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Geology and Structure of Gjerøy, Nordland

RICHARD J. LISLE

Lisle, R. J. 1981: Geology and structure of Gjerøy, Nordland. *Norges geol. Unders 370*, 1–9.

The island of Gjerøy is made up of acid gneisses with subordinate mixed gneisses which are tightly interfolded with banded gneisses and associated metasediments. The early tight folds are refolded to produce a large-scale interference pattern. Basic sheets which cut the migmatitic banding in the acid gneisses are themselves metamorphosed at amphibolite facies and folded by the earlier set of tight folds. The relationship of the banded gneisses and metasediments to the basic dykes is unclear and the problem of the stratigraphic ages of these rocks unresolved.

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Introduction

The island of Gjerøy (approx. 66°36'N, 12°20'E) lies 10 km north of the Arctic Circle and approximately the same distance west of the Norwegian mainland. Its geology has never been described but immediately adjacent areas have been mapped by Holmsen (1932), Rutland & Nicholson (1965) and Vreeken (1979). On large-scale compilation maps the area has been assigned to the basement (with parautochthonous cover) by Rutland & Nicholson (1965).

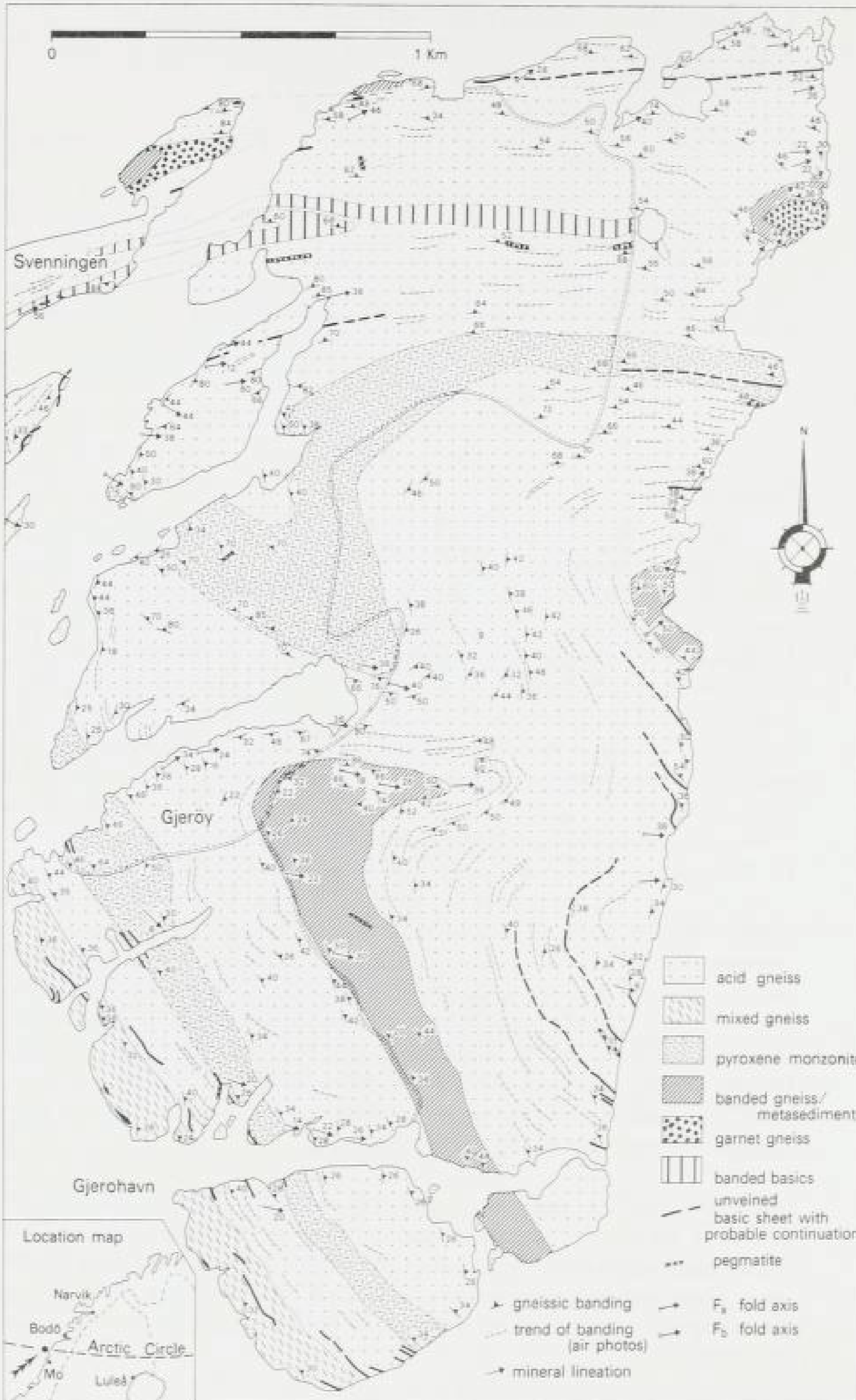
Reconnaissance mapping by the writer during the summer of 1979 shows the island to consist of veined granitic gneisses interfolded with paragneisses of similar metamorphic grade. A large-scale pattern of interference is produced by two sets of folds.

This paper with the accompanying map (Fig. 1) provides a short account of the lithologies, structures and geological history of the island.

ROCK TYPES

Acid gneisses: Most of the island is made up of coarse-grained microcline-rich quartzo–feldspathic gneiss with small amounts of khaki-brown biotite and often with a blue-green amphibole as well. Sodic plagioclase and microcline commonly occur together as perthites. A plane-parallel arrangement of quartz–feldspar veining sometimes produces a banding (Fig. 2f) which is itself paralleled by a weak mineral foliation defined largely by mica and amphibole. Elsewhere, veins are ptygmatically folded (Fig. 2a).

Mixed gneisses: This group of gneisses occupying the SW coastline and the NW tip of the island is more heterogeneous than the granitic gneisses, and has darker components occupying small pods and thin bands with diffuse boundaries. These mafic components contain a blue-green amphibole often accompanied by clinopyroxene. This rock type is transitional to the banded basics described below.


0
1 Km
Svenningen
Gjeröy
Gjerohavn
Location map
Narvik
Bodö
Arctic Circle
Mo
Luleå
N
acid gneiss
mixed gneiss
pyroxene monzonite
banded gneiss/ metasediment
garnet gneiss
banded basics
unveined basic sheet with probable continuation
pegmatite
gneissic banding
trend of banding (air photos)
mineral lineation
F_a fold axis
F_b fold axis

A string of pods of larger size runs parallel to the SW coast following the strike of the rocks and reappears on the island of Svenningen and in the NW tip of the island (Fig. 1). Some of them are feldspar-free and consist of hornblende, clinopyroxene and sometimes biotite. Another pod on the SW corner of the main island has a mottled appearance. In thin-section this appearance can be seen to be produced by large labradorite crystals which have enclosed equant epidote and clinopyroxene crystals.

Pyroxene monzonite: A distinctive rock because of its homogeneity and lack of veining, this monzonite/monzodiorite forms an important marker on the map (Fig. 1). The monzonite is often foliated and locally shows a faint banding (Fig. 2b). The dark minerals defining the banding are clinopyroxene and hornblende. In thin-section, plagioclase, alkali-feldspar, quartz, clinopyroxene and hornblende are interpreted as magmatic minerals whilst hastingsite is secondary and has replaced the pyroxene.

Banded gneisses and metasediments: Within fold cores in the central part of the island and on the east coast, a group of rocks occur which are made up predominantly of banded gneiss (Fig. 2d, 4a and 4b). The banding is defined by an alternation of quartz–biotite-rich and hornblende ($\pm$garnet)-rich layers. Calc-silicate rocks, together with sillimanite–biotite gneisses, garnet–biotite gneisses and occasional thin quartzites, make up the rest of this group. The banded gneisses and metasediments are grouped together by virtue of their close field association. The nature of the source material of the former rocks which show an extremely regular banding is obscure. On a mesoscopic scale the are S-tectonites (Flinn 1965, Lisle 1977) and resemble gneiss-types whose field appearance is deduced to have been brought about by the imposition of a finite strain of flattening-type and of large magnitude (Myers 1978).

The boundary between these rocks and the acid gneisses is everywhere sharp and concordant with the banding in rocks on both sides of the boundary, but a few isolated layers of garnetiferous gneisses occur well within the acid gneisses (Fig. 1).

Veined basics: A particularly thick sheet of basic gneiss of this type extends E to W across the north end of the island and is composed essentially of a hornblende, andesine and clinopyroxene (Fig. 2e). Other thin layers with similar field characteristics fall into this group and contain biotite. The occasional presence of small amounts of microcline and of veining in these basics suggests that they are older than the migmatization which took place in the granitic gneisses.

Unveined basics: The rocks for which this field name was used occur in thin sheets (Fig. 2c, 2f), sometimes folded or boudinaged and lacking the veining and internal lithological banding typical of the rocks described above. They are hornblende–biotite schists. The hornblende now has less of a blue tint than in

Fig. 1. Geological map of Gjerøy.

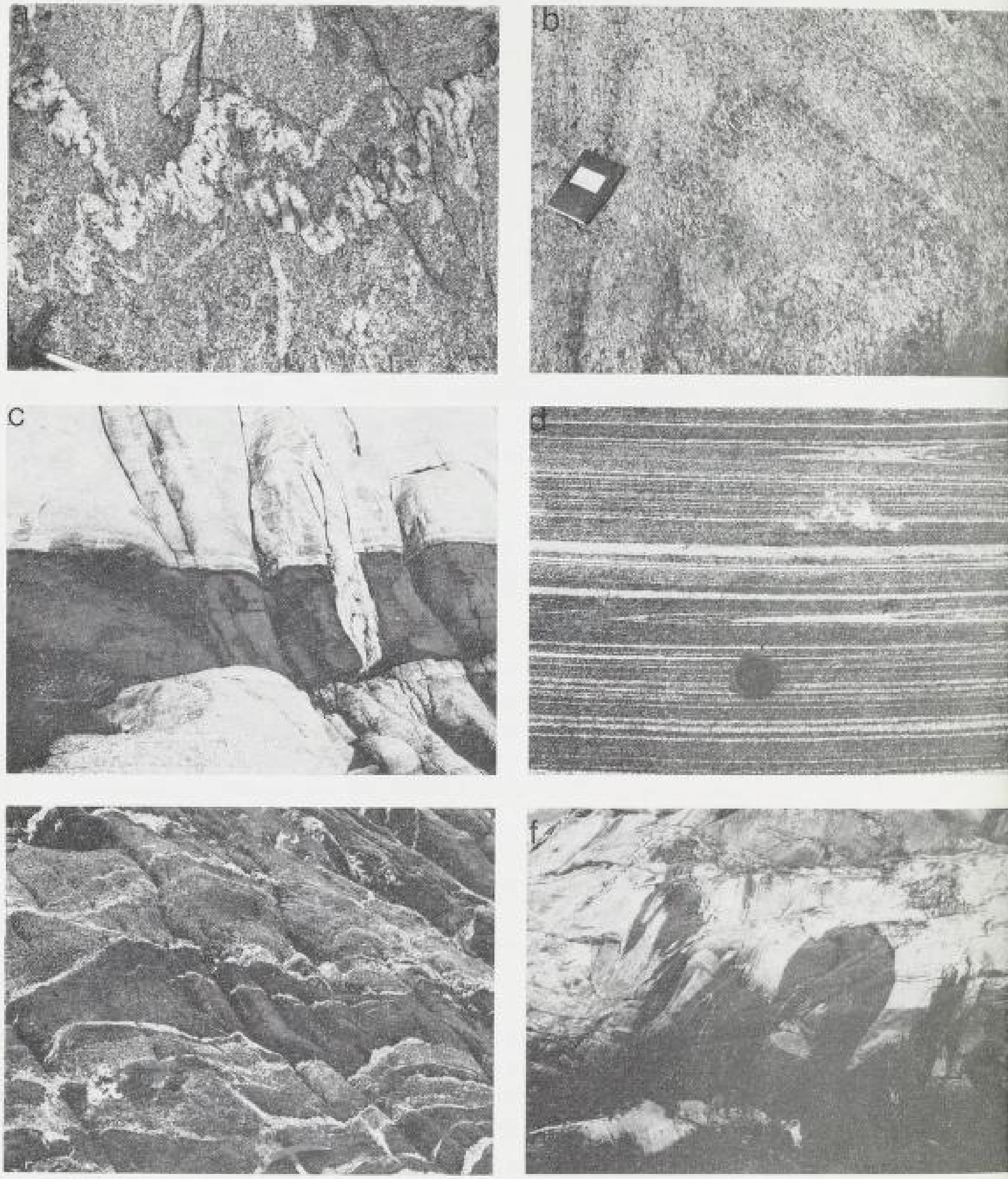


Fig. 2. Lithologies and structures. (a) Acid gneiss with ptygmatic veining. (b) Strongly foliated pyroxene monzonite. (c) Concordant unveined basic sheet. (d) Banded gneiss with isoclinal Fa folds. (e) Basic sheets with internal banding and veining. Svenningen. (f) Discordant unveined basics cutting acidic gneisses.

the veined basics and the plagioclase is andesine. In rare cases these sheets are discordant to the migmatic banding in the veined gneisses (Fig. 2f). These are

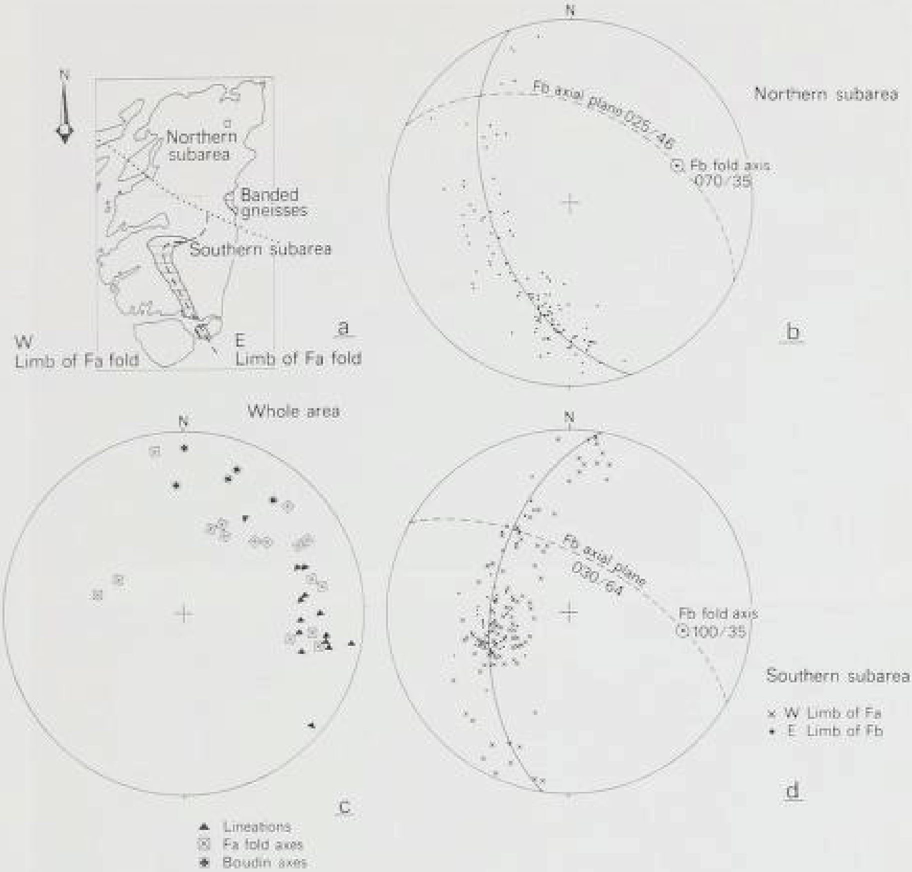


Fig. 3. Stereograms showing structural data, Gjerøy. (a) Location of subareas. (b) & (d) Equal-area plots of gneissic banding in the northern and southern subareas respectively. (c) Plot of other structures from the island as a whole.

interpreted as dykes intruded into the acid gneisses after migmatization of the latter. These rocks are finer grained than the basics mentioned earlier and lack pyroxene and microcline.

STRUCTURES

The configuration of the various geological units of the map is the result of the superimposition of a set of folds (termed Fb folds) with SE-trending axial surfaces upon a set of tighter folds (called Fa folds) whose axial surfaces are consequently of variable orientation.

Fa folding: Large-scale tight folds have produced a piling-up and interfolding of the lithological units. Although the axial planes of these folds have later been folded, they show a sheet dip in the direction of their axes. The geometry of the folds as seen in true profile can be appreciated by viewing the map (Fig. 1) down the plunge of the folds (eastwards with an angle of plunge of

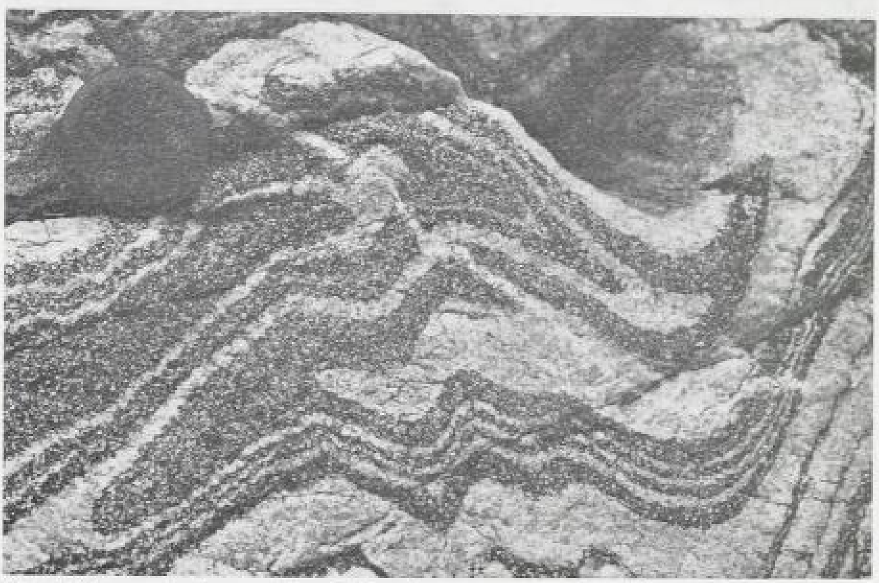


Fig. 4. Mesoscopic Ramsay Type 3 interference patterns (banded gneisses, Svenningen) comparable to the large-scale pattern of Fig. 1. The gentle to open Fb folds have steep axial planes and refold the tight to isoclinal recumbent Fa folds.

30°). On an outcrop scale these are tight to isoclinal folds with axial surfaces subparallel to the dominant gneissic banding (Fig. 2d, 2f). Equal-area plots of gneissic banding (Fig. 3a, 3b) therefore also reflect the variability of Fa axial surface orientations. This folding postdates the intrusions of the unveined basics as the latter show Fa folding (Fig. 2f) and it is possible that some folds in the acid gneisses (e.g. Fig. 2f) could be older than the Fa folds. A mineral foliation (and lineation) invariably seen parallel to the gneissic banding is assigned to a deformation which predates the Fa folding, whereas the foliation developed in the unveined basics is attributed solely to Fa deformation. Boudinage of these basic bodies is also assigned to the later deformation.

Fb folding: This folding is mesoscopically developed almost exclusively in the banded gneiss/metasediment lithologies (Fig. 2f). It deforms the lithological banding and mineral foliation without the development of a new axial plane foliation. The Fb axes and their axial planes vary considerably (cf. Fig. 3a, 3b) while those of the Fa folds (Figs. 1 and 3) vary with them and within the same range, indicating that their interference is of a coaxial type (Ramsay Type 3). In the banded gneisses, a rock type apparently suited for the development of mesoscopic folds, small-scale interference patterns occur equivalent to those developed on a large scale (Fig. 4a and 4b).

GEOLOGICAL HISTORY

The unveined basics have proved useful for distinguishing metamorphic events on this island. These sheets are discordant to a migmatitic banding in the granitic gneisses and these gneisses, which are muscovite-free, are considered to have attained at least amphibolite facies before the intrusion of the basic sheets. The metamorphism of the sheets themselves is also at amphibolite facies; a second metamorphism separated from the first by dyke intrusion. The second metamorphism would have to be vewed as simultaneous with Fa folding if the foliation in the dykes were regarded as associated with Fa folding itself. A third metamorphic event of low-grade character is recorded in the pyroxene–monzonite with the growth of stilpnomelane and hastingsite.

The age of the metasediments/banded gneisses poses a problem. The rocks are clearly older than the Fa folding but their relationship to the unveined basics is unclear. The lack of quartzo–feldspathic veining and K-feldspar in these rocks may be an indication of their age (i.e. younger than the migmatization) but the occurrence of sillimanite (one thin-section possesses two generations of this mineral) could be related to either of the first two metamorphic events. The age of the pyroxene–monzonite is similarly problematical. The foliation in this rock, seen in one instance to be axial planar to Fb folds (Fig. 2b), developed during the lower grade metamorphism. Complementary structural evidence of the age of the pyroxene–monzonite comes from the fact that the sheet itself is folded by Fb folds (Fig. 1), and the latter post-date the higher grade metamorphism as they fold the dominant foliation is most rock-types which is defined by amphibolite facies minerals.

These deductions about the relative age of structural, metamorphic, igneous and sedimentation events are set out in Fig. 5 where tie-lines are used to join two events whose relative ages are known.

REGIONAL COMPARISONS

Vreeken (1979), working on the Nordvernes area on the mainland due east of Gjerøy, argues for a Precambrian age for similar rocks which include meta-

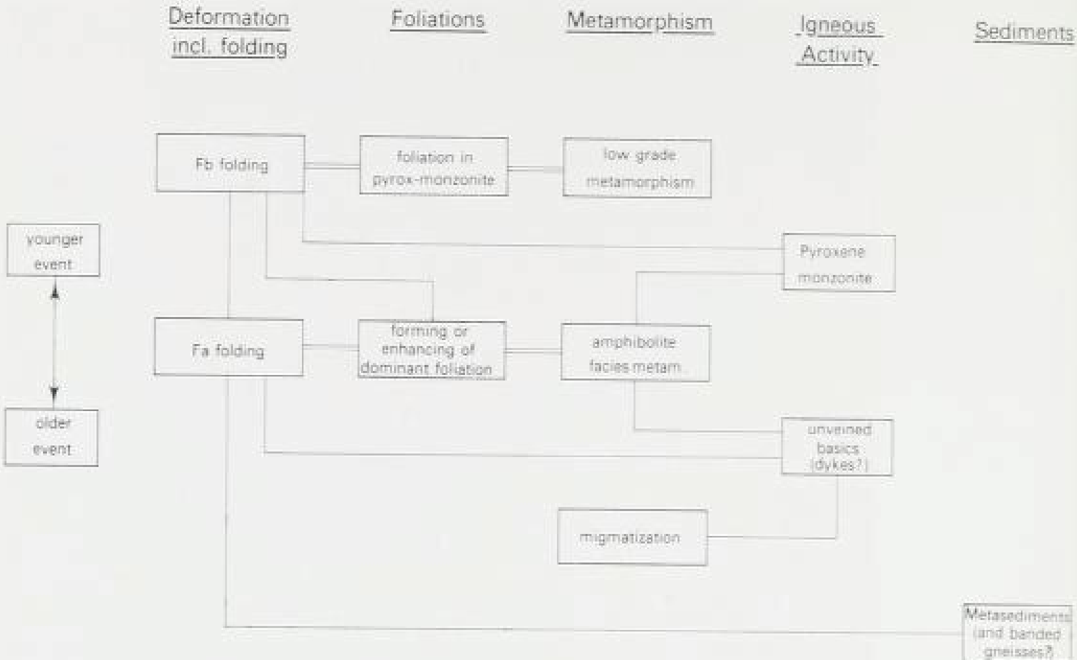


Fig. 5. Geological event network for Gjerøy. Pairs of geological events whose age relationships have been observed are linked by tie-lines. The younger member of such a pair occurs higher in the network than the older member of the pair. Events linked by a double tie-line are equivalent in age. Pairs of events not directly linked by tie-lines are events whose relative ages could not be determined by direct observation. In some cases however, the relative age of a pair of events can be inferred if events are indirectly linked by tie-lines. The relative height in the network of such events cannot be used as an indication of relative age unless the path defined by the tie-lines linking them are monotonic (always increase or decrease in age). Paths involving minima or maxima link events whose relative ages cannot be deduced.

sediments on the basis of the associated ultrabasic bodies. He does this by comparing the characters of these bodies with descriptions of such bodies from Precambrian and Palaeozoic rocks given by Moore & Qvale (1977). As similar bodies do not occur on Gjerøy, similar criteria cannot be applied. However, rocks of the Meløy Group, which according to Rutland & Nicholson (1965) are of Lower Palaeozoic age, are shown on the map of these authors to occur on the south end of Rødøy (the island immediately north of Gjerøy). The Gjerøy banded gneisses and metasediments differ from Meløy Group descriptions in being less psammitic and lacking calcareous lithologies. The possibility of a Precambrian age for the Gjerøy metasediments thus remains open.

Acknowledgements. – I thank Prof. P. F. Williams for introducing me to the rocks of the Svartisen area and for reading the manuscript; Ir. G. A. E. Hermans and Dr. A. Senior kindly assisted with the interpretation of some of the thin-sections. Mr. and Mrs. Rolf Hagland are thanked for their warm hospitality during my stay on Gjerøy.

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Lithostratigraphy of the Storvann Group, East Hinnøy, North Norway and its Regional Implications

JOHN M. BARTLEY

Bartley, John M. 1981: Lithostratigraphy of the Storvann Group, east Hinnøy, North Norway, and its regional implications. *Norges geol. Unders. 370*, 11–24.

The Storvann Group is a sequence of metasedimentary rocks which constitutes the autochthonous Cambrian(?) sedimentary cover of the Precambrian Lofoten terrane on east Hinnøy, North Norway. The Storvann Group is composed of impure quartzite overlain by progressively more aluminous quartz-rich schists, with subordinate calcite marbles, comprising a transgressive sequence. The general stratigraphic similarity of the Storvann Group to autochthonous, unmetamorphosed sedimentary rocks of the Scandinavian platform supports a recent hypothesis that the Lofoten terrane was part of the Baltoscandian continental block prior to Caledonian orogenesis.

The composition of the Storvann Group suggests a platformal or miogeoclinal environment of deposition. However, without some means of determining the original thickness and stratigraphic continuity of this sequence, the position it represents relative to the pre-Caledonian Baltoscandian continental edge remains unknown.

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Introduction

The island of Hinnøy is located in North Norway at 68° to 69°N latitude (Fig. 1). The author's objective on Hinnøy was to examine the contact relationships and structural geology of the boundary between the Lofoten–Vesterålen Precambrian terrane to the west (hereafter called the Lofoten terrane) studied by Griffin et al. (1978), and the Caledonian nappe terrane lying to the east (Gustavson 1966, 1972, Binns 1978). Three tectonically distinct rock associations were recognized: (1) pre-Caledonian crystalline basement rocks, which are considered to be an eastward extension of the Lofoten terrane; (2) the Storvann Group, a sequence of metasedimentary rocks in depositional or modified depositional contact with the pre-Caledonian basement, and which is believed to be its Cambrian or Vendian sedimentary cover; and (3) Caledonian allochthons, which in the study area include at least three distinct associations of metasedimentary and lesser meta-igneous rocks. This report focuses on the lithostratigraphy and significance of the Storvann Group.

The Caledonian structural position of the Lofoten terrane, and the pre-Caledonian relationship to the Baltoscandian craton, have been uncertain. In general, the exposures of Precambrian gneiss along the west coast of Norway have been interpreted to represent pre-Caledonian Baltoscandian basement under-

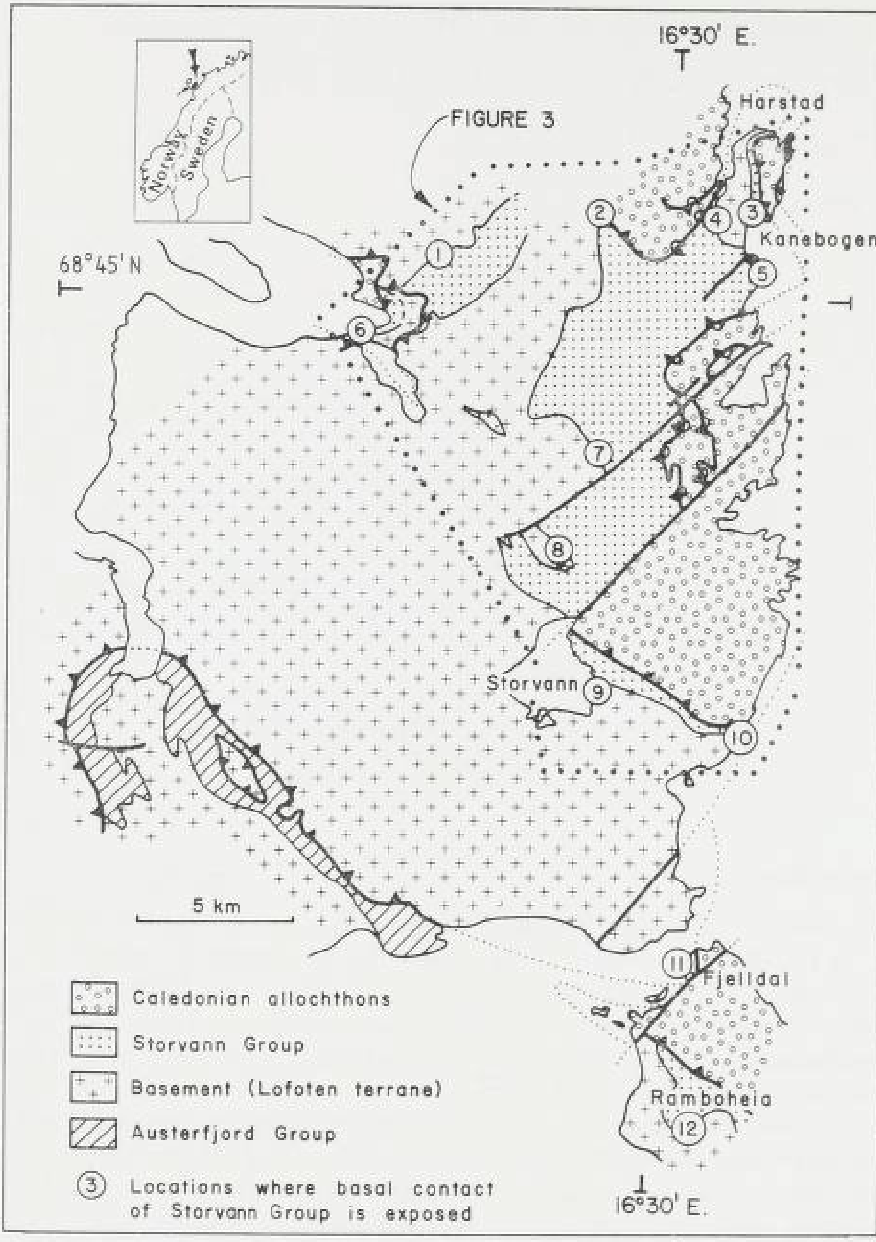

Fig. 1. Simplified geologic map of East Hinnøy, North Norway, showing distribution of Storvann Group and locations of exposed basal contact.

lying the nappes (Wilson & Nicholson 1973, Gee 1975, Binns 1978). However, Brueckner (1971) and Griffin et al. (1978) inferred a shallow structural level for the Lofoten terrane on the basis of a lack of petrological and geochronological evidence for Caledonian metamorphism or deformation. Tull (1977) and Hakkinen (1977) presented structural evidence for the absence of Caledonian penetrative deformation in most of the Lofoten terrane (but cf. Griffin & Taylor 1978). To explain the lack of Caledonian effects, Hakkinen (1977) proposed that the Lofoten block was a far-travelled, high-level nappe. Griffin et al. (1978) considered the Lofoten terrane to underlie the nappes, but nonetheless to have remained at a shallow crustal level (3–9 km).

A shallow structural level for the Lofoten terrane is precluded by the fact that the (meta)sedimentary cover of the block experienced amphibolite facies (kyanite grade) metamorphism synchronous with emplacement of the Caledonian nappes. The lack of Caledonian effects in the Lofoten basement terrane is considered in a separate paper (Bartley, in prep., see also Bartley 1979). The lithostratigraphic similarities of the Storvann Group to the autochthonous Cambrian cover of the Caledonian foreland to the east (Vogt 1967) support the hypothesis of Griffin et al. (1978) that the Lofoten terrane was a part of the Baltoscandian craton in pre-Caledonian time. This also favors a position beneath the nappes for the Lofoten block, consistent with structural observations on and near Hinnøy (Bartley 1980, in prep.).

The Storvann Group consists of a sequence of kyanite- to sillimanite-grade metasedimentary rocks of mainly quartz-rich terrigenous protoliths. It is exposed extensively in the eastern half of the study area from the southern outskirts of Harstad to Storvann, and in the extreme southeast of the study area near Fjelldal and on Ramboheia (Fig. 1). Outcrops along the eastern shore of Storvann are proposed as a type section (location 9, Fig. 1), where a relatively complete and coherent (though still strongly deformed) sequence of rock is exposed, from the basal contact with pre-Caledonian basement to the thrust-truncated top (Fig. 2). Another excellent set of exposures, tectonically thinned but essentially complete, is present along the shoreline and in road-cuts east of Kanebogen to the Harstad NAF campground (location 3, Fig. 1). The name Storvann Group is here considered as an informal designation. It has been divided into 5 unnamed formations (see below).

Lithostratigraphy

CONTACT RELATIONSHIPS

The basal contact of the Storvann Group is exposed at several places (Fig. 1). In all cases, the compositional layering of the basal quartzite is parallel to the contact. Although the rocks are tectonized, there is no suggestion of concentrated strain at the contact. On the map scale, lithologic units of the Storvann Group carry through parallel to the basal contact. Three of the exposures of the basal contact (2, 11, and 12) are anomalous in that little or none of the basal quartzite unit is present, and the quartz–garnet schist (Fig. 2 and below) is in direct contact with the basement or separated from it by only 10 centimetres to a metre of quartzite. At these places, it is suggested that the contact may have been sheared early during the penetrative deformation and thrusting so that no mylonitic fabric was preserved, or perhaps mylonites were never formed. Rocks at the base of the Storvann Group would have been removed

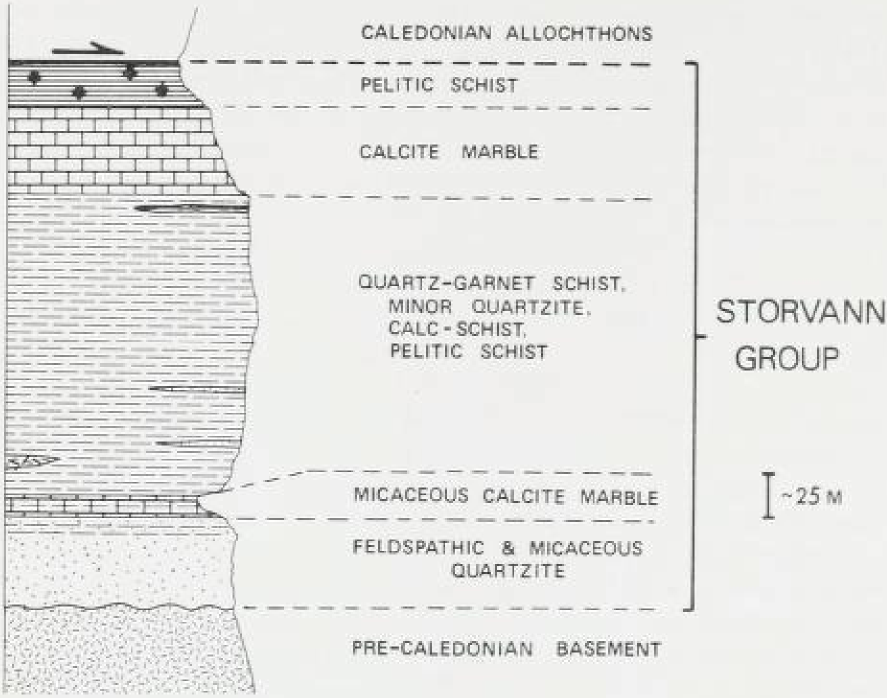

Fig. 2. Stratigraphic column for the Storvann Group. Absolute thicknesses, and to a lesser extent relative thicknesses, of the lithologic units are highly variable due to complex strain patterns. In this figure, approximate typical thicknesses are indicated.

in the process; these contacts are thus considered tectonic slides in the sense of Fleuty (1964). Strictly speaking, it might be preferable to consider the Storvann Group rocks at these places, at Ramboheia (location 12) in particular, as parautochthonous rather than autochthonous.

The Storvann Group constitutes a transgressive sequence. At the base is an impure quartzite which is overlain by progressively more aluminous (though still quartz-rich) rocks, interrupted by two marble horizons. This compositional change is inferred to reflect fining upward of the terrigenous clastic input to the sedimentary basin, consistent with a transgressive setting. How much of the sequence has actually been preserved is unclear, because the stratigraphically upper contact is everywhere a thrust fault.

The combination of the above relations, that is, (1) lack of discordance of lithological units at the map and outcrop scales, (2) consistency of rock sequence above the pre-Caledonian basement, (3) lack of evidence of concentrated strain at the basal contact, and (4) the transgressive nature of the sequence, is taken to indicate that the basal contact of the Storvann Group is an unconformity.

The upper contact of the Storvann Group is always marked by a thrust fault, but this is not necessarily the same thrust everywhere. On the Stangnes peninsula (Fig. 3), the Storvann Group is overlain by amphibolite with an intervening zone of complex tectonic mixing of schist, marble, and amphibolite. On the east shore of Storvann, as entirely different assemblage of rock overlies the Storvann Group. Above the highest schist unit (Pelitic Schist below) of the Storvann Group is an assemblage of mixed lithologies similar to that on the Stangnes peninsula, which is in turn overlain by a slice of Narvik Group (Strand 1960) pelitic gneiss and schist. The Narvik Group slice is overlain by gray calcite marble of the Salangen Group (Gustavson 1966). The significance of these relationships for the geometry of the nappe stack will be considered in a separate paper (Bartley in prep., see also Bartley 1980).

LITHOLOGIES

The distribution of Storvann Group lithologies mapped on east Hinnøy is shown in Fig. 3. The structural complexity of the rocks is striking. However, despite polyphase folding and high-angle faults, a consistent sequence of five formations is recognizable away from the contact with the Precambrian basement.

1. Quartzite

The basal unit of the Storvann Group is a heterogeneous mixture of micaceous quartzite, vitreous quartzite, and quartz–feldspar–biotite schist, with local occurrences of quartz–biotite schist and garnet–mica schist. A separate upper member consisting of the latter lithologies was mapped separately in one area (center of Fig. 3).

The vitreous quartzite is usually present as compositional bands 2–10 centimetres thick, separated by more micaceous layers 1–20 cm thick. The vitreous layers are fine-grained and range in color from white to bluish gray and rarely dark gray. The miceceous quartzite is similar, but contains mica in discontinuous films and as disseminated grains defining the schistosity. The compositional banding on a mesoscopic scale has been transposed by isoclinal folding and in general has little primary significance.

Rocks with higher feldspar contents than the vitreous quartzite are present in this unit, especially in its lower parts, on the north side of Finnslettheia in particular (Fig. 3). It is locally difficult to locate the basement/cover contact here because of the compositional similarity of the feldspathic cover (here a meta-arkose) to the granitic basement. In such cases, the contact was drawn on the basis of: (1) the appearance of vitreous or micaceous quartzite interlayers, and (2) the better layering and finer grain-size of the metasedimentary rocks (in most cases). In the northwestern part of the map area, the basement is dominated by metasedimentary rocks and the identification of the basal contact can become difficult. However, in general the basal contact can be traced into areas where the lithologies are more distinctive and the contact more easily located, so that its geometry can be determined with confidence.

In thin-section, the feldspathic quartzites contain both microcline and plagioclase in subequal amounts, forming from a few percent up to perhaps 25% of the rock. In general, microcline is dominant in less feldspathic rocks, while plagioclase is more important in the meta-arkoses. A change from muscovite to

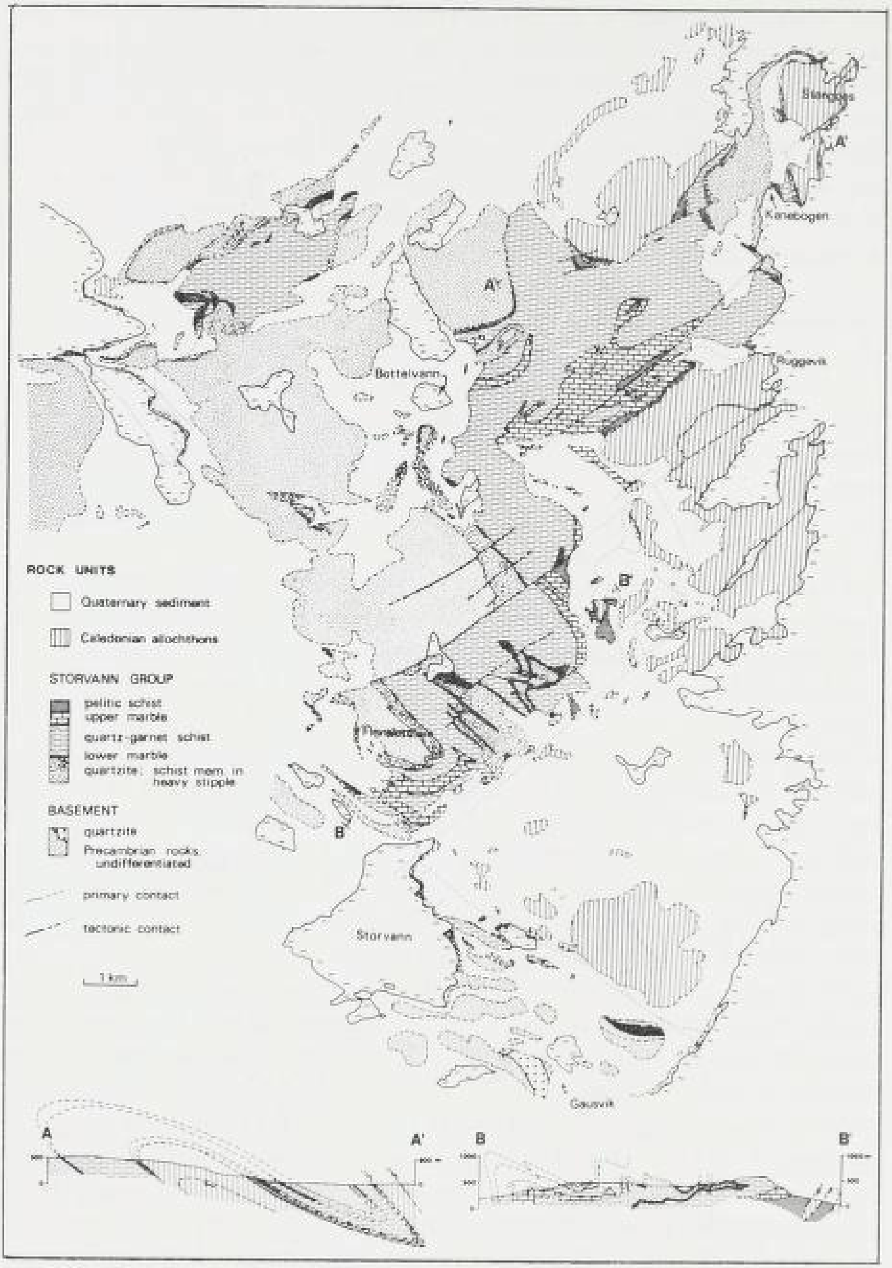

Fig. 3. Distribution of Storvann Group lithologies on East Hinnøy, with illustrative geologic cross-sections.

biotite as the dominant mica also accompanies the transition from less to more feldspathic compositions. The micas, though concentrated along compositional bands, are commonly individually rotated into an orientation parallel to axial planes of late folds. The rotation of micas and the resulting intersection lineation are commonly present even in outcrops which show no mesoscopic late folding.

The basal 2–20 metres of the quartzite unit usually consist of vitreous quartzite or meta-arkose, but on the east shore of Storvann and at the north end of the Stangnes peninsula, about 5 m of quartz–biotite schist with 5 mm feldspar porphyroblasts is present at the base. This is considered to be a local facies of the basal Storvann Group. It is equally possible that these are lenses of Precambrian metasedimentary rock, preserved locally below the basal unconformity of the Storvann Group. Volumetrically, these rocks represent a minor element of the sequence, so the resolution of this question is of modest importance.

A layer of quartz–biotite–garnet schist is locally present at the top of the quartzite unit. On the east shore of Storvann, about 5 m of a schist very similar to the quartz–garnet schist unit (see below) is present at the top of the quartzite. Further north, 50 m or more of this unit occurs as part of a separate, upper member of the quartzite unit (Fig. 3). This lithology is in a sense transitional to the quartz–garnet schist present higher in the sequence, although generally separated from it by the lower marble unit.

The other lithology present in the upper member of the quartzite unit is a quartz–biotite schist. This dark brown or gray, fissile schist is an atypical rock type in the Storvann Group, but clearly occurs stratigraphically above vitreous and feldspathic quartzite more typical of the quartzite unit and below the lower calcite marble. It is thus considered a member of the basal quartzite formation.

The thickness of the quartzite formation varies, probably mainly due to variations in finite strains, from zero to about 250 m northwest of Sørvikfjell. A typical thickness for the unit would be about 50 m. The possibility that some of the thickness variations in the quartzite, including its absence in some areas, may be due to facies variations (for instance, due to infilling of topography) cannot be ruled out.

The protolith of the quartzite unit was of variable composition, but was generally a moderately mature sandstone with subordinate shaly facies. The absence of a basal conglomerate is notable, and unfortunate because the presence of locally-derived pebbles would reinforce the interpretation that the basal contact is an unconformity. It is important that the Storvann Group basal quartzite is readily distinguished from quartzites of the basement complex, on the basis of lithology and associated rock types, as well as structural position.

2. Lower calcite marble

Above the basal quartzite unit in many areas is a thin (5–10 m) gray calcite marble. It is commonly impure, containing several percent white mica, minor quartz, and occasional accessory pyrite and/or graphite. Compositional banding is defined by varying concentrations of mica and by a subtle shading of color from light to medium gray. Intrafolial isoclinal fold hinges are moderately common, indicating transposition of bedding by folding to produce the present compositional banding. Texturally, the calcite is granoblastic, with lepidoblastic mica films. Quartz occurs as disseminated equant grains. Calcite is generally intensely twinned, indicating late to post-metamorphic strain.

A somewhat different facies of this unit is present on the west flank of Finnslettheia. Outcrop is moderate to poor, but normal marble appears to pass laterally into a calc-silicate rock composed of tremolite, calcite, plagioclase, quartz, epidote, and sphene. Phlogopite occurs in trace amounts in one sample.

The lower marble formation is not present everywhere. This is at least in part due to tectonic slides between the quartzite and quartz–garnet schist units. In the inverted section on top of Finnslettheia (Fig. 3), only three lenses of this marble occur, while the intervening areas where the marble is absent commonly show more intense foliaton. This could result from either large-scale boudinage of the marble or tectonic sliding. From what is known of the relative ductility of calcite- and quartz-rich rocks at amphibolite facies conditions, it seems unlikely that marble would be the boudin-forming lithology. Hence, a tectonic sliding interpretation is preferred.

The protolith of the calcite marble was presumably a shallow-water biogenic limestone with some terrigenous component. The reducing conditions which produced the pyrite and graphite may have resulted from incomplete oxidation of organic carbon. The calc-silicate facies was in part dolomitic (as evidence by Mg-bearing phases such as tremolte), and bears no mineral indicating low f_{O_2}. It is possible that local secondary dolomitization occurred under oxidizing conditions to form the protolith of the calc-silicate facies, at the same time oxidizing all remaining carbon in the rocks.

3. Quartz–garnet schist

The quartz–garnet schist is the areally most extensive unit of the Storvann Group, and one of the most distinctive lithologies. In outcrop it is a massive, resistant light gray rock with foliation defined by lensoidal or sigmoidal clots of mica one centimetre or less in length. Quartz veins a few mm to a few cm thick, and several cm to several tens of cm long, are common. Easily confused with, but distinct from, these veins are bands of quartzite on the same scale. The quartzite bands are more laterally continuous and contain mm-scale color bands parallel to the schistosity.

Quartz generally constitutes 70 to 80% of the rock, occasionally ranging up to more than 90% (forming essentially a garnetiferous quartzite). Quartz generally shows only weak dimensional preferred orientation, mainly appearing as granoblastic aggregates. Sutured grain boundaries are not uncommon. Undulatory extinction is locally present. The textures suggest an annealing recrystallization after schistosity formation, followed by late to post-metamorphic strain.

Garnet is ubiquitous as small, subhedral porphyroblasts (1–3 mm, occasionally up to 5 mm), composing up to 10% of the rock. The garnets are randomly dispersed throughout the rock. Where late spaced-cleavage planes intersect garnet, rims of chlorite ± epidote are developed. Although the garnet is commonly sieved with equant blebs of quartz, trails of aligned inclusions have not been observed within this unit.

Both biotite and muscovite are present, but in relative minor amounts, together comprising no more than 10–15% of the rock as a rule. Proportions of the two micas vary, biotite usually dominating. Continuous mica-films are seldom present, making the schistosity often abscure in outcrop. A late spaced-cleavage defined by mica concentrations is often more prominent than the earlier foliation. In one specimen, random-oriented, interkinematic biotite porphyroblasts were observed.

Green tourmaline is an ubiquitous accessory phase, occurring as small (<1 mm) euhedral or subhedral grains. Other accessory minerals include zircon, sphene, and apatite.

Three rock types with somewhat different compositions occur locally in this formation. On Finnslettheia, 10 m of plagioclase–porphyroblastic, graphitic schist is interlayered with more typical quartz–garnet schist at the base of the formation. The prominent foliation is marked by lepidoblastic intergrowths of muscovite and graphite, with subordinate biotite. Sparse, albite-twinned plagioclase porphyroblasts are a few mm across, anhedral, and randomly oriented.

A hornblende–porphyroblastic facies of this formation is present near its base in the area 1 km south of Bottelvann (Fig. 3). Hornblende occurs as stubby, randomly-oriented porphyroblasts up to 1 cm across. In the shoreline exposures at Fjelldal (Fig. 1), the stratigraphically higher part of the quartz–garnet schist unit becomes more pelitic, with the minerals kyanite and staurolite appearing. The rocks here are more strongly retrograded in late events, developing much secondary white mica, biotite, and chlorite. Coexistence of kyanite, staurolite, and garnet can be documented petrographically. Due to the intense retrogression, it cannot be demonstrated that biotite coexisted with these other phases, though this seems likely.

The thickness of this unit ranges from about 150 to more than 700 m. To what extent these thicknesses relate to original stratigraphic thickness is not certain, but it is clear that this was and is the thickest formation within the Storvann Group.

The protolith of this unit was probably a chemically mature siltstone with some intermixed sand. The presence of abundant garnet in a relatively mica-poor rock suggests high alumina relative to alkalis, and hence a clay-rich, feldspar-poor protolith. The feldspar- and amphibole-bearing lithologies may reflect a variation in provenance, or a moderate amount of carbonate (as cement?) which reacted with the silicate phases to form Ca-bearing silicates during metamorphism.

4. Upper calcite marble

The upper gray calcite marble formation is generally similar to the lower, but may usually be distinguished by its purer composition and more prominent color banding. A few percent white mica and minor quartz are present, but graphite and sulfides are absent and the rock is generally more than 95%

calcite. While the lower marble unit is generally homogeneous in color, the upper marble is commonly banded gray and buff on a scale of 1–5 cm. Textures are granoblastic; early tectonic foliation can be recognized by the color bands and by sparse micaceous layers.

The thickness of the upper marble is difficult to estimate because the upper contact is generally poorly exposed or tectonic. Only 5 metres of marble are present in the Kanebogen shoreline section (Fig. 3), while its apparent thickness is considerably increased by late folds in most other areas of favorable exposure. Perhaps the most representative thickness is encountered southwestward along strike from Ruggevik, where the marble is about 100 metres thick.

5. Pelitic shrist

This highly garnetiferous schist is the stratigraphically/structurally highest unit assigned to the Storvann Group. It probably constituted a mechanically favorable level for detachment, since it commonly occurs as a thin (10–20 m) layer immediately beneath the lowest allochthonous rocks of the area. In some areas it appears to be involved in the zone for tectonic mixing mentioned above (p. 14–15). The consistent presence of a pelitic schist, often rusty-weathering and bearing coarse garnet porphyroblasts, at the top of the Storvann Group section, regardless of the overlying units, leads the author to believe that this schist forms an integral part of the Storvann Group.

The rock has a well-developed schistosity with the foliation defined by mica films which anastomose and wrap around the subhedral and euhedral garnets. The garnets range in size from 0.5 to 2 cm in diameter. In thin-section, major minerals include quartz, muscovite, biotite, garnet, and often kyanite. One specimen contains fibrolitic sillimanite. Accessory minerals include green tourmaline, zircon, and sphene.

The garnet porphyroblasts include trails of elongate quartz grains, recording a two-stage growth history. The inner portion includes abundant linear trails, suggesting growth post-dating formation of an early schistosity. These inclusion trails continue into the outer zone where inclusions are more sparse, and curve into a sigmoidal pattern, indicating synkinematic growth. Where late spaced-cleavage intersects the garnets, truncating the internal fabrics, retrograde chlorite is developed.

Biotite occurs as two growths. It is present as 1–3 mm, randomly-oriented porphyroblasts which are bent and truncated by the late cleavage, and also in secondary quantities to muscovite within the mica films which define that cleavage.

Muscovite forms both the through-going mica films and also locally remains in the matrix preserving an earlier foliation in short segments. It is not clear how this older matrix foliation relates to the internal foliation of the garnet, because of strong disruption by the late cleavage. However, these older fabrics are assumed to be related.

Kyanite occurs as anhedral relics, replaced in part by white mica, quartz, and chlorite. This retrogression (?) is interpreted to be related to late cleavage-forming events, although it is not as clearly related to cleavage traces as in the case of garnet retrogression.

Sillimanite occurs as fine-grained intergrowths with muscovite in one kyanite-bearing thin-section. Since the kyanite is replaced in part by muscovite, it may be that the kyanite to sillimanite reaction ocurred by two coupled reactions involving muscovite, similar to the relationships described by Carmichael (1969). This is the first occurrence of sillimanite grade Caledonian metamorphism to be described from the Scandinavian Caledonides of southern Troms and northern Nordland.

On the east shore of Storvann, the pelitic schist unit is partly calcareous. Calcite occurs as discontinuous lensoid interlayers a few mm thick, composed of fine-grained granoblastic aggregates. No calc-silicate minerals were observed.

The structural thickness of this unit is highly variably, and in most areas very uncertain. It is very non-resistant to weathering so that exposures tend to be rare except along shorelines. Areas of the map showing apparently large thicknesses of the pelitic schist unit are largely the result of generalization from a few outcrops which were impossible to represent individually on the scale of the study. It is probable that in these wooded areas of poor exposure, the structure is far more complex than that shown in the map. However, the data do not permit a more sophisticated interpretation.

The protolith of the pelitic schist unit was a shale, deposited in part under reducing conditions, and in part in environments of some carbonate deposition. It appears to be an expression of continued transgression of the Scandinavian continental margin in early Paleozoic time.

Correlation

The impure, often feldspathic composition of the basal quartzite, the dominantly terrigenous character of the section, and the gross sequence of lithologies encourage correlation with the autochthonous Vendian/Cambrian rocks of the foreland of the Caledonian mountain belt (Dividal Group of Pettersen 1878, see Fig. 4). At this latitude, the autochthonous sequence has been described by Moberg (1908) and Vogt (1918, 1967). Gustavson (1966) has described autochthonous sequences from tectonic windows through the nappe stack. The sections are all largely sandstone and shale, overall fining upwards, with thin calcareous horizons (Fig. 4). Vendian and early Cambrian faunas are reported from several of the foreland sections (Moberg 1908, Vogt 1967, Ahlberg & Bergström 1978, Ahlberg 1979, 1980). Gustavson's section from the Dividalen window is similarly dominated by sandstone and shale. Other sections from windows are too incomplete for reasonable comparison, and in some cases (e.g., Rombak window) are structurally dismembered (K. V. Hodges, pers. comm. 1980).

Two correlation schemes seem possible: (1) the two marble formations of the Storvann Group correlate with the two calcareous horizons of the foreland sequence, with the quartz–garnet schist equivalent to the green shale; or (2)

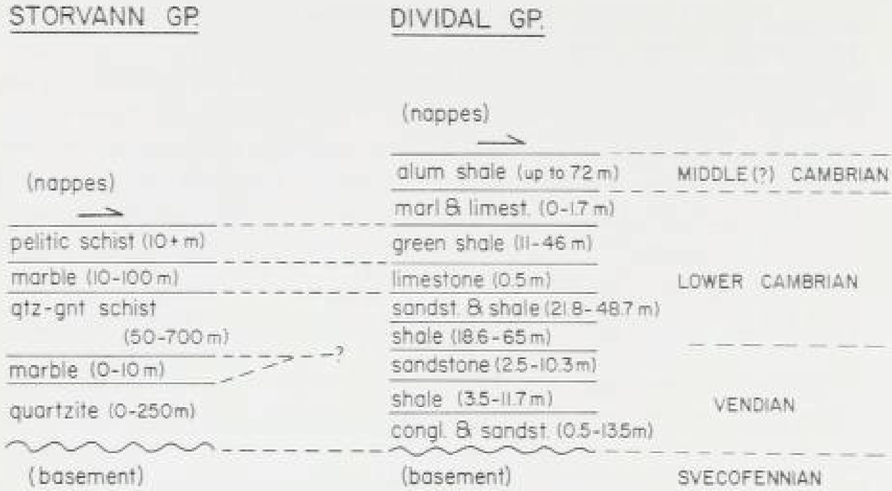

Fig. 4. Comparison and speculative correlations of formations of the Storvann Group to the autochthonous Vendian/Cambrian formations (Dividal Group) of the Caledonian foreland in Norway. Foreland section generalized from Vogt (1967) and Ahlberg (1980).

the upper marble formation of the Storvann Group correlates with the lower limestone horizon of the foreland section, and the quartzite, lower marble, and quartz–garnet schist formations are together equivalent to the basal sandstone and shale formations of the foreland. The latter is preferred for the following reasons: (1) the presence of metasandstone layers in the quartz–garnet schist seems incompatible with the lithology of the green shale, which Vogt (1967) describes as a pure shale, containing no sandstone; (2) the thinness and discontinuity of the lower marble of the Storvann Group suggest that no equivalent need appear in other sections; and (3) the thickness and purity of the upper marble appears inconsistent with correlation to a discontinuous thin, marly horizon.

The correlations shown on Fig. 4 are admittedly speculative, especially considering (1) geographical separation (more than 100 km), and (2) the difference in metamorphic grade (unmetamorphosed versus amphibolite facies). Nevertheless, the author considers a correlation of the Storvann Group with the autochthonous sequence of the Caledonian foreland very likely. This is consistent with the view of Griffin et al. (1978) that the Lofoten terrane was indeed part of the Baltoscandian craton in pre-Caledonian time. It is also consistent with structural arguments (Bartley 1979, in prep.) that the Lofoten block does not constitute a far-travelled nappe (cf. Tull 1977, Hakkinen 1977).

Discussion

The Caledonian cover rocks of east Hinnoy were described by Gustavson (1972) as a single stratigraphic succession. The present study concludes that these cover rocks include elements of at least four distinct stratigraphic sequences, including both autochthonous and allochthonous units. The Storvann Group as defined here is the autochthonous portion of the cover rocks. Allochthonous rocks include parts of: (1) the Narvik Group (Strand 1960, Gustavson 1966), consisting mainly of pelitic schists and gneisses with minor intrusions of granitoids and amphibolites, and in the study area present only in two tectonic slivers at the base of the nappe pile; (2) the Stangnes Group, a newly-distinguished, thrust-bounded slab of layered amphibolite intruded by a semi-concordant tonalite gneiss pluton, the Ruggevik Tonalite Gneiss (these units will be described in detail in a future paper, pending completion of work in progress; see also Bartley 1980); and (3) the Salangen Group (Gustavson 1966), a sequence of marbles and mica schists which comprise the highest unit of the nappe stack exposed on east Hinnøy. All of the above have been metamorphosed at amphibolite facies conditions during the Caledonian orogeny. Juxtaposition of the allochthons and the autochthon occurred prior or to synchronous with metamorphism, so that no abrupt breaks in metamorphic grade are observed (Bartley 1980, and in prep).

The lithostratigraphy of Hinnøy is thus far more complex than previously recognized. Furthermore, the rocks considered autochthonous in this study are not those considered to be autochthonous cover by Gustavson (1972). One of the quartzite units considered autochthonous by Gustavson (west of Gausvik, see Fig. 3) has proven to be intruded by a granite which has given a Rb/Sr whole-rock intrusive age of 1559 ± 155 Ma (Bartley 1980), indicating that this quartzite is different from the basal quartzite of the Storvann Group and lies within the Precambrian basement. This is consistent with the observation that this Precambrian quartzite is associated with amphibolite and calc-silicate marble rather than the schists and calcite marbles of the Storvann Group. Further, this eliminates the principal evidence which Gustavson (1972, p. 12) used to infer that the rocks here termed the Storvann Group, and their subjacent basement, are allochthonous. The structural repetition of lithologic units north from Gausvik reported by Gustavson (1972) has not been confirmed during the present study. A full discussion of the structural geology of this area, and an account of the evidence bearing on the differences in mapping and structural interpretation shown by the present investigation in relation to Gustavson's work, will be presented in a subsequent paper (Bartley, in prep.).

The composition of the Storvann Group metasedimentary sequence, i.e., quartzose terrigenous clastic rocks with laterally continuous carbonate horizons, implies a platformal or miogeoclinal depositional environment. However, without fossil control or some other means of estimating the original thickness and stratigraphic continuity of the Storvann Group, one cannot distinguish between these two possibilities. As a consequence, no inferences can be made about the position of east Hinnøy relative to the pre-Caledonian continental edge. One can only say that the edge of the continent was located somewhere to the west of the modern limits of the Lofoten terrane, roughly 60 km northwest of the present study area.

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The Northern Kongsberg Series and its Western Margin

I. C. STARMER

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In the Kongsberg Series a major fold interference pattern was established during the Svecofennian orogeny before westward thrusting over the Telemark block developed a margin of ultramylonite (preserved from Flesberg to lake Soneren): further north (to Ådal) this was cut out by a later mylonite. A subsequent, basic intrusive phase was associated with rifting, which was aborted by the onset of the Sveconorwegian (Grenvillian) Regeneration. Late reactivations of the whole margin caused brittle faulting and developed the so-called 'friction breccia'. The early regional structure and its relics could explain many features of the Kongsberg Series; in particular, the Modum area complex may represent Bamble Series rocks infolded into the Kongsberg Series during the Svecofennian orogeny.

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Introduction

The northern part of the Kongsberg Series (Fig. 1) presents a number of problems related to the subdivision of the southern Fennoscandian Shield. The Geological Map of Norway (Holtedahl & Dons 1960) shows the Kongsberg Series with a north-south lithobanding and a western junction against the Telemark Series, marked by a fault or breccia line. This 'friction breccia' (of Bugge 1928) trends north from Kongsberg (Fig. 1) for about 50 km, bends at Haugesjø and runs northeast for some 60 km, disappearing at Sokna, where the Kongsberg and Telemark Series merge. Smithson (1963) showed it continued north of Sokna, splitting towards the Flå (Ådal) granite, but he commented that 'the rocks are similar on both sides of the fault'. Recently, Hageskov (1979, 1980) has suggested it might continue north of the Flå granite around Bagn. Additional problems are associated with the occurrence, in the centre of the northern Kongsberg Series, of a 'Modum area' subdivision containing very different lithologies.

The present study concentrates on the western part of the Kongsberg Series, north of Haugesjø, and the geology of the Series from its margin eastwards to the Modum area complex. It covers a tract north of the Flesberg area of Bugge (1937) and west of the Modum district of Jøsang (1966). Summary results of a regional reconnaissance and tectonic synthesis (Fig. 1) cover a wide area including the Modum district and the tract northwest of Tyrifjorden studied by Hofseth (1942): the southern margin was included on the 1:250,000 map sheet 'Skien' (Dons & Jorde 1978) and the northern margin was covered by Smithson (1963) in his study of the Flå granite (comprising the Ådal granite and a more northerly Hedal granite).

Cataclastic rocks are described using the lithological nomenclature of Higgins

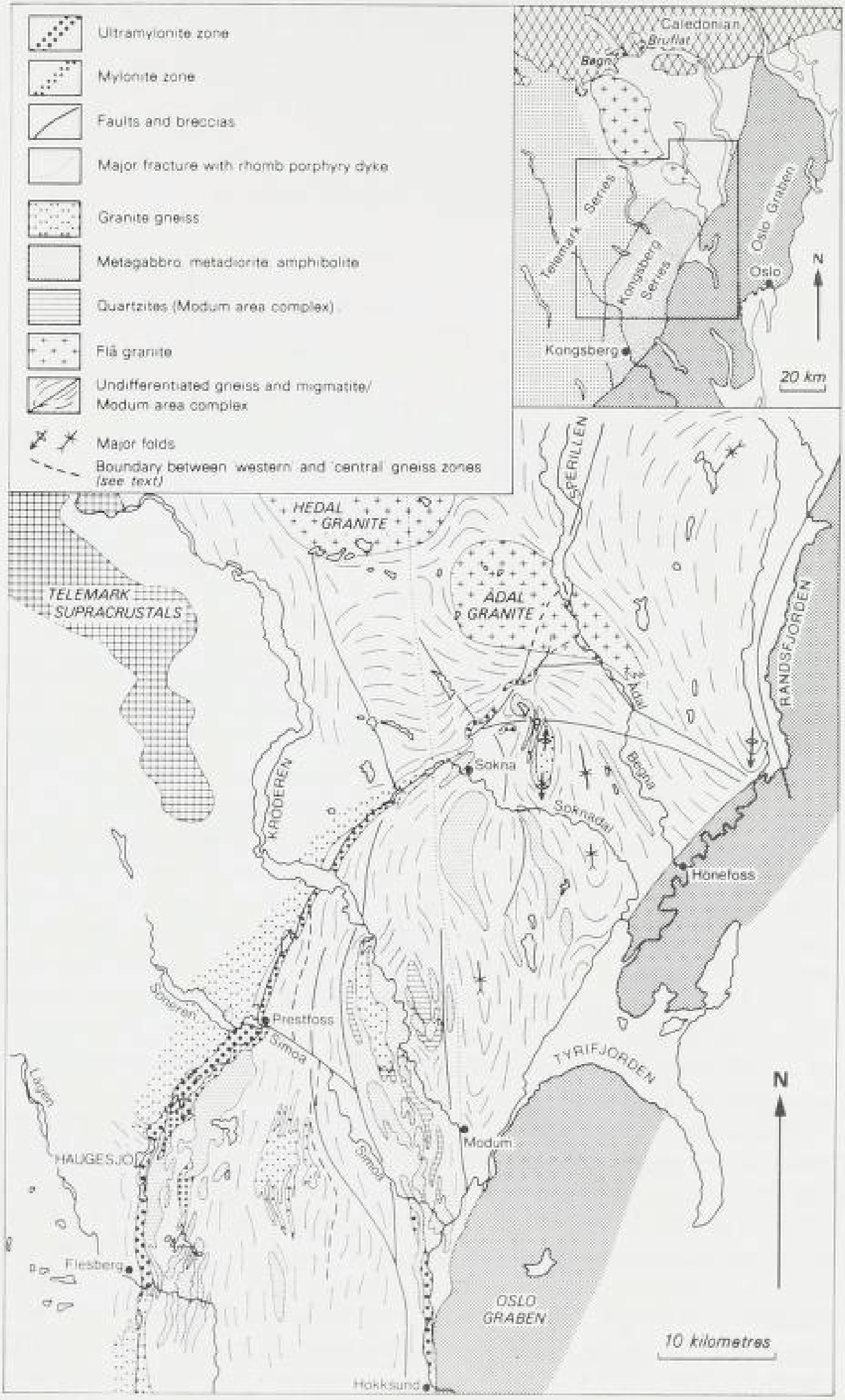

Fig. 1. General geology of the region. The boundaries of the Ådal granite are taken from Smithson (1963) and the geology of the Modum area from Jøsang (1966).

(1971). In terms of the nomenclature of Zeck (1974) they show some myloblastic recrystallisation (synchronous with cataclasis) and dominant blastomylonitic recrystallisation (after cataclasis).

Regional geology of the northern Kongsberg Series

A north-south lithobanding and foliation in the west results from cataclasis. Eastwards, a few remnants of early fold closures are preserved, although the tectonic pattern becomes less ordered around the later basic intrusives and Flå granites. Several closures, apparent on a regional scale (Fig. 1), are seemingly remnants of a 'closed' interference pattern, broadly conforming to the 'dome and basin' type. The structures are difficult to distinguish, due to original complexities, steep foliations, transposition and later folding. Smaller major structures are sporadically preserved (e.g. the granitic dome, east of Sokna, Figs. 1, 4 & 5), as are related minor folds.

The complexity of the elongate domes and basins was partly due to their development from the interference of three fold phases. They were elongated with steep N to NNW striking axial planes (with some axial planar foliation and transposition at their hinges). Associated tight to subisoclinal minor folds showed some transposition and syntectonic garnet growths. They refolded an earlier foliation with a few minor isoclines, which were often intrafolial. A third fold phase produced open to tight, minor flexures, with steep ENE to NE axial planes. All these deformations interacted to produce a major interference pattern (referred to as 'the early regional structure' in the present study) and developed prior to the intrusion of gabbros and the cataclasis associated with the western mylonite zones. The interference patterns were, therefore, much older than the late concentric folding, which was dominant in the Telemark Series and in the western Kongsberg Series.

LITHOLOGICAL DISTINCTIONS BETWEEN MAJOR STRUCTURES

There is some lithological distinction between rocks of the individual major structures. West from Randsfjorden to the Flå granites (Fig. 1), the rocks are heavily migmatised gneisses with amphibolites and some protomylonites. They generally have shallower dips than the Kongsberg Series to the southwest, and are severely affected by open flexuring (and some doming) related to the late concentric folding. One shallow, major synform west of Randsfjorden (in the northeast of Fig. 1) could reflect this later folding.

The southward synformal closure at the north end of Tyrifjorden (and its continuation through Soknadal) occurs in plagioclase–quartz–biotite gneisses and schists ($\pm$ garnet and muscovite) with some biotitic, graphitic and sulphidic schists, thin quartzites and thin amphibolites. In the west of this structure, about 10 km west of Hønefoss and just south of Soknadal, Hofseth (1942) recorded the development of sillimanite gneisses. The other structures closing along the northwest of Tyrifjorden (Fig. 1) occur in a mixed series of garnetiferous biotite–plagioclase–quartz gneisses, hornblende-rich gneisses and abundant amphibolites.

To the west, the location of the Modum complex correlates with the regional fold pattern and is discussed below. Further west, in the area of the detailed survey (Figs. 2–5) another different series of gneisses and protomylonites (towards the western ultramylonite) have a general north-south banding and have been divided into a 'western' and 'central' zone (Figs. 1 & 2); in addition there is an 'eastern' zone representing the edge of the Modum complex: these zones were probably inherited from the early regional structure.

THE MODUM COMPLEX

The Modum complex, the centre of which was surveyed by Jøsang (1966), is somewhat of an enigma in the Kongsberg Series. It contains quartz–plagioclase–mica gneisses and schists (± cordierite and almandine), sillimanite-rich rocks (± quartz–sillimanite 'nodules') and thick, pure quartzites: it also has later scapolitised gabbros, orthoamphibole–cordierite rocks, large pegmatites and albites. It has been likened to the Bamble Series of the Skagerrak coast (Bugge 1926, Jøsang 1966) and it was recently noted (Starmer 1980) that all these lithologies are typical of the Bamble Series and contrast with those of the surrounding Kongsberg Series. Isolated sillimanite-bearing gneisses were found by Hofseth (1942) in the Kongsberg Series west of Hønefoss and they also occur at Eiker (6 km WSW of Hokksund). A few, thin quartzitic layers occur east of Sokna. However, these are all isolated developments and differ from the constant association of thick quartzites with sillimanite-rich rocks, which is a feature of both the Modum complex and the Bamble Series.

The development of the Modum complex correlates with the lithological changes between major units in the early regional structure. It seems to represent an elongate domal area, modified by transposition and brought up by a combinaton of folding and later faulting. Its western margin (Figs. 1, 2 & 3) is defined by a late fault from Krøderen to Simoa, and further south (to Hokksund) by an earlier mylonite zone along which an upthrust of the Modum complex has occurred (Starmer 1980). Around Modum, Jøsang (1966) identified two southward-closing, major folds (which appear to have N to NNW axial planes, consistent with the regional pattern). Both are depicted on Fig. 1: one closure is in quartzite, north of Modum and the other is a 'syncline' at the southern end of the large granitic mass. Elsewhere and particularly further south (ie. west of Modum), no definite major folds were recognised, possibly due to the transposition effects observed elsewhere in the early regional structures.

In the present study, the detailed survey of the western part of the Kongsberg Series (Figs. 2–5) was extended eastwards to cover part of the Modum complex. The western margin of this complex consists of quartz–plagioclase–biotite gneisses with hornblende and almandine and with layers of biotite/phlogopite schist and thin quartzites: a little further east sillimanite gneisses, sillimanite granites and thicker quartzites occur. Almandine is common in all rocks. A few thin protomylonites and mylonites are developed. In the centre of the complex is the Brennåsen granite gneiss, which is largely homogeneous, but has marginal granitised patches of quartz–plagioclase–biotite gneisses, augen gneisses, sillimanitic and 'nodular' rocks and cataclastites. Thin granitic and granitised layers are developed around the main mass and these extend northwards (to the east of Krøderen village). East of the Brennåsen mass, sillimanite gneisses and schists (commonly reaching 2 mm grain size) frequently contain quartz–sillimanite lensoids (5–15 cm in length and 5 mm in width) often intensively retrogressed to muscovite. These lithologies are identical to the 'nodular rocks' in the Bamble Series of the Skagerrak coast.

The western part of the Kongsberg Series

VARIABLE QUARTZ–PLAGIOCASE–BIOTITE GNEISSES

These gneisses show great modal variations, with hornblende and almandine becoming more abundant eastwards. Textures are granoblastic and lepidoblastic to cataclastic, with grain sizes from 5 mm to < 0.2 mm, respectively. (Rusty 'fahlbånds', common further south and in the Tyrifjorden–Soknadal tract to the east, are conspicuously absent in this area.) Isolated thin bands of protomylonite and mylonite occur throughout the area, becoming more common westwards towards the ultramylonite zone, where several major bands have been distinguished (Fig. 2). Layers of amphibolite occur: some concordant, coarse-grained bodies represent basic layers or early intrusives within the original supracrustals, but thin, finer-grained amphibolites form numerous concordant and discordant bodies and are related to the younger 'Vinor' intrusions.

The north-south striking gneisses undergo gradational, but significant changes eastwards across their strike: on a regional scale, this partly reflects the early regional structure and partly reflects the presence of a series of thrust slices, passing upwards to the east. Between Haugesjø and Krøderen (Fig. 2) the changes are sufficiently marked to separate the gneisses into a 'western', a 'central' and an 'eastern' (Modum complex) zone.

The 'western zone' consists of fine- to medium-grained quartz–plagioclase–biotite protomylonites and gneisses. Some layers are biotite-poor or biotite-free and a few contain muscovite or hornblende and/or almandine. One major band of coarser protomylonite and gneiss in the southwest (Fg. 2) contains 1 cm-size plagioclase porphyroclasts and reflects early banding, being cut discordantly by cataclased layers.

The 'central zone' is distinguished by the alternation (on a scale of a few cm) or m) of 'western zone' type gneisses and protomylonites interlayered with coarser variants containing 1 cm-size plagioclase porphyroclasts and augen. Biotite, hornblende and almandine are more abundant than in the western zone. The coarser layers become less common northwards towards Krøderen (Fig. 2) and very sporadic further north (Fig. 4).

The 'eastern zone' of gneisses (Fig. 2) belongs to the Modum complex, considered above. Although its western boundary is partly defined by a late fault, the change to this eastern zone is marked by developments of biotite schists and thin quartzites in the quartz–plagioclase–biotite gneisses.

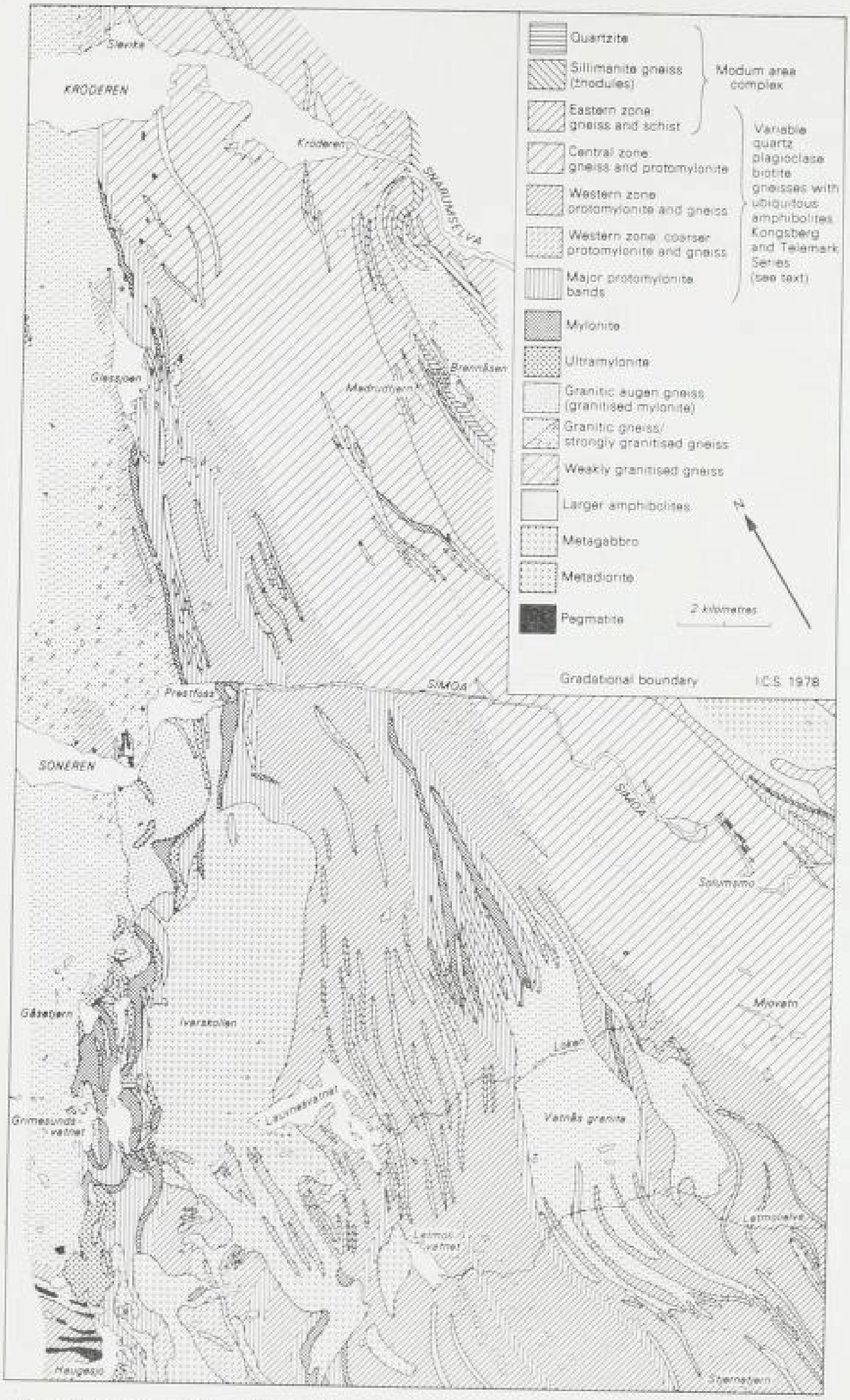

Fig 2. Lithological map of the area from Haugesjø to Krøderen.

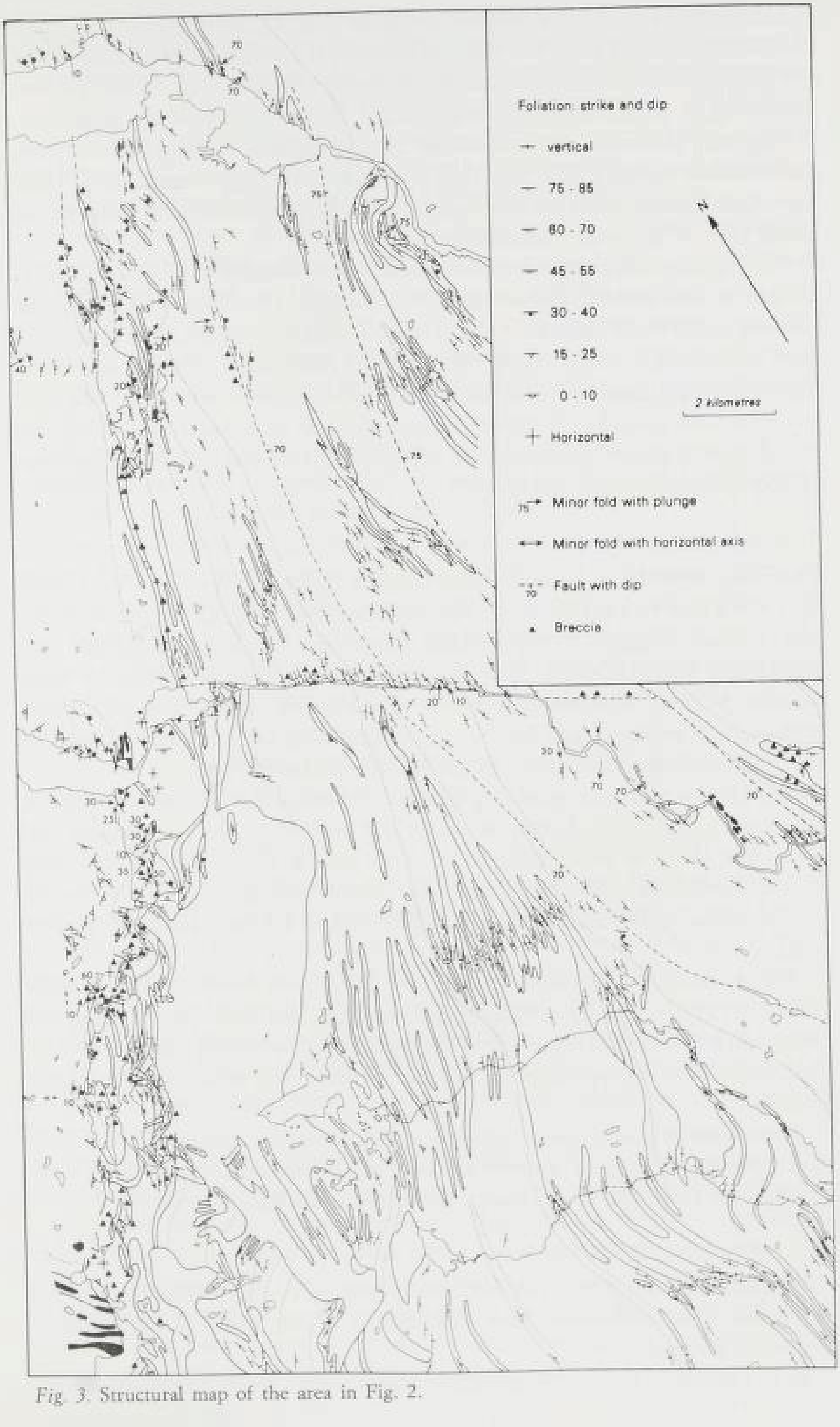

Fig. 3. Structural map of the area in Fig. 2.

North-eastwards from Krøderen to Sokna (Fig. 4), most gneisses resemble those of the central zone, but the coarser-grained types with plagioclase augen are less common (see above). East and northeast of Sokna (Fig. 4), the gneisses contain more hornblende and almandine and thin amphibolites are commoner: further east (between Vælsvatnet and Begna) variable gneisses contain some biotite schist layers, a few very thin quartzites and some amphibolites. These are subtle changes probably inherited from lithological differences between the major units of the early regional structure.

Throughout the area, almandines grew across cataclastic fabrics and were broken by later faulting and brecciation. Some early almandines (notably in the western zone) were broken by the cataclasis. Shadowy microclines (associated with granitisation effects from the Telemark, Brennåsen, Vatnås and Vælsvatnet granites) overgrew cataclastic fabrics and, together with later clinozoisite–muscovite growths ($\pm$ biotite, pistacite), were granulated by faulting and brecciation. Extreme retrogressions produced quartz–plagioclase–biotite–muscovite–epidote gneisses and schists.

THE 'VATNÅS GRANITE'

This body of pink-red granitic gneiss (usually called the 'Vatnås granite') lies between 6 and 14 km east of the ultramylonite zone (Fig. 2). It was an intrusive granite, elongated in the regional structure with thick, concordant apophyses and internal fabrics. It is a medum-grained granitic gneiss with some massive patches. Subequal proportions of microcline, quartz and plagioclase, with some hornblende and biotite, are accompanied by occasional almandine. Late muscovite and clinozoisite are often associated with biotite.

The granite cuts, and variably granitises, quartz–plagioclase–biotite gneisses and a north-south belt of protomylonites and mylonites. Thin, concordant and discordant sheets of amphibolite occur in the granite and its country rocks and all these lithologies may have later shear fabrics and mortar textures cut by granitic veins, aplites and pegmatites (which extend 2.5 km west of the granite margin).

The intrusion of the Vatnås granite was dated at $\sim$ 1370 m.y. by Jacobsen & Heier (1978). It cuts a mylonitised zone, but has undergone subsequent cataclasis, and was then reworked to form granite veins, aplites and pegmatites. It is therefore of critical importance in the synthesis of a sequence of geological events.

Thin veins of garnetiferous granitic gneiss occur just southwest of Simoa, near Solumsmo. They lie about 4 km east of the northern tip of the Vatnås granite, but are of the same relative age as the main body of granite.

THE 'TELEMARK GRANITE'

The 'Telemark granite' (or 'Telemark gneiss–granite') forms a large body in the south, but thins northwards from Krøderen to Sokna, where it disappears in granitised gneisses.

It is a medium- to coarse-grained, red-pink granitic gneiss containing microcline, oligoclase, quartz and biotite with sporadic hornblende, muscovite and almandine. Despite some variations from granitic to granodioritic, monzonitic and adamellitic compositions, this large body is reasonably homogeneous over wide areas: it seems to be of igneous origin, causing extensive granitisation. It transgressed and granitised the ultramylonite zone and the early mylonite band, north of Prestfoss; shadowy microcline porphyroblasts and augen (reaching 1 cm and rarely 3 cm size) formed in mylonites and ultramylonites, with more intensive effects producing granitic augen gneisses, south of Prestfoss. Some early amphibolites were apparently metasomatised during emplacement of the granite. North of Prestfoss, the granite included some of the early mylonite band (e.g. north of Langevatnet) but subsequent severe cataclasis affected it. The development of foliation and the growth of garnets reflect the Sveconorwegian regeneration which also reworked the granite producing veins and pegmatites.

On the north shore of Soneren, variably granitised plagioclase–quartz–biotite–hornblende gneisses have very shallow and variable dips and probably represent the roof levels of the Telemark granite body.

Ampibolite sheets cut the granitised rocks and a few cut the Telemark granite. Microcline porphyroblastesis in the amphibolites is related to later granitic pegmatites, aplites and thin granitic veins, representing a regeneration of the Telemark granite and cutting all lithologies (and the concentric folds). Larger, discrete pegmatite bodies (e.g. on the south shore of Soneren) may carry almandine, magnetite, zircon, orthite and sphene, all of which occur in the granite itself.

Friction breccia effects were intensive near the margin of the Telemark granite and frequently disrupted the pegmatites.

THE VINOR METAGABBROS, METADIORITES AND AMPHIBOLITES

This intrusive phase consisted of a series of injections, now forming some large bodies comprised of coronite, metagabbro, amphibolite and minor metadiorite. A separate series of later amphibolite dykes cut the larger bodies and their country rocks over the whole area. All these intrusives cut the ultramylonite zone and the mylonite band (north of Prestfoss), but may also show later shearing. Microcline and plagioclase porphyroblasts developed adjacent to pegmatites associated with the Telemark, Vatnås, Brennåsen, Vælsvatnet and Ådal granites. The late amphibolite dykes cut all these bodies, except the Ådal granite.

A number of large gabbro bodies are shown on the maps (Figs. 2–5). Throughout the area, elongate, subconcordant sheets of amphibolite and metagabbro have the same age relationships and are cut by the late amphibolite dykes. North of Simoa are a number of thick bodies of coarse, segregated amphibolite rich in garnet. They occur in Krøderen village, in Sokna, southwest of Rådalen, south of Nedre Vælsvatnet and in numerous sheets east and north of Vælsvatnet. They are cut by the amphibolite dykes and may represent earlier intrusions than the large metagabbro–coronite masses.

South of Prestfoss, in one large outcrop at its northwestern corner, the Ivarskollen gabbro overlies mylonite, with a contact dipping 30°E. Although this western margin is complex, the body may have been partly lopolithic, producing the extensive surface outcrop. On its eastern side (east of Lauvnesvatnet) the main body passes into metagabbro interlayered with gneisses on all scales. Southwards, the body develops an irregular outcrop representing the roof of the intrusion: some of the gneiss, protomylonite and mylonite southwest of Lauvsnesvatnet may be displaced, rafted material.

West of Prestfoss, on the north shore of Soneren (Fig. 2) a small metagabbro–amphibolite body cuts the Telemark granite and granitised gneiss. It was clearly emplaced after the Telemark granite, but is itself cut by amphibolite dykes and thick granitic pegmatites, the latter producing porphyroblasts of plagioclase (up to 1 cm in size) and some of microcline.

Small patches of metadiorite are massive or foliate. They occur in the metagabbros in a number of places (e.g. east of Haugesjø, along the west of the Ivarskollen body, at Vælsvatnet and east to Begna). Rarely they contain xenoliths of metagabbro, in which they have caused plagioclase porphyroblastesis. More commonly, foliate varieties contain amphibolite lenses (20–100 cm long) which are orientated parallel or at a small angle ($< 25°$) to the foliation. Plagioclase porphyroblastesis has converted some lenses to ghost relics. Often these rocks are also cut by the late amphibolite dykes. Both the host rock and the lenses may be similarly garnet-rich or garnet-free and occasionally the almandine is developed only as a rim on the lenses. This suggests garnet grade metamorphism after their inclusion. In a few cases, small dioritic patches (not shown on Fig. 2) within the metagabbros resulted from contamination by 'mobilised gneisses', which are more extensively developed south of the present area.

The late amphibolite dykes are usually 3–100 cm wide, but rarely reach several metres. Some are massive, but commonly they are foliated parallel to their margins and contain some biotite and more rarely garnet.

Dykes of several generations are present and in places they cut one another. Their orientations were partly random and partly controlled by the structure of the gneisses. In many cases they have caused the growth of poikiloblastic hornblendes in cataclased rocks (including the mylonites and ultramylonites). Boudinage of some bodies occurred during the concentric folding. Late fabrics of biotite, clinozoisite and pistacite are often strongly developed.

THE ÅDAL (FLÅ) GRANITE

In the north of the area (Fig. 4), the post-kinematic Ådal 'granite' (a quartz monzonite) was emplaced after the intrusion of the late amphibolite dykes and contains xenoliths of them. Granitic (and pegmatitic) injections, from the Ådal granite, cut rocks as far south as the Vælsvatnet metagabbro (Fig. 4): associated microcline porphyroblastesis in many places post-dated the development of clinozoisite. The granite intruded after the concentric folding and modified surrounding structures.

The western margin of the Kongsberg Series

From Kongsberg to Flesberg the margin of the Kongsberg Series is an early mylonite zone, dipping to the east, with the 'friction breccia' (due to late faulting) adjacent to it (Starmer 1977, 1979). From Flesberg, the mylonite zone continues north to Haugesjø (Fig. 1) with a layer dip decreasing from around 60° to ca. 30°E and becoming increasingly deformed by concentric folding. The thinning may be ascribed either to tectonic or to lithological controls. At Haugesjø, a complex basin structure and a break in the mylonite zone seem largely responsible for a major bend in the Kongsberg Series' margin, which then trends northeast into the present area.

From Haugesjø to Prestfoss (Figs. 1–3) the ultramylonite is intensely deformed by concentric folds, but continues northeast, having been sandwiched prior to the folding between a coarse 'Telemark granite' to the west and a large gabbro to the east. At Prestfoss, it is deformed into another major basin structure. Further north, the margin changes character, with a narrower band of mylonite and protomylonite, diminishing northwards to Sokna and branching south of the Ådal granite (Figs. 1 & 4). Later faults and breccias follow the entire margin and some occur north of the Ådal and Hedal granites.

Hageskov (1979, 1980) suggested that the 'real brecciated part' of the shear zone disappeared around the Sperillen (Ådal) granite, but that the 'heavily sheared and blastomylonitic rocks of the shear zone' continued north to Bagn. From Kongsberg to Sokna, 'a strongly deformed linear belt' followed northwards along the shear zone: just west of Sokna, Hageskov suggested both were deformed by a rather tight, late fold and changed orientation from northeast to northwest, subsequently turning north (approximately along the line of the fault shown on Fig. 1) to be cut by the Flå (Hedal) granite and reappearing on its northern side at Bagn. The folded shear zone and accompanying linear belt striking from NW to NNW between Sokna and the Ådal granite are not consistent with structural trends observed by the present author (Figs. 1, 4 & 5) or by Smithson (1963).

The present study suggests that the main shear zone (the 'mylonite band') dissipated northeastwards towards Sokna, from where it continued with the same general trend, splitting along several lines and being broken by later folding. The 'heavily sheared and blastomylonitic rocks' found by Hageskov to the northwest of Sokna are not considered to represent the main shear zone, but may reflect the branching on several movement planes and may be partially related to later faulting.

The cataclased zone at Bagn (about 70 km northnorthwest of Sokna) continues for 8–10 km northeastwards through Bruflat. This zone (shown on Fig. 1) was described by Strand (1954) and Smithson (1963) and consists of sheared granitic augen gneisses and mylonites with a N to NW dip and a lineation down-dip: its orientation is related to that of the Caledonian belt which it underlies. The zone forms a boundary between two major divisions in the Precambrian basement, with quartz–dioritic gneisses to the north and granodioritic gneisses to the south.

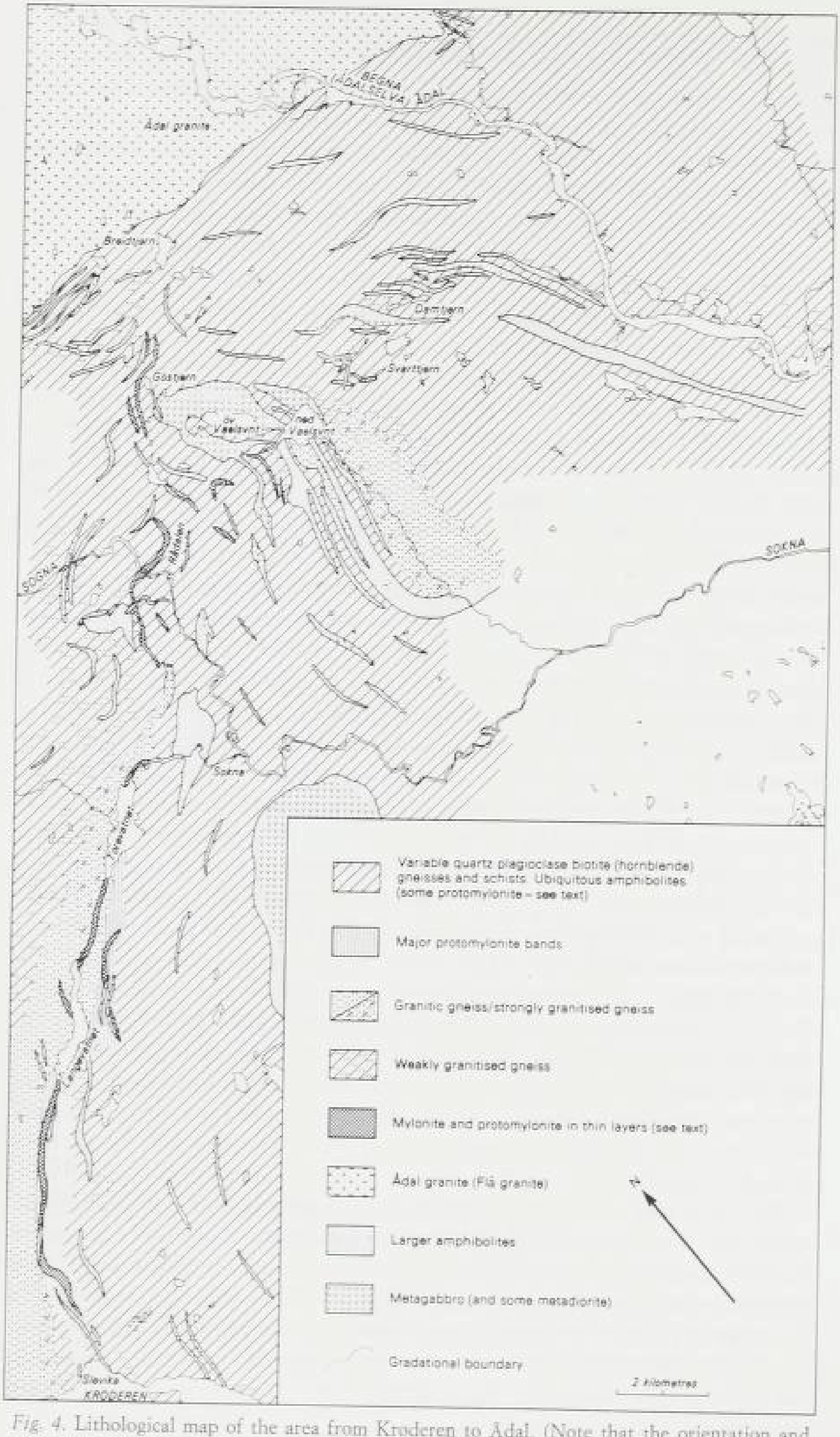

Fig. 4. Lithological map of the area from Krøderen to Ådal. (Note that the orientation and scale are not identical to those of Figs. 2 & 3.)

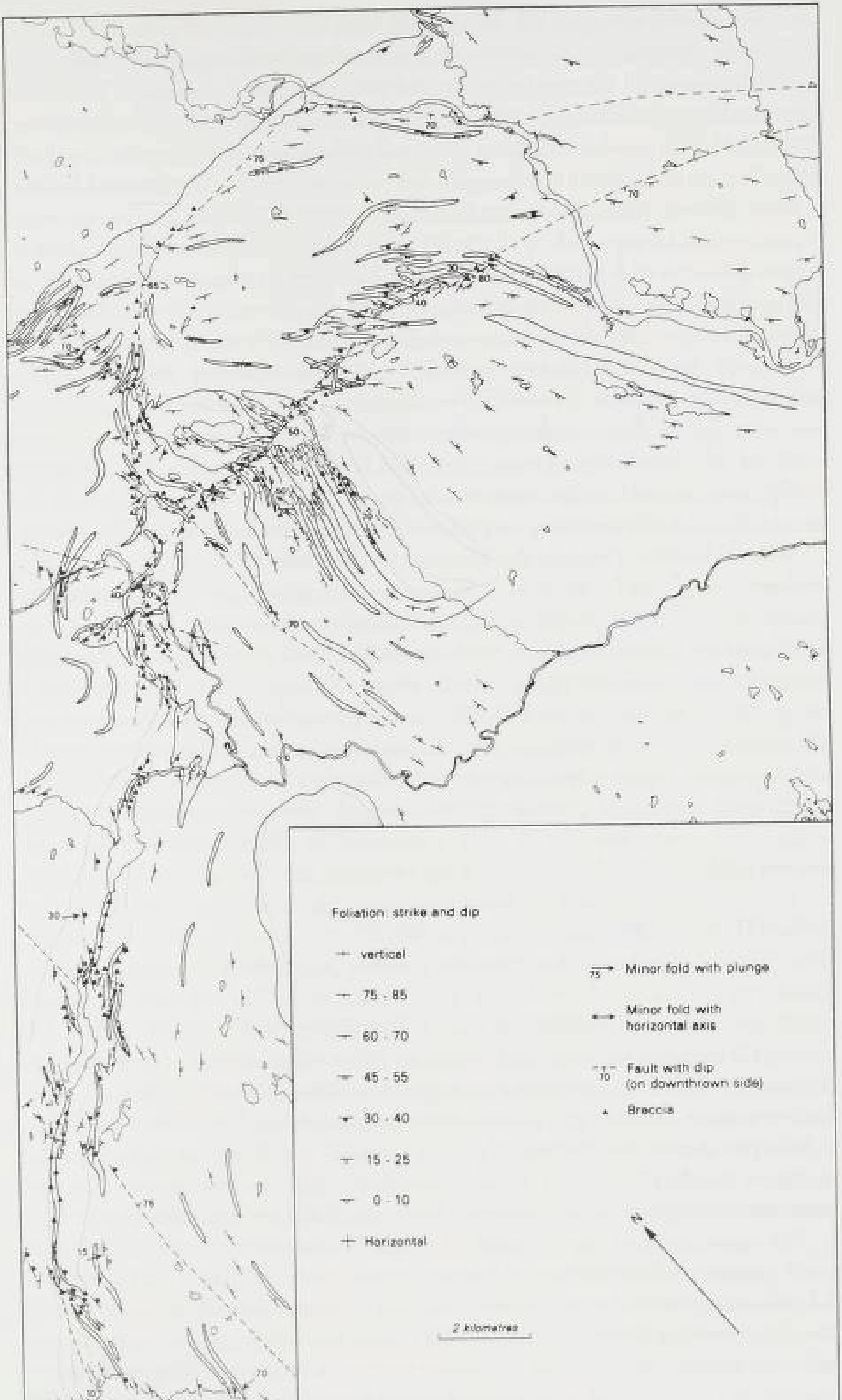

Fig. 5. Structural map of the area in Fig. 4.

THE ULTRAMYLONITE ZONE SOUTH OF PRESTFOSS

This zone, discussed in detail from the area immediately to the south (Starmer 1979), consists of ultramylonite (< 0.1 mm grain size), mylonite and some protomylonite, which were partially granitised to augen gneisses by the underlying Telemark granite. The zone was sandwiched between the granite and the large Ivarskollen gabbro (to the east) before being complexly deformed by concentric folding. Complex major basin structures developed in the north at Prestfoss and in the south at Haugesjø. At Prestfoss, the main basin formed in augen gneisses and granitised mylonites, but just to the east the structural pattern is simpler with mylonites and thin metagabbros concentrically folded together in separate folds on axes plunging moderately southeast and northeast.

East of the ultramylonite zone, there are numerous thin bands of protomylonite and mylonite. West of Letmolivatnet, protomylonite contains lensed-out amphibolite sheets and may reflect late, strike-slip movements, the importance of which will be discussed later. The Vatnås granite cuts thin mylonites within protomylonite (generalised on Fig. 2): these are often granitised and recrystallised and sometimes overgrown by coarse muscovite and clinozoisite.

Where the effects of late deformations can be reliably removed in the ultramylonite zone (and where there is no steepening against the gabbro), porphyroclasts record NW–SE stretching lineations in an early foliate layering, with minimal flattening perpendicular to it. The shear planes dipped moderately southeast (at around 40–60°) with a down-dip sense of movement. This is compatible with the present NE–SW orientation of the ultramylonite zone and the general westward upthrust found in areas to the south. The intense granulation occurred before microcline porphyroblastesis, but sometimes the K-feldspars were later rolled (by dip-slip movements). Subsequent growths of clinozoisite, muscovite and biotite were granulated by the friction breccia movements, with sporadic later overgrowths of calcite.

THE MYLONITE BAND NORTH OF PRESTFOSS

The western margin of the Kongsberg Series, north from Prestfoss to Glessjøen (Fig. 2), is a thin mylonite band (generally < 150 m and sometimes < 50 m wide) locally containing thin ultramylonite layers. North from Glessjøen, it diminishes and the band shown on Fig. 4 often consists of thin mylonite layers (< 5 m thick) alternating with protomylonite. North of Sokna, it splits into a number of thinner, more weakly cataclased bands broken by folding. Although close to the Telemark granite margin from Prestfoss to Sokna, the mylonite band is commonly bounded on both sides by similar gneisses and protomylonites (traditionally separated into the Telemark or Kongsberg Series).

The mylonite band was originally formed before emplacement of the Telemark granite and is incorporated as large enclaves in the latter (e.g. north of Langevatnet). Subsequent, severe cataclasis occurred before the late brittle faulting, crushing microcline porphyroblasts in the mylonites and shearing the adjacent granite. All the later movements were concentrated along the granite margin, the mylonites dissipating and branching where the granite terminated north of Sokna. Microcline porphyroblastesis in the mylonite was usually associated with late pegmatites, aplites and thin granitic veins. In a few places (e.g. around Glessjøen), granitisation (presumably from the adjacent Telemark granite) occurred before intrusion of amphibolites which were subsequently sheared and cut by pegmatites. The mylonite band was intruded by the gabbros and the amphibolite dykes, deformed by concentric folding, injected by pegmatites and aplites, overgrown by clinozoisite and broken by the friction breccia movements.

The mylonite band was deformed by minor concentric folds and a few larger, open flexures which, north of Sokna, become tighter, giving major folds (which disrupted the mylonite band). From Sokna south to Krøderen, the mylonite curves, partially following the major structure, and generally dips 30–50° SE or ESE; further south, from Krøderen to Prestfoss, it dips 50–80° ESE and the lack of folding facilitates an analysis of movements.

Cataclasis in the mylonite and adjacent protomylonite occurred on these dipping planes, with early dip-slip movements and subsequent, minor displacements which were either strike-slip or oblique, pitching 20–40°N. In some places, monoclinal microfolds (deforming the early cataclasis fabric and lineations and cut by pegmatites) were formed during the second movement and show that the eastern (Kongsberg Series) side moved south and in places obliquely upwards, with little compressive stress normal to the shear planes. Some intruded amphibolites show only the strike-slip or oblique-slip lineation (with some microfolds) suggesting that they intruded after the earlier movements. Just north of Slevika, thin, subconcordant amphibolites intruded mylonite, but their apophyses were streaked out by strike-slip displacements during injection. Amphibolites thus intruded before, during, and after later movements within the mylonite.

East of the mylonite band, thin cataclased bands resulted from various movements. In the 'central zone' gneisses, southwest of Medrudtjern (Figs. 2 & 3), plagioclase porphyroclasts record strike-slip movements and nearby lineations suggest oblique displacements, pitching 40–60°S on N-S planes.

LATE FAULTING AND 'FRICTION BRECCIA' MOVEMENTS

The friction breccia, traditionally taken as the boundary to the Kongsberg Series, is represented by a diffuse faulting and brecciation of the earlier ultramylonite and its adjacent rocks south of Prestfoss. Northwards, from Prestfoss to Sokna, the breccia largely followed the mylonite band, but north of Sokna it split along several lines, partly following the mylonites (which had branched here) but cutting their folded structures. Some very thin mylonites were associated with this late, brittle faulting.

The brecciation affected all rocks (including the pegmatites): minor faults cut the Ådal granite and there was some shearing around its margins. The displacements produced granulation and mortar textures, breaking overgrowths of microcline and clinozoisite. A veining of quartz, calcite, chlorite and epidote is also sometimes disrupted by the late movements.

The friction breccia resulted from a series of movements, with an overall normal displacement and downthrow of the southeast (Kongsberg Series) side. It was part of a regional system of faults (showing a series of movements) and is linked to the perpendicular Simoa fault, trending southeast from Prestfoss and downthrowing the Kongsberg Series to the north. The Simoa fault continues southeastwards in a breccia north of Solumsmo (Figs. 2 & 3), before being cut out by a NNW–SSE fault which here forms the western margin of the Modum area complex: small drag folds show that at least one movement involved an upthrow of the eastern (Modum) side. No westward continuation of the Simoa fault was found in the Telemark granite west of Prestfoss, and its maximum movement was in the sector east of the friction breccia with which it was obviously linked. However, there may have been parallel faults westwards in lake Soneren. Along the Simoa fault, granulation and mylonitic fabrics were overgrown by clinozoisite, which was cracked by later movements, indicating a series of displacements similar to those on the friction breccia.

North of Sokna, several major faults are distinguished (Fig. 5) and intervening minor faults form a mosaic. The main breccia trends northeast from Sokna across the Sogna river into Rådalen, where the southern side is downthrown. It then splits, with a number of branches trending from due north to due east. In Begna, just south of the Ådal granite, a brecciated fault dips steeply southwest and has drag folds showing a late strike-slip, with the southwest side displaced southwards. Many of the brecciated minor faults are either parallel to the major friction breccia lines or are markedly transverse, sometimes subparallel to the Begna fault and showing the same late strike-slip.

Folding and general structure of the western Kongsberg Series

The Kongsberg Series structurally overlies the Telemark Series along a cataclased junction. In the Kongsberg Series, the earliest recognisable foliation is a penetrative fabric parallel to transposed lithobanding: rare intrafolial isoclines are preserved. The foliation was overprinted (particularly in the extreme west) by almost coplanar cataclastic fabrics and the structure was later complicated by concentric folding and accompanying deformation around large gabbros.

In the southern part of the area (Figs. 2 & 3), the north-south structure and discontinuous lithobanding of the western Kongsberg Series continue as far north as Krøderen. East of the ultramylonite zone (and the Ivarskollen gabbro), the foliation and banding generally dip very steeply east to vertical. The dips change to become steeply west to vertical across the Vatnås granite and to the east of it. The change occurs just west of the Vatnås granite and can be traced further south (to the south of the present area): this seems to represent a major N-S synform developed prior to the formation of the ultramylonite zone and the intrusion of the Vatnås granite. North of the Simoa fault to Krøderen, the general structure dips 55–75°E.

Further north, the major structure turns northwest from Krøderen to Sokna, with a moderate dip southeast, probably partly related to deformation around a large metagabbro. North of Sokna, a complex folded pattern has developed towards the Ådal granite.

Concentric, buckle folds and flexures developed after intrusion of the late amphibolite dykes, but before the pegmatite activity. Two sets of folds interfered on all scales, giving domes and basins in the Telemark Series (including the previously foliated Telemark granite), in the mylonite zones and in the Kongsberg Series where the folds become less common eastwards. Many were small, open flexures (or domes and basins) on foliation planes and did not significantly alter the layer dip. Others were minor flexures or better defined concentric folds with wavelengths from 1 m to 10 m and much smaller amplitudes. Larger flexures and folds affected the major structures, notably at Haugesjø and Prestfoss, forming basins in the ultramylonite, and north of Sokna, breaking the mylonite band and producing disharmonic folds with considerable transposition. Axial plunges vary and were largely controlled by the previous structure, but many are low to moderate to the northeast and southeast (becoming N-S and E-W in places). Axial planes are normally subvertical, but in places north of Krøderen, are overturned to the east and in rare cases are subhorizontal.

Synthesis and discussion

Away from its western margin, where a later, N-S cataclastic banding was variably superimposed, the Kongsberg Series retains remnants of an 'early regional structure' with major domes and basins developed during upper amphibolite facies metamorphism. Even 5–10 km east of the ultramylonite zone (at the western margin) there seems to be a relict, major synform on the west side of the Vatnås granite.

The early regional structure may have contributed towards a number of features (e.g. the distribution of fahlbands in the supracrustal gneisses). The bend in the ultramylonite at Haugesjø is now largely controlled by a basin structure (Starmer 1979), but further north the northeastward trending ultramylonite was sandwiched between the Telemark granite and the large gabbros before the concentric fold phase, which produced the basin. Some early curvature of the ultramylonite zone may therefore have existed and could have developed in quartz-rich rocks deformed around the early regional structures.

It has already been noted that the rocks of the Modum complex differ from those in the surrounding Kongsberg Series and are remarkably similar to those in the Bamble Series, and that the location of this complex could be explained by the early regional structure. Although the Bamble and Kongsberg Series contain rather different lithologies, they underwent the same sequence of major events (Starmer 1977): they may have been deformed together in the Svecofennian orogeny, the Modum complex representing Bamble Series rocks infolded into the Kongsberg Series. There has also been a relative uplift of the Modum complex along its western margin, the movement producing an early mylonite zone with an adjacent, later fault. The constant association of thick, pure quartzites and sillimanite-rich rocks in early domes and basins, with transposition and isolation of some structures, is seen in Svecofennian folding of both the Bamble Series (Starmer 1978) and the Modum complex. Later events in both the Bamble and the Modum (before and during the Sveconorwegian Regeneration) included scapolitisation of newly-intruded gabbros and the development of orthoamphibole–cordierite rocks, large pegmatites and albitites, indicating possible retention of contiguity at that time.

The Kongsberg Series was thrust westwards, its earlier dome and basin structure becoming partially modified to a north-south banding in its western parts. The frontal thrust and original western margin was the ultramylonite zone, preserved only in the widest part of the Kongsberg Series and cut out north of Prestfoss by the 'mylonite band'. Brecciation of this mylonite band (producing the 'friction breccia') and movement on the Simoa fault downthrew the Kongsberg Series north of Simoa, leading to the exposure of higher structural levels. This partly cancelled the original displacement on the mylonite band, especially further north where the movement had been less. This hingeing effect, probably influenced by the early regional structure, may have affected only the western margin, since there were numerous movement planes to the east. The hingeing and the late faulting, coupled with a decreased dip and retention of some of the early regional structure, led to exposure of similar rocks on either side of the friction breccia: there is no significant difference between supposed 'Telemark' and 'Kongsberg Series' rocks on either side of this breccia zone.

The ultramylonite zone was cut out at Prestfoss by a combination of the steeper, brecciated 'mylonite band' and the Simoa fault. It could have terminated here originally due to lithological changes related to the early regional structure (e.g. the disappearance of quartz-rich rock necessary for its development). The formation of the later basin structure in the ultramylonite at Prestfoss was partly controlled by its position at the northern end of the Ivarskollen gabbro.

The mylonite band north of Prestfoss developed after the ultramylonite zone, but synchronously with other cataclased belts further south, from Prestfoss to Haugesjø. These include mylonites along the eastern side of the ultramylonite and augen gneiss, late shears in the ultramylonite, and mylonite–protomylonite zones lying southwest of Lauvnesvatnet. Some of the mylonite band's later movements were synchronous with the shearing of metagabbro layers east of Lauvnesvatnet and the strike-slip displacements correlated with the lensing-out of amphibolite in protomylonite, west of Letmolivatnet.

Radiometric work places some contraints on the age of major events. The Rb–Sr data of Jacobsen & Heier (1978) recorded two major metamorphisms in the southern Kongsberg Series at about 1600–1500 m.y. and 1200–1100 m.y. B.P. ($\lambda = 1.39 \times 10^{-11} \cdot yr^{-1}$). The 'Telemark granite' of the present area has a similar lithology to the Helgevatnet granite, intruded at around 1200 m.y., and reconnaissance of the latter reveals a similar relative age (in terms of cataclasis, amphibolite dyke intrusions, etc.). The radiometric data of Jacobsen

& Heier can be combined with the sequence of events in the present area to produce the following history:

1) Supracrustals (now variable quartz–plagioclase–biotite gneisses and some amphibolites) were deformed into major dome and basin structures during Svecofennian upper amphibolite facies metamorphism (1600–1500 m.y.).
2) Westward upthrusting of the Kongsberg Series produced the ultramylonite zone (south of Prestfoss) during middle to upper amphibolite facies metamorphism.
3) Initial formation of the 'mylonite band' (north of Prestfoss) during amphibolite facies metamorphism.
4) Vatnås granite intruded (cutting a cataclased belt) (~ 1370 m.y.).
5) 'Telemark granite' intruded (~ 1200 m.y.).
6) Renewed cataclasis (particularly around the 'mylonite band' north of Prestfoss).
7) Vinor gabbros intruded (one body cutting the 'Telemark granite'). (~ 1200 m.y.).
8) Upper amphibolite facies metamorphism.
9) Intrusion of Vinor dykes (now amphibolites). (Probably before 1070 m.y.).
10) Weaker cataclasis (including strike-slip movements).
11) Concentric folding (in places forming domes and basins) on all scales.
12) Formation of granitic veins, aplites and pegmatites by reworking of the Telemark, Vatnås, Brennåsen and Vælsvatnet granites.
13) Epidote–amphibolite facies metamorphism giving overprints of epidote and muscovite in all rocks.
14) Ådal granite intruded (~ 913 m.y.) (date of Killeen & Heier 1975).
15) Brittle faulting (initially at lower to middle greenschist facies) producing the friction breccia and the Simoa fault.

In this sequence of events, elements 8 to 13 inclusive constitute the Sveconorwegian Regeneration.

Periodic cataclasis occurred over an extended period of time. Some early movements took place before the intrusion of the Vatnås granite (~ 1370 m.y.) and the western ultramylonite zone formed before the Telemark granite was emplaced (~ 1200 m.y.). Subsequent severe cataclasis, particularly in the 'mylonite band' north of Prestfoss, affected the Telemark granite, and weaker, later shearing occurred during and after the intrusion of amphibolite dykes.

The earlier movements, prior to the emplacement of the Vatnås and Telemark granites, have been considered to represent a westward thrusting of the Kongsberg Series over the Telemark block. Southwards from the present area, the dip of the western mylonite zone steepened from around 60°E at Flesberg to some 70–80°E where it disappears beneath the Oslo Graben. (The latter orientation is taken from Starmer (1977) with removal of the subsequent southeastward tilting of the Oslo Graben.) The orientation of the western mylonite zone indicates that, at the levels exposed south of the present area, vertical displacements predominated over horizontal motions. Thrust zones commonly steepen in depth due to a re-orientation of the stress trajectories, and the lack of flattening perpendicular to the layering of the mylonites may suggest that (at the exposed levels) east-west compressive stresses were not the dominant forces. The movement, therefore, could also be considered as a net downsinking of the Telemark region, leading to anatexis and production of large volumes of Telemark granite.

Subsequent gabbro intrusions and dykes indicate a period of rifting which was aborted by the onset of the Sveconorwegian Regeneration.

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Some Features of Tectonic Deformation of Old Red Sandstone Sediments on Hitra, West Central Norway

DAVID ROBERTS

Roberts, D. 1981: Some features of tectonic deformation of Old Red Sandstone sediments on Hitra, West Central Norway. *Norges geol. Unders. 370*, 45–48.

Minor tectonic structures affecting Downtonian Old Red Sandstone sediments on Hitra include folds parasitic to a major syncline, and an associated axial surface spaced cleavage. The cleavage is most prominent in mudstones, and appears to represent dissolution foliae; no neocrystallization has been detected. In these rocks carbonate nodular concretions show evidence of mechanical rotation from bedding-parallelism towards the plane of the cleavage. Younger structures are represented by crenulations and kink bands.

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Parorogenic tectonic deformation of the Old Red Sandstone (ORS) deposits of southern Norway first became widely known following Vogt's (1928) synthesis of late Caledonian crustal events throughout the North Atlantic region. Since that time, workers in the several ORS basins have recognised a variable degree of deformation of the intermontane molasse sediments, generally in the form of open folding and faulting (e.g. Holtedahl 1960, Holmsen 1963, Nilsen 1968, Siedlecka & Siedlecki 1972, Roberts 1974) but also reflected in thrusting (Nilsen 1968, Høisæther 1971), local pebble stretching in conglomerates (Nilsen 1968, Indrevær & Steel 1975) and in one area cleavage development with very low grade metamorphic recrystallization (Roberts 1974a).

The Downtonian (and possibly also Ludlovian?) to Lower Devonian sediments on Hitra have been a subject of modern stratigraphical and sedimentological study by Siedlecka & Siedlecki (1972). The sedimentary succession is divided into 5 informal members, A to E, which together constitute the Hitra Formation. Alternating sandstones and mudstones compose members B and D, but otherwise the formation is characterized by conglomerates and sandstones with a basal breccia (Siedlecka & Siedlecki 1972). These authors described the major structure affecting the apparently unmetamorphosed sediments; an asymmetrical open syncline of NE–SW to ENE–WSW trend with generally southeastward-dipping axial plane, which is transected by both longitudinal and transverse to oblique fault sets.

An assessment of the strain state of the Hitra sediments from Siedlecka & Siedlecki's descriptions was not possible since no minor tectonic structures relatable to the major syncline had been reported. Observations by the present author in 1977 showed, however, that mesoscopic folds are present locally in favourable multilayered pelite–arenite lithologies and that a prominent cleavage is developed in mudstones in the southeasternmost parts of the ORS outcrop (for locations, the reader is referred to Fig. 1 in Siedlecka & Siedlecki 1972, p. 2).

In the coastal tract east from Furuholmen, and especially in the small bay Bosvikvågen (1:50,000 map-sheet 'Hemne'), dark grey mudstones of member

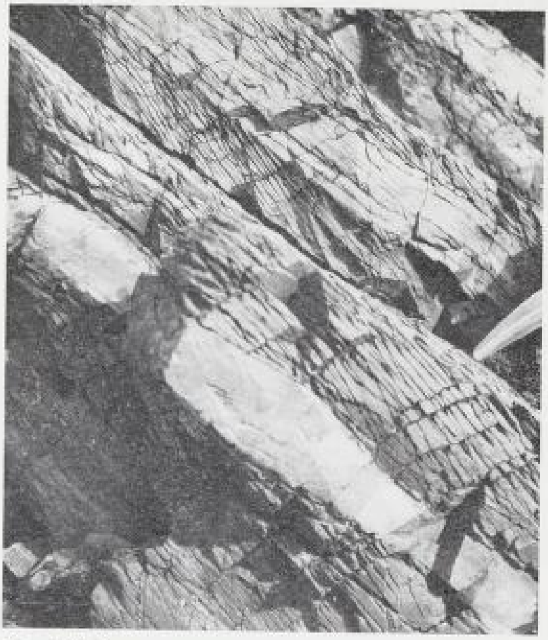

Fig. 1. Cleavage development in mudstones of alternating pelite–arenite sequence, member D, Hitra Formation, Bosvikvågen. Photo taken looking northeast approximately normal to the bedding/cleavage intersection lineation: part of hammer-head as scale.

D exhibit a conspicuous spaced cleavage (Fig. 1). Bedding/cleavage disposition and low-angle WSW-plunging intersection lineations indicate a close relationship to the major syncline. Minor parasitic folds are rare here. Some thicker sandstone units display a coarse fracture cleavage.

Curiously, the cleavage has not developed equally pervasively in all mudstone beds; in some pelites it is difficult to trace, or may be almost absent. The reason for these variations is not known. They cannot be ascribed solely to varying degrees of strain, but may associate with slightly differing mudstone mineralogies; for example, a higher ratio of quartz and calcite to phyllosilicates may facilitate dissolution processes and thus promote cleavage development. Even the most prominent cleavage is not a truly penetrative structure but rather a series of subparallel, slightly sinuous and anastomosing, closely spaced fracture surfaces separating domains of non-cleaved mudstone. Microscopic examination of the cleavage is extremely difficult as the rock splits so readily along these very surfaces. The examples that have been observed to date show the cleavage surfaces as dark seams of insoluble residues, largely clay minerals and oxides, the foliae apparently representing dissolution zones. No new mineral growth has been observed. Inter-cleavage domains reveal clastic quartz, feldspar, muscovite, sericite, chlorite and some carbonate as the main mineral constituents, with a crude preferred orientation of phyllosilicates paralleling the primary lamination and probably representing a compactional fabric. This latter seems a reasonable assumption in view of the likelihood that some 2–2.5 km of Devonian sediments may once have overlain the Hitra Formation in this coastal district of Norway (Siedlecka 1975).

An interesting feature of these mudstones is the presence of carbonate concretions and nodules. Siedlecka & Siedlecki described them as ellipsoidal or subdiscoidal and lying parallel to bedding. Siedlecka (1977) grouped the con-

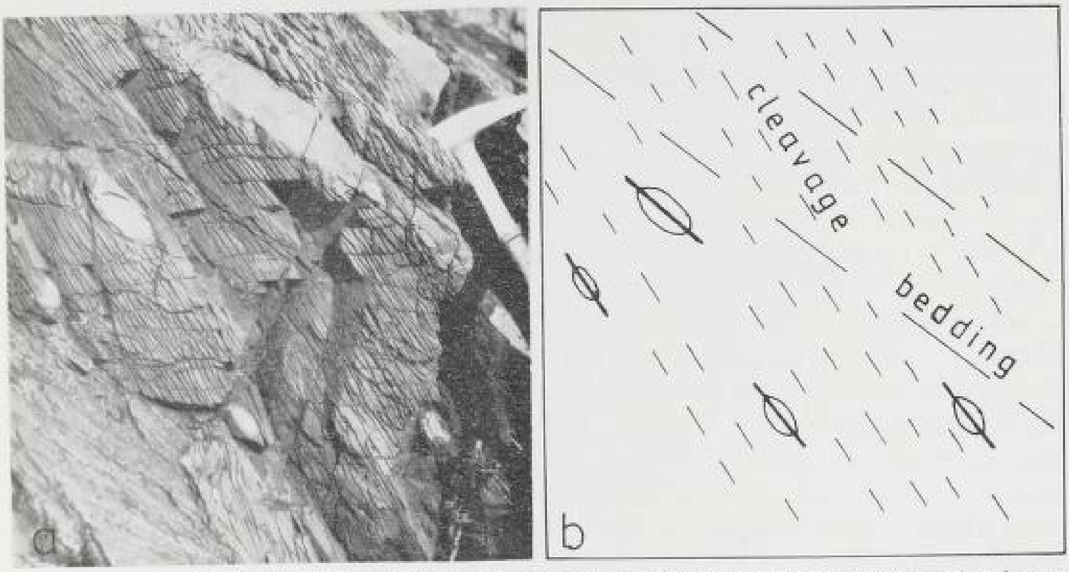

Fig. 2. (a) Carbonate nodular concretions in member D mudstones showing varying degrees of rotation from initial bedding-parallel orientation towards subparallelism with cleavage. (b) Tracing from (a) to help clarify the bedding/cleavage/nodule ellipse relationships. A heavy line is drawn through the long axis of each nodule. Bosvikvågen. Looking east-northeast on joint surface at right-angles to bedding/cleavage intersection.

cretions into two types: one which developed prior to the completion of sediment compaction, and another group which post-date this compaction. Where the cleavage is better developed it is clear that the nodular concretions have suffered a mechanical rotation, with final attitudes disposed at varying angles between bedding and cleavage (Fig. 2), the smaller concretions showing the greater propensity for tectonic rotation. In the present case it would be dangerous to attempt any finite-strain analysis in view of the apparently wide range of initial to post-compactional shape exhibited by these bodies, and also because of problems inherent in determining the amount of compactional strain suffered by these rocks. It can be noted, though, that some 20–30% transverse shortening across the ORS basin is indicated by the macroscopic folding of the Hitra Formation sediments, based on simple stratal rotation or unfolding.

Folds parasitic to and coaxial with the Hitra syncline are better developed on the southeastern limb of the major fold. At Selnes an incipient spaced cleavage is axial planar to close to tight mesoscopic folds in alternating mudstones and silty sandstones which Siedlecka & Siedlecki (1972) found difficult to place in their ORS stratigraphy. Microscopically, this cleavage is identical to that in member D; and the generally stronger deformation here is probably associated with strike-parallel, dip-slip faulting. In this area, traces of a later, N-dipping, incipient crenulation cleavage are detectable, and from here westwards towards Bosvikvågen there are NNW–SSE-trending kink bands with consistent westerly downstep in many mudstone horizons. A pilot TEM study of a cleaved member D mudstone has revealed micro-kink bands affecting clastic micas without any indications of recrystallization of phyllosilicates along the kink planes (S. White & B. A. Sturt, pers. comm. 1980).

The age of the fold-deformation affecting the ORS sediments on Hitra is generally considered as Svalbardian (early Upper Devonian), following Vogt (1928), although there are no precise stratigraphical constraints on this tectonism in this region of Norway. It follows that the cleavage would also be of the same age. Faulting in these coastal districts would appear to have been polyphasal (cf. Roberts 1974b, Siedlecka 1975) — initially syndepositional, then Svalbardian, but also with important dislocatory movements in Mesozoic (Oftedahl 1975) and Tertiary time. The syndepositional character of the faulting in western Norway in ORS time is well documented with strike-slip as well as normal dip-slip components (e.g. Bryhni 1964, Nilsen 1968, Steel 1976). Thrusting of ORS rocks in some areas in late Devonian time has probably been facilitated by a reversal of the slip vector along the earlier normal faults bounding the ORS basins (Steel et al. 1978). The age of the post-cleavage minor structures is open to speculation; any time from Svalbardian to Tertiary may be conjectured. As the kink bands denote downstep to the west, a relationship to either Permo–Jurassic rifting or the Tertiary uplift of Fennoscandian is quite within reason.

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