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

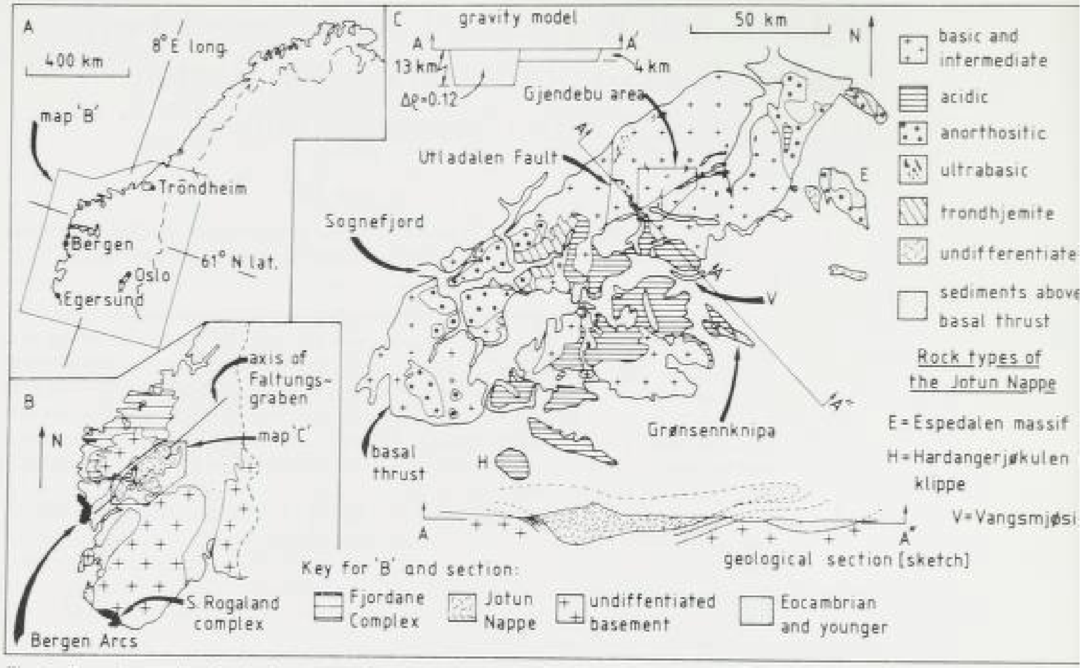


Fig. 1. Location and geological setting of the Gjendebu region. Compiled from Holtedahl & Dons (1960) and Carswell (1973). Geophysical profile after Smithson et al. (1974).

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,

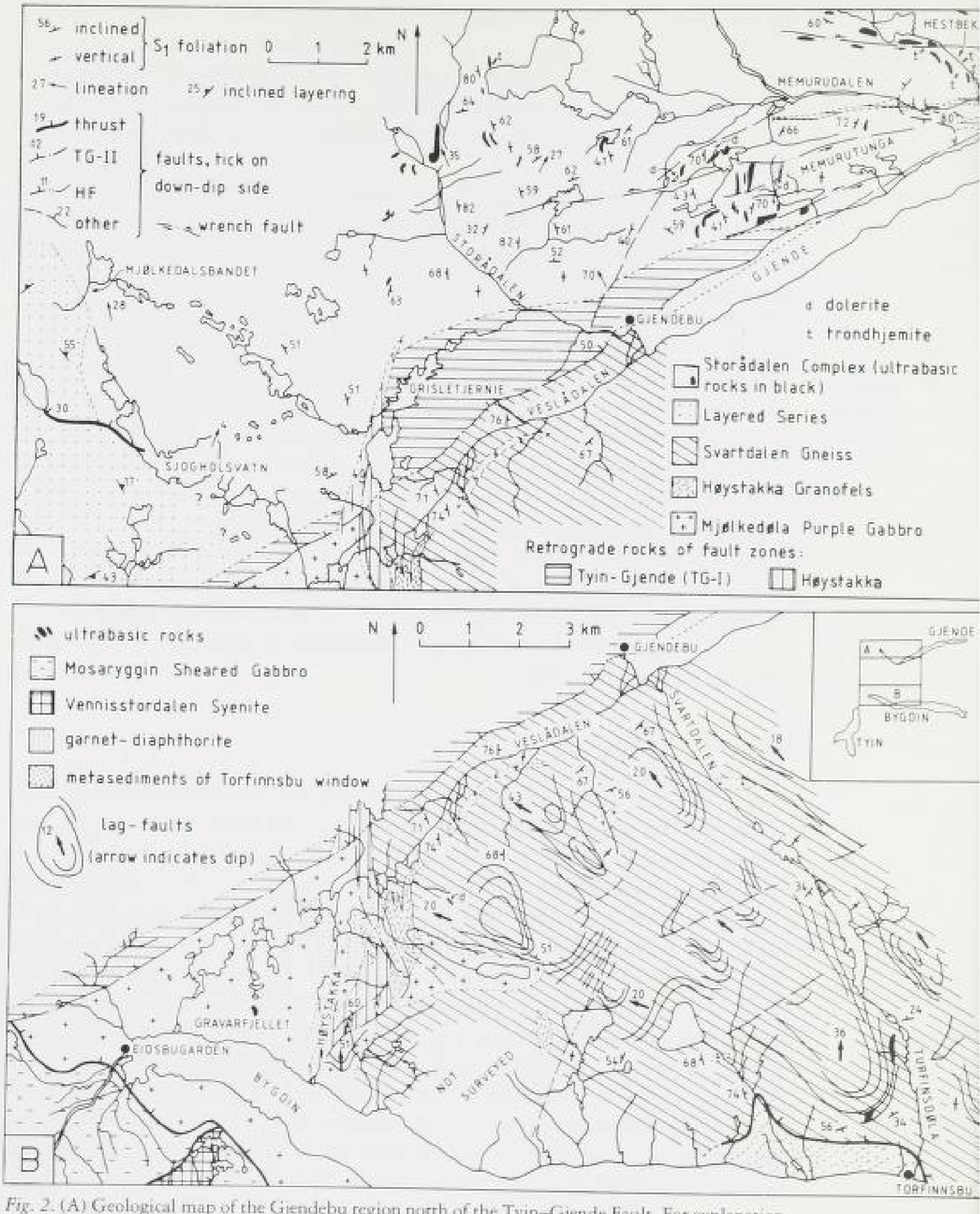


Fig. 2. (A) Geological map of the Gjendebu region north of the Tyin–Gjende Fault. For explanation of T.G.–I. and T.G.–II see text. HF = Høystakka Fault.
(B) Geological map of the Gjendebu region south of the Tyin–Gjende Fault. Other symbols as in Fig. 2A. Inset shows location and relative position of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

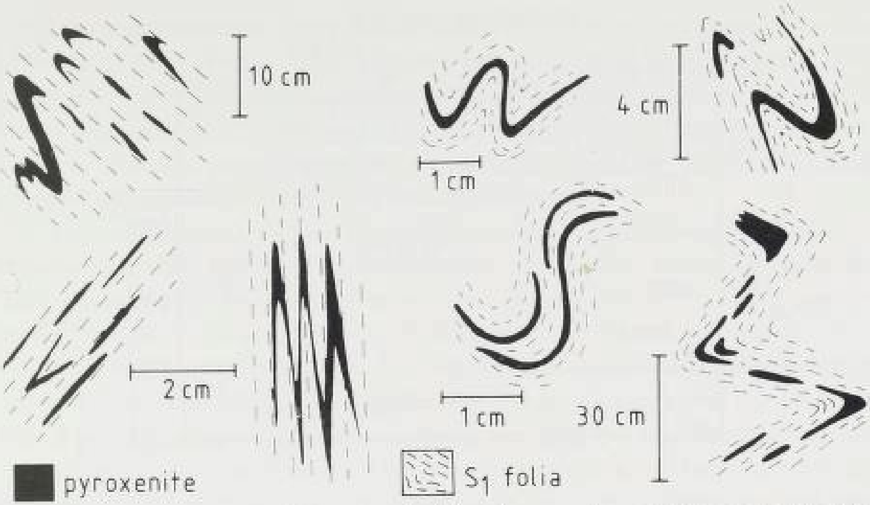


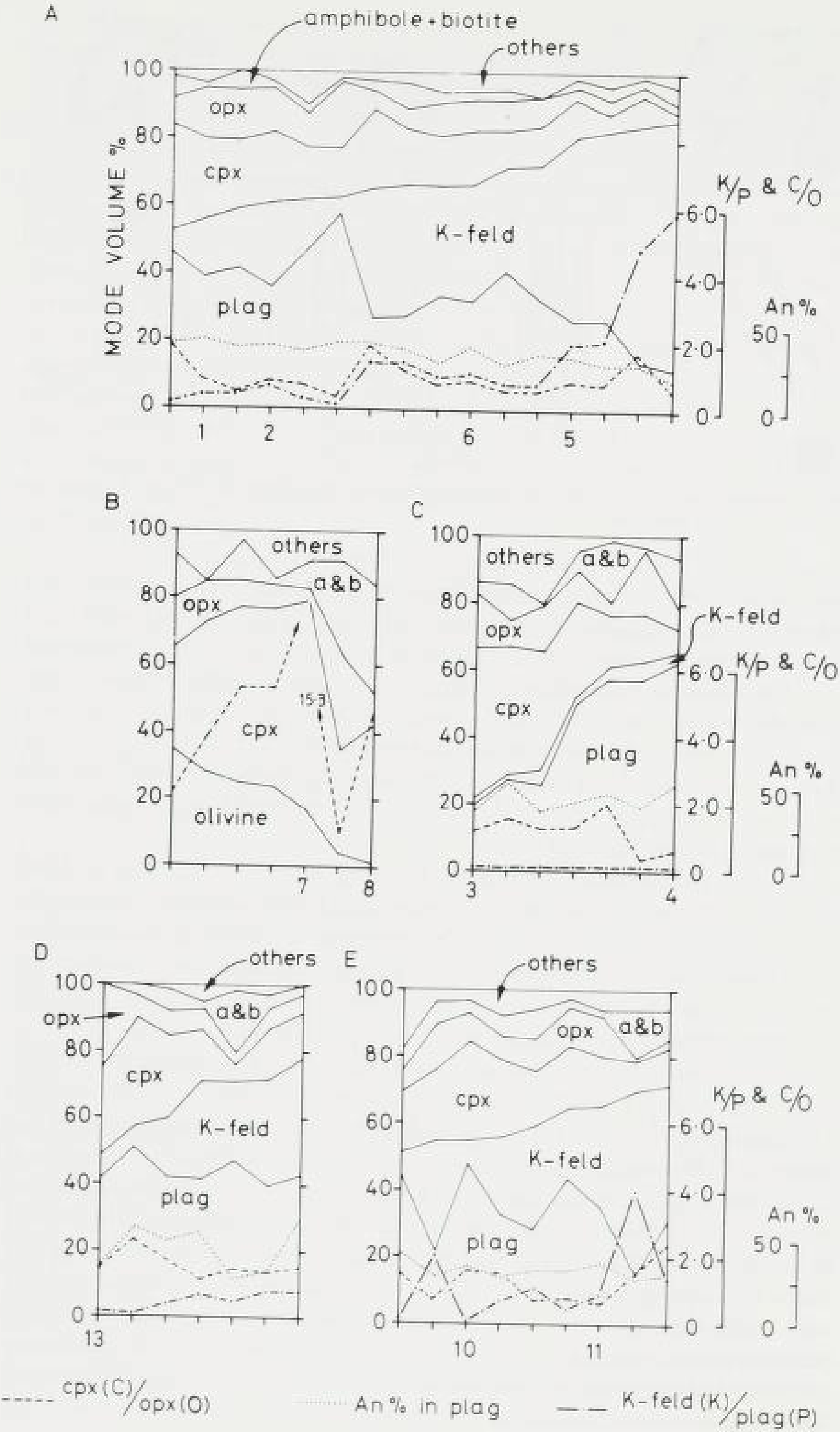
Fig. 3. Structures within the Storådalen Complex. The three examples on the left are D_1 structures, the others are D_{2A} structures.

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
MODE VOLUME %
opx
cpx
K-feld
plag
K/P & C/O
An %
B
others
a&b
opx
cpx
olivine
15·3
C
others
a&b
opx
cpx
plag
K-feld
D
others
opx
a&b
cpx
K-feld
plag
E
others
opx
a&b
cpx
K-feld
plag
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

---

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-

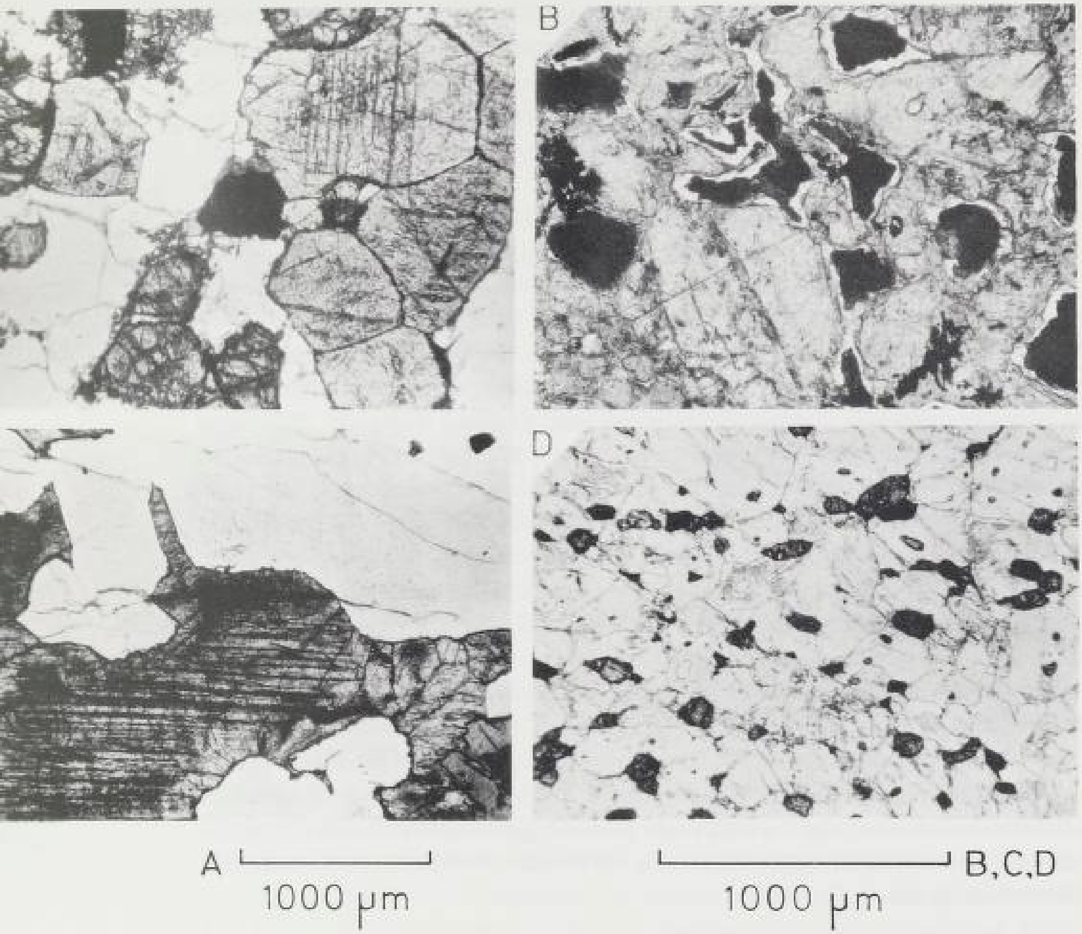


Fig 5. (A) Typical texture of pyriclasite from the Storådalen Complex. Note the smooth gently curved grain boundaries meeting at symmetrical triple points. Minerals present are plagioclase, pyroxene, opaque phases and a little biotite.
(B) Opaque grains, mainly ilmenite, surrounded by 'haloes' of plagioclase (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

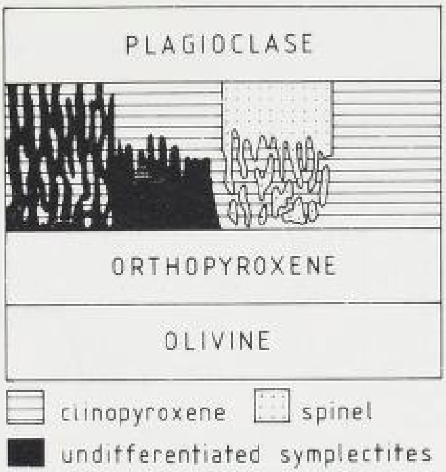
The Høystakka Granofels is distinct in both its mineralogy and its distribution, and it is surprising that its existence has not been noted previously. It occurs as numerous subvertical dykes which vary in thickness from a few centimetres to several metres, the dykes splitting and coalescing to give 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

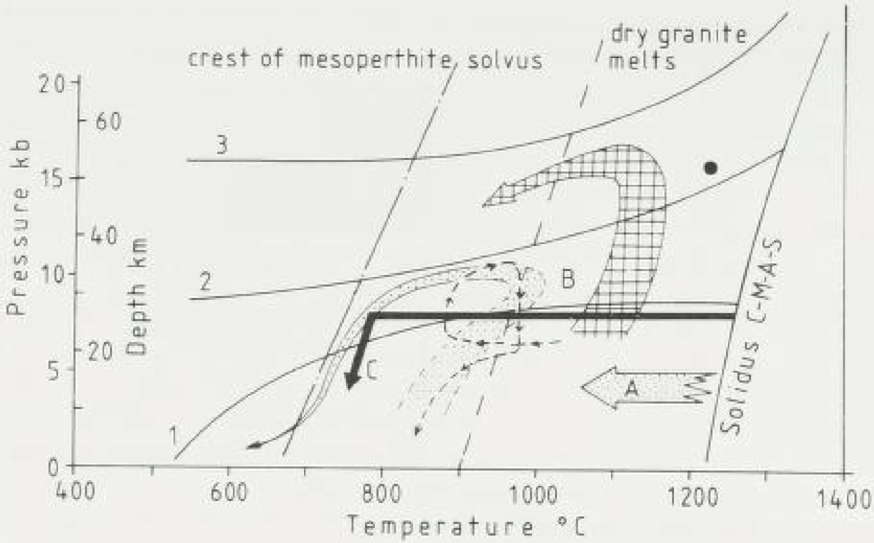


Fig. 7. The stippled arrow shows a proposed path through P–T space for the Storådalen Complex. A = approximate conditions of initial magmatic consolidation. B = climax of post–D_1 high-grade metamorphism. C = terminal decompression event. C–M–A–S is the $CaO–MgO–Al_2O_3–SiO_2$ system (O'Hara 1967). The 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	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

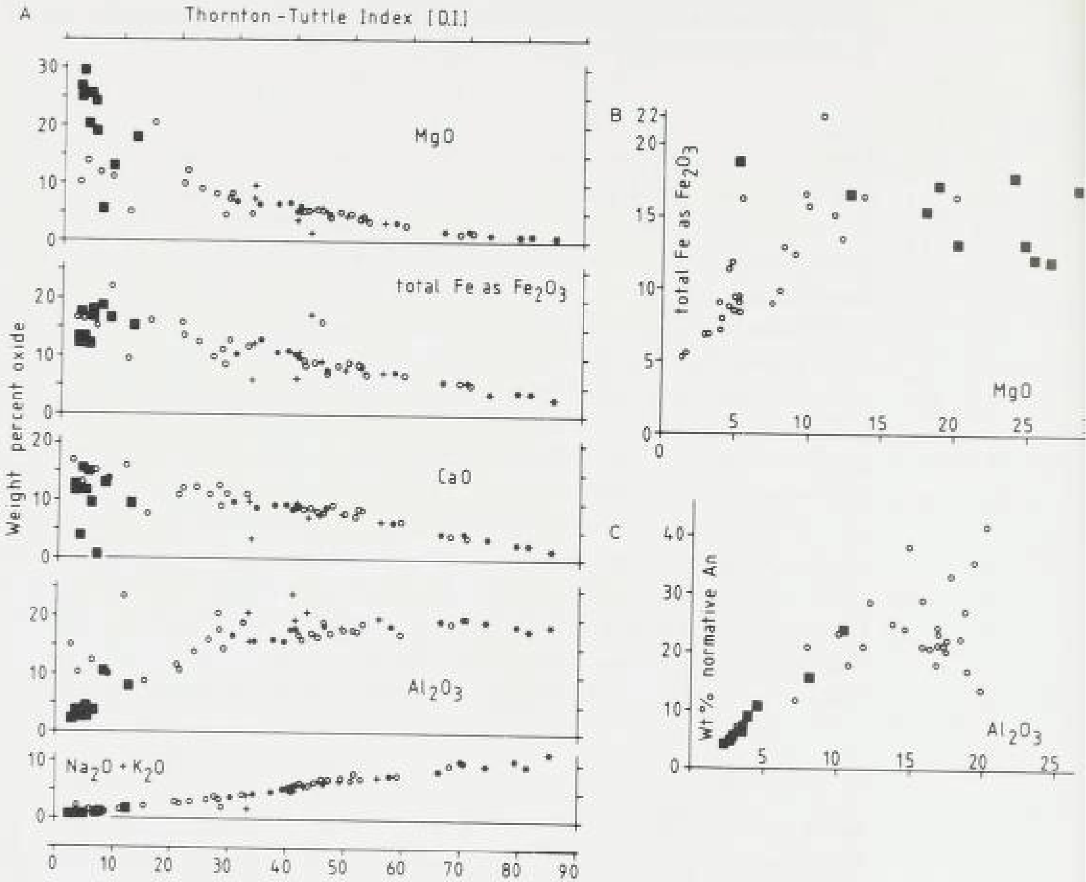


Fig. 8. (A) Variation diagrams for rocks from the Gjendebu region. The Thornton–Tuttle differentiation index is from Thornton & Tuttle (1960).
(B) Fe–Mg relationships in the Storådalen Complex.
(C) Alumina-normative An relationships in the Storådalen Complex. B and C suggest that the geochemistry of 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

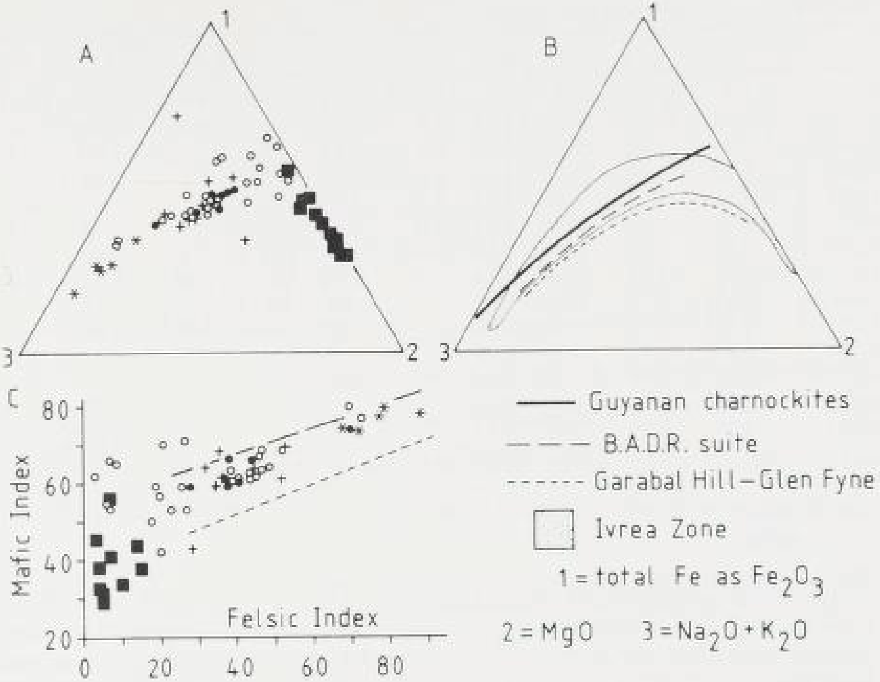


Fig. 9. (A) AFM diagram for rocks of the Gjendebu region.
(B) Comparative trends on an AFM diagram: Garabal Hill–Glen Fyne Complex from Nockolds (1941), Guyanan charnockites from Singh (1966), the trend for the B–A–D–R volcanic suite (see text) is from Turner & Verhoogan (1960), and that for the Ivrea 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	}7.2*	}9.0*	3.66	5.00
FeO	5.53	5.30	5.25	5.36	6.0	8.68			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			}0.81**	0.44
H_2O^-	0.16	0.25	0.18	0.20						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

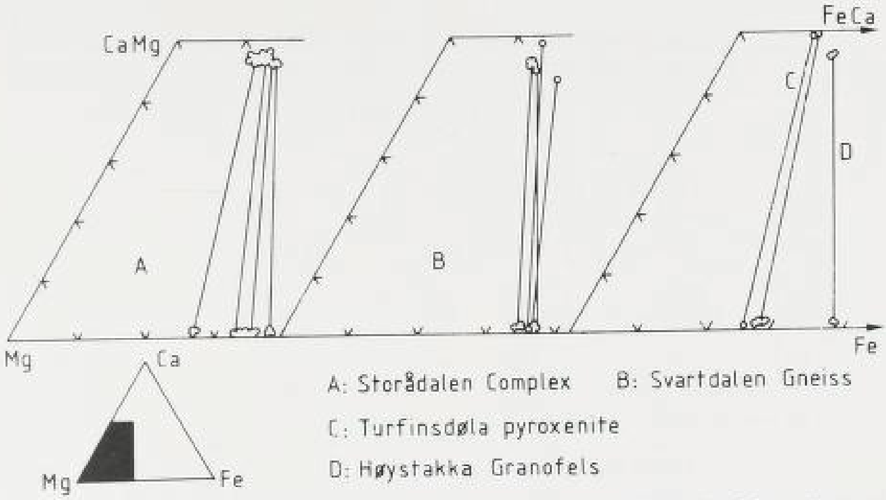


Fig. 10. Quadrilatral diagrams for co-existing pyroxenes from rocks of the Gjendebu region.

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.

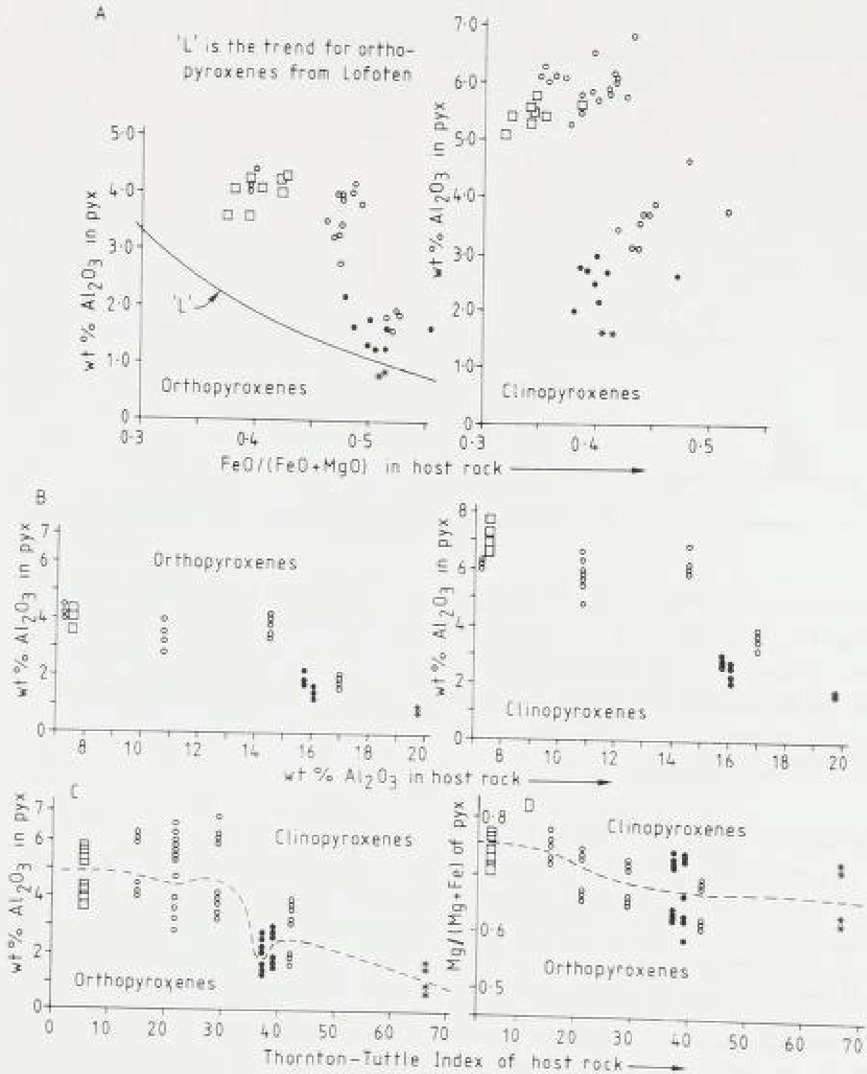


Fig. 11. Geochemical trends of analysed pyroxenes.
(A) Alumina content of pyroxenes against the ratio FeO/(FeO + MgO) in host rock. The trend from Lofoten is taken from Griffin & Heier (1969).
(B) Alumina content of pyroxenes against alumina content of the host rock. Note the strong dependence of the 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

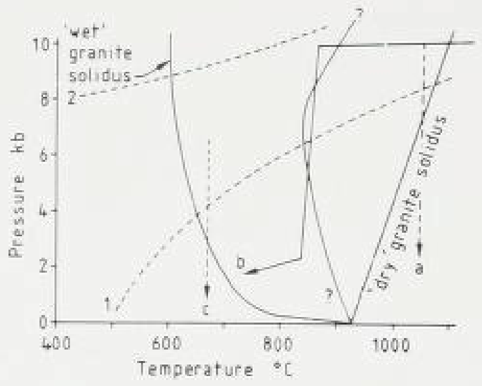


Fig. 12. Model for the generation of the mesoperthosites. The reaction curves 1 and 2 are the same as in Fig. 7. The granite solidi are from Huang & Wyllie (1975). The line labelled '? – ?' is a hypothetical solidus for $P_{H_2O} > O < P_{total}$ (cf. Eggler & Kadik 1979). The lines 'a' and 'c' represent 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(?)
		(emplacement of ultrabasic bodies)	
high grade metamorphism	Intermediate pressure granulite facies	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.

Acknowledgements. I thank Dr. M. H. Battey for much valued discussion and guidance concerning the geology of Jotunheimen. He, Dr. R. G. Park, and Lieutenant G. F. Wing all contributed to the improvement of early versions of this paper, the work for which was financed by N.E.R.C. and carried out at the University of Newcastle-upon-Tyne. I also thank Professors T. S. Westoll, J. R. Cann, and G. Kelling for the use of facilities in their respective departments. Analytical work was supervised by P. Oakley and W. Davison, and the figures were prepared for publication by D. Kelsall and Ms. P. J. Douglass.

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: Parageneses 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.