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The sub-Caledonian Unconformity on Hjelmsøy – New Evidence of Primary Basement/Cover Relations in the Finnmarkian Nappe Sequence

M. RAMSAY, B. A. STURT & T. B. ANDERSEN

Ramsay, D. M., Sturt, B. A. & Andersen, T. B. 1979: The sub-Caledonian unconformity on Hjelmsøy – new evidence of primary basement/cover relations in the Finnmarkian nappe sequence. *Norges geol. Unders. 351*, 1–12.

A new and important occurrence of a primary stratigraphic unconformity beneath the Caledonian cover in the allochthon of the Finnmarkian segment of the North Norwegian Caledonides is described. This unconformity, on the island of Hjelmsøy, is excellently although only locally preserved as a geological accident. The basal lithology above the unconformity is an orthoquartzite. This is overlapped by meta-arkoses containing clasts of the crystalline basement rocks. The effects of Caledonian tectonothermal processes produce a structural and tectural convergence of both members along the cover/basement interface, which may blur the precise contact. The Caledonian structural and metamorphic development of Hjelmsøy is also briefly discussed, as is the regional significance of the unconformity.

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Introduction

Recently, Ramsay & Sturt (1977) have described a major unconformity beneath the Klubben Psammite Group on the island of Kvaløy in West Finnmark. The substrate to the Klubben Psammite is a multicomponent gneiss complex of Precambrian age. This complex had undergone a protracted sequence of Precambrian deformational and metamorphic events, followed by uplift and erosion prior to the deposition of the Caledonian cover sequence. The cover sequence, according to Ramsay and Sturt, had been subjected to polyphasal deformation and metamorphism in the upper greenschist facies during the Finnmarkian Phase of Caledonian development. The basement gneisses also experienced both structural and metamorphic reworking during this Caledonian deformation. In analogous situations where Caledonian strains were high and the unconformity is no longer preserved, the gneisses were flattened and a variety of mylonitic gneisses and banded blastomylonitic lithologies were produced. Ramsay & Sturt (1977) drew attention to how, in the wider Finnmark region, the contact between the basal part of the Klubben Psammite Group and the subjacent gneisses has generally been the focus of very high shear strains and mylonite production during Caledonian deformation. In consequence, displacement frequently occurred at such contacts, resulting in the effective uncoupling of the cover sequence from its pre-Caledonian substrate. At this stage the uncoupled portions of the cover sequence became independent units and continued their translation histories as separate thrust nappes.

Ramsay et al. (1979) have discussed this particular problem in some detail and have demonstrated that the entire Finnmarkian nappe sequence, from Lyngenfjord in the south to North Cape in the north, is characterised by repetitions of a tectonostratigraphic pairing of basement and cover units (see also Zwaan & Roberts 1978). In the allochthon the pattern is a simple stacking of nappes each of which contains representatives of the same Caledonian cover sequence on a variable pre-Karelian basement. The cover sequences contain a well-differentiated metasedimentary sequence which although not uniformly developed throughout the region, can be correlated with the type Finnmarkian stratigraphy of the island of Sørøy (Ramsay 1971).

The thin and less extensive parautochthonous nappes at the base of the nappe pile exhibit a more diverse lithostratigraphic content, reflecting the geology of the nappe substrate. In West Finnmark the Vendian-Cambrian autochthon is less than 300 m thick and many thrusts have penetrated this cover. These nappes, therefore, comprise either single assemblages of Caledonian or Karelian rocks or basement-cover pairings in which the basement is pre-Karelian gneiss while the cover is either low-grade Karelian rocks or Caledonian metasediments. The Pre-Karelian gneiss complexes contain variable contents of paragneiss and orthogneiss, the latter showing a considerable composition range, and some of the gneiss units reveal indications of different patterns of tectono-metamorphic evolution.

Until recently the only well-preserved example known of a primary stratigraphic unconformity between basement and cover was the Stangnes locality on Kvaløy (Ramsay & Sturt 1977). During reconnaissance work in the most northern part of the Finnmarkian Zone, in the summer of 1978, the authors discovered another excellently preserved example of the unconformity at the base of the Klubben Psammite Group on the northern part of the island of Hjelmsøy (Fig. 1).

Description of the Hjelmsøy unconformity

The northwestern part of Hjelmsøy comprises a series of high-grade gneisses (Fig. 1). Reconnaissance studies reveal these rocks to be essentially a series of granitic and granodioritic orthogneisses, which are variably garnetiferous. These gneisses form part of a multicomponental complex and contain xenoliths of earlier, often basic phases. They are themselves cut by granitic pegmatites, aplites and dykes and sheets of basic igneous rocks, now in the condition of amphibolites bearing both biotite and garnet. The gneisses preserve evidence of polyphasal Precambrian deformation and a recrystallisation history which pre-dates the deposition of the overlying cover sequence. In places, however, as the contact with the cover sequence is approached, considerable Caledonian reworking of the gneissose basement can be identified with attendant

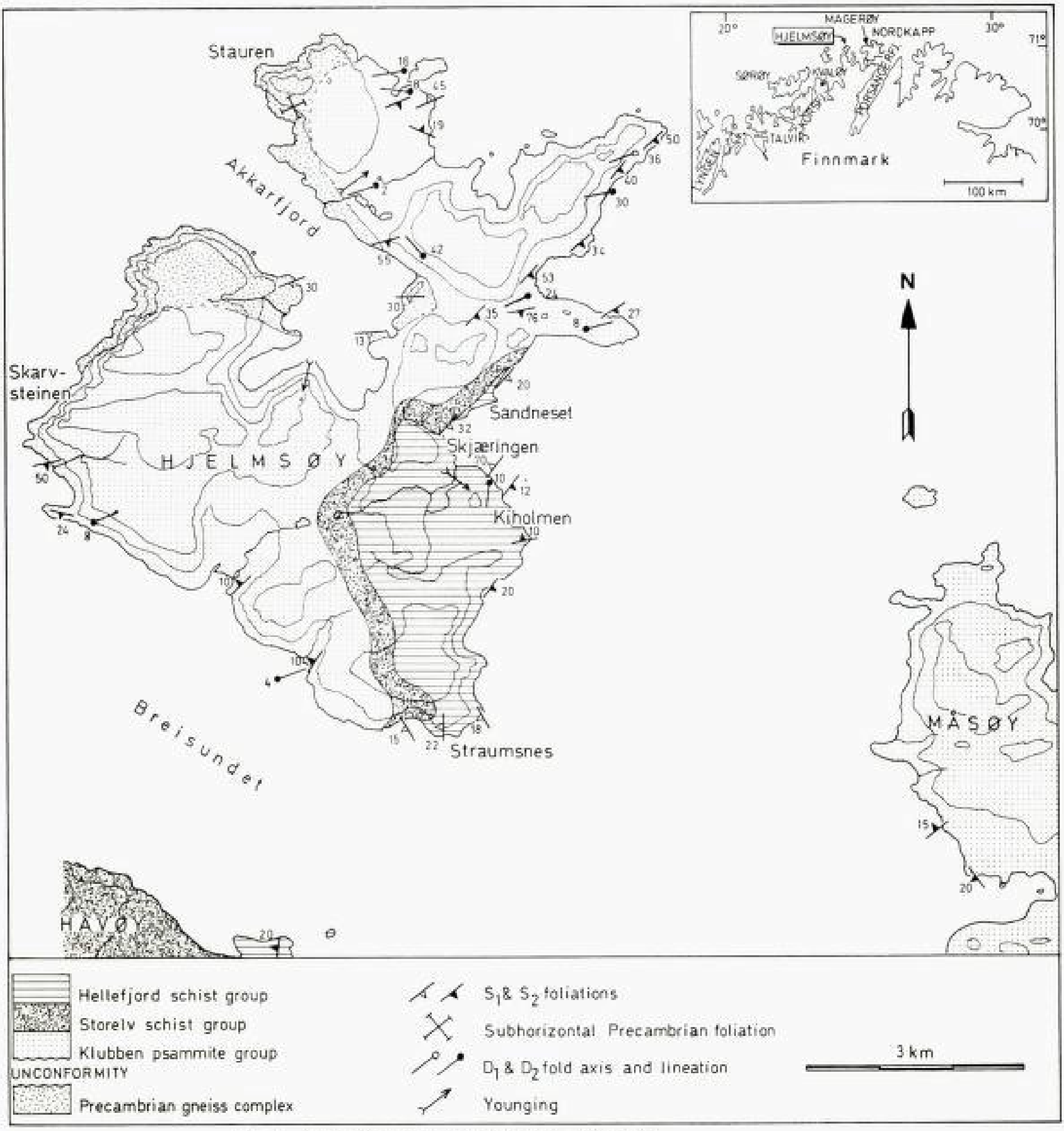

Fig. 1. Geological map of Hjelmsøy, Finnmark.

recrystallisation in amphibolite facies. In this situation both the cover assemblage and the basement gneisses have developed strong Caledonian fabrics, with the result that it is extremely difficult to pin-point the precise boundary surface, e.g. the contact zone east of Akkerfjord.

When the contact is traced west of Akkerfjord, it descends the cliffs at Skarvsteinen (Fig. 1), where it is dramatically exposed as a well-preserved primary stratigraphic unconformity (Fig. 2). The surface of the unconformity

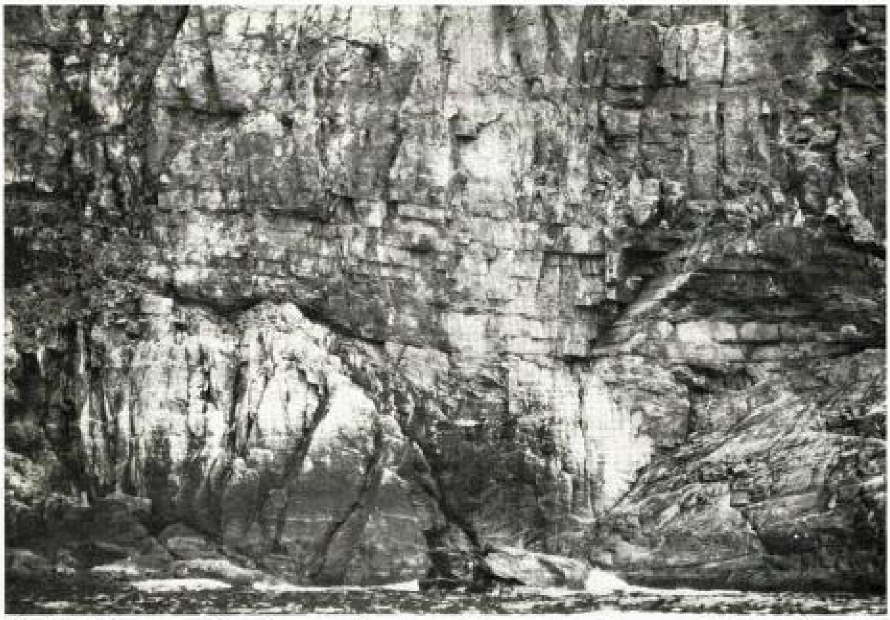

Fig. 2. Primary stratigraphic unconformity at Skarvsteinen, between Klubben Psammite Group (above) and Precambrian orthogneiss complex (beneath). Note that the contact is very sharp, but represents an undulating erosion surface. The orthogneiss has undergone Precambrian reworking and flattened streaks and lenses of pegmatite are cut by the unconformity. Above the unconformity the lower part of the cover sequence is a well-bedded orthoquartzite with progressive overstep onto upstanding irregularities in the erosion surface. In the photograph the upper part of the cover sequence is of meta-arkose and irregular developments of migmatitic neosome can be discerned. The cliff section shown in the figure is approximately 25 m in height.

is gently undulating, cuts into the basement gneisses and abruptly truncates the structures in this substrate. The gneisses at this locality are a series of variable granitic gneisses with more mafic horizons. These rocks have a strong foliation defined by parallel orientation of biotite and muscovite and appear to have been deformed prior to the injection of a suite of granitic pegmatite sheets. The pegmatites display the effects of strong flattening strains in the form of marked pinch-and-swell structures, and in some instances are now seen as isolated lenses lying in the foliation. This deformation sequence clearly predates the unconformity, as flattened pegmatite lenses are truncated at angles of up to 30 degrees by the overlying basal quartzite beds of the Klubben Psammite Group. The lowest member of the Klubben Psammite at this locality is a fairly pure orthoquartzite. This has a well-bedded aspect with the lowermost 10 m comprising beds up to 1 m in thickness which progressively overstep on to higher levels of the irregular erosional surface (Fig. 2). In other localities, where the quartzites have been examined in detail, bedding is well-preserved and current-bedding has been observed in two places, the sequence younging upwards away from the unconformity (Fig. 1).

On the eastern side of Akkerfjord a gritty facies of the lower part of the Klubben Psammite Group is preserved in a small recumbent infold within the

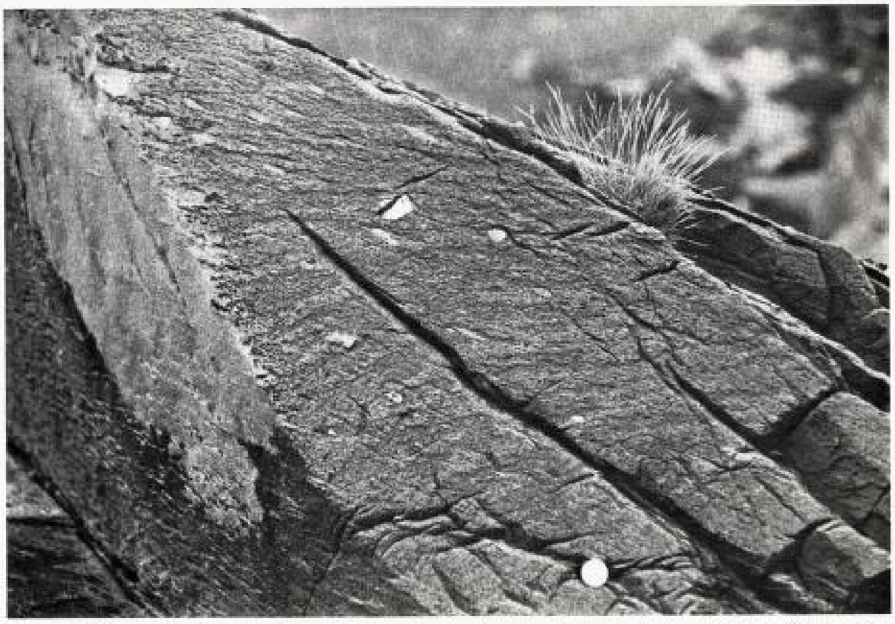

Fig. 3. Gritty lithology in the Klubben Psammite on the east side of Akkerfjord. This contains small pebbles of gneiss, granite and vein quartz. Coin measures 2.5 cm across.

gneisses, 10 m below the main contact. This lithology includes many small pebbles of material derived from the underlying gneisses in addition to vein quartz (Fig. 3). Numerous quartz veins do, in fact, cut the basement gneisses and many of them appear to have been emplaced prior to the deposition of the cover sequence. Occasional cobbles, up to 20 cm in diameter, are present and these vary from gneisses bearing the imprint of a pre-Caledonian foliation to virtually undeformed granite (Fig. 4).

From the above, there would appear to be little doubt that the original unconformable contact between Late Precambrian sediments of the Caledonian depositional cycle and the older crystalline basement is preserved in the island of Hjelmsøy. This forms a somewhat irregular surface such that different lithologies of the cover sequence are in contact with the basement. Pebbles derived from the basement complex contribute to a conglomeratic facies in the lower part of the cover sequence.

General characteristics of the metasedimentary cover sequence

The metasedimentary succession occurring above the Hjelmsøy unconformity is comparable in part with the stratigraphic sequence of the island of Sørøy. On Hjelmsøy, however, the standard Sørøy succession is incomplete. The Klubben Psammite, Storelv Schist and Hellefjord Schist Groups are present in typical developments, but both the Falkenes Marble Group and the Åfjord Pelite Group are absent. The omission of these two units is also a feature of the stratigraphic pattern in the Kobbfjord – western Magerøysund area some 30 km

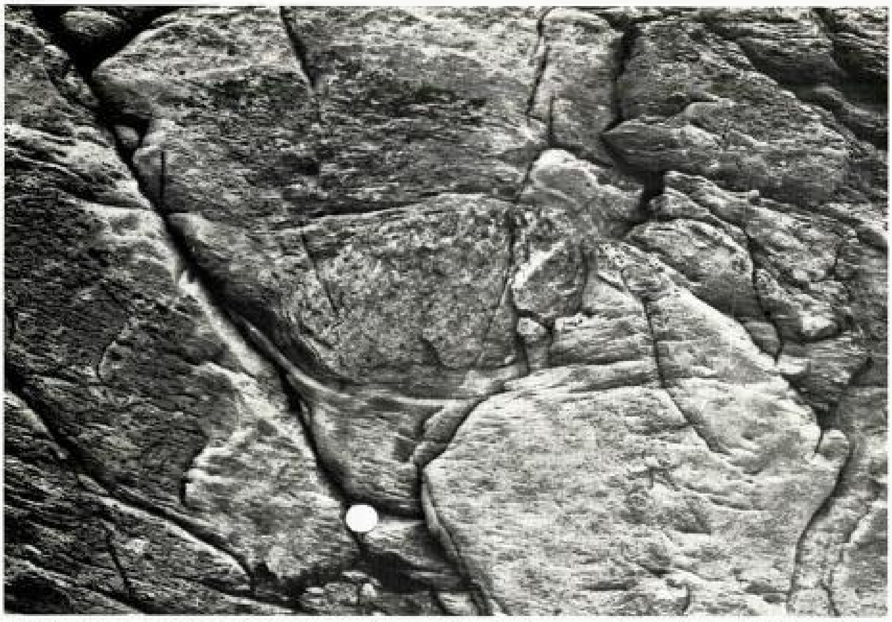

Fig. 4. Large cobble of basement granite in gritty horizon near base of the Klubben Psammite, east side of Akkerfjord. Coin measures 2.5 cm across.

further east, and would appear to represent a primary sedimentary facies variation in this northeastern part of the depositional basin. Although geographically only some 60 km along strike from Sørøy, the Hjelmsøy – Kobbfjord rocks probably occur in a major nappe which underlies the Seiland/Sørøy Nappe (Ramsay et al. 1979). Originally, therefore, they had a much greater separation.

The *Klubben Psammite Group* on Hjelmsøy consists of an alternating sequence of quartzites and meta-arkoses with subordinate variably garnetiferous pelitic schists. The quartzites are in many cases relatively pure orthoquartzites, with occasional heavy-mineral concentrations along bedding planes and with two observed examples of current bedding (Fig. 1). Transitions are observed between the orthoquartzites and meta-arkoses, although the latter tend to dominate the sequence. Within the meta-arkoses, especially where they are coarse-grained, clastic grains of feldspar, quartz and occasional small pebbles of granite and gneiss have been observed. Generally, however, as the result of elevated metamorphic grade (high amphibolite facies) during the Finnmarkian orogenic phase, the meta-arkoses show significant effects of migmatitization and considerable quantities of quartzo–feldspathic neosome have been produced. The neosomal material ranges from vague rootless and topless veins and segregations through discrete, medium-grained cross-cutting granite veins to pegmatite sweats, segregations and intrusive bodies. The migmatites developed in the Klubben Psammite of Hjelmsøy are very similar to those described from the Gjesvær area of Magerøy (Ramsay & Sturt 1976). On the neighbouring island of Måsøy (Fig. 1), strongly migmatized Klubben Psammite is again found and the neosomal material is seen to post-date foliated amphibolite sheets.

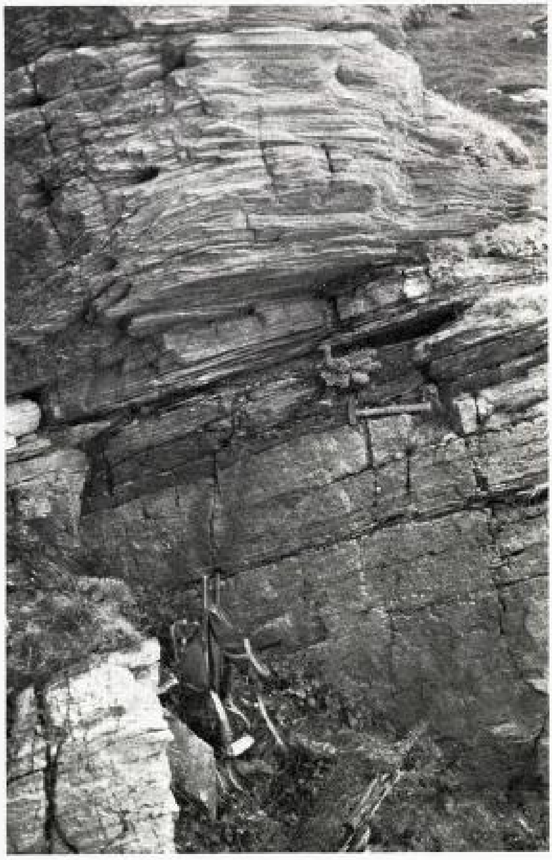

Fig. 5. The unconformity in a highly flattened state. The basement gneisses (bottom) have acquired the Caledonian foliation (S_1), which is well-developed in the orthoquartzite at the base of the Klubben Psammite (top); east side of Akkerfjord.

Tracing the contact with the gneissic substrate eastwards from Skarvsteinen meta-arkoses overlap the basal quartzite and rest directly upon the underlying crystalline basement. As the meta-arkoses are now almost invariably in a migmatitic condition it is difficult to locate the precise position of this contact over much of its length. Another factor which plays a significant role in the obliteration of the contact is the magnitude of the Finnmarkian strains in the Klubben Psammite and the subjacent gneisses (Fig. 5). During the late stages of the main, syn-foliation deformation phase on Hjelmsøy, here termed D_1, the rocks experienced strong flattening strains which had the effect of producing a marked reworking in the gneisses near the contact. This in turn produced a prominent foliation and a new gneissic banding which was subsequently overprinted by the Caledonian migmatization; and there was a similar neosomal development to that observed in the cover meta-arkoses. In the contact zone east of Akkarfjord there has therefore been a structural and metamorphic convergence between the migmatized reworked basement gneisses and the meta-arkoses of the cover sequence. The basement gneisses away from the reworked zones might also be expected to bear the imprint of superimposed Caledonian migmatization, but such effects are difficult to distinguish from neosomal material produced during Precambrian metamorphism.

On Hjelmsøy, the preservation of the unconformity in unequivocal terms is due to the fortunate combination of a refractory orthoquartzitic lithology and a somewhat reduced level of straining. Elsewhere, although mylonite is not developed as is generally the case over the region, the products of the high grade of regional metamorphism obscure the original nature of the contact. In the one other published example of the basement-cover unconformity, on Kvaløy, the psammites of the cover sequence display only upper greenschist facies metamorphism.

The *Storelv Schist Group* directly overlies the Klubben Psammite Group and the contact is of a transitional nature. The rocks progressively take on a darker colour and a more schistose appearance with increasing biotite content. Typically, the Storelv Schist is a rusty-weathering, coarse-grained mica schist composed essentially of varying proportions of quartz, biotite, muscovite and garnet. Flakes of mica several centimetres across are not uncommon. Garnet is frequently idioblastic and ranges up to 5 cm in size, as can be seen at Skjæringen. The Storelv Schist is variably affected by the migmatization and pegmatitic neosomal sweats are quite common. The lower part of the Storelv Schist is more quartz-rich, whilst the facies with large garnets is better developed in the upper parts of the unit. A clear division into a Lower and Upper Storelv Schist, which can be made on Sørøy, the more southern parts of the Finnmarkian zone and in the Kobbfjord area, cannot be made on Hjelmsøy.

On Hjelmsøy, the Storelv Schist Group is directly overlain by the metaturbidite sequence of the *Hellefjord Schist Group* (Roberts 1968). The precise contact with the Storelv Schist Group is not exposed, but as one traces the approximate boundary inland no evidence for the presence of the intervening Falkenes and Åfjord Groups (Ramsay 1971) can be found. As no evidence has been found for a tectonic break at this level, the lack of these two groups can probably be explained in terms of regional variations in the environment of sedimentation. The Hellefjord Schist on Hjelmsøy is typically a flaggy, green-coloured sequence of metagreywackes and pelites (Fig. 6), very similar to that of Sørøy. The flaggy aspect relates to the pronounced interbanding of greywacke layers with pelitic and semi-pelitic mica schists, and the sub-parallelism of the S_1 schistosity and the bedding. Both the metagreywackes and the mica schists display considerable thickness variations. None of the metagreywacke beds observed was thicker than 1.5 m although generally they range from 30–60 cm in thickness, while the interbanded mica schist horizons are usually somewhat thinner (10–20 cm). Graded bedding occurs in the metagreywackes, and shows a consistent sense of younging upwards from the Storelv Schist Group.

The rocks of the Hellefjord Schist Group show few effects of the migmatization so characteristic of the underlying lithologies. They are, however, very rich in garnet. No evidence of minerals such as staurolite or aluminium silicates has been recorded either from the Hellefjord Schist or from the underlying lithologies; it must be noted, however, that all observations were made during a relatively rapid field reconnaissance and thin-section studies have not been carried out.

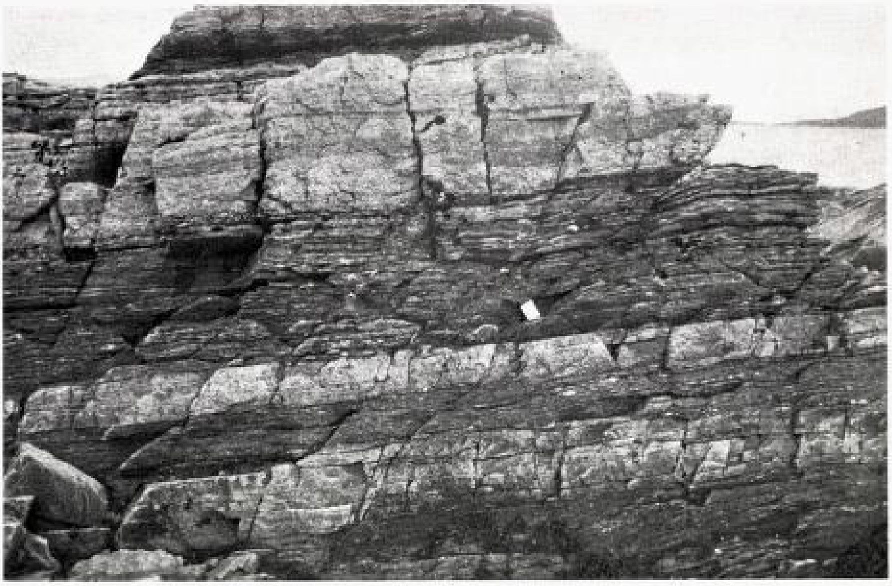

Fig. 6. Interbedded greywacke and semi-pelite at the base of the meta-turbidite sequence of the Hellefjord Schist Group at Sandneset (scale, matchbox 5 cm).

Structural observations

Evidence for a polyphasal pattern of Precambrian structural evolution in the basement gneisses has already been given, and the Caledonian reworking of these gneisses has been discussed. Two principal phases of Caledonian deformation have been recognised on Hjelmsøy, both of which were of a penetrative nature. Minor folds of the first phase (D_1) are only rarely observed and these are tight to isoclinal folds of bedding with a pronounced axial planar mica foliation (S_1). One reason for the lack of preserved D_1 structures is the strong transposition produced by D_2 structures. The anatexis in the Klubben Psammite and Storelv Schist lithologies can be shown to be both syn-tectonic and post-tectonic with respect to D_1. Where migmatization is highly developed, this too has had the effect of obliterating or partially obliterating the D_1 structures.

Excluding the large D_2 Hjelmsøy Synform discussed below, the map (Fig. 1) presents a simple homoclinal pattern of southeasterly dipping S_1 foliation. No evidence of macroscopic D_1 folds has been discerned. The existence of a mesoscopic, D_1, coupled fold of the surface of unconformity can be identified on the eastern shore of Akkerfjord (Fig. 1). All that now remains is an elliptical enclave of Klubben Psammite Group rocks, some 400 m in length, within the gneisses, a remnant of the synclinal part of the fold. Strong attenuation of the long limb of the complementary anticline created markedly fissile fabrics in both basement and cover which, as previously described, became the focus of later injection by leucocratic neosome.

The second major deformation (D_2) produced a pronounced crenulation cleavage (S_2) which in many areas strongly transposes earlier planar structures,

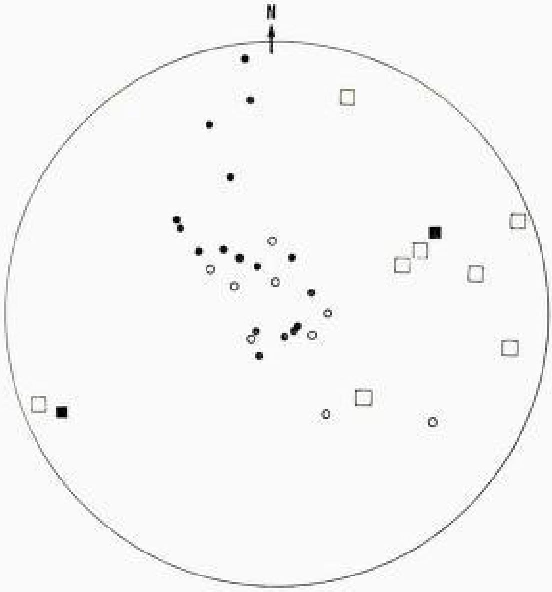

Fig. 7. Stereographic plot of structural elements, Hjelmsøy.
○ Poles to S_1
● Poles to S_2
□ Minor fold axes
■ Intersection lineations S_1/S_2

and effectively becomes the dominant planar structure in the rocks. A plot of the minor structures recorded from Hjelmsøy is given in Fig. 7. Study of D_2 minor folds suggest that fairly elevated metamorphic temperatures continued well into the D_2 deformation phase as minor developments of syn-tectonic neosomal materials are frequently encountered in association with the structures of this phase. This is also a feature which has been noted from the Gjesvær area of Magerøy (Ramsay & Sturt 1976).

The megascopic structural pattern of Hjelmsøy is dominated by a large recumbent synform of D_2 age. This fold has an axis which plunges gently towards SSW and an axial plane inclined at about 15° towards the east. The fold, which we propose to call the *Hjelmsøy Synform*, closes westwards and gives rise to an extensive tract of inverted strata in the southern part of Hjelmsøy and on the nearby island of Havøy. The D_2 minor folds bear a non-coaxial relationship to the Hjelmsøy Synform (Fig. 7), although they show a normal consistent vergence towards the hinge of the main fold on both the right-way-up and the inverted limb. The hinge-zone of the main structure is characterised by a steeply dipping to vertical enveloping surface of S_0/S_1, particularly well illustrated in the area of Kiholmen.

Conclusions

On Hjelmsøy a primary stratigraphic unconformity between the Finnmarkian cover sequence and the underlying pre-Caledonian basement gneisses is preserved. The relationships, in places, are rendered complex through the combined effects of Caledonian D_1 strains and migmatization, affecting both cover sequence and reworked basement. The Caledonian lithostratigraphic sequence of Hjelmsøy is not as complete as elsewhere in Finnmark, and this is explained in terms of primary regional variations in the sedimentary environment.

The migmatization can also be observed in the cover sequences of the islands of Havøy and Måsøy, and would appear to correlate with the migmatization of the Klubben Psammite in the Gjesvær area of Magerøy. The Caledonian structure of Hjelmsøy is dominated by an eastward-facing recumbent synform of D_2 age which produces an extensive zone of inverted strata.

The Hjelmsøy unconformity represents only the second published record of such primary contact in the allochthonous nappes of the Kalak Nappe Complex, although P. Bowden and D. M. Ramsay (pers. comm. 1977) have discovered a further example near Talvik in Altafjord. At the latter locality, although quite strong strains have affected the contact zone, rocks of the Klubben Psammite Group truncate the foliation in dioritic gneisses of the substrate. Such examples of primary stratigraphic relationships between cover and basement, especially where the superimposed effects of Caledonian deformation and metamorphism are clearly displayed, are highly significant.

In recent years it has become apparent that the Kalak Nappe, once regarded as a single major unit, comprises several nappes and tectonic slices derived from Caledonian, Karelian and pre-Karelian assemblages (Roberts 1974, Jansen 1976, Ramsay & Sturt 1977, Zwaan & Roberts 1978). Ramsay & Sturt (Sturt et al. 1978, Ramsay et al. 1979) have demonstrated that this complex pattern of Caledonian nappes falls into a rational pattern, characterized in the allochthon by repetitions of older and younger sequences. These couplets are interpreted as being originally slices of older basement with a cover of Caledonian sediments, despite the common occurrence of tectonic junctions between them. This was confirmed by the finding of a section of the original unconformable contact preserved on the island of Kvaløy (Ramsay & Sturt 1977). This primary contact has, of course, suffered some modification in the ensuing Caledonian deformation, but not to the extent where mylonite was generated, as in the more common situation. Further support for this interpretation came from pre-Caledonian ages obtained from one of the basement gneiss units at Korsfjord (Ramsay & Sturt 1977). The unconformity described in this paper, together with the one reported from Talvik, add further support to this thesis.

Six allochthonous Finnmarkian nappes comprising such pre-Karelian basement with Caledonian cover have been identified between Lyngenfjord and Porsangerfjord, together with six smaller and less extensive parautochthonous nappes (Ramsay et al. 1979). The latter have a more varied composition reflecting the geology of the local autochthon and include non-metamorphosed Caledonian sediments, Karelian rocks and couplets of pre-Karelian basement with Karelian or Caledonian covers.

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Geology of the Altenes Area, Alta–Kvænangen Window, North Norway

EIGILL FARETH

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Precambrian supracrustal and intrusive rocks of the Raipas Group of supposed Karelian age are unconformably overlain by Vendian to ?Lower Cambrian sedimentary rocks, which have in turn been overridden by allochthonous rocks of the Kalak Nappe Complex. The Raipas rocks are divided into 3 formations; the lithologies of these units and of the autochthonous cover sequence are described. The structural history of the area includes a Precambrian, post-Karelian to pre-Vendian, deformation episode as well as Caledonian events. It is shown that in the north-western part of the window, the Raipas 'basement' together with its autochthonous cover has been affected by high-angle, SE-directed thrusting and imbrication during the Caledonian orogeny.

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Introduction

The Altenes peninsula is situated on the northern side of Altafjord, Finnmark (Fig. 1), and covers an area of about 130 km^2. Most of the ground lies between 200 and 400 m a.s.l. The degree of outcrop is excellent, except in the easternmost part, but exposures of good quality are mostly confined to the coast.

The area was mapped in 1972–73 for ore prospecting purposes as part of the North Norway project of the Geological Survey of Norway (NGU) on a scale of 1:21,500 (Fareth 1975), in connection with a geochemical survey (Krog & Fareth 1975). Some of the regional implications of this mapping and of investigations in a neighbouring area were discussed by Roberts & Fareth (1974). Additional data from Altenes were collected by the author on a short visit in 1977. Previous accounts of the geology are confined to brief remarks by Holtedahl (1918, p. 105–108), and rock sample descriptions and a fragmentary map by M. G. Legg (1961, unpublished data in NGU's files).

A 400^g compass was used in the present investigation; for dip values, 100^g = vertical.

Geological setting

This description concerns the north-eastern part of the Alta–Kvænangen window of Precambrian and overlying autochthonous rocks of Vendian – ?Lower Cambrian age in the Altenes area (Fig. 1). The major part of the area is underlain by low-grade sedimentary and volcanic rocks which are correlated with the Raipas Group rocks in Alta, of supposed Karelian age (Barth & Reitan 1963). Similar rocks are found in the Komagfjord window to the north-east (Reitan

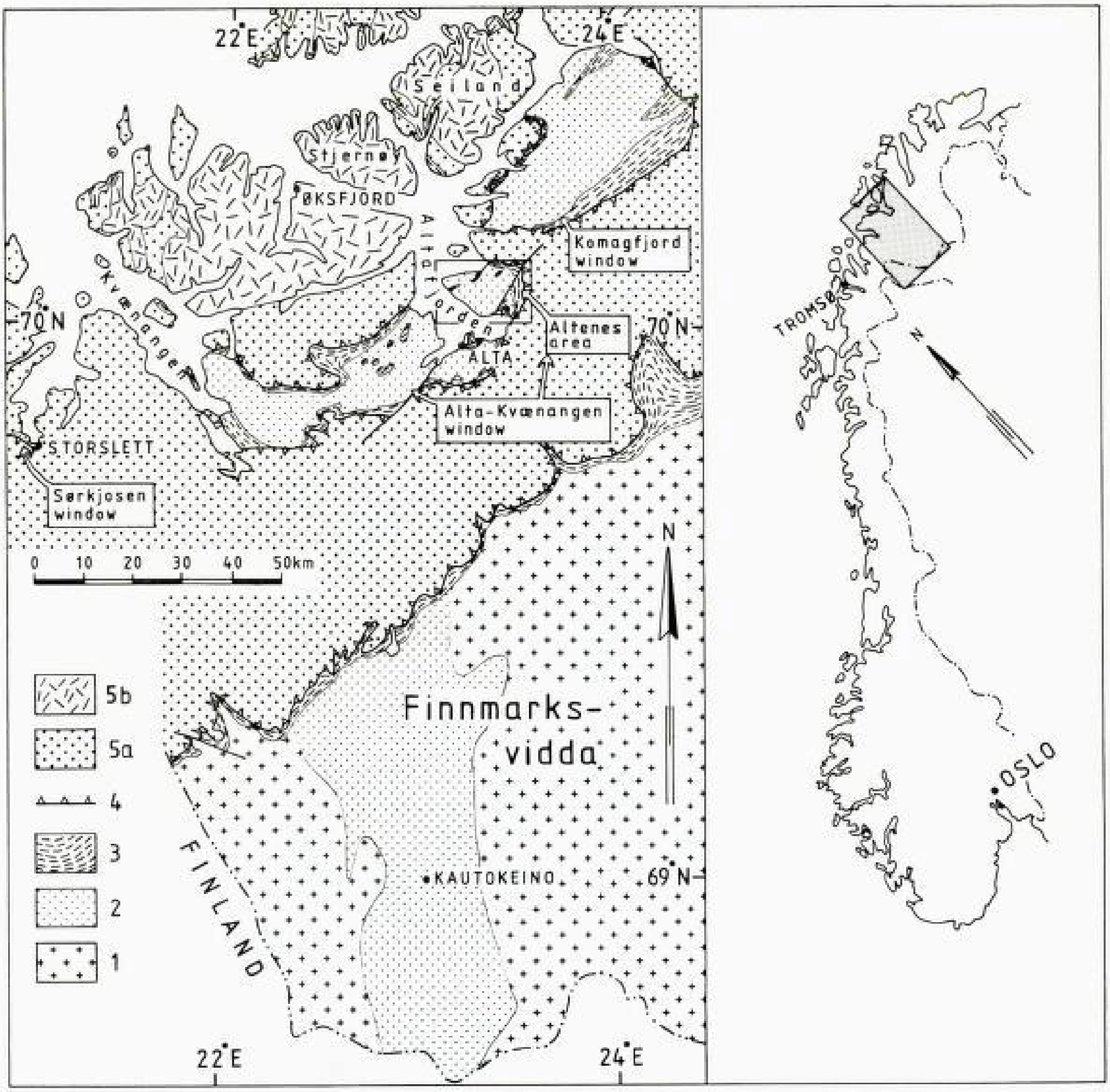

Fig. 1. Regional geology and location map. (1) Precambrian basement rocks older than Raipas Group; (2) Raipas Group and probable correlatives; (3) Autochthonous/parautochthonous Late Precambrian–Cambrian rocks (Dividal Group, Bossekop Formation, Borras Formation and probable correlatives); (4) Thrust plane; (5a) Kalak Nappe Complex (= Reisa Nappe Complex), locally including Tierta and Gaissa Nappes at base: mainly supracrustal rocks; (5b) Kalak Nappe Complex: igneous rocks of the Seiland petrographic province. Based on Skålvoll (1972), Bjørlykke & Fareth (1973), Roberts (1973), Fareth & Lindahl (1974), Roberts & Fareth (1974), Jansen (1976), Zwaan (1976) and T. Torske (pers. comm. 1977).

1963), the small Sørkjosen window to the south-west (Zwaan 1976), and in a N–S trending belt on Finnmarksvidda. On Finnmarksvidda, the Raipas rocks are bordered by other, presumably older, Precambrian rocks. The Precambrian rocks in the windows and on Finnmarksvidda are overlain unconformably by

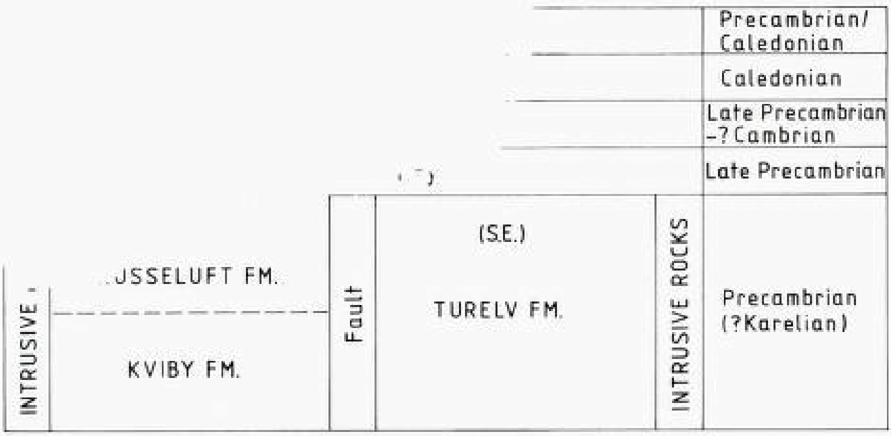

Fig. 2. Schematic representation of the principal tectono-stratigraphical elements in the Altenes area.

weakly metamorphosed sedimentary rocks including tillite, ranging in age from latest Precambrian (Vendian, in some areas possibly older) to Cambrian (Roberts & Fareth 1974). The uppermost tectono–stratigraphic unit is the Kalak Nappe Complex, equivalent to the Reisa Nappe Complex west of Kvænangen. These complexes are composed of plutonic and metamorphic rocks of Precambrian to Lower Paleozoic age (Roberts 1974, Zwaan 1976).

Tectono-stratigraphy

The tectono–stratigraphical relationships between the rock units are shown in Fig. 2.

The Precambrian rocks below the unconformity can in broad terms be correlated with the Raipas rocks at Alta. The latter have been variously named Raipas System (Dahll 1868), *Avdeling* (Division), Formation, Group, Series and Suite. The present author feels that the usage of Holtedahl & Dons (1960), Zwaan et al. (1973) and Gautier (1977) is preferable, and will extend their Raipas Group to include the equivalent rocks in the Altenes area.

A major fault trending SW–NE from Russeluft to Leirbotnvatnet splits the Raipas Group into two outcrop areas (Plate 1). On the north-western side of the fault two formations are recognised, the names of which are informal: the *Kviby formation* composed of basic volcanics, and the *Russeluft formation* composed of epiclastic sediments. The boundary between the two formations is tectonised, but in one coastal exposure (described in a later section) there is evidence suggesting a depositional relationship between them. The Kviby and Russeluft formations can be correlated with parts of Gautier's (1977) Lower and Upper Raipas sub-groups, respectively.

The Raipas rocks on the south-eastern side of the fault are more poorly exposed and have more complicated structures than those on the north-western side. Lithologies corresponding with those of both Lower and Upper Raipas

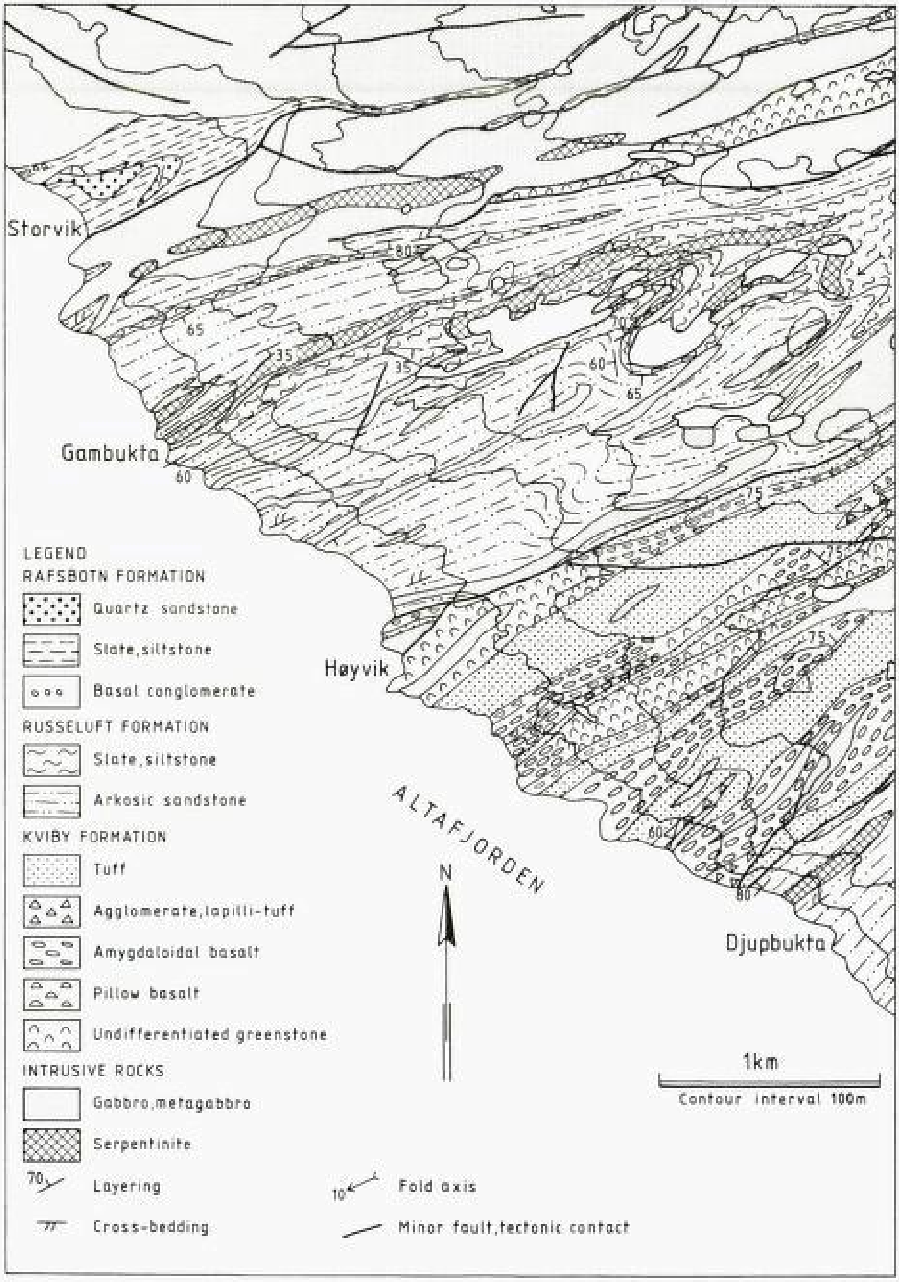

Fig. 3. Detailed geological map of the area around Høyvik (see Plate 1 for location).

have been found, but it has not been possible to map them separately. The informal term *Turelv formation* is used to denote all the Raipas rocks in this part of the area.

Due to marked changes in lithology and structures across the Russeluft-Leirbotnvatnet fault, it is assumed that the fault has a Precambrian origin, as

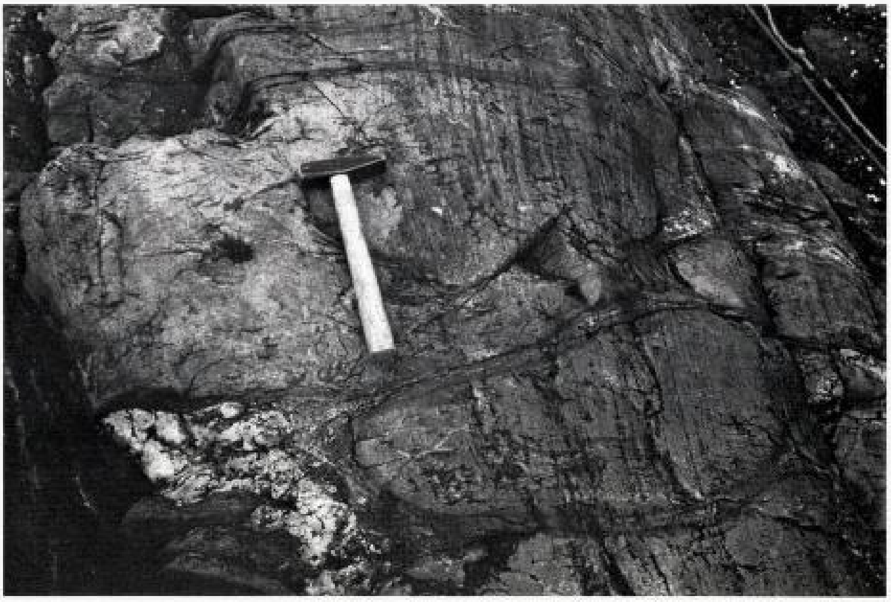

Fig. 4. Pillow lava, Høyvik. Note fine-grained marginal zone. Length of hammer handle 30 cm.

illustrated in Fig. 2, but definite proof for this is lacking. It has also been active in Caledonian time, deforming both the Rafsbotn formation autochthon and the Kalak Nappe Complex.

Description of rock units

KVIBY FORMATION

The Kviby formation is composed of low-grade metamorphic, extrusive volcanic rocks. From their mineralogy and field appearance the rocks have been classified as basaltic, although no analyses of bulk chemistry were made for this investigation. The formation crops out in three belts which are considered to represent approximately equivalent stratigraphies which have been tectonically separated. The thickness of the south-eastern belt, not accounting for possible repetition of layers, varies from about 600 m in the north-east to about 1500 m in the south-west, including a 100 – 200 m thickness of intrusive bodies in most sections. The other belts are thinner, and the rocks are generally more deformed there than in the south-eastern belt. The distribution of the rock-types varies in different areas. In the north-east part of the south-eastern belt pyroclastic rocks predominate, whereas lavas constitute a substantial portion of the coastal section in the south-west (Fig. 3).

Pillow basalt occurs at several levels. The most extensive layer, which is about 50 m thick, has been traced for about 3 km along the contact against gabbro east of Langvatnet. The width of the pillows is generally 0.3–1.0 m. In the most favourable localities, primary features such as fine-grained marginal zones (Fig. 4) and amygdules are fairly well preserved. In general, the pillow structure is rather obscure due to tectonic deformation. It has not been found suitable for geopetal determination.

Amygdaloidal basalt without pillow structure has a fine-grained, dark green matrix, and amygdules filled with quartz. The amygdules are ovoid or irregular, their size is 2–30 mm and they compose 5–30% of the rock.

Agglomerate, lapilli-tuff and tuff grade into each other. The coarser pyroclasts are predominantly greenstone, seldom diabase or gabbro. Maximum size observed is about 30 cm, but usually the pyroclasts are less than 10 cm in length (Fig. 5). The proportion of lapilli and blocks to tuff material varies greatly. Tuffs with only scattered lapilli are widespread.

To a large extent, especially in the mountains, poor exposures and tectonic deformation preclude the observation of primary structure in the Kviby formation, and the rocks have been mapped as undifferentiated *greenstone*.

The mineralogy of the volcanics is of low-grade metamorphic origin. The matrix of the rocks is composed of albite + amphibole + epidote ± chlorite ± biotite ± leucoxene. The amphibole is in most cases colourless to light green tremolite/actinolite. Megacrysts in the pyroclastic rocks include plagioclase, epidote, olive-green amphibole and quartz.

RUSSELUFT FORMATION

The Russeluft formation is composed of medium- to thick-bedded sandstone of chiefly arkosic composition, with subordinate slate, siltstone and conglomerate. Like the Kviby formation, outcrops are located in three separate belts, the easternmost of which is the most extensive, its thickness being around 3500 m. Tectonic repetition of layers may, however, have occurred in this belt. Cross-bedding structures, which are widespread throughout the formation, invariably show younging directions towards the south-east.

Arkosic sandstone. The sandstone is light grey, frecuently with a reddish tinge. Somewhat darker varieties with a shade of green also occur. Quartz, plagioclase and microcline, in descending order of abundance, are the major minerals. The amount of phyllosilicates is small. Occasional samples are rich in quartz and grade into quartz sandstone. The rocks have an undeformed appearance in many localities, but microscopic examination usually reveals some micro-brecciation. Quartz as a rule shows undulatory extinction. Average grain size is commonly about 0.5–1.0 mm. Bed thickness is between 0.1–1.0 m. Depositional structures include ripple marks, graded bedding and cross-bedding. Soft-sediment deformation of foresets in cross-bedded units can be seen in some localitites (Fig. 6). Cross-bedding has been used to determine way-up in all the outcrop areas of the Russeluft formation. Examples of graded bedding are few and show fining upwards. In a few zones scattered pebbles of quartz and coarse-grained gneiss are found in the sandstone. The pebbles are rounded and about 2–6 cm across.

Conglomerate (not shown on the map) occurs between Kvibyvatnet and Bannasgamvatnet. The most extensive zone is about 1 km long and up to 30 m thick. It consists of rounded to subangular, sometimes flattened, clasts of quartz

Fig. 5. Agglomerate with large clast of amygdaloidal basalt, Tømmerholmen. (Note miniature crag-and-tail caused by glacial erosion). Length of pencil 14 cm.

Fig. 6. Cross-bedded sandstone with foresets deformed prior to consolidation, 150 m southeast of Gambukta.

and feldspar. Average grain size is 0.5–1.0 cm. Quartz pebbles may be up to about 5 cm across. The rock grades into arkosic sandstone.

Slate, siltstone. Fine-grained sediments occur sparingly throughout the Russeluft formation. The most extensive zone is found in the central belt (Fig. 3) and attains a thickness of about 100 m. The rocks have a greyish colour and are composed of alternating layers of clayey and silty material varying in thickness from about 0.2–10.0 cm. They are frequently somewhat calcareous. Cleavage cutting layering is developed to a varying degree.

Contact metamorphic features are noted near intrusives, the sedimentary rocks taking on a hornfelsic appearance. Slates and siltstones are commonly partly assimilated by intrusive gabbros, particularly in the area between Kviby and Langvatnet. Growth of epidote and tremolite/actinolite in the sedimentary rocks is also noted as a contact metamorphic feature.

Dolomitic breccia (not shown on the map) is found in a small area 1.5 km north-west of Russeluft. It is composed of dolomitic and quartzo-feldspathic fragments in a dolomitic matrix. The outcrop area measures about 50 m × 400 m. Layering is not seen. The contacts with the surrounding arkoses have not been observed, and the origin of the breccia is unknown.

TURELV FORMATION

The term 'Turelv formation' is used for convenience for what might more correctly be termed 'undifferentiated rocks of the Raipas Group south-east of the Russeluft–Leirbotnvatnet fault'. This rock unit consists of both volcanic and sedimentary rocks, which may be stratigraphical equivalents of parts of the Kviby and Russeluft formations, respectively. The outcrops are separated into an eastern and a western belt by the large gabbroic body east of Turelva. The northernmost part of the western belt is included in Fig. 7. The stratigraphical thickness of the Turelv formation has not been determined.

Greenstone is fine-grained, massive or laminated. No recognisable primary volcanic structures have been observed. In many cases, it is difficult to distinguish between lavas and intrusives in the field. The main minerals of the greenstones are albite, light green amphibole and epidote. Chlorite, ore minerals, leucoxene and calcite occur in varying amounts. Some greenstone samples display relics of an ophitic or subophitic texture.

Dolomite. Massive and laminated dolomite grade into each other and occur most extensively in the hillside west of Grareinelva. In the massive dolomite, breccias resembling karst fillings occur. Minor chert bands up to 1 cm thick have been observed in the dolomite.

Phyllite. Dark-coloured pelitic rocks occur in zones up to 50 m thick. The rocks are distinguished from the slates in the Russeluft formation by having a phyllitic lustre, probably due to a slightly higher degree of metamorphism. The phyllosilicates in two samples were sericite and chlorite.

Quartz sandstone occurs in contact with dolomite, in zones up to 100 m thick. It is medium-grained, has a reddish colour and contains occasional zones of quartz conglomerate up to 10 m thick.

Arkosic sandstone has its main occurrence on the eastern side of the large gabbroic body east of Turelva, where a maximum apparent thickness of 700 m is exposed. The rocks are less well preserved than the arkoses of the Russeluft formation, and primary structures have not been found. Contact metamorphic effects such as hornfels, growth of epidote and tremolite/actinolite, and partial assimilation by gabbro, are seen in many localities.

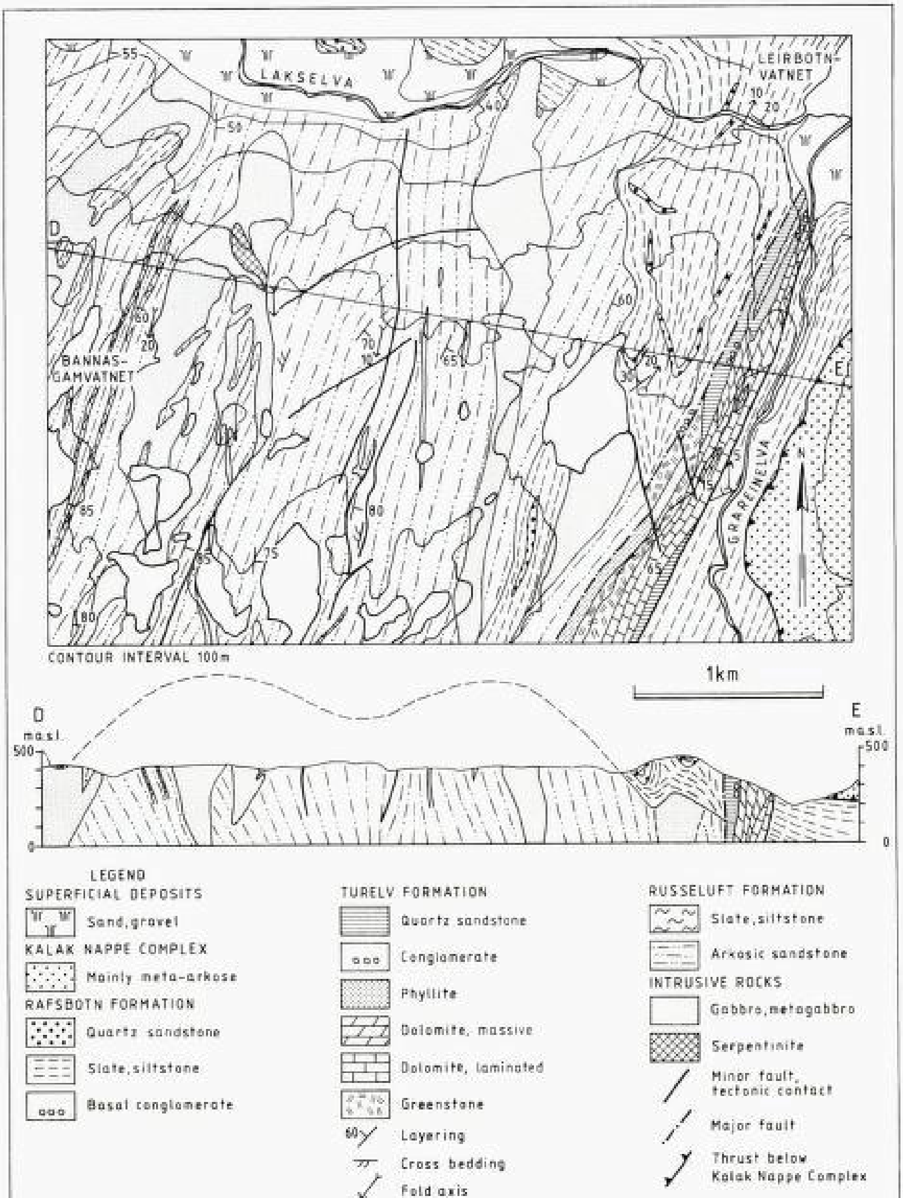

Fig. 7. Detailed geological map and profile of the Bannasgamvatnet–Leirbotnvatnet area. (See Plate 1 for location).

INTRUSIVE ROCKS

Bodies of intrusive, dominantly gabbroic rocks, constitute about a quarter of the exposed area of Precambrian rocks in the Altenes window. They are usually lenticular sheets or sub-concordant dykes and vary in dimension from the metre scale up to 12 km in length and several hundred metres thick. Individual bodies have nowhere been observed to cross formational boundaries.

Fig. 8. Columnar jointing in diabase, Tømmerholmen.

Gabbro, metagabbro, diabase. The rocks are mostly medium-grained, with a sub-ophitic texture. Saussuritisation and uralitisation have taken place to a varying degree. A small portion of the samples have unaltered pyroxene and plagioclase. A particularly fresh specimen from a locality 400 m north-west of Trollvatnet has orthopyroxene and clinopyroxene together with slightly zoned basic andesine. A diabase dyke cutting the layering in pyroclastic rocks on Tømmerholmen displays well-preserved columnar jointing (Fig. 8).

Serpentinite. Ultramafic rocks have intruded the sedimentary sequence of the Russeluft formation. They occur as lenses with longest dimension up to 1.5 km, mainly in the fine-grained sedimentary rocks west of Langvatnet. Other ultramafic bodies are contained in gabbroic rocks. Most samples of ultramafic rocks have a mineralogy dominated by serpentine and chlorite. Relict textures suggest that the rocks were originally composed of olivine and poikilitic pyroxene. A sample from the ultramafic body north-east of Bannasgamvatnet contains about 3% plagioclase.

Tonalite (not shown on the map). A swarm of dykes, each dyke between 200–800 m long, up to 20 m thick and oriented approximately NW-SE, occurs in a restricted area in the north-eastern part of Altenesfjellet. The rocks have a porphyritic texture with euhedral crystals of quartz and zoned plagioclase set in a fine-grained quartzo-feldspathic matrix; compositionally they correspond to tonalite. The dykes cut through the bodies of metagabbro.

RAFSBOTN FORMATION

This formation is composed of low-grade sedimentary rocks, mainly slate and siltstone. The sediments were deposited unconformably on a substrate of

Fig. 9. Basal conglomerate, Rafsbotn Formation, Djupvik. Person standing on underlying gabbro. Dark fragments in conglomerate are gabbro; light fragments are quartz and quartzo-feldspathic rock. The contact between conglomerate and gabbro is approximately vertical.

deformed and eroded Raipas Group supracrustals and intrusives, and are autochthonous with respect to this substrate. The formation occurs in two different settings (see chapter 'Structural history'): —

1. The areas along the edges of the window in the north and east (Kviby – Leirbotnvatnet, Leirbotvatnet – Rafsbotn, Rafsbotn – Turelva) and two small outliers north of Bannasgamvatnet. Here the basal contact has gentle to moderate dips (Plate 1, profile segment B–C), except where disturbed by the Russeluft–Leirbotnvatnet fault. There is no evidence to indicate that the rocks of the Rafsbotn formation in these areas were deposited outside the present area.
2. Narrow zones in the north-western part of the Altenes area, between Storvik and Djupvik. The basal unconformable contact is vertical or steeply inclined in an inverted position (Plate 1, profile between A and Langvatnet). These occurrences of the Rafsbotn formation have been tectonically transported, together with their substrate, from a position probably not very far to the north-west.

The stratigraphical thickness of the Rafsbotn formation is estimated to be between 50 and 100 m.

Tillite is considered to represent the lowermost stratigraphical unit. It is confined to a single locality, described by Roberts & Fareth (1974), where it rests on Raipas rocks. Its maximum exposed thickness, according to observations made in 1977, is 4 m.

Conglomerate is found as a discontinuous layer along the basal contact of the Rafsbotn formation in localities where the tillite is absent. It is chiefly a quartz conglomerate, up to 1 m thick, with well-rounded pebbles which are 1–3 cm in size. The zones between Storvik and Djupvik, which rest on gabbro, have at their base a generally less well-sorted conglomerate with less well-rounded pebbles. Maximum exposed thickness is about 2 m. Many subangular blocks are found, frequently more than 10 cm in size. A large percentage of the clasts, in several localities more than 50%, is composed of gabbro (Fig. 9).

Slate, siltstone. These rocks, which in many localities are interbanded, constitute the dominant lithology of the Rafsbotn formation. The rocks may be laminated on a mm to cm scale, but homogeneous pelites also occur. The coloration is grey, green or red. A mottled red-green alternation is particularly characteristic. Slaty cleavage is prominent in most localities. Trace fossils have been found in siltstone at Mosenes (Roberts & Fareth 1974).

Quartz sandstone is medium-grained, grey or red and occurs in horizons which range in thickness from a few decimetres to several metres. These seem to be most frequent in the lower part of the formation. Layers of *conglomerate* with quartz and quartzite pebbles about 1 cm in size are found at a few localities. The layers are up to 1 m thick.

KALAK NAPPE COMPLEX

The Rafsbotn formation is overthrust by rocks of the Skillefjord Nappe of the Kalak Nappe Complex (Roberts 1974). The nappe rocks directly bordering the Altenes area are dominantly grey to greenish grey, flaggy meta-arkoses with subordinate mica schist. They contain pegmatite lenses, and isoclinal folds are common. No further account of these rocks will be presented here.

REGIONAL METAMORPHISM

From the mineralogy of greenstones and pelitic rocks it can be concluded that the degree of regional metamorphism in both the Raipas Group and the Rafsbotn formation is about the same and does not exceed low greenschist facies. There may be a slight difference between the Turelv formation and the other formations, as the pelitic rocks of the former have a phyllitic appearance, in contrast to the slates of the Russeluft and Rafsbotn formations. However, more refined investigations are necessary to determine whether this is significant. Whether the metamorphic mineral assemblages of the Raipas Group rocks date from Precambrian or Caledonian metamorphic events, or both, is not known.

Structural history

The Altenes area can be divided into three subareas which show different structural histories (*a–b–c*, Fig. 10). The diving line between subareas *a* and *b* is the Russeluft–Leirbotnvatnet fault; that between *b* and *c* is less precisely defined. It has been drawn at the north-western boundary of the south-eastern belt of the Kviby formation, but it could be somewhat further north-west.

In subareas *a* and *b* the sediments of the Rafsbotn formation overlie their

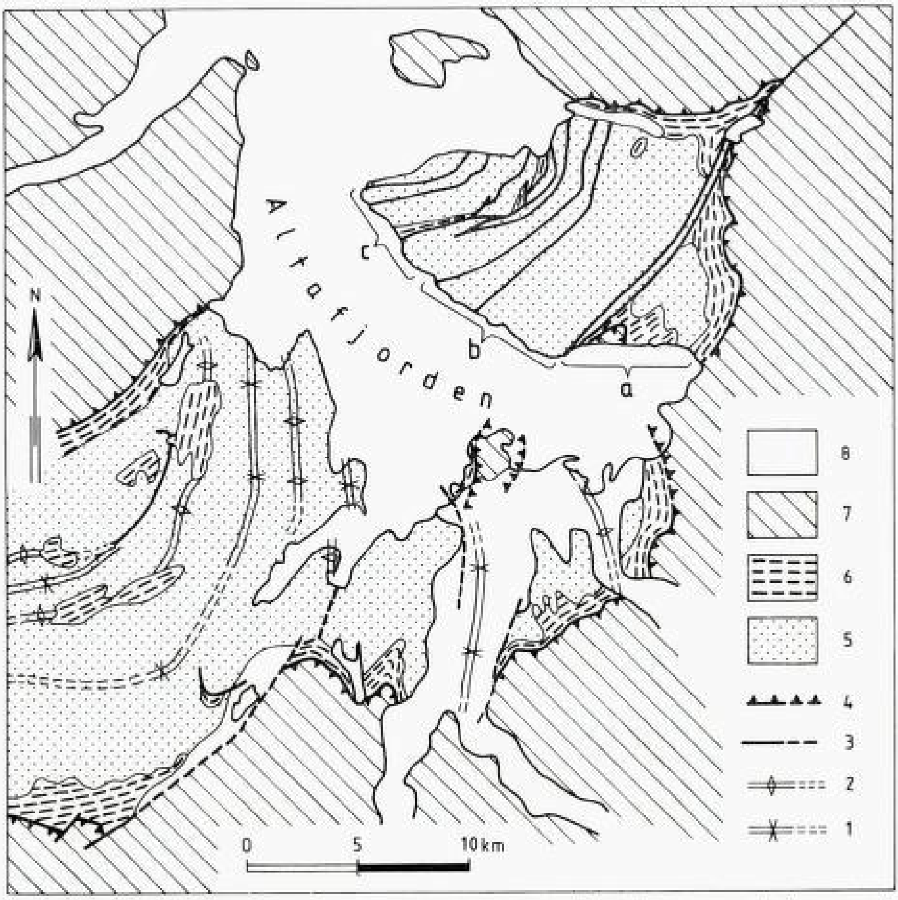

Fig. 10. Simplified map showing structural subdivision of the Altenes area (subareas *a*, *b* and c), and main structural features of the Altafjord region. (1) Syncline, dashed where inferred; (2) Anticline; (3) Fault, tectonic contact; (4) Thrust plane below Kalak Nappe Complex; (5) Raipas Group; (6) Bossekop, Borras and Rafsbotn Formations; (7) Kalak Nappe Complex; (8) Superficial deposits. Data from the southern side of the fjord modified from Gautier (1977).

substrate in a normal position, whereas in subarea *c* they occur in vertical or steeply inclined, overturned zones, sandwiched between slices of the substrate upon which they were deposited.

The layered rocks of the Raipas Group in general have steep, often vertical dips. In subarea *a*, large-scale folding is probably the main factor controlling the present distribution of the Raipas rocks. Open to tight folds with NE-SW to N-S trending axes and steep axial planes, and with amplitudes ranging from 0.1 m to 50 m have been observed, but poor outcrop has prevented mapping of individual large-scale folds.

In subareas *b* and *c*, mesoscopic folds in the Raipas rocks are rare and restricted to zones with extensive outcrops of slate, and to local areas in the vicinity of minor faults. The largest folds are found in slate-siltstone lithologies 1–2 km west of Langvatnet, where tight folds with steeply dipping axes have

Fig. 11. Contact between Russeluft and Kviby Formations, northern side of Djupbukta. East is towards the right. Tuff (Kviby Formation) immediately to the left of the person, who is standing on sandstone (Russeluft Formation). Thin bed of sandstone (light-coloured) in tuff is present in the left part of the picture.

amplitudes up to 500 m. No major isoclinal folds exist in the rocks belonging to the Russeluft formation in these two subareas. This is proved by the widespread cross-bedding structures, constantly showing younging direction towards the south-east.

Cleavage is poorly developed in the Raipas rocks. The sandstones and the bulk of the volcanic rocks are more or less massive. Cleavage transecting layering has been noted in slate and tuff. The scattered observations yield younging directions which correspond with those decuded from sedimentary structures.

In order to understand the major structure of subarea *b*, it is important to determine the nature of the contact between the Russeluft and Kviby formations. This is sharp wherever exposed. As might be expected in a tectonically disturbed area, the contact between the competent sandstones and the relatively less competent effusives shows signs of movement, the effusives being in general sheared and shattered immediately adjacent to the contact. The contact has accordingly been drawn as 'tectonic' on the geological map. However, observations at the coastal exposure of the contact on the northern side of Djupbukta between Høyvik and Russeluft suggest that the contact may have a depositional origin (Fig. 11). Here, the inverted basal lithology of the Russeluft formation, dipping 90° to the north-west, is an arkosic sandstone, below which is a tuff belonging to the Kviby formation. Within this tuff, 3 and 5 m away from the contact, are found two beds of arkosic sandstone, each half a metre thick. This kind of contact could tentatively be interpreted as a transitional one, the two sandstone beds being forerunners of the Russeluft formation and representing temporary cessation of a waning volcanic activity. If this interpretation is correct, the Raipas rocks in subarea *b* can be envisaged to constitute the north-western limb of a syncline whose south-eastern limb has been cut out by tectonic movements.

When discussing the major structures of the Raipas rocks in subareas *a* and *b*, we are mainly – although not exclusively – dealing with the effects of events that took place before the deposition of the Rafsbotn formation. In subarea *c*, the mode of occurrence of the Rafsbotn formation shows that in this north-western area, Caledonian events have played a major part. Here, slices of the Raipas substrate, partly with zones of the Rafsbotn formation sandwiched between them, have been piled upon one another in some kind of imbrication structure. This implies that the rocks in this subarea are not autochthonous. Probably their source area was not very far to the north-west. The generally NW-dipping tectonic contacts in subarea *c* are interpreted as minor, high-angle thrusts.

The structures described above differ from those on the opposite side of Altafjord. Here, Gautier (1977) has mapped a number of major synclines and anticlines, whereas complete folds on this scale are absent from subareas *b* and *c* in Altenes.

In both the Altenes area and the rest of the Alta–Kvænangen Precambrian window, the Raipas rocks display a varying strike trend. This may be due to some extent either to two separate phases or to a primary non-cylindricity of Precambrian folding. However, in the Altenes area, the curving trend of the thrusted zones involving Rafsbotn formation rocks in subarea *c*, which is reflected in a less conspicuous strike change in subarea *b* from NNE-SSW in the north to NE-SW in the south, could indicate that part of the deformation of the Raipas rocks is Caledonian, not only in subarea *c* but also in *b*.

Reitan (1963) found thrusts in the north-western part of the Komagfjord window which are similar in attitude to those described from Altenes. The structural position of his Kvalsund formation is similar to that of the Rafsbotn formation in subarea *c*, although no basal conglomerate has been described from the Kvalsund formation.

Folding is widespread in the Rafsbotn formation. The scale varies from about 1 cm to 100 m, the largest folds being found in thick quartzite intercalations. The folds vary in style from open to tight, curvilinear to chevron, and symmetrical to inequant. Also, directions of fold axes and axial planes show great variability. In large parts of the area the pelitic rocks have a steep to vertical cleavage trending NE-SW to N-S, which is axial planar to small-scale folding. The spacing between cleavage planes is in the interval 0.5 cm–2 cm. The observations in the outcrop areas of the Rafsbotn formation are not sufficient to provide a detailed structural history for this formation. The thrust plane beneath the Kalak Nappe Complex truncates folds in the Rafsbotn formation, proving that the folds pre-date this thrusting (D. Roberts, pers. comm.).

The following sequence of events is envisaged to have led to the present disposition of the rock formations in the Altenes area:

Precambrian: Deposition of Kviby, Russeluft and Turelv formations. Folding, intrusion, faulting, including the Russeluft–Leirbotnvatnet fault.

Vendian – ?Lower Cambrian: Deposition of Rafsbotn formation on a peneplaned surface.

Caledonian I: Folding of Rafsbotn formation rocks. Approximately SE-directed compression in Raipas basement, causing high-angle thrusting/imbrication in subarea *c*. In this subarea the Rafsbotn formation took part in the thrust movements. Compression possibly also caused some deformation of the Raipas rocks in subarea *b*. Younger folding in the cover rocks; this includes folds which deform the thrust plane beneath the Kalak Nappe Complex.

Caledonian II: Re-activation of Russeluft–Leirbotnvatnet fault, the block on south-eastern side sinking at least 300 m near Russeluft, displacement decreasing towards Leirbotnvatnet. The fault dies out north-eastwards in the Kalak Nappe Complex, outside the map area.

The Altenes area is thus not just a simple 'window' where erosion has exposed a basement. The Raipas rocks have taken part in the structural development of the area not only in Precambrian time but also after they had acquired their cover of younger sediments.

Finally, it should be noted that Torske (1978) has proposed an aulacogen-rift model for the Raipas rocks in the Alta–Kvænangen window and their correlatives on Finnmarksvidda. The studies which are now in progress in both areas will probably throw new light upon this part of the Baltic shield margin to which the Altenes area belongs.

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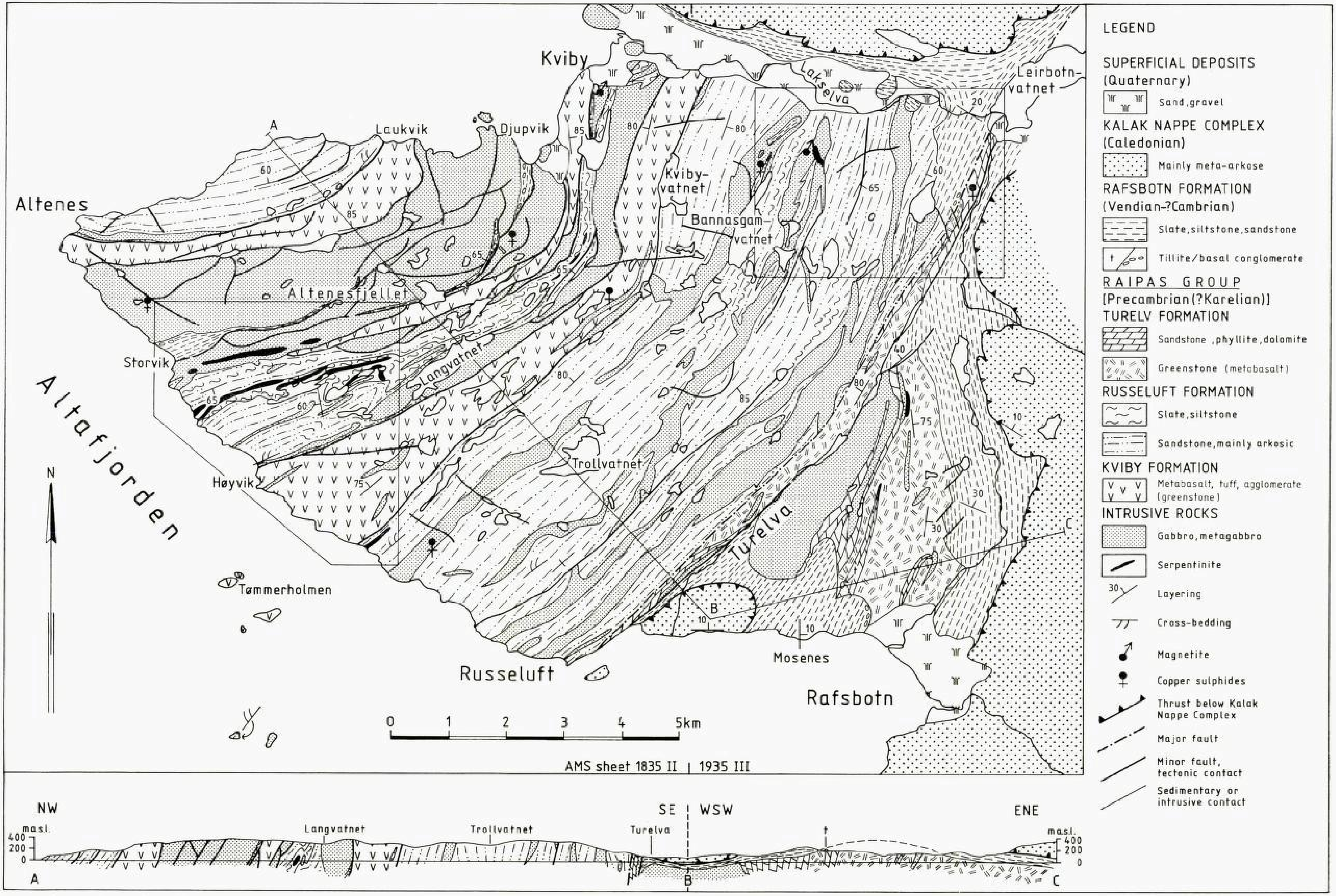

Plate 1. Geological map of the Altenes area.

The Geology of the Sorjusdalen Area, Nordland, Norway

M. A. COOPER, G. M. BLISS, I. L. FERRIDAY & C. HALLS

Cooper, M. A., Bliss, G. M., Ferriday, I. L. & Halls, C. 1979: The geology of the Sorjusdalen area, Nordland, Norway. *Norges geol. Unders 351*, 31–50.

The rocks of the Sorjusdalen area are divided into two distinct tectono–stratigraphic sequences, the lower termed the Pieske/Vasten Nappe and the upper the Gasak Nappe. The mutual contact of the nappes is imbricate in character and is marked by several distinctive lithologies and tectonic features. The Sulitjelma Gabbro was intruded across this contact resulting in intrusive relationships with both tectono–stratigraphic sequences. There are three phases of folding in the area, but only F_1 and F_2 have any significant effect on the distribution of lithologies. F_1, a phase of NE–SW trending isoclinal folds with an axial planar schistose foliation, is associated with nappe emplacement. The open, E–W trending F_2 folds deform the F_1 fabric and appear to be on the southern limb of a major F_2 synform. Medium-pressure regional metamorphism reached its peak during F_1, producing a pattern of inverted isograds; the easternmost structurally lower units are greenschist facies and the westernmost structurally highest units are middle amphibolite facies. The Sulitjelma Gabbro has a thermal aureole and yet is locally profoundly affected by regional metamorphism, indicating pre- to syn-metamorphic intrusion. The emplacement of the Gasak Nappe during F_1 was likewise syn-metamorphic and just prior to the intrusion of the gabbro. Regional correlations show that the Pieske/Vasten Nappe is merely a structurally lower level of the Gasak Nappe, and that their contact is an early slide associated with the emplacement of the Gasak Nappe.

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Introduction

The Sorjusdalen area described in this paper lies approximately 13 km north-east of Sulitjelma, spanning the border between Norway and Sweden. The Bodø-Sulitjelma region (Fig. 1) in the Central Scandinavian Caledonides lies in a tectonic depression of Caledonian cover, between the major basement culminations of the Tysfjord and Nasafjell massifs. The region has been discussed in various tectonic syntheses (Rutland & Nicholson 1965, Nicholson & Rutland 1969, Nicholson 1974, Cooper 1978). The major deformation of the area occurred during the climactic stages of the Caledonian orogeny as a result of the closure of the Iapetus Ocean during the Lower Palaeozoic and the consequent collision of the Baltic and Laurentian cratons (Wilson 1966).

The Caledonian cover consists of a sequence of metasediments, intrusives and effusives which are divisible into a series of nappes (Fig. 1) that are conjunctive in the west and disjunctive in the east (Nicholson & Rutland 1969). The varied fold structures in these rocks show a general NE-SW Caledonide trend and are polyphase in character. Using fossils found at Sulitjelma, part of

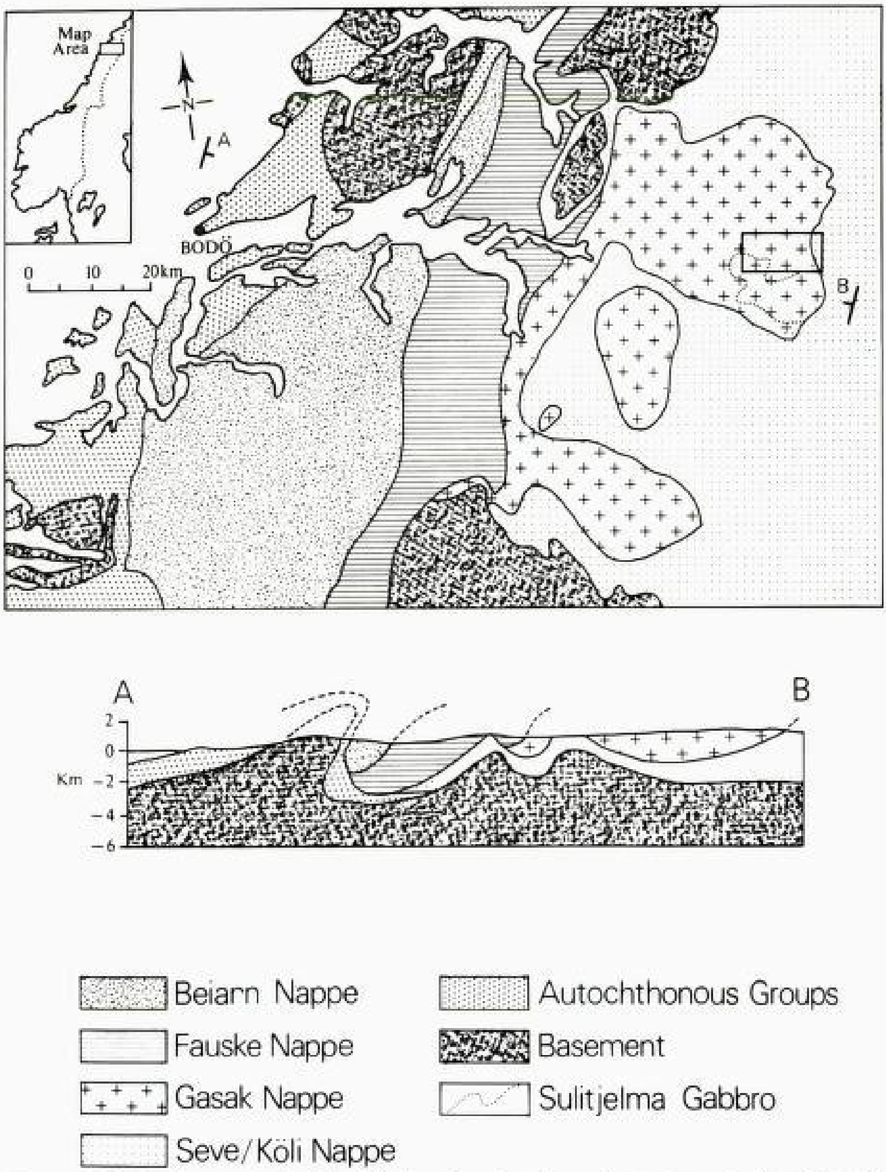

Fig. 1. Geological map of the Salta Region showing the major tectono–stratigraphic units (after Nicholson & Rutland 1969). The box locates the Sorjusdalen area.

one of the lower nappes has been dated as Middle Ordovician (Sjøgren 1900). The limits of penetrative Caledonian deformation in the basement culminations are marked by a line trending NNE-SSW (Nicholson 1974).

Since the early investigations of Sjøgren (1900) the Sulitjelma area has been considered one of the classic areas of the Scandinavian Caledonides. The tectonic and metamorphic features of the region are given an additional economic significance by virtue of the bodies of basic rocks (gabbro and amphibolite) and the stratiform pyritic ore bodies which form important parts of the rock sequence in the area (Wilson 1973).

The early investigations are reviewed in a memoir by Vogt (1927), who concluded that the gabbro and amphibolites were all part of one syn-tectonic intrusion. Kautsky (1953) challenged this interpretation based on his work on the Swedish side of the border. Kautsky divided the rocks into three allochthonous nappes, which together form the Seve Nappe complex. The highest he termed the Gasak Nappe, the middle one the Vasten Nappe and the lowest the Pieske Nappe. Kautsky interpreted the gabbro as being intruded into the Gasak Nappe and having a basal tectonic contact with the amphibolites of the Vasten Nappe. Nicholson & Rutland (1969) have fully discussed the two alternative hypotheses and concluded that Kautsky was correct in principle, but they maintained certain reservations regarding the nappe interpretation. Other discussions of the two hypotheses are given by Mason (1967) and Henley (1970). Boyle et al. (in press) present evidence for large-scale stratigraphic inversion in the Sulitjelma area and suggest a modified interpretation of the nappe structure.

The rocks in the area discussed here are divided into two sequences which, following the nomenclature of Kautsky (1953), comprise the Gasak Nappe above and to the west, and the Pieske-Vasten Nappe below and to the east. Both nappes have been intruded by the Sulitjelma gabbro. No conclusive evidence of stratigraphic way up is present in the Sorjusdalen area and hereafter the terms 'overlain' and 'underlain' are used in a purely tectonostratigraphic sense.

This study of the stratigraphy and structure of the nappes was intended to aid the understanding of the nature of the contact between the two nappes, the timing of nappe emplacement with respect to deformation history and the timing of the gabbro intrusion.

Stratigraphy

The stratigraphy of the area is described in ascending tectonic sequence as lithological units, divided into three tectonic groups as shown on the lithological map (Fig. 2). The lowest group is the upper part of the Pieske/Vasten Nappe (Kautsky 1953) the middle one is the so-called nappe junction group and the upper one is the lower part of the Gasak Nappe (Kautsky 1953). The stratigraphy, which is diagrammatically represented in Fig. 3, indicates that the Pieske/Vasten Nappe is composed of phyllites and marbles, the nappe junction group mostly of ultramafic rocks and the Gasak Nappe dominantly of amphibolites and pelites. The Sulitjelma Gabbro, which cuts across the nappe boundaries, is described separately.

THE PIESKE/VASTEN NAPPE

The Phyllite Unit

This is a greenish chloritic phyllite with cleavage parallel to bedding, and is often kink-banded and chevron-folded. Pyritic mineralisation is locally developed on a small scale. The other main minerals present are sericite, chlorite, quartz, feldspar, calcite and biotite with accessory zircon and epidote.

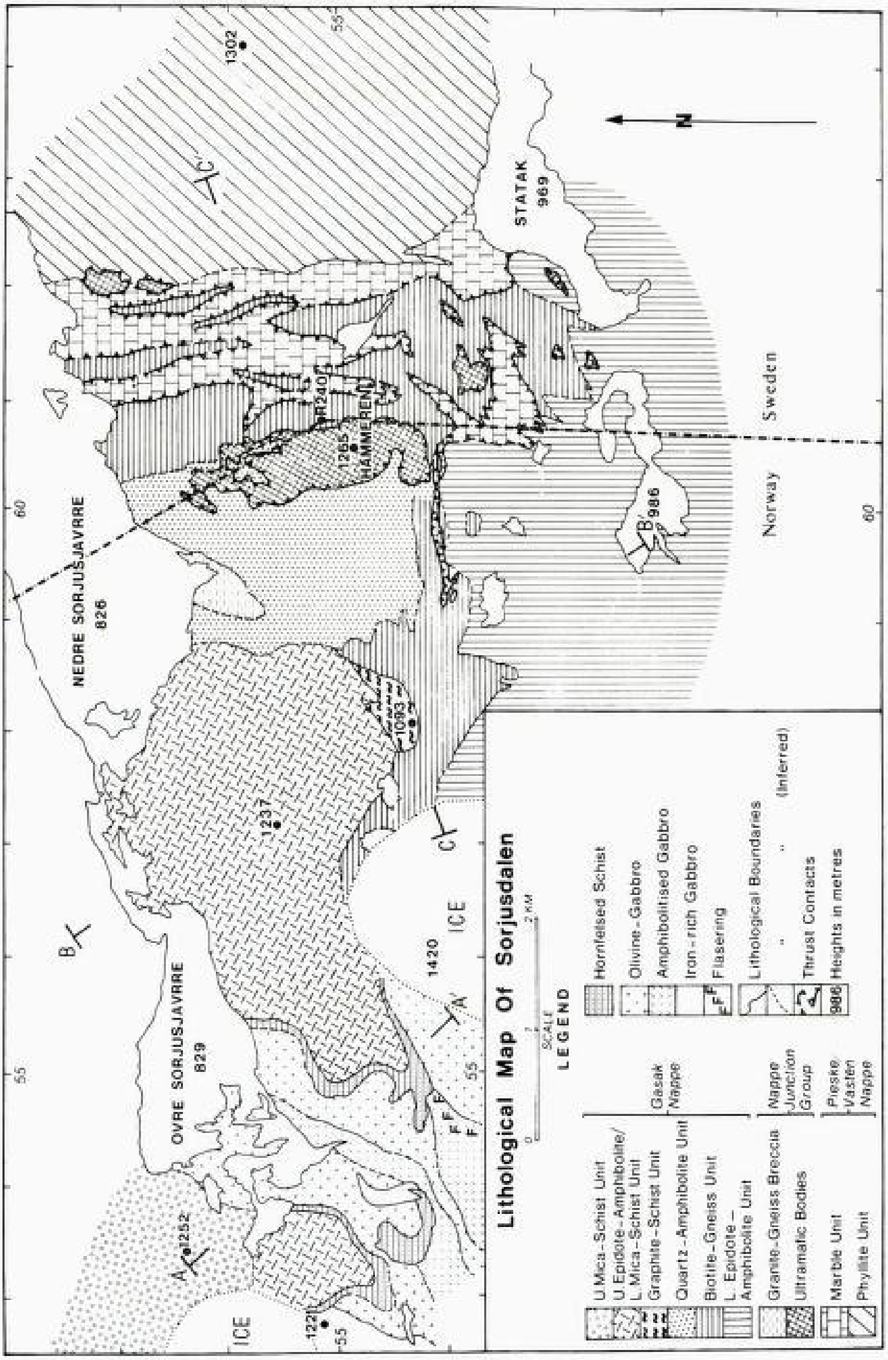

Fig. 2. Lithological map of the Sorjusdalen area.

Interbedded with the phyllite are creamy-brown calcareous, arenaceous metasediments and thin porphyritic flows of basic metavolcanics apparently thinning southwards.

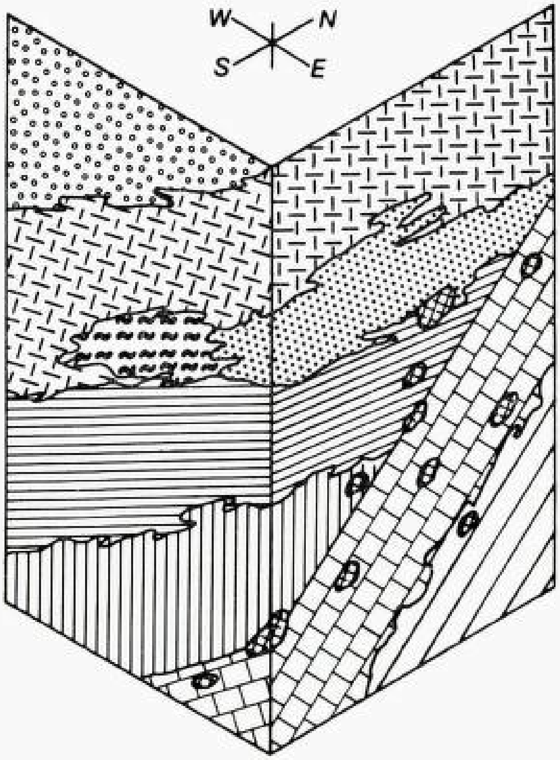

Fig. 3. Diagrammatic tectono-stratigraphy of the Sorjusdalen area. Key as for Fig. 2.

The Marble Unit

The creamy-yellow marbles vary from coarsely granular with a poorly developed planar foliation to fine grained and schistose in appearance depending on the proportions of mica and quartz present. Near their contact with the phyllite unit the marbles contain folded and boudinaged felsic bands and thick micaceous quartzites. Mineralogically the marbles comprise calcite, quartz, phlogopite, magnetite, rutile and small amounts of plagioclase.

THE NAPPE JUNCTION GROUP

Certain of the lithologies occurring in the Sorjusdalen area strongly suggest the presence of a major tectonic dislocation. These are the granite-gneiss breccia and the numerous pods of ultramafic rock. We believe that it is significant that these lithologies are broadly located between the rock sequences of Kautsky's (1953) Gasak Nappe above, and his Pieske/Vasten Nappe below.

The Granite-Gneiss Breccia

A small outcrop of the granite-gneiss breccia is present south of Nedre Sorjusjavrre; it is thought to be continuous with a far larger outcrop north of Nedre Sorjusjavrre described by Nicholson (1971).

Spheroidal blocks of granite-gneiss varying from 10 cm to 1 m in diameter are set in a finer-grained matrix of similar composition. A coarse foliation is developed, and the rock has the appearance of a metamorphosed breccia. Following Kautsky (1953) and Nicholson (1971) we interpret this rock as being a sedimentary breccia of basement gneiss clasts associated with a small area of coherent gneiss (present north of Nedre Sorjusjavrre) and associated with the basal thrust of the Gasak Nappe. One of us (I.L.F.) located a block of sheared granite-gneiss within the marble unit a short distance below the biotite-gneiss unit of Gasak Nappe.

The Ultramafic Bodies

The ultramafic bodies are found at various structural levels in the lowest parts of the Gasak Nappe and the highest parts of the Pieske/Vasten Nappe, and all show flattening in the plane of the regional foliation. The bodies generally form well-exposed topographic humps due to the hard, compact nature of the rock. Their area varies from 100 m^2 up to 150,000 m^2 in the case of Hammeren, the largest and most prominent body in the area. The dark green ultramafic rock does not developed the regional planar foliation. The mineralogy is: actinolite 60–70%, calcite 5–10%, chlorite 5–10% with talc, muscovite, plagioclase and spinel as accessories.

Associated with, and enclosing the ultramafic bodies are envelopes of altered rock derived from the original ultramafic material. Generally these consist of a coherent calcareous talc-magnetite schist, though a tectonic breccia may locally develop, as for example, around Hammeren. A typical section into an ultramafic body is as follows:

Country Rock
Calcareous talc-magnetite schist
Talc-rich actinolite schist with magnetite
Medium to coarse grained talc-actinolite rock
Fine-grained actinolite ultramafic rock.

The light-grey calcareous talc-magnetite schist is widely developed as a thin envelope enclosing the ultramafic bodies. It often contains small octahedra or larger aggregates of magnetite and occasionally becomes dark blue-grey due to finely disseminated magnetite. The mineralogy is talc, quartz, calcite and magnetite.

The tectonic breccia has a variable composition, containing blocks of ultramafic rock and carbonate, frequently fractured and veined by quartz, with a talc-rich matrix.

Within these lithologies large crystal aggregates are locally developed, for example, diopside crystals up to 5 cm long, sideritic rhombs up to 3 cm across, talc in radiating crystalline form, prismatic clinozoisite, almost perfect spheroids of pyrite and other more deformed pyrite grains.

There is no evidence of the presence of thermal metamorphic aureoles around the bodies, and we consider the alteration envelopes described above to represent a dynamic metamorphic aureole. We therefore suggest that the bodies were tectonically emplaced in the solid state, their movement being facilitated by the lubricating nature of their aureoles. The bodies are therefore allochthonous with respect to the host nappes and were probably derived from the tectonic break-up of a larger ultramafic mass.

THE GASAK NAPPE

The Lower Epidote-Amphibolite Unit

The lower epidote-amphibolite is medium grey in colour, compositionally banded and fine grained with occasional coarser patches. A planar schistose foliation is usually developed parallel to the compositional banding. Numerous pale green lenses of epidote and occasional pale brown lenses of carbonate occur, flattened in the plane of the foliation. The local presence of matrix carbonate causes the development of solution pits. The rock contains hornblende 50–60%, sodic andesine 25–30%, quartz 10%, opaques 5% and calcite 0–5%. The mafic and felsic minerals occur in discrete bands and within the former the hornblendes have a strong preferred orientation.

The Biotite-Gneiss Unit

The biotite-gneiss is a well-foliated rock with alternating discrete bands 2–3 mm wide of felsic and mafic minerals. Small red almandine garnets (1–2 mm) are common. The mineralogy is biotite 30–40%, quartz 25–30%, sodic plagioclase 15–20%, muscovite 10–15%, garnet 0–5% and kyanite 0–2%; accessory minerals include tourmaline and opaques. The garnet and kyanite form post-tectonic porphyroblasts that truncate the micas of the gneissose foliation. In its outcrop east of Hammeren the biotite-gneiss contains numerous quartz and feldspar augen. Here it is less homogeneous than in the western outcrop, with some mineralized horizons which are discussed in detail in the section on mineralisation.

The Quartz-Amphibolite Unit

This rock is dark green-grey, fine grained amphibolite with a well-developed planar foliation. Almond-shaped lenses (amygdaloidal?) of quartz, feldspar or calcite and small pyrite cubes are common. At the southern contact with the biotite-gneiss unit the amphibolite becomes very schistose, whereas at the contacts with the calcareous talc-magnetite schist it is apparent that localised Fe/Mg metasomatism has occurred during regional metamorphism. The mineralogy is hornblende 80%, quartz 5–10% and plagioclase 5–10%, with accessory epidote, pyrite, calcite and secondary chlorite. The amphiboles show a high degree of preferred orientation, and also show uneven extinction indicating post- or syn-crystallisation stress.

Within the quartz-amphibolite, lenses 100×100 – 500 m of metafelsitic rock occur. There are complex interveining and mixing relationships between the two lithologies and reaction rims around xenoliths can be observed, suggesting that the unit was originally igneous.

Nicholson (1971, p. 157) included these rocks, by extrapolation, within his granite-gneiss breccia. We believe that the granite-gneiss breccia thins southwards across Nedre Sorjusjavrre, only occurring as a small outcrop and that the quartz-amphibolite unit (presumably not present north of the lake) should be considered as a separate lithology.

The Graphite Schist Unit

The graphite schist is a black, well-foliated, rusty-weathering rock composed principally of graphite and pyrite. This presumably indicates sedimentation under strongly reducing conditions producing a black mud rich in FeS and organic matter. Nicholson (1971, p. 156) notes the presence of the graphite schist in the area, but it is more discontinuous than he implies.

The Upper Epidote–Amphibolite Unit

The rock is medium grey and fine-grained with a strong planar foliation which is often crenulated. Compared with the lower epidote–amphibolite it contains fewer epidote lenses and more carbonate lenses, some of which are very large. In areas of high matrix carbonate content the rock becomes mottled grey and brown with numerous solution pits. The mineralogy is: hornblende 60–75%, (with a marked preferred orientation), quartz 15–25%, plagioclase 5–10%, epidote 0–5% and calcite 0–5%.

The Lower Mica Schist Unit

The lower mica schist unit occurs as small concordant bands within the upper epidote–amphibolite on the northern slopes of hill 1237. The schist is well-foliated, fine/medium-grained and varies from grey to dark grey in colour with some coarser mica flakes and clusters of red garnets. The matrix carbonate content must be locally quite high as solution pits are often observed on weathered surfaces. However, the carbonate is not seen in the available thin-sections, which contain biotite 35–40%, quartz 10–20%, plagioclase 5–15% and opaques. The micas have a marked preferred orientation and are concentrated into discrete bands.

The Upper Mica Schist Unit

The field appearance of this rock is very similar to the lower mica schist unit; it has been reported by Mason (1967)to contain kyanite, but none was observed in the field or in the thin-sections studied.

THE SULITJELMA GABBRO

The north-western corner of the Sulitjelma Gabbro outcrops within the Sorjusdalen area. The gabbro is layered, and in the field is divisible into four major facies; olivine–gabbro, amphibolitised gabbro, flaser gabbro and iron-rich gabbro. The flaser gabbro is more commonly found at the gabbro/schist contact, while the other three facies occur as broadly E–W trending bands parallel to Sorjusdalen.

Olivine–Gabbro

A detailed account of the olivine–gabbro facies can be found in Mason (1971), with which our own petrological observations concur. Contacts with the amphibolitised gabbro are gradational, with amphibole replacing olivine over a distance of 8–10 m. The contact with the country rocks is primary.

Amphibolitised Gabbro

The amphibolitised gabbro contains plagioclase (An_{45-55}), cummingtonite, clinozoisite and small amounts of chlorite and pale-brown biotite. Locally the amphibole is green hornblende in place of cummingtonite. The amphiboles tend to form large aggregates of small fibrous grains with larger crystals sometimes developed in the centres. Many specimens retain relict ophitic textures, and occasionally relict pyroxene and olivine occur. The amphibolitisation is the result of processes similar to those which formed the corona structures described by Mason (1971), but occurred later as water was introduced during regional metamorphism. The amphibolitised gabbro occurs as two broad layers sandwiched between layers of olivine gabbro, and is concordant with primary igneous banding. The development of amphibolitisation thus appears to have been controlled by the layered nature of the intrusion. Shear zones are localised within the amphibolitised gabbro indicating preferential failure within this layer.

Flaser Gabbro

The word 'flaser' is used to describe rocks in which crystal lineation is developed. There are both amphibole- and olivine-rich flaser gabbros and there is no visible mineralogical difference between the flasered and non-flasered facies. In the amphibolitised gabbro where the flasering phenomenon is best developed, segregation of the amphibole and plagioclase has occurred and preferred orientation is pronounced even in the plagioclase layers. Plagioclase seems to be more resistant to the deformation, although the plagioclases are slightly fractured, well rounded and sub-spherical to elliptical. The olivine gabbros behave in a similar fashion, crystals becoming strained and elongate during flasering. However, the olivine tends to fragment into anhedral aggregates. The flasering facilitates movement of water and subsequent recrystallisation to an amphibolitised gabbro; however, this does not always occur and some rare flasered olivine gabbros are preserved. The formation of the flaser rocks occurred during the regional metamorphism, primarily in the later stages due to hydration as the gabbro cooled.

Iron-rich Gabbro

The sulphide enriched facies occurs on the western margin of the Sorjusdalen area. Its contact is concordant with the other primary layering within the gabbro. Some smaller localised bodies also occur. The presence of such enrichments suggests the possibility of economic Cu–Ni cumulus mineralisation within the Sulitjelma gabbro. Details of the textures and compositions of the opaque minerals are given in the mineralisation section.

Deformation history

The Caledonian deformation within the two nappes and the intervening nappe junction group consisted of three phases of folding of varied importance, style

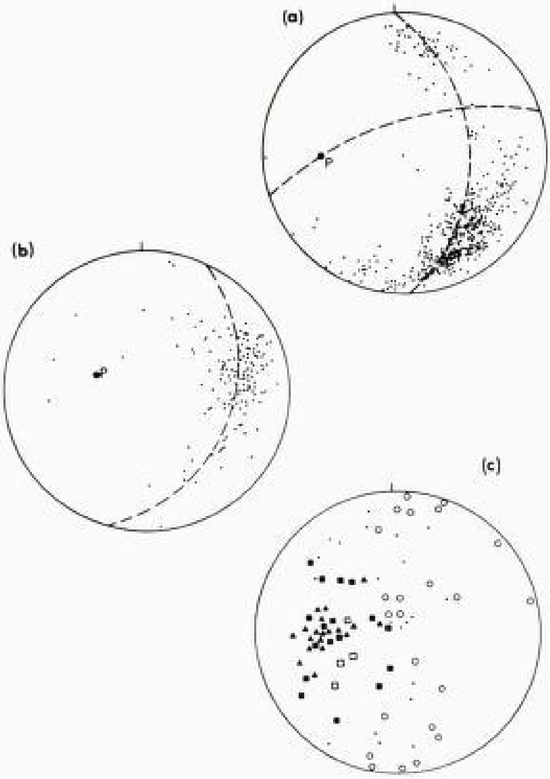

Fig. 4. (a) Stereogram of poles to S_1 in the Gasak Nappe, P is the pole to the best fit π circle, and the great circle containing P the approximate orientation of the axial plane; (b) stereogram of poles to S_1 in the Pieske/Vasten Nappe, P is the pole to the best fit π circle; (c) stereogram of F_1 minor fold axes ·, F_2 minor fold axes ■, F_3 minor fold axes ○, mineral lineations ▲, and boudin neck directions □.

and distribution, with the later formation of small shear zones and areally important faults. Coeval with the first fold phase (F_1) regional metamorphism occurred at greenschist facies in the east, rising steadily to lower amphibolite facies in the west and producing the regional foliation (S_1).

FOLDING

Within the area three distinct phases of folding can be recognised by their different styles, scales and their relationship to S_1. Folds ascribed to the earliest of these phases, termed F_1, are present throughout the area, although in the western and central parts they are strongly modified by subsequent F_2 folding. The F_1 folds are isoclinal with approximately N–S axes and axial planar foliation (S_1) produced during the coeval regional metamorphism. Within the phyllite unit, where the effects of F_2 are negligible, S_1 is preserved in its original orientation, striking N–S with a dip of 60° to the west (Fig. 4b). F_1 minor folds are common in the eastern part of the area, but in the central and western sectors they are comparatively rare (Fig. 5). They are predominantly isoclinal to tight and have a random orientation (Fig. 4c). Application of the techniques of Ramsay (1967, p. 450) for removing the effects of later deformations does not, however, produce a concentration of data. Major F_1 folds are best displayed east of Hammeren where the biotite–gneiss unit and the marble unit are isoclinally interfolded (Fig. 7).

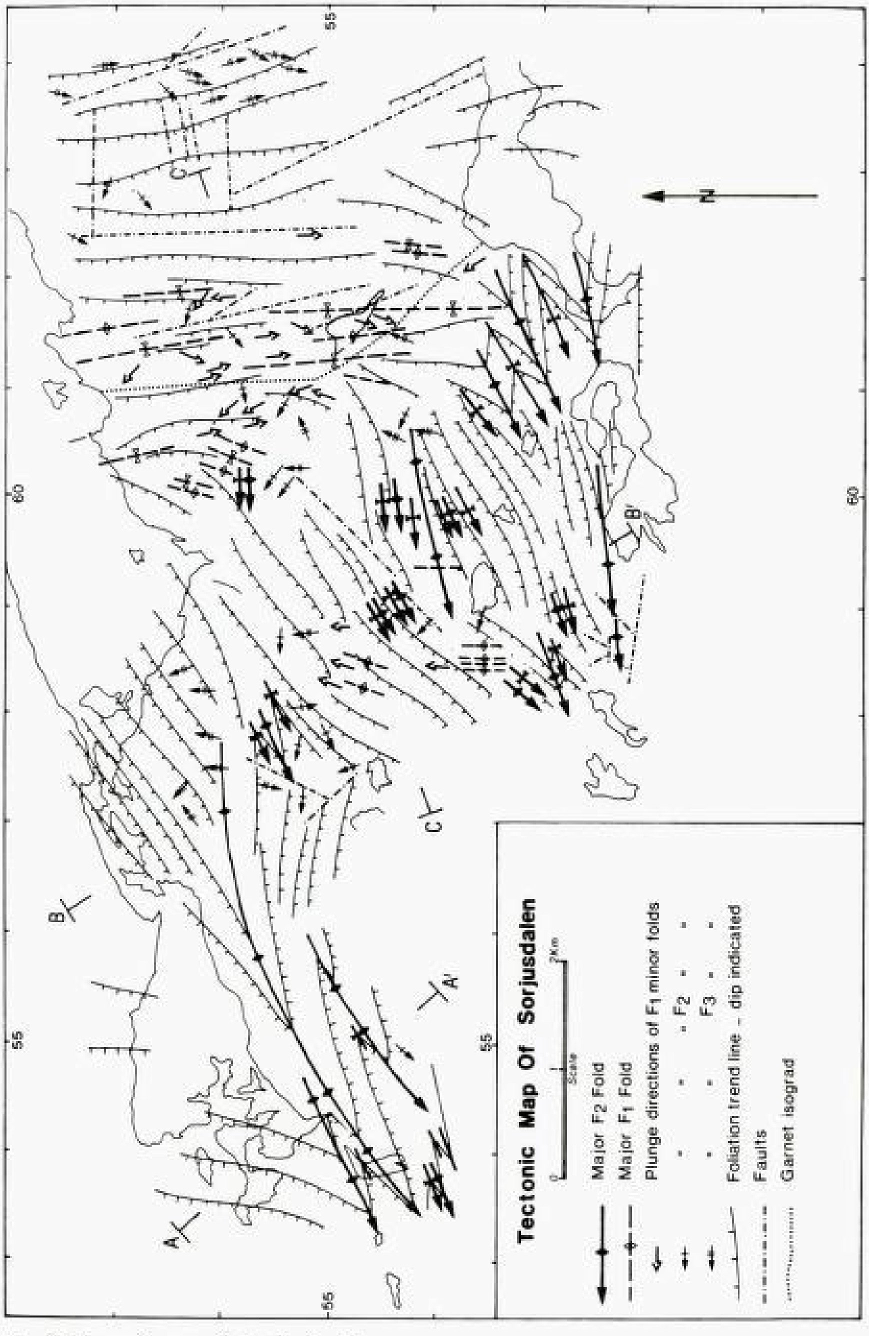

Fig. 5. Tectonic map of the Sorjusdalen area.

F_2 folds are present at all structural levels and deform S_1 in an open asymmetric style on E–W fold axes without development of an axial planar cleavage. F_2 folding appears to die out eastwards, being completely absent in the phyllite unit, thereby causing the areal swing in S_1 (Fig. 5). Southern limbs dip approximately 65° to the south-west; the spread of data for the northern limbs, however, suggests a curved profile with dips varying from 60° to the north-west to 70° to the north-north-west. Fold axes plunge 50° to the west and the axial planes strike 075° with a dip of 70° to the north-north-west (Fig. 4a). F_2 minor folds are common, frequently displaying S and Z symmetries related to their position on larger F_2 structures. They vary considerably in style but always fold S_1 and usually plunge west (Fig. 4c). The largest and best example of an F_2 fold in the area occurs on the ridge just to the south of Øvre Sorjusjavrre (Fig. 5). The symmetry of the major F_2 folds indicates that the area is situated on the southern limb of a regional F_2 synform. Boudinage apparently developed during F_2, the boudin neck directions being broadly parallel to the F_2 fold axes (Fig. 4c), however, much of the boudinage has a chocolate-tablet structure and it is difficult to discriminate between the effects of F_1 and F_2. A mineral lineation is developed apparently parallel to the F_2 axes, but there is no evidence to support a metamorphic event during F_2; it is therefore considered to be an F_1 lineation. In the central part of the area F_1 and F_2 folds are superimposed producing interference patterns.

Major F_3 folds are not recognised in the area, but minor folds occur which are not obviously compatible with the known geometries of F_1 and F_2 structures, and therefore are assumed to be F_3. In the amphibolitic units these are often ptygmatic in style but in the phyllite unit chevron folding and kinking of S_1 has occurred on gently southward plunging axes (Fig. 4c).

METAMORPHISM

One main phase of regional metamorphism is recognised in the area, broadly coeval with the F_1 fold phase and producing S_1. From east to west there is a general increase in metamorphic grade rising from the chlorite zone up to the kyanite zone.

The easternmost part of the phyllite unit is in the chlorite zone, the proportion of chlorite relative to sericite increasing westwards. The marbles contain the first occurrence of green biotite, but this has no bearing on the location of the biotite isograd. As the contact of the marble with the biotite–gneiss unit is approached knots of acicular amphibole and epidote appear. The garnet isograd can, however, be fixed with comparative certainty (Fig. 5). Moving west from the garnet isograd the biotite/chlorite ratio changes from 7.0 to 25.0 and muscovite appears at the expense of sericite. The amphibolites which occur in the almandine zone contain calcic oligoclase, garnet, green hornblende and epidote, an assemblage which indicates epidote–amphibolite facies metamorphism (Miyashiro 1973). A kilometre or so to the west of the garnet isograd, kyanite begins to appear as small grains but the position of the kyanite isograd cannot be located accurately. The biotite within the pelitic rocks becomes brownish green and tourmaline appears. The amphibolites within the kyanite zone contain a similar assemblage to that in the almandine zone except that the plagioclase present is a sodic andesine. This is the highest grade of regional metamorphism present in the area.

The S_1 fabric developed during the metamorphism varies from phyllitic through schistose to gneissose, with no evidence of an earlier fabric preserved in the porphyroblasts. The garnet porphyroblasts are idioblastic, inclusion-free and truncate the S_1 foliation. The kyanite porphyroblasts have similar relationships to the S_1 foliation, and both garnet and kyanite porphyroblastesis is therefore regarded as post-S_1 and hence post-F_1. The plagioclase porphyroblasts are somewhat ragged but clearly contain inclusions of mica that parallel the external foliation. It is concluded that although the regional metamorphism was broadly coeval with F_1, the highest grade in fact post-dates F_1.

During metamorphic retrogression and cooling, a large number of quartz veins seem to have formed at various times. Their relative chronology can be deduced in relation to the fold phases, one set forming post-F_1 pre-F_2, another post-F_2 pre-F_3, and the youngest set post-F_3. The rocks must have cooled down considerably by F_2 times as no evidence of diaphthoresis is apparent, even in F_2 hinges.

The effect of regional metamorphism on the gabbro has been to produce amphibolitisation. However, as noted earlier regional metamorphic effects are incomplete, and certain parts of the gabbro retain their primary mineralogy. In agreement with Mason (1971) it is thought that this is probably due to the availability and mobility of water during the metamorphism of the gabbro. The gabbro shows the effects of regional metamorphism but it also has a thermal aureole preserved as a hornfelsic envelope varying in width from 30 to 40 m and dying out gradually as shown by the progressive appearance of the regional foliation.

SHEAR ZONES AND FAULTS

Shearing and faulting occur at all structural levels, cutting and therefore post-dating all the above-mentioned structural elements. There is, however, no apparent relationship between the shear zones and the faults. The shear zones cause sharp changes in strike of schistosity, and have occasionally developed sigmoidal tension gashes infilled by quartz. Thin-sections taken near the shear zones show post-crystallisation strain effects.

The faults shown on the tectonic map (Fig. 5) are not directly detectable in the field by displacement of foliation or juxtaposition of lithologies. However, fractures are readily seen on the available aerial photographs and help to explain several apparent paradoxes of structural trend (Fig. 5). The faulting is thought not to be coeval with the shearing, as the former is a brittle fracture compared with the more ductile development of the shear zones.

Mineralisation

Mineralisation of both magmatic and sedimentary origins occur within the area. Magmatic mineralisation is represented by the occurrence of the pentlandite–pyrrhotite association together with magnetite and chalcopyrite in the gabbro and by magnetite–pyrite–pyrrhotite–chalcopyrite–chromite mineralisaton in

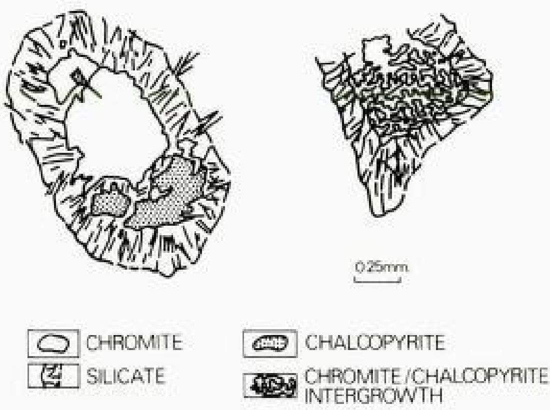

Fig. 6. Ore mineral textures from the Sorjusjavrre ultramafic body (620571).

bands within the ultramafic pods. Sedimentary mineralisation is represented by minor concentrations of pyrite, with rare chalcopyrite, within the biotite–gneiss unit and by pyrite–graphite in the graphite schist unit.

Evidence for the presence of a nickel-bearing monosulphide liquid at certain stages during the crystallisation of the gabbro magma is provided by the occurrence of the rusty weathering iron-rich facies. The opaque minerals are located in the interstices of the silicate cumulus fabric making up approximately 20% by volume of the rock. The opaque minerals occur in the following relative proportions pyrrhotite (with alteration products) 80%; magnetite 10%; chalcopyrite 5% and pentlandite $<2.5\%$. The pentlandite was shown by electron probe microanalysis to be a cobaltian pentlandite with the following composition:

Fe 30.3%, Co 3.3%, Ni 35.1%, S 29.8%.

Within the two largest ultramafic bodies, Hammeren and the body just to the south of Nedre Sorjusjavrre, mineralisation involving chromite, chalcopyrite, pyrite and cobaltian pyrite occurs. The mineralisation in the Sorjusjavrre body (620571) occurs as crude bands of small grains (0.5 mm) dispersed through the fibrous rock. Individual bands are 3–4 mm thick, and the texture appears to be the result of the cataclasis of larger grains. In polished section chromite forms 10–15 modal % of the rock, occurring as large grains with decussate boundaries in intimate association with chalcopyrite (Fig. 6). Electron probe scans across several chromite grains produced an average figure of 4–6% Cr. Pyrite and cobaltion pyrite are dominant, however, occurring as large (1–2 mm) sub-angular to rounded grains.

Within the biotite–gneiss unit, to the north-east of Hammeren, there occurs a distinctive and unique horizon of about 1 m thickness, composed of approximately 80% garnet, 10% quartz and feldspar and accessory chlorite, biotite, carbonate and magnetite. This lithology is pink in colour, very fine-grained and is pervasively dissected by fine fracture veinlets filled by quartz and carbonate. At the time of mapping this horizon was considered to represent an anomalously iron–alumina-rich sedimentary intercalation.

Since 1973, investigations carried out in the Grong region of the Central

Norwegian Caledonides (Halls et al. 1977) have revealed a striking similarity between the Sorjusjavrre garnetite and certain volcanogenic exhalative (chemical–sedimentary) horizons which commonly occur within the volcano-stratigraphy of the Skorovas area near Grong. In the Skorovas area, laminated horizons of pink garnetite and epidotite, laterally gradational into hematite/magnetite cherts, occur above and stratigraphically peripheral to the Skorovas massive sulphide orebody. Such iron–aluminium–calcium rich horizons are considered to be the concentrated products of precipitation under appropriate physio-chemical conditions from fluids rich in elements leached from submarine extrusives, following exhalation and dispersal of the fluids into the submarine environment. Pink spessartite-rich horizons of similar appearance have been reported from the Trondheim region by Oftedahl (1967).

In the Sulitjelma region, garnetite horizons have recently been found associated with pillow lavas of the Sulitjelma amphibolite (A. Boyle, pers. comm. 1978), being essentially very similar both in field appearance and petrography to the exhalative garnetite horizons of the Grong region. Geological evidence thus suggests that the garnetite horizons within the Sulitjelma amphibolite are exhalative in origin and similarly there is little doubt that the Sorjusjavrre garnetite is exhalative.

Discussion

THE TIMING OF GABBRO INTRUSION

The presence of a thermal aureole around the gabbro together with regional metamorphism of the gabbro indicates that intrusion of the gabbro occurred syn-metamorphically. The mineral assemblages described from the units adjacent to the gabbro indicate a minimum pressure of 5 kb and temperatures in excess of 500°C (Winkler 1974). The temperature of the gabbro during and immediately preceding intrusion must have been considerably higher than that of the country rocks. The metamorphism of the gabbro, mainly flasering and amphibolitisation, occurred during the latter stages of the regional metamorphism as the gabbro slowly cooled to a temperature at which water could enter, because as shown by Mason (1971, p. 137) the metamorphism is isochemical apart from H_2O.

REGIONAL SIGNIFICANCE OF THE SORJUSDALEN AREA

The division of the lithological units into the Gasak Nappe above the nappe junction zone (from here on referred to as the Hammeren slide) and the Pieske /Vasten Nappe beneath follows the interpretation of Kautsky (1953). Recently, major stratigraphic inversions in the Sulitjelma area have been recognised and used to reinterpret the structure and stratigraphic relationships of the area (Boyle et al. in press). The inversions are largely based on pillow lavas and require the interpretation of the Gasak Nappe as a fold-nappe with an extensive inverted limb of which the sequence east of Hammeren is a part. The granite–gneiss breccia and the ultramafics are considered by Boyle et al. (in press) to occupy the core of the fold-nappe. Such an interpretation presents difficulties, for as the evidence presented above suggests an early slide is present in the Hammeren area. A further problem is that the stratigraphy west of Hammeren does not repeat that found to the east and thus it is difficult to accept that Hammeren occupies the core of a large fold-nappe. Boyle et al. (in press) appreciate the latter point, correlating the rocks west of Hammeren with those of the Duoldagop area (Fig. 8), known to belong to a structurally higher level of the Gasak Nappe. They do not, however, insert the early slide in the Hammeren area that this stratigraphic argument would appear to require.

The correlation of the rocks west of Hammeren with those of the Duoldagop area must be questioned for two reasons. The first is the presence in the biotite–gneiss unit of the banded exhalative garnetite described earlier which is similar in appearance to the garnetiferous horizons of the Vaknahelleren schist described by Boyle et al (in press, Fig. 1) as 'commonly graphitic, manganiferous or keratophyric'. This correlation is supported by the presence of the graphite schist unit on the northern contact of the biotite gneiss unit. The second point is that many of the rocks west of Hammeren are amphibolitic and locally display igneous characteristics. It is therefore possible to correlate the biotite gneiss, graphite schist, quartz amphibolite and lower epidote–amphibolite units with the Sulitjelma Amphibolite Group. The upper epidote–amphibolite and the lower and upper mica schist units are, however, correlated with the rocks of the Duoldagop area. This substantially reduces the amount of stratigraphy cut out by the Hammeren slide, necessitating only the removal of the marble and phyllite units. If the correlation with the Sulitjelma Amphibolite Group is refuted a major slide of even greater displacement must be postulated.

A consequence of these correlations is that an early synclinal fold must be proposed in the Hammeren area (Fig. 8) similar to that demonstrated south of Duoldagop by Boyle et al. (in press). Also, a second slide must be inserted to account for the juxtaposition of the above synclinal structure and the structurally higher rocks of the Gasak Nappe which form hill 1237. This strengthens the analogy with the structure south of Duoldagop.

The stratigraphy requires that both of these slides must cut out early anticlinal closures to which the proposed syncline is complementary. These relationships are indicated on the cross-section X–Y (Fig. 8). The Hammeren slide must be accepted in view of the stratigraphic arguments presented. The other structures, however, are speculative and the interpretation presented here is not unequivocal. The units attributed to the Pieske/Vasten Nappe are therefore structurally lower elements of the Gasak Nappe, and the nappe junction units regarded by Kautsky (1953) as marking the base of the Gasak Nappe in fact represent a slide within the Gasak Nappe. This revised interpretation does not prejudice the conclusion of Boyle et al. (in press) that the Gasak Nappe is a fold-nappe but modifies it to accord with the stratigraphic relationships.

THE HAMMEREN SLIDE

The Hammeren slide is very complex (Fig. 7 C-C') but, significantly, no

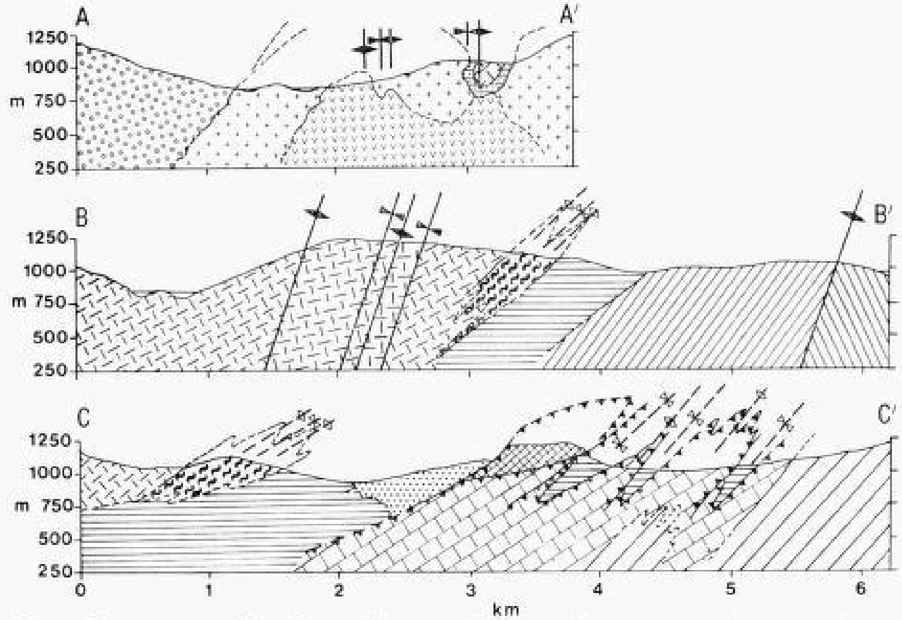

Fig. 7. Cross-sections of the Sorjusdalen area. The lines of section are indicated on Fig. 2.

evidence of mylonitisation is present. Evidence of high tectonic strain is often preserved in the lithological units in contact with the marble unit. Slices of the biotite–gneiss unit occur within the marble, indicating anastomosis of the slide (Fig. 7 C-C'). The complex structures within the marble beneath the slide zone suggest that a great deal of flow has occurred in the marble, which was presumably in a highly ductile state.

In the field, the Hammeren slide is marked by the change from one lithological sequence to another and by the very characteristic lithologies described previously. These include the ultramafic bodies and the granite–gneiss breccia. The envelope of calcareous calc–magnetite schist surrounding the ultramafic bodies is thought to be a dynamo–metamorphic alteration of the ultramafic rock and would act as an excellent lubricant suggesting solid emplacement of the ultramafic bodies (Heard & Rubey 1966). The bodies are therefore the remnants of a larger ultramafic body fragmented and tectonically emplaced during the development of the Hammeren slide.

The granite–gneiss breccia has been correlated by Nicholson (1971) with similar rocks on the basement gneiss of the Nasafjell culmination who considered that: ". . . its discovery adds weight to the proposals that the Sorjusvann assemblage also is of basement origin . . ."; the implication being that the breccias formed uncomformably on the basement (not necessarily on the Nasafjell culmination) together with the overlying comformable sequence prior to emplacement as part of the Gasak Nappe. The rocks now in contact with the granite–gneiss breccia have been tentatively correlated with the Sulitjelma Amphibolite Group. The regional stratigraphy, however, requires that in a conformable sequence the Pieske marble and the Furulund and Sjønstå Groups

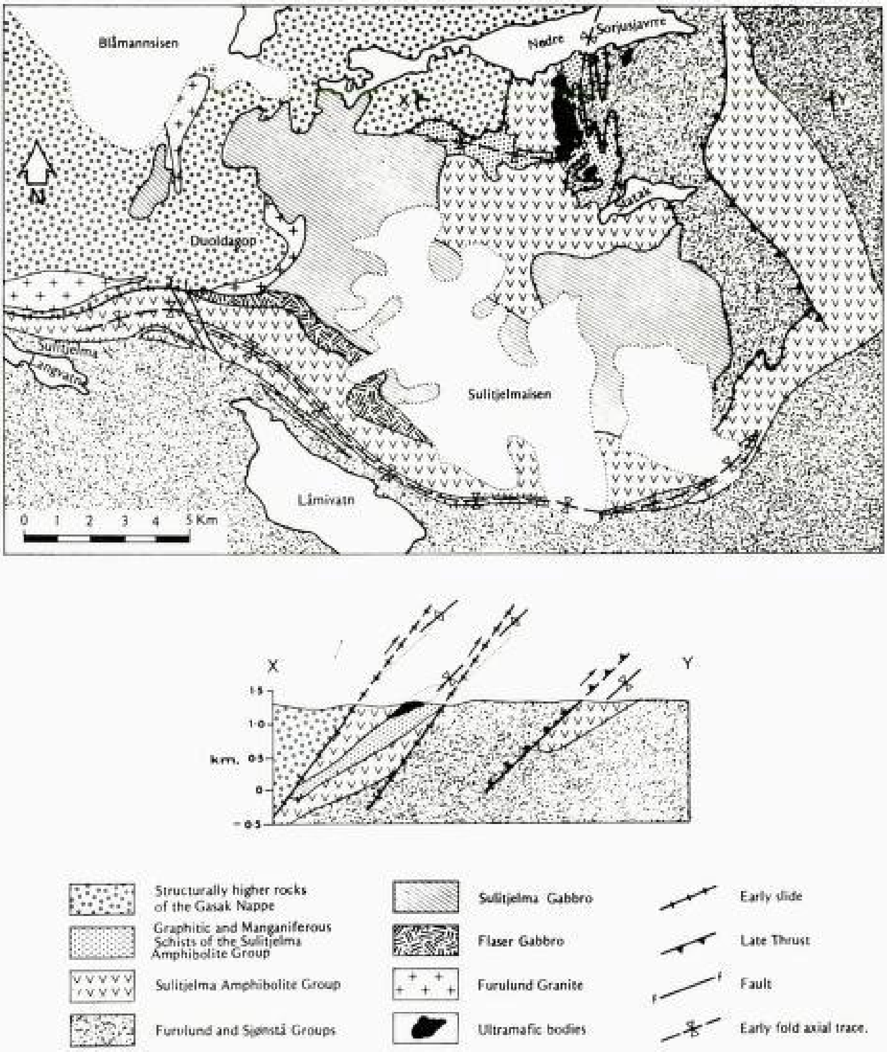

Fig. 8. Speculative geological map of the Sulitjelma district based on that of Boyle et al. (in press).

should separate the two (Nicholson & Rutland 1969). It must be concluded, therefore, that the Hammeren slide is an imbricate tectonic slide within the Gasak Nappe.

All the lithological units were involved in F_1 folding and the coeval S_1 foliation has overprinted the Hammeren slide. This, together with the ductile style of the slide, suggests development during the early stages of F_1 deformation.

TIMING OF NAPPE EMPLACEMENT

It is known from other studies in the Nordland region that nappe emplacement occurred during F_1 deformation (Nicholson & Rutland 1969, Wells & Bradshaw 1970). The association of the development of the Hammeren slide with F_1

deformation in the Sorjusdalen area therefore suggests both were linked with the emplacement of the Gasak Nappe. This is supported by the conclusions of Boyle et al. (in press) who point out that the Gasak Nappe is a fold-nappe emplaced during the earliest deformation phase.

Following Henley (1970) and Mason (1971) we attribute the inversion of metamorphic grade in the area to the syn-metamorphic nature of the large Sulitjelma Gabbro which must have supplied a considerable amount of heat to the country rocks. The regional trend of increasing metamorphic grade westwards must not, however, be forgotten, and we believe that even if the thermal effects of the gabbro were removed there would still be a general westward increase in grade. The effect of the thermal energy supplied by the gabbro has been to enhance and exaggerate this trend. The gabbro must have been intruded after the emplacement of the Gasak Nappe and the associated development of the Hammeren slide as there is no evidence to suggest that these early structures affect the gabbro.

Henley (1970) made a detailed examination of the area to the south-east of Sulitjelma and concluded that the nappe emplacement is syn-metamorphic. However, he noted that the observed penetrative fabric was created by a flattening modification (his D_2) of an earlier schistosity axial planar to isoclinal folds (his D_1). We equate our F_1 fold phase with Henley's D_1, and suggest that subsequent (D_2) flattening, which we have been unable to differentiate as such, may have produced the congruency of structure and the fabric above and beneath the Hammeren slide.

We conclude that the Gasak Nappe was emplaced syn-metamorphically, at or just prior to the metamorphic peak, with the intrusion of the gabbro occurring shortly afterwards. The nappe is internally disrupted by the anastomosed Hammeren slide which is further complicated by the ultramafic bodies which were tectonically emplaced as the slide developed during the translation of the Gasak Nappe. Considering the nature of the deformation occurring at this time the Gasak Nappe must be regarded as a disjunctive fold-nappe of considerable size.

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Petrology, Geochemistry and Genesis of the Type Area Trondhjemite in the Trondheim Region, Central Norwegian Caledonides

WILLIAM B. SIZE

Size, W. B. 1979: Petrology, geochemistry and genesis of the type area trondhjemite in the Trondheim Region, Central Norwegian Caledonides. *Norges geol. Unders. 351*, 51–76.

Trondhjemite from the Follstad district of the Trondheim region is present within allochthonous Lower Paleozoic eugeosynclinal and partly miogeosynclinal rocks of the Caledonide orogen. Four periods of trondhjemite intrusive activity have been recognized, with the main Follstad trondhjemite a product of late tectonism and metamorphism in the upper greenschist facies. Field relations, mineralogy and major and trace element abundances all attest to the high degree of homogeneity of the trondhjemite, which is of the high-Al_2O_3 type (average 16.2%). Trondhjemites of the district average 71.5% SiO_2, 5.3% Na_2O_3, but only 1.3% K_2O. The REE abundances are depleted and highly fractionated. Barium greatly exceeds rubidium in abundance. Distinctive compositional characteristics of the Follstad trondhjemite support designation of the district as a type locality for trondhjemite. Petrogenetic models for the origin of the Follstad district trondhjemite that are based on anatexis of granitic crust or partial melting of greywacke are unlikely. Distinctive compositional gaps in the gabbro–trondhjemite suite, when compared with the other associated magmatic rocks in the Follstad district, cannot easily be explained by a fractional crystallization model. The petrogenetic model that best fits the data from the Follstad trondhjemite is based on equilibrium melting of a low-K_2O tholeiitic basalt during anatexis in an orogenic zone. Published experimental work on the partial melting of basalt shows that at 5 kb PH_2O and between 825°–850°C, a melt of trondhjemite composition is produced. Residual hornblende would produce compositional trends similar to those of the Follstad district trondhjemite.

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Introduction

Trondhjemites have been called by other names in the literature including soda-granite, plagiogranite and leucogranodiorite. They are distinctive rock types and are abundant in continental Archean age rocks (Arth & Hanson 1975). The association of trondhjemites with younger mobile belts at continental margins and subduction zones indicates that the production and growth of continental crust may be associated with the genesis of trondhjemitic rocks.

The purpose of this paper is to quantify the mineralogic, chemical and petrologic characteristics of the type area of trondhjemite as first proposed by Goldschmidt (1916) and to provide the data necessary for this area to be used as a standard against which to compare other trondhjemite-bearing areas. Of equal importance in this study is an examination of the different petrogenetic models proposed for the origin of trondhjemite to determine which, if any, are consistent with the data from the type area trondhjemite.

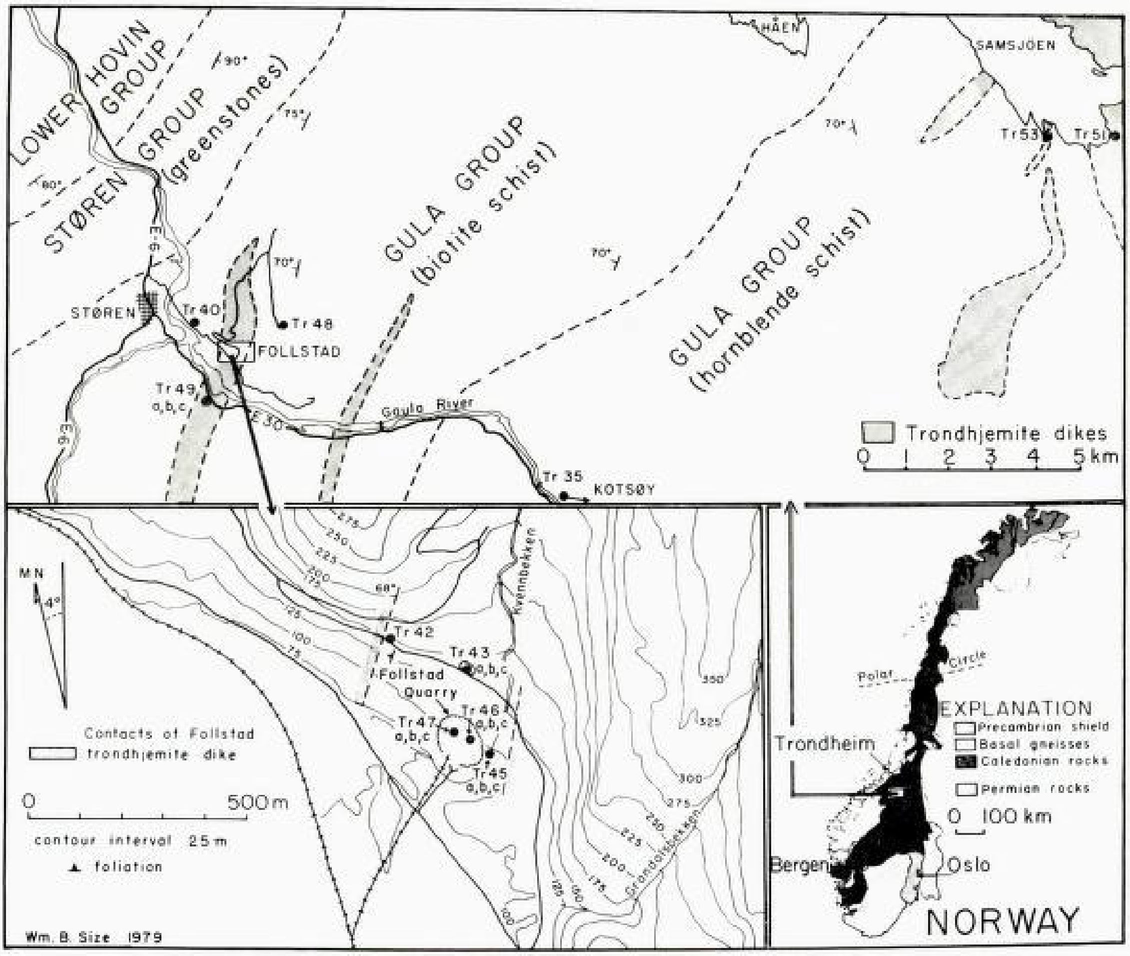

Fig. 1. General geologic maps and two sample location maps of the study area in the Central Norwegian Caledonides, south of the city of Trondheim. Geology after Wolff (1976). Sample number locations listed below Table 1.

The general study area is in the Trondheim region (Fig. 1), which is considered to represent the type area of the 'eugeosynclinal zone' of the Norwegian Caledonides (Strand & Kulling 1972). Petrochemistry of the eugeosynclinal magmatic rocks and interpretation of their relationships to the tectonic history of the Trondheim region have been discussed by Gale & Roberts (1972, 1974), Loeschke (1976) and Dypvik (1977).

The area reported in this study is located 45 km south of Trondheim (Fig.1) near the town of Støren (10°20′E. Long., 63°N. Lat.). The major area of interest is 2 km east of Støren at the Follstad quarry (Fig. 1). Replicate trondhjemite samples were collected from the quarry and from the same instrusive body north and south of the quarry. For comparison, other trondhjemite bodies were sampled from the surrounding area (Fig. 1); these and the quarried trondhjemites will hereafter collectively be called the 'Follstad district trondhjemite'.

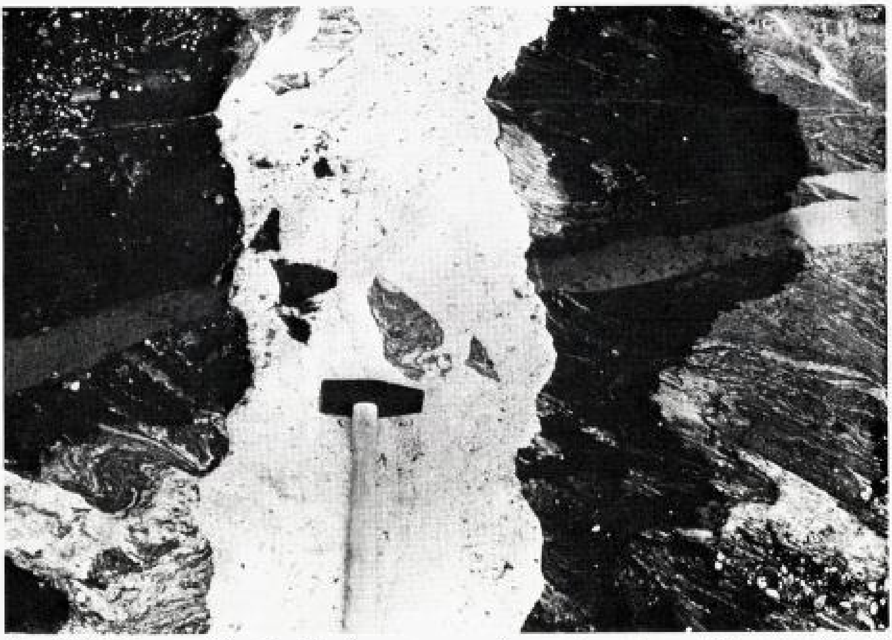

Fig. 2. Multiple trondhjemite dikes in outcrop on the southwest shore of Samsjøen (Tr 53, Fig. 1). Cross-cutting dikes contain xenoliths and are not chilled against the Gula schist country rock. The largest dike represents a composite intrusion.

Stratigraphic and tectonic setting of the Follstad district

Rocks of Late Precambrian to Silurian age in the Trondheim region Caledonides are allochthonous and rest with tectonic discontinuity on the Precambrian granitic and gneissic basement of the Baltic Shield to the east. These Caledonian nappe rocks are mainly eugeosynclinal-type basic volcanic rocks associated with a variety of sedimentary rocks which have undergone a polyphase tectonometamorphic history (Roberts 1967, Wolff 1967, 1976, Roberts et al. 1970, Rui 1972, Olesen et al. 1973). Rock assemblages in the principal nappe unit, the Trondheim Nappe (Wolff 1967), are collectively termed the Trondheim Supergroup (Gale & Roberts 1974) and comprise the basal Gula Schist Group followed by the Støren, Lower Hovin, Upper Hovin and Horg Groups.

In the study area, almandine–amphibolite facies pelitic and quartz schists and subordinate amounts of quartzites of the Gula Schist Group lie tectonically beneath Støren Group greenstones (Fig. 1). Traditionally, the Gula Group has been considered Cambrian (Wolff 1967, Guezou 1975), but a Precambrian age has also been suggested (Roberts 1978). The Follstad trondhjemite bodies intrude these rocks (Fig. 2) and are late-tectonic with respect to the principal Silurian deformation. The overlying Støren Group, in tectonic contact with the Gula, consists of a 2.5 km pile of basic volcanic rocks, locally with pillow structure, which are generally thought to be of Tremadocian to early Arenig age (Roberts et al. 1970). These volcanic rocks represent metamorphosed tholeiite-type basalts indicative of both ocean floor and, in part, island-arc tectonic set-

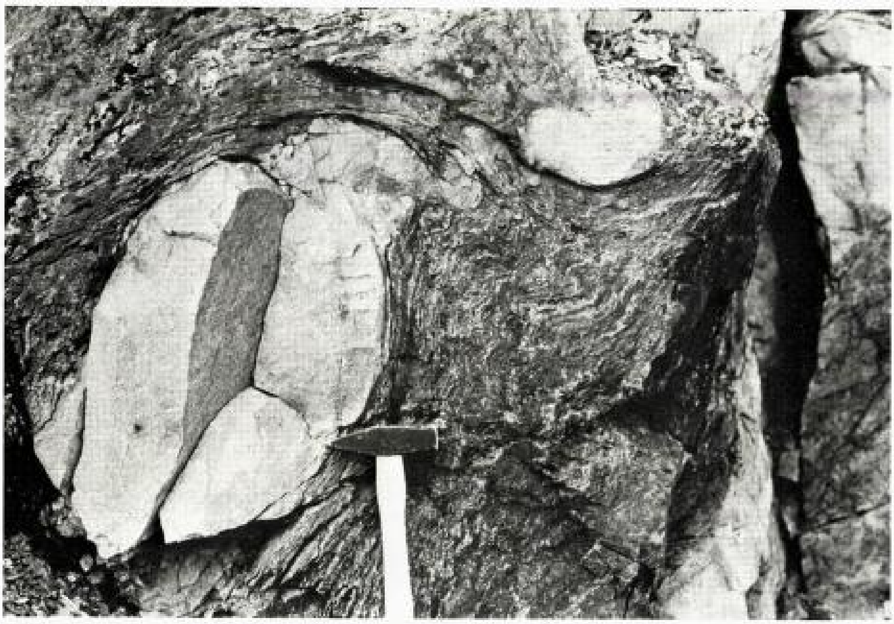

Fig. 3. Deformed trondhjemite dike northeast of the Follstad quarry (Tr 48, Fig. 1). The dike is boudined and sheared with the plane of shear conformable to the adjacent Gula schist country rock foliation. Later small-scale folding is also evident.

tings (Gale & Roberts 1974, Loeschke 1976). The overlying Lower Hovin Group is of Lower to Middle Ordovician age and the Upper Hovin of Upper Ordovician age. These groups are mainly flysch-type sedimentary rocks metamorphosed in lower to upper greenschist facies.

The climactic Caledonian deformation and metamorphism in the region occurred in Middle Silurian (Roberts et al. 1970). Four major phases of deformation have been documented (e.g., F_1–F_4), with the pervasive metamorphic fabric associated with the F_1 isoclinal fold episode. Major thrusting is thought to have occurred towards the end of the F_2 deformation, forming the Trondheim Nappe, but important pre-metamorphic thrusting is also recognized in the obduction of the Støren Nappe (Gale & Roberts 1974). A minimum age for the last metamorphic phase is given as 438 ± 12 m.y. B.P. (Wilson et al. 1973, p. 58). This date is also assigned as a minimum age for the F_3 period of folding.

Folds of F_1 age, together with the associated regional schistosity and local shearing, are present in the Gula Schist Group in the study area, but this F_1 fabric is not present in the trondhjemite bodies. Some of the smaller trondhjemite dikes in the Follstad district were boudinaged and sheared prior to later folding (Fig. 3). However, the main intrusion at the Follstad quarry shows only weak metamorphism and very little mechanical deformation. Therefore, it appears that the trondhjemite dikes in the Follstad district were emplaced at different times with the older, more deformed ones being post-F_1 deformation and the Follstad quarry trondhjemite being possibly post-F_3 deformation. The radiometric ages for these rocks lend support to this conclusion.

Age relationships of trondhjemitic rocks in the Follstad district to country rocks have been outlined by Wilson et al. (1973). The last thermal peak of metamorphism of the Gula Schist, which is the country rock for the Follstad trondhjemite, gives K/Ar dates of 418 and 415 m.y. B.P., whereas the trondhjemite itself yields ages of 407 and 403 m.y. B.P. This trondhjemite could therefore post-date the F_3 folding period.

Field relations of the Follstad district trondhjemite

There were three or four periods of trondhjemite intrusive activity in the Follstad district during the Lower Paleozoic (Roberts 1978). The oldest trondhjemite bodies tend to lie within the regional foliation (Fig. 3). Younger trondhjemite intrusions cut across the foliation and tend to be contorted into later, open folds. An even younger phase is represented by the Follstad body, which has little mechanical deformation, but contains a metamorphic mineral assemblage. The youngest phase of trondhjemite emplacement is represented by the trondhjemite pegmatites. No examples of this last type are found in the study area, but such rocks occur in the Verdal and Stjørdalen valleys as described by Wolff (1960, 1967) and Roberts (1967), and some have been dated by Wilson et al. (1973).

Most of the trondhjemite bodies in the study area are discordant and contain xenoliths of country rock, as can be seen at Samsjøen (Fig. 2) and in the Kvennbekken stream just west of the main Follstad quarry. At contacts with the country rock, trondhjemite is not chilled and shows a fluxion foliation. Many of the dikes show evidence of being multiple intrusions, judging from the sharp discontinuities in grain size and mineral proportions across the dike (Fig. 2). Most of the dikes in the area are not as wide as the Follstad body, and even the smaller ones do not show any chilling, a good indication that the wall rock was at a reasonably high temperature at the time of intrusion.

The Follstad trondhjemite body (Fig. 1) is approximately 375 m wide in the vicinity of the main quarry. It has a general strike of N20°–25°E and dips approximately 68° to the northwest. Its western contact with the Gula phyllite is very sharp with slight silicification and hornfelsic texture as the only contact metamorphic effects. Iron-staining and a weakly developed cataclastic zone were observed at the contact.

Sampling and analytical methods

Thirty samples of trondhjemite were taken from dikes in the Follstad district (Fig. 1; Table 1). Other trondhjemite intrusives from the surrounding region were also sampled for comparison (numbers Tr 35, 40, 42, 48, 51, 53). However, Sample Tr 40 was ultimately determined to be a quartz keratophyre and Sample Tr 51 is a leuco-monzogabbro.

Rock samples for statistical analysis of variance were taken from fresh exposures of trondhjemite in the Follstad district. At each outcrop location (Fig. 1) three rock samples, each weighing 2 kg, were taken 5 m from one another in a triangular arrangement. Two thin-sections were cut perpendicular to one another from each rock sample. The remainder of the sample was then trimmed to remove any surface alteration and then powdered in an

Fig. 4. Block of trondhjemite from the main quarry at Follstad showing biotite streaking and faint foliation. The trondhjemite is usually very homogeneous and massive.

agate mill for chemical analysis. The powders were split for major and minor element analysis.

A detailed petrographic analysis, including the modal analysis method of Chayes (1956), was performed on each thin-section. At least 1000 counts were made on each thin-section and usually the slide was turned around and an additional 1000 counts were made. Results of these modal analyses are presented in Table 1 as means and standard deviations of replicate samples.

Chemical analyses of SiO_2, TiO_2, Al_2O_3 and CaO were done by X-ray fluorescence spectrometry using the method of Padfield & Gray (1971) (J. Sandvik, analyst); total iron, MnO, Na_2O and K_2O were determined by atomic absorption spectrophotometry; FeO by titrimetry; H_2O (total), CO_2, and H_2O^- by gravimetry; and P_2O_5 by spectrophotometry. Results of the chemical analyses are given in Table 2. All analyses listed are believed accurate to $\pm$ 2% or better of the amount present. CIPW norms based on these chemical analyses are given in Table 3.

Trace element analyses were done by neutron activation at the Nuclear Research Laboratory at Virginia Polytechnic Institute and State University (T. F. Parkinson, analyst). Samples were irradiated in a thermal flux of 1.3×10^{12}n/cm^2-sec, using high resolution Ge-(Li) detectors. U.S. Geological Survey standard rocks, BCR-1, AGV-1, G-2 and GSP-1 were used as standards. Results of trace element analyses are given in Table 4.

Petrography

TEXTURE AND STRUCTURE

Follstad trondhjemite is whitish in color, speckled with dark brown biotite. In hand-specimen it is homogeneous and massive, having only a weak foliation which is defined by biotite and muscovite (Fig. 4) and which strikes N40°E and dips to the northwest. Prominent jointing strikes N5°–12°E and dips 45°–47° to the northwest.

Trondhjemite samples from dikes in the Follstad district are holocrystalline, medium-grained (0.5–1.5 mm) and hypidiomorphic–granular in texture (Fig. 5). As the degree of alteration increases, the rock texture grades into allotriomorphic–granular. Slightly metamorphosed varieties of trondhjemite have a weakly developed granoblastic texture. Further details of metamorphic recon-

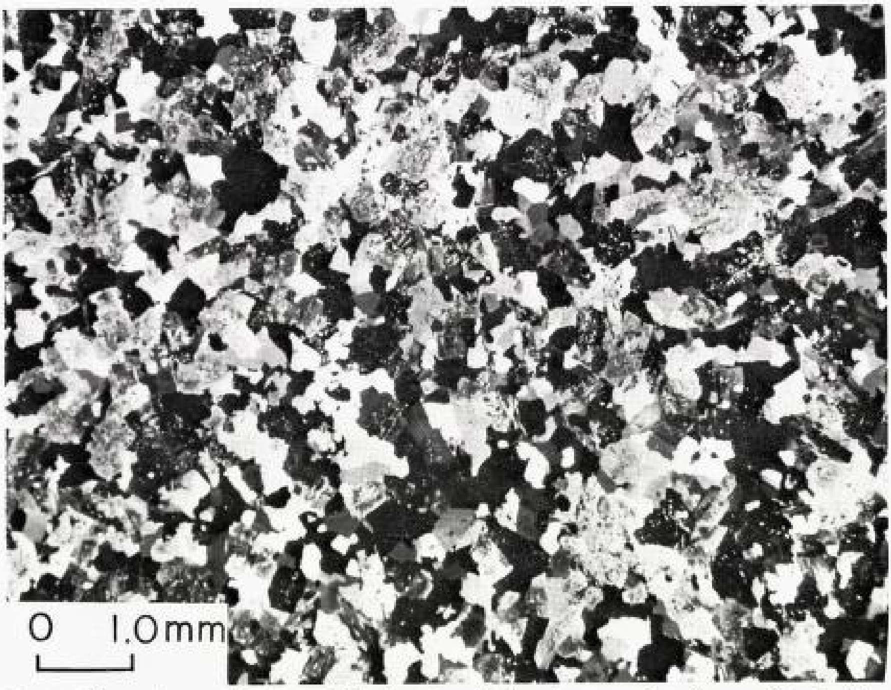

Fig. 5. Photomicrograph of trondhjemite from Follstad quarry showing typical hypidiomorphic–granular texture with slight overprinting of a granoblastic, recrystallized texture. Crossed-polars.

stitution of the rocks are given in the section on evidence of metamorphism. Alteration of trondhjemite ranges from slight to moderate and is best seen in thin-section. Cores of plagioclase are commonly partially replaced by sericite, epidote and calcite.

Three different types of mineral clusters are present in the rock: (1) mafic clusters, consisting of epidote, biotite, muscovite, sphene and zircon; (2) quartz clusters; and (3) clusters of plagioclase laths with small angular–interstitial spaces filled with albite. These clusters appear more indicative of magmatic synneusis than metamorphic segregation.

In thin-section, biotite and muscovite are commonly bent and crudely aligned to define a faint foliation (Fig. 4). Locally quartz and plagioclase have been microsheared into a mortar structure. Grain boundaries rarely interlock; rather, most are straight to slightly curved, indicating recrystallization. Overgrowths of albite on plagiclase are conspicuous, even in hand-specimen.

MODAL MINERAL DESCRIPTION

Trondhjemites from the Follstad district show a very small degree of modal variation. The composite average for 30 samples (Table 1) can therefore be viewed as representative of the entire intrusion. A triangular plot of the relative modal percentages of quartz, alkali feldspar and plagioclase (Fig. 6) shows that most samples plot in the tonalite field and none of them in the trondhjemite

Table 1. Means and standard deviations (in parentheses) of replicate modal analyses (1000 counts) of trondhjemitic rocks from the Follstad district, central Norway. Sample locations listed below and shown on Fig. 1

	Follstad Quarry				
Series	Tr 43	Tr 45	Tr 46	Tr 47	Tr 49
Number of Samples	6	6	6	6	6
Quartz	25.5 (3.9)	27.2 (1.3)	26.0 (0.8)	25.3 (1.3)	26.3 (1.2)
K. Feldspar	1.0 (0.2)	1.8 (1.2)	2.4 (0.8)	2.7 (0.2)	8.6 (1.7)
Plagioclase	59.4 (4.1)	58.4 (2.2)	57.2 (2.3)	60.5 (2.4)	53.4 (6.4)
Muscovite	6.2 (1.2)	5.8 (0.8)	6.2 (1.2)	4.5 (0.9)	7.1 (3.6)
Biotite	2.4 (1.1)	1.8 (1.1)	2.6 (0.8)	2.0 (0.5)	0.4 (0.3)
Magnetite	tr.	tr.	tr.	tr.	tr.
Apatite	tr.	tr.	tr.	0.1	tr.
Sphene	0.2	0.3	0.4	0.3	tr.
Zircon	tr.	tr.	tr.	tr.	tr.
Epidote	4.8 (0.8)	4.6 (0.8)	5.1 (0.8)	4.5 (0.9)	3.1 (1.2)
Chlorite	tr.	tr.	tr.	—	0.6
Calcite	tr.	tr.	tr.	—	tr.
TOTAL	99.5	99.9	99.9	99.9	99.5
Density g/cm^3	2.69 (0.01)	2.68 (0.01)	2.69 (0.01)	2.68 (0.01)	2.67 (0.01)

Location of trondhjemite samples (grid references refer to 1:50,000 map sheet M711 1621 III 'Støren'):

Tr 43 (a, b, c) — Follstad, small quarry located 125 m to the north of main quarry (670 901).

Tr 45 (a, b, c) — Follstad, 25 m east of main quarry in Kvenbekken Stream (670 899).

Tr 46 (a, b, c) — Follstad, main quarry, center section of quarry wall face (669 899).

Tr 47 (a, b, c) — Follstad, main quarry, western margin of quarry wall face (669 899).

Tr 49 (a, b, c) — Follstad area, road outcrop approximately 3 km southeast of Støren, on Highway 30 (665 887).

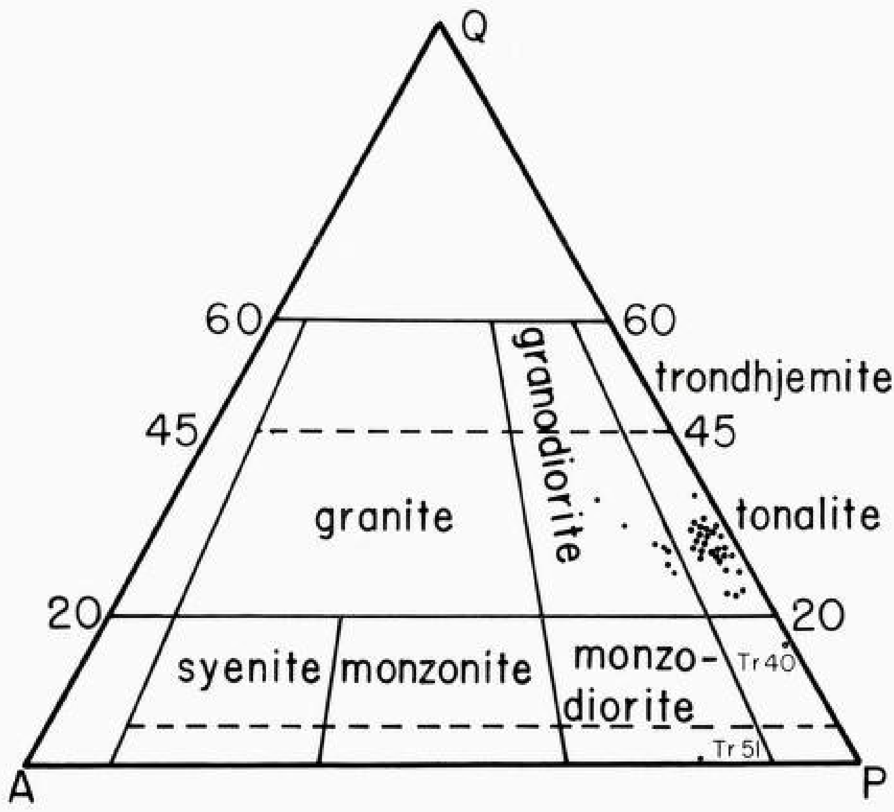

Table 1 (Continued)

Pooled Tr 43–49 30	Surrounding Region					
	Tr 35	Tr 40	Tr 42	Tr 48	Tr 51	Tr 53
26.1 (2.2)	25.1	12.3	26.0	15.8	tr.	21.9
3.3 (2.9)	4.2	—	2.9	2.1	17.5	3.9
57.8 (4.3)	60.3	63.8	60.2	50.9	81.0	69.2
5.9 (2.0)	6.5	10.3	4.2	20.8	1.1	0.3
1.9 (1.0)	3.2	—	1.8	—	—	tr.
tr.	tr.	0.4	tr.	tr.	—	tr.
tr.	tr.	—	tr.	—	—	tr.
0.2	—	0.9	tr.	0.1	0.1	—
tr.	—	—	tr.	—	—	—
4.4 (1.1)	—	4.8	4.9	—	—	—
tr.	—	3.4	—	4.8	0.2	4.7
tr.	0.7	4.0	—	5.6	0.1	—
99.6	100.0	99.9	100.0	100.1	100.0	100.0
2.68 (0.01)	2.66	2.70	2.66	2.66	2.61	2.62

Tr 35 — Kotsøy, approximately 0.5 km south of the village in a road outcrop.

Tr 40 — Follstad area, road outcrop at intersection of new bridge over Gaula River and road tunnel from Støren to Follstad (660 904). Quartz keratophyre.

Tr 42 — Follstad district, approximately 8 m from western contact of trondhjemite with Gula phyllite (668 901).

Tr 48 — Follstad district, approximately 1.2 km northeast of main quarry. Small sheared dike (679 905).

Tr 51 — Samsjøen Lake, southeast shore, near water outlet tunnel (867 953). Leuco-monzogabbro.

Tr 53 — Samsjøen Lake, southwestern shore, near water outlet tunnel (848 955).

field. Chemically and mineralogically, however, the rocks from the Follstad district are true trondhjemites; in fact, the quarry at Follstad is viewed as the type-locality of trondhjemite, as first proposed by Goldschmidt (1916). The boundary between tonalite and trondhjemite (45% quartz), as shown in the diagram, is not truly representative of the natural system. Most trondhjemitic rocks from other areas would also plot in the tonalite field of Streckeisen (1967). The mafic index of tonalite is higher than that of trondhjemite and the SiO_2 content is lower, so there is actually a real difference between the two rock types.

Plagioclase: Plagioclase is present as subhedral laths ranging from 0.1–2.0 mm and averages 58% by volume of the rock. Crystals show no apparent orientation apart from their tendency to form clusters or glomeroclasts. Composition of plagioclase ranges from An_{21} to An_{30} with an average composition of An_{25} (calcic oligoclase). Compositional zoning is present, but is oscillatory and not strongly developed. Twinning is also present, but is not abundant.

Fig. 6. Triangular plot of the relative percentages of modal quartz (Q), alkali feldspar (A) and plagioclase (P) for the trondhjemitic rocks in the Follstad district listed in Table 1. Rock name field boundaries after Streckeisen (1967, p. 160).

Two stages of plagioclase genesis have been distinguished in the trondhjemites. Most plagioclase laths have a mantle of albite that is usually unzoned and untwinned. Albite may be a reaction mineral developed during metamorphism, as discussed below. In the plagioclase rimmed by albite, only the core of the lath is altered. Alteration is moderate with sericite, calcite, and epidote being the most common replacement minerals. There is a strong negative correlation between plagioclase and microcline ($r = -.88$); as the anorthite content of the plagioclase decreases, the microcline content increases. This variation indicates that some differentiation has taken place in these rocks.

Alkali feldspar: Alkali feldspar is present as small, angular, interstitial grains ranging from 0.1 to 0.7 mm. Average modal content is only 3.3%. Grains are mainly microcline, although some have a patchy or braid–perthitic intergrowth with albite. Microcline composition is $Or_{62} : Ab_{38}$ determined by X-ray diffraction. Alteration of alkali feldspar is not extensive and sericite is the most common secondary mineral. The strong negative correlation between alkali feldspar and both epidote and biotite ($r = -.92$) indicates a reaction between them.

Quartz: Quartz is present as anhedral interstitial grains ranging from 0.1 to 1.5 mm. Quartz is abundant, with an average modal content of 26%, and tends to form in clusters commonly showing a faint granoblastic texture with straight to slightly curved grain boundaries meeting at 120°. Some grains have undulatory extinction but most grains are unstrained, demonstrating that the quartz may be partially recrystallized.

Biotite: Biotite is the major mafic mineral in trondhjemite. It is present as subhedral plates ranging from 0.2 to 1.6 mm. The modal content averages only 1.9% of the rock. Grains are commonly bent and often found associated with muscovite in subhedral arrangement. Biotite is also found in mafic clusters dispersed unevenly throughout the rock. Alteration is variable with chlorite as the most common secondary mineral. There are a few chlorite pseudomorphs of biotite. Epidote and biotite are closely associated in the trondhjemite and sometimes show a mutual reaction texture. The correlation coefficient is positive between epidote and biotite ($r = .99$).

Muscovite: Muscovite is also present as small subhedral plates in the rock. Its modal average is 5.9%. Some of the secondary sericite in the plagioclase has a recrystallized texture and a sagenitic structure. Where muscovite shows a mutual reaction texture with other minerals it has a spongy texture.

Epidote: Epidote, which is a metamorphic mineral in these rocks, is present as well-crystallized, subhedral to euhedral prisms locally with compositional zoning. Optical properties show it to be an iron-poor epidote having about 10 wt.% Fe_2O_3. As already noted there is a mutual reaction texture between epidote and biotite. Epidote is also associated with small grains of magnetite and has been observed as an alteration product after plagioclase. Epidote modal variability is related to the degree of recrystallization in the trondhjemite. The trondhjemite that is more highly recrystallized typically contains greater amounts of epidote.

Accessory minerals: Where present, magnetite is present in trace amounts and commonly forms small subhedral grains associated with epidote. Sphene is present as subhedral to euhedral prisms, up to 1 mm across: it has also been observed included in biotite. Zircon is present as small, slightly rounded grains commonly included in biotite and showing pleochroic haloes. Apatite is ubiquitous as small, euhedral grains evenly dispersed throughout the rock. In rare instances apatite has a very pale green pleochroism.

ANALYSIS OF MINERAL VARIANCE

The Follstad district trondhjemite samples were taken from a variety of locations to permit a statistical analysis-of-variance test. Modal variation for the most abundant minerals was determined first within each of the sample outcrops. This comparison of local versus regional variation permits the degree of homogeneity of the trondhjemite to be determined. An F-test, which is a test for equivalence of means, was used to determine if variation at each level of comparison is significant or is due to random variation in the samples. The calculated F-value for each mineral was compared with the tabular F-value at the 5% level of significance. Although details of the analysis-of-variance tests are not included here, they are available from the author on request.

Within each outcrop (local fluctuation) major mineral variation was determined to be insignificant in the trondhjemitic rocks from the Follstad district. On the regional scale of comparison (among outcrops) the mineral also have an insignificant amount of variation. This shows that the trondhjemite is homogeneous throughout its extent. Means and standard deviations for the modal percentages given in Table 1 also show only a small amount of mineral fluctuation in the trondhjemites. This homogeneity demonstrates that no older or younger fractionates or differentiates are present in the sampled population. A high degree of compositional homogeneity also seems to be characteristic of trondhjemites worldwide. The possible genetic importance of this homogeneity will be discussed in the section on petrogenesis.

Evidence of metamorphism

Textural and mineralogical features of the Follstad trondhjemite collectively indicate that the rock has undergone metamorphism. Quartz occurs in granoblastic clusters with slightly curved to straight grain boundaries meeting at 120°. Biotite, muscovite, epidote and microcline show mutual reaction textures between grains. Small, euhedral, incipient epidote shows good crystal form, with the amount of epidote greater in the trondhjemites that show evidence of advanced recrystallization. Albite rims on plagioclase are untwinned and unzoned. The amount and perfection of twinning in the plagioclase is less in trondhjemites showing advanced recrystallization.

Table 2. Means and standard deviations (in parentheses) of replicate major element chemical analyses in weight per cent from trondhjemite samples listed in Table 1. (J. Sandvik, analyst)

	Follstad Quarry									
Series	Tr 43		Tr 45		Tr 46		Tr 47		Tr 49	
No. of Samples	3		3		3		3		3	
SiO_2	71.58	(0.14)	71.54	(0.25)	71.24	(0.26)	70.84	(0.09)	72.14	(0.33)
TiO_2	0.25	(0.01)	0.23	(0.02)	0.24	(0.01)	0.23	(0.01)	0.20	(0.0)
Al_2O_3	16.38	(0.12)	16.15	(0.27)	16.34	(0.20)	16.46	(0.22)	15.70	(0.25)
Fe_2O_3	0.64	(0.03)	0.52	(0.03)	0.60	(0.04)	0.66	(0.03)	0.48	(0.08)
FeO	0.79	(0.02)	0.76	(0.05)	0.82	(0.02)	0.81	(0.03)	0.76	(0.04)
MnO	0.020	(0.002)	0.022	(0.002)	0.019	(0.002)	0.020	(0.002)	0.024	(0.002)
MgO	0.51	(0.02)	0.46	(0.02)	0.50	(0.01)	0.52	(0.02)	0.40	(0.02)
CaO	2.83	(0.26)	2.71	(0.04)	2.87	(0.02)	3.04	(0.13)	2.16	(0.07)
Na_2O	5.35	(0.07)	5.37	(0.15)	5.38	(0.18)	5.37	(0.16)	5.19	(0.12)
K_2O	1.27	(0.04)	1.42	(0.07)	1.33	(0.04)	1.30	(0.01)	1.62	(0.17)
P_2O_5	0.09	(0.01)	0.08	(0.01)	0.08	(0.01)	0.09	(0.0)	0.08	(0.0)
H_2O^+	0.52	(0.02)	0.50	(0.10)	0.45	(0.0)	0.70	(0.11)	0.73	(0.26)
H_2O^-	0.02	(0.01)	0.03	(0.01)	0.03	(0.01)	0.07	(0.02)	0.04	(0.01)
CO_2	—		—		—		—		—	
TOTAL	100.25		99.79		99.90		100.11		99.52	

Table 3. CIPW normative mineral percentages calculated from chemical analyses listed in Table 2. Non-normative CO_2 and H_2O subtracted from totals. The differentiation index is that of Thornton & Tuttle (1960)

	Follstad Quarry					Pooled
Series	Tr 43	Tr 45	Tr 46	Tr 47	Tr 49	Tr 43–49
Number of Samples	3	3	3	3	3	15
Q	28.75	28.37	27.86	27.32	30.81	28.62
C	1.26	1.06	1.01	0.91	2.02	1.25
or	7.49	8.37	7.88	7.66	9.57	8.19
ab	45.30	45.41	45.55	45.47	43.97	45.14
an	13.49	12.88	13.71	14.48	9.22	12.76
hy	1.82	1.77	1.90	1.88	1.70	1.81
mt	0.93	0.76	0.87	0.96	0.70	0.84
il	0.47	0.44	0.46	0.45	0.38	0.44
ap	0.20	0.19	0.19	0.21	0.19	0.20
Differentiation Index	81.54	82.15	81.29	80.45	84.35	81.95

As biotite and epidote contents increase, the other minerals — microcline, chlorite, muscovite and quartz — show a corresponding decrease These characteristics suggest that the trondhjemite has been subjected to regional metamorphism and converted into the assemblage muscovite–biotite–quartz–albite–epidote, an assemblage representative of the upper greenschist facies (Winkler 1965). The degree of metamorphism was first reported by Goldschmidt (1916). At the top of the biotite zone of regional metamorphism, albite can coexist with plagioclase of composition An_{23}. This feldspar coexistence is displayed in the Follstad district trondhjemite as albite rims on plagioclase laths of An_{25}. The upper greenchist facies assemblage present in these rocks is of lower grade than the almandine–amphibolite facies assemblage present in the Gula Group, but of higher grade than the lower greenschist facies assemblage present in the Hovin Groups.

able 2 (Continued)

pooled Tr 43–49 15	Surrounding Region Tr 35	Tr 40	Tr 42	Tr 48	Tr 51	Tr 53
71.47 (0.48)	73.89	64.98	72.33	67.29	65.83	71.19
0.23 (0.02)	0.07	0.30	0.17	0.26	0.05	0.25
16.20 (0.34)	15.72	17.37	16.17	16.08	19.94	16.36
0.58 (0.08)	0.37	0.52	0.52	0.14	0.03	0.21
0.78 (0.04)	0.73	1.08	0.62	1.43	0.08	1.23
0.021 (0.002)	0.029	0.025	0.022	0.021	0.002	0.026
0.48 (0.05)	0.23	1.09	0.36	1.14	0.02	0.82
2.72 (0.33)	2.11	3.94	1.91	3.57	2.15	2.57
5.33 (0.14)	4.62	5.92	5.66	4.42	7.34	5.68
1.39 (0.15)	1.89	1.82	1.81	2.00	4.25	0.91
0.08 (0.01)	0.06	0.10	0.08	0.09	0.05	0.11
0.58 (0.16)	0.31	1.28	0.52	0.94	0.04	0.69
0.04 (0.02)	0.01	0.06	0.03	0.06	0.03	0.11
—	—	1.46	—	1.92	—	—
99.90	100.04	99.94	100.20	99.36	99.81	100.16

Surrounding Region Tr 35	Tr 40	Tr 42	Tr 48	Tr 51	Tr 53
34.64	17.28	27.86	29.11	2.66	27.46
2.38	2.12	1.62	4.82	—	1.62
11.17	10.76	10.70	11.82	25.12	5.38
39.09	50.09	47.89	37.40	62.11	48.06
10.08	9.66	8.95	4.99	8.91	12.03
1.55	3.82	1.37	4.96	0.19di / 0.50 wo	3.76
0.54	0.75	0.75	0.20	0.04	0.30
0.13	0.32	0.32	0.49	0.09	0.47
0.14	0.19	0.19	0.21	0.12	0.25
84.90	78.13	86.45	78.33	89.89	80.90

Major element geochemistry

Major element analyses and CIPW norms for the Follstad district trondhjemite showed relatively high contents of silica and sodium compared with the low content of potassium (Tables 2 & 3). The Na_2O/K_2O ratio averages 3.8. The high Na_2O content relative to K_2O is in part due to alteration and metasomatism (sericitization, albitization). The trondhjemite averages 16.2% Al_2O_3, which classifies it as the high alumina type (greater than 15% Al_2O_3) as defined by Barker & Arth (1976). In contrast to trondhjemites worldwide, the Follstad district trondhjemite has relatively low amounts of total iron (average 1.36%) and the FeO/Fe_2O_3 ratio is relatively high at 1.34. The total iron-to-magnesium ratio is also low (average 2.83) in comparison to that in trondhjemites from

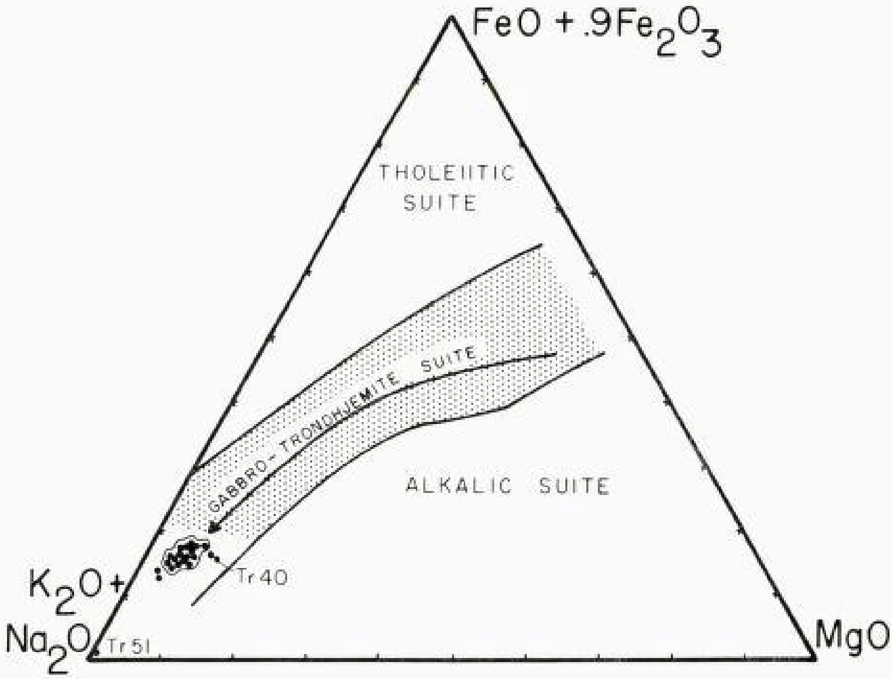

Fig. 7. AFM plot for the Follstad trondhjemites. Gabbro–trondhjemite trend after Barker & Arth (1976, p. 598). Shaded area represents locations for the eugeosynclinal magmatic rocks in the Trondheim region (Loeschke 1976, p. 45). The shaded area is also representative of the calc-alkaline rock suite. Group of points outlined represents trondhjemite samples from the Follstad quarry.

other regions (Arth & Hanson 1975). The low iron-to-magnesium ratio, together with the low potassium content, indicates that these rocks represent a primitive magma type that has not further differentiated to leucogranodiorite or leucomonzonite by the removal of plagioclase.

A comparison of the K_2O content versus SiO_2 shows that the Follstad trondhjemites plot in the continental trondhjemite field of Coleman & Peterman (1975, p. 1105), with more K_2O than the oceanic plagiogranites. There is not enough range in the compositions of these rocks to show any trend in crystallization. However, the correlation coefficients for major oxides show that as silica and potassium increase, there is a corresponding decrease in calcium, iron, magnesium, sodium and aluminum.

Trondhjemites plotted on an AFM diagram (Fig. 7) are at the low temperature end of the gabbro–trondhjemite trend (Barker & Arth 1976), which is a sub-trend of the more general calc–alkaline suite of Green & Ringwood (1968). Other eugeosynclinal magmatic rocks from the Trondheim region are represented by the shaded area of Fig. 7 (Loeschke 1976, p. 45). This field also shows a calc–alkaline trend, with Follstad trondhjemite at the most differentiated end. This demonstrates that Follstad district trondhjemite may represent the lowest temperature melt of a basaltic magma.

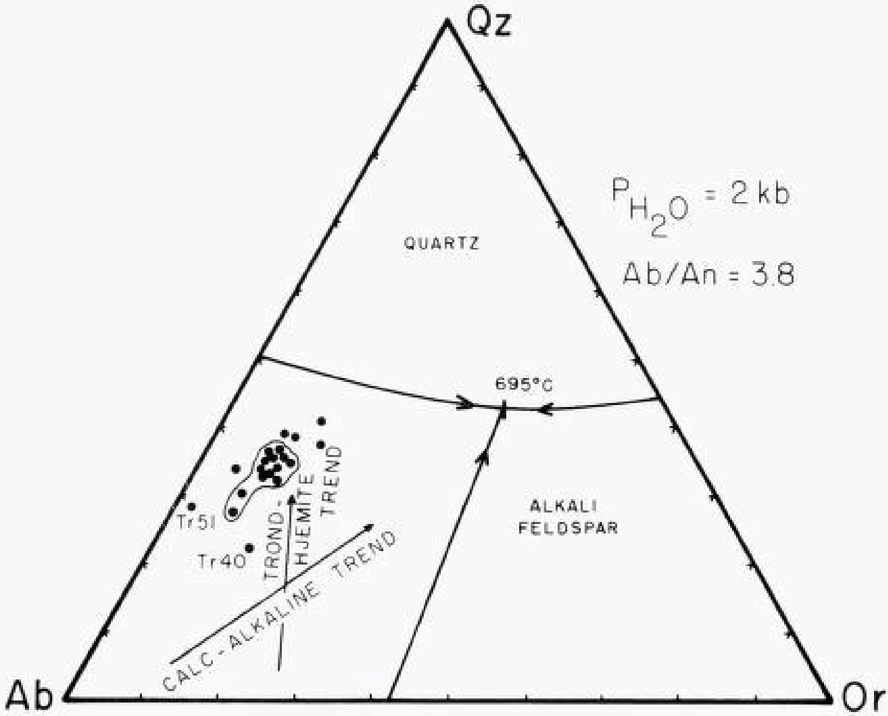

Fig. 8. Triangular plot of the relative percentages of normative quartz (Q), orthoclase (Or) and albite (Ab) for the Follstad trondhjemites. Eutectic position for Ab/An = 3.8, P_{H2O} = 2 kb, projected from the Q–Ab–Or–An–H_2O system (Winkler 1965, p. 188). Calc-alkaline trend and gabbro–trondhjemite trend after Barker & Arth (1976). Group of points outlined represents trondhjemite samples from the Follstad quarry.

The triangular plot of normative Qtz–Or–Ab (Fig. 8) shows that all of the trondhjemites plot away from the eutectic melting composition in the Or–Ab–An–Qtz–H_2O system. The trondhjemites, as plotted, show a slight decrease in An content that corresponds with distance away from the minimum melt composition, as would be expected in a fractional crystallization process or a partial melting process. However, rocks so poor in K_2O, yet rich in SiO_2, cannot be easily accounted for unless a low K_2O-bearing parent material is invoked. Also shown in Fig. 8 are the eutectic point and cotectic lines for $P_{H_2O} = 2$ kb at an Ab/An ratio of 3.8, which is near the Follstad district trondhjemite ratio of 3.5. If the trondhjemite represented a melt composition, it would have crystallized at approximately 750°–800°C. The calc–alkaline crystallization trend shows potassium enrichment towards the lower temperatures (Fig. 8), whereas the gabbro–trondhjemite trend does not (Barker & Arth 1976). Arth & Barker (1978) would identify this as the calc–alkaline trondhjemitic trend. This deviation is explained by differences in parental material (tholeiitic versus alkalic basaltic magma).

On a triangular plot of normative Or–Ab–An (Fig. 9) all the samples from the Follstad district are in the trondhjemite field as defined by O'Connor (1965). Sodium enrichment and potassium depletion in these rocks is obvious.

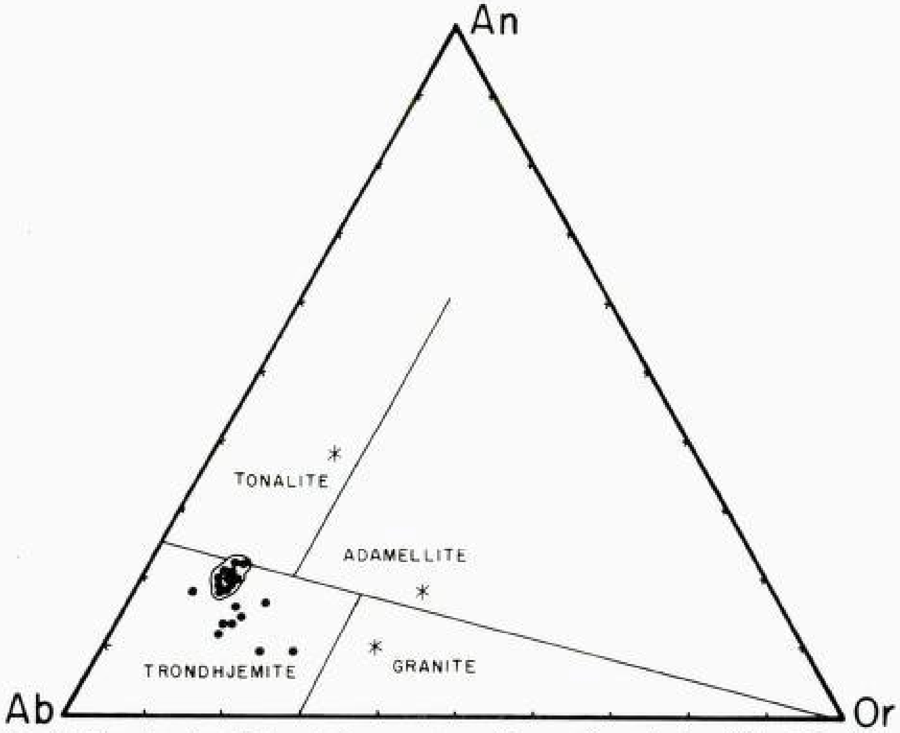

Fig. 9. Triangular plot of the relative percentages of normative orthoclase (Or), albite (Ab) and anorthite (An) for the Follstad trondhjemites. Rock name field boundaries after O'Connor (1965). Asterisks represent average composition for each major rock from LeMaitre (1976). Group of points outlined represents trondhjemite samples from the Follstad quarry.

Average tonalite (LeMaitre 1976) has a much higher An content than the trondhjemitic rocks (Fig. 9).

Almost all of the trondhjemites are peraluminous, having corundum in their norms (Table 3). This indicates that trondhjemites have modal mica or that the rocks have undergone chemical alteration, or both. The trondhjemites with the highest proportion of normative corundum (Tr 48, Tr 49a, b, c,) show the highest degree of alteration in thin-section. However, partial fusion of a tholeiitic basalt yields a melt of trondhjemitic composition that is also corundum-normative. This alternative will be examined more closely in the section on petrogenesis.

Trace element geochemistry

Selected trondhjemite samples from the Follstad district were analyzed for trace elements and rare earth elements (Table 4). Barium content ranges from 240 to 350 ppm, whereas the rubidium content is much lower, ranging from 20 to 43 ppm. This enriched proportion of barium to rubidium indicates a more primitive trondhjemite, as reported by Bouseily & El Sakkary (1975). As the Or/Ab ratio increases in Follstad trondhjemite, the Ba content increases. This

Table 4. Trace element and rare earth element analyses, in ppm, from trondhjemitic samples listed in Table 1. Analytical methods given in text

	Follstad Quarry				Surrounding Region	
Sample	Tr 43	Tr 46	Tr 47	Tr 49	Tr 51	Tr 53
La	8.4	5.5	6.4	6.5	1.2	9.9
Ce	27	19	19	21	4.5	24
Sm	5.0	4.3	4.4	5.3	1.1	7.3
Yb	0.30	0.20	0.15	0.13	0.27	0.63
Lu	0.038	0.05	0.05	0.03	0.03	0.03
Rb	20	39	43	28	60	49
Ba	240	290	280	350	600	92
U	0.43	0.41	0.69	0.58	0.56	0.69
Th	4.4	2.7	3.2	4.2	0.87	3.2
Cr	1.7	ND	3.0	ND	0.78	5.8
Ni	62	91	ND	ND	30	340
Ba/Rb	12.0	7.4	6.5	12.5	10.0	1.9
K/Rb	470	246	279	428	600	186
La/Lu	221	110	128	216	40	330

correlation has also been reported by McCarthy & Hasty (1976). The Ba/Rb ratio in the trondhjemite from the Follstad district ranges from 6.5 to 12.5, which is similar to that in other trondhjemites located in similar tectonic settings. For example, in Nova Scotia, trondhjemite from the Shelbourne pluton has a Ba/Rb ratio from 4.1 to 4.9 (Albuquerque 1977, p. 6). In the Uusikaupunki complex of southwest Finland the Ba/Rb ratio is from 3.1 to 11.6 for trondhjemites (Arth & Barker 1978). The K/Rb ratios are also similar in these three areas: Follstad, K/Rb = 246–470; Nova Scotia, K/Rb = 207–222; Finland, K/Rb = 261–320.

A plot of rare earth element (REE) abundances normalized to the average chondrite abundances (Haskin et al. 1968), shows that the Follstad trondhjemite is fractionated with HREE depleted relative to the LREE (Fig. 10). The highly fractionated nature of REE in Follstad district trondhjemite is shown in the La/Lu ratio, which ranges from 110 to 221. The LREE are from 20 to 30 times chondrites, whereas the HREE abundances in the trondhjemites are almost chondritic. This pattern of highly fractionated REE is characteristic of the high Al_2O_3-type trondhjemite defined by Barker & Arth (1976). One analysis of REE in a trondhjemite from the Follstad quarry has no Eu anomaly (Barker, pers. comm., 1978). This indicates that plagioclase was not an important cumulate phase if trondhjemite was derived by fractional crystallization; or it indicates that plagioclase was remelted if trondhjemite had its origin due to partial anatexis. The REE element pattern for Follstad district trondhjemite is similar to the pattern in trondhjemites from Nova Scotia (Albuquerque 1977) and Colorado (Barker et al. 1976), except that these trondhjemites from Nova Scotia and Colorado have a negative Eu anomaly.

Crystal fractionation of amphibole would deplete a magma in all REE except

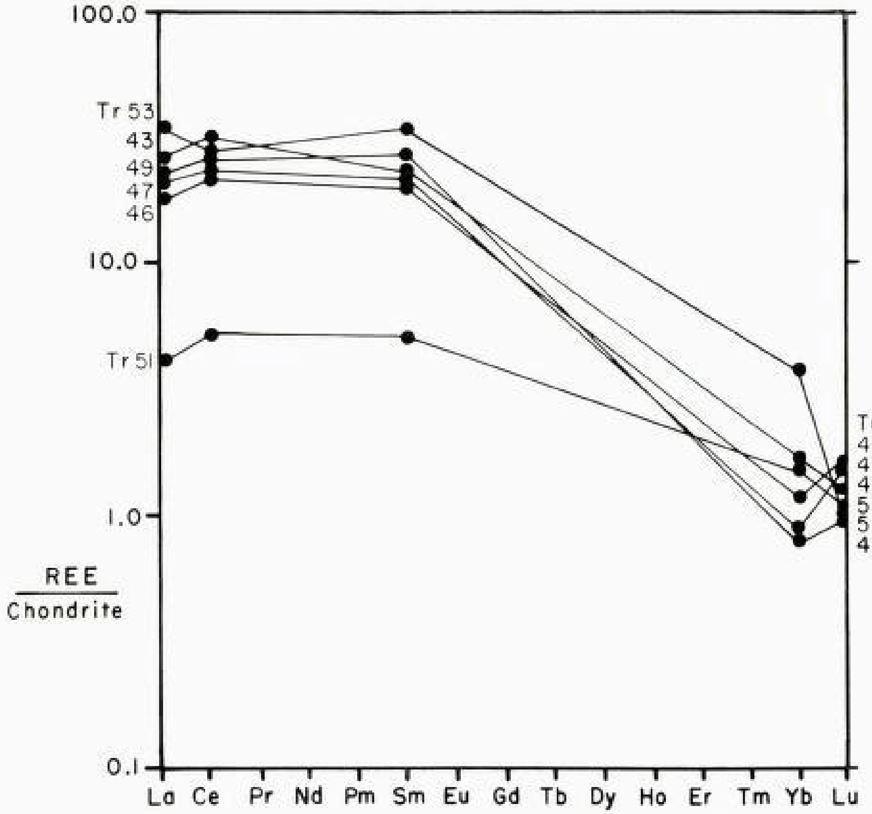

Fig. 10. Rare earth element plot for the Follstad trondhjemites. REE values are normalized to average chondrite values of Haskin et al. (1968). Numbers Tr 43, 46, 47, 49 from the Follstad quarry area; Tr 51, 53 from Samsjøen.

cerium (Arth & Barker 1976, p. 535). This pattern is present in the Follstad district trondhjemite, although no associated amphibole-bearing cumulate rocks are present in the area. In Finnish trondhjemites, where amphibole is present, a cerium enrichment is again present in the REE pattern (Arth & Barker 1978).

Relationship of trondhjemite composition to tectonic setting

Interpretation of the Caledonian tectonic history of the Trondheim region in a plate tectonic context has been given by Gale & Roberts (1974), whereas Gee (1978) has presented a broader scheme for development of the entire nappe system in the region. Only the more salient points pertaining to the origin of trondhjemitic rocks need be mentioned here.

Of major importance is recognition of the allochthonous character of Lower Paleozoic rocks in the Trondheim region. Development of the eastward-moving nappe pile can be dated stratigraphically to post-Llandoverian but prior to the Old Red Sandstone molasse sedimentation (Strand 1960, Roberts et al. 1970, Gee & Wilson 1974, Gee 1978). These allochthonous rocks now rest upon continental gneiss and granite of the Baltic Shield. Estimated amounts of east-southeastward displacement of the nappes range from 200 to 250 km for the Støren Nappe, to 400–500 km of cumulative nappe movement (Gale & Roberts 1974), to over 1000 km for the entire nappe sequence in the region (Gee 1978). Gee suggests that a considerable amount of this apparent displacement may be due to stretching.

Trondhjemites are found entirely within the allochthon, and structural studies and radiometric ages show that they both pre-date and post-date the time of nappe formation and thrusting. This apparent contradiction indicates there were several periods of trondhjemite intrusive activity in the region. Field relations support this by the presence of trondhjemitic intrusives showing differing degrees of deformation.

Gale & Roberts (1974) have shown that the Støren Nappe was obducted upon the Gula rocks pre- to syn-F_1. Therefore, Follstad district trondhjemites, which are virtually undeformed and intrude the Gula Schist Group, post-date the juxtaposition of the Støren Nappe upon the Gula rocks. However, the Gula rocks may also represent an allochthonous unit, provisionally termed the Gula Nappe by Roberts (1978). Later the Støren Nappe and Gula moved as one unit, the Trondheim Nappe.

Translation of the nappes above their respective basal overthrusts has produced a telescoping of possible time-correlative rock sequences consisting of, from east to west: continental-margin facies, back-arc facies, island arc-trench facies and ocean floor volcanic rocks. As noted earlier, the thick sequence (2.5 km) of the Støren Group consists mainly of basaltic greenstone with subordinate clastic rocks, representative of ocean floor and partly island-arc tholeiite (Gale & Roberts 1974, Loeschke 1976). Quartz keratophyre, which is intercalated in the greenstone, may be a differentiate of these basalts or a product of partial fusion of the basalts. The quartz keratophyres are interpreted as being dikes, flows, tuffs and reworked volcano-clastic rocks. They are similar in bulk chemistry to the trondhjemites and may represent their nearer surface equivalents.

Interpretation of tectonic development of the region (Gale & Roberts 1974) shows that from Late Precambrian to latest Cambrian time, the Iapetus Ocean was opening, producing ocean ridge basalts at the spreading center. After this early opening, initial stages of closure in Early to Middle Ordovician time led to development of an eastward-dipping subduction zone with the production of an island-arc and back-arc sequence of volcanic rocks and sediments close to the continental margin. Flysch-type sediments were deposited in the back-arc. Continent to continent collision occurred in Mid-Silurian time, resulting in large-scale obduction of the primitive ocean floor based island-arc and back-arc pile onto the continental margin and the Gula Group rocks. Before that, the subduction of oceanic basalt and possibly some sediments beneath the back-arc and continental margin set the stage for partial remelting and fusion of basalt, producing trondhjemitic magma. There was possibly an appreciable time interval (from earliest Ordovician to Mid-Silurian) between the commencement of subduction and the ultimate intrusion of trondhjemite into the nappe pile.

Petrogenetic interpretation

Magmas are mainly derived from fusion of pre-existing rocks. However, the important petrogenetic questions remaining are how far the remobilized magma migrated from its source and what has modified it along the way (fractionation, contamination). This history is difficult to interpret if the source rock, the residual phases or the cumulus phases are not known. Such is the case for the origin of Follstad district trondhjemite.

Several tectonic and genetic settings could have produced trondhjemitic rocks as described from the Follstad district. Three general models, each with refinements, are the basis for the origin of all magmatic rocks: (1) perfect equilibrium melting and/or crystallization, (2) perfect fractional crystallization, and (3) perfect fractional melting. In a natural system none of these models is acceptable, but rather some modification or non-equilibrium process. The distinction between a trondhjemite magma produced by partial melting and one produced by fractional crystallization can be determined from the relative abundance of the derivative rock types. However, Follstad trondhjemites show little evidence of differentiation.

A downgoing slab of oceanic crust in a subduction zone is favorable to equilibrium melting. This setting provides for continual movement of new material into the region of magma generation and the removal of residium by sinking or lateral convection. Characteristics of equilibrium melting include: (1) no separation of melt from crystals once the magma is generated, and (2) little difference among compositions of successive magma generated. These features are common to trondhjemitic rocks in general and to Follstad district trondhjemite in particular. Plagioclase in the rock shows evidence of being an early phase that did not separate from the magma. In addition, the bulk composition of Follstad district trondhjemite has been shown to be extremely homogeneous.

Early ideas on the origin of trondhjemitic rocks included fractional crystallization of basaltic magma, as first proposed by Goldschmidt (1916) for this region of the Caledonides. This model was also proposed for the origin of the gabbro–trondhjemite suite in the Uusikaupunki–Kalanti area, southwest Finland (Hietanen 1943). However, a fractional crystallization model is difficult to apply in the case of Follstad district trondhjemite, because there are significant compositional gaps in the magmatic rock series from the Trondheim region. Loeschke (1976, p. 48), for example, has shown a compositional gap between 55–65% SiO_2 for the magmatic rocks of the region. Also, there are no examples of more highly differentiated rocks than the trondhjemite in the region, even though exposure is relatively good. Compositional gaps in this rock series can better be explained by equilibrium melting of a more primitive source rock.

The trondhjemite trend, as shown in Fig. 8, can be produced by partial melting of rocks in which residual hornblende was the predominant K_2O-bearing phase (Helz 1976, p. 179). Fractional crystallization of basaltic magma would produce a trend towards higher K_2O which is not seen in the gabbro–

trondhjemite trend. Increasing the P_{H_2O} or f_{O_2} wold not change the trend, only the volume of the melt and the composition.

Several models have been proposed for the origin of trondhjemite rocks. However, some of these models as applied to the Follstad trondhjemite are not well supported by the evidence and will be dispensed with first. These include the model that the trondhjemite is an integral part of an ophiolite complex (similar to the oceanic plagiogranite of the Troodos Complex, Cyprus; Coleman & Peterman 1975, p. 1105). This model is unlikely for the simple reason that the trondhjemites are not pre-tectonic. However, the Støren volcanic rocks may still represent a fragment of a dismembered ophiolite (Roberts, pers. comm. 1976). In addition, the K_2O content in an ophiolite oceanic plagiogranite is significantly lower than even the low K_2O content of Follstad trondhjemite. Also, the HREE show an enriched pattern in the Troodos plagiogranite whereas the Follstad trondhjemite shows a depleted HREE pattern.

Another unlikely model for the origin of Follstad district trondhjemite is partial fusion of gneiss and granite of the continental crust margin. The trondhjemite is situated within the Trondheim nappe, which partly rests on such basement, and magma may therefore have been generated or contaminated by fusion of basement rocks. However, the composition of the low temperature melt of such source rocks is enriched in K_2O and even with progressive melting of these rocks at higher temperatures the K_2O content in the melt would not be diluted to resemble the composition of a trondhjemitic magma. Also, data from Peterman & Barker (1976) show a $^{87}Sr/^{86}Sr$ = .7039 from Follstad trondhjemite, which is indicative of a less radiogenic source than continental crustal rocks.

A model of partial melting of greywacke as the source for trondhjemitic magma is not so easily discounted. Flysch-type greywacke and associated volcanoclastic rocks are abundant in Lower Paleozoic rocks of the Trondheim region, and they show differing effects of metamorphism and recrystallization. Experimental studies by Kilinic (1972) on partial melting of greywacke show that a trondhjemitic magma can be produced by this process. Albuquerque (1977) used this model to explain the origin of trondhjemites in the Shelbourne pluton of Nova Scotia. He estimated that trondhjemite melts formed from partial fusion of greywacke above the minimum melt temperature in the Qtz–Ab–Or system at about 715°–730°C, and at a pressure of 5.5–6.5 kb. However, in both of these examples the composition of the trondhjemites is significantly higher in both SiO_2 and K_2O than the Follstad district trondhjemite. Average greywacke, which has about 67% SiO_2 and 2% K_2O, would produce abundant quantities of 'granitic melt' before changing towards a melt of trondhjemitic composition. These granitic rocks are not evident in the Trondheim region. In addition, partial melting of greywacke would produce a melt showing HREE enrichment. Follstad region trondhjemite has a depleted HREE pattern. Wyllie (1977), in his crustal anatexis experiments, concluded that source rocks such as shale and greywacke cannot produce melts of tonalitic or dioritic composition under conditions of normal regional metamorphism.

The high Al_2O_3 content and low K_2O content of Follstad region trondhjemite require a more primitive source such as island-arc basalts or ocean floor, low-K_2O tholeiites. Approximately 10–35% partial melting of low-K_2O basalt would produce a high-Al_2O_3 trondhjemitic magma. This model of partial melting of basaltic rock is of special interest, because greenstones are very abundant in the Follstad district. These greenstones are metamorphosed and recrystallized to differing degrees and thus may have undergone partial remelting at deeper levels.

Green & Ringwood (1968) proposed a two-stage development for the calc-alkaline rock series, of which trondhjemites constitute a sub-trend. First, partial melting of mantle source rock below a spreading center would produce a high-Al_2O_3 tholeiitic magma, leaving garnet and clinopyroxene as residiuum. In the second stage this basalt crust is subducted under an island-arc type continental margin. At greater depths along the Benioff zone, the basalt partially melts, and if at shallow depths where hornblende would be residual it would produce a magma of trondhjemitic composition. Such a model produce not only the major element characteristics of trondhjemites, but also the trace element and rare earth element characteristics of the Follstad district trondhjemite (low abundance of REE, highly fractionated REE pattern, HREE abundances near chondritic, no Eu anomaly, slight Ce enrichment).

Critical to the origin of trondhjemitic magma is the influence of the mafic phases. In the trondhjemite system (Hb–Bi–Pl–Q), early removal of hornblende (or its being a residual phase) almost seems necessary to produce not only the K_2O and Al_2O_3 abundances in trondhjemites, but also the REE pattern characteristic of these rocks. The depletion of HREE is distinctive in these trondhjemitic rocks. Arth & Barker (1976, p. 535) have shown that removal of hornblende from a magma produces such a pattern in the REE. Removal of plagioclase enriches the melt in all REE except europium.

Fractionation of hornblende, plagioclase and some biotite from a basaltic melt yields about 24% by volume of trondhjemitic magma ($SiO_2 = 71\%$). However, a 20–30% partial melting of a low-K_2O tholeiitic basalt yields the same trondhjemitic magma with the added possibility of being able to produce greater proportions of trondhjemitic rocks over the more basic types (as observed in the Uusikaupunki–Kalanti complex of southwest Finland; Hietanen 1943). The temperature increase necessary to melt hornblende and plagioclase produces a gap between the melting points of tonalitic and granodioritic rocks (Wyllie 1977, Piwinskii 1967). This increase partially explains why trondhjemites are so deficient in mafic minerals. With excess water in the system most rocks begin melting just below 700°C (upper almandine–amphibolite facies range), and most minerals will melt within a narrow temperature range (about 100°C). However, a much wider temperature interval is necessary to melt mafic minerals to produce a resultant magma of tonalitic composition (Piwinskii 1968). Such magmas do not reach their liquidus until about 950°–1000°C (Wyllie 1977). This temperature is too high to be developed during normal regional metamorphism. This fact may explain why trondhjemites are more common in such tectonic settings than tonalites.

The composition of the initial melt during anatexis depends on many factors of which the source material is one of the more important. In the granite system an increase in water pressure can move the composition of the eutectic melt towards the Ab component, whereas an increase in the An content of the parent material can shift the minimum melt composition towards a greater Or component, counteracting the effect of change in pressure (Winkler 1965, p. 188). In Fig. 8 the eutectic melt composition is plotted at $P_{H_2O} = 2$ kb, with an Ab/An ratio of 3.8. The minimum melt temperature is 695°C. A temperature increase of approximately 70° could produce a magma of the composition shown in the Follstad district trondhjemite plotted in Fig. 8.

Models for the origin of trondhjemitic magmas are clearly polygenic, but the model that most closely fits the data for the Follstad district trondhjemite is one of equilibrium melting of a low-K_2O tholeiitic basalt during anatexis in an orogenic zone. Helz (1976) performed melting experiments on tholeiites in the low temperature melting range (680°–1000°C, $P_{H_2O} = 5$ kb) and determined that within the hornblende stability field, partial melts of all the starting basalts are strongly quartzo–feldspathic and corundum-normative (the Follstad trondhjemites are also corundum normative). The composition of the low temperature melts are quite consistent and insensitive to differences in starting materials, P_{H_2O} or f_{O_2}. One exception is that the relative abundances of calcium, sodium and potassium vary directly from the source rock to the magma.

The results of Helz' experiments can be directly related to the Follstad district trondhjemite. Støren Group greenstones are similar in composition to the basalts used by Helz (1976) as starting material, except that the K_2O is lower in the greenstones (K_2O = 0.16–0.22%) than in the basalts (K_2O = 0.49–0.97%). The solidus for the basalts is at about 690°C and the melt compositions are all similar up to the upper stability field of hornblende. The Or is relatively constant as long as hornblende is the only alkali-bearing phase. The melts move out towards the center of the Q–Or–Ab ternary only when plagioclase begins to separate from the melt (Stern 1974). The closest natural analog to the melts derived in the partial melting experiments of the tholeiitic basalts is the trondhjemite suite. At 825°–850°C and $P_{H_2O} = 5$ kb, the composition of the partial melts from the 1921 Kilauea tholeiite and the Picture Gorge tholeiite (Helz 1976) had almost the same bulk composition as Follstad district trondhjemite. The slight difference is that K_2O content is lower in these trondhjemites than in the basaltic melt. This difference could be accounted for by the lower K_2O content of Støren greenstones, if they were comparable to the parent material for the trondhjemite.

Summary

Trondhjemite of the Follstad district intrudes a sequence of associated Lower Paleozoic metasedimentary and metavolcanic rocks of the Caledonide orogen in central Norway. There were four periods of trondhjemite intrusion in this district, with the oldest intrusives the most deformed. Trondhjemite at the Follstad quarry near Støren is late-tectonic, and has an upper greenschist facies

mineral assemblage. Follstad district trondhjemite is homogeneous in mineral and chemical composition, and shows little evidence of any earlier or later differentiation. Its composition is at the low temperature minimum of the gabbro–trondhjemite trend, which is a sub-trend of the calc-alkaline series. However, distinctive compositional gaps in the trondhjemite compared to other associated magmatic rocks in the Trondheim area are difficult to explain by a simple fractional crystallization model.

Composition of the Follstad trondhjemite is characteristic of trondhjemites of other regions associated with mobile belts and greenstones (Shelbourne pluton, Nova Scotia, Albuquerque 1977; Twillingate pluton, eastern Newfoundland, Strong et al. 1974). Trondhjemites associated with basic volcanic rocks are peraluminous; high in SiO_2, Al_2O_3 and Na_2O, but low in K_2O; usually late-tectonic; extremely homogeneous in composition; depleted and show a highly fractionated REE pattern; and have low abundances of Rb compared with Ba.

The petrogenetic interpretation for the origin of the Follstad trondhjemite is based on the model of equilibrium melting of a K_2O-poor tholeiitic basalt during anatexis in an orogenic zone. The regional tectonic setting of the Follstad district and the character of the associated rocks support this model. Other models, such as anatexis of continental crust or partial melting of greywacke, would not produce the composition of Follstad trondhjemite.

The composition of the Follstad trondhjemite plots away from minimum melting temperatures in the Or–Ab–An–Qtz–H_2O system. Melting experiments of Piwinskii (1968) and Wyllie (1977) show that below 2 kb pressure, temperatures necessary to produce an anatectic melt of trondhjemitic composition are too high to be produced during normal regional metamorphism. However, in the range of 5–10 kb P_{H_2O} a trondhjemitic melt can be produced between 760°–840°C. This work agrees with the work of Helz (1976) on the partial melting of tholeiitic basalts, which produce a trondhjemitic melt at 825°–850°C at $P_{H_2O} = 5$ kb. Composition of the trondhjemite produced would show major element, trace element and REE abundances as seen in the Follstad trondhjemite if hornblende is a residual phase.

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Structures and Ore Genesis of the Grimsdalen Sulphide Deposits, Southern Trondheim Region, Norway

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Pedersen, S. A. S. 1979: Structures and ore genesis of the Grimsdalen sulphide deposits, southern Trondheim region, Norway. *Norges geol. Unders. 351*, 77–98.

The central part of the Grimsdalen area of the southern Trondheim region is occupied by a sequence of metavolcanics with subordinate metasediments, termed the Folla Group, which are correlated with the Støren Group of the central and western parts of the region. The Folla Group is thrust above arkosic sparagmites and is itself overthrust by metasediments of the Mesæterhø Group. A tectonic model for the Grimsdalen area, which has been affected by two major phases of deformation, is presented. The sulphide deposits of the area occur in a single pyritite horizon within the Folla Group, and 4 ore types have been differentiated. The metamorphic evolution of the pyritite and the different generations of pyrite are discussed in relation to the structural and metamorphic history of the Folla Group sequence.

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Introduction

The Grimsdalen area lies 10 km south-west of the Folldal ore field in the southern part of the Trondheim region. The thrust boundary between the deformed mica schists and greenschists of the Trondheim Nappe (Wolff 1967) and the sparagmites of the underlying Kvitvola Nappe (Heim 1971, Strand 1972) passes through the southern part of the area (Plate 1). The sulphide ore deposits in Grimdalen form a continuation of the known ore field around Folldal (Foslie 1926), which belongs to the large group of stratiform sulphide deposits of the Scandinavian Caledonides (Oftedahl 1958, Vokes 1968, 1976).

The area has been affected by two main deformations, both characterized by greenschist to amphibolite facies conditions. In this paper a model for the structural development is presented. It is shown that the concentration of the stratiform sulphides into ore bodies occurred during the first major deformation while the actual present-day location of the ore bodies is governed by the second phase of deformation. The structural model makes it possible to predict the subsurface position of the ore bodies.

Lithology and field relations of the major units

The rocks of the Grimsdalen area can be divided informally into three larger tectono–stratigraphic units: the sparagmites, the Folla Group and the Mesæterhø Group (Fig. 1).

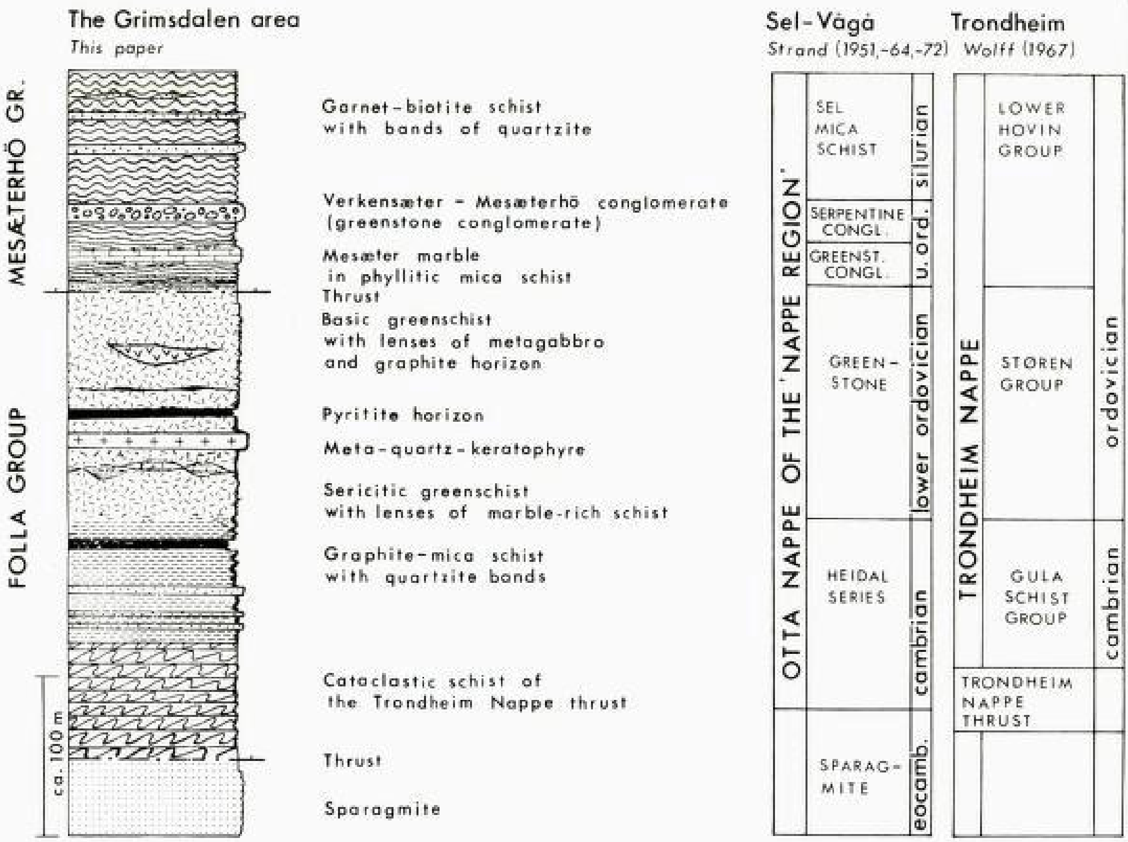


Fig. 1. Tectonostratigraphic column for the Grimsdalen area.

THE SPARAGMITES

The sparagmites of the Grimsdalen area lie along the northern border of the Rondane mountains. The rocks are grey to pale grey quartzites or meta-arkoses frequently containing epidote and specular hematite. Quartz is often segregated into quartz rods parallel to the main fold axes. The general dip is 20°–30° to the north-west and fold axes plunge 20° to the west.

THE THRUST ZONE BETWEEN THE SPARAGMITES AND THE FOLLA GROUP

The border between the sparagmites and the overlying Folla Group is part of the major thrust beneath the Trondheim Nappe (Wolff 1967). The thrust is marked by a sequence of mylonitic rocks which in accordance with Higgins (1971) may be termed mylonitic schists, mylo–sparagmite, magnetite–porphyroblast mylonite with banded intercalations of a black phyllonitic schist with ilmenite porphyroblasts, breccia mylonite and metagabbro mylonite. From this it is evident that the mylonite zone is a lithologically variegated zone.

The general dip of the thrust plane is to the NNW east of Sjøberget, and to the NW west of Sjøberget. Overturned antiforms of the Folla Group rocks have been upthrusted against the mylonite zone along thrust planes dipping ca. 50°NW.

THE FOLLA GROUP

The metavolcanics extending from Kakelldalen in the east to the watershed between Grimsdalen and Gudbrandsdalen to the west were named by Heim (1971) the Folla Group after the river Folla in Folldal. The group has been correlated with the Lower Ordovician Støren Group (Vogt 1945) of the central part of the Trondheim region (Heim 1971).

The Folla Group makes up the central part of the Grimsdalen area and has a thrust boundary with the overlying Mesæterhø Group and with the underlying sparagmites (Plate 1). In the Grimsdalen area the Folla Group can be divided into the following units (Fig. 1).

Basic greenschist (structurally highest)
(containing the pyritite horizon)
Meta–quartz–keratophyre
Sericitic greenschist
Graphite–mica schist (structurally lowest)

For reasons that are given later, the sequence is believed to be inverted.

Basic greenschist

This unit consists of alternating layers of amphibolite, fine-grained hornblende greenschist and hornblende–chlorite greenschist. A plane-parallel orientation of hornblende forms the schistosity in all the rocks. Plagioclase is generally an albite, but andesine has been found in a few samples of amphibolite and as porphyric grains in a metabasite from the hinge zone of a fold formed during the early fold phase (F_{D2}, see later). On fold flanks this metabasite has been transformed into an albite–chlorite greenschist with chlorite pseudomorphs after hornblende and zoisitization of the plagioclase grains. Within the lowest 10 m of the basic greenschist occurs a rock consisting of more than 50% pyrite, hereafter called pyritite (Schermerhorn 1970). The present thickness of the pyritite varies from 10 cm to 2 m, and in isoclinal early fold closures it forms ore bodies with an average thickness of 4 m.

The pyritite is seldom exposed in the field, but it is known from the abandoned Grimsdal mine and from a series of diamond drill cores drilled by Folldal Verk A/S in the years 1970–1976 (Fig. 2).

Metagabbro is found in the middle of the basic greenschist. The gabbro was intruded as sills or dykes and has locally been transformed into lensoid actinolite aggregates up to 5 metres in size. At Sjøberget a larger body of metagabbro occurs. To the southwest this body is strongly mylonitized where it abuts against the Trondheim Nappe thrust.

Graphitic schists occur as up to 30 cm-wide bands in the fine-grained horn-

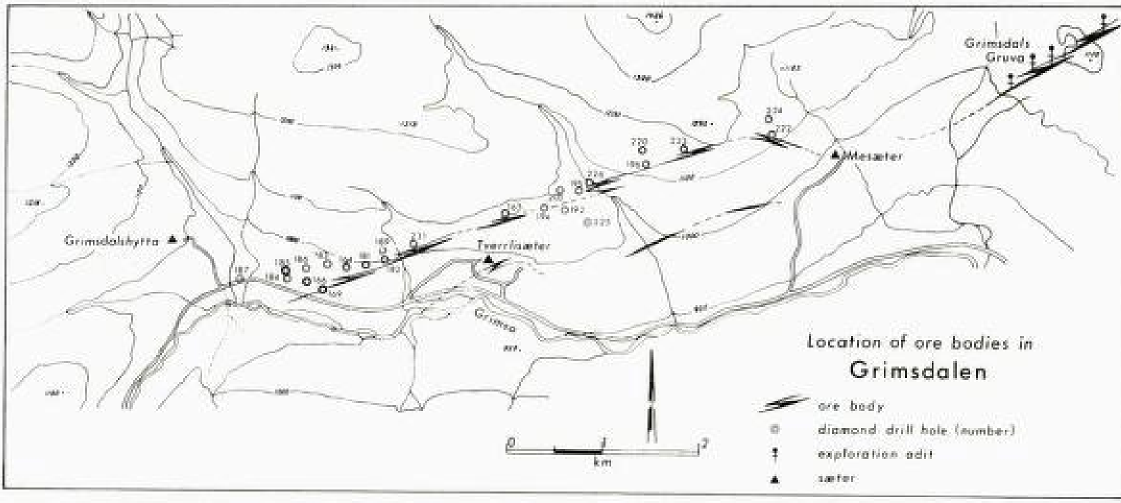


Fig. 2. Location of the ore bodies in Grimsdalen. Based on data from diamond drill cores (Folldal Verk A/S), geophysical investigation and geological mapping.

blende–chlorite greenschist. This alternation between metasediments and metavolcanics makes it probable that the whole sequence of amphibolites and greenschists represents a series of submarine metavolcanics.

Meta–quartz–keratophyre

This rock crops out as a light, resistant rock often forming terraces or ridges in the landscape. The rock is whitish, but can have a yellowish tint due to corrosion of small (0.5 mm) pyrite idioblasts.

Hornblende porphyroblasts form star-like aggregates, and at the contact to the basic greenschist these may develop into garbenschiefer. Garnet porphyroblasts are abundant, and along the contact to the sericitic greenschist the meta–quartz–keratophyre is developed as a snowball-garnet schist, where the rotated garnets can reach a size of 5 cm in diameter. The main minerals are quartz and albite forming a fine-grained granoblastic matrix which is a result of mylonitization. A few larger grains of oligoclase have been found; these contain a lot of inclusions and have corroded boundaries indicating a pyroclastic origin for the rock.

The submarine–sedimentary origin of the lithology is documented by the existence of 1–10 cm-wide graphite bands interbedded with meta–quartz–keratophyre at the transition to the sericitic greenschist.

Tectonically the meta–quartz–keratophyre behaves like a competent layer between the basic and the sericitic greenschists. It often forms characteristic buckle-type folds in hinge zones, while it is boudined on fold flanks. Boudinage combined with compression, shearing-out and metamorphic transformation can locally have transformed the meta–quartz–keratophyre to a foliated quartz–hornblende–plagioclase schist.

Sericitic greenschist

This unit consists of light green or grey-green, fine-grained greenschists characterized by their content of sericite and by the bright green mineral fuchsite. The greenschists are strongly tectonised and kink folds are common. Yellowish to pale brown carbonaceous mica schists are interbedded in the unit, and close to the contact to the meta–quartz–keratophyre a carbonate-banded grey mica schist appears locally. These rocks are interpreted as metasediments, representing sedimentary intercalations in a series of metavolcanics.

Graphite–mica schist

A graphite–mica schist which grades into a quartzite-banded mica schist has a concordant contact to the sericitic greenschist. The quartzite-banded mica schist is seldom exposed, but the graphite schist is often found intensively infolded in the sericitic greenschist.

Wolff (1967) and Heim (1971) correlate the Mesæterhø Group (this paper) with the Gula Group. As the graphite–mica schist in Grimsdalen tectonically belongs to the Folla Group, which is separated by a thrust zone from the Mesæterhø Group, it is proposed that the unit belongs to the Folla Group.

THE THRUST ZONE BETWEEN THE FOLLA AND MESÆTERHØ GROUPS

The mylonitic rocks of this zone can be related to either the overlying Mesæterhø Group or the underlying Folla Group. In the Folla Group the basic greenschists have been altered to cataclastic greenschist with hornblende cataclasts surrounded by chlorite forming a mylonite schistosity. Near the thrust the meta–quartz–keratophyre has been transformed into leucocratic breccia mylonite. A phyllonitic greenschist with a green silky sheen is regarded as being the mylonitic equivalent of the sericitic greenschist, but as the meta–quartz–keratophyre is boudined and sheared-out, it is impossible to differentiate the tectonically mixed greenschists.

The rocks of the upper contact are mostly fine-grained, grey mica schists (phyllonitic) or red-grey, strongly sheared, biotite–chlorite schists. North of Mesæter a silver-grey to dark grey mylonite schist appears with intercalations of Mesæter marble (see later).

The general orientation of the thrust is shown in stereogram 2 in Plate 1. The fold axes have been sheared out during the thrusting and lie on a great circle, which corresponds to the plane representing the orientation of the thrust.

THE MESÆTERHØ GROUP

The Mesæterhø Group makes up the northern part of the mapped area and has previously been described by Heim (1971) under the name Gula Group. As the lithostratigraphic correlations between the southern Trondheim region and the type area of the Gula Group have been much discussed (Bugge 1954, Strand 1951, 1960, 1972, Wolff 1967) the present author prefers to use a more local designation. The name Mesæterhø Group is strictly informal.

The main rock-type of the Mesæterhø Group is a grey-brown to rust-coloured garnet–biotite schist. Garnets frequently form up to 0.5 cm-sized porphyroblasts that have been rotated, flattened, and often altered to chlorite. Staurolite idioblasts are locally abundant, and kyanite has been found as sheaf-like aggregates up to 5 cm long. Due to retrograde metamorphism staurolite and kyanite may be altered to sericite pseudomorphs.

In the structurally highest part of the mica schists metre-thick intercalations of arkosic quartzite occur. In the structurally lowermost part of the group two important units appear: a meta-conglomerate and a marble horizon, both previously described by Bjørlykke (1905) from Verkensæter in Grimsdalen.

The meta-conglomerate varies in lithology laterally, but as the matrix generally consists of amphibole $\pm$ chlorite it may be termed a greenstone conglomerate (Strand 1951). The clasts are dominated by quartzo–feldspathic rocks which may contain hornblende needles in which case they resemble meta–quartz–keratophyre. Amphibolite–greenstone clasts are common and holes with a rim of hematite/goethite indicate that clasts with a high content of sulphides have been present. Thus, most of the rocks of the Folla Group are represented in the conglomerate.

As a result of deformation the pebbles are flattened and elongated, and in strongly deformed meta-conglomerates the quartzo–feldspathic pebbles form a leucocratic banding in a cataclastic hornblende schist.

The marble unit ('der Mesæter Marmor' of Heim, 1971) forms a 1–12 m-thick layer in a series of grey phyllitic mica schists in the lowest part of the Mesæterhø Group. It is a grey-white to yellow calcite marble that shows intensive folding with hinge thickening and shearing-out along fold limbs.

STRATIGRAPHIC POSITION OF THE FOLLA AND MESÆTERHØ GROUPS

The pyritite in Grimsdalen is a stratiform horizon in a series of greenschists which represent metavolcanics intercalated with thin layers of sedimentary origin. The pyritite is therefore interpreted as being of submarine exhalative origin. This interpretation has already been applied to the belt of sulphide deposits associated with greenstones that characterizes the Scandinavian Caledonides (Oftedahl 1958, Geis 1960, Page 1964, Anger 1966, Vokes 1968, 1976, Rui 1973, Juve 1974).

The sequence basic greenschist – pyrite – quartz-keratophyre – sericitic greenschist can be regarded as representing a magmatic evolution starting with an initial series of basic eruptives followed by a period during which fractionation of a hydrothermal and an acid phase took place. The hydrothermal phase migrated away along cracks and fissures, and the heavy metals were precipitated on the sea floor (exhalative phase). The magmatogenic origin of the pyritite is supported geochemically by the occurrence of the element Te which is not found in measurable concentrations in sedimentary rocks (Sindeeva 1964). After the exhalative phase the volcanism continued with an acid extrusive phase with eruption of quartz–keratophyre. The volcanic activity ended with the deposition of a series of spilitic volcanics represented by the sericitic greenschists. According to this interpretation the Folla Group is inverted.

The composition of the clasts in the greenstone conglomerate indicates that the conglomerate was formed by erosion of the Folla Group and older rocks. It is therefore assumed that the true stratigraphical position of the Mesæterhø Group is above the Folla Group.

On the basis of the presence of the greenstone conglomerate and the marble, the Mesæterhø Group may be correlated with the Sel Mica Schist (Strand 1951) in the Sel–Vågå area. The Folla Group can then be correlated with the greenstone series of the Sel–Vågå area.

Wolff (1967) and Heim (1971) correlate the Folla Group with the Støren Group (Vogt 1945) of the Trondheim Nappe and consider the Mesæterhø Group as representing the Gula Group of the central Trondheim Region. Heim (1971) argues that the Folla Group and the Gula Group in the southern part of the south-eastern Trondheim Region are part of the same inverted nappe unit. However, as the Folla Group and the Mesæterhø Group are separated by a major thrust the present author suggests that the Folla Group is inverted, while the overlying Mesæterhø Group is lying in its original stratigraphical position in the Grimsdalen area.

The pyritite and its ore minerals

The pyrite is rather massive and ore minerals (pyrite, pyrrhotite, magnetite, sphalerite, chalcopyrite and galena) constitute more than 70% by volume. The gangue minerals are mainly quartz and calcite while chlorite, plagioclase and hornblende are usually present but only as minor constituents.

In the few fields exposures, the pyritite forms rusty bands often covered by jarosite. In fresh drill cores and hand specimens the rock is brass-yellow due to the presence of the fine-grained pyrite. Pyrrhotite in the matrix gives the rock a buff grey hue, and sphalerite is visible as red-brown stains. Magnetite forms thin black bands or schlieren in a pyritite with green-yellow colour, due to chalcopyrite in the matrix.

On the basis of the mineral parageneses (Table 1) four ore types may be distinguished:

1) pyrrhotite–pyrite ore
2) magnetite–pyrrhotite–pyrite ore
3) magnetite–pyrite ore
4) sphalerite–pyrite ore.

Pyrite is the dominant ore mineral in all ore types and constitutes 50–70 vol% of the rock. Chalcopyrite is always present in amounts of generally about 2–3%. Pyrrhotite, magnetite and sphalerite in varying amounts form characteristic constituents of the different ore types.

TYPES OF ORE

The *pyrrhotite–pyrite ore* is characterised by its large amount of pyrrhotite —

Table 1. The composition (vol.%) of the ore types as determined by microscopic point-counting analysis of polished sections with a 25 points grid integration occular. The reliability of the point-counting analysis is calculated from the method given by Kalsbeek (1969). In most samples 1000 points were counted. Sample 3078 represents cataclastically deformed ore type 4. The samples, with the numbers given, are stored at the Institut for Almen Geologi, University of Copenhagen. In addition to the minerals listed in this table all the samples contain silicates and carbonates

Ore type	Sample number	Pyrite	Pyrrhotite	Chalcopyrite	Sphalerite	Galena	Nativ Gold	Molybdenite	Cubanite	Altaite	Hessite	Marcasite	Magnetite
pyrrhotite -pyrite ore ore type 1	2795	68 ± 2	22 ± 1	2.5 ± 0.4	0.4 ± 0.2	+	+	+					1.3
	2792	54 ± 5	35 ± 3	2.4 ± 0.5	0.9 ± 0.6		+	+		+			
	2790	52 ± 5	36 ± 3	2.7 ± 0.6	0.2 ± 0.1		+	+					+
magnetite -pyrrhotite -pyrite ore ore type 2	2612	60 ± 4	13 ± 2	2 ± 0.8	2 ± 0.8			+					8 ± 2
	2794	57 ± 4	22 ± 3	2 ± 0.8	0.5 ± 0.4	+	+	+					16 ± 2
	3979	57 ± 3	3 ± 1	2.5 ± 0.8	0.7 ± 0.6			+				0.2	7 ± 2
magnetite -pyrite ore ore type 3	3085	60 ± 4		3 ± 1	0.5 ± 0.3			+					18 ± 3
	3086	49 ± 4	0.4 ± 0.2	4 ± 1	0.8 ± 0.3	+		+			+		29 ± 3
	3094	58 ± 5	0.3 ± 0.2	3 ± 1	0.3 ± 0.2			+		+			12 ± 2
sphalerite -pyrite ore ore type 4	3081	73 ± 3		0.5 ± 0.2	12 ± 0.2	0.5 ± 0.2		+	+	+	+	0.2	
	3768	63 ± 4		0.4 ± 0.2	14 ± 2	3 ± 1		+	+	+	+		
	3769	63 ± 4	0.1 ± 0.1	3 ± 1	5 ± 1	0.3 ± 0.2		+	+	+	+		0.2 ± 0.1
	2717	63 ± 3	+	1 ± 0.2	9 ± 1	2 ± 0.5		+	+	0.05 ± 0.02	0.05 ± 0.02		
	3078	54 ± 3	+		21 ± 3	+		+	+		+	2.2 ± 0.8	

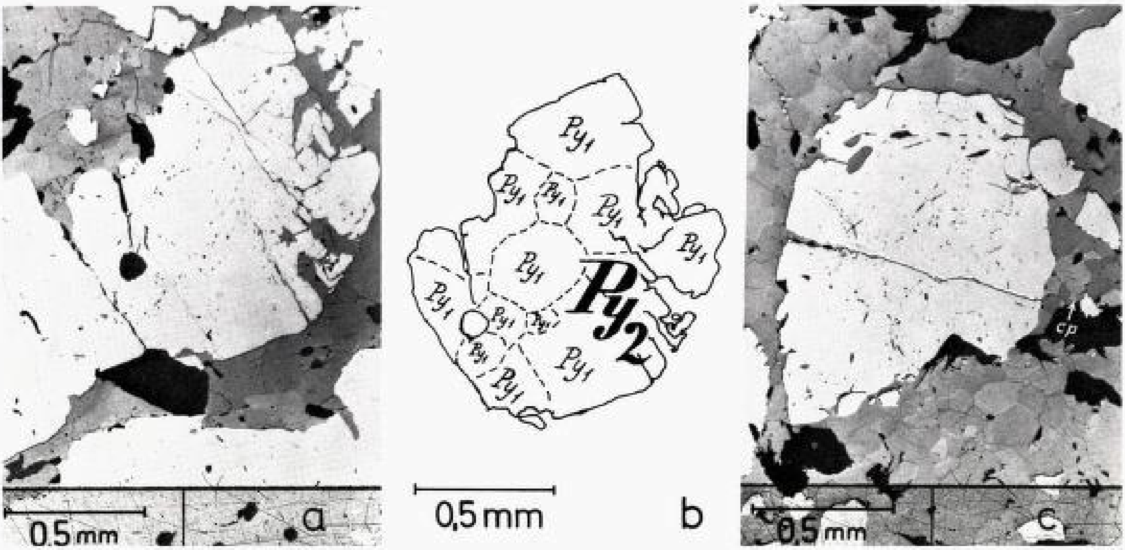


Fig. 3. Pyrrhotite-pyrite ore, ore type 1. White = pyrite, grey = pyrrhotite, black = silicates, Cp = chalcopyrite.
a) Large pyrite (py_2) grain containing traces of boundaries of py_1 grains.
b) Outline of the py_1 grains in the py_2 porphyroblast of Fig. 3a.
c) Py_2 porphyroblast in granoblastic pyrrhotite (po_2) matrix.
Optical conditions: reflected, plane polarised light, partly crossed Nicols and Nomarski interference-contrast.

some 30 vol%. A granoblastic pyrrhotite matrix contains pyrite porphyroblasts with an average size of 1–2 mm (Fig. 3), occasionally up to 1–2 cm. The pyrites form cubes which are often partly rounded and corroded by pyrrhotite (Fig. 3). Chalcopyrite and sphalerite are intergrown with the pyrrhotite matrix.

In the *magnetite–pyrrhotite–pyrite ore* a considerable amount of the pyrrhotite is substituted by magnetite. The pyrite and pyrrhotite closely resemble the pyrite and pyrrhotite of ore type 1, and the magnetite is found as 0.1 mm idioblasts that have grown preferentially in the pyrrhotite matrix (Fig. 4). The magnetite idioblasts are concentrated in small lenses (1 × 0.1 cm) and layers up to 0.5 cm thick.

In the *magnetite–pyrite ore* pyrrhotite is almost absent and apart from pyrite, magnetite is the dominant mineral. There is an enrichment of chalcopyrite (up to about 4%) which occurs as interstitial matrix between pyrite grains (Fig. 5). Ore type 3 has, in general, a cataclastic structure. The magnetite is concentrated in 1–5 mm-wide layers and lenses that cut through aggregates of pyrite. The pyrite forms grains, 0.5 mm in size, that are flattened and elongated to give the ore type a characteristic fabric. Magnetite occurs as rounded grains c. 0.1 mm in size.

The *sphalerite–pyrite ore* contains a relatively large amount of sphalerite (some 10%), whereas pyrrhotite and magnetite are absent. Pyrite aggregates (0.5 mm across) show triple junction patterns, with sphalerite as matrix between the aggregates (Fig. 6). A number of accessory ore minerals occur in this ore

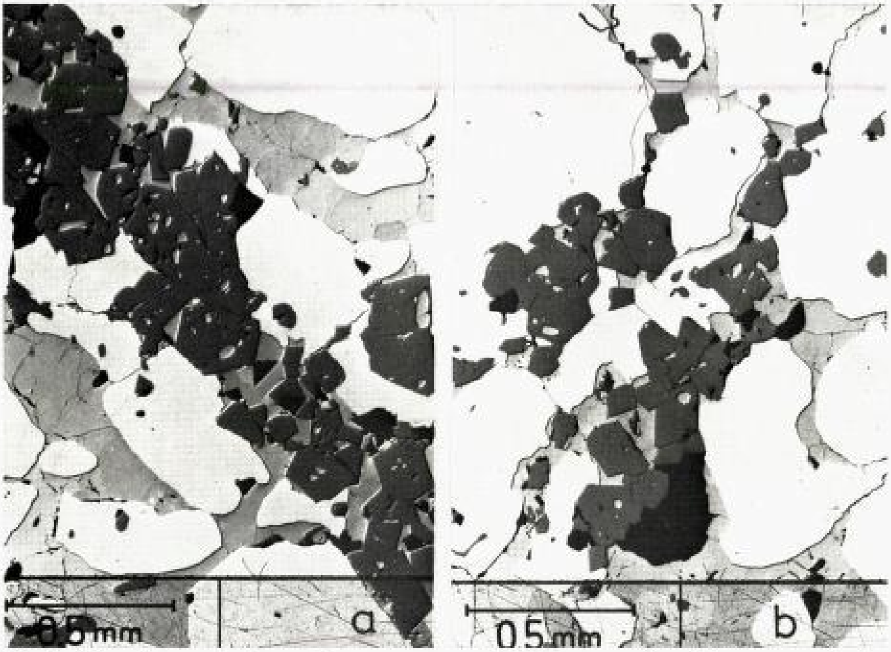
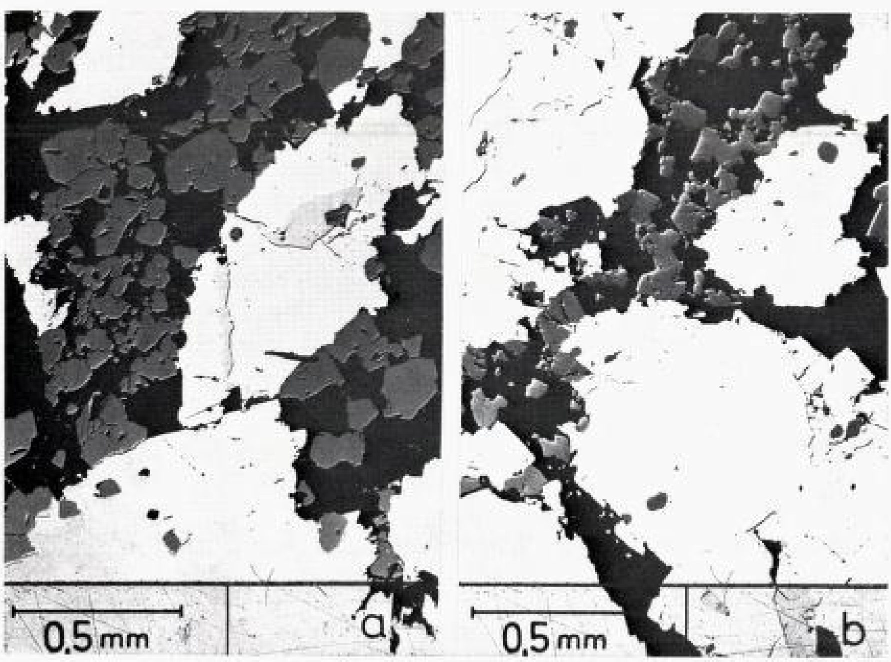


Fig. 4. Magnetite–pyrrhotite–pyrite ore, ore type 2. White = pyrite, light grey = chalcopyrite, dark grey = magnetite, black = silicates. Optical conditions: reflected, plane polarised light, partly crossed Nicols and Nomarski interference-contrast.

a) Magnetite idioblasts form a crude foliation in pyrrhotite–pyrite ore. Magnetite grew at at the expense of pyrrhotite, which occurs as inclusions in the magnetite idioblasts.

b) Magnetite idioblasts (mainly octahedra) form a diffuse foliation, from upper right to lower left, and occur predominantly in the pyrrhotite matrix.



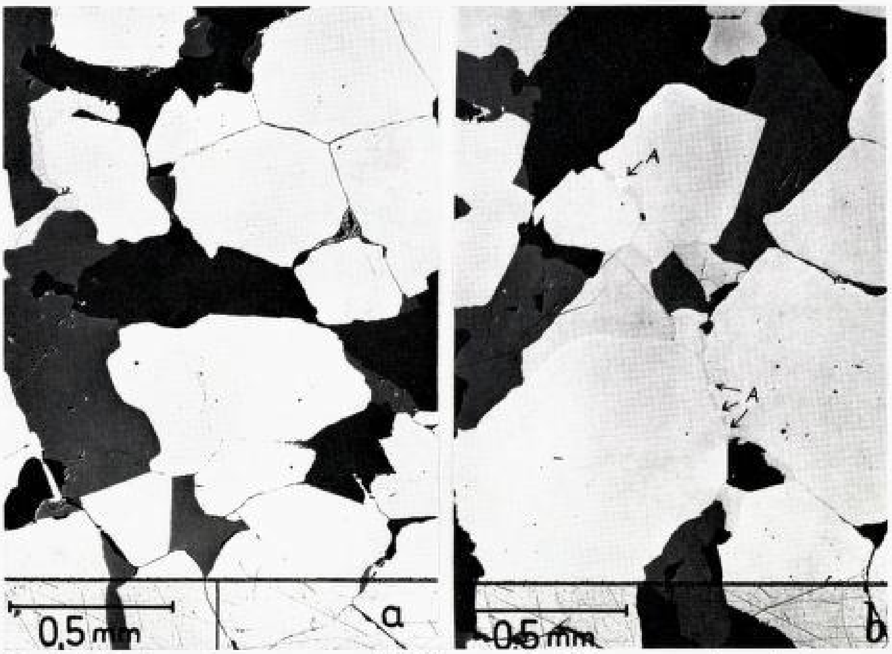


Fig. 6. Sphalerite–pyrite ore, ore type 4. White = pyrite, light grey = galena, dark grey = sphalerite, black = silicates and carbonates.

a) Typical triple junction pattern of py_3.

b) The location of altaite (A, arrowed) in galena-filled cracks between pyrite grains. Sphaerite shows weakly convex intergrown pattern with galena.

Optical conditions: reflected, plane polarised light, partly crossed Nicols and Nomarski interference-contrast.

type. Galena is the most important, besides tellurium minerals, molybdenite and cubanite. The ore type is characterized by alternation between 1–3 cm-wide bands rich in sphalerite (15–20%) and galena (2–3%) and less sphaleritic pyritite (sphalerite 5–10%, galena 1%, with chalcopyrite and pyrrhotite present, e.g. sample 3769, Table 1). Ore type 4 appears in parasitic folds on the limbs of larger isoclinal early folds (F_{D2}). Locally the sphalerite–pyrite ore is cataclastic and contains marcasite (sample 3079, Table 1; also see Fig. 8a).

ORE MINERALS

Pyrite occurs as aggregates intergrown with other sulphides or as individual grains among calcite and silicates. In aggregates with pyrrhotite and chalcopyrite, pyrite often forms cubes, but due to tectonic deformation these may be fractured and the cracks filled with chalcopyrite.

Fig. 5. Magnetite–pyrite, ore type 3. White = pyrite, light grey = chalcopyrite, dark grey = magnetite, black = silicates and carbonates. Optical conditions: reflected, plane polarised light, partly crossed nicols and Nomarski interference-contrast.

a) Elongation and flattening of pyrite due to cataclastic deformation. Magnetite grains are rounded and inclusion-free. The foliation in the rock is oriented diagonally in the picture.

b) Cataclastically deformed pyrite with interstitial chalcopyrite.

Fig. 7. Cataclastic sphalerite-pyrite ore.
White = pyrite, light grey = galena, dark grey = sphalerite, black = carbonates and silicates. Pyrite is cataclastically deformed and sphalerite has migrated into cracks and fissures.
Optical conditions: reflected, plane polarised light, partly crossed Nicols and Nomarski interference-contrast.

Four generations of pyrite formation have been recognized, termed py_1, py_2, py_3 and py_4. The outlines of py_1 are preserved by inclusions in py_2 grains. Py_2 occurs as porphyroblasts in ore types 1 and 2 (Figs. 3 and 9), and has partly been corroded by the pyrrhotite. Py_3 is found in ore type 4 as 0.5 mm grains showing 120° triple junction patterns (Smith 1964, Stanton 1972; Fig. 6), which together with the absence of inclusions indicates recrystallization under stable conditions. Py_4 occurs interstitially betwen cataclastic py_2 and py_3 grains (Fig. 8), and *marcasite* forms a concentric pattern in the central parts of py_4.

Droplets of native *gold* 0.005 mm in size and 0.01 mm-long *molybdenite* laths are found as exotic inclusions in py_2. Molybdenite is present as scattered grains up to 0.1 mm long in the ore types 3 and 4.

Small inclusions of *pyrrhotite* following the py_1 grain boundaries within the py_2 grains constitute the oldest pyrrhotite, termed po_1 . Po_2 forms the matrix between py_2 grains and shows granoblastic textures (Fig. 3). Single po_2 porphyroblasts show undulating extinction and the po_2 textures resemble blastomylonites (Hobbs et al. 1976) with py_2 cataclasts.

In the magnetite-pyrrhotite-pyrite ore, pyrrhotite occurs as inclusions in magnetite idioblasts. The inclusions form drops or laths which are often orientated in the (111) directions in the magnetite.

In ore type 3 *magnetite* is concentrated in bands and lenses together with the gangue minerals quartz and plagioclase (Fig. 5). The magnetite grains have

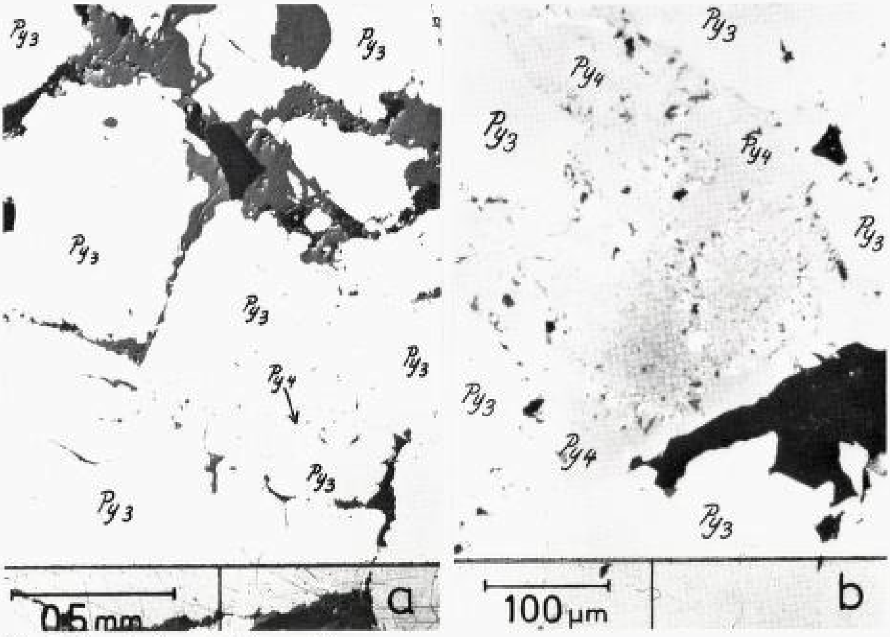


Fig. 8. Cataclastic sphalerite–pyrite ore.
a) Interstitial pyrite 4 (py_4) between pyrite 3 (py_3) grains. Dark grey = sphalerite, black = silicates and carbonates.
Optical conditions: reflected, plane polarised light, partly crossed Nicols and Nomarski interference-contrast.
b) Concentric pattern of marcasite in pyrite 4. Py_3 = pyrite 3, Py_4 = pyrite 4, dark and white angular minerals = marcasite, black = silicates and carbonates.
Optical conditions: reflected, plane polarised light, crossed Nicols and oil immersion.

a diameter of 0.01–0.1 mm and may be more or less rounded due to cataclastic deformation. In this ore type the magnetite is always clean and free of inclusions in contrast to the pyrrhotite-inclusion-rich idioblastic magnetite of ore type 2.

The content of *sphalerite* varies very much in the pyritite. In ore types 1 and 2 sphalerite occurs as an accessory mineral in the pyrrhotite matrix, and in ore type 3 it is associated with the chalcopyrite matrix. Sphalerite has been found as inclusions in py_2 where it forms inclusions together with pyrrhotite (po_1).

Apart from pyrite, sphalerite is the main mineral in the sphalerite-pyrite ore. Sphalerite behaves as a typical interstitial matrix mineral, filling spaces between py_3 grains and calcite. In the cataclastic sphalerite-pyrite ore sphalerite shows signs of great mobility, having migrated into cracks and fissures in the deformed py_3 grains (Fig. 7).

Galena is always present together with sphalerite in ore type 4, but is rarely observed in the other ore types. Sphalerite is frequently seen convexly grown into galena, a relationship which according to Stanton (1972) and Juve (1974) is due to a 'stronger' recrystallization strength of sphalerite.

The *chalcopyrite* content is greatest in the magnetite-pyrite ore; here, chalcopyrite generally occurs as the only mineral besides pyrite and magnetite. This indicates a positive correlation between chalcopyrite and magnetite (which means that when the pyritite is magnetite-banded one will always find a chalcopyrite-rich sulphide ore). In ore types 1 and 2 chalcopyrite occurs as a matrix mineral intergrown with pyrrhotite.

In the sphalerite-pyrite ore, chalcopyrite appears together with galena in the matrix. Characteristic inclusions in py_3 contain chalcopyrite + cubanite ± pyrrhotite. These inclusions reflect an exsolved intermediate solid solution (Kullerud 1969). Mackinawite/valleriite and enargite have been observed in chalcopyrite inclusions in pyrite (mainly py_3).

Tellurides occur as inclusions within, or along the boundaries of the galena grains in ore type 4. *Hessite* (Ag_2Te) and *altaite* (PbTe) form 0.01 mm grains in corners and fissures between galena and adjacent sulphides (sphalerite, pyrite and chalcopyrite). Hessite is found as 0.01 mm droplets within galena.

Tellurobismuth (Bi_2Te_3) forms small wedge-shaped lammellae in altaite or lath-shaped individual grains. The identification of altaite, hessite and tellurobismuth was based on reflectivity measurements and supported by semiquantitative microprobe scanning element determinations.

Ilmenite is common (up to 5%) in the basic greenschist, but is only present in the pyritite where the greenschist has been infolded in the pyritite. The ilmenite grains are 0.1–0.3 mm in size and often lath-shaped. In a few cases *rutile* inclusions have been found in the ilmenite. In addition, rutile occurs as single grains together with magnetite in ore type 3.

Metamorphic evolution of the pyritite

Primary structures, such as gel- or framboidal structures, that indicate formation in seawater at ordinary temperatures have not been observed in the investigated samples. It is suggested that the oldest sulphide grains $py_1 + po_1$ (Fig. 9, stage I) were generated during diagenesis or early metamorphism. The reaction $py_1 + po_1 \rightarrow py_2$ (Fig. 9, stage II) requires increasing S_2 pressure (Holland 1965, Barton 1970). Less metamorphosed stratiform exhalative sulphide deposits (Meggen, Rammelsberg and Kuroko) often contain gypsum, anhydrite or baryte (Anger et al 1966, Tatsumi 1970, Sato 1972). Therefore, it is supposed that the required increase in S_2 fugacity resulted from the breakdown of syn-depositional $CaSO_4$ which becomes unstable at higher temperatures under contemporaneous formation of other Ca-minerals, e.g. $CaCO_3$. Calcite is always present in the pyritite.

The evolution from stage II to stage III (Fig. 9) may be explained by an increase in temperature (Barton 1970), and the first three stages thus represent increasing metamorphism up to amphibolite facies conditions.

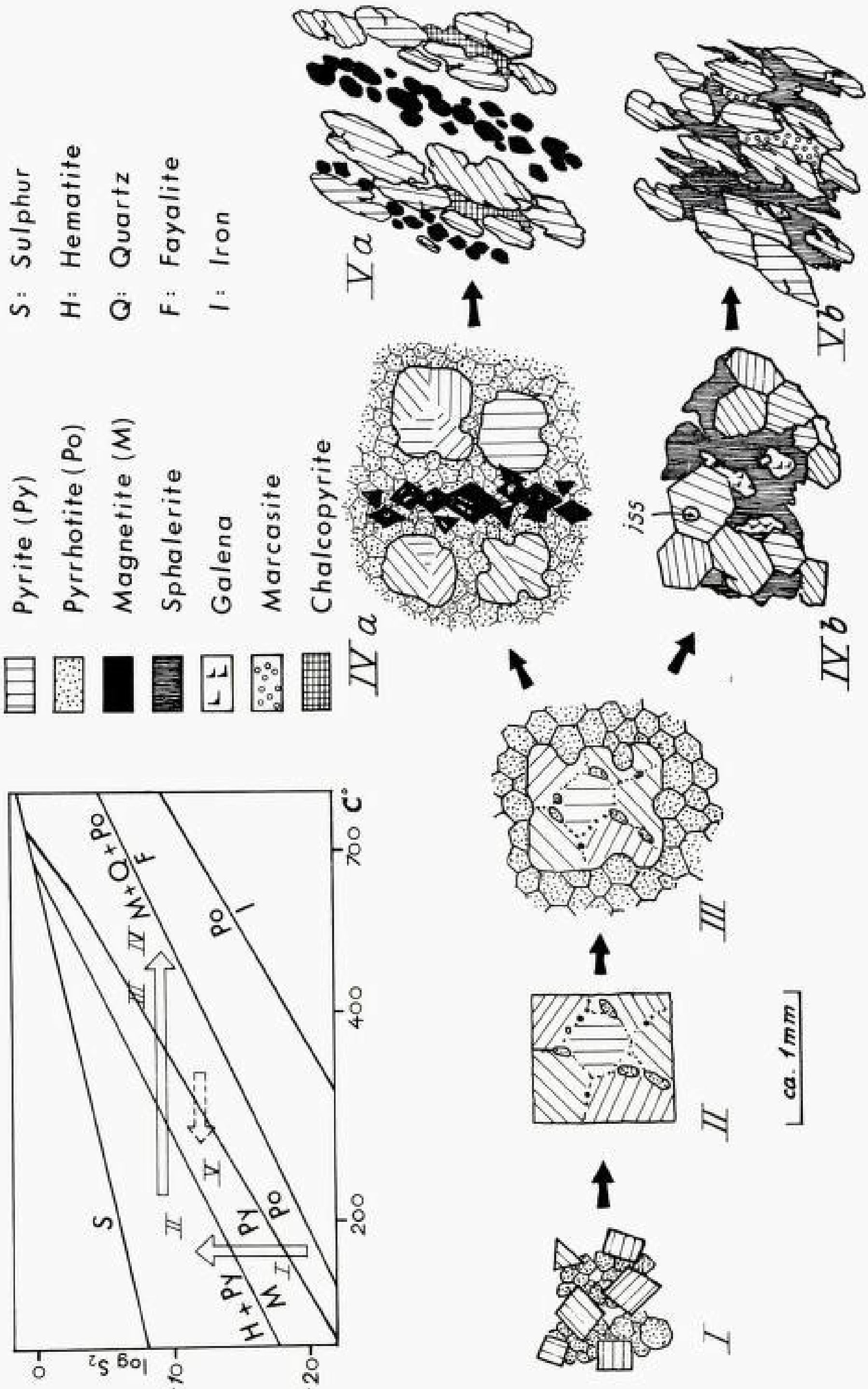


Fig. 9. Development of the ore types. The diagram is modified after Barton (1970) and shows the development of the ore types in stages in relation to temperature and sulphur fugacity (log S_2). Stage III = ore type 1, stage IVa = ore type 2, stage IVb = ore type 4, stage Va = ore type 3 and stage Vb = cataclastic ore type 4. In stage IV 'iss' refers to intermediate solid solution inclusion (chalcopyrite + cubanite ± pyrrhotite).

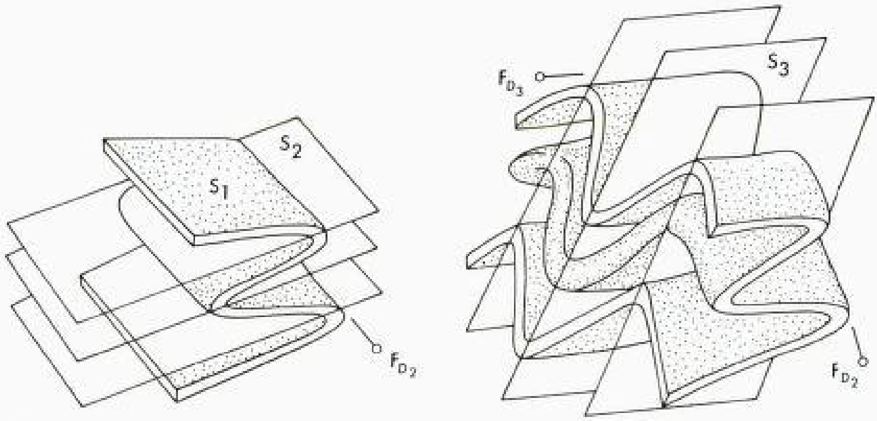


Fig. 10. Diagram to show the relationship between the structural surfaces developed during the two main deformations.

The magnetite blastesis (stage IVa, Fig. 9), and the formation of ore types 2 and 3 (stage IVa and Va), is interpreted as the result of an increase in O_2 fugacity. As pyrite was not altered during the magnetite blastesis, the S_2 fugasity must have been stable (Holland 1965, Figs. 26, 27), but the O_2 fugacity was increasing. O_2 was introduced along cleavage planes in the pyritite formed during the early fold phase, and in agreement with this magnetite layers and lenses define a foliation in the pyrite (Fig. 9, stage IVa, Fig. 4).

The sphalerite–pyrite ore (ore type 4) is concentrated in parasitic folds. The ore was recrystallized under stable conditions as shown by the triple junction pattern of py_3 (Smith 1964, Stanton 1972). Py_3 contains inclusions of chalcopyrite + cubanite ± pyrrhotite corresponding to intermediate solid solution inclusions. The estimated composition of the intermediate solid solution shows that the inclusions formed at temperatures above 400°C (Craig & Scott 1974). This supports the idea that the concentration of the sphalerite–pyrite ore took place under amphibolite facies conditions.

A cataclastic deformation affected the pyritite whereby ore type 2 was transformed into ore type 3 (stage IVa to Va, Fig. 9). The disappearance of pyrrhotite is ascribed to decreasing temperature (Barton 1970) and the introduction of O_2 along developing cleavage planes. Chalcopyrite is separated from the disappearing po_2 matrix and remains as the only matrix mineral between the pyrite grains.

In ore type 4 cataclastic deformation (stage Vb, Fig. 9) resulted in elongation and jointing of py_3 grains, and the cracks in py_3 were filled with sphalerite. Py_4 was formed as a metastable phase which under retrograde metamorphic conditions altered to marcasite. (A supergene origin for the py_4 and marcasite can hardly be supported as the investigated samples with py_4 and marcasite were all taken from a fresh drill core below 30 m depth).

Structural and metamorphic development

The general strike in Grimsdalen is ENE and dips between 50°–70° to NNW prevail. Three main deformations have affected the area. The first deformation,

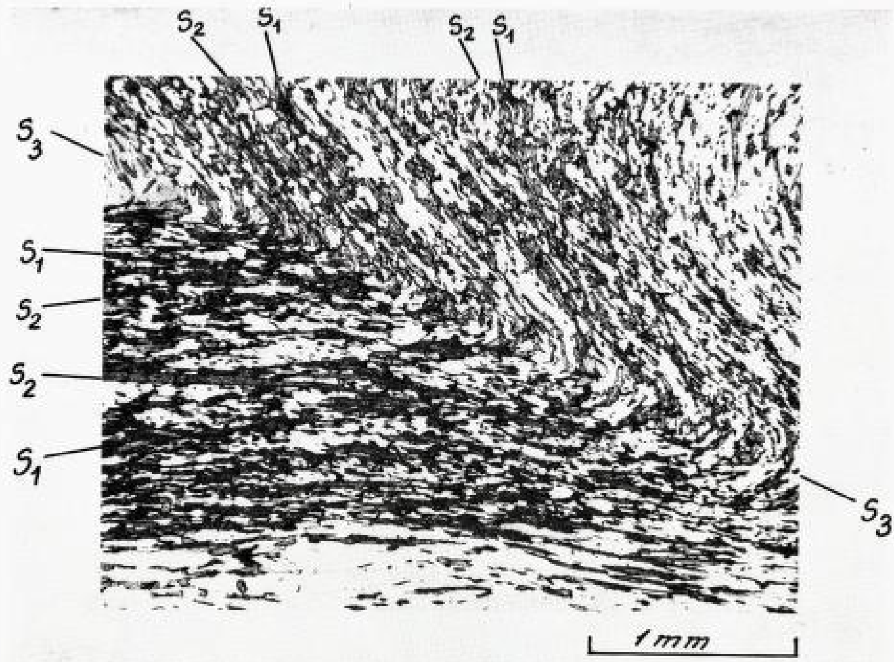


Fig. 11. Basic greenschist. The relationship between the schistosities S_1 and S_2 and the incipient cleavage S_3. S_1 and S_2 are defined by hornblende. S_3 is a poorly developed crenulation cleavage corresponding to the axial plane of the F_{D3} deformation. Neocrystallization of chlorite is just perceptible along S_3. Thin-section, plane polarised light.

D_1, is represented by the earliest recognizable schistosity, s_1. During the D_2 deformation s_1 was folded, and the axial plane schistosity s_2 developed. Finally the s_1 and s_2 schistosities were folded together by the structures belonging to the D_3 deformation, whereby the s_3 axial plane cleavage was formed (Fig. 10); this deformation is responsible for the steep dip of the rock layering to the north–northwest.

THE EARLY FOLD PHASE, F_{D2}

The earliest recognizable *folds* are isoclinal, recumbent folds with axes trending between N and NW with an overlap of up to several kilometres. The axes were more or less horizontal prior to the later folding. During the D_3 deformation the early axes were folded so that they now plunge in accordance with their position on the limbs of F_{D3} folds, generally to between N and NW (Fig. 10). A large nappe fold closure has been shown to be situated in Kakelldalen c. 10 km northeast of Folldal.

F_{D2} mesostructures can be seen in most larger exposures. Interference patterns such as arrow-head and lunar structures occur where the F_{D2} folds appear in favourable F_{D3} crests. This is especially the case where meta–quartz–keratophyre is infolded in the basic greenschist. In the sericitic greenschist F_{D2} crenulation folds have often been preserved on the limbs of F_{D3} crenulation folds, but have been wiped out on the crests.

F_{D2} microfolds are often seen in thin-section. In the mica schist the folds are defined by the bending of a biotite s_1 schistosity and in the basic greenschist of a hornblende s_1 schistosity (Fig. 11). The development of the s_2 schistosity in the garnet–biotite schist is accompanied by the formation of staurolite, kyanite and garnet. In the basic greenschist almandine occurs as idioblasts while hornblende defines the s_2 schistosity which is characterized by fewer but larger hornblende laths than s_1 (Fig. 11). As a result of the F_{D3} folding the s_1 and s_2 schistosities have been pressed together so as to be nearly parallel. Hornblende prisms also outline a lineation parallel to the F_{D2} axis. Andesine (An_{30}) has been found in a single amphibolite unit and as megacrysts in an F_{D2} fold closure. According to Turner & Verhoogen (1960) and Winkler (1967) these minerals indicate almandine–amphibolite facies metamorphism, and hence it is concluded that the early F_{D2} folding took place under similar metamorphic conditions.

THE LATE FOLD PHASE, F_{D3}

The late fold phase produced the dominant structures in the Grimsdalen area. The folds are tight to close folds which are overturned to the south with axial planes dipping c. 60°N. The general plunge of the fold axes is 10°–20° towards WSW.

The most prominent F_{D3} macrostructure is a synform with a core of basic greenschist which can be followed from Grimsdals Gruva to Mesæter where it has been displaced by a fault with both vertical and dextral displacement (Plate 1). Along the south-west side of the fault the axis of the synform shows a drag effect which indicates the direction of movement on the fault plane (subarea 10, Plate 1). South-west of the fault the synform continues to the area north of Tverrlisæter and on to the south-west below Grimsdalshytta to another fault. On the other side of this fault the structure can be followed to Verkensæter. The faulting is regarded as the latest deformation phase in the area, D_4.

The style of the fold structures seen in exposures depends very much on the lithology. The mica schist and the basic greenschist form angular folds, while the meta–quartz–keratophyre is typically buckle-folded. The sericitic greenschist is often kink-folded.

The F_{D3} microfolds are very obvious. The bending of s_1 and s_2 is locally accompanied by the development of an axial plane cleavage s_3 (Figs. 10, 11). In the greenschist chlorite forms the s_3 schistosity and it is inferred that greenschist facies conditions prevailed during the D_3 deformation. Due to this greenschist facies metamorphism sericite forms pseudomorphs after staurolite and kyanite, while garnet, hornblende and biotite are often altered to chlorite; albite is the generally occurring plagioclase. Accordingly, the deformations can be listed as in Table 2.

A tectonic model (Fig. 12) has been established for the area around Tverrlisæter (Plate 1). Field work has shown that the model is valid throughout Grimsdalen and that with few modifications it may be applied in the Folldal–Kakelldalen area. To clarify and simplify the model all units are shown as

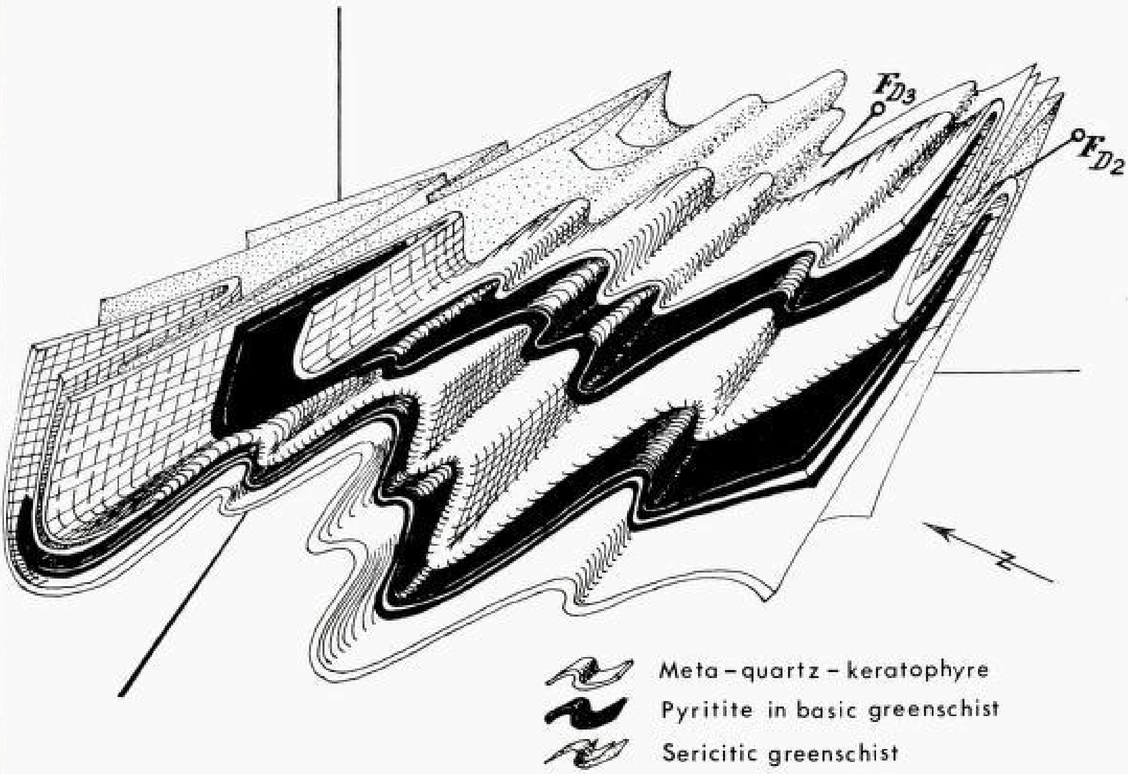


Fig. 12. Tectonic model of the fold structure at Tverrlisæter, Grimsdalen.

having constant thicknesses. In actual fact there are numerous examples of thinning of lithological units on fold limbs and increased thickness in fold hinges. These variations in thickness are most strongly developed within the pyritite.

able 2. The structural and mineralogical characteristics of the different phases of deformation

eformation phase	Folding	Schistosity, lineation	Metamorphism	Other deformations
D_1	None observed	s_1: hornblende, biotite	upper greenschist facies?	
D_2	F_{D2}: recumbent, isoclinal	s_2: hornblende, muscovite	amphibolite facies	thrusting along F_{D2} fold limbs, boudinage
D_3	F_{D3}: overturned, tight to close	s_3: chlorite, crenulation cleavage	greenschist facies	overthrusting to the SE
D_4	drag folding of F_{D3} axes along faults			faulting

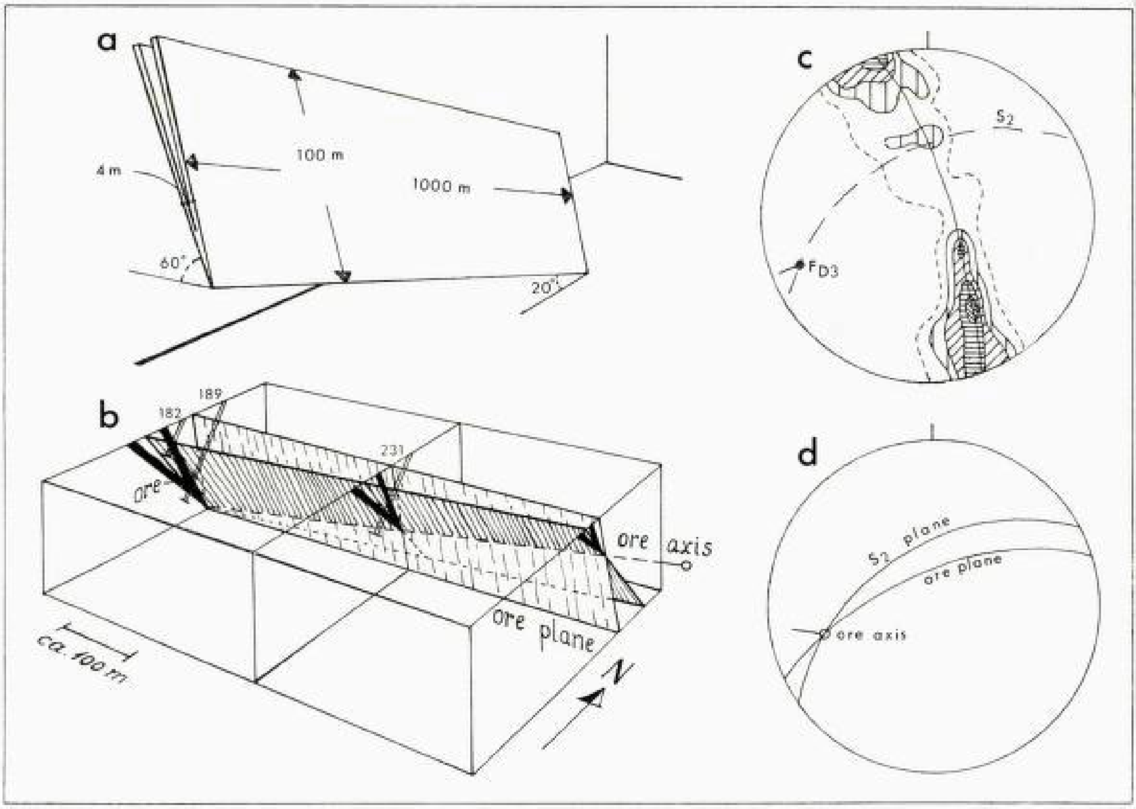


Fig. 13. Construction of the position of an ore body in the area around Tverrlisæter.
a) Idealized shape and orientation of the ore body.
b) Spatial relationship between the ore plane (dashed lines) and the general schistosity (lined). The construction of the ore axis is shown. The numbers refer to diamond drill cores (see Fig. 2).
c) The poles to the schistosity planes from the Tverrlisæter area are plotted stereographically and a general structural plane is constructed. Lower hemisphere, Wulff net. The normal to the great circle through the schistosity poles is the latest fold axis F_{D3}.
d) The intersection between the ore plane, taken from stereogram 8, Plate 1, and the S_2 plane from Fig. 13c gives the ore axis, i.e. the edge of the wedge-shaped ore body (Figs. 13a and 13b).

Tectonic control of the ore

Investigations at the Grimsdalen mine show that the ore is concentrated in the closures of isoclinal F_{D2} folds (Figs. 2 and 12). The fold cores are dominated by ore type 1 while ore type 4 is concentrated in parasitic folds on the limbs of the big F_{D2} folds. Where the pyritite has been strongly affected by the D_3 deformation the cataclastic ore types of stage V (Fig. 9) are developed.

Field evidence thus etablishes that the pyritite is concentrated in F_{D2} fold hinges. Location of ore bodies is then dependent on the distribution of F_{D2} folds within the F_{D3} pattern. The F_{D2} axes which are bent around the F_{D3} axes lie approximately on a great circle on a stereographic net as shown by diagram 8 in Plate 1. This great circle defines a plane that, for ores concentrated in early folds in refolded areas, is designated the ore plane (Stauffer 1968) (Fig. 13).

The ideal shape of an ore body will be a narrow wedge with the edge parallel to the F_{D2} axis (Fig. 13). The direction of the edge is given by the intersection of the ore plane and the dominating planar structure of the area.

Geophysical investigations carried out by Folldal Verk A/S show a complex pattern, but the anomalies caused by the pyritite can be related to the F_{D2} fold axes and to the tectonic model (Fig. 12). On the basis of the geophysical investigations diamond drilling was carried out which intersected an ore body west of Tverrlisæter. The continuation of this ore body was predicted from the tectonic model (Fig. 12), and a diamond drill core later proved the existence of the pyritite at the estimated depth.

Conclusions

The sulphide deposit in Grimsdalen is a stratiform mineralization belonging to a single pyritite horizon and related to the big group of Cu and Zn sulphide deposits of the Scandinavian Caledonides (Vokes 1976). The ore is regarded as being of exhalative–sedimentary origin and occurs in a series of metavolcanics, the Folla Group of Heim (1971), which may be correlated with the early Ordovician Støren Group. The ore-forming pyritite horizon is found in the following structural succession: (top) grenschist, pyritite, meta–quartz–keratophyre sericitic greenschist and graphite–mica schist (bottom). This sequence of the Folla Group is considered to represent an inverted stratigraphical succession, while the Mesæterhø Group lies in its original stratigraphical position above the Folla Group, though now with thrust contact.

The pyritite has been subjected to two major deformation phases. The earliest recognizable fold phase (F_{D2}, Table 2) formed recumbent, isoclinal folds under amphibolite facies conditions. During this deformation the ore types 1, 2 and 4 were formed. The pyritite was concentrated in wedge-shaped bodies in the closures of isoclinal folds.

The later fold phase (F_{D3}, Table 2) characterized by greenschist facies conditions created overturned tight to close folds. During cataclastic deformation ore type 3 and a cataclastic ore type 4 (Stage V, Fig. 9) developed. The position of the ore bodies is determined by the F_{D3} folding.

A structural model for the Grimsdalen area has been presented. On the basis of this model the position of the ore bodies may be predicted.

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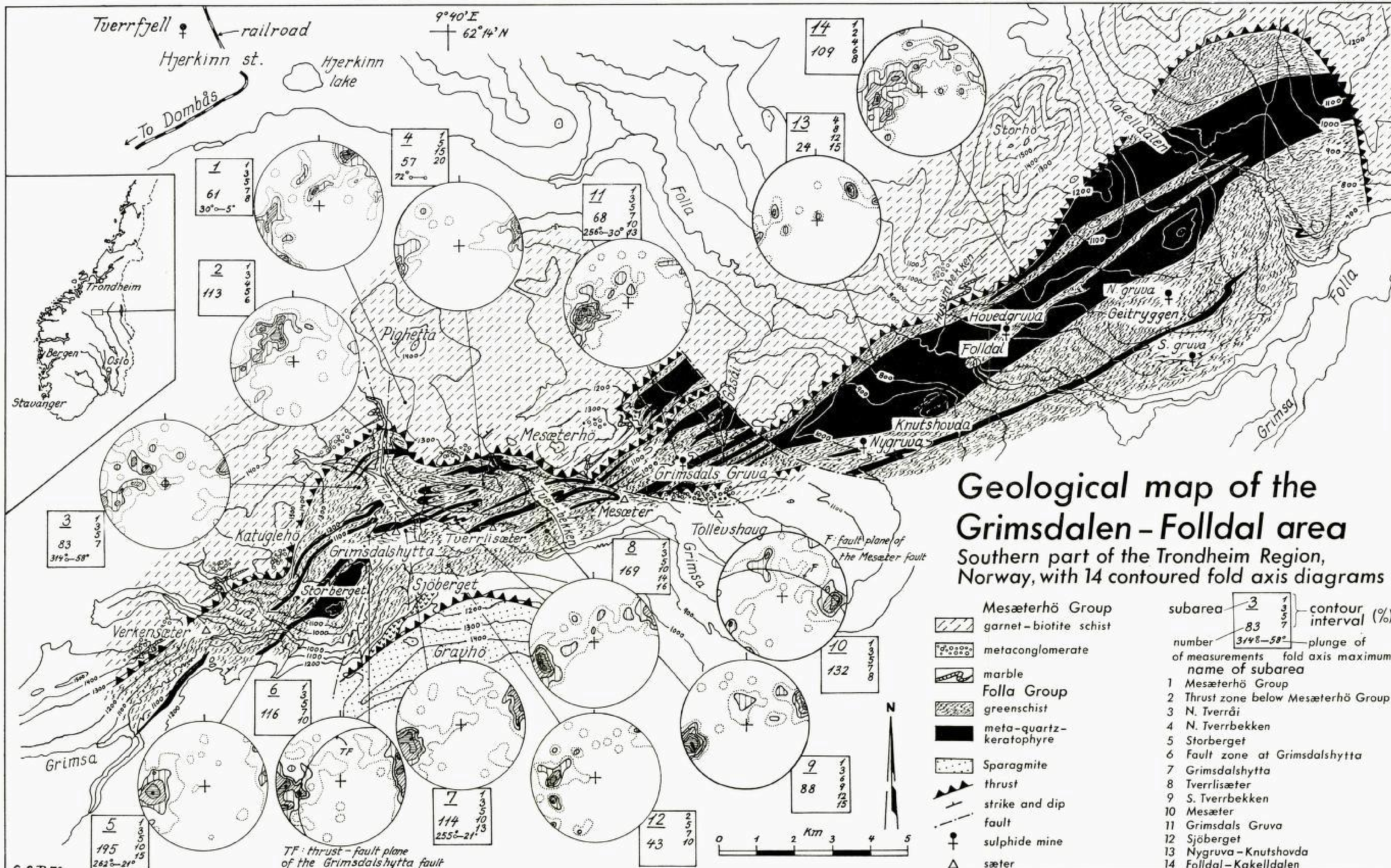

Plate 1. Geological map of the Grimsdalen – Folldal area. The stereograms show the axial orientations of F_{D3} folds (largest concentration) and a more diffuse great circle pattern due to the bending of the F_{D2} axes. The ornamentation representing the Folla Group greenschist depitcs the style and vergence of the F_{D3} folds (looking down-plunge).

The Kongsberg Series Margin and Its Major Bend in the Flesberg Area, Numedal, Buskerud

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Starmer, I. C. 1979: The Kongsberg Series margin and its major bend in the Flesberg area, Numedal, Buskerud. *Norges geol. Unders. 351*, 99–120.

This major boundary is shown to be defined not by the traditional 'Friction Breccia', but by a much earlier and more extensive mylonite zone. The mylonitisation was produced by a complex westward upthrust of the Kongsberg Series over the Telemark Series; both blocks had a common history thereafter. Spatial and chronological relationships between the upthrusting, an extensive series of gabbroic intrusions and an increased regional heat flow, suggest association with major crust – mantle processes. The only major bend in the Kongsberg Series margin occurs in this area and its location and development are explained.

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Introduction and general geology

The area lies some 20 km north of Kongsberg, around the villages of Flesberg, Lampeland and Lyngdal (Fig. 1) and is part of that described by A. Bugge (1937) in the memoir 'Flesberg og Eiker'. It is also included on the 1 : 250,000 map sheet 'Skien' (Dons & Jorde 1978). Bugge (1937, p. 111) considered that 'a great friction breccia 100–300 m wide, locally even wider, marks the boundary between the Telemark Formation and the Kongsberg Formation'. This 'rivningsbreccie' (friction-breccia) was described as a typical brittle-fracture, crushing the adjacent 'Kongsberg-granitt' which formed a continuous marginal strip to the Kongsberg Formation.

The present survey shows that the 'friction breccia' was a late, brittle fracturing, which was not of regional significance. The Kongsberg Series (to the east) and the Telemark Series (to the west) were separated by a much earlier, more major zone of mylonite (exceeding 1 km width) which was subsequently deformed and metamorphosed, with ubiquitous microcline porphyroblastesis. More intensive granitisation produced sporadic augen gneisses, probably equivalent to the 'Kongsberg-granitt (Øiegranitt)' of Bugge (1937). These augen gneisses, however, do not form a continuous marginal strip on the Kongsberg Series and were initially part of the mylonite zone. The term 'Kongsberg granite' has not been retained, since it has also been variously used, in areas to the south, for different bodies both east and west of Kongsberg. Moreover, in the present area, these granitic augen gneisses cannot be considered unequivocally as part of the Kongsberg Series.

Within the present area, some 10 km north of Flesberg (Fig. 1), the Kongsberg Series margin undergoes its only major bend from a N–S trend in the southern part to a NE–SW orientation further north. This bend (see key map

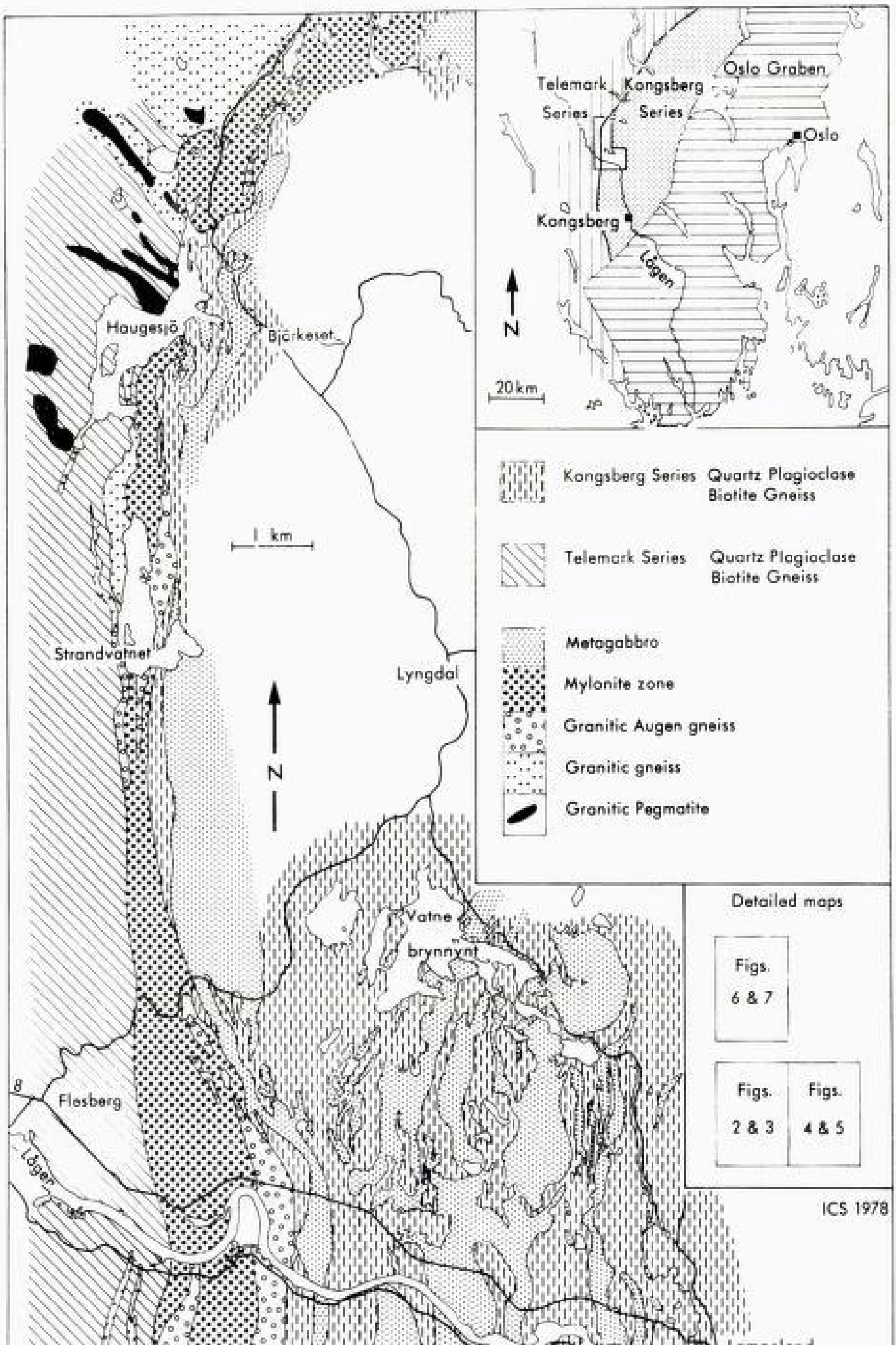

Fig. 1. Location and general features of the area.

of Fig. 1) is a gradual deflection of the boundary on larger scale maps (Figs. 1 & 6) and is produced by a tectonic break in the mylonite zone, which is deformed by a complex basin structure. Northeast of the present area, the boundary (of heavily granitised mylonite) continues for about 9 km with an essentially planar outcrop pattern, despite considerable internal deformation by minor structures (Starmer, in prep.).

The Kongsberg Series consists of supracrustal gneisses intruded by a series of basic bodies, the so-called 'Vinor' intrusions. In the Flesberg–Lampeland district, numerous gabbroic outcrops have the surrounding gneisses deformed around them and represent protrusions from the irregular roof of a larger mass. This is probably the northern end of the large 'Vinoren' body of C. Bugge (1917) and A. Bugge (1937). Metamorphism altered the gabbros and, enhanced by an increased heat flow around the large Vinoren body, caused some mobilisation of the supracrustal gneisses.

The mylonite zone cuts across this complex with a planar outcrop, although it has been deformed and even broken in the north, around Haugesjø. Late, brittle (friction-breccia) fracturing and associated hydrothermal veining were particularly concentrated around this mylonite zone.

The Telemark Series consists of cataclased supracrustal gneisses, granitic gneisses, amphibolitised intrusives and granitic pegmatites. These rocks have a much shallower dip than those of the Kongsberg Series, although lower angle dips also occur in the mylonite zone.

Thick drift occurs in the Lågen valley and has been shown on the maps Figs. 2 & 4) where extensive lack of exposure prevents accurate interpretation of bedrock. (The most probable interpretation is shown on a small-scale map of the general geology — Fig. 1).

Cataclastic rocks are of major importance and are decribed using the lithological nomenclature of Higgins (1971). In terms of the rationalised nomenclature of Zeck (1974) they show limited myloblastic recrystallisation (simultaneous with ruptural deformation) with dominant blastomylonitic recrystallisation (after mechanical degradation).

In the Kongsberg district (some 20 km south of the present area) reconnaissance radiometric work by O'Nions & Heier (1972) suggested that the rocks had undergone an early metamorphism around 1700 ± 100 m.y. and a second metamorphism around 1260 ± 40 m.y. ($\lambda = 1.39 \times 10^{-11} . yr^{-1}$). More recently, Rb–Sr isotope data presented by Jacobsen & Heier (1978) for rocks around Kongsberg, showed a maximum age for the crust of about 1600 m.y. ($\lambda = 1.39 \times 10^{-11} . yr^{-1}$). Two metamorphic episodes at about 1600–1500 m.y. and 1200–1100 m.y. were recognised, with a series of granitic rocks formed by anatexis of pre-existing crust. An intrusion age of 1200 m.y. was defined for the large gabbro at Vinoren. A sequence of geological events in the Kongsberg district, suggested by the present author (Starmer 1977) correlates completely with the radiometric data of Jacobsen & Heier (1978) and is compatible with the history of rocks in the Flesberg area. In particular, cataclasis at the

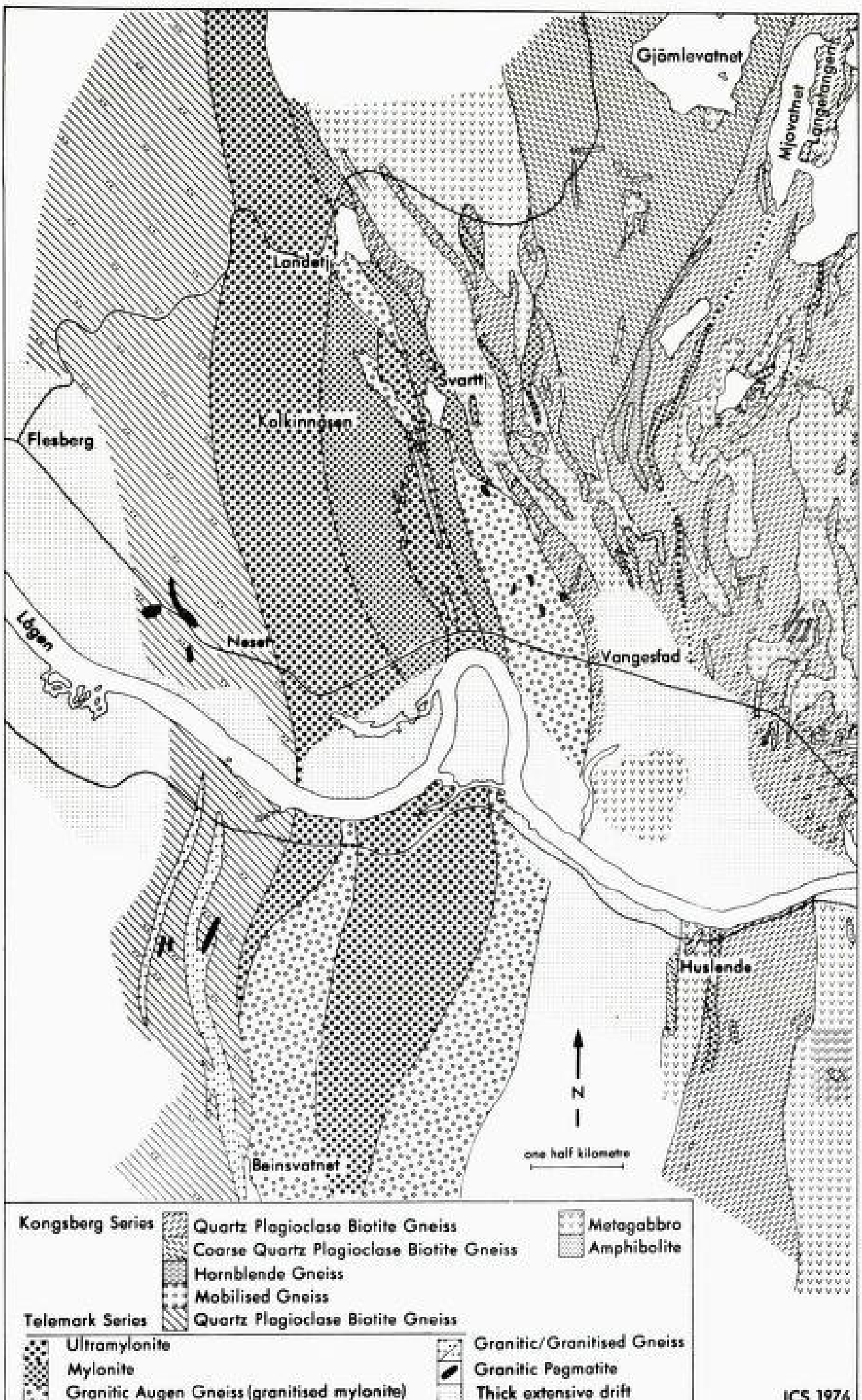

Fig. 2. Lithological map of the Flesberg area.

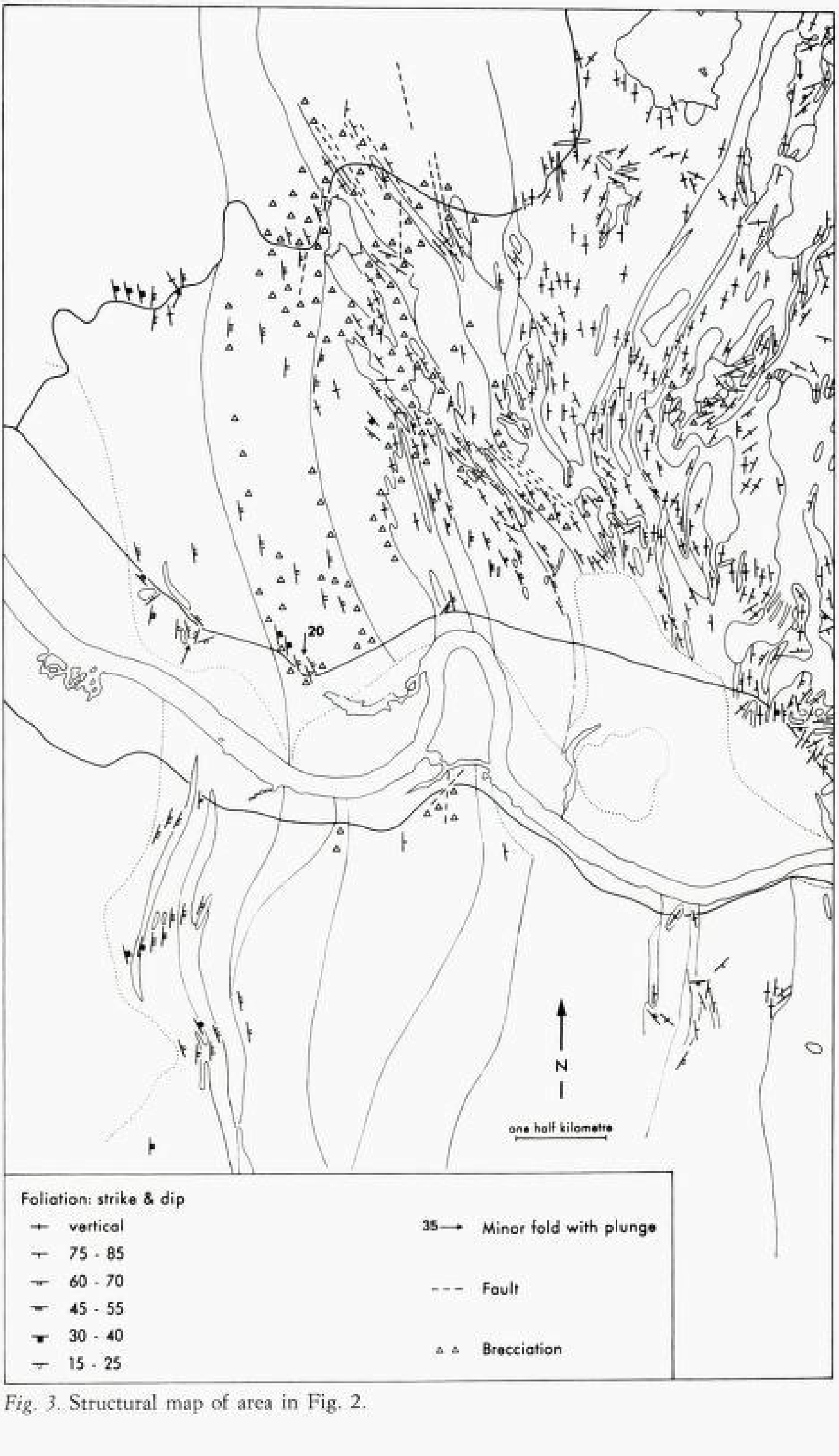

Fig. 3. Structural map of area in Fig. 2.

margin of the Kongsberg Series was thought to reflect uprise relative to the Telemark block, with besic Vinor intrusions emplaced during and after the closing stages of movement.

The mylonite zone

MYLONITE AND ULTRAMYLONITE

This zone (which really also includes the granitic augen gneisses, described subsequently) has the shallower dips, cross-folds and granitic pegmatites characteristic of the Telemark Series. In places, however, it contains less-altered enclaves of both Kongsberg and Telemark Series gneisses and it therefore seems inappropriate to assign it to either complex.

The main zone of ultramylonite (with sporadic mylonite and protomylonite) exceeds 1 km width near Flesberg and is gently flexured. It thins northwards towards Haugesjø, where it has a shallower dip, complicated by both major and minor folding: here, it is also more dominantly ultramylonitic, although less severe cataclastic effects are spread over a wider area than in the south.

The mylonite zone has a thoroughly transposive cataclastic fabric forming S-tectonites but, in many places, prominent stretching lineations or rolled porphyroclasts produce L–S tectonites. There were a number of movement phases and work is in progress to try to define these more precisely. The main mylonite–ultramylonite zone has fairly sharp boundaries, but on both sides the rocks show milder cataclasis, partly due to late movements taken-up by the surrounding gneisses after the competent mylonite zone had formed.

In the Kongsberg Series, protomylonites and sporadic mylonites are common over a distance of 1–2 km east of Flesberg (Figs 2–5). More isolated cataclasis occurs further east, particularly in a belt of mylonite just west of Lampeland where contemporaneity with the main zone is shown by a narrow ultramylonite cut by a thin Vinor amphibolite in Lyngdalselva. In the Haugesjø area (Figs. 6 & 7) the Kongsberg Series supracrustals are all mylogneiss and protomylonite, with occasional bands of mylonite. They become more severely cataclased westwards, towards the main mylonite zone.

The Telemark Series around Flesberg shows only weak cataclasis near the main zone, but further north in the Haugesjø area, the mapped rock are predominantly protomylonite and mylogneiss.

Everywhere within the ultramylonite and mylonite, isolated microclines (up to 1 cm size) grew across the lepidoblastic groundmasses of quartz ($\pm$plagioclase) which had been finely granulated (often to 0.1 or 0.05 mm size). Some quartz has recrystallised (to 0.2–0.5 mm size) in ribbons which rarely also contain microcline, suggesting contemporaneous crystallisation. Biotite (up to 1 mm) recrystallised after cataclasis, but is often ragged. Isolated plagioclase porphyroclasts (usually up to 3 mm, rarely 1 cm size) are variously sericitised and later intensely saussuritised after microcline growth. Occasionally the porphyroclasts are multi-crystal augen and may contain chloritised biotite. Some apparent augen are aggregates of fine, granulated groundmass (with or without a central plagioclase) surrounded by recrystallised streaks of coarser quartz and forming sites for microcline nucleation. Although there are some early garnet porphyroclasts, small (<5 mm) euhedral crystals have occasionally grown across the cataclastic fabrics.

The microclines are single or multi-crystal porphyroblasts (up to 1 cm or rarely 2 cm) forming isolated, rounded masses or irregular augen, with no preferred orientation. They are often replacive, shadowy perthite and antiperthite, nucleating on plagioclase porphyroclasts or forming poikiloblasts in the groundmass. In a few cases, weak cataclasis had caused minor granulation of the microclines.

In the mylonites and protomylonites, the effects were similar to those in the ultramylonites but finely granulated material developed only in certain bands, plagioclase porphyroclasts were more abundant and hornblendes recrystallised without retrogression.

The cataclased rocks all underwent late saussuritisation of plagioclase to clinozoisite. A late fabric of epidote associated with biotite and/or muscovite developed at this time. All the above features are cut by 'friction-breccia' veinlets of quartz, calcite and epidote.

The age of the cataclasis is thus well-defined, but minor late movements probably correspond to the diachronous activity in the Kongsberg area (Starmer 1977). Late cataclasis affected some of the early Vinor intrusions west of a line from Mjovatnet to Huslende (Fig. 2). On the margin of a metagabbro body, about 0.5 km north of Vangestad a thin ultramylonite has lenses of chloritised Vinor amphibolite included in a granulated groundmass. Late epidote and muscovite grew across both ultramylonite and basic lenses. Just to the west, actinolitic metagabbros have protomylonitised bands, but some of these are in composite bodies which just cut the edge of the mylonite zone. Other gabbros further east, towards Lampeland, include cataclastic xenoliths.

In a number of places around Haugesjø (Fig. 6) ultramylonite with microcline porphyroblasts, developed a late, random growth of prismatic hornblende. This phenomenon is particularly characteristic of granitic gneisses in the adjacent Telemark Series. Its origin is not known, but some occurrences are adjacent to exposed metagabbro or amphibolite.

GRANITIC AUGEN GNEISSES

Although, where well-developed, these are rather heterogeneous, granitic augen gneisses, they represent the end-product of the process of microcline porphyroblastesis and variable granitisation seen throughout the mylonite–ultramylonite zone. They contain large microcline porphyroblasts as euhedral crystals or augen, reaching 2 cm or even 3 cm size. Plagioclase porphyroclasts form single or multi-crystal augen. The groundmass consists of fine, granulated quartz and plagioclase (sometimes with later microcline). This is traversed by biotite, which, in a few cases, has partially formed from fine-grained hornblende. The development of these rocks is essentially the same as that of the mylonites. After cataclasis, they had a strong fabric defined by flaser trails of finely gran-

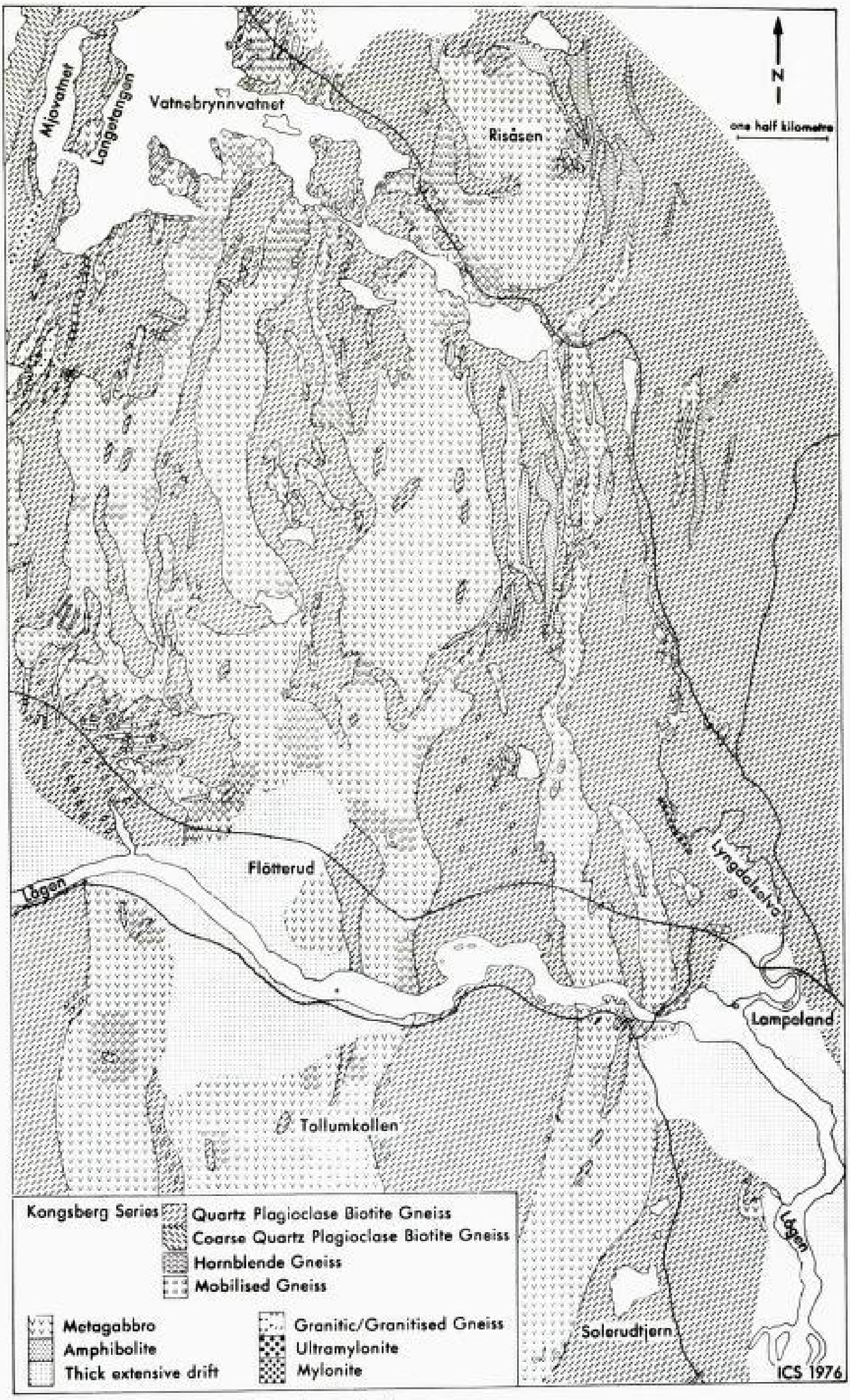

Fig. 4. Lithological map of the Lampeland area.

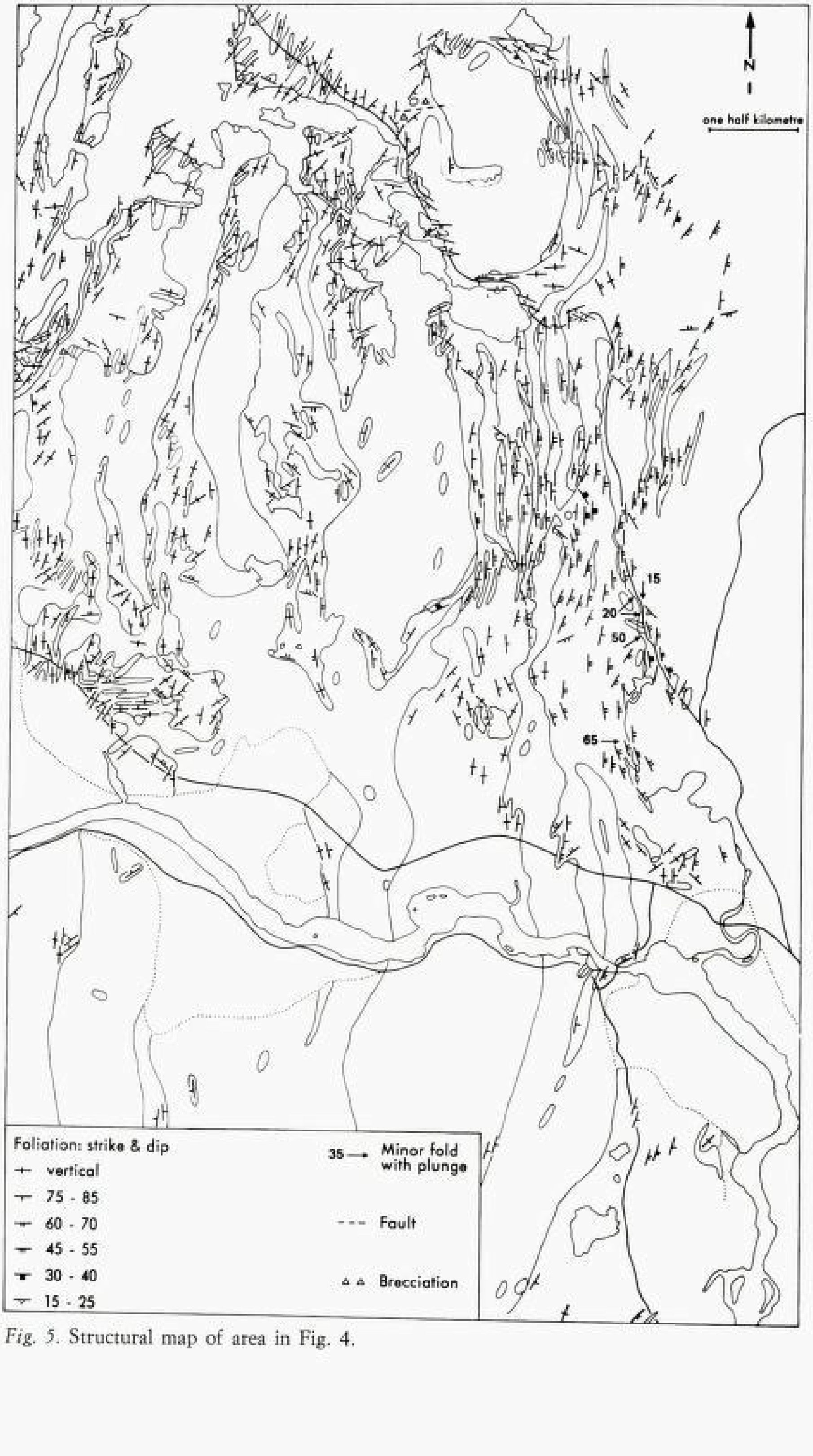

Fig. 5. Structural map of area in Fig. 4.

ulated quartz and plagioclase, often with streaks of biotite. Commonly they had numerous porphyroclastic augen of plagioclase. The microclines subsequently grew in a well-orientated manner, but were also larger and more abundant than in the rest of the mylonite zone. Rarely, microclines in certain bands were slightly granulated or even rolled, but later microcline growths traversed the fabric.

Basic Vinor dykes cut the augen gneisses, but both lithologies are occasionally dissected by thin granitic veins which also cut the Telemark Series amphibolites and are connected with a weak regeneration of granitic rocks and development of pegmatites.

The Kongsberg Series

QUARTZ–PLAGIOCLASE—BIOTITE GNEISSES

These gneisses represent metasediments and possibly some metavolcanics. Cataclasis increases westwards towards the mylonite zone and the rocks are commonly mylogneiss or protomylonite in the western part of the Flesberg–Lampeland tract (Figs. 2–5) and throughout the Haugesjø map area (Figs. 6 & 7).

The gneisses show all modal variations, with some developments of quartz–plagioclase gneiss (± muscovite). Some layers of quartz–plagioclase–hornblende gneiss may be related to intercalated hornblende gneisses. Quartz and plagioclase (oligoclase–andesine) are normally granoblastic or elongate (up to 2 mm size). Biotite forms a lepidoblastic fabric, sometimes partially emanating from hornblende.

A distinctive, coarse protomylonite type forms a discrete, major band, tracing south from Gjømlevatnet across Lågen (Fig. 2). A few thin layers also occur east of this main band. Further north, around Haugesjø (Fig. 6) the coarser and finer types alternate without developing discrete, major units. The coarse types contain sericitised plagioclase porphyroclasts, often developed as augen (0.3–1.0 cm in size). The groundmass commonly consists of flaser trails of finely granulated quartz and plagioclase (0.1 mm size) with ribbons of coarser, recrystallised quartz: biotite occurs in streaks or in clusters with hornblende.

All types of gneiss frequently contain almandine. They also show microcline porphyroblastesis and granitisation. Late fabrics of epidote and recrystallised biotite (± muscovite) occur in many rocks. Rusty 'fahlbånds', produced by disseminations of pyrite and minor chalcopyrite, are noticeably rare in this area compared to tracts further south towards Kongsberg. Their only major development is in the extreme east, between 2 and 3 km north of Lampeland (Fig. 4).

HORNBLENDE GNEISSES

These are part of the supracrustal sequence and are essentially well-segregated, plagioclase-rich amphibolites, although they are quite distinct from the later, Vinor amphibolites. They probably represent intercalated basic volcanics and occur everywhere as thin layers or lenses (usually less than 1 m wide and too small to be shown on the maps). Rarely, much broader irregular bands have somewhat transgressive contacts and interfingering apophyses, suggestive of intrusive relationships.

All types contain green-brown hornblende, andesine and quartz (up to 2 mm size) usually with some biotite. Almandine is common. Retrogressions formed biotite-rich layers and later chlorite, epidote and clinozoisite. These gneisses also contain pyrite ($\pm$chalcopyrite) and form part of fahlbånds. They are equivalent to the 'Hornblende Gneisses' unit of the Kongsberg area (Starmer 1977).

The Vinor injections cut all types of gneiss and, even when completely amphibolitised, are noticeably less-segregated and more melanocratic.

MOBILISED GNEISSES

These are extensively developed in the Flesberg–Lampeland area (Figs. 2–5) and formed by mobilisation of the supracrustal gneisses. They are particularly spectacular where they enclose xenoliths of metagabbro or have veined bodies of the latter. Elsewhere, totally within the supracrustals, more acidic (quartz–feldspar-rich) gneisses are mobilised around more mafic bands (hornblende gneisses, biotite-rich layers or fahlbånds) which have either remained intact or broken into lenses. Although there was sometimes a partial assimilation of basic rock and diffusion of margins, contacts often remained sharp. In thin-section, amphibolite and coronite show some marginal assimilation and mixture of minerals, with an increase in biotite content (and enlargement of biotite coronas around iron-ore crystals).

Veins and stockworks of mobilised material have intruded metagabbro, amphibolite and more mafic gneiss. In some cases, several generations of metagabbro and later amphibolitised dykes are all veined. Some later dykes cut through mobilised gneisses, but a few of these have lensed-out along their length. This suggests that the mobility continued, in some places, until after these dykes were intruded, with local variations due to changes in pp H_2O.

The mobilised gneisses are usually white in colour, with grain sizes of 0.5 to 5.0 mm, but occasionally grading to pegmatitic segregations (over 1 cm crystal size) in discrete veins or immediately adjacent to xenoliths. They contain granoblastic quartz and oligoclase–albite, with randomly oriented biotite, but they may become weakly lepidoblastic adjacent to xenoliths or vein margins. Microcline (up to 5 mm size) with good cross-hatching, occurs in varying quantities, but may be entirely absent; where it is more abundant, the gneisses can assume a granodioritic composition. An epidote–biotite fabric, with associated clinozoisite growth in plagioclase, developed after mobilisation and is often lepidoblastic.

In the early stages of mobilisation, networks of thin veinlets (usually 1.0–1.5 mm wide) traverse the rock and often concentrate around xenoliths. The networks contain embayed quartz and sericitised twinned plagioclase with diffuse margins. These crystals are invaded by vermicular, lobate and bleb-like quartz and plagioclase which seem to have generated from them: the blebs (usually 0.02–0.75 mm size) are neither isotropic nor fully ordered and commonly have undulose extinction with dark marginal zones. The blebs also invade biotite and muscovite and the microcline porphyroblasts of granitised specimens.

The mobilised gneisses can be seen to generate gradually from the more acidic supracrustals and are not due to introduced, anatectic material. Increased fluid-vapour activity caused the sweating-out of these segregations from the gneisses under locally increased heat-flow. This thermal upgrading, around the roof levels of the large gabbro body, was overprinted on the regional amphibolite facies conditions. In some cases, almandines have grown across amphibolite and mobilised gneiss junctions and fabrics, reflecting a continuation of these conditions after mobilisation and before the epidote–amphibolite facies fabrics were superimposed.

GRANITIC GNEISSES

In the Kongsberg Series, granitic gneisses and granitised supracrustals (with sporadic augen developments) are related to activity of the same general age as that producing augen gneisses in the mylonite zone. They are concentrated in the west, near the mylonite zone, but occur throughout the area.

These lithologies normally consist of granoblastic quartz and albite–oligoclase, with lepidoblastic biotite and later microcline and microperthite (up to 2 mm size). In some rocks, microcline overprints granulated bands or replaces plagioclase porphyroclasts, giving augen up to 2 cm in size. Late effects include development of epidote and muscovite, saussuritisation of plagioclase and sometimes a new growth of biotite.

The Telemark Series

The Telemark Series differs from the Kongsberg Series in having a much shallower easterly dip and abundant granitic pegmatites. Concentric folds are also more common and often interfere to produce major and minor doming. Amphibolites occur as dykes, but no plutonic bodies were found (akin to the Kongsberg Series metagabbros).

QUARTZ—PLAGIOCLASE–BIOTITE GNEISSES

These gneisses show all modal variations, with minor hornblende in some layers. In the Flesberg area (Fig. 2) they are weakly cataclased towards the mylonite zone, but further north, in the mapped area around Haugesjø (Fig. 6), they are predominantly protomylonite and mylogneiss. In addition, a lithological change northwards from Flesberg to Haugesjø may reflect a different stratigraphical level adjacent to the mylonite zone.

Around Flesberg, the gneisses do not contain augen and are superficially not greatly different from those of the Kongsberg Series, although they are generally slightly coarser (up to 5 mm size crystals) and richer in biotite and epidote. Coarse biotite schists are often developed. Banded gneisses are commonly formed with a series of early biotite amphibolites, which are too small to be shown on the maps. (They would seem to be equivalent, in many ways, to the hornblende gneisses of the Kongsberg Series). The gneisses and biotite amphibolites are variously granitised and sometimes cut by later, intrusive amphibolites.

The mylogneisses and protomylonites of the Haugesjø area, tend to be coarser than the adjacent Kongsberg Series supracrustals (or the Telemark Series around Flesberg). They often have plagioclase porphyroclasts or augen (up to 1 cm size) although these developments are random and sporadic. A very characteristic lithology, with large white augen (reaching 2–5 cm size) is also developed sporadically, but particularly in a belt stretching from the west of Strandvatnet northwards to Haugesjø. The augen are of plagioclase (with subordinate microcline in some specimens) within a fine, granulated groundmass. Layers and lenses of quartz–biotite–plagioclase–hornblende gneiss may represent early amphibolitic rocks (broadly equivalent to the biotite–amphibolites of the Flesberg area). All types of gneiss underwent variable granitisation after cataclasis.

The rocks have a groundmass of quartz and plagioclase which varies from granoblastic to very finely granulated (<0.5 mm size) in the more cataclased specimens. The groundmass is traversed by lepidoblastic biotite laths or ragged growths in cataclased rocks. The augen (where present) are aggregates of granulated quartz and plagioclase (often 0.5 mm size) with, or without, a central plagioclase porphyroclast, which has often partially recrystallised to the surrounding smaller crystals, late in the cataclasis. Biotite streaks, bent around these aggregates, produce the augen appearance. In some specimens, shadowy microcline (white in hand specimen) forms single or multi-crystals porphyroblasts and augen, often as patch perthites replacing plagioclase, or as poikiloblasts enclosing granulated groundmass.

GRANITIC GNEISSES AND GRANITIC AUGEN GNEISSES

In the Telemark Series, granitic and partially granitised gneisses are common and formed after cataclasis, although weak shearing occurred later in some. Early biotite amphibolites were partially granitised, but later, intrusive amphibolites cut the granitised rocks and all were deformed by concentric folding. Subsequent minor granitic activity and veining are considered in the following section. Many of the granitic rocks developed late hornblende growths as random, prismatic crystals up to 1 cm length.

The granitic and granodioritic gneisses contain granoblastic quartz, oligoclase and replacive, perthitic microcline (up to 2.5 mm grain size) with isolated biotites defining a weak foliation. Hornblende and muscovite rarely occur and minor components include epidote, sphene and magnetite. In some rocks, microcline overprinted layers of granulated quartz and plagioclase.

In a few places, the granitic gneisses have local developments of microcline and plagioclase augen (up to 1 or 2 cm size) but grade into the only major granitic augen gneiss unit northwest of Strandvatnet (Fig. 6).

GRANITIC PEGMATITES AND GRANITIC VEINS

Granitic pegmatites containing quartz, microcline, plagioclase and biotite

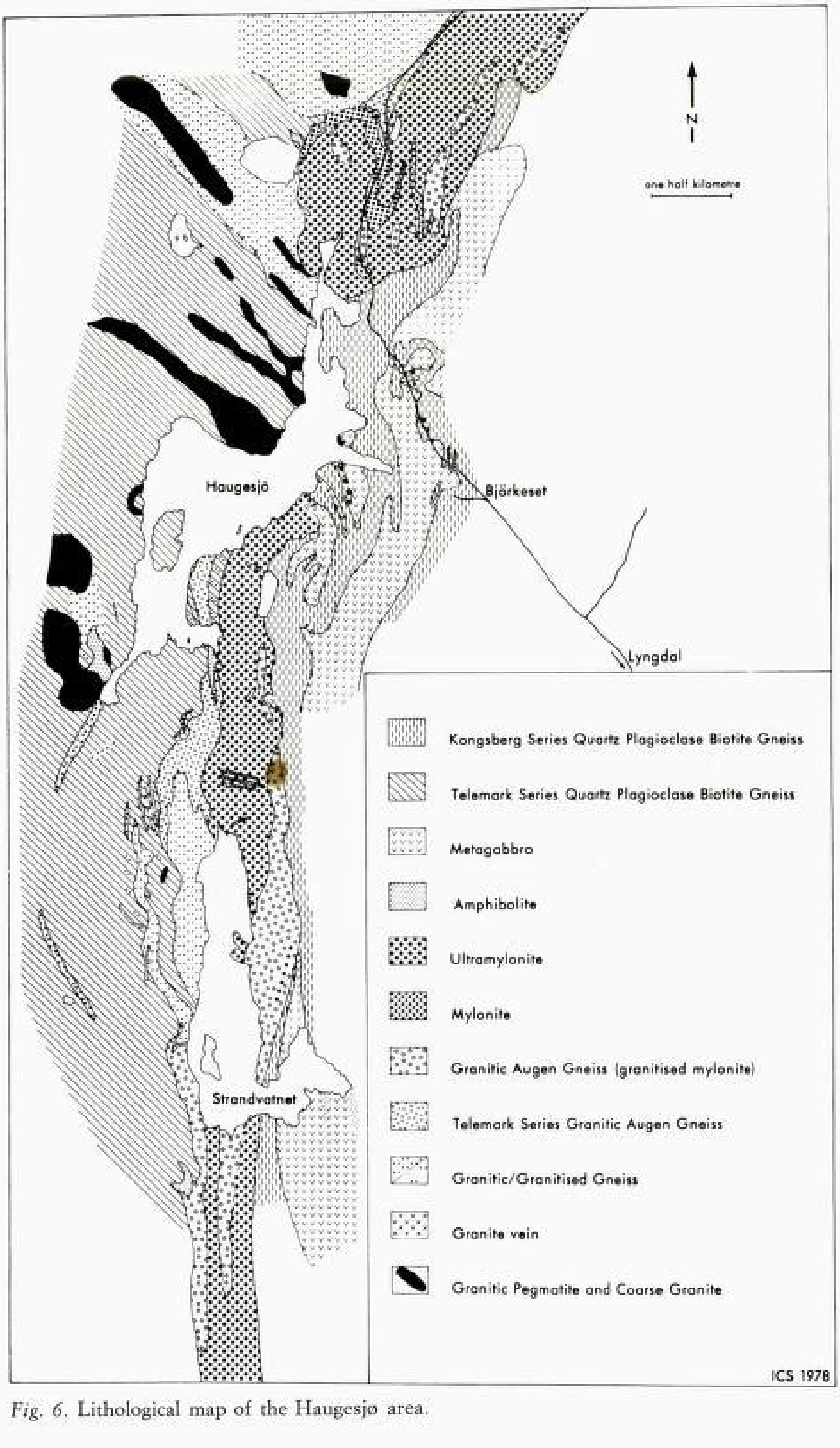

Fig. 6. Lithological map of the Haugesjø area.

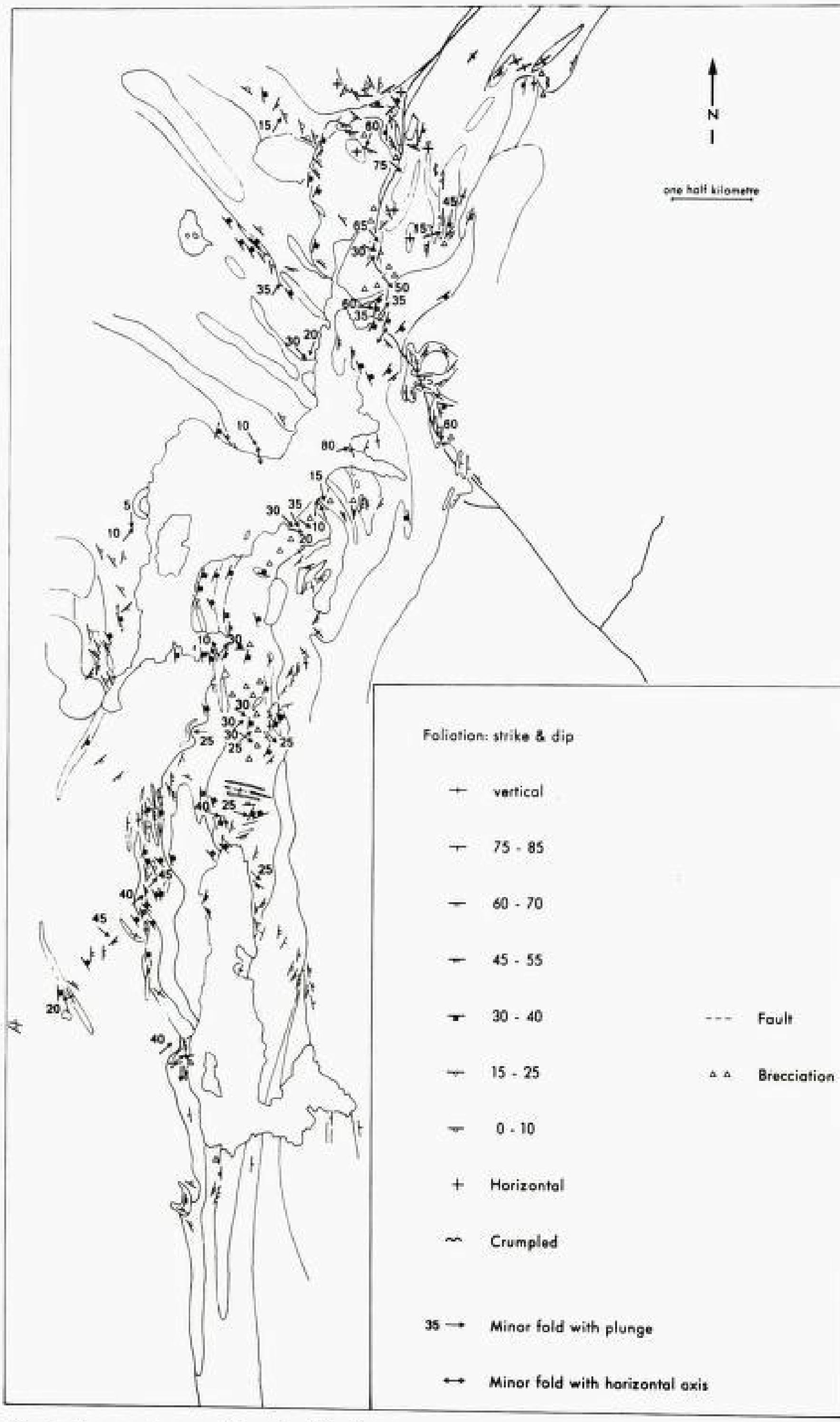

Fig. 7. Structural map of area in Fig. 6.

(± muscovite) cut all rocks in the Telemark Series. They also vein the mylonite zone and granitic augen gneisses. West of Haugesjø (Fig. 6) large outcrops of pegmatite and pegmatitic granite contain xenoliths of cataclased gneisses and of the intruded amphibolites. Around them, supracrustals and granitic gneisses are often soaked in pegmatite veins.

The pegmatites are contemporaneous with a weak regeneration of granitic rocks, giving thin (< 1 cm wide) veins and localised microcline porphyroblastesis, affecting the intruded amphibolites and the concentric folds. Rarely, wider veins have formed. A 5 m wide granite body trends N–S in Telemark gneisses (developed here as the type with large plagioclase augen) just northwest of Strandvatnet. This has small apophyses, but its contact is concordant to the foliation in the gneiss and the granite itself has developed a weak fabric near its margin.

The intruded gabbros and amphibolites

The 'Vinor' intrusions of the Kongsberg Series and the amphibolitised dykes of the Telemark Series are grouped together in this description since they belong to the same general period of basic injections.

VINOR METAGABBROS AND AMPHIBOLITES

The Vinor intrusions represent a series of basic injections into metamorphosed and cataclased supracrustals. They were subsequently partially amphibolitised, particularly in the east, but still often preserved their igneous textures. A number of injections can be distinguished, but across the area as a whole three distinct phases are separable. These are now represented by (i) early olivine gabbro and olivine norite coronites, (ii) later coarse metagabbros and (iii) a separate phase of finer grained, heavily amphibolitised dykes.

The earliest intrusions of olivine gabbro and olivine norite (usually 5 mm grain size) commonly have ophitic textures, with corona growths around mafics and andesine rims on the original labradorite-bytownite laths. The rocks are melanocratic, often with a purple coloration which is also common in the Bamble coronites (or 'hyperites') where it has been attributed to dust inclusions in the feldspars (Starmer 1969). More intensive metamorphism of these bodies formed metagabbros consisting of actinolite or actinolitic hornblende (often as felted masses almost completely replacing pyroxenes) and granoblastic andesine (sometimes with relict cores of labradorite-bytownite). All the plagioclases are saussuritised, concomitant with a growth of chlorite and epidote in the rest of the rock. Late retrogressions in some have caused complete diapthoresis to greenschist assemblages.

A second set of intrusions is now represented by more felsic, green-coloured metagabbro with a coarse, decussate growth of hornblende, actinolite and andesine. In some the plagioclase is purple. Grain size is usually greater than 5 mm. Widespread felsic segregations are developed and, in a few cases, these may have resulted from contamination by mobilised gneiss. This second phase of intrusions often veins the earlier coronites and may enclose xenoliths with both sharp and diffuse margins, suggesting that the earlier gabbro was not totally consolidated during incorporation.

Towards the mylonite zone (particularly west of a line from Mjovatnet to Huslende – Fig. 2) more felsic metagabbro is more common and parts of the bodies show effects of very late cataclasis and weak granitisation. A few dioritic patches occur, but they are not as extensive as in areas further south, close to the Kongsberg Series margin. Although the bodies are cataclased in some parts, they also cut the edge of the main mylonite zone. They are composite masses which combine intrusions of both the first and second phases. Further east (between Flesberg and Lampeland and east of Haugesjø) other composite bodies, comprising both first and second phase intrusions, contain cataclased xenoliths. Apart from the isolated occurrences of late cataclasis in the west (mentioned above) the large gabbroic bodies generally have a post-tectonic form and cut cataclastite fabrics throughout the area.

The third, late phase of finer-grained, heavily amphibolitised intrusions is represented by small bodies, many of which cannot be shown on the maps. They usually occur as thin dykes or concordant sheets (10 cm to 0.5 m wide) and their fine grain size (usually $<$0.5 mm) has facilitated severe amphibolitisation, even during waning metamorphism. Rarely, they contain small ($<$3 mm) almandines. The dykes frequently cut the earlier metagabbros and dissect mobilised gneisses with incorporated metagabbro xenoliths. Often, intrusion trends through metagabbros are controlled by the foliation of the surrounding gneisses. Some dykes which cut mobilised gneisses are lensed-out further along their length. Occasionally, two generations of these dykes can be seen in supracrustals and mobilised gneisses.

The amphibolitised dykes cut discordantly through granitised mylonites and their associated augen gneisses. A spectacular example of this is seen just west of Svarttjern (Fig. 2) where small apophyses inject around and between microcline porphyroblasts. A very weak and late phase of granitic veining in pluton margins and amphibolite dykes has already been discussed.

The amphibolites carry nematoblastic green-brown hornblende, with granoblastic andesine, quartz and late epidote ($\pm$biotite). Some contain remnant labradorite laths. Occasionally crushing is accompanied by complete chloritisation.

The major gabbro bodies (Figs. 2–7) are usually composite, derived principally from the first and second phases of intrusion. Although they show a tendency to elongation, parallel to the regional structure, their complex outcrop patterns represent protrusions from a much larger mass at depth. Supracrustal roof pendants have complex and interfingering margins. The large mass at depth is probably the northern end of the huge 'Vinoren' body, which is exposed for some 8 km south of Lampeland (Bugge 1937). Within the present area, the main mylonite zone marks the western limit of large gabbros outcrops. In the east, they also disappear; north of Lampeland, several small gabbro outcrops occur in Lyngdalselva but not at a slightly higher level in the adjacent road, suggesting that the large gabbro mass may be at greater depths in that area.

Mobilised gneisses are extensively developed in the east, but are less common towards the western mylonite zone in the Flesberg and Haugesjø areas. Surveys around the southern end of the large 'Vinoren' body (at Dronningkollen, some 8 km south of Lampeland) have shown developments of mobilised gneiss with xenoliths of metagabbros and fahlbånds. The mobilisation, therefore, is associated with the roof of this huge gabbro mass. The gneisses involved are more acidic than many supracrustals in other parts of the Kongsberg Series and their plastic behaviour probably facilitated the relatively passive emplacement of this large basic body and the development of some globulithic features. Their mobility resulted from increased fluid-vapour activity and a greatly increased heat flow associated with the gabbro uprise, but not emanating entirely from the cooling magma. The gneisses were mobile when some of the earlier gabbro intrusions were partially or totally consolidated. The gabbros were held at high subsolidus temperatures for a prolonged period and cooled very slowly under elevated metamorphic conditions. The coronite pyroxenes often show some exsolution features which, in some places, became extreme (e. g. at Tollumkollen, south of Lågen – Fig. 4).

TELEMARK SERIES AMPHIBOLITES

The Telemark Series has numerous, thin, intruded amphibolites, but in the mapped areas they never develop the plutonic forms found in the Vinor intrusives of the Kongsberg Series. They resemble the amphibolitised, late Vinor dykes and sheets cutting Telemark gneisses, granitic gneisses and the mylonite zone. They may represent several intrusive phases, since some, which are mildly sheared and biotite-rich, are adjacent to unsheared bodies. Late injection of granitic pegmatites and thin granitic veins into these amphibolites has been discussed previously.

The amphibolites generally have nematoblastic green-brown hornblende (up to 3.5 mm size) with granoblastic quartz and saussuritised andesine (usuallly up to 2 mm size). Biotite follows the fabric and is partially retrograde from hornblende. In sheared rocks, it may form sheaves (up to 6 mm size) of acicular biotite and remnant hornblende. Extreme retrogression has produced greenschist facies assemblages. Some amphibolites have a late, random overgrowth of large hornblendes (up to 2 cm size) possibly related to similar developments in granitised gneisses and ultramylonites.

Structure of the area

The Kongsberg Series structurally overlies the mylonite zone and the Telemark Series. The regional foliation and lithobanding strike N–S with an easterly dip which decreases from east to west, through the Kongsberg Series and mylonite zone into the Telemark Series. This is complicated in the Haugesjø area (Figs. 6 & 7) by concentric cross-folding in the Telemark Series and by the development of a complex basin in the ultramylonite.

The earliest foliation observed in the Kongsberg and Telemark gneisses is a penetrative fabric parallel to the lithobanding, which is considered to represent transposed bedding. Sporadic development of asymmetric, intrafolial, shear folds (within non-cataclased rocks) may result from this process.

Later, almost coplanar cataclastic fabrics overprinted the gneiss foliations, particularly in the west, towards the mylonite zone. All the above structures were locally deformed around the subsequent Vinor intrusions, partly during their emplacement and partly during a later phase of concentric folding.

A few isolated, minor folds post-date the intrusion of the Vinor and Telemark amphibolites, but pre-date a phase of regional concentric folding. These tight to sub-isoclinal shear folds (usually <1 m wavelength and amplitude) have moderate to shallow plunges to the south and occasionally deform some intruded amphibolites without affecting adjacent dykes. On the Langetangen peninsula in Vatnebrynnvatnet (Figs. 2–5) the shear folds are deformed by N–S concentric folds.

Open, concentric folds on subvertical axial planes developed after the intrusion of most Vinor bodies and Telemark amphibolites, but before the injection of granitic pegmatites. These structures formed on NE–SW and NW–SE axes (which were sometimes more N–S and E–W, particularly near resistant gabbros): they occurred as minor folds throughout the area, but also developed as major structures and were most intensive in the Telemark Series, becoming more sporadic and open in the mylonite zone and less common in the Kongsberg Series.

Concentric cross-folding was most strongly developed in the north, around Haugesjø, where it produced domes and basins on all scales and helped to form the major bend in the Kongsberg Series margin. Figs 6 and 7 show a major, open basin in ultramylonite to the north of Haugesjø: this is a complex structure (with shallow plunging, minor cross-folds) and is located immediately north of a discontinuity in the ultramylonite zone. A major structure, complementary to the basin, occurs just to the north, in the Telemark Series. The combination of the discontinuity in the main mylonite zone with these structures formed the only major bend in the Kongsberg Series western margin. Further south, in the Flesberg area, effects were less marked with isolated, minor folds (usually without cross-folding) and gentle flexuring of the mylonite belt on a major scale. Local deformation occurred around protruding gabbros and west of Fløtterud (Fig. 4) a large flexure has associated minor folds which deform some Vinor dykes but are cut by others. A concentration of open concentric folds and domes occurs along Lyngdalselva, about 1.5 km north of Lampeland (and about 5 km east of the mylonite zone).

Where cross-folding produces doming, the two sets of interfering structures usually seem contemporaneous, but rarely folds with NE–SW axes deform those on NW–SE axes. This is seen both to the north of Lampeland and in the Haugesjø area. Wavelengths range upwards from 20 cm and the respective amplitudes are much smaller. The style and tightness varies, to some extent, with lithology and previous structure. Axial plunges are commonly low to moderate to the NE or SE, being largely controlled by the pre-existing easterly layer dip. A few folds in the Kongsberg Series are markedly asymmetric (up to 5 m amplitude) with NE–SW to N–S axes and steep axial planes: vergences suggest an upward westerly movement of the eastern side. Isolated recumbent folds of the same phase deform Telemark amphibolites west of Neset (Figs. 2 & 3) and affect the eastern side of the ultramylonite zone between Haugesjø and Strandvatnet (Figs. 6 & 7).

The 'friction breccia' movements and later faulting

Brittle fracture phenomena corresponding to the 'friction breccia' movements of A. Bugge (1928, 1937) are developed randomly over a wide area, in and around the mylonite zone. The brecciated fractures are not continuous and are not of regional significance in separating the two complexes: they result from a number of tectonic events, including intensive developments of conjugate jointing on concentric flexures (particularly extensively developed in the ultramylonite between Haugesjø and Strandvatnet — Figs. 6 & 7).

Where discrete, well-defined fault planes are observed, they often post-date the main brecciation. The faults vary somewhat in orientation, but generally strike N–S to NW–SE and dip steeply east. Where definable, the displacements show a consistent pattern, being reverse to oblique and more rarely dextral strike-slip (with the eastern sides moving upwards and/or southwards). The strike-slip faults occur throughout the area (e.g. west of Landetjern, on the southwest side of Risåsen and along the west shore of southern Haugesjø) and may represent a separate displacement to the oblique-reverse faulting. A few random shear planes around gabbro margins have no coherent pattern.

Retrogressions associated with the brecciation and faulting show lower to middle greenschist facies assemblages. The 'friction breccias' are commonly accompanied by hydrothermal veins, developed on all scales from microscopic to 20 cm width and containing quartz, calcite and epidote (either alone or in combination). In some cases, pure epidote veinlets are cut by calcite veinlets. Late stage drusy quartz occurs locally.

The 'friction breccias' and veins transect all rocks and structures, including the pegmatites, the concentric folds and the ubiquitous saussuritisation. Examples of disoriented xenoliths of granitised ultramylonite within 'friction breccia' veins, show clearly the separation of syn-metamorphic cataclasis and post-metamorphic brecciation. Many of the 'friction breccia' movements were concentrated around the mylonite zone because of the relatively planar and brittle nature of this belt of instability at the junction of the Kongsberg and Telemark Series. Brecciation of the same age also occurs in restricted zones throughout the area.

Summary

In the present area, the boundary between the Telemark and Kongsberg Series is a zone of mylonite and ultramylonite which is granitised in places to form augen gneisses. The boundary is not the so-called 'friction breccia' which is the result of brittle fracturing developed long after the syn-metamorphic cataclasis.

The only major bend in the Kongsberg Series margin occurs in the present area. It resulted from the development of a major basin in the mylonite zone, with complementary structures in the Telemark Series. It was located at this point because of a tectonic break in the mylonite zone and because this area may have been a locus of instability between the huge Vinoren gabbro to the south and large gabbros to the north around Lauvnesvatnet (Starmer, in prep.). The mylonite zone was also thinner and less steeply inclined at this point.

From the foregoing descriptions and discussions, the sequence of events in this area is as follows:

(1) The Kongsberg and Telemark Series consisted of supracrustals (metasediments and basic volcanics) which underwent middle to upper amphibolite facies metamorphism and developed a penetrative foliation.
(2) A complex upthrust of the Kongsberg Series (at middle to upper amphibolite facies grade) produced a major mylonite zone along the junction with the Telemark Series.
(3) Granitic activity confined largely to the Telemark Series, caused ubiquitous microcline porphyroblastesis in the mylonite zone and sporadic developments of granitic augen gneiss (at middle to upper amphibolite facies).
(4) Several intrusions of 'Vinor' gabbro produced composite plutonic bodies (including the large Vinoren body) in the Kongsberg Series. The earliest injections (in the west) were slightly cataclased by late movements and weakly granitised but later ones include granitised cataclastite xenoliths. Metamorphic conditions were middle to upper amphibolite facies and acidic gneisses were mobilised by increased heat flow around the large gabbro mass, a mobilisation which partially overlapped stage (5).
(5) The late Vinor dykes and sheets and the Telemark amphibolites were then emplaced (middle to upper amphibolite facies); these cut the mylonite zone.
(6) Major and minor concentric folding, on two axial trends, produced domes and basins and caused the major bend in the Kongsberg Series margin at at point where there was a tectonic break in the mylonite zone.
(7) Granitic pegmatites invaded the Telemark Series and the mylonite zone. Thin granitic veins probably represent a weak regeneration of granitic rocks.
(8) Prolonged metamorphism at epidote amphibolite facies produced overprints of epidote, clinozoisite, biotite and muscovite in all rocks.
(9) Brittle-fracture ('friction-breccia') movements at lower to middle greenschist facies were followed by hydrothermal veining. Late faults show a consistent displacement pattern, with the eastern side moving upwards and/or southwards.

Many of the stages in this sequence are shown particularly well within a very small area on the east shore of Haugesjø (Fig. 6). Metagabbro there intrudes granitised ultramylonite (with inclusion of xenoliths) and is cut by amphibolitised dykes. Interbanded ultramylonites and amphibolites have been folded about subhorizontal NNE and NW axes and all lithologies and structures are cut by granitic pegmatites. Later 'friction breccia' veins cut all rocks.

The two complexes were effectively welded together at the end of the thrusting and cataclasis and subsequently had a common history, although granitic activity and later pegmatites were concentrated in the Telemark Series and plutonic gabbros in the Kongsberg Series. The relative age of the cataclasis is well-defined and the overall displacement was an upthrust of the Kongsberg Series, but it involved a series of movements. Work is now in progress on the continuation of the mylonite zone and on fabric analysis of the movement directions.

If the large Vinoren gabbro of Bugge (1937) is one entity, it is exposed intermittently over a N–S distance of 13 km and an E–W width of up to 3 km. Its size and the extent of mobilised gneisses around it, both reflect a zone of increased heat flow and fluid-vapour activity. The evidence of prolonged cooling of the gabbro also supports the idea that the magma was not the sole source of the increased heat flow. Gabbro bodies extend westwards in the Kongsberg Series to the mylonite zone, although later dykes cut through it and are also represented in the Telemark Series. The gabbros pierced exposed levels at the end of the upthrusting which produced the cataclasis and show some tendency to more felsic compositions near the mylonite zone. All of these features suggest a connection between the upthrusting, the gabbro intrusions and the greatly increased heat flow. The granitic activity at the end of the cataclasis may also be linked to high regional heat flow and anatexis at depth. Thus, the present area displays evidence of a number of major petrological and tectonic processes which are linked to large-scale crust-mantle interactions in the Proterozoic.

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