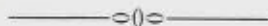# NORGES GEOLOGISKE UNDERSØKELSE NR. 207

PETROLOGY AND GEOCHEMISTRY OF HIGH-GRADE METAMORPHIC AND IGNEOUS ROCKS ON LANGØY, NORTHERN NORWAY

BY

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OSLO 1960

Contents.

Abstract.

The rocks on Langøy have been considered as representing an igneous rock series including metamorphic sedimentary rocks of "Caledonian type". The present work shows that they are for the most part high-grade metamorphic rocks which are intruded by a few post metamorphic igneous rocks. The petrography of the rocks is discussed. They range in composition from ultrabasic to granitic. Rocks in the 50—60 per cent silica range predominate areally.

The metamorphic rocks on Langøy are divided into: the banded series, the charnockite border series, the amphibolite facies veined gneisses, porphyroblastic monzonitic granulites, old red granites, red augengneisses (of monzonitic composition), metasedimentary rocks, and diaphtoritic gneisses. The Eidet-Hovden intrusion, young red granites and the Bø gabbro are considered as post metamorphic igneous intrusions.

The position of the anorthosite with associated monzonite east of Eidsfjorden is doubtful but may be related to the Eidet-Hovden intrusion.

The metamorphic facies and mineral equilibria are discussed. A metamorphic boundary which separates amphibolite facies rocks to the east from granulite facies rocks to the west coincides along the strike with the observed thrust plane under the anorthosite east of Eidsfjorden. The granulite facies rocks are charnockites and especially the banded series rocks are virtually identical to the type charnockites from India in the field appearance as well as in thin sections.

The two feldspar geothermometer (Barth, 1956) shows the amphibolite facies → granulite facies transition to take place at about 510° C. Temperatures above 700° C are found for the Eidet-Hovden monzonites and the young granites.

43 rocks, 26 potash feldspars, 8 plagioclases and 7 biotites were chemically analysed both for major and trace elements. This was done with two purposes in mind:

1. To investigate whether or not the chemical data could be used as indicators of the origin of the rocks.
2. To investigate whether or not some of the elements show variations which could be related to the change in metamorphic grade.

It was concluded that the chemistry does not give any conclusive information about the origin of such highly metamorphosed rocks except in some few

cases (lack of Cr in amphibolites derived by metamorphism of impure limestones).

Some of the volatile elements: e. g., Li, Na, Rb, Cs, Pb, and Tl vary with the metamorphic grade.

The possible use as a geothermometer of the distribution of some of the elements between minerals in equilibrium, e. g., Pb, Sr, and possibly Ba in feldspars, is discussed.

The Cs concentration in rocks is probably significantly altered by metamorphic processes. However, to be used to advantage the detection limit of this element by standard d. c. arc spectrographic methods must be lowered to about 0.1 p. p. m. Cs.

The petrogeneses and geological history of the area are discussed. The nature of the monzonitic and granitic rocks is uncertain. The banded series, charnockite border series, and amphibolite facies veined gneisses, which all can be termed banded or veined gneisses, are areally most important and differ only in metamorphic grade. Their field appearance and close association to metasedimentary rocks (marbles and graphite schists) indicate that they represent geosynclinal sediments and lavas. The amphibolite facies veined gneisses are identical to rocks on Hinnøy to the east which are considered as basal gneisses affected by the Caledonian orogeny.

The young red granites are regarded as consolidated anatectic magma. The Eidet-Hovden intrusion and the Bø gabbro are regarded as igneous rocks intruded after the period of granulite facies regional metamorphism.

The anorthosite east of Eidsfjorden is associated with a monzonitic rock similar to the Eidet-Hovden monzonites. This may indicate a relation in petrogenesis between the anorthosite and the Eidet-Hovden gabbro. The age of the rocks remains unknown. In best agreement with the geological information known today is the conclusion that they have all recrystallized or formed during the Caledonian orogeny.

Introduction.

General descriptions.

The island of Langøy lies in the county (fylke) of Nordland in northern Norway between approximately 68° 35′ to 68° 50′ N latitude at 15° E longitude. On the north and west Langøy is surrounded by the ocean. To the east the islands Andøy and Hinnøy are near neighbours separated from Langøy by the narrow Gavlfjorden and Sortlandssundet. Hadseløy is the nearest neighbour to the south, separated from Langøy by the narrow Børøysundet, which is at the narrowest about 1500 metres wide. Further south are parts of Hinnøy and Austvågøy. The map, Fig. 1, shows the general situation of the area. Langøy belongs to the group of islands generally known as the Lofoten and Vesterålen group. The group is composed of the major islands (listed from the south to the north): Mosknesøy, Flakstadøy, Vestvågøy, Austvågøy, Hadseløy, Langøy, Hinnøy and Andøy, with a number of smaller islands in between and further south. The last four islands constitute the Vesterålen group.

With an area of 860 km.² Langøy is the third largest Norwegian island, and the second largest in the group (surpassed only by Hinnøy). It is divided in five administrative districts or parishes (kommuner): Hadsel, Sortland, Langnes, Øksnes and Bø.

Langøy has a very irregular coastline with long fiords which typically are separated by narrow areas of low ground (eid).

The interior of the island is mountainous and rugged and reaches altitudes of 760 metres. A characteristic landmark is Reka, 607 metres (Fig. 2).

Most vegetation is restricted to near the coast and along the valleys. Rock exposures are therefore plentiful and good.

It is possible to drive around most of the island with a car, and at only one place, between Sandset and Kråkberget, is it necessary to use a car ferry. The central part of the island between Eidsfjorden and
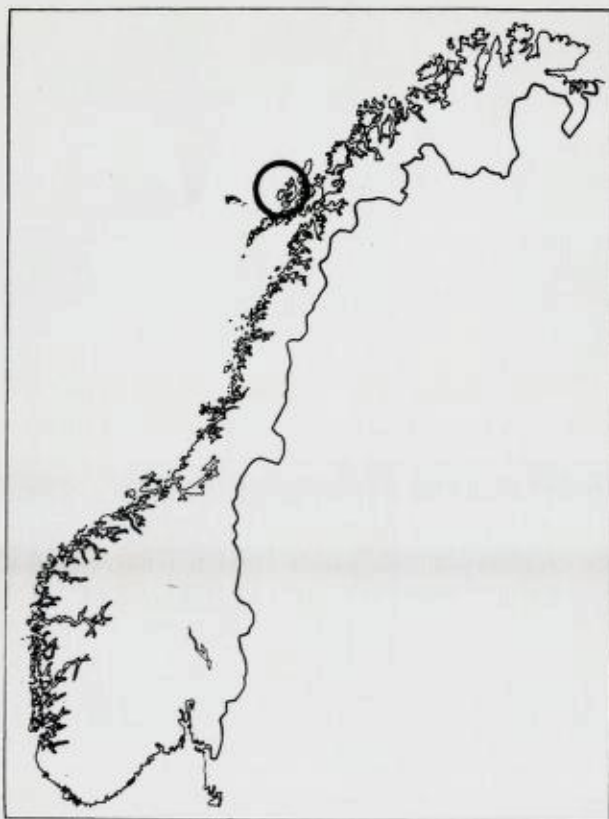
Fig. 1. Map showing the location of Langøy.
Kart som viser beliggenheten av Langøy.

Sandset in the northeast and Kråkberget and Jørgefjord in the west is very rugged with only a few houses close to the sea and no road connections.

Langøy has 15,700 inhabitants (1950), most of whom live along the coast and concentrated in small villages. The largest of these is Sortland.

Hotel accomodation can be obtained in Sortland (on the east side of the island) and Steine in Bø (on the west side). There also is a hotel in Stokmarknes on Hadseløy; a car ferry runs frequent routes between Stokmarknes and Sannes on Langøy.

Small scale farming generally combined with fishing is the main

Fig. 2. Area underlain by granulite facies rocks west of Eidsfjorden;
"Reka" to the left.

Område som består av bergarter i granulit facies vest av Eidsfjord;
"Reka" sees til venstre.

occupation on Langøy. Fishing is most important for the economy of
the island as a whole, agriculture is important especially in the eastern
part. The importance of agriculture is manifest in the modern school of
agriculture for northern Norway at Kleiva some kilometres south of
Sortland.

As grain farming is not possible in this part most of the farming
is concentrated upon dairy products and sheep farming. A central dairy
for the whole island is located at Sortland. Apart from this and some
fish refineries there are no other industries. A graphite mine at Jen-
nestad (10 kilometres north of Sortland) is in the process of being
developed.

Approach to the island is made by a daily boat from Bodø (about
12 hrs. Bodø-Sortland), or by bus from Harstad on Hinnøy. On the
island transportation is taken care of by bus routes. The climate is the
typical coast climate with much rain in normal summers. The sun is
above the horizon 24 hrs. a day between the 27th of May and the
27th of July.

Previous work in the area.

The following is a chronological list of published papers in which the geology of the island (or parts of it) is discussed. (These papers are not included in the general reference list unless specifically referred to in the text).

Keilhau, B. M. (1844), Über den Bau der Felsenmasse Norwegens. Gaea Norwegica, 2, pp. 303—312.

— (1861), Beretning om en geognostisk reise til Nordlandene i 1855. Nyt Mag. f. Natur. 11, pp. 209—219.

Pettersen, K. (1874), Geologiske undersøkelser inden Tromsø Amt og tilgrænsende Dele av Nordlands Amt IV. Kgl. Norske Videnskabers-Selsk. Skr. I, pp. 257—444.

— (1875), De gneis-granitiske dannelser langs det nordlige Norges kyststrøg. Geol. För. Stockh. Förh. 2, pp. 450—468.

— (1881—1882), Lofoten og Vesterålen, 1 and 2. Archiv f. Math. og Naturv. 5, pp. 369—435, and 6, pp. 81—111.

Vogt, J. H. L. (1897), Norsk marmor. Norges Geol. Unders. 22, pp. 1—364.

Helland, A. (1897), Lofoten og Vesterålen. Norges Geol. Unders. 23, pp. 1—536.

Kolderup, C. F. (1898), Lofotens og Vesterålens gabbrobergarter. Bergens Museums Aarbog, pp. 1—56.

Vogt, J. H. L. (1900), Weitere Untersuchungen über die Ausscheidung von Titanerzen in basischen Eruptivgesteine. Zeitschrift für practischen Geologie. 8. pp. 233—242.

— (1907), Über die schrage Senkung und die spätere schrage Hebung des Landes in nordlichen Norwegen. Norsk Geol. Tidsskr. 1, no. 2, pp. 1—47.

Helland, A. (1907—1908), Nordlands amt, 1 and 4, Norges land og folk.

Vogt, J. H. L. (1909), On Labradorite-Norite with Porphyritic Labradorite Crystals: a contribution to the Study of Gabbroidal Eutectic. Quarterly Journal of Geol. Soc. 65, pp. 81—103.

Vogt, Th. (1909), Om eruptivbergartene på Langøen i Vesterålen, Norges Geol. Unders. 53, pp. 1—39.

Vogt, J. H. L. (1910), Norges Jernmalmforekomster, Norges Geol. Unders. 51, pp. 1—225.

Vogt, Th. (1915), Kysteruptiverne i Tromsø Amt, Norsk Geol. Tidsskr. 3, pp. 66—67.

Vogt, Th. (1922), Bidrag til fjellkjedens stratigrafi og tektonik. Geol. För. Stockh. Förh. *44*, pp. 714—739.

— (1941), Trekk av Narvik—Ofoten-traktens geologi, Norsk Geol. Tidsskr. *21*, pp. 198—213.

Holtedahl, O. (1944), On the Caledonides of Norway, Skrifter, Det Norske Videnskaps-Akademi i Oslo, 1. Mat.-Naturv. Klasse. No. *4*, pp. 1—31.

— (1953), Norges geologi, Norges Geol. Unders. *164, 1*, pp. 399—401.

In addition to the published papers listed above there are available a number of unpublished reports on various occurrences of graphite schists, especially from the one at Jennestad, and on the magnetite-ilmenite ore at Selvåg.

Summary of previous geological interpretations of the area.

Keilhau and Pettersen considered most of the island to be under-lain by gneisses with granitic affinities. According to the latter even the granites of the area are stratified. Later Helland (1897) and Vogt (1897) visited the island and found that granites are subordinate in amount, and that most of Langøy is underlain by gabbros and related rocks. According to Helland (1897, 1907, 1908) the only place where rocks of granitic affinities are important is on the eastern part, in Sortland.

Helland also stressed the similarity between the rocks of the Lofoten-Vesterålen group and the "Jotun rocks" (gabbroic and anorthositic rocks of charnockitic affinities of central Norway). Vogt compared them with the Egersund suite of rocks (southern Norway) which is similar to the "Jotun rocks". The latter view was supported by Kolderup (1898) who made a petrographical study of rock samples earlier collected by Helland. Kolderup had earlier made an extensive study of the Egersund rocks, and was quite familiar to the rock types that most geologists today term granulite facies rocks or charnockites. According to him the whole western and northern part of Langøy is underlain by gabbros, whereas gneisses underlie the eastern parts. Helland also concluded that the rocks have not been affected by re-gional metamorphism of any importance, and regarded the minerals of the crystalline rocks as primary igneous ones. He stated, however,

that pressure effects are found to be much more important in Lofoten and Vesterålen than in the corresponding Egersund area.

In 1909 J. H. L. Vogt pointed out that the rocks on Langøy formerly named gabbros by Helland, Kolderup and himself can be divided into a whole series of rocks ranging from gabbroic to granitic types, but he concluded that the basic rocks are quantitatively most important.

The same year (1909) Th. Vogt published the results of one month's field work in the summer of 1908 covering the western half of Langøy, including a geological map of the area. He was able to separate the rocks into different groups which he termed: gabbros (herein included a number of different types ranging from ultrabasics to normal gabbros), dioritic monzonites (a series characterized by alternating bands of dioritic and monzonitic composition), monzonites, and granites. Within the dioritic monzonites "inclusions" of crystalline limestones and graphite schists occur frequently. All the crystalline rocks, except the inclusions, were considered by Vogt as products of fractional crystallization from the same parent magma. It is interesting to see that whereas Pettersen, who belonged to the school of transformists, was able to describe the rocks as granitic gneisses (even when he was obviously dealing with gabbros he described them as "peculiar varieties of granites"), Th. Vogt, who was influenced by the modern theories of fractional crystallization, which in this country were strongly supported by his father Professor J. H. L. Vogt and by Professor W. C. Brøgger, was able to describe the very same rocks as crystallization products of a gabbroic magma. After Th. Vogt's visit in 1908 no other geologist has reexamined the geology on Langøy, though the above mentioned graphite schists and iron ores have been investigated repeatedly from a commercial point of view. The results of these examinations are available as unpublished reports in the files of the Geological Survey of Norway.

Up to now the knowledge of this area has been based on the data given by Th. Vogt and the earlier geologists, and the understanding of the petrogenesis has largely been based on Vogt's views. Opinions as to the age of the rocks have been highly diverse, and ages ranging from pre-Cambrian, to "Caledonian" and post-Silurian have been mentioned. A post-Silurian (Devonian?) age was held by Th. Vogt (1909) as he believed the rocks to be unmetamorphosed igneous, and the graphite schists and limestones occurring in them as inclusions of contact metamorphosed sediments of a Caledonian age. The most generally held

opinions have, however, been that they represent either pre-Cambrian or "Caledonian" intrusions. No radioactive age determinations have been carried out on the Langøy rocks. On the geological map of Norway by Holtedahl and Dons (attached to Holtedahl, 1953) the Lofoten and Vesterålen rocks are indicated by a yellow colour quite distinct from the "Caledonian" and pre-Cambrian colours and are designated "plutonic rocks of Lofoten". The fairly extensive area of metasedimentary rocks at Jennestad (graphite schists, limestone, quartzites in gneisses) has on this map been given the blue colour of the metamorphosed Caledonian sedimentary rocks.

Present investigations.

The field work was carried out during five months of the summers of 1955, 1956 and 1958 with a field allowance from the Geological Survey of Norway. Some of the microscopic work was carried out at the Mineralogisk-Geologisk Museum, Oslo, (1955—1956), but most of it and all of the spectrographical analyses were done at the Department of Geology and Mineralogy, Oxford (1956—1958). Compilation and writing up of the results were done at the Mineralogisk-Geologisk Museum, Oslo (1958—1959).

The purpose of this study was to examine the petrography and geochemistry of the rocks in order to elucidate their mineral paragenesis and petrogenesis. The regional mapping was restricted to the amount necessary to know the field relations between the different rock types, and the appearance of the rocks in the field.

The general strike and dip directions as they are indicated by the foliation in the sedimentary rocks and gneisses and also by the banding in the latter are indicated on the geological map (Fig. 3, attached as appendix). The mapping was done on the scale 1:50,000. The veined gneissic rocks and the metamorphic sedimentary rocks of the area can be seen to be extensively folded. A structural study was beyond the scope of this paper. The work is confined to the rocks on Langøy. The area north of an east-west line between Gåsfjorden and Godvikbotn is not included although the geology there is comparable to the districts further south. Thus only a comparatively small part of the area occupied by the "plutonic rocks of Lofoten" has been examined in detail (roughly one fifth of the total area). However, there is no reason to believe that the conclusions drawn from the data obtained by this in-

Fig. 4. Map showing the field relations between the "plutonic rocks of Lofoten" and the "basal gneisses" containing metamorphic Caledonian sedimentary rocks.

Kart som viser "Lofoteruptivenes" forhold til de omgivende "bunn-gneiser" med Kaledonske metamorfe sedimenter.

vestigation are not valid for the whole district. All the characteristic rock types included in the term "plutonic rocks of Lofoten" are displayed within the mapped area on Langøy. Reconnaissanse work has been done in the neighbouring districts on Hinnøy and Hadseløy, and on Vestvågøy in the Lofoten group.

General geology.

The general setting of the "plutonic rocks of Lofoten" as shown on the geological map of Norway by Holtedahl and Dons is given in Fig. 4.

The geology on the parts of Langøy mapped by the author is shown in Fig. 3. Early during the first summer's field work it became apparent to the author that most of the "plutonic rocks of Lofoten"

Fig. 5. Amphibolite facies veined gneisses, Sigerfjord, Hinnøy.
Åre gneiser i amfibolit facies, Sigerfjord, Hinnøy.

that occur on Langøy are high grade metamorphic rocks. This was confirmed later by the study of thin sections. The metamorphism ranges from that of the amphibolite facies (in the east) to that of the granulite facies (in the west). The granulite facies rocks have charnockitic affinities, as these are defined by Holland (1900).

Within this metamorphic series of rocks are other rock types of less areal importance which may have been emplaced during a late stage of, or after, the general regional metamorphism. These are the Eidet-Hovden intrusion in the northwest, which is comprised of gabbroic and monzonitic rocks and "young red granites" which occur as small patches and a few larger massifs scattered through the granulite facies area. The gabbro in the south of Bø also may be of postmetamorphic origin. This suggestion is based mainly on the assumption that the retrograde metamorphism of the gneisses north of the mass was caused by escaping volatiles during its crystallization.

The rocks on Langøy have in this work been divided into the following groups:

Fig. 6. Veined gneisses of the charnockite border series; roadcut north of Stor-
vann between Frøskeland and Steinlandsfjorden.

*Åre gneiser som tilhører "the charnockite border series"; vegskjæring nord av
Storvann mellom Frøskeland og Steinlandsfjorden.*

1. The amphibolite facies veined gneisses.

   This series is a direct westward continuation of the basal gneisses
   on Hinnøy and consists mainly of irregular, alternating bands and
   veins of acidic (granitic-monzonitic) and basic (amphibolitic)
   rock types (Fig. 5). The westward continuation of this series is
   represented by:

2. The charnockite border series.

   These rocks appear in the field to be similar to the amphibolite
   facies veined gneisses (Figs. 6, 7), but they are mineralogically
   different in that hypersthene occurs in both the acid and basic
   types. This series grades into:

3. The banded series (charnockites proper).

   This is a banded or veined series of rocks similar to the other two
   (Fig. 8). It is distinguished from the charnockite border series
   because of its subordinate amount of biotite.

Fig. 7. Folded rocks of the charnockite border series; roadcut at
Indre Strømfjord.

*Foldete bergarter tilhørende "the charnockite border series"; veg-
skjæring ved Indre Strømfjord.*

Hypersthene is the dominant femic silicate mineral in all rocks
within this series, with the possible exception of the extremely
silica deficient rocks. The veining of this series is not readily ap-
parent except on strongly weathered surfaces because of the dark
colour of quartz and feldspar in granulite facies rocks.

These three rock series which may be classified as veined or
banded gneisses are the most frequent rock types on Langøy. The
banded series and the charnockite border series were termed

Fig. 8. Veined gneisses of the banded series (weathered surface), Vågsbøgen (271 m.a.s.l.), Eidet.

Åre gneiser tilhørende "the banded series" (forvitret overflate), Vågsbøgen (271 m.o.h.), Eidet.

"dioritic monzonites" by Vogt (1909) because of the dioritic affinities of the basic bands and the monzonitic to quartz monzonitic affinities of the acid bands. They were regarded by him as intermediate rocks in an igneous series. They are regarded by the present author as representing, for the most part, severely metamorphosed geosynclinal sediments and lavas.

4. Metamorphic sedimentary rocks occur interbanded with the banded gneisses of each of the three above mentioned series. The sedimentary rocks consist of bands and lenses of crystalline limestones and graphite schists. Typically their foliation is parallel to the banding of the gneisses. In the amphibolite facies gneisses occur quartzites which tend to be feldspathized; in the granulite facies areas quartzites, if they occur, cannot be distinguished easily from normal granulites.

Metamorphic "quartz banded iron ores" also occur within the three series, and also were probably originally sedimentary.

Fig. 9. Massive red augengneiss (monzonitic composition); roadcut west of Sannes.

Massiv rød øyegneis, vegskjæring vest av Sannes.

5. Porphyroblastic monzonitic granulites.

This rock type occupies an area in the far western part of Langøy. The unit has an overall homogeneity and is not foliated. Except for the large potash feldspar porphyroblasts which characterize them are they mineralogically and chemically identical to the intermediate rocks in the banded series with which they have a gradational contact.

6. Red augengneisses in the southeast.

These rocks are in several respects similar to the porphyroblastic monzonitic granulites, but have an amphibolite facies mineral paragenesis. They are characterized by red potash feldspar porphyroblasts (Fig. 9) and are typically but slightly foliated. They contain sporadic amphibolitic inclusions commonly as small "ghost-like relics". A coarse grained red granitic gneiss occurs close to the amphibolite facies veined gneisses and red potash

feldspars are characteristic in the veined gneisses several hundred metres away from the contact.

7. The granulite facies counterpart of the red augengneisses occurs in a narrow zone southeast of Eidsfjorden. The potash feldspars in this rock are gray in colour.

   These rocks are in many ways similar to the monzonites within the Eidet-Hovden intrusion.

8. The old red granite.

   This is a very coarse grained hypersthene carrying granitic rock which occurs in a restricted area northwest of Eidsfjorden. It has a gradational "migmatitic" contact against the surrounding banded series. Inclusions of the latter are common in the granite. The veining of the inclusions is parallel to that of the surrounding banded series.

9. The Bø gabbro and the retrograde metamorphic gneisses.

   The gneisses occur in the area from the south of Bø (the gabbro) up along the west side of Jørgenfjord. The gabbro is a massive two pyroxene (ortho- and clinopyroxene) gabbro. The retrograde metamorphic gneisses include types chemically similar to the porphyroblastic monzonitic granulites and the banded series. Sillimanite gneisses are also present locally.

10. Anorthosites.

    Anorthosites occur at two places on Langøy. One at Sunnan at the extreme north end of the peninsula between Malnesfjord and Møklandsfjord and the other between Grønning and Strømfjord on the southeast side of Eidsfjorden.

11. The Eidet-Hovden intrusion.

    This is an intrusion of gabbroic and monzonitic rocks that occupies the peninsula between Malnesfjord and the ocean on the west side of Langøy.

12. Young red granites.

    These occur as numerous small patches some of which are too small to be indicated on the map (Fig. 3), and a few larger massifs within the area occupied by the porphyroblastic monzonitic granulites and the banded series.

In the amphibolite facies area in the east, the gneisses strike about E-W, whereas they strike more N-S in the granulite facies area. The strike as defined by the banding or veining of the gneisses is conformable around the Eidet-Hovden intrusion.

Fig. 10. Anorthosite thrusted above red augengneisses, Grønning, southeast side of Eidsfjorden.

Anorthosit som er skjøvet over røde øyegneiser, Grønning, sydøst siden av Eidsfjorden.

The anorthosite between Grønning and Strømfjord has been thrust over the red augengneisses in the southeast. The thrust zone strikes E-W and dips 40° to the N (Fig. 10). It is marked by mylonite about 40 metres thick. The mylonite zone does not crop out along the strike within the gneiss area except close to the gneiss-anorthosite contact. The thrust plane, which describes an arc in outcrop defines the metamorphic facies boundary between the amphibolite and granulite facies gneisses to the north and southeast. A solid line on the map (Fig. 3) indicates the thrust plane where it was observed as such; a dashed line marks its continuation as the metamorphic facies boundary.

The rocks on both sides of the facies boundary are chemically similar. The amphibolite facies veined gneisses in the east succeeded by the veined charnockite border series to the west are found north of the anorthosite. The strike changes from E-W to a more N-S direction.
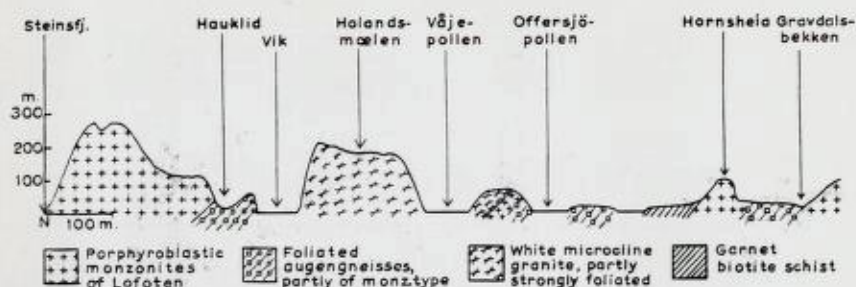
Fig. 11. Schematic prophile showing the geology on the west side of Vestvågøy in Lofoten.

Skjematisk profil som viser geologien på vestsiden av Vestvågøy i Lofoten.

The red augengneisses in amphibolite facies succeeded by a gray augengneiss (monzonitic type) in granulite facies occur south of the anorthosite. The chief difference observable in the field is the change in colour from pink to gray of the potash feldspars.

Some significant geographical features in the area run parallel to the arc defined by the metamorphic boundary and the thrust plane. Sortlandssundet and Gavlfjorden, which form the eastern side of Langøy, describe together closely parallel arcs. A similar arc is described by Eidsfjorden, the valley (natural depression) between Frøskeland and Steinlandsfjorden north of Eidsfjorden and out along Steinsfjorden. The area between this arc and the metamorphic boundary roughly coincides with the area of the charnockite border series, between the banded series (charnockites proper) to the west and the amphibolite facies gneisses to the east.

The narrow Øksfjorden on Hinnøy is parallel to Eidsfjorden and Sortlandsfjorden and closely marks the southeast boundary of the Lofoten rocks on Hinnøy.

The author is of the opinion that imbricate overthrusts, along which granulite facies rocks were thrust over chemically similar rocks of the amphibolite facies, are especially significant features in the Lofoten-Vesterålen area (Hadseløy is largely, if not completely, underlain by amphibolite facies rocks similar to those on the southeast end on Langøy). Skjeseth (personal communication) has found that the basal gneisses were thrust over the Caledonian sedimentary rocks near Harstad on the east side of Hinnøy. The presence of overthrusts is also indicated by structures observed by the author on Vestvågøy in Lofoten.

Fig. 12. Topography of an area underlain by amphibolite facies rocks south of Vikbotn; Jennestad and Ånstadblåheia are seen in the background.

Topografi i et område som oppbygges av bergarter i amfibolit facies, syd av Vikbotn; Jennestad og Åndstadblåheia i bakgrunnen.

Fig. 11 is a schematic profile between Steinsfjorden and Buksnes on the west side of Vestvågøy. The coarse grained monzonites are similar to the porphyroblastic monzonitic granulites in the west of Langøy, with minerals characteristic of the granulite facies, whereas the gneisses and micaschists belong to a lower metamorphic facies. The similarity of the rocks on both sides of the thrust plane (metamorphic boundary) on Langøy indicates that the distance over which the thrusting took place, is probably not very great.

There is a marked difference in the topography between the areas underlain by the amphibolite versus the granulite facies rocks. The former are characterized by rounded surfaces, whereas the latter are characterized by rugged relief (Figs. 12, 13 — see also Fig. 2).

In the following chapters the petrography of the rock types and rock series is treated in some detail first. A chapter dealing with the metamorphic classification of the rocks and the probable metamorphic reactions based on the petrographical data follows. Next comes a discussion of the chemistry of the rocks, and thereafter a discussion of the petrogeneses of the rocks. The last section presents a summary, and the author's attempt to unravel the history of the rocks.

Fig. 13. View of Langøy from Kliftra (541) on Hadseløy. The photograph shows the typical rugged relief of the area underlain by granulite facies rocks northwest of Eidsfjord (to the left and in the background) in contrast to the more smooth and even relief of the area underlain by amphibolite facies rocks (to the right).

Utsyn over Langøy fra Kliftra (541) på Hadseløy. Billedet viser det typiske ujevne relief av området som oppgygges av bergarter i granulit facies nordvest av Eidsfjord (til venstre og i bakgrunnen) sammenlignet med de mer jevne og avrundede former i området som består av amfibolit facies bergarter.

Acknowledgements.

Petrography.

The porphyroblastic monzonitic granulites.[1]

These rocks occur only in the southwest part of Bø parish. Their eastern boundary is defined by the N-S striking zone of lower grade metamorphic gneisses and towards the north they are bordered by the banded series rocks. The latter contact is not sharp and porphyroblastic monzonites occur sporadically within the banded rocks. Their abundance within this series diminishes rapidly towards the north. Where they occur inside the banded series, partly assimilated fragments of the basic members of the latter can be seen within them (Veggfjeld). The porphyroblastic monzonites are both chemically and mineralogically similar to the intermediate rocks of the banded series. The distinguishing feature is the presence of large sized potash feldspars in the former. However, all gradations exist between the two rocks. A fairly straight line striking ESE from Bøklubben (west on Skårvåg) and directly north of Hole and Veatind to Poll, essentially coincides with their extension northwards. The porphyroblastic monzonites are massive, unfoliated and homogeneous rocks. The only other rocks present within the monzonite area are the young red granites, which occur as small irregular patches, one larger massif (Veatind-Nåla), and a few well defined crosscutting dykes (Søberg).

The rocks, like the typical charnockites, are rather dark gray. Plagioclase, the lightest constituent, is light grayish brown. Quartz is generally absent and where present it is dark gray. The most conspicious and characteristic feature of the rocks is the presence of large potash feldspar porphyroblasts. These are, in the extreme, more than one cm. in largest dimensions. The best examples occur on the extreme west of the Søberg peninsula. They are dark gray brown and are remarkable clear and free from inclusions.

[1] For the sake of convenience these rocks are often referred to as western monzonites and porphyroblastic monzonites in the text.

Table 1.

*Modal composition of the porphyroblastic monzonitic granulites
(point counted thin sections).*

Minerals	1 280/55	2 281/55	3 293/55	4 55/56	5 225/55	6 251/55	7 263/55
Quartz.............	6	–	tr.	–	tr.	–	6
Potash feldspar.....	19	13	28	24	24	12	13
2Vx	45		50–55	50			
Plagioclase	64	73	62	63	60	76	66
Hypersthene.......	3	4	tr.	2	5	5	12
Diopside	3	5	2	4		2	–
Hornblende	∠1	1	–	4	tr.	tr.	–
Biotite............	∠1	tr.	6	∠1	1	2	1.5
Iron ore..........	3	4	2	3	9	2	1.5
Zircon	tr.	–	tr.	–	–	tr.	–
Apatite	tr.	tr.	tr.	tr.	1	∠1	–
Calculated mol per cent SiO₂ ...	55.8	52.2	53.3	53.7	49.6	52.3	55.9

1. West of Søbergfjell, 2. South of Søbergfjell, 3. West side of peninsula at Vinje, 4. Roadcut, top 115, Steine, 5. Nåla, 6. Veatind, 7. Strømsjøen.

Modal compositions of 7 of the rocks are given in Table 1. (No. 7 actually occurs in the banded series north of the monzonite proper). The similarity among these is good considering their porphyroblastic nature and the consequent difficulty in making representative thin sections. Chemical analyses, C. I. P. W. norms, and calculated modes of two monzonites are shown in Table 2.

The main difference is that hornblende [1] and biotite are present in the mode instead of olivine and nepheline in the norm. In no. 57/56 hornblende is not present. The calculated mode was obtained by substituting olivine for biotite. This was achieved by correcting for the silica deficiency of the analysis with the hypothetical mineral reaction:

$$3 SiO_2 + K(Mg,Fe)_3 AlSi_3 O_{10}(OH)_2 \rightleftharpoons KAlSi_3 O_8 + 3(Mg,Fe)SiO_3 + H_2O \qquad 1$$
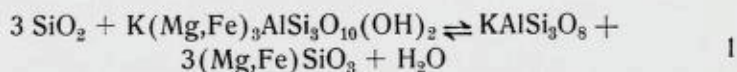
(see Ramberg, 1952, p. 152).

$$3Q. + 8Bi. = 5Or. + 6Hy. \qquad 1b$$
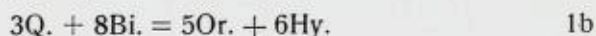
[1] The name hornblende is used for any common amphibole in this paper.

Table 2.

Chemical anlcyses, C.I.P.W. (molecular) norms and calculated modes of two porphyroblastic monzonitic granulites.

Weight %	57/56 (Skårvåg)	55/56 (Steine)
SiO_2	56.38	57.46
TiO_2	0.87	0.80
Al_2O_3	18.86	20.02
Fe_2O_3....................	3.15	2.53
FeO	3.78	2.64
MnO	0.11	0.09
MgO	2.46	1.38
CaO	5.53	4.81
Na_2O	5.22	5.55
K_2O	2.96	4.00
H_2O^-	0.08	0.25
H_2O^+	0.19	0.20
CO_2	n.d.	n.d.
P_2O_5....................	0.38	0.45
Sum....................	99.97	100.18

C. I. P. W. molecular norm.

K-feldspar (or)............	17.5	23.5
Na-feldspar (ab)...........	46.5	48.5
Ca-feldspar (an)...........	19.0	17.8
Nepheline (ne)	–	0.3
Olivine (ol)	0.3	3.0
Hypersthene (hy)..........	6.6	–
Diopside (di)	4.8	2.4
Magnetite (mt)	3.3	2.7
Ilmenite (il)	1.2	1.0
Apatite (ap)...............	0.8	0.8

Calculated mode

K-feldspar (or)............	17.3	21.7
Na-feldspar (ab)..........	46.5	49.0
Ca-feldspar (an)	19.0	5.8
Hypersthene (hy)..........	6.8	0.4
Diopside (di)	4.8	2.4
Hornblende (hb)	–	3.0
Biotite (bi)	0.3	2.8
Magnetite (mt)............	3.3	2.7
Ilmenite (il)	1.2	1.0
Apatite (ap)	0.8	0.8

(anal. E. Christensen)

instead of the common normative correction:

$$4En. = 3Fo. + 1Q. \quad \text{(Barth, 1952, p. 80)} \qquad 2$$
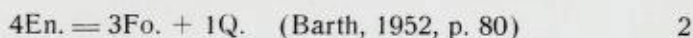
In no. 55/56 the ratio hbl./bio. is about 4/1. The silica deficiency may here be accounted for by combining reactions 1b and 4b (below) in suitable proportions instead of using equations 2 and 3.

$$5Ab. = 3Ne. + 2Q. \quad \text{(Barth, 1952, p. 80)} \qquad 3$$

$$SiO_2 + Ca_2(Mg,Fe)_3Al_2Si_6Al_2O_{22}(OH)_2 \rightleftharpoons$$

$$2CaAl_2Si_2O_8 + 3(Mg,Fe)SiO_3 + H_2O \quad \text{(See Ramberg, 1952 ,p. 68)} \quad 4$$

$$1Q. + 15 \ Hbl. = 10 \ An. + 6 \ Hy. \qquad 4b$$
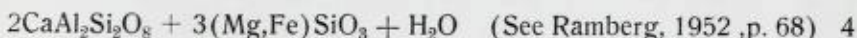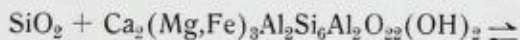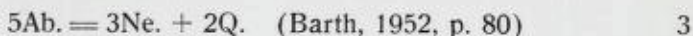
Quartz is generally absent. Where present it has a marked undulatory extinction. Spindles of myrmekitic quartz inside plagioclase which is in contact with potash feldspar are, however, present in all of these rocks.

Potash feldspar is characteristically present as large rectangular porphyroblasts. They appear to be monoclinic, or very nearly so, both optically and on X-ray powder diagrams. The (131) reflection of the latter is typically slightly diffuse. This has been found to be the general case for natural feldspars of the orthoclase type (Heier, 1957).

$2V_x = 45°—55°$, usually $50°—55°$, $X \wedge a = 6°—8°$, $AP \wedge (010) = 89°—90°$.

There is in general no sign of the undulatory extinction that characterizes potash feldspars of intermediate triclinicities. The low $2V_x$ (45°) was obtained from the feldspars in the most western monzonites, 1 of Table I. It should be noted that this smaller 2V is most characteristic of the smaller sized potash feldspars in the banded series rocks.

All of the potash feldspars are perthitic. Their chemistry is discussed in a separate section in this paper.

The perthites are of the string type, i. e., the exsolved albite lamellae have the form of strings or elongated drops. The elongation is parallel to (010) and at an angle of about 72° to (001). Strings in two directions at a 60° angle are less common. Along the contacts against plagioclase, or any other mineral, the potash feldspars are not perthitic in the same manner (Fig. 14). The exsolved albite lamellae are nearly sub microscopic. The same effect occurs where stringers of potash feldspar penetrate plagioclase (a common occurrence in these rocks). The size of the exsolved plagioclase lamellae of these stringers varies directly with the width of the stringers. Within the relatively broad stringers the plagioclase lamellae are identical in size and orientation to the plagioclase lamellae within the main potash feldspar grains with which the stringers are connected. Although mesoperthites occur in these rocks, they are not characteristic. Apart from the perthitic nature the potash feldspars are free of inclusions, including dust of any kind.
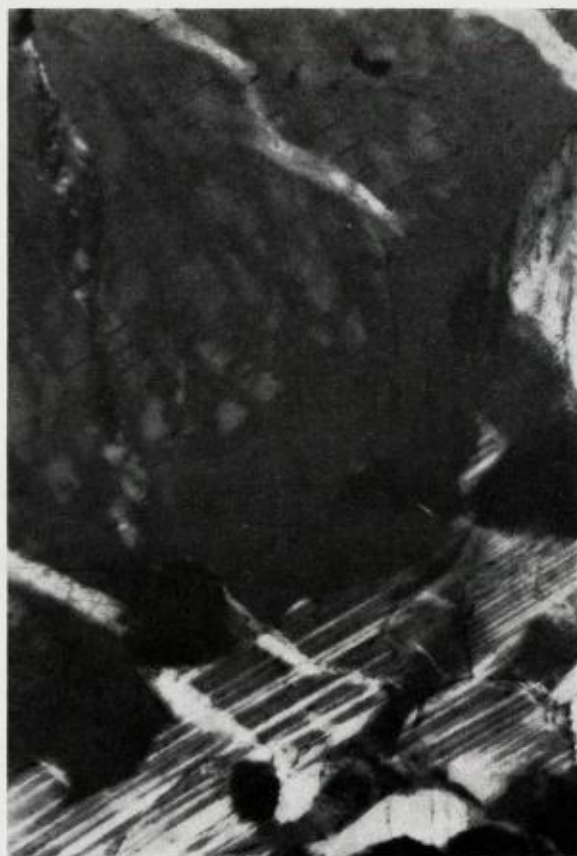
Fig. 14. Potash feldspar perthite which shows no signs of exsolved plagioclase lamellae close to the contact with plagioclase. (+n. 100×).

Perthitisk kalifeltspat hvor utskilte plagioklas lameller ikke kan sees nær kontakten mot plagioklas. (+n. 100×).

Plagioclase is the dominant mineral in the rocks. The grain size varies within rather large limits but no plagioclase grains are porphyroblastic in appearance. The larger grains especially are antiperthitic. The potash feldspar lamellae generally have rectangular shapes, but in some cases have the form of small drops.

Like the potash feldspar the plagioclase is free from inclusions and dust of any kind.

All plagioclase is twinned. Twinning according to the albite law is most common, but pericline twins and combinations of the two types are also common. More complex twins are rare. Typically individuals of one set of lamellae differ greatly in size, the one lamellae being broad and the other narrow.

Plagioclase compositions were determined on the universal stage. The parameters determined were: 2V, optic plane and normal, (010) and (001) cleavages, composition planes, and extinction angles. For twinned grains both lamellae were measured and thus values for $X \wedge X$, $Y \wedge Y$, $Z \wedge Z$, were obtained as well. Refractive indices on cleavage flakes were also measured for some specimens. The readings were plotted on the curves published by Trøger (1952) and Turner (1947). The compositions vary between An. 25—32 and cannot, except for the specific cases, be given more accurately; they are probably close to An. 30.

Compositional zoning was not observed.

Orthopyroxene (hypersthene) was seen to occur in all thin sections of the rocks examined. It is non pleochroic. $2V_x$ was measured in a number of grains and found to range between 50°—56°, and typically be about 55°. The refractive indices were checked for grains of one rock (no. 6, Table 1): n_y, 1,708 ± 0.001 and n_z, 1,711 ± < 0.001. ($2V_x$ 50°). These optical data indicate hypersthene about En. 65—70.

Clinopyroxene (diopsidic pyroxene) is typically present in about the same amounts as hypersthene. It is colourless or very light greenish in thin sections. It is slightly pleochroic where coloured. $2V_z$ ranges between 50°—57°. The intermediate refractive index was determined in one case (no. 6, Table 1): $n_\beta = 1,693 ± 0.001$ ($2V_z$ 57°). This suggests Di. 70—80.

Hornblende is present in very small amounts or absent. It has the same appearance in all thin sections. ($2V_x = 62°$, $Z \wedge C = 18°$, pleochroism: X, yellow brown, Y, colourless — faint green, Z, greenish brown). It forms discrete grains with no preferential association and therefore, is probably not a secondary mineral.

Biotite is present in all specimens examined but only rarely in amounts more than about 2 per cent. It does not appear to be an alteration product of any other mineral. Pleochroism: X, colourless, Y, yellow brown, Z, yellow brown.

Magnetite is a rather important femic mineral and occurs in amounts similar to the pyroxenes.

Zircon and *apatite* are accessory minerals in the rocks.

The banded series (or charnockites proper).

Areally this is the most important rock series on Langøy. It covers the area from the porphyroblastic monzonites in the west to the metamorphic facies boundary in the east. However, within this area it is possible to delineate some rock types which do not belong to the banded series; these are described separately. These are: 1. the Eidet-Hovden intrusion in the northwest; 2. the retrograde metamorphosed zone of diverse gneisses which extends from Maars (in the far south of Bø parish) along the west side of Jørgenfjord to Ravnås and Kråkberget in the north; 3. the old red granite on the northwest side of

Eidsfjorden between Olderfjord and Grytting; 4. the young red granites most of which occur as small sporadic patches but also which constitute two major massifs (the Saltberggranite between Skjelfjord and Romsetfjord, and the north-south extending granite body between Skorpen (Jørgenfjord) and Rygge); 5. the anorthosites and 6. several highly metamorphosed sedimentary rocks and amphibolites that occur as bands within the series.

This section does not include considerations of the part of the series which is situated east of a line drawn from Steinlandsfjord across Sandskaret to Frøskeland and further along Eidsfjorden, i. e., the part close to the metamorphic boundary. (It should be noted that this is a completely arbitary division. The reason for separating this eastern boundary zone is explained later). Actually rocks different from the banded series rocks underlie only a small part of the total area (see map Fig. 3).

The banded series are charnockites proper. Vogt (1909, p. 16) draws attention to their veined character ("schlieren") which is especially conspicious along the shores and on highly weathered surfaces. He states that this veining (schlieren) commonly is so regular that one gets the impression that they are layered (sedimentary) rocks.

This series on Langøy appears to be essentially identical to the type charnockites of India. In fact, Pichamuthu's (1953, pp. 8 and 13) description of the Indian charnockites could be used without modification to describe the Langøy series. These rocks are banded (or veined) but the banding (or veining) cannot be traced for any considerable distance. The banding is everywhere parallel to the layering of sedimentary rocks which occur within the charnockites. The banded series rocks are not foliated. The quartz has a blue gray colour, and the feldspars are gray to brown. Thus even the silica rich rock types are dark and, in the field, are easily confused with the varieties possessing the mineral composition of typical norites (Holland, 1900, p. 124). The chemical and mineralogical difference between adjacent bands results in differential erosion and the banding is more noticeable on weathered surfaces. On fresh surface it is almost impossible to recognize any banding.

Typically the banded series rocks are equigranular. Locally the potash feldspars are larger than the other minerals; the rocks grade into the western monzonites. Typically, however, the feldspars in the potash feldspar rich banded series rocks are relatively small. The "type

Table 3.

Modal composition of the banded series rocks.

Minerals	1 69/56	2 61/56 b	3 249/55	4 355/55	5 47/56	6 58/56	7 145/56	8 137/55
Quartz........	–	–	–	–	–	tr.	–	–
Potash feldspar	–	–	12	∠1	19	4	–	28
2Vx			(51)		(44)			(40–45)
Plagioclase ...	52	32	65	66	55	70	46	51
An. in plag. ..	(45)	(67)	(42)	(40)	(34)	(35)	(47)	(35)
Hypersthene..	2	36	2	17	–	19	17	5
2Vx		(68)	(53–58)	(52)		(57)	(53–55)	(60)
%En.		(77)	(67)	(65)		(68)	(67)	(72)
Diopside	8	10	5	7	9	tr.	23	4
2Vz	(54)	(62)	(53)	(54–55)			(52)	(52–54)
Hornblende ..	18	–	–	–	–	–	10	–
2Vx	(75)						(74)	
Biotite	17	21	10	6	12	tr.	1	4
Ore	3	∠1	5	3	4	6	3	7
Apatite	tr.	–	1	tr.	1	tr.	tr.	2
Calcite	–	–	–	–	–	tr.	–	–
Zircon	–	–	–	–	–	–	–	–
Calculated mol per cent SiO₂..	42.6	44.0	47.5	48.7	48.8	48.9	49.4	49.5

Minerals	9 21/55	10 38/56	11 an.12	12 348/55	13 63/56	14 86/56	15 59/56	16 61/56a
Quartz.......	1	–	–	4	∠1	–	6	12
Potash feldspar	20	21	23	–	11	1	2	–
2Vx	(50)	(46)	(40)		(44)			
Plagioclase ...	52	60	59	72	77	92	76	57
An. in plag....	(30)	(32)	(30)	(38)	(38)	(28)	(34)	(38)
Hypersthene..	4	7	2	20	10	6	14	28
2Vx	(60)	(53–54)	(56–60)	(51)	(55)	(50)	(58)	(59)
%En	(72)	(67)	(70)	(65)	(68)	(63)	(70)	(71)
Diopside	11	5	7	–	∠1	–	–	–
2Vz	(57)	(53–54)	(56)	–	–	–	–	–
Hornblende ..	–	–	–	–	tr.	–	–	–
2Vx								
Biotite	7	2	3	2	tr.	tr.	tr.	∠1
Ore	5	4	5	2	1	1	2	2
Apatite	∠1	∠1	∠1	∠1	tr.	tr.	–	tr.
Calcite	–	–	–	–	–	–	tr.	–
Zircon	–	–	–	–	–	–	–	–
Calculated mol per cent SiO₂..	50.7	51.2	51.4	52.4	53.1	53.6	54.8	55.9

Table 3. (cont.).

Minerals	17 338/55	18 201/55	19 287/55	20 231/55	21 348/55	22 205/55	
Quartz........	12	14	15	25	14	30	
Potash feldspar	16	6	30	9	21	30	
$2V_x$	*(42)*		*(55)*	*(46)*	*(47)*	*(52)*	
Plagioclase ...	58	65	54	57	49	30	
An. in plag....	*(30)*	*(30)*	*(25)*	*(29)*	*(28)*	*(29)*	
Hypersthene..	7	12	–	5	12	5	
$2V_x$	*(60)*	*(57)*		*(54)*	*(57)*	*(58–65)*	
%En.........	72	68		67	68	73	
Diopside	–	–	-	–	–	–	
$2V_z$							
Hornblende ..	–	∠1	–	–	–	–	
$2V_x$							
Biotite	∠1	∠1	tr.	2	2	tr.	
Ore	6	2	∠1	2	2	5	
Apatite	tr.	tr.	–	tr.	tr.	∠1	
Calcite	–	tr.	–	–	–	–	
Zircon	–	–	–	–	tr.	–	
Calculated mol per cent SiO_2..	56.7	59.0	62.8	65.5	65.5	66.7	

Table 3, localities.

1. Close to Grytting, 2. West side of Melfjord, 3. Gåsland, 4. West of small lake close to road west of Gråholmene, 5. Uvåg, 6. Roadcut at Solheim, close to gabbro contact, 7. Top 228, north of Rekvann, 8. Between Selvåg and Eidet, 9. Top 76, Sandset, 10. Road towards Uvåg, 11. Roadcut Rotteåsen, 12. South end of Ryggepollen, 13. Strømmen, Olderfjord, 14. North of Langstrand, 15. Roadcut at Solheim, close to gabbro contact, 16. West side of Melfjord, 17. Dalbugten, 18. 300 metres above sea level, northeast of Åsentind, 19. Ridge southeast of Lynghaug, 20. Jennfjell, 21. South end of Ryggepollen, 22. Top 604, Vikanfjell.

specimen" of the potash feldspar rich type was taken from a roadcut at Rotteåsen, Eidet. It is chemically similar to the western monzonites (an. 12, Tables 3 and 4).

The modal mineral compositions of 22 rocks belonging to this series (point counted thin sections — one section of each rock) are given in Table 3. The rocks are arranged in series from basic to acid based on the SiO_2 content calculated from the modes. The rocks are well distributed over a wide silica range. Four of the rocks (nos. 6, 7, 11, 15 in Table 3) were chemically analysed. The chemical analyses, C. I. P. W. norms and calculated modes of these four rocks are given in Table 4. A close resemblance exists between the norm and observed mode for three of the rocks (nos. 58/56, an. 12, and 59/56). However, olivine is present in the norms and not in the modes. This may be cor-

Table 4.

Chemical analyses, C.I.P.W. (molecular) norms and calculated modes of four rocks from the banded series.

Weight %	58/56	145/56	an.12	59/56
SiO_2	48.87	51.85	54.62	59.27
TiO_2	2.14	0.69	0.95	0.42
Al_2O_3	16.92	13.23	18.54	17.80
Fe_2O_3	5.23	0.85	3.60	1.56
FeO	8.86	9.25	4.51	5.28
MnO	0.14	0.15	0.15	0.11
MgO	4.55	11.39	2.96	4.40
CaO	6.68	8.89	6.07	4.99
Na_2O	4.30	2.35	4.91	4.68
K_2O	0.87	0.81	3.00	1.17
H_2O^-	0.10	0.14	0.18	0.17
H_2O^+	0.23	0.23	0.18	0.18
CO_2	–	n.d.	–	–
P_2O_5	0.94	0.34	0.51	0.13
Sum	99.83	100.17	100.15	100.16

C. I. P. W. (molecular) norm.

	58/56	145/56	an.12	59/56
Quartz (q)	–	–	–	6.6
K-feldspar (or)	5.5	4.5	17.5	7.0
Na-feldspar (ab)	39.5	21.0	44.0	41.5
Ca-feldspar (an)	24.5	23.0	19.5	23.7
Olivine (ol)	4.5	8.4	4.5	–
Hypersthene (hy)	12.8	25.6	2.8	18.6
Diopside (di).............	2.8	14.8	5.6	–
Magnetite (mt)	5.6	0.9	3.8	1.7
Ilmenite (il).............	3.0	1.0	1.2	0.6
Apatite (ap)..............	1.8	0.8	1.1	0.3

Calculated mode.

	58/56	145/56	an.12	59/56
Quartz (q)	–	–	–	6.6
K-feldspar (or)...........	2.9	–	14.9	7.0
Na-feldspar (ab)..........	39.5	21.0	44.0	41.5
Ca-feldspar (an)	24.5	21.0	19.5	23.7
Hypersthene (hy)	15.8	30.4	5.8	18.6
Diopside (di)	2.8	14.8	5.6	–
Hornblende (hb)	–	3.0	–	–
Biotite (bi)...............	4.1	7.2	4.1	–
Magnetite (mt)	5.6	0.9	3.8	1.7
Ilmenite (il)	3.0	1.0	1.2	0.6
Apatite (ap)..............	1.8	0.8	1.1	0.3
analysed mol % SiO_2	46.1	47.5	50.4	54.5
calculated mol % SiO_2	48.9	49.4	51.4	54.8

(anal. 58/56, an.12, 59/56, E. Christensen, 145/56, L. B. Bolkesjø).

rected by introducing modal biotite (eq. lb., p. 29). In no. 59/56 all modal minerals are represented in the norm (except for a trace of biotite), but their relative abundances differ slightly. This is not surprising when it is remembered that the observed mode is based on the study of one thin section only and that the normative mineral formulae are simplified.

In the case of no. 145/56 recalculation of the norm into the modal minerals was more difficult. In order to obtain the relative abundances listed in Table 4 all the Or. was first recalculated into biotite (eq. lb. p. 29). The remaining surplus quartz was then accounted for by recalculating the necessary amounts of anorthite and hypersthene into silica and hornblende (eq. 4b p. 31). By doing this all the observed minerals are represented in the calculated mode; but, the ratios between the femic minerals Hy., Di., Hbl., and Bio. are wrong. This is probably chiefly the result of a much more complex hornblende composition than expressed in eq. 4.

Mol percentages of SiO_2 obtained from the chemical analyses and calculated from the mineral mode are given at the bottom of Table 4. A fair agreement between the two exists, especially within the more acid members of the series. That the analysed and calculated SiO_2 contents of the more basic rocks deviate is probably also mainly due to the presence of more complex femic minerals than those suggested by the formulae.

The variation of the rock forming minerals as plotted against the calculated mol per cent SiO_2 on a semi logarithmic scale is illustrated diagramatically in Fig. 15. The minerals show no regular variation with the SiO_2 content.

Quartz is present in all rocks with mol percentages of SiO_2 greater than 54. At lower SiO_2 concentrations it is present sporadically in small amounts, but it is typically absent, except as a constituent of myrmekite. The quartz has undulatory extinction and is of the same general size as the other minerals in these rocks. However, where the concentration is high (more than 20 per cent) the grains are commonly large. Because of their dark gray colour the quartz grains are not easily recognizable in handspecimen. Only in those rocks in which its concentration is above about 10 per cent is its presence easily ascertained.

Potash feldspars. It is only when the mol per cent of SiO_2 is above about 56 that potash feldspar is a typical mineral in the rocks. However, it is present in large quantities in some much more basic types as well.

The potash feldspars in this series are orthoclase perthites. The exsolved albite lamellae are normally of the "drop" type. "Mesoperthites" are locally
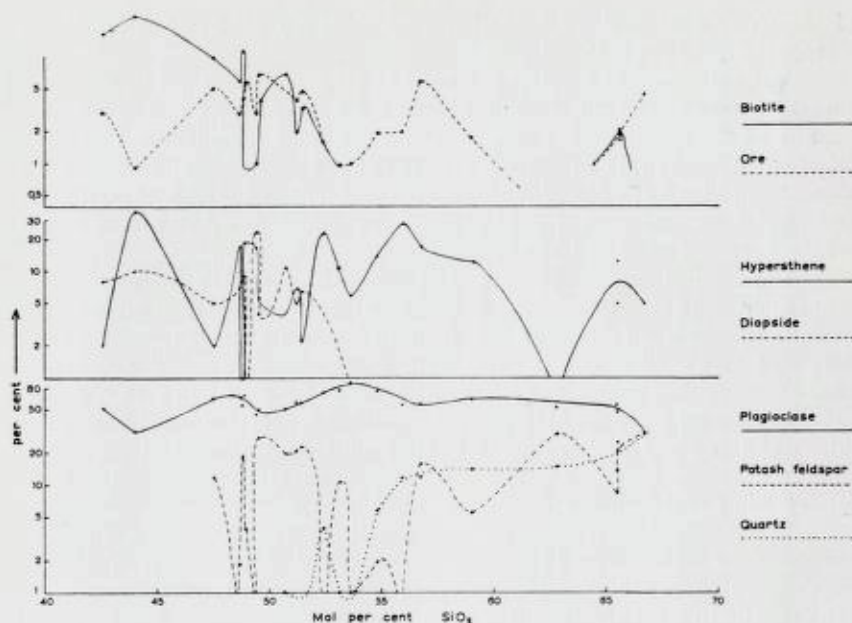
Fig. 15. Diagram showing the variations of the minerals in the banded series rocks vs. SiO₂ content calculated from the mode.

Diagram som viser mineralvariasjonene i bergartene fra "the banded series" plottet mot SiO₂ innholdet beregnet av moden.

present but are not typical. On the whole the feldspars are optically very similar to those in the western monzonites into which rocks of this series grade. The main difference between the two types is the very much larger size of the feldspars in the western monzonites. The feldspars are monoclinic, both optically as well as when measured on X-ray powder diagrams. The $2V_x$ for each specimen measured is listed in Table 3, it is typically smaller in the orthoclases of the banded series rocks than of the western monzonites (p. 31). The largest angles (nos. 19 and 22, Table 3) were obtained from feldspars from rocks directly adjacent to the western monzonites. Where potash feldspars are present in small amounts they typically have the form of small interstitial grains between plagioclase grains. Even where it is not present as independent grains it may occur as lamellae in plagioclase antiperthites.

Plagioclase is present in all of these rocks. Its concentration in the rocks shows no regular variation with the SiO₂ content (Table 3 and Fig. 15). The plagioclase composition as determined optically is remarkably constant over a wide silica range, but tends to be somewhat more anorthite rich in some of the lowest silica members. The slight increase in anorthite content (An₃₈) in nos. 12 and 13 is believed to be particularly interesting because the SiO₂ content

of these rocks reaches the level where diopside, the other major Ca mineral, disappears from the mode.

Antiperthites are rather typical except in the most anorthite rich varieties. However, even within the same thin section the antiperthitic nature cannot be seen in all the plagioclase grains, even with the same orientations. The potash feldspar lamellae of the antiperthites typically have rectangular forms with their elongations parallel to (010). Where plagioclase contacts potash feldspars myrmekite is commonly seen to penetrate into the plagioclase. Apart from this the plagioclases are clear and free of inclusions.

The plagioclase compositions were obtained by normal Fedorov methods on the universal stage. During the course of these determinations it was possible to establish the nature of the different twins. Twinning according to the following laws were proved to be present; albite, albite/ala, pericline and manebach. The possible presence of other twins was not excluded and no systematic determinations of the twin laws were undertaken. For the same reason it is difficult to give any reliable estimates of the relative anbudances of the different twins. It seems, however, that albite and albite/ala twins predominate (the former being slightly more frequent than the latter).

The plagioclases are low temperature types, though some of the stereographic plots do deviate from the low temperature composition curves published by Trøger (1952). (This may, however, in several cases be ascribed to erroneous readings. In any case high and low temperature plagioclases of this composition are difficult to distinguish optically).

Hypersthene is the one essential dark mineral throughout the charnockite series (Howie, 1955, p. 751). Only two out of the 22 samples listed in Table 3 contain no hypersthene.

The hypersthenes are typically completely unaltered in these rocks and are of the same granular form as the other minerals. They are colourless or very faintly coloured and slightly pleochroic in thin sections. A certain relationship between $2V_x$ and colour exists. In Table 3 nos.: 1, 3, 6, 8, 9, 11, 15, 18, 19, 22, (range of $2V_x$ 57°—65° average 59°) are colourless; nos. 2, 10, 13, 17, 20 are pleochroic: X, faint reddish brown, Y and Z, colourless, (range of $2V_x$ 53° 68°, average 58°. Excluding no. 2 ($2V_x$ 68°) they average 55°); nos. 4, 7, 12, 14, 16, 21 (range of $2V_x$ 50°—59°, average 54°) pleochroic: X, faint reddish brown, Y and Z, very weak yellowish green. It is apparent from Table 4 that the hypersthene composition can be calculated fairly accurately from the rock analyses in the cases of nos. 58/56 and 59/56 (Table 5). The table shows that the normative hypersthene composition tends to be more iron rich than what is deduced from the size of the 2V. The same trend was shown by Howie (1955, Table 8, p. 754).

The chemical composition as indicated by the size of $2V_x$ is given in Table 3.

Clinopyroxene (diopsidic pyroxene) is a characteristic mineral in the silica deficient members of the series, and is not present in the rocks with excess silica, *i. e.,* rocks containing free quartz in the mode (Table 3).

The clinopyroxenes may be seen in thin sections to be colourless to very weak greenish. $2V_z$ ranges from 52°—62°, but is typically 53°—54°. The size of the 2V suggests the clinopyroxenes to have a higher Mg/Fe ratio than the

Table 5.

*Normative composition of hypersthene as
calculated from rock analyses (a) and
from size of $2V_x$ (b) (Trøger, 1952, p. 59).*

Rock no.	a mol per cent En.	b mol per cent En.
58/56	62.5	68
59/56	65.6	70

coexisting orthopyroxenes. This was confirmed in the cases where refractive indices were checked.

Hornblende, a rare mineral in these rocks, is only present as a major mineral in two of the more basic varieties of Table 3. The optical data are: no. 1, X-light yellow green, Y-yellow green, Z-green; $2V_x$ 75°; no. 7, X-very light green, Y-brown green, Z-brown green, $2V_x$ 74°.

Biotite has an interesting distribution in these rocks. It is a major mineral in the more basic varieties (nos. 1—5), and is down to a minimum (nos. 6—7) and up again to a smaller maximum in the intermediate potash feldspar rich types (nos. 8—11). Otherwise its concentration is low, but shows a tendency to be relatively higher in the extremely silica rich types. Biotite occurs as an independent primary mineral. The pleochroism varies as follows:
X-colourless, Y-yellow brown, Z-light yeollow (nos. 1, 2, 12, 13, 17, 20)
X-light yellow, Y-brown, Z-yellow brown (nos. 3, 7, 9, 10, 21)
X-light yellow green, Y, Z-yellow green (nos. 5, 8, 14)
X-light yellow green, Y-strong brown, Z-yellow brown (no. 4).

The remaining minerals in these rocks will not be discussed. Their distribution can be read from Table 3. It should be mentioned however, that the calcite traces reported in nos. 6, 15, 18 probably are secondary and related to the Eidet-Hovden gabbro intrusion.

The charnockite border series.

This designation is used for rocks of the eastern part of the banded series. The series extends southwards to the anorthosite of the thrust plate over Slåttnestind and eastward to the metamorphic boundary (map, Fig. 3); a line from Steinlandsfjord, across Sandskartind to Frøskeland and further southwest along Eidsfjorden defines the western boundary.

Actually the western border is quite arbitrary as the banded series and the charnockite border series grade into each other and petrologically they should be regarded as one unit. The rocks within the border

series have their exact counterparts in the banded series. However, one important difference exists. The banded series is characterized by rocks rather evenly distributed over a wide silica range (see p. 36) and except for extreme cases it is difficult to distinguish between the diverse types in the field. On the other hand the charnockite border series tends to be marked by the separation of the dark (basic, amphibolitic) and light (acid, granitic) bands or veins which can be easily distinguished in the field. Modal analyses of 21 samples (point counted thin sections) are presented in Table 6.

It is obvious from this table that the rocks are not evenly distributed over the whole silica range as they were in the banded series (Table 3) but are more confined to two groups, one silica deficient and one with excess silica, i. e., in contrast to the rocks of the banded series the intermediate rock types are lacking in the charnockite border series. This tendency is still more pronounced in the amphibolite facies gneisses and is discussed later (p. 126).

Chemical analyses, calculated norms and modes of six chemically analysed rocks from this series are given in Table 7. It may be seen that the calculated modes deviate rather markedly from the observed and that the deviation is especially great in the acid members. This has its explanation in the very nature of these rocks. Although the dark and light bands tend to be better separated than in the banded series rocks, the light members especially are rather inhomogeneous, in that thin, irregular dark bands (mostly biotite) commonly occur within them. Therefore, the counted modes should not be considered as representative of mineral concentrations for any considerable bulk of rock. However, they do give a correct picture of the mineral parageneses into which the chemical analyses should be calculated (the analysed samples are more representative of the light bands including the biotitic parts than are the thin sections). General and textural remarks are omitted in the following mineral description as those given in the description of the banded series pertain. Features which serve to distinguish the border series rocks from the banded series rocks are stressed here.

The quartz is not of the same dark colour as in the rocks of the banded series. Otherwise it is similar.

Potash feldspar of these rocks is also generally an orthoclase (optically monoclinic). Undulatory extinction was observed in nos. 8, 10, 11 and 18 in Table 6, although in nos. 10 and 18 this feature is not very marked. This and

Table 6.

Modal composition of the charnockite border series rocks.

Minerals	1 160/56	2 148/56 b	3 124/55	4 151/56 b	5 150/56 b	6 102/55	7 55/56	8 138/56
Quartz	-	1	-	-	5	-	-	4
Potash feldspar	-	-	-	-	-	-	-	13
2Vx								(40–69)
Plagioclase	13	32	32	46	39	65	63	45
An. in plag	(68)	(40)	(52)	(50)	(53)	(41)	(36)	(32)
Hypersthene	-	tr.	10	20	20	4	2	7
2Vx			(60)	(52–55)	(59)	(52)	(49)	(48)
%En			(71)	(67)	(70)	(65)	(62)	(61)
Diopside	19	6	21	6	11	13	16	8
2Vz	(52–55)	(56)	(58)			(55)	(53–56)	(54)
Hornblende	46	59	34	9	tr.	9	-	18
2Vx	(79)	(68–70)	(75)	(73)		(70)		(61–67)
Biotite	22	tr.	tr.	16	23	10	18	2
Ore	tr.	2	3	2	2	∠1	1	3
Apatite	tr.	-	tr.	∠1	tr.	tr.	tr.	tr.
Calcite	-	-	tr.	-	-	-	-	-
Zircon	-	-	-	-	-	-	-	-
Spinel?	-	-	-	-	-	-	-	-
Calculated mol per cent SiO₂	38.3	41.6	43.5	43.8	46.5	47.6	47.7	49.9

Minerals	9 7/56	10 141/56	11 56/55	12 158/56	13 152/56	14 148/56a	15 150/56a	16 28/55
Quartz	1	1	8	19	23	27	20	24
Potash feldspar	19	32	12	10	-	10	18	5
2Vx	(46–50)	(51–55)	(52–62)	(42–45)		(60)	(50)	(40)
Plagioclase	61	47	57	49	60	45	50	59
An. in plag	(48)	(27)	(30)	(33)	(41)	(33)	(28)	(29)
Hypersthene	-	8	-	8	10	2	3	2
2Vx		(60)		(51)	(52)			(61)
%En		(72)		(50)	(48)			(38)
Diopside	∠1	-	5	-	-	tr.	-	-
2Vz			(51–53)					
Hornblende	8	-	tr.	-	-	-	-	-
2Vx	(65–67)							
Biotite	6	8	14	13	7	14	9	9
Ore	3	4	3	1	tr.	tr.	tr.	1
Apatite	∠1	-	1	-	tr.	-	tr.	tr.
Calcite	-	-	-	-	-	-	-	tr.
Zircon	-	-	-	tr.	-	-	-	-
Spinel?	-	tr.	-	-	-	-	-	-
Calculated mol per cent SiO₂	51.0	52.4	52.7	59.1	61.3	62.2	62.2	62.4

Table 6 (cont.).

Minerals	17 25/55	18 27/55	19 142/56	20 156/56	21 151/56a	
Quartz.......	21	20	29	31	59	
Potash feldspar	43	37	5	4	2	
2V$_x$	(46)	(53)	(40)	(40–45)	(50)	
Plagioclase ...	26	37	48	59	33	
An. in plag....	(31)	(26)	(32)	(32)	(30)	
Hypersthene..	–	2	11	2	4	
2V$_x$		(62)	(55–58)	(60)	(60)	
%En.........		(38)	(45)	(40)	(40)	
Diopside	–	–	–	–	–	
2V$_z$						
Hornblende ..	–	–	–	–	–	
2V$_x$						
Biotite	8	5	5	3	2	
Ore	1	∠1	2	∠1	tr.	
Apatite	∠1	tr.	–	–	–	
Calcite	–	–	–	tr.	–	
Zircon	–	tr.	tr.	–	–	
Spinel?	–	–	–	–	–	
Calculated mol per cent SiO$_2$..	63.7	64.5	64.7	66.9	80.6	

Table 6, localities.

1. Roadjunction, Frøskeland, 2. Close to top 421, 330 m.a.s.l., west of Gåsebøl, 3. Roadcut, Sildpollen, 4. Roadcut, small lake north of Storvann, Frøskeland—Steinlandsfjord, 5. West of top 547, between Storvann and Gåsfjord, 6. Top 455, Sandskaret, 7. Between Trollvann and Hornvann, 8. 200 m. east of road, Nyg, 9. Small lake east of Hornet, 10. Roadcut, southeast end of Valfjord, 11. Hornet, 12. Roadcut 1000 m. south of Storvann, road between Frøskeland and Steinlandsfjord, 13. Sandskaret, 14. Close to top 421, 330 m.a.s.l., west of Gåsebøl, 15. West of top 547, between Storvann and Gåsfjord, 16. Roadcut, 1 km. north of Åsen, Steinlandsfjord, 17. Roadcut, Frøskeland, 18. Storbringen—Natmålsfjell, 19. Roadcut, Sildpollen, 20. Roadcut, south end of Storvann, road between Frøskeland and Steinlandsfjord, 21. Roadcut, small lake north of Storvann, Frøskeland—Steinlandsfjord.

the large 2V$_x$ determined for these suggest that they are more triclinic than the typical feldspars of these rocks (*i.e.*, they are intermediate microclines characterized by varying triclinicities from spot to spot within the same "crystal", — Goldsmith and Laves, 1954 b; Heier 1957). Typical microcline twinning was not observed in any of these except possibly in no. 11. The feldspar in no. 14 (2V$_x$ 60°) shows straight extinction and is optically monoclinic. It is of interest to note that nos. 8, 10, 11 and 14 are from rocks close to the metamorphic facies boundary. Chemical analyses of feldspars from nos. 15 and 17 are discussed later.

Plagioclase is similar to that in the banded series rocks. The normative plagioclase composition of the three silica rich rocks in Table 7 is more albitic

Table 7.

*Chemical analyses, C.I.P.W. (molecular) norms and calculated
modes of six rocks from the charnockite border series.*

Weight %	151/56 b	150/56 b	158/56	150/56 a	142/56	151/56 a
SiO_2	49.20	51.37	57.44	67.98	62.79	72.26
TiO_2	1.16	1.11	0.83	0.43	0.33	0.22
Al_2O_3	15.02	14.94	18.33	16.06	17.52	14.41
Fe_2O_3	1.19	0.92	1.46	1.24	1.28	0.52
FeO	9.08	8.36	7.08	2.42	4.23	1.26
MnO	0.22	0.14	0.13	0.02	0.07	0.03
MgO	10.26	11.08	4.05	1.47	2.32	0.69
CaO	8.52	7.71	3.76	1.84	3.00	2.03
Na_2O	2.37	1.41	3.92	3.60	4.06	3.26
K_2O	1.58	1.99	2.89	4.73	3.96	4.83
H_2O^-	0.13	0.13	0.14	0.18	0.12	0.05
H_2O^+	0.42	0.46	0.42	0.32	0.22	0.24
CO_2	n.d.	–	n.d.	n.d.	–	n.d.
P_2O_5	0.70	0.79	0.04	0.10	0.08	0.12
Sum	99.85	100.41	100.49	100.39	99.98	99.92
C.I.P.W. (molecular) norm						
Quartz (q)	–	0.2	5.0	20.5	10.6	27.4
K-feldspar (or)	9.5	11.5	17.0	28.0	23.5	29.0
Na-feldspar (ab)...	21.0	12.5	35.0	32.5	36.5	29.5
Ca-feldspar (an)...	25.8	28.5	17.5	8.0	14.0	9.0
Corundum (c).....	–	–	2.5	2.3	1.5	0.6
Olivine (ol)	17.1	–	–	–	–	–
Hypersthene (hy)..	13.4	39.4	20.0	6.4	11.8	3.2
Diopside (di)......	8.8	3.6	–	–	–	–
Magnetite (mt)	1.2	1.1	1.5	1.4	1.4	0.6
Ilmenite (il)	1.6	1.6	1.2	0.6	0.4	0.4
Apatite (ap).......	1.6	1.6	0.3	0.3	0.3	0.3
Calculated mode						
Quartz (q)	–	6.7	7.3	22.8	10.4	27.4
K-feldspar (or)	–		6.5	18.0	19.8	27.4
Na-feldspar (ab) ...	21.0	12.5	35.0	32.5	36.5	29.5
Ca-feldspar (an) ...	22.8	28.5	17.5	8.0	14.0	9.0
Hypersthene (hy) ..	23.5	26.4	15.5	1.7	12.1	3.2
Diopside (di)......	8.8	3.6	–	–	–	–
Hornblende (hb)...	5.0	–	–	–	–	–
Biotite (bi)	14.8	17.9	6.5	6.7	–	–
Muscovite (mu) ...	–	–	8.8	8.1	5.3	2.1
Magnetite (mt)	1.2	1.1	1.5	1.4	1.4	0.6
Ilmenite (il)	1.6	1.6	1.2	0.6	0.4	0.4
Apatite (ap).......	1.6	1.6	0.3	0.3	0.3	0.3
analysed mol % SiO_2	45.4	47.5	53.2	63.2	58.1	67.7
calc. mol % SiO_2 ..	43.8	46.5	59.1	62.2	64.7	80.6

(anal. L. B. Bolkesjø, no. 142/56 by E. Christensen).

Table 8.

*Normative composition of hypersthene as
calculated from the rock analyses (a) and
from the size of $2V_x$ (b) (Trøger, 1952, p. 59).*

Rock no.	a mol per cent En.	b mol per cent En.
151/56 b......	70	67
150/56 b......	74	70
158/56	56	50
150/56 a......	66	–
142/56	54	45
151/56 a......	63	72 (or 40)

than determined optically. All of the normative albite should not, however, be calculated as plagioclase because a substantial amount of it forms alkali feldspar with orthoclase. Twinning after the following laws was observed: albite, pericline, normal carlsbad, normal manebach, albite/carlsbad and albite/ala. Normal albite- and albite/ala twins were dominant and present in about equal amounts.

Hypersthene is a stable mineral throughout the series and shows no signs of secondary alteration. The pleochroism is as follows:

X, Y, Z, colourles — ($2V_x = 48°$—$52°$, average $50°$, nos. 6, 7, 8, 13).

X, reddish brown, Y, colourless Z, light greenish ($2V_x = 51°$—$61°$, average $57°$, nos. 4, 12, 16, 19, 20, 21),

X light reddish brown, Y, Z, colourless ($2V_x = 59°$—$62°$, average $60°$, nos. 3, 5, 10, 18).

This is an opposite relation between optic axial angle and colour from that observed in the banded series rock. The compositions of the normative hypersthenes in Table 7 have been calculated (assuming the same Fe/Mg ratios in all the normative femic minerals). In Table 8 these are compared with the compositions derived from the size of the $2V_x$. Contrary to what was found for the hypersthenes in the banded series rocks (Table 5) the normative hyperthene composition in the charnockite border series rocks tends to be more Mg-rich than is deduced from the size of the 2V.

The agreement is fairly good except in the case of no. 21 (151/56 a). The optically derived compositions are probably more nearly correct as the other femic minerals, especially biotite, probably have a higher Mg/Fe ratio than hypersthene. Tentative chemical compositions based on the $2V_x$ are given in Table 6.

Clinopyroxene (diopside) is optically similar to that in the banded series rocks. It also shows the same distribution pattern, *i. e.*, it is absent where quartz becomes a major mineral phase (i. e., in excess silica rocks).

Hornblende is a more consistent mineral in the silica deficient rocks of this series than in the banded series rocks. Its distribution pattern is similar to that of

diopside in that it also is absent in rocks with excess silica (*i. e.*, rocks in which free quartz is present in appreciable quantities).
It is pleochroic as follows:
X-weak light green, Y-green to brownish green, Z-light green to green (nos. 2, 3, 4, 6).
X-colourless to very light green, Y, Z-light green (nos. 1, 8, 9, 11).
The general relationship is that colour decreases with decreasing $2V_x$ and increasing silica content of the rocks (except no. 1).

Biotite is a common mineral throughout the whole series and is generally present in larger amounts than in corresponding rock types of the banded series. Corundum is a normative mineral in rocks with excess silica (Table 7) which suggests that the biotite is Al rich. In the calculated modes excess Al is accounted for by muscovite:

$$K(Mg,Fe)_3AlSi_3O_{10}(OH)_2 + 3SiO_2 + Al_2O_3 \rightarrow \qquad\qquad 5$$
$$KAl_2AlSi_3O_{10}(OH)_2 + 3(Mg,Fe)SiO_3$$
$$4Bi. + 1.5 Q. + 1Corund = 3,5 Musc. + 3 Hy. \qquad\qquad 5 b$$
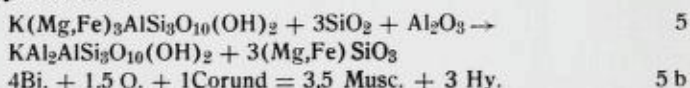
The muscovite and the biotite in the calculated mode constitute the mica in the rocks. No muscovite occurs in the rocks. In this respect the calculated modes for nos. (142/56) and (151/56 a) Table 7 are misleading, but they do serve to illustrate the point that at least the silica rich types of this series contain an Al rich biotite. The pleochroism of the biotites vary in degree.
X-colourless, Y, Z-light brown (nos. 3, 10, 13, 17);
X-colourles, Y-brown, Z-yellow brown (nos. 1, 2, 5, 8, 11, 15, 16, 18, 19, 20, 21);
X-light yellow, Y-green brown, Z-light green brown (nos. 9, 14);
X-light yellow, Y-dark brown, Z-dark brown to green brown (nos. 4, 6, 12).

The remaining minerals will not be discussed. Their distribution is evident from Table 6. The "spinel" in no. 10 was not identified definitely. It is a green isotropic mineral which was seen only as one or two grains.

The amphibolite facies veined gneisses.

These gneisses extend eastward from the metamorphic boundary which separates them from the similarly appearing charnockite border series rocks. They grade into the red augengneisses in the southwest, south of a line between Sortland and Holmstad. Certain metamorphic sedimentary rocks (similar to those in the banded series and the charnockite border series) also are common within these gneisses. They, and amphibolites occurring with them, are considered separately. Their distribution may be seen on map Fig. 3. The amphibolites described in this chapter are mineralogically and chemically similar to those occurring with the known metasedimentary rocks. Nontheless, they occur enclosed in the gneisses without any obvious relation to the other. On

the other hand the part of this area across Ånstadblåheia, Karihaugen to Jægtbøgen can be characterized as being underlain by coarse granodioritic gneisses; amphibolites are less common here and where present have the form of relatively large lenses and bands. In other words, these are not so thoroughly mixed as are the banded series and the border series. (This is even more evident on Hinnøy). As the border series is approached from the east the amphibolite facies gneisses are progressively more mixed and become similar in appearance to the border series rocks.

At Gåsebøl, in the north of the mapped area, garnets occur in the gneisses. Apart from some places in the banded series at its contact with the old red granite this is the only place where garnets have been observed.

South of the road between Ånstad and Holmstad the gneisses are more thoroughly mixed and are similar in appearance to the banded series rocks. Red potash feldspars and quartz-feldspar veins are very common in this southern part. South of the metalimestone and associated "gabbro" band that strike eastward from Sortland the amphibolite facies veined gneisses grade into the red gneisses and augengneisses of the southwest. 13 modal analyses derived from point counted thin sections are given in Table 9.

Chemical analyses, C. I. P. W. (molecular) norms and the mesonorms (Barth, Calculation of the common mesonorm of rocks with excess SiO_2 — unpublished[1]) of the rocks within these gneisses which are chemically analysed are given in Table 10. In the mesonorm excess Al_2O_3 is listed as corundum. This is not present in any of the rocks examined and it, together with the corresponding amount of Or., was calculated as muscovite in Table 10. It forms the "mica" together with the normative biotite. It is evident from the table that the C. I. P. W. norms have no resemblance whatsoever to the actual mineral composition of the rocks. The mesonorms describe the rocks much better. No significance should be given to the deviation in the mineral percentages between the observed and calculated mode (on the basis of the mesonorm method). Because a modal analysis of one thin section of such heterogeneous rocks as these cannot be expected to express the correct

[1] Barth, T. F. W. (1959)Principles of classification and norm calculations of metamorphic rocks. J. Geol., 67, pp. 135—152, appeared after this paper had gone to press.

Table 9.

Modal composition of the amphibolite facies veined gneisses.

Minerals	1 3/56b	2 147/56b	3 8/55b	4 13/55a	5 36/55	6 8/55a	7 147/56a	8 37/55
Quartz	–	1	1	2	7	2	8	18
Potash feldspar	–	–	9	3	–	21	28	22
$2V_x$			(66–68)	(60–84)		(56–62)	(70–80)	(60–72)
Plagioclase	45	42	48	49	55	63	48	44
An. in plag.	(30)	(33)	(32)	(25)	(25)	(35)	(28)	(26)
Hornblende	33	53	21	40	17	6	–	1
$2V_x$	(65)	(70)	(60)	(74)	(58)			
Biotite	18	3	14	4	19	7	16	13
Ore	4	1	7	2	3	1	tr.	tr.
Titanite	–	∠1	–	tr.	tr.	–	–	∠1
Apatite	tr.	∠1	–	tr.	tr.	tr.	–	tr.
Calcite	–	–	–	–	–	tr.	–	∠1
Calculated mol per cent SiO_2	41.7	42.3	43.5	46.3	48.2	52.2	55.7	60.3

Minerals	9 14/56	10 61/55	11 35/55	12 an.1	13 3/56a
Quartz	24	24	19	31	35
Potash feldspar	10	–	32	31	31
$2V_x$	(69–76)		(55–80)	(72–81)	(60)
Plagioclase	45	58	38	27	27
An. in plag.	(30)	(32)	(20)	(25)	(28)
Hornblende	–	∠1	2	–	–
$2V_x$					
Biotite	21	17	7	11	7
Ore	tr.	tr.	2	tr.	tr.
Titanite	tr.	tr.	tr.	tr.	–
Apatite	tr.	tr.	tr.	–	–
Calcite	–	–	–	tr.	–
Calculated mol per cent SiO_2	60.6	60.8	62.3	67.6	70.4

Table 9, localities.

1. Amphibolite in gneiss, Lafjellet, 2. Amphibolite in gneiss, 120 m. a. s. l., Gåsebøl, 3. Amphibolite, Sandstrand, 4. Amphibolite, brook between Litlholmen and Storholmen, 5. Amphibolite in gneiss, Ånstadblåheia, 6. Fine grained gneiss, Sandstrand, 7. Light gneiss, 120 m. a. s. l., Gåsebøl, 8. Gneiss, Karihaugen, 9. Gneiss, Steirodalen, 10. Gneiss, Litlvannet, 11. Gneiss, Ånstadblåheia, 12. Gray gneiss, quarry, Jægtbøgen, 13. Light gneiss, Lafjellet.

percentage, probably the calculated mode based on a chemical analysis of a large and thus more representative sample is more nearly correct. The mesonorm was constructed especially for rocks with excess SiO_2 and when used on rocks where silica is deficient (3/56 b, 147/56 b,

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Table 10.

Chemical analyses, C.I.P.W. (molecular) norms and mesonorms
of five rocks from the amphibolite facies veined gneisses.

Weight %	3/56b	147/56b	147/56a	an. 1	3/56a
SiO$_2$	46.87	49.45	61.80	67.58	73.37
TiO$_2$	1.81	0.72	0.40	0.36	0.29
Al$_2$O$_3$	16.75	17.13	18.91	15.69	14.37
Fe$_2$O$_3$	4.45	2.87	1.00	0.50	0.58
FeO	9.29	7.15	3.49	3.65	1.52
MnO	0.15	0.14	0.01	0.02	0.01
MgO	5.55	7.24	1.68	1.78	0.43
CaO	7.69	9.64	2.88	2.11	2.04
Na$_2$O	3.55	3.48	4.37	3.89	3.23
K$_2$O	2.26	0.89	4.45	3.79	3.96
H$_2$O⁻	0.17	0.15	0.08	0.08	0.11
H$_2$O⁺	0.99	0.99	0.67	0.48	0.13
CO$_2$	n.d.	n.d.	n.d.	0.08	n.d.
P$_2$O$_5$	0.70	0.11	0.55	tr.	0.09
Sum	100.23	99.96	100.29	100.01	100.13

C. I. P. W. (molecular) norm

Quartz (q)	–	–	9.3	19.8	32.2
K-feldspar (or)	13.5	5.5	26.5	22.5	23.5
Na-feldspar (ab)	26.7	30.5	39.0	35.0	29.5
Ca-feldspar (an)	23.5	28.5	10.5	10.5	9.0
Nepheline (ne)	3.5	0.6	–	–	–
Corundum (c)	–	–	3.3	1.5	1.7
Olivine (ol)	15.3	15.8	–	–	–
Hypersthene (hy)	–	–	8.6	9.6	2.8
Diopside (di)	8.8	14.8	–	–	–
Magnetite (mt)	4.8	3.0	1.1	0.5	0.6
Ilmenite (il)	2.6	1.0	0.6	0.6	0.4
Apatite (ap)	1.3	0.3	1.1	tr.	0.3

Mesonorm

Quartz (q)	–	–	13.6	24.6	33.6
K-feldspar (or)	7.5	2.3	11.1	10.7	16.8
Na-feldspar (ab)	32.5	31.5	39.0	35.0	29.5
Ca-feldspar (an)	12.5	12.5	10.5	10.5	9.0
(% An. in plag.)	(28)	(28)	(21)	(27)	(23)
Hornblende (hb)	33.0	48.0	–	–	–
Biotite (bi)	9.6	5.1	11.5	12.8	3.7
Muscovite (mu)	–	–	11.5	5.3	5.9
Titanite (ti)	–	1.5	–	–	–
Magnetite (mt)	4.8	3.0	1.1	0.5	0.6
Ilmenite (il)	2.6	–	0.6	0.6	0.4
Apatite (ap)	1.3	0.3	1.1	tr.	0.3

In the mesonorms of nos. 3/56b and 147/56b it was used 3.8 % and 4.2 % SiO$_2$ respectively in excess of the silica concentration given by the chemical analyses.

analysed mol % SiO$_2$	44.2	45.9	51.7	63.3	69.0
calc. mol % SiO$_2$	41.7	42.3	55.7	67.6	70.4

(anal. L. B. Bolkesjø, no. an. 1, E. Christensen).

Table 10) too much SiO_2 will be used in the early formed minerals. Nevertheless even in low silica rocks it lists the modal minerals in remarkably correct proportions. Further the Ab./An. ratio corresponds closely to the measured plagioclase composition. Though potash feldspar is not present in the observed mode whereas it is in the mesonorm, it is (at least in no. 147/56 b) not present in the mesonorm in any significantly higher concentrations than might be expected to enter the plagioclase lattice. It does not appear possible to construct any reasonable silicate mineral reaction which will account for the extra SiO_2 used in calculating the mesonorms in nos. 3/56 b and 147/56 b and which will still be in reasonable agreement with the mode. The only solution appears to be that Ti^{4+} and Fe^{3+} (in the norms calculated as ilmenite and magnetite respectively) enter the biotite and/or hornblende lattices, thus allowing more Al^{3+} to substitute for Si^{4+}. That this probably is the case is suggested by the lower percentage of modal "ore" than normative magnetite and ilmenite (compare with Tables 1 and 2, 3 and 4, 6 and 7). This point will be elaborated further in the chapter dealing with the metamorphism of the area. Table 9 illustrates the same irregular mineral variations as noted in the hypersthene gneisses.

Quartz is commonly present in small amounts even in the amphibolitic types of these gneisses. It is characterized by undulatory extinction. Especially where present in somewhat high concentrations the grain size is large. It is clear and colourless throughout and never of the darkish colour typical of this mineral in the banded series.

Potash feldspar is typically a microcline perthite in these rocks. The triclinic nature of these feldspars is suggested in Table 9 by the sizes of $2V_x$. Some of the variations of 2V indicated in the table were measured on different parts of the same grain which suggest that the feldspars are of the intermediate triclinicity type (Goldsmith and Laves, 1954 b, Heier, 1957).

The microcline twinning is normally present. Nos. 3, 6 and 13 in Table 9 are the only cases in which the feldspars appear to be optically monoclinic in thin sections and these have the smallest 2V's of the feldspars measured. Even so their values also suggest a triclinic symmetry and it may be stated safely that all of the potash feldspars in these rocks are microclines.

The size of the feldspars ranges from very small "interstitial" grains to large crystals which can be recognized easily in hand specimen.

On weathered surface some of the potash feldspars are reddish, but typically they have a clear gray-white colour.

In thin sections they may be seen to be unaltered and the only inclusions are exsolved albite lamellae.

Partial chemical analyses of some microclines from within this series (nos. 147/56 a, 37/55, 35/55, an. l, 3/56 a) are discussed in a later chapter.

Plagioclase is present in rocks of all parts of the silica range just as in the rocks from the banded series and the charnockite border series. It differs from these by being more albitic. This is especially marked in the basic (amphibolitic) types, but is also generally evident in the relatively silica rich varieties. Some of the plagioclases show signs of having been slightly saussuritized (nos. 2, 4, 8, 9, 11, 12 Table 9); this alteration is nowhere great. The plagioclase in no. 2 (An. 33) is most altered. At the calculated mol per cent SiO_2 of these rocks where hornblende is about to disappear, the An. content of the plagioclase is slightly higher than typical. This is similar to the diopside: An. content relationship of the banded series rocks. Plagioclase and hornblende are the only important Ca-containing minerals in these rocks. Albite, pericline, albite/ala and carlsbad twins occur. (The presence of other types of twins was not excluded). The impression is that pericline twins are most frequent with albite and albite/ala next in abundance.

Myrmekite quartz spindles are common within plagioclase next to microcline borders.

Hornblende is a common mineral in the amphibolitic rocks and also is sporadically present in the silica rich gneisses. As a rule, however, it is absent where quartz is a major mineral. The $2V_x$ varies from $58°—74°$ and a very slight change in the pleochroism can be related to this variation;

X-colourless, Y, Z-light green (no. 4, $2V_x$ 74°);

X-colourless to very light green, Y-light green, Z-green (no. 2, $2V_x$ 70°);

X-light green, Y-dark green, Z-light green (nos. 1, 3, 5, 6, 11, $2V_x$ 58°—65°).

The slight increase in colour together with the decrease of the $2V_x$ may be related to an increase of iron in the same direction.

Biotite is present in appreciable concentrations over the whole silica range. It shows varying pleochroism;

X-light yellow, Y, Z-deep reddish brown (no. 1);

X-colourless, Y,Z-light brown (nos. 2, 3, 4, 5);

X-very light green, Y,Z-brown green to green (nos. 6, 7, 8, 9, 10, 11, 12, 13).

As is apparent the colour varies systematically from the basic to the acid rocks. Refractive index (n_z) was determined for three of them:

$n_z = 1,630 \pm 0.003$ (nos. 7, 12), $n_z = 1,645 \pm 0.003$ (no. 8).

It is difficult to relate these optical data to definite chemical compositions. However, the red brown colour of no. 1 suggests a rather high concentration of TiO_2 in the lattice.

This agrees with the deductions made from the normative calculations (p. 51).

The remaining minerals will not be discussed. Of importance is the fact that titanite is commonly present in small amounts and that the observed "ore" minerals[1] are always present in significantly smaller amounts than the normative magnetite and ilmenite (p. 51).

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[1] The names "ore" and "iron ore" are used for the opaque minerals observed in thin sections of the rocks. The "ore" has been shown to be magnetite in the cases where the rocks have been ground for mineral separations.

The old red granite.

This rock type occurs within the banded series and underlies a small area between Daljorden and Hokkabogen (Grytting) on the west side of Eidsfjorden. Towards the west it extends towards top 574. Its approximate extent is indicated on the map Fig. 3. As far west as Olderfjordvann (north of the north end of Olderfjorden) it may be traced as pegmatite-like veins in the banded series to which it gives a migmatitic appearance. As the main granite mass is approached along the carriageroad between Sølferstranden and Grytting the coarse red granitic material gradually becomes more abundant. Thus the granite is surrounded by a widespread migmatite aureole. Even within the main part of the granite inclusions of basic (amphibolitic) rocks and types similar to those of the surrounding banded series are common. The "inclusions" have the appearance of lenses and "fishes" with gradual contacts with the surrounding granite. They appear to represent partly assimilated remnants of granitization processes. Along the upper part of the ridge down from top 574 towards Grytting red garnets are rather common in the granite close to its border against the banded series.

The granite is a massive reddish unfoliated rock. The most distinguishing feature is the size of the potash feldspars. Many are more than one cm. in largest dimension: they are responsible for the colour of the rock. The large sized feldspars may suggest association of the granite with the western monzonites. However, the rocks are mineralogically and colourwise different.

Some modal analyses (point counted thin sections) of the granite and associated rocks are given in Table 11. Nos. 5 and 6 represent rocks from the surrounding banded series and are repeated from Table 3. That the granite is in the same mineral facies as the surrounding rocks is indicated by the fact that hypersthene is the characteristic dark silicate mineral for both, and potash feldspar is an orthoclase with small $2V_x$. A partial chemical analysis of a feldspar from rock no. 1 is listed as no. 64 in Table III. It is characterized by a somewhat higher Na content than is typical for feldspars from the banded series and, in this respect, more closely resembles the feldspars from the western monzonites.

The granite (granodiorite?) is also characterized by a high quartz content as compared to that typical of the banded series. The quartz occurs as very large grains with strong undulatory extinction.

Table 11.

Modal composition of the old red granite and associated rocks.

Minerals	1 65/56	2 68/56	3 64/56	4 65/56 b	5 63/56	6 69/56
Quartz	25	24	34	9	<1	—
Potash feldspar ...	30	7	34	3	11	—
2Vₓ	(47—53)	(35—45)	(45)	(42)	(44)	
Plagioclase	39	60	30	59	77	52
An. in plag.	(30)	(35)	(30)	(50)	(38)	(45)
Hypersthene	5	5	tr.	27	10	2
2Vₓ	(62)	(60)		(56)	(55)	
Diopside	—	—	tr.	—	<1	8
Hornblende	—	—	—	—	tr.	18
Biotite	1	3	tr.	tr.	tr.	17
Ore	<1	1	tr.	2	1	3
Apatite	—	—	2	tr.	tr.	tr.
Calculated mol per cent SiO₂	66.6	64.1	70.6	53.8	53.1	42.6

Table 11, localities,

1. Granite, shore east of Troldøen, 2. Granite, ridge between top 574 and Grytting, 3. Granite-migmatite, Olderfjordvann, 4, Inclusion of basic rock in granite no, 1, 5. Banded series rock, Strømmen, Olderfjord, 6. Banded series rock, ridge between top 574 and Grytting.

No. 4, which represents a basic inclusion in the granite, probably was modified mineralogically by granitization processes. Quartz is present as very few large grains. The potash feldspar is present in very small grains quite different from the feldspars of the granite. The plagioclase is more basic (An. 50) than plagioclases from similar rocks in the banded series (Table 3), and may represent a basification that accompanied the granitization.

The Bø gabbro and the gneisses between Maars and the north end of Jørgenfjord.

The rocks which are grouped together in this chapter cover a rather narrow NS-trending zone along the shore from Maars on the south northward along the west side of Jørgenfjord. On the north end of Jørgenfjord it joins the NS-striking zone of sedimentary and associated rocks which, together with young red granites, underlie most of the area north to Kråkberget.

Most of the rocks west of these gneisses are of the porphyro-

blastic monzonite type, but in the northern part members of the banded series are present along the border. All rocks east of the red granites and sedimentary rocks on the east side of Jørgenfjord belong to the banded series. The rocks between Maars and the north end of Jørgenfjord are mineralogically different for similar compositions (Table 12).

In Table 12 the rocks are arranged according to their geographic position from south to north irrespective of their chemistry and/or mineralogy. Vogt (1909) indicated on his map a gabbro (olivine hyperite) in the south of Bø. The major outlines of this mass indicated by him are correct (compare with map Fig. 3) but in detail more complex relationships exist. The western monzonites occur as far east as the north end of the bay at Maars in the southern part of Bø. No. 1, Table 12, is from the west side of this bay, some 100 m. south of the road. Although in the minerals present it appears to be related to the monzonites, in mineral proportion and grain size it is more closely related to certain members of the banded series. On the east side of the bay, close to the shore on the south end of Maarsundneset, rocks identical to the porphyroblastic monzonites occur (no. 2, Table 12, cf. e.g., no. 2, Table 1). According to Vogt (1909) this is gabbro. About 200 m. further along the shore towards the northeast the rock appears megascopically to be a gabbro but in thin sections it may be seen to be more nearly similar to the monzonites (no. 3, Table 12). Hornblende is, however, typically present in lieu of the pyroxenes. Some diopsidic pyroxene is present sporadically but even this is in the stage of alteration into green hornblende. On the other hand, in the monzonite no. 2 the diopsidic pyroxene is quite unaltered although the hypersthene is notably altered into biotite and green hornblende (in fact, all the hornblende (3 per cent) in this rock occurs as reaction rims around hypersthene). The writer believes that it is reasonable to assume that these rocks represent different stages in a retrograde metamorphism of the western monzonites. Attention is also called to the gradual increase in the $2V_x$ of the potash feldspars in the three rocks so far described in this section. This relationship indicates an increase in triclinicity of feldspars as the retrograde metamorphism proceeds. Although microcline twinning was not observed in any of them, the large $2V_x$ in no. 3 clearly indicates a fairly ordered feldspar. The introduction of titanite (which in all the rocks in Table 12 may be seen, where present, to surround magnetite grains) is another indication of the change in mineral facies. Further northeast at the end of the little bay directly north of Strand-

Table 12.

Modal composition of rocks belonging to the Bø gabbro and the gneiss series between Maars and the north end of Jørgenfjord.

Minerals	1 301/55	2 303/55	3 304/55	4 309/55	5 310/55	6 313/55	7 358/55	8 51/56
Quartz	–	–	tr.	–	tr.	–	–	17
Potash feldspar	6	13	29	–	19	2	tr.	46
2Vx	(40–45)	(45–53)	(60–65)		(60–62)			(55)
Microcline twinning					+			
Plagioclase	72	70	60	69	59	70	70	32
An. in plag.	(40)	(34)	(33)	(38)	(25)	(40)	(39)	(30)
Signs of sec. alt.		+	+		+			
Hypersthene	5	2	–	3	–	3	2	2
2Vx	(54)			(50)		(50–52)	(50–52)	(50–52)
Signs of sec. alt.		+		+			+	
Diopside	7	4	tr.	14	–	10	12	–
2Vz	(52)			(53)		(52)	(52)	
Hornblende	–	3	7	1	10	–	–	–
2Vx								
Biotite	4	5	4	7	10	11	5	tr.
Sillimanite	–	–	–	–	–	–	–	–
Scapolite	–	–	–	–	–	–	•	–
Calcite	–	–	–	–	–	tr.	–	–
Epidote	–	–	–	–	–	–	–	–
Titanite	–	–	tr.	–	2	–	–	–
Apatite	1	tr.	tr.	1	tr.	tr.	6	tr.
Zircon	–	–	–	–	–	–	–	tr.
Ore	5	3	tr.	5	tr.	4	5	3

Minerals	9 53/56 b	10 54/56	11 318/55	12 44/56	13 42/56	14 321/55 a	15 321/55 b	16 36/56
Quartz	tr.	1	1	–	18	tr.	tr.	14
Potash feldspar	35	27	21	8	44	tr.	8	67
2Vx	(55–60)	(60)	(60)	(60–65)	(70–73)		(58–61)	(63)
Microcline twinning			+		+	+		
Plagioclase	59	59	64	58	33	54	74	1
An. in plag.	(30)	(36)	(23)	(29)	(24)	(35)	(30)	(35)
Signs of sec. alt.			+					
Hypersthene	1	–	–	–	–	–	–	–
2Vx	(48)							
Signs of sec. alt.								
Diopside	3	tr.	–	–	–	–	–	–
2Vz	(53)							
Hornblende	–	11	11	14	–	24	3	–
2Vx		(60)		(40)		(48)		
Biotite	tr.	–	1	16	5	20	8	5
Sillimanite	–	–	–	–	–	–	–	7
Scapolite	–	–	–	3	–	–	6	–
Calcite	tr.	–	tr.	–	–	tr.	–	–
Epidote	–	–	–	–	tr.	–	<1	–
Titanite	–	–	tr.	tr.	tr.	tr.	tr.	–
Apatite	tr.	tr.	1	tr.	tr.	1	tr.	–
Zircon	–	tr.	–	tr.	–	–	–	–
Ore	2	2	1	1	tr.	1	1	6

skjærene the gabbro (hyperite) no. 4, Table 12, occurs. Near the border it is strongly foliated, nearly schistose. In these foliated parts the minerals are plagioclase, green hornblende, biotite, epidote, titanite and "ore minerals". Away from the tectonized border zone the high temperature mineral paragenesis of no. 4 is characteristic (no. 4 was collected only 4 m. from strongly foliated gabbro with the above mentioned mineral paragenesis).

Some 100 m. further towards the northeast another monzonitic rock with large potash feldspar porphyroblasts occurs. It is rather dark and may be easily mistaken for a gabbro in the field (no. 5, Table 12). Lenses of a dark amphibolitic rock occur within it locally. No. 5 is rather similar to no. 3, in mineral composition. Although the $2V_x$ of the potash feldspar is not larger than in the latter, microcline twinning is here clearly developed. A chemical analysis of this rock is given in Table 13 (310/55). There is a close chemical similarity between this and the western monzonite from Steine (no. 55/56, Table 2).

Exactly were the gabbro (hyperite) contact is, is difficult to establish. However, the gabbro may be seen in intensely weathered exposures along the road from Bø towards Snarset. No. 6, Table 12, is representative of the gabbro at Snarset. It is a dark massive rock with a typical ophitic texture. Most of the potash feldspar listed in the table occurs as exsolved irregular lamellae within plagioclase (antiperthite). No. 7, in Table 12, is a similar rock from Kringlevann north of this gabbro area. There is a close similarity between the three gabbros listed in the table (nos. 4, 6 and 7). Vogt (1909, p. 12) reported olivine to be present in these rocks as well and named the rock an olivine hyperite. Olivine, however, was never observed to occur in these rocks by the writer.

Table 12, localities.
1. Charnockitic monzonite, 300 m. south of north end of bay at Maars, west side, 2. Dark monzonitic rock, south end of Maarsundneset, 3. Dark monzonitic rock, 200 m. northeast of no. 2, 4. Medium grained gabbroic rock, north of Strandskjærene, 5. Dark coarse grained monzonitic rock, east of peninsula north of Strandskjærene, 6. Medium grained dark gabbroic rock, 200 m. south of road, Snarset, 7. Gabbroic rock, north end of Kringlevann, 8. Granodioritic dyke in gabbro, south point of peninsula south of Snarset, 9. Monzonitic gneiss, 1.5 m. east of gabbro, peninsula south of Snarset, 10. Monzonitic gneiss, 20 m. east of gabbro, peninsula south of Snarset, 11. Monzonitic gneiss, 1000 m. east of gabbro, roadcut at Ramberg, 12. Dioritic-monzonitic gneiss, south end of Pollen, 13. Red granodioritic gneiss, south end of Pollen, 14. Basic (amphibolitic) band in gneiss, roadcut 500 m. south of junction with road to Uvåg, 15. Light band in gneiss, roadcut 500 m. south of junction with road to Uvåg, 16 Sillimanite gneiss, west of Gåslandsvann.

Table 13.

Chemical analyses and C.I.P.W. (molecular)
norms of three retrograde metamorphosed gneisses

Weight %	310/55	318/55	235/55
SiO$_2$	57.4	56.7	64.7
TiO$_2$	0.83	0.92	0.22
Al$_2$O$_3$	19.8	16.9	16.1
Fe$_2$O$_3$	1.72	3.52	2.46
FeO	3.60	4.30	1.00
MnO	0.09	0.17	0.04
MgO	1.66	2.16	0.41
CaO	4.76	4.62	2.66
Na$_2$O	5.34	5.03	2.83
K$_2$O	3.85	4.19	9.05
C. I. P. W. (molecular) norms			
Quartz (q)	–	–	8.4
K-feldspar (or)	22.5	25.0	54.0
Na-feldspar (ab)	47.5	45.5	25.5
Ca-feldspar (an)	18.5	11.3	4.5
Olivine (ol)	4.5	2.4	–
Hypersthene (hy)	–	1.0	–
Diopside (di)	4.0	9.6	2.4
Magnetite (mt)	1.8	3.8	2.1
Ilmenite (il)	1.2	1.4	0.4
Hematite (hm)	–	–	0.3

(spec. anal. by K. S. Heier).

South on the Snarset peninsula the gabbro is penetrated by several narrow granodioritic dykes. They are typically from 10—50 cm. thick and can be traced with even thicknesses and sharp contacts for several tens of metres along the strike to where they are covered by overburden. A sample of one of them, which strikes N 45° W and dips 30° NE, is listed as no. 8 in Table 12. Vogt (1909, pp. 22—23) described these dykes and stated that although they are more fine grained than the monzonites in the west they are macroscopically similar to them. However, by comparing no. 8 (which is believed to be representative of these dykes) with the western monzonites in Table 1 it is evident that one important difference exists: the dyke has a high quartz content. The quartz is present in these rocks as small rounded grains which is a quite different mode of occurrence from that in any other rocks exposed on the island. The high quartz content also distinguishes these dykes from the lower grade metamorphic monzonites to the north.

Irregular red granite- pegmatite- like dykes are also rather common within the gabbro.

The contact between the gabbro and the monzonites in the east and northeast can be followed in a nearly straight NS direction from the Snarset peninsula to the main road. It dips about 70° W. Further to the north heavy overburden covers the rocks. Both the gabbro and the monzonite are rather strongly foliated close to the border and up to a few metres away from it. The gabbro is locally schistose here just as was the case along the western horder (p. 57). Bands of monzonites that strike parallel to the contact occur locally within the gabbro close to the main contact. The gabbro between these bands is especially strongly foliated.

No. 9, Table 12, represents the monzonite that crops out, 1.5 m. east of the main gabbro contact. The gabbro close to the contact here is nearly identical to no. 6 from Snarset. The high grade mineral paragenesis of this monzonite agrees with that of the gabbro whereas only 20 m. to the east a lower grade metamorphic mineral assemblage may be observed (no. 10, Table 12). Macroscopically this latter rock is characterized by a redder colour of the rocks (dependent upon the potash feldspars).

Because of heavy overburden north of the main road it is difficult to obtain a good impression of the true form of the gabbro. There are some indications that it is at least two rather than only one gabbro lens, i. e., monzonitic rocks (no. 5) crop out along the shore beyond the first gabbro outcrops. The total outcrop area of the gabbro is, in any case, rather small. The gabbro is shown as one single intrusion on the map Fig. 3. It is of interest to note that Vogt's original descriptions of this gabbro (1909, p. 12) and of certain dyke rocks (ibid, p. 24) occurring elsewhere in the Lofoten-Vesterålen group are very similar. It is probable that this gabbro is represented by hyperitic dykes or lenses. That they probably were intruded into the gneisses rather late, is indicated by their stable high temperature mineral assemblage despite their occurrence within monzonitic gneisses which tend to occur chiefly in a lower metamorphic facies within this area. The strong foliation of the gabbro and recrystallization of the femic minerals into hornblende and biotite close to the contacts probably indicates mechanical movements subsequent to solidification of the gabbro. (Recall that during this work it has been shown that thrusting took place after the regional metamorphic recrystallization of the rocks).

The slightly reddish somewhat foliated monzonitic rocks that border the Bø gabbro on the east extend for some distance further towards the northeast. No. 11, Table 12, is from a roadcut at Ramberg about 100 m. northeast of the gabbro contact and is fairly typical of the rocks that occur here. The potash feldspars occur in these rocks as medium sized porphyroblasts (microcline twinning is observed in no. 11). It may be that the rocks represent lower grade metamorphic equivalents of the western monzonites. That the rocks have undergone a retrograde metamorphism is suggested by the transitional state of some of the minerals within these rocks and by the fact that the feldspar geological thermometer (see later) indicates temperatures of crystallization of the feldspars similar to those obtained for the western monzonites. This is regarded as a relict temperature. The rock no. 11 was chemically analysed (318/55, Table 13). The chemical analyses of rocks no. 310/55 and 318/55 show considerable similarities with the western monzonites (Table 2).

As far northeast as Pollen more banded rocks occur. In the field they are very similar to rocks from the banded series although they have a lower grade metamorphic mineral assemblage (nos. 12, 14 and 15, Table 12). The very low 2V of the hornblendes in nos. 12 and 14 should be noted. The colour is clear light green.

The zone of the low grade metamorphic rocks is rather narrow here. Thus the high grade banded series is found only 50 m. west of the main road west of Gråholmene (no. 4, Table 3), and in a roadcut on the road to Uvåg about 200 m. east of the junction with the main road (no. 10, Table 3). In the east the high grade metamorphosed banded series is found along the shore east of Uvåg. Thus the low grade metamorphosed zone here has a maximum width of about 2000 metres. It is interesting to note that at this place where the banded series occurs bordering the low grade gneisses to the west and east, the retrograde metamorphosed rocks are, by appearance in the field, extremely similar to the banded series rocks.

At Pollen gneisses are interbedded with red granitic rocks which have a low grade metamorphic mineral paragenesis (no. 13, Table 12). The potash feldspar is a microcline perthite ($2V_x$ about 70°) and epidote and titanite are common constituents. A chemical analysis of the rock is given (no. 235/55, Table 13). Although it contains some green hornblende it is very similar to no. 13, Table 12. Although these granites are in appearance rather similar to the young red granites, they

differ from the young red granites chemically as well as mineralogically. They are probably best understood as low grade metamorphosed derivates of the old granites (pp. 53—54).

Further northwards to the north end of Jørgenfjord these red granitic types are common in the gneisses. The gneiss types are somewhat variable but are typically similar to the banded series despite their occurrence in a lower metamorphic facies than that of the banded series.

The sedimentary rocks and the rocks associated with these that strike from Jørgenfjord towards Kråkberget in the north occur in a high metamorphic facies. The red granites that occur with them are representative of the young granites. Though the retrograde metamorphic zone seems to end here at the north end of Jørgenfjord it is interesting that this complex of young granites and sedimentary rocks forms the continuation of the zone towards the north. It is possible that the zone continues directly to the west of this complex in which case it is covered by overburden all the way north to Kråkberget. Some support for this suggestion is that at Kråkberget, gneisses that contain scapolite occur[1]. These rocks are described in the chapter dealing with the metamorphic sedimentary rocks and associated rocks.

Before leaving the subject of the low grade metamorphosed gneisses the occurrence of some sillimanite containing gneisses along the west side of Gåslandsvann between Ringstadåsen and Langmovann should be mentioned (no. 16, Table 12). This is the only occurrence of sillimanite gneisses known on the whole island. The gneiss is strongly foliated and strikes N 10° E and dips 45° E. It is white on weathered surfaces and thus quite distinct. On some fresh surfaces it somewhat resembles the monzonites in the west, but it is lighter in colour. As seen from Table 12 it is a potash feldspar dominant rock (67 per cent Or.) and plagioclase, which typically is a major mineral in the gneissic rocks on Langøy, is in this rock present to the amount of 1 per cent. The potash feldspar forms porphyroblasts. Microcline twinning was not observed, but the fairly large $2V_x$ indicates a high degree of order (intermediate triclinicity). The chemistry of this rock as reflected by the mineral content does not correspond to any normal igneous rock (it is extremely rich in K). In the field, however, it may have an "igneous" appearance in that partly assimilated "amphibolitic" rocks

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[1] See Fyfe et al. 1958 p. 160 for a discussion on the stability of scapolites in the granulite facies.

occur included within it. These are not bands as for instance in the banded series, but rather small lenses. These are also rather potassium rich, in that biotite is the dark silicate mineral and occurs in amounts equal to plagioclase, the other major mineral in them. The sillimanite gneisses most probably represent metamorphic (and metasomatic?) products of originally sedimentary rock types (e. g., shales).

Sedimentary and associated rocks.

Rocks of undoubtedly sedimentary origin occur sporadically on the island. They occur in the west together with the banded series and the charnockite border series, and in northeast within the veined gneisses, but are not known elsewhere. The association of these gneisses with recognizable sedimentary rocks suggests that they originated through extreme metamorphism of original geosynclinal sediments and lavas. In general the gneisses occurring next to the sedimentary rocks can not be seen to differ from the typical gneisses in the series.

Vogt (1909) assumed that the sedimentary rocks represent contact metamorphic inclusions in igneous rocks. However, their general appearance in the field contradicts this view. They extend as narrow zones for considerable distances (often several kilometres) in the direction of the strike, and their strike and dip are everywhere parallel to the foliation of the gneiss. The sedimentary rocks are intensely folded and tectonized. The tectonic relationships can be solved only by very detailed mapping. In the present study this was not attempted because the maps available (1:50.000) do not warrant such detailed mapping.

The area of sedimentary rocks between Jennestad and Koven about 10 km. north of Sortland was examined in some detail by Skjeseth (1952) and Vokes and Padget (1955). (The area has interest for possible graphite production.) Here in an area of dominantly amphibolitic gneisses with "granitic" veins a sedimentary series with graphite schists occurs overlain stratigraphically by metalimestone (marble), quartzite and calcareous sandstones. (According to Skjeseth, 1952, crossbedding occurs in the latter.) Iron ore is found together with sedimentary rocks west of Kovvann.

Skjeseth (1952) suggested that the graphite represents a horizon in a primary sedimentary series which is repeated by folding. The strike is E—W.
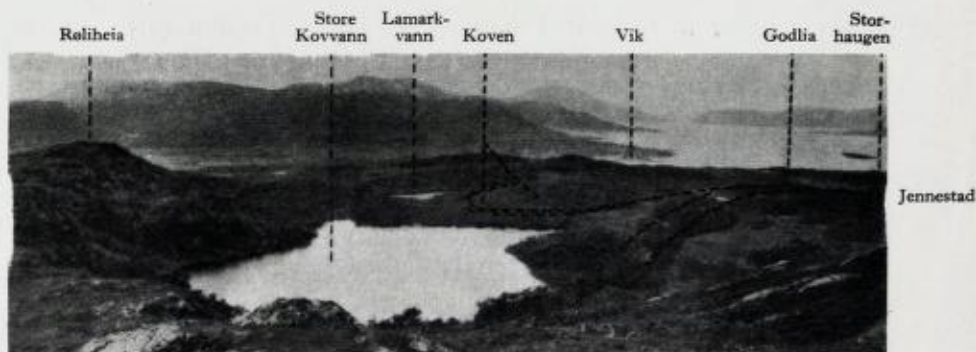
Fig. 16. Trends of the graphite schist in the Jennestad area (dashed lines).
Reproduced from Skjeseth, 1952.

Strøket av grafitskiferen i Jennestad området (brutte linjer). Gjengitt etter
Skjeseth, 1952.

South of Koven the general strike is N—S and then turns to E—W
south of Lamarkvannet. (Fig. 16 shows the trends of the graphite
schist in the Jennestad area as given by Skjeseth, 1952.) The structural
picture is an anticlinorium consisting of several "graphite anticlines".
The major fold axes dips 20° towards the west. Skjeseth (1952) also
found a major foldthrust (foldningsforkastning) striking E—W from
Koven to the sea at Jennestad. A relative uplift has taken place north
of the thrust. The thrust plane dips 60° N. The graphite quality appears
to be related to the tectonics. It is completely mylonitized ("amorphous
graphite") along the fold thrust, whereas it is very good quality gra-
phite ("flake graphite") in the less tectonized parts.

Graphite schists also occur at Vik, Strand, and Evatn within the
amphibolite facies veined gneisses (Fig. 3). The general strike is paral-
lel to that in the Jennestad area and the schists may represent the same
zones repeated by folding. In the west (the granulite facies area) gra-
phite occurs between Troldvann and Hornvann, Selneset and in a
several kilometres long, strongly tectonized zone between Jørgenfjord,
Kråkberget and Møkland. In the latter area the strike is essentially
N—S. Metalimestone (marble) characteristically occurs in close con-
nection to the graphite, but is absent in the Jørgenfjord-Kråkberget
area. It is interesting that the banded gneisses in roadcuts next to the
shop at Kråkberget contain abundant scapolite together with plagio-
clase, diopside, titanite, apatite and ore. The rock is in field appear-

ance identical to the banded charnockites (of unknown origin), and its mineral paragenesis suggests that it is a product of regional metasomatism of impure limestones.

Isolated bands of metalimestone occur at several other places. A band several kilometres long extends from Sortland towards the west. The strike is E—W and the dip is towards the south. Veined gneisses of the Jennestad type occur in the foot wall of the metalimestone. Directly in the hanging wall, there is a massive amphibolite ("gabbro") several metres broad. The amphibolite is followed by a zone some metres broad of a rather pure quartzite which again is followed by an amphibolite similar to the first one. A homogeneous red gneiss granite follows next. The exact contact between the amphibolite and the metalimestone is nowhere exposed; the metalimestone occupies a depression in the area with the amphibolite occurring along the ridges above it. Some trace element data suggest that this amphibolite was formed from the metalimestone as a result of metasomatic processes. This is further discussed in the chapter dealing with the geochemistry. (Amphibolites believed to be of a similar origin are also found elsewhere in the Jennestad area.)

Metalimestone also occurs in a number of isolated places in both the banded series and the charnockite border series. (Knutlitind, Hornvann, Tuenelva, Kavåsen, Skålebø-Ryggedalsvann.)

The zone between Skålebø and Ryggedalsvann was especially highly tectonized and metasomatized. The band was broken up into disconnected lenses, the long axes of which strike in a N—S direction. Some of the limestones were completely metasomatized close to the border. One specimen consists only of green diopside and green spinel and another rock which occurs between two metalimestones contains olivine, diopside, tremolite, Mg rich biotite, and apatite.

The normal metalimestones consist essentially of calcite. Olivine (generally serpentinized along cracks and borders) and spinel (green to colourless) are also present. Diopside (though present in the impure contact rocks) has never been seen to occur in the pure limestones from the amphibolite facies area; it may occur within the charnockites, though everywhere very subordinate to olivine. As few thin sections of the marbles were studied this cannot be said to be a general rule.

Vogt (1909 p. 27) also reported the presence of a brown, rounded mineral which he assumed to be chondrodite. Such a mineral occurs in small concentration in most of the metalimestones. Because of the low

concentration it was not attempted to separate it; it may well be chondrodite.

The quartzites that occur with the other sedimentary rocks on Langøy are typically impure and feldspathized. The quartzite band between the two amphibolites associated with the metalimestone that strikes west from Sortland consists of almost wholly quartz. (A rather pure quartzite also occurs along the road on the west side of Gullesfjorden, S of Bømarken, in the amphibolites facies gneisses on Hinnøy. It is about 50 m. broad and can be seen along the road for about 65 m. It disappears under overburden. It strikes N—S and the dip is 70° E. Towards the west it becomes feldspathized and grades into a fine grained granitic rock.)

Pure quartzites do not occur within the granulite facies area. Several types termed quartzites in the field have been found in thin sections to be indistinguishable from quartz rich varieties of the banded charnockites.

Plagioclase is commonly the dominant feldspar in these rocks. Hypersthene is the dominant (and in some rocks the only) femic mineral.

Mica schists proper are not present anywhere on the island. Some of the amphibolite facies gneisses in the east have the appearance of biotite schists, but their gnessic character is always marked. Rocks resembling mica schists are absent in the charnockite area. The sillimanite gneiss at Gåslandsvann within the zone of retrograde metamorphic gneisses (no. 16, Table 12) with its high potassium and aluminium content is possibly best understood as a high grade metamorphic product of clay sediments.

Another rock which may be termed a mica schist occurs at Vikelva. A chemical analysis together with the normative composition and the calculated and observed mode of this rock is listed in Table 14. The high magnesium content of this rock suggests that it represents a metamorphic derivate of basic lavas or tuffs.

Quartz banded iron ores of sedimentary origin occur scattered all over the island. Like the other sedimentary rocks they occur embedded in the veined gneisses in the east and in the banded granulite facies rocks in the west, but not in the other rocks. The quartz banded ores are generally not associated directly with the other sedimentary rocks, but occur isolated in the gneisses. West of Kovvann (within the Jennestad sedimentary complex) iron ore is found in a typical sedimentary environment. The rock sequence is here: gray granodioritic gneiss, iron

Table 14.

Quartz-plagioclase-biotite-schist from Vik.

	Weight %	Cation %	C. I. P. W. (mol) norm	Calculated mode	Observed mode	Minerals
SiO_2	58.3	54.1	2.8	17.5 (15.9)	14.0	Quartz (q)
TiO_2	0.83	0.6	26.0	–	–	K-feldspar (or)
$AlO_{1\frac{1}{2}}$	13.9	15.2	28.0	28.0 (25.4)	23.8	Na-feldspar (ab)
$FeO_{1\frac{1}{2}}$	1.86	1.3	11.0	12.0 (10.9)	10.2	Ca-feldspar (an)
FeO	5.15	4.0	26.2	–	–	Hypersthene (hy)
MnO	0.05	–	2.8	–	–	Diopside (di)
MgO	7.96	11.1	–	41.5* (47.0)	50.0	Biotite (bi)
CaO	2.94	2.9	1.2		1.0 (Ore)	Ilmenite (il)
$NaO_{\frac{1}{2}}$	3.11	5.6	2.0	1.0 (0.9)		Magnetite (mt)
$KO_{\frac{1}{2}}$	4.44	5.2	n.d.		1.0	Apatite (ap)
H_2O	n.d.					
CO_2	n.d.					
P_2O_5	n.d.					

*) The amount of biotite given was calculated without water as this element was not taken into account in the sum of the cation proportions. Nos. in brackets give the recalculated mode when water was included. The calculated mode was obtained by making all the Na into Ab. Enough Ca was then made into An. to make a plagioclase of composition An. 30. 1% magnetite was made out of FeO and Fe_2O_3.
The remaining elements were then made into biotite.
The biotite formula obtained is:

$$K_{5.2} (Mg_{11.1}, Fe^{2+}_{3.7}, Ca_{0.5}, Al_{0.2}) (Al_{4.6}, Fe^{3+}_{0.6}) (Si_{15.0}, Ti_{0.6}) (OH)_{10.4} = 51.9$$
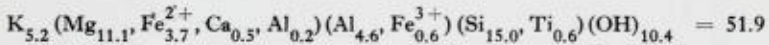
which recalculated to 100 % is:

$$K_{10.0} (Mg_{21.4}, Fe^{2+}_{7.1}, Ca_{1.0}, Al_{0.4}) (Al_{8.8}, Fe^{3+}_{1.2}) (Si_{28.9}, Ti_{1.2}) (OH)_{20.0}$$
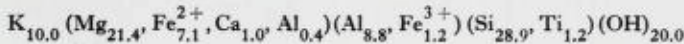
ore, amphibolitic gneiss, impure metalimestone, amphibolitic gneiss, and coarse granodioritic gneiss. The ore of this sequence is not, however, exactly similar to the quartz banded ores and more probably represents a skarn type. (Some trace element data suggest that the associated amphibolitic gneisses are of a metasomatic origin.)

An ore similar to the one from Kovvann occurs on the farm Flåtten on the north side of the road between Vik and Frøskeland (about 1 km. E of Trollvann). Graphite occurs on the same farm. No exposures can be seen because of overburden but the ore and graphite are probably associated with sedimentary rocks (metalimestone and graphite schists in gneisses) that strike west from Vik.

Quartz banded ore is found in Steirodalen on the east side of Langøy, and Lahaugen, Strømfjord and Åsanfjord in the granulite facies

area. The ores exhibit the same metamorphic facies as the gneisses in which they occur. Hypersthene is a typical femic constituent of the ores in the granulite facies area whereas hornblende and biotite occur within the amphibolite facies parts.

These iron ores generally do not extend over long distances and none of them has been found to be of economical importance (nearly all of them have been quarried). Similar ores occur at many places on Hinnøy (in both the granulite and amphibolite facies areas) and characteristically are present on most of the islands in the Lofoten-Vesterålen group. It is tempting to consider that the ores represent one original horizon which has been broken apart by subsequent orogenic folding. A study of them and of the graphite schists and metalimestones may go a long way towards solving the problems of stratigraphy and tectonics for this area of dominantly crystalline rocks.

Amphibolites.

It was stated that the sedimentary rocks occur within mixed gneisses where acid bands or veins alternate with basic (amphibolitic) bands. The amphibolitic rocks which so occur are included in the descriptions of the various gneiss types, as it is impracticable to treat them separate and all transitions are present. These amphibolites are typically irregular and intimately intermixed with granitic material.

Amphibolites of a much more massive appearance are associated with the true sedimentary rocks; e. g., the graphite schists are everywhere between massive amphibolites. Pyrrhotite mineralization of the amphibolites close to the graphite bands is characteristic and the mineralization is probably caused by a metamorphic activation of sulphur which originally occurred within the carbonaceous sediments. It is tempting to assume that these amphibolites are metamorphic basalt lava flows. Nos. 77/56 and 81/56, Table 15, give the norms and modes of amphibolites that occur with graphite schists at Jørgenfjord and Kråkberget, respectively.

Amphibolitic rocks also occur associated with metalimestones (nos. 13/56 and 33/55 j, Table 15). These are mineralogically and chemically (major elements) comparable to amphibolitic types that occur within the gneisses. Certain trace element data, however, suggest a metasomatic origin (of impure limestones) for these rocks. No. 29/55, Table 15, also belongs to the latter group. It occurs in ordinary gneisses not clearly associated with metalimestones, but is found in the

Table 15.

Chemical analyses, C.I.P.W. (molecular) norm and observed
mode of five amphibolites associated with sedimentary rocks.

Weight %	77/56	81/56	29/55	13/56	33/55 j
SiO₂	46.2	42.5	45.9	45.6	48.7
TiO₂	1.50	1.67	2.17	1.17	1.17
Al₂O₃	13.6	11.2	12.8	15.2	17.7
Fe₂O₃	4.98	7.15	7.62	3.97	5.32
FeO........................	8.65	11.82	10.25	8.65	5.50
MnO........................	0.22	0.18	0.22	0.21	0.13
MgO	6.63	7.46	5.97	8.79	5.80
CaO	11.75	12.73	10.07	11.89	7.70
Na₂O........................	4.07	3.11	2.84	2.60	4.50
K₂O	0.87	0.67	0.67	0.45	2.02
C. I. P. W. (molecular) norms.					
K-feldspar (or)	5.5	4.0	4.0	3.0	12.0
Na-feldspar (ab)	16.5	8.0	27.0	17.0	28.7
Ca-feldspar (an)	16.5	15.3	21.0	28.5	22.3
Nepheline (ne)	12.3	12.6	–	4.2	7.1
Olivine (ol)	7.3	10.0	5.1	16.3	9.9
Hypersthene (hy)	–	–	6.2	–	–
Diopside (di)...............	34.4	40.0	25.2	25.2	12.8
Ilmenite (il)...............	2.2	2.4	3.2	1.6	1.6
Magnetite (mt)	5.3	7.7	8.3	4.2	5.6
Observed mode.					
Quartz	–	–	2.0	4.0	–
Plagioclase	32.0	35.0	43.0	33.0	62.0
(An. in plag.)	(30)	(30)	(42)	(54)	(43)
Hypersthene	1.0	–	–	–	–
Diopside...................	6.0	15.0	14.0	1.0	9.0
Hornblende	60.0	42.0	31.0	61.0	10.0
Biotite.....................	tr.	2.0*	–	–	16.0
Titanite	–	–	–	tr.	–
Apatite	tr.	tr.	tr.	tr.	tr.
Ore	1.0	6.0	10.0	1.0	3.0

*) The biotite appears to be an alteration product of hypersthene.

(spec. anal. by K. S. Heier)

77/56, amphibolite with graphite schist, Jørgenfjord.
81/56, amphibolite with graphite schist, Kråkberget.
29/55, amphibolite in gneiss, between Jennestad and top 70.
13/56, amphibolite in contact with marble, close to Prestetjern.
33/55 j, amphibolite in contact with marble, S. of Kovvann.

Table 16.

$\dfrac{Na + Ca}{Al - Al'}$ *ratios of the amphibolites from Table 15*

(Al'; aluminium allotted to orthoclase).

	77/56	81/56	29/55	13/56	33/55 j
$\dfrac{Na + Ca}{Al - Al'}$	1.5	1.6	0.9	1.1	1.0

Jennestad area of sedimentary rocks, and metalimestones are present not far away.

The amphibolites presented in Table 15 are from the granulite (nos. 77/56, 81/56) and amphibolite facies areas (nos. 29/55, 13/56, 33/55 j). The environmental difference is indicated to some extent by by the minerals present. The variations in the plagioclase are dependent upon the varying chemistry (see Table 16) and do not reflect the degree of metamorphism. Hypersthene (characteristic of the granulite facies) is present in no. 77/56. It is not stable in no. 81/56, but it is seen in traces nearly completely altered to biotite. (The rocks at Kråkberget are the elongation of the retrograde metamorphic gneisses that strike from Maars to Jørgenfjord. The instability of the hypersthene and the presence of scapolite together with plagioclase, diopside, titanite, and ore in some of the nearby gneisses (p. 61) suggest some retrograde metamorphic activity.)

Hornblende occurs in all the amphibolites, but its colour reflects the difference in metamorphic facies. It is deep brown in nos. 77/56 and 81/56 and green in the remaining three rocks.

The anorthosites.

The anorthosite at Sunnan is marked on the map, Fig. 3, but was not examined in the laboratory and is not discussed in this paper. The other anorthosite massif covers an area on the east side of Eidsfjorden between Skogstrand and Strømfjord. Towards the east it extends as far as the tops of Slåttnestind (651) and Ole Hansatind (734) (see map Fig. 3). Here it has been thrust into its present position above the red augengneisses. The thrust plane is distinct (Fig. 10) and an intense mylonitization of the underlying red augengneisses can be ob-

served. The anorthosite, itself, does not seem to have been mylonitized, but crushing and distortion of the plagioclases can be seen in thin sections. The strike of the thrust plane at Slåttnestind is EW and it dips about 40° N. The anorthosite is a massive gray rock. Its outcrop surface is barren of vegetation other than sporadic moss except very close to the shore where a mixed vegetation is present.

It is a very coarsely crystalline rock. The individual plagioclase crystals are up to several centimetres in greatest dimension. Some of the pyroxenes (both ortho- and clinopyroxenes) are also large. (The largest hypersthene crystal seen was 20 cm. long.) The contacts between the anorthosite and the rocks to the northeast and southwest are covered by vegetation. It seems that to the north near Middagstuva, the amount of both pyroxenes increases and the anorthosite becomes more gabbroic. The anorthosite seems to be more fine grained along the shore from Grønning to Skogstrand. The road between Fleines and Kjørstad seems to have been built along the contact between the anorthosite and the monzonite east of it. The actual contact is covered by vegetation, but it must be relatively sharp. There are no visible signs of thrusting here. The mineralogy of the anorthosite is simple. Plagioclase is the major mineral and in large parts of the mass is essentially the only mineral present. The plagioclase composition ranges from An. 45 to An. 55 (no zoning has been observed).

In most thin sections pyroxenes can be seen. Hypersthene is the most common pyroxene. It is slightly reddish brown and $2V_x$ varies between 60° and 65°. The clinopyroxene commonly occurs as a rim around the hypersthene but is also common as independent grains. It typically has a faint greenish colour. In one rock (130/56) from near Middagstuva, where the anorthosite is more gabbroic, exsolution lamallae of orthopyroxene can be seen in the coarse clinopyroxenes. Small lamellae of exsolved magnetite (?) can be seen within the exsolved orthopyroxenes. The $2V_x$ of the host clinopyroxene was measured as 58° and $2V_x$ of the exsolved orthopyroxene lamellae is about 60°, whereas it is 65° in the independent orthopyroxene grains. Brown biotite is present in parts of the anorthosite in small amounts. The amount increases as the rock becomes more gabbroic. Some green to brown hornblende also is present in the more gabbroic anorthosite. Iron ore minerals are present as accessory minerals and locally apatite also occurs.

Lenses of yellowish, rusty pyroxenites occur in the anorthosite.

Fig. 17. Band of anorthosite in pyroxenite; roadcut 100 m. north of farms at Ytre Strømfjord.

Bånd av anorthositlinser i pyroksenit; vegskjæring 100 m. nord av gårdene ved Ytre Strømfjord.

The major mineral in these is hypersthene with clinopyroxene in somewhat smaller amounts. Some few grains of plagioclase are also commonly present and iron ore minerals are always present as accessory minerals.

An exposure of a dense pyroxenitic lens which consists of hypersthene (faintly pleochroic) and clinopyroxene (strongly pleochroic) with a few small grains of iron ore minerals, is seen in a roadcut about 100 m. north of the northernmost farm in Ytre Strømfjord. The contacts of this with the anorthosite mass are covered by overburden but the impression is that the contacts are gradational because the pyroxenite becomes gradually more anorthositic towards them. Two narrow parallel bands with anorthosite lenses of size from 20 cm. to less than one cm. traverse the pyroxenite across the entire length of the roadcut. The inclusions are typically lens shaped but some are irregular in form. The two bands seem to combine into one to the north. In the south the

lower band is about 30 cm. broad and the higher somewhat smaller. The largest sized anorthosite inclusions occur in the central part of the exposure, and sizes decrease rapidly upwards. Fig. 17 shows one of the bands. It is difficult to explain the anorthosite lenses other than as xenoliths in a pyroxenitic magma which consolidated to form the pyroxenite.

Monzonites east of Eidsfjorden (southwest of Grønning).

These rocks were not studied in any detail. They are of interest because of their similarity to the monzonites in the Eidet-Hovden intrusion and because of their geographical relationships to the anorthosite. The general field impression is of sharp contacts with the anorthosite, but none are exposed.

In the field the transition between these monzonites and the red augengneisses east of them is best marked by a change in the colour of the rocks; whereas the red augengneisses in amphibolite facies are characterized by their red colour, the monzonites in the west are gray. Apart from this the two rock types are extremely similar macroscopically. Both are coarse grained rocks in which the alkali feldspars commonly are present as irregular porphyroblasts. Amphibolitic inclusions are rare in both of them.

On Slåttnesfjell in the northeast the anorthosite is separated from the red augengneisses only by an approximately 40 m. thick mylonite zone. Towards the southwest the monzonites here discussed occur between the anorthosite and the red augengneisses. This "contact" defines a metamorphic boundary which, in the field, is apparent only as a colour change of the alkali feldspar from gray to red.

As was mentioned in the introduction and will be discussed later, the position of the metamorphic boundary is so closely associated with the continuation of the thrust plane beneath the anorthosite that a direct relation between them seems probable.

A slight crushing of the rocks may be observed close to the metamorphic boundary separating the monzonites and the red augengneisses. This is especially apparent in some thin sections because the tectonization has led to formation of small biotite aggregates.

The rocks resemble in both hand specimens and thin sections the monzonites within the Eidet-Hovden intrusion.

Though no chemical analyses of these rocks have been made the

red augengneisses in the southeast and the monzonites discussed here must be rather similar chemically. Mineralogically, however, they are very different. The mineralogical difference is of the same kind as that between the charnockite border series and the amphibolite facies veined gneisses *i. e.*, that of different metamorphic facies.

Quartz is, except for myrmekite quartz spindles in plagioclase, normally absent or present only as traces.

Alkali feldspar (potash feldspar) is most commonly present as mesoperthites. All gradations exist, however, between fairly large normal orthoclase perthites and typical mesoperthites. In one rock (115/56) the ratio between the feldspars was counted in thin section and found; Or.-perthite, 20 %, alk. feldspars (mesoperthite), 45 %, plagioclase, 30 %; all gradations between the three types are present. $2V_x$ of the potash feldspars ranges between 55° and 60°. Microcline twinning was not observed in any grains.

Plagioclase ranges in composition between An. 25 and An. 35; zoning occurs.

Hypersthene is typically present in all these monzonites. Practically none of it shows signs of any secondary alteration. $2V_x$ was measured in one rock (115/56) and found to be 54°. Some of it is very slightly coloured.

Diopsidic pyroxenes are sporadically present. In one rock very close to the thrust plane (114/56) the amount counted was 14 %. This is, however, an abnormally high concentration.

Biotite is present. The amount ranges from traces up to a few per cent. Close to the thrust plane, where small secondary biotite grains have developed in the rocks, the amount is between 6 and 10 %. The pleochroism of the biotites ranges from X-light yellow and Y, Z-dark brown to red brown.

Iron ore minerals are the most common accessories and are typically present up to a few per cent.

The other accessory mineral in these rocks is apatite. The differences in mineral compositions between these monzonites and the red augengneisses east of them are shown in Table 17.

Red augengneisses in the southeast.

Along a profile from Storvann (by the road between Sortland and Holmstad) towards the southwest, the first rock encountered is the normal, amphibolite facies veined gneisses. Thence red granitic material, mostly as quartz-feldspathic veins parallel to the general foliation, is present in large amounts. South of the EW striking metalimestone band and associated amphibolite (metasomatized limestone) the amount of quartz-feldspathic material rapidly increases. Across Bøblåheia the main rock is a coarse, red granitic gneiss with sporadic inclusions of amphibolitic bands.

Table 17.

Table showing the mineralogical differences between the monzonites east of Eidsfjord and the red augengneisses.

Minerals	Monzonites	Red augengneisses
Quartz	trace	trace
Potash feldspar	orthoclase	microcline
(mesoperthite)	(always present)	(sometimes present)
Plagioclase	present	present
Hypersthene	present	—
Diopside	present	present, but partly altered into green hornblende
Hornblende	—	present
Biotite	present in small amounts	present as major minerel
Titanite	—	present
Epidote	—	may be present
Apatite	present	present
Ore minerals	present in some per cents	present as traces

At the top of the slope down towards Bødalen (Bøskaret) a rock type is encountered which is unknown elsewhere within the mapped area. It is a nonfoliated light, nearly monofeldspathic (plagioclase) rock (diorite). Its colour on weathered surface is light yellowish-reddish. (The red colour becomes more intense with increasing amounts of potash feldspar; the concentration of this mineral increases over a very short distance towards the southwest.) The most characteristic feature of this rock type is the spotted appearance caused by the presence of large amounts of green hornblende and yellow brown biotite both of which are distributed with essentially random orientation throughout the rock. Because of the rather large size of the hornblende and biotite grains and their random orientation the rock is tough and even difficult to obtain samples from.

The rock strikes in an eastnortheast direction from Bø, across Oshaug and up along the north side of Valfjord. Here its relations to the rocks of the charnockite border series can be examined. No sharp contacts are evident; the one rock occurs as inclusions in the other and it seems impossible to say which is the younger.

This rock occurs on both sides of the metamorphic boundary. Therefore a difference in mineral paragenesis can be observed within the rocks. Modal thin section analyses from two localities are listed

as nos. 1 and 2 in Table 18. No. 2 is from a roadcut at Oshaug west of the boundary) and no. 1 from Bøskaret well to the east of the boundary. The difference in mineral facies is obvious. The presence of orthorhombic pyroxene (with no signs of secondary alteration) and abundant diopsidic pyroxene in no. 2 correspond to the mineral paragenesis of chemically similar rocks of the charnockite border series within which it occurs. The green hornblende in this rock occurs either inside or along the borders of the diopsidic pyroxene.

In no. 1 orthorhombic pyroxene is absent and the small amount of diopsidic pyroxene present may be seen to have been highly altered into hornblende. This, together with the presence of microcline twinning of the alkali feldspar (not to be seen in no. 2), agrees with the mineral paragenesis of the surrounding rocks in the eastern part.

Further southwest the amount of potash feldspar increases and gives the rocks their red appearance. This red, rather potassium feldspar rich rock occurs over all of the rest of the southeast area. Inclusions of amphibolitic rocks occur within it (no. 3, Table 18) but are relatively rare.

Southwest of a line between Rise and Skata augengneisses are typical (large red potash feldspar porphyroblasts), although medium grained types also occur. The rocks are not foliated except where the thrust plane beneath the anorthosite is approached. Here an intense shearing of the rocks may be observed (for instance along the ridge between Skata, Ole Hansatind and Slåttnestind). Complete mylonitization of the rocks directly beneath the anorthosite can be observed. The absence, or at least the very low content, of quartz in these feldspathic rocks (Table 18) seems to separate them from the amphibolite facies veined gneisses north of them. Other distinguishing features are: the red colour, the porphyroblastic size of the potash feldspars and the presence of but few inclusions of basic (amphibolitic) rocks. The mineralogy of the rocks is evident from Table 18.

The potash feldspar is a microcline with typical microcline twinning. In no. 1 ($2V_x = 63°$) the twinning can be seen in only some of the grains in the thin section. Twinning was never observed in the few potash feldspars present in no. 2, which is from the granulite facies area.

The grain size ranges from small irregular grains within plagioclase and along contacts between plagioclase grains to large porphyroblasts some of which are strongly perthitic and approach the mesoperthite type. In this respect they are related to the monzonites across the boundary and to the Eidet-Hovden

Table 18.

Modal composition of the red augengneisses and some amphibolitic inclusions.

Minerals	1 95/56	2* 98/56	3 97/56	4 100/56	5 109/56	6 105/56	7 110/56	8 113/56	9 99/56
Quartz	1	–	–	–	2	–	3	1	10
Potash feldspar ..	5	2	12	–	28	27	34	30	39
2Vx	(63)		(70)		(71)	(73)	(82)	(78)	(80)
Plagioclase	72	51	59	58	46	51	50	55	40
An. in plag.	(42)	(32)	(25)	(32)	(30)	(27)	(30)	(20)	(33)
Hypersthene	–	2	–	–	–	–	–	–	–
2Vx		(48)							
Diopside	1	15	4	17	–	–	–	–	–
2Vz			(50)	(55)					
Hornblende	9	4	10	8	1	tr.	7	–	–
2Vx	(56)	(71)	(53)	(63)					
Biotite	10	26	14	14	18	18	5	12	10
Titanite	–	–	1	tr.	3	3	1	tr.	1
Epidote	–	–	–	–	1	–	–	1	–
Apatite	1	tr.	tr.	1	1	tr.	tr.	-	tr.
Zircon	–	–	–	–	tr.	-	–	–	–
Ore	1	tr.	tr.	2	tr.	1	tr.	tr.	tr.
Calculated mol per cent SiO_2	49.1	47.9	51.4	48.1	51.5	51.0	55.8	54.2	58.1

* from within the charnockite border series area (see p. 75).

Table 18, localities.

1. Light dioritic rock, Bøskaret, 2. Biotite diorite, roadcut Oshaug, 3. Amphibolitic inclusion in gneiss, Holandsknuten, 4. Biotite gneiss, 750 m. southeast of Skata, 5. Monzonitic augengneiss, Kvalsøfjell, 6. Red monzonitic gneiss, Hallartind, 7. Medium grained gneiss, brook from Rødvann, 8. Augengneiss, Vetheia, 9. Gneiss, Rishatten.

monzonites. None of the small grains are mesoperthitic. In fact, very small grains cannot be seen to be perthitic at all. The irregular contacts against plagioclase are typical. The small grains embedded in plagioclase commonly include small bits of the same plagioclase. Myrmekite quartz spindles occur inside much of the plagioclase where it is in contact with somewhat larger potash feldspar crystals.

Some of *the plagioclase* shows signs of secondary alterations (no. 3, Table 18), but most is quite fresh. Their compositions are indicated in Table 18. Attention should be given to the more anorthitic (An. 42) plagioclase in no. 1 as compared with its higher metamorphic equivalent no. 2 (An. 32). In this connection the much higher concentration of diopside in no. 2 (15 versus 1 per cent) should also be noted.

Albite/ala and normal albite twins are most typical. Pericline, normal manebach, and normal carlsbad twins were also observed.

Diopside is colourless and in all the rocks shows signs of alteration into hornblende. The instability of diopside was also observed in no. 2 even though the associated orthopyroxene shows no signs of secondary alterations.

Hornblende is of a clear green colour in all the rocks. The much larger $2V_x$ in the higher metamorphic type, no. 2 (71°) as compared with the rest (53° to 63°), is of interest (the hornblende colour is the same in both types). The variation in 2V of hornblende with metamorphism is discussed on p. 135 —138.

Biotite is the "type" femic mineral of these rocks. The pleochroism is as follows:

X-colourless to light yellow, Y-green to green brown, Z-yellow green to green.

About the remaining minerals it should be noted that titanite is typically present in these rocks. It commonly forms rather large mineral aggregates and occurs associated with the hornblende.

Epidote is present as independent grains in two of the rocks. In some rocks the opaque mineral is hematite.

The Eidet-Hovden intrusion.

There is a series of igneous rocks in Bø parish, north of Eidet. Although they have no true counterparts within the examined part of the island, and thus are recognized as a distinctive intrusive series, the previously described anorthosite and associated monzonite may be related to them. In reality the rocks do not form one intrusion. No direct relationship through continuous fractional crystallization from the basic through the most acid members appears to agree with the field evidence.

The petrogenesis of these rocks constitutes a problem quite separate from that of the surrounding rocks with which this paper is especially concerned. However, because the relations between the intrusion and the surrounding banded series (or rather, whether the intrusion was temporally pre or post the general regional metamorphism of the area) are not readily apparent, and because dykes which occur within the gabbro are in appearance similar to certain rocks of the banded series, a description of these rocks is needed. The occurrence of an associated magnetite-ilmenite deposit, by far the most important prospective mineral resource on the island, also necessitates a geological description of the intrusion.

The map Fig. 18 gives a somewhat more detailed picture of the peninsula than the main map Fig. 3. The rocks that occur may be divided into the following groups:

1. gabbro and associated derivatives
2. monzonite intrusions
3. dyke rocks
4. granite pegmatite.

Simplified geological map of the EIDET - HOVDEN intrusion

■ Ultrabasic rocks

▨ Magnetite-ilmenite ore

□ Gabbro

Banded gabbro, Fig. 19

Monzonite

The banded series

1, 2 Nos. of samples, Table 19

60/ Strike and dip of banding and/or foliation

564 Altitude in metres

Fig. 18. Simplified geological map of the Eidet—Hovden intrusion which shows the gabbro zones and the position of the samples listed in Table 19.

Geologisk oversiktskart over Eidet—Hovden intrusjonen som viser gabbro sonene. Prøvenumrene fra Tabell 19 er inntegnet.

The gabbro and associated rocks.

These rocks occupy the largest area and can be subdivided:

    a. altered contact zone
    b. normal gabbro
    c. ultrabasic lenses
    d. magnetite-ilmenite deposit.

The altered contact zone.

Although the exact contact between the gabbro and the surrounding banded series is nowhere exposed the contact can safely be assumed to be sharp. The intrusion is nearly surrounded by the sea, the Malnesfjord to the east, the Åsanfjord in the south and the ocean proper towards north and west. Only over a relatively short distance in the southeast is the contact above the sea level. Where the contact traverses the island it can be found to occur along a narrow valley trending southwards from just south of Selvåg along the west side of the small top Grubklatten. To the west the gabbro is exposed on steep, nearly inaccessible cliffs with heights of about 500 metres. The contact trends N—S further south to just north of Grimstad where it takes a rather sharp southwest turn and continues in this direction across to Åsanfjord. Although the contact itself is nowhere exposed it is in several places possible to observe the rocks but a few metres away from it. None of the observed rocks of the banded series show evidences of having been mechanically or mineralogically affected by the gabbro intrusion. Rocks no. 6, 8, and 15 in Table 3 were all close to the gabbro contact. They have the mineral compositions and textural characteristics typical of the banded series. The only effect of the intrusion which can be observed in the banded series is a large scale bending of the strike of the banded series rocks around the gabbro (Fig. 3).

In the case of the gabbro, an intense pressing may be observed which extends in places up to 50 m. from the contact (Røshagen). Along the eastern side (along Malnesfjord) this squeezing is present as far north as Røshagen. About one km. south of Røshagen on an eastwards extending point the gabbro is so strongly foliated that it has the appearance of a greenschist (the green colour being dependent upon hornblende and serpentine). This foliated gabbro can be traced discontinuously southwards to Selvåg where the strike turns slightly westwards and on towards Grubklatten. At Selvåg the green, recrystal-

lized and strongly foliated gabbro is intercalated with the gray, less altered gabbro, i. e., zones of recrystallized gabbro alternate with normal gabbro — this is typical of the contact zone.

The dip of the foliated gabbro may be parallel to the dip of the contact. At Røshagen it ranges between 80° W and vertical, and at Selvåg it is about 70° W to vertical. The mineralogy of the gabbro in the contact zone differs with distance from the contact (greater alteration nearer the contact). Farthest away from the contact the gabbro can be seen to be mechanically affected and slightly foliated but without any recrystallization of the minerals. Olivine appears to have been the first mineral to be altered completely; it now is represented by a serpentine aggregate. The pyroxenes appear to have altered subsequently. The characteristic minerals of the green foliated gabbro near the contact are: plagioclase, hornblende, biotite, scapolite, serpentine, apatite and opaque minerals. In addition clinopyroxene occurs in the rather strongly altered rocks (but is normally absent).

The plagioclases are scapolitized in nearly all of this rock. The anorthite content is among the highest observed in the whole gabbro. Four of these rocks found to occur between Røshagen and Selvåg had An. contents of between 70 and 80 (determined on the universal stage) and one from Grubklatten showed An. 65. These high An. contents are assumed to be a primary feature; the unaltered gabbros close to the eastern border tend to have plagioclases in the An. 65 to 70 range which is slightly higher than those in the normal gabbros. It suggests that the deformation of the contact rocks took place under PT conditions where very basic plagioclases were stable in contact with H_2O.

The scapolite commonly occurs as well defined crystals either inside, or along the borders of plagioclase grains. Calcite (or perhaps a Mg rich carbonate mineral), which is present as traces in nearly all of the normal gabbros, has not been observed in any of the scapolite bearing contact rocks. Thus the formation of this scapolite may represent a metamorphic reaction according to the equation:

$$3 \text{ CaAl}_2\text{Si}_2\text{O}_8 + \text{CaCO}_3 \rightleftarrows \text{Ca}_4\text{Al}_6\text{Si}_6\text{O}_{24} \cdot \text{CO}_3 \text{ (Barth, 1952, p. 283): } \qquad 6$$
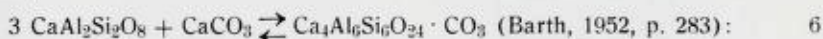
Because the concentration of scapolite is not very high, and because it is a rather brittle mineral, all attempts to separate it from the rocks have failed. Its refractive index relative to that of the coexisting plagioclase can be given. In one of the rocks (158/55) the plagioclase composition is An. 75 and the refractive indices range between 1,566—1,576. The refractive indices of the coexisting scapolite is $n_o >$ plagioclase and $n_e <$ plagioclase. The magnitude of $\dfrac{n_o + n_e}{2}$ may then be assumed to be approximately 1,570 which according to Tröger (1952) indicates the ratio Me./Ma. is closely similar to the An./Ab. ratio in the plagioclase.

Fig. 19. Banded gabbro from ridge between Dalvann and Lundvann.

Båndet gabbro fra ryggen mellom Dalvann og Lundvann.

This suggests that the scapolitization of basic plagioclases took place when the equivalent plagioclases were stable in contact with H_2O. In the contact rocks the PTX conditions were in the interval between the two reactions (scapolitization and saussuritization). Unfortunately the PT conditions are as yet not very well known for any of these reactions.

The hornblende is typically green but in some rocks is colourless (tremolite).

The biotite likewise has a normal pleochroism; yellow-light brown except where the hornblende is colourless; here, the biotite is also virtually colourless.

The gabbro along the contact between Grubklatten and Åsanfjord was nowhere seen to be foliated like that described above. This may be because it is not exposed close enough to the contact.

The normal gabbro.

The bulk of the intrusion is made up by the "normal gabbro" which is a gray, medium to coarse grained rock (in places pegmatitic). Banding is best developed in the central parts of the gabbro between the lakes Dalvann and Lundvann (Fig. 19) in the valley between

6

Table 19.

*Variations in mineral composition within the different gabbro zones
(arrows indicate presence of minerals).*

Zone	No	Plagioclase	Olivine	Hypersthene	Diopside	Hornblende	Biotite	Spinel
5	1	An. 48		2V$_x$59	2V$_z$53	2V$_x$72		
	2	48						
	3	50	2V$_x$84	71	53			
	4	54	84	72		85		
	5	54						
	6	54						
4	7	57						
	8	57	84	73	53			
	9	57						
	10	57						
	11	58						
	12	58	86	76	56	91		
	13	59						
	14	58						
	15	58						
3	16	60	90	82				
	17	61		76	55			
	18	63	88			90		
	19	64	sec. alt.	sec. alt.	sec. alt.			
2	20	68	90	81	55	86		
	21	68	88	77	55			
1	22	72		82				
	23	75	87	80	56			
	24	75	sec. alt.	sec. alt.	sec. alt.			
	25	77	sec. alt.	sec. alt.	sec. alt.			
ultra basic lenses	26	74	89	(85-92)		2V$_x$55		
	27		94	86				

Selvågtind (570) and Trihyrna (618), and in the gabbro associated
with the iron ore between Selvåg and Selvågtind. The banding is char-
acterized by bands rich in plagioclase (nearly anorthositic) intercalated
with pyroxenitic bands. Carstens (1957, p. 10) presented a photo-
graph of the banded gabbro of Napp, Flakstadøy in Lofoten, which
is very similar to the banded gabbro from Langøy. The dip of the
bands is nearly vertical (80° to 90°). This banding together with a
certain lamellar parallel orientation of the plagioclase grains defines
a strike and dip of the gabbro. These are marked on the map Fig. 18.

Plagioclase is the dominant mineral of all of the normal gabbros. In some cases it makes up about 90 per cent of the bulk and the "gabbro" becomes anorthositic. The plagioclases are typically free of any dust and none is anti-perthitic. Typically no signs of zoning are present. The plagioclase composition varies between An. 75 and An. 45. The An. content seems to decrease regularly from the outer contact towards the center (i. e., the area occupied by the monzonite intrusion between Tussen and Selvågtind). On the basis of this variation in the plagioclases the gabbro is divided into five zones. The characteristics of the different zones are illustrated in Table 19 and the positions of the samples examined are plotted in the map Fig. 18. The divisions are tentative as the number of samples on which they are based is small. However, the general trend is clear. Zones 1 and 2 define a narrow strip close to the outer contact of the gabbro. Zone 1 especially is to a large extent defined by the strongly altered contact rocks. However, gabbros containing unaltered olivine also occur. The alteration has not affected the plagioclase composition. Zone 5 which contains the most albitic plagioclases constitutes a narrow zone around the monzonite intrusion. The bulk of the gabbro occurs within zones 3 and 4. The interstitial plagioclases sometimes present in the ultrabasic lenses which are discussed later has a composition An. 70—75, comparable to the highest An.-content gabbros.

Olivine is typically present in all the gabbro zones (except in the secondarily altered contact rocks and in parts of zone 5 close to the monzonite contact). The amount of olivine in the rocks varies. It is the dominant femic mineral in some of the rocks and nearly absent in others. No regular variation in the olivine concentration throughout the different zones has been recognized. The size of the olivine grains varies greatly but no relationship between this and the total amount of olivine or to the zone in which they occur has been recognized.

The olivine is quite fresh except in the highly altered contact rocks. However, a rim of serpentine is commonly present on it where it contacts plagioclase. Nearly all the other femic silicate minerals locally form reaction rims around the olivine. In several cases more than one rim is present. The following sequences of rims have been observed;

olivine-plagioclase (no reaction rim)
olivine-serpentine-plagioclase
olivine-orthopyroxene-plagioclase
olivine-orthopyroxene-hornblende-plagioclase
olivine-clinopyroxene-plagioclase
olivine-hornblende-plagioclase
olivine-biotite-plagioclase

More than one type of rim is commonly present within the same rock.

The compositions of the olivines as determined by the size of the 2V on the universal stage show a fairly regular variation with iron content increasing from the outer contact (zone 1) towards the central part (zone 5), (Table 19). The variation covers a range between Fo. 88 and 75.

Orthopyroxene (bronzite) is present in most rocks. Its abundance varies greatly and in some rocks it is present only as reaction rims around olivine. It is present in all the gabbro zones (1—5) with a respective variation in composition from En. 85 to 80. (Deduced from the variation in the 2V.) In zone 5 the $2V_x$ drops from 71° to 59° over a distance of about 35 metres; this corresponds to a change in composition from about En. 80 to 70. The iron rich pyroxene is from a rock collected at a distance of about 4 m. from the monzonite contact. This rock is also characterized by a lack of olivine and is much finer grained than the normal gabbros.

Clinopyroxene (augite) is present in all of the gabbros and locally is the major femic silicate mineral. Determinations of refractive indices have not been carried out and little can be said about the variation in the chemical composition of the clinopyroxene from zone to zone. The $2V_z$ varies from 56° (zone 1) to about 52° (zone 5) which indicates a change in composition of the clinopyroxenes from the outer contact towards the centre (Table 19). The presence of exsolved magnetite (or ilmenite) lamellae is characteristic of the augites from all the zones (though not always present). These lamellae have the form of elongate rectangular rods with the elongation slightly inclined to the (110) cleavage.

Hornblende is present locally in all of the zones but is not common nor regularly present. The pleochroism is typically from colourless to yellow brown.

Biotite is present in insignificant amounts in nearly all varieties. The pleochroism is typically from colourless to yellow brown, but locally is more marked.

Green spinel is present in all of the gabbros throughout the whole series. Typically it occurs in amounts less than one per cent.

Iron ore minerals are also present in very small amounts in all varieties.

Calcite (or another carbonate mineral optically similar to calcite, Mg rich?) is likewise present in nearly all of the gabbros. It occurs as traces even in the more fine grained varieties close to the monzonite contact (inner part of zone 5). Its concentration in the rocks is typically very low, but in at least two of the normal central gabbros the amount was counted as between 1 and 2 per cent. The presence of calcite is interesting in that it indicates that the necessary amount of CO_2 to cause the scapolitization of the plagioclases in the outer contact rocks may have been available within the gabbro itself.

Three chemical analyses have been made of the gabbro (nos. an. 16, an. 8, and 32/56d, Table 20, from zones 2, 3, and 5 respectively). Two analyses given by Vogt (1909 p. 7) are also listed, but specimen localities are not known.

The ultrabasic lenses.

The outcrops of the ultrabasic lenses range from some cm.² aggregates to large bodies of some 100 m². All are typically very irregular in outline and elongated in one direction locally giving the impression that they are bands in the gabbro. The ultrabasic rocks are

Table 20.

*Chemical analyses, C. I. P. W. (molecular) norms, and observed
minerals in the Eidet- Hovden gabbro.*

Weight %	an. 16	an. 8	32/56 d	Vogt (1909)	
				a	b
SiO_2	46.65	48.72	46.78	45.56	49.43
TiO_2	0.13	0.23	0.69		0.54
Al_2O_3	23.07	20.56	16.18	19.06 [1])	19.01
Fe_2O_3	1.31	0.30	0.69	4.32	4.05
FeO	4.73	4.68	10.98	5.30	4.56
MnO	0.07	0.08	0.12		
MgO	9.01	9.81	12.65	14.55	7.57
CaO	10.69	11.83	7.36	9.01	11.59
Na_2O	2.69	2.60	3.11	1.90	2.27
K_2O	0.38	0.40	0.67	0.40	0.56
H_2O^-	0.16	0.13	0.08		
H_2O^+	0.97	0.37	0.39		
P_2O_5	n.d.	n.d.	0.35	0.016	0.013
CO_2	0.55	0.30	n.d.		
Sum	100.41	100.01	100.05	100.12	99.59

C. I. P. W. (molecular) norm.

Quartz (q)	–	–	–	–	0.3
K-feldspar (or)	2.0	2.5	4.0	2.5	3.5
Na-feldspar (ab)	23.0	21.7	21.2	16.5	20.5
Ca-feldspar (an)	49.0	42.2	27.8	40.7	39.8
(An. in plag.)	*(68)*	*(64)*	*(57)*	*(71)*	*(66)*
Nepheline (ne)	0.6	0.8	3.4		
Calcite (cc)	1.4	0.8			
Olivine (ol)	23.1	20.9	34.9	28.2	
Hypersthene (hy)				5.2	16.4
Diopside (di)		10.8	6.0	2.4	14.0
Magnetite (mt)	1.4	0.3	0.8	4.5	4.7
Ilmenite (il)	0.2	0.4	1.0		0.8
Apatite (ap)			0.8	tr.	tr.

Observed minerals.

Plagioclase	74.0	88.0	+		
(An. in plag.)	*(68)*	*(63)*	*(48)*		
Olivine	18.0	5.6	+		
Hypersthene	2.4		+		
Augite	4.0	3.1	+		
Hornblende		1.1	+		
Biotite	tr.	1.3	+		
Spinel	tr.	tr.	+		
Calcite	1.6	0.9	+		
Ore	tr.	tr.	+		

[1]) Herein included some TiO_2

(anal. an. 16, an. 8, E. Christensen, 32/56 d, L. B. Bolkesjø, a and b from Vogt, 1909 p. 7)

(nos. an. 16, an. 8, 32/56 d, refer to nos. 21, 18 and 4, Table 19)

consentrated in the part of the gabbro just south of and grading into the major magnetite-ilmenite deposit. From the shore at the small bay south of Selvåg they can be traced towards the southwest across the ridge between tops 470 and 381, just south of the lowest depression on the ridge, and farther in a nearly straight eastwest direction towards the south end of Troldalsvann where they apparently disappear. An ultrabasic lens occurs north of this in the depression on the ridge between top 381 and Selvågtind (570), but the gabbro to the north and west of this area (the major part of the intrusion) is free from ultrabasic rocks. Ultrabasic rocks are also absent south of this area across Spikerheia and southward to the southern contact, but in the far southwest on Gustadtuva there is a fairly large lens, and on the small island of Torvøen there are small lenses ranging in size from that of a matchbox up to ½ metre in largest dimension. This is of special interest because here in the southwest there is a low grade magnetite mineralization which has been traced from Gustad to Gustadtuva, top 305. At the roadcut close to the shore the impregnated zone is 150 metres broad but is narrower towards the top. This geographical relationship between the iron dominant magnetite-ilmenite deposits and the magnesium dominant ultrabasics may be genetically important.

All gradations between the ultrabasic lenses and the normal gabbro exist. This is shown by a variation in the plagioclase concentration. Plagioclase is localy completely absent but is more commonly present as small interstitial grains which become more abundant as the normal gabbro is approached. The mineral phases present are the same as in the normal gabbros but their relative proportions are different. Two examples of the ultrabasic rocks are listed in Table 19. The one without plagioclase is typical of the most basic of these and the other has plagioclase as small interstitial grains.

The plagioclase composition is about An. 70 which corresponds to the plagioclases of zone 1.

Olivine is the dominant mineral in these rocks. Its composition varies with the plagioclase content and it is extremely Mg rich in the plagioclase free types ($2V_z$ 86°, Fo. 100 per cent in no. 27). Where some plagioclase is present the olivine has a composition comparable with that of the olivines in zone 1 ($2V_x$ 89°, Fo. 87 per cent in no. 26). A much greater serpentinization of the olivines in these rocks reflects their more Mg rich character. The olivines in no. 27 are nearly completely altered and typically only the central cores are fresh.

Orthopyroxene is present in small amounts. $2V_x$ is about 88° suggesting a somewhat more Mg rich pyroxene than that of the normal gabbro (about En. 88) in the ultrabasic rocks. Some of it is pleochroic.

Clinopyroxene is present in very small amounts. For one of them (no. 26) the $2V_z$ was measured and found to be 54°.

Hornblende is present in small amounts. It is typically colourless (tremolite).

Biotite is present in small amounts. The pleochroism is from colourless to yellow.

Spinel is commonly present. It is colourless to very light green which suggests a much more Mg rich spinel than the darker green types in the normal gabbro.

Iron ore minerals are present as very small interstitial grains in negligible concentrations (the nature of the iron ore in these rocks was not determined).

Calcite (or a carbonate mineral, probably magnesite) is present as ill defined grains. It is commonly found inside the serpentinized olivines which suggests magnesite rather than calcite.

The magnetite-ilmenite deposit.

At one stage during the crystallization of the gabbro magma, conditions were such as to allow large quantities of iron and titanium oxides to crystallize.

The heaviest concentration of these oxides is found in the Selvåg valley about 600 to 800 m. west of the shore. It has been customary to divide the occurrence in two major lenses (here referred to as lens N and lens S, — see map Fig. 18). It is possible that the two lenses are connected by a zone of a low grade mineralization. (Extensive cover by large rock boulders in the critical area makes this point difficult to prove.) Carstens (1948) estimated the area of the largest lens (lens N) to be about 80 000 m². The lens N is elongate in a N—S direction with a length of about 540 m. and an average width about 150 m. The mineralization continues, however, as a low grade mineralized gabbro for about 2 kms. in a northerly direction and can be traced as far north as the ridge between Trollan (537) and Malnestind (528). The lens S is about 50 000 m.² in area and is elongate in a nearly straight NE direction. The length is about 500 m. and the maximum width roughly 200 m. A several metres broad band of normal gabbro occurs in this lens and reduces the effective "economic" area considerably. Lens S can be traced as a very low grade mineralization across the ridge east of Troldvann (of type similar to the low grade bands in the gabbro north of lens N). The direction of the bands is here more nearly E—W. The mineralization cannot be traced further in this direction. A low grade mineralization of the gabbro can be traced from Gustad in an eastnortheasterly direction to the top of Gustadtuva (305). It cannot

Table 21.

Minerals present in the Selvåg magnetite-ilmenite deposit.

Oxides		
	magnetite-ilmenite intergrowth green spinel ilmenite	decreasing amounts
Sulphides		
	pyrrhotite marcasite pyrite (chalcopyrite ?)	decreasing amounts
Silicates		
	clinopyroxene (green) plagioclase hornblende (brown) (biotite)	decreasing amounts

be traced any further and is not connected with the Selvåg deposits.

The impression is that the magnetite-ilmenite mineralized gabbro and the pyroxenite banded gabbro between Lundvann and Dalvann, together, form an arc. The mineralization defines the east "leg" with the heaviest ore concentration at the sharp curvature of the arc. The pyroxenite banded gabbro defines the western leg of this arc. Thus a relation between the "ore banded" and the "pyroxene banded" gabbro is indicated. The dips of both "legs" are nearly vertical; in lens N it is about 60—70° W, and in lens S about 70° NW to vertical.

Typically the ore is banded. This is the case in the low grade mineralized parts and in lenses N and S. The width of the bands ranges from less than one cm. to several m. In the two major lenses the unmineralized gabbro bands are generally rather narrow; however, several narrow bands may occur within small areas, thus notably lowering the iron content of the mineable ore. All gradations exist between gabbro and ore. The increase of ore minerals is accompanied by a proportional decrease of plagioclase. The most ore rich bands are nearly plagioclase free.

A large portion of ore from the central part of lens N was sampled. This sample was crushed and the minerals separated by help of heavy liquids and a Frantz magnetic separator. The fractions were

Table 22.

Fe and Ti content in the Selvåg magnetite-ilmenite ore.

analyses of ore	Fe/TiO$_2$	%	
		TiO$_2$	Fe
1, quarry in the lower part of lens S:	5.7	3.9	22.1
2, quarry in the central part of lens S:	7.3	3.2	23.2
3, quarry in the upper part of lens S:	6.2	4.2	26.2
4, quarry 22, lens N:	6.1	5.2	31.4
5, from the upper part of lens N:	6.9	3.6	24.8
6, general sample, lens N:	6.3	4.7	29.8

	ore			concentrate			waste		
	Fe/TiO$_2$	% TiO$_2$	% Fe	Fe/TiO$_2$	% TiO$_2$	% Fe	Fe/TiO$_2$	% TiO$_2$	% Fe
sample: 6	6.3	4.7	29.8	11.2	5.2	58.1	3.5	4.2	14.8
on sieves:									
+ 70 mesh	6.6	4.13	27.15	9.1	5.3	48.0	4.1	2.7	11.1
+ 100 mesh	6.6	4.21	27.80	10.0	5.4	53.8	3.5	3.5	12.2
+ 200 mesh	6.6	4.56	30.25	11.6	5.0	58.1	3.1	4.3	13.3
÷ 200 mesh	6.8	4.54	31.10	12.3	4.9	60.0	3.5	4.4	15.3
sample 5:									
on sieves:									
÷ 200 mesh	6.9	3.6	24.8	14.0	4.0	56.1	2.5	4.2	10.5
÷ 325 mesh * ...	14.0	4.0	56.1	17.5	3.6	63.0			13.6

* the concentrate from the ÷ 200 mesh fraction of sample 5 ground further down.
(analyses by Christiania Spigerverk).

studied through a binocular microscope and the minerals present identified on X-ray powder diagrams. Table 21 lists the mineral phases present.

The magnetic-ilmenite is the most dominant mineral. The Ti content of the magnetite (originally existing in solid solution in the mineral?) is exsolved into fine ilmenite lamellae (ilmenomagnetite, Buddington et al. 1955). Table 22 gives some Fe and Ti values of ore and concentrates. The table shows the Ti content in the iron concentrate to be higher than in the ore, and even a grinding down to 325 mesh before magnetic separation does not improve the iron

Table 23.

Analysis of ilmenite concentrate.

TiO₂:	30.0 %
Fe:	46.2 %
S:	14.57 %
Insoluble in conc. H₂SO₄:	1.0 %

Ilmenite (FeOTiO₂): 30.0 % TiO₂ + 26.98% FeO = 56.98% ilmenite
Rest Fe: 46.20 ÷ 20.97 = 25.23% Fe
Pyrrhotite (FeS)   : 14.57%S + 25.38%Fe = 39.97% pyrrhotite

(analyses by Christiania Spigerverk).

concentrate significantly. The width of the exsolved ilmenite lamellae is about 4—10 μ. Ilmenite is also present as independent grains (diam. 0.2—0.3 mm.) though in insignificant amounts. An ilmenite concentrate was obtained by treating the ilmenite rich fraction from the Franz magnet separation with Clerici solution. The heavy fraction which consisted of ilmenite, spinel and pyrrhotite, was analysed (Table 23). As can be seen in Table 23 all the iron (in fact a bit more than present) can be accounted for by converting the amounts of TiO_2 and S into ilmenite and pyrrhotite of ideal composition. This very pure ilmenite composition suggests that also the independent grains of this mineral were exsolved at low temperatures from one originally homogeneous magnetite-ilmenite solid solution mineral (titanmagnetite, Buddington et al. 1955). By taking the TiO_2 content in the Selvåg ore as roughly 5 % (Table 22) a temperature of 750—800° C ($\pm$ 100°) is obtained for the crystallization of the ore by plotting on Fig. 2 (p. 514) and Table 10 (p. 518) in Buddington et al. (1955). This is 100—200° C lower temperature than obtained by Buddington et al. for similar systems. However, a criterion for the validity of the thermometer is that ilmenite originally was present as independent crystals besides titanmagnetite. As stated above this is not definitely known in the case of the Selvåg ore. The use of this geothermometer has been criticized by Heier (1956 b).

The concentration of vanadium is about 0.17 % V in the ore and 0.40 % V in the iron concentrate.

The sulphides.

The ore assays between 0.1 and 0.7 % S (analyses by Christiania Spigerverk). Pyrrhotite is by far the most abundant sulphide. Marca-

Table 24.

Refractive indices of clinopyroxene in the Selvåg ore.

1: n_x; 1.680 ± (0.002), n_y; 1.690 ± (0.002), n_z; 1.710 ± (0.002). Strongly magnetic with numerous exsolved magnetite rods.

2: n_x; 1.683 ± (0.002), n_y; 1.693 ± (0.002), n_z; 1.713 ± (0.002). Very few exsolved magnetite rods.

site and pyrite are present in smaller amounts; marcasite is more abundant than pyrite. Whether marcasite is a primary or secondary mineral in the ore is not known.

Chalcopyrite has not been recognized in the ore, in either polished sections or through careful study of different ore concentrates. The ore contains about 0.03 % Cu. No. 5 Table 22 contained 0.19 % S and 0.032 % Cu. By primary flotation of this ore a concentrate was made with 2.06 % S and 0.91 % Cu (impurity was mainly green spinel). The Cu content of the ore suggests that chalcopyrite may be present.

The silicates.

Clinopyroxene is the dominant silicate mineral. The amount of plagioclase (An. about 55) varies inversely with the amount of magnetite. The clinopyroxene is an augite with exsolution lamellae of elongated magnetite rods parallel to (110). The amounts of exsolved magnetite in the pyroxene ranges within the same sample.

Actually the optic properties of pyroxene with and without magnetite lamellae differ. This variation is indicated in Table 24.

The monzonite intrusions.

Two separate monzonite intrusions occur within the gabbro. One of them can be studied over only a rather small area on Engenyken west of Bufjell and little can be said about the size of the intrusion. The second one occupies a comparatively large area from Selvågtind (570) in the south across Trollantind to Hovden in the north. Its extension towards south, east, and west is known; the northern boundary is covered by the sea. No exact contacts between either of the monzonite bodies and the gabbro have been seen, so it is not known whether the

monzonites are flat lying bodies on top of the gabbro or intrusions into it. A quartz-feldspar pegmatite exposed about 400 m. above sea level on the west side of the valley, between Røshagen and Selvågtind, may offer an answer. The monzonite-gabbro contact is not exposed, but the pegmatite is undoubtedly along the contact. This 3 m. broad pegmatite is mostly of pure quartz with red microcline and patches of biotite along the outer rim. The pegmatite strikes E—W and appears to be horizontal or with a slight dip towards the west. This suggests that the monzonite-gabbro contact is also close to horizontal and that the monzonite occupies a position on the top of the gabbro. Of course, this is a strictly local relationship.

The monzonite cannot be seen to change in appearance from the central parts towards the contact. The gabbro, however, is distinctly finer grained a few metres away from the contact than it is farther from it, and olivine tends to disappear close to the contact. This may be a contact metamorphic effect, but the disappearance of olivine is more likely but a manifestation of chemical variation of the gabbro from the outer contacts towards the centre.

The monzonite is definitely younger than the gabbro as small pegmatitic veins and irregular dykes of it occur in the gabbro. They contain the mesoperthitic feldspar which is characteristic of the monzonites. The dykes and veins are restricted in their occurrence to the vicinity of the parent monzonite bodies. A close association seems to exist between these veins and the fine grained dykes next discussed.

These monzonites are different in several respects from the porphyroblastic monzonitic granulites in the west of Bø (this was also noted by Vogt, 1909 p. 14) and the two rocks are probably not related temporally or petrogenetically.

In Table 25 two chemical analyses of each type of the monzonites are compared; there is a close similarity in chemistry between the two types. The variations between them do not appear to be greater than variations within the same units. Some significance may be put on the slight difference in degree of oxidation as expressed by the $FeO/FeO + Fe_2O_3$ ratios, the slight increase in MgO over $FeO + Fe_2O_3$ and of Na_2O over K_2O in the western monzonites versus the Eidet-Hovden monzonites. These differences, except possibly the last, appear to be too slight to be of any importance.

The two types of rocks are also rather similar megascopically. The monzonites within the gabbro intrusion are somewhat lighter in

Table 25.

Comparision between chemical analyses (major elements)
of monzonites in the Eidet-Hovden intrusion (1 and 2)
and the porphyroblastic monzonitic granulites (3 and 4).

Weight %	1 an. 15	2 an. 9	3 57/56	4 55/56
SiO$_2$	54.88	60.59	56.38	57.46
TiO$_2$	1.67	0.75	0.87	0.80
Al$_2$O$_3$	17.17	19.43	18.86	20.02
Fe$_2$O$_3$	3.72	1.22	3.15	2.53
FeO	4.69	2.40	3.78	2.64
MnO	0.20	0.08	0.11	0.09
MgO	1.96	0.68	2.46	1.38
CaO	5.51	3.38	5.53	4.81
Na$_2$O	4.84	5.86	5.22	5.55
K$_2$O	3.98	5.14	2.96	4.00
H$_2$O$^-$	0.19	0.17	0.08	0.25
H$_2$O$^+$	0.05	0.01	0.19	0.20
P$_2$O$_5$	1.17	0.32	0.38	0.45
Sum	100.03	100.03	99.97	100.18
FeO/FeO+Fe$_2$O$_3$:	0.56	0.66	0.54	0.51
MgO/FeO+Fe$_2$O$_3$:	0.23	0.19	0.35	0.27
CaO/Na$_2$O+K$_2$O:	0.62	0.37	0.68	0.50
Na$_2$O/K$_2$O:	1.22	1.14	1.76	1.39

C. I. P. W. (molecular) norm:

Quartz (q)	0.2	–	–	–
Potash feldspar (or)	23.5	30.0	17.5	23.5
Albite (ab)	43.5	52.0	46.5	48.5
Anorthite (an)	13.5	11.3	19.0	17.8
Nepheline (ne)	–	–	–	0.3
Olivine (ol)	–	1.3	0.3	3.0
Hypersthene (hy)	5.4	–	6.6	–
Diopside (di)	5.2	2.8	4.8	2.4
Magnetite (mt)	3.9	1.2	3.3	2.7
Ilmenite (il)	2.4	1.0	1.2	1.0
Apatite (ap)	2.4	0.5	0.8	0.8
Sum salic	80.7	93.3	83.0	90.1
Sum femic	19.3	6.8	17.0	9.9

(anal. E. Christensen)

Localities: 1. Selvågtind, 2. Roadcut, Tussen, 3. Skårvåg, 4. Roadcut, Steine.

colour as their feldspars are not so dark as those in the western mon-
zonites. Large alkali feldspar crystals are characteristic of both types.
However, whereas the potash feldspars constituting the porphyroblasts
of the western monzonites are similar in type to the smaller potash

feldspar grains in the rock, this is not the case for the monzonites within the gabbro intrusion. In these latter rocks it is necessary to distinguish between three different coexisting feldspar types. The phenocrysts are mesoperthites, and orthoclase perthites and plagioclase occurs in the groundmass.

The mesoperthitic feldspars consist of stringers of potash feldspar and plagioclase of nearly equal size.

Where the lamellae are broad enough twinning can be seen in the plagioclase phase. The orientation is the same in all the stringers within the same crystal and the feldspar has the appearance of being an antiperthite. The twin plane (010) is normal to the elongation of the stringers. On the other hand, where the potash feldspar stringers are broad enough it can be seen that they are perthites; the exsolved albite lamellae have the form of slightly elongated drops. Mesoperthites also occur locally as smaller crystals in the groundmass but they tend to be phenocrysts.

In one sample from the small monzonite on Engenyken phenocrysts were not seen; small mesoperthite crystals are the single feldspar phase. A few plagioclase crystals are present, but even these have stringers of potash feldspar penetrating into them in a direction normal to (010).

It was impossible to obtain reliable optical data on the potash feldspar and plagioclase phases of the mesoperthites. Microcline twinning or undulatory extinction was not noted in the potassic members; they appear to be pseudomonoclinic orthoclases.

In one rock (monzonite from Engenyken, 178/55) another type of alkali feldspar besides the mesoperthites forms phenocrysts. They are perthites of the string type, but not so thoroughly unmixed as the mesoperthites (in this case the alkali feldspar can easily be recognized as the host and albite as the exsolution product, which is not true of the mesoperthites). The alkali feldspar host has a relief (refractive index) closely similar to that of the exsolved albite lamellae, i. e., much higher than the potassic members of the mesoperthites and the normal perthites of the groundmass. The $2V_x$ was measured on three different grains (nos. 1, 2 and 3, Table 26) and found as a rule to be much higher than the $2V_x$ of the perthites from the groundmass in other samples of the monzonite (nos. 4, 5, 6, 7, same table).

In spite of the high 2V microcline twinnig cannot be observed although the crystals do show undulatory extinction. The 2V may range

Table 26.

*Optic axial angle of «anorthoclase» from Engenyken compared
with those of normal perthites in the monzonite groundmass.*

| | «Anorthoclase» | | | | | Groundmass perthites | | | |
	1		2		3	4	5	6	7
	a	b		a	b				
$2V_x$	57	70	78	71	76	45	48	56	52

optic plane $\perp$ (010).

1. Monzonite, west on Engenyken (a and b measured on different spots on the same «crystal»), 178/55.
2. Monzonite, as no. 1.
3. Monzonite, as no. 1.
4. Monzonite, roadcut, Tussen (an. 9).
5. Monzonite, as no. 4.
6. Monzonite, 440 m.a.s.l., Selvågtind (143/55).
7. Monzonite, Selvågtind (an. 15).

greatly within the same "crystal" which may be explained by a range
in triclinicity within the same "crystal" or also by assuming different
stages of unmixing. That these feldspars are not completely unmixed
is indicated by their higher refractive index (see above). Assuming
them to be chemically roughly equal to the mesoperthites of these rocks
(Or. 30—40, Ab. + An. 70—60, Table 38) and plotting the relation
between optic axial angles and composition of alkali feldspars on a
graph, a position intermediate between sanidine-anorthoclase and
orthoclase-low albite series is obtained (Fig. 20).

A similar position was obtained by Tuttle and Keith (1954) for
feldspars from the Beinn an Dubhaich granite (Skye, Inner Hebrides),
(higher in Or. because of the more granitic system). They consider
that granite to be "a missing link between granites and rhyolites"
(Tuttle and Keith, 1954 p. 70), meaning that it probably was intruded
into shallow depths.

The small potash feldspar grains of the ground mass are normal
orthoclase perthites in which the exsolved albite lamellae have the form
of elongated drops. The perthitic nature disappears towards the borders
of the feldspars where they appear optically homogeneous. They are
optically monoclinic (orthoclases) and their $2V_x$ ranges between 45°
and 56° (nos. 4—7, Table 26). It was impossible to separate the feld-
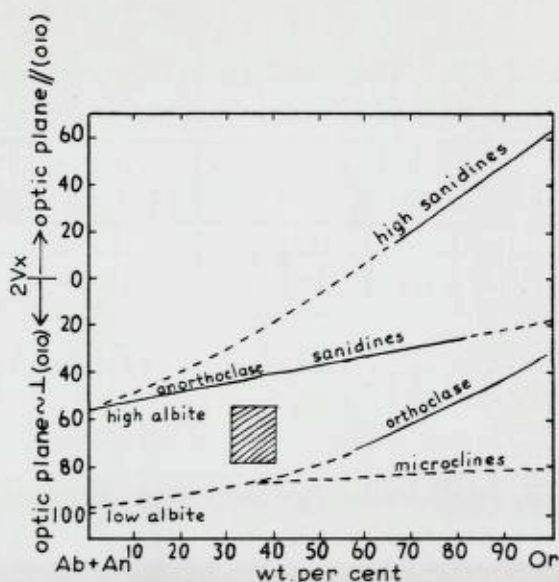
Fig. 20. Graph showing the relation between optic axial angles and composition of alkali feldspars (MacKenzie and Smith, 1955, after Tuttle, 1952). Shaded area: assumed position of feldspars from the Engenyken monzonite.

Diagram som viser forholdene mellom aksevinkler og sammensetning av alkali feltspater (MacKenzie and Smith, 1955, etter Tuttle, 1952). Det skraverte feltet viser den antatte beliggenheten av feltspatene fra Engenyken.

spars in the groundmass from the feldspar phenocrysts. Therefore, the chemical composition of the former is not known. As they are optically similar to the alkali feldspars of the charnockites they are probably similar chemically.

The plagioclases occur mainly in the groundmass surrounding the mesoperthite phenocrysts and are in general of a small size. This, together with the fact that they are nearly always zoned and have thin albite twin lamellae, mens that only a few crystals are suitable for exact optic determinations. (All the plagioclase determinations were made on the universal stage and diagrams and tables of Trøger (1952) and Turner (1947) were used). It was hoped to be possible to establish whether or not high temperature phases of the plagioclases were present, but because of the characteristics of the plagioclases mentioned this proved difficult. The plagioclase compositions and the different parameters determined are listed in Table 27. In most cases all of the parameters plot on curves and stereographic projections corresponding

Table 27.

Optics of plagioclases from monzonites within the Eidet-Hovden intrusion.

	1		2		3		4	5		6	7	8
	core	rim	core	rim	core	rim		core	rim			
X $\wedge \perp$ (010).........	90	88	90	87	94	94	92			89	88	90
Y $\wedge \perp$ (010).........	101	98	80	100	73	100	86			109	80	81
Z $\wedge \perp$ (010).........	11	8	10	12	14	12	5			12	10	9
X $\wedge \perp$ (001).........	92	91			97	97	85	94	94	87	90	83
Y $\wedge \perp$ (001).........	15	13			18	13	8	8	6	16	18	8
Z $\wedge \perp$ (001).........	74	78			104	78	96	96	90	75	80	96
max. ex. $\perp$ (010)	6	8	7	12	18	12	4			11	8	10
max. ex. $\perp$ (001)	18	12	13		18	18	12			25	15	14
$2V_x$	76	85	81	76	78	76	80	83	86	78	82	80
twinning, A; albite law, P; pericline law,......	A/P	A	A/P		A/P	A/P	P	P	P	A/P	A/P	A/P
derrived approx. comp. % An.	28	25	25	25	28	25	22	27	20	28	27	26

io low temperature symmetry. Those with the closest approach to high temperature symmetry are nos. 2 (rim) and 3 (core and rim). However, the optic axial angles ($2V_x$) are too large even in these cases. As the differences in the optics between high and low temperature plagioclases at this composition (An. 25 to 30) are small, and really exact measurements were difficult to make, it must be concluded that there is no evidence for the presence of the high temperature form.

Though this is not directly evident from Table 27 nearly all of the plagioclases are zoned. The zoning is dependent upon a normal variation in composition from core to rim. The difference in composition between core and rim may be as much as 7 per cent An. (The plagioclases of the surrounding banded series rock and the western monzonites are, as a rule, not zoned.)

The characteristic antiperthitic nature of the plagioclases of the granulite facies rocks is also absent. Irregular inclusions of potash feldspar may be present and stringers of potash feldspar locally penetrate them and give them a mesoperthitic appearance. However, none of the typical rectangular exsolution lamellae of potash feldspar have been seen. The femic silicate minerals in these rocks are hypersthene, diopsidic pyroxene, hornblende, and biotite.

7

Hypersthene is typically present in small amounts or absent altogether. It is rather abundant in one sample from the monzonite body in the west (Glimmerbugten 179/55). It is nowhere coloured and the optic axial angle which was measured in three cases was found to range between 49° and 52° (Table 28).

Diopsidic pyroxene (clinopyroxene) is typically the most abundant femic silicate mineral and in some places is the only one present. It is colourless or very light green. Some optical data are given in Table 29.

The pyroxenes richest in iron (as indicated by the optical data, nos. 4 and 5, Table 29) are very light coloured and have exsolved lamellae of magnetite parallel to the (110).

The diopsidic pyroxenes are commonly zoned (nos. 5 and 6, Table 29) so that there is a gradual increase in the iron content from the core towards the outer rim.

Hornblende was observed to occur in one rock only (Glimmerbugten, 179/55) where it is the most abundant femic silicate mineral. This rock is also characterized by having mesoperthite as the only feldspar phase (p. 94). Optical data are as follows: X-light brown, Y-very light brown, Z-brown-green, $2V_x = 82°$, $Z \wedge c \sim 20°$ (hypersthene which may be secondarily altered, diopside and biotite occur in the same rock).

Biotite is present in extremely small amounts in all the rocks examined.

Iron ore, apatite and zircon are accessory minerals in all of the rocks. The first two are present in amounts up to some percentages locally and iron ore is locally the dominant dark mineral. The amount of zircon is negligible.

The modal and normative compositions of these rocks correspond fairly well. Quartz is, in the majority of the cases, not present and it was nowhere seen to exceed one per cent of the rock.

The ratio between salic and femic minerals in the norms agree with the modes; the latter are subordinate in amount. It should be remembered that the feldspar composition cannot be derived from the norms because three feldspar phases, mesoperthite, orthoclase perthite, and plagioclase are present. The normative olivine of rock no. 2, (Table 25), does not agree with the mode but because biotite occurs, the silica deficiency can be corrected for by using equation 1 b, p. 29.

At the beginning of this chapter it was stated that although chemically similar, these monzonites are rather different from the western monzonites. The differences may be summed up:

Mesoperthites as phenocrysts in a groundmass of orthoclase perthite (drop-string type) and zoned plagioclase are characteristic of the monzonites within the gabbro intrusion. Mesoperthites are the dominant feldspar phase. In the western monzonites plagioclase antiperthite is the most abundant feldspar. Potash feldspar (drop-string perthite) is present as large rectangular porphyroblasts which appear

Table 28.

Optic axial angle of hypersthene.

2Vx	Localities
49 ..	Glimmerbugten, 179/55.
51 ..	Trollan-Malnestind, 185/55.
52 ..	440 m.a.s.l., Selvågtind, 143/55.

Table 29.

Optic axial angles and Z ∧ c of some diopsidic pyroxenes.

	1	2	3	4	5		6	
					core	rim	core	rim
2Vz:......	50	52	53	54	58	60	47	53
Z ∧ c:.....	43	42	39	43	42	45	43	46

Localities:

1, 2, 3, and 6. Monzonite, Trollan-Malnestind, 185/55.
     4. Monzonite, Glimmerbugten, 179/55.
     5. Monzonite, Selvågtind, an. 15.

to be optically rather similar to the groundmass orthoclase of the Eidet-Hovden monzonites.

The difference between the alkali feldspars of the two rocks are illustrated on the diagram Fig. 21. The alkali feldspars of the Eidet-Hovden monzonites (dots, nos. 50—53, Table 38), plot in the Ab.-An. rich field, whereas the potash feldspar porphyroblasts of the western monzonites plot (open circles, nos. 55—58, Table 38) in the Or. field. (The one feldspar from the latter group that plots on the cotectic curve was handpicked from the sample and is probably different from the rest because of plagioclase contamination.)

A further difference between the two rocks is the common zoning of the clinopyroxenes in the Eidet-Hovden monzonites and lack of this in the western monzonites.

In short it can be stated that the Eidet-Hovden monzonites are characterized by a mineral paragenesis not in equilibrium whereas the western monzonites appear to be in equilibrium.
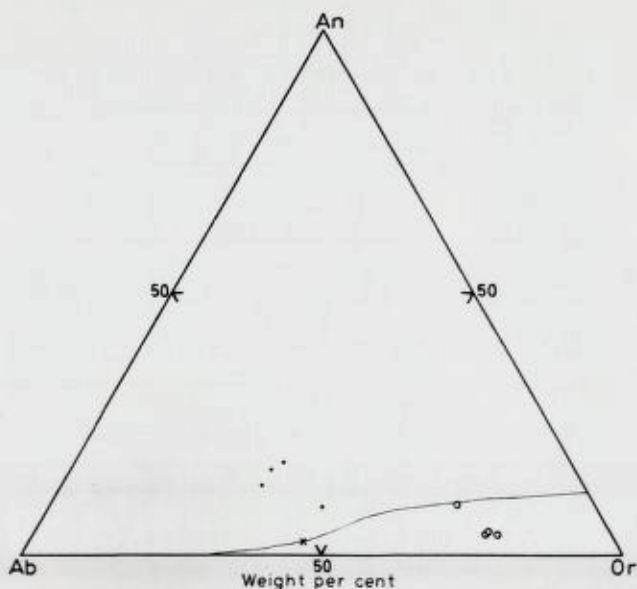
Fig. 21. Diagram showing the difference in composition between the alkali feld-
spars of the Eidet—Hovden monzonites (dots) and the porphyroblastic monzon-
itic granulites (open circles). Cross shows the position of the normative feld-
spar in the dyke no. 60/56.

*Diagram som viser forskjellen i sammensetning mellom alkali feltspatene fra
Eidet—Hovden monzonitene (prikker) og fra "the porphyroblastic monzonitic
granulites" (ringer). Kryss angir den normative feltspaten i gang bergarten
no. 60/56.*

The dyke rocks,

Dykes are abundant in the gabbro south of a line between Selvåg-
Selvågtind-Trihyrna and Sandvikknatten. North of this line dykes are
rare. Some can be observed along the coast between Selvåg and Røs-
hagen, and a small pegmatitic dyke of a type like the monzonite intru-
sion was observed between Ravnvann and Toftevann.

The dykes are too small to be shown on the map Fig. 18. Vogt
(1909, p. 22) stated that no corresponding massif (of deep seated
rocks) is found in the neighbourhood and that they probably were de-
rived from the large area of monzonitic rocks to the south (the banded
series and the porphyroblastic monzonites).

For reasons, which will be discussed later, this does not seem
probable. If these dykes are related to any deep seated rocks exposed,
it must be to the monzonites within the gabbro intrusion.
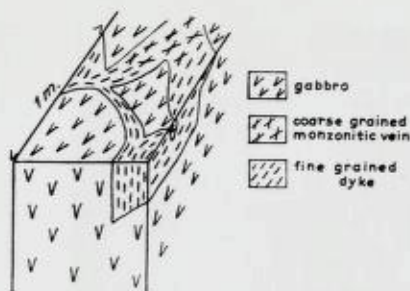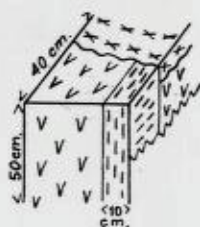
Fig. 22 a.   Fig. 22 b.

Fig. 22 a. Gradual transition between a "pegmatitic" vein of the "monzonite type" and a fine grained monzonitic dyke; top 381, south of Selvågtind.

Gradvis overgang mellom en "pegmatitisk" gang av monzonitisk sammensetning og en finkornet monzonit gang; top 381, syd av Selvågtind.

Fig. 22 b. Fine grained monzonitic dyke in gabbro which appears to be cut by a "pegmatitic" vein of "monzonitic type"; top 381, south of Selvågtind.

Finkornet monzonitisk gang i gabbro som synes å bli gjennomsatt av en "pegmatitisk" gang av monzonitisk sammensetning; top 381, syd av Selvågtind.

In fact, a certain relation between the two can in some cases be observed. Fig. 22 a shows a gradual transition between a "pegmatitic" vein of the "monzonite type" and a fine grained monzonitic dyke. However, in some places it appears that the former is younger and cuts the fine grained dykes (Fig. 22 b).

None of the dykes can be traced across the contact into the banded series rocks. On the contrary, the strike of the dykes seems to be roughly parallel to the gabbro contacts just as the strikes of the banded series rocks (as defined by the banding) appear to bend around the gabbro massif.

Most of the dykes cannot be traced over large distances. Although this is not necessarily because they pinch out, the impresion is that they do not reach significant lengths.

The thickness of the dykes ranges from a few centimetres to several metres. The thickest dyke occurs across Spikerheia. At top 517 it is 20 metres thick but further northeastwards it appears to split up into several smaller dykes. It can be traced for a distance of about 1500 m. (the longest distance any of these dykes could be traced) in an eastnortheasterly direction and dips 45° SE. At both ends it disappears beneath overburden and it was not seen along the continuation

Fig. 23 a.

of the strike. It has sharp contacts against the gabbro and gabbro inclusions are seen within it. Figs. 23 a and b show two smaller dykes. One of them (b) can be seen to end towards the left and one may get the impression that it represents a block inclusion and not a dyke. However, a careful study of numbers of these has convinced the writer that they are not inclusions of the country rocks.

The dykes have a dark greenish brown appearance on fresh surfaces; they resemble the banded series rocks, except that they lack banding. This resemblance between the two types is not surprising as they are chemically and mineralogically closely similar. The dykes are, however, more fine grained than the banded series rocks and their alkali feldspars never attain porphyroblastic size or shape.

Just as the banded series rocks are characterized by a heterogeneous chemical composition, the dykes also differ considerably (no variation was observed within any single dyke, however). The SiO_2 content of three analysed dykes differ by 14 wt. per cent SiO_2 (Table 30).

The quartz in no. 60/56 is present as small rounded grains with no sign of undulatory extinction. Because the quartz in the banded

Fig. 23 b.

Fig. 23 a and b. Fine grained dykes in gabbro; close to road between Solheim and Gustad.

Finkornede ganger i gabbro; nær vegen mellom Solheim og Gustad.

series rocks typically has a very irregular form and strong undulatory extinction, the appearance of the quartz in no. 60/56 is another feature that appears to favour the possibility that no relation exists between the two rocks. The dykes appear to be post metamorphic!

The high quartz content of this rock is quite extraordinary. A number of thin sections from other dykes were examined and quartz is normally not present in them. Nos. 50/56 and 91/56 are therefore, believed to be more representative of the normal dyke composition.

An alkali feldspar rich in Ab. and Or. is essentially the single feldspar phase in no. 60/56 although a few independent plagioclase grains also occur. Optically the plagioclase composition was determined to be pure albite, which is in good agreement with the norm. The alkali feldspar is a perthite (or antiperthite); the norm indicates about equal proportions (in weight per cent) of albite and orthoclase. This is also confirmed by the appearance in thin section. The perthite does not,

Table 30.

Chemical analyses, C.I.P.W. (molecular) norms of three dykes.

Weight %	50/56	91/56a	60/56
SiO_2	54.81	55.26	69.62
TiO_2	1.03	1.07	0.52
Al_2O_3	16.69	17.01	13.91
Fe_2O_3	4.21	4.13	1.41
FeO	5.25	4.93	1.75
MnO	0.20	0.16	0.04
MgO	3.16	2.38	0.55
CaO	5.58	4.93	1.64
Na_2O	5.00	5.35	4.24
K_2O	3.03	3.51	5.82
H_2O^-	0.13	0.18	0.17
H_2O^+	0.17	0.15	0.12
P_2O_5	0.73	0.88	0.12
Sum	99.99	99.94	99.91

(anal. E. Christensen)

C.I.P.W. (molecular) norm.

Quartz (q)	–	–	18.0
K-feldspar (or)	18.0	21.0	34.5
Na-feldspar (ab)	45.0	48.0	38.5
Ca-feldspar (an)	13.3	12.0	1.8
Olivine (ol)	3.9	4.2	–
Hypersthene (hy) ...	3.2	1.6	0.6
Diopside (di)	9.2	5.6	4.0
Apatite (ap)	1.6	1.8	0.3
Ilmenite (il)	1.4	1.4	0.8
Magnetite (mt)	4.4	4.4	1.5

Observed mode.

K-feldspar perthite	34.0
Plagioclase.....................	50.0
Hyp. + Diop....................	11.0
Biotite	1.0
Apatite........................	tr.
Ore............................	4.0

Localities

50/56, Spikerheia.

91/56a, roadcut, Gustad.

60/56, roadcut between Solheim and Gustad.

however, correspond morphologically with the normal mesoperthites; lamellae of the former are much more irregular both in form and size and the feldspar commonly has the appearance of a "film perthite". According to the norm the feldspar also has considerably less An.

Table 31.

*Counted mode of dyke rock no. 91/56a,
Table 30 compared with the counted mode of one rock from the banded series
which is taken close to the gabbro contact, no. 137/55, Table 3.*

	91/56a	137/55
Potash feldspar............	34	28
Plagioclase	50	51
Hypersthene, $2V_x$	(53)	(60)
	11	9
Diopside, $2V_z$........	(52)	(52—54)
Biotite	1	4
Ore.....................	4	7
Apatite	tr.	2
Calcite...................	tr.	–
Sum salic	84	79
Sum femic	16	22

content than that typical of the mesopherthite. It plots on the eutectic curve close to the feldspar eutectic on Fig. 21.

The typical dykes, as represented by nos. 50/56 and 91/56 a, are mineralogically closely similar to certain types of the banded series rocks. Normal drop-string perthitic orthoclase and plagioclase in the oligoclase-andesine range are typical of them both. The plagioclases in the dykes may exhibit undulatory extinction (zoning?). Hypersthene and diopsidic pyroxene normally occur together in both rocks. One difference between the two types is that biotite is in general not present in the dyke rocks, whereas in the banded series biotite is nearly always present in small amounts. The counted mode of no. 91/56 a, Table 30 is repeated in Table 31 and compared with one of the rocks from the banded series (the mode of no. 50/56, Table 30 is closely similar to that of no. 91/56 a).

It is interesting that the one member of the banded series which compares best with the dyke is from the area between Eidet and Selvåg (just west of Survikstrømmen) close to the gabbro contact. Contrariwise the two charnockites, nos. 58/56 and 59/56, Table 3 and 4, which are closer to this dyke, are chemically and mineralogically quite different from it.

Young red granites.

Unfoliated granites occur rather frequently in the granulite facies area. Although they have not been found to occur with rocks of lower metamorphic grade, small pegmatitic veins with a reddish feldspar which may be related to them do so occur. The veins are small and rather sparsely distributed in the gneisses; in general they are comformable with the foliation of the gneisses, but some crosscut the foliation.

Although the red granites occupy appreciable areas locally within the granulite facies rocks, they most commonly occur as patches of some tens m.^{2} or less in size. In three places they cover more extensive areas which appears to justify the use of the term granite massif (see map Fig. 3).

Although several of the smaller areas of the red granites are marked on the map, for practical purposes (scale) some are not. It must suffice to state that they are more frequent than the map picture indicates.

The granite contact with the surrounding rocks are sharp in the sense that the two rocks are distinct. It is not sharp in the sense of there being structural discordance between the two rocks. In the field one gets clear impression of a sudden "evolution" of the granites from the surrounding rocks (no hypothesis of origin is meant to be implied by this statement). The contacts can also be said to be gradational in the sense that by approaching the larger granite bodies from a distance an increasing amount of small pegmatites and granite veins and patches is apparent with the granulite facies rocks.

At one locality (Søberg) granite dykes can be seen to cut straight through the western monzonites. The strike is N 30° W and the dip 30° SW. One of these is as much as 1.5 m. wide. Nothing can be said as to their extension along the strike because they are exposed only close to the shore. They have the same colour and mineral composition as the normal granites but are more fine grained. Chemically they are extremely similar to the other young granites (no. 282/55, Table 32).

The red granites are typically very homogeneous, but locally xenoliths of the surrounding rocks occur within them.

No thermal contact metamorphic effects can be seen even in connection with the largest granite bodies. This appears to imply either that a subsequent regional metamorphism has destroyed the primary

Table 32.

Chemical analyses, and calculated modes of eight young granites.

Weight %	1 71/55	2 an. 5	3 347/55	4 211/55	5 208/55	6 282/55	7 254/55	8 270/55
SiO₂	69.9	70.90	72.9	73.4	74.0	73.8	74.1	75.6
TiO₂	0.20	0.22	0.22	0.20	0.12	0.18	0.18	0.17
Al₂O₃	15.7	14.20	13.8	13.9	13.9	13.7	13.6	13.1
Fe₂O₃	1.53	0.84	0.94	0.73	0.61	0.57	0.87	0.59
FeO	0.55	1.05	0.57	0.50	0.35	0.90	0.50	0.50
MnO	0.02	0.03	0.01	0.01	0.01	0.05	0.02	0.01
MgO	0.41	0.17	0.50	0.23	0.07	0.08	0.07	0.07
CaO	1.12	0.42	0.84	1.12	1.26	0.70	0.70	0.98
Na₂O	4.95	4.31	3.28	3.15	3.05	3.63	3.71	3.49
K₂O	5.11	5.75	6.40	6.23	6.16	5.83	5.70	5.00
H₂O⁻	n.d.	0.08	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
H₂O⁺	n.d.	0.03	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
P₂O₅	n.d.	–	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.

(an. 5, anal. E. Christensen, remaining seven samples, spec. anal. K. S. Heier).

Calculated modes.

Quartz (q).....	17.7	21.6	25.4	27.1	28.0	27.1	27.2	32.1
K-feldspar (or).	29.0	33.5	37.5	37.0	37.0	34.0	34.0	29.5
Na-feldspar (ab)	44.5	39.5	30.0	28.5	27.5	33.0	34.0	32.0
Ca-feldspar (an)	5.5	2.5	4.0	5.5	6.5	3.5	3.5	5.0
Biotite (bi)	2.0	2.0	2.2	1.0	tr.	2.0	tr.	0.7
Ilmenite il)	0.4	0.4	0.4	0.2	0.2	0.2	0.2	0.2
Magnetite (mt).	0.6	0.9	0.6	0.9	0.7	0.6	1.0	0.6
Hematite (hm) .	0.7	–	0.3	–	–	–	–	–
Sum	100.4	100.4	100.4	100.2	99.9	100.4	99.9	100.1

71/55, large granite massif, Stenheia.

an. 5, large granite massif, Torset.

347/55, small granite body, n. Rygge.

211/55, small granite body, W. of Eidspollen.

208/55, small granite body, Rissjøen.

282/55, granite dyke in porphyroblastic monzonite, Søberg.

254/55, minor granite massif, Veatind-Nåla.

270/55, small granite body, Bergodden, Skårvåg.

effects or that the regional PT conditions during emplacement of the granites were similar to those under which granitic magmas can be stable. The latter interpretation is favoured by the author. Slight molybdenite mineralization can be seen in the gneiss about 100 m. north of Nipa, adjacent to the contact with the granite between Romsetfjord and Skjelfjord (Saltberggranit, Vogt, 1909). Apart from this can no mineralization be traced from the granites.
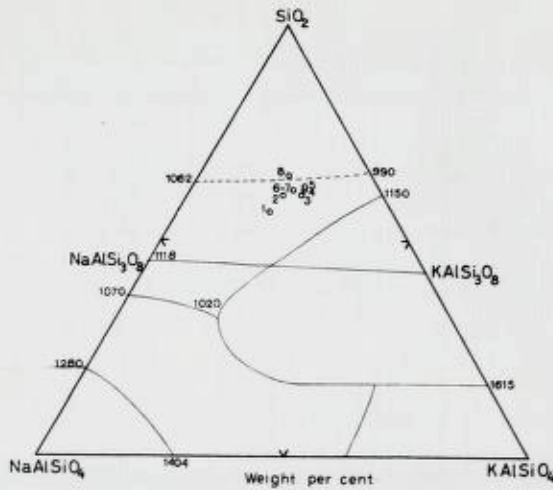
Fig. 24. Plots of red granites in the system NaAlSiO₄-KAlSiO₄-SiO₂
(Schairer, 1950).

Rød granitter plottet i systemet NaAlSiO₄-KAlSiO₄-SiO₂
(Schairer, 1950).

The grain size varies from aplitic to pegmatitic. The medium grained nonporphyritic types are most common. They commonly exhibit graphic granitic textures. In the more coarse grained varieties the quartz often has the form of irregular veins of a grayish blue colour (Torset).

Chemical analyses of eight granites are given in Table 32. The calculated modes as listed in the same table correspond closely to the observed modes. (C. I. P. W. (molecular) norms are not presented for these rocks but they would correspond closely to the modes apart from the fact that the small amounts of biotite would be expressed as hypersthene plus some additional potash feldspar.)

They are dominantly quartz-feldspar rocks with the femic minerals (biotite + iron ores) less than 4 mol per cent. The analysed granites are plotted on the equilibrium diagram for the system NaAlSiO₄ — KAlSiO₄ — SiO₂ (Schairer, 1950) in Fig. 24. It can be seen that they fall in the low temperature trough in the system (Bowens "residua" system).

The mineralogy of the red granites, as can be seen in the Table 32, is simple and uniform.

Quartz has marked undulatory extinction. The grain size ranges from minute grains to irregular stringers two cm. long. It is typically colourless to gray, but locally bluish. In the latter case, thin rod like inclusions (rutile?) can be seen to give the quartz the so-called "sagenite" structure (no. an. 5). Myrmekitic quartz can be seen in rocks in which independent plagioclase grains are relatively abundant. All myrmekite occurs inside plagioclase close to potash feldspar grains (in Table 32 myrmekitic quartz is noted as occurring in rocks nos. 211/55 and 254/55).

The alkali feldspar is mesoperthite in the young red granites and this serves to distinguish them from other, older red granitic rocks that occur with the granulite facies rocks. The symmetry of the potash feldspar is difficult to determine because of its lamellar structure. In some rocks microcline twinning can be seen through the mesoperthite texture (nos. 270/55, 282/55, 208/55, 211/55 in Table 32); in others the alkali feldspar has an undulatory extinction characteristic of those with intermediate triclinicities (nos. 254/55 and 347/55). In rocks in which the potash feldspars at a first glance seem to be optically monoclinic (orthoclases) further investigations have so far always shown these also to be of intermediate triclinicities (nos. 71/55 and an. 5). The feldspar from no. an. 5 was examined by X-ray and a triclinicity (Goldsmith and Laves, 1954 a) of 0.39 was measured. $2V_x$ in one crystal measured was found to range between $67° - 73°$ and $x^1 \wedge (001)$ on (010) was about $10°$. This feldspar was chemically analysed (no. 75, Table 39). Plotted on the diagram Fig. 25 (cross) it can be seen that it falls in the potash feldspar field to the right of the feldspar eutectic point.

Plagioclase is present in small amounts relative to the alkali feldspar. It is rarely more than 5 per cent in the mode and typically less than this. In some it is absent and the mesoperthite is the only feldspar phase (no. 71/55, Table 32). Perhaps the rocks should, therefore, be termed quartz syenites? The plagioclases are not antiperthitic but potash feldspar veins can be seen to penetrate into some of them. Their grain size is small. Because of this and the small amount present the optical determinations of their compositions were rather difficult. One might out of general considerations believe them to be albitic. However, careful determinations on the universal stage of the plagioclase in no. 254/55 and an. 5 suggest An. 10—15 ($2V_x \sim 95°$, n, slightly less than quartz, but definitely higher than balsam, max. ex. $\perp (010) \sim 10°$). A good idea of the plagioclase composition can be obtained by comparing the analyses of the alkali feldspar (no. 75, Table III) and the mode of the corresponding rock (no. an. 5, Table 32). By relating the two An. 12.5 is obtained for the plagioclase composition (disregarding possible solid solution of potash feldspar in the plagioclase). This satisfactorily corroborates the composition An. 10—15 deduced optically.

Biotite and iron ore are the other two minerals present in all the rocks. The biotite is typically present only in traces and has a pleochroism colourless to yellow brown or greenish.

Muscovite and zircon are present as accessory minerals in some sections.

In Fig. 25 the feldspar components of the red granites from Table 32 are recalculated to 100 weight per cent and plotted on the projection
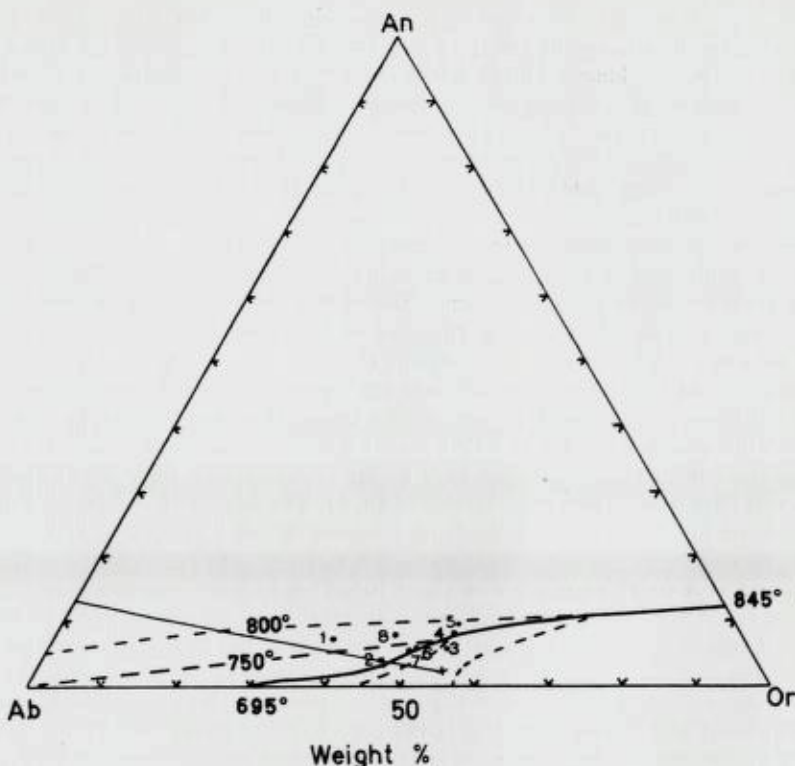
Fig. 25. Plots of normative Or+Ab+An of the young red granites on the projection of the quaternary system $NaAlSi_3O_8(Ab)$-$KAlSi_3O_8(Or)$-$CaAl_2Si_2O_8(An)$-H_2O at 5000 bars water pressure (Yoder et al., 1957. x, separated alkali feldspar from no. 2; tieline connects it with the coexisting plagioclase (optically determined).

Plot av normativ Or+Ab+An for "the young red granites" på projeksjonen av firekomponent systemet $NaAlSi_3O_8(Ab)$-$KAlSi_3O_8(Or)$-$CaAl_2Si_2O_8(An)$-H_2O ved 5000 bars vann trykk (Yoder et al., 1957). x, utseparert alkali feltspat fra no. 2. Tynn linje forbinder denne med den tilhørende plagioklas (optisk bestemt).

of the quaternary system Ab.-Or.-An.-H_2O at 5000 bars H_2O pressure (Yoder, Stewart and Smith, 1957). The cotectic curve defining the composition of magmas yielding potash feldspars and plagioclase as independent phases upon cooling is indicated. It is interesting that the one granite which plots at an appreciable distance away from this curve is the only one where mesoperthite is the only feldspar phase (no. 71/

55, Table 32). The remaining plots cluster closely along the curve. This should indicate that the rock composition is closely identical to that of the crystallizing magma.

The 750° and 800° isobars are also indicated on Fig. 25. It is seen that the rocks correspond to crystallizing temperatures between 725° and 800°.

Using the feldspar geological thermometer (Barth, 1956) a temperature of 780° was derived for no. an. 5. (The tie lines between the rock and the two feldspars are indicated on Fig. 25). As read from Fig. 25 this temperature should be about 725° at 5000 bars. The confining rock pressure during metamorphism in granulite facies is probably less than 5000 bars (3000—4000 bars was suggested by Barth (1952 p. 349)). A lower pressure would increase the temperatures in Fig. 25. The similarity between the two temperatures is in any case quite satisfactory.

The crystallizing temperatures for these rocks are at least 100° higher than that what was determined for the granulite facies rocks in which they occur. This agrees with the field evidence that they are late or post metamorphic intrusive rocks.

The metamorphic facies.

General discussion.

The rocks on Langøy exhibit diverse mineral facies for the same chemical compositions. The island may be divided into two major facies areas separated by a line that extends from Varvik in the southwest to the west end of Gåsfjord in the north. (Although the northernmost part of Langøy north of an E—W line from Gåsfjord to Utskår was not included in this work, reconnaissance work therein indicates a N—S extension from Gåsfjord of this division line.) The division line (facies boundary or isograde, Tilley, 1924) coincides with the thrust plane beneath the anorthosite over Slåttnesfjell (651) and Ole Hansatind (734), (Fig. 10). The facies boundary is marked on the map Fig. 3 as a solid line where it can be observed as a thrust zone with intense mylonitization beneath the anorthosite and for a short distance to the northeast where the charnockite border series rocks overlie the lower grade metamorphic red gneisses over Knutlitind and towards Oshaug. The rest of the facies boundary is dashed. Here no mylonitization was observed in the field, and the position of the boundary was established on the basis of thin section studies. That this part of the boundary is a result of faulting is also indicated by a change in the strike (as defined by banding and foliation of the gneisses) across the boundary and by the fact that the change in mineral facies is abrupt. Crushing accompanied by crystallization of fine grained biotite is seen in thin sections of some rocks close to the facies boundary.

The irregularity of the boundary line on the map (Fig. 3) reflects the dip of the thrust plane and the irregular topography of the area. The thrust plane proper strikes about E—W and dips towards the north.

The rocks east of the boundary exhibit mineral parageneses corresponding to the amphibolite facies and those west of the boundary, granulite facies mineral associations. The latter area, by far the larger,

is intersected by a zone of retrograde metamorphic gneisses (with mineral associations corresponding to amphibolite facies) that strikes from Maars up to the north end of Jørgenfjord, and possibly as far north as Kråkberget.

The granulite facies.

It was implied throughout the petrographical descriptions that the mineral associations of the rocks are metamorphic and not primary magmatic (the Eidet-Hovden intrusion is not included in this discussion except where specifically so stated). That this is the case for the banded series and the charnockite border series is obvious from their general appearance in the field. Vogt (1909, p. 16), however, assumed that the banding or veining reflected in situ magmatic differentiation processes, or that the acid bands were injected into the rocks at a late stage in the differentiation. Some in situ metamorphic differentiation probably took place in these ultrametamorphosed rocks; but, it also appears possible that the heterogeneity of the rocks could to a large extent reflect inherited primary sedimentary differences. The occurrence of undoubted sedimentary rocks within the gneiss series suggests that the major portion of the banded rocks were formed at the expense of graywackes, shales, sandstones, lavas, etc. which were originally deposited in the geosyncline with the still recognisable metasedimentary rocks (Ramberg, 1951, p. 29). Vogt, 1909, pp. 25—32, suggested that the metasedimentary rocks are contact metamorphic inclusions in an igneous rock, but in the author's opinion the field evidence clearly shows that they were folded and metamorphosed together with the gneisses.[1]

The massive porphyroblastic monzonitic granulites do not show any signs of original heterogenities and could be assumed to have formed by crystallization of a magma. Their mineral parageneses are,

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[1] The "recognizable metasedimentary rocks" are graphite schists and marbles which are extremely resistant to regional metamorphic alterations (see Barth, 1952, Table 72, p. 358). Quartzites are recognizable as such in the eastern amphibolite facies area, whereas they are not (if they occur) in the granulites in the west. Mica schists are not present in either area. Holtedahl (1944 p. 10) wrote: "Typical constituents of parts of the northern areas just mentioned (outer districts of Troms) are deposits of graphite which may be regarded as remains of alum shales, left after the granitization processes had thoroughly altered the character of adjacent sediments of a more normal type".

however, identical to those of chemically similar rocks within the banded series. In any case irrespective of their origin (igneous or sedimentary), today they are metamorphic rocks.

The same data apply to the old red granite.

The evidence that indicates that the rocks in question are metamorphic includes the generally observable granular texture, the freshness of the minerals, and the occurrence of antiperthite with regularly oriented blebs of potash feldspar commonly together with perthite. Furthermore, hornblende (and biotite) are characteristic of the basic rocks and rhombic pyroxene of the acid rocks which is the opposite of what is to be expected according to Bowen's reaction series.

Parras (1958) in the most recent survey of charnockitic rocks, found (p. 32) that before 1920 the view that charnockites are primary magmatic was dominant, stimulated by Holland's (1900) original view, but that the majority of descriptions since 1920 have favoured a metamorphic origin.

It is difficult on the basis of one sample to decide whether the mineral paragenesis is igneous or metamorphic. The theoretical catanorm agrees fairly well with the mineral associations obtained by crystallization of a "dry" magma; however, this corresponds, at least qualitatively, to the mineral parageneses for rocks with excess SiO_2 which are stable under the conditions of granulite facies. For the rocks deficient in SiO_2, one important difference exists. Olivine is not generally considered stable in the metamorphic assemblages but is in the igneous. It has not been observed to occur in any of the rocks of proper chemical compositions; the silica deficiency is accounted for by hornblende and biotite.

Some mineralogical variations.

When the rocks are regarded as members of a series some interesting mineralogical features become apparent. In Fig. 26 are given the arithmetic means of the concentrations of the various femic minerals in the granulite facies rocks over 5 per cent SiO_2 intervals (from Tables 1, 3, 6, 11, 15 nos. 77/56 and 81/56) and calculated as percentages of the sum of femic minerals considered as 100 per cent. Although the number of samples is, of course, too small to treat statistically, the trends are considered suggestive.
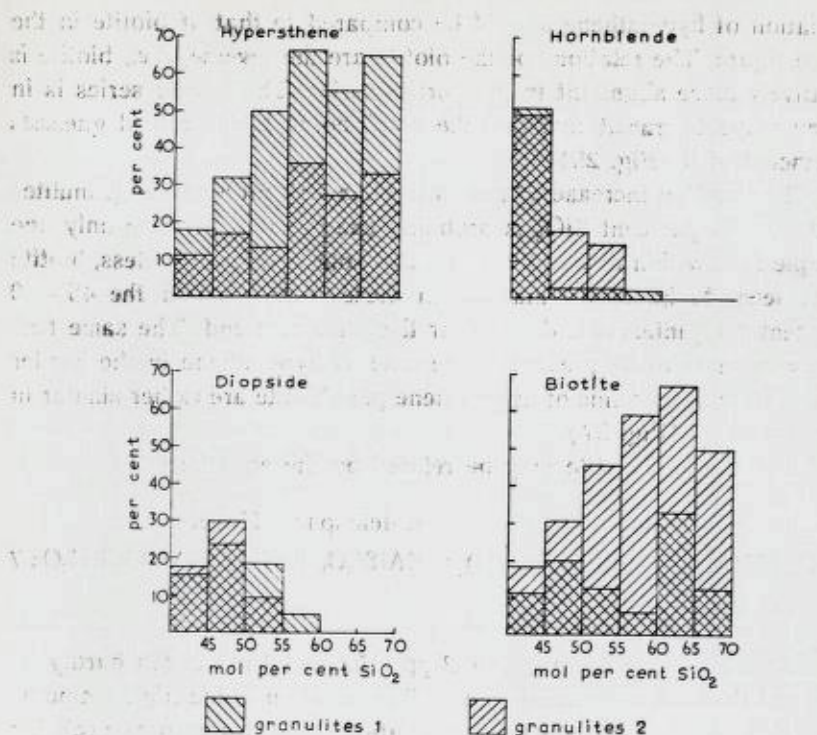
Fig. 26. Variation of femic minerals in granulite facies rocks. Diagrams show the arithmetic mean per cent for each of the femic minerals with respect to the sum of the femic minerals (reckoned as 100 %) as a function of the SiO$_2$ content calculated from the mode. (Granulites 1 and 2, see text.)

Variasjonen av femiske mineraler i granulit facies bergarter. Diagrammene viser det aritmetiske middel i % for hvert av de femiske mineralene med hensyn på summen av de femiske mineralene (regnet som 100 %) som en funksjon av SiO$_2$ innholdet beregnet av moden. (Granuliter 1 og 2, se teksten.)

The amount of hypersthene relative to the sum of the femic minerals shows a regular increase with increasing SiO$_2$ in the rocks. Above the 55—60 per cent SiO$_2$ interval the concentration is more nearly constant. This is the composition at which free quartz appears in the banded series. Within the charnockite border series (granulites 2) the concentration of hypersthene is lower than it is in the remaining granulite facies rocks (granulites 1), although the variation of hypersthene shows the same trend in both cases (Fig. 26). The

variation of hypersthene should be compared to that of biotite in the same figure. The relations of the biotite are the reverse, *i. e.,* biotite is relatively more abundant in the border series. The border series is in many respects transitional into the amphibolite facies veined gneisses northeast of it (Fig. 29).

The sudden increase in the biotite concentration of the granulites 1 at 60—65 per cent SiO_2 is probably exaggerated because only one sample falls within this interval (no. 2, Table 11). Nevertheless, biotite does seem to have two maxima in these rocks; one in the 45—50 per cent SiO_2 interval and one near the silica-rich end. The same tendency appears to be present in the case of hypersthene in the border series. In fact, the sums of hypersthene plus biotite are rather similar in the two cases (Fig. 27).

Hypersthene and biotite may be related by the equation:

<div align="center">

Biotite   K-feldspar   Hypersthene

</div>

$$3SiO_2 + K(Mg,Fe)_3AlSi_3O_{10}(OH)_2 \rightleftharpoons KAlSi_3O_8 + 3(Mg,Fe)SiO_3 + H_2O : 7$$
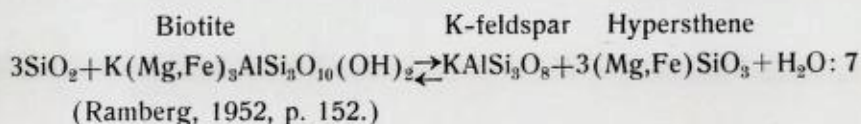
(Ramberg, 1952, p. 152.)

As stated by Ramberg (1952 pp. 152 and 158) it can hardly be doubted that the right hand side of this equation is the high temperature side. Assuming constant temperature (and water pressure) the direction of the reaction depends upon the SiO_2 content and the Mg/Fe ratio of the rocks (Ramberg, 1952, p. 158). An increase in both quantities lowers the equilibrium temperature of the reaction (thus favouring minerals of the right side at constant temperatures). The SiO_2 values (wt. per cent) and some relevant element ratios of the chemically analysed granulites are listed in Table 33. By comparing pairs of closely similar SiO_2 concentrations (e. g., 145/56 with 151/56 b and 150/56 b, 55/56 with 158/56, and 59/56 with 142/56) it becomes evident that the relative increase in biotite in granulites 2 is not due to any difference in the Mg/Fe ratios. The evidence appears to favour a displacement of the equilibrium towards the left in eq. 7, governed by variations in the regional PT conditions; thereby the border series would be transitional into the amphibolite facies gneisses.

The term granulite facies is here used to designate rocks in which hypersthene is a stable mineral phase. Equation 7 shows that an increase in SiO_2 will favour the formation of hypersthene. This means that a lower part of the granulite facies may be recognized as belonging
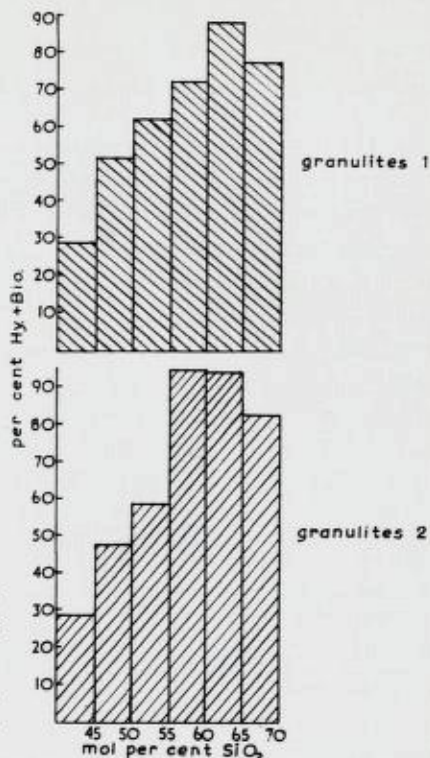
Fig. 27. Arithmetic means of hypersthene + biotite in the granulite facies rocks (calculated as for Fig. 26).

Aritmetisk middel av hypersthen + biotit i granulit facies bergarter (beregnet som for Fig. 26).

to the PT conditions where hypersthene is stable in rocks with excess SiO_2 only. (Highest half of amphibolite facies, Ramberg, 1952, p. 156.) An intermediate group may be defined as where hypersthene and biotite are both stable and occur together in the more basic members. Based on the data of Fig. 26 one can assume that a higher part of the granulite facies could exist where biotite is unstable even in rocks low in SiO_2. It is difficult to ascertain from the literature whether this stage has ever been reached in large volumes of rocks which have undergone "regional metamorphism".

Some information given by the feldspar geological thermometer (see later) and the distribution of the young red granites may be taken to indicate that the intermediate part of the granulite facies grades upwards into the region of anatexis.

Both granulites 1 and 2 should then represent intermediate granulite facies conditions. It is evident from the petrographical tables that

the Hy./Bio. ratios even for rocks within the same groups range over wide limits around the arithmetic mean illustrated diagramatically in Fig. 26. On the basis of their areal distribution it is difficult to ascribe this to variations in regional temperature and pressure.

On the basis of the information given in Table 33 it is also difficult to ascribe the variations to differences in the element ratios. However, the equilibrium conditions of eq. 7 are also dependent upon the water pressure. In geological considerations this is generally identified with the effective external pressure P or else the most of it, as the partial pressures of other gases are considered to be relatively low. However, the vapour pressure certainly may range from place to place within the same region due to differences in permeability and to structural control. Therefore, water should be incorporated in the X part of the physico-chemical PTX system (Yoder, 1952).

Small differences in the water content may explain the variations in the mineral ratios observed within the same metamorphic groups but is unlikely alone to be responsible for the regular variations in metamorphic grade from the granulites 1 via the granulites 2 to the amphibolite facies gneisses.

Hornblende is the dominant femic mineral in the most basic rocks of granulites 1 and 2. Its frequency of occurrence diminishes rapidly with increasing SiO_2 concentrations in the host rocks and it is not present where quartz occurs as an independent mineral phase (Fig. 26). Ramberg (1952, p. 70) states that a deficit in SiO_2 in the rock will expand the stability field of hornblende towards a higher degree of metamorphism. This is because the hornblendes, which are stable at high temperature and pressure, have two of the eight silicon positions occupied by aluminium, and therefore, behave as if they were unsaturated with reference to silica. The lower metamorphic grade of the amphibolite facies gneisses in the east is thus reflected in the stability of hornblende in rocks of higher SiO_2 content (Fig. 29).

The arithmetic means of the hornblende concentrations in the most basic granulites 1 and 2 (40—45 per cent modal SiO_2) are closely similar. This is because the two amphibolites of the sedimentary series (nos. 77/56 and 81/56, Table 15) were incorporated in granulites 1. If the charnockite border series is compared with the banded series alone a much higher average hornblende concentration is found in the former (compare Tables 3, and 6).

Table 33.

*Tabulation of some elements and element ratios which govern
the stability of hornblende in granulite facies rocks.*

No.	SiO₂	K₂O	Na₂O/CaO	MgO/FeO	MgO/(FeO+Fe₂O₃)	TiO₂/(MgO+FeO+Fe₂O₃)	Hy.[1]/Bio.	Hy.[1]/Hbl.
Granulites 1.								
81/56	42.5	0.67	0.24	0.63	0.39	0.063	0/3.1	0/64.6
77/56	46.2	0.87	0.35	0.77	0.49	0.074	1.5/tr.	1.5/88.2
58/56	48.87	0.87	0.64	0.51	0.32	0.11	76/tr.	76/0
145/56	51.85	0.81	0.26	1.23	1.13	0.032	31.5/1.9	31.5/18.5
an. 12	54.62	3.00	0.81	0.66	0.36	0.086	13/19.5	13/0
57/56	56.38	2.96	0.94	0.65	0.35	0.093		
55/56	57.64	4.00	1.15	0.52	0.27	0.12	14.3/7.1	14.3/28.6
59/56	59.27	1.17	0.94	0.83	0.64	0.037	87.5/0	87.5/0
Granulites 2.								
151/56 b	49.20	1.58	0.28	1.12	1.00	0.057	37.7/30.2	37.7/17
150/56 b	51.37	1.99	0.18	1.33	1.19	0.055	35.7/41.1	35.7/tr.
158/56	57.44	2.89	1.04	0.57	0.47	0.066	36.4/59.1	36.4/0
142/56	62.79	3.96	1.35	0.55	0.42	0.042	61.1/27.8	61.1/0
150/56 a	67.98	4.73	1.96	0.61	0.40	0.084	25.0/75.0	25.0/0
151/56 a	72.26	4.83	1.61	0.55	0.39	0.089	66.7/33.3	66.7/0

[1] The contents of Hy., Bio., and Hbl. are given in per cent of the sum of the femic minerals (reckoned as 100 per cent).
Nos. 81/56, 77/56 are from Table 15.
  58/56, 145/56, an. 12, 59/56 are from Tables 3 and 4.
  57/56, 55/56 are from Tables 1 and 2.
The granulites 2 are from Tables 6 and 7.

In the granulites 1 hornblende is also stable in some rocks in the 55—60 per cent SiO₂ range, less than 1 per cent in one of the western monzonites (no. 280/55, Table 1), and 0.5 per cent in the granulite no. 201/55, Table 3. The latter rock belongs to the banded series near the western monzonites. By comparing the western monzonites with the rocks of the banded series (Tables 1 and 3) it becomes apparent that hornblende tends to be stable in the former in the intermediate silica range but not in the latter. Ramberg (1952, pp. 68—70) lists a number of factors influencing the stability of hornblende in granulite facies rocks. The relevant element concentrations and ratios for hornblende are listed in Table 33. Nos. 57/56 and 55/56 in this table are western monzonites. Hornblende is present in

small amounts in both of them. It is difficult to relate the chemical data to the stability of hornblende. The $TiO_2/MgO + FeO + Fe_2O_3$ ratios in the western monzonites are relatively high and this should expand the stability field of hornblende in the higher temperature range. This ratio is, however, comparable to that of the ordinary granulite no. 58/56 (same table) in which hornblende has not been observed even though this rock is lower in SiO_2, K_2O and the Na_2O/CaO ratio, factors each of which should tend to stabilize hornblende in the granulite facies. The MgO/FeO ratios are roughly similar. One gets the impression that the grain size of the rock may have influenced the mineral stabilities to some extent. This possible control is also suggested in the case of the potash feldspars where those with the lowest axial angles have the finest grain sizes. However, grain sizes alone cannot govern mineral stabilities; but its possible effect on PTX conditions warrants examination. A point to consider in this regard is the permeability of rocks to water vapour; e. g., equation 8 shows that increase of water in the rocks will expand the hornblende field in the higher temperature range.

$$SiO_2 + Ca_2(Mg,Fe)_3Al_2Si_6Al_2O_{22}(OH)_2 \overset{T\rightarrow}{\underset{\leftarrow}{\rightleftharpoons}}$$
$$2\,CaAl_2Si_2O_8 + 3(Mg,Fe)SiO_3 + H_2O: \qquad\qquad 8.$$
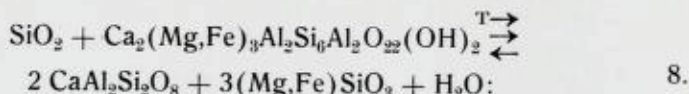
The smaller grain size the more numerous the intergranular contacts, and thus the greater the permeability to water vapour. Thus coarse grained rocks might tend to keep the vapour under a relatively higher pressure and stabilize the hydrous mineral phases. Although no systematic study of the influence of grain size on the observed mineral parageneses of chemically similar rocks has been attempted in this study the point seems worth considering.

Diopside has a distribution rather similar to hornblende and is absent in the rocks with excess SiO_2. It was pointed out in the petrographic descriptions that the anorthite content of the plagioclase tends to increase at approximately the same SiO_2 values as where diopside is no longer stable (Tables 3 and 6), and in the amphibolite facies veined gneisses where hornblende disappears. That diopside is stable only in rocks deficient in SiO_2 may indicate that diopsidic pyroxenes which are stable under granulite facies conditions have like hornblende some of the Si positions occupied by Al. None of the clinopyroxenes in these rocks were chemically analysed; however, it may be noteworthy
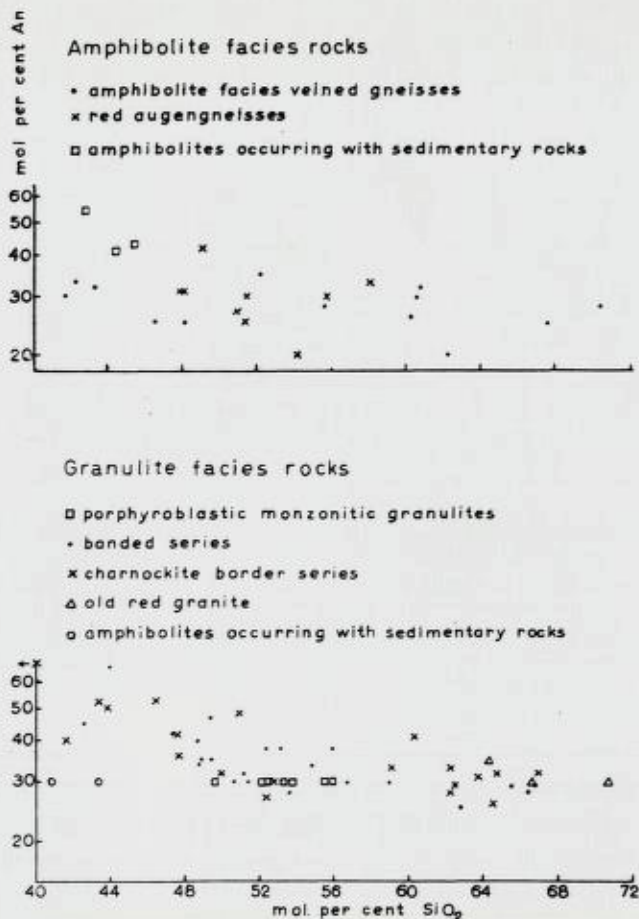
Fig. 28. Relationship between mol per cent. An. in the plagioclases and calculated
mol per cent SiO₂ of the Langøy rocks.

*Forholdet mellom mol % An. i plagioklas og beregnet mol % SiO₂ i bergartene
fra Langøy.*

that Howie (1955) who has given a number of chemical analyses of
ciinopyroxenes from the Madras charnockites (which are similar to the
granulite facies rocks on Langøy), found a rather high Al concentration
in those of the more basic members.

Plagioclase. In Fig. 28 the An. content in the plagioclases is plot-
ted against increasing mol per cent SiO₂ (calculated from the modal

analyses) in the rocks. It can be seen that the majority of the feldspars between 49 to 70 mol per cent SiO_2 are of the An. 28 to 35 range, and it is only in the rather basic rocks that the plagioclases are more anorthitic. This small range in the plagioclase composition throughout so wide a silica range is indicative of the metamorphic (rather than igneous) character of the rocks.

The amphibolite facies.

The rocks in the east of Langøy are characterized by mineral associations corresponding to amphibolite facies (Tables 9 and 18). These eastern gneisses may be divided into two major groups: the amphibolite facies veined gneisses in the north and the red augengneisses (with some amphibolite inclusions) in the south. No chemical analyses are available of the red augengneisses. It is suggested that they are low metamorphic derivatives of the monzonites associated with the anorthosite in the granulite facies area east of Eidsfjorden. As can be deduced from their mineral compositions they must be chemically similar to the porphyroblastic monzonitic granulites and the Eidet-Hovden monzonites.

The amphibolite facies veined gneisses north of the red gneiss area are on the other hand strikingly similar in appearance to the rocks of the charnockite border series and the banded series (Figs. 5—8). In both the amphibolite facies gneisses and the border series rocks banding is conspicious. In the banded series proper the veining is not readily apparent on fresh surfaces but is quite evident along the shores and on strongly weathered surfaces because of the differential resistance to erosion of the acid and the basic bands. The amphibolite facies veined gneisses are on the other hand essentially identical macroscopically as well as in mineral parageneses to the gneisses on Hinnøy directly east of them.

The old concept of regarding the Lofoten-Vesterålen rocks as a completely independent formation is misleading. The work done on Langøy shows the major difference between "the plutonic rocks of Lofoten" and "the basal gneisses" on Hinnøy, to be that of metamorphic facies. There is no sign that the amphibolite facies mineral paragenesis that occurs in the east was caused by any retrograde metamorphism of the granulite facies rocks (except possibly for the red augengneisses). It is in this respect that these rocks differ from the

Table 34.

$$\frac{CaO}{AlO_{1\frac{1}{2}}-KO_{\frac{1}{2}}-NaO_{\frac{1}{2}}} \quad \text{ratios for some amphibolites (cation per cent).}$$

	147/56 b	3/56 b	33/55 j	29/55	13/56
$\dfrac{CaO}{AlO_{1\frac{1}{2}}-KO_{\frac{1}{2}}-NaO_{\frac{1}{2}}}$:	0.84	0.82	0.87	1.20	1.05
Diopside present:	−	−	+	+	+

147/56 b, 3/56 b, amphibolite bands in the amphibolite facies veined gneisses, Table 9.
33/55 j, 29/55, 13/56, amphibolites associated with sedimentary rocks in the amphibolite
facies veined gneisses, Table 15.

retrograde metamorphic gneisses on the west side of Jørgenfjord. But
the rocks in question are obviously of a lower metamorphic facies than
the granulite facies rocks to the west of them. Hypersthene is not present
even in the most silica rich types. Titanite which was nowhere observed
to occur in the granulite facies rocks, is a common accessory constituent
in the two types of gneisses east of the boundary (Tables 9 and 18).

Because the two rocks are chemically similar this is indicative of
the lower metamorphic facies of the former (see Ramberg, 1952, pp.
72—73).

Typically diopside is absent in the amphibolite facies veined gneis-
ses; however, it does occur in the amphibolitic inclusions within the red
augengneisses (though in all cases in various stages of alteration into
green hornblende) and also in the amphibolites associated with sedi-
mentary rocks (Table 15). The lowermost stability border of diopside
in impure limestones indicates the amphibolite to epidote amphibolite
facies transition (eq. 9) (Ramberg, 1952, p. 150):

$$2SiO_2 + 3CaCO_3 + Ca_2Mg_5Si_8O_{22}(OH)_2 \rightleftharpoons 5CaMgSi_2O_6 + 3CO_2 + H_2O, \qquad 9.$$
$$\text{(epidote amphibolite facies)} \qquad \text{(amphibolite facies)}$$
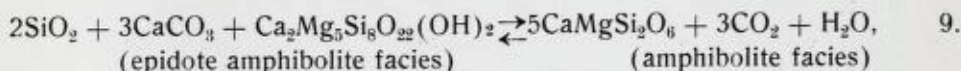
Because the field relationship indicate that the veined gneisses and
the amphibolites that occur with the metasedimentary rocks recrystal-
lized under the same PT conditions it appears that the absence of diop-
side in the former must be related to the chemistry of the rocks. The
various $CaO/AlO_{1\frac{1}{2}} - KO_{\frac{1}{2}} - NaO_{\frac{1}{2}}$ (cation per cent) ratios for the
chemically analysed amphibolitic rocks concerned are listed in Table
34. Although diopside would be present in the catanorm of all these

rocks (otherwise the ratio should be less than 0.5) diopside actually is present only in the rocks which are relatively low in free Al which could combine with Ca.

Epidote is present in two of the red augengneisses (Table 18) as independent mineral grains. The plagioclase compositions in these two rocks are An. 30 and An. 20 and if the equilibrium association epidote (zoisite) → plagioclase An. 30 (±) is taken as the lowest boundary for amphibolite facies (Ramberg, 1952, p. 150) this may indicate epidote amphibolite facies conditions.

However, as can be seen in Tables 9 and 18, a plagioclase more basic than An. 30 is commonly present. This, together with the fact that epidote was only found to occur in two of the rocks probably indicates that the rocks are of the amphibolite facies (probably of its lower part).

Heier (1956 a) found that "orthoclases" are the stable potash feldspar in the higher part of amphibolite facies and upwards (the term "orthoclase" was used for feldspars that appear monoclinic optically from the standpoint that microcline twinning could not be seen). In most of the amphibolite facies gneisses on Langøy microcline twinning is developed; this would agree with the suggestion that the rocks belong to a lower part of the amphibolite facies. However, microcline twinning was not observed in the feldspars in some of these rocks, especially of rocks close to the metamorphic boundary. However, the $2V_x$ in these is larger than that observed within the feldspars of the neighbouring rocks of the charnockite border series. The variations of the alkali feldspars as related to the metamorphic grade is discussed in a later section of this chapter and it is shown that the temperature drop acoss the boundary (as determined by the feldspar geological thermometer, Barth, 1956), though marked, is not of any significant order of magnitude.

The variations of hornblende and biotite (as per cent of the sum of femic minerals recalculated to 100) compared to the SiO_2 content in the rocks are plotted for the amphibolite facies gneisses in Fig. 29. Hornblende shows very much the same pattern as in the granulite facies rocks but is also stable in the higher SiO_2 range.

Although biotite shows a similar pattern to that in the charnockite border series its relative concentrations are higher — actually reaching 100 per cent in the most silica rich rocks.
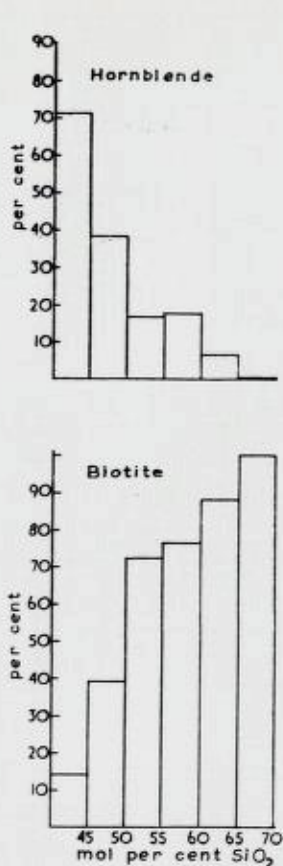
Fig. 29.
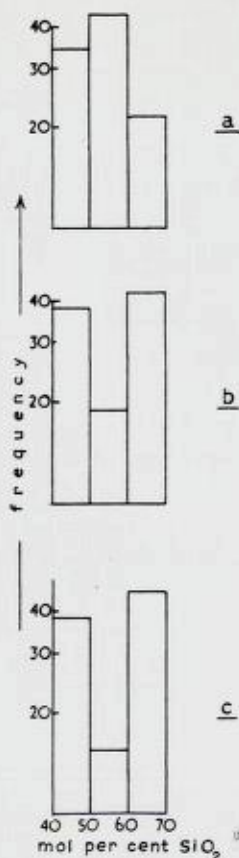
Fig. 30.

Fig. 29. Arithmetic means of hornblende and biotite in amphibolite facies rocks (calculated as for Fig. 26).

Aritmetisk middel av hornblende og biotit i amfibolit facies bergarter (beregnet som for Fig. 26).

Fig. 30. Frequency distribution of rocks plotted against calculated mol per cent SiO_2 in 23 rocks from the banded series (a), 21 rocks from the charnockite border series (b), and 13 rocks from the amphibolite facies veined gneisses (c).

Fordelingen av bergarter over forskjellige intervaller av beregnet mol % SiO_2 i 23 bergarter fra "the banded series" (a), 21 bergarter fra "the charnockite border series" (b), og 13 bergarter fra "the amphibolite facies veined gneisses" (c).

The frequency distribution of the three rock series; the banded series, the charnockite border series and the amphibolite facies veined gneisses are shown on Fig. 30. The intervals are taken at 10 per cent modal SiO_2, (it should be repeated that the number of samples point counted in each group is much too small to be treated strictly statistically. A 10 per cent interval is also rather too much — but although ideally a much smaller interval should be used (see also Ahrens, 1954a, pp. 123—124), it is not possible here because of the small number of observations made.) Fig. 30 reveals, however, that the amphibolite facies veined gneisses (c, Fig. 30) may be nearly completely divided into two separate groups with very few rocks in the intermediate SiO_2 range. The same division was observed within the charnockite border series (b, Fig. 30), although in it a greater number of rocks are in the intermediate range. In the case of the banded series (a, Fig. 30) a distribution peak occurs in this range. On the basis of these data one might draw the conclusion that increasing metamorphism has led to homogenization of the gneisses. (It should also be remembered that the coarse grained homogeneous western monzonites which were undoubtedly metamorphically recrystallized, occur in this intermediate SiO_2 interval). On the other hand some care should be made not to draw to wide conclusions on the basis of Fig. 30 because the sampling error may be significant. For the amphibolite facies veined gneisses, where the dark amphibolitic and light "granitic" bands are clearly distinguishable, generally from each occurrence a thin section was made of each type, not across the contacts. For the banded series rocks, where the quartz and feldspars are characterized by dark colours and the basic and acid bands are not so easily distinguished, thin sections may cut across the bands. In the course of the microscope work, however, it was never noted that one end of any thin section was significantly different from the other. Ramberg (1951, pp. 31—33) found on West Greenland that the main amphibolite facies gneiss (Egedesminde gneiss) is less homogeneous and more "acid" than the synmetamorphic main granulite facies gneiss (Isortoq gneiss). On Langøy the bulk chemistry of the banded gneisses in amphibolite and granulite facies is probably similar.

The retrograde metamorphic gneisses.

These rocks are heterogeneous with respect to their chemistry and mineral parageneses. Rocks of the group have mineral assemblages stable in all facies between and including those commonly designated granulite and epidote amphibolite. It is obvious that the high grade rocks are not retrogressive. However, they are so closely associated in the field with the lower grade gneisses that any explanation of the nature of the retrograde metamorphism must also explain this close association. The group of rocks is by no means isofacial.

It is the author's intent in this chapter to show that the bulk of these gneisses are low grade metamorphic equivalents of the surrounding rocks and that their low temperature mineral assemblages were formed as the results of retrograde alteration of granulite facies rocks.

Chemical similarities between the western monzonites and the porphyroblastic types of the retrograde metamorphosed gneisses have been stressed previously. These chemical similarities can be seen by comparing analyses nos. 55/56 and 57/56 of the western monzonites (Table 2) with nos. 310/55 and 318/55 of the retrograde metamorphosed gneisses (Table 13). Further north in this zone banded gneisses occur (nos. 44/56, 321/55 a and 321/55 b, Table 12). No chemical analyses have been made of the retrograde metamorphosed banded rocks, but there can be no doubt that before diaphthoresis they were petrographically similar to the banded series rocks.

Modal analyses of the series of feldspathic rocks (nos. 301/55, 303/55, 304/55, 310/55) illustrate mineralogical changes in rocks brought about by retrograde metamorphism of chemically similar rock types. The mineral compositions are, for the sake of convenience, repeated here in Table 35. The metamorphic grade decreases from left to right in the table. The variations of the different minerals are interesting. The $2V_x$ of the potash feldspar shows a gradual increase from left to right, and the microcline twinning was seen in no. 310/55. Thus the symmetry of the potash feldspars apparently adjusted itself to changing external conditions and became most highly ordered in the low temperature assemblages. It will be shown later that the chemistry of the feldspars does not change accordingly. This indicates that they originally crystallized at a high temperature, and although their crystal symmetry could adjust itself to the changing external conditions during

Table 35.

The effect of retograde metamorphism on the porphyroblastic monzonitic granulites (localities in Table 12).

Minerals	301/55	303/55	304/55	310/55
Quartz	–	–	tr.	tr.
Potash feldspar...........	6	13	29	19
2V$_x$......................	(40—45)	(45—53)	(60—65)	(60—62)
Microcline twinning	–	–	–	+
Plagioclase...............	72	70	60	59
An. in plag...............	(40)	(34)	(33)	(25)
Signs of saussuritization ...	–	+	+	+
Hypersthene	5	2	–	–
Diopside	7	4	tr.	–
Hornblende	–	3	7	10
Biotite	4	5	4	10
Titanite	–	–	tr.	2
Apatite	1	tr.	tr.	tr.
Ore	5	3	tr.	tr.

retrograde metamorphism the readjustment of chemical equilibrium could not be so easily (or quickly) achieved (see Heier, 1957, p. 478).

The plagioclase composition (as determined by measuring the maximum extinction angle on the universal stage) also shows such an adjustment to low temperature conditions. A decrease in An. content and an increasing amount of saussuritization occur. If the equilibrium association epidote (zoisite) $\rightleftarrows$ plagioclase An. 30 ± determines the lowest boundary of amphibolite facies (Ramberg, 1952), no. 310/55 should represent epidote amphibolite facies conditions.

The disappearance of hypersthene and diopsidic pyroxene together with the introduction of hornblende and increasing amounts of biotite from left to right in Table 35 likewise reflects the differences in metamorphic grade. All of the hornblende in no. 303/55 was found as reaction rims around hypersthene whereas the diopsidic pyroxene in the rock was found to be unaltered. Rock no. 304/55 which illustrates the next step of the retrograde metamorphic action has no hypersthene remaining and the diopsidic pyroxene can be seen to represent a stage of alteration into green hornblende. Textural criteria are difficult to interpret, but in these rocks it seems obvious that a low grade assemblage has gradually taken the place of the high grade one (see also Fyfe et al., 1958, p. 14).

However, relationships between titanite and ore in the rocks in Table 35 appear to be good indicators of the changing metamorphic

grade. Titanite is absent in the granulite facies rocks. As stated on p. 55 titanite where present in the retrograde metamorphosed rocks surrounds iron ore. This is so even when iron ore is present as traces in the rock. As elaborated upon further on p. 139 the concentrations of iron ore are characteristically lower in the amphibolite facies gneisses than in their counterparts in granulite facies. The ore is partly transferred to titanite, but as this mineral is generally present only as traces (see Table 9) and does not contain appreciable amounts of iron it is obvious that most of the titanium and iron must be contained in the hornblende and biotite; these minerals are capable of higher concentrations of Ti and trivalant iron than are hypersthene and diopsidic pyroxenes, typical of the granulite facies gneisses.

The minerals present in the retrograde metamorphic gneisses are not characteristic of the maximum metamorphic temperatures and pressures which have affected them. When considering the nature of this retrograde metamorphism one should keep in mind that complete equilibrium was not reached. This is indicated by the common occurrence of "unstable relics" and by the fact that the rocks were not affected equally over the whole area. A number of different factors may be responsible for a retrograde metamorphism of this kind and for that of the Langøy area there is no obvious explanation. The author has in a previous study of metamorphosed gneisses in the Ørsdalen district in Southern Norway found that zones of retrograde metamorphism were connected with the tectonics of the area (Heier, 1956 a, p. 205).

A purely structural explanation is difficult to conceive for the retrograde metamorphic rocks on Langøy; the rocks in question cannot be seen to be more highly tectonized than rocks of the banded series in general.

Apart from the retrograde metamorphic rocks of the banded series and the western monzonites the rock types found in the retrograde metamorphosed area are: metasedimentary rocks (sillimanite gneisses, graphite schists, etc.), red granites and the Bø gabbro.

The Bø gabbro is probably a late intrusion in the gneisses and it is the only rock which could have possibly caused the retrograde metamorphism. The gabbro has a stable high temperature mineral assemblage. It occurs in the far south end of the retrograde metamorphosed zone. If the gabbro has caused the metamorphism it must be explained why the effect has been largely in the northnortheast while to the west and northwest, the retrograde effect is only observed in the

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monzonites close to the gabbro contact. The reason may be the different nature of the rocks surrounding the gabbro. The massive western monzonites represent the country rock to the west and north while banded rocks are more typical towards northnortheast. The former has no preferred directions where vapour and gases from the crystallizing gabbro can easily escape, and thus the metamorphic effect is here restricted to the immediate contacts only. The gases may find easier access through the banded rocks parallel to the directions of the bands. If the metamorphism was caused by the escaping gases this explains the disequilibria observed in these rocks.

Scapolite was formed when plagioclase reacted with escaping gases. (In fact, scapolite is more common than may be suspected from Table 12, where it is listed in only two of the rocks.) Scapolitization is a well known metasomatic effect associated with gabbroic intrusions. In the present case it is not known whether the scapolite is predominantly of the Cl or CO_2 type. However, it should be recalled that calcite was observed to occur as traces in the gabbro and in some of the neighbouring gneisses.

Although it is not proved definitely, it appears that processes associated with the intrusion and crystallization of the Bø gabbro caused the retrograde metamorphism; at a stage when the rocks in the west of Langøy were no longer under granulite facies PT conditions the Bø gabbro was intruded and caused secondary metamorphism of the surrounding gneisses.

The Eidet-Hovden intrusion related to the period of regional metamorphism.

In the descriptions of the petrography of this intrusion it was stated that its relation to the period of regional metamorphism was not directly obvious.

Vogt (1909) classified all the crystalline rocks on Langøy (or rather all the rocks within the granulite facies part) as one igneous series derived through fractional crystallization of a single parent magma. He believed the mineral parageneses and rock textures to be primary magmatic and did not recognize any subsequent metamorphism. As noted above, it has been shown that a metamorphic recrystallization must have taken place in the rocks dealt with in this chapter. The problem here is whether the Eidet-Hovden intrusion was pre or post the regional metamorphism, which in these parts took place under the PT conditions of the granulite facies.

The dykes.

The youngest rocks within the intrusion are the dykes. Vogt (1909) found no corresponding massifs of deep seated rocks in the neighbourhood, but assumed that they were derived from the large area of monzonitic rocks to the south, i. e., the rocks termed the banded series and porphyroblastic monzonitic granulites in this paper. Such an association would require that the Eidet-Hovden intrusion is of pre- or syn-metamorphic age. All that favours such an association between the two rocks are the chemical and mineralogical similarities between some of the dykes and certain members of the metamorphic rocks. The chemistry is discussed further in a subsequent chapter, but it should be mentioned here that the banded series contains rocks of widely diverse origins and their chemistries accordingly vary greatly. Therefore that some of them approach the dyke rocks in chemical composition is not surprising and little weight should be attributed to such similarities.

Mineralogical similarities are controlled, among other things, by chemical similarities. The granulite facies mineral parageneses of this area are extremely similar to the igneous gabbro facies parageneses of rocks with excess SiO_2. (Most of the dykes do not contain free quartz but are in the intermediate silica range where this relationship still holds.)

Opposing any association between the dykes and the country rocks are:

1. In spite of the large number of dykes in the southern part of the gabbro none has been seen to cross the contact into or to occur within the surrounding banded series.

2. The strike of the banded series (as defined by the banding) is not cut by the gabbro; rather it parallels the boundary of the intrusion. This probably indicates that the gabbro was intruded into the banded series under PT conditions under which the gneisses could be deformed plastically. In any case, it is rather against a younger age for the banded series rocks.

3. The dykes strike roughly parallel to the gabbro-banded gneiss contacts.

4. At least one dyke contains rounded grains of quartz with straight extinction, no. 60/56, Table 30. As undulatory extinction has been observed universally in quartz affected by the stress a lack of this

is considered to be of great importance so far as suggesting a post metamorphic origin for the dyke.
5. The apparent genetic relationship between the dykes and the Eidet-Hovden monzonites.

The monzonites,

It was stated on p. 100 that a genetic relationship may exist between the dykes and the monzonites within the gabbro (Figs. 22 a and b).

It was concluded that the monzonites are characterized by a nonequilibrium mineral paragenesis. If the mesoperthitic alkali feldspar phenocrysts within them were an originally Na and Ca rich homogeneous alkali feldspar phase (anorthoclases) three types of feldspars coexisted in these rocks: mesoperthites (anorthoclases), normal orthoclase perthites, and zoned plagioclases (see p. 98). Evidence that the mesoperthite originated as a homogeneous Na, K and Ca solid solution feldspar is afforded by chemical data (see p. 168) and by the fact that not all the feldspars are unmixed to the same extent (pp. 94—95). However, it would be extremely difficult to disprove the possibility of simultaneous crystallization at different stages in evolution of the system.

The lack of equilibrium is also indicated by the zoning of the clinopyroxenes.

On the other hand, the rocks which recrystallized under the PT conditions of granulite facies are characterized by a paragenesis with a high degree of equilibrium.

It is difficult to conceive how the non equilibria of the monzonites could have lasted through the processes of granulite facies metamorphism. Migration of Na and K is known to have taken place under the conditions of regional metamorphism. K metasomatism is assumed to be important under PT conditions of the lower part of epidote amphibolite facies, and Na metasomatism (spillitization) active even at lower grade stages. Considering these it is especially difficult to see how the inequilibria observed in the alkali dominated feldspars could have lasted through the granulite facies conditions.

These nonequilibria, as contrasted with the high degree of equilibrium in the true granulite facies rocks are, therefore, also evidence supporting a post metamorphic age for the monzonite intrusion.

The gabbro.

The gabbro which occcupies the main part of the intrusion is older than the monzonite and the dykes.

There is no proof of a direct cogenetic association between the gabbro and the monzonites.

The gabbro is much finer grained along the monzonite contacts than elsewhere and this may be the result of some contact metamorphic effect of the intrusive monzonite on the already solidified gabbro (p. 92). The fact that there probably is a flat lying contact between the two rocks (p. 92) may, on the other hand, indicate a "rest magmatic" origin for the monzonites.

Because of the relationships between the two rocks are rather controversial, a post metamorphic age of the gabbro cannot be assumed on the basis of this association alone.

The secondary alterations observed in the gabbro close to the contacts with the banded series rocks are of interest with this regard. The banded series rocks cannot be seen to have been metamorphosed by the gabbro. But, as the granulite facies minerals are themselves stable at high temperatures this is not remarkable. Gabbro intrusions are seldom surrounded by the metasomatized aureole which is typical of more acid intrusions and any heat effect alone would probably not have affected the high temperature (granulite) mineral parageneses significantly.

The formation of the hydrous minerals in zones within the gabbro close to the contact is a secondary or retrograde effect probably produced by the alteration of preexisting olivines and pyroxenes. As such, it might be of a completely secondary nature and not in any way related to the crystallization history of the gabbro. As mentioned previously, although scapolite is in equilibrium with plagioclase in the altered contact zones, calcite (or some other carbonate mineral) is in stable equilibrium with the plagioclase in the unaltered gabbro. Therefore the reaction of the calcite with plagioclase to form scapolite cannot be the result of any regional metamorphism because such should have affected the rest of the gabbro and some of the country rocks as well. The alterations were local and probably of hydrothermal nature; it is tempting to assume that the hydrothermal alterations were caused by the escaping volatiles after the solidification of the gabbro. In any case the regional PT conditions must then have been such that the hydrous minerals could be stable and subsequently they have never been such

Table 36.

Chemical analyses of three gabbros (1—3)
and four granulite facies rocks (4—7) with similar SiO_2 contents.

	1 (an. 16)	2 (an. 8)	3 (32/56 d)	4 (58/56)	5 (77/56)	6 (81/56)	7 (151/56 b)
SiO_2	46.65	48.72	46.78	48.87	46.2	42.5	49.20
TiO_2	0.13	0.23	0.69	2.14	1.50	1.67	1.16
Al_2O_3	23.07	20.56	16.18	16.92	13.6	11.2	15.02
Fe_2O_3	1.31	0.30	0.69	5.23	4.98	7.15	1.19
FeO	4.73	4.68	10.98	8.86	8.65	11.82	9.08
MnO	0.07	0.08	0.12	0.14	0.22	0.18	0.22
MgO	9.01	9.81	12.65	4.55	6.63	7.46	10.26
CaO	10.69	11.83	7.36	6.68	11.75	12.73	8.52
Na_2O	2.69	2.60	3.11	4.30	4.07	3.11	2.37
K_2O	0.38	0.40	0.67	0.87	0.87	0.67	1.58
H_2O^-	0.16	0.13	0.08	0.10	n.d.	n.d.	0.13
H_2O^+	0.97	0.37	0.39	0.23	n.d.	n.d.	0.42
CO_2	0.55	0.30	n.d.	n.d.	n.d.	n.d.	n.d.
P_2O_5	–	–	0.35	0.94	n.d.	n.d.	0.70

C. I. P. W. (molecular) norm

K-feldspar (or).....	2.0	2.5	4.0	5.5	5.5	4.0	9.5
Na-feldspar (ab) ...	23.0	21.7	21.2	39.5	16.5	8.0	21.0
Ca-feldspar (an)....	49.0	42.2	27.8	24.5	16.5	15.3	25.8
Nepheline (ne).....	0.6	0.8	3.4	–	12.3	12.6	–
Calcite (cc)	1.4	0.8	–	–	–	–	–
Olivine (ol)	23.1	20.9	34.9	4.5	7.3	10.0	17.1
Hypersthene (hy)...	–	–	–	12.8	–	–	13.4
Diopside (di)	–	10.8	6.0	2.8	34.4	40.0	8.8
Magnetite (mt).....	1.4	0.3	0.8	5.6	5.3	7.7	1.2
Ilmenite (il)	0.2	0.4	1.0	3.0	2.2	2.4	1.6
Apatite (ap)	–	–	0.8	1.8	n.d.	n.d.	1.6

that the hydrous minerals became unstable. Therefore the gabbro intrusion must be younger than the granulite facies metamorphism.

Although no unequivocal proof has been found to relate the gabbro to the regional metamorphism in time, it appears that the mineralogy and petrographic characteristics of the gabbro are probably of primary magmatic origin. Chemical analyses and norms of three of the gabbros (nos. an. 16, an. 8, and 32/56 d) and four of the granulite facies rocks (nos. 58/56, 77/56, 81/56, 151/56 b) are compared in Table 36. Olivine is present in the mode of all the three gabbros but is absent in the metamorphic rocks (Tables 3, 6 and 15). The reaction rims around the olivines in the gabbro are generally thin and olivine does not appear to be stabilized in the gabbro as armoured relics.

It is difficult to conceive how the primary magmatic minerals could have persisted if the gabbro has subsequently been brought under the conditions of regional granulite facies metamorphism. Therefore a late or post metamorphic age appears most likely.

The young red granites.

The young red granites were also probably intruded after, or during a late stage of the regional metamorphism under granulite facies conditions. This is indicated by the presence of biotite in a paragenesis where hypersthene should be expected if they were metamorphosed under granulite facies conditions.

Characteristic variations in the stable minerals of the diverse rocks.

On Langøy certain minerals, e. g., quartz, potash feldspar, plagioclase, hornblende, biotite, and iron ore, are stable constituents of rocks of each metamorphic grade. This does not imply, however, that each mineral is of identical chemical and/or structural composition in all the rocks.

Thus it is possible to determine how certain members of some mineral series demand certain restricted PTX conditions for their formation.

The alkali feldspars have been studied in detail for this specific purpose. A few of the remaining minerals have been found to show some interesting trends (on the basis of optical analyses).

The data for the various minerals are given in the pertinent petrographic sections.

Hornblende.

The value of $2V_x$ of hornblende in the diverse rock types are plotted in Fig. 31. There is a general tendency for the axial angle to decrease with decreasing metamorphic grade of the enclosing rock. Within each rock series there is a tendency for the $2V_x$ to decrease with increasing SiO_2 of the host rock (Fig. 32). For similar SiO_2 contents the hornblendes of the low grade metamorphic rocks tend to have lower 2V angles than their high grade metamorphic counterparts. An especially interesting example is provided by rocks nos. 95/56 and 98/56, Table 18. The hornblendes of these two similar rocks are joined
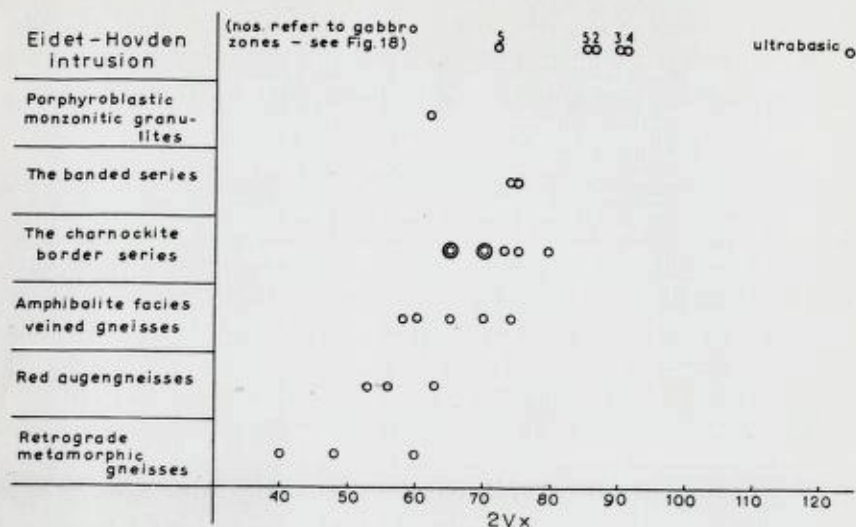
Fig. 31. Variation of $2V_x$ in hornblende in the Langøy rocks.

Variasjonen av $2V_z$ i hornblende i bergartene fra Langøy.

with a line in Fig. 32; the one which occurs within the charnockite border series has a much larger angle than the one from east of the metamorphic boundary. The colour of the two hornblendes is similar (green) and, in general, the colour of the hornblendes does not differ appreciably in the differently metamorphosed rocks. However, there is a tendency for hornblendes in the granulite facies rocks to be brown or brownish green whereas those in the amphibolite facies are characteristically green.

The chemistry of the hornblendes is too complex to permit determination of the relation of the variations of the axial angle to variations in chemical compositions. Actually the hornblendes of the granulite and those of the amphibolite facies rocks may belong to two different series of amphiboles. Amphiboles of the granulite facies rocks are typically hastingsitic and thus characterized by a relatively large amount of Al substitution for Si in the silica tetrahedra. This does not mean that hastingsites are enriched in total Al as compared to the common hornblende of the lower metamorphosed rocks; it means only that there is an increased tendency for Al to occupy four fold co-ordination positions in the lattices with increasing metamorphism (see Foslie, 1945, p. 95). Trøger (1952) indicates that $2V_x$ in the hastingsite series varies
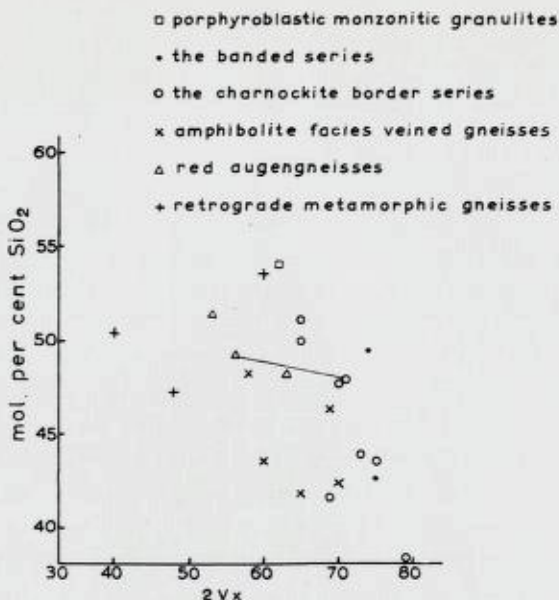
Fig. 32. Relationship between $2V_x$ in hornblende and calculated mol per cent SiO_2 of the enclosing rocks.

Forholdet mellom $2V_z$ i hornblende og beregnet mol % SiO_2 av bergartene.

in the range 68° to 25° (Fig. 184, p. 75) and common hornblendes in the range 88° to 64° (Fig. 187, p. 77). This shows clearly that the value of the axial angle alone cannot be used in determining identities of hornblendes because use of Trøger's Tables of axial angles would lead to the probably erroneous conclusion that the amphiboles in the amphibolite facies rocks tend to be more hastingsitic than those in the granulite facies, and furthermore that the hastingsitic character increases with increasing SiO_2 in the host rocks. Trøger (1952) also lists a series of titaniferous amphiboles (Fig. 185 p. 76) with $2V_x$ ranging between 83° and 79°. It seems most probable to the author that the amphiboles in both granulite and amphibolite facies are hastingsitic with a substantial amount of Al in four fold co-ordination. Harry (1950) shows a substantial substitution of Al for Si in hornblendes from amphibolite facies to exist. Charge balance is achieved in granulite facies amphiboles by Ti^{4+} substituting for Fe^{2+} and Mg^{2+} in the octahedral positions, whereas it is achieved in the amphibolite facies rocks chiefly through an increase of Fe^{3+}. This is suggested because

an introduction of TiO_2 in the system would expand the stability field of hornblende so far as the uppermost part of the hornblende field is concerned (Ramberg, 1952, p. 69). Thus the difference between amphibolite and granulite facies amphiboles may depend chiefly on the $FeO + Fe_2O_3/TiO_2$ ratio.

Biotite.

The small variations in pleochroism of biotite from one rock series to the other probably does not indicate significant chemical variations. As was stated the biotites in rocks with excess silica in both the border series and the amphibolite facies gneisses must be richer in Al than indicated by the general biotite formulae $K (Mg,Fe)_3AlSi_3O_{10}(OH)_2$. This excess amount of Al could not be taken up by hornblende because hornblende is not present in these rocks with excess SiO_2. It appears that a certain amount of solid solution between biotite and muscovite exists in these rocks. Ramberg (1952, p. 45) has written "Muscovite forms at lower temperatures than biotite and does not persist stably up to a higher degree of metamorphism than the border line between amphibolite and granulite facies". A comparative study of the Al content in the biotites of chemical similar rocks from the different areas on Langøy might be of great interest. According to Larsen and Berman (1934, Table 16, pp. 237—238) muscovites have lower refractive indices than biotites. Table 37 shows that the biotites in the amphibolite facies veined gneisses (except no. 37/55) tend to have lower β refractive indices than the biotites in the granulite facies rocks.

Table 37.

Refractive indices of biotites from rocks
of different metamorphic grade.

	nβ
293/55, porphyroblastic monzonitic granulite	1.645 ± 0.003
224/56, slightly retrograde metamorphosed monzonite	1.638 ± 0.003
25/55, charnockite border series rock..............	1.634 ± 0.003
150/56, charnockite border series rock..............	1.645 ± 0.003
37/55, amphibolite facies veined gneiss............	1.645 ± 0.003
147/56, amphibolite facies veined gneiss............	1.630 ± 0.003
an. 1, amphibolite facies veined gneiss............	1.630 ± 0.003
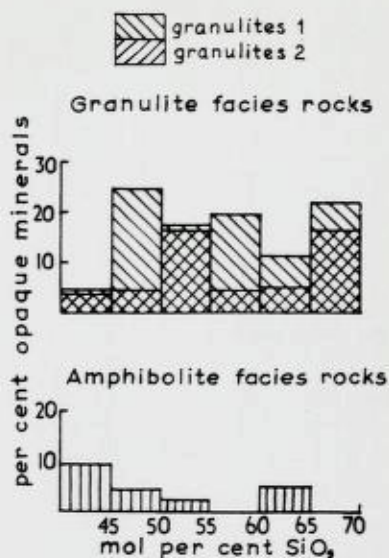
Fig. 33. Arithmetic means of opaque minerals in the Langøy rocks (granulites 1 and 2, see p. 115, calculated as for Fig. 26).

Aritmetisk middel av opake mineraler i bergartene fra Langøy (granulit 1 og 2, se s. 115, beregnet som for Fig. 26).

Opaque minerals (iron ore).

Biotites may contain fairly high amounts of Fe_2O_3, and high temperature biotites would seem to require a rather high percentage of Ti in order to remain stable (Ramberg, 1948 a). When biotite becomes unstable in the granulite facies and reacts with quartz to form potash feldspar and hypersthene TiO_2 is liberated and contributes to the formation of titaniferous iron ore. Fig. 33 shows the frequency distribution of iron ore in the granulite and amphibolite facies rocks at different SiO_2 intervals. The data of this figure should be compared with those for biotite in the same rocks (Figs. 26 and 29). It can be seen that the concentration of iron ore in the border series rocks is lower than in the higher metamorphosed banded series. In the most silica deficient rocks in the amphibolite facies the concentration of iron ore is higher than in the corresponding higher grade metamorphic rocks but its concentration decreases with increasing SiO_2 content and iron ore tends to be absent in the silica rich types. When Fig. 33 is compared with the corresponding frequency diagrams of biotite an interesting relationship is seen. The silica deficient rocks in amphibolite facies which are rich in iron ore contain roughly the same amounts of biotite as the
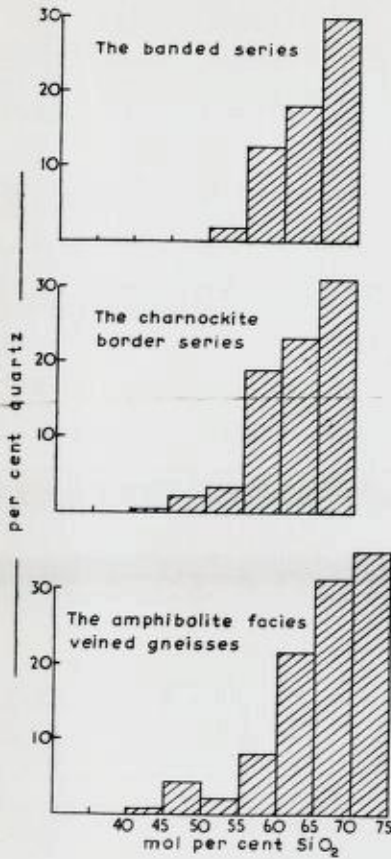
Fig. 34. Arithmetic means of quartz in the Langøy rocks.

Aritmetisk middel av kvarts i bergartene fra Langøy.

corresponding granulite facies rocks. If the concentration of TiO_2 is higher in the biotite of the latter (increasing with increasing metamorphic grade) this could explain why the concentration of iron ore in this silica interval is lower in the granulite facies rocks. With increasing SiO_2 concentrations in the rocks the biotite content of the amphibolite facies rocks increases rapidly and significantly exceeds that of the higher grade metamorphosed equivalents. The opposite is true for the iron ore. The same relationship occurs between the banded series rocks and the border series rocks which are of much more similar metamorphic grade.

Thus isochemical metamorphic reactions lead to the liberation of iron ore in granulite facies rocks with excess SiO_2 (see also Ramberg, 1948 a, p. 564).

Quartz.

The major variation in the quartz is in its colour. The quartz in the granulite facies rocks is characterized by a bluish gray colour.

Fig. 34 shows the frequency distribution of quartz plotted against total SiO_2 (as calculated fram the mode) in the rocks for the three mutually comparable rock series on the island. The feature which should be noted is the increase of free quartz in the low silica range (40—50 % SiO_2 in the rocks) with decreasing metamorphic grade. This is a consequence of the relative preponderance of biotite and hornblende in the amphibolite facies rocks (see eq. 1 and 8).

The feldspars.

Structural and chemical work in the feldspar group of minerals during the last about 10 years indicates that the feldspars may be used to explain diverse petrological aspects of metamorphic rocks, especially those of the amphibolite and granulite facies.

The calculated compositions of some alkali feldspars from rocks within the different groups on Langøy are listed in Table 38. The calculations are based on analyses of the cations outside the Al-Si tetrahedra (given in Table III). The distribution of minor elements in the feldspars is discussed together with the geochemistry of the rocks and

Table 38.

Calculated values of theoretical feldspar components (weight per cent) in the potash feldspars from Langøy (for analyses see Table III).

No.	Or	Ab	An	Ba	Sr	Pb	Rb	Tl	Sum
				Group I.					
50	34.9	48.1	18.0	1.1	0.24	0.002	0.02	–	102.362
145/55 . . .	34.1	47.0	17.6	1.1	0.23	0.002	0.02	–	100
51 (5)	34.4	51.3	16.7	0.82	0.20	0.001	0.02	–	103.441
an. 9	33.3	49.6	16.1	0.79	0.19	0.001	0.02	–	100
52 (5)	33.4	54.1	13.2	0.82	0.22	0.001	0.02	–	101.761
an. 9	32.8	53.2	13.0	0.81	0.22	0.001	0.02	–	100
53	47.4	47.4	9.7	0.96	0.19	0.003	0.12	tr.	105.773
179/55 . . .	44.8	44.8	9.2	0.91	0.18	0.003	0.11	tr.	100
				Group II.					
54	76.0	23.6	0.63	0.04	0.03	0.002	0.20	0.0001	100.5021
an. 7	75.6	23.5	0.63	0.04	0.03	0.002	0.20	0.0001	100

Table 38. (cont.).

No.	Or	Ab	An	Ba	Sr	Pb	Rb	Tl	Sum
				Group III.					
55 (9)	65.2	21.6	10.4	3.0	0.63	0.002	0.12	–	100.952
271/55 ...	64.6	21.4	10.3	3.0	0.62	0.002	0.12	–	100
56	75.6	20.6	3.8	2.7	0.45	0.002	0.13	0.0001	103.2821
293/55 ...	73.2	19.9	3.7	2.6	0.44	0.002	0.13	0.0001	100
57	75.8	18.5	3.5	2.7	0.45	0.002	0.25	0.0001	101.2021
255/55 ...	74.9	18.3	3.5	2.7	0.44	0.002	0.25	0.0001	100
58	73.5	19.4	4.2	2.7	0.48	0.002	0.14	0.0001	100.4221
251/55 ...	73.2	19.3	4.2	2.7	0.48	0.002	0.14	0.0001	100
				Group IV.					
59 (11) ...	72.1	25.5	3.8	0.96	0.30	0.003	0.19	0.0001	102.8531
318/55 ...	70.1	24.8	3.7	0.93	0.29	0.003	0.18	0.0001	100
60	73.9	22.0	2.4	0.68	0.24	0.002	0.28	0.0002	99.5020
238/55 ...	74.3	22.1	2.4	0.68	0.24	0.002	0.28	0.0002	100
61	78.5	17.5	2.1	2.7	0.52	0.002	0.20	0.0002	101.5220
224/55 ...	77.3	17.2	2.1	2.7	0.51	0.002	0.20	0.0002	100
				Group V.					
62 (16) ...	81.6	14.3	3.5	2.7	0.59	0.002	0.18	0.0001	102.8721
an. 12	79.3	13.9	3.4	2.6	0.57	0.002	0.17	0.0001	100
63	82.1	15.6	2.8	1.4	0.37	0.004	0.35	0.0002	102.6242
348/55 ...	80.0	15.2	2.7	1.4	0.36	0.004	0.34	0.0002	100
				Group VI.					
64	71.1	20.5	3.5	1.1	0.28	0.016	0.22	n.d.	96.7160
65/56	73.5	21.2	3.6	1.1	0.29	0.017	0.23	n.d.	100
				Group VII.					
65	81.2	17.2	1.4	1.6	0.37	0.005	0.27	0.0001	102.0451
25/55	79.6	16.8	1.4	1.6	0.36	0.005	0.26	0.0001	100
66	78.3	19.4	2.1	1.9	0.52	0.005	0.25	0.0001	102.4750
150/56 ...	76.4	18.9	2.0	1.9	0.51	0.005	0.24	0.0001	100
				Group VIII.					
67	86.7	11.1	1.0	2.25	0.41	0.004	0.30	0.0001	101.7141
147/56 ...	85.2	10.9	0.98	2.2	0.40	0.004	0.29	0.0001	100
68	83.0	14.9	1.0	1.8	0.45	0.004	0.32	0.0001	101.4741
3/56	81.8	14.7	0.99	1.8	0.44	0.004	0.32	0.0001	100
69	81.2	13.6	1.4	0.99	0.19	0.002	0.24	0.0001	97.6221
35/55	83.2	13.9	1.4	1.0	0.19	0.002	0.25	0.0001	100
70	84.4	12.8	1.7	2.2	0.32	0.003	0.20	0.0001	101.6231
37/55	83.0	12.6	1.7	2.2	0.31	0.003	0.20	0.0001	100
71 (30) ...	89.1	9.1	1.4	1.2	0.20	0.004	0.14	0.0001	101.1441
an. 1	88.1	9.0	1.4	1.2	0.20	0.004	0.14	0.0001	100

Table 38 (cont.).

No.	Or	Ab	An	Ba	Sr	Pb	Rb	Tl	Sum
					Group IX.				
72	90.1	8.4	1.0	3.3	0.22	0.002	0.12	0.0001	103.1421
109/56 ...	87.4	8.1	0.97	3.2	0.21	0.002	0.12	0.0001	100
73	83.9	15.6	1.4	0.82	0.20	0.002	0.07	0.0001	101.9921
99/56	82.3	15.3	1.4	0.80	0.20	0.002	0.07	0.0001	100
74	90.1	9.0	1.0	0.52	0.16	0.003	0.29	0.0001	101.0731
E-vatn ...	89.1	8.9	0.99	0.51	0.16	0.003	0.29	0.0001	100
					Group X.				
75 (37) ...	53.5	42.8	1.4	0.25	0.05	tr.	0.09	0.0002	98.0920
an. 5	54.5	43.6	1.4	0.25	0.05	tr.	0.09	0.0002	100

Group I: 50—53, monzonites within the Eidet-Hovden intrusion
    II: 54   , pegmatite on border between gabbro and monzonite (Eidet-Hovden intrusion).
   III: 55—58, porphyroblastic monzonitic granulites
   IV: 59—61, retrograde metamorphic gneisses
    V: 62—63, charnockites, banded series
   VI: 64   , old red granite
  VII: 65—66, charnockite border series
 VIII: 67—71, amphibolite facies gneisses
   IX: 72—74, red augengneises
    X: 75   , young red granite

only the major feldspar components, Or., Ab. and An., are considered here. These components are given as weight and mol percentages in Table 39 and are plotted on the projection of the quaternary system Or.-Ab.-An.-H_2O at 5000 bars H_2O pressure (Yoder et al., 1957) in Fig. 35. The inclined dotted line close to the Or. corner in this diagram separates the feldspars of the amphibolite facies area (to the right) from those of the granulite facies (to the left). The amount of plagioclase in the alkali feldspar increases markedly with increasing metamorphic grade.[1] The large pegmatite, no. 54, Table 39, is the only exception so far as the increase in An. is concerned. This is not a surprising exception because the pegmatite, which consists of quartz,

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[1] The alkali feldspars are either orthoclase perthites or microcline perthites. Ab. and An. are, therefore, no longer in solid solution in the feldspar. However, it is believed that they were originally in all of the feldspars considered.
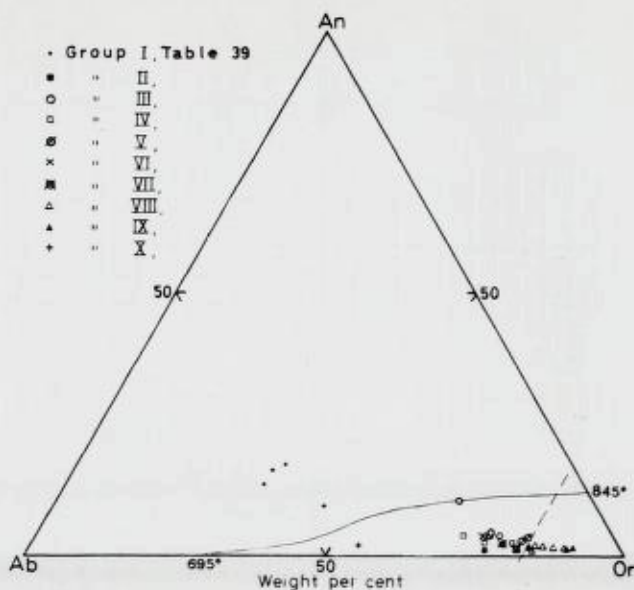
Fig. 35. Alkali feldspar compositions from Langøy plotted on the projection of the quaternary system $NaAlSi_3O_8(Ab)$-$KAlSi_3O_8(Or)$-$CaAl_2Si_2O_8(An)$-H_2O at 5000 bars H_2O pressure (Yoder et al., 1957). Dashed line separates between feldspars from amphibolite (right) and granulite (left) facies rocks.

Alkali feltspater fra Langøy plottet på projeksjonen av firekomponent systemet NaAlSi₃O₈(Ab)-KAlSi₃O₈(Or)-CaAl₂Si₂O₈(An)-H₂O ved 5000 bars H₂O trykk (Yoder et al., 1957). Brudt linje skiller mellom feltspater fra amfibolit (høyre) og granulit (venstre) facies bergarter.

microcline, and a little biotite, is extremely low in calcium as compared to the rocks from which the rest of the feldspars crystallized.

The feldspars from the young red granite and the Eidet-Hovden monzonites occupy a unique position in the diagram. It has previously been contended that the rocks have not been metamorphically recrystallized (unlike the other rocks from which the feldspars were separated), and the different positions of the feldspars on the diagram should thus correspond to a difference in origin. The fact that these feldspars are perthites (or mesoperthites) and not homogeneous is not contradictory to a magmatic origin, "for all natural feldspars having compositions near Ab_{50} Or_{50} consist of mixtures of nearly pure albite and pure potassium feldspar" (Tuttle and Bowen, 1958, p. 138). (The feldspar from the group III which plots on the cotetic line is contaminated with plagioclase, see pt. 4 Table 39.)

Temperature of formation of the alkali feldspars.

Barth (1934, 1951, 1956), has proposed to use the distribution

$$\text{coefficient; } k = \frac{\text{mol fraction of albite in alkali feldspar}}{\text{mol fraction of albite in plagioclase}}$$

as a geological thermometer; although the coefficient depends chiefly upon the temperature it also depends on pressure (Yoder et al., 1957, Bowen and Tuttle, 1950) and it may be influenced by chemical environment (Winkler and von Platen, 1958).

As to the work of Winkler and von Platen some comments may be made, the most significant of which is based on the interpretation of their experimental results: far from using a "normative" composition, they added increasing amounts of NaCl to their original clay sample (containing 0.42 per cent Na_2O) and thereby they found that by adding 4.8 per cent NaCl only 3.2 per cent took part in the reactions while the rest (1.6 per cent) remained as NaCl; no experiments were carried out in the interval between addition of 1.8 per cent (where all the NaCl took part in the reaction) and 4.8 per cent NaCl or at still higher Na concentrations. It seems that an addition of 3.2 per cent NaCl could give the same result as an addition of 4.8 per cent NaCl, and that the distribution coefficient k may be independent of the total amount of Na at still higher Na concentrations. The critical value then would be 2.11 wt. per cent Na_2O. In most normal potash feldspar bearing rocks the Na concentrations are higher than this and the chlorine content is nil. Along this line, Barth (1956, p. 14) also took care to point out that "the feldspar geologic thermometer is not accurate but should be specially adjusted for different compositions of the feldspar phases" . . . "Usually plagioclase is in the range 10—30 An., and the alkali feldspar in the range 10—30 Ab. In such rocks the distribution law would be approximately valid, and the K-ratio thus becomes an accurate relative temperature indicator."

The dotted line on Fig. 35 separates the feldspars east and west of the metamorphic boundary. The Ab. concentration in the plagioclases is in all cases similar (Table 39), it is usually slightly more *albitic*, however, in the amphibolite facies gneisses.

The temperatures of formation of the feldspars, based on "the 1956 values for the feldspar crystallization temperatures" (Barth, 1956 p. 15), are listed in Table 39. It is interesting to see that the temperatures of feldspar formation (i.e., the temperatures at the final establish-

Table 39.

Composition (Or, Ab, An), temperature of formation[1]), optical axial angle,

No	Wt. %			Mol %			Mol %		Ab in coex. plag.	k[3])	T°C
	Or	Ab	An	Or	Ab	An	Or	Ab			
Group I.											
50 (145/55)	34.6	47.6	17.8	33.6	49.1	17.3	40.6	59.4	73—77	0.771 0.814	(1110)
51 (an. 9)	33.6	50.1	16.3	32.6	51.6	15.8	38.7	61.3	73—80	0.766 0.840	(1150)
52 (an. 9)	33.2	53.7	13.1	32.1	55.2	12.7	36.8	63.2	73—80	0.790 0.866	(1170)
53 (179/55)	45.4	45.4	9.2	44.2	46.9	8.9	48.5	51.5	no independent plag. phase.		
Group II.											
54 (an. 7)	75.9	23.5	0.6	74.8	24.6	0.6	75.3	24.7	pegmatite with no plagioclase phase.		
Group III.											
55 (271/55)	67.1	22.2	10.7	66.2	23.2	10.6	74.0	26.0	70	0.371	625[4])
56 (293/55)	75.6	20.6	3.8	74.7	21.6	3.7	77.6	22.4	70	0.320	590
57 (255/55)	77.5	18.9	3.6	76.6	19.8	3.6	79.5	20.5	70	0.293	570
58 (251/55)	75.7	20.0	4.3	74.8	21.0	4.2	78.1	21.9	70	0.313	585
Group IV.											
59 (318/55)	71.1	25.1	3.8	70.1	26.2	3.7	72.8	27.2	77	0.353	610
60 (238/55)	75.2	22.4	2.4	74.2	23.5	2.3	75.9	24.1	70	0.344	605
61 (224/55)	80.0	17.8	2.2	79.1	18.7	2.2	80.9	19.1	68	0.281	560
Group V.											
62 (an. 12)	82.1	14.4	3.5	81.3	15.2	3.5	84.2	15.8	70	0.226	510
63 (348/55)	81.7	15.5	2.8	80.9	16.3	2.8	83.2	16.8	72	0.233	515
Group VI.											
64 (65/56)	74.8	21.5	3.7	73.8	22.5	3.7	76.6	23.4	70	0.334	595
Group VII.											
65 (25/55)	81.4	17.2	1.4	80.6	18.0	1.4	81.7	18.3	69	0.265	545
66 (150/56)	78.5	19.4	2.1	77.6	20.3	2.1	79.3	20.7	72	0.288	565
Group VIII.											
67 (147/56)	87.8	11.2	1.0	87.2	11.8	1.0	88.1	11.9	72	0.165	440
68 (3/56)	83.9	15.1	1.0	83.1	15.9	1.0	83.9	16.1	72	0.224	505
69 (35/55)	84.4	14.1	1.5	83.7	14.8	1.5	85.0	15.0	80	0.188	470
70 (37/55)	85.4	12.9	1.7	84.7	13.6	1.7	86.2	13.8	74	0.186	465
71 (an. 1)	89.5	9.1	1.4	89.0	9.6	1.4	90.3	9.7	75	0.129	400
Group IX.											
72 (109/56)	90.6	8.4	1.0	90.1	8.9	1.0	91.0	9.0	70	0.129	400
73 (99/56)	83.1	15.5	1.4	82.3	16.3	1.4	83.5	16.5	67	0.246	525[5])
74 (E-vatn)	90.0	9.0	1.0	89.5	9.5	1.0	90.4	9.6	100	0.090	300
Group X.											
75 (an. 5)	54.8	43.8	1.4	53.3	45.3	1.4	54.1	45.9	88	0.522	780

Table 39.

and triclinicity ($\triangle$)² of a selected group of alkali feldspars from Langøy.

$2V_x$[7]	$\triangle$[8]	Comments[9]
	0.10	mesoperthite
	0.10	mesoperthite, small orthoclase perthites in the groundmass have $2V_x$ 40—48.
	0.14	mesoperthite
	0.09	mesoperthite
	0.23	microcline twinning
50—55	0.24	optically monoclinic
	0.10	optically monoclinic
	0.09	optically monoclinic
	0.14	optically monoclinic
60	0.19	some grains exhibit microcline twinning, others appear optically monoclinic.
	0.09	optically monoclinic
63—65	0.10	no microcline twinning, but slightly inclined extinction, $X \wedge$ (010) on (001) = 9
40	0.24	optically monoclinic
47	0.13	optically monoclinic
47—53	n.d.	optically monoclinic
46	0.24	optically monoclinic
50	0.19	optically monoclinic
70—80	0.86	microcline twinning
60	0.25	optically monoclinic
55—80	0.91	no microcline twinning visible, but strong undulatory extinction. One «monoclinic» (131) reflection on 2.990 Å in between the «triclinic» (131) and (131) reflections, clearly resolved.
60—72	0.21	microcline twinning
72—81	0.78	microcline twinning
71	0.84	microcline twinning
80	n.d.	microcline twinning
84	1.01	microcline twinning
68—73	0.39	mesoperthite, (131) and (1$\overline{3}$1) reflections are clearly resolved.

Table 39 (footnotes).

[1]) 1956 values for the feldspar crystallization temperatures (Barth, 1956, p. 15).

[2]) Triclinicity: $\triangle = 12.5 \ (d(131) - d(1\bar{3}1))$ (Goldsmith and Laves, 1954 a).

[3]) $k = \dfrac{\text{mol fraction of albite in alkali feldspar}}{\text{mol fraction of albite in plagioclase}}$ (Barth, 1956, p. 6).

[4]) Some of the large potash feldspar porphyroblasts were handpicked from the sample whereas the other feldspars were separated from the rocks with heavy liquids. The relatively high Ab and An content in the former is due to slight contamination by plagioclase; nos. 56, 57, 58 are probably more representative of the alkali feldspars from group III.

[5]) Inclusions of the surrounding plagioclase in the potash feldspar are seen in thin sections of this rock. Contamination by plagioclase of the potash feldspar is, therefore, probably responsible for the high Ab content in this feldspar, and, thus, also for the indicated high temperature of formation.

[7]) A.P. = (010) in all samples.

[8]) (1$\bar{3}$1) and (131) reflections are not resolved when the triclinicity is less than 0.25 (Guinier-type focusing camera, Nonius Co., Delft, Holland, Fe rad., 14 mA, 40 KV, Kodirex X-ray films, 20 hrs. exp.). Triclinicity values less than this indicate the width of one single reflection i.e., its diffuse character. For triclinicity values between 0.20 and 0.25 the feldspars may or may not be optically monoclinic (exhibit microcline twinning); for values less than 0.20 the feldspars are optically monoclinic (see also Heier, 1957, p. 472).

[9]) In the monoclinic and close to monoclinic potash feldspars the extinction angles were difficult to measure exactly on the universal stage because of the low birefringence ($n_y \approx n_z$). Therefore, they are not listed in the table except for some special cases.

Group I—X, see Table 38.

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ment of equilibrium in the rocks, see Dietrich, 19—), west of the metamorphic boundary (granulite facies) range between 510—625° C (excluding the high temperature Eidet-Hovden monzonites and the young granites). East of the boundary they lie in the interval 300—505° C (excluding no. 73 which includes pieces of the surrounding plagioclase and is thus contaminated, pt. 5, Table 39). No. 74 (T = 300° C) represents a young red granite pegmatite vein not directly related to the country rock. The amphibolite facies gneisses proper are thus found in the interval 400—500° C. According to this the transition between amphibolite and granulite facies should take place at a temperature a little above 500° C. This is in good agreement with the temperature usually assigned to this facies transition. Using Barth's thermometer temperatures not less than 1000° C are obtained for the mesoperthitic feldspar from the monzonites within the Eidet-Hovden intrusion if they are considered in conjunction with the groundmass plagioclase. It appears that these mesoperthites were not in equilibrium with any pla-

gioclase, i. e., they are believed to have crystallized before the plagioclase and normal orthoclase perthites of the groundmass. The distribution coefficients given in Table 39 are related to the plagioclases of the groundmass, and the temperatures should thus be minimum temperatures. There may be some doubt as to whether this thermometer is valid in such cases to give even approximate temperatures. These feldspars plotted on the quaternary system Ab.-Or.-An.-H_2O at 5000 bars H_2O pressure (Yoder et al., 1957, Fig. 41, p. 211), fall in the 800°—900° C range. It was mentioned on p. 95 that these rocks may have been intruded at shallow depths in which case the pressure would be considerably lower and the temperatures correspondingly higher to the point that they might approach the temperatures indicated by the feldspar thermometer.

In any case it can be assumed that the temperature of crystallization of these feldspars was much higher than the maximum measured metamorphic temperatures in the area.

The temperature of formation of the red granites (780° C) was discussed on p. 111; this also is higher than the metamorphic temperatures reached.

The temperatures of formation of the retrograde metamorphosed gneisses (group IV, Table 39) are found to be similar to those of the western monzonites (Group III). The temperature given is apparently that reached when the rocks were under granulite facies conditions and not that of the retrograde metamorphic conditions.

No. 59 (318/55) is listed in, Table 12; nos. 60 and 61 are not.

No. 60 (238/55) contains, quartz, K-feldspar, plagioclase, green hornblende, biotite, apatite, sphene and ore; no. 61 (224/55) contains, K-feldspar, plagioclase, serpentinized hypersthene, diopsidic pyroxene partly altered into green hornblende, biotite, apatite and ore.

The reasons for the comparatively large temperature ranges within the two major facies areas are not easy to explain. It seems probable that other factors than temperature may control the Na distribution between the feldspar phases. The western monzonites (group III), the retrograde metamorphic gneisses (group IV) and the old red granite (group VI) probably existed together within a narrow temperature range at about 585 ± 25° C (the temperature determined for no. 55 is influenced by contamination with plagioclase). This is in contrast to the neighbouring veined granulite facies rocks at 535 ± 25° C (groups

V and VII). The difference is not likely dependent upon different Na contents of the respective host rocks. Chemical analyses of the western monzonites, Table 2, show 5.22 per cent and 5.55 per cent Na_2O. The host rocks of the feldspars nos. 62 (an. 12) and 66 (150/56), Table 39, were analysed (Tables 4 and 7), and showed 4.91 per cent (T = 510° C) and 3.60 per cent Na_2O (T = 565° C) respectively. The Na content of one of the retrograde metamorphosed gneisses (no. 59 (318/55), Table 39) is 5.03 per cent Na_2O (Table 13) (T = 610° C).

The significant difference between the porphyroblastic and banded rocks is the presence of much larger potash feldspar in the former. A possibly similar effect was illustrated by Barth (1956, Fig. 7, p. 33); he showed the small and large pegmatites to have been formed at temperatures 150°—200° C higher than the country rock gneisses. Still higher temperatures were found for the augengneisses. (These latter temperatures may be somewhat exaggerated as most of the augengneiss feldspars used in Barth's study were handpicked from the samples and not separated with heavy liquids and thus they may have been contaminated by included plagioclase grains as was the case with no. 55, Table 39. The augengneiss feldspars may, therefore, have been formed at roughly the same temperatures as the pegmatites, i. e., temperatures closely similar to those of the porphyroblastic rocks on Langøy). Barth (1956, pp. 32—34) ascribed the corresponding variations in the distribution coefficient entirely to variations in temperature of formation. But, if this were true steep temperature gradients must exist in the deep zones of high grade regional metamorphism. In any case the pegmatites and augengneisses would have to have been superheated with respect to their country rocks. The author believes, however, that whereas this is not impossible and indeed agrees with the more orthodox views of the formation of these rocks, it is not an entirely satisfactory explanation. The lower Na content of the fine grained alkali feldspar crystals most likely may reflect a relatively greater loss of Na from these to the surroundings by exsolution than from the large porphyroblasts (see also Orville, 1958, p. 206).

Variations in the symmetry of the alkali feldspars.

In southern Norway it has been found that, "where it has been possible to be certain, it has turned out that orthoclase is never found in rocks where the metamorphism is lower than amphibolite facies, while microcline is the alkali feldspars in these rocks and also in the

lower part of amphibolite facies" (Heier, 1956 a, p. 199). (Orthoclase was therein defined as a feldspar appearing optically monoclinic; the same definition is used in this paper). Heier (1957) also found that "The inversion monoclinic $\rightleftarrows$ triclinic potash feldspar seems to take place at (PT) conditions a little below the transition between granulite and amphibolite facies" and further "When the (PT) conditions fall below those of high-grade amphibolite facies, the complete ordering of the potash feldspars seems to take place within a narrow (PT) field. Thus, for rocks of undoubted amphibolite facies, nearly all ranges of triclinicities may be found" (p. 477).

The feldspars in the Langøy rocks are in complete agreement. As mentioned in the petrographic sections the granulite facies rocks contain "orthoclases" whereas the amphibolite facies gneisses typically contain microclines of varying triclinicities (variations in the size of $2V_x$). "Orthoclase" was also found in the latter but the sizes of the $2V_x$ as compared to those in the granulite facies feldspars suggest a high degree of ordering in these as well.

As the microcline structure exhibits the highest degree of Al-Si order it is generally accepted that this is the alkali feldspar stable at low temperatures. Increasing temperatures will create a higher degree of Al-Si disorder and gradually the monoclinic, completely disordered form, sanidine, is attained and is then the stable high temperature alkali feldspar. Sanidine and orthoclase are not synonymous; the latter is pseudomonoclinic rather than monoclinic (Goldsmith and Laves, 1954 b) whereas the former is truely monoclinic. The major difference between sanidine and orthoclase is to be found in the distribution of Al and Si in the tetrahedral groups: Si_1 (O), Si_1 (m), Si_2 (O), Si_2 (m), (notations by Megaw, 1956). Each group contains a sum of $Si + Al =$ 4. The Al distribution pattern in the 4 sets is listed in Table 40.

The stability fields of the three modifications are not quite known. Sanidine is the high temperature phase and microcline is the low temperature phase. Whether orthoclase has any real stability field is not known. Sanidines are, so far as is known by the author, not known in regionally metamorphosed rocks. The transition series which should be considered in these rocks is the one from microcline to orthoclase, or to a phase between orthoclase and sanidine. Microcline has the most ordered structure and should, therefore, according to normal considerations be the low temperature phase. In studying the electrostatic charge distribution in the alkali feldspars, Ferguson et al. (1958, p.

Table 40.

Ideal distribution of Al atoms (per cent)
in the four tetrahedral groups in alkali feldspars.

	$Si_1(O)$	$Si_1(m)$	$Si_2(O)$	$Si_2(m)$
sanidine (disordered monolinic)........	25	25	25	25
orthoclase (partly disordered monoclinic)	50	50	0	0
microcline (completely ordered, triclinic)	0	100	0	0

348) concluded: "....., it seems highly probable that an enhanced degree of Al-Si ordering need not necessarily be associated with greater stability in the potassium feldspars; and that there is not one series increasing continuously in stability from sanidine to maximum microcline, but two series of which one begins with sanidine and runs to a partially ordered orthoclase as both degree of Al-Si order and stability increase, while the other begins with a highly ordered maximum microcline and also runs to the same monoclinic orthoclase but with stability increasing as order decreases". According to these authors the microcline lattice is stabilized by Na substituting for K. However, available data and also the data of Table 39 show that increases in triclinicities correspond to decreases in sodium concentrations. In fact all empirical geological data suggest microcline to be the low temperature modification of the alkali feldspars.

Data on the variations of the $2V_x$ and triclinicity Δ (Goldsmith and Laves, 1954 a) are given in Table 39, and plotted against temperature in Fig. 36 a and b. In rocks of the amphibolite facies area the $2V_x$ has not been found to be less than 55°—60°, and in rocks of the granulite facies it has not been found to exceed 50°—55° and is typically less than 50°. Especially in the amphibolite facies gneisses $2V_x$ has been found to vary considerably even within the same "crystal" (55°—80°, no. 69, Table 39). This is characteristic of intermediate microclines (Goldsmith and Laves, 1954 b). The feldspars from the retrograde metamorphosed gneisses which yielded the original granulite facies temperature, have adjusted their symmetry, as reflected by the 2V, to the changing conditions. 2V seems to be more sensitive to symmetry variations than triclinicity, i. e., the 2V undergoes a rather marked change before the splitting up of the $(131—1\bar{3}1)$ reflections is noticeable.

The latter show no splitting up above 470° C, but directly below
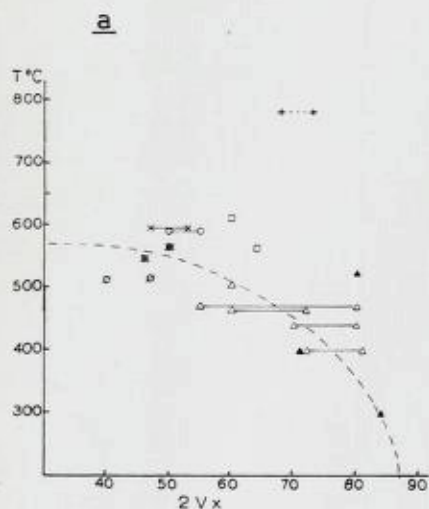
Fig. 36 a.
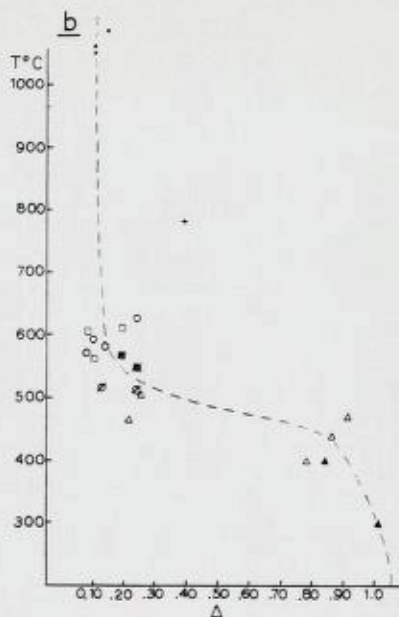
Fig. 36 b.

Fig. 36 a. Relationship between $2V_x$ and temperature of formation (Barth, 1956) of alkali feldspars from the Langøy rocks. Symbols as in Fig. 35.

Forholdet mellom $2V_z$ og dannelsestemperatur (Barth, 1956) av alkali feltspater i bergartene fra Langøy. Symboler som på Fig. 35.

Fig. 36 b. Relationship between triclinicity (Δ) and temperature of formation (Barth, 1956) of alkali feldspars from the Langøy rocks. Symbols as in Fig. 35.

Forholdet mellom triklinisiteten (Δ) og dannelsestemperaturen (Barth, 1956) av alkali feltspater i bergarter fra Langøy. Symboler som på Fig. 35.

that a high triclinicity obtains. Intermediate triclinicities are rare and it seems that ordering of the potash feldspars takes place within a narrow field.

The size of 2V is not dependent upon symmetry alone but also on the total amount of Na contained in the feldspar (Fig. 20). However, the change in the 2V of highly ordered microclines with changing Na content is negligible within the concentration range exhibited by the Langøy feldspars, whereas the 2V change of the orthoclases within this range is marked. Table 41 compares the analysed concentrations of

Table 41.

A comparision between chemically analysed Ab. + An.
contents and contents derived from the size
of the $2V_x$ of the orthoclases.

No	Ab. + An. (analysed)	$2V_x$	Ab. + An. (derived from the $2V_x$)
56.........	24.4	50—55	17—23
62.........	17.9	40	9
63.........	18.3	47	16
64.........	25.2	47—53	16—20
65.........	18.6	46	15
66.........	21.5	50	17

Ab. + An. in the orthoclases with those derived by reading on the graph, Fig. 20.

The optically derived Ab. + An. concentrations are systematically lower than those determined chemically. This shows the 2V to be an unreliable indicator of chemical composition. Nontheless the variations of 2V in the orthoclases may reflect differences in the Na content rather than variations in the symmetry.

The change in symmetry and the increased ability to take Na into solid solution at increasing temperatures have opposite effects on the variation of the 2V. As the amount of Na in the alkali feldspars decreases with decreasing metamorphism the observed increase in 2V must be related to the change in symmetry in the general series from orthoclase to microcline. The dotted line on Figs 36 a and b indicates that approximate maximum ordering of alkali feldspars was achieved at temperatures a little less than 300° C (these contain also the minimum amount of Na) and that the partly disordered orthoclase structure is attained at temperatures above approximately 510° C. By hydrothermal heat treatment of a microcline from Madagascar, Goldsmith and Laves (1954 a) found the monoclinic phase to develop at about 500° C after 500 hrs. They call this phase a sanidine but it is not known whether it corresponds to sanidine or orthoclase as these were defined in Table 40. In any case, this temperature corresponds closely to the temperature at which orthoclase is stable in regionally metamorphosed rocks as determined by the feldspar thermometer. The orthoclase-microcline transformation further defines closely the transition between granulite and amphibolite facies.

Although the above conclusions agree with earlier studies within

other areas (Heier, 1956, 1957), their general validity is somewhat uncertain according to data presented by Howie (1955) for the Madras charnockite series. (The striking similarity between these rocks and the granulite facies rocks on Langøy has been mentioned.) Howie (1955, p. 746) wrote: "The potash feldspars from the Madras charnockite series are dominantly microcline perthites and microperthites, although, as has been mentioned earlier, closer investigation often shows them to be a variety intermediate between orthoclase and microcline." His Table IV (p. 745) indicates clearly the much higher 2V of the Madras charnockite feldspars, and only one (no. 2270, $2V_x$ 52°—60°) is within the upper range reached by the granulite facies rocks on Langøy. The feldspar thermometer indicates a temperature of formation of 485° ($2V_x$ 52°—60°) and 540° C ($2V_x$ 69°) for nos. 2270 and 6436 respectively (the only two rocks from which the coexisting plagioclase compositions were given). It is difficult to offer any feasible hypotheses for the deviating behaviour of the Madras charnockites. One explanation could be that they originally had a lower water content than the Langøy rocks (the temperature of the transition amphibolite facies → granulite facies is dependent upon the water pressure). A low water content would lower the temperature of the facies transition. No evidence of an originally lower water content in the Madras charnockites was offered. On the basis of geochemical evidence Howie assumes an original igneous origin. The typical charnockites of the banded series and the border series on Langøy are considered by the author to represent original geosynclinal sediments and lavas. Thus a higher water content may have been present in these. The amphibolite facies to granulite facies transition would then take place at a higher temperature, possibly one which would correspond to the temperature where orthoclase becomes stable.

According to Wilson (1954) orthoclase is characteristic of the charnockitic rocks in Australia.

Finally the position exhibited by the feldspars from the young red granite should be pointed out (Fig. 36 a and b). In spite of a high temperature of formation its feldspar has an intermediate triclinicity ($\Delta = 0.39$) and shows a splitting up of the (131) and ($1\bar{3}1$) reflections. The triclinicity varies from place to place within the same intrusion (the example given was measured on the sample used for chemical analyses). The significance of this has been discussed previously by the author (Heier, 1957, p. 476), and it was found to agree with a post metamorphic intrusion of the granite.

Geochemistry.

Sample preparation and analytical methods.

The rocks.

A total number of 43 rock samples were analysed for both major and trace elements. Most samples were of "hand specimen" size, approximately 300—500 gms. weight. A few larger samples of several kg. were analysed.

The samples were crushed in a steel mortar to a size of — 80 mesh and finally ground in an agate mortar to — 120 mesh.

Major elements.

For most rocks ordinary "wet" chemical methods were used to determine the major elements. These were carried out in the laboratory of the Geological Survey of Norway, Oslo.

For the rest of the rocks (marked with asterisks in Table I) the major elements, except Na, K and ferrous iron, were determined spectrographically. Na and K were determined by flame photometer and ferrous iron by titration, of a sulphuric acid solution with ceric sulphate, o-Phenantholine-ferrous sulphate complex as indicator. All the spectrographic work was done in the Department of Geology and Mineralogy, Oxford; a Hilger large quartz and glass optical spectrograph was used. G-1 (granite) and W-1 (diabase), Fairbarn et al. (1951), were used as the primary standards in all the spectrographic determinations of both rocks and minerals. The adopted concentrations of the various elements in G-1 and W-1 are given in Table 42.

Spectrographic procedures are as follows:

Ca, Mn, Fe, Ti.

λ 3650—4850 A, glass optics, slit 10 microns, wedge 11, step sector; lower electrode (anode) 2.4 mm. int. diam. $\times$ 2.5 mm. depth Johnson Matthey 4 B graphite; upper electrode (cathode) "Ship" carbon 5 mm. diameter; Ilford ordinary plates; current 6.8 amps. d.c. Samples were burned to completion (about 60 sec.); 1 part sample mixed with 5 parts graphite/paladium complex salt mix, mix: 9 parts graphite to 1 part $((NH_3)_4Pd)$ $(NO_3)_2$.
Lines read: Ca (4425 and 4435), Mn (4034), Fe (4251), Ti (3390); internal standard line Pd (4473).

Table 42.

Values used for primary standards
G- 1, and W- 1.

	W- 1	G- 1
Al ..	8.3 %	7.9 %
Ba ..	150 p. p. m.	1300 p. p. m.
Ca ..	7.9 %	1.0 %
Co ..	49 p. p. m.	(4) p. p. m.
Cr ..	130 p. p. m.	20 p. p. m.
Cs ..		2.5 p. p. m.
Cu ..	110 p. p. m.	11 p. p. m.
Fe ..	7.9 %	1.42 %
Ga ..	20 p. p. m.	15 p. p. m.
K ...	0.53%	4.5 %
Li...	9 p. p. m.	22.5 p. p. m.
Mg..	4.0 %	0.24 %
Mn .	1300 p. p. m.	210 p. p. m.
Na ..	1.5 %	2.4 %
Ni ..	90 p. p. m.	(5) p. p. m.
Pb ..	7 p. p. m.	27 p. p. m.
Rb ..	25 p. p. m.	220 p. p. m.
Si ...	24.4 %	33.7 %
Sr...	180 p. p. m.	240 p. p. m.
Ti ..	0.67 %	0.15 %
V ...	200 p. p. m.	25 p. p. m.

Si, Al, Mn, Mg.

(If not directly stated the conditions were as above). λ 2470—3560 Å, quartz optics, samples were burned to completion (about 80 sec.);
1 part sample mixed with 2 parts graphite/paladium mix: 10 gm. graphite: 0.015 gm. $((NH_3)_4(NO_3)_2$.
Lines read: Si (2532), Al (2652), Mn (2801), Mg (2783 and 2803); internal standard line Pd (3421).

Si and Al are not too well determined by this method. To obtain better values for these elements the following calculations were made as suggested by Taylor (1953).

1. Sum of Fe_2O_3, MgO, CaO, Na_2O, K_2O, TiO_2, k = X (k represents the sum of P_2O_5, H_2O and trace elements and is taken as 1.5 per cent for the amphibolites and gneisses and 0.5 for the red granites).

Then 100 — x = SiO_2 + Al_2O_3

2. The Al (2652)/Si (2532) ratio is obtained from the spectrographic data. This is converted to the ratio Al_2O_3/SiO_2 (R) from a graph drawn from the standard sample values.

3. Al_2O_3 and SiO_2 are then obtained by solving the equations 1 and 2 below.

$$(1) \quad SiO_2 + Al_2O_3 = 100 — X$$
$$(2) \quad Al_2O_3/SiO_2 = R$$

Minor (trace) elements.

The minor elements in the rocks were all determined spectrographically. Spectrographic details are as follows.

Li, Rb and Cs.

Because Na and K of all the rocks were independently determined by flame photometer the spectrographic determinations of Li, Rb and Cs were made by using the variable internal standard method described by Ahrens (1954 b, pp. 52—59) with Na as the internal standard. λ 4600—9600 A; glass optics; Kodak IRER plates; lower electrode (anode) 1.6 mm. int. diam. x 5 mm. depth, J. M. 4 B graphite; cathode as for Pd run; current: 4.4 amps. d.c., slit 20 microns; wedge 11; exposure: samples were arced until the alkali metals had completed destillation. The average time necessary was about 120 sec. Developing was as previously described.

Lines read: Li (6707), Rb (7800) and (7947), Cs (8521); internal standard line: Na (5682).

Ga, Cu, V, Ni, and Co. λ 2750—4680 A; quartz optics; slit 10 microns; wedge 11; stepsector; current 6.8 amps. d.c., Ilford ordinary plates; lower electrode (anode) 2.4 mm. int. diam. x 3 mm. depth J. M. 4 B graphite; cathode as previously; samples mixed 1:2 with carbon; samples were burned to completion (about 90—100 sec.); development as previously.

Lines read: Ga (2944), Cu (3274) with internal standard line Na (3302); V (3185), Ni (3414), Co (3453) with internal standard line Fe (3287).

Ba, Sr, Cr, V. λ 3650—4850 A; glass optics; lower electrode (anode) 1.6 mm. int. diam. x 2 mm. depth J. M. 4 B graphite; rest of conditions as in the previous run; burning time 40—50 sec.

Lines read: Ba (4554), Sr (4607) with internal standard line Ca (4425); Cr (2454), V (4379), with internal standard line Fe (4251).

In the rocks where the major elements were determined spectrographically the minor elements were determined together with these using Pd as an internal standard.

V, Cu, Ni, Co, Ga were determined with Si, Al, Mn, Mg. Lines read were: V (3185), Cu (3274), Ni (3414), Co (3453), with internal standard line Pd (3421). Ga (2944) was determined using Na (3303) as a variable internal standard.

Ba, Sr, Cr were determined with Ca, Mn, Fe, Ti. Lines read were: Ba (4554), Sr (4607), Cr (4254), internal standard line Pd (4473).

Pb and Tl. λ 2750—4680 A; quartz optics; slit 20 microns; wedge 11; step sector; current 4.4 amps. d.c.; Ilford ordinary plates; samples unmixed; lower electrode (anode) 3.2 mm. int. diam. x 5 mm. depth J. M. 4 B graphite; upper electrode (cathode) National Carbon Company special carbon 3.2 mm. diam.; development as previously. Samples were arced until the alkali metals had completed destillation. Burning time varied with the alkali content in the rocks. From 2 min. 50 sec. for some of the gabbros to 8 min. for the most alkali rich rocks.

Line read: Pb (2833). Tl 3775.7) was not determined in the rocks because of interference by Ni (3775.6). Internal standard line Na (3303).

The minerals.

Only the potash feldspars, plagioclases and biotites were studied individually by spectrochemical means and in no case was a complete chemical analysis made

Potash feldspar.

The rocks were ground to — 80 mesh and the dust washed away. Mineral separation was then carried out with acetylene tetrabromide which was gradually diluted with acetone. The concentrates were controlled with immersion oils to make certain that no plagioclase or quartz were present. The concentrates thus obtained were all run through a Franz electromagnetic separator to take away minor amounts of dark minerals, especially biotite, which adhered to the feldspars in the liquid. Very pure concentrates of alkali feldspars were obtained in this way.

Analytical procedure.

Na and K were determined by flame photometer.

Rb, Li and Cs were determined spectrographically. The method was the same as previously described for the rocks, but the slit opening was set at 30 microns instead of 20 and slightly bigger anodes were used (2.4 mm. int. diam. x 3 mm. depth). The samples were arced until the alkali metals had completed distillation (about 100 sec.). In addition to the primary basic standards G-1 and W-1 a feldspar (M-2) containing 1800 p.p.m. Rb (Taylor & Heier, 1958 a) was used as a standard.

Lines read: Rb (7800), 7947), Li (6707); (Cs 8521) was not detected in any of the samples). Internal standard line Na (8183).

Ca, Sr and Ba. The method is described by Heier and Taylor (1959 a).

Cu, Pb and Tl. With the exception of a slightly smaller anode (internal diam. 2.4 mm. x 5 mm. depth) the conditions were identical with those for the Pb determinations of the rocks. Working curve for Tl was based on two standards kindly supplied by Dr. D. M. Shaw, (1952), McMaster University, Canada. S-98 cleavelandite 6.2 p.p.m. Tl. 7.52 % Na, 0.88 % K.

S-100 microcline perthite 84 p.p.m. Tl. 0.82 % Na, 11.09 % K.

Lines read: Cu (3247), Tl (3775), Pb (4057).

Internal standard line: Na (3303).

Plagioclase.

For the sake of comparison with the potash feldspars a small number (9) of coexisting plagioclases were also partially analysed. While one with repeated separations in heavy liquids can obtain a plagioclase concentrate fairly free of potash feldspars (the composition of the plagioclases is about An. 30), the separation of plagioclase from quartz is a much more difficult task. For the purpose of the analyses, however, this separation was not considered necessary. The purifications of the fractions were done with bromoform gradually diluting with benzene. The fractions were controlled in immersion oils to ascertain that no free grains of potash feldspars were present. The pure fractions were all run through a Franz electromagnetic separator to remove traces of dark minerals.

Na and K were determined by flame photometer.

Li, Rb and Cs. The conditions for these elements were as for the rocks. A slightly bigger anode was used (int. diam. 2.4 mm. x 5 mm.).

Burning time 180—225 sec. (end of alkali distillation).

Lines read: Li (6707), Rb (7800—7947); Cs (8521) was not detected; internal standard line: Na (5688).

Ca, Ba and Sr. The conditions were indentical with those used for the potash feldspars with the exception of slightly bigger electrodes (2.4 mm. int. diam. x 3 mm. depth). The samples were burned to completion (80 sec.).

Cu, Pb and Tl. These elements were determined in the plagioclases as in the potash feldspars. Tl (3775) was not detected in any of the plagioclases.

Biotite.

Biotite is a third mineral in which the alkali elements are concentrated. For the sake of comparison with plagioclase and potash feldspar a small number (7) of coexisting biotites were partially analysed.

The biotites were concentrated with a Franz electromagnetic separator from the heavy fractions after the feldspar separations. Pure biotite fractions were obtained in this way though slight contamination by hornblende and diopsidic pyroxene was difficult to avoid.

After the separations the biotites were ground in a mica cutter (Neumann, 1956) until they passed through a 120 mesh bolting cloth.

Na and K were determined with a flame photometer.

Li, Rb and Cs. The spectrographic conditions for these elements were identical with those used for the plagioclases. Because of the low concentration of Na in the biotites K was used as the variable internal standard.

Lines read: Li (6707), Rb (7800 and 7947), Cs (8521).

Internal standard line: K (6939).

Tl was not read because of the possible interference by Ni as mentioned under the analytical description of the rocks. The concentration of Pb was too low to register. (The experimental conditions were as for the rocks.)

Presentation of data.

The major elements of the rocks are given as oxides in the petrographic text. For the sake of facility of reference they are also assembled in the attached Table I. The cation percentages used for calculating the molecular norms are given in the same Table.

In Table II in the appendix the major and trace elements are given as p.p.m. of the metal. The elements are listed in order of increasing ionic radii (Goldschmidt values).

The analyses of the potash feldspars, plagioclases and biotites are given in Tables III, IV, and V, respectively.

In the discussions of the distribution of each element attention should be given to attached Tables (nos. I to V). The numbers, e. g., 1, 2, 3 . . . etc., and (an. 16), (an. 8), (32/56 d) . . . etc. refer to the designations given in these tables.

The major elements.

The metal concentrations of the elements are used both for the major and trace elements, in the following discussions.

Variation diagrams of the major elements (except Ti, Mn, and P) are given in Fig. 37. The plotting method used is the one devised by Larsen (1938) as modified by Nockolds (1953). The elements are plotted against the "position" of the rocks which is defined by the term $(\frac{1}{3} Si + K) \div (Ca + Mg)$. In order to show iron enrichment Fe is omitted from the base.

There is no established method for plotting the variations of major elements in a heterogeneous series of high grade metamorphic rocks such as the Langøy rocks. Most existing variation diagrams are meant to show variations in igneous rock series. The chief reason why the Larsen—Nockolds method is used in this paper is because it was used by Howie (1955) in discussing the geochemistry of the Madras charnockite series which, as mentioned, is similar to at least some of the rock types on Langøy. By comparing Fig. 37 with Text-fig. 1, of Howie (1955, p. 735) it can be seen that almost identical curves obtain for the different elements in the rocks of the two areas.

The Langøy rocks tend to be richer in Al than the Madras charnockites which are slightly low in this element. Apart from this the two curves are extremely similar. Two of the gabbros (nos. an. 16 and an. 8) from the Eidet—Hovden intrusion are much higher in Al than the "position" of the rocks should imply. This is because of the high content of plagioclase in these two rocks (72 and 63.9 per cent normative Ab + An, respectively). The quartz-plagioclase-biotite schist from Vik (43/55) is low in Al; this is reflected by its somewhat peculiar mineral composition (Table 14). The same rock is higher than the average in Mg but is normal in Fe. The curve for Na is interesting in that it reveals that the lowest grade metamorphic rocks (charnockite border series, amphibolite facies gneisses, and associated sedimentary rocks) tend to be relatively low in this element. This tendency is most marked in the intermediate and acid composition rocks but is also apparent in

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Fig. 37. Variation diagrams for the major elements in the Langøy rocks.
Variasjonsdiagrammer for hovedelementene i bergartene fra Langøy.

the basic members. A similar systematic variation does not occur in the case of K even though two of the rocks from the banded series (nos. 58/56 and 59/56) are low in K. Therefore the total effect is that the Na/K ratio increases with increasing metamorphic grade. As this is mainly dependent upon variations in the Na concentrations it may be related to the increased ability of Na^+ (0.98 Å) to substitute for K^+ (1.33 Å) in alkali feldspars as the temperature increases. One of the retrograde metamorphosed rocks (235/55) is extremely rich in K. The rock is regarded as a retrograde metamorphosed equivalent of the old red granites and the extreme K content may indicate K metasomatism.

The Fe^{3+}/Fe^{2+} ratios, listed in Table 43, show no regular variation trends with the metamorphic grade. One group of comparatively high ratios (> 0.5) includes the dykes and monzonites within the Eidet—Hovden intrusion, the porphyroblastic monzonitic granulites, the retrograde metamorphosed gneisses, the young red granites, and most of the amphibolites associated with sedimentary rocks in both facies areas. Ratios less than 0.5 include most of the banded and veined gneisses and the Eidet—Hovden gabbro. Howie (1955, p. 734) remarked on the low Fe^{3+}/Fe^{2+} ratios of the Madras charnockites (0.2—0.5) and considered them to suggest high temperature metamorphism. The very low ratio of 70/56 (intermediate rock associated with sedimentary rocks) of the Langøy rocks appears to depend upon the presence of most of the Fe as pyrrhotite. The Fe^{2+}/Mg^{2+} ratios are also given in Table 43. The young red granites tend to have a high ratio, as was expected. The monzonites of the Eidet—Hovden intrusion (especially an. 9) also tend to have high ratios. Apart from this no significant trends are obvious.

The fact that the major elements of the diversely originated rocks on Langøy all plot on relatively smooth curves on the variation diagrams of the "Larsen type" (Fig. 37) shows especially well the futility of using them as indicators of the origin of these highly metamorphosed rocks.

The minor (trace) elements.

Rubidium (Rb+ 1.49 Å).

A plot of percentage K (K^+ 1.33 Å) versus p.p.m. Rb for the Langøy rocks superimposed on the K/Rb curve of Ahrens et al. (1952, Fig. 2) is given in Fig. 38. This curve has been adjusted by Taylor et

Table 43

Fe^{3+}/Fe^{2+} and Fe^{2+}/Mg^{2+} ratios of rocks from Langøy.

No.	Fe^{3+}/Fe^{2+}	Fe^{2+}/Mg^{2+}	Rock type	
1 (an. 16) ...	0.17	0.68	Gabbro	
2 (an. 8)	0.06	0.61		
3 (32/56 d) ..	0.06	1.12		Eidet–
4 (an. 15) ...	0.71	3.09	Monzonite	Hovden
5 (an. 9)	0.45	4.56		
6 (50/56) ...	0.72	2.15	Dykes	intrusion
7 (91/56 d) ..	0.76	2.68		
8 (60/56)	0.73	4.12		
9 (57/56)	0.75	1.99	Porphyroblastic monzonitic	
10 (55/56) ...	0.86	2.47	granulites	
11 (318/55) ..	0.74	2.57	Retrograde metamorphosed	
12 (310/55) ..	0.43	2.80	gneisses	
13 (235/55) ..	2.21	3.12		
14 (58/56) ;..	0.53	2.51	The banded series	
15 (145/56) ..	0.08	1.05		
16 (an. 12) ...	0.72	1.96		
17 (59/56) ...	0.27	1.55		
18 (81/56) ...	0.54	2.04	Rocks associated with	
19 (77/56) ...	0.52	1.68	metamorphic sedimentary	
20 (70/56) ...	0.02	2.35	rocks in the banded series	
21 (151/56 b) .	0.12	1.14	The charnockite border	
22 (150/56 b) .	0.10	0.97	series	
23 (158/56) ..	0.19	2.25		
24 (142/56) ..	0.27	2.35		
25 (150/56 a) .	0.46	2.11		
26 (151/56 a) .	0.37	2.34		
27 (3/56 b) ...	0.43	2.16	The amphibolite facies	
28 (147/56 b) .	0.36	1.27	veined gneisses	
29 (147/56 a) .	0.26	2.68		
30 (an. 1)	0.12	2.65		
31 (3/56 a) ...	0.35	4.54		
32 (13/56)....	0.41	1.27	Rocks associated with the	
33 (29/55)....	0.67	2.21	metamorphic sedimentary	
34 (33/55 j) ..	0.87	1.22	rocks in the amphibolite	
35 (43/55)....	0.33	0.83	facies veined gneisses	
36 (71/55)....	2.49	1.73	Young red granites	
37 (an. 5)	0.72	8.20		
38 (347/55)...	1.50	1.47		
39 (211/55)...	1.31	2.79		
40 (208/55)...	1.59	6.75		
41 (282/55)...	0.57	14.00		
42 (254/55)...	1.56	9.75		
43 (370/55)...	1.05	9.75		
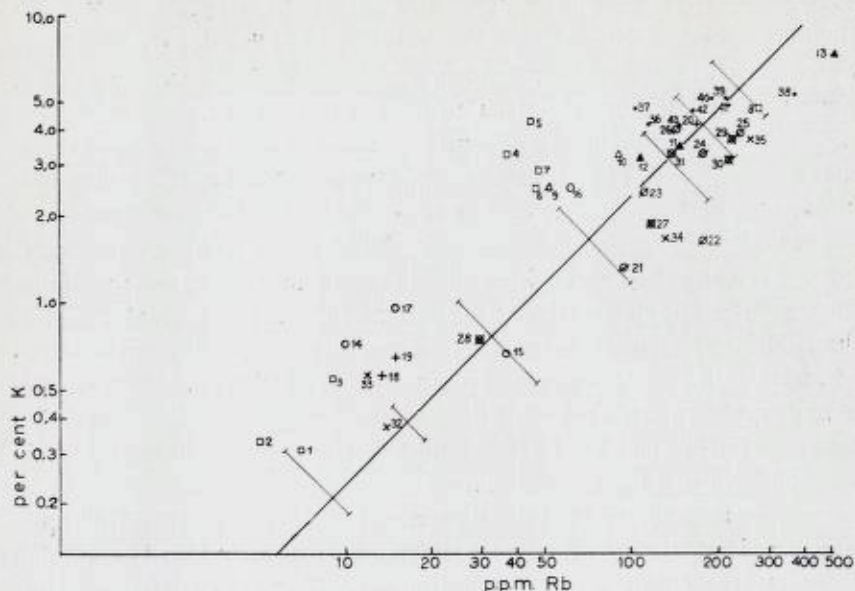
Fig. 38. Relatioship between K and Rb in the Langøy rocks; nos. refer to Table II, symbols as in Fig. 37.

Forholdene mellom K og Rb i bergartene fra Langøy; no. refererer til Table II, symboler som på Fig. 37.

al. (1956) to conform to more recent Rb determinations on G-1 and W-1, which were used as standards by Ahrens. The vertical lines show the scatter limits of Ahrens' original data; most subsequent analyses fall within these; therefore, they probably indicate the limits of "normal" K/Rb ratios. Relative enrichment in Rb is known in large pegmatite bodies (Goldschmidt, 1954, Heier and Taylor, 1959) and in rhyolites and granites which have crystallized from highly fractionated magma (Taylor et al. 1956). This has usually been ascribed to extreme magmatic or metamorphic differentiation which caused the 10 per cent difference in ionic radius between K^+ and Rb^+ to become effective. Depletion of Rb has been noted by Taylor and Heier (1958 b) in some alkali feldspars of the "mesoperthite" type. It has also been noted in some syenitic rocks (Ahrens, Upton, personal communications).

Fig. 38 shows that the gabbros, monzonites and two of the dykes from the Eidet—Hovden intrusion have high K/Rb ratios. The two

monzonites are lowest in Rb. The only analysed rock from the Eidet—
Hovden intrusion which has a "normal" ratio is the dyke (8—60/56)
which is rich in Si and K and thus may be assumed to have crystallized
from a relatively more fractionated part of the magma.

The porphyroblastic monzonitic granulites and the banded series
rocks, except one, (15—145/56), likewise have high K/Rb ratios. It is
interesting that the one exception (15—145/56) is located close to the
suggested dividing line between the banded series and the charnockite
border series. The charnockite border series and all of the remaining
rock groups are characterized by normal ratios. The normal ratios of
the young red granites agree with the hypothesis that they are products
of anatexis of the charnockite rocks and not highly fractionated rest
magmas. If the K/Rb ratios in Fig. 38 are compared with the variation
diagram for Na in Fig. 37 it is found that the rock groups low in Na
tend to have the highest Rb values.

It was suggested that the variation in Na was dependent upon
the increased ability of potash feldspars to accomodate Na at high
temperatures (high grade metamorphism). Because of its large ionic
radius, Rb will tend to be enriched relative to K in low temperature
potassium minerals. Thus Na and Rb should oppose each other in
potassium minerals.

The K/Rb ratios of the alkali feldspars, of some coexisting biot-
ites and plagioclases, and of the host rocks analysed are plotted in
Fig. 39.

The Fig. 39 shows similar ratios for the rocks, plagioclases, and
potash feldspars, whereas the coexisting biotites are preferentially
enriched in Rb. This may be explained by the higher co-ordination of
the alkali metals in the micas; this would favour the larger Rb ion
(12 co-ordination in micas as compared to 10 in orthoclases (Bragg,
1937)).

However, the feldspars do not show variations of the ratios with
metamorphic grade similar to those of the rocks. Fig. 39 indicates that
the increasing amounts of biotite in the amphibolite facies gneisses
(and the charnockite border series, Figs. 26 and 29) are largely
responsible for the relative enrichment in Rb in the latter groups.

Two of the feldspars from the red augengneisses (109/56,
99/56) and one from the amphibolite facies veined gneisses (an. 1)
are conspiciously low in Rb. In the latter case the rock, the plagioclase,
and the biotite were analysed as well. These have normal ratios al-
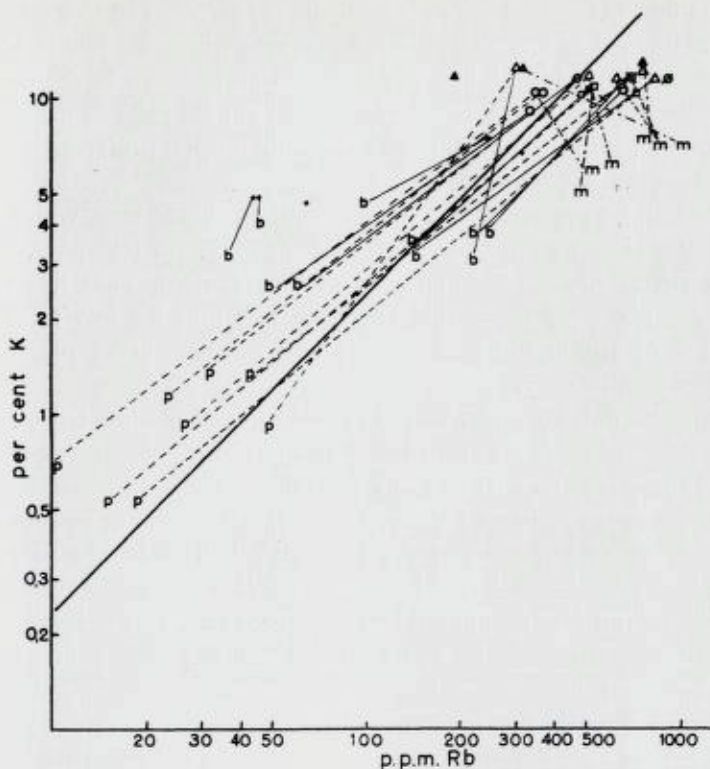
Fig. 39. Relationship between K and Rb in alkali feldspars (symbols as in Fig. 35), plagioclases (p), biotites (m), and some enclosing rocks (b). Tielines connect coexisting minerals and/or rocks.

Forholdene mellom K og Rb i alkali feltspater (symboler som på Fig. 35), plagioklaser (p), biotit (m) og en del bergarter (b). Forbindelseslinjer er trukket mellom sammenhørende mineraler og/eller bergarter.

though the biotite may be considered to be somewhat enriched in Rb. The abnormal slopes of the tie lines between the potash feldspar and the other phases (Fig. 39) would seem to indicate that the analysis of the former is wrong. The calculated mode (mesonorm) based on chemical analysis of this rock is given in Table 10. It contains 10.7 per cent Or. (K/Rb, 408), 45.5 per cent Plag. (K/Rb, 162), and 18.1 per cent Mica (K/Rb, 78). These are the major K and Rb carrying minerals in the rock and considered equal to 100 per cent give a K/Rb ratio of 177 of the total rock which is close to the actual ratio (143) found; this

suggests that the potash feldspar analysis is correct. The two feldspars from the low grade metamorphic red gneisses which are low in Rb are from rocks rich in biotite (18 per cent and 10 per cent in the counted mode respectively for nos. 109/56 and 99/56 in Table 18) so the relative depletion of Rb in these is probably dependent upon preferred entry of the element into the biotite lattice.

It might be expected that the coexisting plagioclases, which have less favourable structures for Rb than the potassium feldspars, would have relatively lower concentrations of this element. That this is not the case is probably dependent upon the fact that the low total concentration of Rb in the plagioclases will not affect the plagioclase structures significantly.

This can not, however, be the explanation of the low Rb contents of the feldspars from the monzonites within the Eidet—Hovden intrusion. The low Rb contents in these feldspars have been discussed in a previous paper (Taylor and Heier, 1958 b); it appears to be related to the unfavourable structure of the Na rich alkali feldspars so far as accomodating the large Rb ion is concerned. (It was concluded that the Rb depletion in these cases could be explained only on the basis of the mesoperthitic structures being formed by unmixing from a once homogeneous phase.)

One of the Eidet—Hovden monzonite feldspars (179/55) has a normal K/Rb ratio, although mesoperthite is the only feldspar phase. This is, by far, the most potassium rich of the feldspars and in Fig. 21 can be seen to plot rather close to the cotectic curve of Yoder et al. (1957). If these rocks crystallized under lower pressures than the 5000 bars, for which this curve was constructed, its position may actually be on the cotectic curve.

The author has previously published results of trace element analyses of potash feldspars from the pre-Cambrian basement complexes of southern Norway. The average K/Rb ratio of the K-feldspars from the gneisses and granites of that area (nos. 6—25, Table 1, Heier and Taylor, 1959) is 245 as compared to 237 (excluding nos. 50—53 and 75, Table III) for the Langøy rocks.

Cesium (Cs^+ 1.65 Å).

Like Rb, Cs replaces K in potassium minerals, but because of its larger size it becomes much more enriched in the late crystallized minerals.

The detection limit of Cs by the spectrographic method used is

about 1 p.p.m. Cs. Cs was not registered for any of the rocks, plagioclases or potash feldspars from Langøy; the concentration of Cs in these compounds is less than 1 p.p.m.

According to Sahama (1945), Cs is absent in rocks of the basement complexes of southern Lapland, Finland. Rankama and Sahama (1950, p. 436) state that the pre-Cambrian rocks of Finland are poorer in Cs than the metamorphic rocks of younger age. The potash feldspars from the basement complex of southern Norway (except those from some of the large pegmatites) likewise tend to be low is Cs (Heier and Taylor, 1959, Table I, nos. 1—49).

This along with the data from Langøy, might be considered to show that Cs has a regional temporal distribution pattern. However, it appears more likely to the author to depend upon the fact that Cs probably is not stable in potassium minerals under severe regional metamorphism conditions. Cs was found to be present in all the analysed samples of biotite from the Langøy rocks. That Cs prefers mica over potash feldspars is probably dependent upon the larger co-ordination of the alkali metals in the biotite. According to Rankama and Sahama (1950, p. 439) biotite from a "pegmatitic granite" from Finland contains 30 p.p.m. Cs_2O ($\sim$ 27 p.p.m. Cs) and in Table 12. 7 (ibid) they gave the average Cs content in biotites from pegmatites as 0.30 per cent Cs_2O.

Table V shows Cs content in biotites from the porphyroblastic monzonitic granulites (293/55) to be 7.2 p.p.m., in those from the retrograde metamorphic gneisses (224/55) 5.7 p.p.m., and in those from the charnockite border series (25/55, 150/56) 7.6 p.p.m. Biotites from three rocks of the amphibolite facies gneisses (147/56, 37/55, an. 1) were found to have between 17 and 27 p.p.m. Cs. The K/Cs ratios are 8470—9700 for biotites in the granulite facies rocks and 2500—4350 for biotites in the amphibolite facies rocks. It is suggested that this difference reflects the different PT conditions under which the rocks recrystallized. It appears that Cs would be a good indicator of geological conditions. Before it can be satisfactorily used, however, the detection limit must be extended to at least 0.1 p. p. m.

Lead (Pb^{2+} 1.32 Å).

Wedepohl (1956, Table 12, p. 110) in his recent investigation of the geochemistry of lead, found the average concentration of Pb in ultrabasic to granitic rocks to be in the range 3—20 p.p.m. and the average content in the lithosphere to be 15 p.p.m. Pb.

The Pb in the Langøy rocks ranges from below the detection limit ($< \sim 3$ p.p.m.) to 27 p.p.m. Further no regular variation in the Pb content with changing metamorphic grade was found. Because of similar size Pb^{2+} (1.32 Å) substitutes for K^+ (1.33 Å) in potassium minerals. Wedepohl (1956) showed an average of 27 p.p.m. Pb in potash feldspars from rocks (Hauptkristallisation) and 100 p.p.m. in pegmatite feldspars. Heier and Taylor (1959) found the Pb in potash feldspars (including feldspars from certain large pegmatites) to range from undetected (< 2 p.p.m.) to 160 p.p.m. (in a pegmatite feldspar). Feldspars from the Telemark basement granite and gneiss range from 6—38 p.p.m., averaging 18 p.p.m. Pb.

The potash feldspars from the Langøy rocks have Pb concentrations in the interval 3.2—74 p.p.m. (Table III). The feldspar from the young red granite (no. an. 5) is extremely low in Pb (3.2 p.p.m.) and the one from the old red granite (65/56) is extremely high in Pb (73.8 p.p.m.). Excluding these two extremes and also the feldspars from the Eidet—Hovden intrusion, the average Pb content in the potash feldspars from the metamorphosed gneisses is 13.5 p.p.m. Pb. This is slightly less than the crustal average (15 p.p.m.) and decidedly less than Wedephol's average of potash feldspars (27 p.p.m.).

Average Pb concentrations and K/Pb ratios in the feldspars from the different rock series on Langøy are given in Table 44.

Because of the similar ionic radius and higher ionic charge Pb^{2+} would be expected to be "captured" and concentrated in the early (high temperature) potassium minerals (Goldschmidt, 1954, p. 399). However, the data presented by Wedepohl (1956) and Heier and Taylor (1959) show that Pb tends to be enriched in the most fractionated (low temperature) potash feldspars. This contradiction of Goldschmidt's rule is believed to be related to the more covalent nature of the Pb-O bonds as compared to the K-O bonds (Ringwood, 1955). (Wedepohl, 1956, p. 137, found concentrations in rocks to increase with increasing metamorphic grade, i. e., with increasing temperature. However, the present author believes that as metamorphic grade is defined by changes in mineral parageneses this enrichment of Pb is not a proof of Goldschmidt's rules with respect to Pb.)

Wedepohl (1956) found 23 p.p.m. Pb as an average in biotites from metamorphic rocks (as compared to 34 p.p.m. in igneous rocks). He also found (p. 135) that the Pb content in biotites increases with increasing metamorphic grade from 1.9 (19 ?) p.p.m. in the epidote amphibolite facies to 32 p.p.m. Pb in granulites.

Table 44.

*Average Pb concentrations and K/Pb ratios of potash feldspars
from the various rock series on Langøy.*

	p. p. m. Pb	K/Pb
Eidet–Hovden monzonites (50—53)	8.5	6362
Large pegmatite (54)	10.0	10640
Porphyroblastic monzonitic granulites (55—58)	9.4	11946
Retrograde metamorphosed gneisses (59—61).......	12.0	8747
The banded series (62—63	13.3	9308
Old red granite (64).............................	(73.8)	(1350)
The charnockite border series (65—66)	22.5	4994
Amphibolite facies veined gneisses (67—71)........	16.4	7542
Red augengneisses (72—74)......................	9.9	13258
Young red granites (75)	(3.2)	(23407)

His suggested high Pb content from granulites is unlike the very
low Pb concentrations in the biotites from the granulite and amphi-
bolite facies rocks on Langøy in which Pb was below the detection
limit in all samples. On the other hand these low values do agree with
the data presented by Howie (Table VII, p. 752, 1955) who found no
detectable Pb in the biotites from the Madras charnockites.

The quartz-plagioclase-biotite schist on Langøy also has no
detectable Pb (43/55).

Wedepohl (1956, p. 88) estimated the average Pb content in
plagioclases as 10 p.p.m., and found no regular variation with anorth-
ite content. Howie (1955, Table V) found Pb to vary between 10—30
p.p.m. in the plagioclases from the Madras charnockites (average
20 p.p.m. for a total of 6 samples). In two out of three cases where the
coexisting potash feldspars were also analysed he found the higher
Pb concentration in the plagioclases.

The nine plagioclases from the Langøy rocks range in Pb from
21 to 52 p.p.m., average 32 p.p.m. (recalculated to 100 per cent feld-
spar). All plagioclases were found to be higher in lead than the co-
existing potash feldspars (Table 45).

That Pb is relatively enriched in the plagioclases is surprising and
contrary to most common tenets. The 1.32 Å radius for Pb^{2+} is Gold-
schmidt's value. Using the radii recomended by Ahrens (1952) the
difference between the radii of Pb^{2+} (1.20 Å) and Ca^{2+} (0.99 Å) and
between Pb^{2+} and K^+ (1.33 Å) are 17.5 per cent and 10 per cent
respectively. On the basis of these it is reasonable to assume that Pb^{2+}
will distribute itself between potassium and calcium minerals in some-

Table 45.

*Ratios of Pb in plagioclase
to Pb in potash feldspar.*

Sample nos.	Pb in plag./Pb in Or
(293/55)	26.4/10.9 = 2.4
(251/55)	21/9.9 = 2.1
(238/55)	32/11.4 = 2.8
(224/55)	25/11.4 = 2.2
(an. 12)	39.9/9.6 = 4.2
(25/55)	37/21.1 = 1.6
(37/55)	26/14.1 = 1.8
(an. 1)	52/20.5 = 2.5

what similar proportions. Actually the properties of Pb as compared to K and Ca might make the distribution coefficient of Pb between coexisting plagioclases and potash feldspars particularly sensitive to temperature variations. Too few data are available, however, to ascertain this although it could be interpreted from Table 45 that the distribution coefficient between Pb in plagioclases: Pb in potash feldspars tends to be low in the charnockite border series and amphibolite facies gneisses (one of the amphibolite facies gneisses has a high ratio) but high in the granulite facies rocks.

It has long been known that Pb^{2+} replaces Ca^{2+} in minerals. Rankama and Sahama (1950, p. 733) listed 50 p.p.m. Pb in apatite, 100 p.p.m. in epidote, and 50 p.p.m. in fluorite. Aragonite may contain up to 18 % of $PbCO_3$ (ibid.).

The general belief that Pb is enriched in potash feldspars relative to plagioclases in rocks (though also partly sustained by Wedepohl's data) may partly be due to an erroneous deduction of the fact that Pb is enriched in granitic rocks (rich in potash feldspars). However, because of its ionic properties Pb will not be camouflaged in feldspars, but will be enriched by fractionated crystallization in the liquid and concentrated in the most fractionated compounds. Thus though Pb is enriched in granitic (most fractionated) rocks this is no indication of its being concentrated in potash feldspar relative to coexisting plagioclases.

However, that Pb may be relatively enriched in the plagioclase phase even in highly fractionated systems (pegmatites) has been shown by Heier and Taylor (unpublished data). The data are reproduced here in Table 46. Only in one case was the potash feldspar found to be richer in Pb than the coexisting plagioclase.

Table 46.

*Some elements in potash feldspars and plagioclases from pegmatites
from southern Norway (Heier and Taylor, previously unpublished data).*

		Li p.p.m.	Na %	K %	Rb p.p.m.	Cs p.p.m.	Pb p.p.m.	Tl p.p.m.
1a,	potash feldspar perthite, Snarum....	0.76	2.71	9.55	344	8.9	4.4	2.4
1b,	purified potash feldspar lamellae from							
	1 a	0.66	1.07	11.83	450	12.0	4.5	3.4
1c,	purified albite lammellae from 1 a...	1.66	6.53	2.99	110	n.d.	10.0	n.d.
1d,	coexisting oligoclase (An$_{11}$)	1.60	7.57	0.58	18	n.d.	6.5	1.0
2a,	potash feldspar from Ås, Evje	1.00	1.97	10.13	280	n.d.	42.0	1.91
2b,	plagioclase « » »	5.80	6.38	0.83	(1—2)	n.d.	105.0	n.d.
3a,	potash feldspar, Mannekleiv, Haver-							
	stad, Iveland.....................	n.d.	1.14	11.54	760	18.0	43.0	2.57
3b,	plagioclase Mannekleiv, Haverstad,							
	Iveland	4.7	7.57	0.79	40	n.d.	20.0	n.d.
4a,	potash feldspar, Varoddbrua Kristian-							
	sand.	0.8	1.82	10.58	825	36	7.0	12
4b,	plagioclase Varoddbrua, Kristiansand.	6.1	7.76	0.23	41	n.d.	26.0	n.d.

That the enrichment of Pb in the plagioclase phase in the Langøy
rocks is not due to analytical errors becomes evident when analyses
of host rocks and potash feldspars are compared (Table 47).

In all cases except one (150/56 a) the Pb concentration is higher
in the rocks than in the potash feldspars and the K/Pb ratio is lower
in all the rocks; this demands that Pb must be concentrated in some
of the other minerals.

Thallium (Tl+ 1.49 Å).

Shaw (1957) has reviewed the geochemistry of Tl and given its
crustal abundance as 1.3 p.p.m.

Tl$^+$ (1.49 Å) substitutes for K$^+$ (1.33 Å) in potassium minerals
and its geochemical behaviour in the lithosphere is similar to that of
Rb$^+$ (1.49 Å).

Tl was not determined in the rocks and biotites from Langøy
because of possible interference of Ni (3775.6) with the suitable
Tl (3775.7) line).

Tl was not detected in the plagioclases from the Langøy rocks.
This is not surprising considering the coherence between Tl and K.
Ahrens (1948) found up to 3 p.p.m. Tl in albites but that in 10 out of

Table 47.

Concentrations of K (per cent) and Pb (p.p.m.)
in host rocks, potash feldspars and plagioclases
on Langøy.

No.		rock	potash feldspar	plagioclase
an. 9	K	4.27	4.82	
	Pb	11.7	6.8	
57/56	K	2.46	9.13	
	Pb	19.4	7.6	
318/55	K	3.48	10.09	
	Pb	16.0	13.4	
an. 12	K	2.49	11.42	1.39
	Pb	17.1	9.6	39.9
150/56a	K	3.93	10.99	
	Pb	21.4	23.90	
147/56a	K	3.69	12.14	
	Pb	25.6	19.4	
3/56a	K	3.29	11.62	
	Pb	25.2	16.8	
an. 1	K	3.15	12.48	0.95
	Pb	26.9	20.5	52.0
an. 5	K	4.77	7.49	
	Pb	9.9	3.2	

13 analysed plagioclases Tl was not detected. Table 46 also shows Tl to be absent in most pegmatitic plagioclases from southern Norway. Some data show Tl to be enriched in extremely fractionated plagioclases, e. g., cleavelandite (Ahrens, 1948).

The potash feldspars from granites analysed by Ahrens (1948) average 0.8 p.p.m. Tl (based on a total of three samples). In the potash feldspars from the Langøy rocks Tl was not detected in the feldspars of nos. 145/55, an. 9, from the Eidet—Hovden mozonites which are

Table 48.

Rb/Tl ratios of potash feldspars from the Langøy rocks.

No.	Rb/Tl	Average Rb/Tl ratio	Rock type
54 (an. 7)...	865	865	pegmatite
56 (293/55) .	389		porphyroblastic monzonitic
57 (255/55) .	836	545	granulites
58 (251/55) .	411		
59 (318/55) .	548		retrograde metamorphosed
60 (238/55) .	734	605	gneisses
61 (224/55) .	532		
62 (an. 12)..	594	757	the banded series
63 (348/55) .	920		
65 (25/55) ..	875	799	the charnockite border series
66 (150/56) .	722		
67 (147/56) .	870		the amphibolite facies
68 (3/56) ...	1037		veined gneisses
69 (35/55) ..	910	781	
70 (37/55) ..	576		
71 (an. 1)...	510		
72 (109/56) .	400	314	red augengneisses
73 (99/56) ..	228		
74 (E-vatn) .	1107		red granitic vein
75 (an. 5)...	245	245	young red granite

low in K and depleted in Rb. In no. 179/55 which is the most K rich of the Eidet—Hovden monzonite feldspars and which has a normal K/Rb ratio, Tl was found to be present as traces. It is also present as traces in one of the potash feldspars of the porphyroblastic monzonitic granulites (271/55, contaminated with plagioclase). In the feldspar from the old red granite (65/56) Tl was not determined because of irregularities in the spectrographic plate, and the fact that more material was not available.

The rest of the feldspars range between 0.6—1.0 p.p.m. in Tl. The arithmetic mean is 0.8 p.p.m. which is identical to Ahrens (1948) average for potash feldspars from granitic rocks. The Rb/Tl ratios of the potash feldspars (with detectable Tl) from the Langøy rocks are listed in Table 48. Ahrens (1948) found a close geochemical coherence between Rb^+ and Tl^+ and the mean value of the Rb/Tl ratio to be 100

(Ahrens, 1948, p. 580). If this is taken as the "normal" ratio, the potash feldspars from Langøy are all low in Tl relative to Rb. The average Rb/Tl ratios of the feldspars from the various rock types on Langøy are given in Table 48. The porphyroblastic monzonitic granulites and the retrograde metamorphic gneisses (which, for a large part, represent originally identical rocks) have similar ratios. A second group of similar ratios occurs in the banded series, the charnockite border series, and the amphibolite facies veined gneisses. (It has previously been stressed that these three rock series are closely related, and that the difference between them is that of varying metamorphic grade.)

It is interesting to note that the two feldspars from the red augen gneisses (109/56, 99/56) have low and more "normal" ratios. These feldspars have Tl concentrations close to the average; the low ratios are due to their being low in Rb.

A close to normal ratio (100) is also shown by the feldspar from the young red granite an. 5.

The feldspar from E-vatn (74) should not be compared with the others in this group because it is from a relatively young granitic vein. The late stage emplacement of this vein is indicated by the low temperature of formation (300° C) and high triclinicity ($\Delta = 1.01$, $2V_x = 84°$) shown in Table 39. This feldspar has a close to normal Rb content but is low in Tl and has a maximum Rb/Tl ratio value.

The data presented are much too scanty to warrant serious discussion of the variations of the Rb/Tl ratio with metamorphic grade. Shaw (1957, p. 201) found no relation of Tl content in rocks to metamorphic grade. The Rb/Tl ratios and the absolute Tl contents are plotted against the temperature of formation of the potash feldspars in Fig. 40.

Though there is a considerable spread of the plots the general tendency is for low Rb/Tl ratios and high Tl contents to be associated with a high temperature of formation.

Increase of the Rb/Tl ratio of potash feldspars with decreasing temperatures (when decreasing temperatures are taken to define amount of fractionation) is apparent from Ahrens (1948, Table 2, pp. 581—582). In the series granite-pegmatite-amazonite feldspars he finds the arithmetic mean of the Rb_2O/Tl_2O ratio to be 85-178-183 respectively. The tendency of pegmatitic feldspars to be richer in Tl than feldspars from other rocks can, however, not be disputed (Ahrens, 1948, Table 2, Heier and Taylor, 1959, Table 1). Pegmatites are believed to represent relatively low temperatures of formation (most fractionated systems)
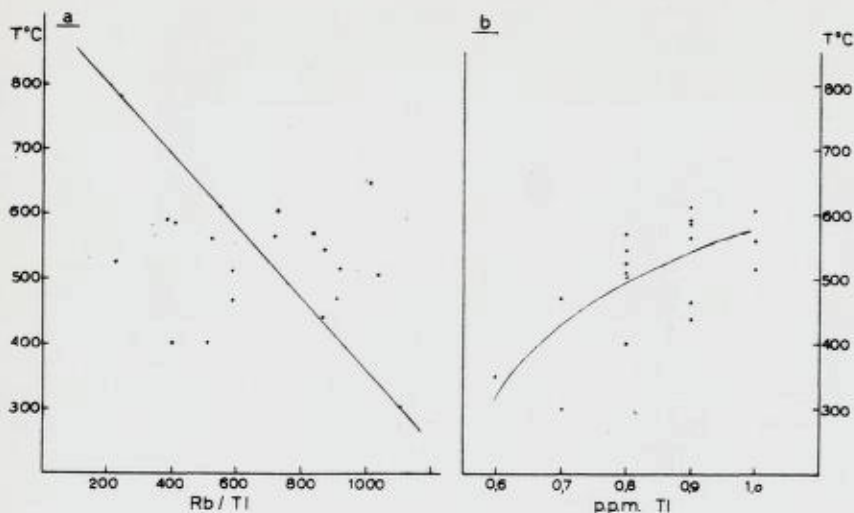
Fig. 40. Plots of Rb/Tl ratios (a) and absolute Tl content (b) of alkali feld-
spars against their temperature of formation (Table 39).

*Rb/Tl forholdene (a) og absolutte Tl verdier (b) i alkali feltspatene plottet mot
dannelsestemperaturer (Tabell 39).*

and their being rich in Tl would seem to contradict the above state-
ment (Fig. 40). It should be remembered, however, that Fig. 40 is
based on feldspars from rocks and that pegmatite samples are not
included. The system from which the pegmatites crystallize may be
abnormally high in Tl but so far as pegmatites of regionally metamor-
phosed areas are concerned, a much lower temperature for the forma-
tion of these than for the country rocks is commonly difficult to
conceive. Barth (1956) suggests even a higher temperature of forma-
tion for large pegmatite bodies than for the nearby gneisses and
granites. Thus, although the most fractionated products in igneous
processes are probably formed at the lowest temperatures, it appears
that increasing fractionation should not be identified with decreasing
temperatures in metamorphic areas.

Lithium (Li+ 0.78 Å).

The crustal abundance of Li is taken as 60 p.p.m. (Rankama and
Sahama, 1950, p. 422). The average of the Langøy rocks is consider-
ably less than this (13 p.p.m. Li), close to the average of andesitic

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Fig. 41. Relationship between Li and Mg in the Langøy rocks (symbols as in Fig. 37).

Forholdet mellom Li og Mg i bergartene fra Langøy (symboler som på Fig. 37).

rocks (15.2 p.p.m.). The maximum value found is 30 p.p.m. Li in a quartz-plagioclase-biotite schist (43/55). This average is close to the mean of the Madras charnockites (14 p.p.m. Li) as calculated from Howie, 1955, Table II, p. 737 (concentration range 5 to 25 p.p.m. Li).

The geochemical behaviour of Li departs markedly from that of the other alkali metals in that it tends to be concentrated in the femic silicate minerals, where, because of its small ionic radius it substitutes for Mg^{2+} (0.78 Å). There is, however, apparantly no geochemical coherence between Li and Mg in rock series ranging from basic to acid. Fig. 41 shows plots of Li versus Mg in the various rock types from Langøy. The figure indicates that the Mg/Li ratio decreases from basic

to acid rocks. The decrease is fairly regular down to a concentration of about 0.8 per cent Mg and below this the ratio decreases at a much higher rate. In fact, in the Mg poor rocks the total amount of Li is approximately constant whereas Mg may vary by a factor of 10. This agrees with the long known fact that Li may be enriched in the femic silicate minerals in granitic rocks (e. g., biotites). The reason why the most silica rich rocks in Fig. 41 tend to be low in Li (less than in the intermediate rocks) appears to be that these (especially the young red granites) are low in biotite and do not contain any other femic minerals into which Li can be readily incorporated. In fact, most of the Li in these granites is probably in the feldspars.

According to available data (Rankama and Sahama, 1950, Howie, 1955) the minerals orthopyroxene-clinopyroxene-amphibole-mica form a series in which there is a regular increase of Li concentration. That Li is concentrated in the biotites from the Langøy rocks is obivous from Table V. Li in the biotites is higher than in the enclosing rocks by a factor of 2 to 3. There also is a marked difference in the Li concentration in biotites from granulite facies rocks (293/55, 224/55, 25/55, 150/56 — 40 to 50 p.p.m.Li) from that in the biotites of the amphibolite facies gneisses (147/56, 37/55, an. 1—66 to 75 p.p.m. Li). The quartz-plagioclase-biotite schist (no. 43/55, Table II) contains 30 p.p.m. Li and consists of 50 per cent biotite (Table 14). Assuming the bulk of the Li content in this rock to be in biotite this mineral contains about 60 p.p.m. Li, which is close to that from the biotites of the amphibolite facies gneisses within which the rock occurs.

Because the biotites from the amphibolite facies gneisses are markedly higher in Li than those from granulite facies rocks and because this mineral is relatively much more abundant in the first group (Figs. 26 and 29) it might be expected that the rocks within the former would be higher in Li than those within the latter. However, in the Mg/Li diagram (Fig. 41) and in the diagram where Li contents are plotted against the "position" of the rocks (Fig. 42) this tendency cannot be seen to be very marked. From Fig. 42 it can be seen that the amphibolite facies gneisses and the rocks associated with them tend to be slightly richer in Li than the average of the other rocks. (43/55 is markedly richer in Li but this is a plot of the rather special quartz-plagioclase-biotite schist).

There obviously must be modifying factors. The data in Fig. 26 show that the only additional femic mineral which is stable
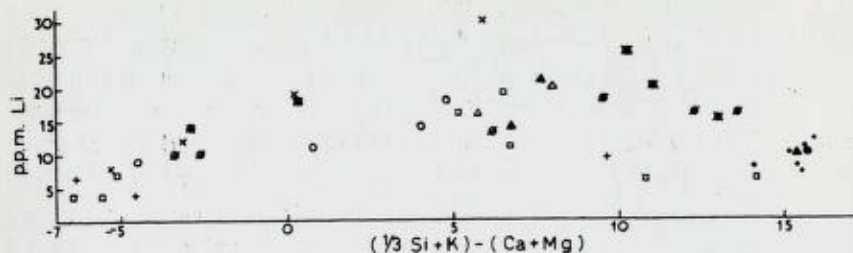
Fig. 42. Li in the Langøy rocks plotted against (1/3Si+K)-(Ca+Mg), (symbols as in Fig. 37).

Li i bergartene fra Langøy plottet mot (1/3Si+K)-(Ca+Mg), (symboler som på Fig. 37).

from the basic to the acid end in the granulite facies rocks is hypersthene. According to Howie (1955, Table IX, p. 755) the content of Li in the orthopyroxenes is negligible (2—7 p.p.m.).

It seems that in the granulite facies rocks the bulk of the Li present must be in the feldspar lattices. A maximum concentration of 17.6 p.p.m. Li was found in the plagioclases (Table IV, recalculated to 100 per cent plagioclase). The data of Table 49 show Li in the plagioclase from the granulite facies area to be very close to the average value of all of the rocks and in one case (the banded series) to be higher in the former. The plagioclases from the amphibolite facies rocks have considerably less Li, so within these the biotite must contain most of the Li present. There is ample reason to believe that plagioclase is the most important Li carrying mineral in the granulite facies rocks (except, of course for the biotite which is typically present in very small amounts). In the Eidet—Hovden monzonites the feldspar phase (mesoperthite) must likewise contain most of the Li present.

That the potash feldspars are lower in Li than the plagioclases appears to be a regular tendency (see Table 46). Lundgårdh (1947) suggested that Li^+ is able to replace Al^{3+} in the plagioclase structure. This is rather hard to conceive because of both the large difference in ionic radius between the two, and the difference in electrostatic charge. It appears more probable that Li^+ substitutes for Na^+ (0.98 Å) in the feldspars. That the plagioclases are higher in Li than the potash feldspars supports this; also corroborative is the fact that the Li content of the potash feldspars increases with increasing metamorphic grade; *i. e.*, with increasing Na content in the feldspars.

Table 49.

Average Li contents (p. p. m.) of plagioclases and potash feldspars in the rocks from Langøy.

Rock type	rocks	plagioclase	potash feldspar
Eidet–Hovden monzonites	8.5	17.4	
Porphyroblastic monzonitic granulites ..	18	14.6	9.8
Retrograde metamorphosed gneisses....	14.5	8.2	4.9
The banded series	13	17.5	8.9
The charnockite border series	14	10.9	4.9
Amphibolite facies gneisses	20	6.0	4.4*

* The average is actually less than 4.4 because no. 71 (an. 1) with only traces of Li is not included.

However, although the potash feldspars are lower in Li than the plagioclases the Li/Na ratio is higher in the potash feldspar. This is surprising because it would be expected that the small Li ion would preferentially substitute for Na where the structure sites tend to be smaller in the plagioclases. The reason for this apparent anomaly may be that only a very restricted amount of Li can enter the feldspar lattices without unstabilizing the structure (the pure Li feldspar is unstable all together, Rankama and Sahama, 1950, Table 5.20, p. 139).

Strontium $(Sr^{2+}\ 1.27\ Å)$.

The crustal abundance of Sr is 450 p.p.m. (Turekian and Kulp, 1956). Sr accompanies Ca $(Ca^{2+}\ 1.06\ Å)$ in minerals and rocks of both igneous and sedimentary origin. Turekian and Kulp (1956) found, however, that the Sr content in basaltic rocks is essentially independent of the Ca content although in granitic rocks a definite relation exists between them.

In Fig. 43 Ca versus Sr is plotted for the various rocks from Langøy. It can be seen that the basic rocks (high in Ca) show a considerable spread of plots whereas the intermediate and granitic rocks have ratios which are much more nearly uniform (two of the young red granites (nos. 282/55 and 254/55) are extremely low in Sr; also note these under Ba).

There is no apparent relationship between the Sr/Ca ratio and metamorphic grade. The two retrograde metamorphic gneisses (318/55, 310/55) which are chemically similar to the porphyoblastic monzonites have relatively high ratios.
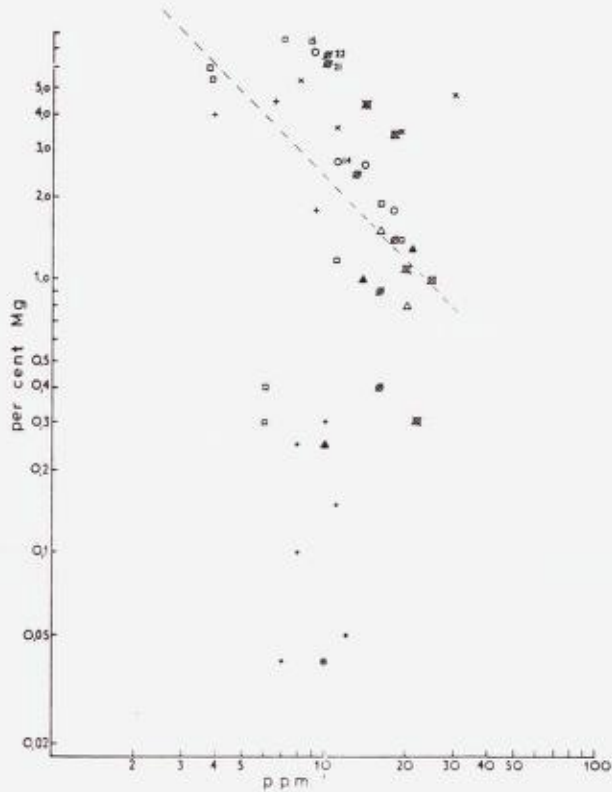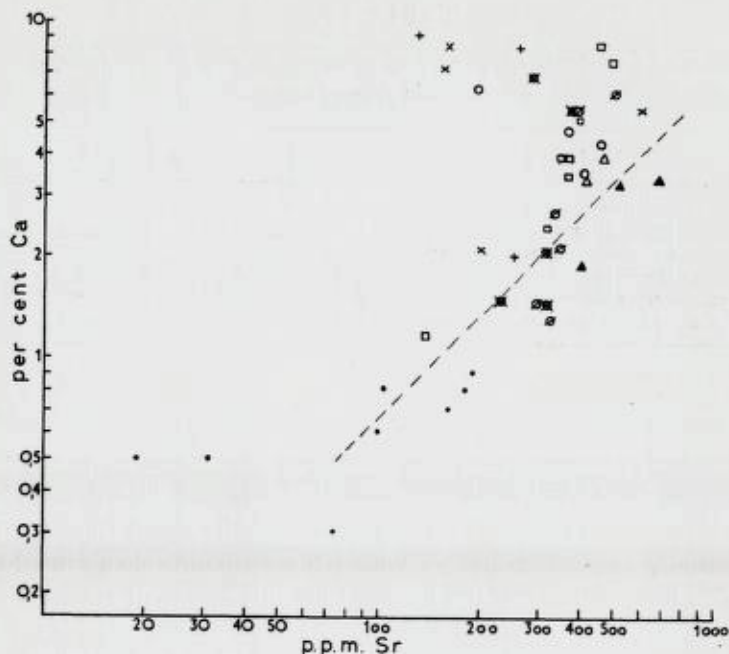
Fig. 43. Relationship between Sr and Ca in the Langøy rocks (symbols as in Fig. 37).

Forholdene mellom Sr og Ca i bergartene fra Langøy (symboler som på Fig.37).

Turekian and Kulp (1956, p. 290) say that it is not probable that ortho-amphibolites and para-amphibolites can be differentiated on the basis of their Sr/Ca ratios. It is interesting that four out of five Langøy amphibolites directly associated with undoubted sedimentary rocks (crosses, Fig. 43) have low ratios. Similar low ratios were obtained for no. 145/56 from the banded series and no. 147/56 b which is an amphibolitic band in the amphibolite facies gneisses. According to Goldschmidt (1954, p. 246) the Sr/Ca ratio increases in the mineral series:

Early apatite < pyroxene, amphibole < labradorite
Sr/Ca   0.0002   0.0002—0.003   0.002—0.005
                  < oligoclase < orthoclase
                   0.007   0.08

Sr, thus, enters the feldspars preferentially. The four amphibolites with low ratios have between 32—43 mol per cent plagioclase (Table

15). The one of them with a normal ratio (33/55 j) has 62 per cent plagioclase. The two gneissic rocks, (145/56 and 147/56 b), have 46 per cent and 42 per cent plagioclase, respectively. Potash feldspar is absent in all of them. Thus the observed ratio variations in the basic rocks are probably related to compositional differences rather than to different origins.

The tendency of the basic rocks to have lower Sr/Ca ratios than the intermediate and acid rocks appears to depend upon the fact that the sum of the feldspars tends to be higher in the latter.

The potash feldspars from Langøy (except those from the Eidet —Hovden intrusion and the young red granite) average 990 p.p.m. Sr. The total Sr content in the potash feldspars is higher than in the plagioclases and the Sr/Ca ratio is several times higher in the potash feldspars than in the plagioclases.

Because of very similar ionic radii Sr^{2+} (1.27 Å) substitutes readily for K^+ (1.33 Å) in potash feldspars. Because Sr does readily substitute for both Ca and K in minerals it is obvious that no strong geochemical coherence will obtain between Sr and either Ca or K in rocks.

The data of Table III show that potash feldspars from the amphibolite facies rocks tend to be lower in Sr (average 720 p.p.m. of groups 8 and 9) than those of the granulite facies rocks (average 1240 p.p.m. of groups 3, 5, 6, 7). This is in accord with Goldschmidt's rule of capturing. Sr^{2+} with the higher valency will tend to be concentrated in the early formed (*i. e.*, high temperature) potash feldspars.

The potash feldspars from two of the retrograde metamorphic gneisses (318/55 and 238/55) have a Sr content similar to that of the amphibolite facies gneisses whereas one (224/55) is very similar in this respect to those from the porphyroblastic monzonitic granulites. It is interesting that this last rock is much less altered than the other two and that it contains relics of hypersthene (p. 149). The Sr content of the potash feldspar from the old red granite (65/56) is less than the average of the feldspars from the granulite facies rocks. The Eidet —Hovden monzonite feldspars are low in Sr (average 275 p.p.m.), but their K/Sr ratios are similar to those of the western monzonites, probably indicating their high temperature of formation. The pegmatite in the Eidet—Hovden intrusion an. 7 and the young red granite an. 5 have feldspars which are especially low in Sr (90 and 130 p.p.m. respectively). The feldspar geological thermometer indicates a high

temperature of formation for the granite (higher than for any of the metamorphosed rocks) and if the pegmatite represents a "rest magma" from the monzonite crystallization (which appears very likely from its position in the field) it probably also crystallized at a fairly high temperature. Thus the low Sr values of these two rocks may appear to contradict the above generalizations. However, both the pegmatite and the young granites are highly fractionated in the sense that they crystallized from systems enriched in volatile elements. This, therefore, is an example illustrating that in metamorphism a fractionated system does not necessarily represent low temperatures. Plots of per cent Ca versus p.p.m. Sr in potash feldspars, plagioclases and rocks from Langøy are given in Fig. 44 a. The plagioclases and rocks have similar Ca/Sr ratios and these are much higher than those of the potash feldspars. The Ca/Sr ratios are higher in the plagioclases than in the potash feldspars by a factor of about 10. This agrees with Goldschmidt's data cited above (though the value of the ratios is different). The meso-perthites from the Eidet—Hovden intrusion have ratios close to those of the plagioclases and rocks. This indicates the plagioclase like struc-ture of these minerals (Taylor and Heier, 1958 b). The slope of the tie lines joining the plagioclases and the coexisting potash feldspars are shown (Table 50) and plotted against the temperature of formation (Fig. 44 b).

The retrograde metamorphic gneiss (238/55) is not included on Fig. 44 b because it plots far off the dotted curve. By using the curve to determine under which temperature the retrograde metamorphism took place a temperature of 450° C is obtained. The other retrograde metamorphic gneiss, 224/55, plots close to the curve, however. It is interesting to note that whereas the former (238/55) is completely recrystallized and exhibts a stable amphibolite facies mineral para-genesis, the latter (224/55) is much less altered and still contains hypersthene and diopside (p. 149). Considered in conjunction with the way in which the Ca and Sr contents in the plagioclases were deter-mined, the curve (Fig. 44 b) appears to indicate that the ratio, $\frac{Ca/Sr \text{ in plagioclase}}{Ca/Sr \text{ in K-feldspar}}$, is a function of temperature of formation of the feldspars.
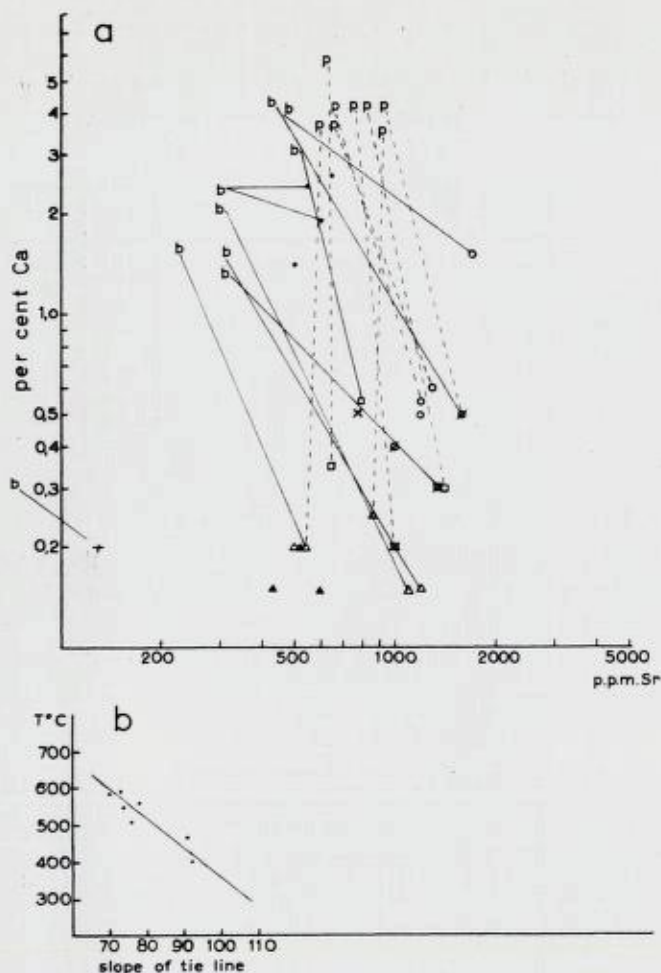
Fig. 44 a. Relationship between Ca and Sr in potash feldspars (symbols as in Fig. 35), plagioclase (p) and rocks (b) from Langøy. Thin lines connect co-existing minerals and/or rocks.

Forholdet mellom Ca og Sr i kalifeltspater (symboler som på Fig. 35), plagioklaser (p) og bergarter (b) fra Langøy. Tynne linjer forbinder sammenhørende mineraler og/eller bergarter.

Fig. 44 b. Slope of tieline between plots of Ca vs./Sr in plagioclase and Ca vs./Sr in potash feldspar plotted vs. temperature of formation of the potash feldspar (Table 39).

Vinkelen til forbindelseslinjen mellom Ca/Sr i plagioklas og Ca/Sr i kalifeltspat plottet mot dannelsestemperaturen av kalifeltspat (Tabell 39).

Table 50.

Slope of tie line between plots of plagioclase and coexisting potash feldspars in Fig. 44a, related to temperature of formation of feldspars (Table 39).

No.	Rock type	Slope	T°C
56 (293/55) .	porphyroblastic monzonitic granulite	107	590
58 (251/55) .	» » »	110	585
60 (238/55) .	retrograde metamorphosed gneiss	91	605
61 (224/55) .	» » »	102	560
62 (an. 12) . .	the banded series	104	510
65 (25/55) . .	the charnockite border series	106	545
70 (37/55) . .	the amphibolite facies veined gneisses	89	465
71 (an. 1). . .	» » » » »	88	400

Barium (Ba²⁺ 1.43 Å).

The crustal abundance of Ba is about 670 p.p.m. (Rankama and Sahama, 1950). Ba is concentrated in potassium minerals in which it replaces K. According to Goldschmidt's rule of capturing, Ba should be concentrated in the early formed (high temperature) potassium minerals. From the K/Ba ratios of the alkali feldspars (Table 51), it

Table 51.

K/Ba ratios of potash feldspars from Langøy.

No	K/Ba	No	K/Ba	No	K/Ba
50 (145/55) . .	12.20	59 (318/55)	28.83	67 (147/56)	15.18
51 (an. 9). . . .	16.07	60 (238/55)	41.36	68 (3/56)	17.60
52 (an. 9). . . .	15.60	61 (224/55)	10.99	69 (35/55)	31.58
53 (179/55) . .	19.00			70 (37/55)	14.77
		62 (an. 12)	11.42	71 (an. 1)	29.02
		63 (348/55)	22.98		
54 (an. 7). . . .	118.22				
55 (271/55) . .	8.30	64 (65/56)	24.90	72 (109/56)	10.52
56 (293/55) . .	10.59			73 (99/56)	39.17
57 (255/55) . .	10.61	65 (25/55)	18.95	74 (E-vatn)	66.42
		66 (150/56)	15.70		
58 (251/55) . .	10.29				
				75 (an. 5)	83.22

can be seen that the ratio tends to increase (decreasing Ba content) with decreasing metamorphic grade. When the K/Ba ratios are plotted against the temperature of formation of the feldspars (Fig. 45 a and b) it can be seen that most of the granulite facies feldspars (western mon-
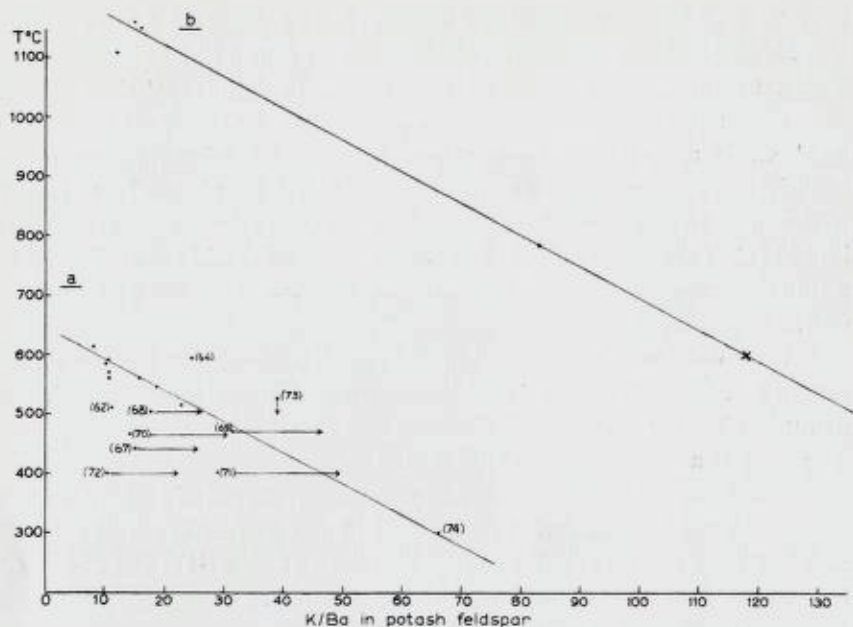
Fig. 45 a. Plots of the K/Ba ratios of potash feldspars from the metamorphic (a) and igneous (b) rocks on Langøy vs. the temperature of formation of the potash feldspars (Table 39); explanation in text.

K/Ba forholdene i kalifeltspatene fra de metamorfe (a) og eruptive (b) bergartene på Langøy plottet mot kalifeltspatenes dannelsestemperatur (Tabell 39); forklaring i teksten.

zonites, the banded series, and the charnockite border series) plot on a fairly straight line. The plot of the feldspar from the least altered retrograde metamorphic gneiss (61—224/55) also falls on this curve. The two others in this group (59—318/55 and 60—238/55) have ratios corresponding to a lower temperature just as was indicated by their Sr contents. This suggests that the Sr and Ba distribution probably have been changed as a result of changing external conditions which have not affected the original Na content. The temperature of recrystallization of 238/55 as read from the "strontium thermometer", is about 460° C; the temperature indicated by the "barium thermometer", Fig. 45 a, is 495° C.

The K/Ba ratios of the feldspars from the Eidet—Hovden monzonites and the young red granite (an. 5) are plotted against their temperatures of formation on Fig. 45 b. The slopes of the two curves,

Fig. 45 a and b, are essentially the same. The pegmatite from within the Eidet—Hovden intrusion (an. 7, cross on Fig. 45 b) is indicated to have crystallized at a minimum temperature of about 600° C.

Feldspars from only two rocks plot above the curve on Fig. 45 a. 525° C. is a too high temperature for no. 73—99/56 because the potash feldspar concentrate was contaminated by plagioclase. If the K/Ba ratio of this feldspar is plotted on the curve the temperature read is 440° C. This temperature is in better agreement with the other feldspars from the amphibolite facies rocks. No. 64—65/56 is a plot of the old red granite.

Only one of the granulite facies rocks (62—an. 12) plots significantly below the curve. The temperature derived for this rock by using the feldspar geological thermometer (510° C) is remarkably low. The rock is a type sample of the banded series rocks and is chemically extremely similar to the western monzonites. The potash feldspars are smaller in the banded series rocks than in the porphyroblastic mozonites but they are chemically nearly identical (Table III). On Fig. 45 a temperature of 580° C is indicated for this rock; with the "strontium thermometer" 560° C is obtained.

Nearly all the feldspars from the amphibolite facies rocks plot below the curve. Exceptions are nos. 69—35/55 (amphibolite facies gneiss) and 74 — E-vatn (red granite vein in gneiss).

In order to try to explain the random distribution of the plots of the amphibolite facies gneisses the occurrence of biotite in the rocks must be considered. In the granulite facies gneisses potash feldspars are almost the only potassium mineral present whereas in the amphibolite facies gneisses biotite also is a major mineral. Thus in the latter rocks K is distributed between two different minerals and if the K/Ba ratio is different in the two, the formation of biotite in addition to potash feldspar, probably would affect the K/Ba ratio of the feldspar. If the feldspars have Ba contents corresponding to the maximum possible at the prevailing external (PT) conditions, more Ba could not enter the structure and would either be incorporated in some other mineral structures or leave the system. The data of Fig. 46 show all of the metamorphosed rocks on Langøy to have roughly similar K/Ba ratios. If there is any trend it is for the amphibolite facies gneisses to be relatively lower than the granulite facies rocks in Ba.

Table 52 presents the modal potash feldspars and biotites of the amphibolite facies rocks. It can be seen that a general relationship
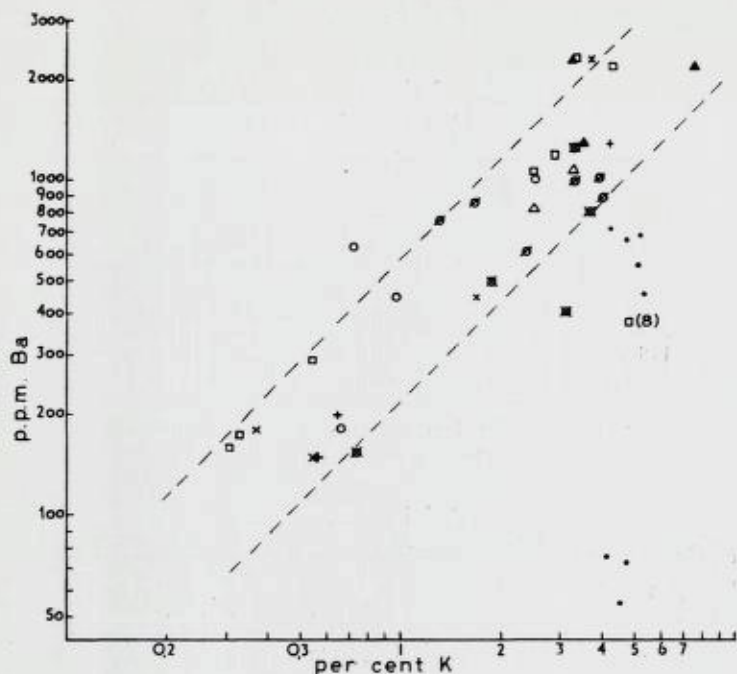
Fig. 46. Relationship between K and Ba in the Langøy rocks (symbols as in Fig. 37).

Forholdet mellom K og Ba i bergartene fra Langøy (symboler som på Fig. 37).

exists between the Or./Bio. ratio and the horizontal distance from the curve on Fig. 45 a. The red granitic vein (74—E-vatn) which plots on the curve consists almost totally of quartz and feldspar and essentially

Tabell 52.

Molecular per cent of potash feldspar and biotite, and the potash feldspar/biotite ratio in the amphibolite facies gneisses.

No	Or.	Bio.	Or./Bio.
67 (147/56)	28	16	1.75
68 (3/56)	31	7	4.43
69 (35/55)	32	7	4.57
70 (37/55)	22	13	1.69
71 (an. 1)......	31	11	2.75
72 (109/56)	28	18	1.55

Table 53.

*K and Ba contents in potash feldspar
and coexisting biotites from the Madras
charnockites (Howie, 1955).*

	Potash feldspar			Biotite		
No.	K %	Ba p.p.m.	K/Ba	K %	Ba p.p.m.	K/Ba
2270....	12.08	5000	24.16	7.20	2300	31.30
36218...	11.5	4500	25.56		1000	72.00*
115.....	11.5	3000	38.33		900	80.00*
4646 A..	11.0	2500	44.00		1000	72.00*

* Howie (1955) does not give K values for these
biotites; it is taken here as 7.20 % K.

Table 54.

*K/Ba ratio in the hypothetical potash feldspars formed by
transforming biotite into potash feldspar and adding to the
feldspar already present. K/Ba ratio in the biotite is assumed
to be five times the K/Ba ratio in potash feldspars.*

No.	Or.	K/Ba	Or. calc. from bio.	K/Bax5	K/Ba in total feldspar
67 (147/56)..	28	15.18	9.6	75.90	30.7
68 (3/56)....	31	17.60	4.2	88.00	26.1
69 (35/55)...	32	31.58	4.2	157.90	46.7
70 (37/55)...	22	14.77	7.8	73.85 .	30.3
71 (an. 1) ...	31	29.02	6.7	145.1	49.6
72 (109/56)..	28	10.52	10.8	52.60	22.2

no mica. If all the K and Ba in the biotites formed potash feldspars a
displacement of the plots to the right would be expected to take place
if the K/Ba ratio in the biotites was higher than in the feldspars.

Howie (1955, Table V, p. 747 and Table VII, p. 752) reported
the K and Ba contents of potash feldspars and coexisting biotites
from the Madras charnockites; data reproduced in Table 53. According
to his data the Ba content is less and the K/Ba ratio is from 1½ to 3
times higher in the biotites. If the K/Ba ratio in the biotites from Lang-
øy is assumed to be as much as five times that in the coexisting feld-
spars the K/Ba ratio in the normative potash feldspar phase becomes
that indicated in the last column in Table 54. The ends of the arrows
mark their position on Fig. 45 a.

Table 55.

The ratio $\dfrac{Ba\ in\ Or.}{Ba\ in\ plag.}$ *related to temperature of formation of the alkali feldspar (Table 39).*

No	$\dfrac{Ba\ in\ Or.}{Ba\ in\ plag.}$	T°C
56, 80 (293/55)	25.8	590
58, 81 (251/55)	45.7	585
60, 82 (238/55)	4.6	(605 – ret. met.)
61, 83 (224/55)	32.5	560
62, 84 (an. 12).....	12.0	510
65, 85 (25/55)	6.2	545
70, 86 (37/55)	11.7	465
71, 87 (an. 1)......	9.5	400

The Ba content in plagioclases is typically rather low (Rankama and Sahama, 1950). Howie (1955) reported a 20 to 60 times higher Ba content in the potash feldspars than in the coexisting plagioclase. The $\dfrac{Ba\ in\ Or.}{Ba\ in\ plag.}$ ratios of the Langøy rocks are given in Table 55. The data clearly indicate that Ba prefers the potash feldspar over the plagioclase structure and that this is especially characteristic of the high grade metamorphic (high temperature) rocks.

The K/Ba ratios of all the rock types from Langøy are plotted on Fig. 46. Except for the young red granites, which are low in Ba there is a distinct interrelationship between K and Ba in the rocks. It is not possible to distinguish between the various rock types on Langøy on the basis of the K/Ba ratios. One of the amphibolite facies gneisses is low in Ba (an. 1). The young red granites are characterized by low Ba contents and three of them (282/55, 254/55, 270/55) have exceptionally low contents. There is no obvious reason (chemical or mineralogical) why these three should be different from the rest. The three are related geographically, however, in that they all occur inside the porphyroblastic monzonitic granulites, one as a dyke (282/55), one as a small "patch" (270/55), and one as a minor massif (254/55).

All of the diverse rocks within the Eidet—Hovden intrusion have normal ratios except for the strongly fractionated dyke 60/56 which is low in Ba and in this respect is related to the young granites.

Gallium (Ga³⁺ 0.62 A).

The crustal abundance of Ga is 19 p.p.m. (Shaw, 1957, Table 1, p. 166). Ga is camouflaged in Al minerals (Al^{3+} 0.57 Å). In the various Langøy rocks Ga ranges between 9 and 33 p.p.m. The minimum value is from one of the young red granites (211/55). These rocks are in general low in Ga (9 to 16 p.p.m., average 13.4 p.p.m.). The maximum value is from one of the amphibolites associated with the metamorphic sedimentary rocks (29/55). The five amphibolites associated with sedimentary rocks (81/56, 77/56, 13/56, 29/55, 33/55 j) range between 20 and 33 p.p.m. Ga with arithmetic average 24 p.p.m. The granulite facies rocks proper (porphyroblastic monzonitic granulites, nos. 9 and 10, the banded series, nos. 14 to 17, and the charnockite border series, nos. 21 to 26) range between 16 and 32 p.p.m. Ga with arithmetic average 22 p.p.m. The Madras charnockites (Howie, 1955, Table II, p. 737) range between 5 and 35 p.p.m. Ga with arithmetic average 16 p.p.m. Disregarding the two most basic types of the Madras charnockites (which contain 5 p.p.m. Ga) the arithmetic average Ga content of this series is 18 p.p.m. The Ga/Al ratios ($\times$ 10 000) of the Langøy rocks (Table 56) range between 1.22 (211/55) and 4.85 (29/55). The average ratio is 2.25. The amphibolites associated with sedimentary rocks have higher ratios than the average whereas the young red granites have lower ratios.

There are very small differences between the ratios of the gabbros (1 to 3), monzonites (4 to 5) and dyke rocks (6 to 8) of the Eidet—Hovden intrusion, or between these rocks and most of the regionally metamorphosed rocks. The banded series rocks (14 to 17) have fairly high ratios (2.65 to 3.76, average 3.08).

That Ga^{3+} also substitutes for Fe^{3+} (0.67 Å) is shown by the high content of Ga in magnetites (10 to 120 p.p.m. Ga in magnetites from the Madras charnockites, Howie, 1955, Table XIV, p. 761). Ga/Al + Fe³ ratios ($\times$ 10 000) of the Langøy rocks are also listed in Table 56. These ratios show the same trend as the Ga/Al ratios but their scatter is much less (1.15 to 2.72, average 1.88).

Because of its larger size Ga will preferentially substitute for Al in 6 co-ordination. This is indicated by the data given in Table 57 in which Ga/Al ratios of various coexisting minerals from the Madras charnockites are given. It can be seen that the Ga/Al ratios decrease in the series: orthopyroxene > clinopyroxene > hornblende > (biotite)

Table 56.

$$\frac{Ga \times 10000}{Al} \ and \ \frac{Ga \times 10000}{Al + Fe^{3+}} \ ratios \ of \ the \ Langøy \ rocks.$$

No	$\frac{Ga \times 10000}{Al}$	$\frac{Ga \times 10000}{Al + Fe^{3+}}$	Rock type
1 (an. 16)..	1.89	1.75	Gabbro
2 (an. 8)...	1.84	1.80	
3 (32/56d).	2.34	2.21	Eidet-
4 (an. 15)..	2.20	1.71	Monzonite Hovden
5 (an. 9)...	1.85	1.71	instrusion
6 (50/56) ..	2.60	1.95	Dykes
7 (91/56a) .	2.00	1.51	
8 (60/56) ..	2.58	2.28	
9 (57/56) ..	2.35	1.89	Porphyroblastic monzonitic
10 (55/56) ..	2.08	1.78	granulites
11 (318/55) .	2.02	1.58	Retrograde metamorphosed gneisses
12 (310/55) .	1.81	1.63	
13 (235/55) .	1.41	1.17	
14 (58/56) ..	3.76	2.54	The banded series
15 (145/56) .	2.71	2.50	
16 (an. 12)..	2.65	2.11	
17 (59/56) ..	3.18	2.85	
18 (81/56) ..	3.56	1.93	Rocks associated with metamorphic sedimen-
19 (77/56) ..	3.54	2.39	tary rocks in the banded series
20 (70/56) ..	1.26	1.25	
21 (151/56b)	2.39	2.16	The charnockite border series
22 (150/56b)	2.15	1.99	
23 (158/56) .	1.96	1.77	
24 (142/56) .	1.94	1.77	
25 (150/56a)	2.12	1.92	
26 (151/56a)	2.10	2.00	
27 (3/56b) ..	2.03	1.50	The amphibolite facies veined gneisses
28 (147/56b)	2.32	1.90	
29 (147/56a)	1.50	1.40	
30 (an. 1)...	2.41	2.31	
31 (3/56a) ..	1.97	1.87	
32 (13/56) ..	2.50	1.86	Rocks associated with metamorphic
33 (29/55) ..	4.85	2.72	sedimentary rocks in the amphibolite
34 (33/55j) .	2.34	1.68	facies veined gneisses
35 (43/55) ..	2.16	1.84	
36 (71/55) ..	1.57	1.39	Young red granites
37 (an. 5)...	1.73	1.61	
38 (347/55) .	2.19	2.01	
39 (211/55) .	1.22	1.15	
40 (208/55) .	1.90	1.80	
41 (282/55) .	1.93	1.83	
42 (254/55) .	1.94	1.79	
43 (270/55) .	2.03	1.92	

Table 57.

$$\frac{Ga \times 10000}{Al}$$ *of coexisting minerals from*

the Madras charnockites (compiled
from Howie, 1955).

Minerals	2270	137	6436	2941	4642A
Potash feldspar	0.2	1.3	1.0	–	–
Plagioclase ...	2.5	1.8	2.0	3.0	1.5
Biotite	3.0	–	–	–	–
Hornblende ..	2.0	9.0	–	4.0	2.0
Clinopyroxene	5.0	–	–	3.0	5.0
Orthopyroxene	5.0	13.0	12.0	23.0	14.0

> plagioclase > orthoclase. Further, according to Shaw, 1957, Table 8, p. 180, olivine contains extremely little Ga. The small variations in the Ga/Al and Ga/Al + Fe³ ratios of the Langøy rocks can easily be ascribed to mineralogical differences.

Copper ($Cu^{2+} \sim 0.83$ Å).

The crustal abundance of Cu is about 60 to 70 p.p.m. (Rankama and Sahama, 1950).

Cu is geochemically mainly chalcophile but it is probable that in some instances it also enters the structures of the rock forming silicates (Rankama and Sahama, 1950, p. 99).

The average Cu content in igneous rocks is:

> Basic igneous rocks ...... 149 p.p.m.
> Intermediate igneous rocks .. 38 p.p.m.
> Acidic igneous rocks ...... 16 p.p.m.
> (Rankama and Sahama, 1950, p. 697).

The Cu content in the Langøy rocks ranges between 2 and 300 p.p.m. The lowest concentrations were found in the young red granites, between 2 and 5 p.p.m. Cu, with arithmetic average 4 p.p.m. Cu. Concentrations above 100 p.p.m. Cu are rare and were observed only in three of the rocks which are closely associated with the metamorphic sedimentary rocks. The maximum amount of 300 p.p.m. Cu was obtained for an amphibolite closely associated with graphite schists (81/56). This high Cu content probably represents metasomatic influx from the graphite; (commonly the amphibolites next to these schists are rich in sulphides).

Table 58.

Cu contents in some unmetamorphosed basalts.

	p. p. m. Cu
Keweenawan lava (Greenstone flow). Cornwall and Rose, 1957	120
Hawai (average, mainly basalt), Goldschmidt, 1954..........	155
Columbia river, basalt, Goldschmidt, 1954	370

Table 59.

Cu contents of amphibolitic bands in the Langøy rocks
(concentrations in brackets refer to Cu contents of one adjacent
acid band).

		Cu p. p. m.
14 (58/56) ...	The banded series	(70) 78
15 (145/56) ..	The banded series	10
21 (151/56b).	The charnockite border series	(11) 14
22 (150/56b).	The charnockite border series	(16) 24
27 (3/56b) ...	The amphibolite facies veined gneisses	(22) 34
28 (147/56b).	The amphibolite facies veined gneisses	(38) 40

The granulite rock (70/56) which is associated with the sediments at Jørgenfjord and is rich in Cu (210 p.p.m.) can likewise be seen to be rich in sulphides (pyrrhotite).

The amphibolitic rocks associated with the metamorphic sedimentary rocks (nos. 81/56, 77/56, 13/56, 29/55, 33/55j) range between 8 and 300 p.p.m. Cu. The last three which contain 63, 167, 8 p.p.m. Cu respectively, are believed to have been formed through metasomatic action on limestones, whereas the other two, which contain 300 and 52 p.p.m. Cu may represent original basaltic lava flows. The extremely low Cu content of no. 33/55 j (8 p.p.m.) indicates that this rock is not a metabasalt. However, Cu, especially of sulphides, is volatile and easily activated through metamorphic action so the severe metamorphism within this area probably altered the original Cu content significantly. Cu concentrations of some basaltic rocks are given in Table 58. As amphibolitic bands in amphibolitic gneisses are sometimes regarded as metabasalts it is of interest to compare the data of Table 58 with those for rocks from Langøy which chemically may be designated metabasalts (Table 59). (The amphibolites occurring with sedimentary rocks discussed above are omitted.) The "amphibolitic"

Table 60.

Average Cu content in the various Langøy rocks.

		p. p. m. Cu
1	Eidet-Hovden gabbro (1—3)	65
2	Eidet-Hovden monzonite (4—5)	30
3	Eidet-Hovden dykes (6—8)	66
4	Porphyroblastic monzonitic granulites (9—10)	33
5	Retrograde metamorphosed gneisses of monzonitic type (11—12)	21
6	Retrograde metamorphosed gneiss of granitic type (13)	4
7	The banded series (14—17)	48
8	Amphibolites associated with sedimentary rocks in the banded series (18—19)	176
9	Granulitic rock associated with sedimentary rocks in the banded series (20)	210
10	The charnockite border reries (21—26)	27
11	The amphibolite facies veined gneisses (27—31)	35
12	Amphibolites associated with sedimentary rocks in the amphibolite facies veined gneisses (32—34)	79
13	Quartz-plagioclase-biotite schist (35)	3.5
14	Young red granites (36—43)	4

bands have conspiciously less Cu than the "average" unmetamorphosed basalts; in fact, their contents are not significantly higher so far as this element is concerned than that of the directly adjacent acidic band. The intermediate rock types on Langøy are commonly higher in Cu than the basic types and only the granitic rocks have lower Cu contents. The Cu contents of the metamorphosed rocks on Langøy (Table 60) are similar to the average value of intermediate igneous rocks.

Lundegårdh (1947, p. 143) found an average of 30 p.p.m. Cu in the pre-Cambrian basic rocks of Roslagen, Sweden. 17.5 p.p.m. Cu is the average of 6 "old amphibolites" from Roslagen (ibid, Table 8, p. 32).

The low Cu content in the basic rocks from Langøy and Roslagen as compared to those of unmetamorphosed basalts is probably related to the activation of Cu during metamorphism and, if so, cannot be ascribed to original differences. The depletion of Cu from such rocks as the result of metamorphic action is interesting when related to the question of the origin of the Cu in sulphide ores which are characteristic mineral deposits in regions of relatively low grade regional metamorphism (see Vokes, 1957, pp. 179—185). The great similarity between the Cu contents in the amphibolitic and acidic rocks on Langøy (Table 59) is interesting. It strongly suggests that Cu has left the basic rocks. However, the amounts found in the more acidic rocks are

not higher than the average of intermediate igneous rocks (see above). Thus, it may be that the Cu which left the amphibolites was either disseminated fairly evenly through the other rocks within the area or left the region altogether. As a proportional amount as compared to that of the amphibolites cannot be judged to have left the intermediate and acid rocks which recrystallized under the same temperature and pressure conditions, this may give a clue to the much disputed problem of whether Cu is present in rocks mainly as sulphides or in replacement of major elements in certain silicate crystal lattices. Current opinions are highly diversified. Rankama and Sahama (1950, p. 99) stated "copper seems to be sulphophile, but this character is not very pronounced".

If we consider 120 p.p.m. Cu as the average copper content in basalts (the average of the various rock types of the Greenstone flow, Table 58) and the average content in the Langøy amphibolites as 24 p.p.m. Cu (rejecting no. 58/56, Table 59 because this rock occurred only a few metres away from the Eidet—Hovden gabbro contact and its high Cu value might reflect contact metamorphic effects) it can be seen that on an average 4/5th of the original Cu content has been driven off. Cornwall and Rose (1957) found between 53 per cent and 66 per cent of the total Cu of the Keweenawan basalts to be contained in the minerals, but they were not able to state whether all the Cu contained in the minerals replaced major elements in the crystal lattices. If it is assumed that all the Cu originally present as sulphides in the Langøy amphibolites was driven off during metamorphism it means that 20 per cent of the total original Cu of the basalts is contained in the structures of the silicate minerals. A higher proportion of Cu may be present in the silicate lattices of the more acid rocks than of the more basic rocks. Cornwall and Rose (1957) found 100 per cent of the total Cu of the granophyre from the Keweenawan to be contained in the minerals. This could be the explanation of why the intermediate and acid rocks on Langøy have Cu contents close to those of the average igneous types of similar compositions.

The gabbros of the Eidet—Hovden intrusion (an. 16, an. 8, 32/56d) are not particularly high in Cu (average 65 p.p.m.). The monzonites (an. 15, an. 9) have Cu contents similar to those of the western monzonites (57/56). The dykes within this intrusion are, however, high in Cu; their range is between 37 and 100 p.p.m. Cu with the highest Cu content in the most basic varieties. Their Cu content is 2 to 3 times

higher than that of the metamorphosed rocks within the same silica range.

Cu has been determined in both potash feldspars and the plagioclases from the Langøy rocks. In the potash feldspars Cu ranges between 0.6 and 19.2 p.p.m. with arithmetic average of 6 p.p.m. The Cu content is higher in the plagioclases than in the coexisting potash feldspars and ranges between 4 and 98 p.p.m. with arithmetic average 29 p.p.m. Cu; very similar to the 70 p.p.m. maximum Cu and 30 p.p.m. average Cu found by Cornwall and Rose (1957) in plagioclase from the Keweenawan lavas.

The Cu contents of both the plagioclase and potash feldspars are irregular and have not been found to be related to the concentration of any other element present. That Cu is enriched in the plagioclases relative to the potash feldspars may suggest that it substitutes for the major elements outside the Al—Si tetrahedra. The radius of Cu^{2+} (0.83 Å) is rather close to the radii of Na^+ (0.98 Å) and Ca^{2+} (1.06 Å), but is rather different from the radius of K^+ (1.33 Å).

Vanadium (V^{3+} 0.65 Å).

The crustal abundance of V is about 150 p.p.m. (Goldschmidt, 1954). The average of the Langøy rocks (excluding the special types of the Eidet—Hovden intrusion and the young red granites) is 157 p.p.m. V. This is the arithmetic mean and most of the Langøy rocks contain less than 150 p.p.m. V. Only five rocks were found to contain 150 or more p.p.m. Four of these are amphibolites associated with sedimentary rocks; nos. 81/56, 77/56, 13/56, 29/55 contain 275, 225, 290 and 500 p.p.m. V, respectively. (A fifth amphibolite from this same group contains 125 p.p.m. V, no. 33/55j). One of the amphibolite bands in the amphibolite facies gneisses is exceptionally high in V; it contains 1022 p.p.m.

The average V content in the metamorphic rocks, excluding those mentioned above, is 85 p.p.m.

The average V contents for the various rock types from Langøy are given in Table 61. The V of magmatic rocks is predominantly of the trivalent type (V^{3+} 0.65 Å) according to Goldschmidt (1954, p. 489) but, Rankama and Sahama (1950, p. 596) claim that quadrivalent V (V^{4+} 0.61 Å) is more plentiful in igneous rocks. This is so because trivalent V is a strong reducing agent which would reduce ferric iron to the ferrous state. Both trivalent and quadrivalent V have

Table 61.

Average V content in the various Langøy rocks.

		p. p. m. V.
1	Eidet–Hovden olivine gabbro (1—3)	37
2	Eidet–Hovden monzonites (4—5)	29
3	Eidet–Hovden dykes (6—8)	73
4	Porphyroblastic monzonitic granulites (9—10)	72
5	Retrograde metamorphosed gneisses of monzonitic type (11—12)	82
6	Retrograde metamorphosed gneiss of granitic type (13)	24
7	The banded series (14—17)	105
8	Amphibolites associated with sedimentary rocks in the banded series (18—19)	250
9	Granulitic rock associated with sedimentary rocks in the banded series (20)	92
10	The charnockite border series (21—26)	77
11	The amphibolite facies veined gneisses (27—31)	93*
12	Amphibolites associated with sedimentary rocks in the amphibolite facies veined gneisses (32—34)	305
13	Quartz-plagioclase-biotite schist (35)	120
14	Young red granites (36—43)	16

* No. 28 with the extremely high V content (1022 p. p. m.) is excluded.

a size which would fit into the structures of several rock forming silicates and oxides. It is believed to be able to replace Fe^{2+} (0.83 Å), Mg^{2+} (0.78 Å), Fe^{3+} (0.67 Å), Ti^{4+} (0.64 Å) and probably also Al^{3+} (0.57 Å) in such minerals. Because of this V does not show any strong coherence with any of the above mentioned elements. However, it is, in general, enriched in the gabbroid and basaltic rocks.

That its concentration in olivine rocks is typically less than 100 p.p.m. (Goldschmidt, 1954, p. 489) agrees with the findings of this study where in the olivine gabbro (an. 16, an. 8, 32/56 d) it was found in the concentration range of between 21 and 54 p.p.m. whereas the amphibolitic rocks in the gneisses contain more than 100 p.p.m. V. (Table 62.) In this table it also may be seen that not only does rock no. 147/56 b, have an exceptionally high V content but also the adjacent acid gneiss is higher than the normal in this element.

In Fig. 47 the concentration of V is plotted against "the position" of the rocks. The irregular distribution of V is obvious. However, a broad pattern of regular variation is distinguishable (indicated by dotted lines). The Eidet—Hovden gabbro (1—3) and the monzonites (4—5) form a separate group, low in V, whereas the above mentioned amphibolitic rocks have decidely high contents of this element.

Table 62.

V contents in the «amphibolitic» and «granitic» bands in the metamorphic rocks.

	p. p. m. V. in «amphibolites»	p. p. m. V in «granitic» bands
14 (58/56) and 17 (59/56) The banded series .	115	70
15 (145/55) The banded series..............	124	
21 (151/56b) and 26 (151/56a) The charnockite border series.............................	110	45
22 (150/56b) and 25 (150/56a) The charnockite border series......................	107	70
27 (3/56b) and 31 (3/56a) The amphibolite facies veined gneisses	140	50
28 (147/56b) and 29 (147/56a) The amphibolite facies veined gneisses	1022	125

Fig. 47. The V concentration in the Langøy rocks plotted against $(1/3Si+K)-(Ca+Mg)$ of the rocks (symbols as in Fig. 37).

Innholdet av V i bergartene fra Langøy plottet mot $(1/3Si+K)-(Ca/Mg)$ av bergartene (symboler som på Fig. 37).

The high V content in some of the amphibolites is interesting when considered in conjunction with the possible origins. Three of them (nos. 13/56, 29/55 and 33/55j) occur in direct contact with meta-limestones. These rocks are exceptionally low in Cr and it is for this

Fig. 48. Relationship between Co and Mg in the Langøy rocks (symbols as in Fig. 37).

Forholdet mellom Co og Mg i bergartene fra Langøy (symboler som på Fig. 37).

reason difficult to consider them as original basalts. The field indications are that they may represent metasomatized limestones. However, most limestones contain very little V (< 10 p.p.m., Rankama and Sahama, 1950, Table 24.5, p. 601).

Limestones containing bituminous substances may, however, be high in V (Goldschmidt, 1954, p. 498) and black marine shales may be exceptionally high in this element. This appears pertinent when it is recalled that graphite schists are abundant in the area and commonly occur close to limestone bands.

The two amphibolites (81/56, 77/56) which contain 275 and 225 p.p.m. V, respectively, are directly associated with graphite schists in the field.

The amphibolite in the gneisses (147/56 b) which contains the maximum amount of V observed (1022 p.p.m.) is not associated with any rock of a definite sedimentary origin.
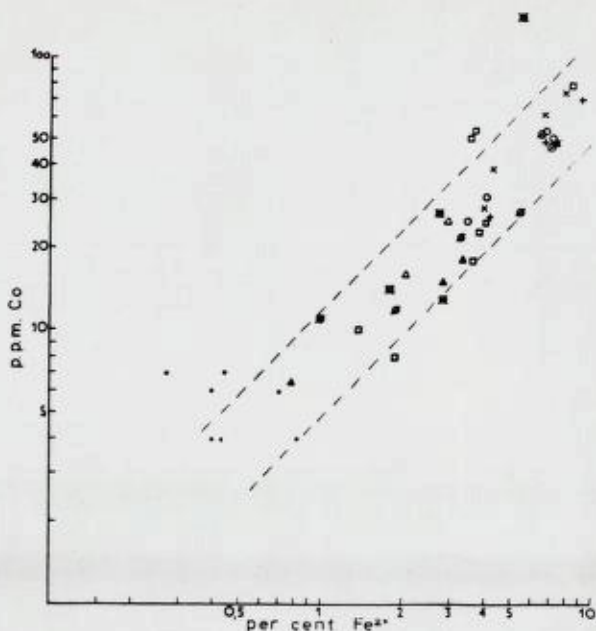
Fig. 49. Relationship between Co and Fe^{2+} in the Langøy rocks (symbols as in Fig. 37).

Forholdet mellom Co og Fe^{2+} i bergartene fra Langøy (symboler som på Fig. 37).

Chromium (Cr^{3+} 0.64 Å).

The crustal abundance of Cr is 200 p.p.m. (Goldschmidt 1954). The average content in the Langøy rocks (excepting the Eidet—Hovden intrusion and the young red granites) is 176 p.p.m. Cr. The distribution of Cr in the Langøy rocks is even more irregular than that of V although Cr is, as expected, typically concentrated in the most basic rocks. Exceptions to this are the three amphibolites associated with sedimentary rocks in the amphibolite facies gneisses (nos. 13/56, 29/55 and 33/55j); these are low in Cr (in no. 29/55 Cr was not even detected, i. e., it is < 10 p.p.m.). The close association between these amphibolites and metalimestone suggests that they represent metasomatized limestones. The low Cr concentrations in these amphibolites could be taken as corroborative of this as limestones are low in Cr (~ 2 p.p.m. (Rankama and Sahama 1950, Table 28. 3, p. 623). However, Lundegårdh (1947), found low Cr contents to be characteristic of the

basic rocks from Roslagen in Sweden, and also reported (ibid. p. 142) extremely low Cr contents for Palaeozoic "plateau-diabase" of western Sweden (6 p.p.m.) and for an Icelandic basalt (23 p.p.m.). Thus it seems that low Cr contents do not constitute proof of an a non-basaltic origin.

Cr was detected in only one of the young granites. It is also very low in the monzonites within the Eidet—Hovden intrusion. In the chemically similar porphyroblastic monzonitic granulites the Cr concentration is three times as high. As a whole the intermediate and acid metamorphosed rocks are higher in Cr than the Eidet—Hovden monzonites. The dyke rocks in this intrusion have Cr contents similar to the metamorphic rocks.

The highest Cr content was found in a basic rock within the banded series (no. 145/56, 1030 p.p.m. Cr). The other rocks within this series range between 15 and 58 p.p.m. Cr. All but one (no. 151/56 a, 50 p.p.m. Cr) of the six rocks from the charnockite border series are comparatively high in Cr. The average for this series is 336 p.p.m. Cr. The average for the petrographically similar banded series is 238 p.p.m. Cr (excluding the extremely high no. 145/56 the average is 34 p.p.m. Cr). That of the amphibolite facies gneisses is 75 p.p.m. Cr.

Cobalt (Co^{2+} 0.84 Å).

Goldschmidt (1954) assumed 40 p.p.m. to be the crustal abundance of Co. Sandell and Goldich (1943) found Co in rocks to vary in a linear fashion with Mg and the ratio, Co/Mg, to be approximately 0.0011. Using this value the average Co content of igneous rocks was predicted to be 23 p.p.m. Co. Nockolds and Mitchell (1948) found the Co/Mg ratio in some Caledonian igneous rocks to be highly variable and without any definite trend. Co is plotted against Mg (Mg^{2+} 0.78 Å) for the Langøy rocks (excluding the most Mg poor red granites) in Fig. 48. The dashed central line indicates the variation of the mean ratio from the basic to the acid rocks. In the basic (Mg rich) rocks this ratio is 0.0011 and in the acid (Mg poor) end it is 0.0023. Thus Co is slightly enriched relative to Mg in the acid rocks. This was also found to be the case for the Madras charnockites (Howie, 1955, p. 738). Fig. 49 shows plot of p.p.m. Co versus per cent Fe^{2+} (0.83 Å) in the same rocks. The dashed lines indicate the limits of "normal scatter". The slope is 45° which indicates a constant average ratio from the basic to the acid rocks. Three of the young red granites
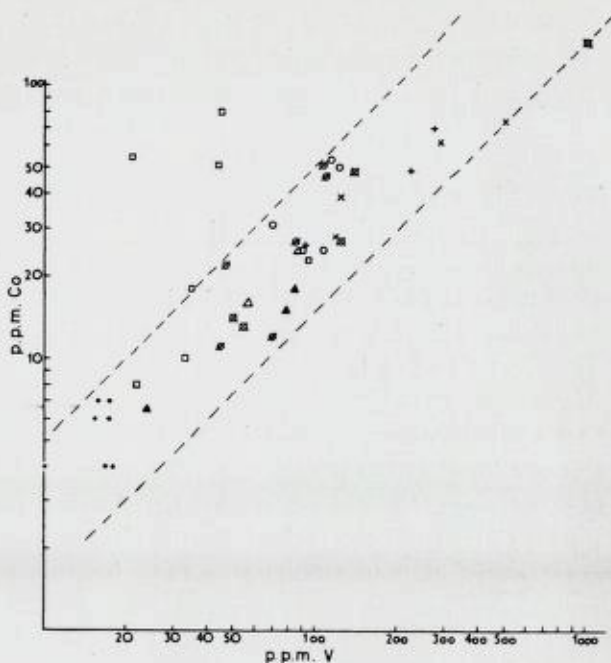
Fig. 50. Relationship between Co and V in the Langøy rocks (symbols as in Fig. 37).

Forholdet mellom Co og V i bergartene fra Langøy (symboler som på Fig. 37).

are enriched in Co relative to Fe^{2+} which may indicate that extremely fractionated rocks are enriched in Co. The enrichment of Co relative to Fe^{2+} is not as pronounced as it is relative to Mg and the data suggest a much closer coherence between Co and Fe^{2+} than between Co and Mg in these rocks. The mean of the Co/Fe^{2+} ratio is about 0.00075.

Rock no. 147/56 b is most enriched in Co relative to Mg and Fe^{2+}, it contains 144 p.p.m. This is an amphibolite band in the amphibolite facies gneisses and it also was found to be enriched in V (1022 p.p.m.). Apart from this it is not possible to distinguish between the various rock types on the basis of these ratios. Fig. 50 shows plots of p.p.m. Co versus p.p.m. V^{3+} (0.65 Å); it is clear that a certain coherence exists between these two elements in the Langøy rocks. The three Eidet—Hovden gabbros (1—3) can be separated on the basis of their Co/V ratios. It is interesting that the Co and V rich amphibolite mentioned above has a fairly normal Co/V ratio.

Nickel (Ni^{2+} 0.78 Å).

Ni is more abundant than Co in the upper lithosphere (100 p.p.m., Goldschmidt 1954, p. 647).

The substitution of Ni for Mg in ferro-magnesium minerals is often cited as an example of the camouflage principle. Ringwood (1955) suggests that Ni^{2+} enters crystal lattices at the expense of Fe^{2+} rather than of Mg^{2+}.

Ni concentrations in the Langøy rocks are presented in Table II. No strong coherence can be seen to exist between Ni and either Mg or Fe^{2+}. Ni tends to be concentrated in the most basic rocks and decreases more rapidly with increasing SiO$_2$ content in the rocks than any of the other elements. This was also found to be the case in the Madras charnockites (Howie, 1955, Fig. 4, p. 739). Rocks with roughly equal concentrations of Mg may show a wide spread of Ni concentrations. However, it does not appear possible to distinguish among the various rock types on the basis of this.

Petrology.

The veined and banded rocks
(including the metamorphic sedimentary rocks).

The basal gneisses on Hinnøy pass undisturbed into the northeast part of Langøy where they are called in this work the amphibolite facies veined gneisses. The continuation of these rocks to the west is represented by "the charnockite border series", and further to the west by "the banded series" which are charnockites proper. The petrochemical similarity of the various rock types of the charnockite provinces to their surrounding rock complex in general has been pointed out by Parras (1958, p. 11). These three rock series have similar appearances in the field and are characterized by alternating veins and bands of basic (amphibolitic or dioritic) and acid (monzonitic to granodioritic) rocks (Figs. 5—8). The major difference between them is that of metamorphic grade. The banded series and the charnockite border series are granulite facies rocks in contrast to the amphibolite facies of the gneisses in the east. The transition between the amphibolite facies gneisses and the charnockite border series cannot be seen to be transitional along the strike, but is rather abrupt. This, together with the fact that the metamorphic facies boundary is found to be coincident with a thrust plane beneath the anorthosite southeast of Eidsfjorden (Fig. 10) has led to the assumption that the observable boundary is related to the tectonics of the area. However, the charnockite border series is in many ways a transitional series between the amphibolite facies gneisses and the banded series (Figs. 26—29), and is thereby related to the amphibolite facies gneisses. The presence of hypersthene in both the acid and basic types of the series shows, on the other hand, that it belongs to the granulite facies.

The temperature of formation, or recrystallization, of the potash feldspars are also rather similar within the three series (400—505° C

in the amphibolite facies veined gneisses, and 510—565° C, average 535° C in the other two series, Table 39).

The amphibolite facies → granulite facies transition is largely shown by the mineral reaction:

$$3SiO_2 + K(Mg,Fe)_3 AlSi_3O_{10}(OH)_2 \rightleftharpoons$$
(amphibolite facies)
$$KAlSi_3O_8 + 3(Mg,Fe)SiO_3 + H_2O$$
(granulite facies).
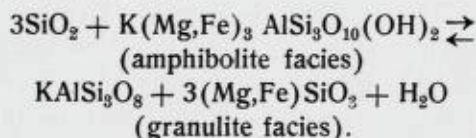
In the charnockite border series biotite and hypersthene are stable together whereas in the banded series the equilibrium is shifted toward the right side of the equation. As a concequence of this reaction free quartz becomes stable at a much lower silica content in rocks of the amphibolite facies than in those of the granulite facies (Fig. 34).

Other mineral reactions are also characteristic: increased stability of hornblendes in rocks of higher SiO_2 contents in the amphibolite facies gneisses (Figs. 26 and 29), variation of the optic axial angle of hornblende with metamorphic grade (Figs. 31 and 32), variations in the concentrations of opaque minerals (magnetite) (Fig. 33), and introduction of sphene in the lower grade metamorphic rocks.

A different type of reaction which is very characteristic of the facies transition is the increased disordering of the potash feldspars with increasing metamorphic grade. In the granulite facies rocks (including the charnockite border series) "orthoclase" is stable whereas microclines of intermediate to maximum triclinicities are stable in the amphibolite facies. According to this the "orthoclase" structure is stable above 510° C (see also Goldsmith and Laves, 1954 a).

That the amphibolite facies to granulite facies transition takes place at about this temperature has long been assumed by geologists. The Madras charnockites, otherwise similar to the granulite facies rocks on Langøy, contain dominantly microclines (Howie, 1955, p. 746), although some feldspars approach the "orthoclase" symmetry. In other granulite facies areas orthoclase is the normal potash feldspar phase (Wilson 1954, Heier 1956 a, 1957).[1]

Sedimentary rocks occur interbedded with gneisses of all the

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[1] Parras (1958) reports both untwinned and cross-hatched potash feldspars in highly metamorphosed rocks from west Uusimaa, Finland, but states (p. 89) that in the charnockitic rocks the untwinned form predominates.

three series. The true sedimentary rocks are represented by graphite schists and crystalline limestones. Quartzites may be present in the amphibolite facies gneisses, but are generally feldspathized. In the granulite facies it is impossible to separate them from normal quartz rich granulites. "Quartz banded iron ores" (see p. 65), likewise assumed to be of a sedimentary origin, are common within the area. A detailed mapping of the sedimentary rocks mentioned above could probably solve the tectonics of the area. This work cannot be carried out before detailed large scale topographical maps, or air photographs become available.

The graphite schists are associated with amphibolites, and amphibolites are typically in contact with metalimestones. These amphibolites are mineralogically and chemically similar to the normal amphibolite bands in the gneisses, but they tend to be broader and are typically more coarse grained. The amphibolites associated directly with the graphite schists may represent original basaltic lavas, but some of the amphibolites occurring with the limestones appear in the field to have been derived from impure limestones. However, there is not much chemical evidence to support this view. The only difference is the very low Cr content in amphibolites associated with limestones as compared to amphibolites in the gneisses. Similar low Cr contents are, however, also reported in some igneous basalts and are also very characteristic of metamorphosed amphibolitic rocks from Roslagen, central Sweden, (Lundegårdh 1947). It is, nevertheless, felt that such low Cr contents, when restricted to a very special group of amphibolites on Langøy, must have a petrogenetic significance.

The amphibolites associated with the graphite schists are rich in pyrrhotite. This is believed to represent a metasomatic influx of sulphides from the schists during regional metamorphism. These rocks are, therefore, not amenable to studies of original element concentrations.

The chemical data cannot be seen to give any clues as to the origin of the veined and banded rocks. Howie (1955), who concludes that the Madras charnockites represent a series of metamorphosed (granulite facies) igneous rocks, uses the chemical data as a proof of their igneous origin and writes (p. 767) ". . . the chemistry of the Madras series affords strong confirmation of the view that these rocks were originally a series of calc-alkali plutonic rocks. The analyses all fall on reasonably smooth curves on both the triangular and Larsen-type

variation diagrams, which could hardly be the case for any random series of sediments. In addition, the distribution of the trace elements, which in general obeys Goldschmidt's rules, would be extremely unlikely, if not impossible, for any group of sediments with variable and haphazard composition".

The present author is opposed to this interpretation of the chemical data, and considers that the data obtained for the distribution of elements in the Langøy rocks indicate the futility of using trace elements as guides to petrogenesis in this way. The plots of major elements of the Langøy rocks, which are of diverse origin, give equally as smooth curves on the Larsen-type variation diagrams as the Madras charnockites (Fig. 37), and the trace elements follow closely, with very few exceptions, the distribution rules established originally by Goldschmidt.

Great thicknesses of sediments with an essentially uniform composition are known. Even in the common cases where thin bands of sediments alternate in a rapid succession, the trace element concentrations will, for each band in the series, be governed largely by the concentrations of the related major elements. Although an alternating sedimentary series may be said to have a varying and haphazard composition not in mineralogical equilibrium as a whole, the same cannot be said to be the case with the individual thin layers. Weathering and sedimentation are chemical fractionation processes just as much as magmatic crystallization. Further complicating factors are layers of graywackes and of acid and basic tuffs which are common geosynclinal sediments along with "other" sediments and lavas. How could it be possible on the basis of their chemistry to separate them from intrusive igneous rocks or lava flows of similar compositions? The same contingencies obtain when independent minerals are analysed. How is it possible in metamorphosed rocks of this kind definitely to know that the chemistry of a mineral is not directly inherited from a previous igneous mineral of the same type? In fact, there is hardly any common igneous rock with chemical composition not comparable to some known sediment.

When it is also remembered that the rocks in question were metamorphosed under severe conditions of regional metamorphism, i. e., conditions grading upwards into and being overlapped by partial anatexis, adjustments of chemical equilibrium, involving short range material transport, cannot be considered as unlikely.

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The veined and banded gneisses, both in the amphibolite and granulite facies on Langøy, which contain bands of recognizable sedimentary rocks such as graphite schists and crystalline limestones, will certainly, to an objective eye, give the impression of a series of highly metamorphosed sediments, possibly with included acid and basic tuffs and lavas. The clay type of sediments, and also partly the quartzites, which probably were present in large quantities in this series, have been altered beyond recognition. Only sediments such as limestones and carbonaceous shales, which are most resistant to metamorphic and metasomatic alterations, have remained recognizable as sedimentary rocks (see Barth, 1952, Table 72, p. 358).

The chemistry of these rocks, which is comparable to the chemistry of calc-alkali plutonic rocks, prevents the use of this kind of data as a proof of their origin.

The porphyroblastic monzonitic granulites.

Although this rock type has no great areal extension on Langøy it is a rather important member of the series of "plutonic rocks of Lofoten". It is the most important plutonic rock type areawise on the west side of Vestvågøy in the Lofoten group. Holland (1900) reported the lack of a porphyritic phase to be a characteristic of the Madras charnockites. The western monzonites on Langøy, with the large potash feldspar porphyroblasts (phenocrysts if an igneous nomenclature is used), are thus foreign to the charnockite series as it was originally defined. The mineral paragenesis of the western monzonites is that of the granulite facies and is similar to that of the associated banded series rocks on Langøy. Both are definitely regionally metamorposed rocks. It also is not significantly different in chemistry from intermediate rocks of the banded series. However, just as it is impossible definitely to relate the chemistry of the latter series to any certain mode of origin, it also is impossible to reverse the argument in the case of the monzonites and say that their chemical similarity to certain rocks of the banded series constitutes a proof of identical origin. On the other hand the chemical similarity and the close geographical association between them may be considered an indication of the existence of such a genetic relationship. The contacts between the monzonites and the banded series are gradational; locally the relationships give the impression of inclusions of banded rocks in the monzonites. The orien-

tation of the banding in these "inclusions" is parallel to the banding of the surrounding rocks. Further the potash feldspars in the banded series sporadically attain appreciable size, though never so large as those in the monzonites.

When chemistry is eliminated as proof of any definite origin for rocks the strongest indication of a magmatic origin of this monzonite is its massive and homogeneous appearance. However, as the mineral parageneses of the rocks are undoubtedly metamorphic, the question naturally arises as to whether the homogeneous and massive character and formation of the large sized potash feldspars are likewise the results of metamorphic processes. The temperature derived for their formation by the two feldspar thermometer is a metamorphic temperature for rocks of this composition ($\sim 600°$ C, Table 39). Considering the possibility of a metamorphic origin for these rocks through homogenization processes of rocks similar to those in the banded series, attention should be redirected to Fig. 30. The figure shows the intermediate rock types (i. e., rocks chemically similar to the western monzonites) to be more abundant within the banded series than in the amphibolite facies veined gneisses. In the text it was pointed out that this tendency towards homogenization by increasing metamorphism may possibly be due to sampling error because the acid and basic bands in the charnockites are not easily separable. However, no proof of heterogenities of the rocks was observed in thin sections.

As the degree of metamorphism on Langøy increases from east to west, the extreme western position of the monzonites on Langøy is especially interesting.

The few observations the author has made within the granulite facies area in the southwest of Hinnøy show this area to be composed largely of banded, including metasedimentary, rocks and quartz banded iron ores, similar to the banded series. As can be seen on the map (Fig. 4) these rocks have an eastern position.

On the other hand, it was stated that in the west of Vestvågøy, southsouthwest of the monzonites on Langøy, monzonites similar to the western monzonites are typical. It must be concluded that nothing definite can be said on the basis of the present studies about the origin of the western monzonites. They belong to the metamorphic series (granulite facies). Their massive character may point to an igneous origin, but may also manifest metamorphic homogenization processes.

The red augengneisses, monzonites east of Eidsfjorden and the nearby anorthosite.

The red augengneisses in the amphibolite facies in the southeast of Langøy are largely of a monzonitic type which is granitic close to the amphibolite facies veined gneisses. No chemical analyses of these rocks are available but based on their mineralogical modes they must approximate chemically the western monzonites. They contain sporadic inclusions of amphibolites, which may represent remnants of the amphibolite facies veined gneisses, and thus are also in this respect similar to the western monzonites in which inclusions of the banded series occur. An original connection between the western monzonites and the red augengneisses would cause difficulties to any theory of formation of the monzonites through metamorphic homogenization as outlined above. However, it seems that the two rocks are related only in that both are monzonitic. On the basis of character, occurrence, and theoretical considerations, the red augengneiss appears to be genetically related to the monzonitic rock of the granulite facies occurring directly northwest of it. Similarily this monzonite appears to be possibly related to the Eidet—Hovden monzonite, but less likely related to the western monzonite. This is supported by studies of thin sections: mesoperthitic alkali feldspar porphyroblasts or phenocrysts are typical in both of these monzonites, but not in the western monzonites; zoning of the plagioclases is also present in both of the two former monzonites, but is characteristically absent in the latter; and on the whole, the microscopic textures of the two former monzonites are very similar and different from that of the western monzonites. It has been stressed repeatedly that the Eidet—Hovden monzonites are probably igneous and that they also were very likely formed at a late stage of or after the regional metamorphism in granulite facies. Because of the similarities between them and the monzonites here dealt with, an igneous origin is favoured for these as well.

It is very interesting that, whereas the Eidet—Hovden monzonites are closely associated geographically with the Eidet—Hovden gabbro, the other monzonites are associated with the anorthosite east of Eidsfjorden. This may indicate an igneous origin of the anorthosite.

The Eidet—Hovden gabbro is in parts very plagioclase rich and appears to be related to anorthosites. The exact contact between the anorthosite and the monzonite has never been observed because of overburden.

Any igneous theory for the formation of the studied monzonites demands that at least the red augengneisses have subsequently been metamorphically recrystallized. Mesoperthites which characterize the Eidet—Hovden monzonites and the monzonites east of Eidsfjorden are also present locally in the red augengneisses (Table 17), and may be taken as a relic of an originally higher crystallization temperature. In several cases it can also be seen how potash feldspar (microcline) replaces plagioclase (p. 76). Rocks similar to the red augengneisses are also common on Hadseløy.

The old red granite.

This rock, which occupies a restricted area on the west side of Eidsfjorden, is similar in mineral facies to the surrounding banded series and regarded as a metamorphic granite. Theories of the origin of this rock will depend upon the general view anyone has about the formation of such rocks. The work carried out during this investigation offers no new data bearing on the granite problem. The granite has an irregular migmatite contact against the banded series in which the numerous "inclusions" of the banded series rocks have their banding parallel to that of the surrounding rocks. In mineral paragenesis the granite belongs to the granulite facies.

The Bø gabbro and the retrograde metamorphic gneisses.

The Bø gabbro which occurs in the south of Bø has a comparatively small areal extension. It is a massive pyroxene gabbro which shows no signs other than strong shearing along the borders, of any subsequent metamorphism. In the petrographic descriptions it was pointed out that the rock may be identical to hyperitic dykes and lenses which, according to Vogt (1909), are rather typical elsewhere in the Lofoten and Vesterålen group. Its unmetamorphosed appearance and its close connection to the zone of diaphtoritic gneisses along the west side of Jørgenfjord indicate that the intrusion took place after the regional metamorphism and that escaping gases and hydrothermal solutions from the crystallizing gabbro are responsible for the mineralogical disequilibria observed in the latter zone.

The retrograde metamorphic gneisses are chemically similar to the western monzonites, the banded series, and the old red granite. However, locally there are apparently anomalous occurrences of such rocks as the sillimanite gneisses.

The Eidet-Hovden instrusion.

The position of the Eidet—Hovden intrusion has been discussed at length in the chapters dealing with the petrography and metamorphism. It is petrographically rather unique on the island though, as is pointed out, it shows some interesting relations to the anorthosite and associated monzonites east of Eidsfjorden. There can be little doubt that it is an original igneous intrusion. The mineralogical evidence shows that the gabbro started to crystallize at the outer border. The centre, which is occupied by a monzonite, cannot be seen to have been derived from the gabbro magma through differential crystallization. The feldspars of the monzonites indicate a magmatic temperature of crystallization of above 1000° C. That there are numerous monzonitic dykes cutting the gabbro also indicates the magmatic character of the rock. The close spatial connection between this special type of monzonite and the gabbro may suggest some sort of comagmatic relationship. The time of the intrusion is rather difficult to ascertain. Arguments for a post metamorphic origin are rather strong (the regional metamorphic facies of the surrounding rocks is granulite facies):

1. Olivine is a characteristic mineral in the gabbro. Though olivine is stable under the PT conditions of granulite facies, it is not a stable metamorphic mineral in silicate rocks of this composition.
2. The change of mineral composition from the outer border to the centre agrees with magma crystallization proceeding in the same direction. Evidence of this trend must have persisted through subsequent metamorphism.
3. In a narrow zone near the contact with the banded series the gabbro occurs in a low grade facies. The occurrence of this zone is thought to be related to autometamorphism caused by escaping volatiles during the final stages of solidification of the gabbro magma and was not overlapped by later regional metamorphism of granulite facies intensity.

4. The mineral disequilibria observed in the monzonites (zoning of the pyroxenes and plagioclases in a direction corresponding to falling temperatures), and the presence of different types of alkali feldspars in the groundmass versus the phenocrysts, are suggestive of igneous rocks.

5. The temperature obtained by using the feldspar geological thermometer was > 1000° C. That the mesoperthites are exsolution products of a once homogeneous feldspar formed at a high temperature is indicated by the depletion in rubidium (Taylor and Heier, 1958 b).

6. Small rounded quartz grains, with straight extinction unlike the normal undulatory extinction of quartz in most metamorphic rocks occur in at least one of the dykes.

The only argument which possibly points towards a syn-metamorphic origin for this intrusion is the lack of dust and cloudy inclusions in the rocks.

The young red granites.

One striking fact became obvious during the study of the distribution of the young red granites. They are all restricted to the region of the highest grade of metamorphism west of the charnockite border series. In this region they occupy considerable areas as very homogeneous massifs and also occur as numerous minor patches within both the banded series and the western monzonites. They are non-existent in the amphibolite facies gneisses but nevertheless irregular red granitic and pegmatitic veins that occur herein may be related to them. However, the latter give very low temperatures by the feldspar geological thermometer (300° C for a red granitic vein from E-vatn, no. 74, Table 39) whereas magmatic temperatures measured are indicated for the red granites (780° C, no. 75, Table 39). As mentioned in the chapter dealing with the petrography, this is close to the temperature read on the experimental three feldspar diagram at 5000 bars H_2O pressure (725° C, Fig. 25). In any case the latter temperature is a minimum temperature, and the similarity between the two is strongly suggestive of an igneous origin for the rocks. It was shown that most of the granites plot on or close to the cotectic curve of the experimental

diagram. The one granite which plots at an appreciable distance from this curve (no. 71/55, Table 32) is the only sample in which mesophertites are the only feldspar phase present. Others carry independent plagioclase grains in subordinate amounts.

The close correlation between the minerals present and those shown experimentally to be present in a crystallizing melt of the same general composition also suggests a high temperature (anatectic?) origin for the rocks. The magmatic temperatures indicated by the feldspars, and the presence of biotite instead of hypersthene as the femic silicate mineral suggests that they have not been subsequently metamorphosed. The highest temperature reached by the regional metamorphism was slightly above 600° C (western monzonites). Thus, the regional temperature approaches temperatures of anatexis. At temperatures close to the partial melting point (i. e., the temperature of crystallization of the banded series and the western monzonites) experiments suggest that a high degree of mobility of the constituting atoms occurs. Such mobilization may cause a homogenization of rocks as discussed previously with respect to the formation of the western monzonite. At temperatures somewhat higher than the maximum measured in the regionally metamorphosed rocks on Langøy melting takes place. The melt would be granitic in composition and if moved would leave behind a solid rather basic rock. It is interesting that Ramberg (1951) found that the granulite facies rocks of western Greenland were basified as compared to the amphibolite facies gneisses. The young red granites are believed to have crystallized from palingenic melts in the above manner. The melts probably formed in regions where the PT conditions were slightly higher than obtained in the metamorphic Langøy rocks. It is believed that they migrated upwards and solidified where they came into regions of PT conditions under which magmas are not stable. As the distance of migration probably was not very great, the melts formed at various places did not coalesce before they crystallized and, therefore, most of the red granites occur as small patches within the granulite facies rocks.

This suggested formation of the red granites requires that the magmas were generated when the intensity of the regional metamorphism was at a maximum. But why then do they not exhibit the mineral parageneses typical of the regions where they crystallized? The areas they intruded were, at the time of the intrusion probably at the highest

temperature attained. Thus if the temperature was raised even slightly higher the now solid granite would probably remelt and migrate further upwards. Thus after the granites crystallized, the regional temperatures and pressures were gradually decreasing. Energy was not introduced and the igneous mineral paragenesis was kept metastable with only small adjustments, such as exsolution processes in the feldspars forming mesoperthites.

Following this hypothesis the granites are late orogenic, i. e., they crystallized when the temperatures were at a maximum but apparently were not affected by orogenic forces.

Summary and conclusions.

Nature of the rocks.

The previous picture of the geology of Langøy was mainly based on the work of Vogt (1909). According to him the Langøy rocks belonged to the series of "plutonic rocks of Lofoten", an igneous differentiation series ranging in composition from ultrabasic to acid types, with the intermediate "diorites" and "monzonites" having the widest areal extents. The metasedimentary rocks (marbles and graphite schists) occurring in them, were believed to be contact metamorphosed sedimentary inclusions in the igneous rocks. Depending upon the age ascribed to the metasedimentary rocks the age of the plutonic rocks was regarded as pre-Cambrian, Caledonian (Cambrian-Silurian) or post-Silurian. The sedimentary rocks were generally believed to be of Caledonian age because similar rocks from this period occur in large quantities on the mainland and on the east side of Hinnøy. Thus a late Caledonian or post-Silurian age of the plutonic rocks was generally favoured. A post-Silurian (Devonian?) age was originally assumed by Vogt (1909) and he correlated the Lofoten and Vesterålen rocks with the Permian igneous rocks of the Oslo region which at that time were believed to be of Devonian (post-Silurian) age.

The present work has shown that the "plutonic rocks of Lofoten" constitute a series of highly regionally metamorphosed rocks into which some igneous rocks are intruded.

The basal gneisses on Hinnøy pass undisturbed into the northeast part of Langøy where they are called in this work the amphibolite facies veined gneisses. The continuation of these rocks to the west is represented by "the charnockite border series" and further west "the banded series" which are charnockites proper. These three rock series together with the metamorphic sedimentary rocks which occur within them are regarded as high grade metamorphic geosynclinal sediments and lavas.

The origin of the porphyroblastic monzonitic granulites and the old red granite is uncertain. They are metamorphosed in granulite facies and may have formed through metamorphic processes or they are pre or syn metamorphic igneous intrusions.

The young red granites are explained as crystallization products of palingenic magmas.

The monzonitic rocks and the anorthosite east of Eidsfjorden are compared with the rocks of the Eidet—Hovden intrusion. This intrusion is regarded as a post metamorphic igneous intrusion.

The Bø gabbro is also assumed to have intruded the granulite facies rocks after the period of granulite facies metamorphism. Escaping gases from the crystallizing gabbro caused the retrograde metamorphism of the surrounding rocks.

The temperatures of the rock forming processes may be arrived at by using the two feldspar geothermometer (Barth, 1956). The temperatures are given in Table 39.

The metamorphic boundary which separates the amphibolite facies rocks (to the east) and granulite facies rocks (to the west) is found to be coincident with a thrust plane beneath the anorthosite southeast of Eidsfjorden (Fig. 10). This has led to the assumption that the observable boundary is related to the tectonics of the area.

Imbricate thrust sheets are believed to be important tectonical features in the whole of the Lofoten—Vesterålen area. The horizontal movements of the thrusts are probably not very significant as far as can be deduced from the similarity in rock types on both sides of the facies boundary on Langøy. The fact that the thrust plane coincides with changes in regional metamorphic facies does not necessarily imply long distances of transfer along the thrust plane.

Chemical data.

A number of chemical analyses both for major and trace elements has been carried out during this work with two chief purposes in mind:

1. To investigate whether or not the element concentrations could be related to the origin of the rocks.
2. To investigate how variations in metamorphic grade of related rocks affect the concentrations of various elements.

The volatile character of the elements Na, K, Rb, Cs, Pb, Tl, and

Cu makes them unfit as indicators of the origin of rock terranes, as their concentrations may be significantly affected by metamorphism. Therefore, for the first purpose attention was especially directed towards the involatile elements V, Ni, Co, and Cr, and for the second purpose especially towards the volatile elements. It must be concluded that, concerning the origin of the rocks in the area studied, this line of approach failed.

Variation of element concentrations with metamorphic grade.

Concentrations of some elements of the rocks, show variations related to the change of metamorphic facies. *The Na/K ratios* are highest in the granulite facies rocks mainly because of variations in the Na concentrations. *High K/Rb* ratios were observed for the rocks from the Eidet—Hovden intrusion (except for the most highly fractionated dyke, no. 60/56). High K/Rb ratios are also characteristic of the western monzonites and the banded series. None of the Langøy rocks show significant enrichment in Rb.

Cesium which was detected only in the biotites, is significantly higher in the amphibolite facies than in the granulite facies micas. The K/Cs ratios in biotite vary by a factor of 2 to 4 in passing from amphibolite facies to granulite facies rocks; possibly the K/Cs ratio is a good indicator of changing PT conditions in regional metamorphism.

Lead. Variations in Pb in the rocks cannot be correlated to the variations in metamorphic grade. Pb was not even detected in the biotites. This is contrary to observations of Wedepohl (1956), but in harmony with data given by Howie (1955) for the Madras charnockites. In the Langøy rocks Pb is somewhat higher in the plagioclases than in the coexisting potash feldspars, and the distribution coefficient of Pb between the two minerals in equilibrium may be used as a geological thermometer (Table 45).

Thallium was determined only in the potash feldspars and it seems that high Tl contents are associated with a high temperature of formation of these feldspars in the Langøy rocks.

Lithium. Biotites from the amphibolite facies rocks are higher in Li than those from granulite facies. And even though biotite is more abundant in the former, Li is essentially the same in both rock types (Figs. 41 and 42). This may be explained by the plagioclases being higher in Li in the granulite facies rocks.

Strontium. The potash feldspars from the amphibolite facies rocks tend, on an average, to be lower in Sr than those from the granulite facies rocks. The data obtained demonstrate that the distribution of Sr between plagioclase and coexisting potash feldspars also may be used as a geological thermometer (Fig. 44 b).

Barium is also apparently relatively enriched in the high temperature potash feldspars. Figs 45 a and b show plots of the K/Ba ratios of potash feldspars plotted against their temperatures of formation as determined by the feldspar geological thermometer. Two parallel curves result. The one is for the igneous rocks comprising the Eidet—Hovden rocks and the young granites. The other is for the metamorphosed rocks. The latter is based on the ratios of the granulite facies feldspars which fit nicely on the curve. The plots of the amphibolite facies feldspars are distributed randomly. This may be because of the introduction of biotite as an aditional potassium mineral phase. The distribution of Ba between minerals in equilibrium may also be used as a geothermometer.

Copper. There is some indication that Cu is removed from rocks during regional metamorphism. However, the tendency cannot be seen to be any more pronounced in ~~the granulite than~~ amphibolite facies rocks.

Other elements determined in the rocks do not seem to vary with changing external PT conditions.

The age of the rocks.

The age considerations will be confined to a discussion of pre-Cambrian vs. Caledonian age possibilities. Ages younger than Caledonian can be excluded safely for the regionally metamorphosed rocks; the post orogenic intrusive rocks may, of course, be of a younger age.

No absolute age determinations have been made on Langøy rocks. The author has made ready for separation a number of samples containing biotite and potash feldspars from the various Langøy rocks, and it is hoped the K/A and Rb/Sr ages can be determined from these in the near future. It may seem premature therefore to discuss the age at this stage. However, the present is a good time to review the strictly geological data known.

A Caledonian age is generally based on two arguments. One is the concept that the graphite schists and marbles on Langøy are Cale-

donian sedimentary rocks. Support for this argument is found in the common occurrence of similar sedimentary rocks with mica schists of undoubted Caledonian age on the nearby mainland. Skjeseth (personal communication) who has studied the graphite schists on Langøy (Jennestad), and on the mainland, strongly advocates the Caledonian (and probably Cambrian) age of these rocks on Langøy. He has also informed the writer about data obtained from uranium prospecting done by the Geological Survey of Norway, in which it has been shown that the Cambrian carbonaceous shales contain appreciable amounts of uranium, whereas definitely pre-Cambrian graphite schists are very low in uranium. The graphite schists on Langøy are comparatively high in uranium.

Obviously identification of the sedimentary rocks on Langøy as of Caledonian age rests strictly on analogy and can by no means be considered proved.

Graphite is sometimes present in the charnockites originally described by Holland (1900). Ramberg (1948 b) describes graphite bearing gneisses and coarse grained marbles from the granulite facies Isortoq complex on Greenland of apparent pre-Cambrian age. Wilson (1957) mentions graphite in pre-Cambrian granulite facies rocks from South-Western Australia. On the whole graphite schists and gneisses containing graphite seem to be rather common in granulite facies areas.

The second and strongest argument for a Caledonian age is based on the general conception of the evolution of the Caledonian orogeny in Norway, i. e., increasing metamorphism of the rocks from the east to the west with deeper erosion levels in the west than in the east. The gneisses typical of the west coast of Norway would be according to this view, regarded as an intimate mixture of pre-Cambrian gneisses and Caledonian sediments, lavas and intrusives all metamorphosed during Caledonian time. Thus the mineral parageneses and structures of the rocks as now seen would be results of the Caledonian orogenic forces and the rocks just as much Caledonian as the Cambrian sediments. The increasing metamorphism from amphibolite to granulite facies in going west on Langøy fits very nicely with the regional westward increase in metamorphism in the same direction during the Caledonian orogeny; the geographical position of the Lofoten and Vesterålen rocks corresponds to an extreme western position in the Caledonian orogenic belt. The basal gneisses on Hinnøy are believed to have formed through processes as outlined above, and they are both

in appearance and in mineral facies exactly similar to the amphibolite facies veined gneisses in the east of Langøy. The same type of metasedimentary rocks occur in both. In the author's opinion the charnockite border series and the banded series which also contain metasedimentary rocks of the same type as those in the amphibolite facies gneisses cannot be distinguished from the gneisses in the east on any basis other than grade of metamorphism. But Herr and Merz (1958) using the Re/Os method determined the age of a molybdenite in the Svolvær granite from Vatterfjord on Austvågøy in the Lofoten—Vesterålen group as 2.3×10^9 years.[1]

This is a strong argument for a pre-Cambrian age of the rocks on Langøy. However, the possibility should be considered as to whether this could be a relict age of pre-Cambrian rocks now intermixed in the Caledonian gneisses. It appears quite likely that rocks belonging to the pre-Cambrian basement of the Caledonian geosyncline and those representing the Caledonian sediments, lavas, and intrusives, were intimately interfolded. This of course, raises the question as to whether a pre-Cambrian molybdenite could possibly have remained essentially unaffected in the pre-Cambrian rocks when, and if these recrystallized during the later Caledonian orogeny.

One geological interpretation on Langøy suggests two periods of metamorphism. The red augengneisses in the amphibolite facies are regarded to be metamorphic rocks related to the monzonites east of Eidsfjorden; these, in turn, are considered to be unmetamorphosed igneous rocks similar to the Eidet—Hovden monzonites. Thus it is possible that a pre-Cambrian metamorphism under granulite facies conditions followed by intrusion of the Eidet—Hovden rocks and the igneous rocks east of Eidsfjorden was followed by a Caledonian metamorphism under conditions of the amphibolite facies which affected only the eastern part of Langøy. On the other hand, the two periods of metamorphism may have occurred during different stages of the same major orogeny.

The concept that all the metamorphic rocks on Langøy were recrystallized during the Caledonian orogeny is the simplest and most easily conceiveable picture of the geological history. It also is in best agreement with the general ideas of the regional geology. Considering

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[1] This age is comparable to the 2.8×10^9 years found for the charnockites from Western Australia (Wilson, 1957) using the K/A method.

this to be correct, it is possible to fit the observed data into the following chronological sequence of events.

1. Deposition of Caledonian sediments upon a "basement" of pre-Cambrian rocks (2.3×10^9 years old) and local igneous activity.
2. Early Caledonian orogenic activity with regional metamorphism of both the sediments and the basement rocks.
3. Formation of palingenic granitic magmas and local migration of these magmas to form masses such as the young red granites.
4. After this metamorphism, intrusion of the Eidet—Hovden gabbros and monzonites, and the anorthosites and associated monzonites southeast of Eidsfjorden.
5. Metamorphism affecting only the eastern Langøy rocks. The red augengneisses were recrystallized during this period of metamorphism.
6. Intrusion of the Bø gabbro with consequent retrograde metamorphism of the gneisses now exposed along the west side of Jørgenfjord.
7. Regional thrusting of rocks from the west towards the east.

Sammendrag.

Petrologi og geokjemi av høymetamorfe og eruptive bergarter på Langøy, Nord-Norge.

Det foreliggende arbeid meddeler feltgeologiske og laboratorie-resultater av forfatterens undersøkelser av bergartene på Langøy i Vesterålen. Feltarbeidet ble utført somrene 1955, 1956, 1958, tilsammen fem måneder. Laboratoriearbeidet ble utført i Mineralogisk-Geologisk Museum, Oslo (1955—1956) og i Department of Geology and Mineralogy, Oxford (1956—1958). Bearbeidelsen av resultatene ble utført i Mineralogisk-Geologisk Museum, Oslo (1958—1959). Hensikten med arbeidet var å skaffe klarhet over hvilke bergarter som opptrer og hvilke prosesser som har ført til deres dannelse.

Avhandlingen er delt i fem hoveddeler.

1. Innledning (Introduction) hvor områdets beliggenhet, klima, kommunikasjoner osv. er beskrevet, og med en kort oversikt over tidligere og eget utført geologisk arbeid innen området, samt en kort oversikt over områdets alminnelige geologi.
2. Petrografi (Petrography) hvor de forskjellige bergartene som opptrer er beskrevet i noen detalj. Dette kapitelet danner et vesentlig grunnlag for den etterfølgende diskusjon.
3. De metamorfe facies (The metamorphic facies) hvor de forskjellige bergartstypers forhold til de regional metamorfe prosesser er diskutert. Variasjonene i de stabile mineralene under vekslende PTX betingelser er spesielt diskutert, særlig variasjonene innen alkali feltspatene.
4. Geokjemi (Geochemistry) hvor bergartenes (og enkelte mineralers) kjemi er diskutert og hvor det er undersøkt om variasjonene i elementkonsentrasjonene kan kaste klarhet over hvilke prosesser som har ført til bergartenes dannelse.

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5. Petrologi (Petrology) hvor de data som har blitt presentert og diskutert i de foregående kapitler er brukt til å diskutere petrogenesen.

Avhandlingen avsluttes med et kort sammendrag hvor også områdets geologiske alder er diskutert.

Et geologisk kart over det undersøkte området er gitt på Fig. 3. Betegnelsen "Lofot-eruptiver" er tidligere benyttet som et gruppenavn for bergartene innen dette området. Det foreliggende arbeid viser at disse bergartene for største delen utgjøres av en serie av høymetamorfe bergarter hvori enkelte eruptive bergarter har blitt intrudert.

"Bunn-gneisene" på Hinnøy fortsetter over på nordøst siden av Langøy (amphibolite facies veined gneisses). Metamorfose graden tiltar mot vest og amfibolit facies gneisene går over i lignende gneiser i granulit facies. Facies grensen er inntegnet på kartet, Fig. 3. Den er funnet i fortsettelsen etter strøket av skyveplanet under en anorthosit som er skjøvet over amfibolit facies bergarter øst for Eidsfjorden. Denne forbindelsen mellom facies grensen og skyveplan indikerer at facies grensen er tektonisk betinget og er av forkastnings eller overskyvnings natur. Gneisene på hver side av grensen er meget like av utseende (Figs. 5—8) og lignende metamorfe sedimentære bergarter opptrer i begge. Det er derfor antatt at gneisene på hver side av facies grensen tilhører den samme geologiske enhet og at overskyvningene ikke har foregått over store avstander. Granulit facies gneisene like vest av facies grensen (the charnockite border series) inntar også mineralogisk en stilling mellom amfibolit facies gneisene og de lignende granulit facies gneisene lenger vest (the banded series).

Alle disse tre gneisseriene er åre eller bånd gneiser (Figs. 5—8). De sikre metasedimentære bergarter som opptrer i dem alle er: grafitskifre, kalkstener (marmor) og kvartsbåndete jernmalmer. Kvartsitiske bergarter kan gjenkjennes innen amfibolit facies gneisene, men er som regel sterkt feltspatførende og kan ikke med sikkerhet skilles fra de vanlige granuliter i granulit facies området. De sedimentære bergartenes foliasjon er alltid parallelt med gneisenes bånding. Hele komplekset av gneiser med de metamorfe sedimentære bergarter er antatt å representere regional metamorfe opprinnelige geosynklinal sedimenter og lavaer.

Massive monzonitiske bergarter med store alkali feltspat porfyroblaster opptrer vestligst på Langøy (porphyroblastic monzonitic granulites). De har en granulit facies mineral paragnese og er meta-

morfe bergarter. De er kjemisk lik enkelte bånd innen bånd gneisene. Endel data kan tyde på at de er dannet ved metamorfe homogeniseringsprosesser av de tilgrensende bånd gneiser, men det er også mulig at de er monzonitiske intrusjoner som er metamorfosert sammen med gneisene.

På vestsiden av Eidsfjorden opptrer innen et lite område en grovkornet rødlig granit (old red granite). Den har migmatiserings kontakter mot de omgivende gneisene. Dette er også en metamorf bergart med en granulit facies mineral paragnese som tilsvarer de omgivende gneisenes.

Røde graniter av en annen type (young red granites) opptrer ganske hyppig innen granulit facies området. Som oftest utgjør de bare små partier på noen meters omkrets, men de kan og forekomme i større massiver (Fig. 3). Mineralogisk er de dominert av kvarts og feltspat. Disse granitene er forklart som krystallisasjons produkter av palingene magmaer.

Halvøya mellom Åsenfjord og Malnesfjord oppbygges vesentlig av olivin gabbro med noen mindre monzonitiske partier (Eidet—Hovden intrusion). Både gabbroen og monzonitene er mineralogisk forskjellige fra de omgivende metamorfe bergartene, og enkelte data indikerer en post metamorf dannelse av intrusjonen.

En stor forekomst av titan-magnetit opptrer i gabbroen ved Selvåg. Gabbroen varierer kjemisk og mineralogisk innen ganske vide grenser. Variasjonene tyder på at gabbroen begynte å krystallisere ved grensen mot de omgivende gneisene.

Monzonitiske ganger opptrer hyppig i gabbroens sydlige del. Anorthositen og monzoniten vest av den metamorfe facies grensen (monzonites east of Eidsfjorden) på øst siden av Eidsfjorden har mange likhetspunkter med Eidet—Hovden intrusjonens bergarter, men er ikke studert i detalj.

De rød øyegneisene på østsiden av facies grensen (red augengneisses) er for største delen av monzonitisk sammensetning, og er antatt dannet under regional metamorfose i amfibolit facies av monzonitene like vest av dem.

Gneissonen som strekker seg langs vestsiden av Jørgenfjord, mellom Bø og Kråkberget, representerer en retrograd metamorf bergartsserie (the retrograde metamorphic gneisses). Bergartene som inngår i denne sonen er for største delen omvandlet av de omgivende porfyroblastiske monzonitene og åre gneisene fra "the banded series".

Sillimanit gneiser, som er ukjent fra andre steder på Langøy, fore-kommer i denne sonen langs vestsiden av Gåslandsvann. Den retro-grade metamorfosen antaes å ha funnet sted ved at gasser trengte gjennom bergartene på et tidspunkt etter regional metamorfosen i granulit facies. Det antaes å være en sammenheng mellom den retro-grade metamorfosen og intrusjonen av gabbroen sydligst i Bø (the Bø gabbro).

Temperaturene for de bergartsdannende prosessene kan anslåes ved hjelp av feltspat geotermometeret (Barth, 1956), og er oppført i Tabell 39.

De kjemiske analysene ble utført vesentlig med hensikt på å undersøke:

1) hvorvidt bergartenes kjemiske sammensetning kunne benyttes som indisium for deres opprinnelige dannelsesmåte,

2) hvorvidt variasjonene i metamorfose grad av beslektede berg-arter innvirker på element konsentrasjonene.

Det ble funnet meget vanskelig å trekke noen forbindelse mellom bergartenes kjemiske sammensetning og deres dannelsesmåte, men enkelte elementer (Li, Na, K, Rb, Cs, Pb, Tl, Sr, Ba, Cu) viser varia-sjoner som kan forbindes med vekslende metamorfose grad.

Alderen av bergartene og tidspunktet for de forskjellige bergart dannende prosesser på Langøy er vanskelig å etablere på grunnlag av de foreliggende data.

Bergartenes tilknytning til "bunn gneis" komplekset på Hinnøy og den tiltagende metamorfose mot vest på Langøy indikerer at berg-artene fikk sitt nuværende preg under den Kaledonske fjellkjede fold-ningen. Imidlertid er en pre-Kambrisk alder av bergartene indikert ved en radioaktiv aldersbestemmelse (Re/Os) på molybdenglans i Svolvær graniten ved Vatterfjord på Austvågøy (Svolvær graniten har tidligere vært innfattet i begrepet Lofot-eruptiver). Denne alderen er gitt av Herr og Merz (1958) til 2.3×10^9 år. Inntil denne alderen er bekreftet av andre radioaktive alders bestemmelser innen området kan en ikke se bort fra muligheten for at den representerer en relikt alder av pre-Kambriske bergarter som er sammenblandet med de Kaledonske gneisene. Det synes å være meget sannsynlig at bergarter som tilhørte det pre-Kambriske underlag og de Kaledonske sedimenter, lavaer og intrusiver er sterkt sammenblandet i disse områdene.

Ut fra et slikt syn er det mulig å oppstille følgende kronologiske skjema for den geologiske utviklingen på Langøy.

1. Avsetning av Kaledonske sedimenter på et underlag av pre-Kambriske bergarter (2.3 × 10^9 år gamle) og lokal magmatisk virksomhet.

2. Tidlig Kaledonsk fjellkjede foldning med regional metamorfose av både Kaledonske og underlagets bergarter.

3. Dannelse av palingene granitiske magmaer. De unge rød granitene ble dannet da magmaene krystalliserte.

4. Intrusjon av Eidet—Hovden bergartene og anorthositen med tilgrensende monzonit øst av Eidsfjorden.

5. Metamorfose av bergartene østligst på Langøy. De rød øyegneisene rekrystalliserte under denne metamorfosen.

6. Intrusion av Bø gabbroen med retrograd metamorfose av gneisene langs vest siden av Jørgenfjorden.

7. Regional overskyvning av bergarter fra vest mot øst.

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Attachments.

Eidet-Hovden intrusion.

1 (an. 16), olivine gabbro, Selvåg, close to the fjord.
2 (an. 8), olivine gabbro, roadcut south of Kobviken.
3 (32/56 d), gabbro, between Selvågtind 570 and 564.
4 (an. 15), monzonite, Selvågtind.
5 (an. 9), monzonite, roadcut, Tussen.
6 (50/56), monzonitic dyke in gabbro, Spikerheia.
7 (91/56a), monzonitic dyke in gabbro, roadcut, Gustad.
8 (60/56), "granitic" dyke in gabbro, between Solheim and Gustad.

Porphyroblastic monzonitic granulites.

9 (57/56), roadcut, Skårvåg.
10 (55/56), roadcut below top 115, Steine.

Retrograde metamorphosed gneisses.

11 (318/55), monzonitic gneiss, 500 m. north of Ramberg.
12 (310/55), monzonitic gneiss, north of Strandskjærene, Maars.
13 (235/55), coarse grained red granitic rock, west of Klaksjorden school.

The banded series.

14 (58/56), "dioritic" granulite, roadcut at Solheim.
15 (145/56), "dioritic" granulite, top 228, north of Rekvann.
16 (an. 12), "monzonitic" granulite, roadcut, Rotteåsen.
17 (59/56), "quartz dioritic" granulite, roadcut at Solheim.

Rocks associated with metamorphic sedimentary rocks in the banded series.

18 (81/56), amphibolite, Kråkberget.
19 (77/56), amphibolite, Jørgenfjord.
20 (70/56), monzonitic granulite, impregnated with pyrrhotite, Jørgenfjord.

The charnockite border series.

21 (151/56b), dark "basic" gneiss, roadcut, small lake north of Storvann, Frøskeland, Steinlandsfjord.

22 (150/56b), dark "basic" gneiss, west of top 547, between Storvann and Gåsfjord.

23 (158/56), light "acid" gneiss, roadcut, 1000 m. south of Storvann, road between Frøskeland and Steinlandsfjord.

24 (142/56), light gneiss ("enderbitic"), roadcut, Sildpollen.

25 (150/56a), light "acid" gneiss, west of top 547, between Storvann and Gåsfjord.

26 (151/56a), light gneiss ("enderbitic"), small lake north of Storvann.

The amphibolite facies veined gneisses.

27 (3/56b), amphibolitic band in gneiss, Lafjellet.

28 (147/56b), amphibolitic band in gneiss, 120 m. a. s. l., Gåsebøl.

29 (147/56a), light "quartz monzonitic" band in gneiss, 120 m. a. s. l., Gåsebøl.

30 (an. 1), gray "granodioritic" gneiss, quarry, Jægtbøgen.

31 (3/56a), light "granodioritic" band in gneiss, Lafjellet.

Rocks associated with metamorphic sedimentary rocks in the amphibolite facies veined gneisses.

32 (13/56), amphibolite in contact with crystalline limestone, close to **Prestetjern.**

33 (29/55), amphibolite in gneiss, between Jennestad and top 70.

34 (33/55j), amphibolite in contact with crystalline limestone, Kovvann.

35 (43/55), quartz-plagioclase-biotite-schist, Vik.

Young red granites.

36 (71/55), large granite massif, Stenheia.

37 (an. 5), large granite massif, Torset.

38 (347/55), small granite body, n. Rygge.

39 (211/55), small granite body, west of Eidspollen.

40 (208/55), small granite body, Rissjøen.

41 (282/55), fine grained granite dyke in porphyroblastic monzonite, Søberg.

42 (254/55), minor granite massif, Veatind-Nåla.

43 (270/55), small granite body, Bergodden, Skårvåg.

Table III, IV and V.

Rock types and localities.

Eidet-Hovden intrusion.

Group 1, 50, (145/55), monzonite, top 570, Selvågtind.

      51, (an. 9), monzonite, roadcut, Tussen.

      52, (an. 9), monzonite, roadcut, Tussen.

      53, (179/55), monzonite, (syenite), Glimmerbugten.

Group 2, 54, (an. 7), pegmatite, between Selvågtind and Trollan.

Porphyroblastic monzonitic granulites.

Group 3, 55, (271/55), roadcut, southwest of top 33, Skårvåg.
   56, 80, 90, (293/55), west side of peninsula at Vinje.
   57, (255/55), Nåla.
   58, 81, (251/55), Veatind.

Retrograde metamorphosed gneisses.

Group 4, 59, (318/55), "monzonitic gneiss", roadcut, 500 m. north of Ramberg.
   60, 82, (238/55), "quartz monzonitic gneiss", between Oppsåte and
    Jørgenfjord.
   61, 83, 91, (224/55), "coarse monzonite", up towards Ravnåsen.

The banded series.

Group 5, 62, 84, (an. 12), "monzonitic" rock, roadcut, Rotteåsen.
   63, (348/55), "granodioritic charnockite", south end of Ryggepollen.

Old red granite.

Group 6, 64, (65/56), east of Troldøen, Eidsfjord.

The charnockite border series.

Group 7, 65, 85, 92, (25/55), "granodioritic charnockite", roadcut, Frøskeland.
   66, 93, (150/56), light "acid" gneiss, west of top 547, Storvann-
    Gåsfjord.

Amphibolite facies veined gneisses.

Group 8, 67, 94, (147/56), light band in gneiss, 120 m. a. s. l., Gåsebøl.
   68, (3/56), light band in gneiss, Lafjellet.
   69, (35/55), light "granodioritic" gneiss, Ånstadblåheia.
   70, 86, 95, (37/55), light "granodioritic" gneiss, Karihaugen.
   71, 87, 96, (an. 1), light "granodioritic" gneiss, quarry, Jægtbøgen.

Red augengneisses.

Group 9, 72, (109/56), "monzonitic" augengneiss, Kvalsøfjell.
   73, (99/56), "granodioritic" gneiss, Rishatten.
   74, (Evatn), red granitic-pegmatitic vein in gneiss, Evatn.

Young red granite.

   75, (an. 5), large granite massif, Torset.

Table I.

Chemical analyses (a, weight per cent; b, cation per cent)
of the Langøy rocks. (*, spectrographic analyses).
Major elements.

	a 1 b (an.16)		a 2 b (an. 8.)		a 3 b (32/56 d)		a 4 b (an.15)		a 5 b (an. 9)	
SiO_2	46.65	42.5	48.72	44.1	46.78	42.5	54.88	51.1	60.59	55.4
TiO_2	0.13	0.1	0.23	0.2	0.69	0.5	1.67	1.2	0.75	0.5
$AlO_{1.5}$	23.07	24.8	20.56	22.0	16.18	17.3	17.17	18.8	19.43	20.9
$FeO_{1.5}$	1.31	0.9	0.30	0.2	0.69	0.5	3.72	2.6	1.22	0.8
FeO	4.73	3.6	4.68	3.5	10.98	8.3	4.69	3.6	2.40	1.5
MnO	0.07	0.1	0.08	0.1	0.12	0.1	0.20	0.2	0.08	0.1
MgO	9.01	12.3	9.81	13.3	12.65	17.2	1.96	2.7	0.68	0.9
CaO	10.69	10.5	11.83	11.5	7.36	7.1	5.51	5.5	3.38	3.3
$NaO_{0.5}$	2.69	4.8	2.60	4.6	3.11	5.4	4.84	8.7	5.86	10.4
$KO_{0.5}$	0.38	0.4	0.40	0.5	0.67	0.8	3.98	4.7	5.14	6.0
H_2O^-	0.16		0.13		0.08		0.19		0.17	
H_2O^+	0.97		0.37		0.39		0.05		0.01	
CO_2	0.55	(0.7)	0.30	(0.4)	–		n.d.		n.d.	
$PO_{2.5}$	–		–		0.35	0.3	1.17	0.9	0.32	0.2
Sum:	100.41		100.01		100.05		100.03		100.03	

	a 6 b (50/56)		a 7 b (91/56 a)		a 8 b (60/56)		a 9 b (57/56)		a 10 b (55/56)	
SiO_2	54.81	50.8	55.26	51.2	69.62	64.8	56.38	51.8	57.46	52.6
TiO_2	1.03	0.7	1.07	0.7	0.52	0.4	0.87	0.6	0.80	0.5
$AlO_{1.5}$	16.69	18.1	17.01	18.6	13.91	15.3	18.86	20.4	20.02	21.6
$FeO_{1.5}$	4.21	2.9	4.13	2.9	1.41	1.0	3.15	2.2	2.53	1.8
FeO	5.25	4.1	4.93	3.8	1.75	1.3	3.78	2.9	2.64	2.0
MnO	0.20	0.2	0.16	0.1	0.04	–	0.11	0.1	0.09	0.1
MgO	3.16	4.4	2.38	3.3	0.55	0.8	2.46	3.4	1.38	1.9
CaO	5.58	5.6	4.93	4.9	1.64	1.6	5.53	5.5	4.81	4.7
$NaO_{0.5}$	5.00	9.0	5.35	9.6	4.24	7.7	5.22	9.3	5.55	9.8
$KO_{0.5}$	3.03	3.6	3.51	4.2	5.82	6.9	2.96	3.5	4.00	4.7
H_2O^-	0.13		0.18		0.17		0.08		0.25	
H_2O^+	0.17		0.15		0.12		0.19		0.20	
CO_2	n.d.		n.d.		n.d.		n.d.		n.d.	
$PO_{2.5}$	0.73	0.6	0.88	0.7	0.12	0.1	0.38	0.3	0.45	0.3
Sum:	99.99		99.94		99.91		99.97		100.18	

	a 11* b (318/55)		a 12* b (310/55)		a 13* b (235/55)		a 14 b (58/56)		a 15 b (145/56)	
SiO_2	56.7	52.9	57.4	52.9	64.7	60.3	48.87	46.1	51.85	47.5
TiO_2	0.92	0.7	0.83	0.6	0.22	0.2	2.14	1.5	0.69	0.5
$AlO_{1.5}$	16.9	18.6	19.8	21.4	16.1	17.7	16.92	18.8	13.23	14.3
$FeO_{1.5}$	3.52	2.5	1.72	1.2	2.46	1.7	5.23	3.7	0.85	0.6
FeO	4.30	3.4	3.60	2.8	1.00	0.8	8.86	7.0	9.25	7.1
MnO	0.17	0.1	0.09	0.1	0.04	0.1	0.14	0.1	0.15	0.1

Table I. (cont.).

	a 11* b (318/55)		a 12* b (310/55)		a 13* b (235/55)		a 14 b (58/56)		a 15 b (145/56)	
MgO	2.16	3.0	1.66	2.3	0.41	0.6	4.55	6.4	11.39	15.7
CaO	4.62	4.7	4.76	4.7	2.66	2.7	6.68	6.7	8.89	8.8
$NaO_{0.5}$ ..	5.03	9.1	5.34	9.5	2.83	5.1	4.30	7.9	2.35	4.2
$KO_{0.5}$ ...	4.19	5.0	3.85	4.5	9.05	10.8	0.87	1.1	0.81	0.9
H_2O^- ...	n.d.		n.d.		n.d.		0.10		0.14	
H_2O^+ ...	n.d.		n.d.		n.d.		0.23		0.23	
CO_2	n.d.		n.d.		n.d.		n.d.		–	
$PO_{2.5}$ ...	n.d.		n.d.		n.d.		0.94	0.7	0.34	0.3
Sum:							99.83		100.17	

	a 16 b (an. 12)		a 17 b (59/56)		a 18* b (81/56)		a 19* b (77/56)		a 20* b (70/56)	
SiO_2.....	54.62	50.4	59.27	54.5	42.5	40.8	46.2	43.5	58.9	54.4
TiO_2	0.95	0.6	0.42	0.3	1.67	1.2	1.50	1.1	0.33	0.2
$AlO_{1.5}$ ...	18.54	20.1	17.80	19.2	11.2	12.7	13.6	15.1	17.9	19.4
$FeO_{1.5}$ ..	3.60	2.5	1.56	1.1	7.15	5.1	4.98	3.5	0.10	0.1
FeO	4.51	3.5	5.28	4.0	11.82	9.5	8.65	6.8	5.44	4.2
MnO	0.15	0.1	0.11	0.1	0.18	0.2	0.22	0.2	0.13	0.1
MgO	2.96	4.1	4.40	6.1	7.46	10.8	6.63	9.4	2.98	4.2
CaO	6.07	6.0	4.99	4.9	12.73	13.1	11.75	11.9	2.80	2.8
$NaO_{0.5}$ ..	4.91	8.8	4.68	8.3	3.11	5.8	4.07	7.4	4.84	8.6
$KO_{0.5}$ ...	3.00	3.5	1.17	1.4	0.67	0.8	0.87	1.1	5.10	6.0
H_2O^- ...	0.15		0.17		n.d.		n.d.		n.d.	
H_2O^+ ...	0.18		0.18		n.d.		n.d.		n.d.	
CO_2	n.d.		n.d.		n.d.		n.d.		n.d.	
$PO_{2.5}$ ...	0.51	0.4	0.13	0.1	n.d.		n.d.		n.d.	
Sum:	100.15		100.16							

	a 21 b (151/56 b)		a 22 b (150/56 b)		a 23 b (158/56)		a 24 b (142/56)		a 25 b (150/56 a)	
SiO_2.....	49.20	45.4	51.37	47.5	57.44	53.2	62.79	58.1	67.98	63.2
TiO_2	1.16	0.8	1.11	0.8	0.83	0.6	0.33	0.2	0.43	0.3
$AlO_{1.5}$ ...	15.02	16.4	14.94	16.2	18.33	19.9	17.52	19.1	16.06	17.6
$FeO_{1.5}$ ..	1.19	0.8	0.92	0.7	1.46	1.0	1.28	0.9	1.24	0.9
FeO	9.08	7.0	8.36	6.4	7.08	5.4	4.23	3.3	2.42	1.9
MnO	0.22	0.2	0.14	0.1	0.13	0.1	0.07	0.1	0.02	–
MgO	10.26	14.3	11.08	15.3	4.05	5.6	2.32	3.2	1.47	2.1
CaO	8.52	8.4	7.71	7.6	3.76	3.7	3.00	3.0	1.84	1.8
$NaO_{0.5}$ ..	2.37	4.2	1.41	2.5	3.92	7.0	4.06	7.3	3.60	6.5
$KO_{0.5}$ ...	1.58	1.9	1.99	2.3	2.89	3.4	3.96	4.7	4.73	5.6
H_2O^- ...	0.13		0.13		0.14		0.12		0.18	
H_2O^+ ...	0.42		0.46		0.42		0.22		0.32	
CO_2	–		n.d.		–		n.d.		–	
$PO_{2.5}$ ...	0.70	0.6	0.79	0.6	0.04	0.1	0.08	0.1	0.10	0.1
Sum:	99.85		100.41		100.49		99.98		100.39	

Table I (cont.).

	a 26 b (151/56a)		a 27 b (3/56b)		a 28 b (147/56b)		a 29 b (147/56a)		a 30 b (an.1)	
SiO_2.....	72.26	67.7	46.87	44.2	49.45	45.9	61.80	57.1	67.58	63.3
TiO_2	0.22	0.2	1.81	1.3	0.72	0.5	0.40	0.3	0.36	0.3
$AlO_{1.5}$ ...	14.41	15.9	16.75	18.6	17.13	18.8	18.91	20.6	15.69	17.2
$FeO_{1.5}$ ..	0.52	0.4	4.45	3.2	2.87	2.0	1.00	0.7	0.50	0.3
FeO.....	1.26	1.0	9.29	7.3	7.15	5.5	3.49	2.7	3.65	2.8
MnO	0.03	–	0.15	0.1	0.14	0.1	0.01	–	0.02	–
MgO	0.69	1.0	5.55	7.9	7.24	10.1	1.68	2.3	1.78	2.5
CaO	2.03	2.0	7.69	7.7	9.64	9.6	2.88	2.8	2.11	2.1
$NaO_{0.5}$ ..	3.26	5.9	3.55	6.5	3.48	6.3	4.37	7.8	3.89	7.0
$KO_{0.5}$ ...	4.83	5.8	2.26	2.7	0.89	1.1	4.45	5.3	3.79	4.5
H_2O^- ...	0.05		0.17		0.15		0.08		0.08	
H_2O^+ ...	0.24		0.99		0.99		0.67		0.48	
CO_2	–		–		–		–		0.08	(0.1)
$PO_{2.5}$ ...	0.12	0.1	0.70	0.5	0.11	0.1	0.55	0.4	tr.	
Sum:....	99.92		100.23		99.96		100.29		100.01	

	a 31 b (3/56a)		a 32* b (13/56)		a 33* b (29/55)		a 34* b (33/55j)		a 35* b (43/55)	
SiO_2.....	73.37	69.0	45.6	42.8	45.9	44.4	48.7	45.4	58.3	54.1
TiO_2	0.29	0.2	1.17	0.8	2.17	1.6	1.17	0.8	0.83	0.6
$AlO_{1.5}$ ...	14.37	15.9	15.2	16.8	12.8	14.6	17.7	19.4	13.9	15.2
$FeO_{1.5}$ ..	0.58	0.4	3.97	2.8	7.62	5.5	5.32	3.7	1.86	1.3
FeO.....	1.52	1.2	8.65	6.8	10.25	8.3	5.50	4.3	5.15	4.0
MnO	0.01	–	0.21	0.2	0.22	0.2	0.13	0.1	0.05	–
MgO	0.43	0.6	8.79	12.4	5.97	8.7	5.80	8.1	7.96	11.1
CaO	2.04	2.0	11.89	12.0	10.07	10.5	7.70	7.7	2.94	2.9
$NaO_{0.5}$ ..	3.23	5.9	2.60	4.8	2.84	5.4	4.50	8.1	3.11	5.6
$KO_{0.5}$ ...	3.96	4.7	0.45	0.6	0.67	0.8	2.02	2.4	4.44	5.2
H_2O^- ...	0.11		n.d.		n.d.		n.d.		n.d.	
H_2O^+ ...	0.13		n.d.		n.d.		n.d.		n.d.	
CO_2	–		n.d.		n.d.		n.d.		n.d.	
$PO_{2.5}$ ...	0.09	0.1	n.d.		n.d.		n.d.		n.d.	
Sum:....	100.13									

	a 36* b (71/55)		a 37 b (an.5)		a 38* b (347/55)		a 39* b (211/55)		a 40* b (208/55)	
SiO_2.....	69.9	64.6	70.90	67.0	72.9	68.2	73.4	68.9	74.0	69.3
TiO_2	0.20	0.2	0.22	0.2	0.22	0.2	0.20	0.1	0.12	0.1
$AlO_{1.5}$ ...	15.7	17.1	14.20	15.8	13.8	15.3	13.9	15.4	13.9	15.5
$FeO_{1.5}$ ..	1.53	1.1	0.84	0.6	0.94	0.7	0.73	0.6	0.61	0.5
FeO.....	0.55	0.4	1.05	0.9	0.57	0.4	0.50	0.4	0.35	0.3
MnO	0.02	–	0.03	–	0.01	–	0.01	–	0.01	–
MgO	0.41	0.6	0.17	0.2	0.50	0.7	0.23	0.3	0.07	0.1
CaO	1.12	1.1	0.42	0.5	0.84	0.8	1.12	1.1	1.26	1.3
$NaO_{0.5}$ ..	4.95	8.9	4.31	7.9	3.28	6.0	3.15	5.7	3.05	5.5
$KO_{0.5}$ ...	5.11	6.0	5.75	6.9	6.40	7.7	6.23	7.5	6.16	7.4
H_2O^- ...	n.d.		0.08		n.d.		n.d.		n.d.	
H_2O^+ ...	n.d.		0.03		n.d.		n.d.		n.d.	
CO_2	n.d.		n.d.		n.d.		n.d.		n.d.	
$PO_{2.5}$ ...	n.d.		–		n.d.		n.d.		n.d.	
Sum:....			98.00							

Table I (cont.).

	a 41 * b (282/55)		a 42 * b (254/55)		a 43 * b (270/55)		
SiO₂.....	73.8	69.3	74.1	69.4	75.6	71.2	
TiO₂	0.18	0.1	0.18	0.1	0.17	0.1	
AlO₁.₅ ...	13.7	15.0	13.6	15.0	13.1	14.4	
FeO₁.₅ ..	0.57	0.4	0.87	0.7	0.59	0.4	
FeO.....	0.90	0.7	0.50	0.4	0.50	0.4	
MnO	0.05	0.1	0.02	–	0.01	–	
MgO	0.08	0.1	0.07	0.1	0.07	0.1	
CaO	0.70	0.7	0.70	0.7	0.98	1.0	
NaO₀.₅ ..	3.63	6.6	3.71	6.8	3.49	6.4	
KO₀.₅ ...	5.83	7.0	5.70	6.8	5.00	6.0	
H₂O⁻ ...	n.d.		n.d.		n.d.		
H₂O⁺ ...	n.d.		n.d.		n.d.		
CO₂.....	n.d.		n.d.		n.d.		
PO₂.₅ ...	n.d.		n.d.		n.d.		

Analyst: nos. 1—2, 4—10, 14, 16—17, 30, 37, E. Christensen
nos. 3, 15, 21—29, 31, L. B. Bolkesjø
nos. 11—13, 18—20, 32—36, 38—43, K. S. Heier

Table II.

Chemical analyses of the Langøy

		1 an. 16	2 an. 8	3 32/56d	4 an. 15	5 an. 9	6 50/56	7 91/56a	8 60/56	9 57/56	10 55/56
Si^{4+}	0.39	217900	227600	218600	256400	283100	256100	258200	325300	263400	268500
Al^{3+}	0.57	122000	108800	85600	90900	102800	88300	90000	73600	99800	105900
Ga^{3+}	0.62	23	20	20	20	19	23	18	19	23	22
Ti^{4+}	0.64	780	1380	4140	10000	4500	6200	6400	3100	5200	4800
Cr^{3+}	0.64	46	231	541	10	tr.	80	40	32	31	29
V^{3+}	0.65	21	44	45	35	22	90	95	33	87	57
Fe^{3+}	0.67	9200	2100	4800	26000	8500	29400	28900	9900	22000	17700
Mg^{2+}	0.78	54300	59200	76300	11800	4100	19000	14300	3300	14800	8320
Li^{1+}	0.78	3.8	3.7	7	11	6	16	19	6	16	20
Ni^{2+}	0.78	272	169	345	(10)	–	31	14	29	23	11
Co^{2+}	0.82	54	51	79	18	8	25	23	10	25	16
Cu^{2+}~	0.83	45	83	67	24	35	100	61	37	37	29
Fe^{2+}	0.83	36800	36400	85300	36500	18700	40800	38300	13600	29400	20500
Mn^{2+}	0.91	540	620	930	1550	620	1550	1240	310	850	690
Na^{1+}	0.98	20000	19300	23100	35900	43500	37100	39700	31500	38700	41200
Ca^{2+}	1.06	76400	84600	52600	39400	24200	39900	35200	11700	39500	34400
Sr^{2+}	1.27	489	448	397	366	314	347	373	137	465	413
Pb^{2+}	1.32	tr.	tr.	tr.	10	11.7	15.2	16.3	18.9	19.4	20.6
K^{1+}	1.33	3150	3300	5560	33000	42700	25200	29100	48300	24600	33200
Ba^{2+}	1.43	160	175	290	2320	2200	1055	1200	380	830	1085
Rb^{1+}	1.49	(7)	(5)	(9)	37	45	47	48	280	52	92
Cs^{1+}	1.65	–	–	–	–	–	–	–	–	–	–

		22 150/56b	23 158/56	24 142/56	25 150/56a	26 151/56a	27 3/56b	28 147/56b	29 147/56a	30 an. 1	31 3/56a
Si^{4+}	0.39	240000	268400	293400	317600	337600	219000	231000	288700	315700	342800
Al^{3+}	0.57	79100	97000	92700	85000	76200	88600	90600	100100	83000	76000
Ga^{3+}	0.62	17	19	18	18	16	18	21	15	20	15
Ti^{4+}	0.64	6700	5000	2000	2600	1300	10900	4300	2400	2200	1700
Cr^{3+}	0.64	800	270	115	215	50	45	113	70	108	40
V^{3+}	0.65	107	85	47	70	45	140	1022	125	55	50
Fe^{3+}	0.67	6400	10200	9000	8700	3600	31100	20100	7000	3500	4100
Mg^{2+}	0.78	66800	24400	14000	8900	4200	33500	43700	10100	10700	2600
Li^{1+}	0.78	10	13	18	16	16	18	14	25	20	22
Ni^{2+}	0.78	336	95	51	18	18	41	136	52	40	25
Co^{2+}	0.82	53	27	22	12	11	48	144	27	13	14
Cu^{2+}~	0.83	24	55	43	16	11	34	40	38	40	22
Fe^{2+}	0.83	65000	55000	32900	18800	9800	72200	55600	27100	28400	11800
Mn^{2+}	0.91	1080	1000	540	150	230	1160	1080	80	155	80
Na^{1+}	0.98	10500	29100	30100	26700	24200	26300	25800	32400	28900	24000
Ca^{2+}	1.06	55100	26900	21400	13200	14500	55000	68900	20600	15100	14600
Sr^{2+}	1.27	380	330	348	325	295	385	285	319	230	315
Pb^{2+}	1.32	tr.	14.6	21.7	21.4	13.8	tr.	11.6	25.6	26.9	25.2
K^{1+}	1.33	16500	24000	32900	39300	40100	18800	7400	36900	31500	32900
Ba^{2+}	1.43	868	620	1007	1025	900	500	155	823	410	1275
Rb^{1+}	1.49	183	113	180	240	145	120	30	225	220	140
Cs^{1+}	1.65	–	–	–	–	–	–	–	–	–	–

Table II.

rocks (all elements in p.p.m.)

11 318/55	12 310/55	13 235/55	14 58/56	15 145/56	16 an. 12	17 59/56	18 81/56	19 77/56	20 70/56	21 151/56b
265000	268000	302000	228300	242200	255200	279000	199000	216000	275000	229900
89000	105000	85000	89500	70000	98100	94200	59000	72000	95000	79500
18	19	12	32	19	26	30	21	25.5	12	19
5500	5000	1300	12800	4100	5700	2500	10000	9000	2000	7000
(18)	(21)	40	58	1030	15	30	135	172	370	570
85	78	24	115	124	109	70	275	225	92	110
24600	11900	17200	36600	5900	25200	10900	50000	34800	700	8300
13000	10000	2500	27400	68700	17900	26500	45000	40000	18000	61900
21	13.6	10	11	9	18	14	6.5	3.9	9.2	10
13	13	(8)	40	136	14	90	230	135	37	280
18	15	6.5	54	51	25	31	70	49	26	47
20	22	3.8	78	10	34	70	300	52	210	14
33400	28000	7800	68900	71900	35100	41000	91900	67200	42300	70600
1325	720	300	1080	1200	1160	850	1400	1700	1000	1700
37300	39600	21000	31900	17400	36400	34700	23100	30200	35900	17600
33000	34000	19000	47700	63500	43400	35700	91000	84000	20000	60900
520	680	400	365	195	455	410	130	260	250	500
16	24.5	7.6	16.6	10.4	17.1	25	–	–	14.7	tr.
34800	32000	75100	7200	6700	24900	9700	5600	6500	42300	13100
1300	2300	2200	640	182	1030	455	150	200	1300	775
150	109	515	10	37	62	15	13.5	15	170	95
–	–	–	–	–	–	–	–	–	–	–

32 13/56	33 29/55	34 33/55j	35 43/55	36 71/55	37 an. 5	38 347/55	39 211/55	40 208/55	41 282/55	42 254/55	43 270/55
213000	214000	228000	272000	326600	331200	340600	342900	345700	344800	346200	353200
80000	68000	94000	74000	83000	75000	73000	73500	73500	72500	72000	69000
20	33	22	16	13	13	16	9	14	14	14	14
7000	13000	7000	5000	1200	1300	1300	1200	700	1100	1100	1000
(23)	–	28	370	–	–	–	(17)	–	–	–	–
290	500	125	120	18	10	16	17	17.5	17.5	17.5	17.5
27800	53300	37200	13000	10700	5900	6600	5100	4300	4000	6100	4100
53000	36000	35000	48000	2500	1000	3000	1400	400	500	400	400
8	12	19	30	8	8	10	11	7	12	10	10
76	26	65	122	–	–	9	(8)	10	6	–	–
62	74	39	28	4	4	7	4	7	6	6	6
63	167	8	3.5	2	5	5	5	4	4	3	3
67200	79700	42800	40000	4300	8200	4400	3900	2700	7000	3900	3900
1650	1700	1000	400	165	180	60	75	38	400	130	40
19300	21100	33400	23100	36700	32000	24300	23400	22600	26900	27500	25900
85000	72000	55000	21000	8000	3000	6000	8000	9000	5000	5000	7000
160	155	600	200	103	73	100	180	190	19	31	160
–	–	13.9	–	14.3	9.9	13.4	15.7	15	18	16.2	14.5
3700	5600	16800	36900	42400	47700	53100	51700	51100	48400	47300	41500
180	150	450	2300	715	670	460	690	560	55	72	75
14	12	134	260	117	105	375	212	192	220	165	153
–	–	–	–	–	–	–	–	–	–	–	–

Table III.

Chemical analyses of potash feldspars from Langøy.
(Na, K, Ca, Ba, and Sr in per cent, Li, Rb, Cu, Pb, Tl in p.p.m.).

No.	Li	Na	K	Rb	Ca	Ba	Sr	Cu	Pb	Tl
50 (145/55) Group 1	9.8	4.23	4.88	44	2.6	0.400	0.065	3.7	7.5	–
51 (an. 9)	26.8	4.51	4.82	47	2.4	0.300	0.055	5.4	6.8	–
52 (an. 9)	4.3	4.76	4.68	65	1.9	0.300	0.060	7.9	6.9	–
53 (179/55)	28.8	4.17	6.64	312	1.4	0.350	0.050	18.5	13.1	tr.
54 (an. 7) Group 2	13.0	2.08	10.64	519	0.09	0.013	0.009	0.6	10.1	0.6
55 (271/55) Group 3	12.0	1.90	9.13	337	1.5	1.100	0.170	2.5	7.6	tr.
56 (293/55)	8.7	1.81	10.59	350	0.55	1.000	0.120	1.3	10.9	0.9
57 (255/55)	9.5	1.63	10.61	669	0.5	1.000	0.120	2.4	9.1	0.8
58 (251/55)	9.0	1.71	10.29	370	0.6	1.000	0.130	8.5	9.9	0.9
59 (318/55) Group 4	4.0	2.24	10.09	493	0.55	0.350	0.080	4.7	13.4	0.9
60 (238/55)	6.0	1.94	10.34	734	0.35	0.250	0.065	15.5	11.4	1.0
61 (224/55)	4.7	1.54	10.99	532	0.30	1.000	0.140	0.7	11.4	1.0
62 (an. 12) Group 5	5.5	1.26	11.42	475	0.5	1.000	0.160	8.2	9.6	0.8
63 (348/55)	12.2	1.37	11.49	920	0.4	0.500	0.100	19.2	17.1	1.0
64 (65/56) Group 6	6.0	1.80	9.96	575	0.5	0.400	0.076	14.8	73.8	n.d.
65 (25/55) Group 7	4.3	1.51	11.37	700	0.2	0.600	0.100	2.4	21.1	0.8
66 (150/56)	5.5	1.71	10.99	650	0.3	0.700	0.140	11.1	23.9	0.9
67 (147/56) Group 8	1.6	0.98	12.14	783	0.15	0.800	0.110	8.3	19.4	0.9
68 (3/56)	6.0	1.31	11.62	830	0.15	0.660	0.120	4.5	16.8	0.8
69 (35/55)	3.4	1.20	11.37	637	0.20	0.360	0.050	5.4	11.3	0.7
70 (37/55)	6.4	1.13	11.82	518	0.25	0.800	0.085	3.7	14.1	0.9
71 (an. 1)	tr.	0.80	12.48	306	0.2	0.430	0.054	0.6	20.5	0.6
72 (109/56) Group 9	4.6	0.74	12.62	320	0.15	1.200	0.060	5.6	9.3	0.8
73 (99/56)	tr.	1.37	11.75	182	0.2	0.300	0.054	1.6	7.0	0.8
74 (Evatn)	3.6	0.79	12.62	775	0.15	0.190	0.043	0.9	13.4	0.7
75 (an. 5) Group 10	5.3	3.77	7.49	245	0.20	0.090	0.013	4.0	(3.2)	1.0

Group 1: monzonites within the Eidet-Hovden intrusion.
Group 2: pegmatite on border between gabbro and monzonite, Eidet-Hovden intrusion.
Group 3: porphyroblastic monzonitic granulites.
Group 4: retrograde metamorphosed gneisses.
Group 5: the banded series.
Group 6: old red granite.
Group 7: the charnockite border series.
Group 8: amphibolite facies veined gneisses.
Group 9: red augengneisses (no. 74, red granitic-pegmatitic vein).
Group 10: young red granite.

Table IV.

Chemical analyses of plagioclases from Langøy
(contamination is restricted to quartz, and possibly subordinate amounts
of potash feldspar). Na and K in per cent,
remaining elements in p.p.m.

		(% Ab[1])	Li	Na	K	Rb	Ca	Ba	Sr	Cu	Pb	Tl
80 (293/55)	Group 3[2]	70	17.5	5.48	1.15	23.5	4.1	385	666	15.3	26.3	–
81 (251/55)		75	11.1	5.85	0.68	10	3.4	209	635	20.5	19.9	–
82 (238/55)	Group 4	60	9.0	3.34	0.35	12.5	3.7	360	425	65.1	21.4	–
83 (224/55)		70	2.8	5.84	0.53	15.2	4.1	305	820	7.6	24.5	–
84 (an. 12)	Group 5	68	17.5	5.25	1.39	31.5	4.3	830	945	9.2	39.9	–
85 (25/55)	Group 7	70	6.2	3.09	0.75	25.5	2.3	550	437	13.6	21.0	–
86 (37/55)	Group 8	75	2.1	4.54	0.70	20	2.6	515	695	3.1	19.7	–
87 (an. 1)		75	5.4	3.59	0.47	29	2.1	267	355	28.7	30.9	–

Calculated for 100 % plagioclase:

			Li	Na	K	Rb	Ca	Ba	Sr	Cu	Pb	Tl
80	Group 3		17.6	5.52	1.16	24	4.1	388	671	15.4	26.4	–
81			11.7	6.14	0.71	10	3.6	219	666	21.5	21	–
82	Group 4		13.5	5.03	0.53	19	5.6	542	640	98.0	32	–
83			2.8	5.90	0.54	15.3	4.1	308	828	7.7	25	–
84	Group 5		17.5	5.25	1.39	31.5	4.3	830	945	9.2	39.9	–
85	Group 7		10.9	5.44	1.32	45	4.1	967	770	24.0	37	–
86	Group 8		2.8	6.05	0.93	27	3.5	686	926	4.1	26	–
87			9.1	6.08	0.95	49	3.6	452	601	48.6	52	–

Calculated values of theoretical feldspar components in plagioclases.

		Or	Ab	An	Ba	Sr	Pb	Rb	
80	Group 3	8.27	62.69	28.66	0.11	0.25	0.006	0.009	
81		5.10	69.79	24.79	0.06	0.25	0.004	0.004	
82	Group 4	3.76	57.15	38.68	0.15	0.24	0.007	0.007	
83		3.83	67.01	28.77	0.08	0.30	0.005	0.006	
84	Group 5	9.93	59.63	29.84	0.23	0.35	0.01	0.01	
85	Group 7	9.44	61.85	28.13	0.26	0.28	0.01	0.02	
86	Group 8	6.66	68.73	24.06	0.19	0.35	0.005	0.01	
87		5.76	69.13	24.74	0.12	0.22	0.02	0.02	

[1]) Optically determined plagioclase composition.
[2]) Group . . — see Table III.

Table V.

Chemical analyses of biotites from Langøy. (Na and K in per cent, Li, Rb, Cs in p.p.m.).

No		Pleochroism	Refractive index nβ	Li	Na	K	Rb	Cs	Pb	Tl
90 (293/55) Group 3*)	X	: colourless-light yellow	1.645±0.003	41	0.52	6.10	490	7.2	–	n.d.
	Y	: light brown-yellow brovn								
91 (224/55) Group 4	X	: light green	1.638±0.003	46	0.50	5.16	490	5.7	–	n.d.
	Y	: yellow green								
	Z	: yellow green								
92 (25/55) Group 7	X	: colourless	1.634±0.003	40	0.19	7.37	850	7.6	–	n.d.
	Y,Z	: light yellow								
93 (150/56)	X	: light yellow	1.645±0.003	49	0.31	7.22	850	7.6	–	n.d.
	Y	: yellow-brown								
94 (147/56) Group 8	X	: colourless	1.630±0.003	66	0.19	7.40	820	26.6	–	n.d.
	Y	: yellow green								
95 (37/55)	X	: light green	1.645±0.003	68	0.37	6.18	600	24.7	–	n.d.
	Y	: dark green								
96 (an. 1)	X	: light green	1.630±0.003	75	0.37	7.43	950	17.1	–	n.d.
	Y	: greenish brown								

*) Group . . — see Table III.

Fig. 3 GEOLOGICAL MAP OF LANGØY

Fig. 3. Geological map of Langøy.   *Geologisk kart over Langøy.*