Geology and Structure of the Beito Window, Valdres

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Three groups of Caledonian nappes sit directly on Precambrian gneisses. The lowest nappe consists of generally upward-facing parautochthonous or allochthonous Cambro–Ordovician quartzites and phyllites. The middle nappe is a complex of repeated slices of inverted Eocambrian sparagmites with associated Jotun igneous rocks. The highest nappe is the Jotun Nappe. The Precambrian basement beneath the nappes is part of a pre-thrusting ridge which has influenced the emplacement of the Caledonian nappes. The nappe emplacement was the first Caledonian structural event in the area but the formation of the main slaty cleavage and stretching lineation would appear to have outlasted thrusting. Some thrust rejuvenation took place during a later tectonic phase and the effects of six Caledonian tectonic phases are evident in the area.

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Introduction

The Beito Window (Fig. 1) is one of the key areas in the Slidre map originally made by Strand (1951). Firstly, the area lies on the southern margin of the large depression beneath the Jotun Nappe. Secondly, erosion has cut right down through the Caledonian nappe pile to expose the Precambrian basement beneath. An adjacent basement window lies to the southwest of Beito in the Vang area. In both windows, the complete sequence of Caledonian nappes can be viewed around the sides.

Three main groups of nappes are evident in the Slidre map sheet and around the Beito Window. The lowest group of nappes consists of Eocambrian and Cambro–Ordovician sediments (mainly quartzites and phyllites) which have been thrust in a southeasterly direction over an autochthonous Cambrian section and the Precambrian rocks of the southeast foreland. These lowest nappes correspond to the Vemdal Nappe of Strand (1954).

Kulling (1961), Loeschke (1967a, 1967b), Nickelsen (1967) and Loeschke & Nickelsen (1968) have shown that the Valdres Sparagmite forms a middle group of nappes which they call the Valdres Nappes. The rocks of these nappes consists mainly of Eocambrian sparagmite which pass up stratigraphically, via a tillite of Varangian age, into Eocambrian, Cambrian, and a possibly Ordovician quartzite-slate sequence called the Mellsenn Group.

The highest nappe is the Jotun Nappe, which was originally described by Goldschmidt (1916) and more recently by Battey (1965) and Battey & MacRitchie (1973). These nappes consist of crystalline rocks of the Bergen–

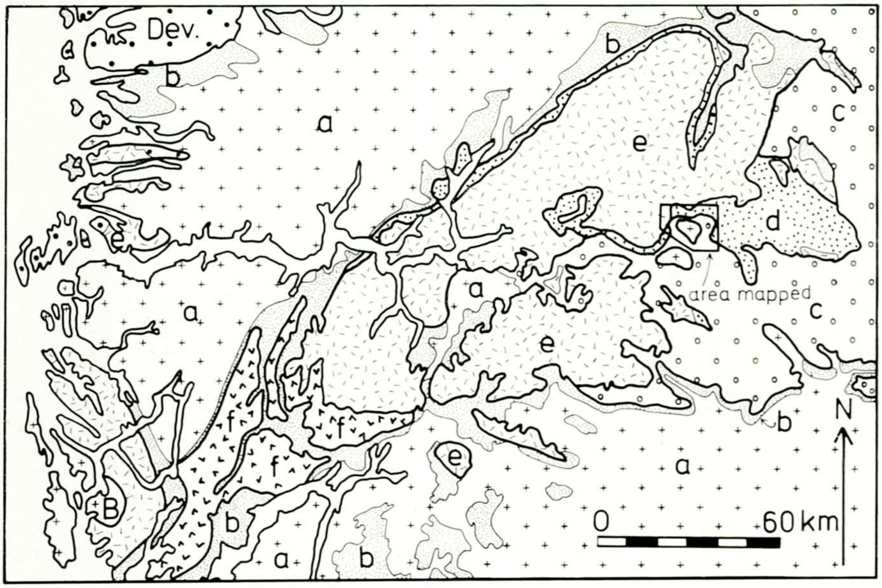


Fig. 1. Regional geology of the Jotunheim. a — Precambrian basement, b — Cambro-Ordovician, c — Vemdal Nappe, d — Valdres Sparagmite Nappe, e — Jotun Nappe, f — Bergsdalen Nappe, Dev — Devonian, B — Bergen. Area mapped is outlined.

Jotun kindred (Goldschmidt 1916), which are probably Precambrian in age. Priem (1968) has produced radiometric ages of 1550 m.y. from rocks of the Bergen–Jotun kindred. Battey & MacRitchie (1973) have shown that the Jotun rocks can be divided into two groups: there is an internal group of two-pyroxene gneisses, which are in the granulite facies of metamorphism, and an external or marginal group of igneous granites and gabbros that have subsequently been metamorphosed in the amphibolite facies. Both these groups are separated by the Gjende–Tyin fault and their age relations to each other are not known.

The area has not been mapped solely by the author. He is responsible for mapping the north, west and south sides of the window. The east side was mapped by two students from the City of London Polytechnic, Mr. P. Thomas and Mr. P. McKenna. In addition, some data have been copied from the published map of Loeschke (1967b) of the northeast side of the window. The Precambrian basement in the centre of the window was mapped by Mr. R. Hammond .Finally, Mr. G. Heard supplied the writer with some data on the Olefjell/Heklefjell Sparagmite contact on the Slettefjell ridge. All the data were transferred to an enlarged photo-mosaic prepared by the writer from aerial photographs, to make the final map. Both aerial photographs and enlarged copies of the 1 : 50,000 M 711 ordnance mapss were used to map the data in the field.

Precambrian basement

Introduction

The basement of the Beito Window has previously been described by Strand (1943, 1951). The western part of the basement consists of gabbros and hornblende schists which have a broad outcrop of acid gneisses to the east (Plate 1). Farther to the east there is a broad north-south outcrop of garnetiferous quartzites, semi-pelites, and pelites. Finally, the eastern part of the window consists of a repetition of acid gneisses. At least three separate periods of deformation can be recognized in the metasedimentary quartzites and pelites and two in the gabbros and hornblende schists. It seems likely that the gabbros, and their associated igneous rocks, were intruded after the first deformation in the quartzites but before the second deformation. The suggested history of these Precambrian rocks is as follows.

F_1 deformation

The earliest recognizable phase was responsible for the formation of the gneissic banding in some of the acid gneisses and the main foliation in the quartzite group. A few folds of this age have been recognized in gneissic xenoliths within the younger gabbros. The early foliation generally strikes east-west and dips northwards.

Gabbro intrusion

It is proposed that the gabbro was intruded after the F_1 deformation because of the presence of older, folded xenoliths of acid gneiss within the gabbro. Also, within the acid gneiss there are basic dykes (which are presumed to be synchronous with the gabbro intrusion) which cross-cut the F_1 gneissic banding and also cross-cut F_1 boudins within thin basic layers in the gneiss.

This post-F_1 igneous intrusive event had a complicated history because the intrusion of the gabbro was followed by the intrusion of granite. The gabbro can often be found as xenoliths floating within a matrix of acid rock. The xenoliths form a gradational sequence with angular, unaltered blocks at one end of the spectrum and vague basic lumps and spots at the other end. It is suggested that the various features can be explained by the assimilation of the basic rock by the acid. The best exposures of the assimilation features are to be seen in the valley of the Rauddola. As the basic xenoliths become more vague, the acid host rock becomes more basic in composition and can range from granite, through granodiorite to diorite composition. There are also a few outcrops of ultramafic rocks which are also veined by an acid host. The igneous history must have been completed with the intrusion of more basic rock because there are basalt exposures with xenoliths of the older xenolithic complex.

The contact between the xenolithic gabbro complex and the central outcrop of the acid gneiss is not sharp. A transition zone of gabbro, diorite, and granodiorite has been mapped between them to the southwest of Raudhorn, and across the contact within the acid gneiss, dykes and lenses of basic rock intrude the gneiss.

F_2 deformation

This deformation was responsible for the folding of the gneissic banding and the quartzite foliation. The F_2 folds are more widespread than the F_1 folds and are generally of similar type. They have northwest-plunging axes and may have a northwest-striking crenulation cleavage. Within the gabbro and the xenolithic complex, the F_2 deformation is the first to be recognized. The deformation has produced a large number of shear zones (Ramsay & Graham 1970) which have deformed the igneous rocks by a process of ductile simple-shear and produced planar bands in which a gneissic foliation has formed with a sigmoidal pattern. The shear zones also deform the basic xenoliths so that they are drawn out from their original equidimensional shapes into long spindle shapes. The shearing has deformed the granular gabbros, diorites, and granites to form new, foliated gneisses in which the original pyroxenes have been replaced by hornblende. This suggests that the F_2 deformation was accompanied by amphibolite facies metamorphism. The hornblendic foliation produced by the shearing has various trends between northwest and northeast. This phase of deformation also folds the basic dykes which intrude the acid gneisses, and in outcrops where the folds are absent, the presence of the deformation can be recognized by the formation of a hornblendic foliation within the dykes.

F_3 deformation

This is the final deformation which can be recognized within the basement. It refolds the F_2 crenulation cleavage and the hornblende foliation of the shear zones. The F_3 folds are generally open ripples which may produce a crinkle lineation upon the earlier cleavages and the plunge direction of these folds and lineations is towards the west or northwest.

The first two phases of deformation in the basement can be shown to have formed prior to the first Caledonian phase in the overlying rocks because at the basement-cover contact, just to the north of the River Mugna, an F_2 shear zone is truncated by the phyllonites at the base of the Caledonian nappes. These phyllonites formed during the first Caledonian structural event, so the first two structural events in the basement must be of Precambrian age. It is not possible to date the last structural event of the basement, but it is also presumed to be of Precambrian age as it cannot be matched in style or orientation with any of the Caledonian tectonic phases that occur in the cover rocks.

The basement ridge

The upper surface of the Precambrian basement (or the base of the overlying Cambro–Ordovician rocks) has been contoured (Fig. 2). Strand (1943) described the irregular nature of this surface. It rises from about 700 m at Øyangen in the southeast, up a northeast-trending slope, to almost 1200 m at

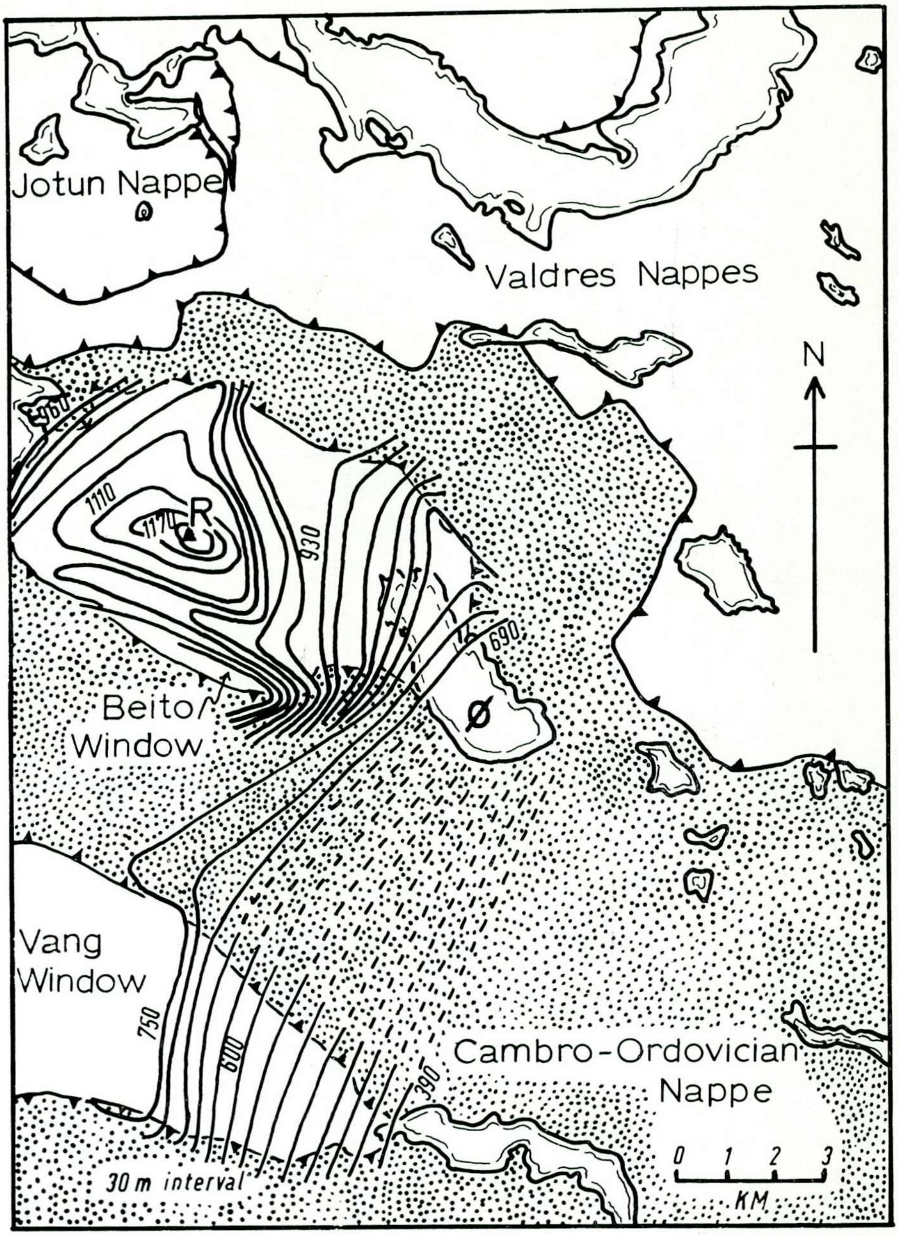


Fig. 2. Structure contours to the top of the Precambrian basement. Dotted — Cambro-Ordovician rocks overlying the basement. 30 metre contour interval. R — Raudhorn. Ø — Øyangen.

Raudhorn. The surface then slopes downwards the northwest at Fleinsendin where it disappears downwards below 960 m. The basement ridge also appears to continue towards the southwest to appear in the Vang Window (Fig. 2). In detail, the contours define an irregular surface with a northwest-trending valley or synformal depression to the south of Raudhorn. This depression is

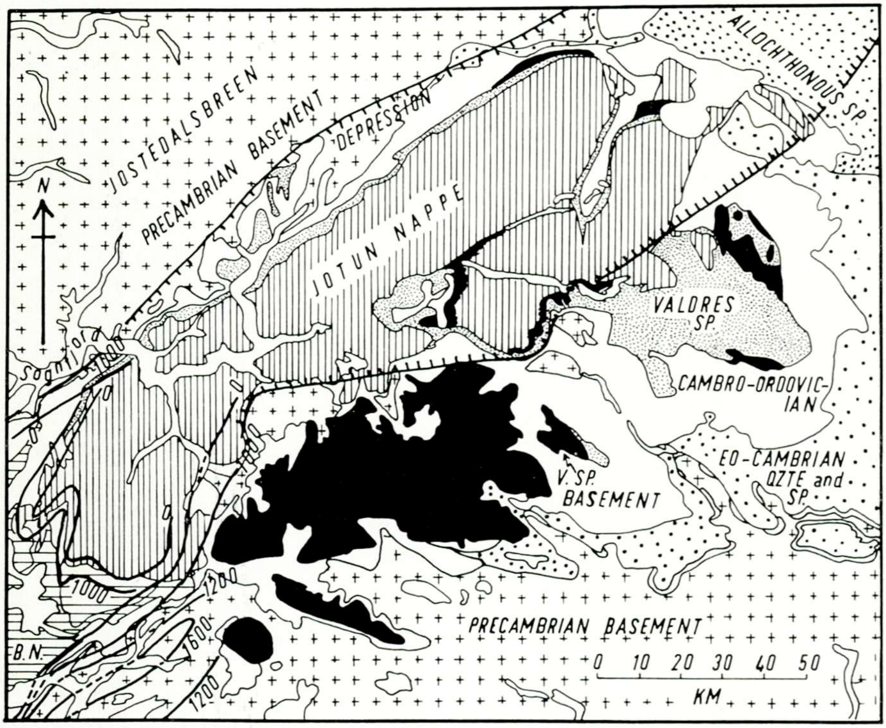


Fig. 3. Outline of the depression beneath the Jotun Nappe. Contours at the southwest end taken from Kvale, in Strand & Kulling 1972, Fig. 41. Sp. — Sparagmite. B.N. — Bergsdalen Nappe. Black — probable basement to Valdres Sparagmite.

followed at the present day by a topographic valley. The only tectonic structure near this depression with the same trend is the Mugna Antiform (see Fig. 8), which, if it has any effect on the basement, should fold the upper surface up into a ridge and not downwards into a depression. Another valley or synformal depression appears in the contours to the north of Øyangen. Again this structure is in line with an antiformal structure in the overlying rocks: the Bygdin Antiform (Hossack 1968a). These valleys do not appear to be of tectonic origin and are probably remnants of the original surface on which the cover rocks were thrust, or even the surface on which these rocks were originally deposited. They may be truly erosional Precambrian valleys which have been exhumed by more recent erosion.

The Precambrian basement also appears in a small window within the Cambrian quartzites on the west shore of Fleinsendin. The presence of this window is due to the irregular surface at the top of the basement, and because it is being brought up in the core of the Fleinsendin Antiform (Fig. 8).

The structure contour map also defines the orientation of the overlying Cambro–Ordovician nappe (Fig. 2). The quartzites at the base of this nappe have been draped over the irregular surface of the basement. To the northwest of the window, the quartzites dip towards the northwest, whereas on the southeast side of the window, they dip eastwards and southeastwards.

On the Geological Map of Norway (Holtedahl & Dons 1960), the northeast-trending basement ridge to the southeast of the Jotunheim is a prominent feature of the thrust belt. For instance, the Precambrian basement also appears to the northeast of the present area in a row of windows to the east and northeast of Rondane, finally crossing the Swedish border at Vigelen. The basement also appears to the southwest of the present area in two windows at Mørkedalen and at Lærdal. These windows, strung out for over 400 km, probably define a ridge which lies to the southeast of the Jotunheim depression. Goldschmidt (1912) and Kvale (in Strand & Kulling 1972, Fig. 41) have constructed structural contour maps of the Jotunheim depression. There is a great depression beneath the Jotun Nappe and to the southeast there is a basement ridge rising to over 1600 m which crosses Eidfjord and Sørfjord. This is in direct line with the string of basement windows and they are likely to form one continuous structure. The ridge appears to form a lip on the southeast side of the Jotunheim depression (Fig. 3). The ridge is generally at a height of 1600–1200 m, but in the Vang and Mørkedalen windows appears to be lower at about 500 m. Kvale's map also shows that the ridge descends towards the northwest to below sea-level in the Jotunheim depression. Extrapolation of the contours suggests that the depression may go down to 3 km below sea-level. Smithson (1964) and Smithson & Ramberg (1970) have postulated on geophysical arguments that the Jotun Nappe is between 6–13 km thick, and more recently Smithson et al. (1974) have suggested that the Jotun Nappe is up to 16 km thick and has a root zone within the depression.

Caledonian stratigraphy

The Caledonian rocks form a gently-dipping sequence of nappes which lie above the Precambrian basement. A tectono–stratigraphic sequence has been erected on the basis of the rock lithologies and by the recognition of intervening thrust planes. All of the units of this sequence have been internally folded and tectonically flattened so it is not possible to calculate any true stratigraphic thicknesses.

The Fleinsendin Quartzite Formation

This formation lies directly above the Precambrian basement and surrounds the window like a frame (Plate 1). The formation consists of alternations of coarse and fine quartzites, slates, slaty quartzites, and sandy quartzites and the colour ranges from white, through grey, to black. The alternation of these bands is on such a fine scale that the different rock-types cannot be separated in any greater detail. Internally, the quartzites and slates are folded by numerous tight or isoclinal flat-lying folds. The best exposed section of the unit lies just to the east of Fleinsendin, where the quartzites rest directly on the Precambrian gabbros and hornblende schists. One km to the east of the lake, just at the base of the formation, there is a quartzite conglomerate, and just above it, a truncated cross-bedding structure, which indicates that the rocks at the structural base of the formation face upwards. The structural thickness of the formation along this section to the east of Fleinsendin is 200 m.

Strand (1951) originally mapped these rocks as Cambrian quartzites, and although they are devoid of fossils at Fleinsendin, they can be traced for 30 km to the southeast into the North Etnedal area, where similar quartzites contain Cambrian fossils (Strand 1938).

The Fleinsendin Quartzite Formation also appears higher up in the structural section. For instance at Beitostølen, the Fleinsendin Quartzite is overlain by black phyllites and above these lie coarse quartzites and slates, identical to the Fleinsendin Quartzite at the base. To the northwest of the window, on each side of Fleinsendin, quartzites again reappear as an upper unit above the phyllites. Here the outcrop of quartzite closes off towards the southeast and the phyllites can be traced right round the end of the closure. This structure has all the appearance of a fold closure, and in the western outcrop, the sense of the minor folds changes from S to Z from the bottom to the top of the exposure, as they should do on opposite limbs of a fold. These probable fold closures are truncated by the sparagmite thrust to the north.

A more complicated section of quartzite and phyllite occurs on the southwest side of the window between the River Mugna and Slettefjell (Plate 1). This will be described after the phyllites have been discussed.

The Beitostølen Phyllite Formation

The Beitostølen Phyllite Formation lies structurally above the Fleinsendin Quartzite Formation (Plate 1) and was originally mapped by Strand (1951) as the Phyllite Division. Throughout the area, the formation consists of a single rock-type, a black graphitic phyllite with a slaty cleavage defined by graphite and muscovite flakes. The cleavage surfaces of this rock are not completely planar but are curved, and this feature more than any other has been used to distinguish the Beitostølen Phyllite Formation from the slaty phyllite horizons within the Fleinsendin Quartzite. In the latter, the phyllitic rocks have a perfectly planar cleavage surface. Another important tectonic feature of the Beitostølen Phyllite which aids its identification in the field are the numerous lineated quartz-rods, which are a characteristic feature of most exposures of the phyllite.

The formation has been named from the section to the east of Beitostølen, where it reaches a maximum structural thickness for the area of 800 m. Elsewhere around the window, the Beitostølen Phyllite is typically 200–300 m in thickness, but at the north end of Fleinsendin, the unit is reduced to less than 20 m in thickness.

Very few bedding layers can be observed in the Beitostølen Phyllite Formation. Occasionally one can see thin quartzite ribs only a few cm thick. However, at most exposures, the main feature is the tectonic cleavage, and originally the Beitostølen Phyllite must have been a rather monotonous shale sequence. The phyllites are not always black and graphitic, however, From Slettefjell east towards Lake Øyangen the phyllites are silvery-grey and lacking in graphite.

There is more direct evidence for the stratigraphic age of the Beitostølen Phyllite Formation than for the Fleinsendin Quartzite Formation. Bjørlykke (1905, in Strand 1951) found Ordovician graptolites in the phyllite near Rogne, just 23 km to the southeast of Øyangen.

The complicated sequence of quartzite and phyllite on the west side of the Beito Window has already been mentioned. The repetition of quartzite and phyllite is best seen in the stream sections up from the Vang road, which runs southwards over Slettefjell. There is little exposure between the streams, but the quartzites can be traced as they form a series of distinctive ridges or scarps in the topography. Five northwest-trending quartzite bands can be crossed upwards from the Precambrian basement (Plate 1). Between these, and on the top of the highest quartzite, there are five bands of black phyllite with the curved cleavage surfaces. Two of the quartzite bands in the middle of the section merge within one of the streams, and although the exposures to the southeast of here are not very good, the quartzite exposure has all the appearance of an early fold core. Beyond here, towards the southeast, the middle quartzite bands disappear, and although the outcrop pattern appears quite complicated on the map, the structural sequence is quite simple. Southwards from the River Gipa, the Precambrian basement is overlain by a lower quartzite with a phyllite band above; both these have a gentle regional dip towards the south. Another quartzite band then appears above. This quartzite forms a closed outcrop on the map (Plate 1) which is the expression of a basin structure with dips inwards towards the centre. The highest phyllite lies above the second quartzite band and is preserved from erosion by the basin, and forms the roughly circular outcrop at the basin centre. The lower phyllite can be traced eastwards towards Øyangen where it reappears to the north and continues to Beitostølen where it links up with the Beitostølen Phyllite Formation. Here it separates a lower quartzite unit from an upper quartzite unit. However, to the northwest of Beitostølen, the phyllite disappears and the lower and upper quartzite units are in contact. The phyllite reappears to the north, where it can be traced round to Fleinsendin. Structurally, the upper quartzite band on Slettefjell is probably equivalent to the upper quartzite unit to the north and south of Beitostølen. Likewise, the lower quartzite units in both areas can be correlated.

Finally, the repeated quartzite and phyllite bands which have been described west of the Vang road are truncated by the overlying sparagmites. The truncation of the highest quartzite band is very well exposed above the three closely-spaced streams which run to the north to form the River Gipa. The phyllite beneath this can be traced to the northwest beneath the sparagmite. However, 1.5 km northwest of the first truncation, this phyllite is cut out and the quartzite band beneath this in contact with the sparagmite. This quartzite band is itself cut off towards the northwest and the next phyllite beneath is then in contact with the sparagmite. The truncations of the middle two quartzite bands beneath this are not exposed, but the continuation of the intervening phyllite bands into the River Mugna and into contact with the overlying sparagmite suggests that the middle two quartzite bands have to be truncated by the sparagmite in a similar manner.

The repeated quartzite and phyllite bands are lithologically identical to the Fleinsendin Quartzite and the Beitostølen Phyllite, respectively. It is suggested that the repetition of these two formations is the result of tectonic interleaving due to folding and thrusting.

The Valdres Sparagmite

The Valdres Sparagmite (Strand 1951) or the Valdres Group (Strand 1964) lies above the Beitostølen Phyllite Formation (Plate 1). The petrography of the sparagmites in this area has already been described by Loeschke (1967b). On the cliffs between Fleinsendin and Skyrifjell, Loeschke recognized two petrographically distinct sparagmites; the Olefjell sparagmite type and the Weissgestreifter sparagmite type. However, he did not name these sparagmites as formal stratigraphic units and it is proposed by the present writer that two formal sparagmite formations can be erected which correspond to Loeschke's two sparagmite types.

The Olefjell Sparagmite Formation

This corresponds to the Olefjell sparagmite type of Loeschke (1967b) and is a fine-grained greenish or grey feldspathic greywacke with violet, white or pink porphyroclasts of alkali and plagioclase feldspars. According to Loeschke, the Olefjell Sparagmite is marked by a high concentration of epidote. The formation is named from a type section which starts from the Mellsenn Group, just south of the eastern Olevatn, northwards through Olefjell up to the Bygdin Conglomerate (Hossack 1968a). The structural thickness in this section is 650 m, but elsewhere in the area the Olefjell Sparagmite Formation us usually 400 m thick.

Within the Olefjell Sparagmite, on the cliffs below Skyrifjell, there is a quartzite conglomerate (mapped by Loeschke) which is identical to the Bygdin Conglomerate (Strand 1951). This can be traced to the north, where it disappears in the area of the Olevatn road. Eastwards it can be traced to the cliffs above Hornestøl. However, in two places to the south of Skyrifjell there are exposures of sheared granite above the conglomerate. Beneath the conglomerate is the Olefjell Sparagmite. The granite exposures are considered to be two pips of basement, and a thrust has been drawn above the conglomerate. The Olefjell Sparagmite Formation reappears above the conglomerate and the thrust. This upper unit of Olefjell Sparagmite passes upwards into the second sparagmite formation, the Heklefjell Sparagmite Formation. The contact does not appear to be a thrust and is consequently drawn as a lithological boundary. The contact between the Olefjell and Heklefjell Sparagmites can be followed round the valley to the north of Fleinsendin, and southwards on the Slettefjell ridge. The Heklefjell Sparagmite always lies above the Olefjell Sparagmite.

Eastwards from the Skyrifjell cliffs, the contact between the two sparagmites can be traced to a north-dipping contact just below Heklefjell (Plate 1) on the Bygdin road. At this contact, the two sparagmite formations are separated by a cream or orange coloured, sugary marble, 0,5 m thick.

The Heklefjell Sparagmite Formation

This formation corresponds to the Weissgestreifter sparagmite type of Loeschke (1967b). It has a higher proportion of quartz and sericite compared to the Olefjell Sparagmite. The quartz is often concentrated into 1 cm-thick pink- and cream-coloured bands which alternate with grey sparagmite bands. This gives the formation a characteristic striped appearance in the field. The type-section of the Heklefjell Sparagmite Formation has been taken to start at the cream sugary marble below Heklefjell and continuous northwards along the line of the Bygdin road up to the first lens of Bygdin Conglomerate which lies below the wedge of Jotun phyllonite (Plate 1). The lithology of this formation is as described by Loeschke (1967b) and in this type section has a structural thickness of 350 m. At the bend in the Bygdin road, halfway along the section, there are inverted cross-bedding structures which indicate that the whole section is probably overturned. Hence the Bygdin Conglomerate lies stratigraphically below the Heklefjell Sparagmite Formation and the sugary marble at the top.

The marble, in addition to its distinctive cream colour, has bright green mica foliae and passes laterally to the west and east into more mica-rich calc-schist. This marble may prove to have an important regional distribution and may be of some considerable stratigraphic interest. A similar marble has been found in the sparagmite rocks around the Vang Window by Michael Heim (pers. comm.). A similar marble has also been described by McAuslan (1967) on the north side of the Jotunheim, near Lom. Here it separates a sparagmite-like porphyroclastic quartzo–feldspathic unit from a striped pelitic and quartzo–feldspathic schist group.

The contact between the Heklefjell Sparagmite and the Olefjell Sparagmite Formations, when traced north of Heklefjell, meets the Bygdin Conglomerate at a triple-point junction within Stavtjern (Plate 1). In the western half of the area, the Heklefjell Sparagmite lies between the Bygdin Conglomerate and the Olefjell Sparagmite, whereas in the eastern half, the Olefjell Sparagmite is in direct contact with the Bygdin Conglomerate. It is proposed that the quartzitic Heklefjell Sparagmite Formation is a lateral facies variation of the quartzitic Bygdin Conglomerate, and at Stavtjern the Heklefjell Sparagmite is wedging out and grading into the Bygdin Conglomerate.

The Mellsenn Group

In several areas, the Olefjell Sparagmite Formation passes structurally downwards, via a sedimentary transition, into a quartzite-slate sequence (Plate 1). The type-area for this type of sedimentary transition is on the southern slopes of Mellane, 25 km to the southeast of Beito (Strand 1938). Recently, Loeschke & Nickelsen (1968) have shown that the rocks are overturned in the type-section and that the stratigrapically older sparagmite passes, via a tillite (which proves its Eocambrian age), into alternations of quartzite and an slate: the Mellsenn Group (Loeschke 1967a). Fossil evidence indicates that the stratigraphic base of the Mellsenn Group is probably in the Eocambrian and probably ranges up into the Ordovician. Loeschke (1967b) also mapped the transition from the Olefjell Sparagmite into the Mellsenn Group at Hornestøl and Smørkollen in the present area. Like the transition at Mellane, the passage from sparagmite into the Mellsenn Group is achieved by a gradually increasing quartz content of the sparagmite. However, in many localities northwards from Mellane, Loeschke & Nickelsen (1968) have mapped thrust contacts between these two units.

The main outcrop of the Mellsenn Group (noted on Plate 1 as quartzite and slate in sedimentary contact with the Olefjell Sparagmite) runs from the east side of Lake Fleinsendin, eastwards through Hornestøl and Smørkollen, to the south side of the eastern Olevatn, where it thins out. Several small patches of Mellsenn Group also appear on the west side of Fleinsendin. In most sections, two quartzite units can be recognized, and these can be correlated with the two thick quartzite units in the type-section (see Loeschke 1967a; Loeschke & Nickelsen 1968).

If the correlation between the quartzite-slate sequence of this area and the type Mellsenn Group is accepted, it means that the Olefjell Sparagmite-Mellsenn Group sequence to the east of Fleinsendin is inverted (Loeschke 1967b). Facing indicators are difficult to find in the sparagmites around the Beito Window because of the large amounts of tectonic flattening that the rocks have undergone. Original cross-bedding layers have been rotated tectonically until the angle between the cross layers is so small that it is difficult to see which layer is doing the truncating. Two facing directions on the east side of Fleinsendin (Plate 1) do indicate that the lower part of the Olefjell Sparagmite, at least, is inverted. The section to the east of Fleinsendin, from the Mellsenn Group, up through the Olefjell Sparagmite, to the Bygdin Conglomerate, and the two basement pips of granite, is considered to be completely overturned. The granite probably represents the original basement on which the conglomerate and the Olefjell Sparagmite were deposited.

Finally, rocks which Strand (1951) mapped as Mellsenn Group to the north and south of Beitostølen, are placed by the present writer into the Fleinsendin Quartzite Formation. The repetition of these quartzites above the Beitostølen Phyllite Formation has already been described. The upper quartzite unit is petrographically more similar to the Fleinsendin Quartzite than the quartzites of the Mellsenn Group. Loeschke (1967b), when he mapped the area north of Beitostølen, also separated these upper quartzites from the Mellsenn Group.

Jotun igneous rocks

The structurally highest rock of this area are the granites and gabbros belonging to the Bergen–Jotun kindred of Goldschmidt (1916). These were considered by Holtedahl (1936) to form a far-travelled nappe, the Jotun Nappe, that was

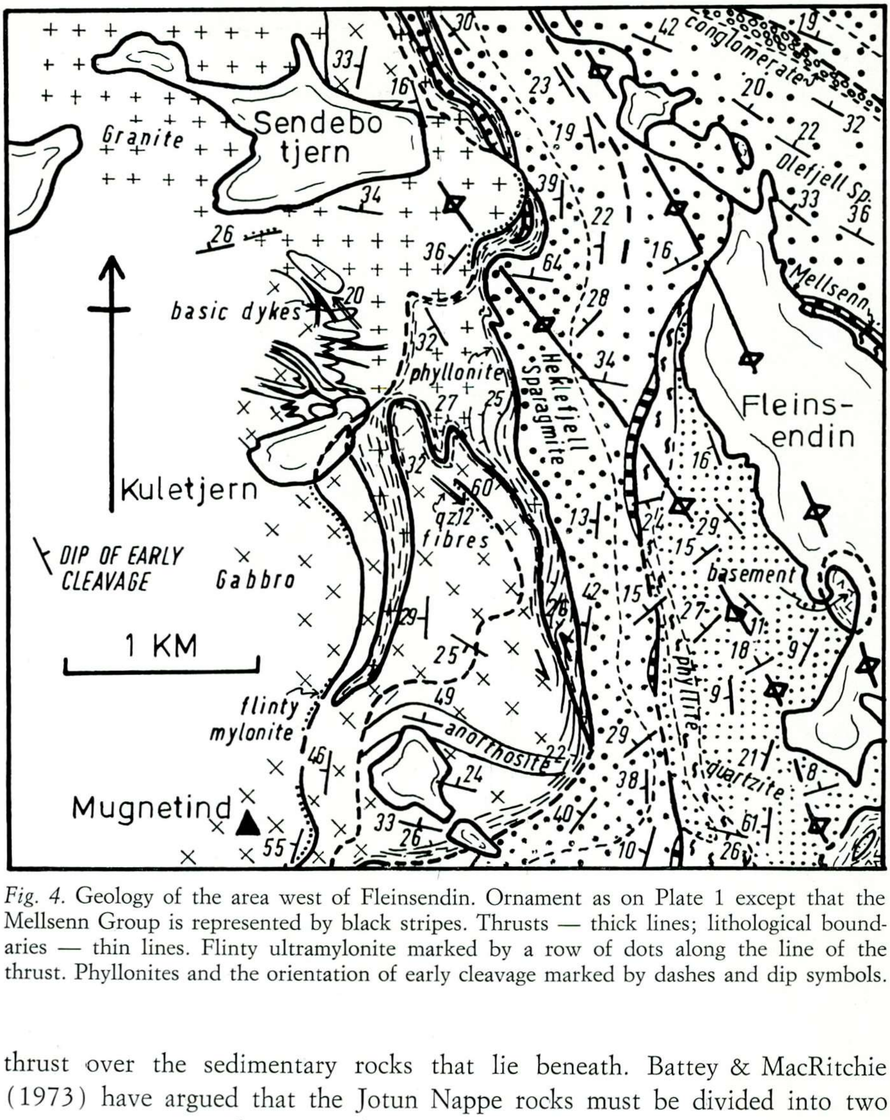


Fig. 4. Geology of the area west of Fleinsendin. Ornament as on Plate 1 except that the Mellsenn Group is represented by black stripes. Thrusts — thick lines; lithological boundaries — thin lines. Flinty ultramylonite marked by a row of dots along the line of the thrust. Phyllonites and the orientation of early cleavage marked by dashes and dip symbols.

thrust over the sedimentary rocks that lie beneath. Battey & MacRitchie (1973) have argued that the Jotun Nappe rocks must be divided into two groups: an external group of igneous parentage (granites, gabbros and syenites) that have been metamorphosed in the amphibolite facies, and an internal group of two-pyroxene granulites. The former igneous group outcrops at Grønsennknipa (Hossack 1972), where it is overlain unconformably by the Valdres Sparagmite, and in the Røsjøkollan–Dokkavatn area (Nickelsen 1974), where Jotun gabbro is overlain by gabbro conglomerate belonging to the Valdres Group. If an Eocambrian age is accepted for the Valdres Sparagmite (Loeschke & Nickelsen 1968) it is likely that the Jotun marginal igneous group is of Precambrian age and constitutes a basement on which the sparagmites were deposited (Fig. 3).

The Jotun rocks round the Beito Window can be divided into two main groups: gabbroic and granitic. There is a lot of evidence here that the gabbroic rocks are older and that the granitic rocks intrude the gabbros. The best localities for interpreting the age relations are between Sendebotjern and Kuletjern (Fig. 4). Here the granite veins the gabbro and in places the granite has xenoliths of gabbro. The veins, granite/gabbro contacts and xenoliths are crossed by a series of ophitic basic dykes which have chilled margin against these country rocks. Most outcrops of granite are separated from the gabbro by thrust planes. However, in this area south of Sendebotjern, the contact between the granite and gabbro is intrusive.

South of Kuletjern, the highest rocks around Mugnetind are mapped as gabbro (Plate 1). This conceals the fact that the gabbros here are heavily veined by pink granite. Time, and the scale of the mapping, did not allow the granites and the gabbros to be separated. As the gabbros are the main country rocks, the outcrop has been labelled accordingly.

Many of the gabbroic rocks around Mugnetind are not simple gabbros, but have a two-feldspar mineralogy. In addition to the usual white plagioclase, the gabbros also contain pink mesoperthites. These rocks are transitional to the jotunites and mangerites of the Jotun suite. Anorthosites are rare, but to the northeast of Mugnetind, and east-west trending anorthosite layer was mapped. Also, some of the granites are poor in quartz and are more properly termed syenites. These, however, have not been separated on the map.

These Jotun igneous rock have a rather remarkable similarity to the Precambrian basement granites and gabbros of the Beito Window. In both cases, igneous gabbros have been intruded by granite to form xenolithic complexes. However, assimilation effects of the gabbro by the granite are more evident in the basement rocks to the southeast of Fleinsendin. In both cases, the xenolithic complexes have been cut by ophitic basic dykes. However, the subsequent Precambrian deformation phases which deform the xenolithic rocks and the dykes are only apparent in the basement rocks of the window. In the present area, the subsequent deformation which deforms the dykes and the xenolithic rocks of the Jotun Nappe can all be attributed to the Caledonian.

Thrust planes

The rocks above the Precambrian basement have been divided up into a number of nappes separated by thrust planes. The thrusts can be recognized from the presence of phyllonite. For instance, beneath the base of the Fleinsendin Quartzite Formation, around most of the frame of the window, the Precambrian gneisses and gabbros have been sheared to finer grained chlorite schists. These schists, which are really phyllonites because of their metamorphic retrogression and reduction of grain-size, extend for only a few metres downwards from the base of the quartzites; the maximum extension is only about 10 m. These phyllonites allow a thrust to be mapped at the base of the quartzites (Plate 1). A Caledonian age for the formation of these phyllonites is confirmed by the presence of sheared lenses and lumps of the

Cambrian Fleinsendin Quartzite within the gabbroic phyllonite in the outlet stream at the south end of Fleinsendin. The Fleinsendin Quartzite and the Beitostølen Phyllite above this thrust form a nappe consisting of Cambro-Ordovician rocks.

The next overlying major thrust is at the base of the Olefjell Sparagmite or the Mellsenn Group (Plate 1). The thrust here has left little tectonic evidence, and it is mainly recognized by stratigraphic arguments. The Mellsenn Group and the sparagmites are mainly Cambrian and Eocambrian in age. In many places they overlie presumed Ordovician phyllites. Hence the superposition of older rocks over younger is easily explained by thrusting. There are, however, some tectonic effects. To the south of the River Mugna, there is a zone of papery sheared sparagmite at the base of the Olefjell Sparagmite. Also, a thrust here helps to explain the truncation of the quartzite and phyllite bands beneath.

Finally on Mellbysfjell to the northeast of Beitostølen, the Mellsenn Group below the Olefjell Sparagmite thins out and disappears. Just before it disappears, there are three high angle reverse faults which bring up the Mellsenn Group over the sparagmite. These were originally mapped by Loeschke (1967b) and were drawn by him as secondary splay faults curving out of the main thrust at the base. The basal thrust also explains why the Mellsenn Group appears to have been cut out in so many places at the base of the sparagmite.

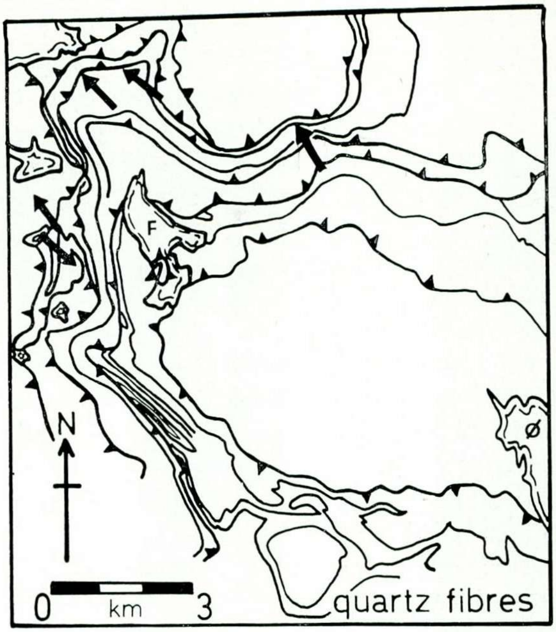
The position of the next highest thrust within the sparagmite has already been briefly discussed. Two pips of Jotun granite occur above the conglomerate band in the cliffs beneath the Skyrifjell ridge. The section of conglomerate, Olefjell Sparagmite and Mellsenn Group beneath this is considered to be completely overturned. The presence of the thrust plane just above the conglomerate and the granite pips can be inferred from the very platy or papery nature of the sparagmite in the vicinity of the contact. This papery sparagmite is interpreted as a phyllonitic rock. Also, the proposed thrust plane is marked by heavy quartz veining. The thrust has been traced to the east across the Bygdin road, via a zone of conglomerate and papery sparagmite much used by local farmers as a source of flagstone. The thrust is traced farther to the east into a conglomerate band and hence into the north side of the eastern Olevatn. The thrust is more difficult to trace towards the west. The conglomerate band marking the line of the thrust can be followed to the west of the road north of Fleinsendin, where it peters out. A papery sparagmite horizon can be traced farther to the south, but finally to the west of Fleinsendin, the trace of the thrust plane was lost. However, a view from the cliffs above the east side of Fleinsendin suggests that the thrust is being cut out by higher levels of sparagmite. In this view, the lower sparagmite (including the thrust) dips more steeply to the northwest, and this dip is defined by large benches (=bedding?) which run diagonally downwards towards the north. The benches in the sparagmite at higher levels are more flat-lying and appear to truncate the lower, steeper benches. The likeliest position for this truncation is at the base of the Heklefjell Sparagmite Formation, and even although this boundary is marked as a lithological boundary (Plate 1), it could be a disguised thrust plane which is truncating the thrust plane beneath.

The next highest thrust is in Heklefjell Sparagmite. Its presence is marked by a 1 cm-thick, red, amorphous, flinty ultramylonite (Loeschke 1967b). The best exposures are to the west of Skyrifjell and it can be traced to the east to just below the Bitihorn. To the west it can be traced to Olevatn and then into a wedge of phyllonitized granite which extends from the main Jotun Nappe, just to the east of Sendebotjern. A similar wedge of phyllonitized granite occurs to the east in the Bygdin area (Hossack 1968a). The flinty ultramylonite probably continues from the cliffs below Bitihorn round to link up with the wedge at Bygdin. It is likely that the wedge of granite phyllonite to the east of Sendebotjern is the same wedge as that at Bygdin.

The thrust plane at the base of the Jotun igneous rocks is marked everywhere by a phyllonite zone formed of sheared granite or gabbro. The origin of the phyllonites can usually be determined from the colour; those phyllonites that formed from the gabbroic rocks are darker in colour than those that formed from the granite. Several thrust planes occur in the Jotun rocks at higher levels. These generally separate granitic rocks from gabbroic. The presence of these higher thrusts can be inferred by the outcropping of phyllonites or red flinty ultramylonites.

The higher thrust relations to the west of Fleinsendin indicate that thrusting occurred in more than one stage. Between Kuletjern and the basal thrust plane, there is a folded thrust between granite and gabbro (Fig. 4). The fold axis plunges at 14° towards 339°. The folds are recumbent isoclinal folds and the folded thrust can be traced downwards to meet the basal thrust of the Jotun rocks, and upwards to Kuletjern where it is cut off by the highest thrust mapped in the area. Heim (pers. comm.) suggests that this thrust is to be correlated with the main thrust at the base of the Jotun Nappe and that all the thrusts and Jotun igneous rocks beneath this are properly part of the Valdres Nappe. It is suggested that the folded thrust was originally planar and that for some reason movement along the thrust ceased, possibly because of the rotation of the stress directions. The resolved shear stresses along the thrust plane would then fall to a low value which could no longer support movement. Continuing movement could only then take place by forming a new thrust, probably at a higher level. The movement along a higher thrust could then fold the lower thrust plane by frictional drag (see Gilluly 1960). With sufficient movement along the higher thrust, the lower thrust could become completely overfolded and over-ridden by the higher thrust. To the southeast of the folded thrust, in the overlying gabbro, there is a planar thrdust running off south and southwest from a flinty ultramylonite band. This truncates the anorthosite layer and finally links up with the main thrust to the south of Mugnetind (Fig. 4). This thrust would also have taken up some of the continuing movement once the lower thrust began to fold. Part of the displaced anorthosite on this middle thrust lies above the thrust on the lake shore south

Fig. 5. Trend and plunge direction of the long axes of syntectonic quartz fibres associated with the ultramylonite bands. F — Fleinsendin. Ø — Øyangen



of Mugnetind. The separation of the anorthosite on both sides of this thrust is about 1 km and this is probably the order of the displacement on this minor thrust. The sequence of locking and folding of the lower thrust so that it becomes over-ridden by higher thrusts is similar to that described by Gilluly (1960) in the Goat Ridge Window, Nevada.

Many years ago there was considerable discussion about the displacement direction of thrust planes and the relations between this direction and the prominent transverse stretching lineations that are common in many thrust zones (c.f. Anderson 1948; Kvale 1953). However, the stretching lineation represents the maximum elongation direction of finite strain and need not coincide with the direction of nappe displacement (Ramsay 1969). Durney & Ramsay (1973) have described the significance of syntectonic fibrous crystals of quartz and calcite. The fibrous crystals grow between points that were originally adjacent in the rock and which are gradually being displaced apart as a result of rock deformation. For instance, many slickenside striations are the result of displacement along bedding planes or fault surfaces, and the striations are long fibrous crystals which are aligned in the displacement direction. Fibrous crystals of quartz have been found in the present area lying on the top surfaces of the ultramylonite bands. They are assumed to display the movement direction of the thrust displacement, certainly its final stages. They are all aligned in a northwest-southeast direction (Fig. 5). Geologists have used for years the rough or smooth feel of the slickensides to try and determine the sense of slip as well as the direction, even although there is some argument about whether the displacement is in the rough or the smooth direction (see, for instance, Billings 1954, Gay 1970). For syntectonic fibres the sense of slip is in the smooth direction (Durney & Ramsay 1973). In the present area, the quartz fibres to the south of Sendebotjern have a smooth sense towards the southeast. Hence an overthrust sense of displacement towards the southeast is proved for one locality .

In the area of the folded thrust, the cleavage in the phyllonites on both sides of the thrust curves into the thrust in the manner of the foliation in shear zones (Ramsay & Graham 1970) (Fig. 4). As the cleavage planes curve into the thrust, they become more closely spaced and the phyllonites become more highly deformed. This is a pattern that is also typical of shear zones. It is proposed that this cleavage, and the phyllonite associated with the thrust, have been formed by the simple-shear process suggested by Ramsay & Graham (1970) and that the cleavage formation is synchronous with the thrusting. The sense of curvature defines the sense of slip and in this case the thrust displacement is an overthrust movement of the upper layer towards the southeast over the lower (the sense of slip is marked by arrows on either side of the thrust in Fig. 4).

Evidence to suggest that the lower part of the sparagmite section is overturned has already been described. In the overlying sparagmites there are only four more facing directions; two inverted and two right way up. However, it is tentatively suggested that the overlying sections are mostly overturned. In the northwest between Olevatn and Skyrifjell (Plate 1) there are several good cross-bedding units, close together in a stream section, which consistently face away from the wedge of Jotun granite within the sparagmite and again indicate overturning. Finally all the facing directions seen by the writer at Bygdin, within equivalent parts of the structural section, are all inverted. Hence it is suggested that the entire section of the Valdres Sparagmite in this area including the Mellsenn Group is inverted. Because the Jotun marginal igneous rocks form a basement to the Valdres Sparagmite at Grønsennknipa (Hossack 1972) at least some of the overlying Jotun rocks in the present area can be envisaged as Precambrian basement which has been overturned and thrust along with the Valdres Sparagmite Nappe.

Caledonian deformation

Introduction

Six phases of deformation can be recognized in the Valdres region. Not all of these can be found in any one area and the author is thankful to Dr. R. P. Nickelsen for supplying him with some of his unpublished data of deformation phases in the Steinsetfjord area, 40 km to the southeast of Beito. The best evidence for the separation of the first two phases is to be found at Steinsetfjord. However, some separation is also possible in the Beito area. As an aid to description, the tectonic effects of the first two phases will be considered together.

Phase 1 and Phase 2

The earliest tectonic structures to be formed in the area are the Phase 1 folds and major thrust planes. Not many minor Phase 1 folds are evident. They can be found locally folding the sparagmite bedding, but because of the homogeneous nature of the sparagmite, the hinges do not weather out so that they can be measured. Some major Phase 1 folds have been mapped in the southeast at Javneberg by Nickelsen (Plate 1) and he has kindly given the author his data. The folds trend northwest and are overturned to the southwest. To the northwest of Javneberg, according to Nickelsen, the style of deformation in the sparagmites changes so that the bedding, which defines the early folding, becomes heavily overprinted by the main slaty cleavage. The sparagmites then become the typical flaggy tectonic schists which are characteristic of most of the area. This change is no doubt due to an increase in the amount of internal tectonic flattening in the sparagmites.

Phase 1 minor folds are more common in the Fleinsendin Quartzite, where they can be seen to fold the bedding into tight or isoclinal recumbent folds. Because the bedding is not generally apparent in the Beitostølen Phyllite, Phase 1 folds are not visible. The trends of the Phase 1 folds are quite variable and may plunge in easterly, northerly or westerly directions; there is no obvious regional pattern.

Most of the major thrust planes are also Phase 1 structures, and the distribution of these has already been described. The folded thrust plane to the east of Kuletjern (Fig. 4) has been formed by a ductile shear-zone mechanism with the slaty cleavage in the associated phyllonites forming during thrusting. Presumably the formation of the phyllonites is also synchronous with the thrusting, and the continuing movement which could have overfolded this early-formed thrust has been mentioned. Throughout much of the area, the main slaty cleavage is parallel to the thrust planes and at first sight would appear to be synchronous with thrusting. However, the cleavage is in fact regionally a Phase 2 structure. Nickelsen has shown at Steinsetfjord that Phase 1 folds are cut obliquely by the main slaty cleavage which must be a younger Phase 2 structure. It can also be shown in a section south of the River Mugna that the main slaty cleavage is generally younger than the thrusting.

In the latter section, the early slaty cleavage (S_1) can be shown to refract up through the section (Fig. 6). In the phyllite, the cleavage has dips of between 25° to 60°, all in a southerly direction. However, when the slaty cleavage passes up into the quartzite bands, it refracts so that the dip flattens out. The dip diminishes to under 20° south and there may also be a small change in the strike orientation generally in a clock-wise sense. The refraction through the phyllite–quartzite contacts is best exposed on the undersides of the quartzite ridges. Eventually the refraction cleavage passes up to the lower contact of the Olefjell Sparagmite Formation, which is a major thrust plane. The cleavage refracts across this surface and so must be younger than the thrust plane. The most obvious refraction of the cleavage across the thrust plane is

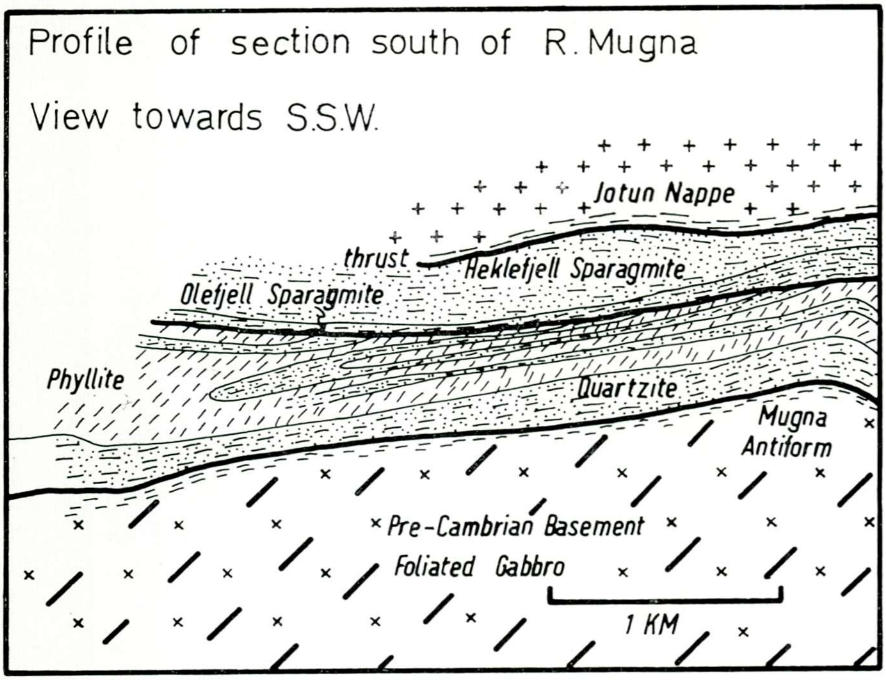


Fig. 6. Tectonic profile of the rocks south of the River Mugna. View towards the south-southwest. Thick dashes — foliation in the Precambrian rocks. Thin dashes — Phase 2 slaty cleavage.

visible when the phyllite is beneath the sparagmite. The dips of the cleavage decrease to less than 20° in the sparagmite, because of the refraction, and the strike swings through 45° to 90°, again in a clock-wise sense.

Sometimes the most prominent cleavage in the phyllites is a Phase 4 crenulation cleavage which dips steeply to the south and gives the appearance of refraction where it dies out at the quartzite and sparagmite junctions. However, the refraction of S_1 across the thrust at the base of the sparagmite has been established at contacts where the Phase 4 crenulation cleavage is only weakly developed.

The Phase 2 slaty cleavage reported by Nickelsen (S_1 in this account) occurs within the Eocambrian and Cambro–Ordovician rocks near the bottom of the Caledonian nappe pile. The cleavage refraction across the thrust in this area is also in a lower nappe. Hence it is proposed that S_1, at least in the lower nappes, was formed after the Phase 1 thrusting and folding. A slightly younger age for the cleavage and the accompanying pebble deformation has already been proposed for the Bygdin area (Hossack 1968b). It was argued there that the deformation of the Bygdin Conglomerate was a result of the static flattening under the weight of the overlying nappes. The evidence of the folded thrust in this area, however, suggests that in the higher nappes the cleavage formation was synchronous with thrusting. Now if the cleavage is the same age everywhere, it would mean that the thrusts formed at slightly different times;

conversely if the thrusts are all of the same age, it would mean that the cleavage formed at slightly different times. It is proposed that in the lower nappes, the Phase 1 folds and thrusts were formed in the rocks as they were moving ahead of the Jotun Nappe. Only when the frontal, lower nappes were finally over-ridden by the Jotun Nappe did the overburden increase so that the lower nappes were flattened under the weight of the upper nappes and the Phase 2 slaty cleavage formed.

The main stretching lineation, defined by the elongation of the rock particles, lies in the Phase 2 slaty cleavage, and hence it must also be a Phase 2 structure. In the Beito area, the Phase 2 lineation is defined by elongate pebbles in the conglomerates, elongated porphyroclasts in the sparagmites, and by the general grain elongation of the minerals of the rock matrix in the sparagmites and quartzites. There is no doubt that this direction represents the maximum finite elongation (X) whereas the slaty cleavage represents the XY-plane of the finite strain ellipsoid (Ramsay 1967). In the phyllites there are often two microcrenulations. One of these can be matched with the Phase 4 minor structures but the other is interpreted as a Phase 2 structure. The Phase 2 microcrenulation is parallel to the grain lineation of the adjacent quartzites. In the absence of strain measurements in the phyllites, it is difficult to comment on the significance of the microcrenulation, but it could be due to a progressive strain that started in the flattening field and gradually moved into the constrictional field (Flinn 1962). This is the kind of strain path that has been recorded in the pebbles of the Bygdin Conglomerate (Hossack 1968b).

The regional distribution of the Phase 2 lineation has a northwest trend around Fleinsendin and in the Olefjell Sparagmite to the north of Olevatn and Beitostølen (Fig. 7). However, in the eastern and southern parts of the window, there is a gradual swing of the stretching lineation into an east-west trend. Note that the upper quartzite unit north of Beitostølen, which was originally mapped by Strand (1951) as belonging to the Mellsenn Group, has the east-west trend characteristic of the lower Cambro–Ordovician nappe rather than the northwest trend which is characteristic of the overlying Mellsenn Group and Olefjell Sparagmite. Briefly then, the Jotun and Valdres Sparagmite nappe have a northwest stretching direction, whereas the Cambro–Ordovician nappe at the bottom of the nappe pile has a northwest stretching direction on the north flank of the underlying basement ridge, but on the crest and southern flank, it has an east-west stretching direction. This suggests that the underlying basement ridge (Fig. 2) has controlled the direction in which the lower nappe was allowed to elongate.

There is some evidence of major Phase 1 folding. The quartzite–phyllite contact on the north side of the Slettefjell basin is strongly interdigitated in the manner of folding (Fig. 8). These are probably Phase 1 folds with northwest axial traces which dip gently southwest. Only two minor Phase 1 folds were mapped in this area; one plunges 6° southeast and the other 3° east. Hence the trend of the major folds could be east or southeast. Several cleavage-bedding intersections were also measured here, but they have had to be

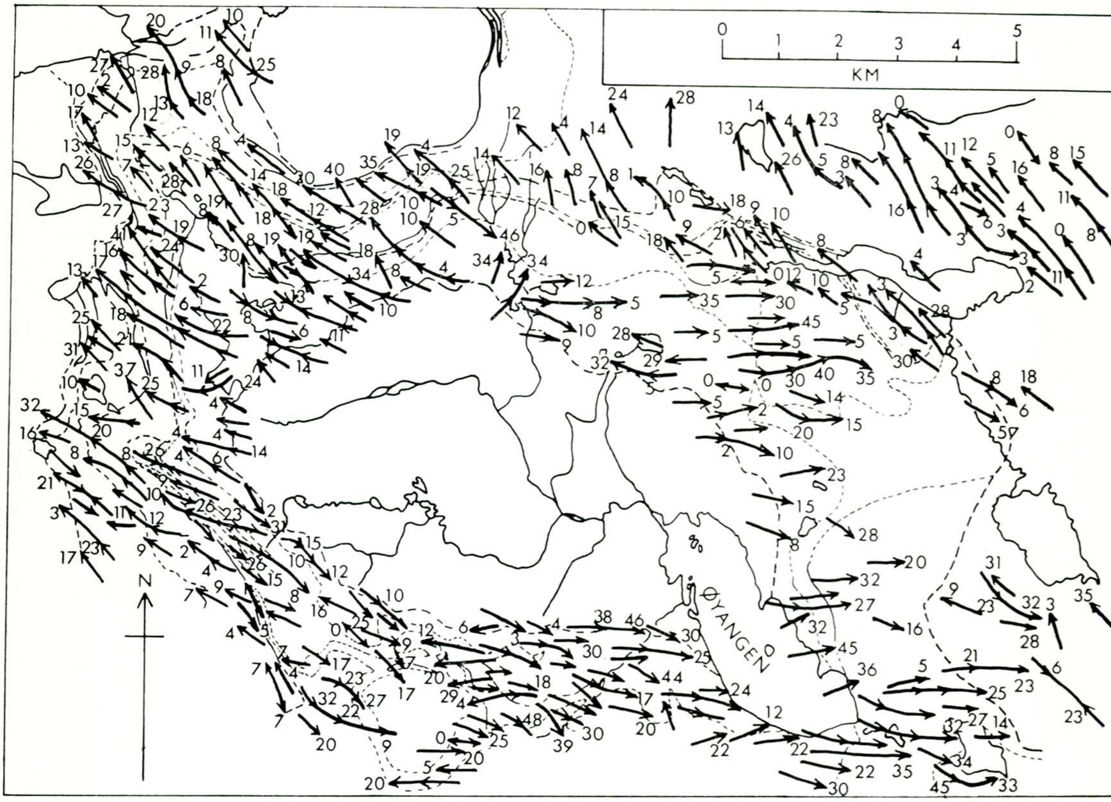


Fig. 7. Phase 2 stretching lineation directions.

discarded because the cleavage is a Phase 2 structure. The quartzite fold core in the section to the south of the River Mugna has been described, and the axial trace of this fold has the same northwest strike as the Slettefjell Phase 1 folds. A possible continuation of this fold is outlined (Fig. 8). It is cut off at its northwest end by the basal thrust of the Valdres Sparagmite Nappe. Towards the southeast it could separate the lower quartzite from the upper quartzite (each representing the opposing limbs of the fold) with the phyllite unit forming the fold core. The repetition of the upper and lower quartzite units at Beitostølen could be the result of the same fold, and a possible continuation of the fold axial trace right round the south and southeast sides of the window is indicated (Fig. 8). However, the evidence for this fold is negligible and the superposition of the upper quartzite plate over the phyllite and the lower quartzite could also be the result of thrusting.

There is a probable higher complimentary fold which also lies in the area of the Slettefjell basin. The highest part of the structural succession is the phyllite which lies in the centre of the basin. Beneath this, there is an upper quartzite band which meets a middle quartzite band in the northwest corner of the basin (Fig. 8). Between this upper and middle quartzite, there is more phyllite. It is likely that this is a Phase 1 fold and that both quartzites cor-

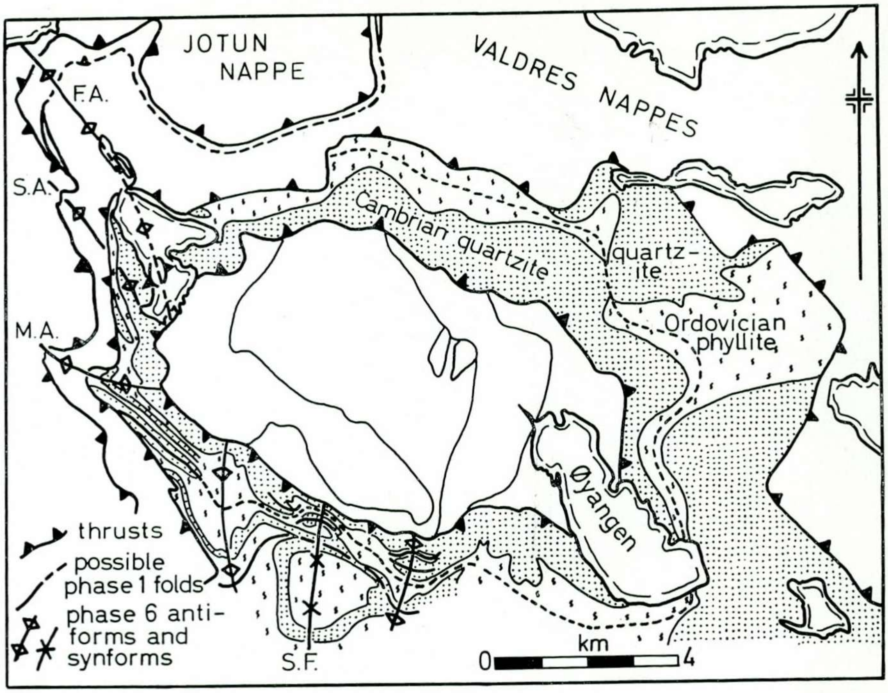


Fig. 8. Phase 1 and Phase 6 major fold axial traces. F.A. — Fleinsendin Antiform; S.A. — Sendebotjern Antiform; M.A. — Mugna Antiform; S.F. — Slettefjell Folds.

respond to the upper and lower limbs. As both quartzites dip gently south, this fold will be a recumbent fold which closes towards the northeast. This fold seems to continue from the western Slettefjell antiform, round the middle synform, to the eastern antiform (Fig. 8). The section up from the basement, south towards the Slettefjell basin, has the appearance of a major fold pair which causes the repetition of the quartzite–phyllite sequence. As the phyllite is probably younger than the quartzite, the lowest fold will stratigraphically be a syncline which is closing downwards the south. Because of the southern dips of the fold limbs, this fold will also be a synform. The overlying complimentary fold will hence be an antiformal anticline closing northwards. The other smaller Phase 1 fold traces in this area could be parasitic folds on these larger folds. The fold traces are all roughly parallel to the strike of the basal thrust above the Precambrian basement, probably because they all formed at the same time.

As indicated earlier, there is little evidence of facing. The lower quartzite is apparently right way up just above the basement to the southeast of Fleinsendin and is also facing upwards to the west of Øyangen (Plate 1). However, parts of the overlying quartzite–phyllite sequence must be inverted because of the overturned nature of the major and minor Phase 1 folding.

Phase 3

Phase 3 structures are not represented in the Beito area, but they do occur at

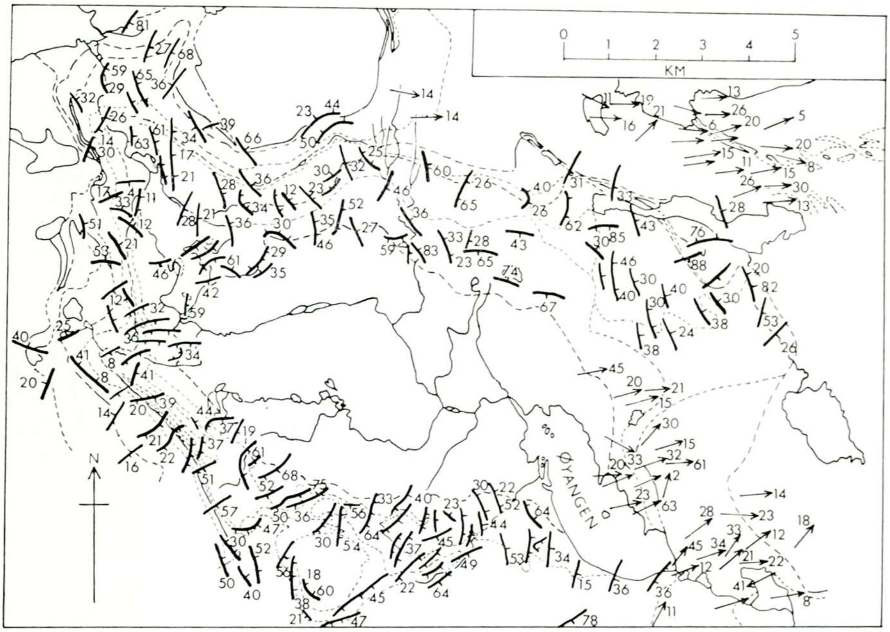


Fig. 9. Phase 4 axial surfaces and crenulation cleavages. Arrows — Phase 4 crenulation linations.

Bygdin. There the Phase 2 slaty cleavage (S_1) and stretching lineation (i.e. the Bygdin Conglomerate elongation direction) are refolded by northwest-trending major and minor folds of the Bygdin Antiform generation (Hossack 1968a). These were originally mapped as F_2 structures.

Phase 4

The Phase 4 structures are the most ubiquitous in the area. They consist of minor folds which deform the earlier Phase 2 slaty cleavage to form a widespread Phase 4 crenulation cleavage. These are synchronous with the F_3 folds mapped at Bygdin (Hossack 1968a) where they refold the Phase 3 (or F_2) structures associated with the Bygdin Antiform.

The most common strike of the Phase 4 crenulation cleavages and axial surfaces is northeast with a southeast dip. In thin-section, the crenulation cleavage has dusty seams parallel to the cleavage planes, and these are probably the result of tectonic pressure solution (Williams 1972). The associated minor folds are generally overturned towards the northwest. However, like the corresponding folds at Bygdin, these folds are often paired conjugate folds (Johnson 1965) so that there is often an accompanying fold which is inclined in a northerly direction (see Fig. 10). The regional pattern of the axial surfaces and crenulation cleavages is quite complex (Fig. 9). Trends other than the usual northeast one are apparent. A northwest, north, or northeast set is apparent to the north of Fleinsendin and in a strip to the east towards Beitostølen. The northeast-trending set is most common in the southern part of the

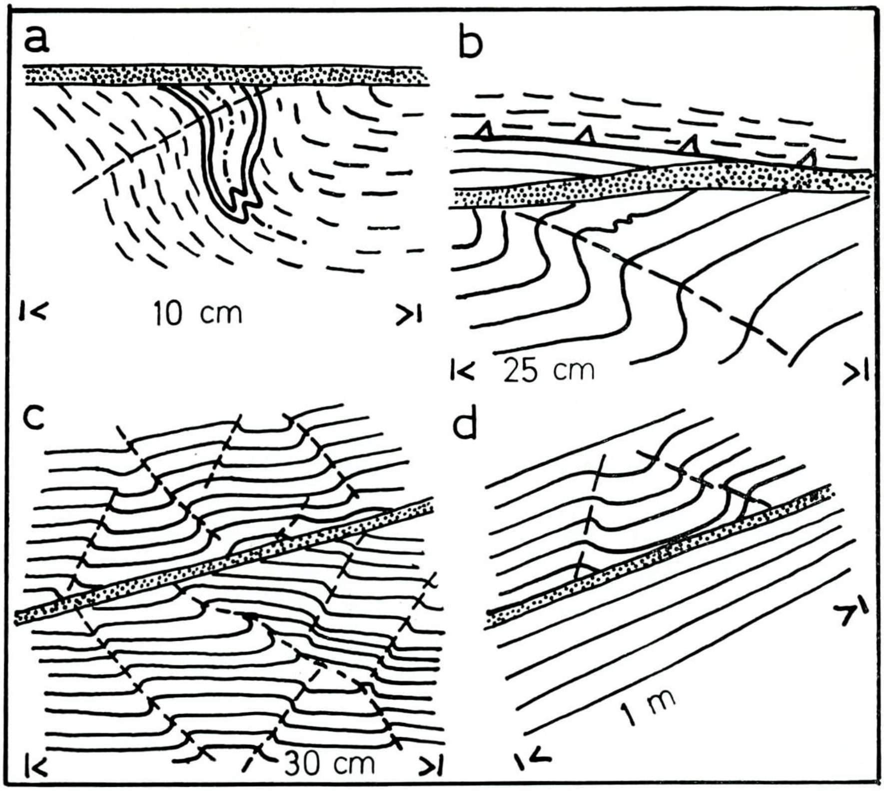


Fig. 10. Field relations between ultramylonite bands (dotted) and Phase 4 folds. (a) Phase 1 fold in foliated gabbro refolded by a Phase 4 fold. Both folds cut by ultramylonite. View towards west-northwest. (b) Gabbro phyllonite (dashed lines) thrust over sparagmite. Phase 4 fold truncated by ultramylonite. View towards north. (c) Conjugate Phase 4 folds in sparagmite. View towards south. (d) Conjugate Phase 4 fold with ultramylonite band. View towards east.

window. Two major trends are apparent on either side of the basement ridge, so the ridge may be controlling the formation of tectonic structures in the overlying nappes.

There are some Phase 4 folds which are large enough to be mapped on Plate 1. These lie near the quartzite–phyllite boundary to the west of Øyangen and are shown by the folding of the slaty cleavage. The folds have a northeast plunge at the west end of the section, but there is a gradual swing round to a northwest plunge at the east end of the folded section. There is another Phase 4 synform defined by the folded phyllite band just to the northeast of these folds.

Phase 5

This deformation phase consisted of renewed thrusting. In several localities thrusts have been found cutting through the Phase 4 folds (Fig. 10). A renewal of thrusting has already been reported at Bygdin (Hossack 1968a). Several new localities have been found around the Beito Window. Generally, it is the flinty ultramylonite bands that truncate the Phase 4 folds. However, sometimes the relations are not clear and the flinty mylonites look as if they could be synchronous with some of the Phase 4 folds (Fig. 10c, d). In some of the thrust planes, phyllonites (which are undoubtedly Phase 1 structures) and the flinty ultramylonites occur along the same thrust planes, so it would appear that the same thrust planes have moved more than once.

Only two of the major thrust planes are represented solely by the flinty mylonites. One is the thrust that extends out of the phyllonitized granite wedge east of Sendebotjern and the other the upper thrust on Mellbysfjell. The latter thrust plane is marked by a flinty mylonitic breccia (Loeschke 1967b) and is undoubtedly a late structure. The mylonite consists of randomly orientated irregular fragments of foliated Olefjell Sparagmite in a finer grained matrix. As the foliation is a Phase 2 structure, it is obvious that the mylonitic breccia must be of post Phase 2 development.

The quartz fibres that were described earlier are to be found on the upper surfaces of the flinty ultramylonite bands (Fig. 5). Hence it is likely that they only give the thrust movement direction for the Phase 5 regeneration and not the main Phase 1 thrusting.

Phase 6

The structures of Phase 6 include major and minor folds. To the northwest of Fleinsendin, there is a large embayment in the traces of the thrust planes and this is the result of major folding along a northwest-plunging antiform, the Fleinsendin Antiform. Two more northwest-plunging antiforms of the same age occur along the River Mugna and to the southeast of Sendebotjern. Between these three antiforms (Fig. 8) there are broad, flat synforms which are left unmarked. The antiforms and synforms were formed after the Phase 5 ultramylonites as the latter are folded around these folds.

Parasitic minor folds occur in association with the Phase 6 folds (Fig. 11). The axial surfaces trend west-northwest and northwest around the Fleinsendin Antiform and generally dip steeply between 60° to 80° to the northeast. The hinge lines plunge between 10° and 30° to the northwest. The folds are mainly open warps, and in the core of the Mugna Antiform, Phase 4 minor folds can be seen to be refolded around the antiformal crest.

Folds of this age also occur in the southern and eastern parts of the area. On the Slettefjell ridge, there are two antiforms with an intervening synform (the Slettefjell folds, Fig. 8). However, the trend of these folds is different from that of the Fleinsendin folds and the axial traces of both the major and minor folds trend north-south (Figs. 8 & 11). Again the folds are upright open warps with axial surfaces dipping steeply east between 60° and 80°. The plunge direction of the fold hinges is not constant. They plunge between 10° and 50° to the south, but a few culminations and depressions of plunge are evident. For instance, there are two culminations which form small domes and which are marked on Plate 1 by closed loops of cleavage dips. These occur on the most western antiform and on the eastern limb of the synform. A major plunge depression forms the basin-like structure of the central synform and

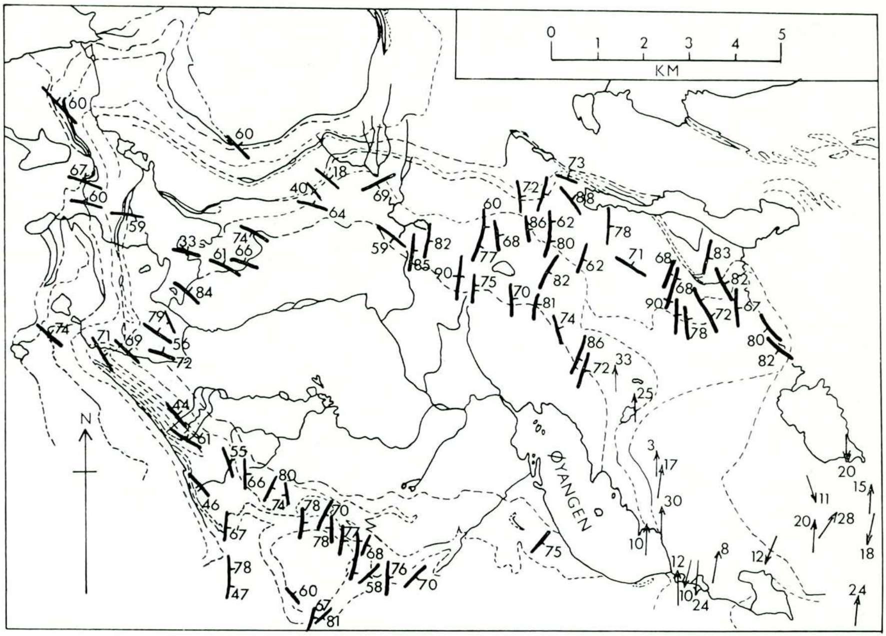


Fig. 11. Axial surfaces of Phase 6 minor folds. Arrows — Phase 6 fold hinges east of Øyangen.

accounts for the closed outcrop pattern of the quartzites. In the north of the basin, the plunge is towards the south, and in the south of the basin, it is towards the north.

Within the area of the Slettefjell folds, there is evidence of interference patterns of refolding of the type described by Ramsay (1962). The patterns are caused by the refolding of the rather nebulous Phase 1 folds by the Phase 6 Slettefjell folds. All three of Ramsay's interference pattern-types are present. The large Slettefjell basin is similar to a type 1 interference pattern, and on its northern edge there is a small offshoot of quartzite which has the arrowhead or mushroom outcrop pattern of type 2; type 3 interference occurs on the eastern antiform of the Slettefjell folds. The type 1 basin is not thought to be a true refolding pattern, but rather the plunge depression is following the irregular surface of the underlying basement. The two other patterns are probably due to refolding. The change in the shape of the refolding pattern between type 2 and type 3 can be explained by a change in the axial direction of the Phase 1 fold, which is being refolded, relative to the Phase 6 folding. In the mushroom structure, the original orientation of the Phase 1 axis was probably southeast, at a high angle to the Phase 6 structure. In the type 3 pattern to the east, the axial direction of Phase 1 would have to be closer to that of the Phase 6 antiform, i.e. nearly north-south.

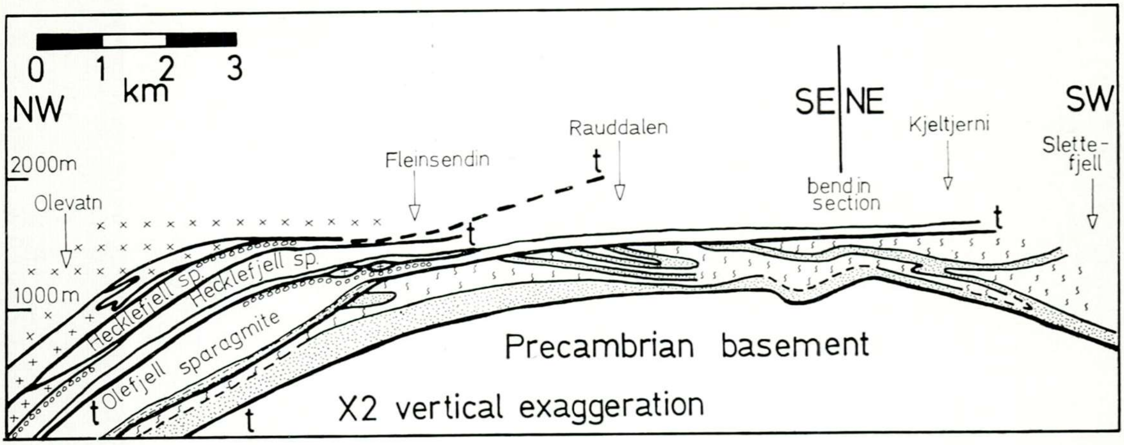


Fig. 12. Geological section through the Beito Window. Ornamentation as in Plate 1, but the sparagmite and Precambrian basement are here left blank. Vertical exaggeration ×2.

Discussion

The major Caledonian structure consists of a pile of sheet-like nappes that have been thrust to the southeast over the Precambrian basement (Fig. 12). The lowest group of cover rocks, consisting of Cambrian quartzite and Ordovician phyllite, have moved relative to the underlying Precambrian basement because of the presence of phyllonites at their interface. Hence an autochthonous cover is no longer present in the area (if it ever existed). The overlying Cambro–Ordovician nappe is parautochthonous or allochthonous. The middle sparagmite nappes and the overlying Jotun rocks are allochthonous.

This thrusting produced overfolding in the lower sedimentary nappes but an overall direction of overthrusting is not evident in this area. The main slaty cleavage and the elongation lineation in the lower nappes were formed mainly during a second phase of deformation after the main nappe emplacement. The slaty cleavage, as a result, cross-cuts the early folds and thrusts. As the main slaty cleavage is rather flat-lying, it is suggested to have resulted from a vertical flattening, probably under the weight of the overlying nappes. The Precambrian basement beneath the cover nappes has not been deformed by the thrust movements or by the flattening of the overlying nappes. Down to the deepest level of exposure in the window, the basement appears to be autochthonous and not part of a nappe beneath the Cambro–Ordovician nappe. Hence the forces which deformed the Caledonian nappes would not appear to have been transmitted to the basement, and a mechanism of sliding of the Caledonian cover rocks under gravity seems likely.

The underlying Precambrian basement has a northeast-trending ridge structure, and it is pertinent to discuss how this ridge formed. It may have been present prior to the emplacement of the Caledonian nappe rocks, or it could have formed after nappe emplacement. Ramberg (1966) has described how there are two great internal and external ridges in the underlying basement beneath the Scandinavian Caledonides which appear in two rows of basal gneiss culminations. He suggests that originally they may have been the result of the primary unevenness of the geosynclinal floor, but that they are mainly due to the uprise of a series of domes, spaced out along the ridges, because of the inversion of density stratification. The Beito Window culmination is part of the line of the external ridge. Ramberg suggests that the uprise of the basal gneiss culminations was a Caledonian event, and that because the rocks at depth were less dense than those above, buoyancy forces caused the basement to rise up in domes into the overlying denser cover rocks.

In the Beito area, however, it seems unlikely that an inversion of density stratification ever occurred, except after thrusting. The cover rocks are less dense than the Precambrian basement rocks, so no buoyancy forces would be set up. Even after the emplacement of the Jotun nappe, when more dense gabbros and granites were thrust over the less dense quartzites, slates, and phyllites, it would be the Cambro–Ordovician sedimentary rocks that would be expected to rise, and not the basement rocks. A mechanism of buoyancy rise for the basement rocks of the Beito Window is therefore rejected.

The uplift of the basement could be the result of folding of the basal interface as a result of horizontal contraction. The general local base level of the top of the basement would appear to be that of the frontal zone of the Caledonian nappes at 400 m above sea-level, 40 km to the southeast of Beito. The basement height at the south end of Øyangen is also probably at 400 m. Hence the amplitude of the folding would have to be of the order of 800 m over a distance of 12 km. However, there are no structures in the basement which can be matched with any of the Caledonian structures in the cover, and a folding of this magnitude would be expected to leave some strain behind in the rocks.

There are some arguments that suggest that the basement ridge is a pre-thrusting structure. For instance, the Cambro–Ordovician cover is draped over the irregular basement surface (Fig. 12). The dips of the quartzite–phyllite sequence are generally parallel to the inclinations of the basement upper surface and form a hump-like structure over the ridge. The overlying sparagmite nappes are more flat-lying and in fact have planed off or cut out the crest of the quartzite–phyllite hump above the ridge axis (Fig. 12). This accounts for the truncation of the quarzite and phyllite bands to the south of the River Mugna. On the down-movement side of the ridge, relative to the thrust displacement, the quartzite and phyllite show an increase in the structural thickness because of the steeper dips to the south relative to the flat, overlying Valdres Nappe. Hence the ridge would appear to have existed, in part, before the arrival of the Valdres Nappe.

A similar planing occurs within the Cambro–Ordovician section. Overfolding or overthrusting has carried the Cambrian Fleinsendin Quartzite up over the the Ordovician Beitostølen Phyllite, to the north and south of Beitosølen. However, over the ridge crest the upper quartzite has planed off the phyllite so that it rests directly on the lower quartzite. A mechanism of overfolding and overthrusting on the down-movement side of the ridge can also be used to explain the repetition of the Fleinsendin Quartzite and the Beitostølen Phyllite Formations in the area south of the River Mugna.

The folding of the thrust plane in the Jotun rocks to the east of Kuletjern can also be ascribed to a pre-thrusting ridge, which would act as a barrier to thrusting. Movement along the Kuletjern thrust could have stopped because it locked against this barrier, and continuing movement could only occur by forming thrusts at higher levels. Like the mechanism suggested by Gilluly (1960), the movement of the higher thrusts could drag the under-lying rocks to form the folded thrust plane. Note, however, that in the Gilluly description, the folding is one the down-movement side of the window relative to the direction of thrusting. In the present area, the folding is on the up-movement side of the window.

Finally, the presence of the pre-thrust ridge is suggested by the fact the phyllite is extremely thin on the north side of the window: only a few metres thick where it finally disappears beneath the sparagmite and the Mellsenn Group. This would correspond to the side towards which the thrusts were actively moving, and the extreme thinning could be due to the squeezing of the incompetent phyllite against the ridge by the moving Valdres Nappe. Hence it seems likely that the ridge in the underlying basement was a pre-thrust structure, and there is no reason to suppose that the ridge was not present during the deposition of the Caledonian cover rocks.

The depression beneath the Jotunheim is also probably a pre-thrusting feature of the irregular upper surface of the Precambrian basement. However, it is likely that this depression probably underwent some modification during the thrusting. The arrival of a thick pile of nappes above the depression would cause some isostatic readjustment. The base of the depression would probably sink under the weight of the overlying nappes.

It is suggested that the Phase 3 Bygdin Antiform was formed by a buoyancy rise, after the completion of the Phase 1 and Phase 2 movements, because of the rather restricted regional distribution of Phase 3. A profile through the Bygdin Antiform (Hossack 1965) indicates that it has many geometric similarities with load structures that are formed at the base of greywacke beds. It is suggested that the antiform is a result of loading of the Jotun Nappe over the Valdres Sparagmite and the Bygdin Conglomerate. With the emplacement of the Jotun Nappe over the Valdres Nappe, there would be a density inversion, and the buoyancy mechanism of Ramberg (1966) would come into operation. The Bygdin Conglomerate and the Valdres Sparagmite probably began to rise up into the base of the Jotun Nappe to form the Bygdin Antiform and its associated minor structures.

The Phase 4 deformation was more widespread and obviously involved considerable deformation of the cover rocks. The Phase 5 rejuvenation of thrusting could have been a late stage of the Phase 4 deformation or may have been completely later. The Phase 4 folds are more common in the sedimentary rocks than they are in the overlying Jotun rocks. Hence the sedimentary nappes would appear to have undergone larger Phase 4 strains than the Jotun rocks. This would mean that the boundary between the former and the latter would have to be a strain discontinuity, and this discontinuity would probably be marked by renewed thrusting.

The Phase 6 deformation was the last Caledonian event and involved contraction of the cover rocks mainly in a horizontal east-west or northeast-southwest direction.

The problems of the root zones and the amount of displacement of each of the nappes cannot really be solved from the geology of the present area. Smithson et al. (1974) have argued that the Bouguer gravity anomalies over the Jotun Nappe are consistent with a model in which the Jotun Nappe is up to 16 km thick and that the nappe is rooted within the Jotunheim depression (see Fig. 3). Hence, according to this model, the amount of translation which could have taken place in the Jotun Nappe, and in the underlying cover nappes, during the thrusting is quite limited. The Jotun igneous rocks are in many ways similar to the igneous rocks of the Precambrian basement in the Beito Window. During mapping, the writer was convinced that they were the same rocks, and that the amount of nappe displacement of the Jotun Nappe relative to the basement need not be large. However, the total length of the cover nappes to the south of the Beito Window must also be considered. Calculations by Nickelsen (in Hossack & Nickelsen 1974) have shown that the total length of the cover nappes between the southeast foreland and the Jotun nappe is about 200 km and this must be the order of the amount of nappe transport for the Jotun Nappe. This distance is of the same order as that estimated by Gale & Roberts (1974) for the obduction of the Støren nappe, and similar to the amount of nappe translation in the section through the Caledonides between Östersund and Trondheim (Gee 1975). Gee calculated that the eastern nappes must have been moved through a distance of 120 km eastwards, whereas the western nappes have undergone a translation of up to 300 km. Hence the model of Smithson et al. (1974) is not accepted by the writer. It is likely that the Valdres Sparagmite nappes have been transported southeastwards from the area of the Sunnmøre–Jostedalsbreen basal gneiss culmination or even farther west from the west coast area of Norway (Hossack & Nickelsen 1974). The overlying Jotun Nappe has probably been thrust for a greater distance and probably originated from an area off the west coast of Norway.

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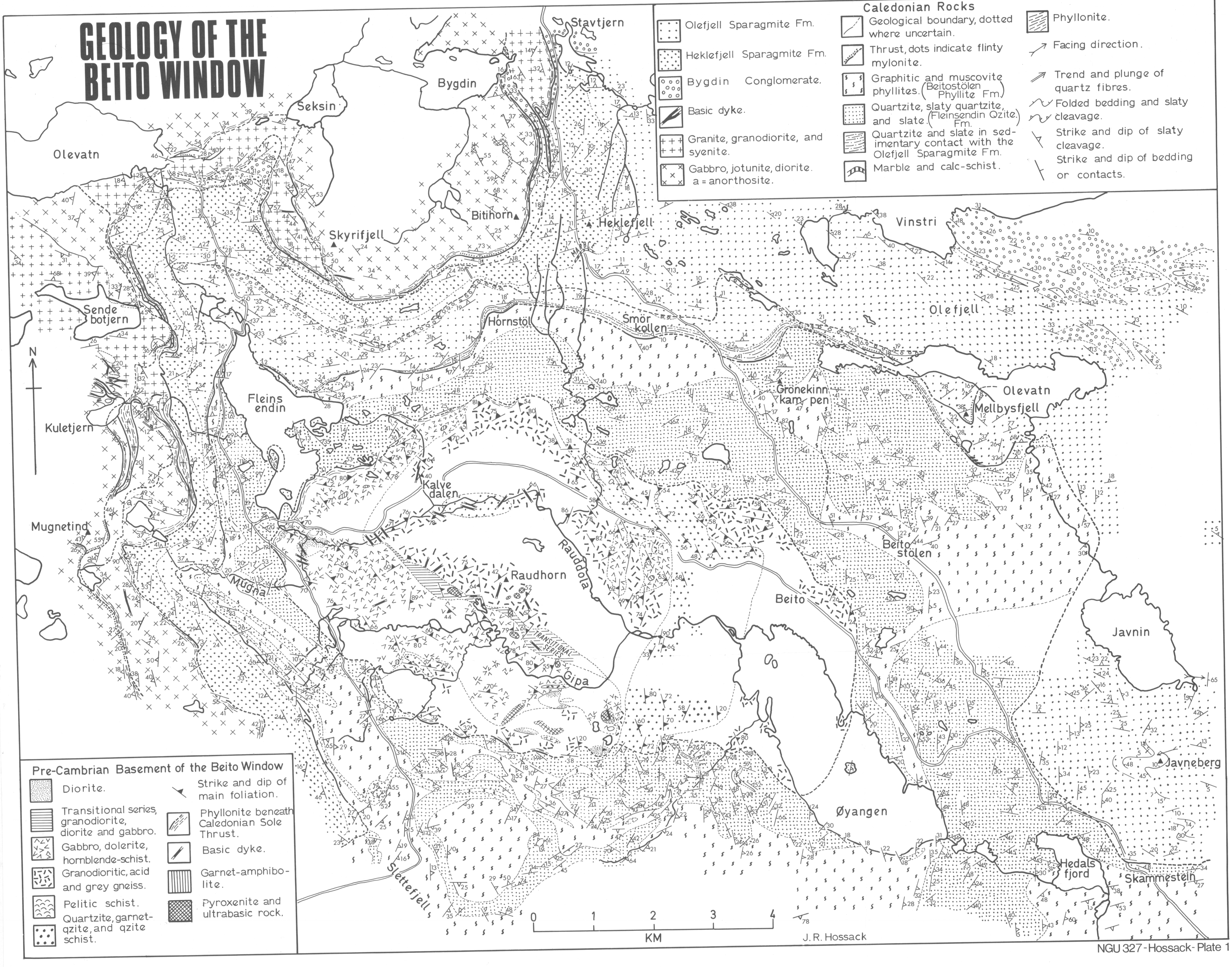

NGU 327-Hossack-Plate 1