NORGES GEOLOGISKE UNDERSØKELSE NR. 239

THE CALEDONIAN MOUNTAIN CHAIN OF THE SOUTHERN TROMS AND OFOTEN AREAS

Part I. Basement Rocks and Caledonian Meta-sediments

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With 2 maps, 38 text-figures 22 tables

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Contents

Abstract.

The geology of the present area will be treated in a series of three publications. In this first paper the petrography and general features of the Precambrian rocks and the so-called "basal gneisses" are described and the interrelations of these rocks are discussed. In addition, description and comments on the stratigraphy, petrography and metamorphism of the Caledonian meta-sediments are presented.

It is shown that chemical, mineralogical and general structural relations point to a close relationship between the Precambrian Rombak window and the basal massif of Tysfjord. Thrust sheets of Precambrian granite in the Caledonian schists within the Rombak profile and the Hinnøy area are described. These observations support the hypothesis of Kautsky that gneiss-granite layers in schists of the Tysfjord area as well as the border gneiss of the Tysfjord granite are thrust sheets of Caledonized Precambrian rocks.

The presence of thrust planes at several levels within the Caledonian meta-sediments renders the original stratigraphy uncertain. The sequence is divided into the following groups, in ascending order: Hyolithes Zone, Storfjell Group, Rombak Group, Narvik Group, Salangen Group and Niingen Group (Table III).

The Hyolithes Zone occurs along the eastern margin of the Caledonian area and, to some extent, in connection with Precambrian windows within the Caledonian schists. It is composed of unmetamorphosed (or very weakly metamorphosed) sandstones and shales and is resting with a basal arkose or conglomerate on weathered Precambrian rocks. From previous investigations and fossil evidence the Zone is known to be of Lower Cambrian age. The Hyolithes Zone is separated by a thrust zone from the overlying Storfjell Group. This latter group consists of feldspar-bearing schists, mostly quartzitic, and dolomite layers or lenses. Clastic textures are retained in parts of the group. The feldspathic rocks show resemblance to sparagmites as described from other parts of Norway. The Rombak Group is composed mainly of marbles and mica schists, metamorphosed in the biotite and the garnet zones. A well-marked thrust plane has been observed at the base of the Rombak Group in several localities. In the Rombak area the Group borders directly on tectonized Precambrian granite. A thrust plane also separates the Rombak Group from the overlying Narvik Group. The latter is mainly pelitic with mica schists and -gneisses as the predominant rocks. Quartzites are present in the lower part of the Group. In eastern areas rocks thought to be of volcanic origin (keratophyres) are common. Agglomerate-like rocks in the Kvæfjord area possibly have a comparable stratigraphical position. The Salangen Group, overlying the Narvik Group without any recognisable tectonic break, consists of marbles, mica schists and meta-sedimentary iron ores and includes the rocks previously termed the Elvenes limestone Group and the Bogen Group. A conglomerate (the Elvenes conglomerate) is present at the base of the Group in the area south of Ofotfjorden: it has also been observed in the Harstad area. The uppermost member of the Caledonian sequence, the Niingen Group, consists of the same rock types as those of the western parts of the Narvik Group. The tectonic position of the Niingen Group is uncertain.

Parts of the Narvik, Salangen and Niingen Groups are metamorphosed in the staurolite and kyanite zones. Metamorphic grade within the meta-sediments thus ranges from the unmetamorphosed autochthonous rocks to almandine-amphibolite facies in some of the allochthonous schists. The variations are shown on a facies map (Fig. 2). The boundaries between different facies are partly coincident with the thrust planes, thus indicating that the thrusting post-dates the main regional metamorphism. Where the thrust plane is also a metamorphic boundary higher grade rocks are everywhere overlying those of a lower grade. Otherwise there seem to be no particular regularities in the regional distribution of the facies', for example along an east-west profile. The metamorphic history, as interpreted in this paper, can be subdivided into the following three phases. 1. Main regional metamorphism, contemporaneous with the first main folding. 2. Porphyroblastic growth of amphibole, plagioclase, garnet (?), staurolite (?) and biotite in a static period subsequent to the first folding. 3. Retrograde metamorphism, mainly chloritization of ferromagnesian minerals, connected with late thrust movements. The mutual stability relations of kyanite and staurolite are discussed and it is concluded that P-T conditions are the determining factors in these relations, chemical features of host rocks being of little or no importance.

The association of amphibole and muscovite in certain schists within the almandine-amphibolite facies area is found to be due to the polyphase metamorphism of the rocks. This mineral assemblage is probably not in equilibrium.

Eruptive rocks, both intrusives and volcanics(?), are present in the Narvik Group and higher units. The eruptives comprise rocks from ultrabasic and basic to trondhjemitic and granitic chemistry, though intermediate types are mostly absent.

Taking into account the presence of eruptive rocks, metamorphism and thicknesses, it appears justified to consider the Narvik, Salangen and Niingen Group as eugeosynclinal sediments, whereas lower tectonic and stratigraphical units are to be looked upon as miogeosynclinal. The chemistry of the meta-sediments has been enlightened by chemical analyses. It appears that the Hyolithes Zone and Rombak Group rocks are mainly of the so-called eastern facies type (residual-type sediments). In the Narvik Group schists of eastern and western (incompletely weathered) type alternate. It is concluded that this alternation is most probably due to differences in chemistry of the source rocks rather than to changes in the climatic or sedimentation conditions.

Structures are mentioned only briefly in this paper: The first (?) folding, contemporaneous with the main regional metamorphism was largely plastic in nature, producing folds with Caledonian trends (NE—SW) and cross folds (axes NW—SE). The Caledonoid folds are partly overturned towards the SE. Later thrusting, towards the SE and ESE, occurred at various levels. Flexural folding on NNE axes may be partly contemporaneous with, or younger than the thrusting as thrust planes are in part folded by these late movements.

INTRODUCTION

Location and topography of the area

The key map (Fig. 3) shows the location of the map area which is situated between 68° and 69° N latitude and between 16° and 20° E longitude, the boundaries are those of the AMS map Narvik (1 : 250 000). About ¾ of the map area are on the Norwegian side of the national border and are to be treated in this paper. Most attention will be paid to the areas mapped by the writer with assistants (Fig. 3).

The roughest topography is undoubtedly found in the area from Rombaken south-southwestwards to Tysfjord, the highest mountain of the map area, Storsteinfjell east of Skjomen, rises to about 1900 metres above sea level. The topography is likewise rugged in the coastal areas between Ofotfjorden and Salangen, smaller parts of which are hardly accessible for inspection. Exposures are mostly good within the coastal districts, even in the lowermost parts, though on steeper hillsides scree may cover large areas. Turning to the inland districts the shape of the mountains is another one: The hillsides are frequently very steep even there, but the upper slopes of the mountains are less inclined, giving a characteristic "rounded' 'shape to the inland mountains. Exceptions to this rule are found, as for instance Kirkestind, 1681 m. above sea level, the highest peak in the map area north of Ofotfjorden. The differences between the rugged topography of the outer districts and the rounded forms of the inland seem, at least partly, to be due to differences in the petrography of rocks forming the higher parts of the mountains: Whereas these are mica schists and gneisses in the peaks along the coast and in mountains like Kirkestind in eastern areas, the inland mountains are frequently built up of marble-bearing sequences in their higher parts. The relatively flatlying layers are, of course, of importance in the development of topographical features, but this seem mostly to be common to the coastal and inland districts. It is probable that differences in glacial erosion played a great part in the formation of the varying topography of the area, but it is outside the scope of this paper to discuss such problems.

Exposures are mostly excellent in the inland districts, exceptions are the valley bottoms, which are heavily covered, in part filled completely by Quaternary terrasses.

Also climatic conditions change when moving inland from the coastal area. The latter typically has wet and chilly, in part foggy weather during the summer months, whereas the inland summers are fairly dry and with pleasant temperatures for field work.

Previous work

A list of the most important older publications on the geology of the area, or parts of it, is given below (Quaternary geology excluded). If not specially referred to in the text, these papers are not included in the reference list.

Pettersen, Karl, 1868. Geologiske Undersøgelser i Tromsø Omegn i Aarene 1865—67 (Kgl.N.Vid.skb.Skr. Bd 5, H2).
— 1874. Geol. Undersøgelser inden Tromsø Amt og tilgrænsende Dele av Nordlands Amt (Kgl.N.Vid.Skb.Skr. Bd. 7).
— 1876. Risehulen ved Lavangsbotten. (Arch. f. Math. og Nat. vit. Bd. 1.)
— 1878. Det nordlige Sveriges og Norges geologi. (Arch. f. Math. og Nat. vit. Bd. 3.)
— 1883. Balsfjordgruppens plads i den geologiske følgerække. (Tromsø Mus. Aarsb. VI.)
— 1886. Notitser vedrørende den nord-norske fjeldbygning. (GFF Bd. 8, H.6.)
— 1887. De geologiske bygningsforholde langs den nordlige side av Torneträsk. (GFF Bd. 9, H.6.)
— 1887. Den nord-norske fjeldbygning I. (Tromsø Mus. Aarsb. X.)
— 1888. Den nord-norske fjeldbygning II, Del 1. (Tromsø Mus. Aarsb. XI.)
— 1889. Den nord-norske fjeldbygning II, Del 2. (Tromsø Mus. Aarsb. XII.)
— 1891. Geologisk kart over Tromsø Amt (posthumous). (Tromsø Mus. Aarsb. XIV.)
Reusch, H., 1892. Nogle bemerkninger om Tromsø Amts geologi (NGU 4).
Vogt, J. H. L., 1897. Norsk Marmor (NGU nr. 22).
— 1910. Norges Jernmalmforekomster (NGU 51).
Vogt, Th., 1916 a. Kysteruptiverne i Tromsø Amt (NGT 3, pp. 66—67).
— 1916 b. Den kaledoniske deformation av grundfjeldstavlen i det nordligste av Skandinavien. (Forh. ved 16. skand. naturforskermøte.)
— 1918. Geologiske studier langs den østlige del av fjeldkjeden i Tromsø Amt (NGT 4, pp. 260—266).
— 1922. Bidrag til fjeldkjedens stratigrafi og tektonikk. (GFF 44, pp. 714—739.)
— 1923. En postglacial jordskjælvsforkastning (Naturen 1923).
— 1942. Trekk av Narvik—Ofotentraktens geologi. (NGT 21, pp 198—213.)
— 1950. Kartblad Narvik 1 : 50 000 (map without description). (NGU.)
Foslie, S., 1920. Raana noritfelt. (NGU 87, pp 1—44, English summary.)
— 1921. Field observations in Northern Norway bearing on magmatic differentiation. (Journ. Geol. XXIX, pp 701—719.)

— 1922. In Bugge & Foslie: Norsk arsenmalm og arsenfremstilling. (NGU 106, pp 10—16.)
— 1926. Norges Svovelkisforekomster. (NGU 127, pp. 114—117.)
— 1931. On antigorite-serpentines from Ofoten with fibrous and columnar vein minerals (NGT 12, pp 219—245.)
— 1941. Tysfjords geologi (NGU 149). (English summary.)
— 1946. Melkedalen grube i Ofoten (NGU 169). (English summary.)
— 1949. Håfjellsmulden i Ofoten og dens sedimentære jernmanganmalmer (NGU 174). (English summary.)

In addition to these publications there is a number of reports on ore occurrences within the area, most of which are available at NGU. Some of these refer briefly to the general geology, as for instance those by J. H. L. Vogt. The present writer has also got informations from diaries by J. H. L. Vogt and Th. Vogt on parts of the Ofoten area, mainly from the map sheets Ofoten and Narvik (1 : 100 000), the diaries are now in the Geological Institute, NTH, Trondheim.

The pioneer work in this area was carried out by Karl Pettersen in the years 1865–1890. His investigations were of regional character, covering large parts of northern Norway, his papers, however, contain a number of detail observations, many of them still have their great interest. Pettersen divided the meta-sedimentary rocks into the following groups:

Tromsø micaschist Group (Tromsø glimmerskifergruppe).
Balsfjord Group (Balsfjordgruppen).
Dividal Group (Dividalsgruppen).

These divisions were based on the variable metamorphic appearance of the rocks in question, and it is not possible to correlate the groups safely with the stratigraphy presented in this paper. The lowermost group (Dividal Group), however, is much the same as the autochthonous rocks of this paper, though parts of the sequence above the latter seem to be included in it in some profiles. The Tromsø mica schist Group covers the higher parts of the mountains, whereas the Balsfjord Group take up the lower areas between the two other groups. Pettersen recognized the increasing "crystallinity" upwards, a feature he found difficult to explain. In that connection we must remember that Pettersen published his papers before Törnebohm had lanced his idea of the great overthrusts within the mountain chain. On the geological map of Troms county ("Tromsø Amt"), published after his death, Pettersen designated as Archean the coastal gneisses ("basal gneisses" of recent times) and some gneissic parts of the Caledonian sequence, for instance parts of the Narvik Group.

Th. Vogt, in the years from 1916 to 1950, published some results of his extensive fieldwork, mainly north and east of Ofotfjorden. Most of the papers, however, are lectures held at various occasions and therefore contains little of details, though they are of the greatest general interest. In the 1942 paper, for instance, a stratigraphy of the Ofoten area was proposed, and in the 1922 paper, already, this area had been compared with other parts of the mountain chain. As the results and views held by Th. Vogt will be commented on in their proper places, we will not here go further into them. The areas mapped by Vogt are shown in Fig. 3.

Parallel with Vogt's mapping north of Ofotfjorden, Foslie investigated areas south of it. The chief results of his detailed and accurate mapping were published in the 1941 and 1949 papers. Of great value, for instance, is the establishing of a fairly detailed stratigraphy of the Håfjell syncline (Håfjellsmulden). Further data given by Foslie will be presented later in this paper. Fig. 3 shows the areas mapped by Foslie.

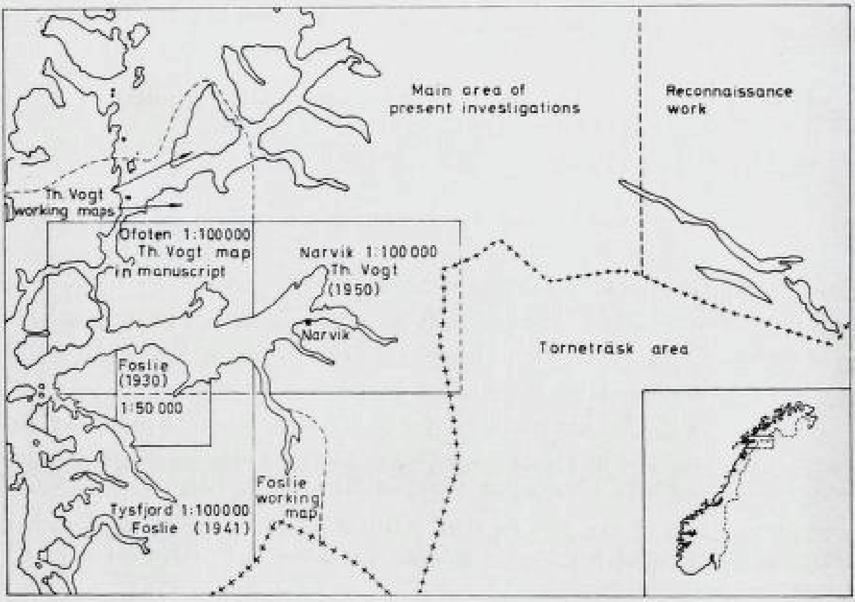


Fig. 3. Map showing areas of previous and recent investigations.
Kart som viser tidligere og nye undersøkelser.

Present investigations

Field work started in the summer of 1959 as a regional investigation of the iron ores of Southern Troms. These investigations implied extensive mapping in order to clarify the relations between the different occurrences of iron ore. The regional mapping continued the following years without any special consideration of ores. In this mapping a lot of interesting problems arose, some of which are special of this area, others of a more general kind. In the field work and indoor treatment of the collected material, the writer chose to concentrate on problems which seemed essential in understanding the total geological history of the area. A fairly thorough petrographical description was thought to be necessary for the discussion of various problems, the more so because great parts of the area were only cursorily known from before.

Important subjects are among others the metamorphism and metamorphic facies distribution, original character of sedimentary and eruptive rocks, as well as structural interrelations between the different rock types.

The results will be presented in a series of three publications: This first paper is mainly concerned with the petrography of the basement rocks and the meta-sedimentary sequences as well as the metamorphism of the latter rocks. The next paper will deal with the petrography and metamorphism of the eruptive rocks, and in the last publication the structural geology of the area will be treated. As it is evident that the stratigraphy, as it is presented in the first sections of this paper, may partly be a tectono-stratigraphy (or an apparent stratigraphy), a discussion of the original stratigraphy and some attempts at correlation with other areas are postponed till a survey of the structural relations has been given.

The area mapped during the present investigations is shown on Fig. 3. Most of it was covered by the writer (with 1 assistant in 1961, 2 in 1962), in the summers 1959, 1961, 1962 and 1963, a little mapping was also carried out in 1960 and 1964. Independent mapping was done by Per R. Lund in the area between Gratangen and Salangsdalen northwards from map sheet Narvik to Sagvannene (P. R. Lund, 1965) and by Tore Mitsem on the peninsula north of Gratangen. The present writer has made additional mapping and observations in the latter areas, and therefore feel responsible of the general picture as it is given on the maps Fig. 1 and 2 of the whole area mapped since 1959.

Some comments on recent investigations by the writer within areas previously mapped are needed: The map sheets Narvik and Ofoten (1 : 100 000), mapped by Th. Vogt, are largely undescribed. As much of the stratigraphy in the areas investigated by the present writer is based on relations within these map sheets, it became necessary to make a lot of investigations also in these areas for the case of comparison of rock sequences and of structures. To some extent even profiles south of Ofotfjorden have been reinvestigated. Concerning areas on Hinnøy some data were available on working maps by Th. Vogt. As information from diaries was scarce, however, much of these areas has been remapped. An exception is the southern area on both sides of Tjeldsund, which is entirely based on maps by Th. Vogt. The island Rolla was also previously mapped by Vogt and has not been investigated during recent years. The southernmost part of the map, the Tysfjord area, has not been visited by the writer, when this area is referred to in the text, it is based on microscopic work and Foslie's publications.

The eastern and southeastern parts of the map area are as yet rather insufficiently investigated, the map is partly based on previous observations by Th. Vogt (diaries) and K. Pettersen.

Methods

Mapping was mainly carried out on topographical maps, scale 1 : 50 000 (AMS series), the quality of which is relatively good. Within the easternmost area, from about 19° longitude eastwards, air photographs were available (scale 1 : 35 000, 1 : 40 000) and in part used during mapping, though the quality of the photographs was variable.

It is obvious that mapping of such an extensive area during some few summers of fieldwork must imply a lot of generalizations and interpolations between profiles. Many profiles have, however, been investigated in great detail, much more detailed than indicated by the map scale. This was in part necessary for a safe correlation of profiles and in the same time they gave information about lateral variations.

The microscopic nature of the rocks has been investigated in about 500 thin sections, a great number of powder preparates as well as a few polished sections (of iron ores). The chemistry and mineralogical composition of the chief rock types have been enlightened by chemical analyses and by calculated and observed modes of the rocks in question. The observed modes were determined by the point counting method with from 500 to 1,000 points in each section. Mineralogical investigations have mostly been restricted to the common microscope methods, composition of plagioclases have frequently been checked on the universal stage, some determinations of axial and extinction angles were also carried out on the universal stage. Indices of refraction were determined by the immersion method, and in a few cases X-ray technique was used in the identification of minerals. In a single case a chemical analysis was performed to determine the exact character of a mineral (an amphibole).

General outline of geology

Generally the area is built up by the following units, from above:

Table I

Western area	Eastern area.	
Niingen Group, pelitic schists and gneisses (thrust plane at base?)	Lacking.	Allochthonous sequences.
Salangen Group, pelitic schists and meta-limestones, (iron ores). Narvik Group, mainly pelitic schists and gneisses, amphibolites (thrust plane at base).	Salangen Group, same rocktypes as in western area. Narvik Group, pelitic schists, acid volcanic (?) rocks, amphibolites (thrust plane at base).	
Rombak Group, mica schists and meta-limestones (thrust plane at base).	Rombak Group, same rocktypes as in the western area. (thrust plane at base).	
Lacking.	Storfjell Group, quartzitic schists, in part arkosic. (thrust zone at base).	Parautochthonous or allochthonous
(Basal conglomerate.) Basal gneisses. (In part mobilized Precambrian? rocks.)	Hyolithes Zone, various shales and sandstones. Conglomerate. Precambrian basement.	Autochthonous

Petrographically the area is dominated by mica schists and micaceous gneisses of various kinds and by marbles, mostly calcitic. The metamorphic grade (Fig. 2) ranges from low greenschist facies to the almandine-amphibolite facies, boundaries between the different metamorphic zones are in part coinciding with thrust planes. Primary clastic textures are retained only within the Hyolithes Zone and small parts of the Storfjell Group. The most high-grade rocks are met with in the three uppermost sedimentary groups, but there is no general increase in metamorphism upwards in the different units, a more irregular lateral variation seem to be present as far as the usual criteria are applied.

The Narvik, Salangen and Niingen Groups constitute the eugeosynclinal rocks of the area, whereas the lower sequences can be assigned to the miogeosynclinal and foreland facies as far as thicknesses and presence of eruptive rocks are considered. Metamorphic eruptive rocks are met with in the three firstmentioned groups. Most common are amphibolites of "normal" composition, some of which are clearly intrusives, and trondhjemitic to granitic rocks. Serpentines are present in the Narvik and Niingen Groups. Soda-rich, finegrained, acid rocks in the Narvik Group are thought to be quartz-keratophyres, probably of volcanic origin.

Structures are not described in this paper. The most prominent features are the thrust planes separating the different tectonic units. It is shown by the table above that the Rombak Group as well as the Niingen Group (?) are thought to form separate thrust masses, whereas the Narvik and Salangen Groups together constitute a tectonic unit. The position of the Storfjell Group is debatable, it may be truly allochthonous but it is also possible to interpret the Group as parautochthonous. Thrusting is a late process in the history of the area, accompanied by retrograde metamorphism within the movement zones. A late episode of rigid or flexural folding is thought to be contemporaneous or subsequent to the thrusting. The fold axes of this phase are mainly orientated about NNE–SSW. Previous to thrusting strong folding of a more plastic nature occurred, in part producing recumbent folds overturned towards the SE. This folding was on axes about NE–SW and NW–SE and is thought to be contemporaneous with the main regional metamorphism of the area.

THE BASEMENT ROCKS

Introduction

The Precambrian basement below the supposed Cambro-Silurian sequence is exposed at several places in the middle and eastern part of the area as erosion has removed the overlying rocks. The Precambrian thus constitutes a series of windows in the topographically lowest part of these districts. In addition to the smaller windows, Precambrian rocks are met with in the large Rombak window east of Narvik and in the area by the eastern end of Altevann, where there is connection with the great Precambrian areas on the Swedish side of the border; The rocks of the immediately adjoining area are of supposed Karelian age (Magnusson et.al. 1957). The area east of Altevann has not been investigated by the writer. Some observations were made by Th. Vogt (1918), who mentions hornblende granites and syenites as the typical rocks. In recent years investigations have been carried out by Danish geologists (Berthelsen a.o.) in connection with ore prospecting (A/S Syd-Varanger), the results of which are as yet unpublished. The northern half of the Rombak window was mapped by Vogt (1942, 1950), who distinguished several rock types, both intrusives and extrusives as well as metamorphic sediments.

On a general scale the so-called "basal granites" in the western areas take a position similar to the Precambrian basement in the eastern, though boundary relations to the Cambro-Silurian rocks are different and the age relations of the gneisses are open to discussion. On the map, Fig. 1, the true Precambrian rocks and the "basal gneisses" have been given the same symbol in order to emphasize their similar structural position.

The Precambrian windows

The Dividalen–Skjold district.

The windows in Dividalen have earlier been described by the writer (Gustavson 1963). The rocks are mostly granites, which seem to be

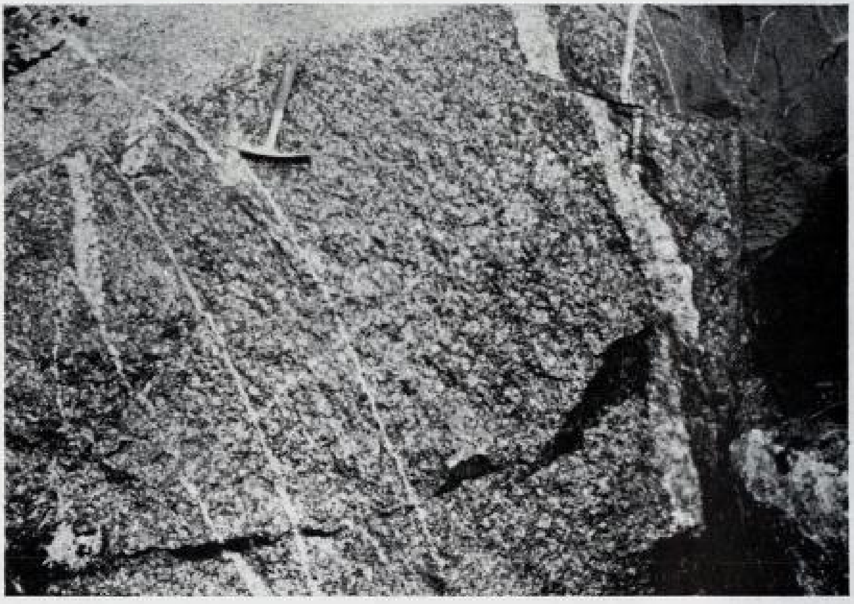


Fig. 4. Precambrian augen-gneiss cut by granitic veins S. of Målselva.
Prekambrisk øyegneis gjennomsatt av granittiske årer S. for Målselva.

of two generations: An older grey or white granite type and a younger reddish one intruding the former. Thin bands of amphibolite were observed in one locality (at Kvernbekken).

West of Skjold, about 20 km north of the Dividalen windows, an area with gneisses, amphibolite and greenstone-like rocks is encountered. The Precambrian age of these rocks is proved by the findings in 1963 of grey sandstone of the Hyolithes Zone type just above the gneisses in the easternmost part of the area. Flatlying quartzites and schists of the low metamorphic Storfjell Group follow upwards. The western part of this window near the farm Reinmelmo consists of an about 500 m thick zone of greenstone striking NW—SE and dipping 70° to the SW. The western boundary of the greenstone is not exposed. To the south it borders against flinty mylonite, which is overlain by rocks of the Storfjell Group. On the east side it has a somewhat tectonized, steeply-dipping boundary against gneisses, which are partly of augengneiss type (Fig. 4). The greenstone is partly rich in calcite with a cloddy, agglomerate-like appearance, partly it is schistose or faintly banded. Quartz lenses or stripes occur in some places. It seems probable that these greenstones represent basic volcanics, lavas or pyroclastics or both in alternation. It should be noted in this connection that greenstones are known from the Precambrian of the Swedish part of the Rombak window (Holmquist 1910); the exact character of the latter greenstones are, however, not known to the present writer.

As mentioned, the gneisses east of the greenstones are partly augengneisses with occasional coarse variants containing augens of white feldspar in the magnitude range 1 to 3 cm. Thin veins of the same feldspar occur, as well as dykes of granitic composition (Fig. 4). The granitic veins are cut by small faults, partly also by later quartz veins.

Eastwards to the eastern boundary of the window basic rocks, partly dyke-like, are abundant in the gneisses. These rocks, dark, and mostly amphibolitic in character, seem to be of two different generations. The older one is clearly affected by feldspathization processes and in part "assimilated" by granitic material. On the other hand, there are basic dykes, cutting the granitic veins, which are themselves unaffected by granitization processes (Fig. 5). The existence of two generations of basic rocks has also been described from the Precambrian of the Rombak window by Th. Vogt (1942).

Pink, microcline-rich veins are encountered in the area SE of Målselva, while the same type of veins has also been observed in the Dividalen and Rombak windows. Very similar veins have been detected within the lower part of the Cambro-Silurian sequence in two or three localities. It is therefore a possibility that they are of Caledonian age. This is supported by observations in the Skjomen profile. As

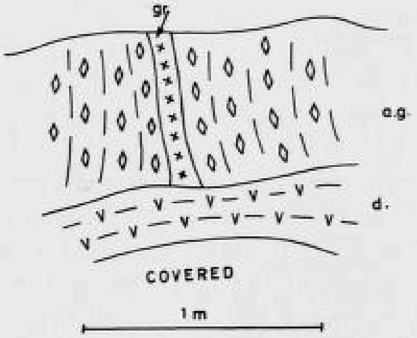


Fig. 5. Basic dyke (d) cutting augen-gneiss (a.g.) and granite vein (gr) S. of Målselva.
Basisk gang som skjærer gjennom øyegneis og granittåre syd for Målselva.

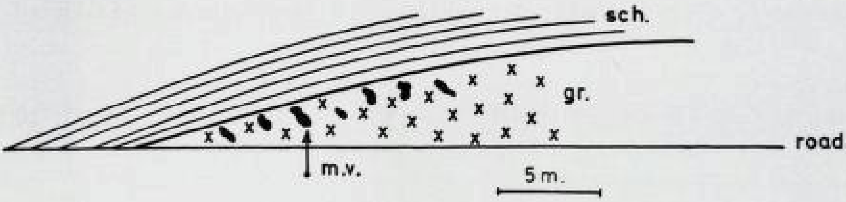


Fig. 6. Microcline veins (m.v.) in Rombak granite (gr.) near the boundary against the Rombak Group (sch.). Skjomen.
Mikroklinårer i Rombak-granitt nær grensen mot Rombakgruppen. Skjomen.

shown by Fig. 6, the microcline veins are here arranged more or less parallel to each other near to the upper boundary of the Precambrian, and in a manner suggesting them to be related to the Caledonian deformation of the uppermost part of the Precambrian (see later). It seems as though, during thrusting of the schists above and shearing of the Precambrian, fissures opened in a "Fiederspalten"-like arrangement, the fissures being filled with feldspar material contemporaneously or subsequently mobilized from the granite. In the same profile a similar, small microcline vein was found within the Rombak group. It is, of course, possible that this and the two similar veins found within the Storfjell Group in Dividalen are feldspar material mobilized from the schists themselves (thin sections show microcline to be present in schists of both groups), and it can not be fully proved that they are contemporaneous with the veins in the Precambrian rocks. In any case, the above-mentioned veins are common neither in the Precambrian nor in the schists, and an eventual mobilization during Caledonian time must have been very restricted in these areas.

The Bardu district.

By the bridge at Straumstad, 5 km south of Setermoen, a grey, medium-grained and slightly gneissic rock crops out. The foliation is steeply dipping and strikes N—S. The boundaries of this rock are nowhere exposed. Some hundred metres east of the bridge are found small exposures of quartz-rich schist of a type common within the Storfjell Group. The schist is folded, but is generally flatlying. It is probable that the gneiss at the bridge belongs to the Precambrian basement. This is supported by the orientation of structures which are divergent from those found within the schists in the neighbourhood and also by the appearance of the rock, which is very similar to the granite at Bonnes, Salangsdalen (see later). In thin section the gneiss is shown to be granitic in composition with microcline, plagioclase and quartz as chief minerals, the accessories are sericite, epidote, titanite, calcite and pyrite. The plagioclase is strongly sericitized, and albite twins are bent or displaced. Microcline occurs as larger grains (1–5 mm) with inclusions of sericitized plagioclase. Myrmekitic feldspar is seen where microcline and plagioclase border against each other. It seems probable that the crystallization of microcline postdates that of the plagioclase feldspar.

At Straumsli, some 15 km further south-southeast, another window is present. The rocks are amphibolite, granite and banded granitic gneisses. Foliation is steeply dipping with strike about N–S, the same as at Straumstad. The boundary relations are not easily studied; however, along the river Straumslitverrelva one get an impression: The upper parts of the gneisses exposed are strongly weathered; some tens of metres above, there are good exposures of quartzites and flinty, banded "hard schists" dipping gently (about 10°) to the north. In exposures nearby the hard schists are found resting upon white and grey quartzitic sandstones, probably belonging to the sequence immediately above the gneisses. The rock types and the divergence between the orientations of planar structures of the gneisses and the schists above strongly suggest that this is another Precambrian window. The quartzitic, massive sandstones probably belong to the Hyolithes Zone, whereas the hard schists above are rocks of the Storfjell Group. Petrographically the granite and granite gneisses contain varying amounts of microcline, sericitized plagioclase and quartz with small amounts of biotite, sericite and chlorite. Granitic veins cross-cut amphibolite bands in the gneisses. The amphibolite has plagioclase, epidote and hornblende as chief minerals with accessory titanite, quartz, sericite and hematite. Epidote and plagioclase occur together in aggregates evidently replacing more basic plagioclases. The present composition of the plagioclase is not easily determined because of the intergrowth with epidote.

Salangsdalen.

On the map sheet Narvik (1 : 100 000, Vogt 1950) a small area between the Rombak window and Bonnes in Salangsdalen has got the same colour as the Precambrian Rombak granite. The window at Bonnes seems to continue northwards some kilometres, mostly beyond thick cover. According to P.R. Lund (1965), who mapped this area in some detail, the rocks are grey, medium-grained granite as well as younger, light granite intruding the former. A biotite-bearing amphibolite is also intruded by the lighter granite. The present writer has briefly studied the relations at this window: Grey, granitic gneiss is here overlain by quartz schist and quartz-mica schist with a foliation differing from that in the gneiss. The exact boundary was not exposed. The relations and rock types, however, support the view that this window is of Precambrian age.

Where the river Salangselva changes its direction from N–S to E–W in the area west of Brandvoll, granitic rocks of possible Precambrian age likewise occur. At the farm Leirbekkmo an outcrop of weathered granite is overlain by flatlying, rusty mica schist. Somewhat higher up occurs fine-grained dolomite of the type typical of the Storfjell Group. The boundaries are nowhere else to be studied because of thick cover. Granite, more or less foliated, is the main rock of this window with smaller inclusions of amphibolite and mica-rich bands. Pegmatitic, pink veins of the same type as in the Dividalen–Skjold windows occur sporadically. Where the granite occasionally is of a more massive, white type with plagioclase as dominating feldspar it may resemble the trondhjemitic dykes occurring within the Caledonian rocks. Most probably the window at Leirbekkmo is of Precambrian age, but it is possible that part of it was intruded or perhaps remobilized during Caledonian times.

The Rombak window and allochthonous Precambrian granite in the Rombak profile.

The Precambrian rocks of the Rombak window are briefly described by Vogt (1942) and will not be much commented on here as the writer knows very little to add to the description by Vogt. The investigations made by the writer are some studies along the road from Treldal to Bjørnfjell from the western boundary of the window and some 15 km eastwards. Most of this area is taken by the Rombak granite (Table II). Vogt compared it with the Revsund granite of Sweden, of Svionian age. On the geological map of Sweden (Magnusson et. al. 1957), however, the granite about Vassijaure (= Rombak granite) has got the same symbol as the Karelian Lina granites.

Fig. 7. Rombak granite, undeformed, normal type, Skjomen. The minerals are microcline, quartz, biotite and sericitized plagioclase (upper left corner) (+n, 35x).
Rombakgranitt, udeformert, normal type, Skjomen. Mineralene er mikroklin, kvarts, biotitt og sericittisert plagioklas (øvre venstre hjørne) (+n, 35x).

The Rombak granite (Fig. 7) is (on the map, 1950) described by Vogt as "microcline granite, coarse-grained, grey." Occasionally the colour may be slightly reddish. The most interesting feature observed by the writer is the sheared appearance of the granite in the vicinity of the western boundary against the Caledonian rocks. As far as can be judged, the foliation attained by the granite is parallel to the boundary and to the schistosity of the deformed schists above. The sheared appearance can be traced some hundred metres eastwards from the boundary, but while it is not easy to say what depth below the boundary this corresponds to, it certainly must be some tens of metres. Farther east the granite becomes massive. It seems obvious that the shearing and foliation are related to the Caledonian movements in the sequence above. It is worthy of note that the basal conglomerate and quartzite of the Cambro-Silurian which, according to Vogt, are present along the northern margin of the window, are absent on the west side, north of Rombaken. The thrusting of the Rombak Group has here taken place along the Precambrian surface and deformed the

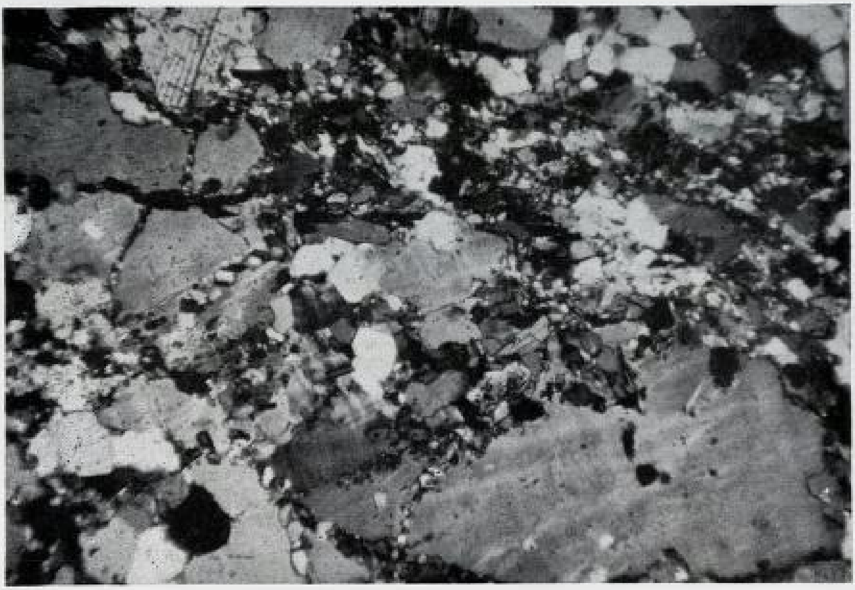


Fig. 8. Tectonized Rombak granite, Beisfjord. Minerals as for the undeformed type + epidote and titanite in small grains (compare Fig. 7) (+n., 35x).
Tektonisert Rombakgranitt, Beisfjord. Mineralene er de samme som i den udeformerte type + epidot og titanitt i små korn (+n., 35x).

upper part of it. A corresponding deformation of the Precambrian granite in the Beisfjord area, south of Narvik, was observed by Kulling (1950) who described the upper part of the Precambrian as fine-striped gneiss ("finstrimmig gnejs"), (Fig. 8).

It appears as if the foliation in the granite has been formed in part by mechanical crushing and rearrangement of minerals, and partly by recrystallization of quartz and greenish biotite. Some further structural aspects of the Rombak window will be discussed in the structural paper.

In connection with the deformation of the Rombak granite some rocks not belonging to the basement in the structural sense, will be described: Along the road Treldal-Bjørnfjell, but still within the schists of the Rombak Group, granitic rocks are met with at several localities. The westernmost outcrop is situated about 3 km east of Treldal (Fig. 9):

The rocks in question are grey, rather coarse-grained granitic gneis-

Fig. 9. Roadcut in Precambrian granite within the Rombak Group, 3 km E. of Treldal.
Skjæring i prekambrisk granitt i Rombak-gruppen. 3 km øst for Treldal.

ses with foliation apparently parallel to the foliation of the adjoining mica schists west of the locality (strike about N 10° E, dip 30° to the W). Some zones are more strongly weathered than others, probably due to cataclastic effects. A certain grooving (lineation) occasionally occurring on foliation planes is also parallel to the predominant lineation of the Caledonian schists of this area. Small quartz veins are rather frequent in the granite gneiss at this locality. The boundaries to the mica schists are not exposed. The next exposure, two or three hundred metres eastwards shows quartz-mica schist with about the same strike and dip as the gneiss. The next outcrop of granite gneiss is found about 1 km E of the former one. The thickness of this gneiss seems to be at least 200 metres: it is resting on folded and wrinkled, somewhat "rotten" mica schist. The boundary dips about 30° to the west. Eastwards to the Precambrian window grey gneisses alternate with strongly deformed dark mica schist but it is not easy to decide whether this alternation is due to repetitions of one layer or to the existence of several slices of gneiss within the schists.

Microscopically the granite gneiss is slightly foliated, in part with mortar textures. Microcline makes up about ¾ of the rock. It occurs mostly in approximately equant grains of 1 to 5 millimetres size. The grain boundaries are full of small curved incisions against the groundmass. Perthite, of the "patch perthite" type, is frequent. Some individuals are crushed.

The groundmass between the microcline grains has a much smaller grain size, 0,01—0,1 mm, with granoblastic to cataclastic textures. It consists of:

Quartz, which is mostly granoblastic, polygonal; most of it probably recrystallized after tectonization.

Biotite in subparallel flakes. Strongly pleochroic with $Z = Y$: dark brown with greenish tinge, X: light yellow. Chloritization is observed in some flakes.

Microcline, in part as crushed fragments of larger individuals.

Ilmenite is a common accessory. To a great extent transformed into a mass of *titanite*.

One of these granite "plates," probably the western one, was noticed by Th. Vogt and interpreted by him as allochthonous Precambrian. This granite is seen on the map to extend southwards from Rombaken some 4 or 5 kilometres and northwards from the fjord about 2 km (Vogt 1950). The interpretation put forward by Vogt is strongly supported by the present investigations: Firstly, the granite gneiss is different in appearance and composition (higher in microcline) from the Caledonian eruptives of the map area. Secondly, with two possible exceptions, eruptive rocks are not known elsewhere from the Rombak Group, and thirdly, the same rock type is known to occur in the Precambrian window, especially as there is a strong resemblance to the tectonized Rombak granite immediately below the Caledonian schists. Thus it seems justified to consider these rocks as thrust slices separated from the Precambrian basement during Caledonian deformation. The granite slices were to some degree foliated and, in part, acquired a linear element; these structures are parallel to the foliation and lineation in the Caledonian meta-sedimentary rocks. Thrusting of Precambrian rocks is a well-known feature within the Torneträsk area on the Swedish side of the border (se f. inst. Kulling, 1930, 1950, 1960 a, b, 1962, 1964). In a recent publication Brown and Wells (1966) described comparable thrust slices in the lower part of the Caledonian sequence from the Vassijaure—Sjangeli area immediately east of the border.

Summary of the Precambrian windows.

The Precambrian age of the described windows is proved by the following facts:

a) The rocks of the different windows show petrographical likenesses both mutually and with the Rombak window, the Precambrian age of which is undoubted (see Vogt 1942, 1950, Holtedahl 1953, p. 139, and Reitan in Holtedahl et.al. 1960, pp. 67–68). Common to them all is f. inst. the prevalence of granitic rocks, in part gneissic granites together with amphibolites (hornblende gabbros) in subordinate amounts. Both in the Skjold and the Rombak windows the basic rocks can be shown to be of two generations.

b) Both microscopical and mesoscopical features indicate that granitization (K-metasomatism) took place at some stage in the development of these rocks, a feature that is absent in the schists immediately above.

c) Foliation in the windows mostly strikes N–S with nearly vertical dips, whereas the schists above are generally flatlying or have low-angle dips.

d) Topographically, all windows, except the Rombak window, have approximately the same positions, a feature that might be expected if they are parts of the Precambrian peneplain. Only smaller disturbances can be recognized in some of the windows (Gustavson, 1963).

The Precambrian rocks are to be considered as mainly autochthonous, simply because there is nothing to indicate otherwise. The relations found west of the Rombak window (and east of it, in the Torneträsk area) with thrust slices of Precambrian rocks, seem not to have the validity of the areas north and northeast of the Rombak area. The discussion of this feature, together with the vertical movements of the Rombak window, is included in the structural paper.

Basal granites and gneisses

Introduction.

The eruptive rocks below the supposed Cambro-Silurian sequence in the Tysfjord area were examined by Foslie (1941) and termed "eruptives of the bottom massif." On the geological map of Norway by Holtedahl and Dons (1953, revised 1960) the rocks are classified as "basal granites." Both terms underline the essential structural feature of the rocks as forming a basement for the Caledonian sequence without taking any decisive standpoint as to their age.

Fig. 10. Normal Tysfjord granite, Æfjord. The minerals are microcline, plagioclase, quartz and biotite (+n., 35x).

Normal Tysfjordgranitt, Æfjord. Mineralene er mikroklin, plagioklas, kvarts og biotitt. (+n., 35x).

A brief description of the Tysfjord granite has been given by Foslie (1941), while other contributions are presented in the papers of Rekstad (1919), Th. Vogt (1922, 1942) and Sverdrup (1962). From areas north of Ofotfjorden information may be gathered from the diaries of Th. Vogt, from his paper 1922 and from the printed discussion following Vogt's lecture on the Narvik–Ofoten area (1942). Of interest also is the discussion by Heier (1960) of the Langøy rocks W. of the map area.

It is not the purpose of this paper to give a thorough account of the basal massifs. However, as it undoubtedly is conducive to the correct understanding of the general structure, some features will be commented on in the following.

The Tysfjord granite.

The main rock-type within the Tysfjord basal massif was described by Foslie (1941) as a "coarse-grained, light grey to pinkish, rather massive granite, but nearly always with a faint lenticular parallel structure, which in some regions may be quite prominent."

Mineralogically (Fig. 10) this granite consists of microcline perthite, quartz, a little plagioclase and accessory hastingsitic amphibole, epidote-orthite, fluorite, biotite, occasionally also garnet, pyroxene, titanite, apatite etc.

In most areas the main granite is surrounded by a more gneissic "border" granite, in part separated by a thin zone of mica schist referred to by Foslie as the "bottom mica schist." This border granite is reddish and finer grained than the central Tysfjord granite. It contains microcline (non-perthitic), plagioclase as discrete grains, quartz and accessory fluorite, epidote-orthite and magnetite, but not hastingsitic amphibole except in occurrences of coarse-grained granite within the border gneiss. The chemistry of the central as well as the border granite is given in table II, columns 2 and 3.

Table II

Chemical compositions, mesonorms, Niggli values, calculated modes[1]) and specific weight of, 1. Rombak granite, 2. Tysfjord granite (normal), 3. Tysfjord granite (fine grained, "border granite"), 4. Sildvik granite.

	1		2		3		4	
	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat.%	Wt. %	Cat. %
SiO_2	72,97	68,3	69,86	65,6	76,81	72,4	75,99	71,6
TiO_2	0,26	0,3	0,45	0,3	0,14	0,1	0,20	0,2
Al_2O_3	13,18	14,6	14,02	15,5	12,30	13,7	12,19	13,6
Fe_2O_3	0,93	0,6	0,76	0,6	0,53	0,4	0,91	0,7
FeO	1,67	1,3	3,01	2,3	0,48	0,4	1,06	0,8
MnO	0,04	0,0	0,07	1,0	0,02	0,0	0,02	0,0
MgO	0,84	1,1	0,38	0,5	0,06	0,0	0,20	0,3
CaO	0,80	0,8	1,79	1,9	0,47	0,5	0,76	0,7
BaO	0,08	0,0	0,13	0,1	—	—	0,04	0,0
Na_2O	3,23	5,9	3,60	6,5	3,46	6,3	3,29	6,0
K_2O	5,68	6,8	5,49	6,6	5,19	6,2	5,14	6,1
H_2O+	0,31	—	0,36	—	0,09	—	0,24	—
P_2O_5	0,05	0,0	0,09	0,0	—	—	0,02	0,0
S	0,03	0,0	0,02	0,0	—	—	0,01	0,0
F	0,10	0,3	—	—	—	—	tr.	0,0
CO_2	—	—	—	—	—	—	0,04	0,0
ZrO_2	—	—	—	—	—	—	0,04	0,0
Sum	100,20	100,0	100,03	100,0	99,58	100,0	100,16	100,0

Mesonorms:				
Quartz	29,5	21,2	34,0	34,1
K-feldspar	30,5	32,2	30,7	29,3
Na-feldspar	29,5	32,5	31,5	30,0
Ca-feldspar	1,0	3,0	2,0	2,5
Ba-feldspar	—	0,5	—	—
Corundum	1,5	—	0,4	0,5
Sum salic.	(92,0)	(89,4)	(98,6)	(96,4)
Hornblende	—	7,5	—	—
Biotite	5,6	1,3	0,5	1,9
Magnetite	0,9	0,9	0,6	1,1
Titanite	0,9	0,9	0,3	0,6
Fluorite	0,6	—	—	—
Sum femic	(8,0)	(10,6)	(1,4)	(3,6)
Niggli values:				
al	42	39½	47½	44
fm	17½	18	5½	12
c	4½	9	3½	5
alk	36	33½	43½	39
si	391	333	502	465
k	0,54	0,50	0,50	0,50
mg	0,37	0,15	0,07	0,16
qz	+ 147	+ 99	+ 228	+ 209
Calculated modes:				
Quartz	31,22	24,75	35,84	Not calculated, mineral comp. not exactly known.
K-feldspar	28,62	29,11	30,07	
Na-feldspar	26,65	29,04	29,27	
Ca-feldspar	2,39	3,04	2,04	
Ba-feldspar	0,15	—	0,35	
Muscovite	1,50	—	0,35	
Biotite	8,20	4,80	0,80	
Hastingsitic amph.	—	7,38	—	
Epidote	—	0,20	—	
Orthite	—	0,20	0,16	
Fluorite	0,26	—	0,06	
Zircon	—	—	—	
Titanite	0,52	0,97	0,07	
Apatite	0,12	0,21	—	
Pyrite	0,06	0,04	—	
Magnetite	0,51	—	0,74	
Ilmenite	—	—	0,18	
Sum	100,20	100,03	99,58	
Specific weight, determined:	2,680	2,693	2,626	2,628

Fig. 11. Fine grained granite gneiss within mica schists, Sitasjavrre, Tysfjord area. The minerals are microcline, plagioclase, quartz, epidote and biotite (+n., 35x).
Finkornet granittgneis i glimmerskifer, Sitasjavrre, Tysfjord.
Mineralene er mikroklin, plagioklas, kvarts, epidot og biotitt (+n., 35x).

[1]) The modes are calculated by S. Foslie. They are given in weight per cent. The rest of calculations are by the present writer. Mesonorms are calculated according to Barth (1959). When this was first printed the writer was not aware of changes in the mesonorm later proposed (Barth, Journ. Geol. 70, 1962). The analyses are previously unpublished. No:s 1—3 completed for S. Foslie, No. 4 for Th. Vogt.

1. Rombak granite (Skjomen granite of Foslie), Kjerringnes, Skjomen. Anal. E. Klüver, 1930.
2. Coarse Tysfjord granite, Lille Hulløy, Tysfjord. Anal. E. Klüver, 1929.
3. Red, fine-grained granite (border granite gneiss), Jernlien, Æfjord. Anal. O. N. Heidenreich, 1916.
4. Fine-grained Sildvik granite, E. of Saltvikelv. Anal. M. Klüver, 1943.

Rocks very much resembling the border gneiss granite are present in the form of layers within the sequences of mica schist above the basal massifs. (Fig. 11.) These gneisses, encountered within the map sheets (1 : 100 000) Tysfjord, Hellemobotn and Linnajavrre, were termed by Foslie, "schistose sills of more or less aplitic granite-gneiss, rich in microcline." Above the gneiss layers follow sequences of mica schist and meta-limestone with amphibolites and serpentines. The parallel arrangement of micas and quartz aggregates is very pronounced

Fig. 12. Boundary between basal gneisses (to the left) and mica schists. W. of Tjelle.
Grensen mellom basalgneis og glimmerskifre. Vest for Tjelle.

within these gneisses; a certain "banding" with alternating stripes of finer and coarser grains has also been observed in some cases. True mylonite textures have, however, not been detected in the thin sections studied by the present writer.

The pegmatites of the Tysfjord granite, of appreciable economic value (quarried since 1907), deserve a special mention. According to Sverdrup (1962 and personal communication) these are structurally of two different types: Those occurring within the main granite are thick and lenticular paralleling the slight foliation, but cross-cutting contacts have been observed. The pegmatites within the border gneisses occur as more elongate bodies with foliated, sugargrained margins and uneven "concretions" of coarse quartz/feldspar material. Most feldspars of the Tysfjord pegmatites seem to be relatively rich in soda and no general difference between the feldspars can be observed within the two types of pegmatite.

When studied in detail the boundary of the Tysfjord granite is found to be concordant. Foliation and schistosity of the overlying schists and gneisses follow this boundary as does the foliation of the border granite. The relations of the western side of the Håfjell synclinorium have been studied by the writer. Red gneisses dip southeast below mica schists and marbles of the synclinorial core. The boundary is sharp and well-defined (see Fig. 12), though a little mica schist is found within the gneiss near to the boundary. Quartz segregations occur along the boundary, mostly within the mica schist. Foliation of the gneiss is defined by an orientation of biotite and amphibole*) (probably hastingsite), and as far as can be judged, the foliation parallels the boundary. Of considerable interest is the finding of gneissic layers, of exactly the same type as the basal gneiss, within the mica schists some hundred metres east of the boundary. The layers are thin, from 0,1 to 1 metre in thickness.

Only when studied on a large scale does the discordant nature of the Tysfjord granite become apparent as noted by Foslie (1941). Northwards from Tysfjord to the Ofoten area the granite reaches successively higher levels in the sedimentary sequence: north of Ofotfjorden its continuation, the Lødingen granite nearly meets the lower limestones of the Salangen Group (Evenes limestones). Vogt (1922) interpreted the boundary in this area as an injection contact. From the observations of the present writer, the field relationships appear to correspond to those described from the area just south of Ofotfjorden.

Basal gneisses on Hinnøy.

Most gneisses of this area are reddish coloured, medium-grained rocks with the mineral assemblage microcline - plagioclase - quartz - biotite - muscovite - epidote with additional small quantities of titanite, ores, apatite, chlorite (after biotite), hornblende, orthite, tourmaline and garnet. For the most part they are microcline-rich and rather homogeneous rocks. Veins, with a mineralogy comparable to the red gneisses, in some areas (Kvæfjord etc.) intrude grey gneisses. Mineralogically these grey gneisses are very similar to the red types. In some thin sections, however, a pronounced saussuritization of plagioclases has been noted; the plagioclase content is also a little higher. The relations of these gneiss variants are much like those found within the Precambrian windows of eastern areas.

The occurrence of a massive monzonitic rock type in the Kvæfjord district some 4 km west of Voktor, is also of considerable interest. The rock in question is present in a small area only, but differs greatly

*) An exception to the rule that amphibole is absent from the border granite.

Fig. 13. Brecciated granite in Caledonian schists. About 5 km W. of Harstad. Photo., O. Jøsang.
Breksjert granitt i kaledonske skifre. Ca. 5 km vest for Harstad. Fot., O. Jøsang.

from the red gneisses in its massive, non-foliated character and dark grey colour. It consists of microcline, plagioclase, amphibole, biotite, garnet, epidote and a little quartz, titanite and apatite. Very conspicuous are the porphyroblastic microclines with string perthites and inclusions of saussuritized plagioclase. The amphibole shows a strong pleochroic scheme—pale green to dark green--and the biotite is dark brown in the Z direction. The rock shows a resemblance to the porphyroblastic monzonitic granulites described by Heier (1960) from the Langøy area, though the latter rocks typically contain pyroxenes. In the present case garnet takes the place of pyroxenes as a ferromagnesian mineral together with amphibole and biotite.

The boundaries of the granite gneiss area on Hinnøy deserve some attention. In the area around Storehornet, Kasfjord, WNW of Harstad, these relations can be studied in good exposures. Bands of mica schist in gneiss and gneiss bands in the mica schist occur near to the boundary. In some localities the granite gneiss is slightly cross-cutting in relation to compositional banding in the biotite schists. In most cases, however, gneiss and mica schists have foliation (schistosity, mica orientation) parallel to each other. It is interesting to note that feldspar porphyroblasts occasionally occur in the biotite schist along the boundary.

In most parts of the Hinnøy area studied the relations described above seem to be valid. Of special interest, however, is the granite belt found within the area of mica schist some 5 or 6 km west of Harstad. The rock is a strongly red-coloured, fine- to mediumgrained microcline granite with a brecciated appearance (Fig. 13). Numerous chlorite veins transsect the rock. The granite is deeply weathered and rotten, the loose material being utilized as road material. Under the microscope the rock is seen to consist mainly of microcline, quartz and chlorite. The transformation of the ferromagnesian minerals into chlorite seems to be complete. The lower boundary of this granite-mylonite is apparently tectonic against mica schist and marble dipping gently below the mylonite. The granite continues south-southwestwards as a broad band above tectonized schists for some 10 kilometres to the Storvann —Steinsåsvann area where it joins up with a larger gneiss area. This granite belt can most probably be considered as a slice of the basal gneiss thrust up into the Cambro-Silurian sequence. The same explanation is proffered for a granite gneiss resting on marble in the Kvæfjord area. More dubious are some gneissic rocks 2—3 km east of Harstad, the boundaries of which are not apparently tectonic and also poorly exposed. Red granite appears to intrude grey mica gneiss in one locality, but it is not known if this mica gneiss belongs to the basal gneisses or to the sequence above.

Interpretations of the basal granites and gneisses.

The problem of the age and origin of rocks and structures within the basal massifs is an old one. Concerning the present area the question was briefly touched by Vogt in his 1922 paper on the stratigraphy and tectonics of the mountain chain. He stated that the Lødingen granite was of Caledonian age as it showed an "injection boundary" against the schists. He also refers to (p. 736) the opinion of Foslie of a central core of Precambrian granite within Caledonian "injection granites" south of Ofotfjorden. In 1942 (pp. 210—212) Vogt stressed the resemblances between the Tysfjord granite and the Rombak granite, suggesting a Precambrian age for both. He explained the Lødingen granite and the border granite of the Tysfjord area as Caledonian intrusives mobilized in deeper levels of the Precambrian area. He also gives this general statement (in translation): "There is much that speaks in favour of the opinion that many, perhaps most of, these granite areas (basal granites in Northern Norway) consist of Precambrian granites, partly with Caledonian, perhaps palingenous, granite injections along the contacts." Foslie, however, in 1941 interpreted all granites of the Tysfjord area as Caledonian. He found proof of this interpretation in the boundary relations (on a large scale) and in the close resemblances between the border granite and the gneiss layers within the overlying mica schists. He appears to have no reservations in his proposing of a similar age for the central Tysfjord granite. This is not in accordance with his original view if this is correctly referred to by Vogt (1922). (The older interpretation by Foslie of the Tysfjord granite as Precambrian seems to be unpublished.)

Kautsky (1946) discussed the relations between the Norwegian coastal area and the front range on the Swedish side of the border. He interpreted the gneiss granite layers as thrust sheets of Precambrian granites which subsequently had been recrystallized or mobilized. Likewise the border gneisses of the coastal area were supposed to be mobilized Precambrian granite, and he suggested that it is impossible to draw any sharp boundary between the true Precambrian and the Caledonian mobilization granites. It is worthy of note that Kautsky thinks this mobilization to be related to a folding episode later than the main thrust movements.

The chronological scheme of events affecting the Langøy area, Vesterålen, as proposed by Heier (1960), was as follows: Firstly, deposition of Caledonian sediments upon a Precambrian basement. Later at least two episodes of metamorphism and orogenic activity occurred, separated by "formation of palingenic granitic magmas and local migration of these magmas to form masses such as the young red granites." Basic, ultrabasic and monzonitic rocks were intruded in the period between the two main phases of regional metamorphism. Subsequent to the later metamorphic phase intrusion of gabbro with accompanying retrograde metamorphism occurred. Regional thrusting from the west towards the east was the latest event in this area. Heier finds that, "The concept that all the metamorphic rocks on Langøy were recrystallized during the Caledonian orogeny is the simplest and most easily conceivable picture of the geological history."

Oftedahl (1964, p. 11) suggests the concordance along the basement contact to be primary, that is, the sediments were deposited on unfolded Precambrian rocks.

Absolute age determinations of the basement rocks within the map area are lacking. Further west, in the Lofoten—Westerålen area, some determinations in recent years gave interesting results (Neumann, 1960): The Re-Os dating on molybdenite from Vatterfjord giving a doubtless Precambrian age is well known. More recent datings from Langøy (not completed when Heier's paper was published) show the following ages:

a) 423 m. y. Pegmatite, Selvågfjord.
b) 430 m. y. Grey gneiss, Jægtbogen.
c) 450 m. y. Mica schist, between Viksfjord and Geirsfjord.
d) 575 m. y. Augen gneiss, between Sandnes and Sandnesodden.
e) 640 m. y. Young red granite, Torset.

A dating on the Svolvær granite, 650 m. y. is comparable to that on young red granite. Unfortunately no determination on the rocks ("banded series") in which the young granite occurs is at hand, except on the pegmatite giving 423 m. y. If the Precambrian age of the young red granite is correct, it seems evident that most of the granulite facies rocks are Precambrian too, because the young granites are frequent within these rocks (Heier, 1960). On the other hand, the determinations on gneiss and mica schist from the amphibolite facies area (b, c, and d) show Caledonian rocks to be present (d is probably indicating a Precambrian age for the augen gneiss). It seems as though the earliest metamorphism and palingenesis of the Langøy area must be Precambrian in age, a possibility that was also taken into consideration by Heier.

It is not possible at present to draw any safe correlations between rocks on Langøy and the present map area. There are for instance no direct counterparts to the unfoliated "young, red granites." However, as the existence of rocks of Caledonian as well as Precambrian age appears to be established by the K-Ar datings from Langøy, it seems justifiable to expect and predict comparable relations within gneisses of the Hinnøy area.

Returning now to the rocks of the Tysfjord area and the question of their age, it is necessary to discuss certain criteria in an endeavour to clarify the problem, the chief ones being:

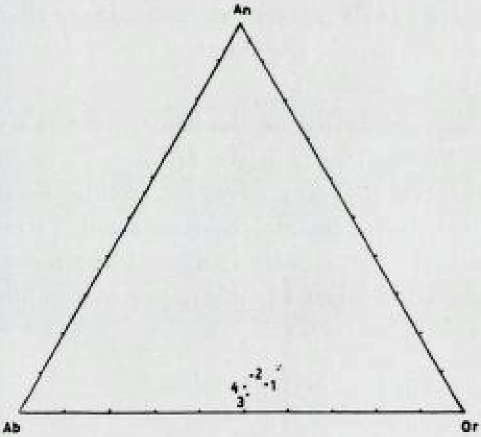


Fig. 14. Plots of normative feldspars of the four rocks from Table II.
Plottinger av normativ feltspat i de fire bergartene i tabell II.

1. Chemistry of the rocks.
2. Mineralogy.
3. Boundary relations.
4. General relations.

1. The chemistry of the two chief rock-types within the Tysfjord basal massif is shown in Table II, columns 2 and 3, together with two analyses of the prevailing granite types of the Rombak window, the Rombak granite and the younger Sildvik granite. As shown by the table, there is a considerable resemblance between the Rombak and the Tysfjord granites, and even more striking is the similarity between the chemistry of the Sildvik granite and that of the border granite of the Tysfjord area. The relations between normative feldspars of the four rocks are nearly identical as seen in Fig. 14. Differences between the granites are mainly in the quartz content. It is also quite clear from Table II that the Sildvik granite and the border granite are considerably lower in the femic constituents than the Rombak and central Tysfjord granite.

In hand specimen the Tysfjord and Rombak granites are very much alike, as emphasized by Th. Vogt (1942, p. 211) and Foslie (1941, p. 43).

2. Mineralogically the Rombak and Tysfjord granites are also very similar to each other, both containing microcline perthite (or microcline) plagioclase and quartz as chief minerals. Common accessories are

biotite, fluorite and epidote. The difference is mainly in the absence of hastingsitic amphibole and orthite from the Rombak granite. This conclusion is, however, based on a few thin sections only and consequently the possibility that this suggested difference is invalid cannot be ruled out. As far as chemical and mineralogical composition and the general appearance of the rocks are concerned, there is thus much that points to a parallelization of the Rombak and main Tysfjord granites and of the Sildvik and Tysfjord border granites too. The two last-mentioned rocks appear, however, to be somewhat different in the relations of the feldspars (perthite in Sildvik granite, microcline in border gr.).

3. The most important concern in discussing the age of the basal granites of the Tysfjord area is undoubtedly the interpretation of the boundary relationships. We have to explain the following main features:

a) The gneissic habit of the border granite.

b) The concordance between foliation structures of the gneiss and the overlying schists.

c) The presence of concordant layers of granite of the basal gneiss type within the lower part of the Caledonian sequence. The layers are present on all scales, from one decimetre to hundreds of metres.

d) The apparent general discordance between the basal massif and the Caledonian sequence when studied on a larger scale.

e) The absence of contact metamorphic phenomena along the boundary.

a) and b): These features may be explained in at least two ways: Either the gneissic structure is a Precambrian parallel texture, flatlying at the time of deposition of the younger sediments (Oftedahl 1964) or it is a Caledonian feature, formed as a consequence of folding and recrystallization during Caledonian deformation. The latter interpretation is the explanation favoured by Th. Vogt and Foslie. If the obviously Caledonian deformation of the Rombak granite along the boundary is brought into mind (p. 23) it certainly becomes the most probable explanation, that is to say that the basal granites of these more westerly areas also were affected by Caledonian deformation.

c) The concordant gneiss-granite layers can be explained either as (1) supracrustal rocks, Caledonian in age, or as (2) thrust sheets of (or with) Precambrian or Caledonian gneisses, or as (3) Caledonian intrusives (including the possibility that they may be remobilized Precambrian granites).

Foslie (1941), who considered both basal granites and gneiss layers to be Caledonian intrusives, rejected the possibility of the gneiss layers being metamorphic sparagmites because of their composition and their resemblance to the basal massif rocks. Foslie's interpretation also explains point d).

Th. Vogt interpreted the border granites as remobilized Precambrian granite, and it seems implicit in this explanation that he also considers the gneiss layers higher up to be mobilized and intruded during the Caledonian orogeny. To the present writer the most critical objection to this explanation is the seemingly complete absence of cross-cutting dykes and connections between the border granite and the gneiss granite layers. But this may be due to mobilization originating in deeper levels (as suggested by Vogt) with the granites intruded as sills from below.

The interpretation of the gneiss layers as thrust sheets of Precambrian granite with basal Cambro-Silurian layers resting on them was suggested by Kautsky (1946). The strongest argument in favour of this theory appears to be the supposed analogy with relations along the front range of the mountain chain on the Swedish side of the border. It is, perhaps, somewhat dubious if such analogies really are to be expected from general considerations; on the other hand, the presence of thrust sheets of Precambrian granites or with such granites at the base has been demonstrated from the Rombak profile and is an argument in favour of the tectonic interpretation of the gneiss layers as is the presence of supposed thrust sheets of gneiss in the Hinnøy area. From what has been said of the textures of the gneiss layers, however, they are not mylonitic. Kautsky thinks that the thrust sheets were mobilized and recrystallized in connection with folding subsequent to thrusting; thus the mylonitic textures could have been destroyed. The present writer finds little evidence of such post-thrusting mobilization and recrystallization in the areas mapped by him (apart from retrograde chloritization), but the possibility of this being the case in the Tysfjord area can not be rejected. It seems, however, difficult to explain the thinner (one decimetre) and apparently extensive gneiss layers as thrust sheets or as occurring at the bases of such sheets.

The hypothesis of the gneiss layers being supracrustals should also be considered. The question can be subdivided into the possibility of their being meta-sedimentary, sparagmitic rocks or recrystallized volcanic rocks. The first possibility was, as mentioned above, rejected by

Foslie, and his reasons seem strong enough. The possibility of these layers being recrystallized volcanic rocks remains to be considered: The extension and concordance of the rocks are arguments in favour of this explanation. Another argument is the presence of probably volcanic (agglomerate and tuff) granitic rocks in the lower part of the sequence in the Kvæfjord area (see forthcoming paper on eruptive rocks). By analogy it might be suggested that at least some or part of the "aplitic sills" of Foslie have the same origin, but have been recrystallized such that primary features are obliterated. The present writer will not, however, postulate an extension of this interpretation to the gneiss granite layers without further investigations. For the thick and rather homogeneous border granite this explanation seem not to be very probable.

d) The successively higher position of the basal granites northwards towards Ofotfjorden has already been mentioned. This can well be interpreted as evidence for Caledonian intrusion of the border granite gneiss as suggested by Vogt and Foslie. The features can hardly be explained as a primary one, because of general sedimentation rules. In point of fact, basal type sediments are absent west of the Håfjell synclinorium. The schists and marbles above the boundary are thick sediments quite dissimilar to the thin and rather rapidly alternating sediments one would expect to find deposited on the basement.

A tectonic explanation of the general discordance of the gneiss/mica schist boundary is perhaps possible. It is, however, little of field evidences in favour of such an interpretation. Quartz segregations in the mica schists along the boundary may indicate some tectonization but nothing is known about the size of order of the movements.

e) The absence of contact phenomena is a well known feature where intrusives of the mountain chain are concerned and need not be taken as evidence of a non-intrusive origin for the basal granites. However, if we regard the quartzite above the Tysfjord border granite as a sedimentary quartzite and not as a quartz-rich facies of the granite ("quartzite-gneiss" of Foslie) then the fluorite content of this quartzite (up to 30 % in one instance) might be considered as a bonafide contact phenomenon.

4. As already pointed out by Vogt (1942, p. 211) the general structural relations with the Rombak granite in the east and the basal granites in the west, both underlying an intervening narrow zone of Cale-

donian sediments, together with the close resemblance of the rocks, points to a probable relationship between those two areas.

As briefly mentioned, the border granite in Tysfjord contains microcline, not perthite as in the other rocks. Amphibole is also lacking as compared with its central granite. It is interesting to note in this connection that Tuttle and Bowen (1958) have remarked upon the unstability of amphiboles in the presence of moderate to high amounts of water and conclude that "the absence of amphiboles in the granite pegmatites and the almost universal presence in the perthite-quartz granites indicate that the pegmatites were produced in a water-rich environment." Applied to the present situation this should indicate a more water-rich environment during the crystallization of the border granite than has been the case with the central Tysfjord granite. This might be taken as an argument in favour of a difference in age, but could also be explained as a consequence of enrichment in water during crystallization of the central granite. In the latter case the age difference would be insignificant.

———————

Concluding points from the foregoing discussion are as follows: Chemical, mineralogical and general structural relations point to a close relationship between the Precambrian Rombak granite and the Tysfjord granite. Whether the central Tysfjord granite is true Precambrian or recrystallized during Caledonian time is, however, somewhat conjectural.

The border gneiss granite is more difficult to interpret. It seems obvious that the foliation and general structure is of Caledonian age. A close connection to the gneiss layers within the mica schists seems evident and a common mode of formation seems probable. The hypothesis of Caledonian intrusion as proposed by Vogt and by Foslie explain the close relations between these gneisses mutually and between them and the central granite. It also explains the large-scale discordance between the gneisses and the meta-sedimentary sequences along the western side of the Håfjell synclinorium. But there is no good explanation to the complete lack of crosscutting dykes and veins. The thrust sheet theory of Kautsky explains the great extension of the gneissic layers and points to a connection with eastern areas (also Rombak profile) and with relations on Hinnøy. But in the latter areas the tectonic nature of the gneiss layers is evident, whereas in Tysfjord no direct proof of thrusting has as yet been demonstrated. Further investigations may do so. If thrusting of the gneisses is assumed it seems natural to consider the mica schist and quartzite between the border gneiss and the central granite as Cambro-Silurian basal sediments as proposed by Kautsky. The relations on Hinnøy indicate that the socalled basal gneisses in part are thrust above mica schists, in part they are of intrusive nature with crosscutting boundaries and feldspathization of the adjoining mica schist (p. 34). It seems probable to the present writer that comparable relations will in the future be detected in the Tysfjord area. Concerning the general features they are perhaps best explained by assuming complete recrystallization in connection with thrust movements. Occurrences of some thin layers of gneiss granite may, however, be better interpreted as sill intrusions. The possibility of volcanic rocks being present among the gneisses should not, either, be completely ruled out till further investigations have been carried out.

STRATIGRAPHY AND PETROGRAPHY OF THE META-SEDIMENTARY ROCKS

Introduction

A tectonostratigraphy of the map area is presented in Table III. Thrust planes (inferred or proved, see structural paper) are marked by small arrows. The sequences constituting the different groups are supposed mostly to reflect original stratigraphical relations, though local differential movements within the formations and along formation boundaries certainly have taken place. The mutual stratigraphical relations of the main tectonic units and their relations to the stratigraphical systems (Eocambrian, Cambrian, Ordovician etc.) are discussed in the structural paper.

In establishing the stratigraphical columns (Table III) the tables published by Foslie (1941, 1949) and Vogt (1942) from the Tysfjord, Håfjell and Ofoten areas, respectively, have been of great value. Some changes have been introduced for these areas, however, in the light of new evidence and modern rules of nomenclature. These changes will be commented on in the relevant passages. Description of stratigraphy and petrography in the areas previously mapped will be restricted to formations which have not been or are incompletely described. For the rest of the region the following scheme of description will be adopted: Short regional descriptions of the stratigraphical groups and their most important formations are given. Within each group the most significant units or rock types are given a petrographical description. Mica schists in different stratigraphical levels are frequently described together. Not all petrographical types discussed are of stratigraphical significance. Particular attention is, however, paid to formations of stratigraphical importance together with rocks which are of special interest in the understanding of the metamorphic or structural history.

Autochthonous sediments

The autochthonous sequence along the mountain front range from Torneträsk to Nordreisa.

The rocks in question belong to the so-called Hyolithes Zone (Hyolithus Series of Svenonius) of supposed Lower Cambrian age. These

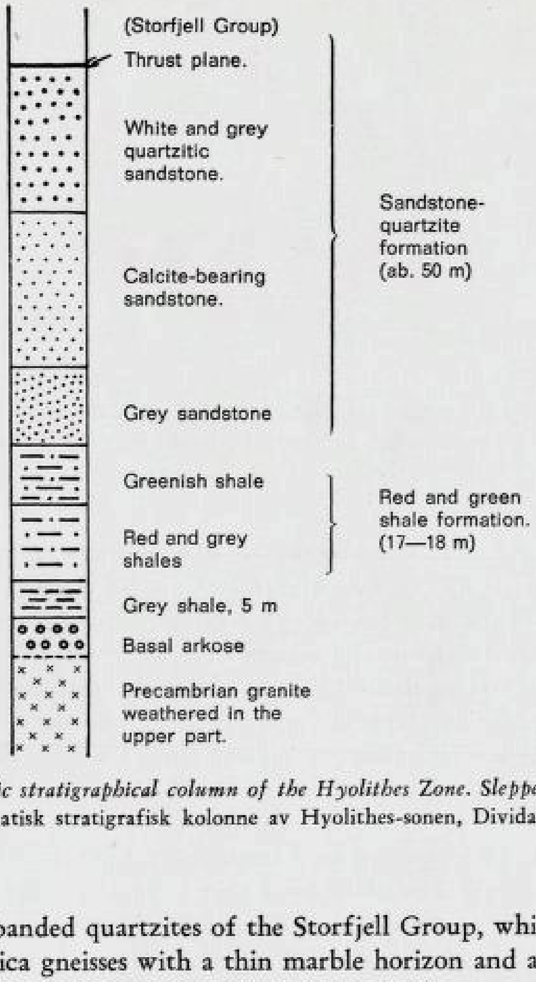


Fig. 15. Schematic stratigraphical column of the Hyolithes Zone. Sleppelva, Dividalen.
Skjematisk stratigrafisk kolonne av Hyolithes-sonen, Dividalen.

found below banded quartzites of the Storfjell Group, which are again overlain by mica gneisses with a thin marble horizon and amphibolites. The autochthonous sandstone and shale probably correspond to the formations B and C in Vogt's scheme. The basal layers are not exposed.

Vogt (1918) describes a locality where only a few metres of basal conglomerate are preserved below the allochthonous rocks. The present writer has not succeeded in finding this locality, the exact position of which, although unknown, is probably in proximity to the southern part of the window.

The relations at the window west of Skjold (northern map boundary) are obscured by thick cover in many places. Unmetamorphosed sandstone, overlain by tectonized banded Storfjell Group schists, is

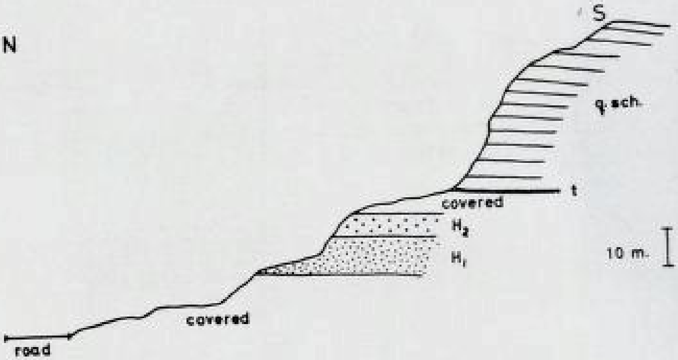


Fig. 16. Hyolithes Zone (H_1 and H_2) and Storfjell Group. W. of Skjold.
Hyolithes-sonen og Storfjell-gruppen vest for Skjold.

present at the bridge west of Skjold, although the lower boundary of this sandstone is unexposed. About 200 metres west of the bridge (a little north of the map boundary) a N–S profile reveals the relations shown in Fig. 16: Some 10 metres of white, quartzitic sandstone is followed upwards by about 5 metres of brownish sandstone with green, shaly layers. Neither the lower nor the upper boundary is exposed, but schists of the Storfjell Group crop out some 5 metres above the sandstone. The total thickness of the Hyolithes Zone at this locality may be in the range 20 to 30 metres. Farther west, north of the mountain Ruten, about 20 metres of grey sandstone follow close above Precambrian granite and is overlain by a thin zone of strongly folded Storfjell Group schists. In the western part of this window the Hyolithes sandstones are lacking and the Storfjell Group rests directly upon the Precambrian basement with a tectonic boundary.

Relations at the Straumsli window have already been briefly mentioned. White and grey massive sandstones possibly belong to the autochthonous sequence; their thickness and extension are certainly very restricted, though any exact measurement cannot be given. The relations seem to correspond to those described from the window west of Skjold with "hard schists" of the Storfjell Group above and Precambrian rocks below. As the boundaries are nowhere exposed it cannot be proved beyond doubt that the sandstones are strictly autochthonous.

Petrography of the autochthonous sediments.

Although the petrography of the sediments within the Dividalen area has already been dealt with in an earlier paper (Gustavson, 1963), it is necessary here to supplement these data with some further observations and comments.

The basal arkose was shown (1963) to consist of the same minerals as the Precambrian granite, but in different proportions. This fact, together with the textural relations (angular mineral fragments etc.), was taken as an indication of an in situ formation of the basal arkose.

Sandstones are the most prominent rocks within the Zone. They are mostly grey-coloured rocks with quartz as the chief mineral. In contrast to the basal arkose the quartz grains are frequently rounded or sub-angular. The clastic character is evident in most cases except in the uppermost layers where recrystallization appears to be related to tectonization features. Feldspar is quite subordinate; when present, it is mainly as a sericitized plagioclase feldspar. Sericite, the most frequent accessory, constitutes the matrix between the quartz grains. An exception is the limy sandstone in the Sleppelva profile where calcite is abundant in the matrix. Ore grains, mostly pyrite (?), are occasionally present. Secondary veins of quartz, in a few cases with calcite, are restricted to the uppermost horizons, but even there they are scarce.

Shales of red, green and grey colours are characteristic members of the Hyolithes Zone. Those of the Dividalen area show transitions into sandstones, while many of them might possibly be classified as siltstones. Sedimentary banding is seldom very pronounced. The boundary between the red and green varieties is characteristically irregular, often with tongues of green shale protruding into the red types. The colour differences were found (1963) to be due to a disparity in the oxidation state of the iron of the two rocks (see Pettijohn, 1957, pp. 346—347). Trivalent iron dominates in the red shale and is present as "diffuse" grains of hematite. Occasionally hematite is also observed as lenses, one decimetre in length in the red shale. In the green shales hematite is absent, though iron is present in chlorite which is one of the chief minerals in this rock. It might be inferred that iron is present mainly in the divalent state (see f. inst. Winchell & Winchell, 1957, on chlorites). Green veins in the red shale consist of chlorite, quartz, titanite and occasionally calcite. As titanite and calcite are not found in the surrounding sediment, it seems justifiable to consider the material as derived from some outside source. The frequency of the veins is undoubtedly related to the tectonization features observed in parts of the Dividalen area. Whereas these green-coloured veins are to be considered as secondary, essentially metamorphic in origin, the age of the colour changes in the various shales is somewhat uncertain.

The chemistry of the normal red shales is illustrated by Table IV; the mineral composition is also expressed in the same table. Comments on the chemical composition are reserved for a later chapter.

For further petrographical data the reader is referred to the 1963 paper.

Table IV

Chemical composition, mesonorm, Niggli values and calculated mode of red, sandy shale[1]) Hyolithes Zone, Dividalen.

	Weight %	Cat. %	Mesonorm		Niggli values	Mode (calculated)	
SiO_2	61,21	60,9	Quartz	36,5	al = 41	Quartz	31,9
TiO_2	0,59	0,5	K-feldspar	23,2	fm = 39	Microcline	2,5
Al_2O_3	17,20	20,1	Na-feldspar	6,5	c = 4	Plagioclase	8,5
Fe_2O_3	3,92	2,9	Ca-feldspar	2,0	alk = 16	Muscovite	41,1
FeO	3,61	3,0	Corundum	11,6	si = 247	(sericite)	
MnO	0,05	0,1	Sum salic	(79,8)	k = 0,83	Chlorite[2])	11,4
MgO	2,52	3,7	Biotite	14,1	mg = 0,38	Hematite	2,9
CaO	0,98	1,0	Magnetite	4,4	qz = +83	Calcite	0,2
Na_2O	0,69	1,3	Titanite	1,5		Titanite[3])	1,5
K_2O	5,04	6,4	Calcite	0,2			100,0
H_2O-	0,28	—	Sum femic	(20,2)			
H_2O+	3,63	—					
CO_2	0,06	0,1					
P_2O_5	0,03	0,0					
Sum	99,81	100,0					

[1]) Specimen 47/62. Analyst P. R. Graff.

[2]) As all Al_2O_3 was needed for muscovite, the chlorite was calculated Al-free (comp. between antigorite and ferroantigorite). The actual composition is unknown.

[3]) Titanite occurs on secondary crack-fillings only.

The Storfjell Group

Introduction.

A general threefold stratigraphical subdivision of the Storfjell Group has been established:

3. Feldspathic quartzites and schists, in part banded, with subordinate quartzite conglomerates.
2. Devdiselv dolomite formation.
1. Quartzites, mostly banded, in part typical hardschists.

The dolomite formation is subordinate when thicknesses are compared, the dominant rocks being those of subgroups 1 and 3. The latter subgroups are variably developed in different parts of the area and where the dolomites are absent, it is difficult to delimit any boundary between them. Nor can boundaries of any regional significance be drawn between different schist types within the subgroups. It has therefore not been practicable to introduce special stratigraphic names for any of the rocks above or below the Devdiselv formation.

The lower and upper boundaries of the Storfjell Group are tectonic, the former against autochthonous sediments or Precambrian basement, the latter against Rombak Group or (in the southeastern area) the Narvik Group.

The thickness of the Group varies in the investigated profiles between some tens of metres and about 300 metres. This may in part be a consequence of primary sedimentation features, but as will be shown in the structural paper, these variations might as easily be attributed to tectonic causes.

Regional description.

The type profile on the northern slope of the mountain Storfjell, Målselv, is shown in Fig. 17. Though the lower parts are to some degree covered by Quaternary deposits, it seems fairly obvious that the Group rests directly on the Precambrian basement in this area. Mylonitic, flinty schists or occasionally a loose-weathering, breccia-like rock type are met with at the base in the terrain south of the Precambrian window. The mylonites are overlain by quartzitic schists which, some few tens of metres above the base, contain thin layers of light grey dolo-

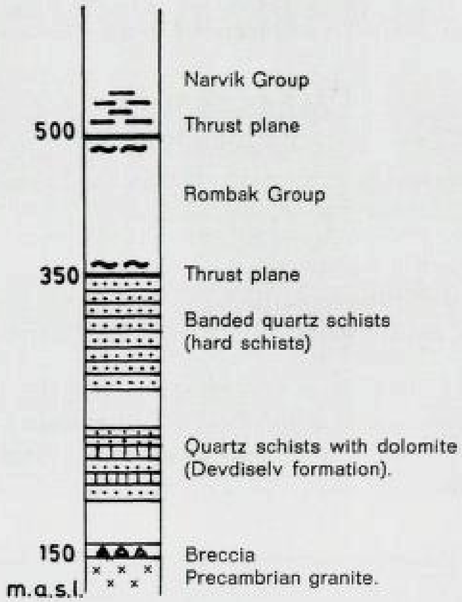


Fig. 17. Schematized profile in the Storfjell Group. Storfjell, Målselv.
Skjematisert profil i Storfjell-gruppen, Storfjell, Målselv.

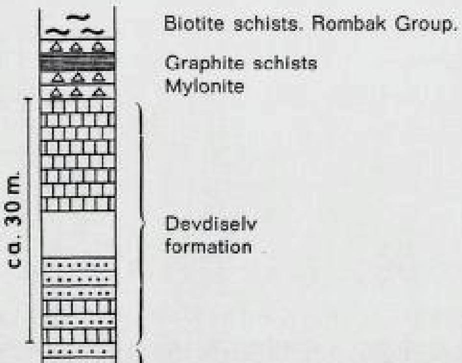


Fig. 18. Schematized profile in the Storfjell Group, Devdiselv, Dividalen.
Skjematisert profil i Storfjell-gruppen, Devdiselv, Dividalen.

mite. Banded, quartz-rich schists follow above. Strong small-scale folding is evident, especially in the lower part of the sequence. The feldspar content of the upper portion of the schists is readily apparent while thin sections from the lower part also contain some feldspar.

Frequently the feldspar, mostly microcline, segregates into thin stripes, mainly in the upper part of the Group.

Eastwards from Storfjell the banded feldspathic schists occur in numerous profiles, an example of which is that of Fig. 16. These schists are partly of the type called "hard schists," in which the banding can, at least to some degree, be ascribed to tectonic causes. However, primary sedimentary banding is also undoubtedly present: this is obvious from profiles in the less tectonized parts of the Group where clastic textures have survived the movements. Dolomite marbles seem to be rare in the area between Storfjell and Dividalen, only small and isolated lenses having been observed.

The dolomite-bearing part of the Group called the Devdiselv formation in this paper, is well exposed along the river Devdiselva, Dividalen. As shown in Fig. 18, the feldspathic schists above the dolomite (Storfjell profile) are lacking at Devdiselv and the uppermost dolomite, about 15 metres in thickness, is overlain by mylonite and metamorphic schists of the Rombak Group. The thin graphite-bearing schist within the mylonitic rocks probably belongs to the Rombak Group too. The upper dolomite is grey, finegrained and transected by numerous, irregular and quartz-filled cracks. The two lower dolomite horizons, 0,1 and 2 metres in thickness, are separated from the uppermost one by some few metres of quartz schist. These dolomites are yellowish to white in colour, finegrained or cryptocrystalline in texture and lack the crackfillings typical of the uppermost horizon. Below there follow some rather monotonous, slightly feldspathic quartz schists. As far as can be judged, the Group is resting on sandstones of the autochthonous sequence. Because of the cover it has not been possible to follow the Devdiselv formation for any distance. Only the quartz schists can be detected in all profiles of the Dividalen area. The thicknesses of these vary considerably from about 30—40 metres to 200—250 metres (west of Høgskarhus). In the area southwards to Altevann it appears from the diaries of Vogt that banded quartz schists frequently crop out above the Hyolithes Zone. These schists undoubtedly belong to the Storfjell Group. On the map, Fig. 1, the Group has been drawn quite schematically in these areas. Dolo-

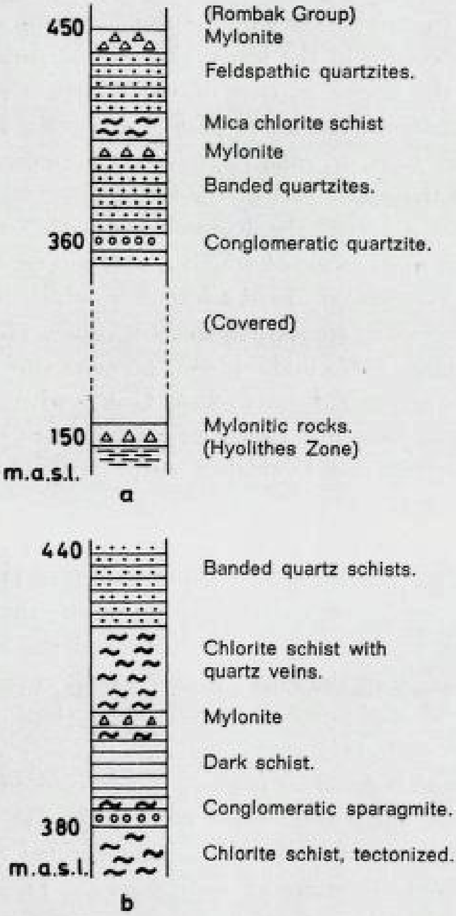


Fig. 19. Schematized profiles in the conglomerate-bearing sequences of the Storfjell Group. (a) W. of Høgskarhus, Dividalen. (b) 5 km S. of Sørmo, Bardu.
Skjematiserte profiler i de konglomeratførende deler av Storfjellgruppen.

mite rocks do not seem to be common south of the Dividalen window.

Conglomeratic layers within the Storfjell Group are encountered in two localities only (Fig. 19, a, b). The first one is met with 360 metres a.s.l. west of Høgskarhus, Dividalen. Over a thickness of some few metres the quartz-rich schist contains what appears to be flattened quartzite boulders and pebbles. There is no marked difference between the schists above and below the conglomerate horizon. The conglomerate has not been detected anywhere else in this part of the map area, and it would seem to be of little stratigraphical importance.

The second locality where a conglomerate has been observed is in Sørdalen, Bardu, a stream section some 5 km S. of the farm Sørmo. As shown in Fig. 19 b, the sequence is different from that west of Høgskarhus. The thickness of the conglomerate horizon is 2—3 metres consisting of small (up to 3 cm in length) pebbles of quartzite in a mediumgrained, grey and sparagmitic groundmass, the larger pebbles having a somewhat flattened appearance. This conglomerate horizon cannot be followed for any distance and like the previous one, is thought to be of local importance only. Elsewhere in the Bardu area the Storfjell Group consists mostly of feldspar-bearing quartz schists, in part banded, generally the same types as found in the eastern areas. Dolomite, for the most part white and fine-grained, occurs in the lower part of the sequence in some localities and is frequently present as lenses in quartz schists. It seems reasonable to think that these dolomite lenses were originally more extensive layers, probably comparable to the Devdiselv formation of the eastern area.

The westernmost outcrop of the Storfjell Group is found in a small area by the Leirbekkmo window (not depicted on the map). The rocks are quartz schists together with a dolomite layer, 0,5 metre thick.

Returning now to the Storfjell area, the Group is there easily traced westwards for several kilometres. Towards the map boundary, however, it disappears below thick and extensive Quaternary terraces. On the adjoining 1 : 100 000 geological map Målselv (Landmark, 1959) a quartzite group is overlying amphibolite, granodiorite and green schists. These last-mentioned rocks occur in continuation of the Precambrian window along the northern boundary of the present writer's map area. According to Landmark's legend, the quartzite group ("Lower quartzite zone") includes sericite-quartzite, microcline-quartzite, quartzite-gneiss, mylonite-gneiss, etc. It appears evident that these rocks, or at least part of them, may correspond to the Storfjell Group.

Petrography.

As noted in the previous description, feldspathic quartzites and related rocks constitute the greatest part of the Storfjell Group. They do however, show a strong variation in their metamorphic development, both in mineralogical and textural relations. In the least deformed types, the clastic character of feldspar and quartz grains may still be recognized in thin section. Generally, however, recrystallization has taken place giving the rocks a pronounced schistosity and obliterating most of the primary textures. The Storfjell Group was metamorphosed under conditions ranging from low greenschist facies to epidote-amphibolite facies (garnet zone), as shown by the facies map, Fig. 2.

In the field many of the rocks have a quartzitic appearance with few visible signs of feldspar content. With a higher mica content they may resemble "normal" mica schists into which they occasionally grade across the strike. However, these mica schists are sometimes feldspar-bearing. In some localities the feldspathic schists have a spotted appearance due to the presence of larger grains of feldspar or quartz in a fine-grained groundmass. Most of the schists show grey or greenish colours, while larger feldspar grains are usually white, rarely red, in colour. Compositional banding is frequently seen in the feldspathic rocks; in most cases this is thought to be of primary sedimentary origin. Other sedimentary features, such as graded bedding, cross-bedding etc., have not been observed. In the more tectonized variants banding (and schistosity) may be even more pronounced.

As mentioned above the feldspathic quartzites may grade upwards or downwards into mica schists. Another feature is that they show transitions across the strike to quartzites of little or no feldspar content. Nothing can be said of the possibility of variations along the strike. Because of cover it has not been possible to follow individual horizons for any distance.

The mineralogical composition of the feldspathic schists varies widely. To illustrate these variations the modal compositions of nine specimens have been determined (600 to 1000 counts in each thin section): these are listed in Table V. For the purpose of comparison, two specimens of sparagmitic rocks from Northern Troms (Padget, 1955), collected by the present writer, have been counted and listed in the same table. Attention is drawn to the total feldspar content which although varying between 6 an 21 per cent, lies mostly in the range 10 to 15 per cent. In most rocks microcline predominates over plagioclase, this feature being apparent also in other thin sections not listed in the table. The mica content varies from 1 to 38 per cent, thus representing both the feldspathic quartzites, nearly free of mica minerals ,and the transitions to the mica schist types. The classification of these rocks from a sedimentological point of view is not discussed in this place, but it is appropriate and timely here to note the resemblance of some of the schists to sparagmitic rock types, for instance those of Northern Troms.

Some microscopic features of rocks from the Storfjell Group are

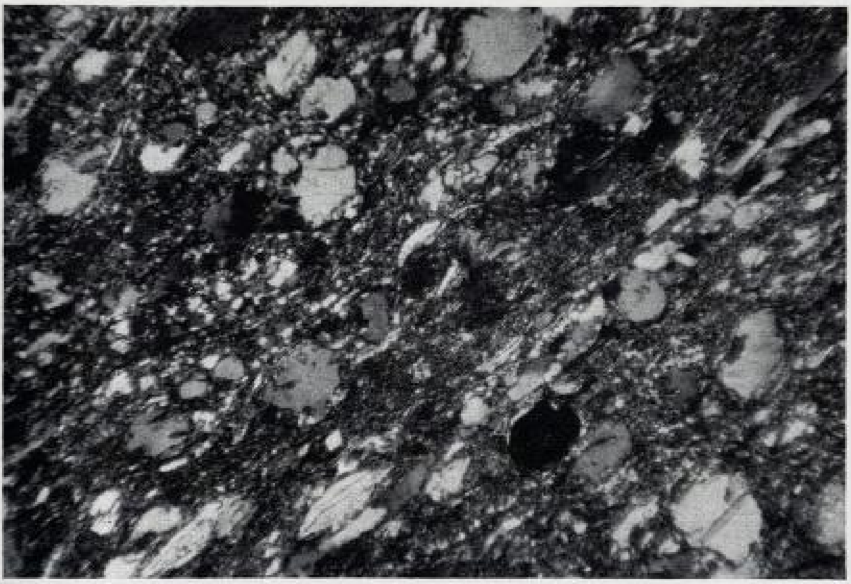


Fig. 20 a.

Sparagmite schist, Evenmo, Målselv. The larger grains are quartz, microcline and plagioclase. Note the lense shape of muscovite flakes in the lower part of the picture and the quartz vein approximately parallel to the schistosity (+n., 35x).

Sparagmittskifer, Evenmo, Målselv. De større korn er kvarts, mikroklin og plagioklas. Legg merke til linseformen på muskovittflak nederst i bildet og kvartsårer tilnærmet parallelt skifrigheten (+n., 35x).

illustrated in Figures 20 a, b, c. Further petrographical data are given below:

Microcline. Most microclines occur as relatively large grains in the range 0,5—5 mm. and only in subordinate amounts in the finegrained groundmass. Both clastic and recrystallized grains have been observed. The clastic microclines are angular to subrounded or rounded in shape. The predominance of ovoid microclines in some tectonized schists indicate that these recrystallized and acquired their present shape during the tectonization processes. Some microcline grains have very irregular boundaries the embayments of which are filled with smaller grains, mostly of quartz. These boundaries can hardly be interpreted as primary and are probably of metamorphic origin. However, as the shapes of the grains show a striking resemblance to those of the clastic variety, it is possible that in this case, only the boundaries have been recrystallized and not the whole feldspar grains. Most of the microcline feld-

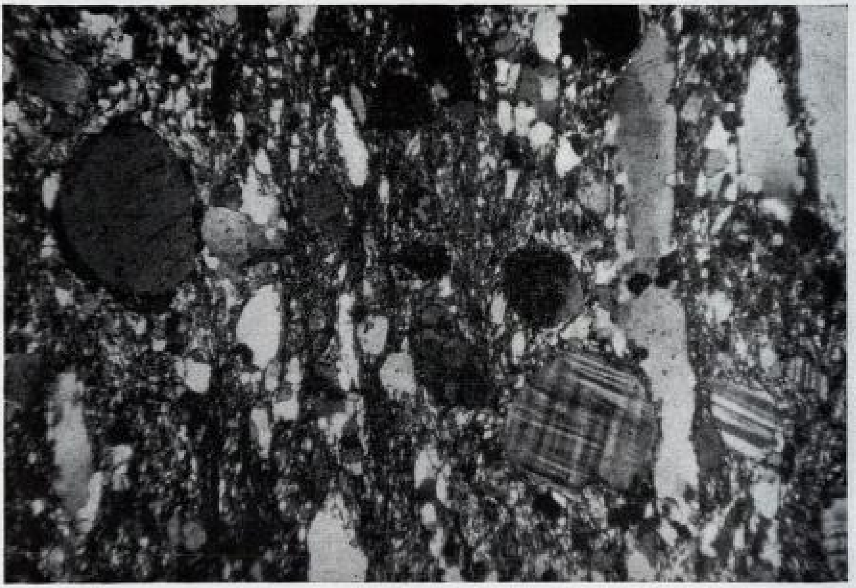


Fig. 20 b.
Sparagmite schist (conglomeratic), Sørdalen, Bardu. Note the well rounded shape of the larger quartz and microcline grains (+n., 35x).
Sparagmittskifer (konglomeratisk), Sørdalen, Bardu. Legg merke til den vel rundete form på de større kvarts- og mikroklinkorn (+n., 35x).

spar are unaltered and have a well-developed twinning lattice. Perthitic microclines have been observed, but they are not common.

Plagioclase. The range of grain sizes is mostly the same as for the microclines. Shapes are often somewhat angular in the clastic grains; good rounding is seldom present. Much of the plagioclase feldspar is probably recrystallized, however. Sericitization is apparent in some cases and occasionally the plagioclase grains exhibit a fine, dusty pigmentation. Twinning according to the albite law is common. The composition appears to be that of albite in most cases as determinations all lie in the range An_0 to An_{10}.

Quartz. In contrast to the feldspars quartz constitutes an important part of the finer groundmass fabric besides occurring as larger grains (0,5—5 mm.). Texturally the quartz is present in the following forms: a) as clastic grains, angular or rounded (only the larger ones), b) as crushed grains with recrystallized quartz along the cracks, c) as granoblastic grains in various aggregate forms, and d) as cross-cutting veins.

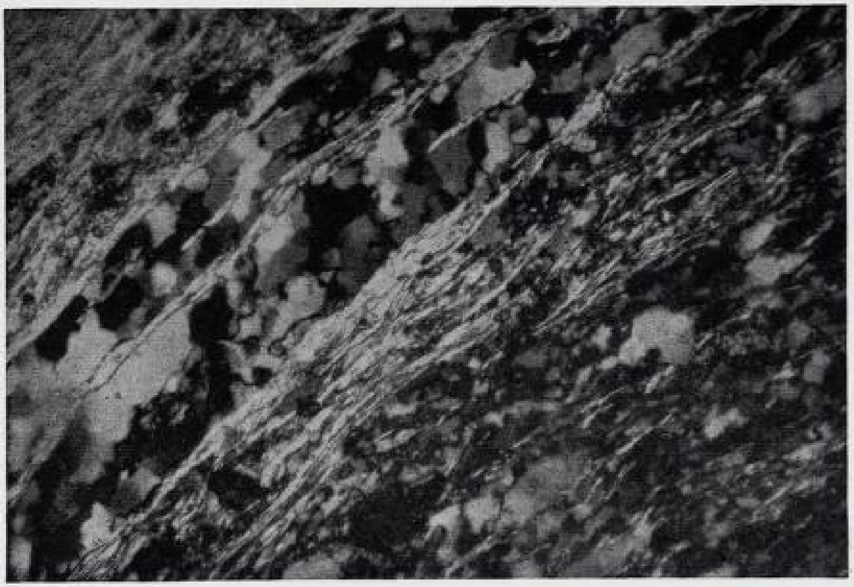


Fig. 20 c.
Quartz-mica schist SW of Skjold, Målselv. The minerals are quartz, muscovite, epidote and a few grains of microcline (not clearly visible on picture) (+n., 100x).
Kvartsglimmerskifer, SV for Skjold, Målselv. Mineralene er kvarts, muskovitt, epidot og noen få mikroklinkorn (ikke synlig på bildet) (+n., 100x).

Muscovite occasionally occurs as large (up to 1 centimetre) wedge-shaped flakes showing undulatory extinction. Inclusions occur in these flakes but are not very frequent. Most of the white micas, however, are small sericite flakes belonging to the fine-grained groundmass. It has not been possible to find any strong indications of the formation of sericite and muscovite from K-feldspar, as has been described from South-Norwegian sparagmites (Barth 1938, Oftedahl 1943). The rocks with clastic microcline are as high in sericite content as any of the types with recrystallized feldspars and textures (see f. inst. nos. 2 and 9 of Table V with 16 and 30 per cent of sericite, respectively).

Biotite occurs in appreciable amounts in many of the feldspathic schists, partly in intergrowth with sericite. The biotite is frequently a greenish variety. Chloritization of the biotite has been observed in some thin sections.

Epidote. In the mica-rich rock types epidote grains are quite common. They are generally only weakly pleochroic and show anomalous (red to yellow) interference colours.

Table V

Modal analyses of feldspathic schists from the Storfjell Group (1–9) compared with two sparagmitic rocks from Northern Troms (10, 11).

	1	2	3	4	5	6	7	8	9	10	11
Quartz	87	69	67	67	61	60	56	51	44	68,5	73
Microcline	11	7	10,5	10	4	8	5	14	3	18,5	9
Plagioclase	+	7	4	+	2	6,5	5	7	7	—	3
Sericite } Muscovite }	1	16	11,5	16	25	8	7	+	30	12	9
Biotite	—	—	3	3	7	15	23	20,5	8	—	3
Epidote	—	—	3	3	—	+	+	5	6	0,5	1
Titanite	—	+	+	+	—	+	+	+	2	0,5	+
Calcite	—	—	+	—	+	+	—	—	—	—	1
Chlorite	+	+	—	—	—	—	—	+	—	—	+
Ores	+	+	+	—	+	+	—	+	+	—	+
Garnet	—	—	—	—	—	+	1	—	—	—	—
Tourmaline	—	—	—	—	+	—	—	—	—	—	—
Zircon	+	—	—	—	—	—	—	—	—	—	—
Sum mica	1	16	14,5	19	32	23	30	20,5	38	12	12
Sum feldspar	11	14	14,5	10	6	14,5	10	21	10	18,5	12
Relation micr.: plag.	∞	1,0	2,6	∞	2,0	1,2	1,0	2,0	0,4	∞	3,0

1. Specimen. 31/62. 360 m.a.s.l. Høgskarhus, Dividalen.
2. — 14/62. 410 m.a.s.l. Storfjell, Målselv.
3. — 88/63. N. of bridge, Straumsørseter, Bardu.
4. — 33/62. 430 m.a.s.l. Høgskarhus, Dividalen.
5. — 202/62. S. of Altevann, Bardu.
6. — 47/63. 6 km S. of Holt, Dividalen.
7. — 4/62. 280 m.a.s.l. Evenmo, Målselv.
8. — 46/63. 6 km S. of Holt, Dividalen.
9. — 13/62. Storfjell, Målselv.
10. — 80/63. 2 km. S. of Helligskogen, Skibotndal.
11. — 81/63. 7 km. S. of Sappen, Reisadal.

Titanite. This occurs as irregular grains, partly with an ilmenite core. Titanite is probably the most frequent accessory mineral.

Garnet. As shown by Fig. 2 a small part of the Group is within the garnet zone. Even there garnet is relatively uncommon occurring sporadically in the more pelitic horizons.

There is little of importance to say about other rock types of the Storfjell Group. Interesting, however, is the preservation of clastic features in the conglomerate in Sørdalen within a highly tectonized sequence (see profile, Fig. 19 b). Brief comments on these features are given in the structural paper.

The dolomite layers or lenses do not appear to be of any particular petrographic interest or significance. The characteristic quartz veins of the Devdiselv dolomite have already been mentioned. Quartz seems to be the chief accessory in most lithologies. In a thin section of a dolomitic marble from the Storfjell profile talc has been observed together with quartz and dolomite.

The Rombak Group

Introduction.

Th. Vogt (1942) introduced the terms "Rombak series" and "Rombak marble" for the rocks in question, the autochthonous basal conglomerate and sandstone being included in the "Rombak series." As "series" is now used for chronostratigraphic units, the term "Rombak series" is therefore not in accordance with modern rules of nomenclature (Henningsmoen 1961 a, b). It is also of course unsatisfactory that allochthonous and autochthonous rocks are included under the same stratigraphical term. Strand (in Strand and Henningsmoen, 1960) used the name "Rombak Group" for the allochthonous Rombak schists and marbles. This usage is adopted in the present paper. As indicated by the presence of Precambrian granite-gneiss in the Rombak profile, the sequence is, however, probably not without tectónical breaks in this area. The importance of these breaks will be discussed in the structural paper. In the areas mapped by the present writer the Rombak Group seems to form a continuous succession and in the following discussion the Group is treated as a structural and stratigraphical unit.

Regional description.

The profile along the road running northeastwards from Treldal, Rombaken, is shown in Fig. 21. The meta-sediments of the Group in this area can be divided into the following four formations:

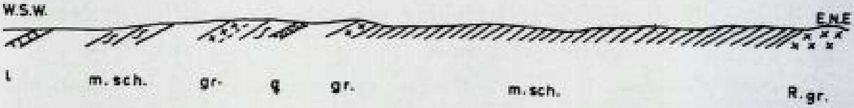


Fig. 21. Simplified profile in the Rombak Group, about 8 km long, from ab. 1 km E. of Treldal to the Rombak granite. (l. = meta-limestone, m.sch. = mica schists, gr. = thrust slices of Rombak granite, q = quartzite.)
Profil i Rombak-gruppen fra 1 km øst for Treldal til Rombaksgranitten.

4. Rombak calcite marble.
3. Upper mica schists.
2. Quartzite, slightly dolomite-bearing with thin layers of calcite marble.
1. Lower mica schists, occasionally graphite-bearing.

Successively higher levels are met with when crossing the sequence from east to west. The lower mica schists are strongly tectonized, fissile rocks, frequently dark-coloured and containing a little graphite. Segregations and veinlets of quartz and calcite are common. In the road profile the mica schist rests directly on the Precambrian basement. Further northeast, however, autochthonous basal conglomerate is present below the Rombak Group according to the map of Vogt (1950). The Precambrian gneiss-granite layers are mostly found within the lower mica schists, mainly in their upper part.

Overlying the mica schists is a quartzite, some tens of metres and possibly as much as one hundred metres in thickness. Folding and cover made it difficult to evaluate the exact thickness. The quartzite is sugary-textured and white on a freshly broken surface; on weathered surfaces it is slightly yellow-coloured. Thin bands (mostly less than 10 centimetres) of coarse-grained, rosy-coloured calcite marble occur within the quartzite. Small lenses of white quartz with chalcopyrite and malachite have also been encountered in one outcrop.

The upper mica schists are in part banded with alternating thin layers of psammitic, semipelitic and pelitic composition (Fig. 22). The uppermost part, below the Rombak marble, consists of monotonous biotite schists, devoid of banding. The boundary against the Rombak marble is not exposed along the road, but observations elsewhere indicate that it is primary and undisturbed by tectonic movements.

The Rombak marble is thick-bedded and medium- to coarse-grained, the colour varying from bed to bed in different shades of grey. Dips show values of about 45 degrees to the west. Thickness probably ex-

Fig. 22. Banded mica schist in the upper part of the Rombak Group. Rombak profile.
Båndet glimmerskifer i den øvre del av Rombakgruppen. Rombak-profilet.

ceeds 100 metres and may be as much as 150 metres inclusive of some mica schist bands in the lower part. The boundary against the overlying Narvik Group is not directly observable in this profile but judging from observations elsewhere (see structural paper) it is tectonic.

In the northeastern part of the 1 : 100 000 geological map Narvik by Vogt (see Fig. 3) somewhat different profiles of the Rombak Group are obtained. The quartzite with thin layers of marble is absent in this area. Instead, above the lower mica schists there occurs a formation of alternating layers of marble and mica schist in approximately the same stratigraphical position. Some of the marbles are dolomitic. An upper mica schist formation corresponding to that of the Rombak profile overlies the marbles. On top of these mica schists follows another formation of alternating marbles and mica schists. According to Vogt's map this formation connects directly with the Rombak marble. There is evidently a splitting up of the marble northwards caused by an increase in the interlayered schistose material: the total thickness of the Rombak marble formation is also increasing northwards. The lowermost of these marbles is bituminous in this northern area.

Approximately the same sequence is present in the areas on both sides of Salangsdalen, mapped by P. R. Lund (1965), though dolomitic marbles are less common in the lowermost marble/mica schist formation. The thickness of the Group is also quite considerable in this part of the map area and eastwards to Sørdalen, Bardu. Along the western side of Sørdalen a quartzite is present; it is thought to belong to the lower parts of the Rombak Group, possibly corresponding to the quartzite of the Rombak profile. In the Neslia area (see map Fig. 1) the sequence above the quartzite is lacking, due to particular tectonic conditions (specified and discussed in the structural paper).

East of Sørdalen the thickness of the Group is considerably reduced. In the Blåberget profile, for instance, it is about 200 metres, consisting of two marbles with a mica schist between them, each marble being 40 to 50 metres thick. The lowermost one is typically pyrite-bearing with brown spots on the weathered surface, while the uppermost marble horizon is severely brecciated with an abundance of schist fragments in the upper part. The same succession is found in a number of profiles in the eastern area resting on feldspathic quartzites of the Storfjell Group (Fig. 23). In some localities, however, there is a mica schist horizon below the lowermost marble belonging to the Rombak Group. It is not possible to decide with certainty whether the two marbles correspond to the two marble-bearing formations on the map Narvik or to one of them. The relations on the Narvik map do, in fact, indicate that lateral facies changes are present, and it would perhaps seem improbable that the marbles are strictly equivalent over such large distances.

Towards the southeastern part of the map area the Group becomes gradually thinner. This is possibly due to primary thickness variations, as shown by the thin marble horizons present in the profiles in the Dividalen area. On the other hand, tectonic nature of the boundaries of the Group also points to the possibility of thrust movements and tectonic squeezing being partly responsible for the thinning in the southeastern areas.

Petrography.

Mica schists.

The schists of the Rombak Group are not especially important in the stratigraphical sense since it is mainly their association with marbles that makes regional mapping of the Group possible. However, some rocks are the only ones within the map area which are dated by fossil findings. Detailed descriptions have been given by Moberg (1908) from the Torneträsk area. Th. Vogt mapped the Zone from Torneträsk to Nordreisa in Northern Troms (outside the map area). Information is presented in a lecture to the Norwegian Geological Society (1918) and in diaries. Investigations by the present writer are restricted to those parts of the Zone encountered in connection with some of the Precambrian windows.

Vogt (1918) subdivided the autochthonous sediments into seven formations ("zones"), which he could follow, from profile to profile, in the area between Torneträsk and Nordreisa. These were, from top to bottom:

G. Alum shale
  3. Thin limestone
F. 2. Green shale.
  1. Thin limestone with phosphorite conglomerate.
E. Sandstones.
D. Red and green shales.
C. Sandstone with layers of shale.
B. Thin green shale.
A. Basal conglomerate and sandstone.

Three of the formations are fossil-bearing (Moberg, 1908, Vogt, 1918). Most important is the D horizon with Platysolenites antiquissimus, Torellella sp. and Hyolithes sp. The two limestone horizons, F_1 and F_3 contain Strenuella.

The alum shale at the top may be absent from some profiles and at Luopakte (Torneträsk) Moberg observed a thrust plane near the base of it, the alum shale thus being parautochthonous in that locality. Generally, however, all the formations A to G are present as one sequence. According to Vogt the basal conglomerate, some few metres in thickness, rests on Precambrian granite along the whole front range. The granite below is frequently weathered and no signs of thrust movements were observed along the boundary. Undoubtedly this boundary is a primary sedimentation feature, and the major parts of the Zone must be strictly autochthonous. As the uppermost horizons are approached, however, folding and small-scale thrust movements become apparent and, according to Vogt, the upper boundary of the Zone is defined by a flat-lying thrust plane against metamorphic rocks. In general, this thrust plane is situated above the alum shale, but occasionally the uppermost part of the Zone has been removed by the thrust movements and the metamorphic rocks are resting on some sandstone or shale horizon from lower levels. The metamorphic schists were described as mylonites and meta-sedimentary rocks (quartzites, dolomitic marbles, black schists etc.).

The thicknesses of two sequences at Luopakte (Luopahta) are given by Moberg as 95 metres and 122 metres, alum shale not included.

According to Vogt the thickness shows a general increase northeastwards, from 113 metres at Torneträsk to 170 metres in Nordreisa (excluding the alum shale). Nothing precise is known about the thickness of the separate formations in this part of the area.

Autochthonous sediments at the Precambrian windows.

Sediments, supposed to belong to the Hyolithes Zone, are encountered at the Dividalen window, the window at Målselva west of Skjold and at the Straumsli window in Bardu.

As far as can be judged from the scarce outcrops there may well be field connection between the Hyolithes Zone along the front range and autochthonous sediments adjacent to the Dividalen window. When studied in detail, however, it is evident that a considerable part of the sequence is lacking in this latter area. A profile along Sleppelva (Fig. 15) shows the presence of formations similar to A to E in Vogt's scheme, these being overlain by tectonized quartz schists belonging to the Storfjell Group (see later). This sequence is valid also for the area immediately north of the window. At Devdiselv for instance, the Hyolithes sandstones (E) are followed upwards by banded feldspathic quartzites with two dolomites, 2 and 15 metres thick, in the upper part. A thin graphitic schist with mylonites below and above, separates the quartzites from overlying biotite schists. Because of the relatively unmetamorphosed appearance of the dolomite/quartzite sequence it was at first thought to be a part of the autochthonous group, but later investigations have shown that it probably belongs to the Storfjell Group, thrust above the autochthonous sediments.

Much less of the Hyolithes Zone is preserved in the southern part of this area. At the brook Kvernbekken, some 4 kilometres south of Sleppelva, only 20 to 25 metres of grey and greenish sandstone and shale are

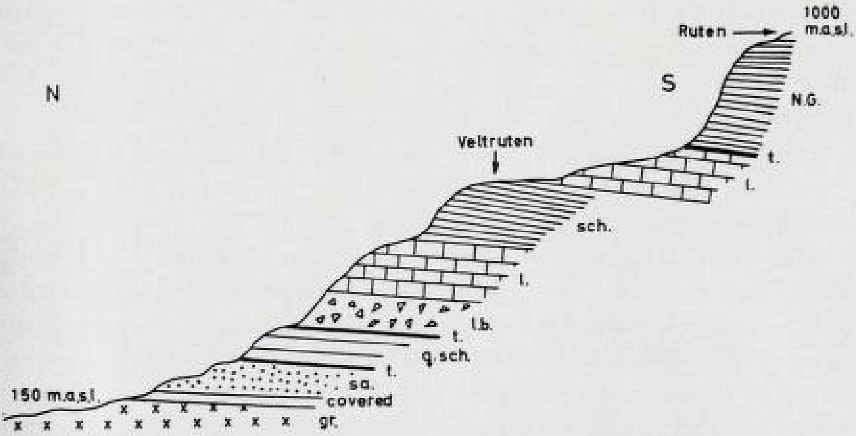


Fig. 23. Section through Hyolithes Zone, Storfjell Group, Rombak Group and the lowest part of the Narvik Group on the northern slopes of Ruten, Målselv. (gr. = Precambrian granite, sa. = Hyolithes sandstone, q.sch. = quartz schist of the Storfjell Group, l. and sch. = meta-limestones and mica schists of the Rombak Group, N.Gr. = Narvik Group, t = thrust planes.)

Snitt gjennom Hyolithes-sonen, Storfjell-gruppen, Rombakgruppen og den laveste del av Narvik-gruppen på nordskråningen av Ruten, Målselv.

differences in metamorphism and mineral assemblage as compared with schists of the overlying groups are apparent. Chlorite minerals are, for instance, more common in the schists of the Rombak Group. Table VI shows the mineral composition of ten mica schists from the Group. As can be seen there is considerable variation. The three rocks of the southern area (1—3) are appreciably higher in lime-bearing minerals than those from northern and eastern areas. This does not mean that schists of the types (4—10) are absent in the Rombak and Skjomen profiles. On the contrary, quartz-muscovite-biotite-chlorite schists are the most frequent type within the whole Group. Lime-silicate-rich schists appear to be restricted to the southern area. As already mentioned in discussion of the Rombak area, compositional banding may be present. For the most part, however, the schists are not banded, but display a well-developed planar schistosity (Fig. 34). Colours are grey to greenish, depending on the chlorite content. Some dark schist types are graphite-bearing, for instance nos. 3 and 6 of Table VI. The graphite content is not restricted to horizons of stratigraphical significance, but can be shown to occur at different levels and with little persistence in the strike direction. Nevertheless, the graphitic schists are of some importance from a stratigraphical point of view, because they are more common in the Rombak Group than in any other group of the mapped area.

Some of the chief minerals of the schists are described below:

Quartz is present as granoblastic grains and frequently in aggregates paralleling the schistosity. The amount is in the range 25 to 50 per cent in most rocks of the Group.

Muscovite flakes in most cases exhibit a marked parallelism while often they are found in parallel intergrowth with chlorite or biotite. In the most tectonized schists lenticular aggregates of muscovite are commonly present which show undulatory or fan-shaped extinction.

Biotite is about equally as common as muscovite, though mostly present in smaller amounts. Generally the biotite is partly transformed into a light green chlorite. The pleochroic scheme is as follows: X: yellow to light brown, Y = Z: brown with reddish or greenish tinge. The greenish brown colour in Z seems to be present more particularly in rocks where chloritization of the biotite is frequent. The reddish brown colours are never very dark. Measurements of refraction index on the red-brown biotites gave n_z = about 1,600, corresponding to about 80 per cent of the phlogopite-eastonite components (Winchell 1957). Pleochroic haloes due to zircon inclusions are seldom met with in these biotites in contrast to their presence in biotites of the overlying Narvik and Salangen Groups; the disparity has not been accounted for.

Chlorite. In some schists of the Group chlorite is an important constituent. It is most common near to the thrust boundaries, frequently containing remnants of biotite. Pseudomorphs of chlorite after garnet have also been observed in some schists from the eastern area. The shape of chlorites is lathlike or fibrous; in most cases they display the same pronounced parallelism as the micas. Chlorite is also found as crack-fillings crossing the schistosity planes together with calcite, quartz, titanite etc.

Albite is a minor constituent in some schists. The grains are small and granoblastic. Porphyroblastic growth of feldspar has not been observed in the Rombak Group. Twinning is mostly according to the albite law.

Microcline has been detected in a few cases, but as a rule it is absent from schists of this group.

Garnet is rare in this group. In the observed cases it is in a state of transformation into chlorite.

Tourmaline has been observed as an accessory mineral in some schists. The colour is greenish, seldom brown. Zoning of the grains is not uncommon; in these cases the central parts of the grains are always darker coloured than the peripheral zones. Pleochroism is distinct in all cases.

Amphibole. A greenish hornblende or actinolite is occasionally present in significant amounts, always with the crystallographic c-axes in the schistosity planes, although no pronounced linear element is developed. Muscovite is notably absent in rocks containing amphibole (see f. inst. Table VI, nos. 1 and 3).

The remaining minerals do not show any features of special interest.

Table VI

Modal analyses of ten mica schists of the Rombak Group.

	1	2	3	4	5	6	7	8	9	10
Quartz	33	49	40	37	51	40	36	27	47	36
Microcline	8	+	—	—	—	—	—	—	—	—
Plagioclase	+	4	+	—	3	—	—	1	2	—
Muscovite	—	—	—	59	32	38	27	48	23	45
Biotite	29	37	22	3	+	4	10	11	26	16
Chlorite	—	—	—	—	14	10	25	9	—	2
Garnet	—	—	—	—	—	—	—	—	+	4
Calcite	—	2	14	—	—	—	—	2	—	—
Epidote	13	8	1	—	—	+	—	+	2	—
Orthite	—	+	—	—	—	—	—	+	—	—
Titanite	—	+	—	—	+	—	+	+	+	—
Amphibole	17	—	21	—	—	—	—	—	—	—
Apatite	—	+	+	—	—	—	—	—	+	—
Tourmaline	—	—	—	—	+	—	—	+	+	—
Zircon	—	+	+	—	—	—	—	—	—	—
Ore minerals	+	+	+	+	+	+	2	2	—	1
Graphite	—	—	2	—	+	8	—	—	—	+

1.	Specimen	103/63.	Sandvik, Skjomen.
2.	—	23/63.	3, 4 km. E. of Treldal, Rombaken.
3.	—	25/63.	7, 5 km. E. of Treldal, Rombaken.
4.	—	441/59.	Å, Lavangen.
5.	—	115/63.	Lundlia, Salangsdalen.
6.	—	100/63.	Sørskogen, main road 6, Bardu.
7.	—	97/63.	Neslia, Bardu.
8.	—	24/61.	Blomlibekken, Bardu.
9.	—	49/63.	6 km. S. of Holt, Dividalen.
10.	—	77/63.	Devdiselv, Dividalen.

Table VII

Chemical composition, mesonorm, Niggli values and mode of mica schist[1]), Rombak Group.

	Weight %	Cat. %
SiO_2	65,74	64,4
TiO_2	0,80	0,6
Al_2O_3	14,59	16,8
Fe_2O_3	0,49	0,4
FeO	6,06	5,0
MnO	0,08	0,1
MgO	2,62	3,8
CaO	0,81	0,8
Na_2O	0,84	1,5
K_2O	4,49	5,8
H_2O-	0,12	—
H_2O+	2,03	—
CO_2	0,67	0,8
P_2O_5	0,06	0,0
Sum	99,42	100,0

Mesonorm	
Quartz	42,5
K-feldspar	15,5
Na-feldspar	7,5
Corundum	9,5
Sum salic	(75,0)
Biotite	21,6
Magnetite	0,6
Ilmenite[2])	1,2
Calcite	1,6
Sum femic	(25,0)

Niggli values	
al	38
fm	41½
c	4
alk	16½
si	292
k	0,77
mg	0,41
qz	+128

Mode (calculated)	
Quartz	41,1
Albite[3])	7,5
Muscovite	28,0
Biotite	14,4
Chlorite	6,8
Magnetite[4])	0,6
Calcite	1,6
Zircon	tr.
Tourmaline	tr.
Sum	100,0

1) Specimen 16/63, E. of Treldal, Rombaken. Analyst: P. R. Graff.
2) Because of Ca-defiency ilmenite was calculated instead of titanite.
3) Calculated as An_0 though compositions up to An_{10} may be present.
4) Some pyrite, FeS_2 is also present, but as the content of S is not determined the pyrite content cannot be calculated. The amounts are, however, small.

As shown by the mineralogical variations illustrated in Table VI, the chemistry of the schists must necessarily also show some variations. However, as already mentioned, the bulk of the schists comprises quartz-muscovite-biotite-chlorite rocks. A specimen typical of these schists has been chemically analysed, the composition of which is shown in Table VII.

Quartzites.

Quartzites are not frequent within the Rombak Group. Occasionally thin layers of sericite-bearing quartzite have been observed as intercalations in the mica schists. More significant are the quartzites described from the Rombak profile and from Bardu. That of the Rombak area is interesting because it contains small amounts of granular dolomite giving the rock a slight yellow colour on weathered surfaces.

Muscovite is an important minor constituent of this rock. The quartzites occurring on the northern side of the Rombak window have not been investigated by the writer and it is not known whether they are similar to the dolomite-bearing quartzite or not.

The quartzite of the western side of Sørdalen, Bardu, and the Neslia area is a "normal" muscovite-bearing quartzite. At Neslia an occurrence of hydrothermal quartz (quarried) is situated close above the quartzite. It is probable that this quartz represents silica mobilized from the quartzite during the metamorphism and tectonization of the area. For further discussion of this matter the reader is referred to the chapter on metamorphism.

Marbles.

The marbles of the Rombak Group do not differ much from marbles of other groups within the map area. They will therefore not be described in any detail. The only feature worthy of special note is the small but quite constant, pyrite content of the uppermost marble in the areas east of Sørdalen. The small brown spots due to weathering of the pyrite have been used (with caution) as a distinguishing feature of this marble during mapping of the eastern area. Graphite is also occasionally present in small amounts in this and other marbles of the Group.

The presence of schist fragments in the uppermost marble (Rombak marble in the southern area) will be further commented on in the structural paper.

The Narvik Group

Introduction.

The term Narvik Group, as used in this paper, was introduced by Strand & Henningsmoen (1960). It comprises the rocks previously called "Narvik schists" and "Narvik Series" (Th. Vogt 1922, 1942); the boundaries are drawn in accordance with those of Vogt (1942) and Foslie (1941, 1949): The lower boundary coincides with the upper boundary of the Rombak Group, that is, the upper limit of the Rombak marble or its equivalents in central parts of the area; in the eastern parts possibly lower formations of the Rombak Group border on the Narvik Group. The tectonic nature of this boundary can be demonstrated in a number of localities (see forthcoming paper on the structural geology of the area). In areas where the Elvenes (Evenskjær) conglomerate is present the upper boundary of the Narvik Group is drawn at the base of the conglomerate. Unfortunately the Elvenes conglomerate has a limited extension, being absent in northern and eastern parts of the map area. In some profiles, mica schist with alternating thin layers of quartzite is possibly an equivalent formation and the boundary has tentatively been placed below this quartzite-bearing horizon. In most cases, however, the boundary has to be drawn at the base of the lowermost marble of the Salangen Group.

Although no exact measurements have been worked out, it is obvious that the thickness of the Group decreases quite considerably from the Narvik area northwards to Salangen; eastwards from Salangen the thickness increases again. Approximately the same formations are resting on the Rombak Group throughout the area and the succession upwards into the Salangen Group seems to be devoid of any notable tectonic breaks. Most probably therefore, the thickness variations are of primary, sedimentary origin.

Some differences in the general sequence of the Narvik Group between the areas south and west of Salangen and the eastern areas are, however, apparent. The characteristic graphite-bearing upper mica schists of southern areas (Ballangen graphite schists) are absent in the east and partly replaced by calcite-bearing schists. Likewise, the sedimentary iron ore horizon (Sjåfjell iron ore) is absent in the eastern areas.

On the other hand, the sequence of the eastern area contains formations different from those found in the Narvik area. Most characteristic are the keratophyres (?) in the lower part of the sequence and the Seternes schist, the latter containing large porphyroblasts of amphibole and possibly corresponding to the Reppi schist of the Tysfjord area (Foslie, 1941).

Despite these differences, the similarities between different parts of the Narvik Group should be kept in mind, especially the dominantly pelitic character of the Group throughout the whole area. The connection in the Salangen area between the eastern and southern development of the Group also appears to be well established (see forthcoming paper on structures). A generalized section through the Narvik Group, valid within most of the area, is tabulated below:

7. Mica schists, in part calcareous in the eastern area, graphitic in the Ofoten area.
6. Sjåfjell iron ore (absent in the eastern area).
5. Mica schists and gneisses.
4. Marble (thin, partly absent).
3. Mica schists and gneisses, with keratophyric (?) rocks in the Bardu area.
2. Quartzites and quartz schists.
1. Mica schists, in part calcite-bearing, in part with amphibole porphyroblasts.

The formations 2, 4 and 6, subordinate as far as thicknesses are considered, are of importance as marker horizons.

Regional description.

In the area between Skjomen and Lavangen the lowermost part of the Narvik Group consists of monotonous biotite schists. Only in a smaller part of this area, where the Djupvik quartzite is present, it is possible to draw an exact upper limit for the lower schists. Where the quartzite is absent, they grade without sharp boundaries into the schists and gneisses above. In the outliers of the Narvik Group east of Lavangen the lowermost part of the Group is represented by banded calcareous mica schist (Fig. 24) and garben schists. The latter rock-type is present as a layer, some few metres thick only, and is probably overlying the calcareous schist. Apart from the above-mentioned schists only some plagioclase-bearing gneissic rocks are present in these outliers, the rest of the Group evidently having been removed by erosion.

Schist layers containing amphibole porphyroblasts are also encountered within the lower mica schist of the Narvik Group in the Salangen area. From Sjøvegan eastwards to the Brandvoll area the lower mica schists are mostly lacking as thick quartzites rest directly upon the Rombak Group.

Amphibole-bearing schists are again met with in road cuts at Seternes, about 3 kilometres W. of Setermoen. The Seternes schist is a slightly greenish grey, lustrous, garnet-mica schist with stout prisms of dark green amphibole, 1–2 centimetres in their long dimension. The amphiboles seem to be randomly orientated. Lenses of quartz and larger (up to 5 cm.s) amphibole needles are occasionally present. Intercalated in the amphibole-bearing horizon which may be some tens of metres thick

Fig. 24. Banded calcareous mica schist belonging to the lowest part of the Narvik Group, Soløyheia, Lavangen.
Båndet kalk-glimmerskifer i den laveste del av Narvik-gruppen, Soløyheia, Lavangen.

are thin layers of mica schist without porphyroblasts. In the main, this schist type is typical of the lower part of the Narvik Group in most of the eastern area. Variations in colour may certainly occur and calcite is visibly present in some cases. Occasionally the rock is a typical garben schist with all amphibole needles in the schistosity planes.

It must be emphasized that the schist with amphibole porphyroblasts is not, even in the eastern area, a continuous horizon and is in fact absent in several profiles. It is, however, to be considered as a special rock type occurring frequently within a restricted part of the sequence. The parallelization, in Table III, of the lower mica schists of the Narvik Group with the Reppi schist of the Tysfjord area is only tentative.

In a number of localities in the eastern area a thin quartzite horizon has been observed below the mica schists described above. More important are the quartzites overlying the lower mica schists in certain parts of the area. In the Rombak profile, for instance, a considerable thickness of quartz schist is encountered at Kalvikneset, 1 km. W. of Treldal. Because of strong folding and local thrusting the precise thickness cannot be given, but 200—300 metres seems to be a reasonable estimate. This horizon was called the Djupvik quartzite by Th. Vogt after a small settlement on the southern side of Rombaken. The quartz schists (quartzites) are typically banded, schistose and very similar to flagstones in appearance. (J. H. L. Vogt, in a diary, called the rock "helleskifergneis"—flagstone-gneiss). The Djupvik quartzite extends southwards to the area south of Beisfjorden. Northwards it thins rapidly and disappears after a few kilometres. The boundaries against the mica schists above and below are everywhere relatively sharp, but alternations of thin bands of mica schist and quartzite are found near to the boundaries as well as elsewhere within the quartzite.

No quartzites of importance are then met with in the Narvik Group until the Salangen area is reached. The Sjøvegan quartzite, the thickness of which exceeds 200 metres at the maximum, crops out on both sides of the inner fjord and on islands in the outer parts. It forms a large anticlinal structure, the main axis trending parallel with the fjord about the NW—SE direction. The Sjøvegan quartzite is less banded

Fig. 25. Sjøvegan quartzite with pelitic intercalations, Narvik Group. Rotvik, Salangen.

Sjøvegan-kvartsitt med pelittiske lag, Narvik-gruppen. Rotvik, Salangen.

than the Djupvik quartzite, but in certain localities, especially on the southern side of the fjord, pelitic intercalations have been encountered (Fig. 25). Amphibolitic layers are also occasionally present within the quartzite. The boundaries of the quartzite are sharp against mica schists above and below, while their mapping is generally facilitated by conspicuous changes in vegetation. The Sjøvegan quartzite has been traced eastwards along the northern side of Barduelva nearly to the map boundary.

Also in the area north of the western end of L. Altevann and west of the Kirkesdal area in Målselv quartzites of a type similar to the Sjøvegan quartzite occur above schists of the Seternes variety. In their upper parts these quartzites frequently contain keratophyric (?) bands, often together with layers of amphibolite.

The Likkafjell quartzite, several (?) hundred metres thick in the north-eastern part of the map area, seems to have a corresponding position in the lower part of the Narvik Group and is tentatively correlated with the quartzites described above.

Rocks thought to be keratophyres occur mainly in the Bardu area and belong to the sequence immediately above the quartzites. In some cases there is a mica schist horizon between the quartzites and the keratophyres, while in other localities the keratophyric rocks lie directly upon the quartzites and may alternate with quartzite layers in the lower part. Alternation with meta-sediments such as mica schists, quartz schists, and with thin amphibolite layers, is, in fact, a characteristic feature of the keratophyric formation (see Fig. 26). Mixtures of keratophyre and sedimentary material are also present. The formation is especially well developed in the Bardu district, but is found as far west as the Salangen area. The thickness including the interbanded metasediments varies from a few metres to two or three hundred metres. There are also in some areas considerable thicknesses of coarser gneisses which are possibly formed by recrystallization of keratophyric rocks. Such rocks are also encountered in the area west of Salangen (Fig. 27); the original extension of this formation may therefore have been quite considerable. The keratophyres have been named after the mountain Blåberget in Bardu where they are well exposed and fairly typically developed. Characteristic of most profiles in the Blåberget formation is the upward increase in amphibolitic layers, both in number and thickness. In many cases the upper boundary is therefore gradational

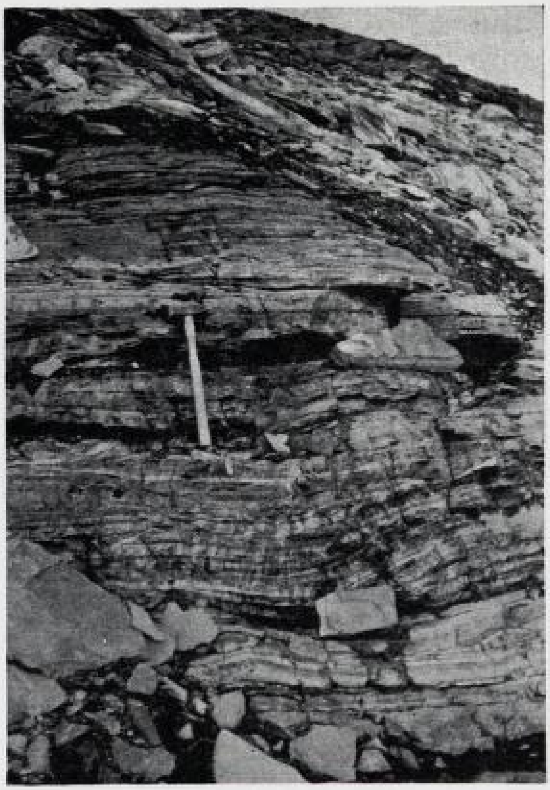


Fig. 26. Keratophyre with intercalated layers of mica schist and amphibolite. Basevardo, Bardu.
Keratofyr med lag av glimmerskifer og amfibolitt. Basevardo, Bardu.

into amphibolites in the eastern areas. In the western areas, keratophyres are present mainly as intercalations in mica schists and mica gneisses.

The rest of the Narvik Group is dominated by mica schists and gneisses which, from a stratigraphical point of view, are of relatively little interest. It has not been possible to separate lithological units mappable over any great distance. An exception is in areas where the thin marble horizon is present and where the mica schists below the marble and those overlying it could be separated. As, however, the schists are generally of the same lithological type above and below the marble, this apparent distinction is of little use. On the map, Fig. 1, they are alotted the same symbol.

It is apposite here to comment on the marble horizon. The Melkedal

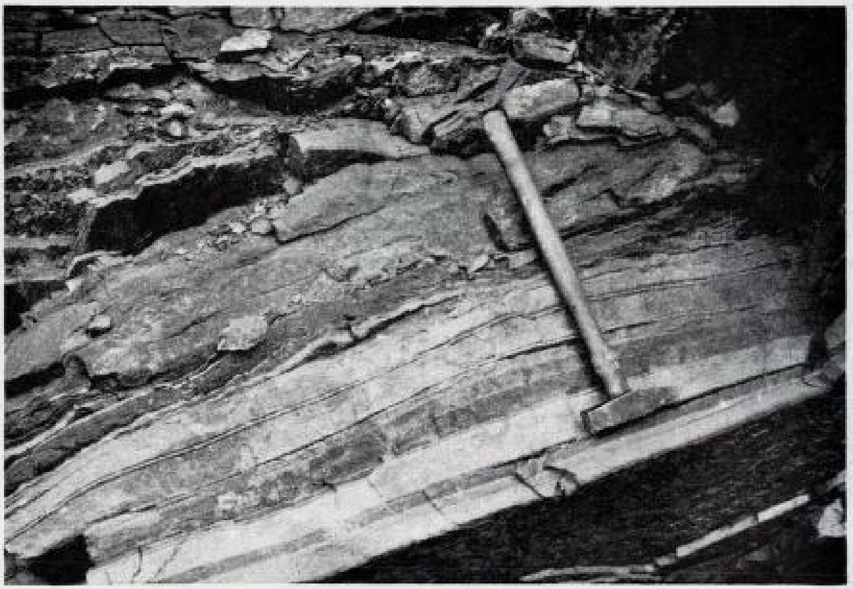


Fig. 27. Alternating layers of plagioclase gneiss and mica schist, Narvik Group, Løkse, Salangen.
Vekslende lag av plagioklasgneis og glimmerskifer, Narvikgruppen. Løkse, Salangen.

limestone, described by Foslie (1941), has an outcrop extension of several tens of kilometres, the thickness varying between 7 and 40 metres. It can be followed northwards to the area just north of Narvik where it wedges out between mica gneisses. Marbles have nowhere been detected further north in this part of the Narvik Group. In the eastern region, however, thin marbles occur mainly in two areas, northwest of Bardu-elva and southwest of L. Altevann (S. of Dittielva). Thicknesses are mostly in the range 5—10 metres, though occasionally higher as a result of folding. Stratigraphically the marbles are placed somewhere in the middle part of the Narvik Group, that is, in approximately the same position as the Melkedal limestone, though of course, reliable correlation is not practicable over such distances and without other marker horizons in the neighbourhood.

The Sjåfjell iron ores of sedimentary origin, are undoubtedly stratigraphically bound to a single niveau in the Håfjell synclinorium. Further to the northeast the ores of the Narvik area were correlated by Th. Vogt with the Sjåfjell horizon. Between the Narvik and Gratangen areas only scattered occurrences of the same ore type are present. There is no evidence that they necessarily belong to the same stratigraphical level as the Sjåfjell ores, although it appears likely that the difference, if any, in position is not great. Ores have not been found in the eastern parts of the Narvik Group.

Petrography.

In this section, the principal schist types of the Narvik Group are described. The keratophyres, together with other igneous rocks of the map area will be treated in a later paper.

Garnet-mica schists.

This is the most common rock type within the Narvik Group. The uniformity of these schists often makes it difficult to judge the extent and influence of folding and therefore the actual thicknesses of the schists.

The typical garnet-mica schist is medium- to dark grey in colour, often with a slightly violet tinge. Veins of white quartz, mostly concordant to sub-concordant, are frequent in many parts of the area. The garnet-mica schists are typically foliated rocks with subparallel mica flakes and quartz aggregates. In most cases the foliation surfaces are somewhat buckled; this is due partly to the deflection of mica flakes around garnet porphyroblasts and partly to small-scale folding. Occasionally, compositional banding of supposed sedimentary origin is seen parallelling the foliation, but for the most part the schists are devoid of such banding.

Mineralogically the schists in question are composed of quartz, biotite, muscovite and garnet in varying proportions. Common accessories are plagioclase, titanite, ore minerals and zircon. Less common are calcite, zoisite and chlorites. In many areas garnet-mica schists grade into gneisses containing plagioclase in considerable amounts. They may also grade into, or be interbanded with, schists devoid of garnet.

Quartz is the dominant mineral in many of these schists, being present in amounts up to 50 per cent by volume. Larger grains or aggregates of grains are frequently lenticular. Undulatory extinction is common but not universal.

Biotite. It is black in hand specimen, in thin section strongly pleochroic, reddish brown to yellow. Haloes around zircon inclusions are generally very strong. In some schists, where late axial planar cleavage is developed, the biotite flakes are seen to follow the earlier foliation (= bedding planes?) whereas muscovites occasionally parallel the cleavage planes. In most cases, however, the parallelism of biotite and muscovite is pronounced. The most common inclusions in biotite are zircon, titanite, quartz and ore grains. The biotite is slightly biaxial, negative. The refringences $n_z \approx n_y$ lie in the range 1,620 to 1,635, values about 1,630 being most common. This indicates a composition intermediate between the phlogopite—eastonite series and the annite—siderophyllite series (Winchell & Winchell, 1957).

Muscovite is common, especially in garnet-rich varieties of these schists. It is mostly present as medium-grained lamellar flakes, fine scaly sericite being rare. The refringence n_z = 1,590—1,600. The optic angle (—)2V is intermediate in value; the acute bisectrix X is nearly normal to 001. From the above data, the muscovite appears to be of a "normal" composition.

Garnets are nearly always present as porphyroblasts, the largest ones about 2 centimetres in diameter. Smaller garnets, those below 5 mm diameter, are generally idioblastic. With increasing size crystal faces tend to be less distinct, the garnets becoming xenoblastic. In some cases there appears to have been two generations of garnet growth as evidenced by differing physical properties and the varying nature of the inclusion fabric from the peripheral to the central parts of any crystal. Occasionally, idioblastic garnets have a rim of irregular garnet containing numerous inclusions. A two-phase development of garnets has also been found by G. Juve (paper in preparation) in the central part of the Håfjell synclinorium. Inclusions of quartz are especially abundant. Chloritization phenomena are uncommon; where present, they are found to be related to the thrust zones. The garnets are mostly of a red-brown colour. Specific gravity was determined for one single specimen to about 4,25, indicating a high content of the almandine molecule. The overall similarity between most garnets within these schists suggests that they are predominantly of almandine composition.

Chemical compositions of the garnet-biotite-schists are indicated by analyses 1 and 2 of Table VIII, both rocks being from the Bjørkåsen area and belonging to the upper part of the Narvik Group. Analysis 3 is a rather special schist type without biotite and garnet but with an uncommonly high muscovite content. No. 4 is high in plagioclase and therefore typical of the rocks described immediately below.

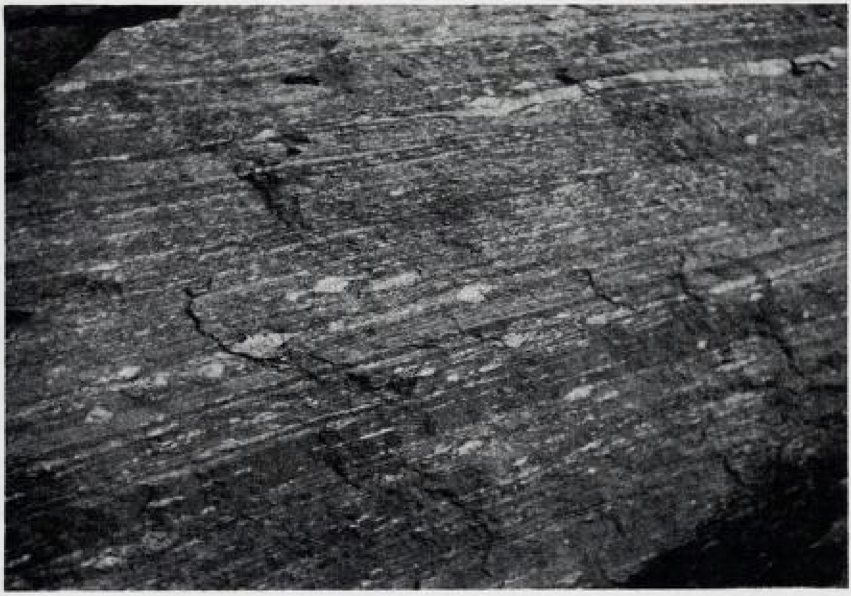


Fig. 28. Gneiss with augens of plagioclase, Narvik Group. Compositional banding is visible. Height of picture about 2 m.

Gneiss med «øyne» av plagioklas og bånd av vekslende sammensetning. Høyden på bildet ca. 2 m.

Plagioclase-mica gneisses, in part kyanite-bearing.

The distinction between "schists" and "gneisses" in this paper is essentially visual, being based on the appearance of the rocks in the field. Accordingly the feldspar content of some schists may be quite considerable (see modes of anal. 1 and 2, Table VIII). As usual, however, with increasing amounts of plagioclase feldspar the gneissic character of the rocks becomes more and more apparent. In general, therefore, rocks termed "schists" very rarely contain more than 12 to 15 per cent by volume of plagioclase feldspar. As previously mentioned, the boundaries between the plagioclase gneisses and garnet-mica schists may in many cases be gradational, but relatively sharp transitions across the strike from gneissic layers to schists poor in feldspar have been observed.

Gneissic rocks, with plagioclase as evenly distributed grains, as augen or in veins, are found in different parts of the map area. All occurrences, however, are situated in those areas with the highest metamorphic grade—almandine-amph. facies—as shown on the map, Fig. 2. Especially abundant are the veined plagioclase gneisses and the augengneisses (Fig. 28). Related to these gneisses, at least in space, are granitic, granodioritic and trondhjemitic lenses, layers and dykes, mostly of relatively small dimensions. The smaller veins, which may be considered as part of the gneiss, consist of quartz, plagioclase feldspar, muscovite, garnet, tourmaline and apatite. Microcline is very sparingly present or totally absent. In a number of cases kyanite is an important constituent mineral of the veins, present as striking light to deep blue needles, the length of which vary from some few millimetres to several centimetres. As a rule, however, the kyanite-bearing veins are poor in, or entirely free of, feldspars, quartz being the chief constituent besides kyanite and micas. These kyanite-bearing quartz veins are not restricted to the plagioclase gneisses, but have also been observed within the ordinary mica schists, although in smaller amounts.

Outside the veins, the mineralogical composition of the gneiss is similar to that of the garnet-mica schists (see Table IX). Plagioclase is, however, present in considerable amounts, while garnets are generally subordinate. Kyanite may be present, but the amounts are small and the crystals almost invisible in hand specimen. Foslie (1913) claimed that kyanite is restricted to the quartz and "granite" veins, a view which seems to be based on megascopical investigations only.

The plagioclase-mica gneisses are easily studied in a multitude of localities, but particularly worthy of note are the new roadcuts at Geisvik (N. of Narvik). The gneiss type encountered in this area is a well-foliated rock with concordant lenses or schlieren, mainly of quartz and kyanite. The quartz-kyanite veins may also contain plagioclase, the amounts of which, though variable, are usually quite subordinate. As already described by Foslie (1913), there is thus no sharp contrasts between the quartz lenses and what he calls "granitic dykes." Common to both types are the accessory minerals muscovite, garnet, tourmaline and apatite, the only notable difference being that of the feldspar content (plagioclase + small amounts of microcline).

In the Geisvik area the kyanite-bearing lenses seem to be confined to certain horizons in the gneiss, these horizons attaining thicknesses of some tens of metres. Similar rocks are met with east of Øsevann, some 15 kilometres to the north, in the approximate strike extension of the gneisses described above. Whether or not these two lithologies are

Table VIII

Chemical analyses, mesonorms, Niggli values and modes, of four mica schists from the Bjørkåsen area, Narvik Group.
(Analyses completed for S. Foslie.)[1]

	1		2		3		4	
	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %	Wt %	Cat %
SiO_2	70,03	68,1	63,28	61,1	56,17	52,7	61,15	56,3
TiO_2	0,75	0,6	0,75	0,6	0,56	0,4	0,78	0,6
Al_2O_3	13,46	15,4	17,74	20,2	19,49	21,6	17,45	19,2
Fe_2O_3	0,56	0,4	0,93	0,6	0,30	0,2	0,71	0,5
FeO	4,64	3,7	5,03	4,1	5,05	3,9	5,52	4,3
MnO	0,07	0,1	0,07	0,1	0,07	0,1	0,08	0,1
MgO	2,17	3,1	2,01	2,9	2,56	3,5	3,20	4,4
CaO	1,35	1,4	1,18	1,2	1,90	1,9	4,41	4,4
BaO	0,05	—	0,07	0,1	0,13	0,1	0,04	0,0
Na_2O	1,07	2,0	0,88	1,6	0,49	0,9	3,56	6,5
K_2O	3,29	4,1	5,21	6,4	6,91	8,2	1,66	2,0
H_2O —	0,08	—	0,10	—	0,24	—	0,09	—
H_2O +	1,38	—	1,73	—	2,96	—	1,05	—
CO_2	0,57	0,7	0,42	0,5	1,55	1,9	0,70	0,9
P_2O_5	0,16	0,1	0,21	0,2	0,09	0,1	0,16	0,1
S	0,15	0,3	0,21	0,4	2,49	4,4	0,40	0,7
CuO	n.d.	—	n.d.	—	0,056	0,1	n.d.	—
Sum	99,78	100,0	99,86	100,0	101,02	100,0	99,96	100,0

Mesonorms:

	1	2	3	4
Quartz	49,2	36,3	24,8	23,2
K-feldspar	9,8	21,3	33,3	—
Na-feldspar	10,0	8,0	4,5	32,5
Ca-feldspar	—	—	—	8,2
Ba-feldspar	—	0,5	0,5	—
Corundum	9,3	12,0	12,3	6,4
Sum salic	(78,3)	(78,1)	(75,4)	(70,3)
Hornblende	—	—	—	7,9
Biotite	17,1	17,1	12,3	16,3
Magnetite	0,6	0,9	0,3	0,8
Pyrite	0,5	0,6	6,3	1,1
Chalcopyrite	—	—	0,4	—
Titanite	1,8	1,8	1,2	1,8
Apatite	0,3	0,5	0,3	0,3
Calcite	1,4	1,0	3,8	1,8
Sum femic	(21,7)	(21,9)	(24,6)	(29,7)

Niggli values:

al	39½	43½	43	34½
fm	38	33½	31	34½
c	7	5½	8	16
alk	15½	17½	18	15
si	348	265	210	203
k	0,67	0,80	0,90	0,23
mg	0,43	0,38	0,46	0,47
qz	+186	+ 95	+ 38	+ 43

Modes (calculated)[2])

Quartz	45,8	33,6	13,5	22,4
Plagioclase	12,7	10,4	4,7	46,5
(% An in plag.)	(21)	(23)	(5)	(30)
Biotite	8,0	11,2	—	11,1
Muscovite	21,9	35,2	70,6	4,3
Chlorite	tr.	—	—	6,7
Garnet	7,5	6,7	—	2,2
Epidote or clinozoisite	tr.	—	tr.	2,1
Pyrrhotite	0,6	0,8	0,2	1,4
Pyrite	—	—	6,2	—
Chalcopyrite	—	—	0,4	—
Ilmenite	1,2	tr.	—	1,2
Rutile	—	0,6	0,4	—
Titanite	tr.	—	—	—
Apatite	0,3	0,5	0,3	0,3
Calcite	1,4	1,0	3,7	1,8

1) Calculations by the present writer.

2) For the sake of simplicity the small amounts of BaO were combined with CaO in calculations, though in actuality Ba is probably substituting for K.

Analysis 1. Material from drilling hole 28, depth 0—58 m., Brattåsen, Ballangen.
— 2. Same drilling hole, depth 76—118 m.
— 3. Mica schist, Bjørkåsen mine.
— 4. Drilling hole 27, depth 17 m., Bjørkåsen mine.

Analyst nos. 1, 2. E. Klüver, 1922.
„ „ 3, 4. M. Klüver, 1933, 1934.

Table IX

Minerals of the plagioclase-mica gneiss and the veins inside the gneiss.[1])

Minerals	In the gneiss	In the veins
Quartz	+	+
Plagioclase	+	+
Microcline	—	(+)
Kyanite	(+)	+
Muscovite	+	+
Biotite	+	—
Garnet	+	+
Apatite	+	+
Tourmaline	+	+
Zircon	+	—
Orthite	+	—
Titanite	+	—
Ores	+	—
Epidote	(+)	—

[1]) +, nearly always present, (+), present in some cases,
—, not detected.

strictly correlatable and belong to the same stratigraphical level is still undecided.

The minerals of the plagioclase-gneisses are described in the same order as presented in Table IX. Unless additional comment is made, the properties also concern the same minerals occurring within the veins.

Quartz occurs either as equant grains with even extinction, probably having crystallized in a later stage of the tectonic history, or as smaller, undulatory extinguished groundmass grains, apparently tectonized. These relations hold for the gneiss outside the veins. Within the veins the grain size of the quartz is larger, frequently exceeding one millimetre; such grains seldom display undulatory extinction.

Plagioclase. The feldspars are clearly porphyroblastic. As regards size, single individuals may measure several millimetres, with aggregates of grains up to 2—3 centimetres in diameter. Inclusions of micas, garnet, quartz, etc. are common, mica flakes being especially abundant in the marginal zones of individual feldspars. Grain boundaries are irregular though the grains themselves are approximately equidimensional. The outer rim frequently has a more soda-rich composition than the core, although the boundary between the core and rim is never very distinct. Twins according to the albite and pericline laws are narrow and frequently wedge out being quite often unsuitable for the determination of composition. A few accurate determinations of refringence gave the compositions shown below:

Specimen 106/63. Core An_{33}. Rim An_{29}.

Specimen 29/63. Core An_{23}. Rim An_{21}.

Specimen 66/62. Core An_{28}. Rim An_{26}.

Incipient sericitization is found in some plagioclases.

Microcline. Only small amounts of K-feldspar with the ordinary microcline twinning is present in the veins.

Kyanite. In hand specimen most kyanites are of bright blue colour, though a greenish blue variety was found in one case. In the road cuts at Geisvik, kyanite lies in the schistosity planes with a rather pronounced, linear element parallelling first generation main fold axes. Under the microscope, the kyanites are seen to be, in part, bent without being fractured or crushed (Fig. 38). The prisms show well-developed cleavages in three directions; twinning on 100 is common. In some cases the kyanites lie within fine, scaly, sericite aggregates, apparently in a state of transformation.

The *micas* and *garnet* do not differ from those described earlier from the garnet-mica schists of the Narvik Group. The same holds for the accessory minerals. It should, however, be noted that the *tourmalines* frequently have a zonal structure with a dark green core and a rim of grass green colour.

The chemical composition of a kyanite-bearing plagioclase gneiss is given in Table X (anal. 1). Comments are reserved for a later section.

Staurolite-bearing mica schist.

Staurolite-bearing rocks have been observed in two small areas. The first of these is in the southern Bardu district, in the high-grade metamorphic area west of Dittielva (see maps, Fig. 1 and 2). The second area is south of Rostadal (northeastern corner of the map) below the Likkafjell quartzite. Both occurrences belong to a relatively low stratigraphical level of the Narvik Group. The staurolite-bearing horizons are in both cases some few tens of metres thick; while their strike extensions are not known, they must evidently be rather restricted. The rocks in question are not dissimilar to the garnet-mica schists. They are greyish coloured schists with crenulated schistosity planes. As well

Table X

Chemical compositions, mesonorms, Niggli values and modes of kyanite-bearing plagioclase gneiss (1), staurolite-bearing schists (2 and 3) and schists with amphibole porphyroblasts (4 and 5) from the Narvik Group.

	1		2		3		4		5	
	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat. %	Wt. %	Cat. %
SiO_2	59,47	56,7	57,80	56,0	57,75	55,9	62,79	60,6	57,69	55,3
TiO_2	1,00	0,7	0,70	0,5	0,61	0,4	0,74	0,5	1,10	0,8
Al_2O_3	19,80	22,2	21,06[1])	24,0	22,11	25,3	16,48	18,8	17,23	19,6
Fe_2O_3	1,88	1,4	2,99	2,2	0,38	0,3	2,13	1,5	2,94	2,1
FeO	6,31	5,0	4,77	3,8	8,05	6,5	5,15	4,2	5,43	4,3
MnO	2,89	4,1	3,74	5,4	3,03	0,1	0,17	0,1	0,13	0,1
MgO	2,89	4,1	3,74	5,4	3,03	4,4	3,40	4,9	5,58	8,1
CaO	1,67	1,7	0,35	0,3	0,24	0,2	2,51	2,6	2,20	2,2
Na_2O	1,86	3,4	0,69	1,3	1,31	2,4	2,05	3,9	1,53	2,9
K_2O	3,68	4,5	4,89	6,1	3,49	4,3	2,15	2,7	3,62	4,5
H_2O-	0,14	—	0,23	—	0,06	—	0,13	—	0,15	—
H_2O+	1,47	—	2,24	—	2,42	—	1,85	—	2,50	—
CO_2	0,06	0,1	0,07	0,1	0,07	0,1	0,11	0,1	0,02	0,0
P_2O_5	0,06	0,1	0,06	0,1	0,06	0,1	0,8	0,1	0,20	0,1
Sum	100,36	100,0	99,81	100,0	99,67	100,0	99,78	100,0	100,32	100,0

Mesonorms:					
Quartz	30,9	33,8	35,8	36,5	29,9
K-feldspar	8,3	17,5	4,2	—	3,5
Na-feldspar	17,0	6,5	12,0	19,5	14,5
Ca-feldspar	3,5	—	—	8,2	6,0
Corundum	12,9	16,6	18,6	8,8	9,8
Sum salic	(72,6)	(74,4)	(70,6)	(73,0)	(63,7)
Hornblende	—	—	—	1,1	—
Biotite	22,7	20,8	27,7	21,6	30,4
Magnetite	2,1	3,3	0,5	2,3	3,2
Titanite	2,1	—	—	1,5	2,4
Ilmenite [2])	—	1,0	0,8	—	—
Apatite	0,3	0,3	0,2	0,3	0,3
Calcite	0,2	0,2	0,2	0,2	—
Sum femic	(27,4)	(25,6)	(29,4)	(27,0)	(36,3)
Niggli values:					
al	40½	43½	46	36	32½
fm	38½	42	41	41	48
c	6½	1½	1	10	7½
alk	14½	13	12	13	12
si	208	202	204	233	183
k	0,57	0,83	0,64	0,40	0,61
mg	0,39	0,47	0,39	0,46	0,55
qz	+ 52	+ 50	+ 56	+ 81	+ 35
Modes (calculated):					
Quartz	22,7	25,3	26,2	33,1	27,8
Plagioclase	24,0	7,5	12,4	23,8	15,2
(% An in plag.)	(29)	(7)	(3)	(24)	(16)
Biotite	12,3[3])	—	7,7[4])	4,0	13,6
Muscovite	21,0	39,2	23,4	15,4	19,6
Hornblende	—	—	—	5,3	5,6
Chlorite	—	tr.	—	9,1[5])	12,3
Garnet	10,7	14,8	14,5	5,0	—
Staurolite	—	7,7	14,5	—	—
Kyanite	7,0	tr.	tr.	—	—
Epidote	tr.	—	—	tr.	tr.
Orthite	tr.	—	—	—	—
Magnetite	2,1	—	0,5	2,3	3,2
Ilmenite	—	—	0,8	—	—
Pyrite	—	tr.	—	—	—
Titanite	—	—	—	1,5	2,4
Tourmaline	tr.	—	—	—	tr.
Apatite	0,3	—	—	0,3	0,3
Calcite	0,2	—	—	0,2	—
Zircon	tr.	tr.	—	—	—

1) Including 0.05 per cent Cr_2O_3.
2) Ilmenite was calculated in nos. 2 and 3 as the amount of CaO was too low to allow the formation of titanite.
3) The relative amounts of muscovite and biotite were estimated from thin sections.
4) The amount of biotite was estimated from thin sections as 7—8 % by volume.
5) The relative amounts of chlorite and garnet were estimated on the basis of thin section examination.

Analysis 1. Sp. 106/63. Lemman, Skjomen.
— 2. Sp. 56/62. Dittitind, Bardu.
— 3. Sp. 43/63. Rostafjell, E. of Dividalen.
— 4. Sp. 474/59. Seternes, Bardu.
— 5. Sp. 204/61. Jormecacca, Bardu.
Analyst for nos. 1—4: P. R. Graff, no. 5, R. Solli.

as stout brown staurolite prisms, up to 1,5 centimetres in length, garnet porphyroblasts are also abundant. Scattered flakes of biotite, up to 5 millimetres in size, are conspicuous in the easternmost area.

Chemical compositions of two specimens of the staurolite-bearing schist are shown in analyses 2 and 3 of Table X, one from each of the two areas. The high alumina and low calcium contents are noteworthy. Further discussion of the chemistry is given in a later section.

Staurolite. The prisms are mostly lying in the schistosity planes, but seem to be randomly orientated within these planes. In thin section, the staurolite prisms are seen to have markedly irregular outlines with numerous inclusions of quartz and some ore grains. Some of the staurolites have a curved shape following small scale folds. A somewhat remarkable feature of these curved staurolites is that their optical orientation is the same throughout the curved prisms. The implications of this observation are discussed in a later section. The staurolites are distinctly pleochroic, the colour ranging from light yellow to golden yellow. The highest refringence $n = 1,745 \pm 0,002$; birefringence is weak. According to Tröger (1959) the refringence would indicate a low content of iron ($Fe^{II} + Fe^{III}$): the examples given by Winchell & Winchell (1957) show no direct relationship between iron content and index of refraction.

Garnets are mostly 3—4 millimetres in diameter. Crystal faces are frequently developed although a certain rounded shape is generally apparent. The arrangement of inclusions, mostly quartz, indicates rotation of the garnets after their formation. In hand specimen the colour is dark brown, in thin section very pale brown. There is no sign of chloritization of the garnets.

Biotite of red-brown colour is present. It is strongly pleochroic with haloes around numerous zircon inclusions. The biotite is partly porphyroblastic with flakes up to 5 millimetres across. Inclusions of ore minerals, muscovite and zircon are common.

Muscovite. Flakes of muscovite, almost without exception, follow the outline of intricate small scale folds. The muscovite is partly in parallel intergrowth with biotite, but occasionally included in the porphyroblasts of biotite.

Quartz. Quartz occurs as fairly equant grains arranged in a more or less granoblastic mosaic. Undulose extinction is usually manifest.

Kyanite occurs as a few small grains within mica aggregates, not visible in hand specimen.

Feldspar. Only a few small grains, probably albite, are present.

Tourmaline is present as small idioblastic crystals of greenish colours. Zonal arrangement of colour is common, the cores of the crystals being darker than the margins. Strong pleochroism in shades of green is always apparent.

Ilmenite with incipient transformation into titanite is present as small grains.

Zircon. Numerous small crystals occur as inclusions within biotite. The haloes are nearly opaque in the strongest absorption direction.

Calcite-bearing mica schists.

Apart from the thin marble in the middle of the Group, calcite-bearing rocks are scarce. Some few occurrences of calcite-bearing schists belonging to the upper part of the Narvik Group are found, especially in the eastern area. Sedimentologically they are to be considered as the initiation of the carbonate sedimentation so richly found in the overlying group.

The banded calcareous mica schists at the base of the Group NE of Lavangen (Fig. 24) are also included in the following description:

Texturally these rocks are not very different from the previously described garnet-biotite schists. Planar schistosity is pronounced. Streaks of calcite (and quartz, in part) may be seen, while frequently calcitic nature of the rock is detected by weathering features. Brownish spots or a more extensive brownish coating may be seen on the surfaces. Mineralogically they are composed of calcite, quartz, biotite, muscovite and in part clinozoisite, as the chief minerals. Common accessories are tourmaline, zircon, titanite and ore minerals; less common are plagioclase, orthite, graphite, garnet and secondary chlorite.

The minerals are here discussed only briefly. Worthy of note is the frequent porphyroblastic development of the *biotite*. The biotite is of the red-brown colour found in most schists of this group. Zircon inclusions with strongly pleochroic (nearly opaque) haloes are common. Strongly undulatory extinction in some of the biotite porphyroblasts indicates deformation younger than, or contemporaneous with the growth of biotites.

Calcite grains are evenly distributed or else segregated into thin streaks concordant to the foliation of the schist. Generally the grains have irregular boundaries, displaying a sutured texture developed during the metamorphic recrystallization of the calcite. *Clinozoisite* is often full of calcite inclusions, and has clearly grown by reaction between calcite and some silicate minerals.

Tourmaline grains are common as accessories in these schists. They are a grass green variety, pleochroic, and in part with a darker green core. Crystals are mostly idiomorphic.

Schists with amphibole porphyroblasts.

The schists in question are mainly of two types: 1. The Seternes schist type with randomly orientated amphiboles and 2. The garben schist type with all porphyroblasts lying in the schistosity planes. There are, however, good reasons for treating them together: Firstly, varieties intermediate between the two main types are encountered in some localities. Secondly, the two types are almost identical in many respects, for instance in chemistry, mineralogy, stratigraphical position and probably also in their metamorphic history. No systematic differences in the areal distribution of the main types related to metamorphism, tectonics etc. have been found. In the following account the features described are valid for both types unless otherwise stated.

The amphibole-bearing schists are grey or greenish in colour, depending on the amount of chlorite present. The groundmass is medium-grained with a well-developed planar schistosity. Occasionally the rock is garnet-bearing with crenulated schistosity planes typical of the garnet-mica schists. The amounts of porphyroblastic amphibole are quite variable: a rough estimate is 5—10 per cent, but locally higher or lower amounts are present. In the Seternes type the amphiboles are frequently seen to grow across the schistosity planes. The size of the porphyroblasts is mostly 2 to 5 centimetres in the longer dimension, the largest ones occurring in the garben schists.

The common mineralogical composition of the schists in question is: Amphibole, garnet, biotite, muscovite, quartz, plagioclase (not always present) and epidote. Common accessories are tourmaline, apatite and ores. Chlorite is frequently present as an alteration product of the other ferromagnesian minerals. Calcite is present in a few rocks, seldom in more than accessory amounts.

As a result of observations under the microscope, the minerals have been divided into three generations: 1. Syntectonic minerals. 2. Post-tectonic (porphyroblastic) minerals. 3. Diaphthoretic minerals. The classification of the porphyroblastic minerals as post-tectonic does not mean that movements have not occurred later than the porphyroblastesis, but that the growth of these minerals is post-tectonic in relation to the strong F_1 folding and associated development of schistosity and is thus also later than the main regional metamorphism (see the section on the metamorphism later in this paper).

First generation (syn-tectonic) minerals.

Quartz of this first generation is found as irregular small grains (less than 0.5 mm) between the mica flakes, as elongate aggregates parallel to the schistosity or as inclusions within the amphibole porphyroblasts. The inclusions are mostly arranged linearly, in part paralleling, in part lying obliquely to the schistosity planes, thus indicating rotation of some porphyroblasts. The quartz commonly has undulatory extinction.

Muscovite is exceedingly common, though occurring in varying amounts, and nearly always showing a strong parallelism of the flakes. Where small-scale folding and crinkling of the schistosity planes is present the muscovite follows the outline of the folds.

Biotite is frequently found in parallel intergrowth with muscovite, but the amounts of biotite are greatest in rocks where muscovite is subordinate. The parallelism of biotites is pronounced, though not to the same degree as that of the muscovite flakes. The strongest interference colour of the biotites ranges from light to dark brown indicating quite notable variations in composition from place to place. Pleochroism is usually strong, especially in the vicinity of inclusions of orthite and zircon. Optic character is always negative with very small axial angle.

Garnet. Small porphyroblasts of 1—3 millimetres diameter are common in some of these schists, but are never present in such amounts as in rocks where amphiboles are absent. Inclusions of quartz, ores and other mineral grains are partly arranged along S-shaped lines thus indicating syntectonic growth and rotation of the garnets. Postcrystalline rotation of garnets with rectilinearly arranged inclusions has also been observed. These observations can be explained either by assuming that garnets are of two generations or that rotation of garnets took place at different times. Garnets have been found as inclusions in amphiboles, whereas the reverse relation is never met with. Chloritization of the garnets is certainly present, although, as a rule, it is not pervasive.

A partial development of crystal faces is seen in a number of cases, but idioblastic crystals are rare.

Epidote-clinozoisite occurs as small, equidimensional grains 0.1—0.2 millimetres) or as elongate, rod-like crystals lying in the schistosity planes. Cores of orthite are present in some cases. The epidote minerals are monoclinic, colourless or slightly greenish and weakly pleochroic. The interference colours vary often within each grain, and are frequently anomalous.

Plagioclase. The feldspars are mostly small and uncommon. Twin lamellae are scarce. The composition was determined for a single specimen (19/63) by measuring the index of refraction: $n'_x = 1{,}540 \pm 0{,}002$ which corresponds to an oligoclase, An_{23}.

Most frequent among the accessories are *ore grains* and *tourmaline.* All the accessory minerals seem to be older than the porphyroblasts of the second generation.

Second generation (post-tectonic) minerals.

Amphibole. The shapes of the porphyroblasts are irregular and variable, ranging from short to long prismatic. Inclusions are numerous; all the groundmass minerals have been found as inclusions though quartz grains are by far the most common. In many cases the schistosity can be traced as lines of inclusions through the porphyroblasts, while in other cases later movements have rotated the amphiboles from their original position. The porphyroblasts are, in part, broken by these movements and chloritized along the cracks. (Figs. 36, 37.) Properties of some of the amphiboles are shown in Table XI.

Table XI

Amphiboles typical of the porphyroblastic schists.

Sample	Colour in Z	$2V_x$ [1])	n_z	n_x	$n_z - n_x$	c/Z
63/62	Light greyish blue-green	80°	1.681	1.662	0.019	21°
2/61	Light blue-green	85°	1.684	1.665	0.019	16.5°
474/59	Greyish blue	n.d.	1.683	1.666	0.017	15.5°
443/59	Blue-green	85°	1.680	1.662	0.018	17°
459/59	Blue-green	80°	1.695	1.675	0.020	20°

Localities of samples:
63/62 Salvasskarfjell, S of Altevann.
2/61 „ „ „
474/59 Seternes, Bardu.
443/59 Soløyheia, Lavangen.
459/59 Otterå, Salangen.

[1]) Only approximately determined.

Table XII

Chemical analysis of amphibole porphyroblasts in mica schist[1]), *Narvik Group.*

	Weight %	Cations	Cations corresponding to 24 O-atoms
SiO_2	41,41	689	6,146
TiO_2	0,48	6	0,054
Al_2O_3	17,08	335	2,988
Fe_2O_3	5,28	66	0,589
FeO	10,37	144	1,284
MnO	0,46	7	0,062
MgO	9,52	236	2,105
CaO	10,65	188 [2])	1,677
Na_2O	2,11	68	0,607
K_2O	0,49	11	0,098
H_2O —	0,03	—	—
H_2O +	1,51	167	1,489
CO_2	0,11	(2)	—
P_2O_5	0,02	0	—
Sum	99,52		

1) Specimen 4/61, Salvasskarfjell, analyst, P. R. Graff.
2) Corrigated for $CaCO_3$.

The chemistry of the amphiboles was determined by the analysis of a typical specimen from Salvasskarfjell (Table XII). The general formula of amphiboles is

$$W_{2-3}\,(XY)_5\,(Z_4O_{11})_2\,(O, OH, F)_2.$$

If we set Z = 8.00, we arrive at the following relations:

$$\begin{aligned} Z &= 8.00 \\ \left.\begin{aligned} Y &= 1.77 \\ X &= 3.45 \end{aligned}\right\} & \; XY = 5.22 \\ W &= 2.38 \\ OH &= 1.49 \end{aligned}$$

and the simplified formula of the amphibole becomes:

$$(Na, K, Ca)_{2,4}(Mg, Fe, Al, Ti, Mn)_{5,2}(Si, Al)_8(O, OH, F)_2O_{22}$$

The ratio Si : Al I (Al substituting for Si) is 6, 15 : 1, 85. The present amphibole is in this and other respects very close to the amphibole no. 22 of Foslie (1945), a femag-hastingsite (Billings, 1928), except for the higher total Al content in the present case. Optical properties of the amphibole (the same as for sp. 2/61 of Table XI) are, however, not in accordance with the assumption that it is a hastingsite if we use the diagram of Foslie (mg against n). In this diagram, it would fall between the tremolite-actinolite series and the "amphiboles from the epidote-amphibolite facies." This would suggest that the alumina not substituting for Si (Al II) has some effect on the index of refraction, or else the diagram is too simplified. According to Tröger (1959) hastingsites also typically have lower axial angles than normal hornblendes. This is not the case with the present amphibole (Table XI). The resemblance in optical characters between amphiboles from different localities within the area indicates that most of them are of approximately the same composition.

Quartz. In some cases the silica of quartz seems to have been remobilized, subsequently crystallizing—as quartz—together with amphiboles in small clots or veinlets. This quartz is coarser grained than the quartz of the first generation. Undulatory extinction is commonly observed.

Albite, composition An_{8-10}, is found in a few cases together with amphibole and quartz.

Third generation (diaphthoretic) minerals.

Chlorite. Some chloritization phenomena have already been mentioned. Chlorite is found as alteration product of amphibole, garnet and biotite and is clearly related to late movements producing small shear planes and mechanical deformation, especially fracturing of the porphyroblasts. This chlorite has lathlike or irregular shape, colours ranging from light green to nearly colourless with low birefringence and with interference colour grey of the 1st order. Index of refraction is about $n_\gamma = 1{,}610$. Optic character is positive. These data indicate that the chlorite is a prochlorite (Winchell & Winchell, 1957).

Ore minerals. In a single thin section ore grains were found within chlorite aggregates formed from broken amphibole porphyroblasts. Their composition was not determined.

Quartz. Some quartz occurs as infillings along cracks in amphibole porphyroblasts and is probably contemporaneous with the chloritization.

The chemistry of two schists with amphibole porphyroblasts is shown in Table X, columns 4 and 5. Attention is drawn to the higher CaO content as compared with the other schists of this table and Table VIII nos. 1—3. The MgO content is also relatively high, especially that of no. 5.

Quartz schists and quartzites.

Attention has already been drawn to the common presence of quartzites and quartzitic rocks in the lower part of the Narvik Group. Because of the discontinuity of field occurrences, local names like Djupvik, Sjøvegan and Likkafjell quartzites have been introduced though these may well be roughly equivalent to each other in stratigraphical position.

A planar schistosity is always well developed in these schists, which are partly banded with alternating layers of darker or lighter colour. In thin section the schists are frequently seen to be feldspar-bearing, microcline and plagioclase both being present. Feldspars are encountered mostly in accessory amounts although specimens containing about 5 per cent feldspar are present in some cases. Microcline is more frequent than plagioclase but rocks with plagioclase as the sole feldspar have been found. Quite considerable variations in mineral composition are apparent in these schists, different thin sections often being representative of only small parts of the formation. In order to give an impression of the mineralogy of the quartz schists some thin sections from different parts of the map area will be described:

Specimen 15/63 from Kalvikneset, Rombaken: Megascopically a fine banding is manifest with darker and lighter laminae from some millimetres to five centimetres in thickness. The thin section is mostly from a dark band and shows quartz and amphibole as the chief minerals together with less amounts of microcline, plagioclase, pyroxene (diopside?) clinozoisite, biotite, titanite and ore grains. The texture is finegrained, typically granoblastic. Some amphibole individuals are poikiloblastically intergrown with quartz grains. The amphibole is pleochroic, slightly brownish green in the Z direction. Though some parallelism of the elongate minerals is seen, it is not especially apparent in this specimen.

Specimen 101/63, main road 6, about 4,5 km N of Brandvoll: Composed of alternating grey and white bands of some millimetres thickness. Small scale folding, with a tendency for thickening of the bands at the fold hinges is observed, though this is never pronounced. Microscopically the lighter bands or laminae are seen to consist of quartz, calcite and microcline with small amounts of biotite, muscovite and ores. The texture is granoblastic with polygonal quartz grains and somewhat larger microclines of ovoid shape. The darker bands are finer grained than the lighter ones and consist of quartz, muscovite, biotite and ore grains.

Specimen 73/61, Iselvdalen, Målselv. This rock was found to be enclosed in the amphibolite massif of Iselvdalen. It is a light grey, flaggy rock of rather homogeneous composition. The chief mineral is quartz. Plagioclase (An_{35}) is present in considerable amounts, whereas microcline is subordinate. Accessories are muscovite, apatite, epidote and rutile. The quartz and plagioclase display remarkable textures, the grain boundaries being very irregular with deep incisions and curved outlines of the grains.

Specimen 41/63, Likkafjell, Målselv. This is a finegrained, tectonized, light grey schist with planar schistosity. Micas occur in thin stripes at several millimetres intervals. Quartz is the chief mineral with muscovite as an important constituent. Accessories are plagioclase, biotite, apatite, titanite and magnetite. Microcline is absent, but occurs in other thin sections from the same area. The texture is granoblastic to cataclastic. Quartz shows strongly undulose extinction with polygonal or irregularly curved boundaries. Parallelism of quartz aggregates and muscovite flakes is pronounced.

The Salangen Group

Introduction.

Rocks of this Group include the Evenes limestones and the Bogen Group (see Strand & Henningsmoen, 1960). The reason for introducing a new stratigraphical term is that over most of the area it is not possible, or practical, to discriminate between the Evenes and Bogen Group lithologies. Both consist of marbles and mica schists in varying proportions; moreover there is no distinct boundary between them.

In a small part of the area a conglomerate occurs at the base of the Salangen Group, but as this already has been discussed, the lower boundary must therefore be drawn at the base of the marbles, above the mica schists of the Narvik Group. The upper boundary below the Niingen Group is drawn above the uppermost marble horizon as marbles are absent within the Niingen Group.

Besides the rocks mentioned above, the Salangen Group comprises meta-sedimentary iron ores, some of which have been mined for short periods or investigated for economic purposes.

Regional description.

For a small area or a single profile the establishing of a stratigraphical succession raises no more problems within the Salangen Group than in any other group of the map area. The sequences of the Håfjell and the Bogen areas, as given by Foslie and Vogt respectively, are shown in Table III. It is to some extent, possible to extend these sequences into the northern continuation of the Håfjell synclinorium up to the Salangen area, though many details are uncertain. Outside the synclinorium precise correlations based on these sequences are nowhere possible. The reason for these difficulties is that the common rocks, the marbles and the mica schists, are very much the same in all parts of the Group. Iron ores are also present at several levels. It is therefore frequently difficult to correlate different profiles with any certainty without following each horizon along the strike. This has not always been possible and the map, Fig. 1, thus invokes a lot of extrapolations and interpolations within this group and should be judged with this in mind.

The middle and upper parts of the Group are not considered in any detail for the reasons given above, but are treated as a marble-mica-schist-iron ore-bearing sub-group of the Salangen Group.

A simplified subdivision of the Group, valid for the greater part of the map area, would be as follows:

Upper and middle parts	Marbles, mica schists and meta-sedimentary iron ores.
Lower part	Quartzite Mica schist Marble(s) Conglomerate, or mica schist with thin quartzite horizons.

This sequence is also, apart from the conglomerate, present in the type area of Salangen.

The conglomerate horizon is only locally developed while other formations of the lower part of the sequence likewise show regional variations.

From regional descriptions by Foslie (1941, 1949) the Elvenes conglomerate is known to occur in the area south of Ofotfjorden extending for some 25 kilometres as a nearly continuous horizon. The conglomerate is situated approximately at the boundary between the lowest marbles of the Håfjell synclinorium (the Ballangen marbles) and the mica schists of the underlying Narvik Group. In some places, however, there is a hornblende schist horizon, up to 30 metres in thickness, between the conglomerate and the marbles. The thickness of the conglomerate is some few tens of metres; according to Foslie (1949) it reaches a maximum thickness of 75 metres in the type area at Elvenes. North of Ofotfjorden a conglomerate is found in a corresponding stratigraphical position at Evenskjær on the western side of the synclinorium. As far as the present writer is aware it has not been observed between Evenskjær and the Harstad area. This may be due to lack of exposures in parts of the intermediate area, but evidently the conglomerate is absent in most of this district. In the town of Harstad, however, outcrops of conglomeratic rocks are again met with. (Gudbrandsen, 1958) Good exposures are, for instance, seen in the park by the Fleischer statue and in adjoining streets (Fig. 29). The boulder material (see petrographical description) is similar to that constituting the Elvenes conglomerate. Interpretation of the structure of this area also indicates a po-

Fig. 29. Conglomerate at the base of the Salangen Group. Harstad.
Konglomerat ved basis av Salangen-gruppen. Harstad.

sition between an underlying mica schist group and an overlying marble formation. A correlation of the Harstad and Elvenes conglomerates is thus strongly supported by the various observations. The conglomerate has also been observed just south of the town. A further extension could possibly be verified by a closer inspection of the Hinnøy area, but with the present state of knowledge it would seem that the conglomerate is restricted to a small area in and around Harstad. On the eastern side of the great synclinorium the conglomerate horizon has not been observed north of Ofotfjorden, except for a single occurrence north of Gratangsbotn (described by Tore Mitsem, thesis Oslo University 1964). In most of this area from Ofotfjorden to Salangen the lowermost part of the Salangen Group (below the lower marble) consists of alternating layers of mica schist and thin quartzite. It is possible that this formation is stratigraphically equivalent to the Elvenes conglomerate. In areas further east the common mica schists of the Narvik Group extend up to the lowermost marble of the Salangen Group.

The marbles immediately above the conglomerate are very thick in the area south of Ofotfjorden. The Ballangen marbles (Foslie 1949) are about 1200 metres in thickness, including subordinate mica schist layers. Dolomite layers and lenses occur in various nivaux, but most of the marbles are meta-limestones. The northern continuation of this formation on the western side of the synclinorium shows similar relations in thickness and composition, though the actual thicknesses are obscured by folding and repetitions in the Tjeldsund area. Further northwest, on Hinnøy, the marbles are less important, though nearly everywhere present at the base of the Group. The average thickness is perhaps a hundred metres. Colour banding is very common in the Evenes—Tjeldsund area and also immediately south of Ofotfjorden (Tjelle area). The variant called Leifset marble with alternating layers of red, yellow, green and white marble is mentioned by J. H. L. Vogt (1897) from the Tjelle area and by Th. Vogt (diaries) from Evenes. Bituminous layers and white dolomite horizons are also present. The present writer has observed comparable relations (colour banding, compositional banding) in the Skånland area.

Along the eastern side of the Håfjell synclinorium the lowermost marbles become less important north of Ofotfjorden. There are from 1 to 3 marble horizons, each with a thickness varying from 10 to some tens of metres. The variations in number and thickness of these marbles are in the opinion of the present writer, best explained as primary, sedimentary changes, even though tectonic disturbances may be important locally, causing repetitions and tectonic squeezing of the marbles. In the easternmost areas where the Salangen Group is represented, it is not possible to say if the lowermost marble is equivalent to those described above, particularly as the important quartzite horizon (Balteskar quartzite) separating the lower and middle parts of the Group is not present in these areas. The explanation of this feature may be that the lower part of the Group is wholly absent in these areas or that only the quartzite is lacking; this latter explanation seems the less probable.

Overlying the lower marbles of the Håfjell synclinorium are garnet-mica schists. In the area from Gratangen to Salangen they are slightly greenish, lustrous rocks, frequently with small (less than 5 millimetres) idioblastic garnets and rather different from garnet-mica schists higher up in the Group. Elsewhere in the area the schists at this level are of a more common type.

The quartzite above is well known from the investigations of Foslie and Th. Vogt—the Bø quartzite of Foslie and the Balteskar quartzite of Vogt. To the present writer's knowledge the latter name was the first one to be used in a publication (Vogt, 1942) and should therefore be preferred. (Balteskar is a farm by Astafjorden at the western side of the great synclinorium.)

The Balteskar quartzite is a typical flaggy rock splitting into slabs approximately an inch in thickness. A considerable sericite content is apparent on all cleavage planes. Considering the outcrop extent of this quartzite, its thickness is fairly constant. According to Foslie (1949) the thickness varies between 25 and 60 metres south of Ofotfjorden. In the area mapped by the present writer the quartzite can be followed along the eastern side of the synclinorial structure to the Salangen area. Thicknesses vary between 20 and 30 metres but decrease to 10–15 metres at Salangen. On the west side of the synclinorium the quartzite was traced by Vogt from Ofotfjorden to Balteskar. It is probably this same horizon that can be seen to cross the island Rolla from south to north. A thin quartzite, too small to be represented on the map, has also been observed on Andørja. There is thus only a relatively short distance over which it is impossible to follow the Balteskar quartzite. It seems highly probable that the horizon was originally continuous from Tysfjord to Salangen prior to the development of the synclinorium. In this respect it is rather unique among the metasedimentary rocks of the area.

The quartzite at Tjeldsundet (Tennevik) is probably the same horizon, but thicknesses appear to be greater in this area.

As already mentioned, the Balteskar quartzite has not been observed in more easterly areas.

The rest of the Salangen Group upwards consist of marbles and mica schists, together with, in the central area (Håfjell–Salangen), iron ore horizons of sedimentary origin. It is not the intention to give a detailed description of this part of the sequence. The marbles are nearly all grey, medium-grained calcite marbles with few, if any, distinguishing features. Dolomitic layers are not common, but occur occasionally, especially in the southern part of the area, an example being the Hekkelstrand marble. Alternating bands with variable admixtures of micaceous or quartzitic material are frequent: likewise, colour banding is common, though colours are nearly always in various shades of grey. The outcrop extent of some marble horizons is undoubtedly quite considerable, but in this respect the map gives too simplified a picture. Splitting of marbles along the strike into two or three horizons or wedging out of horizons is fairly common. Locally this can be ascribed to tectonic causes, though this is the exception rather than the rule. Changes in sedimentation both laterally an vertically are thus regarded as having been fairly frequent.

The iron ores have a distribution comparable with that of the marbles, with which they are closely associated. Generally the ore is found at the base of a marble horizon or as "impregnations" in mica schist within marbles. In some areas, such as those of Bogen, Årstein (Gratangen) and Salangen a number of ore bands are present, a fact that can be attributed partly to repetitions due to folding but also without doubt to primary sedimentation features. A closer inspection of many ore layers reveals intercalations of schist material and variable mixtures of mica schist and ore. Much of the "ore" is only mica schist with a few per cent of magnetite, this being the case especially where the thickness is greater than the average. The richest ore is found in layers of some decimetres, or down to a few centimetres thickness. The persistence of such thin layers is, however, remarkable. In some cases ore horizons, some decimetres thick, can be traced for kilometres; they may then wedge out, reappearing in the vicinity at a level a little higher or lower in the sequence.

From a stratigraphical point of view the strike extension of the ore horizons is of very limited value because of the possibilities of meeting such layers at nearly all levels within the Salangen Group (above the Balteskar quartzite). The nature of the ore is also everywhere the same —rather poor magnetite ore with thin streaks of pinkish coloured garnet. Where the manganese content amounts to some few per cent this is apparent as brown or dark bluish weathering colours. (The manganese is unfortunately bound in silicates, mostly in the garnet.)

The iron ores are absent east of the great central synclinorium except for the siderite ore of Rubben, Målselv. In this area the Salangen Group is represented by marbles and mica schists of various common types.

Petrography.

Conglomerates.

The Elvenes conglomerate has already been described by Foslie (1941). According to this author it is a calcareous mica schist with scattered pebbles or boulders of quartzite and trondhjemite. In the Ballangen area, where the present writer has made some investigations, the conglomeratic character is generally quite impressive. In addition to the two types of pebbles mentioned by Foslie a third type can be added, namely dolomite pebbles. Foslie mentions dolomite lenses, but in the opinion of the present writer, there can be no doubt that some of them, at least, are boulders or pebbles, some of which are not even deformed to any notable degree. The dolomite pebbles are white, sugary-grained and confusingly similar to the quartzite pebbles in hand specimen and on freshly broken surfaces. The trondhjemite boulders consist of plagioclase (comp. unknown), quartz, green biotite, chlorite, epidote and small amounts of sericite, zircon and ore grains. Plagioclase is partly intergrown with quartz in a skeletal or "schriftgranit"—like texture which is not common in the investigated trondhjemites of this area. The plagioclase is saussuritized and also contains irregular calcite grains.

The groundmass of the conglomerate is coarser grained than the trondhjemite boulders and consists of calcite, quartz, biotite, oligoclase, epidote and chlorite. Calcite and quartz are particularly important constituents. In some localities hornblende porphyroblasts have been observed.

The boulder material in the conglomerate of Harstad is partly the same as that of the Elvenes conglomerate, but trondhjemitic or granodioritic boulders are more predominant. Pebbles of calcite marble (partly a red variety) are relatively uncommon. There are also quartzite boulders and rare pebbles of a reddish coloured gneiss-granite resembling some of the basal gneisses. Most pebbles in this area are strongly flattened and elongated, the largest dimension ranging from a few centimetres to some decimetres. The ratios of the three dimensions vary between 1 : 2 : 3 and 1 : 5 : 30 within the same outcrop. Primarily most boulders must have been less than 10 centimetres across.

The groundmass of the Harstad conglomerate is now a quartz-biotite-schist.

The conglomerate at Evenskjær is mainly a quartzite conglomerate according to Vogt (1922), though it also contains calcite boulders. It is in part strongly deformed with considerable elongation of the boulders.

Quartzites.

The Balteskar quartzite and the thin quartzites at the base of the Group are here treated together. They are not especially interesting from a petrographical point of view. At least 80 to 90 per cent of the rocks is quartz in granoblastic intergrowth, frequently as elongate aggregates parallel to the schistosity (and cleavage) planes. Feldspar, mostly albite-oligoclase is never more than an accessory, in contrast to some of the Narvik Group quartzites. Calcite and chlorite are found in small amounts in some localities. Sericitic mica is undoubtedly the most important constituent besides quartz. The flakes nearly always exhibit a parallel planar orientation, the concentration in certain planes giving the characteristic cleavage to these quartzites.

Intercalations of quartz-mica schists are quite common, especially in the thin quartzites of the base.

Mica schists.

The schists of the Salangen Group generally resemble some of those described from the Narvik Group. Garnet-mica schists are especially frequent. Feldspar-bearing rocks are not as common as in the Narvik Group, while kyanite-bearing schists are also relatively rare. The mica schists of the upper parts of the Group are often darker and rich in biotite, though variations across the strike are apparent in most profiles. Because of the similarities to previously described types, the mica schists of the present group will not be discussed further here. The chemical composition of a typical garnet-mica schist is given in Table XIII.

Marbles.

The marbles are medium—to coarse-grained rocks, mostly grey in colour. As already mentioned, colour banding occurs. Impurities and intercalations of mica schist are frequent. In a few cases concentrations of galena, sphalerite, chalcopyrite and pyrite have been found within these intercalations, but these are also partly located along dolomitized slide zones or breccias within the marbles. The mineralizations are unimportant from an economic point of view.

In the eastern area (Bardu—Dividalen), the meta-limestones occasionally contain small clots (standing out on weathered surfaces) which are found to consist of calcite, quartz, muscovite and graphite, these same minerals occurring in the marbles themselves, though in different proportions. The aggregates probably represent pieces of more silicate-rich layers broken up by tectonic movements. In addition to the above-

Table XIII

Chemical composition, mesonorm, Niggli values and mode of garnet-mica schist. Salangen Group.[1]

	Weight %	Cat. %	Mesonorm		Niggli values		Mode (calculated)	
SiO_2	62,42	60,1	Quartz	37,3	al	30½	Quartz	32,2
TiO_2	0,79	0,6	K-feldspar	0,7	fm	46½	Plagio-	
Al_2O_3	14,58	16,5	Na-feldspar	12,5	c	11	clase An_{29}	17,5
Fe_2O_3	1,04	0,8	Ca-feldspar	7,5	alk	12	Epidote	2,0
FeO	7,27	5,8	Corundum	7,1	si	222	Muscovite	13,4
MnO	0,14	0,1	Sum salic	(65,1)	k	0,61	Biotite	15,9
MgO	4,12	5,8			mg	0,47	Almandine	14,2
CaO	2,94	3,0	Biotite	30,1	c/fm	0,24	Magnetite	1,2
Na_2O	1,32	2,5	Magnetite	1,2	qz	+74	Titanite	1,8
K_2O	3,15	3,9	Titanite	1,8			Calcite	1,8
H_2O-	0,12	—	Calcite	1,8				
H_2O+	1,27	—	Sum femic	(34,9)			Sum	100,0
CO_2	0,64	0,9						
P_2O_5	0,07	0,0						
Sum	99,88	100,0						

1) Specimen 521/60, Samueljord, Salangen. Analyst: R. Solli.

Table XIV

Chemical compositions of marbles within the Salangen Group.

	1	2	3	4	5
CaO	48,65	33,07	52,48	49,82	53,44
MgO	2,40	16,83	0,12	2,24	0,96
$CaCO_3$	86,83	59,02	93,67	88,92	95,38
$MgCO_3$	5,02	35,21	0,25	4,69	2,01
Sum carbonates	91,85	94,23	93,92	93,61	97,39

Localities of analysed specimens:

1 Sp. 406/59. Gratangsbotn.
2 Sp. 437/59. Gamvik, Lavangen.
3 Sp. 540/60. Kråkrøhamn, Andørja.
4 Sp. 523/60. 0,5 km E. of Samueljord, Salangen.
5 Sp. 510/59. Rubben, Kirkesdal, Målselv.

Analyst: M. Gustavson.

mentioned minerals scattered grains of sodic plagioclase are present in some thin sections. Clinozoisite, actinolite and diopside are common in certain layers within the marbles and along the marble boundaries, their mutual relations and appearance depending on the grade of metamorphism: this is discussed in a later chapter.

Dolomitic layers or lenses are relatively common in the southern areas (south of Ofotfjorden and in the Tjeldsund area). Further north and east true dolomites are rare in this group. The compositions of 5 marbles have been determined by chemical analysis and are shown in Table XIV. Except for no. 2 they are all typical calcite marbles.

Iron ores.

The iron ores of the Håfjell synclinorium have been described by Foslie (1949). The present writer has visited most of the occurrences north of Ofotfjorden and some of those on the southern side. The detailed observations made will, however, be reserved for a later publication, and accordingly the ores are only briefly described here.

In most areas the ore horizons are present as very thin layers, from less than 1 metre up to a few metres in thickness. Locally, thicknesses and concentrations are greater and the ores have temporarily been worked or investigated for economic purposes. This is the case with the Bogen mine (worked with several interruptions 1906—39), Salangsverket (worked 1909—12), Andørja (investigated by Christiania Spigerverk in recent years), Lavangen (drilling 1915—19), Rolla and Årstein (recent investigations). For the most part the ores seem to be too poor in iron or else the layers are too thin for exploitation. The total iron content generally lies between 20 and 25 per cent while in addition, some of the ores are manganese-bearing up to 10 per cent or more. The Mn-content is present mainly in stripes and clots of garnet and is thus of little or no economic value (Fig. 30). Phosphorus varies in amount, some ores being relatively rich in apatite. The content of sulphur seems always to be low.

Petrographically the main feature is that the ores south of Ofotfjorden contain hematite as the dominant ore mineral (hematite: magnetite = about 2 : 1 or 3 : 1) whereas the ores north of Ofotfjorden are magnetite ores with up to a few per cent hematite. Additional minerals are quartz, calcite, biotite, garnet, epidote and hornblende.

A different type of iron ore is found in Kirkesdalen in the eastern-

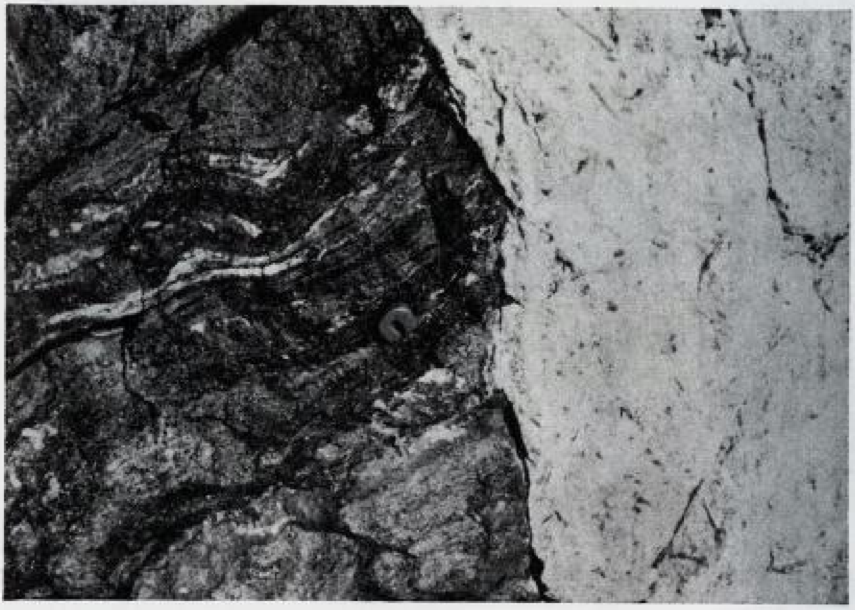


Fig. 30. Banded iron ore cut by granitic dyke, Salangen Group. Storhaugen, Salangen.
Båndet jernmalm gjennomsatt av granittgang, Salangen-gruppen. Storhaugen, Salangen.

most area. This ore has been described earlier (Landmark 1952, Gustavsson 1960). The chief mineral is a sideritic carbonate containing some few per cent of Mn and small amounts of Ca and Mg. Parts of the ore contain a few per cent of magnetite. Garnet is the most important silicate mineral. The iron content of the ore is between 40 and 50 per cent while manganese amounts to 5—8 per cent.

In connection with the magnetite ores, together with those of the Narvik Group, there occur interlayered stripes of red and green minerals. Where quartz veins cut the layers the grain size of these minerals is increased, sometimes to 2—3 centimetres. The minerals in question are red almandine garnet and diopsidic pyroxene. Quartz and ore minerals are present as well. The stripes of garnet are frequently found to consist of a multitude of small polygonal or rounded garnet individuals, partly without other interjacent mineral grains. The size of the garnets is about 0,05 millimetres or even less. The pyroxene is slightly greyish green, optical positive and with extinction angles c/Z about 38°. Grains are mostly less than 1 millimetre across.

The aggregates of diopside and garnet are to be considered as a type

Table XV

Optical data for amphiboles, pyroxenes and plagioclases of the meta-sedimentary amphibolites.

a) Amphiboles.

Specimen	$2 V_z$	Extinction c/Z	Colour (Z)	n_z
458/59	78°	20°	Blue-green	n.d.
502/59	72°	21°	„ „	„
417/59	85°	16°	Brownish green	1,655 ± 0,002
422/59	86°	21°	Pale green	n.d.
539/60	82°	18°	„ „	„
472/59	78°	17°	„ „	„
568/60	80°	22°	Brownish green	„

b) Pyroxenes.

Specimen	$2 V_z$	Extinction c/Z	Alteration
502/59	58°	39°	None
417/59	n.d.	39°	„
539/60	54°	38°	To hornblende along rims.
472/59	57°	35°—40°	To hornblende.
522/60	60°	n.d.	None.

c) Plagioclases.

Specimen	Composition	Twinning	Amounts
502/59	An_{22-28} (zoned)	Albite law	5—10 % (vol.)
517/59	An_{30-35} (variable)	Albite + pericline	30 %
422/59	An_{10}	„ „	5 %
539/60	An_{42}	„ „	5—10 %
472/59	n.d.	Albite law	5—10 %
568/60	An_{29}	„ „	25—30 %

Localities of specimens:

458/59 Garnes, Salangen.
502/59 Meby, Salangen.
417/59 Okshammeren, Gratangen.
422/59 Sandnes, Gratangen.
539/60 Kråkrøhamn, Andørja.
472/59 Skårvikelv, Salangen.
568/60 Nappen, Andørja.
522/60 Samueljord, Salangen.

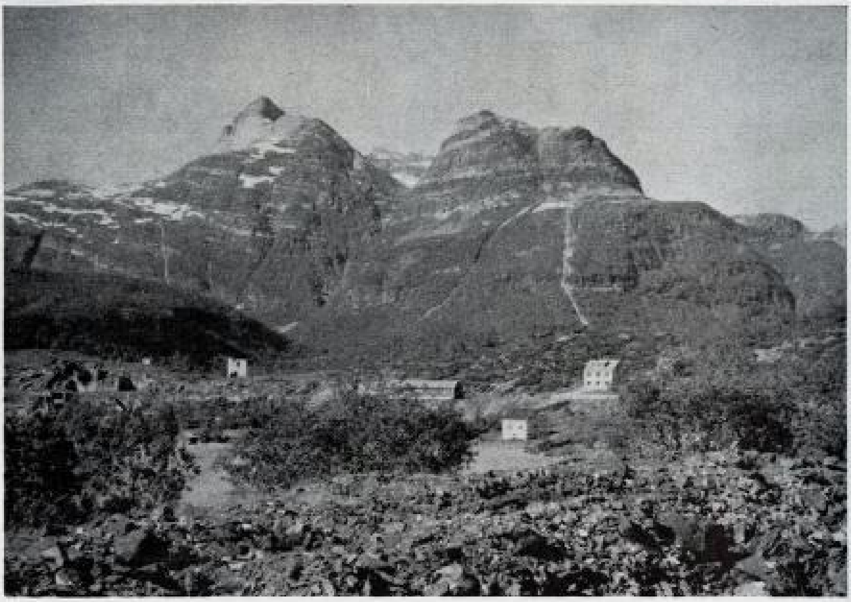


Fig. 31. Outlook from Bogen towards Niingen. The mountains consist of Niingen schists and -gneisses with dykes of granodioritic to trondhjemitic composition (light-coloured).

Utsyn fra Bogen mot Niingen. Fjellene består av Niingengruppens skifre og gneiser med trondhjemittiske til granodiorittiske ganger (lyse på bildet).

of reaction skarn. Younger quartz veins evidently promoted further growth of these minerals to a pegmatitic grain size.

Meta-sedimentary amphibolites.

Lime-rich amphibolites occur frequently as lenses, boudins and concordant layers. The abundance of lime silicates and the frequent association with meta-limestones indicate that these rocks were once impure calcareous sediments.

The texture of the amphibolites is generally characterized by a strong parallelism of minerals. The most common minerals are hornblende, plagioclase (not always present), diopsidic pyroxene, epidote-clinozoisite, titanite, biotite, calcite and quartz. Less frequent are scapolite, garnet, tourmaline, chlorite, apatite, zircon and ore minerals. Some properties of the amphiboles, pyroxenes and plagioclases are shown in Table XV. It is observed that considerable variations occur in optical properties, and consequently also in composition.

The Niingen Group

This group is represented within the central part of the Håfjell synclinorium and its continuation from Bogen to Salangen. The rocks of the Niingen Group form the uppermost part of the sequence in the Ofoten–Southern Troms area.

A quartzite, the Butind quartzite, occurs within the lowest part of the Group in the south. The rest of the Group is the Niingen Schist of Th. Vogt and was designated by him as "mica schist and injection gneiss' (on map in manuscript). Though the present writer prefers not to use the term "injection gneiss," it is evident that the rocks found by Vogt as being typical of the Niingen area are the same as those investigated by the writer in the areas of Gratangen and Salangen. The sequence seems to be composed of quite uniform mica schists or garnet-mica schists, frequently with amphibolites and younger trondhjemites and granodiorites cutting across the schistosity (Fig. 31). The grano-

Table XVI

Chemical composition, mesonorm, Niggli values and mode of kyanite-bearing gneiss, Niingen Group[1]).

	Weight %	Cat. %	Mesonorm		Niggli values		Mode (calculated)	
SiO_2	60,65	57,9	Quartz	35,4	al	44½	Quartz	28,0
TiO_2	0,90	0,6	K-feldspar	8,3	fm	39	Plagioclase	
Al_2O_3	20,56	23,1	Na-feldspar	12,5	c	2½	An[illegible]	16,0
Fe_2O_3	0,78	0,6	Corundum	15,8	alk	14	Epidote	tr.
FeO	6,68	5,3	Sum salic	(72,0)	si	222	Muscovite	18,3
MnO	0,12	0,1			k	0,66	Biotite	18,5
MgO	2,95	4,3	Biotite	25,1	mg	0,41	Almandine	5,0
CaO	0,73	0,7	Magnetite	0,9	qz	+ 66	Kyanite	12,3
Na_2O	1,35	2,5	Titanite	1,8			Magnetite	0,5
K_2O	4,00	4,8	Calcite	0,2			Ilmenite	1,2
H_2O-	0,10	—	Sum femic	(28,0)			Calcite	0,2
H_2O+	1,36	—					Zircon	tr.
CO_2	0,07	0,1						
P_2O_5	0,06	0,0					Sum	100,0
	100,31	100,0						

1) Specimen 416/59, Bratberg, Gratangen. Analyst: R. Solli.

Table XVII

Chemical analyses, mesonorms, Niggli values and modes of three schists from the Tysfjord area (analyses completed for S. Foslie) [1].

	1		2		3	
	Wt %	Cat. %	Wt. %	Cat. %	Wt %	Cat. %
SiO_2	92,46	91,2	72,68	71,0	55,37	52,2
TiO_2	0,12	0,1	0,72	0,5	0,94	0,7
Al_2O_3	3,52	4,1	10,53	12,1	17,61	19,6
Fe_2O_3	0,16	0,2	3,03	2,2	1,29	0,9
FeO	0,48	0,4	2,23	1,8	6,23	4,9
MnO	0,01	—	0,13	0,1	0,05	—
MgO	0,15	0,2	2,89	4,2	5,83	8,2
CaO	0,16	0,2	3,32	3,4	5,68	5,8
BaO	0,09	—	0,06	—	0,06	—
Na_2O	0,73	1,4	0,92	1,7	2,20	3,9
K_2O	1,79	2,2	2,18	2,7	2,95	3,5
H_2O —	0,05	—	0,09	—	0,11	—
H_2O +	0,24	—	0,76	—	1,37	—
CO_2	0,02	—	0,16	0,2	0,14	0,2
P_2O_5	0,02	—	0,17	0,1	0,20	0,1
S	0,02	—	0,01	—	tr.	—
Sum	100,02	100,0	99,88	100,0	100,03	100,0

Mesonorms:

	1	2	3
Quartz	80,1	52,3	18,3
K-feldspar	10,2	5,2	—
Na-feldspar	7,0	8,5	19,5
Ca-feldspar	0,5	12,5	18,2
Corundum	0,3	2,7	3,9
Sum salic	(98,1)	(81,2)	(59,9)
Hornblende	—	—	7,9
Biotite	1,3	13,3	28,0
Magnetite	0,3	3,3	1,4
Titanite	0,3	1,5	2,1
Apatite	—	0,3	0,3
Calcite	—	0,4	0,4
Sum femic	(1,9)	(18,8)	(40,1)

Niggli values:

	1	2	3
al	42½	30	29½
fm	16	42	42
c	3½	17	17½
alk	38	11	11
si	1895	353	156
k	0,61	0,60	0,48
mg	0,31	0,50	0,59
qz	+ 1643	+ 209	+ 11

Modes (calculated):

	1	2	3
Quartz	81,0	52,9	18,2
Microcline	8,7	—	—
Plagioclase	6,0	7,6	28,3
(% An in plag.)	(5)	(32)	(39)
Biotite	0,8	17,0	31,7
Muscovite	2,4	7,0	—
Chlorite	0,6	—	0,3
Garnet	—	—	2,0
Epidote	—	8,9	—
Zoisite	—	—	4,5
Hornblende	—	4,0	13,2
Pyrrhotite	tr.	—	—
Hematite	—	1,7	—
Magnetite	0,2	—	0,3
Rutile	—	0,2	—
Titanite	0,3	0,2	0,7
Apatite	tr.	0,4	0,5
Calcite	tr.	—	0,3

1) All calculations except the mesonorms by S. Foslie. Analyst: E. Klüver. (1930—1931.)

Analysis 1. Quartz schist, Mannfjellvann, Tysfjord.
„ 2. Mica schist, Brynsvann, Tysfjord.
„ 3. Mica schist, S. of Rusløkvann, Tysfjord.

diorites may also be concordant or sub-concordant, but mostly they are clearly crosscutting. Feldspar-bearing schists are met with in a number of localities. Petrographically the Niingen Group thus markedly resembles the Narvik Group. During mapping the question therefore has arisen as to whether the Niingen Group could be an overthrust nappe of Narvik Group rocks. Such a possibility is further supported by the findings of serpentinite bodies within the Niingen Group. Within the map area serpentinites are restricted to the Narvik and Niingen

Groups. Occurrences of serpentinite (peridotite) have in the Caledonides of Norway been taken as an indication of old age for the host meta-sediments (see f. inst. Strand, 1960, p. 176). The tectonic evidence found within the present area has, however, not quite convinced the writer that over-thrusting of the Niingen schists has taken place. Because of their close similarity to the mica schists of the Narvik Group, further description of the Niingen schists will not be given. The chemistry of a typical kyanite-bearing gneiss from the Narvik Group, shown in Table X, compared with Table XVI, emphasizes the inherent similarities.

Schists of the Tysfjord area

As these schists have been described by Foslie (1949) they will be commented on only briefly here. On the map, Fig. 1, they have been given a special symbol to denote that the stratigraphical position is uncertain. It is obvious, however, that they belong to a relatively low part of the sequence when compared with the rest of the map area. There are marbles and mica schists in the lowermost part which resemble those of the Rombak Group while the gneiss-granite layers suggest parallelization with the Precambrian thrust slices of the Rombak profile. The upper schists of the Tysfjord area for instance the Reppi schist (calcareous mica schist, partly with amphibole porphyroblasts), the Gicce gneiss (feldspar-bearing "injection-gneiss") and mica schists with basic and acid eruptives strongly indicate that Narvik Group rocks are present. It is not possible, however, to draw any boundaries between supposed Narvik Group and Rombak Group schists in this area without extensive remapping.

Table XVII contains three chemical analyses of schists from the Tysfjord area. These analyses were completed for Foslie in 1930—31 but have never been published.

Comments on the chemistry of the meta-sediments

Nine chemical analyses, carried out as a part of the petrographical investigation, have been presented in the preceding chapters together with seven analyses completed for S. Foslie, but as yet unpublished. Except for one, that of a quartzite, these analyses are listed together in Table XVIII for comparison and discussion. As the analysed rocks are few in number and since their purpose was mainly to illustrate some of the common schist types, the analyses are not well suited to solving problems of lateral or stratigraphical variation. However, some interesting features appear in the light of earlier investigations on the chemistry of Caledonian schists in other areas.

Th. Vogt (1927) classified the Scandinavian Caledonides into the "eastern" and "western" sedimentary facies. According to him the eastern facies is composed of typical residual sediments, high in Al_2O_3 and K_2O. The western facies sediments were supposed to be more incompletely weathered and consequently higher in the constituents Na_2O, CaO and MgO. Most of the sediments of the Sulitjelma area, investigated by Vogt, and the Nordland sediments in general were assigned to the western facies, whereas the "phyllite formation" of central Southern Norway (e.g. the Stavanger area) was taken as typical of the eastern facies. A distinct difference in thicknesses between the two facies' was also suggested, the western facies sediments being present in much thicker piles than those of the eastern facies. From these features Vogt inferred a northwestern origin for the sediments with increasing decomposition eastwards away from the source.

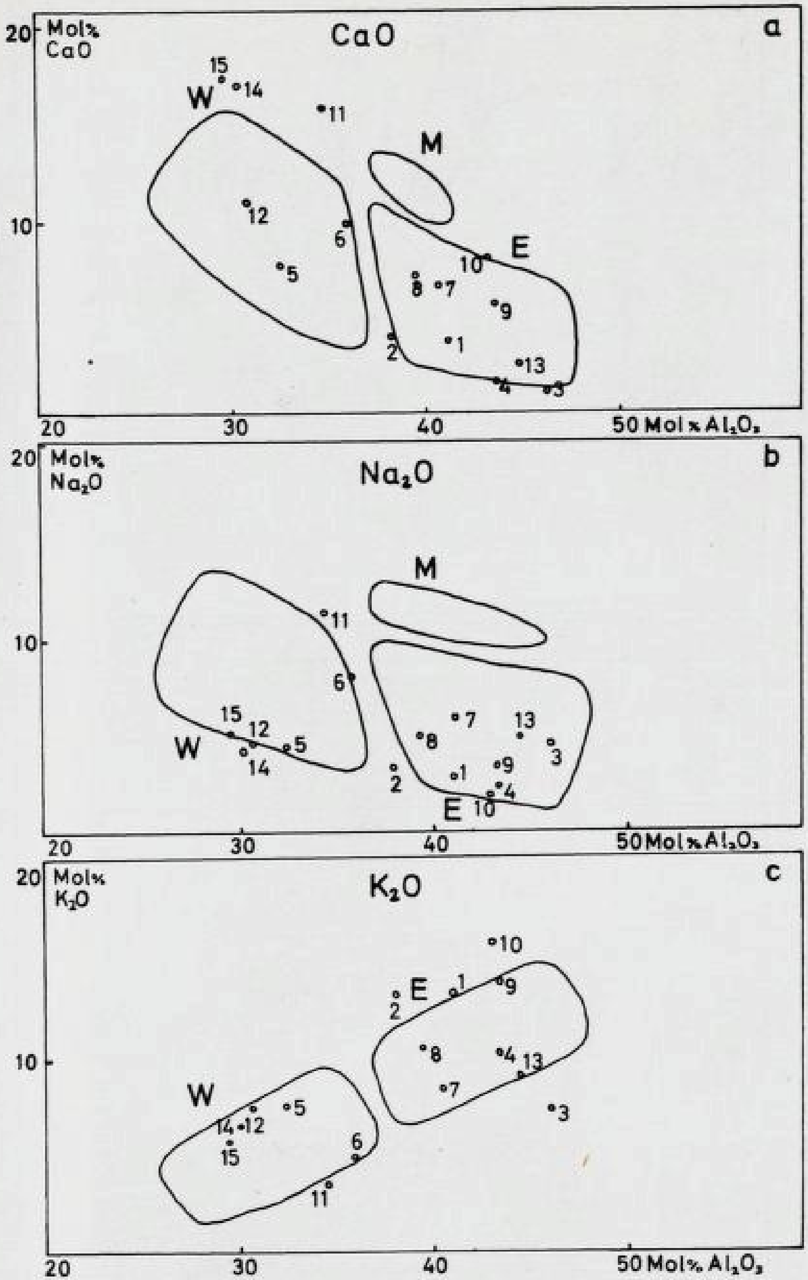
The unweathered or incompletely weathered character of many sediments from the central part of the Scandinavian Caledonides has later been emphasized by several authors, including C. W. Carstens (1928), Kulling (1933) and Strand (1951). In a recent publication K. Bjørlykke (1965) has shown from the Oslo area, that the chemical differences in a single stratigraphical section may be just as great as the regional differences between eastern and western facies. The changes were supposed to be related to orogenic events and a high MgO content was interpreted as due to the derivation of sediments from basic volcanic rocks and the nappes of basic composition in central Southern Norway.

Strand (1951) also emphasized the importance of source rocks on the mg value of the geosynclinal sediments, but suggested a metasomatic origin for sodium and lime in many schists and gneisses of the Caledonides. In a later paper (Strand, 1960) the possibility of metasomatism is expressed with special reference to plagioclase gneisses and micaceous gneisses within the Nordland sediments. It is of interest also in this connection that Strand (1961) divides the sediments of western (eugeosynclinal) facies into Trondheim region facies and Nordland facies. Typical of the former are thick intercalations of basic as well as acid volcanics at various levels. The metamorphism varies from weak to strong. The Nordland sediments are divided into, "a lower division mainly of pelites and an upper division with thick limestones and dolomites," according to the old descriptions by J. H. L. Vogt (1897). A further feature is the presence of sedimentary iron ores whereas the importance of volcanics is insufficiently known. Coarse clastic sediments are characteristically absent. Metamorphism is high-grade in most areas.

Bugge (1948) emphasized the residual nature of the sediments of the iron ore fields of Dunderlandsdalen. His conclusions are therefore in contradiction to Vogt's supposition of western facies types being characteristic of the Nordland sediments.

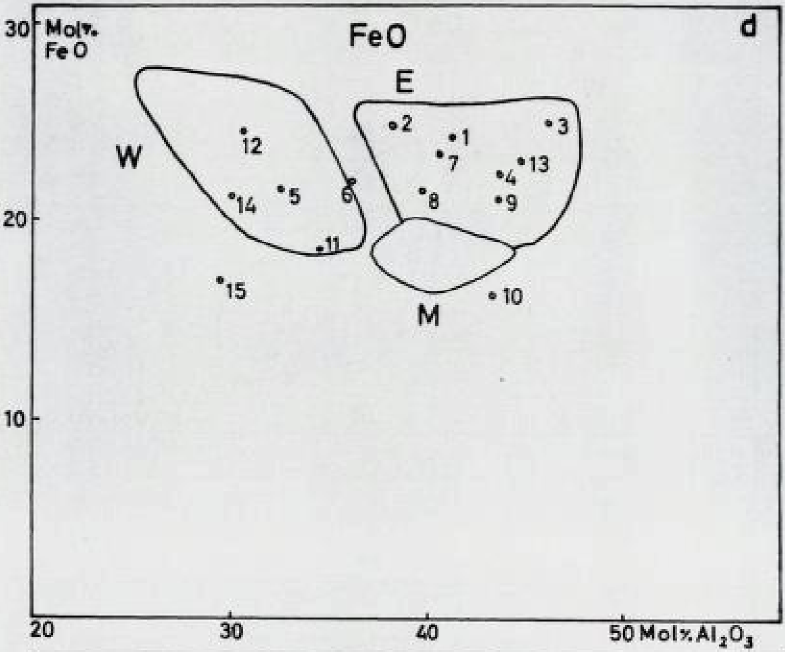
In order to place the schists of the map area into the general picture based on our present knowledge, the chemical analyses have been recalculated and plotted in the same way as Vogt (1927) did for the Sulitjelma area. In Fig. 32 a—e the groups of western and eastern facies as well as metasomatic rocks have been redrawn from the diagrams of Vogt. (The parameters used in Figures 32 and 33 are shown in Table XVIII.) As shown in this figures ,the meta-sediments are of both eastern and western facies type. Nine of the analyses (no:s 1, 2, 3, 4, 7, 8, 9, 10, 13) fall within, or close to, the fields of eastern facies, whereas the other six analyses (no:s 5, 6, 11, 12, 14, 15) belong to the western type. The latter rocks include four amphibole-bearing schists (5, 6, 14, 15) and two plagioclase-micagneisses (11, 12). The amphibole-bearing schists are subordinate in amount over most of the area, whereas the plagioclase-mica gneisses, with or without garnet, are relatively common in the Narvik Group and higher tectonic units. Among the eastern facies sediments it is interesting to note that the mica schist typical of the Rombak Group belongs to this type, as does the Hyolithes Zone shale. The staurolite and kyanite-bearing schists of the Narvik Group are all typical eastern facies rocks; of these, the kyanite-bearing rocks are rather frequent within the area. Three of the four schists from the Ballangen area (8, 9, 10) belonging to the upper part of the Narvik Group, also fall in the eastern facies group, all being relatively common rock types within the area.

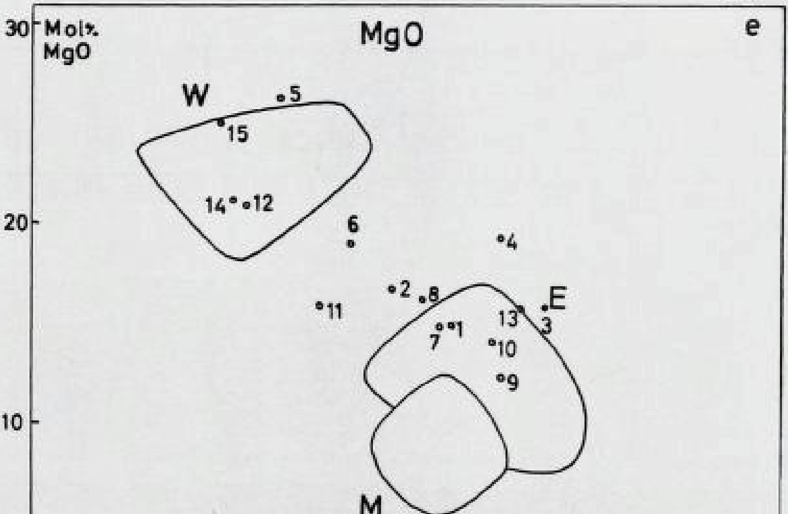
There seem to be no regional or stratigraphical regularities in the distribution of the two facies types. On the contrary, rocks of the two facies appear to alternate at various levels. For instance, plagioclase-mica schists or gneisses of the western type (11, 12), frequently alternate with mica schists and kyanite gneisses of eastern facies chemistry (8, 9, 10 and 7, 13). Amphibole-bearing schists (5, 6, 14, 15) likewise are intercalated with "ordinary" mica schists of supposed eastern type.



Figs. 32 a—e. Plotting of chemical analyses from the area on the diagrams of Vogt. E = eastern facies, W = western facies, M = metasomatic rocks.

Plotting av kjemiske analyser i Vogts diagrammer. E = østlig facies, W = vestlig facies, M = metasomatiske bergarter.


30
Mol%
FeO
FeO
d
W
E
M
20
10
20
30
40
50 Mol% Al_2O_3
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15


30
Mol%
MgO
MgO
e
W
E
M
20
10
20
30
40
50 Mol% Al_2O_3
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15

The field relations suggest that the latter alternation is probably a primary sedimentation feature. Where the plagioclase gneisses are concerned, a metasomatic origin must also be considered. Taking for example, analysis 11 from the Ballangen area—a gneiss associated with schists of the eastern facies type (10)—Fig. 32 shows a very high Na_2O + CaO content, whereas MgO is closer to the rocks of eastern type. A transformation of the schists into plagioclase-mica gneisses by sodium/calcium metasomatism seems a probable explanation. In other cases, association of plagioclase-bearing gneisses with "granites" lends further support to this view. A primarily high Na_2O + CaO content of certain layers cannot, however, be entirely ruled out. Admixture of keratophyric material is a possible explanation of such a feature, more so as rocks thought to be keratophyres are present in the eastern parts of the Narvik Group. It seems difficult, at present, to favour either of the two possible explanations given above.

Concerning the amphibole-bearing rocks there is little evidence pointing to a metasomatic origin. The alternations of these schists with mica schists of eastern (?) type can be explained in one of two ways, either (1) by relatively rapid changes in the sedimentation or weathering conditions, or (2) by changes in the type of source rock. The former explanation does not seem probable to the present writer. As emphasized by Strand (1951), sediments with high values of both fm and mg (western facies type) can be considered as largely derived from gabbroid rocks. It is possible that this is the case with the amphibole-bearing schists of the present area. This suggestion is consistent with the fact that such sediments are absent in the Rombak Group where also basic rocks are absent. On the other hand, no evidence has as yet been found to suggest that rocks of gabbroic composition were in a position of erosion at the time of sedimentation of the Narvik Group and the higher tectonic units. Basic igneous rocks are, however, frequent in these sedimentary groups. An explanation along these lines is also in accordance with that given by Bjørlykke (1965) for the sedimentary changes at the Lower Ordovician/Middle Ordovician boundary in the Oslo area and in some Swedish sediments. A plotting of the analyses from the present area in Bjørlykke's diagram reveals no special relations to rocks of that area. There is, however, a marked difference between rocks of the eastern and western types from the present map area, although this grouping shows no simple relation to stratigraphy (Fig. 33).

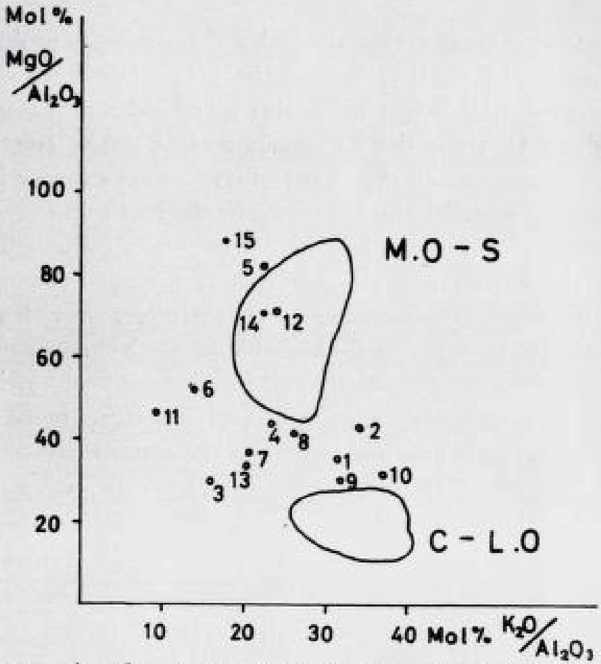


Fig. 33. Plotting of analyses on a diagram after K. Bjørlykke (1965). C - L.O. = Cambrian and Lower Ordovician of the Oslo area and Sweden. M.O. - S = Middle Ordovician and Silurian of the same areas.

Plotting av analyser i et diagram etter K. Bjørlykke (1965). C - L.O. = kambriske og underordoviciske bergarter fra Oslo-feltet og Sverige. M.O. - S = mellomordoviciske og siluriske bergarter fra de samme områder.

The Storfjell Group is not included in the considerations above as analyses of these rocks are not available. It seems obvious, however, that they are of a less residual type than the Hyolithes sediments because of the feldspar content and are in this respect closer to the western facies type.

In conclusion it can be said that a distinction between sediments of eastern and western facies is possible also in the present area, but the occurrence of the two types is not in accordance with Vogt's theory of a regional east—west distribution pattern. It is possible that all rocks of the Rombak Group together with the Hyolithes sediments belong to the eastern facies. In the higher tectonic (and stratigraphic) units, probably originating from more westerly areas, the two types alternate. In this respect the conditions prevailing during sedimentation seem to have been less regular than those of the Oslo area. The possible influence of metasomatic processes must, however, still be kept in mind.

Table XVIII

Chemical composition of 15 schists from the map area.

	1	2	3	4	5	6	7
SiO_2	61,21	65,74	57,75	57,80	57,69	62,79	59,47
TiO_2	0,59	0,80	0,61	0,70	1,10	0,74	1,00
Al_2O_3	17,20	14,59	22,11	21,06	17,23	16,48	19,80
Fe_2O_3	3,92	0,49	0,38	2,99	2,94	2,13	1,88
FeO	3,61	6,06	8,05	4,77	5,43	5,15	6,31
MnO	0,05	0,08	0,08	0,22	0,13	0,17	0,11
MgO	2,52	2,62	3,03	3,74	5,58	3,40	2,89
CaO	0,98	0,81	0,24	0,35	2,20	2,51	1,67
BaO	—	—	—	—	—	—	—
Na_2O	0,69	0,84	1,31	0,69	1,53	2,05	1,86
K_2O	5,04	4,49	3,49	4,89	3,62	2,15	3,68
H_2O —	0,28	0,12	0,06	0,23	0,15	0,13	0,14
H_2O +	3,63	2,03	2,42	2,24	2,50	1,85	1,47
CO_2	0,06	0,67	0,07	0,07	0,02	0,11	0,06
P_2O_5	0,03	0,06	0,07	0,06	0,20	0,08	0,06
S	—	—	—	—	—	—	—
CuO	—	—	—	—	—	—	—
Sum	99,81	99,42	99,67	99,81	100,32	99,78	100,36

8	9	10	11	12	13	14	15
70,03	63,28	56,17	60,15	62,42	60,65	72,68	55,37
0,75	0,75	0,56	0,78	0,79	0,90	0,72	0,94
13,46	17,74	19,49	17,45	14,58	20,56	10,53	17,61
0,56	0,93	0,30	0,71	1,04	0,78	3,03	1,29
4,64	5,03	5,05	5,52	7,27	6,68	2,23	6,23
0,07	0,07	0,07	0,08	0,14	0,12	0,13	0,05
2,17	2,01	2,56	3,20	4,12	2,95	2,89	5,83
1,35	1,18	1,90	4,41	2,94	0,73	3,32	5,68
0,05	0,07	0,13	0,04	—	—	0,06	0,06
1,07	0,88	0,49	3,56	1,32	1,35	0,92	2,20
3,29	5,21	6,91	1,66	3,15	4,00	2,18	2,95
0,08	0,14	0,24	0,09	0,12	0,10	0,09	0,11
1,38	1,73	2,96	1,05	1,27	1,36	0,76	1,37
0,57	0,42	1,55	0,70	0,64	0,07	0,16	0,14
0,16	0,21	0,09	0,16	0,07	0,06	0,17	0,20
0,15	0,21	2,49	0,40	—	—	0,01	tr.
—	—	0,056	—	—	—	—	—
99,78	99,86	101,02	99,96	99,88	100,31	99,88	100,03

		1	2	3	4	5	6
al	Al_2O_3	41	38	46	43½	32½	36
fm	FeO	24	24½	25	22½	21½	22
	MgO	15	17	16	19½	26½	19
c	CaO	4	4	1	1½	7½	10
alk	K_2O	13	13	7½	10½	7½	5
	Na_2O	3	3½	4½	2½	4½	8
si	SiO_2	247	292	204	202	183	233
k	$\frac{K_2O}{K_2O + Na_2O}$	0,83	0,77	0,64	0,83	0,61	0,40
mg	$\frac{MgO}{MgO + FeO}$	0,38	0,41	0,39	0,47	0,55	0,46
$M =$	$\frac{Al_2O_3 + K_2O}{MgO + CaO + Na_2O}$	2,2	2,1	2,5	2,3	1,0	1,1
$100 \cdot$	$\frac{MgO}{Al_2O_3}$	36,7	44,7	34,8	44,8	81,5	52,8
$100 \cdot$	$\frac{K_2O}{Al_2O_3}$	31,7	34,2	16,3	24,1	23,1	14,0

Types and localities of the analysed rocks:

1. Sandy shale, Hyolithes Zone, Dividalen.
2. Mica schist, Rombak Group, Rombaken.
3. Staurolite schist, Narvik Group, Rostafjell, Dividalen.
4. Staurolite schist, Narvik Group, Dittitind, Bardu.
5. Schist with amphibole porphyroblasts, Narvik Group, Jormecacca, Bardu.
6. Schist with amphibole porphyroblasts, Narvik Group, Seternes, Bardu.

7	8	9	10	11	12	13	14	15
40½	39½	43½	43	34½	30½	44½	30	29½
23½	21½	21	16½	18½	24½	23	21	17
15	16½	12½	14½	16	22	16	21	25
6½	7	5½	8	16	11	2½	17	17½
8½	10½	14	16	3½	7½	9	6½	5½
6	5	3½	2	11½	4½	5	4½	5½
208	348	265	210	203	222	222	353	156
0,57	0,67	0,80	0,90	0,23	0,61	0,66	0,60	0,48
0,39	0,43	0,38	0,46	0,47	0,47	0,41	0,50	0,59
1,8	1,8	2,7	2,3	0,9	1,0	2,3	0,9	0,7
37,0	41,7	29,0	33,7	46,4	72,1	36,0	70,0	84,7
21,0	26,6	32,2	37,2	10,0	24,6	20,2	21,7	18,6

7. Kyanite-bearing gneiss, Narvik Group, Lemman, Skjomen.
8. Mica schist, Brattåsen, Narvik Group, Ballangen.
9. Mica schist, Brattåsen, Narvik Group, Ballangen.
10. Mica schist, Narvik Group, Bjørkåsen, Ballangen.
11. Mica gneiss, Narvik Group, Bjørkåsen, Ballangen.
12. Garnet-mica gneiss, Salangen Group, Samueljord, Salangen.
13. Kyanite-bearing gneiss, Niingen Group, Bratberg, Gratangen.
14. Hornblende mica schist, Brynsvann, Tysfjord.
15. Calciferous mica schist, Rusløkvann, Tysfjord.

METAMORPHISM AND MINERAL FACIES OF THE SEDIMENTARY ROCKS

Introduction

The nomenclature and classification of mineral facies show noticeable variations in different papers and textbooks, and the boundary lines between them are drawn on different premises. This seems to some extent to result from the fact that the various authors have gained their experience from different metamorphic terrains. It is evident from the literature that a facies classification constructed for one area does not always fit the rocks of other areas.

Miyashiro (1961), in an attempt to establish a system of universal application, divided the rocks of regional metamorphism into facies series on the basis of various pressure types, characterized by the stability of certain minerals at appropriate temperature conditions. The chief types of facies series are: the kyanite-sillimanite type, the andalusite-sillimanite type and the jadeite-glaucophane type. In addition, two intermediate types, a low pressure and a high-pressure type, are proposed. According to Miyashiro the kyanite-sillimanite series, representing somewhat intermediate pressure conditions, is the common type within the Appalachian fold belt of North America and the Caledonian belt of Scotland and Scandinavia and is thus of particular interest to us in the present area. The facies series of this type of metamorphism is as follows: Greenschist facies, epidote-amphibolite facies, amphibolite facies and granulite facies, or in terms of mineral zoning, the chlorite, biotite, almandine, staurolite, kyanite and sillimanite zones of pelitic rocks.

A further sub-division of the traditional facies of Eskola (1921) into sub-facies' has been proposed by Turner & Verhoogen (1960). In this system the greenschist facies includes three sub-facies', the uppermost of which extends up to the lower boundary of the amphibolite facies (almandine-amphibolite facies of T & V). The amphibolite facies has been divided into four subfacies'. Though the present writer thinks that the extending of the greenschist facies to include also epidote-amphibolite rocks, as compared with the original system of Eskola, is unfortunate, he finds the division into sub-facies' very useful. The system adopted in this paper is therefore a combination of that of Eskola and the sub-facies system of Turner & Verhoogen, as shown by table XIX.

Table XIV

Facies classification used in this paper.
(Modified after Eskola and Turner & Verhoogen.)

Main facies	Sub-facies	Mineral zone
Greenschist Facies	Muscovite-chlorite sub-facies	Chlorite zone
Epidote-amphibolite facies	Epidote-biotite sub-facies	Biotite zone
	Epidote-almandine sub-facies	Almandine zone
Amphibolite facies	Staurolite-almandine sub-facies	Staurolite zone
	Kyanite-almandine-muscovite sub-facies	Kyanite zone
	Sillimanite-almandine-muscovite sub-facies	Sillimanite zone
	Sillimanite-almandine-orthoclase sub-facies	

General survey of metamorphic facies within the area

The main regional metamorphism within the map area is of the kyanite-sillimanite type as defined by Miyashiro (1961). This is shown by the common occurrence of kyanite in pelitic rocks. The metamorphism has nowhere been as high as to stabilize sillimanite. Thus the kyanite-almandine-muscovite subfacies represents the highest grade rocks within the area.

In terms of mineral zoning the chlorite, biotite, almandine, staurolite and kyanite zones are encountered. The lowest grade rocks (chlorite zone) belong to the quartz-albite-muscovite-chlorite subfacies of Turner & Verhoogen, corresponding to the greenschist facies of Eskola (1921) and Barth (1962). The next step of progressive metamorphism, the biotite zone, corresponds to the quartz-albite-epidote-biotite subfacies of T. & V. and to the lower part of the albite-epidote-amphibolite subfacies of Barth. The almandine zone is equivalent to the upper part of Barth's albite-epidote-amphibolite subfacies and the quartz-albite-epidote-almandine subfacies of T. & V. High-grade rocks of the staurolite and kyanite zones belong to the lower part of the amphibolite facies of T. & V. and to the oligoclase-epidote-amphibolite subfacies of Barth. As staurolite is rare within the area it has not been possible to discriminate between the staurolite-almandine subfacies and the kyanite-almandine-muscovite subfacies on map scale.

The map of metamorphic facies, Fig. 2, shows the distribution of the muscovite-chlorite subfacies (1), the epidote-biotite subfacies (2), the epidote-almandine subfacies (3) and the staurolite-almandine + kyanite-almandine-muscovite subfacies' (4). Within the area of the last-mentioned subfacies (4) oligoclase is the common feldspar of the pelitic schists. Andesinic plagioclases are occasionally met with in schists and diopside in calcareous rocks. It is thus probable that the highest metamorphism extends up to the lower part of amphibolite facies as defined by Barth (epidote with plagioclase An > 35). Areas with such rocks always lie within the kyanite zone, but with the data available it is impossible to distinguish them on the map.

Porphyroblastic development of feldspars and ferromagnesian minerals is not uncommon. Evidence of syntectonic as well as post-tectonic growth of the porphyroblasts has been detected.

Retrogressive metamorphism is restricted to certain parts of the sequence. Indications of a relationship with thrusting have been found in most cases.

Possible metasomatic changes have already been discussed shortly in connection with the chemistry of the rocks. It can be stated that within the bulk of metasediments metasomatism has not been of any great importance. An exception could be parts of the Narvik and Niingen Groups (plagioclase gneisses).

The muscovite-chlorite subfacies

Rocks of this subfacies are present in three small areas in the eastern districts from Bardu to Dividalen (see map, Fig. 2). Stratigraphically they belong to the lover sequences, mainly the Storfjell Group. It is, furthermore, uncertain whether parts of the Hyolithes Zone should be included in this subfacies. Because of the fine-grained textures of the latter rocks it is difficult to decide whether chlorites and micas are of clastic or low metamorphic origin. Some veins within the Hyolithes

Zone, consisting of quartz, chlorite and titanite, are certainly metamorphic products. Titanite has not been detected in the sediments outside the veins. It is of some interest to note that the veins are apparently most frequently encountered in the tectonized parts of the Zone. As there are no other indications of a higher metamorphic grade for the tectonized shales, the frequency of the veins is, therefore, probably a reflection of the increased possibility of solutions circulating in cracks and fissures of the tectonized rocks.

Observations on the sandstones of the upper part of the Hyolithes Zone also appear to demonstrate the role of tectonization in metamorphic recrystallization. The uppermost part of these sandstones border with a thrust zone against the overlying Storfjell Group. The clastic character of the sandstones disappears and they become increasingly quartzitic and granoblastically textured as the thrust zone is approached. The mineralogical composition is unchanged. It is not known whether the recrystallization was promoted by differential movements, by increased temperature, by a higher water content of the thrust zone or by these factors in combination, but it seems evident that it is in some way related to the thrusting. It follows from the above-mentioned facts that it is difficult to draw a precise boundary between unmetamorphosed sediments and rocks of the muscovite-chlorite subfacies. The boundary between this subfacies and the epidote-biotite or epidote almandine subfacies is partly coincident with the thrust plane above the Storfjell Group. South of Målselva, however, increasingly higher grade metamorphic rocks appear westwards in the Storfjell Group, the facies boundaries thus cutting obliquely across the Group in that area.

Within the present subfacies the association chlorite-muscovite-albite-quartz is very frequent in pelitic schist types. Epidote is an important constituent in some variants. The most conspicuous variations in mineralogical composition are found when moving across the strike and are therefore thought to reflect primary variations in the sedimentation. Chlorite is characteristically present in nearly all schists. This chlorite is evidently a product of progressive metamorphism as remnants of higher grade minerals have never been detected. This is in contrast to the relations found within the biotite and almandine zones, where retrogressive chloritization of ferromagnesian minerals is common.

In the quartzo-feldspathic rocks of the muscovite-chlorite subfacies clastic grains of feldspar (albite and microcline) and quartz are occasionally present. In most cases, however, these minerals are also assumed to have recrystallized. This is shown by the elongate or oval shape of the grains and by grain boundaries. The shape is obviously an adjustment to the local stress conditions; even large aggregates of muscovite flakes may have oval cross sections with undulatory extinction. Porphyroblastic growth of minerals has not been observed in any of these rocks, and it is believed that most of the larger feldspar and quartz grains have retained their original size, whereas their shapes have been changed by solution and redeposition according to the principle of Riecke. Growth of new minerals appears to be restricted to the "clay fraction," the most finegrained material, with the formation of muscovite (partly as sericite), chlorite, epidote, titanite, quartz etc. Reactions between minerals may, of course, have taken place. Ramberg (1952, p. 145) thinks that the reaction between K-feldspar and kaolinite, according to the equation

$$\text{potash feldspar} + \text{kaolinite} \rightarrow \text{muscovite} + \text{water} + \text{silica}$$

occurs at very low temperatures, perhaps during diagenesis. It is then, as emphasized by Barth (1962, p. 316) the relative proportions of kaolinite and potash feldspar that determine which of the two minerals will remain from this reaction. If potash feldspar outweighs kaolinite in the original sediment it seems probable that it will remain stable in the muscovite-chlorite subfacies. This will be possible only in quartzo-feldspathic sediments where the content of clay material is low. The presence of microcline in the quartzo-feldspathic schists and its complete absence from pelitic schists of the present subfacies within the map area is thus in accordance with the reasoning followed above. The reaction between potash feldspar and chlorite to form biotite has taken place at a higher temperature and belong to the epidote-biotite subfacies (epidote-amphibolite facies).

The deficiency in clay material which could react with the feldspars is a possible reason for the better preservation of clastic textures in coarse quartzo-feldspathic rocks as compared with the pelitic schists. It is probable, however, that the different mechanical properties of pelites and psammites are of greater importance. This is, for instance, demonstrated in the conglomerate locality in Sørdalen, Bardu. (Fig. 19 b and 20 b.) The feldspathic and conglomeratic bands have their clastic character relatively well preserved, whereas the micaceous and chloritic layers are strongly tectonized, slickensided and recrystallized with no sign of original clastic texture. Stresses seem to have been released mainly in the pelitic layers while the coarser clastics remained relatively unaffected.

Except for the relatively subordinate dolomites of the Storfjell Group, calcareous rocks are absent within the muscovite-chlorite subfacies. The assemblage dolomite + quartz + talc was found within one of the marbles. Actinolite (or tremolite) has not been observed in rocks of this facies.

The epidote-biotite subfacies

In terms of mineral zoning the present subfacies belongs to the biotite zone of pelitic rocks. Within the map area it has a great field extension. From Bardu westwards it covers much of the area underlain by the Rombak Group and is there restricted to rocks of this group. In eastern areas (see map, Fig. 2) parts of higher structural units are encountered in the epidote-biotite subfacies, for instance parts of the Narvik Group south of L. Altevann and parts of the Salangen Group in the northeastern corner of the map. A small part of the Storfjell Group is likewise belonging to this subfacies.

As shown on the facies map, boundaries of the epidote-biotite subfacies frequently, though not always, coincide with thrust boundaries.

In pelitic rocks the common mineral assemblages are biotite-muscovite-chlorite-albite-quartz and biotite-muscovite-epidote-albite-quartz. The main difference from the muscovite-chlorite subfacies is the presence of biotite. Most probably the biotite was formed by reaction between muscovite and chlorite, according to the equation:

(1) muscovite + $\text{chlorite}_1 \rightarrow$ biotite + chlorite_2 + quartz + water,

where chlorite_2 is more aluminous than chlorite_1.

In quartzo-feldspathic rocks another possibility of biotite formation is

(2) potash feldspar + chlorite → biotite + quartz + water.

Barth (1962) suggests that the reactions (1) and (2) both take place at the lower boundary of the epidote-amphibolite facies. Within the present area it would appear that reaction (1) has taken place at temperatures lower than those for reaction (2): As already mentioned the boundary between the chlorite and biotite zones crosses the Storfjell Group south of Målselva. The feldspathic schists west of this boundary, belonging to the epidote-biotite subfacies, do not contain any lesser amount of microcline than those metamorphosed in the chlorite zone. Table V of this paper also shows that there is no correlation between high biotite contents and low contents of potash feldspar. The chlorite content, however, is greatly reduced when biotite appears and in most cases chlorite is absent from quartzo-feldspathic rocks in this subfacies. At the same time the amounts of muscovite diminish. Table V serves to illustrate these relations. It is therefore reasonable to conclude that biotite formation was effected mainly according to equation (1) above, and that potash feldspars did not take part in the formation of biotite, at least not at this stage of metamorphism. We cannot, of course, extrapolate these observations claiming them to be general rules, because the temperature of reaction, especially of equation (1), is largely dependent on chemical relations, for instance the Fe/Mg/Al proportions of the chlorites.

The biotites of the epidote-biotite subfacies frequently have greenish brown colours in the Z direction in contrast with most biotites found at higher grades of metamorphism. Tilley (1925) in his work on metamorphic zones in the Scottish Highlands used what he calls the "common brown biotite" as an index mineral. According to this author "a green to brownish green pleochroic biotite arises at an earlier stage of metamorphism — — —," but he claims that this mineral is "not usually a constituent of the normal pelitic rock, but is characteristic of certain varieties of Green Beds in which this green biotite and epidote are common associates." In contrast to the relations found by Tilley, greenish biotites in the present area really are constituents of normal pelitic rocks. Other colour variants are not rare; the "common" brown type is frequent in parts of the subfacies, though no decisive evidence has been found that these colour variations represent changes in metamorphic grade.

The chlorites of the epidote-biotite subfacies occur in flakes of greater size than those of the chlorites of the muscovite-chlorite subfacies and also occasionally as porphyroblasts. Optical properties are variable and no systematic variations in these properties in relation to metamorphic grade have been found.

Epidote is present in many rocks of this subfacies, together with feldspar of albite composition.

Calcite marbles are common within the present subfacies. White or light grey tremolite is a conspicuous accessory mineral in many of these marbles. The assemblage calcite-quartz-tremolite with clinozoisite as an occasional member appears to be stable in the greater part of the subfacies.

The epidote-almandine subfacies

Parts of the Salangen, Narvik, Rombak and Storfjell Group belong to this subfacies which has a wide extension over the map area. The boundaries against higher and lower subfacies in part coincide with the thrust zones as shown in Fig. 2.

At the lower limit of the subfacies almandine garnet becomes stable. This isograd is very practical in mapping because garnets are easily detected and seem to develop in rocks of a fairly wide range of composition.

The most common mineral association in this subfacies is

almandine—biotite—muscovite—quartz.

Albite is stable with epidote to the upper limit of the subfacies and these two minerals may be added to the assemblage in many cases. Chlorite is relatively common together with garnet in the lower part of the epidote-almandine subfacies.

According to different authors (Ramberg 1952, Fyfe, Turner and Verhoogen, 1958) chlorites attain higher amounts of Al and Mg with increasing metamorphic grade. It is not evident in the present area that chlorites of the epidote-almandine subfacies are especially rich in Al. In most cases they have optical negative character which implies relatively low Al-contents if it is supposed that the Fe content is not high. The pale green colours and low refractive index ($n \angle 1{,}600$), indeed, seem to indicate a fairly low content of iron. This suggests that chlorites at this stage of metamorphism within the present area are high in Mg, low in Fe and low to moderate in Al content.

According to Ramberg (1952), the Mg/Fe ratio is higher in chlorites than in coexisting garnets. If garnet is formed from chlorite it is thus probable that some iron would have to be acquired to give the garnet a stable composition. A possible equation for garnet formation is therefore:

chlorite + iron oxide + quartz → almandine + water.

It is also probable that micas take part in the formation of garnets in many cases, for instance according to the equations:

muscovite + chlorite → biotite + almandine + water

and

biotite_1 + chlorite → biotite_2 + almandine + water.

In the latter equation biotite_2 must have a higher Mg/Fe ratio than biotite_1 if there is to be an increase in iron content of the almandine as compared with the chlorite taking part in the reaction. This would be in accordance with the decrease in iron content of biotites with increasing metamorphic grade as has been found by some petrographers (Barth 1936, Engel & Engel 1960). The relative and absolute reaction temperatures of the proposed equations are not known but they probably vary considerably, depending on such factors as H_2O-pressure, Mg/Fe ratio etc. It is therefore impossible to decide which of the above reactions actually occurred in the present rocks.

The association microcline—almandine—muscovite—biotite—epidote—albite—quartz has been observed in quartzo-feldspathic rocks of the Storfjell Group. Chlorite is absent.

In calcareous rocks tremolite or actinolite is present together with calcite and quartz, frequently also with clinozoisite.

The almandine-amphibolite facies

The boundary between the epidote-almandine subfacies and the almandine-amphibolite facies is not everywhere easy to map. Staurolite, which is the characteristic mineral of the lowermost subfacies of the almandine-amphibolite facies, is not common within the map area. Besides the possible appearance of staurolite, the main change at the boundary between the epidote-almandine and the staurolite-almandine subfacies' is the increase in An content of plagioclase (in equilibrium with epidote) from albite to oligoclase composition. The recognition of this change is dependent on microscope studies and is therefore not very suitable as a base for facies mapping. In some parts of the map area, however, the only possible way of determining the boundary has been to investigate a great number of specimens in powder or thin section. Kyanite is much more common than staurolite in the present area. A great part of the area underlain by almandine-amphibolite facies rocks evidently belongs to the kyanite-almandite-muscovite subfacies. These are the most highgrade rocks of the area. On the map, Fig. 2, the latter subfacies and the staurolite-almandine subfacies have been given the same symbol. Even though they are thought to represent slightly different metamorphic conditions it has not been possible to draw any boundary between them. In the following they are largely treated together as rocks of the almandine-amphibolite facies.

The stability of oligoclase and andesine in the present facies does not imply that more sodic plagioclases are totally absent. In rocks of very low Ca-content, as for example the analysed staurolite-bearing rocks (Table X, 2 and 3), albite is the only possible composition. This is shown by the absence of epidote and other calcium-bearing minerals. Schists with the assemblage staurolite-almandine-muscovite-(biotite)-(kyanite)-albite-quartz are found in close association with almandine-muscovite-biotite-oligoclase-epidote-quartz schists. The latter assemblage is also present within the areas where kyanite-bearing gneisses occur but the plagioclase there is occasionally of andesine composition. A common mineral assemblage in these gneisses is that of kyanite-almandine-muscovite-biotite-oligoclase/andesine-(epidote)-quartz. The relative stability ranges of staurolite and kyanite will be discussed later in this paper.

Within calcareous rocks diopside becomes stable in the high-grade parts of this area approximately parallel with the appearance of andesine in pelitic gneisses. This is in good agreement with the proposals of for instance Ramberg (1952), Barth (1962) and Turner & Verhoogen (1960).

Rocks of the almandine-amphibolite facies are common within the Narvik, Salangen and Niingen Groups. The varying chemistry of the schists is reflected in the mineral assemblages. Kyanite-bearing associations are much more common in the dominantly pelitic Narvik and Niingen Groups than in the Salangen Group where a higher Ca content generally prevents the development of Al-silicates.

Observations on the crystallization of the principal metamorphic minerals in relation to the deformation phases

In the petrographical descriptions we have to some extent dealt with textures and the interrelationships of minerals. There is abundant evidence favouring the view that the present mineral assemblages are the result of more than one phase of metamorphism. The purpose of the following chapter is to consider and summarize those features which seem to be of importance in an understanding of the metamorphic history of the area.

In this respect the relations between structures and the growth of minerals are of special significance when age relationships and metamorphic phases are to be investigated. Both microtextures and field relationships will be considered.

Q u a r t z is a common mineral in the meta-sediments at all the represented stages of metamorphism. In parts of the autochthonous sequence it is found as clastic grains. With increasing deformation the clastic quartz rapidly recrystallizes. In the metamorphic rocks quartz occurs as evenly distributed grains, as lamellar aggregates parallel to the schistosity (see Fig. 34 and 35) and in cross-cutting veins. Undulatory extinction is commonly observed. According to some experimental studies by Carter, Christie and Griggs (1964) undulatory extinction in quartz is not the result of crushing but caused by recrystallization during deformation. Applied to the present area this signifies a syntectonic growth for most quartz grains. This does not, of course, imply that all quartz is of one age. On the contrary, several generations can frequently be recognized, both under the microscope and in the field.

The youngest generation of quartz is present in veins crossing the schistosity and in lenses from which small veins protrude into the adjoining schists. Lenses or veins of considerable size are known, some of which are utilized or are being investigated for economical purposes. The quartz bodies are located in the neighbourhood of thrust zones, especially where these are situated in the vicinity of meta-sedimentary quartzites. It thus seems probable that the quartz represents silica mobilized and segregated from the sediments during the thrusting episode. Small amounts of calcite and chlorite are found associated with the quartz. Other examples of solution and redeposition of silica are demonstrated in innumerable localities, for instance in connection with boudinage structures, rotation of porphyroblasts etc. Some of these features may well be related to the thrust movements.

The age of the quartz found as lamellar aggregates and scattered grains in the schists is not easy to determine. Much of it probably belong to the period of strong, regional metamorphism which is thought to be contemporaneous with the main (first) folding episode.

C h l o r i t e is regarded as having been formed by regional, progressive metamorphism when it occurs together with minerals with which it is commonly in stable equilibrium and if signs of pseudomorph development are absent. Such chlorites occur in schists from the lowest grade up to the lower part of the epidote-almandine subfacies. Of equally great interest is the chlorite of diaphthoretic origin. This is found as pseudomorphs after garnet and biotite and as transformation product of amphibole. The retrograde chlorites appear to be related to the late thrust movements, as evidenced by their abundance along thrust planes. Amphibole porphyroblasts, for instance, are broken and altered to chlorite along the cracks (Fig. 37). The chloritization may thus be dated as later than the growth of amphibole porphyroblasts, which in its turn is later than the main metamorphic (and folding) phase. The occurrence of chlorite in late quartz veins is probably of about the same age. Though the retrograde chlorite is most frequently developed in thrust zones, it has also been observed where no sign of thrusting is present. Areal or stratigraphical limits of the chloritization are impossible to draw.

P l a g i o c l a s e. In rocks with clastic textures plagioclases are relatively rare; where present they are frequently saussuritized and their composition is consequently difficult to determine. In the regionally metamorphosed schists albite is the common feldspar up to the almandine-amphibolite facies. The albites are always present as small grains, porphyroblastic development being restricted to oligoclase and more basic feldspar. Inclusions of albite are found in micas, garnet, staurolite and amphiboles.

In plagioclases of oligoclase or andesine composition porphyroblastic growth is common in the present area. The feldspars occur as augen and irregular porphyroblasts. Granitic to trondhjemitic veins or lenses are frequently associated with the augen-gneisses.

The porphyroblastic plagioclases are mostly 1 to 3 centimetres in their greatest dimension but may occasionally attain a larger size. Inclusions are numerous. Where micas are enclosed they frequently have the same orientation as micas outside the feldspars, that is parallel to the foliation. This feature indicates that the foliation was already present when the porphyroblastesis occurred. The textures thus point to a late-tectonic or post-tectonic growth of the feldspars with respect to the first folding phase.

The composition of porphyroblastic plagioclases varies from about An_{23} up to An_{40}. Zonal growth is often encountered; it is mostly an irregular zonal arrangement with the core of a more basic plagioclase than that of the margins (see the petrographical description). As epidote minerals are also commonly present in the groundmass, the zoning probably indicates a decline in temperature during formation of the porphyroblasts. This is another feature which suggests that the feldspar porphyroblastesis outlasted the main phase of regional metamorphism and took place when stresses were moderate and the temperatures still high, but declining. It is also possible that the porphyroblast formation belongs to a separate metamorphic phase subsequent to the first regional metamorphism. In both cases the result would probably be the same.

The relationship between granite (or trondhjemite) bodies and porphyroblastic growth of feldspar in the gneisses has already been briefly discussed. Consequently the problem of the origin of the granite material will not be commented on here.

Potash feldspar. Apart from its occurrence in granitic veins K-feldspar is absent from pelitic rocks within the area. In psammitic sediments microcline is a common constituent, though mostly in minor amounts. The recrystallization of microcline into ovoid grains in psammitic rocks of the Storfjell Group has been described earlier, this process seemingly having taken place without involving the feldspar in any reaction with other minerals. Transformation of potash feldspar, for instance into muscovite, appears to be of little importance in the psammitic schists. In pelites any clastic microclines present in the original sediments would react with excess alumina and form muscovite (sericite) at an early stage of metamorphism. Porphyroblastic growth of potash feldspar has nowhere been observed.

Epidote minerals, of which clinozoisite and epidote are the most common, are present in schists throughout the area frequently as rods or stringers lying in the schistosity planes. Orthite is a relatively frequent accessory in pelitic rocks, but is of little value in a discussion of the metamorphism. Zoisite, present but in a few cases, is also relatively unimportant. Clinozoisite is especially abundant in calcareous rocks, whereas epidote is more frequent in pelites. The clinozoisite is commonly filled with calcite and is probably formed by reaction between calcite and some silicate mineral, for instance chlorite or the anorthite component of plagioclase (see Ramberg, 1952, p. 53 and 140).

(1) Calcite + chlorite + silica → clinozoisite + actinolite + water + carbon dioxide.
(2) Calcite + anorthite + water → clinozoisite + carbon dioxide.

The association of clinozoisite and actinolite in many calcareous rocks suggests that the reaction of equation (1) above might have taken place in some cases.

Judging from thin-section study, saussuritization is not extensive in the pelitic schists. Epidote minerals mostly occur as separate grains: their occurrence as inclusions in plagioclase is no more common than in other minerals. The abundance of epidote in some of the low-grade rocks as compared with parts of the same rocks of higher metamorphic grade indicates a progressive digestion of epidote with increasing metamorphism. Much of the Ca and Al from the epidote could naturally have entered the plagioclases as more basic compositions became stabilized, but it is also likely that some of it entered other Ca-bearing minerals, as for example amphiboles in certain schists.

Some plagioclase porphyroblasts are saussuritized in the core showing that they were originally more basic in their central parts but have since been adjusted to the changing P-T conditions. The lack of saussuritization in some of the zonal plagioclases is best explained as a kind of "frozen" dis-equilibrium.

Epidote minerals are occasionally associated with retrograde chlorite.

M u s c o v i t e, in part sericite flakes, is present from the lowest to the highest grade of metamorphism within the area. In the low-grade rocks it appears together with chlorite, evidently formed from the clay material of the sediments. In feldspathic psammites (Storfjell Group and feldspar-bearing quartzites of the Narvik Group) formation of sericite could also occur by transformation of the feldspars, especially the potash feldspars. Such sericitization has been described from sparagmitic rocks of Southern Norway by Barth (1938) and Oftedahl (1943). The sericitization could be a direct hydration of the K-feldspar or, as proposed by Oftedahl, a reaction between the feldspar and clay material of the groundmass. A participation of the clay material seems necessary to explain the high percentage of sericite in some of the sparagmites. In the present map area there is little evidence in favour of a sericitization of potash feldspars. There is also no correlation between large amounts of sericite and low contents of potash feldspar (see Table V). Sericitization in plagioclases has been observed in a number of cases but this process cannot account for the high percentages of sericite in some feldspathic rocks. A growth of sericite directly from the clay matrix must therefore be supposed. Muscovite formation at the higher grades of metamorphism is probably largely a process of rearrangement of the earlier formed sericite into larger flakes.

Fig. 34. Typical Rombak schist, 3.4 km E. of Treldal showing pronounced parallel texture due to orientation of biotite, epidote and quartz aggregates (35x).
Typisk Rombakskifer, 3,4 km øst for Treldal med utpreget parallelltekstur som skyldes orientering av biotitt, epidot og kvartsaggregater (35x).

Most muscovite flakes show a pronounced parallelism, thus contributing to the foliation present in all meta-sedimentary rocks in the area (Fig. 35). It seems highly probable that most of the muscovite was formed syn-tectonically during the main regional metamorphism. A possible exception is the muscovite developed at the expense of kyanite in some of the gneisses (Fig. 39). An eventual explanation is that the muscovitization was promoted by the reaction with granitic material, schematically according to the equation:

Fig. 35. Porphyroblastic mica schist from the lower part of the Narvik Group, Soløyheia, Lavangen. The biotite porphyroblasts are partly growing across the pre-existing planar structure (orientation of muscovite, biotite, epidote and quartz aggregates) (35x.)
Porfyroblastisk glimmerskifer fra den undre del av Narvikgruppen, Soløyheia, Lavangen. Biotittporfyroblastene vokser delvis på tvers av den eldre planstruktur (orientering av muskovitt, biotitt, epidot og kvartsaggregater) (35x).

$$\text{potash feldspar} + \text{kyanite} + \text{water} \rightarrow \text{muscovite} + \text{quartz.}$$

In this case the transformation of kyanite may be roughly contemporaneous with the formation of plagioclase porphyroblasts as this latter process has also been shown to be possibly related to the granitic veins and dykes. Another possibility is that the muscovitization is a late event, for instance coeval with the chloritization during the thrust movements. The latter possibility is contradicted by the absence of chloritization phenomena in the kyanite-bearing gneisses. With the present state of knowledge, therefore, the first alternative, outlined above, is favoured.

B i o t i t e. Some features of biotite have already been briefly mentioned, particularly the variations in colour in the different sedimentary groups. A parallel arrangement of biotite flakes is equally as frequent as in the case of muscovite, (Fig. 34) and intergrowths of the two mi-

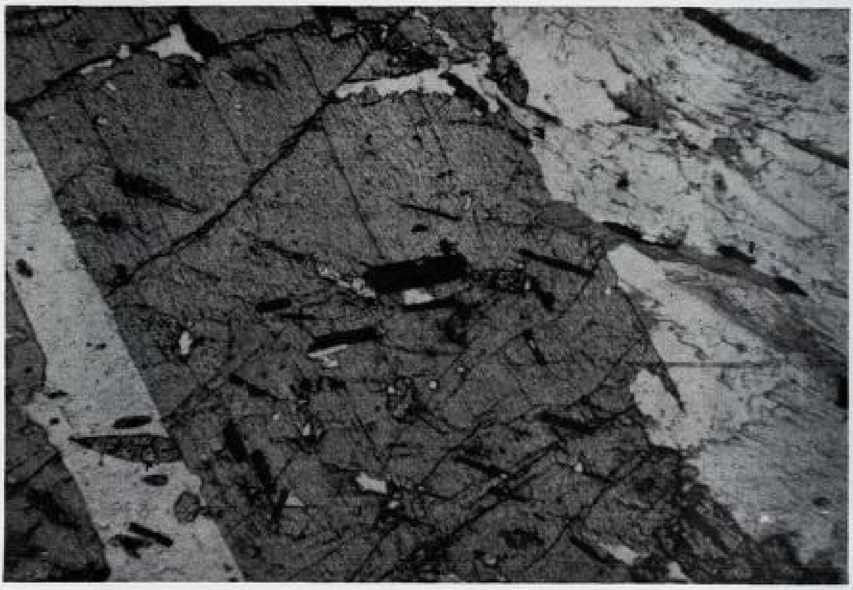


Fig. 36. Amphibole porphyroblasts growing across the planar structure. The earlier structure is shown by parallel arrangement of micas and ore grains, the latter is present as inclusions within the porphyroblasts. Salvasskarfjell (35x).

Amfibolporfyroblaster som vokser på tvers av planstrukturen. Denne sees som en parallellorientering av glimmer og ertskorn. Ertsen sees også som orienterte inneslutninger i porfyroblastene. Salvasskarfjell (35x).

nerals are, in fact, rather common. Porphyroblastic biotite is occasionally found, (Fig. 35) for instance in some calcite-bearing schists. The porphyroblasts show red-brown colours under the microscope, pale in the calcareous rocks, darker in normal pelites. Usually these biotites, too, are orientated parallel with the schistosity. In a few cases, however, it has been observed that biotite porphyroblasts grow across the schistosity in rocks where muscovite and smaller flakes of biotite are parallel to this schistosity. (Fig. 35) The porphyroblasts are clearly younger than the schistosity. Whether or not they are related to any deformation phase is, as yet, unknown. As these schists have partly been observed in close association with garben schists it is possible that the porphyroblastesis was simultaneous in both rock types.

Garnet, mostly present as porphyroblasts, is common in pelites of the epidote-almandine subfacies and in higher grade rocks. Idioblas-

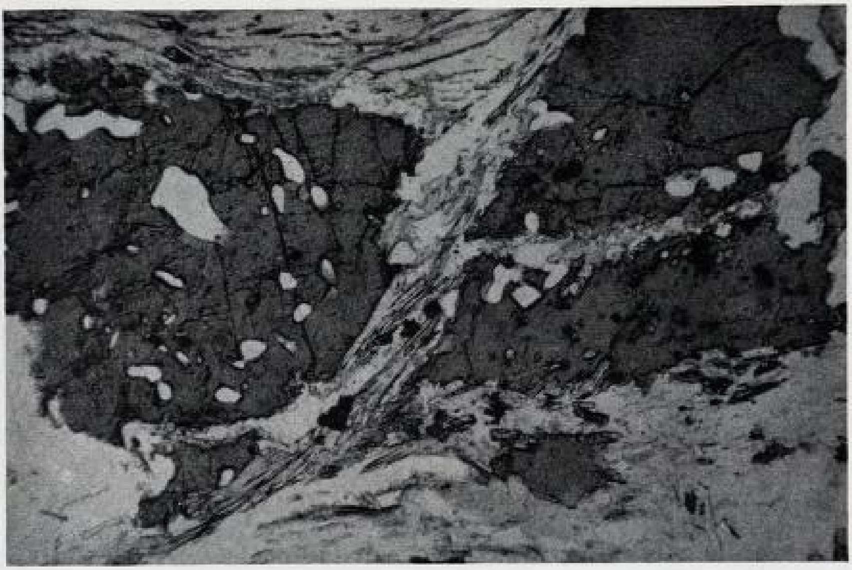


Fig. 37. Amphibole porphyroblasts in Seternes schist broken by late tectonic movements. Crackfillings of chlorite + quartz (35x).

Amphibolporfyroblaster som er brutt istykker ved sene tektoniske bevegelser. Sprekkefyllinger av kloritt + kvarts (35x).

tic development is a conspicuous feature in small-sized garnets, up to a few millimetres in diameter. Signs of rotation have not been observed in the case of these garnets. With increasing size the crystal outlines become irregular and inclusions of quartz, ores, micas etc. are numerous. In some cases helicitic inclusions show that the garnets rotated during their growth. In other cases the inclusions are arranged along straight lines deviating in trend from the groundmass schistosity, thus indicating rotation after the crystalloblastesis. Such garnets are evidently younger than the development of the schistosity represented by the inclusion fabric but older than the last differential movements. It is probable that most garnets belong to the main regional metamorphism and deformation phase, though some garnets may belong to a later stage, namely the period between the first folding phase and the last thrust movements. A two-phase growth of some garnets has also been observed and has already been described in an earlier section (petrography of garnet-mica schists, Narvik Group). This last mentioned observation confirms the assumption of two generations of garnet porphyroblasts.

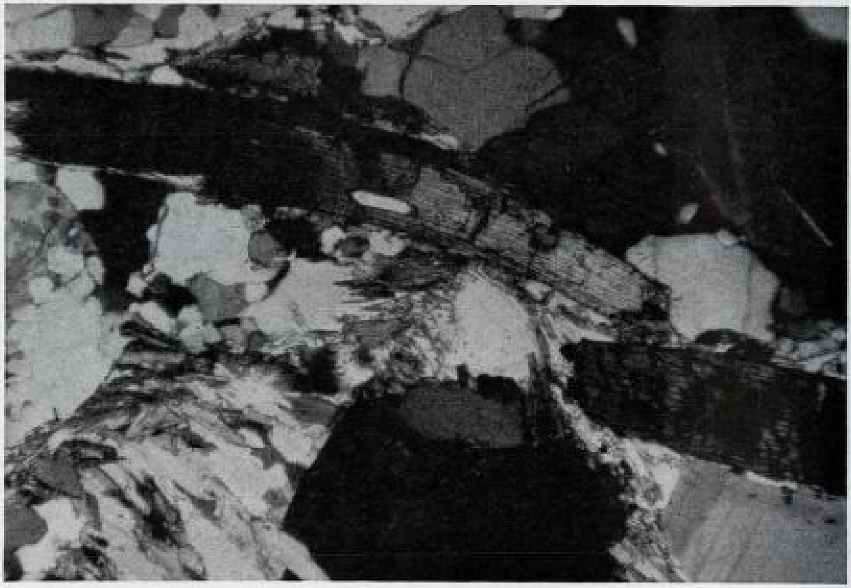


Fig. 38. Kyanite and garnet in plagioclase gneiss, Lemman, Skjomen. One of the kyanites is bent and is showing uneven extinction. The kyanite prisms are parallel or subparallel to the foliation (+n. 35x).

Disthen og granat i plagioklas-gneis, Lemman, Skjomen. Ett av disthen-prismene er bøyet og viser ujevn utslukning. Prismene er parallelle eller sub-parallelle med foliasjonen (+n. 35x).

Amphiboles are uncommon in the pelitic schists, though in calcareous rocks tremolite and actinolite are frequently observed. In the few cases in which amphiboles (actinolite, common hornblende) are present in pelites they are found as small needles, usually arranged in the schistosity planes. They probably belong to the main regional phase of metamorphism. An exception appears to be that of the porphyroblastic amphibole in rocks described as Seternes schist or garben schists. (Figs. 36, 37) The mineral association of these rocks, amphibole—muscovite—biotite—garnet—chlorite—(plagioclase)—epidote—quartz, indicates a state of disequilibrium (see f. inst. Strand 1958). The reason for this is probably polymetamorphism, as the minerals were divided, mainly on the basis of microscopic work, into three groups: (1) Syn-tectonic. (2) Post-tectonic and (3) Diaphtoretic minerals (see pp. 90—94). The amphiboles belong to group (2), the porphyroblastic, post-tectonic minerals, which means that they postdate the main regional metamor-

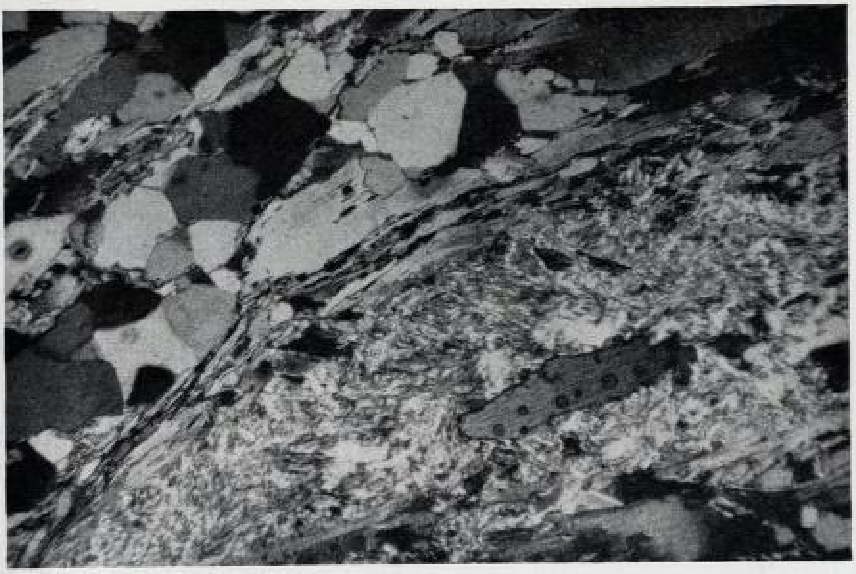


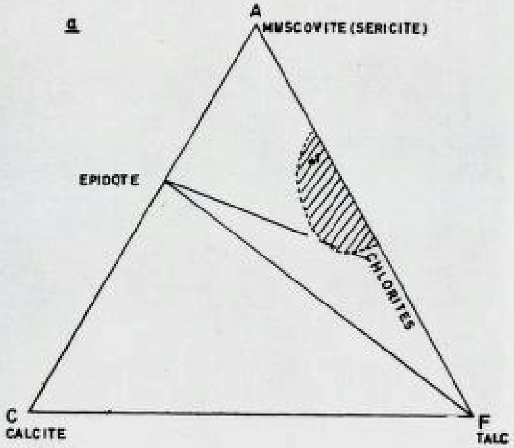
Fig. 39. Muscovitization of kyanite. Only a small part of the kyanite is left in the centre of the muscovite masses. Other minerals present are plagioclase, quartz and biotite. Kvernmo, Gratangen (+n. 35x).

Muskovittisering av disthen. Bare en liten del av disthenen er tilbake i midten av muskovittmassene. Andre mineraler i slipet er plagioklas, kvarts og biotitt. Kvernmo, Gratangen (+n. 35x).

phism. Further discussion of possible causes of the formation of amphibole porphyroblasts is given in a later section where mineral associations and rock chemistry are compared.

Tourmaline. The zonal growth of tourmalines as described from some schists, may indicate a two-phase development. As the significance of the colour differences is not known further conclusions cannot be drawn from this feature.

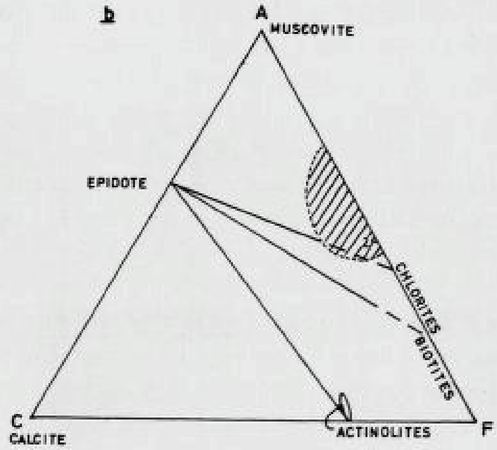
Staurolite is present as somewhat elongate but irregular prisms containing numerous quartz inclusions. The general outline of the individuals frequently follows the limbs and curves of microfolds belonging to the first deformation episode, but even the curved prisms show uniform extinction in all parts. This remarkable feature cannot be explained by a bending of the staurolites subsequent to their formation, nor can it be attributed to rotation during the growth. The only



Figs. 40 a—d. ACF diagrams showing minerals stable in the different metamorphic facies represented in the map area. Numbers show the positions of analysed rocks (see Table XXI). Hatching = field of analysed rocks.

(a) Muscovite-chlorite subfacies. (b) Biotite-epidote subfacies. (c) Almandine-epidote subfacies. (d) Staurolite-almandine and kyanite-almandine-muscovite subfacies.

ACF diagrammer som viser stabile mineraler i de metamorfe facies som er representert i området. Tallene refererer til analyserte bergarter (se tabell XXI). Skravert = området for de analyserte bergarter.



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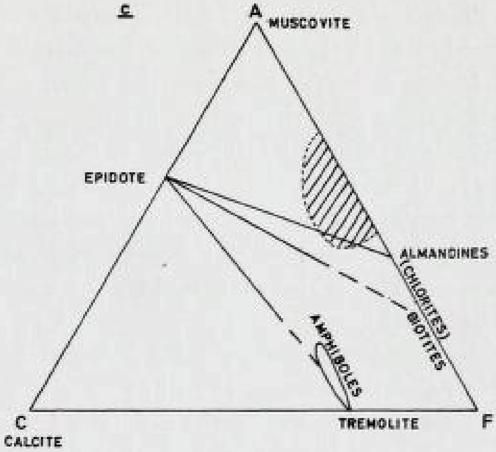

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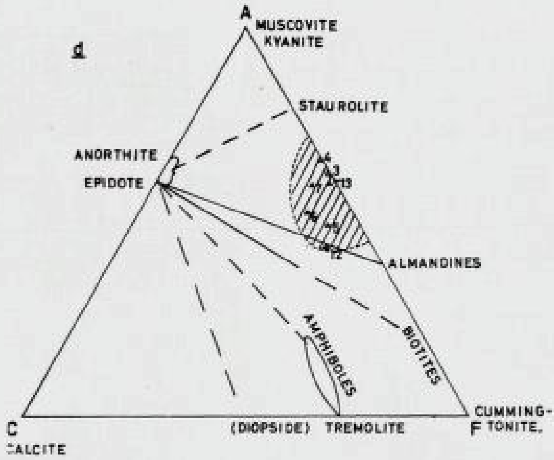

feasible explanation seems to be that the staurolite was formed after the microfolding, crystallization being essentially mimetic with individual staurolite crystals following, and being strongly influenced by, the curves of the pre-existing microstructures. Some staurolites, however, grow across previous microfolds, thus supporting a post- or late-tectonic formation.

Slight chloritization of the staurolite has been observed in one thin section.

K y a n i t e needles are common in the gneissic rocks. As already described the needles are oriented more or less parallel to the main, first fold axes. Microscopically, however, the kyanite prisms are frequently seen to be curved with the extinction appearing to pass through the crystal from one extremity to the other as the microscope stage is rotated (Fig. 38). There is no sign of a mechanical breaking of the crystals, and the bending is thought to have been more or less contemporaneous with crystallization. The growth of kyanite is therefore probably syntectonic. This is also suggested by the tendency towards a linear arrangement of the needles.

The subsequent muscovitization of some kyanites has already been commented on in an earlier section. (Fig. 39)

The main conclusions concerning the relations between growth of the respective minerals and the deformational and metamorphic phases as outlined in the preceding discussion, are summed up in Table XX.

T a b l e XX

Relations between formation of some of the principal minerals in pelitic rocks and deformational/metamorphic phases. (+: mineral formed, —: mineral not formed).

Deformation phase	Main (first) folding Development of schistosity	Static period	Thrusting Weak folding
Metamorphic process	Main regional metamorphism	Granitization? Porphyroblastesis	Retrograde metamorphism
Quartz	+	+	+
Chlorite	+	—	+
Epidote	+	—	+
Muscovite	+	(+)[1])	—
Biotite	+	+	—
Plagioclase	+	+	—
Garnet	+	(+)[2])	—
Amphibole	+	+	—
Staurolite	(+)[3])	(+)[3])	—
Kyanite	+	—	—

1) The age of the muscovitization of kyanite is uncertain.

2) The last phase of garnet formation is of uncertain age.

3) Whether staurolite development belongs to the later stages of the main regional metamorphism or to a later period is undecided.

Comparison of chemistry and mineral assemblages in the meta-sediments Discussion of particular assemblages

Introduction.

A thorough discussion of relations between rock chemistry and mineral associations would have demanded numerous chemical analyses in addition to those presented in this paper. Only nine analyses are available from the areas where metamorphic facies have been investigated by the present writer. Seven of these analyses are of schists from the almandine-amphibolite facies areas. The following discussion is therefore mainly concerned with some mineral assemblages from this facies.

ACF-diagrams depicting the minerals regarded as stable in the respective metamorphic zones or subfacies, as well as the positions of analysed rocks are shown in Figures 40 a—d. Plotting is in accordance with the rules proposed by Barth (1962, p. 311).

The ACF diagrams.

Typical representatives of the muscovite-chlorite subfacies have not been analysed. The analysis, no. 1, plotted in Fig. 40 a belongs to the Hyolithes Zone rocks and should probably not be considered as a metamorphic rock. Clastic grains of oligoclase (?) and microcline occur together with muscovite (sericite) and chlorite. The position of this analysis in the diagram indicates epidote to be the stable calcium mineral. In the present rock, however, epidote is absent while calcite is present in small grains. The position of the analysed schists, and probably the bulk of pelitic rocks, is indicated by the hatched area on the ACF-diagrams. It can be seen that most rocks would tend to have produced the association epidote—chlorite—muscovite at the lower grades of metamorphism. Additional minerals not represented in the diagram are quartz and albite. In fact, the above-mentioned association is very common in the pelitic members of the Storfjell Group.

The next diagram, Fig. 40 b, serves to illustrate the common minerals of the biotite—epidote subfacies. Because of the presence of K_2O as an important constituent the relative positions of biotite and chlorite are misleading. The associations biotite—chlorite—muscovite—albite—quartz and biotite—epidote—muscovite—(chlorite)—albite—quartz are common in this facies; which of them will occur is dependent on the calcium content. The analysed rock, no. 2, is poor in CaO; for some reason this small amount of CaO is present as calcite, not as epidote. As shown in Table XXI, this is also the case with some schists of higher grade.

Minerals of the epidote-almandine subfacies are shown in Fig. 40 c. No analyses are available from this facies. The association almandine–biotite–muscovite–quartz–albite–epidote occurs frequently. This is in good agreement with the chemistry of the analysed rocks from the other subfacies, if the limitations of the ACF diagram are taken into account. (K_2O, Na_2O not represented). Chlorite is present only in the lower part of the subfacies. In some calciferous schists of the Rombak Group (Table VI) the association amphibole–biotite–epidote–quartz is present, partly with calcite as an additional component.

The remaining analyses, seven in number, are from the almandine–amphibole facies areas (Fig. 40 d). Six of these fall within the epidote–almandine–staurolite triangle. Because of the presence of K_2O, muscovite and biotite are to be expected as additional minerals together with quartz and oligoclase feldspar. Staurolite is however, present in only two of the rocks (3 and 4). Kyanite is found as small remnants (?) in these same two rocks, while in two of the other analysed specimens kyanite is an important constituent (7 and 13). In the remaining two schists (5 and 6) within the epidote–almandine–staurolite triangle amphibole is present as porphyroblasts. The association amphibole + muscovite is not to be expected taking into account the position in the diagram. This relationship as well as the staurolite–kyanite stability relations, are discussed in the following sections.

One analysis, no. 12, falls within the almandine–epidote–biotite triangle. These three minerals are also frequent in the schist but with calcite, muscovite, quartz and oligoclase as additional members.

Cummingtonite has been detected in one case only, in a special iron-rich skarn rock found in connection with the siderite iron ore at Rubben, Kirkesdalen. It is associated with almandine garnet, diopsidic pyroxene and quartz.

Table XXI

Mineral assemblages of the nine rocks[1]) plotted in diagrams 40 a—d.

1. Quartz-oligoclase (?) —microcline—sericite—chlorite-(calcite).
2. Quartz-albite-muscovite-biotite-chlorite-(calcite).
3. Quartz-albite-muscovite-biotite-almandine-staurolite-(kyanite).
4. Quartz-albite-muscovite-almandine-staurolite-(kyanite).
5. Quartz-oligoclase-muscovite-biotite-chlorite-amphibole-(epidote).
6. Quartz-oligoclase-muscovite-biotite-chlorite-amphibole-almandine-(epidote)-(calcite).
7. Quartz-oligoclase-muscovite-biotite-almandine-kyanite-(epidote)-(calcite).
12. Quartz-oligoclase-muscovite-biotite-almandine-epidote-calcite.
13. Quartz-oligoclase-muscovite-biotite-almandine-kyanite-(epidote)-(calcite).

1) The numbers refer to Table XVIII. Further data for number 1. in Table IV; no. 2, Table VII; nos. 3—7, Table X; no. 17, Table XIII no. 13, Table XVI.

The stability relations of kyanite and staurolite.

One of the more interesting questions pertaining to the almandine-amphibolite facies is whether any differences in metamorphic grade between staurolite-bearing and kyanite-bearing rocks really exist or whether the mutual relations are governed by chemical variations in the host rock. Most authors seem to agree that some differences in metamorphic grade are present between schists with staurolite, and those with kyanite, in the sense that kyanite appears to indicate a slightly higher grade than staurolite. Examples of kyanite appearing at lower stages than staurolite have, however, been reported, for instance by Francis (1956). Most authors claim that certain chemical conditions must be fulfilled, especially concerning the formation of staurolite; kyanite seems to have a wider chemical range of stability. With regard to the special chemical conditions favouring staurolite growth there is, however, little agreement between the different authors.

Barth (1936) expresses the view that the relationship between kyanite and staurolite is governed by variations in the water content of the rocks. In a table of the stability ranges of some metamorphic minerals staurolite is shown to be stable both in "facies of cyanite" and in "facies of sillimanite gneisses," whereas kyanite is stable in the first one only and is replaced by sillimanite in the higher facies. Williamson (1953) investigated chloritoid- and staurolite-bearing rocks from Kincardineshire, Scotland. He found that "a $Fe_2O_3 : Al_2O_3$ ratio of about 0,40 is critical for the formation of chloritoid and staurolite." If this value "exceeds a figure somewhere between 0,47 and 0,55 staurolite cannot form." A minimum temperature was also found necessary; below this temperature chloritoid will form. The conclusions reached by Williamson were based on chemical analyses of eleven rocks of which only three were staurolite-bearing (one with both staurolite and kyanite).

Ellitsgaard—Rasmussen (1954) described a staurolite schist from Eqûtit, West Greenland and discussed the stability relations between staurolite and the minerals with the composition Al_2SiO_5. He plotted analyses from the literature in a A-K-FM-H(H_2O) diagram and in six of seven cases a connection between the position in the diagram and the mineral association was found. He states that "biotite is the first mineral to adapt itself to a composition relative to the P-T conditions." There must therefore be (Fe, Mg)O and H_2O left in suitable amounts after the formation of biotite if staurolite is to appear. If these conditions are not fulfilled some Al_2SiO_5 mineral will appear in the Al-rich rocks. Francis (1956) discussed kyanite-staurolite relations in connection with facies boundaries. Referring to Suzuki, Williamson and Ellitsgaard—Rasmussen he states that as a pre-requisite to staurolite formation, both alumina and iron must be high, but total iron (as Fe_2O_3) must not exceed about half the alumina figure. The lime content of staurolite-bearing rocks is low and potash is usually, but not always, low, according to Francis.

Vogt (1927) described the association staurolite—almandine—kyanite in the Sulitjelma area and states that according to the equation:

$$\text{quartz} + \text{staurolite} \rightarrow \text{almandine} + \text{kyanite} + \text{water}$$

the three minerals are not stable together provided that SiO_2 and H_2O are present in sufficient amounts. It is also stated that the right-hand side of the equation is the high-temperature association.

Snelling, in two papers (1957, 1958), dealt with rocks bearing kyanite and staurolite from the Scottish Highlands. His investigations are concerned more especially with the relations of chloritoid and biotite, but the chemical analyses are nevertheless of interest in the following discussion.

In order to evaluate the various proposals outlined above a number of analyses from the literature have been compiled, and listed together with analyses of kyanite- and staurolite-bearing rocks from the present area in Table XXII. Instead of listing all the weight percentages of the different oxides the Niggli values, H_2O contents, total iron content and the ratios $\frac{FeO + Fe_2O_3}{Al_2O_3}$ are compared. These cover the chemical characteristics thought to be of prime importance by earlier investigators. The average shale composition as given by Clarke (1924) is presented for comparison in column 17 and the last column of the table shows the variation range of nine pelitic rocks from the present area. The mineralogical composition of these rocks is shown in Table XXI and further data are presented in Table XVIII and in the petrographical descriptions.

Numbers 1—10 of the table are staurolite-bearing rocks, 11—15 are kyanite-bearing and no. 16 has both minerals in significant amounts.

Table XXII shows the high al values of the rocks in question; all of them are above the average shale value and within the present area (columns 9—12) they are the rocks highest in al. An al value above 40 appears necessary for the formation of kyanite and staurolite. There is no regular difference between staurolite- and kyanite-bearing rocks, in considering the al values. The fm values may be higher or lower than the value of the average shale. The rocks from the present area lie within the same range as schists devoid of staurolite and kyanite, the highest values occurring in rocks not carrying these minerals. There is no significant difference between staurolite- and kyanite-bearing schists.

A low c value is typical for all the sixteen rocks; all are considerably below the average shale value. In the present area the rocks in question are the poorest in calcium, especially the two staurolite schists. All rocks of Table XXII taken into account, kyanite- and staurolite-bearing schists vary approximately within the same limits of c values.

The alk values group around the average shale value, except for small variations. Within the present area they are not different from the other pelitic schist types.

The k values vary from 0,28 to 0,83 in staurolite-bearing rocks and between 0,36 and 0,66 in kyanite schists. They thus seem to be of little or no significance or assistance to a solution of the problem.

Concerning the mg values the rocks of the present area show little variation between the analysed schist types. Except for one of the analyses all values lie below the average shale value. The range of variation in staurolite/kyanite rocks is 0,39—0,47. In rocks from other areas the mg values are lower—between 0,28—and 0,37 in staurolite-bearing rocks and from 0,17 to 0,37 in those which are kyanite-bearing; all values are thus considerably below the average shale value.

Except in one case the ratios $\frac{FeO + Fe_2O_3}{Al_2O_3}$ all lie between 0,30 and 0,46, with a mean about 0,38 (no. 5 excluded). The ratios thus group around that of the average shale and are not far from the values found to be "critical" by Williamson (1953). The value 0,46 may be near to the upper limit (a value between 0,47 and 0,55 was suggested by Williamson). The lower limit is less well defined, and the value 0,19 of column 5 may be incorrect. (As a too low Fe_2O_3 value will result in a correspondingly higher Al_2O_3 value according to the classical methods of chemical analysis, the ratio $\frac{FeO + Fe_2O_3}{Al_2O_3}$ will be lowered considerably with only a relatively small error.) On the other hand, low ratios have been reported from other areas too, according to Williamson.

Total iron content is relatively high in most rocks. The mean for staurolite-bearing rocks is 9,10 with a variation range from 6,49 to 12,06. Corresponding values for kyanite-bearing schists are 8,53 and 7,46—9,89. Practically all values are thus above that of the average shale (6,5). Comparing the data for kyanite- and staurolite rocks there seems little reason to believe that high iron contents favour staurolite at the expense of kyanite. Several kyanite-bearing rocks are higher in iron than the staurolite schists with the lowest values.

The water content varies considerably. The lowest values are found in kyanite-bearing rocks, thus apparently supporting the opinion of Barth and Ellitsgaard-Rasmussen concerning the significance of the water content. On the other hand, the kyanite-bearing schists 13, 14 and 15 are as high in H_2O as any of the staurolite schists. The lowest value, 1,29 per cent, is found in no. 16, the specimen containing both staurolite and kyanite. It is interesting to note the high water content in no. 5, the rock with the especially low $\frac{FeO + Fe_2O_3}{Al_2O_3}$ ratio. The possibility that the formation of staurolite was promoted by a high water content in this rock, where the chemical relations otherwise were not favourable(?), must be considered. On taking into account all the analyses, however, the importance of water content seems doubtful.

The conclusions which we may draw from the preceding discussion are as follows: High al values and low c values are features common to kyanite- and staurolite-bearing schists. The mg values are relatively low, but not exceptionally low as compared with the other pelitic schists in the present area. The k, alk and fm values seem to be of little importance in the development of staurolite and kyanite. $\frac{FeO + Fe_2O_3}{Al_2O_3}$ ratios vary between 0,30 and 0,46; higher and lower values appear unfavourable. The importance of the water content is doubtful, as the various evidence is somewhat contradictory. The necessity of high iron contents for the formation of staurolite as claimed by different authors is to some degree disproved by the present analyses. It is probable, however, that as a general rule a high iron content is favourable.

Concerning the interrelations of kyanite and staurolite the table shows that no special chemical feature is present in staurolite schists that is not also present in most kyanite-bearing schists, and vice versa. An exception is the low water content of the two kyanite-bearing rocks from the present area. The conclusion that the appearance of one or the other of the two minerals, or of both, depends chiefly on the pressure-temperature conditions seems unavoidable. As mentioned previously in this paper thin sections appear to indicate that kyanite was formed syn-tectonically. Staurolite was probably formed either late syn-tectonically or post-tectonically, at a later stage than kyanite. It is thus probable that conditions were somewhat different during growth of the two minerals. Pressure and possibly also temperature must be supposed as having been a little lower during formation of staurolite than during crystallization of kyanite. This view is in accordance with the subfacies divisions of Turner & Verhoogen where the staurolite-almandine subfacies belongs to a slightly lower grade of metamorphism than the kyanite-almandine-muscovite subfacies.

The porphyroblastic amphibole schist assemblage.

The main features of these rocks have been dealt with in the petrographical description and are as follows: The rocks form extensive layers intercalated with mica schists, occasionally also with amphibolites.

Table XXII

Niggli values, water content, total iron and ratio $\frac{FeO+Fe_2O_3}{Al_2O_3}$ *of some kyanite- and staurolite-bearing rocks from different areas.*

	1	2	3	4	5	6	7	8	9	10
al	50,1	41,7	50,8	45,4	57,2	40,2	42,8	45,9	43,5	46,0
fm	34,9	39,1	29,9	31,3	24,2	38,9	38,8	39,2	42,0	41,0
c	2,9	2,8	4,0	6,8	2,5	4,1	5,9	3,9	1,5	1,0
alk	12,1	16,4	15,3	16,5	16,1	16,8	12,6	11,0	13,0	12,0
k	0,28	0,50	0,67	0,49	0,44	0,47	0,43	0,31	0,83	0,64
mg	0,28	0,35	0,26	0,35	0,37	0,33	0,31	0,31	0,47	0,39
$\frac{FeO + Fe_2O_3}{Al_2O_3}$	0,41	0,36	0,38	0,30	0,37	0,46	0,45	0,41	0,37	0,38
$FeO + Fe_2O_3$	10,84	12,06	6,71	8,60	6,49	8,61	10,87	10,62[1])	7,76	8,43
H_2O +	2,76	3,59	2,79	2,43	4,53	2,50	2,70	2,43	2,24	2,42

	11	12	13	14	15	16	17	18
al	40,5	44,5	51,9	55,8	47,4	56,7	39,7	30,5 —46,0
fm	38,5	39,0	36,8	27,8	32,1	29,1	31,5	38,5 —46,5
c	6,5	2,5	2,3	2,0	5,8	2,8	14,4	1,0 —11,0
alk	14,5	14,0	9,0	14,4	14,8	11,4	14,4	12,0 —16,5
k	0,57	0,66	0,56	0,44	0,36	0,64	0,62	0,40— 0,83
mg	0,39	0,41	0,28	0,17	0,28	0,37	0,51	0,38— 0,55
$\frac{FeO + Fe_2O_3}{Al_2O_3}$	0,38	0,46	0,32	0,32	0,19	0,38	0,42	0,37— 0,57
$FeO + Fe_2O_3$	8,19	7,46	9,48	7,65	9,89	9,60	6,5	6,55— 8,43
H_2O +	1,47	1,36	3,17	3,39	2,77[2])	1,20	5,00[2])	1,27— 3,63

1. Anal. (8) in Williamson (1953). Staurolite-schist.
2. (11) „ „ „ „ „
3—7. Anal. (7)—(11) in Snelling (1957). Staurolite-garnet-schists.
8. Anal in Ellitsgaard-Rasmussen (1954). Staurolite-schist.
9—10. Present work. (Sp. 56/62 and 43/63) Table X. Staurolite-garnet-schists.
11—12. Present work. (Sp. 106/63 and 416/59 Tables X and XVI). Kyanite-garnet-schists.
13—14. Anal. (1) and (2) from Snelling (1958). Kyanite-chloritoid-schists.
15. Anal. (12), Snelling (1957). Kyanite-garnet-schist.
16. Anal. (10), Williamson (1953). Kyanite-staurolite-schist.
17. Average shale, Clarke (1924).
18. Variation range of pelitic rocks in the present area.

[1]) Total iron as FeO.
[2]) H_2O + and H_2O —

It is thought that the layers in different parts of the area belong to approximately the same stratigraphical horizon. Amphibole porphyroblasts are of alumina-rich composition. They are partly orientated at random in relation to the schistosity planes and partly as "garben" within these planes. The porphyroblasts seem to be younger than most of the other minerals, retrogressive chlorite excluded. (Figs. 36,37) It was suggested that the mineral assemblage of these rocks is the result of a polyphase development. In the first syntectonic phase muscovite, biotite, garnet, epidote, oligoclase and quartz were formed. A pronounced schistosity dates from this metamorphism. During the next post-tectonic phase porphyroblastic development of amphibole occurred, in rare cases together with albite. In a third phase chlorite, quartz and ore minerals were formed along shear planes and cracks in the amphibole porphyroblasts. This phase is regarded as contemporaneous with thrust movements.

The schists in question are closely associated with rocks belonging to the almandine-amphibolite facies. They have therefore been plotted in the ACF-diagram 40 d (nos. 5 and 6). It is obvious from a comparison between this diagram and the mineral associations shown in Table XXI that the assemblages are not in accordance with their chemical composition while the association of an amphibole with muscovite is particularly unexpected in this facies. Strand (1958) described greenschists from Helgeland with the association amphibole-muscovite occurring together with biotite-epidote-chlorite rocks. The occurrence of eitherone of the two assemblages was supposed to be dependent on the Mg/Fe relations, the muscovite-amphibole-bearing rocks being the most magnesia-rich. It seems probable that the rocks described by Strand belong to the biotite-epidote subfacies. In the case of the present rocks they are associated and alternating with pelitic rocks which are partly in the almandine-amphibolite facies. If judgment is based on mineral assemblage only, rocks of different metamorphic grade are interbanded with each other. The textural relations, however, as described previously, show that the minerals have developed during several epochs. It seems probable that the discrepancy between the actual assemblage and that which is to be expected from the position in diagram 40 d results from this polyphase development. The second and third metamorphic phases did not entirely reconstruct the assemblage and textures attained during the first main metamorphism. The association thus probably represents a state of disequilibrium in the schists in question. Chloritization in the last phase is not restricted to this rock type but found in several other schists. The growth of amphibole porphyroblasts was evidently promoted by the somewhat special chemistry of these schists within the area. There is nothing to indicate special thermal or tectonic conditions for the porphyroblastic rocks. Moreover the association with amphibolites is not so regular as to suggest any thermal or chemical effects. The field relations also seem to contradict the possibility of metasomatic processes having been active. Fold style is not in any way different from that found in other schist types. It is evident, therefore, that the porphyroblastic growth must have taken place in the static period subsequent to the main folding and metamorphism.

Summary of metamorphic phenomena in the meta-sedimentary rocks

From the discussion in the preceding chapters it follows that the metamorphism of the metasediments can be sub-divided into the following phases:

a) Progressive metamorphism. Only one phase of this kind can be recognized within each locality; whether this occurred at exactly the same time everywhere within the whole area is undecided. Minerals formed during this progressive metamorphic phase include chlorite, epidote, muscovite, biotite, garnet, kyanite and feldspars. This metamorphic phase has been shown to be coeval with the main (first) episode of folding.

b) Post-deformational porphyroblastesis. Amphibole, staurolite (?), garnet (?), plagioclase and to some extent biotite porphyroblasts were formed, mainly within the almandine-amphibolite facies schists. It is possible that granite intrusion (?) and granitization of sediments within the Narvik Group and higher tectonic units occurred during this phase. The possibility of the feldspar material being inherited from the sediments cannot, however, be quite discarded. The muscovitization of kyanite is probably due to a reaction between feldspar material and kyanite at this stage of metamorphism.

c) Retrogressive metamorphism. In connection with thrust movements chloritization of other ferromagnesian minerals occurred. The present boundaries between the metamorphic facies were, to some extent, determined during these late movements.

In addition to the general discussion particular attention has been paid to certain mineralogical and related chemical features: The mutual stability relations of staurolite and kyanite were found to be governed mainly by the changing metamorphic conditions; most chemical characteristics of the two types of rocks are common to both. Rocks with amphibole porphyroblasts were found to be in a state of disequilibrium with respect to the mineral assemblages.

SAMMENDRAG

Dette er den første av tre publikasjoner som beskriver geologien i det sydlige Troms. Den omfatter stratigrafi og metamorfose i de sedimentære bergarter og en omtale av basalgneisene og de prekambriske deler av området. De to neste publikasjoner vil behandle henholdsvis petrografi og metamorfose i de eruptive bergarter og strukturgeologien i området.

Lagrekken innen kartområdet er fremstilt i Tabell III. En rekke skyveplan opptrer (merket med piler i tabellen) og gjør den opprinnelige stratigrafi usikker. En videre diskusjon av denne blir derfor ikke presentert før i den struktur-geologiske publikasjon. Det samme gjelder et forsøk på korrelering med andre deler av den kaledonske fjellkjede.

Underst i lagrekken er Hyolithes-sonen som bare opptrer i de østligste strøk (se kartet, fig. 1). Hovedbergartene er sandstener og sandige leirskifre. De øvre deler av Hyolithes-sonen mot Storfjell-gruppen er tektonisert. Ved de nordvestligste grunnfjellsvinduer er bare mindre deler av sonen bevart, antagelig på grunn av tektoniske bevegelser.

Storfjellgruppen, som hviler på Hyolithes-sonen med tektonisk eller tektonisert grense, består av feltspatførende kvartsskifre med tynne lag eller linser av dolomitt. Klastiske teksturer er bevart i en del av Storfjellgruppens skifre, men i store trekk er teksturene metamorfe. Tektoniseringen er sterk, og en del av bergartene er utviklet som typiske «hårdskifre» med tektonisk bånding. Grensen mellom Storfjellgruppen og den overliggende Rombakgruppe er markert ved et skyveplan.

Rombakgruppen består av glimmerskifre, glimmerklorittskifre og kalkmarmorer, og dessuten i mindre utstrekning kvartsitter.

Over den øvre kalk i Rombak-gruppen følger, igjen med tektonisk grense, Narvik-gruppen. Denne utgjøres i dominerende grad av glimmerskifre og gneiser. Kvartsitter forekommer hyppig i den lavere del, og en tynn kalkmarmor opptrer unntaksvis i den midtre del av gruppen. I det østlige området er sure, Na-dominante bergarter en viktig bestanddel. De er tolket som keratofyrer av vulkansk opprinnelse.

Evenes-gruppen og Bogen-gruppen overleirer Narvik-gruppen i vest. Den første består av mektige kalkmarmorer med et konglomerat, Elve-

nes-konglomeratet ved basis. Bogen-gruppen utgjøres av glimmerskifre, kalkmarmorer og sedimentære jernmalmer. Lenger nord og øst er det ikke mulig å sette noe skille mellom de to gruppene, og de er der sammenfattet under betegnelsen Salangen-gruppen.

Over de kalkførende grupper følger Niingen-gruppen, i stor utstrekning bestående av de samme bergartstyper som Narvik-gruppen. Det er mulig at Niingen-gruppen er en overskjøvet ekvivalent til Narvik-gruppen. Helt avgjørende beviser på dette er ikke funnet.

Kjemisk sett viser sedimentene varierende karakter. Både sedimenter typiske for langt fremskreden forvitring, såkalt østlig facies, og mindre forvitrede typer (vestlig facies) er representert. De kan være vekselagret med hverandre, men noen lovmessighet i opptreden er ikke oppdaget. Variasjonene kan sannsynligvis best forklares ved forskjelligheter i utgangsmaterialet for sedimentet. En del plagioklasgneiser kan tolkes som et resultat av metasomatisk tilførsel av Na og Ca til de vanlige glimmerskifre, men muligheten for et primært høyere innhold av disse elementer må også tas i betraktning, f. eks. ved en tilblanding av keratofyrisk materiale.

Eruptivbergarter, omfattende basiske og ultrabasiske til trondhjemittiske og granittiske typer, forekommer fra Narvik-gruppen og videre oppover i lagrekken. Både intrusiver og mulige extrusiver er tilstede.

Selvom ikke strukturgeologien hører hjemme i denne publikasjonen, er noen trekk nevnt innledningsvis. En sterk plastisk folding både etter NØ—SV-lige og NV—SØ-lige akser er antatt å være samtidig med den sterkeste regionalmetamorfosen. Senere ble bergartene skjøvet mot ØSØ eller SØ. En stiv fleksurfolding er samtidig eller noe yngre enn skyvningen, da skyveplanene kan være svakt foldet om NNØ-lige akser.

Metamorfosen kan henføres til en fase av sterk regional omvandling samtidig med den plastiske folding. Porfyroblastisk vekst av enkelte mineraler (amfibol, staurolitt (?), granat (?), plagioklas, biotitt) fant sted etter selve regionalmetamorfosen. Også enkelte andre metamorfe prosesser som f. eks. muskovittisering av disthen hører antagelig hjemme i denne fasen, altså etter hovedmetamorfosen. Retrograd klorittisering av enkelte ferromagnesiamineraler som biotitt og granat henger sammen med de sene skyvebevegelser. Dette ser man av at klorittiseringen stort sett er begrenset til bevegelsessonene og de nærmest omgivende bergarter. De nåværende grenser mellom de metamorfe facies er i stor utstrekning bestemt ved disse skyvebevegelsene.

Forholdet mellom staurolitt og disthen i visse skifre i området og i sin alminnelighet er diskutert i relasjon til bergartenes kjemi: Konklusjonen er at stabilitetsforholdene mellom de to mineraler i hovedsak er bestemt ved metamorfosegraden og ikke ved variasjoner i kjemisk sammensetning. De viktigste kjemiske karakteristika er felles for staurolitt- og disthen-førende bergarter.

Bergarter med amfibolporfyroblaster ble funnet å være i ulikevekt med hensyn til mineralparagenesen.

Med hensyn til metamorfe facies er variasjonene innen området store, fra nærmest umetamorf i Hyolithes-sonens mest uforstyrrede deler til almandin-amfibolittfacies i endel av de høyere tektoniske enheter. Variasjonene er dels i form av sprang i omvandlingsgraden ved de tektoniske grenser og da alltid med de mest metamorfe bergarter øverst, dels er det en mer uregelmessig variasjon lateralt innenfor de enkelte sedimentære grupper. Det er altså ingen generell økning i metamorfosen fra øst mot vest bortsett fra at de laveste og minst metamorfe tektoniske og stratigrafiske enheter er representert bare i østligere strøk av området.

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Table III

Tectono-stratigraphical columns of three different parts of the map area. (Arrows indicate position of the thrust planes.)

	Ofoten	Håfjell—Tysfjord	Salangen—Bardu—Dividalen	
Niingen Group	Mica schists and gneisses Butind quartzite Mica schist	Djupviknes quartzite Mica schist	Mica schists and gneisses	Niingen Group
Bogen Group	Bogen marble Bogen iron ore Mica schist Laksåvann marble Mica schist Grønli marble Grønli iron ore Mica schist Upper Osmark marble Mica schist Osmark mangan- iferous ore Mica schist Lower Osmark marble Mica schist Balteskar quartzite Mica schist	Mica schist Hekkelstrand dolomite and lime- stone marble Mica schist with iron ore Fuglevann marbles with iron ores Bø quartzite Mica schist	Mica schists with marbles and iron ores Quartzite Mica schist	Salangen Group
Evenes Group	Evenes limestone- and dolomite marble Evenskjær conglo- merate	Ballangen marbles Elvenes conglo- merate	Marble (Mica schist with thin quartzites)	
Narvik Group	Mica schist Graphite schist Mica schist Narvik iron ore Mica schists and gneisses Øyjord marble Mica schist and gneisses Djupvik quartzite Mica schist	Mica schist Ballangen graphite schist Sjåfjell iron ore Mica schists and gneisses Melkedalen marble Mica schists and gneisses Reppi schist	Mica schists, in part calciferous Mica schists and gneisses (Marble) Blåberget kerato- phyre Sjøvegan quartzite Seternes schist	Narvik Group
Rombak Group	Rombak marble Mica schist Quartzite, dolomite- bearing with marble layers Precambrian granite	Marble Mica schist Marble Mica schist Gneiss-granite layers	Marble Mica schist Marble (Mica schist)	Rombak Group
	Mica schist	Mica schist and quartzite		
			Feldspar-bearing quartz schists with local quartz conglo- merates Devdiselv dolomite Quartz schists	Storfjell Group
	Basal conglomerate and sandstone		Sandstones Red and green shales Basal arkose	Hyolithes Zone
	Rombak granite	Tysfjord granite	Precambrian granites	Basement Rocks

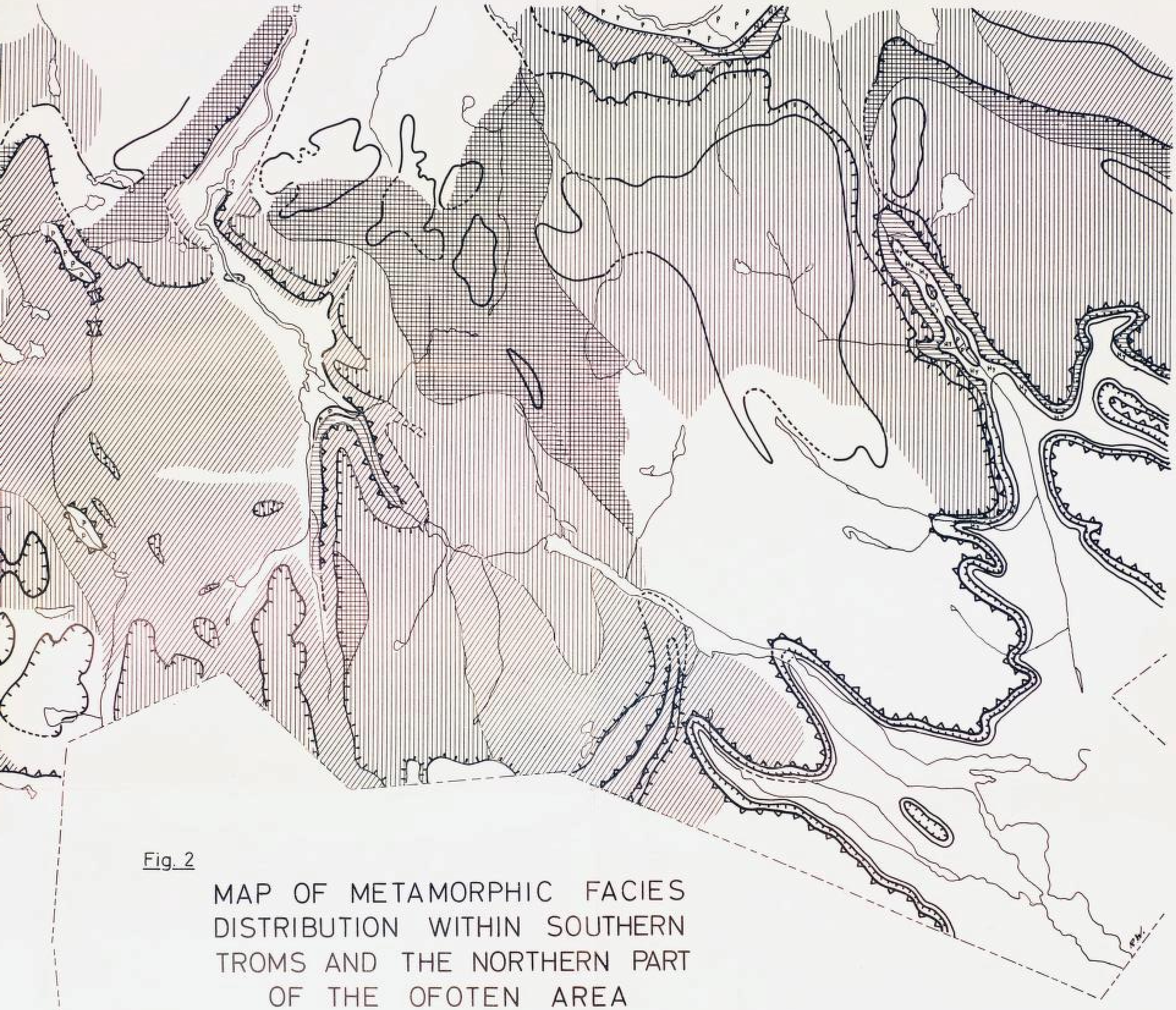


Fig. 2

MAP OF METAMORPHIC FACIES DISTRIBUTION WITHIN SOUTHERN TROMS AND THE NORTHERN PART OF THE OFOTEN AREA

scale

0 5 10 15 km.

LEGEND.

Muscovite-chlorite subfacies. (Chlorite zone.)

Epidote-biotite subfacies. (Biotite zone.)

Epidote-almandine subfacies. (Almandine zone.)

Almandine-amphibolite facies. (Staurolite and kyanite zones.)

Thrust zone between the Hyolithes Zone and the Storfjell Group.

Thrust plane below the Rombak Group.

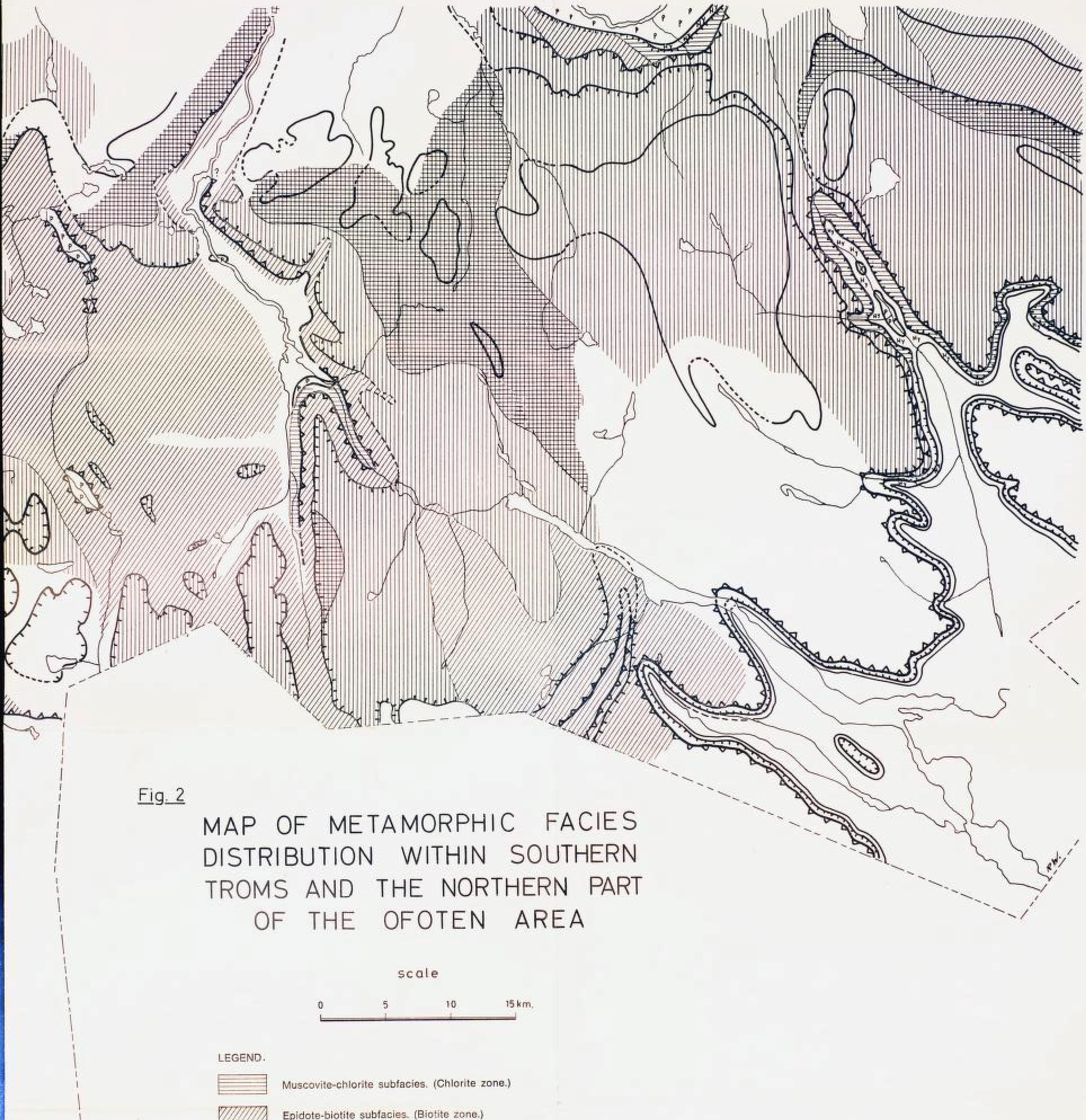
Thrust plane below slice of Precambrian granite (Rombak-Skjomen area) or «basal gneiss» (Hinnøy).

Thrust plane below Narvik Group.

Thrust plane (?) below Niingen Group.

Boundary between the Narvik and Salangen Groups.

P = Precambrian rocks. Hy = Hyolithes Zone.



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- Muscovite-chlorite subfacies. (Chlorite zone.)
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