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The Caledonian Mountain Chain of the Southern Troms and Ofoten Areas

Part II. Caledonian Rocks of Igneous Origin

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Contents

With 24 text figures (including one geological map) and 7 plates. 19 tables in text.

Abstract

Metamorphosed rocks of igneous origin are described from the Caledonian part of the Southern Troms and Ofoten areas. Based on megascopic and microscopic appearance, on chemistry and on field relations they can be divided into two separate groups whose age and geologic history differ from each other:

The first group comprises serpentinites. amphibolites, meta-trondhjemites and metamorphosed quartz keratophyres. It is suggested that these rocks represent early peridotite intrusions, basalts, gabbros, basalt tuffs and tuffites, trondhjemite intrusives, and quartz keratophyre pyroclastics and lavas, respectively, all rocks having been metamorphosed by progressive regional metamorphism occurring during the Caledonian orogeny. Their mineral assemblages are generally in good accordance with the distribution of metamorphic facies' as this is revealed by the writer's previous investigations in adjoining meta-sediments. This first group represent the initial Caledonian magmatism in the area, largely belonging to the geosynclinal stage.

The second group of igneous rocks includes ultrabasics (peridotites and troctolites), norites, gabbros, diorites, granodioritic and granitic rock types (to a smaller extent also trondhjemites). These rocks were only to a small degree altered by metamorphic processes and primary magmatic textures are common, especially in gabbros and diorites. Included in the younger group is the well-known Råna norite massif. This second group of igneous rocks is thought to be late-tectonic or post-tectonic in relation to the strongest deformation in the area, but older than the last fold movements.

For both groups of igneous rocks it is suggested that the granitic types (including granodioritic and trondhjemitic rocks) are not necessarily comagmatic with the basic and ultrabasic members, and an anatectic origin may seem more probable than a formation by magmatic differentiation.

In the last section of the paper parallels are drawn with previous results on Caledonian igneous rocks from different parts of the Norwegian Caledonides, and a brief discussion of age relations is given. A comparison with the alpine ophiolites is included.

I. Introduction

Scope of the present investigations.

This work is part of a regional investigation of petrology and structure of Caledonian rocks in the Southern Troms and Ofoten areas, carried out by the writer in recent years. The field work was done mainly in the period 1959—1964, partly with field assistants (1-2 each year).

A paper including petrology of the meta-sediments as well as a petrographical description of the Precambrian basement has previously been published (Gustavson, 1966). This first publication, especially the survey of the metamorphic history and the distribution of metamorphic facies, is important for an evaluation of the results achieved in the present paper. Field data for the present publication have been gathered mainly in the central part of the map area from Ofotfjorden to Salangen and Bardu (map, Fig. 1). Some field work has also been carried out in the easternmost parts of Troms (area without symbols on the map) and this area is briefly taken into consideration. Some observations on amphibolite massifs in this area are of interest from a genetical point of view.

The igneous rocks have been studied in thin sections and powder preparates. In addition, chemical analyses of the major elements have been completed. Twelve of these are full analyses, while the rest include alkali and SiO_2 determinations. Six full analyses, completed for S. Foslie, previously unpublished, have also been included in the description.

As the igneous rocks in question are partly or wholly metamorphosed after their emplacement, their genesis and primary relations present special problems and uncertainties. It has been tried, in this paper, to discriminate as much as possible between the metamorphic and the igneous aspects of their formation: A petrographical description is given in chapters II and III together with a discussion of the mineral facies and metamorphic alteration of the rocks. Considerations of primary, igneous relations which are, of course, of a more debatable nature, are reserved for chapter IV.

Previous work on the igneous rocks of the area.

The most important publications dealing with the general geology of the area have been listed in the first paper of this series (Gustavson, 1966, p. 10-11). Papers on the igneous rocks are restricted to Foslie's descriptions of the Råna massif (Foslie, 1921, 1922, 1941). Brief notes on amphibolites, serpentinites and "granites" were also presented in these papers. The main results of Foslie's investigations are referred to in their proper places.

The amphibolite massifs of Eastern Troms were mentioned by Th. Vogt (1922). They were interpreted as intrusions which had their "roots" in the central and western parts of the area. These rocks have also been briefly commented on quite recently by Kalsbeek and Østerby Olesen (1967).

Terminology

Definitions of some rock names used in the following are given below. Most of them are in accordance with the proposals of Streckeisen (1967) in his "Classification and Nomenclature of Igneous Rocks", but a few terms are taken from other sources because of the metamorphic nature of the rocks in question:

Peridotite. Composed of olivine (30-90 per cent) and pyroxene, mainly. The sum of mafic minerals should be more than 90 per cent (Streckeisen, p. 160 and 174). Dunites contain more than 90 per cent olivine.

Serpentinite. Due to strong variations in the degree of alteration the terms is used in a non-quantitative manner for rocks where a high content of serpentine minerals is clearly visible in hand specimen.

Amphibolite. "Metamorphic rocks composed principally of hornblende and plagioclase. Foliation, due to alignment of amphibole prisms, is less conspicuous than in schists". (Turner and Verhoogen 1960, p. 454).

Gabbro and diorite. Plutonic rocks composed of plagioclase + mafic minerals. Among the feldspars, plagioclase must amount to more than 90 per cent, that is, feldspar ratio plagioclase/alkali feldspar + plagioclase (P/A + P) should be higher than 90. (Streckeisen, 1967). The distinction between gabbro and diorite is determined by the plagioclase, in gabbro having An content generally above 50, in diorites below 50. In both rocks quartz amounts to less than 20 per cent of the total quartz + feldspar, types with 5 to 20 per cent quartz are termed "quartz-bearing" (Streckeisen, p. 160). Diorites usually should contain more hornblende than pyroxene, but this is not essential for the classification used.

Norite and quartz norite. A norite is a gabbro with more orthopyroxene than clinopyroxene. The rocks of Råna, for which this term was used by Foslie (1921, 1922, 1941), seem mostly to conform to this definition. In the system of Streckeisen a quartz norite is a norite in which more than 20 per cent of the light minerals are quartz. From examples given by Foslie it seems as the "quartz norites" of Råna rarely contain more than a few per cent quartz. These rocks would possibly therefore be better designated as norites or quartz-bearing norites. In some of them clinopyroxene (diallage) dominates over hypersthene and these are thus normal gabbros or diorites (plagioclase: An_{40-55}). As, however, the description of the Råna rocks is entirely based on Foslie's work, his terminology is used also for these rocks.

Granite. A rock with feldspar ratio (plagioclase/plagioclase + alkali feldspar) above 10 and below 65, quartz content more than 20 per cent of the light minerals (Streckeisen, 1967). It should be noted that in the Streckeisen system albite (An_{0-5}) is reckoned as alkali feldspar. Rocks with lower contents of plagioclase (feldspar ratio 0-10) are termed alkali granite.

Granodiorite. In the Streckeisen system a rock with feldspar ratio (P/A + P) between 65 and 90, quartz content higher than 20 per cent of the light minerals.

Trondhjemite. According to Streckeisen a trondhjemite is a leucoquartz-diorite with oligoclase (less frequently andesine) as sole feldspar and colour index below 15 (usually 3-10).

Quartz keratophyre. As Johannsen (1955, p. 260) says, the term is "often applied to albitized extrusives". In Norwegian geological litterature this is also the common usage, an exception being N.-H. Kolderup (1929) who applied the term also for a potash-dominant effusive rock, in agreement with the original definitions by von Gümbel, according to whom keratophyre is "the paleovolcanic equivalent of alkali syenite, that is, an older trachyte". (Cited from Johannsen (1955, p. 260)). In the present area all rocks thought to be quartz keratophyres are soda dominant so there is no problem as to whether potash-rich rocks should be included or not.

In the Streckeisen system the present quartz keratophyres are mostly quartz andesites (plagioclase of oligoclase composition) while in a few occurrences the feldspar is albite with An 5 per cent or less and the rocks would accordingly be classified as alkali rhyolites, supposed that the feldspar were primary magmatic. This, however, is not always the case (see description p. 44).

Outline of general geology and igneous rocks

In a previous paper the sequence of meta-sedimentary rocks has been described (Gustavson, 1966). The following groups were distinguished:

6. Niingen Group
5. Salangen Group
4. Narvik Group
3. Rombak Group
2. Storfjell Group
1. Hyolithus Zone

The meta-sediments rest on Precambrian basement, partly with their primary sedimentation contact. The Hyolithus Zone is autochthonous, while overlying groups are all considered allochthonous. Thrust planes separates groups 1 and 2, 2 and 3, 3 and 4 and possibly also 5 and 6. Groups 1 and 2 are found along the eastern margin of the mountain range and in connection with windows of basement rocks in the central and eastern areas and are not represented in the westernmost areas.

Since the printing of the author's 1966 paper, the following publications arrived, dealing with the easternmost parts of the AMS 1 : 250 000 map Narvik: Berthelsen (1967) on the Målselv (Mauken) and Dividal windows, Landmark (1967) on the Målselv window, Kalsbeek and Østerby Olesen (1967) on the area between Altevann and Målselva, and Vogt (post-humous paper, 1967) on the Hyolithus Zone between Torneträsk and Nordreisa. As these districts had been only cursorily investigated by the writer the mentioned papers have caused important changes in the general geological picture of these areas. This has, however, little influence of the following treatment of igneous rocks as the areas in question have mostly been left out of consideration. For the rest of the map area, the general description given by the writer in the 1966 paper, is still the most up to date one.

Igneous rocks or metamorphic rocks of igneous origin are present in the sequence from the lowermost part of the Narvik group upwards. They are most abundant in the Narvik and Niingen Groups, but granitoid intrusive rocks are frequent also in the Salangen Group (see map, Fig. 1).

The igneous rocks have been subdivided into two groups:

1. An older group (Table I), including volcanic rocks emplaced during the geosynclinal stage and intrusive rocks with a chemistry mostly comparable to the volcanics. The group embraces ultramafics (peridotites) of the alpine type, basalt, basaltic tuff and gabbro, now present as amphibolites, meta-trondhjemites (to a certain extent also granodiorites) and

bolites, meta-trondhjemites (to a certain extent also granodiorites) and metamorphosed quartz keratophyres. All rocks of the older group are metamorphosed by progressive regional metamorphism ranging from the epidote-amphibolite facies to the lower part of almandine-amphibolite facies and their assemblages are, in general, in accordance with the facies distribution within the meta-sediments (Gustavson, 1966). The ultrabasic/basic rocks of the older group can, from a general point of view, be regarded as the ophiolites of the Caledonian mountain chain in the present area.

2. A younger group of igneous rocks (Table I), including the Råna massif, is considered as late-tectonic or post-tectonic in relation to the period of strongest folding. Rocks of this group are only partly altered by metamorphic processes and show, in places, well-preserved igneous textures. Rock types included in this group are ultramafic rocks (peridotites and troctolites), norites, gabbros and diorites, granodioritic (in part also trondhjemitic) and granitic dykes and veins.

Some quartz veins may possibly belong to the igneous rocks, but no detailed investigation have been undertaken of these veins.

Table 1.

Igneous rock types in relation to geological processes

	Older igneous group		Younger igneous group	
Stage:	Geosynclinal	Early orogenic	Orogenic	Late orogenic
Processes:	Sedimentation. Volcanism.	Early folding. Sedimentation.	Folding and metamorphism.	Late folding and thrusting
Supracrustal rocks:	Basalt lavas and tuffs, tuffites Quartz keratophyre pyroclastics and lavas			
Intrusive rocks:	Peridotites → ? Gabbros → ? Diabases → ? Trondhjemites → ?		Peridotites Gabbros (norites) Diorites and quartz norites Granodiorites and granites (trondhjemites subordinate)	

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II. Petrography and metamorphism of the older group of igneous rocks

Ultrabasic rocks

Occurrence

As described by Foslie (1941) the ultrabasics south of Ofotfjorden occur at two different levels in the stratigraphical sequence: Those related to the Råna massif of noritic rocks are situated within the Narvik Group. They are comparable in mineral and chemical composition to the Orta peridotite of the Bardu area and will be treated in the next chapter. The rest of the ultrabasics, that is, the bulk of them, are in this southern area situated at a lower level, below the Reppi schist which has been tentatively correlated with the Seternes schist near the base of the Narvik Group (Gustavson, 1966, p. 89). As meta-dunites and meta-peridotites north of Ofotfjorden mostly seem to occur in the upper parts of the Narvik Group, above a thin marble horizon (Gustavson, 1966, p. 71), there are evidently differences in the stratigraphical positions north and south of Ofotfjorden. As, however, the fold tectonics render stratigraphical relations somewhat conjectural, this statement should be considered as a preliminary one.

The rest of the ultrabasics occur within the Niingen Group of meta-sediments. On the map 1 : 100.000 by Th. Vogt (1950) two ultrabasic bodies are shown to occur in the Dudalselv area, one within the Salangen Group, the other within the Niingen Group. The boundary between the two groups is drawn as a straight line. The investigations by the present writer have shown this interpretation to be incorrect: As the boundary below the Niingen Group has to be drawn above the upper marble horizon of the Salangen Group, represented in the Dudalselv area by a limesilicate schist or impure meta-limestone, the ultrabasics are within the Niingen Group (Fig. 2). Furthermore, there are three ultrabasic bosses in this area and not two as shown on the map by Vogt.

The following occurrences (Fig. 1), investigated by the writer, are included in the following petrographical description: Salvasskaret (east of

Sørdalen), Storvann (at Gratangseidet), Vasskaret (at Andørja), Dudalselv (south of Gratangen) and Hesjeberg (north of Gratangen). The two last-mentioned localities belongs to the Niingen Group, while the other ones are situated within the Narvik Group.

The peridotites in the Niingen Group are apparently situated at a much higher stratigraphical level than the other ultrabasics within the area. The possibility of this meta-sedimentary group being an equivalent to the Narvik Group has been mentioned in the 1966 paper and will be further discussed in a forthcoming structural paper. The presence of similar ultrabasic rocks in those two groups is one of the arguments in favour of such an interpretation.

Field relations

The occurrences of ultrabasic rocks are small and scattered. In each locality there are usually 2 or 3 bosses together, either arranged in the same strike line or in an *en échelon* fashion with a small displacement in stratigraphical position from one to the next. The serpentinites generally possess a thick lense shape in outcrop; the size varying from about 10 by 20 to 30 by 80 metres. At the Dudalselv locality they are situated in anticlines as shown on Fig. 2. Most common among the country rocks are mica gneisses and amphibolites.

The boundaries are mostly covered but has been observed in a few cases: At the Salvasskaret locality (Fig. 1) an apparently concordant boundary against marble was detected. The marble is greyish white, containing tremolite needles. No sign of contact metamorphism or other influence from the serpentinite was observable. In another part of the same locality a concordant boundary against hornblende schist can be seen.

One of the Dudalselv serpentinites was found to border upon mica gneiss in two outcrops. As far as it could be judged the boundary is concordant in relation to the gneiss foliation. Changes in the appearance of the gneiss towards the boundary could not be detected.

Apart from varying degree of serpentinization the rocks seem to be homogeneous in composition and texture. Banding or parallelism of minerals have not been megascopically observed. An exception is the serpentinized borders of some rocks, as for instance of the Dudalselv serpentinites. In these cases a pronounced orientation of the serpentine minerals is partly visible. In the central parts of the bodies the serpentinization is more irregular, forming veinlets of variable orientation. As a rule, the alteration becomes pronounced towards the borders.

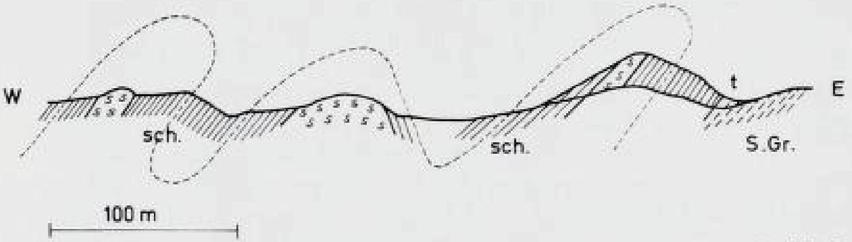

Fig. 2. Profile in serpentinites (s) and adjoining mica schists (sch) at Dudalselv, Niingen Group. Stippled line indicates the tectonic interpretation.
t = possible thrust plane.

Mineralogy and texture

The only minerals which seem to be of primary igneous origin are olivine, orthorhombic pyroxene (enstatite) and, in some cases, magnetite. Secondary alteration minerals found by means of microscopic work are antigorite, chrysotile (subordinate), actinolite, talc, magnesite and haematite. Olivine constitutes fram 50 to 90 per cent by volume of the rocks, the rest being enstatite and alteration minerals. Enstatite varies in amount from 0 to 25 per cent.

The texture of the serpentinites are determined by the variable appearance of olivine: Variations are mainly in the great size; coarse-, medium- and fine-grained types all being present. In the ultrabasics of the Niingen Group a parallel texture is visible (Pl. Ia). This is due to the orientation of the poor cleavages of olivine, to orientation of elongate grains and of antigorite which has been formed along grain boundaries and cleavages in olivine. In this case alternating zones of strongly altered and less altered olivine were seen to form a banding or lamination, each "band" being only some few millimetres across. Some features of the individual minerals are described below:

Olivine: In the Vasskaret serpentinite olivine is found as large grains. up to 2 centimetres in diameter. transsected by numerous irregular cracks filled with alteration minerals. Most common are veinlets of chrysotile + haematite (Pl. Ib). These veinlets are cut by later cracks along which talc has been formed. The olivine shows optic positive character and an axial angle close to 90°. Consequently the fayalite content is low, less than 12 per cent (Tröger, 1959).

The olivine of the Dudalselv and Hesjeberg serpentinites has the same optic data but are extremely fine-grained (0,1 - 0,3 millimetres). Alteration to antigorite along rims and cracks is observable. Cleavages are more frequently developed than in the coarser grained rocks and, as it has already been noted, a certain orientation of olivine grains and their cleavages is observable. Olivine in other rocks than those described above mostly resembles those of the Vasskaret serpentinite but are less coarse-grained.

Enstatite: Orthorhombic pyroxene with the optical characteristica of enstatite is found in most rocks. The grain size is generally smaller than for the olivine, in the coarse-grained rocks enstatite fills the interstices between the olivine grains. Alteration is mostly to antigorite, in one case also to an unidentified mineral (see below).

Magnetite is found as octahedral or equant grains, partly in clusters, in some specimens. The magnetite has been found enclosed in olivine in some cases and are then thought to be a primary mineral.

Alteration minerals: Among the alteration minerals *antigorite* is most frequent. It is found as crackfillings, along rims of olivine and pyroxene and in the interstices between these minerals. It is present as flakes rather than fibres and is slightly green-coloured and pleochroic. The elongation is positive, birefringence moderate to low, (-) 2 V large to moderate. *Chrysotile* has been observed only in the Vasskaret serpentinite. In this rock it forms veinlets in olivine together with *haematite*. These minerals are zonally arranged as shown by Pl. Ib with haematite in the central part of the veinlet. The chrysotile-haematite veinlets are cut by cracks on which *talc* has been formed. Talc also form rims on olivine and pyroxene in some thin sections where it can be observed that these minerals have been altered to antigorite on cracks prior to the formation of talc.

Carbonate minerals which according to their optical properties must be either magnesite or dolomite have been observed in some specimens. Most probably it is magnesite. This conclusion can be drawn from the fact that Ca-bearing minerals are generally absent from serpentinites of the present group of igneous rocks.

Actinolite prisms have been detected in one of the Dudalselv serpentinites (determination confirmed by X-ray identification) as scattered unorientated needles. The actinolite has been slightly altered; on cracks to antigorite and, apparently in a later stage, to talc along the rims.

An unknown mineral, possibly a member of the chlorite group occur as veinlets in olivine and pyroxene grains in the Vasskaret serpentinites. It is present as light greyish-brown masses with very low birefringence and weak pleochroism. Its relation to other alteration minerals can not be stated. A similar mineral has been described by Foslie (1931, p. 223) from serpentinites in the southern part of the Ofoten area.

Chemistry

No chemical analyses have been performed on ultrabasic rocks during the recent investigations. The microscopic investigations have, however, revealed mineral assemblages, and amounts of the minerals involved, comparable to those described by Foslie (1931, 1941) from the area south of Ofotfjorden for ultrabasics not related to the Råna massif. A chemical analysis from Foslie (1931, p. 221) has therefore been quoted in Table II. Molecular norm (catanorm) and Niggli values have been added. Characteristic are the extremely low values for Na_2O, Al_2O_3 and CaO and a high MgO/FeO ratio. The validity of these features for all ultrabasics described in this chapter is shown by the general scarcity or absence of minerals containing Na, Ca and Al (as for instance plagioclase and monoclinic pyroxene) and by high amounts of olivine rich in the forsterite molecule. These characteristica are identical with those found by Hess (1955) to be essential in *alpine-type peridotites*.

Metamorphism and mineral facies

The relative age relations of the alteration minerals can be established on the base of thin section study (see description above). In general it seems justified to divide the alteration of the peridotites into 1. *Serpentinization*, leading to the production of antigorite and, occasionally, chrysotile, this process predating 2. The formation of talc (*steatitization*). Magnesite and haematite are less important minerals formed during the serpentinization process. Formation of actinolite seems to be older than the serpentinization but is of little importance in the present rocks. The conclusion reached about the age relations of serpentine minerals and talc is in agreement with the relations found by Foslie (1931) by study of two serpentinites south of Ofotfjorden. Another conclusion drawn by Foslie (p. 241) is that antigorite was formed before chrysotile during the serpentinization period. The observations made by the present writer in the Vasskaret rocks show that this feature does not have general application as the age relations are reversed in these serpentinites. In the couple of ultrabasics described by Foslie magnetite was obviosly formed as a consequence of the serpentinization. As already described magnetite appears to be a primary mineral in some of the ultrabasics recently investigated while haematite in one case was found to be a secondary mineral, occurring together with chrysotile. The conditions during serpentinization, therefore, obviously must have been somewhat different in different parts of the area.

It has been mentioned how serpentinization in some rocks (e.g. the Dudalselv ultrabasics) is most pronounced in the outer parts. Like Du Rietz (1935, p. 255) we must conclude that this feature indicates that "the serpentinizing waters were not produced by a dunitic magma out of which the rocks themselves crystallized, but that solutions effecting the transformation came from outside sources." The same view is expressed by Foslie (1931, p. 227) though a more irregular distribution of the serpentine minerals were found in the serpentinites investigated by him. Bowen and Tuttle (1949), on the base of experimental investigations, also express the view that the serpentinizing solutions in general derive from external sources.

As to the age of the serpentinization it is of interest to note that Foslie ascribed it to late-tectonic processes while formation of talc was considered post-tectonic. It will be shown later in this discussion that a pre- to syntectonic serpentinization and talc formation in these rocks is a more probable solution.

Table II

Chemical composition, molecular norm, mode and Niggli values of serpentinite, Ofoten.

	Wt %	*Cat %*	*Molecular norm*		*Mode*[3]):		*Niggli values*	
SiO_2	45,45	36,8	Or	0,5	Olivine[4])	52,3	al	0,8
TiO_2	0,00	—	Ab	1,0	Magnetite	2,7	fm	99
Al_2O_3	0,67	0,7	An	—	Chromite	0,5	c	0
Fe_2O_3	3,34	2,3	C	1,4	Pyrrhotite	0,1	alk	0,2
FeO	4,15	3,1			Magnesite	2,5	si	60
MnO	0,13	0,1	Hy	34,4	Talc	1,6	k	0,20
MgO	40,21	54,4	Ol[1])	56,1	Antigorite	40,0	mg	0,91
CaO	0,00	—	Mt	3,4	(Col. index)	(100,0)	qz	—41
Na_2O	0,11	0,2	Chr[2])	0,5				
K_2O	0,05	0,1	Ptl[2])	0,3				
CO_2	1,41	1,7	Ms[2])	3,4				
P_2O_5	0,00	—	(Colour					
Cr_2O_3	0,48	0,3	index)	(98,1)				
NiO	0,07	0,1						
S	0,08	0,2						
$H_2O > 200°C$	8,51	—						
$H_2O < 200°C$ $> 110°C$	0,08	—						
$H_2O < 110°C$	0,20	—						
	99,94	—						
— O for S	0,04	—						
Sum	99,90	100,0						

Locality of specimen: E. of Rauvatn, Melkedalen.
Analyst: E. Klüver, for S. Foslie (1931).

[1]) The proportion Fo:Fa in the olivine calculated is 27:1. (Norm calculation by the present writer).

[2]) Chr = Chromite, Ptl = Pentlandite, Ms = Magnesite (Magnesite was calculated instead of calcite as CaO is absent).

[3]) Calculated by Foslie. The mode is given in volume per cent.

[4]) Part of the olivine has undergone a transformation, with increase in water and silica, decrease in iron, according to Foslie (1931, p. 224).

A consideration of metamorphic facies within the serpentinites reveals interesting relations: Most of them are situated within schists belonging to the lower part of the almandine-amphibolite facies (Gustavson, 1966). According to Turner and Verhoogen (1960) the typical assemblages of ultrabasic rocks at this stage of metamorphism are cummingtonite (or anthophyllite) - hornblende - almandine or cummingtonite (anthophyllite) - tremolite. Talc may appear in the epidote - almandine subfacies together with actinolite (tremolite) and chlorite, and in the epidote-biotite subfacies together with serpentine and actinolite. The mineral assemblages of two serpentinites from the map area are listed in table III.

Table III

Mineral composition of two ultrabasic rocks from the map area.

Vasskaret	*Dudalselv*
Olivine	Olivine
Enstatite	Enstatite
Chrysotile	Antigorite
Antigorite	Talc
Talc	Actinolite
Magnesite	Magnesite
Haematite	
Magnetite	

It is seen that the compositions are quite different from the amphibolite facies assemblages as outlined by Turner and Verhoogen. Olivine is not expected to appear until the upper part of this facies. Orthorhombic pyroxene belongs to the granulite facies. On the other hand, talc and serpentine minerals indicate metamorphism in a lower facies than for the adjoining schists. Disequilibrium relations thus obviously are present. This supposition is further emphasized when the assemblages are considered in the light of the phase rule: According to Barth (1962, referring to Korzhinskii) the mineralogical phase rule can be formulated in the following way:

$$p \leqq c - n$$

where p is the number of phases (minerals), c is the total number of components and n is the number of mobile components. The phase rule thus expresses that the number of minerals should not, in equilibrium assemblages formed in an open system, exceed the number of components subtracted for the mobile components. In the present serpentinites (Table II) the components are as follows: $(Mg, Fe^{II})O$, SiO_2, Fe_2O_3, H_2O and CO_2, that makes $c = 5$. If H_2O and CO_2 are considered mobile during metamorphism, n = 2 and the number of minerals in an equilibrium assemblage should not exceed 3, eventually 4, if MgO and FeO are treated as separate components. In the two rocks listed in Table II 6, respectively 8, minerals are present in significant amounts.

If it is assumed that H_2O and CO_2 were not mobile during the whole process, that, is, the system was closed, the number of possible minerals increases to 5. But it is seen that there are still too many minerals present for an equilibrium assemblage. Thus it must be concluded that the assemblages are discordant with the phase rule and consequently in disequilibrium.

In the opinion of the present writer this lack of adjustment can be explained in either of two ways: 1. The metamorphism took place by declining temperatures, perhaps late-tectonically and late in the metamorphic history of the area, or 2. The ultrabasics were metamorphosed during the main metamorphism, but for some reason greater parts of the ultrabasic bodies were prevented from complete reconstruction.

The former alternative is supported by the fact that recrystallization seems to be slow when it takes place in the absence of stress and by declining temperatures (Turner and Verhoogen, 1960). It is also in accordance with the view expressed by Foslie (1931, p. 241) that the alteration postdates the main folding. However, this interpretation is contradicted by observations on the age sequence of the minerals as well as by general considerations: It has been shown that talc was formed later than the serpentine minerals. According to experiments by Bowen and Tuttle (1949), chrysolite will, by rising temperatures, break up giving talc, forsterite and water vapor. The reaction temperature is nearly independent of the hydrostatic pressure. It is also implyed in the facies classifications of Turner and Verhoogen that talc belongs to a higher metamorphic state than serpentine minerals. In a recent discussion of the alteration of ultrabasic rocks, Turner (1968, pp. 160-165) confirms the previous statements about the stability fields of serpentine and talc. As the temperature thus evidently increased after the serpentinization it seems probable that *serpentine as well as talc formation belongs to the period of progressive regional metamorphism*. It is of interest to the discussion of this problem that the frequent field association of ultrabasics with amphibolite indicates a certain genetic connection. As the amphibolites have been found to have suffered complete reconstruction during the progressive phase of metamorphism this is another indication that the peridotites were emplaced before the main regional metamorphism. Alternative 2. above is therefore preferred. In determining why the serpentinites have been incompletely adjusted to the metamorphic conditions the following features may be of importance:

a) Alteration is most pronounced along the borders.

b) Deformation of the ultrabasics is observable in some cases, but as a rule the tectonic influence is only slight.

From point a) above it can be inferred that the serpentinizing waters, derived from the adjoining sediments, after having altered the outer parts, had not easy access to the central part of the bodies. This may perhaps be due to the protecting effect of the rock already altered. As noted by Turner and Verhoogen (1960, p. 486) "the low permeability of completely recrystallized metamorphic rocks is a factor operating against late hydration and carbonation". Resistance of the ultrabasic rocks against folding and shearing movements would act in the same direction as shear planes and fissures giving access for H_2O and CO_2 were available only to a small degree. As the temperature increased talc was stabilized, but, for the same reasons as above, the formation was relatively restricted. When the temperature was further increased and the metamorphism reaching its maximum, the serpentinites remained largely unchanged due to their shell of alteration minerals. At this time the H_2O (and CO_2) of the sediments had also to a great extent been bound in newly formed minerals like mica and amphiboles within schists and amphibolitic rocks.

The reasoning above implyes that the ultrabasic rocks during their metamorphism partly acted as a closed system, also with respect to H_2O and CO_2.

Amphibolite dykes

Field relations

Among the amphibolites of intrusive origin the dykes form a small group as observations of cross-cutting boundaries against the metasediments are rare. Such rocks have been encountered in a single locality only: The dykes are met with close to the farm Rotvik, Salangen, where they can be studied in new roadcuts (Fig. 3). The country rocks are impure quartz schists, belonging to the Narvik Group (Gustavson, 1966, pp. 73-74) in part with a distinct banding, obviously relict sedimentary layering. The dykes are steeply inclined, 1-2 metres in thickness and cut the banding of the schists under an acute angle. Deformation of the amphibolites is shown by small-scale folds and the development of axial plane schistosity. The axes of the folds are roughly parallel with axes of folds in the quartz schist not far from the dykes. The amplitude of the folds in the amphibolite is somewhat less than within the quartz schist.

Fig. 3. Three amphibolite dykes in quartz schist at Rotvik, Salangen. The steeply dipping banding (layering) in the quartz schist meets the amphibolite boundary at an acute angle. Subhorizontal jointing is observable.

Mineralogy and texture

Texturally the dykes are characterized by pronounced planar orientation of minerals and mineral aggregates. Some features of the minerals are given below.

Hornblende: Crystal forms are regularly developed. They are typical hornblende forms and uralite has not been detected. Ample inclusions of plagioclase, ore minerals and sphene testify to the metamorphic origin of the hornblende. Cracks in the hornblende prisms, caused by post-crystalline deformation, are in some cases filled with quartz. Optic data are: n_z = 1,680 (ca.). 2 V_X = 72°, extinction angle c/Z = 14°. The pleochroic scheme is as follows: Z: Blue-green. Y: Greyish green, X: Light greyish yellow.

Plagioclase. The grains are irregular in shape. Crystal faces have not been observed. Only a few of the feldspar grains show distinct twinning (albite law) and the determination of the composition based on the extinction angles is not very exact in this case. The approximate composition is about An_{30}, perhaps a little higher.

Biotite. The flakes show planar orientation parallel with the hornblende prisms. Inclusions of sphene, hornblende, ore grains and zircon are common. The pleochroic scheme is from reddish brown to nearly colourless.

Sphene. The grains occur in the form of irregular drops or as chains of such grains parallel with the schistosity.

Magnetite, apatite, calcite and *zircon* are present in small amounts. A single grain of *garnet* was observed, partly transformed into *chlorite.*

Chemistry

A chemical analysis of one of the dykes shown on Fig. 3 is presented in Table IV. When choosing the specimen for analysis caution was taken to avoid stripes of quartz, evidently introduced from the adjoining quartz schists.

A comparison with Table V shows a great similarity with the chemistry of amphibolite massifs of intrusive origin occurring in eastern areas (pp. 26-27).

The mode calculated from this analysis seems to be close to the actual composition as observed by thin section study. Estimates, made before the analysis was available, were about 60 per cent hornblende and 30 per cent plagioclase.

Mineral facies

The assemblage: Plagioclase (An_{30}) + hornblende + biotite + almandine, shows the metamorphism to have taken place under the conditions of the almandine - amphibolite facies. The association quartz + muscovite of the adjoining quartz schists, is stable over a wide range of P-T conditions and it is, therefore, not possible to decide whether there is full accordance between the mineral facies of the amphibolite dyke and that of the adjoining schists.

Amphibolite massifs

Occurrence

Amphibolite as greater massifs is mainly confined to the easternmost areas. Brief comments on these rocks have been given by Vogt (1921) by whom they where considered as intrusive rocks injected from central parts of the mountain chain in the west. During recent years some observations have been made by assistants of the writer in the Rokkomborre massif and by the writer on the Bangfjellet amphibolite, the latter forming part of a greater massif (Vogt, 1967, Kalsbeek and Østerby Olesen, 1967). In addition to the two mentioned occurrences the writer has mapped the Iselvdal amphibolite massif (Fig. 1). The Kistefjell massif (see maps Vogt, 1967, and Kalsbeek and Østerby Olesen, 1967) has not been investigated by the writer.

The amphibolite massifs seem to occur at a relatively low stratigraphical or tectonical level: The Iselvdal and Rokkomborre amphibolites are situated within schists of the Narvik Group and probably within the lower part of

Table IV

Chemical composition, molecular norm, mode and Niggli values of amphibolite dyke, Rotvik, Salangen

	Wt %	*Cat. %*	*Molecular norm*		*Mode (calculated)*		*Niggli values*	
SiO_2	47,80	45,5	Or	4,5	Plagioclase	26,1	al	21
TiO_2	2,34	1,7	Ab	22,5	(% An)	(30)	fm	47
Al_2O_3	15,44	17,3	An	29,7	Hornblende	59,3	c	25,5
Fe_2O_3	1,01	0,7	(% An)	(57)	Almandine	tr.	alk	6,5
FeO	11.31	9,0	Hy	14,0	Biotite	7,2	si	111
MnO	0,27	0,2	Di	15,2	Magnetite	1,0	k	0,17
MgO	6,42	9,2	Ol	8,4	Sphene	5,1	mg	0,48
CaO	10,13	10,4	Mt	1,0	Apatite	0,5	qz	—15
Na_2O	2,49	4,5	Il	3,4	Calcite	0,8		
K_2O	0,74	0,9	Ap	0,5	(Colour index)	(73,9)		
CO_2	0,29	0,4	Cc	0,8				
P_2O_5	0,30	0,2	(Colour index)	(43,3)				
H_2O+	1,02	—						
$H_2O—$	0,03	—						
Sum	99,59	100,0						

Analyst: P.-R. Graff.

this group. The position of the other massifs is possibly comparable to the two, though frequent marble inclusions in the Bangfjellet amphibolite may indicate a somewhat different position of this massif. It is obvious, however, that it belongs to the lowermost parts of the sequence (see also Kalsbeek and Østerby Olesen, 1967, on this matter).

Field relations

The amphibolite massifs seem to have the shape of thick lenses. This is for instance the interpretation of the Iselvdal massif. This amphibolite is rather thick in the central part, some few hundred metres, and it is successively thinning north- and southwards. It has been traced nearly to the northern map boundary as a sill within mica schists, the thickness decreasing to some few metres in the northernmost part. Discordant boundaries have not been detected. The lower boundary of the Iselvdal amphibolite has been observed south of the farm Iselvmo. Quartz - mica schists crop out below the amphibolite, strongly folded and compressed, seemingly concordant. The upper boundary is exposed in the profile at Skredbekken in the central part of the amphibolite. From about 700 m.a.s.l. the amphi-

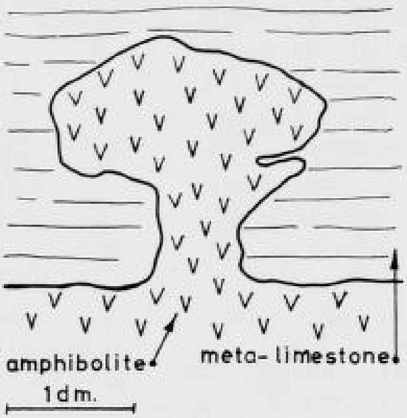

Fig. 4. Boundary detail of amphibolite and meta-limestone inclusion in the Bangfjell amphibolite.

bolite occurs in layers alternating with mica schists, obviously concordant with the mica schistosity. The schists are partly of a quartz-rich type.

Boundary relations in other amphibolite massifs are investigated only to a small extent. As far it can be judged they are comparable to those described above.

Inclusions of meta-sediments are present in the amphibolite massifs at Iselvdal, Rokkomborre and Bangfjellet: In the Iselvdal and Rokkomborre massifs quartz schist is, up to now, the only rock type found as inclusions. In the Rokkomborre amphibolite these quartz schists seem to have an extension of at least several hundred metres and thicknesses of some tens of metres. In the Skredbekken profile in the Iselvdal massif quartz schists crop out in the central part of the amphibolite. These have an observed extension of some few metres, and are cross-cut by veins of amphibolite. It is of interest to note that these quartz schist inclusions are of a type similar to that found in schists of the Narvik Group west of Iselva (Gustavson, 1966, map, Fig. 1). Inclusions of quartz schist, mica schist and marble have been observed by the writer in the Bangfjellet amphibolite, some of which are intruded by amphibolite veins (Fig. 4). Inclusions of schists and marble are also mentioned by Kalsbeek and Østerby Olesen (1967, p. 255). The size of the inclusions must be considerable and their frequency seems to increase upwards in the massif.

Mineralogy and texture

Plagioclase and hornblende are the dominant minerals, but epidote is occasionally present in significant amounts. Minor constituents are sphene, calcite, garnet, quartz and ore minerals. Chlorite is found only in veins.

From a textural point of view variations from pronounced planar orientation of minerals to nearly non-directional textures are found, the latter types are commonly observed in relict porphyric varieties with plagioclase phenocrysts. *The relict porphyric types*, which have been observed mainly in the upper part of the Iselvdal amphibolite, show close resemblance to certain amphibolite layers described later in this paper (p. 28 and Fig. 7). Locally well preserved "gabbroic" textures are reported from the Bangfjellet massif by Kalsbeek and Østerby Olesen (1967, p. 257).

The schistose types are relatively finegrained in most cases. This does not mean that all fine-grained types are schistose, however, in the Rokkomborre amphibolite, for instance, only slight parallelism of minerals is visible in the investigated thin sections. The fine-grained type is, in fact, very common in the Rokkomborre massif.

Some features of the main minerals are given below:

Hornblende: In the medium-grained rocks of even grain size, as well as in the relict porphyric types, relatively short prismatic hornblende is present. The outlines are mostly irregular. No signs of original pyroxene have been recognized. The pleochroic scheme is usually as follows: Z = Y: Green with brownish or bluish tinge, X: Greenish yellow. In the Rokkomborre amphibolite a pale green actinolite is occasionally present though a blue-green amphibole is most common. The two amphiboles have not been observed in the same thin section. In schistose amphibolite types within the Iselvdal massif the amphibole is long prismatic of the bluish green type.

Plagioclase. The composition varies from albite in the Rokkomborre massif to andesine in the massive part of the Iselvdal amphibolite. Slight zoning with composition varying from An_{38} in the core to An_{35} in the rim, is observed in the latter rock. It is of some interest to note that the plagioclase of the more schistose parts of the Iselvdal massif is considerably more sodic than in the massive parts, about An_{25}. Inclusions of epidote within the plagioclase are especially conspicuous in the relict phenocrysts. Flakes of sericite are found in these feldspars as well. The strong alteration of the phenocrysts is in some contrast to the white and relatively fresh appearance they exhibit in hand specimen. Below the microscope the boundaries of the phenocrysts are also seen to be diffuse, as clinozoisite needles frequently protrude across the boundaries. This may partly be the case also with some hornblende prisms. Alteration to epidote minerals is frequently more pronounced in the central part of the phenocrysts than in their rim. This tend to show that they were originally zoned, with the most basic composition in the core (Pl. IIA).

Epidote minerals. In some thin sections the epidote mineral is a colourless clinozoisite while in other cases it is common epidote, greenish and slightly pleochroic.

Garnet and *quartz* have been observed only in the schistose variants. *Sphene* occurs commonly in the ampibolites, partly as grains of droplike habit.

Chemistry

The chemical compositions of two specimens from amphibolite massifs are shown in Table V. The specimen from the Iselvdal massif (column a) is a massive variant from the central part without obvious igneous textures. From the Bangfjell amphibolite a variant with remains of porphyric texture was selected (column b). Resemblances in composition to the analysed dyke specimen (Table IV) from Salangen are evident. Further comments on the chemistry are given in a later section (p. 78).

Table V

Chemical compositions, molecular norms, modes and Niggli values of two specimens from amphibolite massifs.

	(a)		(b)	
	Wt %	Cat %	Wt %	Cat %
SiO_2	48,92	46,4	48,87	45,8
TiO_2	1.71	1,2	0,20	0,2
Al_2O_3	14,39	16,0	16,60	18,4
Fe_2O_3	1,22	0,9	0,02	—
FeO	10,58	8,4	10,04	7.8
MnO	0,22	0,2	0,27	0,2
MgO	6,94	9,9	6,68	9,4
CaO	11,44	11,6	10,96	11,0
Na_2O	2,56	4,7	3,31	6,0
K_2O	0,50	0,6	0,73	0,9
CO_2	0,06	—	0,09	0,1
P_2O_5	0,13	0,1	0,20	0,1
H_2O +	1,21	—	1,63	—
H_2O —	0,02	—	0,09	—
Sum	99,91	100,0	99,69	99,9

Molecular norms:

	Or	Ab	An	(% An)	Ne	Hy	Di	Ol	Mt	Il	Ap	Cc	(Colour index)
(a)	3,0	23,5	28,8	(53)	—	9,6	24,0	9,0	1,4	2,4	0,3	—	(46,7)
(b)	4,5	24,7	29,0	(54)	3,2	—	19,2	18,3	—	0,4	0,5	0,2	(38,6)

Modes:	Plagioclase	(% An)	Hornblende	Biotite	Epidote min.	Sphene	Apatite	Calcite	Ore min.	(Col. index)
(a)	31,5	(35-38)1)	61,5	—	2,0	3,6	—	tr.	1,4	(68,5)
(b)	32,3	(?)	58,2	2,2	5,0	0,5	0,1	—	1,4	(67,7)

Niggli values:

	al	fm	c	alk	si	k	mg	qz
(a)	19,5	46,5	28	6,5	111,5	0,11	0,51	— 14,5
(b)	22,5	42,5	27	2	111,5	0,13	0,54	— 20,5

1) Zoned, see description.

Localities and numbers of specimens:

(a) Specimen 71/61, E. of Iselva, central part of the Iselvdal amphibolite. (Anal. R. Solli.)

(b) Specimen 69/63, 700 m.a.s.l., northern slope of Bangfjellet, Dividal. (Anal. P.-R. Graff.)

Mineral facies

The mineral association of the Iselvdal amphibolite, plagioclase (oligoclase/andesine) + epidote + hornblende ± almandine ± quartz shows that almandine - amphibolite facies metamorphism occurred after the emplacement. Zoning, in some specimens, may indicate a certain degree of disequilibrium, but in general there is *good agreement between the assemblages found in the amphibolite and those of the meta-sediments* of the adjoining area. (Gustavson, 1966, map Fig. 2). Investigations of assemblages in other massifs are too scarce for any safe conclusions to be drawn, but from a few thin sections of the Rokkomborre amphibolite it seems as low epidote - amphibolite facies or greenschist metamorphism occurred, with formation of the following association: Albite + actinolite (or blue-green hornblende) + chlorite (on veins) + epidote. From the few data available from adjoining schists a similar metamorphic grade might be inferred.

Amphibolite layers of various kinds

Subdivision of the amphibolite layers

Based mainly on their megascopic appearance the amphibolite layers have been subdivided into the following groups:

A. Massive-textured amphibolite layers.
B. Relict porphyric amphibolite layers.
C. Thin, schistose amphibolites with admixed sedimentary material.
D. Greenstones.
E. Amphibolites with relict pillow structures.

Amphibolite layers are also encountered in the area south of Ofotfjorden. (Foslie, 1941). They are to a great extent comparable to the layers mapped by the present writer but are not included in the following discussion except when chemistry is concerned. Four analyses completed for Foslie, previously unpublished, are included in Table VIII. A brief description of the specimens, given by Foslie (from archives of NGU) is cited in connection with this table.

Occurrence and field relations

Representatives of all types of amphibolite layers are found within the Narvik Group. In addition, massive-textured amphibolites are frequent in the Niingen Group. Layers, most commonly belonging to group C, are met with also in the Salangen Group. In the latter group amphibolites thought to have been formed from calcareous sediments are also encountered. They have been briefly treated in a previous paper (Gustavson, 1966, p. 108).

Greenstone-like rocks have been observed in two localities only: One of them is found at Straumslitverrelva, Bardu, above the meta-limestones of the Rombak Group. The lower boundary is not exposed. The greenstone is concordantly overlain by mica schists. Total thickness of the greenstone is probably about 15 metres. The second occurrence is west of the farm Leirbekkmo, Salangen, in a similar structural position and with a comparable thickness. It seems that both occurrences are within the lowermost part of the Narvik Group.

The relict porphyric layers are found as intercalations (not very frequent) in mica schists in the lower part of the Narvik Group within the Bardu district. More rarely they have been observed interlayered in the quartz keratophyre formation (Gustavson, 1966, p. 74 and p. 41, this paper).

Massive-textured layers and amphibolite layers with admixed sedimentary material are found within most parts of the Narvik Group sequence. However, it seems as they are especially abundant within the quartz keratophyre formation and the sequence immediately above this formation. As noted in the 1966 paper there is an upward increase in the frequency of amphibolite layers within this formation.

Of great interest is the type E, amphibolite with relict pillow structures in the Kvæfjord area. These rocks have been investigated by Flood (Thesis, Oslo University, 1961) and will be further treated by him. Texturally these amphibolites are similar to the massive-textured layers. The thickness is considerable, but pillow structures are found only within a limited area in their upper part. As far as the writer's present interpretation of stratigraphical relations is correct, the amphibolites in Kvæfjord also belong to the Narvik Group.

All types of amphibolite layers are folded together with the meta-sediments (Fig. 5). Boudins and pieces of disrupted amphibolite layers (Figs. 5, 6) are common. The boudins are mostly belonging to the amphibolite types A and C and are included in the following petrographical description and considerations.

The extent of individual amphibolite layers seems to be considerable in some cases, but a closer inspection of the extensions is mostly prevented by cover or by the presence of several layers of similar appearance within a restricted part of the sequence.

The concordant nature of the layers in relation to the schistosity of adjoining rocks is mostly obvious. When sedimentary layering is observable in the schists the amphibolites have also been found concordant with this structure. The parallel textures visible in many amphibolites seem in all cases to be parallel with the boundaries.

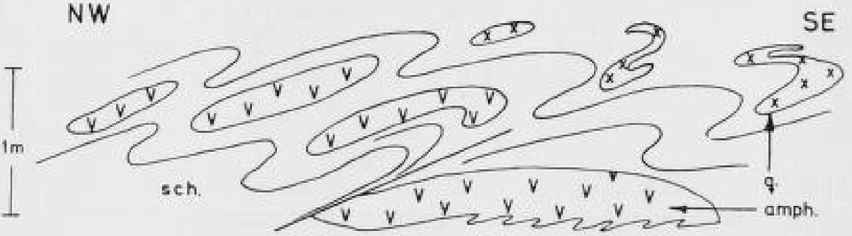

Fig. 5. Amphibolite (amph.), folded and disrupted together with the adjoining mica schist (sch.), Blåberget, Bardu. (q = quartz veins.)

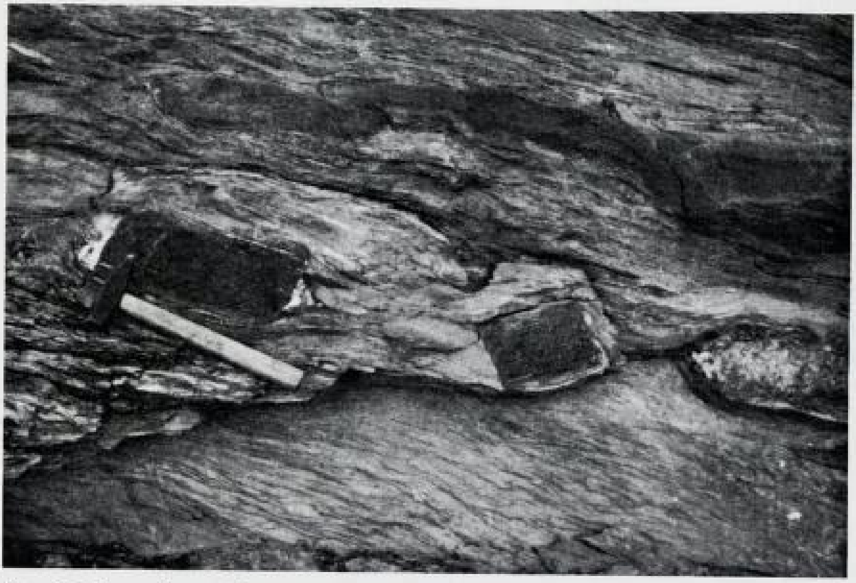

Fig. 6. Disrupted amphibolite layer, Garsnes, Salangen. Quartz segregations at the ends of the pieces. Note compression in lime silicate layer above amphibolite.

In general the amphibolite layers are too small to be given separate symbols on the map (Fig. 1). Exceptions are some amphibolites, mostly of the massive-textured type, above the quartz keratophyre formation in the Bardu district (e.g. in Bangskletten) and the amphibolites mapped by Vogt in the mountains north of Rombaken (Vogt, 1950). These layers may attain thicknesses as high as 100 - 200 metres.

Mineralogy and texture

The variations in appearance of the amphibolite layers are to some extent indicated by the subdivisions introduced above. Some further comments on their megascopic appearance are given below:

The massive-textured amphibolites are fine- to middle-grained rocks where parallel textures are mostly weak or absent. (Pl II b.) Colours are greenish or greyish black. The layers are homogeneous in the sense that compositional banding is absent. Differences in mineralogy between the central parts and the borders are restricted to the occasional occurrence along the borders of considerable amounts of biotite, a feature which can be explained either by admixture of sedimentary material or by influx of chemical components during metamorphism.

In the relict porphyric amphibolites parallelism of minerals is usually not visible. The main difference from the massive-textured layers is the presence of white and mostly rectangular plagioclase phenocrysts (Fig. 7). Irregularly outlined feldspars occur together with the euhedral ones.

The "thin amphibolite layers" (group C), is mostly schistose rocks, fine- to medium-grained with colour varying from grey to greenish. Parallel stripes of quartz, biotite and occasional garnets are frequently visible. In some few cases calcite may be present. The greenstones are also schistose and fine-grained in appearance. They differ from the other basic rocks in their light green colour. Conspicuous small, elongated spots or stripes of white colour are visible. Veins or clots of quartz transsect the greenstone in the mentioned Salangen locality.

Mineralogically the amphibolite layers are composed of hornblende and plagioclase, in most cases amounting to 80 - 90 per cent by volume, and minor amounts of the following minerals: Epidote (or clinozoisite), biotite, quartz, sphene, garnet, calcite, chlorite, clinopyroxene (rare) and ore minerals. The modes of 15 amphibolite layers have been determined by point counting in thin section and listed in Table VI. The presence of quartz and biotite in all the thin, schistose layers and of epidote in the relict porphyric rocks should be noted.

Fig. 7. Relict porphyric amphibolite, Bangskletten, Bardu. Nat. size.

Properties of the minerals are given below:

Plagioclase. Apart from the phenocrysts occurring in some layers, the plagioclase is present as irregular grains in the size range 0,05 to 0,5 millimetres. Albite and pericline twins are mostly present. The composition varies from albite to andesine (about An_{45} is maximum). Weak *inverse zoning* is observable in some amphibolites, especially in the massive-textured type. The compositional variations in plagioclase from one locality to another seem mostly to be related to the areal variation in metamorphic grade as demonstrated by assemblages in the adjoining schists (see Gustavson, 1966, pp. 123-124).

Under the microscope the phenocrysts of the relict porphyric rocks are seen to be composed of an aggregate of plagioclase and clinozoisite, frequently with the original outlines of the phenocrysts well preserved. Metamorphic recrystallization has, however, caused smaller changes in this outline. Hornblende needles may, for instance, protrude into phenocrysts. The present plagioclase is untwinned and filled with rods and grains of clinozoisite, especially within the central parts of the phenocrysts, exactly as it was found in the porphyric variant of the Iselvdal massif (Pl. IIa). Originally the phenocrysts must have been considerably more basic in composition than the present plagioclase which is albite or oligoclase. The size of the phenocrysts is in the range 0,5 to 1,5 centimetres.

Hornblende. Some features of the hornblendes in amphibolite layers are listed in Table VII. Only the thin, schistose layers and the massive-textured layers are represented in this table. Hornblendes in the greenstones and the relict porphyric layers are of the long prismatic, blue-green type like that found in specimens 3 and 6 of Table VII. The optic data determined are consistent with the assumption that all amphiboles are

Table VI
Observed modes of 15 amphibolite layers of various types.[1]

	1	*2*	*3*	*4*	*5*	*6*	*7*	*8*	*9*	*10*	*11*	*12*	*13*	*14*	*15*
Hornblende	63,3	64,3	65,2	53,0	63,4	64,7	60,0	57,0	58,0	64,0	73,0	61,5	72,2	71,0	56,0
Plagioclase	33,7	32,7	23,8	33,0	25,8	22,0	24,7	31,0	24,0	15,0	3,0	7,5	16,4	12,3	27,0
Epidote minerals	—	—	—	—	4,4	7,7	9,3	tr.	15,5	—	3,0	tr.	3,0	0,4	8,5
Calcite	—	—	—	—	—	tr.	2,0	—	1,0	—	—	—	—	—	—
Diopside	—	—	—	1,6	—	—	—	—	—	—	—	—	—	—	—
Garnet	—	—	0,3	3,5	—	—	—	—	—	—	3,4	—	—	—	—
Biotite	—	—	5,4	1,0	—	—	—	2,0	tr.	9,3	0,7	1,5	2,0	tr.	—
Sphene	tr.	0,2	1,0	2,8	2,0	—	—	—	tr.	1,7	0,6	tr.	1,2	4,7	3,0
Chlorite	—	—	—	—	—	0,3	0,3	—	—	—	—	—	—	—	5,5
Quartz	—	—	4,3	—	—	—	—	—	—	10,0	16,3	24,5	5,2	11,6	—
Ores	3,0	2,8	—	5,1	4,4	5,3	3,7	10,0	1,5	—	—	5,0	—	—	tr.

---

[1] No:s 1-4: Massive-textured amph.s. 5-9: Relict porphyric amph.s. 10-14: Thin, schistose layers. 15: Greenstone.
No:s 1, 5, 13 and 15 are the analysed rocks a-d of Table VIII.
Localities: 1. - sp. 462/59: Høgfjell, Salangen. 2. - 439/59: Gamvikskar, Lavangen. 3. - 515/60: Røyrbakkvann, Salangen. 4. - 418/59: Foldvik, Gratangen. 5. - 55/61: Bangskletten, Bardu. 6. - 64/61: Basevardo, Bardu. 7. - 65/61: Basevardo, Bardu. 8. - 36/62: W. of Høgskarhus, Dividalen. 9. - 216/61: Østerdalen, Bardu. 10. - 558/60: Vasskaret, Andørja. 11. - 544/60: Straumskarheia, Andørja. 12. - 447/59: N. of Lavangen. 13. - 68/61: Ravddovaras, N. of Altevann. 14. - 8/61: Leirvassfjell, Bardu. 15. - 46/61: Skinskaret. Bardu.

Table VII
Properties of hornblendes in thin, schistose amphibolites (1-6) and in massive-textured amphibolite layers (7-11).

	Specimen no.	Shape	Colour (Z)	$2V_x$	Extinction angle c/Z	n_z
1	558/60	Long prismatic	Brownish green	80°	16°	1.678
2	544/60	Short prism.	Grass green	85°	25°	n.d.
3	447/59	Long prism.	Blue-green	73°	18°	n.d.
4	505/59	» »	Light green	86°	17°	1.656
5	562/60	Short prism.	Brownish green	76°	14°	n.d.
6	68/61	Long prism.	Blue-green	75°	22°	1.674
7	462/59	Short prism.	Brown	86°	13°	1.680
8	439/59	» »	Brownish green	84°	20°	n.d.
9	515/60	» »	» »	71°	14°	1.678
10	418/59	» »	Grass green	71°	16°	n.d.
11	66A/63	» »	Slightly greenish	95°	19°	n.d.

Localities:
558/60: Vasskaret, Andørja. 544/60: Straumskaret, Andørja.
447/59: Roadcut, N. of Lavangen. 505/59: Rubben, Kirkesdal, Målselv.
562/60: Trolla, Andørja. 68/61: Ravddovaras, N. of Altevann.
462/59: Høgfjell, Salangen. 439/59: Gamvikskar, Lavangen.
515/60: Røyrbakkvann, Salangen. 418/59: Foldvik, Gratangen.
66A/63: Kvernelva, Dividal.

hornblende, containing some Na and Al, although nothing precise can be said about the composition. Of the hornblendes listed, no:s 3, 4, 6 and 11 (blue-green and light green types) are found in rocks from the almandine zone of the epidote-amphibolite facies (Gustavson, 1966), while the rest (green and brown types) are from the almandine - amphibolite areas.

Miyashiro (1961, 1968) has given the following sequence of stable hornblendes by increasing metamorphic grade: Blue-green, green, brownish green and brown types. In the kyanite - sillimanite type of facies series, which is also the one present in the Southern Troms area (see Gustavson, 1966, p. 123), blue-green hornblendes will appear in metabasites approximately at the almandine isograd (Miyashiro, 1968, p. 810). Green and brown types are stable in the almandine - amphibolite facies. In the greenschist facies actinolite amphibole is stable.

As far as it can be judged the variation in hornblende types within the present amphibolites is determined by the changing metamorphic conditions. The hornblendes encountered are those to be expected from a consideration of the facies distribution in the adjoining meta-sediments.

Some of the short prismatic hornblendes (Table VII) could certainly by uralitic in origin but decisive proofs are lacking.

Epidote minerals. Clinozoisite is most common, appearing as rods or irregular grains, colourless in thin section. Its presence in the phenocrysts has already been commented on. Greenish epidote occurs within the white spots and stripes in the greenstones mentioned in the previous description. In addition these stripes contain *calcite* as a main component together with albite.

Pyroxene has been found in a couple of thin sections. It is greenish *diopside*, probably of metamorphic origin.

Biotite occurs in the thin, schistose layers and a few other amphibolites. It is mostly red-brown under the microscope.

Sphene as drop-like or wedge-shaped grains is the most common accessory in the amphibolite layers.

Ore minerals. *Magnetite*, *pyrite* and *ilmenite* are the most frequent ore minerals. The ilmenite grains are usually in a state of transformation into sphene.

Quartz grains are, like the biotite, most common in the thin, schistose layers, occurring in stripes parallel with the schistosity.

(The text continues on page 40, after the table).

Table VIII

Chemical compositions, molecular norms, modes and Niggli values of 8 amphibolite layers.

	a		b		c		d		e		f		g		h	
	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %
SiO_2	49,36	45,8	49,57	47,0	52,57	49,0	47,09	44,1	50,20	46,7	49,12	46,9	49,93	46,3	49,47	47,1
TiO_2	1,46	1,0	2,73	1,9	0,72	0,5	1,67	1,1	1,50	1,1	2,46	1,8	0,26	0,2	1,66	1,2
Al_2O_3	16,29	17,8	15,36	17,2	12,70	14,0	14,55	16,1	16,62	18,2	13,70	15,3	20,65	22,6	15,39	17,3
Fe_2O_3	1,08	0,8	2,71	1,9	2,15	1,5	3,82	2,7	0,82	0,5	1,95	1,4	2,44	1,7	3,88	2,7
FeO	8,39	6,5	8,38	6,6	5,89	4,6	7,89	6,2	7,56	5,9	10,67	8,5	2,59	2,0	6,79	5,4
MnO	0,17	0,2	0,23	0,2	0,17	0,1	0,19	0,2	0,12	—	0,18	0,2	0,12	0,1	0,17	0,1
MgO	7,28	10,9	7,44	10,6	11.06	15,5	7,06	10,1	8,41	11,6	5,19	7,4	6,22	8,6	7,02	10,0
CaO	10,16	10,1	9,28	9,5	9,64	9,6	10,75	10,9	9,72	9,7	11,02	11,0	13,13	13,1	10,84	11,0
BaO	—	—	—	—	—	—	—	—	0,03	—	tr.	—	—	—	—	—
Na_2O	3,49	6,3	2,36	4,3	2,51	4,5	3,47	6,2	2,39	4,3	2,58	4,8	2,91	5,2	2,48	4,8
K_2O	0,51	0,5	0,49	0,6	0,57	0,6	0,26	0,4	1,04	1,3	0,44	0,5	0,13	0,2	0,27	0,4
CO_2	0,03	—	—	—	—	—	1,55	2,0	0,22	0,3	1,21	1,6	—	—	—	—
P_2O_5	0,10	0,1	0,26	0,2	0,10	0,1	0,13	—	0,18	0,1	0,22	0,2	—	—	0,21	0,02
V_2O_5	—	—	—	—	—	—	—	—	0,04	—	—	—	—	—	—	—
Cr_2O_3	—	—	—	—	—	—	—	—	0,07	—	—	—	—	—	—	—
S	—	—	—	—	—	—	—	—	0,16	0,3	0,19	0,4	0,01	—	0,02	—
H_2O +	1,12	—	1,43	—	0.61	—	1,35	—	1,28	—	1,37	—	1,52	—	1,64	—
H_2O —	0,07	—	0,05	—	0,04	—	0,04	—	0,03	—	0,01	—	0,05	—	0,11	—
Sum	100,14	100,0	100,29	100,0	99,73	100,0	99,82	100,0	100,39	100,0	100,31	100,0	99,96	100,0	99,95	100,0
— 0 for S									0,07		0,10		—		0,01	
	100,14		100,29		99,73		99,82		100,32		100,21		99,96		99,94	

Descriptions of specimens (e) - (h) of Table VIII, by S. Foslie (from archives of NGU):

(e): "Green hornblende schist". The hornblende is a light green variety with weak pleochroism. Opt. data: 2 V_x = 88° ± 6°; Extinction c/Z = 15°; n_z = 1.660, n_y = 1,649, n_x = 1,637. The plagioclase is zoned with the most basic composition along the rim, the core is frequently irregular in outline. Variation within a single grain from An_{21} in the core to An_{34} in the rim. The biotite is light brown in the Y and Z directions, n_z = n_y = 1,615. Rutile occurs as independent grains, in one case a rim of sphene has been observed. The chlorite is a transformation product of biotite.

(f): "Garnet hornblende schist" (anal. published in Foslie (1921)): The amphibole is long prismatic with pleochroic scheme, Z: Bluish green: Y: Olive green: X: yellow green. Opt. data: 2 V_y = 79° ± 3°; c/Z = 13,5°; n_z = 1,681, n_y = 1,671, n_x = 1,658. The plagioclase is inversely zoned, the core about An_{20}, the rim from An_{22} to An_{26}. The sphene occasionally has small cores of rutile. The garnet is typically porphyroblastic. It is surrounded by a rim of plagioclase. quartz and small biotite flakes as well as some calcite.

(g): "Light hornblende schist, rich in feldspar". The hornblende shows pale colours, blue-green in the Z direction. Opt. data: 2 V = 85° ± 5°; c/Z = 15°; n_z = 1,657, n_y = 1,646, n_x = 1,633. The epidote is zoned, with the most iron-rich composition in the core. The plagioclase shows marked inverse zoning with the basic rim quantitatively dominating. The most acid core is An_{19} with a rim of comp. An_{28}. The parallel texture of this rock is pronounced.

(h): "Hornblende schist". Pronounced parallel textures. The hornblende is long prismatic with pleochroism from strong blue-green to yellow-green. Opt. data: 2 V_x = 77° ± 3°; c/Z = 16°; n_z = 1,673, n_y = 1.664, n_x = 1,652. The plagioclase show weak inverse zoning with composition from An_{20} in the core to An_{22} in the rim. The epidote is partly zoned as in (g) above. Rutile is present in considerable amounts. The chlorite is probably clinochlore.

(Table VIII continued).
Molecular norms:

Q	—	2,4	1,0	—	—	3,5	—	2,3
Or	2,5	3,0	3,0	2,0	6,5	2,5	1,0	2,0
Ab	30,7	21,5	22,5	31,0	21,5	24,0	26,0	23,0
An	27,5	30,8	22,3	23,8	31,5	25,0	43,0	30,8
(% An)	(47)	(59)	(50)	(43)	(59)	(51)	(62)	(57)
Ne	0,5	—	—	—	—	—	—	—
Hy	—	23,2	28,0	6,4	19,8	18,6	5,2	17,0
Di	17,6	12,0	19,6	16,4	11,6	16,4	18,0	18,0
Ol	17,7	—	—	10,2	4,8	—	3,9	—
Mt	1,2	2,8	2,3	4,0	0,7	2,1	2,5	4,0
Il	2,0	3,8	1,0	2,2	2,2	3,6	0,4	2,4
Py	—	—	—	—	0,5	0,6	—	—
Ap	0,3	0,5	0,3	—	0,3	0,5	—	0,5
Cc	—	—	—	4,0	0,6	3,2	—	—
(Colour index)	(38,8)	(42,2)	(51,2)	(43,2)	(40,5)	(45,0)	(30,0)	(41,9)

Modes:

Quartz	4,1	5,6	0,5	5,0	—	—	5,2	—
Plagioclase	22,8	22,5	34,0	12,7	33,7	25,8	16,4	27,0
(% An in plag.)	(20-34)	(20-26)	(19-28)	(19-25)	(38-40)[1]	(n.d.)	(22-25)[2]	(5-10)[2]
Hornblende	57,3	58,5	40,4	65,2	63,3	63,4	72,2	56,0
Biotite	12,7	0,2	—	0,6	—	—	2,0	—
Epidote min.	0,9	—	25,10	13,8	—	4,4	3,0	8,5
Garnet	—	5,3	—	—	—	—	—	—
Chlorite	0,1	—	—	1,4	—	—	—	—
Sphene	0,7	5.6	—	—	tr.	2,0	1,2	3,0
Rutile	0,7	—	0,1	1,2 Ore				
Pyrrhotite	0,7	0,1	—	— min.	3,0	4,4	—	—
Ilmenite	—	0,7	—	—				
Apatite	—	0,3	—	0,1	tr.	tr.	tr.	tr.
Calcite	tr.	1,2	—	—	—	—	—	5,5
(Colour index)	(73,1)	(71,9)	(65,5)	(82,3)	(66,3)	(74,2)	(78,4)	(73,0)

Niggli values:

al	20	21,5	17	19,5	23	19,5	28,5	21,5
fm	45	48,5	53	46	46	45	31,5	45
c	25	23,5	23,5	26	24	29	33	27,5
alk	8,5	6,5	6,5	8,5	7	6,5	7	6
si	112,5	118	120	107	117	120	117,5	117
k	0,08	0,11	0,13	0,05	0,22	0,09	0,03	0,07
mg	0,59	0,55	0,72	0,52	0,64	0,43	0,70	0,55
qz	— 21,5	— 8	— 6	— 27	— 11	— 6	— 10,5	— 7

[1] Slight, irregular zoning, most basic in the rim.

[2] Inaccurate determinations.

Localities and numbers of specimens a-d:
- (a) Massive textured amphibolite layer, Høgfjell, Salangen. (Sp. 462/59).
- (b) Relict porphyric amphibolite layer, Bangskletten, Bardu. (Sp. 55/61).
- (c) Thin, schistose amphibolite layer, Ravddovaras, N. of Altevann. (Sp. 68/61).
- (d) Greenstone, Skinskaret, Bardu. (Sp. 46/61).

Localities of specimens e-h:
- (e) From drilling core, Brattåsen, Ballangen.
- (f) From drilling core, Bjørkåsen mine.
- (g) Height 855, Geitvann, Ofoten.
- (h) Height 855, Geitvann, Ofoten.

Analysts: (a-d): P.-R. Graff, (e-f): O. Røer, (g): M. Klüver, (h): E. Klüver. Analyses (e-h) completed for S. Foslie. All calculations by the present writer.

Chemistry

Specimens typical of each of the classes A to D of amphibolite layers have been chemically analysed. The analyses are listed in Table VIII together with four analyses (completed for S. Foslie) of layers from the area south of Ofotfjorden. The composition of the thin, schistose amphibolites (column c) as compared with the other rocks is worthy of note. Further comments are reserved for the later discussion (pp. 77-78).

Mineral facies

Amphibolite layers occur within schists belonging to the epidote-amphibolite and the almandine-amphibolite facies (see map, Fig. 2 in Gustavson, 1966). Within amphibolites of the epidote-almandine subfacies areas the following assemblages are met with: Hornblende + albite/oligoclase + epidote/clinozoisite ± biotite ± quartz ± sphene ± calcite ± "ores" ± chlorite. Almandine is hardly present. Representatives of amphibolites from areas of the epidote-almandine subfacies are for instance number 5, 6, 7, 8, 9, 12, 13, 14, 15 of Table VI. In these amphibolites oligoclase feldspar is about equally as common as albite. Epidote minerals are universally present. According to Turner and Verhoogen (1960, p. 540) albite is the stable feldspar together with epidote in this subfacies, also in basic rocks. In the almandine-amphibolite areas the following assemblages are found in amphibolite layers: Hornblende + oligoclase-andesine ± epidote ± diopside ± almandine ± biotite ± quartz ± "ores" ± sphene. Chlorite and calcite are not present in this facies. The maximal An content of the plagioclase (about An_{45}) is a little higher than in the pelitic schists. Held together with the presence of oligoclase in some amphibolites of the epidote-almandine subfacies as contrasted with albite in adjoining pelitic rocks, this underlines *a general tendency for the plagioclases being more calcic in the metabasites than in the meta-sediments of the same metamorphic grade*. Similar relations have previously been described by Billings and White (1950) from the Appalachians. The reason for this may perhaps by sought in the chemical differences of the rocks or in different yielding to stresses. In general, however, it can be little doubt that *the metamorphic grade of the amphibolites vary areally with the same pattern as it does for the pelitic rocks*. It is therefore assumed that the metamorphic history, at least the most important part of it, is the same for the amphibolite layers as for the meta-sediments.

Little is known about variations in the composition of minerals other than plagioclase. As already noted (p. 34) it is evident, however, that the type and, consequently, the composition of hornblendes vary with metamorphic grade within the area. The nature of the chemical variations is insufficiently known, but it is generally thought (Miyashiro, 1968, p. 818-820) that the contents of Na and Al in hornblende tend to increase with metamorphic grade at this stage of metamorphism (epidote-amphibolite to low amphibolite facies). It is also believed that H_2O and the Fe^{3+}/Fe^{2+} ratio are higher in the blue-green hornblendes than in other ones. The brown colours of higher-grade hornblendes may be due to a higher content of Ti (Gustavson and Grønhaug, 1960, p. 42; Miyashiro, 1968, p. 818).

Metamorphosed quartz keratophyres

Occurrence and field relations

Fine-grained, mostly light-coloured rocks, which are interpreted as metamorphosed quartz keratophyres, are abundant in the Bardu district, stratigraphically within the Narvik Group. In the previous paper (Gustavson, 1966, p. 45 and 74) the term "Blåberget keratophyre" was introduced. As the bulk of the rocks show high quartz contents a better designation would be "Blåberget quartz keratophyre". In the following they are simply spoken of as "the quartz keratophyres" or "the quartz keratophyre formation".

The quartz keratophyres occur *concordantly intercalated with the metasediments of the Narvik Group, that is, mica schists and quartzites and with amphibolite layers.* An example of the latter case is shown in Fig. 8. Intercalated amphibolites comprise the types described as "massive-textured", "relict porphyric" and "thin, schistose layers". In a single locality a thin marble horizon was found within the quartz keratophyre succession.

The boundaries of the formation are not well defined. In some parts of the Bardu district the lowernmost quartz keratophyre layers alternate with quartzite (Gustavson, 1966, p. 74), while in other localities they are interlayered with mica schists in the lower part. Amphibolites become increasingly important upwards in the formation, while the quartz keratophyre layers there are less frequent and at last disappear.

The thickness of the quartz keratophyre formation varies within wide limits. Maximal thicknesses, inclusive of numerous layer of amphibolite and meta-sediments, may be some 300 metres in the central Bardu district (see map, Fig. 1). Outside this area the quartz keratophyres are usually found within a smaller part of the sequence. Individual layers may vary from a few centimetres to several metres, in the latter case the layers

Fig. 8. Alternating layers of amphibolite and quartz keratophyre. Near Innset, Bardu.

frequently show fine *internal banding,* as that shown on Fig. 9. The normal type of quartz keratophyre is relatively fine grained, especially as compared with metasediments and most of the amphibolites. Deviations from this normal type are, however, frequent in some parts of the area: Small-scale mobilization has occurred in places producing increase in grain size, veinlets of quartz + feldspar and slightly discordant behaviour of light-coloured "bands" in relation to darker variants. As already mentioned in the 1966 paper (p. 74 and Fig. 27, p. 76) some coarse gneisses in the area west of Salangen may possibly be equivalent to the quartz keratophyres of eastern districts. The differences can eventually be ascribed to the higher grade of metamorphism within this part of the sequence west of Salangen.

On the map, Fig. 1, in the writer's 1966 paper, some rocks in the Kvæfjord area were termed "agglomeratic rocks". This is a conglomerate with abundant boulders of a fine-grained, "leptitic" rock type resembling the quartz keratophyres, although of a more potash-rich composition. Relict porphyric texture in some boulders has been observed. Later, however, it has been detected that this conglomerate is not so monomictic as first thought. Metasedimentary boulders may locally play an important part.

The conglomerate in Kvæfjord is of considerable interest as a possible witness of acid volcanic activity also in this western areas. Whether it has an agglomeratic origin or is a sedimentary conglomerate which derived

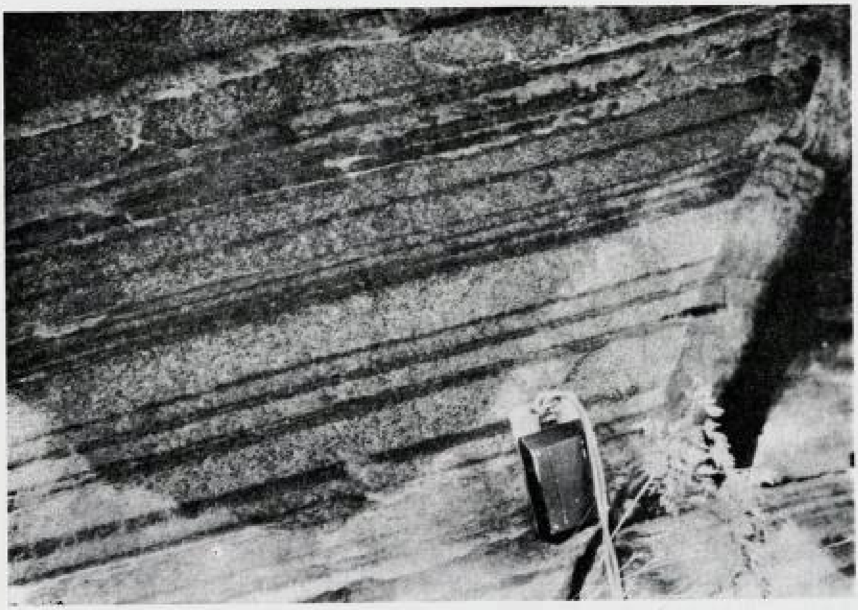

Fig. 9. Banding in quartz keratophyre due to varying amounts of dark minerals. Leirvassfjell, Bardu.

its material largely from acid volcanic (?) rocks, can not be decided at present. The actual area will be further investigated.

Mineralogy and texture

Quartz and sodic plagioclase are the main minerals in all specimens investigated and variations in the assemblages are therefore due to different minor constituents and to variations in their amounts. The banding which is common in these rocks is due to varying amounts of dark minerals, among which micas are most important. Other minor constituents occurring are epidote, chlorite, hornblende, garnet, calcite and ore minerals. Potash feldspar is present only in a few of the thin sections.

Some variations are also found in the textural relations: Primary igneous textures are mostly absent (see below) and the rocks usually exhibit granoblastic textures. Parallelism of minerals or of mineral aggregates has been observed in several cases, but is also frequently absent. Further textural details are given below as part of the description of individual minerals:

Plagioclase. Most plagioclases are granoblastic and in the size range 0,02 - 0,1 mms. In a couple of thin sections, however, plagioclases have been detected which are subhedral or nearly euhedral and of greater size, usually 1-2 millimetres (Pl. III a). It seems reasonable to assume that they are relict phenocrysts. One of these thin sections is from the analysed specimen (b) of Table IX, from Salangen: The phenocryst-like feldspars of this rock commonly are tablets parallel with (010). Twinning according to the Carlsbad law or the combined Carlsbad and albite laws is frequent. Determinations of refractive index on cleavage pieces gave $n'_x = 1{,}545$, corresponding to An$_{35}$. An outer rim of oligoclase composition is sometimes present. The outer boundary of these rims is irregularly outlined although the euhedral shape of the feldspar is approximately retained. Twinning is less common in the fine-grained groundmass feldspars of this rock and their composition has not been determined. The average composition of the plagioclase within the rock, as calculated from the analysis, is An$_{23}$. It is thus evident that the groundmass plagioclase is considerably more sodic than the (supposed) phenocrysts.

Saussuritization has not been observed in the mentioned specimen. Sericitization is, however, pronounced, especially within the phenocrysts (Pl. III b). The sericite flakes are mostly orientated parallel with crystallographic directions, mainly along the basal cleavage. Biotite occasionally grows along the same directions. Probably these micas are secondary alteration products and not inclusions. The general absence of inclusions in the larger plagioclases. together with their nearly euhedral outlines, are strong evidences for their being phenocrysts.

Similar feldspars are observable in a specimen from Salvasskaret (Pl. IV a) although this rock is tectonized and the larger feldspar crystals partly scattered by crushing of the rock. The composition is considerably more sodic than in the Salangen specimen as it has been determined to An$_5$. Small grains of clinozoisite are included in the plagioclase. Judging from the amounts they indicate a previous composition of the plagioclase not more basic than oligoclase. It is of interest to the discussion of the primary viz. secondary origin of these feldspars that Carlsbad twins are present. Such twins have been encountered only in the phenocryst-like feldspars from the two mentioned localities. In the bulk of quartz keratophyres investigated the plagioclases are granoblastically intergrown and of about equal size within each specimen (Pl. IV b). They are mostly of albite or oligoclase composition, exact determinations (refraction on cleavage pieces) from 10 localities gave the following values: An$_4$, An$_5$, An$_5$, An$_8$, An$_9$, An$_{21}$, An$_{26}$, An$_{28}$, An$_{30}$, An$_{36}$. Saussuritization as well as sericitization occur but are not always pronounced. Apart from clinozoisite and sericite, inclusions are uncommon. Twinning is mostly scarce. When present it seems mostly to be albite twins.

Quartz. Granoblastically intergrown grains are present in significant amounts in all thin sections. They are in the same range of size as the feldspars. Quartz grains of phenocryst-like habit have not been observed.

Potash feldspar. In most of the quartz keratophyres plagioclase seem to be the only feldspar. This is supported by the low potash contents of the analysed specimens (see later). In a few cases, however, small amounts of microcline is present. This is especially the case with some specimens from the area between Setermoen and Salangen. The microcline grains show the typical grid twinning.

Biotite is present in all specimens, mostly in small quantities. In some cases, as for instance in one of the chemically analysed rocks, column (b) of Table IX, it is found in considerable amounts. The biotite usually shows dark colours in the Z direction, reddish or greenish brown. Parallelism of the flakes is pronounced in some rocks, faint or absent in others.

Muscovite is rarely more than an accessory, an exception is some rocks from the easternmost areas with exceptionally high amounts of muscovite as well as calcite (10-15 per cent of each mineral). Muscovite is present as separate flakes, about 0,1 millimetre in size, or as small scaly sericite within the plagioclase.

Chlorite. In about one half of the investigated specimens chlorite is an accessory. It is found as well-bounded flakes, partly orientated at random, partly in an approximately radial arrangement. In a couple of thin sections the chlorite is obviously formed from biotite flakes or hornblende needles. The chlorite shows about the same optical properties in all specimens: Negative optic sign, 2 V small and the highest index of refraction about 1.600 or a little below. The birefringence is low (interference colours are always grey of the 1st order and never anomalous). According to Winchell and Winchell (1951) these data correspond to a relatively iron-poor chlorite, probably a penninite or a delessite.

Hornblende. Needles of hornblende are found in a few specimens. Appearance and properties seem to be about the same as for hornblendes in the adjacent amphibolites.

Epidote minerals. Epidote (or clinozoisite) and orthite are frequent accessories. *Orthite* is a characteristic member of nearly all assemblages, partly as a core within epidote grains. Euhedral, nearly equant crystals of epidote, slightly yellow in colour, are conspicuous in some thin sections.

Garnet, in the form of small euhedral crystals, has been observed in a few cases, mainly within the Blåberget area.

Accessories, less commonly present than those described above, are *sphene, tourmaline, zircon, apatite* and *ore minerals* (mostly magnetite).

Chemistry

Two full chemical analyses and six partial analyses (Na_2O, K_2O and SiO_2) of quartz keratophyres have been completed (Tables IX and X). In table XVII the two complete analyses are compared with five analyses of Scandinavian quartz keratophyres taken from literature (see discussion p. 80).

Attention is drawn to the high proportion of Na_2O/K_2O, especially in those from the central Bardu district ((a) of Table IX and (a), (b), (c) of Table X).

Mineral facies

The quartz keratophyres are present within areas of the following metamorphic facies' as shown by assemblages in the pelitic meta-sediments (Gustavson, 1966, pp. 122-131): The epidote-biotite subfacies, the epidote-almandine subfacies and the kyanite-almandine-muscovite subfacies. In these three subfacies' the following parageneses are to be expected in quartzo-feldspathic rocks (Turner & Verhoogen, 1960, pp. 537-540 and 545-548):

Quartz-albite-microcline-biotite-muscovite-epidote.
Quartz-albite-microcline-epidote-almandine-muscovite-biotite.
Quartz-microcline-plagioclase-biotite-muscovite-epidote.

Apart from the absence of microcline in most specimens the assemblages of the quartz keratophyres are close to the cited ones. The chlorite present can, in some cases, clearly be ascribed to retrograde alteration of hornblende or biotite. It may, however, in some rocks have been formed during the main metamorphism as a consequence of low potassium contents of these rocks, preventing the formation of biotite from the Fe and Mg available.

In a few of the quartz keratophyres from the epidote-almandine subfacies areas oligoclase is present instead of albite which is the feldspar to be "expected". As a conclusion, however, it can be stated that there is a fairly good correlation between the metamorphic grade as indicated by assemblages in the quartz keratophyres and that found in intercalated or adjoining meta-sediments.

Table IX

Chemical compositions, molecular norms, modes and Niggli values of two quartz keratophyres from Southern Troms.

	(a)		(b)	
	Wt %	Cat %	Wt %	Cat %
SiO_2	70,51	65,1	68,65	63,3
TiO_2	0,38	0,3	0,06	—
Al_2O_3	15,69	17,1	16,85	18,3
Fe_2O_3	1,18	0,8	0,47	0,3
FeO	1,35	1,0	2,63	2,0
MnO	0,04	0,1	0,10	0,1
MgO	0,94	1,3	1,24	1,7
CaO	3,06	3,0	2,99	3,1
Na_2O	6,11	11,0	5,42	9,7
K_2O	0,21	0,2	1,26	1,4
CO_2	n.d.	—	0,05	—
P_2O_5	0,12	0,1	0,17	0,1
H_2O+	0,33	—	0,58	—
H_2O-	0,05	—	0,02	—
Sum	99,92	100,0	100,49	100,0

Molecular norms:

	Q	Or	Ab	An	(% An)	C	Hy	Mt	Il	Ap	(Col. index)
(a)	24,2	1,0	55,0	14,0	(20)	0,3	3,4	1,2	0,6	0,3	(5,5)
(b)	20,6	7,0	48,5	14,5	(23)	1,4	7,2	0,5	—	0,3	(8,0)

Modes:	Quartz	Plagioclase	(% An)	Amphibole	Biotite	Chlorite	Epidote min.	Muscovite	Sphene	Apatite	Ores	(Col. index)
(a)	26,0	65,5	(21)	3,5	1,5	2,5	tr.	tr.	tr.	—	1,0	(8,5)
(b)	23,2	63,0	(23)[1]	—	13,5	—	tr.	tr.	—	0,3	—	(13,8)

Niggli values:

	al	fm	c	alk	si	k	mg	qz
(a)	42	16	15	27	319	0,02	0,40	+ 111
(b)	42	19	13	26	290	0,13	0,42	+ 86

[1] The calculated average (see text, p. 44).

Localities and numbers of specimens:

(a) Sp. 67/61 Ravddovaras, Bardu. Analyst R. Solli.
(b) Sp. 485/59 Skårvikelven, Salangen. Analyst P.-R. Graff.

Table X

Na_2O, K_2O, SiO_2 and Niggli k-values of six quartz keratophyres from Southern Troms.

	(a)	(b)	(c)	(d)	(e)	(f)
Na_2O	4,24	5,24	5,67	5,67	5,43	6,00
K_2O	0,63	0,43	0,56	1,30	0,67	1,31
SiO_2	77,51	73,01	74,53	70,94	71,59	74,89
Niggli k-values	0,08	0,04	0,06	0,13	0,07	0,13

Localities and numbers of the analysed specimens: (a) - Sp. 61/62 Salvasskaret, Bardu, (b) - 208/62 Salvasskarfjell, Bardu, (c) - 87/63 Blåberget, Bardu, (d) - 6/61 Leirvassfjell, Bardu, (e) - 20/61 Moseter, Bardu, (f) - 594/61 Orrefjell, Salangen.

Analyst: P.-R. Graff.

Meta-trondhjemites

Occurrence and field relations

Trondhjemites belonging to the present group of igneous rocks occur stratigraphically within the Narvik and Niingen Groups. Furthermore, they seem to be more abundant in areas of the highest metamorphic grade (amphibolite facies areas, see map, Fig. 2 in Gustavson, 1966) than in other parts of these groups. Most occurrences are too small to be depicted on the map, Fig. 1, but a few, for instance in the Tysfjord area, have greater dimensions. Several occurrences, however, have not been visited by the writer, especially in the central part of the Ofoten area, mapped by Th. Vogt. The informations are here too scarce for an exact classification. On map copies in the archives of NGU there are rocks termed "Niingen granite". From the brief descriptions it seems evident that they are not granites, but granodiorites, and some of them may be trondhjemites. On the map (Fig. 1) they are termed "granitoid rocks, not classified".

The trondhjemites occur mainly in the following forms:

A. Lense-like bodies, from a few to some tens of metres in their greatest dimension.
B. Small veins or schlieren.
C. Extensive bands along the strike of the meta-sediments.

The latter type is found mainly within the Tysfjord area.

An example of one of the typical lense-shaped bodies is shown on Fig. 10. Some small veins may occur along the boundaries of larger lenses as shown on Fig. 11, but this is no rule. On the other hand, veins or schlieren are found in abundance in a number of localities without any obvious relation to larger bodies. In some of the gneisses where trondhjemite lenses and veins occur, plagioclase augens are also frequent (Gustavson, 1966, pp. 79-84).

An intrusive nature of the trondhjemites is indicated by observations of *cross-cutting relations of dykes and veins to compositional banding of probable primary sedimentary origin* (Fig. 12). In the Høgfjell area, Salangen, meta-trondhjemitic dykes in massive-textured amphibolite layers were observed (Fig. 20).

The classification of the present trondhjemites to the older igneous group is based on studies of their metamorphism and their relation to deformation: Folding, assumed to belong to the earliest deformational phases, is in several localities observed to be younger than the trondhjemites (Fig. 12). *Axial plane schistosity, formed during these folding phases, is usually parallel with the axial planes of folds affecting the trondhjemites*

Fig. 10. Trondhjemite lense at Bratberg, Gratangen. The surrounding rocks are kyanite-bearing mica schists of the Niingen Group.

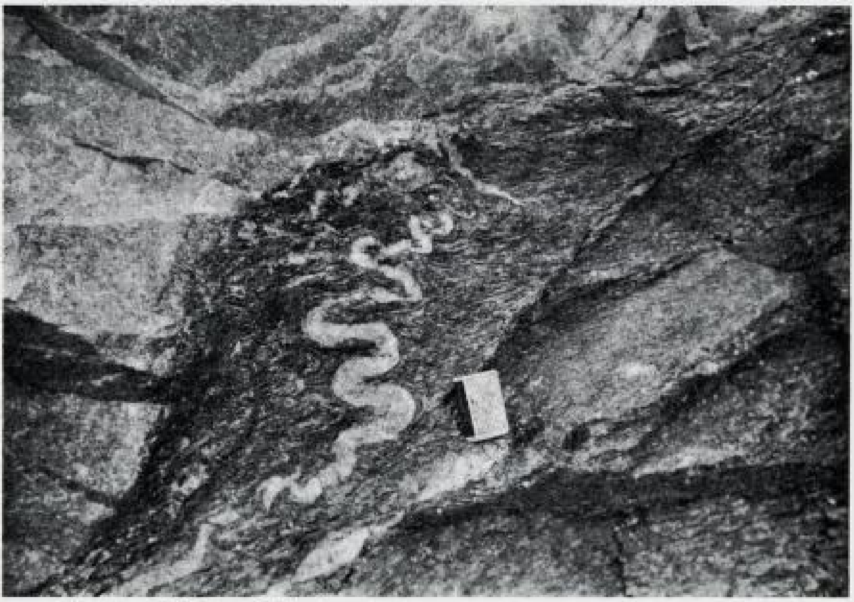

Fig. 11. Folded trondhjemite vein in mica schists at the boundary of the trondhjemite lense shown in Fig. 10.

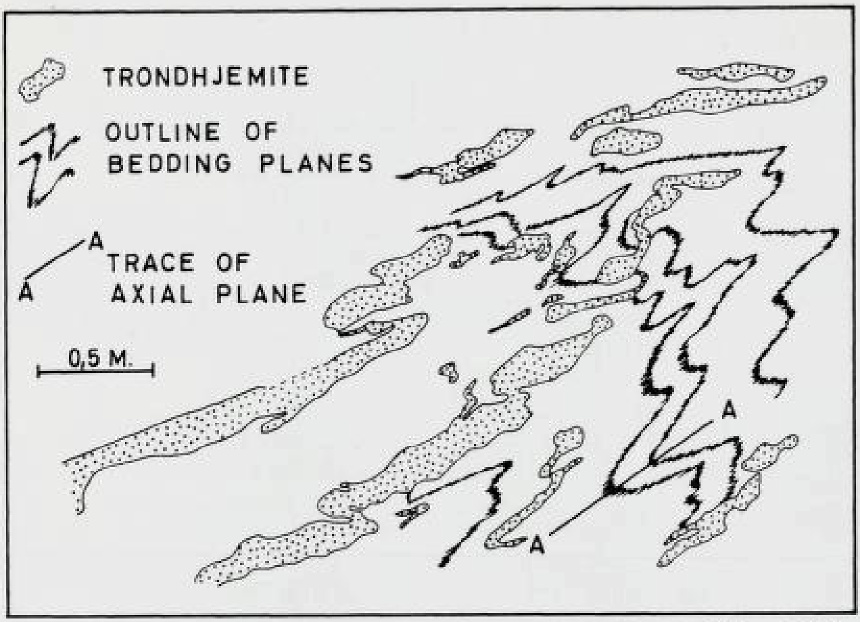

Fig. 12. Sketch of trondhjemite veins in banded quartz-mica schists, Seterfjellet, Målselv. The veins are cross-cutting in relation to the banding (probably primary bedding), but are folded (and boudinaged) together with the schists. Axial plane schistosity is developed in the quartz-mica schist.

(Figs. 11, 12). Boundaries of the larger trondhjemite lenses seem to be parallel or, at least, subparallel with the schistosity of adjoining metasediments. A foliation, primarily due to the parallel orientation of micas, is sometimes observed in the trondhjemites, especially within the border zones. Usually, however, the content of micas is too low to make this foliation particularly pronounced.

Another fact of the utmost importance to the question of the age of the trondhjemites is the presence of *trondhjemite boulders in the Elvenes conglomerate zone*. In the boulders investigated when the petrographical description were written (Gustavson, 1966, p. 101-102) the trondhjemites were texturally of a somewhat special type. Later work has shown that trondhjemites of the usual type for the area is also frequent in the Elvenes conglomerate. In the Harstad conglomerate trondhjemite boulders are predominant and of the same type as trondhjemite intrusives met with east of the town. It is thus evident that some of the trondhjemites, probably most of them, were emplaced before the deposition of the Elvenes-Harstad conglomerates, that is, they were intruded during the geosynclinal stage.

Mineralogy and texture

Most trondhjemites are medium- or coarse-grained rocks. Pegmatitic variants are rare. Foliation is, as already described, faint and confined to the border zones of the bodies. Shear planes with fine-grained and pronounced parallel textures are occasionally observed. The grain size may sometimes show strong variation with large feldspar augens in a fine- or medium-grained groundmass. Microscopic investigations show this feature partly to be due to crushing and recrystallization of a formerly more evengrained and coarse rock, but it can also in some cases be attributed to porphyroblastic growth of feldspars (usually plagioclases). This porphyroblastesis may reasonably be assumed contemporaneous with the porphyroblastic growth of plagioclase in the plagioclase-mica gneisses which has been shown (Gustavson, 1966, p. 133) to be, in part at least, younger than the development of mica schistosity in the gneisses.

As plagioclase and microcline are both frequently white-coloured, it is partly difficult to discriminate between trondhjemitic and granodioritic rocks in the field. Occasionally the potash feldspars are faintly red in colour, but microscopic work is usually necessary for determining the relative amounts of the feldspars. There seem within some of the bodies to exist gradations from rocks devoid of microcline to granodioritic types. It can, however, be stated that the bulk of the investigated leucocratic intrusives in the older group of igneous rocks are trondhjemitic in composition. Granites proper seem to be quite rare in this group.

The mineral assemblages of the meta-trondhjemites can be generally written:

Plagioclase + quartz + epidote/clinozoisite ± microcline ± muscovite ± biotite ± garnet ± orthite ± tourmaline + apatite + sphene + ore minerals.

Micas and epidote are especially abundant among the minor constituents. Epidote (or clinozoisite) and sericitic muscovite are commonly observed as alteration products of the plagioclase.

Under the microscope the common type of trondhjemite shows granoblastic textures, partly with a visible foliation (Pl. V a). Euhedral mineral grains are extremely rare and confined to the minor components like epidote and tourmaline.

Plagioclase. The amounts of plagioclase are in the range 50 to 75 per cent of the rock volume. The grains are anhedral and mostly equant. The grains commonly enclose quartz, micas and accessories and are clearly of metamorphic (and not magmatic) origin. Saussuritization is mostly pronounced. The result of this process is either a fine-grained mass of clinozoisite, sericite and feldspar, or larger grains (partly euhedral) of clinozoisite within the plagioclase.

The composition of the plagioclase range from oligoclase to low andesine. Higher An contents than 35 per cent have not been measured. The values obtained are in agreement with those found in adjoining schists and gneisses and this testify to the premetamorphic emplacement of the trondhjemites. The measurements are based on the maximum extinction on polysynthetic albite twins. These are, together with pericline twins, scarce and the only types of twinning observed. Zoning is not present in the plagioclases, but irregular variations in composition within single grains have been detected. These variations are never amounting to more than a few per cent An.

Evidence of post-crystalline deformation of the plagioclases are present in some bodies. Such evidences are, for instance, fragments formed by disintegration of larger grains, quartz-filled cracks within the feldspars and bent lamellae of albite twins.

Microcline. By definition, potash feldspar never amounts to more than a few per cent in trondhjemites. As these rocks, however, are in some cases gradational into granodiorites, higher microcline contents are found in some thin sections. The amounts rarely exceed 10 - 15 per cent by volume even in the latter rocks, and it is not uncommon to find specimens devoid of potash feldspar. The grains are irregularly outlined and larger individuals may contain inclusions of quartz and accessories. The typical grid twinning is usually well developed. Perthites are not common, but irregularly distributed string perthites are encountered in rare cases. An irregular "flame" perthite, protruding from the boundaries into the microclines, has been observed in a couple of thin sections from a tectonized trondhjemite. As with the plagioclases, granulation due to internal movements can in several cases be seen in microcline grains, partly with development of mortar textures. Muscovitization (sericitization) is observed in a few thin sections.

Quartz grains mostly exhibit strong undulatory extinction. An exception to this rule is the quartz of (apparently) late veins found within some trondhjemite bodies.

Muscovite is the most important minor constituent as far as the amount are considered. The muscovite has two modes of occurrence within these rocks:

a) As separate flakes of some millimetres' grain size. These flakes frequently show parallel orientation contributing to the foliation of the trondhjemites (Pl. V a). Orientation is more at random in the coarser variants than in finer grained rocks.

b) As sericitic mica within plagioclases and, occasionally, in microcline. These flakes are haphazardly orientated or parallel with one of the crystallographic directions of the feldspars.

Biotite is present in some meta-trondhjemites, especially in microcline-bearing ones, but is seldom as abundant as muscovite. The colours and pleochroism are in various shades of brown and red-brown. Partial chloritization has been observed.

Epidote minerals. Epidote, or more commonly, clinozoisite is present in all the investigated specimens. Most grains are small and occurring within the plagioclase but some are 1-2 mms across and found along grain boundaries or within plagioclase. These larger grains are occasionally euhedral crystals. In some cases cores of *orthite* are present. Zonal growth is not rare. Orthite is not as common as in the granodiorites described in chapter III.

Among the accessories *apatite* seems to be commonly present in small amounts. Usually some ore mineral is also encountered. At least in some rocks the ore mineral is *pyrite*. *Sphene*, minute *garnets* and black *tourmaline* are found in some specimens. Tourmaline seems to be restricted to quartz-rich veins and some coarse variants of the trondhjemites.

Table XI

Chemical composition, molecular norm, mode and Niggli values of trondhjemite,[1]) *Ballangen.*

	Wt %	*Cat %*	*Molecular norm*		*Mode*		*Niggli values*	
SiO_2	68,64	64,1	Q	25,1	Quartz	26,8	al	45
TiO_2	0,26	,02	Or	11,5	Microcline	tr.	fm	13
Al_2O_3	16,98	18,7	Ab	42,5	Plagioclase	50,0	c	16
Fe_2O_3	1,14	0,8	An	13,5	(% An in		alk	26
FeO	0,71	0,6	(% An		plag.)	(15)	si	308
MnO	0,02	—	in plag.)	(24)	Epidote	4,0	k	0,21
MgO	0,87	1,2	C	2,5	Biotite	4,5	mg	0,47
BaO	0,06	—	Hy	2,4	Muscovite	12,9	qz	+ 104
CaO	3,24	3,2	Mt	1,2	Magnetite	0,3		
Na_2O	4,69	8,5	Il	0,4	Sphene	0,6		
K_2O	1,91	2,3	Ap	0,3	Apatite	0,3		
CO_2	0,27	0,3	Cc	0,6	Calcite	0,6		
P_2O_5	0,11	0,1	(Col.		Orthite	tr.		
S	0,01	—	index)	(4,9)	Pyrite	tr.		
H_2O+	0,86	—			(Col. index)	(23,2)		
H_2O-	n.d.	—						
Sum	99,77	100,0						

Locality of the specimen: Bruksjord, Ballangen. Analysis completed for S. Foslie. Analyst: M. Klüver.

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[1]) Foslie's designation. The rock is actually a quartz diorite, as the amount of dark minerals is a little too high for a trondhjemite.

Chemistry

Meta-trondhjemites have not been analysed during the present study. A previously unpublished analysis, completed for S. Foslie, has been listed in Table XI. It is, in all components, similar to earlier trondhjemite analyses published from the Norwegian Caledonides (see for inst. Goldschmidt, 1916).

Mineral facies

Mineralogy as well as textures testify to the strong influence of regional metamorphism on these trondhjemites. The assemblage, quartz + microcline + oligoclase/andesine + epidote + muscovite (± biotite), place the rocks in the lower part of the almandine-amphibolite facies. As the bulk of the meta-trondhjemites are encountered within gneisses and schists of the kyanite and staurolite zones (Gustavson, 1966) there is general agreement between the regional metamorphic pattern and the trondhjemite assemblages. It therefore seems obvious that the present trondhjemites were all emplaced previous to the main regional metamorphism.

III. Petrography and metamorphism of the younger group of igneous rocks

The Råna massif

The Råna igneous massif embraces a group of rocks varying in composition from ultrabasic to granitic types. The reason for treating them together here is that they have previously been thoroughly described by Foslie (1921, 1922, 1941). In the following a summary of the rocks, based on Foslie's papers, is given.

The shape of the large, synformal body and its present outcrop seems largely to be determined by E-W orientated fold structures. (Fig. 13 and Foslie 1921, p. 707). In the opinion of the present writer these belong to a late fold episode, post-dating, for instance, the period of progressive regional metamorphism. The lense-shaped body (Foslie, 1921, p. 704) is composed of a central and upper mass of quartz norite and a marginal and lower zone of norite with bodies of various ultrabasic rock types within (Fig. 13). The boundary between norite and quartz norite is one of "gradual, though rapid transition". Dykes of aplitic granodiorite and trondhjemite as well as irregular veins of pegmatitic potash granite transsect the main rock types of the massif.

The ultrabasics within the norite comprise bands or bosses of troctolite, lherzolite and olivine norite. Minerals present are olivine, enstatite, diallage, bytownite (about An_{80}) and accessory amounts of spinel, magnetite and pyrrhotite.

The norite is composed of plagioclase An_{50} to An_{80}, frequently zoned, with hypersthene, diallage and some biotite. Secondary actinolite, formed by alteration of pyroxene are found in some parts of the mass. Variations in the composition of the norite are due to variations in the relative proportions of plagioclase to pyroxene. The plagioclase content may be as low as 20 per cent, but is usually higher, 30 to 40 per cent. Hypersthene amounts to 50 per cent, or more, in most cases.

The quartz norite can be subdivided into several types due to variations

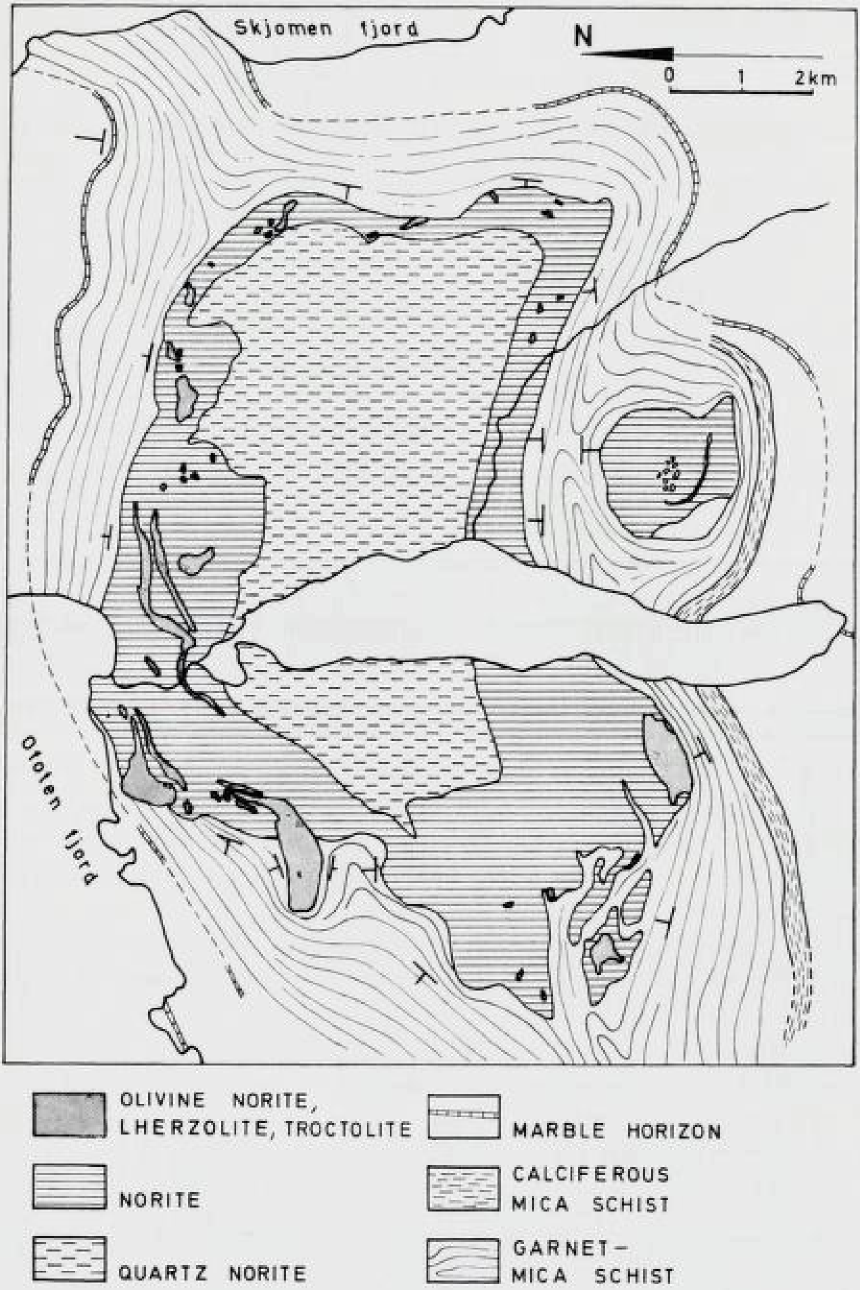

Fig. 13. Map of the Råna massif, redrawn from S. Foslie (1921).

in texture and, to some degree, also to different degrees of uralitization of the pyroxene. Mainly, however, the quartz norite contains zoned plagioclase An_{40} to An_{55}, hypersthene, diallage, biotite, quartz and small amounts of orthoclase. While the hypersthene of the norite has a composition with about 25 per cent iron silicate, the hypersthene in the quartz norite contains 40 to 45 per cent ferrous silicate. The plagioclase content is considerably increased in the quartz norite and can be as high as 60 per cent. Contrary to the relations in the norite, the diallage content is higher than the hypersthene content, amounting to 15-20 per cent. The amounts of biotite is generally increased.

The younger, aplitic, dykes are composed of oligoclase An_{26}, microcline, quartz, muscovite and biotite. Judging from the brief descriptions given (Foslie, 1921, 1922) they vary in composition from trondhjemitic to granitic types. The pegmatitic veins usually have the composition of micro-

Table XII

Mean chemical composition of the main rock types of the Råna massif, as given by Foslie (1921).

Wt %	(a)	(b)	(c)	(d)
SiO_2	41,53	50,80	55,90	72,20
TiO_2	0,10	0,44	0,45	0,39
Al_2O_3	5,29	16,99	17,50	14,22
Fe_2O_3	2,40	0,44	0,08	0,55
FeO	9,64	6,47	6,97	2,00
MnO	0,24	0,14	0,09	0,03
MgO	36,98	13,20	5,98	1,03
CaO	1,85	8,43	8,32	2,60
BaO	0,00	tr.	tr.	0,02
Na_2O	0,59	1,43	2,75	2,55
K_2O	0,49	0,60	1,50	3,78
P_2O_5	0,07	0,03	0,05	0,08
Vd_2O_3	n.d.	0,09	n.d.	n.d.
S	0,20	0,06	0,05	tr.
Cr_2O_3	0,08	0,09	0,00	n.d.
Sum	99,46	99,21	99,64	99,45

(a) Olivine-bearing rocks (ultrabasics).
(b) Norite.
(c) Quartz norite.
(d) Aplitic dykes.

cline granites. In addition to microcline, oligoclase and quartz they characteristically contain black tourmalines. The mean chemical compositions of the various rock types were calculated by Foslie (1921, pp. 709-710). Two analyses of ultrabasic rocks, three of norites, one of quartz norite and one of an aplitic dyke were used for this purpose. These compositions are listed in Table XII.

According to Foslie (1921) the Råna massif is an example of rocks formed by magmatic squeezing differentiation. The "driving force" in this process was supposed to be orogenic folding and lateral pressures operating during the supposed syntectonic or late-tectonic crystallization of the magma. We shall return to the genetic aspects of the Råna massif in chapter IV of this paper.

Peridotites outside the Råna massif

Occurrence and field relations

Alpine-type serpentinites have been described as part of the older igneous group. Younger ultrabasics can be subdivided into a) Ultrabasics within the Råna massif, and b) Peridotites outside the massif, supposed to be of corresponding age. The former have been briefly mentioned above, while some statements of the latter are given below. These peridotites are restricted to five occurrences in the area south of the Råna massif and one in the Bardu district, here called the Orta peridotite:

The southern occurrences (Foslie, 1941) are situated at approximately the same stratigraphical level as the Råna massif, shortly below the Melkedal limestone, and are encountered some few kilometres south of the massif (map, Fig. 1). They are relatively small, the largest being 150 by 45 metres in outcrop. The country rocks in general are mica schists, but one of the bosses is described by Foslie (1941, p. 149) as surrounded by "actinolite gabbro". It is of some interest that the latter peridotite, as well as the actinolite gabbro, is transsected by numerous veins of plagioclase-rich pegmatite, thus indicating similar age relations between basic and acid rocks as found within the Råna massif.

The Orta peridotite is encountered N. of Setermoen, Bardu, about one kilometre W. of the mountain Storala. Approaching the locality from south 3 small peridotites are first met with, the boundaries of which are obscured by thick cover. Shortly to the east of these a fourth body is encountered. This one is important from a genetical point of view since it grades across the strike into gabbroic and dioritic rocks without sharp contacts between them. It is interesting to recall in this connection the statement of Foslie

(1921, p. 705) that the basic and ultrabasic rocks of the Råna massif "form a stepwise, but nearly continuous, series without definite eruptive contacts, . . .".

At Orta the thickness across the strike of ultrabasic + basic rocks may reach about one hundred metres, while the strike extension is unknown due to cover. The boundaries are against coarse and steeply dipping plagioclase - mica gneisses on the southern side and fine-grained, partly mylonitic, schists on the northern side. The country rocks have been assigned to the Narvik Group: Some 2 kilometres east of the peridotite and approximately in its strike direction, a thin marble horizon occurs, tentatively correlated (Gustavson, 1966, p. 76) with the Melkedal Limestone in the middle of the Narvik Group. The stratigraphical position, therefore, is approximately the same as for the other peridotites of the younger group.

Mineralogy and texture

The most ultrabasic rock type of the Orta locality is dunite and contains nearly 100 per cent olivine. Other variants consist of olivine, augitic pyroxene, plagioclase, iron ore and alteration minerals. These variants grade into hornblende gabbro and diorite. The transformation of pyroxene into hornblende increases from the basic to the intermediate members. In the ultrabasic types amphibole is not present. Some features of the minerals are given below:

Olivine. Grain size is about 1 - 2 mms. The grains have a rounded appearance. They are transsected by numerous irregular fractures, but the degree of alteration is unusually low as compared with peridotites belonging to the older group. Serpentine (antigorite) occurs sparingly along cracks and grain boundaries. The optic sign is (+) or (—), the axial angle about 90°. These data show the composition to be about Fo_{88} Fa_{12} (Trøger, 1959).

Clinopyroxene occurs in small pockets between olivine grains and is also partly enclosing round olivines. On a minute scale the boundaries against olivine are irregular. The immediate impression is that the pyroxenes are growing into the olivine grains and are replacing them. Frequently, however, there is a narrow zone of antigorite and talc between pyroxene and olivine. The pyroxene is light brown and slightly pleochroic. Birefringence is medium strong. The prismatic cleavages are well developed, while the extinction is incomplete at an angle between 40 and 45 degrees. These data taken into account, it seems probable that the clinopyroxene has an augitic composition.

Plagioclase occurs in small grains and is relatively scarce. The composition is probably that of bytownite, though more exact measurements are lacking.

Ore grains are present as inclusions in olivine and as minute grains within talc masses. It is probably magnetite in the former case, while the identity of the latter grains is unknown.

The *alteration minerals* are *talc* and *antigorite.* No definite age difference between the two minerals can be detected. They are about equally important, occurring along rims and fractures in olivine and pyroxene grains. The antigorite is seen as well-bounded flakes of 0.2 - 0.5 mms size, while the talc is present in irregular clusters of minute flakes.

There are no evidence of strong deformation in the present peridotite, neither on the megascopic nor on the microscopic scale. This feature it shares with the gabbroid and dioritic rocks.

Chemistry

Chemical analyses of the present peridotites are not available. As Ca-bearing minerals (for instance clinopyroxene and plagioclase) seem to be more frequent in the younger ultrabasics than in those of the older group (see Table II) they must obviously contain more CaO. This conclusion seems to be valid also for the peridotites of the Råna massif as compared with older peridotites of that area: The figures given in Table XII are the average of two analyses, showing 1,48 and 2,22 weight per cent CaO, while the CaO content in the analysed rock of Table II is 0. It seems probable that also Na_2O and Al_2O_3 are higher in the younger group as compared with the older ultrabasics. It should be noted that the chemical differences correspond with those found by Hess (1955) to be valid for alpine-type peridotites as compared with the Bushweld type ultrabasics.

Mineral facies

Only slight alteration of the primary minerals into serpentine and talc can be recognized in the Orta peridotites. Any age relations of the secondary minerals, like those found in the older group of ultrabasics, can not be established in the present case, in fact, it seems probable that they have been formed contemporaneously, as part of the same process. The presence of serpentine and talc together seems to indicate that alteration took place at temperatures below those prevailing during the metamorphism of the adjoining schists, as these show almandine-amphibolite facies assemblages. Thus alteration must have occurred either before the peak of regional metamorphism (like it was assumed for the older group of serpentinites) or subsequent to the main metamorphism. The incomplete metamorphic reconstruction of gabbros and diorites, including the small bodies with which the peridotites are closely associated, indicates that all these rocks were held at elevated temperatures for a relatively short period. Tectonic influence seems relatively unimportant, although the steep dip of the composite Orta body may be due to folding about E-W axes, possibly the same phase which influenced the Råna norite massif. It may thus be in best accordance with the observed facts to assume that the slight *alteration of the present ultrabasics occurred after the peak of regional metamorphism and the most important fold movements.*

Diorites and related rocks

Occurrence and field relations

a) The most important occurrence of dioritic rocks is the Istind diorite (map, Fig. 1), situated southeast of the Bardu river, some kilometres northeast of Setermoen. In addition, similar rocks are known from a few occurrences in Bardu, Salangen and Andørja, usually too small to be represented on the map. The similarities of the diorites with rocks found within the Råna and Sørfjorden norite massifs, especially with some of the quartznorites, will be obvious from the following description, as will be the apparent differences from amphibolitic rocks described in chapter II.

The Istind diorite forms the upper part of the Istind mountains. Its lower boundary is seen as a slightly curved line from about 1000 m.a.s.l. in the steep northern slopes of W. Istind (1490 metres high) to the peak of St. Istind (1487 metres high), approximately. The boundary against the meta-sediments has not been directly observed in outcrop but in general it seems to run parallel with the foliation of mica gneisses below. The gneisses belong to the Narvik Group.

The main rock of the massif, as it can be studied along the path to the peak of W. Istind, is a massive and fairly homogeneous, middle grained, green- and white-spotted rock. The textures, as observed in hand specimen, are non-directional. Breccia-like structures, with pieces of a darker gabbroic variant "floating" in the main rock type, are occasionally met with. Dark, micaceous garnet-bearing portions with diffuse boundaries against diorite are found within the lower part of the massif and possibly represent xenoliths of the adjoining mica gneiss. In the upper part of W. Istind small and scattered inclusions of peridotite are met with. That of Fig. 14 is typical of these. The boundaries between peridotite and diorite are sharp, and, as shown in Fig. 14, further emphasized by weathering.

b) It has already been mentioned how the Orta peridotite is connected by gradual transition with gabbroic and diorite rocks. The diorite shows close resemblance to the (main) Istind rock. In hand specimen the two rocks seem nearly identical. As described later the similarities also concern microscopic details.

c) Massive, dioritic rocks were also observed on the two mountains Nap-

Fig. 14. Peridotite inclusion in the Istind diorite, W. Istind, about 1300 m a.s.l.

pen, Andørja and Ratberget, Bardu. In hand specimen they show resemblances to the Istind diorite and differ from amphibolites of the same local areas in their undeformed, nondirectional structure. These bodies are obviously small, although their boundaries have not been observed. The adjoining gneisses are in both places assigned to the Narvik Group.

d) More dubious in origin are some occurrences in Salangen of a massive, unfoliated rock composed of diopside, actinolite, biotite and minor amounts of plagioclase, microcline and clinozoisite. In one locality, at Skårvikelven, it occurs as a 5-6 metres thick zone within mica schists of the Salangen Group. In two other localities, along the road to Bekkebotn, the same rock occur as a 7-8 m thick zone in contact with granite. The granite and the basic rock both follow the local dip and strike of the mica schistosity in adjoining metasediments. In the two localities it can also be observed that the granite intrudes the more basic rock (Fig. 15) and thus is younger. Small portions of the basic rock are composed of biotite + plagioclase and have a pegmatitic grain size. The pegmatitic portions are mostly less than one decimetre across and occur irregularly distributed. The pyroxene-amphibole-biotite rock may be related to the basic and ultrabasic rocks under discussion in this chapter, but it may also possibly be a sort of scarn rock eventually related to granite intrusion. Its non-directional textures point to it being younger than amphibolites in the vicinity as these are all strongly foliated rocks.

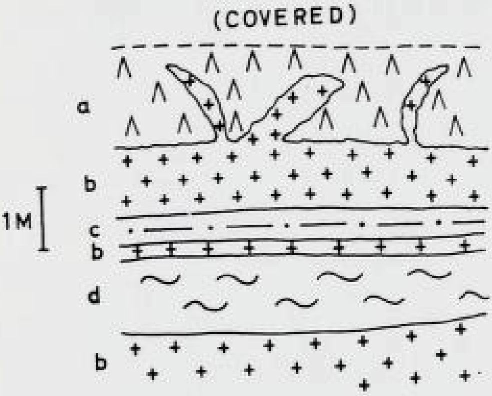

Fig. 15. Granite dykes (b) intrusive in relation to "diorite" (a)(at Høglund, Salangen. The granite, in this locality, is concordant against the metasediments. (c = garnet-mica schist. d = magnetite-bearing mica schist).

Mineralogy and texture

As already mentioned, the Istind diorite is a massive, middle-grained rock. The mineral composition is plagioclase (An_{40-47}) + clinopyroxene + amphibole ± biotite ± quartz ± epidote ± sphene + ore minerals. The three first-mentioned minerals are discernible in hand specimen. Under the microscope non-directional, "igneous" textures are obvious. In general these textures can be classified as "hypautomorphic-granular", which according to Johannsen (1955, p. 37) are "the most common in the plutonites".

The ultrabasic inclusions are composed of olivine, orthorhombic and monoclinic pyroxenes, green spinel and an ore mineral (pyrrhotite?). As the olivine occurs as roundly shaped grains within pyroxene, the texture is probably best described as "poikilitic" (see Johannsen, 1955, p. 50) (Pl. V b).

Some features of the minerals of diorite and ultrabasics are given below:

Olivine occurs as "rounded" grains either as poikilitic inclusions (Pl. V b) in clinopyroxene or surrounded by a rim of orthorhombic pyroxene which again may be enclosed in clinopyroxene. The age sequence is thus: Olivine → orthorhombic pyroxene → clinopyroxene. Spinel is partly found within the olivine grains and may be the first mineral that crystallized. The olivine has $2V_X$ between 80 and 90 degrees which serves to classify it as a chrysolite (Trøger, 1959). No signs of serpentinization of the olivine are observed, a rather remarkable feature.

Orthorhombic pyroxene has been observed only in the ultrabasic inclusions. In most cases it is found as a rim of irregular grains, or in a radial arrangement, on olivine. This mode of occurrence is the same as described by Foslie (1922, p. 18) from the ultrabasics of Råna. The orthorhombic pyroxene is optically positive and colourless and is probably *enstatite.*

Clinopyroxene. Irregular (rarely euhedral) crystals, up to 5 mms. in length, enclose olivine and orthorhombic pyroxene in the ultrabasics. In the diorite it is the only pyroxene. It is partly found enclosed in, partly enclosing, plagioclase and it is probably of about the same age as the feldspar. Transformation of the clinopyroxene into amphibole is pronounced along rims (Pl. VI a) and in several cases only small remnants of the pyroxene are left within the amphibole. Opt. data: $2V_z$ is about 60°, extinction c/Z about 40°, axial dispersion clearly visible and $r > v$. These data are typical for *diopsidic* pyroxenes.

Plagioclase. The grains are equidimensional to tabular with flattening parallel to 010 (Pl. VI a). Range of grain sizes is 0,1 - 3 mms. in the investigated thin sections. Crystals faces as well as irregular boundaries are observable. The crystals show no preferred orientation. Twinning is present in all crystals. Albite twins are the most common type, but pericline as well as combined albite/Carlsbad twins are encountered. Zonal growth is common with the most basic composition in the core. The composition determined on extinction angles in sections perpendicular to 010 lies between An_{47} and An_{40}.

Amphiboles seem always to be *uralitic* in the Istind diorite. The shape of pyroxene is partly retained even where no remnants of the primary mineral is left. The colour of the amphibole is pale yellow brown (occasionally greenish), with pleochroism ranging from yellow brown to colourless. Extinction c/Z is 19°, 2V close to 90° and axial dispersion is visible. The grains are length-slow. The highest refractive index is approximately 1.640, and the birefringence moderate. It seems probable that the amphibole belongs to the *actinolites*. A pale brown amphibole is mentioned by Foslie as an uralitization product in the Råna norite and he arrives at a similar conclusion about the composition.

Biotite flakes show no preferred orientation. Strong red-brown colour in the Z direction is characteristic. The amounts vary from zero to about 10 per cent.

Quartz grains frequently occur together with the amphibole and is then obviously related to the uralitization process. However, quartz probably was also primarily present, as it is part of the norm (Table XIII).

Sphene is found as rims on ore grains which probably are *ilmenite*. The amounts of both minerals are fairly small.

Epidote grains are present in the analysed specimen (Table XIII). It shows no direct relation to the plagioclases but occurs most commonly along the borders of altered pyroxenes. The grains are small and of irregular shape.

The other occurrences of dioritic rocks show similarities to the Istind diorite in texture and mostly also in mineral composition. The Salangen "diorites" are different in that they contain microcline in significant amounts. The microcline encloses remnants of plagioclase and it seems possible that the K-feldspar was introduced in connection with intrusion of the younger granite dykes (Fig. 15).

Chemistry

In Table XIII the chemical composition of the Istind diorite is compared with a diorite from the Sørfjorden "norite" massif and a quartz norite from the Råna massif. The close similarities of the three rocks in nearly all chemical components are striking. From a genetic point of view, this is a fact of great interest as will be further discussed.

Table XIII

Chemical composition, molecular norms, modes and Niggli values of Istind diorite (a), Sørfjorden diorite (b) and Råna quartz norite (c).

	(a)		(b)		(c)	
	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %
SiO_2	54,19	50,4	53,48	50,1	55.90	51.6
TiO_2	0,96	0,7	0,45	0,3	0,45	0,3
Al_2O_3	18,04	19,8	17,16	18,9	17,50	19,0
Fe_2O_3	0,64	0,4	2,47	1,8	0,08	0,1
FeO	5,95	4,6	3,99	3,2	6,97	5,4
MnO	0,11	0,1	0,13	0,1	0,09	0,1
MgO	5,64	7,9	6,97	9,7	5,98	8,2
CaO	9,50	9,5	11,41	11,3	8,32	8,2
Na_2O	2,71	4,9	2,28	4,2	2,75	4,9
K_2O	1,29	1,6	0,31	0,3	1,50	1,7
CO_2	—	—	—	—	0,20	0,3
P_2O_5	0,12	0,1	0,04	—	0,05	0,1
S	n.d.	—	0,04	0,1	0,05	0,1
H_2O+	0,77	—	1,29	—	0,50	—
H_2O-	0,11	—	0,04	—	0,07	—
Sum	100,03	100,0	100,06	100,0	100,41	100,0

Molecular norms:			
Q	3,3	6,4	4,7
Or	8,0	1,5	8,5
Ab	24,5	21,0	24,5
An	33,3	36,0	31,0
(% An in plag.)	(58)	(63)	(56)
Hy	18,2	15,2	23,4
Di	10,4	16,4	6,0
Mt	0,6	2,7	0,2
Il	1,4	0,6	0,6
Py	—	0,2	0,2
Ap	0,3	—	0,3
Cc	—	—	0,6
(Colour index)	(30,9)	(35,1)	(31,3)

Modes[1]):

Quartz	6,1	10,0	2,3
Orthoclase	—	—	5,1
Plagioclase	49,5	26,2	63,5
(An in plag.)	(40-47)[2])	(29-43)[2])	(40-55)[2])
Clinopyroxene	12,1	—	14,5
Hypersthene	—	—	9,5
Amphibole	16,3	51,3	tr.
Biotite	9,4	0,5	3,6
Epidote	2,8	12,0	—
Sphene	2,8	—	—
Apatite	0,4	—	—
Ore min.	0,6	—	1,5
(Col. index)	(42,4)	(63,8)	(29,1)

Niggli values:

	al	fm	c	alk	si	k	mg	qz
(a)	27,5	36,5	26,5	9	141	0,24	0,60	+ 5
(b)	25	39	30	6	132	0,07	0,66	+ 8
(c)	27,5	39,5	23,5	9,5	148	0,26	0,60	+ 10

Localities of the analysed specimens:

(a) W. Istind, Bardu. Anal. R. Solli.
(b) Mellomslåttfjell, Indre Tysfjord. Anal. M. Klüver.
(c) Stemnes, Råna. Anal. O. Røer.
Analyses (b) and (c) completed for S. Foslie.

[1] Calculations by the present writer.

[2] Zoning.

Mineral facies

Uralitization of the pyroxene is described by Foslie (1922, p. 19) as a very common process in the Råna and Sørfjorden norites. It is less common in the quartz norites but occurs also there. The mineralogical differences between the three rocks listed in Table XIII are therefore, when the nearly identical chemistry are taken into account, best explained by varying degrees of alteration in the three rocks. Uralitization is, as described previously in this paper, relatively frequent in the Istind diorite, although nearly unaltered parts have been observed. As orthorhombic pyroxene seems to be uncommon most of the amphibole were evidently formed from clinopyroxene.

Diopsidic pyroxene seems also to be the only main mineral taking part in the alteration, and the uralitization can probably be expressed adequately by a simple equation as

$5\,Ca\,(Mg,Fe)Si_2O_6 + H_2O \rightarrow Ca_2(Mg,Fe)_5Si_8O_{22}(OH)_2 + 2\,SiO_2 + 3\,CaO.$

This is in accordance with the observation that much of the quartz is related to the formation of amphibole. The CaO and SiO_2 released by this process may partly have been absorbed in the production of small amounts of sphene from ilmenite:

$FeTiO_3 + CaO + SiO_2 \rightarrow CaTiSiO_5 + FeO$

and in the formation of epidote minerals (simplified):

$4\,CaO + 6\,SiO_2 + Al_2O_3 + H_2O \rightarrow 2\,Ca_2Al_3Si_3O_{12}(OH)$

Iron in the epidote may eventually originate from the alteration of ilmenite into sphene. It has previously been mentioned that the epidote shows no obvious relation to plagioclase in the investigated specimens.

The incompleteness of metamorphic alteration in the Istind massif is shown by the following observations:

1. Uralitization of pyroxene is only partial.
2. The composition of normally zoned plagioclases varies from An_{47} to An_{40} which is considerably higher than to be expected from the metamorphic facies of the adjoining schists (low amphibolite facies). The plagioclases in usual must be assumed to be primary magmatic.
3. In ultrabasic inclusions the association olivine + orthorhombic pyroxene + clinopyroxene shows no sign of metamorphic transformation.

In summary it seems difficult to avoid the conclusion that *the emplacement of the Istind massif was later than the main regional metamorphism* of this area. This is further supported by the general absence of deformation phenomena within the massif. Furthermore, the amphibole formed by uralitization of the pyroxene is actinolite. Actinolite has a fairly wide range of stability and its presence gives no exact information about the conditions at the time of its formation. It seems probable, however, that it was formed at lower temperatures than those prevailing during the strong regional metamorphism which has been recognized by study of the meta-sediments (Gustavson, 1966, p. 154).

Granodiorites and granites

Occurrence and field relations

These rocks are encountered in the meta-sediments from the Narvik Group upwards. An occurrence of granite/granodiorite at Tennevoll, Lavangen (P. R. Lund, 1965, pp. 92-95) is situated within schists which have been reckoned to the Rombak Group. Together with a couple of dykes in the Brandvoll area these are the only occurrences of Caledonian igneous rocks found in the Rombak Group.

Most bodies depicted on the map are granodioritic in composition although gradations to granitic types occur, as for instance in the mentioned Tennevoll occurrence. True granites are also found as relatively small dykes and veins within the same areas as the larger granodiorite bodies. Trondhjemite is a very subordinate member of the younger igneous group. Foslie (1922, 1941) describes some trondhjemites cutting basic rocks of the Råna massif but in the rest of the map area most trondhjemites seem to belong to the older group.

The greater granodiorite/granite bodies are present either as

a) Elongate, sill-like intrusions, or

b) Irregularly shaped bosses.

The former of these modes of occurrence is rather common: Some of them can be followed along the strike for several hundred metres, the thickness seldom exceeding some few tens of metres. It is not rare to find these bodies slightly cross-cutting in relation to stratigraphy. Examples of this type are, for instance, met with at Storhaugen. Salangen (below the iron ore horizon) and in the eastern slopes of Rundfjellet, Lavangen. The boundaries are concordant or subconcordant with the mica schistosity, but a slightly discordant behaviour of these rocks are occasionally shown as a sort of interfingering with adjoining mica schist. A generalized sketch of these relations is shown on Fig. 16.

Dykes of smaller dimensions are usually discordant in relation to the schistosity as well as to bedding planes (Figs. 17, 18). These planar structures are cut at all angles. Slight folding of the dykes are occasionally evident. However, it seems obvious that *deformations subsequent to dyke*

Fig. 16. Boundary relations (interfingering) of granite and mica schist, W. of Storhaugen, Salangen.

emplacement were less penetrative than previous phases. Observations in some localities (Fig. 17) clearly prove the younger age of granodioritic rocks as compared with meta-trondhjemites. That they are also later than basic rocks of the younger group is indicated by relations already referred to, in Råna and Salangen (Fig. 15).

Mineralogy and texture

The granodiorites and granites are medium- to coarse-grained rocks Pegmatite-textured granitic dykes are rare. Parallel orientation of mica (mostly biotite) may be visible in hand specimen but is usually absent. Under the

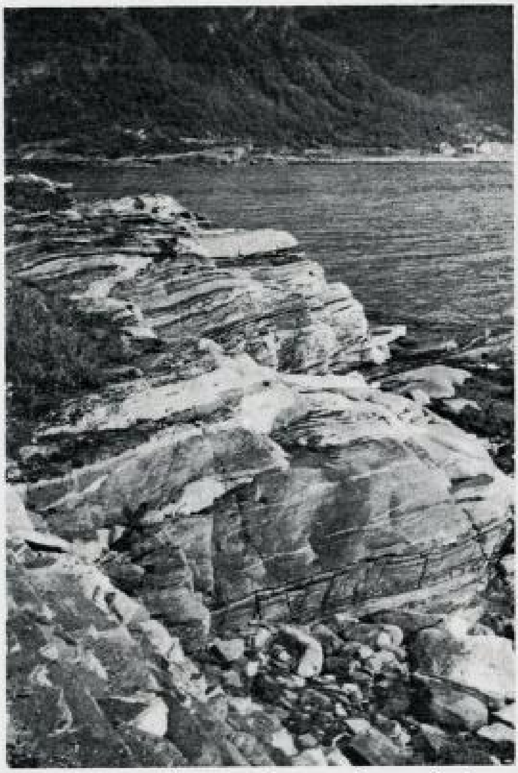

Fig. 17. Meta-sediments with concordant and subconcordant trondhjemite dykes cut by younger granodiorite dyke (x-x), Faksfjord.

Fig. 18. Granite dyke cutting chistosity and fold structures in magnetite-bearing mica schist at the Storhaugen iron mine, Salangen.

microscope some preferred orientation is visible but as a rule little pronounced. Post-crystalline crushing of minerals is occasionally observed. However, primary magmatic features are more common than in acid rocks of the older group. This concerns especially the plagioclase feldspar as will be described below.

The assemblages can be generally written as:

Microcline + plagioclase + quartz + epidote + orthite + biotite ± muscovite ± pyrite ± magnetite ± sphene ± apatite.

Minerals found only in the pegmatitic variants are calcite and tourmaline. Remarkable is the relative abundance of yellow orthite (Pl. VI b) in all thin sections, mostly as cores within epidote. Biotite is more abundant than muscovite and is present in all granodiorites and granites.

The common presence of orthite and the relative proportions biotite: muscovite are different from meta-trondhjemites of the older group. Garnets are generally absent from the present rocks.

Microcline. The grains are irregulary outlined and may contain inclusions of the minor constituents. Typical microcline twinning is present. Perthite has not been observed. Neither has sericitization, as found in microclines of the meta-trondhjemites played any role in the present rocks.

Plagioclase. Most grains are irregular in shape, but nearly euhedral laths are occasionally preserved as shown on Pl. VII a. Carlsbad twins, combined with albite twins, are found in the latter type of plagioclase. Albite twins are met with in the anhedral plagioclase. Some saussuritization can be observed in euhedral plagioclase as well as in anhedral ones. Uneven extinction or normal zoning is partly present, mainly in the euhedral or subhedral types. Compositions are mostly oligoclase (see modes of analysed rocks, Table XIV) but albite, $An_{8\text{-}10}$ has been determined in some cases, for inst. in (c) of Table XIV).

Epidote is present as an alteration product in plagioclase and as separate grains outside the feldspars. In the latter grains cores of orthite are commonly present (Pl. VI b). This feature seems to be characteristic for the granodiorites but is not equally common in the granitic dykes.

Biotite of strong red-brown, brown, or (more rarely) greenish colours are present in considerable amounts in all specimens. In most cases it dominates over muscovite. (Exceptions are, for instance, the analysed rocks (b) and (c) of Table XIV.)

Muscovite, together with biotite, partly shows a preferred orientation.

Other minerals show no features of special interest. *Tourmaline* (black) is locally abundant in pegmatitic granite. Zoning, as found in tourmalines of the meta-sediments, is not encountered. *Calcite* is partly present in the same rock type.

Chemistry

Three full chemical analyses are shown in Table XIV. Alkali and SiO_2 determinations of three additional specimens are listed in Table XV. Although K_2O is higher than Na_2O in three of the six specimens, plagioclase dominates over potash feldspar also in these rocks because part of the K_2O is bound in micas.

Mineral facies

The assemblages of the granodiorites and granites are: Quartz + microcline + albite/oligoclase + biotite + muscovite + epidote. This is not inconsistent with the metamorphic grade of the adjoining schists which is in the epidote - amphibolite and the almandine - amphibolite facies (Gustavson, 1966, pp. 129-130 and map, Fig. 2). However, if the plagioclase compositions in equilibrium with epidote in the granodiorites are compared with those of older meta-trondhjemites in the vicinity, somewhat lower An contents seem to be present in the former rocks. This indicates that the saussuritization of the granodiorites took place at lower temperatures than those prevailing at the peak of metamorphism in the respective areas. Zoning of the plagioclases in some granodiorites points to a certain disequilibrium (Barth, 1956). Alteration at declining temperatures and resulting low reaction rates may be a reasonable explanation of this disequilibrium. The relations to the phases of strong folding as already referred to (p. 68) also indicate that the emplacement of these rocks occurred at a relatively late stage in the history of the area.

Table XIV

Chemical composition, molecular norms, modes and Niggli values of two granodiorites (a, b) and one granite (c).

	(a)		(b)		(c)	
	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %
SiO_2	67,77	63,3	66,86	61,7	71,92	67,7
TiO_2	0,64	0,4	0,60	0,4	0,26	0,2
Al_2O_3	16,06	17,6	17,94	19,5	14,29	15,9
Fe_2O_3	0,47	0,3	0,68	0,4	0,39	0,3
FeO	2,29	1,8	0,98	0,7	1,99	1,6
MnO	0,05	0,1	0,02	—	0,01	—
MgO	1,19	1,7	0,93	1,3	0,44	0,6
CaO	2,41	2,4	2,86	2,8	1,37	1,4
Na_2O	4,10	7,4	5,16	9,2	3,31	6,0
K_2O	3,98	4,8	3,31	3,9	5,16	6,2
CO_2	0,05	—	0,01	—	0,07	0,1
P_2O_5	0,19	0,2	0,15	0,1	0,08	—
H_2O+	0,69	—	0,43	—	0,47	—
H_2O-	0,07	—	0,08	—	0,08	—
Sum	99,96	100,0	100,01	100,0	99,84	100,0

Molecular norms:			
Q	19,5	15,8	26,7
Or	24,0	19,5	31,0
Ab	37,0	46,0	30,0
An	10,5	13,0	6,5
(% An in plag.)	(22)	(22)	(18)
C	1,2	1,2	1,1
Hy	6,0	2,8	3,6
Mt	0,5	0,6	0,5
Il	0,8	0,8	0,4
Ap	0,5	0,3	—
Cc	—	—	0,2
(Col. index)	(7,8)	(4,5)	(4,7)

Modes:			
Quartz	23,7	17,0	29,0
Microcline	15,0	11,5	22,9
Plagioclase	45,5	57,0	33,5
(% An in plag.)	(19	(19)	(10)
Epidote	tr.	tr.	1,6
Biotite	10,9	5,4	5,4
Muscovite	2,7	7,0	6,3
Magnetite	0,5	0,6	0,5
Pyrite	—	tr.	—
Sphene	1,2	1,2	0,6
Orthite	+	+	+
Calcite	tr.	—	0,2
Apatite	0,5	0,3	tr.
(Col. index)	(15,8)	(14,5)	(14,6)

Niggli values:

	al	fm	c	alk	si	k	mg	qz
(a)	42	17,5	11,5	29	302	0,39	0,45	+ 86
(b)	45	11,5	13	30,5	284,5	0,30	0,52	+ 62,5
(c)	44,5	14	7.5	34	379	0,51	0,25	+ 143

Localities and numbers of specimens:

(a) Lihammeren, Salangen (sp. 701/61). Anal. R Solli.
(b) Melkevann, Salangen (sp. 461/59). Anal P.-R. Graff.
(c) Tennevoll, Lavangen (sp. 259/59). Anal. K. Haugen.

This analysis (c) is previously published by P. R. Lund (1965). All calculations by the present writer.

Table XV

Partial chemical analyses of three granodiorites.

	(a)	(b)	(c)
Na_2O	4,80	4,36	3,27
K_2O	3,73	4,34	4,21
SiO_2	72,51	72,86	71,26
Niggli k values	(0,34)	(0,40)	(0,46)

Localities and numbers of specimens:

(a) L. Grøsnes, Gratangen (sp. 611/61).
(b) Kvitflåget, Salangen (sp. 512/60).
(c) Storfossen, Gratangen (sp. 102/63).

Analyst: P.-R. Graff.

IV. Genetic relations of the igneous rocks

Introduction

In the preceding chapters a description of petrography, chemistry and metamorphic alterations have been given. Some conclusions which can be drawn from these data are discussed and summarized in the following. The mode of emplacement is, for some rock types, quite clear, for others a more thorough discussion is needed. Further problems to be treated are the interrelations and possible genetic connections between rocks in each group. A brief comparison between the two igneous groups will be given. At the end of the chapter comparisons with alpine ophiolites and with igneous rocks from other parts of the Scandinavian Caledonides are presented.

Mode of emplacement of the different rock types in the older igneous group

The ultrabasic rocks

The present ultrabasics are of the alpine type as defined by Hess. Considerable variations in the stratigraphical positions of the different occurrences seem to exist.

The metamorphic alteration can be subdivided into 1. Serpentinization, and 2. Steatitization. The serpentinization predates the process of steatitization. As it is supposed that talc forms at higher temperatures than serpentine minerals it can be inferred that metamorphism took place by rising temperatures and therefore most probably, during the period of progressive, regional metamorphism. This leads to the conclusion that the ultrabasics were emplaced before, or at an early stage in, the metamorphic history of the area. The alterations were, however, incomplete, leading to disequilibrium assemblages being in discordance with the mineralogical phase rule. This feature can be ascribed to insufficient water supply after the outer parts of the ultrabasic bosses had been altered. It is implicit in this explanation that the serpentinizing waters originated from sources outside the peridotites.

The amphibolite dykes

Dyke intrusions of basic rocks are observed in a small area in Salangen. The adjoining meta-sediments have been assigned to the Narvik Group. Subsequent to intrusion the dykes were folded together with the sediments in the earliest fold phases of the area. Metamorphism in the lower part of the almandine-amphibolite facies affected the dykes (and adjoining sediments) and signs of primary textures or minerals have not been observed.

The deformation and mineral facies show that intrusion took place previous to the main phase of regional metamorphism.

The amphibolite massifs

Already from a consideration of the size of the amphibolite massifs an igneous origin seems most probable. Remnants of igneous textures show that this interpretation is correct. Furthermore, an *intrusive* nature of these amphibolites is proved by large inclusions of meta-sediments, transsected by amphibolite veins. From the presence of porphyric textures in the upper part of the Iselvdal amphibolite it might be inferred that intrusion, at least for this body, occurred at a relatively shallow level.

The metamorphic grade, as shown by the mineral assemblages, varies from the epidote-biotite subfacies of the epidote-amphibolite facies in the Rokkomborre massif to the almandine-amphibolite facies in the Iselvdal massif. As this seems to be in accordance with the facies distribution within the meta-sediments, the time if intrusion must be previous to the main phase of regional metamorphism.

The present variations in texture seem, at least partly, to be due to varying influence of directive pressures during metamorphism.

The amphibolite layers

Relict porphyric textures in some amphibolite layers and preserved pillow structures in others point to an igneous origin for the amphibolites in question. For the rest of amphibolite layers primary features are absent and their origin, consequently, less obvious. It is the opinion of the writer, however, that the absence of significant variations across the layers (except in some thin, schistose layers) speaks in favour of the rocks being ortho- rather than para-amphibolites. The chemical relations are not in opposition to this

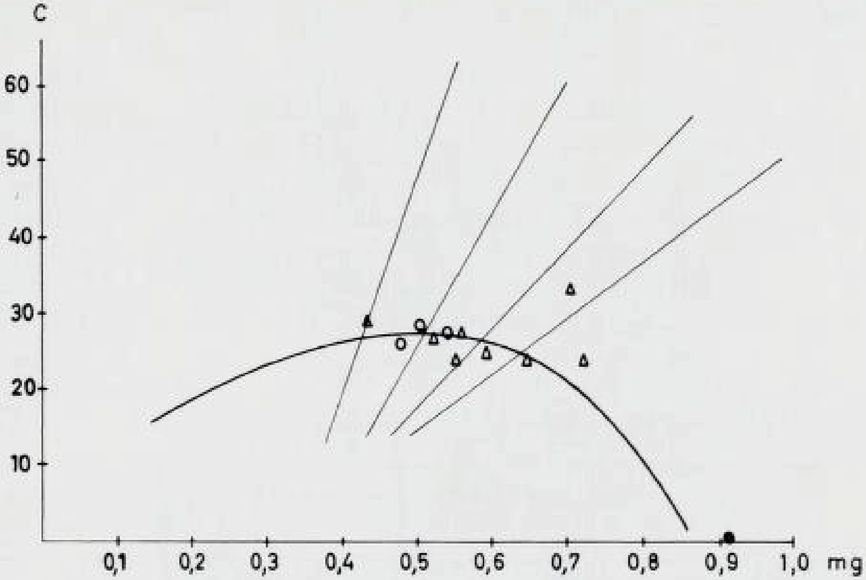

Fig. 19. Amphibolites of the map area plotted onto a c/mg diagram. The straight lines, indicating variation trends of para-amphibolites, and the curved line, representing the Karroo dolerites, are redrawn from Leake (1964). Triangles: amphibolite layers (Table VIII). Circles: amphibolite massifs and dyke (Tables IV and V). Dot: Peridotite (Table II).

view: Table VIII shows that the layers, including the relict porphyric ones, are of mutually similar compositions (an exception being the thin, schistose layers). At the same time they show close resemblance to basic rocks of igneous parentage from other areas. This is demonstrated by Table XVI.

Leake (1964) found the trends of chemical variations to be of significance in discriminating between ortho- and para-amphibolites: When, for instance, the Niggli c values are plotted against mg, ortho-amphibolites of an area will be distributed along a curved line like that of Fig. 19. (The curved line of this figure represents the Karroo dolerites). Para-amphibolites, on the other hand, tend to vary along straight lines (indicated on Fig. 19). Plotting of the analysed amphibolites from the present area in this diagram shows that the variation trend is consistent with an igneous parentage. (One layer deviates considerably from the main trend; this rock is a hornblende schist from the area south of Ofotfjorden; se column (g) of Table VIII).

Shaw and Kudo (1965) introduced the following function for 9 major oxides in order to discriminate between ortho- and para-amphibolites:

$$X = 7.07 \log TiO_2 + 1.91 \log Al_2O_3 \div 3.29 \log Fe_2O_3 + 8.48 \log FeO + 2.97 \log MnO + 4.81 \log MgO + 7.80 \log CaO + 3.92 \log P_2O_5 + 0.15 \log CO_2 \div 15.08.$$

In case X is positive, the amphibolite has an igneous origin, while para-amphibolites give negative values for X. The probability of incorrect classification was calculated to 5.7 %.

X were calculated for the 11 amphibolites given in Tables IV, V, and VIII of this paper: In the same order they are listed in these tables, the results were as follows: X = 6.97, 0.61, 1.92, 1.78, 4.27, ÷ 2.10, 0.94, 2.70, 4.43, ÷ 7.86, 1.09. It is seen that all, except two, have positive values for X and accordingly should be considered as ortho-amphibolites. The two amphibolites with negative X-values are those listed in column (c) and (g) of Table VIII. The former is one of the layers classified as "thin amphibolite layers with admixed sedimentary material", while the latter is the one deviating also from the igneous trend in the Leake diagram, Fig. 19. It is probable that these two amphibolites are of meta-sedimentary origin. Concerning the "thin layers" an origin by mixture of basalt tuff and terrigeneous sediment material may be the best explanation, as discussed below. In general, however, the calculation of the discriminant function support the view that most amphibolite layers have an igneous parentage.

The next problem to be discussed is whether the amphibolite layers are intrusives or metamorphosed supracrustals:

The metamorphic assemblages and relations to folding are consistent with both possibilities. In case they are intrusions, the emplacement must predate the main metamorphic phase and the strongest folding.

The position and spread in the stratigraphical column would normally be an important concern in deciding this problem: Supracrustals tend to occur of definite levels or within restricted parts of the sequence, while considerable spread in the stratigraphical position is to be expected when intrusives are considered. Although the bulk of amphibolite layers are situated within the Narvik and Niingen Groups, too many uncertainties are, however, involved in the interpretation of the internal stratigraphy of these groups for any safe conclusions to be drawn from their position.

The relict pillow structures in the Kvæfjord area show that at least some of the amphibolite layers were lavas.

A further consideration of the rock chemistry gives interesting results: In Table XVI the chemistry of amphibolites from the present area is compared with the following basic igneous rocks: Støren greenstone (col. 3), massive and schistose amphibolites from Holleindalen, Jotunheimen (4a, 4b), average gabbro and basalt (5a, 5b). The average composition and variation ranges of the proven intrusive amphibolites (amphibolite massifs and dykes) of the present area are given in column 1a, 1b, while the data for 8 amphibolite layers are listed in columns 2a and 2b. A close resemblance in most respects between the layers and the proven intrusives is evident from this table. The most important differences are found when the proportions of ferric to ferrous iron are considered, that is, in the oxidation ratios of the rocks.

Elliott and Cowan (1966) used the oxidation ratios in discriminating between lavas and intrusives among the Holleindalen greenstones. It was shown that in areas where basalt and gabbros are present, the latter have markedly lower oxidation ratios than the former. The use of this criterium for metamorphic basic rocks is based upon the assumption that the oxidation ratios are not significantly changed during metamorphism. Eugster (1959) and Chinner (1960) have given arguments for believing this to be the case.

A distinctly higher oxidation ratio in the amphibolite layers than in the intrusives is obvious from a look at Table XVI. While the intrusives have ratios in the range 1 - 10 with an average of 6,17, the layers have most ratios between 20 and 30 with average 24,05. (One value, 8,88, is considerably lower than the rest.) It is of special interest in this connection to note that the oxidation ratios of amphibolites with pillow structures (not listed in the table) lie in the same range as other amphibolite layers. Three specimens have the following values: 19,64, 23,12 and 36,09. (Data on FeO and Fe_2O_3 for these rocks given by B. Flood, personal communication.) Following the argumentation of Elliott and Cowan (1966), therefore, it seems reasonable to assume that most amphibolite layers were once lavas.

When the thin, schistose layers are concerned an origin from basic tuffs may be more probable than a formation from lavas. This because of the small thicknesses, great extensions and interlayering with meta-sedimentary material. The single analysed specimen of such a layer is the one with the highest oxidation ratio, 45,85 (Table XVI), which is in accordance with the generally higher oxidation state of tuffs as compared with lavas and intrusives (Elliott and Cowan, 1966, p. 322).

It is evident from the preceding descriptions that the groups of amphibolites with high, respectively low, oxidation ratios is not, in the present area, following the distinction between schistose and massive-textured rocks. The latter grouping will clearly depend, not only on primary grain

Table XVI

Chemistry of amphibolites from the present area (1a, b, 2a, b) compared with greenstones from Støren (3), amphibolites from Holleindalen, Jotunheimen (4a, b) and the average gabbro and basalt (5a, b).

	1a	1b	2a	2b	3	4a	4b	5a	5b
SiO_2	48,53	47.80—48,92	49.66	47,09—52,57	47,78	47,21	48,50	48,24	49.06
TiO_2	1.42	0,20— 2,34	1,56	0,26— 2,73	1,40	2,41	2,05	0,97	1,36
Al_2O_3	15,48	14,39—16,60	15,66	12,70—20,65	14,95	14,40	15,00	17,88	15,70
Fe_2O_3	0,75	0,02— 1,22	2,36	0,82— 3,88	3,45	2,17	3,39	3,16	5,38
FeO	10,64	10,04—11,31	7,27	2,59—10,67	5,26	9,65	8,44	5,95	6,37
MnO	0,25	0,22— 0,27	0,17	0,12— 0,23	0,13	0,12	0,11	0,13	0,31
MgO	6,68	6,42— 6,94	7,53	5,19—11,06	5,38	7,82	7,12	7,51	6,17
CaO	10,84	10,13—11,44	10,57	9,28—13,13	11,87	10,20	9,44	10,99	8,95
Na_2O	2,79	2,49— 3,31	2,77	2,36— 3,49	2,95	2,77	2,65	2,55	3,11
K_2O	0,66	0,50— 0,74	0,46	0,13— 1,04	0,21	0,56	0,51	0,89	1,52
CO_2	0,15	0,06— 0,29	0,38	0,00— 1,55	3,30	0,38	X[1]	—	—
P_2O_5	0,21	0,13— 0,30	0,15	0,00— 0,26	0,11	0,22	0,29	0,28	0,45
H_2O+	1,29	1,02— 1,63	1,30	0,61— 1,64	3,06	2,03	2,52[1]	1,45	1,62
H_2O-	0,05	0,03— 0,09	0,05	0,01— 0,11	0,06				
Sum	99,74	—	99,89	—	99,91	99,94	100,02	100,00	100,00
Ox. ratios[2] $\frac{2Fe_2O_3 \times 100}{2Fe_2O_3 + FeO}$	6,17[3]	1,43— 9,38	24,05[3]	8,88—45,85	37,46	16,78	26,58	32,45	42,96

1a. Mean composition of intrusive amphibolites from the present area (Tables IV, V).
1b. Variation ranges of amphibolites in **1a.**
2a. Mean composition of 8 amphibolite layers from the map area (Table VIII).
2b. Variation ranges of amphibolite layers.
3. Greenstone, Haga bridge, Støren. Goldschmidt (1916, p. 15).
4a. Mean composition of 13 massive amphibolites from Holleindalen, Jotunheimen (Elliott and Cowan, 1966, p. 316).
4b. Mean composition of 14 schistose amphibolites, same area as 4a.
5a. Average gabbro after Daly.
5b. Average basalt after Daly.

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(1) $H_2O + CO_2 = 2{,}52$.
(2) Oxidation ratios calculated according to Chinner (1960).
(3) Mean of oxidation ratios (not oxidation ratio of average rock).

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size, but also on the varying influence of directive pressures during metamorphism. The present investigations, therefore, tend to show that the reservations taken by Elliott and Cowan (1966, p. 323) about the significance of textural relations were correct.

In conclusion it can be stated that field relations and general chemistry are in accordance with an intrusive as well as a supracrustal origin. The oxidation ratios and presence of pillow structures within a local area taken into account, however, *an origin by submarine lava outpourings seems most probable.* Some thin and extensive layers may be best interpreted as basic tuffs. Mineral assemblages were determined by regional metamorphism, varying from epidote-amphibolite facies to low almandine-amphibolite facies. Variations in the present textures may partly be due to inherited primary variations, partly to differences in the intensity of directive pressures during metamorphism.

The metamorphosed quartz keratophyres

The reasons for believing that the described layers are metamorphosed igneous rocks and not meta-sediments of terrigeneous origin are summarized and briefly discussed in the following:

Tables IX and X of this paper (p. 46 and 47) generally indicate a constancy in composition which seems to be in best accordance with an igneous origin. Quartzo-feldspathic sediments of terrigeneous origin would normally tend to show stronger variations in the relative proportions of the main minerals and, at least locally, higher SiO_2 contents than found in the present rocks. The chemistry also shows strong similarity to that of quartz keratophyres from other parts of the Scandinavian Caledonides (Table XVII). Most analyses cited in the literature show a marked soda domin-

Table XVII

Chemical composition of some quartz keratophyres from different parts of the Scandinavian Caledonides.

Wt %	(a)	(b)	(c)	(d)	(e)	(f)	(g)
SiO_2	70,51	68,65	77,80	74,97	76.80	72,96	66,58
TiO_2	0,38	0,06	n.d.	0,43	0,25	0,18	0,25
Al_2O_3	15,69	16,85	»	12,09	11,87	15,90	15,79
Fe_2O_3	1,18	0,47	»	2,31	0,48	0,86	0,89
FeO	1,35	2,63	»	1,27	1,66	0,37	1,03
MnO	0,04	0,10	»	0,04	0,04	0,01	0,05
MgO	0,94	1,24	»	0,33	0,21	0,56	0,71
CaO	3,06	2,99	»	0,81	0,20	0,46	0,73
BaO	n.d.	n.d.	»	0,02	0,02	0,06	0,57
Na_2O	6,11	5,42	6,08	6,48	3,75	5,31	1,73
K_2O	0,21	1,26	0,21	0,54	4,03	2,53	9,88
CO_2	n.d.	0,05	n.d.	0,26	0,01	0,13	0,56
P_2O_5	0,12	0,17	»	0,14	0,03	0,03	0,06
S	n.d.	n.d.	»	n.d.	n.d.	0,02	0,02
H_2O+	0,33	0,58	»	0,22[1]	0,49[1]	0,67	1,02
H_2O-	0,05	0,02	»			0,05	0,04
Sum	99,92	100,49	—	99,91	99,94	100,10	99,91

(a) Quartz keratophyre, Ravddovaras, Bardu.
(b) » » Skårvikelven, Salangen.
(c) Schistose quartz keratophyre, Varaldsøy, W. Norway. (Foslie, 1955, p. 19).
(d) Quartz keratophyre, N. of Mortojåkko, S. Storfjället, Sweden. (Beskow, 1929, p. 91).
(e) Microcline-bearing quartz keratophyre, N. of Mortojåkko, S. Storfjället, Sweden. (Beskow, 1929, p. 93).
(f) Quartz keratophyre, Sagelva, Hølonda, Trondheim region (Vogt, 1945, p. 469).
(g) Quartz keratophyre, S. Leiro, Huglo, W. Norway. (Kolderup, 1929, p. 11).

[1] Total H_2O.

ance, as do the present rocks. The analyses, cited in Table XVII, were, however, partly selected for a demonstration of the variations in the Na/K ratio.

Field associations with amphibolites of igneous (probably volcanic) origin is characteristic of the rocks under discussion. This feature they share with quartz keratophyres of different areas, where these rocks are associated with basalts or the metamorphic equivalents of these, greenstones and amphibolites. The banding observed in a number of outcrops of the present rocks can be taken as evidence neither for an igneous nor for a sedimentary origin, as such structures may occur in both categories of rocks.

Euhedral or subhedral plagioclases with Carlsbad twins detected in a couple of occurrences are evidence of an igneous parentage, the more so as these feldspars occur in a finer grained matrix thus giving the larger feldspars a phenocryst-like habit and a relict porphyric appearance of the rocks in question. Such textures might appear also in metamorphosed arkosic sediments, but as it has been shown above, the small variations in chemical composition are speaking against this possibility. All arguments taken into account, it seems reasonable to assume that the present rocks are corresponding to the quartz keratophyres described from several areas within the Caledonides, but now present in a more or less metamorphic state. In discussing more precisely their mode of formation a short review of literature on the subject is apposite in this place.

Reusch (1888) described quartz porphyries from the island of Bømlo and som adjacent areas in West Norway. These porphyries clearly correspond to rocks later termed quartz keratophyre by N. H. Kolderup (1929) and Foslie (1955). Reusch (p. 372) interpreted them as lavas, pyroclastics ("quartz porphyry tuff") and shallow dyke intrusions. An effusive (volcanic) mode of formation for a potash-dominant quartz keratophyre (Table XVII, (g)) from the same area was suggested by Kolderup (1929).

Foslie (posthumous paper, 1955) gave a description from parts of the same region (Varaldsøy-Ølve area). In this area soda-dominant and, subordinate, potash-dominant quartz keratophyres were shown to occur in close association with greenstones. A discussion of genetic relations is not included in Foslie's paper, but a volcanic origin is clearly assumed.

Th. Vogt (1945) mentions quartz keratophyres from the Hølonda-Horg area, S. Trøndelag. According to the brief description they are minor intrusions in greenstone, but details of field relations are not presented.

H. Carstens (1955) described quartz keratophyre and keratophyre agglomerate in association with basic volcanics in the Snåsa syncline, N. Trøndelag. The quartz keratophyres were interpreted as lava flows.

Intrusions of quartz keratophyre in greenstone within the Mostadmarka - Selbustrand area, S. Trøndelag, have recently been treated by Torske (1965). Based on mineralogical details, several types were distinguished. Soda dominance seems to be a general feature in this area, although chemical analyses were not presented. In addition to intrusions, part of the material in pyroclastic beds was thought to have a quartz keratophyre parentage.

In a recent paper from the Meråker area, N. Trøndelag, Chaloupsky and Fediuk (1967, pp. 10-11) describe alternating beds of quartz keratophyre and greenstone. A pyroclastic origin is supposed for both rock types: "As the layers of both basic and acid volcanics are traceable over a distance of several hundred metres, they can hardly be interpreted as lava sheets, particularly the acid types (due to the high viscosity of acid magmas), and a pyroclastic origin for at least part of the sequence should be postulated".

Within the Scandinavian Caledonides the most thorough descriptions of quartz keratophyric rocks are those given by Beskow (1929) from the Södra Storfjället area, Västerbotten, Sweden. Keratophyre and quartz keratophyre lavas as well as agglomerate and tuff layers of corresponding composition are present in this area. Most types are clearly soda dominant, but microcline-bearing quartz keratophyres are occasionally met with (see (d) and (e) of Table XVII). Genetic connections between the keratophyric supracrustals and trondhjemitic intrusives are emphasized by Beskow.

From Northern Norway, quartz keratophyres have not, up to now, been described. However, a rock from the Bjørkåsen mine, south of Ofotfjorden, described by Foslie on a map from 1930 as "fine-grained protoclastic trondhjemite", is referred to by Oftedahl (1958) as quartz keratophyre. The writer has studied specimens and a couple of thin sections of this rock (collections of Foslie and J. Færden) and it undoubtedly shows some resemblance to metamorphosed quartz keratophyre of the Bardu area.

Th. Vogt (1927) described "albite porphyries" from the Sulitjelma area which possess some characteristica typical of keratophyres. A chemical analysis shows a pronounced soda dominance (10, 34 wt % Na_2O, 0.48 % K_2O). From the same area was reported a keratophyre conglomerate. An agglomeratic origin of the latter rock has recently been suggested by Nicholson (1966, pp. 145 - 146).

Because documentations are partly insufficient, it seems difficult to evaluate all the different conclusions reached at in the cited literature. In general, however, it can be concluded that lavas and pyroclastics, as well as intrusives, are represented amoung the quartz keratophyres of the Scandinavian Caledonides. Alle these types of origin should therefore be considered possible for the quartz keratophyres of the present area.

A summary of the most important observations on the present quartz keratophyres and the conclusions which can be drawn from them, are given below:

Observations	Conclusions
The mineral assemblage is in accordance with the distribution of regional metamorphic facies within the meta-sediments.	The quartz keratophyres were emplaced before the main regional metamorphism.
The boundaries seem to be concordant against meta-sediments and amphibolites.	The boundary relations are in accordance with a supracrustal origin or formation by sill intrusions.
The extension of the quartz keratophyres is considerable but thicknesses vary within wide limits.	The extension and thicknesses are consistent with an intrusive as well as a supracrustal origin.
The quartz keratophyres in part show fine banding due to variations in the relative amounts of the minerals.	Some sort of sedimentation process may have played a role in the formation of these rocks.
Some quartz keratophyres are high in muscovite and calcite.	Mixing of sediments and quartz keratophyric material probably occurred in some cases.
Most of the quartz keratophyres are fine-grained. Supposed phenocrysts of sodic plagioclase are retained in a couple of localities.	The supposed phenocrysts show that primary grain sizes have been retained in some rocks although the textures now are largely metamorphic. The usual type of quartz keratophyre may therefore have been originally fine-grained.
The quartz keratophyres are associated in the field with metamorphosed marine sediments and amphibolites interpreted as originally basaltic lavas.	If of supracrustal origin the quartz keratophyres were deposited on a sea floor.
The quartz keratophyres have a high SiO_2 content (approximately 70 per cent).	A magma with the composition of quartz keratophyre would probably be highly viscous.
The dominance of soda over potash is pronounced.	

The sum of observations seem to be best explained by assuming a supracrustal origin. Absence of observations favouring an intrusive nature of the rocks is negative evidence of the same.

It is near at hand to interpret the banded type as a pyroclastic rock, probably deposited as waterlaid tuffs. This explains the layering as well as the admixture of calcitic and micaceous material in some layers.

Explosive volcanic activity as a general explanation for the often thin and extensive keratophyric or granitic layers is clearly expressed by Oftedahl (1958, 1959). He also thinks the action of sea water to be responsible for the soda dominance typical of most of these layers.

The viscosity of highly siliceous magmas seems to be in disfavour of an origin by lava outpourings for extensive layers (see for instance Rittmann (1962)). As noted by Carstens (1955), however, a high water content of the magma may possibly have lowered the viscosity sufficiently for the formation of extensive quartz keratophyre layers by lava flows and so may other dissolved volatiles (Rittmann, 1962, p. 169). This possibility can not be quite discarded. The large thicknesses of the quartz keratophyres in part of the area may also indicate that some layers are lavas rather than pyroclastics. All thick beds are not banded, either.

As a conclusion it can be stated that *the quartz keratophyres of the present area probably are volcanic rocks; some are almost certainly metamorphosed pyroclastics while others may be lavas.* Concerning the special chemical nature of these rocks it is more properly discussed together with the problems of consanguinity between the igneous rocks. At this place it shall only be stated that the soda dominance is not in any way in contradiction with the origin outlined above.

The meta-trondhjemites

In the description given it has been shown that the meta-trondhjemites occur as intrusive bodies of relatively small sizes. The time of emplacement is previous to the strong folding episodes and the main regional metamorphism, and at least some of them are also previous to the deposition of the Elvenes conglomerate. In other words, the *intrusion occurred in the early orogenic stage and, at least partly, in the geosynclinal stage.*

As a result of progressive metamorphism the rocks were mineralogically and texturally reconstructed. Some changes (porphyroblastic feldspar growth) probably postdate the main progressive phase as plagioclase porphyroblasts in gneisses areally related to the intrusions have previously (Gustavson, 1966) been shown to belong to a later, static phase. It is

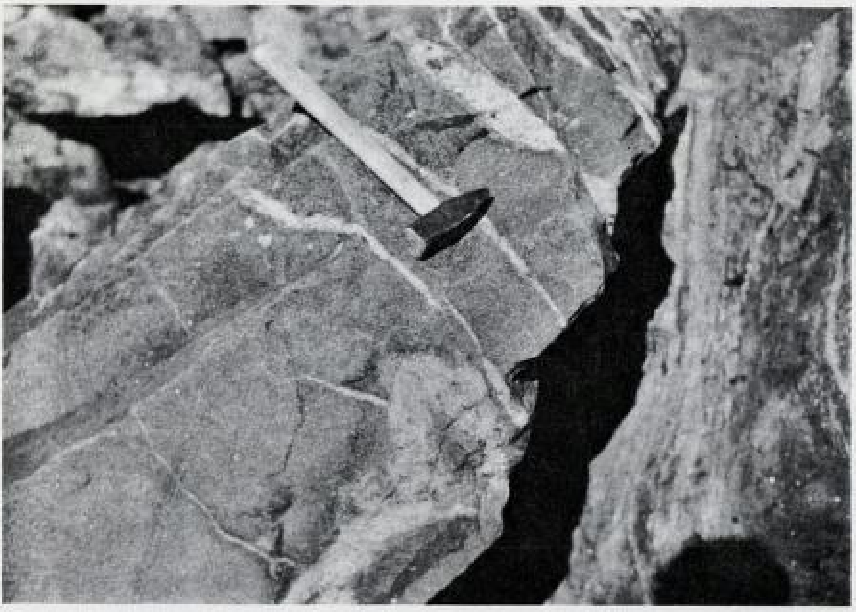

Fig. 20. Trondhjemite veins in amphibolite, Høgfjell, Salangen.

possible that chemical components mobilized as a consequence of almandine-amphibolite facies conditions, migrated within and out from the trondhjemites, causing porphyroblastesis in the meta-trondhjemites and adjoining gneisses subsequent to the main, syntectonic, reconstruction of the rocks. Even younger than the porphyroblastesis is crushing, possibly due to late folding and thrust movements.

Interrelations of rocks in the older igneous group

The relative ages

Interpretations of the age relations of the supposed volcanic rocks is clearly dependent on the structural and stratigraphical solutions for the areas in question. From the quartz keratophyre formation in Bardu has been described an upward increase in amphibolite layers. The interlayering of acid and basic rocks, however, and the occurrence of amphibolites below the formation as well, show that there can be no significant age differences, provided that a supracrustal origin of basic and acid rocks is accepted.

The intrusive amphibolite massifs and dykes seem to occur at a lower stratigraphical level than the supposed basic lavas. For the dykes as well as the Iselvdal massif the level of intrusion is within a quartzite belonging to the sequence below the quartz keratophyres (Gustavson, 1966, pp. 71-74). Relict porphyric textures in the upper part of the Iselvdal amphibolite indicate a shallow intrusion level. The various data are therefore probably best interpreted by assuming that the amphibolite massifs and dykes were formed approximately contemporaneous with the outpourings of basic lavas. At the moment nothing seems directly to contradict this interpretation.

A younger age of the trondhjemites relative to the amphibolites is shown by observations referred to (p. 48) from the Høgfjell area, Salangen: Veins of meta-trondhjemites were seen to intrude massive-textured amphibolite layers (Fig. 20). How important the age difference is, cannot, however, be safely stated. Both rock types are definitely older than the main regional metamorphism, as already noted in the petrographical descriptions.

Direct field observations bearing on the age relationships of serpentinites relative to other igneous rocks of the older group are lacking. Frequent association with amphibolites may, however, indicate close connection also concerning time relationships.

The chemical and mineralogical classification of the rocks

The major oxides of the main rock types have been plotted against SiO_2 in Figs. 21A and B. A distinct threefold grouping of the rocks is apparent from these diagrams. Group I consists of the single analysis of an ultrabasic rock. From microscopical investigations of the peridotites it seems obvious that all ultrabasics of the older group possess similar chemical characteristics as the analysed specimen. Group II, with SiO_2 between 47 and 52 weight per cent, is formed by the amphibolites while group III includes trondhjemites and quartz keratophyres with approximately 70 per cent SiO_2. The chemical similarity between the two latter rock types is striking. Within each of the three groups of Fig. 21 there is again a marked grouping of each oxide. The most pronounced variations is found in the contents of total iron.

A further inspection of the basic rocks is needed for a more exact classification purpose: According to Yoder and Tilley (1962, p. 352) the normative minerals can be used in the classification of basaltic rocks in the following chematic way:

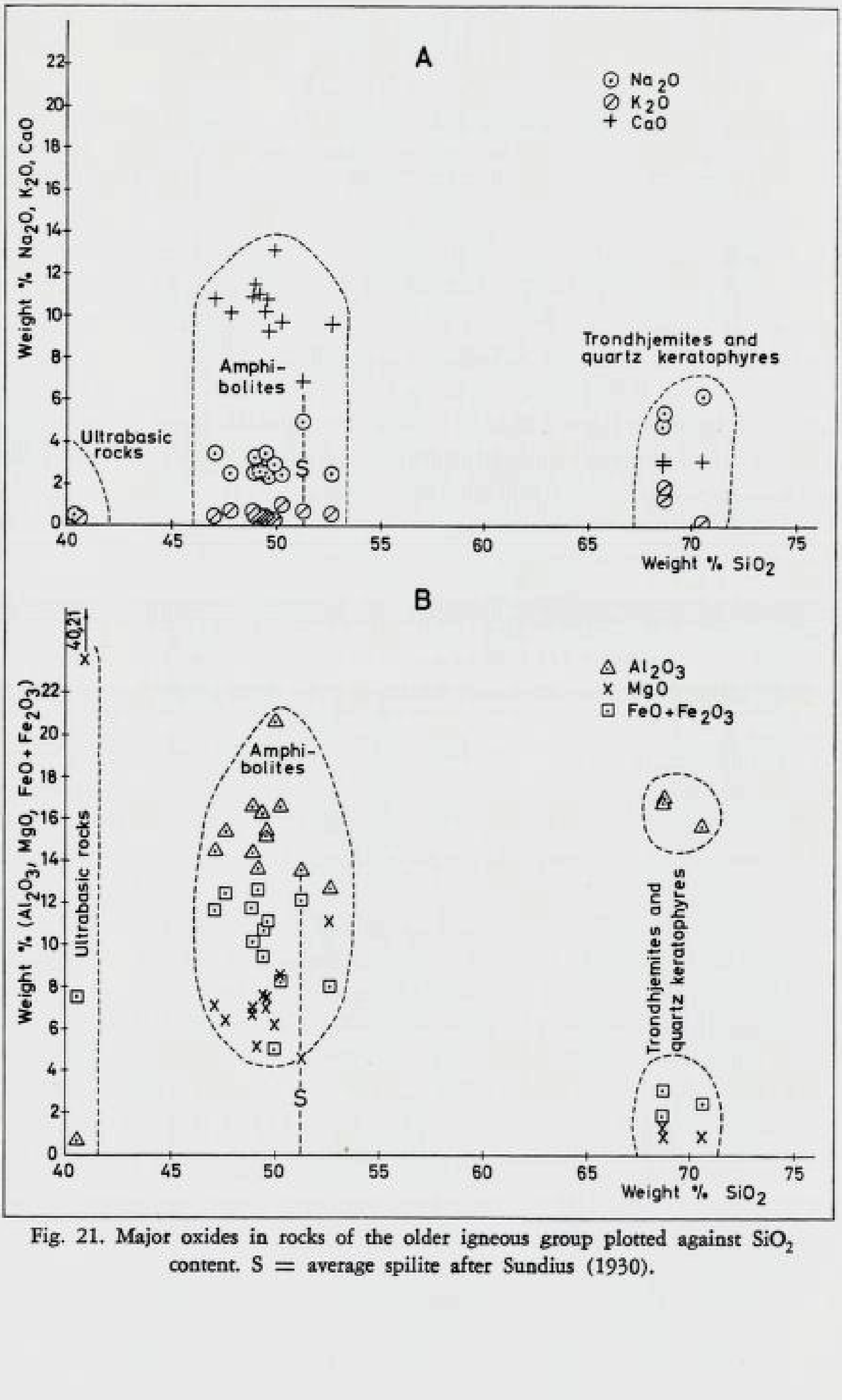

Fig. 21. Major oxides in rocks of the older igneous group plotted against SiO_2 content. S = average spilite after Sundius (1930).

1. Presence of normative quartz and hypersthene: Tholeiite (oversaturated).
2. Normative hypersthene: Tholeiite (saturated).
3. Normative hypersthene + olivine: Olivine tholeiite (undersaturated).
4. Normative olivine: Olivine basalt.
5. Normative olivine + nepheline: Alkali basalt.

Table XVIII demonstrates a classification of the present amphibolites according to these rules. It is seen that the rocks could not have crystallized directly from one single type of basalt magma, as *quartz tholeiitic, olivine tholeiitic as well as alkali basaltic compositions are represented.*

It should be noted in this place that the above conclusion as well as the value of the following discussion is depending on the correctness of the assumption that metamorphism of the present basic igneous rocks was generally isochemical. There is nothing in composition, microscopic features or field observations that indicate important metasomatic changes, and in the following this possibility is left out of consideration.

In Table XVIII is also given the calculated values of

$$\sigma = \frac{(Na_2O + K_2O)^2}{SiO_2 - 43},$$

the socalled suite index of Rittmann (1962, p. 110). According to Rittmann's classification scheme, all the present basic rocks belong to the calc-alkaline suite, with variations from strong to weakly calc-alkaline types. (The one with the lowest σ value, 0,99, corresponding to extreme calc-alkaline composition, should be left out of consideration as this is the one thought to be contaminated with sedimentary material, as discussed on page 77).

The non-spilitic nature of the basic rocks should be noted. For comparison the average spilite composition (Sundius, 1930) is plotted onto the diagram, Fig. 21. The generally higher Al, Mg and Ca and lower Na contents of the present rocks as compared with the spilitic composition[1]) are evident.

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[1]) In a recent paper Amstutz (1968) doubts the existence of a spesific spilitic chemistry as far as the rock as a whole is considered. The chemical features termed "spilitic" could be the result of variations in composition within the rock bodies and sampling errors due to these variations.

Table XVIII

Magma types of basic rocks in the older igneous group

Mode of occurrence	Locality (see map)	Norm. minerals[1])	Magma type	σ[2])
1. Amph. dyke	Rotvik	Hy + Ol	Olivine tholeiite	2.17
2. Amph. massif	Iselvdal	Hy + Ol	» »	1.58
3. » »	Bangfjellet	Ol + Ne	Alkali basalt	2.78
4. Amph. layer	Høgfjell	Ol + Ne	» »	2.52
5. » »	Bangskletten	Hy + Q	Tholeiite	1.24
6. » »	Ravddovaras	Hy + Q	»	0.99
7. » »	Skinskaret	Hy + Ol	Olivine tholeiite	3.44
8. » »	Brattåsen	Hy + Ol	» »	1.63
9. » »	Bjørkåsen	Hy + Q	Tholeiite	1.49
10. » »	Geitvann	Hy + Ol	Olivine tholeiite	1.33
11. » »	»	Hy + Q	Tholeiite	1.17

[1]) For further chemical and mineralogical data on these rocks, see Tables IV, V and VIII.
[2]) See the text.

Discussion of possible genetic connections between rocks in the older igneous group

In the Japanese islands Sugimura (1968) have described a distinct zonal arrangement of the basic volcanics; the tholeiites and associated calc-alkali series occurring in one belt, alkali basaltic and associated calc-alkali series in an adjoining belt. It is also shown to similar relations in other parts of the Pacific area. In the present case such an areal distribution have not been detected. On the contrary, basalts of tholeiitic and alkali basaltic composition both occur within the same area. This is of considerable interest although it is not yet known if this feature is one of regional significance, and eventually, to what extent.

O'Hara (1965) proposed a petrogenetic scheme for the relations between various types of basic magmas, based on crystal fractionation processes. This scheme seems of special interest to us as it readily explains a close connection in time and space between basalts of different composition. According to the proposals of O'Hara the nature of the basalt magmas and their differentiation products largely depends upon the depth of origin (melting level) of the magmas and on their rate of migration to the surface. Furthermore, it will greatly influence their composition whether they originate under wet or dry melting conditions. Wet melting in the intermediate part of the socalled low pressure regime, as thought to be characteristic of orogenic regions, may, for instance, give rise to andesitic parental magmas and to andesitic, dacitic and rhyolitic rocks. A hypersthene - normative picrite magma originates by dry melting in the high-pressure regime. By steady migration towards the surface such a magma may change to alkali basaltic, higher up to high alumina basaltic and, at shallow depths, to a quartz tholeiitic basalt composition. Tholeiitic basalt may pour out on the surface, or, if this magma is held for some time at shallow levels, give rise to tholeiitic silica saturated fractionation series. In the same way as a magma from the deepest levels changes its composition as it rises through the crust, melting in anyone pressure regime produces basalt of a composition characteristic of that level. By rapid ascent each of these basalt types may reach the surface or, if held for some time at shallow depths, differentiate into its own characteristic series. Applied to the Southern Troms and Ofoten areas the reasoning outlined above gives as a result that formation of magmas occurred at least as deep as in the deepest part of the low pressure regime because alkali olivine basaltic compositions are present among the basic rocks (Table XVIII). Most of the magmas generated, however, changed composition towards a tholeiitic chemistry during their rise to the surface. It is also possible that the latter magmas formed by melting in the upper part of the low pressure regime although this alternative seems less probable. Another conclusion which can be drawn from this reasoning is that melting in the present case (in part at least) occurred under higher confining pressures and lower H_2O partial pressures than those thought by O'Hara to be characteristic for orogenic regions in general. This is shown by the absence of andesitic rocks in the present area.

So far only the basaltic rock types have been considered. Although nothing definite can be concluded, *it seems reasonable to assume that ultrabasic and basic rocks are genetically connected, probably formed by fractional crystallization of basalt magmas.* This is implicit in the reasoning above which explains the occurrence of different basalt types as a result of crystal fractionation. Field evidence also, to some degree, points to a connection between amphibolites and serpentinites as both rock types occur within the same stratigraphical groups.

In a recent paper, Thayer (1967) presented arguments in favour of a genetical connection between ultramafic and gabbroic rocks of the alpine ophiolite suite and he is arguing against views of genesis based on Hess' (1938) theories of a primary peridotite magma. As will be shown later there are strong similarities between the alpine ophiolites and older group rocks from the present area.

Another matter of interest is the relations between basic rocks and the mostly soda-dominant acid members of the older group: A review of the current literature convinces about the common association of certain basic volcanics with acid intrusives and extrusives of soda-dominant type. This is the socalled "spilite-keratophyre association" of orogenic zones. According to Turner and Verhoogen (1960, p. 258) "Most eroded geosynclines show evidence of igneous activity approximately synchronous with at least the later part of the filling and sinking of the trough. Prominent among the products of such activity and almost confined to the geosynclinal environment, are submarine lavas, tuffs and equivalent intrusions of sodic composition. These constitute the spilite keratophyre association. This association may also include basalts of normal composition". Beskow (1929) likewise pointed to the fact that keratophyres may frequently be associated with basic volcanics of a "normal", nonspilitic character. As already noted (p. 89), Amstutz (1968) doubts the existence of a special spilitic chemistry. It is beyond the scope of the present paper to discuss the existence or non-existence of spilitic magmas, especially as the present basic rocks are definitely not spilitic in character. It can, however, be stated that in the present case, at least, such a magma evidently did not exist and the sodic character of the acid members must be ascribed either to secondary and essentially non-magmatic processes, or have been derived by some sort of differentiation from non-spilitic basaltic magmas.

The possibility of fractional crystallization of basaltic magma leading to sodic differentiates was suggested already by Goldschmidt (1922) in his description of the biotite diorite series of igneous rocks. Later Battey (1956) and Kuno (1968), for instance, have stressed the importance of water in the development of soda-dominant acid rocks. Battey, however, ascribes this process to the post-magmatic stage, as albitization of feldspars is thought to be due to a redistribution of alkalies during subsequent burial of rhyolites. Kuno (1968, p. 632) states that: "The calc-alkali rock series appears to start from each of the three parental magmas. The condition necessary for the formation of this series is enrichment of water in the magma which would result in a rise of the oxygen partial pressure of the magma. This enrichment may occur in some way in connection with assimilation of granitic or siliceous sedimentary rocks by the basalt magmas, or it may occur independently."

Foslie (1921) considered amphibolites, serpentinites, trondhjemites and granites from part of the present area as comagmatic igneous rocks formed by differentiation due to differential squeezing during crystallization. Objections can, however, be raised against a comagmatic origin, not least concerning the acid rocks:

Oftedahl (1959) states that the extreme soda dominance and potash deficiency of many keratophyres can hardly be of magmatic origin and suggests that metasomatic action of sea water on rhyolitic ashes is the prime cause. Such an explanation seems applicable to quartz keratophyres of the present area as these are thought to be water-laid ashes or submarine lavas. It is also possible to ascribe the soda dominance to the passage of granitic magma through wet geosynclinal sediments. Such sediments must be assumed to contain considerable amounts of sea water and soda as emphasized by several authors (see Gilluly (1935) for a discussion and for references). The compositional considerations concerning the quartz keratophyres can also to a certain extent be applied to the trondhjemitic rocks, although most of these are less extreme in their potash deficiency.

Even if soda/potash relations are due to secondary processes and the acid extrusives and intrusives were originally granitic or granodioritic in composition, there is the possibility that these primarily were derived from basaltic magmas and are essentially comagmatic with the basic and ultrabasic rocks. However, the large compositional gap from about 50 per cent to nearly 70 per cent SiO_2 (Fig. 21) is in disfavour of this interpretation. It could be suggested, of course, that intermediate rocks by chance are not exposed in the area, but this solution seems unsatisfactory because absence of intermediate rocks can be seen from descriptions to be a common feature in the Scandinavian Caledonides, at least when rocks altered by regional metamorphism (older group of igneous rocks) are considered. With this in mind it seems more reasonable to assume an anatectic origin for the acid members. As the chemistry of these rocks differ considerably from the minimum melting composition (Fig. 23) some secondary process is probably responsible for the conversion of granodioritic and granitic melts into soda-dominant types. As rocks in the Precambrian basement are largely of granitic composition, it seems reasonable to believe that the anatectic melts originated from this substratum. It is tempting to assume that anatexis was promoted by the ascent of, and heating by, basalt magmas. This would explain the close association in space and time of basic and acid igneous rocks of this group. During their slow passage through the geosynclinal sediments the granitic melts then came in contact with soda-rich solutions which may be responsible for their conversion into trondhjemitic and quartz keratophyric compositions. The soda-rich solutions could have been derived mainly from sea water entrapped in the sediments, but partly also from graywacke-like sediments. It has been shown in a previous paper (Gustavson, 1966, p. 117) that metasediments relatively high in Na and Ca are present in the lower part of the Narvik Group. More extreme composition of the quartz keratophyres as compared with the trondhjemites may be ascribed to the possibility that they moved a longer way through the sediments or perhaps to the additional effect of the direct contact with sea water after extrusion.

Barth (1962, p. 222) suggested a formation of trondhjemitic magmas or liquids by melting of graywackes and similar sediments during metamorphism and deep burial. This is an alternative to the genesis outlined above, similar, however, in that it assumes anatexis instead of fractional crystallization as responsible for these rocks. In the present case, however, melting of sediment is a less probable origin because boulders of trondhjemite in some conglomerates (p. 50) show the emplacement of most trondhjemitic rocks to belong to the geosynclinal stage. It is not probable that the sediments at this stage were sufficiently deeply buried for any anatectic melting to occur within them.

Interrelations of rocks in the younger igneous group

Field and age relations

The age sequence established by Foslie (1921, 1922) in the Råna massif is the one usually found in igneous rock complexes: Ultrabasic rocks crystallized before the basic and intermediate types which are again older than the acid rocks. These relations are probably valid also within other parts of the map area. It has been described, for instance (p. 61), the occurrence of peridotite inclusions in the Istind diorite, the ultrabasics evidently being the older rock, and from Salangen have been reported observations of granitic veins intrusive into diorite (p. 62 and Fig. 15).

Chemical relations

The major chemical constituents have been plotted against SiO_2 in Fig. 22. A threefold grouping, similar to that found for the older group, is clearly recognized. It is seen, however, that the gaps between ultrabasic, basic and acid rocks, respectively, are smaller in the present case. It is also known that rocks transitional between peridotites and gabbros do occur (Råna and Orta, see pp. 58-59) and these rocks would probably plot in the SiO_2 interval 42 - 50 if they were analysed. The gap from about 57 per cent to 66 per cent SiO_2 seems, however, to be a real one as rocks transitional between diorites and granodiorites (or trondhjemites) are unknown in the area.

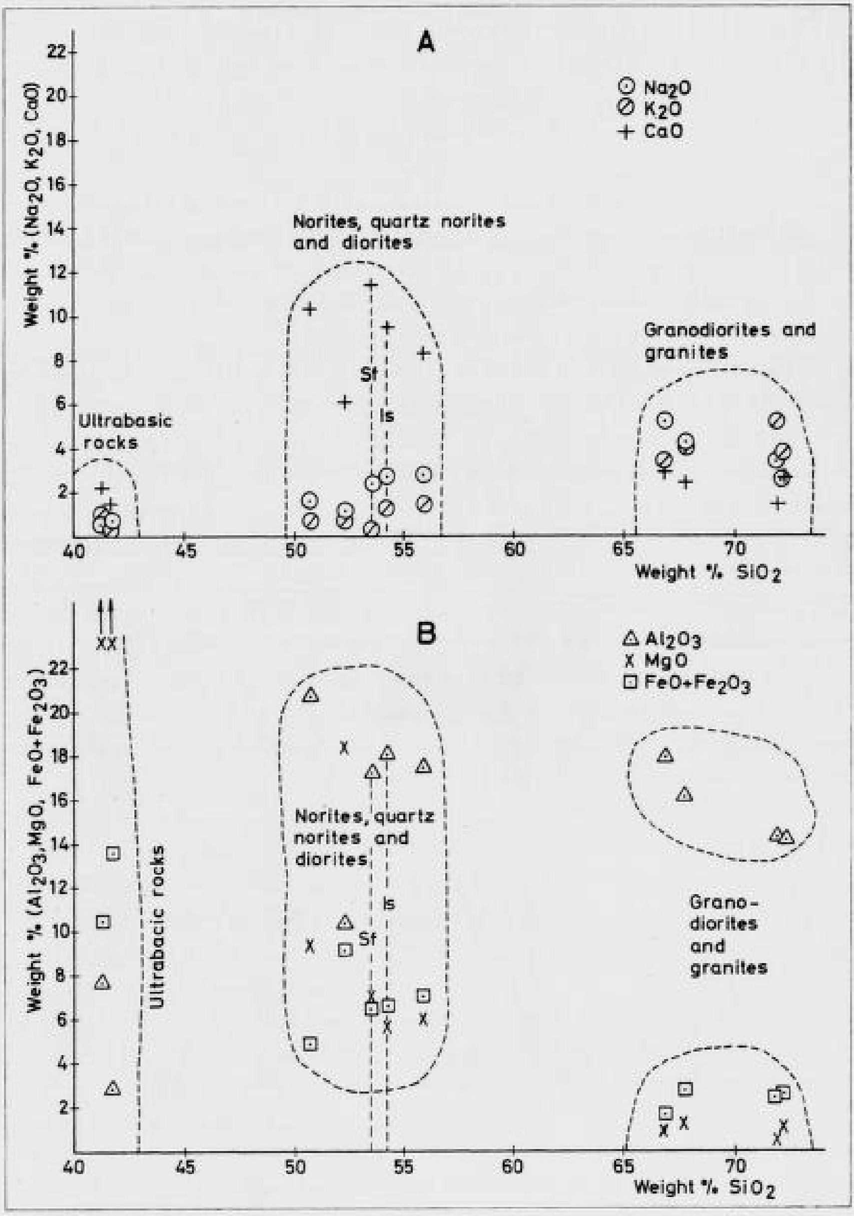

Fig. 22. Major oxides in rocks of the younger igneous group plotted against SiO_2 content. Sf. and Is. = Sørfjorden and Istind diorites.

Foslie (1921) calculated the composition of a possible parental magma for the Råna massif based on the available analyses and the field frequency of the various rock types. The normative composition (Table XIX, col. a) shows this magma to have been tholeiitic, as further discussed below.

Genesis of the younger igneous group

Interrelations between rocks belonging to the younger group have, in part, already been pointed out in the petrographical description. Similarities in petrography and chemistry are evidences for a similar origin for the various occurrences of the younger igneous rocks.

A natural base for a discussion of these rocks is the investigations by Foslie (1921, 1922) of the Råna massif. The opinions about origin advocated by this author have already been briefly referred to (p. 58):

The various rock types of Råna formed as a result of fractional crystallization of a basalt magma. The fractionation process were supposed to have been promoted by lateral pressures operating during the crystallization period, the massif thus being considered syntectonic or late-tectonic. The differentiation took place in situ, without supply of new magma during the crystallization and also without any important influence of gravitative settling. At the time of formation of the granite dykes fissures had opened in the solidified basic rocks due to consolidation and contraction of the main norite/quartz norite mass. In this fissures the granite dykes were emplaced, according to Foslie. It seems implied in this explanation that the lateral compression was less active at the late stages of crystallization.

The present writer agres with Foslie that crystal fractionation is the main cause for differentiation of the Råna massif. This opinion seems to be safely based on field relations and the mineralogical-chemical details described. What is not so clear, however, is what is the "driving force" in this fractionation process. It seems obvious, from structural features and mineralogy, that the massif is not syntectonic in relation to the main phases of orogenic folding and that it is postdating the regional metamorphism. Obviously some less penetrative fold movements affected the massif and it thus seems justified to consider the massif as late-tectonic. The squeezing theory therefore has gained little support by the present investigations. In the writer's opinion gravitative settling, although discarded by Foslie, *may* have played a part in the differentiation process. This is, for instance, indicated by the rock distribution within the massif: The heavy ultrabasic and basic rocks are found along the marginal zone which according to Foslie's own interpretation (1921, p. 707) forms the lower part of the body, while the central and upper parts are formed by the intermediate and lighter rock types. It will be an interesting task for future detail investigations to evaluate the importance of gravitational forces on rock distribution and in the differentiation process.

Apart from the mentioned similarities in the rock types there are few direct evidences for the genetial connections between the Råna massif and the other occurrences of the younger rocks. The Istind massif could be an erosional rest of a similar (but smaller?) igneous body, but there are no direct proofs. Such an explanation was used by Foslie for the Sørfjorden "norite" (Foslie, 1941, p. 47), but any convincing evidences were not presented. The other occurrences of the younger rocks are all small bodies without the variation in composition found in Råna. However, a close connection between basic and ultrabasic rocks are evident in the Istind and Orta occurrences. Field relations and the crystallization sequence of the minerals support the assumption that these rocks are related and probably formed by fractional crystallization of a basalt magma. There is nothing to indicate that this differentiation either was promoted by squeezing. There is, for instance, no signs of shearing or pressure effects in the first crystallized rocks, that is, the ultrabasics. It has been shown that in the small body of Orta there is a gradual transition from peridotite at the one side to gabbro and diorite on the other side. Although this body is now steeply dipping it is not improbable that the body were once horizontal or subhorizontal. In such a case gravitative crystal settling would seem the most probable cause of differentiation. Such an explanation is not uncommonly used for sills and sheets with ultrabasic facies'. Drever and Johnston (1967, p. 61) states that "In many cases there is convincing evidence in support of gravitative differentiation, . . . ". In the case of the Orta rocks multiple intrusion seems a less probable explanation because of the lack of sharp contacts between the various rock types.

The ultrabasic rocks in the Istind massif are small inclusions with sharp contacts and at the first glance they might be regarded as xenoliths. However, similar bodies are present in the Råna massif and evidences which link them genetically to the gabbroic rocks are too strong to be overlooked. These evidences are: Compositional variations in minerals, sequence of crystallization and interrelations of the main minerals, and the presence of transitional types gabbro — periotite (Foslie, 1921, 1922). In general, therefore, it must be assumed that rocks of the younger group with compositions varying fram peridotite to diorite originated by differentitation from basalt magma.

While genetic connections between ultrabasic, basic and intermediate rocks seem clear, it is less obvious that the acid rock types belong to the same magmatic series. In discussing the genesis of these rocks the following observations are of importance:

a) Acid rock types seem to be absent from the Istind and Orta occurrences.

b) There is clearly a compositional gap between the dioritic rocks on the one side and the young dykes on the other. (Fig. 22).

c) Granodiorites and granites are common also within parts of the area where basic members of this group seem to be absent. In most of the map area granitic rocks are far more abundant than the basic types. Foslie (1921, p. 705) stated that the granite dykes are confined to the norite massif. However, observations by the present writer, as well as later descriptions by Foslie (1941, p. 165) show that such dykes (with varying Na_2O/K_2O proportions) occur also outside the massif and in part far away from it.

On background of these observations it would not seem unreasonable to assume a more independent origin for the acid dyke rocks. It is not at all evident, at least, that they were formed by differentiation from the same basalt magma as the basic members. It is of interest to see what consequences it brings about for the genetic theories on the Råna massif if the granitoid rocks are excluded from the comagmatic series: According to Foslie (1921) the "granitic" rocks constitute about 3,9 per cent (by weight) of the whole mass. If this is subtracted from the total and the rest recalculated to 100 per cent, the composition of the undifferentiated magma becomes as listed in Table XIX, column (b). A comparison with the parental magma composition as given by Foslie (col. (a)) shows the differences to be small and relatively insignificant. Olivine and hypersthene are normative minerals in both cases, and the original magma is olivine tholeiitic irrespective of in what way the acid dykes are considered. It can be concluded, therefore, that an "independent" origin for the granitic rocks is possible without rejecting the crystal fractionation theory for the ultrabasic, basic and intermediate rocks.

Like it was proposed for the acid rocks of the older group an anatectic origin can be suggested for the granites and granodiorites. Fig. 23 shows that these younger dyke rocks plot in, or close to, the field of minimum melting composition, and an anatectic mode of formation is thus in this case possible without postulating secondary processes (an exeption may be some trondhjemitic variants). It is again possible that this melting occurred within the Precambrian basement and, like it was presumed in the case

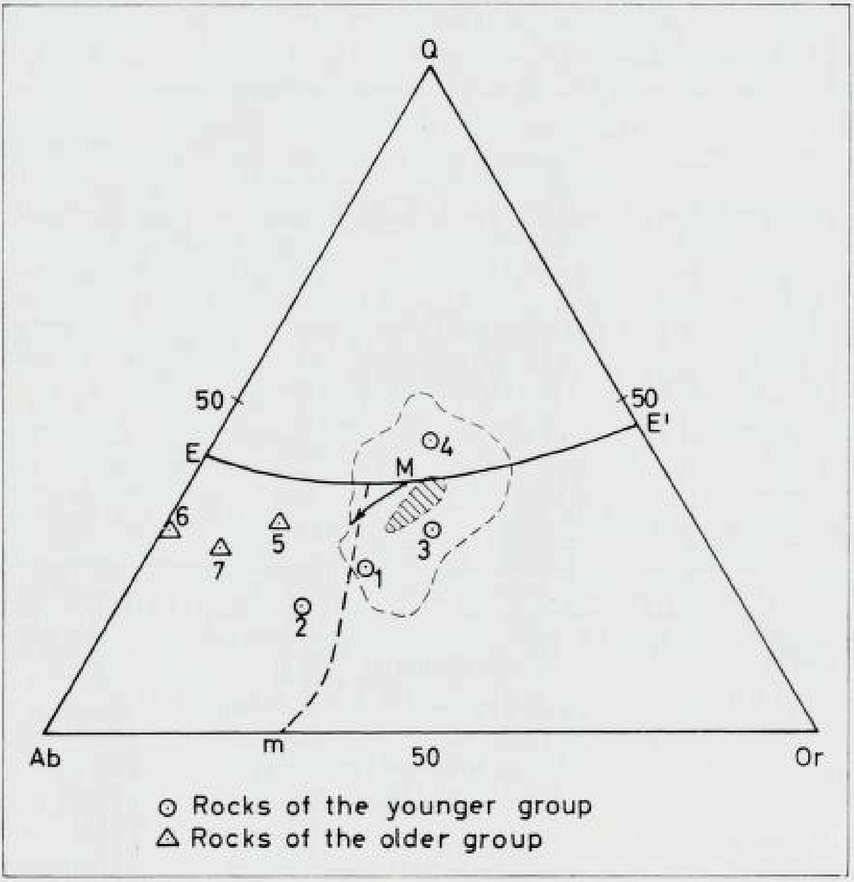

Fig. 23. Acid rocks of the area plotted onto a Q-Ab-Or diagram. The "thermal valley" (m) and minimum (M) at 1000 kg/cm^2 water vapor pressure are redrawn from Tuttle and Bowen (1958, p. 65). The line EE' is the boundary for quartz and feldspar separation. Position of granitic plutonic rocks is delimited by the stippled line (around M) and the maximum concentration indicated by the hatched area. The different position of younger and older group rocks in relation to the minimum melting composition is evident from this diagram.

of the older group rocks, that this anatexis was promoted by the ascent of hot basalt magmas. In this way the supposed time connections between acid and basic rock types are explained without their being part of a comagmatic series.

Following the reasoning of O'Hara (see p. 90) it can be assumed that the tholeiitic basalt magma rose to relatively shallow depths before its differentiation into ultrabasics, gabbros and dioritic rock types occurred.

Table XIX

The composition of the undifferentiated magma of the Råna massif, (a) including and (b) excluding the granitoid rocks.

	(a)		(b)	
	Wt %	Cat %	Wt %	Cat %
SiO_2	52,6	48.2	51,7	47,2
TiO_2	0,4	0,3	0,4	0,3
Al_2O_3	16,2	17,5	16,3	17,6
Fe_2O_3	0,5	0,3	0,5	0,3
FeO	6,7	5,1	6,9	5,3
MnO	0,1	0,1	0,1	0,1
MgO	12,2	16,7	12,7	17,3
CaO	7,7	7,5	7,9	7,7
Na_2O	1,8	3,2	1,8	3,2
K_2O	1,5	1,1	0,9	1,0
Sum	99,2	100,0	99,2	100,0

Molecular norms:	(a)	(b)
Or	5,5	5,0
Ab	16,0	16,0
An	33,0	33,5
(% An)	(67)	(68)
Ol	0,6	6,0
Hy	40,2	34,4
Di	3,6	4,0
Mt	0,5	0,5
Il	0,6	0,6
(Col. index)	(45.5)	(45,5)

Brief comparison and summary of the two igneous groups

A comparison of the main rock types of the two igneous groups gives the following results, briefly summarized:

a) Ultrabasic rocks of both groups are considered as intrusives. Connection with basic rocks in field is more obvious for those of the younger group than for the older ones. In general the younger ultrabasics seem to be higher in Ca-bearing minerals, like plagioclase and clinopyroxenes, than the older group peridotites. Serpentinization is observed in both groups. but complete alteration of parts of the bodies is restricted to the older group.

b) The amphibolites of the older group are to a great extent considered as effusive rocks; lavas and, in part, basic tuff layers. Intrusives are present in the form of dykes and larger massifs. Metamorphism have altered the older basic rocks strongly and primary textures are mostly obliterated, but occasionally porphyric textures are present. The basic rocks of the younger group are less altered and primary minerals and textures are common. All basic rocks of this group are intrusives.

c) Acid rocks of the older group are mostly soda-dominant or soda-extreme and comprise trondhjemitic and quartz keratophyric types. The latter are considered to be of effusive origin, mostly tuffs, but in part perhaps also lavas, while the trondhjemites are best interpreted as intrusives. Acid members of the youngest group are mostly granodioritic, but trondhjemites and true granites are encountered. They are all of them clearly intrusives.

It has been suggested for acid rocks of both groups that they are not directly genetically related to the basic and ultrabasic rocks of their respective groups. An anatectic origin seems more probable. In the case of the older group secondary processes are postulated as being responsible for the strong soda dominance. The acid rocks of the younger group mostly show a chemistry close to the minimum melting composition. In both cases melting may have been promoted by the heating effect of ascending basalt magmas.

General considerations and comparisons with other parts of the Scandinavian Caledonides

As already noted, the rocks of the older group are representatives of the initial magmatism in the area. From a general point of view they can be regarded as Caledonian ophiolites. The younger group is evidence of a late-orogenic igneous activity which can be correlated with rocks of the mica diorite group in certain areas (see below). The acid rocks of this younger group in the present area are mostly granodioritic in composition, while in other areas trondhjemites are more frequent.

Half a century ago, Goldschmidt (1916) subdivided the igneous rocks of the mountain chain in Southern Norway into 3 groups or kindreds:

a) The group of green lavas and intrusives
b) The opdalite - trondhjemite group
c) The Bergen - Jotun rocks.

The Bergen - Jotun kindred, composed of rocks of charnockitic affinitites, is now considered to be Precambrian in age (Nickelsen, 1967, Loeschke, 1967a, 1967b). Similar rocks in the Lofoten - Vesterålen areas (Heier, 1960) have also recently been dated by Rb-Sr isotope determinations to the Precambrian (Heier and Compston, in press). In the following, therefore, only groups a) and b) above are discussed. These are both of them of undoubted Caledonian age (Strand, in Holtedahl et al., 1960, p. 175).

The group of "green lavas and intrusives" are mainly metamorphosed rocks of basaltic composition, including lavas (Støren greenstones) agglomerates and tuffs on the one side, intrusive massifs and dykes on the other. Ultrabasic rocks (as serpentinites) are not uncommon. Acid members are less abundant; examples given are the quartz porphyries described from Bømlo, W. Norway by Reusch (1888). *It is apposite here to point to the similarities between the "green lavas and intrusives" and the older group of igneous rocks in the Southern Troms and Ofoten areas.* It is not difficult to find differences in appearance and mineralogy, but these can obviously be ascribed to metamorphic facies' differences between the areas in question.

Acid rock types would seem to be more common in the present map areas, but recent investigations in the eastern part of the Trondheim Region (Chaloupsky and Fediuk, 1967) show that abundant quartz keratophyres are associated with metamorphosed basic lavas in that area, too. The occurrence og trondhjemites obviously older than the opdalite-trondhjemite group has previously been recognized (Goldschmidt, 1916, p. 85). In the view of the present writer these could well be associated in time with group of "green lavas and intrusives", thus being comparable to trondhjemites of the older group in Troms.

The opdalite- trondhjemite group is displaying a varied association of rocks from peridotites and pyroxenites to granitic and trondhjemitic types. The basic and intermediate rocks include norites, quartz norites, diorites and opdalites. *There is thus in the rock types close similarities to the younger group in Ofoten and Southern Troms. Also when textures and structural relations are considered, obvious similarities between the two groups exist.* The opdalite-trondhjemite group is composed of nontectonized rocks were metamorphism is insignificant. Magmatic textures are commonly present. The acid dykes of the group are, however, frequently somewhat tectonized and it is obvious, therefore, that all rock types must be older than the last fold movements. It would seem natural, then, to assume a late-orogenic emplacement of rocks in this group in the same way as it has now been proposed for the younger igneous rocks in Troms. To make the picture more complete the general absence of volcanic rocks in the two groups should be noted.

It has been argued by some authors (Kolderup, 1941; Strand, 1960) that it may be difficult to discriminate between the opdalite - trondhjemite group and the group of green lavas and intrusives. It seems probable that these problems will become less as detailed mapping and structural and petrologic studies of igneous rocks advance. At the present state of knowledge, therefore, it may be most correct to maintain the subdivision.

Recent investigations in other parts of N. Norway have proved the existence of more than one phase of igneous activity: Mason (1967) in a contribution to the geology of the Sulitjelma area, have shown that the Sulitjelma gabbro, previously thought to form a single intrusive body with variable degrees of metamorphism (Vogt, 1927), in fact consists of two different complexes; a layered gabbro and a sequence of basic metavolcanics. As there are considerable differences in the degree of alteration and no obvious connection between the two complexes it would seem probable that they belong to two different phases of igneous activity with the layered gabbro complex as a result of the younger one.

A representative of a younger phase of igneous activity is also the Haldit massif, at the Finnish border in Northern Troms, described by Hausen (1936, 1942). The massif is composed of ultrabasic and gabbroic rocks which are completely unmetamorphosed and mostly non-tectonized. Only some jointing and faulting took place after their crystallization. Although only brief descriptions are given in these papers, it is evident that some ultrabasics (e.g. the troctolites) are banded and that large-scale layering is present within the massif (see descriptions and Fig. 15 in Hausen, 1942). Ultrabasic rocks are abundant in the lower parts of the massif and gabbros in the upper part. Hausen compared the Haldit rocks with the Lyngen, Vaddas and Råna massifs and found considerable analogies. To the present writer it seems near at hand to assign the Haldit massif to the same late phase of igneous activity as the younger group in the Southern Troms and Ofoten areas.

Basic/ultrabasic igneous complexes in the coastal distrikts of Sørøy, West-Finnmark, have been investigated in recent years. (Stumplf and Sturt, 1965; Sturt, Miller and Fitch, 1967). At least three phases of intrusive igneous activity are recognized (Sturt, Miller and Fitch, 1967, p. 258) in this area. The latest of these phases is the emplacement of alkaline rocks during the late stages of the second deformation phase (Sturt and Ramsay, 1965). These latter rocks are of special interest because they were used for K-Ar dating by Sturt, Miller and Fitch (1967). The apparent ages fall in two different categories, 480 - 491 m.y. and 384 - 420 m.y. (determinations on nepheline and micas, respectively). The orogenic events contemporaneous with the emplacement of the igneous rocks (highest apparent ages) were tentatively correlated with the Lower Ordovician Trondheim disturbance of South Norway (Vogt, 1928, 1945) and with the 500 m.y. phase of the Caledonian orogeny of the Dalradian series. The second group of ages (end-Silurian?) fits in with several K-Ar datings from the Caledonian mountain chain in Norway (Broch, 1964), as also pointed out by Sturt, Miller and Fitch (1967, p. 268).

Age determinations are lacking from the present map area (except on Precambrian basement rocks), and it is not possible to place the igneous groups exactly in the time scale. It could be suggested, however, that the emplacement of the older group is roughly contemporaneous with the Lower Ordovican greenstones and associated rocks in South Norway. The Elvenes conglomerate with its pebbles of trondhjemite rock types shows that movements occurred before its deposition and probably rather early in the orogenic history of the area. It is tempting to compare this phase with the 490 m.y. phase of West Finnmark and with the Trondheim disturbance in South Norway. The younger group postdates the main metamorphism but the exact age of this metamorphism is as yet unknown. Most K-Ar datings from different parts of the Norwegian Caledonides (Broch, 1964) lie in the range 370 - 425 m.y. (Silurian) and it is possible that these ages mark approximately the culmination of regional metamorphism in the different areas. The latest Caledonian movements is the Svalbardian phase (Vogt, 1928) in the Devonian. This is obviously the minimum age of the younger group as some rocks of the group are definitely affected by fold movements. It is of interest in this connection that K-Ar age determination on biotite from a hypersthene-mica diorite in the southern part of the Trondheim Region (the rock belonging to the opdalite-trondhjemite group) gave 355 m.y., that is, early Devonian age (Broch, 1964).

Brief comparison of the igneous rocks with alpine ophiolites

The present brief comparison is based upon the comprehensive description of the alpine ophiolites given by Burri and Niggli (1945). It is stated by these authors that the ophiolites comprise basic and ultrabasic rock types, while acid rocks are practically absent. Despite strong variations in the degree of metamorphic alteration the rocks involved show restricted variations in chemical relations. Their chemistry is also comparable to known unmetamorphosed igneous rocks. Metasomatic processes and transport of material during metamorphism therefore seem to be only of local importance. The two authors conclude that their investigations show that it is possible to a certain degree to draw conclusions about primary genetic processes also in metamorphosed igneous rocks. This conclusion is of interest as the same assumption has been made in the present paper for the rocks of the Southern Troms and Ofoten areas. Its correctness is, of course, crucial for our discussion of primary relations. The fact that most rocks correspond well with common magma types is an indication that the supposition is fulfilled.

Comparison of chemistry with that of the alpine ophiolites is achieved by plotting in a diagram (Fig. 24), redrawn from Burri and Niggli (1945, p. 129): *The basic/ultrabasic rocks of the older group all fall within the fields of the alpine ophiolites as far as c, al, fm and alk values are considered. The younger group rocks are partly within, partly outside, these fields.* Differences are most pronounced when *c* and *alk* values are compared.

Burri and Niggli (ibid., p. 570) states that a spilitic-keratophyric chemistry is occasionally found among the ophiolite rocks but as a rule

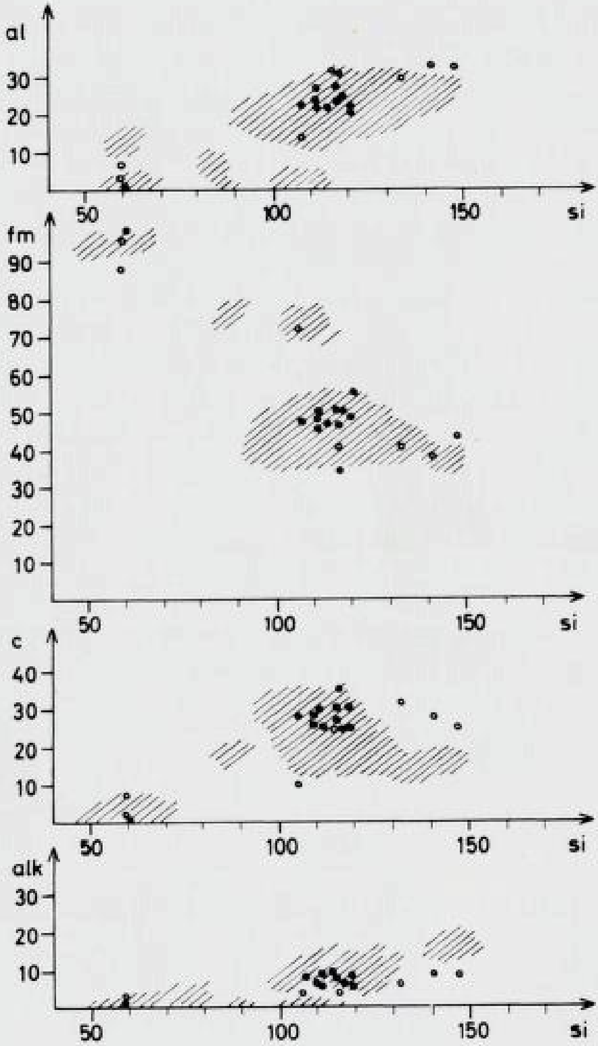

Fig. 24. Plots of Niggli values of basic and ultrabasic rocks from the present area. Dots: Older group rocks, the same as listed in Tables II, IV, V and VII. Rings: Younger group, represented by the three analyses of Table XIII and analyses 1, 2, 3 and 5 from Foslie (1921, p. 709). The hatched areas show concentration of alpine ophiolite analyses, redrawn after Burri and Niggli (1945, Fig. 43).

a more "normal" composition is found. That some variations occur is, however, obvious from the following statement: "Im grossen gesehen ergibt der Vergleich des Chemismus unserer Ophiolithe mit den vergleichsweise herangezogenen magmatischen Provinzen, dass die Ophiolithmagmen sowohl Glieder enthalten, die sonst mit typisch pazifischen Provinzen vergesellschaftet sind, als auch Glieder der Olivinbasaltisch-alkalibasaltisch - mugearitischen Vergesellschaftung". *The similarity in the scheme of variation with that found in the older group of Southern Troms is evident.* It is clearly suggested in the cited paper that ultrabasic and basic rocks are comagmatic, an opinion later also advocated by Thayer (1967).

Concerning field relations certain differences between the alpine ophiolites and the rocks of the present area are evident: While the ophiolites of the Alps and the Mediterranean area frequently occur as large igneous bodies, the amphibolites and serpentinites of the present map area are small occurrences. Moreover, while the older basic rocks in the S. Troms area have been shown to be of supracrustal origin in most cases, the alpine ophiolites are, to a great extent at least, considered as intrusive rocks (Burri and Niggli, 1945; Thayer, 1967).

It is suggested by Burri and Niggli (ibid. p. 567) that the ophiolites were emplaced during the geosynclinal stage and, eventually, during the first fold movements. This interpretation of the time relation to orogeny corresponds well with what has been concluded in the present paper (p. 101) about the rocks of the older igneous group.

In conclusion it can be stated that strong similarites exist between the alpine ophiolites and the older group rocks of the present area. These similarities are so important that it seems justified to consider the latter rocks as *the Caledonian ophiolites* of the Southern Troms and Ofoten areas.

Lastly it should be noted that syn- and late-tectonic diorites, quartz diorites and granites are described also from the Alps. These are not included with the ophiolites but are distinctly younger. They seem, in some respects at least, to be a parallel to the younger group in the present map area.

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Abbreviations:

NGU = Norges Geologiske Undersøkelse.
NGT = Norsk Geologisk Tidsskrift.

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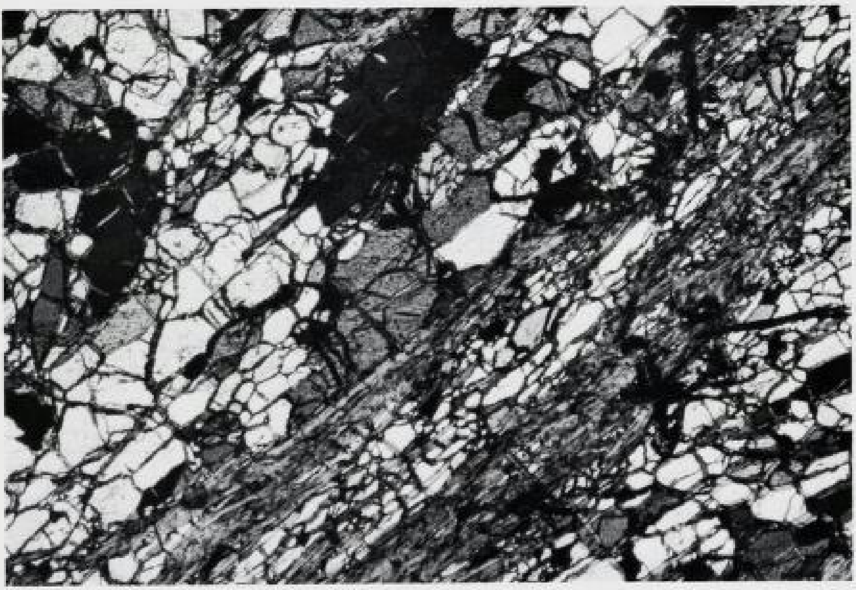


Pl. I a. Parallel texture in serpentinized olivinite from Hesjeberg, Gratangen. (Crossed nicols, 35 x).

Pl. I b. Zoned veinlets of haematite + chrysotile in olivine, Vasskaret peridotite, Andørja. (Plane polarized light, 35 x).

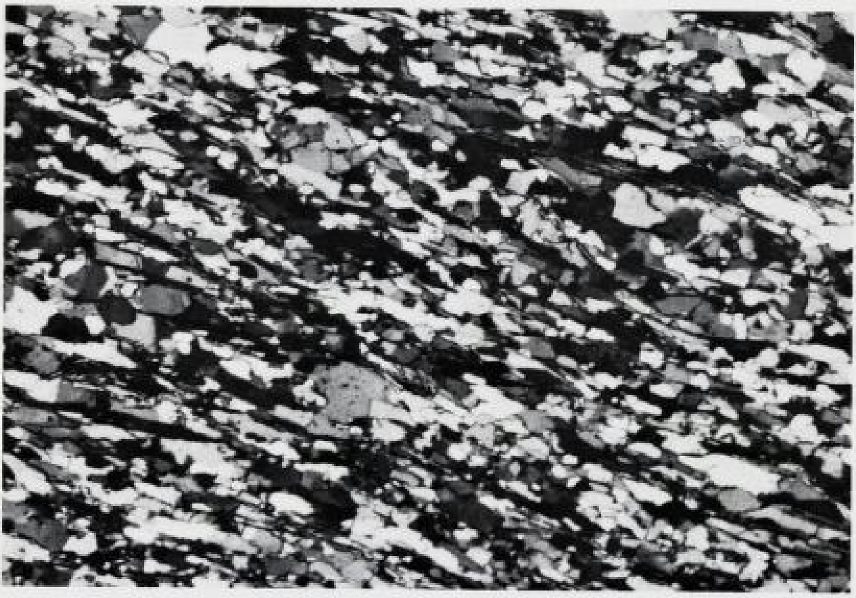
Pl. II a. Altered plagioclase phenocrysts in the Iselvdal amphibolite. Note that clinozoisite needles are mostly absent from the outer rim. The surrounding groundmass consists of hornblende, plagioclase and ore minerals. (Plane polarized light, 35 x).

Pl. II b. Massive-textured amphibolite from Gamvikskar, Lavangen. Brownish hornblende + plagioclase. (Plane polarized light, 35 x).

Pl. III a. Euhedral plagioclase phenocryst (?) in quartz keratophyre, Skårvikelven, Salangen. (Crossed nicols, 35 x).

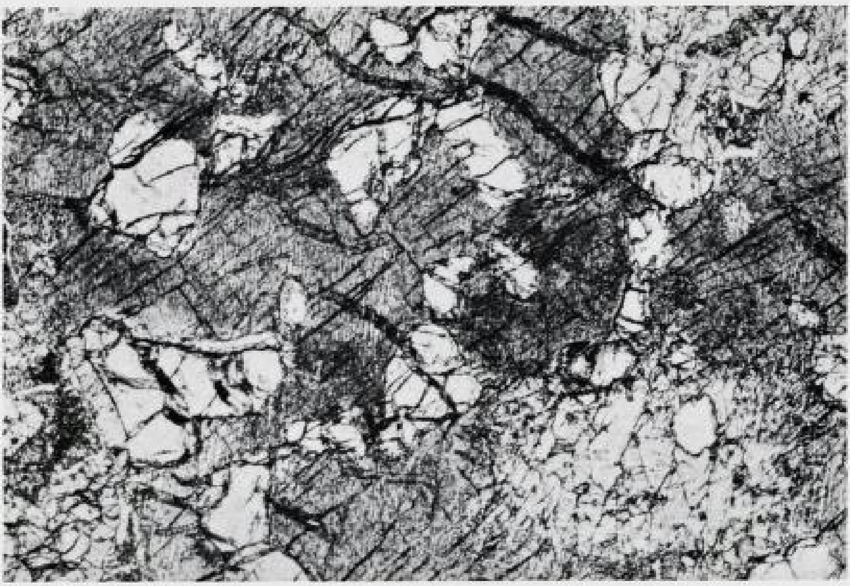
Pl. III b. Sericitization and biotite formation along cleavage directions in plagioclase. Deviating trends of some biotite flakes (dark), are visible. (Crossed nicols, 100 x). Same specimen as on Pl. III a.

Pl. IV a. Quartz keratophyre from Salvasskaret, Bardu. The larger subhedral plagioclases are twinned according to the Carlsbad and Baveno (?) laws. (Crossed nicols. 35 x).

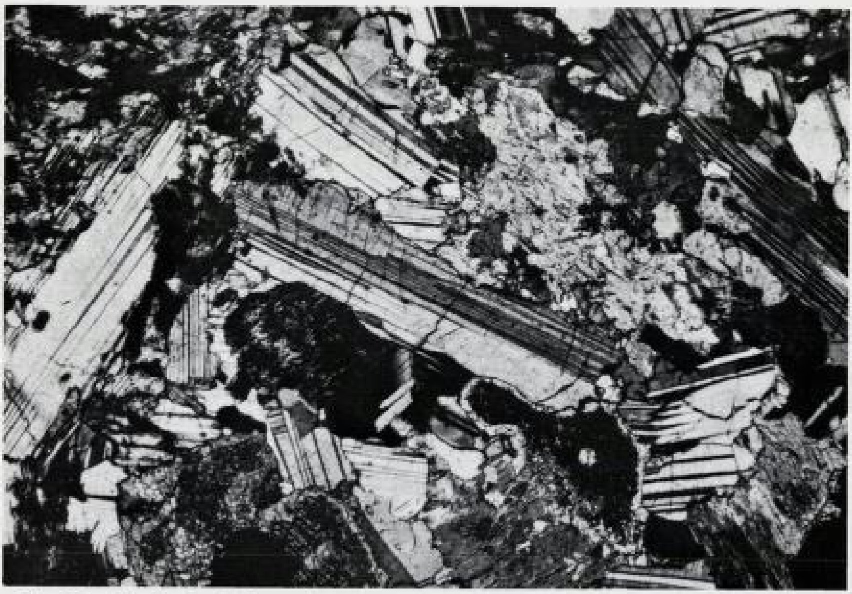


Pl. IV b. Common type of metamorphosed quartz keratophyre, Salvasskaret, Bardu. (Crossed nicols, 35 x).

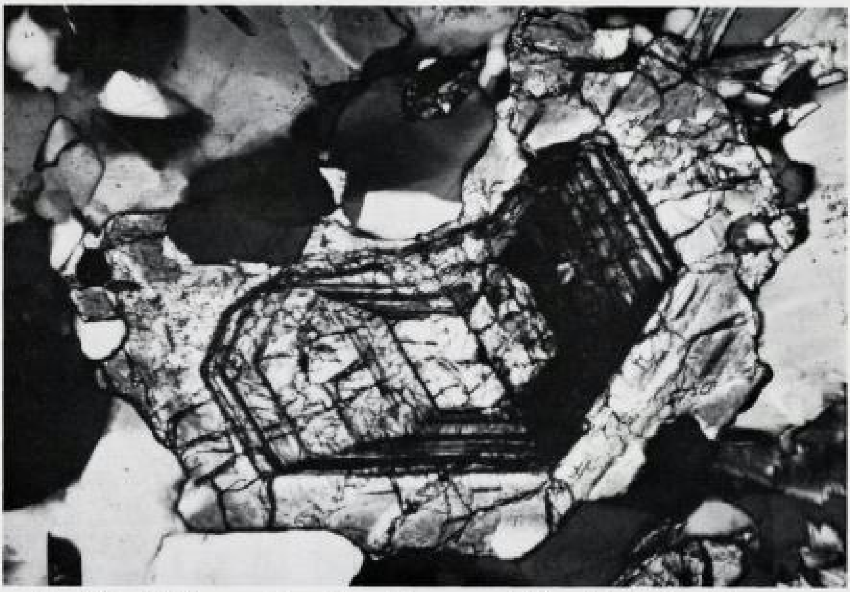
Pl. V a. Foliated trondhjemite, Rundkollen, Lavangen. Parallel orientation of biotite, muscovite and elongate feldspar grains. (Crossed nicols, 35 x).



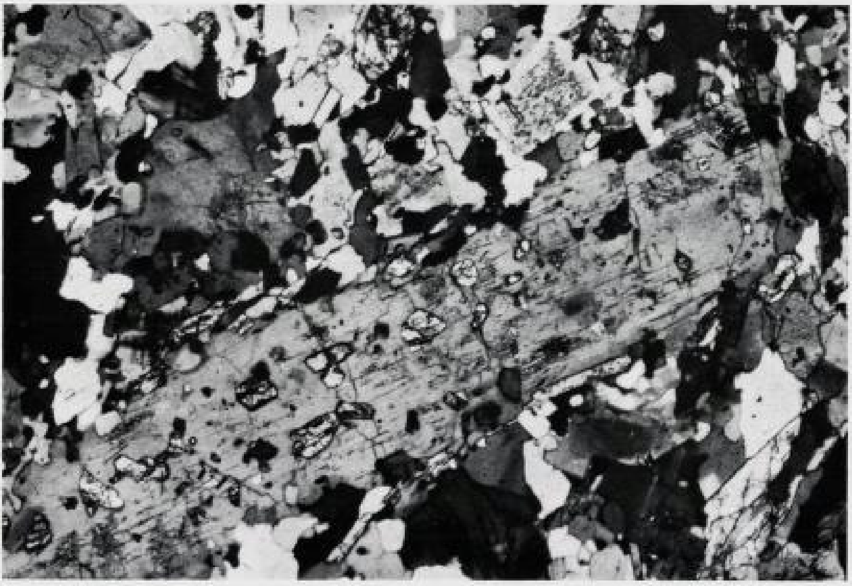
Pl. V b. Poikilitic inclusions of olivine in clinopyroxene. From ultrabasic inclusion in the Istind diorite. (Plane polarized light, 35 x).



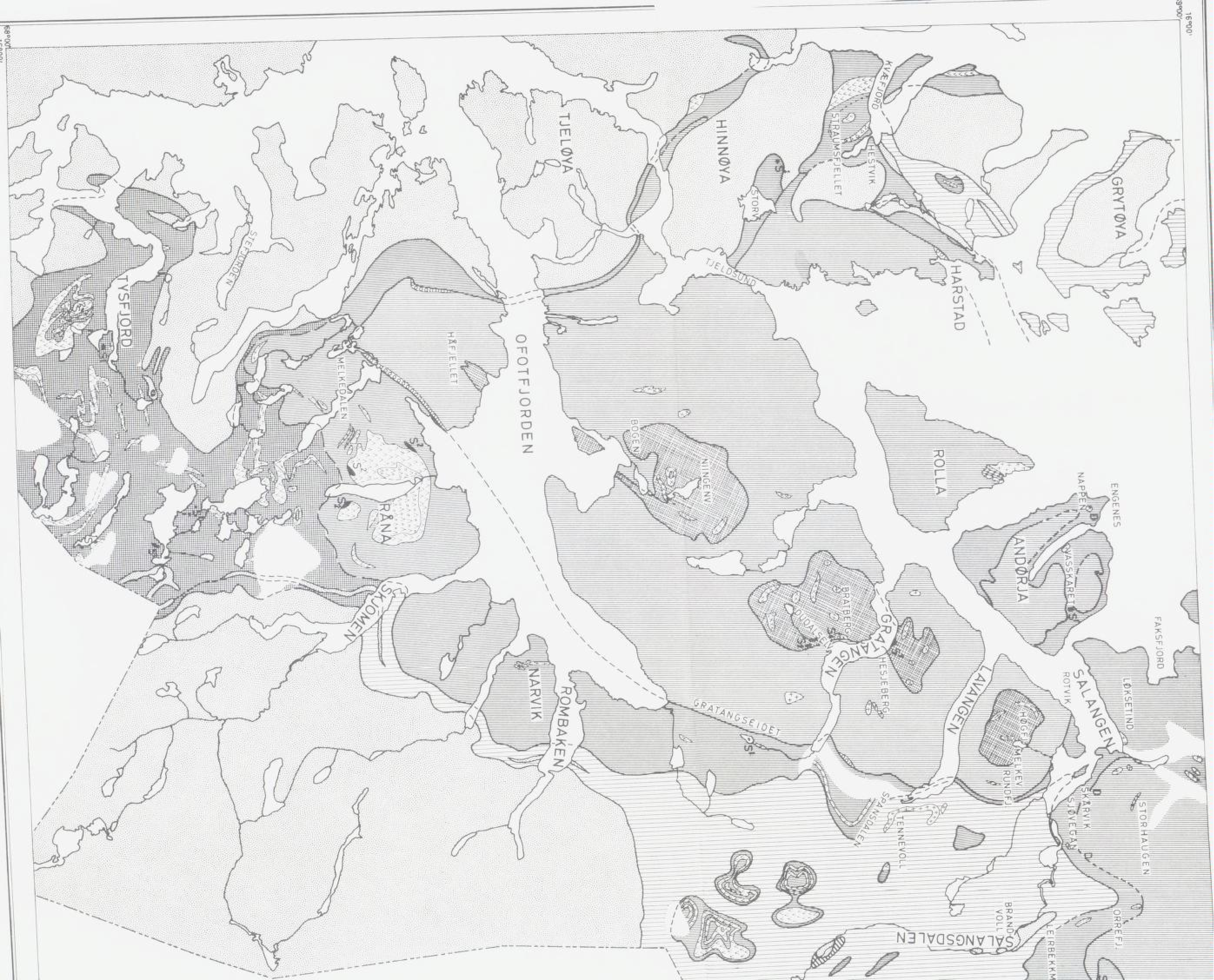
Pl. VI a. Plagioclase laths and partly uralitized clinopyroxene in the Istind diorite. (Crossed nicols, 35 x).



Pl. VI b. Orthite core in epidote, from granodiorite, Rundkollen, Lavangen. (Crossed nicols, 35 x).



Pl. VII a. Saussuritized plagioclase lath in granodiorite, Rundkollen, Lavangen, same specimen as on Pl. VI b. (Crossed nicols, 35 x).


16°00'
16°00'
TYSFJORD
STEFJORDEN
TJELØYA
HINNØYA
KVÆFJORD
STRAUMSFJELLET
HESTVIK
STORV.
S
GRYTØYA
HARSTAD
TJELDSUND
OFOTFJORDEN
HÅFJELLET
MELKEDALEN
RÅNA
S¹
S²
S³
BOGEN
NINGEN
ROLLA
ENGENES
NAPPEN
ANDØRJA
VASSKARET
FAKSFJORD
SALANGEN
ROTVIK
LØKSETIND
LAVANGEN
HØGFJ.
MELKEV.
RUNDFJ.
GRATANGEN
BRATBERG
DUDALSELV
HESJEBERG
GRATANGSEIDET
SKJOMEN
NARVIK
ROMBAKEN
SPANSDALEN
TENNEVOLL
SKARVIK
SJØVEGAN
STORHAUGEN
BRANDVOLL
SALANGSDALEN
LEIRBEKKM
ORREFJ.

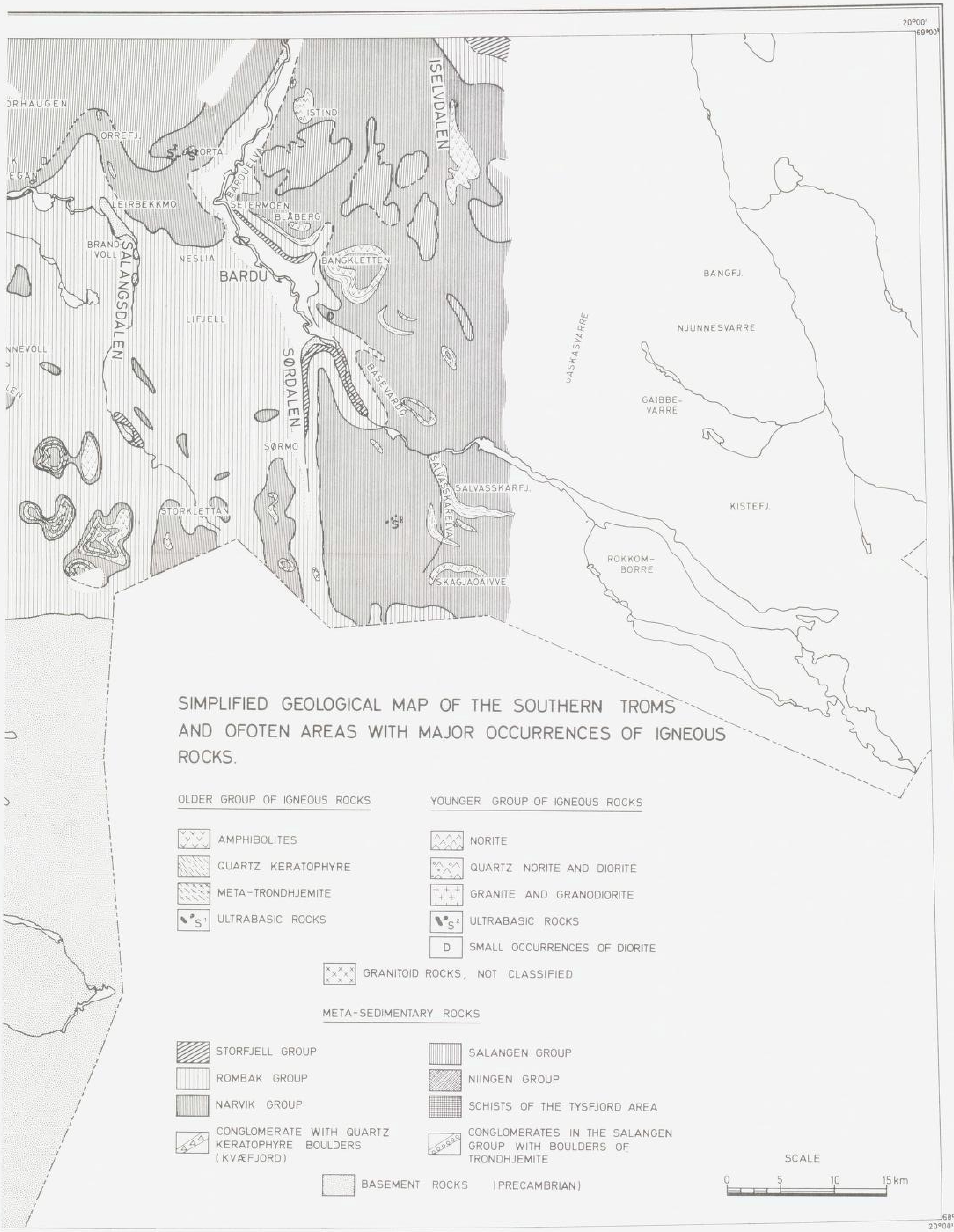

SIMPLIFIED GEOLOGICAL MAP OF THE SOUTHERN TROMS AND OFOTEN AREAS WITH MAJOR OCCURRENCES OF IGNEOUS ROCKS.
OLDER GROUP OF IGNEOUS ROCKS
AMPHIBOLITES
QUARTZ KERATOPHYRE
META-TRONDHJEMITE
S1 ULTRABASIC ROCKS
YOUNGER GROUP OF IGNEOUS ROCKS
NORITE
QUARTZ NORITE AND DIORITE
GRANITE AND GRANODIORITE
S2 ULTRABASIC ROCKS
D SMALL OCCURRENCES OF DIORITE
GRANITOID ROCKS, NOT CLASSIFIED
META-SEDIMENTARY ROCKS
STORFJELL GROUP
ROMBAK GROUP
NARVIK GROUP
CONGLOMERATE WITH QUARTZ KERATOPHYRE BOULDERS (KVÆFJORD)
SALANGEN GROUP
NIINGEN GROUP
SCHISTS OF THE TYSFJORD AREA
CONGLOMERATES IN THE SALANGEN GROUP WITH BOULDERS OF TRONDHJEMITE
BASEMENT ROCKS (PRECAMBRIAN)
SCALE
0
5
10
15 km
20°00'
69°00'
68°0
20°00'
ORHAUGEN
ORREFJ.
EGAN
PORTA
ISTIND
ISELVDALEN
LEIRBEKKMO
BARDUELVA
SETERMOEN
BLÅBERG
BRAND-VOLL
SALANGSDALEN
NESLIA
BARDU
BANGKLETTEN
LIFJELL
NNEVOLL
SØRDALEN
BASEVARDO
SØRMO
SALVASSKARFJ.
SALVASSKARELVA
STORKLETTAN
SKAGJAOAIVVE
GASKASVARRE
BANGFJ.
NJUNNESVARRE
GAIBBE-VARRE
KISTEFJ.
ROKKOM-BORRE

Fig. 1.