Petrography and Structural History of the Caledonian Rocks North of Haukelisæter, Hardangervidda

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The main features of the geology of the Caledonian nappe front north of Haukelisæter are described. The Precambrian basement in this area is overlain by an autochthonous/parautochthonous sequence in the east and by Caledonian nappes in the west. The rocks above the Precambrian basement are divided into five tectono-stratigraphic units; the autochthonous Cambro-Ordovician sequence, the Holmasjø Formation, the Nupsfonn Complex, the Dyrskard Group and the Kvitenut Complex.

At least five episodes of deformation may be distinguished (F_0–F_4) in the area. The F_0 structures in the Dyrskard Group and the pre-F_1 structures in the Nupsfonn and Kvitenut Complexes are thought to be of Precambrian age, whereas the last four episodes (F_1-F_4) are of Caledonian age. The F_1 folds trend ENE-WSW and are most probably related to the main episode of thrusting from the geosyncline southeastwards onto the Baltic Shield. During the F_2 episode thrusting occurred towards the NE causing the formation of NW-SE trending structures. F_3 is recorded only in the phyllitic quartz schist of the Holmasjø Formation, and the F_3 fold axes are subparallel to the F_2 fold axes. F_4 produced an intense crenulation cleavage within the pelitic rocks during a 'backward' movement of the nappes towards the north-west.

The rocks situated above the Precambrian basement show a discontinuous upward increase in metamorphism from low greenschist facies in the autochthonous sequence to upper almandine-amphibolite facies in the Kvitenut Complex.

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Introduction

This paper presents the results of reconnaissance studies on the south-western part of Hardangervidda (Fig. 1) carried out for a Cand. real. thesis. It also forms part of a larger project involving geological mapping, petrological study and structural investigation of the Caledonian rocks in the Hardangervidda–Ryfylke area. A preliminary account of the nappe succession in the Haukelisæter–Røldal area has already been published (Naterstad et al. 1973). It is the purpose of the present paper to present more detail concerning the petrography and structures of the nappe rocks as these data provided the basis upon which the nappe succession was established.

The eastern part of the area is an undulating surface 1200 to 1300 m above sea-level. It is an exhumed Precambrian peneplain with local remnants of overlying Cambro-Silurian basal layers. The central and western areas consist largely of a partly glaciated plateau about 300 m above the peneplain. The whole area lies above 1000 m with Sandfloeggi as the highest peak (1719 m,

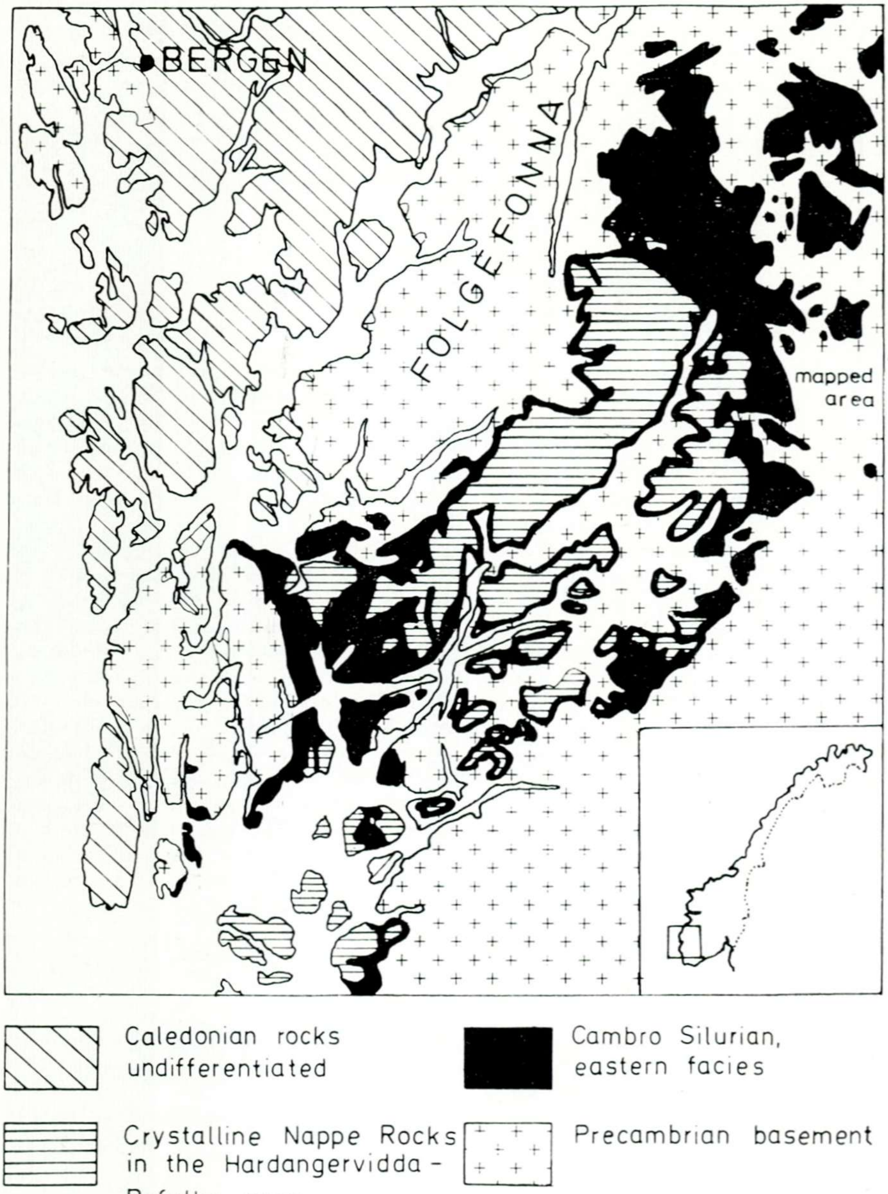


Fig. 1. Simplified geological map of south-west Norway, showing the location of the mapped area.

and is generally well exposed except for areas permanently covered by snow and ice.

The mapping was carried out on a scale of 1:25,000 using enlargements of the AMS M 711 1414 IV Haukelisæter map. The area is also covered by aerial photographs on a scale of approximately 1:37,000. The fieldwork was carried out during the summers of 1969 and 1970 and part of 1971. A total of 10 weeks was spent in the field.

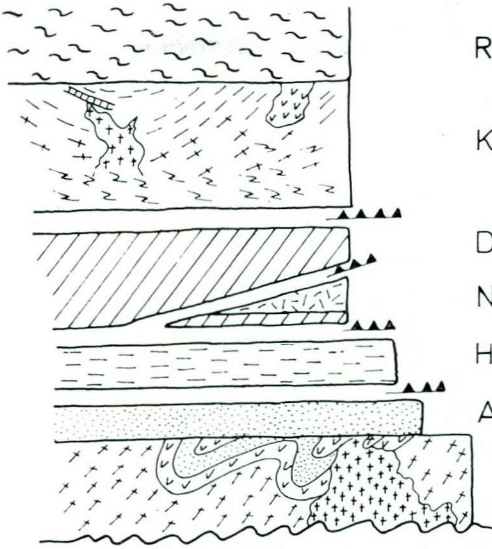


Fig. 2. Tectono–stratigraphic succession in the Haukelisæter–Røldal area (after Naterstad et al. 1973).

Earlier literature on the geology of Hardangervidda is very restricted. In 1861 Dahll described fossil finds (*Dictyonema flabelliforme*) from Hullberget (north of the mapped area), but Brøgger (1983) was the first to try to establish a stratigraphic column. He concluded that the whole rock-sequence above the basement was of Cambro-Silurian age or younger. Reusch, Rekstad & Bjørlykke (1902) published the first and, until now, only map from Hardangervidda. In this publication the authors made a threefold division into Precambrian basement, autochthonous Cambro-Silurian and overthrust gneisses ('høifjeldskvartsen'). The boundary between the Cambro-Silurian sediments and the overlying gneiss was marked as a thrust plane on the geological map (p. 29). This interpretation has since been accepted by most Norwegian geologists.

In a recent publication, Naterstad et al. (1973) have proposed a further subdivision of these uppermost allochthonous rocks, and the tectono-stratigraphic units and names adopted by Naterstad et al. will be used here (Fig. 2).

The rock units mapped in the Haukelisæter area continue to the north and to the south. Strand (1960, 1972) has reviewed the geology of these areas.

Petrography and field relationships

PRECAMBRIAN BASEMENT

The Precambrian basement dominates the northern, eastern and southern part of the mapped area. Basement rocks are also exposed in the U-shaped valleys in the western part. Lenses of basement rocks are also observed as thrust units

Table 1. Modal analyses of granodioritic augen gneiss (71.5), granodiorite (71.71) and granites. Based on point-counting of thin sections

Minerals / Samples	71.5	71.71	70.18	70.71	70.72	70.89	70.119	71.72
Plagioclase	44.2	44.4	39.2	29.5	42.1	36.7	38.1	38.5
Microcline	8.1	15.7	26.7	34.7	32.6	28.1	24.8	32.5
Quartz	27.6	13.2	27.0	32.3	18.4	24.2	27.4	25.5
Biotite	14.8	14.2	4.1	1.5	20.2	9.7	5.0	3.2
Amphibole	1.1	5.5	–	–	–	–	–	–
Chlorite	x	x	x	x	x	–	x	x
Sphene	x	2.5	0.4	x	3.1	0.8	1.5	x
Epidote	2.7	2.5	x	x	0.5	x	1.1	1.0
Muscovite	x	x	x	x	x	x	1.0	x
Zircon	x	?	?	?	x	x	x	–
Rutile	x	x	x	x	0.6	0.2	1.2	x
Ore minerals	x	x	–	x	x	x	x	x
Apatite	x	1.0	x	x	0.4	1.5	0.5	x

x: present in small amounts
–: not observed

in the overlying autochthonous sediments. Only the largest lenses are marked on the map.

Augen-gneisses and granites dominate in the basement, but supracrustal rocks and intrusives other than granites are also found.

Granodioritic augen-gneiss

A granodioritic augen-gneiss dominates the area around Hedlevatn. It contains thin layers of biotite gneiss and biotite-amphibole gneiss, and is transected by veins and dykes of granite, monzodiorite and dolerite. All the intrusives were emplaced after the pre-Caledonian deformation and metamorphism of the augen-gneisses.

The granodioritic augen-gneiss has a well-defined foliation due to strong preferred orientation of biotite. The augen are usually made up of euhedral microcline crystals in the core, surrounded by a biotite-free aggregate of quartz, plagioclase and K-feldspar. The augen, which have rod-like shapes, define a lineation in the plane of foliation. The size of the augen varies, but they are seldom more than 2 cm in diatmeter and 5–6 cm in length. Except for this small-scale inhomogeneity the augen-gneiss is fairly homogeneous throughout the area. A modal analysis of an augen-gneiss is given in Table 1.

The matrix of the augen-gneiss has a granoblastic texture. Plagioclase is an essential constituent of the groundmass and also occurs in the augen zones. It is a sericitized and saussuritized oligoclase (An_{20-24}) with frequent pericline and albite twins. Microcline is concentrated in the augen and the surrounding biotite-free zones. The euhedral microcline found in the core is sometimes up to 1 cm in diameter. The microclines display typical grid twinning, and are mostly developed as string-perthite. Biotite is the dominant mafic mineral and is partly altered to chlorite. It has pleochroism: X=yellow brown, Y=dark brown, Z=olive green. Most grains have numerous rutile inclusions arranged in a sagenitic pattern. Other major constituents are quartz and amphibole. The optical data for the amphiboles indicate common hornblende. White mica and epidote occur as alteration products of plagioclase. Apatite, zircon (subhedral), rutile, sphene and ore minerals are present in minor amounts.

From field evidence alone it is difficult to determine the origin of the strongly foliated granodioritic augen-gneiss. However, the homogeneity throughout the kilometre-wide rock body makes a magmatic origin likely. The occurrence of euhedral to subhedral zircons also suggests a magmatic origin. The present texture would then have to be explained by short-range diffusion within the granodioritic body with Mg^{++}, Fe^{++} and OH^{-} diffusing out of, and K^{+} and Al^{+++} into, the domains now occupied by the augen.

The foliated biotite-amphibole gneisses which occur as isolated layers within the augen-gneiss may then represent a still older complex which has been intruded by the granodiorite.

Telemark Suite

Supracrustal rocks of the Telemark Suite constitute the basement in Slettedalen, Trossovdalen, and around the western part of Ulevåvatn in a NW to NNW-trending synclinorium. These supracrustals can be divided into upper, middle and lower divisions (Naterstad et al. 1973). No contact between the augen-gneisses and the Telemark supracrustals has been observed within the mapped area, but from Valldalen (Nyastøl Bridge) Naterstad et al. have reported a sedimentary contact, with sandstones and conglomerates along the boundary showing the Telemark supracrustals to be the younger.

Acid volcanic rocks, characterized by Naterstad et al. (1973) as being typical of the lower division, are found in Trossovdalen. The supracrustal rocks in Slettedalen and around Ulevåvatn are mainly basic volcanics, typical of the middle division. Quartz schists of the upper division have not been observed. Both at Ulevåvatn and at Trossovdalen the supracrustals are intruded by post-tectonic, pre-Caledonian granites.

The Telemark Suite of rocks has not been studied in detail, but a few thin-sections have been examined from the metavolcanics and quartz schists of Ulevåvatn. The dominant rocks here are basic volcanics, but thin layers of quartz schist occur locally. Two types of basic metavolcanics are found: a fine-grained dark rock interpreted as a metabasalt, and a light-spotted rock with a relict coarse-grained texture. The latter rock-type is interpreted as a metagabbro or metadiorite.

The metabasalts show an indistinct foliation defined by parallel-oriented biotite and amphibole. Zones with biotite-chlorite aggregates are found in some places. These probably represent original vesicle fillings. Amphibole (ca. 40%) and biotite (15–20%) are the dominant minerals. The amphibole occurs as hypidioblastic prisms with a pale blue-green to green colour in Z direction. This together with extinction angle: $Z/c = 19°$, indicates actinolitic hornblende. Biotite (X colourless, Z, Y pale olive-green) is present as allotrioblastic to hypidioblastic flakes, partly altered to chlorite. The chlorite has an anamalous brown-green interference colour and is length-slow, a characteristic of Mg, Fe-chlorite (Trøger 1969). Plagioclase (4–10%) is strongly saussuritized and occurs as allotrioblastic grains. Its refractive indices indicate a basic oligoclase composition. Zoned epidote, with clinozoisite core and pistacite rim, occurs as an alteration product of plagioclase. Accessory minerals are black iron ore grains, quartz, sphene and apatite.

The metabasalts from Ulevåvatn are mineralogically different from those in Valldalen, which contain albite and chlorite. Chlorites from the metabasalts in Valldalen are not alteration products after biotite. This change in metamorphic mineralogy must reflect different conditions during breakdown of the original pyroxene-plagioclase mineralogy. Whether this indicates an increasing regional metamorphic gradient towards the east or is a contact-metamorphic phenomenon caused by the post-tectonic granite intrusion is unknown.

The light-spotted green rocks with relict coarse-grained texture have a distinct contact against the fine-grained metabasalts. Field observations have provided no definite evidence as to whether the rock is a porphyritic effusive or an intrusive gabbro or diorite. The zones of light-spotted rocks are sometimes seen to thin out. The rock usually displays a weak foliation. Amphibole is the dominant mineral, and is found as 1 mm-long subhedral porphyroblasts showing sieve texture. These have grown in aggregates within which the grains are parallel, but each aggregate has a different orientation. A few amphiboles are zoned. The amphibole has pleochroism: X=blue-green, Y=yellow green, Z=green, and extinction angle Z/c=17°–18°, indicating common hornblende. Plagioclase is strongly saussuritized oligoclase/andesine. Chlorite is present in considerable amounts. The optical data indicate Mg-chlorite (Trøger 1969). Biotite has not been found. Epidote, sphene, leucoxene and iron ore grains are present as accessory minerals.

It is possible that this rock is a metagabbro, as the metabasalt/metagabbro association is well-known both from Skånevik (Mortensen 1942) and Sørfjorden (Kvale 1945).

A thin band of feldspathic quartz schist with basic metavolcanics on both sides is found in Middyrelva. The schist is laminated and fine-grained. The lamination is defined by concentration of amphibole and epidote in thin zones, most probably reflecting a sedimentary compositional feature. Grain-size varies between 0.05 mm and 0.2 mm. Primary sedimentary textures have been obliterated by recrystallization, which has produced a granoblastic texture. Quartz and plagioclase are the dominant minerals. Plagioclase (An_{18-25}) occurs as small recrystallized grains in the groundmass, and as larger relict clastic grains. A few microcline grains are found. Amphibole is present as weakly parallel-orientated, idioblastic prisms (less than 0.1 mm). Pleochroism is X=pale green, Y=Z=pale green to colourless, and extinction angle Z/c=16–17°, indicating actinolite. Epidote is associated with amphibole. Pyrite is present as cubes. Other minerals found are sericite, chlorite, zircon (rounded) and tourmaline.

Intrusives

Monzodiorite: Undeformed monzodiorites intruding the augen-gneiss have been found east of Årmoteggi and north of Hellevassbu. As seen on the geological map (Plate 1) the monzodiorites have a very local distribution. At the locality north of Hellevassbu the monzodiorite is cut by dykes of younger granites, thus setting an upper age limit for intrusion of the monzodiorite.

On weathered surfaces 4–6 mm long aggregates of amphibole and biotite in fine-grained matrix give the rock a characteristic appearance. The matrix is equigranular with some euhedral plagioclase phenocrysts (up to 3 mm). The plagioclase is strongly sericitized. Biotite and amphibole are the only mafic minerals present in appreciable quantities. Biotite occurs as brown, hypidioblastic, weakly deformed flakes with abundant rutile and haloed zircon inclusions. The amphiboles occur as unoriented, poikilitic, twinned porphyroblasts with pleochroism: X=pale yellow-green, Y=green, Z=olive-green, and extinction angle X/c=18°, indicative of common hornblende. Quartz and microcline are found as 0.1 mm anhedral grains between the plagioclase phenocrysts. The microcline is perthitic with a typical quadrille grid pattern. The perthites can be classified as 'rod-perthite'. Accessory minerals are euhedral apatite, zircon, sphene, epidote, sericite and tourmaline.

Granites: Except for the area round Nyastøl bridge (Naterstad et al. 1973) the Telemark supracrustals are in every case separated from the basement gneisses by granite intrusions. The granites clearly intrude both the Telemark supracrustals and the granodioritic augen-gneisses. Fine-grained, coarse-grained and porphyritic varieties of granites have been observed. Since little attention has been given to the Precambrian basement rocks it is at present uncertain whether or not these texturally different types belong to different periods of intrusion. In this publication all the granites will be treated as local varieties within the same pluton and will be described together.

In the east and north homogeneous medium-grained granites predominate, while the area south of Store Nup is dominated by porphyritic types. The porphyritic type often shows phenocrysts of microline-perthite up to 3 cm in length. Along highway E 76 a coarse-grained type is common. There seems to be a gradual transition between these three types, as sharp contacts have never been observed. Sharp contacts do occur, however, between the granites and some clearly younger aplitic and pegmatitic veins which transect them.

Quartz, microcline, plagioclase and biotite are the main minerals. Plagioclase, An_{25} to An_{32}, is saussuritized and often shows normal zoning. Myrmekite is observed along the microcline/plagioclase contact. The only mafic mineral is biotite, which is brown and rich in rutile inclusions arranged in sagenite texture. Modal compositions of the granites are shown in Table 1.

Granodiorite: At the northern shore of Ulevåvatn a small body of unfoliated granodiorite is found within the granite. The granodiorite is homogeneous and medium-grained and, in the field, resembles a biotite-rich variety of the granites. The modal analysis (Table 1) shows, however, that the rock has a granodioritic composition. Besides this, the granodiorite contains amphibole, a mineral not present in the surrounding granite.

Unfortunately the contact between the granodiorite and the granite has nowhere been found. It has therefore not been possible to determine whether the granodiorite represents a separate episode of intrusion, or is simply an assimilation phenomenon in the granite. The author favours the first possibility.

If one accepts the granodiorite body as representing a separate intrusive episode, the intrusion must have taken place after the deformation of the Telemark supracrustals. This is indicated by the fact that the Telemark supracrustals, which are found close to the granodiorite body, are strongly folded and foliated, while the granodiorite is completely undeformed.

Amphibolites: Dykes of an aphanitic, dark rock are found within the granite in the road-cuts along highway E 76. These dykes are from a few centimetres up to 3 m in width. Although the rock is altered now, and consists of saussuritized plagioclase, amphibole and chlorite, a relict sub-ophitic texture can be seen; it was therefore most likely a dolerite dyke. These amphibolite or metadolerite dykes are assumed to be the youngest rocks within the Precambrian basement.

AUTOCHTHONOUS CAMBRO–ORDOVICIAN SEQUENCE

Recent regional-scale mapping has shown that the autochthonous Cambro–Silurian sequence described by Brøgger (1893) and Reusch et al. (1902) can be divided into two tectonic units in the investigated area with a distinct thrust plane between (Naterstad et al. 1973). One unit is autochthonous or partly parautochthonous and will, in this publication, be reffered to as 'the autochthonous Cambro–Ordovician sequence'. The other unit is allochthonous and is called the Holmasjø Formation, a name proposed by Naterstad et al. (1973). As the autochthonous sequence has chlorite marble and chlorite schist as the uppermost units, most probably of Ordovician age (3c?) (Brøgger 1893), the designation Cambro–Ordovician is used instead of Cambro–Silurian for this unit.

The autochthonous sequence is exposed in the northern and eastern parts of the mapped area (see map, Plate 1), lying disconformably upon the Precambrian basement. Owing to peneplanation in late Precambrian time this contact is almost horizontal, and as a result of differing degrees of resistance to erosion the sub-Cambrian peneplane can be seen as a shoulder in hillsides where the contact is exposed. Intense deformation with folding and thrusting has destroyed the autochthonous sequence in the south and here it is assumed to be in a parautochthonous position.

The stratigraphic sequence established by Brøgger (1893) (Fig. 3) is found only NW of Hellevassbu, but even here the deformation is too intense for detailed stratigraphic or sedimentological studies.

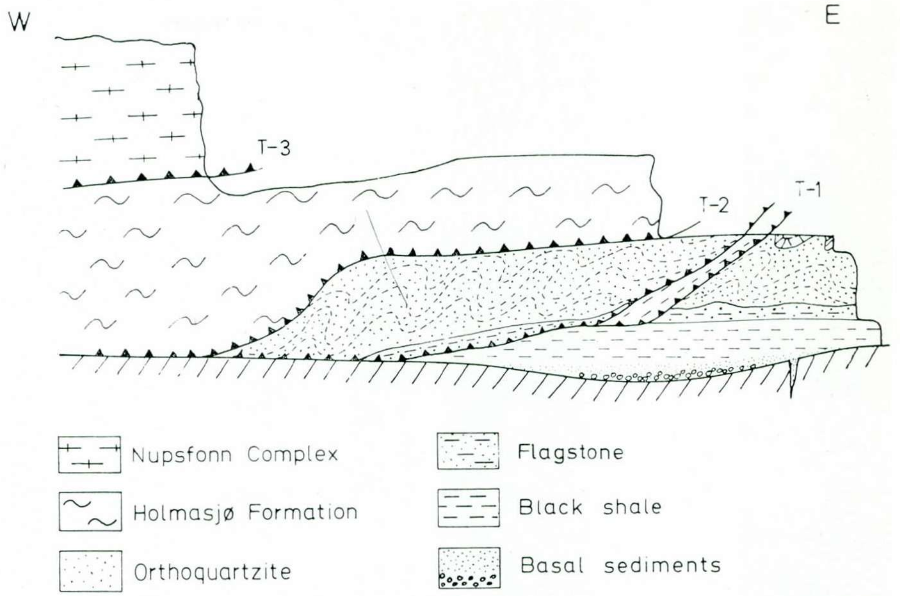


Fig. 3. Schematic section from the Precambrian basement in the east to the crystalline nappes in the west. Note the imbrication structures indicated in the autochthonous sequence.

The subdivision of the autochthonous sequence used in the following pages is taken from Brøgger (1893), but new names are given for some of the units in accordance with modern terminology (Fritsch et al. 1967).

Basal sediments

Conglomerates, sandstones and siltstones are found as basal sediments deposited directly upon the sub-Cambrian peneplain. The thickness of these sediments is everywhere less than 3 m, and usually only a few decimetres. Conglomerate is the most common lithology, and good exposures are found in the area around Hedlevatn. The thickness varies from 0 to 2 m. Such variation in thickness has been observed in a single exposure over a distance of 100 m, and seems to be related to the filling of depressions in the peneplain (Fig. 3)

Variation in the size of clasts gives the conglomerates a distinct layering. The individual beds vary in grain-size from sand to gravel, with pebbles up to 5–6 cm across. The basal conglomerates are polymictic with fragments of quartz and feldspar as well as quartzite and granitoid pebbles. In some places the conglomerates show a grain-supported texture, while in other areas the pebbles are distributed in a fine-grained matrix. Most fragments are rounded to well-rounded. The fine-grained clastic grains of the matrix usually have the same mineralogical composition as the larger clasts. However, in some places phosphate minerals dominate the matrix. South of Hedlevatn calcite has been found as a matrix between the clasts. A dark sandstone and mudstone unit is found as a transitional zone between the basal conglomerate and the overlying black shale. Where the conglomerates thin out and disappear the sandstone

Fig. 4. Fractures in the sub-Cambrian peneplain filled with silt and fine sand of (?) Lower and Middle Cambrian age. Lake 1302 NE of Hellevassbu.

and mudstone have depositional contacts with the Precambrian basement, as illustrated in Fig. 3. Good examples of this are seen at lake 1302, north of Hedlevatn. At this locality the sandstone is not only deposited on the basement, but is also filling cracks in the Precambrian peneplain (Fig. 4). Fragments of fossils (*? Torellella*) have been found in one of these sand-filled cracks, and trace fossils occur in the sandstone about 1 cm above the basement. Galena impregnation has been observed in the basal sediments around Haukelisæter.

Black shale

The dark sandstone and mudstone unit shows a gradual upward transition into a black shale. In earlier literature (Dahll 1861, Brøgger 1893) this shale has been designated 'Alum shale' or 'Dictyonema shale'.

The appearance and thickness of this incompetent black shale are governed by its stratigraphic position between the underlying Precambrian basement and the overlying competent orthoquartzite. At its greatest thickness, in the north, it is nearly 100 m, but in most places it is less than 10–15 m and locally it can be completely squeezed out. Brøgger (1893) assumed that the original thickness of these beds was about 40 m. Owing to intense isoclinal folding and imbrication caused by décollement type deformation, thin zones of black shale are locally found within the overlying orthoquartzite (Fig. 3).

Typical features of this shale are its black colour, well developed secondary cleavage and lack of sedimentary variation. Thin beds and laminae of black sandstone are only found in the uppermost part of the shale.

Carbonaceous material and quartz make up 90% of the shale. The carbonaceous material occurs along the cleavage planes, while quartz is found in small, elongated, recrystallized lenses with mosaic texture. White mica occurs both as large deformed and broken detrital flakes, and as small flakes of porphyroblastic origin. Microcline is also found as clastic grains. Pyrite is a common mineral.

At Hullberget some tens of kilometres towards the east *Dictyonema flabelliforme* was reported by Dahll (1861) in the uppermost part of this black shale. In the present area, however, intense secondary cleavage in the shale leaves no hope of finding fossils.

Flagstone

The next unit in the Cambro–Silurian sequence described by Brøgger (1893) is a 10–15 m thick unit of flagstone ('storhellet skifer'). Rocks typical for this unit are found at a few localities within the area, but the observed thickness is only 4–5 m. The unit consists of centimetre- to decimetre-thick layers of weakly metamorphosed sandstones with a variable content of pelitic material. Most of the sandstones show good sorting, and the clastic grains, predominantly quartz with some plagioclase and microcline, are well-rounded. Corrosion phenomena are common along grain boundaries. The matrix consists of quartz, magnetite and white mica, together with some chlorite and carbonaceous material. The characteristic slaty cleavage is developed in the mica-rich layers. In a few beds calcite has been observed as the cementing material.

Orthoquartzite

An orthoquartzite is the dominant rock unit in the autochthonous sequence, covering large areas south and west of Hedlevatn. Cliffs of bluish quartzite more than 100 m high can be seen in these areas. Because of this blue colour the rock has been named 'blue-quartz' in earlier literature (Brøgger 1893). In the weakly folded areas north of Hellevassbu, observations indicate that that the real thickness of the orthoquartzite is less than 40 m, in accordance with observations made by Brøgger.

The great difference in competence, together with intense deformation, has produced tectonic boundaries between the orthoquartzite and the underlying rocks in most places. Good examples of this are seen at Nupsflott and in the southern part of Årmoteggi, where the orthoquartzite lies directly upon the Precambrian basement.

The orthoquartzite is massive and is without sedimentary structures except for occasional interlayering of some calcareous lithologies. The rock is completely recrystallized, showing a granoblastic texture with grain-size 0.05–0.1 mm. Distribution of a finely disseminated ore mineral most probably along primary clastic grain boundaries, indicates good sorting and grain-size 0.6–

0.8 mm in the original sediment. Quartz makes up 90–95% and feldspar 5–10% of the rock, with microcline as the dominant feldspar. The finely disseminated ore mineral (? magnetite) is thought to be responsible for the blue colour of the rock.

The orthoquartzite contains a great number of post- or late-tectonic, white, hydrothermal quartz veins, often rich in perfectly shaped quartz crystals.

Chlorite marble

A gradual change from orthoquartzite to chlorite marble can be observed in the weakly deformed rocks north-west of Hellevassbu. Here the chlorite marble can also be found folded into the orthoquartzite. A parallel orientation of white mica and chlorite gives the rock a distinct bedding foliation. The content of chlorite varies, and locally the rock can be classified as a marble (Fritsch et al. 1967).

The chlorite marble has a granoblastic to lepidoblastic texture with average grain-size 0.05–0.15 mm, although some coarser grained marbles are observed. Calcite (55–90%) is the dominant mineral. Chlorite occurs as lepidoblastic flakes which often show signs of a weak deformation (bending). Both Mg-chlorites and Fe-Mg-chlorites are found. The Fe-Mg-chlorites occur in rocks rich in ore minerals. Albite (An_{3-4}) and quartz (18–20%) are evenly distributed, while white mica (1–12%) is enriched in the carbonate-poor zones. Two generations of mica are observed. The first generation is represented by deformed anhedral flakes (clastic ?), while the second generation comprises smaller, undeformed idioblastic flakes. Accessory minerals are tourmaline (euhedral prisms), rutile, zircon (rounded) and iron ore minerals.

Chlorite schist

The chlorite marble is overlain by a chlorite schist; the only place where this rock has been found is west of Buadalen. Here the transition from orthoquartzite via chlorite marble to chlorite schist is preserved. It is important to stress that this chlorite schist belongs to the autochthonous Cambro–Ordovician sequence and not to the chlorite schists of the Holmasjø Formation. The most pronounced difference between this chlorite schist and the schists of the Holmasjø Formation is seen in the higher content of chlorite and lower content of quartz in the former. The autochthonous chlorite schist has calcite as a minor mineral, while this mineral is seldom observed in the Holmasjø Formation.

Holmasjø Formation

Brøgger (1893) and Reusch et al. (1902) grouped all the rocks lying above the Precambrian basement but below the crystalline nappe rocks into an autochthonous Cambro–Silurian sequence. However, the uppermost unit of this autochthonous sequence, the grey phyllite rich in quartz segregations, has a distinct thrust contact against the underlying autochthonous rocks, and has been given the separate name Holmasjø Formation (Naterstad et al. 1973). The type locality is around lake Holmasjø east of Nupseggi.

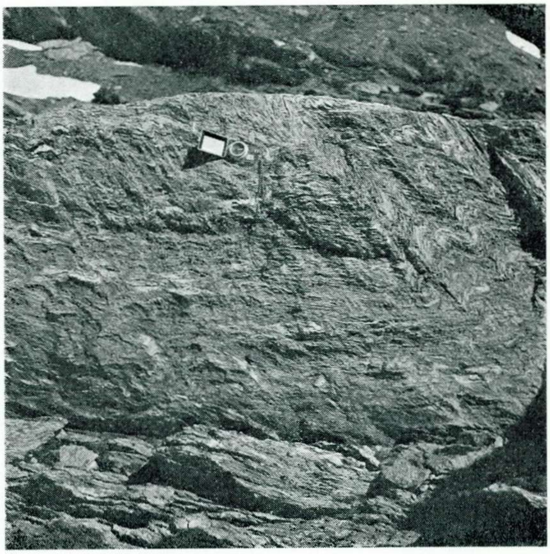


Fig. 5. Phyllitic quartz schist of the Holmasjø Formation with two generations of planar structures (S_2 and S_4). Asymmetric kink folds (F_4) are developed in the zone rich in quartzite layers (upper part), while an S_2 transposition foliation dominates in the lower part. South of Holmasjøen.

The thrust plane below the Holmasjø Formation is rather flat-lying. Along Årmoteggi, however, where the Holmasjø Formation overlies the orthoquartzite, the thrust plane strikes NW–SE and dips steeply to the south–west. As the NW–SE orientation of the thrust plane is rather constant over a considerable distance, the direction of thrusting is thought to be from SW to NE. South-west of Årmoteggi, the autochthonous sequence is not present except for some local remnants, and the Holmasjø Formation there lies directly upon the Precambrian basement (Fig. 3).

In the area north of Langevatn the thrust zone is not a single one, but consists of several thrust planes, which has produced a complex pattern of wedges of orthoquartzite, phyllonitic quartz schist and some black shale. The wedges of orthoquartzite vary in size, but all are concentrated along the thrust zone. The same phenomenon is also observed along the contact between the Precambrian basement and the Holmasjø Formation, where mylonitized basement wedges are found within the overlying Holmasjø Formation.

The thickness of the Holmasjø Formation is rather variable as the phyllitic rocks of the formation have acted as an incompetent unit between the basement and the crystalline nappe rocks during the Caledonian orogeny.

The Holmasjø Formation is dominated by low-grade metamorphosed and intensely deformed, laminated pelitic and semi-pelitic rocks. Massive beds of meta-arkose are also found. Zones of black shale (alum shale) have been observed below Store Nup. Whether these black shale horizons represent primary deposits of the Holmasjø Formation or have been introduced during thrusting or folding is at present an unanswered question.

Intense multiple folding, with the formation of at least two sets of secondary planar structures (Fig. 5), has destroyed the primary bedding and lamination in most places. Locally, the deformation has been so intense that the rocks may be classified as phyllonite. Even the semi-pelites often have a phyllitic appearance.

White mica, chlorite and quartz are the dominant minerals, the relative proportions varying from rock to rock. Most of the rocks studied in thin-sections, however, can be classified as white mica–chlorite–quartz schists. There is a gradual transition from this rock-type to chlorite–muscovite schists. This subdivision of the schists cannot be indicated on a map owing to the transitional contacts. There are strong colour variations from grey to green to black, depending on the content of iron ore minerals, carbonaceous matter and chlorite, but again these have proved useless as a basis for subdivision during mapping.

The high content of phyllosilicates gives the rocks a well-developed secondary cleavage. The white mica is twisted and bent parallel to the plane of foliation. Quartz usually occurs in zones and lenses with a recrystallized granoblastic texture. Clastic grains have only been observed in the meta-arkoses.

Hypidioblastic grains of chlorite, 0.01 mm across, occur in all rocks; they are usually evenly distributed by may be enriched in zones parallel to the secondary foliation. The chlorite flakes are undeformed or only weakly deformed. Chlorite has been found in only one thin-section as an alteration product of biotite. The optical data from the chlorites indicate variable chemical compositions.

Biotite has been identified in only one thin-section. In this it is strongly deformed and chloritized, and may have been of clastic origin. Most rocks show a high content of iron-ore minerals. Other accessory minerals are sphene, tourmaline, apatite, zircon, albite (?), stilpnomelane, carbonate, and some carbonaceous matter. Albite (An_{0-2}), together with chloritized garnets, was found in some semi-pelites collected for radiometric age determinations a few km west of the area under investigation.

THE CRYSTALLINE PARTS OF THE HARDANGERVIDDA–RYFYLKE NAPPE SYSTEM

The allochthonous rocks above the Holmasjø Formation have been treated as a single tectonic unit in earlier literature (Reusch 1913, Reusch et al. 1902, Strand 1960), and Strand (1972) introduced the name Hardanger Nappe for these rocks. Recent mapping by Naterstad et al. (1973), however, has shown that four tectonostratigraphic units can be identified within the allochthon (Fig. 2). These authors have proposed the name 'Hardangervidda–Ryfylke Nappe System' for these rocks, as they have been traced southwestwards to the Boknfjord–Ryfylke area. The uppermost part of the Cambro–Silurian sequence described by Brøgger (1893) (the Holmasjø Formation of this publication) is included in this nappe system.

The four tectonostratigraphic units above the Holmasjø Formation are the Nupsfonn Complex, the Dyrskard Group, the Kvitenut Complex and the Revsegg Formation. Of these four units the lowermost three occur in the present area, while the uppermost unit, the Revsegg Formation, is situated 2–3 km to the west (see map in Naterstad et al. 1973).

The boundary between the Holmasjø Formation and the overlying crystalline rocks is easy to map as it invariably forms a marked topographical feature. The crystalline rocks form 200–300 m high cliffs above the Holmasjø Formation, and in most places the boundary can be traced directly from aerial photographs. The boundary is easily recognized as a thrust plane were the mylonitized intrusives of the Nupsfonn Complex are in contact with the phyllitic quartz schists of the Holmasjø Formation. Some difficulty arises when trying to identify a thrust plane between the supracrustals of the Dyrskard Group and the phyllitic quartz schists, as intense deformation has taken place here not only along the contact but on both sides of the formational boundary with the resulting development of concordant structures. However, both the petrographical and the structural data, presented below, reveal that a major break is also present between the Dyrskard Group and the Holmasjø Formation.

Nupsfonn Complex

The Nupsfonn Complex, the lowermost unit within the crystalline part of the Hardangervidda–Ryfylke Nappe System, occurs as a local wedge with restricted distribution lying between the Holmasjø Formation and the Dyrskard Group, and covering the area around Nupsfonn. Five klippe southeast of Nupsfonn are also composed of rocks of the Nupsfonn Complex. As may be seen from the map and profiles the wedge of Nupsfonn Complex rocks trends NE–SW, with its greatest thickness in the southeast. Towards the southwest, along Nupsdalen valley, the complex suddenly disappears. From a thickness of about 300–400 m around Nupsfonn it is reduced to zero within a distance of 2–3 km, and the Dyrskard Group is then found in contact with the Holmasjø Formation. On the northeastern side of the Nupsdalen valley it can be seen that this sudden change in thickness is partly a result of the presence of an overturned, isoclinal F_2 antiform (Fig. 22). The Holmasjø Formation is here locally folded up and over the Nupsfonn Complex. The Nupsfonn Complex is absent in the adjacent areas to the west and north, where either the Dyrskard Group or the Kvitenut Complex lies directly above the Holmasjø Formation. It may have originally extended further to the south and east and been removed by erosion. The distribution and tectonostratigraphic position of the Nupsfonn Complex will be discussed in a later chapter.

The Nupsfonn Complex is heterogeneous, consisting of metasediments, metavolcanics, migmatites, gneisses and various types of intrusives. At the present stage of investigation these rocks have been grouped into metasupracrustals, banded gneisses and intrusives. A characteristic feature of most of the rocks is a more or less pronounced, brittle, cataclastic texture. This often makes it very difficult to decide whether one is dealing with a metamorphic sediment or a strongly mylonitized intrusive with cataclastic layering. The cataclasis is not restricted to the thrust zone bordering the complex, but is found throughout the whole complex. The cataclastic texture is especially well developed in the quartzo–feldspathic rocks, while amphibolites and metagabbros show only minor signs of cataclastic deformation, although they too are quite clearly pre-cataclastic intrusives. This variation in resistance to cataclasis is in agreement with observations made by Hossack (1966) from the Jotun Nappe. Since thin-section studies have been necessary to distinguish the petrographically different rock-types, the author has not been able to subdivide them on the map.

These rocks of the Nupsfonn Complex show post-cataclasis low-grade metamorphism, in contrast to the post-mylonitic low almandine–amphibolite facies metamorphism along the Dyrskard Group/Kvitenut Complex boundary.

Metasupracrustals: Assumed metasupracrustal rocks occur nearly everywhere directly above the thrust plane that separates the Nupsfonn Complex from the Holmasjø Formation. The thickness is usually less than 30 m, except in the area around Nupstjern, where the thickness increases to about 100 m. In other places these beds thin out and disappear.

Around Nupseggi and Vesle–Nup the metasupracrustals are dominated by feldspathic quartzites and meta-arkoses, while chlorite-rich and fine-grained metagreywackes predominate farther west. Calcareous rocks are also observed. The metagreywackes are very fine-grained with some larger possibly clastic grains (0.4 mm) of plagioclase (An_{22-35}). Plagioclase (An_{25-27}) is also found together with quartz in the matrix. Rotated and deformed pre-cataclastic amphibole porphyroblasts with hornblende cores and actinolitic rims occur together with brown biotite. The biotite flakes are chloritized. Chlorite also occurs as undeformed neocrystallized flakes. Other minerals present are sphene, calcite, white mica and rounded zircon.

Layers of greenschist with plagioclase, actinolitic hornblende, epidote and chlorite are found interbedded with the metagreywackes, indicating the presence of volcanic material in the metasupracrustals. Thin horizons of black schist have also been observed.

Banded gneiss: Banded gneiss covers large areas northwest and south of Nupsfonn. This rock-type may also be of supracrustal origin, although it differs in many ways from the metasupracrustals already described. The banded gneiss is more coarse-grained than the metasupracrustals, and primary sedimentary layering has not been recognized. Intrusive rocks are never found in the metasupracrustals, whereas both granites and amphibolites are common in the banded gneiss. Intense intrafolial folding can be observed within a single gneiss band (Fig. 6); in this case the axis of folding is not measurable.

Quartzo–feldspathic gneiss is the predominant lithology, this varying in composition from granitic to dioritic. Except for amphibolite bands the gneiss is poor in mafic material. Migmatitic zones are sometimes observed in the gneisses (Fig. 7).

The texture is dominated by the cataclastic deformation so typical of the Nupsfonn Complex, giving the gneisses a secondary cataclastic layering. Bent and broken plagioclase and K-feldspar grains indicate the brittle nature of the cataclastic deformation. Quartz has crystallized in a mosaic texture in cracks in the broken feldspar porphyroclasts. Both sharp-edged and rounded por-

Fig. 6. Banded gneisses of the Nupsfonn Complex with intrafolial folding. The faults are normal faults with strike parallel to the axes of F_4 folds. Between Nupsfonn and Nupsdalen.

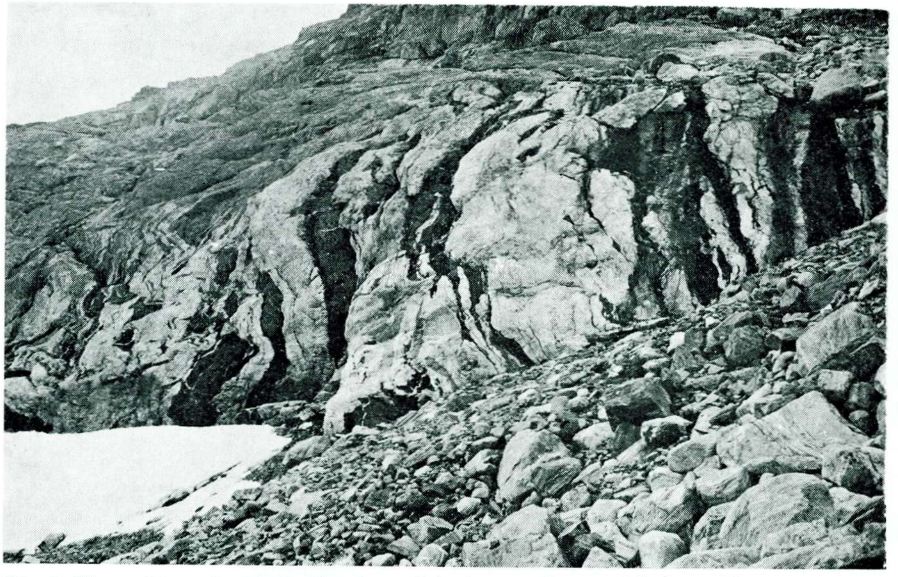


Fig. 7. Migmatitic rocks of the Nupsfonn Complex, southwest of Lake 1486.

phyroclasts are found ,sometimes up to 3–4 mm in length. The K-feldspar is now a microcline–perthite. The exsolved plagioclase has a rod-like shape, with refractive indices indicating oligoclase. The main plagioclase is an oligoclase (An_{16-30}) in which albite-twins are common. White mica is enriched in thin zones, these possibly representing relicts of pre-cataclasis muscovite from which the new generation of white mica has grown. Deformed, brown biotite is the dominant mafic mineral, and both the biotites and the white micas show parallel orientation. Other minerals are epidote, often zoned with an orthite core, sphene, zircon, apatite, chlorite and ore minerals.

Less common than the quartzo-feldspathic gneisses are bands of amphibolite and biotite amphibolite. The amphibolites are nowhere found cutting the quartzo-feldspathic layers. Amphibole and biotite are the dominant minerals and occur in variable proportions .The amphibole porphyroblasts are pleochroic with X = yellow-green, Y = green and Z = olive-green and Z/c = 18–19°, indicating common hornblende. Biotite is found as weakly parallel-oriented porphyroblasts, often with sagenitic texture. Two generations of biotite growth are observed in the biotite amphibolites. The first generation consists of large, deformed brown flakes with exsolved rutile and leucoxene. Small flakes of undeformed green biotite characterize the second generation. Accessory constituents are sphene, apatite, rutile, leucoxene, chlorite and ore minerals.

Igneous rocks: Igneous rocks, varying in composition from granite to quartz diorite (Fig. 9), occur predominantly around Store Nup and Nupseggi. Differentiation of the various types of intrusives in the field is impossible due to the intense cataclastic deformation, which gives the rocks a uniform appearance. All types of intrusive have been allotted the same symbol on the map, and they will be described together here.

The igneous rocks are often found in contact with the metasupracrustals, but the intense cataclastic deformation makes interpretation of age relationships difficult. Intrusives are never observed within the metasupracrustals, a feature which may indicate a tectonic contact between the latter and the overlying igneous rocks, but an inverted basement/cover contact cannot be excluded. As already mentioned, the banded gneisses are intruded by granites This, together with the gradual transition from gneisses via migmatites to intrusives, indicates a close relationship between these rocks. Unfortunately the area between the intrusives along Nupseggi and the gneisses around lake 1486 is covered by permanent snow and ice, so detailed study there is impossible.

A gneissic texture (cataclastic layering), as described for the quartzo–feldspathic gneisses, is common for all the igneous rocks. Primary magmatic textures can, however, be identified in some rocks, these indicating fine- to medium-grained rock-types with quartz, K-feldspar and plagioclase as essential minerals. The only major mafic mineral observed is biotite. The K-feldspar is a microcline perthite with occasionally saussuritized flame-shaped plagioclase component (acid oligoclase) as shown in Fig. 8. The main plagioclase is oligoclase, but albite has also been observed. Two generations of biotite are found; the first is represented by large, strongly deformed, brown flakes thought to be of primary magmatic origin, while the second is seen as small, unoriented, undeformed flakes of green biotite. Stilpnomelane is found in some rocks, but never together with second-generation biotite. The stilpnomelane has grown as radiating clusters of thin needles.

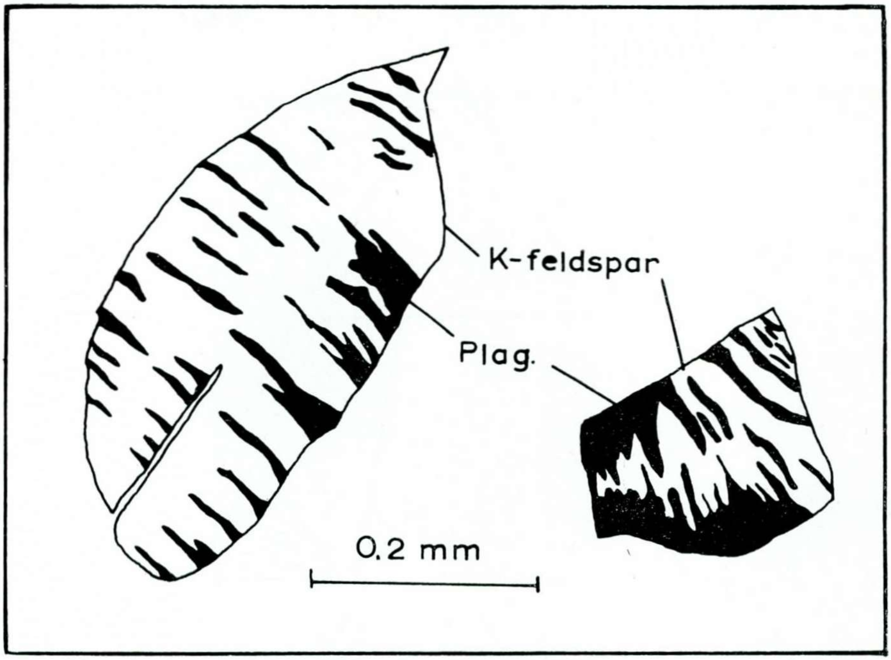


Fig. 8. Flame-like perthites from a granite within the Nupsfonn Complex.

The igneous rocks group into three different categories (Fig. 9): granite, granodiorite and quartz diorite. Quartz dioritic veins are observed cutting through the granodiorite in the area round Nupseggi. The age relations between the granite and the granodiorite/quartz diorite have not been determined.

The quartz diorites have only been found along Nupseggi, while the granites and granodiorites occur along Nupseggi and Store Nup. Southeast of the summit of Store Nup the granodiorite is intruded by amphibolites. These amphibolites are in some places found to connect with small bodies of foliated metagabbro.

Dyrskard Group

The Dyrskard Group is the main tectonostratigraphic unit in the northern and western parts of the mapped area. Two small klippe are also situated above the Nupsfonn Complex just west of the Nupsfonn glacier. The Dyrskard Group has a tectonic boundary against the underlying Nupsfonn Complex. The thrust plane between the two units dips gently north–westward and the boundary is easily recognized on passing from the strongly mylonitized gneisses and intrusives of the Nupsfonn Complex into the fine-grained metasediments of the Dyrskard Group.

In its northern and western parts the Dyrskard Group has a thrust contact against the Holmasjø Formation. It has never been observed in direct contact with the autochthonous Cambro–Ordovician sequence. The thrust plane is folded by F_2 and F_4 folds.

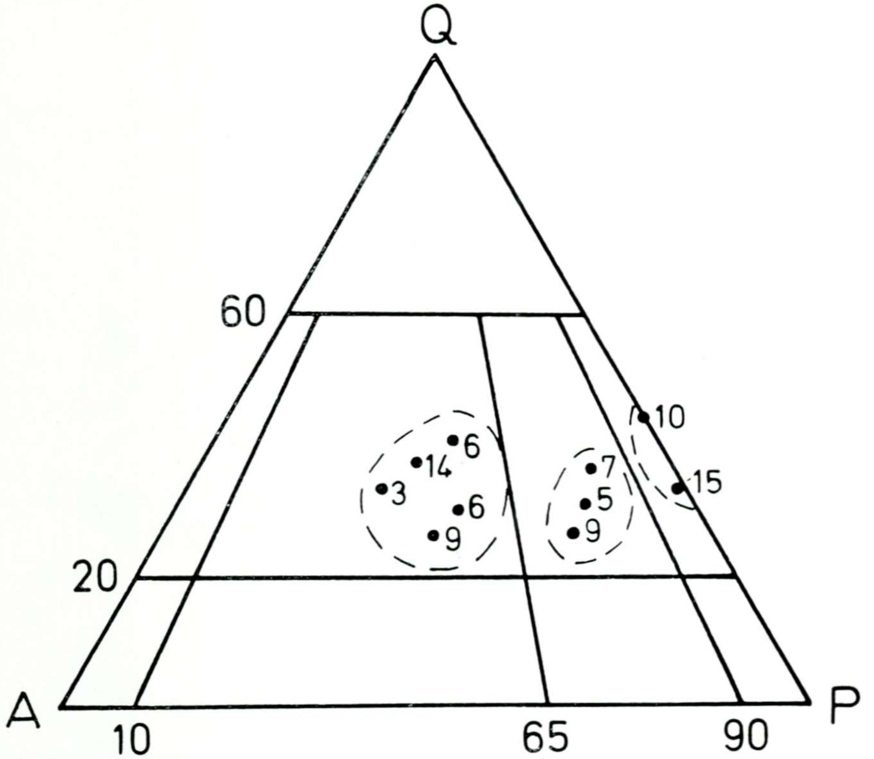


Fig. 9. Igneous rocks from the Nupsfonn Complex plotted on a Streckeisen (1967) classification diagram; based on modal analyses. Numbers indicate per cent biotite.

In contrast to the local occurrence of the Nupsfonn Complex, the Dyrskard Group is rather widespread. It has been mapped as a continuous unit far to the west (Naterstad et al. 1973), and the author has mapped it as far south as Sauda. The thickness is usually between 100–200 m, though locally up to 300–350 m, as at the type-locality at Dyrskard. Several phases of folding and thrusting are thought to be responsible for these variations.

The Dyrskard Group is composed only of supracrustals, both metasediments and metavolcanics. Meta-arkoses, quartzites and amphibolites are the dominant rock-types, but calcareous meta-arkoses, thin marble horizons, biotite amphibolites and garnet–quartz schists are also found. Intrusives have not been observed within the mapped area, but further west Naterstad et al. (1973) have reported metagabbroic rocks in the Dyrskard Group. None of the intrusives observed in the Nupsfonn Complex or the overlying Kvitenut Complex intrude the metasupracrustals of the Dyrskard Group.

The intense multiple folding and lack of good marker horizons make it difficult to establish a detailed litho-stratigraphic succession. However, from profiles and sections throughout the unit, a rough three-fold division seems logical. The subdivision described below is best demonstrated in the type-locality at Dyrskard.

A grey-green strongly foliated gneiss with thin layers of both feldspathic quartzite and pure quartzite dominates the lower part of the Dyrskard Group. The gneiss is often rich in small feldspar augen. In the upper part of this lowermost division two 3–5 m thick horizons of calcareous meta-arkoses are found. In contrast to the grey-green gneiss, which is restricted to Dyrskard, Middyrdalen and a few other places, the calcareous meta-arkose has been traced nearly continuously from Dyrskard to Sandfloeggi. In view of its position at the base of the Crystalline Nappe Rocks, it seems reasonable to interpret the grey-green gneiss as a mylonite gneiss. The high content of feldspar may indicate an original volcanic rock.

The middle part of the Dyrskard Group is composed of amphibolites in a characteristic alternation with quartzite layers. The thickness of the individual amphibolite and quartzite layers varies from less than one centimetre to several metres, and may even be as great as 50–100 m. In most sections through the Dyrskard Group quartzites and meta-arkoses dominate in this middle part, but on Dyrskardnut and south of Sandfloeggi amphibolites are the predominant rocks.

Orthoquartzites and feldspathic quartzites with a few horizons of black quartz-rich schists constitute the uppermost part of the Dyrskard Group. Thin layers of pure marble, never exceeding 50 cm in thickness, have been observed in the lower horizons of this division. A garnet–muscovite–quartz schist is frequently observed just below the mylonite gneiss of the overlying Kvitenut Complex, but well within the Dyrskard.

The following is a short petrographic description of the main rock-types within the Dyrskard Group.

Quartzites and meta-arkoses: These lithologies constitute at least 60% of the Dyrskard Group, and are found in all parts of the unit. A more or less pronounced compositional banding is typical for the quartzites and meta-arkoses, this being due to a variable content of mica, epidote and other minor minerals. The thickness of individual layers varies from a few millimetres to some metres, and both sharp and gradual transitions are common. The lithological banding is thought to be relict sedimentary bedding modified by later deformation (Fig. 10).

The quartzites and meta-arkoses are usually equigranular, fine-grained and without visible clastic grains. However, in some of the feldspar-rich layers, large, well-rounded, feldspar grains (1–5 mm) are present in an unsorted matrix. At first sight these rocks resemble poorly sorted arkoses or greywackes, but a closer inspection shows broken and sheared grains which indicate a cataclastic rather than an epiclastic origin for the clasts. The feldspar augen may, of course, represent original clastic grains, but their present shape is due to cataclasis.

Feldspar, quartz and mica all show a distinct mineral orientation. Quartz occurs as small recrystallized grains (0.01 mm) with undulating extinction. Various types of K-feldspar are observed. In the equigranular fine-grained

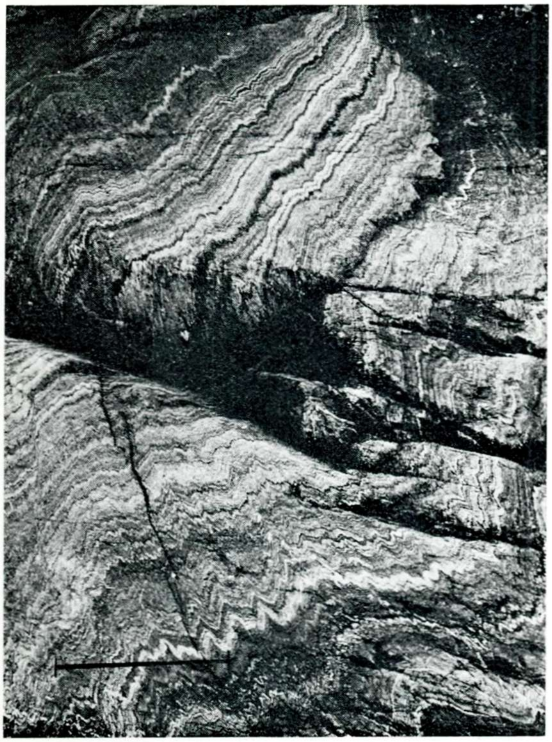


Fig. 10. Characteristic alternation of pelitic, semi-pelitic and psammitic layers from the Dyrskard Group. The folds are F_2 folds. Scale ½ m. Verjesteinsnuten.

layers microcline with or without twinning occurs together with quartz and (?) albite. The larger grains of K-feldspar (? epiclastic grains), often modified by cataclasis, show perthite patterns resembling those shown in Fig. 8. Refractive indices of the plagioclase component in these perthites indicate acid oligoclase. Plagioclase occurs both as large augen and as small recrystallized grains in the fine-grained matrix. Plagioclase in the matrix is untwinned albite, while the larger clasts are saussuritized, sericitized and clouded oligoclase with polysynthetic twinning after both the albite and pericline laws. Sphene is often found as large, deformed, clastic grains, while zircon is observed as small well-rounded grains.

Metamorphic minerals in addition to quartz and feldspars are muscovite, biotite, epidote and garnet. The muscovite occurs as large deformed flakes, usually concentrated in zones and showing a distinct parallel orientation. Biotite and garnets occur together in zones in slightly more pelitic meta-arkoses. The biotite occurs as small, weakly deformed flakes varying in colour from green-brown to dark brown, and is locally altered to chlorite. Garnet occurs as hypidioblastic, poikilitic grains (0.3 mm) with inclusions of quartz and white mica. Accessory minerals are tourmaline, calcite, epidote, orthite and opaques. Epidote is enriched in the garnet–biotite-rich zones.

Albite together with chlorite predominate in the grey-green mylonite gneisses low in the Dyrskard Group. In the calcareous meta-arkoses of the upper part of the lower division, actinolite and chlorite are common minerals together with calcite. Plagioclase in this rock is a basic oligoclase with An content up to 32%.

Garnet–muscovite–quartz schists: The garnet–muscovite–quartz schist observed in many places close to the Dyrskard Group/Kvitenut Complex boundary is of special petrographic interest, since the mineral assemblage garnet–muscovite–zoisite–kyanite is observed. This particular schist alternates with horizons of black quartz schist and pure orthoquartzite. The thickness seldom exceeds 2 m and the rock is characterized by garnet porphyroblasts ranging in size from 0.1 mm to 1 cm.

Quartz occurs as elongate, oriented grains with sutured grain boundaries. White mica seems to occur as two generations. First generation micas are large deformed flakes (4–5 mm) while those of the younger generation are undeformed or only weakly deformed small flakes (0.2 mm). The garnet porphyroblasts are hypidioblastic to idioblastic. Some of the large garnets have textures which are interpreted as representing two generations of growth. The core is poikilitic with the inclusions sometimes arranged in an S-shaped pattern, indicating syntectonic growth. The rim is non-poikilitic and is responsible for the idioblastic shape. The smaller garnet porphyroblasts are all non-poikilitic. Muscovite is flattened around the larger garnet porphyroblasts. Kyanite occurs as fibrous aggregates intergrown with the muscovite.

Zoisite is found as 0.1 mm idioblastic prisms. Optical data indicate pseudozoisite (Trøger 1959).

Accessory minerals are thin needles of amphibole (? hornblende), brown biotite, chlorite, rounded zircon and opaques.

Amphibolites and amphibole–biotite schists: Both the amphibolites and the amphibole–biotite schists occur as conformable layers in the metasediments (quartzites and meta-arkoses), thicknesses varying from a few millimetres to several metres. Locally, the thickness increases to more than 100 m, as in the Dyrskard area and around Sandflotjørn. Neither the amphibolites nor the amphibole–biotite schists can be followed over long distances. The amphibolites are more frequent than the amphibole–biotite schists.

The amphibolites are dark green rocks without distinct foliation. Partly parallel-oriented, large, amphibole porphyroblasts up to 1 cm in length are often present. Primary volcanic textures and structures have not been observed. The amphibolites have sharp boundaries against the surrounding rocks, and nowhere have the amphibolites been seen to cut through the metasediments.

The amphibole–biotite schists have biotite instead of amphibole as the dominant mafic mineral, producing a distinct planar structure. In contrast to the rather massive amphibolites, the amphibole–biotite schists often show a distinct banding. In some places a gradual transition into biotite- and quartz-rich metasediments can be observed. These two features, together with a high proportion of quartz in the amphibole–biotite schists, indicate that at least some of these rocks originated either as epiclastic or pyroclastic sediments.

Amphibole occurs as hypidioblastic, poikilitic prisms. The grain-size is usually less than 1 mm, except for the larger porphyroblasts already mentioned. The amphibole porphyroblasts show a weak parallel orientation. Pleochroism is: X = yellow-green, Y = green, and Z = olive green, and extinction angle Z/c = 17–19°, most probably indicative of common hornblende. Biotite is found as hypidioblastic (0.5 mm) flakes, sometimes showing a weak chloritization. Colour in the Z direction varies from green (amphibole–biotite schist) to brown (amphibolites. Both the amphiboles and the biotites are deformed by F_2 folds. Plagioclase (An_{20-30}) occurs as sericitized allotrioblastic grains. Quartz comprises up to 10–15% of the amphibole–biotite schists, but is not present in the amphibolites. Euhedral garnet porphyroblasts are observed in an amphibolite zone between lake 1486 and lake 1470. Other minerals found are sphene, epidote, apatite, sericite and opaques.

Since the amphibolites have nowhere been observed in transgressive relationship to the metasediments an extrusive origin is most likely. The massive, several metres thick amphibolites may very well be lava flows, but for the mm- to cm-thick layers a pyroclastic origin seems more reasonable.

Kvitenut Complex

The Kvitenut Complex lies above the Dyrskard Group, and represents the highest tectonostratigraphic unit within the investigated area. However, only 2–3 km west of Stavsnuten the Kvitenut Complex is overlain by the Revsegg Formation (Naterstad et al. 1973). The contact between these two units has been interpreted as a tectonized sedimentary contact by Naterstad et al. The Kvitenut Complex is thus thought to form the basement for the mica gneisses of the Revsegg Formation. The type locality is west of the present area (Naterstad et al. 1973).

Both field observations and thin-section studies have revealed that an augen gneiss zone at the base of the Kvitenut Complex represents a recrystallized mylonite (Fig. 11). The Kvitenut Complex is markedly heterogeneous, with supracrustals, migmatites and various types of gneisses and intrusives. A petrographic description of all these rock types is beyond the scope of this paper.

As seen on the map published by Naterstad et al. (1973), the Kvitenut Complex has a broad regional distribution; the present author has also mapped it still further to the south and west (Andresen, unpubl.).

Augen gneisses (blastomylonites): In most places there is a distinct change in lithology when crossing the Dyrskard Group/Kvitenut Complex boundary. As already mentioned the uppermost part of the Dyrskard Group is usually made up of pure quartzites with zones of black schists and garnet–muscovite–quartz schists, quite different from the augen gneisses which dominate the lowermost part of the Kvitenut Complex. Augen gneisses, however, are not restricted to the lowermost part of the Kvitenut Complex but are also found higher up in the unit.

The thicknesses of the basal augen gneiss zone varies from a few metres to

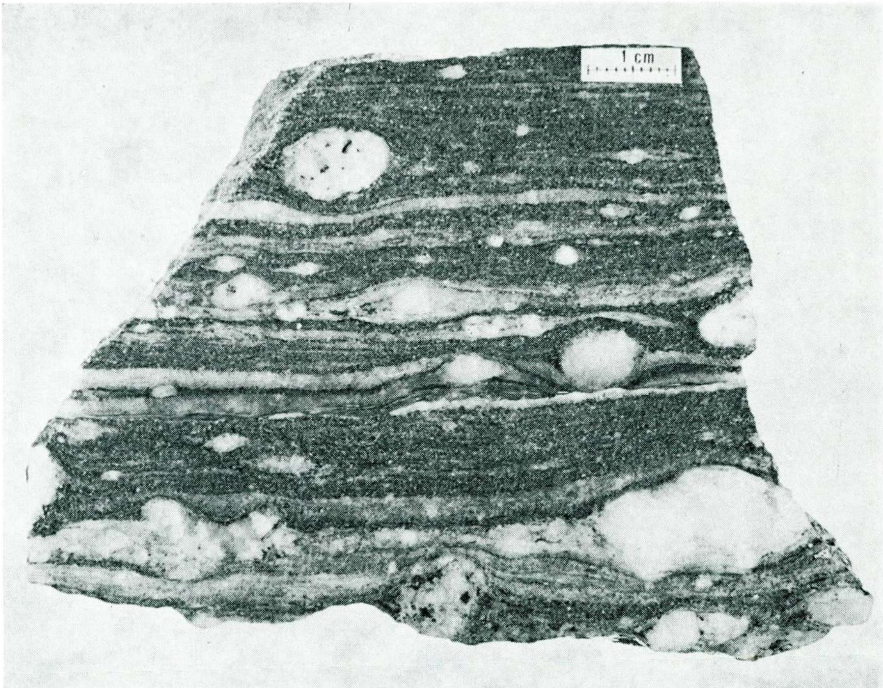


Fig. 11. Blasto-mylonite from the Dyrskard Group/Kvitenut Complex boundary. Stavsnuten.

several tens of metres. Both banded and massive augen gneisses are common. The feldspar augen, consisting of both plagioclase and K-feldspar, vary from a few millimetres to several centimetres. Feldspar-schlieren are often developed instead of feldspar-augen. The schlieren are orientated parallel to the gneiss foliation. Tectonic pseudo-conglomerates are also sometimes observed along the Dyrskard Group/Kvitenut Complex boundary.

The augen gneiss has both sharp and gradational contacts with the other rocks of the Kvitenut Complex. Sharp contacts are well demonstrated at Stavsnuten, where intricate migmatite structures grade into horizontally foliated augen gneisses within a distance of 10–20 cm. Large isolated boudins of this migmatite can be found locally within the augen gneiss. Around Sandfloeggi, no sharp contact between the augen gneiss and the overlying biotite gneiss is observed.

The augen gneiss is characterized by large feldspar porphyroblasts or porphyroclasts in granulated matrix, with K-feldspar dominating over plagioclase in the augen. The mineralogical composition of the augen gneiss varies from granitic to granodioritic, and in most cases biotite is the main mica. Thin-section studies indicate that at least some of the augen have been through a post-porphyroblastic mylonitization, as some of the augen consist of broken and granulated K-feldspar metacrysts (Fig. 12), where the feldspar fragments have the same optical orientation. The present mineral assemblage of plagioclase, microcline, white mica, biotite and epidote is post-mylonitization;

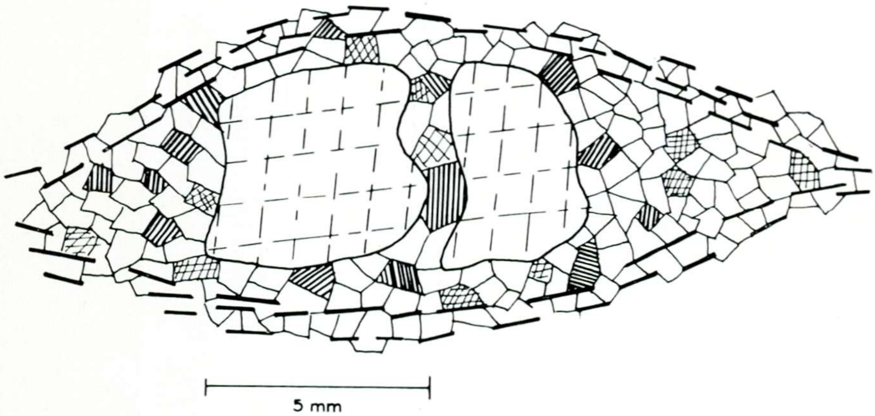


Fig. 12. Simplified drawing of an augen in the blasto-mylonite showing a large, broken, K-feldspar clast in a recrystallized matrix of quartz, microcline, plagioclase and mica.

these minerals have crystallized in between the K-feldspar clasts and are only weakly deformed, indicating a post-mylonitizational upper greenschist or lower almandine–amphibolite facies metamorphism. It is worth noting that, in the Nupsfonn Complex, the cataclasis is younger than the almandine–amphibolite facies metamorphism.

Banded gneiss: Above the augen gneiss in Verjesteinsnuten and Sandfloeggi are various types of banded gneiss. Distinct lithological banding and lack of large feldspar augen are characteristic of these gneisses, and serve to distinguish them from the augen gneiss. However, a distinct contact between the augen gneiss has not been observed. The thickness of the individual layers varies, but it is usually less than 0.5 m. The mineralogical composition varies from amphibolite and biotite–plagioclase gneiss to granitic gneiss.

Thin (0.5 m) granodioritic veins are sometimes observed cutting through the gneiss banding. These veins are folded.

High-grade metamorphic rocks of Stavsnuten: While augen gneiss and other quartzo–feldspathic gneisses dominate at Sandfloeggi and Varjesteinsnuten, quite different rocks are found on Stavsnuten. Above a 30–50 m-thick zone of augen gneiss occur high grade metamorphic rocks such as sillimanite-bearing gneiss, hornblende gneiss, calc–silicate gneiss and ultramafic rocks. Migmatites and intrusives are also found, especially in the lower part. The textures and structures of these rocks are quite different from those of the underlying augen gneiss (Fig. 13). There is a sharp boundary between the augen gneiss and the overlying migmatite, but isolated boudins of migmatite can be found locally within the augen gneiss. The size of the migmatite boudins varies, but the largest axis is usually less than 10 m. The migmatite boudins show internal structures which are discordant to the horizontal foliation of the augen gneiss which bends around them. The boudins probably represent mega-cataclast associated with the development of the augen gneiss.

Fig. 13. Intricate pre-F_0 (?) migmatite structures from the Kvitenut Complex, Stavsnuten.

Various types of granitic to quartz–dioritic gneisses with migmatites and intrusives are found just above the augen gneiss. These gneisses have a pronounced foliation defined by the preferred orientation of biotite and partly amphibole. The quartz–dioritic gneisses are equigranular, medium- to coarse-grained rocks. Plagioclase (50–60%) with An_{27-28} is the dominant mineral; this is weakly sericitized and shows albite and pericline twinning. Quartz (20–30%) occurs as allotrioblastic grains with undulating extinction. Biotite 12–15%) dominates over amphibole (6–7%). Biotite has pleochroism: X = pale yellow-green, Y = Z = green to olive-green. Small grains of sphene are found along the rims, indicating exsolution from a more Ti-rich biotite. The amphiboles are hypidioblastic prisms of green hornblende. In most of the amphiboles ore minerals, too fine-grained for identification, have been exsolved along certain crystallographic directions. K-feldspar (2–3%) is also present and apatite, epidote, zircon (rounded), sphene and magnetite are accessory minerals.

Foliated and banded granitic gneisses alternate with the quartz–dioritic types. The banding is due to a variable content of mafic minerals. The granitic gneisses are usually medium-grained, but sometimes contain coarse-grained layers. The major minerals are quartz (30%), K-feldspar, brown biotite (15–20%) and plagioclase (10%). Plagioclase is saussuritized oligoclase/andesine. Accessory minerals are sericite, clinozoisite, zircon (rounded) and ore-minerals.

These gneisses and migmatites grade upwards into amphibolites and leucoamphibolites, with layers of calc–silicate, biotite–amphibole gneiss, biotite–orthoclase gneiss and ultramafic rocks. These rocks are readily distinguishable from the surrounding gneisses on aerial photographs.

The amphibolites, which are the dominant rock-type of the high-grade metamorphic rocks of Stavsnuten, show compositional banding. Grain-size varies from fine-grained to coarse-grained. Amphibole porphyroblasts 1–2 cm long can sometimes be found. The amphibolites are rich in quartz–plagioclase schlieren, which have irregular shapes with diffuse transitions into the host rock. The grain-size of these schlieren, which were most probably formed by segregation processes during metamorphism (? partial melting), is the same as that of the surrounding rock.

Amphibole occurs as hypidioblastic poikilitic prisms with pleochroism: X = yellow-green, Y = green, Z = olive-green and Z/c = 18–20°. Plagioclase varies in composition from An_{28} to An_{38} with the highest An content in amphibolites bordering calc–silicates. The biotite is brown and shows a weak parallel-orientation, as do the amphibole porphyroblasts. Quartz occurs in variable amounts, and shows an increase towards the quartz–plagioclase schlieren. Other minerals observed are epidote (Fe-rich), apatite, zircon, ore-grains, chlorite and sericite.

Thin layers of medium-grained granitic gneiss are found in some places low down in the amphibolite sequence. Modal analysis showing a high percentage of quartz (40–50%) and the occurrence of large well-rounded zircons, indicate a sedimentary origin for this rock. In this rock an interesting mineral reaction seems to have taken place. Biotite, which is the dominant mafic mineral, is partly replaced by orthoclase, as illustrated in Fig. 14a. This may indicate that the following reaction has taken place: biotite → K-feldspar (+ ? garnet) which according to Turner (1968) is characteristic of the transition between upper almandine–amphibolite facies and hornblende–granulite facies.

Higher in the sequence, zones of fine-grained biotite–K-feldspar gneiss are often present. Plagioclase is usually present only in small amounts (Table 2); an exception to this is specimen B.O., but this is explained by its occurrence as a 10-cm thick horizon bordered on both sides by calc–silicates. Optical and X-ray data show that most of the K-feldspar is orthoclase.

In a medium-grained variety of the biotite–orthoclase gneiss, occurring close to the calc–silicates, sillimanite was identified; this is intergrown with muscovite. In a third type of orthoclase gneiss (a nearly biotite-free type) the following minerals were identified: orthoclase (55%), clinopyroxene (30%), quartz (10%), amphibole, biotite, apatite, sphene, zircon and rutile. The optical data for the clinopyroxene; $2V_z = 60°$, Z/c = 42–45° and olive-green to green pleochroism, indicates augite. The amphibole, which is a green hornblende, is always associated with clinopyroxene and is most probably secondary after clinopyroxene (uralite) (Fig. 14b).

The calc–silicates occur as a ca. 10-m thick zone composed of thin marble layers, diopside-fels, phlogophite–muscovite schist and other, more common, plagioclase–diopside–epidote rocks.

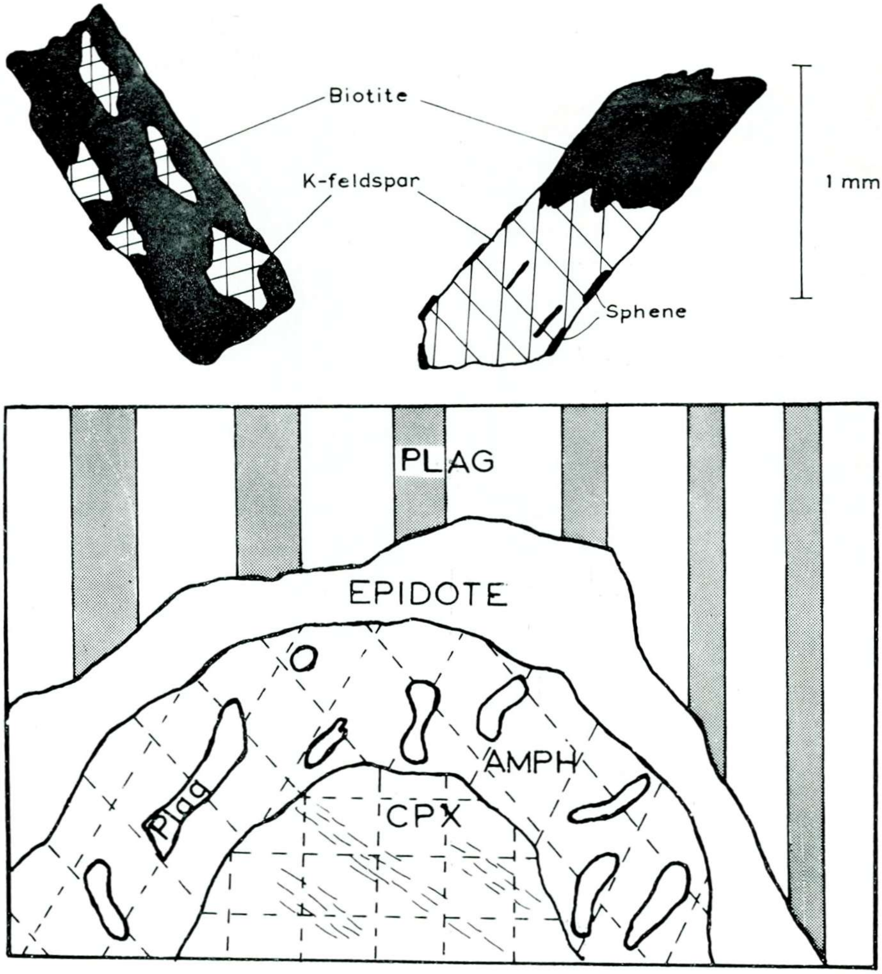


Fig. 14. Simplified drawings showing a) replacement of biotite by K-feldspar and b) a retrograde reaction between clinopyroxene and plagioclase in the calc-silicate gneisses. Kvitenut Complex, Stavsnuten.

Table 2. Modal analyses of biotite–K-feldspar gneisses from Stavsnuten (Kvitenut Complex)

Minerals / Samples	71.67	B.0	B.4	B.5
K-feldspar	58.4	44.0	65	52.5
Biotite	14.0	22.8	20	18.6
Quartz	24.2	18.8	9	18.0
Plagioclase	4.8	14.0	1	x
Apatite	x	x	x	x
Zircon	x	x	x	x
Chlorite	x	–	x	–
Epidote	–	x	x	11.0
Muscovite	–	x	–	8.1
Tourmaline	–	–	x	–
Ore minerals	x	x	x	x

x: accessory mineral

–: mineral not identified

A 5–6 m-thick layer of a brown-weathering ultramafic rock is observed in the upper part of Stavsnuten. It has sharp contacts against the surrounding gneiss, and can be followed as a conformable layer for several hundred metres. Characteristic of this rock are enstatite/bronzite porphyroblasts, up to 3 cm long, in a fine- to medium-grained granoblastic matrix. The following minerals have been identified, partly with the help of X-ray diffraction and U-stage studies: tremolite (40%), enstatite/bronzite (35–40%), forsterite (10%), pleonast (5%) and magnetite (5%). In microfractures serpentine minerals are observed as alteration products. The mineralogy thus indicates an extremely magnesium-rich rock.

According to Turner & Verhoogen (1960) this mineral assemblage is characteristic of magnesium-rich rocks metamorphosed in the upper part of the almandine–amphibolite facies (sillimanite–almandine–orthoclase subfacies) or higher metamorphic facies.

Structural geology

The structures described in this chapter are mainly those of the rocks involved in the Caledonian orogeny; very little attention has been given to the structures of the Precambrian basement.

To simplify the description and discussions, linear and planar structures which are thought to be of the same age are given the same subscript number in all the different tectonostratigraphic units. The structural symbols used and described within the different units are summarized in Table 3. The relative ages of the various structural elements have been determined mainly from observations of interference structures, but the rather constant orientation of linear structures and the uniform style of folding have also been used where interference structures are lacking.

PRECAMBRIAN BASEMENT

At least two phases of pre-Caledonian deformation can be recognized in the Precambrian basement. The writer assumes that the oldest deformational phase was a synmetamorphic folding event producing the augen gneisses and the observed foliation in the basement complex below the Telemark supracrustals. The second phase of deformation involved the folding of the Telemark supracrustals into a NNW–SSE trending synclinorium. Whether this second episode of deformation can be subdivided into two or three minor fold phases equivalent to those identified in the Telemark supracrustals further east (Dons 1960) has not been determined and is outside the scope of this work.

The two main phases of deformation are thought to be clearly separated in time, as the metamorphic conditions which dominated the first phase were not present in the second. Further evidence of a long hiatus here is provided by a conglomerate at the base of the Telemark supracrustals, which indicates uplift and erosion prior to their deposition.

With the exception of faulting, all Caledonian structures are restricted to

Table 3. Summary of tectonic events and structural features in the various tectonic units

	Precambrian (?) deformational episodes	Assumed Caledonian deformational episodes		
		Longitudinal struct.	Tranverse struct.	Longitudinal struct.
Kvitenut Complex	Pre-F_2 structures: Plastic folding and migmatic formation. Formation of mylonite gneisses.	F_1: Not identified.	F_2: Gentle to open folds. Axes trend NW–SE. S_2: Mainly vertical, but seldom developed as a planar structure.	F_4: Gentle to open large-scale folds. S_4: Vertical, but not developed as a planar structure.
Dyrskard Group	F_0: Isoclinal folds. Axes trend NW–SE. Development of a pronounced mineral lineation (L_0). S_0: Subhorizontal.	F_1: Close to tight folds. Axes trend WSW–ESE to E–W. S_1: Variable dip towards N and NNW.	F_2: Open to tight folds of variable scale. Axes trend NW–SE to NNW–SSE. S_2: Variable dip toward SW.	F_4: Open folds. Axes trend NE–SW. S_4: Developed only in the pelitic layers, dipping SE.
Nupsfonn Complex	Pre-F_2 structures: At least one generation of folding. Intrafolial folding in the gneiss. Migmatite formation.	F_1: Mesoscopic structures not identified. S_1(?): Cataclastic layering. Subhorizontal.	F_2: Open to tight folds of variable scale. Axes trend NW–SE. S_2: Variable dip towards SW, but seldom developed.	F_4: Gentle folds. Axes trend NE–SW.
Holmasjø Formation		F_1: Identification uncertain.	F_2: Close (quartzite) to isoclinal (phyllite). Axes trend NW–SE. S_2: Horizontal transposition foliation in the phyllite. Thrusting. F_3: Open folds. Axes trend NW–SE. S_3: Steeply SW dipping, fracture or crenulation cleavage.	F_4: Minor open to close asymmetric folds. Axes trend NE–SW. S_4: Crenulation cleavage dipping SE.
Autochthonous Sequence		F_1: Not identified.	F_2: Open to tight small- and large-scale folds. Axes trend NNW–SSE. S_2: Variably dipping towards SW. Thrusting.	F_4: Gentle to open folds. Axes trend NW–SE.

the uppermost part of the basement within the area investigated. The Caledonian influence is most intensely developed in the south and west, but even here Caledonian structures are seldom observed more than 5–15 m below the peneplain. The basement in these areas is strongly mylonitized in the upper part, and wedges from the basement are occasionally found thrust into the overlying sediments. These wedges are deformed by F_2 folds, indicating a tectonic emplacement prior to, or synchronous with, the F_2 event.

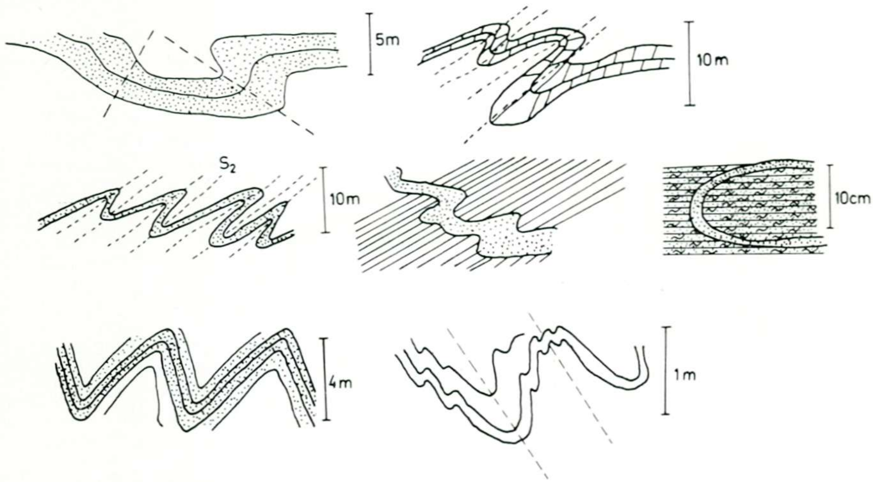


Fig. 15. Profiles of mesoscopic F_2 folds in the autochthonous Cambro–Ordovician sequence. All profiles are from quartzite layers except the upper right sketch which is of a chlorite marble bed.

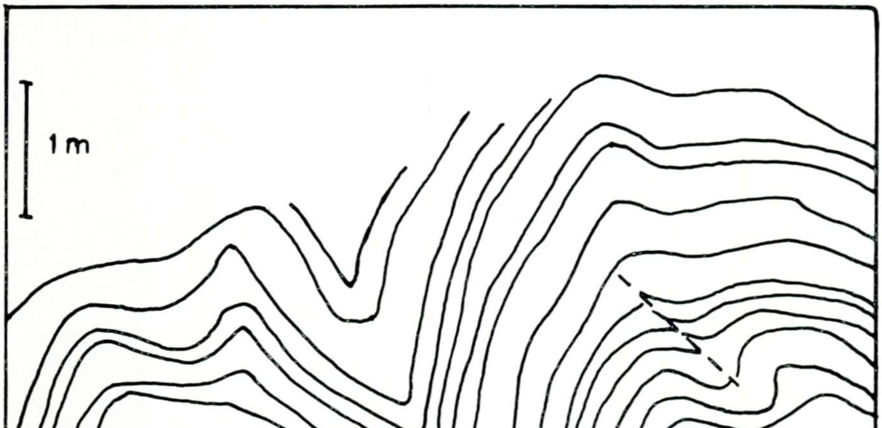


Fig. 16. A flexural-slip fold from the transition zone between black shale and orthoquartzite. Note the parasitic folds and the fault in the fold core. From a photo. East of Simletindvatn.

AUTOCHTHONOUS CAMBRO–ORDOVICIAN

Two episodes of deformation, F_2 and F_4, can be recognized in the autochthonous Cambro–Ordovician sequence as a folding of bedding (S_s) and/or bedding foliation (S'_s). The bedding is observed in virtually all lithological units except the black shale, and it is best developed in the transitional zone from black shale to orthoquartzite. At localities where this transition is present the alternation of pelitic, semi-pelitic and psammitic layers offers an ideal lithology for structural observations and analyses. A pronounced bedding foliation, defined by parallel orientation of muscovite and chlorite, is common in the chlorite marble.

F_2 deformation

The oldest folds in the autochthonous sequence are NNW–SSE-trending structures (F_2 folds). These folds show variations in scale, style and type. Structures vary from microscopic to mesoscopic, with a maximum amplitude of 50–60 m.

Most of the F_2 folds studied in detail are from the transition zone between the black shale and the orthoquartzite. Flexural-slip folds are the most common type, often with parasitic folds and with small faults in the fold core (Fig. 16), features typical of this type of fold (Whitten 1966). Folds developed in sequences rich in pelitic or semipelitic layers grade into slip folds or shear folds. At one locality asymmetric box-folds have been observed in the orthoquartzite (Fig. 15). Tight to close folds (Fleuty 1964) dominate the quartz schist, pelitic rocks and chlorite marbles. They show thickening of the folded layer along the hinges and attenuation along the limbs. Pronounced axial plane foliation and transposition foliation in the pelitic rocks indicate slip-folding in these lithologies. The F_2 folds developed in the autochthonous sequence thus provide good examples of how fold style and fold type are governed by the competence between the alternating layers, as has been shown experimentally by several authors (Ramberg 1961, Biot 1961, 1965). The F_2 folds are asymmetrical and, with a few exceptions, overturned towards the NE (Fig. 17c), indicating tectonic transport in this direction.

The well-exposed cliffs around lake Simletindvatn which trend at right-angles to the F_2 fold axes show that F_2 folds here occur along a décollement surface. The incompetent black shale has acted as a lubricating layer, on which thrusting and flow have taken place.

L_2 linear structures other than fold axes are present as a small-scale folding or crenulation of micas, and as an intersection lineation between S_s and S_2 in the pelitic and semipelitic rocks. Of these two types of lineation, both of which are parallel to the F_2 fold axes, micro-folding is the most frequently seen. The trend of these L_2 lineations and the F_2 fold axes is fairly constant, but a great variation in plunge is caused by the later F_4 folds.

The S_2 planar structures are developed parallel or sub-parallel to the axial surface of the F_2 folds. In the semi-pelitic layers they occur as a crenulation cleavage with muscovite bent parallel to S_2. These planar structures are sometimes developed as a fracture cleavage if the content of psammitic material is

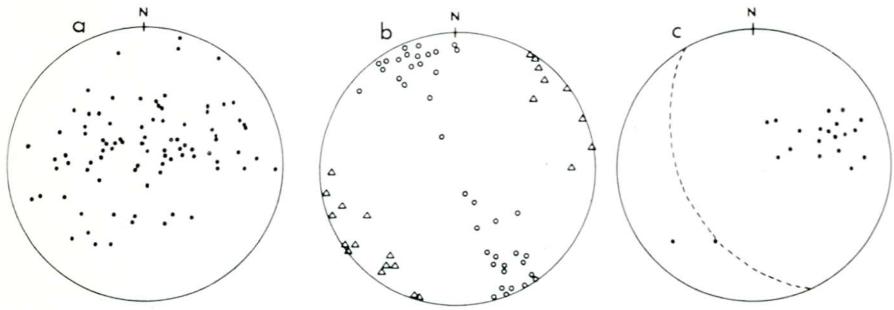


Fig. 17. Stereographic projection (Schmidt net, lower hemisprere) of structural data from the autochthonous Cambro-Ordovician sequence.
a) dots – poles to S_0 (bedding)
b) circles – F_2 fold axes and L_2 lineations; triangles – F_4 fold axes
c) dots – poles to axial planes of F_2 folds

high, but they have never been observed in the orthoquartzite. In the black shale they are developed as a more or less horizontal transposition foliation with carbonaceous matter and clastic muscovite concentrated along the foliation plane.

The S_2 foliation has a fairly constant NW–SE strike and always dips to the southwest (Fig. 17c). The actual angle of dip is variable often depending on the character of the lithology.

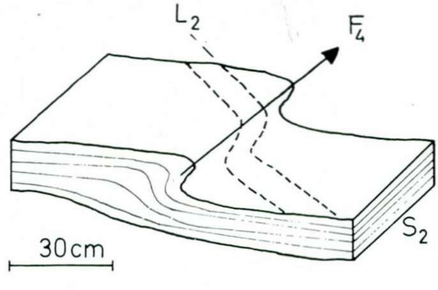
F_4 deformation

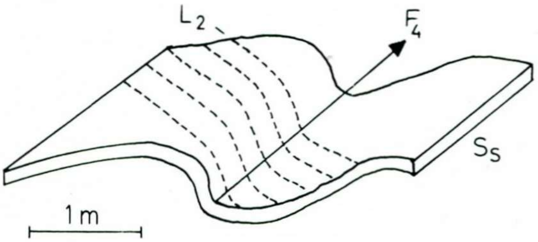
F_3 structures, as they are described in the Holmasjø Formation, have not been identified in the autochthonous sequence; F_4 structures are the only structures overprinting the F_2 linear and planar structures in this unit.

F_4 folds are open, undulating folds trending approximately at right-angles to the F_2 folds. In most places where the F_4 folds are developed they refold F_2 folds and L_2 lineations (Fig. 18). The F_4 fold axes show a much greater variation in trend than the F_2 folds, but there is no way of knowing if this variation is caused by a regional variation in the orientation of stress axes during deformation, or if it simply reflects the variable orientation of S_s prior to the F_4 deformation (Fig. 17b). While wavelengths and amplitudes are approximately equal in magnitude in the F_2 folds, the wavelength is often 5 to 10 times the amplitude of large-scale F_4 folds. The wavelength is often several hundred metres. In contrast to the asymmetrical and overturned F_2 folds, the large-scale F_4 folds are upright and symmetrical. Small-scale F_4 folds in the pelitic and semi-pelitic rocks are, however, asymmetrical and overturned towards the NW.

Planar structures (S_4) associated with the F_4 deformation are not very frequent. However, where developed they transect all other structures. With a fairly constant southeasterly dip of about 40–50ᵍ, they serve to distinguish the S_4 foliation from the older planar structures. Like the S_2 foliation the S_4 foliation is developed both as a crenulation cleavage and as a fracture cleavage, depending on lithology. The S_4 foliation is developed as an axial cleavage in

Fig. 18. Sketches showing the relationship between S_S, L_2 lineation and F_4 folds. East of Simletindvatn.





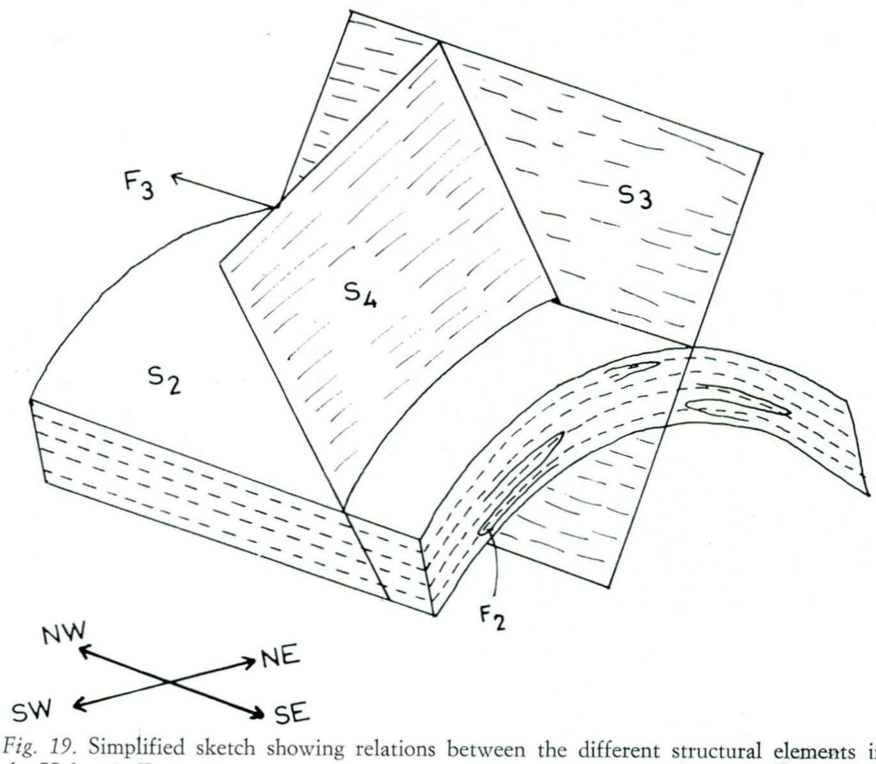


Fig. 19. Simplified sketch showing relations between the different structural elements in the Holmasjø Formation.

small-scale F_4 folds but is not related to the axial plane of the upright large-scale open F_4 folds.

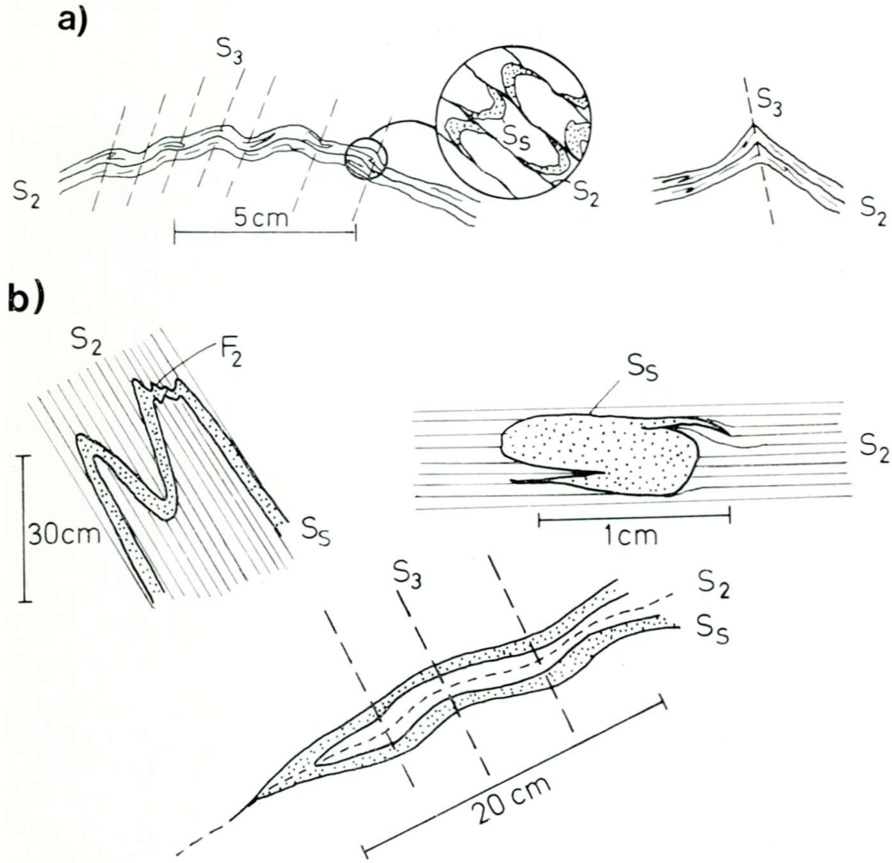


Fig. 20. Fold profiles of F_2 and F_3 folds and their mutual relationship in a) pelitic layers, b) psammitic layers (dotted) from the Holmasjø Formation.

HOLMASJØ FORMATION

From observation of interference structures three episodes of deformation (F_2, F_3 and F_4) have been identified in the Holmasjø Formation (Fig. 19). In most localities, therefore, two or three sets of secondary planar structures are developed. The original bedding, S_s, which is defined by alternating quartz-rich and mica-rich layers, usually less than 2 cm thick, has been strongly modified by the secondary planar structures, and its orientation is in most places impossible to obtain.

F_2 deformation

The oldest set of folds (F_2) developed in the phyllitic quartz schists dominating the Holmasjø Formation are isoclinal folds with flat-lying axial planes (S_2). The S_2 foliation is developed as penetrative transposition foliation sub-parallel to the strongly modified bedding, and it can in fact be easily misidentified as primary bedding. Close observation, however, reveals its secondary origin: muscovite is bent parallel to and is concentrated along the plane of foliation, and locally it has recrystallized along this plane. As S_2 is often developed as transposition foliation, the F_2 folds occur as rootless and intrafolial folds (Fig. 20a). Large scale recumbent folds associated with these structures have not been observed. F_2 fold axes from the phyllitic quartz schist have a variable orientation (Fig. 21a), a feature which is discussed below.

F_2 folds developed in the more competent layers of psammite are open to close and are overturned towards the NE. The NW–SE trending folds are present both in the tectonically included wedges of the underlying ortho-quartzite and in the autochthon beds of meta-arkose.

The quartz segregations which are so typical of the Holmasjø Formation are also deformed by the F_2 folds, indicating pre- or syn-F_2 hydrothermal activity. A striation parallel to the isoclinal F_2 fold axes in the quartz segregations is commonly observed. Pronounced and measurable lineations are, however, seldom found elsewhere in the Holmasjø Formation due to the development and intersection of up to three sets of planar structures.

Although the F_2 fold axes in the Holmasjø Formation show a wide distribution they tend towards concentration in a NNW–SSE direction (Fig. 21a). Since all the measured fold axes in the psammite layers trend NNW–SSE and since these folds are only slightly disturbed by later folding, this trend has been taken to be the direction of the tectonic B-axis during the F_2 folding. As no regional trend seems to be responsible for the greater spread of F_2 folds in the phyllitic quartz schists other possibilities have to be explored. The three most reasonable explanations are listed below:

a) The measured fold axes (Fig. 21a) represent two-fold episodes (? F_1 and F_2)

b) The spread is caused by inhomogeneous deformation (Borradaile 1972)

c) The variation or spread in F_2 fold axes is caused by refolding by F_3 folds. If there is a small angle between the trend of F_2 and the overprinting F_3 folding is of a flexural-slip type, then the F_2 fold axes will be distributed around a small circle with F_3 as the axis.

Detailed work, now in progress, is necessary to decide which of these possibilities is the most likely one.

F_3 deformation

In some localities a second set of NNW–SSE trending folds and lineations is developed in the phyllitic rocks of the Holmasjø Formation. These folds (F_3) are coaxial or nearly coaxial with the isoclinal F_2 folds which they are clearly refolding; thus a lower limit can be put on their age. The F_3 folds are transected by the S_4 crenulation cleavage.

The F_3 folds are open and usually slightly overturned toward the NE. Amplitudes and wavelengths are of a similar magnitude, varying in size from a few millimetres to several metres (Fig. 20). In pellite-rich layers F_3 structures are sometimes developed as kink folds. The axial surface is developed either as a crenulation cleavage or as a fracture cleavage. The L_3 lineation is the most frequently observed linear structure in the Holmasjø Formation. It is developed both as an intersection lineation (S_2 and S_3) and as a micro-folding. Intersection

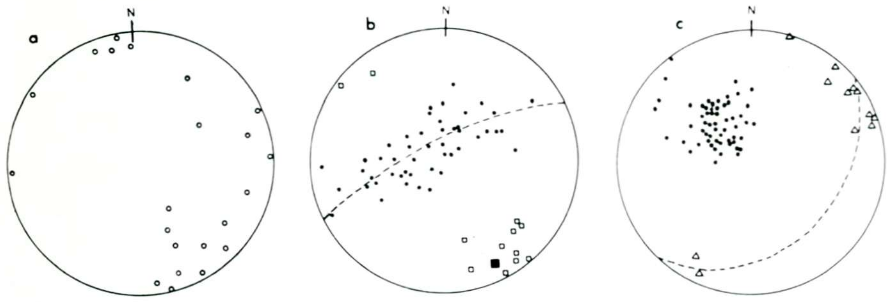


Fig. 21. Stereographic projection (Schmidt net, lower hemisphere) of structural data in the Holmasjø Formation.
a) Circles – F_2 (and ? F_1) folds
b) dots – poles to S_2; open squares – F_3 folds solid square – F_3 girdle axis
c) dots – poles to S_4 crenulation cleavage; triangles – F_4 fold axes; dashed line-average orientation of S_4.

lineations are usually found in the quartz-rich rocks; micro-folding usually occurs in the phyllites.

F_4 deformation

The dominant structure related to this episode of deformation is an intense crenulation cleavage (S_4). As in the autochthonous sequence, S_4 has a fairly constant orientation throughout the whole area, striking NE with a moderate dip towards the SE (Fig 21c), a phenomenon recorded previously by Reusch et al. (1902) for most of the grey phyllites on Hardangervidda. Muscovite is bent parallel to the plane of foliation. Chlorite is also oriented parallel to this plane. Mesoscopic F_4 folds associated with the S_4 crenulation cleavage are seldom observed. Small-scale F_4 folds with amplitudes less than 2–3 cm are frequently found, however, where one has exposures perpendicular to the fold axes and S_4 foliation (Fig. 5). These folds are asymmetrical, with the S_4 parallel to the axial planes; this indicates a relative movement of the overlying rocks towards the NW as already observed in the pelitic and semi-pelitic rocks of the autochthonous sequence. This conclusion is in agreement with the interpretation given by Kvale (1960) of the youngest episode of deformation from the Finse–Hallingskeid district.

NUPSFONN COMPLEX

The structural histories of the Nupsfonn Complex, the Dyrskard Group and the Kvitenut Complex are much more complex than those of the two units already described. One finds several planar and linear structures here which were developed prior to the oldest set of structures (F_2 and S_2) observed in the underlying units. In the Nupsfonn Complex (and the Kvitenut Complex) these structures have been grouped together as pre-F_2 structures. A direct correlation with the two pre-F_2 deformational episodes (F_0 and F_1) established for the Dyrskard Group is impossible at the present stage.

Pre-F_2 deformation

Both in the banded gneisses and in the metasediments low down in the Nupsfonn Complex pre-F_2 intrafolial, isoclinal folds can be observed. In the few places where it is possible to measure them, these pre-F_2 linear structures seem to be parallel or sub-parallel to the axes of F_2 folds. However, the style of folding and refolding of these folds by F_2 folds clearly indicate that there were two separate fold episodes.

Muscovite, biotite, amphibole and feldspar are oriented parallel to the lithological boundaries (? S_s), and they are assumed to have developed during or after the isoclinal folding. Both the intrafolial, isoclinal folds and the dis harmonic folds in the migmatites developed prior to the cataclastic layering. The cataclastic layering is most characteristically developed in massive igneous rocks, but it is also observed in the gneisses deforming the feldspar and micas which define the foliation in these rocks. This puts a lower limit on the age of the cataclasis. Similarly, deformation of the cataclastic layering by F_2 folds places an upper age limit on this event. These facts suggest that the cataclasis may be connected with F_1 folding observed in the Dyrskard Group (p. 44), which is interpreted as a fold episode associated with the major thrusting of the nappes from the central zone of the orogen southeastward onto the Baltic Shield. An alternative explanation is that the cataclastic layering was formed during an early stage of the F_2 deformation, and that it has been folded during the main F_2 stage. It is worth mentioning here that large-scale thrusting has also taken place in the F_2 deformational episode, and this could very well be responsible for the cataclasis.

The poles to the pre-F_2 planar structures in the Nupsfonn Complex show a great circle distribution defining a NE-trending b-axis which coincides with the observed F_2 axes (Fig. 23a). The only exceptions to this pattern are found in the northwestern part of the Nupsfonn Complex (around Nupstjørn) where the poles to the pre-F_2 structures show a much greater spread (Fig. 23b). This feature is probably due to a stronger influence of F_4 folding, a fact also indicated by the distribution of the F_2 fold axes and the frequency of F_4 folds (Fig. 23d).

F_2 deformation

NW- to NNW-trending folds (F_2) and associated lineations (L_2) are the most frequently observed linear structures in the Nupsfonn Complex, and these are easily seen on aerial photographs. Fold style varies from open to tight, and the amplitude from a few millimetres to several tens of metres; the largest fold is the overturned antiform east of Nupsdalen (Fig. 22). The F_2 folds show vergence towards both the NE and the SW, although folds overturned towards the NE are the most common. Some of the observed F_2 folds have curved axial surfaces, probably indicating an influence of the F_3 deformation, as in the Holmasjø Formation.

The F_2 linear structures have a fairly constant NW to NNW trend and a general low angle of plunge towards the NW. The influence of F_4 folding is

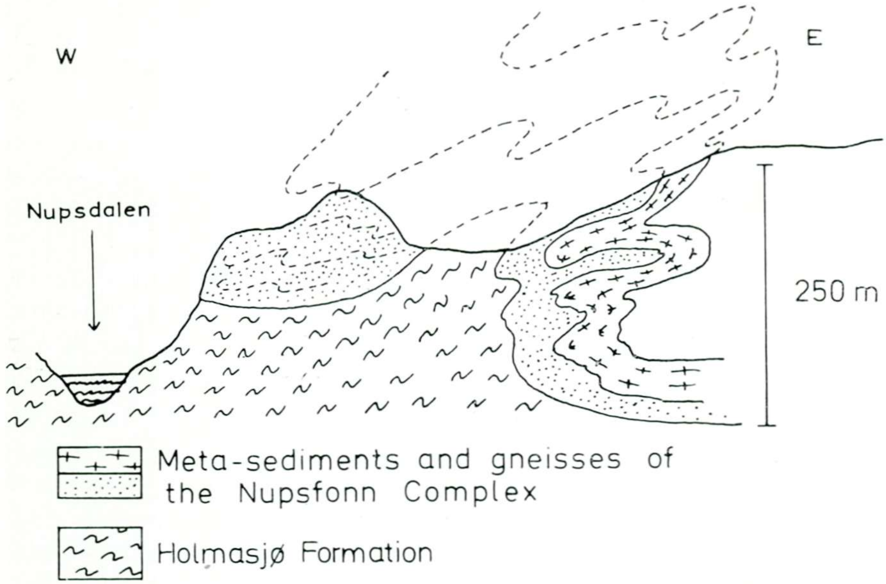


Fig. 22. Profile from Nupsdalen eastward towards Nupsfonn showing a large overturned F_2 fold with associated parasitic folds. (After photo). The Holmasjø Formation is folded into and partly above the Nupsfonn Complex.

usually weak (Fig. 23c) except in the area northeast of Nupsridet, where F_2 linear structures show a great circle distribution with a b-axis coinciding with the measured F_4 fold axes.

Axial surface cleavages or foliations associated with the F_2 folding are not present in the banded gneisses of the Nupsfonn Complex, but a locally developed crenulation cleavage is present in the metasediments at the base of this unit.

F_3 deformation

F_3 folds have not been observed with certainty in the Nupsfonn Complex. However, the curved F_2 axial surfaces and the NW trending undulating folds deforming the thrust plane below the Nupsfonn Complex may represent F_3 deformation in this unit.

F_4 deformation

The youngest folds observed in the Nupsfonn Complex are open, NE–SW-trending folds plunging at low angles towards the NE. Most folds are symmetrical with vertical axial planes, and they do not show the overturning towards the NW which is so common for F_4 folds in the Holmasjø Formation. Planar structures, except faults, related to this episode of deformation are not developed.

DYRSKARD GROUP

The Dyrskard Group, with its distinct lithological variations and preserved bedding (S_s), is the unit within the crystalline part of the Hardangervidda–

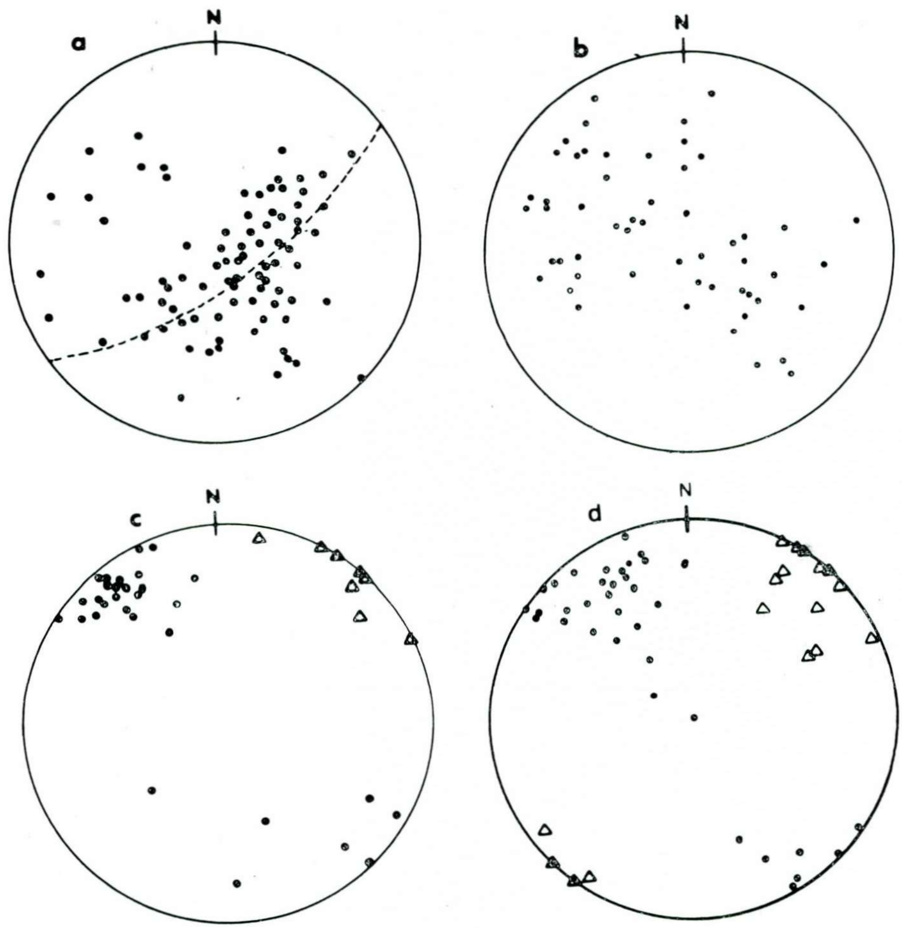


Fig. 23. Stereographic projection (Schmidt net, lower hemisphere) of structural data from the Nupsfonn Complex.

a) Poles to pre-F_2 planar structures (except for the area around Nupstjern) showing an ill-defined girdle indicating a b-axis coinciding with the F_2 fold axes.

b) Poles to pre-F_2 planar structures from the area around Nupstjern. The strong influence of F_4 folding together with F_2 folding are responsible for the great spread.

c) Fold axes from the Nupsfonn Complex (except for the area around Nupstjørn). Circles – F_2 fold axes; triangles F_4 fold axes.

d) Fold axes from the area around Nupstjørn. F_2 fold axes show a greater spread due to a stronger F_4 influence.

Ryfylke Nappe System, which is best suited for structural investigations. In this unit two episodes of folding (F_0 and F_1) are identified prior to the F_2 deformation.

F_0 deformation

The earliest folds to be recognized in the Dyrskard Group are isoclinal, flat-lying structures (F_0) (Fig. 25). Amplitude varies from a few centimetres to several tens of metres in large recumbent folds. Since the F_0 folds occur in massive meta-arkoses and quartzites their trend is usually difficult to determine

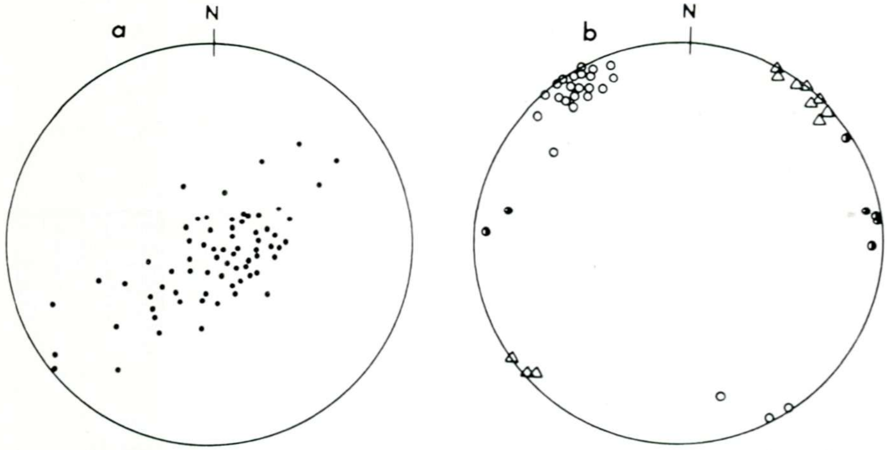


Fig. 24. Stereographic plot of a) poles to S_S (S_0) and b) fold axes in the Dyrskard Group. Semi-open circles – F_1 fol daxes, open circles – F_2 fold axes, triangles – F_4 fold axes.

directly. During the F_0 deformational episode, however, a pronounced mineral lineation was developed parallel to the F_0 axes, and this provides some information about the trend of the F_0 folds. The general trend of this mineral lineation, L_0, is NW–SE and thus parallel or sub-parallel to the trend of F_2 fold axes. When isoclinal structures are developed, however, this lineation is clearly seen to be related to the F_0 folds and not to the F_2 folds. Observation of non-parallelism between the L_0 lineation and the F_2 fold axes confirms this conclusion. Where the lineation is oblique to the F_2 fold axes, it always trends more westerly than the F_2 folds; the angle of separation between the two trends is nowhere greater than 35^g. A secondary planar structure, S_0, was developed, associated with the formation of the isoclinal F_0 folds. Owing to the isoclinal nature of the F_0 folds, S_0 is parallel or sub-parallel to the bedding (S_S), except in fold closures where the muscovite is oriented parallel to the axial surfaces.

F_1 deformation

In incompetent layers such as marbles and biotite–amphibole rocks, or where quartzite layers are surrounded by phyllite, ENE- to E-trending folds (F_1) are locally developed (Fig. 24a). These folds are seen to deform F_0 structures (Stavsnuten area), but they are in turn deformed by F_2 structures. This confirms the existence of two fold episodes prior to the F_2 deformation. The F_1 folds are minor folds with amplitudes of less than 50 cm. Fold style varies from close in the psammites to tight in the pelitic and calcareous layers. The axial surfaces dip NNE to N, but are seldom developed as a secondary foliation. From their orientation, paralleling the main Caledonian structures in this region, and their SSE vergence, the F_1 folds are interpreted as structures formed during the main thrusting of the nappes from the central zone of the orogen onto the Baltic Shield.

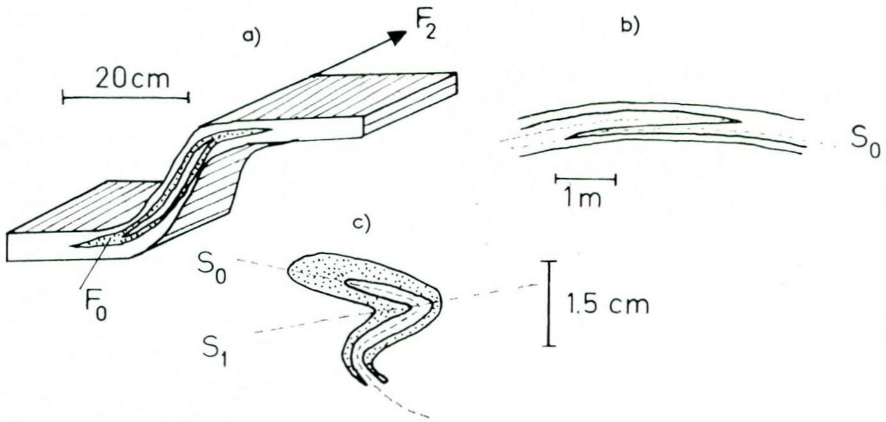


Fig. 25. Fold profiles from the Dyrskard Group.
a) An eyed F_0 fold refolded by an F_2 fold,
b) an isoclinal F_0 fold refolded by an open F_2 fold,
c) an isolated F_0 fold (quartzite) which is refolded by an F_1 fold.

F_2 deformation

The dominating structures within the Dyrskard Group, are the NW to NNW trending F_2 structures. The scale of folding varies from a few millimetres to several tens of metres, and isoclinal F_0 folds are often observed along the limbs of major F_2 folds. The F_2 folds vary in style from open to tight (Fig. 25), and almost everywhere are overturned towards the northeast; monoclinal folds are also observed. The thrust zone below the Dyrskard Group is also deformed by F_2 structures. The axial surfaces (S_2) of F_2 folds are developed only in the pelitic layers as a foliation surface.

The stereographic plot of poles to the S_0 and S_s planar structures shows a great circle distribution with a b-axis which coincides with the observed F_2 fold axes (Fig. 24). This corroborates the observations made nearly everywhere in the field that the NW–SE-trending F_2 structures are the most prominent ones.

F_4 deformation

Gentle to open folds (F_4), trending NW–SE and clearly refolding F_0, F_1 and F_2 structures are the only deformation episode identified as post-F_2. The relationship between F_4 and older structures is well demonstrated on the northeastern slopes of Slettedalen, where gentle F_4 folds with wavelengths of 200–300 m deform all other structures (Fig. 26). The F_4 folds generally have horizontal axes and vertical axial surfaces. Only the minor F_4 folds in the pelitic layers near the base of the Dyrskard Group show overturning towards NW. An S_4 crenulation cleavage is locally developed in these layers.

The gentle nature of the F_4 folds is seen in the fairly constant orientation of F_2 fold axes (Fig. 24b).

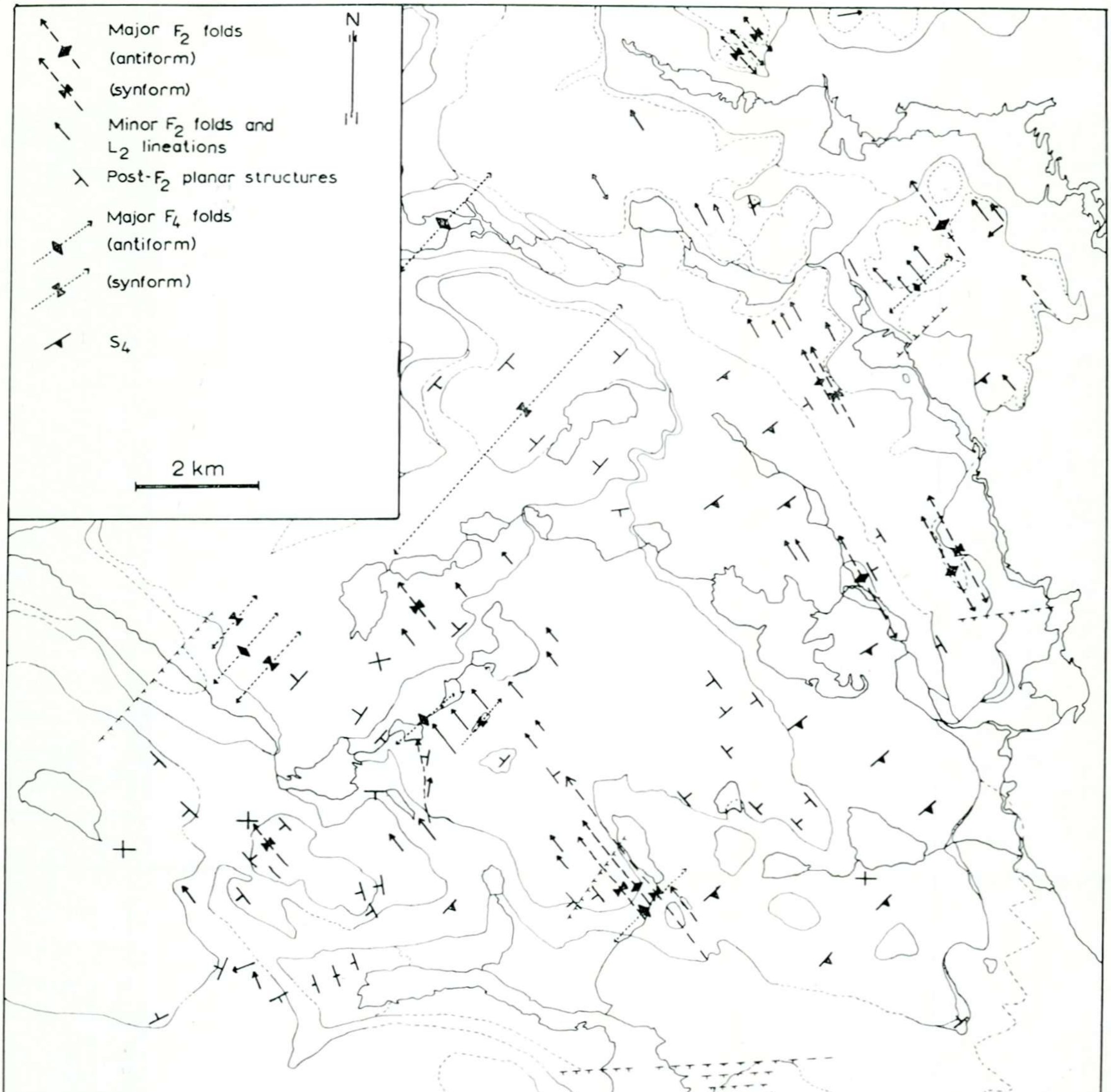


Fig. 26. Structural map of the area north of Haukelisæter showing the trend of major F_2 and F_4 folds, and the orientation of dominant planar structures.

KVITENUT COMPLEX

Pre-F_2 deformation

Most of the linear and planar structures observed in the Kvitenut Complex, including those in the high-grade metamorphic supracrustals at Stavsnuten, the migmatite structures, the structures of the augen gneiss and the gneiss banding observed elsewhere in the complex, are of pre-F_2 age. The foliation and linear structures observed in the augen gneiss are probably the youngest pre-F_2 structures, but the age relation between the different types of structures has not yet been studied; a detailed investigation is now being carried out by a graduate student (Roy Gabrielsen) at the University of Oslo. From the Rb/Sr whole rock ages obtained on mylonite gneisses from the Stavanger area, 1140 m.y. (Heier et al. 1972) and from granitic gneisses below Hardangerjøkulen, 1643 m.y. (Priem 1968) a Precambrian age seems likely for the structures developed prior the F_2 deformation.

F_2–F_4 deformation

Mesoscopic structures related to the last three episodes of deformation F_2–F_4 are few. The only observed folds are open NW–SE-trending structures clearly refolding all other structural elements such as the gneiss banding and the augen gneiss foliation. The Kvitenut Complex around Verjesteinsnuten is situated in a F_2 synform (Fig. 26).

F_3 structures have not been identified. F_4 structures occur only as very gentle, non-measurable megascopic folds; they are probably responsible for the variable plunge of the F_2 fold axes.

COMPARISON OF STRUCTURES FROM THE DIFFERENT TECTONIC UNITS

Two separate fold episodes, F_2 and F_4, can be distinguished in the autochthonous Cambro–Ordovician sequence, trending NNW–SSE and NE–SW, respectively. The same two fold episodes can also be identified in the tectonically overlying unit, the Holmasjø Formation. In addition to these two main fold episodes another phase of folding, F_3, is observed locally in the Holmasjø Formation. The F_3 folds are developed post-F_2 and pre-F_4, but they do not occur in the autochthonous sequence; an explanation for this absence may lie in the difference in competence between the phyllitic quartz schist of the Holmasjø Formation and the quartzites which dominate the autochthonous sequence. During the F_3 deformation (a late phase of the F_2 event?) the stress built up was sufficient to initiate folding of the S_2 foliation in the incompetent phyllitic quartz schist but not in the more competent quartzites of the autochthonous sequence.

Even though the autochthonous sequence and the Holmasjø Formation seem to have had different deformational histories, the differences are thought to be nowhere near as great as those observed between the Holmasjø Formation and the overlying units. Evidence in the latter of at least one pre-F_2 deformational episode in addition to the F_2 and F_3 fold episodes, indicates a structural break above the Holmasjø Formation. As F_2 linear and planar structures are the first structures to be identified with certainty in the autochthonous sequence and the Holmasjø Formation, the appearance of pre-F_2 structures in the overlying crystalline units is strong evidence favouring an allochthonous position for the latter.

If all pre-F_2 structures are accepted to have formed outside the Haukelisæter area, the next questions to be answered are: where have these structures been developed, and what is their age? Are they Precambrian or Caledonian structures? These questions cannot be answered with certainty now, but geochronological work at present in progress will possibly shed some light on the problem. In the meantime some suggestions can be made, and these are outlined below.

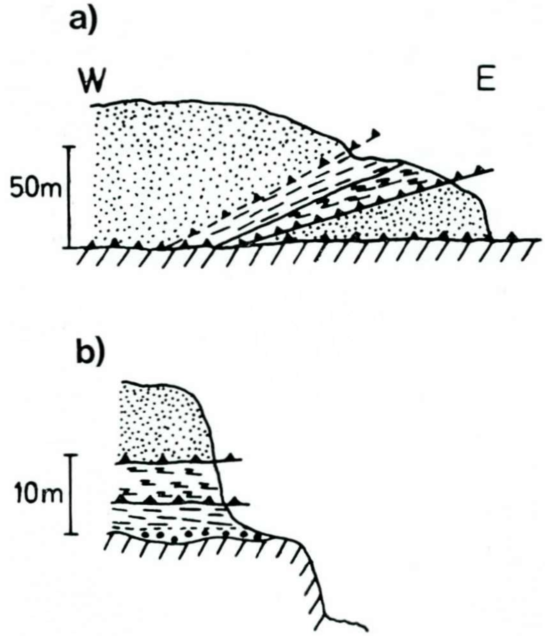
The structures (and metamorphism) observed in the high-grade metamorphic rocks at Stavsnuten are mostly Precambrian in age, as mentioned by Naterstad et al. (1973). Greater uncertainties arise, however, when one tries to fix an age for the augen gneiss formation at the base of the Kvitenut Complex, and for the F_0 and F_1 structures observed in the Dyrskard Group. The trend and south-easterly overturning of the F_1 folds, which are restricted to the incompetent layers in the Dyrskard Group, indicate that these folds *may* have been formed during the main thrusting of the nappes from the central zone of the orogen south-southwestwards onto the Baltic Shield. The brittle cataclasis with formation of cataclastic layering in the Nupsfonn Complex may have been formed during this oldest episode. If one accepts this deformational episode to be responsible for the development of first Caledonian structures in the area, then the isoclinal F_0 folding in the Dyrskard Group is almost certainly of pre-Caledonian (Precambrian) age. The augen gneiss (mylonite gneiss) may have been formed in connection with the intense (?) pre-Caledonian F_0 folding. A Precambrian age is also most likely for the intrafolial pre-F_2 folds observed in the gneisses of both the Nupsfonn Complex and the Kvitenut Complex. These structures may be of F_0 age, but a direct correlation with those in the Dyrskard Group is not possible since they belong to three separate tectono-stratigraphic units. A detailed structural analysis of the three units would probably help to solve this problem.

The model presented above accepts the first main Caledonian tectonic episode (F_1) as a thrusting of the nappes from NNW to SSE in agreement with Kvale (1960). If this is the case, however, one would expect to find structures of this episode in the autochthonous sequence and in the Holmasjø Formation. So far, definite structures of this age have not been observed in these particular units, although it is not impossible that some of the F_2 folds observed in the Holmasjø Formation may represent F_1 structures. The only explanation that seems reasonable to the author is that in these rocks pre-existing F_1 structures have been completely destroyad or masked by the later deformational episodes, especially that of F_2.

THRUSTING

In earlier descriptions from Hardangervidda (Reusch et al. 1902) only one major thrust plane was reported and mapped. This thrust plane was placed at the base of what is now referred to as the crystalline part of the Hardangervidda–Ryfylke Nappe System. All rocks below this thrust plane, including the Precambrian basement, the Holmasjø Formation and the autochthonous Cambro–Ordovician sequence, were assumed to be autochthonous. The rocks above this thrust plane were treated as a single allochthonous unit. Recent investigations by Naterstad et al. (1973), however, have shown that several thrust units and thrust planes can be distinguished in the Haukelisæter–Røldal

Fig. 27. Section through imbrication structures
a) south of Årmoteggi, b) east of Årmoteggi. Dots – orthoquartzite; z-mylonitized basement rocks; dashed – black shale; circles – basal conglomerate.



area. Naterstad et al. did not discuss the structural history and the relation between folding and thrusting, apart from stating that thrusting between the different tectonic units must have taken place in both Precambrian and Caledonian times.

Five major thrust planes (T–1, T–2, T–3, T–4 and T–5) and several minor ones, with local imbrications, have been identified. There are some minor thrust zones in the Precambrian basement, but the lowermost major thrust plane (T–1) is found below the orthoquartzite of the autochthonous Cambro–Ordovician sequence. The thrust plane is easily identified east of Årmoteggi and further south, where the orthoquartzite lies directly upon the Precambrian basement. It is less pronounced further north, where black shale is found below the orthoquartzite. Here the thrust plane very often splits up into minor thrust planes with local imbrication of black shale and orthoquartzite. Sometimes the Precambrian basement is also involved in the thrusting (Fig. 27). The imbrication wedges always thin out to the southwest or south-southwest, indicating thrusting from the south-west towards the north-east. The T–1 thrust plane is deformed by F_4 folds, indicating a pre-F_4 age. The north-west trend of the F_2 folds would suggest a syn-F_2 age for the thrusting.

In the central and western parts of the mapped area, the T–1 thrust plane joins up with the thrust plane (T–2) separating the Holmasjø Formation and the autochthonous Cambro–Ordovician sequence. As mentioned earlier and illustrated in Fig. 3, the T–2 thrust plane transects the autochthonous sequence along a NW–SE-striking line. Along this line the thrust plane dips steeply towards the SW, and south-west of Årmoteggi where the thrust plane is exposed the Holmasjø Formation lies directly upon the Precambrian basement. The constant orientation of the thrust plane along the contact against the autochthonous sequence most probably indicates that the cutting off of the latter and the formation of the thrust plane is caused by thrusting from the SW towards the NE. In front of this thrust plane the autochthonous sequence is deformed by a 'disharmonic' type of folding. These folds (F_2) are arranged to the strike of the T–2 thrust plane and have their axial planes (S_2) overturned towards the NE (with just a few exceptions), also supporting the theory that the direction of tectonic transport was from SW to NE during the F_2 deformation episode. The folding of the thrust plane by F_4 folds puts an upper limit on the age of the thrusting. The intense isoclinal F_2 folding with formation of transposition structures in the Holmasjø Formation was probably contemporaneous with the thrusting. Both the T–1 and the T–2 thrust planes are therefore thought to be of syn-F_2 age.

The next major thrust plane, T–3, is developed between the Nupsfonn Complex/Dyrskard Group and the Holmasjø Formation, and is the same as that mapped by Reusch et al. (1902) separating the crystalline nappe rocks from the Cambro–Silurian sediments. The thrust plane is close to horizontal, except where local folding is present. The thrust plane is deformed by F_2, F_3 and F_4 folds. Even if the F_2 folding sets an upper limit for the start of thrusting, movement must also have taken place during the F_3 or F_4 folding as minor thrust planes located along the major T–3 thrust plane deform S_2 foliations. The deformation of the S_2 structures indicates a relative movement of the crystalline rocks towards the NW. Such a relative movement of the crystalline nappes explains the well-developed S_4 crenulation cleavage in the Holmasjø Formation, and it agrees with the 'backward' movement of the nappes reported from the Finse district by Kvale (1960). Movement along the T–3 thrust plane has probably also taken place during the F_2 episode, but before the complete history of movement along the T–3 thrust zone can be unravelled more detailed work must be carried out.

The thrust plane between the Nupsfonn Complex and the Dyrskard Group (T–4) is of special interest, since the Nupsfonn Complex occurs as a local wedge showing strong brittle cataclasis. Naterstad et al. (1973) interpreted the Nupsfonn Complex as an imbricated wedge of rocks derived from the Dyrskard Group and Kvitenut Complex. This interpretation was based on the lithological similarities between the metasupracrustals at the base of the Nupsfonn Complex and the metasupracrustals of the Dyrskard Group, and on the similarities between the gneisses and intrusives in the Nupsfonn Complex and Kvitenut Complex. According to Naterstad et al. (1973) the Nupsfonn Complex has been cut away from its original position in the nappe and subsequently overridden by the overlying part of the nappe, thereby acquiring its cataclastic texture and tectonic position. The idea is fascinating, since the Nupsfonn Complex wedges out north-westwards and could thus have been cut off during overthrusting from the NW. Other possible explanations for the tectonic position and cataclastic texture of the Nupsfonn Complex are:

a) The Nupsfonn Complex is a separate thrust unit within the Hardangervidda–Ryfylke Nappe System, without any connection to the Dyrskard Group and the Kvitenut Complex, and with its root zone far to the west.
b) The Nupsfonn Complex represents a wedge derived from the Precambrian basement just west or northwest of the present area which was 'picked up' by the overlying nappes (Dyrskard Group and Kvitenut Complex) during their movement from the NW.

Both the cataclastic layering and the T–4 thrust plane are deformed by F_2 folds placing an upper limit on the age of the thrusting. The cataclastic layering and the thrust plane could have been formed during the F_1 folding. The regional dip of the thrust plane also indicates this, but more data are needed before a final interpretation can be made.

The uppermost thrust zone, T–5, shows a post-mylonitic upper greenschist/lower almandine–amphibolite facies metamorphism in contrast to the underlying thrust planes which are metamorphosed in the lowest part of the greenschist facies. This difference indicates that the movement and metamorphism in this thrust zone took place far outside the investigated area and probably also earlier in the Caledonian orogeny, perhaps in pre-Caledonian time. The movements in this zone are thought to have taken place when the F_0 structures were developed in the Dyrskard Group.

The structural history of the Caledonian rocks can be summarized as follows. In early Caledonian or Precambrian times the Kvitenut Complex with the Revsegg Formation on top was brought into contact with the Dyrskard Group, with the formation of T–5 and the F_0 structures in the Dyrskard Group.

These three units, and possibly also the Nupsfonn Complex, were then thrust onto the Baltic platform from the NW. During this phase of movement F_1 structures were developed in the incompetent zones in the Dyrskard Group; the cataclastic layering in the Nupsfonn Complex also possibly dates to this period. In a late stage of this main Caledonian thrusting the Holmasjø Formation, situated somewhere north-west of the Hardangervidda area, became involved in the nappe system and was thrust south-eastward together with the crystalline nappes.

The next period of tectonic activity resulted in the thrusting of all tectonic units towards the NE with formation of the F_2 structures. During this episode the autochthonous Cambro–Ordovician sequence was removed from the south-western part of Hardangervidda, and thrust north-eastwards. In a late stage of this phase of folding, open folds (F_3) were developed in the Holmasjø Formation. After this, intense crenulation cleavage (S_4) was developed in the incompetent rocks, indicating a relative movement of the crystalline nappes towards the NW.

Metamorphism

The petrographic descriptions (pp. 3–30) clearly show that the metamorphic grade within the area is rather variable, and that the metamorphism must have taken place at different times in different places.

PRECAMBRIAN BASEMENT

Except for secondary alteration, the minerals hornblende–biotite–microcline–quartz–plagioclase represent the characteristic assemblage in the granodioritic augen gneisses. The occurrence of plagioclase instead of albite indicates almandine–amphibolite facies metamorphism (Winkler 1967). As no index minerals are present a closer designation of metamorphic grade is impossible. The secondary alteration of hornblendes and biotites to chlorite, and the saussuritization of plagioclase indicate a later retrogressive metamorphism.

The mineral association chlorite–albite–epidote is typical of the metabasalts in the central part of the Telemark supracrustals (e.g. Austmannlia). According to Winkler (1967) this is the characteristic mineral assemblage of the lowest part of the greenschist faces. In the vicinity of Ulevåvatn, the assemblage biotite–hornblende–acid oligoclase–epidote–chlorite is found in the metabasalts, and quartz–acid oligoclase–actinolite–epidote–chlorite in the feldspathic quartz schists, indicating a distinctly higher metamorphic grade. It has been impossible to decide whether this apparent increase in metamorphism towards the east is a regional feature or is caused by the intrusive granites, as only a small area of the Telemark supracrustals is exposed. As granite intrusions disturb the contact between the Telemark supracrustals and the granodioritic augen gneisses in this part of the Valldalen synclinorium, it is difficult to demonstrate a metamorphic break here. However, in other places, for example at Nyastøl Bru in Valldalen, there is a distinct metamorphic break between the gneisses and the low-grade supracrustals with primary structures preserved. This strongly supports the interpretation put forward by Naterstad et al. (1973) that the granodioritic gneisses represent a basement for the supracrustals.

CALEDONIAN UNITS

There is also a metamorphic break between the Precambrian basement and the discordantly overlying autochthonous Cambro–Ordovician sediments, as the metamorphic mineralogy of the latter is characterized by muscovite and chlorite. In chlorite marbles, albite also is observed.

More interesting from a metamorphic point of view is the allochthonous Holmasjø Formation. Chlorites in the Holmasjø Formation, in contrast to those in the autochthonous sequence, are sometimes found as alteration products of biotite, a feature which may indicate a slightly higher metamorphism in this formation. The observed chloritized biotites and garnets a few km west of the mapped area (collected for age determination) also indicate this, although it could also be explained by an increasing metamorphism westwards.

The crystalline nappes are characterized by polymetamorphism. The pre-cataclasis mineral assemblage in the Nupsfonn Complex, plagioclase–hornblende–brown biotite–epidote, indicates almandine–amphibolite facies metamorphism, while the post-cataclasis assemblage chlorite–albite and green biotite–albite indicate low green schist facies. An absence of rocks of suitable composition makes it impossible to give a more exact determination of the almandine–amphibolite facies metamorphism.

Both the Dyrskard Group and the Kvitenut Complex, with the exception of Stavsnuten, show almandine–amphibolite facies metamorphism with mineral assemblages such as plagioclase–hornblende–biotite–epidote (garnet) and kyanite–zoisite–garnet–muscovite–quartz in the Dyrskard Group, and different oligoclase/andesine-bearing gneisses in the Kvitenut Complex. As pointed out in the petrographic description, the mylonite gneisses along the Dyrskard Group/Kvitenut Complex boundary also show almandine–amphibolite facies mineral assemblages. The grey-green mylonitized rocks in the lower division of the Dyrskard Group are exceptions to this almandine–amphibolite facies metamorphism; here, albite and chlorite are the dominant minerals, indicating low greenschist facies metamorphism. This metamorphism may be correlated with the chloritization observed elsewhere in the Dyrskard Group and the Kvitenut Complex, and with the post-cataclasis metamorphism in the underlying Nupsfonn Complex.

Even though the mapping and petrographical investigations of Stavsnuten are not yet complete, the rocks there clearly display a higher metamorphic grade than that in the surrounding rocks. This higher metamorphism is characterized by orthoclase instead of microcline, more basic plagioclase and the occurrence of pyroxene and sillimanite. These minerals indicate a metamorphism in the upper part of the almandine–amphibolite facies or maybe the lower part of the hornblende–granulite facies. The latter is indicated by the reaction biotite → orthoclase (Turner & Verhoogen 1960).

From the mineral reactions in some rocks, pyroxene → hornblende, plagioclase → epidote + acid plagioclase, together with exsolution of Ti minerals in biotite, it is deduced that these high-grade rocks have been through a younger retrogressive almandine–amphibolite facies metamorphism.

It is obvious that the medium- and high-grade metamorphism in the crystalline nappes must have taken place outside the present area, as the underlying autochthonous Cambro–Ordovician sequence is only weakly metamorphosed.

The relationship between the upper almandine–amphibolite/hornblende–granulite facies rocks in Stavsnuten and the other rocks of the Kvitenut Complex can be explained as follows. Before the Kvitenut Complex and Dyrskard Group were brought into contact, the Kvitenut Complex or parts of it were subjected to high-grade almandine–amphibolite or hornblende–granulite facies metamorphism. The two units were then juxtaposed, an event which took place together with strong mylonitization of the Kvitenut Complex, and only small parts of the complex retained evidence of the earlier high-grade metamorphism. Stavsnuten is thought to be one of these areas. Later almandine–amphibolite facies metamorphism took place in both the Dyrskard Group and the Kvitenut Complex. Both of these two periods of metamorphism took place in Precambrian or early Caledonian times. The only evidence of the main Caledonian (? Silurian) metamorphism of the crystalline nappes in their present position is seen in the growth of chlorite, green biotite and albite.

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Plate 1

GEOLOGICAL MAP OF THE AREA NORTH OF HAUKELISÆTER