Structural history of the Bygdin area, Oppland

By

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Abstract.

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

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

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

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

1. Introduction.

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

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

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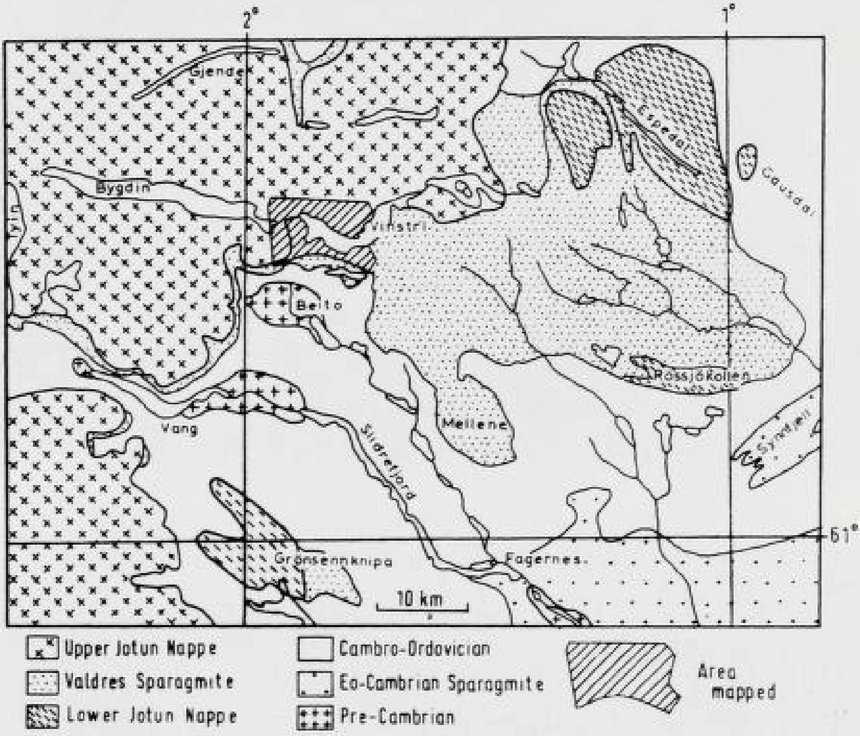

Fig. 1. Geological map of the Valdres district.

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

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

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

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

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

2. Stratigraphy and Petrography.

(A) Jotun Nappe Rocks.

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

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

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

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

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

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

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

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

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

(i) Gabbro crystallization and emplacement of the ultrabasic body.
(ii) ?Intermediate igneous rock crystallized.
(iii) Intrusion of granite.
(iv) Intrusion of dolerite dyke.

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

(B) Valdres Sparagmite and Bygdin Conglomerate.

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

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

3. Minor structures of the Bygdin area.

(A) Introduction.

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

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

(B) First Movement Phase (F_1).

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

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

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

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

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

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

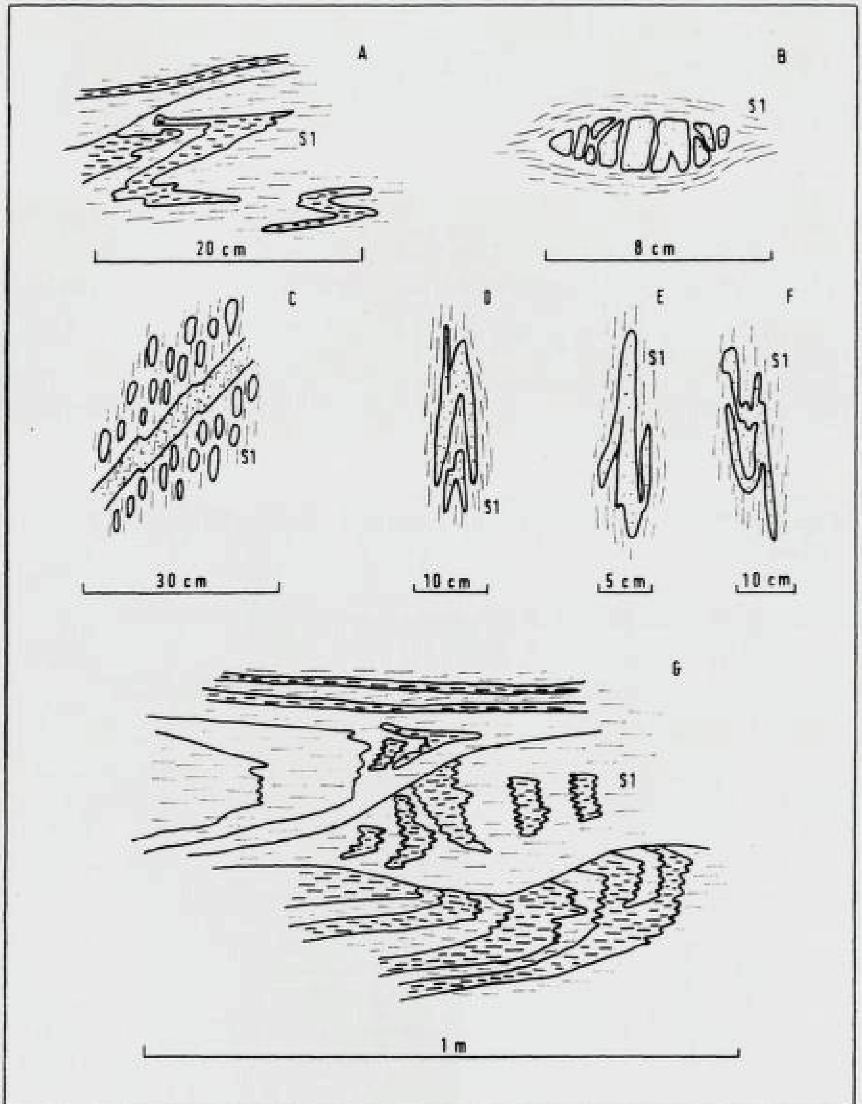

Fig. 2. Minor structures of the First Movement Phase.

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

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

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

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

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

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

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

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

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

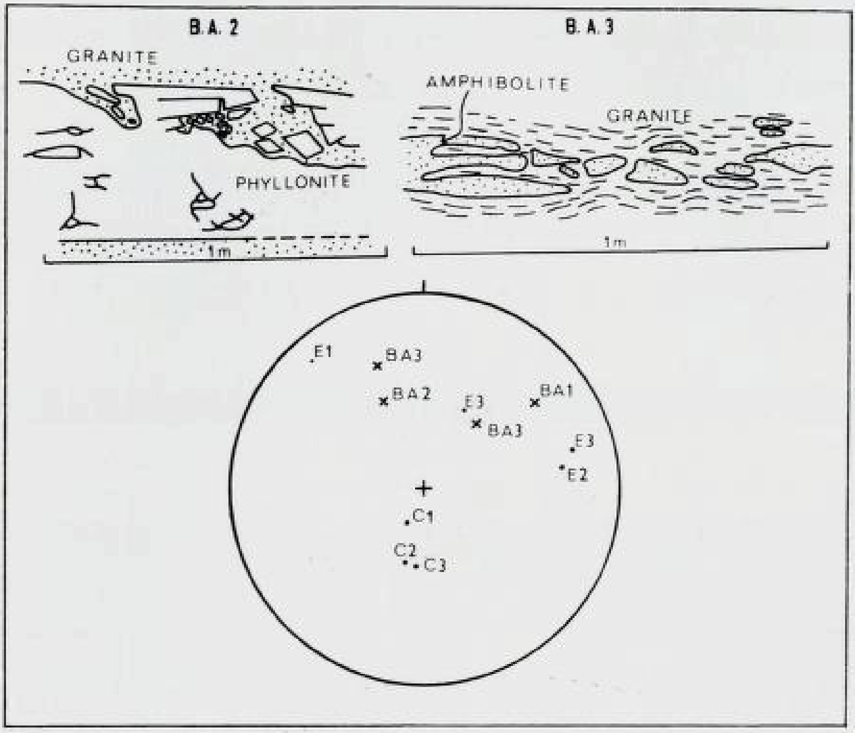

Fig. 5. F_1 boundinage structures and their calculated extension (E) and contraction axes (C). *Crosses* - long axes of the boudins.

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

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

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

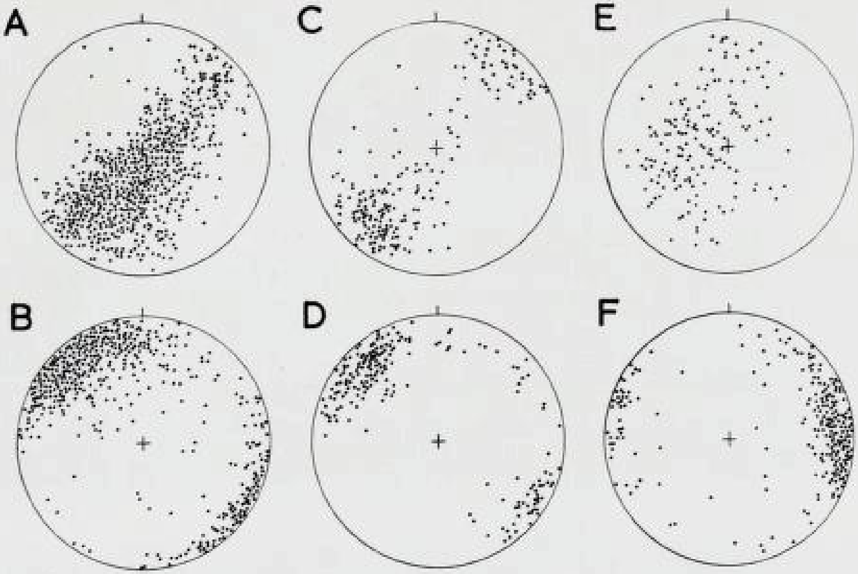

Fig. 6. Stereographic plots of the minor structures of the Bygdin area. A. S_1 poles. B. F_1 fold axes and lineations. C. S_2 poles. D. F_2 fold axes and lineations. E. S_3 poles. F. F_3 fold axes and lineations.

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

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

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

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

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

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

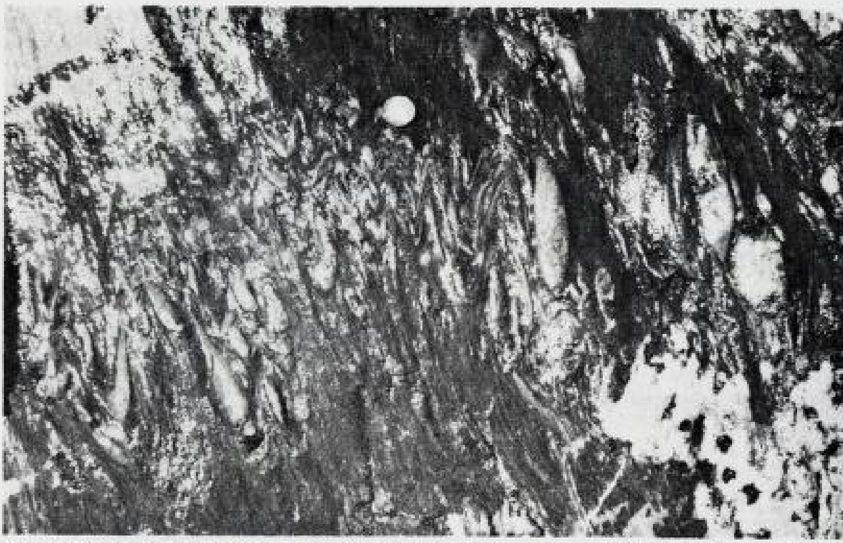

Fig. 7. Plane of pebble flattening (S_1) folded by F_2 minor folds with the production of an S_2 axial plane crenulation cleavage.

(C) Second Movement Phase (F_2).

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

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

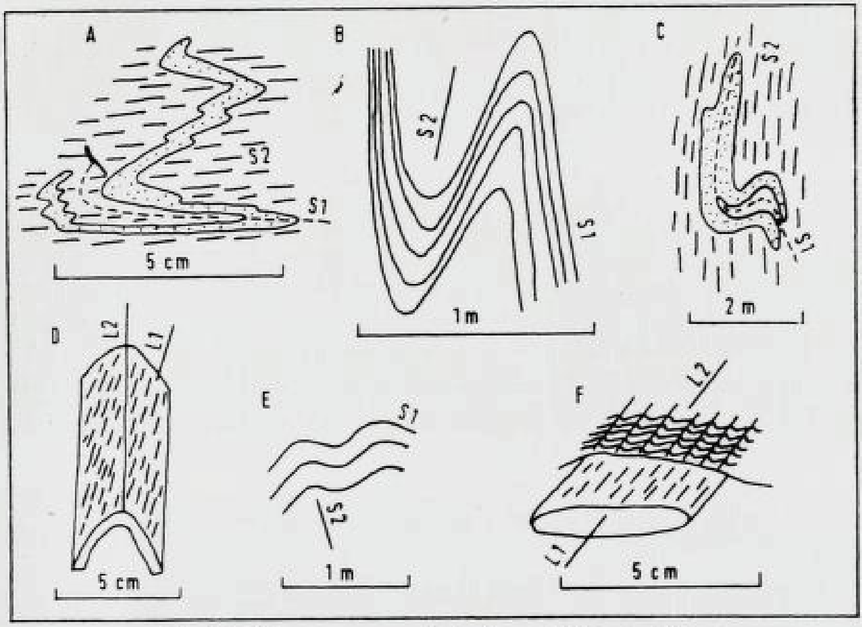

Fig. 8. Minor structures of the Second Movement Phase.

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

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

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

(ii) L_1 has controlled the orientation of L_2.

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

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

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

(D) Third Movement Phase (F_3).

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

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

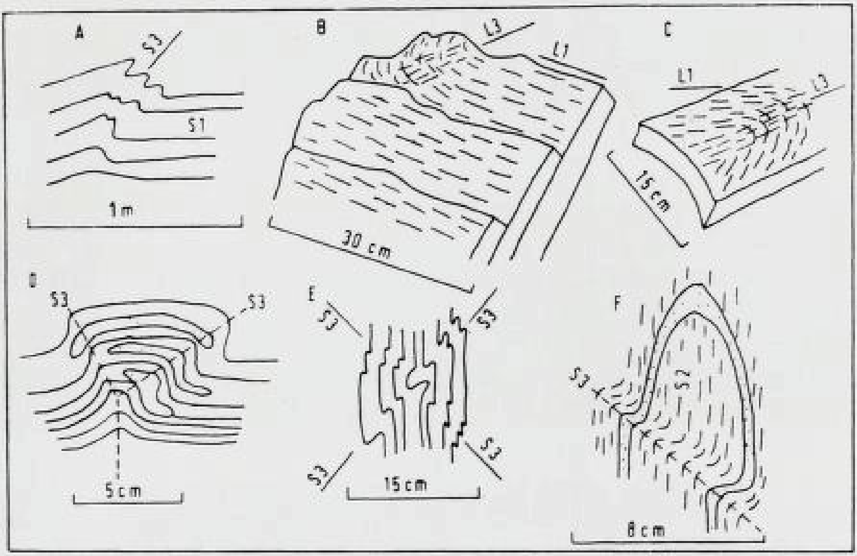

Fig. 9. Minor structures of the Third Movement Phase.

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

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

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

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

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

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

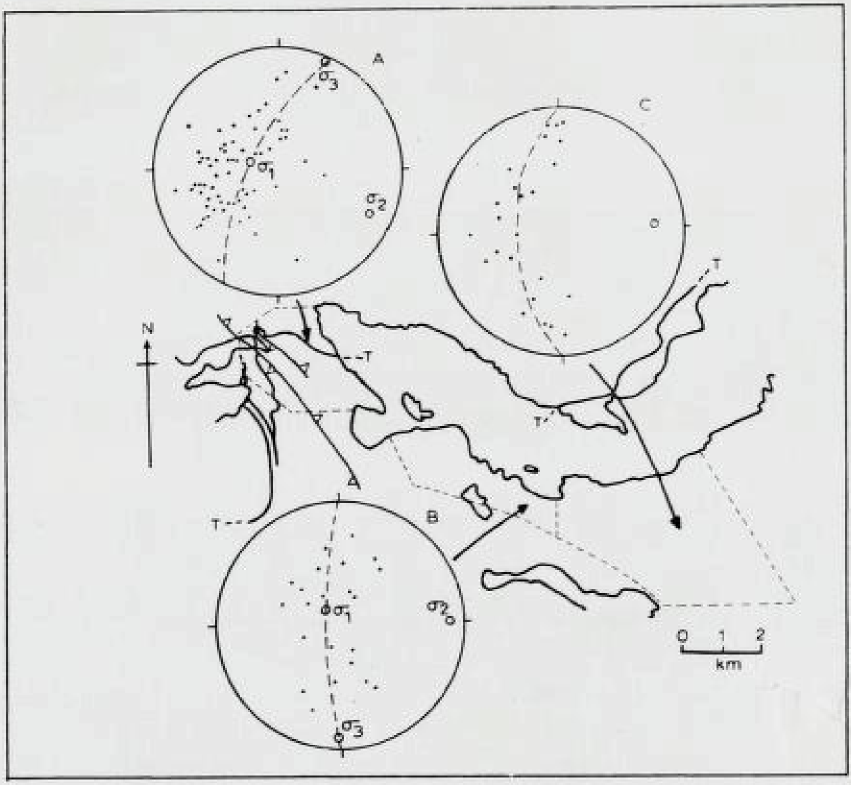

Fig. 10. F_3 pole girdles from three sub-areas at Bygdin with their calculated stress axis.

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

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

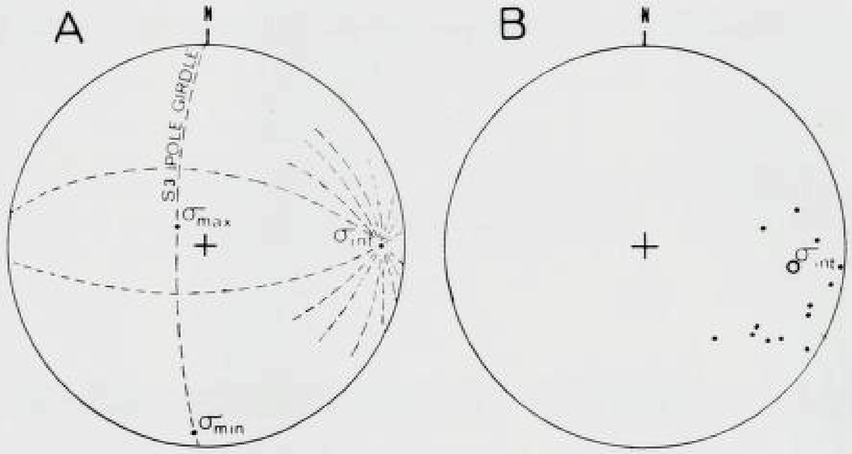

Fig. 11. A) Calculation of the stress axes from the S_3 pole girdle. B) Stereographic plot of the a_3-directions and the regional intermediate principal stress axis (σ_{int}) of the Third movement phase.

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

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

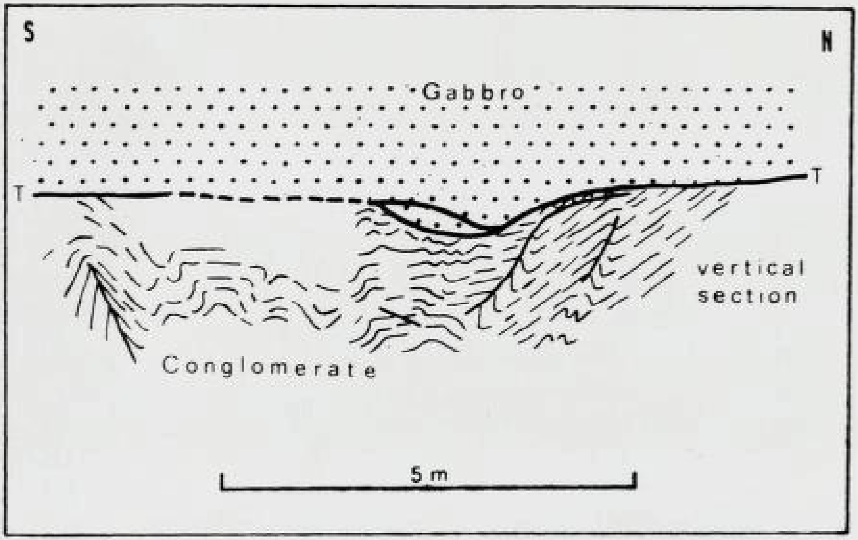

Fig. 12. Basal thrust plane truncating F_3 box folds on the Bitihorn crags.

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

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

(E) Fourth Movement Phase (F_4).

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

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

5. Major Structure

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

(A) Basal Thrust Zone.

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

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

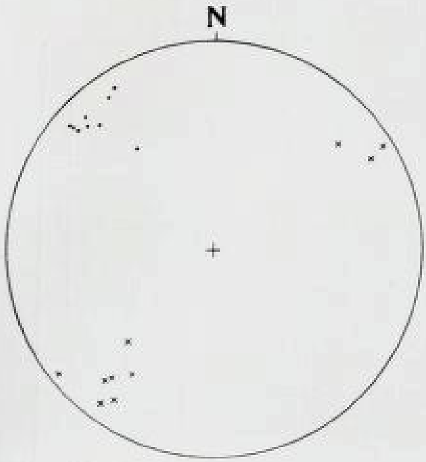

Fig. 13. Stereographic plot of the F_2 major structures. *Dots* - fold axes. *Crosses* - poles to axial planes.

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

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

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

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

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

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

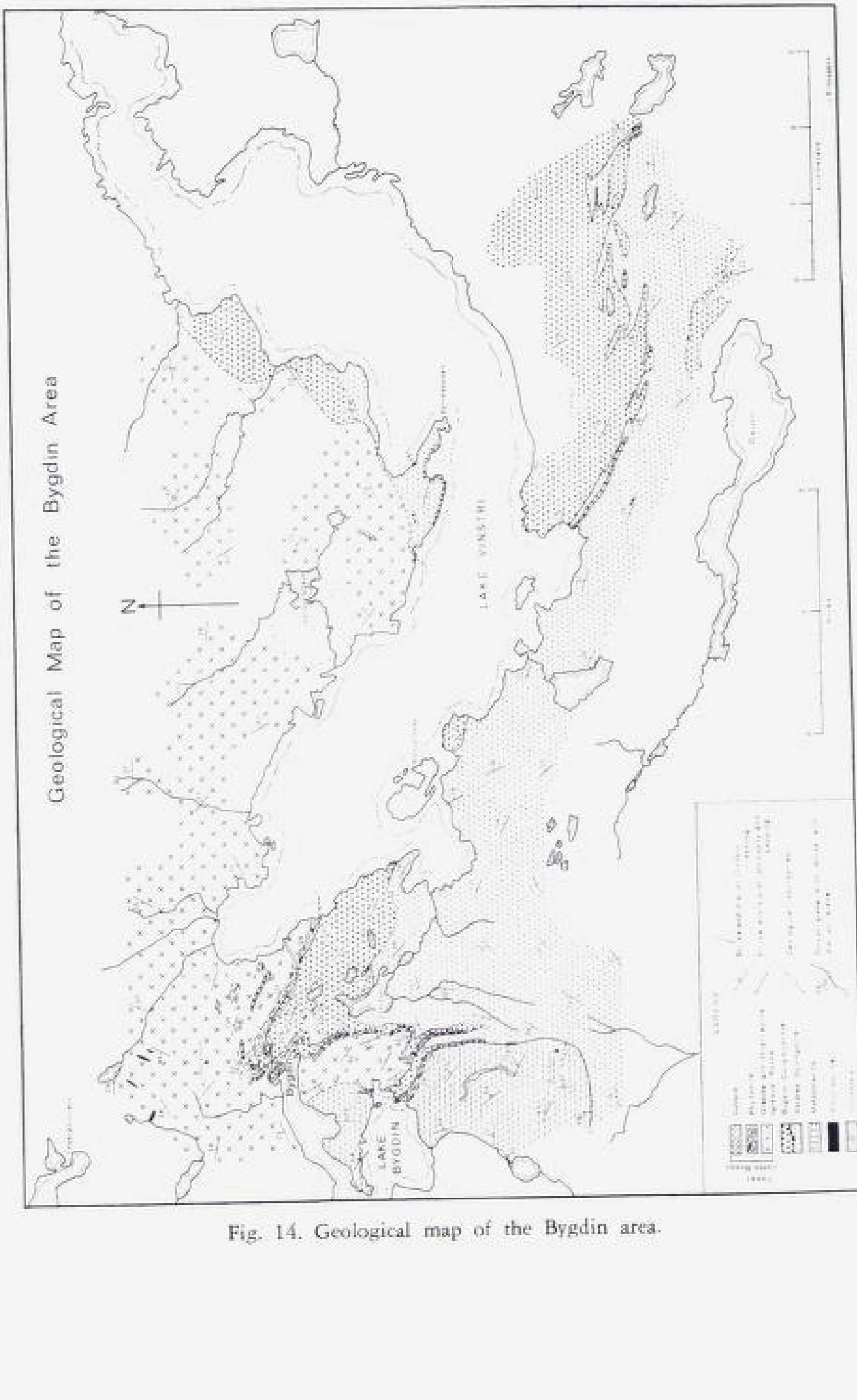

Fig. 14. Geological map of the Bygdin area.

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

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

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

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

(B) Major Second Folds.

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

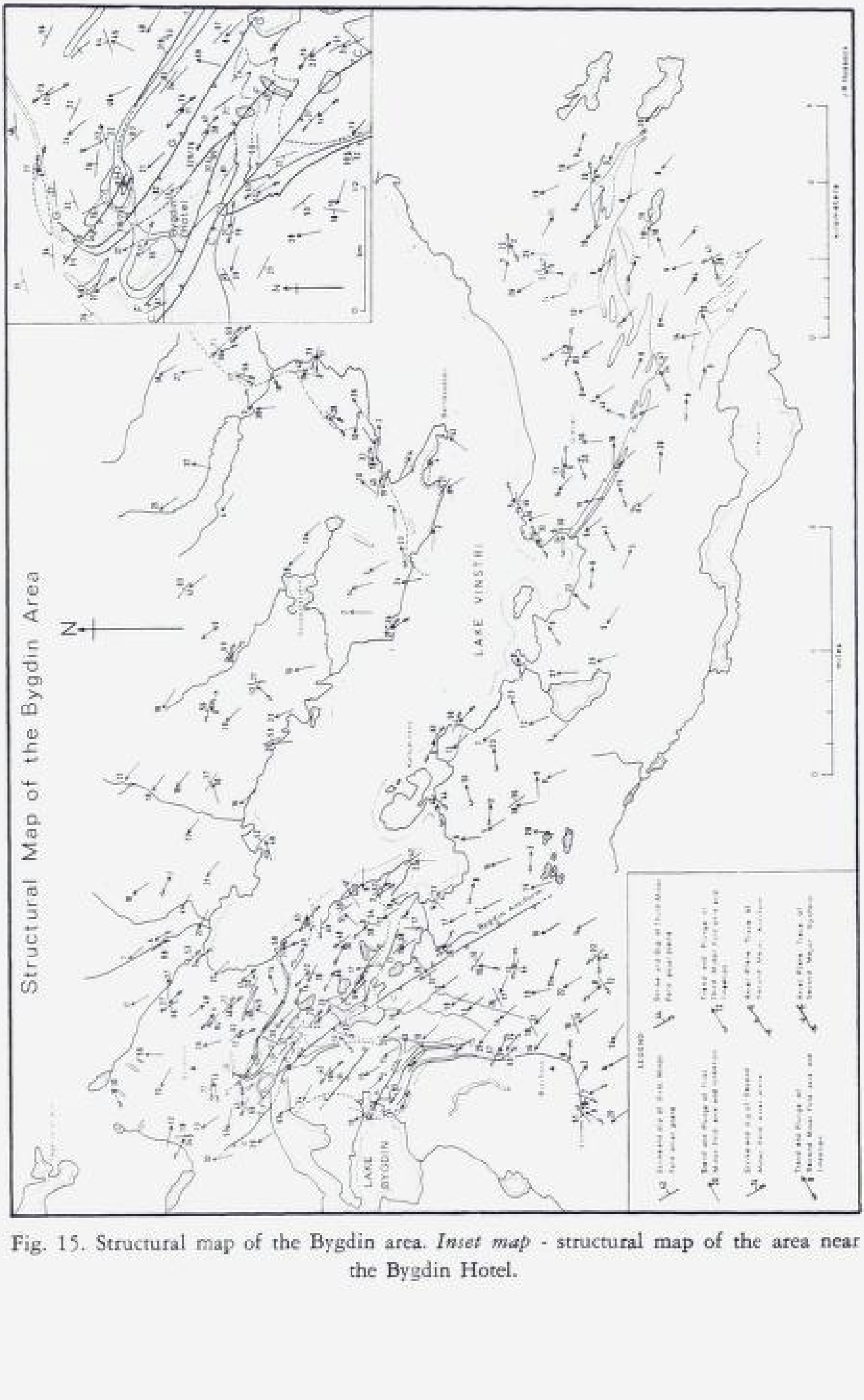

Fig. 15. Structural map of the Bygdin area. *Inset map* - structural map of the area near the Bygdin Hotel.

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

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

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

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

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

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

6. Metamorphic history

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

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

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

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

7. Conclusions

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

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

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

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

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

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

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