NGU

Norges geologiske undersøkelse

Nr. 378

Bulletin 69

Universitetsforlaget 1983

Trondheim - Oslo - Bergen - Tromsø

NGU
Norges geologiske undersøkelse

Geological Survey of Norway

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PUBLISHER
Universitetsforlaget, P.O.Box 2959, Tøyen, Oslo 6, Norway.

DISTRIBUTION OFFICES
Norway: Universitetsforlaget, P.O.Box 2977, Tøyen, Oslo 6, *United Kingdom:* Global Book Resourses Ltd., 109 Great Russell Street, London WC1B, 3ND.
United States and Canada: Colombia University Press, 136 South Broadway, Irvington on Hudson, New York 10533.

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The Husfjord Plutonic Complex, Sørøy, Northern Norway

DAVID L. SPEEDYMAN

Speedyman, D.L. 1983: The Husfjord plutonic complex, Sørøy, northern Norway. *Norges geol. Unders. 378,* 1–48.

The Husfjord area of Sørøy is underlain by a plutonic complex which has been emplaced into Vendian to Cambrian metasediments during the Finnmarkian phase of the Caledonian orogeny.

The metasedimentary envelope of the complex consists mainly of a sequence of psammites, pelites, semipelites, calc-silicate schists and marbles, which have undergone prolonged regional metamorphism and two principal episodes of deformation. The regional metamorphic event commenced before the first folding episode, reached its peak in the almandine-amphibolite facies between the deformation episodes, and waned during the second period of folding. The various members of the igneous complex were emplaced synchronously with these metamorphic and tectonic events, and thermal metamorphic effects produced by some members have been superimposed upon those of the regional metamorphism.

The earliest intrusion, the Husfjord metagabbro, was emplaced towards the end of the first deformation episode, and has been affected by the highest grades of regional metamorphism. A suite of diorites, monzonites and quartz-syenites was emplaced during the second deformation episode, and these have only suffered low-grade regional metamorphism. The Husfjord metagabbro and the diorite – monzonite – quartz-syenite complex were emplaced essentially by a mechanism of permissive intrusion. The latest members of the igneous complex were the Vatna gabbro and a number of minor intrusions including perthosite sheets, basic dykes, and nepheline-syenite-pegmatites. The gabbros are thought to have had their genesis in a long-lived mantle diapir in the asthenosphere above a subduction zone dipping eastwards beneath the Baltic plate. The diorite – monzonite – quartz-syenite suite developed from a dioritic magma which was generated deep in the crust by the fusion of pelitic material by the heat of metamorphism and the mantle diapir. Metamorphic mineral parageneses in pelitic hornfelses indicate that the development of the Husfjord plutonic complex took place at depths greater than 20 km.

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Introduction

Sørøy is an island off the coast of West Finnmark, near the town of Hammerfest (Fig. 1). The Husfjord area, in the southeast of Sørøy, comprises a plutonic complex, mainly gabbros and diorites, whose emplacement and subsequent deformation and metamorphism are associated with events of the Caledonian orogeny.

The Husfjord igneous complex has been intruded into Vendian (Eocambrian) to Cambrian metasediments which constitute the major part of Sørøy and which have been described by Ramsay & Sturt (1963), Appleyard (1965), Roberts (1968a, 1968b), Holland & Sturt (1970) and Ramsay (1971a). The metasediments are part of a regionally extensive stratigraphic succession and occur within the Kalak Nappe Complex of the northern Norwegian Caledonides (Sturt et al. 1975, Ramsay & Sturt 1977, Binns 1978, Sturt et al. 1978, Roberts & Sturt 1980). The geology of neighbouring parts of West Finnmark has been described

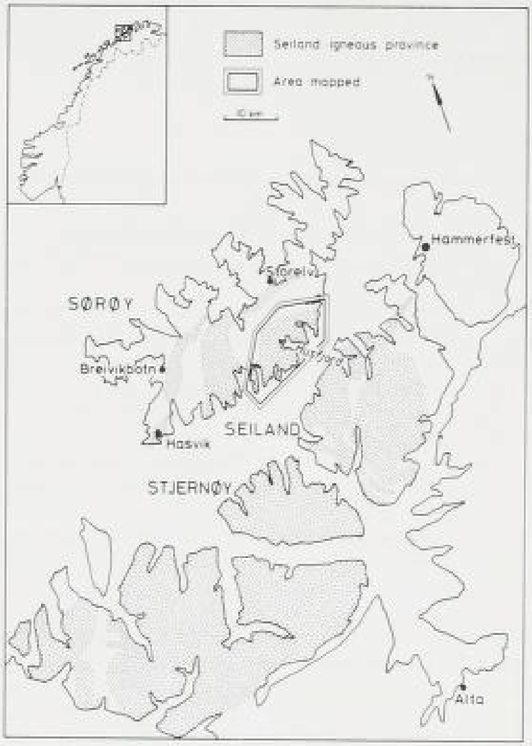

Fig. 1. Map showing the area mapped, in its setting within the Seiland igneous province.

by Holtedahl (1918), Holmsen et al. (1957), Strand (1960), Ball et al. (1963), Reitan (1963) and Hooper & Gronow (1970), and a discussion of the Caledonian nappe sequence and the timing of orogenic deformation and metamorphism of Finnmark has been gived by Sturt et al. (1975, 1978). Norges Geologiske Undersøkelse has produced a geological map (Roberts 1974) and an aeromagnetic map (N.G.U. 1971), both on the scale 1:250 000, and Brooks (1970) has made a gravity survey of West Finnmark. Ramsay (1973) indicated a possible plate tectonic setting for the region.

A number of igneous bodies have been emplaced into the metasediments of Sørøy (Fig. 1). The Storelv gabbro has been described by Sturt & Taylor (1972) and Stumpfl & Sturt (1965), and the latter authors made a mineralogical and geochemical comparison with the Breivikbotn gabbro. Aspects of the Hasvik gabbro have been studied by Sturt (1969), Gardner (1972) and Robins & Gardner (1974). Near Breivikbotn is an alkaline complex which has been described by Sturt & Ramsay (1965). The igneous rocks of Sørøy form the northern-most part of the Seiland petrographic province; most of the rocks in this province are basic or ultrabasic, and some exhibit layering. Accounts of various parts of the province have been given by Strand (1952), Barth (1953, 1961), Krauskopf (1954), Oosterom (1954, 1956, 1963), Heier (1961, 1964, 1965), Ball et al. (1963), Hooper (1971), Robins (1972, 1973, 1974, 1975), Bennett (1974), Gardner & Robins (1974), Robins & Gardner (1974, 1975), Robins & Takla (1979), Robins & Tysseland (1979) and Sturt et al. (1980).

N. & W. SØRØY[1]	N.E. SØRØY[2]	S.E. SØRØY[3]
Hellefjord Schist Group	Hellefjord Schist Group	Phyllitic schists
Åfjord Pelite Group Falkenes Marble Group	Falkenes Limestone Group	Calc-silicate schists with marbles
Storelv Schist Group	Storelv Schist Group Transitional Group	Garnet-mica schists
Klubben Psammite Group	Klubben Quartzite Group	Psammites and semi-pelites

Fig. 2. Comparative metasedimentary successions of Sørøy. 1. Ramsay & Sturt 1963, Ramsay 1971a, Ramsay & Sturt 1973a. 2. Roberts 1968a. 3. Speedyman, this paper.

The present paper is an account of the plutonic complex in the Husfjord area of Sørøy, and its material has been derived from a Ph. D. thesis presented at the University of London (Speedyman 1968). Mapping was done using aerial photographs on a scale of 1:15,000.

I. The Country Rocks

A. Stratigraphy and Petrography

The Husfjord igneous complex is bounded by the sea in the southeast, but to the northwest has an envelope of metasedimentary country rocks which have been overturned by folding. The earliest intrusion is the Husfjord metagabbro, and its intrusive contact is preserved along the northwestern margin of the complex. This metagabbro is a discordant sheet-like body, its margin slightly transgressing the stratigraphic succession of the country rocks (geological map, Plate I).

The generalized stratigraphic succession is as follows:

Youngest	phyllitic schists
	calc-silicate schists with marbles
	garnet-mica schists
Oldest	psammites and semipelites (migmatized in part)

The succession in other parts of Sørøy and West Finnmark has been described by Ball et al. (1963), Ramsay & Sturt (1963), Sturt & Ramsay (1965), Roberts (1968a), Ramsay (1971a), and Ramsay & Sturt (1973a), and the succession in the Husfjord area is similar. A probable correlation with some of these areas is presented in Fig. 2.

The only fossils recorded in Sørøy are archaeocyathids in impure limestones of the Falkenes Marble Group, which are considered to be Lower to early Middle Cambrian in age (Holland & Sturt 1970). This provides, for this region, a lower age limit to the Caledonian orogeny, a protracted event punctuated by several phases of magmatic activity. The youngest igneous intrusion within the late-D_2 Breivikbotn alkaline complex of west Sørøy is a syenite dyke, giving an upper Rb/Sr age limit of 490 ± 27 Ma (recalculated to $\lambda\ ^{87}Rb = 1.42 \times 10^{-11}a^{-1}$) to the orogeny (Sturt et al. 1978). This major orogenic phase, of late Cambrian to early Ordovician age, is called the Finnmarkian.

The lowest members of the succession are fine-grained feldspathic psammites and semipelites; the former are greyish-white to buff with small reddish-brown garnets, while the latter contain abundant large garnets overgrowing and replacing biotites. These rocks have undergone extensive migmatization along a broad belt having its maximum intensity at the northern end of Kobbefjord.

Above these rocks are grey and rusty weathering, garnet-mica schists, often containing quartzose bands and streaks, probably due to metamorphic segregation. Small red garnets and retrogressive muscovite overprint kyanite, sillimanite and early biotite. In the upper part of the group are occasional calc-silicate schists containing actinolite which occur as bands and lenses within the semipelites; towards the top the psammites and semipelites are of less importance, and the calc-silicate schists become more abundant, forming a gradation into the overlying group.

The overlying calc-silicate schists are mainly well-bedded and fine-grained, with numerous pale greenish actinolite-rich bands. Interbedded with these schists, near the base of the group, are occasional diopside marbles which occur as long tectonic lenses, each several tens of metres in length and about ten metres in thickness, elongated parallel to the general layering of the rocks.

The phyllitic schists are light grey, well-banded, flaggy rocks, some feldspathic and slightly micaceous, while others are rich in biotite and poikiloblastic muscovite.

B. Tectonics

Structures belonging to two complex and protracted episodes of Finnmarkian deformation are recognized in the metasediments. In the Husfjord area, the general attitude of the metasedimentary layering is due principally to the second phase of folding.

The first deformation episode (D_1) in Sørøy is characterized by large-scale recumbent folding in which movement was towards the east and southeast (Ramsay & Sturt 1963, Roberts 1968a, Ramsay 1971b), and the metasediments in the Husfjord area all lie on one limb of a large D_1 fold. The earliest D_1 minor folds are tight to isoclinal, lying within the general metasedimentary layering, and the axes of these folds are curved within their axial planes. A detailed analysis of noncylindrical folds on Sørøy has been given by Ramsay & Sturt (1973a, 1973b). These D_1 folds are refolded by later folds, and are cut by amphibolitized basic sheets.

In the second major episode (D_2) folds of more open style were formed, and these tend to have orthorhombic or monoclinic symmetry (Ramsay & Sturt 1963). They generally have axial trends close to those of the D_1 structures, and in the case of the monoclinic folds the eastern- or southeastern-facing limbs are often slightly

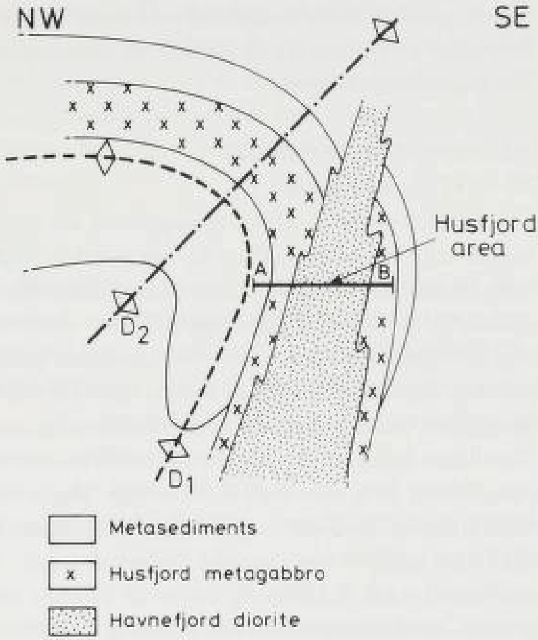

Fig. 3. Simplified NW-SE cross-section across the Husfjord area showing the relationship between the major intrusions and the major structures. Line of section A–B shown on Plate I. After Speedyman 1972.

overturned. The metasediments of the Husfjord area lie within one of these steep, overturned, southeast-facing limbs. The D_2 folding had two main phases, and in the Husfjord area there was considerable igneous activity during and after the earliest D_2 movements. Late D_2 folding was responsible for the present arcuate form of the igneous complex and its envelope, and the latest D_2 deformation at the end of the regional metamorphism was brittle.

The major structures of Sørøy have been discussed by Sturt & Ramsay (1963), Roberts (1968a), Ramsay (1971b) and Ramsay & Sturt (1973a), the Falkenes Marble proving a useful marker horizon in elucidating the structure of the island. Relics of the Falkenes Marble occur as trains of rafts within the igneous complex of Husfjord, where the stratigraphic succession youngs towards the southeast, away from the major D_1 fold axial trace.

The emplacement of the Husfjord igneous complex spanned the period of time between the waning stages of D_1 and the latest phases of D_2 The complex has two major plutons: the Husfjord metagabbro which was intruded discordantly along the upper limb of a large recumbent isoclinal D_1 fold, and the Havnefjord diorite which was emplaced into the steep overturned limb of a major D_2 fold which refolded the D_1 isocline and the Husfjord metagabbro (Fig. 3).

II. The Igneous Complex

A. Field Relationships

The Husfjord igneous complex comprises major plutonic intrusions, principally gabbros and diorites, and a number of minor intrusions, including monzonites, quartz-syenites, basic dykes and perthosite sheets (Plate I). These bodies have been emplaced at various times during the development of the Finnmarkian stage of the Caledonian orogeny.

1. HUSFJORD METAGABBRO

Facies of the metagabbro

The outcrop of the Husfjord metagabbro, the earliest intrusion, runs in an arc from southwest to northeast (Plate I), but much of the central zone is now occupied by later intrusions. The gabbro is a melagabbro which has undergone metamorphism and is fairly uniform, although there are some slight variations.

The typical gabbro is fine- to medium-grained but is sporadically coarser-grained, a primary feature of the gabbro. There is also a finer-grained facies which sometimes has a poorly formed foliation subparallel to the margin of the gabbro. This latter facies is not near the contact, but mainly in the central part of the body, particularly in a zone rich in later thin coarse-grained diorite sheets which have been intruded into the metagabbro. This foliation is a fluxion texture, suggesting that the gabbro was intruded synchronously with slight tectonic movement associated with the waning stages of D_1; the similar, contemporaneous Storelv gabbro, to the north, also has a strong D_1 foliation (Sturt & Taylor 1972). This zone of finer-grained, slightly foliated gabbro provided structural weakness into which the thin coarse-grained diorite sheets were later readily emplaced.

In the southwest the metagabbro is variably contaminated by metasediment, developing a gabbronoritic facies containing metasediment rafts. In places the gabbro is extremely contaminated, forming xenolithic norite, as seen to the north of Kobbefjord and on the Fella peninsula. In Kobbefjord, where the country rocks are migmatized psammites and semipelites, this noritic facies, developed along the marginal zone, is a fine-grained rock containing dark brown garnets. It is extremely xenolithic, inclusions varying from large rafts of migmatite several metres long to small fragments of psammite.

The contact between the norite and the country rocks consists of a zone about a metre wide in which the rocks are a mixture of contaminated garnetiferous norite and assimilated migmatite. Away from the contact the garnets, which are abundant in the neighbourhood of xenoliths in the marginal zone, diminish in both number and size, and it is clear that the formation of garnet was associated with the assimilation of aluminium-bearing semipelites by the gabbroic magma. Several features of this xenolithic norite are similar to those of the contaminated norites of Aberdeenshire (Read 1935, 1936, Read et al. 1965, Gribble 1967, Gribble & O'Hara 1967), although there is an absence of cordierite and more garnet in the Husfjord rocks, presumably due to greater pressures. On the Fella peninsula the garnetiferous norite is similar to that in Kobbefjord, but the rafts are mainly of hornfelsed psammite, semipelite and calc-silicate schist.

The Husfjord metagabbro contains sporadic lenses and layers of troctolite, ranging up to about 5 m in length, elongated parallel to the trend of the gabbro sheet. Usually the lenses are isolated and dispersed throughout the gabbro, but there is one horizon on Husfjordnes in which there is a swarm of closely packed tectonic lenses of troctolite within a zone of D_2 shearing. Husfjordnes is the only locality where continuous troctolite layers have been found, usually only a few centimetres in thickness. One layer dipping steeply to the north is graded, having a sharp ultramafic upper margin and grading downwards into leucocratic rock and if this represents primary gravitational layering it would confirm that the gabbro sheet has been inverted by D_2 folding.

Metasedimentary rafts and xenoliths

As well as the noritic facies, the normal metagabbro contains a number of large elongate rafts and small xenoliths of metasediment, particularly in the north, though here the gabbroic magma was not contaminated by the inclusions. The rafts, generally a few tens of metres in length, include marbles, mica schists and psammites, while the xenoliths, which are usually less than a metre in length, are of mica schist, calc-silicate schist and psammite.

The largest and most abundant of the rafts consist of bluish-grey marble, and although these sometimes occur as isolated bodies they usually appear as groups or trains of rafts composed of several individuals. Occasional rafts of mica schist, up to a few tens of metres in length, have provided weaknesses in the metagabbro, for they are almost invariably intruded by later sheets of coarse-grained quartz-syenite. On Ramnes the metagabbro contains xenoliths and rafts of migmatized psammite and semipelite. The migmatization postdated the emplacement of the gabbro, for the latter has undergone extensive feldspathization and veining by quartzofeldspathic material derived from partial anatexis of the metasediments. In the northwestern part of Havnefjordfjell the marginal zone of the metagabbro is rich in small metasedimentary xenoliths, sometimes only a few centimetres in length. Thes are of several different lithologies including psammite, semipelite, calc-silicate schist, basic hornfels, and blocks of acid pegmatite. They are usually angular and sharp-margined, and are randomly orientated with different rock-types brought into juxtaposition with one another.

The trains of metasedimentary rafts form a relict stratigraphy through both the Husfjord metagabbro and the Havnefjord diorite, a large fine-grained diorite which has been emplaced into the central part of the metagabbro sheet (Plate I). The marble rafts form a good marker horizon, and these occur well within the metagabbro body near to the Havnefjord diorite in the southwest, but close to the outer margin of the metagabbro in the northeast. The trend of these rafts is approximately parallel to the general strike of the country rocks outside the pluton. The preservation of this relict stratigraphy across a large part of the metagabbro sheet is a result of the emplacement mechanism discussed below.

Early intrusions

The finer-grained foliated facies of the Husfjord metagabbro has been susceptible to feldspathization and the introduction of dioritic material. The diorites are coarse-grained pyroxene-mica diorites with a fluxion texture and generally occur as parallel sheets less than a metre in width (Fig. 4), although some irregular veins are present. Many have been sheared and jointed during D_2, and the large-scale orthogonal swing in the regional strike of the country rocks, due to the late D_2 movements, is reflected by a similar swing in strike in these early diorite sheets.

The foliated facies of the Husfjord metagabbro has also been prone to feldspathization in the neighbourhood of the dioritic sheets. The metagabbro contains streaks and bands replete with feldspar porphyroblasts which are always elongated parallel to the foliation of the metagabbro. In some cases the porphyroblasts are relatively dispersed and the borders of the feldspathized areas are diffuse, while in others the porphyroblasts are closely spaced and the margins of the bands of feldspathization are sharply defined.

The emplacement of the diorites postdates the feldspathization since they cut across the feldspathized patches, but it is possible that the two events were closely linked in time, the feldspathization being a precursor to the diorite intrusion. Petrography shows that the diorite intrusion followed the peak of the regional metamorphism, whilst it is likely that the feldspathization could have been associated with higher grades of the regional metamorphism. The zones containing diorites have been prone to shearing, probably due to internal failure in the metagabbro during D_2 folding along planes of weakness formed by these diorite sheets. Sometimes only the diorites have been sheared, but locally the metagabbro is also affected.

Late intrusions

A complex of diorites, monzonites and quartz-syenites, of variable grain-size, was emplaced into the central zone of the Husfjord metagabbro. The largest body is the fine-grained Havnefjord diorite, which thermally metamorphosed the Husfjord metagabbro, the width of the aureole being 300–400 m. The metagabbro becomes more amphibolitic in the neighbourhood of the diorite, and sometimes feldspar porphyroblasts develop, occasionally containing minute inclusions of mafic minerals in their cores. Some of the early coarse diorite sheets in the metagabbro occur within the aureole of the Havnefjord diorite, and these have also been amphibolitized. There is no evidence in the metagabbro that the Havnefjord diorite caused any marked deformation of its envelope during its emplacement.

A number of perthosites have been emplaced into one of the more intense shear-zones in the Husfjord metagabbro around the margin of the Vatna gabbro. They are pink or cream coloured, varying in grain-size from fine to coarse, and ranging in size from sheets tens of metres thick to small irregular streaks and veins only a few millimetres wide. The mechanism of emplacement of these perthosites has been discussed in detail by Speedyman (1973).

Emplacement

The distribution of the metasedimentary rafts provides the clue to the emplacement mechanism of the gabbro sheet. This mechanism must account for the preservation of a relict stratigraphy in the rafts within the metagabbro. It must also explain the presence of metasedimentary rafts in the metagabbro near to the Havnefjord diorite, well away from the gabbro's outer contact.

The raft distribution suggests that on emplacement the gabbro sheet bifurcated,

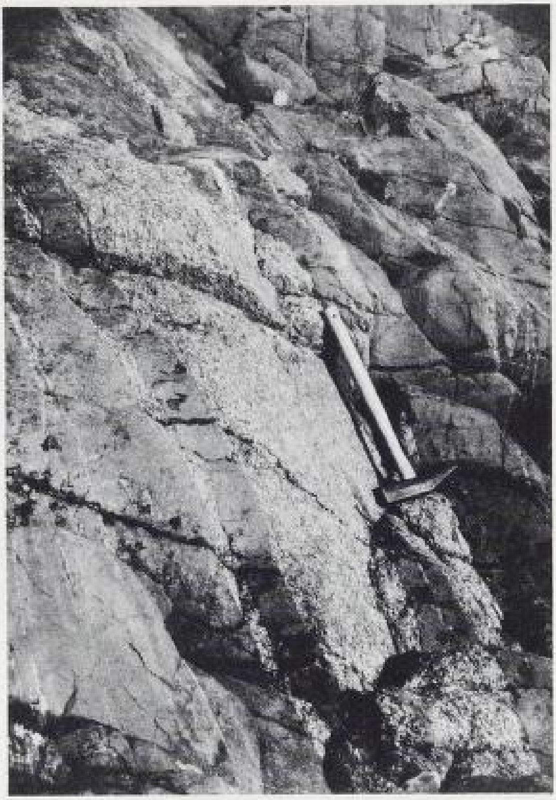

Fig. 4. Early diorite sheets in the Husfjord metagabbro. Husfjord.

preserving a large central lenticular region predominantly of country rock, consisting of large screens of metasediment alternating with sheets of gabbro (Speedyman 1972). This central lenticular complex is now occupied by the Havnefjord diorite, accounting for the general alternation of metagabbro and metasediment rafts within the diorite. The emplacement of the gabbro was syntectonic, associated with the last stages of the D_1 folding, and Speedyman (1972) considered that the intrusion of the Husfjord gabbro was by a mechanism of syntectonic permissive emplacement.

2. HAVNEFJORD DIORITE

This is a large, fairly homogeneous body of fine-grained pyroxene-mica diorite, emplaced into the central zone of the Husfjord metagabbro (Plate I).

It has not undergone the high-grade regional metamorphism that affected the metagabbro, and thus postdates the peak of the regional metamorphic event. Its outcrop is arcuate, about 12 km in length and nearly 2 km in width at its centre, and takes the form of a sheet, both contacts dipping towards a northwesterly direction.

The diorite is almost invariably aphyric and is fairly uniform in grain-size throughout, although sometimes near the margins, particularly the southern, there are zones parallel to the contact in which the diorite is slightly porphyritic exhibiting a fluxion texture.

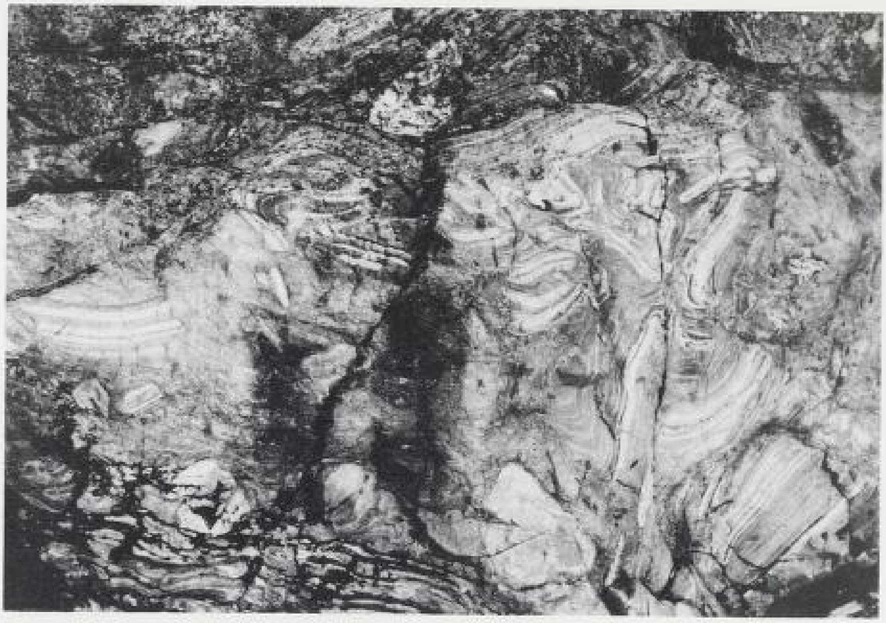

Fig. 5. Mobilized hornfels in a metasediment raft in the Havnefjord diorite. Havnefjord.

Rafts and xenoliths

A common feature of the Havnefjord diorite is the occurrence of numerous inclusions, varying from occasional xenoliths a few centimetres long to numerous large rafts up to several tens or hundreds of metres in length. The xenoliths are of metasediment but the rafts consist of various lithologies including metagabbro, psammite, semipelite, basic schist and marble. The rafts occur throughout the diorite, but the various rock-types are restricted to different zones, reflecting a relict stratigraphy comparable with that of the country rocks. The small xenoliths mainly occur near the northern margin, and are not obviously related to the rafts.

The metagabbro rafts are of the order of a few metres or tens of metres in length, and are lenticular, with their longest axes parallel to the margins of the diorite. Occasionally they contain early diorite sheets and zones of feldspathization, and have been thermally metamorphosed.

Much of the central part of the Havnefjord diorite, especially to the north of Husfjord, is occupied by planar rafts of psammite, rusty-weathering semipelite and basic schist, ranging from a few metres long up to about 1 km in length and 200 m in width. They are orientated parallel to the margins of the diorite, and sometimes occur as trains which trend in this direction. The layering in the rafts is parallel to the raft margins, which are usually subvertical or dipping steeply to the north.

In some places D_1 minor folds are refolded by D_2 folds, indicating that the diorite emplacement postdates at least the beginning of D_2. The margins of the rafts are fairly sharp, and there is no significant assimilation by the diorite. The spatial arrangement of rafts of similar composition indicates that they are close to

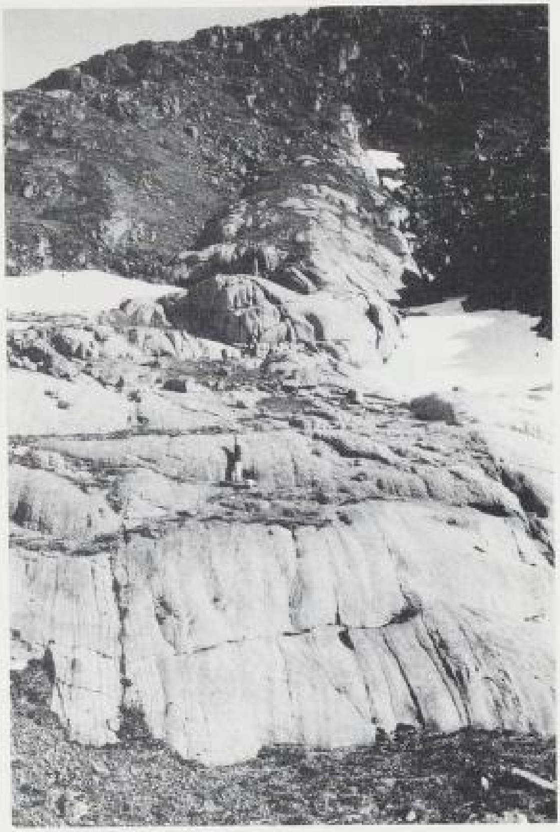

Fig. 6. Marble raft in the Havnefjord diorite. Havnefjordfjell.

their original pre-diorite positions, since a relict stratigraphy can be traced across the body. The size of many rafts also indicates that they have probably not changed their positions very much. The psammitic and semipelitic rafts are very abundant in the central zone of the diorite, and here sporadic metagabbro rafts occur between and parallel to the metasedimentary raft trains. Away from the central zone, and particularly towards the south, the metasedimentary rafts decrease in size and number, and the frequency of metagabbro rafts increases.

In one large raft hornfelsed metasediments have become mobilized by the diorite, and form broken and deformed blocks in a dioritic matrix (Fig. 5). The blocks have sharp margins, sometimes with leucocratic reaction rims, and the diorite in the neighbourhood of these mobilized hornfelses has been net-veined by quartzofeldspathic material, the source of which appears to be the metasediment.

The northern limit of the rusty-weathering semipelitic rafts is fairly clearly defined, and to the north of this the Havnefjord diorite contains occasional rafts of marble. Most of these are large, often many tens of metres in length (Fig. 6), and are elongated parallel to the trend of the psammite and semipelite rafts and the diorite contact.

Fig. 7. Steep northward-dipping contact between monzonite (left) and Havnefjord diorite. Havnefjordfjell.

Fig. 8. Feldspathization of metasediment near quartz-syenite sheet. Havnefjord.

A few randomly-orientated psammite and semipelite xenoliths occur in the northern part of the diorite. These are almost invariably partly assimilated by the diorite, and their margins are generally diffuse, sometimes only remaining as dark schlieren. Locally, tight isoclinal D_1 folds have been preserved in the xenoliths. Such small, highly-digested xenoliths must have been brought up by the diorite from deeper levels.

Emplacement

As in the case of the Husfjord metagabbro, the distribution of the rafts within the Havnefjord diorite provides the key to its emplacement mechanism. The preservation of a relict metasedimentary stratigraphy throughout the diorite must be explained, and the presence of metasedimentary rafts within the diorite at such a great distance from the metasediments of the country rocks must be accounted for. Moreover the alternation of metagabbro and metasediment rafts across the diorite must be explained.

As already mentioned, when the Husfjord gabbro was emplaced it bifurcated forming a central lenticular region consisting of sheets of gabbro alternating with metasediment screens. It was into this sheeted screen complex that the Havnefjord diorite was emplaced synchronously with late D_2 folding (Speedyman 1972). Both metagabbro and metasediment were incorporated as rafts, and the pre-existing sheeted nature of the screen complex resulted in the alternation of metagabbro and metasediment rafts. This also explains the relict stratigraphy and the preservation of the metasediments within the heart of the diorite so far from the country rocks. Like the Husfjord metagabbro, the Havnefjord diorite is considered to have been intruded by a mechanism of syntectonic permissive emplacement (Speedyman 1972).

3. MONZONITES

The Havnefjord diorite and in some places the Husfjord metagabbro have been intruded by a number of coarse-grained pyroxene-mica monzonite sheets (Plate I). These vary in width from a few metres to many tens of metres, the largest having a length of about 1,5 km, although it is likely that many of the smaller ones link up with longitudinally adjacent ones beneath the present erosion surface. The sheets dip steeply towards the north (Fig. 7), and their trends are subparallel to the margins of the Havnefjord diorite and its metasedimentary rafts.

The contacts of the monzonites against their host are always sharp, but there is no chilling, and the monzonites commonly contain hornfelsed blocks of Havnefjord diorite at their margins.

The monzonites are frequently located adjacent to metasedimentary rafts in the Havnefjord diorite, although in some cases the rafts have been completely enveloped by the monzonites. A common feature of the monzonites, whether or not they are associated with rafts, is the presence of small metasediment xenoliths. Psammitic xenoliths usually have fairly well-defined margins, but pelitic and semipelitic inclusions are generally diffuse, and have been extensively assimilated by the monzonite, forming vague schlieren. In some places semipelitic horizons in the larger xenoliths can be traced along strike from normal metasediment through stages of progressive assimilation until they become indistinguishable from monzonite.

4. QUARTZ-SYENITES

These occur as sheets, a few metres in width, emplaced into the Havnefjord diorite, the Husfjord metagabbro, and occasionally the country rocks (Plate I), but their age relative to the monzonites is uncertain in the field. They are coarse-grained, and some contain numerous small reddish garnets. These quartz-syenites are usually associated with metasedimentary rafts which have sharply defined margins, but there are none of the small digested xenoliths characteristic of the monzonites.

In the neighbourhood of the quartz-syenites, the Husfjord metagabbro and the metasediments have generally been extensively feldspathized, the latter being much more prone to this than the former. Feldspar porphyroblasts up to about 2 cm in length are randomly orientated, and in some places the feldspars are relatively sparse, whereas in others they are very dense (Fig. 8). The feldspathization is sometimes so dense that the rock begins to resemble the quartz-syenites, and it is possible that the relationship between the quartz-syenites and the feldspathization is comparable to that between the early diorite sheets in the Husfjord metagabbro and the feldspathization associated with them.

5. VATNA GROUP

The Vatna gabbro occupies the area around Vatna, and makes contact with the Husfjord metagabbro to the north and west (Plate I). It has not undergone the amphibolite-grade regional metamorphism which has affected the metagabbro, and it causes the formation of a metamorphic aureole in the latter. It comprises a number of facies, mostly olivine-bearing, having various grain-sizes, but with a general tendency for the rocks in the southern part to be finer-grained, and in places a fluxion texture is developed. Dioritic patches are fairly common, especially near its western margin and these, together with granular pyroxene-hornfels inclusions, are probably relics of Husfjord metagabbro and its associated diorites. The Vatna gabbro has metamorphosed the Husfjord metagabbro, the aureole being about 1 km wide to the west but narrowing to a few hundred metres in the north. The metagabbro becomes variably amphibolitized, and in places there are feldspar porphyroblasts.

Southward-dipping sporadic and variable layering is present in the gabbro. Two types occur: one is a gradation in grain-size, and the other is a gradation in the proportion of mafic minerals present. In the former, olivine-gabbro with an ophitic texture gradually passes up into a coarser-grained variety before reverting rapidly to the finer-grained facies; the mineralogy remains the same throughout and units are about 0.5 m thick. The second type of layering is fine-scaled with units 0.5–2.0 cm in thickness, and is defined by variation in the ratio of plagioclase to pyroxene, olivine and opaque minerals across the units. In many cases layers are well defined and continuous for many metres, varying little along the strike, whereas in others the layers are diffuse and discontinuous laterally. Although many of the units grade from mafic up to felsic, the direction of grading is often conflicting in neighbouring units. The graded units are not generally adjacent to one another, but are usually separated by several centimetres or metres of homogeneous olivine-gabbro.

Inclusions and veins

Inclusions are fairly common in the gabbro, especially around Vatna; they tend to occur in zones, individual inclusions ranging up to a few tens of centimetres in length. Sometimes they are randomly orientated, and neighbouring ones are frequently of different rock-types. The most common type is a finer-grained facies of the host-rock itself, and these are generally elongate and aligned parallel to the attitude of the fluxion structure in the host. They are autoliths, representing an early phase in the emplacement of this facies of the Vatna gabbro complex, and their origin may be similar to the inclusions in the hypersthene-gabbro of Ardnamurchan, described by Wells (1953).

At Vatna the gabbro commonly has a streaky appearance with bands and veins of coarse pink-weathering feldspathic material. These are parallel to the autoliths and fluxion structure, and some have sharp margins while others are diffuse and mix with the gabbro. The veins and streaks are of perthite, and in those cases where they merge with the gabbro the latter is pink-weathering and in thin-section is seen to be a mixed rock containing both olivine and antiperthitic feldspar. Thus it appears that this facies of the Vatna gabbro had an alkali magma associated with it, either during or shortly following its emplacement.

The gabbro contains a few veins rich in opaque minerals, principally magnetite, which occur mainly in the streaky antiperthitic facies. Their attitude is parallel to the perthosite veins, dipping steeply south, suggesting that an ore body may exist beneath Sørøysund. It is significant that the 1:250 000 Hammerfest aeromagnetic map produced by Norges Geologiske Undersøkelse (1971) shows a magnetic maximum over the Vatna area.

Intrusions

Minor intrusions are abundant in the Vatna gabbro, and include perthosite sheets, basic dykes and nepheline-syenite pegmatites. The perthosites consist almost entirely of hair-perthite, are variable in grain-size from fine to medium, weather to a pink or yellowish colour, and form large sheets up to several tens of metres in width trending NW-SE. In many cases the ends of the sheets are diffuse and merge with pink-weathering antiperthite-gabbro. There is a gradation from olivine-gabbro through antiperthitic olivine-gabbro and melaperthosite to practically pure perthosite, and it appears in places that the perthosite and olivine-gabbro magmas must have become mixed at the time of emplacement; the perthosite may, in fact, be a late fraction of the alkaline olivine-gabbro magma. It is possible that the perthosites were emplaced before the Vatna gabbro was completely solid, so that those parts of the perthosite sheets in contact with the gabbro became mixed with the neighbouring crystallizing host. Pitcher & Read (1960) have described dykes in the Donegal Granite which display mixing of granite and dyke material at their margins and a merging of textures, and which are considered by these authors to have been emplaced into an embryonic joint system in the granite before the latter was entirely solid. The Vatna perthosites have many features in common with the Donegal dykes, and may have had a similar mode of emplacement.

The Vatna gabbro and perthosites are cut by numerous basic dykes, mainly amphibolites, some of which are porphyritic. They range in width from about 10 cm to 2 m, although a distinctive white-spotted olivine-leucogabbro reaches several tens of metres in width. At Vatna a few nepheline-syenite pegmatites have been emplaced into the gabbro. They are invariably sheared, having provided planes of weakness in the massive gabbro along which shearing could take place during the late D_2 brittle deformation, the intensity of the shearing being variable, even within a single body. Feldspars form broad elliptical-sectioned augen, while nephelines form long, sinuous, streaked-out granulated augen, and shearing has also causes the formation of small overturned isoclinal folds. Similar features in western Sørøy have been described by Sturt (1961).

B. Petrography

1. HUSFJORD METAGABBRO

Facies of the metagabbro

The Husfjord metagabbro is essentially a clinopyroxene-gabbro (Table 1) which has been variably amphibolized during the regional metamorphism. The main facies is fine- to medium-grained, typically has a relict subophitic texture, and is sometimes slightly foliated. As a result of regional metamorphism, plagioclase has

Table 1. Modal analyses of the facies of the Husfjord metagabbro.

Mineral	1 36/3A	2 36/139E	3 36/119C	4 36/141A	5 05/38B	6 38/5A	7 46/11A	8 05/34B	9 11/47F	10 56/1A	11 60/47F	12 60/47G	13 60/47H
Qu	–	–	0.5	–	–	–	–	0.3	–	–	–	–	–
Plag	33.2	35.4	44.3	52.6	45.4	34.1	40.6	31.7	77.3	58.2	51.9	37.4	65.8
Cpx	13.7	6.6	2.8	13.5	22.7	2.4	9.5	–	5.4	–	–	–	–
Hyp	–	–	–	0.3	1.8	–	–	–	4.4	3.9	16.7	20.7	21.2
Hbl	46.3	52.7	42.2	29.4	23.0	–	–	–	0.8	2.1	2.8	3.0	2.5
Act	–	–	–	–	–	60.6	43.4	64.9	–	–	–	–	–
Bi	3.2	3.1	7.6	2.8	5.6	0.8	3.7	0.4	4.6	10.7	3.4	3.4	5.6
Sph	0.5	0.3	0.6	–	0.1	0.8	0.3	2.1	–	–	–	–	–
Ap	0.1	0.1	0.2	0.2	0.2	0.4	0.4	0.3	0.5	0.4	0.8	0.7	1.0
Opaq	3.0	1.8	1.8	1.2	1.2	0.9	2.1	0.3	7.0	2.9	4.6	5.3	3.6
Gnt								–	–	21.7	19.8	29.5	0.3
Zir								–	–	0.1	–	–	–

1–2, fine-grained facies; 3–5 coarse-grained facies; 6–8, actinolitic facies; 9, contaminated facies; 10–13, noritic facies. Based on 3,000 points per thin-section.

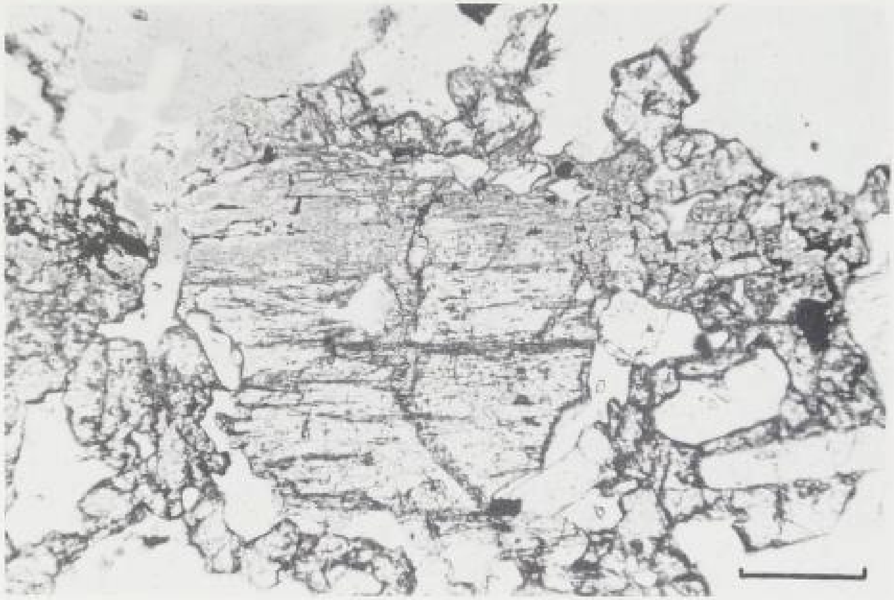

Fig. 9. Garnet corona around hypersthene; note clear zone of quartz. Noritic facies of the Husfjord metagabbro. Plane polarized light. Bar scale = 0.1 mm.

recrystallized and clinopyroxene has become partially altered to green amphibole. Hypersthene occasionally occurs in this facies, (Table 1), but is less prone to amphibolitization than the augite. There are two generations of biotite; early laths altering to hornblende, and later large fresh crystals which have developed retrogressively from both hornblende and early biotite.

In the noritic facies the only pyroxene is strongly pleochroic hypersthene. Ragged biotite laths form from some of the hypersthenes, and there is usually a narrow zone of quartz between the two minerals with vermicular inclusions of quartz in the biotites as a result of the reaction. Corona textures are ubiquitous in this facies, the most common being rims of granular garnet around hypersthene generally with a clear zone of quartz between the two (Fig. 9). These garnet rims also occur around hypersthenes which have altered to biotite containing vermicular quartz inclusions, and where garnet overgrows biotite it sometimes inherits the vermicular quartz inclusions from the biotite, indicating that the formation of garnet is secondary, postdating the metamorphic alteration of hypersthene to biotite. Garnet coronas around orthopyroxene in noritic and related rocks, often with an intermediate zone of quartz, have been described by many workers (Brøgger 1934, Shand 1945, Gjelsvik 1952, Friedman 1955, Murthy 1958, Reynolds & Frederickson 1962, Engels & Vogel 1966, Frodesen 1968, Glaveris 1970, Griffin 1971, Griffin & Heier 1973, and Whitney & McLelland 1973). Although some of these authors believe the reactions to be deuteric, the majority consider them to be metamorphic. In this facies garnet also occurs as large porphyroblasts, particularly at the contact with the country rock migmatites where the norite is strongly contaminated by metasediment. The close association of garnet formation with assimilated metase-

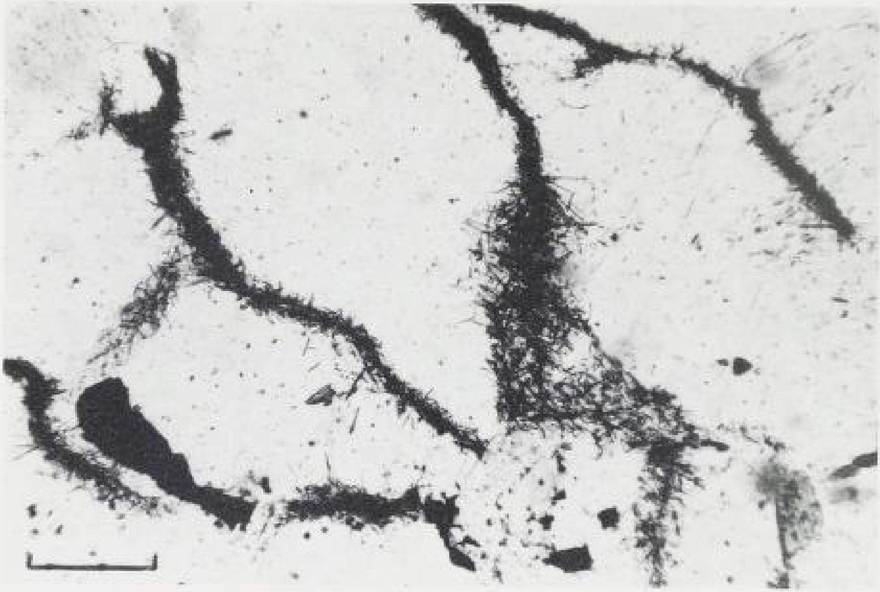

Fig. 10. Fibrolite along feldspar grain boundaries. Migmatized psammite raft in the Husfjord metagabbro, Ramnes. Plane polarized light. Bar scale = 0.1 mm.

diment is not a primary feature, as a result of direct contamination, but a secondary one. It is due to the metasediment providing excess aluminium, creating chemical metastability in the more contaminated parts of the norite which became stabilized during metamorphism by the formation of garnet.

The Husfjord metagabbro has been thermally metamorphosed by the Havnefjord diorite. Towards the contact pyroxene, which has already been slightly altered to amphibole during the regional metamorphism, becomes progressively more amphibolitized, and although it is almost completely replaced by amphibole a relict subophitic texture is preserved. Plagioclase is sodic andesine, and there are sporadic porphyroblasts of antiperthitic oligoclase.

Metasedimentary rafts and xenoliths

The marble rafts consist mainly of coarsely recrystallized calcite forming a subequigranular mosaic in which rounded grains of diopside are common, and small amounts of alkali feldspar, quartz, garnet and idocrase are sometimes present. The thin calc-silicate horizons have a fine-grained granular texture, and consists of diopside, calcite, quartz and feldspar. Locally, the metagabbro adjacent to the marble rafts contains a few anhedral buff-coloured garnets, due to contamination. In the mica schist rafts garnets overprint the folded D_1 schistosity, but have themselves undergone D_2 deformation, with secondary biotite forming along cracks in the porphyroblasts. The psammite and semipelite rafts have been migmatized after being enveloped by the metagabbro. The psammites are quartzofeldspathic, with varying amounts of biotite, sillimanite and garnet, the last sometimes forming large porphyroblasts which are altering to fibrolite and

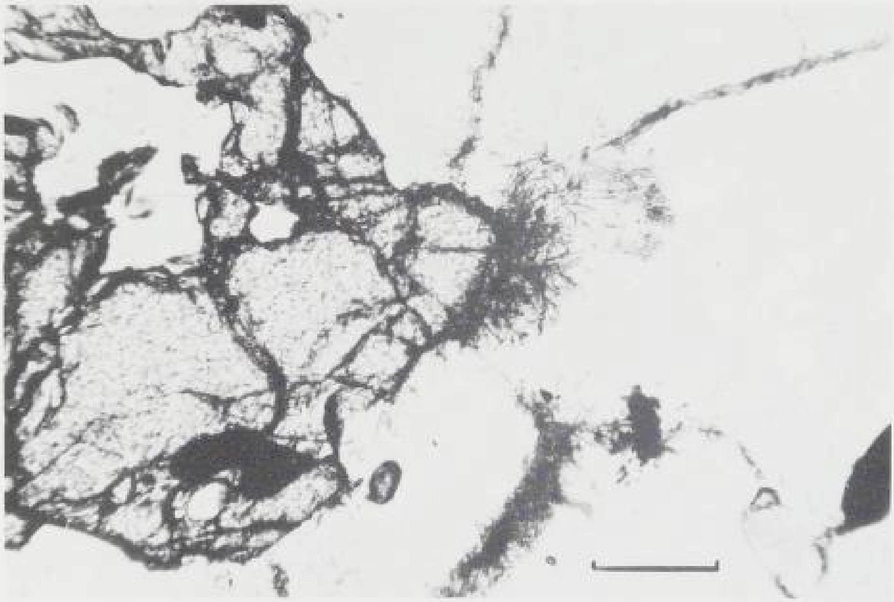

Fig. 11. Fibrolite nucleating on garnet porphyroblast. Migmatized psammite raft in the Husfjord metagabbro, Ramnes. Plane polarized light. Bar scale = 0.1 mm.

opaques, especially at their margins. Along nearly all the grain boundaries between adjacent alkali feldspars, and between alkali feldspar and quartz are aggregates of fibrolite (Fig. 10), a mineral which also nucleates around the margins of garnet porphyroblasts (Fig. 11). The growth of this fibrolite appears to be related to the migmatization, and its significance is discussed in Part III. In the semipelites quartz and alkali feldspar form a subequigranular mosaic, often with biotites along grain boundaries, and some horizons contain garnet which is breaking down to late regrogessive biotite.

Four principal lithologies are presesented in the small xenoliths: psammite, semipelite, calc-silicate schist and basic hornfels. The psammites consist mainly of quartz which, together with a few plagioclase grains, forms a fine-grained mosaic. A relict early foliation is preserved by layers of biotite, garnet, opaques, kyanite, sillimanite and rutile. Kyanite is in the process of altering to sillimanite, and fan-shaped aggregates of fibrolite needles nucleate on garnet. In the semipelites a hornfelsic texture overprints a relict schistosity, the main minerals being plagioclase, biotite, garnet and quartz. The principal mineral in the calc-silicate schists is diopside, occurring as rounded grains, which together with plagioclase forms a granular hornfelsic texture overprinting an early regional metamorphic fabric. Tremolite, quartz and poikiloblastic biotite form a D_2 schistosity, and enclose the rounded grains of diopside and plagioclase. The basic hornfelses represent basic metasediment inclusions, and have a fine-grained granoblastic texture consisting of orthopyroxene and plagioclase, the former beginning to alter retrogressively to amphibole and biotite. In xenoliths in the noritic facies, orthopyroxenes are rimmed by garnet.

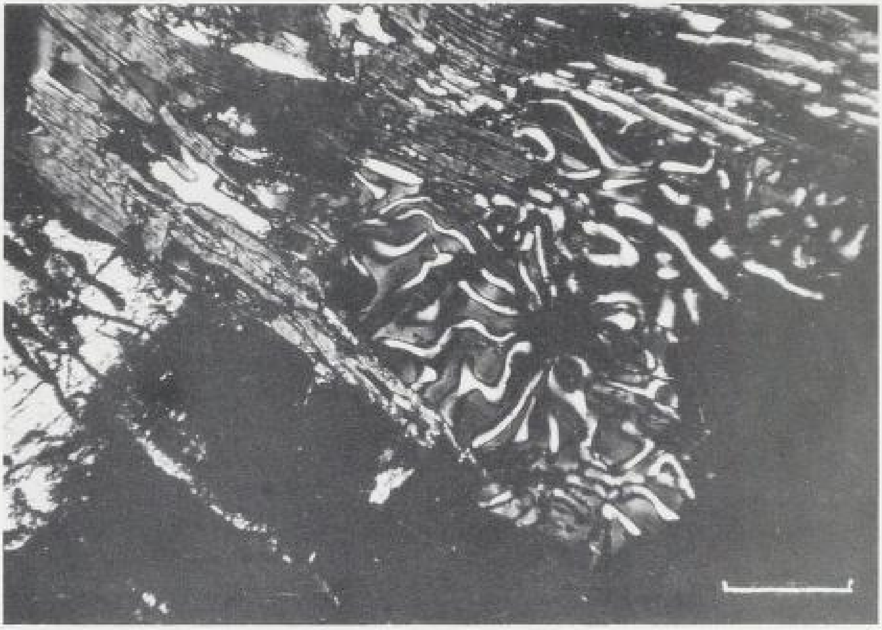

Fig. 12. Biotite with vermicular quartz associated with myrmekite developing from alkali feldspar. Early diorite intrusion in Husfjord metagabbro, Husfjord. Crossed polars. Bar scale = 0.1 mm.

Early intrusions

These are pyroxene-mica-diorites, containing a little alkali feldspar, in which hypersthene and phenocrysts of plagioclase tend to have a preferred orientation parallel to the sheet margins. At the contacts between hypersthene and alkali feldspar biotite often forms at the expense of hypersthene. The biotite usually contains vermicular inclusions of quartz, and this symplectite is frequently closely associated with the development of myrmekite from alkali feldspar (Fig. 12). This is a late metamorphic effect, discussed in Part III. Hornfelsing of the metagabbro by the early diorites postdates the peak of the regional metamorphism, since the regional metamorphic texture and mineralogy are overprinted by a fine-grained granoblastic texture of rounded grains of hypersthene, clinopyroxene and andesine. In places the early diorites have been metamorphosed by the Havnefjord diorite; hypersthene is pseudomorphed by turquoise-green amphibole having a diablastic texture with small rounded blebs of quartz, and this amphibole in turn is regrogressed to biotite containing vermicules of quartz.

2. HAVNEFJORD DIORITE

This is a pyroxene-mica-diorite, sometimes monzodioritic, with a fine-grained subequigranular xenomorphic texture, in which the principal pyroxene is usually hypersthene (Table 2). Although generally non-porphyritic, it occasionally has feldspar phenocrysts which tend to be aligned in a fluxion structure, particularly in the marginal zone. These phenocrysts consist of antiperthite and sometimes perthitic alkali feldspar, the latter generally bordered by myrmekite. When

Table 2. Modal analyses of the Havnefjord diorite

Mineral	36/92B	36/117C	36/126A	36/159B	36/233C	38/48A	05/58A
Qu	–	–	–	2.9	–	0.3	–
Kfeld	37.2	3.5	4.7	0.8	49.7	6.2	5.5
Plag	37.6	59.1	49.6	67.6	21.9	43.5	67.2
Hyp	17.2	23.8	10.8	13.4	15.4	14.7	15.6
Cpx	–	2.5	18.5	–	–	23.2	1.6
Bi	3.6	6.6	12.4	7.8	5.8	6.1	1.7
Hbl	1.8	–	0.8	0.1	0.3	0.2	0.3
Ap	0.2	0.2	0.3	0.9	1.4	0.7	1.3
Zir	–	–	–	0.1	–	–	0.2
Opaq	2.4	4.3	2.9	6.4	5.5	5.1	6.6

3,000 points per thin-section.

diopsidic augite occurs it is often being altered to green hornblende, which in turn is altering to retrogressive poikiloblastic biotites containing vermicular quartz. At contacts with the later monzonite sheets, the Havnefjord diorite is frequently hornfelsed, developing a fine-grained granoblastic texture, mainly consisting of hypersthene, diopside and plagioclase.

Rafts

Some rafts are of recognizable metagabbro, variably hornfelsed, but hybrids are often developed in which the diorite has been contaminated by metagabbro inclusions, forming a rock of intermediate character. These hybrids have a xenomorphic texture like that of the diorite, but the mineralogy has similarities with that of the metagabbro in that the principal pyroxene is clinopyroxene, although some hypersthene does occur, and the opaque minerals are the same as in the metagabbro.

At the margins of the psammite and semipelite rafts a hornfelsic texture develops in which quartz and perthite form polygonal grains, with decussate biotites. Thermal metamorphic garnets overgrow and form from biotite and kyanite, and diffuse relics of radiating kyanite prisms remain in the garnet (Fig. 13). In some rafts, regional metamorphic garnets have become unstable at the raft margins during hornfelsing, and all stages in their pseudomorphing by aggregates of fibrolite, iron oxide, and sometimes biotite can be observed. Some of the more pelitic schists have been slightly feldspathized and contain occasional perthite porphyroblasts.

Basic schist rafts have become two-pyroxene hornfelses due to the thermal effects of the diorite. Hypersthenes are usually large and poikiloblastic with so many inclusions of plagioclase that a diablastic texture develops, similar to the sieve-textured porphyroblastic hypersthenes in sedimentary xenoliths in the hypersthene-gabbro of Ardnamurchan (Wells 1951). Plagioclase is poikiloblastic and appears to be becoming more basic during hornfelsing; the excess silica that cannot be incorporated into the more calcic plagioclase is exsolved as vermicular quartz inclusions (Fig. 14). Long poikiloblastic biotites having a preferred orientation overgrew the pyroxene and plagioclase during regional metamorphism.

Fig. 13. Radiating relict kyanite prisms within garnet. Semipelite raft in Havnefjord diorite, Havnefjord. Plane polarized light. Bar scale = 0.2 mm.

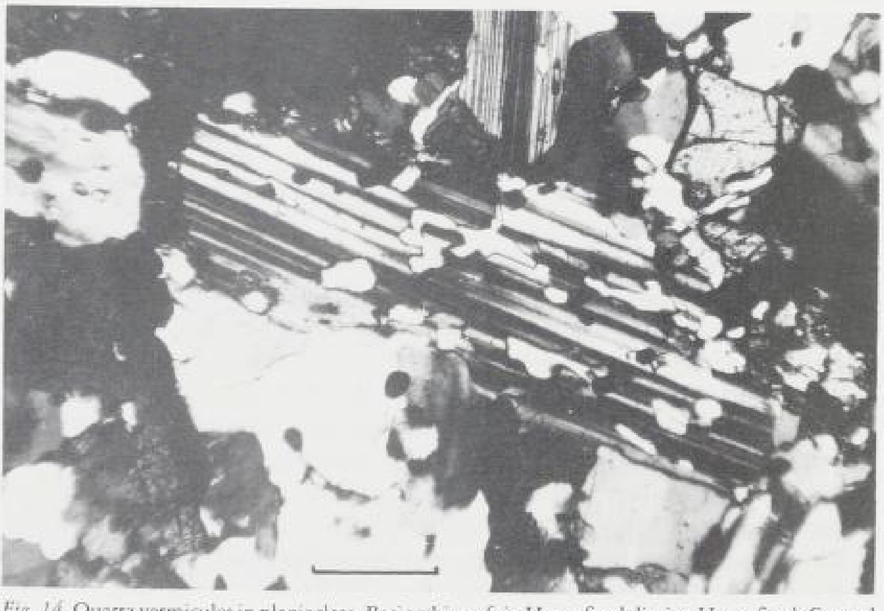

Fig. 14. Quartz vermicules in plagioclase. Basic schist raft in Havnefjord diorite, Havnefjord. Crossed polars. Bar scale = 0.2 mm.

In the marble rafts calcite forms a subequigranular mosaic with fairly straight grain boundaries, and diopsides, apatites, alkali feldspars and opaques occur as interstitial grains. The calc-silicate horizons in the marbles consist of diopside, andesine, opaques and rarely zircon.

Table 3. Modal analyses of the monzonites

Mineral	36/112B	36/237A	36/301A	36/301B	38/18B	38/38A	38/48E
Qu	1.4	–	–	–	–	1.2	–
Kfeld	21.3	32.6	29.6	17.5	50.5	37.8	35.2
Plag	50.6	54.5	53.5	64.4	22.7	43.1	51.5
(Myrm)	–	–	–	–	–	1.9	0.8
Hyp	18.7	7.0	13.0	12.9	13.3	8.9	7.7
Bi	3.6	3.5	1.5	1.7	9.0	4.9	2.3
Hbl	0.5	0.7	0.6	0.2	0.5	0.1	0.2
Ap	–	0.2	0.3	0.5	0.7	–	–
Zir	–	–	–	0.1	–	0.1	–
Opaq	3.9	1.5	1.5	2.7	3.3	2.0	2.3

7,000 points per thin-section.

3. MONZONITES

These are coarse-grained pyroxene-mica-monzonites, are sometimes monzodioritic (Table 3), and have a porphyritic xenomorphic texture in which the pyroxene is strongly pleochroic hypersthene. Biotite, commonly containing vermicules of quartz, frequently envelops hypersthene and sometimes forms from it. Anhedral phenocrysts of perthite and anti-perthite are generally randomly orientated but sometimes form a crude fluxion texture, and the perthite phenocrysts are commonly lobed by myrmekite.

Xenoliths

These range from blocks a few tens of centimetres in length to small fragments distinguisable within a single thin-section, and many are partly assimilated by the monzonite. They are principally semipelites, but there are also some basic hornfelses.

In the semipelites a granoblastic hornfelsic texture is developed with polygonal grains of alkali feldspar, usually hairperthite, and ragged decussate biotites, although a relict schistosity is sometimes preserved. Towards the contacts with the host, biotite is replaced by fibrolite and iron oxides, but in some rocks large fresh poikiloblastic biotites have grown during the waning stages of the regional metamorphism. At some contacts biotite is altered to small hypersthenes, and as the contact is crossed hypersthenes and feldspars gradually become coarser and the amount of biotite decreases until the rock resembles the monzonite.

The basic hornfelses are mostly of metasediment but a few are of Husfjord metagabbro. They have a fine-grained granoblastic texture, generally consisting of hypersthene, labradorite, decussate opaques, and sometimes biotite and clinopyroxene. At the contacts hypersthenes are often large and poikiloblastic, approaching in size those in the host monzonite, and hypersthenes in the monzonite near the xenoliths tend to be very strongly pleochroic suggesting contamination by the metasediment. It appears that these hornfelses too are being converted to monzonite at their margins; further evidence for this transition is described below.

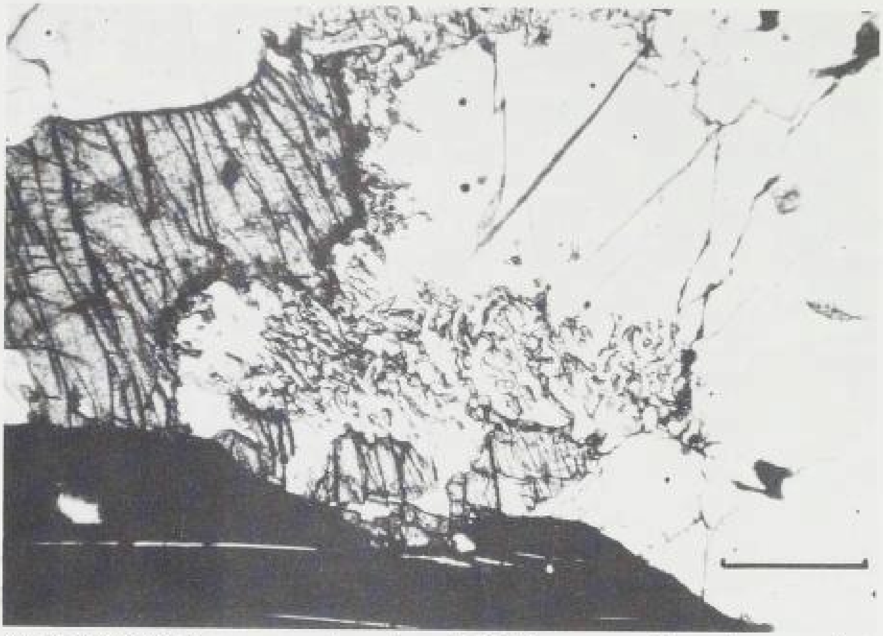

Fig. 15. Vermicular clinopyroxene at hypersthene-alkali feldspar contact. Monzonite next to a basic hornfels xenolith, Havnefjordfjell. Plane polarized light. Bar scale = 0.4 mm.

Symplectites and coronas

These develop in the monzonite next to those xenoliths which are being assimilated. Symplectites form next to the basic hornfels inclusions, whereas coronas occur in the neighbourhood of the semipelitic xenoliths. In both cases chemical instability, apparently as a result of the presence of xenoliths, is indicated.

The symplectites are confined to boundaries between hypersthene and alkali feldspar, the hypersthene becoming frittered with the development of diopside-oligiclase symplectites (Fig. 15). The formation of these is associated with contamination by basic xenoliths, since hypersthene and alkali feldspar are stable together away from the xenoliths. Another phenomenon occurring in the monzonite next to the basic hornfels xenoliths is the presence of opaque inclusions, sometimes vermicular, within hypersthene. These are magnetite and ilmenite, and as the former does not normally occur in the monzonite contamination by the xenoliths is suggested. In the basic hornfels xenoliths themselves small hypersthenes also contain opaque inclusions, often vermicular, closely resembling those in the monzonite hypersthenes. At contacts with xenoliths the monzonite contains large, strongly pleochroic hypersthenes containing a number of small separate inclusions of opaque, each of which resembles the inclusions within the small hornfels hypersthenes (Fig. 16). The large monzonite hypersthenes appear to have formed by the amalgamation of a number of hornfels hypersthenes, each retaining its central opaque inclusion, and under cross-polars the large monzonite hypersthenes are seen to consist of aggregates of small domains. Furthermore, in places the actual annexing of a small hornfels hypersthene to a large monzonite

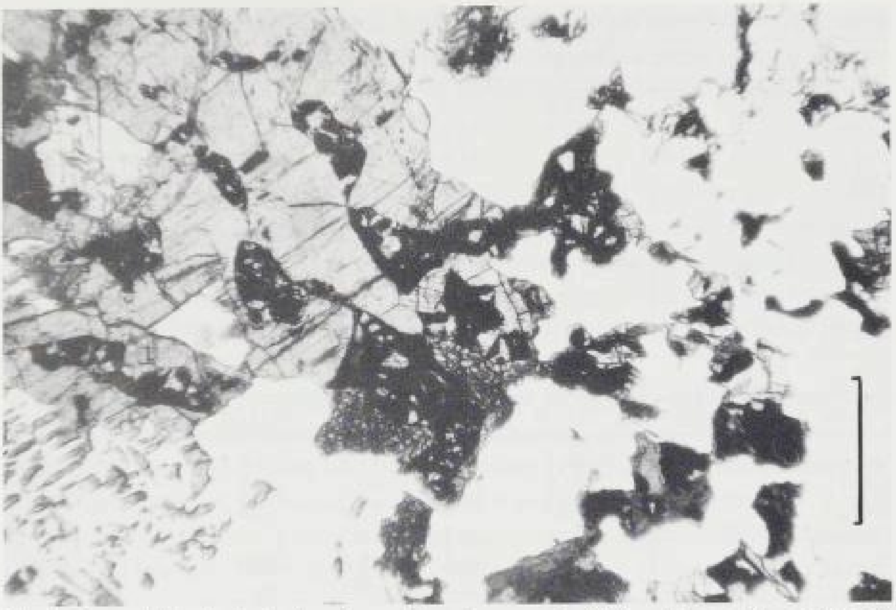

Fig. 16. Iron oxide inclusions in large hypersthene in monzonite (left) and in small hypersthenes in hornfels. Contact of basic hornfels xenolith in monzonite, Husfjord. Plane polarized light. Bar scale = 0.4 mm.

hypersthene, with opaques along the join, can be seen (Fig. 16). Thus some of the minerals in the monzonite appear to have developed from the hornfelsic material which it assimilated.

Coronas involving several minerals form in the monzonite in the neigbourhood of some of the semipelitic xenoliths. Opaques have reaction coronas of dark green hornblende or hypersthene against plagioclase, and sometimes a double corona is formed with hypersthene on the inside against the opaque mineral and hornblende outside against the plagioclase. Some opaques have narrow coronas of granular, colourless garnet against plagioclase, and sometimes there is a shell of hypersthene between the garnet and the opaque. Garnet also forms narrow coronas around some biotites, while other biotites are rimmed by dark green hornblende. The coronas are not ubiquitous, even in the vicinity of semipelitic xenoliths, but occur in irregular fine-grained patches which appear to be assimilated inclusions. The semipelitic hornfels xenoliths themselves sometimes have corona textures very similar to those just described but on a finer scale, particularly the rimming of biotites and hypersthenes by garnet. In addition, some of the more pelitic hornfelses contain green spinel which have coronas of hypersthene or garnet, or both, with garnet on the outside. These coronas in the xenoliths must have formed during hornfelsing, and the fine-grained patches in the monzonite where coronas occur are diffuse relics of assimilated metasediment. Similar symplectites and coronas in the immediate vicinity of xenoliths have been described from the Insch norite of Aberdeenshire (Read 1966). Read considered that the xenoliths were being converted directly into norite by assimilation, and that the norite in the neigbourhood of xenoliths developed as a result of contamination by the hornfelses.

Table 4. Modal analyses of the quartz-syenites

Mineral	38/52A	38/56C	48/29B	03/15C
Qu	22.3	19.6	2.2	9.4
Kfeld	45.4	62.5	47.1	33.0
Plag	20.6	12.7	32.8	39.0
(Myrm)	1.2	–	–	–
Bi	9.3	3.6	17.7	16.0
Musc	0.2	1.5	–	2.2
Hbl	0.2	–	–	–
Ap	0.4	–	–	–
Zir	0.1	0.1	0.2	0.3
Opaq	0.3	–	–	0.1

Mineral	38/13A	38/56C	48/8B	46/4A	46/17A
Qu	6.1	28.1	7.0	11.5	3.0
Kfeld	66.5	37.7	72.7	60.6	40.5
Plag	15.8	20.3	7.8	12.9	33.2
(Myrm)	0.8	–	0.7	0.5	–
Bi	9.7	7.3	10.3	11.6	18.0
Musc	0.4	1.4	0.5	0.2	–
Gnt	0.2	1.2	0.4	1.3	4.8
Ap	–	–	–	0.1	0.1
Zir	0.2	0.2	0.2	1.0	0.2
Sph	–	3.0	–	–	–
Rut	–	–	0.3	–	–
Opaq	0.3	0.8	0.1	0.3	0.2

7,000 points per thin-section.

4. QUARTZ–SYENITES

These coarse-grained quartz-syenites sometimes grade into quartz-monzonite or granite (Table 4), and have a porphyritic xenomorphic texture. Phenocrysts are usually of perthite, commonly with myrmekite developing at their margins, although some are of sodic andesine. A few quartz-syenites contain small anhedral garnets, some of which are altering to biotite; the significance of these garnets is discussed in Section D.

Associated with the quartz-syenite is a coarse feldspathization which affects both the Husfjord metagabbro and its metasedimentary rafts. The metagabbro contains xenoblastic alkali feldspar porphyroblasts, rimmed by myrmekite, and its plagioclase has recrystallized to a polygonal mosaic of sodic andesine. The pyroxenes have completely altered to actinolite and biotite, some of the latter being chloritized. Metasedimentary rafts also contain xenomorphic feldspar porphyroblasts in a matrix of quartz and biotite.

5. VATNA GABBRO

In the southern fine-grained facies, which generally have xenomorphic textures, the principal mafic minerals is pale greenish-buff augite which tends to have blebs and fine schiller needles of opaques, and in places is altered to biotite. Rounded fresh olivines and basic labradorites are sometimes elongated parallel to one another forming a fluxion texture. The northern coarse-grained facies in places exhibits ophitic textures, in which augite containing opaque schiller lamellae is slightly altering to brown hornblende. Some of the coarse rocks contain olivines which generally have narrow coronas of hypersthene or fibrous cummingtonite.

Inclusions and veins
Inclusions are of two kinds: autoliths and xenoliths. The autoliths are inclusions of fine-grained early marginal facies of the gabbro enclosed within the main body, and have a strong fluxion texture. They have the same mineralogy as typical fine-grained facies Vatna gabbro, but are finer grained. The xenoliths comprise a variety of lithologies, mainly coarse-grained pyroxenites but also brown amphibolites.

At Vatna the gabbro contains small irregular veins and streaks of coarse pink perthite which usually merge gradually into normal gabbro. In the transitional lithology the olivine-gabbro becomes progressively enriched in hair-antiperthite, which itself becomes perthite as the veins are approached. The antiperthitic streaks in the gabbro occasionally have mafic aggregates containing considerable amounts of opaque minerals. These aggregates consist of rounded grains of clinopyroxene, olivine, apatite, antiperthite and large euhedral zircons, with the opaques being interstitial. These opaques, principally magnetite but also ilmenite, pyrite, pyrrhotite and chalcopyrite, are often rimmed by biotite where they make contact with antiperthite.

Intrusions
The perthosites are composed almost entirely of hair-perthite, accessory minerals including aegerine-augite, diopside, hypersthene, hornblende, apatite and zircon; the hypersthene is strongly corroded and variably pseudomorphed by iddingsite. The rock has a xenomorphic texture, with perthite grains having highly sutured margins. These sutures become quite complex so that neighbouring crystals are interlocked, and early stages in the development of swapped rims (Voll 1960) can be seen. At the margins of most perthite grains, especially where incipient swapped rims are forming, the perthite lamellae peter out and a single-phase zone occurs; this also happens around some of the inclusions within the perthites. Both the highly sutures margins and the marginal zones indicate that the perthosites have undergone some recrystallization during waning stages of the regional metamorphism.

The basic dykes vary greatly in grain-size but the majority are fine-grained and ophitic textures are common. Some are non-porphyritic whereas others have phenocrysts of zoned plagioclase. Augite, which also sometimes forms phenocrysts, commonly contains opaque schiller lamellae and shows incipient alteration to brownish hornblende and biotite.

The nepheline syenite pegmatites consist primarily of alkali feldspar and nepheline with sporadic sodic plagioclase and mafic minerals such as hornblende and biotite. They have been extensively sheared, and various stages through mortar textures up to a true mylonite can be observed. Nepheline was the mineral least resistant to shearing, and lenses of sericitized nepheline streaked out along the foliation are common.

Table 5. Chemical analyses of the Husfjord metagabbro

Major elements	1 36/1A	2 36/3A	3 36/54A	4 36/11F	5 36/11G	6 11/64A	7 60/47H	8 36/118F	9 36/225A
SiO_2	48.35	47.50	48.20	46.93	47.54	46.40	46.47	50.93	49.56
TiO_2	2.31	2.42	1.72	2.32	2.10	2.35	3.45	2.05	2.28
Al_2O_3	16.85	15.87	17.44	16.31	16.69	16.94	16.30	18.21	16.27
Fe_2O_3	2.21	2.64	1.53	2.30	2.87	2.36	1.02	3.35	3.67
FeO	8.52	8.70	8.49	9.07	7.09	8.03	15.16	7.28	8.88
MnO	0.11	0.11	0.10	0.11	0.09	0.09	0.17	0.09	0.11
MgO	6.34	6.53	6.94	7.52	7.28	6.04	3.12	4.24	6.48
CaO	10.19	10.98	10.84	9.30	9.86	11.87	7.27	6.44	6.86
Na_2O	3.40	3.02	2.74	2.12	2.90	3.23	3.40	3.29	2.34
K_2O	1.24	0.73	0.92	2.45	1.13	0.71	1.20	2.43	2.27
P_2O_5	0.15	0.17	–	0.15	0.14	0.16	0.29	0.10	0.08
H_2O^+	0.71	0.68	0.56	0.67	0.95	0.59	0.81	1.03	2.33
H_2O^-	0.11	0.07	0.08	0.10	0.29	0.17	0.19	0.12	0.19
	100.49	99.42	99.56	99.35	99.74	98.94	98.85	99.56	101.32
CIPW norms									
q	–	–	–	–	–	–	–	–	0.07
or	7.33	4.20	5.38	14.07	6.50	4.20	7.09	14.07	13.42
ab	23.96	23.71	22.05	15.79	24.02	19.45	28.77	28.01	19.80
an	27.05	26.42	31.58	26.71	28.13	29.63	25.67	26.77	27.19
ne	2.61	1.09	0.71	1.21	0.42	4.27	–	–	–
di	18.37	20.75	17.27	13.98	15.06	22.97	7.27	3.13	5.15
hy	–	–	–	–	–	–	4.09	14.41	23.34
ol	12.41	11.52	14.89	16.49	14.08	9.41	16.25	1.30	–
mt	3.20	3.83	2.22	3.33	4.16	3.42	1.48	4.84	5.32
il	4.39	4.60	3.27	4.41	3.99	4.46	6.55	3.89	4.33
ap	0.35	0.39	–	0.35	0.32	0.37	0.67	0.23	0.19

1–3, fine-grained facies; 4–6, coarse-grained facies; 7, noritic facies; 8–9, feldspathized metagabbro
Analyst: the writer, by XRF

C. Chemistry

1. HUSFJORD METAGABBRO

Chemical analyses of the Husfjord metagabbro are presented in Table 5 and it can be seen that, except for the contaminated noritic facies and the two slightly feldspathized specimens, they are nepheline-normative. Total alkalies are plotted against silica in Fig. 17A, showing that the metagabbro falls in the alkaline field of Irvine & Baragar (1971). The two specimens of feldspathized metagabbro, however, have slightly higher values of silica and alkalies.

In northern Sørøy there are two gabbro sheets, the Storelv and Breivikbotn gabbros, which have several field, mineralogical and chemical characteristics similar to the Husfjord metagabbro (Stumpfl & Sturt 1965, Sturt & Taylor 1972). These are plotted on Fig. 17A for comparison with the Husfjord metagabbro; although they have a greater scatter for points they too fall mainly in the alkaline field.

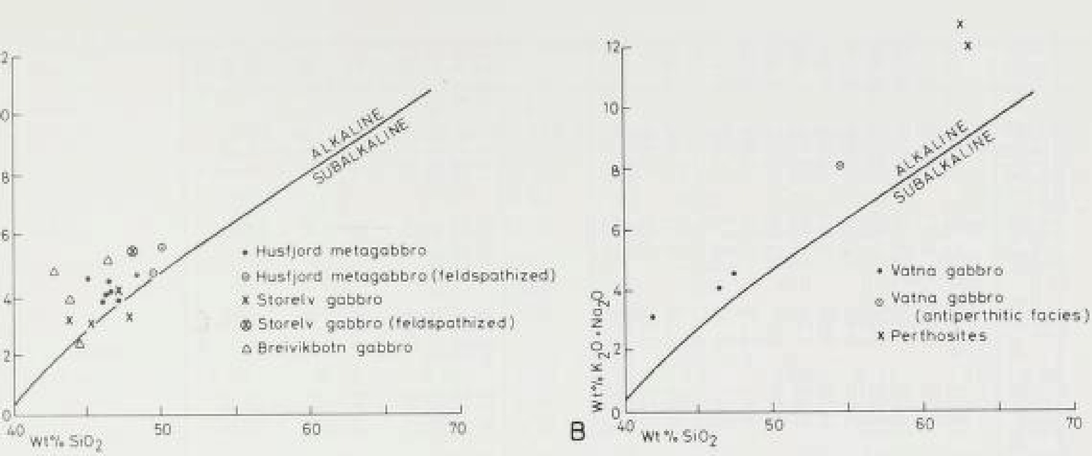

Fig. 17. Plot of total alkalies v. silica. A. Husfjord metagabbro, and the Storelv and Breivikbotn gabbros (data from Stumpfl & Sturt 1965). B. Vatna gabbro and perthosites. Alkaline/subalkaline boundary from Irvine & Baragar (1971).

Table 6. Chemical analyses of the Vatna gabbro and perthosites

Major elements	1 11/29I	2 07/25D	3 07/25G	4 11/38K	5 11/1C	6 11/25A
SiO_2	47.58	46.78	42.67	54.55	63.46	63.26
TiO_2	1.78	2.93	4.81	1.91	0.26	0.44
Al_2O_3	22.17	16.66	15.31	17.57	18.66	19.10
Fe_2O_3	1.56	2.62	1.45	1.80	0.56	0.47
FeO	6.18	8.64	12.41	8.30	1.31	1.53
MnO	0.08	0.12	0.12	0.12	0:02	0.03
MgO	4.08	6.42	7.21	1.67	0.84	0.06
CaO	11.52	11.30	11.17	5.18	1.41	1.44
Na_2O	4.20	3.46	2.73	6.01	6.55	7.42
K_2O	0.28	0.64	0.34	2.01	5.39	5.30
P_2O_5	0.18	0.16	0.42	0.09	–	–
H_2O^+	0.55	0.61	0.70	0.58	0.98	0.15
H_2O^-	0.57	0.12	0.16	–	0.21	0.16
	100.24	100.36	99.50	99.79	99.65	99.36

CIPW norms						
or	1.65	3.78	1.95	11.88	31.56	30.97
ab	24.28	21.73	9.70	47.43	55.59	54.03
an	40.82	28.04	25.23	15.03	5.19	2.57
ne	6.10	4.09	7.30	1.86	–	4.55
di	12.41	21.91	22.47	8.01	1.44	3.60
wo	–	–	–	–	–	0.16
hy	–	–	–	–	1.68	–
ol	8.29	10.35	15.25	7.96	0.93	–
mt	2.26	3.80	2.10	2.61	0.81	0.68
il	3.38	5.56	9.14	3.63	0.49	0.84
ap	0.42	0.37	0.97	0.21	–	–

1–3, Vatna gabbro; 4, Vatna gabbro, antiperthitic facies; 5–6, perthosite.
Analyst: the writer, by XRF

Table 7. Chemical analyses of the diorite – monzonite – quartz-syenite suite

Major elements	1 36/99A	2 36/116A	3 36/117C	4 36/233C	5 36/237A	6 38/1A	7 38/18B	8 38/38A	9 38/13A	10 38/31A	11 38/56C	12 46/4A	13 48/8B	14 48/29B	15 03/15C
SiO_2	54.60	48.69	52.77	51.55	55.31	56.54	58.32	56.96	69.51	63.44	59.52	63.91	65.35	63.39	67.06
TiO_2	1.16	2.27	1.89	2.35	1.44	1.38	1.09	1.36	0.53	0.81	1.27	0.95	0.73	0.90	0.57
Al_2O_3	18.78	17.61	19.90	19.40	18.56	18.40	18.88	17.45	15.94	17.34	16.23	16.61	18.02	18.23	16.38
Fe_2O_3	1.32	2.25	1.26	1.18	1.31	1.91	1.20	1.31	0.35	1.18	2.09	0.87	0.36	0.58	0.44
FeO	7.27	8.11	8.59	10.46	6.76	4.94	4.86	6.59	2.23	3.35	5.60	4.47	2.64	2.95	2.93
MnO	0.11	0.15	0.15	0.17	0.11	0.08	0.07	0.11	0.03	0.06	0.10	0.05	0.02	0.04	0.05
MgO	3.17	5.29	3.23	2.24	1.61	2.48	1.55	2.24	0.92	1.49	2.41	1.47	1.37	1.52	1.08
CaO	6.91	11.34	7.25	5.81	5.00	4.93	4.15	4.17	1.94	1.34	2.78	2.05	2.64	2.17	1.65
Na_2O	4.28	2.39	2.97	2.53	4.20	3.79	3.94	3.68	2.76	3.15	3.41	2.75	3.78	2.89	2.58
K_2O	1.47	1.11	2.08	3.87	4.22	4.42	5.19	4.89	5.33	6.51	5.13	6.32	5.12	6.67	6.17
P_2O_5	0.09	0.14	–	–	–	0.06	0.05	0.06	–	–	–	0.02	–	–	0.01
H_2O^+	0.45	0.76	0.39	0.58	0.60	0.95	0.50	0.69	0.71	0.96	0.61	0.18	0.71	1.04	0.85
H_2O	0.11	0.14	0.11	0.08	0.14	0.12	0.17	0.13	0.04	0.12	0.21	0.09	0.06	0.19	0.17
	99.72	100.25	100.36	100.22	99.26	100.00	99.97	99.64	100.29	99.75	100.36	99.74	100.80	100.57	99.94
CIPW norms															
q	0.79	–	–	–	–	5.99	1.42	1.90	21.55	7.89	8.12	12.38	18.89	13.97	20.53
or	7.98	6.74	12.29	19.15	20.86	17.08	27.60	29.37	31.14	37.76	27.42	33.39	19.86	28.84	34.10
ab	36.22	22.76	30.46	25.64	34.82	32.07	31.14	33.34	28.52	33.09	28.85	27.50	31.99	29.70	25.55
an	28.05	30.48	32.00	26.49	19.43	22.18	15.23	17.41	9.33	6.70	13.10	9.69	13.10	10.77	7.92
ne	–	–	–	–	0.39	–	–	–	–	–	–	–	–	–	–
cor	–	–	–	0.06	–	0.26	–	–	0.48	0.99	0.44	0.98	2.00	2.71	2.26
di	2.35	15.75	3.31	–	3.05	–	3.09	1.15	–	–	–	–	–	–	–
hy	16.83	2.00	4.92	7.68	–	11.14	12.64	9.04	5.65	7.59	12.35	9.99	6.78	7.71	7.04
ol	–	–	10.84	9.23	8.14	–	–	–	–	–	–	–	–	–	–
mt	1.91	3.26	1.83	1.71	1.87	2.70	1.86	1.64	0.51	1.55	2.91	1.15	0.52	0.84	0.64
il	2.20	4.31	3.59	5.49	2.73	2.62	2.58	2.07	0.93	1.50	2.41	1.79	1.39	1.69	1.08
ap	0.21	0.32	–	–	–	0.14	0.14	0.12	–	–	–	0.05	–	–	0.02

1–4, Havnefjord diorite; 5–8, monzonite; 9–15, quartz-syenite.
Analyst: the writer, by XRF

Table 8. Partial chemical analyses of mineral separates of coexisting feldspars in the diorite – monzonite – quartz-syenite suite

Alkali feldspar	1 36/117C	2 36/233C	3 36/237A	4 38/18B	5 38/38A	6 38/56C	7 38/13A	8 38/31A	9 46/4A	10 48/8B
CaO (wt %)	3.29	1.22	2.18	0.96	2.25	1.57	0.28	0.40	0.22	1.48
Na_2O	2.55	1.87	2.60	1.62	2.60	2.50	1.92	1.92	1.67	2.80
K_2O	8.55	13.48	10.20	11.52	9.68	8.80	13.10	12.00	14.60	8.37
Ba (ppm)	2300	4300	7250	4200	5550	2100	5100	3700	4550	9100
Rb	60	140	120	140	130	180	325	210	250	170
Sr	500	575	540	500	535	430	425	395	405	635
Plagioclase										
CaO (wt %)	9.04	9.12	8.19	7.40	8.26	4.50	2.65	4.14	3.44	3.85
Na_2O	5.84	6.15	6.50	6.41	6.47	5.17	3.57	5.06	4.81	5.45
K_2O	0.62	0.50	0.57	0.78	0.50	0.88	1.15	0.58	1.22	0.95
Ba (ppm)	530	200	140	530	95	590	510	100	440	570
Rb	14	–	2	11	2	24	57	14	22	32
Sr	650	580	540	500	580	420	210	280	280	460

1–2, Havnefjord diorite; 3–5, monzonite; 6–10, quartz-syenite.
Analyst: the writer, by XRF

Robins & Gardner (1974, 1975) have divided the Seiland petrographic province into subprovinces, in which there is a general trend from early subalkaline gabbros through transitional rocks to later alkaline gabbros. The Breivikbotn, Storelv and Husfjord gabbros are included in the pre-D_2 subalkaline subprovince by Robins & Gardner. The chemical data, however, suggest that these early gabbros are fairly alkaline, lying chemically (but not temporally) between the pre-D_2 tholeiitic Hasvik gabbro of southwestern Sørøy and the late-D_2 Rognsund clinopyroxene-gabbro on Seiland (Robins & Gardner 1974).

2. VATNA GABBRO

Chemical analyses of the Vatna gabbro and its associated perthosites are presented in Table 6. This gabbro is strongly nepheline-normative which, together with a plot of total alkalies against silica (Fig. 17B), indicates its alkaline nature. The antiperthitic facies plots in an intermediate position between the normal facies and the perthosites.

On Seiland a syn-D_2 syenogabbro with associated perthosites may be approximately contemporaneous with the Vatna gabbro, and the two gabbros have been placed in a transitional (high-K tholeiite) subprovince by Robins & Gardner (1975).

3. DIORITE – MONZONITE – QUARTZ-SYENITE SUITE

Whole rock analyses of the diorite – monzonite – quartz-syenite suite, for both major and trace elements, are presented in Table 7, and partial analyses of coexisting feldspars in this suite in Table 8. The analyses were carried out by the author using X-ray fluorescence (and classical wet methods for the determination of FeO and H_2O). The chemistry of this suite suggests that these rocks may form a petrogenetic series, as discussed below.

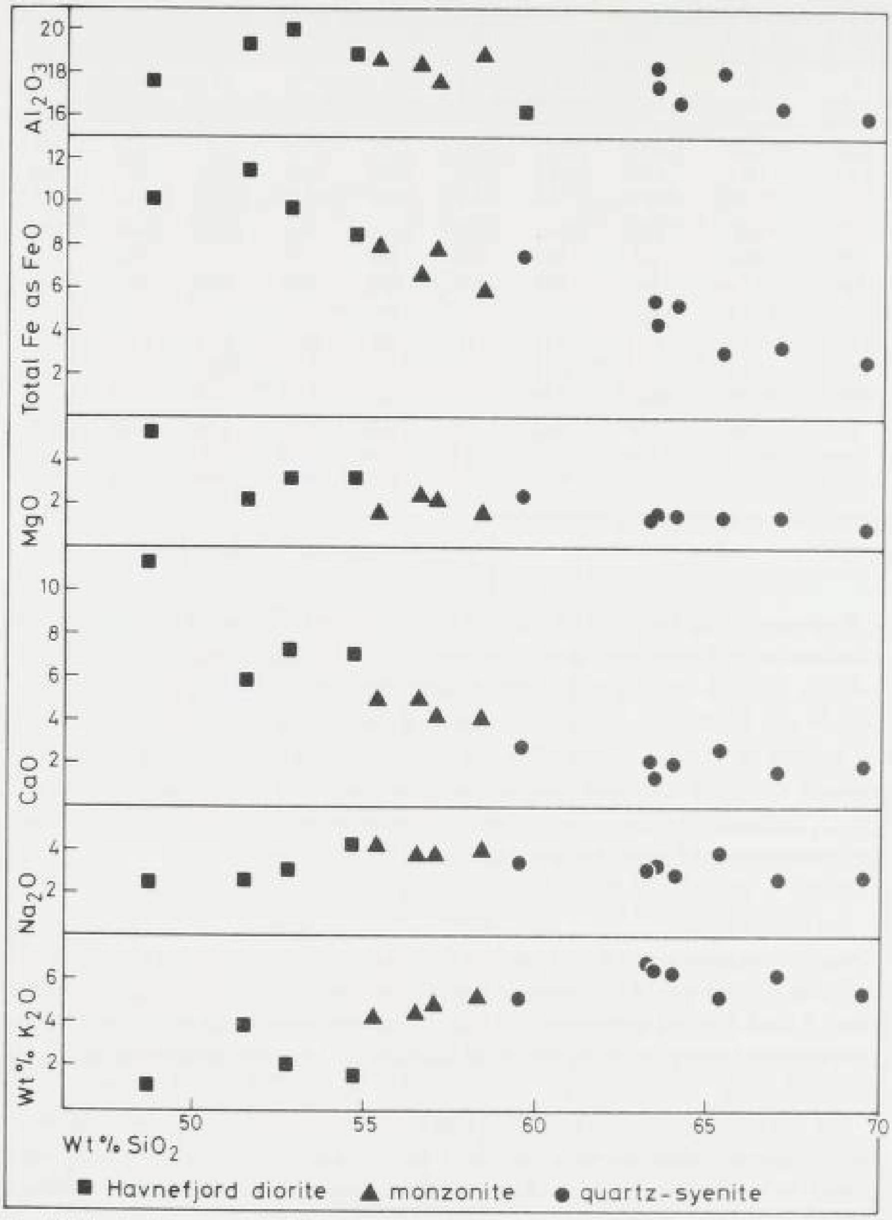

Fig. 18. Harker diagram of the diorite – monzonite – quartz-syenite suite.

Whole rock analyses

The chemical data for the diorite – monzonite – quartz-syenite suite are presented on a Harker diagram (Fig. 18), and various triangular diagrams (Fig. 19). These show a progressive trend from fairly basic Havnefjord diorite, through monzonite, to relatively acid quartz-syenite. MgO, FeO and CaO decrease in absolute amounts (Fig. 18), but remain in fairly constant proportion with respect to each other (Figs.

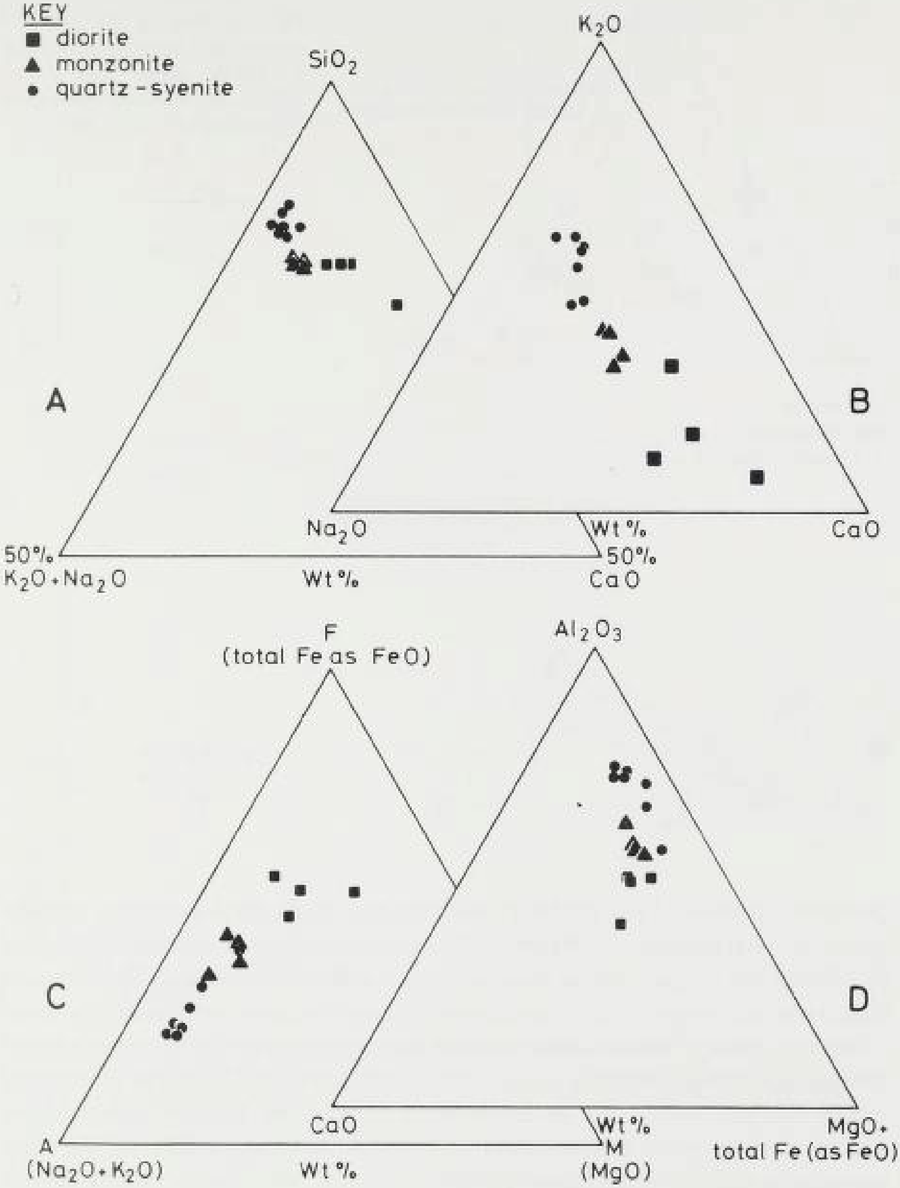

Fig. 19. Plots of the diorite – monzonite – quartz-syenite suite. A. System SiO_2 – (K_2O + Na_2O) – CaO. B. System K_2O – Na_2O – CaO. C. AFM diagram. D. System Al_2O_3 – CaO – (MgO + total Fe).

19C & D) while there is a corresponding increase in SiO_2 and the alkalies (Fig. 19A), with the K/Na ratio progressively increasing (Fig. 19B). The AFM diagram (Fig. 19C) shows that as alkali enrichment takes place the Fe/Mg ratio remains constant at about 3, with no Fe-enrichment during differentation. Although care must be taken while interptreting trends in AFM diagrams (Robinson & Leake 1975), all these various trends suggest that this suite of rocks may form a

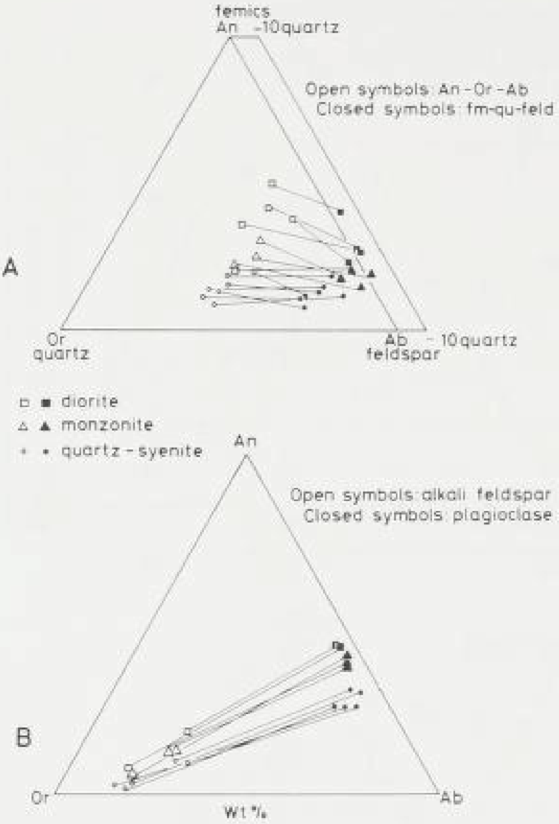

Fig. 20. Plots of the diorite – monzonite – quartz-syenite suite. A. CIPW norms after Larsen (1938). B. Coexisting feldspars in the system An – Or – Ab.

petrogenetic series. Field evidence confirms that both the monzonites and the quartz-syenites postdate the Havnefjord diorite, but there is no evidence in the field to indicate the relative ages of the monzonites and quartz-syenites. The chemical data, however, suggest that the quartz-syenites are the latest members of the suite.

An expression of the saturation of the series is given by the CIPW norms plotted on a triangular diagram (after Larsen 1938) presented in Fig. 20A. The Havnefjord diorite lies just within the undersaturated field in the femics-quartz-feldspar system showing that it is quite a silica-poor diorite, but later members of the series are entirely within the oversaturated field.

Feldspar analyses

The whole rock trend away from calcium towards the alkalies is reflected in the major elements in coexisting feldspars in the diorite – monzonite – quartz-syenite suite (Fig. 20B).

The variations of trace elements with respect to each other in alkali feldspars are sympathetic with those in the corresponding coexisting plagioclase, and these trends are presented by plotting total concentrations of Ba, Rb and Sr in the coexisting feldspars against each other (Figs. 21A–C). During evolution of the series absolute amounts of Ba and Rb increase while Sr decreases, and there is an

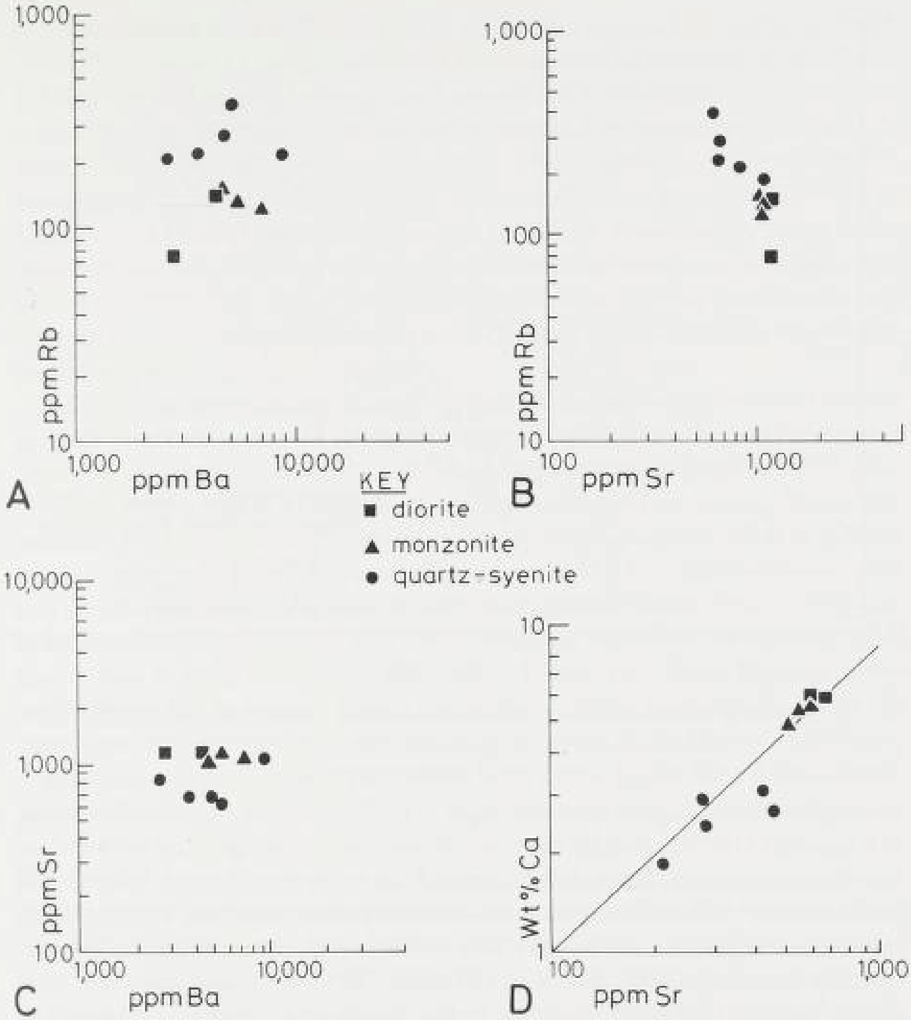

Fig. 21. Graphs of the diorite – monzonite – quartz-syenite suite. A. Total Rb v. total Ba in coexisting feldspars. B. Total Rb v. total Sr in coexisting feldspars. C. Total Sr v. total Ba in coexisting feldspars. D. Ca v. Sr in plagioclase. Squares – diorite, triangles – monzonite, circles – quartz-syenite.

increase in the radios Rb/Sr, Ba/Sr and Rb/Ba, indicating that Rb is concentrated in the later members and Sr in the earlier rocks.

Similar trends are reported for the Rb/Sr ratio by Sen et al. (1959) and Taylor et al. (1968), and for the Rb/Ba ratio by Herz & Dutra (1966) and Taylor et al. (1968), and are due to the incorporation of Rb in late phases because of its large ionic radius and low charge. The Ba/Sr ratio is more problematical in that El Bouseily & El Sokkary (1975) describe an increase in this ratio with fractionation, while others report decreases (Heier & Taylor 1959, Taylor & Heier 1960, Heier 1962). The latter authors point out that Ba has a slightly lower electronegativity than Sr and forms a more ionic bond, apparently favouring its entry into lattice sites before Sr. Nockolds & Allen (1953) on the other hand maintain that Ba is

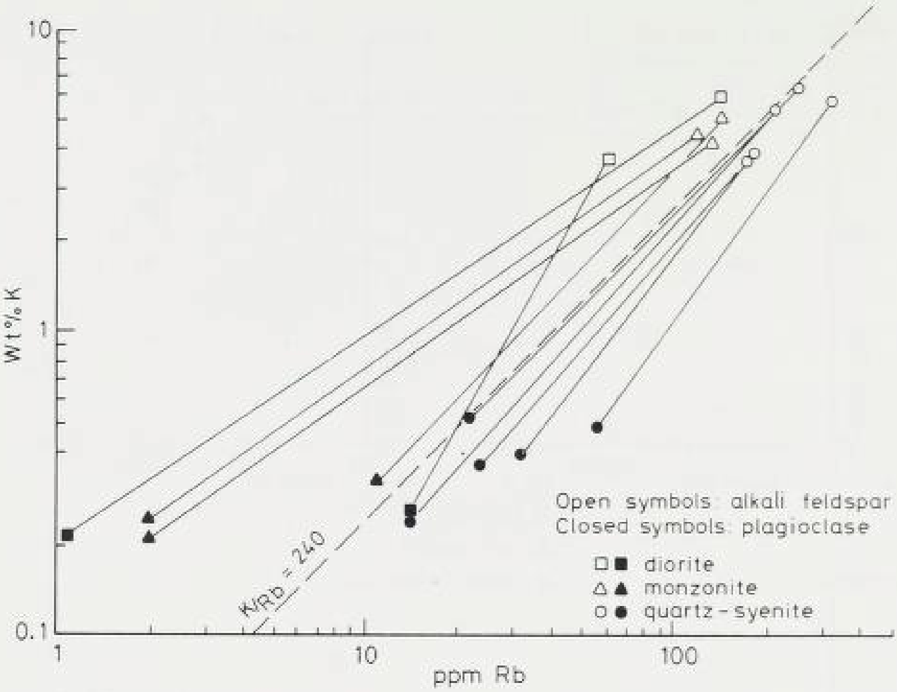

Fig. 22. Graph of K v. Rb in coexisting feldspars in the diorite – monzonite – quartz-syenite suite.

not depleted in the magma until late stages in a differentiation sequence. According to Ringwood (1955), provided that the difference in electronegativity between two ions does not exceed 1, Goldschmidt's rules of diadochy should apply. In the presnt case it seems that the difference in electronegativity between Ba and Sr is insufficient to prevent the smaller ionic radius of Sr from taking Precedence, ensuring early entry for Sr. According to Smith (1974), Sr tends in general to enter plagioclase while Ba tends to remain in the crystallizing liquid. Sr replaces K in alkali feldspar as well as ca in plagioclase, whereas Ba only significantly replaces K in alkali feldspar, and thus Sr has affinities for plagioclas and Ba for alkali feldspar (Sen 1960). Therefore in a suite of rocks as those under discussion, in which alkali feldspar becomes progressively more dominant with evolution, an increase in the Ba/Sr ratio is to be expected. This is supported by the fact that in granitic rocks there is a weak positive correlation between Ba/Sr ond Or content, and that syenitic rocks in alkali feldspar have unusually high Ba/Sr ratios (Smith 1974).

Each of the trace elements Ba, Rb and Sr have been plotted against K, Na and Ca. The most significant trends are the behaviour of Sr in plagioclase, and variations in K/Rb ratios in coexisting feldspars. With differentation of the diorite – monzonite – quartz-syenite suite absolute values of Sr and Ca in plagioclase fall, with the Ca/Sr ratio remaining almost constant at about 100 (Fig. 21D). The constant, linear relationship between these elements has been commented on by Herz & Dutra (1966), and the value of the average partition coefficient for the

Husfjord rocks is close to the 120 for granodiorites reported by Kolbe & Taylor (1966). In basic rocks the Sr content of plagioclase rises as Ca content falls (Butler & Skiba 1962), while Sr has its greatest abundance in plagioclase in intermediate rocks (Wager & Mitchell 1951), and the Sr content falls sympathetically with Ca in acidic rocks (Sen et al. 1959, Hall 1967). This is because calcic plagioclase is less efficient than sodic plagioclase at extracting Sr from the melt (Korringa & Noble 1971). Another factor is that acid rocks contain more alkali feldspar which removes from the magma some of the Sr which would otherwise go into plagioclase (Sen et al. 1959, Hall 1967, El Bouseily & El Sokkary 1975). This accords well with the trends of the Husfjord suite, in which alkali feldspar plays an increasingly important part during differentation.

K/Rb ratios for coexisting feldspars in the diorite – monzonite – quartz-syenite suite are presented in Fig. 22. There is an enrichment in Rb both in absolute terms and with respect to K with differentiation, the series having an average K/Rb ratio of about 240, a figure in agreement with the average K/Rb ratio for normal igneous rocks (Taylor et al. 1956), although the possible range of values in igneous rocks is quite wide (Herz & Dutra 1966, Shaw 1968). In a differentiation series it is common for the K/Rb ratio to drop during fractionation (Smith 1974) due to the late incorporation of Rb into the K lattice site because of the relatively large ionic radius of Rb (Nockolds & Allen 1953, Taylor et al. 1956, Taylor & Heier 1960, Heier 1962, Herz & Dutra 1966, Shaw 1968, Taylor et al. 1968). In agreement with this principle, the coexisting feldspars in the diorite – monzonite – quartz-syenite suite show a progressive drop in the K/Rb ratio with differentiation.

Summary

In summary it can be stated that the chemical trends exhibited by both the whole rock and the feldspar analyses indicate that the diorite – monzonite – quartz-syenite suite forms a petrogenetic series. This suite was emplaced just after the peak of the regional metamorphism and its petrogenesis is associated with that event, as discussed below.

D. Petrogenesis

The members of the Husfjord igneous complex span a time interval from late-D_1 (Husfjord metagabbro) to syn-D_2 (Vatna gabbro and its associated alkaline rocks). During this interval the Finnmarkian orogeny reaches its structural and metamorphic peak, and the complicated igneous history is closely related to the development of the orogen.

In this part of the Seiland province, intrusive igneous activity evolved from low-K tholeiitic magmas through high-K calc-alkali and possibly transitional high-K basaltic magmas to alkali olivine basalt, and finally highly differentiated alkali magmas and carbonatites (Robins & Gardner 1975). This general chemical trend in the development of the Seiland igneous province is linked by Robins &

Gardner to a long-lived mantle diapir complex in the asthenosphere above a subduction zone dipping eastwards beneath the Baltic plate, and becoming progressively steeper with time. This mantle diapir would have been the source of the essentially basic and ultrabasic Seiland plutonic complex, the upper part of which has been displaced to the southeast by nappes which decapitated the mantle diapir (Ramsay 1973). The root of the diapir may underlie S. W. Sørøy where there is a high positive gravity anomaly approaching 100 mgal (Brooks 1970). This positive anomaly, indicating the presence of mafic and ultramafic rocks, is thought to continue southwestwards along the coast of Lofoten (Brooks 1970).

The Husfjord metagabbro is a product of the first phase of magmatic activity associated with the mantle diapir, late in D_1 Field, mineralogical and chemical characteristics of the Storelv and Breivikbotn gabbros of Sørøy (Stumpfl & Sturt 1965, Sturt & Taylor 1972) indicate that these are probably contemporaneous with the Husfjord metagabbro. The later Vatna gabbro and its associated syenitic rocks, however, are likely to be coeval with the syn-D_2 syenogabbros of Seiland (Robins & Gardner 1975).

The emplacement of the mantle diapir into the crust would have caused considerable heating of the metasedimentary pile. It is significant that the peak of the regional metamorphism was closely followed by the emplacement of the diorite – monzonite – quartz-syenite complex, and this suite was probably generated from lower crustal material fused by the heat of metamorphism and the mafic mantle diapir.

Gastil (1975) considers that magmas from the mantle intruded into the crust overlying a subduction zone could cause the generation of welts of tonalitic magma, from which plutons might arise. Dioritic magmas generated by anatexis of crustal rocks would possibly be in the form of crystal mushes under conditions of normal regional metamorphism, but would contain a higher proportion of liquid if additional heat is introduced by the emplacement of basic or ultrabasic magmas in the region (Wyllie 1977). In the Husfjord area considerable additional heat was introduced by the rising mantle diapir which was the parent of the abundant basic and ultrabasic plutons of the Seiland province. The volume of melt which can form by anatexis is also dependent upon the quantity of water available, which is essentially controlled by the amount of hydrous minerals, such as biotite and amphibole, in the crustal rocks (Fyfe 1973); these minerals are abundant in the schists of the Sørøy succession. One of the products of the breakdown of biotite on melting of paragneisses can be hypersthene (Büsch et al. 1974), which is the main mafic mineral in the Havnefjord diorite and monzonites, and the extensive breakdown of biotite to form pyroxene lowers the solidus temperature of the gneiss (Büsch et al. 1974). The intrusive hypersthene-tonalites in the Rio de Janeiro region are thought by Leonardos & Fyfe (1974) to represent extreme products of progressive melting of lower crustal rocks. Hoschek (1976) found that a gneissic assemblage of biotite,plagioclase and quartz began melting between 650° and 725°C at 4 kb P_{H_2O}, according to the composition of the plagioclase, although these temperatures would be lower at higher P_{H_2O}, and Lappin & Hollister (1980) produced a tonalitic melt by melting a hornblende-plagioclase-biotite-quartz gneiss between 675° and 750°C at 6–8 kb P_{H_2O}.

Wyllie (1977) suggests that dioritic plutons can result from the more refractory components during crustal anatexis, the more siliceous and alkali-rich volatile-bearing fractions producing smaller bodies of granitic rocks at higher levels. Studies of the system Ab-An-Or-SiO_2-H_2O by Presnall & Bateman (1973) confirm that it is possible to produce a suite of granitic rocks, as in the Sierra Nevada batholith, by fractional crystallization of a parental dioritic magma produced by equilibrium fusion at the base of the crust (Presnall 1979). The Husfjord monzonites and quartz-syenites are small, late, coarse-grained bodies produced by fractionation of the dioritic magma, and developed as late-stage pegmatitic fluid-rich phases, probably due to vapour saturation at points where crystallization of the diorite was more advanced (Whitney 1975). This could explain their common occurrence alongside the larger metasedimentary rafts in the diorite where cooling and slight contamination would promote crystallization. The small diffuse xenoliths in the diorite and monzonites may represent relict metasedimentary material which had resisted melting (Presnall & Bateman 1973, White & Chappell 1977). Experimental work on the fusion of sediments shows that the melting curve of shales is only about 20°C higher than the minimum melting curve of granite in the presence of water, and assuming a geothermal gradient of about 30°C/km, shales would melt in the depth range of 20–25 km (Wyllie & Tuttle 1960, 1961). It is deduced from the mineral paragenses of the Husfjord rocks that the approximate depth of the regional metamorphism at its peak was greater than 20 km, and the temperature in excess of about 620°C; temperatures at greater depths where the dioritic magma developed would have been higher.

If shales are completely melted they would produce magmas of intermediate composition but which have different chemical characteristics from melts produced from igneous parents by fractionation (Wyllie & Tuttle 1961). The main difference is that rocks formed from fused shales are richer in aluminium, and this can sometimes cause the crystallization of aluminous phases (Wyllie & Tuttle 1961). Members of the diorite – monzonite – quartz-syenite suite are plotted on Fig. 23, and compared with average diorites and monzonites (Nockolds 1954) and with the Plauen quartz-syenite (Johannsen 1932). Each group of rocks from Husfjord is, on the whole, more aluminous than its equivalent from the literature, suggesting that they may have originated from the melting of metasediments. This relatively high aluminium content may account for the presence of occasional garnets in some of the quartz-syenites, for the occurrence of garnet in granitic rocks is considered by Green (1976) to suggest an origin by equilibrium fusion of pelites. The fact that garnet rather than cordierite has formed indicates that the depth of generation of the magma from metasedimentary material was in the order of 25 km or more (Green 1976), although the presence of Mn can lower this a little Green 1977). The average of 235 shales (Pettijohn 1949) is also plotted on Fig. 23, and this falls very close to the average composition of the whole diorite – monzonite – quartz-syenite series, which must represent the bulk composition of the parent material of the suite. Thus the diorite – monzonite – quartz-syenite

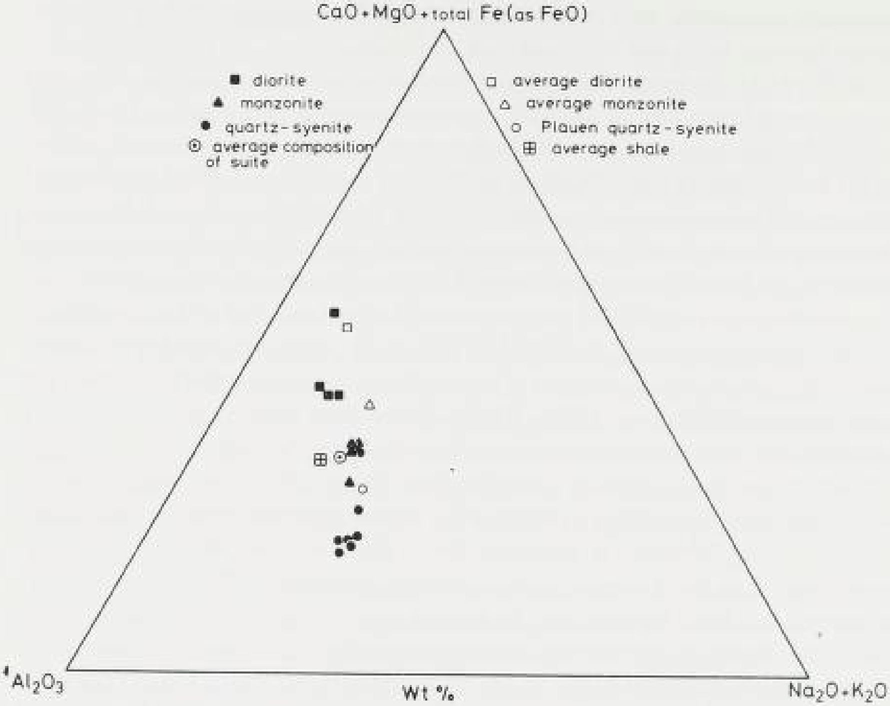

Fig. 23. Composition of the diorite – monzonite – quartz-syenite suite compared with average igneous rocks and shales. Average of 50 diorites and 46 monzonites from Nockolds (1954), Plauen quartz-syenite from Johannsen (1932), and average of 235 shales from Pettijohn (1949).

series could have been generated by fusion of metasediments at about 25 km depth in the crust at the peak of the regional metamorphism, with additional heat being provided by the rising mantle diapir above the subduction zone.

III. Metamorphism

The rocks of the Husfjord area have undergone two major episodes of Finnmarkian deformation, D_1 and D_2; the regional metamorphism began during D_1, increased to its peak between D_1 and D_2, and waned during D_2. Thus the character of the metamorphism varied from being syntectonic during the more intense deformation phases, to essentially static between D_1 and D_2, and superimposed upon this regional metamorphism were thermal metamorphic effects of some of the igneous intrusions. These metamorphic phases are discussed below chronologically, and the relationships between the metamorphic, tectonic and igneous events are summarized in Fig. 24.

The earliest D_1 fabric recorded is a weakly developed schistosity, which is folded around later isoclinal D_1 minor folds. The mineralogy and textures formed in the metasediments during the syn-D_1 metamorphism have been destroyed in places

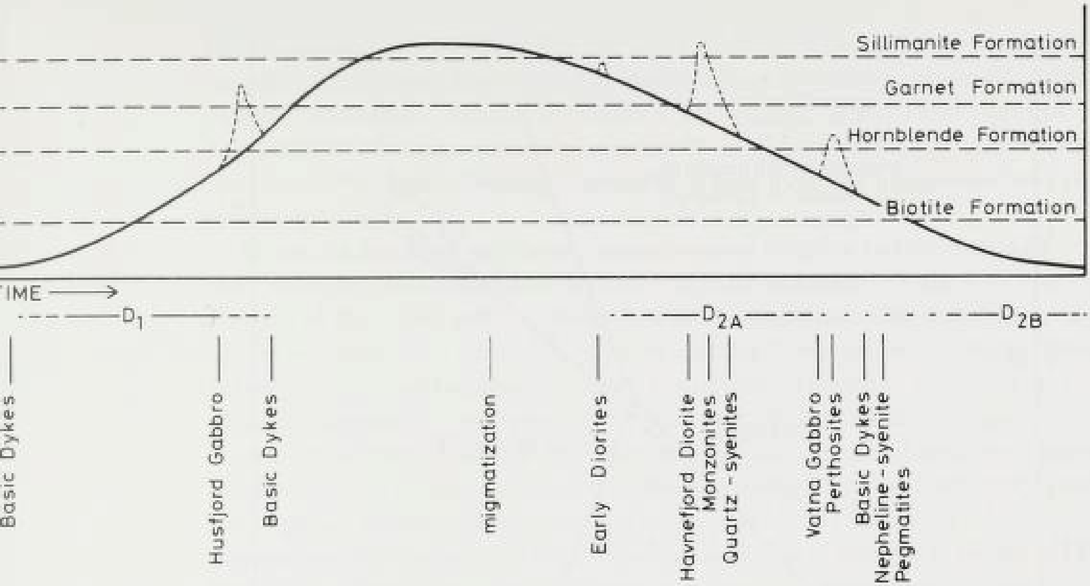

Fig. 24. Diagrammatic summary of the metamorphic, tectonic and igneous events in the Husfjord area during the Caledonian (Finnmarkian) orogeny. Solid temperature curve represents regional metamorphism; dotted temperature curves represent thermal metamorphisms superimposed upon regional metamorphism.

by thermal metamorphism by the Husfjord metagabbro, and in many cases the hornfelsic texture has been overgrown by a later regional metamorphic fabric.

The regional metamorphism reached its peak in the sillimanite-almandine subfacies of the almandine-amphibolite facies (Turner & Verhoogen 1960), between the two major episodes of deformation. This period of static recrystallization in the country rocks is characterized by development of porphyroblasts of garnet, kyanite and sillimanite. Garnet overgrows the D_1 schistosity, and is occasionally augened by the D_2 foliation; in some cases rotation has occurred. Fibrolite has nucleated on some garnets during progressive metamorphism, while other later garnets have formed retrogressively from kyanite and biotite. Kyanite and sillimanite appear to coexist stably in many rocks, suggesting that the prevailing conditions were near those at the kyanite/sillimanite boundary, although reaction between the polymorphs was sluggish. In the Husfjord metagabbro clinopyroxene has altered to green hornblende, with a concomitant decrease in the An-content of the plagioclase. The psammites and semipelites of the country rocks in the southwest have undergone migmatization associated with the peak of the regional metamorphism. In the most intensely migmatized parts fibrolitic sillimanite sometimes occurs along alkali feldspar grain boundaries (Fig. 10), and appears to be related to the migmatization. These fibrolites could either form from ions migrating along grain boundaries, or from constituents which were derived from the neighbouring crystals; the latter seems more likely for three reasons. First, fibrolite is essentially restricted to feldspar/feldspar contacts; second, fibrolite needles commonly penetrate into the neighbouring grains; and third,

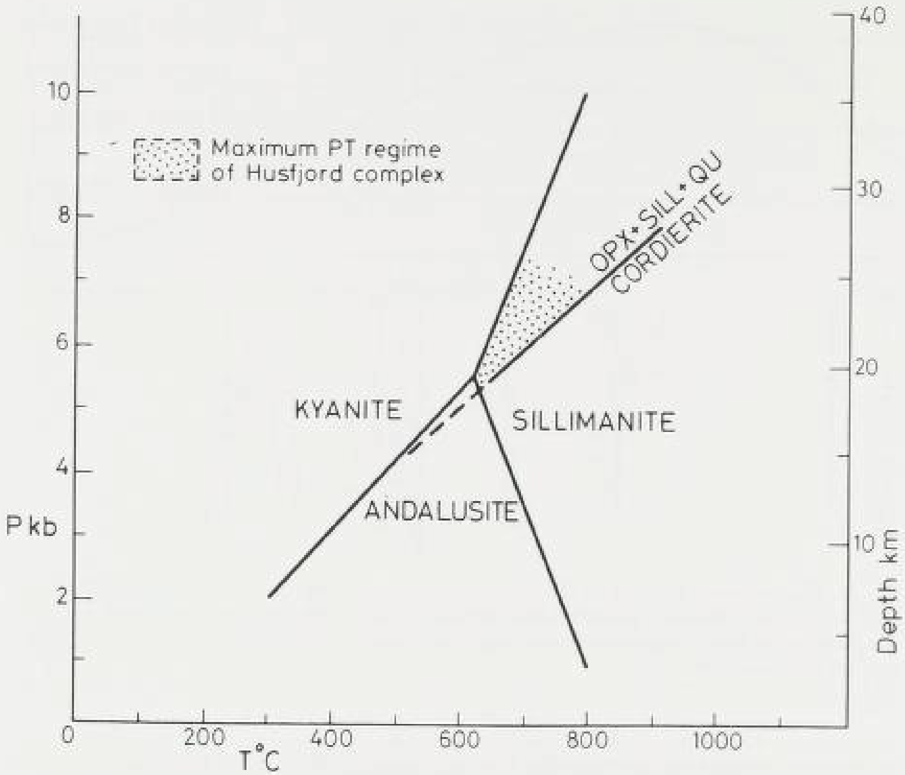

Fig. 25. Stability fields of Mg-cordierite (Newton et al. 1974) and the Al_2SiO_5 polymorphs (Richardson et al. 1969). Depth scale after Wyllie (1979).

fibrolite is occasionally involved in myrmekitic intergrowths which have formed between alkali feldspar and plagioclase. Sturt (1970) has described the development of fine-grained sillimanite at feldspar/feldspar boundaries, usually associated with myrmekite, in the aureole of the syn-orogenic Hasvik gabbro, in southwestern Sørøy. He proposed that these formed from exsolution of excess Si and Al from the feldspars, and suggested that the location of sillimanite nucleation at certain sites may be influenced by the level of thermal stress set up in the minerals during metamorphism.

After the Husfjord metagabbro had been amphibolitized during regional metamorphism, it was locally thermally metamorphosed by the early pyroxene-mica diorites. The diorites themselves have not been amphibolitized, and clearly postdate the peak of the regional metamorphism. The Havnefjord diorite has metamorphosed the Husfjord metagabbro up to the hornblende hornfels facies, and many of the metasedimentary rafts in the diorite have been hornfelsed at their margins, particularly the semipelites. In many cases, the D_2 schistosity has been obliterated, while in others it remains as a relict texture only discernible by the presence of biotite-rich bands. Regional metamorphic garnets have broken down to fibrolite, biotite and iron oxides, and the absence of new garnet indicates that the temperature of regional metamorphism had dropped beneath that of the garnet isograd by this time. The monzonite sheets thermally metamorphose the neighbouring Havnefjord diorite, in which a fine-grained granoblastic hornfels of hypersthene, clinopyroxene and plagioclase is developed. Hornfelsic textures also develop in the metasedimentary xenoliths within both the monzonites and the quartz-syenites.

While the diorite – monzonite – quartz-syenite complex and the Vatna gabbro were being emplaced the regional metamorphism continued to wane, so that by the end of D_2 it had reached greenschist facies conditions. The members of the diorite – monzonite – quartz-syenite suite and the Vatna gabbro, therefore, show a low grade of metamorphism, while retrogressive alteration is evident in the Husfjord metagabbro and some of the country rocks and metasedimentary rafts. In the calc-silicate schists of the country rocks the granular hornfelsic texture produced by the thermal effects of the Husfjord metagabbro has been overprinted by a regional metamorphic texture and mineralogy. In the Havnefjord diorite and monzonites, hypersthenes are sometimes in the process of altering to biotite where they make contact with alkali feldspar, and a similar phenomenon occurs in the early diorites within the Husfjord metagabbro. These biotites generally contain vermicular inclusions of quartz (Fig. 12); similar textures, in which biotite/quartz symplectite has developed at the expense of hypersthene, have been described by Sederholm (1916). The Vatna gabbro is only slightly metamorphosed with olivines scarcely altered and augites sometimes fringed by a little biotite. The perthosites show early stages in the development of swapped rims at the boundaries of perthite grains.

Discussion

The inter-relationships between the regional metamorphism and the thermal metamorphic effects of the intrusions show that the emplacement of the Husfjord plutonic igneous complex was protracted, and took place synchronously with metamorphic and structural events during the Finnmarkian phase of the Caledonian orogeny.

The regional metamorphic mineral parageneses indicate that pressures continued to be relatively high throughout the period of metamorphic recrystallization. The absence of andalusite and the coexistence of kyanite and sillimanite, the former sometimes in the process of altering to the latter, indicate that pressures were in excess of the aluminium silicate triple point at 5.5 kb (Richardson et al. 1969), representing a depth of about 20 km (Wyllie 1979), see Fig. 25.

Temperatures during the thermal metamorphism caused by the Havnefjord diorite, which was emplaced soon after the peak of the regional metamorphism into already hot rocks, must have exceeded 622°C (Richardson et al. 1969) since sillimanite was formed in the hornfelses. Although the composition of the rock may have had some control over the formation of cordierite, the absence of cordierite from these hornfelses is probably mainly due to high pressures. Fig. 25 shows the upper pressure stability limit of anhydrous Mg-cordierite (Newton et al. 1974), although this stability is lowered as Fe replaces Mg (Holdaway & Lee 1977). Cordierites in pelitic hornfelses usually contain a considerable percentage of the Fe molecule (Leake 1960), so the stability limit shown in Fig. 25 should be taken as a maximum for the present situation. At higher pressures, instead of cordierite, orthopyroxene, sillimanite and quartz would form (Newton et al. 1974), an assemblage which is present in the pyroxene-hornfels xenoliths within the Havnefjord diorite. This cordierite reaction curve forms an upper temperature limit within the sillimanite field for the thermal metamorphism ascribed to the intrusion of the Havnefjord diorite.

As deduced from the metamorphic mineral parageneses, Fig. 25 shows the approximate peak of the PT regime that must have prevailed in the environment of the Husfjord plutonic complex during its emplacement.

Acknowledgements. – The writer would like to express his grateful thanks to Prof. B.A. Sturt for his helpful advice during the course of the work, and for critically reading the manuscript. He would also like to thank Mrs. J. Rowland and Mrs. M. Rundle for typing the manuscript, and Miss J. Bees and Mrs. A. Gregory who draughted the diagrams. Thanks go to the technical staff at Bedford College in the University of London for technical assistance, particularly Mr. J. Keith and Mr. J. Holbrook, and at Bristol University, especially Mr. S. Waterman for printing the photographs.

The work was originally carried out with the support of a Natural Environment Research Council (formerly Science Research Council) research studentship, which the writer acknowledges with thanks.

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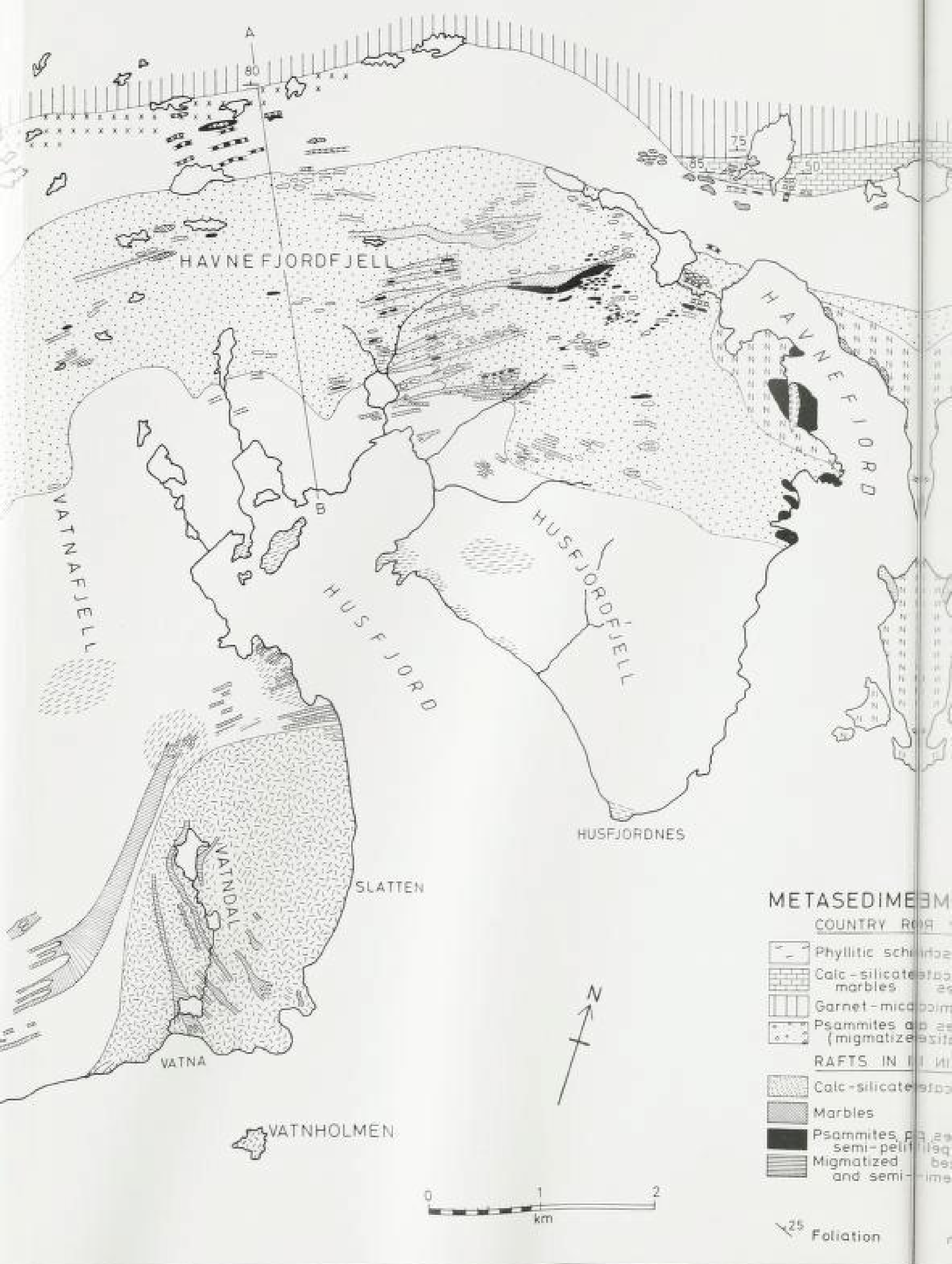
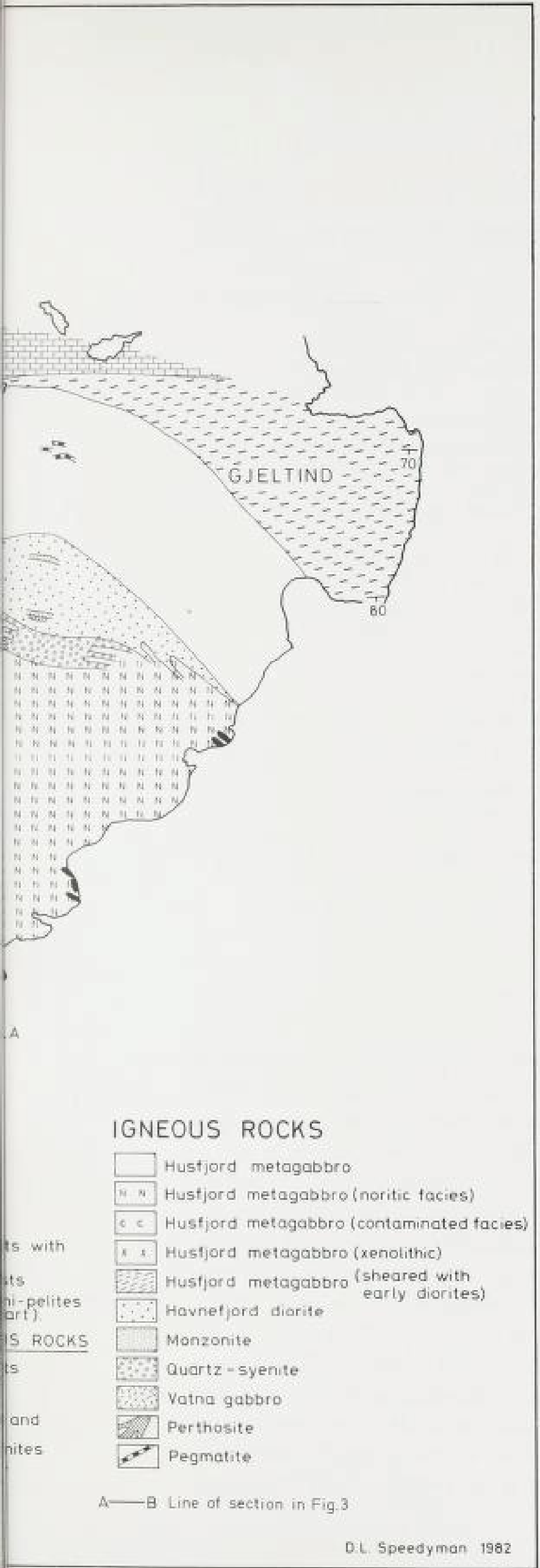
GEOLOGICAL M

MAP OF THE HUSFJORD AREA, SØRØY




GJELTIND
70
80
A
IGNEOUS ROCKS
Husfjord metagabbro
Husfjord metagabbro (noritic facies)
Husfjord metagabbro (contaminated facies)
Husfjord metagabbro (xenolithic)
Husfjord metagabbro (sheared with early diorites)
Havnefjord diorite
Monzonite
Quartz - syenite
Vatna gabbro
Perthosite
Pegmatite
A——B Line of section in Fig.3
D.L. Speedyman 1982

Amphibolites and Metagabbros from the Proterozoic Telemark Suite of Setesdalsheiene, South-central Norway

T. PRESTVIK & F.M. VOKES

Prestvik, T. & Vokes, F.M. 1982: Amphibolites and metagabbros from the Proterozoic Telemark Suite of Setesdalsheiene, south-central Norway. *Norges geol. Unders. 378*, 49–63.

Proterozoic supracrustal rocks of the Telemark Suite, including large thickness of basic metavolcanites, occur as a large-scale isolated outlier within the older gneiss-migmatite-granite complex on the mountain plateau of Setesdalsheiene. Numerous and often large bodies of metagabbro occur mainly as concordant lenses, but may locally show intrusive relations to the supracrustals. Late Precambrian post-orogenic granites cross-cut all the other rock types.

Reviews are given of the lithostratigraphy and structure of the Telemark Supracrustals as a background for a more detailed study of the petrology of the basic meta-igneous rocks. Petrographical investigations of these rocks show that the amphibolites (metabasalts) are almost completely recrystallized, while the metagabbros have usually retained the outlines of primary hypidiomorphic plagioclase crystals. Mineral compositions and assemblages indicate that the area suffered metamorphism of uppermost low-grade conditions. Chemically the rocks are metabasalts of transitional tholeiitic/alkaline character. The geological environment and the chemical data indicate that the amphibolites and metagabbros originally were formed in a within-plate (continental) environment. These findings are compared with present-day tectonic regimes and with published plate tectonic models for southern Norway. It is suggested that the amphibolites represent rift volcanism in a subsiding basin not very far from a continental margin. The supracrustal rocks were deformed, metamorphosed and intruded by gabbro magmas during later adjacent orogenesis which was related to subduction and calc-alkaline volcanism.

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Introduction

Supracrustal rocks considered to belong to the Precambrian Telemark Suite (Dons 1960, 1963; Sigmond 1978) occur as large outlier, in the form of a synclinorium, in the area of Setesdalsheiene, south-central Norway (Plate 1). The location and extent of the outlier is shown on the recently published 1:250 000 geological map-sheet Sauda (Sigmond 1975, 1978); it measures some 15 km by 10 km.

Geological investigations were carried out in parts of the area during the years 1965 to 1968 by F.M. Vokes and T. Vrålstad, centred on small copper-molybdenum and copper deposits lying in the area between Skyvatn and Langvatn (Plate 1). An account of the mineral exploration and deposit assessment methods employed in the Langvatn area has been given by Vokes et al. (1975).

During the course of the investigations, detailed mapping of the rocks in the area Skyvatn–Holmavatn–Mjåvatn was carried out at a scale of 1:12,500, in order to provide a geological framework for the investigation of the mineral deposits. The present report gives the results of the above geological field work and of subsequent studies carried out at Geologisk Institutt, NTH, Trondheim.

General geology

The area under consideration lies between Mjåvatn in the east and Sandvatn and Holmavatn in the west (Plate 1). It covers roughly 135 km^2 and represents the southwesterly two-thirds of the synclinorium mentioned in the introduction. The area is largely occupied by Precambrian supracrustal rocks probably correlatable with the Bandak Group of the central area of the Telemark Suite (Dons 1960, Sigmond 1978). The highest mountains of the area (Plate 1) are capped by parautochthonous to allochthonous Palaeozoic rocks comprising basal grits, phyllites, schists and gneisses (Sigmond 1978). These rocks will not be discussed in this paper.

The Telemark Suite rocks include metabasalts, with quartz-mica and calcareous schists, arenites, conglomerates and other clastic sedimentary rocks; and possibly also some units of originally acid volcanic rocks. Large masses of metagabbro occur, usually in the form of apparently concordant intrusions into the supracrustal units. The rocks underlying the Telemark Suite rocks, the Gneiss-Migmatite-Granite Complex of Sigmond (1978), were not investigated in the present study.

In the extreme north of the area mapped, along the shores of Holmavatn, there is a non-foliated granite showing irregular, intrusive contacts against the Telemark supracrustals and their intruding metagabbros. These outcrops appear to represent the southeastern margin of a very large granitic batholith which extends northwestwards towards the Folgefonna area, a distance of some 60 km. This granite is of Precambrian age, but is younger than the deformation and metamorphism of the Telemark supracrustals of the region. The possibility that this granite might be present at depth under the Langvatn area has relevance to discussions on the genesis of the Mo-Cu deposits there. Its likely presence is tentatively suggested in the vertical section, Fig. 1.

Telemark suite rocks

LITHOSTRATIGRAPHY

The present study has established a lithostratigraphy for the Precambrian supracrustal rocks of the Langvatn area, based on field observations and the structural interpretation shown in Fig. 1. The writers have adopted the formational names proposed by Sigmond (1978) for these rocks. A brief summary is given below, starting with the uppermost unit exposed.

5. *Svartepodd metasandstone*. Calcareous sandstones, generally shistose in the northeast; possibly tuffites. Amphibolite layers (metabasalts) in the southwest.
4. *Båstogvatn conglomerate*. Conglomerate with clasts of metabasalt. Much detrital epidote. An impersistent unit.
3. *Skurefjell metabasalts*. Metavolcanites. A unit of homogeneous, fine-grained and porphyritic ('speckled amphibolite') metabasalts.
2. *Skyvatn conglomerate*. Conglomerate with predominantly sedimentary clasts.
1. *Veggine formation*. Three members distinguished:

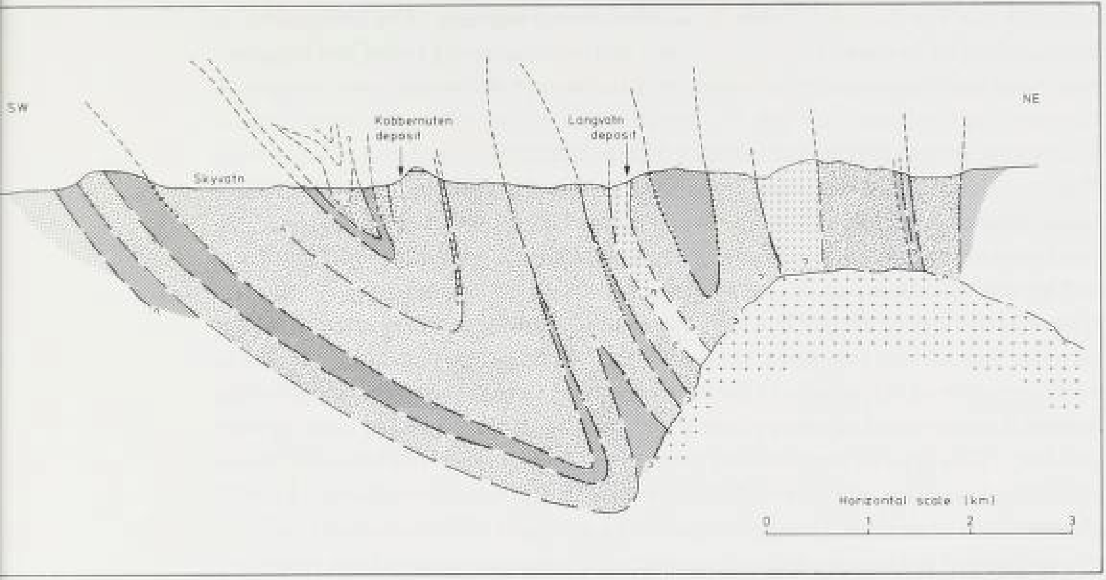

Fig. 1. Interpreted vertical section through the Kobbernuten and Langvatn deposits. Vertical relief not to scale. Legend as for map, Plate 1.

c) Mixed clastic sedimentary rock – metabasalt unit. Lithologies are either interbedded, or metabasalts occur as semiconcordant sills in sediments. Current-bedded sediments are abundant. Metagabbro layers in the south-east.

b) Homogeneous metabasaltic volcanites.

a) Mixed clastic sedimentary rock – metabasalt unit. Lithologies are either interbedded, or metabasalts occur as semiconcordant sills in sediments.

Metagabbros of varying field appearance and structural relations occur within all units.

The present account will deal mainly with aspects of the metabasalts of the Skurefjell and Veggine formations, and of the intrusive metagabbros of the area.

STRUCTURE

The supracrustal rocks of the Langvatn area, as well as the intrusive metagabbros, are strongly folded along NW-SE trending axes. The intensity of the folding increases in a north-easterly direction across the area. A line running NW-SE through Storheddervatn and Øydeskyvatn divides the area into two structurally distinct parts. Southwest of this line, the folding is less intense and the supracrustals are disposed in a single syncline with a slightly overturned NE limb (the Båstogvatn syncline). The rocks are little deformed and sedimentary structures, such as cross-bedding and graded bedding, are abundant in the clastic units.

Northeast of the Storheddervatn–Øydeskyvatn line the folding becomes isoclinal and an axial plane schistosity is prominent in the supracrustal rocks, especially in the belt immediately northeast of Langvatn. The differences in structural styles between the southwestern and northeastern parts of the Langvatn area have made correlation between the sedimentary units somewhat uncertain.

TELEMARK SUITE LITHOLOGY – FIELD OBSERVATIONS

The metavolcanite units (units 1b, and 3) can reach considerable thicknesses of homogeneous lithology (Plate 1). The Skurefjell metabasalts (unit 3 above), for example, are at least 100 m thick in the southwestern part of the area mapped.

The *metabasalts* are dark green massive to schistose amphibolites. Two main types can be distinguished. The rather more abundant of these is massive and has an even, fine to very fine grain size and occurs in units from a few decimetres to a few metres thick. Especially in the southwestern part of the area, this amphibolite variant is massive and in many places it is difficult to distinguish any layering with certainty. This type of amphibolite commonly shows a vesicular structure where possible amygdales are now filled with calcite, epidote and quartz. Another characteristic feature of this rock type is the presence of lens-shaped bodies of coarse-grained epidote, quartz (± calcite) with maximum sizes of the order 1–2 metres. These lie parallel to the formational contacts of massive and schistose amphibolites. No obvious pillow structures were observed, although the above-described epidote-rich lenses may possibly represent altered pillows.

The second, less abundant, amphibolite type has a speckled or flecked appearance due to an abundance of lens-shaped spots, 2–5 mm in length, that are aligned parallel with the planar structure of the rock. These spots, consisting mainly of dark green amphibole, occur in a fine-grained, light green groundmass. The two types of amphibolite occur in intimate association. The units of speckled amphibolite are generally in the form of impersistent lenses from 10 to 100 m in lengt. The two types were not differentiated during the mapping of the area on 1:12,500 scale.

Metagabbros. These rocks vary in character from apparently fresh, massive, coarse-grained, equigranular metagabbros to foliated coarse- to medium-grained types.

The contact relationships shown by the metagabbros are somewhat variable. The majority take the form of large sill-like bodies and lenses intruded into the metavolcanic and sedimentary rocks more or less parallell to their original layering or bedding. In the northeastern part of the area, however, where the folding is tight to isoclinal, and where axial plane schistosity is developed in the supracrustal rocks, the possibility exists that some of metagabbros may have been emplaced parallel to this schistosity and may cross-cut the layering of the rocks at depth. If so, there are two episodes of gabbro emplacement.

A particularly large mass of metagabbro present in the area southeast of Storhedderfjell shows cross-cutting relationships and a chilled margin against the surrounding supracrustal rocks of the Svartepoddvatn metasandstone formation. These relations are best seen in the area between Skyvatn and Storheddervatn (Plate 1), where the intrusion of the gabbro produced a considerable disturbance of its country rocks. The layering in the supracrustals is partly overturned, while a large raft of amphibolites and fine-grained metasediments lies structurally above, and is partly enclosed by, the gabbro. This is interpreted as part of the core of the Båstogvatn syncline which was lifted up from the underlying rocks due to the forcible intrusion of the gabbro. The layering in the supracrustals has been torn apart, and fingers and xenoliths of these rocks can be found along the margins of the metagabbro body. Similar effects can be seen west of Mjåvatn in the extreme southeast of the mapped area, where large lenticular bodies of metagabbro occur parallel to the layering of the supracrustal rocks and finger out into these in both directions along the strike. Such relationships indicate that much of the gabbro intrusion occurred either simultaneously with the folding of the supracrustal rocks or subsequently to this.

Petrology of the basic rocks

About ninety thin-sections and polished sections of volcanic rocks and metagabbros have been examined. Plagioclase (18) and amphibole (18) compositions were determined using the automatic ARL microprobe at the Institutt for Røntgenteknikk, University of Trondheim-NTH. Furthermore, selected samples of volcanic rocks (17) and metagabbros (8) were analysed for major and selected trace elements.

PETROGRAPHY

Volcanic rocks (amphibolites). Primary textures are usually obliterated. Only in one sample has a porphyrite-like texture been observed, and possible amygdales were found in rocks from only a few outcrops. The present metamorphic mineral assemblages are characterized by a texture where amphibole porphyroblasts occur in a 'matrix' of more fine-grained, granoblastic and commonly untwinned plagioclase. In the speckled amphibolites the amphibole crystals are quite large (up to 5 mm) or consist of many smaller grains clustered together, thus giving the characteristic speckled appearance. The main minerals are plagioclase, amphibole, chlorite and epidote with subordinate biotite, rutile, calcite, sericite and opaque minerals. Apatite has not been observed. Albite, quartz, chlorite, epidote and calcite are the principal minerals in secondary veins and in amygdales. The plagioclase is a basic oligoclase (An 22–26.5), except in one sample (LV 176) where andesine (An 35) was found. The amphibole is usually homogeneous and transitional between actinolite and pargasite/ferropargasite in composition (Fig. 2). However, in a few cases a zoned amphibole with a core of actinolite is found. The common amphibole seems to be in a equilibrium relations with green chlorite. The opaques comprise ilmenite with hœmatite lamellae, pyrite, and subordinate chalcopyrite.

Systematic variations in petrographic characteristics have not been recorded within or between the two main volcanic units. The textural differences between the massive and speckled varieties of amphibolite are not easily understood. A tentative explanation is that they may reflect different conditions for mineral growth during metamorphic recrystallization.

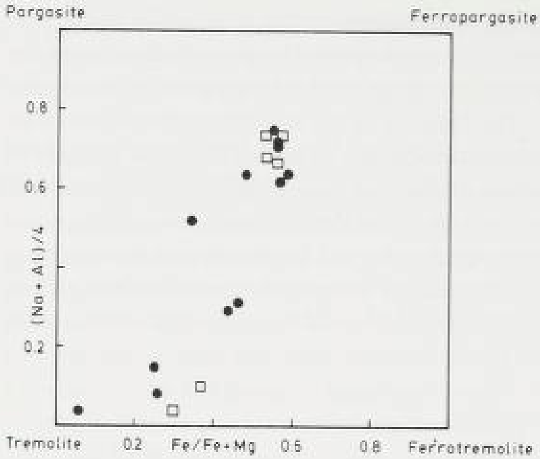

Fig. 2. Distribution of Langvatn amphiboles in the composition range: tremolite-ferrotremolite-pargasite-ferropargasite. □ = from metagabbros, ● = from amphibolites.

Metagabbro. This rock type is medium- to coarse-grained. Primary textural features such as hypidiomorphic plagioclase crystals are commonly preserved; however, incipient recrystallization of plagioclase is prominent in some samples. In these cases the minerals usually form equidimensional grains with typical granoblastic texture similar to that found in the amphibolites. The plagioclase is now a basic oligoclase (An 23.5 – An 30) and is always saussuritized. The number of small epidote grains is greatest in the central parts of the plagioclase crystals indicating original normal zoning. The amphibole is sometimes poikiloblastic with abundant quartz inclusions. In one sample an amphibole crystal probably is a pseudomorph after pyroxene. Some of the amphiboles are zoned with a core of pale actinolite and a pleochroic yellow-green to blue-green rim of hornblende. The common, homogeneous amphibole is of the latter type. In many samples there is abundant apatite in grains up to 2 mm in size. Other minerals are green chlorite, olive-brown biotite, epidote, sericite and opaques. When chlorite occurs, it seems to be in equilibrium relations with amphibole. The principal opaque mineral is ilmenite, which characteristically has exolved hœmatite lamellae. Magnetite and sulphides (pyrite, chalcopyrite and pyrrhotite) are subordinate phases in most samples. The biotite is commonly associated with ilmenite and its brownish colour is probably due to a relatively high Ti/Fe ratio. Epidote and sericite are breakdown products of plagioclase. Accessory minerals are quartz, calcite and sphene.

METAMORPHIC GRADE

Some of the minerals present in the metamorphic basic rocks from Langvatn may be used to infer the metamorphic grade. According to Winkler (1974) the transition from actinolite to hornblende and the coexistence of oligoclase (An $>$17) and hornblende is typical in the upper part of low-grade metamorphism. The appearance of oligoclase (An 17) is estimated to take place 20°–40°C lower than the transition between low-grade and medium-grade metamorphism. The slightly higher An-content of the Langvatn plagioclases thus indicates a somewhat higher metamorphic grade. However, Winkler (1974, p. 165) further states that

Table 1. Major element and CIPW-norm compositions of amphibolites and metagabbros from Langvatn

	AMPHIBOLITE																	METAGABBRO							
		UNIT 1b										UNIT 3					UNIT 5								
	LV1	LV2	LV40	LV140	LV143	LV193	LV11	LV12	LV20	LV44	LV46	LV98	LV109	LV113	LV116	LV122	LV86	LV15	LV51	LV75	LV80	LV89	LV108	LV110	LV161
SiO_2	45,18	43,63	46,27	44,59	43,75	46,39	47,14	45,79	45,44	46,68	43,67	49,79	44,79	45,24	44,29	48,12	44,94	46,04	44,64	44,92	45,85	44,40	42,90	45,37	42,87
TiO_2	2,95	1,26	2,33	3,70	2,33	1,95	2,90	1,39	2,87	1,15	3,03	2,25	1,60	1,11	1,39	1,40	3,02	1,20	2,95	1,83	2,95	2,76	3,90	2,56	3,93
Al_2O_3	15,76	15,35	15,33	15,09	16,99	16,79	16,02	16,64	15,91	9,78	16,25	15,68	16,71	17,80	17,32	16,31	15,44	17,13	18,30	18,47	16,66	15,62	16,00	17,28	15,73
Fe_2O_3	5,44	4,54	4,31	3,76	3,18	4,81	2,30	7,08	4,16	5,01	4,72	5,03	2,51	3,38	5,13	4,51	9,59	3,36	2,98	1,80	2,02	2,48	2,78	2,35	2,64
FeO	9,60	7,71	9,03	10,91	10,43	8,61	10,62	5,13	10,67	5,21	9,86	7,72	10,74	7,93	7,15	7,80	6,22	7,48	9,64	9,73	10,30	11,07	13,97	10,66	12,69
MnO	0,23	0,23	0,17	0,19	0,26	0,23	0,17	0,13	0,25	0,23	0,23	0,19	0,23	0,21	0,22	0,18	0,22	0,19	0,17	0,17	0,16	0,16	0,24	0,15	0,16
MgO	6,10	9,20	7,86	6,30	7,39	7,35	5,95	7,51	6,28	12,13	8,28	5,20	9,53	8,83	8,37	7,11	6,64	7,73	6,58	8,08	6,74	8,10	5,86	7,63	6,79
CaO	9,24	10,44	9,67	7,33	8,79	9,69	8,53	10,25	9,15	15,69	7,24	8,89	8,58	9,89	8,43	9,57	6,85	9,74	8,50	8,29	7,52	8,23	8,45	8,68	7,61
Na_2O	2,77	4,35	4,20	2,53	3,07	2,25	2,50	3,09	2,79	2,09	3,33	3,36	2,84	2,91	2,80	2,98	2,68	2,84	2,28	3,10	2,98	3,28	2,51	2,43	2,59
K_2O	0,60	0,29	0,18	0,22	0,42	0,11	0,85	0,64	0,55	0,09	1,00	0,34	0,33	0,32	–	0,67	0,35	0,97	0,67	0,75	0,42	0,81	0,93	0,80	0,47
P_2O_5	0,20	0,12	0,29	0,59	0,17	0,13	0,48	0,13	0,20	0,05	0,22	0,17	0,12	0,06	0,12	0,19	0,41	0,08	0,29	0,18	0,47	0,61	0,45	0,41	0,33
H_2O^+	1,93	1,74	1,36	3,22	2,82	1,59	0,94	1,27	1,78	1,18	2,53	1,13	2,85	2,32	3,47	1,22	2,03	1,67	2,44	2,26	2,45	2,06	1,69	1,75	2,55
H_2O^-	0,04	0,08	0,10	0,10	0,02	0,01	0,14	0,13	0,04	0,07	0,07	0,03	0,08	0,02	0,09	0,08	0,08	0,10	0,10	0,10	0,12	0,08	0,03	0,07	0,08
CO_2	0,07	–	–	0,32	0,04	0,16	0,04	0,20	0,01	0,12	–	0,01	0,04	0,05	0,32	0,12	0,12	0,04	0,04	0,05	0,20	–	0,02	0,40	0,14
Total	100,11	98,94	101,10	98,86	99,66	100,07	98,57	99,38	100,10	99,48	100,43	99,97	100,95	100,07	98,81	100,26	98,58	98,57	99,58	99,73	98,84	99,66	99,73	100,54	98,56
Q	–	–	–	3,3	–	–	2,0	–	–	–	–	1,9	–	–	–	–	–	–	–	–	–	–	–	–	–
Or	3,5	1,7	1,1	1,3	2,5	0,7	5,0	3,8	3,3	0,5	5,9	2,0	2,0	1,9	–	4,0	2,1	5,7	4,0	4,4	2,5	4,8	5,5	4,7	2,8
Ab	23,4	11,7	25,7	21,4	24,0	19,0	21,2	21,9	23,6	9,4	23,8	28,4	23,7	21,0	23,7	25,2	22,7	21,4	19,3	24,5	25,2	26,2	21,2	20,6	21,9
An	28,8	21,5	22,5	29,2	31,3	35,4	30,0	29,7	29,3	17,0	26,4	26,7	31,9	34,6	34,7	29,2	29,1	31,4	37,7	34,3	30,8	25,5	29,7	33,9	29,9
Ne	–	13,6	5,3	–	1,1	–	–	2,3	–	4,5	2,4	–	0,2	2,0	–	–	–	1,4	–	0,9	–	0,8	–	–	–
Cpx	12,5	24,0	19,2	1,0	8,8	8,9	9,2	15,6	12,1	47,9	6,5	13,3	7,9	11,2	3,5	13,3	1,2	13,3	1,8	4,3	1,7	9,2	7,6	3,2	3,8
Opx	8,5	–	–	22,7	–	21,4	19,2	–	9,0	–	–	16,2	–	–	9,3	8,1	27,4	–	18,8	–	22,6	–	7,8	16,0	16,8
Ol	8,5	17,8	15,2	–	18,6	3,7	–	16,8	8,8	12,3	20,0	–	24,6	20,9	16,3	11,4	–	17,2	2,6	20,2	–	18,3	10,2	7,9	4,5
Mt	6,5	4,0	5,6	7,5	5,6	5,0	6,4	4,2	6,3	3,8	6,6	5,4	4,5	3,8	4,2	4,2	6,6	3,9	6,5	4,8	6,5	6,2	4,8	5,9	7,9
Il	5,6	2,4	4,4	7,0	4,4	3,7	5,5	2,6	5,5	2,2	5,8	4,3	3,0	2,1	2,6	2,7	5,7	2,3	5,6	3,5	5,6	5,2	7,4	4,9	7,5
Ap	0,5	0,3	0,7	1,4	0,4	0,3	1,1	0,3	0,5	0,1	0,5	0,4	0,3	0,1	0,3	0,5	1,0	0,2	0,7	0,4	1,1	1,4	1,1	1,0	0,8
Cc	0,2	–	–	0,7	0,1	0,4	0,1	0,5	–	0,3	–	–	0,1	0,1	0,7	0,3	0,3	0,1	0,1	0,1	0,5	–	–	0,9	0,3
H_2O^+	1,9	1,7	1,4	3,2	2,8	1,6	0,9	1,3	1,8	1,2	2,5	1,1	2,8	2,3	1,2	1,2	2,0	1,7	2,4	2,3	2,4	2,1	1,7	1,8	2,6

Analysts: P.R. Graff, NGU (LV1, LV20, LV46, LV51, LV75, LV98, LV108, LV113, LV143, LV193)
I. Rømme & I. Vokes, NTH (LV2, LV11, LV12, LV15, LV40, LV44, LV80, LV86, LV89, LV109, LV110, LV116, LV122, LV140, LV161).

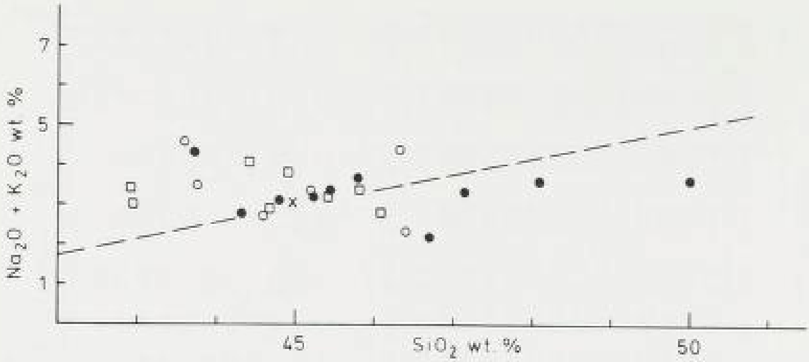

Fig. 3. Alkalis versus silica plot of Langvatn rocks. Dividing line after Irvine & Baragar (1971). □ = metagabbros, ● = amphibolites of units 1b & 3, x = amphibolite of unit 5.

«the presence of chlorite in the paragenesis is diagnostic of low grade». The parageneses of the Langvatn rocks thus formed under upper low-grade metamorphism very close to the transition to medium-grade.

The actinolitic cores of amphiboles of the metagabbros and in a few amphibolites might indicate that the rocks were metamorphosed under other conditions previously. The significance of this feature however, is not easily understood since actinolite may exist both in low-grade (Winkler 1974) and in medium-grade metamorphism (Miyashiro et al. 1971). One possible explanation is that these zoned amphiboles represent uralitized clinopyroxenes in which equilibrium relations were not attained during the metamorphism. The low-Al actinolitic cores, then, may represent more closely the original clinopyroxene composition, whereas the outer hornblende rims may reflect the actual metamorphic conditions.

These findings correspond well with the general conclusion of Sigmond (1978) who found that supracrustal rocks of the Telemark Suite in the Sæsvatn – Valldal area have been subjected to conditions corresponding to the transition between low- and medium-grade metamorphism. The interpretation that abundant chlorite represents a later low-grade retrogradation (Sigmond 1978) is, however, not supported by the present investigation. In that case one would also have expected some adjustments of the plagioclase composition.

CHEMISTRY

Major and trace element data for the analysed rocks are presented in Tables 1 and 2. The major element chemistry (Table 1) is 'basaltic' for rocks of all units. A general feature is the quite low silica content. The rather broad variation of TiO_2 displayed by all groups of rocks is probably a result of some kind of fractionation or accumulation process. K_2O also shows considerable variation, but this may, at least partly, be a result of secondary processes. However, the rather high TiO_2 and K_2O values found in many of the rocks are typical of basalts from plateau lavas, from some oceanic islands and from anomalous oceanic ridge segments (Wood et al. 1979). In the alkalis-silica diagram (Fig. 3) the rocks straddle the division line between the alkaline and subalkaline (tholeiitic) fields. Molecular norms vary

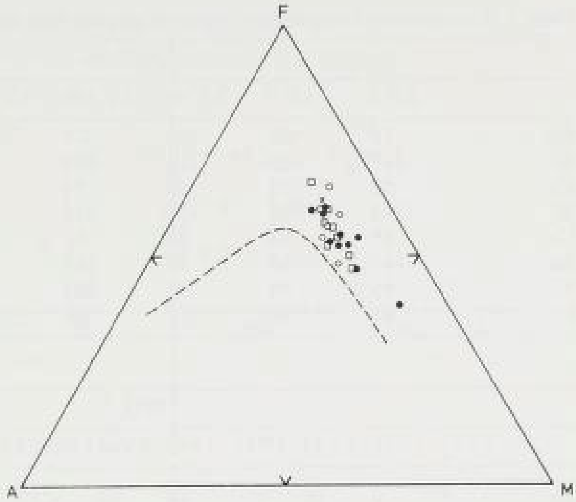

Fig. 4. AFM diagram. Dividing line after Irvine & Baragar (1971). Symbols as in Fig. 3.

from slightly ne-normative via ol + hy-normative varieties to slightly q-normative (Table 1). Both these methods of classifying rock types are, however, sensitive to secondary processes and may thus not be conclusive as to the original chemical affinity. In the AFM diagram (Fig. 4) the samples plot within the area of alkaline and tholeiitic rocks.

Trace element data are presented in Table 2 and in variation diagrams (Fig. 5). The variation patterns of Ni, Cr, Y, P_2O_5, K_2O and TiO_2 versus Zr do not show significant separate trends for any of the groups of rocks. The metagabbros have a much higher Sr mean value than the amphibolites (Table 2), but this is probably an effect of plagioclase accumulation because the highest Sr contents are found in samples especially rich in Al_2O_3 (Tables 1 & 2). The more incompatible (and immobile) elements P_2O_5, TiO_2 and Y show semi-linear trends with Zr. This feature indicates that a certain range of fractionation is represented or that the rocks represent magmas formed as a result of various degrees of partial melting of a mantle source which was homogeneous for these elements. The irregular groupings of Ni, Cr, Sr and K_2O may perhaps represent primary features, but are more likely of secondary origin because these elements are generally thought to be rather mobile during alteration and metamorphism (Pearce 1976).

In the Zr-Ti-Y diagram (Fig. 6) most of the Langvatn rocks plot in the fields of within-plate basalt and ocean-floor basalt. The Zr-Ti-Y diagram does not always successfully discriminate between the tectonic setting of basaltic rocks (Prestvik 1982). It was found, e.g., that Upper Miocene plateau basalts in Northern Spitsbergen (Prestvik 1978), just as the Langvatn rocks, plotted in the same two fields of the Zr-Ti-Y diagram. Based on this similarity, the fact that no geological features indicate that the amphibolites (or metagabbros) of the Langvatn area represent an oceanic environment, and other geochemical/geological features referred to above, it may be concluded that the Langvatn rocks are of transitional tholeiitic/alkaline character and were formed in a within-plate (continental) environment.

Table 2. Trace-element compositions (ppm) of amphibolites and metagabbros from Langvatn

	UNIT 1b						
	LV 1	LV 2	LV 40	LV 140	LV 143	LV 193	Mean Unit 1b
Rb	28	<5	<5	<5	12	<10	–
Sr	225	199	218	303	185	198	221
Y	52	25	46	74	47	35	46.5
Zr	218	23	156	243	185	112	156
Cu	35	32	55	59	12	34	38
Zn	147	36	55	186	110	103	106
Ni	45	44	35	103	–	70	49.5
Cr	85	99	157	99	143	119	117

	UNIT 3											UNIT 5
	LV11	LV12	LV20	LV44	LV46	LV98	LV109	LV113	LV116	LV122	Mean Unit 3	LV86
Rb	5	9	19	<5	19	11	<5	17	<5	25	–	<5
Sr	336	269	210	239	123	288	241	220	223	243	239	289
Y	50	27	54	27	56	45	24	27	17	18	34.5	42
Zr	224	64	225	64	246	182	41	66	76	97	128.5	174
Cu	66	24	37	16	15	34	19	17	63	205	49.5	31
Zn	15	43	138	8	176	114	73	88	149	39	84	86
Ni	81	91	51	39	58	47	62	115	98	91	73	82
Cr	39	67	88	39	116	80	39	141	98	99	80.5	39

	META GABBRO								
	LV 15	LV 51	LV 75	LV 80	LV 89	LV 108	LV 110	LV 161	Mean Gabbro
Rb	6	25	23	<5	<5	30	9	<5	–
Sr	191	435	405	348	287	250	406	272	324
Y	31	43	30	34	38	65	39	28	38.5
Zr	65	218	132	186	205	340	151	155	181.5
Cu	35	17	21	43	31	66	39	67	40
Zn	42	135	118	82	55	143	82	137	99
Ni	61	106	150	109	71	44	94	102	92
Cr	38	149	159	124	98	18	67	67	90

Analysts: J. Sandvik, I. Rømme & I. Vokes, NTH [Rb, Sr, Y, Zr, (XRF); Cu, Zn, Ni, Cr (AAS)]
G. Faye & M. Ødegård, NGU [Ni, Cr (XRF)]

Discussion

If the Langvatn amphibolites (and younger metagabbros) are continental, in what kind of geological setting did they form? The sedimentary rocks of the area do not conclusively point to a continental milieu, but the alternation between quartzites, conglomerates and volcanic rocks together with calcareous sandstones, often cross-bedded, is a strong indication that the rocks were deposited in a shallow water, marine environment, possibly in an intracratonic basin. On the other hand, the lack of obvious pillow structures in the amphibolites may indicate that these were erupted as subaerial flows.

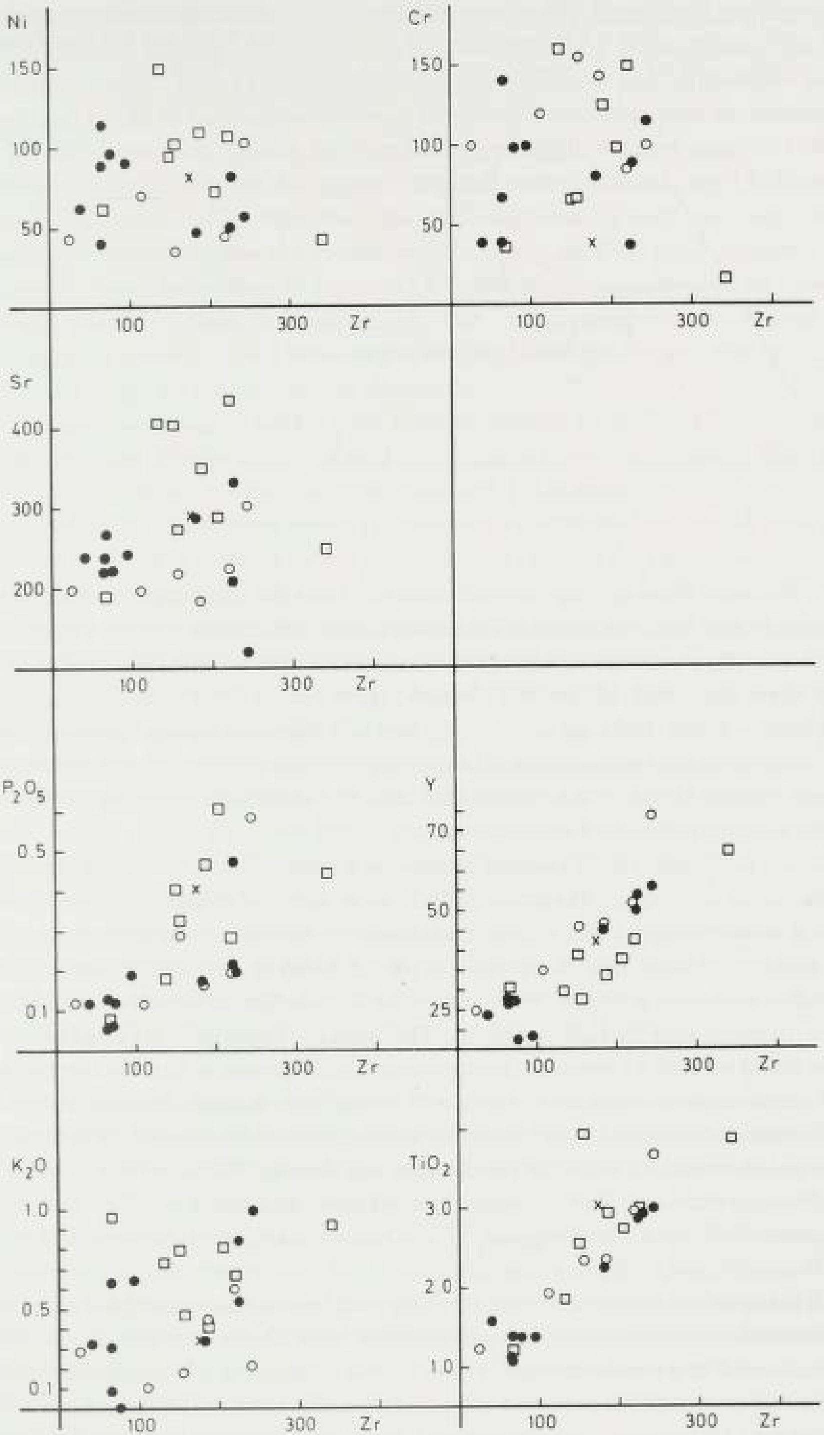

Fig. 5. Minor and trace elements of Langvatn rocks plotted against Zr. Symbols as in Fig. 3.

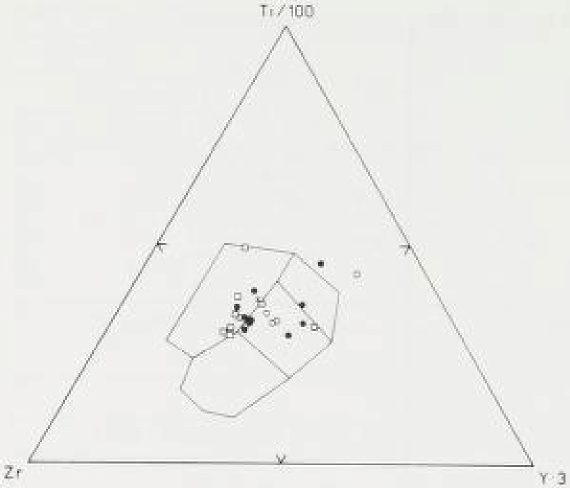

Fig. 6. Discrimination diagram using Ti, Zr and Y. (After Pearce & Cann 1973.)

In southern Norway, supracrustal rocks of about the same type as those in the Langvatn area occur in central Telemark (Dons 1960) and on the Folgefonna peninsula (Kvale 1945) as well as in some areas within the Sauda (1:250,000) map-sheet (Sigmond 1975). In Telemark, three groups have been distinguished, and basic volcanic rocks occur in the Rjukan and Bandak Groups. Quartzites and quartzite conglomerates occur in all the groups but are predominant in the Seljord Group. Singh (1968, 1969) concluded that the quartzites were deposited in a shallow sea to intertidal flat environment.

Moine & Ploquin (1972) made a geochemical study of the rocks of the Telemark supracrustal rock suite. Their conclusions were that the volcanics of the Rjukan Group indicate a thick sialic crust in the area at the time of deposition and that the basaltic rocks of this group show a partial trend of iron enrichment and are probably tholeiitic in origin. The rocks of the Rjukan Group were thought to have been contaminated by sialic material. The volcanic rocks of the Bandak Group were found to have a chemical development similar to that of the Rjukan Group, and these authors suggested that the Rjukan and Bandak Groups could be equivalents. More recent work, however, indicates that there are evident differences between the volcanic rocks of the Rjukan and Bandak Groups (Sigmond, oral communication) and that metabasalt/sandstone sequences west of the main Telemark Suite area, the Langvatn lithologies included, are best correlated with the Bandak Group (Sigmond 1978).

Lithologies similar to those in the Langvatn area occur some 30 km to the south-southwest in the Nesflaten–Grjotdokki area where they pre-date a thick succession of meta-andesitic and related rocks (Sigmond 1978). Furthermore, Naterstad et al. (1973) reported supracrustal rocks from Valldalen, 30–40 km northwest of Langvatn, and interpreted them as equivalents to the Telemark supracrustals. The lithologies described by these authors resemble those from Langvatn, and they include both metavolcanites, conglomerates and intrusive gabbros. In fact, Sigmond (1978) correlates the rocks from the Langvatn and Valldalen areas directly. However, no chemical data have been reported for the Valldalen rocks. The Langvatn, Nesflaten–Grjotdokki and Valldalen areas (Sigmond 1978) all have abundant basic volcanics correlatable with the continental Bandak Group; but only in the Nesflaten–Grjotdokki area are these rocks (the Blåbergås Group) overlain by typical orogenic meta-andesites, the Heddevatn Group (Sigmond 1978). However, the time relations between the large-scale continental volcanism and the formation of the calc-alkaline meta-andesites are not quite clear. According to Sigmond (1978) the meta-andesites are clearly younger than the basic volcanic rocks, but no unconformity or change in metamorphic grade is found between the Blåbergås and Heddevatn Groups. This indicates a more or less continuous deposition.

The vast outpouring of mainly basic volcanic rocks in a continental environment (Langvatn) was most probably related to a tensional or rifting episode (Baker et al. 1978). On the other hand, orogenic volcanism (Heddevatn Group) is probably connected with a compressive stress regime along a continental margin. A principle question is: do such different stress patterns occur more or less simultaneously in a restricted area, or could the tectonic relations be of another kind than indicated here? If the answer to the first question is *no*, it implies that the stage of tension or rifting preceded the compressive regime produsing the orogenic volcanism. Even thouth no unconformable relations are found between the Blåbergås and Heddevatn Groups, Sigmond (1978) mentions that the two groups represent well-defined and separate sedimentological and volcanological environments. This observation in itself indicates that a break between the deposition of the two groups is not unlikely. On the other hand, the answer may be *yes*, because there are well-known examples of more or less simultaneous continental and orogenic volcanism. The Patagonian plateau of South America, e.g., formed in a kind of continental back-arc environment behind the already active orogenic Andes Range (Baker et al. 1981). In the U.S.A. the Columbia River Basalts formed in a complicated structural situation (Hooper & Camp 1981) very close to the simultaneously active orogenic Cascade Range (Prestvik & Goles, in prep).

Even though most traces of original tectonic features in the Langvatn and surrounding areas are now obliterated, the features and similarities referred to above seem to support the view that this area of southern Norway developed rather close to an active continental margin where also orogenic volcanism took place.

Several models for the plate tectonic evolution of the Proterozoic of southern Norway have been proposed. Torske (1977) proposed an arcuate zonation with an *Interior Province* comprising the Telemark Suite, a *Central Belt* with gneisses and granitoid plutons and subordinate calc-alkaline volcanics, and a *Marginal Zone* where the volcanism was inferred to be of theoleiitic type. Falkum & Petersen (1980) suggested that the Telemark Suite rocks formed in epicratonic basins reflecting tensional stresses slightly before subduction and orogenic magmatism started along the continental margin. Furthermore, Berthelsen (1980) presented a model (which concentrated on the more easternmost segments of the Sveconorwegides), where he mentioned that the deposition of the Telemark supracrustals preceeded orogenesis related to eastward subduction west of the Rogaland–Agder area.

It is not the aim of this report to discuss the large-scale tectonic picture of the Late Proterozoic in southern Norway. It may be stated, however, that the conclusions reached above concerning the formation and environment of the rocks of the Langvatn area are in relatively good agreement with (at least, not in serious conflict with) all the plate-tectonic models referred to above. The Langvatn supracrustals were probably folded and metamorphosed during the period of orogenesis that was much stronger closer to the continental margin. The intrusive gabbros were probably emplaced during and partly after the main phase of this type of activity.

Acknowledgements. – The authors acknowledge the criticisms of two unknown reviewers of the first draft of this paper. The final draft also benefited considerably from suggestions from Dr. David Roberts, NGU, on the first draft, and Statsgeolog Ellen Sigmond, NGU, on the second draft.

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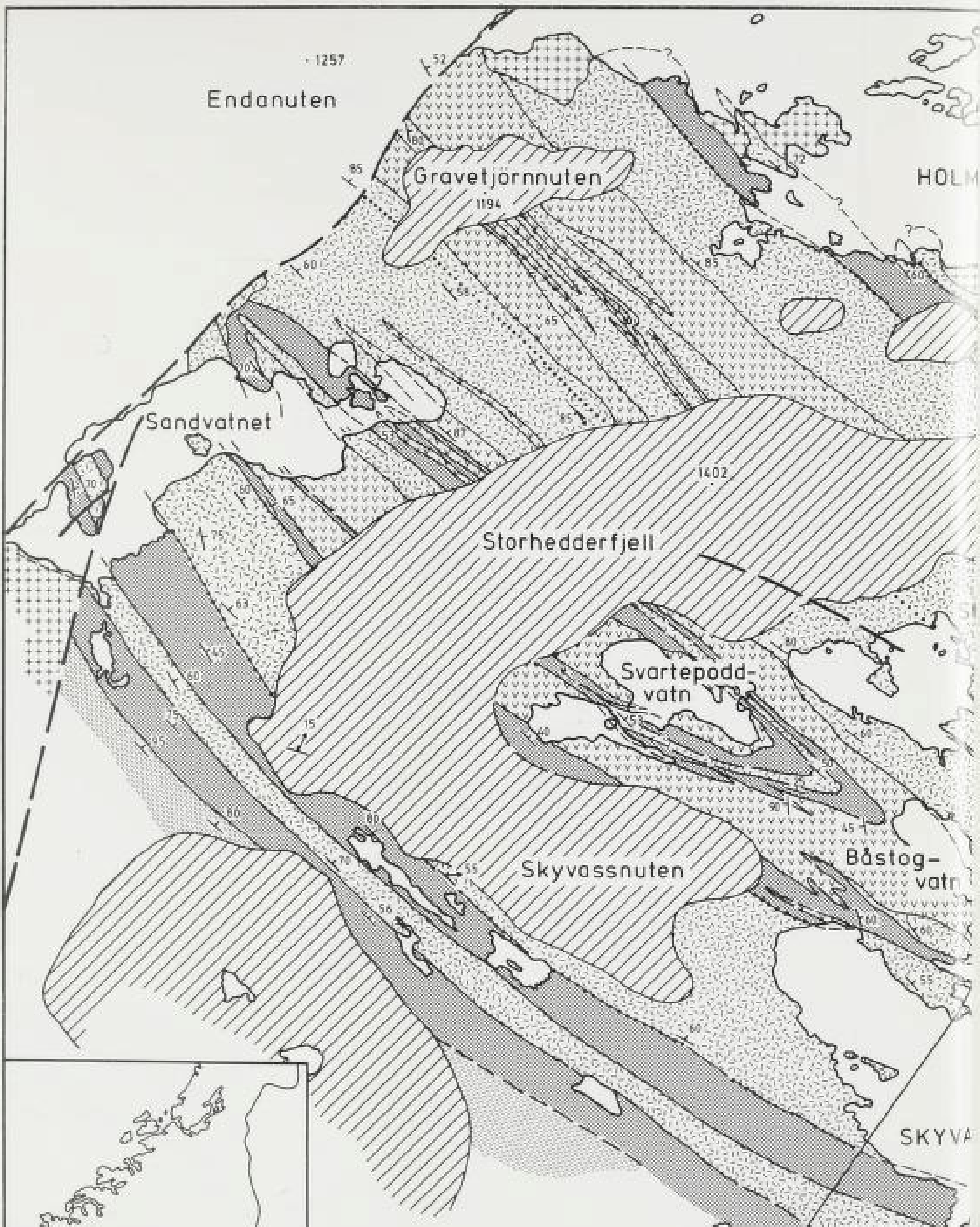

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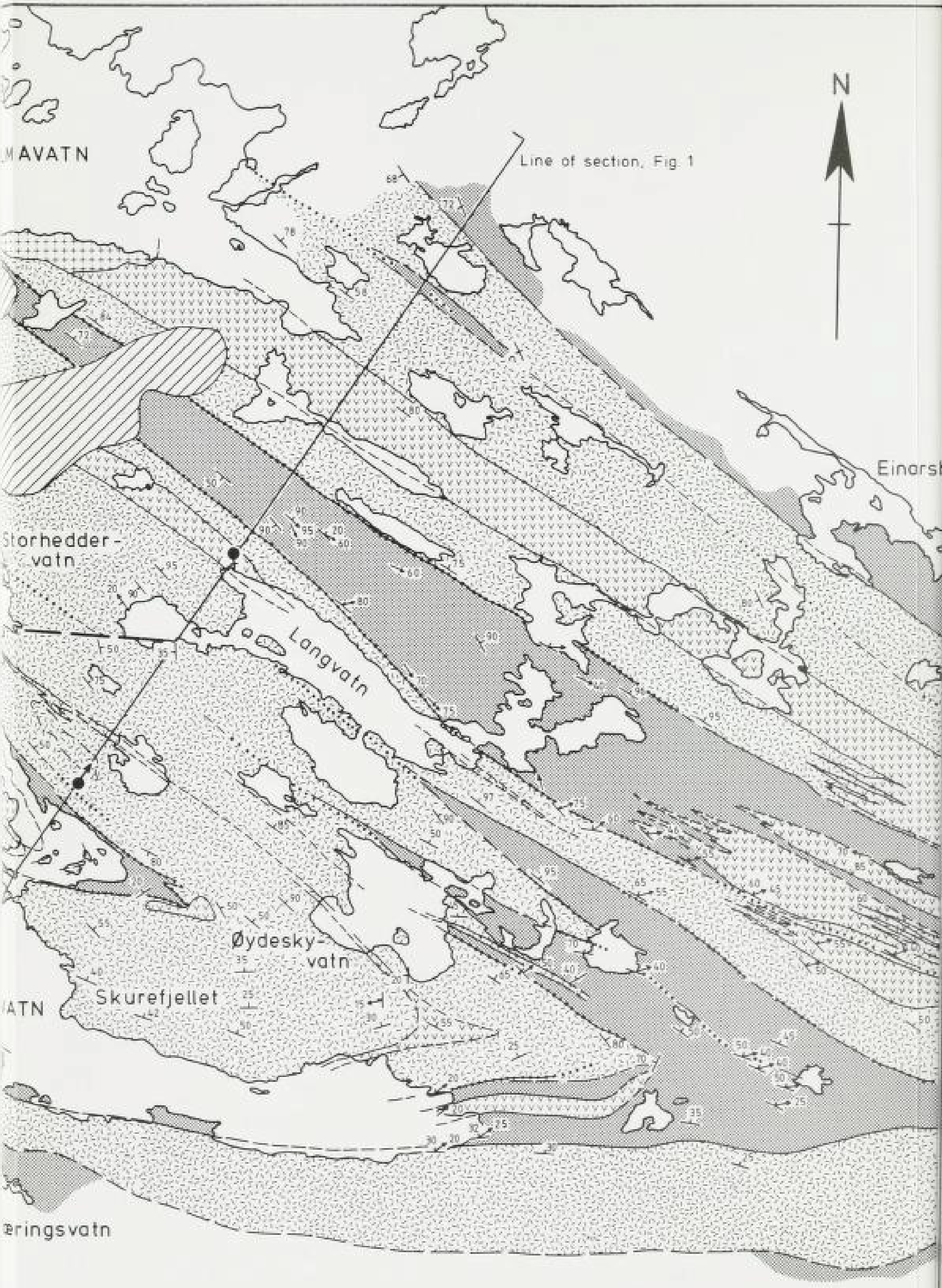
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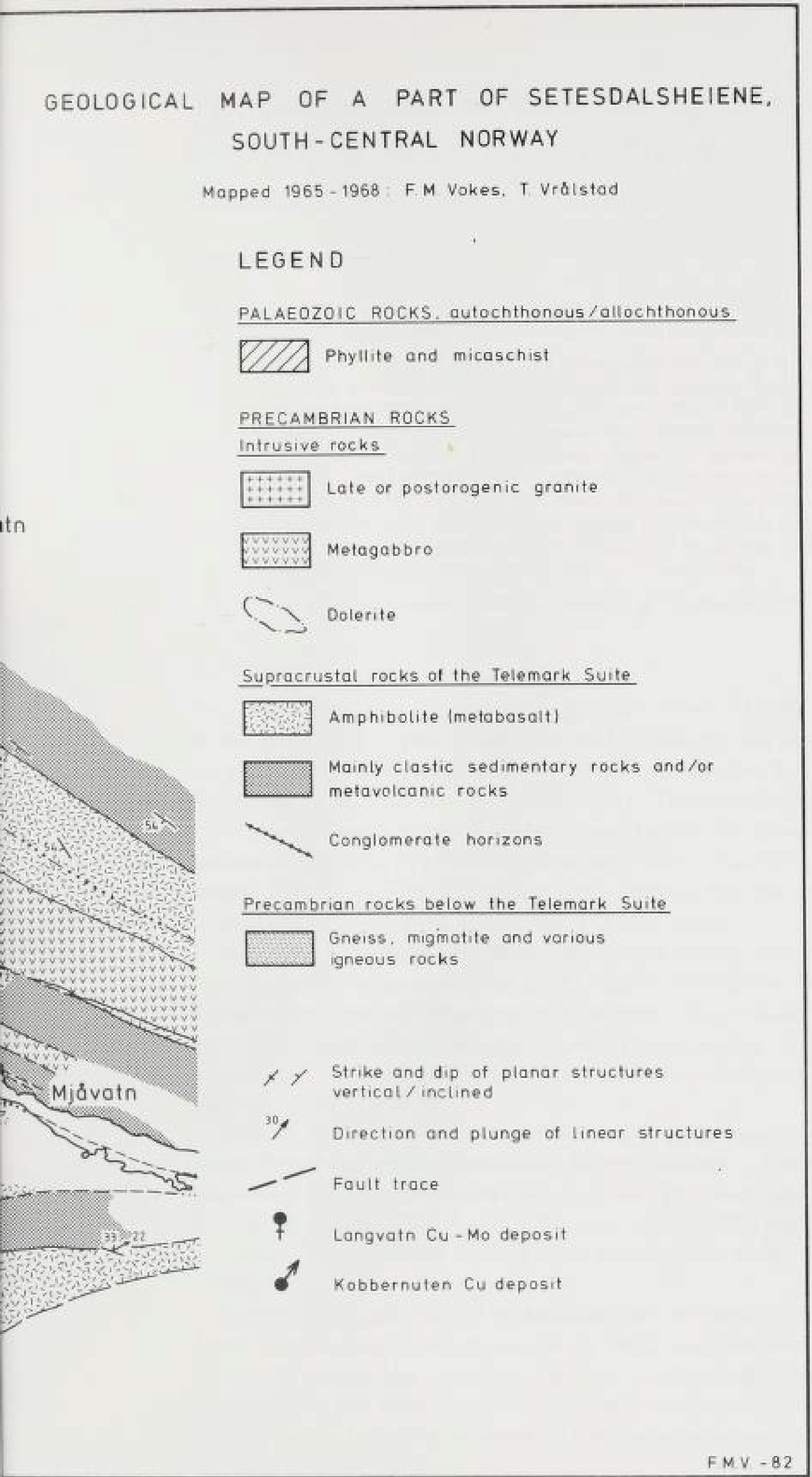

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The Nature and Tectonic Setting of Mélange Deposits in Soknedal, near Støren, Central Norwegian Caledonides

ODD NILSEN

Nilsen, O. 1983. The nature and tectonic setting of mélange deposits in Soknedal, near Støren, Central Norwegian Caledonides. *Norges geol. Unders. 378*, 65–81.

A sedimentary mélange occurs as a mega-lens just east of the disjunctive boundary between the Støren Nappe and the Gula Nappe within the Trondheim Nappe Complex in the Soknedal–Støren area. Packets, blocks and shreds of pervasively sheared Gula and Støren lithologies occur as fragments within a virtually undeformed siltstone-greywacke matrix. The lithology is considered to represent an olistostrome with matrix material chiefly derived from the volcanic and volcanoclastic units of the Hovin and Horg Groups, but also from the Støren and Gula Groups. Evidence points to deposition after obduction, folding, thrusting and gravitational sagging with small-scale renewed thrusting of the Støren Nappe. The petrology of the mélange and the tectonic evolution of the adjacent nappe units suggest a late Silurian age for the olistostrome.

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Introduction

The Trondheim Nappe Complex (Wolff 1967, Guezou 1978) covers a large part of the central Norwegian Caledonides and constitutes the upper allochthonous nappe complex within the Caledonian tectonostratigraphy (Roberts 1978, Wolff & Roberts 1980, Roberts & Wolff 1981). The stratigraphy of the Støren–Hølonda–Horg area within the west-central part of the Trondheim Region was established by Vogt (1945) and comprises a series of metavolcanic and metasedimentary units spanning an age range from probable late Precambrian/Cambrian (Gula Group) to assumed early Silurian (Horg Group).

The Støren Group, of presumed Cambrian to earliest Ordovician age, constitutes the main volcanogenic unit of the western Trondheim Region. The group forms the basal part of a separate nappe, the Støren Nappe (Gale & Roberts 1974), within the upper allochthon, together with the succeeding metasedimentary and metavolcanic assemblages of the Hovin and Horg Groups (Vogt 1945, Chaloupsky 1970) (Table 1).

A study of the trace element geochemistry of the basaltic greenstones of the Støren Group has revealed ocean-floor tholeiitic affinities. Eastward translation or obduction of the metavolcanics upon the underlying Gula Group was postulated by Gale & Roberts (1974) and Roberts & Gale (1974); evidence favouring an obduction in pre-Middle Arenig time has been forwarded by Furnes et al. (1980).

In view of the earlier scanty evidence for a tectonic break between the Gula and Støren units (Roberts 1967, Guezou et al. 1972) and their overall conjunctive and apparantly gradational boundary relations in the Innset–Oppdal area of the south-western part of the Trondheim Nappe (Rohr-Torp 1972), some doubts were initially cast on the obducted nature of the Støren volcanites by the present author

Table 1. Nappe succession of the Central Trondheim region (after Wolff & Roberts, 1980 and Roberts & Wolff, 1981) with subdivisions and equivalent Swedish tectonic units. The column Singsås Formation to Sandå Formation refers to the present author's subdivisions

			Horg Gr.	Sandå Fm.	
			U. Hovin Gr.	Hovin ss.	
UPPER ALLOCHTHONOUS COMPLEX		Støren Nappe	L. Hovin Gr.	Dicranograptus shale Fm.	Køli
				Krokstad Fm.	
	Trondheim Nappe Complex		Støren Gr.	Ilfjell Fm.	
				Elgsjø Fm.	
		Gula Nappe	Gula Gr.	Undal Fm.	Seve (?)
				Singsås Fm.	
	Levanger Nappe				Køli
LOWER ALLOCHTHONOUS COMPLEX	Skjøtingen Nappe				Seve
	Leksdal Nappe	Leksdalsvatn Nappe			Särv
		Hærvola Nappe			Offerdal
AUTOCHTHON	Precambrian Basement				

(Nilsen 1978). However, with the present recognition of a structural break between the two units the tectonic emplacement of the Støren Group upon the Gula now seems evident.

The present paper gives an account of the geology of the area immediately south of Støren where the nature of the disjunctive Gula/Støren contact has been investigated and where the occurrence of a local mélange unit immediately below the Støren Nappe has been mapped in some detail (Figs. 1 & 2).

General Geology

The boundary between the Gula complex or nappe (Roberts 1978) and the Støren Nappe trends NE-SW, more or less parallel with the Soknedal valley from Støren to Rennebu (Fig. 1). Near Støren, the valley bottom is heavily covered with glacial

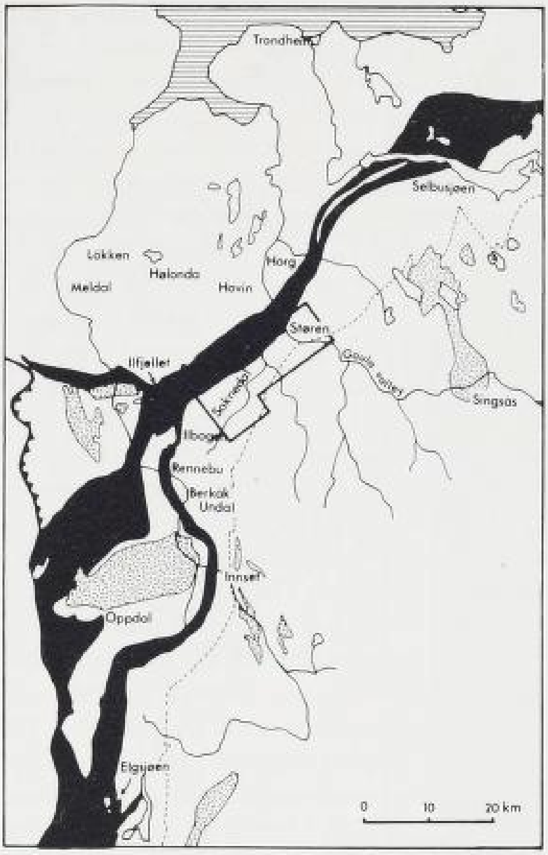

Fig. 1. Key map of the western Trondheim area showing localities mentioned in the text. Støren Group in black; the boundary between the Undal Formation and the Singsås Formation of the Gula Group is stippled. Major intrusions are cross-hatched. The boxed area is that of Fig. 2.

and alluvial deposits which tend to obscure the boundary relations. Detailed mapping through the years 1977 and 1980–1981, however, has revealed the nature of the boundary between these two tectonostratigraphic units (Nilsen 1980) in this particular area.

Geological maps have been published by Bugge (1910), Lindberg (1971), Nilsen (1978) and Rohr-Torp & Nilsen (1979) from the southern (Soknedal) parts of the area, while the northern part is covered by the 1:250,000 map-sheet 'Trondheim' (Wolff 1976).

Gula Group. A twofold division of the Gula complex in its western parts was proposed by Nilsen (1978). To the east, a lower, psammitic unit, the *Singsås Formation,* which occupies the central part of the extensive outcrop of the Gula, is exposed in the hills in the easternmost part of the present map area (Fig. 2). Lithologies range from low-grade calcite-bearing chlorite-quartz schists in the Soknedal–Hauka area to banded calc-silicate biotite schists and gneisses further south and in the Gaula valley where transitions into high-grade diopside-microcline rocks are associated with dioritic plutons in the Singsås district (Fig. 1). The formation is deformed by tight to isoclinal folds with steep, N–S-striking axial planes.

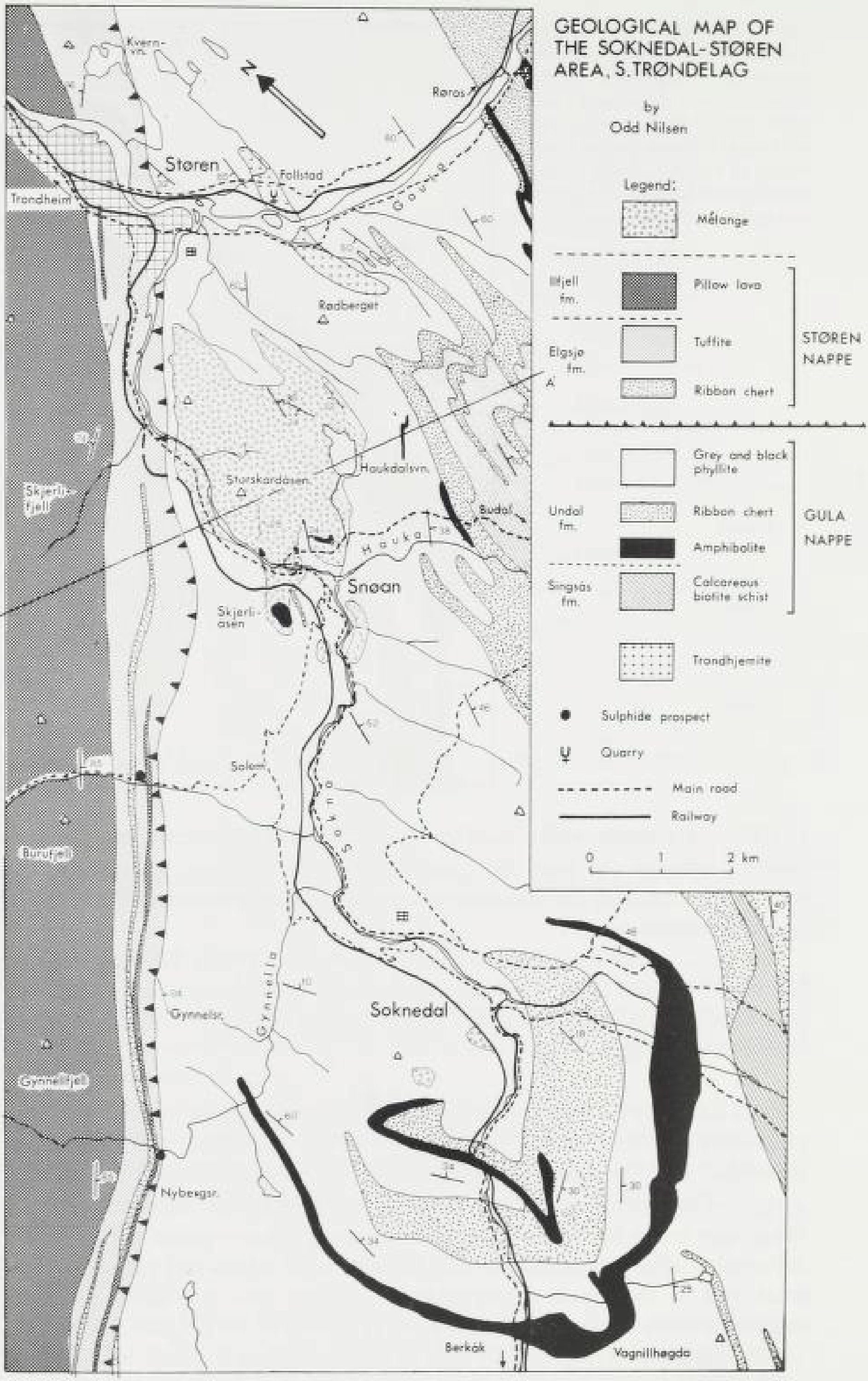

Fig. 2. Geological map of the Soknedal–Støren area.

An extensive zone of ribbon chert, grading into graphitic quartzites, marks the transition into an overlying unit, the *Undal Formation*, to the west. The Undal Formation consists mainly of graphitic metapelites with a variable carbonate content. They are chiefly developed as fine-grained, grey and black chlorite-sericite phyllites, often with thin intercalations of ribbon chert. Sulphide mineralizations, chiefly of pyrite with subordinate pyrrhotite, are ubiquitous constituents within the metapelites and occur as thin laminae and schlieren parallel to the prevailing schistotisy (Nilsen 1978).

Discontinuous, thin horizons of mafic metavolcanics occur within the Undal Formation. These are the *Gula greenstones* (Nilsen 1974) which occur mainly near the Singsås/Undal Formation boundary, but also within the central part of the Singsås Formation where they are associated with horizons of black schists and thin manganiferous iron-formations, usually in conjunction with minor cupriferous stratabound sulphide horizons (Nilsen 1978). In the Soknedal area the Gula greenstones are developed as strongly sheared and schistose, calcareous chlorite-bearing amphibolites and are intimately associated with graphitic ribbon cherts. Ultramafic pods, developed as fine-grained talc-chlorite-tremolite rocks or as medium-grained biotite hornblendites, occur within the Gula greenstones in the Hauka valley and in the Vagnillhøgda area. They were interpreted by Nilsen (1974) as cognate, cumulate, meganodules within the Gula greenstone.

The rocks of the Undal Formation show a strong axial-plane foliation, the regional S_2 schistosity, with a N–S to NNE–SSW strike and a moderate to steep eastward dip. The Undal Formation is folded into a tight, post–S_2, recumbent, SE-verging synform in the Soknedal area (the Soknedal synform), as revealed by the outcrop of the prominent amphibolite horizon (Figs. 2 & 3).

Støren Group. This major unit can be divided into two main formations in the southern Trondheim Region – the *Elgsjø Formation* (Nilsen 1978) and the *Ilfjell Formation* (new name) after the summit Ilfjellet, 8 km SW of Gynnelfjell. These have been traced more or less continuously from Støren southwards to the Elgsjøen area (Fig. 1). The *Elgsjø Formation* comprises a series of pelitic to semipelitic volcanoclastics and are chiefly developed as pastel-green, fine-grained quartz-sericite-chlorite phyllites which include thin intercalations of magnetite-bearing cherts, greyish green metagreywackes, felsic tuffs, ribbon cherts and thin (50–100 m) horizons of greenstone. A correlation of these rocks with the 'Quartz-schist/Upper phyllite unit' of the 'Lower Sedimentary Unit' of Torske (1965), bordering the Støren Group pillow lavas in the Selbusjøen area, seems reasonable.

In the Soknedal–Støren area the different units of the Elgsjø Formation have a wide lateral extent and dip steeply (75–90°) to the northwest. This pervasive (S_2) foliation (Fig. 3C) becomes overprinted by a subhorizontal (S_3) foliation towards the Gula/Støren boundary (Figs. 3E & 4) with the progressive development from a crenulation cleavage into a zone of cataclasites and mylonites (Fig. 5), approximately 1 km wide, which has been traced from Støren along the Støren/Gula boundary to the Berkåk area. Here the thermal influence of the Oppdal–Innset intrusive complex (Rohr–Torp 1974), internal thrusts and a higher

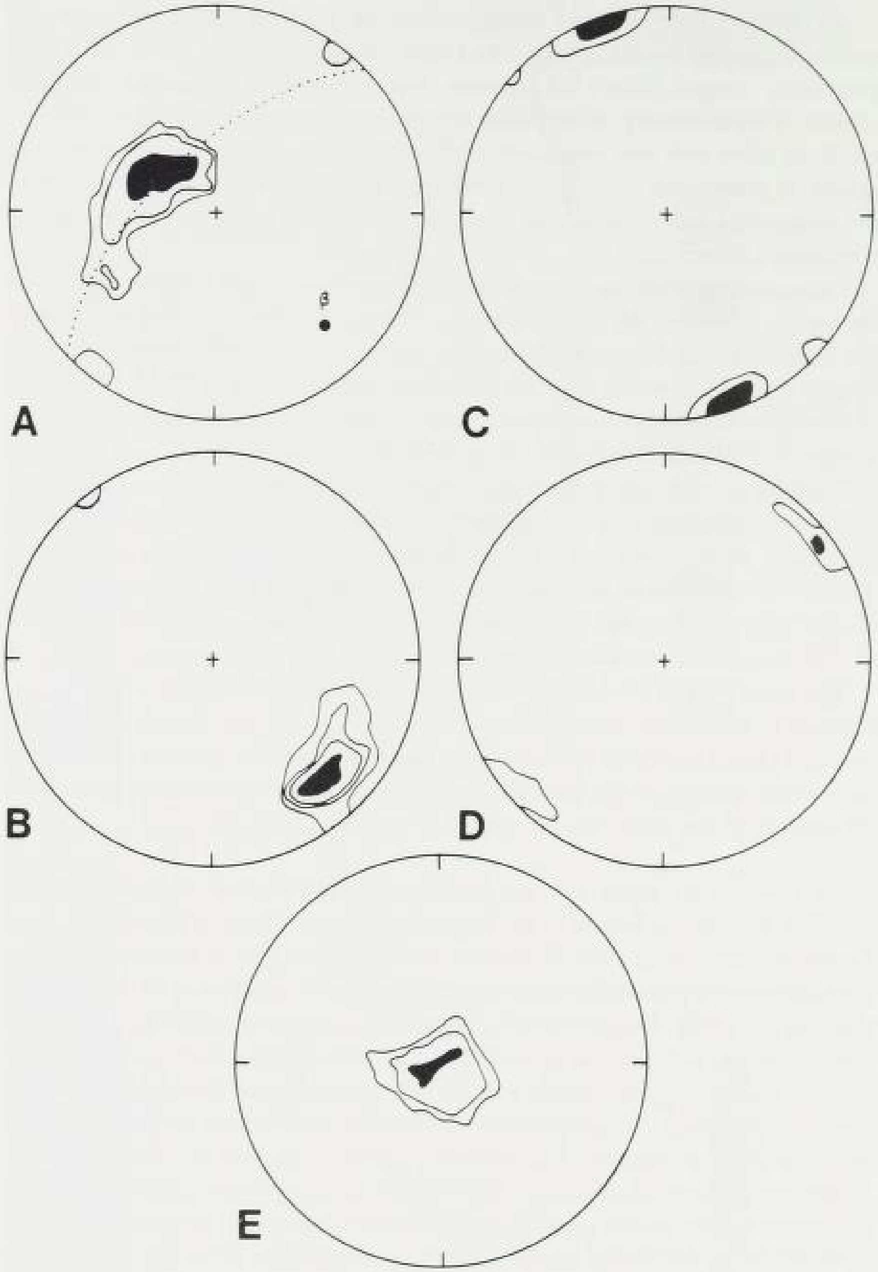

Fig. 3. Stereographic contour diagrams of structures from the Soknedal–Støren area. Equal area projection; lower hemisphere. (A) Pi diagram of S_2 foliation from the Soknedal synform (80 obs.). Contours: 10–5–3% per 1% area. Point ß denotes the statistical (F_3) fold axies of the synform. (B) Fold axes (crenulations and puckerings) of the Soknedal synform (42 obs.). Contours: 25–15–10–5% per 1% area. (C) Pi diagram of S_2 foliation in the Støren Nappe (30 obs.). Contours: 20–10% per 1% area. (D) Fold axes (F_3) in the Støren Nappe (20 obs.). Contours: 20–10% per 1% area. (E) Pi diagram of S_3 foliation from the Støren–Gula boundary zone (40 obs.). Contours: 20–10–5% per 1% area.

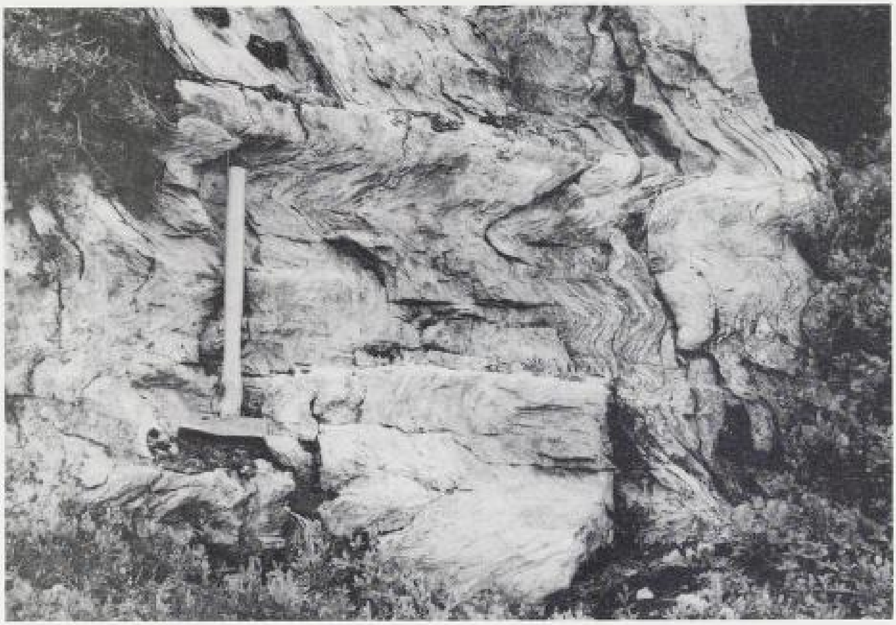

Fig. 4. A sub-horizontal (S_3) crenulation cleavage overprinting the steep (S_2) regional schistosity in tuffite from the Elgsjø Formation (sericite–chlorite–quartz phyllite) from W. Burufjell.

grade Barrovian-type regional metamorphism has obscured the disjunctive, cataclastic nature of the boundary zone. The sub-horizontal (S_3) shear-zones have affected the upper part of the Undal Formation, as revealed by the imbrication of minor stratabound sulphide horizons in the Ilbogen district (Nilsen 1978), but the effect of the locally penetrative S_3 foliation rapidly decreases to the south-east.

The *Ilfjell Formation* constitutes the major part of the Støren Group and has been the subject of much geological research over the years (Bugge 1910, Goldschmidt 1916, Carstens 1920, 1924, Oftedahl 1968, Gale & Roberts 1972, 1974, Loeschke 1976, 1977, Loeschke & Schock 1980). The name Ifjell Formation is here introduced as an informal term for a rather heterogeneous sequence of mafic pillow lavas and pillow breccias with minor intercalations of chert and a few disrupted sulphide horizons. In the upper, western part of the formation, near the border to the overlying Lower Hovin Group, there are some distinct horizons of coarse volcanoclastics. The formation also contains some small lensoid bodies of gabbro and ultramafite.

The pillow lavas of the formation in the map-area are developed mainly as massive, fine-grained, pastel-green chlorite-epidote-amphibole rocks, locally with variolitic structures as first described by Bugge (1910). In the Soknedal–Støren area the formation attains its maximum thickness of ca. 4 km, which was attributed to the existence of a local 'Soknedal volcano' by Oftedahl (1968). To the south, the pillow-lava horizon wedges out into the volcanoclastics of the Elgsjø Formation in the Oppdal area (Nilsen 1978).

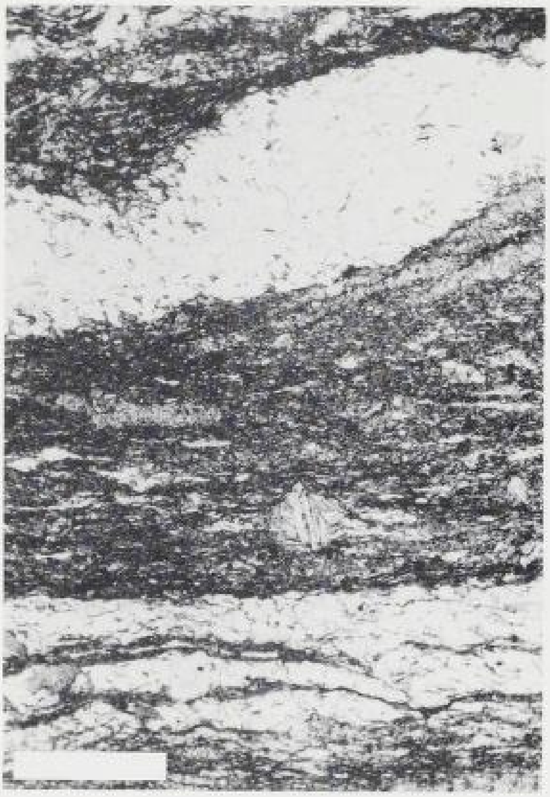

Fig. 5. Mylonite from Øverøyan, 1 km NW of Storskardåsen (UTM grid ref. 623 875). Plane polarised light. Scale bar: 0.5 mm.

INTRUSIVE ROCKS

Swarms of medium- to fine-grained trondhjemite dykes intrude the rocks of the Soknedal–Støren area. They occur as more or less composite, conformable sheets, less than 10 m in thickness, and as a few larger, lenticular bodies within the Undal Formation. Trondhjemite dykes also occur sporadically within the Ilfjell Formation, but are quantitatively much less prominent than within the Gula Group outcrop. Cross-cutting relationships as revealed by apophyses are common. Dykes of trondhjemitic type also cut the mélange described below.

A large lensoid body of trondhjemite occurs in the Follstad–Rødberget area in the Gaula valley, and has been the object of extensive quarrying at Follstad for more than a century. The Follstad locality (Fig. 2) is the type locality of trondhjemite as described by Goldschmidt (1916) and its petrology and geochemistry have recently been re-examined by Size (1979).

MÉLANGE

Weakly metamorphosed sedimentary rocks of pelitic to psammitic composition occur locally adjacent to the Gula–Støren boundary in the western Trondheim region. These have been described by Torske (1965) from the Selbusjøen area and were included as members of his Lower Sedimentary Unit (Gula Group). These

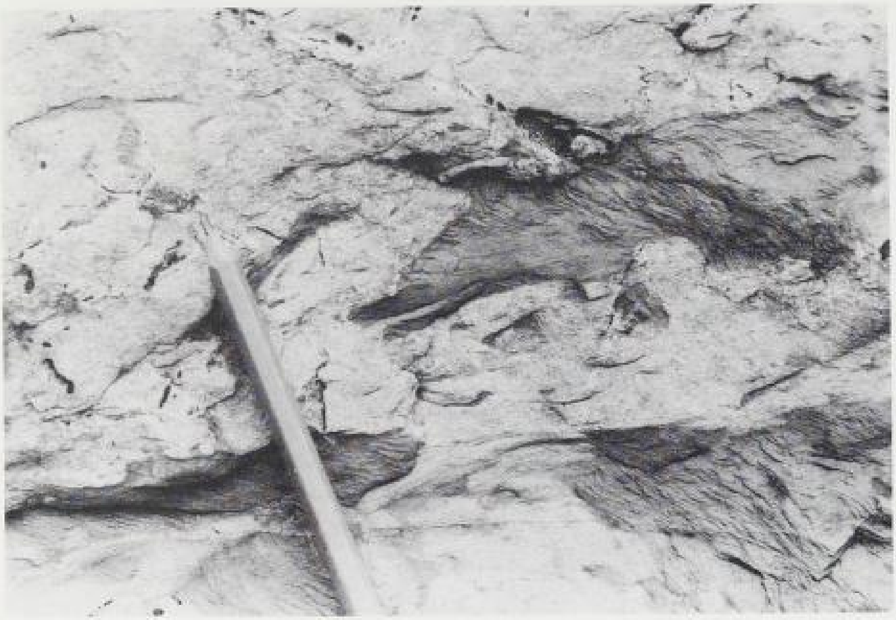

Fig. 6. Fragments of black phyllite (Undal Formation) in greywacke, Storskardåsen.

rocks have recently been reinvestigated by Horne (1979) and described as part of a mélange. Horne interpreted the mélange unit as being associated with the development of a forearc accretionary prism. Particular attention will be given in the following to the apparently corresponding deposits in the Støren–Soknedal area.

The deposits in question occur as a mega-lens of sedimentary rocks deposited upon and locally intermixed with the strongly tectonized rocks of the adjacent Undal Formation. Laterally the sedimentary rocks occur as irregular pockets and lenticles, filling in space between rotated fragments of strongly sheared and fragmented phyllite and ribbon chert of the Undal Formation (Figs. 6 & 7). Virtually undeformed, poorly sorted siltstones and greywackes constitute the main rock types in the central Storskardåsen area. They occur as massive, fine-grained rocks of greenish-grey colour. Well preserved intraformational slumps with internal bedding are present in some places, but in general the sedimentary rocks are devoid of stratification. Due to the extensive glacial and alluvial cover in the Soknedal area the boundary between the polymict sedimentary unit under consideration and the Undal Formation is not exposed. Locally, a bedded structure of the sediments is revealed by indiscrete horizons of coarser material. Where observed, bedding surfaces have in general a N–S strike and a moderate dip of 15–30° to the east and the west revealing an apparent discordant nature of the sediments in relation to the steeply dipping (60–85°) phyllites of the adjacent Undal Formation.

Near the boundary of the sedimentary complex the presence of a great variety of mega-clasts reveals a chaotic mixed lithology. Large blocks, boulders and

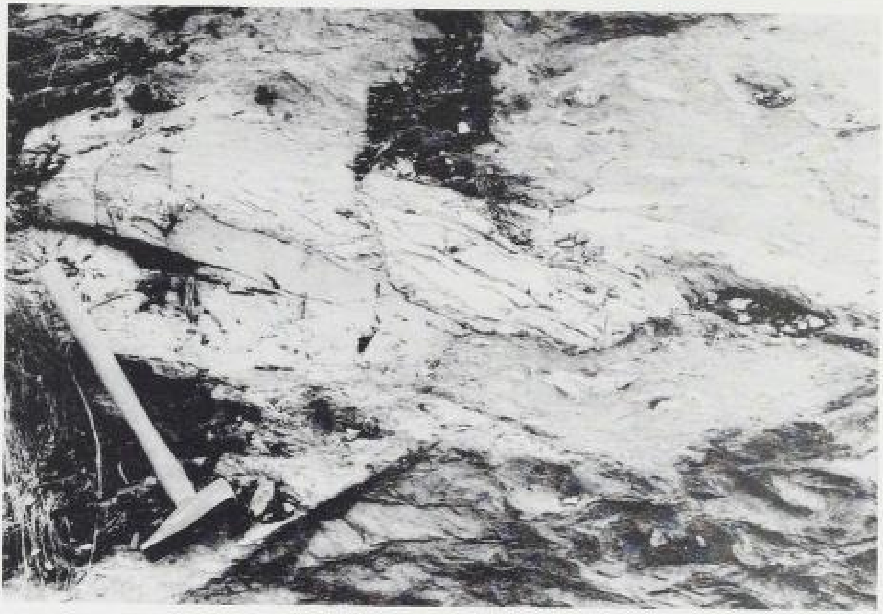

Fig. 7. Fragment of ribbon quartzite in greywacke, Snøan.

packets of strongly sheared phyllite, ribbon chert and greenstone occurring within the undeformed greywacke matrix constitute a mélange fabric which is exposed in roadcuts along the Hauka valley near Snøan (Fig. 7). The mega-clasts range from fist size to blocks hundreds of metres in maximum dimensions. A greenstone mega-block exposed at the hillock Skjetliåsen covers approximately 400 m^2, but usually the fragments occur as slightly rounded and rotated elongate bodies 2–200 m in thickness with fragmented margins. Internally the mega-clasts show a complex history of deformation displaying a sheared and in part mylonitic fabric.

PETROLOGY

The greywackes and siltstones of the mélange matrix are composed of a fine-grained heterogranular quartzo-feldspathic material with a grain-size of 0.03–0.1 mm. Chlorite, sericite, clinozoisite and sphene occur in minor amounts and magnetite, pyrite, sphalerite, limonite and zircon are accessory clastic constituents. Coarser grit deposits contain subangular clasts with a grain-size of 0.6–2 mm in the fine-grained quartzo-feldspatic matrix, with a low clast/matrix ratio (Fig. 8). The clasts display no preferred orientation and reveal a polymict assemblage of the following rock-types and minerals: quartz, plagioclase, albite-quartz symplectite, quartzite, rhyolite, rhyodacite, biotite-chlorite rock, greenstone, chlorite-sericite phyllite (tuffite) and graphitic phyllite.

Quartz occurs as clear, generally strained single grains. Glomeroclasts composed for two or three subgrains usually show amoeboidal grain boundaries. Medium-grained heteroblastic calcite-quartz glomeroclasts are probably derived from quartz-calcite veins intersecting the Gula Group assemblages.

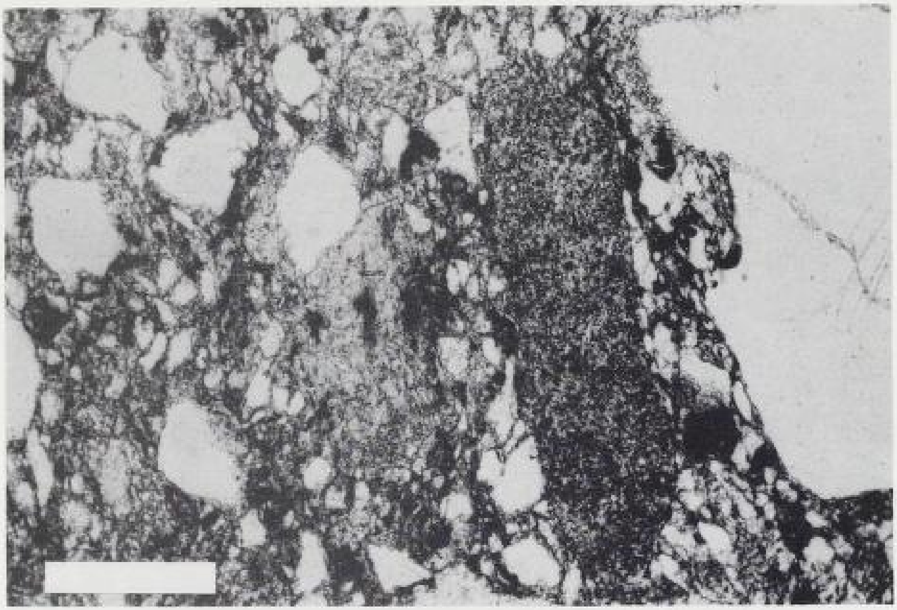

Fig. 8. Greywacke, Haukdalsmyra, NW Haukdalsvatn, (UTM grid ref. 640 876). Fragments of quartz, quartzite and tuffite (the large, grey fragment) embedded in a fine-grained chlorite-sericite-quartz groundmass. Plane polarised light. Scale bar: 0.5 mm.

Plagioclase occurs either as single grains with well developed albite twinning or, less commonly, in a symplectite intergrowth with quartz (Fig. 9). Chess-board twins are common (Fig. 10), and some grains show incipient sericitization and saussuritization.

Quartzite occurs in two different varieties: 1. As a microgranular recrystallized chert with accessory magnetite and chlorite; 2. as granoblastic equigranular orthoquartzite, probably representing clasts of the Undal Formation ribbon cherts.

The clasts of the felsic metavolcanics comprise at least three different varieties, based on textural and compositional features: 1. A porphyritic variety with resorbed and shattered phenocrysts of chess-board albite in a microgranophyric albite-quartz groundmass. 2. An aphyric albititic variety composed of a decussate, interlocking mosaic of lath-shaped albite (Fig. 11). Some varieties may contain interstitial chlorite and clinozoisite in an intersertal texture. 3. Equigranular rhyolite with an interlocking anhedral mosaic of micropoikilitic albite/quartz symplectite ('snowflake' texture).

Biotite-chlorite rock occurs as irregular fragments, with the biotite occurring as discrete flakes in a felty chlorite matrix together with clinozoisite. Greenstone clasts are composed of a fine-grained aggregate of chlorite, albite and sphene with accessory biotite. Tuffites possibly deriving from the Elgsjø Formation (chlorite-sericite phyllite) are less common clastic components (Fig. 8). Black phyllitic fragments are composed of sericite and chlorite with abundant graphite and sulphide ore minerals. Scattered, rounded grains of magnetite and zircon may be present in some of the greywackes.

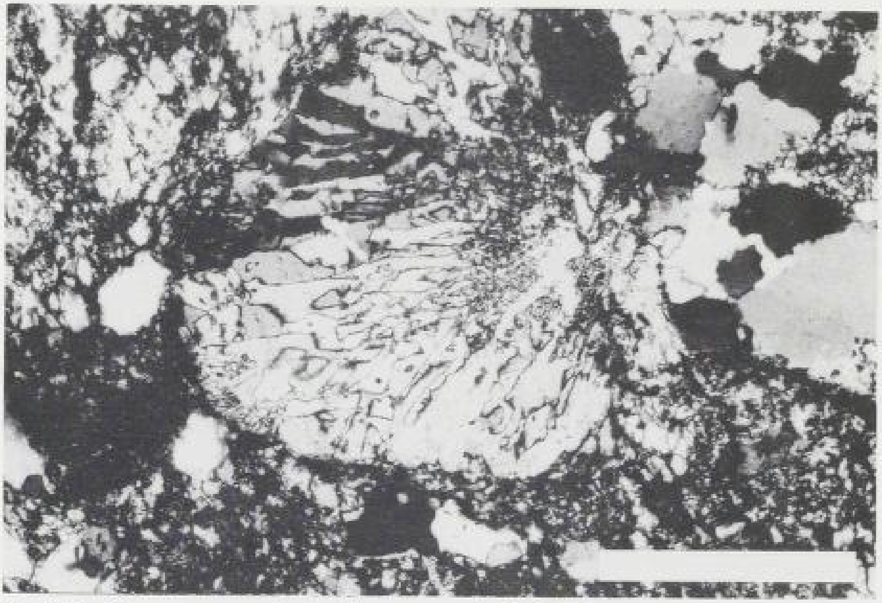

Fig. 9. Plagioclase-quartz symplectite. From greywacke, Haukdalsmyra, same locality as in Fig. 8. Crossed nicols. Scale bar: 0.5 mm.

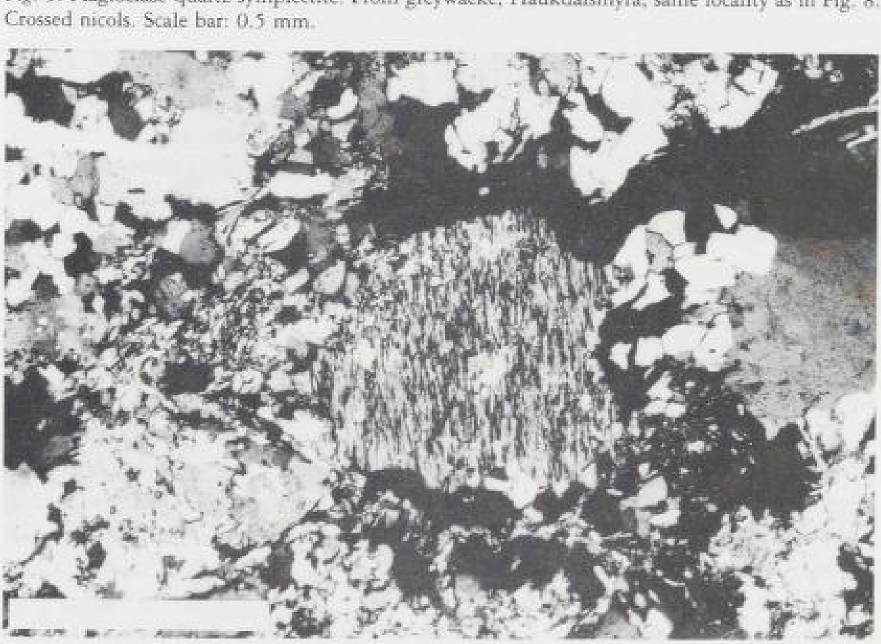

Fig. 10. Chess-borad albite. Clast in greywacke, Snøan (UTM grid ref. 865 631). Crossed nicols. Scale bar: 0.5 mm.

The lithology of the sedimentary clasts in the greywacke matrix corresponds fairly closely with those of the Lower Hovin Group metasediments as described by Vogt (1945) and Chaloupsky (1975). The clasts of the coarser, gritty greywackes reveal a great similarity to the clasts encountered in the lowermost

Fig. 11. Clast of aphyric rhyolite (albititic) in greywacke, Snøan. Same locality as in fig. 10. Crossed nicols. Scale bar: 0.5 mm.

psammitic and psephitic members of the Lower Hovin Group (i.e. the Krokstad beds, as described by Chaloupsky (1963)), with abundant recrystallized jasper and fragments of chess-board albite and micropegmatite. Chess-board albite has also been described as a constituent of quartz-dioritic pebbles in the Lyngestein conglomerate of the Horg Group (Vogt 1945).

The petrology of the felsic fragments corresponds well with the composition and texture of the many units of acidic to intermediate lavas and pyroclastics in the Hovin Groups, e.g. the Grimsås rhyolite and the Hølonda porphyrite in particular, and clasts of the porphyritic granophyres resemble the quartz keratophyres intercalated within the Støren Group equivalents in the Meldal (Rutter et al. 1968) and Selbu (Torske 1965) areas. Here, the observed myrmekitic rims around phenocrysts appear to be a characteristic textural feature. However, a more detailed petrography of the felsic metavolcanics and hypabyssal rocks of the Støren Nappe is essential in order to reveal the true palaeogeographical and stratigraphical position of the source rocks for the Soknedal mélange.

DISCUSSION

The gravity data reported by Åm et al. (1973) from the Støren–Hovin area have been taken to indicate that the Støren Group extends vertically down to a depth of perhaps 10 km or more in the Støren area, forming the eastern, lowermost limb of a complex synformal structure (the Hovin synform) within the Støren Nappe (Vogt 1945, Chaloupsky 1963, 1970, Roberts 1968, Olesen et al. 1973, Oftedahl 1979). The low-angle thrusting of the Støren Group rocks as revealed by the development of a cataclasite zone at the Gula/Støren boundary in the Soknedal area can thus be considered as a local late (S_3) event with a minor low-temperature, sub-horizontal displacement of the earlier obducted, folded and weakly metamorphosed Støren Nappe. This late tectonic event corresponds in all probability with the post-thrusting gravitational collapse after the main, Silurian nappe emplacement recognized by Roberts (1971, 1978) both within and outside the Trondheim region. Late, low-temperature dislocation zones occur quite commonly within the Støren Nappe sequence between Støren and Oppdal and may be contemporaneous with this D_3 tectonic event.

As the greywacke matrix of the Soknedal mélange is devoid of any apparant imprints of the D_3 and earlier tectonic events, the deposition of the mélange in question apparently post-dates the Silurian folding and metamorphism of the Støren Nappe and adjacent nappe units. It seems therefore unlikely that the formation of the Soknedal mélange could have occurred in a tectonic-sedimentary setting related to a destructive plate boundary as a member of a fore-arc accretionary prism between the Gula and Støren Nappes as proposed by Horne (1979).

It is pertinent in this context to define the term 'mélange', as it has been applied with a genetic as well as non-genetic connotation in the literature. Following the recommendations of a Penrose Conference Report (Silver & Beutner 1980, p. 32) the term has a non-genetic meaning, referring to «. . . rock mixtures formed by tectonic movements, sedimentary sliding or any combination of such processes with no mixing process excluded». As such, the polymict sedimentary rock unit under consideration falls well within the concept of a mélange. However, one may distinguish between mélanges occurring as composite bodies of metamorphic tectonites and olistostromes (or 'sedimentary mélanges'). The former connotation has been applies to the pervasively tectonized members of the Undal Formation in general, according to the map by Horne (1979, p. 268); while the latter connotation can be applied to the rock unit considered here. The distinction between these different usages of the term 'mélange' has been discussed by Raymond (1975) who stresses the presence of exotic blocks in a fragmented matrix and not the presence of tectonites as a main criterion for distinguishing a mélange from other rock units. Horne's model for the development of the Trondheim Nappe Complex has been rejected by Roberts (1980) largely on grounds of the incompatibility of a fore-arc accretionary prism involving nappe units of great age ranges with the composition and derivation of a fore-arc subduction complex.

The disjunctive nature of the Gula–Støren boundary in the Soknedal–Støren area may be composite feature, attributed to a primary, destructive plate obduction/thrust contact subsequently folded, metamorphosed and then locally displaced along flat-lying (S_3) shear planes with the renewed development of cataclasites and mylonites along the contact (Fig. 12).

The mélange deposition must have taken place after the last major tectonic event (D_3), incorporating angular meta-clasts of the internally sheared and in part mylonitized fragments of the Undal Formation, the greywacke matrix filling in local depressions and fissures in the folded Gula Group basement. The occurrence of abundant mega-clasts of crenulated black phyllite and granoblastic ribbon chert

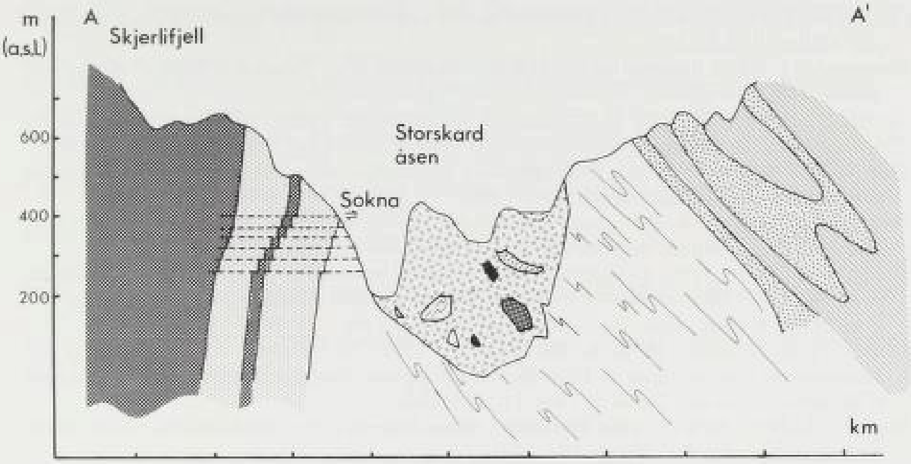

Fig. 12. Schematic section A – A' (Fig. 2) across the Soknedal valley through Storskardåsen. The ornament is the same as on the geological map, Fig. 2. Note that the vertical scale is exaggerated 5 times that of the horizontal.

within the greywacke points to a minor influx of lithified Gula Group material during deposition.

The Soknedal mélange may have been deposited as a debris flow or by sedimentary sliding of material dervied from the Støren and Hovin Group lithologies, incorporating mega-clasts or olistoliths derived from the adjacent Undal Formation which formed a highly fractured irregular floor to the local olistostrome basin. The olistoliths are mostly of local origin, being subjected to only a short-range transport during the deposition. The Soknedal mélange can thus be classified as an olistostrome formed by the mixing by submarine sliding of a variety of lithified rocks in an unlithified fluid sediment matrix forming a 'broken formation' or an 'endolistostrome' in the sense of Raymond (1978). The deposition of the Soknedal mélange was thus of a relatively young age, postdating the obductional event of the Støren Nappe, the main Silurian folding and metamorphic events, the thrusting of the Trondheim Nappe Complex, and the D_3 (or D_4) 'sagging' event. The evidence presented suggests a possible late Silurian age for the formation of the mélange.

Acknowledgements. – I am grateful to 2 anonymous referees and to the editor for constructive criticism and helpful comments on the manuscript. The views expressed, however, are solely the responsibility of the author.

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Isbevegelser i Lillehammer-området, SØ-Norge, under siste nedisning

LARS OLSEN

Olsen, L. 1983: Ice movements in the Lillehammer area, south-east Norway, during the last glaciation. *Norges geol. Unders. 378*, 83–91.

The ice movements from the last glaciation in the northern part of the Lake Mjøsa district are presented in an ice phase model. Four main ice phases are reconstructed mainly based on striation, but fluted surfaces, poorly developed drumlins, marginal moraines, erratics and till stratigraphy have been recorded and are also incorporated in the model. The ice movement directions in the inland ice phase are found to vary between SSW and SSE in the investigated area. No evidence was found for an ice divide located in or south of the Lillehammer area as previously proposed. Ice movement patterns in the deglaciation period are reinterpreted. Earlier, the ice movements in the deglaciation period were believed to have been directed towards south-east in the northern part of the Lake Mjøsa district. This is found to be true for the last part of the deglaciation period, whereas the early part was dominated by an ice movement direction towards east to east-southeast in the investigated area.

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Innledning

I forbindelse med mitt hovedfagsarbeide (Olsen 1979) ble det konstruert en isfase-modell for isbevegelsesmønsteret over nordvestre deler av Mjøs-området. Isfasemodellen er underbygget med flere observasjoner i 1980, og presenteres derfor her i en modifisert utgave (Fig. 1). Modellen bygger på rundt 250 skuringslokaliteter med skuring som antas i hovedsak å stamme fra siste nedisning. Faseutviklingen viser derfor forløpet av siste nedisning i området. Ufullstendig utbygde drumliner og fluted surface, samt randmorener, flyttblokker og morenestratigrafi er også kartlagt og bygget inn i isfase-modellen. Morenestratigrafien er under bearbeidelse i en egen artikkel (Olsen, in prep.).

Undersøkelsesområdet for Garnes & Bergersen's (1980) regionale isbevegelsesstudier strekker seg i SØ ned til Lillehammer-traktene, ved nordenden av Mjøsa. Undersøkelsen som her presenteres er et detaljarbeide i utkanten av det ovennevnte regionale undersøkelsesområdet. Isfase-modellen presenteres som et bidrag til korrelasjonen av isbevegelsesmønsteret i Mjøs-området med tilsvarende i Øst-Jotunheimen og Gudbrandsdalen.

Isbevegelsene over Lillehammer-området er delt inn i 4 hovedfaser, benevnt A, B, C, D1, og D2 (Fig. 1). Isbevegelsesmønsteret i fase A, B og C stemmer godt overens med tilsvarende faser i Garnes & Bergersen's modell (Garnes 1975, Garnes & Bergersen 1980). Isbevegelsene under fase C er funnet å ha hatt et noe annet og mere sammensatt forløp enn tidligere antatt. Deglasiasjonsperioden, fase D1 og D2, har også fått en annen utvikling enn det som tidligere har vært foreslått. I en rekke artikler (bl.a. Garnes op.cit. og Garnes & Bergersen op.cit.) er isbevegelsene under delglasiasjonsfasen(e) antatt å ha vært rettet mot SØ over nordlige del av Mjøs-området. Dette er bekreftet ved denne undersøkelsen, men bare for siste del av perioden.

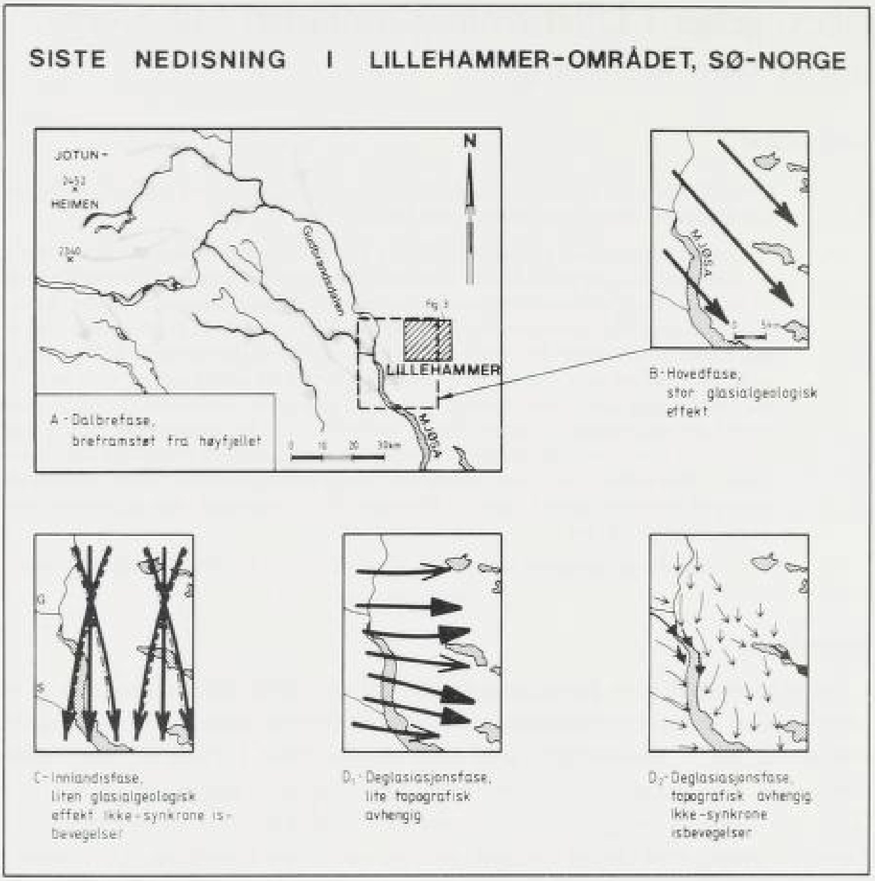

Fig. 1. Hovedisbevegelser under siste nedisning, Lillehammer, SØ-Norge. G-Gausdal (-dal = valley). S-Saksumdal. Nøkkelkart: se fig. 2. Fylte piler markerer kraftigere bevegelser. Dalbrefasen (A) vesentlig etter Garnes & Bergersen (1980). Området for fig. 3 er innrammet og skravert.
Main ice movements during the last glaciation in the Lillehammer area, south-eastern Norway. Location map: see Fig. 2. Filled arrows indicate stronger movements. Valley glacier phase, A, mainly according to Garnes & Bergersen (1980). The area of Fig. 3 is framed and shaded.
Phase A: *Valley glaciers moving out from the mountains.* Phase B: *Major ice phase with strong glacial-geological influence.* Phase C: *Inland ice phase, little glacial-geological influence, arrows showing non-synchronous movements directed southwards.* Phase D1: *Deglaciation period, initial phase. Minor topographical dependence.* Phase D2: *Deglaciation period, last phase. Arrows showing non-synchronous topographically dependent movements.*

Fase A. Dalbrefase

Fase A tolkes som nedisningsfasen med dalbreutløpere fra høgfjellsområdene i V og NV (Fig. 1). De få spor som foreløpig er knyttet til denne fasen finnes i bunnen av hoveddalene og enkelte sidedaler, hvor smeltevann under isavsmeltingen har spylt vekk store bunnmorene-masser, og derved blottet isskurt berggrunn under.

De dalrettede skuringene som er eksponert på grunn av smeltevannsavspyling, er antatt å stamme fra samme isbevegelsesfase fordi de alle før eksponeringen var tildekket av den samme relativt homogene, blågrå basalmorenen. Ved hjelp av morenestratigrafi knyttes denne morenen til en tidlig del av siste nedisningsperiode (Olsen 1979, og in prep.).

Fase B. Dominerende hovedfase

Fase B er den av hovedfasene som har virket med størst glasialgeologisk effekt på landskapet (Fig. 1). Isbevegelsene under denne fasen var lite topografisk avhengig, og var rettet mot SØ over undersøkelsesområdet. Spor etter isen fra denne fasen finnes mer eller mindre overalt, fra nedre deler av dalsidene til de høyeste områdene i regionen.

I deler av hoveddalene og stedvis i viddenivå, har isbevegelsen også under fase D2 vært rettet i sørøstlig retning. Følgelig kan disse to fasene vanskelig adskilles umiddelbart i de nevnte kritiske områder. Dette gjelder f.eks. i store deler av Ø. Gausdal, som er forlengelsen av Mjøsdalen i N, samt i området NØ for Lillehammer. Skuringer fra fase D1 har med sine divergerende retninger gitt grunnlag for å skille mellom skuringer fra fase B og D2 der disse har hatt sammenfallende retninger.

Skuringene fra fase B er som oftest grov, og er dominerende i høydeområdene. Mer enn 50% av morenematerialet i området er trolig avsatt i denne perioden (Olsen 1979).

Fase C. Innlandsisfase

Fase C opptrer med liten glasialgeologisk effekt i området (Fig. 1). Bremassene beveget seg mot SSØ–SSV i hele undersøkelsesområdet. Markert sørlig rettet skuring knyttet til fase C er registrert flere steder. Spesielt tydelig er denne skuringen i de høyeste områdene i N, samt i de aller sørligste deler av området.

Skuringsbildet fra denne fasen er meget komplisert med tilsynelatende motstridende aldersforhold for noen av de målte retningene. Best forenlig med resultatene vil det være om isbevegelsen har dreiet først i vestlig deretter i østlig retning innenfor sektoren SSØ–SSV. En bevegelse mot SSØ kan derfor være både eldst og yngst innen fase C.

Skuringenes intensitet og grovhetsgrad i forskjellige områder med tilnærmet samme berggrunn og topografi, antyder at det har vært varierende betingelser for skuringen under fase C. I området mellom Ø. Gausdal og Gudbrandsdal foreligger eksempelvis skuring rettet mot SSØ–S som tynn til middels grov skuring. Det samme gjelder i traktene N–NØ for Lillehammer. Sør for Saksumdal opptrer sørlig rettet skuring tolket til fase C i hovedsak som middels grov til grov skuring.

Isskillet lå trolig rettet tilnærmet Ø–V over Gudbrandsdalen under innlandsisfasen (Garnes & Bergersen 1980). I deler av fasen kan isdelersonen ha strukket seg sørover mot Lillehammer-traktene (Fig. 2, Vorren 1977). Dette kan forklare eventuelle regionale variasjoner i skuringsbildet, samt den relativt beskjedne effekt innlandsisfasen har virket med i utformingen av landskapet.

Det observasjonsmaterialet som til nå er innsamlet gir ikke grunnlag for

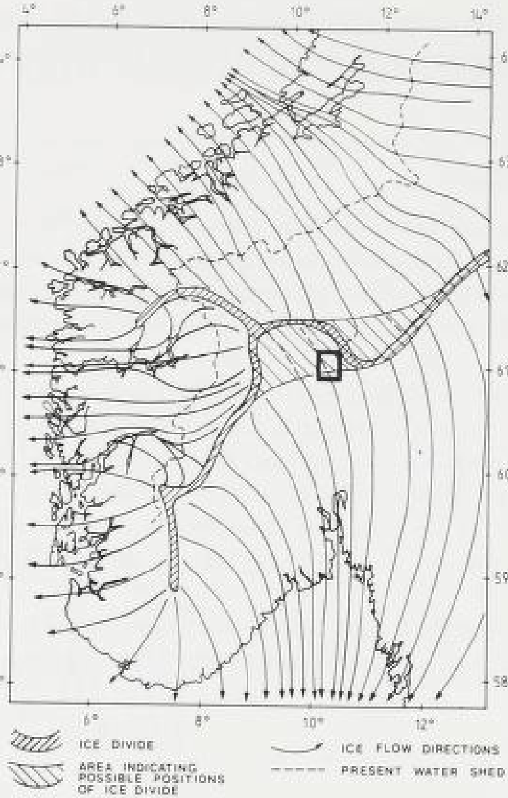

Fig. 2. Posisjon for isskillet og isbevegelsesmønsteret under fase C, som sannsynligvis daterer seg fra 18 000–20 000 år B.P. Isskillet ligger i en betydelig avstand øst og sør for vannskillet. Lillehammerområdet er innrammet med tykk strek. Etter Vorren (1977: fig. 3).

Position of ice divide and ice-movement pattern during Phase C, which probably dates from 18 000–20 000 yrs. B.P. The ice divide is situated a considerable distance east and south of the water-shed. The Lillehammer area is framed with thick lines. After Vorren (1977: Fig. 3).

entydige konklusjoner om isdelerens beliggenhet under hele innlandsisfasen. Vorren (1977: Fig. 3) antyder at isdelersonen kan ha dekket Lillehammerområdet under innlandsisfasen. Dette sistnevnte problemet kan eventuelt bare løses ved detaljerte stratigrafiske studier kombinert med skuringsanalyser, samt en utvidelse av undersøkelsesområdet mot N.

Morenemateriale korrelert til fase C kan lokalt nå opp i minst 6–7 m mektighet (Olsen 1979, og in prep.). Mektigheten er liten sett i forhold til varigheten for isfasen, som antas å ha vart fra før siste istids maksimum (Garnes 1979) og til Preboreal tid.

Fase D1. Deglasiasjonsfase

I denne fasen beveget ismassene seg stort sett i retning ØSØ over området (Fig. 1). Skuringene som knyttes til fasen er i hovedsak tynne, og har retninger som ligger i eller rundt sektoren Ø–ØSØ. Ismassene var fremdeles relativt mektige over området under første del av fase D1. Store ismasser kunne derfor bevege seg langs nedre del av Ø. Gausdal, tvers på Mjøsdalen og videre innover vidda mot Ø–ØSØ. Brebevegelsen mot ØSØ–Ø ville ikke vært mulig uten et større istrykk fra Øst-Jotunheimen enn fra Gudbrandsdalsregionen.

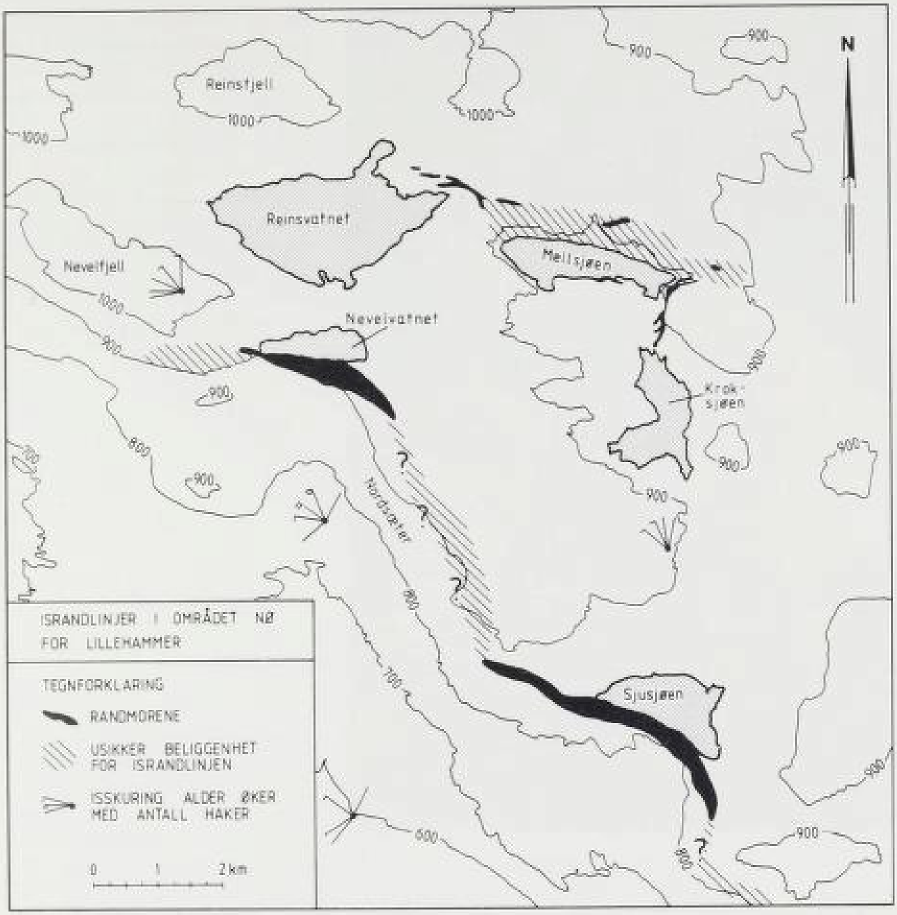

Fig. 3. Israndlinjer NØ for Lillehammer. Nøkkelkart: se Fig. 1. Sjusjøen morenen og Nevelryggen er markert med de største svarte feltene.
Ice-marginal zones NE of the town of Lillehammer. Location map: see Fig. 1. The largest black fields in the Sjusjøen–Nevelfjell area indicate the Preboreal Sjusjøen and Nevelryggen Moraines.

I området NØ for Lillehammer er det registrert randmorener/lateralmorener, som tidligere ble knyttet til en isbevegelse ut Gudbrandsdalen (Holtedahl, 1953:782-786), men som nå korreleres med fase D1 (Fig. 3). Randmorenene er usammenhengende og opptrer i veksling med andre randfenomener langs isens lateralsone. Moreneryggene antas derfor ikke å være betinget av klimatiske forhold, men av økt bregradient grunnet rask avsmelting i Romerikstraktene og sørlige del av Mjøs-bassenget, samt oppover langs Glåmavassdraget i SØ og Ø, (Holtedahl 1953, Garnes 1978, Olsen 1979, Bargel 1983). Skuringer som assosieres med isbevegelsen under dannelsen av randmorenene var rettet skrått ut mot randsonen de siste 2–3 km fra iskanten. Denne type skuring er registrert helt oppe i 1080 m o.h. (Nevelfjell) (Fig. 3). Isoverflaten over Nevelfjell må derfor ha nådd minst 1110 m o.h. i første del av fase D1.

Fig. 4. Flyttblokk av Bygdin-konglomerat. Lokalitet: ca. 5 km V for kartblad 1817 II, midtre del. Avstand til kildeområdet i V–VNV: ca. 50–55 km.
Erratic boulder of Bygdin conglomerate. Locality: about 5 km W of map-sheet 1817 II, middle part. Distance to the provenance area to the W–WNW: about 50–55 km.

Etter teoretiske betraktninger med en postulert bregradient på ca. 1% har de høyeste toppene i iskulminasjonsområdet i Øst-Jotunheimen stukket opp som nunataker i fase D1 (Olsen 1979). Nunatak-fasen, fase Da, i Øst-Jotunheimen (Garnes & Bergersen 1980) overlapper trolig i tid med fase D1.

Fase D2. Deglasiasjonsfase

Etterhvert som ismassene tynnet ut, ble mindre og mindre is ført fra V inn over viddelendet Ø–NØ for Lillehammer. I selve Mjøsdalen ble isbevegelsen mere dalrettet, og det skjer en suksessiv overgang mot fase D2, fig. 1. I nedre deler av viddeområdet vendte isoverflatens gradient seg inn mot Mjøs-bassenget fra V, N og NØ. Spor etter fase D2 kan tyde på at fase D1 dør ut mens isoverflaten fremdeles når opp til 780 m o.h. ved Nordseter (Olsen 1979).

Lav gradient i delgasiasjonsperioden førte til at store areal i viddeområdet ble isfrie innenfor en kort tidsramme. Skuring og glasitektonikk viser at ismassene fremdeles var dynamisk aktive i dalene når isen forlengst var vekksmeltet eller forelå som isolerte dødisrester i viddenivå. Tilsvarende konklusjoner trekkes av Garnes & Bergersen (1980) om ismassene i Gudbrandsdalen og sidedaler.

Fluted surface og drumlinoide formelementer er registrert og tolket til å være dannet under fase D1/fase D2, i hovedsak under fase D2 (Olsen op.cit.).

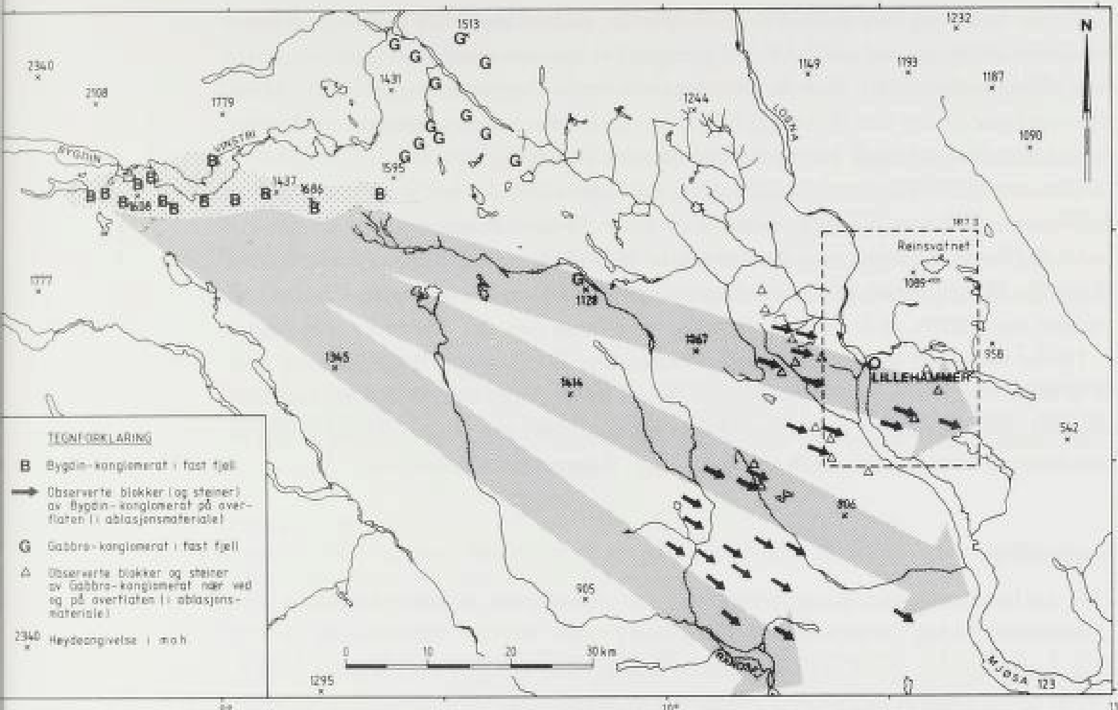

Fig. 5. Blokktransport under unge isbevegelser fra Ø-Jotunheimen. Transportretningene er anvist med store piler. Lillehammer-området er innrammet med stiplete linjer.
Erratics transported during late ice movements from the East Jotunheimen area (Bygdin – Vinstri area on the map). The transport directions are indicated by large arrows. The Lillehammer area is framed with broken lines.
B – Bygdin conglomerate in bedrock. Black arrows – Erratics of B. conglomerate.
G – Gabbro conglomerate in bedrock. Open triangles – Erratics of G. Conglomerate.

Flyttblokkstudier har gitt meget gode informasjoner om materialtransport og isbevegelsesretninger under deglasiasjonsfasene, spesielt under fase D1. Bygdin-konglomeratet som er et karakteristisk deformert kvartsittkonglomerat, står i fast fjell i et begrenset område Ø for Bygdin i Øst-Jotunheimen (Dietrichson 1945). Flyttblokker av Bygdin-konglomeratet (Fig. 4) er registrert i ablasjonsmateriale over et vidt område som strekker seg helt inn i Mjøs-traktene. Nordøst-grensen for spredningsviften til blokkene krysser over Lillehammer området (Fig. 5). Dette antyder en betydelig Ø–ØSØ-lig bevegelses-komponent for ismassene under deglasiasjonsperioden. Transporten av Bygdin-konglomerat blokkene må derfor ha omfattet også første del av perioden, det vil si fase D1.

Fase D2 kan trolig korreleres med fase Db og yngre faser i Øst-Jotunheimen (Garnes 1978, Garnes & Bergersen op.cit., Olsen op.cit.).

Konklusjoner

Isbevegelsesmønsteret i Øst-Jotunheimen og Gudbrandsdalen (bl.a. Garnes 1975, 1978, Garnes & Bergersen 1980), er knyttet sammen med isbevegelsesmønsteret over nordlige del av Mjøs-området.

Det er foreløpig bare funnet noen få spor fra dalbrefasen eller innledningsfasen for siste nedisning over området. Til gjengjeld er sporene etter isbevegelsene under den påfølgende isfasen, fase B, rikt representert i observasjonsmaterialet. Mens isbevegelsene under fase A strengt fulgte hoveddalene, hadde ismassene under fase B en klart dominerende bevegelseskomponent mot SØ (Fig. 1).

Mer enn 50% av morenematerialet i området antas å være avsatt under fase B (Olsen 1979) som således er den dominerende hovedfasen.

Under fase C beveget ismassene seg mot SSØ–SSV over Lillehammer-området (Fig. 1). Mektigheten på moreneavsetningene som er korrelert med fase C er mindre enn ventet ut fra antatt varighet for isfasen (op.cit., Olsen in prep.).

Under første del av deglasiasjonsperioden, fase D1, beveget ismassene seg mot Ø–ØSØ over Lillehammer-traktene. Siste del av deglasiasjonsperiodene, fase D2, er karakterisert av isbevegelser som fulgte topografien i store trekk. Mjøsa fungerte som hoveddreneringsvei både for is og vann i denne sluttfasen.

Summary

The author has tried to correlate the ice movement pattern found in the Lillehammer area, northern Mjøsa district, with the ice movements in the East-Jotunheimen and Gudbrandsdalen region described by e.g. Garnes (1975, 1978) and Garnes & Bergersen (1980). An ice movement model based on striation analyzed at about 250 localities is presented. The model shows 4 major ice movement phases with a subdivision of the youngest phase (Fig. 1).

The oldest phase, A, is assumed to derive from the beginning of the last glaciation (i.e. the Mid-/Late Weichselian glaciation; Olsen 1979). The striation which is correlated with this initial phase has been found under a very compact homogeneous bluish grey till on bedrock surfaces exposed through the action of meltwater erosion. The bluish grey till is correlated with an early period of the last glaciation (Olsen 1979). While the ice movements during phase A strictly followed the main valleys, the ice movements during the next phase, phase B, were directed towards south-east (Fig 1). More than 50% of the materials in the area assumed to derive from phase B which apparently is the dominant ice phase during the last glaciation.

During the inland ice phase C the ice masses moved towards SSE–SSW over the investigated area. At one time during this phase, the ice-shed might have migrated southwards to the Lillehammer area (Fig. 2, Vorren 1977). This investigation can neither reject nor support this hypothesis with conclusive data. The thickness of the tills which correlate with phase C are small compared with the duration of the ice phase, assumed to be from before the maximum extension of the last glaciation (Garnes 1979) to the Preboral Chronozone. In some places the thickness of the associated till units is 6–7 m or more, but more frequently the thickness is much less (Olsen in prep.).

In the first part of the deglaciation period, phase D1, the ice masses moved towards E–ESE over the area (Fig. 1). The first nunataks in this area penetrated through the ice cover during phase D1, or maybe just before. Before the transition

to the last part of the deglaciation period, phase D2, the ice margin was situated in the area between Sjusjøen and Nevelfjell (Fig. 3). The Nevelryggen Moraine and the Sjusjøen Moraine are lateral moraines (Holtedahl 1953) which probably derive from this time.

Ablation till containing erratics transported from the East Jotunheimen area across the valley of Lake Mjøsa (Figs. 4 and 5), clearly demonstrates the eastward directed ice movements during the first part of the deglaciation of the area. The ice movements during phase D2 followed the overall topography. Lake Mjøsa was the major drainage channel for both the ice and the meltwater in this final stage of glaciation.

Etterord. – Forfatteren ønsker å rette en takk til Ole Fredrik Bergersen for hovedfagsveiledning, til Bjørn Bergstrøm, Oddvar Longva og Jan Mangerud som har lest kritisk gjennom manuskriptet, og til David Roberts som har hjulpet til med den engelske teksten. Gunn Sandvik har tekstbehandlet manuskriptet, Irene Lundqvist har rentegnet figurene, og Lars Holiløkk har utført repro-arbeidet.

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Morenestratigrafi og isbevegelser fra Weichsel, sørvestre Finnmarksvidda, Nord-Norge

LARS OLSEN & MARTIN HAMBORG

Olsen, L. & Hamborg, M. 1983: Weichselian till stratigraphy and ice movements on southwestern Finnmarksvidda, northern Norway. *Norges geol. Unders. 378*, 93–113.

At least 6 different till units exposed in a 50 m high section along the Vuolgamasjåkka river. Starting from the top there are three basal tills resting on sorted sediments, and below that there are tills alternating with sorted sediments. At the bottom the exposed sequence ends with sorted sediments. The ice movements corresponding to the different till units are interpreted based on stone orientation. The three upper tills were deposited under ice moving towards the N (youngest), NNW (intermediate), and NNE (oldest), respectively. Preliminary results indicate that the two oldest till units were deposited as a result of ice moving towards the NE and NW, respectively. The two oldest of the three upper tills are found in many sections in the area. At two localities in Kautokeino, at Vuolgamasjåkka and at Vuoddasjavri the till bed complex rests on glacifluvial sediments. The sediments may possibly represent the beginning or the end of an interstadial which we have, informally, named the Eiravarri interstadial. The basal tills which correspond to the youngest regional ice movement, having mainly a northerly component, are only occasionally found. We assume that the two underlying till units represent the last glaciation from the ice advance after the last interstadial (Eiravarri interstadial) to the deglaciation after maximum ice extension. Reorientation of particles in the underlying till with regard to the youngest or the second youngest regional ice movement can be demonstrated at 5 of the 6 described localities on Finnmarksvidda. Without organic content in the subtill sediments and without any indications of ice movements from the mountain range to the northwest, we assume that the exposed part of the stratigraphical sequence at Voulgamasjåkka is younger than the last interglacial (Eem).

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Innledning

Artikkelen gir en oversikt over resultatene fra kvartærstratigrafiske undersøkelser på SV-lige Finnmarksvidda 1981 (Fig. 1) i forbindelse med malmprospektering i regi av A/S Sydvaranger og Norges geologiske undersøkelse. Formålet med undersøkelsene er å finne de regionale isbevegelsene og de tilhørende stratigrafiske enheter, for videre å benytte dette i malmprospekteringen. Resultatene er foreløpige, men er valgt å presenteres på et tidlig stadium som en informasjon og som et diskusjonsgrunnlag for videre arbeider i dette og tilstøtende områder.

Av de beskrevne lokalitetene framheves skjæringen ved Vuolgamasjåkka. Den har sannsynligvis den mest komplette stratigrafien (Fig. 1, 8). Med denne stratigrafien som utgangspunkt er lokalitetene forsøkt knyttet sammen i et korrelasjonsdiagram (Fig. 9).

Alle isbevegelsesretninger angis i nygrader ^{g} i breens bevegelsesretning.

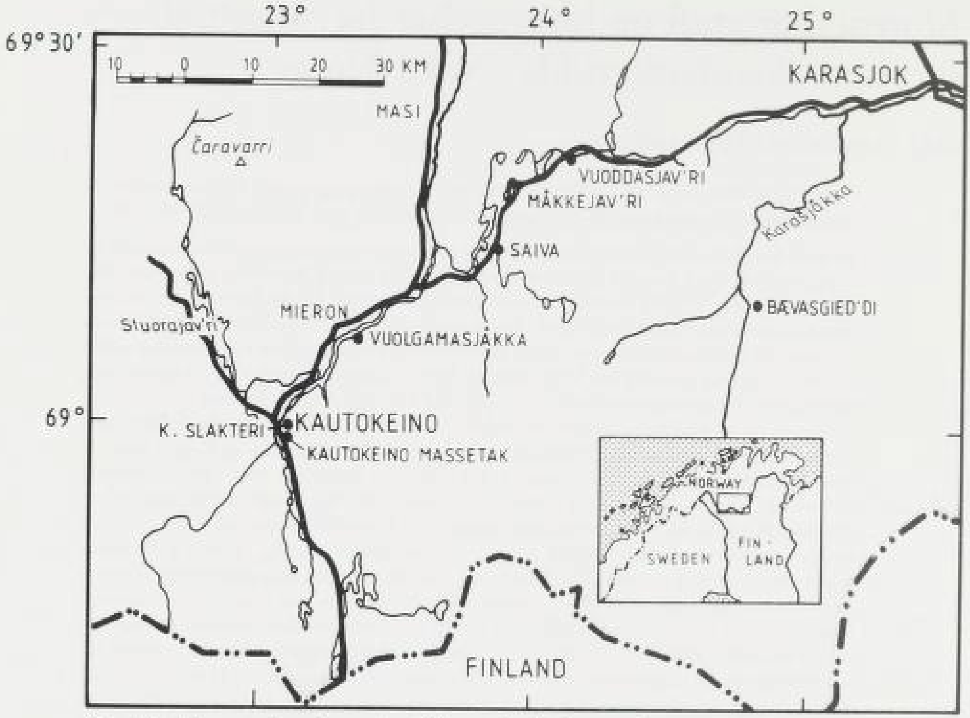


Fig. 1: Lokalitetene hvor de stratigrafiske undersøkelsene er foretatt. Ved Bæivasgieddi ligger en stratigrafisk lokalitet som undersøkes av A. Read, Univ. i Oslo.
Location map of stratigraphical sites. A site not far from Bæivasgieddi is currently being investigated by A. Read, Univ. of Oslo.

Steinorienteringer

Det er ikke på noen av lokalitetene funnet organisk materiale som kunne egne seg til relative eller absolutte dateringer. Pollen er heller ikke funnet i sedimentene. Det er derfor ikke funnet grunnlag for å opprette en kronostratigrafi, men på bakgrunn av lithologi og steinorientering (s.o.) i morener har vi forsøkt å etablere en lokal klimastratigrafi. Undersøkelsene blir etterhvert komplettert med petrografiske analyser og kornfordelingsanalyser.

Topografisk sett gir viddeområdet et særlig gunstig utgangspunkt for bruk av steinorienteringer som morenestratigrafisk metode. Det er lavt relieff og rolig undulerende overflate over et stort areal. Isdelersonen under siste nedisning lå hele tiden sør for området. Under slike forhold vil en forvente at regionale isbevegelser gir regionalt ensartet orientering i de respektive morener. Resultatene fra våre analyser bekrefter dette.

I snittene er orienteringene alltid tatt på horisontale flater hvor morenen er homogen og uten blokker. Hver orientering er forsøkt begrenset til å dekke minst mulig i vertikal retning. Størrelsen på partiklene som måles, varierer etter hvilken clast-fraksjon som dominerer i morenene. For å unngå avhengige (sympatetic) orienteringer har vi forsøkt å holde et snevert størrelses-intervall for partiklene ved hver analyse. For de fleste målte partiklene har lengste akse vært ca. 2,5–6 cm, og forholdet mellom lengste og mellomste akse er alltid større enn 1,5:1 og mindre enn 2,5:1. Under bestemmelsen av retning og fall på lengste akse er skjønnsmessig vektaksen i partikkelen benyttet. På et fåtall partikler har dette vært vanskelig å bestemme. Disse er ikke tatt med i undersøkelsen. Steiltstående partikler som heller mer enn 40^g er registrert, men ikke benyttet i analysen. Dette er i tråd med tester av operatørfeil og andre målefeil utført for steinorienteringsanalyser (f.eks. Krüger, 1970, 1973).

Parallelliteten mellom orienteringsmaksimum for partiklenes lengste akse og isbevegelsesretning, er etterhvert meget godt dokumentert for basalmorener (Goldtwait, 1971). Det er likevel mange moreneundersøkelser som viser at en orientering på tvers kan dominere. Slike orienteringer er vist å opptre i trykksoner i marginale deler av ismassene og i støt for obstruksjoner i terrenget (Boulton, 1971). Våre lokaliteter omfattes så vidt vi vet, ikke av slike forhold. Transversale orienteringsmaksima kan også opptre der hvor steiner med tavle- og kileform med a:b akseforhold mindre enn 1,5, diskosformer med a:b akseforhold mindre enn 2–1, 5, og lange stenger er overrepresentert i materialet (Andrews & King, 1968). Med vårt krav til akseforhold forsvinner de to første form-klassene, og de to siste reduseres til et minimum. Vi bruker derfor steinorienteringer til å bestemme isbevegelsesretningene på stedet.

Karakteristisk for steinorienteringene er de doble orienteringsmaksima som generelt opptrer i øvre del av de underliggende morener (Fig. 2: s.o. 28B, 2). Vi har ikke undersøkt spesielt hvor langt ned i morenen disse doble maksima opptrer, men dybden synes å være av størrelsesorden ca. 30 cm. Det er grunn til å anta at disse doble maksima skyldes reorientering under yngre isbevegelser.

Fra sub-marginale deler av recente breer har Boulton (1979:29) påvist deformasjon 0,6 m ned i underliggende morene. Målingene viser at deformasjonen avtar etter en tilnærmet hyperbolsk funksjon ned til nevnte dyp.

MacClintock og Dreimanis (1964) har vist ved steinorienteringer (fabric analyser), at is av kontinental tykkelse kan forårsake reorientering i ufrossete underliggende avsetninger til et dyp på minst 10 m. De samme forfatterne foreslår tre mulige mekanismer som gir reorientering:

1. Morenen forflyttes langs assosierte relativt inkompetente leir- og siltlag.
2. Morenen utsettes for skjærpåvirkning. Slike skjærplan er sjelden mulig å registrere unntatt på steder der leire og silt er involvert i skjærprosessen (op.cit.: 141).
3. Den tredje mekanismen blir forklart med en viskøs strømning som kan oppstå når skjærkreftene omlagrer eller elter morenen, og omdanner den til en seigtflytende masse.

Sistnevnte hypotese er i tråd med Glen, Donner og West's (1957) forklaring på hvordan partikkelorienteringen oppstår ved skjærbevegelser i isens basaldeler. MacClintock og Dreimanis har imidlertid ikke påvist noen tilfeller der denne deformasjonen har skjedd.

Ramsden og Westgate (1971) har undersøkt fabric i doble morener i Alberta, Canada. De mener å ha påvist eksempler på reorientering som skissert under pkt.

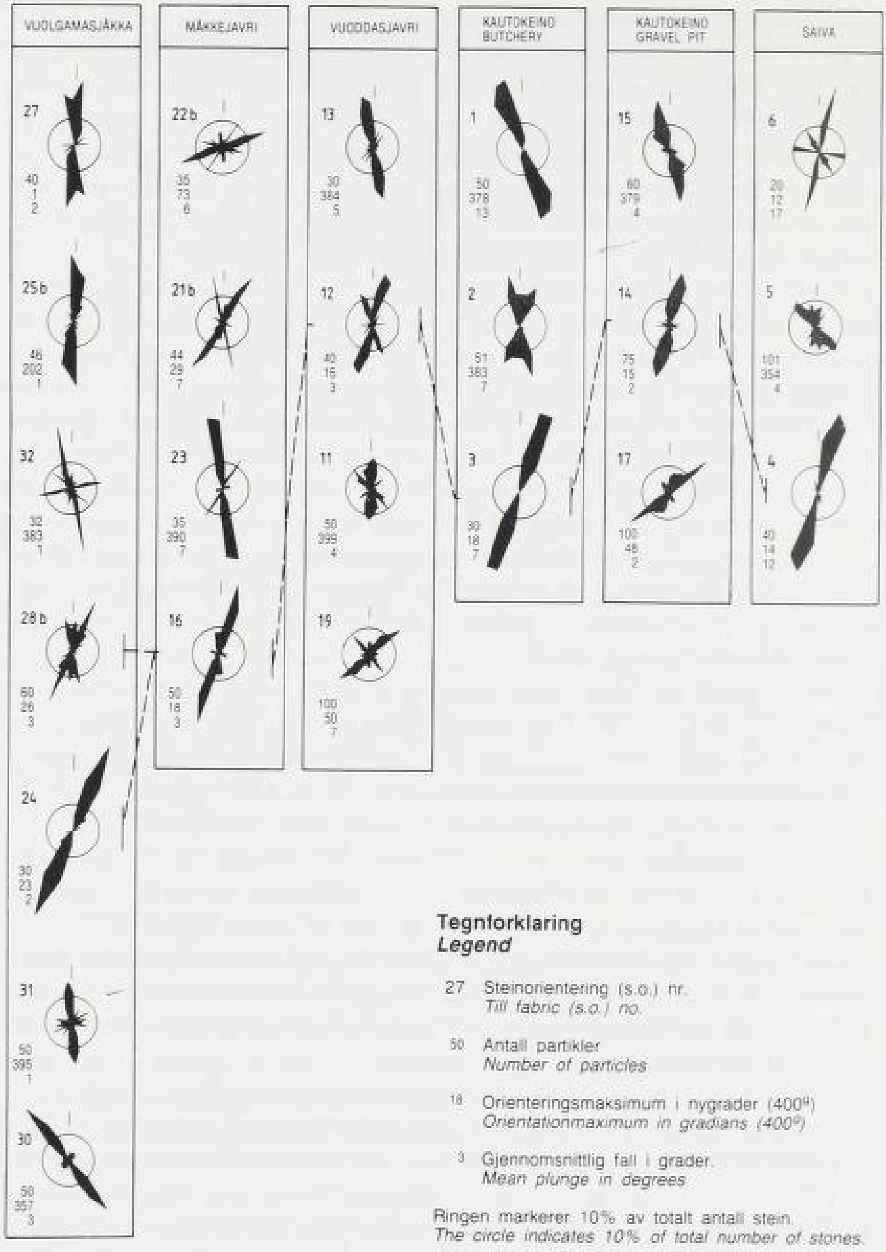


Fig. 2: Steinorienteringer. Deres stratigrafiske posisjon er avmerket i Fig. 8 og 9: Hovedmaksimum er beregnet etter formen på rosediagrammet. Gjennomsnittlig fall er beregnet innenfor sektoren som utgjør hovedmaksimum. Transportretningen for morenen er angitt i nygrader (^{g}), basert på orienteringsmaksimum og at det gjennomsnittlige fall på partiklene er det motsatte av bevegelsesretningen. Mellom steinorienteringer som er korrelate er det trukket en stiplet linje.

Till fabrics. Their stratigraphical locations are marked on Figs. 8 and 9. The primary maximum is calculated based on the shape of the mirror image rose diagram. Mean plunge is calculated for the dominant modal group, i.e. the maximum class plus adjacent classes. The till transportation direction is indicated in gradians (^{g}), based on the primary maximum orientation and the up-glacier plunge principle. A broken line is drawn between fabrics which correlate with till D.

1 og 2. På alle steder der det er funnet reorientering i undre morene, er denne transportert *en bloc* under den yngre isbevegelsen ved skjærbevegelser i den basale leirige sonen (op.cit.: 341). Reorienteringene er registrert ved doble maksima i øvre del av underliggende morene. Det ene maksimum er parallelt med maksimum i overliggende morene, det andre med maksimum i undre morene. Lignende resultater har kommet fram i analysemateriale fra Finnmarksvidda. Ved våre analyser framstår reorienteringen noen steder som hovedmaksimum (Fig. 2; s.o. 2 og 5), andre steder er det omvendte tilfelle (Fig. 2: s.o. 12 og 28b).

Ramsden & Westgate (1971: 343) har i sin undersøkelse påpekt at avlange partikler som skjæres av et av skjærplanene som knyttes til den yngre isbevegelsen, har en tendens til å bli reorientert til den yngre isbevegelsesretningen. Dette mener vi har stor betydning for tolkningen av analysedata fra underliggende morener som er tett gjennomsatt av skjærplan assosiert med yngre isbevegelser.

Beskrivelse av stratigrafiske lokaliteter

Det er undersøkt seks lokaliteter som ligger tilnærmet i et ØNØ–VNV profil fra Kautokeino mot Karasjok (Fig. 1). De fleste lokalitetene er framkommet ved uttak av masse til veibygging. Ulempen med dette er at skjæringene ikke er blitt stående i lengre tidsrom, men lukket igjen når behovet for masse avtok. Tilfredsstillende undersøkelser vanskeliggjøres på denne måten.

I tillegg til de beskrevne lokalitetene arbeider A. Read, Geol. inst., U. i Oslo (pers.medd. 1981), med en lignende stratigrafi ved Bæivasgieddi (Fig. 1).

Steinorienteringenes plassering i dybde og stratigrafi sees i Fig. 9.

KAUTOKEINO SLAKTERI

Her ligger to morener adskilt av en 2–8 cm tykk diamikton av brunlig farge (Fig. 3). Under morenene i en forsenkning i fjelloverflaten ligger en 0–0,3 m tykk lomme med sorterte sandige sedimenter. Eventuelle primære strukturer i dette sedimentet er ødelagt. De forskjellige enhetene er alle adskilt av skarpe erosjonsgrenser.

Den undre morenen er 1–1,5 m tykk og har en sterk orientering mot ca. 20^g (Fig. 2: s.o. 3). Like under grensen til den overliggende morenen, dvs. umiddelbart inder den brune diamiktonen, viser rosediagrammet (s.o. 2) et sekundært maksimum på ca. 25^g og et dominerende maksimum på $380–390^g$. Den stratigrafiske posisjon til steinorientering 2, ca. 25 cm under erosjonsgrensen, indikerer at isbevegelsen som assosieres med avsetningen av den øverste av morenene, under denne prosessen delvis reorienterte de øverste partier av den underliggende morenen. Den eldste bevegelsen framkommer således som et sekundært maksimum.

Overliggende morene er ca. 1,2 m tykk og orienteringen viser isbevegelse mot 380^g.

Fig. 3: Kautokeino slakteri. Grensen mellom undre (20^g) og øvre (380^g) morene er stiplet. Mellom morenene ligger en 2–8 cm tykk brunfarget diamikton.
Kautokeino butchery. The boundary between lower (20^g) and upper (380^g) till is marked with a broken line. A 2–8 cm thick brownish diamicton separates the two tills.

Fig. 4: Kautokeino massetak. Grensen mellom undre (15^g) og øvre (380^g) morene er stiplet. Underst i snittet ligger deformerte glasifluviale sedimenter.
Kautokeino gravel pit. The boundary between lower (15^g) and upper (380^g) till is marked with a broken line. The basal tills rest on deformed glacifluvial sediments.

KAUTOKEINO MASSETAK

Partier av avsetningen er tektonisert i flere deformasjonsfaser.

Forkastninger og overskyvninger viser deformasjon fra SV–SSØ. Aldersforholdet mellom deformasjonsfasene basert på tektonikken er ikke tilstrekkelig undersøkt.

Steinorienteringene er foretatt i deler av avsetningen som tilsynelatende er uforstyrret.

Nederst i grustaket ligger en mer enn 8 m mektig glasifluvial lagserie (Fig. 4). I lagserien ligger flere 0,5–2 m mektige lag av morenisert glasifluvium. De underste 3 m av sedimentet viser en mer eller mindre rytmisk lagfølge av sand og grusig sand i tilnærmet flattliggende lag. Regelmessigheten er stedvis forstyrret av steinholdige lag og yngre deformasjonsbevegelser. Underlaget er ikke blottet. Variasjonen i kornstørrelse og sortering oppover i sedimentet er stor. I enkelte partier består massene av godt sortert sand. I andre partier vitner sedimentet om svært brenære forhold med ekstremt dårlig sortering i en blanding fra blokker til sand. Skrålagning og kryss-sjiktning viser at avsetningen skjedde fra en sørøstlig retning.

Over sanden følger en 1–6 m mektig undre morene som i nedre deler har en klar steinorientering mot ca. 50^g (Fig. 2, s.o. 17). Lenger opp i den samme morenen viser s.o. 14 at bevegelsen er mot ca. 15^g.

Den nordøstlige orienteringen i basaldelen av morenen kan skyldes lokale dalrettede isbevegelser eller en regional isbevegelse da breen rykket fram fra en sørlig posisjon.

Skjærplan og deformerte sediment-bånd som kan følges ned i den underliggende sand markerer grensen til den øvre morenen. Denne morenen er 0–3 m tykk og orienteringen (s.o. 15) indikerer en bevegelse mot ca. 380^g.

Øverst ligger 0–1 m med ablasjonsmateriale og bresjøsedimenter fra isavsmeltingen.

VUOLGAMASJÅKKA

Stratigrafien er her blottlagt i en ca. 50 m høy elveskjæring (Fig. 8). Undersøkelsene er foreløpig ufullstendige og må betraktes som orienterende. Steinorienteringene som er foretatt gir imidlertid en foreløpig oversikt over (hoved)isbevegelsene innen de enheter hvor disse er tatt.

Av spesiell interesse er orienteringene i de basale deler av morenene som ligger over sedimentene. Man kan her få en indikasjon på fra hvilken retning breen rykket fram etter en interstadial eller interglasial.

Den laveste delen av snittet er dekket av nedrast materiale. Underst i den blottede delen av snittet ligger mer enn seks meter med sorterte sedimenter, sand J. Lagning og kryss-sjikting viser transportretning fra sør. Sedimentet er sannsynligvis en proglasial dannelse.

Den overliggende morenen, I, er ca. 3 m mektig. Den har et tilnærmet horisontalt orienteringsmønster med en klar orientering ca. 155^g $\pm$ 200^g (Fig. 2, s.o. 30). Dragfolder i grensen med sand J viser entydig at isbevegelsen var mot ca. 355^g (Fig. 5). Dette kan være en lokal isbevegelse for området omkring

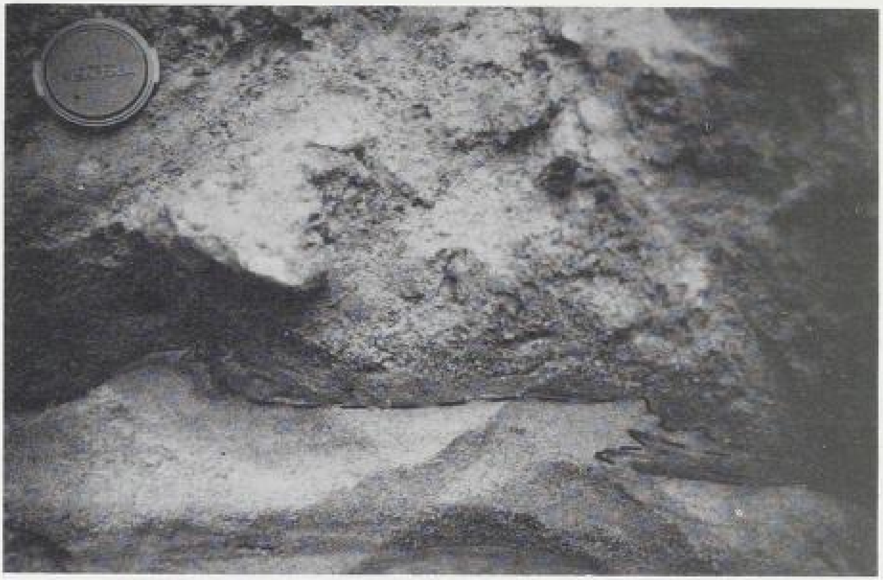


Fig. 5: Vuolgamasjåkka. Dragfolder i overgangen mellom morene I (355^g) og sand J viser en deformasjon fra venstre mot høyre (SØ mot NV).
Vuolgamasjåkka. Drag folds in sand J just beneath till I (355^g) indicate a deformation phase directed from left to right (SE to NW).

lokaliteten, eller den representerer en noe mer regional isbevegelse som viser isens framrykning i området fra en S–SØ-lig beliggenhet.

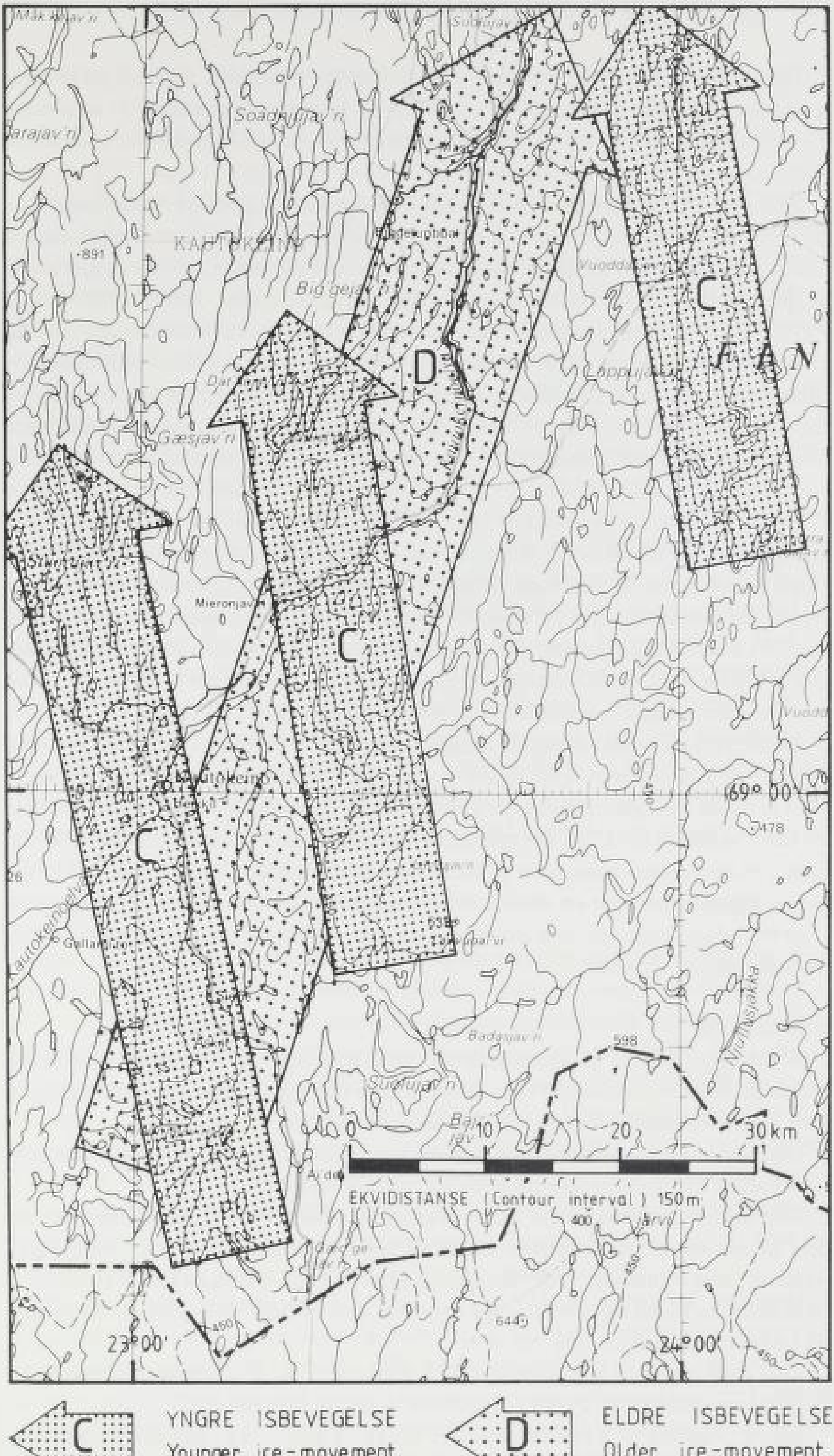
Mellom morene I og den overliggende morenen, G, ligger det en 0,5–1 m tykk pakke med sand og grus, H, uten sedimentære strukturer.

Morene G er ca. 4,5 m mektig. En orientering ca. 1,5 m over undergrensen viser en transportretning mot ca. 395^g (s.o. 31).

Morene G avgrenses oppover av en ca. 0,5 m tykk sand og grus uten sedimentære strukturer. Denne sanden har vi latt gå inn i den overliggende morenepakke F, som gjennomskjæres av flere grus- og sandlag. Felles for de sorterte sedimentbåndene er at de ikke har sedimentære strukturer. Det er heller ingen visuelle forskjeller i den totalt ca. 9 m mektige moreneserien.

Over morenen ligger ca. 12 m med sorterte sedimenter dominert av sand. Sanden er bygget opp av skrålag som er avsatt fra sør. I den sørlige delen av snittet veksler lagene mellom grov grus og sand, mens den nordlige, distale delen har en jevn veksling av sortert sand. Sedimentet er antatt å være en glasifluvial dannelse, avsatt i en isfri periode før den overliggende morenepakken ble avsatt. Vi kaller denne sanden for Eiravarri Sand. Den tilsvarende isfrie perioden kaller vi uformelt Eiravarri interstadial.

Den øvre morenepakken kan deles i tre. Den undre morenen, D, er mellom 0,5 og 1,5 m tykk. En steinorientering midt i morenen viser en isbevegelse mot ca. 25^g (s.o. 24).



Soadnjujav'ri
KAHTOKEINO
Biggejav'ri
Gæsjav'ri
Mieronjav'ri
D
C
C
C
F I N
69° 00'
Badasjav'ri
Suolujav'ri
0
10
20
30 km
EKVIDISTANSE (Contour interval) 150 m
23°00'
24°00'
C
YNGRE ISBEVEGELSE
Younger ice-movement
D
ELDRE ISBEVEGELSE
Older ice-movement

Den undre morenen er mer enn 3,5 m mektig, homogen, normalt pakket og steinene er orientert mot ca. 15^g (Fig. 2, s.o. 4). En steinorientering (5) i de øverste 20 cm av morenen viser at orienteringen her er ca. 355^g, med to sekundære maksima mot ca. 385^g og ca. 15^g.

Den øverste morenen er ca. 1 m tykk, og preget av forskjellige typer flytestrukturer og sedimenter. Vi tolker denne morenen son en basal utsmeltingsmorene, basert på beskrivelser av denne morenetype (f.eks. Dreimanis, 1976). Strukturene og lithologien har karakteristika som Sveg till (Lundquist, 1969, 1977; Shaw, 1979). Orienteringen i morenen viser et dominerende maksimum mot ca. 10^g, og et sekundært mot ca. 360^g. I tillegg er det et lite maksimum ved ca. 310^g, tilnærmet normalt på hovedbevegelsen 10^g. Steinorienteringen er tatt i et «flow» lag ved bunnen av øvre morene og behøver ikke være representativt for moreneenheten i sin helhet (Boulton, 1971: 54). Fallet på a-aksen er stort for partiklene langs de to maksima. Gjennomsnittsfallet er beregnet til henholdsvis 17° og 25°. Bare 20 partikler er målt i dette tilfellet, men alle målingene ligger samlet rundt ett maksimum. Resultatet er av den grunn trolig representativt lokalt i den undre delen av morenen. Det relativt høye fallet på partiklenes a-akse er typisk for bunndelene av en flow till (Boulton, 1971:41), men er også forenlig med en utsmeltingsmorene, «melt-out» till, som ofte avspeiler den englasiale fabric (op.cit.:61). Relativt høyt fall på partiklenes lengste akse er for øvrig påvist i undre deler av basale utsmeltingsmorener av typen Sveg tills (Shaw, 1979:413).

På denne lokaliteten mangler moreneenheten som normalt korreleres med den NNV-lige isbevegelsen som følger etter $15–20^g$ bevegelsen. Steinorientering 5 viser at denne bevegelsen kan være gjenspeilet i form av en reorientering i de øverste 20 cm i morenen hvor 20^g bevegelsen er dominerende. I analyseresultatene finnes også 15^g bevegelsen igjen som et sekundært maksimum. Uten å ha NNV-bevegelsen representert som en moreneenhet mener vi allikevel at den er representert i stratigrafien som en reorientering av den opprinnelige orientering i den underliggende morene.

MÅKKEJAVRI

I en drumlin som er orientert mot ca. 6^g er det laget en skjæring normalt på lengdeutstrekningen. Snittet viser at drumlinen er bygget opp av tre bunnmorener som igjen overlagres av et tynt utsmeltningsmateriale.

Den undre morenen er mer enn 3 m mektig, «normalkonsolidert» og homogen. Den har en klar orientering mot ca. 20^g (Fig. 2, s.o. 16).

Den midtre morenen er ca. 1,5 m tykk og har ellers samme karakteristika som den undre morenen. Morenen avsluttes med et blokklag øverst. Blokklaget framtrer tilsynelatende som en overgangssone til overliggende materiale. I underkant av blokklaget er steinene orientert mot ca. 390^g (s.o. 23), med et svakt sekundært maksimum mot $30–40^g$.

Den øvre morenen som overlagrer blokklaget er ca. 2 m mektig, er lysere av farge, løsere pakket og utgjør den øverste delen av drumlinen.

Like i overkant av blokklaget er det bevart et sekundært maksimum mot ca. 385^g, mens den dominerende retningen nå er ca. $30–40^g$ (s.o. 21b). Midt i denne morenen er det et sekundært maksimum på 30–40^g, mens den dominerende retningen er mellom 70 og 80^g (s.o. 22b).

Vi tolker denne utviklingen slik at den undre morenen er avsatt av en isbevegelse mot ca. 20^g. Deretter endret bevegelsen seg til ca. 390^g.

Dannelsen av blokklaget kan sannsynligvis knyttes til omleggingen av isbevegelsen. I dette tilfellet fra 390^g til en yngre NØ-lig (eventuell N-lig) bevegelse. Blokklag i morener behøver ikke alltid bety en endring av isbevegelsesretningen (Dreimanis, 1976; Hole & Bergersen, 1981), men er ofte knyttet til en overgangssone mellom to isfaser (Garnes & Bergersen, 1977). Det sekundære maksimum på 385^g i s.o. 21b antyder at deler av morenen ble avsatt av denne bevegelsen, men ble siden reorientert av isbevegelsen(e) i påfølgende fase. Tilsvarende reorienteringer i drumliner er registrert andre steder (Gravenor & Meneley, 1958). Det er heller ikke uvanlig å finne eldre løsavsetninger representert i undre deler av drumliner (f.eks. Garnes, 1976; Andersen et al., 1981).

Tverrsnittet i drumlinen ved Måkkejavri viser at de to undre morenene inklusive blokklaget er en erosjonsrest fra eldre isfaser, mens den øvre morenen kan se ut til å være akkumulert i forbindelse med dannelsen av drumlinen. Det er imidlertid uklart om de yngste orienteringene er lokale bevegelser innen dannelsen av drumlinen, eller om de representerer egne regionale bevegelser på skrå av drumlinens lengdeutstrekning. Disse unge NØ-lige bevegelser er imidlertid ikke gjenfunnet i andre tilsvarende unge avsetninger eller som skuringsstriper. Dypet som undersøkelsene er gjort på, ca. 2–3,5 m, gjør det lite sannsynlig at drumlinformen skulle blitt bevart under påvirkning av en sterkt reorienterende diagonaltgående bevegelse. Foreløpig betrakter vi derfor disse NØ-lige orienteringene som uttrykk for de lokale transportretningene under oppbyggingen av drumlinen.

VUODDASJAVRI

Under ablasjonsmateriale ligger to moreneenheter over sorterte sedimenter (Fig. 7). Under sedimentene ligger nok en morene, derunder sorterte sedimenter.

Den underste pakken med sorterte sedimenter er mer enn seks meter mektig og veksler i kornstørrelse og sortering. Underlaget er ikke blottet. De underste delene er sandige og godt sortert. Lagningen er horisontal og kryss-sjiktning viser transport mot nordøst. Oppover øker sedimentene i kornstørrelse. En 0,5–1 m mektig skrålagspakke av grus markerer en oppgrunningsfase i sedimentasjonssyklusen. De øverste 0,5–1 m av sedimentet består av tilnærmet horisontalt lagdelt siltig sand med enkelte grusige, steinige partier. Sanden er sterkt deformert mot overgangen til den overliggende morenen. Sedimentsekvensen ligger over det laveste viddenivået og er sannsynligvis en proglasial dannelse.

Den underste bunnmorenen er ca. 1,5 m mektig, siltrik og hardpakket. Orienteringen i den øvre delen er ca. 50^g (Fig. 2, s.o. 19). Steinorienteringen viser et sekundært maksimum normalt på hovedretningen. Det er ikke undersøkt om dette relativt betydelige maksimum skyldes partikler avsatt normalt på isbevegelsen eller om dette kan representere en tidlig bevegelse. Gjennomsnittlig helling på disse steinene viser en svakt foretrukket retning mot sørøst. Foreløpig betraktes

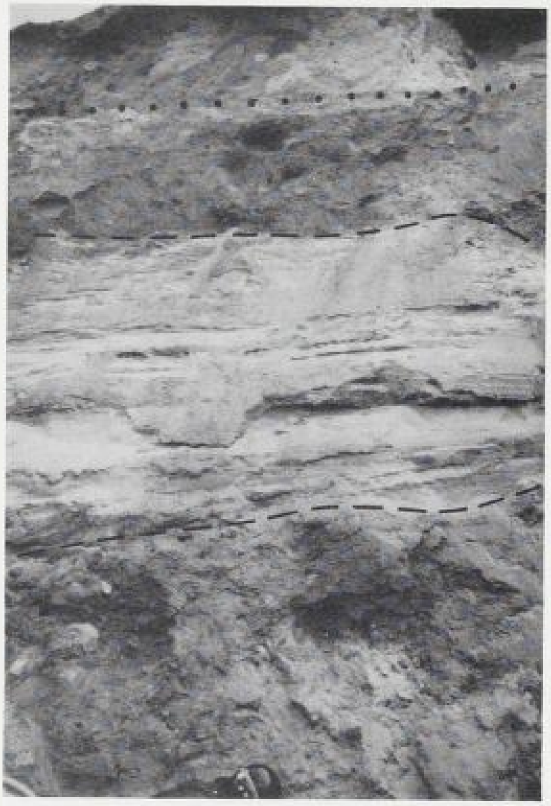


Fig. 7: Vuoddasjavri. I den øvre morenepakken er grensen mellom den undre (15^g) og den øvre (385^g) morenen svært tydelig (prikket linje). Nederst på bildet sees den underste morenen (50^g). Mellom morenepakkene ligger glasifluviale sedimenter (mellom stiplete linjer) med subhorisontal lagning. *Vuoddasjavri. In the upper till sequence the boundary between the lower (15^g) and upper (385^g) till is very distinct. The lowermost till (50^g) is separated from the upper till sequence by a subhorizontally bedded, glacifluvial sediment (between broken lines).*

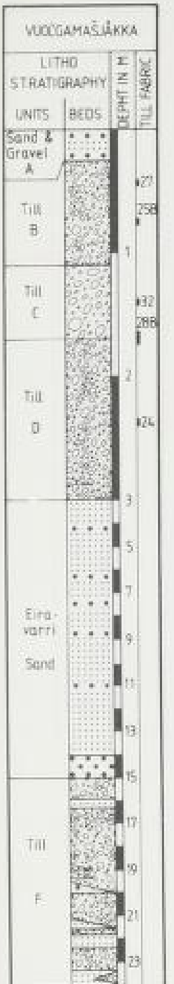
dette sekundære maksimum å representere tverrstilte steiner etter rullende transport.

Den overliggende sedimentpakken er drøye 4 m mektig. Lagningen er tilnærmet horisontal. Avsetningen domineres av sand med enkelte grusige lag som gjennomsetter avsetningen. Kryss-sjikting viser at strømretningen var mot nord. De øverste ca. 0,5 m av sedimentet er usortert med kornstørrelse fra sand til små stein. Vi antar at også denne sekvensen er dannet proglasialt.

Den øvre morenepakken kan deles i to morener etter en tydelig erosjonsgrense. Mektigheten av den undre morenen er ca. 2 m. Enkelte 0–3 cm tykke bånd av sandig grusig materiale gjennomsetter morenen. Orienteringen i basaldelen (s.o. 11) viser et maksimum mellom ca. 380^g og 20^g. Tyngdepunktet ligger på ca. 5^g. I den øvre delen er orienteringen (s.o. 12) dominerende mot 15^g med et sekundært maksimum mot ca. 380^g.

Den øvre morenen er ca. 1,5 m mektig, sandig og stedvis lagdelt i basaldelen. I den undre delen er bevegelsen mot ca. 385^g (s.o. 13).

Denne sekvensen tolkes til å innledes av lokale bevegelser fra sør mot nord i sektoren 380^g–20^g. I den samme moreneenheten forandrer isbevegelsen seg til en regional bevegelse mot ca. 15^g. Deretter er det skifte i bevegelsen til ca. 385^g, og den øverste morenen på denne lokaliteten avsettes.



Det sekundære maksimum (s.o. 12) mot ca. 380^g antyder at bevegelsen som har avsatt den øverste morenen også har foretatt en markant reorientering av den underliggende morenen ned til ca. 25 cm eller mer fra erosjonsgrensen.

Korrelasjoner

Stratigrafien ved Vuolgamasjåkka (Fig. 8) er benyttet som utgangspunkt for et korrelasjonsdiagram mellom de nevnte lokalitetene som bygger på stratigrafisk posisjon, lithologi og steinorientering (Fig. 9).

Det er flere argumenter mot å bruke steinorienteringer i den utstrekning vi har gjort. De regionale variasjoner i steinorientering innen samme moreneavsetning kan være så store at forsøk på å skille mellom morener av forskjellige aldre ved regional prøvetaking er vanskelig (West & Donner, 1956; Boulton, 1971). Dessuten at enkeltstående analyser fra hver lokalitet ofte vil gi misvisende resultater (Boulton op.cit.). Men med de forutsetninger som foreligger, presisert på side 94, mener vi at en kritisk korrelering mellom nærliggende lokaliteter på vidda skulle være mulig. Vi har derfor gjort et forsøk på å bygge opp en lokal stratigrafi der steinorienteringer er korrelasjonsgrunnlaget.

Bokstavinndelingen som er brukt på Vuolgamasjåkka-lokaliteten er benyttet ved korreleringer mellom de forskjellige lokalitetene. Korrelerte lag er gitt samme bokstav.

Ved korreleringen har vi tatt utgangspunkt i morene D, både på grunn av dens karakteristiske orientering og stratigrafiske plassering. Den har en stabil orientering mellom 15^g og 25^g, og den gjenfinnes på alle lokalitetene (sannsynligvis også ved Bæivasgieddi: Fig. 1 (Read, pers.medd. 1981)). Videre overlagrer morene D på fire av lokalitetene sorterte sedimenter. I Kautokeino slakteri er disse deformerte og sannsynligvis para-autoktone. Morene D overlagres på fem av lokalitetene av en moreneenhet som har en orientering mellom 380^g og 390^g. Denne moreneenheten mangler ved Saiva. Her er derimot antatt at den korresponderende isbevegelsen er bevart som en reorientering av morene D, enten i form av et sekundært maksimum mot 385^g eller som hovedmaksimum mot 355^g.

Den øverste morenen, morene B, mangler på flere av lokalitetene. Karakteren på denne morenen varierer noe fra lokalitet til lokalitet, flere steder har den preg av en ablasjonsmorene. Korreleringen er derfor for det meste basert på stratigrafisk posisjon og i noe mindre grad på steinorientering.

Vi regner korrelasjonen mellom de tre øvre morener som relativt sikker. Denne konklusjon baseres i hovedsak på den stabilitet som steinorienteringene viser fra lag til lag og fra lokalitet til lokalitet. Korrelasjonen mellom de underliggende sedimenter bygger på korrelasjonene mellom de overliggende morener, og er derfor i prinsippet en korrelasjon på grunnlag av minimumsalder.

Korrelasjon mellom enheter eldre enn lagpakke E er foreløpig umulig på grunn av manglende opplysninger.

Fig. 8: Stratigrafien ved Vuolgamasjåkka skjematisk framstilt. Meter-skalaen er ikke lineær.
The stratigraphy at Vuolgamasjåkka shown in outline. The metre scale is not linear.

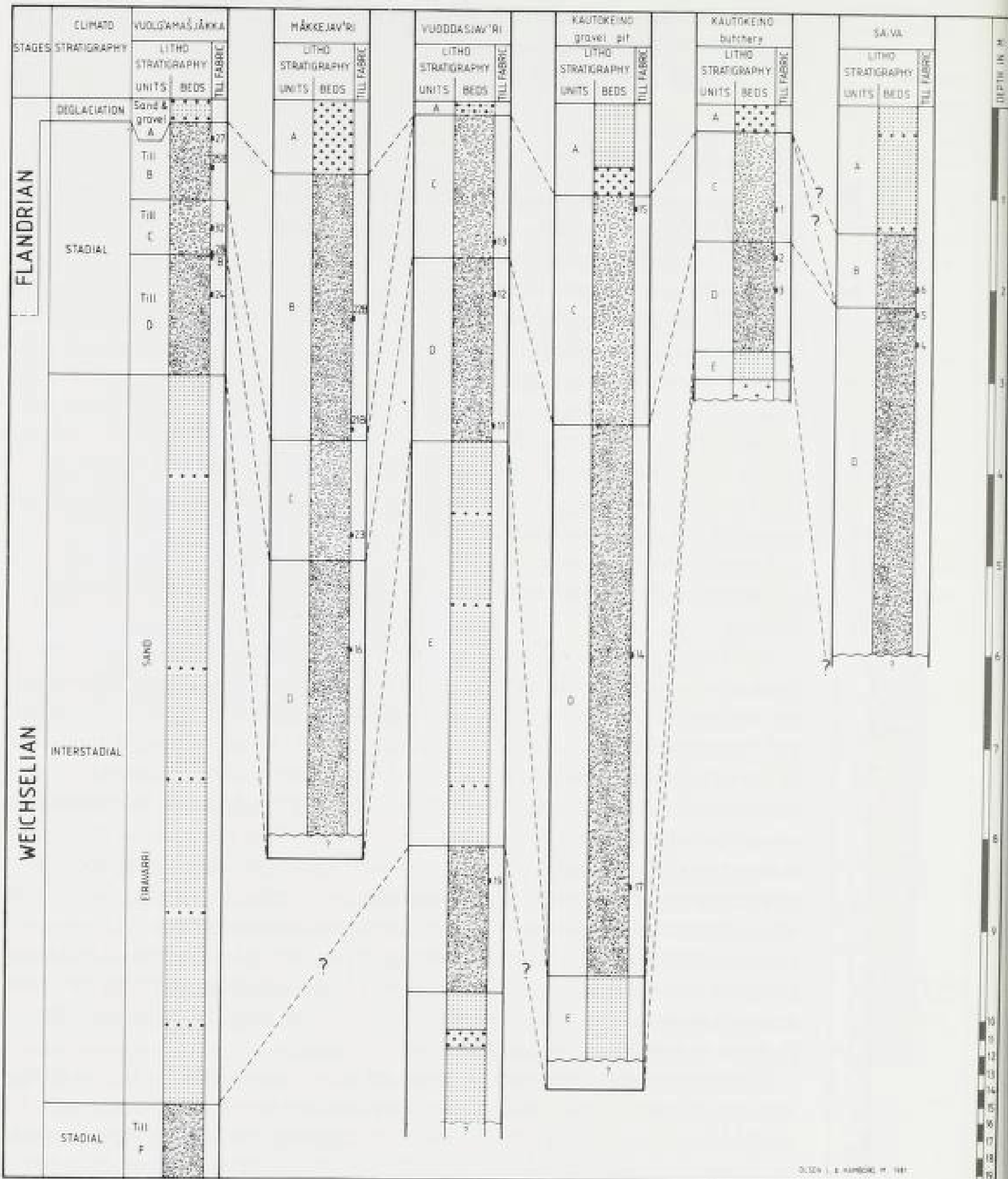


Fig. 9: Korrelasjonsdiagram basert på lithologi, stratigrafisk posisjon og steinorienteringer. Meterskalaen er ikke lineær.

Correlation diagram based on lithology, stratigraphical position and till fabric. The metre scale is not linear.

På fem av de seks beskrevne morenelokalitetene har det tilsynelatende forekommet reorientering under yngre isbevegelser. Lignende forhold i Nord-Finland er omtalt av Kujansuu (1967: 28). Erkjennelsen av reorientering er av særlig betydning for tolkningen av steinorienteringsanalysene og dermed for korrelasjonsforsøkene.

Isbevegelser rekonstruert på bakgrunn av stratigrafien

Med utgangspunkt i orienteringen i de to utholdende moreneenhetene D og C, som varierer i retning henholdsvis mellom 15–25^g og 380–390^g, har vi markert orienteringene i de forskjellige morenelagene yngre enn Eiravarri interstadial i Fig. 10.

ELDSTE ISBEVEGELSER

Fordi datagrunnlaget for isbevegelsene er svært mangelfullt for morenene eldre enn Eiravarri interstadial kan retningene bare betraktes som punktobservasjoner. Vi har derfor valgt å framstille bare de regionale isbevegelsene som er yngre enn denne interstadialen i Fig. 11.

ELDRE BEVEGELSE

Denne bevegelsen følger etter Eiravarri interstadial. Bevegelsen ser ut til å innledes av isbevegelser av lokal karakter, fra SV mot NØ, før det opprettes en NNØ-lig bevegelse med sikker regional utbredelse. Morene D blir avsatt i denne fasen.

YNGRE BEVEGELSE

Deretter er det et tilsynelatende brått skifte i isbevegelsesretningen til NNV med delvis reorientering av morene D. Morene C blir avsatt i denne fasen. Vi har ingen indikasjoner på om denne retningsforandringen foregikk som dreining eller om skiftet i bevegelsen foregikk raskt. I steinorienteringene er imidlertid ikke mellomliggende retninger registrert.

YNGSTE BEVEGELSE

Isbevegelsen blir nordlig, og vi antar at det er denne som utformer drumliner og flutings over store deler av området og avsetter morene B. Materialet i denne morenen har stedvis preg av avsmeltingsmateriale og steinorienteringen kan variere svært. Bakgrunnsmaterialet for denne morenen er for lite til at vi har satt sammen et regionalt bilde for denne fasen. Ved hjelp av skuringsstriper, drumliner og beslektede former, vil det være mulig å rekonstruere denne fasen med stor nøyaktighet (Sollid et al., 1973).

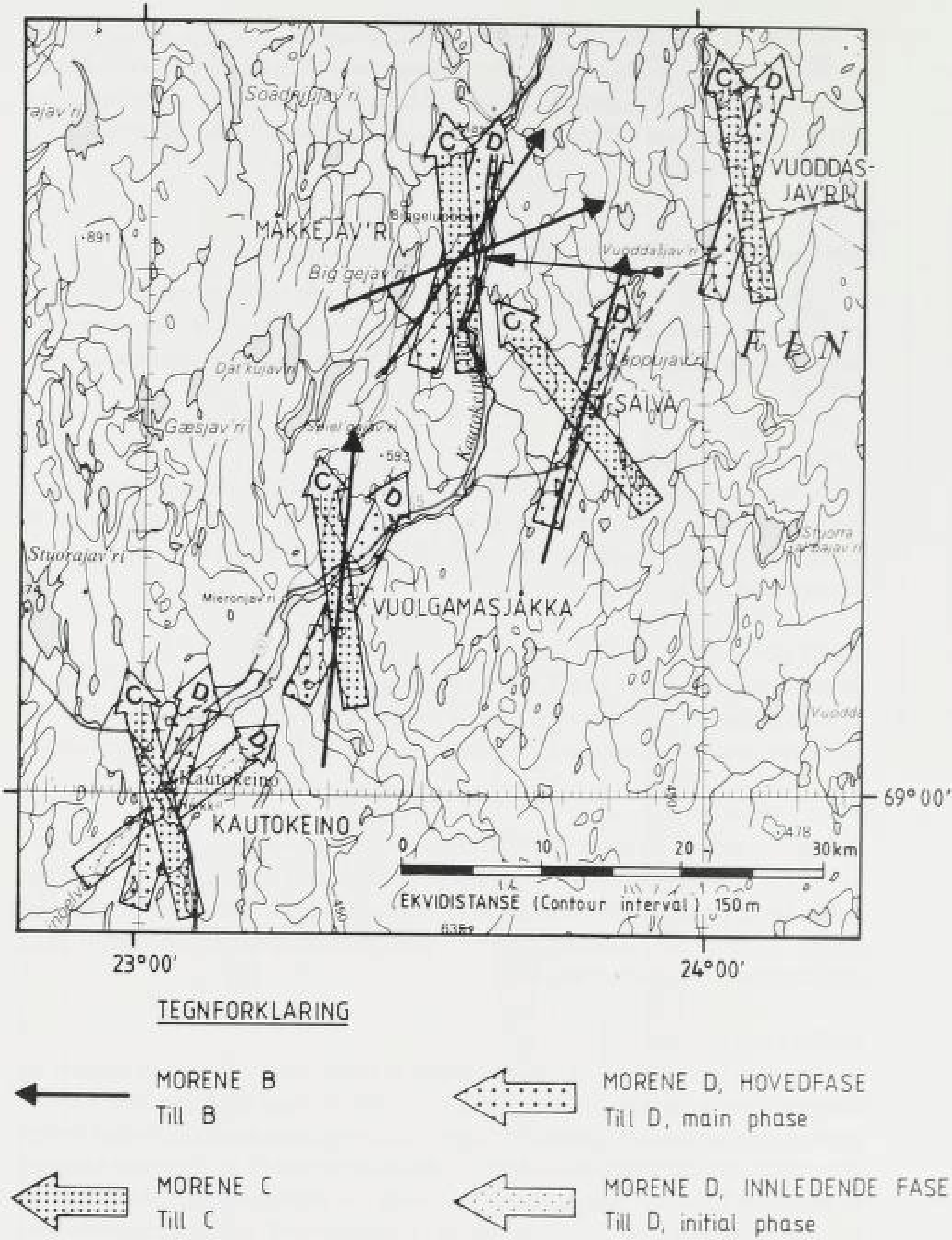


Fig. 10: Avsetningsretning er vist for de to eldste morenene yngre enn Eiravarri interstadial. Korrelate morener (Fig. 9) har fått felles signatur på pilene.
Directions of deposition for the two oldest tills younger than the Eiravarri interstadial are indicated. Tills which are correlated in Fig. 9 have a mutually equal signature on the arrows.

Fig. 11: En tolkning av de to eldste regionale isbevegelsene som er yngre enn Eiravarri interstadial, basert på Fig. 10.
An interpretation of the two oldest regional ice movements younger than the Eiravarri interstadial, based on Fig. 10.

Fig. 6. Saiva. Grensen mellom en pålagringsmorene (15^g) og en overliggende basal utsmeltingsmorene er stiplet. Til høyre for bildet går utsmeltingsmorenen gradvis over i en homogen basalmorene. Over utsmeltingsmorenen (prikket linje) ligger glasifluvialt materiale.
Saiva. The boundary between a lodgement till (15^g) and an upper basal melt-out till is marked with a broken line. Outside the photo, to the right, the basal melt-out till grades into a homogeneous basal till. Glacifluvial sediments rest on the melt-out till (dotted line).

Den midtre morenen, C, er ca. 1 m mektig. Morenen er gjennomsatt av silt-sandbånd og strukturer som vi antar skyldes skjærpåvirkning fra yngste isbevegelse (jfr. s. 4, pkt. 1 og 2). Mye av morene C synes å være reorientert av yngste bevegelse, og bare unntaksvis finnes tilsynelatende uberørte partier av morenen. I grensen mellom morene D og C framkommer et sekundært maksimum mot ca. 385^g, mens 25^g fortsatt er hovedretningen (s.o. 28b). Den NNV-lige retningen dominerer morene C (s.o. 32). Diagrammet viser et klart sekundært maksimum normalt på denne retningen og er antatt å representere en rullende transport i morenen. Årsaken til det noe mindre, men allikevel klare maksimum på ca. 345^g er ikke klarlagt. På grunn av at rosediagrammet baserer seg på få stein kan maksimumet skyldes statistisk støy.

På denne lokaliteten er også en øvre morene, B, representert. Den skiller seg ikke klart lithologisk fra morene C, men har en diffus grense mot denne. Steinorienteringene (s.o. 25b, 27) skiller seg imidlertid klart fra morene C, og viser at morenen ble avsatt mot nord.

Reorientering av partikler i underliggende morene har tilsynelatende skjedd under både nest yngste og yngste isbevegelsesfase.

SAIVA

I et massetak ligger to morener under glasifluvialt materiale (Fig. 6).

Diskusjon og korrelasjon

Det er interessant å merke seg at alle isbevegelsene som er registrert i denne stratigrafien er fra sør mot nord. Spesielt er dette interessant for morenene som ligger i kontakt med og overlagrer sorterte sedimenter. I denne sammenheng er det verdt å legge merke til basaldelen av morene I ved Vuolgamasjåkka, D ved Vuoddasjavri og Kautokeino grustak. Under forutsetning av at sandpakkene under morenene representerer isfrie perioder, og at lagserien ikke har hiatus av betydning i de kritiske partiene, viser alle lokalitetene at isen rykket fram fra sør mot nord.

Allerede Tanner (1930) viser på bakgrunn av ledeblokk-registreringer av Dekkebergarter i den østlige del av viddeområdet, en tidlig isbevegelse fra fjellkjeden mot sør. Read (pers.medd.) finner med basis i steintellinger, kornfordeling og leirmineralogi at det eldste morenematerialet ved Bæivasgieddi (Fig. 1) er avsatt under isbevegelser fra vest mot øst, yngre morenemasser er avsatt under isbevegelser mot NNØ. Ut fra innholdet av forvitringsmineralet kaolin antar han at den eldste morenen er av pre-Weichsel alder, men utelukker ikke en Tidlig-Weichsel alder for morenen.

Kujansuu (1967), Hirvas (1977), Hirvas & Kujansuu (1979) og Hirvas et al. (1981) knytter ved hjelp av stratigrafien en isbevegelse fra fjellkjeden til nedisningen etter Eem.

I Nord-Finland finner Hirvas (1977) og Hirvas & Kujansuu (1979) at nedisningen etter interstadialen(e) Peräpohjola ikke ble initiert fra fjellkjeden, men fra et isskille i Lappland (sør for Kittilä). De første isbevegelsene fra denne perioden er således fra sør mot nord i grensetraktene sør for Kautokeino.

I vår modell når stratigrafien ved Vuolgamasjåkka tilsynelatende lengst tilbake i tid, men selv i basaldelene av underste morene er isbevegelsen *mot* nordvest. For å være i tråd med de finske undersøkelsene bør de første isbevegelsene etter Eem ha vært rettet mot Ø-SØ. Det kan derfor antydningsvis se ut til at vår stratigrafi ikke går så langt tilbake i tiden som Eem.

Lagfølgen ved Vuolgamasjåkka og Vuoddasjavri antyder at Weichsel er representert med minst to interstadialer. Det er foreløpig uklart hvordan denne stratigrafien kan korreleres med Nord-Finland (Hirvas et al., 1981).

I stratigrafien i Kautokeino-distriktet er det langt flere moreneenheter enn den tilsvarende i Finland der det inngår tre moreneenheter over Eem.

Årsaken til dette kan være at SV-lige Finnmarksvidda ligger distalt for isskillene i Midt- og Sein-Weichsel og at forholdene for å få utviklet en mer nyansert og komplett lagfølge derfor er til stede. Innebygget i dette ligger også mulighetene for et lengre Midt-Weichsel interstadialkompleks (Mangerud, 1981; Mørner, 1981).

Karakteristisk for Nord-Finland er Till 2 som overlagrer sedimenter fra Peräpohjola Interstadial. Moreneenheten har stor arealmessig utbredelse og stabil steinorientering i finsk Lappland (Hirvas, pers.medd. 1982). Dette er de samme karakteristika som morene D tilsynelatende har på SV-lige Finnmarksvidda. I en så tidlig fase vil vi bare antydningsvis framholde en korrelasjon mellom morene D og Till 2. Videre korrelasjoner med den nåværende databakgrunn synes spekulativ.

Summary

The authors have tried to correlate some of the main till units (Fig. 9) on the basis of investigations of stone long axis orientation (s.o.) in tills at 6 different stratigraphical sites (Fig. 1). The correlation diagram is based on the stratigraphical sequence exposed in the almost 50 m-high cutting made by the river Vuolgamasjåkka. This locality gives the best opportunity to distinguish between the different stratigraphical units (Fig. 8). Reorientation of till particles caused by overriding ice, as demonstrated by MacClintock & Dreimanis (1964), has been recognized in 5 of the 6 investigated stratigraphical sequences. The reorientation is recognized through the appearance of double maxima in long axis orientation. One of the maxima is parallel to the particle orientation in the undisturbed till; the other maximum tends to be parallel to the orientation in the younger till (Fig. 2: 28b, 5, 12, 2). The reorientation in most cases is common down to about 25–30 cm under the base of the younger till.

Till unit D rests on sand beds in 4 of the 6 investigated sequences. The sandy sediments are interpreted to be of glacifluvial origin and most probably derived from the wastage of the second last glaciation or the growth of the last glaciation. The sediments probably represent one or both of the extremes of an interstadial phase, which we informally have named the Eiravarri interstadial.

Neither pollen nor other kinds of organic contents have hitherto been found in the subtill sediments. Therefore, the possibility of a subglacial origin for the sediments cannot be definitely excluded. However, a subglacial origin seems to be very unlikely because of the regional distribution and the stable stratigraphical position of the sediments.

The low relief, the gently undulating landscape, the wide areal extension and the geographical position about 100 km north of the main ice divide zone during the Weichselian glaciation, provide ideal conditions for development of regionally stable fabrics in the basal tills on Finnmarksvidda. Based on the stratigraphical investigations we have tried to reconstruct ice movement directions which correspond to the different till units younger than the Eiravarri interstadial (Figs. 10, 11). The reconstruction is founded on the well documented supposition that the a-axis orientation maximum of particles in basal tills usually tends to be parallel to the associated ice movement direction. The secondary maximum often found normal to the major orientation direction (Pettijohn 1976) is weakly developed in the investigated tills, and therefore of minor importance in our analyses. Dominant transverse orientation maxima have not hitherto been recognized.

Based on investigations in northern Finland, Hirvas et al. (1976, 1981) conclude that the first ice movement after Eem came from the mountain range. The ice must, therefore, have been crossing Finnmarksvidda from NW–W towards SE–E during the initial stage of the Weichselian glaciation. Our investigations have hitherto revealed no traces of this initial ice phase on the southwestern part of Finnmarksvidda, even though the stratigraphy at Vuolgamasjåkka covers several till units and sediments from at least one interstadial phase older than the Eiravarri interstadial (Fig. 8).

On the eastern part of Finnmarksvidda, Read (pers.comm. 1981) has recognized a till, transported from a westerly direction, which he assumes to be older than Weichsel. An Early Weichselian age, however, cannot be excluded.

Hirvas (1977) and Hirvas & Kujansuu (1979) reported that the glaciation after the interstadial phase(s) named Peräpohjola, started from a culmination zone south of Kittilä in northern Finland. Thus, the first ice movements from this phase were directed from south to north around the national border south of Kautokeino. Till 2 in the Finnish stratigraphy rests on sediments from the Peräpohjola Interstadial. The till unit has a wide lateral extension and a stable fabric orientation (Hirvas, pers.comm. 1982). Apparently, till unit D on Finnmarksvidda has similar characteristics. As a preliminary hypothesis we assume that till D may correlate with till 2 in the Finnish stratigraphy. Further discussion on this subject must be postponed until more data are available.

Etterord. – Vi vil rette en spesiell takk til dr. J. Mangerud som kritisk har lest manuskriptet og foreslått mange forbedringer. I. Lundquist og B. Svendgård har tegnet figurene. L. Holiløkk har utført reproarbeidet. I. Venås har maskinskrevet manuskriptet. O. Sæther og R. Boyd har korrigert den engelske teksten, og våre feltmedarbeidere har lagt inn en god porsjon realisme.

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