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Norges geologiske undersøkelse (Geological Survey of Norway), Leiv Eirikssons vei 39, Trondheim. Telephone: national (075) 15860, international + 47 75 15860. Postal address: Box 3006, N-7001 Trondheim, Norway.

Administrative Director: Dr. philos. *Knut S. Heier*

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EDITOR
Førstestatsgeolog Dr. *David Roberts*, Norges geologiske undersøkelse, P.O.Box 3006, N-7001 Trondheim, Norway.

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Structure and Petrology of the Bergen–Jotun Kindred Rocks from the Gjendebu Region, Jotunheimen, Central Southern Norway

TREVOR F. EMMETT

Emmett, T. F. 1982: Structure and petrology of the Bergen–Jotun kindred rocks from the Gjendebu region, Jotunheimen, central southern Norway. *Norges geol. Unders. 373*, 1–32.

The Gjendebu region is located within the Jotun Nappe, the highest tectonic unit of the southern Norwegian Caledonides. The structure of the area is dominated by the NE-trending Tyin–Gjende Fault, a major zone of repeated movement. To the northwest of this fault occurs a series of granoblastic ultrabasic, basic and intermediate gneisses (the Storådalen Complex) which show polyphase deformation and an intermediate-pressure granulite facies grade of metamorphism. To the southeast occurs an igneous-textured gabbro (Mjølkedøla Purple Gabbro) which grades eastward into the partially recrystallised Svartdalen Gneiss. Differences in major element composition between these units are minimal and they are believed to be comagmatic. Their geochemistry is broadly of calc-alkaline type, though all the rocks are anomalously potash-rich. The Storådalen Complex contains a complete differentiation sequence with both cumulate and liquid-descent trends apparent. Cumulate rocks do not occur in any quantity in the other units. Preliminary studies of pyroxene geochemistry also indicate the importance of igneous differentiation in the origin of these rocks.

After initial crystallisation, the Storådalen Complex was intensely deformed and then progradely metamorphosed, with conditions at the peak of metamorphism estimated at 1000°C, 9 kb. Preserved olivine + plagioclase assemblages in the rare ultrabasics southeast of the Tyin–Gjende Fault indicate that the Purple Gabbro and Svartdalen Gneiss have not exceeded low-pressure granulite facies grade. These high-grade metamorphic and deformational events are Precambrian in age. The Jotun rocks were partially exhumed prior to the Eocambrian, but uplift and thrust transport to their present position did not occur until an early phase of the Caledonian orogeny, in pre-Middle Ordovician time.

T. F. Emmett, Department of Science, Cambridgeshire College of Art and Technology, Collier Road, Cambridge, CB1 2AJ, England

Introduction

The Jotun Nappe forms the highest structural unit of the Caledonide Orogen of central southern Norway (Sturt & Thon 1978, and Fig. 1), and it is preserved as an erosional remnant within a regional synform known as the *Faltungsgraben* (Goldschmidt 1912, and Fig. 1B). The thrust upon which it rests, the so-called basal thrust (Hossack 1968),is a major tectonic feature af the orogen. The nappe itself (Fig. 1C), is composed mainly of basic and intermediate orthogneisses of amphibolite and granulite facies (Strand 1972), though in certain areas anorthosites (Hødal 1945, Gjelsvik 1947, Lacour 1969) and ultrabasic rocks (Carstens 1920, Battey 1960) are also common. Goldschmidt (1916) regarded this diverse suite of rocks as genetically unified and named it the Bergen–Jotun kindred. Recent geochronological work (Sturt et al. 1975, Schärer 1980, Austrheim & Råheim 1981) has demonstrated the antiquity of these rocks, with major magmatic activity extending over the period c. 1200 to 900 Ma b.p.. Pre-1700 Ma crustal relicts may also be present (Schärer 1980). A metamorphic event has been recognised at c. 900 Ma b.p., this being of a low amphibolite grade in

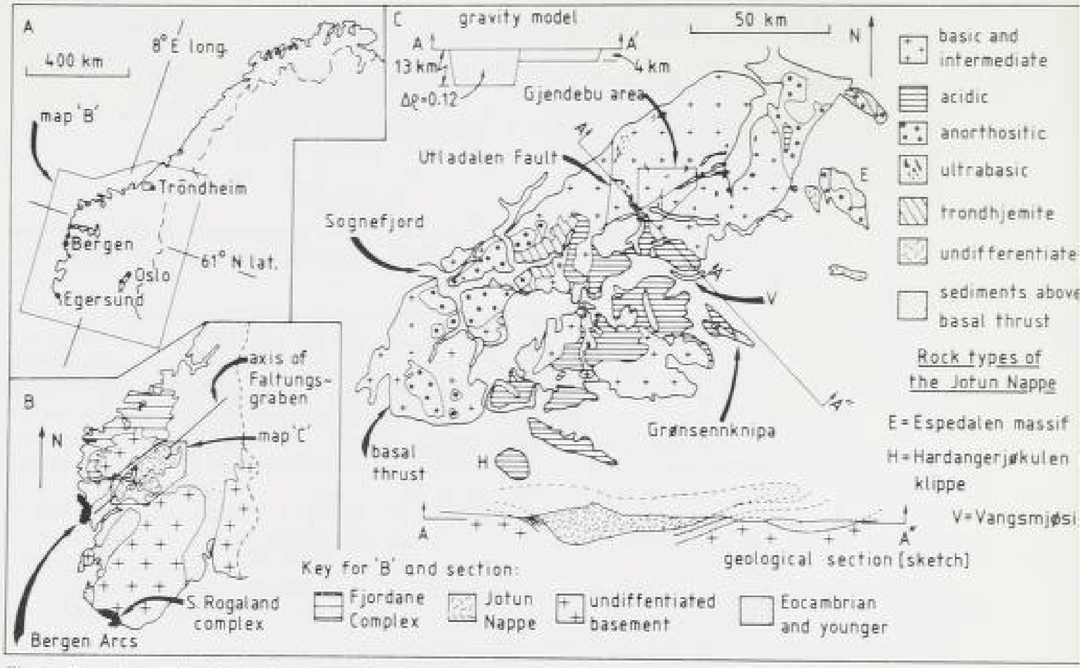


Fig. 1. Location and geological setting of the Gjendebu region. Compiled from Holtedahl & Dons (1960) and Carswell (1973). Geophysical profile after Smithson et al. (1974).

southern Jotunheimen (Schärer 1980) and granulite facies in the Bergen Arcs (Austrheim & Råheim 1981). At some localities outside the nappe, e.g. Grønsennknipa (Hossack 1972), Eocambrian (Vendian) and Lower Palaeozoic sediments rest unconformably upon the gneisses, but the low-grade Caledonian metamorphism which affects these does not penetrate far into the older crystalline plutonic rocks. Obvious Caledonian retrogression is confined to narrow zones immediately associated with faulting, thrusting and pegmatite injection (Emmett 1980b), although re-heating without apparent mineralogical effects did occur, as shown by the K-Ar age pattern (Battey & McRitchie 1973).

In erecting his Bergen–Jotun kindred, Goldschmidt (1916) drew a distinction between hypersthene-bearing gabbroic rocks of 'normal' igneous type (and usually with ophitic textures) and mineralogical similar rocks which lacked such features. This latter group he called the Jotun-norites, a name subsequently contracted to *jotunite* by Hødal (1945), who also provided a rigorous definition. In the region of central Jotunheimen, around the western ends of the lakes Gjende and Bygdin (Fig.1), this two-fold subdivision was elaborated upon by Battey (1965) and Battey & McRitchie (1973). These authors recognised a central body of totally recrystallised high-grade rocks ('axial' rocks) surrounded by several smaller bodies of lower grade rocks which, in places, have igneous (ophitic) textures. The two rock types, axial and peripheral, were considered to be everywhere separated by faults, the Tyin–Gjende Fault to the south and the Utladalen–Gravløyfti Fault Zone to the west (Fig. 1C). When this field evidence was compared to the gravity anomaly profile produced by Smithson et al. (1974), it was clear that the axial rocks formed the deep root (up to 13 km thick) of the nappe, and the peripheral rocks the much thinner 'pan-handle' projection to the southeast (Fig. 1). The break from deep root to thin sheet lay along the line of the Tyin–Gjende Fault (Battey & McRitchie 1973).

This paper will describe the area immediately to the east of that mapped by Battey (1965) and Battey & McRitchie (1973). It straddles the Tyin–Gjende Fault and so contains examples of high-grade axial rocks (the Storådalen Complex) and lower grade peripheral rocks (Mjølkedøla Purple Gabbro and Svartdalen Gneiss). The basal thrust occurs at one locality in the area, near Torfinnsbu, of which a description has already been published (Emmett 1980b). Rocks from the area have been briefly described by Sjøgren (1883) and Rekstad (1904), the latter author describing them simply as 'gabbro rocks'. Carstens (1920) examined the ultrabasic rocks and presented a crude map of their distribution. Axial rocks from the area immediately to the west, the Layered Series, have been described by Battey & McRitchie (1975). The area also contains a suite of minor intrusives of which dolerites are the most common (Battey et al. 1979, Emmett 1982). The other minor intrusives are very rare and include trondhjemite, oxide-rich gabbros, and rocks composed entirely of calcite and biotite. All these minor rocks are undeformed and only weakly metamorphosed, if at all. Their age is unknown but it may range from Precambrian to late Caledonian. Details of these minor rocks are given in Emmett (1980a).

Structural setting

Previous work in areas adjacent to the Gjendebu region (Battey & McRitchie 1973) has shown that petrographically distinct units of the Jotun kindred are usually separated by faults. An exception may be the ultrabasic 'pods' which appear to be cognate with the Layered Series and whose margins are «. . . movement surface welded by recrystallisation» (Battey & Davison 1977). In the Gjendebu region itself, the northeastward extension of the Tyin–Gjende Fault (Battey 1965) separates the gneissic Storådalen Complex (to the northwest) from the less deformed and less recrystallised Mjølkedøla Purple Gabbro and Svartdalen Gneiss to the southeast. This fault is cut and off-set by the Høystakka Fault, which locally juxtaposes the Purple gabbro and Svartdalen Gneiss. These two major faults therefore break up the Gjendebu region into three fault-bounded blocks (Fig. 2A and B). The Tyin–Gjende Fault must, in part at least, be of Caledonian age since it forms the northwestern boundary of a series of Caledonian nappes which underlie most of the Vangsmjøsi district (Heim et al. 1977, and Fig. 1).

STEEP FAULT ZONES

The Tyin–Gjende Fault has been traced from northwest of Tyin to the west end of Gjende, at which point it becomes more easterly in strike and runs out into the lake (Fig. 2A). It is recognised on the ground by a zone of intense deformation,

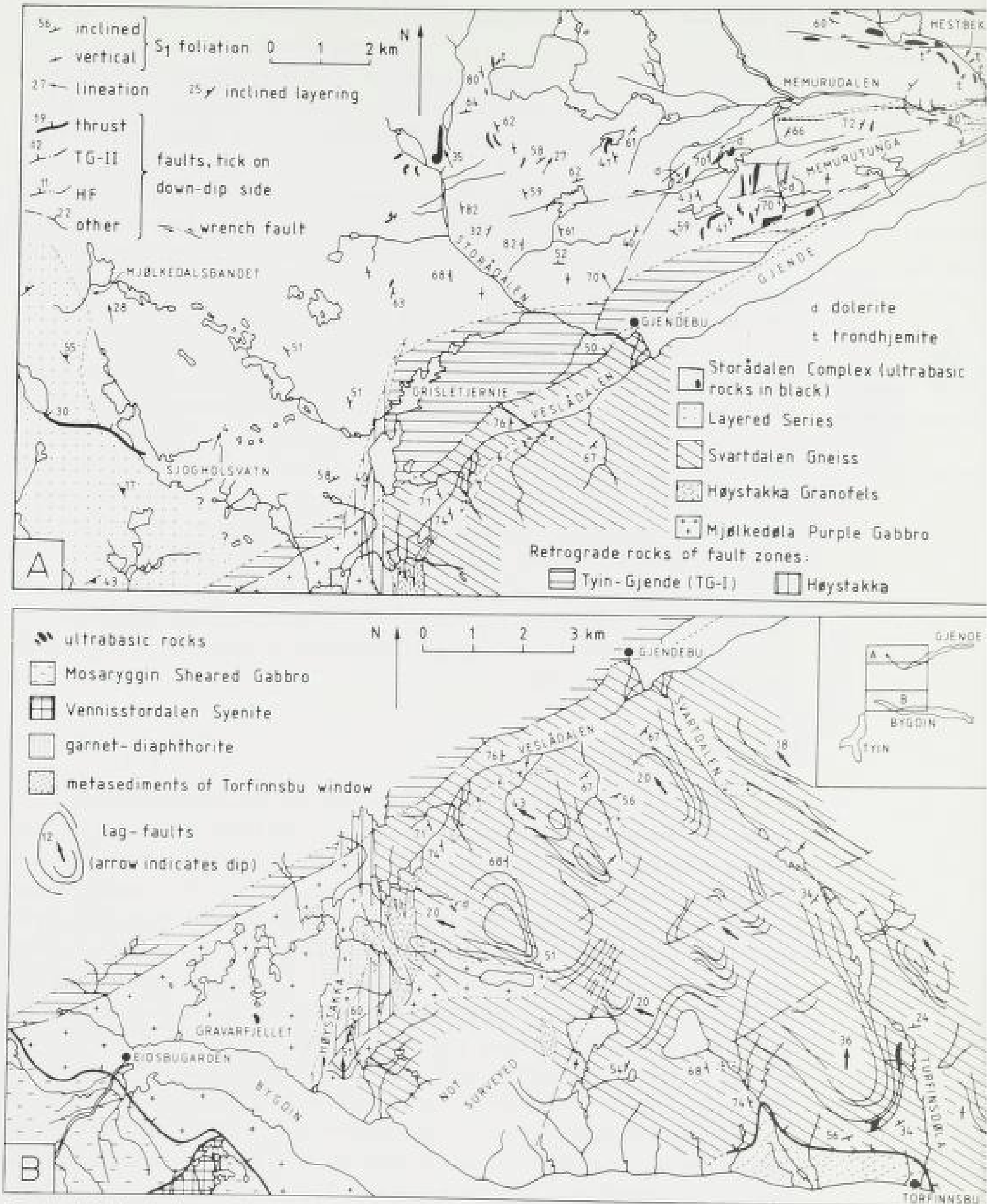


Fig. 2. (A) Geological map of the Gjendebu region north of the Tyin–Gjende Fault. For explanation of T.G.–I. and T.G.–II see text. HF = Høystakka Fault.
(B) Geological map of the Gjendebu region south of the Tyin–Gjende Fault. Other symbols as in Fig. 2A. Inset shows location and relative position of Figs. 2A and 2B.

retrogression and pegmatite injection which may be up to 3 km wide. The rock types within this zone are mylonites and ultramylonites in the sense of Sibson (1977). Several periods of movement can be recognised; for example, pegmatites may cut across the mylonitic fabric of the host rocks yet within themselves may have protomylonitic textures developed. Though undoubtedly a simplification, the movement on the fault zone can be divided into two phases. The earlier, designated T.G.–I, produces structures with a vertical or subvertical dip. The latter, T.G.–II, produces structures with a shallower dip, 40–60° to the northwest, that cut across those of T.G.–I. The products of late T.G.–II movement include rocks more strictly referred to as cataclasites and ultracataclasites. Presumably this transition from mylonite series to cataclasite series rocks reflects movement taking place at progressively shallower levels within the crust (Sibson 1977).

The two periods of movement on the Tyin–Gjende Fault are separated in time by the formation of the Høystakka Fault, a kilometre-wide zone of rocks and structures similar to those produced during T.G.–I but with dips of 40–60° to the west. Within the Høystakka Fault, mylonites and ultramylonites are well developed along discrete shear planes and these combine to give the fault zone an overall structure analogous to a stack of rock slabs dipping west with intense deformation along their tops and bases but comparatively undeformed in their central parts. Mention may be made here of the Utladalen–Gravløyfti Fault Zone which strikes parallel to the Høystakka Fault and which represents a major zone of east-southeastward dipping deformation described by Battey & McRitchie (1973). This zone must represent an earlier phase of movement since it is cut at its southern end by the Tyin–Gjende Fault (M. Heim, pers. comm. 1977) and at its northern and by the basal thrust. These observations imply a pre-Caledonian age for the Utladalen–Gravløyfti Fault Zone, but an early Caledonian age is also possible.

Both the Tyin–Gjende Fault and the Høystakka Fault have assotiated with them numerous minor faults. Unusually well developed examples occur along Memurudalen and Hestbekken (Fig. 2A). The Gjendebu region as a whole is criss-crossed by countless small faults and shear zones which form a grid with trends NE-SW and ESE-WNW. These faults cross-cut all other structures and do not have much alteration or intense mylonite/cataclasite associated with them. Though the earlier major faults have undoubtedly influenced the development of this grid, its overall appearance gives the impression that it belongs to a late, possibly the latest, period of comparatively superficial tectonics.

THRUSTS, LAG-FAULTS AND WRENCH FAULTS

Battey (1965) and Battey & McRitchie (1973) have described a series of thrusts underlying Gravarfjellet, Sløtafjellet and the high ground east of Tyin, all of which are truncated to the northwest by the Tyin–Gjende Fault. Thrusts are not common in the rest of the Gjendebu region. The base of the Mjølkedøla Purple gabbro is a thrust (McRitchie 1965), as is the boundary of the Jotun rocks as seen at Torfinnsbu (Emmett 1980b). A prominent northeast-dipping shear which outcrops northwest of Sjogholsvatnet was regarded as the boundary between the

Layered Series and what is now termed the Storådalen Complex, but the author has shown that this thrust only locally juxtaposes the two units (Emmett 1980a).

Lag-faults (shallow-dipping normal faults) are the distinctive structural element south of the Tyin–Gjende Fault and east of Høystakka, though they have been reported from elsewhere in Jotunheimen (Battey & McRitchie 1973). In Svartdalen, the lag-faults are parallel to the basal thrust as seen at Torfinnsbu (i.e. they dip to north or northwest); the sense af movement is quite clear from field data (e.g. deflection of foliation), namely that the unit above the dislocation has moved down dip. North of the Tyin–Gjende Fault, lag-faults are much less well developed, though good examples are seen in Hestbekken. Here, the lag-faults dip south and the sense of movement can be discerned from displaced pegmatites and trondhjemite bodies. In general, the lag-faults of the Gjendebu region do not have pegmatites intruded along their movement surfaces, unlike those in Visdalen (Battey & McRitchie 1973). On a general scale, all the lag-faults so far observed dip inwards towards the Tyin–Gjende Fault, an observation which further emphasises the importance of the Tyin–Gjende Fault in the regional structure of the Jotun Nappe.

Two E-W-trending wrench faults have been identified on Memurutunga; each shows a sinistral displacement of at least 4 km (Emmett 1980a) and is terminated in the east by fractures of T.G.-I age. These are the first wrench faults to be identified in Jotunheimen, though McRitchie (1965) suggested that the Tyin–Gjende Fault may have a component of wrench movement.

Petrography and internal structure

STORÅDALEN COMPLEX

The suite of rocks northwest of the Tyin–Gjende Fault and east of the Layered Series described by Battey & McRitchie (1975) were termed 'undifferentiated jotunites' by these authors. This name is inappropiate as the following description will show and so the term *Storådalen Complex* is proposed. The 'type locality' is the large valley extending north and northwestward from Gjendebu and the following may be taken as the formal description of this complex.

The Storådalen Complex consists predominantly of orthopyroxene-bearing feldspathic gneisses of jotunitic to mangeritic composition which enwrap and enclose lenses of lherzolitic and websteritic ultrabasic rocks. Apart from these latter bodies the complex shows no mineralogically distinct layering at outcrop scale or larger, but it may show either streaks or discontinuous bands produced by the concentration of dark minerals. The fabric af all the rocks is purely metamorphic but it is most clearly seen in the feldspathic rocks. The fabric is composite, consisting of both linear and planar elements of variable intensity, with the more feldspathic rocks tending to be lineated. For convenience, this fabric is denoted S_1 and it is defined by flattened fusiform aggregates of pyroxenes which were presumably formed as augen-type features that have subsequently recrystallised. The recrystallisation has affected all the rocks of the Complex and has produced an equigranular polygonal texture (Moore 1970) in which even-grained xenob-

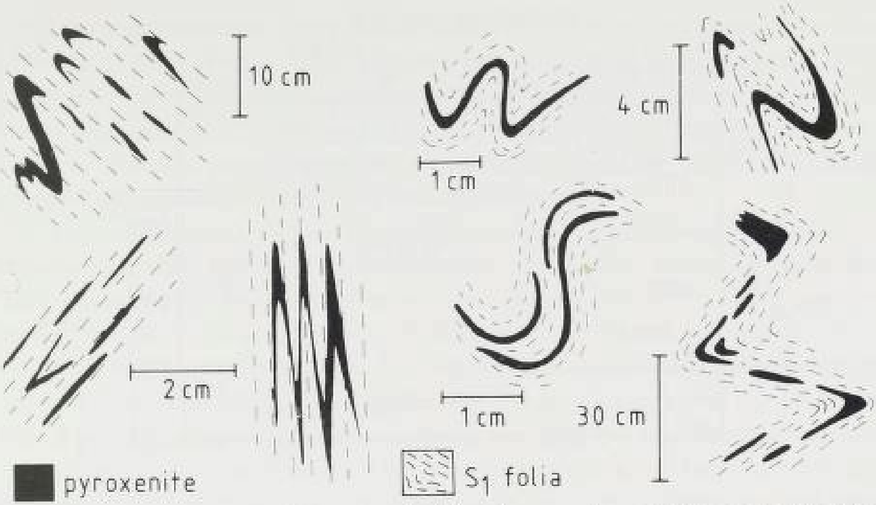


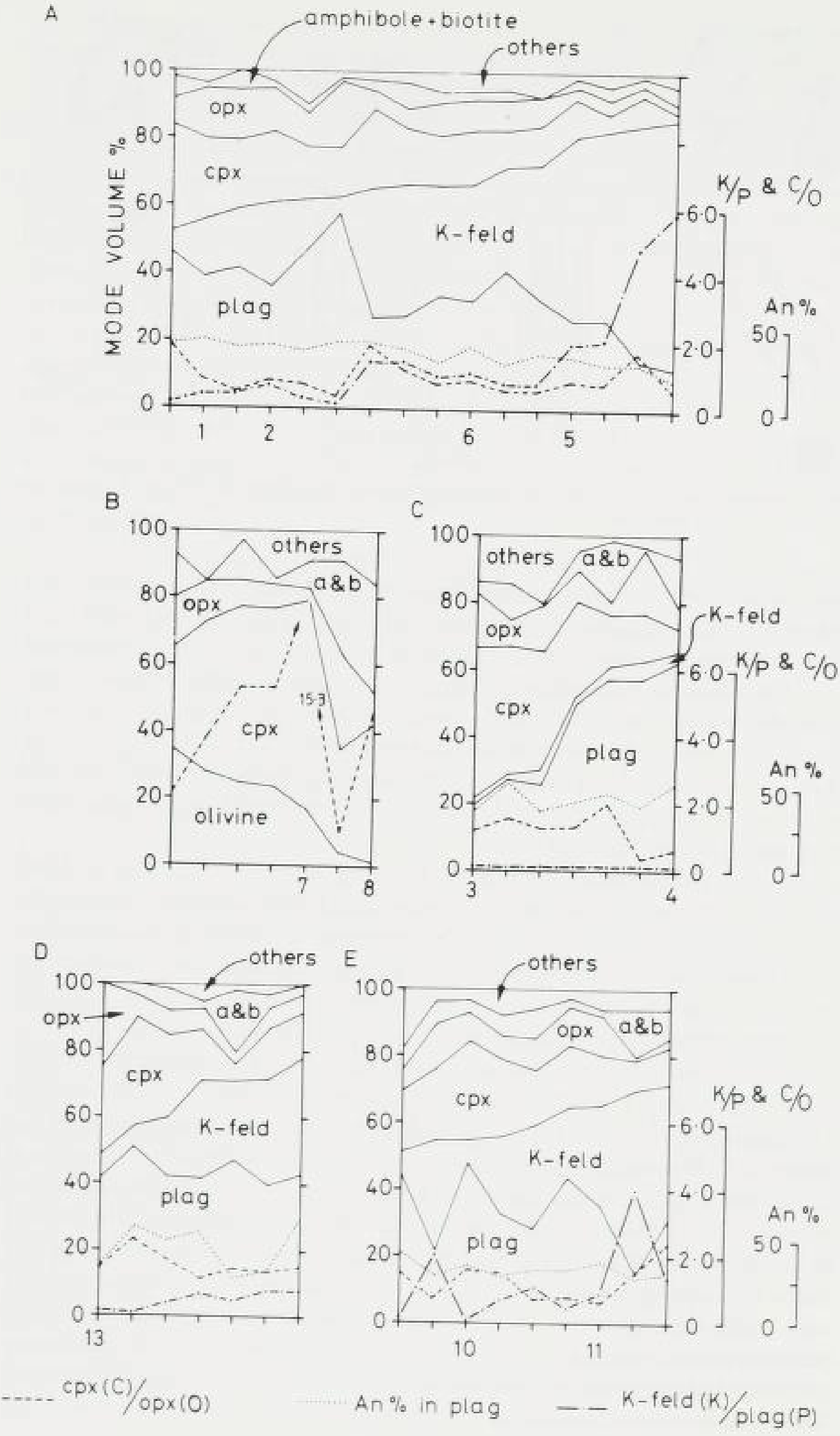
Fig. 3. Structures within the Storådalen Complex. The three examples on the left are D_1 structures, the others are D_{2A} structures.

lastic crystals of the major minerals meet along smooth, gently curved grain boundaries that originate at symmetrical triple points (Fig. 5A). This texture, the triple junction mosaic of Emmett (1980a), is considered to be the product of complete annealing recrystallisation (Kretz 1966, Vernon 1968). In many specimens this mosaic is seen to be in the process of being mylonitised, though there is usually no association retrogression. Battey & McRitchie (1975: 8–9) regarded this type of deformation as occurring at elevated temperatures ($\geqq$ 400°C) and/or in the «virtual absence» of water. Grain size within the Storådalen Complex varies from about 700 to 1200 μm.

The petrography of the Storådalen Complex has been described in detail elsewhere (Emmett 1980a), the following being a brief summary only. Modal analysis was carried out on representative samples and the results are listed in Table 3 and shown graphically in Fig. 4. On the basis of these results, the Complex can be divided into three series of the following parageneses:

(1) orthpyroxene + clinopyroxene ± olivine ± amphibole
(2) orthopyroxene + clinopyroxene + plagioclase (+ minor perthite)
(3) orthopyroxene + clinopyroxene + plagioclase + perthite

The plagioclase-dominated rocks (assemblage 2, called 'pyroxene gneisses' by Battey & McRitchie (1975) and 'anorthositic jotunites' by Emmett (1980a)) and the ultramafic rocks (assemblage 1) are regarded as cumulates and the plagioclase + perthite rocks (assemblage 3, the jotunite–mangerite series) as representatives of liquid descent. These conclusions are supported by geochemical evidence (see later). Considering the jotunite–mangerite series rocks and using colour index as a crude differentiation index, note the fairly constant clinopyroxene/orthopyroxene ratio, the smooth decrease in An content of the plagioclase and the increase in the perthite/plagioclase ratio with increasing differentiation (decreasing colour index). These trends are taken to indicate an igneous origin for these rocks.


A
amphibole + biotite
others
opx
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MODE VOLUME %
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opx
a&b
cpx
K-feld
plag
E
others
opx
a&b
cpx
K-feld
plag
K/P & C/O
An %
cpx (C)/opx (O)
An % in plag
K-feld (K)/plag (P)

Of the major minerals, clinopyroxene is usually augitic (see microprobe analyses in Table 1) and orthopyroxene, a variety pleochroic from neutral to pale pink, varies from En_{58}–En_{65} (Table 1). Plagioclase is usually antiperthitic, the mineral which Goldschmidt (1916) regarded as characteristisc of the Bergen–Jotun kindred, shows a large degree of petrographic variability. Emmett (1980a) grouped the perthite morphologies into three classes:

(1) mosaic perthites, the commonest group with regularly disposed exsolution lamellae and often with a clear rim occupying c. 10% of the width of each crystal. They are so named because they occur as an integral part of the triple junction mosaic, and they appear to be fairly pure albite-in-orthoclase perthites.

(2) perthite porphyroblasts, which are usually amoeboid in shape with irregular, often diffuse, margins and lacking the clear marginal portions so typical of the mosaic perthites. Lamellae of the two feldspar phases are equally abundant and extend right to the edge of the grains. Refractive index measurements indicate that the plagioclase component has a composition of An_{12}–An_{17}.These perthites have all the properties usually assigned to mesoperthosites (Michot 1961).

(3) patch or replacement perthites (Smith 1974). These have ragged or irregular 'islands' of plagioclase (similar in composition to the normal plagioclase in the host rock) set in a 'sea' of perthitic orthoclase. The morphology of the host perthite is similar to that of the mosaic perthites. Patch perthites are the least common of the three groups of perthite.

Accessory minerals include biotite and apatite in the feldspathic rocks, whilst zircon appears in the most felsic members. Green spinel (pleonaste – hercynite) is commom in ultramafic rocks and anorthositic jotunites. Biotite, a pale yellow to deep brown pleochroic variety, commonly forms granular rims around opaque grains embedded in feldspar aggregates, a feature which Sederholm (1916) and Parsons (1980) ascribe to sub-solidus reaction under localised conditions of high iron and water contents. Amphibole, usually pargasitic in composition, is common in rocks with little or no perthite.

The ultramafic rocks within the Storådalen Complex occur in all sizes from thin discontinuous bands a few centimetres thick to large lens-shaped masses several tens of metres wide and hundreds of metres long. Where olivine-bearing rocks occur against feldspathic rocks, a zone of spinel–pyroxenite is developed. Though plagioclase is rare in the mode of olivine-bearing rocks, when the two minerals do occur in contact they are seen to be in a reaction relationship, an observation which has important petrogenetic significance. An unusual occurrence of plagioclase in the ultramafic rocks is as granular rims around ilmenite grains. These haloes

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Fig. 4. Variations in the modal composition of rocks from the Gjendebu region. A: jotunite-mangerite rocks from the Storådalen Complex. B: ultrabasic rocks from the Storådalen Complex. C: anorthositic jotunites (Emmett 1980a) from the Storådalen Complex. D: Mjølkedøla Purple Gabbro. E: Svartdalen Gneiss. Specimens are arranged horizontally in order of decreasing colour index. Numbered examples correspond to analyses given in Table 3 (full data available from the author). Abbreviations: a, amphibole, b, biotite, cpx, clinopyroxene, K – feld, potash feldspar (including perthite), opx, orthopyroxene, plag, plagioclase (including antiperthite). 'Others' includes alternation products, opaque phases and spinel.

TABLE 1. Representative analyses of co-existing pyroxenes

	1	2	3	4	5	6	7	8
Orthopyroxenes								
SiO_2	51.4	52.9	52.4	53.1	52.1	51.1	51.7	52.1
Al_2O_3	1.28	1.82	3.68	2.78	4.44	3.83	1.99	0.84
FeO^*	23.30	22.17	17.04	20.83	17.01	21.35	23.84	22.35
MgO	22.54	22.07	26.08	23.19	25.53	22.05	21.87	21.47
CaO	0.67	0.43	0.50	0.28	0.52	0.49	0.55	0.63
MnO	0.72	0.72	0.45	0.56	0.45	0.06	0.70	2.09
Na_2O	–	–	0.29	–	0.07	0.49	–	0.08
TiO_2	–	–	n.d.	–	0.10	0.09	–	–
Cr_2O_3	–	–	–	–	–	0.03	n.d.	–
total	99.92	100.10	100.34	100.70	100.18	100.03	100.65	99.58
Ca^{2+}%	1.34	0.89	1.01	0.57	1.06	1.02	1.11	1.31
Mg^{2+}%	62.44	63.38	72.44	66.11	72.02	64.14	61.34	62.30
Fe^{2+}%	36.22	35.73	26.55	33.32	26.92	34.84	37.55	36.39
Clinopyroxenes								
SiO_2	49.4	49.3	48.9	51.0	51.3	49.2	50.4	52.2
Al_2O_3	2.52	2.69	5.46	5.33	6.15	5.84	3.57	1.66
FeO^*	9.25	11.86	7.32	7.65	6.67	8.72	9.67	9.27
MgO	14.07	13.34	13.48	12.77	12.54	12.65	12.46	13.24
CaO	21.87	21.42	23.75	21.33	20.52	21.58	21.44	21.95
MnO	0.30	0.21	0.18	0.25	0.18	0.29	0.27	0.82
Na_2O	–	–	0.54	1.32	1.64	0.97	0.70	0.74
TiO_2	–	–	n.d.	0.40	0.49	0.73	0.35	–
Cr_2O_3	–	–	–	–	–	0.05	–	–
total	97.41	98.83	99.61	100.08	99.48	99.98	98.86	99.90
Ca^{2+}	44.96	42.31	49.26	47.35	47.54	46.94	46.27	46.11
Mg^{2+}	40.18	38.48	38.89	39.40	40.40	38.27	37.40	38.69
Fe^{2+}	14.86	19.21	11.86	13.25	12.06	14.79	16.33	15.20

Table 1. Representative analyses of co-existing pyroxenes. All figures are in weight % and all the analyses were made by electron microprobe (full details in Emmett 1980a). The full data set is available from the author. Host rocks are as follows:
Column 1: Svartdalen Gneiss, sample 12. From near Langedalsbre, grid reference 7246 1017.
Column 2: Svartdalen Gneiss, sample 19. From Svartdalen, grid reference 7353 1205.
Column 3: pyroxenite from Turfinsdøla. Grid reference 7830 0540. Sample 17.
Column 4: anorthositic jotunite from Høgtunga, sample 48. Grid reference 7140 1647.
Column 5: anorthositic jotunite from Høgtunga, sample 50. Grid reference 7089 1678.
Column 6: jotunite, sample 61, from north of Grisletjernie. Grid reference 6806 1405.
Column 7: mangerite from northwest side of Storådalen, sample 64. Grid reference 6930 1441.
Column 8: Høystakka Granofels sample 83, from Høystakka. Grid reference 6810 0835.
Columns 4 to 7 are samples from the Storådalen Complex.
FeO^* = total Fe as FeO.

of plagioclase (Fig 5B) are very calcic (c. An_{90}) and are thought to be derived by exsolution of plagioclase from pyroxene (Emmett 1980a), but why the feldspar prefers to nucleate on oxide grains is not understood. All the ultramafic rocks carry a deep grass green spinel, again of the hercynite type, which occur as discrete crystals or as vermicular intergrowths with orthopyroxene. Amphibole (pargasite) is a major phase in some of the ultramafic rocks.

Structurally, the most conspicuous feature of the Storådalen Complex is the S_1 fabric. This seems to have formed by isoclinal folding and subsequent limb-

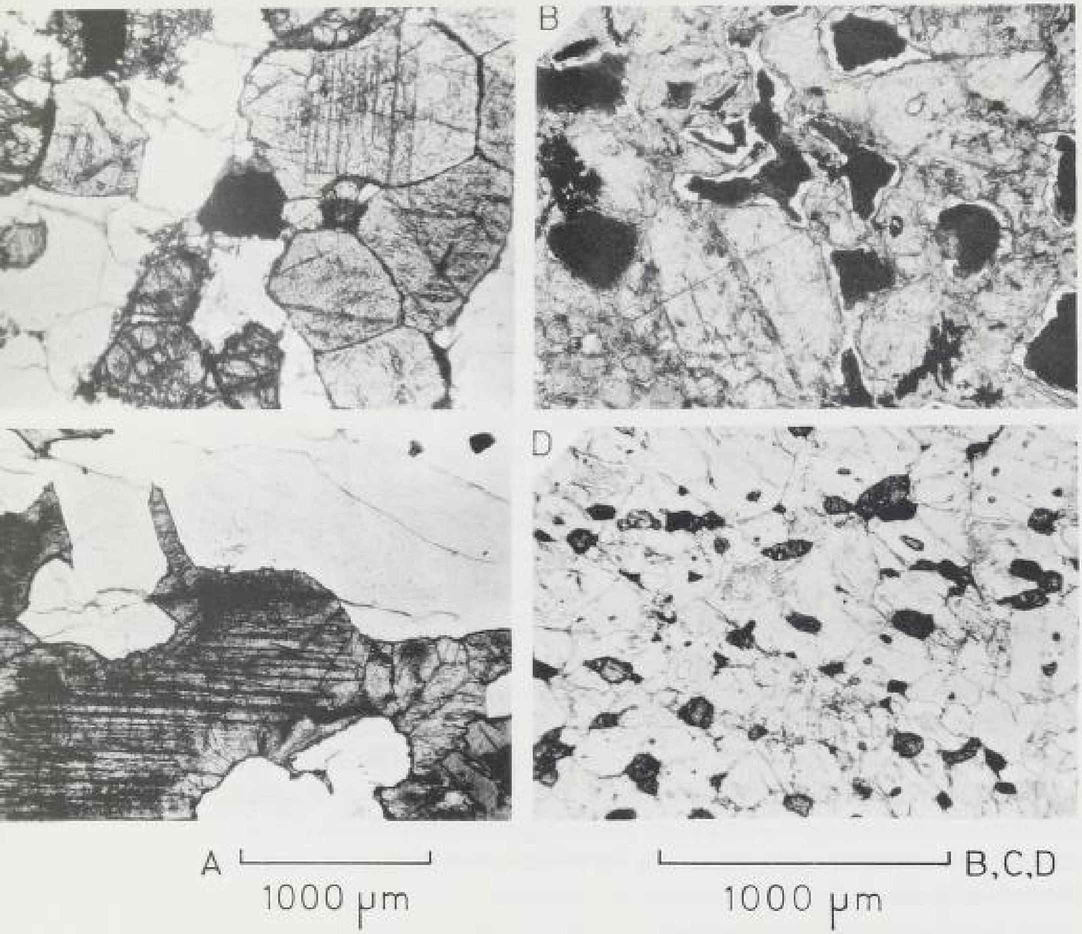

Fig 5. (A) Typical texture of pyriclasite from the Storådalen Complex. Note the smooth gently curved grain boundaries meeting at symmetrical triple points. Minerals present are plagioclase, pyroxene, opaque phases and a little biotite.
(B) Opaque grains, mainly ilmenite, surrounded by 'haloes' of plagioclase (approximately An_{90}). These are embedded in a pyroxene mosaic, with a little biotite and amphibole also present. From a pyroxenite within the Storådalen Complex. (C) Texture of the Mjølkedøla Purple Gabbro showing euhedral to subhedral plagioclase with interstitial pyroxene.
(D) Texture of Høystakka Granofels, showing crystals of pyroxene dispersed througout a perthite + quartz mosaic. Apatite and opaque grains are also visible. All photographs taken in plane-polarised light.

attentuation of a very poorly preserved pre-S_1 planar fabric, S_0. This earlier fabric is seen only in rare fold noses between discrete S_1 folia (Fig.3), and its exact nature is unknown. It may be original igneous layering (Battey 1965, Battey & McRitchie 1975) or an earlier metamorphic foliation. The pods and streaks of mafic and ultramafic material occur in arrays which give the impression of being intensely boudinaged layers, and S_1 is always concordant with such bodies. Accordingly, S_0 is interpreted as initial igneous layering disrupted during the S_1-producing event, D_1, and enhanced (at least on a small scale) during subsequent high-grade metamorphism. However, S_0 relics, including possibly the larger ultrabasic bodies, are now totally recrystallised within the S_1 fabric and the two planar elements are transposed. Emmett (1980a) has demonstrated that, if the ultrabasic bodies do represent disrupted layers then the metamorphism which generated the characteristic pyroxenitic sheaves between olivine-bearing and plagioclase-bearing rocks (Carstens 1920, Battey 1960, Battey & McRitche 1975) must have post-dated their disruption. Post-D_1 recrystallisation has preserved the S_1 fabric but individual crystals develop a xenoblastic habit. In the Storådalen Complex, the mineralogy shows that this recrystallisation occurred under pyroxene–granulite facies conditions.

Structures of two separate events which fold S_1 can be recognised, neither of which generate a post-S_1 fabric. Both are poorly developed and they are regarded as contemporaneous, though there is no real evidence for this. This latter deformation, D_2, has not caused any retrogression of the high grade D_1 assemblages. Folds termed D_{2A} have amplitudes and wavelengths of a few centimetres only and form intrafolially to S_1 (Fig. 3). They may be distinguished from S_1 folds by their less tightly appressed nature and by the fact that the S_1 fabric is not necessarily axial planar to them. D_{2B} folds are steeply plunging and appear to control the field disposition of S_1 (see Fig. 2A). Though persistent marker horizons are absent, the wavelength of these folds is estimated to be up to several hundred metres. They may well be equivalent to McRitchie's F_3, though on a somewhat reduced scale (McRitchie 1965, Battey 1965).

Following D_2, the rocks of the complex were subjected to a period of tension during which pink and white feldspar pegmatites were injected. These cut across all D_1 and D_2 structures and they are possibly associated with the pegmatite injections developed in conjunction with the formation of the Tyin–Gjende and Høystakka Faults.

SVARTDALEN GNEISS

The rock units south of the Tyin–Gjende Fault, the Svartdalen Gneiss, Mjølkedøla Purple Gabbro and the Høystakka Granofels, are much less intensely deformed than the Storådalen Complex. In the Svartdalen Gneiss, no folds comparable to D_1 or D_{2A} structures have been observed. There is a generally steeply inclined foliation and it is possible that variations in the strike of this (Fig. 2B) may be due to presence of gentle D_{2B} folds. The only structure recognised in the Purple Gabbro is a weak foliation developed in proximity to its junction with the Svartdalen Gneiss. The late tensional pegmatite injection phase is present in all the units but is not so extensive as in the Storådalen Complex. The foliation in the Svartdalen Gneiss is termed S_1 and its production is correlated with D_1. It is believed that the variation in intensity of D_1 is related to the crustal level of the rocks at the time when they were deformed.

The Svartdalen Gneiss is a foliated, two-pyroxene, two-feldspar gneiss of jotunitic to mangeritic composition which outcrops east of Høystakka to at least the line of Turfinsdøla (Fig. 2B). It is aphyric and equigranular, but lacks the pervasive annealed texture of the Storådalen Complex. Its fabric is purely planar but is commonly irregular and discontinuous. Representative modal analyses are listed in Table 3 and shown graphically in Fig. 4E. A full description appears in Emmett (1980a), but of particular note here is the occurrence of perthite which is analogous to the mosaic perthites of the Storådalen Complex, though there is a tendency in some specimens for it to be amoeboid in habit and to fill the interstices formed by the mis-fit of plagioclase grains. Biotite and apatite are the main accessory minerals, but amphibole and green spinel are absent.

The transition from Svartdalen Gneiss to Mjølkedøla Purple Gabbro occurs gradually and irregularly east of Høystakka, but its expression is complicated by the presence of the Høystakka fault and the Høystakka Granofels. Small enclaves of Purple Gabbro are found within the Svartdalen Gneiss and the fresh and unaltered nature of these mitigates against their being tectonic intercalations. The conversion of Purple Gabbro into Svartdalen Gneiss is essentially a textural transmutation, involving the reduction of phenocrysts and the production of a foliation. The clear inference is that the Svartdalen Gneiss represents a deformed portion of the Mjølkedøla Purple Gabbro body.

MJØLKEDØLA PURPLE GABBRO

The Mjølkedøla Purple Gabbro was first described by Battey (1965), though specimens of it were used by Goldschmidt (1916) as examples of his 'normal' gabbro. Battey's brief description was subsequently expanded by McRitchie (1965) and Emmett (1980a). The unit outcrops from Sløtafjellet in the west to Høystakka in the east, where it begins its transition into the Svartdalen Gneiss. Its base, a zone of shearing interpreted by Battey & McRitchie (1973) as a thrust, is seen only in the most extreme southwest corner of the Gjendebu region (Fig. 2B). Full petrographic descriptions of the Purple Gabbro will be found in the works cited above, but representative modal analyses are presented in Table 3 and Fig. 4D. Note that Battey does not mention perthite in his description, hence 'gabbro', but the author has found K-feldspar-bearing varieties of essentially jotunitic composition intimately associated with true gabbros. Rather than rename this unit and thus introduce into the literature two names for the same unit, Battey's name is retained though it is conceded that it is not entirely gabbroic in composition. The K-feldspar is usually perthitic and occurs interstitially, occasionally poikilitically enclosing small grains of pyroxene. The association of orthopyroxene and clinopyroxene is thought to be igneous, not metamorphic (Battey 1965), and apatite is an abundant accessory mineral. Igneous textures are characteristic of the 'gabbro' and include ophitic and subophitic intergrowths, and euhedral phenocrysts of plagioclase with interstitial pyroxene (see Fig. 5C and Emmett 1980a). Biotite and, more rarely, garnet, may occur as thin rims growing on oxide grains. Garnet is only common in extensively retrogressed examples (see later).

ULTRABASIC ROCKS SOUTH OF THE TYIN–GJENDE FAULT

The bodies of ultrabasic rock associated with the Mjølkedøla Purple Gabbro in the region of Eidsbugarden have been described by McRitchie (1965), but the bodies on Gravarfjellet and Turfinstindane (Fig 2B) are not included in this work. The body on the south flank of Turfinstindane has an apparently unsheared contact with the Svartdalen Gneiss which cuts across the S_1 fabric in the host. This suggests that the body was emplaced (?intruded) after the main D_1 event, but the presence of disrupted and boudinaged compositional layering within the body indicates that deformation continued after its emplacement. This observation constitutes a major point of distinction between this body and those within the Storådalen Complex since the latter appear to have been emplaced before the onset of D_1.

There are two ultrabasic bodies on Turfinstindane and they are thought to represent a single mass disrupted by lag-faulting. They are composed mainly of wehrlite (orthopyroxene + clinopyroxene), though olivine and amphibole may be abundant locally. Plagioclase occurs as finely granular rims around opaque grains or, significantly, as small discrete subhedral grains, an occurence not recorded in the ultrabasic rocks of the Storådalen Complex. Green spinel is a rare trace mineral. Texturally, the wehrlites have a coarse-grained granoblastic mosaic with well-developed triple junctions. The Gravarfjellet body is composite in nature, being formed of several coalescing smaller bodies. The dominant rock type is lherzolite. A typical modal analysis is listed in Table 3.

HØYSTAKKA GRANOFELS

The Høystakka Granofels is distinct in both its mineralogy and its distribution, and it is surprising that its existence has not been noted previously. It occurs as numerous subvertical dykes which vary in thickness from a few centimetres to several metres, the dykes splitting and coalescing to give a body of very complex form. The representtion of this on the map (Fig. 2B) is only approximate. The granofels occurs entirely to the east of the Høystakka Fault and none has been found north of the Tyin–Gjende Fault. The granofels is not foliated and the dykes cut across the S_1 fabric in the Svartdalen Gneiss. Within the Høystakka Fault, slabs of Purple Gabbro veined with granofels have been found.

A full description has been given elsewhere (Emmett 1980a), but representative modes are given in Table 3. The rock conforms with Goldsmith's original definition of a granofels (Goldsmith 1959) and it consists mainly of quartz and perthite with minor orthopyroxene and clinopyroxene (Fig. 5D). Apatite and zircon are rare accessory minerals. Coarse myrmekitic textures are common and often develop at the expense of perthite. Plagioclase occurs as sparse, much corroded relics only. The pyroxenes are low in alumina and apparently quite free of inclusions and exsolution lamellae.

Metamorphism

It is possible to divide the metamorphic evolution of the rocks of the Gjendebu region into two stages; an earlier high-grade event and a later retrogressive event. The retrogression is very patchy in its representation, resulting from shearing and hydration during fault development or during Caledonian thrusting (Emmett 1980b). Full details are given in Emmet (1980a), the following listing the most important points and conclusions only.

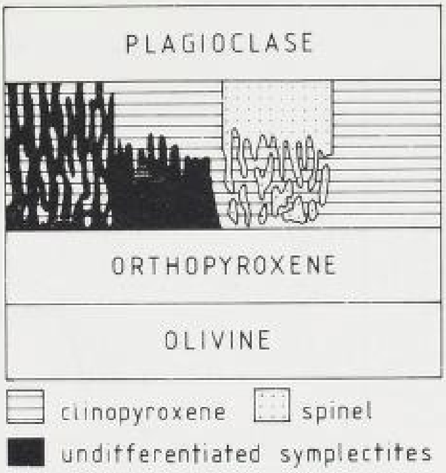

Fig. 6. Schematic diagram showing the form of the four types of corona present in the Storådalen Complex.

HIGH-GRADE METAMORPHISM

Evidence of mineralogical reaction during the earlier metamorphic event is confined to the ultrabasic and basic representatives of the Storådalen Complex. As noted by Carstens (1920) and Battey (1960), olivine-bearing rocks never occur in direct contact with plagioclase-bearing rocks. Where they are seen in close proximity, a zone of pyroxenite intervenes (Fig. 6). Carstens proposed that this was the result of some form of eutectoid crystallisation but Battey (1960) recognised it as the product of subsolidus reaction, a view not subsequently challenged. In addition, major element geochemistry suggests that the more basic of the anorthositic jotunites were originally olivine-bearing cumulates. It is clear that the high-grade metamorphism has resulted in the removal of olivine + plagioclase assemblages and the generation of spinel + orthopyroxene + clinopyroxene assemblages. These observations, taken together with the fact that there are normally no garnetiferous assemblages present in the Storådalen Complex, indicate that the conditions of metamorphism are constrained by two well known mineralogical reactions:

(1) olivine + plagioclase → orthopyroxene + clinopyroxene + spinel.

(2) spinel + orthopyroxene + clinopyroxene → garnet + pyroxene.

Of these, (1) places a limit on the minimum values of P (and T), whilst (2) constrains the maximum. Griffin & Heier (1973) considered these reactions in detail and found that progress from left to right is accompanied by a reduction in specific volume, a result which suggests that they are driven in that direction by an increase in pressure. The positive slope to the reaction boundaries in P–T space thus suggested was supported by early experimental work (e.g. Green & Ringwood 1967) and petrographic data (Gardner & Robins 1974). Reaction boundaries with such attitudes permit the generation of the mineral assemblages of central Jotunheimen by simple post-emplacement isobaric cooling, with the production of garnetiferous assemblages forestalled by a terminal decompression event (Griffin 1971a, and Fig. 7). Herzberg (1975) determined the reaction boundaries as being essentially isobaric; curves of this disposition require an

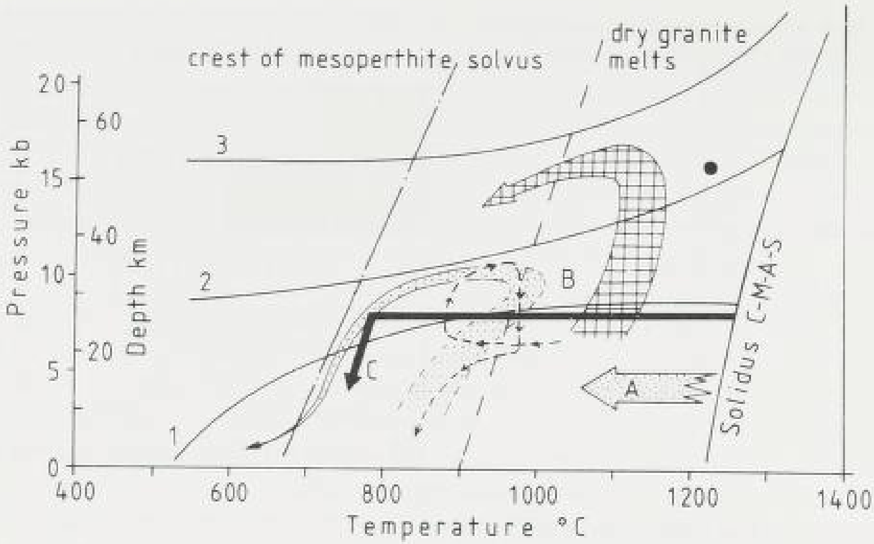

Fig. 7. The stippled arrow shows a proposed path through P–T space for the Storådalen Complex. A = approximate conditions of initial magmatic consolidation. B = climax of post–D_1 high-grade metamorphism. C = terminal decompression event. C–M–A–S is the CaO–MgO–Al_2O_3–SiO_2 system (O'Hara 1967). The reaction boundaries are from Herzberg (1975, 1978) and represent the following reactions:

1. olivine + plagioclase → plagioclase + orthopyroxene + spinel
2. plagioclase + orthopyroxene + spinel → garnet + orthopyroxene
3. clinopyroxene + orthopyroxene + spinel → garnet + olivine

The solid arrow represents the simple decompression model of Griffin (1971a) whilst the broken arrow is the P–T path suggested by Battey (1978). The hatched arrow is the P–T path for the Indre Sogn anorthosites proposed by Herzberg (1975). The black circle represents the peak of Scourian metamorphism, as suggested by O'Hara (1977).

increase in either P alone (Battey 1978) or in both P and T (Herzberg 1975) to generate the observed assemblages. Subsequent work by Herzberg (1978) has reintroduced positive slopes to the curves, albeit somewhat shallower than those required by the earlier work. In this discussion it is important to note the conclusion of Emmett (1980a) that the main D_1 deformation precedes the main high-grade metamorphism. With this in mind it is unlikely that simple isobaric cooling could generate the features seen since a period of intense deformation must intervene between initial magmatic consolidation and the metamorphism.

In Fig. 7 a proposed path in P–T space is inscribed on the reaction boundaries of Herzberg (1978). For comparison, the paths suggested by Griffin (1971a) and Battey (1978) are also shown. The new curve follows that of Battey (1978) in that it envisages metamorphism beginning at rather low P (c. 7kb) and T (c. 850°C), though Battey's curve is rather unrealistically precise. Both curves place the peak of metamorphism at c. 1000°C, 8–10 kb. It is assumed that post-climactic annealing combined with the D_1 deformation removed all evidence of the original textural condition of the rocks. The mosaic produced presumably contained homogeneous, i.e. unexsolved, feldspars (which were eventually to unmix to form the mosaic type perthites and antiperthites) and pyroxenes.

TABLE 2. Analyses of garnets, amphiboles, and relict clinopyroxenes from retrogressed rocks.

	1.	2.	3.	4.
SiO_2	38.08	38.0	42.0	51.6
Al_2O_3	21.34	22.51	13.56	2.40
FeO^*	26.32	22.93	14.97	8.51
MgO	5.49	9.00	10.87	13.40
CaO	7.75	5.76	11.23	22.19
MnO	1.17	1.48	0.06	0.42
Na_2O	–	–	1.82	0.74
TiO_2	0.15	–	1.48	0.27
K_2O	–	–	1.68	–
Cr_2O_3	–	–	0.03	–
total	100.35	99.68	97.54	99.53
alm %	55.75	47.93		Ca^{2+}% 46.70
pyr%%	20.71	33.53		Mg^{2+}% 39.26
gros %	21.03	15.41		Fe^{2+}% 14.04
spes %	2.51			

FeO^* = total Fe as FeO.

Table 2. Average analyses of garnet, amphibole, and relict clinopyroxenes from retrogressed rocks. All analyses by electron microproble, values in weight %.

Column 1: Average of four analyses of garnet from retrogressed Mjølkedøla Purple Gabbro, sample 82; from near Brønebergtjern, grid reference 6785 0780.

Column 2: garnet (average of two analyses) from retrogressed anorthositic jotunite. Sample 48, from Høgtunga, grid reference 7140 1647.

Column 3: secondary amphibole (pargasite) from retrogressed Purple Gabbro. Average of two analyses. Sample 82.

Column 4: relict clinopyroxene (average of two analyses) from retrogressed Purple Gabbro, sample 82.

The only evidence extant for the reversal of reaction (1) with the subsequent exhumation of the Storådalen Complex consists of the limited development of symplectites by the re-equilibration of aluminous pyroxene to form either low-Al pyroxene, spinel and plagioclase (Griffin 1971a), or to form olivine and plagioclase (Battey & Davison 1977). The restricted nature of these decompressive reactions is widely believed to be the result of the high-grade rocks being carried upward too rapidly to allow 'normal' mineralogical reactions to proceed (Griffin 1971a and b, Battey & McRitchie 1975, Battey 1978, Emmett 1980a). Battey (1978) has deflected his decompression curve to allow for several thermal perturbations that have affected the rocks during uplift. Some of these events are of only local significance and so need not be considered further. However,one such event, the generation of feldspathic partial melts, will be consider later.

In contrast to the Storådalen Complex, the rocks to the south of the Tyin–Gjende Fault show a relative lack of diversity and this, especially the lack of cumulate rocks, makes it difficult to define their initial high-grade metamorphism. What cumulate rocks there are – the deformed ultrabasic bodies of Gravarfjellet and Turfinstindane and those associated with the tectonised base of the Mjølkedøla Purple Gabbro – appear to contain stable olivine + plagioclase assemblages (McRitchie 1965, Emmett 1980a). Taken together with the typical granulite facies assemblage of the Svartdalen Gneiss, this suggests that these rocks have not exceeded low-pressure granulite facies grade. Battey (1965) proposed that the Mjølkedøla Purple Gabbro is of amphibolite facies grade. Overall, it seems

TABLE 3. Representative whole rock and modal analyses of rocks from the Gjendebu region

	1	2	3	4	5	6	7	8	9	10	11	12+	13	14
SiO_2	48.4	53.4	40.6	46.5	58.5	55.1	46.3	43.4	45.4	51.8	51.5	51.6	54.6	57.6
Al_2O_3	14.56	16.98	10.14	20.81	16.80	17.37	3.63	10.42	7.51	16.05	15.74	23.93	19.73	19.77
Fe_2O_3	4.51	3.46	10.20	5.15	2.27	2.55	3.51	6.50	5.01	3.92	3.39	1.93	2.65	2.28
FeO	7.40	5.33	10.72	5.61	4.07	4.88	8.71	9.60	6.68	6.46	6.59	3.73	4.08	2.96
MgO	8.26	5.24	10.96	4.57	2.95	4.15	20.20	12.84	13.55	6.46	6.42	3.47	3.02	1.86
CaO	11.01	8.74	13.74	12.60	6.47	7.44	15.50	13.47	19.72	9.14	9.25	8.47	6.55	4.21
Na_2O	2.62	3.65	0.90	3.13	3.98	3.63	0.44	0.81	0.97	3.63	3.21	4.11	4.69	5.28
K_2O	1.20	1.92	0.15	0.28	3.36	3.61	0.20	0.23	0.13	1.14	2.04	0.89	2.68	3.28
TiO_2	0.81	0.77	1.39	0.70	0.91	0.74	0.39	0.88	0.74	0.71	0.87	0.36	0.75	0.76
MnO	0.34	0.33	0.26	0.12	0.11	0.14	0.19	0.20	0.19	0.20	0.19	0.07	0.15	0.14
P_2O_5	0.20	0.16	0.64	0.88	0.20	0.36	0.02	0.03	0.03	0.40	0.45	0.22	0.57	0.25
H_2O^+	0.13	0.45	0.38	0.55	<0.05	0.14	0.33	0.66	0.25	0.29	0.11	<0.05	0.43	<0.05
H_2O^-	0.33	0.14	0.12	0.18	<0.05	0.21	<0.05	0.47	0.20	<0.05	0.43	0.28	0.38	<0.05
(CO_2)	0.20	0.18	0.48	0.14	0.51	0.16	0.53	0.89	0.48	n.d.	0.15	0.45	n.d.	1.00
total	99.97	100.47	100.20	100.53	100.23	100.48	100.00	100.40	100.86	100.20	100.34	99.56	100.42	99.49
D.I.	22.29	42.35	8.49	27.89	59.49	52.00	4.19	8.19	5.04	37.38	39.21	40.56	55.46	66.15
opx	13.8	12.6	17.7	6.0	3.6	8.6	4.0	9.2	9.0	9.0	12.0	–	1.5	0.7
cpx	24.4	20.9	44.2	8.0	9.9	15.8	61.0	40.9	69.0	29.0	15.0	1.0	4.4	2.1
plag*	38.8	36.2	19.8	62.0	25.7	32.4	olv=	olv=	olv=	48.0	35.0	36.6	47.0	–
K-feld**	17.0	24.9	1.2	3.0	56.2	35.2	18.0	1.1	1.0	7.0	31.0	14.1	24.4	65.5
amph + biot	2.0	1.5	1.5	14.0	2.0	2.9	13.0	33.7	17.0	3.0	1.0	29.5	19.6	4.4
opaques + spinel	3.4	3.6	11.1	5.0	2.5	3.7	4.0	8.0	3.0	3.0	4.0	1.8	1.8	3.8
An% in plag	41	38	35	51	32	38	–	–	–	35	37	34	24	–
cpx/opx	1.8	1.7	2.5	1.3	2.8	1.8	15.3	4.5	7.7	3.2	1.3	–	2.9	3.0
K-feld/plag	0.4	0.7	0.1	0.1	2.03	1.1	–	–	–	0.1	0.9	0.39	0.5	–

* includes antiperthite
** includes perthite
\+ heavily retrogressed
\+\+ only in 12, 14 and 15

Additional Components++:

	12	13	14
quartz	2.9		35.0
garnet	10.5		–
apatite	1.1		1.2
		zircon	0.4

Table 3. Representative whole rock and modal analyses of rocks from the Gjendebu region. The full data set is available from the author. The analytical methods, a combination of A.A.S., colorimetric, and gravimetric techniques, are described in Emmett (1980a). All oxide values are in weight %, the modes are in volume %. The (CO_2) figure is nominal and is, strictly speaking, the loss on ignition at 850°C not accounted for by H_2O (determined separately) and corrected for the oxidation of FeO. D.I. is the differentiation index of Thornton & Tuttle (1969).

Storådalen Complex

1. jotunite, sample 61. See Table 1.
2. mangerite, sample 64. See Table 1.
3. anorthositic jotunite, sample 28. From near Høgtungatjern. Grid reference 7123 1597.
4. anorthositic jotunite, sample 33, from Høgtunga. Grid reference 7240 1261.
5. mangerite, sample 137, from near Memurubu. Grid reference 7935 1780.
6. mangerite, sample 62, from north of Grisletjernie. Grid reference 6773 1448.
7. lherzolite, sample 5, from Hellerfossen. Grid reference 6908 1690.
8. websterite, sample 236, from Lågtungetjern. Grid reference 7549 1653.

Other rocks

9. pyroxenite, sample 17. See Table 1.
10. Svartdalen Gneiss, sample 12. See Table 1.
11. Svartdalen Gneiss, sample 19. See Table 1.
12. Mjølkedøla Purple Gabbro, sample 82. See Table 2.
13. Mjølkedøla Purple Gabbro, sample 329. From near Langedalstjern, grid reference 7483 0517.
14. Høystakka Granofels, sample 83. See Table 1.
15. Høystakka Granofels, sample 131. From near Uksedalstjern. Grid reference 6866 0735.

unlikely that the rocks south of the Tyin–Gjende fault have exceeded P and T conditions delimited by reaction (1) above.

RETROGRADE METAMORPHISM

Retrogression of the high-grade assemblages occurs in association with (1) fault and shear zone development, (2) thrusts and lags, and (3) the margins of pegmatites. The widest zones occur in conjunction with the Tyin–Gjende Fault and immediately to the west of the Høystakka Fault (Fig. 2B). Full details of the diapththoretic rocks are given in Emmett (1980a), and a brief description of the Caledonian retrogression of the Svartdalen Gneiss is given in Emmett (1980b).

In general, retrogression is characterised by the progressive replacement of pyroxene, especially orthopyroxene, by amphibole and the development of epidote-bearing aggregates from plagioclase. Garnet (rich in almandine and pyrope, see Table 2) and, rarely, scapolite may also be developed, usually after the main amphibolisation of the host rock. In the zone of retrogressed Puple Gabbro west of Høystakka, clinopyroxene is preserved only as ragged relics and euhedral porphyroblasts of garnet are common. The diaphthoretic assemblages are typically those of the epidote–amphibolite (transitional greenschist-amphibolite) facies (Turner 1968), though amphibolite facies may be developed in places. There are probably several periods of retrogression ranging in age from Precambrian to late Caledonian, but criteria for distinguishing individual events are not yet available.

Geochemistry

WHOLE ROCK ANALYSES

Representative whole rock analyses are listed in Table 3, the complete data set being available in Emmett (1980a). Fig. 8a plots oxide abundance against the Thornton–Tuttle differentiation index (= D.I., see Thornton & Tuttle 1960). For the Storådalen Complex, the curves are continuous and indicate the presence of both cumulate and liquid descent trends. Scatter of the points in the CaO, MgO, and total Fe (as Fe_2O_3) curves at D.I. <30 is correlated with a steep initial increase in alumina up to D.I. = 30 and an initially antithetic relationship between total Fe and MgO (Fig. 8B). It is considered that rocks with D.I. <30 represent original calcic plagioclase + olivine cumulates. (Fig. 8C). Rocks with indices $>$ 30 plot along a well defined linear treend believed to be the line of liquid descent. Feldspathic rocks from south of the Tyin–Gjende Fault have no representatives with D.I. <30, but the liquid descent trends are identical. Possibly these southern rocks represent a congealed portion of the original magma which separated from the main mass after the accumulation of low D.I. rocks. This apparent lack of feldspathic cumulate rocks from south of the Tyin–Gjende Fault is one of the very few major points of distinction between the two terrains either side of the fault. The Høystakka Granofels plots at the high D.I. end of the liquid descent trends.

The total alkali–total iron-MgO (= AFM) diagram (Fig. 9A) shows the initial cumulate trend of iron enrichment followed by a typical calc-alkaline trend for the liquid descent (Irvine & Baragar 1971). This trend is similar to that of the

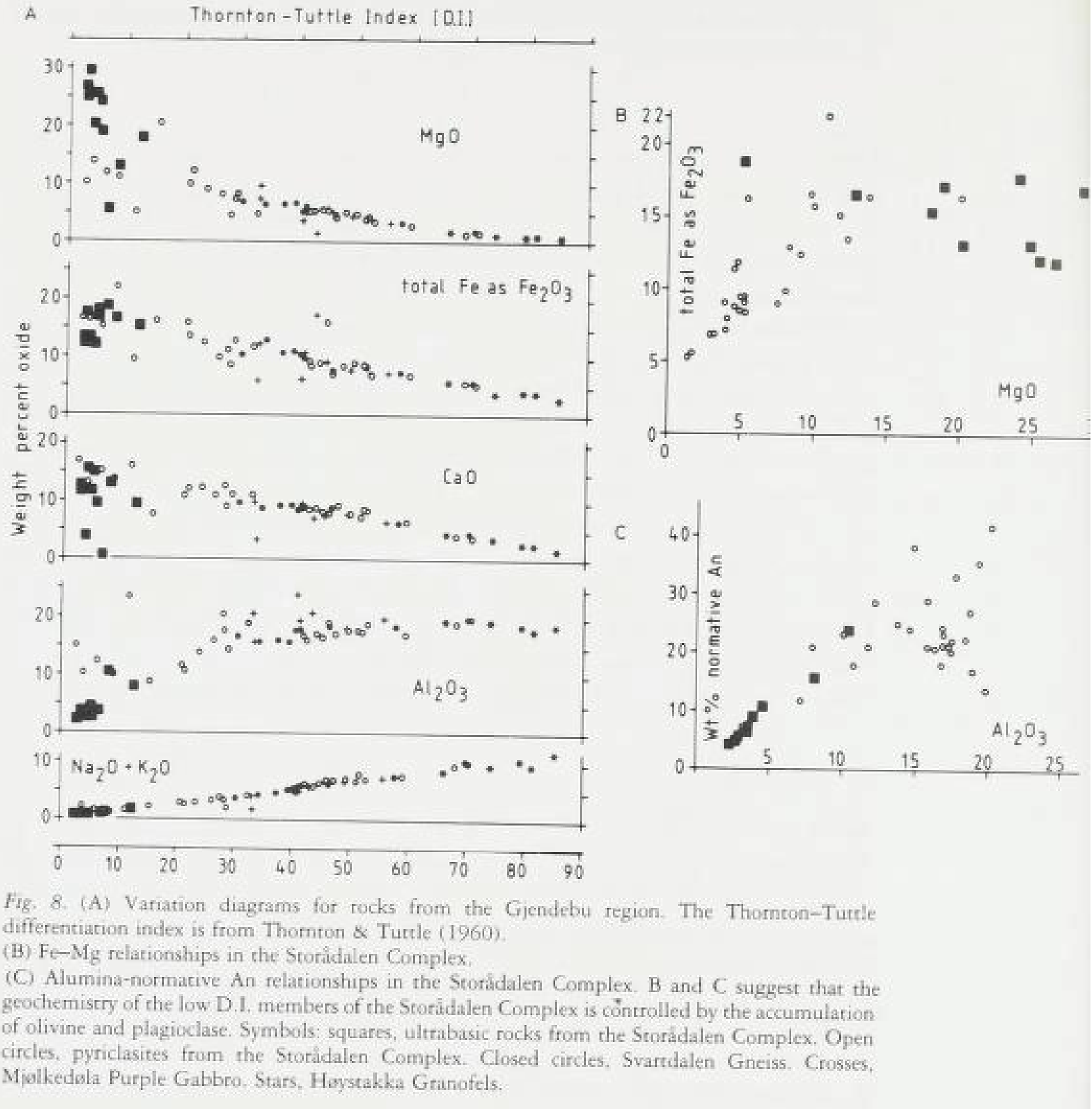

Fig. 8. (A) Variation diagrams for rocks from the Gjendebu region. The Thornton–Tuttle differentiation index is from Thornton & Tuttle (1960).
(B) Fe–Mg relationships in the Storådalen Complex.
(C) Alumina-normative An relationships in the Storådalen Complex. B and C suggest that the geochemistry of the low D.I. members of the Storådalen Complex is controlled by the accumulation of olivine and plagioclase. Symbols: squares, ultrabasic rocks from the Storådalen Complex. Open circles, pyriclasites from the Storådalen Complex. Closed circles, Svartdalen Gneiss. Crosses, Mjølkedøla Purple Gabbro. Stars, Høystakka Granofels.

Layered Series reported by McRitchie (1965). The small degree of iron richness of the Jotun kindred trend over the more typical calc-alkaline trend shown by, say, the classical basalt–andesite–dacite–rhyolite (= B–A–D–R) volcanic suite of the western U.S.A. (Turner & verhoogan 1960, chap. 10), is presumably due to differentiation occurring under conditions of low P_{H_2O}–high P_{load} conditions (cf. Philpotts 1966). Such calc-alkaline trends are exhibited by many granulite facies terrains (e.g. the charnockites of Guyana, Singh 1966, and the ultrabasic, gabbroic, and dioritic rocks of the Ivrea Zone, Rivalenti et al. 1981; see Fig. 9B). The Gjendebu rocks also give a typical calc-alkaline trend on the F–M diagram of Simpson (1954), the trend lying as it does between the curves for the Garabal

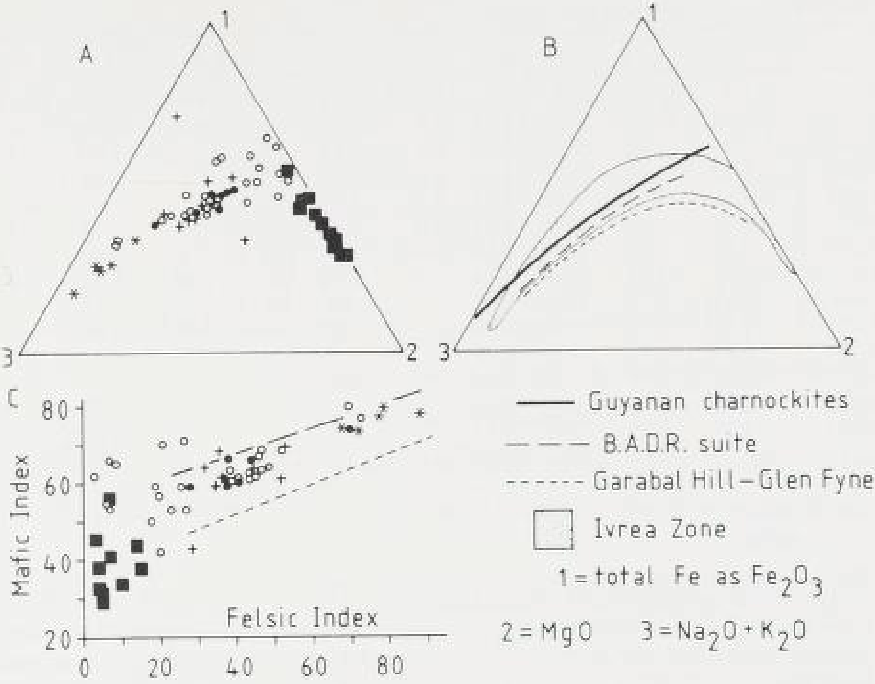

Fig. 9. (A) AFM diagram for rocks of the Gjendebu region.
(B) Comparative trends on an AFM diagram: Garabal Hill–Glen Fyne Complex from Nockolds (1941), Guyanan charnockites from Singh (1966), the trend for the B–A–D–R volcanic suite (see text) is from Turner & Verhoogan (1960), and that for the Ivrea Zone from Rivalenti et al. (1981).
(C) F–M diagram, after Simpson (1954), for rocks from the Gjendebu region. All symbols as in Fig. 7.

Hill–Glen Fyne Complex and the previously mentioned B–A–D–R suite (Fig. 9C).

The average bulk compositions of the Storådalen Complex, Svartdalen Gneiss and Mjølkedøla Purple Gabbro were determined using a simple weighted average technique similar to that described by Eade et al. (1966). Full details will be found in Emmett (1980a). The results, listed in Table 4, indicate that there are no significant differences between the three units. The average bulk composition for the whole area examined has been similarly determined and the result is almost identical to the value obtained by a different method for the 'parental magma' of the Layered Series by Battey & McRitchie (1975, and Table 4). This concurrence of the bulk compositions of the Layered Series and the rocks of the Gjendebu region strongly suggests that these units are comagmatic.

Table 4 also lists selected analyses for comparative purposes, including calc-alkaline rocks from island arcs (Jakês & White 1969, Lowder & Carmichael 1970), and some average granulite facies terrains. The bulk composition of the Gjendebu rocks compares most closely with the island arc rocks, a point also made by Battey & McRitchie (1975), and metamorphosed calc-alkaline rocks, e.g. granulites from the Sao Francisco craton of Brazil (Sighinolfi 1971). However, the

TABLE 4. Comparative whole rock analyses

	1	2	3	4	5	6	7	8	9	10
SiO_2	51.9	53.0	52.5	52.5	51.5	51.39	60.06	54.0	50.59	53.83
Al_2O_3	16.76	17.45	18.66	17.62	18.0	16.67	15.4	19.0	16.29	15.42
Fe_2O_3	3.40	3.28	2.49	3.06	3.5	2.84			3.66	5.00
FeO	5.53	5.30	5.25	5.36	6.0	8.68	7.2*	9.0*	5.08	6.74
MgO	5.45	4.90	4.92	5.09	5.0	5.03	3.9	4.1	8.96	4.36
CaO	8.79	8.04	8.48	8.44	10.0	8.40	5.7	9.5	9.50	8.83
Na_2O	3.49	3.69	3.73	3.64	3.8	3.14	2.8	3.4	2.89	2.99
K_2O	2.11	2.30	1.99	2.13	1.0	1.23	2.6	0.6	1.07	0.92
MnO	0.18	0.76	0.64	0.53		0.191	0.2		0.17	0.20
TiO_2	0.76	0.17	0.14	0.36		1.32	0.9	0.9	1.05	1.01
P_2O_5	0.31	0.45	0.40	0.39					0.21	0.21
H_2O^+	0.47	0.32	0.24	0.34		0.90				0.44
H_2O^-	0.16	0.25	0.18	0.20					0.81**	0.10
(CO_2)	0.37	0.16	0.47	0.21						
total	99.64	100.07	100.07	99.87	98.80	99.791	99.30	100.50	100.28	100.05

* total Fe as FeO
** total volatiles

Table 4. Comparative whole rock analyses

1. average bulk composition of Storådalen Complex.
2. average bulk composition of Svartdalen Gneiss (n = 11).
3. average bulk composition of Mjølkedøla Purple Gabbro (n = 10).
4. average bulk composition of all rocks in Gjendebu region.
5. 'parental magma' of Layered Series (Battey & McRitchie 1975).
6. average Brazilian basic granulite (Sighinolfi 1971).
7. average intermediate-pressure granulite facies rock, Musgrave Range, Australia (Lambert & Heier 1968).
8. average composition of lower continental crust (Taylor & McLennan 1979).
9. calc-alkaline pyroxene-basalt from Mt. Trafalgar, East Papua (Jakês & White 1969).
10. basaltic andesite from Cape Hollmann, Talasea (Lowder & Carmichael 1970).

The figures in column 1 were derived by dividing the complex up into ultrabasic types, anorthositic jotunites, and jotunite-magerites. The outcrop area of each of these three rock types was estimated from the map and a *pro rata* weighting assigned to each on this basis. A simple average was taken of all the analyses available for each group and the values obtained were combined together (after applying the appropiate weighting) to give column 1. This method is similar to that used by Eade et al. (1966) to estimate the average composition of part of the Canadian Shield. Columns 2 and 3 are simply averages of all the appropiate analyses, whilst column 4 is the average of columns 1, 2 and 3.

Gjendebu rocks are generally more alkalic and more basic than any of these, a fact which is most noticeable when the Storådalen Complex is compared with other intermediate pressure granulites, e.g. from the Musgrave Ranges of Australia (Lambert & Heier 1968). This observation is significant because lower crustal granulites are commonly regarded as depleted in potassium and other lithophile elements (Taylor & McLennan 1979), a result of either partial melting (Fyfe 1973) or some as yet poorly understood CO_2-initiated metasomatism (Newton et al. 1980). Most of the potash in the Storådalen Complex is carried by mosaic-type perthites, with the small amounts present in antiperthite and biotite generally insignificant. These perthites presumably formed as homogeneous crystals before or during D_1, with post-metamorphic annealing subsequently incorporating them

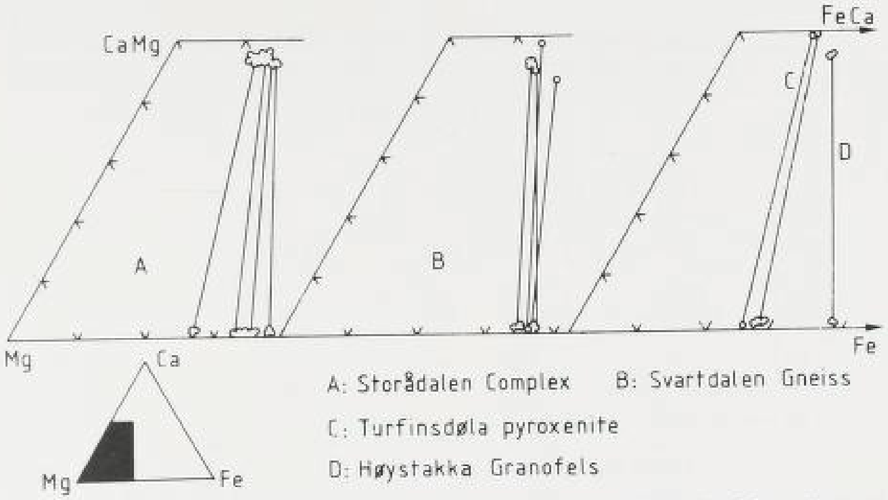

Fig. 10. Quadrilatral diagrams for co-existing pyroxenes from rocks of the Gjendebu region.

into the triple junction mosaic. This observation does not preclude the possibility of pre-D_1 metasomatism, but considering the smooth trend of the Na_2O+K_2O curve on the variation diagram (Fig. 8A), and the fact that the Svartdalen Gneiss and the Mjølkedøla Purple Gabbro are also rich in potash, it is considered that the high potash contents of these rocks is a reflection of the original composition of the parental magma.

CO-EXISTING PYROXENES

A reconnaissance study of the geochemistry of co-existing pyroxenes has been completed, the analyses being made by electron microprobe and fully corrected for ZAF effects (for full details see Emmett 1980a). The pyroxene pairs analysed and the results obtained are listed in Table 1. As is usual with microprobe analyses of pyroxenes, little or no account has been taken of inhomogeneities produced in crystals by exsolution of plagioclase and/or oxide phases (cf. Howie & Smith 1966). No zoning was detected and tie-line arrays sweep smoothly across the quadrilateral diagrams (Fig. 10), both features indicating that equilibrium conditions have been obtained. The clinopyroxenes are broadly augitic in composition whilst the orthopyroxenes vary from En_{63} to En_{74}. Both pyroxene species from the Storådalen Complex and the Svartdalen Gneiss are quite aluminous. Plots of pyroxene composition (Al_2O_3 and Mg/(Mg + Fe)) against whole rock composition (Al_2O_3, D.I., and FeO/(FeO + MgO)) all indicate that bulk composition is having a strong control on pyroxene chemistry (Fig. 11). The partition coefficient K_D (as defined by Kretz 1961) varies from 0.67 to 0.89, all higher than the value of 0.53 thought to typify 'metamorphic' pyroxenes. All these features suggest that the compositions of co-existing pyroxenes are still reflecting relationships established during magmatic differentiation, a similar conclusion to that reached by Battey & McRitchie (1975) for pyroxenes from the Layered Series.

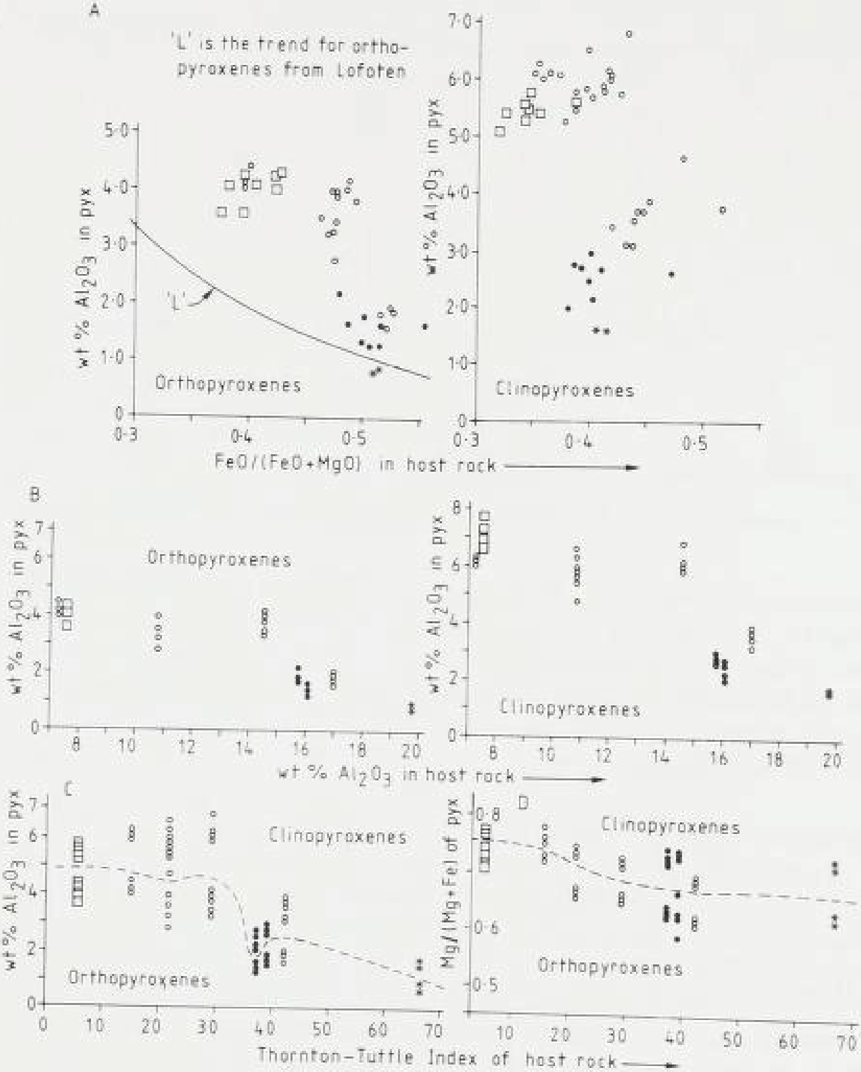

Fig. 11. Geochemical trends of analysed pyroxenes.
(A) Alumina content of pyroxenes against the ratio FeO/(FeO + MgO) in host rock. The trend from Lofoten is taken from Griffin & Heier (1969).
(B) Alumina content of pyroxenes against alumina content of the host rock. Note the strong dependence of the former on the latter.
(C) and (D) Alumina content and the ratio $Mg^{2+}/(Mg^{2+} + Fe^{2+})$ of pyroxenes plotted against the differentiation index of the host rock. Symbols for host rock types: Squares, Turfinsdøla pyroxenite. Open circles, Storådalen Complex. Closed circles, Svartdalen Gneiss. Stars, Høystakka Granofels.

It should be stated that the trends in pyroxene chemistry apparent from this study (for more complete discussion see Emmett 1980a) may, in part, reflect inadequacies in sampling and the small number of data points. However, the results are mutually consistent and are similar to those obtained for the Layered

TABLE 5. Chemical analyses and modes of mesoperthosites

	116	117		C.I.P.W. norms 116	117
SiO_2	77.0	64.5			
Al_2O_3	12.50	18.97	Q	36.02	4.70
Fe_2O_3	0.51	1.24	Or	28.36	35.97
FeO	0.37	0.75	Ab	29.34	46.85
MgO	0.12	0.43	An	4.28	8.37
CaO	1.29	1.76	Di	0.70	0.0
Na_2O	3.47	5.54			
K_2O	4.79	6.08	Hy	0.43	1.21
TiO_2	0.14	0.20	C	0.0	0.15
MnO	0.01	0.06			
P_2O_5	0.03	0.06	Mt	0.74	1.18
H_2O^+	0.33	0.37	Ilm	0.27	0.38
H_2O^-	0.07	<0.05	Ap	0.0	0.13
total	100.63	100.01			

Modes (n = 1000 points)

	Q	opx	cpx	K-flds	plag	amphib.	biot.	opaques
116	54.3	tr.	n.d.	42.9	1.8	0.1	n.d.	0.9
117	4.1	n.d.	0.8	89.9	1.6	1.6	0.1	1.9

Table 5. Chemical analyses and modes of mesoperthosites. Both samples are from the Langeskavlen province. Analyses and norms are in weight %, the modes are in volume %.

Series by McRitchie (1965) and Battey & McRitchie (1975), and so are considered to be fairly reliable. Further data are required to refine the trends and possibly to allow the application of geothermometric methods.

Mesoperthosites

Battey & McRitchie (1975) reported the discovery of transgressive veins of a pale non-foliated rock, termed *mesoperthosite* by McRitchie (1965), at Langeskavlen, 2 km northwest of Eidsbugarden. Emmett (1980a) reported other occurences around Olavsbu, on Mjølkedalsbandet, and in Memurudalen (cf. Rekstad 1904). In these latter localities, the mesoperthosite may form veins, blebs, or ill-defined patches, but in all cases the bodies of this rock cut across the S_1 fabric and are unfoliated. No mesoperthosite has been found south of the Tyin–Gjende Fault.

Mesoperthosite consists dominantly of quartz and mesoperthite, with minor pyroxene and plagioclase, and accessory zircon and apatite. Representative modes are given in Table 5, and full descriptions of their petrography and geochemistry will be found in the works cited. Battey & McRitchie (1975) and Battey (1978) have proposed that these rocks are the product of partial melting of feldspathic rocks during uplift of the Layered Series (and Storådalen Complex). Fig. 12 shows how simple decompression could generate the mesoperthosites. Note that in the absence of volatiles (line 'a' in the figure) temperatures in the order of 1000°C are necessary, whilst in the water-saturated system (line 'c') simple decompression cannot cause melting. It is proposed that, following the experimental work of Huang & Wyllie (1975) and Eggler & Kadik (1979), small amounts of a volatile phase containing both H_2O and CO_2 was present during uplift and that this causes

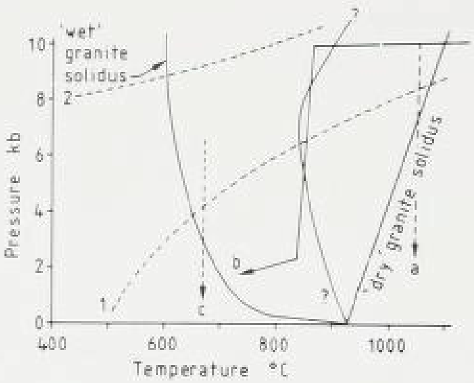

Fig. 12. Model for the generation of the mesoperthosites. The reaction curves 1 and 2 are the same as in Fig. 7. The granite solidi are from Huang & Wyllie (1975). The line labelled '? – ?' is a hypothetical solidus for $P_{H_2O} > O < P_{total}$ (cf. Eggler & Kadik 1979). The lines 'a' and 'c' represent decompression paths for $P_{H_2O} = O$ and $P_{H_2O} = P_{total}$ conditions respectively. In the latter case, simple decompression is incapable of causing partial melting. Line 'b' is the proposed decompression path (Fig. 7) which, combined with the hypothetical solidus, will produce limited amounts of partial melts within the Storådalen Complex.

the granite solidus to have the form of the line '?' in the figure. Though there are insufficient data to locate this solidus accurately in P-T space, it can be seen that simple decompression will generate feldspathic melts at a much more geologically reasonable temperature. Given a solidus of the proposed shape, the decompression will carry the host rock into then out of the zone of partial melting. This, coupled with volatile-scavenging by the melts produced would account for the limited amounts of partial melting produced. Mesoperthosite production would be restricted to those areas of locally increased $P_{volatile}$, but there would be insufficient volatiles present to cause wholesale partial fusion.

Status of the Høystakka Granofels

The status of the Høystakka Granofels is uncertain. Emmett (1980a) regarded it to be of migmatitic aspect, being generated by partial melting of the Svartdalen Gneiss during a re-heating event. This is now considered unlikely. The analysed examples of the Høystakka Granofels plot on the same differentation trends as the other rocks in the area (Fig. 8A), and this can be cited as evidence, not conclusive, indicating that the granofels is a re-mobilised acidic differentiate of the Jotun igneous suite. This would require the re-mobilising process, presumably the high-grade metamorphism, to be essentially isochemical with respect to the major elements. These conclusions are tentative; the true nature of the Høystakka Granofels is not, as yet, understood.

Evolution of the Gjendebu region

The major tectonic problem in the Gjendebu region concerns the relationship of the Storådalen Complex to the rocks south of the Tyin–Gjende Fault. Battey & McRitchie (1973) suggested that the peripheral gabbros were essentially younger than the axial rocks and that their intrusion around the margin of a plug of axial rocks aided the uplift of the latter. This model is supported by (i) the field disposition of the rocks and (ii) the presence of rare xenoliths of granulite in marginal gabbros (Twist 1979). However, Emmett (1980a) noted the geochemical similarity of the Layered Series, Storådalen Complex, Svartdalen Gneiss and Mjølkedøla Purple Gabbro and concluded that these units were comagmatic and originally formed a single body, the Jotun parental body. Similar conclusions were reached by Dietrichson (1958) and McRitchie (1965). Also to be considered now are the radiometric data of Schärer (1980), which indicate that the magmatic evolution of the Jotun orthogneisses extended over a period of 400 to 500 Ma. It seems, therefore, that the Jotun parental body should be regarded as being of composite nature and possibly formed by the congealing of an indeterminate number of pulses of compositionally similar parental magma. At the present time it is not possible to recognise or discriminate between individual magma pulses since metamorphism and deformation have, in part, homogenised the body. In this modified model it is considered that the variations in deformation and grade of metamorphism now apparent represent original vertical variations within the parental body.

Though more than one magma pulse is indicated, each pulse differentiated and solidified under similar physical conditions. The evidence from the ultrabasic rocks within the Storådalen Complex requires that the Jotun parental magmas initially precipitated olivine + plagioclase assemblages, and this would be possible only at pressures of c. 7 kb or less (Emslie 1970, Presnall et al. 1978). This pressure corresponds to a depth of emplacement of about 25 km, and Emmett (1980a) has suggested that the magmas were intruded into stable continental crust of this approximate thickness and then proceeded to differentiate. The fractionation trends thus established indicate dry calc-alkaline magmas evolving under conditions of a small but progressive reduction in partial pressure of oxygen (Philpotts 1966). After consolidation, deformation affected the roots of the Jotun parental body, resulting in a prograde increase in P and T and subsequent metamorphism. The grade of metamorphism would, of course, diminish upwards through the body. The climax of metamorphism was followed by a period of recrystallisation and concomitant mineralogical adjustment which was terminated by a rapid decompression (= uplift) event. During this uplift the ingress of small amounts of water into the Layered Series and Storådalen Complex caused partial melting and the generation of mesoperthosites. Other results of this decompression include the unmixing of homogeneous feldspars to form perthites, the exsolution of Fe-Ti oxides, plagioclase and complementary pyroxene from pyroxene, the exsolution of spinel-magnetite solid solutions, and the symplectitic breakdown of high-pressure phases (mainly aluminous pyroxenes). Full details of these phenomena will be found in Emmett (1980a). The presence of Eocambrian and Lower Palaeozoic sediments resting upon eroded Bergen–Jotun kindred rock (e.g. as at Grønsennknipa, Hossack (1972)) clearly indicates that most of this uplift and related mineralogical changes were complete prior to the onset of Eocambrian sedimentation.

After receiving their cover of sediments, the exhumed Jotun kindred rocks were involved in Caledonian orogenesis. Prior to initial nappe formation, at about 450 Ma b.p. (Berthomier et al. 1972), the Jotun rocks were invaded by trondhjemitic

TABLE 6. Summary of the geological history of the Gjendebu region

Stage	Storådalen Complex	Svartdalen Gneiss	Mjølkedøla Purple Gabbro
magmatic c. 1200–900 Ma	Cumulates and rocks of liquid descent. Production of S_O (igneous layering?)	liquid	descent
D_1	Production of lineated and foliated tectonites (S_1). Disruption of S_O	Weak foliation (S_1).	Some cataclasis(?)
high grade metamorphism	Intermediate pressure granulite facies	*(emplacement of ultrabasic bodies)* Not exceeding low pressure granulite facies *(emplacement of Høystakka Granofels?)*	
D_2	Mainly folding	Folding(?)	
post – climactic	Cooling at high subsolidus temperatures. Production of the triple junction mosaic.	Some recrystallisation	
uplift	Ingress of small amounts of volatiles causes limited partial melting (mesoperthosites) Exsolution of oxides from pyroxenes, unmixing of perthites etc. Intrusion of granites* and dolerites. Erosion, followed by deposition of Eocambrian and Lower Palaeozoic sediments*.		
Caledonian Orogeny K/Ar clocks 'close' c. 450 Ma	Trondhjemite emplacement (c. 430–450 Ma) Caledonian thrusting and low grade metamorphism. Lag faulting T.G.–I Høystakka Fault T.G.–II Minor faulting		

Table 6. Summary of the geological evolution of the Gjendebu region. The absolute ages given are tentative and based on the work of Battey & McRitchie (1973) and Schärer (1980). Events marked with an asterix are not recorded in the Gjendebu region, and those in italics are only approximately located.

magmas which were possibly derived by partial melting at depth (Henry 1977, Size 1979). Basement and cover were then formed into a large nappe, the Jotun Nappe, and transported to their present position. There is still some controversy concerning the nature of this displacement (see Smithson et al. 1974 for summary) but a displacement of up to c. 300 km in a southeastward direction is indicated by some data (Hossack 1978). During this nappe-forming event the Jotun rocks underwent feeble re-heating that was sufficient to re-set their K-Ar 'clocks' (Battey & McRitchie 1973), though mineralogical retrogression is not always readily apparent far from the basal thrust (Emmett 1980b). The emplacement of the nappe along a curved thrust front generated the wrench faults now observed on Memurutunga.

Events subsequent to nappe emplacement and recorded in the Gjendebu region include:

a) gravitative sagging along inwardly dipping lag – faults. This causes a substantial thickening of the central portion of the nappe. The small but definite density excess of the Jotun rocks (Smithson et al. 1974) compared to the underlying units would have assisted this process.

b) renewed southeastward compression which drives the thickened core of the nappe up and against the thinner portion along the line of the Tyin–Gjende Fault. This generates the structures assigned to the T.G.–I period (Emmett 1981).

c) formation of north-south trending faults (e.g. the Høystakka Fault).

d) continued movement on existing fault lines (e.g. T.G.–II structures) and the formation of minor faults.

These events are summarised in Table 6.

In conclusion, it is believed that the Jotun kindred rocks of the Gjendebu region represent a consanguineous suite of plutonic calc-alkaline igneous rocks which were variably deformed and metamorphosed prior to a pre-Eocambrian exhumation. After receiving a cover of Eocambrian and Lower Palaeozoic sediments the rocks were involved in the Caledonian nappe-forming event, but the Jotun gneisses themselves suffered only limited Caledonian deformation and retrogression.

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REFERENCES

Austrheim, H. & Råheim, A. 1981: Age relations within the high grade metamorphic rocks of the Bergen Arcs, western Norway. *Terra Cognita 1* ,33 (abstract only).

Battey, M. H. 1960: Observations on the peridotites and pyroxenites of the Jotunheim Complex in Norway. *XXI Int. Geol. Congr. Norden 13*, 198–207.

Battey, M. H. 1965: Layered structures in rocks of the Jotunheim Complex, Norway. *Mineral. Mag., 34*, 35–51.

Battey, M. H. 1978: Some data for a timetable of events in the crystalline rocks of Jotunheimen (to April 1978). *In* Cooper, M. A. & Garton, M. R. (eds.) *Tectonic Evolution Of The Scandinavian Caledonides.* Proceedings of a conference held at the Dept. of Geology, City of London Polytechnic, on April 27, 1978: 10–14 (abstract only).

Battey, M. H. & Davison, W. 1977: Exsolution of plagioclase from clinopyroxene in a pyroxenite from Jotunheimen, Norway. *Mineral. Mag. 41*, 513–518.

Battey, M.H., Davison, W. & Oakley, P.J. 1979: Almandine pseudomorphs after plagioclase in a metadolerite dyke from the Jotunheim, Norway. *Mineral. Mag. 43*, 127–130.

Battey, M. H. & McRitchie, W. D. 1973: A geological traverse across the pyroxene–granulites of the Jotunheim in the Norwegian Caledonides. *Norsk geol. Tidsskr. 53*, 237–265.

Battey, M. H. & McRitchie, W. D. 1975: The petrology of the pyroxene–granulite facies rocks of the Jotunheim, Norway. *Norsk geol. Tidsskr. 55*, 1–49.

Berthomier, C., Lacour, A., Leutwein, F., Maillot, J. & Souct, J. 1972: Some trondhjemites of Norway: geochronological and geochemical study. *Sci. Terre Nancy 17*, 341–357.

Carstens, C. W. 1920: Norske peridotiter. *Norsk geol. Tidsskr. 5*, 43–73.

Carswell, D. A. 1973: The age and status of the basal gneiss complex of NW Norway. *Norsk geol. Tidsskr. 53*, 65–78.

Dietrichson, B. 1958: Variation diagrams supporting the stratiform magmatic origin of the Jotun eruptive nappes. *Norges geol. Unders. 203*, 4–34.

Eade, K. E., Fahrig, W. F. & Maxwell, J. A. 1966: Composition of crystalline shield rocks and the fractionating effects of regional metamorphism. *Nature 211*, 1245–1249.

Eggler, D. H. & Kadik, A. A. 1979: The system $NaAlSi_3O_8$–H_2O–CO_2 to 20 kb pressure I. Compositional and thermodynamic relations of liquids and vapours co – existing with albite. *Am. Mineral. 64*, 1036–1048.

Emmett, T. F. 1980a: The metamorphism and tectonic relationship between gabbros and granulites around Gjendebu, Jotunheimen, central southern Norway. *Unpublished Ph.D. thesis, University of Newcastle-upon-Tyne.*

Emmett, T. F. 1980b: Geology of the Torfinnsbu window. *Norsk geol. Tidsskr. 60*, 255–261.

Emmett, T. F. 1981: The Caledonian evolution of the Jotun Nappe of central, southern Norway. *J. Structural Geology 3*, 187 (abstract only).

Emmett, T. F. 1982: The petrography and geochemistry of coronite dolerites from the Jotun Nappe, central southern Norway. *Mineral. Mag. 46*, 43–48.

Emslie, R. F. 1970: Liquidus relations and subsolidus reactions in some plagioclase-bearing systems. *Ann. Rep. Geophys. Lab., Carnegie Inst. Yearb. 69*,148–155.

Fyfe, W. S. 1973: The granulite facies, partial melting and the Archaean Crust. *Phil. Trans. Roy. Soc. Lond. A, 273*, 457–461.

Gardner, P. M. & Robins, B. 1974: The olivine–plagioclase reaction: geological evidence from the Seiland petrographic province, nothern Norway. *Contrib. Mineral. Petrol. 44*, 149–156.

Gjelsvik, T. 1947: Anorthosittkomplekset i Heidal. *Norsk geol. Tidsskr. 26*, 1–57.

Goldschmidt, V. W. 1912: Die Kaledonische deformation der Südnorwegischen Urgebirgstafel. *Skr. Norske Vidensk.-Akad. i Oslo, Mat.-Naturv. Kl.*, No.19.

Goldschmidt, V. W. 1916: Übersicht der eruptivgesteine im Kaledonischen gebirge zwischen Stavanger und Trondhjem. *Skr. Norske Vidensk.-Akad. i Oslo, Mat.-Naturv. Kl.*, No.2.

Goldsmith, R. 1959: Granofels, a new metamorphic rock name. *J. Geol. 67*, 109–111.

Green, D. H. & Ringwood, A. E. 1967: An experimental investigation of the gabbro to eclogite transition and its petrological applications. *Geochim. Cosmochim. Acta 31*, 767–833.

Griffin, W. L. 1971a: Mineral reaction at a peridotite–gneiss contact, Jotunheimen, Norway. *Mineral. Mag. 38*, 435–445.

Griffin, W. L. 1971b: Genesis of coronas in anorthosites of the Upper Jotun Nappe, Indre Sogn, Norway. *J. Petrol. 12*, 219–243.

Griffin, W. L. & Heier, K. S. 1969: Paragenesis of garnet in granulite facies rocks, Lofoten, Vesterålen, Norway. *Contr. Mineral. Petrol. 23*, 89–116.

Griffin, W. L. & Heier, K. S. 1973: Petrological implications of some corona structures. *Lithos 6*, 315–335.

Heim, M., Schärer, U. & Milnes, A. G. 1977: The nappe complex in the Tyin – Bygdin – Vang region, Oppland, south Norway. *Norsk geol. Tidsskr. 57*, 171–178.

Henry, A. 1977: La Nappe du Jotun et les Nappes de socle connexes. *Rev. Géogr. phys. Géol. dynam. 19, fasc. 5*, 433–442.

Herzberg, C. T. 1975: Phase assemblages of the system $CaO - Na_2O - MgO - Al_2O_3 - SiO_2$ in the plagioclase–lherzolite and spinel–lherzolite mineral facies. *Unpublished Ph.D. thesis, University of Edinburgh.*

Herzberg, C. T. 1978: Pyroxene geothermometry and geobarometry: experimental amd thermodynamic evaluation of some subsolidus phase relations involving pyroxenes in the system $CaO - MgO - Al_2O_3 - SiO_2$. *Geochim. Cosmochim. Acta 42*, 945–958.

Hødal, J. 1945: Rocks of the anorthosite kindred in Vossestrand, Norway. *Norsk geol. Tidsskr. 24*, 129–243.

Holtedahl, O. & Dons, J. A. 1960: Geological map of Norway. *In* Holtedahl, O. (ed.) Geology of Norway. *Norges geol. Unders. 208*.

Hossack, J. R. 1968: Structural history of the Bygdin area, Oppland. *Norges geol. Unders. 147*, 78–107.

Hossack, J. R. 1972: The geological history of the Grønsennknipa Nappe, Valdres. *Norges geol. Unders. 281*, 1–26.

Hossack, J. R. 1978: The correction of stratigraphic sections for tectonic finite strain in the Bygdin area, Norway. *J. Geol. Soc. Lond. 135*, 229–242.

Howie, R. A. & Smith, J. V. 1966: X-ray emission microanalysis of rock – forming minerals V: orthopyroxenes. *J. Geol. 74*, 443–462.

Huang, W. L. & Wyllie, P. J. 1975: Melting relations in the system Ab – Or – SiO_2 + to 35 kb, dry and with excess water. *J. Geol. 83*, 737–748.

Irvine, T. N. & Baragar, W. R. A. 1971: A guide to the chemical classification of the common volcanic rocks. *Canadian J. Earth Sci. 8*, 523–548.

Jakês, P. & White, A J. R. 1969: Structure of the Melanesian Arcs and correlation with distribution of magma types. *Tectonophysics 8*, 223–236.

Kretz, R. 1961: Some applications of thermodynamics to co-existing minerals of variable composition. Examples: orthopyroxene – clinopyroxene and orthopyroxene–garnet. *J. Geol. 69*, 361–387.

Kretz, R. 1966: Interpretation of the shape of mineral grains in metamorphic rocks. *J. Petrol. 7*, 68–94.

Lacour, A. 1969: Données petrologiques sur la Nappe de Sogn–Jotun et son substratum dans la région de Sogndal.*Bull. Soc. geol. Fr. 51*, 354–366.

Lambert, I. B. & Heier, K. S. 1968: Geochemical investigations of deep seated rocks in the Australian Shield, *Lithos 1*, 30–53.

Lowder, G. G. & Carmichael, I. S. E. 1970: The volcanoes and caldera of Talasea, New Britain: geology and petrology. *Bull. Geol. Soc. Am. 81*, 17–38.

McRitchie, W. D. 1965: Structure and geochemistry of layered metamorphic rocks near Eidsbugarden, Jotunheimen, Norway. *Unpublished Ph.D. thesis, Kings College, University of Durham.*

Michot, P. 1961: Struktur der Mesoperthite. *N. Jb. Mineral. Abhandl.* 96, 213–216.

Moore, A. C. 1970: Descriptive terminology for the textures of rocks in granulite facies terrains. *Lithos 3*, 123–127.

Newton, R. C., Smith, J. V. & Windley, B. F. 1980: Carbonic metamorphism, granulites, and crustal growth. *Nature 288*, 45–50.

Nockolds, S. R. 1941: The Garabal Hill – Glen Fyne igneous complex. *Q. J. geol. Soc. Lond.* 96, 451–510.

O'Hara, M. J. 1967: Mineral parageneses in ultrabasic rocks. *In* Wyllie, P. J. (ed.) *Ultramafic and Related Rocks*. Wiley, New York, 393–403.

O'Hara, M. J. 1977: Thermal history of excavation of Archaean gneisses from the base of the continental crust. *J. Geol. Soc. Lond. 134*, 185–200.

Parsons, I. 1980: Alkali–feldspar and Fe–Ti oxide exsolution textures as indicators of the distribution and subsolidus effects of magmatic 'water' in the Klokken layered syenite intrusion, South Greenland. *Trans. Roy. Soc. Edin.: Earth Sci. 71*, 1–12.

Philpotts, A. R. 1966: Origin of the anorthosite–mangerite rocks in southern Quebec. *J. Petrol. 7*, 1–64.

Pressnall, D. C., Dixon, S. A., Dixon, J. R., O'Donnell, T. H., Brenner, R. L., Schrock, R. L. & Dycus, D. W. 1978: Liquidus phase relationships on the join Di–Fo–An from 1 atmosphere to 20 kilobars and their bearing on the generation and crystallisation of basaltic magma. *Contrib. Mineral. Petrol. 66*, 203–220.

Rekstad, J. 1904: Fra det nordøstlige af Jotunfjeldene. *Norges geol. Unders. 37.*

Rivalenti, G., Garuti, G., Rossi, A., Siena, F. & Sinigoi, S. 1981: The existence of different peridotite types and of a layered igneous complex in the Ivrea Zone of the western Alps. *J. Petrol. 22*, 127–153.

Schärer, U. 1980: U – Pb and Rb – Sr dating of a polymetamorphic nappe terrain: the Caledonian Jotun Nappe, South Norway. *Earth Planet. Sci. Lett. 49*, 205–218.

Sederholm, J. 1916: On synantectic minerals. *Bull. Comm. geol. Finl. 48*, 1–148.

Sibson, R. H. 1977: Fault rocks and fault mechanisms. *J. Geol. Soc. Lond. 133*, 191–213.

Sighinolfi, G.P. 1971: Investigations into deep crustal levels: fractionation effects and geochemical trends related to high – grade metamorphism. *Contrib. Mineral. Petrol. 35*, 1005–1021.

Simpson, E.S.W. 1954: On the graphical representation of differentiation trends in igneous rocks. *Geol. Mag. 91*, 238–244.

Singh, S. 1966: Orthopyroxene – bearing rocks of charnockitic affinities in the South Savanna–Kanuka complex of British Guiana. *J. Petrol. 7*, 171–192.

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

Sjøgren, A. 1883: Mikroskopiska studier: et par gabbroartar från Jotunfjällen i Norge. *Geol. För. Førhandl. Bd. 7, häft 9*, 370–375.

Smith, J. V. 1974: *Feldspar Minerals*, Vol. 2. Springer Verlag, Berlin, 401–516.

Smithson, S.B., Ramberg, I.B. & Grønlie, G. 1974: Gravity interpretation of the Jotun Nappe of the Norwegian Caledonides. *Tectonophysics 22*, 205–222.

Strand, T. 1972: The Norwegian Caledonides: the Jotun Nappe. *In* Kulling, O & Strand, T. (eds.) *The Scandinavian Caledonides*, Wiley Interscience, London, 38–42.

Sturt, B.A., Skarpenes, O., Ohanian, A.T. & Pringle, I.R. 1975: Reconnaissance Rb – Sr isochron study in the Bergen Arc system and regional implications. *Nature 253*, 595–599.

Sturt, B.A. & Thon, A. 1978: The Caledonides of Southern Norway. *Geol. Surv. Canada Paper 78–13*, 39–47.

Taylor, S.R. & McLennan, S.M. 1979: Discussion on Tarney & Windley (1977). *J. Geol. Soc. Lond. 136*, 497–500.

Thornton, C.P. & Tuttle, O.F. 1960: Chemistry of igneous rocks, part I. Differentiation index. *Am. J. Sci. 258*, 664–684.

Turner, F.J. 1968: *Metamorphic Petrology*. McGraw – Hill Book Company, New York. 403 pp.

Turner, F.J. & Verhoogen, J. 1960: *Igneous and Metamorphic Petrology*. McGraw – Hill Book Company, New York. 2nd. ed., 602 pp.

Twist, D. 1979: The structure and geochemistry of rocks along the NW margin of the Jotunheim massif, southern Norway. *Unpublished Ph. D. thesis, University of Newcastle-upon-Tyne.*

Vernon, R.H. 1968: Intergranular microstructures of high grade metamorphic rocks at Broken Hill, Australia. *J. Petrol. 9*, 1–22.

Forekomster av tillitt på nordsiden av Atnsjøvinduet, Syd-Norge

A. SIEDLECKA & S. ILEBEKK

Siedlecka, A & Ilebekk, S., 1982: Forekomster av tillitt på nordsiden av Atnsjøvinduet, Syd-Norge. *Norges geol. Unders. 373*, 33–37.

A previously reported tillite-like conglomerate occurring on the northern side of the Atnsjø Window rests with an erosional contact on Precambrian granite and gabbro and is overthrust by feldspathic metasandstones of the Rondane Nappe. The conglomerate is matrix-supported and shows lateral variation in stone size and composition. It is analogous to the tillites occurring on the southeastern side of the Atnsjø Window and in the Snødøla, Spekedal and Tufsingdal Windows, and is correlative with the Late Precambrian Moelv Tillite of the Lake Mjøsa district.

Anna Siedlecka, Norges geologiske undersøkelse, Postboks 3006, N–7001 Trondheim, Norway.
Svein Ilebekk, Saga Petroleum a.s. & Co., Postboks 550, N–1301 Sandvika, Norway.

Innledning

Basalkonglomeratet i Sølnsjøhøgda (Sølnsjøkrabbane, Fig. 1) ble, ifølge Oftedahl (1945) registrert av Marlow under hans kartlegging i området i 1935. Marlows dagbok fra 1935 finnes ikke i NGU's kartarkiv, men på hans feltkart (Sølnkletten 1:100 000) er sparagmitt avmerket i Sølnsjøhøgdas nordlige topp (1416 m.o.h.) og gabbro i den sydlige topp (1388). Oftedahl (muntlig meddel. 1981) har i 1941 undersøkt disse stedene og har i sin dagbok notert at: « . . . I Sydskråningen av toppen 1410 (siden rettet til 1416 – A.S.) er det en serisittrik kvartsittisk skifer med boller av hvit kvartsitt og granitt. Størrelsen varierte fra nøtt – hodestørrelse.» I sin avhandling av 1945 har Oftedahl tolket dette konglomeratet som presset tillitt.

Under berggrunnskartleggingen i området sommeren 1980 har vi iakttatt flere blotninger av grunnmassebårne konglomerater (diamiktitter) som etter all sannsynlighet er stedegne senprekambriske isavsetninger jevnførbare med Moelvtillitten i Mjøstraktene.

Beskrivelse

En mørk grå diamiktitt er blottet i den sydlige skråning av toppen 1416 m.o.h. av Sølnsjøskrabbane. Grunnmassen i diamiktitten er forskifret og består hovedsakelig av kvarts og fyllosilikater. Bollene er delvis kantet til delvis rundet og er lite deformerte. Bollene er av tre typer: lys grå til mørk grå dypbergarter av granittisk til kvarts-diorittisk sammensetning, mørk grå og lys grå kvartsitt og hvit kvarts. Bollenes tverrmål er omtrent 2–60 cm, men boller på 15–20 cm er mest alminnelige. Den blottede mektighet av diamiktitten er på ca. 20 m.

Gabbro er blottet i den sydlige (1388 m.o.h.) toppen av Sølnsjøskrabbane, men grensen mellom diamiktitten og gabbroen er ikke synlig.

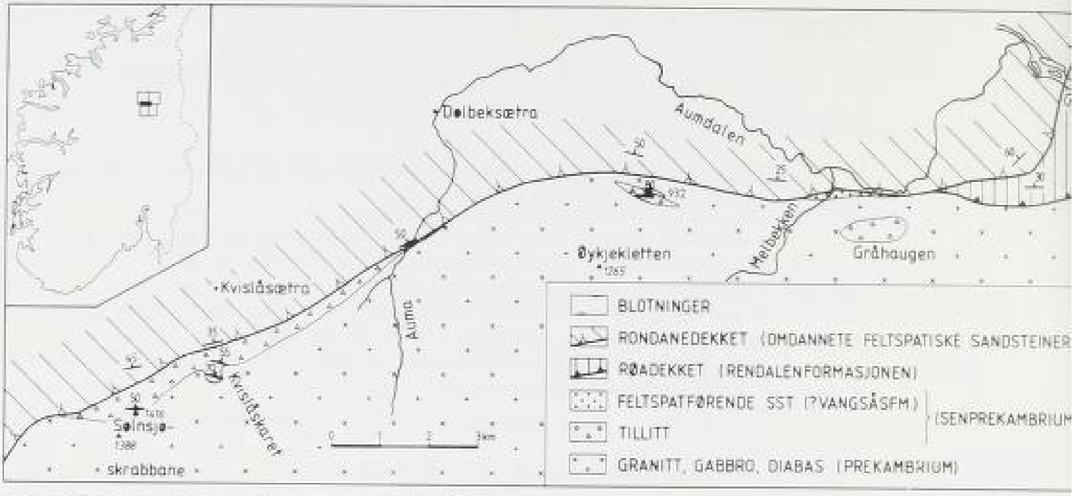

Fig. 1. Geologisk kart over den nordligste del av Atnsjøvinduet.
Geological map of the northern part of the Atnsjø window.

Det samme grunnmassebårne konglomerat er bevart i en forsenkning i grunnfjellet i Kvisläskaret. Her hviler konglomeratet med en klar sedimentær grense på gabbro og granitt. (Fig. 1.) Konglomeratet består av noen få centimeter store (maksimalt 30 cm), kantete bruddstykker av kvartsitt, feltspatisk metasandstein, granitt og metadoleritt (?) spredt i en grunnmasse av kvarts, kloritt, serisitt og feltspat. Mot nord er diamiktitten overleiret av feltspatiske metasandstener; nær kontakten er det en 5–10 m mektig sone hvor konglomeratets grunnmasse er kraftig foliert, bollene er mer eller mindre utvalset og stedvis kan det, på grunn av deformasjon, være vanskelig å skjelne mellom den omdannete feltspatiske sandsteinen og metadiamiktitten.

Sterk foliert diamiktitt kan videre iakttas i Auma. Denne har boller av kvarts, kvartsitt og granitt som er inntil 20 cm i største tverrmål. Her ligger også en sterkt forskifret feltspatisk metasandstein over diamiktitten, og bergartene har sammenfallende skifrighet.

Videre forekommer diamiktitt nordøst for Øykjekletten, på sydvestsiden av et lite vann 932 m.o.h. (Fig. 1). Bergarten er lite deformert og grunnmassen er svakt foliert. Kantete bruddstykker av grå- og rødlig arkose, lys grå kvartsitt, granitt og porfyr (?) er uregelmessig spredt i grunnmassen. Bruddstykkene har et tverrmål som varierer fra noen millimeter til 1,5 m. Grunnmassen kan best beskrives som en slambergart sammensatt av kvarts, feltspat og fyllosilikater. Innen det usorterte konglomeratet forekommer det stedvis lag med noe lavere innhold av grunnmasse og antydning til sortering av bollematerialet. Forekomsten ser ut til å være en utligger (isolert erosjonsrest) på det krystallinske underlaget (Fig. 1). Det samme gjelder forekomsten som ligger lengst mot øst, på nordskråningen av Gråhaugen (Fig. 1). Her er det flere, mer eller mindre sammenhengende blotninger langs hele skråningen og på toppen. Diamiktitten er fullstendig udeformert. I en mørk grå grunnmasse «flyter» det 2–30 cm store, noe kantete, boller av overveiende rød kvartsitt, og noe mindre av grå kvartsitt og granitt. På sydsiden av toppen av Gråhaugen står det en diabasgang, men grensen mellom bergartene er overdekket.

Teksturen av diamiktitten (lav modenhet) tyder på at den ikke ble dannet ved vanlig vanntransport. Mulige dannelsesmåter er enten som morene (tillitt) eller som en bruddstykkestrøm (debris flow) avsetning.

Bruddstykkestrøm-avsetningene forekommer i tykke terrigene lagfølger som er dannet enten som elvevifter i et horst-graben landskap eller som undersjøiske vifter. Elveviftene består vesentlig av ulike konglomerater mens de undersjøiske vifter består av turbiditter. Den beskrevne diamiktitten opptrer ikke som en del av en slik mektig lagfølge, men er et særskilt tynt lag. Derfor syntes det rimelig at diamiktitten er en forsteinet morene. Tolkningen støttes av det faktum at sikker tillitt forekommer i samme tektonostratigrafisk nivå som diamiktitten på sydsiden av Atnsjøvinduet, i Snødøla og på Spekedals- og Tufsingdalsvinduene lenger øst (Nystuen 1978, Siedlecka 1979, 1981). Jevnføring av disse forekomster med den senprekambriske Moelvtillitten er godt begrunnet gjennom flerårig regionalkartlegging og studier av lagfølger (f.eks. Nystuen 1976).

Grunnfjellet i Atnsjøvinduet består vesentlig av granitt, gabbro (i den nordlige delen av vinduet) og diabas. Feltiakttakelser tyder på at samtlige krystallinske bergarter er prekambriske (eldre enn tillitten), men selve kontakten er sjelden å se og det kan ikke utelukkes at en del av de basiske ganger er yngre.

Bollene av granittiske bergarter er alminnelig i tillitten og kan lett føres tilbake til denne typen dypbergarter som opptrer i mesteparten av Atnsjøvinduet. Boller av basiske bergarter, som også danner en del av underlaget, er imidlertid ikke funnet i tillitten. Årsaken kan være at disse bergarter forvitret lettere og nå inngår som en del av grunnmassen av tillitten, eller at det i kildeområdet for tillitten, ikke var basiske bergarter til stede.

Opphavsbergarter for de røde sandsteiner, som er det vanligste bollemateriale i tillitten på Gråhaugen, er ikke kjent i Atnsjøvinduet. Sandsteinen ligner imidlertid på de prekambriske Trysil-Dala sandsteiner.

I Moelvtillitten i den østlige delen av sparagmittbassenget har Nystuen (1976) og Nystuen & Sæther (1979) iakttatt tallrike boller av røde sandsteiner av Trysil-Dala-type, mens boller av basiske bergarter er sjeldnere. Granittiske boller er alminnelig. Resultater av disse undersøkelser sammen med observasjoner fra den nordlige delen av Atnsjøvinduet tyder på at kildeområdet for tillittene var en granittisk provins med enkelte basiske gangbergarter og sandige kontinentale sedimenter som overleiret deler av granittområdet.

På sydsiden av Auma, ved Melbekken, finnes en grovkornet forholdsvis udeformert feltspatisk sandstein (Fig. 1). Sandsteinen er lik den som på sydsiden av Atnsjøvinduet hviler med en sedimentær kontakt enten på tillitten eller direkte på granitten og er anført til Vangsåsformasjonen (Siedlecka 1979, Siedlecka et al. 1979). Samme stilling i lagfølgen er antatt for sandsteinen ved Melbekken.

Grensen mellom tillitten og underlaget er i den nordlige del av Atnsjøvinduet klart sedimentær. Grensen mot den overliggende metasandsteinen derimot, er tektonisk. Begge bergarter er sterkt folierte i kontaktsonen, den feltspatiske

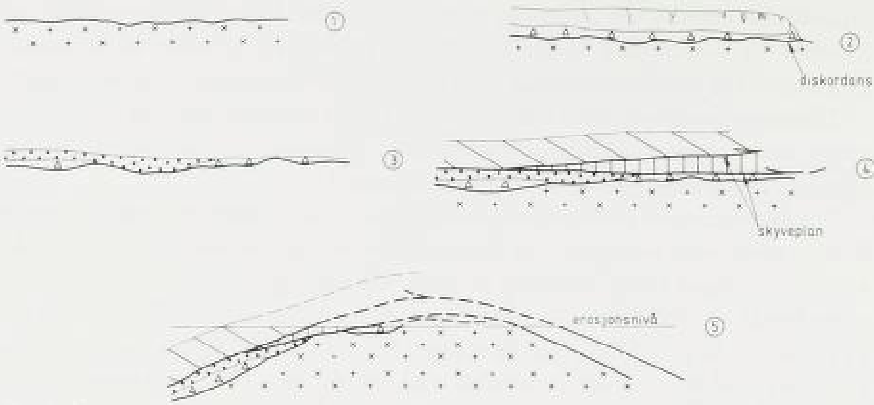

Fig. 2. Skisse over de geologiske hendelsesforløp i Atnsjøvindu-området.
Sketch showing the postulated geological development of the Atnsjø window area.

sandsteinen viser tegn til mylonittisering og bollene i tillitten er utvalsede og utstrakte. Deformasjonen avtar gradvis med avstanden fra kontakten. Denne grensen er tolket som et skyveplan. Den overskjøvne feltspatiske metasandsteinen danner bunnen av en mektig og utholdende lagpakke som kan følges uavbrutt vestover i Rondane og østover til Tylldal og som hører til Rondanedekket (Siedlecka 1979, 1981 og upublisert materiale).

Foliasjonen synlig i kontaktsonen mellom Rondanedekket og tillitten har, stort sett, et steilt fall mot nord (NNV, NØ, Fig. 1) slik at de stedegne bergartene stuper under dekkebergartene. Dette ser ut til å være et alment trekk rundt Atnsjøvinduet, trolig knyttet til den senkaledonske antiformdannelse av grunnfjellet (Siedlecka et al. 1979, Nystuen & Ilebekk 1981).

Slutninger

Den tidligere iakttatte forekomst av et konglomerat i Sølnsjøskrabbane i den nordlige delen av Atnsjøvinduet ble bekreftet av oss og flere nye blotninger ble funnet. Opptreden og tekstur av konglomeratet tyder på at dette er en morene dannet under den senprekambriske istid; denne tolkningen er i samsvar med Oftedahls konklusjon fra 1941. Kontaktforholdene og deformasjonsgrad av tillitten og de tilstøtende bergarter tyder på følgende geologiske hendelsesforløp i Atnsjøvindu-området (Fig. 2):

1. Erosjon av grunnfjellet og dannelse av det prekambriske peneplan.
2. Senprekambrisk glasiasjon, morenedannelse (tillitt) fulgt av:
3. Dannelse av elveavsetninger (Vangsåsformasjonens sandsteiner) etter tilbaketrekning av isbreen.
4. Overskyvning av de kaledonske dekkene fra nord-nordvest.
5. Oppdoming av grunnfjellet og erosjon av dekkebergarter – dannelse av Atnsjøvinduet.

ANERKJENNELSE

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LITTERATURLISTE

Nystuen, J. P., 1976: Facies and sedimentation of the Late Precambrian Moelv Tillite in the eastern part of the sparagmite region, southern Norway. *Norges geol. Unders. 329*, 70 pp.

Nystuen, J. P., 1978: Holøydal, berggrunnsgeologisk kart 1719 III M. 1:50 000. *Norges geol. Unders.*

Nystuen, J. P. & Ilebekk, S., 1981: Stratigraphy and Caledonian structures in the area between the Atnsjøen and Spekedalen windows, Sparagmite Region, southern Norway. *Norsk geol. Tidsskr. 61*, 17–24.

Nystuen, J. P. & Sæther, T., 1979: Clast studies in the Late Precambrian Moelv Tillite and Osdal Conglomerate, Sparagmite Region, south Norway. *Norsk geol. Tidsskr. 59*, 239–251.

Oftedahl, Chr., 1945: Om tillitene i det Central-norske sparagmitområde. *Norsk geol. Tidsskr. 25*, 285–293.

Siedlecka, A., 1979: Atnsjøen, berggrunnsgeologisk kart 1818 IV–M. 1:50 000. *Norges geol. Unders.*

Siedlecka, A., 1981: Sollia, berggrunnsgeologisk kart 1818 I–M. 1:50 000. *Norges geol. Unders.*

Siedlecka, A., Nystuen, J.P., Ilebekk, S., 1979: Stratigrafi og tektonikk ved Atnsjøvinduet, sparagmittområdet, Sør-Norge. *14. Nordiske Geol. Vintermøte, Bergen, Abstracts*, p. 64.

The Deglaciation and Vegetational History of a Former Ice-dammed Lake Area at Skåbu, Nord-Fron, Southern Norway

INGVILD ALSTADSÆTER

Alstadsæter, I. 1982: The deglaciation and vegetational history of a former ice-dammed lake area at Skåbu, Nord-Fron, southern Norway. *Norges geol. Unders. 373*, 39–43.

Cores obtained from a small peat bog within the area of a former ice-dammed lake show that 1.7 m of silt were deposited during the ice lake phase. The pollen in the silt is dominated by grasses and other NAP. A small transitional zone with coarse gyttja above the silt was radiocarbon dated at 8780 ± 210 yrs B.P. The pollen diagram shows Betula and Hippophäe maxima at the dated level. This date gives a minimum age for the drainage of the lake, and indicates that it existed for only a short period. Contrary to earlier suggestions, a valley fill of till and sediments at Skåbu was not able to maintain the lake after the deglaciation. In a 2.1 m-thick peat layer, above the gyttja, Pinus is the dominating pollen, indicating the presence of a dense pine forest. The Alnus Rise is radiocarbon dated at 7870 ± 80 yrs B.P.

Ingvild Alstadsæter, Postboks 7, 5427 Urangsvåg, Norway

Introduction

A large ice-dammed lake, Storsjøen, is known to have existed in the Vinstra river system during the last deglaciation (Rekstad 1898, Ramsli 1947, Bergersen 1971). Delta terraces at the mouths of inflowing rivers indicate a lake water level corresponding to the lowest point of the watershed to the south (Fig. 1).

The lake was probably dammed by remnants of the Scandinavian ice sheet. There has, however, also been discussion as to whether the thick valley fill of till and waterlaid sediments at Skåbu (Fig. 1) could have dammed the lake even after the ice had disappeared (Ramsli 1947, Mangerud 1963, Bergersen 1971). The latter problem has now been solved, and the ice lake phase dated, by coring the ice lake sediments in a small depression on the former bottom of the ice-dammed lake. Sedimentological and palynological analyses and radiocarbon dating were carried out on the core, which includes the peat on the top of the ice lake sediments. The investigation has therefore also provided information on the vegetational history of the area.

Field and laboratory methods

The borings were carried out with a 54 mm piston corer. For pollen analysis the peat sampled were prepared by Erdtman's acetolysis method; the minerogenic samples were also prepared by acetolysis and then treated with HF acid (Faegri & Iversen 1975). Grain size distributions from silt were obtained using the pipette

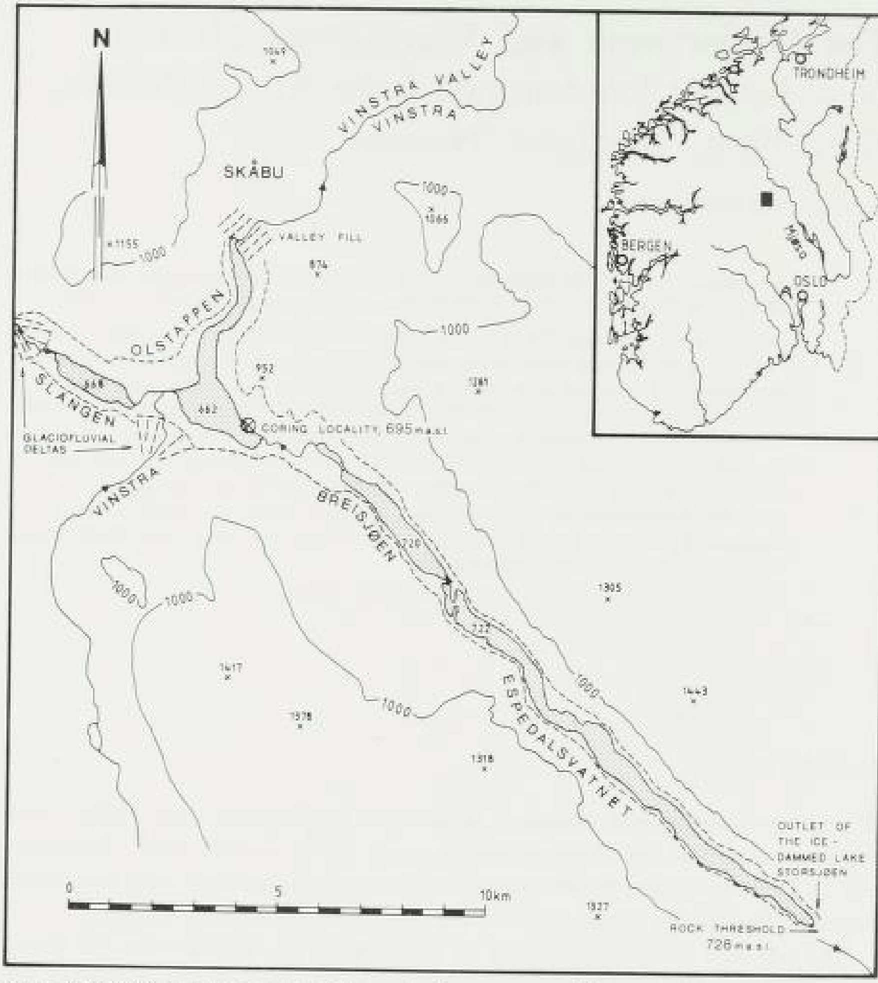

Fig. 1. Map of the ice lake area of the Vinstra drainage system, with present-day drainage direction (dark arrows) and lakes. During the existence of 'Storsjøen' ice lake, glaciofluvial material accumulated in waters reaching the rock threshold to the southeast. The approximate outline of Storsjøen (ice remnants excluded) is indicated by the dashed line. The location of the area is shown on the index map of southern Norway.

method. The radiocarbon datings were carried out at the Radiological Dating Laboratory, University of Trondheim.

Lithostratigraphy

The coring locality is situated near the top of a small rocky headland (Fig. 1). Soundings showed the thickness of sediment in the bog to be about 5 m. Resting on bedrock there is one metre of coarse, minerogenic material, either till or glaciofluvial gravel, which is overlain by 1.7 m of glaciolacustrine silt. The piston corer penetrated only one metre of the silt, and consequently only this part is

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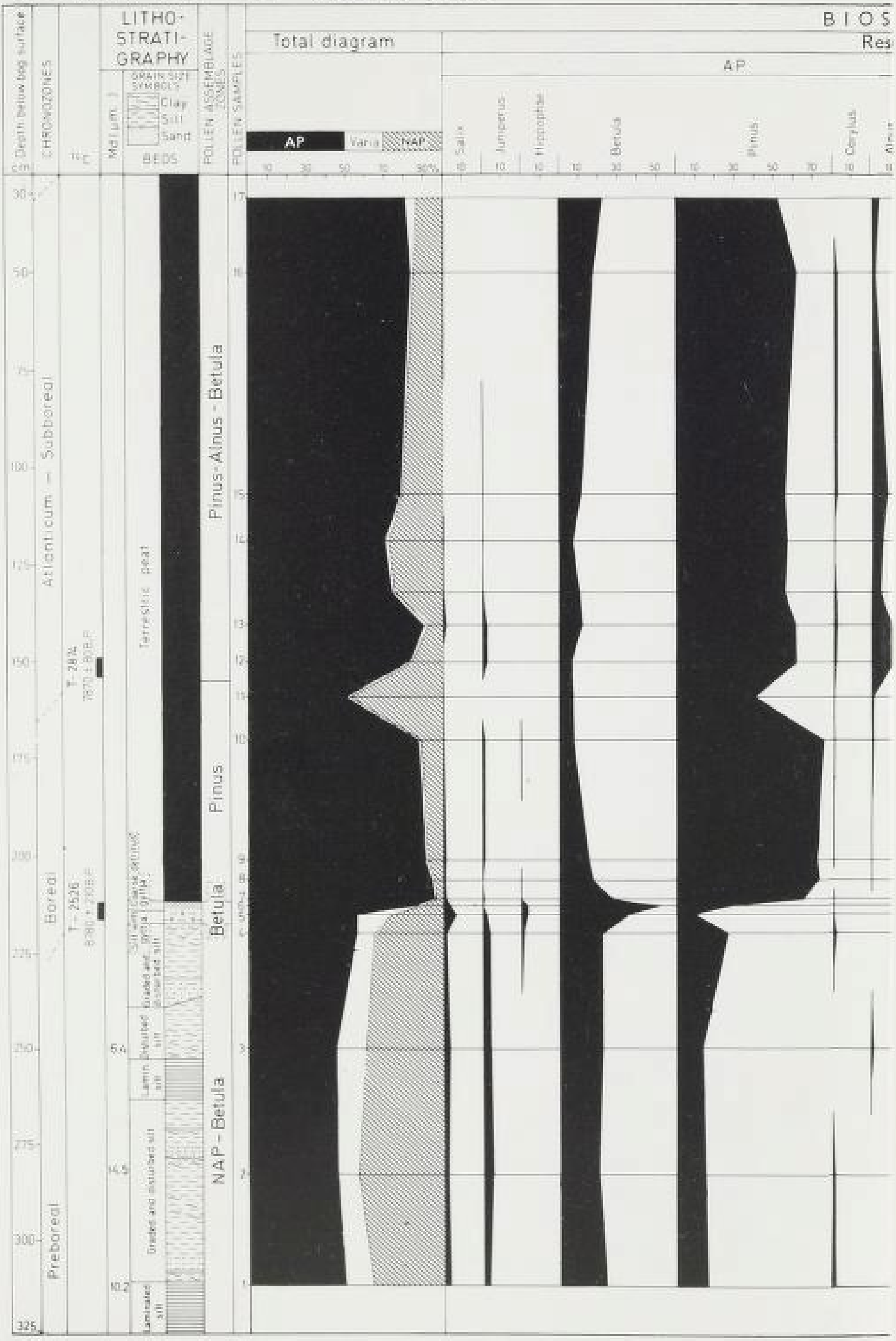
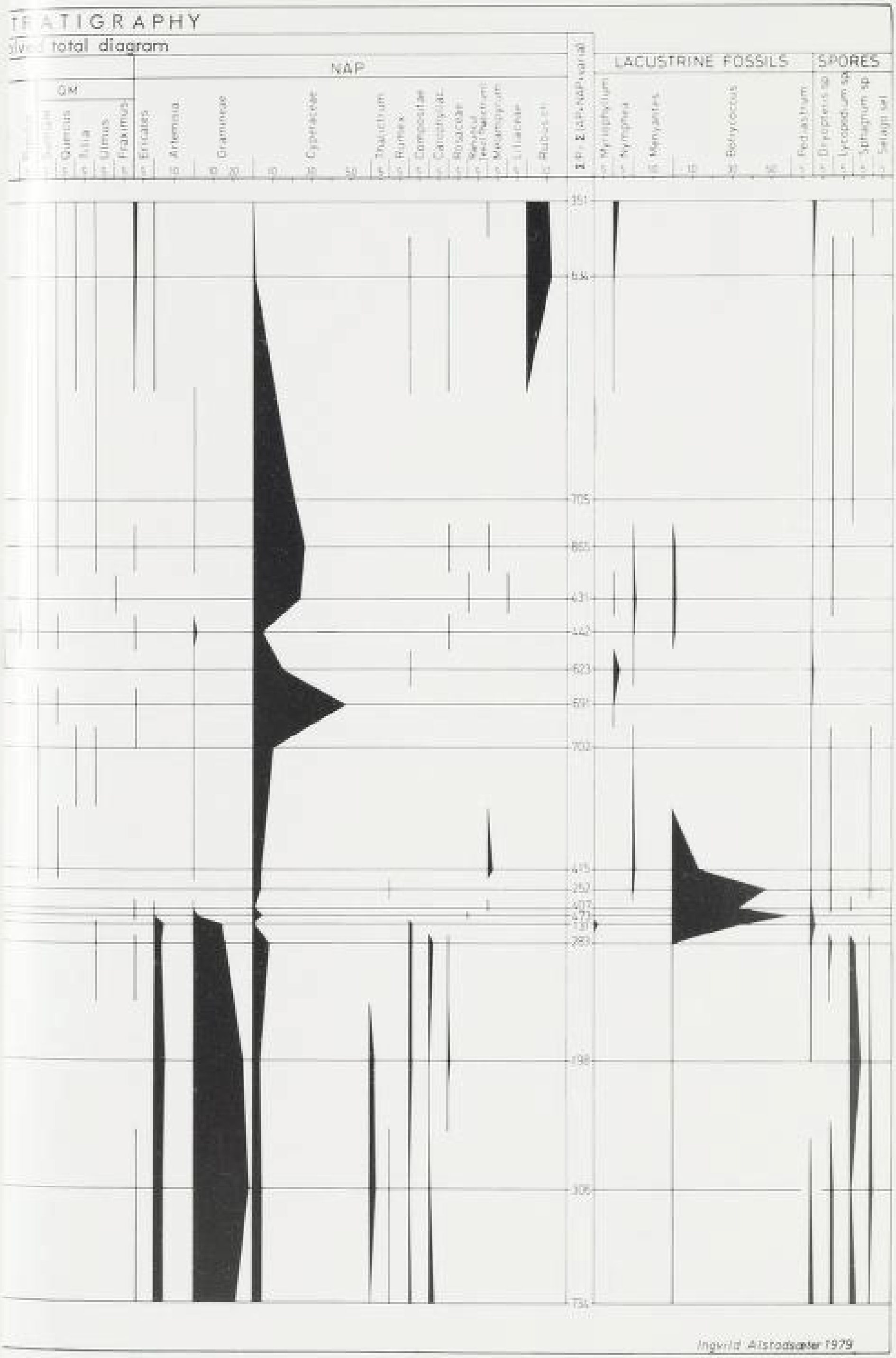



TRATIGRAPHY
olved total diagram
NAP
LACUSTRINE FOSSILS
SPORES
QM
Quercus
Tilia
Ulmus
Fraxinus
Ericales
Artemisia
Gramineae
Cyperaceae
Thalictrum
Rumex
Compositae
Caryophyl.
Rosaceae
Melampyrum
Liliaceae
Rubus
Myriophyllum
Nymphea
Menyanthes
Botryococcus
Pediastrum
Dryopteris sp.
Lycopodium sp.
Sphagnum sp.
Selaginella
Ingvild Alsrodsæter 1979

recorded in Fig. 2. Above the silt is 5 cm of silty gyttja, and a 5 cm zone of coarse gyttja with 2.1 m of peat on top. The bog basin has been completely filled in.

The glaciolacustrine silt consists of regularly repeated thin graded bedding, and therefore appears laminated. Grain size distributions (Fig. 2) gave an Md in the middle silt fraction. About 10 clayey laminations were found in the sequence. At irregular intervals 4 prominently graded beds were passing from sandy to clayey silt. These particular beds contain both the coarsest and the finest material in the core. At least one of these layers rests on an erosional surface.

The very limited drainage area of the small depression, with no brooks leading into the bog basin, points to the deposition of the thick glaciolacustrine silt unit in a larger lake, presumably the ice-dammed Storsjøen. The sandy-clayey silt beds could have been deposited by turbidity currents, triggered e.g. by spring floods. The repeated graded laminae may reflect some rhytmical change in discharge during the ablation season. None of the beds or laminae can be proved to represent annual varves, even though some of them probably are.

The silty gyttja was probably deposited by a redeposition of silt while a small lake or pond occupied the basin. The conditions of sedimentation soon stabilized, with very low input of minerogenic material, during which time the coarse detritus gyttja was deposited. As the lake became shallower peat started to accumulate.

Biostratigraphy and chronostratigraphy

The pollen sequence (Fig. 2) is subdivided into biostratigraphic assemblage zones, as outlined by Fægri & Iversen (1975). The chronozones are according to Mangerud et al. (1974). In this context, the two oldest pollen zones are of special interest; however, a few comments are also given on the younger zones.

1. NAP – Betula zone

This assemblage zone is restricted to the minerogenic part of the sediment sequence. The pollen from the silt were well preserved and devoid of corrosion. The NAP is dominated by grasses and by Artemisia, and the open vegetation indicates that the zone was deposited during or soon after the deglaciation. The very minor changes in pollen composition within the zone may indicate that the assemblage was deposited during a very short time interval.

2. Betula zone

The Betula zone has a vertical extent of about 10 cm at the contact between the silt and the organic sediment. It is characterized by a rapid rise in the Betula curve and corresponding decline in the NAP constituents. Hippophäe has a distinct maximum within the zone.

A sample covering most of the zone was radiocarbon dated at 8780 ± 210 yrs B.P. (T–2525) on the NaOH-insoluble fraction of the gyttja. A sample from another core at the same level yielded 9080 ± yrs B.P. (T–2875) on the NaOH-soluble fraction.

Moe (1979) reported a short Betula-Hippophäe phase, similar to the one described in this paper, characteristic of the oldest organic sediment of the Hardangervidda mountain plateau. There, the deglaciation is dated at 8900–9000 B.P., which is approximately the same age as that obtained in the present investigation.

The light-demanding Hippophäe probably thrived on exposed lake bottoms and along the earth slides of the Vinstra valley fill, while being 'shadowed' out by birch on more stable ground. The birch itself was soon overtaken by the pine forests of the next zone.

3. Pinus zone

Pinus expanded rapidly and reached very high percentages just above the dated level. The pine must have constituted the main element in the forest vegetation from the early Boreal Chronozone. This implies that the bio- and chronostratigraphic zones were more or less coincident.

4. Pinus–Alnus-Betula zone

The Alnus Rise at the bottom of this zone was radiocarbon dated at 7878 ± 80 yrs B.P. (T–2874), which is the early Atlanticum Chronozone. The Q.M. constituents have been recorded only sporadically, as could be expected at this altitude. The dating of the Alnus Rise, which probably reflects the time of migration, corresponds well with the nearest lowland sites in eastern Norway (Henningsmoen 1975). Within this central area it might be taken as a fairly constant level.

The increase of Rubus ch. and decline in Cyperaceae in the upper part of the peat show a development towards the present vegetation on the bog surface. The diagram (Fig. 2) is not intended to cover the last part of the Holocene sequence.

Conclusions

The dated coarse gyttja and the terrestric peat show that the ice-dammed lake was drained before the early Boreal Chronozone. This dating indicates that the valley fill at Skåbu could not have dammed Storsjøen after the disappearance of the ice, and the existence of lake Storsjøen was thus entirely dependant on the presence of an ice dam.

The Storsjøen ice lake was situated between the main watershed and the vertically downwasting ice remnants (Garnes & Bergersen 1980). In these central parts of southern Norway the deglaciation and drainage patterns have mainly been investigated by the abundant geomorphological features rather than by stratigraphy. The dates from Skåbu indicate that stratigraphical investigations could also provide significant contributions, not least from the many sites with ice-dammed lake sediments in eastern Norway. It consequently provides a possible link between the dates from the frontal ice-recession of west Norwegian fjords (see review in Andersen 1980) and eastern Norway (Sørensen 1979), and the inland area of downwasting ice.

The vegetational succession of forest elements corresponds well with that from the nearest lowland sites in eastern Norway (Hafsten 1974, Henningsmoen 1975). Pine forests were establishes after a comparatively short-lived Betula zone. This development, and the Alnus Rise, are elements recognizable even in the mountain valley districts.

Acknowledgements. The results of this study were first presented as a lecture at the Uppsala Quaternary Symposium 1979, and form part of the author's cand.real thesis at the University of Bergen (Alstadsæter 1979, 1979 a) under the guidance of University lecturer O.F. Bergersen. Lecturer E. Sønstegaard gave valuable suggestions. O. F. Bergersen and Dr. J. Mangerud critically read through drafts of the manuscript. Miss E. Irgens of the Geological Institute, University of Bergen, drew the Pollen diagram, and Miss I. Lundquist, Norges geologiske undersøkelse, the map of Fig. 1. Dr. D. Roberts corrected the English text. I am very grateful to all these persons.

REFERENCES

Alstadsæter, I. 1979: Vegetasjonsutvikling og stratigrafi i myrbasseng 700 m.o.h. innen Vinstravassdragets bresjøområde.
Lecture at the Uppsala Quaternary Symposium, 1979.

Alstadsæter, I. 1979 a: Kvartær stratigrafi og sedimentologi i Vinstradalen, Oppland. Unpubl. thesis, Univ. of Bergen 131 pp.

Andersen, B.G. 1980: The deglaciation of Norway after 10.000 B.P. *Boreas, 9*, 211–216.

Bergersen, O. F. 1971: Kvartærgeologien i Sør-Gudbrandsdalsregionen. Report, Geol. Institute, University of Bergen, 68 pp.

Fægri, K. & Iversen, J. 1975: *Textbook of Pollen Analysis.* Munksgaard, Copenhagen, Denmark. 259 pp.

Garnes, K. & Bergersen, O.F. 1980: Wastage features of the inland ice sheet in central South Norway. *Boreas, 9*, 251–269.

Hafsten, U. 1974: Vegetasjons-, klima og jordbrukshistorie. *In* Follestad, B.A.: Tangen. Beskrivelse til kvartærgeologisk kart 1916 II. *Nor. Geol. Unders. 313*, 47–58.

Henningsmoen, K. 1975: Elghornet fra Hov i Løten. *Norsk Skogbruksmuseum, årbok nr. 7.*

Mangerud, J. 1963: Isavsmeltningen i og omkring midtre Gudbrandsdal. *Nor. geol. Unders. 223*, 223–274.

Mangerud, J., Andersen, S.T., Berglund, B.E. & Donner, J.J. 1974: Quaternary Stratigraphy of Norden, a proposal for terminology and classification. *Boreas, 3*, 109–128.

Moe, D. 1979: *Studier av vegetasjonsutviklingen gjennom Holocene på Hardangervidda.* Unpubl. thesis, Univ. of Bergen.

Ramsli, G. 1948: *Kvartærgeologiske undersøkelser i Skåbu og Espedalen.* Unpubl. thesis, Univ. of Oslo.

Rekstad, J. 1898: Mærker efter istiden i Gudbrandsdalen II. *Archiv. Math. Naturv. 20.*

Sørensen, R. 1979: Beskrivelse til kvartærgeologisk kartblad Elvdal. Nor. geol. Unders. 346, 48 pp.

Precambrian and Caledonian Tectonometamorphic Evolution of Northeastern Seiland, Finnmark, North Norway

JOHN AKSELSEN

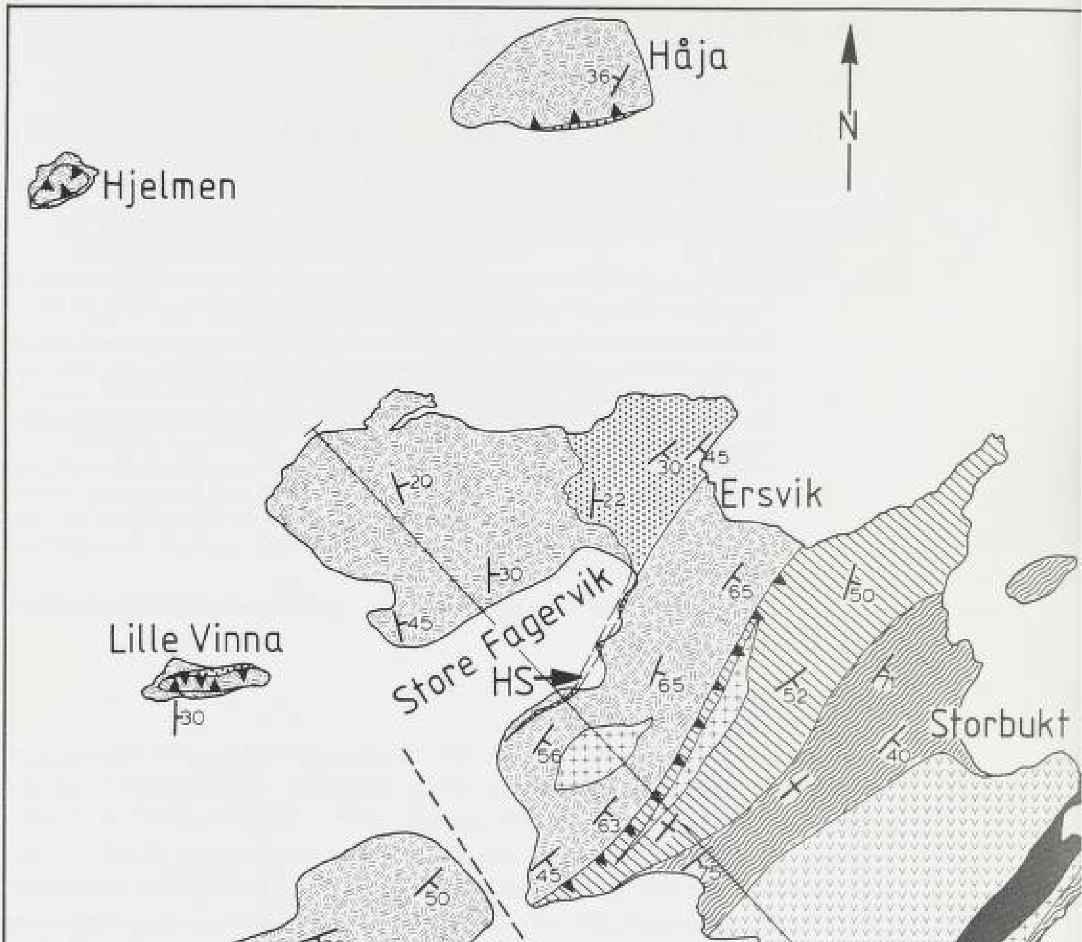
Akselsen, J., 1982: Precambrian and Caledonian tectonometamorphic evolution of Northeastern Seiland, Finnmark, North Norway. *Norges geol. Unders. 373*, 45–61.

The metamorphic rocks of northeastern Seiland are predominantly Precambrian gneisses and schists. The Fagervik infracrustal complex, composed of tonalitic gneisses, constitutes a basement to the Late Precambrian-Cambrian Klubben Psammite Group, while the Precambrian Eidvågeid supracrustal sequence occurs as the uppermost allochthonous nappe. The Precambrian units contain evidence of Precambrian deformation and metamorphism. The subsequent Caledonian tectonism is characterized by two phases of deformation, D_1 and D_2. A metamorphic climax occurred in the interkinematic period. D_1 produced large isoclinal folds and large-scale translation of the nappes. The major D_2 structure is a large asymmetric synform, the Storbukt Synform, with a vergence towards the southeast.

John Akselsen, Norsk Hydro a.s., Research Center, Lars Hilles gate 30, 5000 Bergen, Norway

Introduction

The island of Seiland, is situated to the southwest of the town of Hammerfest, within the Kalak Nappe Complex of the Finnmark Caledonides of northern Norway (Fig. 1, Roberts 1974, Sturt et al. 1975). Worthing (1971 a) made a detailed study of the lithologies, structure and metamorphism of the eastern part of the island and described a conformable sequence of pelitic rocks sandwiched between psammite units (Worthing 1971 b). This sequence constitutes the eastern envelope of the predominantly basic and ultrabasic plutons of the Seiland Petrographic Province (Robins & Gardner 1974 and references therein). Worthing (op.cit.) attempted to correlate these units with the Late Precambrian-Cambrian metasedimentary succession on Sørøy (Holland & Sturt 1970, Ramsay 1971), but could not demonstrate an equivalence. Most of the rocks in the Kalak Nappe Complex had been interpreted as Caledonian in origin (Armitage et al. 1971, Ramsay 1971), but Rb-Sr studies on perthosite instrusions and surrounding rocks suggested that at least some of the rocks in the Øksfjord area had a Precambrian origin (Brueckner 1973). Although Pringle (1975) obtained the same Rb-Sr results from similar rocks on Seiland, he regarded them as spurious because the country rocks to the dated instrusions had been correlated with the Eocambrian Klubben Psammite Group. These correlations were, however, purely lithological and may have been in error. Subsequently Ramsey & Sturt (1977) found a sub-Caledonian unconformity on Kvaløy (Fig. 1) where the Klubben Psammite Group rests unconformably on granitic gneisses. Well preserved unconformable relationships were later discovered on the island of Hjelmsøy (Ramsay et al. 1980). Following the discovery of the Kvaløy unconformity, the pelitic lithologies on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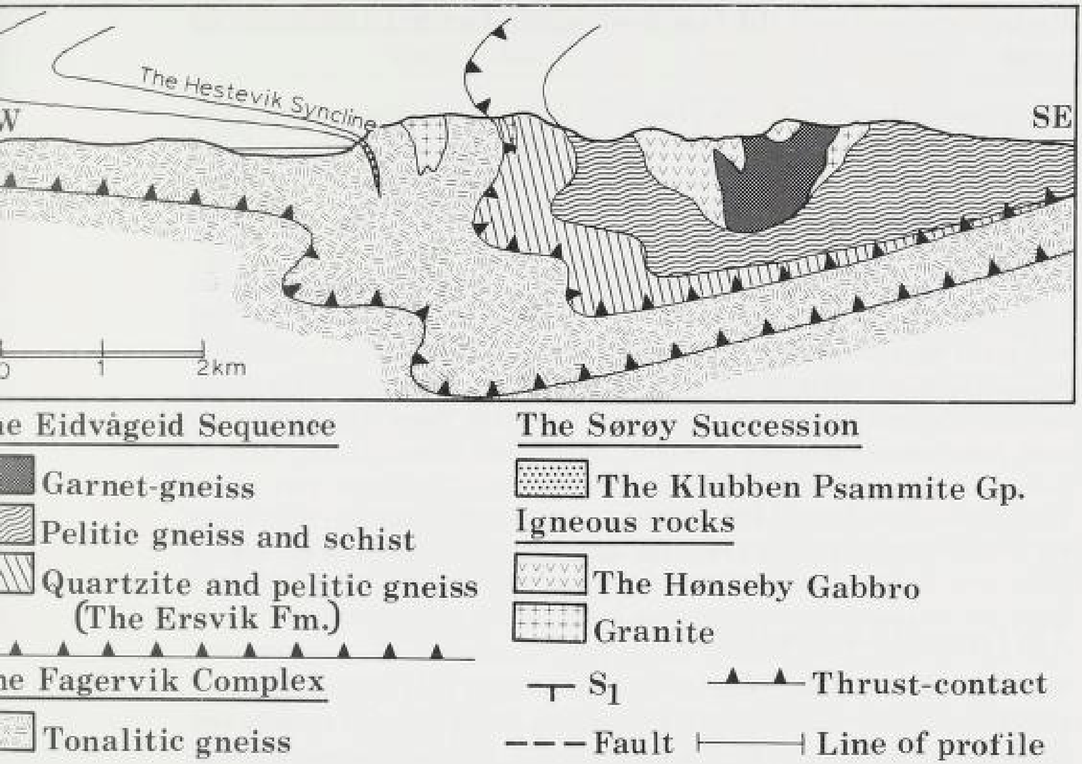

Fig. 2. Geological cross-section of Fig. 1 showing the Storbukt Synform and the refolded Hestevik Syncline.

Seiland were reinterpreted as mylonitic Precambrian gneisses (Ramsay & Sturt 1977), but the structurally lowermost and uppermost units were still correlated with the Klubben Psammite Group (Zwaan & Roberts 1978).

Mapping of northeastern Seiland by the present aothor has shown that Precambrian tectonometamorphic events are even more important on this island than was previously recognized.

Lithotectonic units

The metamorphic rocks of northeastern Seiland have been subdivided into three lithotectonic units:

1) Rocks correlated with the Late Precambrian-Cambrian metasedimentary succession on Sørøy.
2) The Precambrian Fagervik infracrustal complex.
3) The Precambrian Eidvågeid supracrustal sequence.

Fig. 1. Geological map of northern Seiland. Lithological boundaries southwest of Eidvågen are taken from Worthing (1971a). Geological cross-section with legend is shown in Fig. 2. H denotes Hammerfest in the location map and HS refers to the Hestevik Syncline.

Rocks correlated with the Late Precambrian-Cambrian succession on Sørøy

Rocks correlated with the Klubben Psammite Group occur as narrow strips in four areas on northeastern Seiland. They lie in a syncline around Store Fagervik (the Hestevik Syncline), while they occur in a series of thin thrust sheets on the neighbouring islands of Store and Lille Vinna, Hjelmen and Håja as well as on eastern Seiland (Figs. 1, 6 and 7).

The rocks of the Klubben Psammite Group are predominantly arkosic psammites and grey quartzites. The content of potash feldspar commonly is high and gives the rocks a characteristic buff colour. Locally there are semipelitic horizons up to one metre thick. Calc-silicate rocks occur along some of the thrusts on eastern Seiland (Worthing 1971a, Fig. 7) and Lille Vinna (Fig. 6); They are dark green schists containing actinolite, chlorite and biotite. Locally there are numerous quartz-lenses and thin marble horizons. The upper part of the imbrication-zone of eastern Seiland contains some rusty-weathering mica schists (Worthing 1971a). The age of the calc-silicates and mica schists is not known.

The Fagervik infracrustal complex

The Fagervik infracrustal complex is dominated by grey quartzo-feltspathic gneisses, commonly with a tonalitic composition. The gneisses are typically very homogeneous and probably have an igneous origin. They are intruded by granite to the south of Store Fagervik. Calc-silicate rocks and quartzite horizons in the gneisses on Store Vinna indicate, however, a sedimentary origin for some of the complex. The Fagervik complex thus appears to be composed of a series of ortho- and paragneisses, but a detailed subdivision has not been possible. The gneisses are lithologically similar to the gneisses which underlie the pre-Caledonian unconformity on the islands of Kvaløy and Hjelmsøy (Ramsay & Sturt 1977, Ramsay et al. 1980).

CONTACT RELATIONSHIPS WITH THE SØRØY SUCCESSION

The Klubben Psammite Group has an unconformable relationship to the Fagervik gneisses in certain areas and thrust-contacts in others. The lower boundary of the Klubben Psammite Group around the Hestvika Syncline is interpreted as a tectonized unconformity. The rocks along the contact are not generally mylonitic, but are considerably flattened, especially in long-limbs of asymmetric folds. The gneisses to the south of Store Fagervik are transformed into mylonites in a 10 m thick zone along the boundary with the Klubben Psammite Group, while they are coarse-grained with abundant cross-cutting pegmatites outside this zone.

The Eidvågeid Supracrustal sequence

A thick sequence of garnet-mica schists and gneisses, the Eidvågeid supracrustal sequence, occurs tectonically above and is separated from the Fagervik infracrustal

TABLE 1. Chemical analyses of pelitic rocks from the Eidvågeid sequence

	P44	P46	P50	P86	G48	E39a	E39b
SiO_2	60.11	64.42	60.26	57.74	61.43	60.93	60.04
Al_2O_3	19.91	18.02	16.92	22.16	18.26	20.16	20.53
TiO_2	1.12	1.07	1.36	1.26	1.13	1.07	1.01
FeO_t	7.05	6.06	8.68	7.95	7.14	7.13	6.70
MnO	0.10	0.10	0.14	0.18	0.12	0.17	0.18
MgO	3.55	2.53	2.78	2.67	3.54	3.28	3.11
CaO	2.04	1.54	2.40	1.35	2.32	1.17	1.22
Na_2O	2.79	2.21	2.16	1.74	2.68	1.59	1.44
K_2O	3.32	4.09	4.01	3.13	2.75	4.54	5.28
P_2O_5	0.27	0.13	0.17	0.33	0.16	0.29	0.29
H_2O	0.58	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Sum	100.84	100.17	98.88	98.51	99.53	100.33	99.80

complex by a major thrust-contact. Quartzite and calc-silicate rocks occur in subordinate quantities in this sequence with the exception of a zone along the boundary against the Fagervik complex where quartzites dominate over pelitic rocks. The rocks within this zone will be referred to informally as the Ersvik Formation (Fig. 1). The quartzites contain less feldspar and mica than the arkosic Klubben Psammite. The garnet gneisses, which occur in the tectonically uppermost part of the Eidvågeid sequence, contain significantly less mica than the underlying pelitic gneisses and schists. The boundary is a transitional one with gradually decreasing biotite contents. Similar lithologies occur in the tectonically uppermost unit on eastern Seiland (The Olderfjord Group of Worthing 1971 a, b).

ORIGIN OF THE GNEISSES AND SCHISTS

The gneisses and schists in the Eidvågeid sequence have high contents of mica, garnet and kyanite indicative of peraluminous bulk compositions. They are also strongly corundum normative (Cor. $= 9.2 \pm 3.1$, $n = 7$, from CIPW norm). In a plot of the Niggli values al-alk against c (Fig. 3, Niggli 1954) all analyses in Table 1 fall outside the field of common igneous rocks, suggestive of a sedimentary origin. This is supported by the presence of quartzite and calc-silicate horizons and the common occurrence of graphite as an accessory mineral.

Stratigraphic correlations

The psammitic rocks correlated with the Late Precambrian-Cambrian Klubben Psammite Group are lithologically similar to the psammites described from Sørøy (Roberts 1968, Ramsay 1971). Distinctly arkosic horizons are common as they are in the psammites on Sørøy. The tectonometamorphic evolution is also similar. This correlation is further supported by the presence of an unconformity below the psammites. This unconformity is also good evidence for a Precambrian origin of the Fagervik infracrustal complex (Ramsay & Sturt 1977). The Eidvågeid sequence is also believed to be of Precambrian age based on metamorphic criteria (see below).

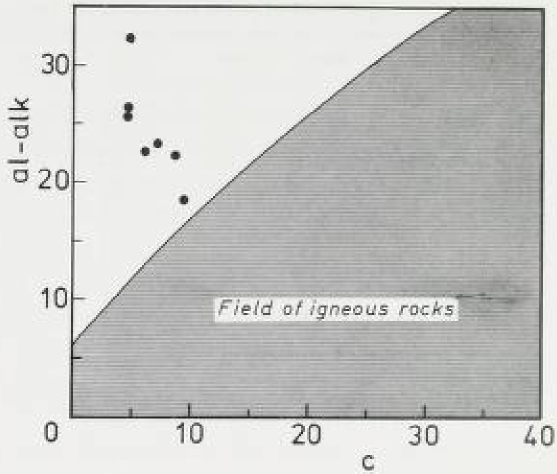

Fig. 3. Niggli plot of schists from the Eidvågeid sequence. The plot clearly shows that these rocks could not have had an igneous origin.

It is frequently difficult to distinguish the gneisses of the Fagervik complex from the Klubben Psammites when they occur in zones of high strain. The presence of abundant feldspar porphyroclasts, the mylonite-foliation and tectonic lenses of less severely deformed gneiss indicate, however, that these rocks had a more coarse-grained parent.

The calc-silicate rocks described from the imbrication zones are similar to the lower calc-silicate unit in the Cambrian Falkenes Marble group on Sørøy (Ramsay 1971). They occur in association with the Klubben Psammite, and might be correlated with parts of the Sørøy succession. The rusty-weathering mica schists in the imbrication zone of eastern Seiland (Fig. 7) may be equivalent to the Storelv Schist group which overlies the Klubben Psammite on Sørøy (Ramsay 1971), but such lithological correlations can only be tentative.

Tectonic structures

CALEDONIAN STRUCTURES

The Klubben Psammite shows evidence for two phases of tectonic deformation, designated as D_1 and D_2. Their associated schistosities are termed S_1 and S_2, respectively. The correlation of these rocks with the Late Precambrian-Cambrian succession on Sørøy implies a Caledonian age for these deformational events.

D_1 minor structures in the Klubben Psammite Group

D_1 folds are uncommon in the Klubben Psammite. Those which have been observed have an isoclinal style with an axial-planar schistosity marked by parallel mica flakes, although in many cases the mica content is so low that no schistosity is visible. Because of the nature of the smooth weathering surfaces, it proved impossible to take accurate measurements of D_1 fold axes. However, they are believed to have an approximate N–S trend like the pre-D_2 lineation in the surrounding gneisses.

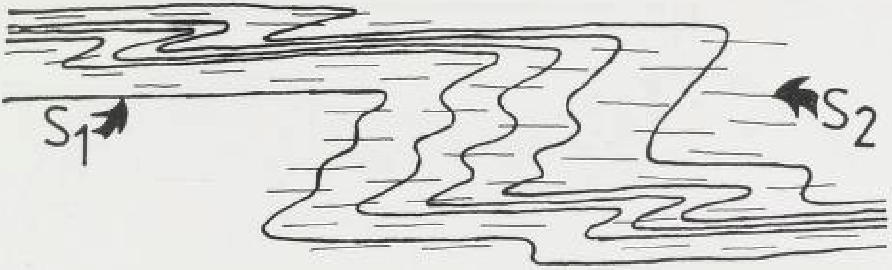

Fig. 4. Schematic figure showing typical style of asymmetric D_2 fold with strongly flattened long-limbs and open buckle-folds in the steep middle-limb.

D_1 minor structures in the Precambrian rocks

Isoclinal folds refolded by D_2 folds are very common in the Fagervik infracrustal complex. Owing to the high strain facies, however, it is impossible in most cases to distinguish Caledonian D_1 structures from possible Precambrian structures. Pre-D_2 linear fabrics have a N–S trend (Fig. 5c).

The main tectonic foliation in the Eidvågeid sequence is folded by D_2 and is considered to correspond to S_1 in the Klubben Psammite Group, but it probably represents a reactivated and transposed Precambrian foliation. It will, for convenience, be referred to as S_1

D_2 minor structures in the Klubben Psammite Group

D_2 folds are more common than D_1 folds. The style of the D_2 folds is variable and depends on their location in larger fold structures and on the intensity of deformation. They are generally asymmetric with steep middle-limbs which are overturned to varying degrees. Parasitic folds in middle-limbs of larger folds are open buckle folds while those in the long-limbs are asymmetric, tighter and show thickened hinge-zones and usually have a penetrative new schistosity (Fig. 4). The D_2 folds are cylindrical with a NNE-NE axial trend (Fig. 5b).

D_2 minor structures in the Precambrian rocks

The D_2 deformation was stronger to the north of Store Fagervik and to the south of the garnet gneisses (Fig. 1) as compared to the area between. This is a consequence of the location of these two areas in flat-lying long-limbs of a major asymmetric D_2 fold, while the area between is situated in a steep middle-limb (Fig. 2).

The D_2 folds in the Fagervik complex have morphologies similar to the D_2 folds in the Klubben Psammite Group (fig. 4). Crenulation folds are frequently developed in the schists of the Eidvågeid sequence. To the south of Store Fagervik the D_2 fold axes have a shallow plunge towards the NE (Fig. 5b) parallel to the axes of D_2 folds in the Klubben Psammite Group. To the north of Store Fagervik the attitudes of the D_2 folds show a large variation within their mean axial surface (Fig. 5a). The variable attitudes of the axes is at least partly caused by noncylindrical folding which frequently can be seen on a small scale (Ramsay &

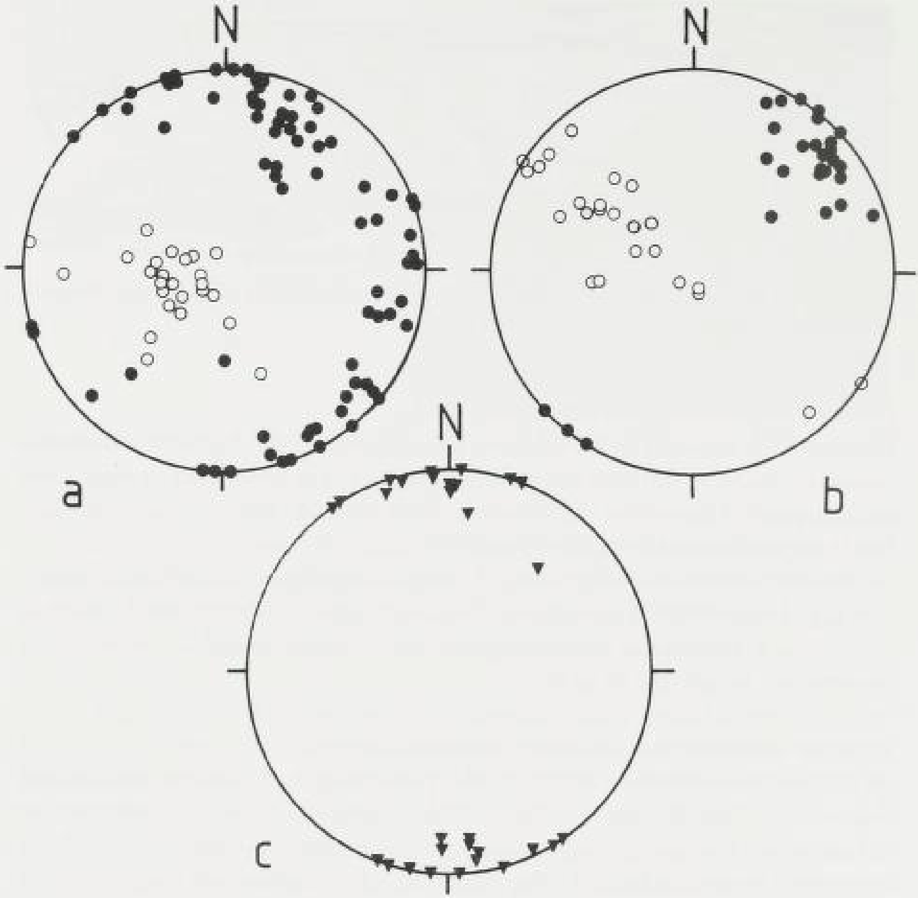

Fig. 5. Stereographic plots (equal area) of D_2 fold axes (filled circles) and axial surfaces (open circles) from (a) the gneisses to the north of Store Fagervik and (b) the Klubben Psammite in the Hestevik Syncline and the gneisses to the south of Store Fagervik. (c) D_1 lineation to the north of Store Fagervik.

Sturt 1973), but some of the folds shown in Fig. 5b might be of an older generation.

MAJOR STRUCTURES

Thrusts

A major thrust-contact occurs on the islands of Lille Vinna, Hjelmen and Håja (Fig. 6). On Lille Vinna it is underlain by seven, thin, discontinuous thrust sheets of the Klubben Psammite, calc-silicate schists and gneisses with a total thickness of 30–80 metres (Fig. 6). The thrust sheets are sandwiched between two thick units of mylonite-gneisses. The mylonite foliation and a NNE-trending linear fabric along the thrusts are folded by small-scale D_2 folds, suggesting that thrusting occurred during D_1.

Worthing (1971a) mapped rocks from eastern Seiland as correlative with the

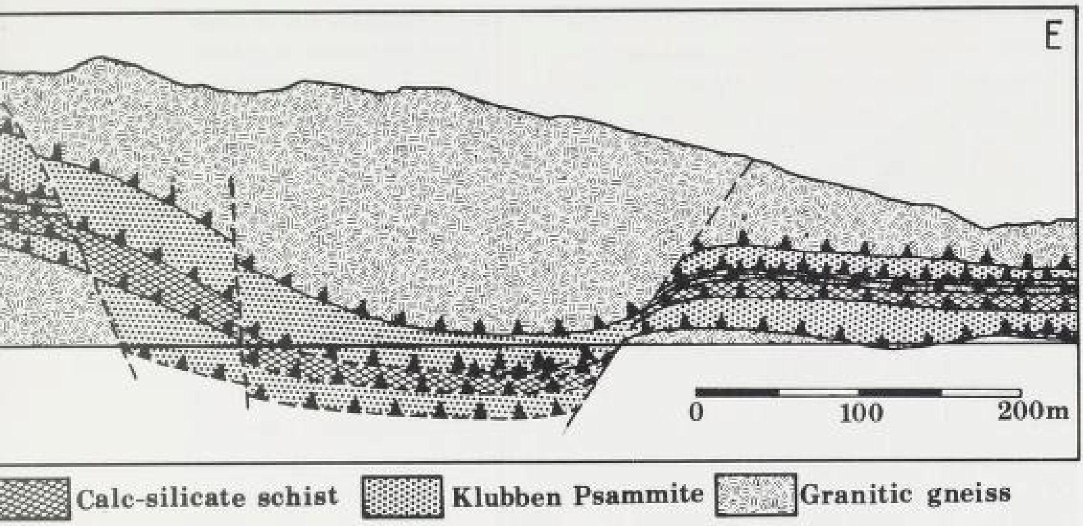

Fig. 6. Geological cross-section of the central part of Lille Vinna, showing the imbrication sheets of Klubben Psammite, calc-silicate schists and gneisses sandwiched between Fagervik gneiss.

Klubben Psammite Group. However, there is a complicated imbrication-zone in eastern Seiland (Fig. 7). The major part of the lowermost 'psammite' unit of Worthing (1971a, b), the Komagnes Group, is composed of grey gneisses similar to those described from the Fagervik infracrustal complex. This area has therefore been remapped and reinterpreted by the present author (Fig. 7) as largely Fagervik equivalent. The Klubben Psammite Group does occur in several thin thrust sheets. Worthing (1971a) mapped one of these and called it the 'pink psammite', but considered it to be an integral part of the surrounding gneisses, all of which he correlated with the Klubben Psammite Group. Some of the sheets shown in Fig. 7 are composed of a series of extremely thin units of alternating psammites and mylonitic and phyllonitic gneisses well below mappable thickness. It thus appears that this imbrication zone consists of a series of thin nappes with large lateral extensions.

The D_1 Hestevik Syncline

The Klubben Psammite Group around Store Fagervik occurs in the core of a large D_1 syncline, the Hestevik Syncline (Fig. 1). The hinge can be seen in a steep cliff to the south of Store Fagervik where the sedimentary layering in the psammite and its contact with the gneisses are isoclinally folded. The hinge zone has been dramatically thinned and refolded by D_2 folds.

The D_2 Storbukt Synform

The Storbukt Synform is the largest D_2 structure on northeastern Seiland (Fig. 2). It has a fold axis with a shallow plunge towards the NE. Several independent observations indicate the presence of this large synform. There is a two-sided areal

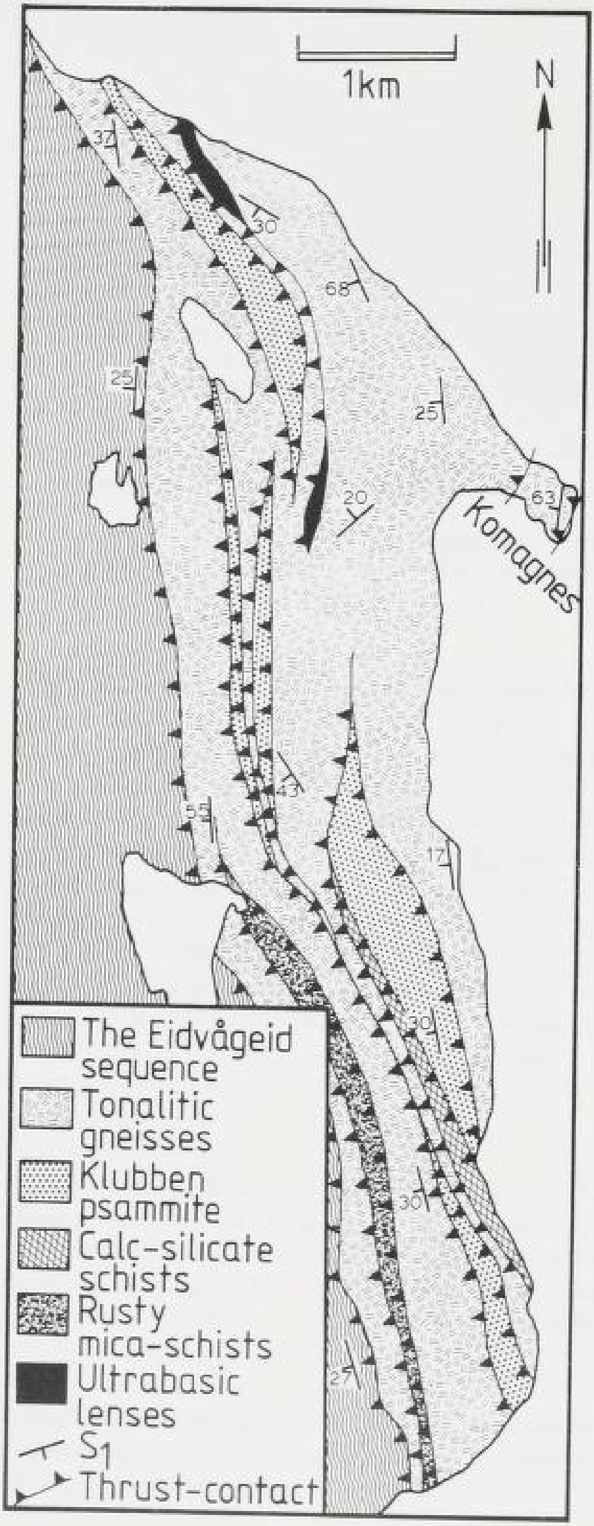

Fig. 7. Geological map of eastern Seiland showing the imbrication sheets. Location map shown in Fig. 1.

distribution of the gneisses of the Fagervik infracrustal complex with the Eidvågeid sequence located between. The northern boundary between these has a steep southeastward dip, while the southeastern boundary has a shallow dip towards the west and northwest where the variation is caused by a strike-swing across Eidvågen (Fig. 1).

The D_2 folds in the area between Ersvika and Store Vinna commonly have an open style and near horizontal axial surfaces as one would expect in a steep middle-limb of an asymmetric fold (Fig. 4). The geometry of the D_2 folds in the area around Fiskebukt and to the north of Store Fagervik indicate, on the other hand, a much stronger flattening and have a vergence towards the SE. All these observations suggest the presence of a large asymmetric synform (Fig. 2). The folds of assumed D_2 age to the south of Eidvågen have, however, a vergence towards the west. This might be a result of different orientations of the layering to the north and south of Eidvågen prior to D_2 (Ramsay 1967, p. 538), which could also explain the strike-swing. Alternatively these folds could be of a younger generation.

PRECAMBRIAN STRUCTURES

Ramsay & Sturt (1977) and Ramsay et al. (1980) described tectonic foliations in gneisses from Kvaløy and Hjelmsøy that are cut by granite dykes which are in turn truncated by the unconformity below the Klubben Psammite Group. This relationship gives evidence for Precambrian deformation. The granitic intrusion to the south of Store Fagervik appears to be younger than much of the deformation and migmatization in the surrounding gneisses, but is affected by the D_1 deformation, especially along its margins. The earliest deformation in the surrounding gneisses is therefore probably of Precambrian age. The deformation of a gneissic layering cut by granitic pegmatites around the D_1 Hestevik Syncline also indicates Precambrian deformation. Internal foliations in garnet and orthoclase porphyroblasts (Fig. 8) give evidence for deformation which predated a high-grade Precambrian metamorphism in the Eidvågeid sequence (see below).

Metamorphism

METAMORPHISM IN THE KLUBBEN PSAMMITE GROUP

Biotite and muscovite have crystallized or recrystallized parallel to the axial surfaces of D_1 and D_2 folds in semipelitic horizons in the Klubben Psammite Group. Xenoblastic garnets have overgrown S_1 while S_2 bends around them suggesting post-D_1/pre- or syn-D_2 garnet growth. The garnet/biotite geothermometer gives temperatures around 585°C from the calibrations of Thompsom (1976) and Ferry & Spear (1978). This temperature must, however, be regarded as only a minimum temperature since the garnets contain up to 37 mole percent grossular + spessartine (Kretz 1958, Albee 1965). From these observations it appears that the P–T conditions increased from D_1 and reached a climax in pre- or syn-D_2 time. This metamorphism must be of Caledonian age based on the correlation with the Late Precambrian-Cambrian Klubben Psammite Group. The tectonometamorphic evolution is similar to that described from Sørøy (Roberts 1968).

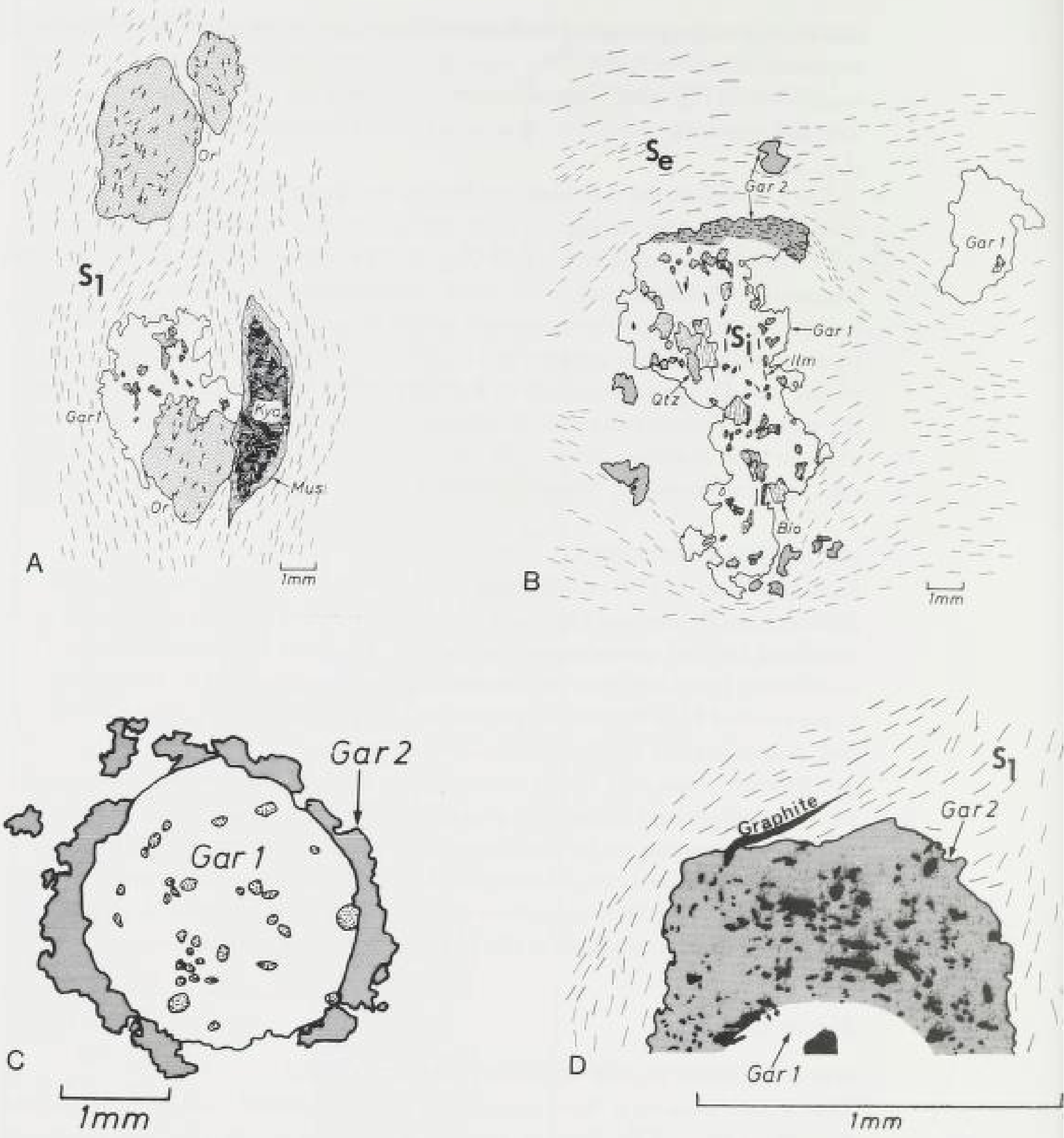

Fig. 8. Textures in pelitic schists from the Eidvågeid sequence. The distinction between garnet 1 and garnet 2 based on textural features and microprobe-analyses (Akselsen 1980 and in prep.).
A) S_1 deflected around pre-D_1 porphyroblasts in the Ersvik formation.
B) Garnet 1 with planar, internal schistosity, S_i, overgrown by interkinematic D_1–D_2 garnet 2.
C) Garnet 1 overgrown by garnet 2.
D) Graphite crystal partly overgrown by garnet 2 and subsequently bent during D_2

METAMORPHISM IN THE EIDVÅGEID SEQUENCE

The Eidvågeid sequence has been affected by two distinct phases of metamorphism which are believed to be of Precambrian and Caledonian age, respectively.

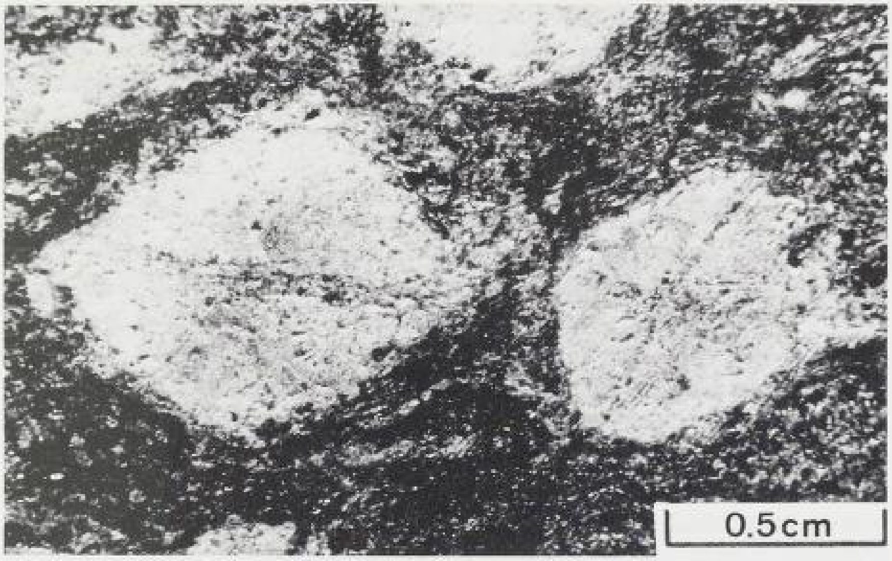

Fig. 9. Kyanite crystals from pelitic schists in the Eidvågeid sequence possibly representing pseudomorphed andalusite.

Precambrian metamorphism

The most pervasive mineralogical and textural reconstitution of the Eidvågeid sequence can be related to a Precambrian metamorphism when coarse-grained and pegmatite-veined lithologies were produced, especially in the tectonically intermediate and upper parts of the sequence. Many of the pegmatites have been disrupted during subsequent deformation which has impressed a cataclastic fabric on the rocks. The Ersvik Formation contains porphyroblastic orthoclase. Unlike the K-feldspar porphyroclasts from the disrupted pegmatites, the porphyroblasts contain numerous inclusions of biotite which define a planar, internal schistosity (Fig. 8a). This schistosity is truncated by S_1 in the surrounding matrix, showing that the orthoclase overgrew a schistosity prior to or possibly at an early stage in the Caledonian D_1 deformation. These porphyroblasts co-existed with pre- or possibly early-D_1 alumino-silicates (Fig. 8a) implying high-grade conditions (Winkler 1979) during this metamorphism. Kyanite locally occurs in prismatic pseudomorphs with rounded, square and diamond-shaped cross-sections (fig. 9). Within the prisms it has recrystallized with (100) in random orientations in the prismatic zone. Many of the diamond-shaped pseudomorphs contain a diagonal zone of quartz grains and in a few cases two such zones have been seen to form a cross (Fig. 9). Bechennec & Herve (1973, 1974) have described similar pseudomorphs from the islands of Arnøy and Laukøy in north Troms and interpreted them as pseudomorphs after andalusite. This is supported by the diagonal cross in some of them, which resembles chiastolite. Alternatively, they may represent pseudomorphed prismatic sillimanite.

Early garnets (garnet 1) in the tectonically intermediate and upper parts of the

Eidvågeid sequence formed pre- or possibly early-D_1 as shown by the deflected S_1 foliation. An internal, planar schistosity defined by parallel biotite and ilmenite has only been observed in one garnet (Fig. 8b). The largest garnets are generally surrounded by felspar-rich mantles.

Caledonian metamorphism

Garnets of a second generation (garnet 2) contain numerous tiny inclusions, mainly quartz and rutile, rather than the coarser inclusions in garnet 1. In many cases they have grown on the first generation garnets (Fig. 8c). Fig. 8b shows an early garnet (garnet 1) with an internal foliation, S_i, overgrown by a later garnet (garnet 2). The latter contains a planar internal schistosity which is discordant to S_i and parallel to the external foliation, S_e, except where S_e is locally deflected around the garnets. S_e is polyphasal due to coincidence of S_1 and S_2 in a long-limb of a D_2 fold (Fig. 4). These relationships show that the two generations of garnets grew during two distinct phases separated by a period when the external foliation was rotated relative to garnet 1. Muscovite which formed simultaneously with garnet 2 (Akselsen 1980 and in prep.) has overgrown S_1, suggestive of post-D_1 growth of garnet 2. Fig. 8D shows a graphite crystal parallel to S_1 partly overgrown by garnet 2. After cessation of garnet growth, the graphite was bent by a D_2 crenulation, which is the only D_2 structure seen in the schists to the north of the Hønseby gabbro where this specimen was collected. This shows that at least this garnet ceased growing prior to the main D_2 deformation in this area. It thus appears that the second generation of garnet grew in the interkinematic D_1–D_2 period.

The major difference in the metamorphic evoluton of the Eidvågeid sequence and the Klubben Psammite Group is the high-grade pre- or possibly early-D_1 metamorphism in the former. This early event is unlikely to be a contact-metamorphic effect of the basic and ultrabasic intrusions in central Seiland, as high-grade assemblages formed in the gneisses as far as 5 km from the intrusions. Such high-grade conditions can also be found on the island of Kvaløy where igneous intrusions only occur in subordinate amounts (Jansen 1979). The post-D_1 tectonometamorphic evolution of the Eidvågen sequence is similar to that described from the Klubben Psammite Group on Seiland and from the island of Sørøy (Roberts 1968). This must, therefore, be of Caledonian age (Sturt et al. 1978), while the early (pre-D_1) metamorphism is considered to be of Precambrian age. This supports the hypothesis of a Precambrian origin for the Eidvågeid sequence, despite its tectonic position above the Klubben Psammite Group.

The presence of kyanite pseudomorphs after andalusite or sillimanite suggests that the Precambrian metamorphism was of a lower pressure type than the Caledonian metamorphism. The detailed evidence supporting this observation will be published elsewhere.

The Hønseby gabbro

The Hønseby Gabbro south of Eidvågen was briefly described by Worthing (1971a). The gabbro also extends northeast of Eidvågen slightly offset by a fault

(Fig. 1). Further to the northeast the gabbro thins and is only found as minor boudins on Kvaløya (Ø. Jansen pers.comm. 1979). Worthing (1971a) concludes from structural relationships that the gabbro is of Caledonian syn-D_1 age. It therefore has been regarded as one of the oldest intrusions in the Seiland Petrographic Province (Robins & Gardner 1975). However, the Hønseby Gabbro on NE Seiland is surrounded by metasediments of the Eidvågeid sequence which have suffered Precambrian deformation and metamorphism. There is consequently no a priori reason to reject Precambrian ages for some of the intrusions on Seiland as argued by Pringle (1975).

The Hønseby Gabbro has a granoblastic texture and contains pyroxenes, plagioclase, hornblende, biotite, quartz and Fe-Ti oxides. Locally it appears to contain two generations of garnet. The largest garnets contain coarse inclusions while the second generation garnets contain numerous tiny inclusions and have grown, together with quartz, as coronas between orthopyroxene and plagioclase. Texturally this is somewhat similar to the garnets described from the surrounding metasediments and may indicate an early (Precambrian) medium-pressure granulite-facies metamorphism followed by a Caledonian metamorphism at lower temperatures and/or higher pressure (Green & Ringwood 1967). Similar textures have been ascribed to polymetamorphic evolution by Wagner & Crawford (1975). However, Griffin & Heier (1973) regarded the formation of such coronas as retrograde features solely due to cooling during uplift.

Conclusions

Northeast Seiland provides a typical example of cover-basement relationships in the Kalak Nappe Complex. The youngest metasediments, correlated with the Vendian-Cambrian Sørøy succession (Holland & Sturt 1970, Ramsay 1971), were affected by two phases of Caledonian deformation, D_1 and D_2 The metamorphic climax was reached in the D_1–D_2 interkinematic period. In some places these rocks rest unconformably on a Precambrian basement complex, and in others occur as thin thrust sheets in complex imbrication zones. It is possible to subdivide the Precambrian basement into two lithotectonic units: The Fagervik infracrustal complex, generally composed of tonalitic gneisses, and the Eidvågeid supracrustal sequence, dominated by pelitic gneisses and schists. Structural, textural and mineralogical relationships indicate that the Eidvågeid sequence suffered a high-grade Precambrian metamorphism prior to a weaker Caledonian metamorphism. This metamorphic evolution clearly suggests a Precambrian age for these metasediments. The main overthrusting event in this area occured during D_1 and the nappes were subsequently folded into a large asymmetric synform, the Storbukt Synform. The latest deformation involved faulting with brecciation and associated diaphthoresis.

The Eidvågeid supracrustal sequence is separated from the Fagervik infracrustal complex by a thrust, and hence the age relationships between these two units is unknown. Ramsey & Sturt (1977) reported preliminary Rb/Sr isochrons of 1469 $\pm$ 70 my. and 2660 $\pm$ 150 my for granitic dykes that cut a high-grade migmatitic gneiss complex in the Skillefjord Nappe on the mainland to the south of Seiland. An Archaean age for these gneisses appears to be conformed by further Rb/Sr isocrons of approximately 3000 my from similar granitic dykes (Austerheim & Sturt pers. comm. 1981), but it is possible that several Precambrian age-provinces are represented in the gneisses of the Kalak Nappe Complex. Further more detailed geochronological work is obviously necessary to solve this problem.

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REFERENCES

Akselsen, J. 1980: Tectonic and metamorphic development of the Precambrian and Caledonian rocks of NE. Seiland, Finnmark. Unpubl. Cand. Real. thesis, Univ. of Bergen, 351 pp.

Albee, A.L. 1965: Distribution of Fe, Mg and Mn between garnet and biotite in natural mineral assemblages. *J. Geol. 73,* 155–164.

Armitage, A.H., Hopper, R.R., Lewis D. & Pearson, D.E. 1971: Stratigraphic correlation in the Caledonian rocks of southwest Finnmark and North Troms. *Norges geol. Unders. 269,* 318–322.

Bechennec, F. & Herve, A. 1973: Etude des batis cristallophylliens des iles d'Arnøy et de Laukøy (Troms, Norvege septentionale), analyse structurale et petrographique. Unpubl. thesis, Orsay.

Bechennec, F. & Herve, A. 1974: Metamorphism.-Decounverte d'un metamorphisme general precoce de type basse pression dans les Caledonides du Nord, Iles d'Arnøy et de Laukøy (Troms, Norvege). *C.R. Acad. Sc. Paris, 279 (D),* 239–242.

Brueckner, H.K. 1973: Reconnaissance Rb-Sr investigation of salic, mafic and ultrabasic rocks in the Øksfjord area, Seiland Province, northern Norway. *Norsk geol. Tidsskr. 53,* 11–23.

Brueckner, H.K. 1975: A. reply. Reconnaissance Rb-Sr investigation of salic, mafic and ultramafic rocks in the Øksfjord area, Seiland Province, N. Norway. *Norsk geol. Tidsskr. 55,* 101–103.

Ferry, J.M. & Spear, F.S. 1978: Experimental calibration of the partitioning of Fe and Mg between biotite and garnet. *Contrib. Mineral. Petrol. 66,* 113–117.

Green, D.H. & Ringwood, A.E. 1967: An experimental investigation of the gabbro to eclogite transformation and its petrological applications. *Geochim. Cosmochim. Acta, 31,* 767–833.

Griffin, W.L. & Heier, K.S. 1973: Petrological implications of some corona structures. *Lithos, 6,* 315–335.

Holland, C.H. & Sturt, B.A. 1970: On the occurence of archaeocyathids in the Caledonian metamorphic rocks of Sørøy, and their stratigraphic significance. *Norsk geol. Tidsskr. 50,* 345–355.

Jansen, Ø. 1979: Kvaløy, Vest-Finnmark – Et typisk eksempel på den tektonometamorfe utvikling i Kaladekkekomplekset. *Abstract, 14. Nordiske geol.v.møte 1980, Norsk geol. Forening, Geolognytt, 13, 34.*

Kretz, R. 1958: Chemical study of garnet, biotite and hornblende from gneisses of southwestern Quibec, with emphasis on distribution of elements in coexisting minerals. *J. geol. 67,* 371–402.

Niggli, P. 1954: Rocks and mineral deposits. Freeman, San Francisco, 559 pp.

Pringle, I.R. 1975: A comment. Reconnaissance Rb-Sr investigation of salic, mafic and ultramafic rocks in the Øksfjord area, Seiland province, northern Norway. *Norsk geol. Tidsskr. 55,* 97–99.

Ramsay, D.M. 1971: Stratigraphy of Sørøy. *Norges geol. Unders. 269,* 314–317.

Ramsay, D.M. & Sturt, B.A. 1973: An analysis of non cylindrical and incongrous fold patterns from Eocambrian rocks of Sørøy, Northern Norway. *Tectonophysics, 18,* 81–107.

Ramsay, D.M. & Sturt, B.A. 1977: The significance of the sub-Caledonian unconformity within the nappe sequence of the Finnmarkian belt. *Norges geol. Unders. 334,* 107–116.

Ramsay, D.M., Sturt, B.A. & Andersen, T.B. 1980: The sub-Caledonian unconformity on Hjelmsøy – New evidence for primary basement/cover relations in the Finnmarkian nappe sequence. *Norges geol. Unders. 351,* 1–12.

Ramsay, J.G. 1967: *Folding and fracturing of rocks.* McGraw Hill, 568 pp.

Roberts, D. 1968: The structural and metamorphic history of the Langstrand–Finnfjord area of Sørøy, North Norway. *Norges geol. Unders. 253,* 160 pp.

Roberts, D. 1974: Hammerfest. Description of the 1:250 000 geological map. *Norges geol. Unders. 301*, 66 pp.

Robins, B. & Gardner, P.M. 1974: Synorogenic Layered basic instrusions in the Seiland Petrograhic Province, Finnmark. *Norges geol. Unders. 312*, 91–130.

Robins, B. & Gardner, P.M. 1975: The magmatic evolution of the Seiland Province and Caledonian Plate boundaries in northern Norway. *Earth and Plan. Schi.Let. 26*, 167–178.

Sturt, B.A., Pringle, I.R. & Roberts, D. 1975: Caledonian nappe sequence of Finnmark, Northern Norway, and the timing of orogenic deformation and metamorphism. *Geol. Soc. Am. Bull., 86*, 710–718.

Sturt, B.A., Pringle, I.R. & Ramsay, D.M. 1978: The Finnmarkian phase of the Caledonian Orogeny. *J. Geol.Soc.Lond. 135*, 597–610.

Thompson, A.B. 1976: Mineral reactions in pelitic rocks: II Calculations of some P–T–X (Fe-Mg) phase relations. *Am.J.Sci., 276*, 653–682.

Wagner, M.E. & Crawford, M.L. 1975: Polymetamorphism of the Precambrian Baltimore Gneiss in southeastern Pennsylvania. *Am.J.Sci. 275*, 653–682.

Winkler, H.G.F. 1979: *Petrogenesis of metamorphic rocks.* Springer–Verlag, 334 pp.

Worthing, M.A. 1971a: Studies in the structure and metamorphic petrology of the Eocambrian rocks of Eastern Seiland, North Norway. Unpubl. Ph.D.thesis, Univ. of London.

Worthing, M.A. 1971b: The stratgraphy of part of eastern Seiland, West Finnmark. *Norges geol. Unders. 269*, 323–327.

Zwaan, K.B. & Roberts, D. 1978: Tectonostratigraphic succession and development of the Finnmarkian nappe sequence, North Norway. *Norges Geol. Unders. 343*, 53–71.

Disparate Geochemical Patterns from the Snåsavatn Greenstone, Nord-Trøndelag, Central Norway

DAVID ROBERTS

Roberts, D. 1982: Disparate geochemical patterns from the Snåsavatn greenstone, Nord-Trøndelag, Central Norway. *Norges geol. Unders. 373*, 63–73.

Greenstone volcanites of Middle Ordovician age from the Snåsavatn area of the Central Norwegian Caledonides can be divided into two distinctive groups on the basis of their geochemical patterns. The bulk of the lavas sampled are tholeiites of ocean floor affinity, though with a hint of transition towards within-plate features in some minor and trace element contents. These OFB rocks are positioned structurally and stratigraphically above the second group of greenstone lavas which are basalts of calc-alkaline chemical character. The CAB lavas are, in turn, stratigraphically underlain by the Snåsa limestone which contains a varied fauna of Lower to Middle Ordovician age. Regional-geological, biostratigraphic and volcanite-geochemical considerations pinpoint similarities between the palaeotectonic situations for the Snåsavatn volcanosedimentary assemblage and that from the island of Smøla some 200 km along strike to the southwest. The influx of OFB volcanites at Snåsavatn, however, is not recorded on Smøla. This upward transition to more primitive tholeiitic basalts is thought to relate to a subduction zone and arc migration, with ocean accretion occurring in an extensional, back-arc marginal basin setting.

D. Roberts, Norges geologiske undersøkelse, Postboks 3006, N–7001 Trondheim, Norway

Introduction

Studies of the geochemical character of Lower Palaeozoic basaltic and andesitic greenstones over the past decade have provided valuable contributions to Caledonide research in the central Norwegian part of the mountain belt. In 1979, systematic sampling of the schistose greenstone volcanites of the Snåsavatn area of Nord-Trøndelag was carried out as a part of an ongoing NGU project 'Grønnstein geokjemi innen de sentrale norske kaledonidene'. Preliminary results of this investigation of the Snåsavatn lavas revealed an interesting bipartite grouping, with the bulk of the volcanites showing ocean floor tholeiite characteristics and a smaller population of samples from a structurally lower zone having a distinctive calc-alkaline chemistry (Roberts 1980a). Subsequent sampling was concentrated on determining the extent of the calc-alkaline association. This short article aims at reporting the principal results of this study and briefly discussing their regional significance.

Geological setting

The greenstone unit sampled forms part of an association of lithologies which Foslie (1958) collectively termed the Snåsa Group. Earlier, Carstens (1956) had noted that these greenschist to epidote-amphibolite facies rocks – schists, limestones, greenstones, conglomerates and sandstones – were deformed in a narrow, NE–SW-trending syncline, his Snåsa Syncline. Further mapping, aided by fossil finds in the Snåsa limestone, led Carstens (1960) to the view that the greenstones and limestones at least (Fig. 1), formed part of the Lower Hovin Group

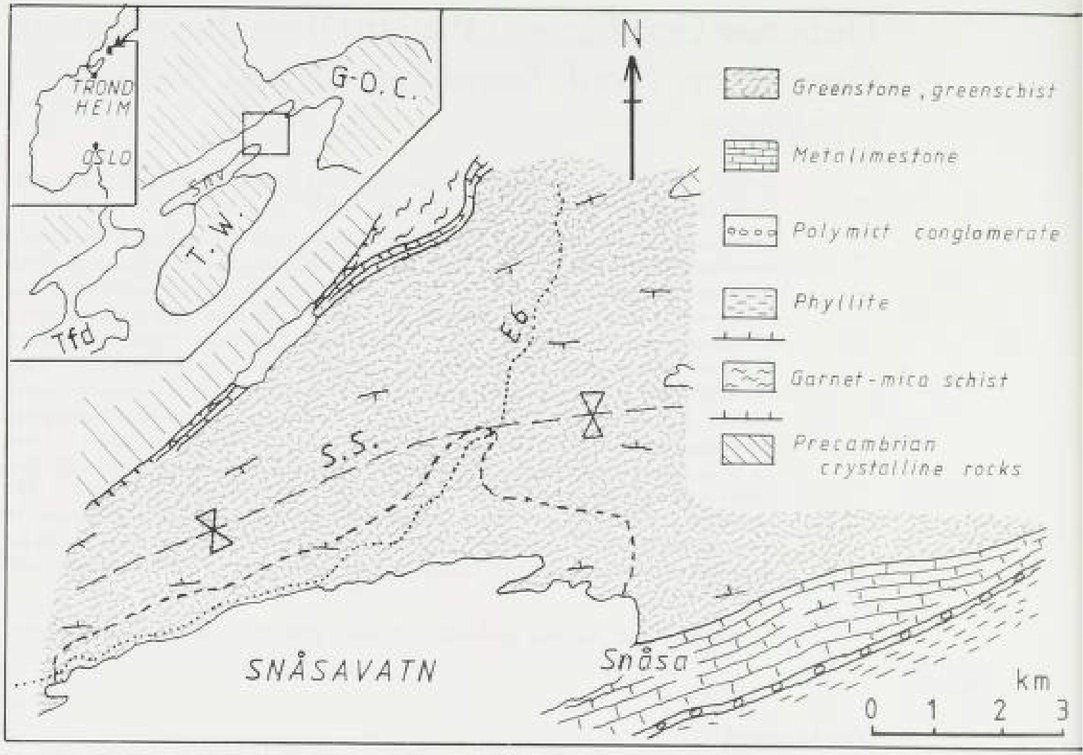


Fig. 1. Simplified geological map of the NE Snåsavatn area. S.S. – Snåsa Synform. All except 2 of the samples analysed in this study are from the area contained by the dashed line between the shores of Snåsavatn and the synform axial trace. In the inset map, T.W. – Tømmerås window; G.–O.C. – Grong–Olden Culmination; Snv – Snåsavatn; Tfd – Trondheimsfjord. The geology is taken from Roberts (1967). The E6 highway is indicated by the dotted line.

of the classical Trondheim region stratigraphy (Vogt 1945), which is part of the Støren Nappe (Gale & Roberts 1974). Both Springer Peacey (1964) and Roberts (1967) adopted the informal Snåsa Group designation in their reconnaissance studies. In the NE Snåsavatn area the sequence is essentially a structural one since no primary structures have yet been found. Moreover, as the major 'syncline' deforms the penetrative regional schistosity and associated sub-isoclinal folds (Roberts 1967) it is more appropriate to refer to it as a synform.

More recent work in the area from the Tømmerås Antiform across the northeastern termination of the Snåsa Synform into the Grong–Olden Culmination (Fig. 1, inset) has pointed to tectonostratigraphical similarities between this region and areas in Sweden (e.g. Gee 1977, Andreasson & Johansson 1982), denoting that Foslie's (1958) original Snåsa Group embraces lithologies from 4, or maybe 5, separate nappes. Similar subdivisions have been traced southwestwards into the Steinkjer and Inderøy–Leksdalsvatn areas by Tietzsch–Tyler (in prep.) and Roberts (unpubl. mapping), respectively. Some elements of this tectonostratigraphy are shown on the 1:250 000 map-sheet 'Trondheim' (Wolff 1976), although amendments have been noted (Roberts & Wolff 1981) and other revisions will follow.

The Snåsavatn greenstone

The predominant lithology of the unit, here informally termed the Snåsavatn greenstone, is a green to pale green moderately schistose rock of generally medium- to medium-fine grain. Some parts are more schistose and may represent mafic tuffs; however, these sometimes correspond to zones of higher strain. Other lithologies represented sporadically within the 1.5–1.8 km-thick unit are intermediate tuffs, keratophyric horizons and agglomerates. Limestone bands and ribs are not uncommon within the basal parts of the volcanic pile (Roberts 1967). Despite an extensive search for pillow structure, nothing has been observed which could be classified as definite pillow form. Carstens (1956) also noted that pillow structure had not been found over an extensive area from Malm to Snåsa. This contrasts with other Ordovician greenstone sequences from other parts of Trøndelag, where pillow structure is sometimes prominently developed (e.g. Grenne & Roberts 1980).

PETROGRAPHY

The greenstones show complete recrystallisation to greenschist to epidote-amphibolite facies assemblages. Primary igneous textural relics are rare, although these may be discerned in some thin-sections through a blanket of saussuritised plagioclase and chloritised and epidotised amphibole. The amphibole is an actinolite which, together with chlorite, defines the foliation in the rock. Other minerals present are epidote, Ti-magnetite, apatite, calcite and assessory leucoxene.

Geochemistry

SAMPLING

The initial sampling was concentrated along a traverse across the southeastern limb of the Snåsa Synform (Fig. 1), with the well exposed road-cuts along the new E6 highway as principal targets (Roberts 1980a). Thirty-six fresh samples of variably schistose greenstone were taken. Follow-up sampling, subsequent to examination of the initial analytical data, provided a further 10 analyses. Analytical procedures are given in an appendix.

MAJOR ELEMENT CHEMISTRY

On initial examination of the raw analytical data from 1979 it was evident that a small number of analyses from one part of the greenstone formation were different from the bulk of the samples in showing higher Al_2O_3, Na_2O, K_2O and P_2O_5 weight percentages and lower MgO, CaO and TiO_2 (Roberts 1980a). These differences were also reflected quite clearly in AFM, FeO *vs.* FeO/MgO and TiO_2 *vs.* FeO/MgO diagrams, with the high-alumina, alkali-rich greenstones displaying calc-alkaline features. Samples collected subsequently served to confirm this chemical disparity (Table 1). Oxide concentrations for the main group of analyses, on the other hand, invite comparison with values for ocean-floor tholeiites (Table 1, group A). Again, this trend appears to be confirmed in graphic representation of the data (Fig. 2).

TABLE 1. Averaged major and trace element contents of the Snåsavatn greenstones (major elements in wt.%, trace elements in ppm). For comparison, mean element concentrations are presented for selected basalt series or types

	Snåsavatn greenstone			*Mean values, diverse basalts*			
	Group A (n=27)	Group B (n=13)	OFB	W.-P. CON	CAB	Smøla CAB	Sunda arc H-K.CAB
SiO_2	45.67	48.01	49.91	48.81	51.31	49.62	49.60
TiO_2	1.66	1.35	1.43	2.47	0.88	0.93	1.15
Al_2O_3	16.01	18.96	16.20	14.41	18.60	18.10	18.20
Fe_2O_3	3.53	4.46	–	13.20[2]	2.91	2.43	–
FeO	6.45	3.39	10.24[1]	–	5.80	5.02	9.80[1]
MgO	7.83	4.72	7.74	5.96	5.95	6.87	5.10
CaO	10.69	6.53	11.42	10.05	10.30	8.64	10.20
Na_2O	2.65	4.92	2.82	2.90	2.93	2.99	3.00
K_2O	0.34	1.58	0.24	0.95	0.74	1.21	2.18
MnO	0.17	0.16			0.15	0.14	0.20
P_2O_5	0.16	0.63			0.12	0.18	0.40
L.O.I.	4.09	4.00				3.69	
Zr	112	230	92	149	106	120	113
Y	31	32	30	25	23	23	28
Sr	246	1143	131	401	375	557	540
Rb	6	26	3	15	23	39	61
Zn	79	128				63	
Cu	46	88	73	99	35	38	110
Ni	102	44	106	68	50	87	19
Cr	305	74	310	139	130	251	39
Ba	68	525	8	338	260	301	480
Nb	5	17	5	25	4	6.5	11
V	323	234	229		174	222	280

Analysts: Gjert Faye and Per-Reidar Graff, NGU.
Data sources for mean values: OFB – Pearce 1975: W.-P.CON – Pearce (1975); CAB – Nockolds & Le Bas (1977), Pearce (1975); Smøla CAB–Roberts (1980b); Sunda arc high-K CAB – Whitford et al. (1979).

1) – Total Fe given as FeO. 2) – Total Fe given as Fe_2O_3

With a total alkali content of 6.5 wt.% for the calc-alkaline basaltic greenstones (CAB), which is more than double that of the ocean-floor type tholeiitic basalts (OFB group), it is tempting to ascribe this relative and absolute increase solely to some form of pre-deformational alteration, notably spilitisation. This is unlikely, however, since the K_2O and Al_2O_3 contents are consistently and sharply higher (quite the opposite of spilitisation trends) in the CAB. Moreover, a Hughes (1973) diagram shows that virtually all the Snåsavatn greenstone samples fall clear of the 'spilite' field (Fig. 3). The depletion of CaO requires some explanation; possibly it may relate to an increase in calcite and epidote veining within this level of the volcanic pile. This needs further investigation.

TRACE ELEMENT CHEMISTRY

Extensive research has shown the value of incompatible trace elements in discriminating between the tectonic settings of diverse magmatic associations. Of

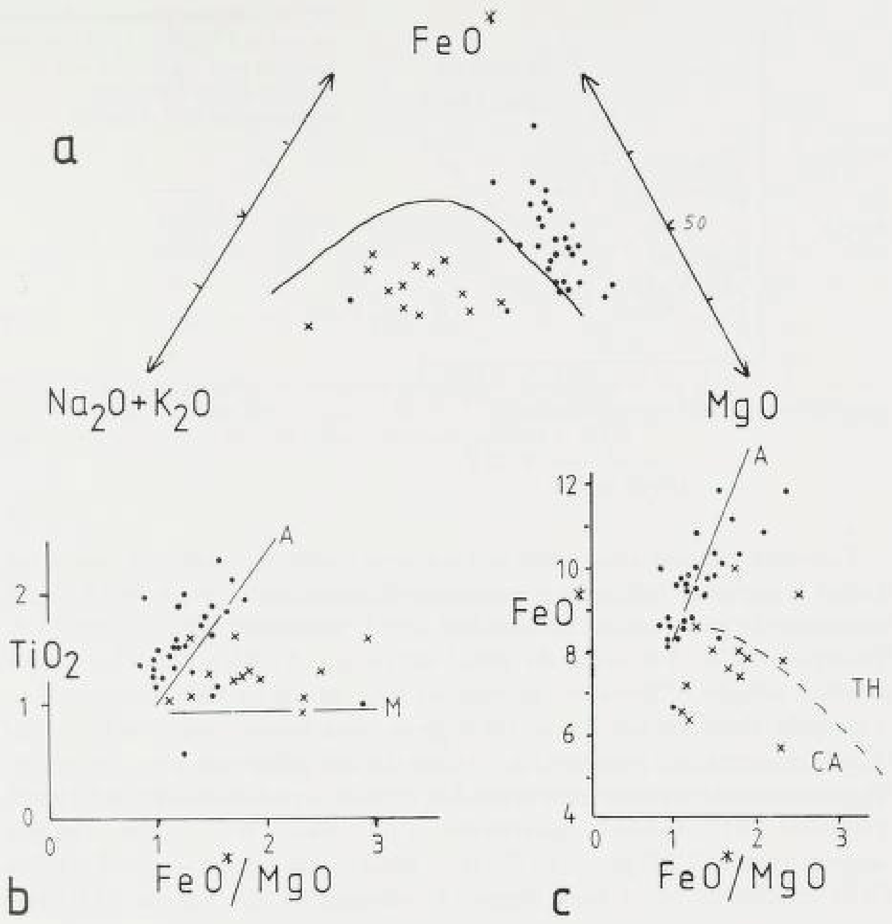


Fig. 2. Plots of the Snåsavatn greenstone analyses on (*a*) an AFM diagram, (*b*) TiO_2 *vs.* FeO/MgO and (*c*) FeO *vs.* FeO/MgO diagram. (FeO^* = total Fe as FeO). Symbols: *dots* – samples of the OFB (group A in Table 1); *crosses* – samples of the CAB (group B in Table 1). In *b* and *c*, A – trend for average ocean floor tholeiites. In *b*, M – trend for Macauley island arc tholeiites. In *c*, the dashed line separates the fields for tholeiites (TH) and calc-alkaline basalts (CA).

these elements, Y, Nb, Zr and the heavy rare-earths (HREE) are the most stable, closely followed by the light REE, and these have proved applicable in distinguishing volcanic settings in both modern and ancient magmatic assemblages (for examples from central Norway, see Grenne & Roberts (1980) and Roberts (1980b)). Here again, plots of particular element ratios provide immediate graphic discrimination, but a visual comparison of mean element concentrations with those from known modern settings is equally valuable (Table 1).

For the Snåsavatn greenstones the dual grouping, OFB and CAB, noted from the major element plots, is even more distinctive from trace element ratio diagrams. In the Ti-Zr-Y plot (Fig. 4), some of the OFB group analyses exhibit an overlap into the field of within-plate basalts, i.e. they show slightly 'transitional' characteristics.

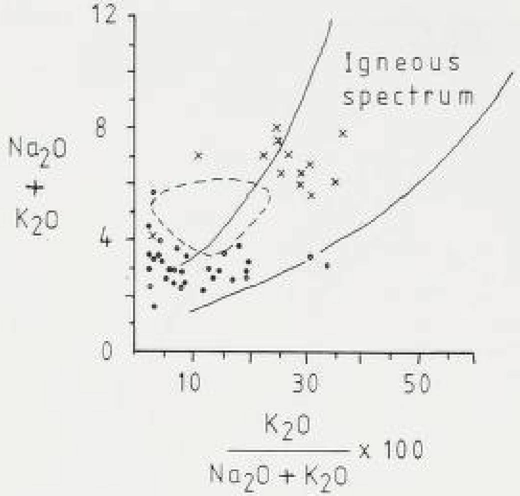


Fig. 3. Snåsavatn greenstone samples plotted in a Hughes' (1973) diagram. Symbols as in Fig. 2. The area enclosed by the dashed line represents the field of spilites.

Rare-earth element abundances and patterns, relative to chondritic values, are known as important indicators of magmatic affiliation, even though the LREE can sometimes be prone to spilitisation and low-T weathering process (Ludden & Thompson 1978). Following the initial sampling and analytical work, 6 representative samples of Snåsavatn greenstones were chosen for REE determination, 4 from the tholeiites and 2 from the high-alumina basalt group (Table 2). The chondrite-normalised patterns (Fig. 5) from the two groups are quite distinctive. The tholeiite analyses show more or less flat, chondritic profiles though with a quite prominent LREE depletion characteristic of mid-oceanic ridge basalts. The two samples from the CAB group, on the other hand, reveal parallel trends of marked LREE-enrichment and a high degree of fractionation. The lightest REE show

TABLE 2. Rare-earth element and Sc, Hf, Ta, Th and U contents (ppm), and selected ratios, from representative samples of Snåsavatn metabasaltic greenstones, Nord-Trøndelag, Norway. The first four samples are classified in group A (Table 1) and the last two in group B.

Sample no.	SN. 5	SN. 8	SN. 16	SN. 36	SN. 23	SN. 24
La	3.36	5.00	5.31	5.67	57.6	70.5
Ce	11.1	14.6	14.6	17.0	124.0	160.0
Nd	10.6	11.6	11.8	13.1	57.8	76.8
Sm	3.59	3.61	3.47	4.03	9.17	12.10
Eu	1.41	1.30	1.29	1.52	2.58	3.44
Tb	0.85	0.78	0.74	0.86	0.87	1.18
Yb	3.01	2.84	2.60	3.06	2.01	2.65
Lu	0.45	0.47	0.41	0.48	0.33	0.41
Sc	46.0	39.3	34.4	38.4	9.46	13.4
Hf	2.42	2.37	2.22	2.70	4.27	5.28
Ta	0.17	0.25	0.16	0.30	0.94	1.05
Th	<0.2	0.46	0.51	0.44	16.3	16.8
U	<0.2	<0.2	0.17	<0.2	3.43	6.85
$(La/Yb)_N$	.67	1.05	1.20	1.11	16.90	15.92
$(La/Sm)_N$	.59	.86	.94	.88	3.84	3.57

Analyst: Dr. Jan Hertogen, Department of Physico-Chemical Geology, Catholic University of Leuven, Belgium.

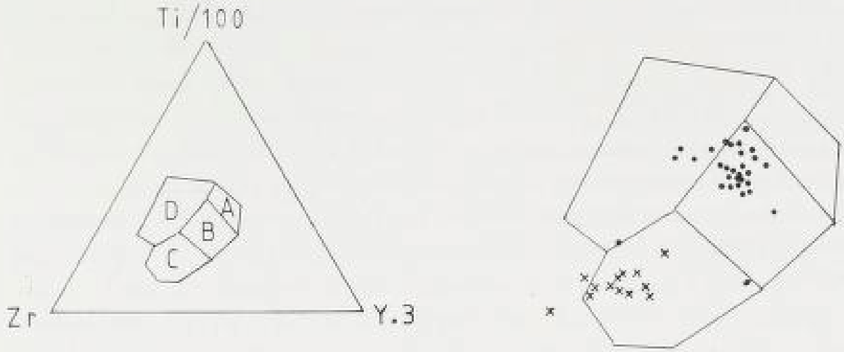


Fig. 4. Ti-Zr-Y diagram for the Snåsavatn greenstones. Symbols as in Fig. 2. Field A – low–K tholeiites; field B – ocean floor tholeiites; field C – calc-alkaline basalts; field D – within-plate basalts. The diagram to the right is an enlargement of fields A to D.

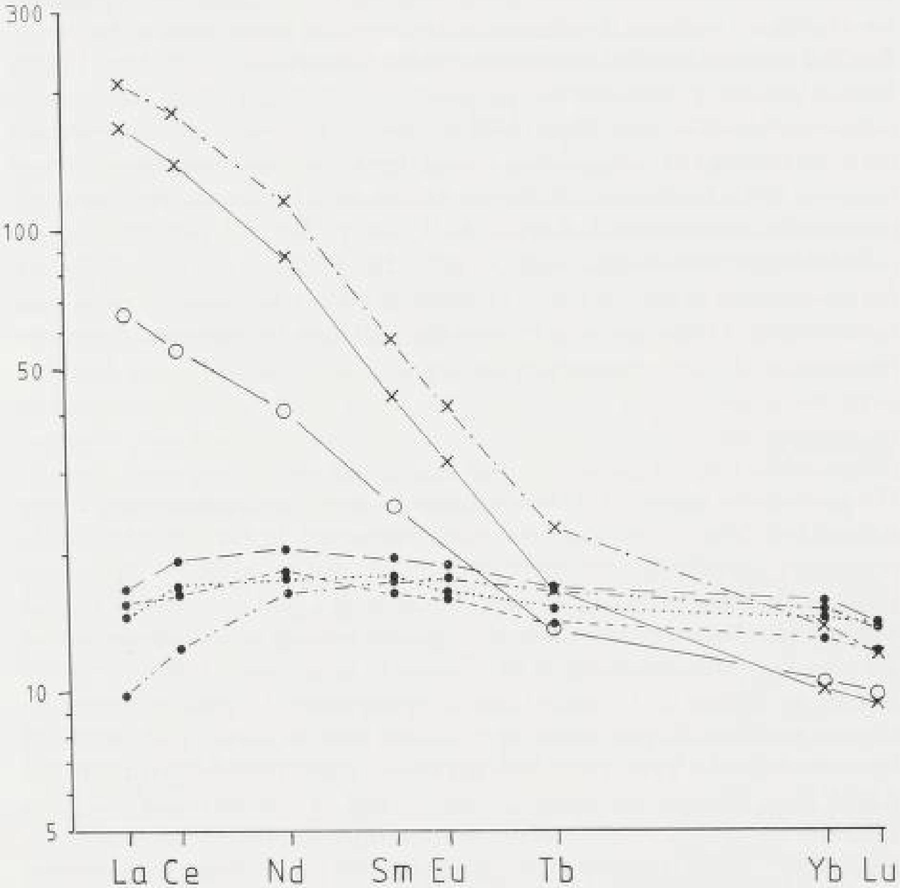
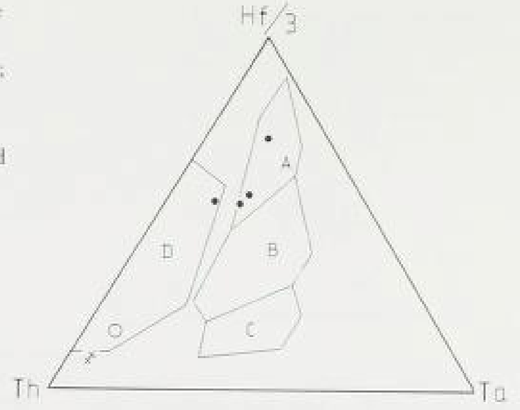


Fig. 5. Chondrite-normalised REE profiles for representative samples of the Snåsavatn lavas. Dot and cross symbols correspond with groups A and B, respectively, as in Fig. 2. Open circles – average values for REE data from the Smøla volcanites. Source of chondritic values: Masuda et al. (1973), Nakamura (1974), Evensen et al. (1978). Details of samples and analyses are contained in an NGU report (Roberts 1981).

Fig. 6. Hf–Th–Ta diagram showing the plots of the REE-analysed samples. Symbols – dots and crosses as in Fig. 2; open circle – average value for basalts from Smøla (Roberts 1981). Field A – normal MORB. Field B – LREE-enriched MORB. Field C – within-plate basalts. Field D – basalts formed at convergent plate margins (both island arc and Andean type).



c. 170–200 times chondritic abundances, which is a somewhat greater figure than the average for modern calc-alkaline basalts, and also higher than in the case of the CAB from the island of Smøla, west-central Norway (Fig. 5) (Roberts 1980b, 1981); but the general profiles are otherwise similar to those from typical calc-alkaline basaltic lavas. High LREE abundances such as these are known from some of the high-K calc-alkaline basalts from the Quaternay Sunda arc of Indonesia (Whitford et al. 1979). In relation to SiO_2 content the Snåsavatn metabasalts, with nearly 1.6 wt.% K_2O, do in fact fall into the high–K calc-alkaline series of volcanic rocks as defined by Peccerillo & Taylor (1976). In Fig. 6, separation of the two groups of Snåsavatn lavas is again clearly visible, with a good degree of correspondance between the CAB from Snåsavatn and those from Smøla.

Discussion

The geochemical signatures of these greenstones show that lavas generated in two separate and distinctive volcanic settings are represented. In this particular area the structurally lowest volcanites are of calc-alkaline affinity, while the bulk of the lavas are tholeiites of ocean-floor type though with a hint of transition towards within-plate character. Aspects of the regional geology are of importance in assessing the palaeoenvironments of effusion of these lavas. Although way-up features are lacking in the studied area, there appears to be good evidence from adjacent areas that the greenstones are positioned stratigraphically above the Snåsa limestone (Carstens 1956, 1960, Springer Peacey 1964, Tietzsch–Tyler, in prep.). Rapid facies changes are present, however, and in one area near southwest Snåsavatn the limestone represents almost the entire Lower Hovin Group (Springer Peacey 1964, p. 19), greenstones interdigitating with the limestone and increasing rapidly in volume northeastwards. It would thus appear that quite evolved calc-alkaline basalts, associated with shallow-marine limestone deposition, were replaced up-sequence by tholeiites of ocean-floor character.

The Snåsa limestone fauna is of importance in that the fossils (Roberts 1980c) show striking similarities to those occurring in the late Arenig-Llanvirn Skjølberg Limestone on the island of Smøla (Bruton & Bockelie 1979) some 200 km further southwest along strike. Since the Smøla volcanites (closely associated with the limestone) are of mature, calc-alkaline type (Roberts 1980b), there are obviously parallels to be drawn with the Snåsavatn situation, at least for the lower CAB lava unit. An allied feature is that mineralisation in the volcanites from Smøla, and northeastwards through an important iron-ore district (Fosdalen, near Malm) to Snåsavatn, is of a similar magnetite-pyrite-chalcopyrite association (Carstens 1956, 1960). There are, however, small differences in certain minor and trace element abundances (e.g. Zr, Nb and LREE) and this may reflect a positioning of the Snåsavatn magmatic arc closer to the palaeo-continental margin than in the case of the Smøla arc rocks.

The OFB volcanites at Snåsavatn, coming in without a visible break above the island arc CAB/limestone association, were clearly not extruded in a major ocean, spreading-ridge setting. On the other hand, the indications of 'transitional' chemistry, towards WPB, are typical of an extensional, marginal basin, back-arc milieu of spreading. Such an interpretation fits very well with what we know of other Middle Ordovician magmatic assemblages in Trøndelag (Grenne & Roberts 1980, Roberts, Grenne & Ryan, in prep.). The actual location of the marginal basin spreading centre in this particular case (prior to nappe translation) could be considered as being related to oceanward migration of the arc and subduction zone with time (Roberts et al. in press), a feature well known from modern island arc/subduction zone situations, for example in the western Pacific, thus facilitating the behind-arc marginal basin distension and ocean accretion. Alternatively, the OFB lavas may have been associated with a splitting of the fairly mature, carbonate-fringed magmatic arc during a changing stress regime – from essentially convergent and compressive to extensional, perhaps itself related to changing vectors of plate motion in which the subduction zone gradually became inactive. Assessing these possibilities will require further detailed mapping and sampling over a much wider area.

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REFERENCES

Andreasson, P.G. & Johansson, L. 1982: The Snåsa mega-lens, west-central Scandinavian Caledonides. *Geol. Fören. Förhandl. 104.*

Bruton, D.L. & Bockelie, J.F. 1979: The Ordovician sedimentary sequence on Smøla, west Central Norway. *Norges geol. Unders. 348*, 21–31.

Carstens, H. 1956: Geologi. Kort geologisk oversikt over jernmalm-distriktet fra Snåsa til Stjørna. *In* 'Fosdalens Bergverk, 1906–1956. Fosdalens Bergverks-Aktieselskab, 149–158.

Carstens, H. 1960: Stratigraphy and volcanism of the Trondheimsfjord area, Norway. Excursion guide A4, XXI Int. Geol. Congr., *Norges geol. Unders. 213b*, 23pp.

Evensen, N.M., Hamilton, P.J. & O'Nions, R.K. 1978: Rare-earth abundances in chondritic meteorites. *Geochim. Cosmochim. Acta 42*, 1199–1212.

Foslie, S. 1958: 1:100 000 map-sheet 'Sanddøla' *Norges geol. Unders.*

Gale, G. H. & Roberts, D. 1974: Trace element geochemistry of Norwegian Lower Palaeozoic basic volcanics and its tectonic implications. *Earth Planet. Sci.Lett. 22*, 380–390.

Gee, D.G. 1977: Extension of the Offerdal and Särv Nappes and Seve Supergroup into northern Trøndelag. *Norsk geol. Tidsskr. 57*, 163–170.

Grenne, T. & Roberts, D. 1980: Geochemistry and volcanic setting of the Ordovician Forbordfjell and Jonsvatn greenstones, Trondheim Region, Central Norwegian Caledonides. *Contrib. Mineral. Petrol. 74*, 375–386.

Hughes, C.J. 1973: Spilites, keratophyres and the igneous spectrum. *Geol. Mag. 109*, 513–527.

Langmyhr, F.J. & Graff, P.R. 1965: A contribution to the analytical chemistry of silicate rocks: a scheme of analysis for eleven main constituents based on decomposition by hydrofluoric acid. *Norges geol. Unders.* 230, 128 pp.

Ludden, J.N. & Thompson, G. 1978: Behaviour of rare earth elements during submarine weathering of tholeiitic basalt. *Nature, 274*, 147–149.

Masuda, A., Nakamura, N. & Tanaka, T. 1973: Fine structures of mutually normalized rare-earth patterns of chondrites. *Geochim. Cosmochim. Acta, 37*, 239–248.

Nakamura, N. 1974: Determination of REE, Ba, Fe, Mg, Na and K in carbonaceous and ordinary chondrite. *Geochim. Cosmochim. Acta, 38*, 757–775.

Nockolds, S.R. & Le Bas, M.J. 1977: Average calc-alkali basalt. *Geol. Mag. 114*, 311–312.

Pearce, J.A. 1975: Basalt geochemistry used to investigate past tectonic environments on Cyprus. *Tectonophysics, 25*, 41–68.

Pearce, J.A. & Cann, J.R. 1973: Tectonic setting of basic volcanic rocks determined using trace element analyses. *Earth Planet. Sci. Lett. 19*, 290–300.

Peccerillo, A. & Taylor, S.R. 1976: Geochemistry of Eocene calc-alkaline volcanic rocks from the Kastamonu area, northern Turkey. *Contrib. Mineral. Petrol. 58*, 63–81.

Roberts, D. 1967: Geological investigations in the Snåsa–Lurudal area, Nord-Trøndelag. *Norges geol. Unders. 247*, 18–38.

Roberts, D. 1980a: Geokjemien av Snåsagrønnsteinene: en framgangsrapport. Norges geol. Unders. report.

Roberts, D. 1980b: Petrochemistry and palaeogeographic setting of the Ordovician volcanic rocks of Smøla, Central Norway. *Norges geol. Unders. 359*, 43–60.

Roberts, D. 1980c: En ny fossil lokalitet i Snåsakalken, Snåsavatn, Nord-Trøndelag. Norges geol. Unders. report.

Roberts, D. 1981: Sjeldne jordartsanalyser av ordoviciske volkanitter fra Smøla- og Snåsaområdene. Norges geol. Unders. report.

Roberts, D. & Wolff, F.C. 1981: Tectonostratigraphic development of the Trondheim region Caledonides, Central Norway. *Journ. Struc. Geol. 3*, 487–494.

Roberts, D., Sturt, B. A. & Furnes, H. 1982: Volcanite assemblages and environments in the Scandinavian Caledonides and the sequential development history of the mountain belt. Proc. IGCP Symp. Uppsala, 1981 (in press).

Springer Peacey, J. 1964: Reconnaissance of the Tømmerås Anticline. *Norges geol. Unders. 227*, 13–84.

Vogt, T. 1945: The geology of part of the Hølonda–Horg district, a type area in the Trondheim region. *Norsk geol. Tidsskr. 25*, 449–527.

Whitford, D.J., Nicholls, I.A. & Taylor, S.R. 1979: Spatial variations in the geochemistry of Quaternary lavas across the Sunda Arc in Java and Bali. *Contrib. Mineral. Petrol. 70*, 341–356.

Wolff, F.C. 1976: 1:250 000 map-sheet 'Trondheim'. *Norges geol. Unders.*

Wood, D.A., Joron, J.L. & Treuil, M. 1979: A re-appraisal of the use of trace elements to classify and discriminate between magma series erupted in different tectonic settings. *Earth Planet. Sci.Lett. 45*, 326–336.

Appendix

ANALYTICAL PROCEDURES

Major element analyses for the first batch (36) of samples were determined by classical wet-chemical methods for SiO_2, Al_2O_3, CaO, MgO and total Fe, and by a method outlined by Langmyhr & Graff (1965) for TiO_2, MnO, Na_2O, K_2O and ferrous iron. Majors for the 10 samples collected in 1981, as well as common trace elements for all 46 samples were analysed on rock powders using an automatic

Philips 1450/20 XRF. All analyses were carried out at the Section for Analytical Chemistry, NGU, Trondheim. Calibration curves were made with international standards.

Rare-earth elements, together with Hf, Ta, Th and U, were analysed by instrumental neutron activation, by Dr. Jan Hertogen at the Department of Physico-Chemical Geology, Catholic University of Leuven, Belgium. Samples were irradiated at the Ghent University nuclear reactor. Standard deviations were calculated from counting statistics and the observed spread among results from different countings and/or different gamma-energies. In cases where counting statistics were smaller than 1%, a realistic standard deviation of 1% was assumed. REE normalisation average values used are as follows: – La (0.34), Ce (0.89), Nd (0.65), Sm (0.209), Eu (0.0806), Tb (0.052), Yb (0.20), Lu (0.035); normalising values from Masuda et al. (1973), Nakamura (1974) and Evensen et al. (1978).

A Proposed Deglaciation Chronology for the Trondheimsfjord Area, Central Norway

ARNE J. REITE, HAAVARD SELNES & HARALD SVEIAN

Reite, A.J., Selnes, H. & Sveian, H. 1982: A proposed deglaciation chronology for the Trondheimsfjord area, Central Norway. *Norges geol. Unders. 373*, 75–84.

Radiocarbon datings of the deglaciation of the Trondheimsfjord area strongly suggest that the coastal areas and the outer part of the fjord were deglaciated before 11000 B.P. During early Younger Dryas the inland ice advanced to the Tautra Moraines (10800–10500 B.P.). Further inland the Hoklingen Moraines (10300–10100 B.P.) and the Vuku Moraines (c. 9800 B.P.) were deposited during marked glacial advances. Other ice-marginal deposits are mostly dependent on local topography. Brief comments are given on the shoreline displacement during the deglaciation.

Arne J. Reite and Harald Sveian, Norges geologiske undersøkelse, P.O.Box 3006, N–7001 Trondheim, Norway
Haavard Selnes, Institutt for kontinentalsokkelundersøkelser, P.O.Box 1883, N–7001 Trondheim, Norway

Introduction

Moraines on the continental shelf off central Norway have been radiocarbon-dated to 13000 B.P. for the Storegga moraines and 12400 B.P. for the Haltenbank moraines (Bugge 1980). Andersen (1979), however, has assumed an approximate age of 15000 B.P. for the inner marginal moraines on the continental shelf outside Trøndelag. Radiocarbon datings from Hemnefjord (Lasca 1969), Bjugn (A. Kjemperud pers.comm. 1982) and Rissa (Bugge 1980, Løfaldli et al. 1981) strongly suggest that the outer part of Trondheimsfjord was deglaciated not later than Allerød. Previous datings from the Ekle–Tiller ice-marginal deposit, situated 10 km to the south of Trondheim, have indicated an early Younger Dryas age (Nydal et al. 1972). This deposit (Figs. 1 & 3) has been correlated with the Malvik moraine, the Tautra moraine and the prominant marginal moraines crossing the Fosen peninsula (Reite, in Oftedahl 1974, Sollid & Sørbel 1975, 1979, Sollid 1976). An early Younger Dryas age for these moraines is supported by radiocarbon datings of limnic sediments from the northeastern part of the Frosta peninsula (Kjemperud 1981). A Younger Dryas age was first proposed by Holtedahl (1928) and Undås (1942). Marginal moraines in the inner part of Trondheimsfjord have been tentatively dated to the Preboral Chronozone (Andersen 1979, Sollid 1979).

In recent years, mapping of Quaternary sediments by the Geological Survey of Norway has provided additional information on the stratigraphy and deglaciation (Reite 1975, 1976, 1977, 1980, 1982, in press., Reite & Sørensen 1980, Sveian 1981a, 1981b, 1981c, in press., Hugdahl 1980 and K. Bjerkli pers. comm. 1982). Shoreline displacements and vegetation history have been studied by the Institute of Botany, University of Trondheim (Kjemperud 1981, Selnes 1982).

The aim of this paper is to present the results of radiocarbon datings and propose a chronostratigraphy for the deglaciation of the Trondheimsfjord area, with special emphasis on the distinct marginal moraines which are considered to have been

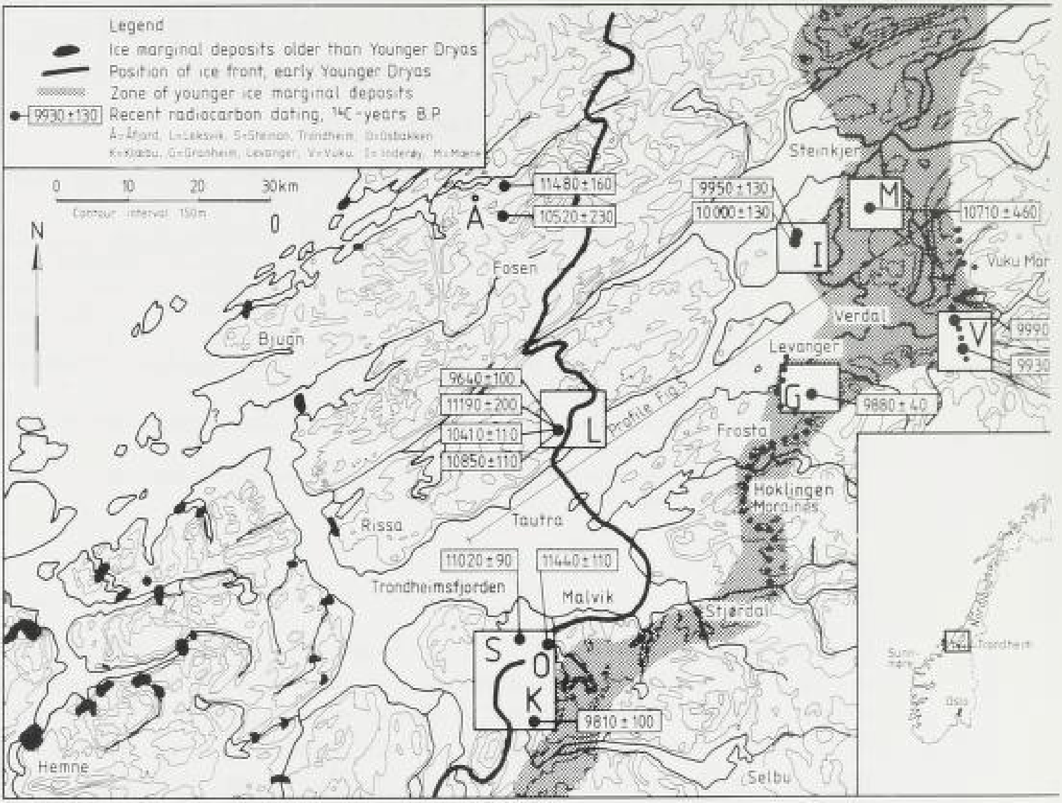

Fig. 1. Map of the Trondheimsfjord area showing ice-marginal features and recent radiocarbon datings. Frames mark the location of the more detailed maps shown in Figs. 2, 3 and 4.

deposited during the Younger Dryas and Preboreal Chronozones. The stratigraphic terminology adopted follows the proposals of Mangerud et al. (1974).

The Fosen Peninsula

Stratigraphical studies of limnic and marine sediments at Åfjord and limnic sediments at Leksvik (Figs 1 & 2) were carried out to date the distinct marginal moraines which cross the Fosen peninsula. The sampling sites at Åfjord are situated 15 km distal to these moraines, while those at Leksvik are in the marginal zone.

Åfjord. In a small basin 25 metres below the marine limit a glacial advance is represented by glaciomarine silt and clay. This bed is underlain by shell-bearing marine clay and overlain by limnic organic sediments. The shells (mostly *Macoma calcarea)* were dated to 11480 ± 160 B.P. (T–3655), and the transition marine/limnic sediments to 10040 ± 100 B.P. (T–3656A). In a small basin nearby, situated above the marine limit, a bottom gyttja sample was dated to 10520 ± 230 B.P. (T–3660A).

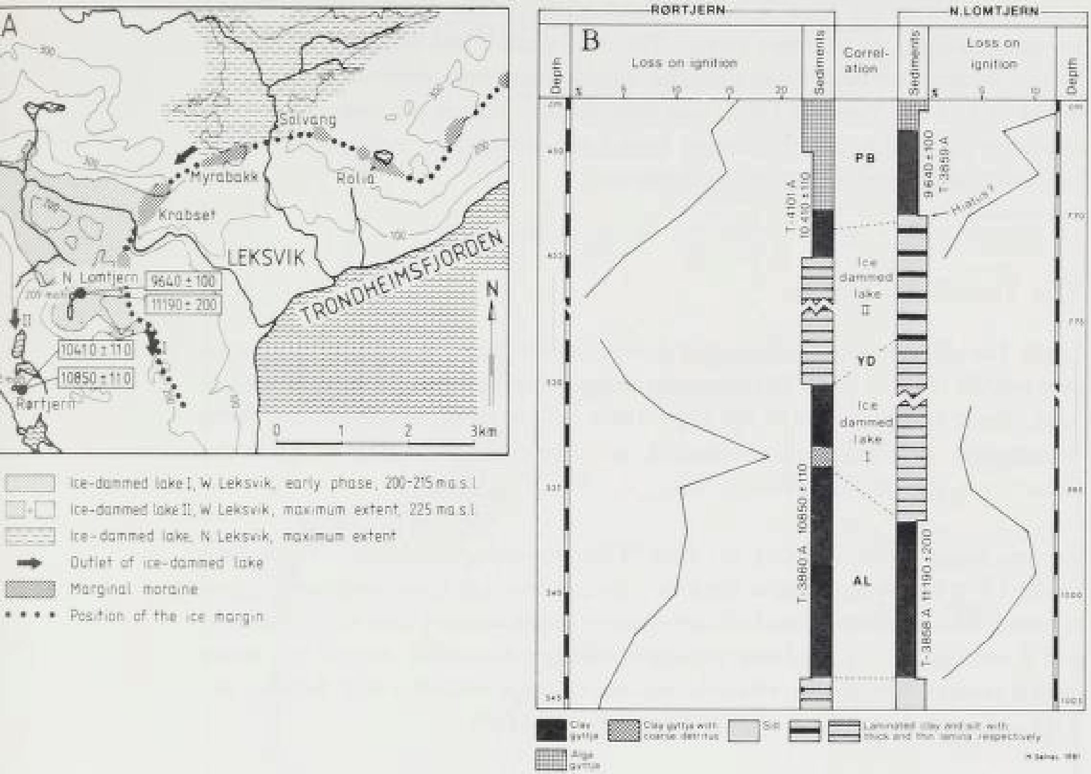

Fig. 2. A: Map of the Leksvik area showing ice-marginal deposits, glacier-dammed lakes and radiocarbon datings.
B: Stratigraphy and radiocarbon datings from the two basins Rørtjern and N. Lomtjern.

Leksvik. At Leksvik the stratigraphy of two small basins, N. Lomtjern and Rørtjern has been studied; both basins are situated above the marine limit. With the exception of a possible hiatus in N. Lomtjern, the basins represent continuous limnic sedimentation from Allerød up to present (Fig. 2). This conclusion is based on the nature of the sediments, the location of marginal moraines, pollen stratigraphy and radiocarbon datings.

Palynological investigations of the organic Allerød sediments suggest a cold steppe/arctic flora, indicating proximity to the inland ice (Selnes 1982). The laminated clay and silt in Lomtjern is interpreted as varves deposited in a glacier-dammed lake which existed during the first half of Younger Dryas (lake I). Rørtjern, on the other hand, was not influenced by lake I. The slightly coarser beds in Rørtjern are considered to be sediments deposited by the river from Lake II, which came into existence when the inland ice experienced a small advance, causing a temporary shift of the threshold. At that time the ice front was situated very close to Lomtjern, where thick laminae of coarser material were deposited.

The datings from *Åfjord* and *Leksvik* and the studies of marginal moraines indicate that the prominent marginal moraines crossing the Fosen peninsula are synchronous with the Leksvik ice-marginal deposits. The advance of the inland ice occured early in Younger Dryas. The stratigraphy and radiocarbon datings further suggest that the average shoreline displacement from the middle of Allerød to the end of the Younger Dryas was only about 2 metres per century. As only one basin situated below the marine limit has been studied, no conclusions can be drawn on the possibility of transgressions.

The Trondheim region

Large, but discontinuous, ice-marginal deposits are found at Ekle, Tiller, Reppe and Malvik (Figs. 1 & 3). Shell-bearing marine clays situated close to the marine limit, distal and proximal to the Ekle–Tiller ice-marginal deposit and overlying this deposit, were radiocarbon-dated. A spinal segment of a whale found in the foreset beds was also dated (Fig. 3B).

Steinan. Some 2 km distal to the Ekle–Tiller ice-marginal deposit, shells were found 15 m below the marine limit in glaciomarine clay containing ice-dropped material. The 50 cm-thick bed of shell-bearing clay is situated close to the bedrock, and is overlain by 1.5 m of clay without ice-dropped material. No till was found in this section. The shells *(Hiatella arctica, Chlamys islandica* and *Balanus sp.)* were radiocarbon-dated to 11020 ± 90 B.P. (T–3296).

Tiller. A spinal segment of a big whale was found at 15 metres depth in the foreset beds of the glaciofluvial ice-marginal deposit at Tiller. Several years later it was radiocarbon-dated to 10990 ± 190 B.P. (T–787).

Ekle. The foreset beds of the glaciofluvial ice-marginal deposit at Ekle are overlain by glaciomarine clay containing a few shells of *Portland arctica*, suggesting arctic water. This bed is overlain by sandy clay with *Mya truncata, Astarte elliptica* and *Balanus sp.* The fauna thus suggest that the inland ice had receded from Ekle–Tiller before this clay was deposited. This is also supported by the grain-size distribution, indicating shallow water conditions. Radiocarbon datings have given results of 10150 ± 100 B.P. (T–854) and 10230 ± 130 B.P. (T–786) for *Mya truncata* and *Balanus*, respectively. The Ekle ice-marginal deposit is underlain by at least 15 metres of marine clay, probably of late Allerød age.

Osbakken. At Osbakken, a few hundred metres proximal to the supposed position of the Ekle–Tiller–Reppe–Malvik ice front, a shell-bearing marine clay was found close to the marine limit. The clay is not overlain by till. The shells (dominated by *Macoma calcarea* and *Balanus sp.)* were radiocarbon-dated to 11440 ± 110 B.P. (T–4242).

Klæbu. Shells were found close to Sørborgan farm, Klæbu, when marine clay was removed from an almost vertical cliff. This locality is situated 15–20 metres below

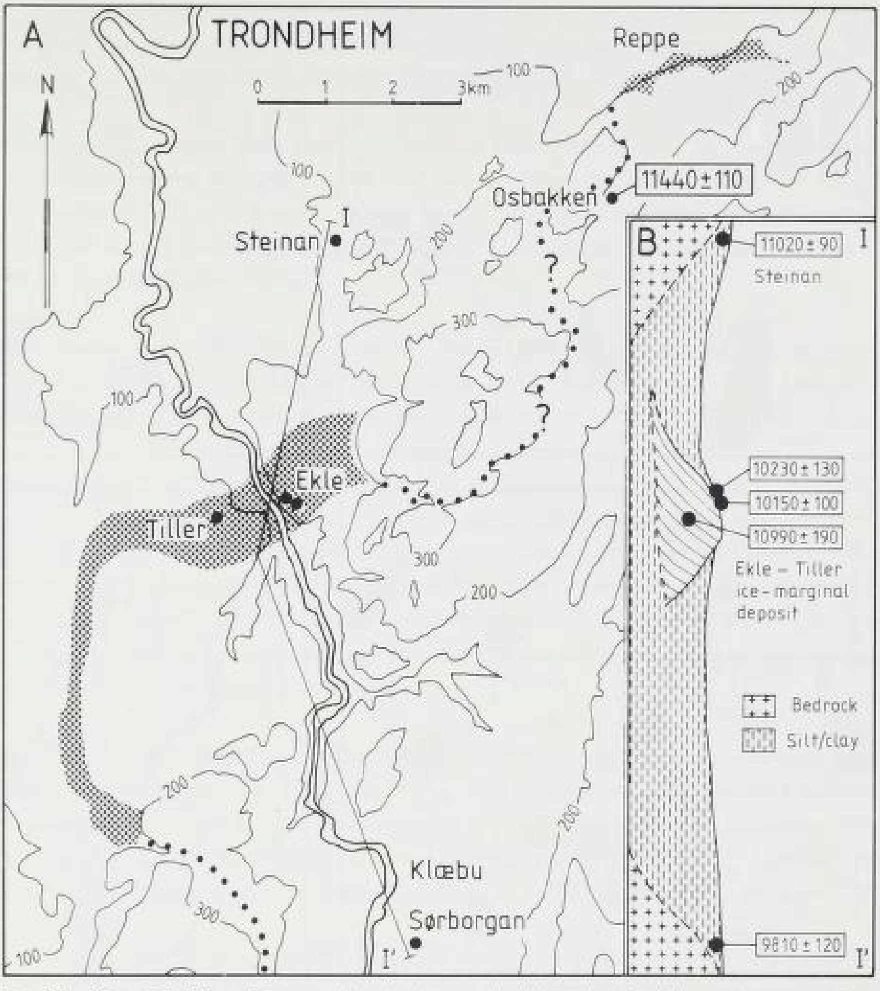

Fig. 3: A: Map of the Trondheim area showing ice-marginal deposits (shaded areas), reconstruction of the ice margin (dotted) and the radiocarbon dating at Osbakken.
B: Profile Steinan–Ekle–Sørborgan with radiocarbon datings.

the marine limit. The shells were radiocarbon-dated to 9810 ± 120 B.P. (T–3113).

The datings from the *Trondheim region* suggest that the Ekle–Tiller ice-marginal deposit was formed in the first part of Younger Dryas. The dating from Osbakken indicates, however, that the marginal moraines at Reppe and Malvik might be slightly older. More radiocarbon datings are needed to solve this question. The datings from several localities situated close to the marine limit indicate that the average shoreline displacement has been less than 2 metres per century during the later part of Allerød and Younger Dryas.

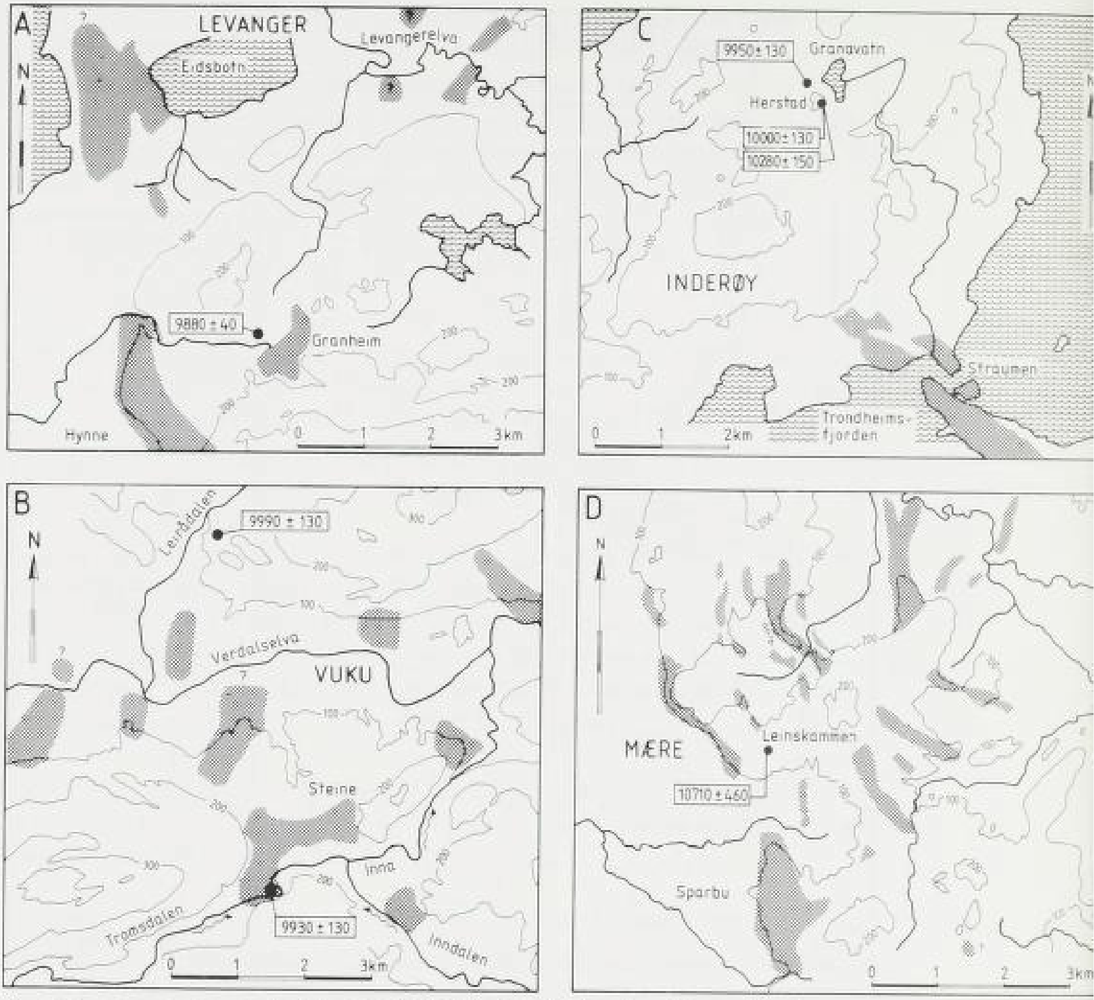

Fig. 4: Maps showing ice-marginal deposits (shaded areas) and radiocarbon datings. A: the Levanger area, B: the Vuku area, C: the Inderøya area and D: the Mære area.

The Verdal region

Numerous ice-marginal deposits occur in this region. A distinct marginal moraine can be traced from Eidsbotn at Levanger to Hynne, lake Hoklingen and further towards Stjørdal (Figs. 1 &4A), and possibly to the western part of the lake Selbusjøen. Evidence of another glacial advance is found at Vuku. This advance can also be traced in a few localities to the north of Vuku and possibly at Steinkjer. Most of the other marginal deposits in this region were formed when the ice front was temporarily halted at bedrock thresholds and other topographic obstacles. They consist mostly of glaciofluvial material. A detailed reconstruction of the ice front is hardly possible for these deposits.

Marine shells situated in the ice-marginal deposits and distally and proximally to them were radiocarbon-dated, as was a peat sample overlying basal till. The datings all give minimum ages for the deglaciation.

Granheim, Levanger. At a locality a few hundred metres distal to the Granheim ice-marginal delta and 2 km proximal to the Hynne moraine, shells were found in marine clay 25 m below the marine limit (Fig. 4A). The shells *(Mya truncata)*, collected in a 3 m-thick clay bed probably close to the bedrock, were dated to 9880 ± 40 B.P. (T–3997).

Leirådalen. Shells *(Mytilus edulis*, with a few *Hiatella arctica* and *Balanus sp.)* were found in a clay 40 m below the marine limit at Leirådalen (Fig. 4B). The shell-bearing clay was radiocarbon-dated to 9990 ± 130 B.P. (T–3999).

Tromsdalen. The Steine ice-marginal delta represents the marine limit, 180–185 m a.s.l. The southern part of the deposit crosses the mouth of Tromsdalen (Fig. 4B). Situated upon the delta is a 20 m-high mound of sandy and gravelly till, containing large blocks of shell-bearing marine clay deposited by a glacial advance. The shells (mostly *Mytilus edulis)* were radiocarbon-dated to 9930 ± 130 B.P. (T–3998); this is a maximum age for this glacial advance and a minimum age for the first ice recession at Vuku.

Herstad. A peat sample was collected in a 2 m-thick bog at the farm Herstad øvre, 40 m above the marine limit (Fig. 4C). The pollen content point to a pioneer vegetation, at least in the lowermost 0–4 cm of the organic sediment. The lowermost part of the sediment was radiocarbon-dated to 10000 ± 130 B.P. for the soluble fraction; and the insoluble fraction to 10280 ± 150 B.P. (T–4259).

Granavatn. Shells were collected from a sandy silt and clay 147 m a.s.l., about 200 m west of lake Granavatn (Fig. 4C). The shells *(Mya truncata)* were found 0.5–1.5 m below the surface in a bed more than 2 m thick. The sediments indicate a sea level higher than 155 m, which is not more than 10–15 m below the marine limit. The shells were radiocarbon-dated to 9950 ± 130 B.P. (T–4257).

Leinskammen. Leinskammen is located between two of the prominent ice-marginal deposits at Mære (Fig. 4D). Only 1.5 g of shells *(Balanus sp.* and probably *Hiatella arctica)* were collected at 168 m a.s.l. in a laminated silt and clay, which is overlain by 1.5 m of coarse-grained beach gravel. The shells date a sea level slightly higher than 170 m, very close to the marine limit. The radiocarbon age is 10710 ± 460 B.P. (T–4258). Unfortunately the sample was too small to give a more precise date.

The datings from the *Verdal region* indicate that Levanger and Inderøy were deglaciated before 10000 B.P. The datings from Mære, Leirådalen and Tromsdalen strongly suggest that Levanger and Inderøy were deglaciated in Younger Dryas, and also that the main valley at Verdal was deglaciated at that time. This

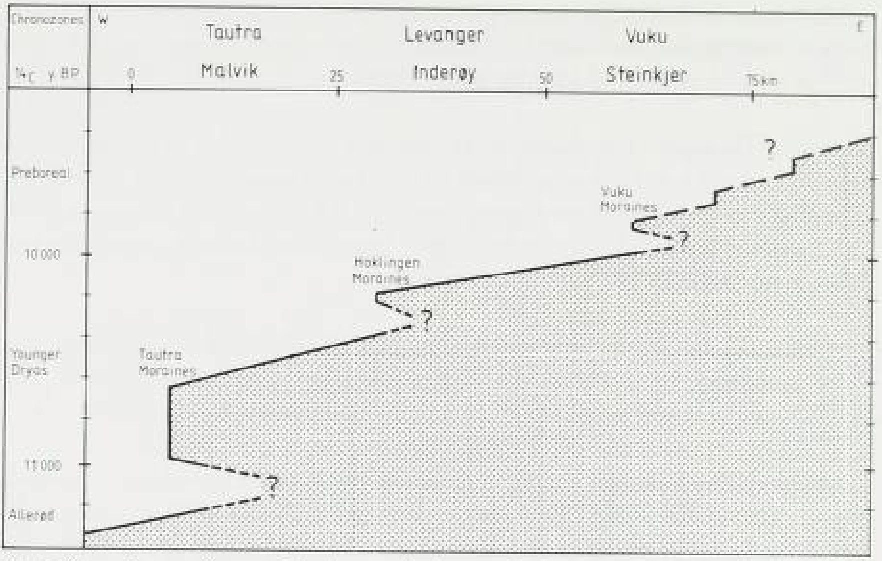

Fig. 5: Time-distance diagram of the ice recession in central and eastern parts of the Trondheimsfjord area.

is in agreement with the datings of gyttja from Frosta, the oldest of which gave Younger Dryas ages (Kjemperud 1981). The shells found in till at Tromsdalen must have been transported by a glacial advance to the Vuku area early in Preboreal.

Conclusions

From the existing data the following preliminary conclusions can be drawn:

- The coastal areas and the outer part of the fjord were deglaciated not later than Allerød, when the front of the inland ice receded to a position to the east of Trondheim, and also at least 10 km to the east of Tautra. The innermost Allerød ice front position is still unknown. (Figs. 1 & 5).
- At the beginning of Younger Dryas a glacial advance took place to the Fosen peninsula, Leksvik, Tautra, Malvik and Trondheim. These ice-marginal deposits are here named the *Tautra Moraines*. The ice recession which followed this advance probably took place in the middle of Younger Dryas.
- The glacial advance to the distinct marginal moraines at Levanger, Hynne, Hoklingen, towards Stjørdal and possibly to the western part of Lake Selbusjøen occurred during the second half of Younger Dryas. We propose to name these ice-marginal deposits the *Hoklingen Moraines*.
- The Younger Dryas chronozone in the Trondheimsfjord area includes the Tautra Moraines and the Hoklingen Moraines, and probably also some ice-marginal

deposits to the east of the Hoklingen Moraines. The deglaciation pattern resembles that found in the Oslo region (Sørensen 1979) and in Nordland (Andersen et al. 1981). This is not in agreement with most previous investigations which have suggested that only the Tautra advance took place during Younger Dryas.

- The glacial advance at Vuku, which can be traced to the north of Vuku and possibly to Steinkjer, took place early in Preboreal. We propose to name these deposits the *Vuku Moraines*. To the south of Vuku the ice margin should be situated to the east of the shaded zone in Fig. 1, where scattered marginal moraines occur.
- With the exception of the Hoklingen Moraines and the Vuku Moraines most ice-marginal deposits in the Verdal region are controlled mainly by local topography. A detailed reconstruction of the ice recession is therefore difficult.
- The average shoreline displacement during the second half of Allerød and throughout Younger Dryas was small compared to that found by Kjemperud (1981) for the Preboreal. A fairly stable sea level has also been found in Nordland (Rasmussen 1981) and in Sunnmøre (Lømo & Lie 1981) for Allerød and the first half of Younger Dryas.

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REFERENCES

Andersen, B.G. 1979: The deglaciation of Norway, 15000–10000 B.P. *Boreas 8*, 79–87.

Andersen, B.G., Bøen, F., Nydal, R., Rasmussen, A. & Vallevik, P.N. 1981: Radiocarbon dates of marginal moraines in Nordland, North Norway. *Geogr. Ann. 63A (3–4)*, 155–160.

Bugge, T. 1980: Øvre lags geologi på kontinentalsokkelen utenfor Møre og Trøndelag. (Shallow geology on the continental shelf off Møre and Trøndelag, Norway). *Continental Shelf Institute, publ. no. 104*, 44 pp.

Holtedahl, O. 1929: Om landisens bortsmelting fra strøkene ved Trondheimsfjorden. *Norsk geogr. Tidsskr. 2*, 95–118.

Hugdahl, H. 1980: Helgådalen, kvartærgeologisk forsøkskart CWX 135136–20, M 1:20 000. *Norges geol. Unders.*

Kjemperud, A. 1981: A shoreline displacement investigation from Frosta in Trondheimsfjorden, Nord-Trøndelag, Norway. *Norsk geol. Tidsskr. 61*, 1–15.

Lasca, N.P. 1969: Moraines in the Hemnefjorden area, Western Norway. *Norges geol. Unders. 266*, 62–69.

Løfaldli, M., Løken, T., Rise, L. and Rokoengen, K. 1981: Kvartærgeologiske undersøkelser av kvikkleireskredet i Rissa ved Trondheimsfjorden. (Quaternary geological investigations of the quickclay slide in Rissa at Trondheimsfjorden). *Continental Shelf Institute, publ.no. 107*, 31 pp.

Lømo, L. and Lie, S.E. 1981: En lito- og biostratigrafisk undersøkelse av marine og limniske sedimenter i Ålesund-området. *Unpubl. Thesis, University of Bergen.*

Nydal, R., Gulliksen, S. and Løvseth, K. 1972: Trondheim natural radiocarbon measurements, VI. *Radiocarbon 14, no. 2*, 418–451.

Oftedahl, Chr. 1974: Norges geologi. *Tapir, Trondheim*, 169 pp.

Rasmussen, A. 1981: The deglaciation of the coastal area NW of Svartisen, Northern Norway. *Norges geol. Unders. 369*, 1–31.

Reite, A.J. 1975: Hølonda, kvartærgeologisk kart 1521 II – M 1:50 000. *Norges geol. Unders.*

Reite, A.J. 1976: Trondheim, kvartærgeologisk kart 1621 IV – M 1:50 000. *Norges geol. Unders.*

Reite, A.J. 1977: Orkanger, kvartærgeologisk kart 1521 I – M 1:50 000. *Norges geol. Unders.*

Reite, A.J. 1980: Snillfjord, kvartærgeologisk kart 1521 IV – M 1:50 000. *Norges geol. Unders.*

Reite, A.J. 1982: Stjørdal, kvartærgeologisk kart 1621 I – M 1:50 000. *Norges geol. Unders.* (in press).

Reite, A.J.: Frosta, kvartærgeologisk kart 1622 II – M 1:50 000. *Norges geol. Unders.* (in press).

Reite, A.J. & Sørensen, E. 1980: Støren, kvartærgeologisk kart 1621 III – M 1:50 000. *Norges geol. Unders.*

Selnes, H. 1982: Paleoøkologiske undersøkelser omkring israndavsetninger på Fosenhalvøya, Midt-Norge. *Unpubl. thesis, University of Trondheim.*

Sollid, J.L. 1976: Kvartærgeologisk kart over Nord-Trøndelag og Fosen. En foreløpig melding. (Quaternary geology of Nord-Trøndelag and Fosen, Central Norway). *Norsk geogr. Tidsskr. 30*, 25–26.

Sollid, J.L. & Sørbel, L. 1975: Younger Dryas ice-marginal deposits in Trøndelag, Central Norway. *Norsk geogr. Tidsskr. 29*, 1–9.

Sollid, J.L. & Sørbel, L. 1979: Deglaciation of western Central Norway. *Boreas 8*, 233–239.

Sveian, H. 1981a: Levanger, kvartærgeologisk kart CST 133134–20, M 1:20 000. *Norges geol. Unders.*

Sveian, H. 1981b: Tromsdalen, kvartærgeologisk kart CUV 133134–20, M 1:20 000. *Norges geol. Unders.*

Sveian, H. 1981c: Stiklestad, kvartærgeologisk kart CUV 135136–20, M 1:20 000. *Norges geol. Unders.*

Sveian, H.: Verdalsøra, kvartærgeologisk kart CST 135136–20, M 1:20 000. *Norges geol. Unders.* (in press).

Sveian, H.: Børgin, kvartærgeologisk kart CST 137138–20, M 1:20 000. *Norges geol. Unders.* (in press).

Sørensen, R. 1979: Late Weichselian deglaciation in the Oslofjord area, south Norway. *Boreas 8*, 241–246.

Undås, I. 1942: On the late-Quaternary history of Møre and Trøndelag. *Kgl. N. Vid. Selskap, Skr. 2*, 1–92.

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