IGNEOUS AND METAMORPHIC GEOLOGY

OF THE HUSFJORD AREA, SØRØY,

NORTHERN NORWAY

A Thesis presented for the Degree of
Doctor of Philosophy
in the
University of London

by

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December 1967

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ABSTRACT

The Husfjord area of Sørøy essentially comprises a plutonic igneous complex which has been emplaced into Eocambrian metasediments during the Caledonian orogeney.

The metasedimentary envelope of the complex consists mainly of a sequence of psammites, pelites, semi-pelites, calc-silicate-schists, and metalimestones, which have suffered a 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 synchronous with these metamorphic and tectonic events. Contact metamorphic effects produced by some members have been superimposed upon those of the regional metamorphism.

The earliest member, the Husfjord metagabbro, was intruded towards the end of the first deformation episode, and has undergone the highest grades of regional metamorphism. A norite and a suite of diorites were emplaced during the second deformation episode and these have only suffered a low-grade regional metamorphism. The Husfjord metagabbro and the diorite complex were emplaced essentially by a mechanism of permissive intrusion. The latest members of the igneous complex were the Vatna gabbro, the Slatten gabbros, and a number of minor intrusions including perthosite sheets, basic dykes, and nepheline-syenite pegmatites.

The main diorite suite appears to have developed from a dioritic

melt, which was generated deep in the crust by the syntexis of sialic material with basic magma emplaced from beneath.

Metamorphic mineral paragenesis in pelitic hornfelses suggests that the depth of emplacement of the Husfjord plutonic complex was in the order of 23 km.

ACKNOWLEDGMENTS

I would like to record my most profound thanks to Dr. Brian Sturt, who constantly offered guidance and advice throughout this work, and who critically read the manuscript of this Thesis.

To Prof. B.C. King go my thanks for providing facilities in the Geology Department, Bedford College, where the work was carried out.

Thanks are proffered to both Dr. Brian Sturt and Dr. Donald Ramsay for valuable discussion in the field, and also to my colleagues at Bedford College for theoretical discussions.

I thank Prof. T.F.W. Barth and Dr. H. Neumann for allowing me to carry out analytical work in the Mineralogisk-Geologisk Museum, Oslo. My thanks also go to Dr. Bernard Leake for permitting me to use the X-ray fluorescent spectroscope at the Geology Department, University of Bristol.

Dr. Eugen Stumpfl and Dr. Frank Vokes introduced me to the techniques of reflected light microscopy, for which I thank them. Some of the polished sections were made at the Geology Department, University College, London, and for this I am grateful to Dr. M.K. Wells.

To Mr. J. Keith and his technicians go my thanks for assistance with technical and photographic aspects of the work, and also to the Science Workshop, Bedford College, for the construction of the macro point counting apparatus.

I wish to thank very much indeed Mrs. Sidsel Sturt who kindly typed this Thesis, and also Mrs. Mary Humphries who typed the plate subtitles.

I am very grateful to various members of my family who assisted with some aspects of the compilation of this Thesis, especially to Miss Valmai Cassam who did the preliminary typing.

I would like to thank all my Norwegian friends on Sørøy and in Hammerfest, particularly fru Henny Jacobsen, fru Ida Johanssen, herr Klemmet Nilssen, and herr Rolf Mauno and family, for their most kind hospitality and friendship.

This work was carried out during the tenure of a Science Research Council (N.A.T.O.) Research Studentship Award, for which I am most grateful.

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INTRODUCTION AND
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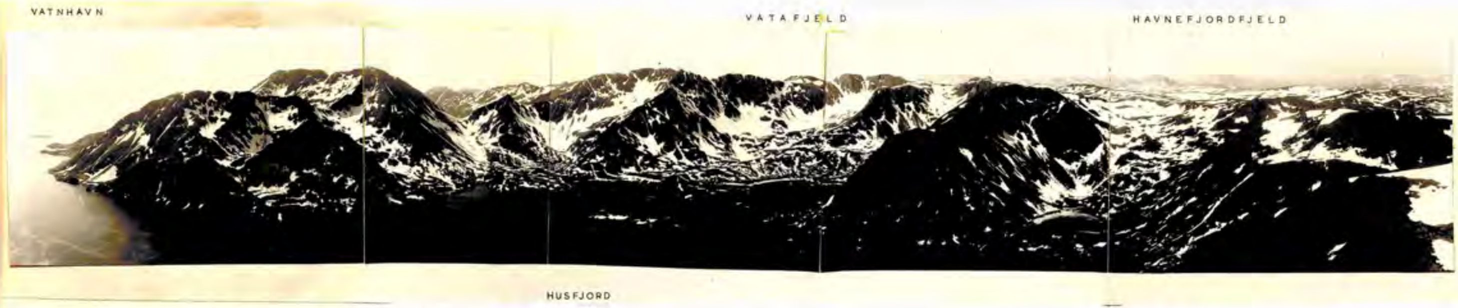


Plate 1. Panorama of the central part of the Husfjord area. The mountains around the fjord consist of Husfjord metagabbro; the high ground of Vatnafjeld and Havnefjordfjeld are of Havnefjord diorite; Vatna gabbro occupies the peninsulas at Vatnhavn.

INTRODUCTION AND REGIONAL GEOLOGY

INTRODUCTION

Sørøy is an island near the town of Hammerfest (Lat. 70° 35' N., Long. 23° 40' E.) which is in Finnmark, Northern Norway, (Fig. 1). The island is about 60 km. in length, and about 30 km. in width across its widest part, and is elongated in a N.E. - S.W. direction. The coastline is highly indented by numerous fjords which give the island a rather stellated shape.

Topographically, the island forms a plateau, on average about 300 - 400 m. above sea-level, rising to over 600 m. in the south-east where the mountains slope steeply down to the sea. Elsewhere in the island, the coastline is characterized by flat-topped headlands which present steep or vertical cliffs to the sea.

Husfjord is on the south-east of Sørøy and is surrounded by high, steep-sided mountains with craggy summits, culminating in Vatnafjeld, 653 m., the highest point of Sørøy. Away from Husfjord, the mountains have a more flat-topped appearance and rapidly drop down to the level of the plateau. Along this part of the coast, the fjords are lengthy, narrow, and closely spaced, leaving long, high, and arete-like headlands between them. Settlements are sparse, the largest being Gåshopen, a mere handful of cottages; others are Vatna, and Husfjord.

The area mapped by the author occupies approximately 100 sq. km. and stretches from Øyfjord in the south-west to Sletnesfjord in the north-east (Fig. 2). The mapping was carried out during the summers of 1964,


MAP SHOWING THE
LOCATION OF SØRØY
SØRØY
Hammerfest
70° N.
70° N.
Tromsø
200 Km.
10° E.
20° E.
Bodø
Arctic Circle
Trondheim
NORWAY
Bergen
60° N.
Oslo
60° N.
10° E.
20° E.

FIG. 1

MAP OF SØRØY SHOWING THE LOCATION
OF THE HUSFJORD AREA

FIG. 2

1965, and 1966. Aerial photographs were used as a base for mapping, and from these the final map was constructed (with a scale of approximately 1 : 15,000).

REGIONAL GEOLOGY

General

The Husfjord area comprises an igneous complex, mainly of gabbros and diorites of varying ages which have a complicated structural and metamorphic history. The intrusion of these bodies and their subsequent deformation and alteration are associated with events that occurred during the Caledonian orogeny. They form the northernmost part of the West Finnmark - North Troms basic and ultrabasic petrographic province, which extends from the Lyngen peninsula in the south to Sørøy in the north.

The Husfjord gabbro complex has been intruded into Eocambrian metasediments which constitute the major part of Sørøy. Detailed mapping of these metasediments was begun by B. A. Sturt and D. M. Ramsay in 1959, the only previous work on Sørøy being that of Petersen (1868, 1883) who surveyed the island and drew a generalized cross-section through it. Ramsay and Sturt (1963) have described the lower part of the succession in northern and western Sørøy, to which the upper part has been added by D. Roberts (1965), who mapped most of the north-eastern part of the island.

Most of Sørøy consists of a sequence of psammites, pelites, and meta-limestones which have undergone complex folding and metamorphism

(Ramsay and Sturt, 1963; Roberts, 1965; Sturt and Ramsay, 1965). The metasediments of Sørøy are a part of the main eocambrian succession of West Finnmark, and lie within the Norwegian Caledonides. In Finnmark, the Caledonides form two main divisions; in north-eastern Finnmark the Eocambrian sediments are unaltered or only slightly altered, whereas in Western Finnmark the Eocambrian and Cambro-Silurian sediments are more strongly metamorphosed and granitized. Both divisions have been thrust towards the south-east over the Pre-Eocambrian Archaean basement of the Baltic Shield (see Holtedahl, 1944).

The main Caledonian thrust-front lies about 100 km. south-east of Sørøy. In south western Finnmark, the highly-metamorphosed division has been thrust over autochthonous Cambrian which rests on the peneplaned Archaean basement. Towards the north-east, the highly metamorphosed division rests with a tectonic contact upon the unaltered Eocambrian rocks, which in turn are thrust as a nappe over autochthonous Cambrian. In eastern Finnmark, the unmetamorphosed Eocambrian and Lower Palaeozoic sediments are in an autochthonous position directly above the Archaean basement (Crowder, 1959: Reading, 1965). The tectonic relationships between these authochthonous unmetamorphosed Eocambrian rocks of eastern Finnmark and the allochthonous unmetamorphosed Eocambrian rocks a short way to the south-west is not yet fully known, but the thrust appears to die away to the north-east. The thrust-plane, which dips gently to the west and north-west, cuts anticlines in the authochthonous Cambrian, and is thus known to post-date the folding which caused buckling in the basement.

The Pre-Eocambrian geology of West Finnmark has been described by Holmsen et alia (1957), and a general account of the Pre-Eocambrian of East Finnmark is given by Bugge in "Geology of Norway" (Ed: Holtedahl, 1960). This latter publication also contains a tentative correlation of all the Pre-Eocambrian rocks of Finnmark by Reitan.

Pre-Eocambrian rocks outcrop in two large tectonic windows in West Finnmark, 30 - 70 km. south and south-east of Sørøy. The southern window stretches from Kvaenangen in the south-west to Altafjord in the north-east; the northern window occupies the ground between Komagfjord and Repparfjord and has been described by Strand (1952) and Reitan (1963). In these windows, a sequence of basaltic pillow-lavas, tuffs, carbonaceous schists, dolomites, and quartzites have been intruded into by lenses of gabbro and serpentinite, and have been slightly metamorphosed to the high greenschist facies on epidote amphibolite facies. This suite of rocks is known as the Raipas formation. For a general account of the geology of Finnmark the reader is referred to Holtedahl (1918), and Strand in "Geology of Norway" (Ed: Holtedahl, 1960).

The West Finnmark - North Troms Petrographic Province

The West Finnmark - North Troms basic and ultrabasic petrographic province has a Caledonian trend, and extends from the Lyngen Peninsula in the South to the islands of Sørøy, Seiland, and Stjernøy in the north. The mainland districts of Loppen and Øksfjord to the south of Stjernøy fall within the province (Fig. 3).

Many of these basic and ultrabasic bodies are banded, and often

DISTRIBUTION OF THE BASIC ROCKS IN THE NORTH TROMS — WEST FINNMARK PETROGRAPHIC PROVINCE

AFTER GEOLOGISK KART OVER NORGE, N.G.U. 1960

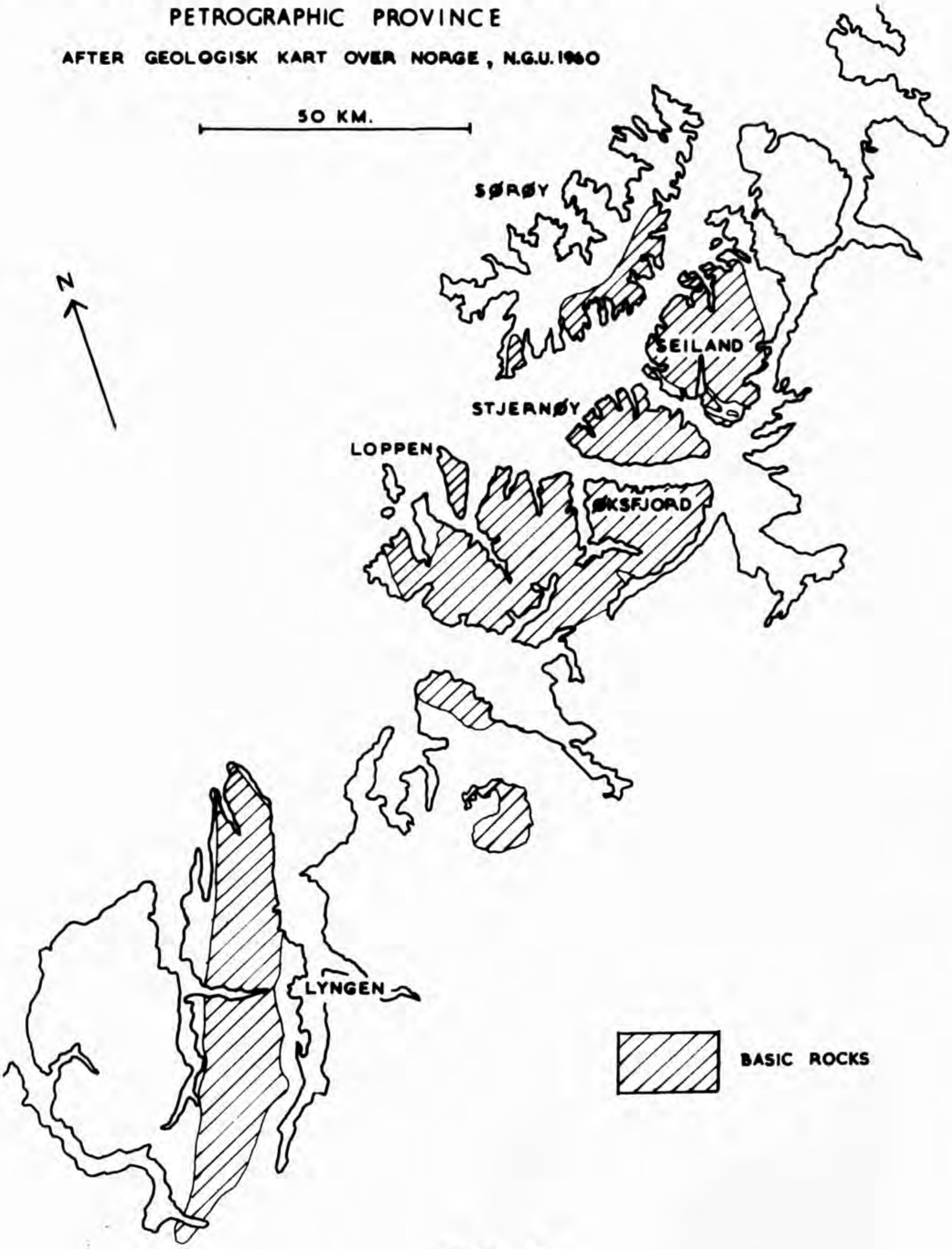


FIG. 3

resemble in many ways the classic examples of Skaergaard, Stillwater, Bushveld, Freetown Complex Sierra Leone, and Rhum, described principally by Wager and Deer (1939), Hess (1960), Daly (1928), Wells (1962), and Brown (1956) and Wadsworth (1961) respectively. Although the Norwegian examples have many similarities with these other bodies, they occur in a rather different tectonic environment. The latter are considered to have formed in stable tectonic conditions, whereas the Norwegian bodies occur along the central zone of the Caledonian orogenic belt and are believed to have been emplaced during the orogeny.

The Lyngen gabbro has been described by Randall (1959) as a hypersthene - gabbro which shows igneous layering in most places. This layering is folded and faulted, and nappe-like structures are sometimes formed, but the layers appear to peter out along their strike. The contacts of the gabbro are described as being tectonic, and Randall deduces that the gabbro was faulted into place in a solid state.

A preliminary account of the Loppen district has been given by Ball et alia (1963). This area contains some gabbros, principally hypersthene gabbro sheets, which in places exhibit banding which is interpreted as being of igneous origin. The gabbros commonly contain numerous metasedimentary inclusions, the smaller ones of which have been disorientated and mobilized by the gabbroic magma. There is good evidence that the magma was very fluid and that it penetrated and reacted with the well stratified country rocks.

Krauskopf (1954) has mapped the layered gabbros and ultrabasic rocks of the Øksfjord area. Here the layers form a broad basin opening northwards with layers on the western margin being overturned. Many layers have cross-cutting relationships and are folded and faulted. In certain horizons within the gabbros there are several small lenses of metalimestone which lie in the plane of the banding in the gabbros. Krauskopf considers that the most likely explanation of the layering is that the gabbros are a product of extreme metamorphism of a sequence of lavas, tuffs, and sediments. He invokes a process of metamorphic differentiation in addition to account for the formation of some of the ultrabasic layers. Small masses of structureless gabbro and of ultra-basic rock, which are intrusive into the layered rocks, are interpreted as possible volcanic centres from which the original volcanic series was derived. In spite of their supposed metamorphic origin, the gabbroic rocks resemble intrusive gabbro both in hand specimens and in thin sections, and is practically unaltered.

The island of Stjernøy is composed of gabbroic and ultrabasic rocks, and these have been described by Oosterom (1963) and Heier (1961). Oosterom has mapped eastern and central Stjernøy where the main rock types consist of an ultramafic sequence of peridotite, dunite, and layered gabbro, and a metamorphic complex of gabbro gneiss and amphi-bolite. The gabbro gneiss contains inclusions of metalimestones and calc-silicate bands and is regarded as a product of high-grade metamor-phism under granulite facies conditions; incipient differential anatexis is considered by Oosterom to have taken place leaving remnants of rocks having high fusion temperatures in these conditions such as limestone and dolomite. The layered gabbro has a layering of igneous appearance, but Oosterom does not favour an igneous origin for the layering. He considers that the gabbroic magma was generated by complete anatexis of pre-existing rocks during granulite-facies metamorphism, and that the layering is a result of tectonic or metamorphic differentiation. He explains the high lime content of the gabbros as being due to contamination by limestone and dolomite. The metamorphic complex and the ultramafic sequence are penetrated by a post-metamorphic metasomatic suite of hornblendite, carbonatite, and nepheline-syenite. According to Oosterom the layered gabbro and peridotite may be considered, from the traditional viewpoint, as products of differentiation of a basaltic magma within the Caledonian orogenic belt. But he thinks that the main difficulty in accepting a co-magmatic hypothesis for the whole of the West Finnmark petrographic province is "the juxtaposition of mafic rocks of igneous and of highly metamorphic appearance".

Heier (op.cit.) has described layered gabbros, gabbro gneisses, and some associated ultrabasic and alkaline rocks in western Stjernøy. Although he is uncertain as to the origin of the banding in the layered gabbros, he considers that it may have been caused by a mechanism similar to that postulated by Wager and Deer (1939) for the Skaergaard intrusion. A period of high-grade regional metamorphism separated the gabbro gneisses from the later ultrabasic and alkaline rocks; it is possible that the layered gabbro is also later than the regional metamorphism. This metamorphism eliminates the possibility of a comagmatic origin for all the rocks described, but Heier considers that there is evidence for a comagmatic relationship among the post-metamorphic rocks, including the layered gabbro.

The south-western half of the island of Seiland is composed mainly of layered gabbros which have been described by Barth (1953) and by Oosterom (1954, 1955). The layering principally consists of alternations of peridotite layers with banded gabbro, and the layered body as a whole does not appear to have definite side walls but seems to continue without a clear break into the neighbouring banded amphibolite-gneiss complex. In describing the layered gabbro of north-western Seiland, Barth says that the metamorphic country rocks, which are at least in part metasedimentary, show successively higher stages of metamorphic-anatectic transformation when traced towards the gabbro, and eventually grade into the layered gabbro. From this it seems that Barth considers that the layered gabbro is possibly a result of high-grade metamorphic-anatectic transformation of the country rocks.

From this survey of some of the gabbroic bodies in the West Finnmark-North Troms petrographic province, it can be seen that many problems are involved in the understanding of the basic rocks in this region. Not least among these is the problem of the genesis of the gabbroic rocks. We have seen that many of the investigators suggest a metamorphic origin for at least some of the layered gabbros, the main evidence being the intercalations and inclusions of metasediments. In some cases the gabbroization is considered to have transformed the country rocks "in situ" as the present layering is supposed to be coincident with the original bedding (as in the gabbros of the Øksfjord area). In other cases, a process of partial and complete anatexis of the country rocks has been invoked (in Stjernøy and Seiland), together with tectonic or metamorphic differentiation to produce the layering where complete anatexis took place (the layered gabbro of Stjernøy).

Another problem is that of the relative ages of the various gabbroic bodies in the province, together with the possibility that they might be essentially comagmatic. It is generally considered that these basic and ultrabasic bodies were formed during the Caledonian orogeny, but in many cases detailed studies of their metamorphic and structural histories have not been carried out. This means that the age relationships between their formation and the metamorphic and structural events that occured during the orogeny are uncertain, and that no conclusions about a possible common parent magma can be made in the case of many of these previously described areas.

In Sørøy, however, there are numerous bodies of basic rocks, together with ultrabasic and intermediate rocks (Fig. 4), which can be demonstrated to have been emplaced at successive stages during the structural and metamorphic development of the Caledonian orogeny in this region. This clearly indicates that the entire suite of igneous rocks cannot have been derived from a common parent magma since their emplacements are separated by long intervals of time and major structural and metamorphic events.

In northern and western Sørøy there are two large gabbros, the Storelv gabbro and the Breivikbotn gabbro. These bodies take the form of sheets which are subconcordant with the country rocks, but they do not show the marked banded layering which is characteristic of many of the gabbros of Seiland, Stjernøy, and Øksfjord. An account of the geochemistry and

MAP OF SØRØY SHOWING THE DISTRIBUTION
OF GABBROIC AND DIORITIC ROCKS

modified after Stumpfl and Sturt, 1965

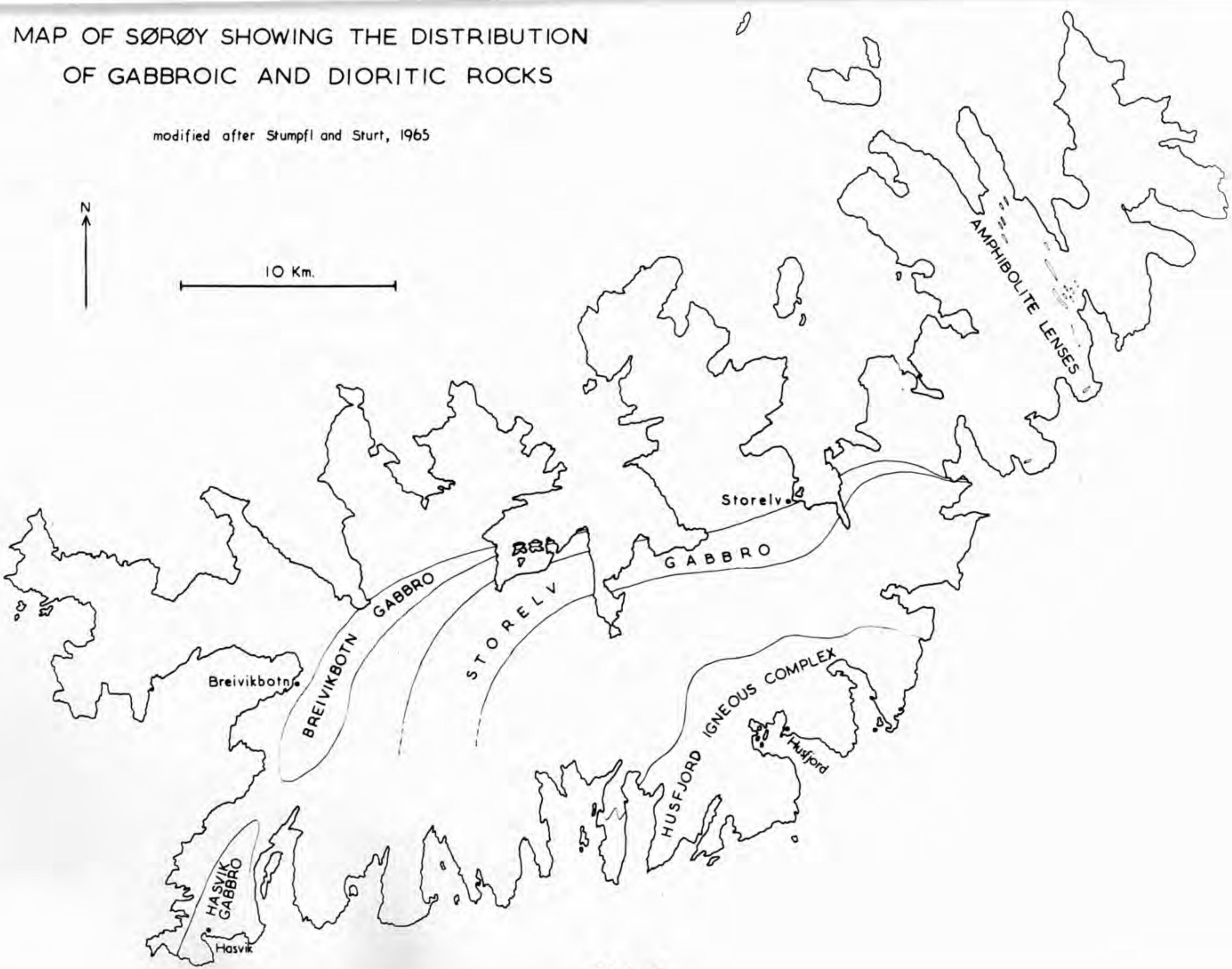


FIG. 4

ore mineral parageneses of these gabbros is given by Stumpfl and Sturt (1965). They were emplaced during the closing stages of the first main phase of folding in this area, and their emplacement was followed by the peak of the main metamorphic event of the region (Stumpfl and Sturt, 1965).

In south-western Sørøy there is another gabbroic body, the Hasvik gabbro, which can be demonstrated to have been emplaced during the main metamorphism (B.A. Sturt, personal communication). This gabbro, which takes the form of a sheet, exhibits a certain amount of layering and contains numerous inclusions of metasedimentary material. In many places the gabbro is seen to penetrate into the metasediment inclusions, and to wedge off blocks, indicating that a mobile gabbroic magma was involved. The metasediment inclusions are regarded as xenoliths, and a metamorphic origin for the gabbro is not invoked (B. A. Sturt, personal communication) as it has been for the gabbros in Seiland, Stjernøy, and Øksfjord. It should be pointed out that no lavas or tuffs have yet been found in the sedimentary succession in Sørøy, although these are hypothetical prerequisites required by Krauskopf for the origin of some of the layers in the Øksfjord gabbros.

At the western end of Sørøy there is an alkaline complex which has been described by Sturt and Ramsay (1965). Age determinations made on the nephelines in nepheline - syenites of this complex by potassium-argon methods reveal that the minimum age of these rocks is about 490 m.y., (Sturt, Miller, and Fitch, 1967). These nepheline - syenites belong to the latest phase of igneous intrusion in the area. They post-date the main regional metamorphism and are syntectonic with respect to the second major period of deformation (Sturt and Ramsay, 1965). Thus the principal episodes of metamorphism and deformation and the emplacement of igneous bodies during the orogeny took place in this region before Upper Arenig times. On the other hand, the Honningsvåg gabbro on the island of Magerøy to the north is intruded into folded rocks containing Silurian fossils. (Reitan, 1960; Henningsmoen, 1961). Thus, the evidence from Sørøy and Magerøy shows that the emplacement of the various igneous bodies in West Finnmark were separated by long intervals of time.

In the Husfjord area of Sørøy, which forms the subject of this present study, detailed structural and petrographic studies of the igneous rocks and the metasedimentary country rocks have also shown that the igneous rocks have been emplaced at successive stages during the development of the orogeny in this area. It can be demonstrated that the various igneous rocks exhibit different degrees of metamorphism and deformation depending upon their position in the sequence of metamorphism and folding. Moreover, many of the intrusions have thermally metamorphosed previous intrusions and metasediments with which they came into contact. Study of the mineralogy and textures of the contact zones has given further information on the relative ages of the igneous rocks with respect to each other and to the main phases of regional metamorphism. In this way the history of the emplacement of the igneous rocks in relation to structural and metamorphic events has been determined, and as a result it seems most unlikely that all the igneous rocks in this area could have had a common parent magma.

PART II

THE COUNTRY ROCKS

THE COUNTRY ROCKS

STRATIGRAPHY

Introduction

The Husfjord igneous complex is bounded by the sea to the south and the east, but to the north and the west it has an envelope of metasedimentary country rocks. The earliest intrusion of the complex is the Husfjord metagabbro, and the original intrusive contact made by this body with the country rocks is preserved along the northern and western margins of the complex. The Husfjord metagabbro is a subconcordant sheet-like body though its margin comes into contact with successive stratigraphic horizons of the country rocks. Thus, if the contact is followed, a stratigraphic succession for the neighbouring country rocks can be built up.

In the western part of the area the strike of the country rocks is approximately N-S, and their dip is towards the west at angles of 70° or more. Towards the north-east the strike of the metasediments changes rapidly to an E-W direction, and here they dip steeply to the north. From this it can be seen that the metasediments have been folded into a large open fold. This belt of metasediments has been overturned by an earlier phase of folding, and detailed discussion of this is more appropriate to the section on the major structures. From various lines of evidence, particularly correlation with neighbouring areas, it is deduced that this overturning has caused the inversion of the true stratigraphic succession in this area. The Husfjord metagabbro makes contact with successively higher stratigraphic horizons as the contact is traced from south-west to north-east. The general stratigraphic succession is as follows:

Quartzo-feldspathic phyllitic schists

Calc-silicate-schists with metalimestones

Kyanite-garnet-mica-schists

Quartz-garnet-mica-schists

Cummingtonite-schists

Psammites and semi-pelites (migmatized)

The migmatized psammites and semi-pelites are the lowest horizons, and occur in the south-west of the area. Due to lack of continuous exposure along the contact zone it has not been possible to build up a complete succession. Only between the Kyanite-garnet-mica-schists and the Calc-silicate-schists with metalimestones is there continuous exposure. This point must be borne in mind when correlation with neighbouring areas is made.

Within the igneous complex there are trains of metasedimentary rafts of different lithologic types, which appear not to have moved far, if at all, from their original positions. The distribution of the various types of raft reflects the stratigraphic succession of the country rocks, and parts of the metasedimentary succession can be mapped within the complex as a relict stratigraphy.

Psammites and Semi-pelites (migmatized)

The lowest members of the metasedimentary succession are a group of fine-grained feldspathic psammites and intercalated semi-pelites. The psammites are greyish-white to buff in colour, and small reddish-brown garnets can be seen with the naked eye. The thickness of the bands are very variable; some are thick and massive and are usually separated from neighbouring massive units by thin semi-pelitic bands. In other places the psammite units are reduced to a few centimetres in thickness, and there are numerous intercalations of semi-pelitic material. In one broad horizon, semi-pelitic rocks are predominant and the psammite units occur as narrow bands only a few centimetres or less in width. The semi-pelites contain abundant garnet, and large crystals up to about 2 cm. in diameter often develop.

The psammites and semi-pelites have subsequently undergone extensive migmatization along a broad belt which has its maximum intensity at the northern end of Kobberfjord. The incidence of migmatite decreases to the west and to the east of Kobberfjord. The migmatites principally take the form of phlebite and agmatite, in which quartzo-feldspathic material occurs along foliation planes in the metasediments. Feldspar porphyroblasts are also abundant. Discussion on the morphology and development of the migmatites appears in the section on migmatization, p. 285.

The psammitic group has been intruded by a number of narrow basic sheets, some of which attain a thickness of a metre or so, while others are only a few centimetres in width. They are usually concordant, but sometimes they are slightly discordant and occasionally they bifurcate. It can be demonstrated that they were not all emplaced during the same phase of intrusion, but that a comparatively long time interval elapsed between the earliest and the latest. Evidence for this is provided by their relationship to episodes of folding, and this is discussed further when the structure of the metasediments is considered, p. 32.

The migmatized psammites and semi-pelites make contact eastwards with the igneous complex. At the north-eastern end of Kobberfjord, they come into contact with a norite, the Kobberfjord norite which forms a norrow sheet between the Husfjord metagabbro and the country rocks in this part of the area. At the contact the country rocks become hornfelsed and mixed up with garnetiferous norite in a narrow zone of slight brecciation.

Cummingtonite-Schists

The next group of rocks is a schist group in which the amphibole cummingtonite occurs. These schists are fine-grained and are usually grey in colour. They are well-banded, and individual bands are quite thin, generally being in the order of a few centimetres or less in thickness. Sometimes they are slightly flaggy, and micas can be seen on the foliation surfaces. Some bands are composed virtually entirely of cummingtonite, and in these the crystals are intergrown to form a solid mat of cummingtonite. The crystals are not very elongate, but there is a general tendency for them to be orientated with their c-axes parallel to the banding. Other bands are more micaceous and contain biotite laths intergrown with the cummingtonites and orientated parallel to the banding. These micaceous bands also contain streaks and pods of quartz which probably segregated out during metamorphism. The effect of weathering on these bands of different mineralogy has been to form a slight surface-ribbing parallel to the banding in which the more quartzose bands stand out. Minor folding of the banding occasionally occurs, and in thin section the foliation is seen to be deformed into small folds.

The cummingtonite - bearing schists possibly represent layers of magnesian shales which have subsequently been metamorphosed.

Quartz-garnet-mica-schists

The next group, which is separated from the cummingtonite-schists by a broad tract of land where exposure is poor, is a group of quartz-garnet-mica-schists. These are dark grey semi-pelites with occasional thin psammite bands. There are occasional quartzose streaks and bands in the semi-pelites, probably due to metamorphic segregation. The schistosity, which is delineated by micas, is rather poorly developed, and in places it is deformed by small folds. Numerous small red garnets are clearly visible in the semi-pelites, and these are slightly more abundant in the more quartzose bands of the semi-pelites. Some of the garnets are sub-spherical, whereas others are elongated parallel to the banding. In places a coarse ac-jointing has developed, particularly in the psammitic bands. Occasionally these joints are occupied by quartz veins.

Kyenite-garnet-mica-schists

Stratigraphically overlying the garnet-mica-schist group is a sequence of well-banded schists with intercalated psammites and occasional calc-silicate-schists. The predominant schists are micaceous semi-pelites which sometimes show a slight rusty weathering. The schistosity is usually fairly well developed, and small garnets are often visible to the naked eye. The psammites are grey or yellowish and contain varying amounts of feldspar. Rocks rich in calc-silicate minerals, particularly tremolite-actinolite, occur as bands and as lenses within some of the semi-pelites. These lenses presumably represent calcareous concretions in the original sediment. A few pegmatites have been emplaced parallel to the banding.

Towards the upper part of the kyanite-garnet-mica-schist group, calc-silicate-schists became more prominent and psammites and semi-pelitic schists are of less importance. The group passes up into the next group, the calc-silicate-schists, which contains limestone horizons near its base.

Calc-silicate-schists with Metalimestones

The rocks in this group are mainly well-bedded, fine-grained schists, sometimes fairly massive. The general colour of the rocks is grey or yellowish-brown, but many bands have a pale greenish colour due to the high actinolite content of certain of the calc-silicate bands. When the psammitic or calc-silicate bands alternate rapidly with semi-pelite, a ribbed appearance results because of the differential weathering.

Interbedded with the calc-silicate-schists near the base of the group there are occasional metalimestones. These do not appear to form continuous bands for any great distance, but outcrop as long lenses, each several tens of metres in length, and about ten metres in thickness, elongated parallel to the general banding of the rocks. It is probable that they originally formed continuous beds in the sedimentary sequence but have since formed tectonic lenses during subsequent deformation.

These metalimestones consist of coarsely-recrystallized calcite which weathers to a bluish-grey colour. It contains occasional thin calc-silicate bands which form thin ribs standing out from the less resistant limestone. The calc-silicate ribs presumably represent more siliceous bands in the limestone and so indicate the original bedding. During subsequent deformation, the metalimestone and the calc-silicate bands behaved in different ways. The siliceous bands, being competent, formed coherent folds, but the less competent limestone deformed in a more plastic fashion and flowed around the folds in the calc-silicate bands. In places the latter have been broken through by limestone, and fragments, especially fold-closures, are now surrounded by a limestone matrix.

In one place the calc-silicate-schists exhibit a structure in the banding which appears to be of sedimentary origin. It has the appearance of a wash-out structure in which churned-up material penetrates downwards into the underlying layers, cutting across them discordantly (Plate 2). The material above the line of discordance is

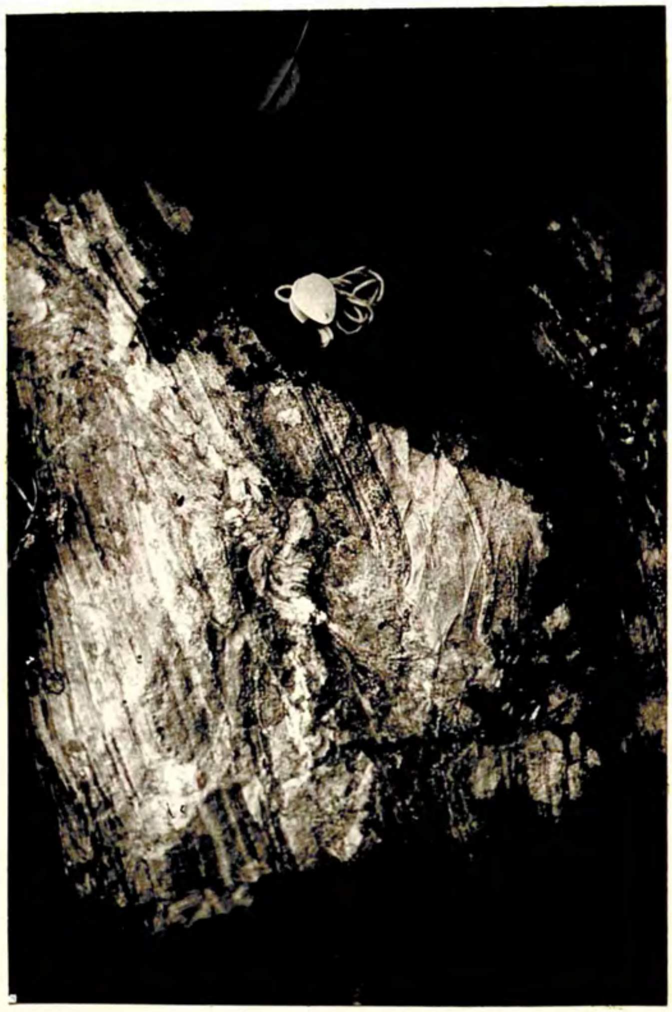


Plate 2. Wash-out type of sedimentary structure; direction of younging is to the left (south). Above Havnefjordbotn.

different from that immediately beneath it. If indeed this structure is of sedimentary origin, it would indicate that the beds young to the left of Plate 2, i.e., to the south, and since the beds dip to the north here, the stratigraphic succession would be inverted.

Occasional pegmatites have been emplaced into the metasediments, usually parallel to the banding. One of these is a medium-grained quartz-rich pegmatite containing large subhedral sphenes which are randomly-orientated and are up to 5 or 6 cm. in length.

At this part of the contact between the igneous complex and the country rocks there has been an introduction of feldspathic material, both into the igneous rocks and into the metasediments. In some places it is seen to be in the nature of a metasomatic introduction of feldspathic constituents in which the metasediments become diffusely feldspathic and contain large feldspar porphyroblasts. In other places a liquid phase is indicated by bands and sheets of coarse-grained porphyritic diorite which contain occasional angular disorientated xenoliths of the metasediment. It is believed that the feldspathization is associated with the introduction of the diorites, and will be further discussed when the suite of diorites is considered.

Quartzo-feldspathic Phyllitic Schists

An unexposed area separates the calc-silicate-schists with limestones from the next group, which are quartzo-feldspathic-phyllitic schists.

These are a sequence of light grey well-banded, somewhat flaggy rocks, in which individual bands do not attain any great thickness. Some of the bands are quite feldspathic and weather to a very light grey or white colour; these are only slightly micaceous. Many bands are less feldspathic and are considerably micaceous, and these weather to a slightly darker grey. The abundance of mica gives the rock a poorly-developed phyllitic schistosity, and considerable white mica can be seen on the surfaces of the foliation planes. Occasional small rounded purply-red garnets occur. A few thin basic sheets have been intruded parallel to the banding.

In places the more massive quartzo-feldspathic bands form boudins around which the foliation of the micaceous bands swing, and the basic sheets are also sometimes boudined. The longest axes of these boudins appear to be subparallel to the axes of the minor folds which commonly deform the banding. A lineation on the banding surfaces, caused by microfolds, is also parallel to these fold areas.

HISTORY OF SEDIMENTATION

As the banding of the metasediments is defined by different lithological units, in which there is an occurrence of an apparent sedimentary structure, it is clear that the banding must be closely related to the original sedimentary bedding. In spite of the subsequent metamorphism and deformation suffered by the metasediments, it is possible to come to some general conclusions concerning the conditions of sedimentation.

The massive psammites at the base of the succession were presumably deposited in shallow water, and it is probable that a large land-mass was rapidly contributing large quantities of material to this area. The abundance of feldspar in the psammites, even where extensive feldspathization has not occurred, indicates that this land mass was probably mainly of a granitic nature. It is likely that it was part of the Pre-Eocambrian granitic basement which had been exposed to erosion.

The thin intercalations of semi-pelitic material within the psammitic group probably represent minor disturbances of the land or temporary changes in sea-level resulting in the slight deepening of the water in the present area. The broader horizons of semi-pelite indicate that the slightly deeper water conditions occasionally persisted for long periods. The presence of numerous thin psammite bands in the predominantly semi-pelitic horizons shows that the conditions during these times were not completely stable, and that the depth of the water was probably not very great.

It is likely that these intermediate depth conditions persisted through the cummingtonite-schists, quartz-garnet-mica-schists and kyanite-garnet-mica-schists groups. These groups are represented by fairly fine-grained rocks, generally argillaceous, with occasional more arenaceous layers.

A general shallowing of the water must have occurred at the time of deposition of the rocks in the calc-silicate-schists with metalimestones group. These schists represent a sequence of calcareous and magnesian arenites which presumably were deposited in fairly shallow water. Rapid alternation of different psammitic units and the presence of sedimentary structures bear witness to the shallow water conditions.

The limestones were presumably deposited in clear, fairly shallow water in which the thin clean calcareous siliceous layers, now represented by the calc-silicate ribs, were also laid down.

The rocks of the quartzo-feldspathic phillitic schists were deposited in water of shallow to intermediate depth. The alternation of quartzo-feldspathic units with phyllitic semi-pelitic material testifies that conditions were unstable, and probably indicates a general deepening of the water after the pre-existing shallower conditions.

The general conditionsof sedimentation implied by this succession of rocks are those of the shelf area of a geosynclinal trough. The depth of the water was constantly changing, but it was probably never very great.

CORRELATION

To the north of the Husfjord area lies the Storelv area which has been mapped by Drs. Ramsay and Sturt. The metasedimentary succession in that area is as follows (Ramsay and Sturt, 1963):

Falkenes Limestone Group

Storelv Formation

Klubben Quartzite

The Klubben Quartzite consists mainly of psammites, often massive, with more pelitic garnetiferous horizons, and the Storelv Formation comprises an alternating sequence of pelitic and psammitic horizons.

There are three main pelitic horizons, and these are principally mica-schists or quartz-mica-schists, sometimes garnetiferous. At the top of the Storelf Formation there are calc-silicate-schists which pass up into the Falkenes Limestone Group. This latter group contains graphitic and mica-schists with a blue-grey limestone at the base.

In the Breivikbotn area to the south-west of the Storelv area a similar succession has been mapped by Drs. Sturt and Ramsay, and they consider that the two successions can be directly correlated (Sturt and Ramsay, 1965). The succession is:

Hellefjord Schist Group

Breivik Group

Falkenes Limestone Group

Storelv Group

Klubben Quartzite

The subdivisions of the Storelv Group here are not so well defined as at Storelv, and the upper part of the Falkenes Limestone Group of the Storelv area is given the name Breivik Group here. The succession is extended upwards into the Hellefjord Schist Group, first named in the Langstrand area in north-eastern Sørøy which was mapped by Dr. D. Roberts.

The succession in the Langstrand area is as follows (Roberts, 1965,1967) :

Hellefjord Schist

Falkenes Limestone Group

Storelv Schist

Transitional Group

Klubben Quartzite Group

The Klubben Quartzite Group consists of three massive psammitic units separated by two semi-pelitic horizons. The psammites are felds-pathic and weather to a light grey or buff colour, but occasionally become more micaceous with a fair amount of biotite, giving a flaggy appearance. The semi-pelitic horizons are rich in garnet. Amphibolite sills and dykes have been intruded into the Klubben Quartzite Group and they pre-date the earliest deformation.

Roberts has correlated the Transitional Group, which consists of alternating thin quartzite bands and pelitic laminae, with the lower part of the Storelv Formation in the Storelv area.

The Falkenes Limestone Group contains two limestone horizons separated by calc-silicate-schists and garnet-kyanite-sillimanite-schists; the calc-silicate-schists underlying the lower limestone are included within this Group. The lower limestone is pale grey on grey-blue and contains thin calc-silicate layers. The upper limestone weathers to a brownish colour and contains frequent calc-silicate and pelitic intercalations which sometimes constitute the bulk of the rock.

The Hellefjord Schists are a series of grey, closely-banded, fine-grained phyllitic schists, which often develop a good flagginess. Small purply-red garnets are common, especially in the more phyllitic bands.

In the Husfjord area the general stratigraphic succession is as follows :

Quartzo-feldspathic phyllitic schists

Calc-silicate-schists with metalimestones

Kyanite-garnet-mica-schists

Quartz-garnet-mica-schists

Cummingtonite-schists

Psammites and semi-pelites (migmatized)

The feldspathic psammites with intercalated semi-pelitic horizons, now extensively migmatized, closely resemble the Klubben Quartzite Group of the Langstrand area. They also contain amphibolite sheets, a common feature in the Klubben Quartzites. It is suggested that the psammites and semi-pelites can be correlated with the Klubben Quartzite Group. One broad horizon of semi-pelite is recognised in the migmatized rocks of the Husfjord area, so that it would appear that the succession goes down beneath the Quartzite 3 of the Langstrand area. Upper parts of the psammites and semi-pelites group probably correspond with the Transitional Quartzite of the Langstrand area, which in turn is equivalent to the lower part of the Storelv Formation of the Storelv area.

The next major group in the succession is the garnet-mica-schists, for the cummingtonite-schists form only a thin, uncommon layer above the psammites. The horizons beneath the garnet-mica-schists are not well exposed, but these schists could be correlated with the upper part of the Storelv Formation, which is principally of garnet-mica-schist. The top of the Storelv Schist is represented by the kyanite-garnet-mica-schists, where a transition into the calc-silicate-schists with metalimestones takes place. This latter group is correlated with the Falkenes Limestone Group. The unexposed tract next to the calc-silicates and limestones group could well be occupied by horizons equivalent to the Breivik Group of S.W. Sørøy. The overlying quartzo-feldspathic-phyllitic-schists of the Husfjord area may be correlated with some of the Hellefjord Schists as described by Roberts in the Langstrand area. Roberts (1965) has made a stratigraphical correlation between the Langstrand, Storelv, and Breivikbotn areas, and has also extended this correlation to include the Loppen area on the Troms-Finnmark border to the south (Roberts 1967). Fig. 5 shows the successions as correlated by Roberts (1967) with the addition of the succession in the Husfjord area.

Thus it appears that across the south-eastern part of Sørøy there outcrops a belt of metasediments which can be correlated with the general Sørøy succession, and which young towards the south-east.

AGE OF THE METASEDIMENTS

According to the 1:1,000,000 geological map of Norway (Holtedahl and Dons, 1960) the metasediments of Finnmark are considered to be of Eocambrian age with extension into Cambro-Silurian in places. Holtedahl (1918) mentions Lower Cambrian fossils in the south-east border of the Caledonian mountain zone, and the Digermul Peninsula, Tana, has provided numerous fossils of Cambrian age (Føyn, 1937; Strand, 1935). In S.E. Magerøy, near the Honningsvåg gabbro, a limestone has yield numerous Silurian fossils including crinoid stems, chain corals, rugose corals, and pentamerids (Henningsmoen, 1961).

COMPARATIVE METASEDIMENTARY SUCCESSIONS OF SØRØY AND THE LOPPEN AREA

LOPPEN[1]	S. W. SØRØY[2]	STORELV[3]	LANGSTRAND[4]	HUSFJORD[5]
	HELLEFJORD SCHISTS		HELLEFJORD SCHIST	QUARTZO-FELDSPATHIC PHYLLITIC SCHISTS
PELITIC SCHISTS	BREIVIK GROUP			
CALCAREOUS SERIES WITH METALIMESTONE(S)	FALKENES LIMESTONE GROUP	FALKENES LIMESTONE GROUP	FALKENES LIMESTONE GROUP	CALC-SILICATE-SCHISTS WITH LIMESTONES
PELITIC SCHISTS.		STORELV FORMATION: PELITE 3	STORELV SCHIST	KYANITE-GARNET-MICA-SCHISTS QUARTZ-GARNET-MICA-SCHISTS
		PSAMMITE 2	TRANSITIONAL GROUP:	
IMPURE QUARTZITES	STORELV GROUP	PELITE 2		CUMMINGTONITE SCHIST
		PSAMMITE 1		
		PELITE 1		
QUARTZITE AND AMPHIBOLITE	KLUBBEN QUARTZITE	KLUBBEN QUARTZITE	KLUBBEN QUARTZITE GROUP: QUARTZITE 3 UPPER SEMI-PELITE QUARTZITE 2 LOWER SEMI-PELITE QUARTZITE 1	PSAMMITES AND SEMI-PELITES (MIGMATIZED IN PART)
FLAGGY QUARTZITE				
MASSIVE AND/OR FOLDED QUARTZITE				

1 Generalized succession by Roberts (1967) from Ball et al (1963) 2. From Sturt and Ramsay (1965) 3. From Ramsay and Sturt (1963)

4 From Roberts (1965, 1967) 5 Present work by present writer Correlation of 1, 2, 3, and 4 by Roberts (1967)

FIG. 5

In Sørøy, the only fossils that have come to light are those found by Dr. B. A. Sturt in a limestone within the Klubben Quartzite Group, which is the lowest member of the succession exposed in Sørøy (Sturt, Miller, and Fitch, 1967). They have been identified by Prof. C. H. Holland as archaeocyathids, of a type restricted to middle Lower Cambrian. It is considered that the metasedimentary succession in Sørøy is partly of Eocambrian age with an extension into the Cambrian (Ramsay and Sturt, 1963; Roberts, 1965, 1967; Sturt, Miller, and Fitch, 1967).

Potassium-argon age determinations made on late nepheline-syenites in western Sørøy indicate that the Caledonian orogeny began in this region well before Upper Arenig times (Sturt, Miller, and Fitch, 1967). This means that the youngest possible age of the metasediments is probably Upper Cambrian.

STRUCTURE

Introduction

The development and structure of the igneous complex is intimately related to major structures of the country rocks. Therefore a consideration of the structure of the metasediments is of great importance.

Two main episodes of folding are recognised in the metasediments, and these were intense, complex, and protracted. In the Husfjord area, the general attitude of the metasedimentary banding is principally due to the second episode of folding.

The first major episode (F_1) in Sørøy, is characterized by large-scale recumbent folding in which movement was towards the east and south-east (Ramsay and Sturt, 1963; Roberts, 1965). The metasediments in the Husfjord area all lie on one limb of a large F_1 fold, and the only visible evidence of F_1 structures is in minor folds.

In the second major episode, F_2, folds of more open style were formed, and these have orthorhombic or monoclinic symmetry (Ramsay and Sturt, 1963). These generally have axial trends close to those of the F_1 structures, and in the case of the monoclinic folds the eastern or south-eastern limbs are often slightly overturned. The metasediments in the Husfjord area lie within one of these overturned limbs.

The F_2 folding episode appears to have had two main phases (Roberts, 1965). In the Husfjord area there was considerable igneous activity during and after the earliest F_2 movements, and before the latest deformation. The late F_2 phase is probably responsible for the open arcuate form of the igneous complex and its metasedimentary envelope.

The key to the structural history of the metasediments lies in the minor structures. These will be considered before the major structures are discussed and correlated with the regional tectonics of Sørøy.

Turner and Weiss (1963) have defined four classes for the scale of geological bodies. These are submicroscopic, microscopic, mesoscopic, and macroscopic. In this thesis, the terms minor structures and major structures correspond with mesoscopic and macroscopic as defined by Turner and Weiss.

MINOR STRUCTURES

F_1 Structures

The earliest folds that are seen are tight isoclinal folds lying within the general banding. They are particularly well preserved in the psammites and semi-pelites group and in the calc-silicate bands in the limestones of the calc-silicate-schists group. The early age of these folds is demonstrated by the fact that they are often refolded by later folds, and also that occasionally they are cut by amphibolized basic sheets. They are minor folds associated with the large-scale recumbent structures produced during the F_1 movements.

This early episode of deformation was long and complex, and must have been composed of several phases of folding rather than of one continuous movement. This is borne out by the fact that these so-called F_1 folds deform a pre-existing schistosity. In one case the fold occurs in a pre-F_1 basic sheet in which there is a faint schistosity which follows around the fold (Plate 3). This fold is typical of the intrabanding isoclinal folds of the early generation, but the presence of the schistosity indicates the occurrence of a phase of deformation prior to the formation of this particular fold, possibly an earlier phase of the F_1 movements.

The form taken by these F_1 folds is tightly isoclinal, with the limbs lying parallel or subparallel to one another, the width across the two limbs usually being in the order of about 10 cm. The amplitude and wavelength of the folds are indeterminate as the complementary folds adjacent to any particular fold are not seen. Also, for the same reason, the sense of overturning or vergence of the folds cannot be determined. Due to the obliquity of the surface of exposure to the fold axes, the fold hinges generally appear considerably thickened. Thus it is difficult to be sure whether the folds are of concentric type or of similar type.

The present trends of the axes of these early minor folds vary according to the attitude of the limb of the latest folding in which they are situated. Furthermore, F_1 antiforms often have a different axial direction from F_1 synforms within the same limb of a later fold. For example, at Kobberfjord, within the steep western limb of a fold of

Plate 3. Folds of two different generations defined by basic sheets; the tight isoclinal fold is of F_1 age, and the monoclinic fold is of F_2 age. Kobberfjord.

the latest generation, F_1 antiform axes trend at 140° - 150° and plunge to the N.W. at angles of 40° - 60°; F_1 synforms in the same limb have axial trends of 010° - 015° and plunges of 20° - 40° to the N.N.E. Since this swing in direction of the F_1 minor folds occurs within the same limb of an F_2 fold, it would appear that it is a primary feature of the F_1 folds.

The most well-preserved F_1 folds occur in metalimestone rafts within the igneous complex at Havnefjord, and here the shapes of the F_1 folds can be clearly seen. Calc-silicate bands are folded into isoclinal folds having axial planes dipping to the north at about 50° (Plate 4). The axes of these folds are seen to curve within the axial planes. Because the axes are curved in a plane subperpendicular to the erosion surface, the calc-silicate bands outcrop in a series of 'eyes' (Plates 5 and 6). These fold-closures in the calc-silicate bands weather out as little 'boats' (Plate 7).

Similar eyed folds have been found in other areas of Sørøy (Roberts, 1965; Sturt and Ramsay, personal communication) and in each case they are due to the curvature of the axes of the F_1 folds. Nicholson (1963) has described some eyed folds from marbles in Northern Norway. He does not completely accept the theory that such structures are due to interference produced by the superposition of folds as advocated by Ramsay (1962). He considers that they may be formed in one phase of folding by differential movement within and along the axial plane of the folds.

In the Dalradians of the Scottish Highlands, some of the early

Plate 4. F_1 fold in calc-silicate band.
Havnefjordbotn.

Plate 5. Eyed fold in a calc-silicate band.
Havnefjordbotn.

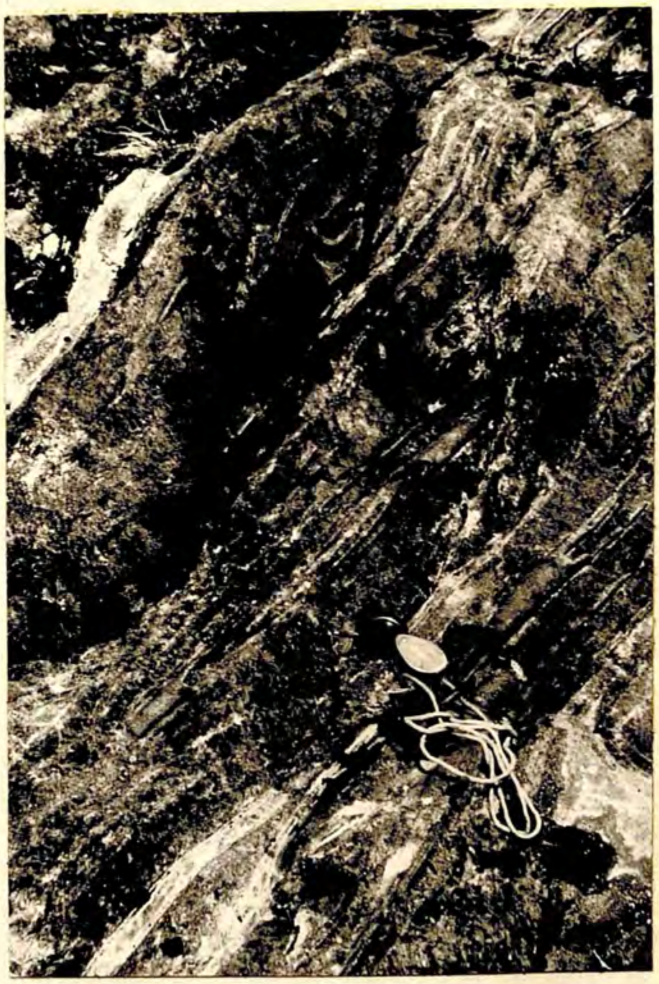


Plate 6. Eyed fold in a calc-silicate band.
Havnefjordbotn.

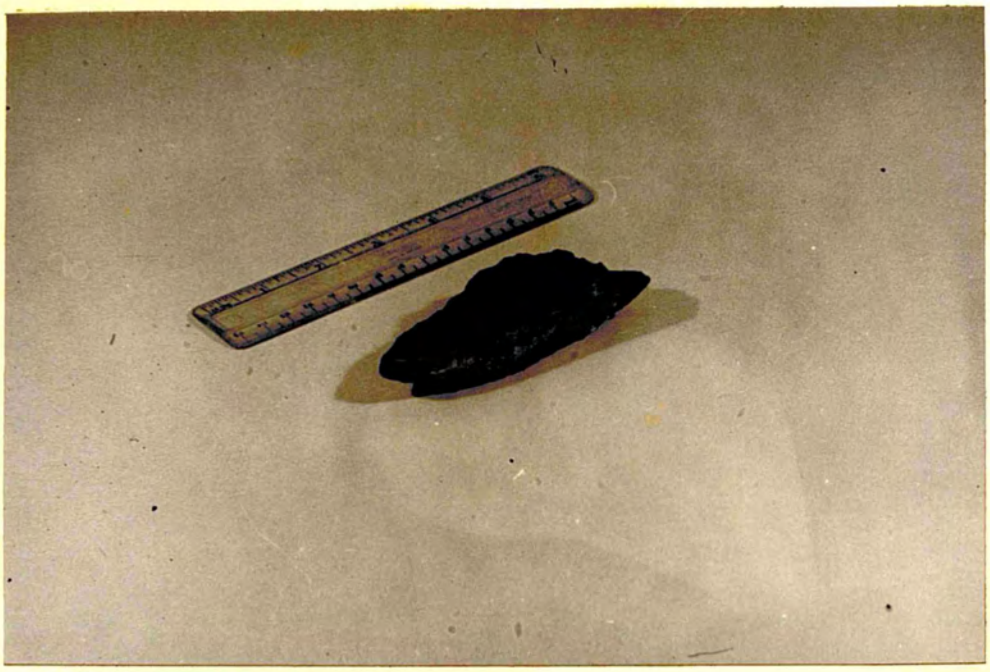


Plate 7. 'Boat-fold' in a calc-silicate band. From Havnefjordbotn.

folds have curving axes (Voll, 1960). Voll lists four possible explanations for the formation of curved axes :-

(1) Unequal stretching along the strike of the axial planes (contemporaneous formation)

(2) Subsequent rotation of detached hinges of folds with normal axes

(3) Bedding (banding) planes buckled before folding

(4) Bedding (banding) planes intersected by axial planes having different attitudes.

He states that (2), (3), and (4) have been observed, but that (1) has not been clearly demonstrated.

Experimental work and mathematical analysis by Bhattacharji (1958), however, supports the view of a contemporaneous origin of curved axes and cross-folds. They are formed by differential flowage within developing folds, especially at the nappe stage of folding. He says they are formed particularly in an association of alternate elastic and plastic materials or materials of different elastico-viscous properties.

Such a difference in elastico-viscous properties exists between the calc-silicate bands and the limestone in the case under present consideration. It seems likely that the swings in the axes of the F_1 folds could be primary features, formed by differential movement of material along the axial planes as envisaged by Nicholson and by Bhattacharji. These early folds were probably essentially flow folds in which deformation was of a plastic kind. In such conditions it is most likely that local heterogeneity would preclude the formation of perfectly linear axes, and would favour the development of curving axes.

In summary, F_1 folds are characterized by recumbent isoclinal folds, probably of flow fold type, which have axes curved in the plane of the axial plane.

Early F_2 Structures

F_2 folds are characterized by a more open style than those of F_1, and are occasionally seen to refold F_1 folds. In the south-western part of the area, where the strike of the metasediments is approximately N-S, the F_2 minor folds have a monoclinic symmetry. They are asymmetrical, with a vergence towards the west and north-west (Plate 3). If Pumpelly's rule is applicable here, this would indicate that the metasediments in this belt are on the inverted limb of an asymmetrical antiform lying towards the west and north-west (i.e. towards the left of Plate 3).

In the northern part of the area, where the general strike of the metasediments is approximately E-W, the F_2 minor folds are not so markedly asymmetrical. In some cases conjugate folds are formed (Plate 8), in which the axial planes of opposing pairs of folds dip in opposite directions. The symmetry of these conjugate folds is not truly orthorhombic since the axes of the conjugate pairs are not exactly parallel, and the kinematic b-axis does not lie in the plane of the banding (Ramsay, 1962a; Ramsay and Sturt, 1963). In the Storelv area, to the north of the Husfjord area, the strike of the metasedimentary

Plate 8. F_2 folds. Havnefjordbotn.

belt is approximately E-W. In that area the F_2 minor folds also commonly form conjugate pairs of orthorhombic and lower symmetries, which were formed by deformation along complementary shears (Ramsay and Sturt, 1963). In the northern part of the Langstrand area, where the metasediments strike approximately N-S, the predominant style of the F_2 minor folds is monoclinic (Roberts 1965). Towards the south of that area, however, the strike swings to an E-W direction and the minor folds become less asymmetrical, and numerous conjugate folds are developed. This change in style of the F_2 folds that occurs in the Langstrand area seems to be reflected in the Husfjord area, where a similar change takes place as the belt of metasediments swings from a N-S strike to an E-W strike.

F_2 minor folds occur in many different sizes, varying from a few centimetres to several metres in amplitude. The shapes of the F_2 folds are almost invariably characterized by thickened hinges, sometimes considerably thickened. In many cases it would appear that the limbs have also been somewhat attenuated, and this is especially true of those folds having a monoclinic symmetry. In these cases, the folds are of similar fold type, and may well have formed by a mechanism involving a certain amount of flattening in the ac-section as described by Ramsay (1962b). (Plate 3 should be compared with Fig. 8 in Ramsay, 1962b).

In the symmetrical folds of the E-W belt the hinges are thickened and the folds appear to be of similar type. In the case of the conjugate folds, it is probable that a mechanism such as that described by Ramsay (1962a) was in operation, although there is generally no sign of the conjugate shears. Absence of marked shears probably indicates that the state of the rocks at the time of formation of the folds was not very brittle.

On many of the banding surfaces, particularly of the psammites, there is a lineation which is due to microfolding in the banding. This lineation is parallel to the axes of the F_2 minor folds and is thus a b-lineation formed during the F_2 movements. Any lineation associated with earlier folding episodes must have since been eliminated.

In those horizons of the metasediments where there are massive bands of a highly competent lithology intercalated with bands of less competent rock, boudinage structures are sometimes developed. These are tectonic structures which are generally considered to be due to a relative stretching of the bands in a direction parallel to the banding (Ramberg, 1955; Rast 1956). The internal tension thus set up along the banding tends to cause rupture of the competent band whilst the incompetent bands on either side yield to deformation in a plastic fashion.

Boudinage structures are fairly common in the psammites and semi-pelites group. Here, the more competent bands form elongate boudins which range up to a metre in length and a few tens of centimetres in thickness. The forms of the boudins vary; those formed in the more massive bands have a length/thickness ratio smaller than those in the less massive bands, and have a lozenge shape. Those occurring in less massive bands are more elongate and are sausage-shaped (Plate 9).

Plate 9. Boudinage in slightly migmatized psammites and semi-pelites. Kobberfjord.

Plate 10. Tectonic inclusions formed by basic sheets. Kobberfjord.

In many cases complete rupture has not taken place, and the boudins have not become separate entities but form a string of connected 'sausages' (Plate 9). In these cases the boudins are in an incipient condition, and the initial stages of the formation of boudinage structures is demonstrated by the 'necking' or extreme thinning of the competent bands (cf. Rast 1956). In some of the boudinage structures where rupture has taken place, there is evidence of movement along the plane of fracture. Adjacent boudins are slightly off-set relative to one another along this plane, which at least in some cases is parallel to a prominent joint direction in the rocks (Plate 9). It is to be noted that this joint direction is not perfectly at right angles to the banding so that there is a tendency for a slight rotation of the boudins to take place, thus causing the off-setting relationship (cf. Rast 1956). There is not usually any significant amount of mineral segregation in the nodes.

A more advanced stage in the development of boudinage is shown by some of the basic sheets. Here, the process has continued beyond the actual rupturing of the sheets, and the boudins have become clearly separated from one another, forming tectonic inclusions (Plate 10). This is the last stage in the evolution of boudins as described by Rast (1956). Some of the thick basic sheets form large boudins which are separated from one another by several metres.

Where a three-dimensional picture of boudins is obtained, it can be seen that they are elongated in a direction which is parallel to the trend of the F_2 folds in that area. Many of the basic sheet boudins exhibit a b-lineation due to the orientation of minerals. It would seem that boudinage structures formed as a result of internal tensional forces set up in the metasediments during the F_2 movements. Boudinage structures of F_2 age are described by Roberts in the Langstrand area (Roberts 1965).

Late F_2 Structures

The right-angled swing in the strike of the metasediments that has already been mentioned is due to a large open warp-fold, which was formed during the last phase of F_2 movements in the area. The axis of this fold plunges at about 60° to the W.N.W. To the north-east of the Husfjord area the strike of the metasediments again swings to a N-S trend (Roberts, 1965). Roberts considers that this orthogonal swing in the Langstrand area is not due to a separate post-F_2 tectonic episode. His argument is that the changes in the symmetry of the F_2 folds are dependent upon whether they are on an E-W belt or a N-S belt. This indicates that the formation of E-W belts and N-S belts could not significantly post-date the main F_2 movements, but must be a part of them. Alternatively, he suggests that the possibility of the swing being a primary feature of the F_1 folding may not be entirely ruled out.

In the Husfjord area, the latter is clearly not the case since the igneous complex, which is entirely post-F_1, is deformed by the open warp-fold. It appears that the folding was associated with the F_2 movements, but it was probably a comparatively late phase of the

F_2 episode since it deforms some members of the igneous complex which were emplaced after the early stages of the F_2 movements. The evidence provided by the symmetry of the F_2 minor folds as discussed above, strongly supports the view that the warp-folding is associated with the F_2 movements, presumably as a late-stage accommodation structure.

The form of this late large warp-fold is shown well by the regional folding of diorite bands that have been emplaced into the Husfjord gabbro (see relevant section under Husfjord metagabbro, p. 72.) These diorites are subparallel to the strike of the metasediments, and a stereographic plot of the fold is given in this chapter since it is of relevance to the general structure (Fig. 6).

The latest stage of deformation was of a relatively brittle nature, and numerous late shear-zones and mylonites formed in connection with it. These are in evidence in the rocks of the igneous complex, and will be considered in more detail when the appropriate rocks are being described. However, it is instructive to insert into the present chapter a stereographic plot of the attitude of the larger of these shear zones and mylonites (Fig. 7). It is to be noted that the general dip of these shears is about 35°-40° to the S.S.W. The strike direction of the shear planes is W.N.W.-E.S.E., which is parallel to the trend of the axis of the late warp-fold. They are low-angle shears, possibly related to the late F_2 folding. They may form one set of a conjugate pair which developed to the exclusion of the complementary set.

The formation of these shears and mylonites appears to be the last tectonic activity to be recorded in the Husfjord area.

STEREOGRAPHIC PLOT OF POLES TO EARLY DIORITE BANDS IN HUSFJORD METAGABBRO

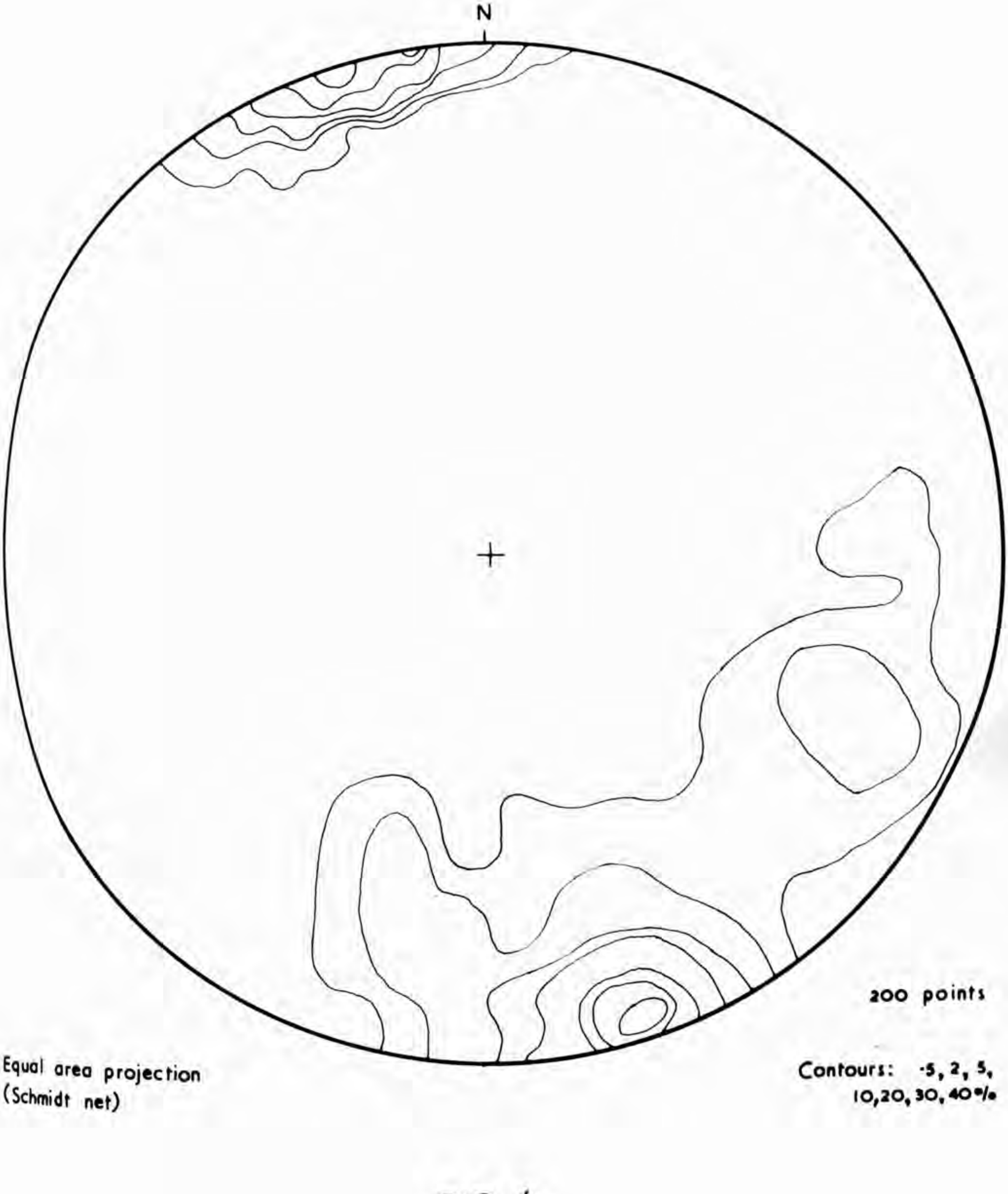


FIG. 6

STEOGRAPH!C PLOT OF SHEAR ZONES AND MYLONITES
HUSFJORD METAGABBRO AND VATNA GABBRO

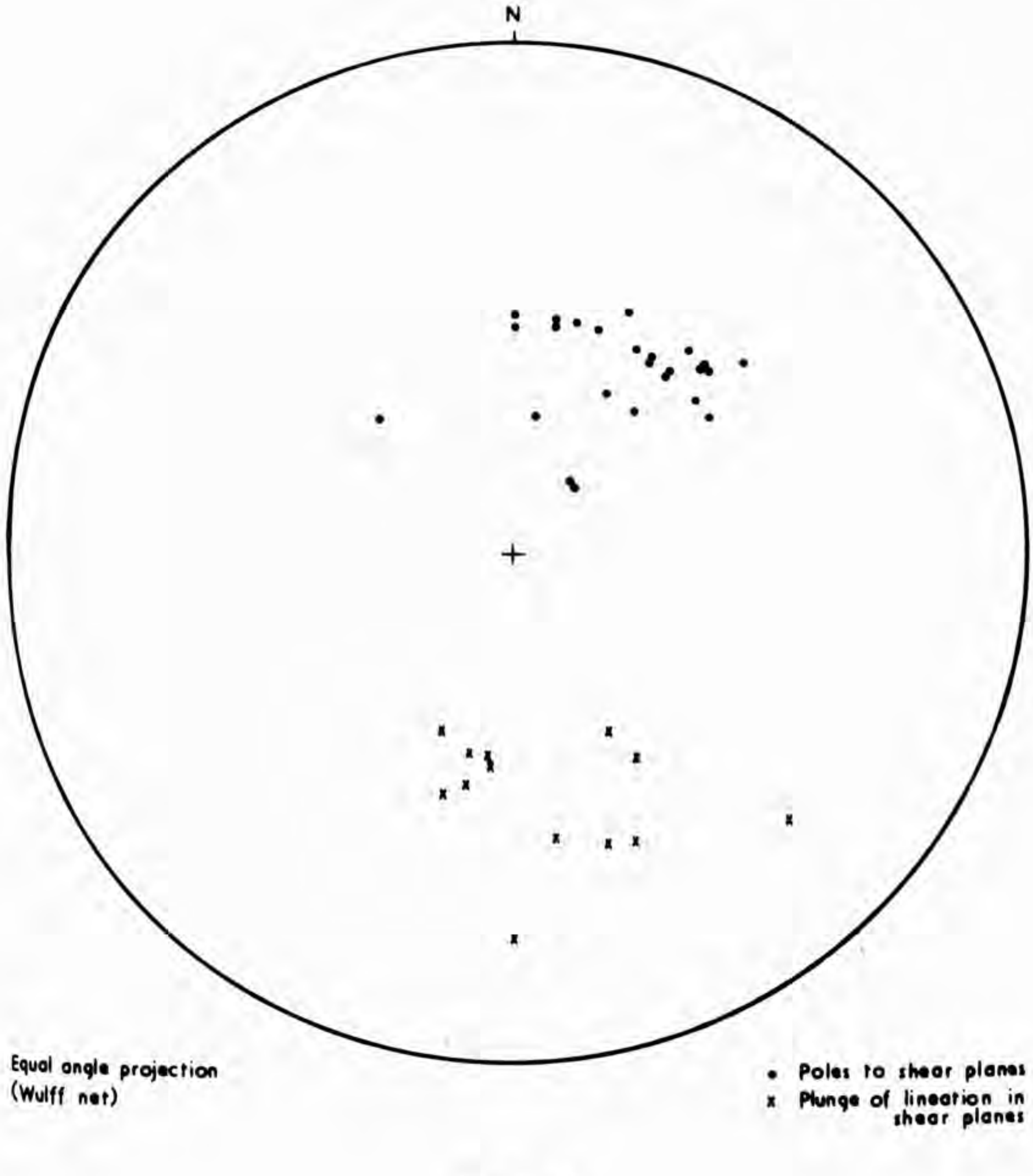


FIG. 7

MAJOR STRUCTURES

Study of the minor structures in the Husfjord area, and the previous work of Ramsay and Sturt (1963) and Roberts (1965) have shown that in Sørøy there have been two major episodes of folding, which are termed for convenience F_1 and F_2. The former was characterized by large recumbent isoclinal folds overturned towards the east and south; the F_2 folds, which are often seen to refold F_1 structures, are either orthorhombic or monoclinic in style, and in the latter case verge towards the south-east.

The present construction of the regional structures has been made from the correlation of the belt stratigraphy by these previous workers. Particularly useful marker horizons have been provided by a limestone belt, the Falkenes Limestone, and by a belt of migmatites (Fig. 8).

The north-western outcrop of the limestone belt is an F_1 synclinal core which closes downwards (Sturt and Ramsay, personal communication). In the Dønnesfjord area it is a complex but unified core which divides into two principal discrete cores when traced either westwards or eastwards (Sturt, personal communication). Towards the east, the southern core dies out at Lotre (Ramsay, personal communication).

The south-eastern outcrop of the limestone lies on the upper limb of the F_1 Hønesby Fold in N.E. Sørøy (Roberts, 1965), and as it is traced to the S.W. appears to run towards the Husfjord area, although the intervening area is as yet unmapped. In the Husfjord area the presumed continuation of this limestone belt has been engulfed by the Husfjord

MAP OF SØRØY SHOWING THE BELT STRATIGRAPHY

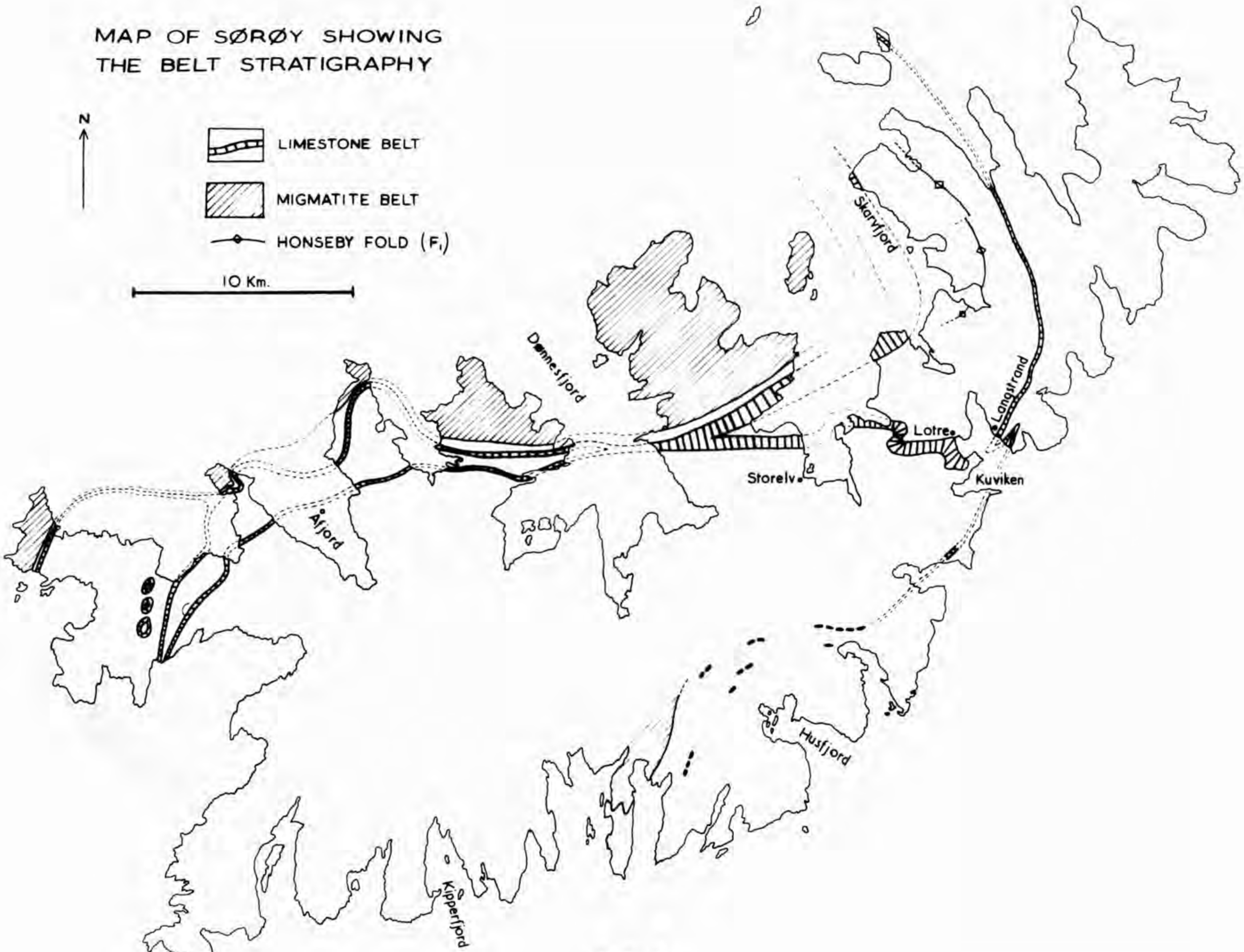


FIG. 8

metagabbro sheet and now occurs as rafts within the metagabbro (see p. 65.) At Langstrand the limestone is folded around the nose of a tight F_1 synclinal fold, probably a continuation of the syncline at Lotre, and a complementary F_1 anticline (see Fig. 8). This small F_1 anticline would appear to open out when traced towards the south-west, and the presence of a large F_1 fold axis through the central part of Sørøy has been reported by Ramsay (personal communication). Large rafts of limestone also occur within the southern extension of the Storelv gabbro to the north of Kipperfjord (Sturt, personal communication).

The rocks occurring between the branches of the divided synclinal core are mainly psammites, the Klubben Quartzite Group, and are only sporadically migmatized. However, the Klubben Quartzite Group outcropping in the central part of Sørøy and to the north of the north-western limestone belt is extensively migmatized (Sturt, personal communication).

For convenience in the following discussion, it is proposed to call the F_1 synclinal core the Storelv Fold, and the F_1 anticline the Kuvik Fold, since it appears to pass through Kuviken (see Fig. 8).

A simplified diagrammatic reconstruction of the structure of N.E. Sørøy, after Roberts (1965), appears in Fig. 9A. The F_1 Hønesby Fold has been refolded by F_2 folds which verge towards the S.E. In the southern part of the area mapped by Roberts, the present erosion surface is at a higher tectonic level than it is in the north of that area (see Fig. 9A). Thus it is reasonable to suppose that further to the south the erosion surface intersects the structure at even higher

SECTIONS SHOWING A RECONSTRUCTION OF THE STRUCTURE OF SØRØY

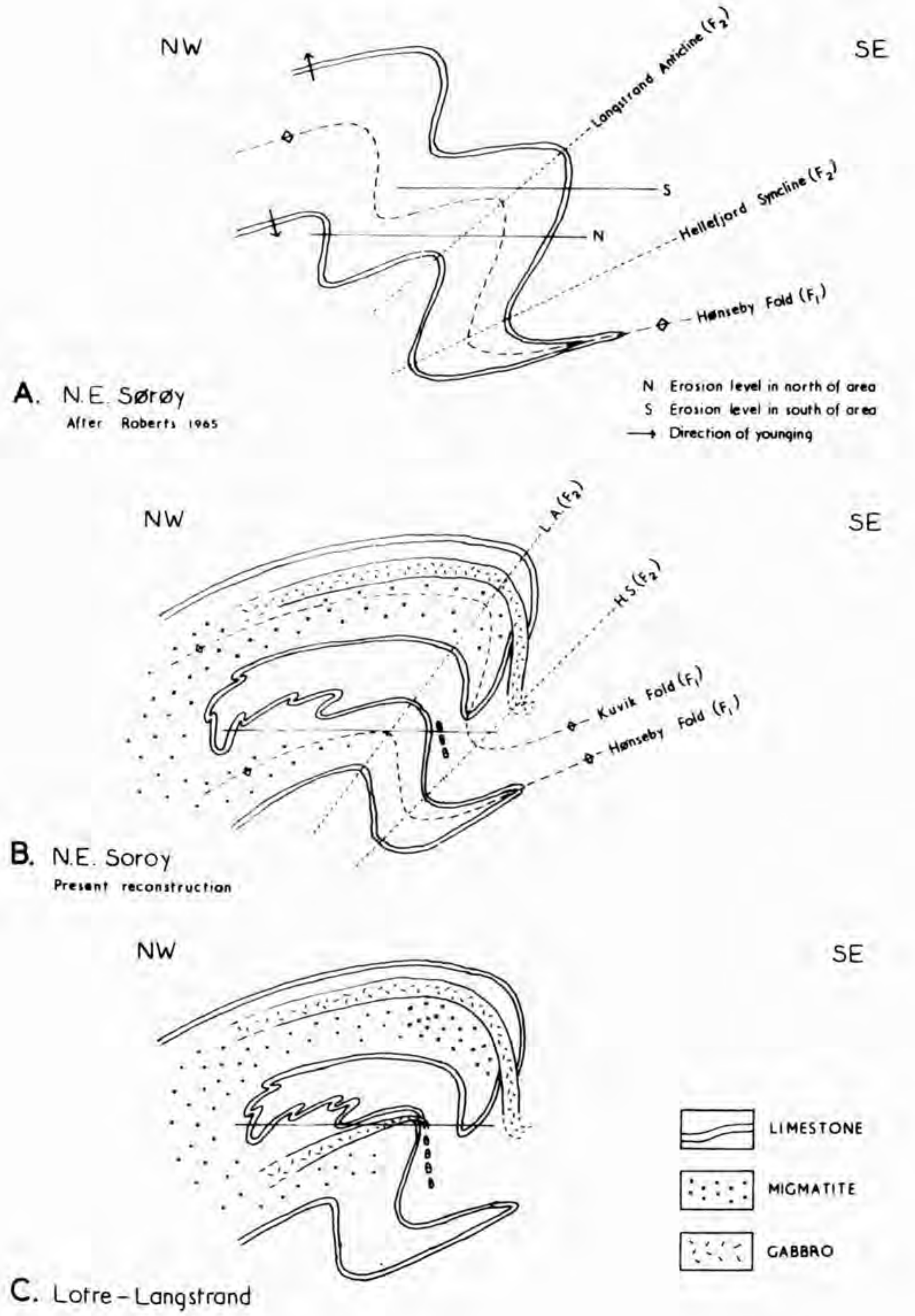


FIG. 9

tectonic levels.

In the present writer's reconstruction it is considered that the Storelv Fold is the complementary F_1 syncline to the upper limb of the Hønesby Fold and the lower limb of the Kuvik Fold (Fig. 9B). In most of Roberts' area the erosion surface is lower than both the Storelv Fold and the Kuvik Fold. It is mainly at Langstrand that these folds are seen, and here their axes are close to one another because of the proximity of the closure of the Kuvik Fold (Fig. 9C). At this tectonic level the southern branch of the Storelv Fold core intersects the erosion surface, forming the limestone belt which peters out at Lotre (Fig. 9C). Further to the west, at higher tectonic levels, this limestone core merges with the northern branch to form a complex unified core at Dønnesfjord (Figs. 8 and 10A).

Meanwhile, the Kuvik Fold has opened out at these higher tectonic levels so that the rocks in its core constitute much of the central part of Sørøy. The Husfjord area lies on the upper limb of the Kuvik Fold which has been overturned by a large F_2 fold, possibly a continuation of the Langstrand Antiform (Fig. 10A). This accounts for the attitude of the metasediments in the Husfjord area, i.e. dipping steeply to the north and west whilst younging towards the south-east.

Further to the S.W. the Storelv Fold again splits into two distinct cores (Fig. 10B). If the limestone rafts occurring within the southern extension of the Storelv gabbro belong to the Falkenes Limestone Group, then it is possible that they may represent the narrow belt of limestone which should occur between Klubben Quartzite of the two F_1 anticlines

SECTIONS SHOWING A RECONSTRUCTION OF THE STRUCTURE OF SØRØY

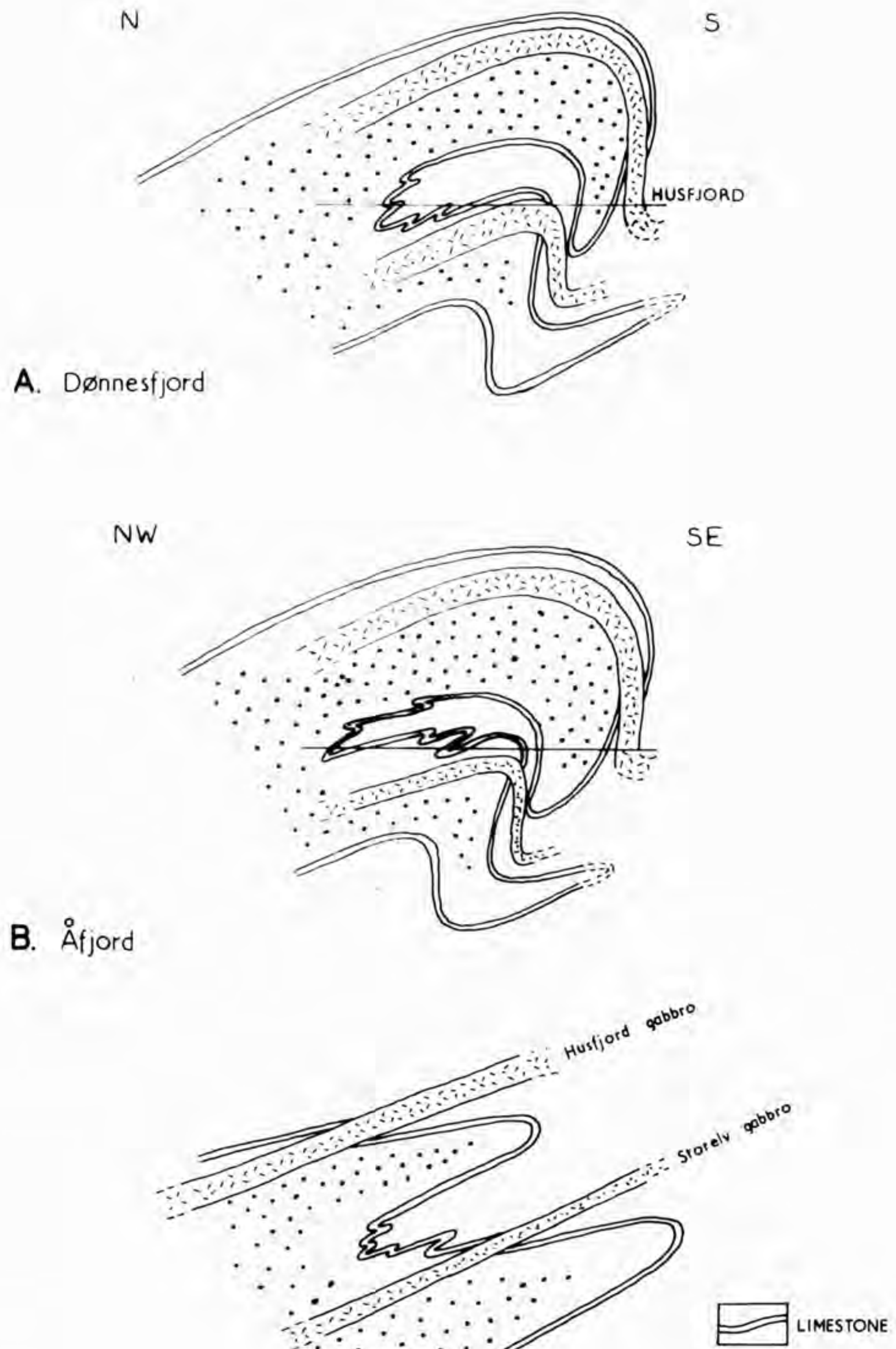


FIG. 10

(see Fig. 10B). If this is the case, then it indicates that the Storelv gabbro sheet moves up through the stratigraphic succession towards the south, since at Storelv it lies beneath a schist group, the Storelv Schist, which underlies the Falkenes Limestone.

The Storelv gabbro also moves up through the succession as it is traced eastwards from Storelv (Sturt and Ramsay, personal communication). Furthermore, it becomes much narrower and considerably sheared in this direction, and is probably represented by the large, sheared tectonic lenses of amphibolite occurring in the Hellefjord Schists in N.E. Sørøy (Roberts, 1965; see also Figs. 9B and 9C for reconstruction).

In the diagrams of the proposed reconstruction of the structure of Sørøy, the representation of the F_2 folds are generalized since there are many minor complications. The pre-F_2 structure probably approximated to that shown in Fig. 10C, which also shows the positions of two principal syn-F_1 gabbro sheets (see Fig. 4). These gabbros were emplaced along the limbs of the F_1 isoclinal folds, probably subparallel to the axial planes. This accounts for the marked discordancy shown by both the Storelv gabbro and the Husfjord metagabbro towards the N.E., where they approach the closures of the Hønesby Fold and Kuvik Fold respectively (see Figs. 8 and 10).

The proposed reconstruction explains the sudden closing of the southern branch of the Storelv Fold core at Lotre, and also the puzzling distribution of the migmatites. It is suggested that the migmatization may have occurred more intensely in the cores of the large F_1 folds.

If this is the case, then it can be seen from the sections in Figs. 9 and 10 that the areas between the branches of the divided parts of the Storelv Fold have not suffered migmatization (cf. Fig. 8). On the other hand, the migmatites occurring in the centre of Sørøy are within the core of the Kuvik Fold, which outcrops again to the north of the Storelv Fold, accounting for the reappearance of the migmatites in northern and western Sørøy (see Fig. 8).

If the proposed hypothesis approximates to the structure obtaining in Sørøy, then a slight adjustment to Roberts' reconstruction (1965) must be made. The subvertical belt of limestone younging to the west which outcrops in the Skarvfjord area (Fig. 8) was considered by Roberts to be the lower limb of the Hønesby Fold (see Fig. 9A). At that time, mapping of the neighbouring areas had not progressed very far, and the synclinal nature of the northern branch of the Storelv limestone belt was not appreciated. In the proposed reconstruction it is suggested that this limestone belt may not be the lower limb of the Hønesby Fold, but rather part of the synclinal core of the Storelv Fold.

Although areas of Sørøy still remain to be mapped in detail, it is believed that the present structural synthesis provides a reasonable working model as a structural background for the Husfjord plutonic igneous complex.

PETROGRAPHY

Psammites and Semi-pelites

The psammites, which show varying degrees of migmatization, are now usually in the condition of gneiss. The most abundant mineral is quartz, which together with feldspar forms an equigranular texture of polygonal, straight-edged grains. The grains are sometimes slightly elongated parallel to the banding in the rock which is produced by zones of small biotite and muscovite laths. The predominant feldspar is plagioclase of oligoclase or sodic andesine composition, although potash feldspar also occurs and in some cases this can be seen to be microcline. The potash feldspar is occasionally fringed by a little myrmekite where contact is made with the plagioclase, and it is common to find the feldspars slightly sericitized. Garnet is a common mineral, and is often fairly large. Accessory minerals include ore, zircon, and a little secondary calcite, the latter mainly associated with muscovite.

The equigranular texture of the groundmass is frequently interrupted by megacrysts of potash feldspar or large areas containing potash feldspar crystals. These are porphyroblasts which have grown during the feldspathization associated with the migmatization that has affected these rocks (see p. 285). These megacrysts are often ovoid in shape, and their margins commonly have a mortar texture, indicating that they have been slightly augened during a post-migmatization episode of deformation. Further evidence of deformation is provided by the plagioclases and the micas which are frequently kinked, and by many of the quartz grains which exhibit undulose extinction.

The garnets vary in size according to their environment. In the quartzo-feldspathic bands where there is very little biotite there are small garnets, whereas larger garnets have grown in the biotite-rich bands. The garnets are colourless and are allotrioblastic to hypidioblastic, with numerous inclusions of biotites and rounded quartz grains. They are usually directly associated with biotite, and in most cases they are overgrowing the biotite laths, from which they can sometimes be seen to be forming.

Although the garnets are fairly fresh-looking, they are sometimes slightly cracked. Small biotite flakes occasionally occur along these cracks, and are probably forming at the expense of the garnet. Occasionally, garnets are slightly augened, indicating that at least some of the garnets pre-date an episode of deformation. In one case there appears to have been rotation of the garnet, since an included biotite lath is subperpendicular to the schistosity outside the garnets (Plate 11).

In the semi-pelitic horizons of the psammites and semi-pelites group, biotite is more abundant than in the psammites. It forms broad zones together with garnet, kyanite, sillimanite, and rutile. Between the zones are quartzo-feldspathic bands in which the plagioclase is of oligoclase composition. Accessory minerals include ore and zircon.

Potash feldspar commonly forms megacrysts which are porphyroblasts formed during the migmatization. These porphyroblasts sometimes contain

Plate 11. Pre-F_2 garnet. Semi-pelite. P.P.L. X 25.

rounded inclusions of biotite and quartz. Many of the potash feldspar/ potash feldspar grain boundaries are highly sutured, and this together with the widespread occurrence of myrmekite at plagioclase/potash feldspar grain boundaries appears to be a later metamorphic effect.

The biotite is altering to colourless kyanite, which overgrows the biotite and contains ghost inclusionsof it. Biotite also nucleates small sillimanite prisms and needles which often grow along the cleavage of the biotite. Rutile is commonly associated with the biotite-kyanite-sillimanite aggregates, and is presumably a result of the breakdown of biotite.

In the most intensely migmatized parts, sillimanite also sometimes occurs as fine fibrolitic needles along feldspar boundaries. They exist particularly at potash feldspar/potash feldspar boundaries, but also occasionally at plagioclase/plagioclase contacts and at potash feldspar/plagioclase contacts. In the latter case they are usually involved with myrmekite. Generally the y form randomly-orientated or fan-like aggregates (Plate 12), although sometimes there is a tendency for them to grow subperpendicular to the feldspar boundaries (see p.288 for further discussion).

The garnets are closely associated with the biotite-kyanite-sillimanite aggregates, and overprint the F_1 schistosity (Plate 13). They are allotrioblastic to hypidioblastic and contain inclusions of quartz, biotite, kyanite, and rutile. In places they can be seen to be forming at the expense of biotite and kyanite, indicating that some of the garnets formation post-dated the growth of kyanite. Many

Plate 12. Fibrolite at feldspar boundaries. Semi-pelite. P.P.L. X 95.

Plate 13. Static-growth garnet. Semi-pelite. P.P.L. X 85.

of the garnets are slightly cracked, and there are indications of insipient alteration to secondary biotite, especially along the cracks.

Basic Sheets

The basic sheets which have been emplaced into the psammites and semi-pelites are now essentially amphibolites.

The early, pre-F_1 sheets consist principally of hornblende which has a preferred orientation giving the rock a crude schistosity. This hornblende has the following pleochroic scheme :-

X pale green

Y yellowish-green

Z yellowish-green

Within the amphibole masses there are relics of a clinopyroxene from which the hornblende has formed. Some of the hornblende, and some of the remaining pyroxene are altering to biotite. The composition of the plagioclase is at the andesine-labradorite border.

The later, post-F_1 sheets are also mainly composed of hornblende which has almost completely replaced the original clinopyroxene. The pleochroism for this hornblende is

X very pale green

Y pale green

Z pale yellowish-green

There are a few green serpentine pseudomorphs after olivine. It is noticeable that the plagioclase, which has a composition of fairly basic labradorite, tends to be orientated with twin planes subparallel to the crude schistosity of the rock. The (010) faces do not form perfectly to give the feldspars a platy habit, but it is possible that potential (010) faces, parallel to the twinning, have provided a structural control during growth, causing the preferred orientation.

Cummingtonite-schists

The predominant mineral in these schists is the colourless amphibole cummingtonite, which occasionally attains euhedral shapes. Some bands are composed entirely of cummingtonite in which the crystals are intergrown in a matted form, with preferred orientations. The mineral is biaxial positive, and basal sections are hexagonal and not rhombic, distinguishing it from the rather similar calc-silicate mineral tremolite.

In other bands, cummingtonite predominates, but there is also quartz and intergrown biotite laths which are elongated parallel to the cummingtonite crystals.

The schistosity, which is delineated by the biotites and cummingtonites, shows slight folding. The folds are gentle undulating buckles which are impersistent when traced both along the schistosity and along the axial planes of the buckles.

Garnet-mica-schists

The schistosity, which is sometimes seen to be buckle~~s~~, is delineated by the micas, which include both biotite and muscovite. The principal felsic mineral is quartz, and this forms a mos~~s~~aic of irregular interlocking grains. There is a tendency for the quartz grains to be slightly elongate in a direction parallel to the schistosity. There is also a little feldspar which is of oligoclase composition.

Poikiloblastic garnets overprint the schistosity and they enclose grains of quartz and biotite. These garnets, which are allotrioblastic, are not altering to any secondary products.

In the less ferromagnesian bands kyanite develops, often forming fairly large crystals and showing strong cleavage and (001) parting. They are frequently poikiloblastic and contain inclusions of muscovite and biotite. Apatite is a common accessory mineral in these bands and is often enclosed within muscovite.

Kyanite-garnet-mica-schists

An F_1 schistosity is formed by strings of micas which have a preferred orientation parallel to the schistosity. Kyanite crystals are commonly associated with the micas. The remaining felsic minerals, which consist principally of quartz together with some potash feldspar and oligoclase, form an inequigranular mosaic in which grain-boundaries are usually irregular. Some of the quartz grains are elongated parallel to the schistosity, and much of the quartz exhibits undulose extinction showing it to have undergone strain. Further evidence of deformation is given by the slight buckling of the schistosity, and the occasional augening of kyanites or aggregates of kyanites.

Kyanite overprints the biotite, which, in turn, is replacing some of the muscovite, particularly along cleavages. It is allotrioblastic to hypidioblastic, and cleavage is only spasmodically developed, but when present it is often very marked, together with a strong (001) parting. Frequently the kyanite is poikiloblastic, containing inclusions of quartz, biotite, and sphenoid rutile. In places kyanite appears to be forming at the expense of biotite. Strings of small sillimanite prisms are also growing from biotite and many develop idioblastic forms. These prisms, together with diffuse relict biotite laths, form aggregates elongated parallel to the schistosity.

Both kyanite and sillimanite are altering to late muscovite, which often grows across the schistosity (Plate 14). The muscovite forms poikiloblastic laths which often contain inclusions of kyanite, and sillimanite as well as occasional quartz blebs (Plate 15). This alteration seems to take place more commonly in the neighbourhood of potash feldspar. Presumably the feldspar contributed potassium to the reaction which must have taken place during a later lower-grade metamorphism post-dating the F_2 deformation, as the muscovites have clearly formed under static conditions, and are undeformed.

The garnets overprint the biotites and early muscovites which delineate the F_1 schistosity (Plate 16). Crystals are rounded and sometimes contain inclusions of quartz, biotite, and sphenoid rutile. Occasionally garnet is overgrowing sillimanite and kyanite (Plate 17).

The occasional calc-silicate bands and lenses that occur in the kyanite-garnet-mica-schist consist almost entirely of pale green tremolite-actinolite. This forms a mat of intergrown allotrioblastic poikiloblastic crystals which are full of inclusions of quartz, sphene, and diffuse biotite. In many examples the biotite inclusions have a reaction halo

Plate 14. Sillimanite altering to muscovite. Kyanite-garnet-mica-schist. X-Pl. X 170.

Plate 15. Poikiloblastic muscovite. Kyanite-garnet-mica-schist. X-Pl. X 95.

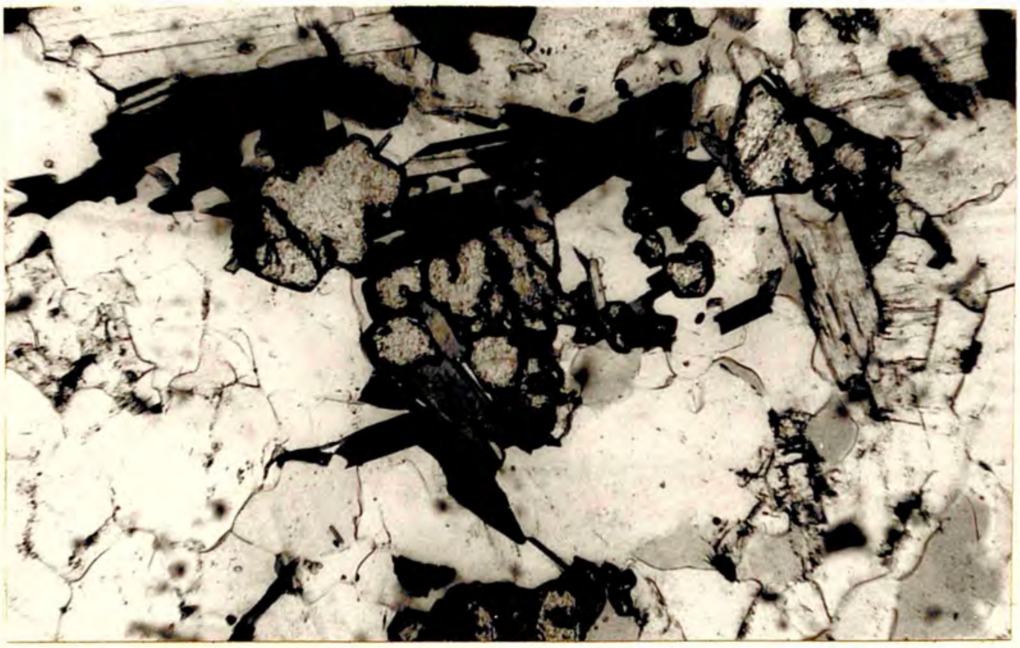


Plate 16. Post-F_1 garnet. Kyanite-garnet-mica-schist. P.P.L. X 95.

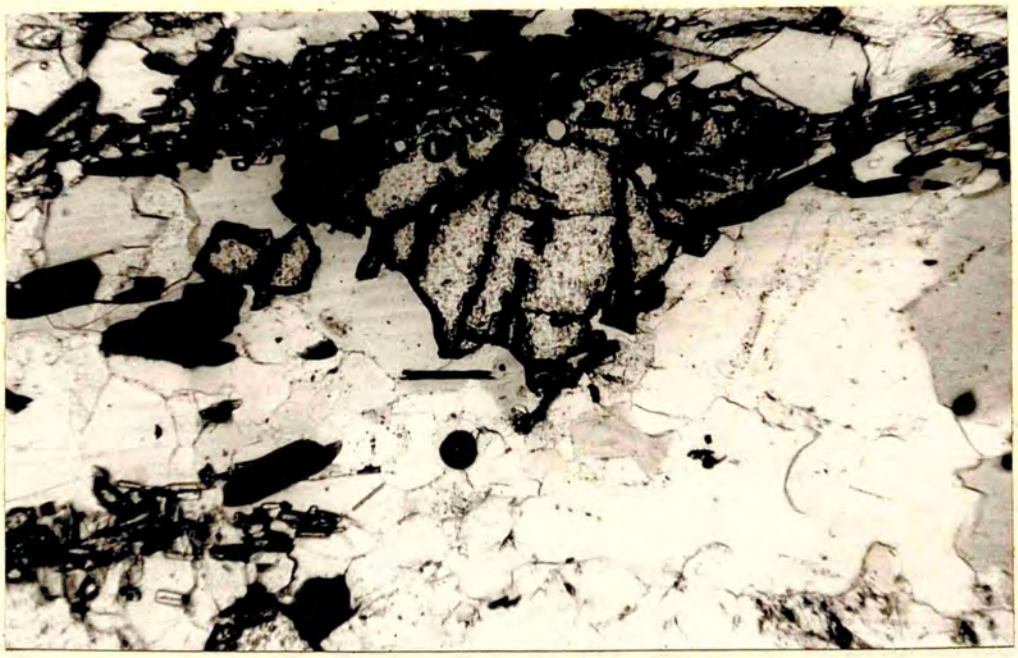


Plate 17. Garnet overgrowing sillimanite. Kyanite-garnet-mica-schist. P.P.L. X 95.

around their margins suggesting that they are altering to the tremolite-actinolite.

Calc-silicate-schists with Metalimestones

It is in the calc-silicate-schists near to the contact with the igneous complex that hornfelsic textures have been preserved. These provide evidence that the country rocks have been thermally metamorphosed by the outer member of the igneous complex, i.e. the Husfjord metagabbro.

Near the contact the calc-silicates have developed a granoblastic hornfelsic texture in which small granular diopside crystals and quartz grains are sometimes accompanied by small scapolite crystals. At the contact, aggregates of buff-coloured garnets develop, and these are intergrown with an equigranular mosaic of scapolite and green diopside.

In many cases, the granular hornfelsic texture shown by these rocks has been overprinted by a regional metamorphic texture and mineralogy, so that the evidence for hornfelsing has been preserved as relics.

Pale green tremolite-actinolite commonly overgrows the hornfelsic texture. These amphiboles are usually poikiloblastic and contain rounded granular inclusions of diopside as relics from the hornfels texture. Sometimes pale phlogopitic biotite laths are also enclosed within actinolite, and together with the diopside appear to be altering to the amphibole. Occasionally small sphenes are included within actinolite.

In one quartzo-feldspathic band containing patches of pale green calc-silicate minerals along one edge, very large green actinolites develop. These are poikiloblastic, with inclusions of rounded diopside, quartz, and feldspar inherited from the hornfelsic texture. The tremolite-actinolite crystals often tend to have a preferred orientation, and in some rocks biotite laths have begun to orientate themselves into a crude schistosity. Also in one less calcic band, new biotite is forming from hornblende and some grains have become elongated parallel to this schistosity. In some bands small poikiloblastic clinozoisites have developed.

The garnets which have grown in the calc-silicate-schists at the contact with the igneous complex have been slightly altered, and cracks and cavities have been filled with calcite.

The limestones in this group consist principally of an inequigranular mosaic of calcite crystals with straight margins. Scattered between the calcite crystals are small rounded diopsides.

Quartzo-feldspathic-phyllitic-schists

Quartz and oligoclase are the predominant minerals in these rocks, and together with a little potash-feldspar form an inequigranular mosaic with very irregular boundaries. Quartz generally exhibits undulose extinction, and the feldspar is slightly sericitized.

Micas are more abundant in the phyllitic schists than in the more massive quartzo-feldspathic bands, and include both biotite and muscovite. The latter is often poikiloblastic, containing inclusions of quartz and of biotite.

Garnets are not common, and are small and allotrioblastic, usually being rounded. They overgrow biotite and appear to be almost unaltered, although in some garnets there is a bire-fringent mica-like mineral growing along cracks.

Pegmatites which occur in these rocks consist of quartz and poorly-twinned sodic plagioclase which is antiperthitic, containing exsolution patches of microcline. A patch-antiperthite is formed with very irregular, randomly-orientated microcline patches of many different sizes enclosed in plagioclase. The feldspars are slightly sericitized, an alteration which affects both the potash-feldspar and the plagioclase.

PART III

THE IGNEOUS COMPLEX

THE IGNEOUS COMPLEX

FIELD RELATIONSHIPS

Introduction

The Husfjord igneous complex comprises major plutonic intrusions, principally gabbros and diorites, and a number of minor intrusions, including basic dykes and perthosite sheets, which have been emplaced at different stages in the development of the Caledonian orogeny in this region. The earliest intrusion of the complex is the Husfjord metagabbro, and this was not emplaced until the end of the earliest phase of deformation F_1. The intrusion of this gabbro was followed by the peak of the regional metamorphic event, which in this area reached the almandine amphibolite grade. After this came the second major phase of deformation F_2 which was followed by the emplacement of a suite of diorites, although some of the diorites were contemporaneous with F_2 movements. The major intrusion of this suite was a fine-grained body, the Havnefjord diorite, which was both preceded and succeeded by smaller diorite bodies, some of which were coarse-grained. During this period after the second deformation phase, the Vatna gabbro and its associated minor intrusions were emplaced. The final stages in the history of the area are characterized by metamorphic and structural events which post-date all the igneous activity. Thus it can be seen that the emplacement of the entire plutonic complex occurred within the period during which the orogeny took place.

THE MAJOR INTRUSIONS

HUSFJORD METAGABBRO

General

The Husfjord metagabbro was the first member of the plutonic igneous complex to be emplaced, and was intruded into the metasediments as a subconcordant sheet.

It is a very resistant rock and forms the high mountains around Husfjord. These mountains are steep sided and are surmounted by dark, steep, rugged cliffs and pinnacles of prominently-jointed metagabbro.

Its outcrop runs from Ramnes in the south-west to Gjelting in the north-east and has an arcuate form. This shape of the outcrop is due to the large-scale open folds of late F_2 age (see p. 39). The maximum width of the outcrop is about 6 km., but much of the central zone is now occupied by later rocks which have been intruded into the metagabbro. Its north-western contact is an intrusive contact against the metasediments of the country rocks, but the original contact against country rocks to the south-east is nowhere to be seen now. The only metagabbro contacts observed in the south-east are those against younger intrusive rocks at Vatna; elsewhere the rocks occurring to the south-east of the gabbro have been eroded by the sea.

The gabbro is a melagabbro which has undergone mineralogical and textural alterations during subsequent metamorphism. It is fairly uniform in composition and mineralogy throughout, and does not contain any leucogabbroic or anorthositic facies, which are present in the

Breivikbotn gabbro to the north-west (Stumpfl and Sturt, 1965).

The gabbro consists essentially of clinopyroxene, plagioclase, and iron ore, although hypersthene occasionally appears in some facies. The pyroxenes show progressive alteration to green hornblende during regional metamorphism, and there is a relict sub-ophitic texture formed by amphibolized pyroxene and recrystallized plagioclase.

The grain-size of the gabbro is slightly variable. Typically it is fine to medium-grained with average grain-size in the order of 1-2 mm., but occasionally it becomes more coarse-grained, and crystals can be seen clearly with the naked eye. The secondary amphiboles appear black or dark green, and plagioclase has often weathered to a yellowish colour. This coarsening of the grain-size is not a metamorphic effect, but a primary feature of the gabbro; in thin-section the original pyroxenes are seen to be larger than in the finer-grained parts. The development of the slightly coarser-grained gabbro is sporadic, and the areas and patches in which it occurs are irregular and do not form any kind of layering. It is possible that these areas represent 'pods' in the cooling magma, in which there was a slight concentration of volatiles enabling larger crystals to grow.

There are also areas in which the gabbro is finer-grained than usual, and in which the crystals rarely exceed about 1 mm. in size. In these finer-grained areas the gabbro sometimes has a very faint and poorly-formed foliation which is orientated parallel to the margin of the gabbro body, although this foliated finer-grained facies does not occur in the contact zone. The main outcrop of this facies is in the central part of the body, to the south of the later Havnafjord diorite, in a zone abundant in narrow coarse-grained diorite bands which have been intruded into the gabbro. The foliation, which is due to sub-parallel orientation of primary minerals, may be a fluxion structure, or may possibly indicate that the gabbro was intruded synchronous with some slight tectonic movement, presumably associated with F_1. If the latter is the case, in which the minerals orientated themselves in the most stable position within the stress-field, then it is clear that the strength of the deformation was not great, for it did not cause shearing of the gabbro at this time. The only shearing that the gabbro has suffered is of F_2 age. The F_1 movements were evidently on the wane, and had almost ceased before the intrusion of the gabbro. It is obvious that this zone of finer-grained, slightly foliated gabbro has provided an area of structural weakness in the sheet into which the later narrow coarse-grained dioritic bands have been readily emplaced.

The north-western contact with the metasediments is well-devined, but the actual contact was not found within a single outcrop because of the differential weathering of the rock types. There is no evidence of appreciable chilling of the gabbro against the country rocks.

As mentioned (p. 14) in connection with the stratigraphy of the metasediments, the gabbro contact is slightly discordant with the banding of the country rocks. At any point along the contact the strike of the actual contact is to the east of the strike of the banding in the meta-sediments. Since the metasediments young towards the south-east, the contact of the gabbro sheet passes up through the succession as it is traced from south-west to north-east.

There is evidence that the gabbro had a thermal effect on the neighbouring country rocks. A hornfelsic texture has been preserved in some of the beds of the calc-silicate-schist group, but it has been overgrown by later regional metamorphic textures and only remains as a relic. This is discussed further in the section on metamorphism (p.280).

For about 4 km. from the northern end of Kobberfjord, the gabbro does not now make contact with the metasediments because a later norite sheet, the Kobberfjord norite, has been intruded along the contact. In the south-east, the metagabbro makes contact with a younger gabbro, the Vatna gabbro, and in this southern part the Husfjord metagabbro is of the massive, non-foliated facies. The Vatna gabbro does not cause a marked thermal aureole to be developed in the Husfjord metagabbro. Just to the north of the Vatna gabbro, near Slåtten, three gabbro sheets apparently of similar age to the Vatna gabbro have been emplaced into the Husfjord metagabbro. Also in this south-eastern area, a few porphyritic dykes have been intruded into the Husfjord metagabbro, and these will be considered in the section on the minor intrusions (p. 132).

Ultrabasic Layers

The gabbro sheet does not show any marked, regular banding; it has neither a small-scale rhythmic layering nor a large-scale cryptic layering. The only kind of layering that is present is that formed by occasional lenses and thin bands of troctolite. These are distinctive in the field, as weathered surfaces are distinctly brown and have a knobbly appearance.

The lenses range up to 5-6 m. in length and are elongated parallel to the trend of the gabbro sheet. Usually they occur as single isolated lenses dispersed throughout the gabbro body, though there is one horizon, which outcrops on Husfjordness, in which there is a swarm of closely-packed troctolite lenses. This horizon is a zone of F_2 shearing in which the troctolites have been boudined, so that boudins laterally adjacent may in fact be part of the same ultrabasic layer which has subsequently been formed into tectonic lenses. These will be described further when the shear-belts are considered. The troctolite lenses frequently show a planar structure, apparently an internal layering, which is often accentuated by weathering (Plate 18). This plane of layering is parallel to the margins of the gabbro sheet.

Husfjordness is the only locality where continuous troctolite bands were found. These bands are usually only a few centimetres in width and maintain a constant thickness for some distance laterally, but sometimes they are only a few millimetres in thickness. In one case, a troctolite band has a sharp boundary at one margin whilst the other margin is diffuse (Plate 19). If this is a primary layering of a gravitational kind, then it is clear that the gabbro sheet has been inverted, since the sharp mafic margin is uppermost. The layering here dips steeply to the north, which means that the top of the gabbro sheet is towards the south. This is in agreement with the deduction made

Plate 18. Banding in troctolite lens. Husfjordness.

Plate 19. Troctolite band (dark band above hammer shaft). Husfjordness.

from the structures of the metasediments that the rocks in the Husfjord area are inverted, younging to the south-east. Since the inversion is due to the F_2 folding, it follows that the emplacement of the metagabbro is pre-F_2.

There is one outcrop, near Gåshopen, of a lens of hornblendite in the metagabbro. This virtually consists of hornblende which is black in hand specimens. The crystals are subhedral and grow up to several centimetres in length. They are randomly orientated with respect to one another and are interlocking. The rock has become very friable due to weathering, and many surfaces are rusty.

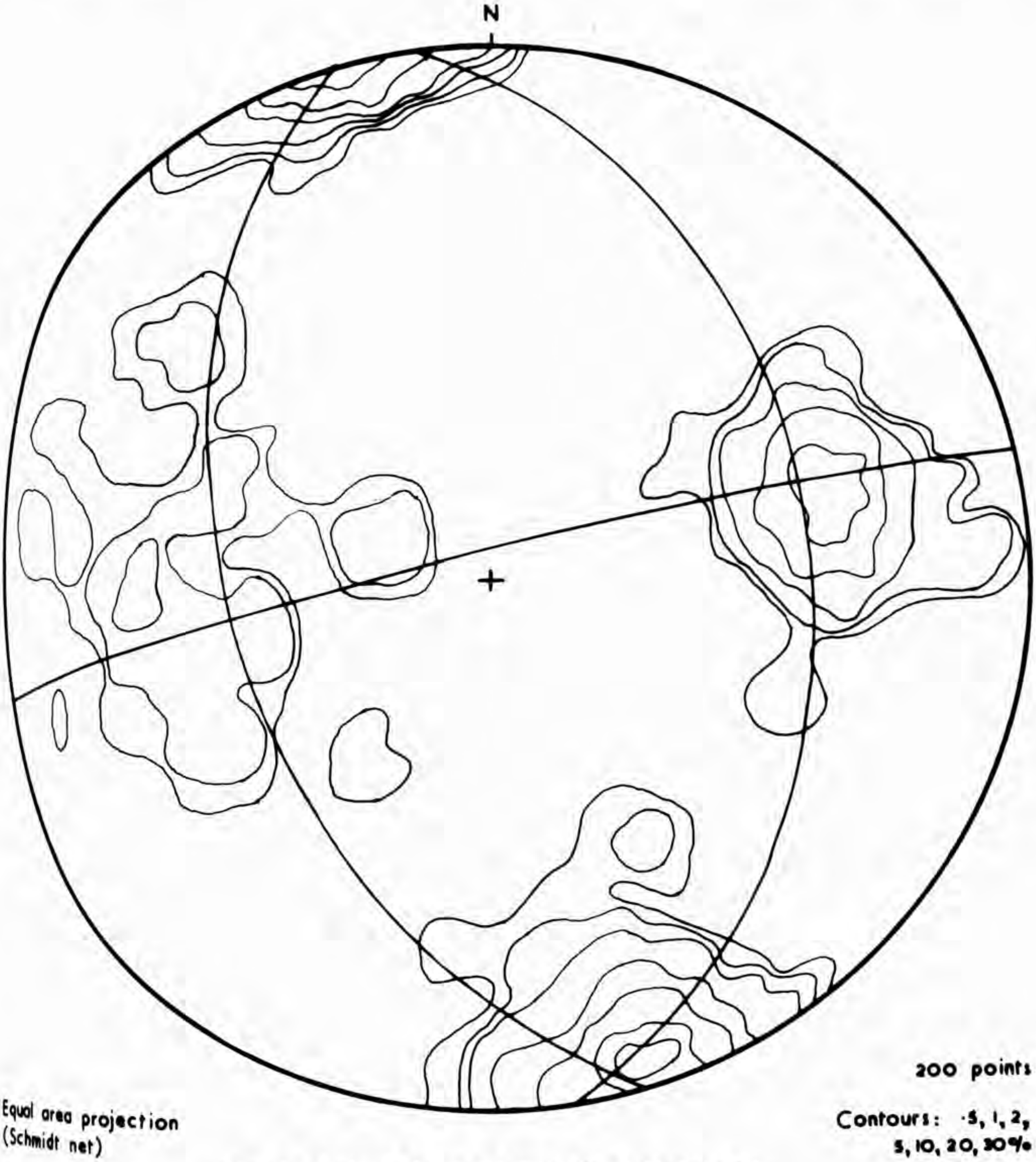
In thin section, it can be seen that there is no relict pyroxene in the rock, and that the amphibole appears to be original. It is greenish-brown and has numerous ore lamellae parallel to cleavage traces in longitudinal sections. It also has fine schiller of ore needles in two directions which intersect at about 80^o, the accute angle being bisected by the lamellae. Amphibole constitutes practically the whole rock, with only a little interstitial plagioclase and some calcite in cavities.

Jointing

The metagabbro is considerably jointed, and the joints fall into two main kinds; an early set and a late set.

The earlier set consists of fracture planes along which there appears to have been no displacement. In this set there are three principal joint-directions which are mutually sub-orthogonal (Fig. 11), and which often give the rock a blocky appearance. This jointing is

STEREOGRAPHIC PLOT OF POLES TO EARLY JOINTS IN HUSFJORD METAGABBRO



The three planes shown are mutually perpendicular

FIG. 11

also responsible for the formation of the crags and pinnacles on some of the mountain tops. The most prominent joint-direction is parallel to the gabbro margin and to the foliation where this is developed. The remaining two joint-directions are subperpendicular to the body walls and to the foliation. Narrow bands of coarse-grained diorite have been emplaced along many of the joints which are parallel to the most prominent direction.

It is not possible to be certain as to the origin of these joints. They could be cooling joints, or they might have been formed during a very late stage of the F_1 folding. In either case they were formed prior to the main F_2 folding, since the early diorite bands are commonly emplaced along these joints, but the diorites themselves have been fractured and sheared during the F_2 movements.

Their mutually suborthogonal nature implies that it is probably more likely that the joints are cooling joints. In this case, the most prominent set would be longitudinal joints, the other two directions being diagonal joints (Balk, 1937).

The later set of joints appears to be associated with late movements of the F_2 folding. They are oblique to the earlier joints and to the diorite bands, but their development is not usually regular enough to give a statistical regional pattern. They could well correspond to F_2 conjugate joints that occur in the metagabbro at Storelv to the north (Sturt and Ramsay, personal communication). They are clearly later than the diorite bands for they cut across the latter, and in places there is evidence that some movement has occurred along these joints, as shown by the offsetting of some of the early diorite bands (Plate 20).

Metasedimentary Rafts and Xenoliths

A fairly common feature of the northern margin of the metagabbro is the occurrence of large rafts and small xenoliths of metasedimentary material. The lithologies of the rafts, which are often as long as a few tens of metres, include metalimestones, mica-schists, and psammites. The xenoliths, which are usually less than a metre in length, are mica-schists, calc-silicate-schists and psammites.

Metalimestone Rafts

The largest and most abundant of the rafts are those of greyish metalimestones. These are generally several tens of metres in length, and often show up well as white scars on the landscape. Sometimes they occur as isolated bodies, but more often they appear as a group or train of rafts composed of perhaps half a dozen individual units. In some cases the individual lenses may join up beneath the surface, the probability of this being shown by the presence of light green grass, a common feature of calcic soils, growing on the ground between the lenses.

The form taken by the rafts seems to be roughly lenticular, the longer axes lying within a plane parallel to the gabbro margins. Gabbro occurs on all sides of the metalimestone, including beneath it (Plate 21), showing that the rafts are truly free-swimming. The raft margins are rounded, and there is often a little brecciation in the contact zone.

Plate 20. Early diorite offset by joint. Husfjord.

Plate 21. Metalimestone raft underlain by metagabbro. N.E. of Kobberfjord.

Plate 22. Bedding in metalimestone raft. N. of Kobberfjordbotn.

Sometimes the contact zone has been intruded by pegmatite or by dioritic sheets belonging to the latest phase of diorite intrusion. Bedding in the metalimestone is usually discernible (Plate 22); frequently it is distinguished by occasional thin calc-silicate or pelitic bands, which represent more siliceous or more argillaceous layers in the original limestone. The general attitude of the bedding is parallel to the plane containing the major axes of the lenticular raft.

Mica-schist Rafts

Rafts of mica-schist are rare, but they may be up to a few tens of metres in length. In the field they resemble patches of sheared metagabbro, but in thin section they are seen to be quartz-biotite-schist with occasional garnets. These schist rafts are elongated parallel to their schistosity and to the margin of the gabbro sheet. In places the schistosity is folded (Plate 23). These foliated rafts have provided weak spots in the metagabbros for they are almost invariably intruded by late coarse diorite sheets. In association with these diorites, there has also been a certain amount of feldspathization of the schist, and this is further discussed in connection with the late diorites.

Psammitic Rafts

On the Ramnes peninsula, the metagabbro makes contact with the psammites and semi-pelites group (see stratigraphic succession, p. 15). At the southern end of Ramnes the metagabbro contains a number of

Plate 23. Folding in mica-schist raft.
Havnefjordfjeld.

xenoliths and a few large elongate rafts of psammites and semi-pelites, the largest of which are several tens of metres in length. The rafts are fairly near to the contact with the country rocks and have probably not moved far if at all from their original positions.

In some places there is sporadic shearing of both the raft material and the metagabbro along the raft borders, particularly at their ends. The lithology of the rafts is principally psammitic, and they are correlated with the psammites and semi-pelites group of the country rocks. Presumably their survival as rafts depended upon their fairly massive, resistant character. Occasional thin basic sheets occur in the psammites, and these have been more or less boudined and deformed.

These psammites and semi-pelites have been migmatized, but not so intensely as those mentioned in the country rocks. It is clear that the migmatization occurred after the emplacement of the gabbro because the gabbro here has also undergone extensive feldspathization and veining by quartzo-feldspathic material. Sometimes it is difficult to distinguish the margin of a raft because it appears to merge into feldspathized metagabbro containing numerous quartzo-feldspathic bands running parallel to the banding in the metasediments of the raft (Plate 24). The introduction of felsic material into the metagabbro appears to be of two kinds; a metasomatic introduction of feldspathic constituents as shown by the patches of feldspar porphyroblasts which have grown in the solid state, and an injection of quartzo-feldspathic liquid as veins and streaks. Both types can be seen in Plate 25. This increase in felsic components only occurs to any great extent in the neighbourhood of the

Plate 24. Raft of psammite and semi-pelite in meta-gabbro. Ramnes.

Plate 25. Feldspar porphyroblasts and quartzo-feldspathic veins in metagabbro.

rafts and xenoliths, and at the contacts with the rafts it is particularly abundant. The gabbro here is not of the foliated facies, so that feldspathic constituents have not been able to penetrate easily and cause widespread feldspathization of the metagabbro. Feldspathization was only possible where there were structural weaknesses in the gabbro, and such weaknesses occur at the margins of the enclosed metasedimentary rafts. The latter are themselves foliated and thus provide an easy passage for the feldspathic constituents. The source material for the quartzo-feldspathic veins in the metagabbro appears to be the psammites of the rafts. Many of the veins join up with the migmatized metasediments and have clearly been injected into the metagabbro as a mobile liquid. This suggests that partial anatexis of the quartzo-feldspathic components of the metasediments may have occurred. The rafts would have acted as traps for heat and volatiles during migmatization due to the more massive nature of the surrounding metagabbro. This concentration of heat and volatiles trapped in the rafts might have facilitated the partial melting of the felsic components of the metasediments, enabling a mobile liquid to form. Evidence of partial anatexis is also provided in the psammites and semi-pelites of the country rocks and this is discussed in detail under migmatization on p.290.

<u>Xenoliths</u>

In the western part of Havnefjordfjeld the marginal zone of the Husfjord metagabbro is very ~~abundant~~ rich in small metasedimentary xenoliths, sometimes only a few centimetres in length. These are of several different lithological types, and in thin section show evidence of being hornfelsed by the gabbro. They include psammites, semi-pelites, calc-silicate-schists, and blocks of pegmatite. The xenoliths have various forms and are usually somewhat angular and sharp-margined (Plate 26). They are randomly orientated, and the various different rock-types are brought into close juxtaposition with one another (Plate 27). Each of the rock-types occurring as xenoliths isrepresented in the country rocks of the envelope, and it is probable that they were derived from this source. It is apparent that the gabbro, whilst it was being intruded, picked up fragments of country rock which had become detached from the walls. This northern contact represents the sole of the gabbro sheet and a similar xenolithic layer also occurs at the base of a gabbro sheet at Hasvik in south-western Sørøy (Sturt, personal communication).

Distribution of Rafts

A study of the distribution of the metasedimentary rafts, in particular the metalimestones which provide a marker horizon, is instructive. It is believed that the trains of metalimestone rafts form a ghost stratigraphy through both the Husfjord metagabbro and the Havnefjord diorite, a large body of fine-grained diorite which has been intruded into the central part of the metagabbro. It can be seen from the geological map that the metalimestone rafts, although rather sporadic, tend to be concentrated along two zones.

The northern zone lies within the metagabbro to the north of the

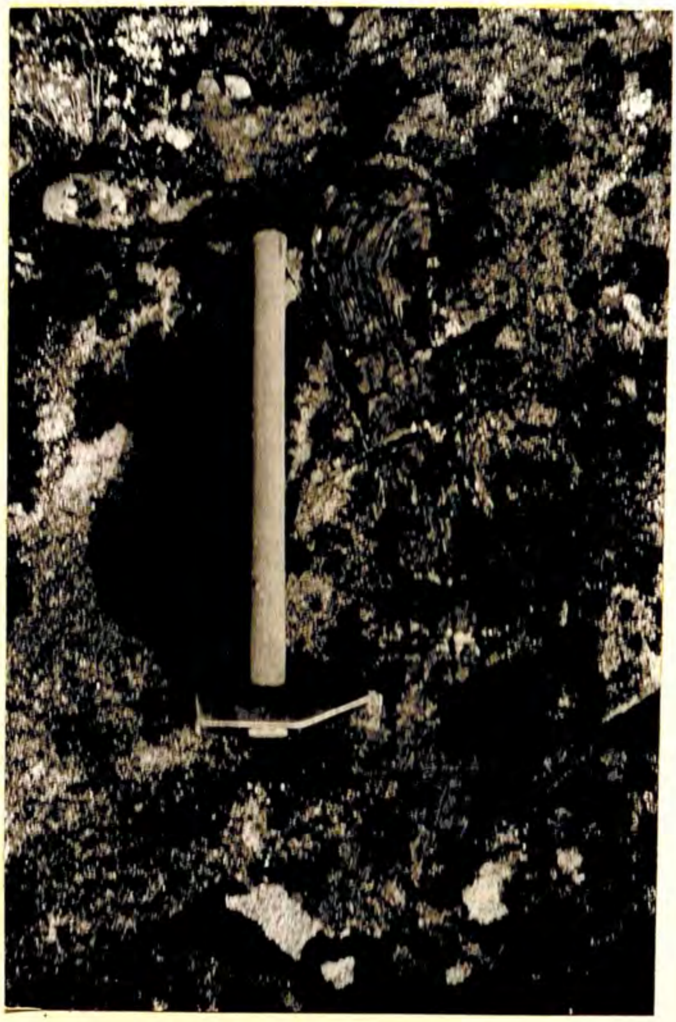


Plate 26. Xenoliths in metagabbro. Havnefjordfjeld.

Plate 27. Xenoliths in metagabbro. Havnefjordfjeld.

Havnefjord diorite, and the line of rafts runs slightly discordantly to the gabbro margin a short way in from the contact. Towards the north-east the rafts become nearer to the contact until above Havnefjordbotn they occur at the contact. To the north-east of this, limestones occur in the country rocks outside the igneous complex. Towards the south-west, the northern raft train occurs further away from the margin, and the number of rafts decreases until in the south-west there are no rafts. The fact that the rafts die out as the distance of the train from the contact increases is presumably due to their removal by the more mobile central part of the incoming gabbro sheet.

The southern train of metalimestone rafts lies mainly within the Havnefjord diorite, but it also occurs in the metagabbro on the western side of the diorite to the north-east of Kobberfjord. This train of rafts runs parallel with the northern one, but is rather sporadic in the diorite. It is believed that this zone may represent the upper limestone horizon mentioned in the stratigraphy section in connection with correlation with the Langstrand area (Roberts, 1965, 1967). Further description of these metalimestones will be given when the Havnefjord diorite is considered, but it is pertinent to mention them here as they yield information concerning the pre-diorite morphology of the metagabbro sheet. It is considered that the region now occupied by the Havnefjord diorite may have previously been a series of large metasedimentary screens in the Husfjord metagabbro (see p. 118). These screens, which included some metalimestones, are thought to have formed an inner subsidiary contact to the gabbro sheet. The metalimestone rafts of the southern train that occur in the gabbro are near to the present diorite contact, and thus were probably near to the original inner metasedimentary screen contact. This could explain the survival of metasedimentary rafts in the metagabbro at such a distance from the metagabbro's outer margin.

The significance of the distribution and attitude of the metasedimentary rafts in relation to the intrusion of the metagabbro will be made more evident when emplacement mechanisms are discussed (p. 118).

Early Diorites and Feldspathization

The finer-grained foliated facies of the Husfjord metagabbro has been particularly susceptible to feldspathization and the introduction of dioritic material.

The diorites are pyroxene-mica-diorites, some containing quartz. They generally occur as narrow bands, which do not exceed about a metre in width (Plate 28), but they also form irregular veins in the metagabbro (Plate 29). In the bands, the minerals tend to have a preferred orientation parallel to the margins, and this is probably a fluxion structure.

The diorite bands are parallel to one another and have been emplaced along the most prominent joint-planes belonging to the orthogonal set. Sometimes small apophyses penetrate side joints, and occasionally the structure of the diorite bands show that they have been emplaced by a dilatational mechanism (Fig. 12A).

Many of the diorite bands have been sheared (Plate 30) and jointed

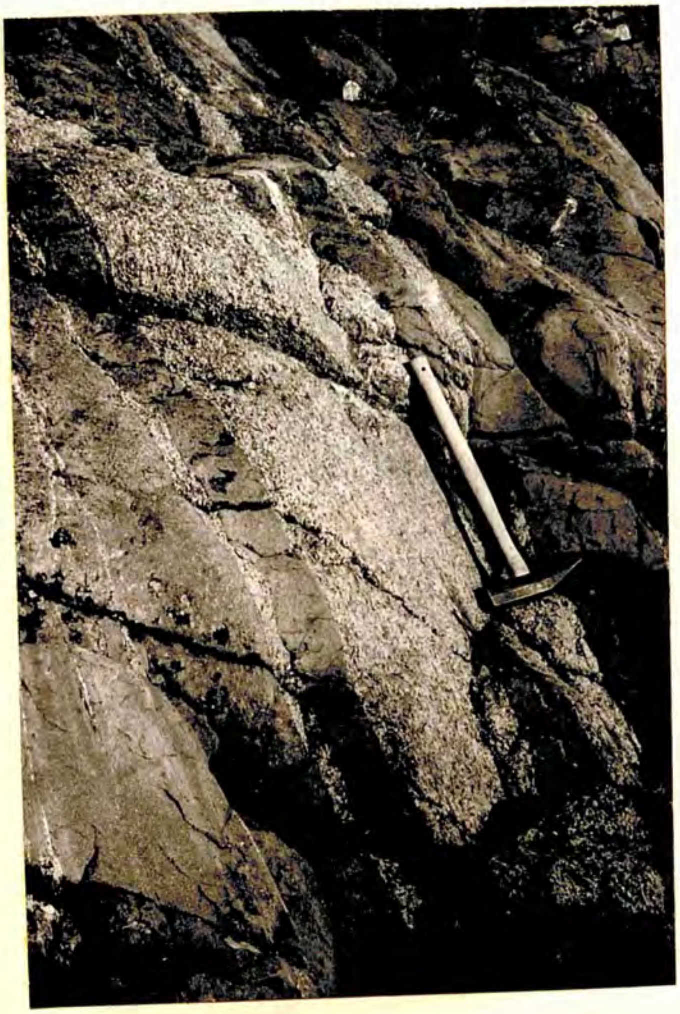


Plate 28. Early diorite bands in metagabbro. Husfjord.

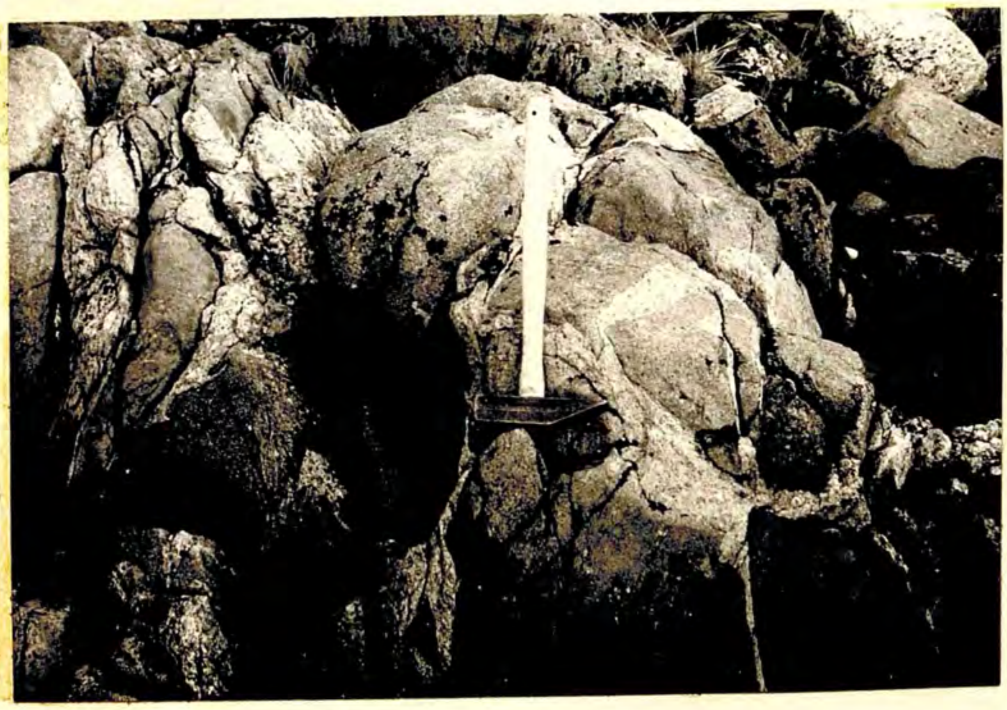
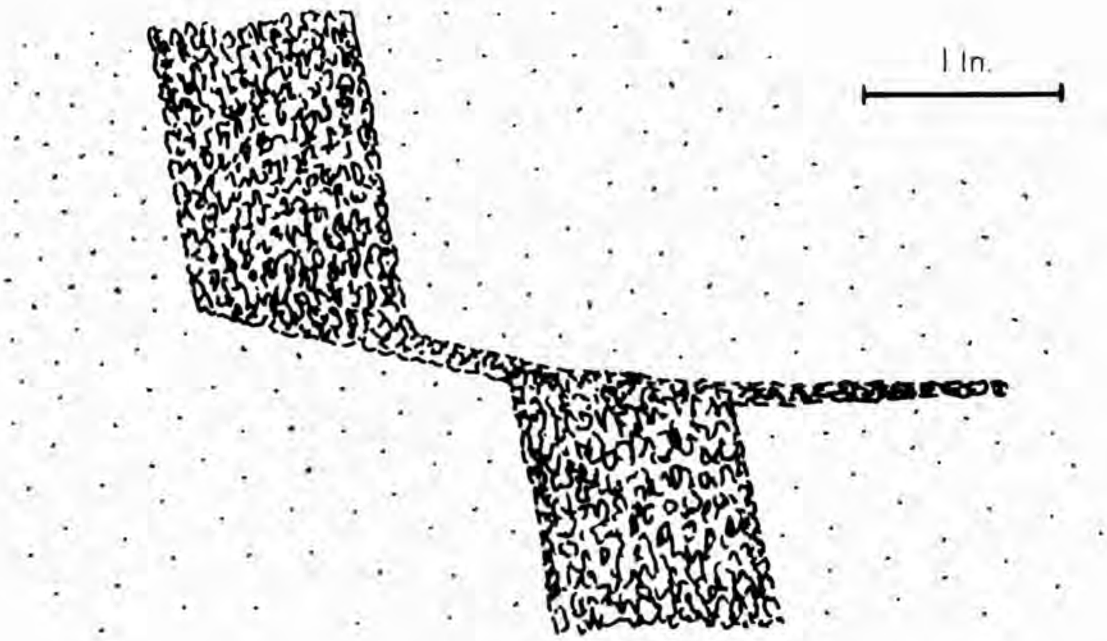
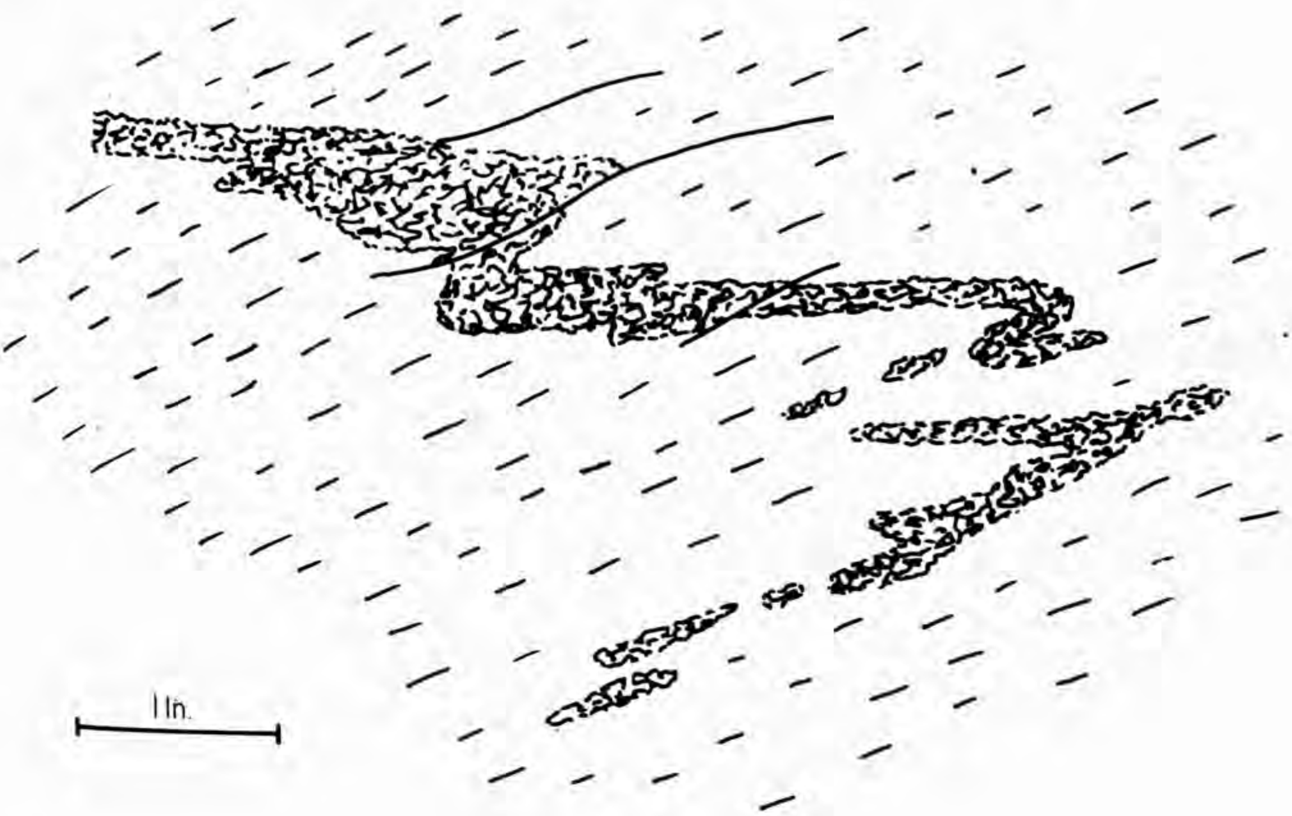


Plate 29. Early diorite veins in metagabbro. Husfjord.



A. DILATATIONAL EMPLACEMENT OF EARLY DIORITE BAND



B. FOLDED AND SHEARED EARLY DIORITE VEIN

FIG. 12

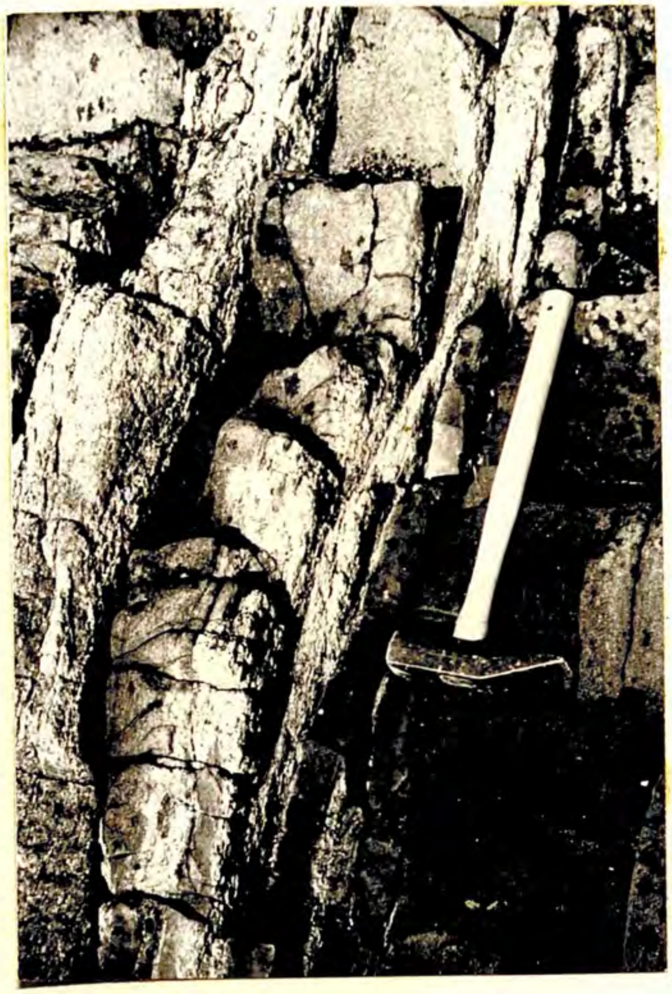


Plate 30. Sheared early diorite bands. Husfjord.

Plate 31. Early diorite bands in feldspathized metagabbro. Husfjord.

during F_2 movements. Some of the narrow irregular veins have been offset by small shears which are orientated parallel to the general shearing direction (Fig. 12B). This plane of shearing is also axial planar to the small folds which form in the veins, and these folds are presumably shear folds formed by differential movement along shear planes (de Sitter, 1956; Hills, 1963).

As described on p. 40, the diorite bands have been folded into a large open arcuate fold by the late F_2 movements (see Fig. 6), reflecting the arcuate form of the metagabbro sheet.

The foliated facies of the Husfjord metagabbro has also been prone to feldspathization, which only occurs in the neighbourhood of the diorite bands and appears to be associated with them (Plate 31). The gabbro contains streaks and bands which are abundant in feldspar porphyroblasts, the feldspars always being elongated parallel to the foliation in the metagabbro. In some cases the porphyroblasts are relatively dispersed and the borders of the feldspathized area are diffuse (Plate 32). In other cases the porphyroblasts are closely spaced and the margins of the bands of feldspathization are sharply defined (Plate 33). There are all gradations between these extremes, and in many bands the spacing of porphyroblasts varies across their widths (seen to some extent in Plate 33). Occasionally the porphyroblasts become so closely spaced that the bands of feldspathization resemble the diorite bands. However, close inspection reveals the presence of small patches of metagabbro between the feldspar porphyroblasts.

The emplacement of the diorite bands post-dated the feldspathization

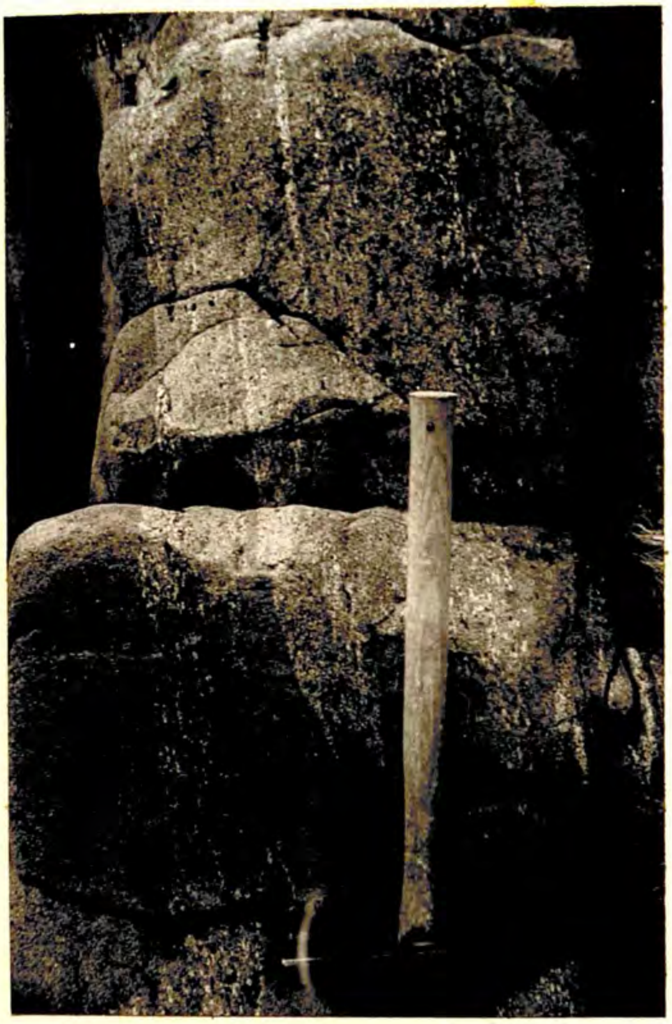


Plate 32. Diffuse felds-
pathization of metagabbro.
Husfjord.

Plate 33. Dense felds-
pathization of metagabbro.
Husfjord.

since they are seen to cut across patches of feldspathization. It is possible that the two events were closely linked in time, the feldspathization being a precursor to the diorite intrusion. Petrographic evidence shows that the diorite intrusion followed the peak of the regional metamorphism, and it is possible that the feldspathization could have been associated with the higher grades of the regional metamorphism.

A similar phenomenon occurs in the Storelv metagabbro, which is a gabbro sheet occurring to the north-west of the Husfjord area (Stumpfl and Sturt, 1965; Sturt and Ramsay, personal communication). Here, the intrusive veins are more irregular in form than in the Husfjord area and are more granitic. Also, extensive diffuse granitization of the metagabbro has taken place, and this is considered by Sturt and Ramsay to be a metamorphic phenomenon.

Late Diorites

Into the central zone of the Husfjord metagabbro has been emplaced a complex of diorites, some of which are fine-grained and some are coarse-grained.

The largest body of diorite is the fine-grained Havnefjord diorite, which has thermally metamorphosed the Husfjord metagabbro. The effects of the contact metamorphism can be observed in the metagabbro for some 300-400 m. from the diorite margin. Certain of the effects of the thermal metamorphism can be observed in the field, though the major part of the evidence results from petrographic studies.

The main observable effect is that the metagabbro becomes more amphibolitic in the neighbourhood of the diorite. Sometimes feldspar porphyroblasts develop, and these very occasionally contain minute inclusions of mafic minerals in their central parts. Some of the early coarse diorite bands in the metagabbro occur within the aureole of the Havnefjord diorite, and these have also been amphibolized and have become crumbly on weathering. Inclusions of both the metagabbro and the early diorites, together with troctolite lenses, occur within the Havnefjord diorite.

There is no evidence in the metagabbro that the Havnefjord diorite caused any marked deformation of its envelope during its emplacement.

Pegmatites

The metagabbro contains numerous acid pegmatites which vary greatly in dimensions, orientation, and grain-size. The smallest are occasional narrow chilled margins, and the largest is a coarse-grained body about 40 m. long and about 3 m. thick. The majority are less than 1 m. in thickness and only a few metres in length and do not usually exhibit chilled margins. They are particularly abundant within the aureole of the Havnefjord diorite, and especially along the contact zone. They are seen to cut across some of the early diorites that have been intruded into the Husfjord metagabbro.

Many of the pegmatites have a runic texture in which graphic intergrowths of quartz and feldspar can be seen on the macroscopic scale, and occasionally quartz crystals are beautifully euhedral.

Normally there is no preferred orientation of the crystals in the pegmatites, but in some cases the feldspars and quartz crystals are aligned parallel to the margins of the pegmatite. This orientation, where it occurs, is not necessarily at the pegmatite margin, but this is the most common site for its development. In these cases where there is a marginal foliation, the gabbro at the contact has developed a schistosity parallel to the margin of the pegmatite and has become augened in the contact zone. In other pegmatites, the marginal zones contain large biotites orientated perpendicular to the contacts.

The grain-size of a particular pegmatite is generally very variable, fine-grained aplitic patches commonly occurring in coarse-grained bodies. If the pegmatite has a schistosity parallel to its margins, then the aplitic patches are elongated parallel to the schistosity. Where no schistosity exists, they are irregular in form.

The metagabbro in the neighbourhood of pegmatites is almost invariably amphibolized, the altered gabbro grading into normal metagabbro away from the pegmatites. The amount of amphibolization varies with the dimensions and grain-size of the pegmatite, being greater next to large or coarse-grained bodies.

This amphibolization is of two kinds. In one kind, where the pegmatite has a sharp contact with the metagabbro, the secondary amphiboles that are developed in the gabbro often have a preferred orientation sub-perpendicular to the contact. This is shown in a hand specimen in which the orientation of the amphiboles has been marked (Plate 34). It would seem that a transfer of heat and chemical

Plate 34. Amphibolization of metagabbro by pegmatite; the pegmatite is white; the black lines are drawn parallel to the orientation of the amphibole crystals. From Husfjord.

constituents, particularly volatiles, from the pegmatite has passed across the contact, and that the secondary amphiboles developing in the metagabbro tended to grow parallel to the thermal and chemical gradient.

The other kind of amphibolization occurs in the majority of cases, where pegmatites, particularly the coarser-grained ones, do not have sharply-defined contacts, but have irregular, ragged margins. Metagabbro is enclosed by pegmatite at the margins, and pegmatitic material lobes into the gabbro. There is no chilling of the pegmatite against its host, and the amphibolization of the metagabbro is often coarse-grained and extensive, the amphibole crystals not having any preferred orientation.

It is generally considered that there are two modes of emplacement of pegmatites and aplites. One is by intrusion, and the other is by metasomatic replacement of the host rock (Ramberg, 1952; Turner and Verhoogen, 1960).

Criteria for distinguishing between dilatational and replacement pegmatites have been described by Ramberg (1952) and King (1948). However, in the Husfjord pegmatites these criteria could not be used, since the absence of banding precludes the study of off-setting relationships.

The only pegmatites that are clearly intrusive are those few micropegmatites which show narrow chilled margins. In the case of the remaining pegmatites, there is no conclusive evidence as to whether they have an intrusive or a metasomatic origin. However, their ill-defined contacts, the absence of chilled margins, the way in which they penetrate into the metagabbro and cause extensive coarse-grained amphibolization all suggest that they are probably of metasomatic origin.

Metasomatic or replacement pegmatites do occur in Sørøy, and these are believed to have been formed by metasomatic replacement of the host rock during regional metamorphism (Roberts, 1965). In these cases the host rock was gneissic, and it was considered that diffusion of felsic constituents from the gneiss into low-pressure zones took place. In the Husfjord pegmatites however, it is clear that such quantities of quartz and alkali feldspar could not have been supplied by diffusion from the metagabbro, and the amphibole-rich patches are not likely to represent complementary basic zones remaining after the removal of felsic constituents, as in examples given by Ramberg (1952). It is more probable that the felsic constituents forming the pegmatites were introduced into the metagabbro from elsewhere, and that the amphibolization is a secondary effect, attendant upon this introduction.

It is necessary to consider a possible origin for this introduced material. The pegmatites post-date the peak of the regional metamorphism. They are especially abundant in the neighbourhood of the Havnefjord diorite, particularly along its northern margin, and may be associated with the diorite's emplacement. The volatiles responsible for the formation of the pegmatites might have been given off at a late stage in the crystallization of the Havnefjord diorite.

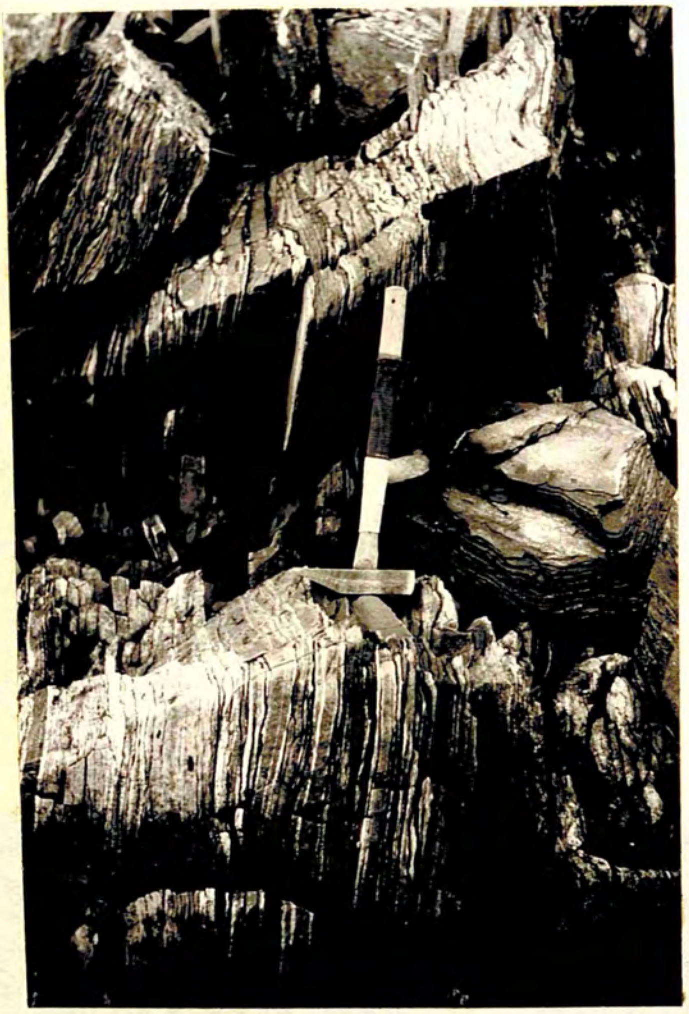


Plate 35. Sheared early diorite and meta-
gabbro. Husfjord.

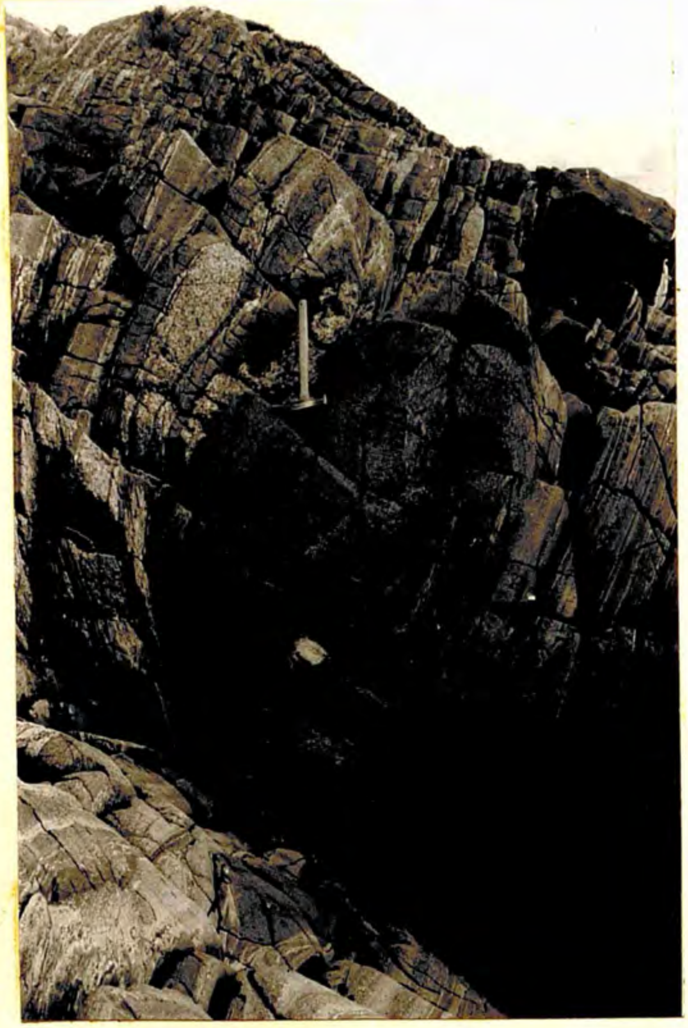


Plate 36. Boudined troctolite in sheared metagabbro. Husfjordness.

Plate 37. Troctolite boudins in sheared metagabbro. Husfjordness.

Shear-belts

There are two zones or belts in the Husfjord metagabbro in which shearing has taken place. The northern shear-belt runs from the east side of Husfjord, across the islands in the fjord, south of Vatnafjeld, and then southwards to the coast at Gåshopen (see geological map). The southern belt is subconcentric with the northern belt, running from Husfjordness in the east to Komagfjord in the south. Each belt is a few hundred metres in width.

In the northern belt only the early diorite bands have been slightly sheared, but in the southern belt the metagabbro as well as the diorite bands have been sheared, and where shearing was most intense a strong foliation was formed (Plate 35). This foliation is parallel to the most prominent joint-direction belonging to the orthogonal set in the metagabbro.

In this latter belt, there are numerous bands, streaks, and lenses of troctolite, particularly well exposed at Husfjordness. The troctolite generally forms lenticular and barrel-shaped boudins, varying from $\frac{1}{2}$ m. to about 10 m. in length (Plates 36 and 37). Sheared metagabbro and sheared diorite bands swing around the boudins, which are also often sheared at their margins. It is clear that the troctolites have been more resistant to deformation, presumably due to their relatively high competence.

The three principal axes of the troctolite boudins are different in length. The shortest axis is perpendicular to the general banding, and the longest is parallel to the strike of the banding. In places there is a strong lineation on some of the joint surfaces parallel to the foliation. This lineation is due to slickensiding, and was presumably produced by relative movement of different parts of the metagabbro sheet during deformation. The lineation plunges steeply to the north, and since the rocks here have an E-W strike, it appears that it is probably an F_2 a-lineation, indicating the direction of transport at least on a local scale.

It is probable that the cause of the shearing was internal failure in the massive metagabbro sheet during folding. It is likely that as folding progressed, that part of the sheet nearest the concave side of the developing fold would have had a compressional strain set up in it, whereas that part on the convex side would have had a tensional strain. These opposing internal strains might have caused internal shearing in the central part of the sheet. Shearing produced in this way, by the release of strain, would have taken place most readily along zones of weakness in the sheet. It is noticeable that the shear-belts are particularly rich in early diorite bands, and it is possible that these provided zones of relative weakness along which shearing could readily have taken place. This is borne out by the fact that the shearing is most intense in the diorite bands.

Perthosites

A number of sheets of perthosite have been emplaced into the Husfjord metagabbro. They are pink or cream coloured, and vary in grain-size from fairly fine to fairly coarse. The size of the sheets is also very variable; the largest ones are tens of metres thick, whereas the smallest are irregular streaks and veins only a few millimetres thick.

The distribution of the perthosites falls into two categories. In one, the sheets have been emplaced into the more intense of the two shear-belts that occur in the Husfjord metagabbro, where their attitude is parallel to the foliation of the shear-belt. In the other category the sheets are orientated radially to this arc and are subvertical. Those of the former group are confined to the shear-belt in the metagabbro, whereas those of the latter group penetrate the Vatna gabbro.

A common feature of the perthosites in the shear-belt is the occurrence of inclusions of sheared Husfjord metagabbro, and also of porphyritic dyke rock. The perthosites of the other group do not contain many inclusions, but are cut by numerous later basic dykes. A full discussion on the perthosites appears on p. 134.

Hydrothermal Veins

Throughout the Husfjord metagabbro there are numerous thin, dark veins which are only a centimetre or less in thickness. They are more resistant than the metagabbro and they always stand out as ribs on weathered surfaces. These veins consist almost entirely of fibrous actinolitic amphibole, sometimes with sphene, and would appear to be the result of hydrothermal alteration along joints, or incipient joints.

The attitudes of the veins are extremely variable and do not form into any statistical pattern. They often occur in swarms in which individual veins may cross or merge with others. They are later than the early diorites and the pegmatites, as they are seen to cut both these.

Occasionally there is evidence that movement had taken place along some of the joints prior to the formation of the veins. Thin early diorite veins have been offset by the movement, but the vein itself is not sheared (Plate 38). It would seem that these veins sometimes formed along some of the late F_2 joints, and probably associated with a post-F_2 phase of regional metamorphism.

Mylonites

The late F_2 deformation is the last tectonic event observed to have taken place in the area, and was of a comparatively brittle nature, and at this time a number of mylonite zones were formed in the Husfjord metagabbro. These mylonites are narrow, and are usually only a few millimetres or centimetres in thickness. In these zones of intense shearing, dynamic metamorphism has advanced far enough to produce a fine-grained, hard, black or greenish mylonite, which has formed from the metagabbro by cataclasis.

The attitude of these mylonites is shown in the stereographic plot of Fig. 7, and it can be seen that the shear planes dip fairly gently to the S.S.W. It was mentioned in the structural section (p. 40) that the strike of these mylonites is parallel to the axis of the late F_2 warp-fold, and that they probably represent shears formed during this phase of defo rmation. Lineation on the shear-planes, apparently due

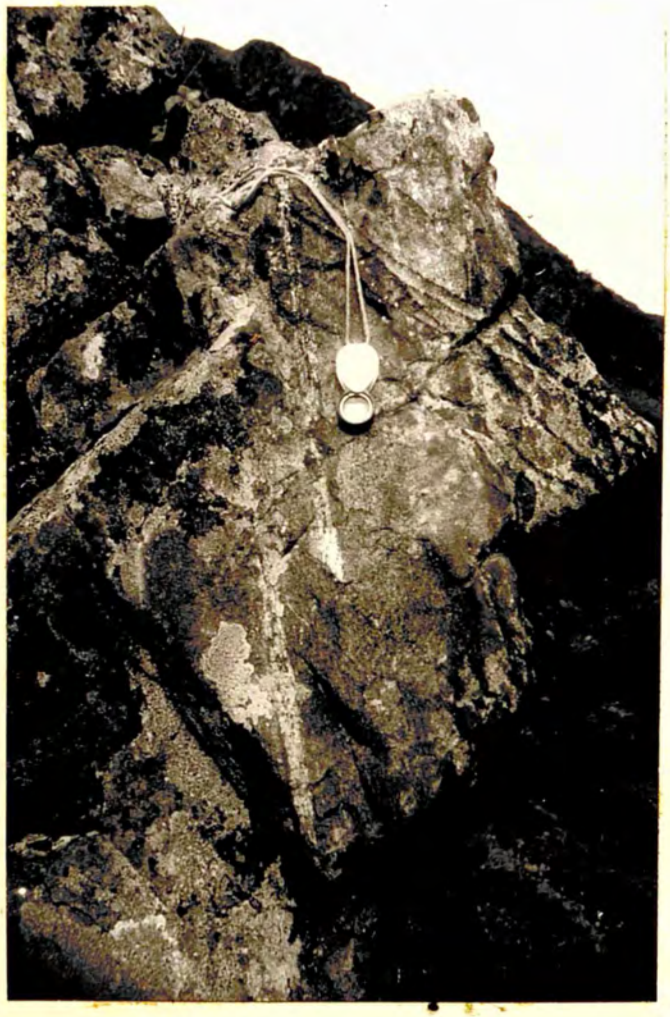


Plate 38. Hydrothermal vein along a joint which has offset an early diorite vein.

to slickensiding, plunges to the south (Fig. 7).

Sone of these mylonite zones are quite complex and comprise several shear-planes which sometimes have slightly different orientations. For example, separate mylonite bands may merge (Plate 39), and the intersection of such bands sometimes causes the formation of another lineation on the shear-planes.

In the thicker shear-zones a gradation can be seen from normal metagabbro into the mylonite. As the mylonite is approached, the metagabbro rapidly becomes progressively augened, with the fo rmation of augen of feldspar and amphibole. Shearing then becomes more intense, and the augen decrease in size until they disappear, and a dark, smooth, banded mylonite is formed.

The mylonites post-date the early diorite bands, as they cut through the latter. Again, progressive augening of the diorite at the margin of the mylonite is sometimes seen.

The mylonites also post-date the pegmatites, as occasionally a micropegmatite vein has been involved in the shearing. In one place, a thin pale pink micropegmatite has been folded into small-scale folds along the upper edge of a mylonite band. The limbs of the folds vary in length from a few millimetres to a few centimetres, and in general they are recumbent, being overturned towards the south. Fold-styles vary; some are chevron-like, some are rounded, and some show a supratenuous form. The overturned limbs are not thinned. In spite of the variation in fold-styles, their axial planes are subparallel, although the trends of their axes within the axial planes are commonly divergent,

Plate 39. Mylonite in Husfjord metagabbro.
Husfjord.

sometimes by as much as 90°. These folds are not shear-folds as described by de Sitter (1956) and Hills (1963), since their axial planes are not parallel to the plane of shearing. The folds must have been formed by the opposing couple which was responsible for the shearing. A microfold lineation is formed on the shear-planes, and this is sub-perpendicular to the slickenside lineation.

It is possible that there might be a complementary shear-zone direction dipping to the north, but which has not developed to the same extent as that dipping to the south.

THE DIORITES

Introduction

An important group of rocks in the igneous complex is a suite of diorites, which are variable in dimensions and in grain-size. The largest diorite body, the Havnefjord diorite, has a maximum width of just under 2 km., and is fine-grained. The ramaining diorites form sheets ranging from a few metres to several tens of metres in width, and most of them are coarse-grained. Some of the diorite sheets have been emplaced into the Havnefjord diorite and thus clearly post-date it. Others do not make contact with the Havnefjord diorite and so there is no field evidence for their relative ages.

The Havnefjord diorite will be described first, as this is the main body, and this will be followed by the remaining diorites.

HAVNEFJORD DIORITE

General

This is a large, fairly homogeneous body of fine-grained pyroxene-mica-diorite. Thin section study shows that it has not undergone the high-grade regional metamorphism that has affected the Husfjord meta-gabbro, and so must post-date the peak of the regional metamorphic event. Furthermore, syn-crystallization deformation of crystals indicate that emplacement was contemporaneous with a phase of folding which must have been of F_2 age.

Its outcrop is arcuate in shape and is about 12 km. in length, running from Strandtind in the south-west to Fella in the north-east. Its maximum width, which is just under 2 km., is in the central part, and it thins out towards either end. It takes the form of a sheet and has been emplaced into the central part of the Husfjord metagabbro. Both its north-western and its south-eastern contacts dip steeply towards a general northerly or north-westerly direction (Plate 40). The contacts are subparallel to those of the Husfjord metagabbro, and it has a subconcordant relationship with the trend of the metagabbro sheet.

It is a fairly resistant rock, forming high ground to the north of Husfjord. It weathers to a light yellowish-grey colour, and surfaces become more rounded than the weathered surfaces of the Husfjord metagabbro. In places it has a coarse jointing, giving it a blocky appearance. The most well-developed directions of jointing are parallel to the margins, and perpendicular to the contacts in a subvertical attitude. There is no evidence of any movement having taken place along these fractures, and they are probably cooling joints, although they may have been formed during the F_2 movements, which apparently had not ceased before the intrusion of the diorite took place.

The diorite is almost invariably non-porphyritic and is uniform in grain-size throughout the body. Sometimes near the margins, particularly the southern margin, there are zones in which the diorite is slightly porphyritic. These zones, whose margins are diffuse, are usually about a metre or so in width, and their trend is parallel to the diorite contacts. In these zones, small feldspar phenocrysts are aligned parallel to the borders of the zones, giving the rock a diffuse

Plate 40. Northern contact of Havnefjord diorite against Husfjord metagabbro; the contact lies along the line of snow at the break of slope, the diorite (on the right) dipping steeply to the left (i.e. northwards).

fluxion structure. These bands are not later intrusions, and most probably represent parts of the diorite sheet in which volatiles were trapped during emplacement.

The emplacement of the diorite has caused the development of a thermal aureole in the Husfjord metagabbro, but has not caused deformation in the host rocks.

Rafts and Xenoliths

A common feature of the Havnefjord diorite is the occurrence within it of numerous inclusions of foreign material. These inclusions vary in size from occasional xenoliths a few centimetres long to frequent large rafts up to several tens or hundreds of metres in length. The xenoliths are of metasedimentary material, but the rafts consist of various rock-types which include metagabbro, troctolite, psammites, semi-pelites, basic schists, and metalimestones. The rafts occur throughout the diorite body, but the different rock-types are restricted to different zones, apparently reflecting a former stratigraphy. The small xenoliths mainly occur near the northern margin of the diorite, and are not obviously related to the rafts.

Metagabbro and Troctolite Rafts

The rafts of metagabbro are not of considerable dimensions, usually being in the order of a few metres or tens of metres in length. Their shape in outcrop is broadly lenticular, with the longest axis oriented parallel to the margins of the diorite. This broad rounded form of the rafts is presumably due to the massive, homogeneous, non-foliated nature of the metagabbro. On the mountain tops they often stand up as knolls due to their resistance to weathering.

These rafts consist of Husfjord metagabbro, generally of the medium-grained non-foliated type, although sometimes they appear quite dioritic. These latter may in fact be hybrids in which the diorite has become contaminated by enclosed metagabbro. In thin section, these hybrids are seen to have a mineralogy and texture similar to that of the diorite, but chemical analysis shows that they are more akin to the metagabbro in chemistry (see Table 12).

Occasionally, the metagabbro rafts contain some of the early diorite bands and zones of feldspathization that occur in the Husfjord metagabbro. These are only found near the south-eastern margin of the diorite body, and in these cases the gabbro is of the finer-grained facies. These rafts have clearly been thermally metamorphosed; the early pyroxene-mica-diorite bands are amphibolized, and the neighbouring metagabbro is often feldspathized, possibly as a result of diffusion of constituents from the early diorite into the metagabbro during metamorphism. From the orientation of the early diorite bands and zones of feldspathization, it is clear that these metagabbro rafts have not been significantly rotated from their original positions; the bands are generally still parallel to the trend of the early diorites occurring outside the Havnefjord diorite.

Many of the metagabbro rafts contain a considerable amount of coarse acidic pegmatite, which penetrates the metagabbro in an irregular fashion. The pegmatites clearly do not follow any jointing in the gabbro. Margins of the pegmatites are generally diffuse and the neighbouring metagabbro often contains feldspar porphyroblasts. The abundance of pegmatite in the rafts supports the suggestion that the formation of the pegmatite might be associated with a pneumatolitic or hydrothermal stage of the emplacement of the diorite. Although pegmatite occasionally occurs in the Havnefjord diorite, no cases were found in which the pegmatite crosses from the diorite into metagabbro rafts.

The distribution of the metagabbro rafts is an important factor in the reconstruction of the pre-diorite morphology of the gabbro, and in the consideration of the probable mechanism of emplacement of the diorite. The metagabbro rafts are especially abundant at the margins of the diorite sheet, but they are not restricted to the marginal zones; they occur sporadically throughout the diorite. In the central part of diorite, metagabbro occurs as disperse trains of rafts lying between and parallel to trains of metasedimentary rafts. This suggests that prior to the emplacement of the diorite, this region was occupied by alternating bands of metagabbro and metasediment. The bearing that this has on the emplacement mechanisms of the major igneous bodies is discussed in detail on p. 118.

It will be recalled that the Husfjord metagabbro contained occasional bands and lenses of troctolite. Two of these troctolite lenses have been preserved as inclusions within the Havnefjord diorite. They form greenish-grey, coarse-grained, knobbly-weathering rafts which are somewhat oval-shaped in outcrop, and they are comparable in size with the larger troctolite lenses in the metagabbro, being a few metres in length. Their greenish colour is due to the abundance of pale green amphibole which has formed during thermal metamorphism by the diorite. There is no sign of any layering that might once have been present in the troctolites, and they only outcrop near to the southern diorite contact.

Psammitic, Semi-pelitic, and Basic Metasedimentary Rafts

Much of the central part of the Havnefjord diorite, especially to the north of Husfjord, is occupied by planar, sheet-like rafts of rusty-weathering semi-pelitic schists and psammites. Some of the rafts are of basic schist material. The smaller of these rafts are a few metres long and a few tens of centimetres wide, but they range in size up to the largest which is about 1 km. in length and about two hundred metres in width. They are elongated parallel to the margins of the Havnefjord diorite, and sometimes the rafts occur as trains which trend in this direction. The attitude of the banding in the rafts is parallel to the raft margins, which in turn are usually subvertical or dipping steeply to the north. The banding is generally fairly planar without much folding, but sometimes irregular minor folds are to be seen. Occasionally F_1 minor folds are refolded by F_2 folds (Plate 41, in which the psammite is a boulder dislodged from a psammite raft). Thus the diorite emplacement must post-date at least the beginning of the F_2 movements.

The margins of the rafts, where seen are fairly sharp, and there appears to be no significant assimilation by the diorite. The diorite at the contact is sometimes a little finer-grained than the usual, and

Plate 41. Isoclinal F_1 folds refolded by F_2 folds. From psammitic inclusion in diorite. Havnefjordfjeld.

probably indicates slight chilling of a diorite magma against the metasedimentary rafts.

Virtually all the exposures of the rafts are such that only subhorizontal sections of the rafts are obtained, and where there are vertical sections, they are inaccessible. Thus it is not possible to be sure of the vertical extent of the rafts, or to say whether the rafts connect downwards (or upwards) to form continuous screens in the diorite, or whether they are completely free-swimming. Thus if the latter is the case, it is not known in which attitude the longest axis of the rafts lie, and so the exact direction of emplacement of the diorite is uncertain.

From the spatial arrangement of rafts having similar composition it would appear that they probably have not been moved far, if at all, from their original pre-diorite positions, since a relict stratigraphy can be traced through the body. The large dimensions of many of the rafts also suggest that they have probably not changed their positions very much. These psammitic and semi-pelitic rafts are very abundant in the central zone of the diorite, and here, occasional metagabbro rafts occur between and parallel to the metasedimentary raft trains. Towards the south, the metasedimentary rafts decrease in size and number, and the number of metagabbro rafts increases, until at the southern contact, only metagabbro rafts occur.

In one of the large rafts on the west side of Havnefjord, the hornfelsed metasediments have become mobilized by the diorite. The diorite has streaked into the metasediments and broken them up into angular fragments and blocks (Plate 42). The fragments have clearly been rotated and mixed up by the diorite magma. However, each block has sharp margins and there is no significant assimilation of the metasediments by the diorite, although the blocks almost invariably have a reaction rim around their margins. The inner part of this rim is of plagioclase and the outer part, which is not always present, is of hypersthene.

The diorite in the neighbourhood of these mobilized hornfelses has been net-veined by quartz-plagioclase material (Plate 43). The veins vary somewhat in thickness, but are usually a few centimetres thick, and they penetrate the diorite in an irregular pattern, often with angular branchings. Some of the hornfels blocks are also veined.

Petrographic evidence (p. 194) suggests that the felsic components of the hornfelses had become slightly mobile during hornfelsing, and it is possible that this mobilized fraction provided the material for the veins in the neighbouring rock. Von Platen (1965) maintains that anatexis of the salic components of metasediments can be caused by heat alone, without the addition of any external fluids (see discussion on migmatization, p. 291). Thus it is feasible for the net-veining to have resulted from the injection of mobilized material from the hornfelses.

Metalimestone Rafts

The northern limit of the rusty-weathering semi-pelitic rafts is fairly clearly defined, and to the north of this the Havnefjord diorite contains occasional rafts of metalimestone. Most of these rafts are

Plate 42. Mobilized hornfelses in a raft in diorite. Havnefjord.

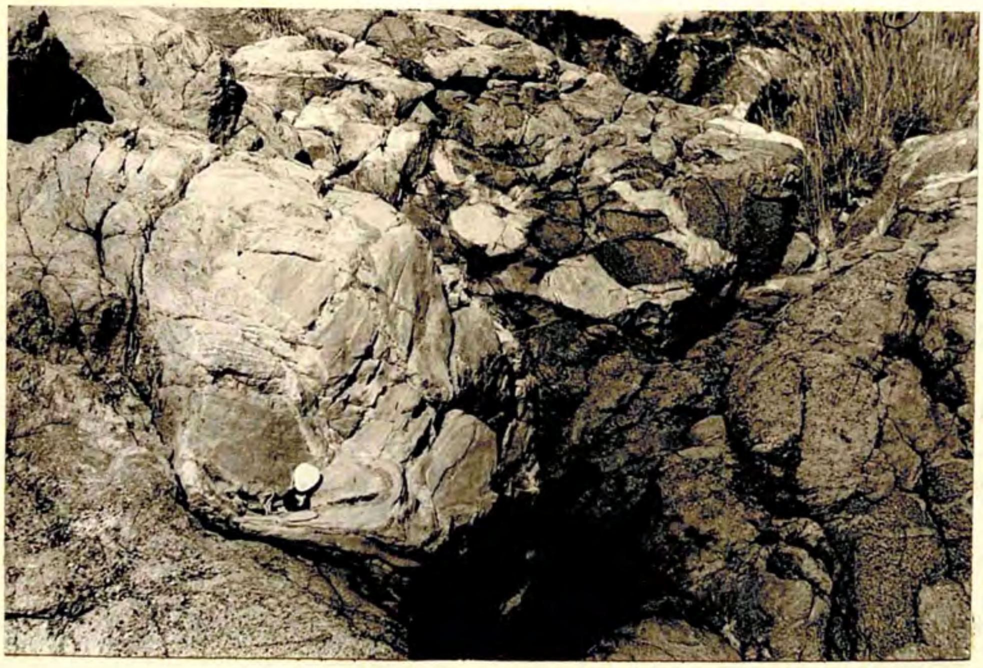


Plate 43. Net-veining in diorite and mobilized hornfels. Havnefjord.

large, often being many tens of metres in length (Plate 44). They are elongated parallel to the trend of the psammitic and semi-pelitic rafts and to the diorite margin.

At least some of these rafts appear to be detached from any country rocks. Diorite occurs above the raft shown in Plate 44, and two rafts outcropping in the cliffs to the west of Husfjord are seen to peter out downwards. In these rafts which seem to be freely suspended in the diorite, the banding in the metalimestone is vertical (see Plate 44). In some of the metalimestone rafts near the northern contact of the diorite at Havnefjordbotn, the banding, which is marked by thin calc-silicate bands, dips towards the north, but in these cases it is not possible to ascertain whether the rafts are detached.

In the metalimestone rafts at Havnefjordbotn, minor folding has been preserved, particularly in the thin calc-silicate bands. In places the competent calc-silicate bands are highly contorted by folds which have discordant axial trends (Plate 45). Some of these folds are tightly isoclinal, and are probably F_1 folds. The limestone itself, due to its low competence, has suffered plastic deformation and has readily recrystallized. Thus the forms of the folds can only be discerned where the limestone contains thin dark pelitic bands, which have preserved the fold shapes (Plates 46 and 47).

In places, it can be seen that the limestone became quite mobile during deformation. It has intruded into cracks in the calc-silicate bands, and has sometimes broken through them, breaking the more competent bands up into fragments, around which the limestone has flowed. There

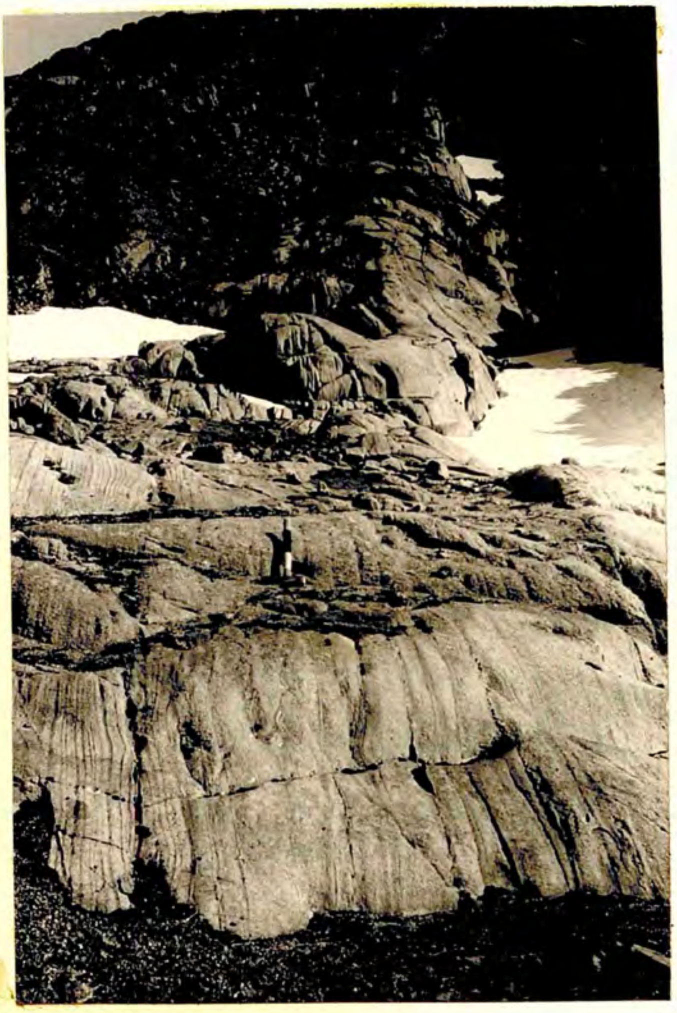


Plate 44. Metalimestone raft in diorite. W. Havnefjordfjeld.

Plate 45. Folded calc-silicate bands in metalimestone raft. Havnefjordbotn.

Plate 46. Thin pelitic band in metalimestone raft. Havnefjordbotn.

Plate 47. Supratenuous folding of calc-silicate bands and pelitic bands in metalimestone raft. Havnefjordbotn.

are many instances of fold closures in the calc-silicate bands which have become completely isolated from their limbs (Plate 48).

Although the metalimestone rafts occur rather sporadically in the Havnefjord diorite, in general they are confined to a zone near the northern margin of the diorite. This arrangement suggests that a broad band of metalimestone may have existed in the zone now occupied by the metalimestone rafts, prior to the diorite's emplacement. Metalimestone rafts occur both on high ground and in the floor of a deep valley running along this zone, indicating that the occurrence of these rafts is a planar feature and not a linear one.

Metasedimentary Xenoliths

The small xenoliths have only occasionally been seen, and appear to occur mainly in the northern part of the diorite body. They do not have any preferred orientation, but have various attitudes in the host rock. They are only a few centimetres in length, and consist mainly of psammitic and semi-pelitic metasediments, which have been hornfelsed by the surrounding diorite. Almost invariably they are being assimilated by the diorite, and their margins are generally diffuse (Plate 49). Sometimes only dark schlieren remain to mark the site of an assimilated xenolith. In Plate 49 the psammitic band of the xenolith has been folded into a tight isoclinal F_1 fold whose axial plane is gently folded, presumably by F_2.

It is inconceivable that such small, highly digested xenoliths could be near their original pre-diorite positions. They are unlikely

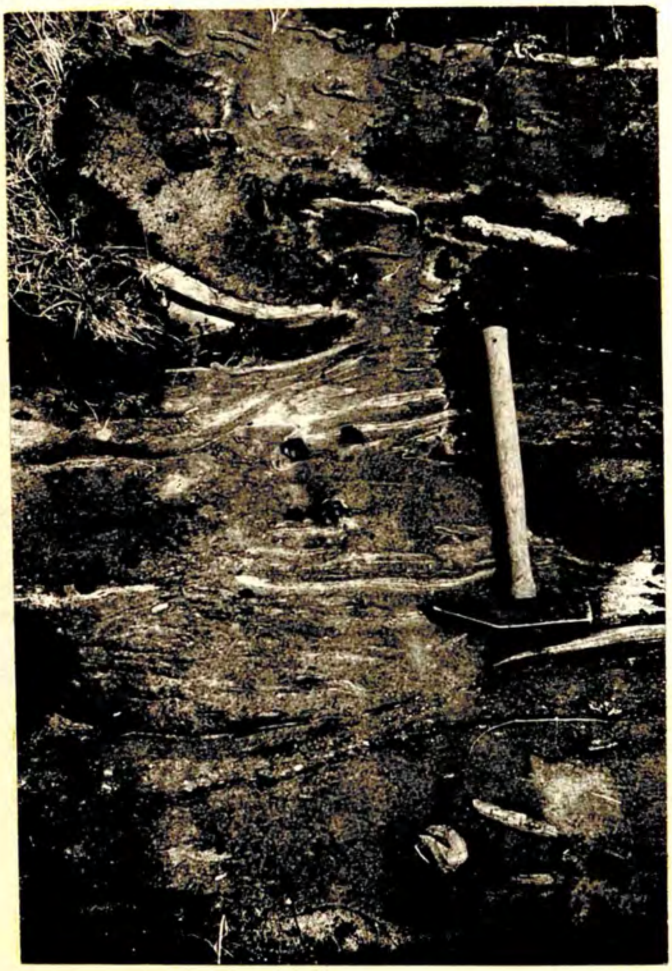


Plate 48. Calc-silicate bands disrupted by mobilized limestone. Havnefjordbotn.

Plate 49. Metasedimentary xenoliths becoming assimilated by diorite. Note isoclinal F_1 fold-closure. Havnefjordbotn.

to represent rock derived locally from the walls of the body, and have probably been brought up by the diorite from depth. The degree to which they are assimilated bears witness to this.

LATE DIORITES

Introduction

The Havnefjord diorite, and in some places the Husfjord metagabbro, have been intruded by a number of later coarse-grained diorite sheets. These are generally several metres or tens of metres in width, and may be up to a few hundred metres in length. They are subvertical, and their trend is parallel to the margins of the Havnefjord diorite and to the trend of the metasedimentary rafts in the diorite. This direction is also parallel to that of the most prominent jointing in the Havnefjord diorite. It is significant that they are commonly closely associated in the field with the metasedimentary rafts.

These late diorites are of three kinds; a pyroxene-mica-diorite, a garnet-rich quartz-diorite, and a garnet-poor quartz-diorite. This latter is not emplaced into the Havnefjord diorite, so there is no direct field evidence that it is later than the Havnefjord diorite. However, because of its close similarity to the other late diorites, it is considered in this section.

Pyroxene-mica-diorites

In general field appearance and in mineralogy, these resemble a coarse-grained version of the Havnefjord diorite, although in thin section they are seen to contain a greater proportion of potash feldspar than the latter. They weather to a yellowish colour, and because of their resistance generally stand out as rocky crags, and on the mountain tops form tor-like knolls. The feldspar phenocrysts are often as large as 2 cm. in length and are usually subhedral in form. Almost invariably they are randomly-orientated, but occasionally there is a preferred orientation parallel to the margins of the diorite. Very occasionally potash feldspar phenocrysts are ovoid, and have a rim of sodic plagioclase around their margins.

These diorites form parallel-side~~s~~, sheet-like bodies emplaced d/ into the Havnefjord diorite, and a few have also been emplaced into the Husfjord metagabbro near its contact with the Havnefjord diorite. They vary in width from a few metres to many tens of metres, and the longest one has been traced for about $1\frac{1}{2}$ km. . However, it is likely that many of the smaller ones link up with longitudinally-adjacent ones beneath the present erosion surface.

Their trend is parallel to the margins of the Havnefjord diorite, and on the regional scale they are deformed into a large arc by the late warp-fold. On the whole, individual sheets maintain a fairly constant thickness when traced horizontally or vertically, although some of the larger ones are seen to vary slightly in an irregular way. The sheets dip steeply towards the north (Plate 50), and this attitude

Plate 50. Late diorite (left) contact with Havnefjord diorite; contact dips steeply to the north. Havnefjordfjeld.

is parallel to the most prominent direction of jointing in the Havnefjord diorite. It is likely that the jointing in the fine diorite was an important factor in the emplacement of the later sheets.

The late diorites themselves often show a prominent jointing, the joints being subvertical in a direction perpendicular to the walls of the sheets. These joints are probably primary cooling joints, although they might be cross joints associated with the deformation which caused the large warp-fold.

The margins of the late diorites against their host are always sharp, but there is no chilling of the coarse diorite (Plate 51 and 52). Thin section study shows that the Havnefjord diorite is hornfelsed by the coarse diorite at the contact. Where the coarse diorite makes contact with the Havnefjord diorite, it often contains hornfelsed blocks of the fine diorite at its margin (Plate 53). These blocks are generally somewhat rectangular, orientated parallel to the diorite margins, and their shape is probably governed by the jointing in the fine diorite. Their margins are sharp and they are not becoming assimilated by the coarse diorite. They have probably been derived locally from the neighbouring Havnefjord diorite.

The coarse diorites are often closely associated in the field with the metasedimentary rafts in the Havnefjord diorite. Where this association occurs, it is always with the rusty-weathering psammitic and semi-pelitic hornfelses, and in many cases the diorites have been emplaced alongside the metasedimentary rafts, particularly the larger rafts. In these cases, the late diorites are not chilled against the

Plate 51. Contact between late diorite (left) and Havnefjord diorite. Havnefjordfjeld.

Plate 52. Late diorite contact with Havnefjord diorite. Havnefjordfjeld.

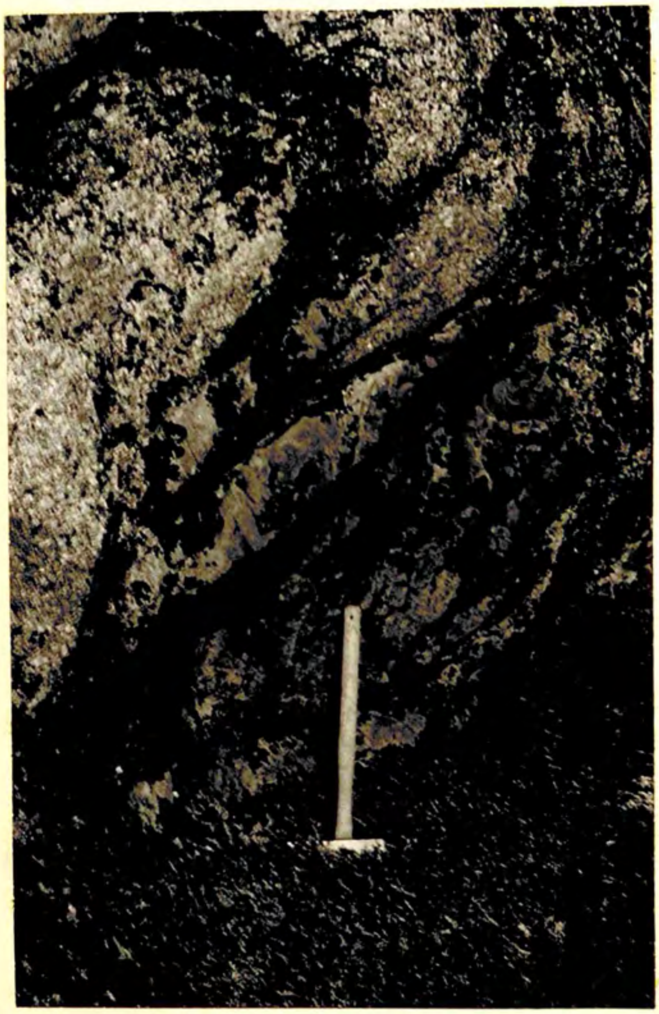


Plate 53. Blocks of fine Havnefjord diorite
in coarse late diorite. Havnefjordfjeld.

metasediment, and they frequently contain blocks of metasediment at their contact with the raft. There are numerous examples of coarse diorite sheets containing blocks of Havnefjord diorite at one margin where it makes contact with that diorite, and undigested blocks of metasediment at the other margin where contact with the raft is made.

In other cases, the late coarse diorite sheets have completely enveloped the rafts, and large enclaves of the rusty-weathering hornfelses occur within the coarse diorite (Plate 54). These enclaves have sharp margins, and this together with the fact that they are of similar lithologies to the metasedimentary rafts in the Havnefjord diorite, suggests that they have probably been derived from nearby rafts.

In one case, one of the coarse late diorites penetrating the Husfjord metagabbro, within the aureole of the Havnefjord diorite, contains a long enclave of the metagabbro (Plate 55). Attached to the sides of this metagabbro enclave there are fragments of an early diorite band. Although this early diorite is also fairly coarse-grained, it can be distinguished from the enclosing coarse diorite; it is amphibolitic, and its feldspars have a preferred orientation whereas those in the surrounding diorite are randomly oriented. There is also an example of a late diorite within the aureole of Havnefjord diorite cutting across one of the early diorite sheets. These examples provide further evidence for the early age of the diorite bands in the Husfjord metagabbro.

Not all the late diorites show a close relationship to the metasedimentary rafts. Many do not make contact with any rafts at the

Plate 54. Hornfels enclave in late diorite. Havnefjordfjeld.

Plate 55. Block of Husfjord metagabbro in late diorite. Above Husfjordbotn.

surface, but it is possible that they might do so at lower levels.

A common feature of all these coarse pyroxene-mica-diorites, whether they are associated with the large metasedimentary rafts or not, is the occurrence in them of small xenoliths. These xenoliths are present throughout the diorite sheets, and are not confined to the margins. They range in size from a few tens of centimetres down to a centimetre or less in length (Plate 56 and 57). Almost invariably they are of metasedimentary material, but occasionally there are xenoliths of dark, fine-grained, basic material resembling Husfjord metagabbro. The orientation of the xenoliths is only approximately parallel to the margins of the diorite sheets; although there is a general tendency for them to have a preferred orientation, there is a slight scatter in their attitudes. It is clear that these xenoliths have been rotated and mixed up.

The basic inclusions and the psammitic varieties of the metasedimentary xenoliths usually have fairly well-defined margins, but the pelitic and semi-pelitic inclusions usually have diffuse margins, and are clearly being assimilated by the diorite. Often, vague schlieren or diffuse dark spots mark the sites of completely digested xenoliths. Occasionally, semi-pelitic bands in the larger xenoliths can be traced laterally from normal metasediment, through stages of progressive assimilation and dioritization, until it becomes indistinguishable from the diorite. These xenoliths must have been immersed in the diorite magma for a considerable time, and were probably brought up with the magma from deeper levels of high thermal and chemical activity.

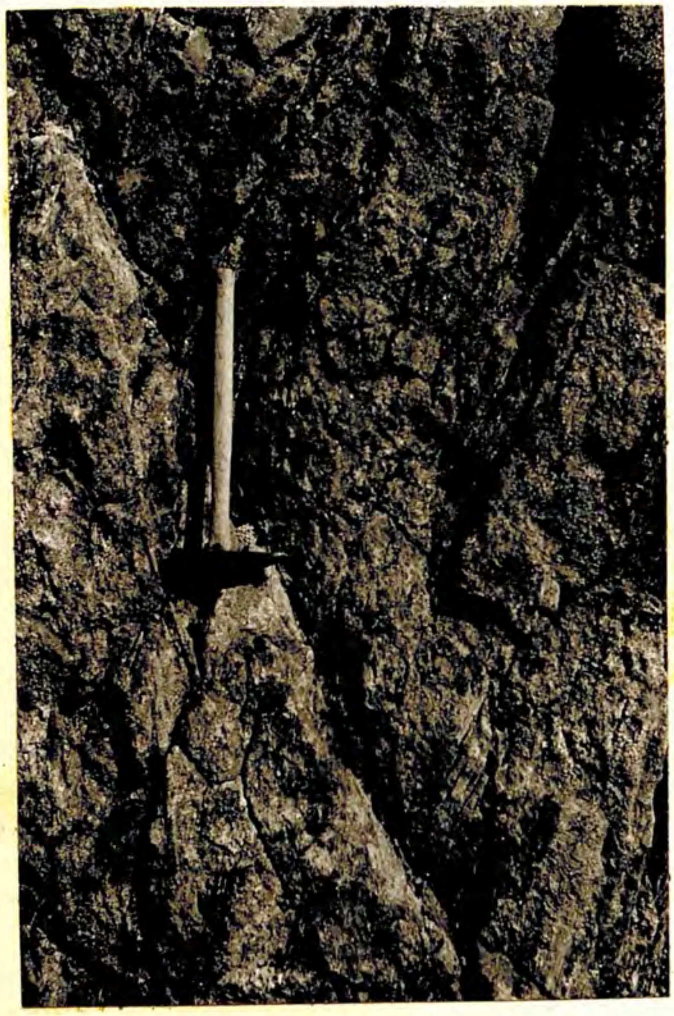


Plate 56. Xenoliths in late diorite. Havnefjordfjeld.

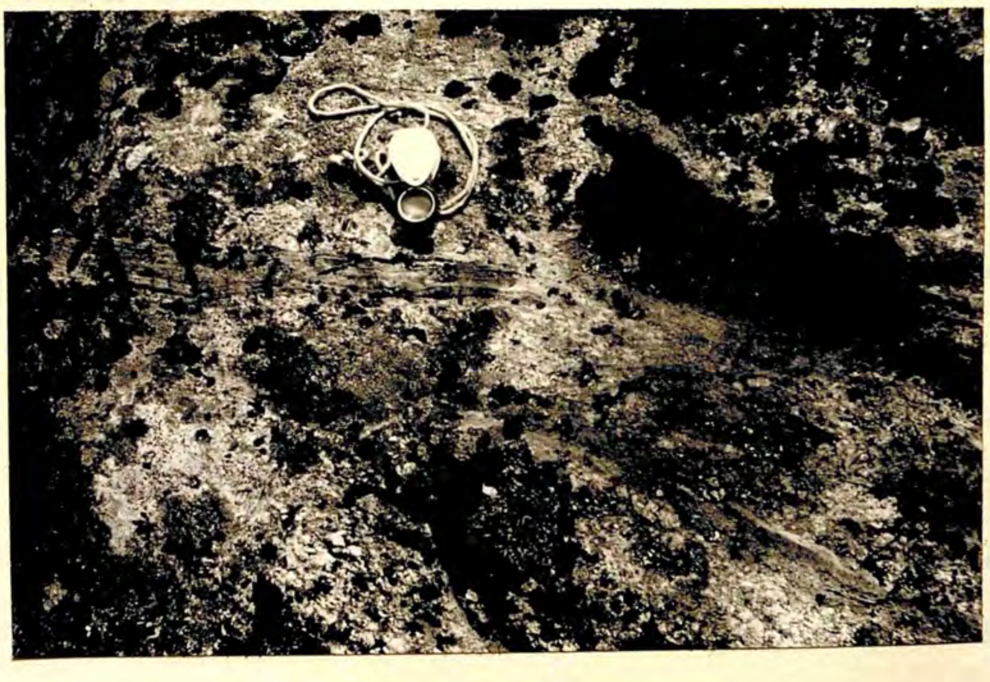


Plate 57. Xenoliths in late diorite. Havnefjordfjeld.

Quartz-diorites

These are coarse-grained, quartz-bearing biotite-diorites in which subhedral potash feldspar phenocrysts range up to about 2 cm. in length. Usually the phenocrysts are randomly orientated, but occasionally there is a crudely-developed preferred orientation parallel to the margins of the body. The feldspars weather to a fairly clean white colour so that the rock has a distinctive white spotted appearance in the field.

The quartz-diorites are of two kinds; one is fairly abundant in small reddish garnets, whereas the other only contains occasional garnets. The former will be distinguished as garnet-rich quartz-diorites, and the latter as garnet-poor quartz-diorites.

Garnet-rich Quartz-diorites

These diorites form sub-vertical sheets which have been emplaced into the Havnefjord diorite. They are not very common, and in general are not as large as most of the coarse pyroxene-mica-diorite sheets, the largest being about 400 m. long and about 50 m. wide. Their trends are usually parallel to those of the coarse pyroxene-mica-diorite sheets, but in one case the trend is discordant with this direction.

It may be significant that they only occur in a zone where the metalimestone rafts are more numerous, to the north of the main pyroxene-mica-diorite belt. They are invariably associated in the field with metasedimentary rafts (Plate 58), often of metalimestone or calc-silicate hornfelses. Rarely, they are associated with the rusty-weathering

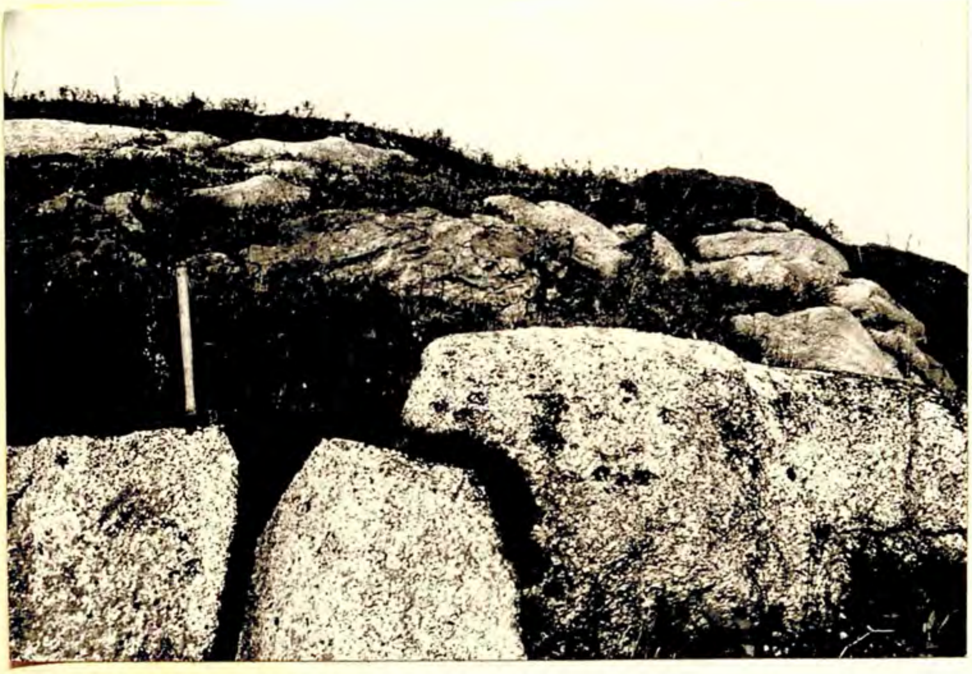


Plate 58. Garnet-rich quartz-diorite associated with metasedimentary raft. Havnefjordbotn.

Plate 59. Metalimestone and calc-silicate block at the margin of garnet-rich quartz-diorite sheet. Havnefjordbotn.

hornfelses, although it is possible that metalimestones may occur nearby beneath the surface.

These diorites have sharply-defined margins, in which there is no marked change in grain-sizes towards the contact, and they are comparatively free from inclusions. They contain occasional blocks of metasediment, but there are none of the small digested xenoliths so characteristic of the pyroxene-mica-diorites. The metasedimentary blocks occur at the margins of the sheet and are often somewhat deformed. They consist of metalimestone and calc-silicate rock (Plate 59).

Garnet-poor Quartz-diorites

These quartz-diorites are also coarse-grained and resemble the more garnetiferous quartz-diorites in the field, particularly as they have a white or light grey spotted appearance. However, the former contain a little more mafic material, and close inspection of the latter reveals the presence of numerous small garnets. Maximum size for the feldspar phenocrysts is about 2½ cm. in length, and there is generally a tendency towards a preferred orientation parallel to the diorite body walls.

These diorites form subvertical sheets, and have been emplaced into the Husfjord metagabbro to the north of the Havnefjord diorite, and also into the neighbouring country rocks. They do not make contact with the Havnefjord diorite, the coarse pyroxene-mica-diorites, or the garnet-rich quartz-diorites, and thus their age relationships with these rocks cannot be established in the field.

The contacts of these diorite sheets are sharply defined, but there is no gradation in the grain-size towards the margins (Plate 60). The diorites often contain blocks of the host rock, which usually have sharp margins, and which are sometimes up to a few tens of centimetres in length. In the case of the metasediments, the xenoliths are generally angular and somewhat rectangular (Plate 61). In some cases, at the margins of the diorites, it is clear that the diorite has penetrated along the foliation planes of the metasediments. In other instances, the blocks have become disorientated and are clearly free-swimming in the diorite. The xenoliths are mainly of calcareous psammites, and sometimes minor folds, apparently of F_2 age, are preserved in them.

In the neighbourhood of these diorites, the Husfjord metagabbro or the metasediments have often been extensively feldspathized, the metasediments of the country rocks being much more prone to this than the metagabbro. Feldspar porphyroblasts have grown in the rocks, apparently in the solid state. They are anhedral, randomly orientated, and grow up to about 2 cm. in length. In some places the feldspars are relatively sparse, whereas in others they are very dense, and there are all gradations between these extremes (Plate 62). The dense parts sometimes follow the planes of the foliation, suggesting that some bands were more susceptible to feldspathization than others. Sometimes the dense feldspathization forms irregular, diffuse patches whose forms are independent of the banding, which it more or less obliterates.

The feldspathization is occasionally so dense that the rock begins to resemble the diorite, but in these cases, there is no indication of

Plate 60. Garnet-poor quartz-diorite contact with metasediments. Havnefjordbotn.

Plate 61. Metasedimentary blocks in garnet-poor quartz-diorite. Above Havnefjordbotn.

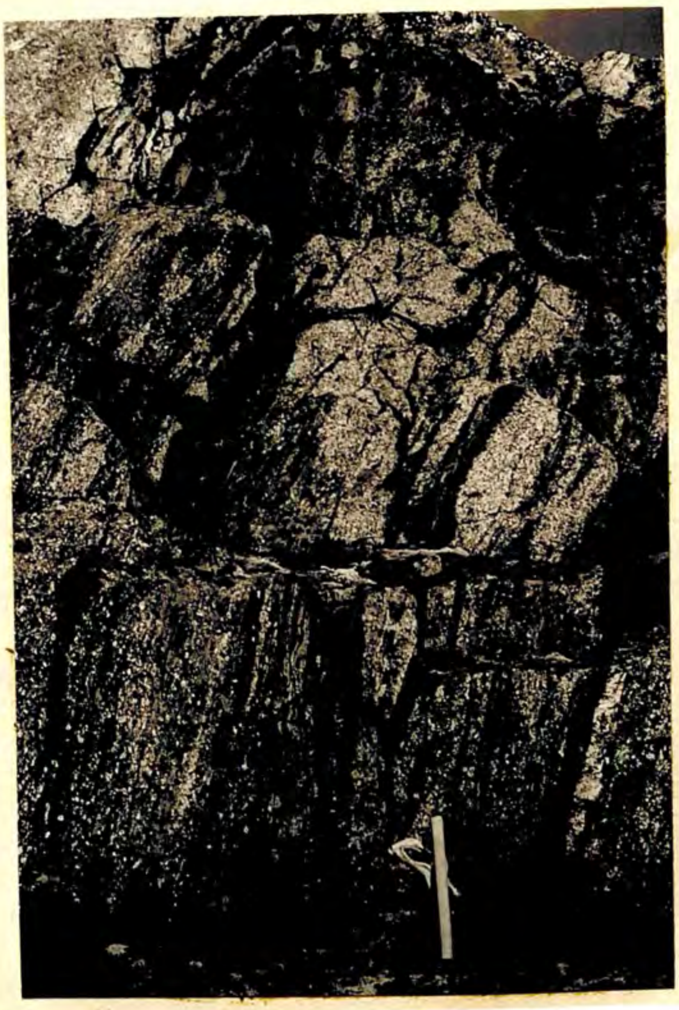


Plate 62. Feldspathization of country rocks adjacent to garnet-poor quartz-diorite. Above Havnefjordbotn.

an intrusive margin. It is possible that the relationship between the diorites and this feldspathization is comparable to that between the early Husfjord diorites and the feldspathization associated with them. In the case of the Husfjord diorites the feldspathization, which preceded the diorites themselves, sometimes became so dense that they were almost indistinguishable from them. In that case it was considered that the feldspathization may have been a precursor to the introduction of the diorite, and a similar mechanism may have been in operation in the present case.

In places, rounded blocks of practically unfeldspathized metasediment occur within what appears to be feldspathized metagabbro (Plates 63 and 64). For much of their boundaries, they have sharp contacts against the surrounding rock, but occasionally they merge completely with the enclosing rock. They are metasedimentary xenoliths in the metagabbro which have largely resisted the feldspathization.

There are several instances near the northern margin of the Husfjord metagabbro where the metagabbro has undergone this feldspathization. It occurs as zones in the neighbourhood of metasedimentary rafts, and usually the rafts themselves are affected as well. Clearly these parts of the metagabbro which contain rafts were points of weakness which were more prone to the feldspathization. Generally, but not invariably, garnet-poor quartz-diorites occur at these places too.

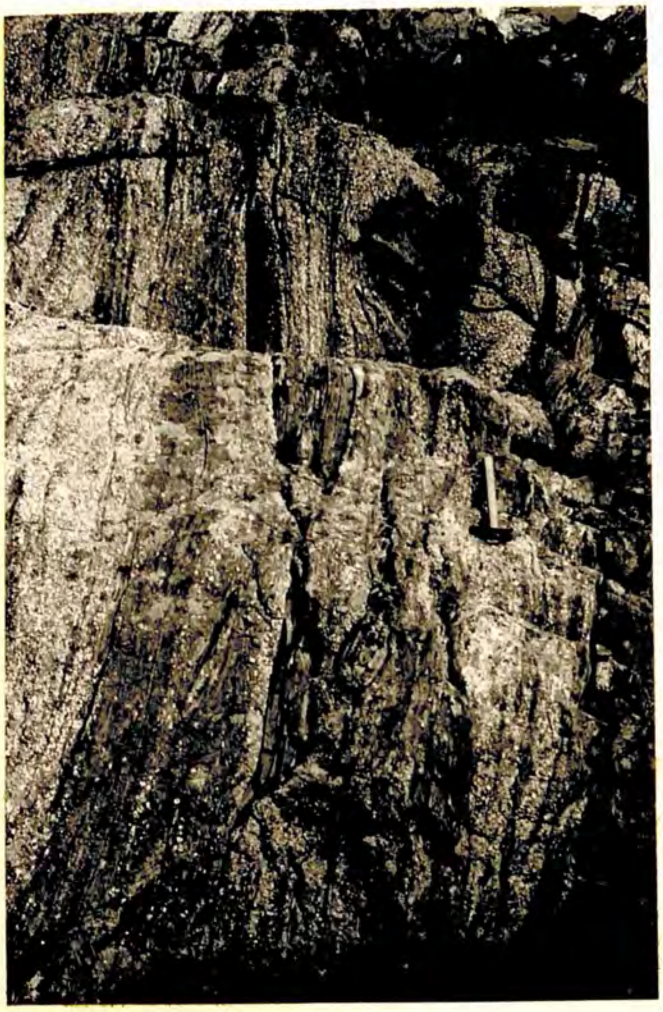


Plate 63. Metasedimentary blocks in feldspathized metagabbro. Above Havnefjordbotn.

Plate 64. Metasedimentary blocks in feldspathized metagabbro. Above Havnefjordbotn.

KOMAGFJORD DIORITE

This is a yellowish-brown weathering, medium-grained, hypersthene-bearing diorite which outcrops on the eastern side of Komagfjord. It forms narrow sheets within the Husfjord metagabbro, but its contact relationships are often obscured by the later perthosites in the Komagfjord area.

It has a slight foliation due to alignment of minerals, especially at the margins where the foliation is parallel to the contacts. Where contact with the Husfjord metagabbro is seen, the diorite occasionally contains blocks of the metagabbro at its margin.

The diorite characteristically contains numerous xenolithic inclusions of both melanocratic material and mesocratic material. The latter are commonly long and narrow, and are elongated parallel to the foliation of the diorite (Plate 65). This foliation and attitude of the xenoliths is sub-parallel to the trend of the foliation in the Husfjord metagabbro. Many of the xenoliths are partially assimilated by the diorite, and a streaky rock full of schlieren results. The melanocratic inclusions are generally fine-grained, granulated basic pyroxene hornfelses, and may or may not be remnants of Husfjord metagabbro. Some of the basic inclusions are fine-grained amphibolites, and appear to be fragments of some very early basic dykes. The fact that some xenoliths are pyroxene hornfelses and some are amphibolites shows that the hornfelsing cannot be due to the enclosing Komagfjord norite. The pyroxene hornfelses resemble the metagabbro which has been thermally metamorphosed by the early diorite sheets, suggesting that they were hornfelsed prior to

Plate 65. Xenoliths in Komagfjord diorite.
Komagfjord.

incorporation in the Komagfjord diorite. Similarly, it is possible that the amphibolite was also in its state of amphibolization before its enclosure within the diorite.

The buff-coloured, mesocratic xenoliths are also fine-grained granulated pyroxene hornfelses, and it is difficult to be sure whether they are of igneous or sedimentary origin. If the latter is the case, they may have been derived from metasediments to the south of the Husfjord metagabbro which are not to be seen now at the surface, due to erosion.

The Komagfjord diorite has been cut by some minor intrusions. These include a grey basic diorite, which forms small sheet-like bodies, and a few dykes. The dykes are of two kinds; a greenish porphyritic amphibolite, and a fine-grained basic dyke which has been emplaced into a sigmoidal joint.

It is also veined by perthosite, and it occasionally occurs as xenoliths and rafts within the perthosites. In one composite raft, the diorite is cut by the later grey basic diorite, which appears to have a dyke-like form and which does not penetrate into the surrounding perthosite.

KOBBERFJORD NORITE

This is a garnet-bearing orthonorite which outcrops in two places in the Husfjord area. One outcrop is at the northern end of Kobberfjord, where it occurs along the contact between the Husfjord metagabbro and the country rocks. The other outcrop is on the Fella peninsula at the eastern end of the area.

Kobberfjord Body

The norite body at Kobberfjord is in the form of a sheet about 200 m. wide, dipping steeply to the N.W. It has been emplaced along the contact between the Husfjord metagabbro and the migmatitic country rocks. It follows this boundary, which is slightly discordant with the banding in the metasediments, north-eastwards for about 3 km. It then swings away from the country rocks and penetrates the xenolithic facies of the metagabbro for about 1½ km., before petering out. In places it has branching tongues which penetrate the host-rock.

The norite is a fine-grained, non-porphyritic rock which weathers to a yellowish-brown colour. Micaceous patches are black, and the garnets are dark brown. It is extremely xenolithic, inclusions varying from large rafts of migmatite several metres long (Plate 66) to small fragments of psammite only a few centimetres or less in length (Plate 67).

The actual contact on the south-eastern side is not seen, but the north-western contact against the migmatite is clearly defined, and is well exposed on the coast. Here, the contact is not sharp, but consists of a zone about a metre in width in which the rocks are a mixture of

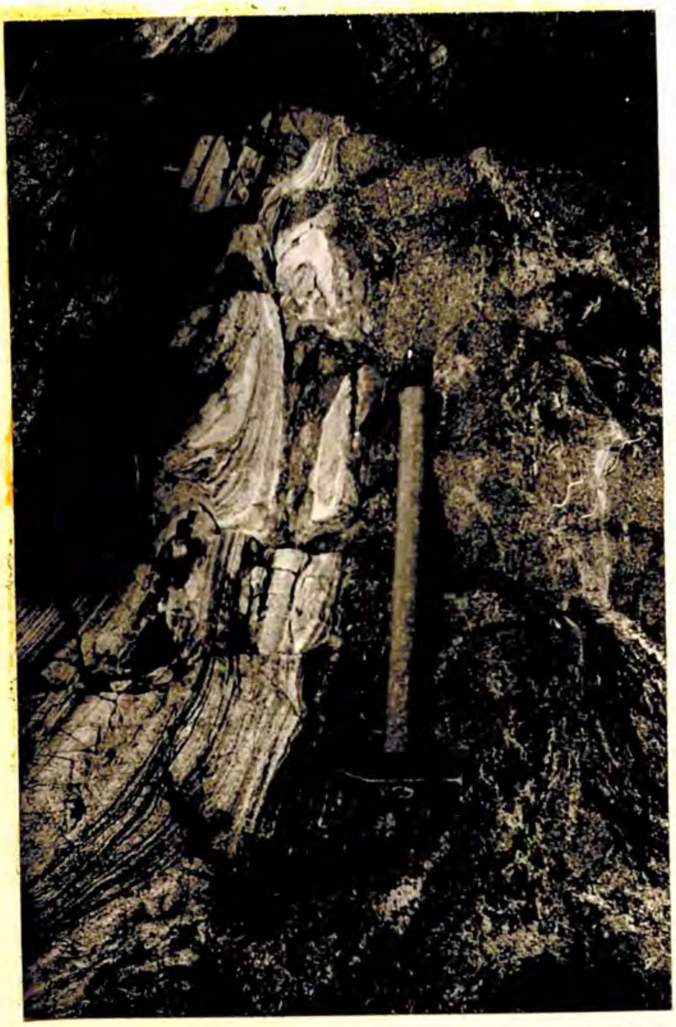


Plate 66. Migmatite raft in Kobberfjord norite. Kobberfjord.

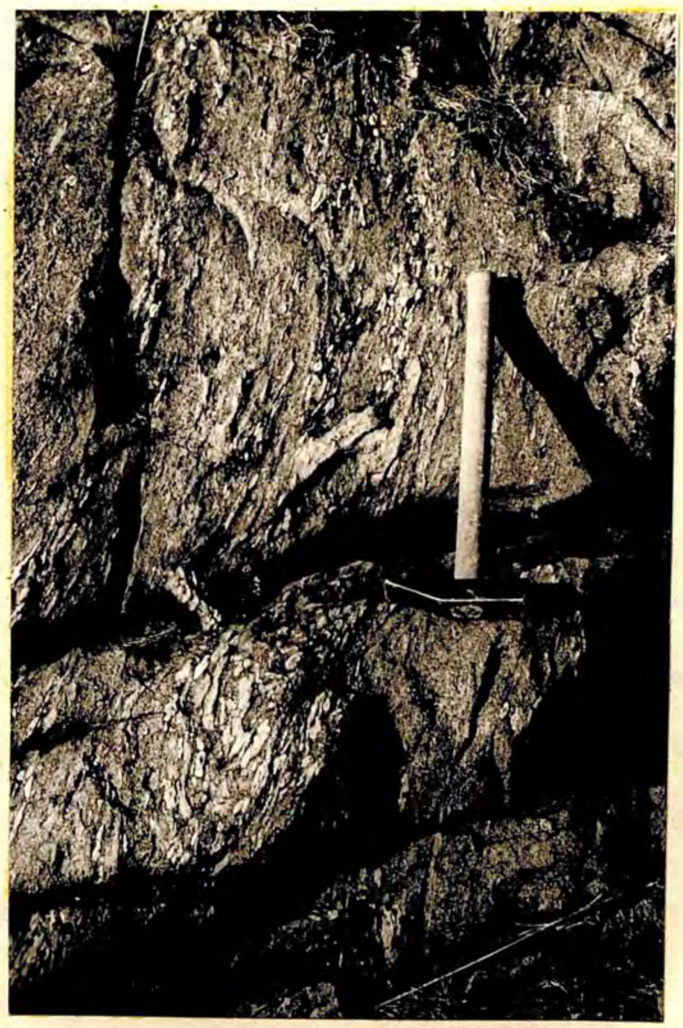


Plate 67. Psammite xenoliths in Kobberfjord norite. Kobberfjord.

contaminated norite and assimilated migmatite. At the margin, the norite is slightly foliated parallel to the contact. Many of the small psammite xenoliths are platy in form, and at the margin the larger of the slivers are orientated parallel to the contact (Plate 68), whilst the smaller ones and the irregularly-shaped ones tend to be randomly orientated. This marginal zone is rich in brown garnets (Plate 68) which are frequently as large as 2 cm. in diameter.

A short way from the contact, the norite contains numerous rafts and blocks of migmatite (as in Plate 66), some of which are clearly being assimilated by the norite. Some of these migmatite rafts contain basic sheets, and occasionally isolated basic sheet material forms xenoliths (Plate 69). All these rafts are generally orientated parallel to the norite margin.

Further away from the contact, the norite is not so foliated, and the small fragmentary psammitic xenoliths become more randomly orientated. The edges of these small xenoliths show all gradations from clean, sharp borders to diffuse, partly-assimilated margins. At greater distances from the norite margin, the number of xenoliths decreases, but the degree to which they are assimilated increases.

In the central part of the norite the metasedimentary blocks, which are usually only a few tens of centimetres in length, are extensively assimilated. In some cases, they are becoming progressively fragmented at their margins, apparently by physical action of the norite (Plate 70). Fragments are being broken off the edges and are being incorporated in the norite, but those adjacent to the xenoliths still retain their

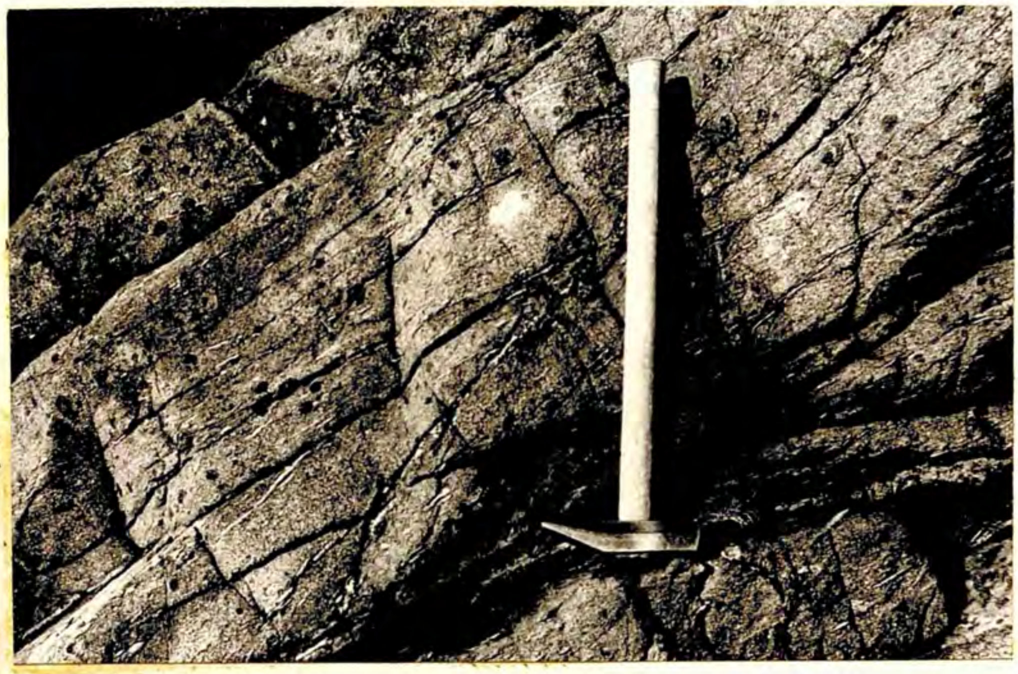


Plate 68. Slivers of psammite in norite orientated parallel to the margin of the norite. Kobberfjord.

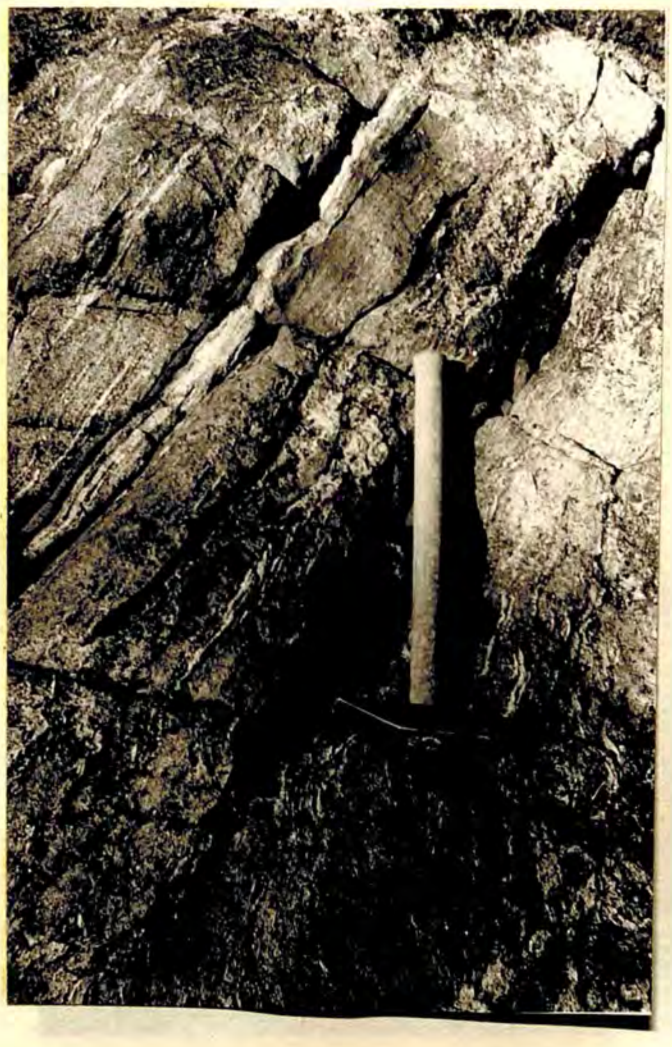


Plate 69. Psammitic and basic sheet xenoliths in norite. Kobberfjord.

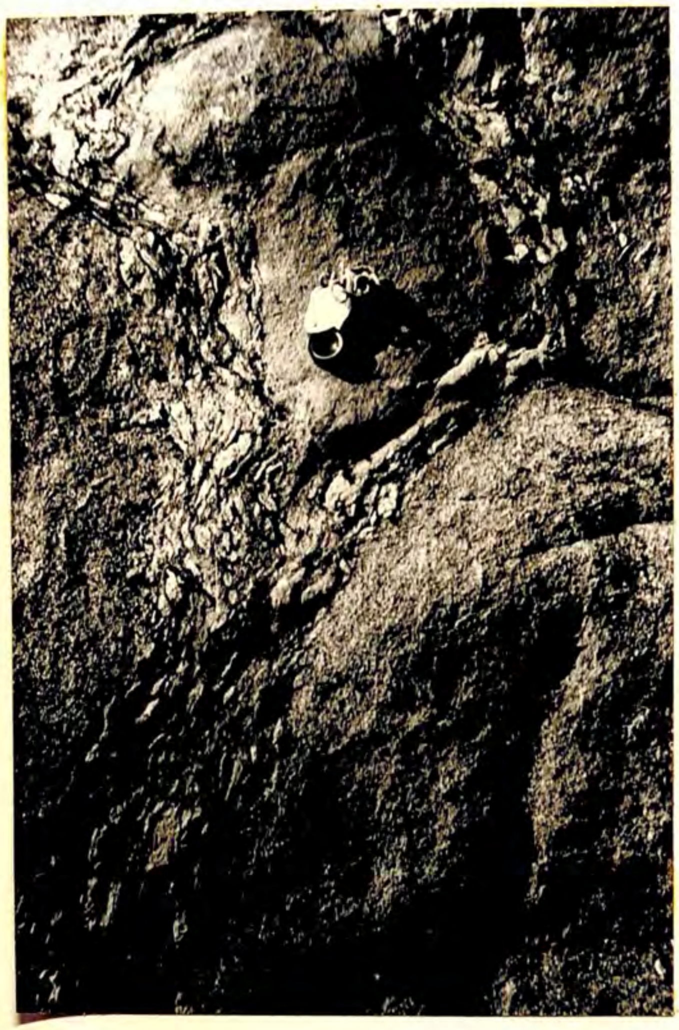


Plate 70. Xenoliths becoming fragmented and assimilated by norite. Kobberfjord.

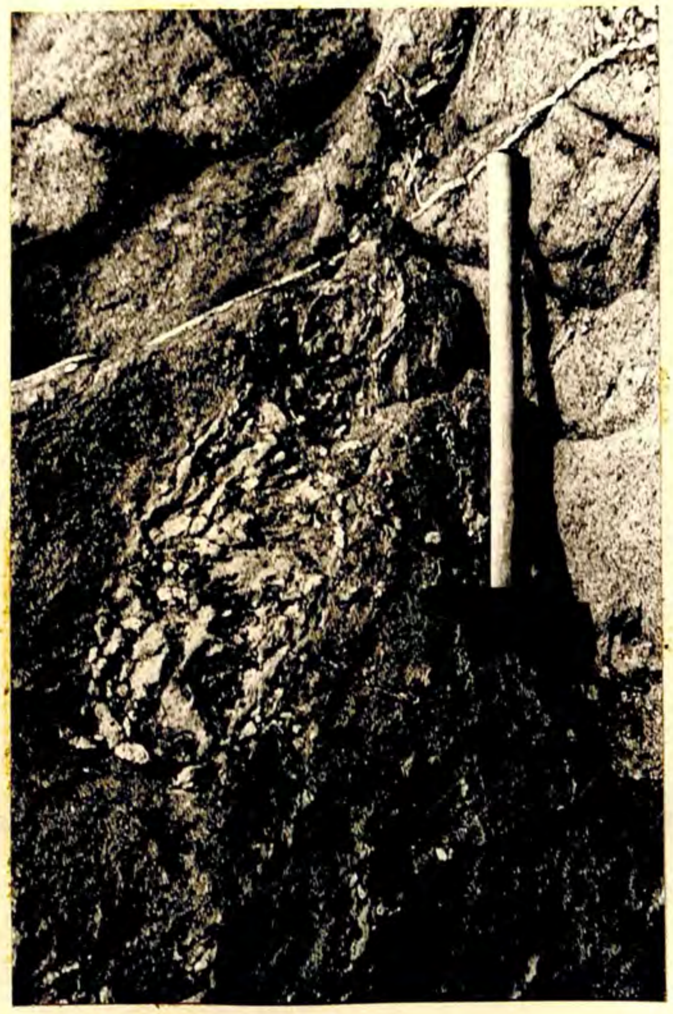


Plate 71. Fragmented rounded xenolith in norite. Kobberfjord.

identity and lie parallel to the margins of the xenolith. With increasing distance from the xenolith, the fragments become progressively more assimilated by the norite until they become very diffuse and merge with the host. In some cases the norite contains ragged patches of psammite fragments which do not have any intact psammite cores (Plate 71). These patches of psammite fragments are sometimes elongate, and it is clear that they represent long xenolithic blocks which have been completely fragmented and partly assimilated by the norite (Plate 72). Occasionally these fragmented blocks are up to a metre or so in width (Plate 73).

Away from the contact zone, the size of the garnets decreases. They are quite abundant in the neighbourhood of the rafts and xenoliths, but never attain the size of those at the margin. Where the norite contains small scattered psammitic fragments, the garnets are quite small and are dispersed irregularly throughout the rock.

The field evidence clearly indicates that the formation of the garnets is associated in some way with contamination of the noritic magma by the metasediments; the amount of garnet is greater where metasedimentary material has been assimilated. The xenoliths that remain are mainly of migmatized psammite with some semi-pelite, and these have obviously resisted assimilation to some extent. The xenoliths were presumably derived from the migmatized psammites and semi-pelites of the country rocks, so it is probable that a considerable amount of semi-pelitic material has been assimilated. This contamination would give the noritic magma an excess of aluminium, possibly enabling the formation

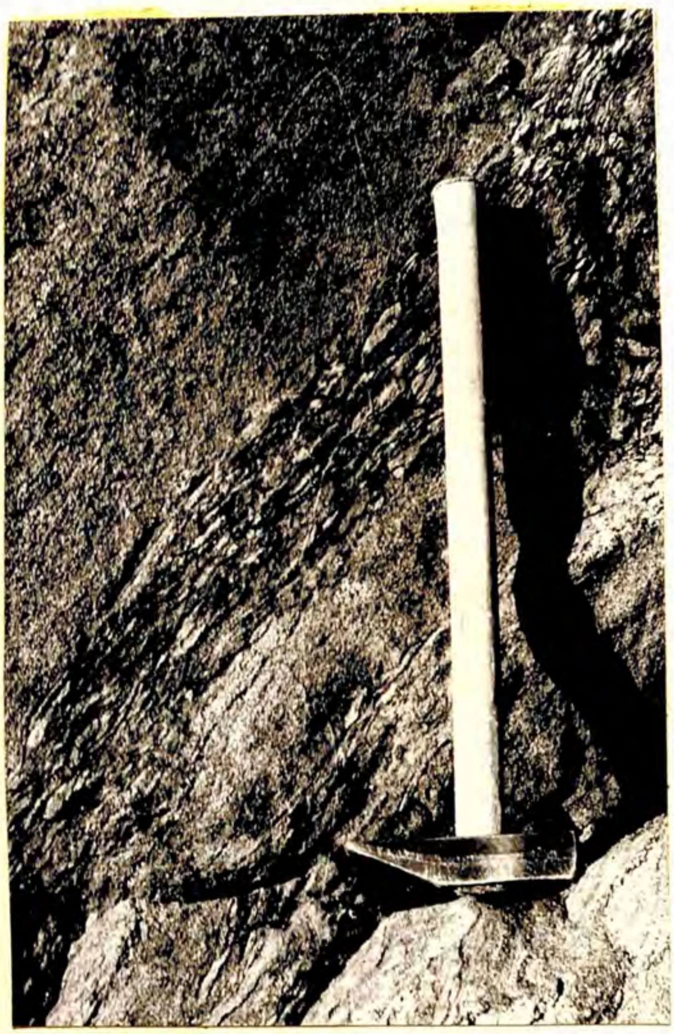


Plate 72. Fragmented elongated xenolith in norite. Kobberfjord.

Plate 73. Fragmented psammite block in norite. Kobberfjord.

of garnet. Whether this garnet was a direct result of contamination or whether it was a metamorphic product will be discussed in the petrographic section (p. 227).

The presence in the norite of migmatite xenoliths and the absence of Husfjord metagabbro inclusions indicate that the norite here was emplaced on the country rock side of the gabbro/metasediment contact. Towards the north-east, the migmatite inclusions become less numerous and disappear when the norite penetrates the metagabbro.

The facts that the inclusions are of migmatized metasediment, and that the norite itself is not migmatized show that the emplacement of the norite post-dated the migmatization.

Fella Body

The Kobberfjord norite rock type constitutes the southern part of the Fella peninsula at the eastern end of the area. The form of this body is not known, as its northern contact has not been observed, and it is bounded on all other sides by the sea. To the north it makes contact with the Husfjord metagabbro, although much of the contact is now occupied by a tongue of the Havnefjord diorite, which may have been emplaced along the contact zone.

This norite closely resembles that of the Kobberfjord body; it is full of metasedimentary xenoliths of various sizes, and contains numerous small garnets dispersed throughout the rock. Many of the metasedimentary blocks are becoming progressively fragmented and assimilated like those at Kobberfjord.

The rafts however are not of migmatite. On the eastern side of Fella the rafts trend E-W, and are mainly of hornfelsed psammites and semi-pelites, often rusty-weathering and resembling some of the rafts in the Havnefjord diorite. At the margins of these rafts considerable assimilation has usually occurred, and they often merge into the norite. On the western side of Fella the majority of the rafts are of white calc-silicate-schists which have been hornfelsed. These trend in a N.W.-S.E. direction and are often several tens of metres in length.

VATNA GABBRO

General

The Vatna gabbro is a fine-grained, olivine-bearing gabbro which outcrops in the Vatna area just south-west of Husfjord. It has been emplaced into the south-eastern part of the Husfjord metagabbro, and forms the slightly lower, craggy ground around Vatna and Vatnhavn.

The gabbro weathers to a rusty-brown colour, which in places is pinkish. In places it has a fluxion structure formed by the preferred orientation of the minerals. The strike of this orientation is about 070° and it dips at 70° to the south.

A number of xenoliths are enclosed within the gabbro and usually occur in zones. The xenoliths are small; the largest are a few tens of centimetres in length. Occasionally the xenoliths are randomly orientated and neighbouring ones are of different rock-types; some of them are fine-grained, and some are coarse-grained (Plate 74). They are mainly fragments of basic dyke, and some of the coarse ones resemble the coarse-grained southern Slåtten gabbro (see p. 117).

Generally however, neighbouring inclusions are all of one kind; these are finer-grained than the host, but they weather to a colour similar to that of the latter. These fine-grained xenoliths are generally elongate and they are aligned parallel to one another and to the attitude of the flusion structure in the gabbro. They usually show fairly sharp margins against the gabbro. In thin section these inclusions are seen to be finer-grained versions of the gabbro itself; their mineralogy and texture are similar to those of the gabbro, but the scale is different.

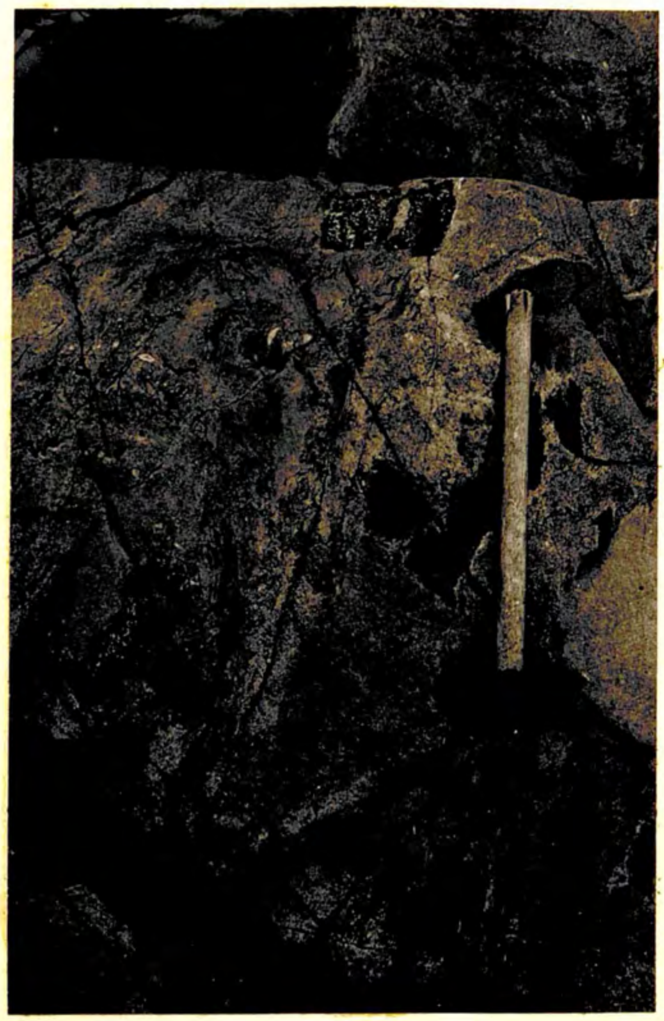


Plate 74. Mixed xenoliths in Vatna gabbro.
Vatna.

Thus a better term for these inclusions might be autoliths since they consist of the same rock type as their host. It would seem that before the main facies of the Vatna gabbro now exposed here was emplaced, a finer-grained facies was formed. This suggests that the emplacement of the Vatna gabbro was complicated and possibly consisted of more than one phase. This is discussed further, and a possible explanation of the autoliths is offered in the section on emplacement (p.129).

The gabbro often has a streaky appearance, in which the rock is streaked through with bands and veins of coarser, pink-weathering feldspathic material (Plate 75). These veins are subparallel to one another, and their attitude is parallel to the attitude of the autoliths and fluxion structure in the gabbro. Sometimes the veins have sharp margins, and sometimes they are diffuse and mix with the gabbro. The veins are of perthite, and 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 at some time during its emplacement the Vatna gabbro became locally mixed with a comparatively sodic magma, possibly perthositic. This complication will be discussed when the emplacement of the perthosites is considered (p.141).

Ore-shoots

In the rocks on the coast about 200 m. to the west of the cottage at Vatna there are a few ore-shoots in the gabbro. These are rare, and the largest are about 30 cm. long and a few cm. wide. They are rusty-weathering, and consist principally of magnetite. They occur mainly in

Plate 75. Streaky felds-
pathic facies of Vatna
gabbro. Vatna.

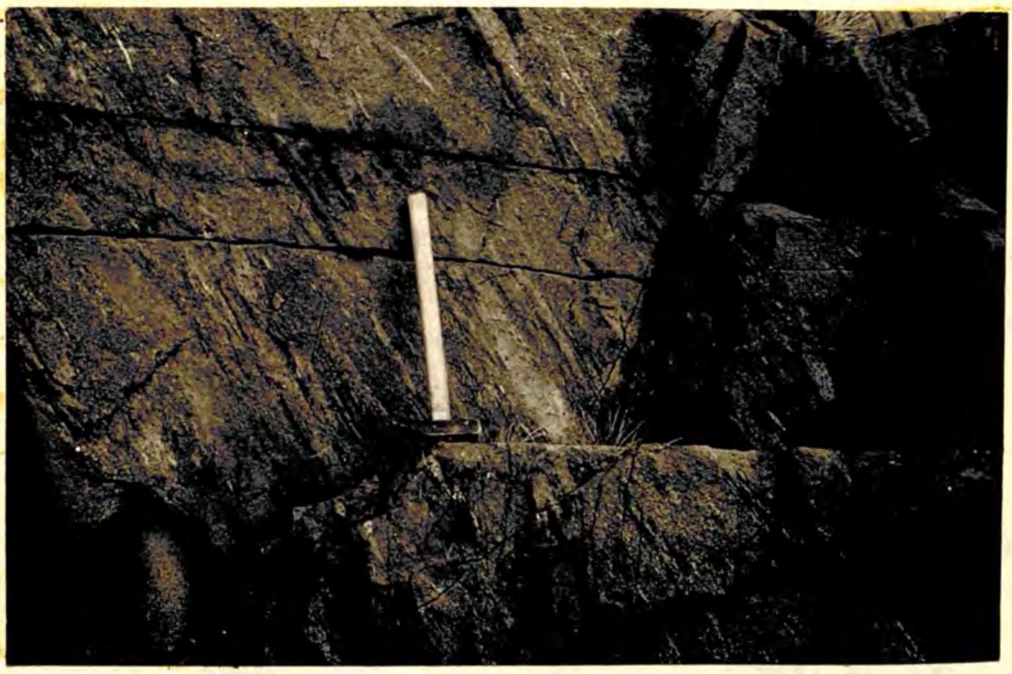


Plate 76. Ore-shoots in Vatna gabbro. Shoots are the dark streaks to the left of the hammer head and the end of the shaft. Other streaks are inclusions and feldspathic veins. Vatna.

the streaky facies and are associated with the antiperthitic veins. Their attitude is parallel to the perthosite veins and the autoliths so that they dip steeply to the south (Plate 76). This is significant, since a geomagnetic survey made by Dr. M. Brooks (personal communication) has shown that there is a high positive magnetic anomaly in Sørøysund just off the coast at Vatna. This confirms magnetic anomalies for this area reported on the navigation charts (Norges Sjøkartverk og Norsk Polarinstitutt (1965)). It is likely that the ore-shoots are associated with a larger source of ore, and their attitude suggests that a large ore body may exist beneath Sørøysund off the coast at Vatna.

The navigation charts also indicate that the magnetic anomaly continues north-eastwards along Sørøysund, suggesting that an ultra-basic or ore-bearing body may continue along this line at depth. The great depths attained in Sørøysund along this line (up to 420 m.) suggest that this line may mark the site of a large fault, up which ultrabasic material may have risen, producing the magnetic anomalies.

Dr. M. Brooks has found that perthosite sheets in Komagfjord, to the west of Vatna, have high positive magnetic anomalies (personal communication), although these rocks are virtually pure perthite. Clearly there must be some highly magnetic rocks beneath the surface, and this provides further evidence of the close association between ore and sodic magma. Furthermore, Sturt and Ramsay (1965) describe magnetite-bearing perthitic pe gmatites in association with the alkaline complex of the Breivikbotn area.

The actual shape of the Vatna gabbro body cannot be determined; the gabbro is bordered to the south-east by the sea, and its contact with the Husfjord metagabbro is not seen because of differential weathering. However, it is likely that the fluxion structure and the plane in which the inclusions lie are subparallel to the margin of the body. The trend of the north-western margin is approximately N.E.-S.W., a direction which is roughly parallel to the strike of the fluxion structure. It is considered likely that this contact dips steeply to the south, parallel to the foliation.

About 2½ km. to the north of Vatna, near Slåtten, there is an outcrop of a fine-grained, olivine-bearing gabbro which forms a sheet about 200 m. wide dipping steeply to the south. This gabbro is rusty-weathering and has a marked fluxion structure due to parallel orientation of the crystals. The fluxion has a strike of 060° and generally dips steeply to the south, although in places it is subvertical.

This gabbro is similar to the Vatna gabbro, although it does not contain the finer-grained autoliths and perthositic streaks. It is considered that it is a similar rock-type as the Vatna gabbro, and may represent a subsidiary sheet to that body. This suggests that the Vatna gabbro itself might be a sheet-like body dipping under Sørøysund to the south.

Minor Intrusions

Minor intrusions are abundant in the Vatna gabbro, and include perthosites, basic dykes, and nepheline-syenite pegmatites. They will be considered in detail under a separate heading (p.132).

The perthosites are pink, and are variable in grain-size although they are usually fairly coarse-grained. They form large dyke-like sheets which are subvertical and are usually several tens of metres in width. Their general trend is approximately N.W.-S.E., although individual sheets tend to waver and branch.

The basic dykes vary in width from a centimetre to about 2 metres. Their average trend is 140° and they are subvertical. The Vatna gabbro does not have a particularly regular jointing, but these dykes were probably emplaced along whatever joints were formed.

The earliest dykes were ultrabasic; these have diffuse margins, are sometimes feldspathized, and are slightly folded. It is possible that they were emplaced before the gabbro had become absolutely rigid, and were deformed by slight movements in the gabbro. Evidence for an early pre-perthosite phase of dyke emplacement is provided by xenoliths of dyke material in some of the perthosites. However, most of the dykes belong to a later, post-perthosite phase since many of them cut the perthosites.

Almost invariably the basic dykes have eroded more readily than the gabbro. This gives the ground a highly dissected appearance, and on the coast the sea has eroded long narrow gullies penetrating into the headlands.

There are a few bodies of nepheline-syenite pegmatite in the Vatna gabbro. These are up to about $1\frac{1}{2}$ m. in width, and have been variably sheared during the later F_2 brittle deformation.

Shears and Mylonites

In the Vatna gabbro there are a number of shears which were produced during the late F_2 movements. The majority of these are very narrow, being only a millimetre or so in width, and they are best seen when they cause the offsetting of basic dykes. Both dextral and sinistral types occur, but they do not provide enough data to enable a statistal pattern to emerge.

There are a few broad shear-zones up to about $1\frac{1}{2}$ m. in thickness in which mylonite are formed. It is in these mylonite zones that the nepheline-syenite pegmatite occurs, and it appears that they have provided planes of weakness along which shearing has taken place. The pegmatite does not always occupy the whole zone, and where the gabbro has become sheared it forms a hard, black, laminated mylonite.

The planes of shearing in these mylonites strike in an E.S.E.-W.N.W. direction and dip at 40°-60° to the south. Thus they are parallel to the mylonites in the Husfjord metagabbro (Fig. 7), and are probably related to the formation of the late F_2 structures.

SLÅTTEN GABBROS

On the west side of Husfjord, near Slåtten, there are three gabbros, each of which appears to form a sheet-like body. The sheets are parallel to one another and dip steeply to the south.

The central gabbro is of the Vatna gabbro rock-type, and was described in connection with the Vatna body (p.113). It is characteristically a brownish, fine-grained rock containing occasional rusty-weathering streaks, which are probably of iron ore. It generally has a marked fluxion structure which usually dips to the south, and is considered to be an outlying sheet of the Vatna gabbro.

The northern gabbro is a medium-grained pyroxene gabbro with only a very little olivine, and it weathers to a black and white speckled appearance. It has been intruded by a number of porphyritic dykes which have white-weathering feldspar phenocrysts giving the rock a spotted character. The trend of the dykes is parallel to that of the gabbro sheet itself, i.e. about 070°. To the north, this gabbro makes contact with the Husfjord metagabbro, but the contact zone has become complicated by the emplacement of numerous perthosite sheets parallel to the contact. These sheets are abundant in the Husfjord metagabbro just to the north of the contact, and they also occur within the olivine-bearing gabbro near its northern margin. The perthosites generally contain xenolithic blocks, and in the case of the perthosites which are intruded into the olivine gabbro, these blocks are of the gabbro itself, and also of the spotted porphyritic dyke. In places the perthosite veins the gabbro in an irregular fashion.

The southern gabbro is a coarse-grained olivine-gabbro. It does not have a fluxion structure; the large black pyroxenes and yellowish-weathering feldspars form a coarse ophitic texture. The grain-size is somewhat variable; generally, large grains are in the order of 4 - 5 mm. in length, but occasionally the grain-size increases so that a pegmatitic facies develops. This latter occurs at the northern margin of the body, where it sometimes veins into the adjacent gabbro. The general trend of the southern margin, which swings southwards along the mountainside, indicates that this margin dips towards the south.

The exact lateral extent of these gabbro sheets is not known. A short way westwards from the coast they traverse extremely steep and rough country, much of which is inaccessible. However, they do not occur in Komagakslen to the west of Vatndal, so they must tongue out before then. They are widening towards the east, and it is likely that they extend for some distance into Husfjord, and may even pass south of Husfjordness.

STRUCTURE AND EMPLACEMENT OF THE MAJOR INTRUSIONS

Introduction

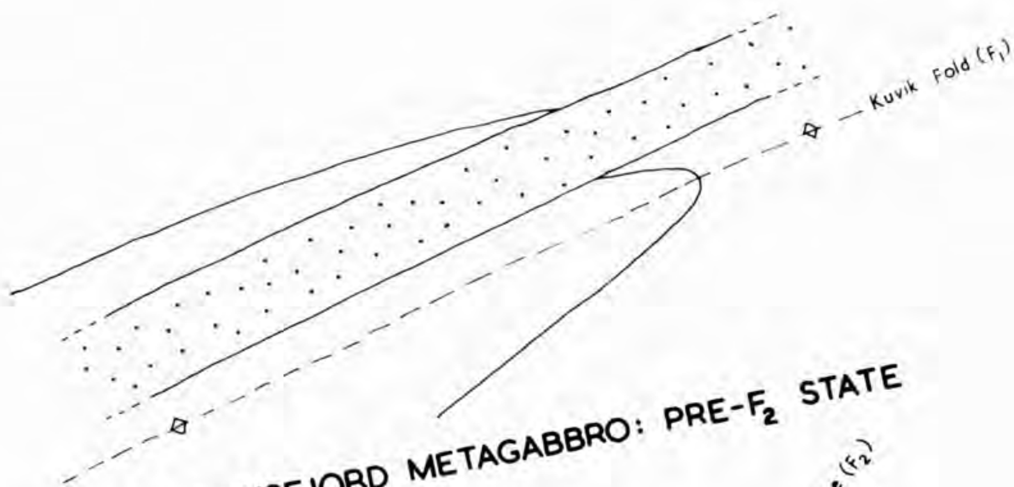
It is important to consider the emplacement mechanisms of the principal major intrusions, and in particular their relationships to the tectonics of the area. These bodies are sheet-like in form, and were emplaced during the Caledonian orogeny, probably at great depths. Buddington (1959) states that conformable sheets (including laccoliths, phacoliths, etc.) are the normal type of syntectonic plutonic intrusion in the catazone (approx. 7 - 12 miles depth).

The emplacement of the plutonic bodies in the Husfjord area is closely related to the major structures in the country rocks, especially so since they were emplaced syntectonically. Thus reference will be made to the major regional structures discussed on p. 41 , and shown in Figs. 8, 9, and 10.

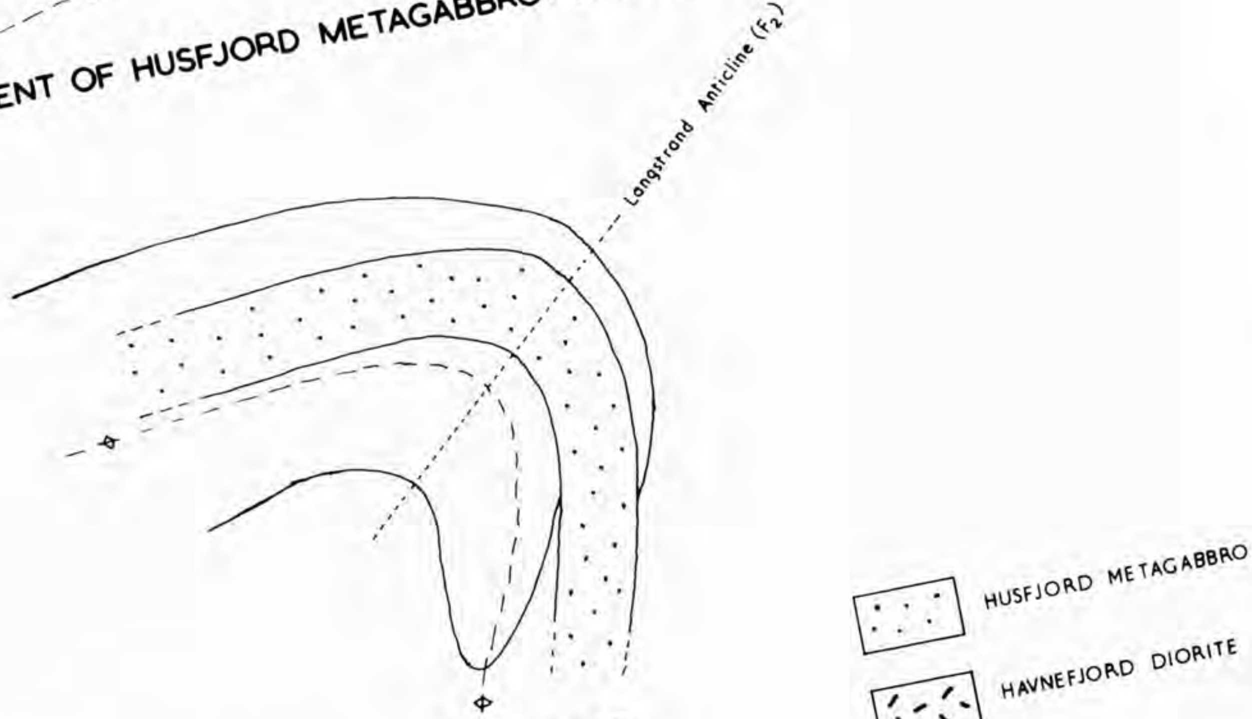
Husfjord Metagabbro

This forms a subconcordant sheet which was emplaced during the very last stages of the F_1 folding. It was emplaced along the upper limb of the F_1 Kuvik Fold (Fig. 13A; cf. Figs. 9 and 10), although its slight discordance shows that it was not emplaced exactly along the banding in the metasediments. The metagabbro passes up through the metasedimentary succession towards the north-east, and the degree of discordancy increases in this direction.

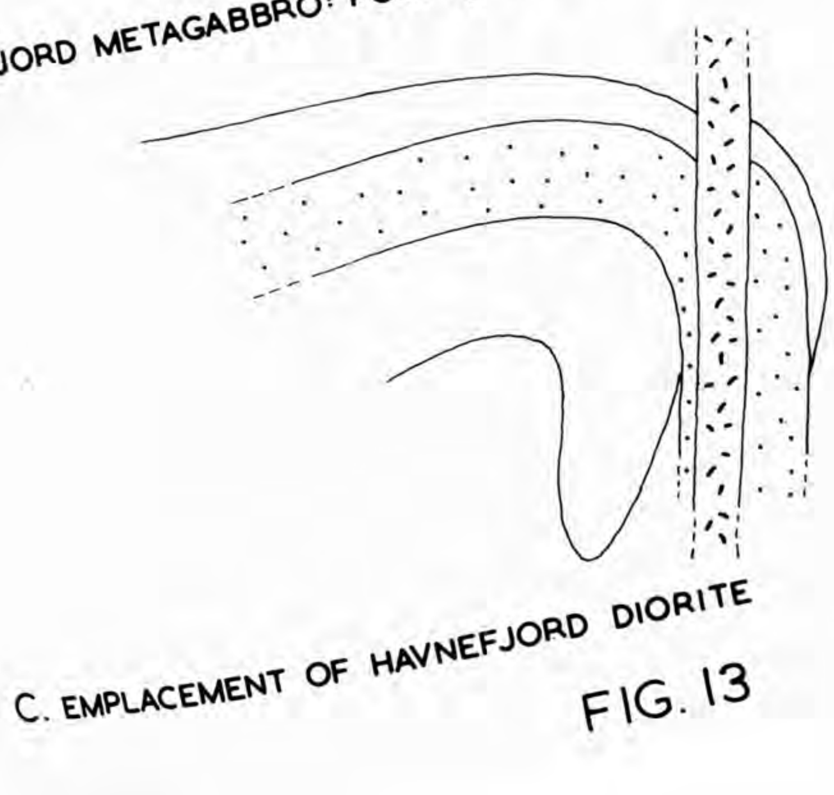
In the Langstrand area the hinge of the Kuvik Fold is seen (Roberts, 1965); thus as the Husfjord metagabbro is traced towards the



A. EMPLACEMENT OF HUSFJORD METAGABBRO: PRE-F_2 STATE



B. HUSFJORD METAGABBRO: POST-F_2 STATE



C. EMPLACEMENT OF HAVNEFJORD DIORITE

FIG. 13

north-east it is approaching the hinge of the F_1 fold. This provides an explanation for the way in which it moves up through the succession and becomes progressively more discordant towards the north-east. This relationship is demonstrated clearly in Fig. 13B, which shows the position of the metagabbro after the first stages of the F_2 folding. Thus it would appear that the metagabbro sheet has been emplaced sub-parallel to the axial plane of the F_1 fold, and because of the tightly isoclinal nature of the fold, the sheet is subparallel to the banding on the limbs of the fold, but discordant at its hinge. Its emplacement clearly post-dates the formation of the F_1 fold since it is not bent round the F_1 fold-closure. This is demonstrated well in Fig. 2 from Ramsay and Sturt (1963) in which a gabbro sheet is shown to be post F_1 since it is not folded by an F_1 synclinal fold.

The Husfjord metagabbro contains a number of xenolithic rafts, mainly of metalimestone, which have been described on p. 65 . It has been suggested by some workers in the general region (e.g. Krauskopf, 1954; Oosterom, 1954) that the presence of metalimestone rafts in gabbro indicates that the gabbro was a product of extreme metamorphism, and that the rafts represent parts which have not become completely gabbroized. In the Husfjord metagabbro there is no evidence to suggest that it is a product of extreme metamorphism, and the rafts are interpreted as xenolithic inclusions enclosed within the gabbro magma. Thus they provide an important clue to the emplacement mechanism of the metagabbro.

The rafts occur near the metagabbro's northern margin and also just to the north of the Havnefjord diorite. It is considered that they had not been moved far, if at all, from their original positions, since it appears that the gabbro magma enveloped these rafts without disturbing them greatly. The Havnefjord diorite is also full of metasedimentary and metagabbroic rafts, and it is considered that the area now occupied by the diorite was formerly a large lens-shaped screen of country rocks. This explains the presence of the metalimestone rafts in the metagabbro so far from its outer contact. The metalimestone belt appears to have been transected by the incoming gabbro magma so that part of it provided rafts at the outer gabbro contact, and part provided rafts at the inner subsidiary contact with the metasedimentary screen. Thus the emplacement of the gabbro involved splitting the metasediments, for the most part approximately along their banding, and shouldering them apart to allow its entry. The actual plane of entry appears to be parallel to the axial plane of the F_1 fold, and its entry must have been guided by the tectonic stress field operating at the time.

There is no evidence of deformation in the country rocks to suggest that the emplacement of the gabbro was very forceful. It has clearly entered along a plane of weakness in the country rocks, and has not had to exert a great deal of pressure to ensure its emplacement. In other words, the gabbro emplacement was of a fairly permissive nature.

A permissive intrusion has been described by Pirsson (1914) as a type of intrusion in which the magma is emplaced by flowing into space that has been opened for it by some force other than one exerted by the invading magma. Such a force could be tectonic, the magma flowing into cavities created by tectonic stresses (Dietrich, 1954). It is considered that such a mechanism operated during the emplacement of the

Husfjord metagabbro.

The large screen of country rock which is considered to have been preserved within the gabbro appears to have been quite complex, and not a simple, single screen consisting entirely of metasediment. Within the Havnefjord diorite, which now occupies this former screen, there are sometimes trains of metagabbro rafts occurring between the trains of metasedimentary rafts. This indicates that the screen was itself a complex of smaller subsidiary screens separated by narrow sheets of metagabbro. These minor metagabbro sheets must have been subconcordant, and were presumably connected with the main metagabbro body. The term Stromatolith was proposed by Foye (1916) for a rock mass consisting of many alternating layers of igneous and sedimentary rocks in sill relationship. Although the rocks described by Foye formed an overall dome-like structure, it is proposed to apply this appropriate term to the pre-diorite form of the intragabbroic screen complex.

In summary then, the Husfjord metagabbro was emplaced as a sheet along the upper limb of the Kuvik Fold, probably parallel to the axial plane of that fold. Its emplacement was of a permissive nature, and a large screen of country rock was formed, which had the form of a stromatolith.

Havnefjord Diorite

The emplacement of the Havnefjord diorite occurred after the peak of the regional metamorphism, and was syntectonic with respect to the F_2 movements. It has been emplaced into the steeply overturned limb of the continuation of the F_2 Langstrand Antiform (Fig. 13C; cf. Figs. 9 and 10), and has been deformed by the late F_2 movements, although the main F_2 folding must have commenced before the diorite emplacement since the diorite has been emplaced along the limb of an already formed F_2 fold.

The diorite contains a large number of rafts as described on p. 86. Again, there is no evidence that the diorite was formed in situ as a product of high-grade regional metamorphism, and the rafts are interpreted as xenoliths enclosed by the invading dioritic magma. It is not possible to ascertain in every case whether or not the raft has been reorientated. In cases where the dip of the metasediments in the raft is parallel to the regional dip in the country rocks, then it is probable that the raft has not been reorientated. In those cases where the dip of the metasediments (which is usually parallel to the plane of the raft) is subvertical, the evidence is more equivocable; the verticality may be due to reorientation, or to a pre-diorite vertical attitude of the metasediments on a local scale. In either case, they probably have not been moved far from their original positions, particularly since the distribution of the different lithologies of the metasedimentary rafts reflects the stratigraphy of the country rocks outside the complex. This latter point is also important in demonstrating that the stromatolithic screen complex enclosed within the Husfjord metagabbro cannot have been displaced much, if at all, from its original position.

Although the rafts in the diorite are slightly mixed, there is in general an orderly succession across the diorite body; metalimestones and calc-silicate-schists in the north, and rusty-weathering psammites and semi-pelites in the south. The former are correlated with the calc-silicate-schist with metalimestones group, and the latter probably with some of the Breivik Group. The slight mixing shows that there was probably some relative movement within the diorite body, which must presumably have been in the form of a magma. However, the relict stratigraphy preserved within the diorite shows that even if the diorite magma itself travelled some distance from its place of formation, its enclosed rafts have not been moved far from their original positions.

The Havnefjord diorite has many characteristics in common with the Main Donegal Granite, described by Pitcher and Read (1959), and also certain features in common with the Thorr Granodiorite, Co. Donegal, described by Pitcher (1953a). Both these bodies are of Caledonian age and were associated with the Caledonian orogeny. However, whereas the Havnefjord diorite was emplaced syntectonically in the middle of the orogeny, the Donegal examples are considered by both Pitcher and Read to have been emplaced just after the main orogenic event, and thus to be post-tectonic.

The Main Donegal Granite contains a large number of free-swimming rafts of the country rock, occurring in narrow zones. These rafts vary in size from a few centimetres to 30 m. or more in length, and can be related to horizons in the wall or roof. The trains of rafts run parallel to the flow pattern in the granite, and although there is slight mixing, there is an orderly succession of lithological types across the granite. Nowhere is there a very strong reaction between the granite and the enclaves, but it is slightly more marked in the central zones. The rafts are interpreted by Pitcher and Read (op.cit.) as material torn from the roof and walls by the magma, which probably moved laterally towards the south-west as suggested by the divergence of the raft trains in this direction.

The Thorr Granodiorite also contains numerous rafts of metasediments and amphibolites which vary from a few centimetres to about 800 m. in length. The larger ones have trends concordant with the strike and dip of the country rocks, but the smaller xenoliths near the rafts are disorientated. Psammites and amphibolites have sharp contacts with the host, but pelitic margins are gradational, and are being assimilated by the granodiorite. Pelites and semi-pelites, especially the small xenoliths, are often becoming feldspathized. Calcareous rafts form thick elongate lenses. Pitcher (op. cit.) considers that there is evidence for a mobile, probably very viscous, magma (e.g. the disorientation of some of the rafts), but also that there is evidence for migmatitic replacement in situ (e.g. the conformity of the large rafts with the regional structure, and the differential replacement of the various lithological types according to their relative resistances to assimilation). He states that it is difficult to determine with certainty which of these mechanisms was in operation, but if the latter was the case, then there must also have been some very slight movement within the body. In any case, it appears to have been a chemically reactive emplacement, in which the magma reacted strongly with its country rocks and inclusions (Pitcher and Read, 1963).

The Havnefjord diorite contains some very large rafts, mainly near its northern margin. These rafts have apparently been slightly mixed, and possibly reorientated, although a general relict stratigraphy is maintained through the body. However, the evidence that there has been relative movement between the rafts within the diorite is considered sufficient to indicate that the diorite must have been essentially a mobile magma.

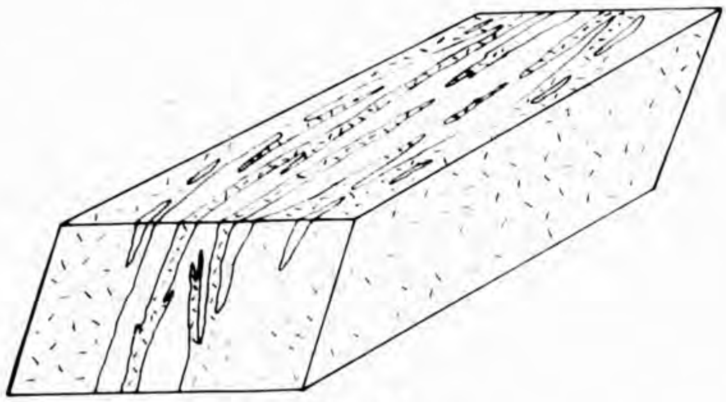
The Havnefjord diorite is regarded as being more analogous to the Main Donegal Granite than to the Thorr Granodiorite, although its mechanism of emplacement is believed to be a little different. In the Donegal Granite, emplacement is considered to be by lateral wedging (Pitcher and Read, 1959), but in the Havnefjord diorite this appears to be an unlikely mechanism. It is difficult to determine the direction of elongation, within the three-dimensional framework, of the rafts in the Havnefjord diorite, but the trends of the rafts do not suggest a lateral movement. Furthermore, the diorite tongues out at either end horizontally, thus making lateral emplacement most unlikely.

Dr. C. F. Tozer in the discussion on Pitcher and Read's paper (1959) suggested that the fact that the rafts in the Donegal Granite represent wall-rocks at the present surface is a criterion for horizontal movement. He considered that vertical movement would bring sediments up from lower levels. This might be true if it were known that the rafts have arrived at their present positions from elsewhere. If, on the other hand, the rafts have not been moved far, then it seems to the present writer that they will indeed represent wall-rocks which occur at the present surface, even if the emplacement of the host was in a vertical direction. Thus it is considered that the relict stratigraphy in the Havnefjord diorite does not prevent a vertical emplacement of the magma being postulated.

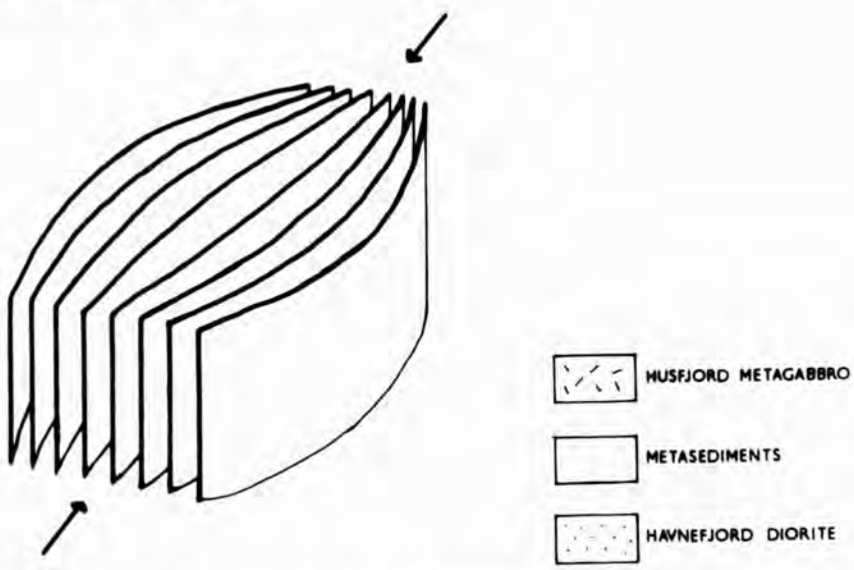
Prof. W. D. Gill, also in the discussion on Pitcher and Read's paper (1959) pointed out that there appeared to be a space problem because the country rocks did not give any evidence of forceful wedging apart by the granite. He favoured major stoping as a better explanation.

There is no evidence in the envelope of the Havnefjord diorite for forceful emplacement of the diorite, but a mechanism of major stoping is not invoked to explain it. This diorite body was emplaced into the steep limb of an F_2 fold during late stages of the F_2 movements. The site of its emplacement was the prismatic lenticular stromatolithic complex in the central part of the Husfjord metagabbro. This stromatolith would have been strongly laminated in comparison with the surrounding more massive metagabbro. During the development of the late F_2 arcuate cross-fold, there would have been a compressional force operating at the nodes of the lenticular stromatolith, and a tensional force at right angles to this direction (Fig. 14A). This would cause the members of the laminated stromatolith to pull apart from one another, leaving cavities or planes of weakness between them. This process is analogous to the behaviour of a sheaf of papers which are compressed in a direction parallel to their planar surfaces (Fig. 14B). As the cross-fold developed, differential movement between members of the

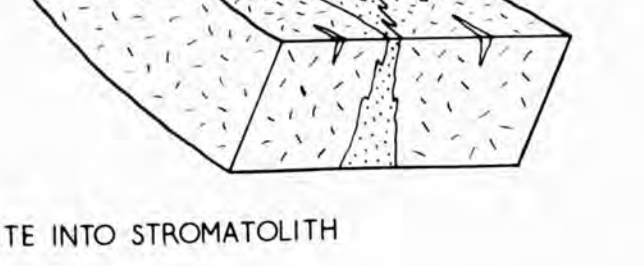
EMPLACEMENT OF HAVNEFJORD DIORITE



A. PRE-DIORITE STROMATOLITHIC METAGABBRO



B. DEFORMATION OF STROMATOLITH ILLUSTRATED BY SHEAF OF PAPERS



C. EMPLACEMENT OF DIORITE INTO STROMATOLITH

FIG. 14

stromatolith would also tend to occur. This environment would provide an ideal setting for the reception of any magma which might be attempting to rise from depth, and in the case under discussion, this zone of weakness appears to have received the Havnefjord diorite magma (Fig. 14C).

Since the space for the magma was created by tectonic forces during the development of a fold, it is clear that the magma did not have to force an entry, and thus there is no evidence to this effect in its envelope. The emplacement of the Havnefjord diorite provides an excellent example of a truly permissive intrusion as envisaged by Pirsson (1914); it rose into the cavities and zones of weakness formed in the Stromatolith by tectonic forces. The larger rafts probably represent relict metasedimentary screens of the original stromatolith, and may well be virtually undisturbed.

Harker (1909) introduced the term phacolith for a concordant intrusive body introduced concurrently with folding. This is a general term, and the habit, magnitude, and form of the phacolith is determined by circumstances of the folding and mechanism of intrusion. In general it has been emplaced into the hinge region of a developing fold, and has a meniscoid or lenticular (concavo-convex) form in cross-section.

The Havnefjord diorite has a concavo-convex cross-section, and it is a concordant body which was emplaced syntectonically. It fulfills the conditions required by Harker's phacolith concept, and it is proposed to classify the diorite body as a phacolith.

A particular type of phacolithic body which is sickle-shaped and which was emplaced contemporaneously with the formation of cross-folds in steeply folded rocks has been named a harpolith by Cloos. However, a harpolith is an essentially discordant body of sickle shape, resembling in form, though not in structural relations, a tilted phacolith. It has been emplaced into previously deformed rocks and has then been stretched together with its host horizontally in the direction of maximum orogenic displacement (Daly, 1933). Thus the Havnefjord diorite is not a true harpolith, and it is maintained that a steeply plunging phacolith is the most appropriate term to describe it.

The mode of emplacement of the Havnefjord diorite and its resulting structure is of importance in the consideration of the emplacement of the late diorites, particularly the pyroxene-mica diorites.

It has been remarked that these coarse-grained late diorites are closely associated in the field with the metasedimentary rafts. It is clear that in the model of emplacement proposed for the Havnefjord diorite, the planes parallel to the banding of the metasediments in the laminated stromatolith will always be planes of weakness, especially if slight deformation is still continuing after the emplacement of the Havnefjord diorite. There will still be a tendency for elongate cavities to form in the neighbourhood of the large metasedimentary screens or rafts, in particular along their boundaries. It is at just such localities that the late diorites are most commonly found. They have manifestly risen alongside the metasedimentary rafts, the magma flowing into cavities created for it by tectonic forces. In many cases the magma has picked up blocks of metasediment on the way, as the abundance of xenoliths shows. The large, well-preserved xenoliths appear not to have been carried far, but the small, highly-digested inclusions were probably brought up from deep levels, where the chemical energy would have been higher.

Thus it is proposed that the late diorites are also examples of permissive intrusion, emplaced synchronous with late F_2 movements.

Vatna Gabbro

This gabbro is probably a sheet-like body dipping steeply towards the south, although its true form is not known in detail. Its emplacement was probably complex, a fact suggested by the presence of fine-grained autoliths.

The hypersthene gabbro of Ardnamurchan contains long, narrow, parallel-sided xenoliths which are an early chilled facies of the gabbro (Wells, 1953). These xenoliths have been metamorphosed, and there is a gradation from original gabbroic textures to entirely granulitized textures in a partially recrystallized xenolith.

In the Freetown layered basic complex of Sierra Leone there are some pyroxene granulite xenoliths, at least some of which are believed to be cognate, having been formed by granulation and recrystallization of the normal gabbroic and troctolitic host rock in situ (Wells, 1962). Wells considers that this supports a hypothesis of multiple intrusion for the complex. Thus in these examples the inclusions are considered to be of the same rock-type as the host, which has been altered by autometamorphism.

The autoliths of the Vatna gabbro may have a similar origin. However, it is difficult to be sure from the texture of the Vatna autoliths whether or not they have been metamorphosed by the surrounding gabbro, for the latter has an open, somewhat granular texture, similar to that of the inclusions. As far as can be ascertained, in spite of the smallness of grain-size, the mineralogy of the autoliths is the same as that of the host, and hence appears not to have been significantly metamorphosed. Furthermore, there are none which are obviously only partially recrystallized.

When a body of magma is emplaced into its host, it will in general begin to solidify first at its margins, particularly its roof. As crystallization proceeds, marginal parts of the body might be solid while central parts are still molten. If crystallization is allowed to be completed undisturbed, a comparatively homogeneous rock should result. However, if at a stage where only partial crystallization has taken place, there is a resurgence of the magma from below, then it is possible for autobrecciation to occur. The second phase of magma intrusion might rupture the early-formed marginal facies so that by the time it reaches its ultimate position of emplacement it contains numerous autoliths of the early facies.

It is suggested that these autoliths are relics of an early, fine-grained, probably marginal, facies of the gabbro which have become autobrecciated and engulfed in a later facies. They are possibly similar in origin to the fine-grained facies inclusions in the hypersthene gabbro of Ardnamurchan, but in this case there appears to be no marked autometamorphic effect. The process envisaged is quite likely to have occurred since the contact of the body is near to the north, where a fine-grained marginal facies might have formed.

Thus it would appear that the emplacement of the Vatna gabbro was not a simple event but consisted of at least two phases, probably following one another quite quickly. The final phase of emplacement was probably accompanied by the emplacement of the perthosites, which became locally mixed with the gabbroic magma. This further complication is discussed in association with the emplacement of the perthosites (p.141).

THE MINOR INTRUSIONS

Introduction

The Husfjord igneous complex contains a number of minor intrusions. These are a comparatively late phase in the history of the emplacement of the complex, and are emplaced into the major bodies that have already been described.

They include basic dykes, perthosite sheets, and nepheline-syenite pegmatites. There are at least two phases of dyke emplacement, one pre-dating the perthosites, and one post-dating them. The minor intrusions will be described in the order in which they were emplaced.

EARLY BASIC DYKES

The early basic dykes are those which pre-date the perthosites, and the evidence for their pre-perthosite age is the fact that they occur as xenolithic inclusions within some of the perthosite sheets. The reason for considering these xenoliths as dyke material is two-fold. First, the morphology of the xenoliths suggests that they were present as dykes in the host rock before the emplacement of the perthosites (see under Perthosites, p.135). Second, similar rock types occur as dykes in the host rock outside the perthosites.

The most common of these early dykes are porphyritic amphibolites. These have numerous white-weathering, subhedral to anhedral feldspar phenocrysts, which give them a distinctive white spotted appearance.

Their occurrence in the Husfjord metagabbro just outside the perthosite sheets confirms their dyke character. They post-date the shearing of the metagabbro since they are not foliated like the metagabbro in which they occur. Relict contacts between these dykes and the metagabbro have been preserved within the perthosites (see under Perthosites, p.140). A similar porphyritic dyke rock occurs in the northernmost gabbro sheet at Slåtten, and here again, both the gabbro and the porphyritic dykes are intruded by perthosite, making their relationships obscure.

Another type of early dyke is a fairly coarse peridotite. This forms a brown-weathering, knobbly rock and occurs as an inclusion within perthosite. It is different from the ultrabasic lenses in the Husfjord metagabbro, which were of troctolite. The peridotite preceded the porphyritic basic dykes since it is cut by one of these dykes. This contact is preserved as a relic in an inclusion within a perthosite sheet.

In the Komagfjord area there are also a number of dyke-like bodies of a grey pyroxene-hornblende diorite which occur both outside and within the perthosites. They post-date the Komagfjord diorite since they cut the foliation in the latter.

There is evidence that there were also pre-perthosite fine-grained basic dykes. This evidence is provided by xenoliths of fine-grained amphibolite in an outcrop of Komagfjord diorite, which in turn forms a raft within perthosite.

Further discussion on the morphology of the early dyke xenoliths appears in the next section.

PERTHOSITES

The perthosites consist almost entirely of hair-perthite, and weather to a pink or yellowish colour, and grain-size is variable from fine to medium. They occur in two areas; one is the southern shear-belt in the Husfjord metagabbro, and the other is in the Vatndal area north of Vatna.

The former area runs from the east side of Komagfjord, over Komagakslen to the west side of Husfjord. In this belt, the perthosites form sheets of greatly varying dimensions. In the coastal regions, they are a few metres or tens of metres in width, although in some places they tongue out into narrow irregular veins (Plate 77). On the top of Komagakslen, there is a broad perthosite which appears to branch into numerous separate sheets towards the south-west; this region is obscured by extensive mountain-top screes of perthosite blocks. Among these blocks there are also some quartzitic metasedimentary blocks, which may represent xenolithic material in the perthosites.

Although perthosites are fairly common in the shear-belt, which must have provided planes of easy entry for the perthosites, they are not confined to this zone, but also occur just to the south-east of the shear-belt. These perthosites frequently contain relics of Husfjord metagabbro, Komagfjord diorite, and pre-perthosite dykes, and these will be described on p.135 .

The second area in which perthosites occur is the Vatndal area. Here, the perthosites form sheets up to several tens of metres in

Plate 77. Thin perthosite veins in Husfjord metagabbro. W. Husfjord.

width trending in a N.W.-S.E. direction. These sheets cut both the Husfjord metagabbro and the Vatna gabbro. Some sheets cross from the former into the latter, and there appears to be a general tendency for them to splay out and branch when they occur in the Vatna gabbro.

In many cases, the ends of the perthosites in the Vatna gabbro are diffuse and merge with pink-weathering, antiperthite-bearing gabbro. Study of thin sections shows that there is a gradation from olivine-gabbro through antiperthitic olivine-leucogabbro and melaperthosite to p ractically pure perthosite. It is clear that in places the perthosite and olivine-gabbro magmas must have become mixed, particularly around the perthosite intrusions. An explanation for this is considered on p. 141 , where the emplacement of the perthosites is discussed.

The Vatna gabbro has been intruded by a number of basic dykes, and these also cut the Vatna perthosites. Hence there was a phase of dyke emplacement post- dating the perthosites.

Xenoliths in Perthosites

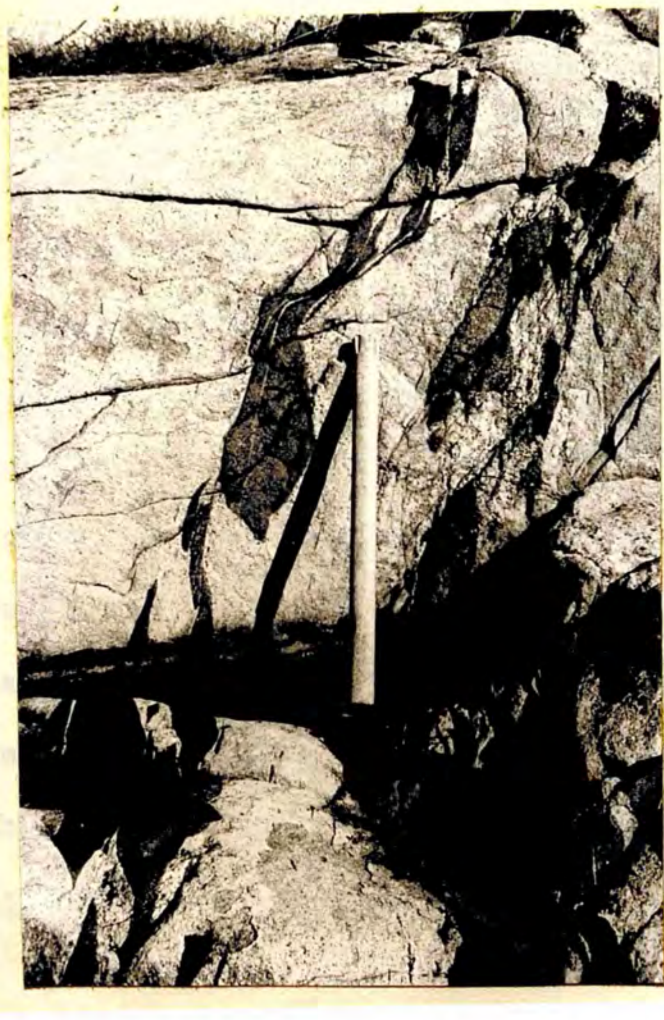
The perthosites which occur in and near the Husfjord metagabbro shear-belt commonly contain xenolithic inclusions of earlier rocks. These are particularly well displayed on a small peninsula in Komagfjord, where there are complex relationships between the perthosites and the inclusions.

Here, perthosite forms branching sheets, several tens of metres wide, which trend at about 050°- 070°. It is pinkish-yellow in colour and generally fairly fine-grained, and in one place the margin is very fine-grained and appears to be chilled against the host-rock. The mafic content of the perthosite is fairly variable; in general it is almost entirely free from mafic minerals, but in places, generally near the margins, mafics are conspicuous. In these parts where the mafics are abundant, there is often a preferred orientation of the mafic minerals parallel to the sheet margins, and this appears to be a flow orientation. The xenoliths include five rock-types; Husfjord metagabbro, porphyritic basic dyke, peridotite, Komagfjord diorite, and basic diorite.

In the central parts of the perthosites, the xenoliths are not large, ranging in size from a metre or so down to a few millimetres in length, Here, xenoliths of different rock types are juxtaposed in a haphazard way (Plate 78). They are usually platy in form, and are orientated parallel to the perthosite margins and the flow lamination, where this latter is evident. They are generally somewhat angular, and at their margins thin veins of perthosite frequently penetrate into them, wedging fragments off the edges of the blocks (Plate 78). The shapes of the blocks appear to depend to a large extent upon the rock types. The Husfjord metagabbro of the xenoliths is commonly foliated, and as a result the xenoliths tend to be long and thin, elongated parallel to the foliation. The peridotite and the porphyritic basic dykes are not foliated, and these tend to form broader, more angular blocks (Plate 78). However, occasionally the porphyritic basic dyke material does form long irregular xenoliths which in places have been split into slivers by the penetrating perthosite (Plate 79).

Plate 78. Mixed xenoliths in perthosite; dark block is peridotite, spotted blocks are porphyritic dyke, plain blocks are Husfjord metagabbro. Komagfjord.

Plate 79. Porphyritic dyke xenolith split by perthosite veins. Komagfjord.



The larger blocks have sharply-defined margins, but this sharpness often diminished as the size of the xenoliths decreases, and this is particularly evident in the case of the metagabbro xenoliths. The smaller ones generally have ragged edges, often rather diffuse, and appear to be becoming assimilated by the perthosite (Plate 78).

In the neighbourhood of the xenoliths, small slivers and clots of basic rock, apparently metagabbro, often occur streaked out parallel to the direction of elongation of the xenoliths (Plates 78 and 79). Sometimes the margins of these small inclusions are sharp, and sometimes they are diffuse. Occasionally these slivers also occur away from the xenoliths.

The perthosite adjacent to the xenoliths frequently appears contaminated (Plate 78), and this may be due to the assimilation of basic material by the perthosite. Diffuse contaminated patches also sometimes occur in the perthosites away from the xenoliths.

Towards the margins of the perthosites, the xenoliths become larger and more coherent. In the larger inclusions relict contacts between the various rock types comprising the xenoliths occur, showing their pre-perthosite relationships.

At the eastern margin of the Komagfjord perthosite, there is a raft of Husfjord metagabbro which has been penetrated and broken up by invading perthosite (Plate 80). The end of the raft is irregular and ragged, and there are numerous detached slivers, some partially assimilated, swimming in the perthosite at the margins. The raft contact is occasionally slightly diffuse, and the neighbouring perthosite

Plate 80. Raft of Husfjord metagabbro in perthosite; note the ragged ends and the neighbouring slivers of metagabbro. Komagfjord.

is contaminated. In places former extension of the raft can be inferred by the areas of contamination in the perthosite. This is well shown in Plate 80; in the centre there is a broad tongue of metagabbro whose left-hand contact has a large embayment in it. However, a former position of this contact is indicated by the extent of the lozenge-shaped dark contaminated perthosite occupying the embayment.

This eastern margin of the perthosite trends at about 055° and its mafic minerals are parallel to this direction. The trend of the metagabbro raft, however, is 030°, and thus the raft is discordant with the flow structure. The foliation has not been traced right up to the metagabbro contact, for the raft is bordered by a zone of contaminated perthosite, with metagabbro slivers, in which the flow structure is not displayed. But it is clear that the raft has not been orientated parallel to the flow, and it has probably not been detached from the host rocks.

Towards the western margin of the perthosite, the metagabbro inclusions become more numerous and much larger, increasing in size as the margin is approached, until there are large rafts of metagabbro in the perthosite. Here the metagabbro is markedly foliated, the foliation striking at 030° and dipping at about 50° to the west. The entry of the perthosite here has clearly been guided by this foliation and this western margin is sub-parallel to it. However, just outside the perthosite the metagabbro is not foliated, and here it is net-veined by perthosite, which is white and slightly coarser-grained than the normal perthosite (Plate 81).

Plate 81. Perthosite net-veining metagabbro just outside perthosite sheet. Komagfjord.

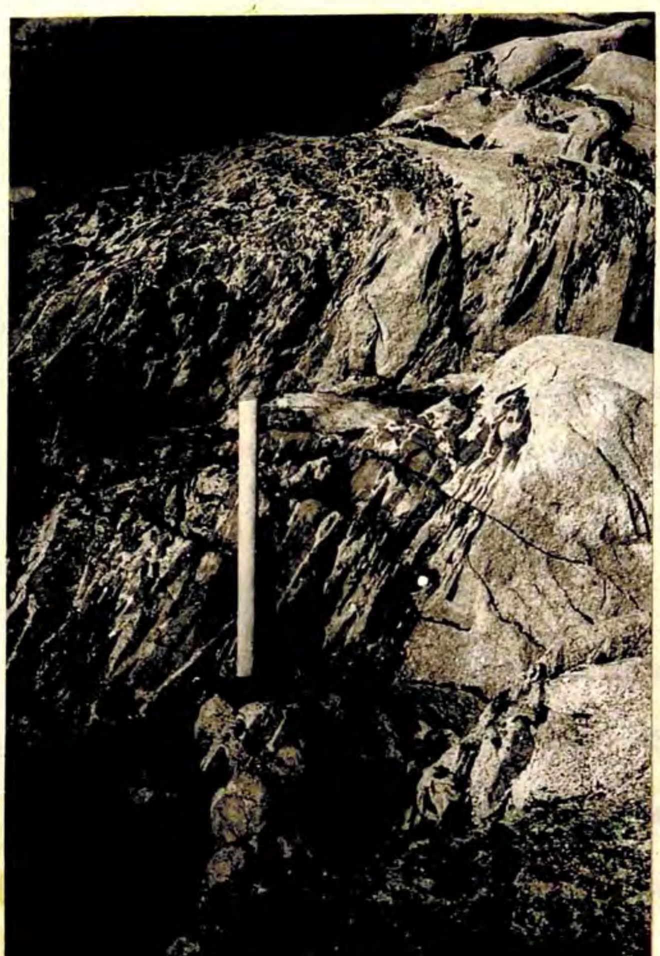


Plate 82. Raft and relict pseudo-dykes of Husfjord metagabbro in perthosite. Komagfjord.

These rafts are generally long and narrow, and frequently occur in trains elongated parallel to the foliation. They are often extensively veined by the perthosite, which penetrates along the foliation planes (Plate 82). As the perthosite penetrates into the cracks it generally becomes coarser-grained, and completely free from mafic minerals.

The morphology of the metagabbro inclusions often causes them to resemble relict dykes which have sometimes become fragmented by the perthosite. Since they consist of Husfjord metagabbro, the host, they cannot be dyke material, and may be termed relict pseudo-dykes (Goodspeed, 1955). Miller (1945) has described some long narrow xenoliths which may be mistaken for dykes, and these pseudo-dykes are also sometimes "cut to pieces" by the host.

Near the perthosite margins the porphyritic basic dyke xenoliths become larger and approach their pre-perthosite dyke form (Plate 83). But even where a fairly coherent dyke occurs, it is invariably fragmented into angular blocks by penetrating veins of perthosite (Plate 83). These dykes are not foliated, so the perthosite veins are irregular since they cannot follow any foliation planes. Further discussion on this veining of the inclusions appears when the emplacement of the perthosite is considered (p. 143).

These porphyritic basic dyke inclusions are known to be of dyke material since a similar rock-type occurs as dykes in the Husfjord metagabbro just outside the perthosite, and furthermore, their trend is subparallel to the dykes outside the perthosite. Their present

Plate 83. Relict porphyritic dyke in perthosite; note disruption of dyke by dilatational mechanism. Komagfjord.

dyke-like character must be a reflection of their original dyke form, and they may be termed relict dykes. The orientation of these relict dykes is generally discordant with both the perthosite margin and the foliation direction of the metagabbro. Presumably they have not been disorientated from their original positions as dykes cutting across the foliation in the metagabbro; this cross-cutting relationship is shown by a dyke outside the perthosite. Slivers of metagabbro sometimes occur near the relict dykes, and these remain parallel to the foliation direction of the metagabbro.

The margins of the porphyritic relict dykes and the porphyritic dyke xenoliths are always sharp; there is no evidence of them becoming assimilated by the perthosite as in the case of the metagabbro inclusions. It is evident that they have been more resistant to attack by the perthosite than the metagabbro, and this may be due to their non-foliated character.

In some of the large rafts the pre-perthosite relationship between the different types of inclusions can occasionally be seen. In one case a veined raft consists of alternating bands of metagabbro and spotted dykes. Although there is a slight mixing of the rock-types along the relict contacts, the general alternation can clearly be discerned (Plate 84). In another case, there is a sharp contact between porphyritic dyke rock and peridotite in a raft (Plate 85). The porphyritic rock forms a dyke cutting the peridotite, and must therefore post-date the latter. The dyke contains inclusions of the peridotite, and both rock types are veined by perthosite in an irregular fashion.

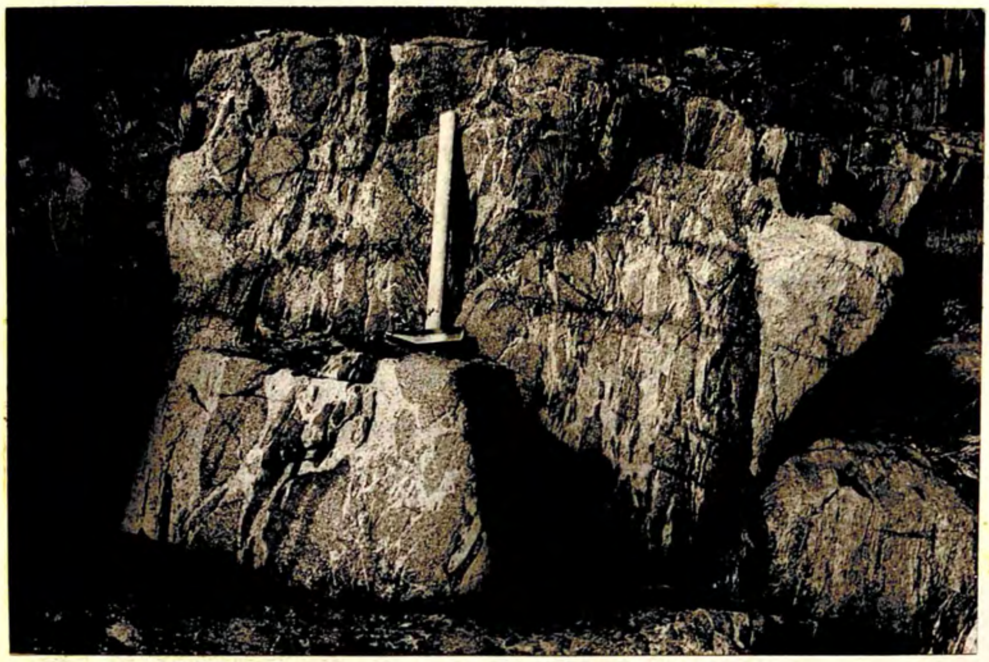


Plate 84. Relict contacts between porphyritic dyke and darker Husfjord metagabbro, veined by perthosite. Komagfjord.

Plate 85. Relict contact between porphyritic dyke and dark peridotite, veined by perthosite. Komagfjord.

Plate 86. Leucocratic reaction zones between peridotite and perthosite veins. Komagfjord.

Where these veins penetrate the peridotite, they are generally flanked by a diffuse leucocratic to mesocratic reaction rim against the melanocratic peridotite (Plate 86).

In one composite raft there is Komagfjord diorite, containing xenoliths of metagabbro, porphyritic basic dyke, and metasediment. This xenolithic diorite is cut by a grey basic diorite; the latter cuts across the foliation in the xenolithic diorite, and is apparently dyke-like in form.

It is important to note that the dykes cutting these rafts do not penetrate into the perthosite, but are confined to the rafts, clearly representing pre-perthosite contacts.

Emplacement of Perthosites

Perthosites occur in two areas; one is in the Vatndal area, principally within the Vatna gabbro, and the other is in the southern shear-belt within the Husfjord metagabbro, particularly well displayed in Komagfjord.

It would appear that these two groups were emplaced in very different environments, and as a result they have rather different characteristics. For this reason, discussion on their emplacement will be in two sections, each dealing with one of the groups.

Vatndal Perthosites

These have somewhat sinuous outcrops, particularly within the

Vatna Gabbro. As they cross from Husfjord metagabbro into Vatna gabbro they also tend to splay out and branch, and have irregular widths.

An important feature of these perthosites is that both in the field and in thin-section they are often seen to have diffuse margins merging into antiperthite-bearing gabbro. It seems probable that the perthosite existed as a magma which became mixed with the olivine-gabbro magma at the time of its emplacement. A possible explanation for this is that the perthosites were emplaced before the Vatna gabbro had become completely solid, so that those parts of the perthosite sheets within the Vatna gabbro became mixed with the neighbouring crystallizing host. The way in which the perthosite sheets splay out and have irregular forms within the Vatna gabbro may be a result of this. If the gabbro was not solid enough to sustain brittle jointing, but was still slightly mobile, then any magma injected into it would probably form irregularly-shaped bodies.

Pitcher and Read (1960) have described some dykes in the Donegal Granite which are discordant with the flow structure of the granite. These dykes have irregular margins, vary in width, and are broken or deformed in places. Their contacts are generally well-defined, but occasionally there is a mixing of granite and dyke material, and a merging of textures. They are considered by Pitcher and Read to have been emplaced into an embryonic joint system in the granite before the latter was entirely solid.

These Donegal dykes have many structural features similar to the

Vatna perthosite, and as indicated, a similar mode of emplacement is envisaged for the latter.

Komagfjord Perthosites

It is assumed that these were emplaced at much the same time as the Vatndal perthosites, but because of their different environment of emplacement, they have different features. They were emplaced into cold Husfjord metagabbro, and thus tend to be finer-grained than the Vatndal perthosites. They do not have diffuse margins merging with the host, but contain numerous xenolithic inclusions of the host and some pre-perthosite minor intrusions. The morphology of the xenolithic relict dykes and relict pseudodykes provide a clue to the emplacement mechanism of these perthosites.

Goodspeed (1955) has described some relict dykes and relict pseudodykes in a granodiorite at Cornucopia, Oregon. The relict pseudodykes are remains of the host-rock, and are elongate, resembling dykes. The relationships between the granodiorite and the inclusions are similar in many respects to those between the Komagfjord perthosites and their inclusions. The Cornucopia relict dykes are cut by veins of granodiorite; they are embayed by the granodiorite; in places there are gradational contacts; dykes have a granoblastic texture; protruberances of dyke penetrate into the granodiorite; the granodiorite sometimes has feldspathic borders against the dykes; the granodiorite surrounding the dykes contain hazy patches or small skialiths which may have been derived from the dykes. All these features are present at Komagfjord with the exception of granoblastic textures in the dykes.

Goodspeed believes that the granodiorite is a product of granitization, and that the relict dykes and pseudodykes are remnants of the rocks which were being granitized. He considers that the relationships between the granodiorite and its inclusions provide evidence for partial replacement of the dykes and their pre-granodiorite host.

His explanation for the survival of fragments of the schistose host as pseudodykes is that during metamorphism certain bands in the schists crystallized to a harder rock than others. When the whole area was granitized, these bands were more resistant to granitization than the rest of the schist and were thus preserved as pseudodykes. It is pertinent to note that in this example, the host rock was schistose; it will be recalled that the host rock at Komagfjord, the Husfjord metagabbro, was also extensively foliated.

In the opinion of the present writer, the evidence for Goodspeed's interpretation of the relationships between the dykes and their host is not unequivocable. The Cornucopia granodiorite may indeed be a product of granitization, but the relict dyke relationships described could equally well occur in a magmatic host, for example the diffuse contacts can be due to progressive assimilation by a magmatic host.

Some relict dykes in the Ilordleq area of southern Greenland have been described by Watterson (1965). These dykes occur in a hornblende granite, which in places veins into the relict dykes causing their fragmentation. In most cases where the relict dykes are deformed and disrupted they are veined by an aplitic granite, which Watterson considers to be metasomatically replacing the dykes, leaving enclaves as relict dykes. In the case of ultrabasic dykes, the feldspathic veining is thought to be a product of metamorphic differentiation of a pre-existing less ultrabasic dyke, although Watterson admits that some of the feldspathic material may have been derived from outside. In some disrupted dykes, considerable amounts of relative movement have occurred between fragments. Clearly there has been movement in the surrounding granite, although there is no evidence of deformation in the granite. Watterson explains this by considering that the granite deformed in a solid or highly viscous state, and then underwent a thorough recrystallization which eliminated all traces of movement. He does not believe that the granite was ever truly liquid.

Many features described by Watterson for the Ilordleq relict dykes are present in the Komagfjord relict dykes and pseudodykes. These include the veining of the more coherent relics by material generally more leucocratic than the bulk of the invading material; the sharp margins of the majority of blocks; the diffuse contacts of some of the blocks; streaked out slivers and schlieren of basic material in the granite between some of the dykes; relict contacts between dykes of different ages.

Watterson believes that the dykes post-date the granite, but that the latter had become migmatized and completely recrystallized after the dyke emplacement, leaving only remnants of the dykes. This may be the case at Ilordleq, but again, the relationships described are not only confined to migmatitic environments.

An idea similar to that of Watterson is invoked by Roddick and Armstrong (1959) for the origin of relict dykes in the Coast Range Mountains near Vancouver. Here, it is considered that the dykes were intruded into rocks which were undergoing granitization under plutonic conditions. Although the dykes are disrupted, there is no sign of fracturing in the enclosing rock. It is thought by the authors that the host was brittle enough to allow fractures to form, up which the dykes came, but recrystallization continued to heal the fractures in the host rock.

Relict dykes which probably closely resemble the Komagfjord examples in mode of origin occur in the Coast Range Batholith near Vancouver, and have been described by Phemister (1945). Here, a granite pegmatite facies of the batholith contains pre-batholithic dykes which have for the most part remained intact during invasion by the granitic magma. The inclusions are penetrated by stringers of the pegmatite, and in some cases relict contacts between dyke and schistose pre-pegmatite host are preserved. The dykes are rarely deformed, and in thin section appear to be little altered. Phemister considers that the dykes must have acted as barriers to the incoming magma, and occasionally protected the neighbouring host rock which is sometimes preserved next to the relict dykes. Most of the host rock has been selectively removed piecemeal, and the dykes have virtually maintained their original positions and directions.

It is important to note that the host rock in this example was schistose, as it is at Komagfjord, and that this was probably an an important factor in determining the ease with which it was removed by the invading magma. The dykes were non-schistose and were thus probably more resistant to piecemeal removal.

It has been noted that the relict dykes and pseudodykes described in the various examples mentioned resemble in many ways those occurring in Komagfjord. In some cases the host was regarded as migmatitic, and in others it was thought to be truly magmatic. However, in all cases it was thought that the host must have been mobile after the formation of the xenolithic inclusions. Thus in the Komagfjord area it is important to look for evidence which might make it possible to determine whether or not the perthosite was a true magmatic liquid.

There are three lines of evidence: the texture of the perthosite itself, the structure of the relict dykes and pseudodykes, and the grain-size of the perthosites.

First, it will be recalled that the perthosite has a foliation formed by the alignment of mafic minerals, especially at its margins. In the central parts of the perthosite sheet, these are parallel to the assorted xenoliths and basic slivers which are clearly free-swimming. Xenoliths and clots are commonly aligned parallel to the foliation planes of igneous rocks. It cannot be assumed that inclusions are rotated into parallelism after consolidation of the host, but only before or during crystallization; thus this foliation must be a primary flow banding (Balk, 1937). Thus the perthosites appear to have a flow structure common in many normal igneous bodies.

Second, where a relict dyke still forms a fairly coherent body, it can be clearly seen that the veins which penetrate the dyke are of the perthosite itself. Furthermore, these veins have disrupted the dyke by a dilatational mechanism. Blocks can be seen to be becoming wedged off the main body by the penetrating veins (Plate 83). In some cases cavities are formed behind the dislodged blocks in which smaller broken fragments are beginning to be assimilated by the perthosite. Inclusions of foliated metagabbro have also been disrupted by a dilatational mechanism, in which the perthosite has streaked through the inclusion along the foliation planes, offsetting it in a direction perpendicular to the foliation. Sometimes fragments of metagabbro remain in the zone of dislocation (Plate 87). Clearly the perthosite must have been in the form of a magma in order to produce these structures.

Third, it has been noted that where the perthosite has entered into cavities and narrow veins, it is almost invariably slightly coarser-grained than the nearby perthosite. A good example is shown in Plate 88, in which the perthosite is markedly coarser-grained, almost pegmatitic, where it is trapped in a cavity between two blocks of porphyritic dyke rock. This probably indicates that the perthosite was fairly rich in volatiles which tended to become trapped in the veins and cavities entered by the perthosite. This would favour the formation of coarser-grained crystals in these areas where the volatiles were trapped.

This demonstration of a high volatile content in the perthosite magma is significant, since the presence of volatiles was probably an

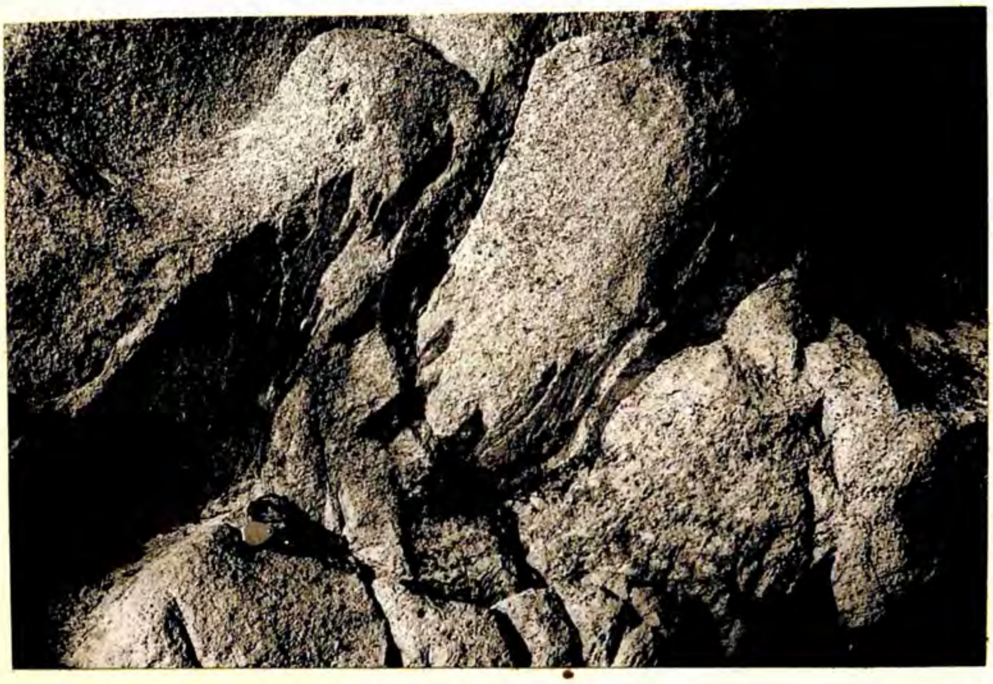


Plate 87. Husfjord metagabbro xenolith disrupted by perthosite along foliation. Komagfjord.

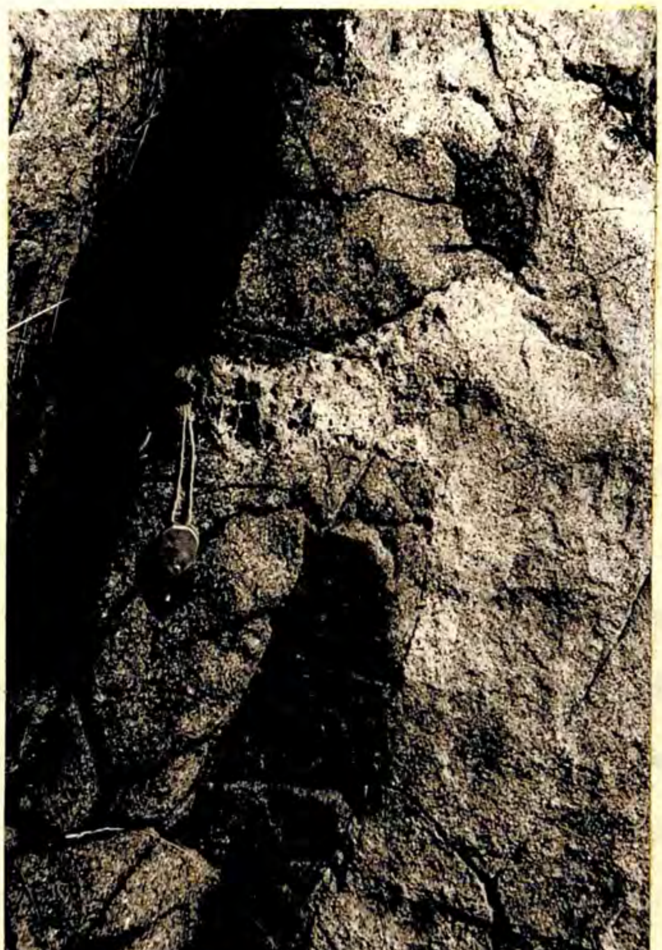


Plate 88. Coarser perthosite in cavity between xenolithic blocks. Komagfjord.

important physical factor in the mechanism causing the disruption of the relict dykes.

The viscosity of a magma depends upon a large number of factors which include its composition, temperature, pressure, and volatile content (Turner and Verhoogen, 1960); thus it is difficult to say how viscous the perthosite magma might have been. Walker and Skelhorn (1966) in a review on net-veined complexes consider that the acid component which veins the basic material must have had a low viscosity, since thin persistent acid veins can often be traced for long distances. They postulate a viscosity of about $10^5 - 10^6$ poises for the acid material, whereas experimental work on acid lavas has given viscosities as high as 10^{10} poises (Friedman, Long, and Smith, 1963). However, the viscosity of a magma is inversely proportional to its volatile content (Shaw, 1965), and the presence of 4% by weight of dissolved water in an acid glass can reduce the viscosity by a factor of about 10^{-3} (Friedman, Long, and Smith, op.cit.; Shaw, 1963). Thus Walker and Skelhorn (op.cit.) believe that the apparently low viscosity of the acid magma in the veins can be explained mainly by a high volatile content of the magma. In the Komagfjord perthosites it appears that the magma had a fairly low viscosity since it veins into long, narrow cracks. However, it has been demonstrated above that it was probably rich in volatiles, and this could account for its comparatively low viscosity.

LATE BASIC DYKES

The Vatna gabbro is intruded by a large number of basic dykes of varying size. Many cut the Vatndal perthosites (Plate 89) and antiperthite-rich Vatna gabbro, and are clearly post-perthosite in age. Those dykes which are not demonstrably post-perthositic are considered to be members of this late suite of dykes because of their similarity in mineralogy and metamorphic history, and their parallel trend to those which are known to post-date the perthosites.

The earliest of these minor intrusions is an olivine-leucogabbro in which the weathered plagioclase grains give the rock a distinctive white spotted appearance. It occasionally contains xenoliths of a mixture of melanocratic rocks (Plate 90). None of these xenoliths appears to be of Vatna gabbro, and they may be fragments of earlier dykes brought up from below.

This leucogabbro forms dykes generally varying greatly in width; some are about 1 m. wide whereas others are several tens of metres in width. The small island of Vatnholm off the coast at Vatna is essentially composed of this white spotted gabbro. In places it sends apophyses into the neighbouring Vatna gabbro.

One of the later dykes which cuts the leucogabbro is a porphyritic amphibolized dolerite, which generally has diffuse contacts with its host. Within the host near the contact, there is an irregular zone of coarse segregations of olivine and pyroxene (Plate 91). This zone is irregular in width and distance from the dyke margin, and may be up to 10 cm. from the dyke. On the dyke side of the zone, the segregations are relatively rich in plagioclase. It would appear that some

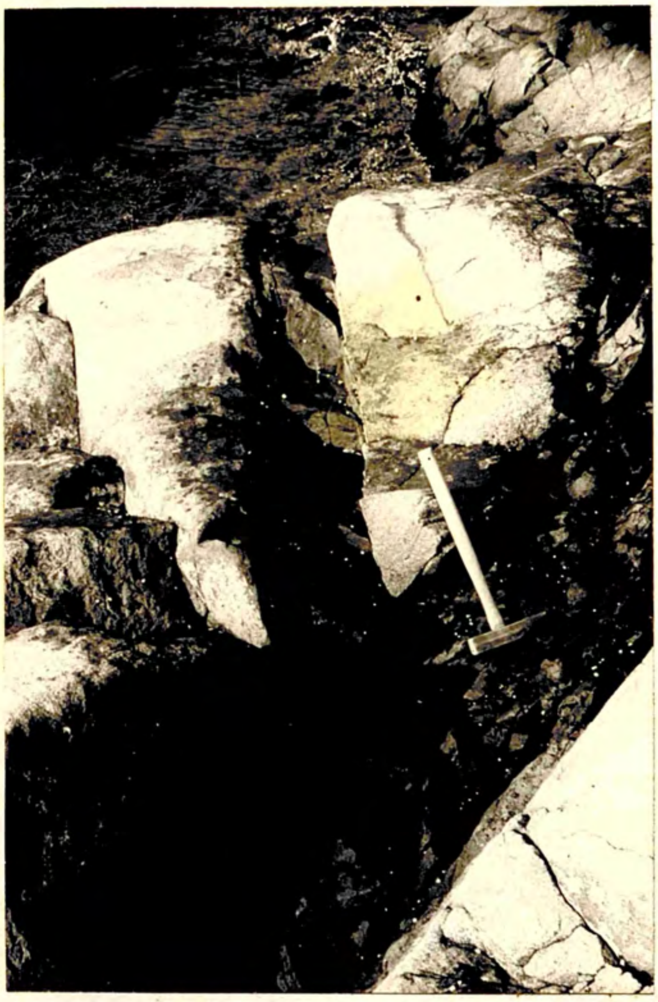


Plate 89. Basic dykes cutting perthosite.
Vatna.

Plate 90. Xenoliths in olivine leucogabbro. Vatna.

Plate 91. Reaction between dyke and its host. Vatnholm.

reaction has taken place between the dyke material and its host, although the rocks here have not undergone a high grade of metamorphism, and none of the other dykes show this relationship. Thus the reaction presumably took place at the time of the dyke's emplacement, and it is possible that this dyke was intruded soon after the emplacement of the leucogabbro, before the latter had completely solidified. A chemical gradient would have been set up between them, and constituents might have migrated across the boundary, forming a reaction zone.

The remainder of the basic dykes are all melanocratic, but they vary somewhat in appearance; some are coarse-grained, and others are fine-grained. Many are porphyritic, containing white plagioclase phenocrysts, while others are non-porphyritic (Plate 92). A number develop pitted surfaces due to the weathering out of olivine phenocrysts, whereas others have smooth surfaces.

Individual dykes maintain a constant thickness, and can often be traced for many tens of metres. The average thickness of the dykes is about $\frac{1}{2}$ m., although they are occasionally as large as 2 m. or as small as 10 cm. Their trends vary between 120° and 170°, and this causes many intersections to take place. This trend of the dykes is subparallel to a prominent joint-direction in the Vatna gabbro. Many dykes are vertical, but others hade at steep angles either towards the west or the east. Occasionally a dyke hades at a more shallow angle as shown by the ultrabasic dyke in Plate 93. The trend of the dykes is roughly perpendicular to the shore-line at Vatna, and they are less resistant to weathering and sea-erosion than the Vatna gabbro. As a

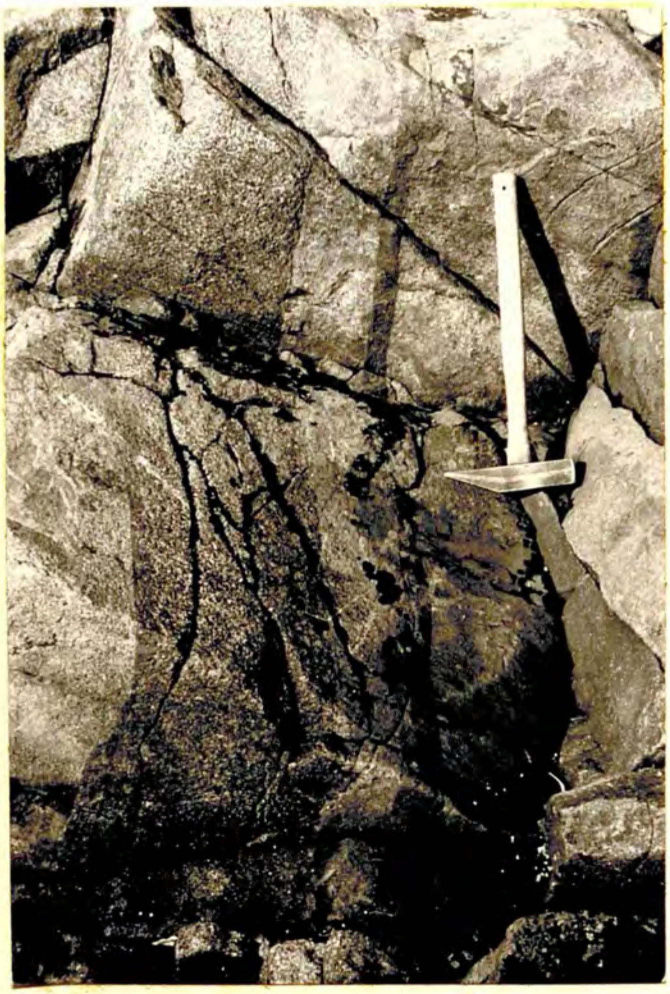


Plate 92. Porphyritic and non-porphyritic basic dykes. Vatna.

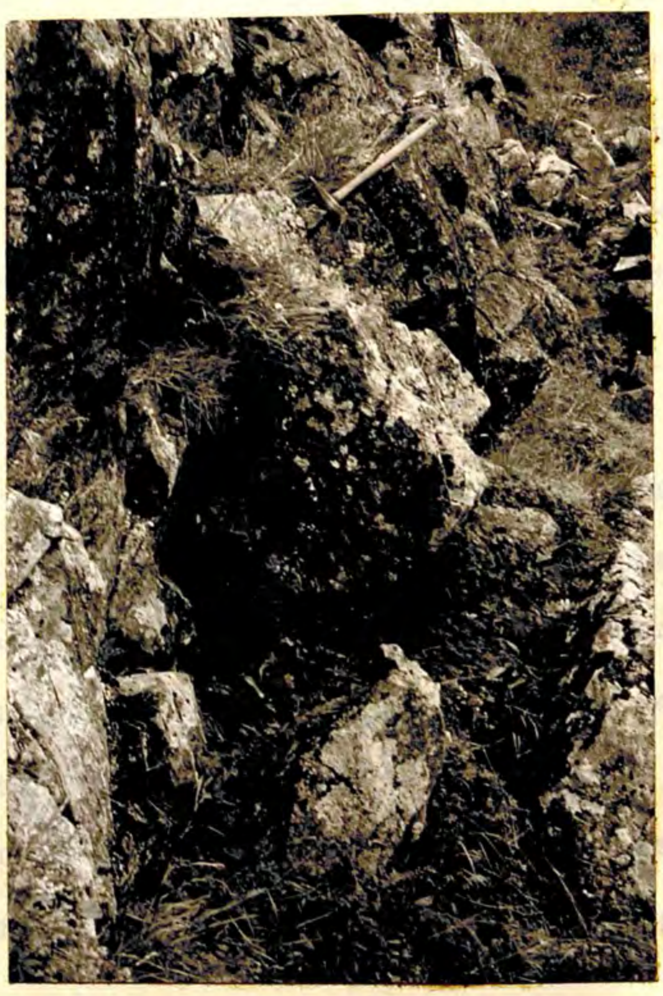


Plate 93. Ultrabasic dyke; contacts are parallel to the hammer shaft. Vatna.

result the dykes have frequently weathered out as subvertical-sided gullies which penetrate into the land from the shore, and which also transect the craggy ground around Vatna.

In many instances it can be seen that the dykes have been emplaced by a dilatational mechanism, shown by the offsetting of earlier dykes or veins. There are numerous examples of chilled margins to the dykes, and these margins are usually about 1 cm. wide, but they have been seen up to 5 cm. in width.

Occasionally a dyke has been emplaced along a shear-plane (Plate 94), but there are also some later small brittle shears which offset some of the latest members of the dyke suite (Plate 95), and which also offset some small micropegmatite veins, which post-date the latest dykes. These small brittle shears occur sporadically throughout the Vatna gabbro, and include both sinistral and dextral types. Displacement along the shears varies from a centimetre or less to over $\frac{1}{2}$ m.

Another late feature which affects the latest dykes is penetration by narrow metasomatic veins of green fibrous amphibole.

In general, the dykes are amphibolized dolerites, some being olivine-dolerites, although there are also occasional troctolite dykes. The porphyritic dykes contain anhedral to subhedral plagioclase phenocrysts, and in the olivine-dolerites the olivines usually form phenocrysts. A few of the fine-grained porphyritic amphibolized olivine-dolerites contain irregular coarse-grained vein-like segregations. These are fairly rich in olivine and amphibole and resemble coarser-grained varieties of the host, and possibly represent volatile-traps in the original magma.

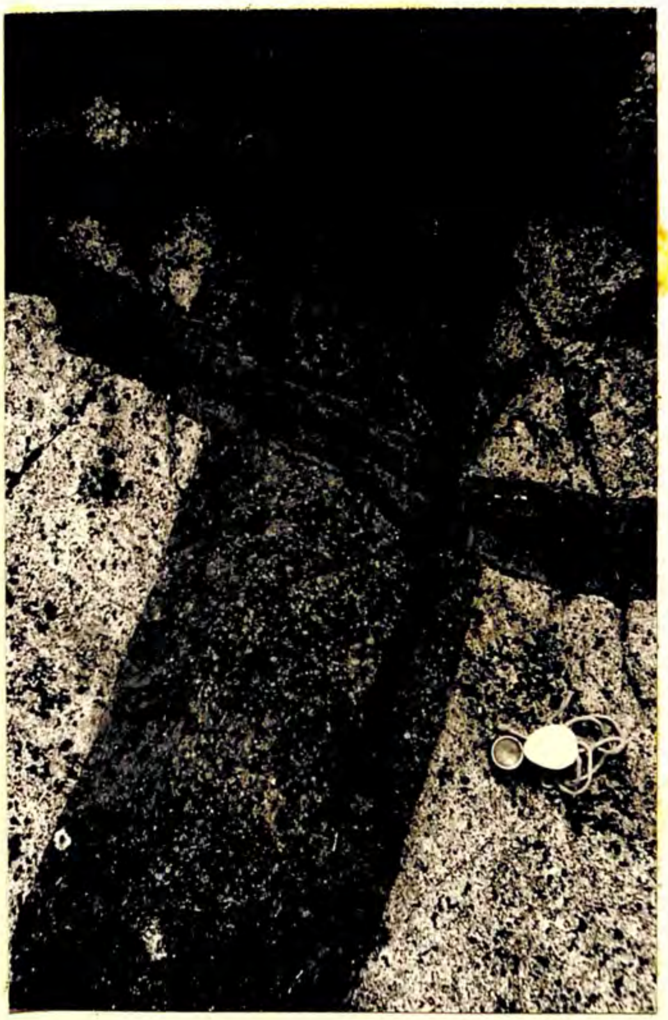


Plate 94. Dyke emplaced along a shear. Vatna.

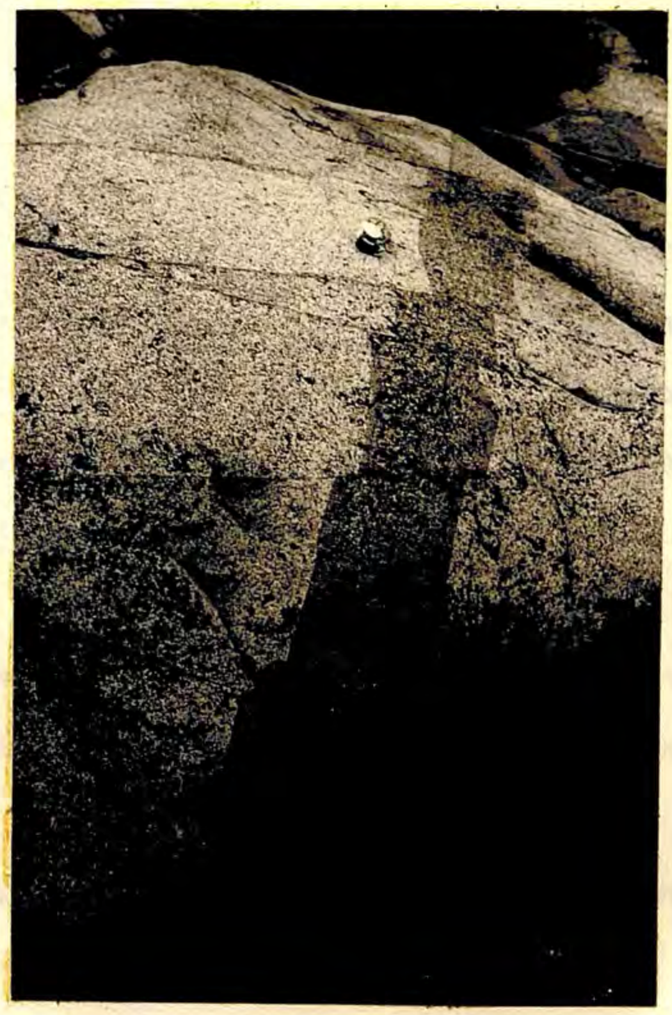


Plate 95. Dyke offset by small dextral shears. Vatna.

By studying the cross-cutting relationships of the various dykes, in conjunction with thin-section studies, it has been possible to build up an overall sequence of emplacement for most of the different dyke types. This sequence is shown in the form of a chart in Fig. 15. Many of the dykes are seen to cut the perthosites since they cut other dykes which in turn cut the perthosites. In those examples where direct field evidence is not available, tie-lines have been omitted from the chart in order to avoid confusion. The fact that at least the majority of dykes are later than the perthosites is indicated by the sequence of tie-lines leading back to the perthosites. The brittle deformation causing the small shears affect the micro-pegmatites, and thus post-date the whole phase of basic dyke emplacement.

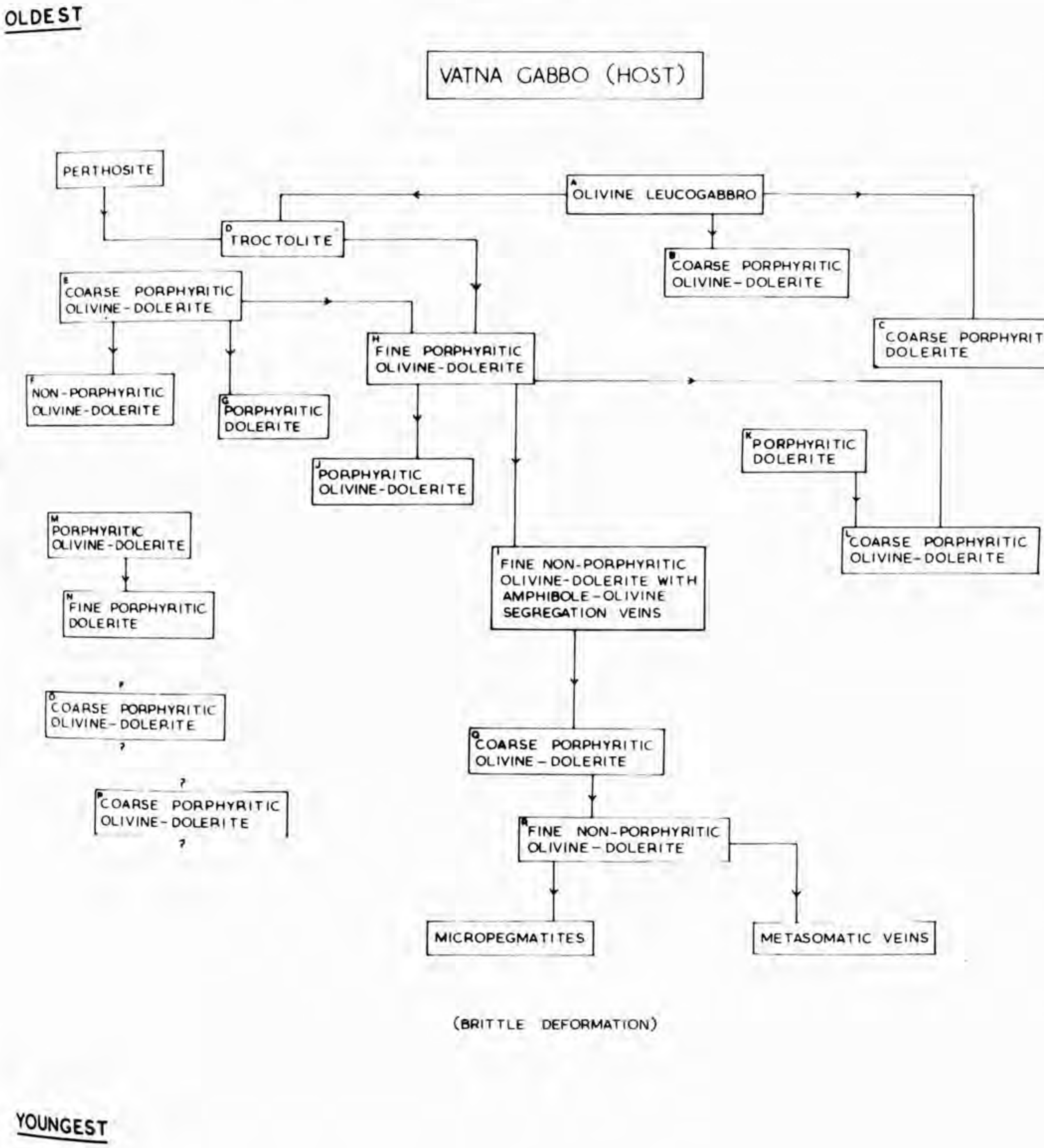
NEPHELINE-SYENITE PEGMATITE

At Vatna there are a few nepheline-syenite pegmatites which have been emplaced into the Vatna gabbro. The nephelines have a grey, pitted, greasy appearance on weathered surfaces. They are invariably sheared, and clearly provided planes of weakness in the massive gabbro along which shearing could take place during the late-stage F_2 deformation.

The pegmatites are fairly thick, the largest being about 2 m. in thickness, and can sometimes be traced intermittently for several tens of metres.

Generally they contain long narrow inclusions of the host rock

SEQUENCE OF DYKE EMPLACEMENT



Connecting tie-lines indicate the presence of field evidence for the age relationships of the dykes, arrows pointing towards the younger dykes. Absence of tie-lines shows that field evidence for their relative ages is lacking. The term 'porphyritic' refers to presence of plagioclase phenocrysts. The letters correspond with those in Table II.

FIG. 15

elongated parallel to the margins of the pegmatites. These mafic inclusions, and much of the host rock next to the sheared pegmatite form a hard, black, fine-grained mylonite (Plate 96).

The shearing of the pegmatites is quite variable, even within a single body. One body is only partially sheared; the central region and part of one margin, together with the neighbouring host is sheared, whereas the rest is unsheared (Plate 97). It can be seen from the unsheared parts that the pegmatite is quite coarse-grained. In other examples, the whole of the body is sheared.

The planes of shearing in these shear-zones in the Vatna gabbro are parallel to those of the shear-zones and mylonites in the Husfjord metagabbro (see Fig. 7), and they are probably related to a regional shear-zone pattern associated with the late brittle deformation. The intersection of the shearing planes with weathered surfaces causes the formation of a marked lineation.

The feldspars and the nephelines become augened during the shearing, but the shapes of the feldspar augen are different from those of the nepheline, presumably due to their difference in competence. The feldspars form broad elliptical-sectioned augen, sometimes quite large (Plate 98), but the nephelines form long, sinuous, streaked-out augen, which in thin-section are seen to consist of partly granulated nepheline.

Augening of nepheline in a nepheline-syenite gneiss during deformation has been described by Sturt (1961). In this gneiss, the nephelines are strongly rodded, parallel to the minor fold-axes, and their c-axes are orientated parallel to this direction. A similar phenomenon has

Plate 96. Mylonite and sheared nepheline-syenite pegmatite. Vatna.

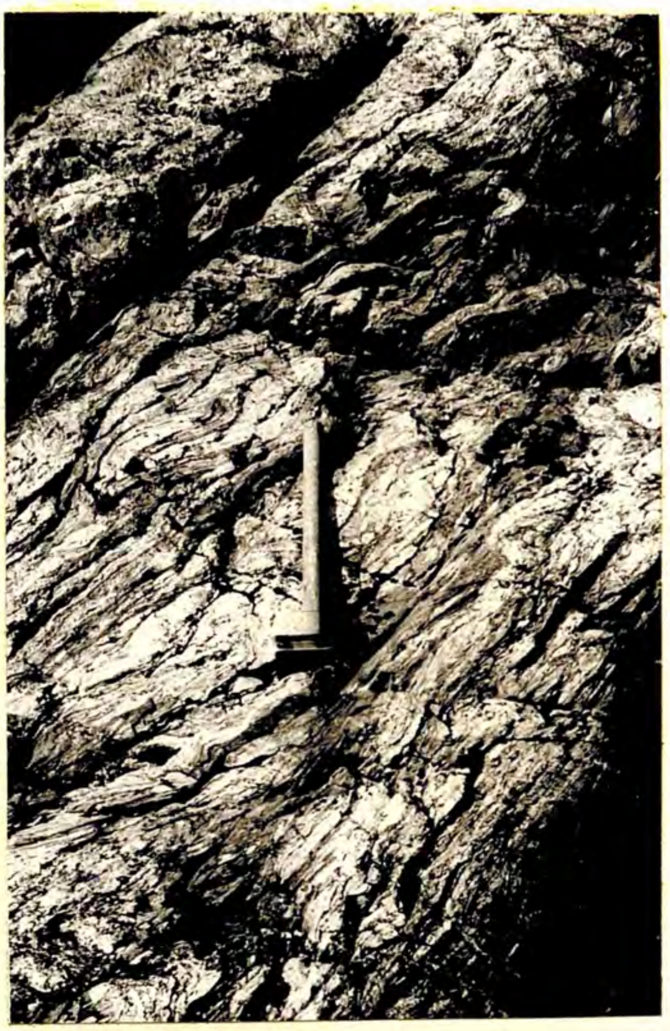


Plate 97. Partially sheared nepheline-syenite pegmatite. Vatna.

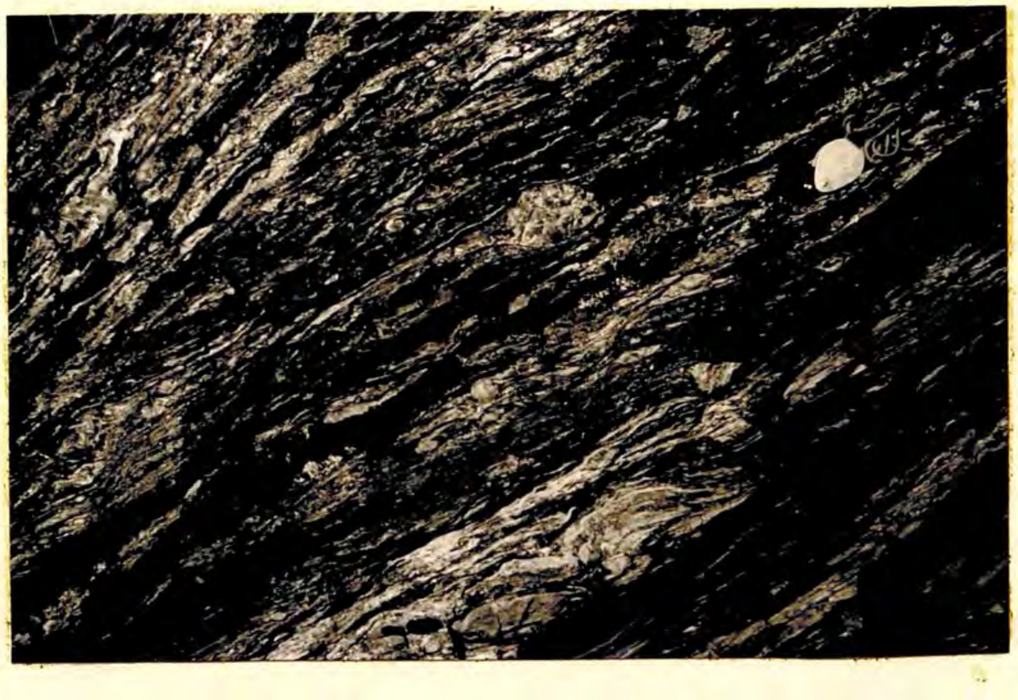


Plate 98. Augen in sheared nepheline-syenite pegmatite. Vatna.

been described in nepheline-syenites from Somali Republic by Gellatly (1964).

In the Vatna examples however, sections perpendicular to one another both show lenticular cross-sections to the augen. Thus the augen are truly lens-like in form, and are not rodded in any direction.

The shearing has also caused small-scale folding in the pegmatite (see Plate 97) in which small isoclinal folds have been formed. These are generally overturned towards the south, which is the same direction of overturning of the folds in the Husfjord metagabbro shear-zones.

Occasionally, fragments of basic dyke are enclosed by the pegmatite, and they are generally highly deformed, together with the pegmatite.

The exact age of the nepheline-syenite pegmatite in relation to the sequence of dyke emplacement cannot be determined unequivocably. Although it is seen to cut a few basic dykes, any evidence of later cross-cutting dykes has been obliterated by the shearing.

PETROGRAPHY

THE MAJOR INTRUSIONS

HUSFJORD METAGABBRO

General

The Husfjord metagabbro is essentially a clinopyroxene-gabbro although hypersthene occasionally occurs in some facies (Table 1). The gabbro has been variably amphibolized during subsequent metamorphism.

Four main facies are recognised within the gabbro; a fine-grained slightly-foliated type, a coarser-grained non-foliated type, a variably-grained non-foliated actinolite-bearing type and a contaminated noritic type.

Fine-grained Facies

The finer-grained, slightly foliated facies of the metagabbro has a relict subophitic texture. Plagioclase has recrystallized to equilateral grains, and the pyroxene has become partially altered to green hornblende (Plate 99). The hornblende is strongly pleochroic, its scheme being :

X light yellowish-green

Y deep olive-green

Z mid-dark green

Hypersthene appears to be practically restricted to this finer-grained facies. It usually occurs as fairly small grains, and is faintly

TABLE I

MODAL ANALYSES OF HUSFJORD METAGABBRO

Mineral	36/3A	36/139E	36/141A	36/199C	05/38B	38/5A	46/11A	05/34B
Plag.	34.01	35.63	52.73	44.30	45.50	34.06	40.73	31.80
Clinopy.	14.55	6.60	13.56	2.76	22.76	2.43	9.60	-
Hypers.	-	-	0.26	-	1.83	-	-	-
Hbl.	46.90	52.90	29.50	42.30	23.03	-	-	-
Act.	-	-	-	-	-	60.68	43.60	65.03
Bi.	3.20	3.13	2.80	7.66	5.60	0.76	3.70	0.36
Ore	2.96	1.83	1.16	1.83	1.23	0.93	2.10	0.30
Sph.	0.50	0.26	-	0.56	0.10	0.76	0.26	2.16
Ap.	0.12	0.13	0.16	0.20	0.23	0.40	0.36	0.26
Qu.	-	-	-	0.50	-	-	-	0.26
	99.24	100.48	100.17	100.11	100.28	99.84	100.35	100.17

Fine-grained facies: 36/3A, 36/139E.

Coarse-grained facies: 36/141A, 36/199C, 05/38B.

Actinolitic facies: 38/5A, 46/11A, 05/34B.

Plate 99. Pyroxene altering to secondary hornblende. Husfjord metagabbro. P.P.L. X 85.

pleochroic from buff to pale green.

The clinopyroxene is augite, maximum extinction angle being $Z \wedge c = 50^{o}$. Alteration to hornblende is predominant around the margins of the pyroxenes, the clinopyroxene being much more prone to alteration than the hypersthene. There is also a certain amount of alteration along cleavage planes of the pyroxenes. The contacts between the hornblende and the pyroxene from which it is forming is usually quite diffuse. In some cases, alteration has proceeded so far that pyroxene only appears as relict cores in the central parts of the secondary hornblende.

Biotite, which generally forms small laths, commonly has diffusely-defined contacts with hornblende, with which it appears to be unstable.

The feldspar is entirely plagioclase, and it is generally slightly cloudy. Albite twinning is well developed, and combined carlsbad and albite twins occur. Pericline twinning is fairly common, and this is usually subperpendicular to the albite twins. The great majority of the plagioclase is andesine in the range $An_{40} - An_{45}$, but some of the plagioclase grains are optically positive, and have compositions as high as about An_{55}. These more basic grains are generally smaller than the average, but contact relationships with the more sodic grains have not been seen. The labradorite probably represents relict original gabbroic plagioclase, whereas the andesine is a metamorphic plagioclase. This secondary plagioclase has a lower calcium content than the original because some calcium has presumably been used up in the alteration of pyroxene to hornblende.

Accessory minerals in this facies of the metagabbro include apatite and ore. The predominant ore in the metagabbro is ilmenite, which forms anhedral grains of variable size, usually contained within gangue crystals, but sometimes occurring at crystal boundaries. There are a few small rounded grains of magnetite. Pyrite occasionally occurs as small aggregates of anhedral grains, but more commonly along cleavages or thin irregular cracks in gangue minerals, indicating a late stage of formation.

Coarse-grained Facies

The coarser-grained facies of the metagabbro also has a relict subophitic texture in which the plagioclase laths have recrystallized and the pyroxene has been extensively altered to hornblende.

Hypersthene occasionally occurs in this facies, and here it is beginning to alter to brown iddingsite along cleavage planes and transverse cracks.

The augite often contains lamellar schiller inclusions of ore, principally ilmenite. This schiller structure is sometimes inherited by the secondary hornblende to which the pyroxene has extensively altered (Plate 100). Although this alteration is most advanced around the pyroxene margins, alteration in patches and along cleavages tends to be slightly more developed than in the finer-grained facies. Again, the contacts between the pyroxene and the hornblende are diffuse. The pleochroic scheme for the hornblende is the same as that for the finer-grained facies.

Some of the biotites form small laths which appear to be altering to hornblende, whereas other biotites are fairly large and in some cases

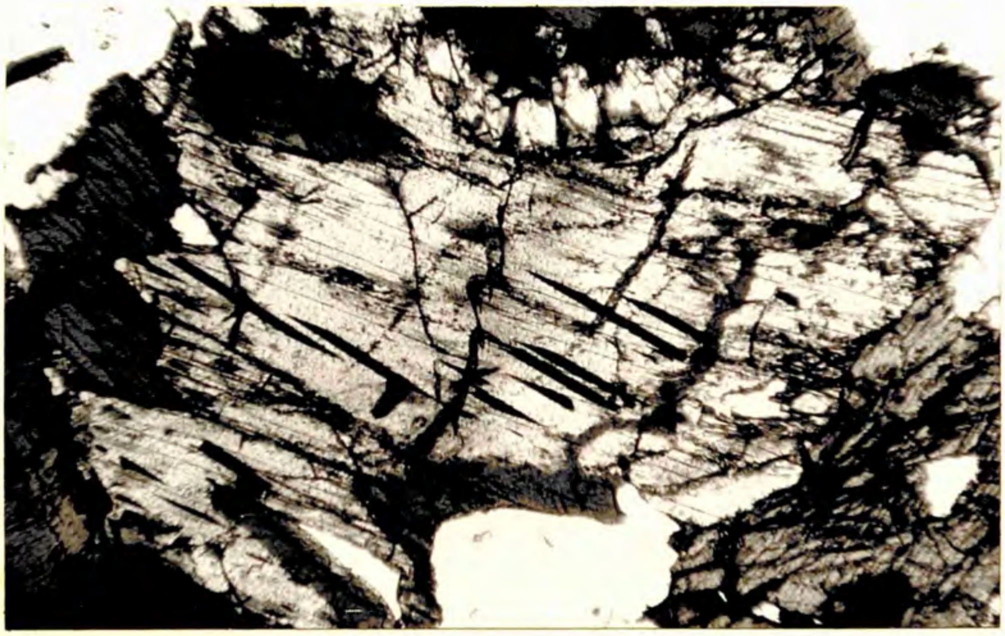


Plate 100. Schiller structure inherited by secondary hornblende. Husfjord metagabbro. P.P.L. X 85.

Plate 101. Actinolitic facies. Husfjord metagabbro. P.P.L. X 95.

form at the expense of hornblende. Thus there are two generations of biotite, and occasionally an early biotite is seen to be recrystallizing to a fresh-looking late biotite.

As in the case of the finer-grained facies, the plagioclase appears to be of two contrasting compositions. Their compositions and the relationships between them are the same as those in the finer-grained facies, and represent both relict original labradorite and secondary metamorphic andesine.

Actinolitic Facies

The third facies of the metagabbro is of variable grain-size, and is characterized by the development of an actinolitic amphibole. This facies is more commonly developed in the northern part of the body, although it does occur sporadically elsewhere.

In this facies, only a trace of the subophitic texture remains, and the rock appears to have recorded a longer recrystallization history than those already described. Much of the rock consists of hypidioblastic to allotrioblastic green actinolitic amphibole, which occasionally exhibits rhombic basal sections. The pleochroic scheme of the actinolite is as follows:

X straw-yellow

Y olivine-green

Z deep sea-green

Generally, the replacement of the pyroxene is complete, but in some cases relics of pyroxene remain in the amphibole masses. Small, rounded quartz blebs are common within the actinolite, and give it a sieve-like or diablastic texture (Plate 101).

Small irregularly-shaped grains of sp hene are common, often contained within the actinolite. Almost invariably the ore grains are rimmed by sphene, to which it appears to be altering. A similar phenomenon has been described in the Breivikbotn gabbro and the Storelv gabbro from northern and western Sørøy by Stumpfl and Sturt (1965). This alteration involves the oxidation of the ilmenite and the addition of Ca and Si, a reaction which may be represented by the following equation:

$$\underset{\text{ilmenite}}{6FeO.TiO_2} + O_2 + 6CaO + 6SiO_2 \rightarrow \underset{\text{sphene}}{6CaO.TiO_2.SiO_2} + 2Fe_3O_4$$

In some cases fresh biotite laths occur, and these tend to have a preferred orientation, forming a crude schistosity. They appear to be coexisting stably with the amphibole, and occasionally they contain inclusions of sphene.

The plagioclase generally forms a subequigranular mosaic in which the grain boundaries are irregular, but not sutured. It is in the composition range An_{35-40}, and no original labradorite remains. Albite twins are common, but pericline twinning is more unusual; when it does occur it is subperpendicular to the albite twins.

Occasionally, plagioclase is slightly altered to clinozoisite. In these areas some of the sodic plagioclase is found to be optically positive, and appears to be albite, although in general the plagioclase is of oligoclase or sodic andesine composition.

In a few cases a fair amount of pale yellow epidote has formed from plagioclase. Here, the plagioclase is oligoclase, and the neighbouring actinolite has a distinctly turquoise tinge, indicating that it is slightly sodic. It is possible that the sodium liberated during the formation of epidote has been incorporated in the amphibole rather than in the recrystallizing plagioclase.

Noritic Facies

At Øyfjord, where the metagabbro contains numerous rafts of metasedimentary material, a contaminated facies is developed. Here, it is often coarse-grained and has a rather poorly-formed subophitic texture. The predominant pyroxene is hypersthene, giving the rock a noritic aspect. Sometimes the hypersthene is variably altered to iddingsite or serpentine.

In most cases, hypersthene is rimmed by green hornblende to which it is altering. The pleochroic scheme of this hornblende appears below:

X pale buff

Y mid olivine-green

Z mid grass-green

Occasionally this amphibole has a turquoise tinge to it.

Accessory minerals include ragged biotite laths, apatite, and ore. The latter is principally ilmenite, which occasionally envelops hornblende, indicating a late stage of formation.

Feldspathization

At the southern end of the Ramnes peninsula the metagabbro has been slightly feldspathized, apparently in connection with the migmatization that has also affected some of the country rocks.

Here, the metagabbro is variably amphibolized, and where this is most extensive the green hornblende has a diablastic texture with numerous small rounded quartz inclusions. Biotite is growing from pyroxene, and its diffuse contacts with hornblende suggest that it may be growing from this mineral as well.

Potash feldspar porphyroblasts are allotrioblastic, generally being ovoid in shape. They are slightly perthitic, and a little myrmekite is sometimes developed at the porphyroblast margins where contact with plagioclase is made. Quartz inclusions also occur in the porphyroblasts. In some cases the potash feldspars have a rim consisting of a mosaic of plagioclase and quartz, and it would appear that the perthitic potash feldspars have recrystallized at their margins, forming a sodic plagioclase and quartz.

The metagabbro here contains numerous quartzo-feldspathic veins and streaks, which in the field are seen to have their origin in the psammitic metasediments of nearby rafts. These veins consist of a mosaic of plagioclase and quartz in which the grain boundaries are irregular, but not markedly sutured. The veins have sharply-defined margins but incorporate a little gabbro at their margins.

Thermal Metamorphism

The Husfjord metagabbro has been thermally metamorphosed by the large Havnefjord diorite. As the contact is approached, the pyroxene, which has already been slightly altered to amphibole during the preceding regional metamorphism, becomes progressively more and more amphibolized.

Although the pyroxene becomes completely or almost entirely replaced by amphibole, the subophitic texture of the metagabbro is still preserved.

The hornblende develops a marked diablastic texture, becoming sieved by small rounded quartz blebs (Plate 102). It is generally pale yellow-green to turquoise in colour, with the following pleochroic scheme:

X straw yellow

Y mid green

Z turzuoise-green

The cores of pyroxene which remained within the hornblende after the regional metamorphism have now altered to an aggregate of fibrous tremolite-actinolite with the following pleochroism:

X almost colourless

Y pale light green

Z pale light green

Plagioclase is essentially sodic andesine in the range An_{35-40}. But there are also some indistinctly-twinned grains whose refractive indices are approximately the same as balsam, and which are optically positive; these are presumably albite or sodic oligoclase.

In some cases greenish-yellow epidote occurs. There is usually considerable biotite, which isoften large and which frequently encloses grains of sphene and epidote. Sphene commonly forms coronas around ilmenite grains.

Occasional feldspar porphyroblasts form in the metagabbro. These are allotrioblastic to hypidioblastic, sometimes being subrectangular in shape. They are usually of oligoclase composition, and are generally poorly twinned. They are often strongly antiperthitic with irregular

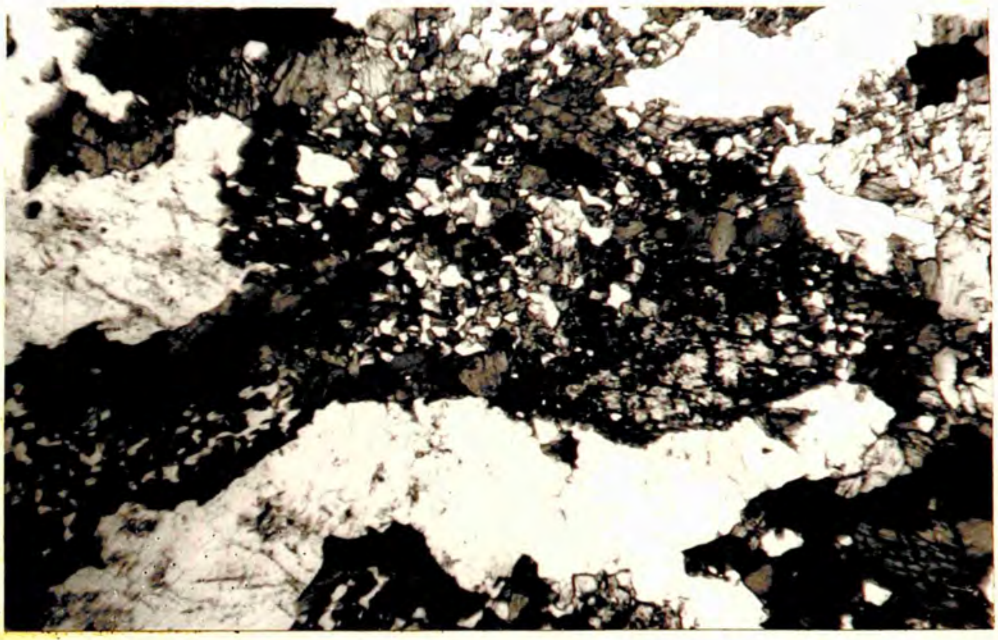


Plate 102. Diablastic texture in secondary hornblende after pyroxene; blebs are of quartz. Husfjord metagabbro. P.P.L. X 85.

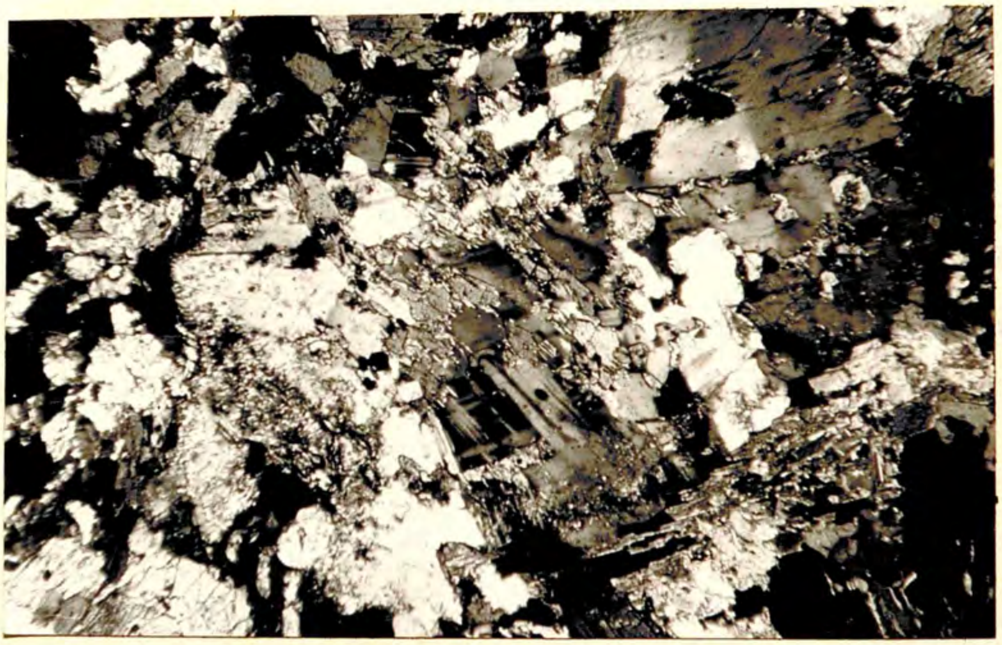


Plate 103. Inclusions in plagioclase porphyroblasts. Husfjord metagabbro. X-Pl. X 25.

patches of potash feldspar, which temselves are sometimes hair perthite. Some of the porphyroblasts contain some fine apatite needles which are randomly orientated, and also a few small quartz inclusions.

A common feature of the porphyroblasts is the inclusion of numerous small pyroxene grains which are in the process of altering to diablastic hornblende (Plate 103). Occasional small biotites are also enclosed within the feldspars.

Troctolites

The troctolite lenses in the Husfjord metagabbro are rich in olivine, which is beginning to alter to ore and antigorite, especially along cracks. Almost invariably the olivine is rimmed either by enstatitic orthopyroxene or a fibrous, colourless amphibole, probably cummingtonite. Orthopyroxene is the more common, although sometimes they both occur. In the latter case, the orthopyroxene lies between the olivine and the amphibole.

Clinopyroxene, which often encloses olivine, is extensively altered to a brown hornblende, which also envelops the orthopyroxene/cummingtonite coronas.

The Hornblende pleochroism is as follows:

X almost colourless

Y khaki-buff

Z dark fawn

The amount of plagioclase varies from body to body, but there is always some present. It is basic labradorite, having a maximum anorthite content of An_{70}, and is twinned on both albite and pericline laws. Combined carlsbad and albite twins also occur. Some crystals are slightly sericitized. In the neighbourhood of the olivines, the plagioclases in particular, but also other minerals, are extensively cracked. The cracks are irregular and closely spaced and radiate out subperpendicularly to the margins of the olivines (Plate 104). These cracks are expansion cracks formed during the alteration of olivine to orthopyroxene. This reaction is accompanied by an increase in volume (Hatch, Wells, and Wells, 1961), which causes the formation of cracks in the neighbouring minerals.

There is only a little biotite, and this appears to pre-date the hornblende.

The ores include both oxides and sulphides. Ilmenite and magnetite both occur as fairly large anhedral grains, which have recrystallized and now partly enclose gangue minerals and occasionally contain inclusions of them. Small grains of pyrrhotite are common and these invariably have patches, lamellae, or flame-like intergrowths of pentlandite, probably due to exsolution. Rarely, these composite grains also contain a little patch of chalcopyrite. They are sometimes partly altered to goethite, which either rims the sulphides or grows from within. When the alteration has proceeded far, only relics of pyrrhotite and pentlandite occur within the goethite, and a few grains have been completely pseudomorphed by goethite. The ore which occurs along the cracks in the altered olivine is magnetite.

The metagabbro next to the troctolites is generally considerably

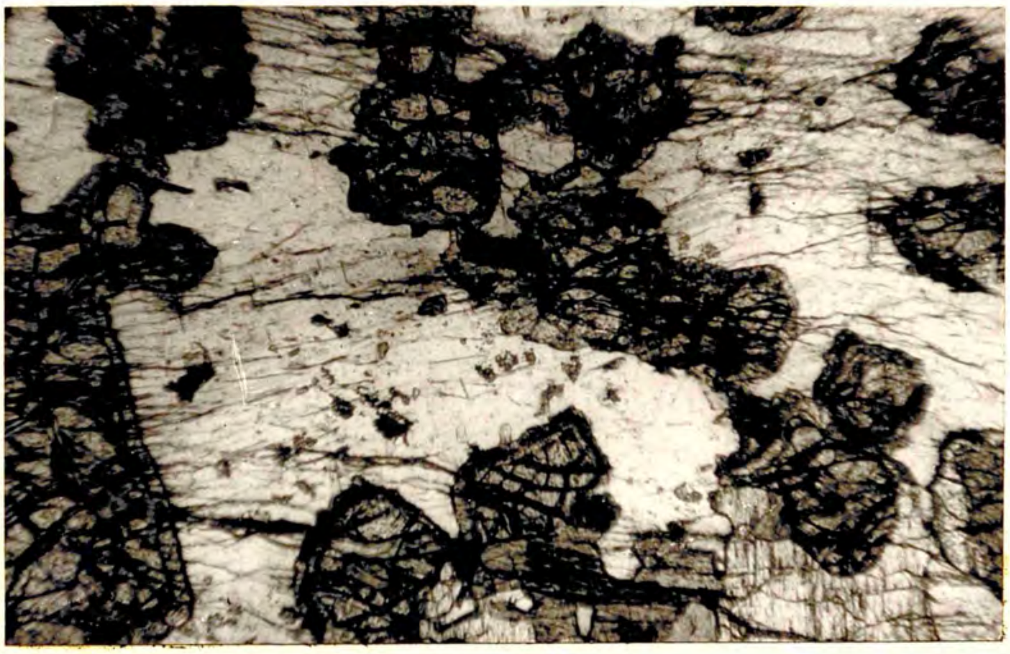


Plate 104. Expansion cracks around altered olivine grains. Troctolite. P.P.L. X 85.

Plate 105. Ore needles in biotite. Troctolite. P.P.L. X 95.

amphibolized, and is fairly rich in biotite, indicating that there has been a considerable amount of water present here during metamorphism. The contact between the troctolite and the metagabbro probably provided a plane along which fluids could have readily penetrated.

Ultrabasic rocks, probably troctolite, occur as relics within the Havnefjord diorite. In these, a relict ophitic texture is preserved in which the plagioclase is labradorite, and only a little clinopyroxene remains.

The rock has been considerably amphibolized, and there are two distinctly different amphiboles, One is a brown hornblende which occasionally contains clouds of finely disseminated ore, and which has the following pleochroic scheme:

X light khaki

Y mid brown

Z greenish-brown

The other amphibole is a pale green actinolite which is seen to be forming at the expense of the brown hornblende. It contains diffuse relics of hornblende, and also of biotite, and its pleochroic scheme is as follows:

X pale yellowish-khaki

Y pale green

Z pale slaty-green

Biotite is of two generations. Early biotites are altering to the green actinolitic amphibole, and sometimes occur as diffuse inclusions within it. The later biotite is often large, fairly pale in colour, and sometimes poikiloblastic. The late biotites commonly contain inclusions of ore which occur as blebs and as minute, fine needles. These needles are generally orientated either parallel to the cleavage planes of the biotite, or subperpendicular (70° - 85°) to one another, the angle between them being bisected by the cleavage. Sometimes both these orientations occur together. On some basal sections the needles trend in three directions exactly at 60° to one another (Plate 105). This p attern resembles the percussion figure characteristic of some strained biotites (Deer, Howie, and Zussman, 1962).

The former presence of olivine is indicated by occasional pseudomorphs of ore and iddingsite, the latter itself now altered to a brown alteration product.

Early Diorites

The early diorite bands have been emplaced into the Husfjord metagabbro and are essentially pyroxene-mica-diorites, occasionally with a little interstitial quartz (Table 2).

The principal pyroxene is hypersthene and this is pleochroic from pinkish-buff to pale green. The hypersthene and the larger feldspar crystals tend to have a preferred orientation parallel to the contacts of the diorite bands. In many cases there has been no post-emplacement shearing, and the orientation would appear to be a fluxion structure.

Feldspar occurs both as potash feldspar and plagioclase of composition An_{40-45}. Both types of feldspar occur as phenocrysts, the large plagioclase crystals sometimes being slightly deformed. The plagioclase phenocrysts are generally antiperthitic, with irregular patches of potash

TABLE 2

MODAL ANALYSES OF EARLY DIORITES

Mineral	36/44A	36/54C	36/201A	36/54D[1]	36/300A[1]
Plagioclase	67.15	73.57	58.03	33.30	48.00
K-feldspar	15.29	9.32	28.06	46.00	38.69
Quartz	0.48	6.12	-	18.59	-
Myrmekite	2.34	-	0.76	-	-
Hypersthene	6.42	9.09	9.06	0.91	8.41
Biotite	3.57	0.08	0.30	0.63	0.68
Ore	3.96	1.56	3.16	0.89	3.48
Hornblende	-	-	0.26	-	-
Apatite	0.55	0.29	0.50	-	0.12
Zircon	0.24	-	-	-	0.34
	100.00	100.03	100.13	100.32	99.72

1 : Modal analysis carried out by macro point counting techniques - See Appendix.

feldspar. Potash feldspar megacrysts are often rimmed by lobes of myrmekite where contact with plagioclase is made.

Biotite is formed from hypersthene, and the reaction occurs mainly at the contacts between hypersthene and potash feldspar. The biotites usually contain numerous rounded and vermicular inclusions of quartz. Sometimes the quartz is so abundant that the biotite has a frittered or dactylitic appearance. This vermicular or dactylitic biotite is frequently closely associated, and sometimes intergrown with, myrmekite (Plate 106). It would seem that the development of this particular myrmekite from potash feldspar is linked with the formation of the biotite from hypersthene. The two reactions involved appear to be mutually assistant, and since they probably took place during late stage metamorphism, are discussed in more detail on p. 302 .

A late metasomatism seems to have affected the early diorites, since occasionally hypersthene is altered completely to aggregates of biotite and turquoise amphibole, and the feldspars become very cloudy and sericitized. Sphene is common, there is a little epidote, and a few rare grains of schorlite. Small remnants of hornfelsed metagabbro enclosed within the diorite are also metasomatized, grains of pyroxene having altered to sodic amphibole and quartz. It appears that the diorite contacts have channelled the metasomatic fluids, since the alteration is confined to the margins of the diorite bands.

The principal ore in these diorites is ilmenite which occurs mainly as large, irregular, often elongate grains. The elongate grains are orientated subparallel to one another and to the fluxion structure of the

Plate 106. Vermicular biotite and myrmekite. Early diorite. X-Pl. X 170.

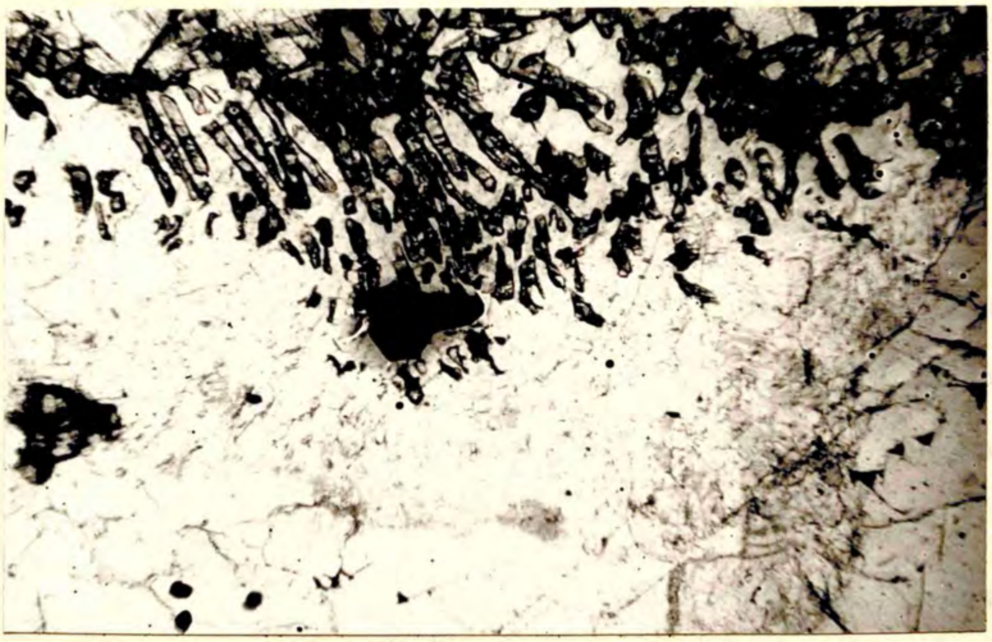


Plate 107. Dactylitic orthopyroxene and myrmekite. Early diorite/ultrabasic contact. P.P.L. X 85.

diorite. The extremities of the long grains are often fragmented into trains of small angular grains, suggesting that they have suffered deformation.

Magnetite also occurs, and this forms anhedral grains of various sizes. Occasionally they contain straight-sided lamellae of ilmenite, presumably an exsolution phenomenon. Pyrite sometimes occurs at the margins of other ores, and it also penetrates into very fine cracks in some of the silicates, suggesting a late crystallization (or recrystallization). Some of the pyrite is altering to goethite.

The Husfjord metagabbro has been hornfelsed by the early diorites, and this hornfelsing post-dates the peak of the regional metamorphism, since a fine-grained granoblastic hornfelsic texture overprints the regional metamorphic texture and mineralogy. A pyroxene hornfels is developed in which the pyroxenes include rounded grains of both hypersthene and clinopyroxene, and the small plagioclase crystals are basic andesine in the range An_{40-45}. At the contact itself there is only pyroxene, plagioclase, and ore, but about 5 mm. from the contact these are accompanied by hornblende and biotite. The biotites form small laths which appear to be forming from hornblende, and these tend to have a preferred orientation indicating that they probably crystallized during the F_2 movements.

An interesting phenomenon occurs where an early quartz-bearing diorite makes contact with a troctolite lens. Where the potash feldspar of the diorite makes contact with clinopyroxene in the ultrabasic rock, the pyroxene is fringed with orthopyroxene and the potash feldspar is replaced by myrmekite. Occasionally vermicular intergrowths between orthopyroxene and myrmekite develop, which in some cases form dactylite (Plate 107). This reaction never takes place where clinopyroxene meets quartz or plagioclase in the diorite. Whether the reaction is a primary feature, forming at the time of the diorite emplacement, or a secondary metamorphic feature is not clear. In either case it is apparent that at some time clinopyroxene-potash feldspar boundaries were chemically unstable and a reaction took place. The equation of the reaction might be simplified as follows:

$$\underset{\text{clinopyroxene}}{CaO.(Mg.Fe)O.2SiO_2} + \underset{\text{perthitic potash feldspar}}{2\left[\tfrac{1}{2}K_2O.\tfrac{1}{2}Na_2O.Al_2O_3.6SiO_2\right]} \rightarrow$$

$$\underset{\text{orthopyroxene}}{(Mg.Fe)O.SiO_2} + \underbrace{\underset{\text{plagioclase}}{2\left[\tfrac{1}{2}Na_2O.\tfrac{1}{2}CaO.Al_2O_3.4SiO_2\right]} + \underset{\text{quartz}}{5SiO_2}}_{\text{myrmekite}} + K_2O$$

Metamorphism

In places the early diorites have been metamorphosed by the Havnefjord diorite, and also by the late coarse-grained pyroxene-mica diorites. In these altered diorites, the hypersthene has been replaced by turquoise amphibole which now forms aggregates pseudomorphing the pyroxene. The amphibole is pleochroic from light yellowish-green to turquoise green. The amphibole in the central parts of the aggregates have a diablastic texture, containing numerous small rounded blebs of quartz.

Biotite is common around the margins of the amphibole aggregates, and it also contains bleb and vermicules of quartz. In places it can be seen to be forming from the amphibole, sometimes containing diablastic amphibole relics within it.

Ore is frequently, but not invariably, rimmed by sphene, and there is occasionally a little epidote.

The potash feldspars are generally cloudy, especially around the borders of the megacrysts, and myrmekite often fringes the potash feldspars.

In two instances relict contacts between early diorites and metagabbro are preserved within the Havnefjord diorite (Plate 108, cf. Plate 109). The pyroxenes of the hornfelsed metagabbro at the contact are altered in a similar way to those of the diorite itself, and are replaced by small aggregates of diablastic turquoise-green amphibole and biotite. The similarity of the mineralogy of the rocks on either side of the contact appears to indicate that the reaction that has occurred represents an example of metamorphic convergence. In this case it seems that sodium in particular has been mobile and has migrated from the diorite to the metagabbro.

Pegmatites

The pegmatites that have been emplaced into the Husfjord metagabbro contain quartz, potash feldspar, and sometimes plagioclase. The plagioclase is generally oligoclase, but in some cases an untwinned plagioclase occurs which is optically positive and has a refractive index lower than that of quartz, and this is presumably albite. When plagioclase occurs it is often antiperthitic, but perthitic potash feldspar is not common. However, myrmekite occasionally occurs at the borders of potash feldspar grains.

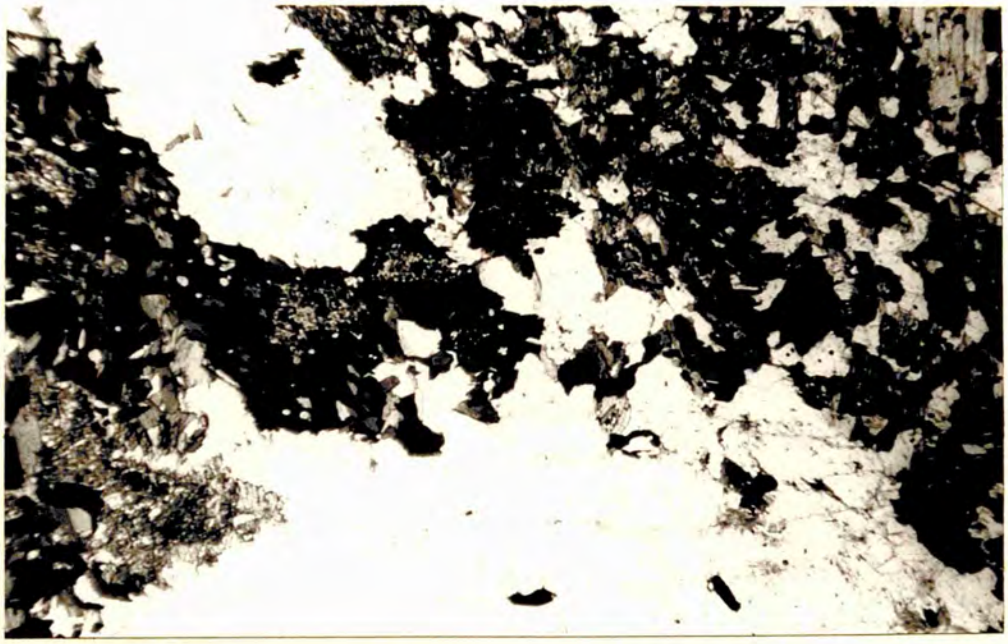


Plate 108. Relict metamorphosed early diorite/ metagabbro contact. P.P.L. X 25.

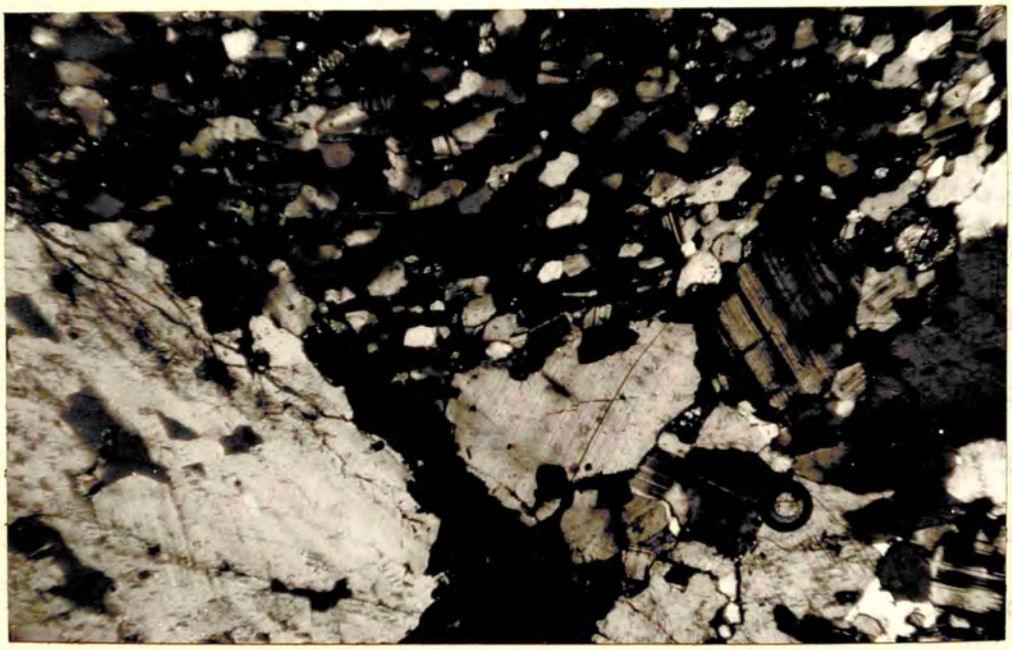


Plate 109. Original unmetamorphosed early diorite/ metagabbro contact. X-Pl. X 25.

Many of the pegmatites contain a little biotite, which is usually rather ragged in appearance. Sometimes there is a little ore, and rarely, zircon, although one red-weathering pegmatite on the Ramnes peninsula is fairly abundant in ore and zircon. In one pegmatite there are a few schorlite crystals, which are strongly pleochroic from buff to slaty-blue.

None of the pegmatites are fresh; all have recrystallized, presumably during the regional metamorphism, and the feldspars are usually cloudy. The myrmekite probably formed during the recrystallization, and many grain boundaries are commonly highly sutured, especially in the case of quartz/quartz boundaries.

Many pegmatites have suffered deformation in which irregularly-shaped quartz crystals exhibit undulose extinction. In one pegmatite not only has the quartz been strongly strained, but the potash feldspar megacrysts have granulated margins, and a mortar texture is developed.

The metagabbro in the neighbourhood of the pegmatites contains a considerable amount of hydrous minerals, and only a little highly-altered pyroxene remains, most of it having been amphibolized. Green hornblende is abundant, and generally forms a diablastic texture, containing numerous quartz blebs. In some cases there is a large number of rounded and elongate grains of ore contained within or occurring around the hornblendes (Plate 110). They are closely associated with the hornblende, and are probably formed from excess ore resulting from the amphibolization process.

In many cases, large poikiloblastic biotites occur containing inclusions of quartz and apatite (Plate 111), and they commonly enclose diablastic hornblende from which they can be seen to be forming (Plate 112).

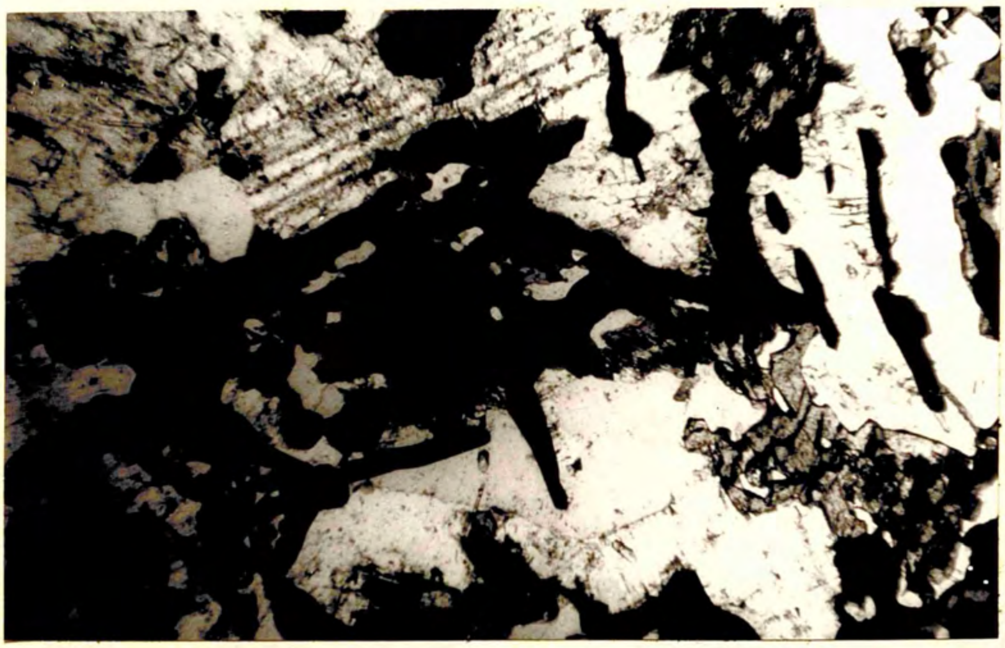


Plate 110. Exsolution lamellae and blebs of ore associated with diablastic hornblende. Husfjord metagabbro. P.P.L. X 85.

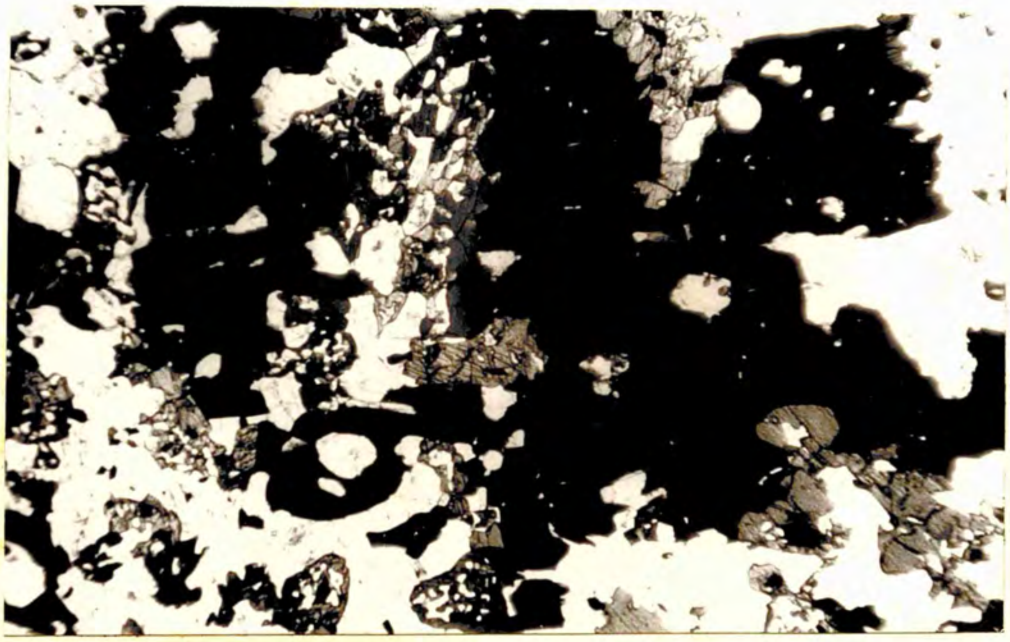


Plate 111. Poikiloblastic biotite. Husfjord metagabbro. P.P.L. X 25.

Plate 112. Relict diablastic hornblende in biotite. Husfjord metagabbro. P.P.L. X 85.

Plagioclase has recrystallized into a polygonal mosaic of grains of oligoclase or sodic andesine. In most cases they are approximately An_{35}, but plagioclase in the range An_{15-20} has been found in some examples. There is usually a fair amount of quartz, usually interstitial, but sometimes also occurring as larger grains.

The presence of biotite in the pegmatites indicates that the pegmatites had a fairly large water content, and that they were probably capable of causing the extensive alteration of pyroxene to amphibole and biotite in the neighbouring metagabbro.

Metasomatic Veins

The narrow, hard, dark metasomatic veins that cut the metagabbro consist primarily of fibrous actinolitic amphibole. The fibres tend to be orientated subperpendicular to the vein walls. This amphibole is pleochroic from pale yellowish-green to turquoise green, suggesting that it is slightly sodic. The margins of the vein are a deeper turquoise than the central parts.

The veins vary in thickness from a fraction of a millimetre up to about 1 millimetre. The former type are impersistent, dying out intermittently, and are often branching. The latter, however, maintain a fairly constant thickness for many centimetres, and alter the neighbouring metagabbro for about 1cm. on either side of the vein. Within this zone of alteration, the pyroxenes of the metagabbro have been progressively replaced by masses of actinolitic amphibole which has the same characteristics as that in the veins. The hornblende has also recrystallized to the new actinolitic amphibole. Plagioclase, which is about An_{30}, has been extensively sericitized, and there is also a little epidote. Sphene occurs within the amphibole masses, and forms coronas around the ores. Sphene is also common in the veins themselves, and often forms crystals as much as 1mm. in diameter (Plate 113).

The nature of the emplacement of these veins can be deduced from study of the vein margins. Although the veins appear to cut across minerals in the metagabbro, they have no continuous sharp margins. The amphiboles and sphenes of the vein frequently cross the vein contacts, and where crystals of green hornblende in the metagabbro cross the vein, they have recrystallized to the sodic amphibole, but in optical continuity with the remainder of the crystal (Plate 114). It would appear that the veins have not been emplaced by the injection of a liquid, but rather by the soaking of liquids along planes of weakness such as joints, metasomatically altering the metagabbro.

Another kind of volatile-rich vein occurring in the actinolitic facies of the Husfjord metagabbro is one bearing scapolite. In the metagabbro next to the vein, feldspar becomes scapolitized, and the actinolite is replaced by secondary diopside. The ores become more extensively rimmed by sphene, and this latter mineral forms large crystals at the contact with the vein.

Shear-belts

The rocks in the two shear-belts do not show extreme crushing or mylonitization, but only the development of a foliation, and occasionally a slight mortar texture.

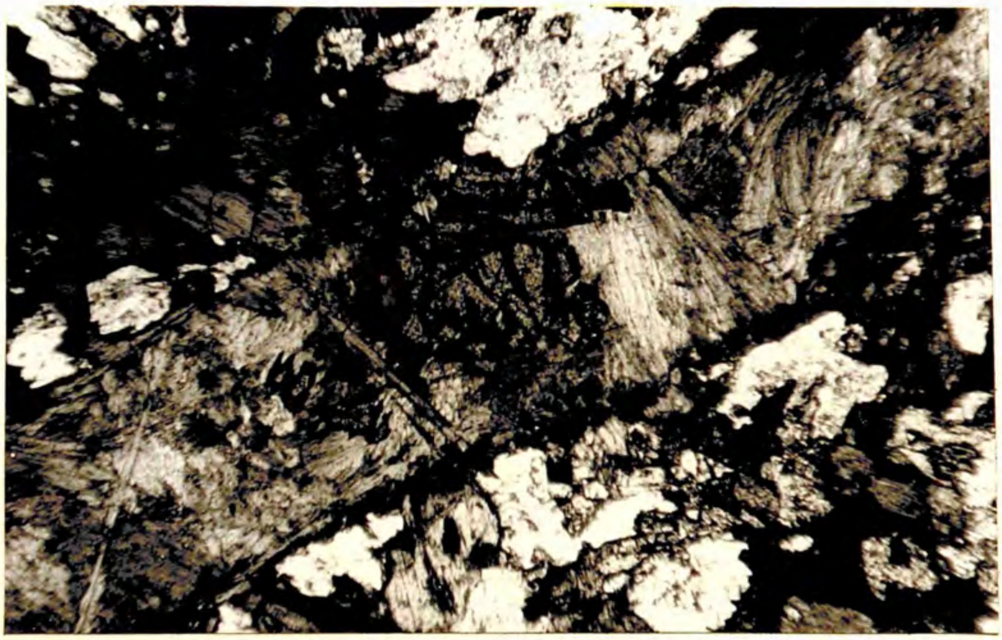


Plate 113. Sphene in actinolitic metasomatic vein in Husfjord metagabbro. P.P.L. X 85.

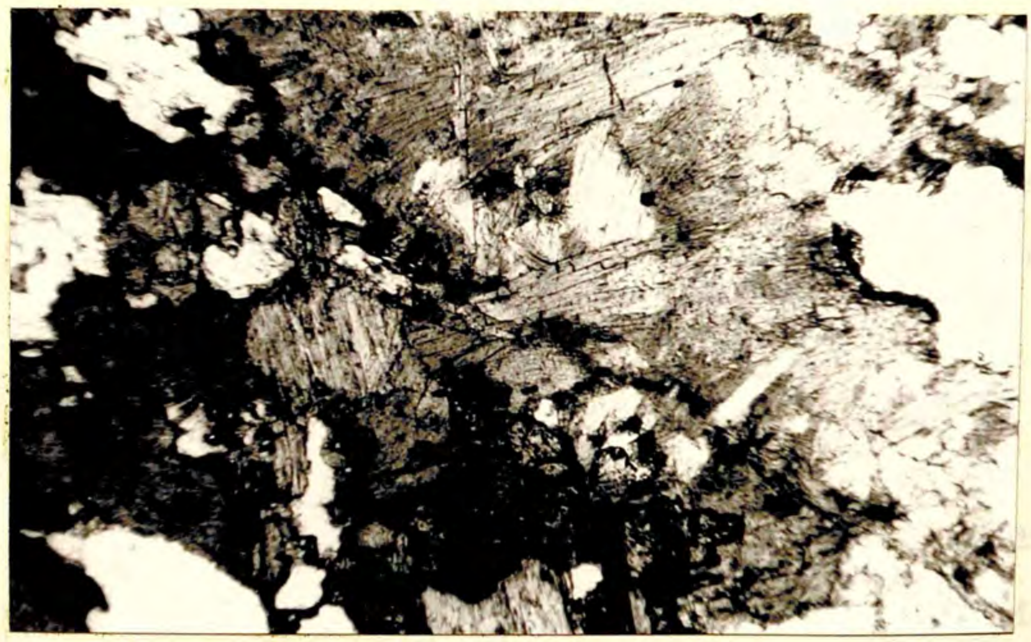


Plate 114. Hornblende recrystallized to actinolite where it is crossed by metasomatic vein. Husfjord metagabbro. P.P.L. X 95.

In the metagabbro a foliation is developed and there is incipient augening; hornblende becomes ovoid in shape and margins are beginning to be granulated (Plate 115).

In the southern shear-belt there are numerous ultrabasic lenses, which now consist principally of brownish hornblende which has formed from clinopyroxene. Within the hornblende masses there are numerous aggregates of granular colourless orthopyroxene, which probably indicate the former presence of olivine. One of these orthopyroxene aggregates contains a greenish-brown serpentine pseudomorph after olivine.

Plagioclase, now basic andesine, has recrystallized and incorporates grains of clinopyroxene and hornblende. Sometimes these inclusions tend to lie along the direction of the twin planes, but occasionally also perpendicular to this direction (Plate 116).

The ultrabasic rocks are sheared at their margins and develop a foliation. The hornblendes have a preferred orientation, and the aggregates of plagioclase grains are elongated parallel to this direction. The aggregates of granular orthopyroxene are also elongated in this direction.

Where the early diorite bands are sheared, a mortar texture is beginning to form, with feldspars forming insipient augen (Plate 117). In the hornfelsed metagabbro at the contacts with the early diorite bands, small biotites have a preferred orientation, forming a crude schistosity.

Mylonites

In the mylonite zones the metagabbro is intensely sheared, and a

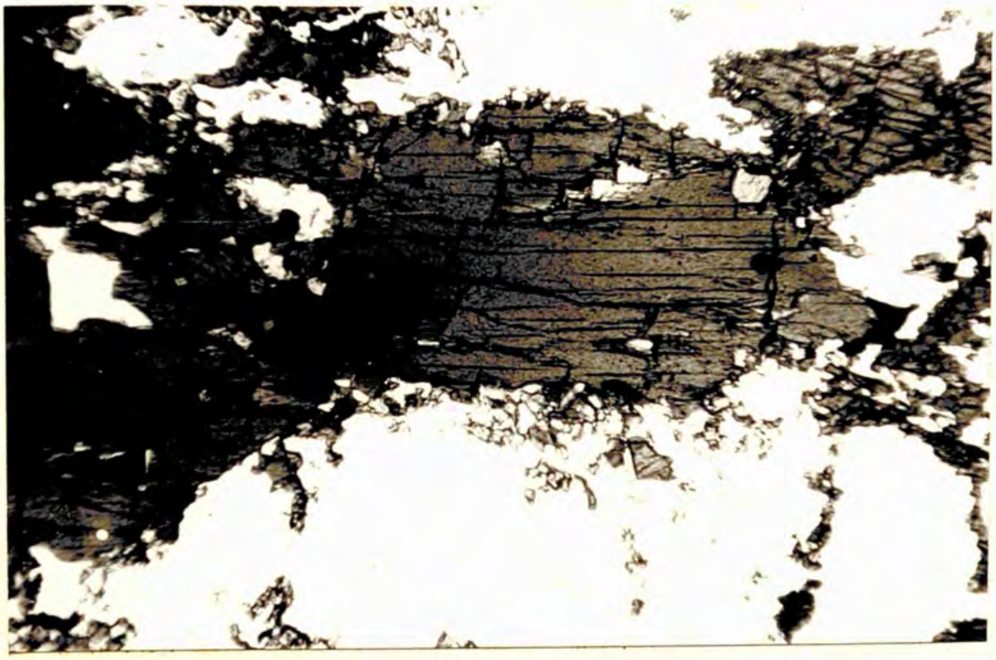


Plate 115. Hornblende with granulated margins. Husfjord metagabbro. P.P.L. X 95.

Plate 116. Inclusions in plagioclase. Troctolite. X-Pl. X 170.

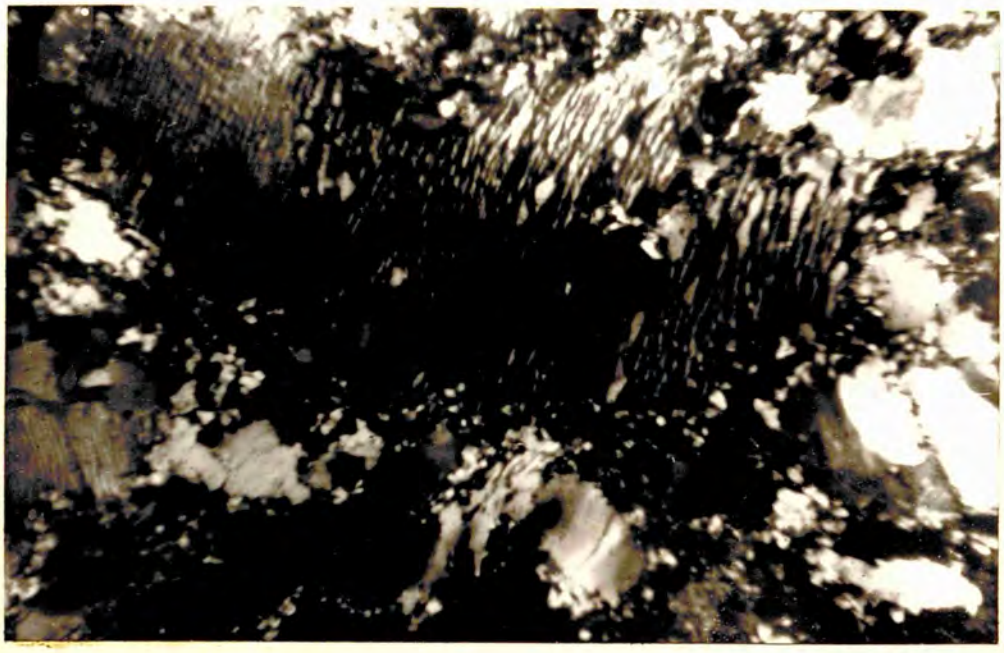


Plate 117. Feldspar with granulated margins. Early diorite. X-Pl. X 85.

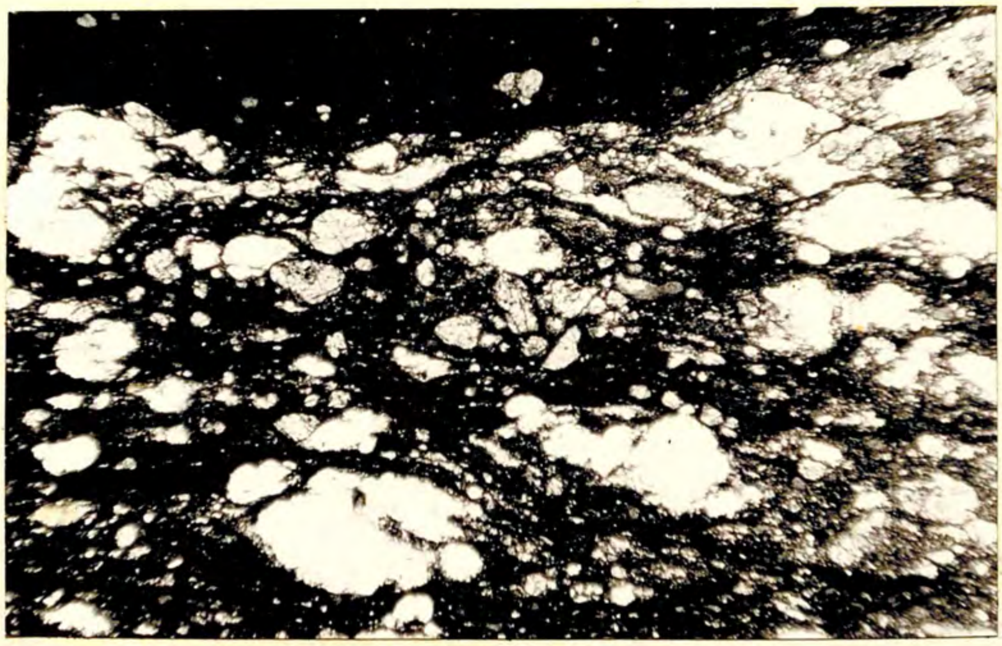


Plate 118. Edge of intense mylonite zone. Husfjord metagabbro. P.P.L. X 2.5.

dense greenish laminated mylonite is formed. The most intense zones of crushing are generally only a few millimetres in width, but the meta-gabbro becomes progressively more and more sheared as these are approached. The edge of an intense zone is shown in Plate 118.

The mineral most sensitive to the shearing appears to be ore which very readily forms strings of fine-grained blebs stretched out along the foliation. As the mylonite is approached, biotite, hornblende, and some of the pyroxene begin to break up and recrystallize with a fine-grained granular texture around the more resistant pyroxenes and plagioclase grains which form augen. At this stage the augen still retain some of the original shape of the minerals concerned.

Adjacent to the mylonite, the foliation is much more marked, and the augen of hornblende and pyroxene are smaller. In the intensely sheared mylonite itself, the augen are lenticular in shape and consist of pyroxene, plagioclase, and occasional hornblende. Their long axes are parallel to the foliation and they are set in a green matrix of hornblende, biotite, rounded granular pyroxene, and streaks of fine ore grains which swirl around the augen. Even in the most intensely-sheared mylonite the augen are not eliminated altogether.

In one mylonite there are two aggregates containing small rounded grains of garnet which appear to be recrystallizing after the mylonitization. Small biotite flakes have also grown after the shearing, indicating that these mylonites probably formed during the F_2 movements.

Rafts and Xenoliths

Pelitic and Semi-pelitic Rafts

These are principally quartz-mica-schists and garnet-mica-schists. The schistosity is delineated by biotite laths, which are often quite long, making the rock fairly coarse-grained. The biotites commonly contain pleochroic haloes, and are sometimes deformed.

Quartz and feldspar form fine-grained subequigranular mosaics between the biotite laths. Feldspar is principally potash feldspar, including microcline, but in a few rocks oligoclase is also present. This latter is antiperthitic, and the potash feldspar lamellae sometimes have microcline twinning. Occasionally feldspar is a little cloudy, but it is usually fairly clear.

A pale green actinolite amphibole is present in some of the semi-pelites and has the following pleochroic scheme:

X buff

Y light mid green

Z pale mid green

Sphene is a common mineral in these schists, and is often quite abundant, usually forming small ragged grains. In one quartz-mica-schist there is a little epidote.

Garnets, which are generally buff-coloured and irregular in shape, overprint the F_1 schistosity, which in places is folded, and contain inclusions of biotite, quartz, sphene, and apatite. The garnets pre-date the F_2 folding, since they have suffered deformation and are considerably cracked, with secondary biotite forming in the cracks.

One richly-aluminous pelite contains some corundum, and another pelite contains a few irregular grains of tourmaline, pleochroic from colourless to pale greenish-blue.

A few pelites have occasional allotrioblastic porphyroblasts of perthitic microcline which sometimes contains a few inclusions. Myrmekite occurs at the microcline margins, and some porphyroblasts have recrystallized to a mosaic texture.

The pelitic and semi-pelitic rafts are usually hornfelsed at their margins by the metagabbro. A hornfelsic texture is developed in which decussate biotite laths lie along the grain boundaries of a polygonal mosaic of quartz and feldspar grains. However, in places there is a relict schistosity, with biotite laths tending to lie subparallel to one another.

Metalimestone Rafts

The metalimestones consist of coarsely-recrystallized marble containing a subequigranular mosaic of cloudy calcite grains which have slightly irregular margins. Small rounded interstitial grains of diopside are common, and occasional interstitial potash feldspars also occur, as well as small amounts of quartz, garnet, and idocrase in some of the rocks.

The thin calc-silicate bands in the metalimestones consist of small rounded grains of diopside, calcite, quartz and feldspar, sometimes with ragged-looking laths of ore. These bands have a fine-grained granular texture, and their margins are well defined.

One raft has some calc-silicate material slightly brecciated along along the contact. In this, there are large aggregates of garnet which incorporate rounded grains of diopside, calcite, and quartz; calcite and quartz also form aggregates and streaks. A green coloured calc-silicate band mainly consists of irregular aggregates of allotrioblastic green diopside intergrown with calcite and andesine.

Occasionally the metagabbro adjacent to metalimestone rafts contains a few irregularly-shaped buff-coloured garnets, possibly due to contamination. The garnets usually contain small quartz vermicules, and is itself breaking down and altering to secondary biotite. This alteration is marked near to pegmatites, which are commonly emplaced along the contacts of rafts.

Calc-silicate-schist Xenoliths

The principal mineral in the fine-grained xenoliths of calc-silicate-schists is diopside. In one case, the diopsides form a mass of rounded grains, and together with some phlogopitic biotite laths are enclosed as inclusions within poikiloblastic scapolite. The scapolite itself is sometimes altering to a brown alteration product along cleavage planes.

In another xenolith, a granular hornfelsic texture comprising rounded grains of diopside and plagioclase has been overprinted by a regional metamorphic texture and mineralogy. Tremolite, quartz, and poikiloblastic biotite form an F_2 schistosity, and enclose rounded grains of diopside and plagioclase.

Basic hornfels Xenoliths

These consist of fine-grained granoblastic pyroxene hornfelses in which labradorite occurs together with non-pleochroic, poikiloblastic orthopyroxenes which contain quartz blebs. The pyroxenes are beginning to alter to a pale yellowish-brown amphibole, and there are also some pale biotite laths which post-date the orthopyroxenes.

These hornfelses appear to represent basic sediment inclusions in the metagabbro which have been thermally metamorphosed by the metagabbro.

Psammitic and Semi-pelitic Rafts

The rocks comprising this group occur as rafts on the Ramnes peninsula, and have been variably migmatized after the emplacement of the Husfjord metagabbro.

The psammites are gneissic and consist mainly of quartzo-feldspathic material, with various amounts of biotite, sillimanite, and garnet. Sometimes psammitic bands contain large garnet porphyroblasts, and these are allotrioblastic, often enclosing quartz (Plate 119). Some of the smaller garnets are slightly altered to fibrolite and ore, especially at their margins, but the relationships are rather obscured by the presence of biotite which surrounds or partly surrounds many garnets. The biotite itself i s also altering to sillimanite.

Along nearly all the potash feldspar/potash feldspar grain boundaries and the quartz/potash feldspar boundaries there are aggregates of fibrolite (Plate 120) similar to those described in some of the migmatites of the country rocks. They do not occur at quartz/quartz contacts.

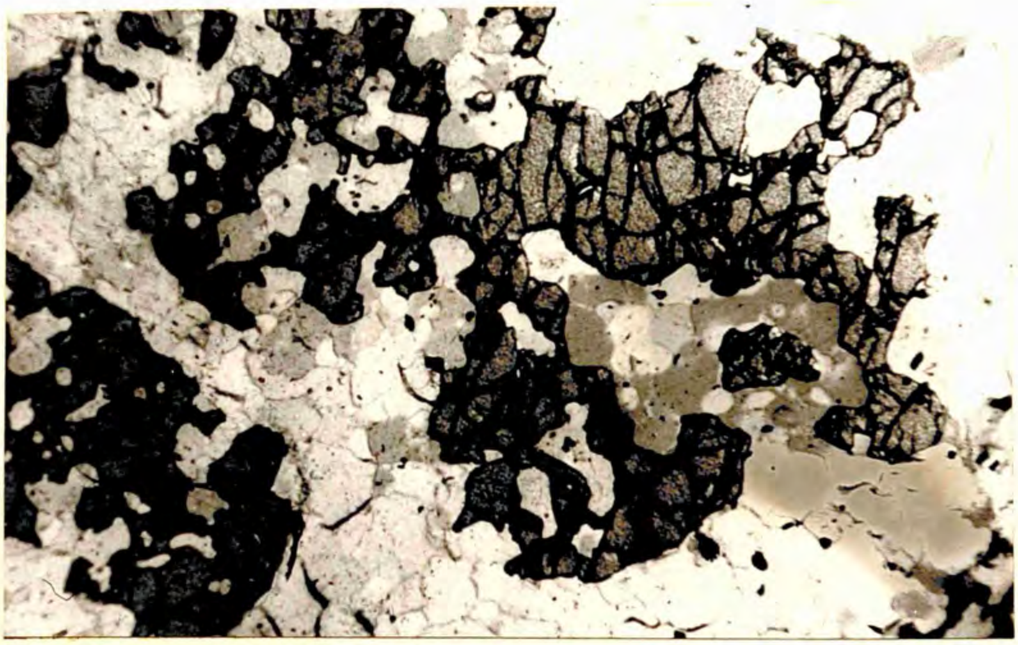


Plate 119. Garnet porphyroblast in psammitic raft in Husfjord metagabbro. P.P.L. X 25.

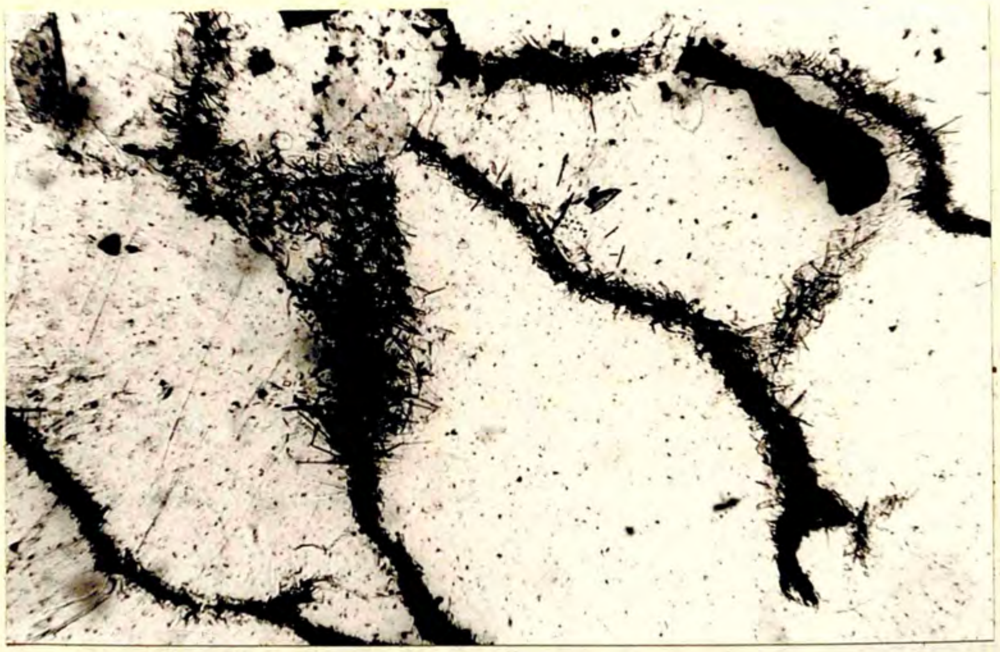


Plate 120. Fibrolite along grain boundaries. Migmatized psammitic raft in Husfjord metagabbro. P.P.L. X 170.

These whiskery fibrolite aggregates also nucleate on the edges of biotite-sillimanite aggregates and on the margins of the garnet porphyroblasts (Plate 121).

In the semi-pelites, quartz and potash feldspar form a subequigranular mosaic with irregular margins. Some of the biotites occur along the mosaic grain boundaries, whereas others form aggregates elongated parallel to a gneissic banding. Biotite is often altering to sillimanite, which generally forms aggregates of small prismatic crystals replacing the biotite. Sometimes sillimanite forms strings of fairly large idioblastic to hypidioblastic crystals occurring in quartz-rich pods (Plate 122).

Some semi-pelites contain garnets which overgrow the biotites, and have minute inclusions of ore, quartz, and biotite, but the garnets themselves are breaking down to secondary biotite. Ore probably recrystallized late, since it occurs along grain boundaries and envelops plagioclase crystals (Plate 123).

One of the semi-pelitic rafts shows evidence of deformation at its margin. Aggregates of granular, strained quartz and small biotite laths tend to swing around ovoid feldspars. Garnets are cracked and are also altered to secondary biotite. The central parts of the raft, however, do not show this deformation.

Basic sheets occurring in these psammitic and semi-pelitic rafts contain hypersthene, which is altering to a yellow iron-stained alteration product along cleavages and cracks. The pyroxene is rimmed by pale green hornblende to which it appears to be altering; sometimes this hornblende has a turquoise tint indicating its sodic content. Grain boundaries of

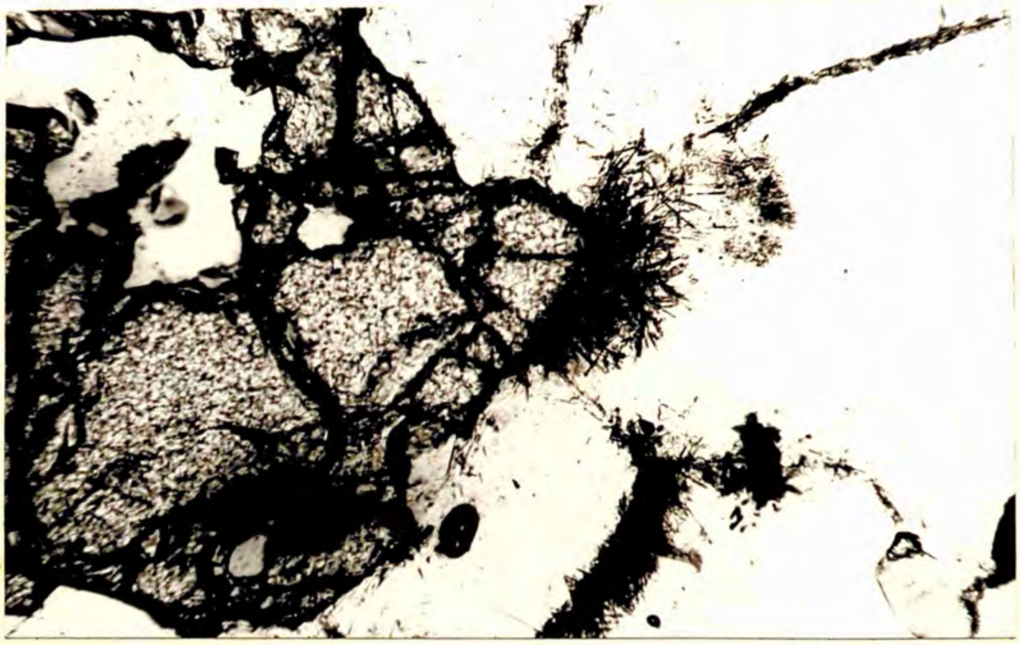


Plate 121. Fibrolite nucleating on garnet. Migmatized psammitic raft in Husfjord metagabbro. P.P.L. X 170.

Plate 122. Idioblastic and hypidioblastic sillimanite. Migmatized semi-pelitic raft in Husfjord metagabbro. P.P.L. X 95.

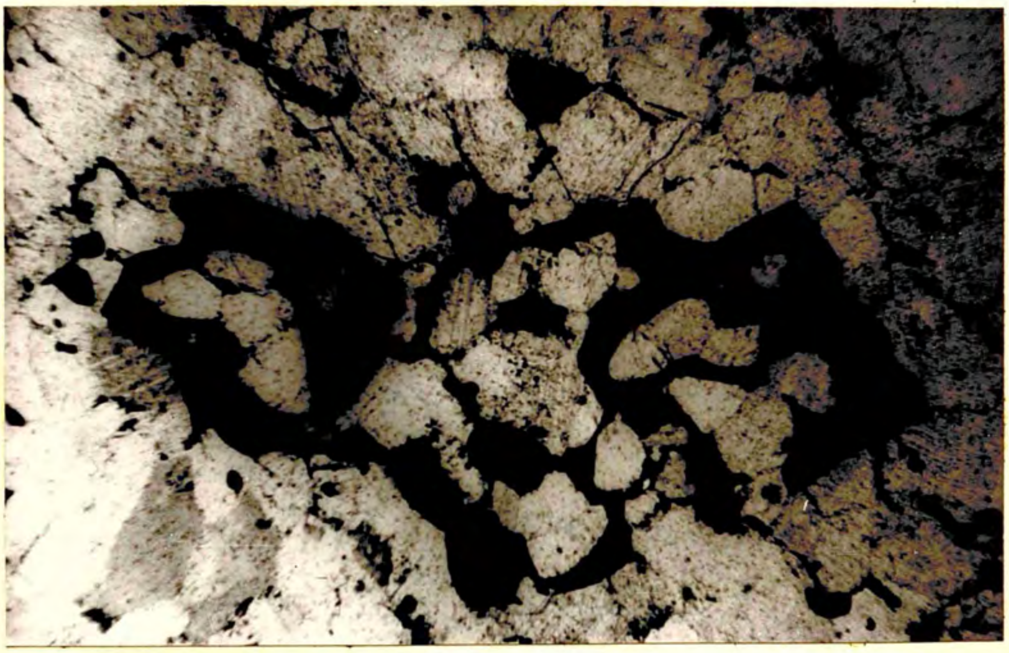


Plate 123. Ore enveloping plagioclase grains. Semi-pelitic raft in Husfjord metagabbro. P.P.L. X 85.

plagioclase and a little quartz are irregular and embayed, and the plagioclase is commonly cross-twinned. In places there is a suggestion of a relict subophitic texture.

HAVNEFJORD DIORITE

General

The Havnefjord diorite is a pyroxene-mica-diorite in which the principal pyroxene is hypersthene (Table 3). The rock has a fine-grained, subequigranular xenomorphic texture (Plate 124). It is generally non-porthyritic, but occasionally has a porphyritic texture in which feldspar phenocrysts tend to be aligned, forming a fluxion structure. The diorite is sometimes more porphyritic in its marginal zone, but even here the phenocrysts are never large. In the slightly coarser-grained parts of the diorite an incipient subophitic texture is developed in which broad plagioclase laths, now slightly recrystallized, are intergrown with hypersthene.

Hypersthene crystals are anhedral to subhedral and vary greatly in size, some of the larger ones being deformed. It is pleochroic from buff to pale green, sometimes strongly so.

Generally, hypersthene is the only pyroxene present, but occasionally it is accompanied by diopsidic clinopyroxene. This latter is pale green and non-pleochroic, and sometimes exhibits twin lamellae at an oblique angle to the cleavage. Quite frequently the diopside contains patches of hypersthene or appears to be intergrown with it. Occasionally, crystals which appear to be all diopside or all hypersthene when parallel to the E-W cross-wire (they have the same shade of green in this position), are seen to be partly diopside and partly hypersthene when parallel to the N-S cross-wire (when hypersthene is buff-coloured). This is presumably an exsolution phenomenon.

TABLE 3

MODAL ANALYSES OF HAVNEFJORD DIORITE

Mineral	36/92B	36/117C	36/126A	36/159B	36/233C	38/48A	05/58A
Plag.	37.56	59.19	49.58	67.61	21.92	43.34	67.14
K-feld.	37.26	3.46	4.72	0.85	49.66	6.20	5.53
Quartz	-	-	-	2.88	-	0.33	-
Hypers.	17.20	23.80	10.85	13.38	15.37	14.61	15.58
Clinop.	-	2.46	18.47	-	-	23.13	1.56
Bi.	3.56	6.60	12.43	7.77	5.81	6.09	1.74
Hornbl.	1.76	-	0.81	0.06	0.28	0.17	0.35
Ore	2.43	4.33	2.85	6.45	5.54	5.10	6.60
Apatite	0.23	0.26	0.30	0.88	1.39	0.73	1.25
Zircon	-	0.03	-	0.11	-	-	0.18
	100.00	100.13	100.01	99.99	99.97	99.71	100.02

Plate 124. Xenomorphic texture of Havnefjord diorite.
X-Pl. X 85.

A few of the hypersthenes are altering to a yellow-stained alteration product along cleavages and cracks. Occasionally it is fringed by a little green hornblende, but diopside, when it occurs, is more prone to alteration to hornblende.

Biotite forms fresh-looking laths, often broad and some times quite large. Frequently their cleavages are kinked (Plate 125). Some biotites are seen to be forming from the fragments of hornblende which fringe the hypersthene, and in places can be seen to be forming from it. In this case it is usually poikiloblastic (Plate 126), containing vermicules of quartz. This alteration may have taken place during late stages of the regional metamorphism. Biotite also contains inclusions of apatite and zircon, the latter being surrounded by pleochroic haloes.

Feldspars are anhedral and their grain boundaries are irregular but not highly sutured. The crystals are usually clear and are rarely altered, but a yellow iron-staining is common along grain boundaries and cracks in the feldspars. The yellow alteration product of some of the hypersthenes may be the source of this staining.

Plagioclase is andesine, normally of composition about An_{40}, but plagioclase as basic as An_{45} does occur in some specimens. Twin lamellae are usually finely spaced, and complex twinning is common. In some of the larger grains twin lamellae are often deformed (Plate 127). When plagioclase occurs as phenocrysts it is antiperthitic, and also contains inclusions of quartz and occasionally of small rounded grains of pyroxene.

Potash feldspar generally occurs as phenocrysts which are anhedral and slightly perthitic, containing small plagioclase patches. Sometimes

Plate 125. Flexed biotite in Havnefjord diorite.
P.P.L. X 85.

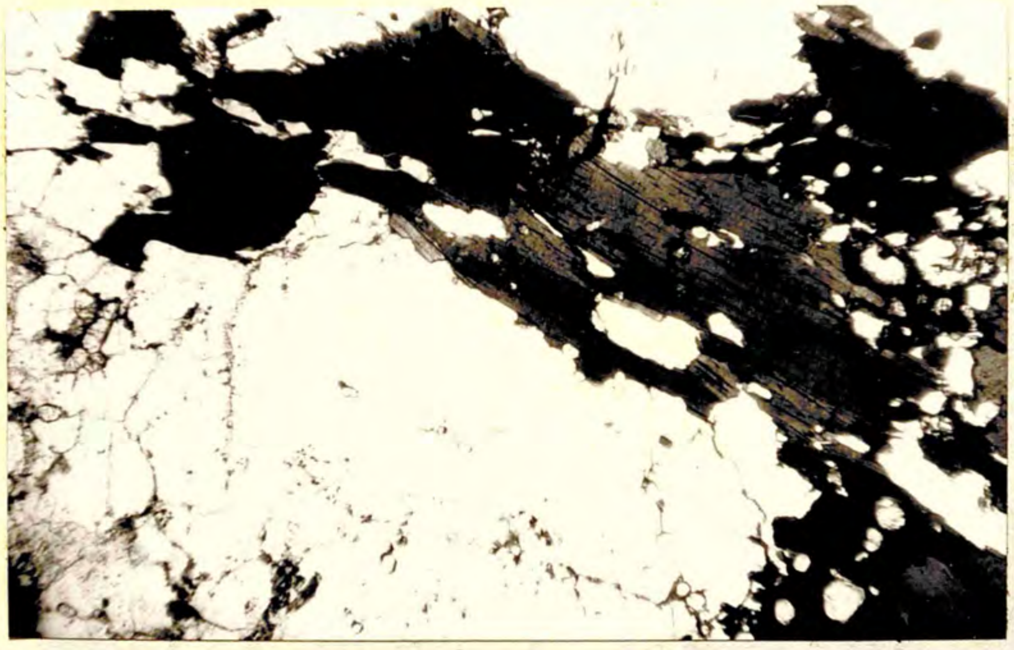


Plate 126. Poikiloblastic biotite in Havnefjord diorite.
P.P.L. X 85.

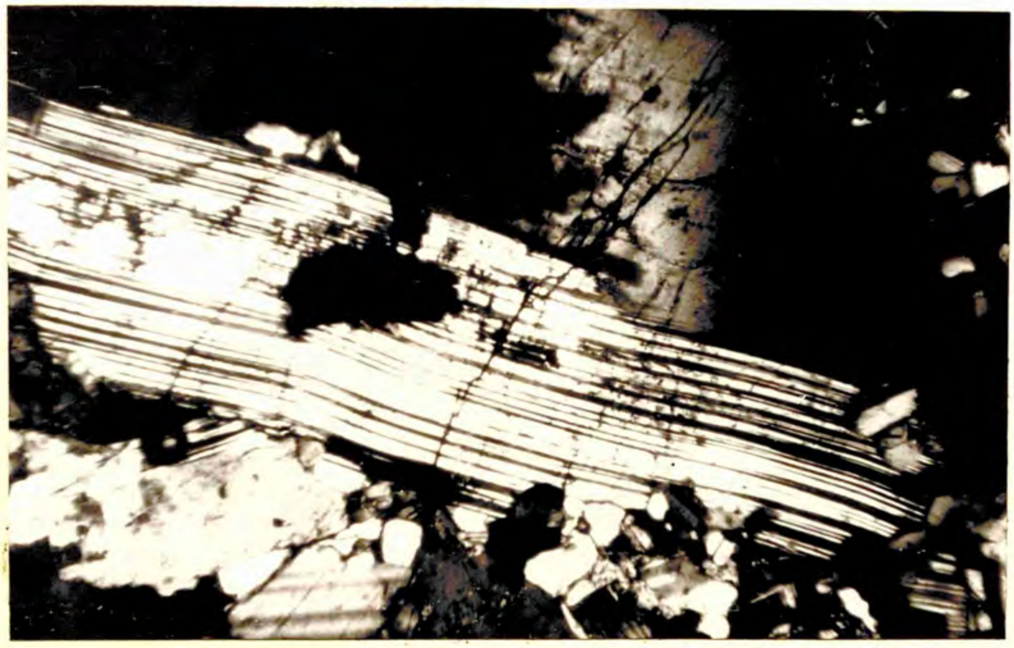


Plate 127. Deformed plagioclase in Havnefjord diorite. X-Pl. X 85.

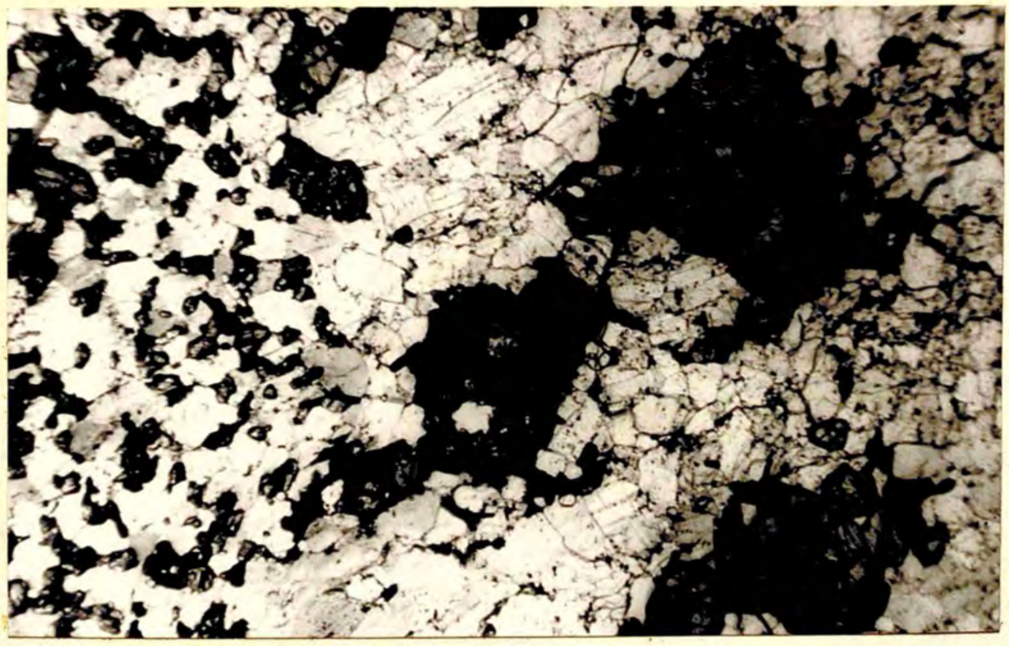


Plate 128. Hornfelsed Havnefjord diorite; note contact with coarse late pyroxene-mica-diorite. P.P.L. X 25.

potash feldspar crystals are bordered by myrmekite.

Ore, which is ilmenite, usually occurs interstitially and along grain boundaries, but it also forms rounded grains enclosed within gangue minerals.

At the contact with the late coarse-grained pyroxene-mica-diorites the Havnefjord diorite has become hornfelsed. The minerals have recrystallized to a fine-grained granoblastic pyroxene hornfels consisting of hypersthene, diopside and plagioclase (Plate 128). There are some diffuse biotites which are altering to pyroxene, but there are also some later, fresh-looking biotite laths which sometimes partly enclose pyroxene. The diffuse biotites are occasionally observed to be recrystallizing to fresh biotite. The coarse diorite is not chilled at the contact.

Pegmatites

There are a few acid pegmatites in the Havnefjord diorite, and these are usually fairly fine-grained. Quartz is abundant and forms an irregular mosaic, and there are a few grains of plagioclase, biotite, and muscovite. Potash feldspar is common and is perthitic, and exhibits microcline twinning, particularly in the neighbourhood of the plagioclase patches. Inclusions of quartz are also common within potash feldspar, and margins of the feldspars are often quite markedly sutured.

In many cases, the pegmatite has suffered slight shearing, and the margins of the large feldspars are often granulated especially where two large grains come into close proximity. They show strain shadows, and their granulated margins are somewhat sericitized. Quartz aggregates become elongated parallel to one another and the crystals of quartz are highly strained.

Metasomatic Veins

The Havnefjord diorite contains a number of dark, metasomatic veins similar to those in the Husfjord metagabbro. They are probably of the same generation, and clearly post-date the diorite.

These veins occur both in very narrow cracks and as thicker veins up to 2-3 mm. in thickness. In the latter case, the neighbouring diorite is affected for up to 5 mm. on either side of the vein.

The veins themselves are of fibrous yellowish-green actinolitic amphibole and closely resemble those in the Husfjord metagabbro. As the vein is approached, pyroxenes and biotites in the diorite are replaced by fibrous actinolite similar to that in the vein. Pyroxene alters more readily than biotite, and at the vein contact all the pyroxene has been replaced, but some diffuse biotite relics remain in the actinolite masses. Grains of ilmenite are progressively rimmed by sphene.

The textures at the contacts of the vein are the same as those in the Husfjord metagabbro, and actinolite crystals often cross the contact from the vein into the host (Plate 129).

Hybrids

The Havnefjord diorite contains large areas, usually several metres or tens of metres in length, in which it has a more basic character than normal. It is considered that in these areas the dioritic magma became contaminated by assimilation of Husfjord metagabbro. They occur more

Plate 129. Actinolite crossing metasomatic vein contact. Havnefjord diorite. P.P.L. X 95.

Plate 130. Xenomorphic texture of hybrid. X-Pl. X 85.

commonly near the margins of the diorite.

The hybrids have a subequigranular, non-porphyritic xenomorphic texture similar to that of the Havnefjord diorite (Plate 130, cf. Plate 124). In the coarser-grained parts an incipient subophitic texture is developed as in the case of the diorite, although this texture is not developed so well as in the Husfjord metagabbro.

In these hybrids the principal pyroxene is diopsidic clinopyroxene, as in the metagabbro ($z \wedge c = 44°$). Some of these clinopyroxenes contain schiller exsolution lamellae of ore orientated parallel to the cleavage, and also in directions perpendicular to one another oblique to the cleavage. Occasionally clinopyroxene has irregular patches of hypersthene, presumably due to exsolution.

The clinopyroxene is altering to green hornblende, and this alteration is most advanced around the pyroxene margins, although it occurs slightly in patches and along cleavages. It was seen that the pyroxenes in the Havnefjord diorite were sometimes fringed by green hornblende, and that the occasional diopsidic clinopyroxene was more readily altered than the hypersthene. The higher content of hornblende in the hybrids is probably explained by the relative abundance of the diopsidic clinopyroxene, which in turn reflects the more calcic character of the hybrids. However, the amphibolization is not so advanced as it is in the Husfjord metagabbro, and the hybrids have not suffered such an extensive metamorphism as the metagabbro.

Hypersthene also occurs frequently in the hybrids, and is markedly pleochroic from pinkish-buff to pale green. It is often altering to a brown alteration product along irregular fractures, and sometimes has a brown staining along cleavages. Furthermore, it is also beginning to alter to biotite.

There are two generations of biotite; the earlier biotites form small diffuse laths which are altering to hornblende, whereas the later biotites are larger and appear to be forming from hornblende and occasionally from hypersthene. Some of the late biotites are poikiloblastic, and the larger ones are often deformed, a characteristic of the Havnefjord diorite.

The only feldspar is plagioclase, which is generally basic andesine in the range An_{45-50}. However, there are also some plagioclase grains which are optically negative, and have diffuse twinning, and these are probably more sodic than An_{40}. Rarely, zoned plagioclase was found, in which the core had a composition of An_{45} whereas the margin was about An_{35}. Twinning in the plagioclase is often diffuse and irregular, and sometimes complex.

The amount of ore in the hybrids is similar to that in the diorite, but the ores include both oxides and sulphides as in the case of the metagabbro. The sulphide is pyrrhotite which occurs as small grains and is almost invariably altering to goethite, particularly around the margins. This alteration is due to oxidation and hydration of the sulphide. Ilmenite is the oxide and forms anhedral, rounded grains of varying size which occasionally enclose grains of pyrrhotite-goethite. It also occurs as lamellae in clinopyroxenes.

The hybrids have a zircon content comparable with that of the diorite.

Thus the petrography shows that the hybrids have characteristics which are intermediate between those of the Havnefjord diorite and the Husfjord metagabbro. It will be seen later that its chemical characteristics also lie between those of the diorite and the metagabbro.

Metasedimentary Rafts

Pelitic and Semi-pelitic Rafts.

These are principally quartz-mica-schists, garnet-mica-schists, and kyanite-mica-schists.

At the margins of the rafts a hornfelsic texture is developed, in which perthitic potash feldspars and quartz form polygonal grains, along the boundaries of which biotite laths lie forming a decussate pattern. Ores tend to occur as long ragged laths which also form a decussate texture (Plate 131).

In one pelite, kyanite forms hypidioblastic to allotrioblastic prisms which often occur in radiating aggregates. Biotite forms randomly-orientated laths which are usually quite diffuse, and interstitial quartz grains are common. Colourless garnet, which overgrows all these minerals, sometimes forms large allotrioblastic lobate crystals, but more commonly they form large aggregates of closely-spaced rounded grains incorporating quartz grains of the matrix. They also contain inclusions of diffuse biotite and kyanite as well as rutile, which is disseminated throughout the rock. The biotite and the kyanite are both seen to be contributing to the formation of the garnet. At the margins of garnet, kyanite prisms can be seen being made over into garnet, and occasionally traces of

Plate 131. Decussate ragged ore laths. Semi-pelitic hornfels in Havnefjord diorite. P.P.L. X 95.

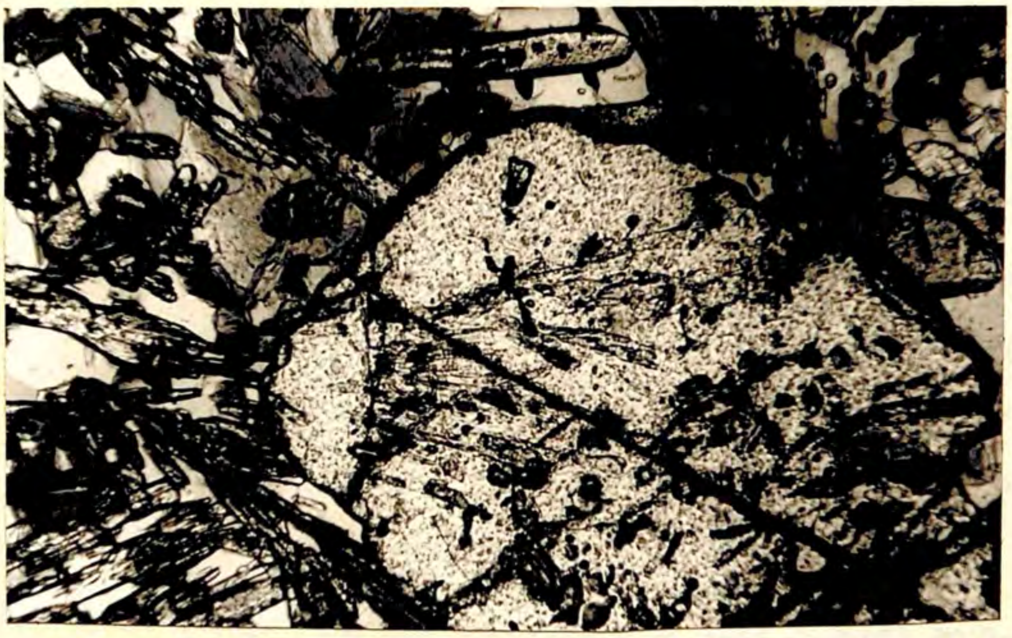


Plate 132. Radiating relict kyanite within garnet; note kyanite altering to garnet at margin of garnet. Pelitic raft in Havnefjord diorite. P.P.L. X 95.

radiating kyanite relics can be seen within the garnets (Plate 132).

In some pelites and semi-pelites, garnet has not been stable at the raft contacts, but has been extensively altered to fibrolitic sillimanite. In one case, garnets have been completely pseudomorphed by solid mats of cloudy fibrolite and ore. Rarely, small relict patches of garnet remain in the central parts of the ovoid pseudomorphs (Plate 133). These pseudomorphs are set in a matrix of fibrolite with small rounded grains of ore and quartz and a few diffuse orange-brown biotites.

In some different pelitic hornfelses, aggregates of biotite, fibrolite, and ore sometimes form ovoid knots. Many have garnet-like shapes (Plate 134) and appear to be pseudomorphed garnets. There are also some garnets which can be seen to be altering to fibrolite, and various stages in the progressive fibrolitization of the garnets may be observed (Plates 135 and 136). In the final stages, no garnet remains. These knots resemble those described by Pitcher and Read (1963), although in the examples under discussion there is no muscovite. Pitcher and Read also interpret their knots as pseudomorphed garnets, and in some cases, garnet was regenerating as small grains. In the present case, no secondary garnet is regenerating, but the relics of garnet in some of the knots indicates beyond doubt that they are pseudomorphed garnets. It follows from this that the grade of the regional metamorphism had dropped beneath the garnet isograd by the end of the hornfelsing. Further discussion of this appears on p.297 .

In some hornfelses, diffuse orange-brown biotites are being replaced by radiating and fan-like aggregates of fibrolitic sillimanite. Occasionally, pelitic hornfelses contain a fair amount of corundum (Plate 137), indicating a high alumina content.

Plate 133. Fibrolite pseudomorphic garnet; note relict garnet. Pelitic hornfels raft in Havnefjord diorite. P.P.L. X 25.

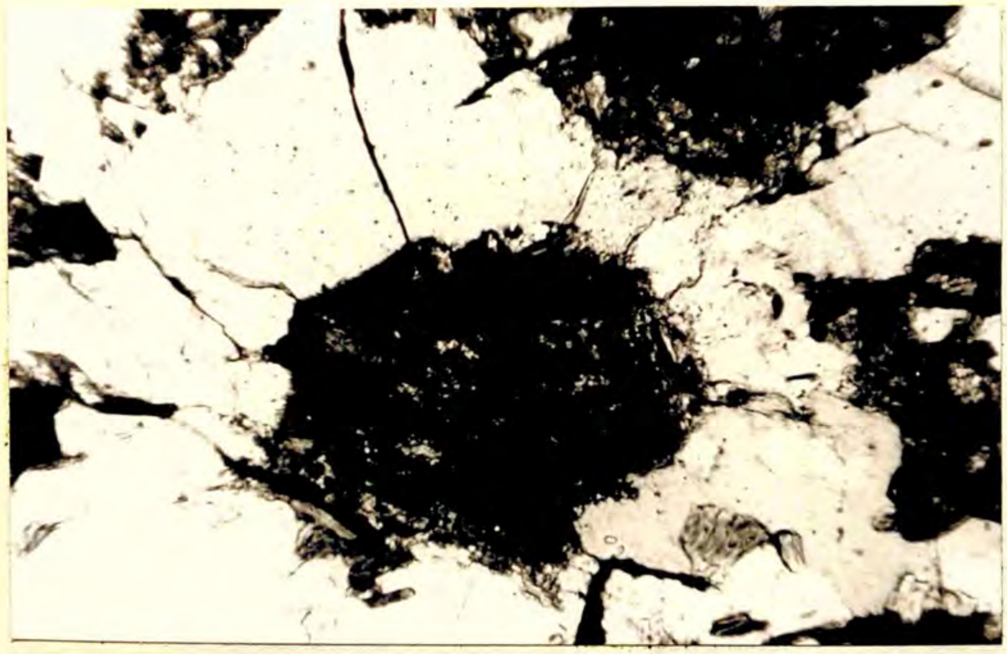


Plate 134. Biotite-fibrolite-ore knots. Pelitic hornfels raft in Havnefjord diorite. P.P.L. X 95.

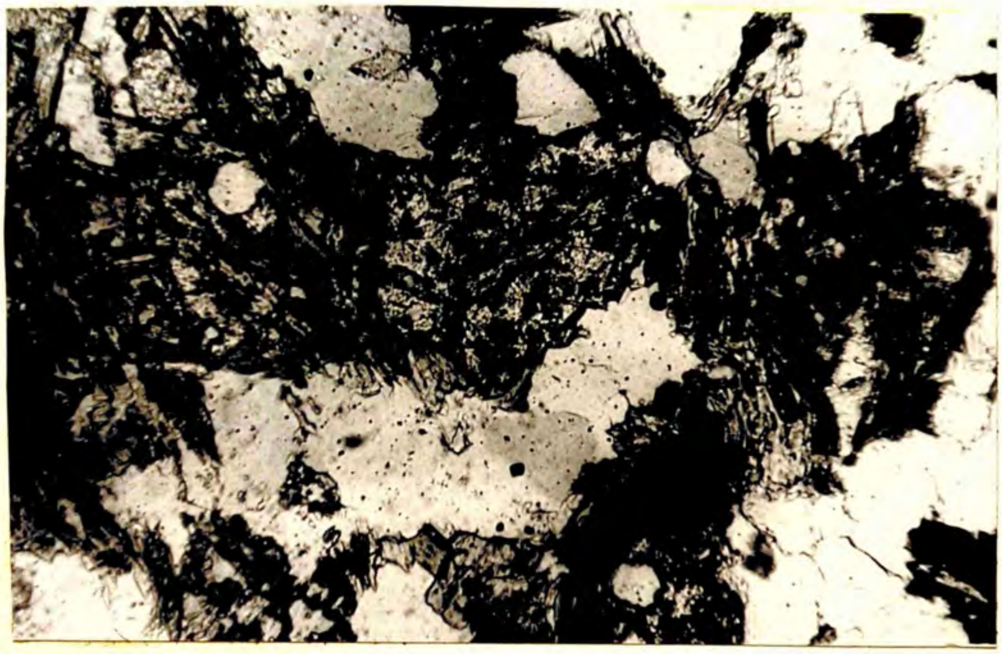


Plate 135. Garnet beginning to alter to fibrolite and biotite. Pelitic hornfels raft in Havnefjord diorite. P.P.L. X 95.

Plate 136. Garnet further altered to fibrolite and biotite. Pelitic hornfels raft in Havnefjord diorite. P.P.L. X 95.

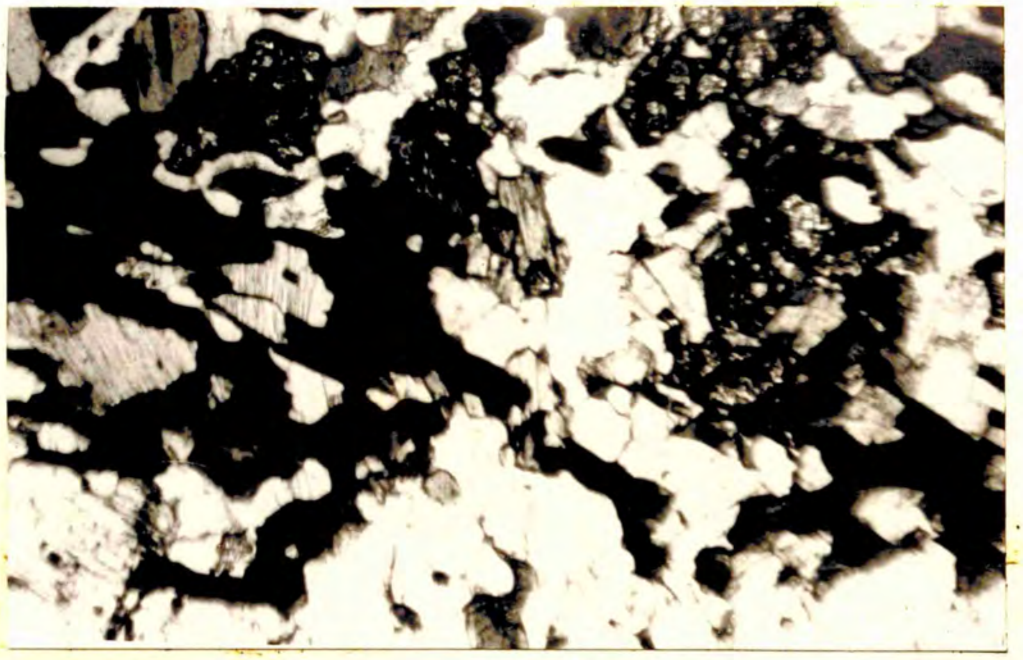


Plate 137. Corundum in pelitic hornfels in Havnefjord diorite. P.P.L. X 85.

Plate 138. Margin of perthite porphyroblast. Pelitic hornfels in Havnefjord diorite. X-Pl. X 85.

Sometimes the pelites have been slightly feldspathized and contain occasional porphyroblasts of braid-perthite containing a few inclusions of quartz. Their margins are irregular, lobing into the host, and there are numerous small inclusions of biotite at the margins (Plate 138).

Metalimestone Rafts

These are coarsely-recrystallized marbles in which the calcite crystals form a subequigranular mosaic with fairly straight, smooth grain boundaries. Apatites, rounded diopsides, some potash feldspars, and a little ore occur as interstitial grains.

In the metalimestones there are a few thin calc-silicate bands which consist of diopside, andesine, ore, and occasional zircons.

In one raft there is a feldspathic patch of which only half a square metre is exposed, and which may represent a leucocratic semi-pelitic area in the limestone. It consists mainly of potash feldspar, which is hair-perthite, and which forms a subequigranular mosaic of polygonal grains. Other minerals include diopsidic clinopyroxene, green hornblende, and biotite, and these occur as isolated grains surrounded by the mosaic.

Along the boundaries between the potash feldspar grains there are bands having a refractive index higher than that of potash feldspars. A few of these bands show very vague albite twinning perpendicular to the grain boundaries, and they probably consist of sodic plagioclase. A few of the potash feldspar megacrysts contain small inclusions of potash feldspar, and these inclusions are also rimmed by sodic plagioclase. It is clearly not a phenomenon that can be attributed to intergranular fluids since it is restricted to potash feldspar/potash feldspar boundaries, even when this occurs within a megacryst. It is probably a recrystallization feature involving only potash feldspar, presumably brought about by thermal metamorphism.

Calc-silicate hornfels Rafts

The principal mineral in these rocks is diopside, which generally forms a fine-grained granular texture with plagioclase. Some bands contain tremolite, which is often irregular and poikiloblastic, and forms a decussate texture. In one band the hornfelsic texture is overgrown by poikiloblastic cummingtonite with the following pleochroism:

X pale straw

Y khaki

Z greenish-khaki.

Basic Hornfels Rafts

There are a few fine-grained granoblastic pyroxene hornfelses consisting mainly of rounded grains of hypersthene, diopside, and plagioclase, with a little ore. In some cases, randomly orientated diffuse biotite laths are altering to pyroxene.

Occasionally bands with a different mineralogy occur, and the contacts between the bands become gradational, presumably due to migration of constituents during metamorphism. For example, one band consists of pale green poikiloblastic actinolites containing inclusions of diopside and plagioclase, and this grades into the pyroxene hornfels by the gradual loss of actinolite and a corresponding gain in hypersthene. In the mixed zone, ores form ragged laths which produce a decussate texture.

In some of the pyroxene hornfelses there are long poikiloblastic biotites which enclose grains of pyroxene and plagioclase. These biotites have a preferred orientation, and must have been formed in post-hornfelsing regional metamorphism.

Towards the contact with the diorite, the hypersthenes in a few pyroxene hornfelses become large and poikiloblastic. Sometimes they are so riddled with inclusions that they appear as separate grains having the same optical orientation (Plate 139). Wells (1951) has described similar sieve-textured porphyroblastic hypersthenes in sedimentary xenoliths in the hypersthene-gabbro of Ardnamurchan.

In the contact zone the texture of these hornfelses begins to take on an igneous aspect. Although there are small patches which still retain a granular hornfelsic texture with sub-rounded hypersthene grains (Plate 140), much of the rock has large poikiloblastic hypersthenes containing randomly-orientated plagioclase grains (Plate 141). The plagioclase is labradorite having a composition of about An_{60}. The hypersthenes and plagioclases are often intergrown in a texture resembling a subophitic texture. Since it is the product of metamorphism, it is proposed to term the texture subophitoblastic (Plates 142 and 143). When the rock is viewed under low power, the predominance of plagioclase over pyroxene gives the rock a diabasic texture, and the term diabasoblastic is used here to indicate its metamorphic origin. It is shown in Plate 144.

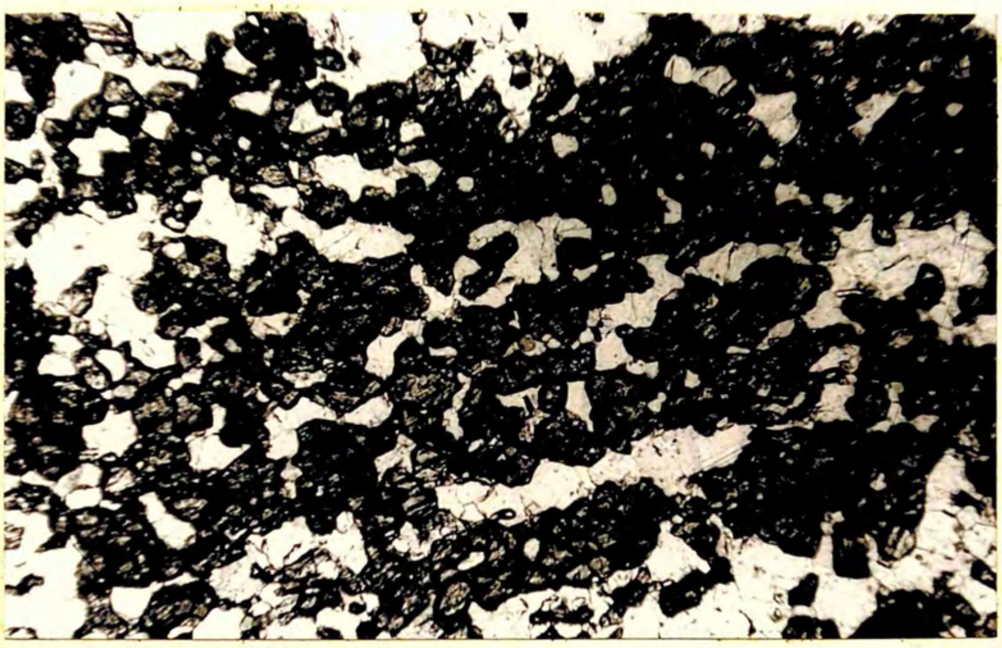


Plate 139. Poikiloblastic hypersthene. Hornfels raft in Havnefjord diorite. P.P.L. X 95.

Plate 140. Relict granular hornfelsic texture. Pyroxene-hornfels raft in Havnefjord diorite. P.P.L. X 85.

Plate 141. Poikiloblastic hypersthene. Hornfels raft in Havnefjord diorite. P.P.L. X 85.

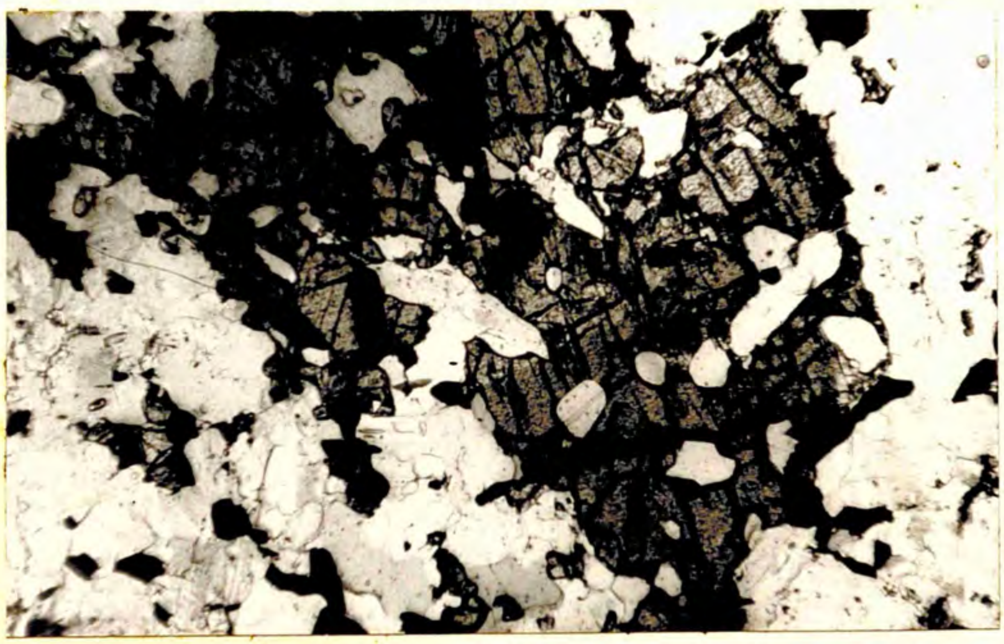


Plate 142. Ophitoblastic texture. Pyroxene-hornfels raft in Havnefjord diorite. P.P.L. X 85.

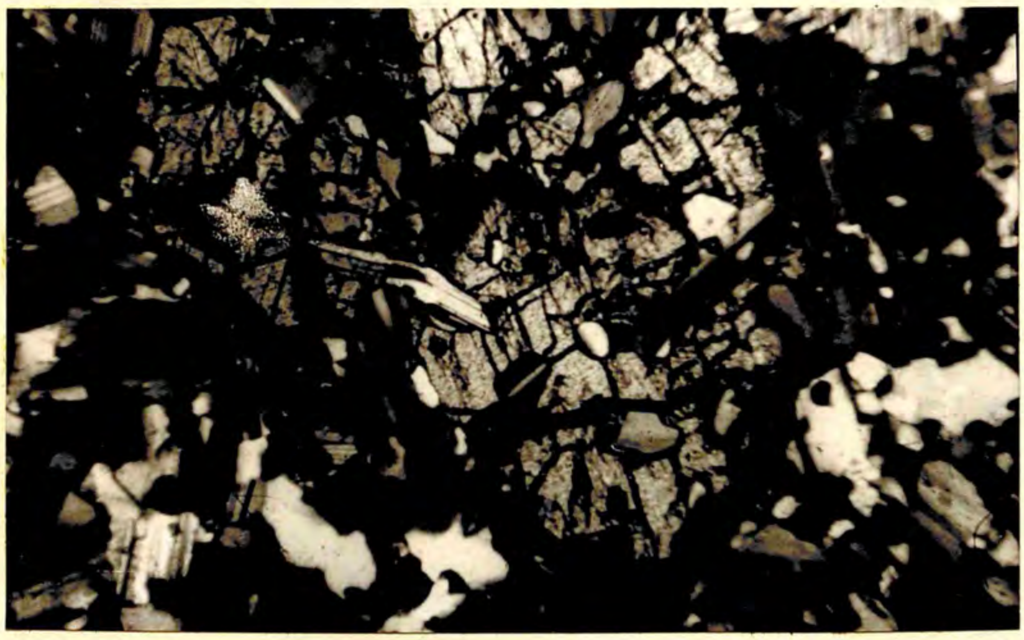


Plate 143. Ophitoblastic texture. Pyroxene-hornfels raft in Havnefjord diorite. X-Pl. X 85.

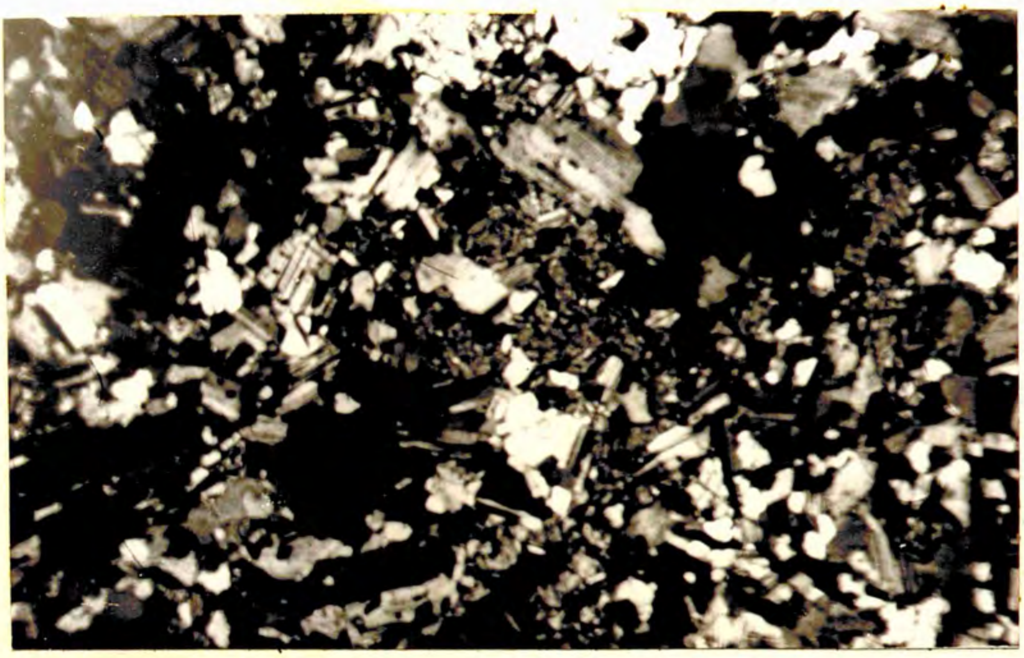


Plate 144. Diabasoblastic texture of pyroxene-hornfels. Raft in Havnefjord diorite. X-Pl. X 25.

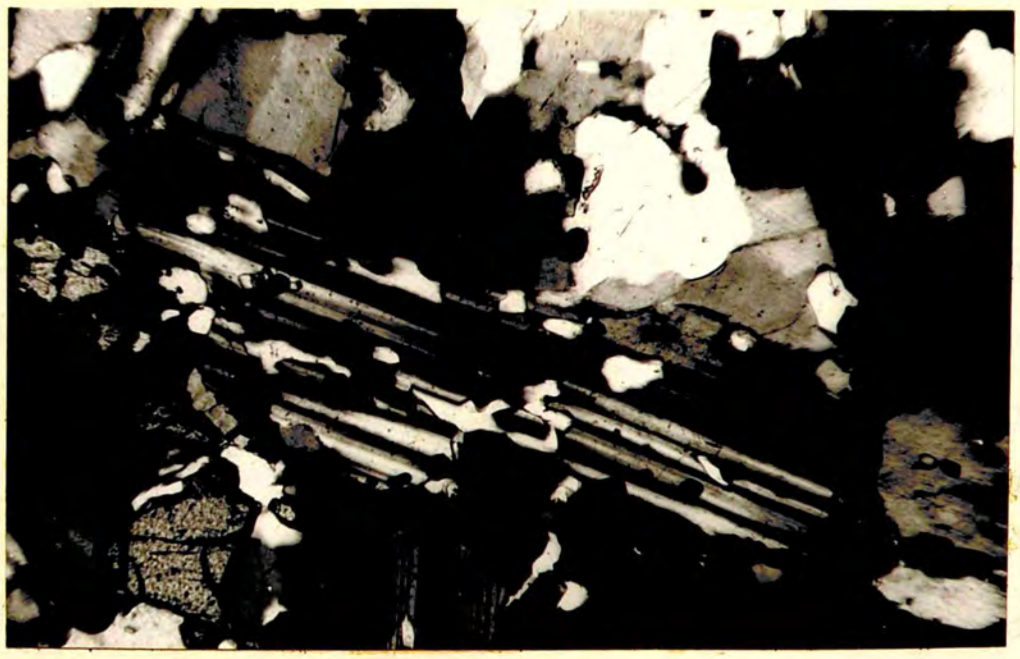


Plate 145. Vermicular plagioclase. Pyroxene-hornfels raft in Havnefjord diorite. X-Pl. X 95.

Plagioclase is commonly poikiloblastic, containing large numbers of quartz blebs and vermicules (Plate 145), although quartz also occurs interstitially. It appears that the quartz was exsolved from the plagioclase, presumably as the plagioclase became more calcic on recrystallization during metamorphism. Silica is clearly being expelled, and the hornfelses are becoming more basic.

Mobilized Hornfels Rafts

At the margin of a metasedimentary raft on the western side of Havnefjord the hornfelses have been mobilized and disrupted by the dioritic magma.

In the semi-pelitic hornfelses, there is a relict schistosity delineated by diffuse biotites and poikiloblastic kyanite prisms. Interstitial grains are of quartz and fairly basic labradorite, An_{60}. Towards the contact, kyanite recrystallizes to fan-like aggregates of fibrolitic sillimanite (Plate 146).

In the basic hornfelses, hypersthene is the predominant mineral, often intergrown with plagioclase and in one case poikiloblastic tremolitic amphibole, the latter probably due to a later low-grade regional metamorphism. Hypersthene is often cracked and is altering along these cracks and cleavages to a brown fibrous micaceous mineral. One basic band has an igneous appearance in hand sample, but in thin section is seen to consist almost entirely of large poikiloblastic orthopyroxene grains (Plate 147). These are colourless but are optically negative, and are presumably magnesian hypersthene. They are randomly orientated,

Plate 146. Kyanite altering to fibrolitic sillimanite. Pelitic hornfels raft in Havnefjord diorite. P.P.L. X 95.

Plate 147. Large poikiloblastic orthopyroxene; one is in extinction. Hornfels raft in Havnefjord diorite. X-Pl. X 25.

and inclusions are mainly of rounded diopside grains. When biotite occurs it does so as rather ragged laths. Ore, which is pyrrhotite, also frequently occurs as decussate ragged-edged laths (Plate 148).

The light, creamy-yellow bands which occur around the margin of most of the hornfels blocks consist of quartzo-feldspathic material. The principal mineral is labradorite, quartz occurring as interstitial grains and small blebs and vermicules in the plagioclase (Plate 149). It appears that the formation of the quartz is a result of the exsolution of silica from the plagioclase. In one case, stages in the exsolution can be observed. Some of the plagioclase grains have diffuse vermicular patches which appear to be less basic than the rest, and may indicate where silica is being concentrated prior to exsolution (Plate 150). In nearby plagioclase grains exsolution has taken place, and there are vermicular inclusions of quartz, which in morphology resemble the diffuse vermicular patches in the other plagioclase grains (Plate 151). Thus it would seem that the plagioclase is becoming more basic in composition, and that the excess silica that cannot be incorporated into the more basic plagioclase is exsolved as quartz inclusions. Leake and Skirrow (1960) have reported increases in the anorthite content of plagioclase in pelitic hornfelses in the Cashel-Lough Wheelaun intrusion in County Galway, due to the extraction of silica from the hornfelses by the magma.

In places recrystallization has been so complete that the texture approaches that of an igneous rock. Large allotrioblastic porphyroblasts of poikiloblastic plagioclase begin to form a xenoblastic texture (Plate 152).

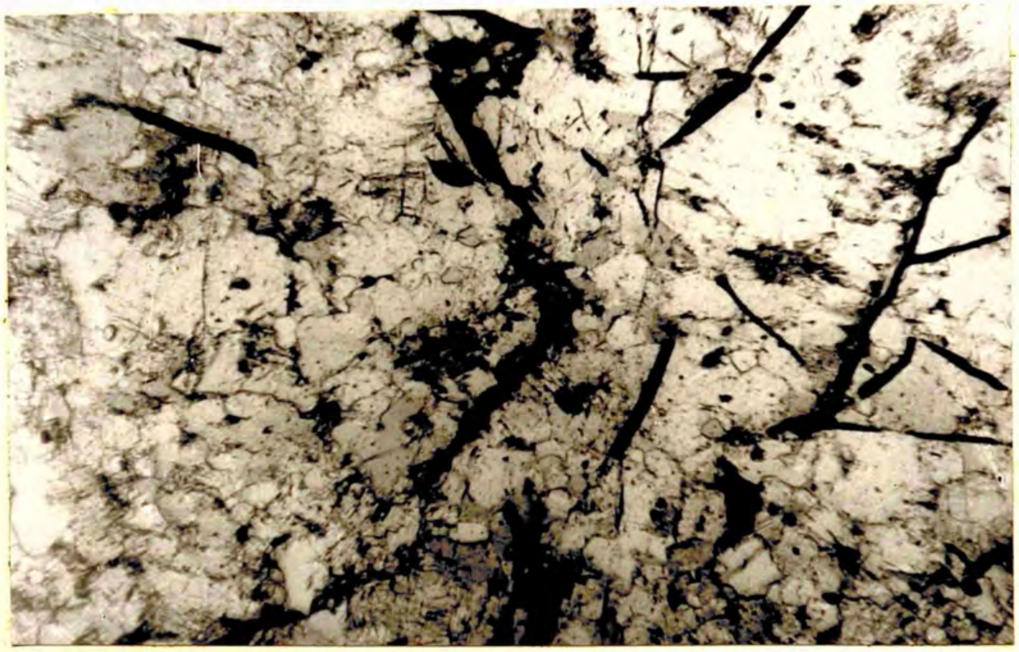


Plate 148. Decussate ragged ore laths. Hornfels raft in Havnefjord diorite. P.P.L. X 85.

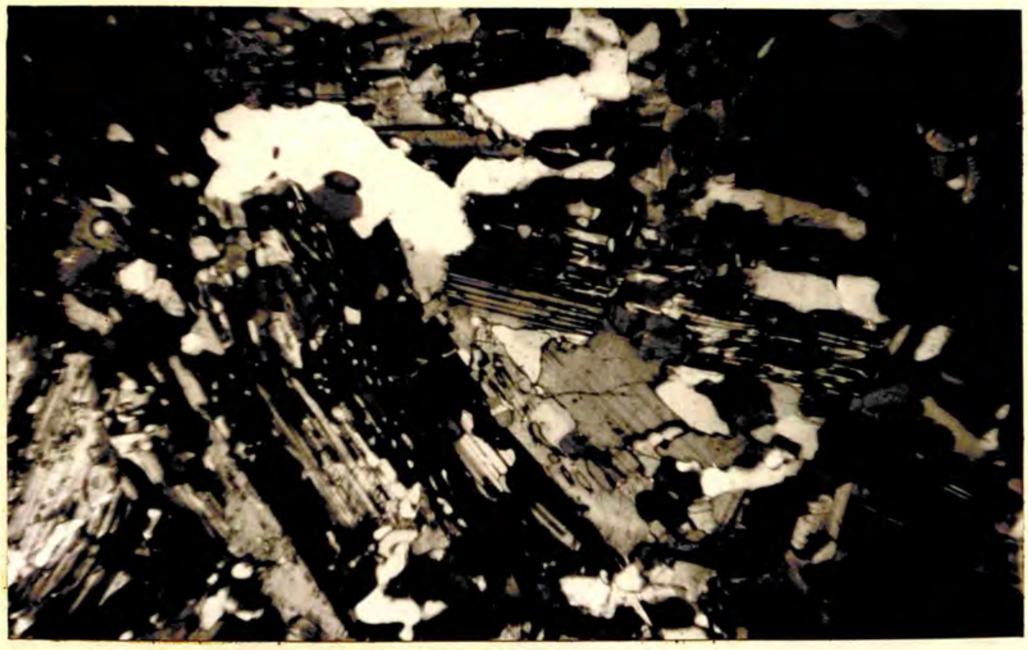


Plate 149. Interstitial and vermicular quartz. Hornfels raft in Havnefjord diorite. X-Pl. X 85.

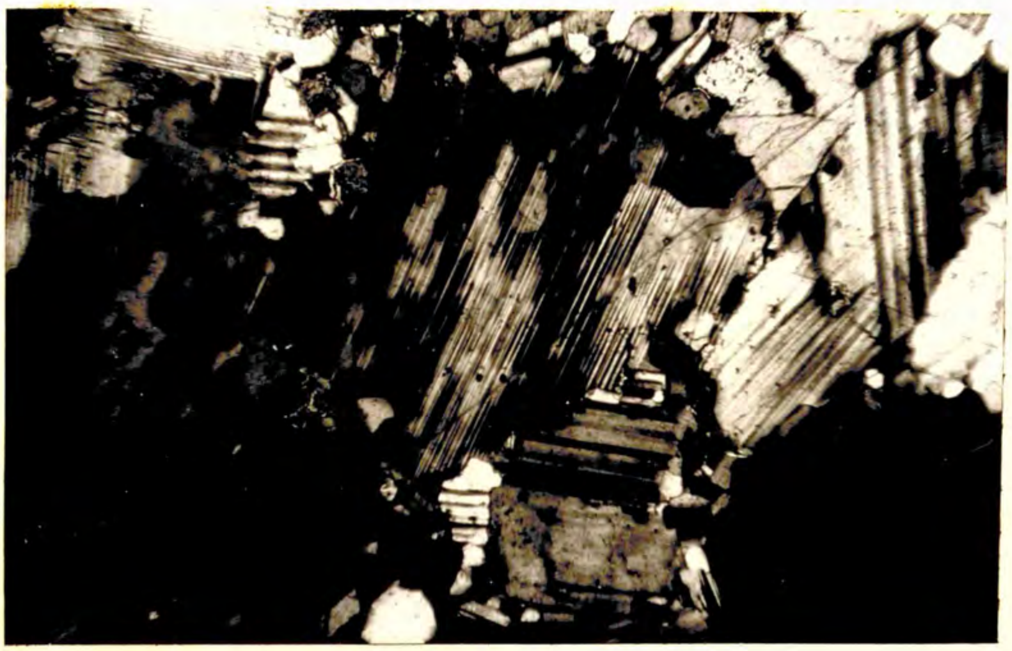


Plate 150. Diffuse vermicular patches in plagioclase. Hornfels raft in Havnefjord diorite. X-Pl. X 85.

Plate 151. Vermicular quartz in plagioclase. Hornfels raft in Havnefjord diorite. X-Pl. X 85.

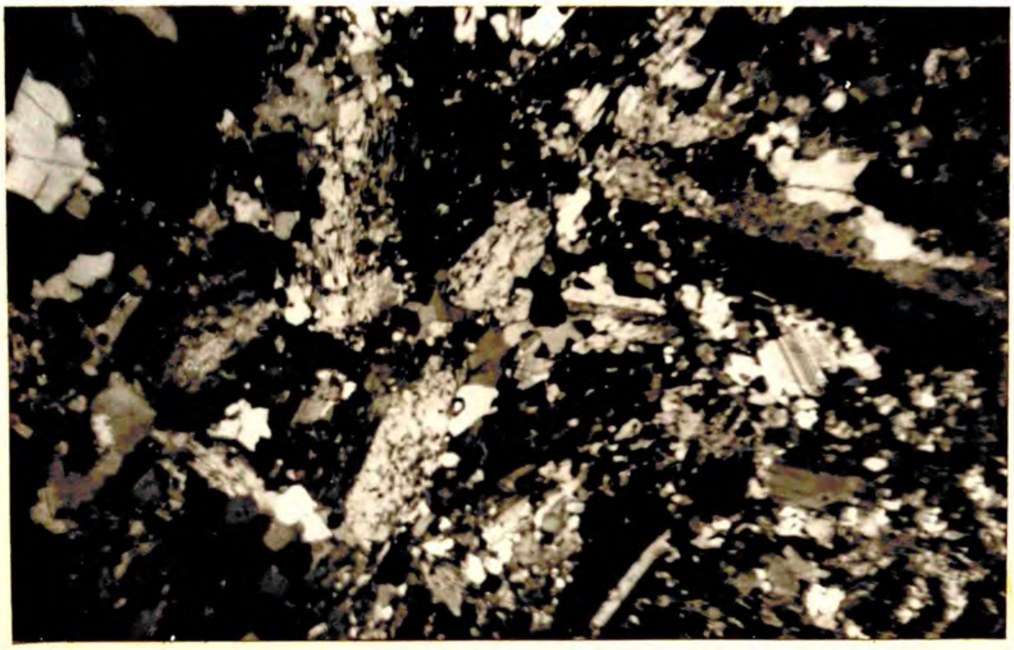


Plate 152. Large plagioclase porphyroblasts. Mobilized hornfels raft in Havnefjord diorite. X-Pl. X 25.

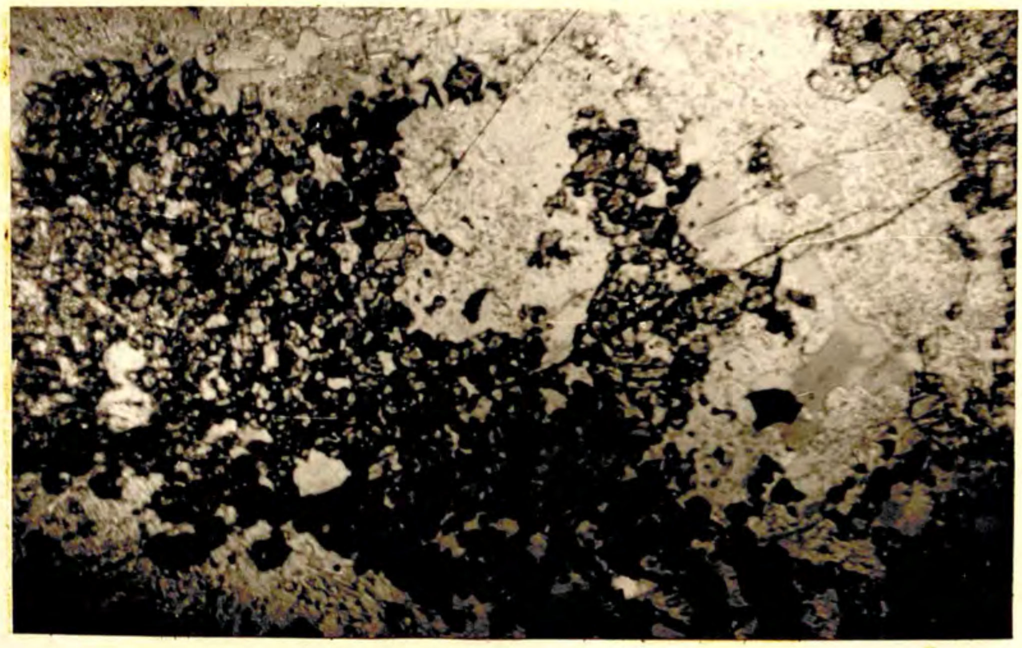


Plate 153. Disruption of granular hornfels by quartzo-feldspathic band. P.P.L. X 25.

This texture suggests that these zones have come very near to being mobile. There is, however, further evidence that mobility was actually achieved in some cases. Textures indicate that some of the basic hornfels bands have been disrupted and veined by these quartzo-feldspathic bands which must have been mobile (Plate 153). It appears that the temperature during the hornfelsing was sufficient to cause mobilization of the quartzo-feldspathic constituents, whilst the more basic constituents remained solid. In places the hornfelses and the diorite have been net-veined by quartzo-feldspathic material consisting of quartz and a little andesine, which form an irregular inequigranular mosaic. It is likely that the source of these veins may be in the quartzo-feldspathic mobilized hornfelses.

The possibility of the mobilization of quartzo-feldspathic material by selective anatexis of metasediments due to heat, probably in the presence of volatiles, is discussed on p.290 .

LATE DIORITES

Pyroxene-mica-diorites

General

This coarse-grained pyroxene-mica-diorite has a porphyritic xenomorphic texture (Plate 154). The principal pyroxene is hypersthene (Table 4), and this is often the only pyroxene present.

The hypersthenes are pleochroic from pinkish-buff to pale green, and the pleochroism is sometimes very strong. They are anhedral, and commonly contain faint yellow-brown schiller inclusions parallel to the cleavage traces in longitudinal sections and also subperpendicular to this direction. There is also a slight alteration along irregular transverse cracks. The occasional diopside grains are often fringed by a little green hornblende.

Biotite sometimes envelops pyroxene and in places can be seen to be forming from it (Plate 155). It is not possible, however, to tell whether or not all the biotite has formed from pyroxene. Biotite crystals are hypidioblastic to allotrioblastic, are sometimes large, and are often deformed. They are commonly poikiloblastic, containing inclusions of pyroxene, plagioclase, apatite, and zircon. Quartz is common both as blebs and vermicules in the biotite, and as interstitial grains in the biotite aggregates.

Feldspars form an inequigranular mosaic with irregularly shaped grains, although the margins are not strongly sutured. Feldspars are usually clear, but yellowish iron staining is common, particularly along grain boundaries.

TABLE 4

MODAL ANALYSES OF PYROXENE-MICA-DIORITES

Mineral	36/122B	38/18B	38/38A	38/48E	36/237A[1]	36/301A[1]	36/301B[1]
Flag.	50.66	22.78	43.12	51.56	54.70	53.60	64.20
K-feld.	21.30	50.51	37.78	35.26	32.70	29.65	17.40
Quartz	1.36	-	1.23	-	-	-	-
Myrmekite	-	-	1.92	0.80	-	-	-
Hypers.	18.66	13.28	8.92	7.66	7.09	13.06	12.89
Biotite	3.60	9.00	4.93	2.30	3.51	1.54	1.75
Hornbl.	0.53	0.49	0.14	0.16	0.66	0.62	0.18
Ore	3.90	3.33	1.98	2.30	1.49	1.46	2.73
Apatite	-	0.68	0.02	-	0.22	0.31	0.45
Zircon	-	-	0.05	-	0.02	-	0.07
	100.01	100.07	100.09	100.04	100.39	100.24	99.76

1 : Modal analysis carried out by macro point counting techniques - see Appendix.

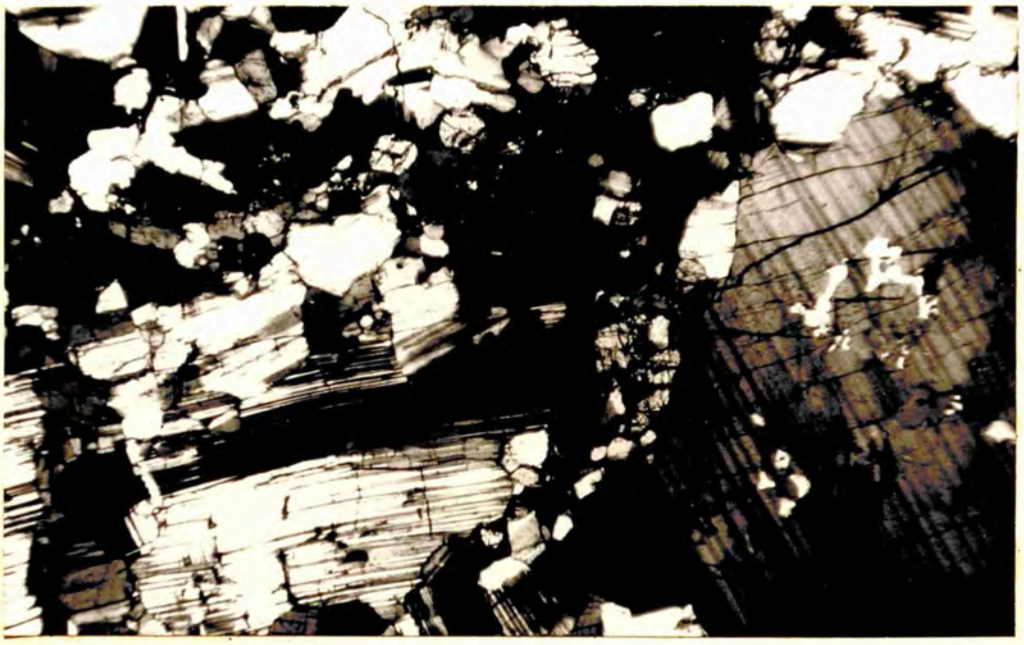


Plate 154. Texture of coarse pyroxene-mica-diorite. X-Pl. X 25.

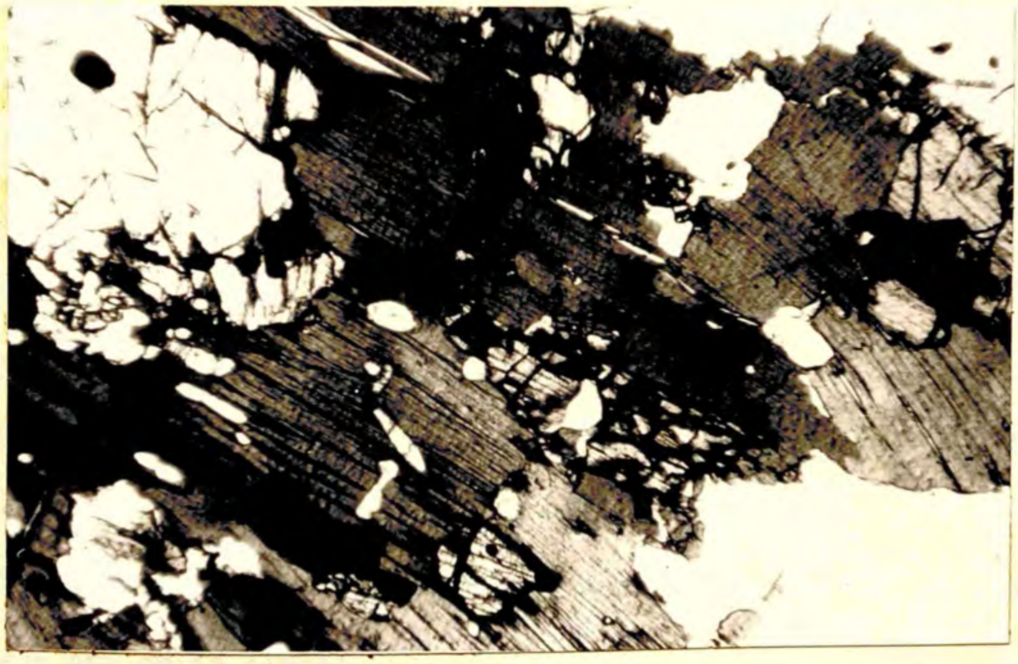


Plate 155. Hypersthene altering to biotite. Pyroxene-mica-diorite. P.P.L. X 85.

Phenocrysts are generally randomly orientated, but occasionally they have a preferred orientation forming a crude fluxion structure. Both potash feldspar and plagioclase form anhedral phenocrysts; the potash feldspar ones are perthitic with small plagioclase inclusions, and the plagioclase ones are usually antiperthitic, sometimes strongly so. The perthitic or antiperthitic inclusions are irregular in shape, and the antiperthites often include irregular quartz grains. Plagioclase is generally irregularly and complexly twinned, and pericline twins are common. The composition of the plagioclase phenocrysts appear to be the same as those in the matrix, i.e. andesine around An_{40}. There are a few large areas of plagioclase mosaic which represent recrystallized phenocrysts.

The potash feldspar phenocrysts are generally lobed by myrmekite in which the plagioclase is about An_{30}. The potash feldspar has an unusually low 2V; measurements by U-stage techniques vary from 40° to 52°, the majority being between 40° and 44°.

These potash feldspars have an Or-content of about 65% (determined by fluorescent X-ray spectroscopy - see Petrochemistry, p.247). This, in combination with the average optic angle value, plots as a feldspar just half-way between the orthoclase-microperthite curve and the sanidine-cryptoperthite curve on the alkali feldspar graphs of Tuttle (1952) and MacKenzie and Smith (1955).

Tuttle and Keith (1954) have found perthitic potash feldspars in this range in the Beinn an Dubhaich Granite of Skye. To account for this they consider that the soda phase has only partly inverted to the low-temperature form and that a part has remained as high-temperature albite, since the value of the optic angle of an alkali feldspar of a certain composition is a function of the ratio of high- to low-temperature albite. MacKenzie and Smith (1955), however, consider that the value of the optic angle may also depend upon the structural state of the potash phase, and that it increases as the triclinicity increases.

Emeleus and Smith (1959) describe alkali feldspars from Slieve Gullion which also fall between the high- and low- temperature curves. They consider that this intermediate position indicates that the rocks probably cooled at intermediate depths where cooling time was not sufficient to allow inversion to the low-temperature state to be completed.

The alkali feldspars in the pyroxene-mica-diorites under present discussion fall in a position intermediate between those of Tuttle and Keith, and those of Emeleus and Smith. Their composition is similar to the former, and their 2V values are similar to the latter. It is probable that they are also in an intermediate state between high- and low-temperature phases, and may indicate that inversion from the high-temperature state to the low-temperature state had not beeen completed.

Ore is principally ilmenite, which forms fairly large irregular grains, often interstitial and along grain boundaries. Goethite also occurs along cracks in the gangue minerals, and is obviously late. Presumably it is a secondary product formed by hydration of an oxide, but there is no evidence as to what the original ore was.

The remaining accessory minerals in these diorites are apatite and zircon, the latter being quite common.

Xenoliths

The coarse pyroxene-mica-diorites contain numerous xenoliths, mainly pelitic and semi-pelitic, but also basic hornfelses. They vary in size from large blocks to small fragments capable of being seen complete in a single thin section, and many are becoming assimilated by the diorite.

Pelitic and Semi-pelitic Xenoliths

In these, a granoblastic hornfelsic texture is developed in which polygonal grains of potash-feldspar, usually hair-perthitic, have small ragged biotite laths along their grain boundaries forming a decussate texture. However, a relict schistosity or gneissic texture is often preserved by bands rich in biotites alternating with feldspathic bands. Towards the contacts, biotite becomes replaced by sillimanite, usually fibrolitic, and ore (Plate 156), but in some pelites large fresh poikiloblastic biotites are probably a later regional metamorphic effect. In many of the hornfelses there is considerable iron staining of the feldspars in cracks and along grain boundaries, particularly at the margins of the xenoliths. At the contact of one hornfels against the diorite small hypersthenes form, apparently at the expense of some of the biotites. At this contact there is considerable iron staining, and this excess of iron may have resulted from the alteration of biotite into hypersthene. In this particular case, as the contact is crossed the hypersthenes and feldspars become coarser and the amount of biotite decreases until the rock resembles the diorite. The contact is completely

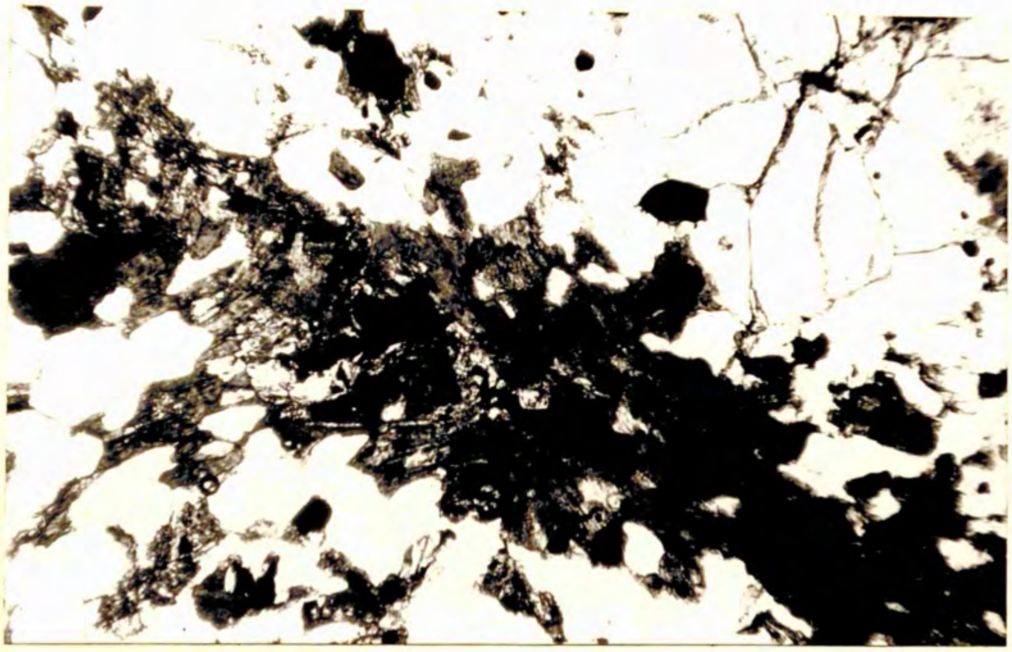


Plate 156. Biotite altering to fibrolite and ore. Hornfels xenolith in pyroxene-mica-diorite. P.P.L. X 85.

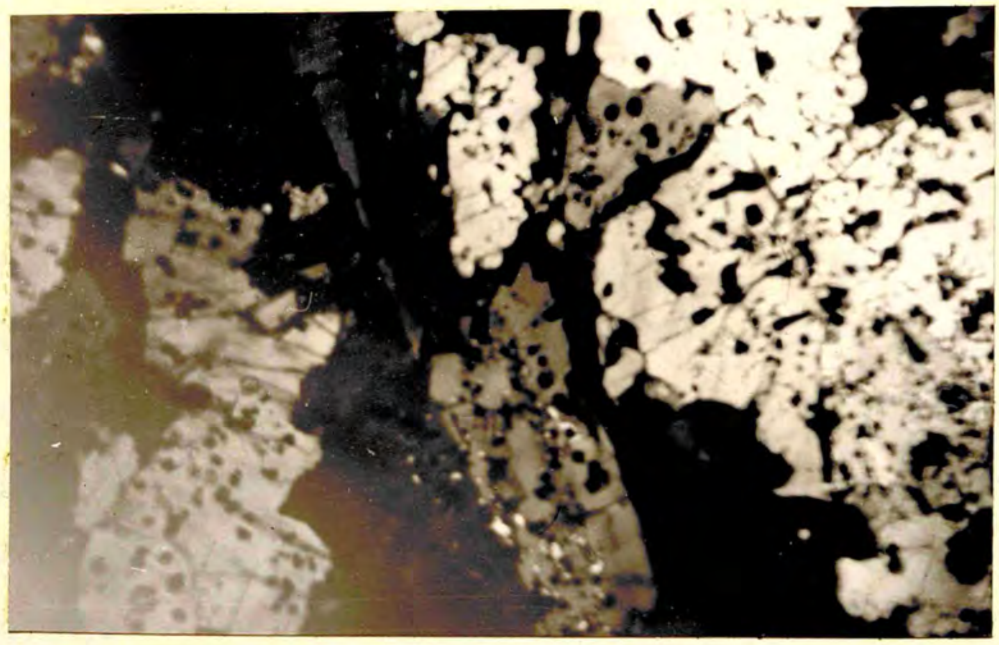


Plate 157. Feldspar porphyroblast. Hornfels xenolith in pyroxene-mica-diorite. X-Pl. X 25.

gradational, and it appears that the hornfels is being made over into diorite.

Small green spinel grains are common in some of the pelites, and occasionally they occur in clusters enclosed in feldspar and biotite. Corundum also occurs in some pelites, indicating a low silica content for these pelites.

In one pelite, large allotrioblastic buff-coloured garnets have formed from biotite. Inclusions of ore, usually very fine-grained, are common in the garnets, and may result from the alteration of biotite. The rock has been intensely deformed after the growth of the garnets, which are severely shattered. Feldspars are strained and have become granulated at their margins.

A few of the pelites have been feldspathized, in which large hair-perthite porphyroblasts have overgrown the hornfelsic texture preserved by small biotite laths (Plate 157). Many of the feldspars contain clouds of finely-disseminated ore.

There are also some large fresh biotites which are obviously late as they lie along the boundaries between the feldspar megacrysts. These biotites are sometimes flexed, and almost invariably contain inclusions of apatite and green spinel, the latter often being large (Plate 158). Spinel is very abundant in these rocks and is disseminated throughout (Plate 159).

Basic Hornfels Xenoliths

It is not always possible to tell whether the basic hornfelses are

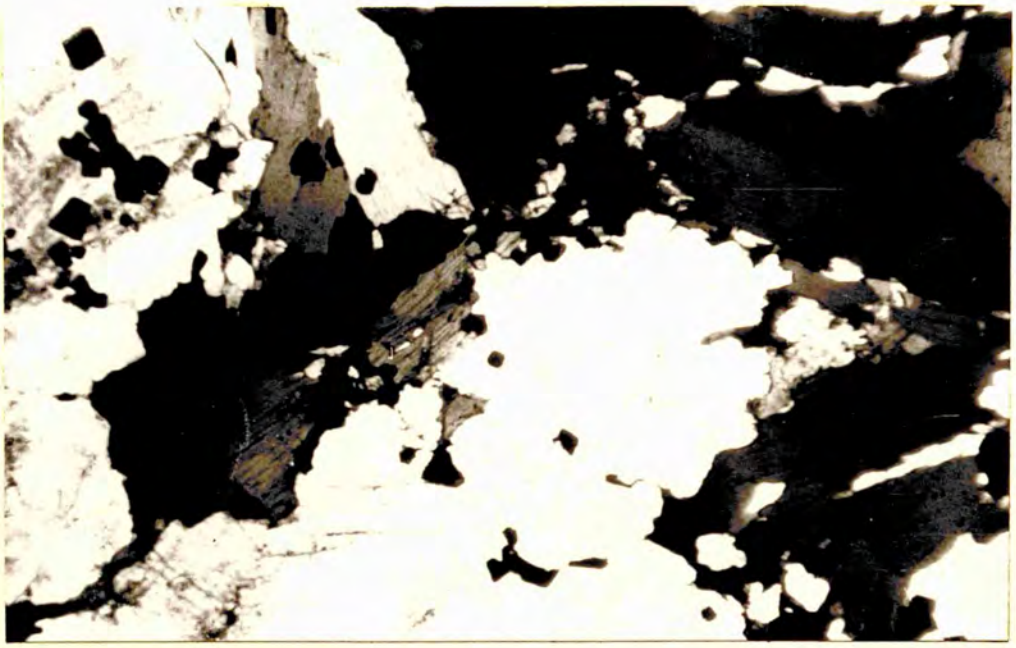


Plate 158. Spinel crystals (dark grains). Hornfels xenolith in pyroxene-mica-diorite. P.P.L. X 25.

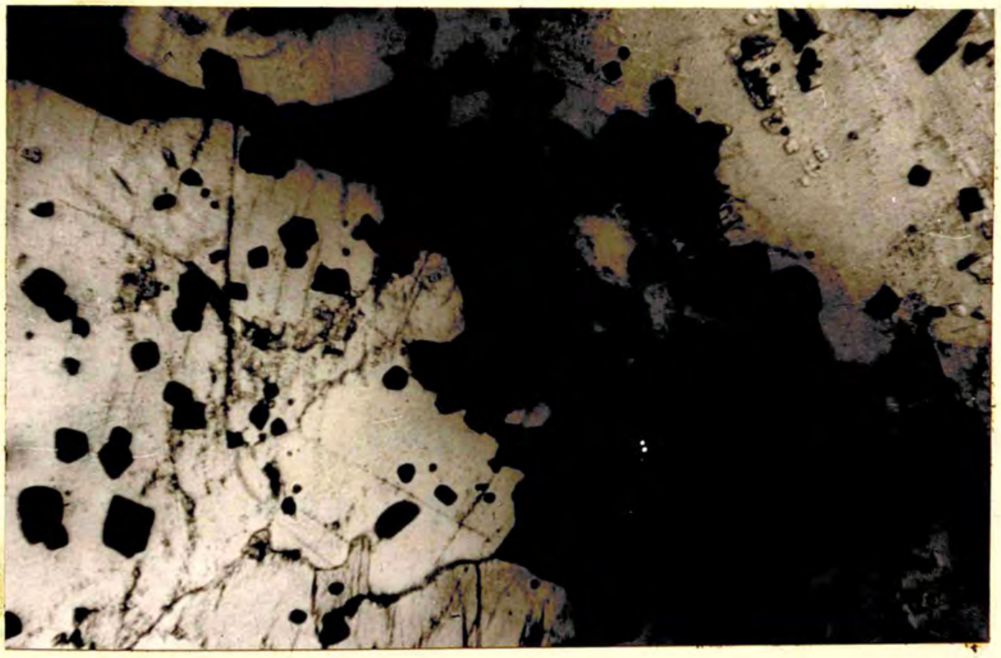


Plate 159. Disseminated spinel grains. Hornfels xenolith in pyroxene-mica-diorite. P.P.L. X 85.

of igneous or metasedimentary origin, although occasionally relict banding shows their metasedimentary origin. They have a fine-grained granoblastic hornfels texture, generally consisting of hypersthene, ore, and labradorite which sometimes has clear, broad twin lamellae. Ore commonly forms laths in a decussate pattern, and in some cases diffuse biotite laths are seen to be altering to hypersthene. Occasionally there are a few large antiperthitic plagioclase grains.

At the contacts hypersthenes are often large and poikiloblastic (Plate 160), and approach those of the neighbouring diorite in size. At these contacts, hypersthenes in the diorite are anhedral and somewhat rounded and there is a considerable amount of iron staining. The diorite-hypersthenes are strongly pleochroic from pinkish-buff to pale green, possibly suggesting contamination by metasedimentary material. It is possible that these hornfelses are being made over to diorite at their margins, and further discussion on this follows in the next section.

One of the basic hornfels xenoliths can be seen in the field to be a large block of Husfjord metagabbro. This is now a pyroxene hornfels with a fine-grained granular texture, consisting primarily of diopside, hypersthene, ore, and plagioclase. Some of the biotites are diffuse whereas others are fresh, and the latter are probably a later regional metamorphic effect.

Symplectites and Coronas

Around some of the metasedimentary basic xenoliths in the pyroxene-mica-diorites, particularly those which are clearly becoming assimilated,

Plate 160. Poikiloblastic hypersthene. Hornfels xenolith in pyroxene-mica-diorite. X-Pl. X 85.

interesting textures develop in the diorite, involving the formation of synantetic minerals. These are secondary minerals produced along the boundaries between two minerals which are reacting with each other (and in general with free migrating ions). The term was introduced by Sederholm (1916), and they may be products either of later metamorphism or of autometamorphism in which they formed at a late magmatic stage. Synantetic minerals may take the form of coronas, sometimes with two or more shells if the reaction takes place in more than one step (Shand 1945), or as complex intergrowths between the synantetic minerals formed. Sederholm termed these intergrowths symplectites when they were vermicular in form, and dactylites when they were finger-like or rod-like in form.

In the rocks under present discussion, symplectites, dactylites, and coronas form, and these are now described in detail and their significance discussed.

The symplectites and dactylites are formed between hypersthene and potash feldspar, indicating that at the time of formation of the synantetic minerals the hypersthene and potash feldspar were unstable together. Where hypersthene makes contact with potash feldspar it becomes highly frittered and breaks up into rounded or elongate grains (Plate 161). The fact that this only occurs against potash feldspar is shown by hypersthenes which make contact with both potash feldspar and plagioclase. In these cases the frittering only develops against the potash feldspar (Plate 162). An extreme example of this selection is given by hypersthenes which have been enclosed within antiperthitic plagioclase phenocrysts. These are quite stable where they make contact with the plagioclase

Plate 161. Pyroxene-plagioclase symplectite at hypersthene/potash feldspar contact. Pyroxene-mica-diorite. P.P.L. X 85.

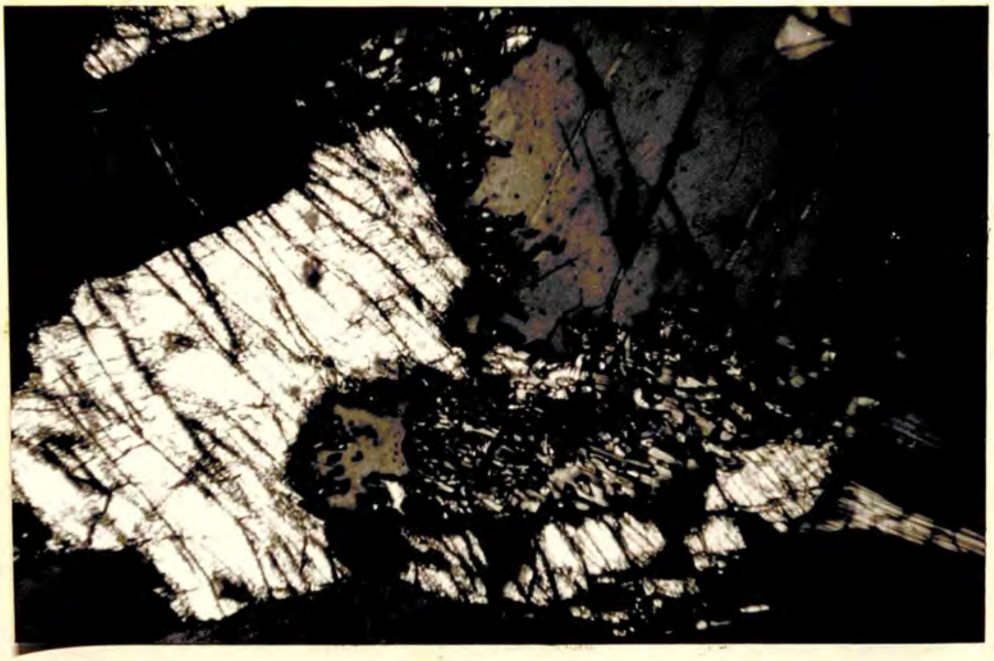


Plate 162. Pyroxene-plagioclase symplectite forming at hypersthene/potash feldspar contact but not at hypersthene/plagioclase contact. Pyroxene-mica-diorite. X-Pl. X 85.

host, but where they are enclosed within potash feldspar exsolution patches they have degenerated to rounded or vermicular grains (Plate 163).

Various stages in the development of the vermiculites can be traced. In early stages, dactylitic hypersthene grains are still joined into the parent crystal and remain in optical continuity with it. In the next step, these dactylitic grains become detached from their parent but are still in optical continuity with it. Meanwhile the parent crystal decreases in size as it provides grains to the vermiculites. The formation of these textures is a prelude to further chemical reactions between hypersthene and potash feldspar. As a result of these reactions the pyroxene becomes pale green, non-pleochroic, and more highly birefringent. It appears to be diopsidic, and now all the grains do not remain in optical continuity with one another. The potash feldspar alters to plagioclase of oligoclase composition, and this encloses the diopsidic grains. Rarely, stages in the evolution of these symplectites can be observed in a single example. In these, the pyroxene dactylites proximal to the parent grain are hypersthene and are in optical continuity with it, whereas distal grains are vermicular diopsides (Plate 164). In the most advanced stages, the parent crystal has been completely replaced by diopside-plagioclase symplectites (Plate 165).

The problem with these symplectites is to tell whether they are late magmatic effects or whether they are of metamorphic origin. It must be remembered that they only occur in the neighbourhood of xenoliths, and that away from these the hypersthenes are quite stable with potash

Plate 163. Vermicular pyroxene in potash feldspar exsolution patch in antiperthite. Pyroxene-mica-diorite. X-Pl. X 170.

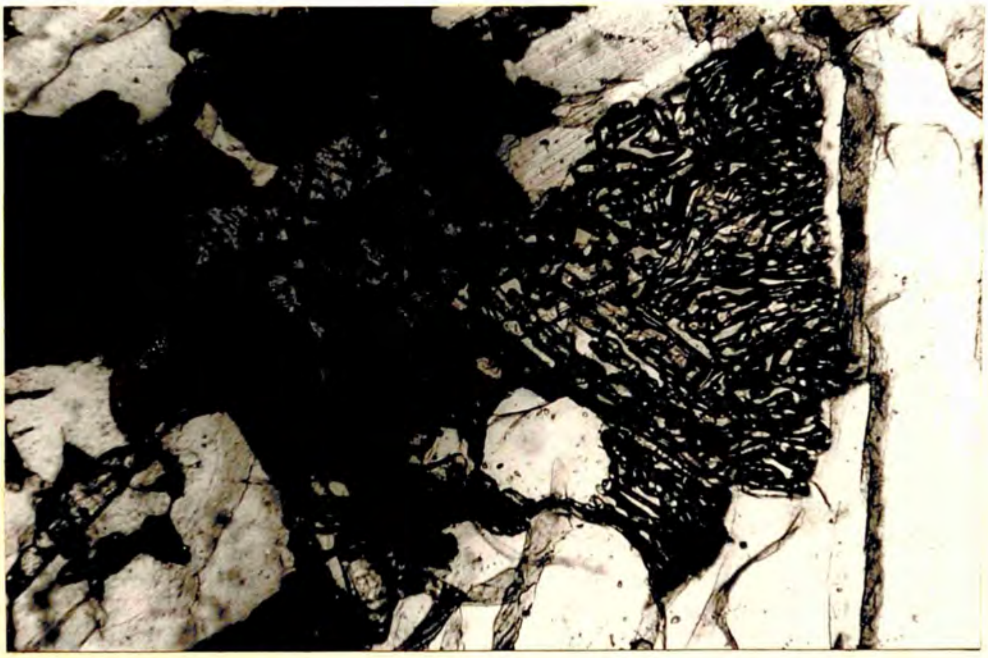


Plate 164. Development of symplectite from hypersthene.
Pyroxene-mica-diorite. P.P.L. X 95.

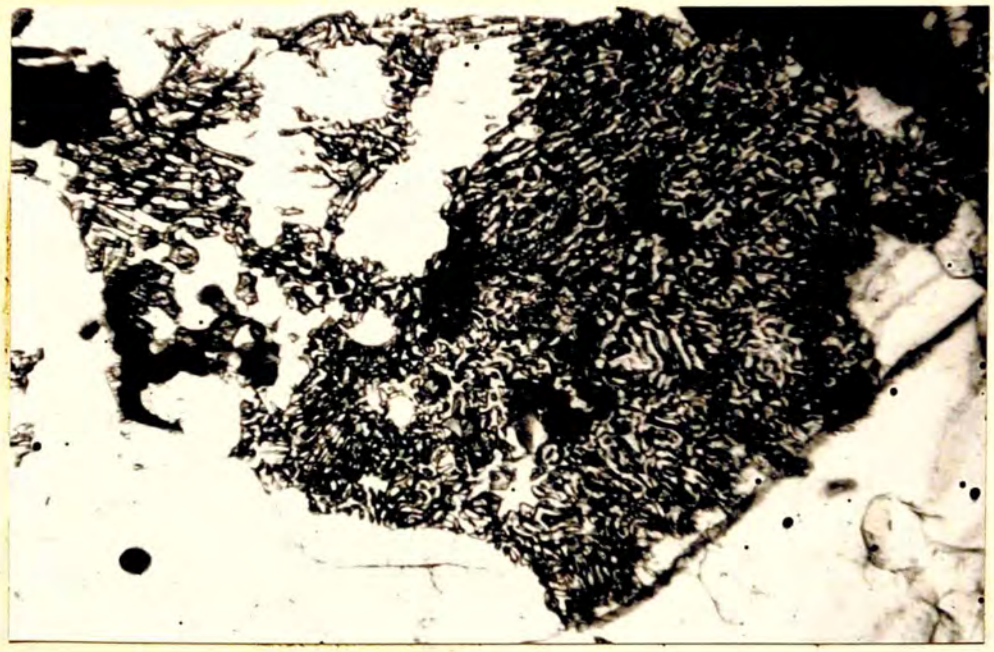


Plate 165. Large pyroxene-plagioclase symplectite.
Pyroxene-mica-diorite. P.P.L. X 85.

feldspar. Thus it is clear that the xenoliths play a significant part in the formation of these textures, presumably due to contamination effects upon the surrounding dioritic material. This contamination could cause the reactions to take place at the time of emplacement of the diorite, or alternatively it could cause a latent chemical instability in the diorite around the xenoliths. In the latter case the instability could become manifest in reactions producing symplectites if the rock underwent a later metamorphism during which metastable relationships were stabilized. Thus the fact that the formation of the symplectites is associated with contamination by the xenoliths does not unequivocably demonstrate at what stage in the diorite's history they were formed.

Considerations of the chemistry of the reactions may give a clue to the likely time of formation. The reactants are hypersthene and perthitic potash feldspar, and the reactions tend to produce plagioclase and a diopsidic pyroxene. An extremely simplified equation of the probable reaction may be written:

$$\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2} + \underset{\text{perthitic potash feldspar}}{(K.Na)_2O.Al_2O_3.6SiO_2} + 2Ca^{2+} \rightarrow$$

$$\underset{\text{diopside}}{(Mg.Ca)O.SiO_2} + \underset{\text{plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + SiO_2 + Fe^{2+} + 2K^{+} + 2O^{2-}$$

The reaction requires the addition of Ca^{2+} since both diopside and plagioclase are formed. During regional metamorphism it is more usual for alkalis to be migrating as free ions than it is for lime, and it seems more probable that the calcium ions were derived from basic material contaminating the diorite. For this reason it is considered that the symplectites were probably formed at the time of emplacement of the diorite rather than during a later metamorphism.

Textural evidence supports this view. The diorite contains late biotites which were probably formed during regional metamorphism. Occasionally these biotites overgrow the symplectites (Plate 166). This evidence does not prove conclusively that the symplectites could not have formed during the metamorphism, but it does show that their formation was followed by the growth of metamorphic minerals.

Another feature common to these contaminated areas of the diorite is the development of vermicular inclusions of ore in the hypersthenes. These textures are not strictly symplectites since no chemical reactions appear to have been involved in their formation, but rather they are exsolution phenomena. However, they are discussed here as their formation is dependant upon the presence of the xenoliths. The ores are ilmenite and magnetite, and these sometimes form delicate vermiculites in the central parts of the hypersthenes (Plate 167). More commonly the vermiculites are rather coarser and ragged, and in many cases irregular inclusions of solid ore are formed (Plate 168). It is to be noted that the ores here include magnetite, which does not normally occur in the coarse diorites, and these must have been derived from some source other than the dioritic magma. It is also significant that the hypersthenes containing this ore are strongly pleochroic from deep pinkish-buff to light green, suggesting contamination by metasedimentary material.

In the basic hornfelses constituting the xenolithic inclusions hypersthene grains also contain ore inclusions, often vermicular

Plate 166. Biotite overgrowing pyroxene symplectite. Pyroxene-mica-diorite. P.P.L. X 85.

Plate 167. Vermicular ore in diorite-hypersthene.
Pyroxene-mica-diorite. P.P.L. X 95.

Plate 168. Ore inclusions in diorite-hypersthene.
Pyroxene-mica-diorite. P.P.L. X 95.

(Plate 169). The textures closely resemble those in many of the diorite-hypersthenes. In the diorite next to the contact, large markedly pleochroic hypersthenes contain a number of separate ore inclusions, each one resembling those in the centres of the small hornfels-hypersthenes (Plate 170). It appears that the large strongly pleochroic diorite-hypersthenes have formed by the amalgamation of a number of hornfels-hypersthenes, each one retaining its central ore inclusions. Although the large hypersthene seen in Plate 170 appears to be a single uniform crystal under plane-polarized light, under crossed polars it is seen that it is not all exactly in optical unity. Areas between the ores have very slightly different optical orientations, perhaps emphasizing that the crystal has been formed by the amalgamation of separate grains and has not quite achieved optical continuity.

This theory for the formation of these large hypersthenes is borne out by the observation of a large diorite-hypersthene actually joined to a small hornfels-hypersthene, with ore occurring in the hypersthene which makes the join (Plate 171). It appears from this that the hornfelses are being incorporated into the diorite at their margins. If the process was continued further, all trace of the hornfelses would probably disappear, and they would be completely replaced by diorite. Thus it seems to be possible for a dioritic rock to form from the assimilation of basic metasedimentary material.

In the neighbourhood of some pelitic xenoliths, synantetic minerals form as coronas around several different minerals.

A corona of dark green hornblende forms around ore grains where

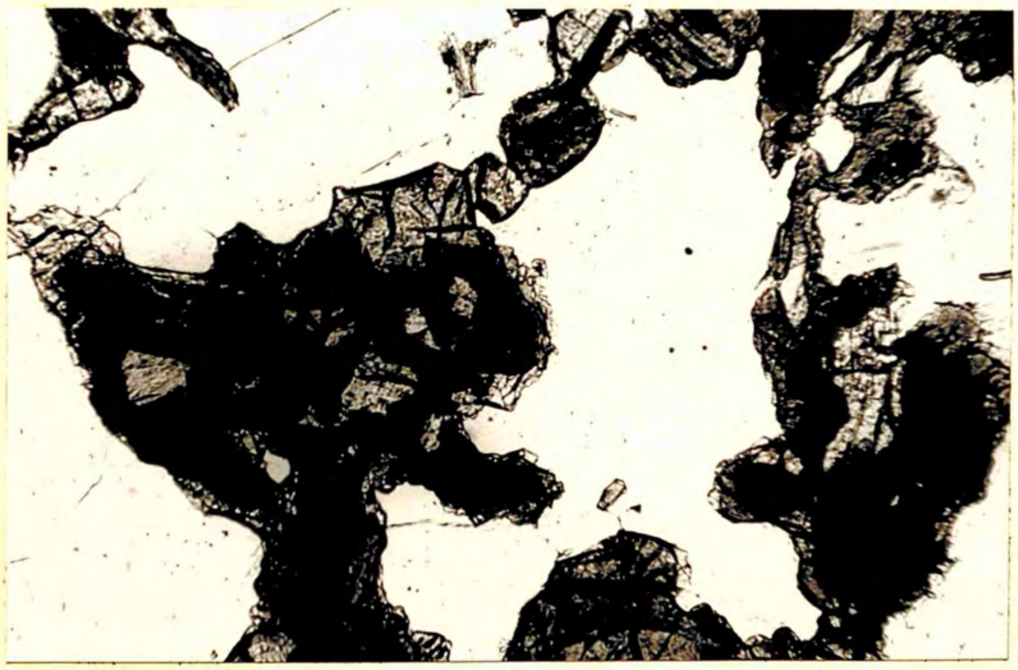


Plate 169. Vermicular ore in hornfels-hypersthene. Pyroxene-mica-diorite. P.P.L. X 170.

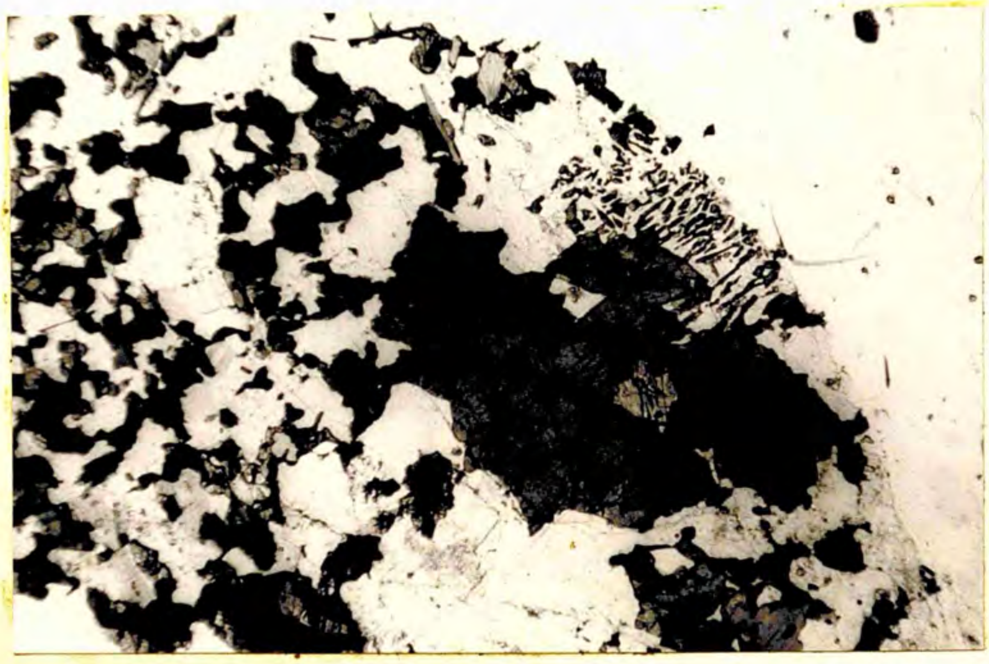


Plate 170. Ore inclusions in diorite-hypersthene (coarse-grained) resembling those in hornfels-hypersthene (fine-grained). Pyroxene-mica-diorite. P.P.L. X 25.

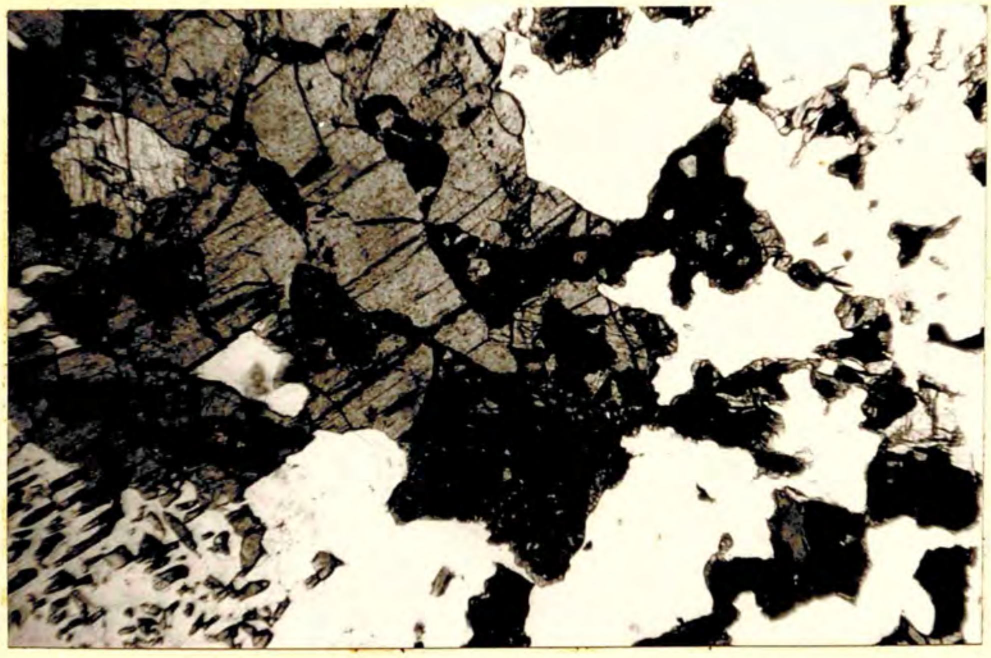


Plate 171. Join between diorite-hypersthene and hornfels-hypersthene; note ore inclusions in diorite-hypersthene. Pyroxene-mica-diorite. P.P.L. X 85.

contact with plagioclase is made. Occasionally this amphibole has a turquoise tinge, indicating a certain sodic content, presumably derived from the plagioclase. In the reaction, iron from the ore has been added to the calcium, aluminium, and silicon of the plagioclase under hydrous conditions to form hornblende; in some cases the excess soda has been accommodated in the hornblende structure. The approximate equation may be represented symbolically as follows:

$$\underset{\text{ore}}{5FeO} + 2\underbrace{[(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2]}_{\text{andesine plagioclase}} + H_2O + Al^3 + 5Mg^{2+} \rightarrow$$

$$\underbrace{2CaO.5(Mg.Fe.Al)O.H_2O.8SiO_2}_{\text{hornblende}} + 4Na^+ + 6O^{2-}$$

Ore is sometimes rimmed instead by orthopyroxene which is faintly pleochroic from buff to pale green, and is presumably hypersthene. This may have the reaction:

$$\underset{\text{ore}}{FeO} + \underbrace{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2}_{\text{andesine plagioclase}} + Mg^{2+} \rightarrow$$

$$\underbrace{(Mg.Fe)O.SiO_2}_{\text{hypersthene}} + Al_2O_3 + 3SiO_2 + Ca^{2+} + 2Na^+ + O^{2-}$$

Double coronas form around some ores in which hypersthene forms the inner shell and hornblende the outer (Plate 172), indicating that the reactions producing the synantetic minerals took place in two stages. If the hornblende formed first by the reaction given above, then the second reaction between this and the ore must have been of the following nature:

$$\underset{\text{Ore}}{FeO} + \underbrace{2CaO.5(Mg.Fe.Al)O.H_2O.8SiO_2}_{\text{hornblende}} + 4O^{2-} \rightarrow$$

$$5[\underbrace{(Mg.Fe)O.SiO_3}_{\text{hypersthene}}] + 2CaO + 3SiO_2 + H_2O + 5Al^{3+} + Fe^{2+}$$

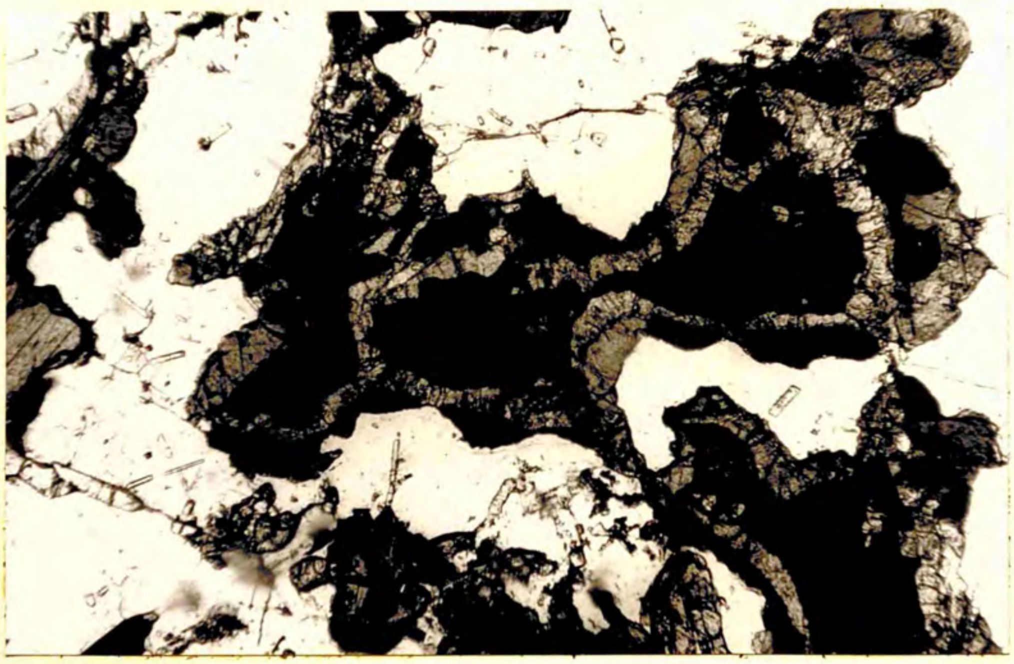


Plate 172. Double coronas around ore; inner shell is of hypersthene and outer is of hornblende. Pyroxene-mica-diorite. P.P.L. X 95.

On the other hand, if the hypersthene formed first, then the second reaction must have taken place between this and the plagioclase as follows:

$$\underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + 5\big[\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2}\big] + H_2O + Al^{3+} + 6O^{2-} \rightarrow$$

$$\underset{\text{hornblende}}{2CaO.5(Mg.Fe.Al)O.H_2O.8SiO_2} + 5SiO_2 + 4Na^{+}$$

The latter reaction is probably the more likely since in the former, there appears to be free iron on both sides of the equation.

In other examples, many ores are rimmed by thin coronas of granular, colourless garnet. Sometimes the garnet is in direct contact with both the ore and the plagioclase, and sometimes there is a shell of hypersthene between the garnet and the ore (Plate 173). In the former case, the reaction is of the following nature:

$$\underset{\text{ore}}{Fe_2O_3} + \underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + 2Ca^{2+} \rightarrow$$

$$\underset{\text{garnet}}{3CaO.(Fe.Al)_2O_3.3SiO_2} + SiO_2 + 2Na^{+} + O^{2-}$$

In the latter case, the reactions may be:

(i) $$\underset{\text{ore}}{FeO} + \underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + Mg^{2+} \rightarrow$$

$$\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2} + Al_2O_3 + 3SiO_2 + Ca^{2+} + 2Na^{+} + O^{2-}$$

(ii) $$\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2} + \underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + 2Ca^{2+} \rightarrow$$

$$\underset{\text{garnet}}{3(Ca.Mg.Fe).Al_2O_3.3SiO_2} + 4SiO_2 + 2Na^{+} + O^{2-}$$

Shand (1945) has recorded that when garnet occurs in complex coronas it occurs on the outside, against the plagioclase.

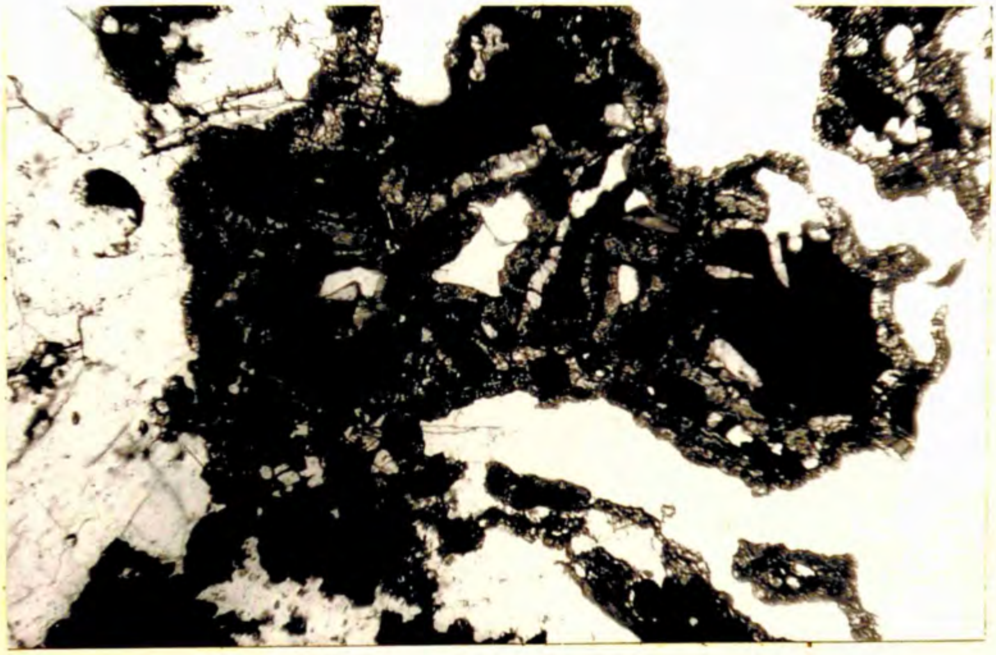


Plate 173. Garnet rimming hypersthene which in turn rims ore. Pyroxene-mica-diorite. P.P.L. X 85.

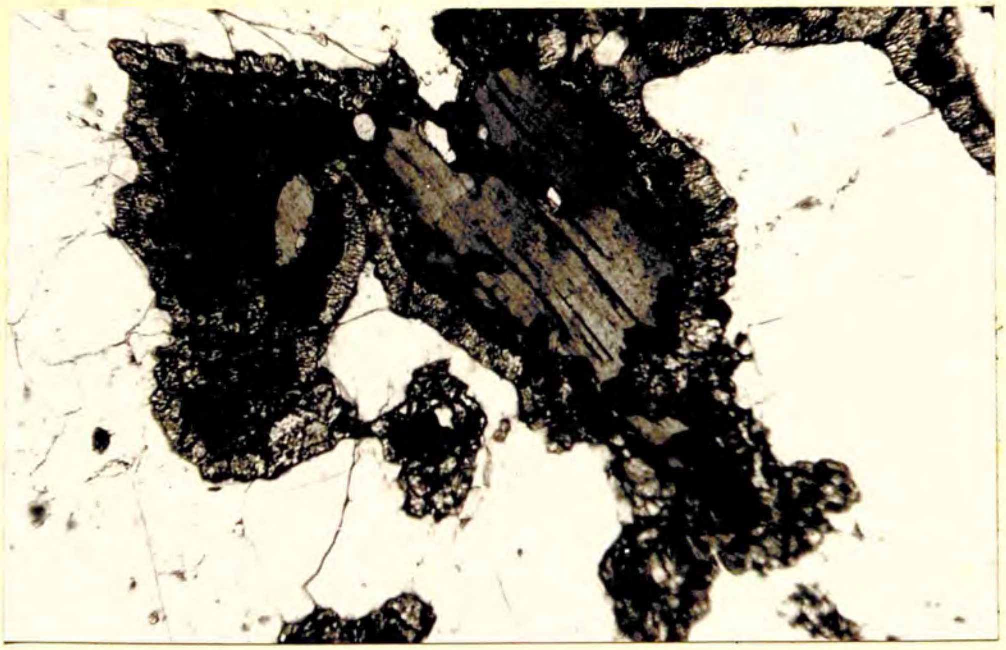


Plate 174. Garnet corona around biotite. Pyroxene-mica-diorite. P.P.L. X 95.

Garnet also forms thin coronas around some of the biotites (Plate 174). The biotite is obviously altering to garnet, and the garnet contains minute birefringent patches where there are relics of biotite, especially along the contact zone. The formation of the garnet was probably in the following way:

$$\underset{\text{biotite}}{K_2O.6(Mg.Fe)O.Al_2O_3.2H_2O.6SiO_2} + \underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} \rightarrow$$

$$2\left[\underset{\text{garnet}}{3(Mg.Fe)O.Al_2O_3.3SiO_2}\right] + 4SiO_2 + 2H_2O + Ca^{2+} + 2Na^{+} + 2K^{+} + 2O^{2-}$$

Some of the biotites have coronas of dark green hornblende to which it is clearly altering. The contacts between them are diffuse and the reaction must be of the following nature:

$$\underset{\text{biotite}}{K_2O.6(Mg.Fe)O.Al_2O_3.2H_2O.6SiO_2} + \underset{\text{andesine plagioclase}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + Ca^{2+} + Al^{3+}$$

$$\rightarrow \underset{\text{hornblende}}{2CaO.5(Mg.Fe.Al)O.H_2O.8SiO_2} + 2SiO_2 + H_2O + Mg^{2+} + Fe^{2+} + 2Na^{+}$$

$$+ 2K^{+} + 12O^{2-}$$

The development of these complex coronas is not ubiquitous through the rock, even in the neighbourhood of xenoliths. However, they occur in irregular patches which are generally somewhat finer-grained than the rest of the diorite and which may represent former fine-grained xenolithic inclusions in the diorite.

Again, the clue to these textures lies in the hornfels inclusions. These pelitic hornfelses sometimes have textures very similar to those just described but on a smaller scale (Plate 175, cf. Plates 173 and 174). In these, biotites and hypersthenes are rimmed by thin coronas of garnet, and in addition there is an abundance of deep green spinel. These spinels, together with ore grains, are rimmed by hypersthene or

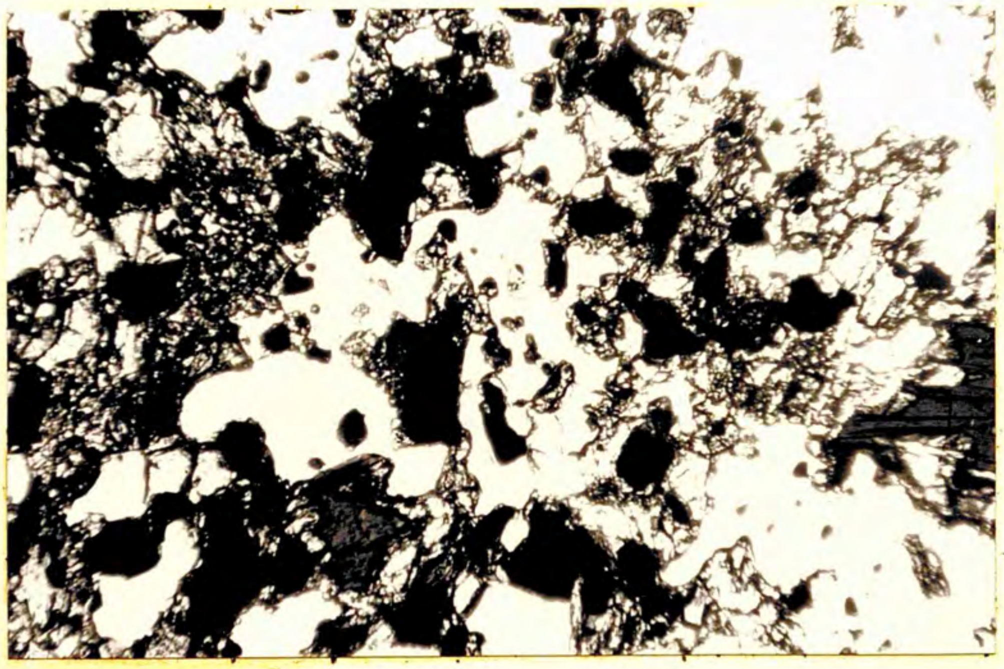


Plate 175. Garnet rimming biotite, hypersthene, ore and spinel. Pyroxene-mica-diorite. P.P.L. X 95.

garnet, or by both with garnet on the outside. The formation of hypersthene or garnet from spinel must be as follows:

$$\text{(i)}\quad \underset{\text{spinel}}{(Mg.Fe)O.Al_2O_3} + SiO_2 \rightarrow \underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2} + Al_2O_3$$

$$\text{(ii)}\quad 3\big[\underset{\text{spinel}}{(Mg.Fe)O.Al_2O_3}\big] + 3SiO_2 \rightarrow \underset{\text{garnet}}{3(Mg.Fe)O.Al_2O_3.3SiO_2} + 2Al_2O_3$$

The formation of these textures in the pelites is probably associated with the hornfelsing by the diorite. The finer-grained patches in which coronas develop in the diorite appear to represent relics of assimilated pelitic hornfelses in which complex coronas had developed before they were made over into diorite.

It can be seen from the study of the symplectites and coronas in the coarse pyroxene-mica-diorites that these diorites are highly contaminated rocks. Evidence provided by the various textures leads towards the conclusion that much of the material constituting the diorite, at least in the neighbourhood of the metasediment xenoliths, is of metasedimentary origin, having been derived from the xenoliths. For further discussion, see p.268 .

Garnet-rich Quartz-diorites

These coarse-grained quartz-diorites are sometimes monzonitic (Table 5), and have a porphyritic xenomorphic texture (Plate 176).

Phenocrysts are nearly always of perthitic potash feldspar and are large and anhedral. Inclusions of irregularly-shaped quartz grains and of small biotites occur in these megacrysts. At their margins there is commonly a development of myrmekite lobing into the potash feldspars.

TABLE 5

MODAL ANALYSES OF GARNET-RICH QUARTZ-DIORITES

Mineral	38/13A	38/56C	46/4A	46/17A[1]
Plagioclase	15.80	12.75	12.95	33.14
K-feldspar	66.47	62.47	60.69	40.35
Quartz	6.12	19.63	11.53	2.97
Myrmekite	0.76	-	0.44	-
Biotite	-	3.59	11.63	17.97
Muscovite	0.36	1.48	0.23	-
Garnet	0.22	-	1.33	4.80
Ore	0.27	-	0.27	0.24
Apatite	-	-	0.05	0.06
Zircon	0.20	0.08	0.96	0.21
	99.96	100.00	100.08	99.74

1: Modal analysis carried out by macro point counting techniques - see Appendix.

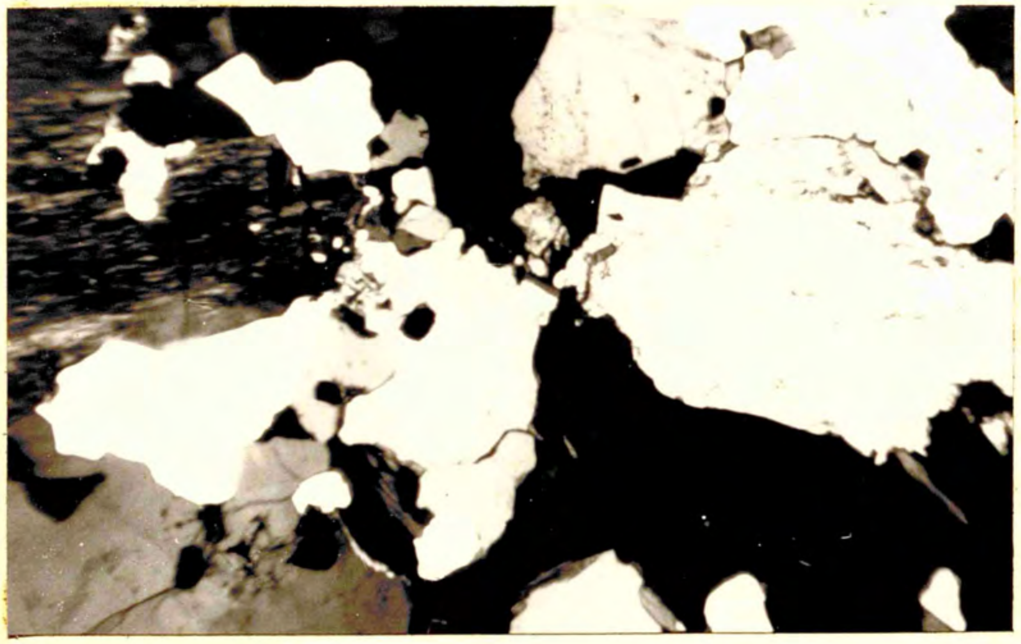


Plate 176. Texture of coarse garnet-rich quartz-diorite. X-Pl. X 25.

In places these potash feldspars exhibit microcline twinning, indicating that they have a certain degree of triclinicity. Potash feldspar also occurs in the matrix.

Plagioclase crystals occur mainly in the matrix, but also occasionally as phenocrysts. They are oligoclase-andesine in composition, being around An_{30}. Some are slightly deformed and cracked, and most of them are cloudy and slightly sericitized. Quartz occurs both as blebs included in some of the plagioclase grains and as large discrete interstitial grains of irregular shape.

Biotite laths are anhedral to subhedral, and are sometimes quite broad compared with their length. They often have a ragged appearance, and some are poikilitic, occasionally containing vermicular inclusions of quartz, and they occur in aggregates in which individual laths are randomly orientated. These aggregates are positioned along the boundaries of the large felsic crystals. Muscovite also occasionally occurs, usually in association with the biotite.

The garnets are colourless, and are irregularly disseminated throughout the rock. They are anhedral and enclose grains of quartz, plagioclase, and biotite, although some garnets are enclosed within large biotite and others appear to be intergrown with biotite. In a few cases, a clear zone of an indeterminate mineral, possibly quartz, occurs between the two minerals, which appear to be unstable together. It appears that the garnets are altering to some of the large biotites, which are clearly late.

These diorites do not contain small xenoliths, but in the field are generally closely associated with metasedimentary rafts. They only occur in or near the area where rafts are seen at the surface, and could well be intimately associated with them beneath the present erosion surface. It is suggested that the formation of sporadic garnets in these diorites is a result of slight contamination of the quartz-diorite by metasedimentary material as the magma rose to its present position (see p. 269).

Accessory minerals in these diorites include apatite, occasional irregular sphenes, frequent zircons which commonly cause the formation of pleochroic haloes in the biotites, and ore. This latter includes ilmenite, and pyrrhotite which is highly altered to goethite.

Garnet-poor Quartz-diorite

General

This coarse-grained quartz-diorite also tends to be monzonitic (Table 6), and has a porphyritic xenomorphic texture (Plate 177). Grain boundaries are sometimes sutured, a phenomenon that suggests that slight reheating may have taken place. Occasionally there is a tendency for the phenocrysts to have a preferred orientation.

Potash feldspar and plagioclase (An_{30-35}) both occur as phenocrysts. The plagioclase phenocrysts are generally free from inclusions, but the potash feldspar is perthitic, and contains quartz inclusions.

Quartz occurs interstitially, but also as large irregular grains which do not usually show any signs of strain.

Micas, principally biotite, generally lie along grain boundaries,

TABLE 6

MODAL ANALYSES OF GARNET-POOR QUARTZ-DIORITES

Mineral	38/52A	38/56C	48/8B	48/29B[1]	03/15C[1]
Plagioclase	20.55	20.34	7.72	32.79	39.10
K-feldspar	45.32	37.64	72.46	46.79	33.10
Quartz	22.79	28.05	6.97	2.13	9.36
Myrmekite	1.14	-	0.69	-	-
Biotite	9.29	7.35	10.29	17.69	16.05
Muscovite	0.15	1.40	0.46	-	2.20
Garnet	-	1.21	0.44	-	-
Amphibole	0.15	-	-	-	-
Ore	0.20	0.79	0.11	1.02	0.08
Apatite	0.30	-	-	-	-
Zircon	0.10	0.24	0.20	0.17	0.32
Sphene	-	2.97	-	-	-
Rutite	-	-	0.33	-	-
	99.99	99.99	99.67	100.08	100.21

1: Modal analysis carried out by macro point counting techniques - see Appendix.

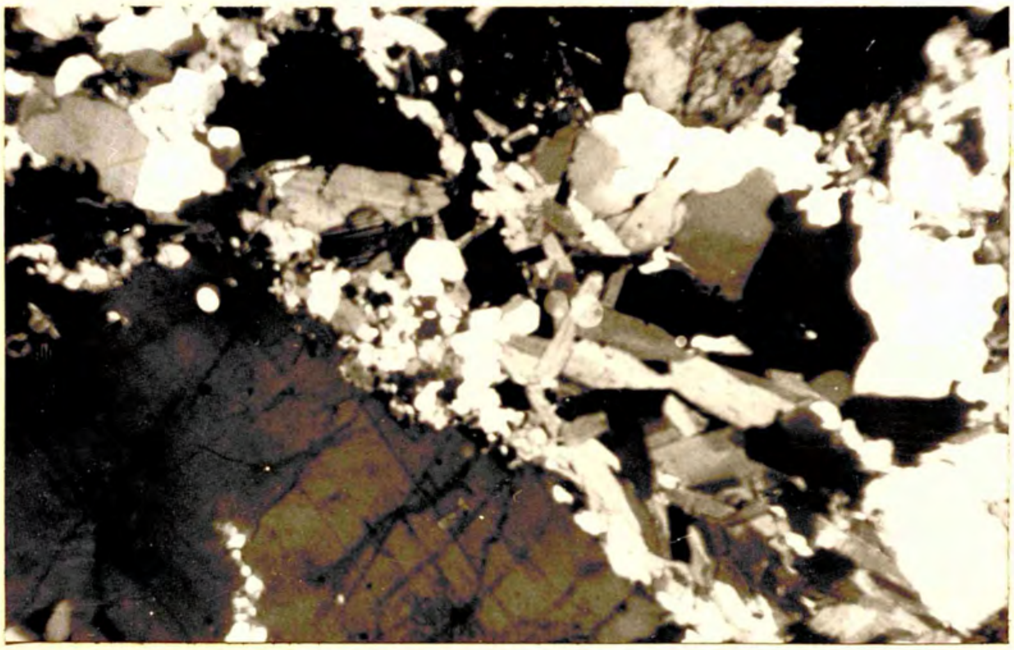


Plate 177. Texture of coarse garnet-poor quartz-diorite. X-Pl. X 25.

or form aggregates of randomly-orientated laths between the large felsic grains. Quartz occurs interstitially between the laths. Rutile and zircon are fairly common accessories in these biotite aggregates.

Garnet rarely occurs, but when it does, it is colourless and forms small irregular grains. Some are altering to lath-like needles or fibrous sheafs of muscovite, which, in turn, are altering to biotite.

There is only a little ore, and this is principally goethite, which forms interstitial grains. Remnants of pyrrhotite usually occur within the goethite, and the latter is clearly an alteration product of the former. A little ilmenite also sometimes occurs.

Xenoliths

Some of these diorite sheets contain xenolithic blocks of metasedimentary material which are usually fairly angular and are not being markedly assimilated by the diorite.

Semi-pelitic Xenoliths

In these, a relict schistosity is preserved by bands rich in biotite which still tend to retain their preferred orientation. The biotite laths are, however, partly reorientated into a decussate pattern with grains of quartz and andesine (An_{35}) to form a granular hornfelsic texture.

Biotites are generally diffuse, and are overprinted by small kyanite prisms, which have not broken down during thermal metamorphism. Some of the kyanites are twinned, with (100) as the composition plane.

Small colourless allotrioblastic garnets, which sometimes occur in aggregates, overgrow both the biotites and the kyanites, and contain numerous inclusions of both these and quartz.

These semi-pelites very occasionally contain grains of a turquoise schorlite variety of tourmaline.

Calc-silicate Hornfels Xenoliths

These have a fine-grained granular hornfelsic texture, consisting mainly of randomly-orientated laths of tremolite-actinolite. This amphibole has the following pleochroic scheme:

X pale buff

Y pale green

Z pale green

It is poikiloblastic and contains inclusions of quartz, and in some bands rounded diopside grains.

Small diffuse biotite laths are in the process of altering to tremolite-actinolite.

Feldspathization

As described in the field account, these quartz-diorites have a coarse feldspathization associated with them, and in places a gradation between this feldspathization and the diorites is apparent. The feldspathization affects both Husfjord metagabbro and metasediments.

Metagabbro

The Husfjord metagabbro which has been feldspathized contains allotrioblastic porphyroblasts of perthitic potash feldspar, some of which enclose quartz grains. They are rimmed by myrmekite having frond-like quartz vermicules, or by fine-grained mosaics of plagioclase and quartz. The perthite porphyroblasts appear to have been slightly re-crystallized at their margins, indicating a slight reheating. The feldspar is often cloudy and has been sericitized.

Plagioclase sometimes forms fairly large crystals, but it has generally recrystallized to a polygonal mosaic having a composition of about An_{35}.

The mafic minerals of the metagabbro now consist of green actinolite and biotite, the latter sometimes enclosing the former. These mafic minerals occur in long aggregates between the porphyroblasts. The biotites form fresh-looking laths which are randomly orientated within the aggregates, and which are sometimes large and poikiloblastic. Interstitial quartz and feldspar occur between the biotite laths, and some of the biotites are altered to fibrous chlorite along cleavage planes. An expansion in volume seems to take place during this change since the biotite has been buckled around the growing chlorite.

Pelitic Rafts in Metagabbro

Feldspathized pelitic rafts also contain allotrioblastic potash feldspar porphyroblasts, which enclose quartz crystals. These megacrysts often have a little myrmekite along their margins, and they have been slightly sericitized. Plagioclase crystals, however, have been more extensively altered to sericite and muscovite, and are very cloudy.

Quartz is common, either interstitially or as large irregular grains, and they do not exhibit any undulose extinction.

Micas, principally biotite, lie along grain boundaries and form aggregates between the megacrysts. There are occasional fragmentary garnets which are cracked and beginning to alter to an indeterminate birefringent mineral, possibly mica, along cracks.

Sphene is fairly common in these rocks. In one case, some very large sphenes have altered to a brown opaque mineral which is creamy-white under reflected light, and is most probably leucoxene. Sometimes the sphenes have been completely pseudomorphed, their characteristic lozenge shapes being retained, but generally, a few relicts of sphene remain (Plate 178). The sphenes evidently were twinned, for partings are reflected across a median twin composition plane giving a herringbone pattern (Plate 179). This alteration is possibly associated with the feldspathization.

Plate 178. Sphene pseudomorphed by leucoxene. Garnet-poor quartz-diorite. P.P.L. X 85.

Plate 179. Twinned sphene being pseudomorphed by leucoxene. Garnet-poor quartz-diorite. P.P.L. X 170.

KOMAGFJORD DIORITE

General

This diorite has a medium-grained, non-porphyritic, xenomorphic texture, and contains both orthopyroxene and clinopyroxene (Table 7). The orthopyroxene is hypersthene, which sometimes contains a little vermicular ore, and the clinopyroxene is diopsidic augite which is beginning to alter to green hornblende, mainly at the margins. There are also some discrete hornblende crystals.

The only feldspar is plagioclase, which is andesine ranging up to An_{40}. Biotite forms laths which have a rather ragged appearance, and are often diffuse.

Ores consist of ilmenite, magnetite, and pyrrhotite. The ilmenite and magnetite both form rounded grains, and also irregular branching grains which penetrate slightly into cracks in the gangue minerals. Ilmenite also forms aggregates of small blebs, and magnetite occurs as fine lamellae in pyroxenes. Pyrrhotite is not common, and is altered at the margins to goethite.

Xenoliths

Small elongate xenoliths are a characteristic feature of this diorite, and these are both dark-weathering and light-weathering. The former include amphibolites and basic pyroxene hornfelses, and the latter consist of more felsic pyroxene hornfelses.

TABLE 7

MODAL ANALYSES OF KOMAGFJORD DIORITE

Mineral	11/47 F
Plagioclase	77.43
Hypersthene	4.43
Clinopyroxene	5.43
Biotite	4.60
Ore	7.00
Hornblende	0.80
Apatite	0.43
	100.12

Amphibolite Xenoliths

These are porphyritic, but have a very fine-grained matrix in which there are relics of a subophitic texture. They appear to represent fragments of dyke material. The amphibole is a greenish-brown hornblende with the following pleochroism:

X greenish-yellow

Y deep greenish-brown

Z deep greenish-brown

The plagioclase phenocrysts have now recrystallized, and some are zoned with less calcic margins. Small inclusions of amphibole in the phenocrysts indicate that the latter must have recrystallized at a relatively late stage.

Basic Pyroxene Hornfels Xenoliths

These consist of plagioclase diopside, hypersthene, and ore, and these minerals form a fine-grained granoblastic hornfelsic texture. A few ragged biotite laths appear to be later than the hornfelsing.

Felsic Pyroxene Hornfels Xenoliths

These have a slightly higher feldspar content than the basic pyroxene hornfelses, and both plagioclase and K-feldspar are present.

Hypersthene is the predominant pyroxene, although a few diopside grains do occur. The pyroxenes and ore form a granoblastic texture with the feldspars.

Minor Intrusions

Besides the perthosites, which will be discussed separately, the xenolithic Komagfjord diorite has been intruded by a number of small minor intrusions. These include grey-coloured basic diorites, green porphyritic amphibolites, and a hypersthene-bearing amphibolite.

Basic Diorite

This is fairly fine-grained, and sometimes shows a slight fluxion structure. A good relict subophitic texture is generally preserved in which plagioclase is in the basic andesine range An_{45-50}.

The predominant pyroxene is clinopyroxene, but some grains of hypersthene also occur. The clinopyroxene is colourless augite which is being altered to green hornblende.

The ores in this rock include pyrrhotite, chalcopyrite, and ilmenite. Pyrrhotite and chalcopyrite are closely associated with one another, and the textures suggest that this association is probably an exsolution phenomenon. The pyrrhotite is altering to goethite, an alteration which does not affect the chalcopyrite. Goethite commonly penetrates the basal cleavages of the platy pyrrhotite, but cannot penetrate the more massive chalcopyrite in this way. Ilmenite is clearly late as it forms irregular grains which enclose the chalcopyrite-pyrrhotite-goethite complexes.

Porphyritic Amphibolite

This green-coloured amphibolite has a relict subophitic texture in which plagioclase has recrystallized to basic andesine An_{45}, and commonly has a cloudy core. Plagioclase also forms the phenocrysts.

The only pyroxene is diopsidic clinopyroxene, which is altering to green hornblende, mainly around its margins.

Hypersthene-bearing Amphibolite

There is only one occurrence of this amphibolite, in which it forms a sigmoidal-shaped dyke. It contains both hypersthene and augite, and the pyroxenes, especially augite, are altering to green hornblende. There are a few late biotites, and plagioclase is generally very diffusely twinned.

KOBBERFJORD NORITE

General

This rock is a pyroxene-mica-orthonorite in which hypersthene is the only pyroxene (Table 8). Occasionally, the hypersthenes form a slight fluxion structure, but generally a medium-grained, non-orientated, xenomorphic texture is formed. In places there is a tendency towards a subophitic texture, and grains often have irregular margins.

Plagioclase is basic labradorite approximately An_{65} in composition. Twinning is usually complex and often diffuse, and some crystals are bent. Hypersthene is not enclosed by plagioclase, but occurs along the plagioclase grain boundaries, and sometimes plagioclase is enclosed within hypersthene, suggesting that the hypersthene in general formed later than the plagioclase. The plagioclase has a high calcium content, and it is probable that the plagioclase, which was first to form, exhausted most of the calcium in the magma so that a calcium-poor orthopyroxene was formed later.

The hypersthene crystals are anhedral, and inclusions of plagioclase, ore, and apatite are fairly common. They are fairly strongly pleochroic from buff to slaty green. Some are cracked and bent, and several are iron-stained.

Biotite laths are ragged and frequently bent, and in places they are apparently growing at the expense of hypersthene. In these cases, there is usually a narrow zone of quartz between the two minerals, and many biotites contain vermicular quartz inclusions (Plate 180).

Occasionally a little greenish hornblende fringes hypersthene and

TABLE 8

MODAL ANALYSES OF KOBBERFJORD NORITE

Mineral	60/47F	60/47G	60/47H	56/1A
Plagioclase	51.43	37.24	65.33	58.16
Hypersthene	16.53	20.64	21.00	3.90
Garnet	19.60	29.42	0.33	21.73
Biotite	3.33	3.42	5.63	10.70
Ore	4.60	5.32	3.60	2.96
Amphibole	2.80	3.04	2.53	2.06
Apatite	0.80	0.64	1.00	0.36
Zircon	-	-	-	0.13
	99.09	99.72	99.32	100.00

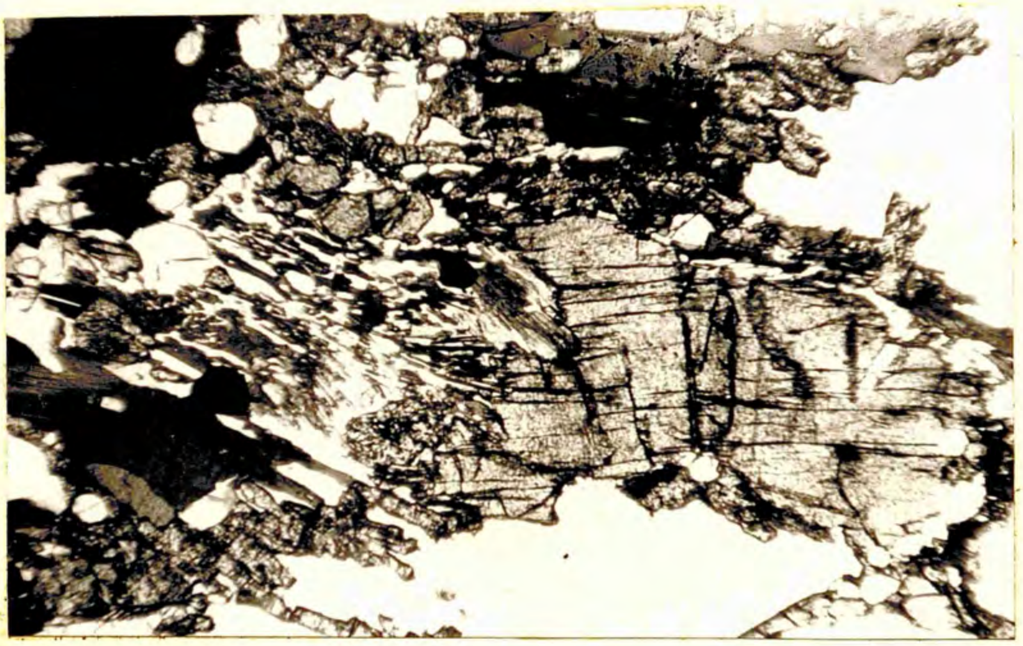


Plate 180. Hypersthene altering to biotite containing quartz vermicules. Kobberfjord norite. P.P.L. X 85.

is forming from it. This hornblende commonly has a turquoise colour, indicating the presence of soda.

Accessory minerals include apatite, which often forms large grains, and ore, the latter including ilmenite, magnetite, pyrrhotite, and pyrite. Ilmenite and magnetite form rounded anhedral grains of various sizes, sometimes in aggregates of blebs. Pyrrhotite is sometimes enclosed within magnetite, but when it is not, it is usually surrounded by haloes of goethite to which it is altering. Alteration takes place by the goethite feathering along cleavages. A few pyrite grains occur, generally contained within pyrrhotite.

Xenoliths

The xenoliths in the Kobberfjord norite vary in size from small rafts down to small partly-assimilated inclusions.

Psammitic and Pelitic Xenoliths

The psammitic xenoliths consist mainly of quartz grains which, together with a few plagioclase grains, form a fine-grained mosaic. A relict gneissic texture is preserved by bands containing biotite, garnet, ore, kyanite, sillimanite, and rutile grains. Biotite is diffuse and altering to sillimanite prisms, and kyanite is also altering to sillimanite. Fan-shaped aggregates of fibrolitic needles nucleate from small garnets.

In the pelites, a relict schistosity is sometimes present, but generally a polygonal-grained hornfelsic texture is evident. The polygonal mosaic is formed by plagioclase, sometimes with quartz, and evidence of strain is common. Diffuse biotites and cracked, buff-coloured garnets constitute the mafic minerals.

Sometimes the pelites contain considerable amounts of green spinel, often occurring in aggregates, and in one case accompanied by a few grains of corundum. In this latter case, many of the spinels have coronas of granular corundum, and also of an indeterminate symplectite (Plate 181). This symplectite consists of two intergrown minerals; one is colourless and the other is pale green. Birefringence is not uniform and the variations do not coincide with the structures of the symplectite. In general it is isotropic, but there are also some irregular slightly birefringent patches; it may be an intergrowth of spinel and corundum.

The margins of the pelites are often diffuse, and appear to be gradational into the norite. The grain-size increases irregularly towards the contact, so that a coarse-grained mosaic containing inclusions of spinel-bearing fine-grained mosaics results. At the contact, many hypersthenes in the norite are rimmed by garnet. This corona development which is characteristic of the norite in the neighbourhood of metasedimentary inclusions is discussed below under a separate heading.

Basic Hornfels Xenoliths

These are fine-grained granoblastic pyroxene hornfelses, consisting essentially of hypersthene and plagioclase. In some cases, hypersthene is slightly altered to green hornblende, and string-like aggregates of hypersthene and hornblende occur along the felsic mosaic grain boundaries.

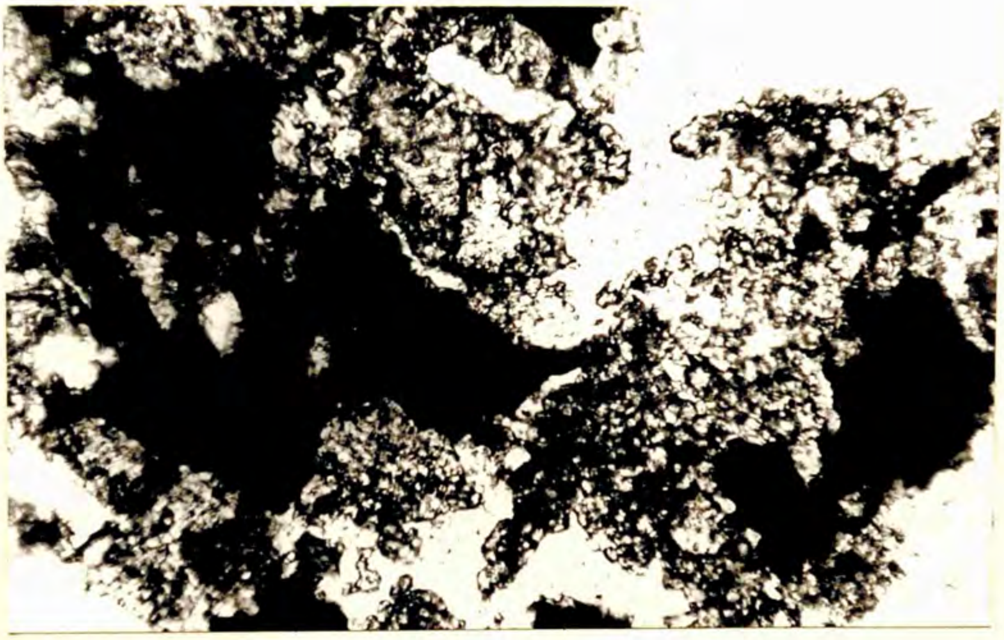


Plate 181. Coronas around spinel grains. Hornfels in Kobberfjord norite. P.P.L. X 170.

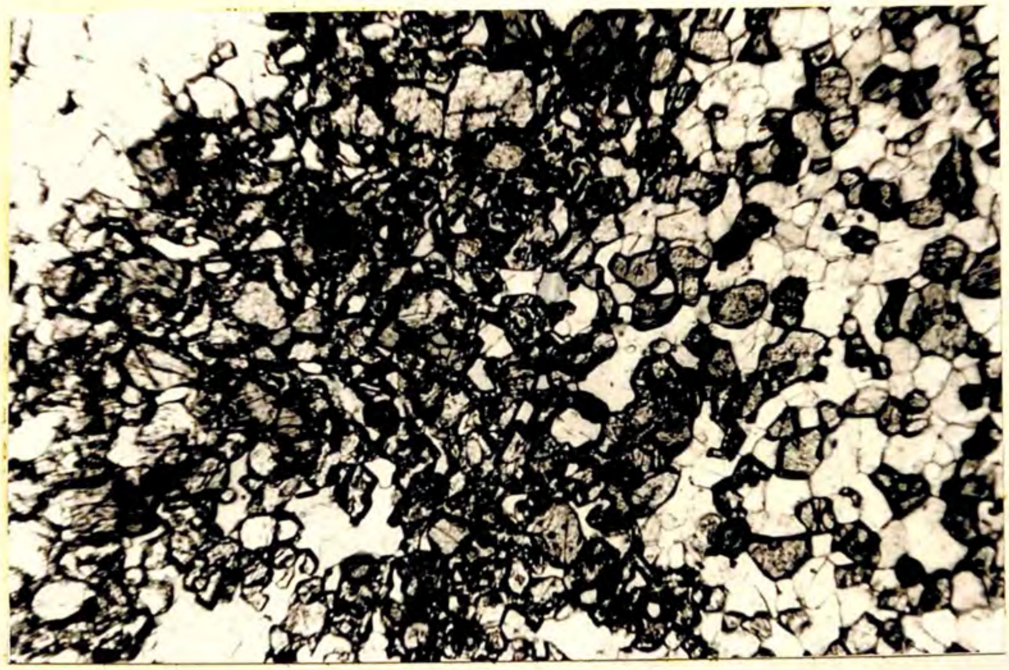


Plate 182. Garnet enveloping hypersthene at margin of hornfels in Kobberfjord norite. P.P.L. X 95.

Ore occasionally forms ragged laths in a decussate pattern.

At the contact with the norite, there is sometimes a narrow zone in which hypersthene grains are enveloped by garnet (Plate 182). The hypersthenes in the norite next to the xenoliths are also rimmed by garnet, and there are also a few discrete garnet grains, containing many inclusions of quartz, biotite, ore, and hypersthene.

The development of these garnet coronas appears to be closely associated with the metasedimentary xenoliths, and are further described below.

Coronas

As in the case of the coarse pyroxene-mica-diorites, complex synantetic textures form, apparently associated with assimilated or partly assimilated metasedimentary xenoliths. These synantetic minerals take the form of coronas.

Some ore grains have dark green hornblende coronas against plagioclase, and others have garnet coronas. Sometimes double coronas occur, with garnet on the outside next to the plagioclase. The relevant reactions were discussed on p. 208 .

Hypersthene is commonly rimmed by pale buff granular garnet, and there is generally a clear zone of quartz between the two minerals (Plates 183 and 184). These garnet rims also occur where the hypersthene is fringed by amphibole. Where hypersthene has altered to biotite-quartz aggregates, these aggregates together with relict hypersthene are also rimmed by garnet (Plate 185). This indicates that the formation of

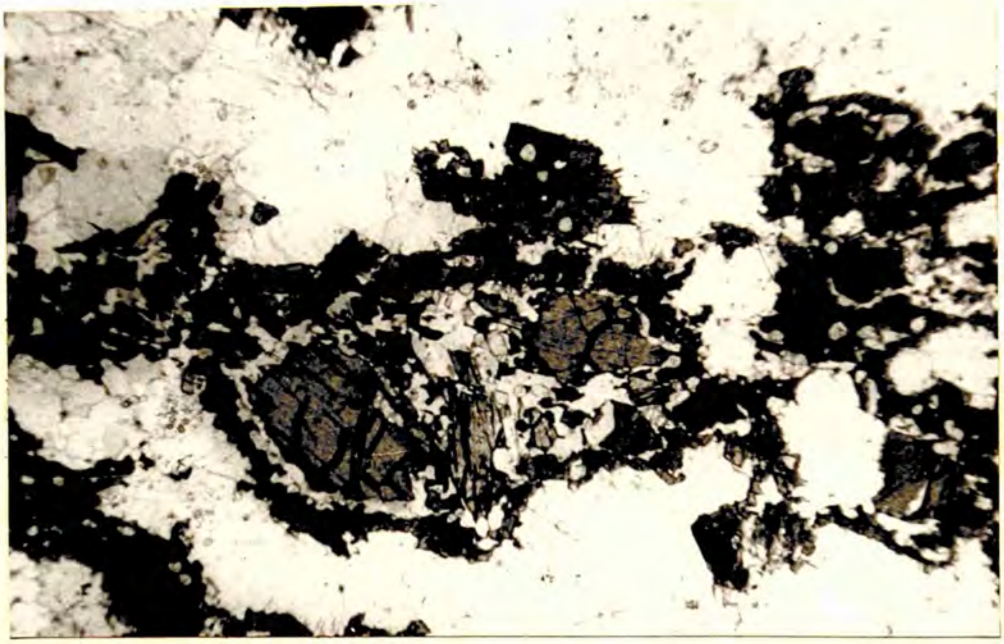


Plate 183. Garnet corona around hypersthene; note clear zone of quartz. Kobberfjord norite. P.P.L. X 25.

Plate 184. Garnet corona around hypersthene; note clear zone of quartz. Kobberfjord norite. P.P.L. X 170.

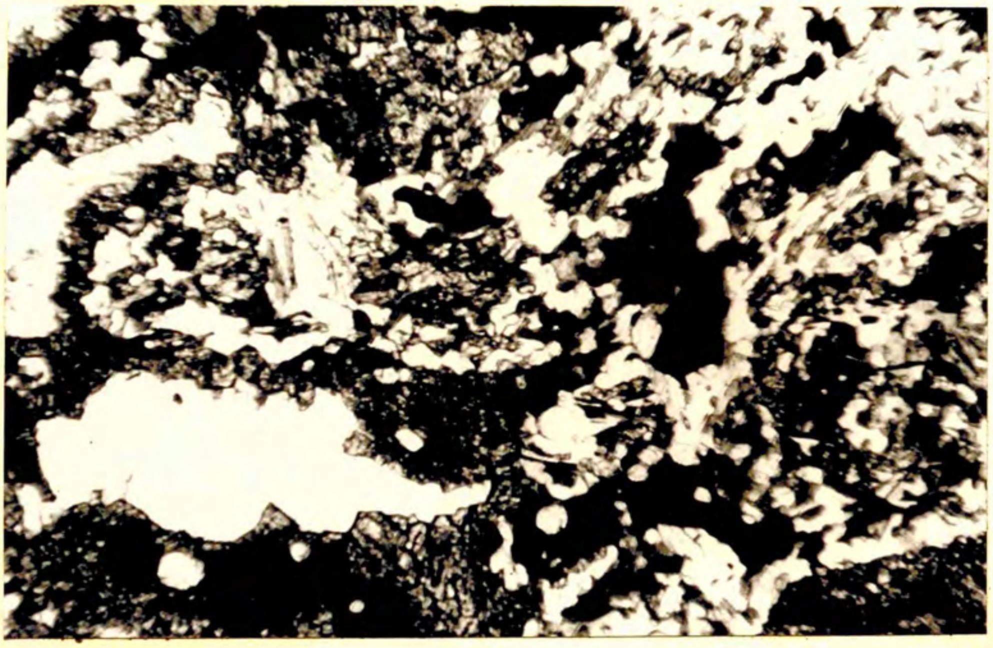


Plate 185. Garnet rimming hypersthene-biotite-quartz aggregate. Kobberfjord norite. P.P.L. X 85.

garnet post-dates the alteration of some of the hypersthenes to biotite.

Garnet also rims and overgrows large biotites (Plates 186 and 187), and occasionally hypidioblastic forms develop. The garnets are full of inclusions of biotite, from which it is clearly forming, ore, quartz, and apatite. Sometimes the garnet has a vermicular texture with numerous quartz vermicules which have been inherited from vermicular biotites (Plate 188). Occasionally where these biotites contain lines of quartz inclusions having a fan-like pattern, this fan shape is preserved as a relict structure by quartz inclusions within the overgrowing garnet (Plate 189).

Garnet coronas around orthopyroxene in norites have been reported by several people.

Friedman (1955) has described norites and olivine-norites from Ontario. In the case of the norites, where pyroxene makes contact with plagioclase a corona of garnet is formed, with a rim of quartz between the garnet and the pyroxene (as observed in the Kobberfjord norite). Hornblende is also locally bluish in colour, like that in the Kobberfjord norite. Friedman considers that these features are metamorphic in origin.

Gjelsvik (1952) described complex coronas in dolerites and olivine-dolerites from southern Norway. Although these are dolerites, they contain hypersthene which is rimmed by multiple coronas against plagioclase. These coronas are more complex than those in the Kobberfjord norite, but it is to be noted that the outermost layer against the plagioclase is always of garnet. In places, these rocks also contain a

Plate 186. Garnet overgrowing biotite. Kobberfjord norite. P.P.L. X 170.

Plate 187. Garnet overgrowing biotite. Kobberfjord norite. P.P.L. X 170.

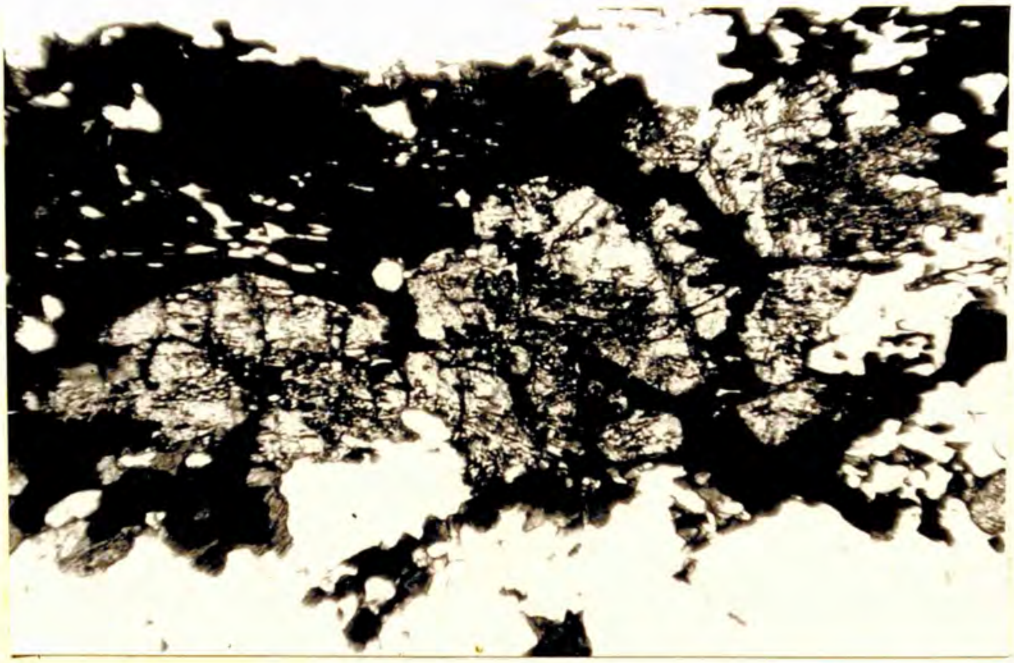


Plate 188. Vermicular garnet overgrowing vermicular biotite. Kobberfjord norite. P.P.L. X 85.

Plate 189. Fan-shaped quartz inclusions inherited from vermicular biotite. Kobberfjord norite. P.P.L. X 85.

bluish-green amphibole. With increase in metamorphism, it was found that discrete garnet porphyroblasts grew.

According to Gjelsvik, corona development may take place basically in one of three ways:

(1) By late magmatic reactions between crystals and residual liquid

(2) By post-magmatic (autometamorphic) reactions

(3) By metamorphic reactions (contact or regional)

Gjelsvik considers that this kind of corona could not form by method (1) because the synantetic minerals are more magnesian than the original minerals. Neither can it be due to (2) since the products are not more hydrous than the original minerals. Gjelsvik therefore considers that they must be due to metamorphic reactions.

A phenomenon very similar to that occurring in the Kobberfjord norite is described by Engels and Vogel (1966) in a metanorite from Spain, in which garnet reaction-rims develop between plagioclase and hypersthene. The hypersthenes have frayed rims and are surrounded by two concentric reaction-rims; the inner is of light bluish-green actinolitic hornblende (only occasionally present in the Kobberfjord norite), and the outer rim is of garnet. Engels and Vogel believe that the reactions took place during metamorphism, and that they were probably as follows:

12 hypersthene + 4 anorthite + 4 albite + 4 water

= 4 hornblende + 4 quartz =

5 garnet + 4 albite + 4 water + 4 quartz

In the case of olivine-norites or hyperites, as in those described by Reynolds and Frederickson (1962), garnet reaction-rims form around the orthopyroxene, which has, in turn, formed from olivine. However, garnet does not form until all the olivine has been converted to orthopyroxene (Brøgger 1934). Reynolds and Frederickson believe that the reason for this is that any available silica is preferentially used in the conversion of olivine to orthopyroxene as this is a simple reaction. When all the olivine has been exhausted, the silica may be used in the formation of garnet, a more complex reaction.

From these accounts it is clear that garnet reaction rims between hypersthene and plagioclase are fairly commonly developed under conditions of metamorphism.

It is deduced that the Kobberfjord norite was emplaced prior to a metamorphic event which was at a sufficiently high grade to cause the formation of garnet. This is discussed further on p. 230 .

Contact Zone

Near the contact with the migmatites of the country rocks, large garnet porphyroblasts form in the norite. These are allotrioblastic and buff-coloured, and markedly cracked. Large inclusions of hypersthene are common (Plate 190), and these are extensively altered. One kind of alteration is to a greenish-yellow serpentinitic substance, mainly along cleavages, and the other kind is to a rusty-brown limonitic alteration product. This alteration only affects those hypersthenes which are enclosed within the garnets. Other inclusions in the garnets are hornblende,

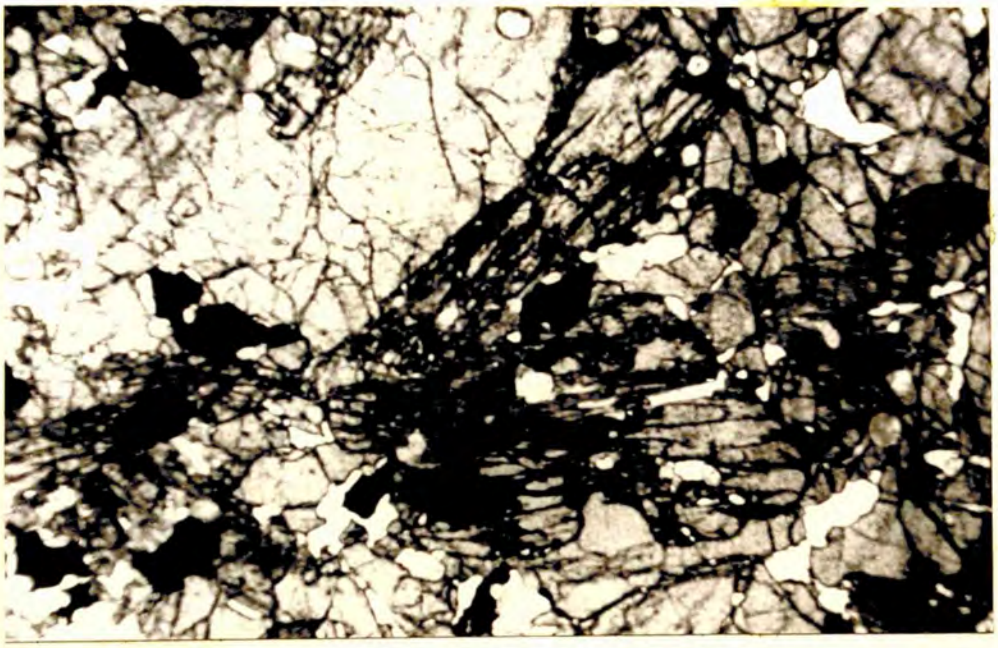


Plate 190. Hypersthene inclusions in garnet porphyroblast. Kobberfjord norite. P.P.L. X 25.

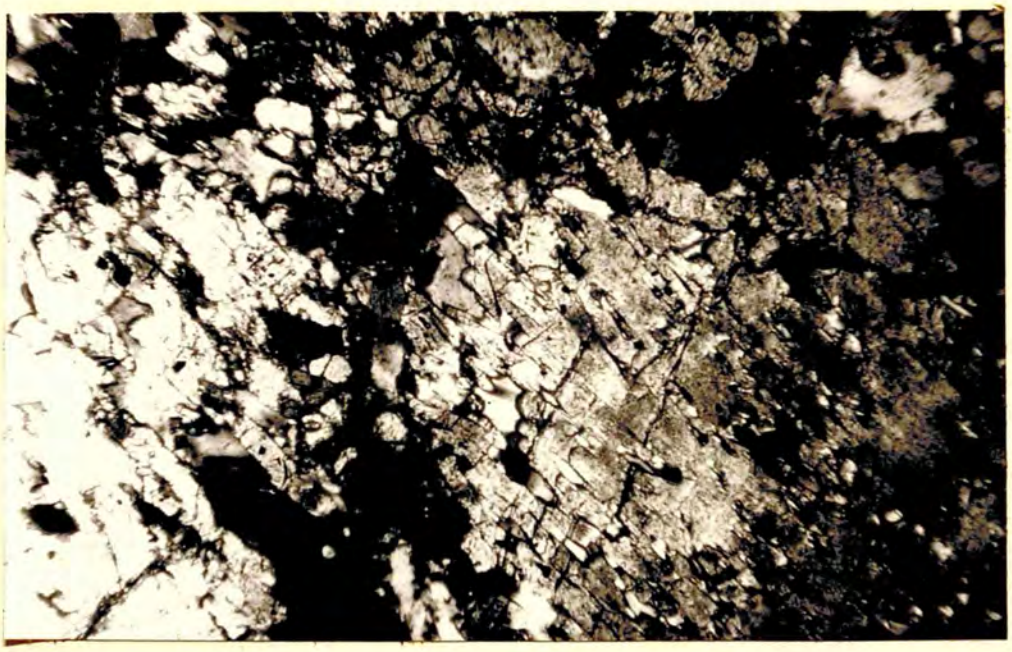


Plate 191. Strained and fragmented sillimanite. Hornfels in thermal aureole of Kobberfjord norite. X-Pl. X 85.

apatite, ore, and occasionally quartz.

In the contact zone itself, the rock is probably a mixture of norite and metasediment, and there is considerable deformation. Pods of quartz aggregates are highly strained and the grain boundaries are sutured. Plagioclase and biotite are also markedly strained, and the latter have a 2V practically equal to zero. Plagioclase has a composition of An_{35}, and forms large grains with diffuse twinning. Many of the crystals are deformed and their margins have sometimes recrystallized.

There is a considerable amount of garnet, which is greatly cracked and iron-stained, and altering to an indeterminate birefringent mineral along cracks. Apatite is quite abundant, and is usually contained within biotite, but it is occasionally enclosed by ore.

The norite has caused a narrow zone of hornfelsing to form in the migmatites of the country rocks at the contact. Here, a hornfelsic texture, with decussate biotite laths, is superimposed upon the foliation, and small green spinels form. The kyanites of the semi-pelites alter to sillimanites, some of which are several millimetres in length. These sillimanites are poikiloblastic and contain inclusions of quartz, feldspar, ore, and spinel. They are highly strained, and have probably suffered crushing from movement along the contact during the F_2 deformation. At their margins they are broken into small rounded fragments which are enclosed within feldspar, but which retain their optical continuity with the parent grain (Plate 191).

Garnet overprints the hornfelsic texture and appears to grow at the expense of biotite, sillimanite, and the remaining kyanite. This indicates that the thermal metamorphism caused by the norite must have been superimposed upon the regional metamorphic event, and that when the temperature of the hornfelsing had dropped to that of the regional metamorphism, the grade of the latter had not fallen beneath the garnet isograd. As a result, garnet re-formed from the sillimanite which was produced during the hornfelsing.

It follows from this that the Kobberfjord norite was emplaced before the grade of the regional metamorphic event had dropped beneath the garnet isograd, and this could explain the presence of garnets in the norite. The close association of garnet formation with assimilated metasedimentary xenoliths is probably not a result of direct contamination, but rather due to the fact that the metasediments provide local sources of the excess aluminium needed for the formation of garnet during metamorphism. This is an example of chemical metastability (see p. 205) in the norite around the xenoliths which became stabilized during metamorphism.

VATNA GABBRO

General

This gabbro has a fine-grained non-porphyritic, xenomorphic texture (Plate 192); a subophitic texture is not developed. The principal mafic mineral is pale greenish-buff augite (Table 9), which generally has irregular blebs and fine schiller needles of ore. These needles are aligned both parallel to the cleavage traces and oblique to them.

The augites, which are anhedral, are occasionally fringed by biotite, which also occurs in patches and along some cleavages. In places the augite can be seen to be altering to biotite.

Plagioclase, which is basic labradorite, has finely-spaced albite twin lamellae, which are often diffuse. Grain margins are irregular and sometimes slightly sutured, and some crystals are bent.

Small rounded or irregular grains of pale green olivine are common, and these are cracked and appear a little cloudy, but are not greatly altered. They are frequently enclosed or partly enclosed by pyroxene.

Ore is interstitial and also occurs as grains enclosed within plagioclase. Ilmenite generally forms large irregular interstitial grains, while magnetite occurs as small somewhat rounded crystals, and the octahedral cleavage of magnetite is commonly seen.

Autoliths

These inclusions in the Vatna gabbro appear to be fine-grained equivalents of the normal gabbro (Plate 193). They have a subequigranular texture consisting of clinopyroxene, diffusely-twinned plagioclase, small

TABLE 9

MODAL ANALYSES OF VATNA GABBRO

Mineral	11/29I	11/38D[1]
Plagioclase	48.46	13.96
Antiperthite	-	67.46
Augite	31.73	12.70
Ore	11.73	4.33
Biotite	4.43	0.20
Olivine	2.66	0.90
Apatite	0.90	0.56
	99.91	100.11

1: Antiperthitic facies.

Plate 192. Texture of Vatna gabbro. X-Pl. X 25.

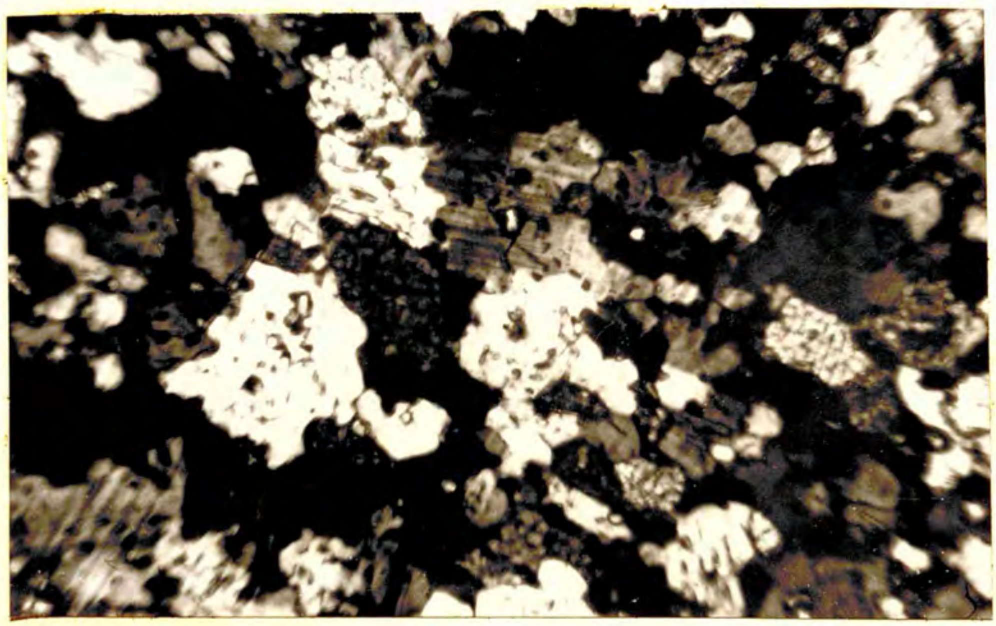


Plate 193. Texture of autolith in Vatna gabbro. X-Pl. X 95.

Plate 194. Fluxion structure of autolith (fine-grained part). Vatna gabbro. P.P.L. X 25.

rounded olivines, and ore.

It is difficult to be sure whether or not they have been significantly thermally metamorphosed by the surrounding gabbro, but the presence of unaltered olivines suggests that they have not.

Some of the autoliths have a foliation which appears to be a fluxion structure (Plate 194).

The margins of these inclusions are fairly well defined, but in places the coarser gabbro penetrates the autolith, and the latter occurs as small fine-grained inclusions in the gabbro (Plate 195). In places they become quite mixed, with the coarser gabbro containing numerous diffusely-bordered, fine-grained patches. Occasionally the autoliths contain plagioclase porphyroblasts, sometimes antiperthitic, and usually containing numerous inclusions.

Xenoliths

The xenoliths are occasional fragments of various rock types.

Some coarse-grained pyroxene-rich inclusions consist mainly of greyish-green augite having strong broad ore schiller lamellae. Some of these lamellae are fairly fine and these are orientated parallel to the cleavage traces, but most of them are quite broad, and intersect the cleavages on longitudinal sections at angles of $23^{\circ} - 27^{\circ}$.

The augite is fringed by a greenish-brown hornblende, which also occurs in patches and along cleavages. There is very little plagioclase, and this has a composition of about An_{50}. It is enclosed within pyroxene, forming a coarse-grained ophitic texture.

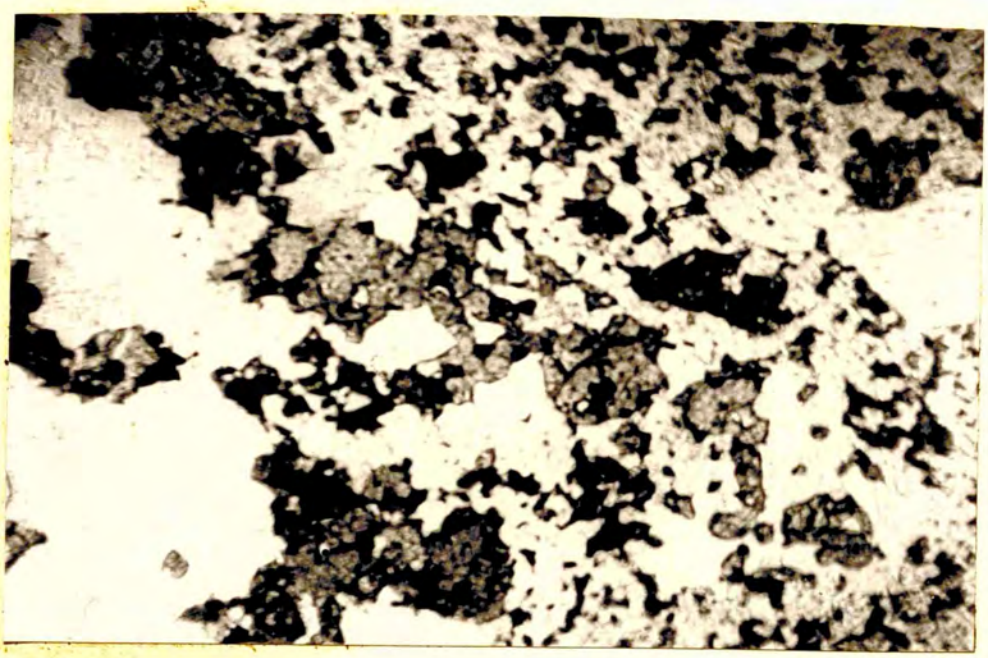


Plate 195. Margin of autolith. Vatna gabbro. P.P.L. X 25.

A few highly mafic, slightly rusty-weathering inclusions consist principally of diopsidic clinopyroxene aggregates with large irregular interstitial grains of ore. A little interstitial plagioclase also occurs.

There are also a few fine-grained amphibolites, in which clinopyroxene has largely altered to a brown hornblende.

All these small mafic inclusions probably represent fragments of different gabbro or dyke material brought up from depth.

Antiperthitic Facies

These bands and streaks of feldspathic material in the gabbro are abundant in antiperthite. This is mainly hair-antiperthite in which the sodic plagioclase contains fine wavy lamellae of potash feldspar. Unfortunately it has not been possible to determine the composition of the plagioclase accurately. Twinning is extremely faint or absent in these grains, although it is more marked in plagioclase crystals which are not antiperthitic. Some crystals are only partly antiperthitic; in these cases twinning is evident in the non-antiperthitic parts but peters out in the antiperthitic areas.

An inequigranular texture is formed, with evidence of recrystallization having taken place at the grain boundaries, which are highly sutured, and show the development of swapped rims in places (Plate 196). Stages in the development of the phenomenon of swapped rims can be seen in the perthosites (p.241), and their formation and significance are discussed further there.

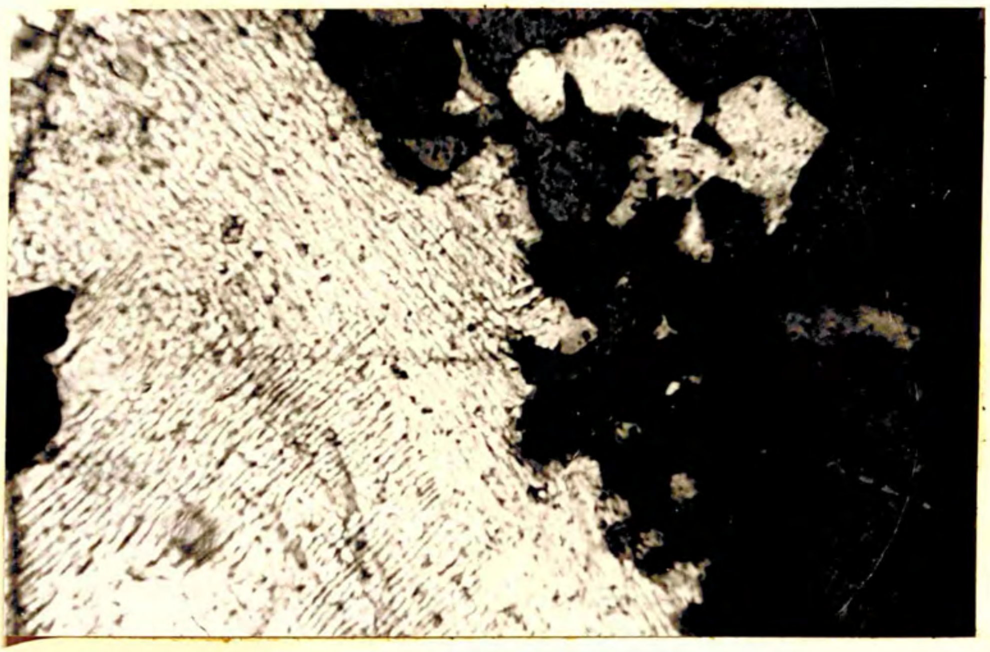


Plate 196. Swapped rims in antiperthite. Vatna gabbro. X-Pl. X 170.

Rounded fragments of olivine are fairly common, and some of the clinopyroxenes show polysynthetic twinning having (100) as composition plane. Ore is often enclosed within pyroxene, and zircon is a common accessory mineral.

Ore-shoots

The antiperthitic streaks in the gabbro very occasionally have clots and shoots of ore. In general, the ore occurs in mafic aggregates in the antiperthitic facies. These aggregates consist of rounded grains of clinopyroxene, olivine, apatite, antiperthite, and zircon which is often large and euhedral. The ore is interstitial to all these minerals, indicating its late formation (Plate 197). Many of the clinopyroxenes are twinned and some of the olivines are pseudomorphed by serpentine or iddingsite. Ore is often rimmed by biotite where it makes contact with antiperthite.

The principal ore is magnetite, with which ilmenite is frequently associated. Although it commonly forms large interstitial masses which send off veins into the gangue minerals, it also forms agglomerations of small blebs and delicate intergrowths with gangue minerals. Other ores which occur in small amounts include pyrite, pyrrhotite, and chalcopyrite.

In some places ore, biotite, olivine, and pyroxene are rimmed by coronas of a cloudy, brownish-grey, translucent, granular material (Plate 198), which often looks as if it might be forming by alteration of pyroxene. It has a low birefringence and has a patchy extinction, and has so far proved indeterminable. Its formation may be associated with some very

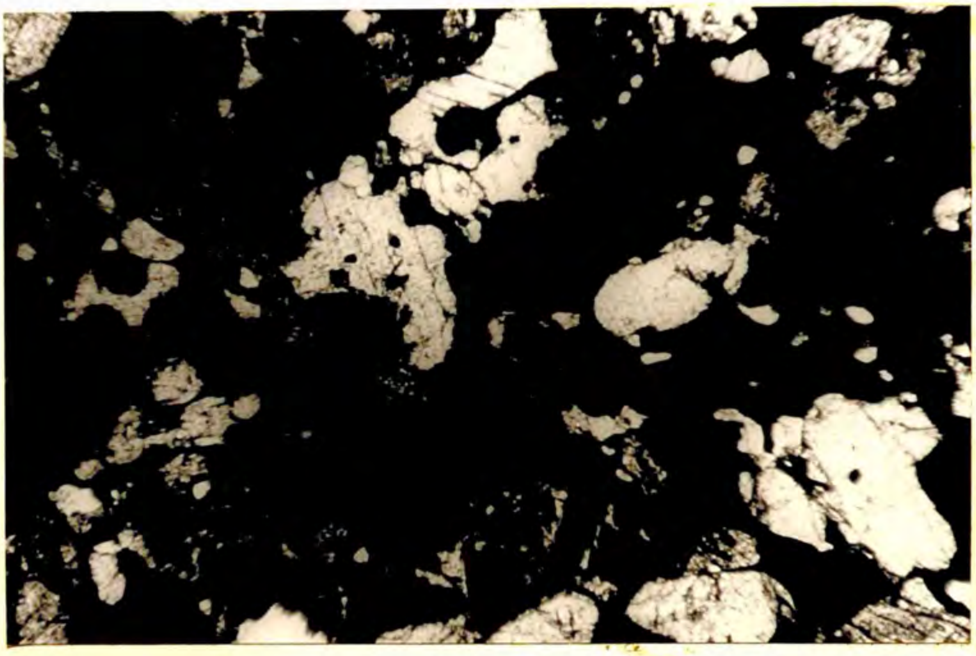


Plate 197. Texture of ore-shoot. Vatna gabbro. P.P.L. X 85.

Plate 198. Coronas around minerals in ore-shoot. Vatna gabbro. P.P.L. X 170.

thin ore-filled veins which penetrate the rock at this point.

Metasomatic Veins

There are a few metasomatic veins in the Vatna gabbro resembling very closely those in the Husfjord metagabbro. They are up to 1 mm. in thickness, and the gabbro is affected for up to 5 mm. on either side.

The veins themselves consist of yellowish-green actinolitic amphibole, which has a slightly turquoise tinge, but they differ from the veins in the Husfjord metagabbro in not having sphene.

Towards the veins, pyroxene of the gabbro becomes progressively amphibolized, and is pseudomorphed by the actinolite. The plagioclase becomes cloudy, but is not markedly sericitized.

The vein margins are ragged but sharply defined. However, as in the Husfjord metagabbro, actinolites of the veins cross the contacts and are joined to those formed in the gabbro. Clearly the mode of formation of these veins is truly metasomatic, like those discussed in the Husfjord metagabbro, and are probably of the same generation.

SLÅTTEN GABBROS

Fine Olivine-gabbro

This resembles some facies of the Vatna gabbro, although in general it is more olivine-rich. Olivines and plagioclase grains are elongated subparallel to one another forming a fluxion structure.

Olivines are fresh and hardly altered internally, although they are often rimmed by hypersthene. Pyroxene is augite and usually contains a number of fine schiller lamellae of ore, which is sometimes so dense that it almost completely occupies the pyroxene. The pyroxene and the olivine are commonly enveloped by a brown hornblende having the following pleochroic scheme:

X straw-yellow

Y mid tan

Z yellow-greenish-brown

This brown hornblende also forms a corona around some of the ore grains, the latter frequently enclosing grains of green spinel, and apatite.

Plagioclase has a composition of about An_{50}, and has complex twinning, pericline twins being common. Crystals are clear and unaltered, but contain clouds of finely-disseminated ore. Sometimes this ore takes the form of fine needles which are generally parallel or subparallel to (010) faces, but are occasionally parallel to (001).

Coarse Olivine-gabbro

This, which is the southernmost of the Slåtten gabbros, has a very coarse-grained ophitic texture (Plate 199). Large augites contain randomly-orientated anhedral laths of plagioclase.

The augite is buff-coloured, and contains numerous ore inclusions and fine ore schiller lamellae. These schiller lamellae occur principally in three directions; parallel to (100), parallel to (010), and subparallel to (101). The augite is beginning to alter to brown hornblende having the following scheme:

X pale straw

Y mid brown

Z greenish-brown

This alteration is taking place mainly at the edges and in irregular patches rather than along cleavage planes. Brown hornblende also rims ore-grains and biotites, some of which are deformed.

Olivines are not common, and when they do occur they show various degrees of alteration. Some are slightly altered to a brownish ore along cracks, whereas others are pseudomorphed by iddingsite, which itself has altered to a reddish-brown iron alteration product. Most olivines, whether pseudomorphed or not, have coronas of faintly pleochroic hypersthene or of fibrous cummingtonite.

Plagioclase is basic andesine, and its twinning is irregular and diffuse, and some crystals are deformed.

Pyroxene-gabbro

This is the northernmost of the Slåtten gabbros, and has a coarse diabasic texture in which broad plagioclase laths are still largely

Plate 199. Ophitic texture of coarse olivine-gabbro at Slatten. P.P.L. X 25.

preserved. However, some recrystallization has probably taken place since a few crystals are zoned. In these the cores are more calcic than the margins, e.g. in one case the core was An_{54} and the margin was An_{30}. Some plagioclase grains contain fine needles of ore.

The pyroxene is pale buff augite containing abundant schiller lamellae of ore. Narrow, needle-like lamellae run parallel to the cleavage traces in longitudinal sections, and short, broad lamellae generally lie orthogonal to the cleavage traces. Ore also forms small rounded inclusions.

Pyroxene is altering to brown hornblende at its margins and also in patches and along cleavages. This hornblende has these characteristics:

X straw

Y deep greenish-brown

Z brownish-green

Hornblende also surrounds ore and biotite, although the latter sometimes envelops the hornblende.

Olivine grains are very rare and small, but fairly well preserved. Apatite is common, and grains are often very large.

THE MINOR INTRUSIONS

EARLY BASIC DYKES

These are relics preserved as xenolithic blocks in the perthosite sheets. They consist of a porphyritic amphibolite, and a peridotite.

Porphyritic Amphibolite

This has a porphyritic texture with numerous anhedral plagioclase phenocrysts (Plate 200). The mafic minerals between the phenocrysts are principally hornblende which can be seen to be forming from clino-pyroxene, which forms relics in the amphibole masses. The hornblende is brownish-green:

X straw

Y brownish olive-green

Z olive-green

Grains of ore are usually rimmed by very thin coronas of sphene.

The original plagioclase was about An_{45}, but many phenocrysts have recrystallized at their margins to a mosaic of more sodic plagioclase (Plate 201). This latter is optically negative and must, therefore, be more sodic than An_{40}. The remaining central parts of the phenocrysts are also slightly zoned, with more sodic margins, but zoning in plagioclase is much more marked where contact with hornblende is made.

Many of the megacrysts, particularly the zoned ones, contain small irregular mafic inclusions, principally of hornblende or pyroxene which is in the process of changing to hornblende. These inclusions are almost

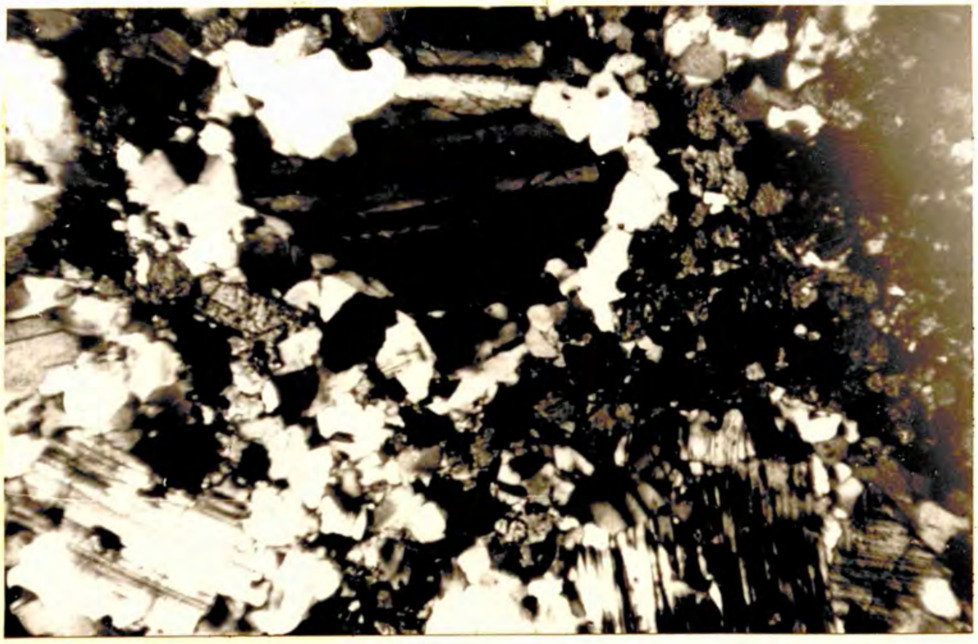


Plate 200. Texture of early porphyritic amphibolite dyke. X-Pl. X 25.

Plate 201. Recrystallized plagioclase phenocryst. Early porphyritic amphibolite dyke. X-Pl. X 25.

invariably surrounded by 'haloes' which extinguish together with the sodic margins if the megacryst is zoned. In these haloes twinning is absent, and at their margins they merge diffusely but rapidly with the rest of the plagioclase (Plate 202). It is possible that their higher sodium content is a result of the 'leaching' away of calcium by the newly-formed hornblende in the inclusions.

Peridotite

There is only a little interstitial plagioclase in this rock, but there is a considerable amount of olivine, which often forms fairly large crystals. Olivine is only just beginning to alter to greenish serpentine along cracks, but ore in the cracks is fairly common. It is usually rimmed or partially rimmed by enstatite.

Much of the rock is occupied by randomly-orientated pale brown hornblende which encloses olivines and relict pyroxenes. The colour scheme of the hornblende is:

X pale khaki

Y light fawn

Z mid khaki

PERTHOSITES

These consist virtually of perthite, although some contain various amounts of mafic minerals (Table 10), and rare sodic plagioclase grains. There are also occasional non-perthitic microcline crystals.

The perthite is principally fine hair-perthite in which potash

TABLE 10

MODAL ANALYSES OF PERTHOSITES

Mineral	11/25A	11/25C	11/42B	07/24A
Perthite	95.70	98.60	92.56	93.20
Ore	2.80	0.16	2.40	1.60
Aegerine-augite	0.46	1.26	-	4.40
Hypersthene	0.13	-	-	0.23
Iddingsite	0.90	-	-	0.43
Clinopyroxene	-	-	-	-
Apatite	0.03	-	4.76	-
	100.02	99.86	99.98	99.86

Plate 202. Haloes around mafic inclusions in phenocrysts. Basic dyke. X-Pl. X 85.

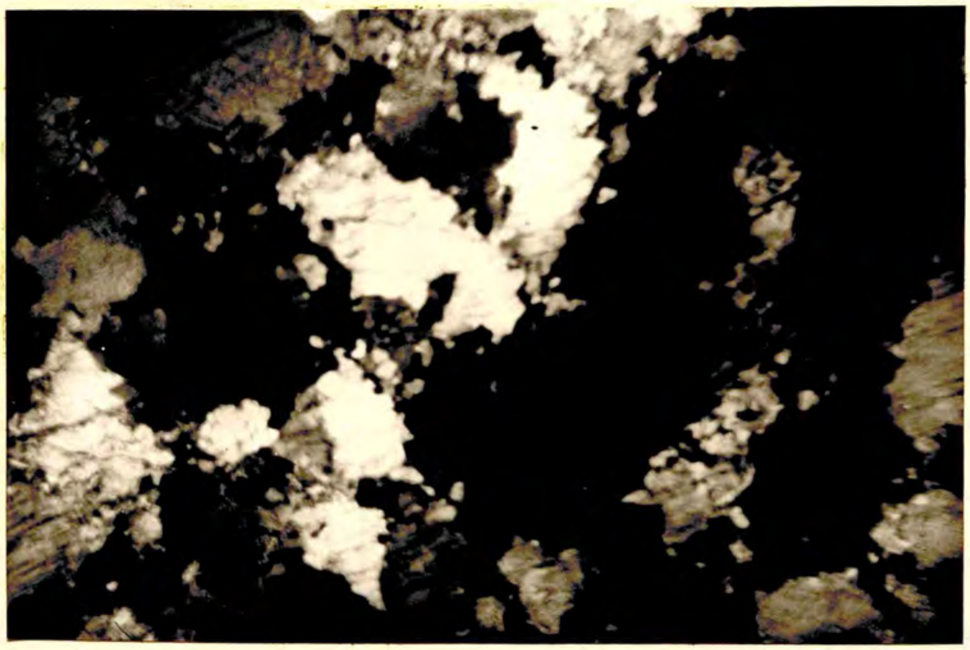


Plate 203. Texture of perthosite; note sutured margins. X-Pl. X 25.

feldspar and sodic plagioclase are in approximately equal proportions. Generally the feldspar grains are elongate, and tend to have a preferred orientation forming a fluxion structure (Plate 203).

An inequigranular texture of anhedral grains is formed, in which the grain margins are highly sutured and swapped rims are commonly seen to be developing. The phenomenon of swapped rims is best described by tracing the steps in its development. In early stages the suturing of the grain boundaries becomes more and more marked until neighbouring crystals have interlocking contacts (Plate 204). As the interlocking becomes more exaggerated, those parts of the crystal which lobe into the neighbouring grain become detached from the parent crystal, and occur as outliers in the neighbouring grain. The lamellae in these outliers extinguish together with those of the parent crystal, and are thus in optical continuity with it (Plate 205). This is the most advanced stage seen in these rocks, but if this process were continued further, then the amount of outlying material would increase until true swapped rims formed as described by Voll (1960).

The development of these swapped rims is clearly a marginal recrystallization phenomenon, and is probably a later metamorphic effect.

At the margins of most perthite grains, especially where the incipient swapped rims are forming, the perthitic lamellae peter out, and a homogeneous zone occurs. This also takes place against some inclusions in the perthites, around which non-perthitic haloes form (Plate 206). This also seems to be a recrystallization feature, especially since it is common where swapped rims are developing. It would appear that due

Plate 204. Interlocking boundaries between perthite grains. Perthosite. X-Pl. X 10.

Plate 205. Development of swapped rims in perthosite. X-Pl. X 10.

Plate 206. Non-perthitic haloes around inclusions in perthite. Perthosite. X-Pl. X 170.

to reheating, the perthite has become homogenized and has not re-exsolved as separate phases. Homogenization of alkali feldspars in the Ardara pluton, Donegal, where it has been reheated by the Main Donegal Granite has been reported by Hall (1965a).

Most of the perthosites contain occasional accessory minerals as interstitial grains or corroded fragments. These minerals include diopside, ore, biotite, augite, apatite, zircon and green hornblende. In some perthosites there are crystals of aegerine-augite with the following pleochroic scheme:

X emerald green

Y yellowish-green

Z pale emerald green

Hypersthene is fairly common, but is clearly unstable in its present environment; it is highly corroded and more or less pseudomorphed by yellowish-brown iddingsite (Plate 207).

Occasionally where perthite meets clinopyroxene of a dyke inclusion, aggregates of vermicular green diopsidic clinopyroxene in a plagioclase mosaic are formed (Plate 208). It appears that there was also a chemical instability here.

LATE BASIC DYKES

Due to lack of space, description of the numerous late basic dykes must necessarily be of a summary nature in order to avoid undue repetition. The minerals that are present in each of the different dykes are shown

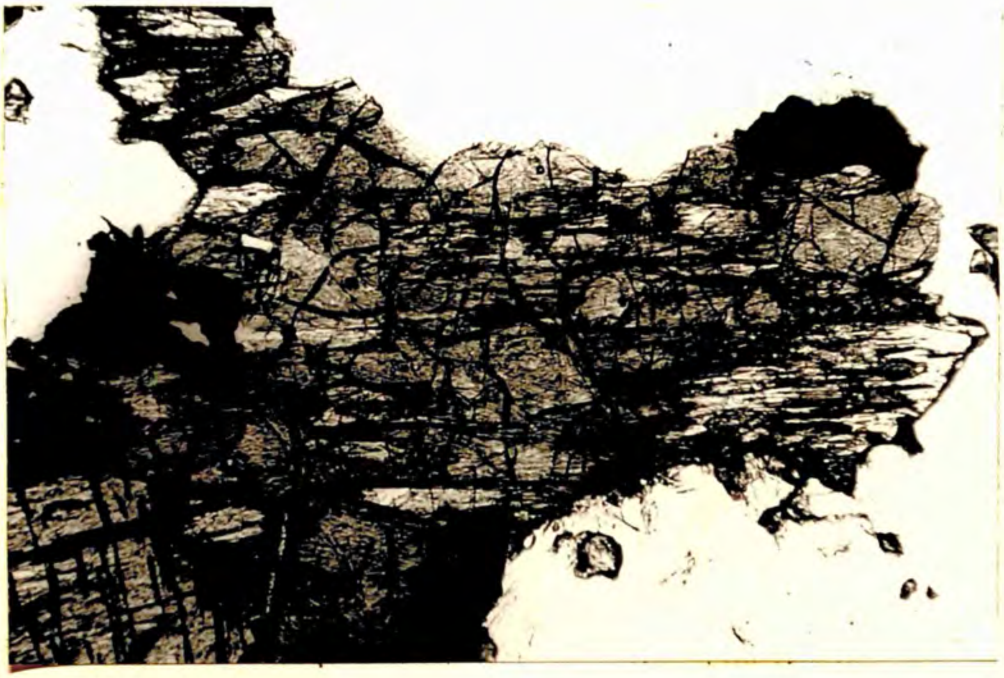


Plate 207. Hypersthene being pseudomorphed by iddingsite. Perthosite. P.P.L. X 170.

Plate 208. Vermicular pyroxene in plagioclase. Vatna gabbro next to perthosite. P.P.L. X 85.

in Table 11 which is to be used in conjunction with Fig. 15.

The dykes vary greatly in grain-size, but the majority of them are fairly fine-grained. Some are abundant in plagioclase phenocrysts. whereas others are non-porphyritic, and many exhibit ophitic textures (Plate 209).

Clinopyroxene, which sometimes forms large phenocrysts (Plate 210), commonly contains schiller lamellae of ore, in which the individual ore needles are generally orientated parallel to (100) and (010) faces and occasionally parallel to ($\bar{1}01$). Sometimes the schiller occurs in concentric zones parallel to the (100) and (010) faces (Plate 211). The clinopyroxene is principally augite, and is invariably altering to brownish-green hornblende, particularly around the margins. Both the pyroxene and the secondary hornblende occasionally alter to biotite.

Some of the dykes contain olivine, and are amphibolized olivine-dolerites. Olivine often forms anhedral phenocrysts, and it is usually fairly fresh. However, in some rocks olivine is rimmed by orthopyroxene and has been variably replaced by serpentine and ore or by iddingsite. In some cases, the olivines have been completely pseudomorphed (Plate 212).

Plagioclase occurs both as phenocrysts and in the matrix, and generally has a composition in the range An_{45-50}. Phenocrysts are anhedral to subhedral, and are sometimes zoned (Plate 213). The cores of these zoned crystals are about An_{50} whilst the margins are about An_{35}, and this is probably a later metamorphic effect.

Some of the plagioclase phenocrysts contain small randomly-orientated

TABLE 11

SUMMARY OF MINERALS PRESENT IN LATE BASIC DYKES

Dyke	OP	CP	O	H	Of	Oa	B	A	Pp	Pn
A		x	x	x	x		x	x		x
B	x	x	x	x		x			x	
C		x	x	x			x	x	x	
D	x	x	x	x	x					x
E	x	x	x	x		x	x		x	
F		x	x	x		x				x
G		x	x	x			x		x	
H	x	x	x	x	x				x	
I	x	x	x	x	x					x
J	x	x	x	x		x			x	
K		x	x	x			x		x	
L	x	x	x	x	x	x	x		x	
M	x	x	x	x	x				x	
N		x	x	x					x	
O		x	x	x	x	x			x	
P		x	x	x	x				x	
Q		x	x	x	x				x	
R		x	x	x			x			x

KEY; OP - Orthopyroxene (around olivine); CP - Clinopyroxene (augite); O - Ore; H - Hornblende (brownish-green); Of - Olivine (fairly fresh); Oa - Olivine (altered to serpentine and iddingsite); B - Biotite; A - Apatite; Pp - Plagioclase (porphyritic texture); Pn - Plagioclase (non-porphyritic texture).

Letters A - R correspond with those in Fig. 15.

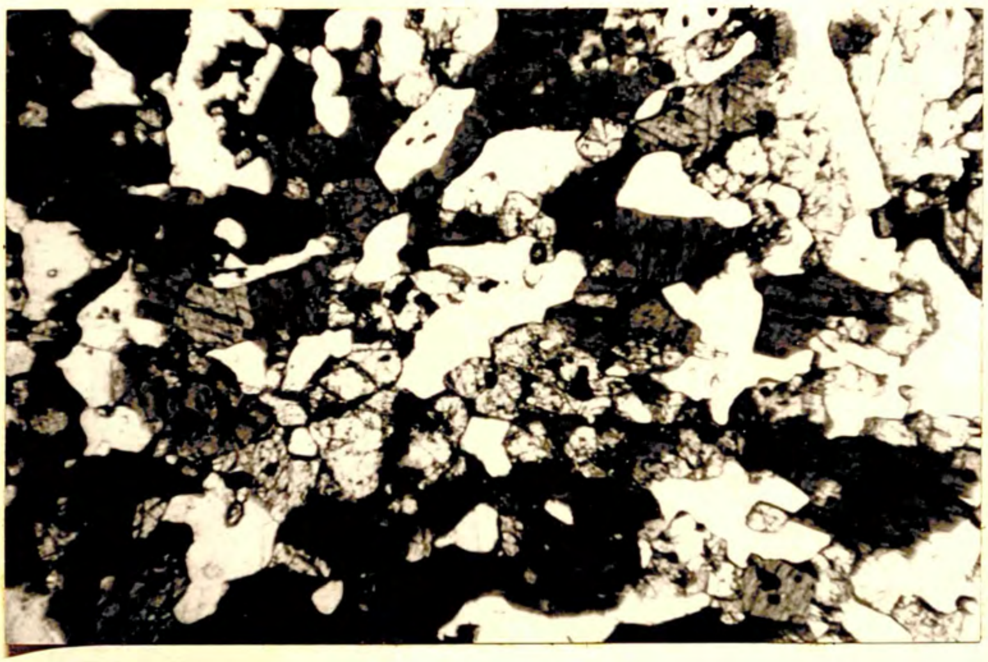


Plate 209. Ophitic texture of late basic dyke.
P.P.L. X 85.

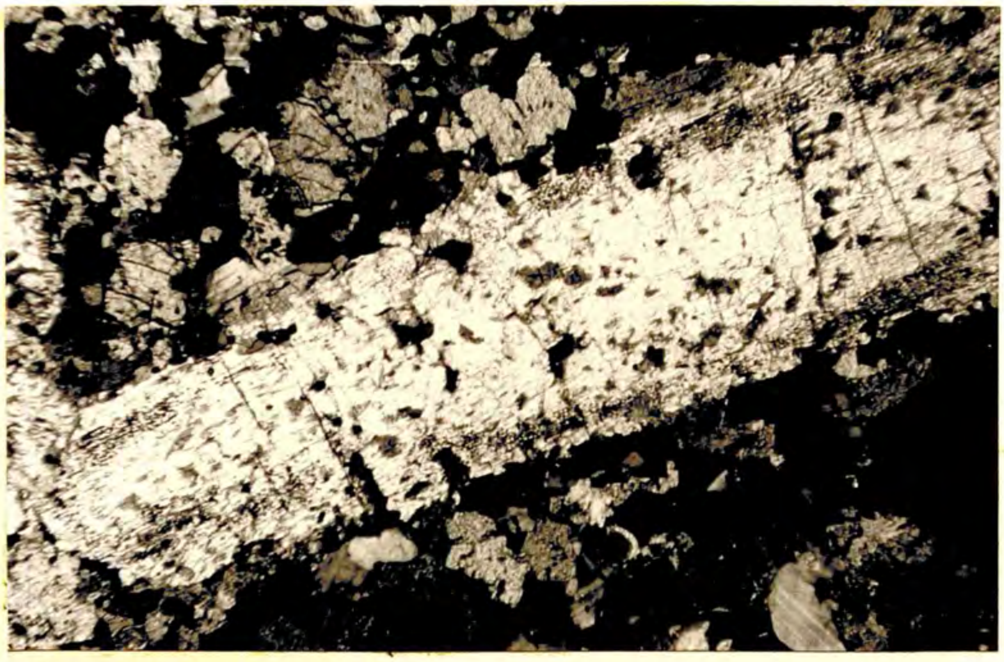


Plate 210. Clinopyroxene phenocryst altering to brownish hornblende. Late basic dyke. X-Pl. X 25.

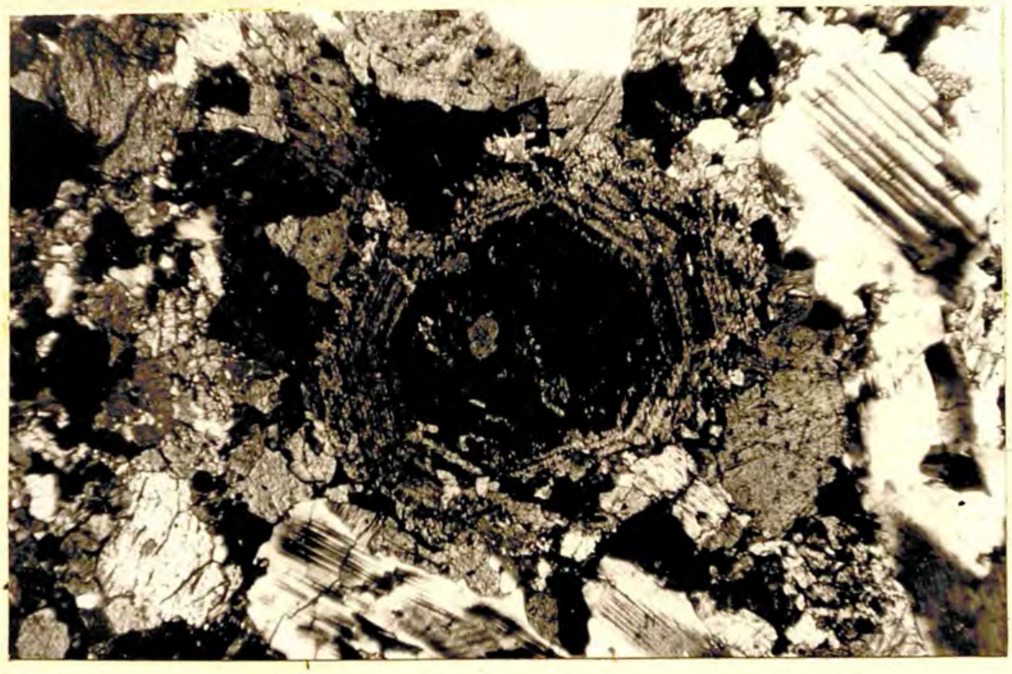


Plate 211. Zones of schiller needles of ore in clinopyroxene basal section. Late basic dyke. X-Pl. X 25.

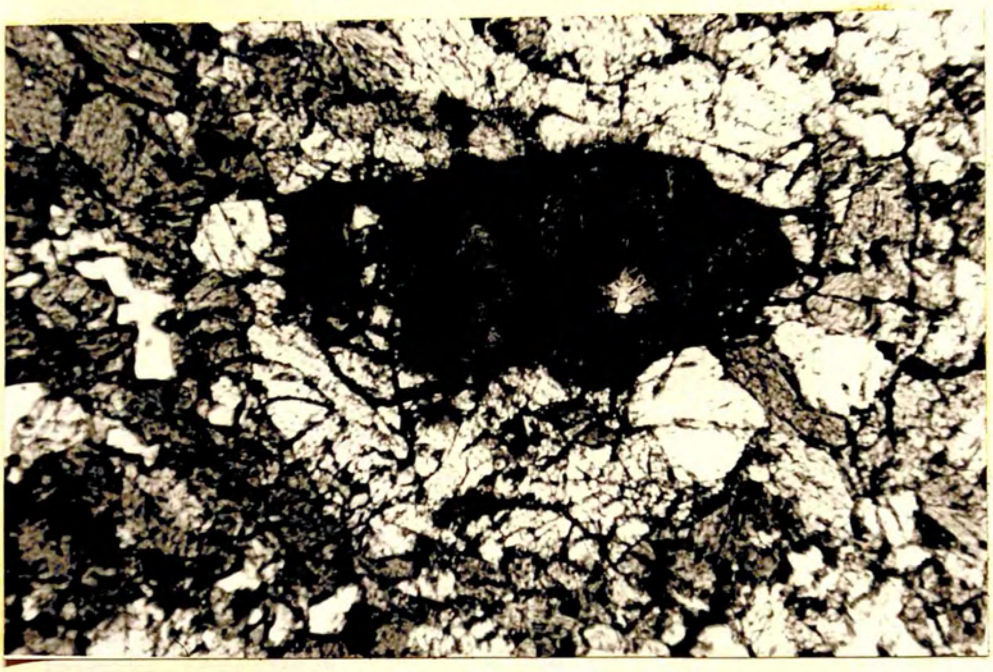


Plate 212. Orthopyroxene rimming iddingsite pseudomorph after olivine. Late basic dyke. P.P.L. X 95.

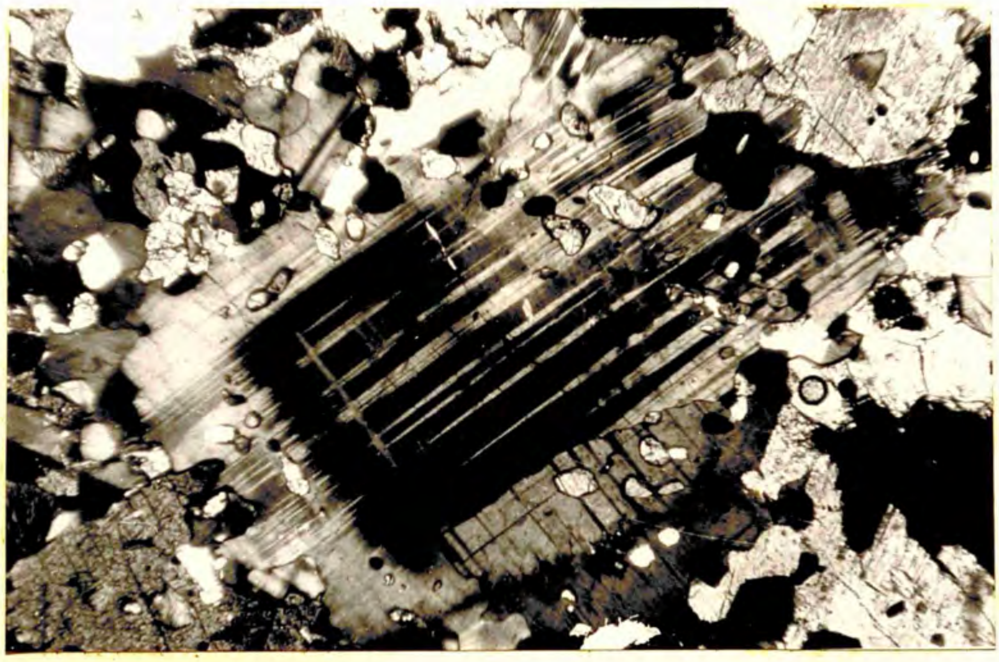


Plate 213. Zoned plagioclase phenocryst. Late basic dyke. X-Pl. X 85.

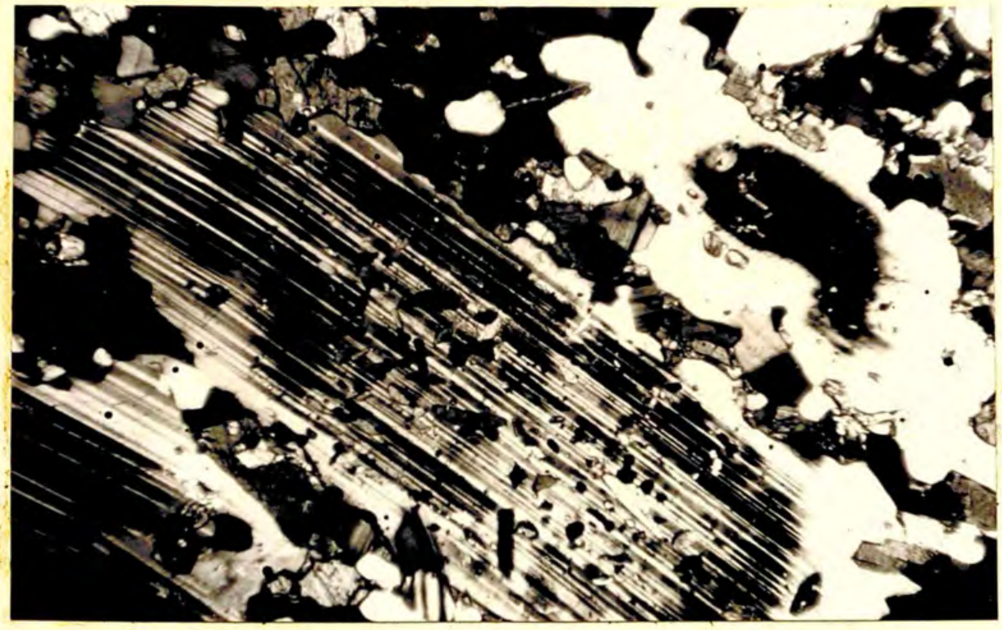


Plate 214. Haloes (light patches) around biotite inclusions in plagioclase. Late basic dyke. X-Pl. X 85.

inclusions of hornblende, biotite, apatite, and ore. The biotite inclusions almost invariably have haloes around them in which the plagioclase is more sodic than in the bulk of the plagioclase crystals (Plate 214). It is possible that during regional metamorphism there was a transference of alkalis from the biotite to the surrounding plagioclase, accounting for the observed haloes.

Although the mineralogy of the late basic dykes is not very variable, there is more variety in the textures and grain-size. This variety, in conjunction with the differences in mineralogy, enable the dykes to be classified into the groups shown in Fig. 15 and Table 11.

The Vatna gabbro and some of the latest dykes have been cut by narrow metasomatic veins resembling those in the Husfjord metagabbro. The veins consist of turquoise actinolitic amphibole whose fibres tend to be orientated subperpendicular to the vein walls. The margins of the veins are sharply defined against hornblende but are diffuse against augite and plagioclase.

Hydrous minerals are formed in the rock adjacent to the veins; olivine is pseudomorphed by green serpentine and fibrous talc, pyroxene is altered to sheafs of fibrous tremolite-actinolite, and hornblende is replaced by chlorite.

It is clear that the emplacement of the veins post-dates the alteration of the pyroxene in the dykes to brownish-green hornblende, since this secondary hornblende has been affected by the metasomatism. Thus the veins are probably associated with a metasomatism which took place at a late stage in the regional metamorphic event.

NEPHELINE-SYENITE PEGMATITES

These are coarse-grained leucocratic rocks consisting primarily of potash feldspar and nepheline, the potash feldspar sometimes showing diffuse microclinic twinning. There are occasional grains of poorly-twinned sodic plagioclase; these are optically positive and have refractive indices less than that of balsam, which indicates that it is albite, or very sodic oligoclase.

The nepheline-syenite pegmatites have been extensively sheared, and even the comparatively unsheared parts have a mortar texture (Plate 215). Crystals of feldspar and nepheline are embedded in a fine-grained granulated matrix to which they are contributing (Plate 216).

The feldspars appear to be more resistant to crushing than are the nephelines. They form ovoid augen, some of which have tails streaked out into the pressure-shadow area on either side of the augen (Plate 217). Some of the feldspar augen are cracked and granulation has taken place along some of the cracks (Plate 218).

The nepheline, however, breaks down more readily and is considerably sericitized, particularly at the margins and along cracks. Lenses of sericitized nepheline become streaked out along the foliation planes of shearing, and sometimes relict fragments of augened nepheline can be seen in the midst of these masses (Plate 219).

Occasional aggregates of small biotites are bent around the augen, and ore has recrystallized along many of the shear planes.

Where mafic material has been caught up in the shearing, it is crushed to a dark green mylonite (Plate 220). This consists mainly

Plate 215. Nepheline in mortar textured nepheline-syenite pegmatite. X-Pl. X 25.

Plate 216. Mortar texture in nepheline-syenite pegmatite. X-Pl. X 25.

Plate 217. Augened feldspar in nepheline-syenite pegmatite. X-Pl. X 85.

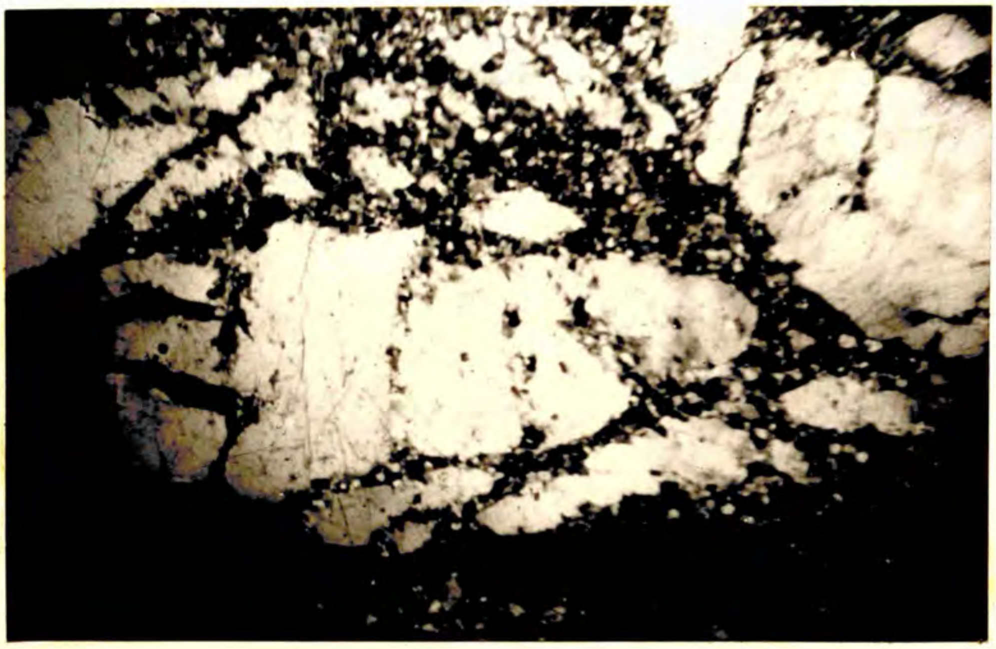


Plate 218. Granulation along cracks in feldspar augen. Nepheline-syenite pegmatite. X-Pl. X 25.

Plate 219. Augened and sericitized nepheline. Nepheline-syenite pegmatite. X-Pl. X 85.

Plate 220. Mylonitized mafic minerals. Nepheline-syenite pegmatite. X-Pl. X 25.

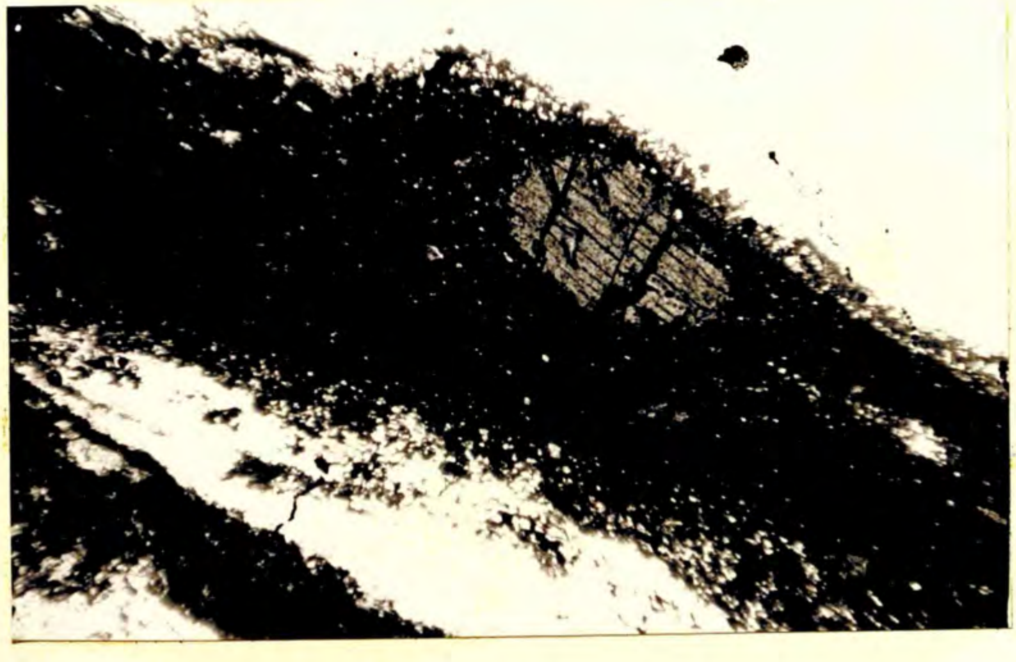


Plate 221. Tension cracks in hornblende augen. Nepheline-syenite pegmatite. P.P.L. X 85.

of hornblende, ore, zircon which is fairly resistant, and fragments of pyroxene, biotite, and sphene. The hornblende, which is yellowish-green in colour, forms small ovoid augen around which the fine granulated mylonite swings. The pleochroic scheme of the hornblende is:

X mid straw

Y mid green

Z mid green

The hornblendes that form the augen commonly have cracks perpendicular to the foliation planes (Plate 221). These cracks are presumably due to tensional forces acting parallel to the foliation planes.

PETROCHEMISTRY

INTRODUCTION

Chemical analyses of 43 rocks and 36 feldspars from the Husfjord area have been made by the writer.

The majority of the rocks analysed belong to the suite of diorites, but a number of Husfjord metagabbro and hybrid specimens were also analysed, as well as a representative from each of the remaining major igneous rock types. The rocks were analysed for the major elements and also for a number of trace elements (Tables 12 - 16).

Coexisting potash feldspar and plagioclase from the diorites were analysed, as well as plagioclase from the Husfjord metagabbro and a number of other rock types. The analyses of the feldspars were partial, only including potassium, sodium, and calcium. The trace elements Ba, Rb, and Sr were also determined (Tables 17 and 18).

The analyses of the rocks were carried out in the Geology Department, University of Bristol, by X-ray fluorescence techniques (with the exception of FeO and H_2O which were determined in the Geology Department, Bedford College, by classical methods).

The feldspar analyses were made in the Mineralogisk-Geologisk Museum, Oslo, the alkalis being determined by flame photometry, and the remaining elements by fluorescent X-ray spectroscopy.

TABLE 12

CHEMICAL ANALYSES OF HUSFJORD METAGABBRO AND HYBRIDS

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
Major elements	36/1A	36/3A	36/54A	36/11F	36/11G	11/64A	36/118F	36/225A	36/146A	36/217A	36/237B
SiO_2	48.35	46.23	47.25	45.65	46.48	46.40	49.86	49.56	45.76	48.00	44.85
TiO_2	2.31	2.42	1.72	2.32	2.10	2.35	2.05	2.28	2.15	1.66	2.46
Al_2O_3	16.85	15.45	17.10	15.87	16.32	16.94	17.83	16.27	16.89	17.00	17.00
Fe_2O_3	2.21	2.64	1.53	2.30	2.87	2.36	3.34	3.67	1.70	2.82	2.02
FeO	8.52	8.47	8.38	8.82	7.72	8.03	7.13	8.88	8.56	7.18	8.44
MnO	0.11	0.11	0.10	0.11	0.09	0.09	0.09	0.11	0.09	0.08	0.10
MgO	6.34	6.57	6.98	7.56	7.32	6.04	4.26	6.48	6.51	6.53	6.85
CaO	10.19	10.69	10.63	9.05	9.64	11.87	6.30	6.86	9.74	8.88	9.63
Na_2O	3.40	3.04	2.76	2.13	2.93	3.23	3.31	2.34	2.44	2.66	2.62
K_2O	1.24	0.71	0.91	2.38	1.10	0.71	2.38	2.27	1.67	2.22	1.38
P_2O_5	0.15	0.17	-	0.15	0.14	0.16	0.10	0.08	0.15	0.12	0.88
H_2O^+	0.71	0.68	0.56	0.67	0.95	0.59	1.03	2.33	0.69	0.98	0.70
H_2O^-	0.11	0.07	0.08	0.10	0.29	0.17	0.12	0.19	0.19	0.23	0.14
Total	100.49	97.25	98.00	97.11	97.95	98.94	97.80	101.32	96.54	98.36	97.07
Trace elements											
S	1486	2067	1827	421	871	362	371	252	203	259	528
V	79	269	387	177	120	257	92	234	212	220	290
Cr	tr.	tr.	-	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Co	31	40	-	16	41	-	-	28	18	28	46
Ni	104	115	-	142	173	108	18	85	55	41	118
Rb	20	10	31	99	28	22	46	35	29	89	30
Sr	365	417	371	347	443	394	476	257	420	378	458
Ce	69	70	-	61	59	49	97	93	64	80	71
CIPW norms											
Q	-	-	-	-	-	-	-	0.07	-	-	-
Or	7.33	4.20	5.38	14.07	6.50	4.20	14.07	13.42	9.87	13.12	8.16
Ab	23.96	23.71	22.05	15.79	24.02	19.45	28.01	19.80	18.98	22.51	22.17
An	27.05	26.42	31.58	26.71	28.13	29.63	26.77	27.19	30.20	27.89	30.55
Ne	2.61	1.09	0.71	1.21	0.42	4.27	-	-	0.90	-	-
Di	18.37	20.75	17.27	13.98	15.06	22.97	3.13	5.15	14.00	12.46	9.33
Hyp	-	-	-	-	-	-	14.41	23.34	-	0.92	0.16
Ol	12.41	11.52	14.89	16.49	14.08	9.41	1.30	-	14.81	12.74	16.23
Mt	3.20	3.83	2.22	3.33	4.16	3.42	4.84	5.32	2.46	4.09	2.93
Il	4.39	4.60	3.27	4.41	3.99	4.46	3.89	4.33	4.08	3.15	4.67
Ap	0.35	0.39	-	0.35	0.32	0.37	0.23	0.19	0.35	0.28	2.04
Total	99.67	96.50	97.36	96.34	96.68	98.18	96.65	98.80	95.66	97.15	96.23
Niggli values											
si	111.75	107.72	108.94	105.92	107.94	105.80	135.17	122.08	108.85	115.79	105.21
al	22.95	21.22	23.24	21.70	22.34	22.76	28.49	23.62	23.68	24.17	23.50
fm	42.37	44.17	42.99	47.49	45.52	40.06	40.40	49.12	43.33	43.24	44.27
c	25.24	26.69	26.26	22.50	23.91	29.00	18.30	18.11	24.83	22.95	24.21
alk	9.45	7.92	7.51	8.31	8.23	8.17	12.82	9.16	8.16	9.64	8.02
qz	-26.05	-29.08	-21.10	-27.32	-24.98	-26.88	-16.11	-14.56	-23.79	-22.77	-26.87

(1) to (3) - fine-grained facies; (4) to (6) - coarse-grained facies; (7) and (8) - feldspathized metagabbro; (9) to (11) - hybrid.

Analyst: the writer

TABLE 13

CHEMICAL ANALYSES OF KOBBERFJORD NORITE, HAVNEFJORD DIORITE, AND PYROXENE-MICA-DIORITES

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)
Major elements	60/47H	36/99A	36/116A	36/117C	36/233C	36/237A	38/1A	38/18B	38/38A
SiO_2	46.47	52.90	46.87	50.79	48.40	50.28	54.63	55.34	56.58
TiO_2	3.45	1.16	2.27	1.89	2.89	1.44	1.38	1.36	1.09
Al_2O_3	16.30	18.78	16.83	19.90	18.26	17.85	17.75	16.69	18.24
Fe_2O_3	1.02	1.32	2.25	1.26	1.18	1.29	1.86	1.28	1.13
FeO	15.16	7.07	8.11	8.59	10.46	6.38	4.73	6.42	4.58
MnO	0.17	0.07	0.09	0.10	0.10	0.07	0.05	0.07	0.04
MgO	3.12	3.17	6.34	3.78	2.55	1.61	2.48	2.74	1.55
CaO	7.27	6.34	10.23	7.25	5.34	4.64	4.55	3.89	3.85
Na_2O	3.40	4.28	2.69	3.60	3.03	4.20	3.79	3.68	3.94
K_2O	1.20	1.35	1.14	2.08	3.24	3.53	2.89	4.67	4.97
P_2O_5	0.29	0.09	0.14	-	-	-	0.06	0.06	0.05
H_2O^+	0.81	0.45	0.76	0.39	0.58	0.60	0.95	0.69	0.50
H_2O^-	0.19	0.11	0.14	0.11	0.08	0.14	0.12	0.13	0.17
Total	98.85	97.09	97.80	99.74	96.11	92.03	95.34	96.52	96.69
Trace elements									
S	1786	499	304	385	531	814	291	188	252
V	-	189	248	217	309	283	182	216	311
Cr	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Co	-	11	26	18	-	-	7	-	-
Ni	27	30	54	15	12	19	18	15	10
Rb	44	21	23	30	45	69	73	114	115
Sr	389	483	398	431	333	447	546	464	514
Ce	110	93	68	85	161	116	47	32	78
CIPW norms									
Q	-	0.79	-	-	-	-	5.99	1.42	1.90
Or	7.09	7.98	6.74	12.29	19.15	20.86	17.08	27.60	29.37
Ab	28.77	36.22	22.76	30.46	25.64	34.82	32.07	31.14	33.34
An	25.67	28.05	30.48	32.00	26.49	19.43	22.18	15.23	17.41
Ne	-	-	-	-	-	0.39	-	-	-
Cor	-	-	-	-	0.06	-	0.26	-	-
Di	7.27	2.35	15.75	3.31	-	3.05	-	3.09	1.15
Hyp	4.09	16.83	2.00	4.92	7.68	-	11.14	12.64	9.04
Ol	16.25	-	-	10.84	9.23	8.14	-	-	-
Mt	1.48	1.91	3.26	1.83	1.71	1.87	2.70	1.86	1.64
Il	6.55	2.20	4.31	3.59	5.49	2.73	2.62	2.58	2.07
Ap	0.67	0.21	0.32	-	-	-	0.14	0.14	0.12
Total	97.85	96.53	96.96	99.24	95.45	91.29	94.17	95.70	96.02
Niggli values									
si	117.06	153.06	111.02	133.08	138.25	164.42	182.42	182.75	
al	24.20	32.02	23.50	30.73	30.74	34.40	34.93	32.48	
fm	45.90	33.82	42.64	36.29	38.62	28.66	30.37	32.13	
c	19.62	19.66	25.97	20.36	16.34	16.26	16.28	13.76	
alk	10.23	14.50	7.90	12.62	14.29	20.68	18.42	21.62	
qz	-23.86	-4.94	20.58	-17.40	-18.91	-18.30	+8.74	-3.73	

(1) - Kobberfjord norite; (2) to (5) - Havnefjord diorite; (6) to (9) - pyroxene-mica-diorite.
Analyst: the writer.

TABLE 14

CHEMICAL ANALYSES OF QUARTZ-DIORITES

	(1)	(2)	(3)	(4)	(5)	(6)	(7)
Major elements	38/56C	48/29B	03/15C	38/13A	38/31A	46/4A	48/8B
SiO_2	57.85	61.82	67.22	68.30	62.04	62.23	62.97
TiO_2	1.27	0.89	0.57	0.49	0.79	0.94	0.73
Al_2O_3	15.87	17.71	16.38	15.15	16.79	15.99	16.65
Fe_2O_3	2.01	0.58	0.44	0.35	1.07	0.79	0.36
FeO	5.44	2.95	2.93	2.23	3.02	4.07	2.64
MnO	0.06	0.03	0.04	0.02	0.04	0.34	0.01
MgO	2.41	1.68	1.16	1.05	1.67	1.64	1.37
CaO	2.64	2.17	1.61	1.88	1.35	1.98	2.64
Na_2O	3.41	3.51	3.02	3.37	3.91	3.25	3.78
K_2O	4.64	4.88	5.77	5.27	6.39	5.65	3.36
P_2O_5	-	tr.	0.01	tr.	tr.	0.02	-
H_2O^+	1.61	1.64	0.85	0.71	0.96	0.18	0.71
H_2O^-	0.21	0.19	0.17	0.04	0.12	0.09	0.06
Total	97.42	98.05	100.17	98.87	98.15	97.17	94.58
Trace elements							
S	303	277	80	117	151	525	242
V	312	204	197	145	178	284	268
Cr	-	tr.	-	tr.	tr.	tr.	-
Co	-	14	-	-	-	2	-
Ni	-	17	-	6	11	13	-
Rb	213	342	284	209	356	253	221
Sr	467	479	336	361	387	343	468
Ce	-	137	-	72	107	154	-
CIPW norms							
Q	8.12	13.97	20.53	21.55	7.89	12.38	18.89
Or	27.42	28.84	34.10	31.14	37.76	33.39	19.86
Ab	28.85	29.70	25.55	28.52	33.09	27.50	31.99
An	13.10	10.77	7.92	9.33	6.70	9.69	13.10
Cor	0.44	2.71	2.26	0.48	0.99	0.98	2.00
Hyp	12.35	7.71	7.04	5.65	7.59	9.99	6.78
Mt	2.91	0.84	0.64	0.51	1.55	1.15	0.52
Il	2.41	1.69	1.08	0.93	1.50	1.79	1.39
Ap	-	-	0.02	-	-	0.05	-
Total	95.60	96.22	99.15	98.11	97.07	96.90	94.51
Niggli Values							
si	205.49	250.17	298.34	320.93	247.54	248.62	274.04
al	33.22	42.24	42.84	41.95	39.48	37.65	42.70
fm	34.47	21.99	20.17	17.43	23.36	26.89	19.71
c	10.05	9.41	7.66	9.47	5.77	8.48	12.31
alk	22.26	26.37	29.33	31.15	31.39	26.99	25.28
qz	+16.45	+44.69	+81.02	+96.33	+21.98	+40.66	+72.92

(1) to (3) - quartz-diorite; (4) to (7) - garnetiferous quartz-diorite

Analyst: the writer.

TABLE 15

CHEMICAL ANALYSES OF EARLY DIORITES AND KOMAGFJORD DIORITE

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
Major elements	36/4B	36/44A	36/54D	36/300A	36/119C	36/207A	36/225C	11/47F
SiO_2	56.27	56.17	64.50	59.54	51.22	49.77	58.69	50.46
TiO_2	1.62	1.35	0.62	1.01	1.91	2.95	1.00	2.62
Al_2O_3	18.25	17.25	14.99	18.38	19.48	20.11	17.41	18.10
Fe_2O_3	2.01	1.38	1.19	1.49	3.20	1.45	1.08	2.74
FeO	5.89	5.66	2.74	4.45	5.68	10.34	4.74	9.51
MnO	0.06	0.06	0.03	0.05	0.05	0.11	0.39	0.13
MgO	2.52	2.27	1.11	1.52	4.07	3.35	2.06	2.46
CaO	4.43	4.02	1.80	2.96	4.57	6.21	3.27	5.88
Na_2O	4.42	4.05	3.06	3.93	3.64	3.95	3.98	5.26
K_2O	3.70	3.07	6.30	5.48	3.11	1.04	4.07	2.09
P_2O_5	-	0.07	-	0.04	0.07	0.10	0.03	0.15
H_2O^+	1.06	0.81	0.34	0.25	1.74	0.31	0.75	0.69
H_2O^-	0.14	0.18	0.12	0.11	0.26	0.13	0.08	0.08
Total	100.37	96.34	96.80	99.21	98.98	100.02	97.55	100.17
Trace Elements								
S	2322	1373	1032	929	832	596	235	1020
V	236	307	221	196	289	-	166	-
Cr	-	-	-	tr.	-	tr.	tr.	-
Co	-	-	-	12	-	-	-	-
Ni	-	-	-	17	-	35	20	-
Rb	60	67	137	99	90	11	101	73
Sr	556	445	248	332	450	421	412	435
Ce	-	-	-	322	-	121	240	-
CIPW norms								
Q	0.78	5.81	15.75	2.62	0.33	-	6.83	-
Or	21.87	18.14	37.23	32.39	18.38	6.15	24.05	12.35
Ab	37.40	34.27	25.89	33.26	30.80	33.42	33.68	36.33
An	19.03	19.49	8.56	14.42	22.21	30.16	16.03	19.61
Ne	-	-	-	-	-	-	-	4.43
Cor	-	0.12	-	0.70	1.98	1.44	0.58	-
Di	2.44	-	0.31	-	-	-	-	7.25
Hyp	11.66	12.79	5.69	12.74	14.86	16.84	12.02	-
Ol	-	-	-	-	-	3.44	-	10.14
Mt	2.91	2.00	1.73	2.16	4.64	2.10	1.57	3.97
Il	3.08	2.56	1.18	1.92	3.63	5.60	1.90	4.98
Ap	-	0.16	-	0.09	0.16	0.23	0.07	0.35
Total	99.17	95.35	96.34	100.29	97.00	99.38	96.73	99.40
Niggli values								
si	173.71	189.99	285.21	194.54	145.69	131.59	206.68	135.67
al	33.20	34.39	39.06	35.39	32.66	31.33	36.13	28.68
fm	31.63	31.14	21.52	30.38	37.74	39.19	28.80	37.08
c	14.65	14.57	8.53	10.36	13.93	17.59	12.34	16.94
alk	20.51	19.90	30.89	23.87	15.68	11.88	22.73	17.30
qz	-8.33	+10.39	+61.65	-0.94	-17.03	-15.93	+15.76	-33.53

(1) to (4) - early diorite; (5) to (7) - thermally metamorphosed early diorite;
(8) Komagfjord diorite.
Analyst: the writer.

TABLE 16

CHEMICAL ANALYSES OF VATNA GABBRO, SLÅTTEN GABBROS, TROCTOLITE, AND PERTHOSITES

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
Major Elements	11/29I	11/38K	07/25D	07/25G	48/10A	36/82B	11/1C	11/25A
SiO_2	47.58	54.55	46.78	39.34	46.45	42.28	62.90	62.50
TiO_2	1.78	1.91	2.93	4.81	2.44	1.51	0.26	0.44
Al_2O_3	22.17	17.57	16.66	14.11	19.78	11.00	18.49	18.87
FeO_2	1.56	1.80	2.62	1.45	0.69	2.19	0.56	0.47
FeO	6.18	8.30	8.54	12.39	8.78	10.92	1.30	1.51
MnO	0.08	0.12	0.12	0.12	0.09	0.13	0.02	0.03
MgO	4.08	1.67	6.42	7.24	6.02	17.07	0.85	0.07
CaO	11.52	5.18	11.30	11.16	9.33	7.69	1.40	1.42
Na_2O	4.20	6.01	3.46	2.74	3.00	1.91	6.57	7.45
K_2O	0.28	2.01	0.64	0.33	0.83	0.71	5.34	5.24
P_2O_5	0.18	0.09	0.16	0.42	0.13	0.11	-	tr.
H_2O^+	0.55	0.58	0.61	0.70	0.49	0.79	0.98	0.15
H_2O^-	0.57	-	0.12	0.16	0.16	0.07	0.21	0.16
Total	100.24	99.79	100.36	94.97	98.19	96.38	98.88	98.31
Trace elements								
S	2079	951	3131	1258	2068	1571	3674	142
V	307	32	-	2193	267	377	34	123
Cr	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
Co	38	27	11	33	16	47	-	-
Ni	45	18	105	15	68	897	15	2
Rb	-	25	14	5	30	19	106	89
Sr	703	547	497	506	448	266	149	432
Ce	57	79	52	70	54	59	25	14
CIPW norms								
Or	1.65	11.88	3.78	1.95	4.91	4.20	31.56	30.97
Ab	24.28	47.43	21.73	9.70	25.05	11.52	55.59	54.63
Ar	40.82	15.03	28.04	25.23	38.06	19.35	5.19	2.57
Ne	6.10	1.86	4.09	7.30	0.18	2.51	-	4.55
Di	12.41	8.61	21.91	22.47	6.06	14.54	1.44	3.60
Wo	-	-	-	-	-	-	-	0.16
Hyp	-	-	-	-	-	-	1.68	-
Ol	8.29	7.96	10.35	15.25	17.36	37.11	0.93	-
Mt	2.26	2.61	3.80	2.10	1.00	3.18	0.81	0.68
Il	3.38	3.63	5.56	9.14	4.63	2.87	0.49	0.84
Ap	0.42	0.21	0.37	0.97	0.30	0.25	-	-
Total	99.41	99.21	99.63	94.11	97.54	95.52	97.69	98.00
Niggli values								
Si	112.88	160.91	105.19	86.49	110.59	79.24	251.95	250.46
al	31.00	30.54	22.08	18.28	27.75	12.15	43.65	44.57
fm	29.63	32.11	42.24	49.13	40.26	68.09	11.18	7.00
c	29.28	16.37	27.23	26.29	23.80	15.44	6.01	6.10
alk	10.08	20.97	8.46	6.30	8.13	4.32	39.16	42.34
qz	-27.44	-22.97	-28.65	-38.71	-22.13	-38.04	- 4.69	-18.90

(1) and (2) - Vatna gabbro; (3) northern Slåtten gabbro; (4) southern Slåtten gabbro; (5) Contaminated gabbro; (6) - troctolite; (7) and (8) - perthosite.
Analyst: the writer.

Partial Chemical Analyses of Coexisting Feldspars in Diorites

Plagioclase	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)
	36/44A	36/54D	36/119C	36/117C	36/233C	36/237A	38/18B	38/38A	38/56C	38/13A	38/31A	46/4A	48/8B
CaO (Wt.%)	7.72	7.01	7.10	9.04	9.12	8.19	7.40	8.26	4.50	2.65	4.14	3.44	3.85
Na_2O	6.57	6.47	6.15	5.84	6.15	6.50	6.41	6.47	5.17	3.57	5.06	4.81	5.45
K_2O	0.71	0.94	0.59	0.62	0.50	0.57	0.78	0.50	0.88	1.15	0.58	1.22	0.95
Ba (p.p.m.)	165	155	145	530	200	140	530	95	590	510	100	440	570
Rb	4	7	51	14	-	2	11	2	24	57	14	22	32
Sr	430	240	590	650	580	540	500	580	420	210	280	280	460

K-Feldspar	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)
	36/44A	36/54D	36/119C	36/117C	36/233C	36/237A	38/18B	38/38A	38/56C	38/13A	38/31A	46/4A	48/8B
CaO	0.80	0.35	2.27	3.29	1.22	2.18	0.96	2.25	1.57	0.28	0.40	0.22	1.48
Na_2O	1.62	1.72	3.80	2.55	1.87	2.60	1.62	2.60	2.50	1.92	1.92	1.67	2.80
K_2O	13.40	13.75	4.60	8.55	13.48	10.20	11.52	9.68	8.80	13.10	12.00	14.60	8.37
Ba	4000	2200	900	2300	4300	7250	4200	5550	2100	5100	3700	4550	9100
Rb	120	190	35	60	140	120	140	130	180	325	210	250	170
Sr	490	265	290	500	575	540	500	535	430	425	395	405	635

(1) and (2) - early diorite; (3) - thermally-metamorphosed early diorite; (4) and (5) - Havnefjord diorite; (6) to (8) - pyroxene-mica-diorite; (9) - quartz-diorite; (10) to (13) - garnetiferous quartz-diorite.

Analyst: the writer.

TABLE 18

PARTIAL CHEMICAL ANALYSES OF FELDSPARS IN MISCELLANEOUS ROCKS

Plagioclase	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)
	36/3A	36/54A	36/82B	11/29I	11/29J	11/13D	48/10A	36/116A	36/237B
CaO(Wt.%)	8.62	11.55	11.60	8.50	10.99	7.57	12.50	9.58	4.62
Na_2O	6.30	4.90	4.61	6.68	5.68	6.50	4.66	5.68	11.13
K_2O	0.41	0.31	0.20	0.54	0.26	0.37	0.27	0.41	5.17
Ba(p.p.m.)	265	105	305	370	225	390	95	145	25
Rb	6	-	-	-	-	9	2	3	-
Sr	1040	920	1540	1100	1020	930	720	850	1050

K-Feldspar	(10)
	11/25A
CaO	1.10
Na_2O	5.52
K_2O	6.40
Ba	1050
Rb	90
Sr	350

(1) and (2) - Husfjord metagabbro; (3) - troctolite; (4) - Vatna gabbro; (5) - olivine leucogabbro; (6) - basic dyke; (7) - gabbro; (8) - basic Havnefjord diorite; (9) - hybrid; (10) - perthosite.

Analyst: the writer.

THE DIORITES

Introduction

Study of the chemical data suggests that the suite of diorites consisting of the Havnefjord diorite, the late pyroxene-mica-diorites, and the two types of quartz-diorites form a petrological series. The chemical features of the Kobberfjord norite indicate that it may represent a basic member of this series.

EARLY DIORITES

Before discussing the main diorite series in detail, it is as well to consider the position of the early diorites as these do not appear to be members of the main diorite series. They plot in positions near to the late pyroxene-mica-diorites and the quartz-diorites, which are late members of the series. Since field evidence shows that the early diorites pre-date the Havnefjord diorite, this would involve a reversal in the trend of the series. This relationship also holds with respect to the trace elements in the feldspars, and it would seem that the formation of the early diorites is not associated with the principal suite of diorites.

MAIN DIORITE SUITE

Whole Rock Analyses

Various aspects of the chemistry of the main diorite suite are plotted on triangular diagrams, Figs. 16 - 19. The Kobberfjord norite

PLOT OF NORITE-DIORITE SERIES IN THE SYSTEM SiO_2-($K_2O + Na_2O$)-CaO

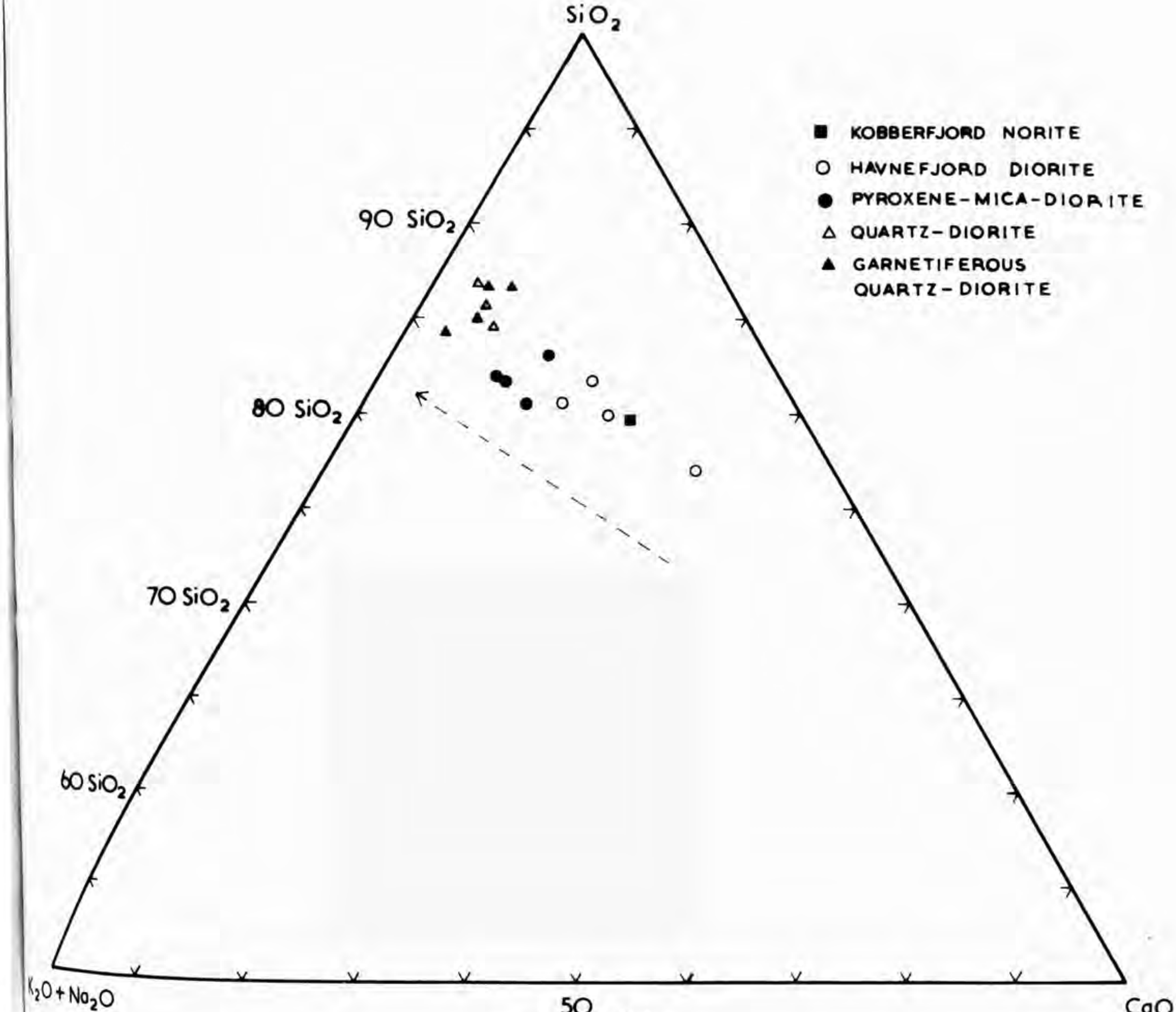


FIG. 16

PLOT OF NORITE-DIORITE SERIES
IN THE SYSTEM $K_2O - Na_2O - CaO$

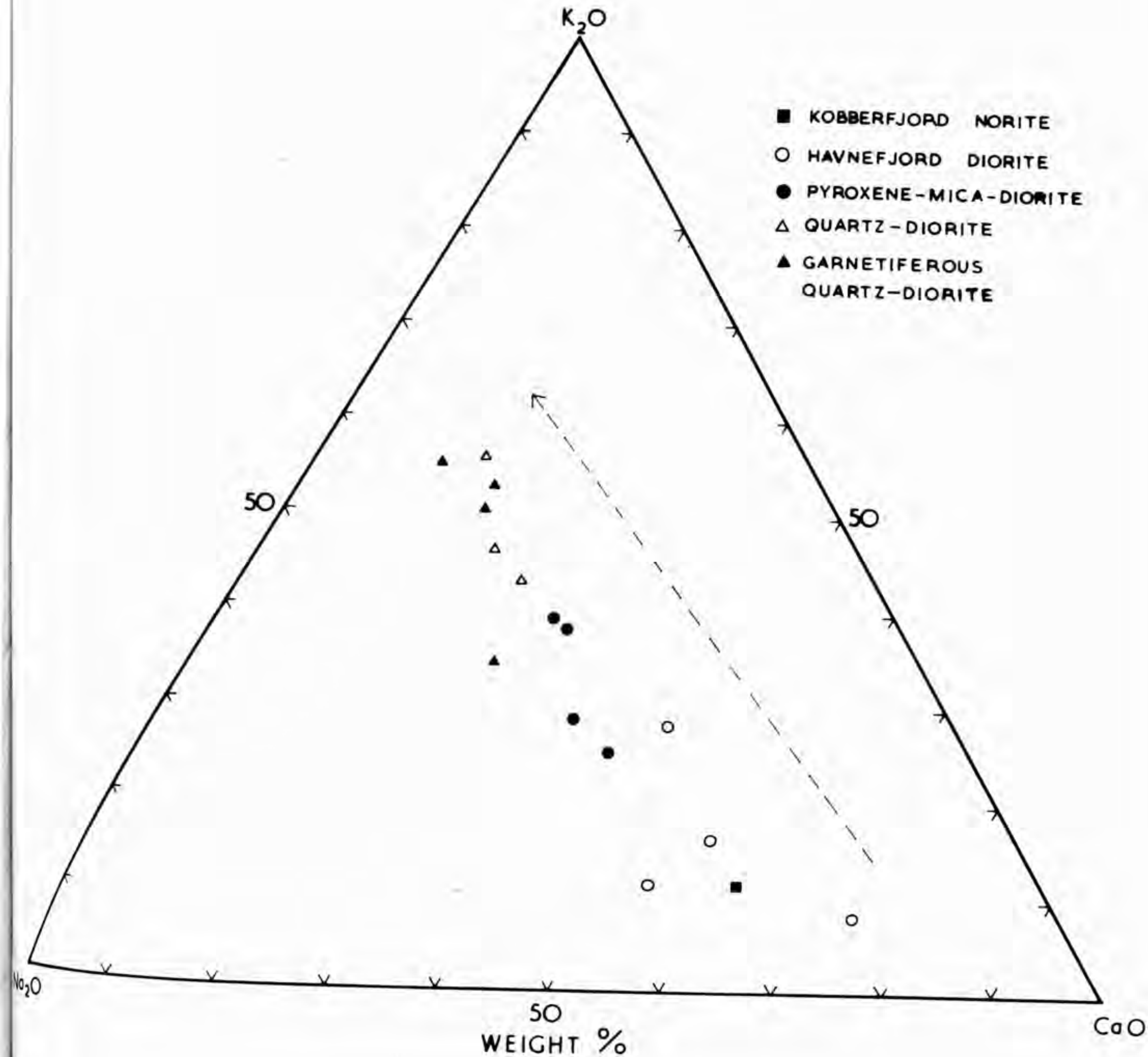
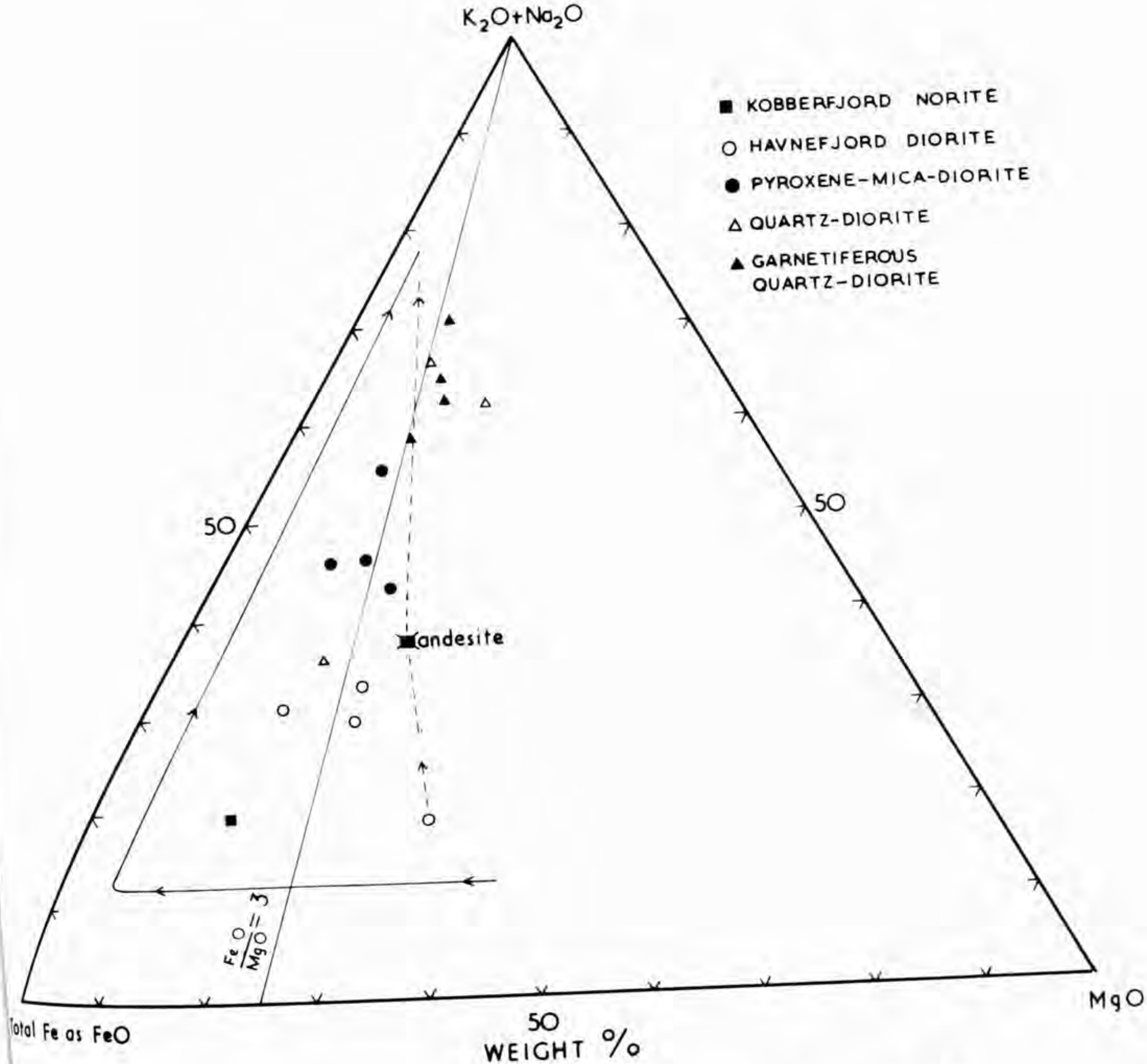


FIG. 17

PLOT OF NORITE-DIORITE SERIES IN THE SYSTEM (K_2O+Na_2O)—Total Fe—MgO



Solid trend line — trend of differentiation due to fractional crystallization in the Skaergaard Complex. Dashed trend line — trend of lavas belonging to the calc-alkali suite. Both after Hatch, Wells, and Wells (1961).

For further discussion see text under Petrogenesis.

FIG. 18

PLOT OF NORITE-DIORITE SERIES
IN THE SYSTEM $Al_2O_3 - CaO - (MgO + Total\ Fe)$

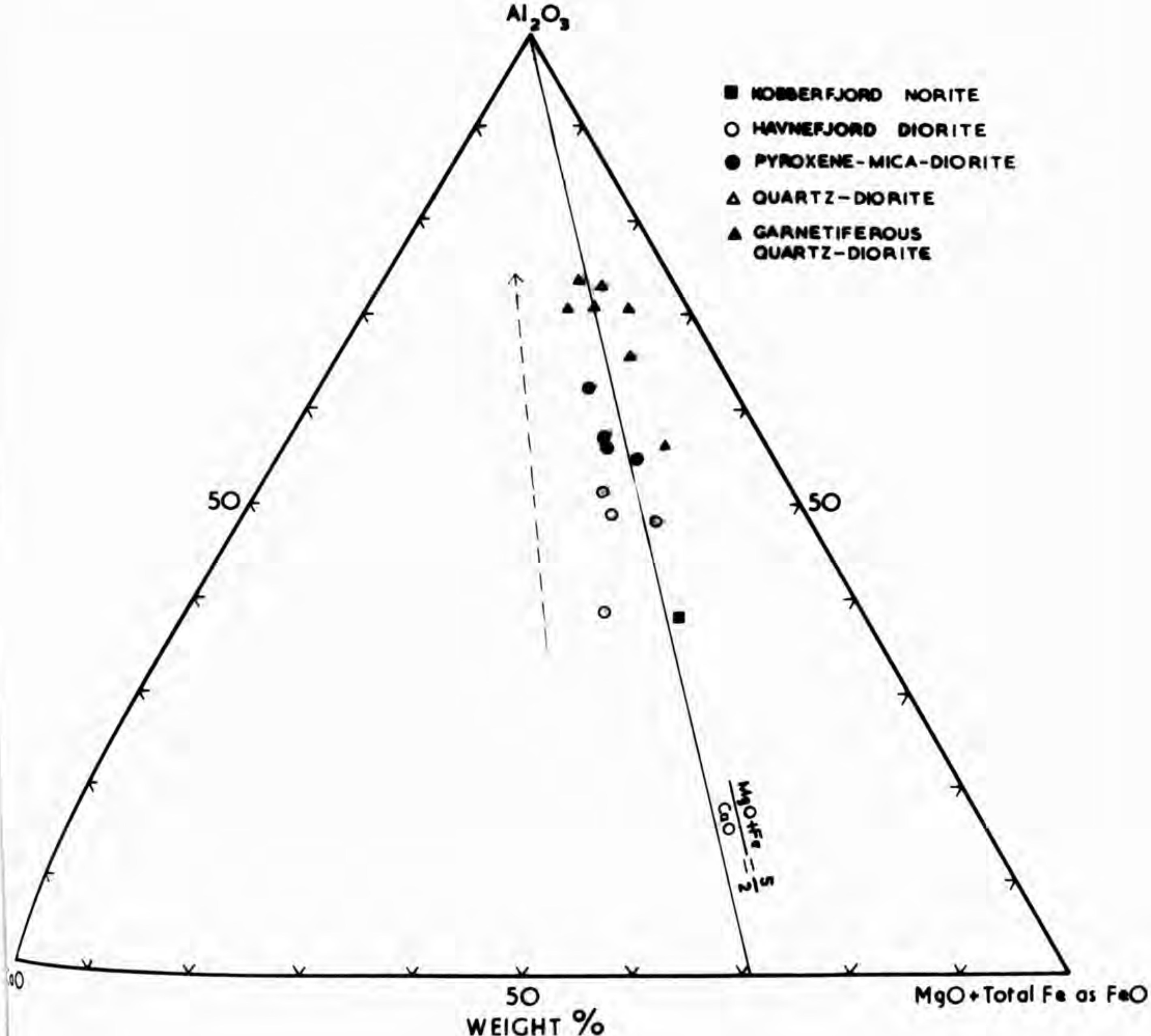


FIG. 19

has also been plotted on these diagrams, and it would appear to form a basic end-member of a series formed by the diorites. It can be seen from these diagrams that this suite of rocks forms a distinct trend with the Kobberfjord norite (together with one exceptionally basic specimen of Havnefjord diorite) at one end, and the quartz-diorites at the other. It is known from field evidence that the pyroxene-mica-diorite post-dates the Havnefjord diorite, and this fact gives the trend a temporal direction. Therefore, it seems reasonable to assume that if these rocks form a series, the Kobberfjord norite represents the earliest member present and the quartz-diorites the latest members.

The diagrams show that there is a trend from fairly basic to relatively acid members. Magnesium, total iron, and calcium decrease in absolute amounts, but they remain in constant proportion with respect to each other. The FeO/MgO ratio remains at about 3(Fig.18), and the (MgO + FeO)/CaO ratio is about $2\frac{1}{2}$(Fig. 19). There is a corresponding increase in the absolute amounts of silicon, aluminium, and alkalis, and in the case of the alkalis the K_2O/Na_2O ratio increases with time.

These trends are similar to those which are characteristic of a suite of rocks which has been formed from differentiation, by fractional crystallization, of a common source magma. Thus it is possible that one explanation of the trends exhibited by these rocks could be that they have all crystallized from a common parent magma.

The variation diagrams (after Larsen, 1938) presented in Figs. 20 and 21 show that the members of this series appear to lie along fairly smooth lines of liquid descent. Features revealed by these diagrams

VARIATION DIAGRAM (AFTER LARSEN) OF NORITE-DIORITE SERIES

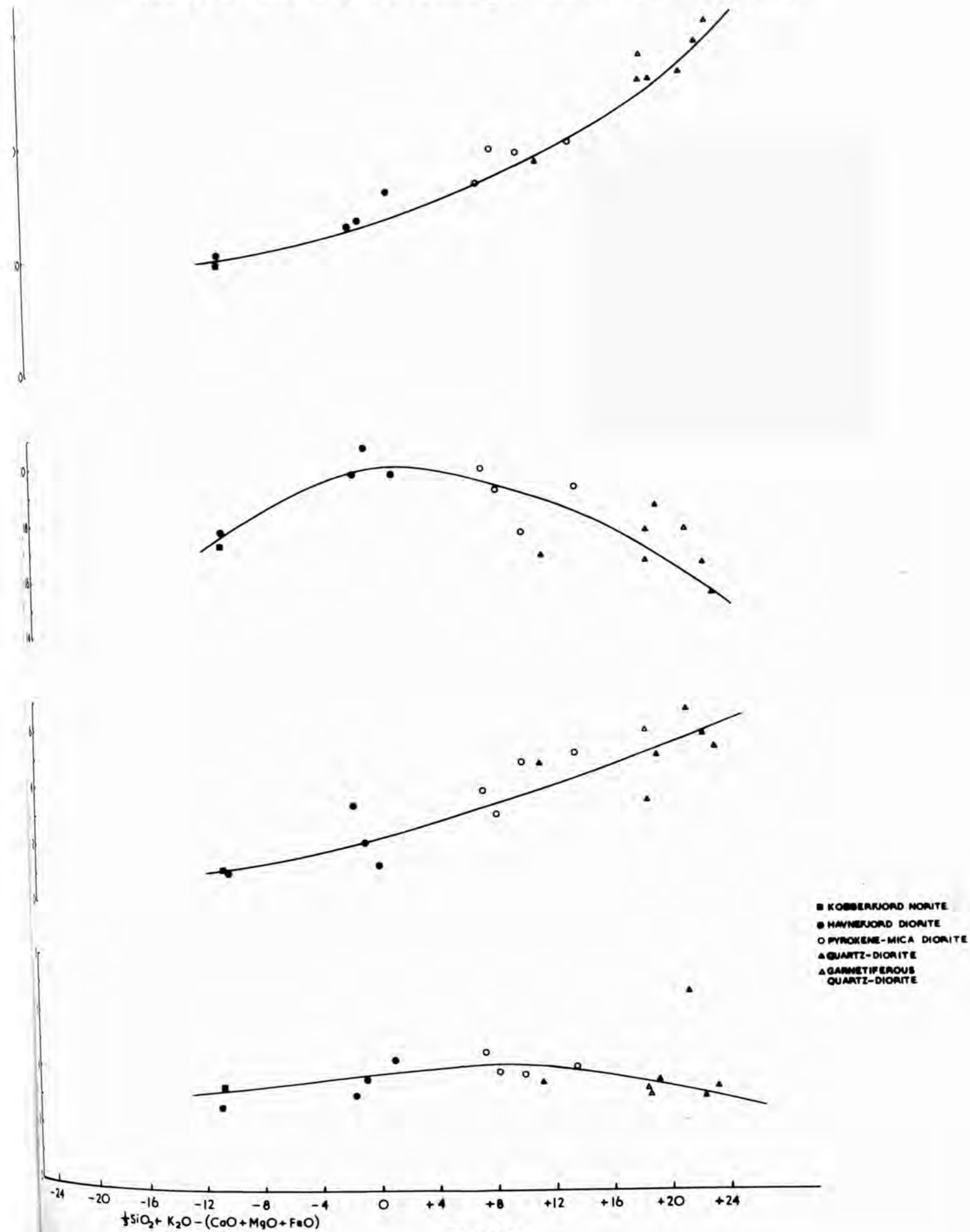


FIG. 20

VARIATION DIAGRAM (AFTER LARSEN) OF NORITE-DIORITE SERIES

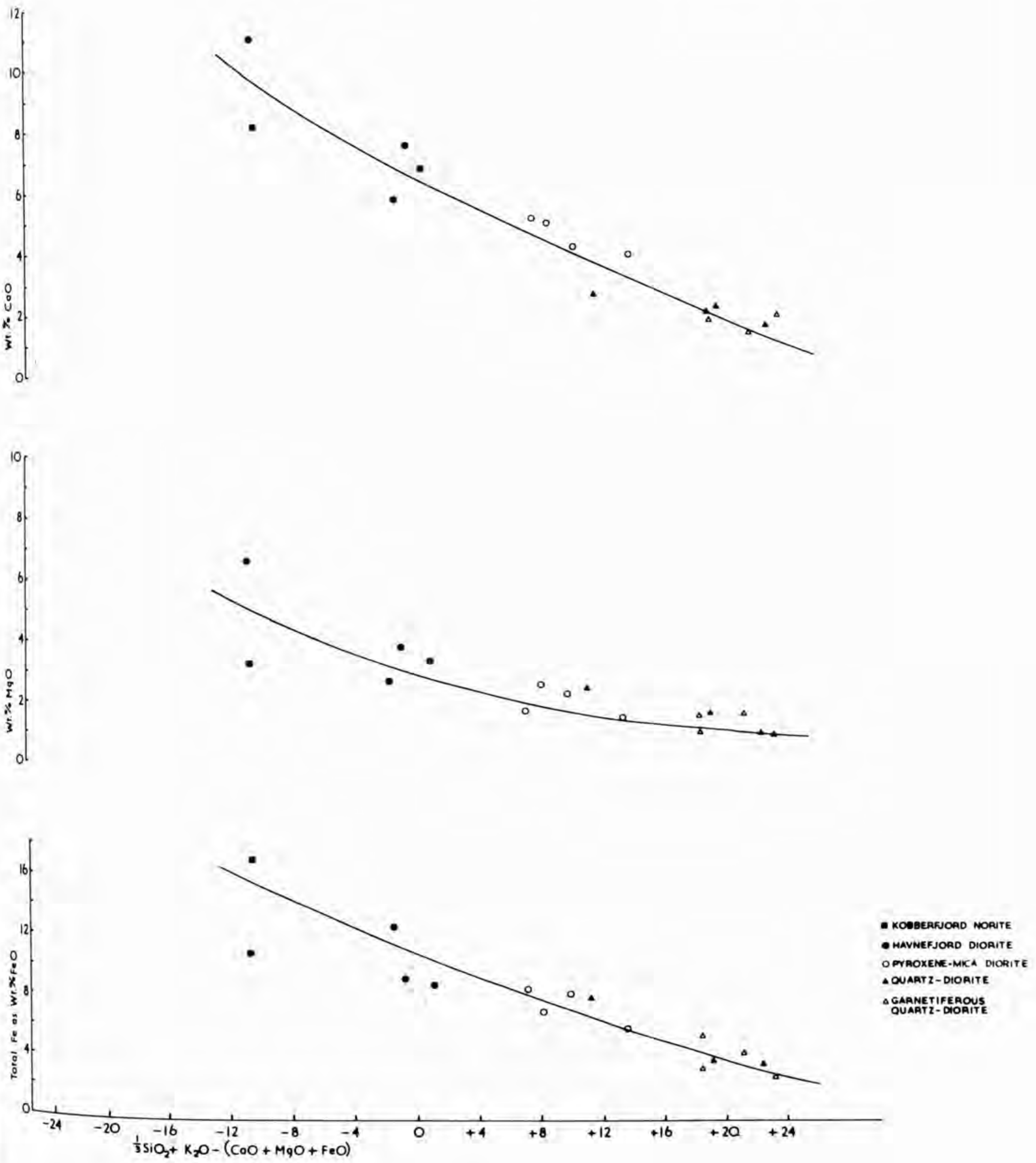


FIG. 21

agree with those of the triangular diagrams, and in addition the reason for the increase in the K_2O/Na_2O ratio is shown to be a slight decrease in amount of Na_2O in the quartz-diorites (Fig. 20). Another feature shown in Fig. 20 is that aluminium has a peak in the Havnefjord diorite and pyroxene-mica-diorites.

Variation diagrams after Harker (1909) show similar trends to the Larsen diagrams (Fig. 22). The aluminium and sodium peaks in the Havnefjord diorite and the pyroxene-mica-diorites are clearly seen. In addition, because K_2O is not involved in the abscissa parameter, a more marked curve in the K_2O line results, showing that there is a fairly rapid increase in K_2O until the formation of the pyroxene-mica-diorites, after which the rate of increase falls off.

The alkali-lime index as determined from the Harker diagram is 51.2, which places the suite in the alkali-calcic class of Peacock (1931). This value for the alkali-lime index makes the suite equivalent to a trachybasalt-trachyandesite-trachyte series (Peacock, op.cit.). Although the analyses show that the Havnefjord diorite is rich in calcium, the rapid fall in the CaO curve in conjunction with the fairly rapid increase in the alkalis already noted gives rise to a comparatively low alkali-lime index.

Niggli values have been computed from the analyses, and have been plotted on a tetrahedral diagram (Fig. 23) by a method which is a variation of that used by Niggli (Johannsen, 1931). In Niggli's method, the four values al, alk, c, and fm, are placed at the corners of a tetrahedron, which is divided into 10% divisions by planes rotating

VARIATION DIAGRAM (AFTER HARKER) OF NORITE-DIORITE SERIES

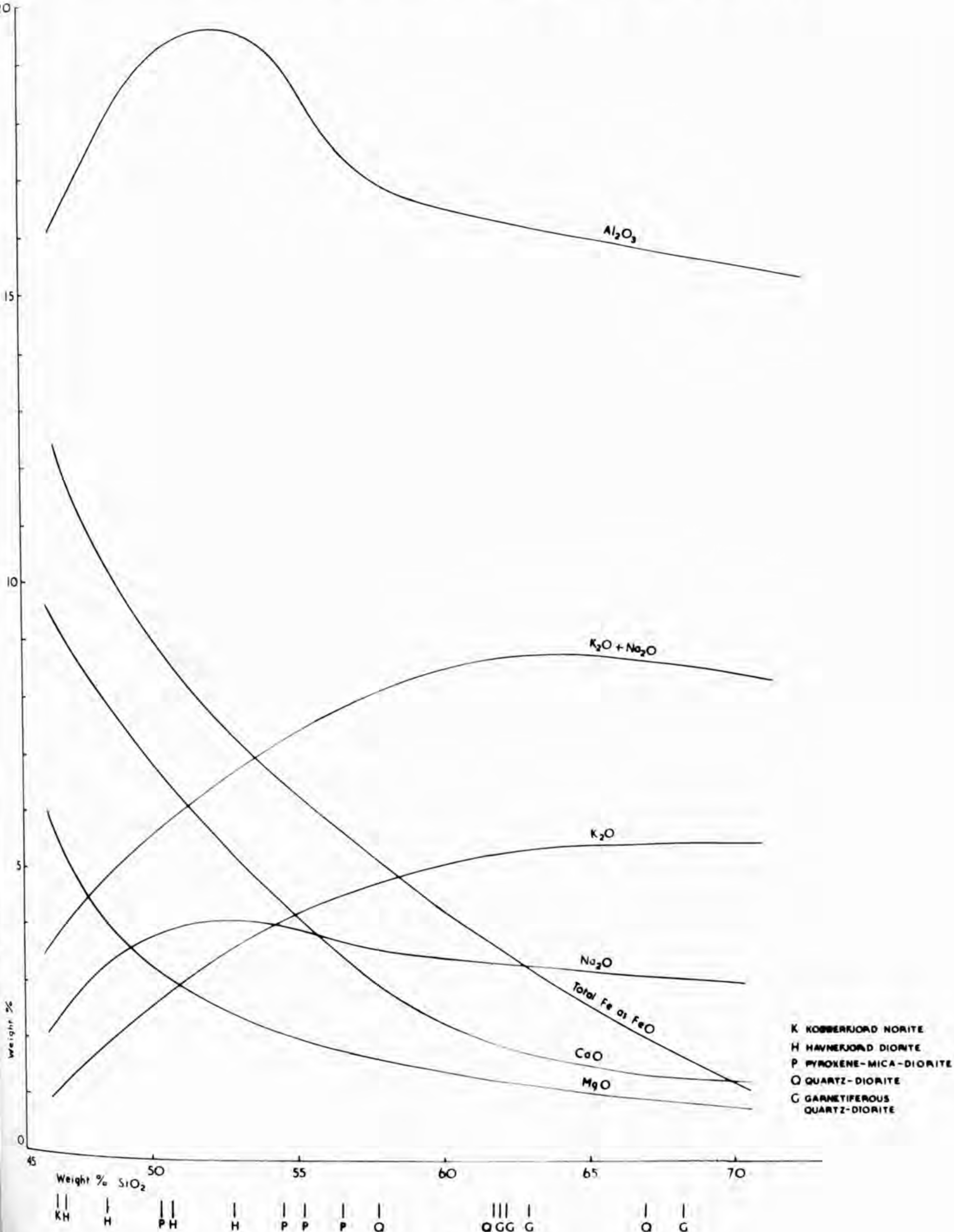


FIG. 22

PLOT OF NIGGLI VALUES FOR NORITE-DIORITE SERIES

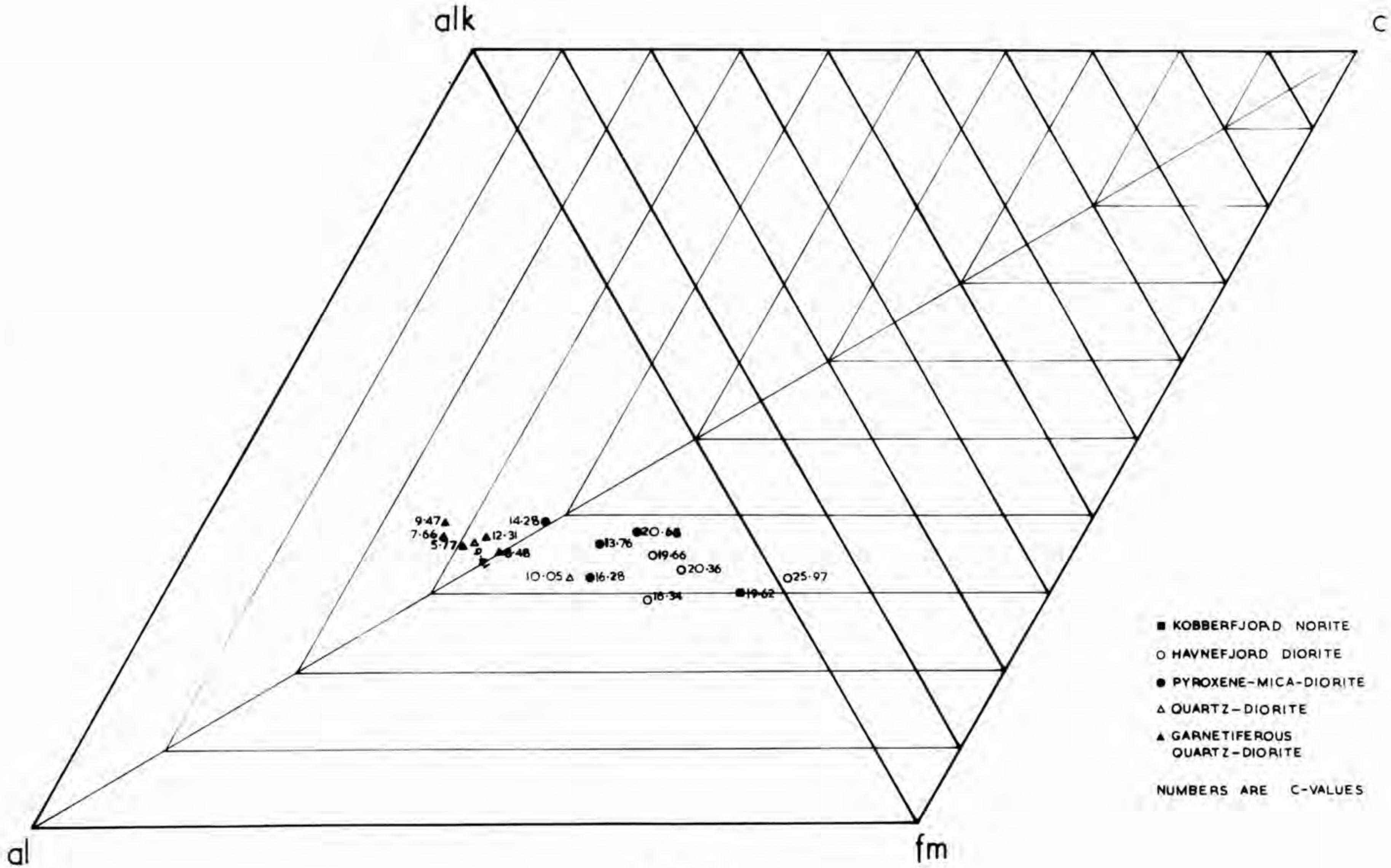


FIG. 23

about the al-alk edge as axis. These triangular planes are taken in pairs and represented as a series of rhombic diagrams with the al-alk line shared by the adjacent triangles. Thus the values are not presented in a single diagram, but in a series of diagrams of varying c/fm ratios, which have to be correlated if an overall picture of any trends is to be obtained.

In the method used by the writer, the al-alk-c-fm tetrahedron is divided into planar divisions parallel to the side opposite the apex at which the Niggli value having the lowest average is represented (which is c in the present case). A projection of such a model has been constructed in Fig. 23, and may either be regarded as a two-dimensional diagram of overlapping triangles having varying c values, or as a perspective diagram of a tetrahedron composed of a 'tunnel' of triangles converging towards the c apex. Thus while the triangles could alternatively have been represented separately, by assembling them into a tetrahedron, a clearer overall three-dimensional effect of trends is possible.

To plot a particular rock, the c value is first subtracted from the total al + alk + c + fm, and the remaining al, alk, and fm values are recalculated to 100. The triangle at the appropriate distance from c is selected, and the al, alk, and fm values are plotted within this triangle in the normal way for triangular diagrams. The c value is written beside the plot to indicate in which triangle it lies.

It can be seen from Fig. 23 that, having regard to the c values, there is a trend away from fm and c and towards alk.

Qz values have been calculated from the Niggli values, and these show a general progression from megative values to positive values with time. This indicates that silica was relatively deficient in early members, and explains why quartz was able to form in the later members.

Another expression of the silica saturation of the diorites is given by the CIPW norms (see Tables 13 and 14). These are presented graphically on a double triangular diagram (after Larsen, 1938) in Fig. 24. It can be seen that the Kobberfjord norite and the Havnefjord diorite lie principally within the undersaturated field of silica deficiency, while the later diorites (with the exception of one pyroxene-mica-diorite) lie within the saturated field. The tie-lines occasionally cross, but they maintain a generally sub-parallel orientation, and a line approximately perpendicular to the tie-lines gives the trend of the series. There is a clear trend away from the basic, undersaturated regions towards the alkaline, saturated areas. The fact that early members of the series are undersaturated is probably connected with the relatively low alkali-lime index of the suite.

Neither the Vatna gabbro nor the perthosites appear to be members of the diorite suite, since they fall off the general trends. An alternative affinity for the perthosites may be found in the syenitic rocks of the carbonatite association in the Breivikbotn alkaline complex of Sørøy (Sturt and Ramsay, 1965). Fig. 25 is taken from Sturt and Ramsay (op. cit., Fig. 69), upon which the perthosites have been plotted. Although the perthosites are a little richer in SiO_2 than the Brievikbotn syenites, the remaining plots fall near to the curves. This suggests

PLOT OF CIPW NORMS (AFTER LARSEN)
FOR NORITE – DIORITE SERIES

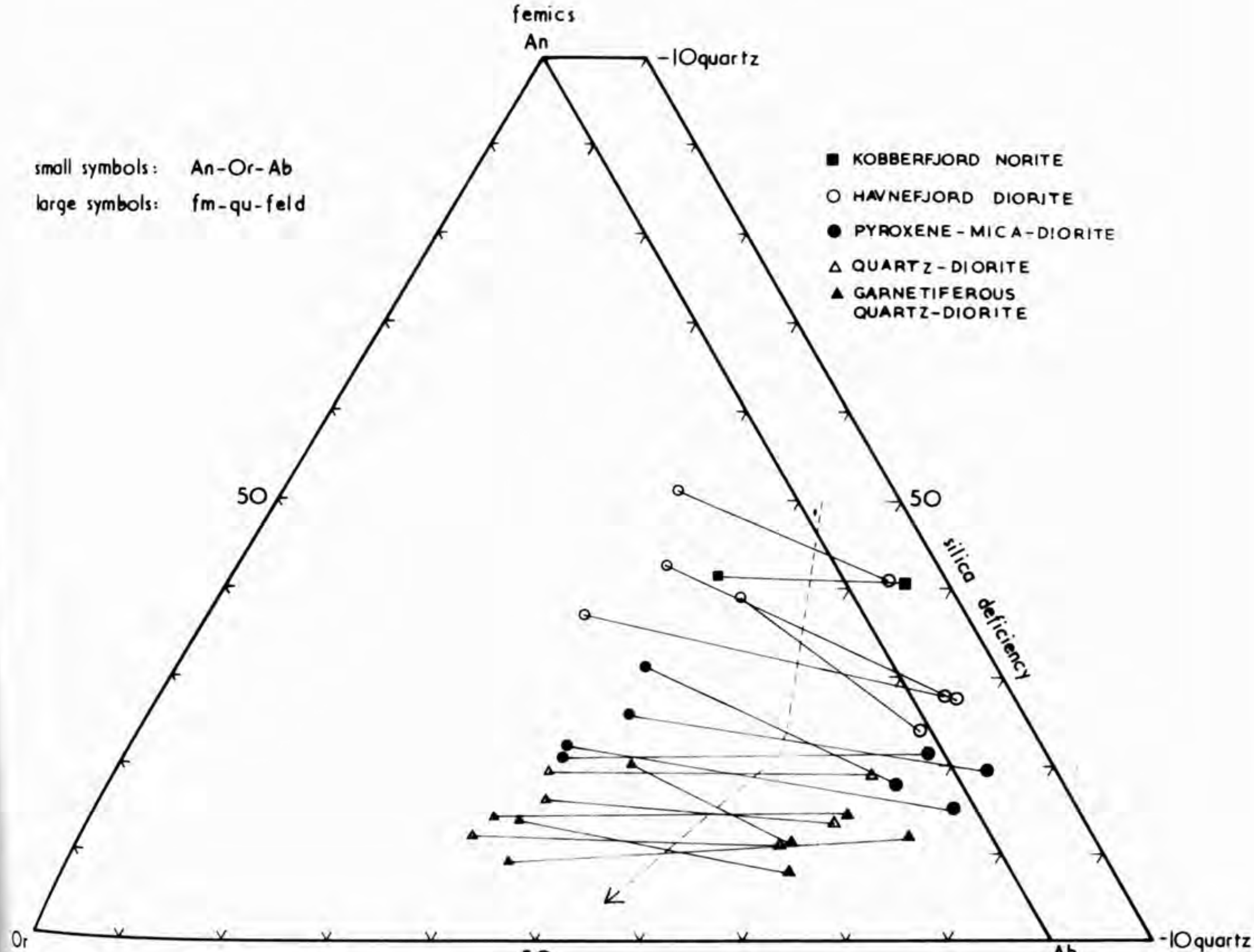
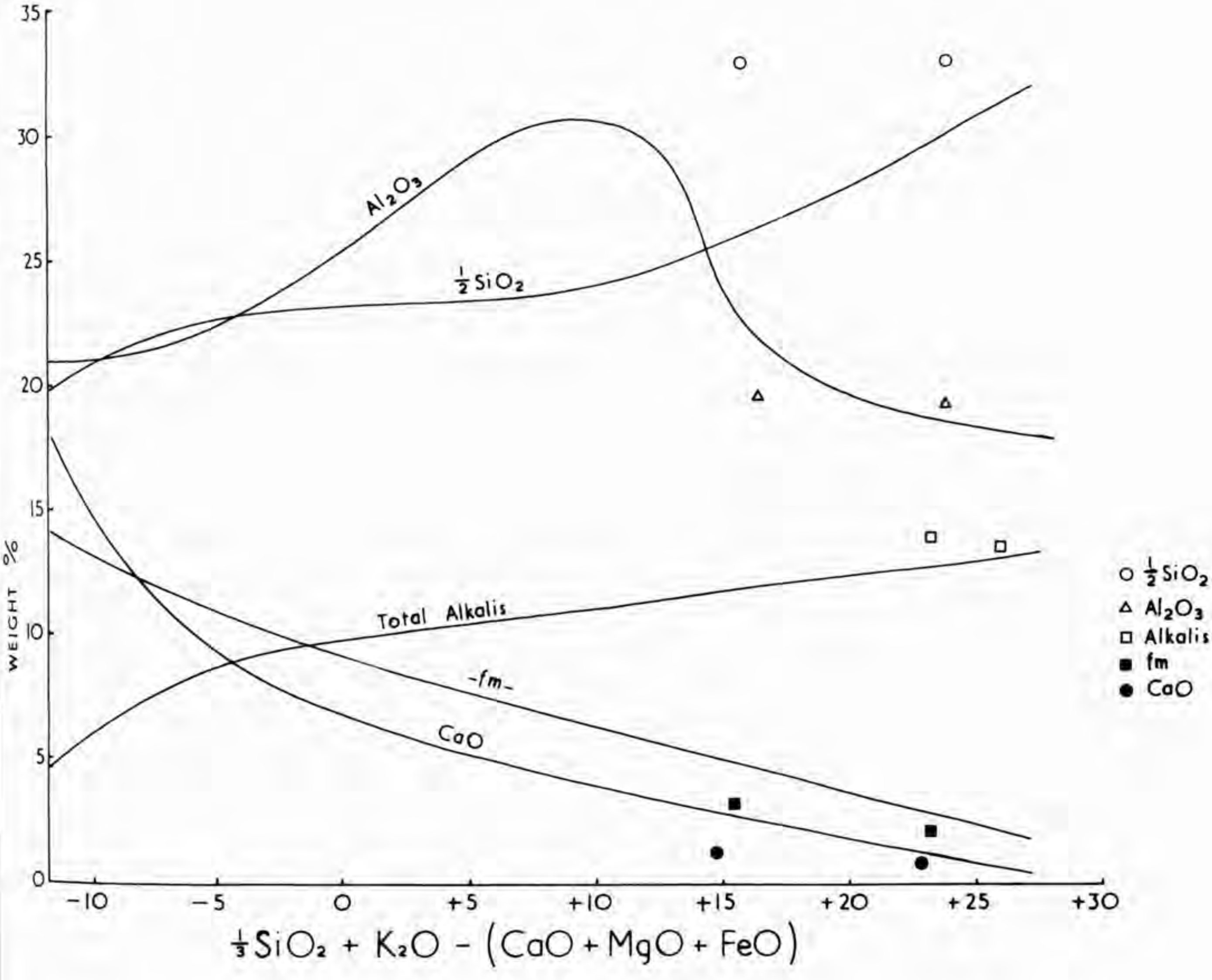


FIG. 24



Vatna perthosites plotted onto the variation diagram (after Larsen) of syenitic rocks of the carbonatite association of the Breivikbotn alkaline complex (Sturt and Ramsay 1965, Fig. 69.)

(fm = total iron as Fe_2O_3 + MgO)

FIG. 25

that the perthosites may have affinities with an alkaline suite of which the nepheline-syenite pegmatites at Vatna might be members. Unfortunately, analyses of these latter rocks are not available.

Feldspar Analyses

Major Elements

The trend away from calcium towards the alkalis that was noted in the data from the whole rock analyses is reflected in the feldspars in the diorites. The Or-Ab-An values of coexisting K-feldspar and plagioclase in the diorites have been plotted on a triangular diagram (Fig. 26). The tie-lines are subparallel, and the trend indicates that the host rock composition became more alkaline with time. This fact is further supported by Fig. 27 which shows that there is a decrease in normative An reciprocal with an increase in Ab and Or in the host rocks.

The partition coefficient of Na_2O between coexisting K-feldspar and plagioclase is illustrated in Fig. 28. The $Na_2O_{plag}/Na_2O_{K\text{-}feld}$ ratio is always greater than unity, but changes from one rock-type to another. These changes are not reciprocated between K-feldspar and plagioclase; only the plagioclase Na_2O concentrations alter significantly, those of the K-feldspar remaining roughly constant. Neither are the changes progressive, but make a reversal with the pyroxene-mica-diorites having the highest Na_2O content in plagioclase.

An explanation of this feature may be found in the variation diagrams, Figs. 20 and 22. In these, the Na_2O curves have a peak in the pyroxene-

PLOT OF COEXISTING FELDSPARS IN DIORITE SERIES IN SYSTEM Or-Ab-An

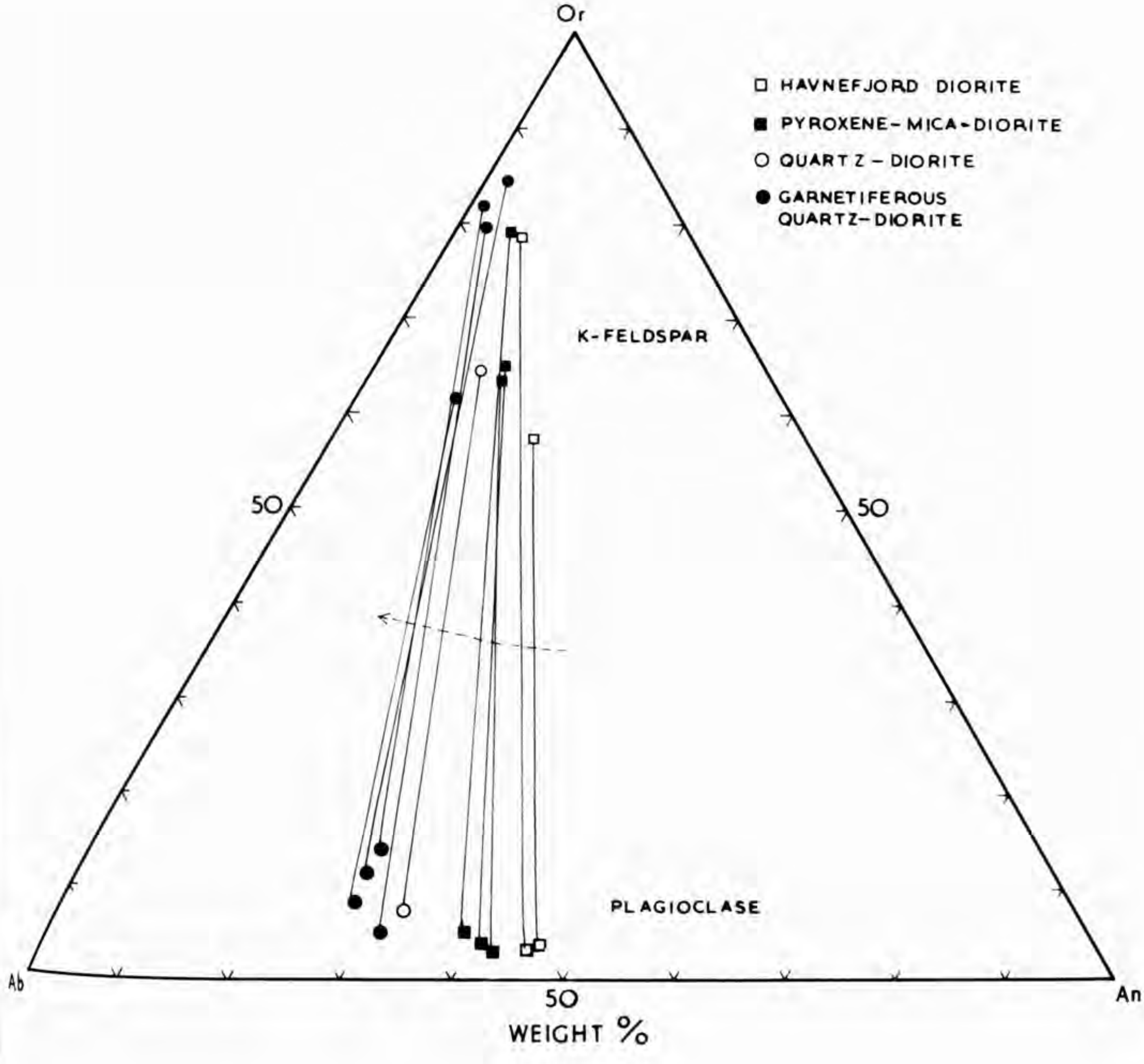
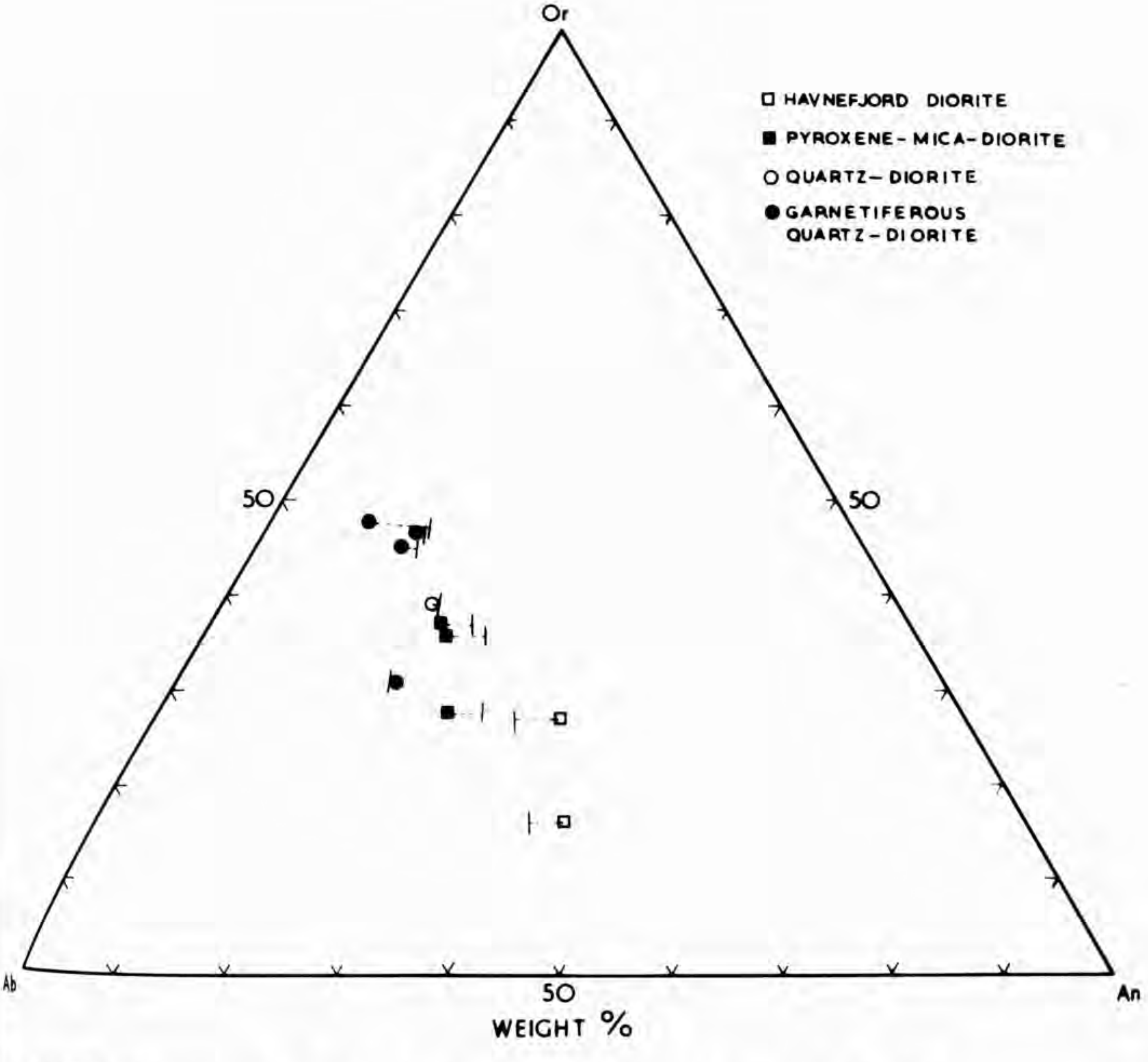


FIG. 26

PLOT OF NORMATIVE Or Ab An IN COEXISTING FELDSPARS IN DIORITE SERIES



DOTTED TIE-LINES JOIN NORM PLOTS TO SHORTENED PROJECTIONS OF THE K-FELDSPAR — PLAGIOCLASE TIE-LINES FROM FIG. 26

FIG. 27

-mica-diorites indicating that these rocks are rich in sodium. The only significant sodium-bearing minerals in the diorites are feldspars, so it is reasonable to expect that practically all the sodium has been used in the feldspars, which therefore reflect the sodium peak. The fact that this peak is only shown in the plagioclase possibly indicates that sodium enters the plagioclase lattice more readily than that of the K-feldspar.

The theoretical minimum temperatures of formation of the coexisting feldspars have been determined from the graph in Barth's paper on the two-feldspar thermometer (Barth, 1962), and the results are presented in Table 19. Although only little data are available, a considerable scatter is apparent, and the meaning of the temperatures appears to be ambiguous.

It is improbable that chemical equilibrium was maintained constant throughout the subsequent regional metamorphism that the rocks have undergone, and it is likely that the K/Na and Na/Ca ratios would have altered. Thus any temperatures recorded in such rocks are equivocable. In the present case, it is possible that temperatures around 700°C. may represent a minimum temperature of crystallization of the original feldspars in the diorite, but the lower temperatures cannot do so, and the meaning of these is uncertain.

Thus it would appear that the two-feldspar thermometer, when applied to metamorphosed igneous rocks, is probably only capable of giving a minimum temperature of formation of the rocks, although even this might have been obliterated by subsequent metamorphism.

TABLE 19

TEMPERATURES OBTAINED BY USE OF THE TWO-FELDSPAR THERMOMETER (AFTER BARTH) ON COEXISTING FELDSPARS IN DIORITES

Havnefjord diorite	Pyroxene-mica-diorites	Quartz-diorites
700°C.	670°C.	640°C.
550°C.	660°C.	530°C.
	530°C.	510°C.
		460°C.

DISTRIBUTION OF Na_2O BETWEEN COEXISTING K-FELDSPAR AND PLAGIOCLASE IN DIORITE SERIES

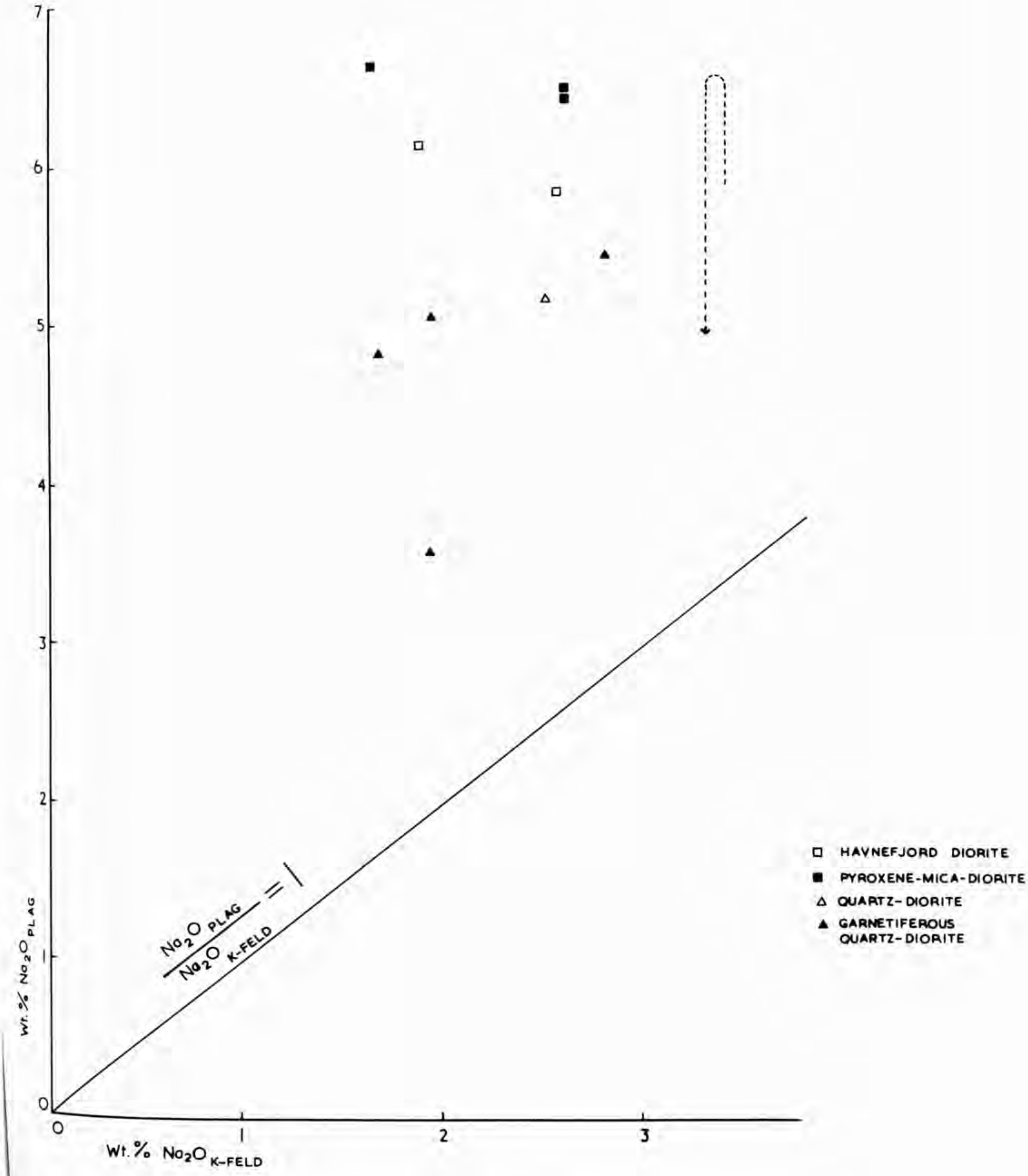


FIG. 28

Trace Elements

The trace elemts Ba, Rb, and Sr have been determined in the co-existing feldspars. The relationships between each of these elements and also their relationships to the major elements have been studied, and the results are presented below. In each case it is found that the trends observed in the whole rock analyses and in the major elements of the feldspars are reflected in the trace elements in the feldspars. There is a progressive series from the Havnefjord diorite to the quartz-diorites; the plagioclase in the Kobberfjord norite has not been analysed. Again it is found that the early diorites cause a reversal of the trend, confirming the suggestion that they do not belong to the main diorite suite.

Graphs of the partition coefficients, showing how the elements distribute themselves between the coexisting K-feldspar and plagioclase, are presented in Figs. 29-31.

The $Sr_{K\text{-}feld}/Sr_{plag}$ ratio is very near to unity, with a slight increase in the quartz-diorites (Fig. 29). Sr^{2+} has the same ionic charge as Ca^{2+}, but has an ionic radius (1.13Å) intermediate between K^{+}(1.33Å) on the one hand, and Na^{+}(0.95Å) and Ca^{2+}(0.99Å) on the other. (All ionic radii which are quoted are those determined by Pauling - see Cotton and Wilkinson, 1962, p.43. The use of these radii is recommended by Ahrens, 1952). All these ions have similar electronegativities (K^{+} = 0.8, Na^{+} = 0.9, Ca^{2+} = 1.0, Sr^{2+} = 1.0; electronegativity values are also Pauling's), and thus Goldschmidt's rules of diadochy are likely to be applicable (Ringwood, 1955). It is to be expected then that Sr^{2+}

DISTRIBUTION OF Sr BETWEEN COEXISTING K-FELDSPAR AND PLAGIOCLASE IN DIORITE SERIES

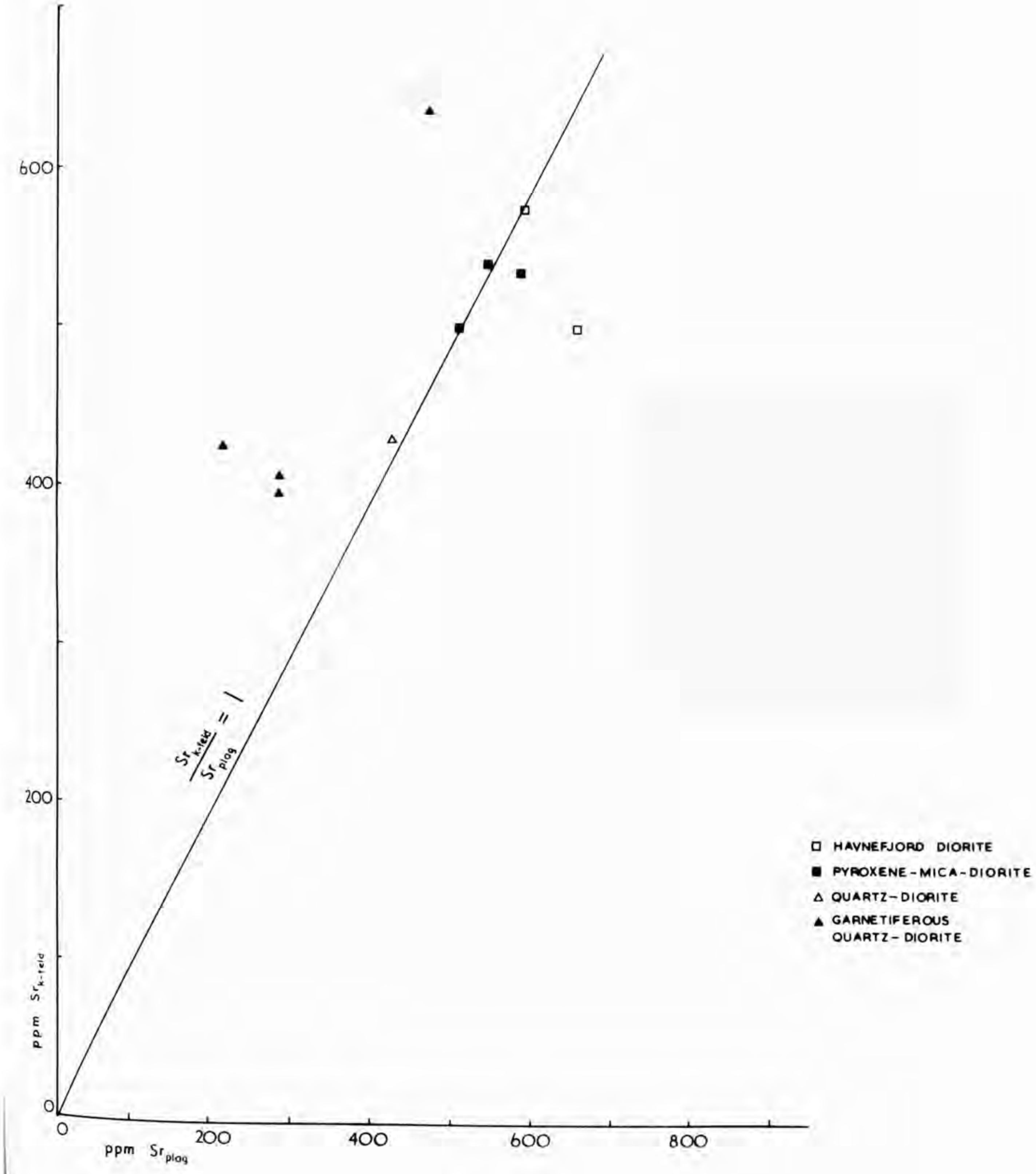


FIG. 29

DISTRIBUTION OF Ba BETWEEN COEXISTING K-FELDSPAR AND PLAGIOCLASE IN DIORITE SERIES

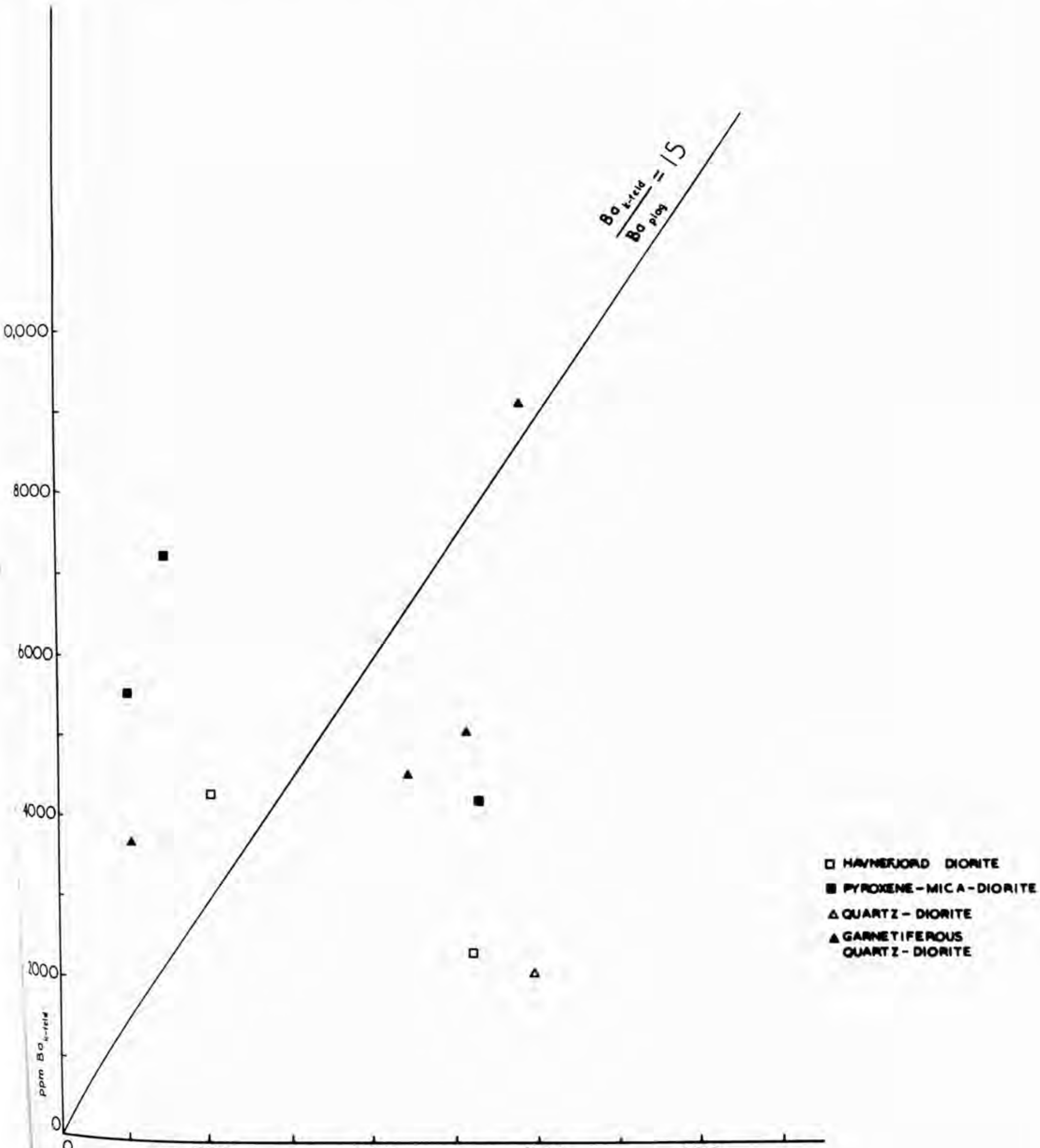


FIG. 30

DISTRIBUTION OF Rb BETWEEN COEXISTING K-FELDSPAR AND PLAGIOCLASE IN DIORITE SERIES

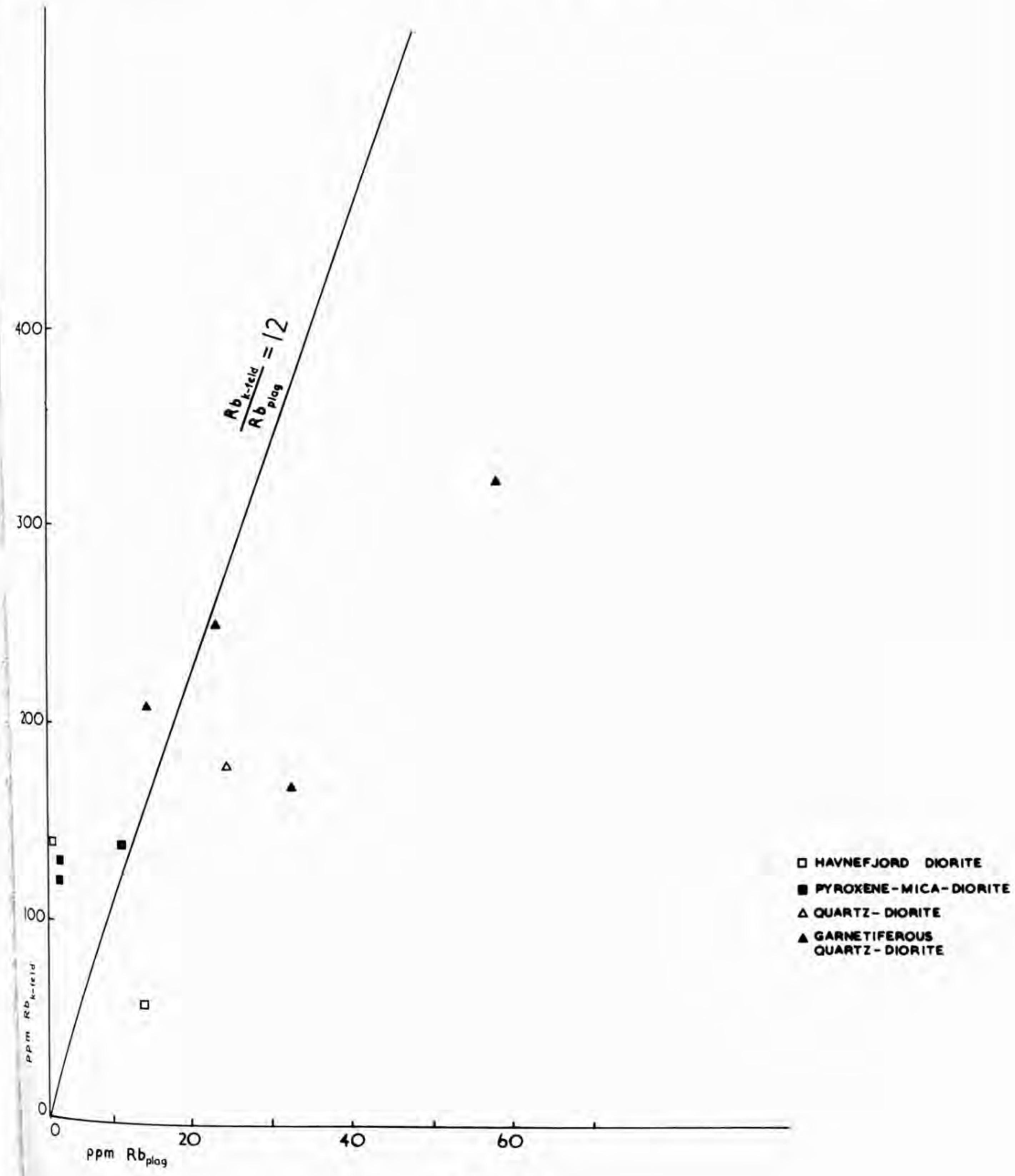


FIG. 31

would replace both K^+ in the K-feldspar lattice and Ca^{2+} in the plagioclase lattice with equal readiness. Hence the equal distribution of Sr between the coexisting feldspars.

In the case of Ba and Rb the distribution is more irregular, but is strongly biased in favour of K-feldspar. The $Ba_{K\text{-feld}}/Ba_{plag}$ ratio is approximately 15 (Fig. 30) and the $Rb_{K\text{-feld}}/Rb_{plag}$ ratio is about 12 (Fig. 31). Again, these ions have similar electronegativities to those of the relevant ions in the feldspars (Ba^{2+} = 0.9, Rb^+ = 0.8). However, the ionic radius of Ca^{2+} differs from those of Ba^{2+}(1.35Å) and Rb^+(1.48Å) by more than 15%, and this makes its replacement by Ba^{2+} and Rb^+ difficult, in spite of the fact that Ba^{2+} has the same ionic charge. Rb^+ has the same charge as K^+, and Ba^{2+} has a very similar ionic radius, and thus it is to be expected that these elements are concentrated in the K-feldspars.

According to Barth (1961), the coefficient of distribution depends upon the temperature. Both Ba and Sr prefer the K-feldspar lattice at high temperatures and the plagioclase lattice at low temperatures. In the case of Sr, the temperature at which the distribution ratio is unity is 450°C. according to Barth. Hall (1967), however, has found that in the Rosses complex, which is believed to be of igneous origin, plagioclase has 30% more Sr than K-feldspar, implying that Sr begins to be incorporated at temperatures above 450°C.

The concentrations of Rb, Ba, and Sr in the K-feldspar have been plotted against each other, and similarly those in the plagioclase. It was found that the trends in the K-feldspar were broadly sympathetic with those in the plagioclase, so a summary of the data has been presented by plotting the total concentrations of Rb, Ba, and Sr in both feldspars against each other (Figs. 32 - 34). The individual trends with respect to time are summarized in Table 20.

Increases in the absolute amounts of Rb and decreases in the amounts of Sr with increase of fractionation have been reported in K-feldspars by Heier and Taylor (1959a) and by Sen, Nockolds, and Allen (1959), and in pegmatites by Taylor, Emeleus, and Exley (1956). However, the latter workers and Heier and Taylor (op.cit.) report a decrease in the Ba content, which is contrary to the case obtaining in the present study. Heier and Taylor (op.cit.) also find that the Ba/Sr ratio decreases with time, whereas in the present case it increases.

It can be seen from the summary in Table 20 that in all cases, i.e. K-feldspar, plagioclase, and total, there is an increase, with respect to time, in the ratios Rb/Ba, Rb/Sr, and Ba/Sr. These three facts considered in conjunction provide the information that Rb is mostly concentrated in the later rocks, Sr is most abundant in the earlier rocks, and Ba occupies an intermediate position. This means that Sr was the first of the three to enter the feldspars, followed by Ba, and then by Rb.

The explanation for this probably lies principally in a combination of the effects of ionic charge and ionic radius, since all the ions involved have similar electronegativities.

In the case of plagioclase, where Ca is the element which is principally replaced, it is clear that Sr^{2+} are the most favoured ions

GRAPH OF TOTAL Rb AGAINST TOTAL Ba IN COEXISTING FELDSPARS IN DIORITE SERIES

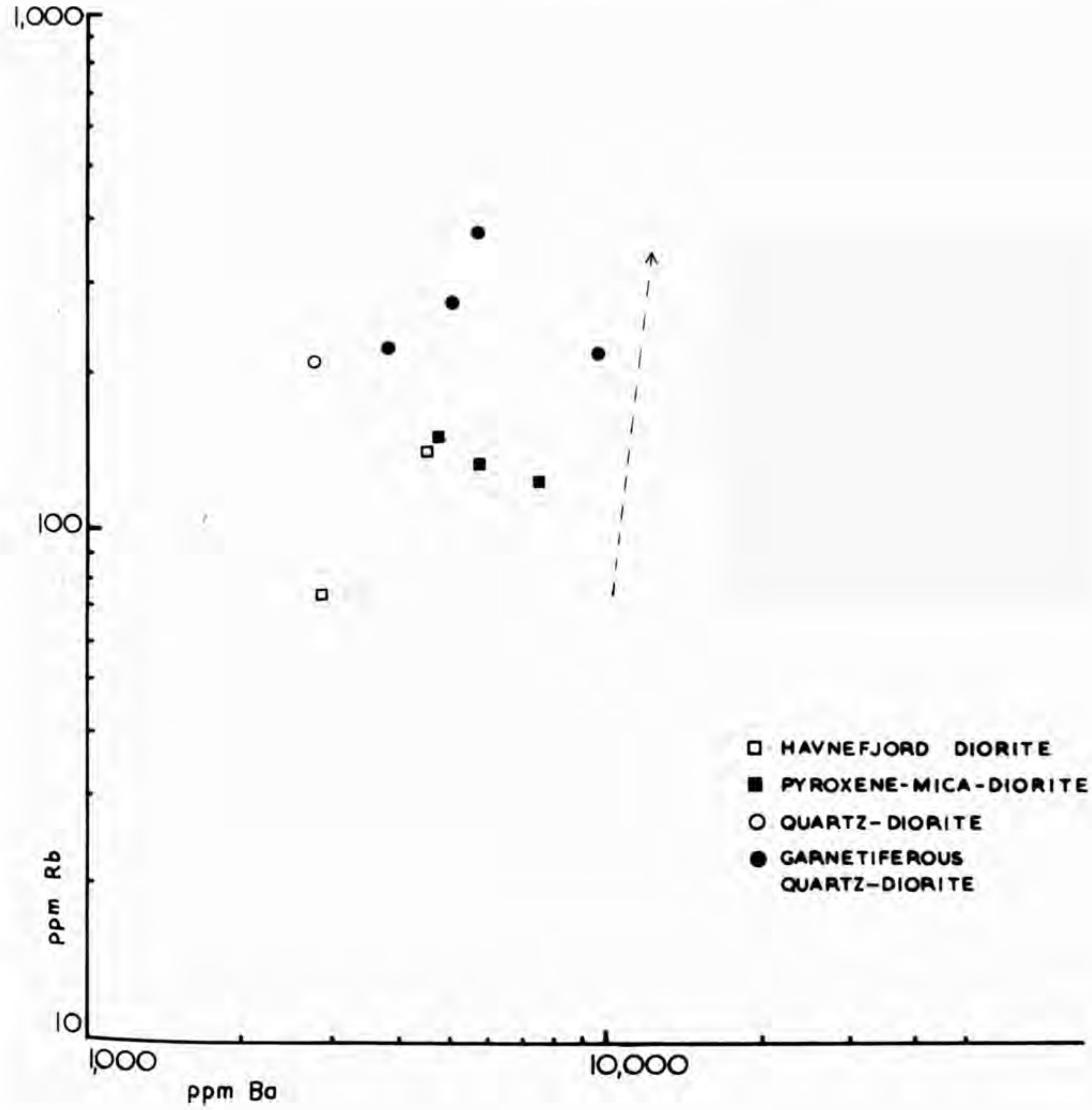


FIG. 32

GRAPH OF TOTAL Rb AGAINST TOTAL Sr IN COEXISTING FELDSPARS IN DIORITE SERIES

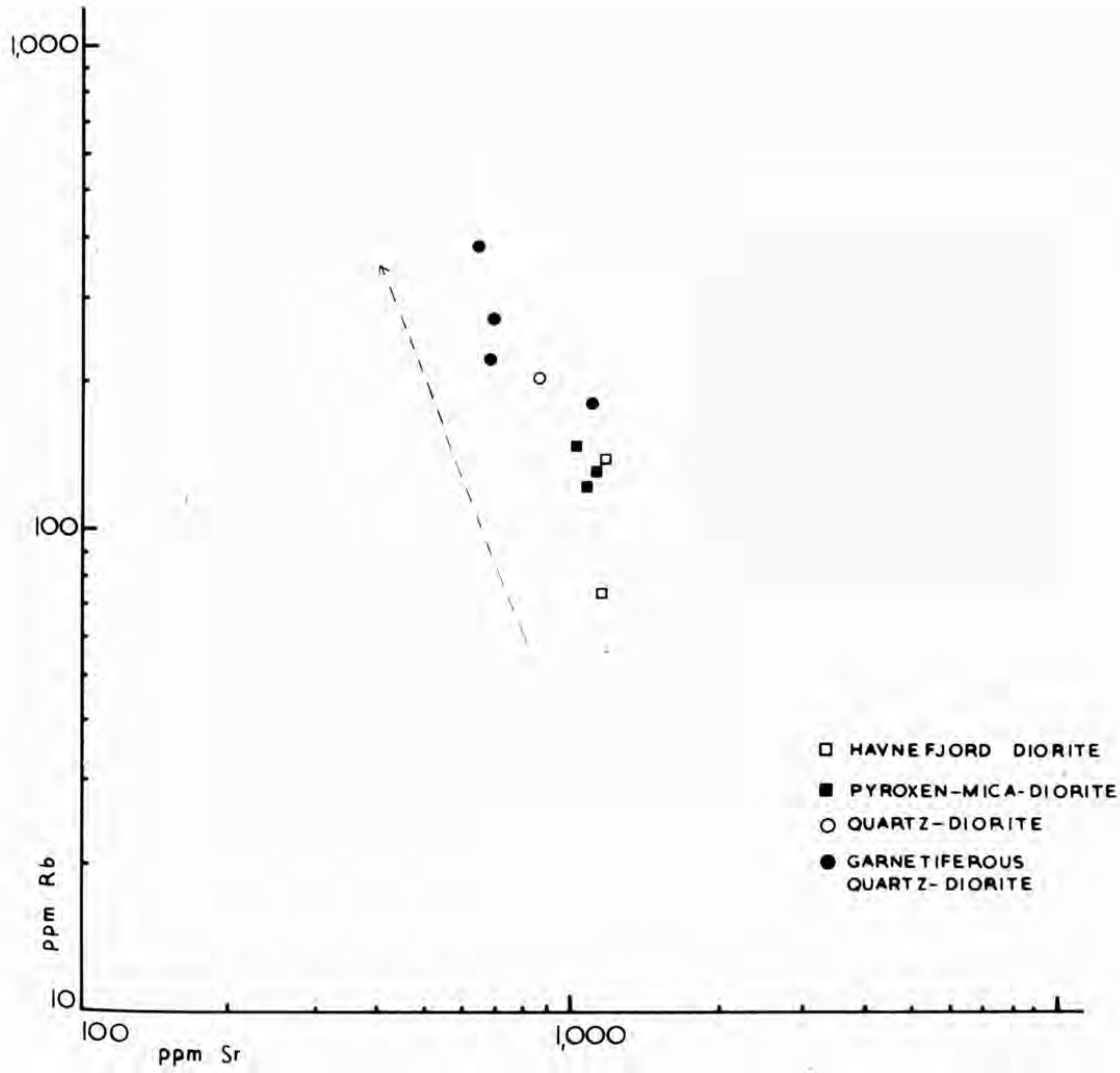


FIG. 33

GRAPH OF TOTAL Sr AGAINST TOTAL Ba
IN COEXISTING FELDSPARS
IN DIORITE SERIES

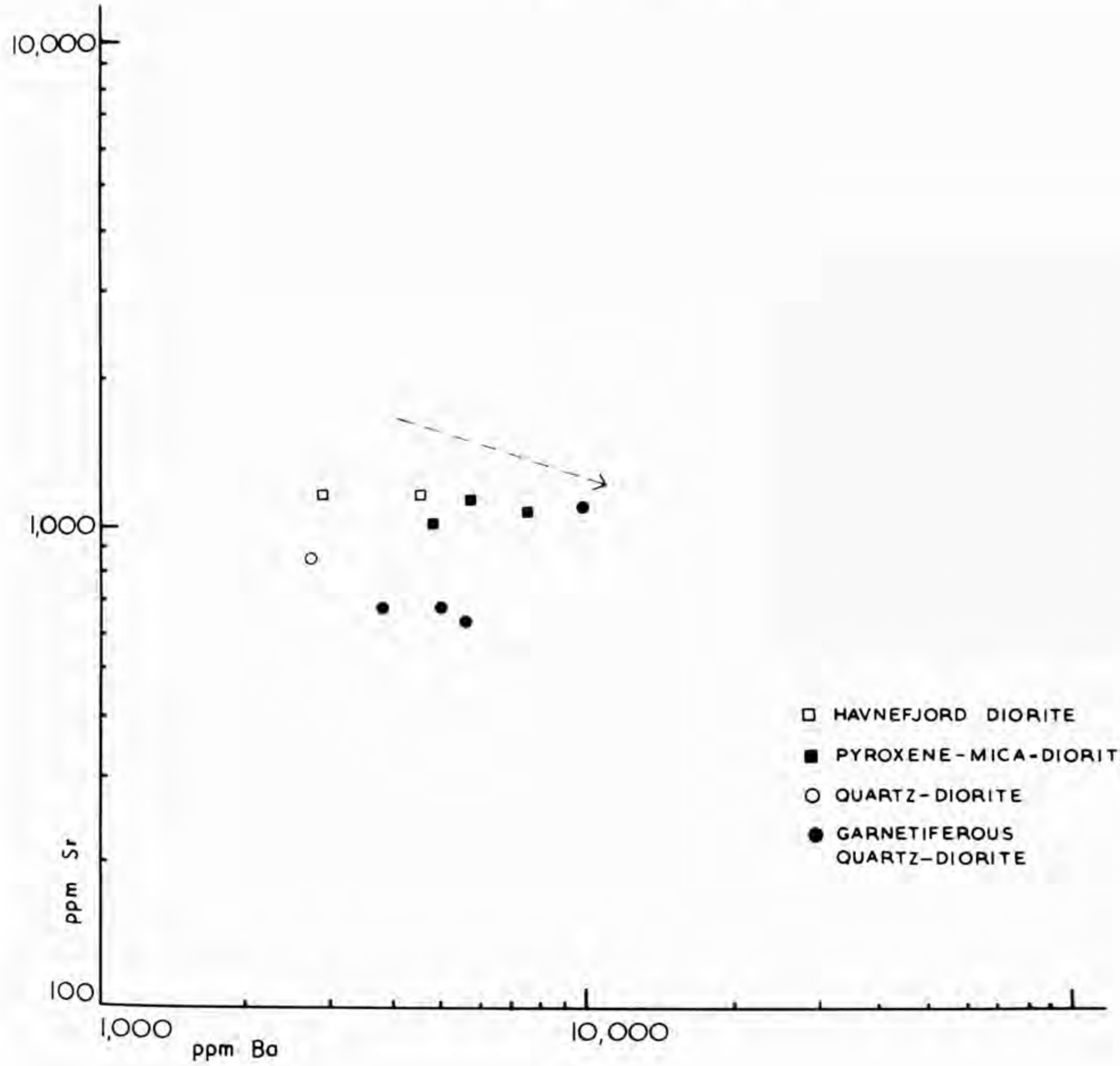


FIG. 34

TABLE 20

SUMMARY OF TRENDS IN TRACE ELEMENTS IN COEXISTING FELDSPARS IN DIORITES

Element	K-feldspar	Plagioclase	Total
Rb	increase	slight increase	increase
Sr	slight increase	decrease	slight decrease
Ba	increase	slight increase	increase
Ratio			
Rb/Ba	slight increase	slight increase	increase
Rb/Sr	increase	increase	increase
Ba/Sr	increase	increase	increase

for incorporation since they have the smallest ionic radius and also have the same charge as Ca^{2+}. Thus Sr^{2+} will be the first of the three to be incorporated. Ba^{2+} has the next largest radius and has the same charge as Ca^{2+}, so it is to be expected that Ba^{2+} will be incorporated next, and Rb^{+} last of the three. Entry of Sr^{2+} and Ba^{2+} in early stages were probably also facilitated by the more open, disordered structure of the feldspars at high temperatures (Sen, 1960; Bray, 1942).

The same laws probably hold good in the case of the K-feldspars. Sr^{2+} and Ba^{2+} are both divalent, and will enter the K-feldspar lattice early at the expense of univalent K^{+} (Heier and Taylor, 1959a). However, Sr^{2+} is the more favoured ion due to its smaller ionic radius, and will therefore enter first. Rb^{+} can also enter K^{+} sites in K-feldspars, but its entry gives precedence to the divalent Ba^{2+} (Heier, 1962). The large ionic radius of Rb^{+} guarantees that it does not enter the K-feldspar lattice until the later stages (Taylor and Heier, 1960).

Heier and Taylor (1959a) maintain that the bond strength of the elements is important, They find that the Ba/Sr ratio in K-feldspars decreases with increase in fractionation, and consider that this is due to the preferential acceptance of Ba^{2+} because it forms a more ionic bo nd with O^{2-} than does Sr^{2+}. However, in the present study, the Ba/Sr ratio increases, indicating a preferential acceptance of Sr^{2+}. Ba^{2+} has a lower electronegativity (0.9) than Sr^{2+} (1.0) and would thus tend to be preferentially incorporated to form a more ionic bond. However, it would seem in this present case that the difference in electronegativity is not sufficient to prevent the smaller size of Sr^{2+} from taking precedence, ensuring an early entry for Sr^{2+}. According to Ringwood (1955), provided the difference in electronegativity does not exceed 1, Goldschmidt's rules of diadochy should apply, as seen in the present case.

Each of the trace elements Ba, Rb, and Sr have been plotted against each of the major elements K, Na, and Ca, the results being presented in Figs. 35 - 40. It can be seen from the graphs that in the case of plagioclase, there is a good correlation between each of the trace elements with each of the major elements. In the K-feldspars however, there is only a marked relationship between Rb and each of the major elements. The trends shown by the graphs are summarized in Table 21.

An average ratio line has been drawn on each graph as a guide to the values of the ratios. Since on logarithmic graphs, lines of varying ratios are represented by parallel lines having positive gradients at 45° to the axes, it is suggested that the rate of change of the ratios may be judged by the angle with which the line of the trend intersects the ratio lines. Thus the ratios change at a maximum rate in series which trend perpendicularly to the ratio lines, and at a minimum rate (zero) in a series trending parallel to the ratio lines. This would mean that the K/Sr ratio in Fig. 36 increases at the maximum rate, and that the Ca/Sr ratio in Fig. 40 decreases at a very slow rate.

Although this present study shows that in plagioclase there is a good correlation between the trace elements and the major elements, studies by previous workers have been mainly concentrated on the relationship between Sr and Ca. In basic rocks it has been found that as the

PLOT OF K AGAINST Sr, Ba, AND Rb IN K-FELDSPAR IN DIORITE SERIES

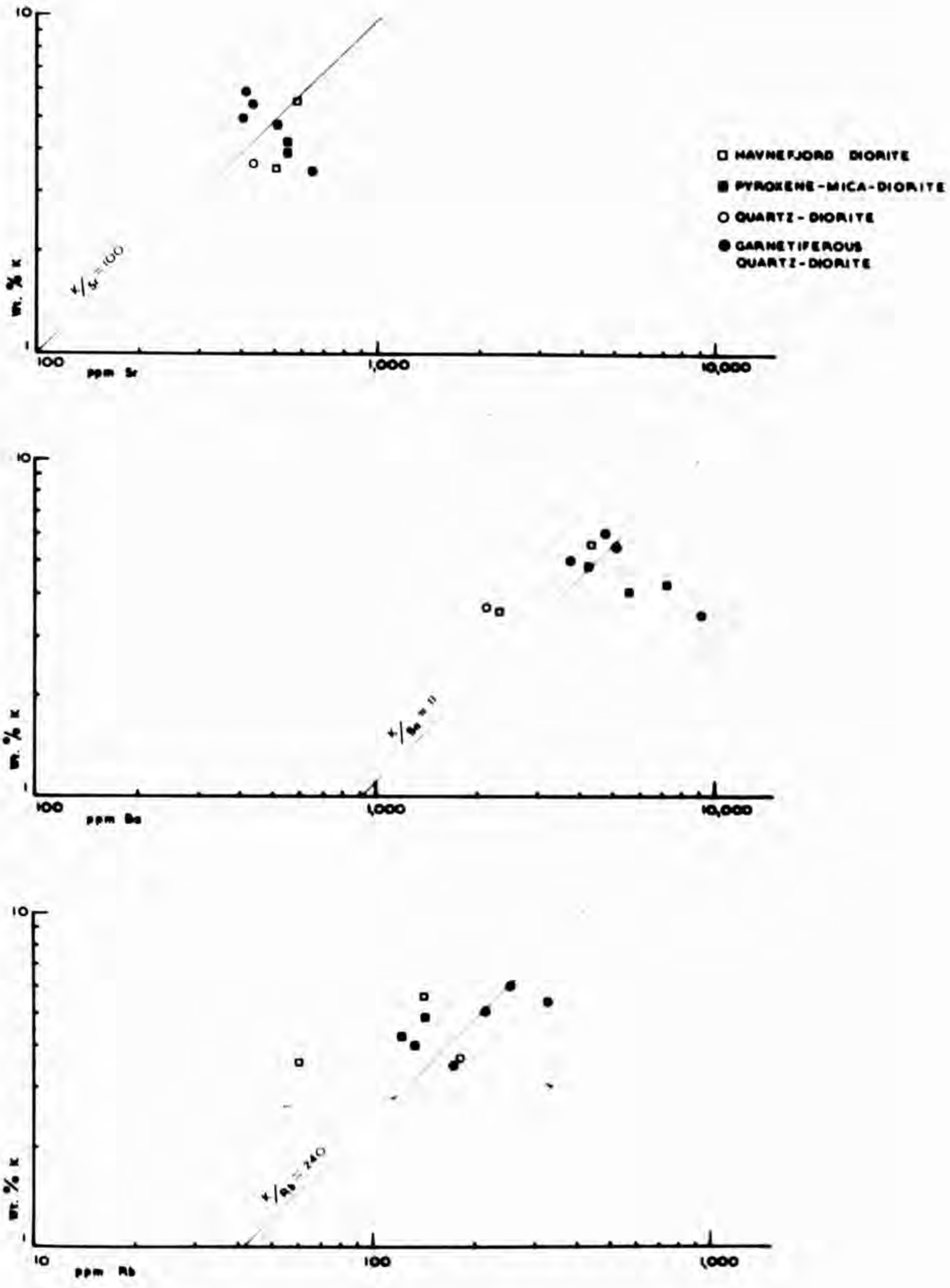


FIG. 35

PLOT OF K AGAINST Sr, Ba, AND Rb IN PLAGIOCLASE IN DIORITE SERIES

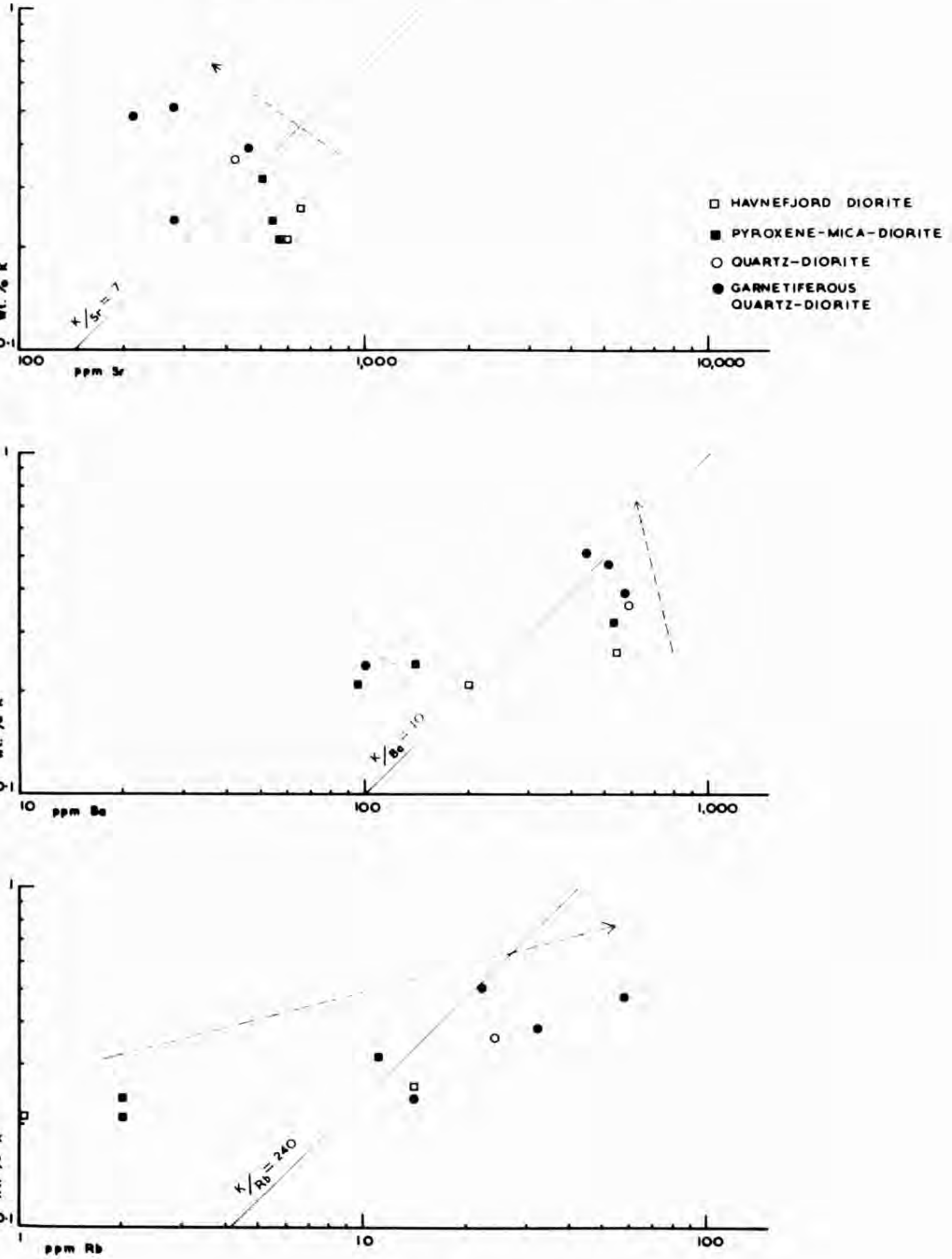


FIG. 36

PLOT OF Na AGAINST Sr, Ba, AND Rb

IN K-FELDSPAR IN DIORITE SERIES

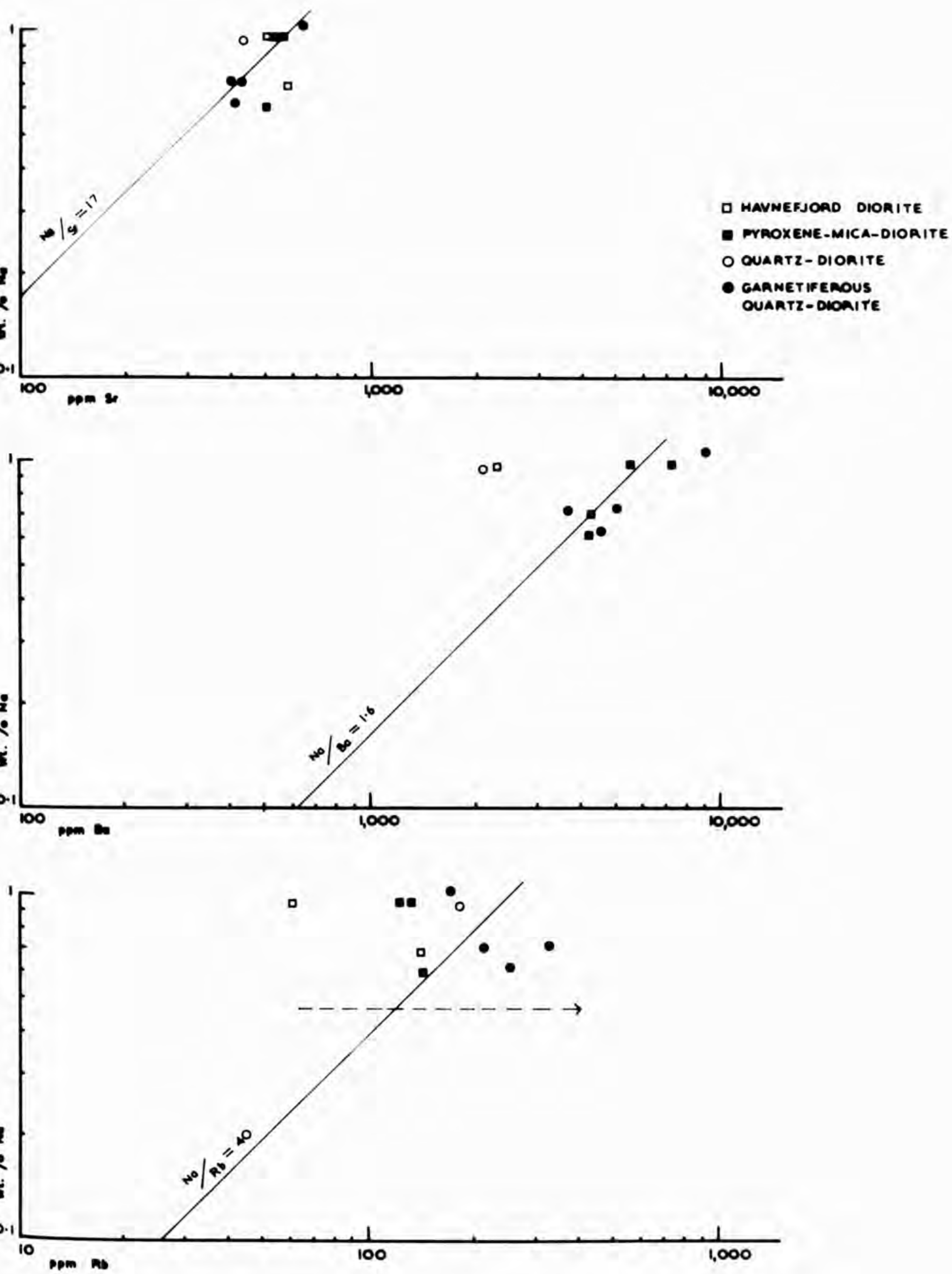


FIG. 37

PLOT OF Na AGAINST Sr, Ba, AND Rb

IN PLAGIOCLASE IN DIORITE SERIES

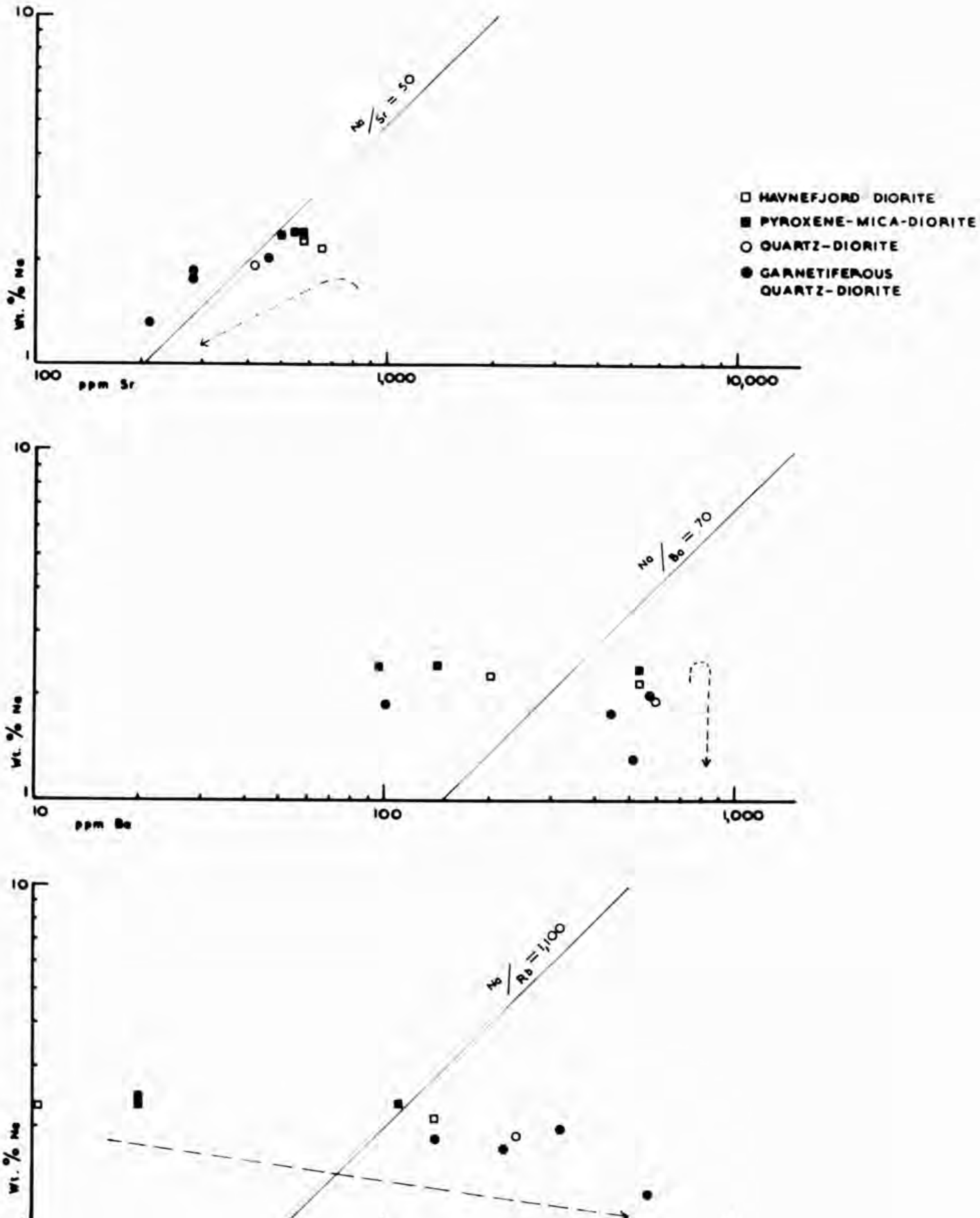


FIG. 38

PLOT OF Ca AGAINST Sr, Ba, AND Rb
IN K-FELDSPAR IN DIORITE SERIES

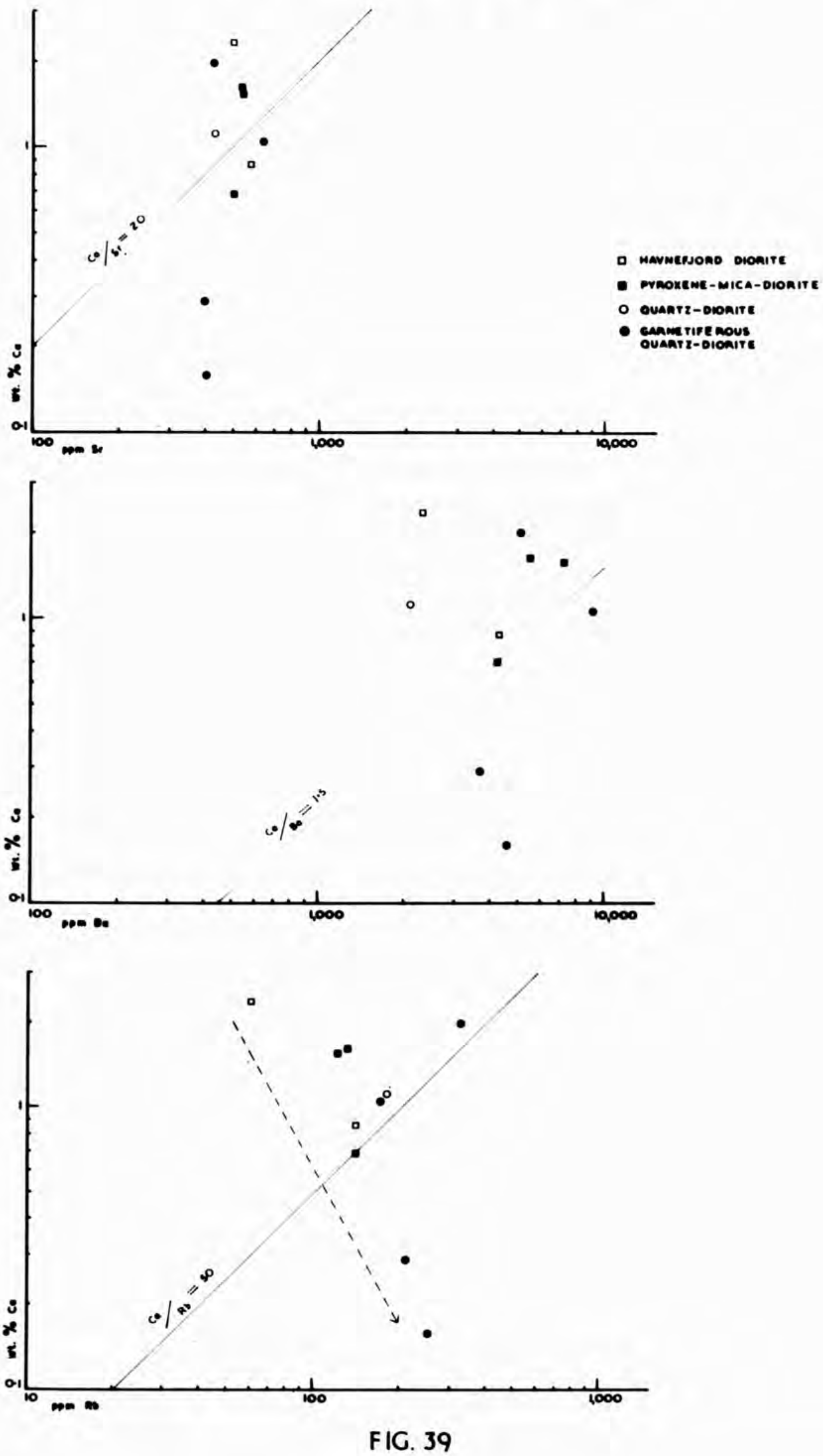


FIG. 39

PLOT OF Ca AGAINST Sr, Ba, AND Rb

IN PLAGIOCLASE IN DIORITE SERIES

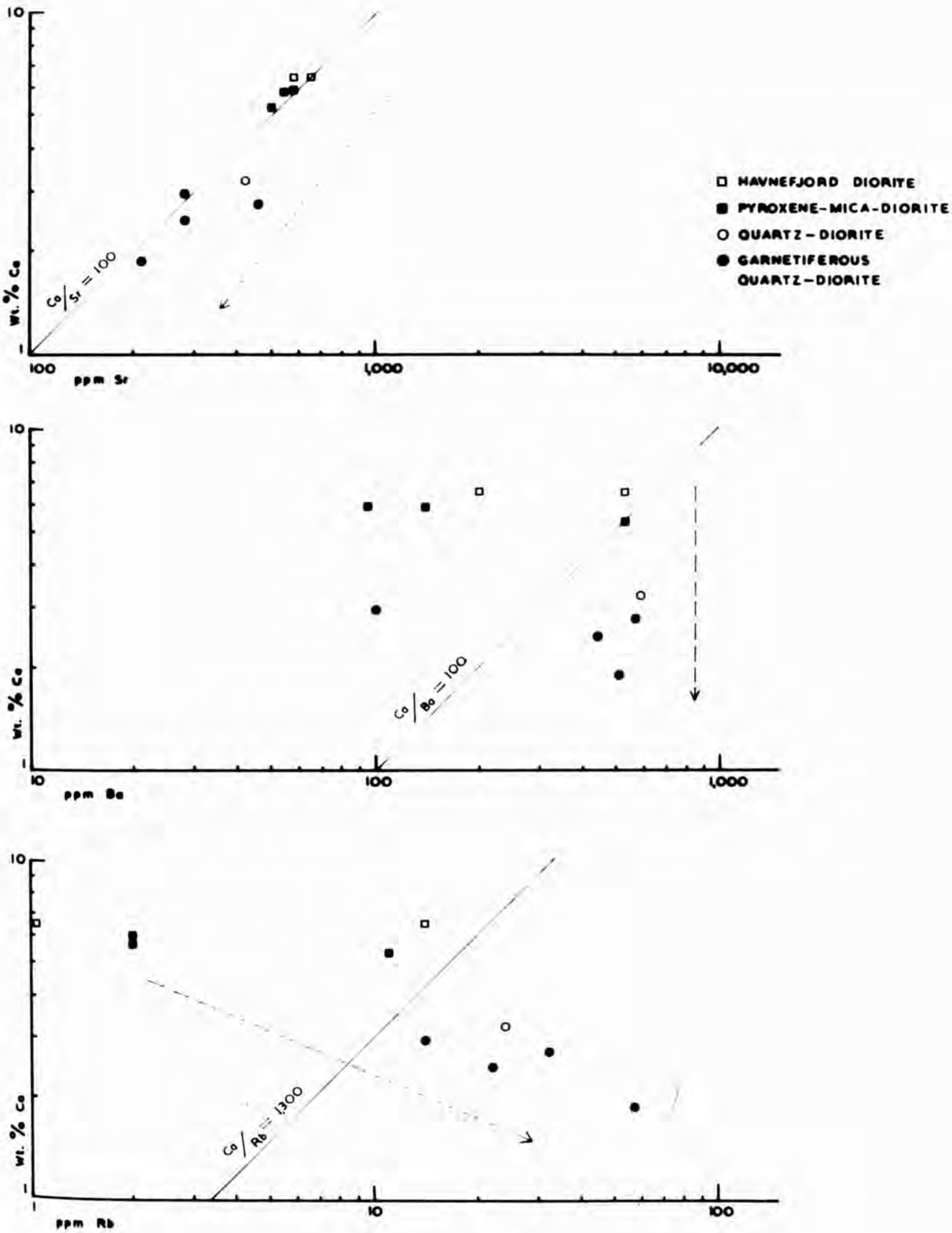


FIG. 40

TABLE 21

SUMMARY OF TRENDS IN TRACE ELEMENTS AND MAJOR ELEMENTS IN COEXISTING FELDSPARS IN DIORITES

Ratio	K-feldspar	Plagioclase
K/Rb	decrease	decrease
K/Ba	none	increase
K/Sr	none	increase
Na/Rb	decrease	decrease
Na/Ba	none	decrease
Na/Sr	none	slight increase
Ca/Rb	decrease	decrease
Ca/Ba	none	decrease
Ca/Sr	none	slight decrease

An content of the plagioclase rises, so the Sr content falls (Butler and Skiba, 1962). According to Wager and Mitchell, 1951, Sr is most abundant in intermediate plagioclase, which is in agreement with the case under present consideration in which there is a decrease in the Sr content from intermediate to more sodic plagioclase. In more acid rocks on the other hand, it has been shown that the Sr content increases with increase in An content (Sen, Nockolds, and Allen, 1959; Hall, 1967), which is in accordance with the present study (see Fig. 40). In fig. 40 it can be seen that the Ca/Sr is almost constant, but falls very slightly with decrease in Ca content (i.e. with increase in fractionation). A similar feature is reported from a teschenite sill in New South Wales by Wilkinson, (1959).

From this it would appear that the relative sizes of Ca^{2+} and Sr^{2+} are not the only factors involved in the substitution process, since the results are clearly different in the basic rocks and the acid rocks. The presence of K-feldspar coexisting with plagioclase in the latter must account for this difference. In the case of the basic rocks, a certain amount of Sr^{2+} could enter Ca^{2+} sites at high temperatures (i.e. early stages) due to the more open structure, but the smaller size of the Ca ion ensures that Ca^{2+} is the main entrant into the lattice. As fractionation proceeds, Ca^{2+} will freely enter the lattice, leaving a concentration of Sr^{2+} which are only captured at late stages when much of the Ca^{2+} has been exhausted. In the case of the more acid rocks, however, the presence of K-feldspar causes a difference in the course of events. Again, some Sr^{2+} would enter the first-formed plagioclase at Ca^{2+}

sites, and as the temperature dropped, less Sr^{2+} would enter in favour of Ca^{2+} so that the concentration of Sr in the plagioclase would begin to fall as fractionation proceeded (i.e. with fall in Ca; see Fig. 40). By now, K-feldspar is starting to form, and the Sr^{2+} which is beginning to become concentrated in the residue is captured by the coexisting K-feldspar. It only enters the Na^{+} sites with difficulty at fairly early stages in the K-feldspar's history, due to the small ionic radius of Na^{+}, and so Sr has a similar relationship with Na as it does with Ca (cf. Figs. 38 and 40). However, Sr^{2+} will occupy K^{+} sites in the K-feldspar very readily, being both smaller and more highly charged. Thus it will enter the K-feldspar lattice rapidly to the exclusion of K^{+} until much of the Sr^{2+} has been used up, leaving a concentration of K^{+} in late stages, which then enters the lattice freely with a rapid increase in the K/Sr ratio shown in Fig. 36.

In the K-feldspars, only Rb exhibits a good correlation with the major elements, particularly with K. A combined plot of K/Rb ratios in coexisting K-feldspars and plagioclases is presented in Fig. 41. Ahrens, Pinson, and Kearns (1952) determined the average K/Rb ratio in normal igneous rocks as about 90. In these determinations, the standards which were used were G-1 and W-1, which on subsequent re-analysis have given different values for Rb. On recalculating the K/Rb ratios with the new values, the average K/Rb ratio for normal igneous rocks becomes approximately 240 (Taylor, Emeleus, and Exley, 1956). The 240 ratio line has been inserted on Fig. 41, and it can be seen that it forms a good average for both feldspars. In this example, there is a Rb

PLOT OF K AGAINST Rb IN COEXISTING K-FELDSPAR AND PLAGIOCLASE IN DIORITE SERIES

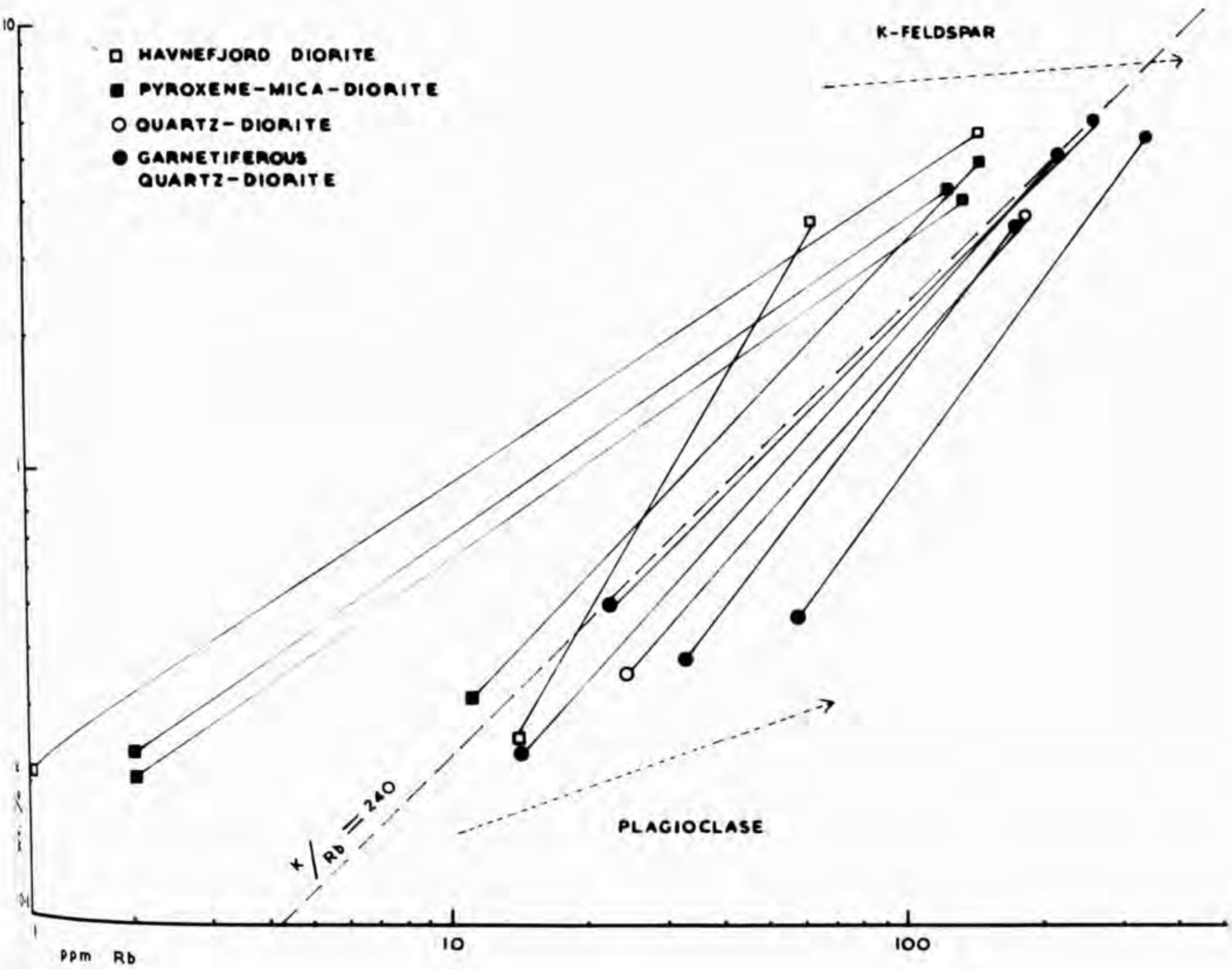


FIG. 41

enrichment both in absolute terms and with respect to K. A similar feature is reported by Herz and Dutre (1966), who find that K/Rb ratio vary from 1,180 in granodiorites down to 124 in late-stage granitic differentiates. Rb enrichment is often found in late-stage granites and pegmatites (Heier, and Taylor, 1959b; Taylor and Heier, 1960). This is probably a fairly straightforward case of the larger ionic radius of Rb^+ preventing it from entering the K^+ sites in the lattice until the later stages. It would app ear that such a mechanism was probably in operation in the present case.

It might be maintained that because both K and Rb are radioactive, the K/Rb ratio would change with the passage of time since the formation of the rocks. However, the half-life of Rb^{87} is about 5.9×10^{10} years which is far greater even than the age of the earth, and the amount of K^{40} (which has a half-life of about 1.3×10^9 years) is very small (now about .011% of total K). Thus Ahrens, Pinson, and Kearns (1952) consider that the total K/total Rb ratio as determined now in rocks is virtually the same as that when the earth was formed. If this is true, then the K/Rb ratio in the rocks under present discussion must have remained constant since the time of their formation, and thus the relating of their K/Rb ratios to their differentiation history is probably valid.

SUMMARY

In summary, it can be stated that all aspects of the chemistry of the diorites and norite, and some of their minerals indicate that the Kobberfjord norite, Havnefjord diorite, pyroxene-mica-diorites, and quartz-diorites form a petrological series. The nature of this series and its possible petrogenesis are considered in the next section.

PETROGENESIS

Introduction

The emplacement of the Husfjord igneous complex was a long and complex event, which took place during the Caledonian orogeny. It is not proposed to discuss the petrogenesis of the complex as a whole, since it seems unlikely that the entire complex is descended from a common source magma. Rather, the petrogenesis of the principal rock-type groups will be considered in turn.

HUSFJORD METAGABBRO

The earliest member of the complex, the Husfjord metagabbro, is similar mineralogically to the Storelv gabbro and the Breivikbotn gabbro, which have been described by Stumpf and Sturt (1965). These latter gabbros occur as sheets in northern and western Sørøy, and were emplaced approximately synchronous with the Husfjord metagabbro, i.E. at the end of the F_1 folding.

It seems reasonable to suggest that these gabbros were derived from the same source body of magma. It is possible that during the F_1 movements, a large body of basic or ultrabasic magma was emplaced into deep levels of the crust. Tectonic forces might well have created spaces at higher levels of the crust into which the magma could have entered from below. Thus these gabbro sheets probably formed by permissive emplacement of magma from this common deep-seated source.

ALKALINE ROCKS

The alkaline rocks of the area comprise nepheline-syenite pegmatites and perthosites. The latter are fairly common in the Vatna area, but there are only a few occurrences of the former.

According to Smyth (1913), alkali pegmatites are the final stage in the concentration of certain elements from a subalkaline magma from which alkaline magmas are locally derived. Nepheline-syenite pegmatites from Haliburton, Ontario, are ascribed by Foye (1915) to the pneumatolitic stage of the intrusion of the Laurentian granite, desilication of the pegmatites being caused by contamination by limestone. Nepheline-syenite pegmatites also from the Haliburton-Bancroft area are considered by Osborne (1930) to be offshoots of a larger body of nepheline-syenite which is closely associated in the field with limestone.

It is possible that the Vatna nepheline-syenite pegmatites could represent offshoots of an alkaline complex which may underlie Sørøy. The presence of such a source of alkaline material is further suggested by the occurrence of alkaline complexes at Breivikbotn (Sturt and Ramsay, 1965) and Dønnesfjord (Appleyard, 1965) in western and northern Sørøy. The Breivikbotn complex includes both nepheline-syenite pegmatites and leucocratic perthitic syenites, and it is considered that the alkaline rocks of the Vatna area are associated in time with the alkaline rocks of the Breivikbotn area.

Alkaline rocks are more commonly found in stable cratonic regions than in orogenic belts. The closest analogy to the synorogenic alkaline rocks of Sørøy is probably the alkaline complex of the Haliburton-Bancroft area, Ontario, described by Foye (1915), Osborne (1930), Tilley (1958), and Tilley and Gittins (1961). Sturt and Ramsay (1965) consider that the alkaline rocks of Sørøy have been derived from a common deep-seated parent magma, possibly alkali peridotite, by crystallization differentiation along separate lines of liquid descent.

NORITE-DIORITE SERIES

The Kobberfjord norite and the main suite of diorites were emplaced synchronous with the F_2 movements, after the peak of the regional metamorphism.

The characteristic mafic mineral of the norite and most of the diorites is hypersthene. According to Bowen (1928, p. 208) the tendency for orthopyroxene to form from a basic magma should be emphasized by the assimilation of aluminous sediments. Close association between norites and aluminous sediments have often been reported, particularly in the Aberdeenshire gabbros (Read, 1923, 1924, 1935, 1966; Read, Sadashivaiah, and Haq, 1965).

Features described by Read (1966) are very similar to some of the characteristics of the Kobberfjord norite and the pyroxene-mica diorites. The orthonorite described by Read has two distinctly different types of inclusions; one kind are sharply-defined xenoliths, and the other are fine-grained diffuse patches and streaks.

The former are pelitic xenoliths which Read considers to have had only slight reaction with the magma. Silica and alkalis are thought to have migrated from the xenoliths into the magma, and alumina, lime, and iron oxides in the opposite direction. The fine-grained diffuse inclusions on the other hand have the same mineralogy as the norite, and are called by Read 'micronorites'. The neighbouring norite appears to be forming from the micronorite inclusions, indicating a strong reaction between the inclusions and the magma.

There are no gradations between the two types, and Read believes that the process of contamination took place in two stages at different depths. He considers that the micronorites were incorporated early at deeper levels, and that the xenoliths were enclosed later and nearer the roof. Thus the magma could have been contaminated with pelitic material before it rose to its present position, where it was further modified by roof material now evident as xenoliths (see also Read, Sadashivaiah, and Haq, 1965).

The xenoliths in the Kobberfjord norite are generally sharply-defined, and in some cases are only just beginning to be assimilated at their margins. It is suggested that these are analogous to the obvious xenoliths in the Insch orthonorite described by Read (1966), and that the magma was already noritic when it arrived at its present position. The fact that many of the xenoliths are primarily of migmatite, the rock occurring outside the norite, suggests that they have a local origin. Although true 'micronorites' were not found in the Kobberfjord norite, patches enriched in garnet may mark the sites of earlier aluminous inclusions in the magma which have become completely assimilated. Thus the Kobberfjord norite may have a similar origin to the orthonorite described by Read (op. cit.).

The late pyroxene-mica-diorites contain andesine plagioclase and potash feldspar, but their principal mafic mineral is hypersthene, and they show certain characteristics similar to those of the orthonorite of Read (op. cit.). These diorite sheets contain both metasedimentary blocks, which are only just beginning to be assimilated, and many small fine-grained diffuse inclusions at an advanced stage of assimilation. These diffuse inclusions have a very similar relationship to the diorite as Read's micronorites do to his orthonorite.

In both cases, the inclusions grade into the surrounding rock, and their mineralogy is principally the same as that of their host, although the diorites contain potash feldspar in addition. The hypersthenes in the orthonorite and the diorites are strongly pleochroic, as are the small hypersthenes in the inclusions in both rock types. Around the borders of some of the inclusions in the diorites, a rim rich in strongly pleochroic hyp ersthene grains with interstitial and enveloping ore is quite common, and Read describes a similar feature in the micronorites. Read also describes an interesting texture in the orthonorite adjacent to the micronorite which appears to be closely analogous to the pyroxene-plagioclase symplectites and dactylites that have been described in the pyroxene-mica-diorites (p.202).

Read sums up the relationship between micronorite and orthonorite by considering that the micronorite is being made over into orthonorite, but states that this does not necessarily mean that all the orthonorite was formed in this way. It was seen on p.211 that a similar conclusion had been reached concerning the relationship between the pyroxene-mica-

-diorite and its diffuse inclusions. It appears that these inclusions are also being made over into the surrounding rock, but again, it is not possible to be sure how much of the diorite was formed in this way.

Although the Havnefjord diorite contains some locally-derived rafts, it has only a few diffuse xenolithic inclusions. If any more foreign material had been assimilated at depth, it has been completely digested to form a fairly homogeneous rock.

Contamination of the quartz-diorites is suggested by the presence in them of sporadic garnets which form part of the igneous paragenesis. Pyrogenetic garnets have been reported in granitic rocks (Brammall and Harwood, 1932; Pitcher, 1953b; Gindy, 1953; Hall, 1965b) and from volcanics (Oliver, 1956). Brammall and Harwood consider that sporadic accessory garnets in the Dartmoor Granite are due to contamination by metasedimentary country rocks or basic igneous rocks. Hall, on the other hand, can see no evidence for assimilation of country rocks to account for the garnet in the Donegal Granite. Analysis of the Donegal garnets shows them to be intermediate between almandine and spessartite, and Hall considers that a high Mn/Fe+Mg ratio would favour the formation of such a garnet. He maintains that a high Mn/Fe ratio tends to be present in late-stage acid rocks of a series because Mn does not enter muscovite at early stages.

In the case under present discussion, it is not possible to make deductions concerning the chemistry of the garnets as these have not been analysed, but it can be stated that the rock is not abundant in MnO (see Table 13). On the other hand field evidence suggests that the quartz-diorites could have been contaminated by metasedimentary material, and it is likely that this has been the cause of the formation of garnet.

Thus it appears that the various members of this suite of rocks have suffered contamination by metasedimentary material, and in some cases can actually be seen to be incorporating metasedimentary inclusions. On the other hand, studies of their petrochemistry show that the series appears to form a petrographic suite with smooth differentiation trends.

Nockolds (1941) found a similar paradox in the Garabal Hill-Glen Fyne igneous complex. There, the members of the complex lay along smooth curves in Larsen variation diagrams, and appeared to form a differentiation series, although some members had been contaminated by earlier rocks. Nockolds considered that the most likely explanation for the formation of the suite was to invoke the intermediate pyroxene-mica-diorite as the parent magma. Those members of the series more basic than this were considered to be accumulates, whereas those which were more acid and lay along the liquid descent lines of the variation diagrams were thought to have formed by the subtraction of the accumulates during differentiation.

It has been suggested by Reynolds (1935), however, that such a suite of rocks could develop from an ultrabasic parent magma of biotite-pyroxenite composition. According to her, the ultrabasic members of the suite would represent the parent magma, whilst the other less basic rock types are formed from the syntexis of the ultrabasic magma with sediments which have already been feldspathized by the magma's advance emanations.

In the norite-diorite suite under present discussion, it can be seen in the field that each member of the series has been contaminated at least locally by metasediment. On the other hand they appear to be along smooth liquid descent lines of variation diagrams.

However, according to Brammal (1933), an assimilation-series between basaltic magma and granitic material would differ little, in composition at least, from a differentiation-series in the line of liquid descent of a basaltic magma. This applies to both variation diagrams and trends on triangular diagrams. Furthermore, Thomas and Smith (1932) describe a whole range of igneous rocks from Côte du Nord which appear to form a normal differentiation-series. They believe, however, that this suite of rocks is a result of hybridization in which olivine-norite, an early member of the series, contaminates granite formed at a later stage. Potash-rich quartz-mica-diorites are considered to be the latest products of hybridization, formed by potash contamination. Poldervaart and Elston (1954) maintain that a calc-alkaline series of igneous rocks is not produced by normal fractional crystallization of basaltic magma, but by assimilation, metasomatism, or partial melting of sialic material.

It would appear from the above that the fact that the members of a suite of associated rocks lie on smooth curves of variation diagrams, does not exclude the possibility that the series has arisen by a process of hybridization. In the case under present discussion, it is manifest that contamination has occurred, and it is suggested that the curves of the variation diagrams do not necessarily represent true lines of liquid descent of a pure magma. A petrogenesis involving hybridization as an alternative hypothesis, however, must account for the early formation of norite followed by the emplacement of a large body of diorite and smaller bodies of quartz-diorite. Furthermore, it must explain the peaks in the Al_2O_3 and Na_2O curves in the variation diagrams in the Havnefjord diorite and the pyroxene-mica-diorites (Figs. 20 and 22). The proposed hypothesis is presented below.

It is suggested that during the peak of the regional metamorphism, which immediately preceded the emplacement of the norite-diorite complex, basic magma penetrated the base of the crust. As this magma soaked upwards, syntexis at the lower levels of the crust might have begun to take place, with crustal material becoming incorporated into the basic magma. Prolonged hybridization could have resulted in the formation of a large mass of fairly homogeneous magma of intermediate composition.

While this extensive soaking was proceeding, it is possible that in places the basic magma itself was intruded into the overlying region. This basic magma would, by now, have been contaminated to a certain extent by crustal metasedimentary material, possibly resulting in a norite as described by Read (1923), Bowen (1928), and Gribble and O'Hara (1967). Significant factors concerning the Kobberfjord norite are that it is relatively heterogeneous with traces of digested inclusions remaining, and that the hypersthenes are not strongly pleochroic. These features suggest that the norite was formed at early stages in the contamination process before homogenization was achieved.

Meanwhile, the hybridization was continuing beneath, and the basic magma was becoming progressively less basic until an intermediate, dioritic magma resulted. It is proposed that this large mass of dioritic magma was emplaced into the immediately overlying region as the Havnefjord diorite. This body is fairly homogeneous and contains strongly-pleochroic hypersthenes, suggesting that prolonged hybridization had taken place.

According to Daly (1928), a syntectic magma could attain to approximate homogeneity and could differentiate prior to its ascent to higher levels in the crust. It is suggested that once this deep mass of dioritic magma had been generated, it was capable of giving rise to a differentiation-series. The first differentiate would have been the mineralogically similar coarse pyroxene-mica-diorite, the coarseness of grain presumably reflecting a concentration of volatiles. These pyroxene-mica-diorites are commonly emplaced alongside the metasedimentary rafts and screens in the Havnefjord diorite, and have clearly assimilated a certain amount of pelitic material.

Later quartz-rich differentiates are represented by the coarse quartz-diorites which subsequently suffered slight contamination. The petrochemical evidence suggests that these quartz-diorites are highly fractionated rocks (p.262), and this probably accounts for the fact that they are not very abundant.

Thus the proposed hypothesis appears to account for the order of emplacement of the suite. It must, however, also explain the Al_2O_3 and Na_2O peaks in the Havnefjord diorite and pyroxene-mica-diorites, and their high iron content.

According to Read (1923), the norite at Arnage gains alumina and alkalis and loses lime and magnesia by reaction with metasedimentary xenoliths. Assimilation of shales by granitic magma can also cause alkali-enrichment of the magma, especially soda-enrichment (Brammall, 1933). If diffusion of alkalis into a basic magma takes place, soda will be incorporated into the magma more readily than potash if the magma is rich in lime (Nockolds, 1933). The Havnefjord diorite and the pyroxene-mica-diorites are fairly rich in lime (see Table 13), and it is suggested that the high Na_2O and high Al_2O_3 content of these two diorites is due to the large amount of pelitic material that the magma has assimilated.

The system proposed above is similar to that described by Reynolds (1934, 1935, 1936) for the Newry Complex. However, in the present case, ultrabasic bodies associated with the suite appear to be lacking, and so it is not possible to say whether an ultrabasic magma such as biotite-pyroxenite was the probable originator of the series.

A mechanism of syntexis is preferred to one of anatexis of sialic material for the present case for three reasons. First, anatexis will not account for the formation of the norite as the first member of the series. Second, in axatexis the earliest melts to form should be granitic, and these are lacking in the Husfjord area, at least at the present erosion level. Third, introduction of a basic magma from beneath could provide extra heat to assist the process of syntexis.

Diorites are considered by some to be the coarse-grained equivalents of andesite, in spite of the fact that diorites are not very common, whereas andesite is a very abundant lava. Although some diorites may have crystallized from an andesitic melt, possibly descended from a basaltic magma, many diorites show evidence that they have formed by hybridization between acid and basic material.

Fig. 18 shows that the trends in the system (K_2O+Na_2O) - Total Fe - MgO of lavas in the calc-alkaline province are different from the trend due to differentiation of a basic magma (after Hatch, Wells, and Wells, 1961). The diorite suite under present discussion falls close to the former trend, and it is significant that the Havnefjord diorite lies near to the position of andesite on the trend line.

Andesites and diorites are closely associated with orogenic belts, where a great thickness of sialic material continues down to considerable depths in the crust. It is considered by many that andesitic melts could be generated at these depths either by anatexis of sial or syntexis of sial with basaltic or ultrabasic magma emplaced from beneath (Buddington, 1959; Hatch, Wells, and Wells, 1961; Mehnert and Büsch, 1967; LeBas, 1967). Emanations rising from the mantle are probably an important factor in the heating up of the sial (Holmes, 1965).

Experimental work on the melting relationships of sediments has been carried out by Wyllie and Tuttle (1960, 1961). They show that the PT curve for the beginning of melting of shales is only about 20°C higher than the minimum melting curve of granite in the presence of water (see Fig. 43). According to Wyllie and Tuttle, a granodioritic melt is produced at about 150°C. above the first melting.

The average geothermal gradient in geosynclinal regions has been measured as 30°C./km. (Wyllie and Tuttle, 1960), and this curve has been added to Fig. 43, although Tuttle and Bowen (1958) consider that the gradient may steepen at a rate of about 6°C./km. with depth. Assuming a gradient of 30°C./km., shales should melt in the depth range of 20 - 25 km. (Wyllie and Tuttle, 1960). It is deduced from the metamorphic mineral paragenesis of the Husfjord rocks that the approximate depth of the regional metamorphism at its peak was in the order of 23 km., and the temperature in excess of 595°C. (see p. 304). Thus it would appear reasonable that with the additional heat provided by the introduction of basic magma from beneath, dioritic rocks could have formed in Sørøy from the syntexis of sedimentary material at deep levels of the crust.

The minimum melting curves of granite and of alkali basalt (after Yoder and Tilley, 1962) have also been added to Fig. 43. The minimum melting curve of diorite presumably lies between the two, probably just over 150°C. above the granite melting curve. It is concluded that the regional formation of a dioritic melt from syntexis of sialic material with basaltic magma is a possibility for the genesis of diorites formed in orogenic belts.

PART IV

METAMORPHISM

METAMORPHISM

Introduction

In the descriptive petrography, the mineralogy and textures that formed in the various rock types as a result of both contact and regional metamorphic events were discussed. The textures of one phase of metamorphism are often superimposed upon those of another, and by studying these relationships, a sequence of metamorphic and structural events in relation to the emplacement of the igneous complex can be determined.

Structural evidence shows that the rocks of this area underwent two major episodes of deformation, designated F_1 and F_2. Petrographic evidence indicates that these deformation episodes were accompanied by regional metamorphism, and that periods of static regional metamorphism occurred between F_1 and F_2. It is considered that the metamorphic event was essentially a continuous process, reaching its peak between F_1 and F_2 and waning during F_2.

The Husfjord metagabbro was intruded during the last stages of the F_1 movements, and the remainder of the complex was emplaced during the F_2 folding episode. The relationships between these igneous, structural, and metamorphic events are summarized in Fig. 42.

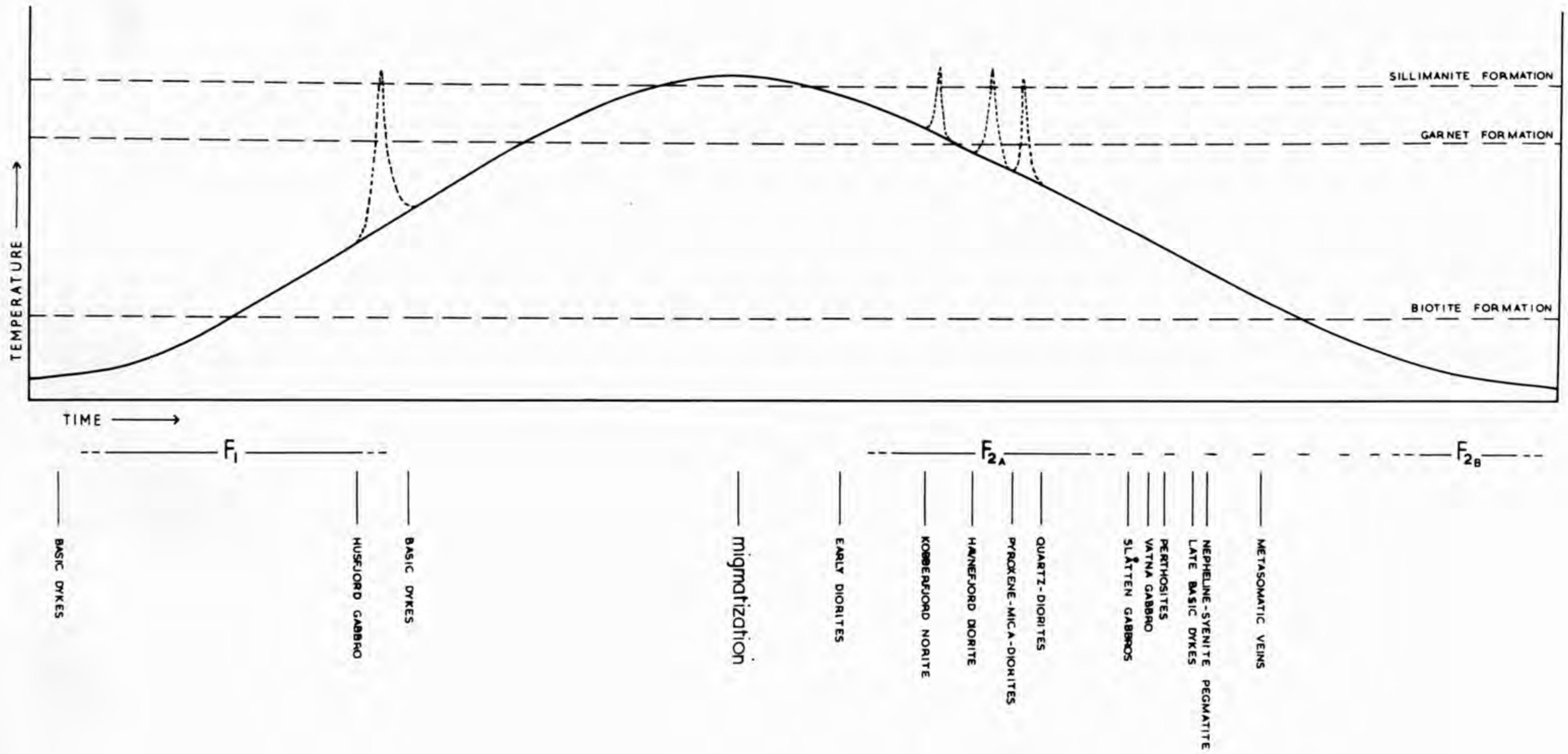
For the present study, the phases of metamorphism are considered in chronological order, under the following headings:

F_1 syn-tectonic regional metamorphism

Contact metamorphism due to Husfjord metagabbro

Post-F_1, pre-F_2 (static) regional metamorphism

SUMMARY OF METAMORPHIC, TECTONIC, AND IGNEOUS EVENTS IN THE HUSFJORD AREA



SOLID TEMPERATURE CURVE REPRESENTS THE REGIONAL METAMORPHISM. DOTTED TEMPERATURE CURVES REPRESENT CONTACT METAMORPHISMS SUPERIMPOSED UPON THE REGIONAL METAMORPHISM

FIG. 42

Contact metamorphism due to early diorites

F_2 syn-tectonic regional metamorphism

Contact metamorphisms due to main igneous complex

Post-F_2 (static) regional metamorphism

F_1 SYN-TECTONIC REGIONAL METAMORPHISM

The earliest F_1 fabric that has been recorded is that preserved in basic sheets which have been deformed by later phases of F_1. In these sheets a metamorphic schistosity is folded around isoclinal F_1 minor folds. The schistosity is not strongly developed, but is delineated by hornblende which has formed from the alteration of clinopyroxene.

Relict F_1 fabrics occur in some of the country rocks, particularly in the semi-pelitic schists. They are also sometimes preserved within the metasedimentary rafts which have been enclosed by the Husfjord metagabbro and the Havnefjord diorite.

The F_1 schistosity is delineated by laths of biotite and muscovite. Biotite is by far the more common mica, and when syn-F_1 muscovite does occur, it is occasionally seen to be altering to biotite, particularly along cleavage planes.

The biotites frequently alter to garnet, kyanite, and sillimanite, and these minerals overprint the F_1 schistosity, and seem to have grown under static conditions. This indicates that the F_1 movements must have ceased before the growth of these minerals.

Some of the pelitic rafts within the Husfjord metagabbro contain actinolite which has a preferred orientation, forming what is probably an F_1 schistosity.

Occasionally the F_1 schistosity, as delineated by orientated biotites, exhibit slight deformation in the form of small folds and kinks.

The only instance where a late syn-F_1 metamorphism can be distinguished is in the fine-grained, slightly foliated facies of the Husfjord metagabbro. At the time of emplacement of this gabbro sheet, the F_1 movements had practically ceased, and the metamorphism was increasing in grade (see Fig. 42). In this facies of the metagabbro, the clinopyroxene is altering to green hornblende, and the hornblende crystals show a slight tendency to have a preferred orientation. This foliation is not very marked, and was presumably formed during the last stages of the F_1 deformation.

F_1 syn-tectonic fabrics have been preserved in the metasediments of the Langstrand area (Roberts, 1965), principally as relics within later porphyroblastic minerals such as garnet and amphibole.

CONTACT METAMORPHISM DUE TO HUSFJORD METAGABBRO

The mineralogy and textures formed in the metasediments during the F_1 syn-tectonic metamorphism have been destroyed in places by the thermal effect of the Husfjord metagabbro's intrusion.

In the country rocks that form the envelope to the metagabbro, it is the calc-silicate-schists that appear to have been the most sensitive to thermal changes. The other metasediments have not been significantly altered.

Towards the contact with the metagabbro, a granoblastic hornfelsic texture is developed in the calc-silicate schists. Quartz, rounded grains of diopside, and occasional scapolite crystals are accompanied at the contact by buff-coloured garnet. In many cases, the hornfelsic texture has been overgrown by a later regional metamorphic fabric, and only remains as a relic.

Hornfelsing also occurs at the margins of the metasedimentary rafts enclosed by the metagabbro. The mineralogy of the calc-silicate-schist rafts is similar to that described above, but with the addition of phlogopitic biotite in some cases. Sometimes a relict F_1 schistosity is preserved in the pelitic schist rafts, but these rafts become completely hornfelsed at their margins. Quartz and feldspar form polygonal grains and biotite laths are orientated into a decussate pattern. A few of the pelitic schist rafts have microcline porphyroblasts, and some of the more aluminium-rich pelites contain corundum.

POST-F_1, PRE-F_2 (STATIC) REGIONAL METAMORPHISM

Between the two major episodes of folding, there was a period of static conditions, during which the metamorphic grade reached its peak. This metamorphism is considered to be a continuation from the F_1 syntectonic metamorphism, and represents a progressive increase in pressure-temperature conditions, which did not begin to wane until shortly before the F_2 movements (see Fig. 42).

A special determination of metamorphic zones has not been possible, but the mineralogy of the semi-pelitic schists indicates that the metamorphic grade attained the sillimanite-almandine subfacies of the almandine-amphibolite facies (Turner and Verhoogen, 1960).

Country Rocks

The period of recrystallization in the country rocks under static conditions is characterized by the development of porphyroblasts. These include garnet, kyanite, and sillimanite, and their textures generally indicate that they have grown in a static environment. Occasionally kyanite has a slightly preferred orientation, and the significance of this will be discussed below.

Garnet

None of the garnets examined showed textures, such as sigmoidal inclusion trails, to suggest that they grew syn-tectonically. In all cases they overgrow the F_1 schistosity which is delineated principally by biotite laths. Almost invariably, garnet forms equidimensional grains, usually containing inclusions of biotite and quartz.

Occasionally, garnets have been slightly augened, and in one example enclosed biotite laths are subperpendicular to the schistosity outside the garnet. These garnets have clearly suffered deformation and even rotation, and are thus pre-F_2.

Garnet is ubiquitous throughout the psammitic and semi-pelitic schists, and it is commonly seen to be forming from biotite.

The nucleation of fibrolite on some garnet porphyroblasts indicates that garnet grew while the metamorphic grade was on the increase. There is evidence, however, that some of the garnet grew while the metamorphism was on the wane. These garnets overprint kyanite fabrics, and can sometimes be seen forming from kyanite, the latter occasionally occurring as ghost relics within the garnet. It is clear that these garnets post-date the kyanites, which grew at the highest grades of metamorphism, and, therefore, they must be of later origin. In many cases, both kyanite and biotite appear to contribute towards the formation of these late garnets. A simplified equation of the reaction is probably as follows:

$$\underset{\text{biotite}}{K_2O.6(Mg.Fe)O.Al_2O_3.H_2O.6SiO_2} + \underset{\text{kyanite}}{Al_2O_3.SiO_2} \rightarrow$$

$$2\left[\underset{\text{garnet}}{3(Mg.Fe)O.Al_2O_3.3SiO_2}\right] + K_2O + SiO_2 + H_2O$$

Kyanite and Sillimanite

Kyanite and sillimanite appear to coexist stably in many rocks, although rarely kyanite is seen to be recrystallizing to sillimanite at its margins. From this it may be deduced that the conditions prevailing in this region were near to those at the kyanite/sillimanite stability boundary.

Both kyanite and sillimanite overprint the F_1 schistosity, and are generally randomly orientated. Occasionally, kyanite porphyroblasts have been augened by F_2 movements. In one of the rafts in the Havnefjord diorite, kyanite forms radiating aggregates which are overgrown by later garnets, but more usually such radiating textures are not formed. Many kyanite porphyroblasts are poikiloblastic, with inclusions of biotite being common. Sometimes kyanite shows a slight tendency to have a preferred orientation, and this suggests that growth may have continued into the beginning of the F_2 movements (see p.294 for further discussion). Sillimanite generally occurs as knots and aggregates of small prismatic needles, but in the migmatites it also occurs as fine fibrolite along grain-boundaries.

Kyanite and sillimanite both form from biotite, the sillimanite nucleating along the cleavage planes of the biotite. Chinner (1961) has described the association of sillimanite with biotite in Glen Clova, Angus, but considers that the constituents of the sillimanite were not derived from the biotite but from nearby kyanite, the biotite merely acting as a nucleus for the growth of sillimanite. In a later paper, however, Chinner (1966) concedes that sillimanite may form from biotite. Watson (1948) believes that sillimanite has formed from biotite in migmatized pelites and semi-pelites of Kildonan, Sutherland, and Pitcher (1965) considers that in many cases textural evidence for the replacement of biotite is difficult to deny. Roberts (1965) has reported the formation of sillimanite from biotite in schists from the Langstrand area of Sørøy, and in the present case, the texture suggests that biotite is being replaced by sillimanite.

Thus in the country rocks, the static regional metamorphism is characterized by the growth of porphyroblasts of garnet, kyanite, and sillimanite as the metamorphic grade increased, and a re-growth of garnet porphyroblasts from kyanite and biotite as the grade began to decrease.

Igneous Rocks

The only igneous rocks that had been emplaced prior to the static period, and therefore affected by this phase of the metamorphism, were the Husfjord metagabbro and some early basic sheets.

The effects are best shown in the Husfjord metagabbro, in which the principal changes in mineralogy involve the alteration of clinopyroxene into green hornblende and a corresponding decrease in the anorthite content of the plagioclase from labradorite to andesine. These two alterations are probably linked in the following way:

$$\underset{\text{clinopyroxene}}{CaO.FeO.2(Al.Fe)_2O_3.2SiO_2} + 2\underset{\text{calcic plagioclase}}{[CaO.Al_2O_3.2SiO_2]} + 6SiO_2 + H_2O + 2Na^+$$

$$\rightarrow \underset{\text{hornblende}}{2CaO.5(Al.Fe)O.H_2O.8SiO_2} + \underset{\text{andesine}}{(Na.\tfrac{1}{2}Ca)_2O.Al_2O_3.4SiO_2} + Al^{3+} + 5O^{2-}$$

Apart from the areas where the metagabbro has a slight F_1 foliation, the metamorphic texture is clearly a static one, and relict subophitic textures are commonly preserved.

The formation of the actinolite in the actinolitic facies of the metagabbro was probably formed at a later lower-grade stage of metamorphism, and is discussed in the appropriate section below.

Migmatization

The psammites and semi-pelites of the country rocks in the south-west of the area have undergone migmatization. This was associated with the attainment of the highest grades in the pre-F_2 static metamorphism, and detailed description and discussion of the migmatites have been reserved for this section.

Feldspathization of the metasediments has occurred in the Langstrand area of Sørøy (Roberts, 1965), and this took place at a late stage in the pre-F_2 static metamorphism. Much of northern and western Sørøy is occupied by migmatites (Sturt, personal communication, but as yet these have not been described in detail.

The migmatites in the writer's area are particularly well displayed at the northern end of Kobberfjord. The principal form taken by the migmatites is that of phlebite, in which feldspathic material occurs as bands and streaks parallel to the foliation planes of the metasediment, so that a 'mixed rock' (Sederholm, 1926) results. Much of the phlebite appears to be venite, while arterite may also be present.

The feldspathic material sometimes has a cross-cutting relationship to the banding, forming diadysite (Plate 222). This feldspathic veining is more common in the semi-pelitic horizons than in the massive psammites, presumably because the veins could penetrate the former more easily. The veins themselves are commonly rich in garnet (Plate 222), and thus these garnets cannot pre-date the migmatization.

Härme (1965) has reported a similar feature from migmatites of southern Finland. Garnet was produced during the normal regional metamorphism before the migmatization, although in the vein formation subsequent to the regional metamorphism, the garnet greatly increased in amount. He considers that the extra aluminium needed for garnet formation was released from the reaction which took place when plagioclase was altered to potash feldspar during the migmatization. It is possible that a similar situation existed in the present example under discussion.

Plate 222. Diadysitic migmatite; note garnets in feldspathic vein. Kobberfjord.

It is clear that in the most intensely migmatized zones, the metasediments have become deformed and mobilized, and the migmatite becomes agmatitic (Plates 223 and 224). The more incompetent semi-pelitic bands, possibly lubricated by feldspathic material, have become mobile much more readily than the more competent psammitic bands and amphibolites. This has resulted in the psammites and amphibolites forming blocks, surrounded by mobilized feldspathic semi-pelitic material which swirls around them (Plates 225 and 226). Sometimes this mobilized material has cross-cutting relationships with the blocks and also within itself (Plate 227).

Almost invariably the psammite and amphibolite blocks and bands have a regular jointing which scarcely penetrates the surrounding mobilized semi-pelite (Plate 228). Commonly only one joint direction is developed (see Plates 225 and 226), and the spacing of these joints increases as the width of the blocks increases. The direction of the joints in one block is practically the same as that in neighbouring blocks (see Plates 222, 223, and 224). This direction is parallel to a prominent F_2-jointing in the non-mobilized migmatites, and this indicates that the migmatization was essentially pre-F_2. Further confirmation of this age is given by the fact that the joints also affect some of the feldspathic veins (Plate 229). The dearth of jointing in the mobilized semi-pelites is probably due to their lower competence compared with the psammites.

Both the feldspathic veins and the mobilized semi-pelites are rich in garnet (Plates 222 and 230), whereas the massive psammitic horizons

Plate 223. Agmatitic migmatite. Kobberfjord.

Plate 224. Agmatitic migmatite. Kobberfjord.

Plate 225. Psammite blocks in mobilized semi-pelite. Kobberfjord.

Plate 226. Psammite blocks in mobilized semi-pelite. Kobberfjord.

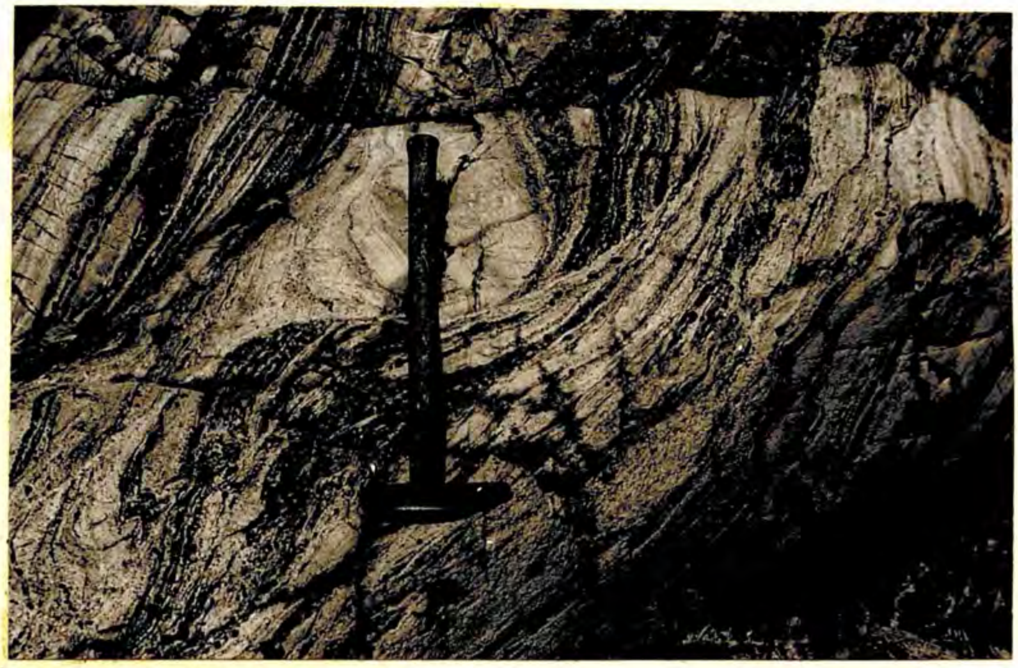


Plate 227. Discordant relationships of mobilized semi-pelite. Kobberfjord.

Plate 228. Joints in psammites. Kobberfjord.

Plate 229. Joints affecting feldspathic veins.
Kobberfjord.

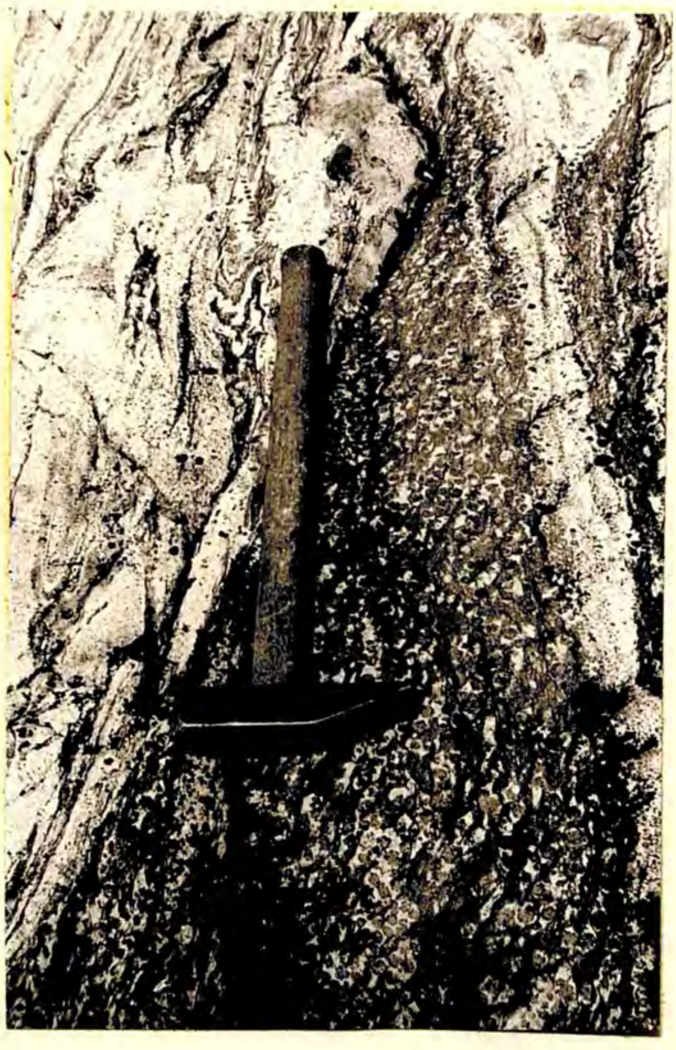


Plate 230. Garnetiferous
semi-pelitic band.
Kobberfjord.

are comparatively poor in garnet. From this it may be deduced that the ease of mobility of constituents is an important factor governing the growth of garnet. It is obvious that the feldspathic veins and the mobilized semi-pelites would provide regions in which constituents could move relatively freely. A few of these garnets have been slightly augened (Plate 231), indicating that they pre-date the main F_2 movements.

The presence of potash feldspar porphyroblasts, mainly in the psammitic horizons, indicates that there was probably also an intergranular migration of feldspathic constituents. These porphyroblasts are often augened by the F_2 deformation, and in thin section are seen to form a mortar texture, with highly granulated margins.

In the descriptive petrography of the migmatized psammites and semi-pelites, a phenomenon involving the occurrence of fibrolite along feldspar grain-boundaries was described (p. 48). This is believed to be an effect associated with the migmatization, and is thus considered in detail here.

It would appear that there are two ways in which these fibrolitic sillimanites might form. They could either be formed from constituents which migrated along grain boundaries, or from constituents which were derived from the neighbouring mineral grains.

The second of these hypotheses seems to be more likely for three reasons. First, fibrolite is restricted to feldspar/feldspar contacts; second, fibrolite needles often penetrate into the neighbouring grains; and third, the fibrolite is occasionally involved in myrmekitic intergrowths which have formed between potash feldspar and plagioclase.

Plate 231. Garnets, some of which are slightly augened. Kobberfjord.

A close temporal and spacial association between growth of sillimanite and the formation of migmatites has often been noted (Watson, 1948; Chinner, 1961; Pitcher, 1965). It is possible that the formation of the fibrolite in these present rocks is associated with the migmatization.

Watson (1948) has described a very similar phenomenon in pegmatitic veins in migmatized pelites and semi-pelites at Kildonan, Sutherland. There, fibrolite occurs as strings of needles around the borders of feldspars, just like those in the Kobberfjord migmatites. In addition, these strings sometimes also penetrate into and cross the feldspar grains, showing that they must have formed by replacement. Another point of similarity with the Kobberfjord migmatites is that both potash feldspar and plagioclase are replaced by fibrolite with equal readiness. However, the alteration appears to have advanced further in the Kildonan rocks. In these, feldspar often only remains as relics between the fibrolite veins, which commonly contain an abundance of quartz, whereas in the Kobberfjord rocks the fibrolite is only just beginning to form in a very sporadic fashion.

Watson considers that the fibrolite was produced from the feldspars at a late stage of the migmatization under the influence of migrating solutions; the reactions are given below:

$$\underset{\text{potash feldspar}}{K_2O.Al_2O_3.6SiO_2} = \underset{\text{fibrolite}}{Al_2O_3.SiO_2} + 5SiO_2 + K_2O$$

$$\underset{\text{alkali plagioclase}}{Na_2O.Al_2O_3.6SiO_2} = \underset{\text{fibrolite}}{Al_2O_3.SiO_2} + 5SiO_2 + Na_2O$$

It would appear that a similar cause could account for the fibrolite in the Kobberfjord migmatites. The larger, prismatic sillimanites which also occur in these rocks formed from biotite at the peak of the regional metamorphism. It is suggested that the sporadic fibrolite, however, did not form until a slightly later migmatitic stage of the metamorphism, when percolating solutions associated with the migmatization could assist the above reactions.

It is clear from the field evidence that there has been a certain amount of mobility in some of the feldspathized bands of the migmatites. However, in the absence of chemical analyses, it is not possible to be sure whether this mobility has arisen from partial anatexis of the metasediments (forming venites) or from the introduction of mobile granitic constituents (forming arterites). Härme (1965) could not always classify veins in some migmatites in southern Finland as venites or arterites. In the present case it is considered possible that both occur.

Evidence that the metasediments have in fact become mobile through partial anatexis is given by the relationships between psammites and some of the larger amphibolites. These amphibolites are veined, sometimes along joints, by quartzo-feldspathic material, consisting principally of quartz with a few sodic andesine grains. When traced towards the margins of the amphibolite, these veins are seen to merge with the migmatized feldspathic psammite of the neighbouring metasediments (Plates 232, 233, and 234). It is clear that the source of the vein material must have been the metasediments, the quartzo-feldspathic constituents of which become mobile through selective anatexis.

Mechanisms by which partial anatexis might take place have been

Plate 232. Amphibolite veined by mobilized migmatized psammites. Kobberfjord.

Plate 233. Amphibolite veined by mobilized migmatized psammites. Kobberfjord.

Plate 234. Amphibolite veined by mobilized migmatized psammites. Kobberfjord.

studied by several workers. Härme (1962, 1965) has described migmatites in Finland in which gneisses were invaded by potash-rich granite. The heat from this granite, aided by the presence of volatiles, caused the salid components (plagioclase and quartz) in the gneiss to melt by anatexis, and in places these back-veined the granite. He considers the presence of potash facilitates anatexis because this converts the bulk composition to the system Ab-Or-Q-H_2O which has a lower minimum melting temperature than the Ab-Q-H_2O system (Tuttle and Bowen, 1958). The wall-rock becomes richer in potash by diffusion from the granite; where a thermal gradient exists, K^+ migrates towards the low-temperature end (in this case the wall-rock), and Na^+ towards the other end (Orville, 1962).

Lundgren (1966) also believes that the presence of potash-bearing minerals facilitates partial melting, particularly muscovite and biotite which also provided water during their breakdown.

On the other hand, von Platen (1965) considers that anatexis can take place in situ without the addition of any extra potash. He maintains that the order in which minerals appear in the melt depends upon the Ab/An ratio rather than the temperature; if the ratio is small then a highly potassic mobilizate is formed. Experiments carried out by Winkler and von Platen (von Platen, 1965), indicate that in general anatexis begins in rocks containing quartz, plagioclase, and potassium minerals at $700 \pm 40°C$. at gas pressures of 2000 bars. Their experiments also showed that the breakdown of muscovite to potash feldspar can occur at lower temperatures at that pressure, thus indicating that the temperatures required for anatexis are attained during regional metamorphism.

The conclusion von Platen draws is that migmatites form solely by the partial anatexis of metasediments, producing granitic or granodioritic mobilizates.

King (1965) however, believes that selective melting alone is not the main factor in the genesis of migmatites, and that in any case, the order of melting of the various salic minerals is not dependent upon the An content of the rock. He considers that metamorphism at high grades, and particularly the formation of migmatites, is probably not isochemical, and that migmatites have reached a granitic composition before they become mobile.

In the Kobberfjord migmatites it is manifest that mobilization through partial anatexis has occurred. It is not possible, however, without chemical data, to determine unequivocably by which mechanism it took place.

It would appear that K^+ was migrating freely, as indicated by growth of potash feldspar porphyroblasts, but it does not follow from this that K^+ was necessarily introduced into the system as required by Härme and by King. The biotite in these rocks is beginning to alter to aluminium silicate, a reaction which results in the liberation of K^+ and H_2O. Thus it is possible that at least some of the potash feldspar could have been derived from the breakdown of mica in the presence of water, as postulated by Lundgren.

An introduction of K^+ does not seem to be essential to the formation of the Kobberfjord migmatites, and it is possible that the process was isochemical, as proposed by von Platen. In this case, the phlebitic veins would be venites, but the possibility of some of them being arterites cannot be definitely ruled out.

Summary

In summary, it can be stated that during the period of static metamorphism between the major folding episodes, the metamorphic grade increased to a maximum and then began to wane before the onset of the F_2 movements. During the upgrading, porphyroblasts of garnet, kyanite, and sillimanite developed in the schists, and the Husfjord metagabbro became amphibolized. Soon after the metamorphic peak, migmatization took place during which potash feldspar porphyroblasts and fibrolitic sillimanite developed. A second generation of garnet began to form from high-grade minerals as the metamorphic temperatures became lower.

CONTACT METAMORPHISM DUE TO EARLY DIORITES

After the Husfjord metagabbro had been amphibolized during the main static regional metamorphism, it was locally thermally metamorphosed by the early diorites. These early diorites are coarse-grained sheets which have been emplaced into joints in the metagabbro.

A fine-grained granoblastic hornfelsic texture is developed in the metagabbro next to the diorites. The principal minerals are hypersthene, clinopyroxene, ore, and plagioclase, and these form small rounded grains. This texture and mineralogy are superimposed upon those produced during the static regional metamorphism. This and the fact that the diorites themselves are not amphibolized indicate that the diorites were emplaced after the peak of the regional metamorphism.

In some p laces, later biotite occurs in these hornfelses, and this must have formed at a later stage of the regional metamorphism.

F_2 SYN-TECTONIC REGIONAL METAMORPHISM

The onset of the F_2 movements brought about a change from static metamorphism to syn-tectonic metamorphism. The regional metamorphic event was still continuing, but its intensity was decreasing during the F_2 deformation episode. F_2 syn-tectonic fabrics are rarely preserved in these rocks, although there are many examples of the effects of the deformation upon minerals.

Although kyanite generally has a random orientation, in a few cases, it shows a slight tendency to have a preferred orientation. From this it may be deduced that kyanite growth continued until the beginning of the F_2 movements. A similar conclusion was reached by Roberts (1965) who found kyanite in the Langstrand area containing sigmoidal inclusion trails, indicating its syn-tectonic growth.

Similarly, Roberts (op.cit.) ascribes sigmoidal inclusion trails in garnets to their syn-F_2 growth. In the rocks of the Husfjord area, no examples of undoubted syn-tectonic garnet have been seen. However, the fact that garnet sometimes forms at the expense of kyanite, suggests that garnet growth probably also persisted into the F_2 deformation.

Effects of this deformation are seen in the augening of migmatitic potash feldspar porphyroblasts, and in the augening and rotation of earlier garnets and kyanites.

The main igneous complex was emplaced during the F_2 movements as the metamorphic temperature was decreasing. The earliest member of this complex, the Kobberfjord norite, was intruded before the temperature had dropped beneath that of the garnet isograd (see Fig. 42). This orthonorite has been contaminated by metasediments, and garnet has grown in the more aluminium-rich parts. The temperature had probably dropped beneath the garnet isograd by the time the main suite of diorites was emplaced. However, if the grade had only been lowered to the staurolite-quartz subfacies of the almandine-amphibolite facies, the presence of potash feldspar in the Havnefjord diorite and the coarse-grained pyroxene-mica-diorites would preclude the growth of garnet (Fyfe, Turner, and Verhoogen, 1958, p. 229). The presence of garnet in the later quartz-diorites is considered to be a result of slight contamination (see p. 269).

CONTACT METAMORPHISMS DUE TO MAIN IGNEOUS COMPLEX

During the course of the F_2 folding episode, the main igneous complex was emplaced (see Fig. 42). Many members of this complex had thermal effects either upon the metasediments of the country rocks and enclosed rafts, or upon earlier members of the complex. Since these thermal metamorphisms took place synchronously with regional metamorphism, study of the mineralogy and texture of the hornfelses can provide valuable information concerning the progress of the regional metamorphic event during the emplacement of the complex.

Kobberfjord Norite

The Kobberfjord norite causes a narrow thermal aureole to form within the migmatites of the country rocks. In this aureole, kyanite recrystallizes to its polymorph sillimanite, which at the contact becomes large. The significance of the absence of andalusite in the contact aureole is discussed on p. 304. In the contact zone, these large sillimanites become severely strained and are beginning to become fragmented. This deformation is presumably due to differential movement along the contact during F_2.

In the thermal aureole, garnet forms at the expense of biotite, kyanite, and sillimanite, indicating that garnet growth continued after the emplacement of the Kobberfjord norite. This confirms the post-intrusion growth of garnet deduced from the evidence provided by the norite (p.230).

Havnefjord Diorite

The envelope to the Havnefjord diorite lies within the Husfjord metagabbro, and it has thermally metamorphosed the latter up to the hornblende-hornfels facies.

The relict subophitic texture of the metagabbro is frequently preserved, but the superimposition of a thermal metamorphism upon the regional metamorphism has caused extensive amphibolization. At the contact, the amphibolization of the pyroxenes has been virtually complete in places, and the hornblende is so full of exsolved quartz blebs that they lose their coherency and form aggregates of small grains. Away from the contact a little epidote occurs, and there are a few oligoclase porphyroblasts.

Relict troctolites occurring as rafts in the Havnefjord diorite have been extensively amphibolized by the diorite.

Many of the metasedimentary rafts enclosed by the Havnefjord diorite have been hornfelsed at their margins. This is particularly well shown by the pelitic and semi-pelitic schist rafts.

In these hornfelsed schists, the F_1 schistosity has been obliterated in many cases, and in other cases remains as a relict texture only discernible by the presence of biotite-rich bands. The hornfelsic texture is typically formed by polygonal grains of potash feldspar and quartz, with biotite laths forming a decussate pattern. Ore minerals often form ragged-edged laths, and these are commonly also in a decussate arrangement. Occasionally potash feldspar porphyroblasts grow in the pelitic hornfelses, and corundum is a common mineral in the more aluminous pelites.

Garnet is unstable in these thermal metamorphic conditions, and breaks down to fibrolite and biotite, or fibrolite and ore. Various stages in the alteration have been observed and in the most extreme examples, the garnet is completely pseudomorphed. A similar phenomenon has been described by Pitcher and Read (1963) in the thermal aureole of the Fanad granodiorite, but in their example the aluminium silicate was andalusite instead of sillimanite, and cordierite was sometimes present. The implications of this difference are discussed on p. 304. There is no new garnet being regenerated in or around these pseudomorphs, and this indicates that the grade of the regional metamorphic event had probably dropped beneath the garnet isograd.

At Hasvik, in S.W. Sørøy, a gabbro has been emplaced synchronous with the peak of the regional metamorphism between F_1 and F_2 (Sturt, personal communication). Garnets in the schists within the contact aureole have been replaced by sillimanite, spinel, hypersthene, and ore, but in this case new garnet has reformed from these breakdown products. Sturt considers that this indicates that the regional metamorphic grade was still at the almandine-amphibolite facies when the thermal aureole cooled down. However, later dioritic rocks, which are probably comparable to those in the Husfjord area, have been intruded into the contact aureole of the Hasvik gabbro. These cause the breakdown of both generations of garnet, and there is no regrowth of later garnet.

In the calc-silicate schist rafts, a granular hornfelsic texture, principally of diopside and plagioclase is developed, and this obliterates the F_1 schistosity.

The granoblastic texture formed in the basic schist hornfelses also destroys the F_1 schistosity. The chief minerals are hypersthene, clino-pyroxene , ore, and plagioclase; occasionally, diffuse biotites are seen altering to pyroxene.

The basic schist hornfelses in one raft develop an ophitoblastic texture (see p.193), indicating that they may have come near to being mobilized. In another case, they have actually been mobilized. In another case, they have actually been mobilized, and the more basic bands have been disrupted by the mobilized quartzo-feldspathic parts. It appears that the plagioclase in these mobilized hornfelses is becoming more basic, exsolving silica as vermicular quartz inclusions (see p. 195).

Late Diorites

The coarse pyroxene-mica-diorite sheets thermally metamorphose the neighbouring Havnefjord diorite. In the latter, a fine-grained granoblastic pyroxene hornfels consisting mainly of hypersthene, clinopyroxene and plagioclase is developed.

Hornfelsic textures develop in the metasedimentary xenoliths within the pyroxene-mica-diorites. Decussate biotites and polygonal grains of hair-perthite characterize the pelitic hornfelses, although porphyroblasts of hair-perthite also occasionally occur. Corundum and green spinel are common in the silica-deficient pelites. Pyroxene hornfelses develop in the basic schist hornfelses, and hypersthene grains are sometimes fairly large and poikiloblastic. It appears that some of the orthopyroxene hornfelses are being made over into the pyroxene-mica-diorite (see p.211).

The garnet-poor quartz-diorites contain occasional metasedimentary xenoliths which have become hornfelsed. In the semi-pelitic xenoliths a hornfelsic texture is seen to have formed during F_1. Biotite laths form a decussate pattern between polygonal grains of quartz and andesine. In the calc-silicate schist hornfelses, tremolite-actinolite has formed from biotite. These amphiboles are randomly orientated, and are frequently poikiloblastic.

POST-F_2 (STATIC) REGIONAL METAMORPHISM

While the main igneous complex was being emplaced, the regional metamorphic event was continuing to wane (see Fig. 42), and these members of the complex show comparatively low-grade metamorphism. Evidence of this metamorphism is also present in the Husfjord metagabbro, and some of the country rocks and metasedimentary rafts.

Country Rocks

The effects of this late stage of the regional metamorphism are well demonstrated in the calc-silicate-schists, in which the granular hornfelsic texture produced by the thermal effect of the Husfjord metagabbro has been overprinted by a regional metamorphic texture and mineralogy. Poikiloblastic actinolite contains inclusions of small rounded diopside grains, and both diopside and biotite are altering to actinolite; there is also a little clinozoisite.

In the kyanite-garnet-schists both kyanite and sillimanite are altering to large poikiloblastic muscovites which often grow across the schistosity, indicating its growth in a static environment. This alteration appears to be more common in the neighbourhood of potash feldspar, which presumably provided potassium to form the muscovite:

$$\underset{\text{kyanite, sillimanite}}{2[Al_2O_3.SiO_2]} + \underset{\text{potash feldspar}}{K_2O.Al_2O_3.6SiO_2} + 2H_2O \rightarrow$$

$$\underset{\text{muscovite}}{K_2O.3Al_2O_3.2H_2O.6SiO_2} + \underset{\text{quartz}}{2SiO_2}$$

In this reaction an excess of silica is produced, and this probably accounts for the numerous blebs of quartz which are commonly enclosed by the muscovite.

In many of the psammites and semi-pelites garnets are beginning to alter, especially along cracks, to fine-grained secondary biotite, and calcite occasionally fills cavities. The formation of myrmekite around the migmatitic potash feldspar porphyroblasts and the sericitization of feldspars probably took place at this stage.

Some of the metasedimentary rafts enclosed by the Husfjord metagabbro or the Havnefjord diorite contain randomly-orientated poikiloblastic biotites which overgrow the hornfelsic texture. These biotites presumably grew during this waning stage of the regional metamorphism.

Husfjord Metagabbro

The effects of this late-stage metamorphism are not ubiquitous throughout the metagabbro. They are evident in the actinolitic facies, for it appears that the formation of actinolite and the alteration of some of the plagioclase to epidote in this facies is a result of this metamorphism.

This facies is not uniformly developed throughout the body, and the late-stage metamorphism has not been markedly recorded by most of the metagabbro. However, late biotites commonly occur, and these are attributed to this metamorphism.

Diorites

In the early diorite bands that have been emplaced into the Husfjord metagabbro, some of the hypersthene grains are altering to biotite where contact with potash feldspar is made. The biotites generally contain vermicular inclusions of quartz, and this may be explained by the presence of excess silica from the reaction:

$$6\,[\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2}] + \underset{\text{potash feldspar}}{K_2O.Al_2O_3.6SiO_2} + 2H_2O$$

$$\rightarrow \underset{\text{biotite}}{K_2O.6(Mg.Fe)O.2H_2O.Al_2O_3.6SiO_2} + \underset{\text{quartz}}{6SiO_2}$$

When the potash feldspar is perthitic, this reaction appears to be connected with the formation of myrmekite, indicated by the intimate association of myrmekite and vermicular biotite. The reaction may be as follows:

$$6\,[\underset{\text{hypersthene}}{(Mg.Fe)O.SiO_2}] + 2\,[\underset{\text{perthitic potash feldspar}}{(\tfrac{1}{2}K.\tfrac{1}{2}Na.)_2O.Al_2O_3.6SiO_2}] + 2H_2O \rightarrow$$

$$\underset{\text{biotite}}{K_2O.6(Mg.Fe)O.2H_2O.Al_2O_3.6SiO_2} + \underbrace{\underset{\text{sodic plagioclase}}{Na_2O.Al_2O_3.6SiO_2} + \underset{\text{quartz}}{6SiO_2}}_{\text{myrmekite}}$$

In both the Havnefjord diorite and the late pyroxene-mica-diorites, some of the hypersthenes alter to biotite which often contain vermicular quartz inclusions. It would appear that in these cases the reaction is similar to that given above.

Vatna Gabbro and Minor Intrusions

It appears that the Vatna gabbro has not suffered a high grade of metamorphism, since olivines are often scarcely altered. However, augite grains are sometimes fringed by a little biotite to which it is altering.

The perthosites show early stages in the development of swapped rims at the boundaries of the perthite grains, and this is probably a late metamorphic effect.

In the late basic dykes, augite is almost invariably altering to brownish hornblende, and occasionally to biotite. Some of the hornblende is also altering to biotite, and many of the plagioclase phenocrysts of the porphyritic dykes axhibit an irregular zoning. The finer-grained dykes are generally more altered than the coarser-grained ones, and this is presumably because the former have a larger overall surface area of grain boundaries, enabling migrating constituent to have a greater mobility.

The hornblende in these basic rocks, particularly those containing olivine, is brownish in colour, and this is probably mainly a compositional effect. Deer (1938) and Leake (1965) have indicated that the colour of amphiboles becomes browner with increase in titanium content. Winchell (1924) considers that the state of oxidation of the titanium may have an influence, the colour becoming browner with increase in oxygen content. The hornblendes in the present case have not been analysed, but it is possible that compositional control was an important factor in the determination of their colour.

The late basic dykes appear to have altered to a slightly greater extent than their host, the Vatna gabbro. A similar feature occurs in the minor intrusions in the Hasvik gabbro in S.W. Sørøy (Sturt, personal communication). However, the dykes cannot have suffered a greater metamorphism than their host, and a possible explanation is that the dykes provided planar discontinuities in the massive gabbro along which alteration might have occurred more readily.

Metasomatic Veins

The metasomatic actinolite-bearing veins that cut the Husfjord metagabbro, the Havnefjord diorite, the Vatna gabbro, and the late basic dykes are all very similar to one another, and they are probably of the same generation. In some cases it can be demonstrated that they post-date the formation of the main F_2 shears, and they were probably formed during a late hydrothermal stage of the regional metamorphism.

DISCUSSION

The inter-relationships between regional metamorphic and contact metamorphic effects as described above show that the emplacement of the Husfjord igneous complex took place synchronous with a long regional metamorphic event during the Caledonian orogeny (see Fig. 42). Thus it is possible to study the effects of thermal metamorphism imposed upon rocks in an orogenic environment.

The most striking feature of the mineral assemblages in the metasediments which have been hornfelsed is the complete absence of andalusite and cordierite in the pelitic rocks.

In the pelites and semi-pelites within the narrow aureole of the Kobberfjord norite, kyanite alters to sillimanite, but andalusite never occurs. Andalusite is the low-pressure polymorph of Al_2SiO_5, and the most recent determination of the triple point in the Al_2SiO_5 system where andalusite, kyanite, and sillimanite coexist stably is 6.5 ± 0.5kb., and $595 \pm 10^{\circ}$C. (Althaus, 1967; Althaus considers that these determinations are more accurate than previous ones because a hydrothermal apparatus was used, giving true hydrostatic pressures). Thus the maximum pressure at which andalusite is stable is about 6.5kb., which is equivalent to a depth in the earth's crust of about 23 km. (Schreyer and Yoder, 1964).

In the Husfjord area, thermal metamorphism occurred synchronous with a fairly high grade of regional metamorphism, suggesting that the emplacement of the igneous bodies took place at considerable depths. Thus it is probable that the high pressures obtaining at great depths precluded the formation of andalusite in the hornfelses. Instead, the kyanite probably inverted directly to sillimanite without passing through the andalusite field because the pressures were higher than that of the triple point.

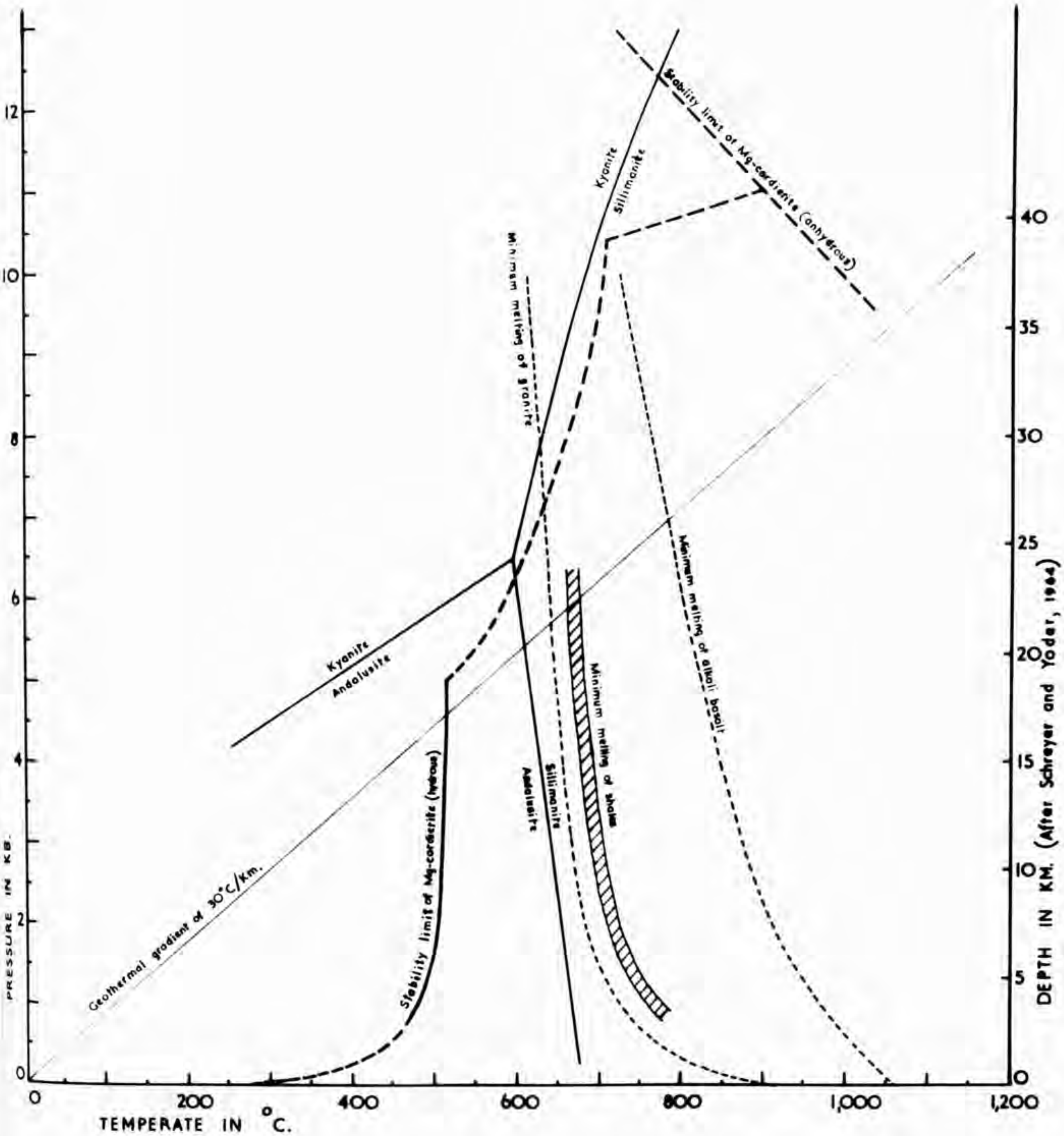
The pelitic and semi-pelitic hornfels rafts enclosed by the Havnefjord diorite contain regional metamorphic garnets, which as a result of the hornfelsing have broken down to knots of fibrolite, biotite, and ore. The breakdown of almandine garnet under thermal metamorphism has been reported by many workers (Tilley, 1924, 1926; Harker, 1939; Miyashiro, 1953; Fyfe, Turner, and Verhoogen, 1958; Leake and Skirrow, 1960; Chinner, 1962). In each case the principal mineral which pseudomorphed the garnet was cordierite, sometimes accompanied by magnetite. The general opinion of these writers is that the major factors determining the stability relationships of garnet and cordierite are pressure and temperature, although Chinner (op. cit.) considers that the bulk composition of the rock provides a more important control. The absence of cordierite from the hornfelses in the Havnefjord diorite presumably indicates that cordierite would not have been stable in the environment in which the diorite was emplaced.

Fig. 43 shows the stability field of Mg-cordierite as it has been determined experimentally by Schreyer and Yoder (1964). (The stability fields of andalusite, kyanite, and sillimanite as determined by Althaus (1967) have been added to Fig. 43). It can be seen from Fig. 43 that Mg-cordierite is confined to a more limited pressure-temperature stability field in the presence of excess water vapour than under anhydrous conditions. However, according to Schreyer and Yoder (1960), Mg-cordierite breaks down at high confining pressures even in the absence of water to denser anhydrous high temperature minerals such as sillimanite. Thus it is deduced by Schreyer and Yoder (1964) that Mg-cordierite is only stable at relatively low pressures, and that anhydrous conditions are more favourable to stability than hydrous ones.

In the orogenic environment in which the Havnefjord diorite was emplaced, hydrostatic pressures must have been considerable, and from the evidence provided by the lack of andalusite it is deduced that they were probably in excess of 6.5kb. The fact that kyanite is altering to sillimanite in the hornfelses indicates that the conditions obtaining were within the sillimanite stability field, and therefore, would probably also be just within the stability field of Mg-cordierite (see Fig. 43). However, cordierite is absent from these hornfelses, and a possible reason for this is considered below.

Natural cordierites are not pure Mg-cordierites but contain variable amounts of Fe-cordierite in solid solution (Leake, 1960), and the effect

STABILITY FIELDS OF Mg-CORDIERITE
AND THE Al_2SiO_5 POLYMORPHS



Mg-cordierite stability curves after Schreyer and Yoder (1964). Al_2SiO_5 polymorphs stability triple point after Althaus (1967). Minimum melting curves of granite and of alkali basalt after Yoder and Tilley (1962). Geothermal gradient and minimum melting of shales after Wyllie and Tuttle (1960).

FIG. 43

of the iron must be taken into account. Preliminary studies on pure Fe-cordierite by Schreyer and Yoder (1959) have shown that Fe-cordierite appears to be stable at similar temperatures but at lower water pressures than Mg-cordierite. Schairer and Yagi (1952) have shown that at atmospheric pressure the melting point of Fe-cordierite is about 250°C. lower than that of Mg-cordierite.

Thus it is deduced by Schreyer and Yoder (1964) that the pressure-temperature field of pure Mg-cordierite is probably the maximum stability field of natural cordierite. The presence of a large percentage of iron in the cordierite may well reduce the stability field significantly to the lower pressure regions.

Leake (1960) has made a compilation of natural cordierite analyses, and those occurring in pelitic hornfelses apparently similar to the hornfelses under present discussion, contain considerable amounts of the Fe-cordierite molecule (see Table 22). Read (1929) has reported an iron-rich cordierite in a hornfels from Arnage, Aberdeen. Thus it seems reasonable to assume that if garnet had formed in the present hornfelses it might have contained more than 50% of the Fe-cordierite molecule. The fact that cordierite is absent from these hornfelses indicates that the pressure-temperature conditions prevailing could not have fallen within the stability field of the potential Fe-rich cordierite. Thus it would appear that the stability field of Fe-cordierite may be restricted to pressures of less than about 6.5kb.

It is considered therefore that the breakdown of garnet to sillimanite, ore, and biotite instead of to cordierite in the pelitic hornfelses in

TABLE 22

IRON CONTENT OF CORDIERITES FROM HORNFELSES

	L32	L40	L41	L43	L45	L50
FeO	7.26	8.55	8.61	7 .99	8.55	9.51
MgO	8.65	7.81	7.56	6.71	7.81	6.99
$\frac{FeO}{FeO+MgO}$ x 100	45.64	52.26	53.25	54.35	52.26	58.48

	L55	L59	Bw1	Bw2	Bw3	R
Fe	10.09	12.39	9.64	10.75	11.12	14.13
MgO	6.89	5.91	5.35	5.60	5.57	5.77
$\frac{FeO}{FeO+MgO}$ x 100	59.40	67.70	64.45	65.70	66.65	71.01

Total iron has been expressed as FeO.

L32 - L59 : from the compilation of cordierite analyses by Leake (1960).

Bw1 - Bw3 : from Best and Weiss (1964).

R : cordierite from Arnage hornfels, calculated from analysis of cordierite-plagioclase residue (Read, 1929).

the Havnefjord diorite is due mainly to the fact that the prevailing pressures were too high for the stable existence of the cordierite that might have formed. The fact that the emplacement of the Havnefjord diorite was synchronous with regional metamorphism not only means that high pressures would obtain, but also that there would probably be an abundance of water. As shown by Schreyer and Yoder (1964) the stability field of cordierite is markedly decreased under hydrous conditions (see Fig. 43), and the presence of water was probably a contributory factor in the prevention of cordierite formation.

It is of interest to note that cordierite is often developed in the hornfelses associated with some of the Newer Plutonic Gabbros of the N.E. Grampians, e.g. Arnage (Read, 1923), Haddo House (Read, 1935), Belhelvie (Stewart, 1947), and Insch (Sadashivaiah, 1950, 1954). In each of these cases, the country rocks into which they have been emplaced are cordierite-bearing or cordierite-andalusite-bearing schists and gneisses. The presence of cordierite and andalusite in the country rocks indicates that they have not been regionally metamorphosed under conditions of high pressure.

It has been demonstrated above, however, that the metamorphic conditions prevailing in Sørøy involved high pressures during regional metamorphism, and that this precluded the growth of the low-pressure minerals andalusite and cordierite, both in the country rocks and in the hornfelses. These minerals are also absent from the aureole of the Hasvik gabbro, S.W. Sørøy (Sturt, personal communication). This gabbro was emplaced synchronous with the peak of the regional metamorphism, and the breakdown of garnet to sillimanite, spinel, hypersthene, and ore instead of to cordierite is thought by Sturt to be due to high hydrostatic pressure as explained above.

Cordierite develops in the hornfelses associated with the Cashel - Lough Wheelaun intrusion, County Galway (Leake and Skirrow, 1960), and the country rocks of this intrusion have previously been regionally metamorphosed to the sillimanite-almandine facies (Leake, 1958). This intrusion thus resembles the Grampian examples in that cordierite develops in the hornfelses, but regional conditions seem to have been comparable to those of Sørøy since cordierite and andalusite are absent from the regional schists. It would appear that conditions in Galway might represent depths which were intermediate between those of Sørøy and the Grampians.

Thus the plutonic intrusions of Sørøy appear to be examples of synmetamorphic emplacement at pressures probably in excess of 6.5kb. Although tectonic overpressures may have been present during regional metamorphism, they were probably not active at the time of emplacement of the plutonic intrusions. In any case, Rutland (1965) considers that the effects of tectonic overpressures are insignificant compared with hydrostatic depth-pressures. Thus, pressures of 6.5kb. probably represent depths of about 23 km. (Schreyer and Yoder, 1964), and this implies that the emplacement of the main igneous complex occurred at depths of this order.

PART V

SUMMARY OF PLUTONIC HISTORY

SUMMARY OF PLUTONIC HISTORY

In the Husfjord area of Sørøy there has been a close relationship between the sequences of igneous, metamorphic, and structural events. This is shown in Fig. 42, and a summary of these events is presented below.

(1) Deposition of Eocambrian sediments, presumably on the south-eastern shelf of the Caledonian geosyncline.

(2) Emplacement of a number of small pre-F_1 basic sheets.

(3) First episode of folding, F_1. This was long and complex and was characterized by recumbent isoclinal folds overturned towards the east and south-east. The regional metamorphism increased in intensity during the F_1 movements, with the production of F_1 schistosities.

(4) Emplacement of Husfjord gabbro, probably during the waning stages of F_1. Emplacement of troctolites.

(5) Emplacement of small post-F_1 basic sheets.

(6) Peak of the regional metamorphism (upper almandine-amphibolite facies) in static period between F1 and F_2, with the formation of kyanite, sillimanite, and garnet static porphyroblasts.

(7) Migmatization and feldspathization associated with highest grades of the regional metamorphism.

(8) Emplacement of early diorites into the Husfjord metagabbro.

(9) Second episode of folding, F_2. This was also long and complex, the main phase being characterized by orthorhombic and monoclinic fold styles.

The regional metamorphism waned during the F_2 movements.

(10) Emplacement of Kobberfjord norite during F_2, before the intensity of the regional metamorphism had dropped beneath the garnet isograd.

(11) Emplacement of Havnefjord diorite at a later stage of the F_2 movements.

(12) Emplacement of coarse-grained pyroxene-mica-diorites, also during the F_2 movements.

(13) Emplacement of quartz diorites.

(The relative age of the Komagfjord diorite is not known).

(14) Emplacement of Slåtten gabbros and Vatna gabbro.

(15) Emplacement of perthosites.

(16) Emplacement of late basic dykes and nepheline-syenite pegmatites.

(17) Last retrogressive stages of regional metamorphism under static conditions, during which biotite and muscovite were formed. Emplacement of narrow actinolitic metasomatic veins.

(18) Late-stage F_2 brittle movements, which under the cooler conditions produced small shears and mylonites.

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APPENDIX

APPENDIX

NOTE ON TECHNIQUES OF MACRO POINT COUNTING

The early diorite bands and the late diorite sheets in the Husfjord area are coarse-grained rocks, with feldspar crystals up to about 2 cm. in length. This means that an accurate modal analysis is difficult to obtain by normal microscopic point counting methods.

It is generally considered that for an accurate modal analysis, a minimum area of 100 times the area of the largest grain in the rock provides a representative sample of a structureless plutonic rock (Larsen and Miller, 1935). Thus a single thin section of these coarse-grained diorites would be most unlikely to show the true proportions in which the various minerals in the rock occur. The coarser the grain-size, the greater the number of thin-sections that need to be counted (Chayes, 1956).

Alternative methods are to analyse a large number of thin sections cut statistically from each hand sample, or to point count large slabs which have been cut from the hand specimen and polished. It was decided to use the latter method for the coarse-grained diorites.

A device for point counting large slabs is not manufactured commercially, but several workers have devised their own apparatus for macro point counting, with varying degrees of success.

Jackson and Ross (1956) used a method which entailed covering an etched and stained polished slab with a Zip-A-Tone dot pattern, and counting the minerals under the dots with a magnifier. They list three limitations to this method: first, it is only applicable to minerals which are identifiable in slabs; second, the grains must not be less than 2 - 3 m.m. in size; third, small grains are concealed under the dots. This third drawback would seem to the present writer to be a rather important one to overcome. Jackson and Ross consider that 1,000 points is the minimum number that should be counted for reliable results.

Fitch (1959) suggested a variation on this method in which a graticule is prepared with a co-ordinate reference grid and cemented onto the stained surface of a polished slab. The minerals under the intersection of the grid lines are counted with a binocular microscope, the slabs being moved under the microscope by hand on the microscope stage.

A more sophisticated technique was devised by Emerson (1958), in which the stained slab is placed on a platform which is moved mechanically by turning a screw. This method is also better than previously described ones because no reference grid is placed on the specimen itself. The point of reference is provided by cross-hairs engraved on a glass plate, upon which a magnifying glass lies, and under which the specimen moves.

Plafker (1956) fitted a microscopic point counter mechanical stage to the adapted stage of a binocular microscope. Polished slabs were placed on the mechanical stage and point counted in the normal way.

Smithson (1963) has developed a machine which is specifically designed for point counting polished slabs. It consists of a stage which can be moved laterally in two directions at right-angles to one another at set intervals by means of knurled knobs. The polished slab is set on modelling clay upon the stage. A binocular microscope bearing cross-wires is erected over the stage, and minerals are counted in the normal manner. The maximum area that can be covered by this apparatus is about 30 sq. cm. Smithson stained the slab for potash feldspar only, and then lacquered it. The potash feldspar showed up yellow, plagioclase frosty white, quartz vitreous clear, and biotite yellowish-green (the rock being a granite). Accessory minerals could not be distinguished.

Barratt and Parslow (1966) have made a piece of apparatus resembling in principle that of Smithson, but having a rather different design. They found that staining the polished slabs was troublesome, and decided to count unstained slabs.

In each of the methods that have been mentioned so far, it is possible to distinguish between potash feldspar, plagioclase, and quartz. However, it is not possible to distinguish confidently between various mafic minerals or to identify accessory minerals. Clearly, the point counting of slabs alone is not satisfactory in determining a full modal analysis of a rock. A method combining the use of slabs and thin sections is needed.

Nesbitt (1964) has used a method in which the phenocrysts in the stained slab of a porphyritic rock are outlined in ink and traced off onto tracing paper. This paper is then placed over graph paper, and the points under the intersections of the grid are counted. In this way potash feldspar phenocrysts, plagioclase phenocrysts, and matrix are counted. Thin sections of the rock are then counted, omitting the phenocrysts. These latter results are recalculated to a total equal to the matrix value in the macro point counting. The two sets of data are then combined to form a complete analysis.

In the technique used by the present writer to analyse the coarse-grained diorites, the apparatus was based on that of Smithson, and the theoretical approach was similar to that of Nesbitt.

The apparatus for making the modal analysis of polished slabs was made in the Science Workshop at Bedford College, London. Plates 235 and 236 show the equipment set up for modal analysis.

The machine consists of a circular brass plate 10 cm. in diameter mounted on a screw 15 cm. in length, which when turned, transports the plate in a 'N-S' direction. This screw, together with the plate, is in turn mounted on another screw, also 15 cm. in length, which is capable of transporting the plate, together with the N-S screw in an 'E-W' direction. The screws are turned by means of knurled knobs, and a ball-and-socket arrangement allows the knobs to be turned in equally-spaced 'clicks'. At each 'click' the brass plate moves laterally $\frac{1}{2}$ mm. The minimum spacing of points is therefore $\frac{1}{2}$ mm., which is small enough for coarse-grained rocks. Larger intervals may be used, simply by making say 2 or 4 jumps between consecutive points counted.

The area that can be covered by movement along these two screws is approximately 100 sq.cm. However, the device can be slid along a groove in the base of the apparatus for a distance of about 30 cm., giving a maximum area of approximately 300 sq.cm. if required.

A binocular microscope, with cross-wires fitted, was mounted over

Plate 235. Macro point counting equipment.

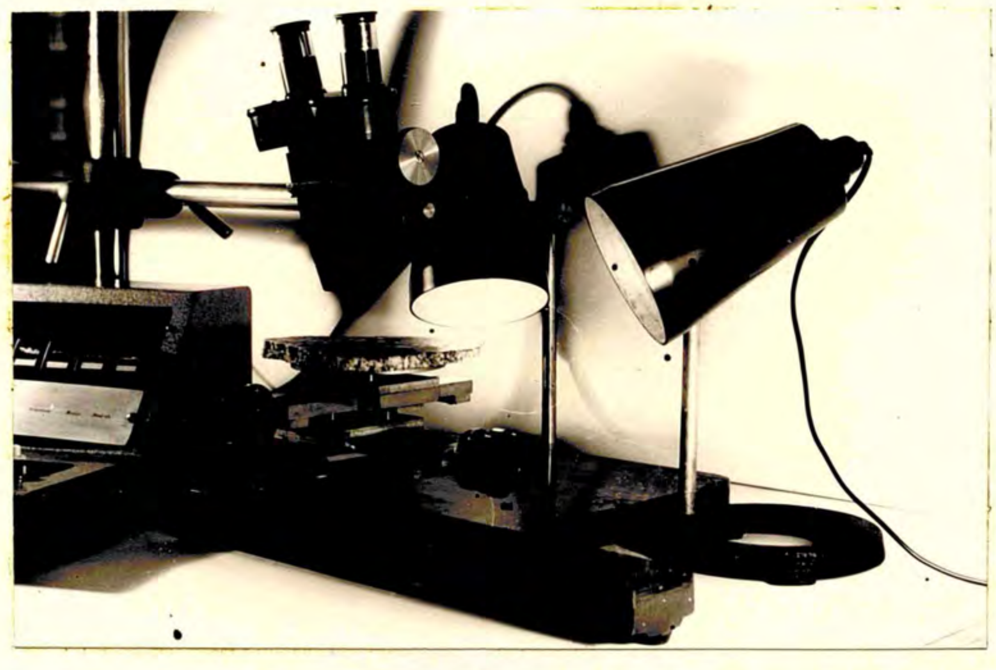


Plate 236. Macro point counting equipment.

the apparatus, and an ordinary point counting tabulator was used to record the points.

The slabs were placed on the brass platform, horizontality being ensured by setting them in plasticine and testing with a spirit level. They were then illuminated by two lamps.

The preparation of the slabs was as follows. After the slabs had been cut with a diamond saw, their porosity was eliminated. This was done by heating them in a Bunsen flame and applying Lakeside, which filled the pores previously occupied by air. The slabs were then ground on steel laps with fine carborundum.

They were then stained for potash feldspar and plagioclase by a method based on that of Bailey and Stevens (1960). Their surfaces were etched by immersing them in 40% HF for about 5 minutes. On removal from the HF they were washed gently in water and then soaked well in 5% BaCl solution for 20 - 30 seconds. After careful washing in distilled water they were immersed in saturated sodium cobaltinitrite solution until the yellow staining of the potash feldspar was of a satisfactory hue. They were then rinsed carefully, and the surfaces covered with potassium rhodizonate (0.05 gm. in 20 ml. distilled water), until the plagioclase was stained a brick red. They were then given a final wash, and when dry, were sprayed with transparent lacquer to preserve the surfaces. The washings were never done in running water, but in a sink full of slowly changing water, in order to preserve the delicate structures on the etched surfaces of the slabs.

The quality of the stain was found to vary from rock to rock.

The quartz-diorites stained well for both potash feldspar (bright yellow) and plagioclase (brick red), but the early diorites and the late pyroxene-mica-diorites were difficult to stain. The potash feldspar stain was pale, and the plagioclase stain was not very satisfactory. In spite of numerous re-grindings and re-etchings for varying lengths of time, no success was achieved with the plagioclase stain for these rocks In these cases it was decided to stain only for potash feldspar, as in the method used by Smithson (1963). To do this, the steps in the staining procedure involving BaCl solution and potassium rhodizonate solution were omitted. After applying lacquer to these slabs, it was found that the plagioclase showed up frosty white.

The slabs themselves were irregular in shape, since trimming to a rectilinear form causes the loss of surface area. The area of the largest slab was about 90 sq.cm. Normally, the slab was counted at intervals of 1 mm. (i.e. two jumps of the mechanism). Two or three slabs were cut from each specimen, and the results of the counts on each slab were averaged to give a value for the specimen as a whole.

As mentioned previously, macro point counting methods alone cannot give a complete modal analysis of a rock containing numerous different mafic minerals and accessory minerals. On the other hand, micro point counting methods alone cannot give a satisfactory modal analysis of a coarse-grained rock unless a very large number of thin sections are analysed. Therefore, a combination of these methods was employed.

The procedure adopted by Nesbitt (1964) is only satisfactory if the difference in size between feldspar phenocrysts and matrix feldspars is marked. In the rocks under present discussion, the feldspar phenocrysts are of varying sizes and grade into the more coarse-grained of the matrix feldspars. Thus the definition of phenocrysts is ambiguous and Nesbitt's method cannot be used.

It was found that the staining of the slabs was delicate enough to indicate the textures of perthite and antiperthite, thus making it possible to make a complete analysis for the feldspars by the macro point counting method. From the counting of the slabs the percentages of potash feldspar, plagioclase, and 'remainder' were obtained.

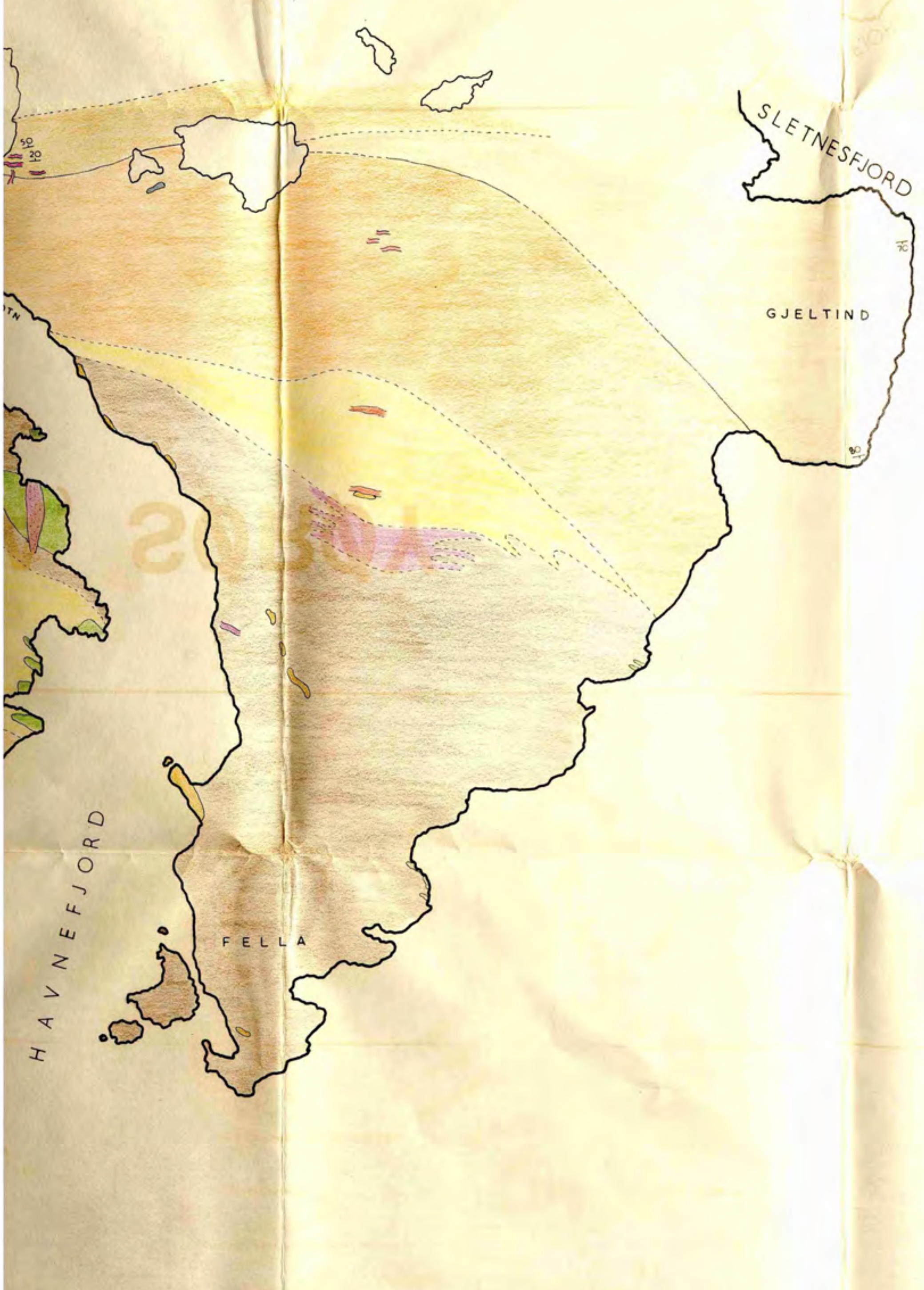
Thin sections, which were made from the same specimens from which the slabs were cut, were then analysed by the normal micro point counting technique. Generally, two or three sections from each sample were counted, the results being averaged. In the micro analysis, potash feldspar and plagioclase were passed over without counting them, and only the minerals constituting the 'remainder' of the macro analysis were counted. In this way, the proportion of each mineral in the 'remainder' was determined as a percentage of the 'remainder'. These values were then recalculated to make their sum equal to the value of the 'remainder' as determined by the macro method. Thus a complete, detailed modal analysis of the rock was obtained.

Each of the coarse diorites was analysed in this way. In addition, some completely microscopic modal analyses of each rock were carried out since the number of large specimens suitable for slab-making was limited.

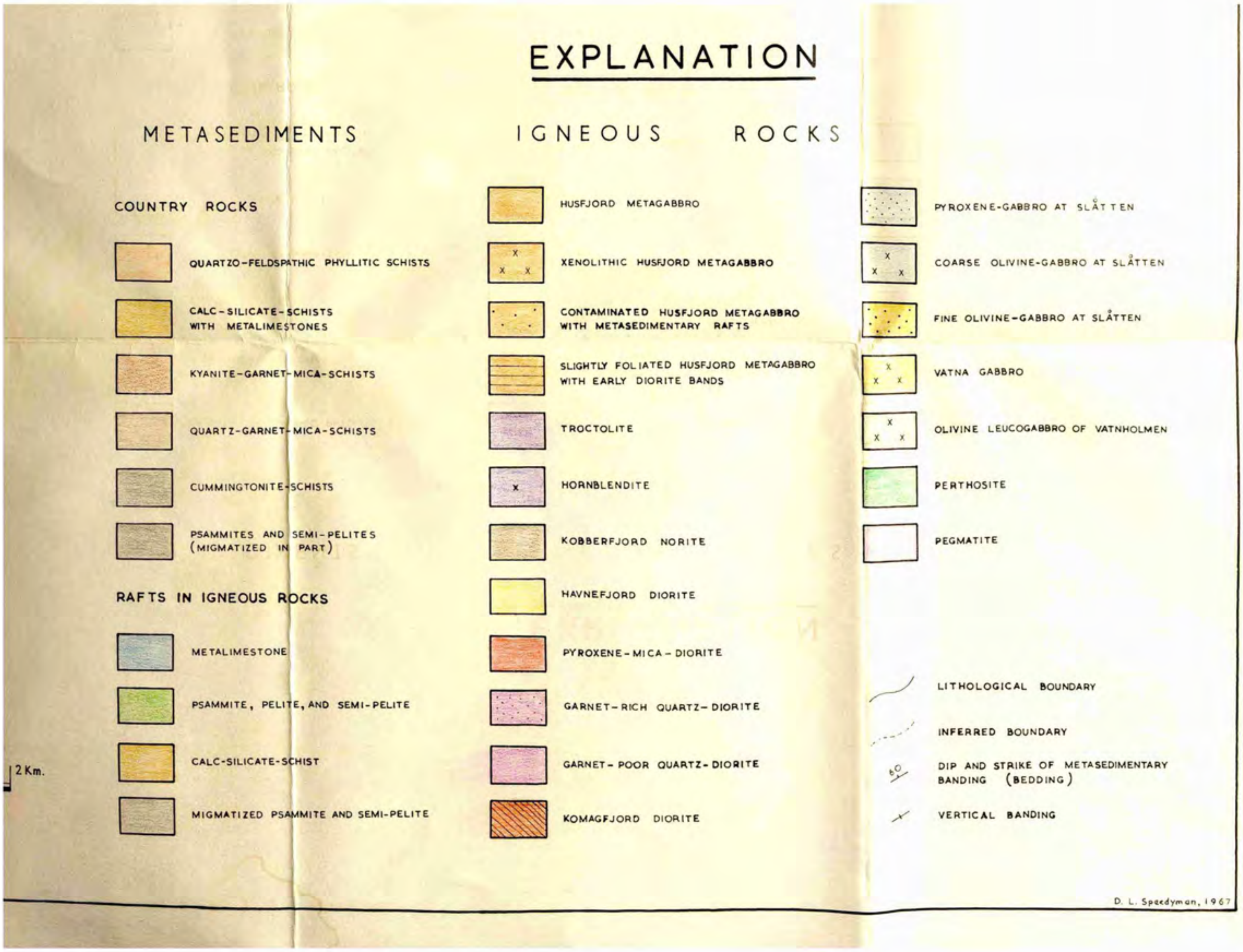
A check on the accuracy of the analyses was made by calculating the percentages after 1,000 points had been counted, and then again after 2,000. If the results differed significantly, then a further 1,000 points were counted and the percentages calculated. This process was repeated until the discrepancies were in the order of 1 - 2 per cent. In most cases 4,000 - 5,000 points were counted, but in the more heterogeneous specimens, 6,000 or 7,000 points were counted.

The writer considers that the method of combined micro and macro point counting can be very accurate for certain coarse-grained rocks with fairly simple textures. The accuracy decreases as the texture becomes more complex because it is difficult to obtain accurate staining with the more complicated textures. In extreme cases, it would probably be more satisfactory to analyse a vast number of thin sections by microscopic methods. However, in any case, the accuracy of any macro analysis is mainly a function of the quality of the stain.

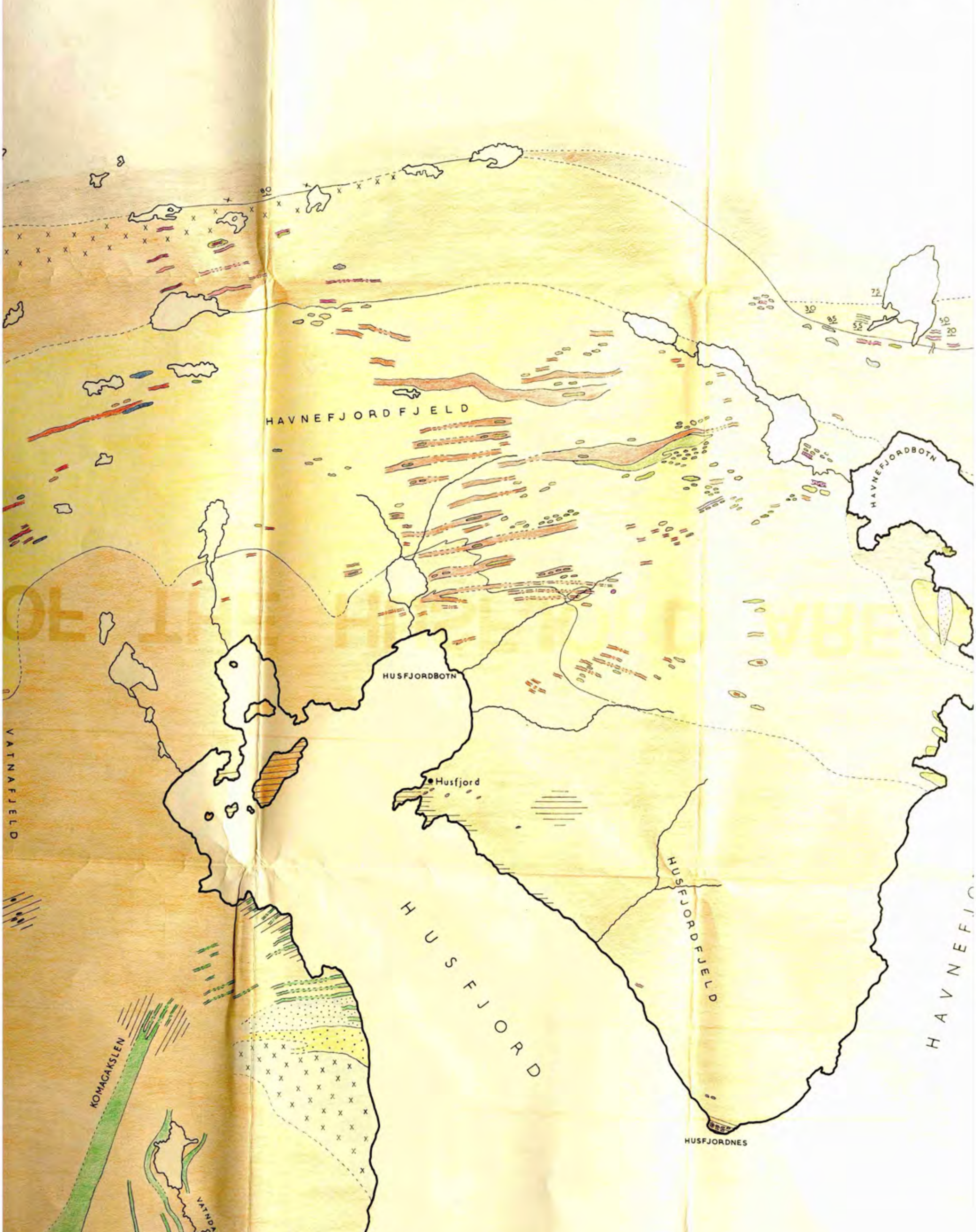
A, SØRØY

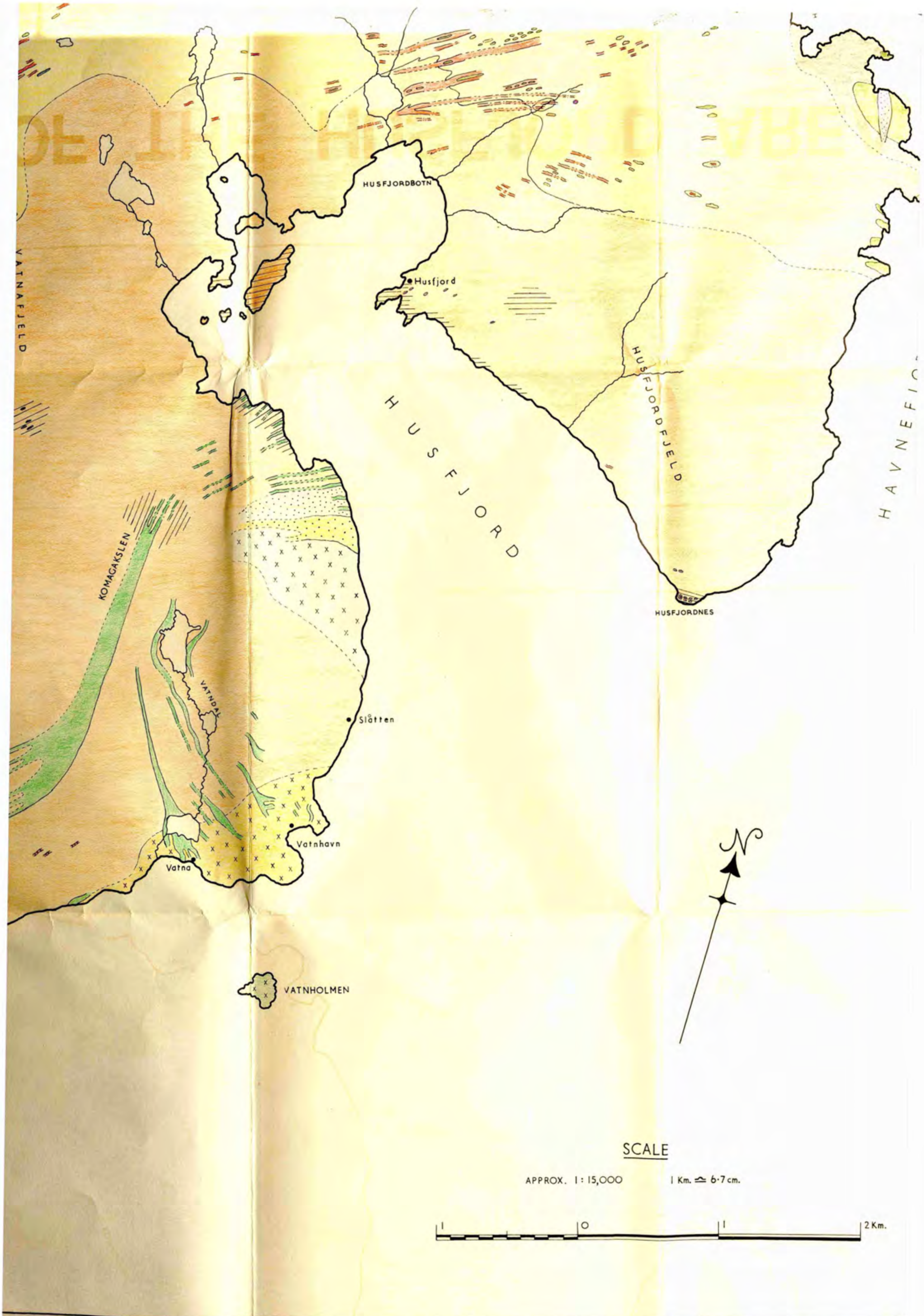


EXPLANATION


EXPLANATION
METASEDIMENTS
IGNEOUS ROCKS
COUNTRY ROCKS
QUARTZO-FELDSPATHIC PHYLLITIC SCHISTS
CALC-SILICATE-SCHISTS WITH METALIMESTONES
KYANITE-GARNET-MICA-SCHISTS
QUARTZ-GARNET-MICA-SCHISTS
CUMMINGTONITE-SCHISTS
PSAMMITES AND SEMI-PELITES (MIGMATIZED IN PART)
RAFTS IN IGNEOUS ROCKS
METALIMESTONE
PSAMMITE, PELITE, AND SEMI-PELITE
CALC-SILICATE-SCHIST
MIGMATIZED PSAMMITE AND SEMI-PELITE
HUSFJORD METAGABBRO
XENOLITHIC HUSFJORD METAGABBRO
CONTAMINATED HUSFJORD METAGABBRO WITH METASEDIMENTARY RAFTS
SLIGHTLY FOLIATED HUSFJORD METAGABBRO WITH EARLY DIORITE BANDS
TROCTOLITE
HORNBLENDITE
KOBBERFJORD NORITE
HAVNEFJORD DIORITE
PYROXENE-MICA-DIORITE
GARNET-RICH QUARTZ-DIORITE
GARNET-POOR QUARTZ-DIORITE
KOMAGFJORD DIORITE
PYROXENE-GABBRO AT SLÅTTEN
COARSE OLIVINE-GABBRO AT SLÅTTEN
FINE OLIVINE-GABBRO AT SLÅTTEN
VATNA GABBRO
OLIVINE LEUCOGABBRO OF VATNHOLMEN
PERTHOSITE
PEGMATITE
LITHOLOGICAL BOUNDARY
INFERRED BOUNDARY
60
DIP AND STRIKE OF METASEDIMENTARY BANDING (BEDDING)
VERTICAL BANDING
2 Km.
D. L. Speedyman, 1967

OF THE HUSFJORD AREA




HUSFJORDBOTN
Husfjord
VATNAFJELD
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VATNDAL
Slåtten
Vatnhavn
Vatna
VATNHOLMEN
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SCALE
APPROX. 1 : 15,000
1 Km. ≏ 6·7 cm.
1
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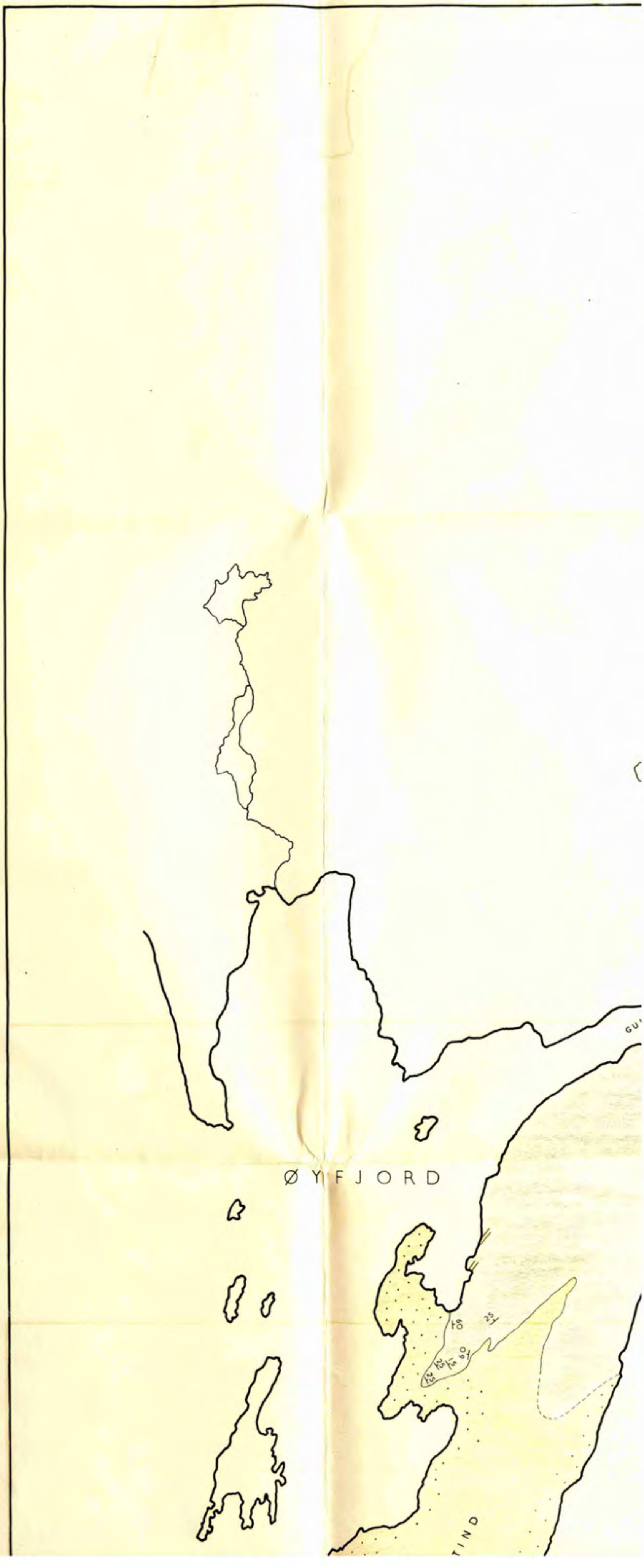
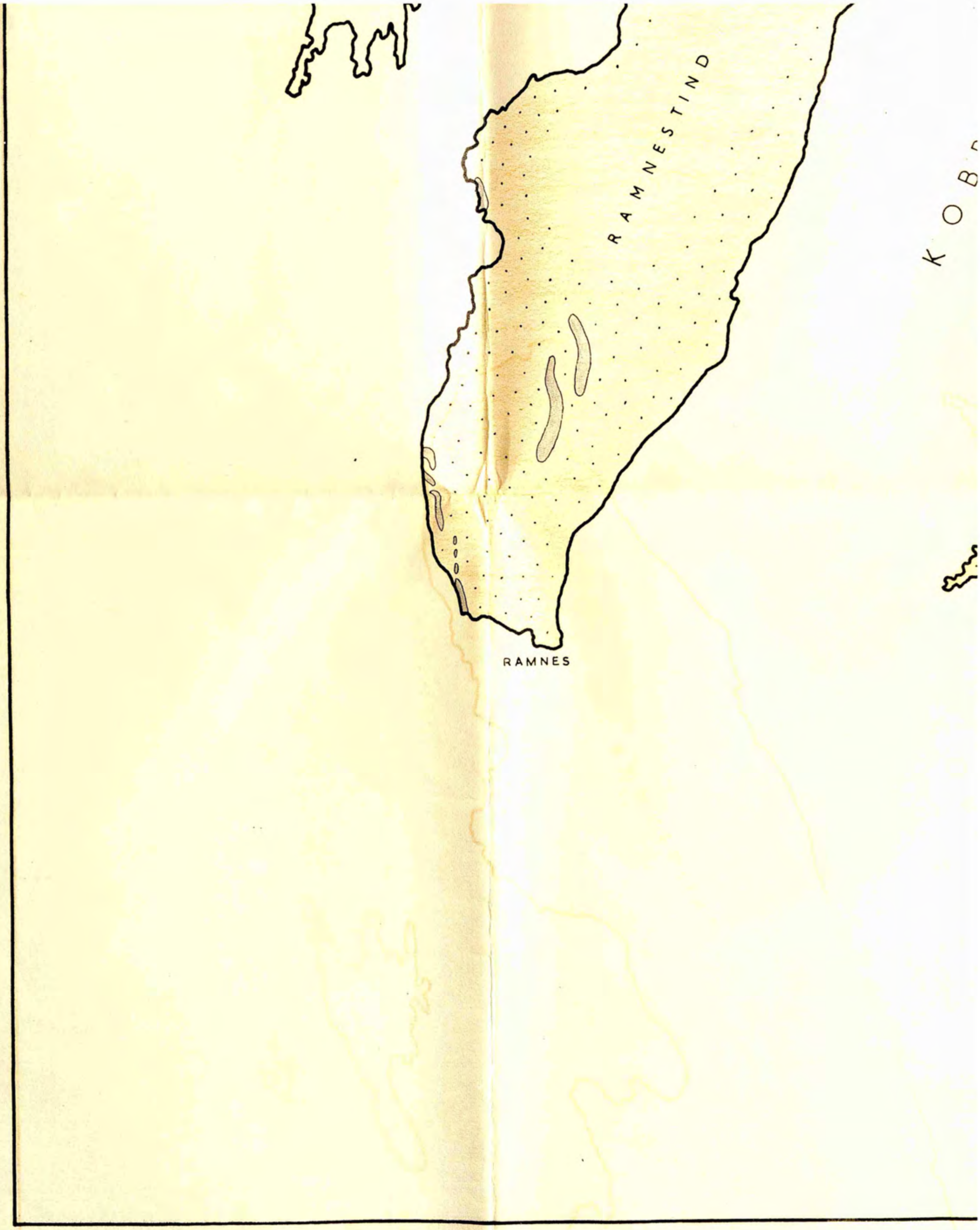
2 Km.

GEOLOGICAL MAP OF


KOBBERFJORD
STRANDTIND
STRØMMEN
Gåshopen
KOMAGFJORD


ØYFJORD


RAMNESTIND
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