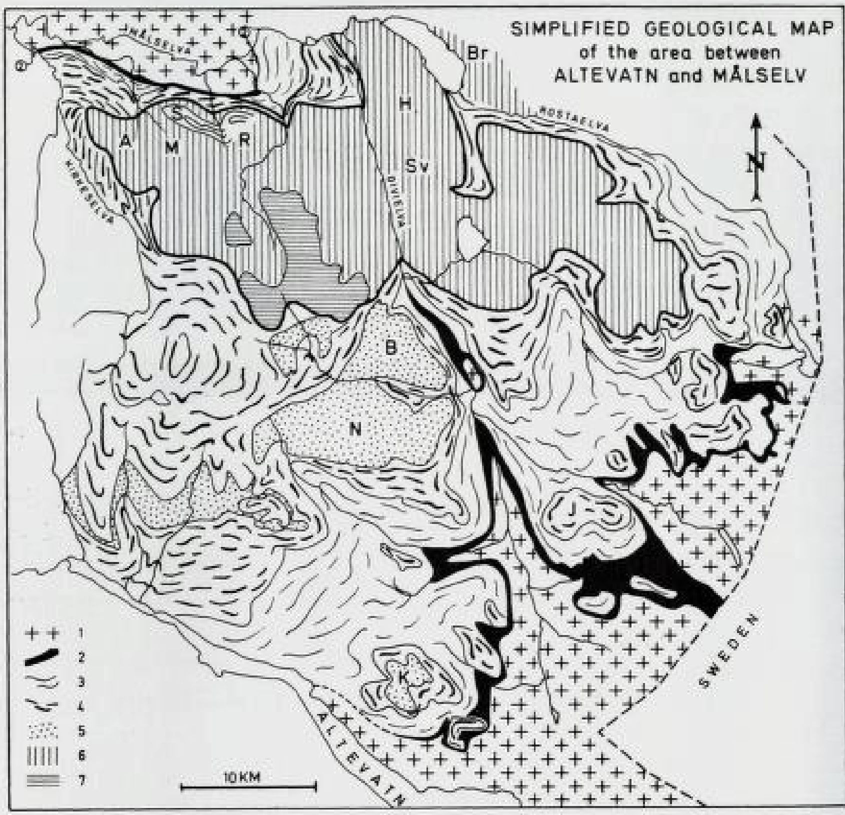

Fig. 1. Geological sketch map of the investigated area.

1. Basement. Mainly granitic rocks and migmatites.
2. Hyolithus zone sediments.

3-5. Rocks of the lower sequence. 3. Low grade rocks (without visible biotite), mainly kakirites, phyllonites and mylonites: probably normal low grade sediments are also present. 4. Rocks of higher metamorphic grade (with biotite). 5. Main outcrops of amphibolite and meta-gabbro.

6. Rocks of the middle sequence.
7. Rocks of the upper sequence.

The signatures for the rocks of the lower sequence roughly indicate the outcrop pattern of the formations mapped. It should be noted that this outcrop pattern is mainly determined by the topography of the area, the bedding mostly being subhorizontal.

Topographical abbreviations: A: Alappen. B: Bangfjellet. Br.: Brattlifjellet. H: Habafjellet. J: Jerta/Litle Jerta. K: Kistefjell. M: Middagsfjell. N: Njunis. R: Ruten. S: Storfjell. Sv.: Svortberget.

long window[1]), then, towards N, they disappear under the valley bottom for a distance of some 20 km, reappearing again in a large window on both sides of Målselva. This window has been described by Berthelsen (this issue) and has been called the Mauken window. Also along Målselva Hyolithus zone sediments have been found on top of the basement rocks, but here they are much thinner (approx. 20 m), and generally also more strongly tectonized than in the southern part of the area. We have good reasons to assume that the Hyolithus zone sediments in the Mauken window and those in the southern part of the area form a continuous horizon, and this would mean that the overlying metamorphic rocks form a continuous sheet, thrust over the underlying sediments for a distance of at least 50 km.

Apart from the main thrust plane between the Hyolithus zone sediments and the overlying rocks, the existence of several other important thrust planes has been proved, and it is probable that many thrust planes of minor importance are present.

During our mapping we have subdivided the rocks in the area into a number of formations which can be arranged into three sequences, each of which locally shows clear tectonic contacts with the underlying rocks. It is possible that we are dealing with three different thrust sheets or nappes. Locally, there is a clear discordance at the bottom of these 'nappes', but elsewhere the contacts with the underlying formations are concordant, and then it is difficult to verify whether one has to do with a normal or a tectonic contact. For the time being, it seems best to call these units neutrally the lower, the middle and the upper sequence. Table I gives an insight into the lithological character and the tectonic style of these three 'sequences'.

In the following pages a more detailed discussion of the stratigraphy and structures within the lower two sequences will be given.

The lower sequence

On Jerta and Litle Jerta mountains in the SE part of the area (indicated with J on the map of Fig. 1) a good section through the lower part of the lower sequence is found.

The following units (formations) can be mapped.

(8) quartzitic schists).

7) 2nd amphibolite, with bands and inclusions of schists and marble.

6) (garnet-) micaschists, fairly coarse grained.

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[1]) This window has been described by Gustavson (1963), but Gustavsons accompanying sketch map, in some respects, does not agree with our findings.

Table I. Main lithological and structural features of the rocks in the three 'sequences' of Caledonian overthrust rocks.

III Rocks of the Upper sequence Folding, on a minor scale, rarely observed.	At the base approx. 100 m micaschists, marbles and amphibolites. Thereover mainly garnet-micaschists, in the lower few hundred metres without marble, higher up with marble bands. Amphibolites rare. Max. exposed thickness approx. 800 m.
— — Thrustplane locally evident — — —	
II Rocks of the Middle sequence Large scale isoclinal recumbent folds with axes trending WNW-ESE. Lineation, with WNW-ESE trend, common. Refolding on N-S, NW-SE and NE-SW trending axes.	Garnet-micaschists (with local occurrence of staurolite and kyanite), marbles and quartzites. Amphibolites rare. Max. thickness approx. 900 m.
— — Thrustplane locally evident — — —	
I Rocks of the Lower sequence Minor thrusts common. Large scale thrusts possibly present. Isoclinal and open folds, on a minor scale, with axes trending in varying directions. Large scale open folds mostly with N-S and WNW-ESE trending axes. Lineation common in WNW-ESE direction.	Granite kakirites, phyllonites and mylonites common in the lower part, together with low-grade schists of probable sedimentary origin in which locally lenses and bands of dolomite occur. Higher up predominantly micaschists and amphibolites. Marbles subordinate. Max. thickness approx. 1,000 m.
— — MAIN THRUSTPLANE — — —	
	Hyolithus zone sediments and basement rocks.

5) 1st amphibolite, strongly foliated and lineated, often containing 'augen' of feldspar and hornblende.
4) augen-schists and -gneisses.
3) (garnet-) micaschists.
2) low-grade schists, partly 'Hartschiefer'.
1) (granite-) kakirites and -mylonites.
0) Hyolithus zone and basement rocks.

In the southwestern part of the area the higher part of the lower sequence is well represented. A formation consisting of schists, marble and amphibolites, which can be correlated with formation 7 in the Jerta area, is here overlain by

10 etc) Several formations mainly consisting of micaschists.
9) 3rd amphibolite.
8) fine-grained micaschists.

The rock types in the different formations have not yet been investigated in detail and the origin of some of them is as yet unclear. The low-grade schists (2) are at least partly of mylonitic origin (Hartschiefer being common), but some seem to be normal low-grade metamorphic sedimentary rocks[1]). The (garnet-) micaschists of the formation 3 locally still have phyllonitic characteristics but those of the formation 6 and higher schist formations are certainly normal metasediments. The augen-schists and gneisses (4) have strongly mylonitic textures, but contrary to the rocks in the formations 1 and 2 they are completely recrystallized. The rocks may contain large amounts of alkali feldspar, and some of them might be orthogneisses. The origin of the 1st amphibolite is not evident but in the amphibolites of the formation 7, and locally in those of formation 9, remnant igneous textures have been found.

In the Jerta area the formations dip towards NW, with a dip steeper than that of the underlying Hyolithus zone sediments and the top of the basement. Towards NW the high-grade formations thin out and disappear above (or into) the formation of the low-grade rocks (see Fig. 2). The latter attain a thickness of approx. 1,000 m in the area N and NW of Jerta.

At many other places in the area the same stratigraphical succession as on Jerta and Litle Jerta has been found. The amphibolite sheet which forms the top of Kistefjell (K on the map of Fig. 1), for example, is underlain by augen-schists and -gneisses and must, therefore, probably be correlated with the 1st amphibolite of Jerta.

In Sandelvdalen, between Njunis and Bangfjell (indicated with N and B resp. in Fig. 1) the whole succession of Jerta is again found. Here, however, the '2nd amphibolite' attains a thickness of up to 500 m, or more, and builds up most of Bangfjell and the Njunis massif. In the amphibolites of Njunis and Bangfjell the igneous (gabbroic) textures are locally well preserved, and it is clear that one is dealing with rocks of intrusive origin.

Also in the northern part of the area, both in Dividalen and south of the Mauken window along Målselva, the Jerta stratigraphy is locally found back. Especially the characteristic association of augen-gneisses with (augen-) amphibolites often serves as a stratigraphic marker. Often the Jerta stratigraphy is not complete, one or several formations being very thin or absent.

As said before, the high-grade metamorphic formations of Jerta and Litle Jerta disappear to the northwest, but some 10 km to the north, on the eastern side of Dividalen, they reappear again at an altitude of approx. 900 m, with fairly steep NW dips (approx. 40°). This repetition of the stratigraphy - and a few analogous cases - are possibly due to thrusting of the high-grade forma-

[1]) Even in thin section it is often difficult to see whether one is dealing with normal phyllites or with rocks of mylonitic origin.

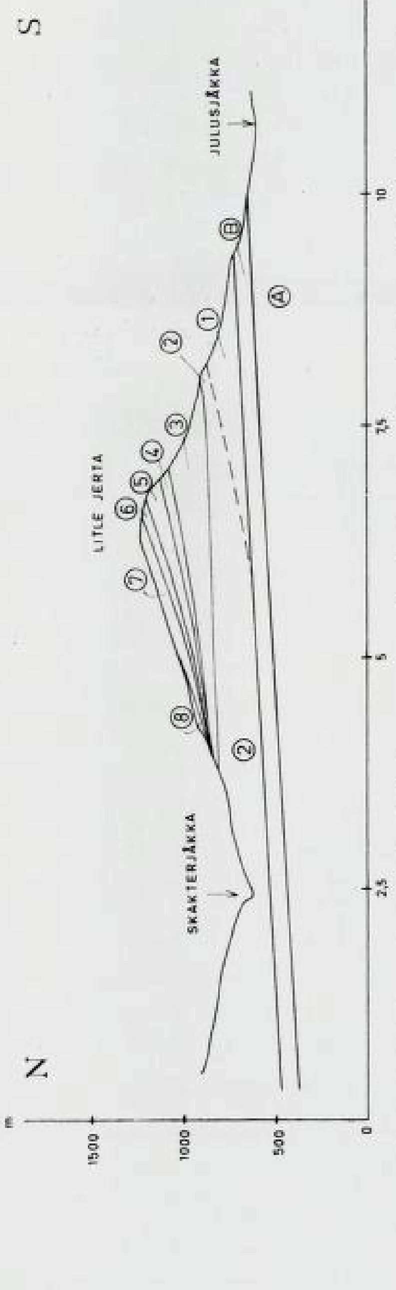

Fig. 2. Profile through the rocks of the lower sequence in the Jerta area. Note that the vertical scale is exaggerated two times.
A: Basement rocks. B: Hyolithus zone sediments.
1-8 overthrust Caledonian rocks, subdivided into the different formations described on p. 255-257.

tions over the low-grade rocks. The wedging out of the high-grade metamorphic formations on Jerta and Litle Jerta (Fig. 2) is also possibly caused by thrust movements. There are several indications that intensive thrusting has taken place at many places in the lower sequence, such as the common occurrence of tectonic discordances and mylonite zones, and the local occurrence of granite kakirites (thrust) on top of meta-sedimentary rocks (marbles and normal mica-schists).

Furthermore, if normal low-grade sediments do occur in the formation 2 of the Jerta stratigraphical section, this might indicate the overthrusted position of the more high-grade rocks which overlie them (comp. Kulling's thrust between his Abisko and Seve-Køli nappes (e.g. Kulling, 1964) and Gustavson's thrust between his Storfjell and Rombak groups (Gustavson, 1966)). In the field the presence of a thrust plane between the low-grade schists and the more high-grade schists is only rarely evident. Generally there seems to exist a gradual transition between the two rock types.

The pattern of folding in the rocks of the lower sequence is complicated. Large scale folds occur locally, but generally the different formations lie subhorizontal, or slightly inclined, over large distances, and do not show signs of large scale folding. On a minor scale both isoclinal folds and more open folds are commonly present. The axes of these folds have low plunges in almost any direction. It has been observed in many instances that the foliation of the schists etc. is an axial plane foliation belonging to some of the isoclinal folds. In other cases, however, the foliation itself is folded, both in isoclinal and in more open folds.

Several large folds with amplitudes of up to several km have been mapped. These have subhorizontal axes with WNW-ESE or N-S trends. In both cases it is the foliation of the schists etc. that has been folded - only locally a new foliation starts to develop in relation to the folds.

Unlike the measured fold axes in the lower sequence, the measured lineations trend WNW-ESE with few exceptions. Commonly there is a clear connection between this lineation and local folds with WNW-ESE axes. Often the lineation is a distinct mineral lineation. In the amphibolites of the formation 5 of the Jerta section, for example, a distinct hornblende lineation in WNW-ESE direction is common. In view of these facts we do not believe that the lineation in this direction is due to the thrusting of the metamorphic rocks over the basement and Hyolithus zone rocks, (comp. Kvale, 1953). It is evident that the metamorphism of the rocks must have taken place before they were thrust on top of the non-metamorphic Hyolithus zone sediments, and a lineation due to the orientation of metamorphic minerals must therefore be older than these thrust movements. The common occurrence of the lineation in rocks folded on WNW-ESE axes indicates a relation between the folding and the lineation.

The middle sequence

The middle sequence consists of metamorphic rocks whose sedimentary origin is evident. Micaschists, marbles and quartzites are the most common rock types. It has been possible to divide the sequence in a number of formations, which can be mapped through parts of the area. Due to the complex structures and to the fact that several of the formations are lithologically more or less identical, however, mapping of the formations is difficult and correlations between different parts of the area sometimes hazardous. The best defined formation in the middle sequence consists of homogeneous, thick-bedded, approx. 200 m thick quartzites, having fairly sharp contacts with the underlying and overlying formations. This quartzite formation can be mapped over long distances, and it is largely due to these quartzites that we have come to a fairly good understanding of the structures in the middle sequence.

The middle sequence seems to be built up of a pile of large scale isoclinal recumbent folds with subhorizontal WNW-ESE trending axes. The quartzite formation very clearly marks one of these folds (see Fig. 3). The hinge of this fold runs through Ruten mountain (indicated with R in Fig. 1), and toward ESE. N of Rostadalen thick-bedded quartzites occur at two levels, separated by some 700 m of schists and marbles. We think that these quartzites form the limbs of a large recumbent fold (see Fig. 3).

Since the rocks of the middle sequence are completely recrystallized, original sedimentary structures are only seldom recognizable. In the schists relict graded bedding has been locally recognized, and in the quartzites structures have been found which possibly may be interpreted as remnants of original cross bedding. Both phenomena indicate that the quartzites in the upper limb of the isoclinal fold lie in their normal position, whereas those in the lower limb are overturned. This conclusion is based on only a few observations in strongly folded rocks and should therefore be treated with caution.

Also on Alappen and Storfjell (indicated with A and S in Fig. 1) large isoclinal folds with the same axial trends are found. Minor isoclinal folds and parasitic folds with axial trends between W-E and NW-SE, but generally in WNW-ESE direction, are common.

The large recumbent folds generally fold a preexisting schistosity of the schists, but a new schistosity, parallel with the axial planes of the isoclinal recumbent folds is often more or less clearly present. A distinct WNW-ESE trending mineral lineation in the different rocks is parallel with the axes of the isoclinal folds, and is obviously related with the isoclinal folding. In a few outcrops a strongly tectonized conglomerate has been found in which the pebbles are clearly stretched (to 5-10 times their original length) in the same direction in which also the mineral lineations are found.

In the middle sequence rather small isoclinal folds locally occur which are older[1]) than the large scale recumbent folds. The folds give rise to an axial plane schistosity in the rocks. The axes of these folds have variable trends.

Folds younger[1]) than the large recumbent folds are of common occurrence. These folds cause a crenulation of the earlier formed schistosity. They range in size from a few metres to several tens of metres. In most cases they are too small to show on a 1 : 50,000 map.

At first sight the axes of these late folds seem to scatter in all directions, but a more detailed investigation seems to justify their subdivision into three groups, with axes trending roughly in N-S, NW-SE, and NE-SW directions

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[1]) Arguments for these age relationships will be given in a later paper.

There are indications that the folds with N-S trending axes are the oldest, and those with NE-SW trending axes the youngest of these later folds. The folds have in most cases much the same style, they generally consist of a subhorizontal and a subvertical flank.

There is a good correspondence between the structures in the middle and the lower sequences. In both folding on WNW-ESE axes gives rise to a strong lineation in the rocks, but large scale recumbent folds on WNW-ESE axes have as yet only been found with certainty in the middle sequence. In both sequences folds have been found that are older than the folding on WNW-ESE axes. Younger folds with different axial trends are ubiquitous in both sequences. These folds may become quite large (up to several km) in the lower sequence, whereas in the middle sequence they are generally much smaller. Also the subdivision of the younger folds into several groups with different axial trends seems to be much the same in the two sequences.

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Lokalglaciasjon på Sunnmøre

(On the Mountain Glaciation of Sunnmøre, West Norway)

Av

Arne J. Reite.

Abstract

Terminal moraines deposited by cirque glaciers are described from Sunnmøre, West Norway.

Radiocarbon dating indicates that the fjord districts were free from inland ice in Allerød time. The mountain glaciation most likely occurred in Younger Dryas time, when the glaciation limit was situated 600 m lower than today.

Innledning

Arbeidet er utført ved Geologisk institutt, Universitetet i Bergen, og ved Norges geologiske undersøkelse. Jeg vil takke dem som har hjulpet meg under arbeidet, både ved stimulerende diskusjoner og med rentegning av kart og diagrammer. En særlig takk til dosent dr. philos. H. Holtedahl og professor dr. philos. U. Hafsten.

Kyststrøkene på Sunnmøre består av et undulerende platå, oppdelt av fjorder og sund. Hvor fjellene er høyere enn 500 m fins godt utviklede botner, særlig i tilknytning til den flere hundre meter høye brattskrenten innenfor strandflaten.

I de høyeste fjellområdene i midtre fjordstrøk har botnbreene erodert seg så langt tilbake at det bare står igjen egger og horn mellom tilstøtende botner. I lavere fjelltrakter og i indre Sunnmøre er botnutviklingen kommet kortere.

Randmorener avsatt av lokalbreer viser at glaciasjonsgrensen en gang i sen- eller postglacial tid ble senket tilstrekkelig til at botnbreer kunne dannes. Noen av bretungene har nådd helt ned til datidens havnivå og lagt opp store randmorener. Det er da oftest mulig på grunnlag av marine terrasser i morenene å foreta en datering av breframstøtet i forhold til bestemte strandlinjenivåer. Disse randmorenene vil bli beskrevet forholdsvis inngående, mens andre blir omtalt mer summarisk.

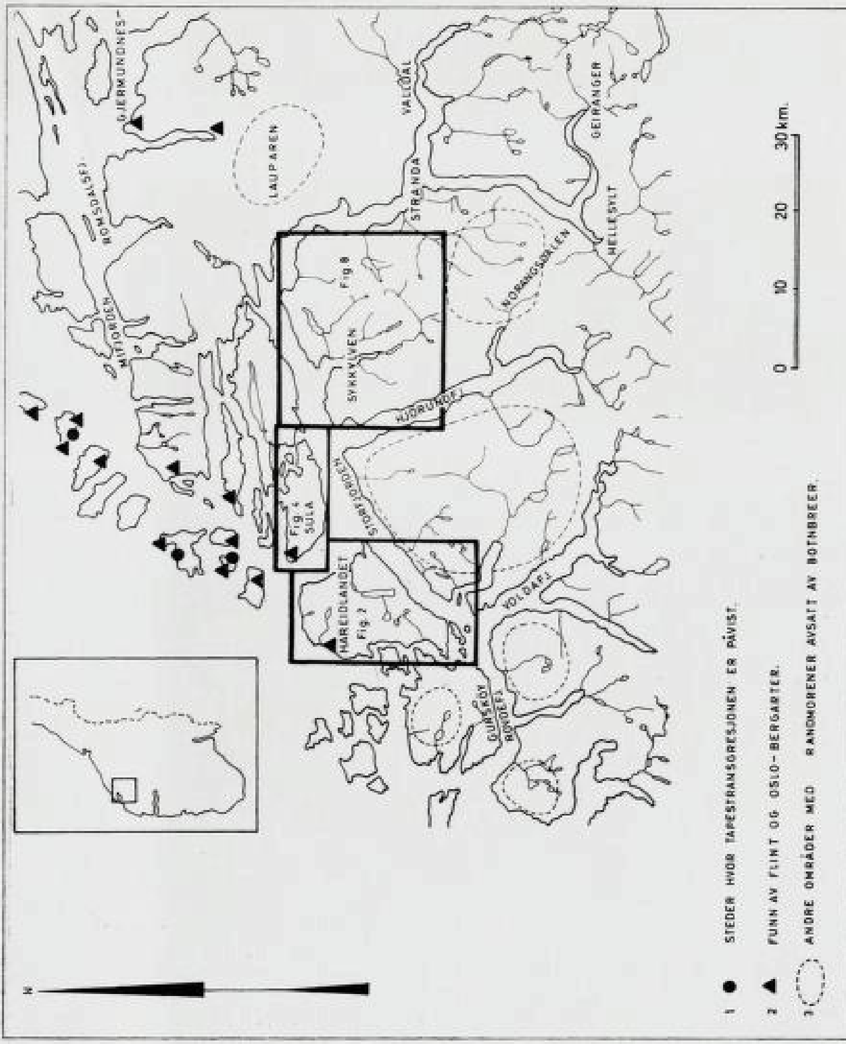


Fig. 1. Oversiktskart over Sunnmøre.

Sketch map of Sunnmøre. Legend: 1. Locality where Tapes transgression has been found. 2. Chert and Oslo rock material found below upper marine limit. 3. Other areas with terminal moraines deposited by cirque glaciers.

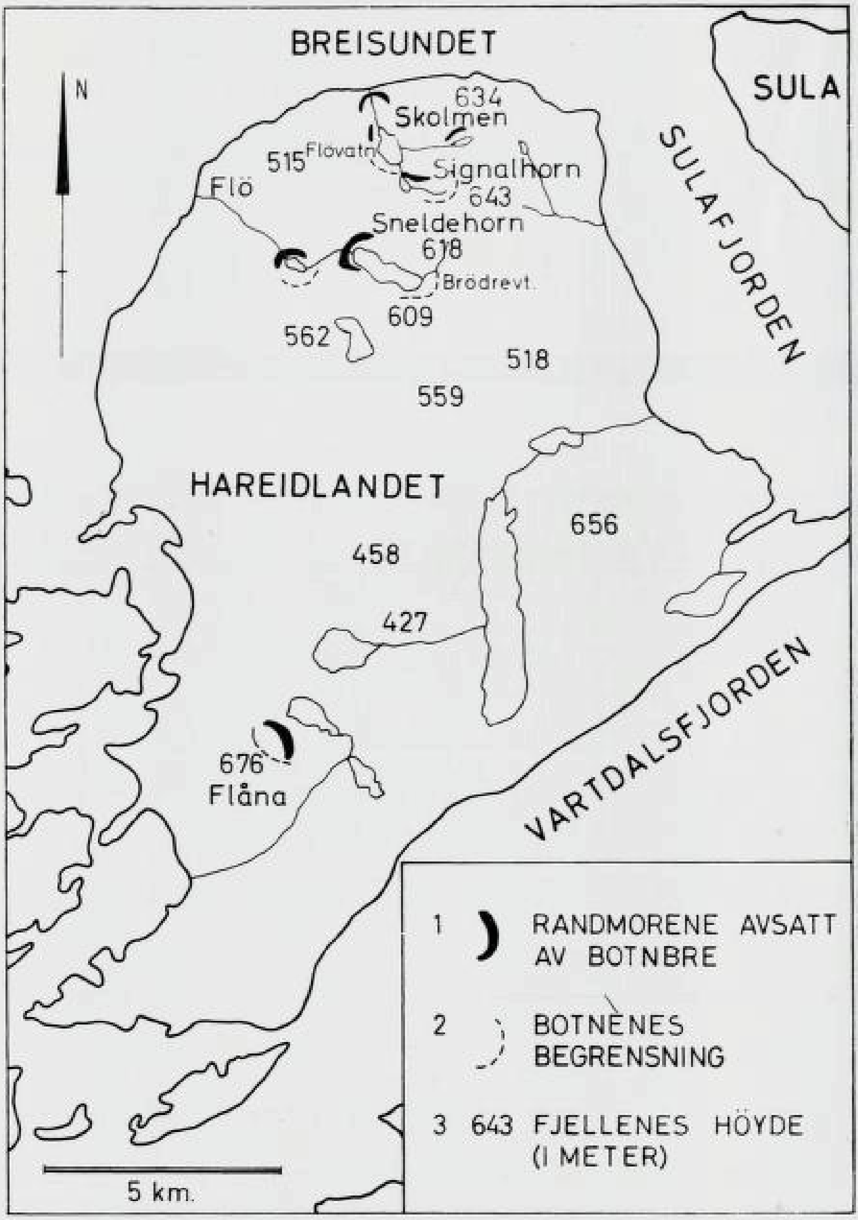


Fig. 2. Kartskisse over Hareidlandet.

Sketch map of Hareidlandet. Legend: 1. Terminal moraine deposited by cirque glacier. 2. Cirque. 3. Altitude of mountains (in metres).

Eldre arbeider.

De fleste kvartærgeologiske arbeider fra Sunnmøre behandler vesentlig terrassemålinger og skjellbankene, og omfatter derfor stort sett bare de deler av dalene som ligger under den senglaciale marine grense.

Kaldhol (1930, 1946) har omtalt randmorener avsatt av botnbreer i Sykkylven (fig. 1). Han mente at disse var avsatt i ratid, som han parallelliserte med Würm-istiden. Han var oppmerksom på at de høyeste marine terrasser i randmorener avsatt av lokalbreer er betydelig lavere enn ellers i dette fjordområdet, og påviste også et plutselig fall i den marine grense i indre deler av Storfjorden.

Heltzen (1948) har behandlet morfologien i Lauparenområdet på nordsiden av Storfjorden (fig. 1). Han finner randmorener avsatt av lokalbreer, men ingen av disse har nådd ned til havnivå.

Strøm (1956 har beskrevet randmorener fra Geirangertraktene, men det er ikke klart om disse er avsatt av botnbreer eller bretunger fra innlandsisen.

Holtedahl (1960) har omtalt en rekke randmorener avsatt av botnbreer på øyene på Mørekysten, og mener at de tyder på så lav glaciasjonsgrense at de neppe kan være avsatt senere enn i Yngre Dryas tid.

Regional beskrivelse

Randmorenene på Hareidlandet.

I dalføret innenfor Flø er to meget tydelige randmorener (fig. 2). Den nederste av disse demmer opp et lite vatn. Den 3—6 m høye, blokkrike moreneryggen kan følges sammenhengende rundt storparten av vatnet. Den andre randmorenen ligger i en bue rundt nedre del av Brødrevatnet og demmer opp dette (fig. 3). Morenen er avsatt av en bre som har ligget i en forsenkning omgitt av 600 m høye fjell.

Mellom Sneldehorn og Signalhorn er en randmorene som sannsynligvis er avsatt av en botnbre. Det samme er tilfelle mellom Signalhorn og Skolmen, men disse morenene er ikke så tydelige som de andre på Hareidlandet. Det ser også ut til å ha vært en bre i botnen hvor Fløvatn ligger. Bretungen har trolig nådd helt ned til havnivå.

Lenger syd på Hareidlandet er der en tydelig randmorene foran en liten botn på nordsiden av det knapt 700 m høye fjell Flåna.

Randmorenene på Sula (fig. 4).

På nordsiden av Sula har tre botnbreer rykket fram helt ned til havnivå og avsatt store randmorener. Den østligste av morenene ligger ved Langevåg. Sidemorenene kan følges nokså langt oppover mot en botn på nordsiden av Tverrfjell.

Fig. 3. Randmorenen ved Brødrevatnet.
The terminal moraine at Brødrevatnet.

Snitt i randmorenen viser at den til dels består av glacifluvialt materiale med regelmessige skrålag som faller mot Hessafjorden (fig. 5). Over det glacifluviale materialet ligger med meget skarp begrensning morenemateriale med 6 m mektighet. Det er ingen forstyrrelser i grenseflaten som ligger 22 m o.h. Kornfordelingen av materialet vil framgå av fig. 6.

Randmorenens beliggenhet viser at det ikke var en fjordbre i Hessafjorden på den tid botnbreen rykket fram. Det glacifluviale materialet kan være avsatt under innlandsisens avsmeltning eller det kan være et delta som ble bygget ut i en tid med voksende botnbreer i Tverrfjellområdet. Senere har breen rykket fram og glidd over en del av deltaet.

Vest for Sandvik har en bre fra en botn i den steile brattskrenten innenfor strandflaten nådd helt ned til havnivå. Sidemorenene er meget tydelige. Morenematerialets kornfordeling framgår av fig. 6.

I den vestligste sidemorenen er en tydelig abrasjonsterrasse 21 m o.h. De skarpt ryggformete sidemorenene som fortsetter ned til denne høyde viser at havet ikke kan ha stått høyere enn dette etter at randmorenen ble avsatt. Også utenom randmorenen er en meget tydelig terrasse i samme høyde.

Noen hundre meter lenger vest er der en randmorene av samme type (fig. 7).

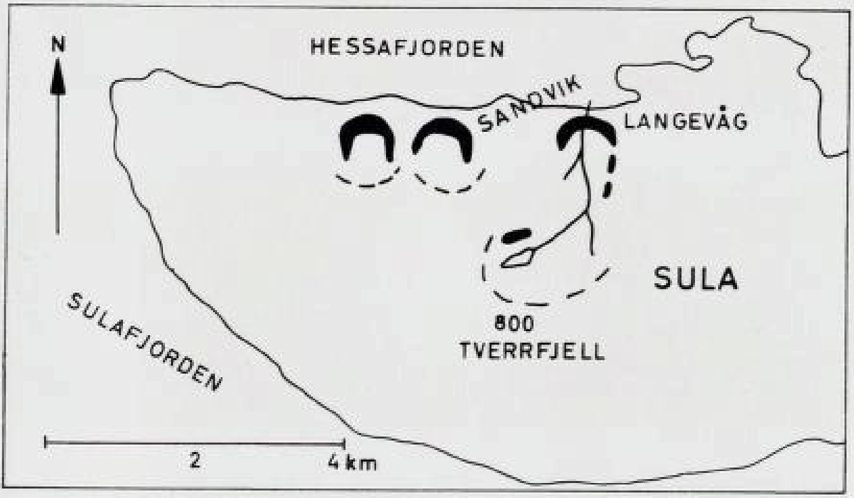


Fig. 4. Kartskisse over Sula.
Sketch map of Sula. Legend: See fig. 2.

Randmorenen har en markert ryggform og når ned til 24 m o.h. Dersom havet etter morenens dannelse hadde nådd så høyt som dette, ville sidemorenene uten tvil ha blitt utjevnet, da havet her står på med stor kraft.

Høyden av abrasjonsterrassen i den midtre randmorenen på Sula er omtrent den samme som høyden av det glacifluviale materialet i randmorenen i Langevåg. Dette kan tyde på at det glacifluviale materialet er et delta som ble dannet kort tid før breen fra Tverrfjell rykket fram.

Randmorenene i Sykkylven.

I dette området er tallrike, meget tydelige randmorener avsatt av botnbreer. Mange av breene har nådd helt ned til havnivå. Beliggenheten av morenene viser at det ikke var en fjordbre i Sykkylvsfjorden under breframstøtet.

Halvøya mellom Sykkylvsfjorden og Hjørundfjorden (fig. 8).

I dalføret innenfor Hundeidvik er en randmorene 2 km fra fjorden, avsatt av en bre som har kommet fra noen botner lenger inne i dalføret. Moreneryggen er 3—10 m høy og består av blokkrikt materiale. Den kan følges minst 500 m.

På nordsiden av Skopshorn finnes store randmorener foran botnene. Randmorenen rundt Sætsvatna er mindre tydelig enn de andre, men det skyldes i hvert fall delvis at det er ganske dyp torvmyr utenfor morenen. Inne i botnen

Fig. 5. Snitt i randmorenen i Langevåg.
Section in the terminal moraine at Langevåg.

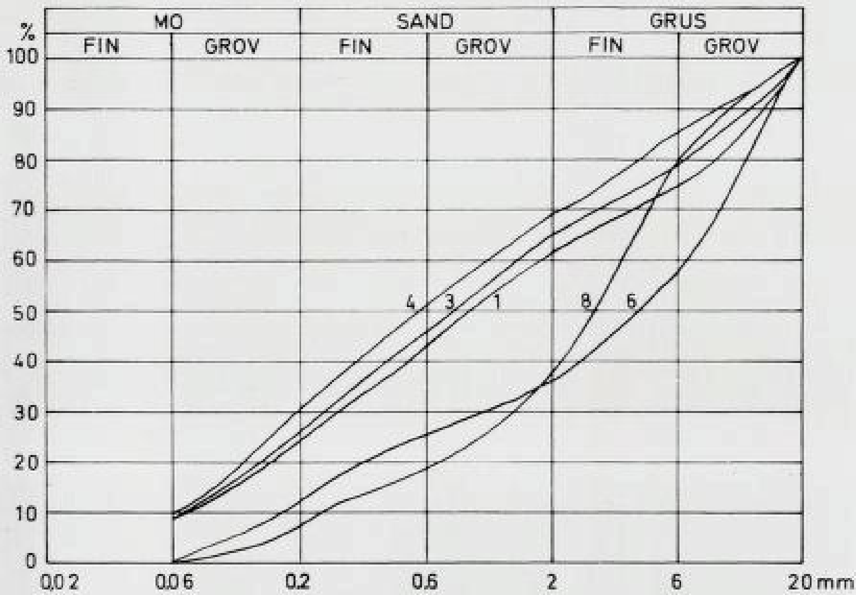
er sidemorenene tydelige. Botnbreens mektighet ser ut til å ha vært mer enn 100 m.

På Riksheim er en stor randmorene avsatt av en 6 km lang dalbre som har kommet fra noen botner helt over mot Hjørundfjorden. Sidemorenene kan følges til 400 m høyde. Videre innover dalen er dalsidene så bratte at morenematerialet ikke er blitt liggende. I dette dalføret er det tydelige skuringsstriper i NØ-lig retning. Godt utformete støt- og lesider viser at denne isbevegelsen har vært langvarig.

Randmorenen består for det meste av blokkrikt, grusholdig morenemateriale (fig. 6), men det er også en del linser av glacifluvialt materiale.

Utenfor den nordligste sidemorenen er et stort delta. Høyden av deltaets framkant er 49 m o.h., toppen på deltaet ligger 51 m o.h. Det består av grus og sand, lagningen viser et regelmessig fall mot øst. Materialtilførselen ser derfor ut til å ha vært fra Riksheimdalen. Deltaet grenser til sidemorenen, men da skikkelige snitt mangler er det ikke mulig å avgjøre om det fortsetter inn under denne.

Riksheimelva har nå skåret seg gjennom morenen. Syd for elva er det i randmorenen en stor terrasse som uten tvil er dannet ved at elva har planert



NR	STED	DYP	Md	So[1]	MERKNAD
1	RANDMORENEN I LANGEVÅG	3 m	0,95	1,39	
3	RANDMORENEN PÅ RIKSHEIM	2 m	0,80	1,47	
4	RANDMORENEN PÅ VELLE	5 m	0,60	1,30	
6	RANDMORENEN VEST FOR SANDVIK	1,5 m	4,7	1,12	NOE UTVASKET
8	LANGEVÅG GL. FLUV.	6 m	3	0,77	

[1] $So = \log \frac{Q75}{Q25}$

Fig. 6. Kornfordelingsdiagram.
Grain-size distribution.

morenen. Høyden av terrassen er 49 m o.h., og da dette stedet ligger like ved fjorden angir terrassen den marine grense den tid elva hadde sitt løp her. Akkumulasjonen av deltaet og dannelsen av elveterrassen har derfor trolig skjedd på omtrent samme tid.

Sidemorenene har en markert ryggform og fortsetter nedover til ca. 50 m o.h. Dette tyder på at havet ikke har stått høyere enn 50 m etter at randmorenen ble avsatt.

På Straumsheim er en randmorene avsatt av en bre som har kommet fra to botner på nordsiden av fjellet Trollkyrkja. I denne morenen er ingen tydelige terrasser.

Fig. 7. Den vestlige randmorenen på Sula.
The westernmost terminal moraine on Sula.

Sidemorenene er opptil 50 m høye og kan følges sammenhengende et par km til 400 m høyde. Lenger inne i botnen er en 200 m lang morenerygg som trolig er en del av sidemorenen.

Randmorenene i Velledalen (fig. 9).

I Velledalen er en rekke randmorener avsatt av botnbreer som har kommet fra de opptil 1600 m høye fjellene vest for dalen. På den tid breframstøtet skjedde, nådde en fjordarm langt opp gjennom dalføret.

Randmorenen på Fet er avsatt av en bre som har kommet fra en forsenkning i fjellsiden innenfor. Den nordligste sidemorenen kan følges til 500 m høyde. Inn mot fjellsiden mellom sidemorenene er en ganske stor forhøyning av storblokkig materiale, som trolig har rast ned fra en hengende botnbre etter at breen trakk seg tilbake fra sin maksimale utbredelse.

I randmorenen er en ganske tydelig terrasse. Kaldhol (1946) har ved nivellering bestemt høyden til 53—54 m o.h.

På Velle er en liknende randmorene som på Fet. Den ene sidemorenen kan følges til 600 m høyde. Kornfordelingen av materialet vil framgå av fig. 6.

Mellom Velle og Brunstad er tydelige randmorener avsatt av to breer som

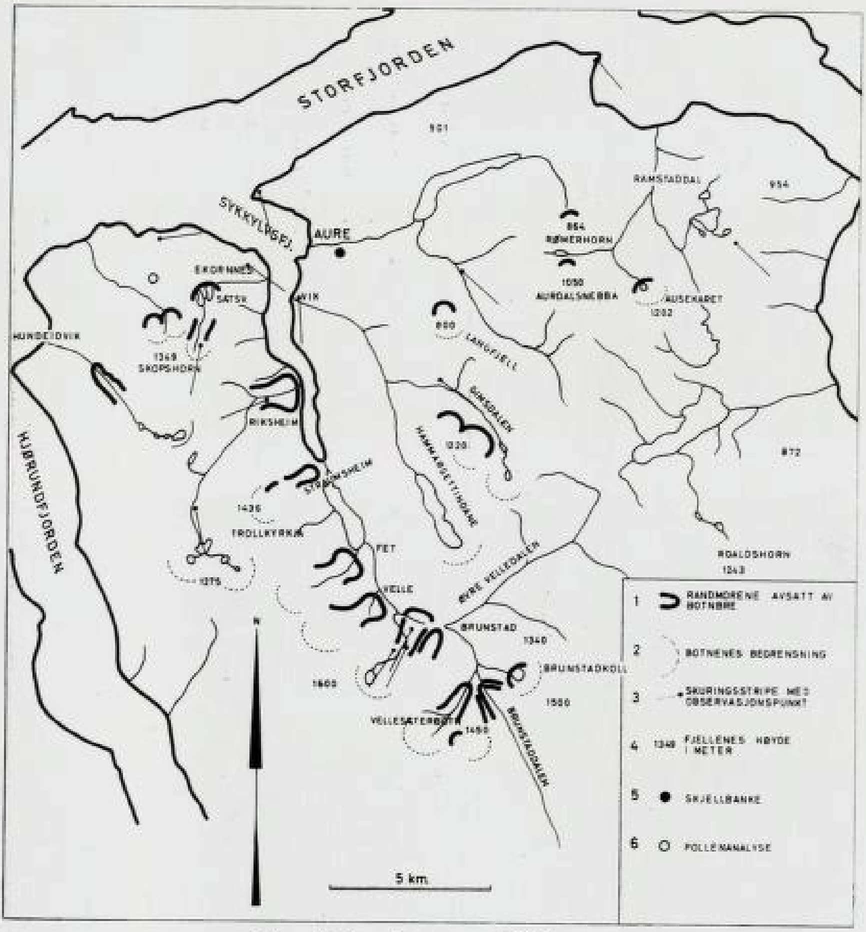


Fig. 8. Kartskisse over Sykkylven.

Sketch map of Sykkylven. Legend: 1. Terminal moraine deposited by cirque glacier. 2. Cirque. 3. Glacial striae with point of observation. 4. Altitude of mountains (in metres). 5. Sites of molluscan shells. 6. Locality for pollen analyses.

har rykket fram fra botner på dalens sydside. Breene har demmet opp et vatn i Øvre Velledalen (fig. 9). I dette isdemte og senere morenedemte vatnet ble det avsatt et stort delta. Deltaets største høyde er 115 m o.h., mens framkanten ligger 93 m o.h. Materialtilførselen har vært fra Brunstaddalen med sidedaler, hvor det på denne tid har vært flere breer ikke så langt fra deltaet. Det kan også ha vært en viss materialtilførsel fra Øvre Velledalen. Snitt i de distale deler av deltaet viser regelmessige skrålag av grus og sand.

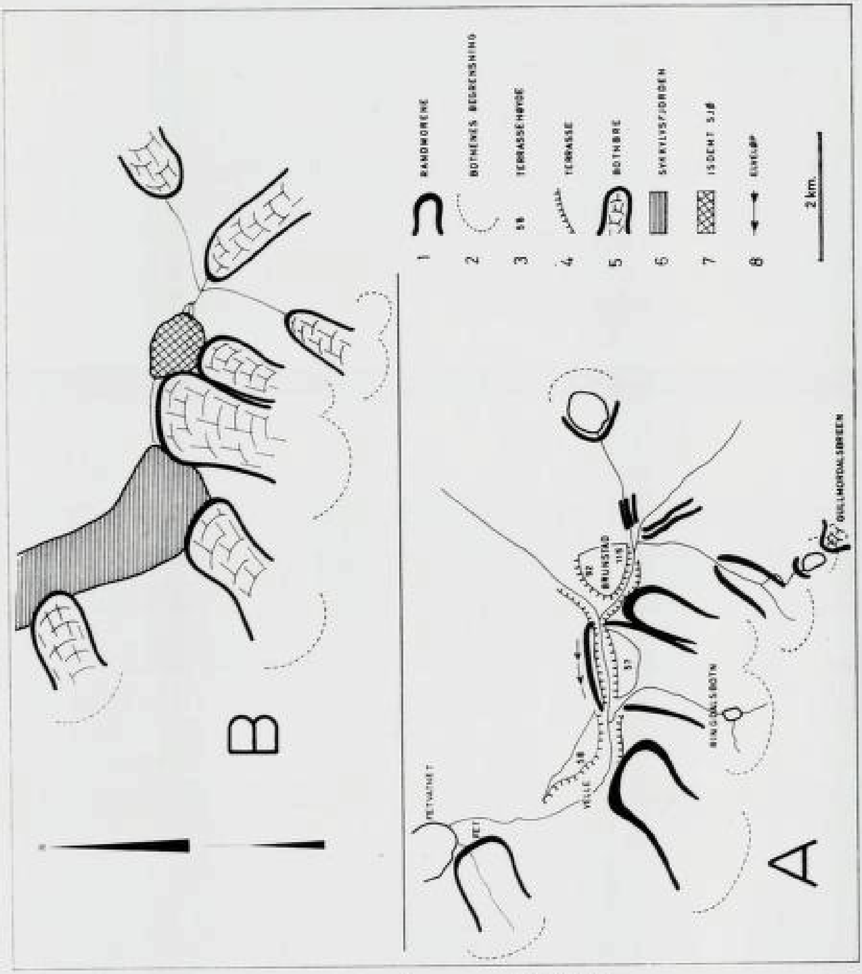


Fig. 9. A. Randmorenene i Velledalen.
B. Rekonstruksjon av forholdene i Velledalen under botnbreenes maksimale utbredelse.

A. The terminal moraines in Velledalen.

B. Cirque glaciers in Velledalen during the mountain glaciation.

Legend: 1. Terminal moraine deposited by cirque glacier. 2. Cirque. 3. Altitude (in metres) of terrace. 4. Terrace. 5. Cirque glacier. 6. Sykkylvsfjorden. 7. Glacier-dammed lake. 8. River channel.

Høyden av deltaet er for stor til at det kan være avsatt i havet, da den senglaciale marine grense ved Sykkylvsfjorden er lavere enn 80 m o.h.

På dalens nordside er en dyp renne i morenemateriale (fig. 9). Hvor rennen begynner er bredden 15 m og dybden 2—3 m. Lenger nede er bredden mer enn 50 m. Høyden ved innløpet er 97 m o.h., men rennen kan tidligere ha vært et par meter dypere da noe materiale kan ha raset ned fra dalsiden.

Rennen kan følges mer enn 1 km og ender i et stort delta på garden Velle. Et 30 m høyt snitt viser at materialet er grus og sand mens topplaget består av stein og blokker. Lagene har et regelmessig fall fra dalsiden utover mot dalbunnen, noe som passer godt med materialtilførsel gjennom rennen.

Bortsett fra materiale i suspensjon ville alt materiale elvene fra Brunstaddalen og Øvre Velledalen førte med seg bli avsatt i den isdemte sjøen. Materialtilførselen til deltaet på Velle må derfor ha vært fra breelver fra Ringdalsbreen, dessuten har elva fra den isdemte sjøen erodert kraftig i morenemateriale breen førte med seg.

På deltaet, som ligger 58 m o.h., sees noen grunne renner etter tidligere elveløp. Høyden angir på det nærmeste den marine grense på den tid avsetningen fant sted. Rester av glacifluvialt materiale langs dalsiden i samme høyde som deltaet tyder på at en del av det er blitt fjernet ved senere elveerosjon.

I Ringdalsmorenen er 57 m o.h. utformet en stor terrasse i morenemateriale (fig. 9). På den tid Sykkylvsfjorden nådde til Velle, måtte dette stedet ligge godt beskyttet mot marin abrasjon. Det er derfor mest sannsynlig at terrassen er dannet ved at elva har planert randmorenen. Høyden av terrassen er omtrent den samme som på deltaet på Velle. Dette tyder på at sidemorenen er blitt gjennombrutt av elva kort tid etter at breen trakk seg tilbake. Det kan også tenkes at havnivået har vært omtrent det samme i lenger tid.

Ringdalsbotnen er nesten helt renskrapet for løsmateriale. Skuringsstriper viser at isen har beveget seg i retningen N 25° Ø, og rundsva med godt utformete støt- og lesider viser samme isbevegelsesretning.

Brunstaddalen med sidedaler (fig. 8 og 9).

Nedenfor Brunstadsæter er en randmorene som består av flere rygger av storblokkig morenemateriale, særlig er den tydelig på dalens vestside.

Under Brunstadkoll er en storblokkig randmorene som går i en bue rundt vatnet (fig. 10). Denne morenen er trolig samtidig med lokalglaciasjonen ellers i området.

En randmorene i Vellesæterbotn viser at der har vært en stor botnbre. Begge sidemorenene er tydelig og kan følges et par km. Isens største mektighet har vært ca. 200 m.

Fig. 10. Randmorene under Brunstadkoll.
The terminal moraine at Brunstadkoll.

Sydøst for Vellesæterbotn ligger en liten randmorene som demmer opp et vatn. Da Gullmordalsbreen ligger like inne i denne trange dalen, kan randmorenen godt være fra et senere tidspunkt enn de andre. Den nesten helt manglende vegetasjon tyder på dette.

Området mellom Sykkylvsfjorden og Storfjorden (fig. 8).

I dette området er fjellene betydelig lavere enn vest for Sykkylvsfjorden, men også her fins randmorener som utvilsomt er avsatt av botnbreer.

I Gimsdalen har to breer rykket fram fra nordsiden av Hammarsettindane og avsatt tydelige randmorener. Lenger inne i dalen er en stor botn med mye storblokkig morenemateriale. Her er ingen tydelig randmorene, men også her har det trolig vært en botnbre.

Under den nordvestlige del av Langfjell er en 4—8 m høy randmorene som kan følges som en sammenhengende bue foran botnen. Fjellet innenfor morenen er bare 800 m høyt, og skilt fra den betydelig høyere sydlige del av Langfjell av en dyp forsenkning. Dette er laveste fjell i Sykkylven hvor det har vært en botnbre etter at innlandsisen smeltet bort.

En bre fra botnen nord for det 1200 m høye Ausekaret har avsatt en randmorene. Under Aurdalsnebba og det 864 m høye Rømerhorn er små rygger av morenemateriale. Beliggenheten av morenene tyder på at de er avsatt av botnbreer.

På sydsiden av Rømerhorn er store rygger som består av blokker og stein. Ryggene ser ut til å være dannet ved at blokker fra det sterkt frostsprengte fjellet innenfor har glidd nedover en snøfonn eller en bre. Avstanden fra ryggen til urda innenfor er ofte ca. 30 m. Fjellskråningen er sydvendt og snøen smelter nå bort tidlig om sommeren. Ryggene må derfor trolig være dannet under kaldere klima enn i nåtiden, og det er nærliggende å anta at de avsatt under klimaforverringen som førte til at botnbreene rykket fram.

Andre områder med randmorener avsatt av lokalbreer

Randmorener foran botner viser at det har vært lokalglaciasjon også i andre deler av Sunnmøre. De viktigste områder er tegnet inn på fig. 1.

I de sydlige deler av Sunnmøre er det randmorener etter botnbreer i området syd for Rovdefjorden og på Gurskøy. På halvøya mellom Voldafjorden og Hjørundfjorden er tallrike randmorener som viser at dette området har vært et viktig glaciasjonssentrum.

Det er også funnet sikre spor etter lokalglaciasjon i Norangsdalen og på Stranda (syd og sydøst for Sykkylven). Som tidligere nevnt er det funnet randmorener avsatt av botnbreer i Lauparenområdet på nordsiden av Storfjorden (Heltzen, 1948).

Da jeg ikke har hatt flyfotografier over de østlige deler av Sunnmøre og feltarbeidet der er blitt begrenset til de største dalførene, har jeg ikke kunnet avgjøre om det har vært lokalglaciasjon i dette området.

Tidspunktet for lokalglaciasjonen på Sunnmøre

Under regionalbeskrivelsen har jeg for en rekke randmorener ment å kunne fastslå havnivået på den tid breframstøtet skjedde. Det viser seg at det høyeste terrassetrinn i randmorenene faller sammen med en meget markert abrasjonsterrasse som kan følges sammenhengende lange strekninger. På øyene består dette terrassetrinn oftest av et belte av frittskylte, godt rundete steiner og blokker. Også lenger inne i Storfjorden er det som regel tydelig. Høyden av denne terrassen er på Sula 21 m o.h. Den stiger jevnt i østlig retning og er på Magerholm 36 m o.h.

Ved munningen av Sykkylvsfjorden er et markert terrassetrinn 44 m o.h. Dette trinn kan følges langs hele Sykkylvsfjorden, og er det høyeste terrassetrinn funnet i randmorener avsatt av lokalbreer. Høyden innerst i fjorden er 54 m o.h.

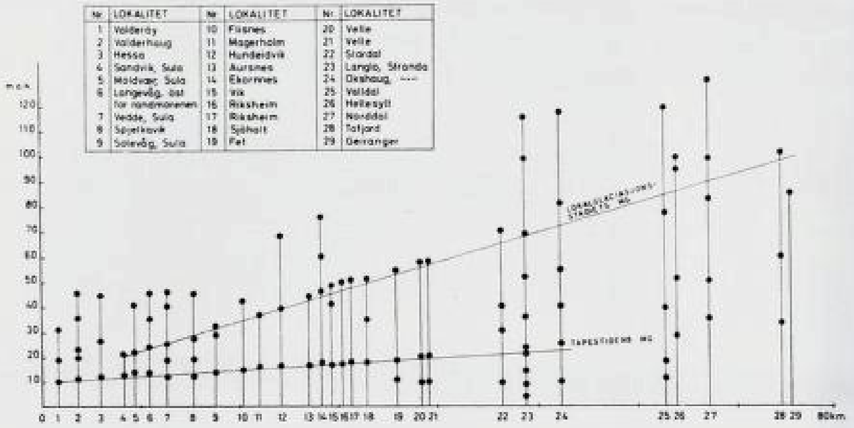


Fig. 11. Ekvidistant strandlinjediagram.
Equidistant shore-line diagram.

Dette terrassetrinn er uten tvil synkront og er derfor brukt til å bestemme isobaseretningen, som i dette området er N 37° Ø. Denne retningen stemmer godt med den en finner for et lavere markert terrassetrinn, trolig fra Tapestid.

Ekvidistansediagrammet (fig. 11) er framstilt ved at terrassehøydene innenfor et smalt belte er projisert på en linje med retningen N 53° V, dvs. vinkelrett på isobaseretningen.

Av ekvidistansediagrammet framgår at det er to terrassetrinn som skiller seg ut. Det laveste av disse har Kaldhol (1946) ment var dannet i Tapestid. Han bygger både på skjellfauna og forekomst av torv overleiret av strandvoller. Slike transgresjonsprofiler er beskrevet flere steder på Mørekysten, og et par nye lokaliteter er funnet under feltarbeidet (fig. 1).

Også lenger inne i Storfjorden er dette terrassetrinn meget markert, og det skulle ikke være noen fare for å forveksle det med andre trinn. Gradienten er bestemt til 0,25 m/km.

Det andre markerte terrassetrinn er det høyeste som fins i randmorener avsatt av lokalbreer. Som tidligere nevnt er dette terrassetrinn meget tydelig også utenom randmorenene. Da de høyeste terrasser i randmorenene på Sula og i Sykkylven faller på samme linje i diagrammet, tyder dette på at de er samtidige.

Når det gjelder randmorener som ikke når ned til havnivå, er det vanskelig å avgjøre om disse er samtidige med randmorenene langs fjorden, men da de synes å tyde på samme glaciasjonsgrense er det grunn til å anta dette.

Det er selvsagt meget vanskelig å avgjøre hvor lang tid det tok fra fjordene ble isfri til botnbreene trakk seg tilbake. Hafsten (1956) har funnet at den gjennomsnittlige strandlinjeforskyvning i Oslo-området i preboreal tid var 11 m/100 år. Nå er den marine grense der omtrent tre ganger så høy som i Sykkylven, så en strandlinjeforskyvning på 25 m må trolig ha tatt noen hundre år, kanskje betydelig lengre.

C-14-dateringer fra Bergenshalvøya (Holtedahl, 1964) og Aure, Sykkylven, synes å tyde på at kyststrøkene på Vestlandet ble isfri i Allerød tid. Dette gjør en slik sammenlikning enda mer usikker enn den ellers ville være, da en kjenner lite til de eustatiske og isostatiske forhold i tidsrommet Allerød — preboreal.

Drivistransportert materiale i Romsdalsfjorden og de ytre deler av Storfjorden.

På Gjermundnes ved Romsdalsfjorden (fig. 1) har Kaldhol (1912, 1946) funnet store mengder bergarter fra Oslo-feltet, særlig rombeporfyr og larvikitt. Han har dessuten funnet atskillig flint. Antallet av slike ledeblokker er ifølge Kaldhol minst 1000 bare på Gjermundnes. Flyttblokker av samme type er også funnet lenger inne i Romsdalsfjorden. Blokkene er funnet i sand og leir over morenemateriale, og bare lavere enn 50 m o.h.

På flere av øyene på Sunnmøre har jeg funnet flyttblokker av liknende type (fig. 1). Blokkene er for det meste funnet i fjæra. De består ofte av rombeporfyr, videre er kalksteiner, sandsteiner og larvikitt forholdsvis vanlige. Det er også funnet noen få blokker av polymikte konglomerater, utvilsomt fra devonfeltene i Sogn og Fjordane.

Oslo-bergartene og flinten må være transportert av isfjell. For transport over kortere strekninger kan nok vinteris være av stor betydning, men nordover langs Norges vestkyst er en slik transport lite sannsynlig.

Funn av Oslo-bergarter og flint i Romsdalsfjorden og de ytre deler av Storfjorden viser at fjordene var helt eller delvis isfri på en tid da innlandsisen nådde havet i en kalvingsfront i Oslo-området. Dette kan ikke ha skjedd senere enn i ratid.

Forholdet mellom den senglaciale marine grense og Tapestidens marine grense.

Nansen (1922) har stilt opp en empirisk formel for å beregne Tapesstrandlinjen når den senglaciale marine grense er kjent, eller omvendt å beregne den senglaciale marine grense når Tapesstrandlinjen er kjent:

$$h = H \times 0{,}315 + 3{,}4$$

hvor h er høyden av Tapesstrandlinjen og H er høyden av den senglaciale marine grense.

Nansen fant at denne formel var tilnærmet riktig langs hele Norges kyst. Største avvik var + 2,8 m og — 2,7 m, og han mente dette skyldtes unøyaktige målinger.

Kaldhol (1946) har funnet at denne formel stemmer meget dårlig for Møre og Romsdal. Ved å sette inn høyden for et lavere markert terrassetrinn (lokalglaciasjonsstadiets marine grense) i stedet for den senglaciale marine grense fant han at overensstemmelsen var god.

Grunnen til at Nansens empiriske formel ikke synes å gjelde i dette området kan være at fjordene på Møre ble tidligere isfri enn de områder Nansen hadde sine terrassemålinger fra. Det høyeste terrassetrinn vil da ikke være synkront, og en skal ikke vente samme enkle forhold mellom den marine grense i senglacial tid og i Tapestid.

C-14-datering av skjellforekomsten på Aure, Sykkylven (fig. 8).

I en skjæring øst for idrettsplassen på Aure fantes skjellførende leire under et stein- og blokkrikt leirlag. Finnestedet ligger ved det for lengst nedlagte Aure teglverk, hvor Kaldhol har beskrevet følgende snitt:

Sandlag	5 — 6 m
Leirlag, ingen fossiler, mengder av stein	1 m
Fossilførende leire	3 m
Leire uten fossiler	2 m
Sand, ukjent mektighet	

Snittet var nå dårlig, slik at bare den skjellførende leira og det stein- og blokkrike laget var synlig, men det er utvilsomt samme lokalitet som Rekstad (1907), Kaldhol (1908) og Øyen (1924) har omtalt.

Finnestedet ligger ca. 14 m o.h. Følgende arter mollusker ble funnet: Mya truncata, Cyprina islandica, Saxicava arctica, Pecten islandicus, Macoma calcaria og Astarte elliptica. Kaldhol og Rekstad angir en langt mer omfattende fossilliste, men under feltarbeidet ble det bare tatt sikte på å samle materiale til C-14-datering.

Foruten skjell har Rekstad av teglverksbestyreren fått tilsendt et trestykke av or som var funnet i leira. Dessuten er det funnet en rygghvirvel av storkobb (Phoca barbata), og Øyen har funnet et ben av spitsbergenalken (Uria arra).

Skjellene er for større arters vedkommende oftest oppknust, men stykkene ligger som regel på plass. De er godt oppbevart.

C-14-datering av en skjellprøve (Cyprina islandica gav resultatet 11620 ± 120 år før nåtiden. Forutsatt at dateringen ikke er beheftet med for store feil tilsvarer dette Allerød tid. Dateringen viser med det forbehold som er tatt ovenfor at Sykkylvsfjorden var isfri på denne tid. Da denne grunne fjordarmen er omgitt av meget høye fjell, er det grunn til å tro at hele Storfjorden var isfri.

Trestykket av or som ble funnet i leira kan ikke skrive seg fra vegetasjonen på stedet i Allerød tid. Det kan være drivved, men det er kanskje vel så sannsynlig at det har kommet inn i leira ved senere utglidninger.

Nå ligger det over leiravsetningen et delta som ser ut til å være avsatt mens havnivået var ca. 30 m høyere enn i nåtiden. Da Tapestidens marine grense her er antatt å være ca. 17 m o.h. (fig. 11), må deltaet være dannet på et tidligere tidspunkt.

Det stein- og blokkførende laget kan enten være morenemateriale eller marin leire med stein og blokker transportert av isfjell under lokalglaciasjonsstadiet, da tallrike botnbreer kalvet i Sykkylvsfjorden.

Øyen (1924) har omtalt basalplater av balaner på stein som fantes i leira, men det framgår ikke klart om dette også gjelder det øvre stein- og blokkrike leirlaget. Han har som Kaldhol funnet en sterk økning av blokkinnholdet i den øverste meter av avsetningen, men han har ikke tatt noe standpunkt til om det er marin leire eller ikke.

På den tid skjellene ble samlet inn var det ikke mulig å foreta større gravinger, da skråningen skulle dekkes med grastorv. Jeg antok da at det stein- og blokkførende laget var morenemateriale (Reite, 1963).

Dersom en bre i Sykkylvsfjorden hadde nådd til Aure skulle en vente å finne markerte randmorener langs fjorden, særlig hvis det hadde vært et kraftig breframstøt. Randmorener fra en fjordbre mangler, men det er store mengder morenemateriale til Ekornnes. Morenematerialet kan imidlertid være avsatt under innlandsisens tilbaketrekning. Den høye marine grense langs Sykkylvsfjorden bortsett fra i randmorener fra lokalbreer tyder også på at det ikke har vært noe sent breframstøt til Aure.

Det er derfor mest sannsynlig at det stein- og blokkførende laget er marin leire med grovere materiale transportert av isfjell. Oppknusingen av skjellene kan ha skjedd ved senere utglidninger i leirmassene. Dette kan også forklare at et trestykke av or er blitt rotet inn i leira.

Øyen henførte fossilene til sitt Mytilusnivå, mens Kaldhol mente de var fra 1. interglacialtid, noe som ikke viste seg å være tilfelle.

Pollenanalyse.

For om mulig å komme fram til tidspunktet for innlandsisens bortsmeltning er det blitt tatt prøver for pollenanalyse. Sunnmøre ligger langt fra områder hvor det er foretatt omfattende pollenanalytiske undersøkelser. En sammen-

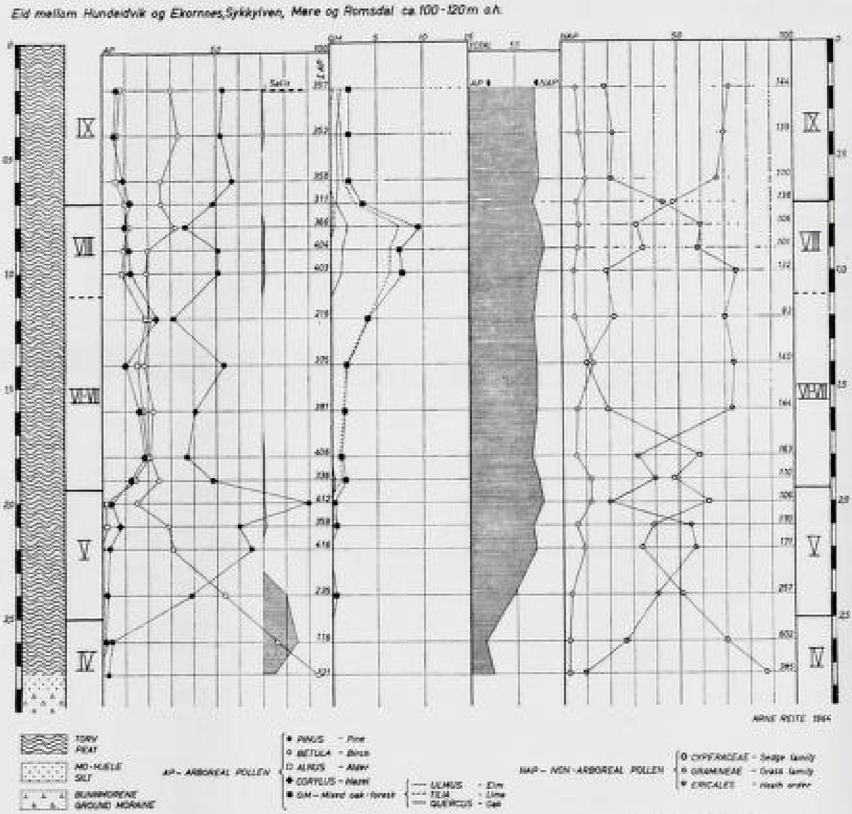


Fig. 12. Pollendiagram fra myr mellom Ekornnes og Hundeidvik, Sykkylven.
Pollen diagram from peat-bog between Ekornnes and Hundeidvik, Sykkylven.

likning med vegetasjonsvekslingen i fjerntliggende steder som Bømlo, Jæren og Oslo-området vil derfor bli usikker.

Prøvene ble tatt i en stor myr på eidet mellom Hundeidvik og Ekornnes (fig. 8). Vegetasjonen på myra i nåtiden består av røsslyng på tuer og rabber, ellers er det mest gras og halvgras. Det er noen få forkrøblete furutrær.

Prøvene ble for den dypeste del av myra tatt med en Hiller prøvetaker, fra de øverste 130 cm i en godt opprensket skjæring.

Prøvene ble behandlet med KOH og acetolyse, og farget med fuchsin. Det ble tatt sikte på å telle minst 400 treslagpollen (AP), men da noen av prøvene inneholdt svært lite treslagspollen i forhold til urtepollen (NAP) ble det talt færre AP i disse. Bortsett fra ett tilfelle har summen av AP og NAP vært mer enn 500.

AP-diagrammet viser variasjonen i pollenmengde av de viktigste treslag. Kurven for Salix er tegnet med større målestokk, og eikblandingskogen (QM) er skilt ut i et eget digram med større målestokk, da pollen av disse treslag er sparsomt tilstede.

Totaldiagrammet viser forholdet mellom treslagspollen (AP) og urtepollen (NAP).

NAP-diagrammet omfatter pollen av lyngordenen (Ericales), grasfamilien (Gramineae) og halvgrasfamilien (Cyperaceae).

Soneinndelingen (sonene angitt med romertall) er etter Jessen (Jessen, 1935). Denne inndelingen er basert på vegetasjonsvekslingen. For å avgjøre om sonegrensene i områder som ligger langt fra hverandre er synkrone er det nødvendig med absolutt datering, men slik datering er ikke foretatt i dette området.

AP-diagrammet viser at bjørk dominerer i de dypeste spektra. Bjørkepollenmengden avtar jevnt, samtidig som furupollen blir mer vanlig. Det fins en del pollen av Salix, og mengden av urtepollen er meget stor. Det er derfor sannsynlig at de dypeste deler av profilet omfatter sone IV — preboreal (Fægri, 1940, 1944, Hafsten, 1956).

Mengden av furupollen øker jevnt og når hele 90 % av AP. Hasselpollen blir mer vanlig, men når først et maksimum omtrent samtidig med or. Det sene hasselmaksimum sett i forhold til furumaksimum et et påfallende trekk ved diagrammet, men det er nødvendig med flere pollenanalyser for å avgjøre om dette gjelder for hele Sunnmøre.

Grensen mellom sone V og VI er trukket hvor pollen av or begynner å bli vanlig. Samtidig øker pollenmengden av hassel, mens eikblandingskogen ser ut til å ha vært spartsomt tilstede.

Grensen mellom sone VII og VIII er stiplet hvor pollenmengden av treslag som tilhører eikblandingskogen, som har vært nesten helt dominert av alm, øker sterkt. Grensen mellom sone VIII og IX er trukket hvor det skjer en markert nedgang i pollenmengden av eikblandingskogen.

Hensikten med pollenanalysen var for å forsøke å finne tidspunktet for innlandsisens tilbaketrekning. Det viste seg at det på denne lokalitet sannsynligvis bare er torv som er avsatt i postglacial tid. Grunnen kan være at prøvene ikke ble tatt i et tjern eller en gjenvoksningsmyr, slik at det gikk lang tid før torvdannelsen ble såvidt betydelig at pollen kunne oppbevares.

Senkningen av glaciasjonsgrensen under lokalglaciasjonsstadiet.

I Sykkylven er det også i nåtiden en rekke breer i de høye fjellene mellom Sykkylvsfjorden og Hjørundfjorden. Glaciasjonsgrensen er ca. 1400 m.

For å bestemme glaciasjonsgrensen under lokalglaciasjonsstadiet har Simonys toppmetode blitt anvendt (Brückner, 1887). Dette er eneste metoden som kan brukes med de kart som har vært til rådighet. For at metoden skal gi riktig verdi for glaciasjonsgrensen må det være fjelltopper omkring glaciasjonsgrensens nivå.

Fjellene ble inndelt i to grupper, en hvor det ikke er randmorener etter botnbreer, og en gruppe hvor morener er funnet. På denne måte har glaciasjonsgrensen i Sykkylven under lokalglaciasjonsstadiet blitt bestemt til ca. 800 m. Det har altså vært en senkning på hele 600 m sett i forhold til nåtiden. Høyeste fjellet på Sula er 800 m, og fjellet innenfor den vestligste randmorenen er betydelig lavere. Glaciasjonsgrensen har trolig vært ca. 700 m.

På Hareidlandet (fig. 2) kan glaciasjonsgrensen bestemmes med større sikkerhet, da det er mange fjell med høyder fra 500 — 700 m. Det er funnet sikre randmorener avsatt av botnbreer i de fleste fjell som er høyere enn 600 m, og dette er trolig på det nærmeste glaciasjonsgrensen under lokalglaciasjonsstadiet.

Da strandlinjeforskyvningen i det undersøkte området etter lokalglaciasjonsstadiet bare er 20—50 m, er det ikke tatt hensyn til denne under beregning av glaciasjonsgrensen.

Andersen (1954) har for ratidens randmorener i Lysefjordområdet i Rogaland funnet at snøgrensen var 400—550 m lavere enn i nåtiden.

Paschinger (1912) har bestemt glaciasjonsgrensens helning mot kysten i det sydlige Norge til 5—7 m/km. På Helgelandskysten har Svensson (1959) funnet en gjennomsnittlig helning mot kysten på 5 m/km, mens den maksimale gradient nær kysten var 13 m/km.

Dersom en regner med en glaciasjonsgrense på 800 m i Sykkylven og 600 m på Hareidslandet blir den gjennomsnittlige gradient 6 m/km, noe som stemmer godt med glacaisjonsgrensens helning andre steder langs kysten.

Det er ikke lett å avgjøre hvor meget lavere temperaturen kan ha vært under lokalglaciasjonsstadiet, da det er mange andre faktorer som er av betydning for bredannelse.

For å bestemme hvor stor temperatursenkning en viss glaciasjonsgrensenedpressing har vært betinget av blir det vanligvis brukt en temperaturgradient. Andersen (1954) bruker gradienten 0,7, dvs. 100 m senkning av glaciasjonsgrensen ved en senkning av sommertemperaturen på 0,7° C. Svensson (1959) bruker gradienten 0,6. Ved bruk av førstnevnte verdi skulle en senkning av glaciasjonsgrensen på 600 m tilsvare en senkning av sommertemperaturen på ca. 4° C, dette under forutsetning av at nedbørforholdene har vært omtrent som nå.

Selv om nedbørforholdene har vært forskjellig fra nåtiden, er senkningen av glaciasjonsgrensen så betydelig at lokalglaciasjonen neppe kan ha funnet sted senere enn i Yngre Dryas tid.

Sammenligning med andre deler av landet.

En så stor senkning av glaciasjonsgrensen som det har vært på Sunnmøre burde også gjøre seg gjeldende i andre landsdeler. Grunnen til at det mangler randmorener avsatt av lokalbreer i mange fjellområder i Syd-Norge kan være at fjellene ennå ikke hadde smeltet fram av innlandsisen under lokalglaciasjonsstadiet. Det må også tas i betraktning at de sentrale fjelltrakter har vært sterkt nedpresset på grunn av istyngden.

Mannerfelt (1940, 1945) har funnet randmorener avsatt av botnbreer i fjellet Fongen i Sør-Trøndelag. Han har ved hjelp av pollenanalyse ment å kunne fastslå at randmorenene er eldre enn den subatlantiske klimaforverring.

På den vestlige del av Hardangervidda er det funnet tydelige randmorener avsatt av innlandsisen. Ofte fins et dobbelt morenetrinn. Liestøl (1963) har funnet at snøgrensen har vært ca. 1550 m under breframstøtet, dette er omtrent 100 m lavere enn i nåtiden. Han mener morenene er avsatt senere enn i ratid.

De samme morenene er beskrevet av Simonsen (1963) og Anundsen (1964).

Sollid (1964) har omtalt randmorener i de vestlige deler av Fokstumyr —Hjerkinn-området, avsatt av en nordøstlig brestrøm fra et glaciasjonssentrum i Jotunheimen. Han mener breframstøtet er yngre enn ratid, og bygger da vesentlig på den høye firngrense randmorenene i området synes å vitne om.

Holmsen (1964) har beskrevet de samme randmorenene. Han antyder at de kan være samtidige med randmorenene på Hardangervidda.

Både på Hardangervidda og Hjerkinn ser det ut til at senkningen av snøgrensen under breframstøtet har vært liten. Det er derfor all grunn til å anta at den meget store senkningen av glaciasjonsgrensen på Sunnmøre må ha funnet sted på et tidligere tidspunkt.

Ratidens randmorener er fulgt temmelig sammenhengende vestover til Ryfylke (Andersen, 1954, 1960). Videre nordover langs kysten har Undås (1963) forsøkt å finne morenenes beliggenhet. Han trekker grensen over Halsnøy i Hardangerfjorden. I området Sognefjorden—Stad mener han morenene fra denne tid ligger langt mot vest, slik at bare de aller ytterste øyer var isfrie. På Mørekysten trekker han grensen for isen i ratid langt vest for de ytterste øyer.

I Bergensområdet er oppknuste skjell i morenemateriale forholdsvis vanlig. Flere C-14-dateringer tyder på at skjellene er fra Allerød tid. Senere — trolig i Yngre Dryas tid — har isen rykket fram, men det er ikke klart hvor langt vest dette breframstøtet har nådd (Holtedahl, 1964).

I motsetning til Undås mener Carlsson (Nydal, 1964) at ratidens brefront i Sognefjorden har ligget ved munningen av sidedalene, hvor det ofte er et dobbelt morenetrinn. En datering av skjell foran et isranddelta i Instevik, Sogn, gav resultatet 10250 ± 440 år før nåtiden.

Nord for Breim i Nordfjord er en meget tydelig randmorene på begge sider av fjorden, men dette breframstøtet har ikke blitt datert. Ut fra beliggenheten er det sannsynlig at det kan være samtidig med lokalglaciasjonsstadiet på Sunnmøre.

Det har vært antatt at randmorenene ved Gikling i Sunndalen og Tautra i Trondheimsfjorden skulle være avsatt i ratid, dette har særlig vært basert på funn av arktiske skjell foran disse randavsetningene. Det foreligger ikke C-14-dateringer fra disse trakter så det er ikke mulig å avgjøre om disse randavsetningene er avsatt samtidig med lokalglaciasjonen på Sunnmøre.

Oversikt — konklusjon.

En så sterk glaciasjonsgrensenedpressing som det har vært på Sunnmøre under lokalglaciasjonsstadiet har trolig også ført til at innlandsisen rykket fram på nytt. Nå har feltarbeidet i de indre deler av Sunnmøre vært lite omfattende, dessuten har ikke flyfotografier fra dette området vært tilgjengelige.

Strøm (1956) har fra Geirangertraktene omtalt randmorener som trolig er avsatt av innlandsisen.

Kaldhol (1946) har fra Stadheim, Hellesylt, beskrevet et snitt hvor morenemateriale av 6 m mektighet overleirer glacifluvialt materiale. Over morenen ligger glacifluvialt materiale. Denne lagfølgen tyder på en oscillasjon av isranden.

De høyeste marine terrasser i Geiranger, Hellesylt og Tafjord faller på omtrent samme linje i ekvidistansediagrammet (fig. 11) som de høyeste terrasser i randmorener avsatt av botnbreer. Dette kan antyde at innlandsisen under lokalglaciasjonsstadiet har nådd ned til fjordbunnene i indre deler av Storfjorden og bygget opp israndavsetninger. Ved videre feltundersøkelser vil det trolig være mulig å fastslå dette.

Funn av flint og Oslo-bergarter viser at Sunnmørskysten ble tidlig isfri. Dette blir også støttet av C-14-dateringen av skjellforekomsten på Aure (11620 ± 120 år før nåtiden), som tyder på at Sykkylvsfjorden var isfri i Allerød tid.

Den meget lave glaciasjonsgrense gjør det sannsynlig at lokalglaciasjonen på Sunnmøre har funnet sted under klimaforverringen i Yngre Dryas tid.

Summary

On the Mountain Glaciation of Sunnmøre, West Norway

Terminal moraines deposited by cirque glaciers are common in the western parts of Sunnmøre (figs. 1, 2, 4, 8). As this area is supposed to have been completely covered by ice during the last glaciation, the mountain glaciation must have occurred in late glacial or postglacial time.

Many cirque glaciers have descended to sea-level, and the position of terminal moraines shows that the fjords were ice-free at that time. The highest marine terraces found on these moraines are considerably lower than those found in other parts of the fjords. On the island Sula (fig. 1) half of the total shore-line displacement was completed when the cirque glaciers receded, in Sykkylven one third.

The glaciation limit during the time of mountain glaciation has been determined as about 800 m in Sykkylven, 700 m on Sula and about 600 m on the island Hareidlandet (fig. 1). The glaciation limit of this area today is about 600 m higher, indicating a severe climatic deterioration during the time of mountain glaciation.

Radiocarbon dating of marine shells (Cyprina islandica) from Aure, Sykkylven (fig. 8), which showed an age of 11620 ± 120 years B.P., seems to indicate that the fjord districts were free from ice during the climatic amelioration in Allerød time.

This is also indicated by stones of Oslo rock material and chert found below the marine limit in the Romsdalsfjord and the outer part of the Storfjord area. This material must have been transported by icebergs at a time when the area was ice-free and the ice reached the sea in the Skagerak, i.e. not later than Younger Dryas time.

Pollen analyses from a bog at Ekornnes (fig. 8) showed a preboreal pollen spectrum, dominated by birch and with high precentages of non-arboreal pollen in the deepest part of the bog (fig. 12).

In the inner part of Storfjorden (fig. 1) the marine limit is much lower than in the middle part of the fjord (fig. 11). The reason may be that the recession of the inland ice was temporarily halted here during the deglaciation, but deltaic deposits of sand overlain by 6 m of till in Hellesylt indicate that there has been a re-advance of ice.

Conclusion: The mountain glaciation of Sunnmøre seems to have taken place during the climatic deterioration in Younger Dryas time.

Litteraturliste

Forkortelser (Abbreviations):
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U.B.Å.: Universitetet i Bergen, Årbok.
N.G.T.: Norsk Geologisk Tidsskrift.
N. Geogr. T.: Norsk Geografisk Tidsskrift.
N.G.U.: Norges Geologiske Undersøkelse.

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Some new aspects of the geology of Varanger peninsula (Northern Norway)

Preliminary report

by

Anna Siedlecka and Stanislaw Siedlecki[1]

Abstract

The authors have established that Varanger Peninsula consists of two different geological regions divided from each other by a large thrust fault trending in NW-SE direction. The south-western, Tanafjord-Varangerfjord region, consists of rocks of the Vestertana Group (Reading, 1965) and of the "Older Sandstone Series" (Føyn, 1937). The north-eastern, Barents Sea region, consists of thick sedimentary sequences which are quite dissimilar to the sedimentary sequences present in the Tanafjord - Varangerfjord region. New groups and formations are described in the Barents Sea region, in which the rock assemblages are considered to have been deposited in a miogeosyncline.

The age of the sedimentary rocks of the Barents Sea region is uncertain; they presumably represent a time period earlier than Cambrian. The Barents Sea region is tentatively regarded as belonging to the Timanian orogenic belt.

Introduction

Thanks to the inspiration of Rektor S. Føyn, and to the organisation and financial assistance of Norges geologiske undersøkelse and Norges Teknisk-Naturvitenskapelige Forskningsråd, geological investigations have been carried out on the Varanger Peninsula during parts of the 1966 and 1967 field seasons. The main object of the investigations has been the study of the stratigraphy and structural geology in those areas of the Varanger Peninsula underlain by unmetamorphosed sedimentary rocks of the "Older Sandstone Series" (Føyn, 1937) and its presumed equivalents. The investigations carried out for 3 weeks in August 1966 during rather inclement weather conditions were necessarily of only a reconnaissance nature. In 1967 more systematic field-studies were undertaken during a period from the end of June until the beginning of Sep-

1) Laboratory of Geology, Polish Acad. of Sci., Cracow (Poland).
Norges geologiske undersøkelse, Trondheim (Norge).

tember. The following descriptions are based mainly on observations made during this time.

With the help of a car, local post-boat, fishing boats and a small boat with an outboard-motor, more or less detailed geological observations along the costal sections of the Varanger Peninsula were eventually completed, with the exception of limited area between Store Molvik, Tanahorn and Skonsvik (in the northernmost part of the peninsula).

In the inner part of the peninsula investigations were at first concentrated along the road crossing the central highland and later in land-areas to the north-west and south-east of this road, especially the area surrounding the big lakes Buevannet, Gednjevannet and Hjordvannet.[1]) The eastern coast of Tanafjord between Leirpollen and Trollfjorden, together with the land-area stretching inland from this coast, has been mapped in detail.

In all, an area of about 600 sq.km. of the inner part of the Varanger Peninsula and of its coastline has been mapped on the scale 1:50.000. Because of lack of time other areas in the interior of Varanger Peninsula were studied in less detail, but the localities thought to be of greater importance to a general understanding of the geology of the peninsula were visited by us during reconnaissance trips.

At the present, however, our field investigations and petrological studies of collected samples are still in the preliminary stage; future studies will supply many details to supplement our present-day knowledge of the geology of the Varanger Peninsula. Consequently only the broad outlines of some new aspects of the geology of this area will be presented here.

Acknowledgements

We wish to express our thanks to Norges geologiske undersøkelse, Trondheim, for defraying the expences of our investigations, and to Norges Teknisk-Naturvitenskapelige Forskningsråd, for providing a fellowship for one of us (A.S.).

We are especially indebted to Rektor S. Føyn, Oslo, for the proposing the present investigations of the "Older Sandstone Series" on Varanger Peninsula, and for his constant advice and valuable discussions.

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[1]) Geographical nomenclature on Varanger Peninsula is based partly on Lappish tradition and partly on Norwegian introduced names. To avoid possible confusion the names used here are those appearing on the "Topografisk kart over Norge", 1:100,000, ed. Norges geografiske oppmåling, 1956-1962. These have been rewritten to conform with modern orthography.

Mr. W. Czajkowski, Mr. T. Dobrowolski and Mr. D. Myklebust were our field-assistants in the summer of 1967.

Dr. D. Roberts, NGU, Trondheim, and Prof. F. C. Vokes, NTH, Trondheim, have kindly corrected the English manuscript.

Summary of previous investigations

Varanger Peninsula and adjacent areas of the eastern Finnmark have been investigated by numerous geologists, the Eocambrian sediments of glacial origin generally attracting the most attention. (Eocambrian is considered by the present authors as a chronostratigraphical term including the time-period between the beginning of deposition of tillites and the first beds containing Cambrian fossils.) Other sediments, particularly those older than the tillites, were not often studied in this part of eastern Finnmark and the monographic paper of Holtedahl (1918) can be regarded as the first and still most important source of information concerning the "pre-tillitic" rock-sequence of the Varanger Peninsula.

The present authors, both during the field investigations and while establishing their stratigraphical and tectonic conceptions, have based their work mainly on the papers of Reusch (1891-b), Holtedahl (1918, 1919, 1932, 1953, 1960-a, 1960-b, 1961), Føyn (1937, 1960, 1964), Rosendahl (1945) and Reading (1965, 1966). In preparing the new stratigraphical scheme (see Fig. 1) account has been taken, as far as possible, of the terminology which have previously been either introduced or suggested by the writers mentioned above.

New data concerning the geology of the Varanger Peninsula

Field investigations carried out by the present authors in 1967 show that a large thrust fault divides Varanger Peninsula into two different geological regions. The thrustfault (or set of parallel faults) runs in an approximate NW-SE direction along Trollfjorden, Trollfjordelven valley, Gednjevannet and Hjordvannet lakes, the NE slopes of the Skipskjølen mountain and along the Komagelven valley (see map, Fig. 2). There is no lithostratigraphical correlation between the two different regions. The region situated SW of the thrust will here be called the Tanafjord - Varangerfjord region and that which extends NE of the thrust the Barents Sea region.

The precise nature of the thrust is not yet known; data collected during the mapping have permitted us only to establish the occurrence of this structure, its direction and extent. Stresses which initiated the thrusting have been directed approximately from NE to SW, causing the rock sequences of the

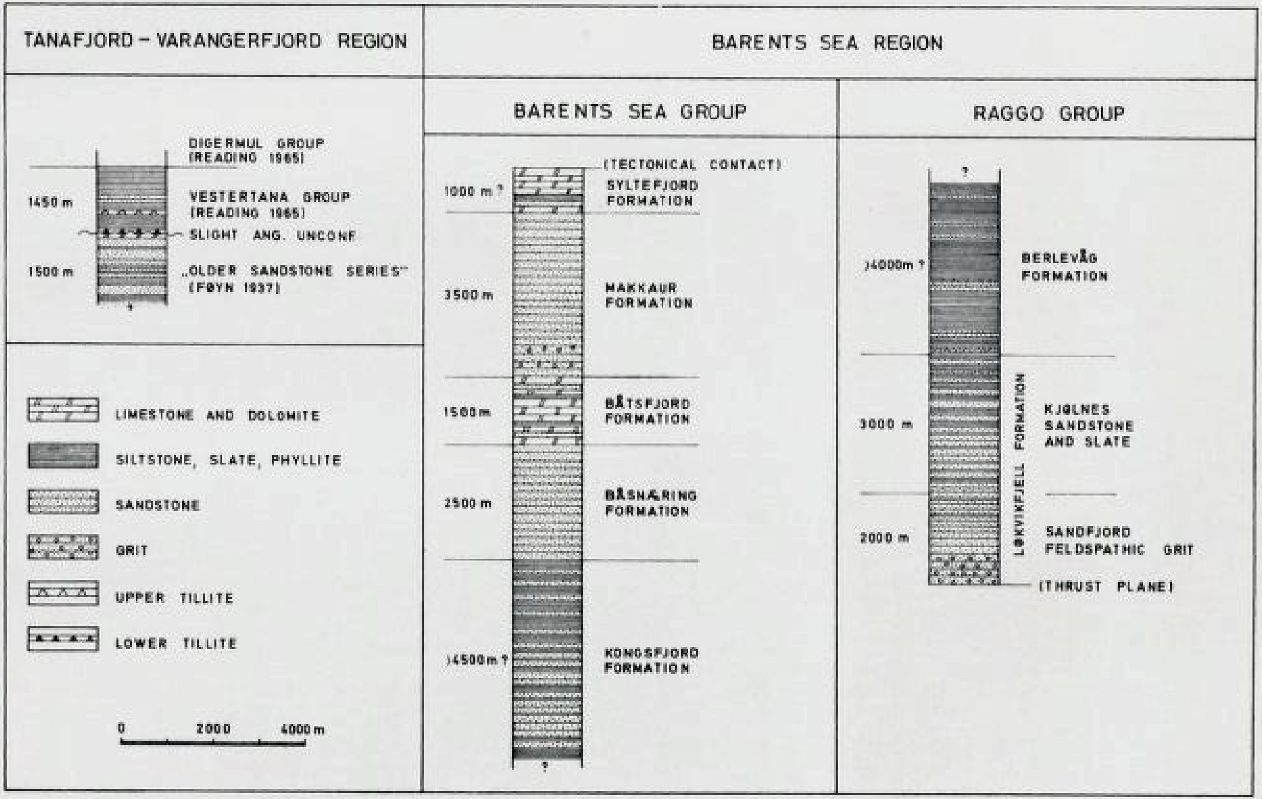

Fig. 1. Lithostratigraphical sections of the rock sequences of the Varanger Peninsula.

Barents Sea region to be thrust over those of the Tanafjord - Varangerfjord region. The thrust will here be called the Trollford - Komagelv thrust.

Short descriptions of the rock sequences observed in the above-mentioned regions follow immediately below.

Tanafjord - Varangerfjord region

The total known thickness of the entire sequence of sedimentary rocks of Hyperborean age (term introduced and defined by Sederholm, 1932) reaches ca. 3000 m. in this region. An Eocambrian succession (ca. 1450 m.), discribed by Føyn (1937) and Reading (1965) from an adjacent area on the Digermul Peninsula, has been termed the Vestertana Group (Reading, 1965). The oldest beds of this group are tillites which, together with slates and sandstones of the Nyborg Formation, represent sediments of Varangian (Varegian) Ice Age. The youngest strata of the Vestertana Group, Breivik Formation, are of Lower Cambrian age (Føyn, 1967). The sedimentary rock sequence underlying the Vestertana Group is that of the "Older Sandstone Series" of Føyn (1937). This reaches ca. 1500 m in thickness, but the base of the series is not known. The main area on the Varanger Peninsula underlain by the "Older Sandstone Series" is the eastern shore of Tanafjord (between Leirpollen and the southern coast of Trollfjord) and the territory to the east of Tanafjord, extending up to Trollfjordelven valley and to the western shores of the Gednjevannet lake. The Trollfjord - Komagelv thrust fault forms the north-eastern border of this area. Toward the SE, along a line running approximately up the Čabmaelven valley and south of Hanglečerro, the "Older Sandstone Series" dips beneath the Vestertana Group. Along the northern shores of Varangerfjord the rocks of the "Older Sandstone Series" crop out again, so that the whole area between the heads of Tanafjord and Varangerfjord can be regarded as a large syncline with Eocambrian and Cambrian sediments of the Vestertana Group in its core and with the limbs composed of sediments of the "Older Sandstone Series" (see cross-section, Fig. 2). The axial plane of the syncline trends approximately in a WSW-ENE direction. The syncline is assymetrical, the north-western (NNW) limb being steeper and more distinctly folded than the strata of the south-eastern (SSE) limb. In addition, evidence of strong deformation is present at the NE side of the syncline, near the plane of the Trollfjord - Komagelv thrust. Outcrops of rocks of the "Older Sandstone Series" in the area near Varangerfjord are not so extensive as in the sections around the Tanafjord area. Problems of detailed stratigraphy and of facial changes which may occur within the rocks of the "Older Sandstone Series" in the Varangerfjord area, i.e. in the vicinity of the border of the Baltic shield, require further investigation.

An examination of the lithologies of the "Older Sandstone Series" in the Tanafjord - Varangerfjord region, and of the preserved sedimentary markings, show that the sediments are mostly shallow water deposits, in part of marine and in part presumably of continental origin. Within these sediments occur thick strata of quartzitic sandstones interbedded with sandy and silty shales. Siltstones often appear in the lower parts of the rock-sequence (e.g. Stangenes shale) while sandstones (quartzites) prevail in its upper part. Some sandstones contain feldspar; typical "sparagmite", rich in coarse grains of feldspar is, however, very rare. Dolomite layers, interbedded with shale and mudstone, are known only from the uppermost part of the rock-sequence. In some places they were eroded before the deposition of tillites. The dolomite occurring in the Tanafjord - Varangerfjord region contains stromatolitic structures and displays many lithological similarities to the Porsanger dolomite, with which it can almost certainly be correlated.

The Tanafjord - Varangerfjord region on Varanger Peninsula should be regarded as the easternmost part of a large area of East Finnmark over which sediments correlated with the "Older Sandstone Series" and Vestertana Group are extensive. Stratigraphical and palaeogeographical correlations within these rock sequences have been discussed by several authors (see Holtedahl, 1918, 1932, 1960; Føyn, 1937, 1964, 1967).

Igneous rocks (occurring in great quantity in the Barents Sea region) are rare in the Tanafjord - Varangerfjord region. They are absent around the inner part of Tanafjord and in the southern part of the Varanger Peninsula (Føyn, 1945). A few diabase dykes have been observed by Holtedahl (1918) along the N coast of the outer Varangerfjord at Krampenes, on Store Ekkerøy and at Kvalneset. The latter locality has been recorded also by Mr. O. Jøsang, Oslo (personal communication).

Barents Sea region

Two different successions of sedimentary rocks occur in this region, the boundary between them being a plane of tectonic disconformity. They will be described below as:

1. Barents Sea Group, and
2. Raggo Group.

1. Barents Sea Group

The sedimentary sequence of the Barents Sea Group crops out most conveniently for observation, in large coastal sections along the north-eastern shore of Varanger Peninsula, approximately between Sandfjordbukten (NW of

Kongsfjord) and Komagvær (S of Vardø). The group reaches a thickness of about 13 000 m and consists of sandstones, siltstones, slates and carbonate rocks; no tillites or tillite-like rocks have been observed. The beds show a constant general strike NE-SW between Kongsfjord and Syltefjord; this direction is followed by the fjords occurring in NE part of the Varanger Peninsula. In the south-east the beds dip mainly 15-30° SE, but northwestwards the dips steepens to about 60-70° SE.

Preliminary geological observations concerning some sediments now referred to the Barents Sea Group were reported by Keilhau (1844) and Reusch (1891 b). More detailed descriptions are to be found in Holtedahl's (1918) monograph, where sediments of the Barents Sea Group have been mostly included in the "Older, dolomite-bearing sandstone division".

The present authors differentiate five formations within the Barents Sea Group using in their descriptions, as far as possible, lithostratigraphical terms introduced by Holtedahl (1918).

Kongsfjord Formation

Holtedahl (1918, pp. 188, 223) introduced term "Kongsfjord skifer" for this assemblage. The "Naalnesets sandstensrække" of Holtedahl (1918, p. 184) is also included here by the present authors.

The Kongsfjord Formation is the oldest in the Barents Sea Group and occurs mainly in the area situated close to Kongsfjord. It consists of dark-grey and blackish mudstones, slates and dark-grey sandstones. The latter, usually fine-grained, become in some places coarse-grained, feldspathic, and contain intercalations of conglomerates. Although the lithology of the Kongsfjord Formation is apparently monotonous, a lot of sedimentary markings have been recorded. Graded bedding, load casts, intraformational conglomerates and indistinct ripple marks are those most commonly observed in this formation.

The base of the Kongsfjord Formation is not known; its contact with the Raggo Group is a thrust surface. At the top, towards the south-east, the Kongsfjord Formation grades into the next, Båsnæring Formation.

The apparent thickness of the Kongsfjord Formation, computed from the geological map approximate 7500 m. However, in reality, the true thickness is much lower; the numerous disharmonic folds occurring in the incompetent slaty beds have resulted in an apparent enlargement. The real thickness of the Kongsfjord Formation may be established, tentatively, as about 4500 m. It seems that in the future it will be possible to differentiate minor lithostratigraphical units within Kongsfjord Formation.

Båsnæring Formation

"Baasnæringens sandstensserie" of Holtedahl (1918, p. 187).

The Båsnæring Formation occurs above the Kongsfjord Formation and there is a transition between these two units. This transition has been recorded by Holtedahl (1918, p. 188) at the SE side of Kongsfjord.

The Båsnæring Formation consists of thick beds of quartzitic sandstones and underlies a peninsula between Kongsfjord and Båtsfjord. In the lower part of the formation medium-grained, grey, massive sandstones are dominant; grey slates are subordinate and disappear gradually upwards. The upper part of the formation consists mostly of red, coarse-grained sandstones, commonly exhibiting cross-bedding, while the uppermost part is characterized by interbedding of red siltstones with beds of medium- and coarse-grained, red sandstones. Then, light-grey sandstones and dark-grey arenaceous shales with interbedded carbonate layers appear. The proportion of the latter increases upwards. These rocks will be regarded as the lowermost beds of Båtsfjord Formation.

Thickness on the Båsnæring Formation is ca. 2500 m (according to Holtedahl, 1918, p. 187 - ca. 2000 m).

Båtsfjord Formation

"Baasfjord dolomit-skifer-sandstensserie" of Holtedahl (1918, pp. 188, 223).

The Båtsfjord Formation underlies the area close to Båtsfjord and the valley forming the SW prolongation of the latter. The formation consists of interbedded dolomitic and clayey limestones, shales, mudstones and sandstones. The dolomitic limestones are grey or light-grey and become yellowish on weathered surfaces. Clastic rocks associated with carbonate rocks are grey, blackish, red, green and yellow. Ripple marks and mud-cracks are common on the bedding surfaces of the clastic rocks; these structures have been described in detail by Holtedahl (1918). In the lower and upper parts of the formation the clastic rocks seem to be dominant while carbonate rocks prevail in the middle. The base of the formation is marked by light-grey sandstones and dark-grey arenaceous shales appearing above the red coarse-grained quartzitic sandstones of the subjacent Båsnæring Formation. At the top of the Båtsfjord Formation the carbonate rocks are interbedded with light-grey or red coarse-grained calcareous sandstones and red silstones. Then, the carbonate rocks gradually disappear and the Båtsfjord Formation grades into the next formation beginning with red, coarse-grained cross-bedded sandstones and conglomerates. This transition from Båtsfjord Formation to the Makkaur Formation has been previously observed by Reusch (1891, p. 41) and by Holtedahl (1918, p. 190, Pl. XI).

Thickness of the Båtsfjord Formation may be established as ca. 1500 m.

Makkaur Formation

"Makurs sandstensserie" of Holtedahl (1918, p. 192).

The Makkaur Formation underlies a large peninsula occurring between Båtsfjord and Syltefjord. This formation consists of mainly medium- and coarse-grained sandstones forming thick layers (1 meter and more) and exhibiting cross-bedding. The sandstones are red-violet or reddish-grey; on the weathered surfaces they are more intensively red stained. The sandstones are feldspathic, especially in the lower part of the formation where quartz-conglomerate layers also appear. These conglomerates and associated sandstones have been observed by Reusch (1891 b, pp. 40-41) at Makkaur and characterized as sparagmite-like rocks. The very monotonous sandstone sequence of the Makkaur Formation reaches about 3500 m in thickness. The uppermost part of this formation occurs at the head of Syltefjord where (according to the description of Holtedahl, 1918, pp. 193-194) dark arenaceous shales and yellowish dolomites rest on coarse-grained sandstone exhibiting cross-bedding. There is a concordance and continuity of sedimentation but the contact is relatively sharp. The arenaceous shales and dolomites have been included in the next formation described below.

Syltefjord Formation

A part of "Syltefjords and Persfjords dolomit-skifer-sandstensserie" of Holtedahl (1918, p. 223).

The Syltefjord Formation crops out near Syltefjord and shows lithological similarity to the Båtsfjord Formation. The Syltefjord Formation consists of ligth-grey dolomitic limestones interbedded with red and greenish sandstones and mudstones often exhibiting ripple marks and mud cracks on the top surfaces. Holtedahl (1918, pp. 193-194, Pl. XVI) observed oolitic and stromatolite-like structures in carbonate rocks of Syltefjord at Veinesodden. Along the SE side of Syltefjord the beds of the Syltefjord Formation dip steeply in a NW direction and are in contact along a thrust fault with reddish and grey sandstones (Båsnæring sandstones). The known thickness of the Syltefjord Formation may be established as ca. 1000 m.

Beds of the Barents Sea Group occur again south-east of the SE coast of Syltefjord as far as the outer part of the coast of Varangerfjord (see Fig. 2). The units present are mainly those of Kongsfjord, Båsnæring and Båtsfjord Formations. Isoclinal dip of the beds which can be observed in the section between Kongsfjord and Syltefjord does not continue east of Syltefjord; in this part of the Barents Sea region the beds are distinctly folded. Axial planes of synclines and anticlines trend in a SW-NE (Caledonian) direction; a distinct syncline, with the Båtsfjord Formation in the core and with limbs consisting of Båsnæring Formation, occurs in Persfjord. This structure has been observed already by Keilhau (1844) and by Holtedahl (1918).

2. Raggo Group

Sedimentary sequences referred to the Raggo Group underlie the northern part of the Varanger Peninsula, which is called the Raggo Peninsula (Raggohalvøya). Beds of the Raggo Group in the coastal section between Sandfjordbukten and Berlevåg dip quite steeply toward NW and are thrusted over the Barents Sea Group. The thrust plane marks the base of the Raggo Group. The top surface of the group is not known. Total known thickness of the Raggo Group may be established as about 9000 m (?). A geological section of the Raggo Group is best exposed along the NE coast of the Varanger Peninsula between the NW side of Sandfjordbukten and Skonsviken. The Raggo Group consists of clastic rocks; similarly as in the Barents Sea Group, tillites have not been observed.

Within Raggo Group two formations may be differentiated:

1. Løkvikfjell Formation,
2. Berlevåg Formation.

Løkvikfjell Formation.

The name "Løkvikfjeldets sandstensserie" was introduced by Holtedahl (1918) for the sandstones which crop out along the sea shore west of Kongsfjord, between Styret and Sandfjordbukten's north side. Løkvikfjell mountain consists of the rocks of this formation.

The present authors divide the Løkvikfjell Formation into two members: (1) Sandfjord feldspathic grit and, (2) Kjølnes sandstone and slate.

Sandfjord feldspathic grit. This member occupies two separate areas: (a) the area extending in a NE-SW direction from the Sandfjordsaksla and Styreaksla mountains on the coast of Barents Sea to the western slopes of Skidnefjell on the north-eastern coast of Tanafjord, N of Trollfjord; (b) the area which occupies the central part of the Varanger Peninsula between Buefjellet mountain, Gednjevannet lake and Ordofjellet mountain and which extends further eastwards op to Skogåsvidden. In addition, isolated klippen of beds of the member under consideration have been observed on the peninsula which divides Båtsfjord from Syltefjord and which consists mainly of Makkaur Formation. The Sandfjord member consists of coarse-grained, pink, light-grey, yellowish or red-violet feldspathic grit with conglomerate intercalations, in layers 0.3-1 m in thickness. Cross-bedding is common in these beds. Pebbles are usually subrounded and reach up to 10 cm in diameter, usually they are about 2-3 cm. Pebbles of white or pink quartz, of red or violet quartzite, red jasper and black siliceous rock have been observed. The psammitic material of the grit consists mainly of rounded quartz grains and quite abundant particles of pink feldspar. Generally, this sediment is reminiscent of the sparagmites.

In a well exposed section between Sandfjordbukten and Kjølnes it may be observed that conglomerate layers occur mainly in the lower part of the member. Upwards, the feldspathic sandstones become medium-grained, shale intercalations appear and then the member passes into Kjølnes sandstone and slate.

Thickness of the Sandfjord feldspathic grit in this section is ca. 2000 m.

Kjølnes sandstone and slate member. This member consists of interbedded sandstones and slates and reaches a thickness of about 2000 m. Medium-grained quartzitic sandstones and coarse-grained feldspathic sandstones, grey and grey-pink, form layers ca. 0.2-0.7 m thick. In some places quartz conglomerates appear. Sandstones are interbedded with dark-grey and black slates and phyllites. Generally the sediments of this member become more and more metamorphosed when traced from east to west. The member does not appear in the SW slopes of Skidnefjell mountain in the northern vicinity of Trollfjorden because the Berlevåg Formation thrust over it in this area and rests with a tectonic contact directly on the Sandfjord feldspathic grit.

Berlevåg Formation

This has been observed in the area near Berlevåg, between Berlevåg and Store Molvik, and N of Trollfjorden. Near Berlevåg this formation is represented by black and black-green phyllites and there is presumably a continuity between the Kjølnes sandstone and slate member and the Berlevåg Formation. Between Berlevåg and Store Molvik the black phyllites are interbedded with quartzites and north of Trollfjorden occur strongly disturbed grey quartzites and grey-green phyllites. In the latter area the phyllites and quartzites of Berlevåg Formation are thrusted directly on the Sandfjord feldspathic grit, as it was already mentioned above.

Although little work has been done on the Berlevåg Formation it may be concluded, in agreement with Holtedahl's (1918) opinion, that a SW-NE running thrust, separating autochthonous (or para-autochthonous) and allochthonous sequences, dies out in a NE-direction and that on the N coast of the Varanger Peninsula there is a continuity and transition from unmetamorphosed to metamorphosed beds.

Igneous rocks

Numerous dykes of basic rocks occur in the Barents Sea region, a feature which seems to be characteristic of this region. The dykes are especially common in the area between Kongsfjord and Syltefjord, intruding sedimentary sequences of the Barents Sea Group. They are distinctly marked in morphology as upstanding ribs within slaty, easily disintegrated beds of the Kongsfjord Formation and apparently are most numerous here. They are, however, equally common in sequences consisting mainly of sandstones or carbonate rocks (e.g. Makkaur Formation, Båtsfjord Formation). The dykes usually do not reach more than about 10 m in thickness; they occupy vertical fractures and, in general, form a set striking in an ENE direction (60-80°). Usually the dykes do not exhibit any signs of tectonic disturbance; in a few cases, however, some deformation appears. This led Holtedahl (1918) to conclude that the intrusions took place before the orogenic disturbance in this region was completely finished.

Dykes of basic rocks do not occur in the Raggo Group. However, Holtedahl (1918, p. 182) observed some thin (2-4 m) basic intrusions concordant to Berlevåg Formation rocks at Skonsvik near Berlevåg.

Origin and age of sedimentary rocks of the Barents Sea region

The thick sedimentary rock sequences in the Barents Sea region, especially those of the Barents Sea Group, would seem to be an unusual feature of the geology of Finnmark. Only the sedimentary rocks of the Raipas Group at Altafjord have been considered by Holtedahl (1918, p. 305) as "...an equivalent of the series, seen on the north coast of the Varanger Peninsula". This comparison was based on similarities between the sandstones and dolomites with stromatolitic structures, present in both these successions. The Raggo Group shows a lithological similarity to the Laksefjord Group described by Føyn (1960) from the Laksefjord district (the first suggestion concerning this similarity was made to us by Føyn, personal communication). In particular, the Løkvikfjell Formation seems to correspond to the Landersfjord quartzite and the Berlevåg Formation to the Friarfjord phyllite. Except for the above-mentioned similarities, no other lithological analogies between the Barents Sea region and other regions of Finnmark have been established up to the present time.

The sedimentary rock sequences of Finnmark, especially the unmetamorphosed ones of eastern Finnmark have many times been compared with those of Ribačij Peninsula, Sredni Peninsula and Kildin Island. Holtedahl (1960, p. 125) even correlated particular lithological units occurring in both these regions. Other comparisons were of rather general character, especially those concerning age. In the opinion of many geologists (Wegmann, 1928, 1929; Lupander, 1934; Polkanov, 1934, 1936; Keller and Sokolov, 1960; Keller et. al., 1963, pp. 103-113) the Ribačij and Srednij Peninsulas may be divided into two geological regions: (1) a northern, allochthonous region, that of the Ribačij Peninsula and (2) a southern, autochthonous region, comprising the Sredni Peninsula an Kildin Island. The sedimentary rock sequences of the northern region are thrust over those to the south. It was only Fieandt (1912) and later Agapiev and Vronko (in Luktevič and Haritonov, 1958, p. 365), who did not observe the presence of a large thrust fault between the Ribačij and Sredni Peninsulas. This latter oponion, however, seems recently to have been repudiated by Russian geologists (Keller et. al., 1963, p. 113); although the thrust is not marked on the "Carte tectonique de l'Europe" (I ed., 1964) or the "Tectoničeskaja Karta Eurazji" (1966). On the "Carte geologique de l'Europe" (1964) a fault (not thrust) is shown.

The Trollfjord - Komagelv thrust occurring on Varanger Peninsula has the same SE-NW trend as the thrust which divides the Sredni and Ribačij Peninsulas and would appear to be a direct prolongation of the latter. A conclusion from this - and that favoured by the present authors - is that there is one large thrust fault, or set of parallel thrust faults, which separates the Sredni and Ribačij Peninsulas from each other and crosses Varanger Peninsula. Consequently, the northern, allochthonous (called "exterior" by Keller and Sokolov, 1960), Ribačij block should correspond to the Barents Sea region of Varanger Peninsula, and the southern, autochthonous ("interior" after Keller and Sokolov, 1960) block of Sredni Peninsula and Kildin Island should be the equivalent of the Tanafjord - Varangerfjord region. However, litho-stratigraphical correlation between the corresponding blocks is not at all easy. Comparison of the lithological sections of the "Older Sandstone Series" and Vestertana Group with those of the "Kildinskaja svita" and "Volokova svita", differentiated and described on Sredni and Kildin by Keller and Sokolov (1960), shows some dissimilarities. It should, however, be emphasized that: (1) in both regions shallow-water sediments are developed in which facies changes can be expected to occur, even over small distances; (2) there is a slight angular unconformity present between the "Older Sandstone Series" and the Vestertana Group (Holtedahl, 1918, Føyn, 1937) and, similarly, the "Volokova svita" on Sredni rests unconformably on a denuded surface of the "Kildinskaja svita"; (3) in the "Kildinskaja svita" dolomites occur (on Kildin with stromatolites) which may be loosely compared with dolomites occurring at the top of the "Older Sandstone Series"; (4) ranges of thickness of sedimentary sequences in both regions are similar.

Tillites, present in the Vestertana Group and constituting an important horizon in any correlation of Eocambrian sediments, are lacking in the "Volokova svita". Beynon et. al. (1967) suggest, however, that the tillite formations of the Vestertana Group observed in the Leirpollen area (Varanger Peninsula) thin eastwards; it is thus probable that these formations disappear in this direction and do not reach the Sredni Peninsula region.

It can also be noted that the considerably thick sedimentary sequences of the Barents Sea Group and the Raggo Group can to a certain extent be compared with the sedimentary succession of the allochthonous Ribačij Peninsula block.

A geological section across the Ribačij Peninsula has been described by many authors (e.g. Fieandt, 1912; Wegmann, 1928, 1929; Lupander, 1924; Tenner, 1936), lately by Agapiev and Vronko (in Lutkevič and Haritonov, 1958, pp. 365-366, and in Keller et. al., 1963, pp. 110-111) and the sequence divided into the following formations, number 1 being the oldest:

1) "Svita Ejna" - coarse-grained, feldspathic sandstones with conglomerates and with peculiar dolomitic concretions up to 0.3 m in diameter. Intercalations of quartzitic shale. Thickness 3600 m(!)
2) "Zubovskaja svita" - feldspathic sandstones and shales with distinct cleavage. Thickness ca. 900 m.
3) "Slates of Cyp Naboloka" - dark-grey and black slates with closely packed cleavage planes; in some places yellow concentrations of jarosite.
   Thickness not less than 500 m.
   Total thickness 5000 m.

From the description of this succession it is apparent that the great thickness of the sedimentary sequence of the Ribačij Peninsula and the predominance of clastic sediments are the features which may best be used in a general comparison of this region with the Barents Sea region; a more precise correlation is not possible. It is probable, however, that e.g. the thick clastic "Svita Ejna" corresponds to one of the thick clastic formations of the Barents Sea region (Løkvikfjell Formation? Båsnæring Formation? Makkaur Formation?).

Deposition of the thick sedimentary sequences of the Barents Sea region seems to be closely associated with the development of a geosyncline. The Barents Sea Group starts with the Kongsfjord Formation which exhibits many features typical of a geosynclinal greywacke facies. This formation passes upwards into mainly red feldspathic sandstones of the Båsnæring Formation which appears to represent a molasse (red-bed) facies. Further upwards a calcareous facies occurs, this being the Båtsfjord Formation. Above that a repeated sucession comprising molasse facies and calcareous facies is represented by the Makkaur Formation and Syltefjord Formation. The general picture, therefore, is of a sedimentary basin, deep at the beginning but becoming gradually shallower, being infilled by sediments the accumulation of which was more rapid than the basin's rate of subsidence. Periods of restriction of clastic sediment supply are recorded by calcareous facies. In general terms this assemblage would seem to be representative of the sediments which usually characterize the cycle of development of a geosyncline (miogeosyncline). The lower part of th Raggo Group (Løkvikfjell Formation) may be interpreted as a molasse facies, also connected with an orogenic cycle. The sedimentary environment of the upper part of this group (Berlevåg Formation) is difficult to elucidate because of secondary, metamorphic changes.

Evidence indicating an Eocambrian and Latest Precambrian (i.e. Hyperborean) age for the sedimentary rocks of the Varanger Peninsula concerns only the sequences present in the Tanafjord - Varangerfjord region and cannot be used in considerations of the age of sedimenary sequences of the Barents Sea region. As mentioned previously, there is no lithostratographical correlation between these regions, separated from each other as they are by a large thrust.

No evidence for the age of any part of either the Barents Sea Group or the Raggo Group has so far been obtained. These rock sequences may indeed be placed anywhere in the vast period of time from Late Precambrian up to Silurian. The Caledonian deformation of the above mentioned group is characterized by a NE-SW trend of many structures and by a slight metamorphism of the upper part of the Raggo Group. Considerations concerning a more exact assignment of age for the rocks in question may at this stage be based only on the unsatisfactory method of comparison with rock sequences in adjacent areas.

Comparison with the Raipas Group, suggesed by Holtedahl (1918), gives unsatisfactory results because the age of the latter is not definitely established, referred either to "Esmarkian" (Rosendahl, 1945; Spjeldnæs, 1964) or to younger Precambrian (Holtedahl, 1960; Reitan, 1960, p. 92; 1963).

A comparison with the allochthonous block of the Ribačij Peninsula, even if correct, still does not solve the problem of age of the sedimentary rocks of the Barents Sea region. An age for the rocks of the former region has been given as Hyperborean by Polkanov (1934) on a basis of comparison with sedimentary rocks of Eocambrian age on Varanger Peninsula(!). However tillites have been recorded (Wegmann, 1928, 1929; Lupander, 1934) on

Ribačij Peninsula, presumably within the sedimentary succession recently termed "Svita Ejna" (description of the geological section after Agapiev and Vronko in Keller et. al., 1963, p. 111). Agapiev and Vronko (l.c.) make no mention of tillites, while Keller and Sokolov (1960) have argued against a glacial origin for conglomerates previously called tillites. Later, however, (Lungersgauzen, 1963, p. 567) the typically glacial character of tillites and tillite-like deposits of Ribačij, Sredni and Kildin has been accepted. This opinion, in all probability based on new observation, can be helpful in establishing the age of sediments from Ribačij as well as the Barents Sea region. However, it should be emphasized that the tillites of Ribačij, if they do exist, may not necessarily correspond exactly in age to those from Finnmark but could represent another, independent horizon.

Absolute age determinations of sedimentary rocks from eastern Finnmark have not yet been carried out, but Polkanov and Gerling (1960) have published data concerning absolute ages of rocks fra. Sredni, Kildin and Ribačij. The absolute age of the glauconitic sandstone of the "Kildinskaja svita" on Sredni came out at 920 mill. years, on Kildin 1030 - 1010 mill. years. An absolute age of phyllites from Ribačij has been established as 887 - 715 mill. years.

In any consideration of age of rock sequences from the Barents Sea region, some previous, broad regional comparisons should be mentioned. Ramsay (1897-99, 1911), Reusch (1920) and Tschernyschew (1902) suggested the occurrence of an old mountain chain, the remnants of which are now represented by the Timan mountains, Kanin Peninsula, Kildin Island, Ribačij (with Sredni) Peninsula and Varanger Peninsula. Holtedahl (1918), however, did not accept the idea of an Timanian chain in Finnmark. Later, Schatsky (1958) defined the Timanides as a region including Timan, Kanin, Kildin and Ribačij, which was deformed during the Baikalian disturbance - older than Caledonian - and Stille (1958) regarded the same region as probably representative of an Assyntian geosyncline extending westwards as far as Finnmark. It would, therefore, seem that the Barents Sea region, which occupies an isolated position in the geology of Finnmark, should be regarded as a part of the Timanian miogeosyncline.

Riphean rock sequences of Kanin and Timan are metamorphosed and cut by large bodies of igneous rocks, and because of this a detailed lithological similarity to the sections across the Barents Sea region cannot be expected. Only the upper, less metamorphosed division of the Riphean of Timan (Zhuravlev and Gafarov, 1959), consisting of conglomerates, quartzites, slates, dolomites and algal limestones may be loosely compared with the rocks of the Barents Sea region. The better preserved, typical sections of miogeosynclinal Riphean are those of the western slopes of the Southern Urals (Schatsky, 1945, 1958, 1960). This ca. 10.000 m thick assemblage consisting of successions of variegated clastic sediments and dolomites is, quite possible, a distant but similarly developed lithostratigraphical equivalent of the sedimentary sequences of the Barents Sea region. It should be mentioned that the Riphean of Kanin and Timan is also regarded as miosynclinal, similar to that of the W slopes of the Southern Urals (Zhuravlev and Gafarov, 1959).

In conclusion, the above considerations concerning the rocks of the Barents Sea region point to an age rather older than Cambrian (i.e. Late Precambrian and Eocambrian, Sparagmitian, Hyperborean, Riphean etc.), although the authors have a clear understanding of the hypothetical nature of this statement.

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Norges geologiske undersøkelse legger hermed frem årsberetningen for 1966, det første hele året styret var i funksjon. Den skiller seg ikke meget ut fra redaksjonen av de foregående års beretninger.

Trondheim, 21. august 1967.

Jens A. W. Bugge
styrets formann

Karl Ingvaldsen
adm. direktør

Innhold.

Årsberetning for 1966

NGU's administrasjon.

Styret for Norges geologiske undersøkelse har i 1966 hatt følgende sammensetning:

Professor Jens A. W. Bugge, Universitetet i Oslo, formann,
Direktør Olav Øverlie, Christiania Spigerverk, varaformann,
Professor Niels-Henrik Kolderup, Universitetet i Bergen,
Administrerende direktør Karl Ingvaldsen, NGU,
Professor Steinar Skjeseth, NLH, Ås.

Varamenn:
Direktør Leiv Løvold, Folldal Verk A/S, Oslo,
Professor Rolf Selmer-Olsen, NTH, Trondheim.

I 1966 ble det holdt 6 ordinære styremøter.

Statutter for NGU, instruks for styret og ny instruks for adm. direktør er fastsatt av Industridepartementet. Nytt ansettelsesreglement vil foreligge først senere.

Den daglige ledelse ved NGU har i 1966 foruten adm. direktør, siv.ing. Karl Ingvaldsen, bestått av direktør dr. philos. Harald Bjørlykke ved Geologisk avdeling, direktør c. r. Inge Aalstad ved Geofysisk avdeling og direktør siv.ing. Aslak Kvalheim ved Kjemisk avdeling. Kontorsjef er c. j. Per Kr. Gundersen. Direktør Bjørlykke ble sykepermittert fra 11. november 1966 og senere har statsgeolog Thor L. Sverdrup delvis dekket førstnevntes funksjoner.

Personale.

Ansettelser i 1966.

Administrasjonskontoret:

*Ristan, Anne Margrethe, kontorfullm. I, 1. januar
Widerøe, Grethe, kontorassistent I, 20. juni, midl.
Møystad, Torill, kontorassistent I, 12. september
Bakken, Solveig, kontorassistent II, 12. mai

Geologisk avdeling:

*Wolff, Christian Fredrik, statsgeolog I, 1. januar
Juve, Gunnar, vitenskapelig assistent I, 1. januar
*Sørensen, Erling, konstruktør II, 1. januar
*Nergaard, Lajla, tegner I, 1. januar
*Holiløkk, Lars, laborant I, 1. januar
Evensen, Bina, tegner II, 17. januar
Johansen, Bjørn, kontorassistent II, 3. februar, midl.
*Forbordsaune, Johan, laborant I, 1. april
Hultin, Ivar, vitenskapelig assistent I, 1. april
Iversen, Bjørn Sverre, kontorassistent II, 1. september

Geofysisk avdeling:

Solvang, Terje, tegner I, 1. januar
*Godø, Rolf, tegner I, 1. januar
*Andreassen, Tore, tegner I, 1. juni
*Sagflaat, Hans, tegner, 1. februar, midl.
*Danielsen, Birgith, tegner, 1. februar, midl.
*Opsahl, Henrik, konstruktør II, 1. mars, vikariat
*Østby, Solveig, tegner II, 1. mai
Dalsegg, Einar, teknisk assistent I, 1. mai, vikariat

Kjemisk avdeling:

*Taftøy, Inger, laboratorieassistent II, 1. februar
*Holmberget, Edna, tegner I, 1. mars

* Disse personer er gått over fra annen stilling eller engasjement ved institusjonen.

Avskjed i 1966.

Administrasjonskontoret:

Skaanes, Kari, kontorfullm. I, 1. juli
Dalsbø, Gunn Helene, kontorassistent I, 11. april
Lunde, Kari, kontorassistent I, 1. august

Geologisk avdeling:

Johansen, Bjørn, kontorassistent II, 10. juli

Geofysisk avdeling:

Gran, Kjell, tegner I, 20. april

Kjemisk avdeling:
Holmen, Asbjørn, lab.ass. II, 1. januar

Ved utgangen av 1966 hadde NGU følgende personale i heldagsstilling: (Den oppførte ansettelsesdato angir tidspunktet da vedkommende ble knyttet til NGU i hovedstilling.)

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Ingvaldsen, Karl, siv.ing., a. 1. januar 1958

Bergingeniør:
Welde, Harald, siv.ing., a. 1. januar 1965

Kontorsjef:
Gundersen, Per Kristian, c. j., a. 1. oktober 1960

Forvalter:
Thorvaldsen, Arvid, a. 1. juli 1956

Bibliotekar:
Ryssdal, Marit, a. 1. oktober 1963

Fotograf:
Aamo, Ingemar, a. 1. august 1962

Regnskapsfører:
Hanssen, Alf, a. 1. august 1955

Kasserer:
Nygård, Hjørdis, a. 17. juli 1961

Kontorfullmektig I:
Ristan, Anne Margrethe, a. 1. mai 1961

Kontorassistent/fullmektig II:
Aursand, Marit, a. 1. august 1965
Widerøe, Grethe, a. 20. juni 1966, midl.
Møystad, Torill, a. 12. september 1966
Bakken, Solveig, a. 12. mai 1966

Vakt- og varmemester:
Wold, Jostein, a. 15. august 1961

Geologisk avdeling:

Direktør:
Bjørlykke, Harald, dr. philos., a. 1. august 1958

Statsgeolog I:
Broch, Olaf Anton, c. r., a. 1. juli 1930
Holmsen, Per, c. r., a. 1. juli 1939
Hagemann, Fredrik, c. r., a. 1. mars 1957, midl. tjenestefri fra 1. desember
Sverdrup, Thor Lorck, c. r., a. 16. november 1958
Bryn, Knut Ørn, c. r., a. 1. januar 1959
Carstens, Harald, dr. philos., a. 1. desember 1963
Wolff, Christian Fredrik, c. r., a. 16. februar 1960

Statsgeolog II:
Skålvoll, Harald, c. r., a. 1. juli 1957
Thorkildsen, Christian Dick, c. r., a. 1. februar 1960
Gustavson, Magne, c. r., a. 1. januar 1961
Hysingjord, Jens, c. r., a. 15. august 1961
Gvein, Øyvind, c. r., a. 11. desember 1963

Midlertidig statsgeolog:
Poulsen, Arthur I., cand. min.
Holmsen, Gunnar, dr. philos.

Vitenskapelig assistent:
Kollung, Sigbjørn Jarle, c. r., a. 1. april 1961
Nissen, August, c. r., a. 1. januar 1964
Englund, Jens Olaf, c. r., a. 3. oktober 1964
Hovland, Roar, siv.ing., a. 1. april 1965
Reite, Arne, c. r., a. 5. april 1965
Rye, Noralf, c. r., a. 1. juli 1965
Kildal, Ellen Sigmond, c. r., a. 1. oktober 1965, midl.
Juve, Gunnar, c. r., a. 1. januar 1966
Hultin, Ivar, c. r., a. 1. august 1966

Laboratorieingeniør I:
Graff, Per-Reidar, c. r., a. 1. april 1964

Konstruktør II:
Klemetsrud, Harald Tidemann, a. 1. juli 1957
Sørensen, Erling, a. 1. mai 1963

Teknisk assistent I:
Hatling, Harald, a. 1. februar 1961
Gust, Johan, a. 1. oktober 1962
Røste, Johannes Rye, a. 9. desember 1963

Preparant I:
Jacobsen, Tom, a. 1. mai 1962
Iversen, Egil, a. 1. august 1965

Laborant I:
Aarsland, Edvard P., a. 1. januar 1959
Holiløkk, Lars, a. 1. juni 1959
Forbordsaune, Johan, a. 1. januar 1961

Tegner I:
Vikholt, Halfrid, a. 1. mars 1955
Nergaard, Lajla, a. 1. januar 1962

Tegner II:
Lund, Astri, a. 1. januar 1962
Evensen, Bina, a. 17. januar 1966
Hemming, Beret, a. 1. desember 1966

Sekretær I:
Møller, Laura, a. 1. april 1961

Kontorassistent/fullmektig II:
Anderssen, Gunhild, a. 1. januar 1962
Teige, Astri, a. 18. februar 1964, midl.
Iversen, Bjørn Sverre, a. 1. september 1966, midl.

En del geologer ved andre institusjoner og viderekomne studenter har vært knyttet til avdelingen som vitenskapelige medarbeidere under sommerens markarbeid. Videre har diverse personell vært ansatt i korttidsengasjementer.

Geofysisk avdeling:

Direktør:

Aalstad, Inge, c. r., a. 1. oktober 1962 (15. juli 1952)

Geofysiker I:

Sakshaug, Gunnar, siv.ing., a. 1. juli 1936
Singsaas, Per, a. 1. september 1937
Hillestad, Gustav, siv.ing., a. 20. januar 1953

Fysiker I:

Breen, Arne, siv.ing., a. 1. desember 1940

Geolog I:

Svinndal, Sverre, c. r., a. 1. juli 1961

Geofysiker II:

Moxnes, Hans Petter, c. r., a. 6. juli 1959
Håbrekke, Henrik, siv.ing., a. 17. august 1959
Sindre, Atle, c. r., a. 24. mai 1961

Geolog II:

Tan, Tek Hong, c. r. (nederl. eks.), a. 23. april 1959
Barkey, Henri, c. r. (nederl. eks.), a. 1. desember 1963, midl.

Konstruktør I:

Uddu, Odd, a. 1. oktober 1952
Brandhaug, Kolbjørn, a. 1. september 1958
Haugan, Arne, a. 1. juni 1961

Konstruktør II:

Dalsaune, Einar, a. 1. juli 1952, midl. tjenestefri
Opsahl, Henrik, a. 21. april 1953, vikar

Borformann:

Bratli, Johannes, a. 1. januar 1953
Gausdal, Odd, a. 20. september 1957
Vassbotn, Sven, a. 1. september 1963

Teknisk assistent I:

Dalsegg, Einar, a. 1. mai 1966, vikar

Tekniker I:
Melleby, Peter, a. 14. november 1955
Blokkum, Oddvar, a. 17. januar 1961

Tekniker II:
Staw, Jomar, a. 18. juni 1956

Laborant II:
Opdahl, Ragnar, a. 23. oktober 1957

Laboratorieassistent II:
Johansen, Hermann, a. 1. april 1963, midl.

Tegner I:
Grønli, Gunnar, a. 12. januar 1956
Solvang, Terje, a. 1. januar 1961
Godø, Rolf, a. 1. januar 1966
Andreassen, Tore, a. 1. januar 1966

Tegner II:
Haugen, Torbjørn, a. 3. juni 1959
Østby, Solveig, a. 14. august 1961

Tegner:
Sagflaat, Hans, a. 1. februar 1966, midl.
Danielsen, Birgith, a. 1. februar 1966, midl.

Mekanikerformann:
Skauge, Ole, a. 1. oktober 1937

Mekaniker:
Brevik, Bjørn, a. 1. mai 1939
Pettersen, Reidar, a. 25. mars 1952
Gravseth, Odd, a. 10. november 1953

Instrumentmaker:
Kirkeby, Kåre, a. 15. september 1951

Verkstedarbeider:
Tetli, Alf, a. 1. oktober 1958

Snekker:

Pettersen, Norman, a. 18. februar 1946

Sekretær I:

Singsaas, Cathrine, a. 1. oktober 1953

Kontorassistent/fullmektig II:

Wettavik, Vigdis, a. 1. mars 1964

Bud og betjent:

Hillesund, Tove, vikar fra 1. januar 1966

Avdelingen har 1 tegner i deltidsstilling. Videre har en del personell vært knyttet til institusjonen på annen måte, f. eks. i korttidsengasjementer som tegnere. Ved avdelingen har det dessuten vært engasjert ekstra personell til feltarbeidet.

Kjemisk avdeling:

Direktør:

Kvalheim, Aslak, siv.ing., a. 1. oktober 1947 (1. oktober 1937)

Laboratorieingeniør I:

Grennes, Johannes, siv.ing., a. 1. mai 1943
Aarvik, Jon, siv.ing., a. 25. august 1950
Faye, Gjert Chr., siv.ing., a. 10. desember 1958
Nilsen, Rolf, siv.ing., a. 1. april 1963
Andreassen, Birger Th., siv.ing., a. 15. februar 1961

Geokjemiker I:

Bølviken, Bjørn, siv.ing., a. 1. mars 1954, midl. tjenestefri

Laboratorieingeniør II:,

Ødegård, Magne, siv.ing., a. 1. mai 1961
Krog, Jan Reidar, siv.ing., a. 1. mai 1964
Stige, Leif, c.r., a. 4. januar 1965

Geokjemiker II:

Hvatum, Ole Ø., siv.agr., a. 1. april 1961

Konstruktør I:
Berner, Beate, a. 4. januar 1955
Næss, Gunnar, a. 16. januar 1960
Solem, Knut, a. 1. januar 1961
Flårønning, Asbjørn, a. 1. juni 1964

Konstruktør III:
Sivertsen, Tove, a. 9. januar 1958

Teknisk assistent I:
Bremseth, Asbjørn, a. 9. november 1959
Wik, John M., a. 23. november 1953

Tegner I:
Holmberget, Edna, a. 1. september 1960

Laborant I:
Horgmo, Birger, a. 1. mars 1953
Ekremsæter, Jørgen, a. 1. september 1960
Wolden, Odd, a. 1. mars 1963
Storvik, Arne, a. 1. mars 1964
Kalvøy, Henry, a. 24. mai 1965

Laboratorieassistent I:
Skarholt, Siri, a. 1. januar 1961

Laboratorieassistent II:
Tan, Brith, a. 1. juni 1963
Taftøy, Inger, a. 1. februar 1966

Sekretær I:
Bersvendsen, Jørgen H., a. 1. juni 1957

Kontorassistent/fullmektig II:
Støren, Erna-Beate, a. 29. november 1965

Kjemisk avdeling har dessuten pr. 31/12 i tjeneste 6 praktikanter.

Ved utgangen av 1966 hadde NGU 139 stillinger, hvorav 122 fast organiserte stillinger og 17 helårsengasjementer. Ved budsjettbehandlingen for 1967 fikk NGU innvilget fast organiserte stillinger som statsgeolog II og laborant I til Geologisk avdeling og laboratorieassistent I ved Kjemisk avdeling ved siden av stilling som teknisk assistent ved Geofysisk avdeling i helårsengasjement.

Bidjovaggeundersøkelsene hadde ved utgangen av 1966 tre medarbeidere i helårsengasjement: Geolog Carl O. Mathiesen, bergingeniør Paul J. Paulsen og sekretær Morten Sandvold.

Geokjemiker B. Bølviken fikk forlenget sin permisjon med ca. 1 år for medvirkning i FN-oppdrag i Equador inntil 1. april 1967.

Statsgeolog F. Hagemann er tilstått permisjon i inntil 1 år fra 1/12 1966 i forbindelse med et stipendium under Industridepartementet for en oljegeolog. Statsgeolog K. Ø. Bryn har under permisjonen overtatt som daglig leder av Oslokontoret.

Budsjett og regnskap.

Statsbudsjettets kap. 3943		Budsjett		Regnskap
Inntekter:				
1. Oppdragsinntekter	kr.	570.000,—	kr.	1.018.100,39
2. Salg av kart og publikasjoner	»	10.000,—	»	16.248,—
3. Salg av instrumenter	»	10.000,—	»	74.781,83
4. Andre inntekter	»	10.000,—	»	141.539,70
	kr.	600.000,—	kr.	1.250.669,92

Statsbudsjettets kap. 943				
Utgifter:				
01. Lønninger	kr.	4.153.600,—	kr.	4.414.979,36
10. Kjøp av kontorutstyr	»	34.500,—	»	34.679,93
11. Kjøp av feltutstyr	»	112.000,00	»	111.357,41
12. Kjøp av instrumenter	»	132.000,00	»	131.802,13
13. Kjøp av maskiner og transportutstyr	»	49.500,—	»	43.378,26
15. Vedlikehold	»	100.000,—	»	94.176,10
29. Andre driftsutgifter				
291. Kontorutgifter	»	145.000,—	»	160.765,07

292. Trykningsutgifter	kr.	115.000,—	kr.	83.023,13
293. Bygningers drift	»	161.000,—	»	205.211,31
294. Reise- og forpleiningsutg.	»	665.000,—	»	672.305,79
295. Forbruksvarer	»	371.000,—	»	393.253,89
296. Ymse driftsutgifter	»	543.000,—	»	545.685,68
	kr.	6.581.600,—	kr.	6.890.618,06

Statsbudsjettets kap. 945

20. Undersøkelser	kr.	1.724.103,50	kr.	1.284.633,64
Hydrologisk dekade	kr.	230.000,—	kr.	22.747,21

Det ordinære utgiftsbudsjett for NGU i 1966 var på kr. 6.581.600,—, mens regnskapet viste en sum på kr. 6.890.618,06. Budsjetterte oppdragsinntekter m. v. for 1966 var kr. 600.000,—. I alt ble det oppnådd inntekter på kr. 1.250.669,92 idet flere grener innen oppdragsvirksomheten og salg av instrumenter og geofysiske kart økte ganske vesentlig.

Ved Stortingets bevilgning 17. juni 1966 fikk Bidjovaggeundersøkelsene 1,6 mill. kroner i tillegg til kr. 124.103,50 overført fra det foregående budsjettår.

Fra virksomheten i 1966.

Bidjovaggeundersøkelsene.

I 1966 fortsatte undersøkelsene i Bidjovagge med diamantboringer på lokaliteter hvor en kunne vente å finne ytterligere malm. Det førte ikke til noen økning i malmberegningen fra januar 1966. En ny malmskjæring på 300 m dyp ble oppnådd i den nordlige del av feltet. Etter feltsesongens slutt er rapportmaterialet stillet til disposisjon for de interessenter som har meldt seg for Industridepartementet.

Oppdragsvirksomhet og nye forskningsoppgaver ved en statsinstitusjon.

NGU har fra gammelt av hatt som en av sine viktigste oppgaver å betjene norsk næringsliv med oppdragsarbeider av forskjellig art. En statsinstitusjon som NGU kan, i motsetning til institutter under forskningsrådene, ikke selv disponere oppdragsinntektene selv om disse overskrider de budsjetterte.

Dette forhold er en betydelig ulempe for institusjonen og hindrer NGU i ikke liten grad å følge med på aktuelle arbeidsfelter. Det medfører et unødig handicap i disposisjonene ved NGU og forholdet ble berørt under Industrikomitéens besøk ved NGU i august. Styret har derfor fremmet en sak om at det ved NGU bør opprettes et disposisjonsfond hvorved en mindre del av budsjettet kan disponeres av styret til f. eks. engasjement av personell, anskaffelse av utstyr og dekning av driftsutgifter ved prosjekter og forskningsoppgaver hvor NGU bør engasjere seg. Uten en slik «modernisering» i forvaltningen ved institusjonen, vil NGU ikke ha muligheter for å bearbeide viktige saker som samfunnet har krav på at NGU forsøker å løse.

Samarbeidsutvalg.

I 1966 hadde Samarbeidsutvalget ved NGU 4 møter. Utvalget hadde ved utgangen av året følgende representanter:

For administrasjonen:

K. Ingvaldsen, nestformann	— varamann:	Per Kr. Gundersen
H. Bjørlykke	»	T. Sverdrup
I. Aalstad	»	G. Hillestad
A. Kvalheim	»	G. Faye

Representanter for de ansatte:

S. Svinndal, formann	— varamann:	M. Gustavson
K. Solem	»	J. Bersvendsen
A. Hanssen	»	O. Uddu
R. Pettersen	»	T. Solvang

Ordningen med regelmessig legekontroll av personalet ved NGU kom i gang i første kvartal, og tiltaket har vist seg å være meget gunstig. Kontrollen utføres av dr. Hans Sejnæs. NGU-nytt — en intern informasjon ved NGU — kom ut med to nummer i 1966. I det påfølgende år regnes med 2 nummer hvert halvår. Bladet er blitt mottatt med interesse både av de ansatte og av personer utenfor institusjonen.

Biblioteket.

Tilveksten på periodisk litteratur var 1.332 bind og samlet antall pr. 31/12 1966 er 32.279 bind.

Boktilveksten var på 235 bind og samlet antall ved utgangen av året er 2420.

NGU har sluttet 10 nye bytteavtaler og har nå i alt ca. 290 slike avtaler.

Møter ved NGU.

Stortingets industrikomité besøkte Norges geologiske undersøkelse 24. august 1966, og fikk herunder, foruten en orientering om NGU og flere av dens aktiviteter, mulighet for nærmere informasjon om personale, utstyr og arbeidsoppgaver. Stortingskomitéen besøkte Geofysisk malmleting og Statens råstofflaboratorium forrige gang i 1958.

Bransjerådet for bergverkene hadde møte i Trondheim 13. og 14. september 1966 og forhandlingene var henlagt til NGU. NTNF's forskningsutredning 1964 og råstoffundersøkelser var fremtredende emner på programmet. Bransjerådet fikk samtidig en demonstrasjon av virksomheten ved NGU.

Utenlandsreiser og møter i utlandet.

Statsgeologene F. Hagemann, T. L. Sverdrup, F. Chr. Wolff, J. Hysingjord, M. Gustavson, H. Skålvoll, Ø. Gvein, geologene S. Svinndal og H. Barkey. geokjemiker Ø. Hvatum samt de vitenskapelige assistenter G. Juve, E. Kildal, A. Reite, J. O. Englund og L. A. Kirkhusmo deltok alle i det VII. nordiske geologiske vintermøte i Åbo 7.—9. januar.

Direktør I. Aalstad, fysiker A. Breen, geofysikerne G. Sakshaug, P. Singsaas, G. Hillestad, H. Håbrekke, H. P. Moxnes og geologene S. Svinndal og H. Barkey deltok i det V. nordiske geofysiske vintermøte som ble avholdt i Stockholm 11.—13. januar.

Statsgeolog F. Hagemann deltok i perioden 28. febr.—3. mars i det I. nordiske møte i den Internasjonale Hydrologiske Dekade i Hesselby, Sverige.

Geofysiker H. P. Moxnes deltok i et møte som komitéen for automatisk databehandling av geofysiske observasjoner avholdt i Stockholm 28. april.

Statsgeolog Chr. D. Thorkildsen foretok i perioden 18.—28. mai en studietur til Sverige og Finland for å se på geokronologiske laboratorier.

De vitenskapelige assistenter A. Reite og N. Rye deltok i tiden 18.—28. mai i en kvartærgeologisk ekskursjon i Danmark, arrangert av Geologisk Institutt, Universitetet i Bergen.

European Association of Exploration Geophysicists avholdt sitt årlige møte i Amsterdam i dagene 15.—17. juni. Geofysikerne G. Hillestad og A. Sindre samt fysiker A. Breen deltok i dette møte.

Direktør H. Bjørlykke samt statsgeologene T. L. Sverdrup og F. Chr. Wolff deltok i et møte vedrørende de internasjonale geologiske kart (geologiske, tektoniske og metallogene). Møtet foregikk i Paris i tiden 20.—30. juni.

Statsgeologene H. Carstens og M. Gustavson deltok i den skandinaviske vulkanologiske ekskursjon på Island 31. juli—15. august.

Adm. direktør K. Ingvaldsen deltok i det årlige nordiske direktørmøte i Kiruna, Sverige, 4.—6. august.

Statsgeologene H. Skålvoll og Ø. Gvein deltok i en ekskursjon i Syd- og Mellom-Sverige i tiden 15.—17. august.

Statsgeolog H. Carstens deltok i et symposium arrangert av IMA i Cambridge 29. august—3. september.

De vitenskapelige assistenter A. Reite og N. Rye har deltatt i en ekskursjon i Jämtland i tiden 29. aug.—4. sept. for å studere de klassiske isavsmeltningsområder.

Likeledes foretok statsgeolog P. Holmsen sammen med dr. Jan Lundqvist, SGU, en lenge planlagt reise til de samme isavsmeltningsområder.

Geolog S. Svinndal deltok sammen med geolog H. Barkey i den kongress som organisasjonen International Society of Rock Mechanics avholdt i Lisboa i tiden 25. september—1. oktober.

Statsgeologene F. Hagemann og K. Ø. Bryn deltok i et symposium over grunnvann, avholdt i Stockholm 11.—13. oktober i forbindelse med den Internasjonale Hydrologiske Dekade.

Vitenskapelig assistent R. Hovland deltok i Bergingeniørenes møte som ble holdt i Stockholm 12.—14. oktober.

Statsgeolog F. Hagemann deltok 17.—19. oktober i Arbeidsgruppe for brønnarkiv, den Internasjonale Hydrologiske Dekade, København.

Direktør A. Kvalheim og laboratorieingeniør J. Krog deltok 9.—10. november i en konferanse arrangert i Stockholm av Sveriges geologiska undersøkning hvor emnet var geokjemisk prospektering, og hvor deltakerne var fra Finland, Sverige og Norge.

Direktør A. Kvalheim besøkte i tiden 12.—16. desember Centre de Recherches Petroprahique et Geochemique i Nancy, Frankrike, for utveksling av erfaringer og arbeidsmetoder.

Publikasjoner.

Redaktør for NGU's publikasjoner er statsgeolog Fredrik Hagemann.

235. Ottar Jøssang: Geologiske og petrografiske undersøkelser i Modumfeltet.
236. Harald Bjørlykke: De alluviale gullforekomster i indre Finnmark.
237. Viggo Wiik: Petrological Studies of the Neiden Granite Complex.
238. Studies on the Latest Precambrian and Eocambrian Rocks in Norway. No. 1. Sedimentary petrology of the Sparagmites of the Rena district, S. Norway. By Knut Bjørlykke. No. 2.

Sparagmittgruppens bergarter ved Fåvang, Gudbrandsdalen. En sedimentologisk og tektonisk undersøkelse. Av Jens-Olaf Englund.

239. Magne Gustavson: The Caledonian Mountain Chain in the Southern Troms and Ofoten Areas. Part I. Basement Rocks and Caledonian Metasediments.
240. Ivar Ramberg: Kongsfjell-områdets geologi — en strukturell og petrografisk undersøkelse i Helgeland, Nord-Norge.
241. Inge Bryhni: Reconnaissance Studies of Gneisses, Ultrabasites, Eclogites and Anorthosites in Outer Nordfjord, Western Norway.
242. Årbok 1965. Innhold: Jens-Olaf Englund: Grunnvann i sparagmittgruppens bergarter i Syd-Norge. Meddelelser fra Vannboringsarkivet nr. 14. Torgeir Falkum: Structural and Petrological Investigations of the Precambrian Metamorphic and Igneous Charnockite and Migmatite Complex in the Flekkefjord Area, Southern Norway. (A preliminary report). Rolf W. Feyling-Hanssen: Geologiske observasjoner i Sandnesområdet. Fredrik Hagemann: Silurian Bentonites in the Oslo Region. Michael Holmes: Structure of the area north of Ørnes, Nordland, Norway. Gunnar Holmsen: Minner fra geologisk feltarbeid i Nordland for 60 år siden. Per Holmsen og Steinar Skjeseth: Trysilhevningen mellom Osensjøen og Jordet i Trysil. En foreløpig meddelelse. Med kort beskrivelse av diamantborkjerne, ved J. P. Nystuen. Ellen Sigmond Kildal: Note on the geology of the archipelago NW of Bergen ("Øygarden"). M. Mortensen: Aldersbestemmelse av "fossil furu" fra Nord-Østerdal. Hans P. Moxnes og Øivind Solvang: Beregning av terrengkorreksjoner med elektronisk regnemaskin og fotogrammetrisk utstyr. Robin Nicholson: On the relations between volcanic and other rocks in the fossiliferous east Lomivann area of Norwegian Sulitjelma, Announcement. Årsberetning for 1965. Fortegnelse over publikasjoner og kart.

De 5 siste publikasjonene ble noe forsinket under trykkingen og forelå i handelen først litt ut i 1967.

Geologisk avdeling.

Statsgeolog Th. L. Sverdrup har fra 11/11—66 vært bemyndiget som daglig leder ved Geologisk avdeling under direktør H. Bjørlykkes sykdom.

Berggrunnskartlegging.

Seksjonens leder er statsgeolog F. Chr. Wolff.

Ved seksjonen har i 1966 følgende arbeidet: Statsgeologene H. Carstens, F. Chr. Wolff, H. Skålvoll og M. Gustavson samt de vitenskapelige assistentene S. Kollung og E. S. Kildal (Bergen). I tillegg har seksjonen hatt følgende engasjerte medarbeidere: Dr. David Roberts, dr. Anna Siedlecka og professor Stanislaw Siedlecki.

Seksjonen har dessuten hatt en rekke sommermedarbeidere fra inn- og utland. En del av geologene fra andre seksjoner innen institusjonen har utført berggrunnskartlegging i sommerhalvåret.

Det er i 1966 utført berggrunnskartlegging innen følgende 1 : 250 000 blad (AMS serie M 515):

Vadsø: På gradteigene (1 : 100 000) Berlevåg, Tana, Makkaur, Båtsfjord av engasjert statsgeolog S. Siedlecki og engasjert vit. ass. A. Siedlecka.

Karasjok: På gradteigene Bæivasgiedde og Isskuras av statsgeolog M. Skålvoll (sammen med den finske geolog Topi Poutto).

Mosjøen: På 1 : 50 000 bladene (AMS serie M 711) Drevja, Elsfjord, Nesna og Mosjøen av vit. ass. A. Nissen.

Trondheim: På rektangelbladene (1 : 100 000) Meråker, Stjørdal og Essandsjø av følgende: Statsgeolog F. Chr. Wolff (prosjektleder), dr. D. Roberts, dr. A. Siedlecka, professor S. Siedlecki, dr. Z. Pelc, dr. M. Fisera. Arbeidet er koordinert med malmgeologiske undersøkelser som drives av Skorovas Gruber A/S. Kartlegging på rektangelbladet Trollhetta er utført av vit. ass. S. Kollung.

Hamar m. Torsby: På 1 : 50 000-bladet Finnskog av statsgeolog F. Chr. Wolff, på blad Flisa av dr. R. Roberts, på blad Strøm av statsgeolog J. Hysingjord og på 1 : 50 000 Kongsvinger av statsgeolog Ø. Gvein.

Lillehammer: De vitenskapelige assistenter J. O. Englund og L. A. Kirkhusmo ved hydrogeologisk avdeling samt stud. real. H. C. Seip har sammen med sommermedarbeidere fortsatt kartleggingen under ledelse av professor S. Skjeseth ved Norges landbrukshøyskole.

Årdal: På 1 : 50 000-bladene Tyin, Øye, Vang og Hurrungane av statsgeolog M. Gustavson, og på bladet Borgund av vit. ass. E. S. Kildal (en kortere tid).

Oslo: Orienterende kartlegging på 1 : 50 000-bladene Rødenes og Øymark ved statsgeologene H. Skålvoll og M. Gustavson. Kartlegging på blad Setskog av M. Gustavson.

Mandal: På gradteig D37V i Setesdal av vit. ass. S. Kollung. Statsgeolog H. Carstens har arbeidet i Egersundsfeltet.

Sauda: På gradteig Sand ved vit. ass. E. S. Kildal.

Måløy: På gradteigene Lavik og Kyrkjebø av vit. ass. E. S. Kildal, som også har vært prosjektleder for det arbeide som, med støtte fra NGU på dette kartblad, utføres av studenter ved Universitetet i Bergen.

Av 1 : 250 000-bladene er Måløy ferdig kartlagt og vil bli trykt i løpet av 1967.

Blad Pechenga (Kirkenes) er ferdig kartlagt, men ennå ikke sammentegnet. Grong-bladet er på det nærmeste ferdig kartlagt.

Kvartærgeologisk kartlegging.

Ved den kvartærgeologiske seksjonen har følgende arbeidet i 1966: Statsgeolog P. Holmsen, vit. assistenter A. Reite og N. Rye. Videre har pensjonert statsgeolog G. Holmsen vært tilknyttet seksjonen.

Statsgeolog P. Holmsen har sammen med assistenter videreført arbeidet på landgeneralkartet Jotunheimen som omfatter 15 norske gradteiger, eller 28 av AMS-kartene (NATO 1 : 50 000). Statsgeolog P. Holmsen har i år vesentlig arbeidet innen kartene Borgund, Hurrungane, Tyin, Sygnefjell og Visdalen.

Vit. ass. N. Rye har fortsatt arbeidet med undersøkelse av økonomisk viktige sand- og grusforekomster i Vest-Norge i samarbeid med dr. G. Holmsen. Feltarbeidet har foregått vesentlig i Rogaland, Møre og Romsdal og har gått ut på å oversiktskartlegge større forekomster.

Vit. ass. A. Reite har fortsatt den kvartærgeologiske kartlegging innen AMS-bladet NP 31, 32, Trondheim.

Pensjonert statsgeolog, dr. G. Holmsen har i inneværende år fortsatt arbeidet med å systematisere norske grusforekomster.

Vit. ass. A. Reite har foretatt en befaring med prøvetaking av morener i Orkdal og Nerskogen/Oppdal.

Lønnet av Trondheim kommune har ingeniørgeolog K. Kowalik arbeidet med undersøkelser av Trondheims byggegrunn.

Hydrogeologi.

Seksjonens leder er statsgeolog F. Hagemann. Under hans permisjon ledes avdelingen av statsgeolog K. Ø. Bryn. Ved seksjonen har følgende arbeidet: Statsgeologene F. Hagemann og K. Ø. Bryn, vitenskapelige assistenter J. O. Englund og engasjert vit.ass. L. A. Kirkhusmo.

Samtlige har vært beskjeftiget med oppdrag i forbindelse med grunnvannsforsyning ved boring i fast fjell og løsavleiringer.

Det er utført en rekke befaringer i forbindelse med planlegging av enkelt og felles vannforsyningsanlegg.

Konstruktør T. Klemetsrud har undersøkt mulighetene for grunnvannsforsyning fra sand- og grusavleiringer. I løpet av året er det bygget en rekke vellykkede rørbrønner forskjellige steder i landet på grunnlag av hans forundersøkelser, eksempelvis Tinfos Jernværks to rørbrønner som gir 7000 liter/min. pr. brønn.

Samarbeid er innledet med Statens Institutt for folkehelse for å undersøke sammenhengen mellom sporstoffer i drikkevann og enkelte sykdommer.

Arbeidet med å samle vannboringsarkivets data på hullkort er kommet i gang i samarbeid med Forsvarets Forskningsinstitutt.

Arbeidet i forbindelse med den Internasjonale Hydrologiske Dekade kom i gang i 1966. De nødvendige midler ble bevilget fra Konsesjonsavgiftsfondet. De fleste større arbeider vil foregå på Romeriksfeltet. Vit. ass. Geir Goffeng og vit. ass. Svein Roar Østmo er ansatt i forbindelse med dette prosjektet. Registrering av grunnvannsbrønner og observasjonshull på Romerike er foretatt, og geofysiske målinger er utført ved geofysiker G. Hillestad. Gjennom IHD er det innledet et skandinavisk samarbeid for fremstilling av hydrogeologiske kart.

Mineralske råstoffer og bygningssten.

Seksjonens leder er statsgeolog Th. L. Sverdrup. Ved seksjonen har i 1966 følgende vært ansatt: Statsgeologene Th. L. Sverdrup, J. Hysingjord, Chr. D Thorkildsen og Ø. Gvein. Vit. ass. I. Hultin har delvis arbeidet ved denne seksjon og delvis ved malmseksjonen.

Statsgeolog Th. L. Sverdrup har foretatt befaringer vedrørende kalksten i Nord-Trøndelag, og NGU har utført diamantboringer i kalken på Hylla. Videre har han foretatt befaringer vedrørende kvarts og sammen med statsgeolog J. Hysingjord har han utført et større undersøkelsesarbeid vedrørende feltspat og kvarts for Feltspatkompaniet, Evje, og Elektrokemisk A/S, Fiskå Verk. Arbeidet med kvarts-feltspatforekomster i Østfold vest for Glomma ble tilnærmet avsluttet sommeren 1966. Statsgeologene H. Skålvoll og Th. L. Sverdrup foretok ut på høsten en befaringsreise i Finnmark. Sand- og skiferforekomster, samt Raipas og Ulveryggen ble besøkt. Sammen med statsgeolog Chr. D. Thorkildsen og teknisk assistent E. Sørensen undersøkte Th. L. Sverdrup bergarter i Vest-Agder og Rogaland med henblikk på sten til vegformål.

Statsgeolog J. Hysingjord har foretatt undersøkelser av feltspat ved Herefoss, gabbro og kragerøitt i Kragerødistriktet, kvarts og feltspat i Rakkestad og olivinsten i Kjølsdalen. Han har i sommerhalvåret arbeidet med kvarts og kvartsitt i Trøndelag, Møre og Nordland, dessuten ledet diamantboringene etter kvarts og feltspat i Evje og nær Lyngdal og endelig foretatt reiser for innsamling av bergarter og elvesand for tungmineralundersøkelser.

Statsgeolog Chr. D. Thorkildsen har foruten bearbeiding av innsamlet materiale, foretatt innsamling av materiale for aldersbestemmelser. Videre har han foretatt befaringer av skiferfelter i Sør-Trøndelag, marmor på Roan i ytre Trøndelag og kleberstensforekomst i Målselv.

Statsgeolog Ø. Gvein har foretatt skiferundersøkelser i Sollia i Atnadal, Vinstra, Lom, Vågå og Snåsa. Den siste befaringen er utført med henblikk på diamantboringer i diverse skiferforekomster for å undersøke spaltbarhet av skifer mot dypet. Han har videre undersøkt mørk labrador ved Vik i Tjølling og gneisbergarter i Nord-Trøndelag.

Vit. ass. I. Hultin har som oppdrag undersøkt kalk og kleberstensforekomster i Lom og klebersten i Målselv. Videre har han fortsatt kartleggingen av serpentin-kromittfeltene i Røros—Feragen-distriktet.

Statsgeolog K. Ø. Bryn har foretatt befaringer for Franzefoss Bruk A/S for å fastslå kalkstenens videre forløp.

Seksjonen har i år hatt følgende oppdragsgivere: Meråker Smelteverk, Kopperå, Hylla kalkverk, Røra, Orkla Grubeaktiebolag, Løkken Verk, Norsk Feltspatkompani, Evje, Elektrokemisk A/S, Fiskå Verk, Kristiansand, Tiltaksrådene i Lom, Vågå, Vinstra og Fosen, Vigsnes Kobberverk, Karmøy, Industridepartementet, Bjørkaasen Gruber A/S, Oslo, og Trondheim Domkirke.

Malm.

Direktør H. Bjørlykke har ledet seksjonen i det forløpne år. Seksjonen har i 1966 hatt følgende ansatt: Vit. ass. R. Hovland, vit. ass. G. Juve og tekn. ass. J. Gust. Vit. ass. I. Hultin har delvis arbeidet innen malmseksjonen.

Vit. ass. R. Hovland har i løpet av året befart en rekke norske ilmenittforekomster i et planlagt arbeid for å systematisere disse forekomster. Han har vært tilknyttet NGU's oppdrag for A/S National Industri i Repparfjord, og har videre foretatt befaring av molybdenførende ganger på Onaholmen i Sør-Frøya.

Vit. ass. G. Juve har fortsatt kartleggingen av Lakselvdalens kobberforekomster. Tidvis har han hatt samarbeid med rektor Sven Føyn, vit. ass. Hultin, tekn. ass. J. Gust samt andre sommermedarbeidere. Samarbeidet med A/S Sydvarangers feltgruppe vedrørende blyprospektering i fjellkjederanden fortsatte.

Tekn. ass. J. Gust har i høst arbeidet med fremstilling av nye anvisningskart over landets malmforekomster.

Malmseksjonen har i løpet av året fortsatt innsamlingsarbeidet for å få en så komplett malmsamling som mulig. Beskrivelse av samlingens preparater har pågått.

Laboratorier og preparantverksted.

Kjemisk laboratorium. Laboratoriets leder er laboratorieingeniør P. R. Graff. Ved laboratoriet har følgende vært ansatt: Laboratorieingeniør P. R. Graff, teknisk assistent I J. Røste, laborant I E. Aarsland samt praktikantene Carsten Qvenild og Anne Karin Buhaug.

I løpet av året er det utført 66 silikatanalyser og 265 andre analyser, vesentlig for institusjonens geologer.

Videre er det utført arbeider som rensing av stoff, konsentrering av mineraler og tillaging av spesielle blandinger for mineralidentifikasjon.

Endelig har en nå startet opp arbeidet med kartotekføring av alle fulle analyser av norske bergarter.

Radiometrisk laboratorium. Daglig leder er statsgeolog Chr. D. Thorkildsen. Det er som tidligere foretatt radiometriske målinger av prøver innsamlet av NGU's geologer og av innsendte prøver.

Mineralseparasjonslaboratoriet. Daglig leder er statsgeolog J. Hysingjord. Ved laboratoriet er arbeidet fortsatt i gang med mineralseparasjon ved hjelp av tunge væsker.

Innsamling av bergarter og mineralseparasjoner for aldersbestemmelser av norske bergarter ble påbegynt. Arbeidet skal foregå i samarbeid med Institutt for Atomenergi.

Røntgenlaboratorium. Daglig leder er statsgeolog J. Hysingjord. I løpet av året er det gjort 500 opptak på røntgenlaboratoriet. Antallet forespørsler og innsendte prøver er som foregående år.

Jordartslaboratoriet. Daglig leder er konstruktør E. Sørensen. Analysearbeidet ved laboratoriet har i vesentlig grad vært utført for å undersøke bergarters anvendbarhet i faste veidekker. Det er utført analyser for sprøhets- og flisighetskontroll av bergarter. Videre har laboratoriet utført kornfordelingsanalyser av sand og grus.

Analyser er utført for å kontrollere borkjernematerialers anvendbarhet som undersøkelsesmateriale for sprøhet og flisighet sett i relasjon til utskutte bergartsprøver.

Preparantverkstedet. Ved verkstedet har preparantene T. Jacobsen og E. Iversen arbeidet i 1966. Det ble i løpet av året fremstilt 1448 stk. tynnslip og 414 stk. polerslip. Antallet kombinerte tynn- og polerslip har øket til 156 stk.

Bergarkivet.

Ansvarshavende for bergarkivet er kontorfullmektig G. Anderssen.

Omleggingen av arkiveringsmåten for rapporter vedrørende «Industrielle mineraler og bergarter» ble avsluttet i 1966.

Arbeidet med nyregistrering og nummerering av kartsamlingen vedrørende malm- og mineralforekomster er påbegynt. Det er i den forbindelse tatt i bruk nye kartotekkort som gir opplysninger om rapporter og karter vedrørende forekomstene. Også tracinger vil bli registrert. Tidligere ble 3 kort benyttet for disse registreringene.

Bergarkivet har i 1966 hatt en tilvekst på 212 rapporter, 99 rapporter vedrørende «Industrielle mineraler og bygningssten», og 113 rapporter vedrørende «Malmer».

Bergarkivet omfatter pr. 31/12 4771 rapporter, hvorav 3813 vedrører malmer og 958 industrielle mineraler og bergarter.

Tilveksten på karter i 1966 er 10, tracinger 6.

Pensjonert statsgeolog A. O. Poulsen arbeider ved Oslokontoret med innsamling av data, delvis for Bergarkivet og delvis for sin publikasjon om glimmer i Norge.

Geofysisk avdeling.

Feltarbeider.

Geofysisk avdeling har i 1966 utført 46 oppdrag med i alt 1340 feltgruppedager. Av disse ble 19 oppdrag med 574 feltgruppedager utført for egne midler, mens 27 oppdrag med til sammen 766 feltgruppedager ble utført for oppdragsgiver utenfor institusjonen.

I det følgende er de viktigste oppdrag kort omtalt, ordnet etter fagområder.

Geofysiske bakkemålinger og borhullsmålinger.

Som en fortsettelse av forrige års undersøkelser ble det etter oppdrag fra A/S Sulitjelma Gruber utført elektromagnetiske målinger i området Anna Grube—Balvatn. Målingene ble utført under ledelse av geofysiker G. Sakshaug.

For A/S Røros Kobberverk ble det under ledelse av geofysiker G. Sakshaug utført elektromagnetiske målinger ved Kongens Grube og over et større område i Vangrøftdalen. For samme oppdragsgiver ble det under ledelse av geofysiker P. Singsaas utført elektromagnetiske målinger i Storwartzområdet og i diamantborhull ved Kongens Grube.

For Folldal Verk A/S ble det under ledelse av geofysiker P. Singsaas utført et vel 3 måneder langt oppdrag med elektromagnetiske målinger av større områder i Lesja, Folldal og Oppdal herred og et mindre område i Alvdal.

Målinger med IP (indusert polarisasjon) ble utført av geofysiker A. Sindre i Repparfjord som oppdrag for A/S National Industri.

Tyngdemålinger av flere mindre områder ble utført for Folldal Verk A/S av geofysiker A. Sindre og konstruktør A. Haugan.

Magnetiske målinger i borhull ble utført for Fosdalens Bergverks-A/S av fysiker A. Breen.

Flymålinger.

De geofysiske målinger utført fra fly ble fortsatt i 1966. Til målingene ble benyttet et 4 motors fly av type Heron som ble leiet fra Nor Flyselskap A/S, Hønefoss. Målingene ble ledet av geofysiker H. Håbrekke med konstruktør K. Brandhaug og tekniker O. Blokkum som assistenter.

Flyet ble benyttet i alt 403 flytimer, og det ble målt en samlet profillengde på ca. 59 000 km.

Som en fortsettelse på den systematiske dekning med profilavstand 500 meter ble i alt ca. 10 000 km^2, fordelt på Nord-Trøndelag, Femundsområdet og Kongsvingerområdet målt med magnetisk, elektromagnetisk og radiometrisk måleutstyr.

Magnetiske målinger med profilavstand 1 km ble utført i et ca. 4000 km^2 stort område over øyene Frøya, Hitra og Smøla.

Som oppdrag for Folldal Verk A/S ble det utført kombinerte magnetiske og elektromagnetiske målinger over et ca. 6000 km^2 stort område mellom Dovre og Alvdal.

Til magnetiske målinger over Kontinentalsokkelen ble det benyttet i alt 145 flytimer. Av dette ble 50 flytimer anvendt til å utvide det i 1965 målte område videre vestover ut over Egga, 33 timer gikk med til å dekke sokkelområdet videre sydover til 62° (Stadt), og 62 timer ble benyttet til målinger i området Brønnøysund—Lofoten (Trænabanken). Som foregående år ble det benyttet Lorannavigasjon og fløyet med en profilavstand på ca. 5 km.

I løpet av året ble 51 aeromagnetiske kartblad offentliggjort. Kartene er i målestokk 1 : 50 000 og har samme bladinndeling som NGO's kartserie i denne målestokk. Sammen med de første 29 kartblad som ble offentliggjort i 1965 er dermed i alt 80 blad tilsvarende et areal på vel 50 000 km^2 utgitt.

Seismiske målinger.

Det ble utført 7 oppdrag med sammenlagt 134 feltdager.

De fleste oppdragene var kortvarige og gikk ut på å bestemme overdekkemektigheten i tilknytning til anleggstekniske problemer.

I Lieråsen ble det for NSB utført et større oppdrag over en tunneltracé hvor det er satt i gang drift fra begge sider. Dette arbeid sorterer under Drammensbanens Dobbeltsporanlegg. Overdekkemektigheten var her av sekundær betydning. Det primære var å registrere lydens forplantningshastighet i fjell for derved å kunne kartlegge svakhetssonene.

Det andre oppdraget av større omfang gjaldt målinger på Romerike i forbindelse med den Internasjonale Hydrologiske Dekade. Det gjaldt her å fastlegge grensene mellom forskjellige typer grusavsetninger, grunnvannsnivået samt beliggenhet av fast fjell.

Vår Interval Timer GT-2 har vært i USA for ombygging, og den fungerte etter modifikasjonen meget godt. De fleste målingene ble utført med denne apparaturen. Geofysiker A. Sindre utførte et mindre oppdrag på Hjerkinn, mens de øvrige oppdrag ble ledet av geofysiker G. Hillestad.

Ingeniørgeologiske arbeider.

Undersøkelser for Statskraftverkene. I forbindelse med Eidfjordanleggene arbeidet geolog H. Barkey i ca. 4 uker på Hardangervidda mens ca. 2 uker gikk med til undersøkelser for Jotunheimanleggene.

Geolog S. Svinndal arbeidet ca. 3 uker ved Sildvik og Skjomen kraftanlegg i Ofoten og ca. 2 uker ved Trollheim kraftanlegg.

Undersøkelser i kraftverkstunneler. Som et ledd i en plan om å få en systematisk geologisk undersøkelse i alle tunneler som drives i forbindelse med kraftverksutbyggingen ble det engasjert en tysk geolog, Bernd Weber. Han arbeidet i 3 måneder og foretok kartlegging ved Uste—Nes-anleggene, Kvænangen kraftanlegg og Ranaanleggene. Geolog H. Barkey deltok delvis ved Uste—Nes og i Rana. I alt ble det kartlagt ca. 23 km tunnel.

Malmgeologiske undersøkelser.

Indre Finnmark. Geologene S. Svinndal og T. H. Tan med tekniker J. Staw som assistent fortsatte undersøkelsene i Kautokeino herred. Det ble i år undersøkt i alt 18 lokaliteter hvor det var fremkommet elektromagnetiske anomalier ved flymålingene i 1959. Undersøkelsene bestod i slingrammålinger, geologisk kartlegging og makro- og mikroblokkleting.

Trøndelag. Geologisk oppfølging av elektromagnetiske anomalier fra flymålinger i 1964 ble utført av geolog H. Barkey i 5 forskjellige områder i Sør-Trøndelag.

Telemark. Geologene S. Svinndal og H. Barkey fortsatte de geologiske undersøkelser i Nedre Telemark for å følge opp magnetiske anomalier som var fremkommet ved flymålinger i 1962.

Diamantboringer.

I løpet av året ble det diamantboret i alt 6677 meter fordelt på 11 forskjellige oppdrag.

Det største borprogram ble også i år utført i Bidjovagge hvor det ble boret 2345 meter.

Som ledd i malmundersøkelser ble det i Repparfjord boret 2132 meter etter oppdrag fra A/S National Industri.

For A/S Røros Kobberverk ble det boret 333 meter i Storwartzområdet.

Ved Skåleseter i Sørli ble det boret 255 meter som en del av fortsatt undersøkelse av et område med geokjemiske og geofysiske anomalier.

I Trysil ble det ved Tenåsen boret 255 meter som bidrag til kartlegging av de geologiske forhold i området.

Oppdrag i forbindelse med grunnundersøkelser ble utført for Statskraftverkene ved Veivatn (175 meter) og for Norges Statsbaner i Lieråsen (350 meter).

I forbindelse med undersøkelse av mineralske råstoffer ble det boret for Hylla Kalkverk på Inderøy (247 meter) og for Elektrokemisk A/S, Fiskå Verk (320 meter).

Diamantboringene ble utført under ledelse av borformennene O. Gausdal og S. Vassbotn.

Verksted- og laboratoriearbeid.

Verkstedet har som vanlig besørget vedlikehold av instrumenter og utstyr, herunder 10 biler, 3 Muskeg beltebiler og diamantborutstyr.

I forbindelse med overgang til nytt og større fly ble måleutstyret for elektromagnetiske flymålinger fornyet og forbedret slik at en oppnår større følsomhet.

Det er fremstillet 6 stk. måleapparaturer for slingrammålinger, 3 stk. susceptibilitetsmålere og 1 stk. likespenningspotensiometer. En del av dette er beregnet for salg.

For salg er videre fremstilt en serie på 100 stk. magnetometre. Av disse ble i alt 39 stk. solgt, herav 33 stk. utenlands gjennom det svenske firma Craelius i henhold til inngått avtale.

Ved verkstedet er det også utført en del arbeide for så vel Geologisk som Kjemisk avdeling.

Kjemisk avdeling.

Spektrografisk og kjemisk analytisk arbeid.

Ledere: Laboratorieingeniør G. Faye (spektrografi) og
laboratorieingeniør B. Andreassen (kjemisk analyse).

Ved kvantometret har en fortsatt arbeidet med tilpassing av rutinemetoder til dette instrument. Hovedbestanddelene i bergarter, unntatt alkalier, kan således nå bestemmes med forholdsvis stor nøyaktighet. Alkalier lar seg mer praktisk bestemme flammefotometrisk. En tilleggsanskaffelse til kvantometret, digitalvoltmeter med en tallskriver, er en betydelig forbedring som har gitt sikrere og nøyaktigere avlesing. En del analyseoppgaver løses best ved anvendelse av både kvantometer og røntgenfluorescens. Røntgenfluorescens har spesielt vært brukt for Cu-Zn-bestemmelse i store serier av opprednings- og borkjerneprøver.

Ved de kjemiske analyselaboratoriene har det som tidligere vesentlig vært utført silikatanalyser og malmanalyser ved siden av utviklingsarbeid. Bl.a. har en hurtig våtveismetode for bestemmelse av svovel gitt øket analysekapasitet av betydning for oppdragsvirksomheten.

Analysevirksomheten fordeler seg slik med antall bestemmelser:

Utført av		Utført for				
Kjemisk lab.	Spektr. lab.	Kjemisk avd.	Geof. avd.	Geol. avd.	Bidjovagge-unders.	Oppdrags-givere
3 333	6 120	1 724	87	299	310	7 033

Oppdragsgivere er for en vesentlig del bergverksselskaper.

Analysene for kjemisk avdeling har for det meste forbindelse med geokjemisk prospektering.

I tillegg til tallene i tabellen kommer ca. 40 000 bestemmelser utført ved geokjemisk laboratorium i forbindelse med geokjemisk prospektering.

Laboratorium for keramiske og ildfaste materialer.

Leder: Laboratorieingeniør Johs. Grenness.

I forbindelse med at Sjøfartsdirektoratet utarbeider nye forskrifter om sikkerhetsregler for transport av malmkonsentrater o. l., er laboratoriet blitt engasjert i et omfattende undersøkelsesarbeid vedrørende fuktighetsgrenser for konsentrater som transporteres til sjøs. I forbindelse hermed er det også laget et apparatur. Bl. a. har Geofysisk avdelings verksted bygget et A.S.T.M. «Flytebord» for disse undersøkelsene.

Forøvrig har laboratoriet som vanlig utført sikte- og slemmeanalyser, spesifikk vekt-bestemmelser, smeltepunktbestemmelser o. l.

Geokjemisk prospektering.

Leder: Geokjemiker B. Bølviken; under Bølvikens permisjon (fra 1/4 1965 til ca. 1/4 1967): Direktør A. Kvalheim.

Geokjemisk prospektering med bekkesedimentmetoden er utført som oppdrag for 3 bergverksselskaper i følgende områder:

1. Et 335 km^2 stort område i Finnmark.
2. Et 720 km^2 stort område ved Hjerkinn.
3. Et ca. 100 km^2 stort område i Meråker. Her har bergverksselskapet selv gjort feltarbeidet, mens analysering og bearbeidelse, med kartrapportering er gitt Kjemisk avdeling som oppdrag.

I forbindelse med Bidjovagge-undersøkelsene ble det gjort orienterende geokjemiske undersøkelser over en geofysosk anomali syd for Časkias, ved prøvetaking og analysering av jord.

Rundt kobberforekomster ved Suovrarappat i Finnmark ble det gjort systematisk prøvetaking av morene som et ledd i studiet av anomalimønstret i morenejord i et område med mineraliserte blokkvifter eller blokktog.

I Oppland fylke har geokjemiker Ø. Hvatum fortsatt detaljundersøkelsene i et tidligere prøvetatt område ved Gjøvik. Arbeidet gjelder jordprøver og går ut på både å studere informasjonsevnen ved forskjellig prøvetetthet, og å studere metallinnholdet i jordprofil over eller nær blymineralisering.

Metallurgisk laboratorium.

Ledere: Laboratorieingeniørene J. Aarvik og R. Nilsen.

Ingeniør J. Aarvik har bearbeidet videre en del kisforedlingssaker, bl. a. reduksjonen av svoveldioksyd med olje og raffinering av kisavbrann.

Ingeniør R. Nilsen har arbeidet videre med oppgaver i forbindelse med reduksjonssmeltet olivin.

Gjestende medarbeidere.

Stud. real. Nils N. Kjøsnes fra Norges Lærerhøgskole har fullført laboratoriearbeidet for sin hovedfagsoppgave i spektrografi.

Amerikansk student Martha Redden og jugoslavisk student Alija Sirbegovic har arbeidet i sommermånedene ved henholdsvis spektrografisk og kjemisk laboratorium som et ledd i praktikantutvekslingen av studenter.

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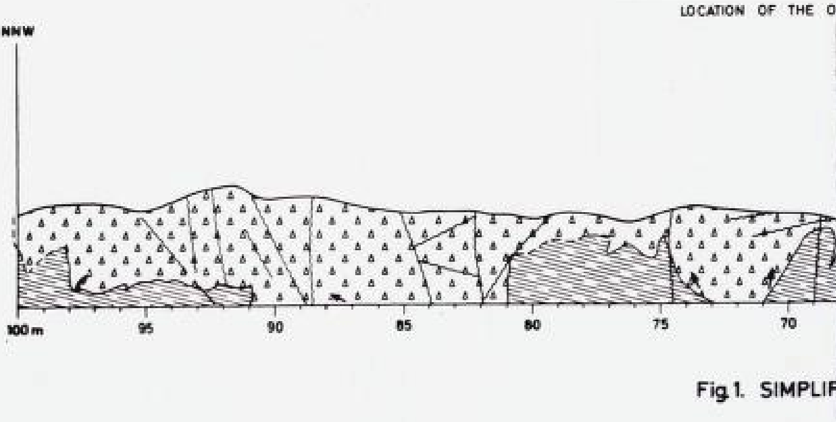
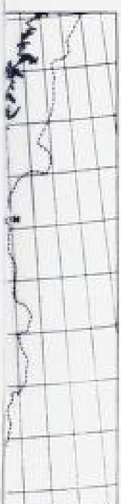


Fig.1. SIMPLIF



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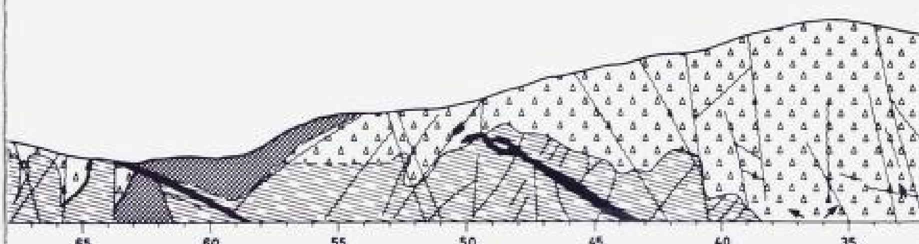
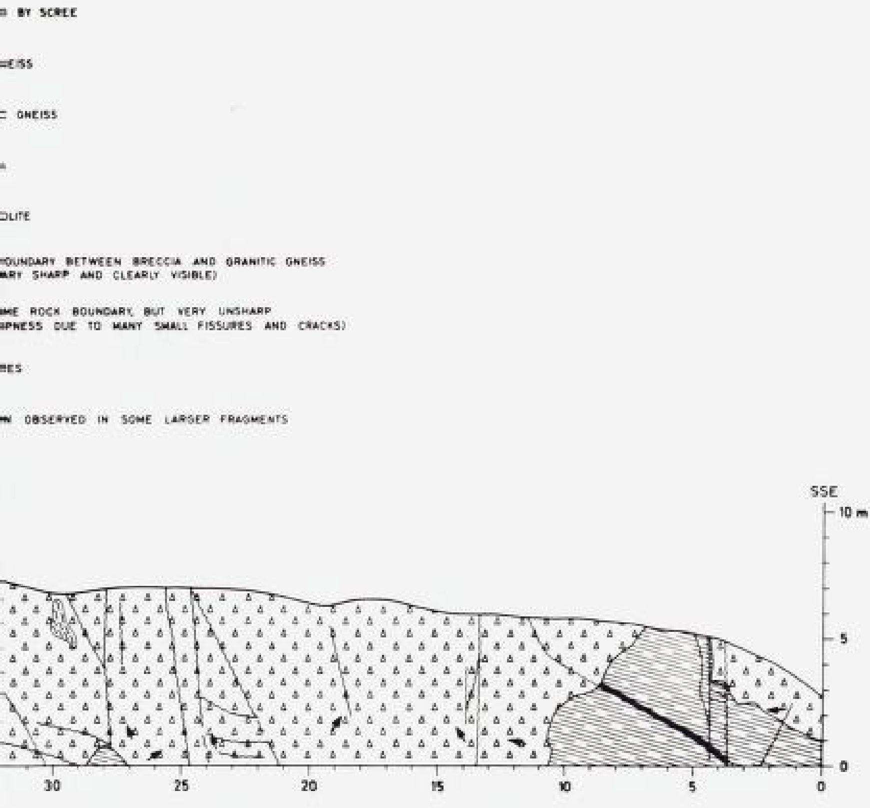


IED SECTION ALONG THE EXPLOSION-BRECCIA IN HJØLMODALEN, HARDANGERVIDD



D BY SCREE
NEISS
C GNEISS
A
OLITE
OUNDARY BETWEEN BRECCIA AND GRANITIC GNEISS
ARY SHARP AND CLEARLY VISIBLE)
ME ROCK BOUNDARY, BUT VERY UNSHARP
PNESS DUE TO MANY SMALL FISSURES AND CRACKS)
RES
N OBSERVED IN SOME LARGER FRAGMENTS
SSE
10 m
5
0
30
25
20
15
10
5
0

DA (NORWAY)

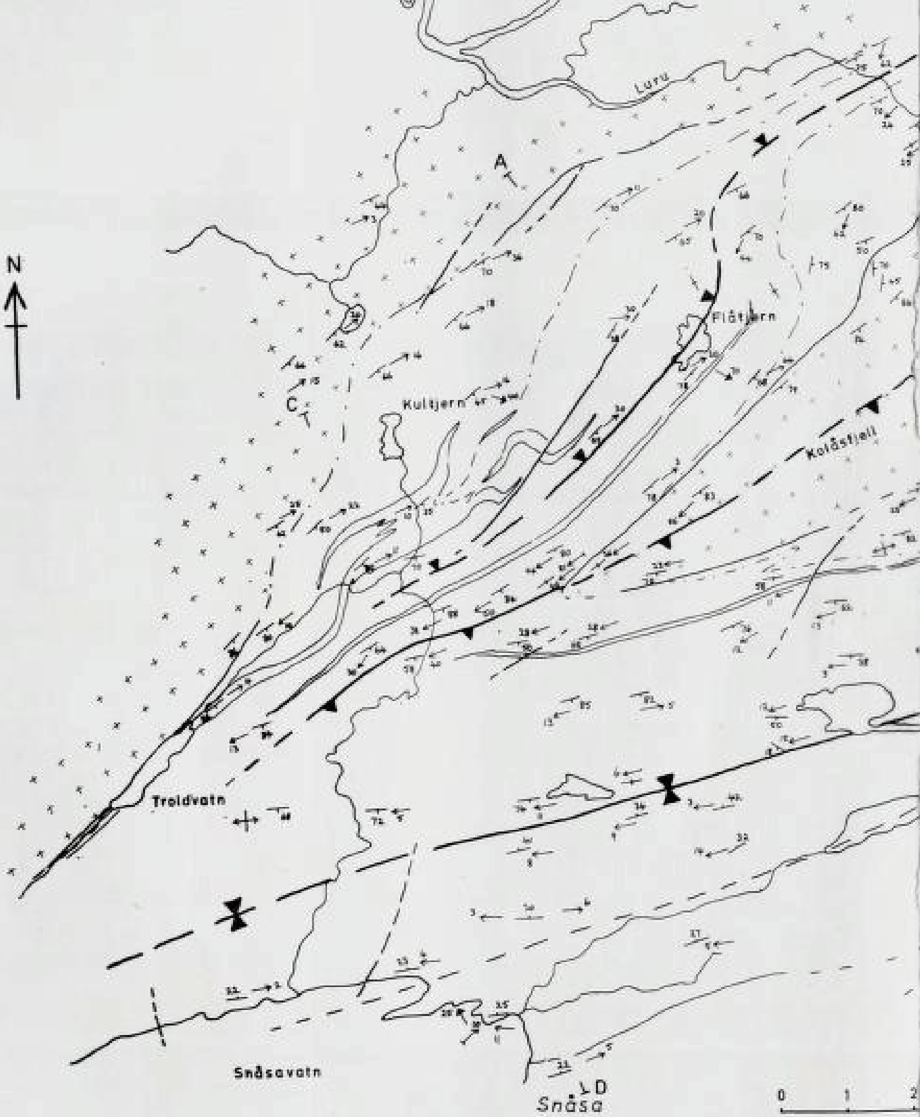
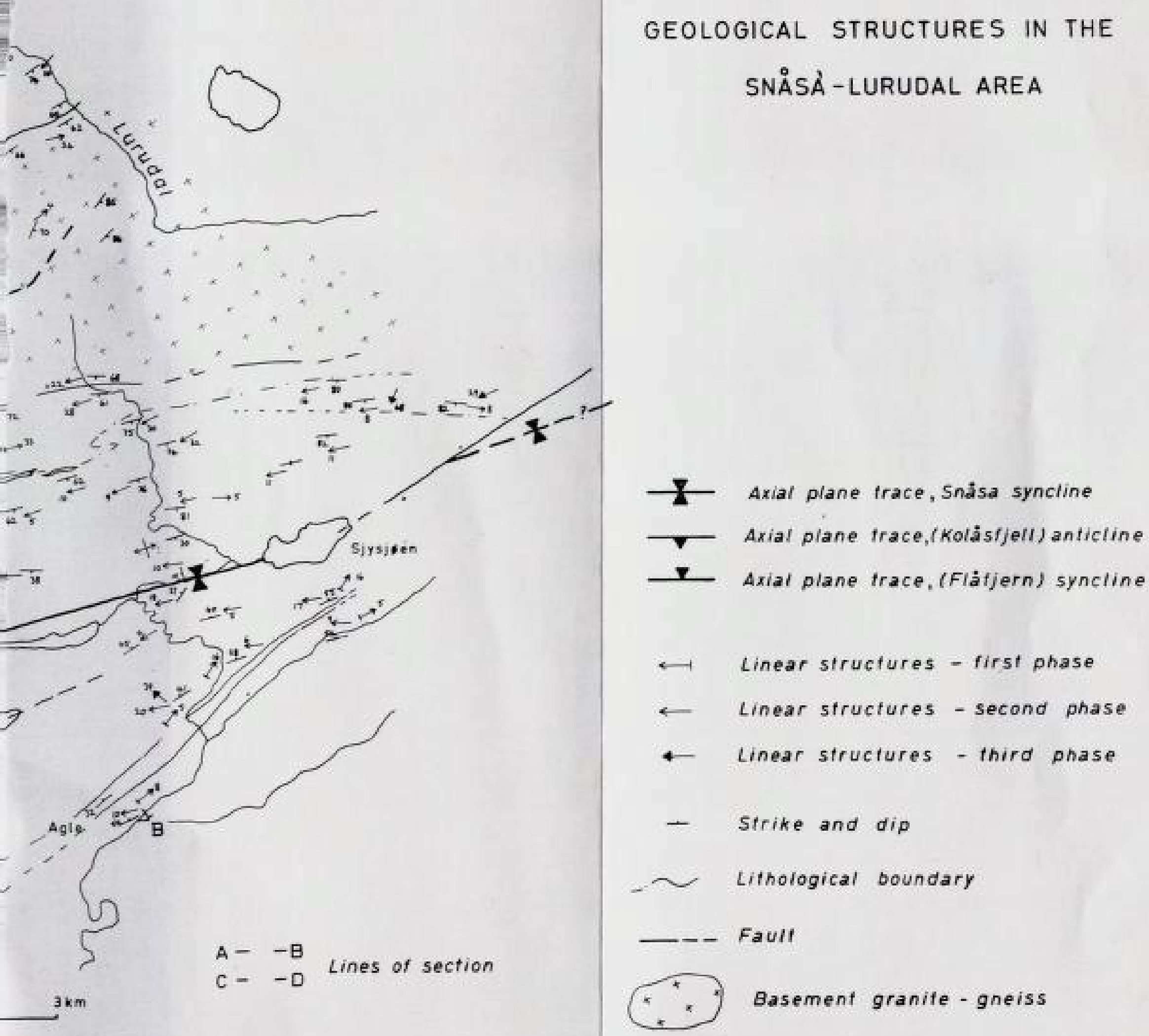
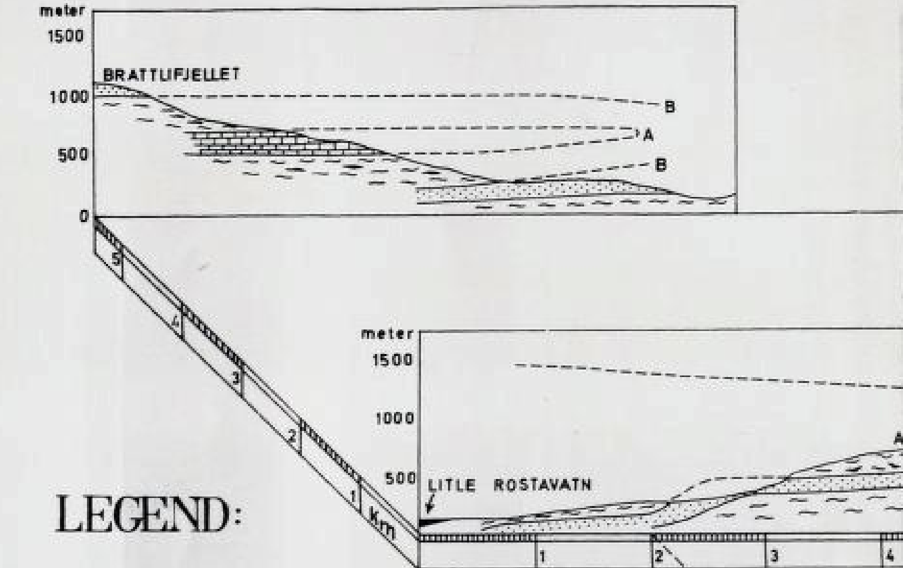


Fig. 1. Map of axi



al plane trends and representative linear elements
in the Snåsa - Lurudal area.



LEGEND:

UPPER SEQUENCE.
- BIOTITE RICH GARNET MICA SCHISTS, IN THE UPPER PART WITH BANDS OF CALCSILICATE ROCKS.

MIDDLE SEQUENCE.
- RED-BROWN WEATHERING MARBLES.
- LIGHT-GRAY WEATHERING MARBLES.
- GARNET-BIOTITE MICA SCHISTS.
- QUARTZITES.
- MICA SCHISTS WITH AMPHIBOLITE-BANDS. } ON HABAFJELLET AND BRATTLIFJELLET.
- RED WEATHERING MARBLES. } ON HABAFJELLET AND BRATTLIFJELLET.

LOWER SEQUENCE.
- CALCAREOUS (GARNET) MICA SCHISTS.
- MIXED FORMATION OF GRAYISH MARBLES, QUARTZITES, MICA SCHISTS AND GARNET-BIOTITE MICA SCHISTS.
- MAINLY AUGEN SCHISTS AND AUGEN GNEISSES.
- LOW-GRADE SCHISTS (MAINLY MYLONITES AND PHYLLONITES).

- HYOLITHUS ZONE SEDIMENTS.
- MAINLY HOMOGENEOUS GRANODIORITES.
- x—x MAIN THRUSTS.

Fig. 3. Profiles through the northern part of the area.

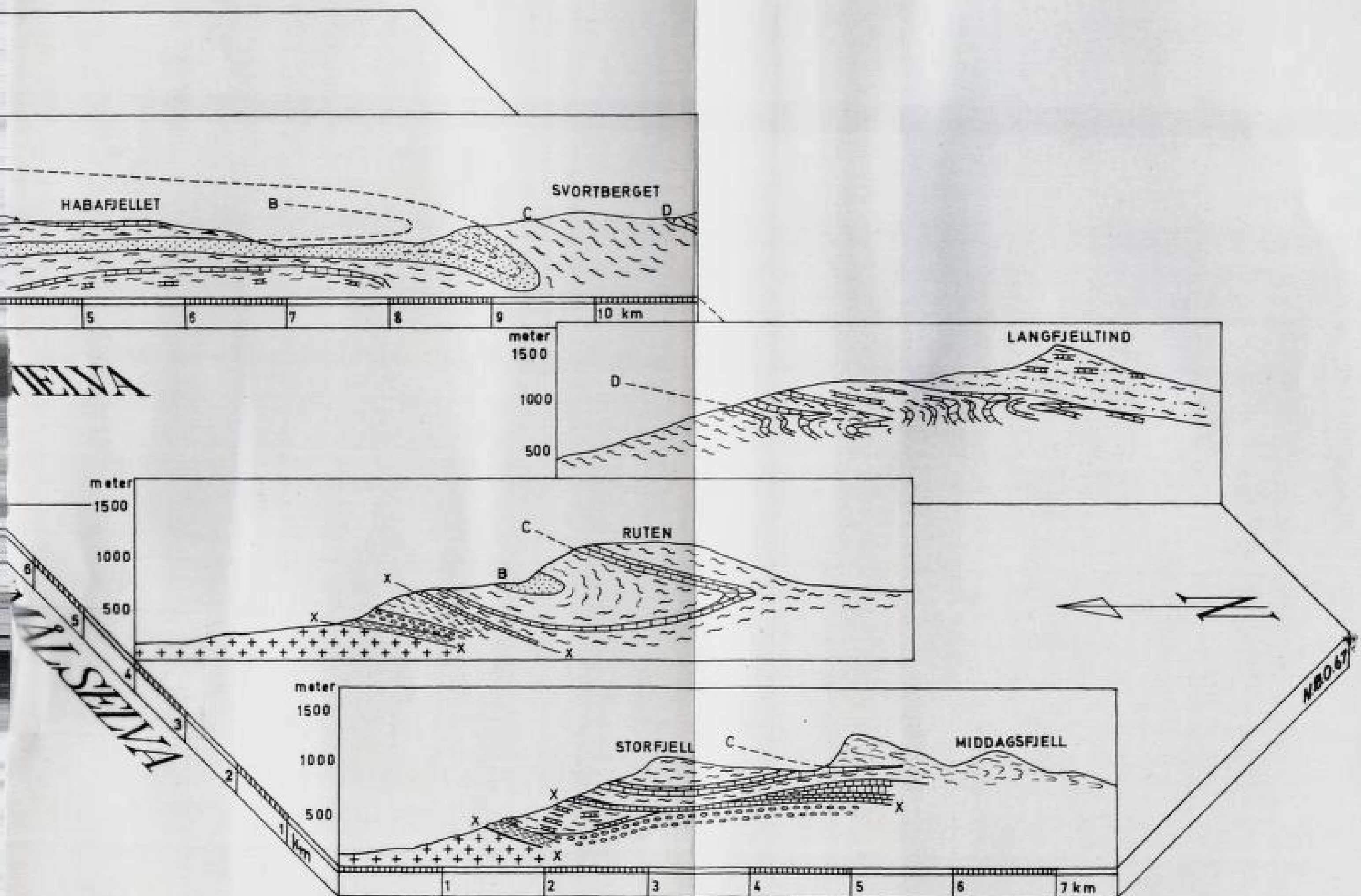
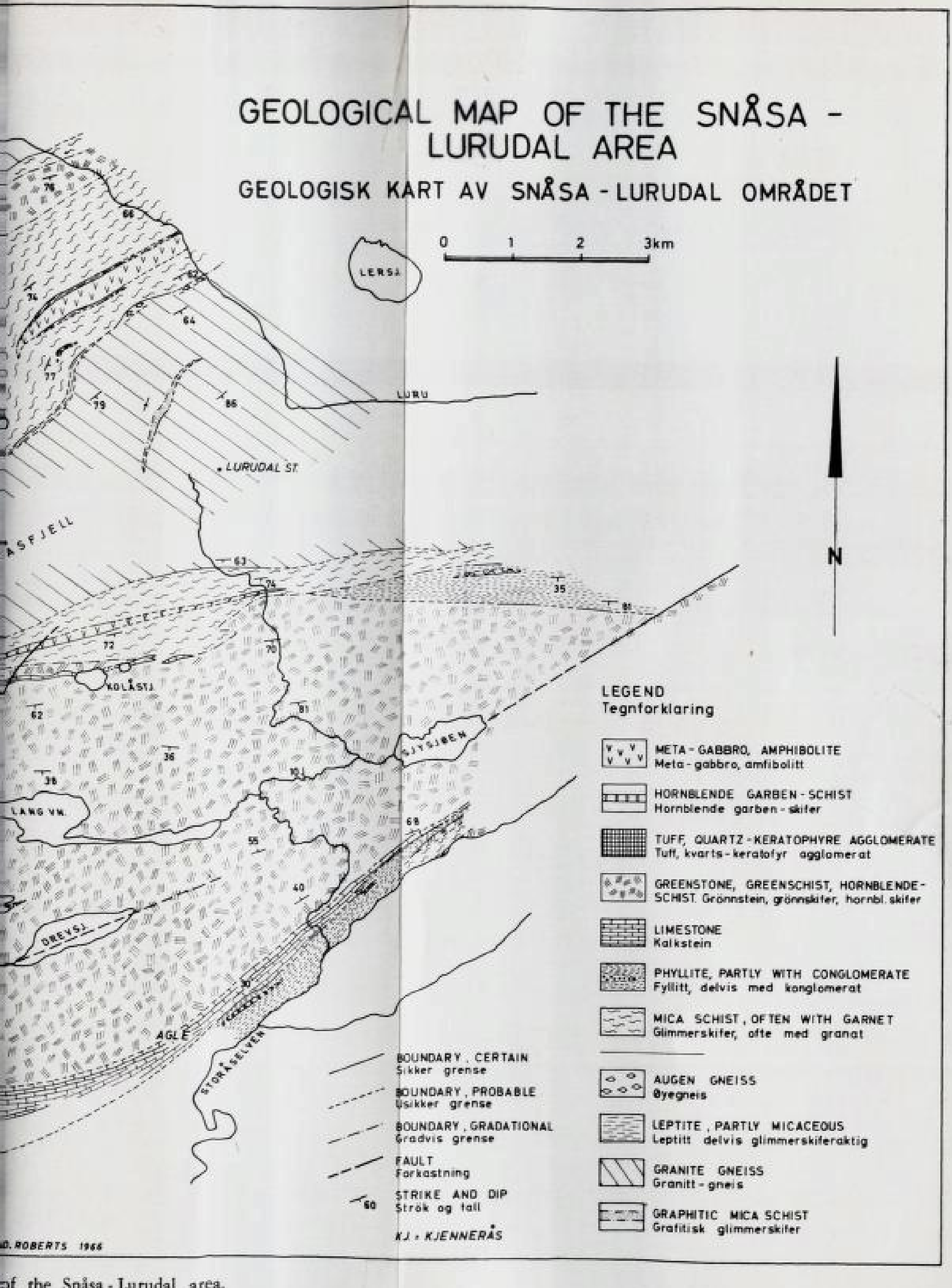
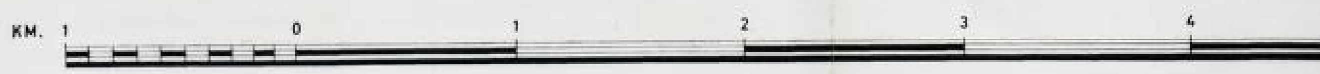


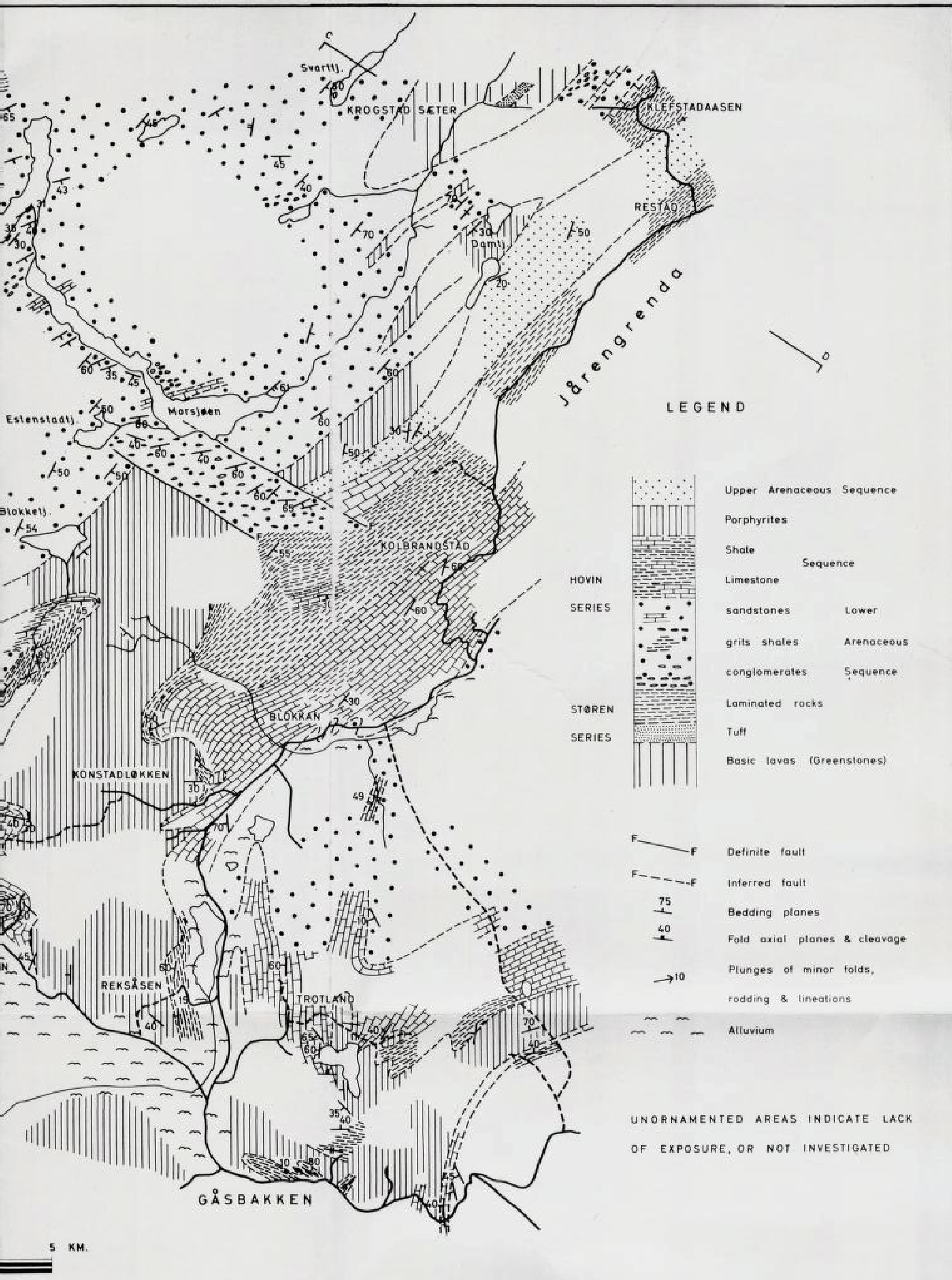
Plate I. Geological map



…of the Snåsa - Lurudal area.

GEOLOGICAL MAP OF AN AREA NORTH OF GÅSBAKKEN

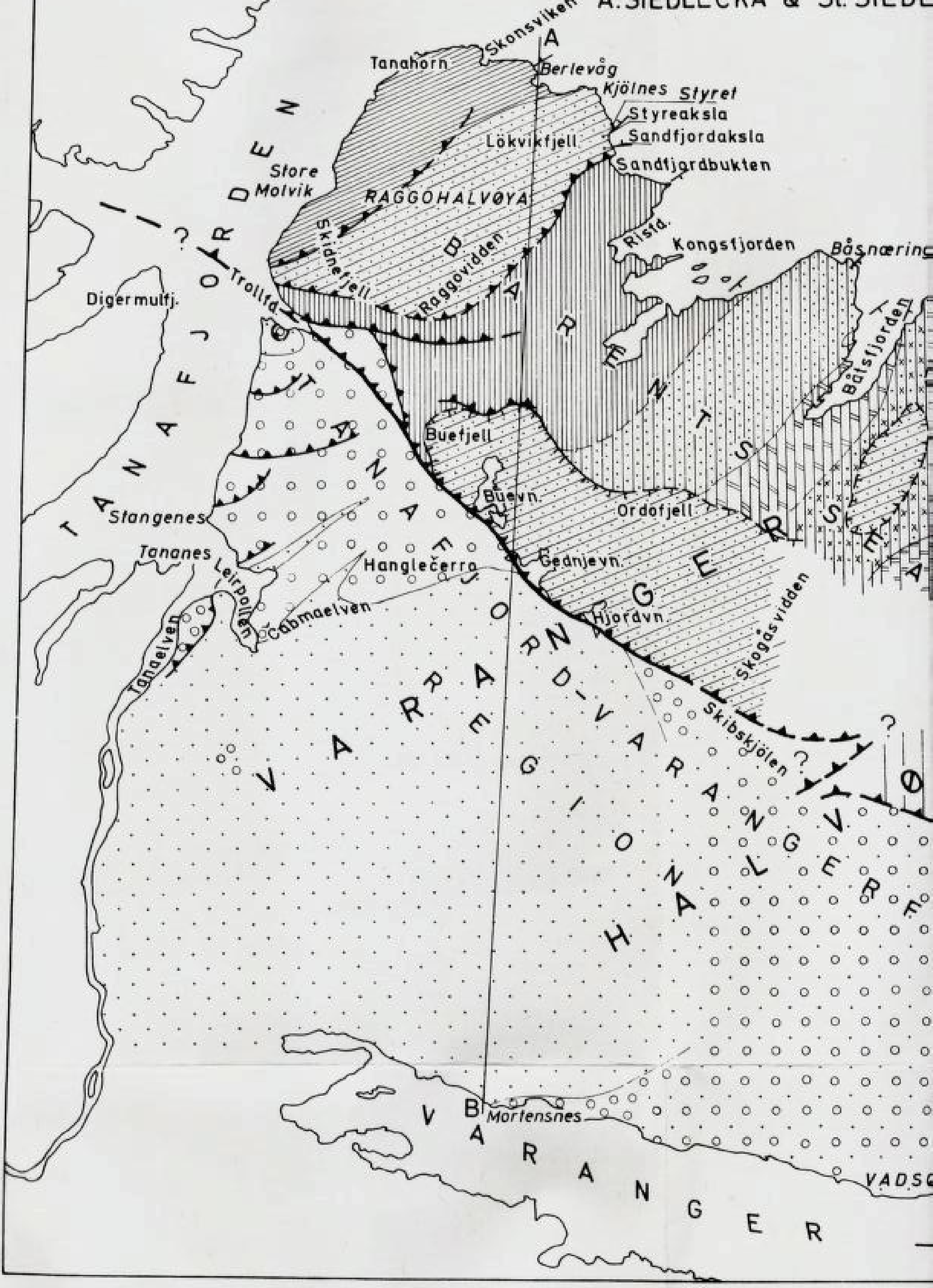


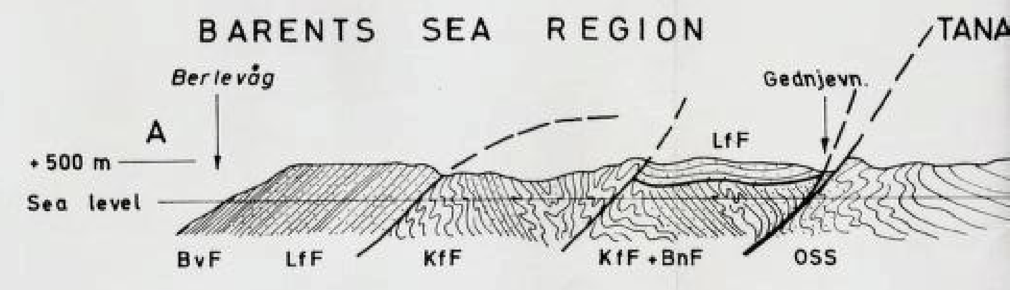


P CARTER JANUARY 1965

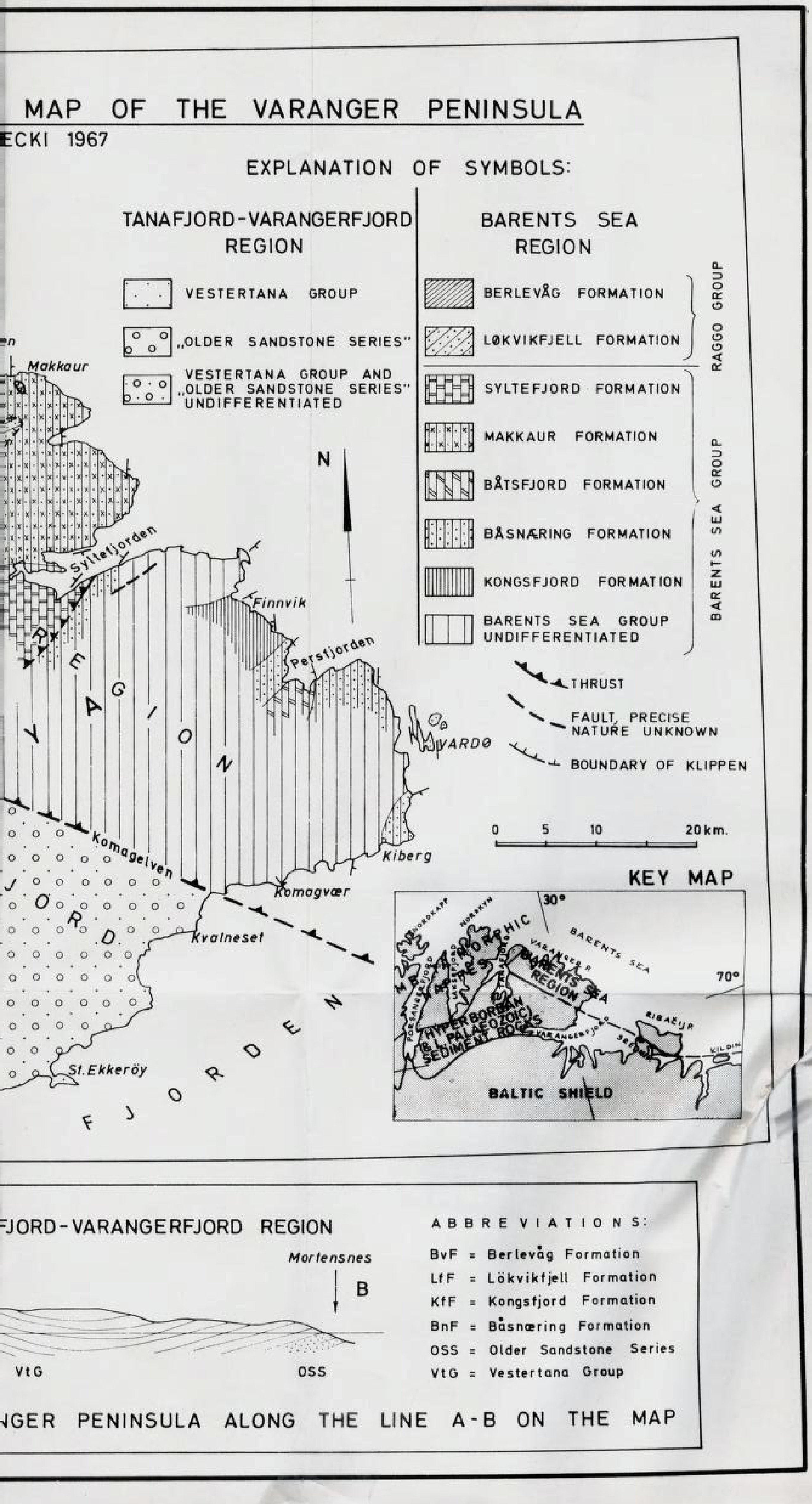
GEOLOGICAL SKETCH

A.SIEDLECKA & St. SIEDL

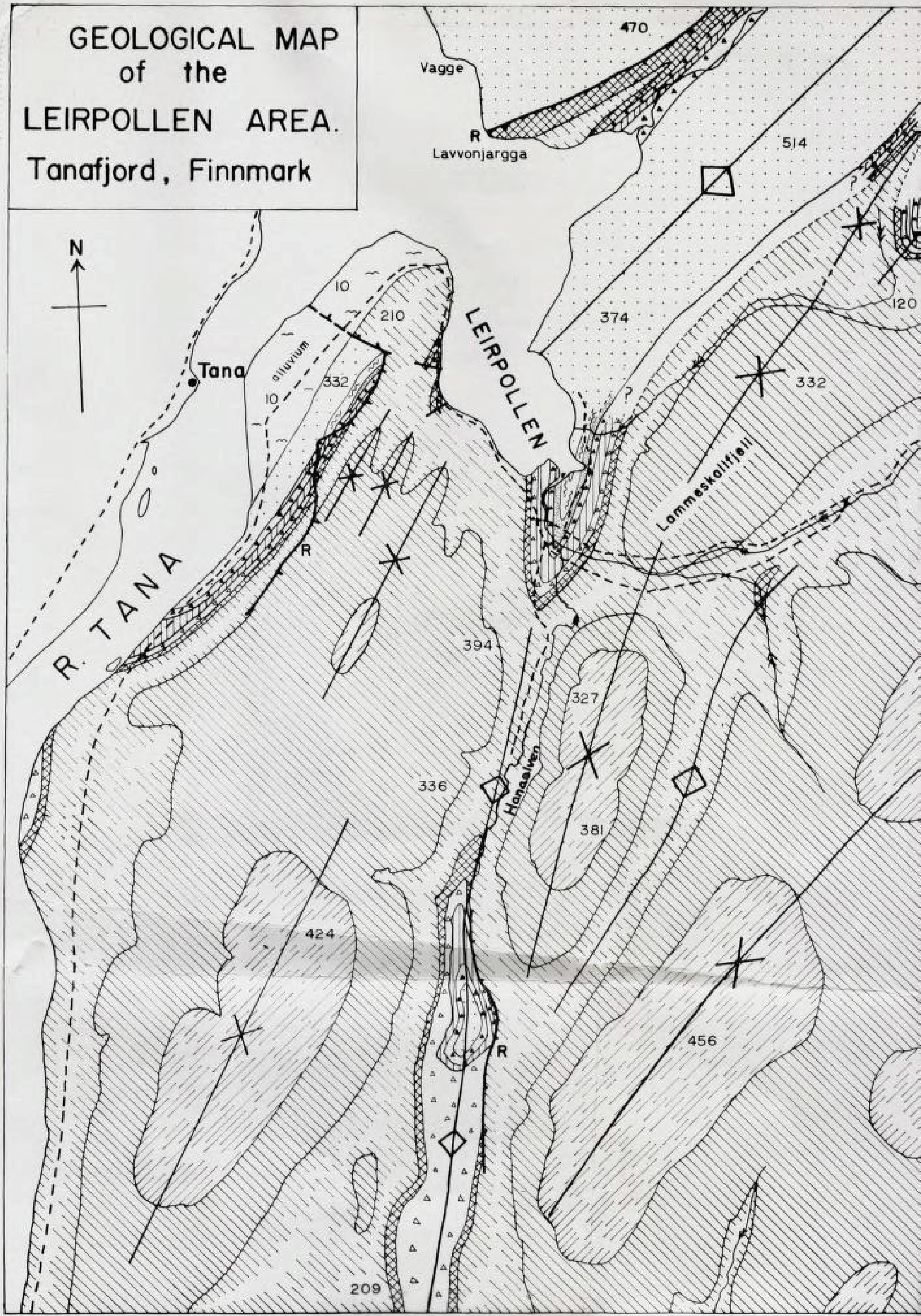




DIAGRAMMATIC CROSS-SECTION THROUGH THE VARA



MAP OF THE VARANGER PENINSULA
ECKI 1967
EXPLANATION OF SYMBOLS:
TANAFJORD-VARANGERFJORD REGION
VESTERTANA GROUP
„OLDER SANDSTONE SERIES"
VESTERTANA GROUP AND „OLDER SANDSTONE SERIES" UNDIFFERENTIATED
BARENTS SEA REGION
BERLEVÅG FORMATION
LØKVIKFJELL FORMATION
RAGGO GROUP
SYLTEFJORD FORMATION
MAKKAUR FORMATION
BÅTSFJORD FORMATION
BÅSNÆRING FORMATION
KONGSFJORD FORMATION
BARENTS SEA GROUP UNDIFFERENTIATED
BARENTS SEA GROUP
THRUST
FAULT, PRECISE NATURE UNKNOWN
BOUNDARY OF KLIPPEN
0 5 10 20km.
N
Makkaur
Syltefjorden
Finnvik
Persfjorden
VARDØ
Kiberg
Komagvær
Komagelven
Kvalneset
St.Ekkeröy
REGION
FJORD
FJORDEN
KEY MAP
30°
70°
BARENTS SEA
BARENTS SEA REGION
HYPERBOREAN (& L. PALAEOZOIC) SEDIMENT. ROCKS
BALTIC SHIELD
FJORD-VARANGERFJORD REGION
Mortensnes
B
VtG
OSS
ABBREVIATIONS:
BvF = Berlevåg Formation
LfF = Lökvikfjell Formation
KfF = Kongsfjord Formation
BnF = Båsnæring Formation
OSS = Older Sandstone Series
VtG = Vestertana Group
NGER PENINSULA ALONG THE LINE A-B ON THE MAP



GEOLOGICAL MAP
of the
LEIRPOLLEN AREA.
Tanafjord, Finnmark
N
R. TANA
Tana
alluvium
LEIRPOLLEN
Vagge
Lavvonjargga
Lammeskallfjell
Hanaelven
R
470
514
374
332
210
10
120
394
327
336
381
424
456
209

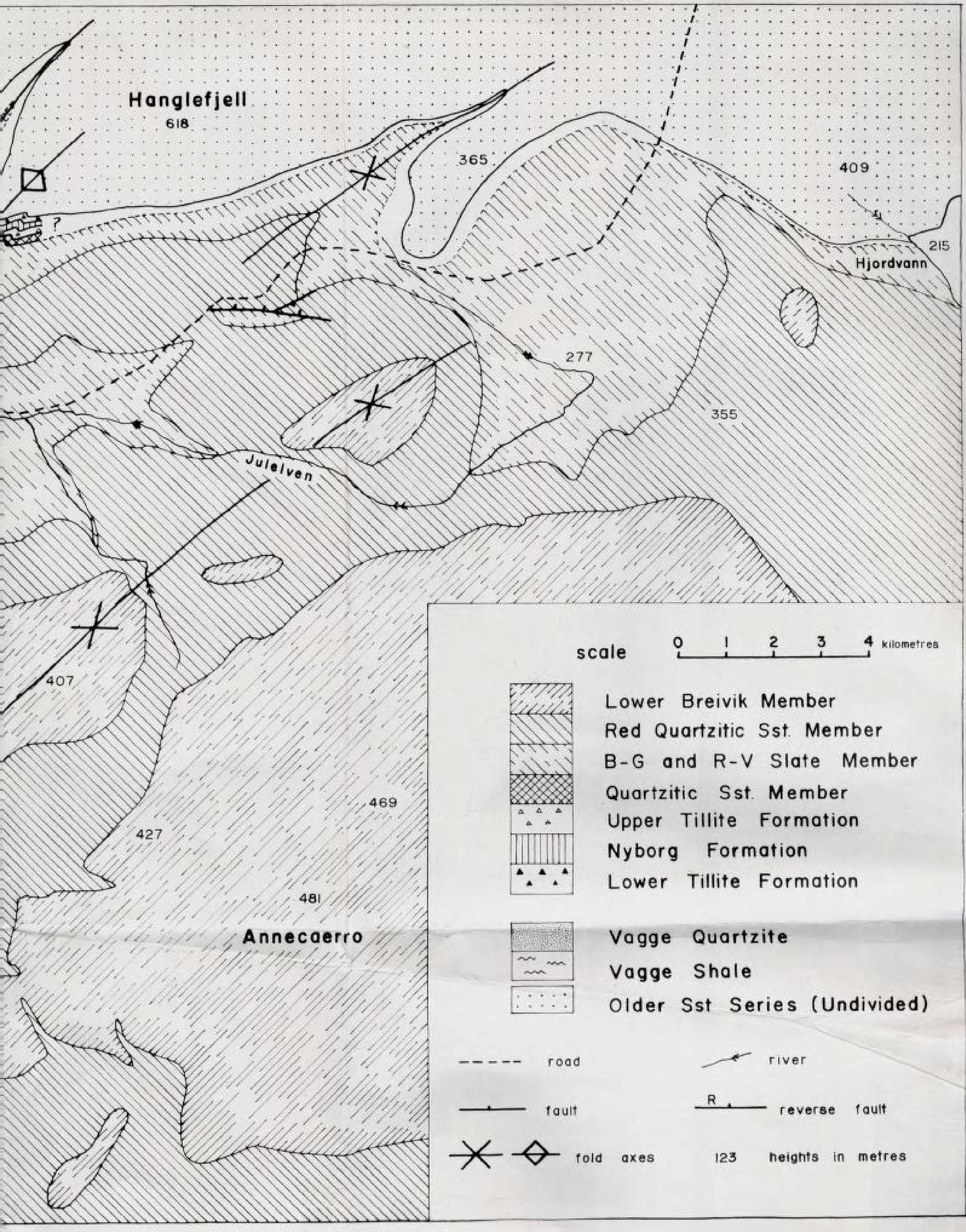

Hanglefjell
618
365
409
215
Hjordvann
277
355
Julelven
407
427
469
481
Annecaerro
scale
0 1 2 3 4 kilometres
Lower Breivik Member
Red Quartzitic Sst. Member
B-G and R-V Slate Member
Quartzitic Sst. Member
Upper Tillite Formation
Nyborg Formation
Lower Tillite Formation
Vagge Quartzite
Vagge Shale
Older Sst Series (Undivided)
road
river
fault
R
reverse fault
fold axes
123 heights in metres

Plate I. **Amphibolite**, specimen 2448 c. Plane polarised light. Green hornblende in a fine grained groundmass of plagioclase and quartz. Diam. of photo = 2 mm.

Plate II a). Granodiorite, specimen 2262. Crossed nicols. Microcline, with plagioclase and quartz. Diam. of photo = 2 mm.

Plate II b). Mylonitic gneiss, specimen 547. Crossed nicols. Crushed, fine grained groundmass of quartz and feldspar. A few large, highly cataclastic plagioclase grains. Diam. of photo = 2 mm.

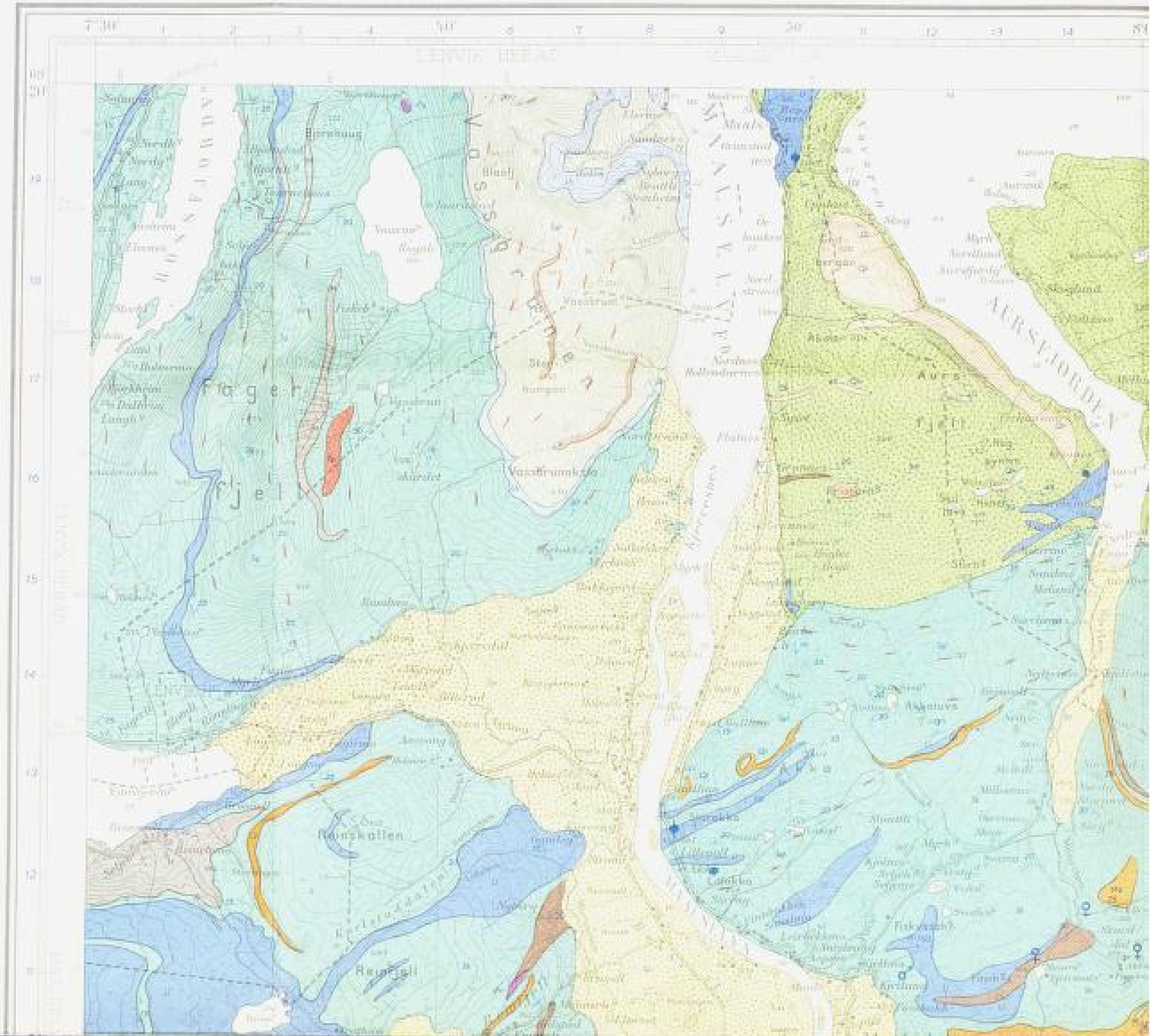
TROMSØ

Geologis

MÅLS

Tromsø

MÅLSELV



Rekkefølgen av de lagdelte bergarter

Stratigraphy

Øvre avdeling
Upper division

Øvre mosone
Upper mylo zone

Øvre kalksone
Upper limestone division

Øvre biotittskifer
Upper biotite schist

Øvre kvartsittsone
Upper quartzite zone

Kleritskiferavdelingen
Chlorite-schist division

Muskovitt-granatskifer
Muscovite-garnet-schist

Undre kalksone
Lower limestone zone

Konglomerater
Conglomerates

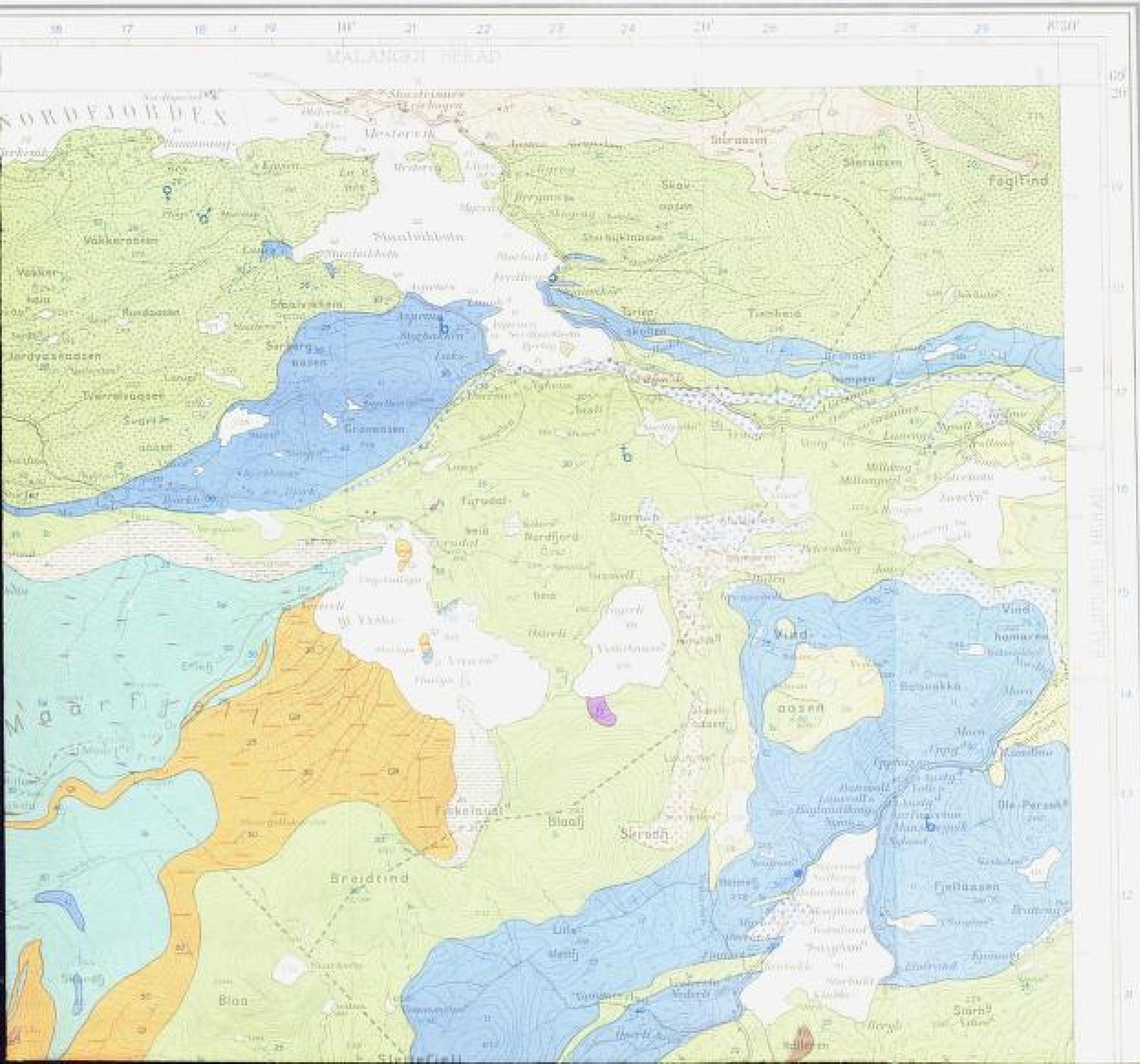
Kloritskifer
Chlorite schist

MUSEUM

k kart

ELV

1959



Kvartære løsavleiringer
Quaternary deposits

Dr — Marine avsetninger — *Marine deposits*

Db — Bresprut — *Morainic drift*

Fjellkjedens bergarter
Rocks of the Caledonian range

Sb — Biotittskifer i Øvre avdeling — *Biotite-schist in the Upper division*

S — Biotittskifer, nivå ikke angitt — *Biotite-schist, level not stated*

Sh — Biotittskifer i Undre avdeling — *Biotite-schist in the Lower division*

Sbl — Biotittskifer, kalk- og hornblendeholdig — *Biotite-schist, calcite- and hornblende-bearing*

Sc — Klorittskifer i Klorittskiferavdelingen — *Chlorite-schist in the Chlorite-schist division*

Sm — Muskovittgranat- (biotitt-) skifer i Klorittskiferavdelingen — *Muscovite-garnet- (biotite-) schist in the Chlorite-schist division*

Sg — Plagioklasrik skifer og glimmerskifergneis, nivå ikke angitt — *Plagioclase-schist and mica-schist-gneiss, level not stated*

Sgn — Plagioklasrik skifer og glimmerskifergneis, Øvre gneissone — *Plagioclase-schist and mica-schist-gneiss, Upper gneiss zone*

Sgl — Plagioklasrik skifer og glimmerskifergneis, Undre gneissone — *Plagioclase-schist and mica-schist-gneiss, Lower gneiss zone*

Lk — Kalkstein (marmor), Øvre kalksone — *Limestone (marble), Upper limestone zone*

L — Kalkstein (marmor), Undre kalksone — *Limestone (marble), Lower limestone zone*

L — Kalkstein (marmor), nivå ikke angitt — *Limestone (marble), level not stated*

Hornblendeskifer

Undre avdeling
Lower division

Undre gneissone
Lower gneiss zone

Undre kvartsitter (mikroklinkvartsitt)
Lower quartzites (microcline-quartzite)

Undre biotittskifer, grønnskifer
Lower biotite-schist, greenschist

Mylonitter
Mylonites

Granodior., amfib. i Andsfjell og Manken
Granodior., amphib. in Andsfjell and Manken

Strøk og fall
Strike and dip

Profil-linje
Section line

Malmforekomster
Ore deposits

Svovelkis
Pyrite

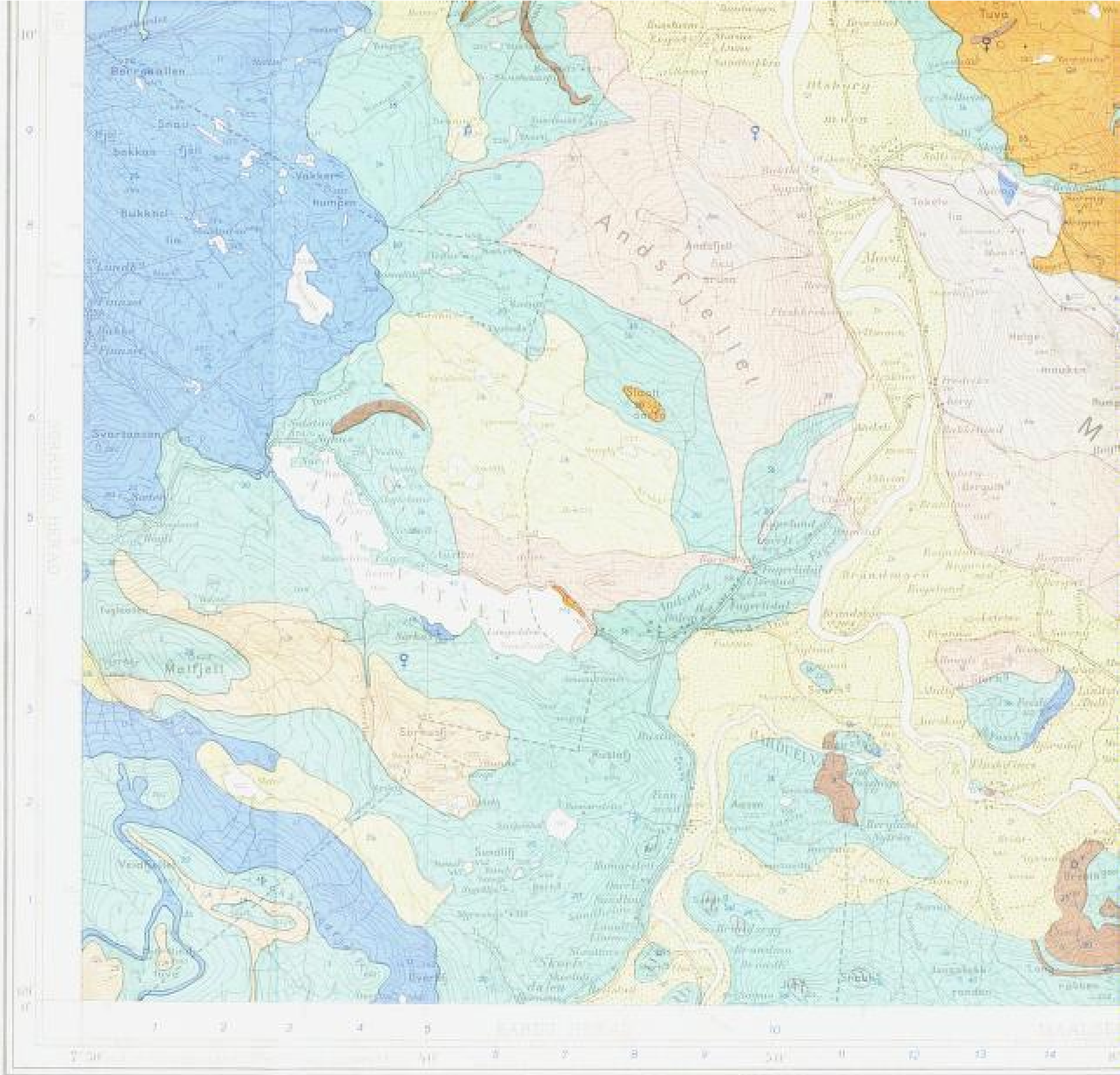
Jern
Iron

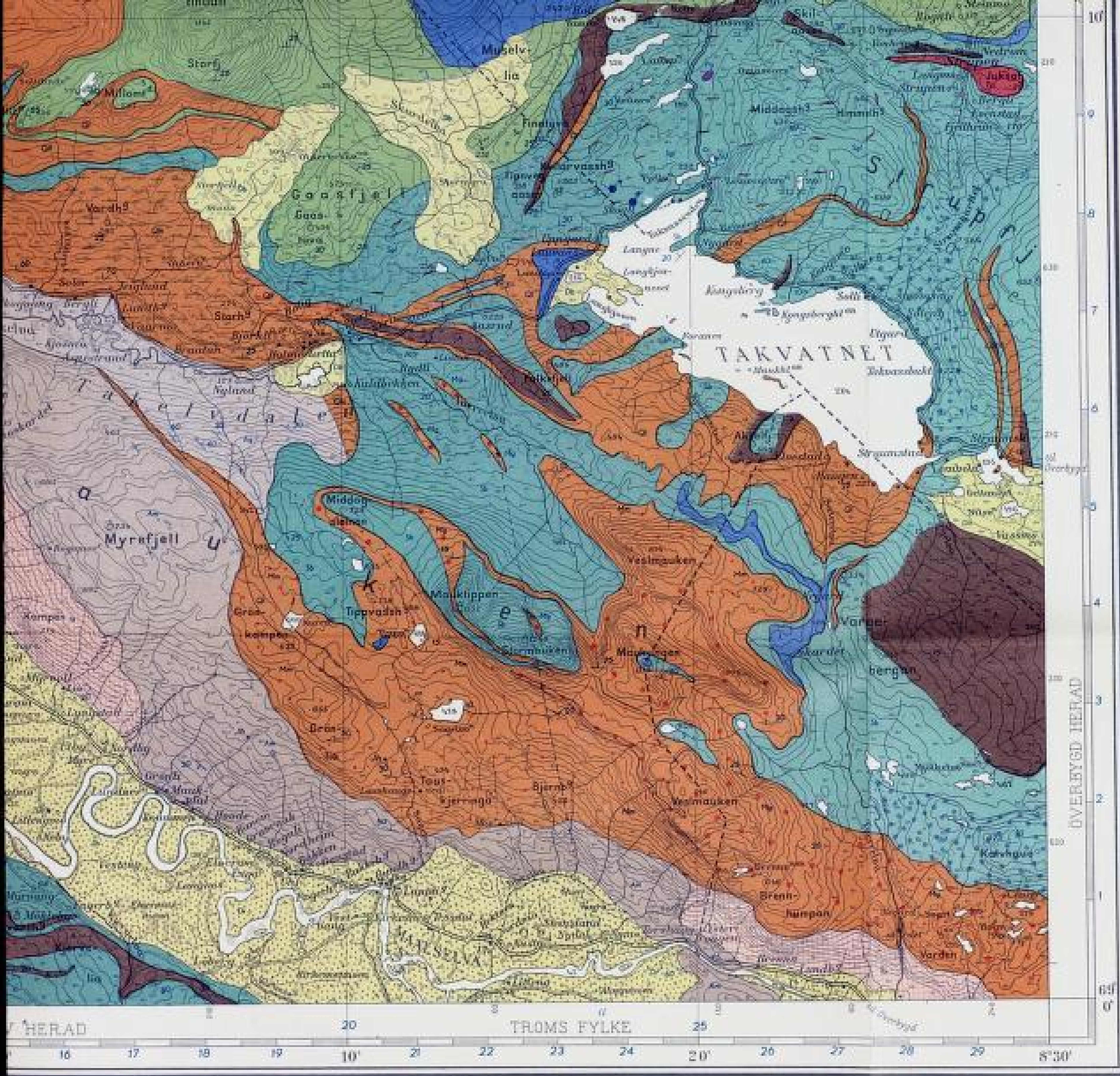
Titan-jern
Titaniferous

Sink og bly
Zink and Lead

Grafitt
Graphite

Kalk
Limestone





1 : 100 000

1 km i landet

5 6 7 8 9 10 km

1 geografisk mil

...nse 30 m

Geologisk kartlagt 1948—54 av Kåre Landmark, med støtte fra Norges Geologiske Undersøkelse. Trykt med bidrag fra Norges almenvitenskapelige forskningsråd.

Qh — Kvartsitt, Øvre kvartsittsone
Quartzite, Upper quartzite zone

Undre kvartsittsone / *Lower quartzite zone*:

Qs — Serisittkvartsitt
Sericite-quartzite

Ql — Kvartsitt med små mengder mikroklin
Quartzite, containing small amounts of microcline

Qll — Mikrolinkvartsitt, kvartsittgneis
Microline-quartzite, quartzite-gneiss

Mg — Mylonittgneis
Mylonite-gneiss

Mm — Mylonitt
Mylonite

(Qll, Mg, Mm:) Arkose, dels også kataklastisk granodioritt Ia
Arkose, in part also cataclastic granodiorite Ia

Cl — Kalk-konglomerat
Limestone conglomerate

Cq — Kvartsitt-konglomerat
Quartzite conglomerate

A — Amfibolitt
Amphibolite

Am — Amfibolitt i Mauken og Andsfjell
Amphibolite in Mauken and Andsfjell

As — Skifrig saussurittgabbro
Schistose saussurite gabbro

Ag — Grønnskifer
Green-schist

Ps — Serpentin
Serpentine

Pg — Sagvanditt
Sagvandite

Tp — Trondhjemitt, pegmatittisk eller grovkornet
Trondhjemite, pegmatitic or coarse-grained

II — Granodiorittisk gneis, vesentlig finkornet
Granodioritic gneiss, mainly fine-grained

Ia — Granodioritt i Andsfjell og Mauken
Granodiorite in Andsfjell and Mauken.