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GEOLOGY OF THE MOSTADMARKA AND SELBUSTRAND AREA, TRØNDELAG

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Fredrik Hagemann.

Contents

Abstract

The bedrock of the area consists of metamorphic epiclastic, volcanic, and intrusive rocks of assumed Lower Ordovician age, assigned to, and lithologically matching the rocks of, the Röros, Stören and Lower Hovin Groups of the Trondheim Region. A detailed petrographic description of the rock varieties is given. The metamorphic grade of all rocks corresponds to the Greenschist Facies. — Small ore deposits occur in connection with the volcanic roks. They comprise pyrite-pyrrhotite ores of the Leksdal Type, mixed pyrite-magnetite ores, and iron oxide ores.—Lithological variation of the sedimentary rocks is tentatively correlated with events during the early periods of the Caledonian Orogeny—The volcanism within the area seems to have started with a phase of explosive activity, followed by extrusion (mostly submarine) of copious amounts of spilitic lavas; towards the end of the main period of volcanism a new episode of explosive activity has produced what is called the upper pyroclastic horizon, comprising rocks ranging from coarse volcanic breccia to finegrained tuffs. —The intrusive rocks consist mainly of saussurite gabbros and quartz keratophyres. The latter are thought to be intrusive rather than effusive, and evidence is offered to support this. In some places the character of pyroclastic material points to a quartz keratophyre parentage. Thus, quartz keratophyres may have occurred both as intrusive and volcanic rocks.—The ore deposits are thought to be of syngenetic exhalative-sedimentary origin. It is argued that the contrasts between sulphide and oxide deposits were caused by differences in the histories of evolution of the metal-producing fumaroles, and that accordingly differences in marine environment (viz. bottom physiography) *per se* were not controlling factors, nor needs biochemical influence be invoked.—Structural observations indicate that the rocks were folded around axes coinciding with the main trend of the Caledonian Mountain Chain in Norway (NE-SW), and that they were subsequently tilted by lateral compression, producing large-scale undulations and deflections from the main Caledonian structural trend.

Acknowledgments

The field work in the area Mostadmarka—Selbustrand was undertaken at the initiative of Geofysisk Malmleting, Trondheim (now a division of the Norwegian Geological Survey, NGU), and the expenses were defrayed by them. The author wishes to express his sincere thanks to the institution for this, and also to its staff for the friendliness and cooperation he received at all times.

The area of the present investigation was proposed to the author by the late Professor Thorolf Vogt.

The author is greatly indebted to Professor Niels-Henr. Kolderup and to Professor Anders Kvale, whose inspiring teaching has greatly stimulated his interest in the science of geology.

Appreciation is extended to Mr. Martin Sögnen, Geological Institute, The University of Bergen, and to Mr. Jomar Staw, NGU, Trondheim, for having made all the thin sections.

To Miss Ellen Irgens is due the credit of having drawn the maps and figures, and of having taken most of the photographs.

Introduction

The investigated area is situated about 30 km ESE of Trondheim, between the valley Leksdalen to the north and the lake Selbusjøen to the south. The area belongs to the counties of Nord-Trøndelag and Sör-Tröndelag, and includes parts of the communities Lånke, Malvik, and Selbu. The situation is shown on the map fig. 1. The mapped area is approximately 140 sq km. The topography is varying, with mountainous, uneven and partly rolling country. Altitudes vary between 160 m asl (Selbusjöen) and 700 m asl (Vennafjell). Vegetation is varying from scarce moss, heather and dwarf birch vegetation in the higher parts to dense spruce forests together with agricultural land in the lower parts. Population is concentrated along the shores of Selbusjöen and Foldsjöen, and in the Venna valley, running eastwards from the southern end of Foldsjöen to just north of Vennafjell.

The area had previously been subject to reconnaissance mapping only, the results being presented on the NGU maps Selbu and Stjördalen, published in 1891 and —81, respectively. Homan (1890) gives observations made during field work by M. Bugge, H. Reusch, and himself.—

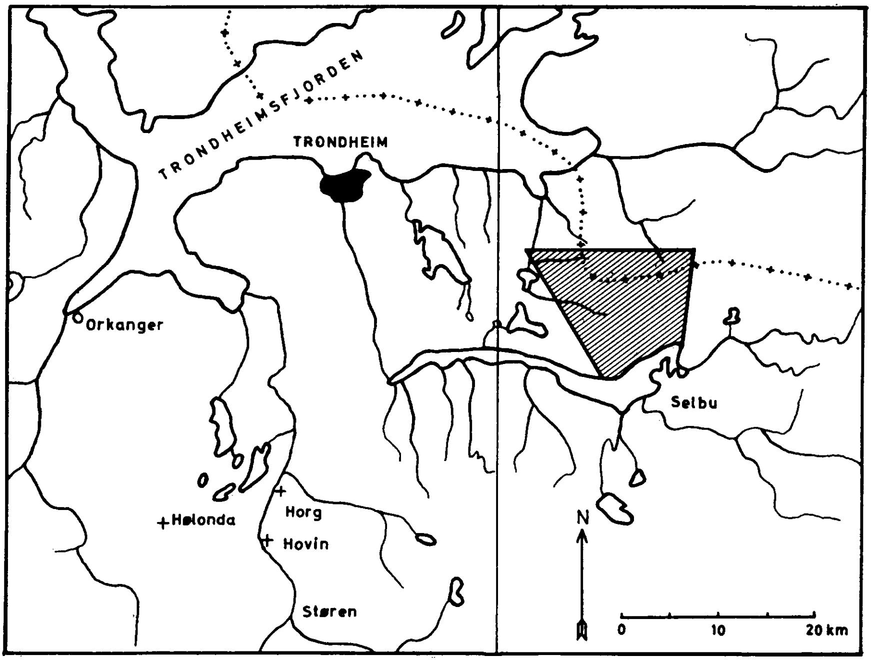

Fig. 1. Shaded: Location of Mostadmarka—Selbustrand area.

Reconnaissance reports on the ore deposits are kept at Bergarkivet, NGU.—The area is also included in the great work of Törnebohm (1896) on the geology of Central Scandinavia.

Parts of the summers 1958 and —59, and late summer and autumn of 1960 were spent in the field by the present author, the time adding up to 18 weeks of field work.

Geological setting

The Trondheim region has, geologically speaking, been one of the main centers of interest within the Caledonian mountain chain in Scandinavia. Its stratigraphy was the subject of differing opinions for some generations, but at present geologists agree on the following age relations, which seem to rest on a sound basis of petrographical and palaeontological evidence. The stratigraphy has been established mainly in the area Gauldalen, Stören, Hovin, Hölonda, and Horg, where regional metamorphism is negligible. The focusing of interest on this area has made it well known, and caused a thorough mapping of it, whereas the rest of the Trondheim region has been subject to more sporadic investigation.

The main features of the stratigraphic column are as follow:

Silurian:	Horg Group:	Shale and Sandstone Conglomerate (Lyngestein cgl.)
Ordovician:	Upper Hovin Group:	Rhyolite and rhyolite tuff Sandstone Conglomerate (Volla cgl.)
	Lower Hovin Group:	Shale Rhyolite tuff Sandstone, shale, limestone Conglomerate (Stokvola-, ex Venna cgl.)
	Stören Group	Greenstones: Lavas, pyroclastic rocks, minor amounts of sedimentary rocks.
	Röros Group:	Mainly phyllites and mica schists.

The earlier mapping on a regional scale has revealed that the rocks of the Stören Group strike through the area Mostadmarka—Selbustrand in northeastward continuation from the type area, and that this rock unit is bounded to the south by the sedimentary rocks of the Röros Group, and to the north by the sedimentary rocks of the overlying Lower Hovin Group. A large-scale petrographic tripartition in this sense is evident within the present field, and there have been found no indications to invalidate the concept of identity with the supracrustal rocks in the type area. Accordingly, the opinion is offered, that within the present field the following rock units occur: The upper part of the Röros Group to the south, the Stören Group in the central parts, and the lower parts of the Lower Hovin Group (Vogt 1945) to the north. The parts of these groups occurring within the map area will, in the following, be designated the Lower Sedimentary Unit, the Greenstone Unit, and the Upper Sedimentary Unit, respectively.

LOWER SEDIMENTARY UNIT

This unit is met with in the southern parts of the field and consists of *phyllites, metagreywackes,* and *quartz schist.* In addition to the sedimentary rocks small acid intrusions occur: *Trondhjemite, quartz-albite rocks* (quartz keratophyres).

Farthest south, and lowermost in the exposed column fine-grained, easily-weathering, thin-splitting grey phyllites occur, in many places containing up to 10 cm long quartz lenses. Layering is occasionally observed in about mm-thick laminae of alternating light and dark grey colours. As a rule cleavage makes a high angle with the bedding; both dip NW-ward, the cleavage always at the higher angle. This indicates that the rocks are not inverted, and that this phyllite thus is older than the sedimentary rocks to the north.

Northwards, and upwards in the sequence the phyllite becomes more coarse-grained, in its upper parts grading into meta-greywackes: Grey- to brownish, medium- to finegrained chlorite-muscovite-albite-quartz schists with varying content of carbonates, and with easily identifiable epiclastic textures. Locally, the metagreywackes may be coarse-grained and gritty, containing rock fragments exceeding 0.5 cm in diameter.

With a sharp boundary the meta-greywackes are overlain by a thin-bedded, white- to light grey-coloured quartz schist, with cm-thick quartzitic layers alternating with mm-thick pelitic layers. The rock is very conspicuous in the field, and was used as a marker bed during mapping. Its intricate pattern of folding and its irregular distribution soon made it clear that tectonic processes have made it a cumbersome task to obtain a reliable and detailed stratigraphic sequence. Törnebohm (1896) presents a profile through the field, from Fuglem in the south to Mostadverket in the north, from which it appears that he assumed the existence of several beds of this quartz schist, some of which he reckoned to the Lower Sedimentary Unit (Törnebohm's Brek schists), and some to the Greenstone Unit (Stören Group). Due to the paucity of outcrops the quartz schist cannot be followed continuously along the strike, but there is evidence to show that it constitutes one bed only. For instance, the quartz schist that in Törnebohm's profile is assigned to the Greenstone Unit does not continue into this unit in the direction of strike to the NE. By a study of the types of folding drag folds were distinguished, and by their pattern and its indications concerning directions of relative movements and transport of material it was possible to determine synclines and anticlines compatible with the concept of one single bed of quartz schist belonging to the Lower Sedimentary Unit.—Drag folds are an exclusive feature of incompetent rocks, and the quartz schist has been rendered incompetent by its closely spaced intercalations of phyllite layers. The incompetence has been accentuated by the close proximity of the Greenstone Unit, although the greenstones are not everywhere immediately adjacent to the quartz schist.— The quartz schist is in places cut by cross-joints, and faulting has taken place along some of them, producing offsets sufficiently large to be revealed by the geological map. However, no direct and undisputable evidence of faulting has been observed at any single outcrop.—The quartz schist withstands erosion somewhat better than the surrounding sedimentary rocks, and its presence is in places revealed by small ridges even though the rock itself be covered.

Above the quartz schist follows a thin bed of a light green-coloured phyllite, termed the Upper Phyllite to distinguish it from the Lower Phyllite previously mentioned. The Upper Phyllite has been found in isolated zones only, along the boundary between the quartz schist and the overlying Greenstone Unit. The irregular occurrence of this rock is probably due to a squeezing out during processes of deformation.

For the Lower Sedimentary Unit the following stratigraphic sequence has thus been etablished:

Upper Phyllite (Youngest)
Quartz schist
Meta-greywackes
Lower Phyllite (Oldest)

Determination of thicknesses within this unit is rendered unreliable due to the intense and partly isoclinal folding of the rocks and the ensuing repetitions of strata. A rough estimate gives these figures: Upper Phyllite 30 m (Max.), quartz schist 100 m, meta-greywackes 100—150 m.

The varied lithology of the upper parts of the Lower Sedimentary Unit indicates relatively rapid changes in relief and in the distribution of land and sea in the source area of the sedimentary material.

GREENSTONE UNIT

The supracrustal rocks of this unit consist of *greenstones* and *-schists, quartzites* and *quartz schist, phyllite,* «jasper»-bearing *greenstone breccia, and ore deposits.* The quantitatively most important rocks are the greenstones, which can be singled out in varieties like *pillow lavas, massive greenstone, porphyritic greenstone, tuffs, agglomerates,* and *tuff breccias.*

The pyroclastic rocks occur mainly in two horizons: Lowermost in the Greenstone Unit a thin bed of strongly deformed agglomerates and tuffs occur, which cannot be traced continuously along the strike. Deformation and metamorphism seem in places to have obliterated the primary structures so that it is impossible in many instances to determine what variety of greenstone is met with in the field. Thus it is not possible to say whether or not this lower pyroclastic bed is only locally developed.—The agglomeratic fragments appear as diffuse, deformed inclusions of a light green-coloured greenstone in a darker green-coloured greenstone matrix. The fragments reach a size of up to 10 cm in their longest dimension.

The upper pyroclastic zone is much thicker than the lower, and is continuously traceable from the eastern shore of Foldsjöen along the entire length of the Venna valley. The rocks of the upper pyroclastic zone are only slightly metamorphosed, and are almost everywhere easily distinguishable from the other varieties of greenstone. The rocks consist of coarse tuff breccia, with angular, exotic fragments measuring up to about 3 m in the longest dimension, tuffs with scattered lapilli, and fine-grained tuffs, partly layered, partly massive, and nearly unmetamorphosed devitrified glassy tuffs. Thin beds of pillow lava are intercalated with the tuff beds, indicating a subaquatic deposition of the latter. The upper pyroclastic beds show a gradual change in the size of fragments, with fragments generally decreasing in size from W to E along the strike. This indicates that during this phase of the volcanism a main center of eruption was situated in a direction just to the west of the Foldsjöen locality.

The several varieties of greenstone will be dealt with later, in the descriptive part of the paper.

The ore deposits may, technically and economically, be divided into pyrite- and iron-ore deposits. The iron-ore deposits were mined from 1657 to 1870, however, with several breaks.

The sedimentary rocks of the Greenstone Unit are quantitatively of little importance. They will be dealt with in a later section.

Estimates of thickness of the Greenstone Unit are not very reliable due to the lack of extensive marker beds. In the northeastern corner of the map area, around Vindsmyr, the maximum thickness is calculated to about 1000 m (1050), which is the shortest horizontal distance between the upper and lower boundary of the unit, and in this area in a vertical position. This figure cannot be used for the central and western parts of the Greenstone Unit area, because volcanic rocks may be expected to vary more widely in thickness over short distances than do sedimentary rocks.

To the north, and upwards in the stratigraphic column, the systematic mapping was limited to the ascertainment of the boundary between the Greenstone Unit and the Upper Sedimentary Unit. For this reason the stratigraphy of the Upper Sedimentary Unit is not known in any detail. The material collected from this unit is dealt with in the petrographic section of the present paper.

Some structural observations indicate that tectonic movements have taken place along the boundary between the Greenstone Unit and the Upper Sedimentary Unit. The present investigations have not been exhaustive enough to bring out whether or not these movements have caused tectonic discordances between the two units.

Small intrusive bodies of *saussurite gabbro, quartz keratophyre, quartz keratophyre-like rocks,* and one small body of an *amphibole-epidote-biotite rock* occur in the greenstone. No contact metamorphic phenomena have been discovered. The saussurite gabbros have been subject to a more penetrative cataclastic deformation than have the acid intrusives, which in part may be without any sign of deformation whatsoever. The saussurite gabbros therefore, are thought to be older than the acid intrusives.

Rock Description

The petrographic part is presented in the following order:

I. Supracrustal rocks (÷ the ore deposits), described in the order of ascending stratigraphical succession.
II. Intrusive rocks.
III. Ore deposits.

The metamorphic facies classification follows the nomenclature suggested by Turner and Verhoogen (1960).

I. Supracrustal rocks

LOWER SEDIMENTARY UNIT

Lower phyllite

This rock is a light grey to dark bluish grey phyllite, cleaving in two directions, the one parallel to the bedding, the other parallel to axial planes of micro-folds. The axial plane cleavage is the more prominent, and along this the rock splits easily into thin flakes. The cleavage planes of the other direction are spaced at a distance of 4—5 cm apart. Bedding is conspicuous in cross section as mm-thick laminae. The phyllite weathers easily, and forms the bedrock of the cultivated land along the Selbustrand area. Farthest to the south, along the shore of Selbusjöen, the phyllite contains numerous quartz lenses, up to 10 cm long.

The mineral composition of the lower phyllite is: Quartz, muscovite, chlorite, carbonate, rutile, tourmaline, albite, zircon, and carbonaceous matter.—The texture is equigranular, granoblastic to lepidoblastic. The carbonaceous matter is arranged in layers, clouding the minerals as a dusty pigment. The layers rich in carbonaceous matter are free from carbonate minerals. The modal composition of the rock is estimated at:

Quartz	30	per	cent
Muscovite	30	,,	,,
Chlorite	22	,,	,,
Carbonate	13	,,	,,
Accessories	5	,,	,,

Metamorphic grade corresponds to the Greenschist facies, Quartz-albite-muscovite-chlorite Subfacies.

Metagreywackes

These rocks have been named in accordance with the nomenclature suggested by Pettijohn (1949). In order to emphasize the gradual transition between the Lower Phyllite and the metagreywackes proper, the designation *greywacke-phyllite* is introduced here to cover the transitional rock type. The metagreywackes may, according to the adopted nomenclature be divided into greywacke, subgreywacke, calcareous greywacke, calcareous arkose, calcareous felspathic quartzite.

Greywacke, subgreywacke and albite-quartzite. (Pl. I, fig. 1)

These varieties of the meta-greywackes are mostly fine-grained. The colours vary from light grey to light greyish green to brownish grey. Grains of carbonate are in many places conspicuous on weathered surfaces as spots of rust. Along the boundary between the meta-greywackes and the overlying quartz schist the meta-greywackes are dissected by thin, irregular quartz veins, 1—2 mm thick. These veins are oriented more or less perpendicularly to the boundary. A few metres from the contact the quartz veins seem to be absent.—The brownish varieties of the meta-greywackes are the richest in carbonate minerals, and they owe their hue to the formation of limonite as a weathering product of carbonate.—The structure is schistose in places (Strain slip cleavage), but lineation is the most prominent structure element, and is caused by parallel orientation of mineral grains (muscovite, chlorite); larger quartz and felspar grains are commonly elongated in the direction of lineation.

In the typical representatives of the meta-greywackes small-scale primary bedding is hardly discernible, whereas this is a conspicuous feature of the more fine-grained greywacke phyllite.

The mineral composition is: Quartz, albite, carbonate, muscovite, chlorite, rutile, tourmaline, sphene, zircon, ore minerals, limonite, ± biotite.

The texture is lepidoblastic-blastopsammitic, with clastic grains of quartz and felspar, and small rock fragments, while the pelitic material is completely recrystallized, and occurs as muscovite and chlorite, in places also as biotite.

Under the microscope, the lineation appears (in sections parallel to it) as an orientation of flakes of muscovite and chlorite around an axis, and to a minor extent as elongation of quartz and felspar grains parallel to the same axis.

The rock fragments consist of quartzite and quartz-albite rocks. In the latter fragments quartz and albite show granophyric intergrowth.

Some of the clastic quartz grains in the greywacke contain unoriented inclusions of hair-thin needles of a mineral with high positive relief, probably rutile. These quartz grains also carry rows of small liquid inclusions.

Some of the larger muscovite-chlorite aggregates are interpreted as recrystallized fragments of pelitic rocks.

Greywacke phyllite. (Pl. I, fig. 2)

This rock is a fine-grained, light grey-coloured schist which splits easily into small flakes and rods. The cleavage is a strain-slip cleavage parallel to the axial planes of micro-folds. The cleavage planes are about 1 mm apart. The cleavage makes angles of 70—90° with the bedding. The axes of the micro-folds, together with the intersection lines between bedding and cleavage constitute a prominent linear structure. Flaky minerals have a grain orientation about axes in the same direction, and quartz and felspar grains show a certain amount of elongation in the same direction. In cross section bedding is seen as alternating white- and grey-coloured laminae, 1—2 mm thick. Euhedral porphyroblasts of carbonate give the rock a characteristic appearance. On weathered surfaces the carbonate is replaced by limonite pseudomorphs. Because of its fineness of grain and the relatively prominent cleavage this rock is thought to constitute a transition between the underlying phyllite and the metagreywackes proper.

The mineral composition is: Quartz, albite, muscovite, chlorite, carbonate, rutile, tourmaline, apatite, ore minerals.

The texture is porphyroblastic (carbonate grains) with a lepidoblastic-granoblastic groundmass. Irregular narrow quartz veins, with quartz grains conspicuously bigger than the quartz of the groundmass, evidently are secondary joint fillings.

The carbonate porphyroblasts consist of siderite, forming euhedral to subhedral rhomb-shaped crystals, averaging about 1 mm in diameter. The crystals are poikiloblastic, with inclusions of small quartz and felspar grains together with dark nearly submicroscopic grains (probably rutile). The inclusions are found in the central parts of the porphyroblasts. In some cases the inclusions are arranged in zones, with concentrations along crystallographic directions in the porphyroblasts. Besides, inclusions show concentrations along the diagonals of the rhombohedral carbonate grains, to produce an hourglass structure.—Along their cleavage planes the porphyroblasts contain limonite, a decomposition product of the siderite. In places the siderite has been almost completely replaced by limonite.

The metamorphic grade of the greywacke phyllite corresponds to the Greenschist facies, Quartz-albite-muscovite-chlorite subfacies.

Calcareous arkose.

This variety occurs as a 10 cm thick layer in the uppermost parts of the Lower phyllite, in the transition zone phyllite-greywacke phyllite. It is underlain by ordinary fine-grained phyllite, and is overlain by greywacke phyllite. The rock is fine-grained, in places cleaving along the weakly indicated bedding. The colour is light brownish grey, brown on weathered surfaces.

The mineral content is: Quartz, albite, calcite, muscovite, chlorite, rutile, sphene, zircon, ore minerals.—The rock contains scattered tiny fragments of quartzite.

The texture is equigranular, granoblastic-blastopsammitic.

Metamorphic grade corresponds to the Greenschist facies, Quartz-albite-muscovite-chlorite subfacies.

Calcareous greywacke.

The larger content of carbonate is the most significant difference between this variety and the carbonate-poor greywacke previously described. The rock is fine-grained, in places with mm-big porphyroblasts of carbonate. The colour is light brownish grey to spotted yellowish and brownish green-grey. The structure varies from weakly schistose to massive. Bedding can be seen locally, as diffuse laminae, about 5 mm thick, of alternating carbonate-rich and more micaceous material. On weathered surfaces the decomposition of carbonate yields abundant limonite.

The mineral composition is: Quartz, ankeritic dolomite, albite, muscovite, chlorite, rutile, limonite, ore.

The texture is lepidoblastic-granoblastic, in places porphyroblastic, with porphyroblasts of carbonate.

Metamorphic grade corresponds to Greenschist facies, Quartz-albite-muscovite-chlorite subfacies.

Quartz schist. (Fig. 2)

This rock is a thin-bedded, light-coloured schist, with alternating light and dark layers. On weathered surfaces the schist is white, and displays a conspicuous selective weathering of alternating layers of quartzite and

Fig. 2. Folded thin-bedded quartz schist. Loc. 139.

phyllite. The quartz-rich layers are 1—2 cm thick, the phyllite layers about 0.1—0.5 cm. Locally, the phyllitic layers may reach thicknesses far in excess of the above-mentioned order of magnitude: In one locality a single phyllitic layer was found to be about 70 cm thick.—The schist in many places is intensely folded, and it has, due to its thin-bedded layering, acted as an incompetent rock.—On fresh surfaces the quartz schist has a light greenish grey colour, with thin grey-coloured phyllitic layers. The quartzite layers as a rule look massive, but frequently a planar structure parallel to the axial planes of folds is to be noticed. In places the rock is dissected by thin, irregular quartz veins.

The mineral composition of the rock is: Quartz, muscovite, chlorite, (±) biotite, albite, calcite, rutile, apatite, tourmaline, ore.

The texture varies from granoblastic mosaic in the quartz-rich layers to lepidoblastic in the mica-rich layers. The rock is also homeoblastic, except where quartz-filled veins have quartz grains of larger dimensions than the pre-ruptural grains. Muscovite (as sericite) shows a well developed parallel orientation, while larger flakes of chlorite are more randomly oriented.—Where biotite is developed it shows no preferred orientation, nor do the biotite flakes show any sign of deformation, and thus the mineral appears to have formed after the cessation of rock deformation. However, jointing and faulting of the rock may be postcrystalline also in relation to biotite.

The quartz schist is interpreted as an originally layered sediment, deposited by cyclic sedimentation of alternating sandy and clayey material.

The paragenesis quartz-muscovite-chlorite shows the rock to belong in the Greenschist facies, Quartz-albite-muscovite-chlorite subfacies. Where biotite enters the paragenesis the schist reaches the Quartz-albite-epidote-biotite subfacies.

Upper phyllite

This rock is a thin-splitting, light grey to greenish grey phyllite. It has a linear structure parallelling small-scale isoclinal folding, the lineation being lines of intersection between primary bedding and schistosity planes. Apart from the above-mentioned isoclinal folding, it shows a small-scale flexuring of the planar structure around axes deviating some 60° from the axes of the isoclinal folds. This flexuring, represented by kinks on the schistosity planes, is a result of a younger deformation than that causing the isoclinal folding. The axial planes of the kinks imparts to the rock a weakly developed cleavage normal to the schistosity.

The minerals of the rock are: Quartz, muscovite, chlorite, albite, tourmaline, sphene, rutile, ore, +/÷ biotite.—The texture is granoblastic-lepidoblastic. In thin section the layering is seen as a composition banding, with alternating quartz- and mica-rich bands. The grain size is mostly less than 0.05 mm, but a few quartz grains reach a diameter of about 1 mm. Some of the chlorite flakes may reach 0.5 mm in their longest dimension, and where biotite occurs, this mineral has a similar porphyroblastic development.—The modal composition of the Upper phyllite is estimated at:

Quartz	30	per	cent
Chlorite	40	,,	,,
Sericite	20	,,	,,
Other minerals	10	,,	,,

Metamorphic grade, as brought out by the mineral composition, corresponds to the Greenschist facies, Quartz-albite-muscovite-chlorite subfacies, with traces of the higher Quartz-albite-epidote-biotite subfacies.

THE GREENSTONE UNIT

The supracrustal rocks of this unit comprise greenstones and -schists, quartzites and quartz schists, phyllite, ore deposits, and greenstone breccia with "jasper" fragments.

Volcanic rocks

The greenstones may be divided into the varieties *pillow lavas* (fig. 3), *massive greenstone, porphyritic greenstones, tuffs, agglomerates,* and *volcanic breccias.* More or less schistose varieties are also met with.

Fig. 3. Pillow lava. Loc. 67.

The mineral composition shows slight variation only, and the petrographic description of the individual types is mainly concerned with structural and textural peculiarities, which have formed the basis for distinguishing the different varieties. The greenstones do not differ appreciably from the metamorphic volcanic rocks of other areas within the Trondheim region (C. Bugge (1910), C. W. Carstens (1918, –19, –24), H. Carstens (1955 a, –60), T. Vogt (1945, –46).

The *pillow lavas* of the present area are quite analogous to the corresponding rocks of the Hölonda–Horg area. The pillows are approximately elliptical in cross-section, and as a rule they have been stretched, in a direction coinciding with the trend of other linear structures in the area. A statistic evaluation of the amount of flattening and stretching of the pillows has not been undertaken, but the dimensional proportions 1 : 2 : 6 would be an approximate of the quantitative relations. The intermediate axes vary from about 10 cm to about 75 cm, with a frequency maximum of 30–60 cm.

The composition of the pillows matches those described by T. Vogt (1945) and H. Carstens (1955 a): A fine-grained central portion, surrounded by a layer about 3–5 cm thick, with epidote as the dominant dark-coloured mineral, and an outer chloritic skin 1–3 mm thick. In a few places reliable criteria for determining the original position of the upper and lower surfaces were found. No inversions were indicated. Locally, small amounts of quartzitic material fill the interstices between the pillows.

One peculiar pillow merits special attention (Loc. 305), as its composition sets it apart from all the other pillows found in the area (Fig. 4): The interior of the pillow is white-coloured with scattered green spots. The surrounding crust, 4–5 cm thick, is green-coloured. The outer skin consists of chlorite, and the green crust has a matrix of small plagioclase grains, with nematoblastically developed needles of amphibole, and grains and aggregates of epidote and calcite. The epidote is distinctly green-coloured in thin section. Birefringence determinations gave a result of 0.045, which corresponds to an epidote with about 30 per cent Fe-epidote (Tröger 1959). The amount of epidote in the crust increases towards the interior of the pillow, and dominates completely just outside the white-coloured central portion. The peripheral part of the interior of the pillow is made up of calcite, and the inner core consists of quartz. The few green spots in the central portion are aggregates and crystals of epidote. The crust is transected by radial cracks, about

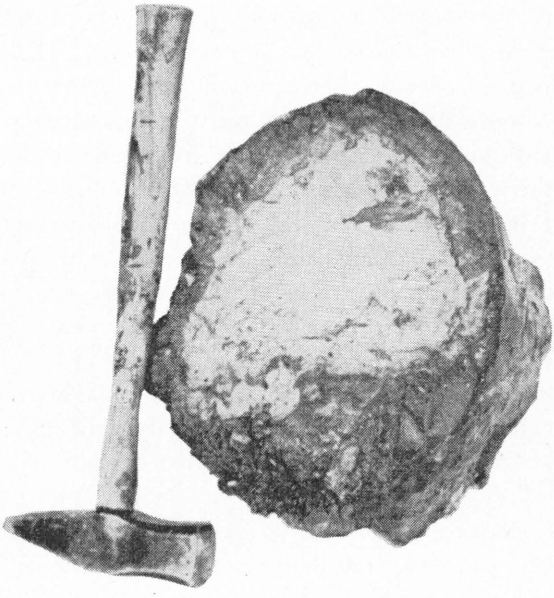

Fig. 4. "Emptied" pillow with quartz- and calcite-filled interior. Loc. 305.

1 mm wide. They are filled with quartz, and some muscovite and epidote. The genetic history of the pillow is interpreted thus: Submarine eruption has caused immediate consolidation of the crust (as in all lava pillows), later, the still liquid lava of the interior, possibly driven by gas pressure, has left the crust a hollow shell. Subsequently the cavity was filled, after a shorter or longer interval, by minerals whose components were introduced hydrothermally. As may be seen in the photo some of the lava was left on the bottom of the pillow. Fuller (1931) have described similar features in pillow lavas from the Columbia River Basalt of the USA. There, the cavities were still empty, and the lava left over rests with a plane surface on the bottom.

Massive greenstones

These rocks are mostly fine-grained, frequently to an extent that renders the determination of their mineral composition very difficult. In other cases the grain size may be comparable to that of saussurite gabbros of small grain. It is not always possible to say whether they originally were lavas or tuffs, or even intrusive rocks. Many workers dealing with the Trondheim region have assumed that the schistose varieties of the greenstones originally were tuffs. Similar reasoning would then naturally lead to the interpretation of the fine-grained, massive greenstones as lavas (and sub-volcanic intrusives). This structural criterion seems to carry little evidential force unless sustained by evidence of the mineral composition or inherited diagnostic textural features. In this connection it may be mentioned that the varieties of greenstone of indisputable pyroclastic origin within the present area are not more schistose than are pillow lavas intercalated with them.—In many instances the massive greenstones under the microscope display a blastophitic texture, indicating a consolidated rock melt, and those are considered to be metamorphosed lavas.

Most of the massive greenstones and the pillow lavas are fine-grained to dense. The colour is green, as a rule relatively light, but in some cases nearly greenish black.—The mineral constituents are: Albite, chlorite, epidote, +/÷ amphibole, calcite, leucoxene, rutile, +/÷ quartz, sericite, +/÷ biotite, apatite, limonite, pyrite, +/÷ magnetite.—Biotite is not common, and quartz usually lacking. Amphibole is lacking in some rocks.—Reliable modal analyses are not easily obtained, but the two presented below are thought to be representative:

Specimen 33:

Amphibole	25	vol per cent
Epidote	30	„ „ „
Chlorite	20	„ „ „
Albite + some quartz	15	„ „ „
Calcite	6	„ „ „
Accessories (Mostly leucoxene)	4	„ „ „
	100	

Specimen 43:

Albite	47	vol per cent
Chlorite	30	„ „ „
Calcite	14	„ „ „
Epidote	3	„ „ „
Biotite	0.5	„ „ „
Accessories (Mostly leucoxene & pyrite)	5.5	„ „ „
	100.0	

The two analyses give an impression of the variation in modal composition of the greenstones, yet they do not represent the extremes.— The textures of these rocks are nematoblastic to lepidoblastic, and frequently blastophitic. Grains of calcite or pyrite may be developed as porphyroblasts.

Porphyritic greenstone

has been found in two varieties, one of which has a macroscopic appearance distinctly different from the non-porphyritic greenstones. It is characterized by white-coloured phenocrysts of felspar, measuring 0.5—2.5 cm in their longest dimension. The phenocrysts are subhedral, mostly with rectangular outlines, but more equant grains are also present. On weathered surfaces the phenocrysts are protruding, giving the rock a nodose surface (Fig. 5 and Pl. II,, fig. 2). The ratio of phenocrysts to groundmass may reach a value of 1 : 1, but is as a rule appreciably smaller than this. The groundmass of this porphyrite variety has a somewhat lighter green colour than the other greenstones. Angular, exotic fragments of this rock are abundantly present in the volcanic breccia at Foldsjöen, to be described later. This mode of occurrence proves that the large felspar crystals are phenocrysts of primary origin, and not porphyroblasts grown during metamorphism.

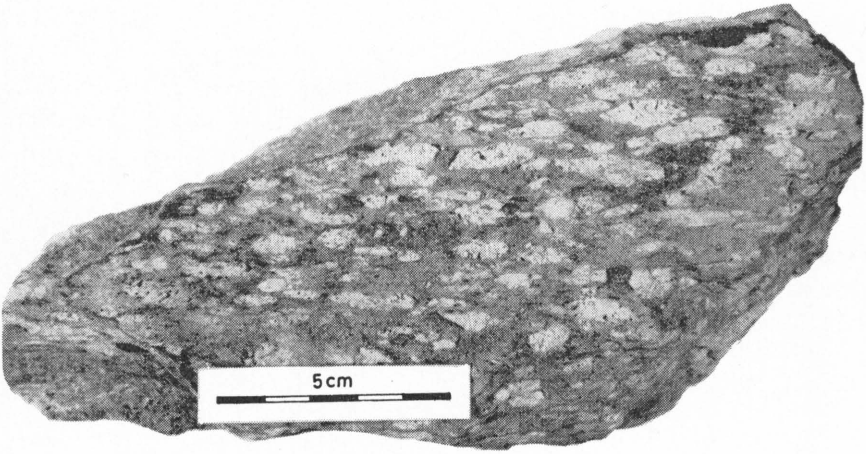

Fig. 5. Porphyritic greenstone. Same as Plate II, fig. 2. Loc. 63.

The mineral composition is: Albite, epidote-clinozoisite, chlorite, sericite, amphibole, leucoxene, rutile, stilpnomelane, pyrite, limonite. A modal analysis gave the following result:

Phenocrysts (Albite)	40	vol	per	cent
Epidote-clinozoisite	41	,,	,,	,,
Chlorite	9	,,	,,	,,
Leucoxene	10	,,	,,	,,
	100			

The abundance of fine-grained epidote readily explains the light green colour of the groundmass.

The texture is porphyritic, with subhedral albite phenocrysts set in a granoblastic groundmass. — Saussuritization phenomena are hardly to be detected, whereas sericitization is very prominent. The sericite is confined to the albite phenocrysts. The other *porphyritic greenstone variety* has a macroscopic appearance much like that of the non-porphyritic massive greenstones. On polished sections, however, the phenocrysts are distinctly observable. This rock also is a felspar porphyrite, with albite phenocrysts. The phenocrysts in this rock are about 1—3 mm in their longest dimension, and their colour as seen macroscopically does not distinguish them from the groundmass. The rock colour is identical so that in the field it is hardly feasible to single it out.

Microscopic examination of this variety shows the same mineral composition as that of the non-porphyritic greenstones, but the texture is significantly different. It is a porphyritic flow texture. The flow texture is delineated by the orientation of small amphibole needles and flakes of chlorite (Pl. III, fig. 1). The phenocrysts are subhedral, and without any visible strain phenomena. Twinning is common. The albite is only insignificantly saussuritized, but has to some extent been subject to sericitization.

The phenocrysts are often clustered in groups of 3 to 4, suggesting a glomeroporphyritic texture. In addition to the phenocrysts of albite, scattered epidote-chlorite aggregates appear to be pseudomorphous after pyroxene phenocrysts, about 0.5 mm in diameter.

Pyroclastic rocks

The varieties of this category of greenstones may be designated tuff-breccias, lapilli-tuffs, and fine-grained tuffs.

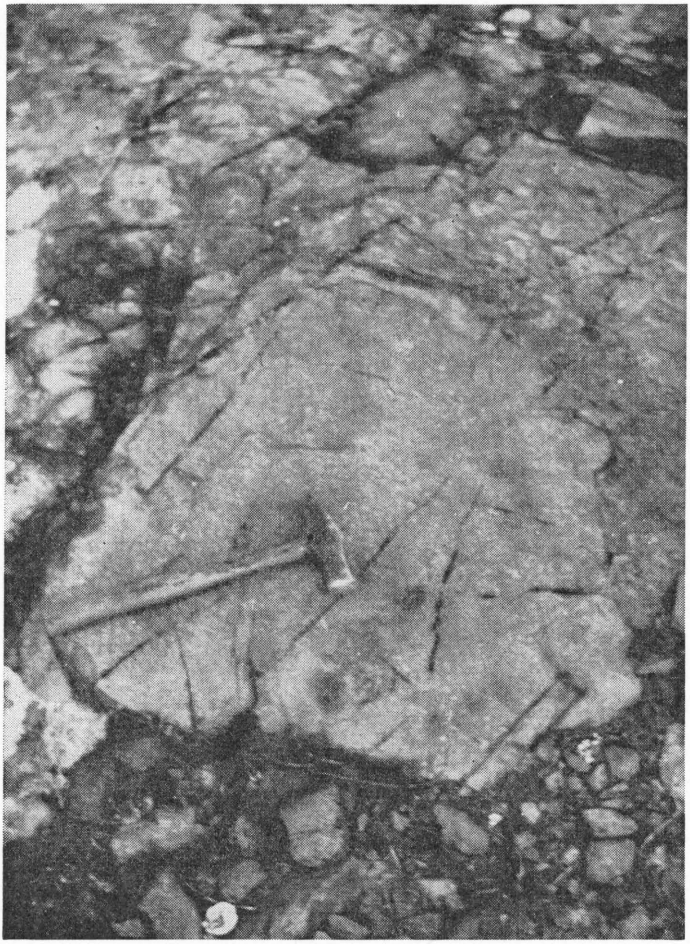

Fig. 6. Coarse tuff breccia. Loc. 251.

The *coarse tuff-breccia* (Fig. 6) occurs in the northwestern part of the mapped area, on and near the eastern shore of Foldsjöen, and it constitutes the western part of the upper pyroclastic horizon of the present area. The most advantageous exposures for field studies of this rock are on a promontory just to the west of the road from Hommelvik.—Angular fragments of different volcanic rocks are set in a dense, green-coloured matrix. Dominating among the exotic fragments is the porphyritic greenstone with large, protruding albite phenocrysts previously described. Other fragments consist of a dense, light grey-coloured, micro-porphyritic keratophyre-like rock. The fragments are completely unsorted, and there is no suggestion of bedding. The fragments are of widely varying sizes, from 2—3 cm up to nearly 3 m in their longest dimensions. Apart from the angular fragments roundish fragments of a light greyish-green rock appears, 5—10 cm in diameter; they are bombs and lapilli of vesicular lava. The tuff breccia is transsected by a joint system, but is otherwise massive. Together with, and to all probability intercalated with the tuff breccia, pillow lavas occur.

Devitrified green tuff.—In close association with the coarse tuff-breccia a dense, massive greenstone is found, which under the microscope is seen to be a devitrified glassy tuff (Pl. III, fig. 2). It displays a completely undeformed vitroclastic texture. Its mineral composition is chlorite, epidote, albite, leucoxene, amphibole, calcite, stilpnomelane, and small amounts of biotite. In addition it contains aggregates of a brownish mass of submicroscopic grains, nearly isotropic in character. Narrow irregular joints transecting these aggregates are thought to be cooling-cracks inherited from glass. Roundish or oval-shaped amygdules no doubt originated as gas bubbles in the glass. They are filled with chlorite, epidote, calcite and albite. Aggregates of extremely fine-grained albite are pseudomorphs after small and narrow lath-shaped crystals.—The different features of the texture are delineated by one, or more frequently by several rhythmically arranged thin bands of leucoxene, which probably originated during devitrification. The texture of this rock fits very well with described and photographed textures of recent basaltic vitreous tuffs (Sideromelane tuffs). Such rocks and their origin have been described by Barth (1937) and others, while their devitrification and further transformation has been thoroughly dealt with by Hentschel (1951). Barth, in describing sideromelane-ashes from Vatnajökull, Iceland, maintains that explosiveness of basaltic volcanism is enhanced by superincumbent loads, in Vatnajökull consisting of glacier ice. In the present area a superincumbent load of sea water may have produced the same effect. Hentschel's lengthy description of sideromelane-tuffs from Germany and their metamorphic equivalents ("Schalstein") is in excellent accord with observations made on the present rock, the only apparent difference being the scarcity of epidote in the German material.

Lapilli tuffs. Lapilli-bearing green tuffs are found along the Venna valley, where they may be studied to the best advantage in and along the river bed.—The rock is a fine-grained greenstone with elliptically

Fig. 7. Tuff breccia with partly corroded fragment of porphyritic greenstone. Loc. same as fig. 6.

elongated pyroclastic fragments, with medium diameters from about 1 to about 5 cm. The fragments are dense and green-coloured, as a rule somewhat darker than the matrix. The relative proportions of the elongated lapilli are about 1 : 1.5 : 5. The elongation constitutes an easterly dipping linear structure. Distinct bedding is frequently seen, with lapilli concentrated in layers with thicknesses varying from 3—4 to about 50 cm. The intervening layers of fine-grained to dense tuff are somewhat thicker, about 50—60 cm. In places the tuffs are intercalated with pillow lavas, mostly 8—10 m thick.—Microscopic investigation reveals a very fine-grained matrix, consisting mainly of albite, epidote, leucoxene, chlorite, and scattered flakes of green biotite. The rock has been somewhat pressed, but the vitroclastic texture is still easily recognizable. The lapilli too have to all appearance been glassy. The prominent leucoxene banding delineates the fragments, and makes the deformation easily visible (Pl. IV, fig. 1).

Sedimentary rocks within the Greenstone Unit

(Exclusive of ore deposits)

These rocks are quantitatively of little importance. They are: Quartzite and quartz schist, and one instance of phyllite.

The quartzite has a peculiar mode of occurrence: Its lateral extent is very small, the bodies forming lenses or rundish lumps, with approximately equidimensional outcrops. The rock is found in scattered occurrences over most parts of the Greenstone Unit area. The bodies are of varying sizes, the smallest being from a few cm's to hand specimen size (Fig. 8), the biggest up to and above 25 m in their longest dimension. The rock is mostly white-coloured, but may be grey to almost black. Invariably it contains magnetite, and in some cases pyrite in addition. The quartzite has no planar structures, and may be tough and hard, or friable and disintegrating, in one locality so friable that in previous times the local inhabitants quarried it for scouring sand. The iron-rich varieties usually have a rust-brown weathering surface.—The mineral composition is quartz, chlorite, biotite, calcite, ore, limonite. The relative proportions of the minerals were estimated in two thin sections from different localities: Spec. 42: 90 per cent quartz, 10 per cent ore. Spec. 134: 90 per cent quartz, 5 per cent ore, 2 per cent biotite + chlorite, 3 per cent calcite.—The texture is a granoblastic mosaic texture, with equidimensional quartz-grains.

A *pyrite-bearing variety* shows this mineral composition: Quartz, calcite, pyrite, magnetite, amphibole, chlorite, stilpnomelane.

Amphibole is developed as needle-shaped crystals, 0.05—0.5 mm long, and occurs in sheaf-like aggregates.

Stilpnomelane is found near the ore grains, as elongate, narrow flakes, with an imperfect cleavage normal to the basal cleavage and transverse

Fig. 8. Quartzite xenolith (To the left, below head of hammer) in massive greenstone. Loc. 313.

to the elongation. Pleochroism is very strong: Dark brown—pale yellow. The elongation is positive.

The inconsiderable lateral extent of these quartzites makes it difficult to visualize their mode of origin. It seems reasonable that their condition is related to their close association with the greenstones. The wide variation in iron content from occurrence to occurrence, revealed by the varying colour, is thought to indicate that iron has been added to them. In that case the surrounding volcanic rocks naturally offer themselves as the source of the iron. Four modes of origin may be put forward: 1) The bodies are sedimentary quartzitic sandstone which have become included in lava flows, either in consolidated or unconsolidated state, and iron added to them during the process of inclusion and transport or later. 2) They may represent the filling-in of lava tunnels, in which case the sand has been transported from above after the consolidation of the lava. 3) The sand was deposited in cavities and depressions on the surfaces of the lava flows, and is analogous to 2). 4) They are of magmatic (Pneumatolytic-hydrothermal) origin.—The author finds the first alternative to be the most likely one. Occurrences of type 2) have been described by Shrock (1948). Their formation is made possible through the collapse of the roofs of tunnels, whereupon sedimentary material may be deposited in the accessible cavities. Sediments of this kind usually are characterized by a distinct bedding. The same might be expected in case 3). Alternative 4) seems improbable because of the pronounced equigranular and "sandy" texture of the quartzites.—A genesis as explained by alternative 1) is sustained by the fact that the quartzites as a rule are surrounded by an aureole of greenstone particularly rich in epidote, but low in the other common dark minerals. These aureoles are similar to the epidote-rich crusts of lava pillows, and have probably originated in a similar way: By rapid cooling, due to the contact with the inclusion, and an introduction of water, supplied by the xenolith. Inclusions of pure quartzite in recent lavas have been described by Ruth Jakob (1958) from Vulcano and Stromboli, and from other parts of the world by Holgate (1954) and Brady and Webb (1943). Holgate writes:

> A characteristic feature of highly siliceous xenoliths in basic igneous magmas, as revealed to observation by the chances of erosion affecting the cooled and consolidated product, is their very marked tendency to retain integrity of form until a late stage of consolidation of their host. The phenomena are such as would suggest the operation of some limitation of solubility of the quartz in the basic host magma. In many occurrences the evidence of reaction are limited to the partial rounding of the quartzose insets, with a smaller-scale marginal suturing of the quartz against what appears to approximate mesostasis material of the host rock.

Brady and Webb describes fossiliferous xenoliths brought to the surface and extruded as cores of volcanic bombs, and find that magnetite has been introduced into the xenoliths, which are otherwise virtually uninfluenced by their host rock.—Thus, it seems reasonable that the quartzites of the Greenstone Unit may be xenolithic inclusions in lava. The varying content of magnetite is also best explained as supplied by the lavas. A similar transport of iron in lavas has been described by Vuagnat (1946), in a paper on pillow lavas from Switzerland, where he demonstrates processes of differentiation to have been operative within and between the pillows, as brought out by the following relations: The crusts of the pillows and the matrix between the pillows are complementary to each other, together they represent the composition of the undifferentiated cores of the pillows. The crusts of pillows are characterized by a low Fe, Mg, and water content, and by a high content of alkali, while the matrix between the pillows are correspondingly richer in Fe, Mg, and water, and poor in alkalis.

The quartzites within the greenstones in the present area have in one respect acted similarly to the matrix of Vuagnat's pillow lavas: As recipients to iron.—The fact that the greenstones in the contact areas around the quartzites are enriched in epidote is evidence that in the aureoles iron has been oxidized to its trivalent state, in which form the element enters epidote. Oxidation of iron in connection with quartzitic xenoliths is described by Holgate (1954). He demonstrates that the host rocks in the contact zones have been impoverished in iron. From his analyses of host rocks and contact zones around xenoliths from different regions of the world it is evident that the degree of oxidation of iron in the rock portions bordering on xenoliths

$$\left(\mathrm{mol}\ \frac{2Fe_2O_3 \times 100}{2Fe_2O_3 + FeO}\right)$$

is markedly higher than in the respective host rocks themselves:

	Fe_2O_3	FeO	Oxidation ratio
1: Vogesite, N. Ireland	1.98	6.31	22.0
2: Felspathic rim of quartzose xenolith in 1	0.87	0.60	56.7
3: Basic markfieldite, Dumfries ..	1.69	6.93	18.0

	Fe_2O_3	FeO	Oxidation ratio
4: Felspathic rim of quartz xenolith in 3	0.96	0.47	64.9
5: Murambite, SW Uganda	3.34	8.67	25.7
6: Glass separated from transfused qt.zite xenolith in 5	0.44	0.38	51.4
7: Murambite, SW Uganda	3.66	8.67	27.5
8: Glass separated from transfused qt.zite xenolith in 7	0.26	0.49	32.0

The generally higher oxidation ratio of iron in aureoles around xenoliths in basic rocks proves, in the author's opinion, that the presence of epidote as the dominating dark constituent in a zone around the quartzites within the Greenstone Unit of the present area is caused by a reaction between lava and quartzitic xenoliths. Further reflections on the formation of epidote and on the degree of oxidation of iron are given in a later section of the present paper.

Quartz schist
occurs within the Greenstone Unit in a zone curving around the eastern side of the lakelet Kleptjern with the direction of strike turning from near NS to WNW northwards along the zone. Estimates of thickness are unreliable, but the quartz schist probably is about 30—40 m thick. The rock is dense, and its colour varies from grey with a tint of red violet, to black. In places bedding is clearly visible, the individual layers being 0.5—1.0 cm thick. The rock may locally be folded and in some places slightly faulted, with displacements on cm-scale. A similar quartz schist occurs in a zone along the river Nævra. Here the schist has two well developed cleavage directions, and splits easily into rods with rhomb-shaped cross sections. In previous times this schist was quarried for whetstones, according to local tradition of a very high quality.

Under the microscope the rock turns out to be a sericite-quartz schist with about 85—90 vol per cent quartz, 15—10 vol per cent sericite + ore minerals. The colouring of the rock is caused by the finely dispersed ore grains. The amount of ore minerals varies, but is hardly anywhere above 5 volume per cent.—The texture of the rock is a granoblastic mosaic texture, the whetstone variety being schistose, with sub-parallel flakes of sericite and a slight elongation of quartz grains. The rock is fine-grained and equigranular, with equant quartz grains varying from sample to sample between 0.01—0.05 mm in diameter.

Phyllite

At one locality within the greenstone area, in an open adit by the old iron mine Grönligruben, a thin-splitting, light grey-coloured phyllite occurs in an about 3 m thick layer. The phyllite shows no folding, but locally fine crenulations may be seen on its schistosity planes. Outside the adit this phyllite is covered, so that it cannot be traced in the field. The rock, to all appearance an ordinary phyllite, has not been subject to microscopical investigation.

Greenstone breccia

In the extreme upper part of the Greenstone Unit, bordering on the Upper Sedimentary Unit, there locally occurs a greenstone variety which is interpreted as a sedimentary breccia. It differs from the overlying conglomerates by the angularity of its rock fragments, by its content of "jasper" fragments, and by its surficial similarity to the massive greenstones, except for the "jasper" fragments, which reveal its identity. The size of the rock fragments may reach about 10 cm, but is mostly about 3—5 cm. The rock is massive and green-coloured.

Besides "jasper" fragments small angular fragments of white-coloured quartzite and of marble are occasionally found. Most of the fragments are of fine-grained greenstones. The matrix of the breccia is composed of quartz, albite, calcite, chlorite, sericite, magnetite, and pyrite. The "jasper" fragments have acquired their designation because they are red-coloured and composed of quartz. This term, however, is a misnomer because the rock lacks the dense, flinty texture of jasper *s. s.* The rock is a quartzite, coloured red by finely dispersed hematite. The texture is a granoblastic mosaic texture composed of anhedral, equidimensional quartz grains, 0.1—0.3 mm in diameter. In addition to quartz a few grains of calcite occur. Hematite is found as a pigment consisting of submicroscopic grains. Apparently, a very small content of hematite is sufficient to give the "jasper" a bright red colour, provided the grain size of the pigment is small enough. In some cases the "jasper" acquires a spotted appearance caused by a patchy decoloration of the quartz brought about by the conversion of disseminated hematite to clusters of magnetite octahedra, averaging 0.5 mm in size. The "jasper" fragments thus transformed accordingly have irregular white and black spots set against a background of bright red.

In one locality (loc 235), on the south bank of the river Nævra, a larger concentration of "jasper" and magnetite was found. The rock is characterized by its relatively high content of *magnetite* occurring as discrete crystals about 0.5 mm in size, or as anhedral, filamentous aggregates. In addition to magnetite the rock is composed of quartz, calcite, chlorite, tourmaline, muscovite, sericite, pyrite, and submicroscopic grains of hematite as a pigment. In the thin section made from this rock, the modal composition is estimated at 40 volume per cent chlorite, 20 per cent quartz, 20 per cent magnetite, 15 per cent calcite, 3 per cent muscovite + sericite, and 2 per cent tourmaline. As in the rest of the breccias, the modal composition is difficult to obtain, owing to the heterogenity of the rock.—The strikingly large amount of magnetite and tourmaline may indicate that this rock represents a breccia that has been subject to the activity of volcanic gases from some nearby fumarole. To what degree volcanic activity has influenced the other greenstone breccias is uncertain.—The greenstone breccias are supposed to represent a weathering residue of the underlying lavas and tuffs, with an admixture of "jasper", which may have been transported from adjacent areas, or may be a rock of local derivation belonging to the original sequence now represented by the breccias.—Volcanic gases possibly have been an active agent in the weathering of the greenstones at this stratigraphic level. The volcanism of the Trondheim region did by no means come to an end at the close of the period of formation of the Stören greenstones, and it seems reasonable that volcanism has been in operation even in the interval between the formation of the rocks of the Stören Group and the deposition of the sediments of the Lower Hovin Group. However, no significant clues as to what kind of weathering has caused the formation of the greenstone breccia have been establisted, but, as mentioned above, the greenstone breccias are thought to represent a practically untransported weathering residue of the underlying greenstones.

THE UPPER SEDIMENTARY UNIT

Conglomerates

This rock unit lies to the north of the Greenstone Unit area, and stratigraphically above it. The boundary between the two units mainly follows the Nævra valley, in an east-westerly direction. This boundary has a much more straightforward and simple course than the boundary between the Greenstone Unit and the Lower Sedimentary Unit, and inasmuch as the primary objective of the mapping was to establish the extent of the ore-bearing rock unit of the Mostadmarka—Selbustrand area (viz. the Greenstone Unit), the boundary between the Greenstone Unit and the non-metalliferous Upper Sedimentary Unit forms a natural and convenient northern limit to the map. For this reason the knowledge of the character and areal distribution of the rocks of the Upper Sedimentary Unit is incomplete, and the following description accordingly rather fragmentary.—The material collected consists of conglomerates, sandstones, slates, tuffites, and tuffs.

The conglomerates are polymictic. They belong to approximately the same stratigraphic horizon, lowermost in the unit, but are macroscopically of different types.—In a small area about 1.5 km NE of Jösås where the river Nævra for a short distance follows a northwesterly course, the conglomerates occur in the following sequence from south to north, i.e. from bottom to top. First, and immediately above the greenstone breccia occurs a green-coloured conglomerate, distinguishable from the breccias by the rounding of its fragments only. Then follows a conglomerate with somewhat larger fragments, and with a carbonate-rich matrix. Next comes a light grey-coloured conglomerate with mostly white-coloured pebbles. The conglomerates have been strongly deformed, the pebbles have been flattened and elongated so that their dimensional proportions may be expressed by the following approximate relations: 1 : 3 : 12. The amount of stretching not uncommonly renders the conglomeratic nature of the rocks hardly discernible. In addition to the flattening and stretching, which is parallel to the trend of linear elements elsewhere in the area, another set of deformations have in places left their imprints on the conglomerates, in the shape of low-amplitude, small-scale flexuring, the axes of which make an angle of 60—90° with the direction of stretching. The thickness of the conglomerate beds is unknown due to the superficial cover in the densely forested area, but the individual varieties hardly exceed a few tens of metres each.

Green conglomerate. (Pl. IV, fig. 2)

The rock is composed mainly of pebbles of fine-grained greenstone, with additional pebbles of quartzite, quartz-keratophyre, and marble. The pebbles are set in a light green-coloured matrix. As an accessory, mm-big pyrite crystals are common. The pebbles have been strongly stretched, and are 1—10 cm long.—Under the microscope the pebbles are seen to consist of *fine-grained greenstone,* with albite, chlorite, and epidote as their main constituents, *porphyritic quartz keratophyre, quartzite,* and *marble.* The first three rock types are well known from the Greenstone Unit (Quartz keratophyre is described in a later part of the present paper).—The matrix of the conglomerate consists of quartz, albite, epidote, calcite, muscovite, sericite, chlorite, leucoxene, sphene, apatite, and ore grains. The texture is blastopsephitic.

Quartz is found in the groundmass either as large, rounded grains, 0.5—2.5 mm in diameter, or as small irregular grains of 0.05—0.1 mm. Considering the strong deformation the rock evidently has undergone, the quartz grains have resisted the stress remarkably well: The larger grains are slightly fractured, and a few show incipient undulating extinction. The smaller quartz grains display no strain phenomena, and there is nothing to indicate their having originated by the tectonic breaking up of larger grains.

Albite grains are mostly subhedral, and between 0.3—1.5 mm across. Occasionally they show undulating extinction, and some have been fractured, but have otherwise not been subject to deformation. The felspar frequently has been heavily saussuritized, the grains being completely filled with small epidote grains, only a narrow rim of clear albite being left, some of the felspar grains have been able to rid themselves of their epidote inclusions along strained or fractured zones.

The large grain size, and the morphology of the subhedral albite grains of the groundmass, and their relatively large amount, makes it probable that the conglomerate matrix has a substantial admixture of pyroclastic material. If the content of calcite were subtracted, the groundmass would probably show the modal composition and textural character of a quartz-rich quartz-keratophyre tuff. In the descriptive nomenclature of sedimentary petrography (Pettijohn 1949), the groundmass presumably would fall within the category calcareous arkose.

Conglomerate with carbonate-rich matrix

This conglomerate is found in the river bed of Nævra just to the north of the previous locality. Macroscopically, the rock is distinguished by its content of carbonate porphyroblasts, occurring both in the pebbles and in the matrix, with a somewhat higher concentration in the latter. On weathering, the carbonate is pseudomorphically replaced by limonite. Secondly, the conglomerate is characterized by the larger size of the pebbles, they are mostly 5—20 cm in their longest dimension, and seem to be less elongated than indicated by the previously presented axial ratio of pebbles.—Microscopical investigation shows the pebbles to be mostly *greenstone* varieties identical to the ones in the green-coloured conglomerate. In addition, scattered pebbles of white *marble* and small rounded *quartzite* pebbles also occur.—The matrix of this conglomerate differs in several ways from that of the former: Carbonate is present in definitely larger amounts, and the carbonate is iron-rich, as evidenced by its transformation to limonite, the proportion of albite relative to quartz is much less, the albite grains are much smaller, and they are not subhedral, but have often been distinctly rounded. The modal composition of the matrix is estimated at:

Carbonate	45	volume per cent
Quartz	33	„ „ „
Albite	12	„ „ „
Muscovite	10	„ „ „

Judged by the texture and modal composition, directly pyroclastic material plays a minor part only in the groundmass of the present conglomerate as compared to that of the underlying, green conglomerate.

Light grey-coloured conglomerate. (Pl. V, fig. 1)

This rock has not been observed in outcrop in the above-mentioned locality by Nævra, but was here found in rock material blasted from local bedrock during road construction. Yet, the conglomerate variety occurs in outcrops further west, in the vicinity of Torshaug (Loc. 228 and 229), with somewhat smaller pebbles.—The pebbles mostly are of *quartzite*, with a granoblastic texture, consisting of equigranular quartz grains, about 0.05 mm in size, and scattered flakes of sericite. Apart from quartzite fragments, pebbles of an *albite-rich greenstone* variety are also present. This rock has a blastophitic texture, with unoriented albite laths making up about 70 volume per cent of the rock set in a matrix of chlorite, calcite, and leucoxene aggregates. In the groundmass of this conglomerate albite is found as large, frequently subhedrally developed grains, in places in large amount. The overall composition of the matrix varies between felspathic quartzite and felspar-rich arkose; the content of calcite also varies.

In this conglomerate variety deformation is the most pronounced as seen under the microscope: Large quartz grains show marked undulating extinction, and some very fine-grained quartz aggregates probably represent crushed quartz grains. Grains of albite also frequently show undulating extinction, and partly crushed albite grains are often met wit. Some grains of albite seem to have reacted to stress by translational gliding: Twin lamellae have been displaced transversely in a zig-zag pattern, but the optic orientation of the grain has not been influenced by the stress. The large albite grains of the matrix of this conglomerate may be pyroclastic in origin.

Conglomerate north of Selbutjern

The colour of this conglomerate is greyish green. The pebbles consist of *quartzite, marble, quartz keratophyre,* and *greenstone* varieties. The pebbles have been stretched along a deeply plunging (60°) direction to the NE. Bedding is visible as pebble-free layers, 2—4 cm thick. The pebbles are mostly 4—10 cm in the longest dimension, but under the microscope pebbles down to 0.3 cm in diameter are also seen. The matrix differs from that of the other conglomerate varieties by its higher content of dark-coloured minerals, and by its fineness of grain, about 0.02—0.05 mm. The dark-coloured constituents, chlorite and leucoxene, have furnished material for the crystalloblastesis of amphibole and biotite.

The matrix of the present conglomerate resembles a fine-grained greenstone, but has a higher content of quartz than the meta-lavas of the Greenstone Unit. No textural criteria were found to establish whether or not pyroclastic material is present in the matrix.

Andesite tuff

Nævra valley, 750 m ENE of Torshaug (loc. 253)

This is a light greenish-grey medium- to fine-grained massive rock. It possesses a weak parting, and a weak linear structure, caused by a sub-parallel orientation of amphibole crystals. The tuff character is in general not macroscopically discernible. Under the microscope, the rock is seen to be a clastic rock, a crystal tuff. It contains large, up to 2 mm long subhedral *albite* grains. As a rule the grains are granulated, and appear as aggregates, with a grain size of the components of about 0.01 mm. However, the individual grains of the aggregates have the same optical orientation, and there can be no doubt that they are fragments of larger grains. In places even the polysynthetic twinning of the parent grain is reflected in the optical orientation of the fragments. Aggregates of chlorite and amphibole apparently are pseudomorphs after pyroxene or amphibole. The aggregates have a diameter of 0.2—0.4 mm.

The larger grains of colourless amphibole in places contain a relict core of a short-prismatic, deeply brown-coloured amphibole, with strong pleochroism: Deep brown—Light yellow brown. The colour indicates that this amphibole is a titaniferous hornblende. In other cases, the colourless amphibole may have a diffuse core of pale green, pleochroic amphibole, with a somewhat larger extinction angle than that of the rim.

In this tuff a *conglomeratic layer* occurs (loc. 254). The layer is about 40 cm broad on the rock surface, the actual thickness somewhat less. Scattered pebbles may be found in the tuff even outside this sharply defined layer. The pebbles have been stretched, and are about 10—12 cm long and about 3 cm across. The pebbles are mostly light-coloured, but scattered specimens of a dense, dark, green-coloured rock are seen also. Under the microscope the pebble material is seen to consist of fine-grained *quartz keratophyre, quartzite,* large *quartz grains,* large, pseudomorph-like *chlorite aggregates, greenstone fragments,* with heavily saussuritized plagioclase laths in a groundmass of chlorite and leucoxene. The texture of the greenstone pebbles is blastophitic. In addition, another green-coloured rock occurs as pebbles. It is a dark, fine-grained *amphibole-epidote-leucoxene-chlorite rock,* the texture of which indicates a pyroclastic origin of the rock: It has chlorite aggregates which evidently forms pseudomorphs after crystals and angular shreds and crystal-fragments of a previous mineral. Some of the pseudomorphs very strongly suggest the crystal habit of pyroxene. Within the chlorite aggregates newly-formed amphibole is found as thin, colourless needles. The chlorite aggregates are closely set in a groundmass mostly consisting of leucoxene and very fine-grained epidote. Some of the pebbles of quartz keratophyre seem to have been angular before the elongation.

However, the original degree of rounding is not easily determined.

A more fine-grained variety of the andesite tuff is found at loc. 257. It is composed of small, angular aggregates of chlorite and of fine-grained albite, closely set in a groundmass mostly consisting of leucoxene. The aggregates are from 0.05 to 0.2 mm in their longest dimension.

This rock variety shows no sign of deformation.

Dacitic tuff with admixed sedimentary material

This rock occurs 1 km NE of Brennfjell, in the north-eastern corner of the map area, in road cuts by highway 665 (Hell—Selbu). The rock is light grey-coloured, massive, and medium-grained. In hand-specimen it is seen to be composed essentially of plagioclase felspar, quartz, and unoriented flakes of biotite, and with accessory pyrite. The clastic character of this rock is evident where it contains up to several metres thick intercalated layers of coarse epiclastic material (conglomerate), with, among others, pebbles of marble. In localities without intercalated conglomerates the origin of the rock is not easily determined. However, in some places one may, even outside the conglomerate layers find traces of bedding, or observe angular fragments of felspar, up to 0.5 cm in size. Under the microscope, the massive rock is seen to be composed of: Albite, quartz, biotite, chlorite, sericite, calcite, epidote, leucoxene, amphibole, sphene, apatite, zircon, ore. The modal composition was found to be:

Albite	45	volume per cent
Quartz	17	,, ,, ,,
Biotite	9	,, ,, ,,
Chlorite	9	,, ,, ,,
Epidote	8	,, ,, ,,
Calcite	9	,, ,, ,,
Accessories	3	,, ,, ,,
	100	

The texture of the massive rock is granoblastic-blastopsammitic, with lepidoblastic biotite and chlorite. The outlines of the clastic quartz and albite grains are delineated by the scarce chlorite-sericite-leucoxene-epidote matrix. The rock evidently has been subjected to a higher degree of recrystallization than have the other pyroclastic rocks of the area.

The *conglomerate layers* within this tuff contain pebbles of *quartzite, keratophyre, greenstone,* and *marble,* the latter apparently in the largest amount. The pebbles are stretched, and they are about 2—5 cm in their longest dimension (Pl. V, fig. 2).

The texture of this rock implies a clastic origin. The modal composition corresponds to that of an igneous rock within the quartz diorite-dacite family. The interpretation of the rock as a tuff conforms to both of these relations. It seems possible that the admixing of epiclastic material, as evidenced by the conglomerate layers, is also responsible for a substantial amount of the quartz of the rock. Thus, the purely pyroclastic material may correspond to a tuff of andesitic composition, closely related to the previously mentioned andesitic tuffs.

Calcareous arkose

N side of the lakelet Svarttjern, 500 m SE of Brennfjell (loc. 142) This rock is a massive, psammitic, grey-coloured rock. Primary bedding is represented by thin (about 0.3 mm) dark layers. The spacing of these layers varies between 1 mm and 2 cm.—Under the microscope the rock is seen to consist of: Calcite, albite, quartz, muscovite, chlorite, tourmaline, zircon, rutile, ore. The modal composition is estimated at:

Calcite	35	volume per cent
Albite	40	„ „ „
Quartz	15	„ „ „
Muscovite	7	„ „ „
Accessories	3	„ „ „
	100	

The texture is granoblastic, equigranular, with grain size about 0.1—0.2 mm. The modal composition places this rock in the category calcareous arkose. The marked dominance of albite over quartz makes it reasonable that even this rock contains a substantial amount of tuffaceous material. However, no corroborating evidence having been found, the descriptive designation calcareous arkose is preferred to the genetic term tuffite.

Fine-grained sandstones and shales

Several varieties of these rocks have been found. The most common, apparently dominating the lithology of the Lower Hovin Group in the area from the above-mentioned conglomerate- and tuff horizons and northwards toward Leksdalen, is a fine-grained, bedded, green-coloured albite-chlorite-sericite-quartz schist. It shows two cleavage directions, one along the bedding, the other a fracture cleavage along the axial planes of small folds (Fig. 9). Bedding is conspicuous as bands of differing shades of green, and the individual layers vary from a few mm to 4—5 cm in thickness. The rock is composed of: *Quartz, albite, muscovite, sericite, chlorite, epidote, sphene, leucoxene, tourmaline, rutile,* and *apatite.* The texture is even-grained, granoblastic. Under the microscope, bedding is seen as bands varying in their relative content of chlorite and sericite, and in variation of grain size of quartz and felspar grains.

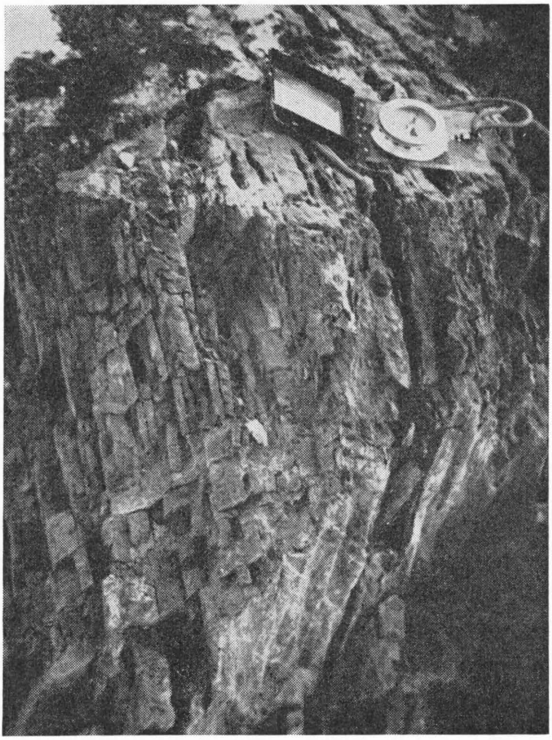

Fig. 9. Fine-grained, bedded green-coloured schist with fracture cleavage. Upper Sedimentary Unit.

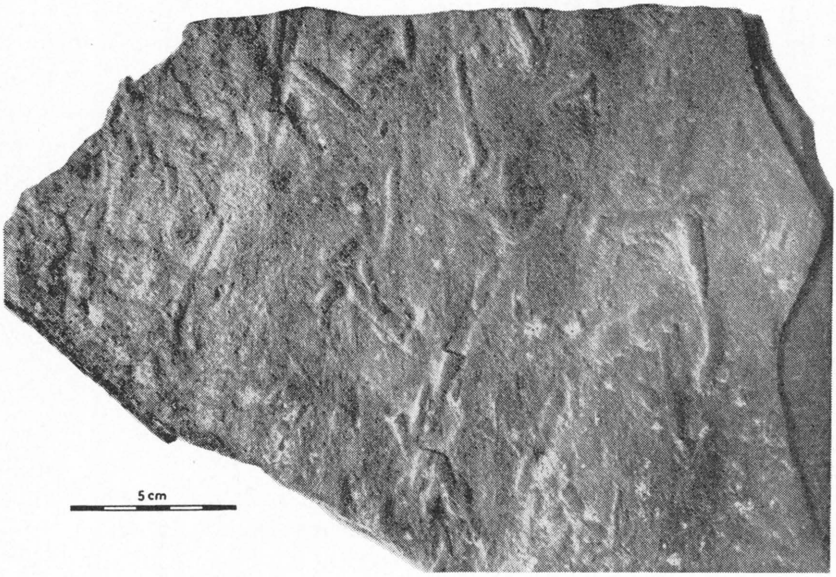

Fig. 10. Worm trails in slate, Loc. 144a, quarry 2 km N of Brennfjell.

The grain size is about 0.05—0.1 mm but is in the mica-rich layers reduced to 0.02—0.05 mm. The relative proportions between quartz + felspar and the micas vary from layer to layer, so that pelitic material makes up about 15 and 25 volume per cent in the coarser and finer-grained layers, respectively. Systematically, the rock is a greywacke shale to arkosic shale (Pettijohn 1949), composed of fragments varying in size from silt to very fine sand.

Another variety is *sericite-albite-carbonate-quartz schist.* This too is fine-grained, but the colour is light greyish. It is bedded, with alternating light and darker grey-coloured layers. It shows the same cleavage directions as the green shale. The grain size is about 0.03 mm. This variety contains a substantial amount of calcite, estimated at about 30 volume per cent, while pelitic material is more subordinate than in the green shale. Pyrite occurs as an accessory mineral, with cubes measuring about 0.5 cm. In connection with the pyrite cubes, "shadow quartz" has developed, growing perpendicular to the crystal planes of the pyrite.

In two localities within the north-eastern part of the map area, minor slate quarries have been in operation, 1 km ESE and 2 km N of Brennfjell, respectively. The slate is non-folded and thin-splitting. The modal composition of the slate corresponds closely to that of the green shale; however, chlorite seems to be present in somewhat larger amounts. One of the slates (loc. 144) contains small rutile needles (0.02—0.1 mm long) in great profusion. The other slate variety is of interest because it has been found to contain fossils (H. Carstens 1960). During the present investigation, imprints, to all probability representing worm trails, were found in this slate (Fig. 10).

II. Intrusive rocks

Saussurite gabbro. (Pl. VI, fig. 1)

This rock type occurs abundantly as small bodies within the Greenstone Unit; particularly in the central parts of the map area, in the vicinity of Vennafjell—Lillefjell. The rocks are relatively coarse-grained, a feature indicative of their intrusive origin. However, cross-cutting relationships to the wall-rock have not been found, nor have contact phenomena. In the field, the contact between the gabbros and the surrounding greenstone is hardly discernible because the former usually lose their coarseness of grain towards the border, eventually to become rather fine-grained, and thus difficult to distinguish from the coarser varieties of the massive greenstones. During the mapping a few instances occurred when coarse-grained greenstones were mapped as fine-grained gabbros until unmistakable pillow structures revealed their extrusive origin.—The rock is a green-coloured, massive, coarse- to medium-grained amphibole-plagioclase rock, with a light-coloured, often purely white, plagioclase. A modal analysis gave the following result:

Plagioclase	39	volume per cent
Amphibole	30	” ” ”
Epidote	20	” ” ”
Chlorite	8	” ” ”
Leucoxene	2.5	” ” ”
Sericite	0.5	” ” ”
	100	

As a rule, *plagioclase* is developed as broad laths, 1—4 mm long. The grains frequently are cataclastically granulated. Polysynthetic twinning is common, with rather broad and diffuse lamellae. Some of the plagioclase grains are densely saussuritized, while others are weakly sericitized. The small grains resulting from the granulation of bigger grains are free from inclusions, and untwinned. The optical determination of the plagioclase gave results varying from An_9 to An_{13}.—In some instances the broken grains have been "healed" by recrystallization of the granulated zones, and in places inclusions of small amphibole needles indicate the cemented ruptures.

Amphibole occurs in two varieties: One distinctly green-coloured and markedly pleochronic, the other colourless in thin section. The colourless variety frequently contains diffuse relicts of the green-coloured variety. Optical determinations gave these values: Colourless amphibole: n_z: 1.625, n_x: 1.650. Extinction angle $Z \wedge c$: 17.7°. This indicates an actinolitic hornblende. Green amphibole: n_z: 1.655, $Z \wedge c$: 23.7°, pleochroism: Z: Bluish green, Y: Green, X: Weak yellow green. The data indicate common hornblende. The amphibole commonly constitutes a matrix to the plagioclase grains, forming a blastophitic texture. In some samples amphibole-chlorite aggregates occur in well-defined units, apparently forming pseudomorphs after short-prismatic pyroxene crystals, i.e. uralite. No pyroxene have been found.

Epidote-clinozoisite grains range in size from 0.01 to 0.2 mm in diameter. The smallest grains are found as inclusions in plagioclase, or are forming narrow veins obviously representing cracks and small-scale faults within the rock. The bigger epidote grains are evenly distributed throughout the rock. The birefringence of the bigger epidote grains was found to be 0.0264, indicating an epidote with c. 19 per cent Fe-epidote molecule.

Chlorite is found in two varieties: One very weakly green-coloured, with a barely discernible pleochronism, greenish grey interference colours, 2V: 0°, and optically positive. The other chlorite variety is strongly green-coloured, with marked pleochroism: Z: Yellow, X, Y: Green. Interference colours are dark brownish to blue violet. Both varieties are associated with amphibole, colourless chlorite together with the colourless amphibole, the other with the green-coloured amphibole. The associations colourless chlorite-colourless amphibole and green chlorite-green amphibole have not been found occurring together in the same rock sample. *Leucoxene* occurs as nearly opaque oval-shaped aggregates, 0.1—0.2 mm in diameter. Sometimes the mineral forms pseudomorphs after skeletal crystals of an ore mineral, probably ilmenite.

This manner of occurrence is found in the gabbro with strongly coloured chlorite and amphibole, and it is tentatively assumed that the pseudomorphs represent ilmenite grains that have been partly resorbed by the rock melt in the later stages of crystallization, the process furnishing iron to the silicates. *Sericite* is found as inclusions in the bigger plagioclase grains.

The metamorphic grade corresponds to the Greenschist facies, quartz-albite-epidote-biotite subfacies, with green hornblende remaining as relics of an earlier somewhat higher facies.

Quartz keratophyre

This designation comprises all the fine-grained to dense or porphyritic intrusive rocks with albite and quartz as their main constituents. Similar rocks have been described both from other areas of the Trondheim Region (Vogt 1945, H. Carstens 1960) and from other regions of the Caledonian mountain chain in Norway (Foslie 1955).

The rocks may show slight variations in their textural relations, and some have mineral compositions that in detail deviate from the majority of the quartz keratophyres, but common to all of them is that they are fine-grained, or porphyritic, and that they are essentially composed of quartz and albite.

The several occurrences are small, hardly any of them exceed 100 m along the strike, and the thicknesses for the most part are from a few up to about fifteen metres. The bodies form sills, and are interpreted as phacolitic intrusives. No sign of contact metamorphism of the landrock has been found anywhere. The limited longitudinal extent of the rock bodies is taken to indicate their intrusive rather than extrusive origin. The wide scattering of the occurrences, from the upper part of the Lower Sedimentary Unit to the top of the Greenstone Unit is also thought to point in this direction. A clearly intrusive contact between quartz keratophyre and greenstone was found just outside the mapped area, about 1 km to the east of the lake Dragstsjöen, by the roadside at the highest point of the road to Amdal (loc. 108) (Fig. 11). One is, of course, not justified in applying the relations at this one locality to all similar rocks in the area, but, on the other hand, no indications of an extrusive origin have anywhere been found. Besides, the textures of some of the quartz keratophyres show a lack of deformation that seems incompatible with the idea of their being contemporary with the surrounding rocks.

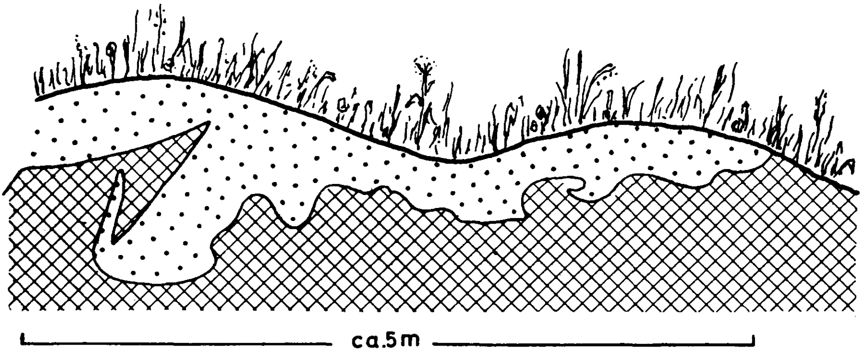

Fig. 11. Quartz keratophyre (stippled) intrusive into massive greenstone (cross-ruled). Loc. 108 (W of map area).

The most common varieties of the quartz keratophyres are white-coloured and dense, or porphyritic with dense groundmass. The non-porphyritic quartz keratophyres have the same texture and mineral composition as the groundmass of the porphyritic varieties. The texture is allotriomorphic-granular, with grain sizes about 0.02—0.05 mm. The mineral composition is: Quartz, albite, muscovite, sericite, epidote—clinozoisite, leucoxene, chlorite, calcite, rutile, apatite, ore, limonite. The modal composition of the rock vary but little, and the following analysis is fairly representative:

Albite	59	volume per cent
Quartz	30	„ „ „
Muscovite (sericite)	7	„ „ „
Chlorite	2.5	„ „ „
Accessories	1.5	„ „ „
	100	

Albite and *quartz* are anhedrally developed, and are evenly distributed through the rock. Some of the albite grains are twinned, but in most cases only the relative relief between the two minerals makes their distinction possible.

The *porphyritic variety* of quartz keratophyre differs from the non-porphyritic variety by its content of quartz and albite phenocrysts. The relative proportion between phenocrysts and groundmass varies considerably, and practically all transitions may be found between the non-porphyritic type and the one with the highest content of phenocrysts (Pl. VI, fig. 2).—The *albite phenocrysts* are subhedral, and as a rule are twinned, commonly after more than one twin law. The grains are always slightly sericitized, and many contain a few small grains of clinozoisite-epidote. The albite phenocrysts are about 1—2 mm in size.—*Quartz* is a much less prominent phenocryst mineral than is albite. The quartz grains have grain diameters of about 0.5 mm, and they are always irregularly roundish and indentated, probably as a result of partial resorption by the rock melt (T. Vogt 1945).

Amphibole-stilpnomelane-quartz-keratophyre (Loc. 147, SE of Vennafjell). Macroscopically, this rock is massive, medium- to fine-grained, with mm-big flakes of unoriented dark brown-coloured brittle mica, and dark green-coloured needles of amphibole set in a white-coloured quartz-felspar groundmass. The rock disintegrates easily, because of the profusion of unoriented mica, along the perfect (001)-cleavage of the mica flakes. Under the microscope (Pl. VII, fig. 1), the rock is seen to have a nemato- to lepidoblastic porphyritic texture, with a fine-grained allotriomorphic-granular groundmass. The mineral composition is: Albite, quartz, stilpnomelane, amphibole, magnetite, leucoxene, sphene.—The modal composition, as determined by the integration stage method, is:

Albite	59	volume	per	cent
Quartz	17	,,	,,	,,
Stilpnomelane	12	,,	,,	,,
Amphibole	8	,,	,,	,,
Magnetite	3	,,	,,	,,
Leucoxene	1	,,	,,	,,
	100	volume	per	cent

The fine-grained groundmass consists mostly of anhedral *albite* and *quartz,* with grain size mostly between 0.01—0.02 mm, some grains reaching 0.05 mm in diameter. The negative relief is evidence of the plagioclase being albite.

Phenocrysts of *albite* are anhedral, commonly partly granulated, and they show incipient undulating extinction. Most of the grains have short irregular twin lamellae, scattered irregularly in the grains.

The phenocrysts of *quartz* occur in small amount, and show rounded, partly resorbed grains. The phenocrysts have grain sizes around 0.5 mm.

Stilpnomelane is a strongly brown-coloured mineral. It occurs as elongate slender flakes, up to 1 mm long. In sections cut parallel to the crystallographic c-axis the mineral takes the appearance of long thin needles, the proportion length : width of which may be up to 50 : 1. The mineral has a perfect cleavage parallel to (001), and an imperfect cleavage perpendicular to this. This is a feature to distinguish stilpnomelane from biotite. In addition, the present mineral lacks the characteristic mottled appearance of biotite near the extinction position. The pleochroism is very strong; Z, Y: Dark brown (nearly sub-opaque), X: Yellow, absorption Z, Y » X. Optic character is negative, $2V_X$ is 0°. Elongation positive. Refractive index measurements gave as a result: (Na-light) n_Z, n_Y: 1.731. According to Tröger (1959) this indicates a stilpnomelane with about 13 mol. per cent ferro-stilpnomelane.

Amphibole is present with in part an extremely long-prismatic crystal habit, the proportion length : width occasionally reaching 30 : 1. The needles are up to 1.5 mm long, and the mineral is strongly green-coloured with strong pleochroism: Z: Bluish green, Y: Green, X: Greenish yellow, absorption Z, Y > X. Maximum extinction angle Z to c was measured at 27.5°. $2V_X$ is about 41° (U-stage measurements), with extreme axial dispersion: $r > v$. The optical data indicate a mineral of the hastingsite group.—*Magnetite* is scattered in the rock as small roundish grains, about 0.02 mm in diameter. In places aggregates of magnetite and leucoxene occur, and they may have shapes indicating pseudomorphs after an earlier mineral, to all appearance an amphibole was the original dark-coloured constituent of the rock. Apparently hastingsite and stilpnomelane have crystallized simultaneously: Stilpnomelane crystals cut through amphibole needles, and amphibole needles cut through stilpnomelane.—*Leucoxene* occurs mostly together with magnetite in aggregates.

SW of Jösåsvann a similar rock occurs (loc. 220). Apart from the association stilpnomelane—amphibole it is texturally and mineralogically identical with the porphyritic quartz keratophyres, containing up to 1 mm big subhedral phenocrysts of albite. All the phenocrysts of this rock are somewhat sericitized, while the groundmass is devoid of sericite or muscovite. The amount of stilpnomelane and amphibole in this variety is distinctly smaller than that of the foregoing rock. An interesting feature of this rock is that between most of the phenocrysts zones of relatively coarse quartz grains are arranged (Pl. VII, fig. 2). Inasmuch as all the phenocrysts lack any sign of deformation this phenomenon is probably a primary fluidal texture of the viscous melt.

Tourmaline-bearing quartz keratophyre (loc. 54). This rock is found in a road cut about 1 km N of Vottafjell. It is a white- to light greyish green-coloured fine-grained rock, occurring as a 4 m thick vertical sill in somewhat schistose greenstone. The sill cannot be followed for more than about 50 m NE-wards in the relatively densely covered and hilly area. A weakly developed lineation, caused by parallel oriented mica can be seen in hand specimen. Thin, dark-coloured needles of tourmaline, up to 0.5 cm long, are sparsely scattered in the rock. The mineral composition is: Quartz, albite, epidote-clinozoisite, muscovite, calcite, tourmaline, chlorite, sphene, and a zeolite. The texture is granoblastic-porphyroblastic, the groundmass mostly consisting of allotriomorphic-granular albite and quartz, with grain sizes about 0.05—0.1 mm, and with megacrysts of tourmaline and epidote-clinozoisite.

Tourmaline is found as prismatic porphyroblasts, making up about 2 volume per cent of the rock. The mineral is poikiloblastically developed, with inclusions of quartz and albite. The colour is very weak for a tourmaline, and so is the pleochroism: The colour is pale green, and the pleochroism is O: Green, E: Apparently colourless. Refractive index measurements gave these values (Na-light): n_O: 1.644 n_E: ca. 1.622.— The rock is transected by narrow veins of an as yet undetermined zeolite mineral.

Near Lövås (loc. 15) occurs a *quartz keratophyre-like rock* with texture and mineral composition differing somewhat from the quartz keratophyres previously described. The rock is a fine-grained muscovite-quartz-albite rock, white-coloured with a yellowish tint. It contains 0.5 mm big garnets, and small spots of limonite as a decomposition product after ore grains. The rock has a lineation, brought about by the orientation of muscovite. The mineral composition is: Quartz, albite, muscovite, garnet, apatite, limonite.—Garnet has not been found in any of the other quartz keratophyres. An additional difference is that quartz is present in larger amounts in this rock, it is estimated to equal albite in volume. Muscovite also is present in somewhat larger amount than in the ordinary quartz keratophyres within the field. The texture is characterized by porphyroblastic garnet, and by phenocrysts of quartz and felspar set in a fine-grained allotriomorphic groundmass of quartz, felspar, and lepidoblastic muscovite.

The felspar phenocrysts of this rock are surrounded by a symplectitic rim consisting of albite and quartz (Pl. VIII, fig. 1). The *albite phenocrysts* are about 1.5 mm in diameter, and they are an- to subhedral. They are sharply bordered against the surrounding symplectite rims.

The origin of this rock is uncertain, but owing to the limited extent in the field, and to the many mineralogical similarities to the other acid intrusives, it is thought to be an intrusive rock. One cannot, however, exclude the possibility of a sedimentary origin, in which case the garnets and the albite-quartz symplectite must be the products of metamorphism. The surrounding rocks are all in a low metamorphic facies, and their recrystallization is not so complete as to suggest the possibility of a retrogressive metamorphism on their part.—The symplectite is thought to have developed during the late-magmatic stage, after the formation of the phenocrysts, but during the crystallization of the groundmass, as indicated by the fact that symplectites have developed around the phenocryst albite only. As to the time of formation of the garnets only this can be said: Apparently small grains of quartz and felspar existed beforehand.

Hornblende porphyrite (loc. 154)

This is a massive, light-coloured, medium- to fine-grained rock, with a greenish tint caused by the presence of small phenocrysts of amphibole. In hand specimen, only amphibole and albite can be distinguished. The amphibole grains are mostly about 1 mm long, but some needles may attain a length of 5 mm. The mineral composition is: Albite, amphibole, epidote, chlorite, quartz, sphene, tourmaline, apatite, zircon, limonite. The modal composition is:

Albite	55	volume per cent
Amphibole	36	„ „ „
Epidote	8	„ „ „
Accessories	1	„ „ „
	100	volume per cent

The texture is porphyritic, with subhedral phenocrysts of amphibole, set in a groundmass mostly consisting of equigranular albite.

Trondhjemite

This rock has been found in one outcrop, about 300 m SSW of the lakelet Haukåstjern (loc. 97). It has been intruded into the quartz schist of the lower sedimentary unit.—The rock is medium grained and massive, on fresh surfaces it is light grey-coloured, with small dark spots of chlorite.—The mineral composition is: Plagioclase, quartz, chlorite, muscovite, calcite, rutile, sphene, leucoxene, apatite.—The texture is hypidiomorphic to allotriomorphic granular, with semi-porphyritic plagioclase grains set in a groundmass of quartz and chlorite. The *plagioclase* grains are mostly rectangular in outline, the grain size mostly between 1 and 2.5 mm. The plagioclase is strongly sericitized, with, as a rule, a clear rim.

The chlorite carries inclusions of rutile, often developed in a sagenitic pattern. This association indicates that the chlorite is a decomposition product after biotite, which is the primary dark constituent of the typical trondhjemites.

The rock appears to be in chemical inequilibrium. The plagioclase is relatively basic (Oligoclase An_{16}), while biotite has been completely altered to chlorite. The felspar lacks any sign of saussuritization, and no traces of epidote have been found.—The metamorphic grade is supposed to be equivalent to the Greenschist facies, Quartz-albite-muscovite-chlorite subfacies, with oligoclase as a metastable relict mineral.

Amphibole-epidote-biotite schist

SE of Vennafjell (loc. 171)

This rock is unique within the present area. It is interpreted as an ultrabasic intrusive rock, completely recrystallized and metamorphosed into the Greenschist facies. No positive criteria for its intrusive character can, however, be brought forward. The schist is greenish black, fine-grained and very tough. It cleaves into thick flags (3—4 cm), and shows no visible folding. The rock is concordant, and scarcely more than a few m thick. In the field it is associated with a porphyritic greenstone of extrusive origin, occurring in fairly close proximity to saussurite gabbro. An outcrop of amphibole-stilpnomelane-quartz keratophyre is also found nearby. The lateral extent of the schist is unknown, the surroundings of the outcrop being rather heavily covered. The rock is spotted with about 2 mm big epidote nodules.

Microscopical investigation reveals a very simple mineral composition: Biotite, epidote, amphibole, leucoxene. The modal composition is:

Biotite	47	volume	per	cent
Epidote	38	,,	,,	,,
Amphibole	8	,,	,,	,,
Leucoxene	7	,,	,,	,,
	100	volume	per	cent

The texture is lepidoblastic-granoblastic, even-grained, with grain size about 0.1 mm.

The *biotite* is strongly pleochroic: X: Light yellow, Y,Z: Olive green. $2V_X$ is 0°. Axial dispersion: $v > r$. Refractive index measurements gave as a result: $n_{Y,Z}$: 1.638. Birefringence was found to be 0.046. According to Tröger (1959) this indicates a lepidomelane.

Epidote grains are often associated to form roundish aggregates. The reason for this aggregation is unknown, but the aggregates probably represent pseudomorphs after an earlier mineral.

Amphibole occurs as scattered subhedral and nematoblastic prisms and needles of bluish green colour. The grains are 0.2—0.3 mm long. The amphibole is distinctly pleochroic, with Z: Bluish green, Y: Green, and X: Light greenish yellow. Maximum extinction angle was measured at 20.5°. These relations indicate that the mineral is common hornblende.

Leucoxene is found as irregular oblong aggregates, 0.05—0.3 mm long. They are evenly scattered in the rock, and have a sub-parallel orientation, curving around the epidote aggregates in a fashion reminiscent of fluidal texture. Inside the epidote aggregates the leucoxene is arranged as rod-shaped aggregates, with an orientation independent of those in the matrix. The rods are arranged in a pattern where they cross each other at an angle of 60°, indicating a sort of palimpsest sagenitic texture. A reasonable explanation to this is that the epidote nodules really are replacing a previous mineral.

In accord with its peculiar modal composition, the rock is very heavy, and by the pychnometer method its specific gravity was found to be 3.22. The most distinctive feature of this rock, as opposed to the other rocks of the area, is the prominent dominance of K over Na, as brought out by the modal composition. The other rocks of the area are Na-rich, and the igneous rocks are generally accepted as belonging to the spilites.

The complete lack of light-coloured minerals makes it difficult to determine the metamorphic facies of this rock, but the identification of the amphibole as common hornblende makes it probable that it belongs in the Greenschist facies, Quartz-albite-epidote-almandine subfacies.

III. Ore deposits

The great majority of the ore deposits are now inaccessible. The pits and adits are filled with water, and many of the old workings have not been found, even the local population having forgotten their whereabouts, since the sites are overgrown with dense spruce forest. A useful aid has been the old reports in the Bergarkivet (Mining archives) from explorations and surveys by mining engineers in earlier times, when features could be studied that today are out of reach.

The ore deposits may, economically and technically, be divided into sulphide deposits and iron ore deposits, none of which, however, are of any economic importance by present-day requirements. In earlier times, from 1657 to 1870, the iron ore deposits were worked by the company Mostadmarkens Jernverk. J. H. L. Vogt (1910) estimated the total amount of ore smelted by this company at 25—30,000 tons.

The iron ores mainly consist of magnetite, but small amounts of hematite are also present. The sulphide deposits are of the kind which C. W. Carstens (1919) designated the Leksdal type, and to which he assigned an exhalative-biochemical genesis. The Leksdal type sulphide deposits are characterized by the following features: The ore minerals are dominantly pyrite and pyrrhotite. Chalcopyrite may be present in insignificant amounts. Magnetite is a ubiquitous constituent of the Leksdal type ores.

The ore deposits within the mapped area show transitions from sulphide deposits of the Leksdal type to pure iron oxide deposits with magnetite as their main constituent and with varying amounts of hematite. An interesting feature is the areal distribution of the ore types. The sulphide deposits are found in the southern part of the greenstone area, while the iron-ores and the mixed iron- and sulphide-ores occur in the northern part. The two latter types seem to be connected to the upper tuff horizon.—The transition ores are bedded, with alternating layers of pyrite and magnetite a few mm to about 1 cm thick. The individual layers as a rule have sharp boundaries against each other. The material collected from the ore deposits have mostly been taken from the waste heaps near the old workings.

Sulphide ores

These occurrences consist of very fine-grained to dense pyrite-pyrrhotite ore bodies of the Leksdal type. The ore mineral paragenesis is: *Pyrite, pyrrhotite,* and smaller amounts of *magnetite.* The colour varies from grey to nearly black in hand specimen. Thin (less than 1 mm) irregular transecting veins of yellow pyrite are ubiquitous. They are of secondary origin. In places bedding is seen as thin laminae of varying colours and grain sizes (Fig. 12). In other cases the ore is massive. The dark colours of this type of ore have been assigned to their content of carbon (C. W. Carstens).—Survey reports dating from the years about the First World War show that the deposits are of no economic value. The ore bodies are too small, and their content of Cu is negligible, ranging from 0.03 og 0.13 per cent. Contents of S vary between 30 and 40 per cent.

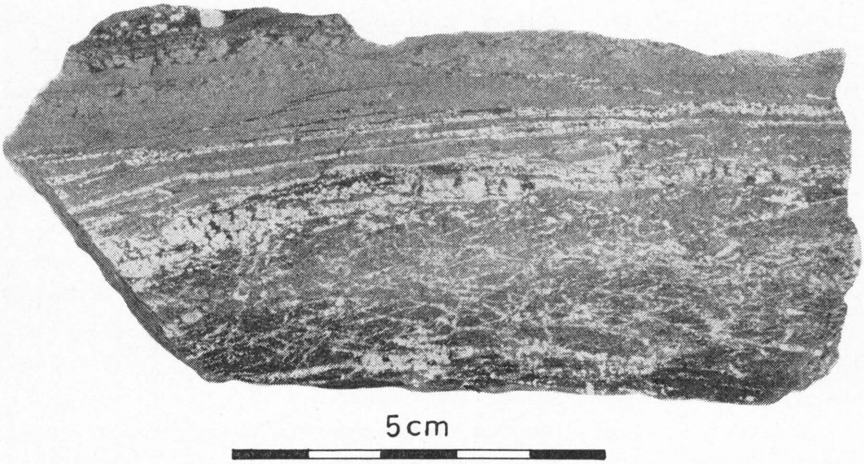

Fig. 12. Finely laminated carbonaceous pyrite-pyrrhotite ore (vasskis), dissected by thin veins of recrystallized pyrite. Loc. 157.

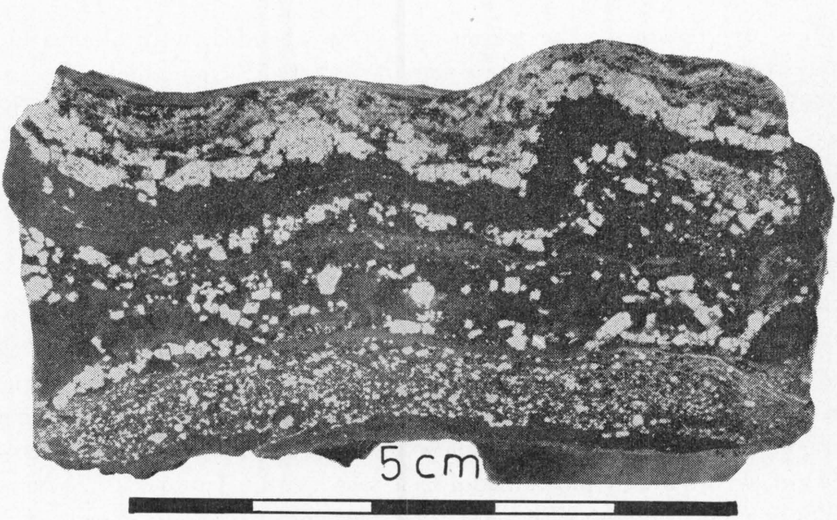

Fig. 13. Layered pyrite-magnetite ore. Loc. 64.

Layered pyrite-magnetite ores

The typical occurrences of this type are found in the vicinity of Værnesvold (Loc. 64, 270). The rock is layered with alternating magnetite- and pyrite-rich layers (Fig. 13). The thickness of individual layers are mostly 0.5—1 cm. In addition these ores are frequently interbedded with thicker layers of so-called black fels (Norwegian: Svartfels), a quartz-magnetite rock, which is also found as single beds and lenses within the greenstones, particularly interbedded with tuffs. In hand-specimens of the pyrite-magnetite rock the bedding is seen as alternating dense magnetite-rich layers and layers of more coarsely crystalline pyrite.

Under the microscope, the magnetite-rich layers are seen to consist of small *magnetite* grains, 0.01—0.02 mm in diameter, and anhedral *quartz* grains, about 0.05 mm in size, with a granoblastic mosaic texture, and scattered small grains of *epidote.*

The pyrite-rich layers have *pyrite* instead of magnetite, with grain sizes about 0.5—1 mm. In these layers mm-big flakes of a strongly green-coloured and pleochroic *chlorite* are also abundant. To all appearance this chlorite is identical with the one mentioned below as a constituent of the black fels.

Black fels (Svartfels)

(Pl. VIII, fig. 2)

This rock has been described as a dense, massive, magnetite-quartz rock (C. W. Carstens 1922, Falkenberg 1914). In several places within the present area there occur rocks that in appearance and mineral composition equal such black fels, and the name is retained here. The rock is heavy, dense, massive and with a choncoidal fracture. The mineral composition is: Magnetite, quartz, chlorite, epidote. The modal composition was found to be:

Magnetite	53	volume per cent
Quartz	41	” ” ”
Chlorite	5	” ” ”
Epidote	1	” ” ”
	100	volume per cent

Magnetite occurs in distinct layers with thicknesses varying from 0.05 to 1.5 mm. The distances between the magnetite-rich layers are 0.1—1.0 mm. The magnetite grains are about 0.01 mm across.

The *quartz* grains vary between 0.03—0.05 mm, and so do the *epidote* grains.

Chlorite is developed as porphyroblasts, with flakes about 0.3—0.5 mm in their longest dimension. The mineral is strongly green-coloured, with pleochroism: Z, Y: Deep green, X: Light yellow. Optic sign is negative, $2V_X$: 0°. Interference colours are blue- to red violet. Refractive index measurements gave as a result $n_{Y,Z}$: 1.635. These relations indicate that the mineral belongs to the prochlorite group, close to the transition prochlorite—Fe-prochlorite (Tröger 1959).

There are no textural or mineralogical indications to distinguish the black fels from the magnetite-rich layers of the mixed sulphide-iron-ore deposits.

An interesting feature of the black fels described here is the fact that the chlorite porphyroblasts evidently have been rotated: They carry helicitically included trains of magnetite grains, arranged in unbroken continuation of the magnetite layers of the matrix. Within the chlorite flakes these layers have an orientation deviating from that in the matrix (Fig. 14). In some cases the magnetite layers change their direction abruptly upon entering the chlorite, while in other cases they

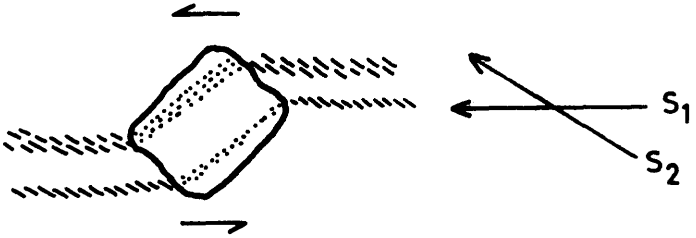

Fig. 14. Rotated chlorite porphyroblast in black fels.

follow a sigmoidal course, indicating slow rotation during the porphyroblastesis. In addition to the bedding, the microscopical investigation reveals the presence of a s-plane of tectonic origin, in the plane of the thin section making an angle of about 20° with the former. It occurs as a parallel orientation of the magnetite grains oblique to the bedding, and a slight elongation of quartz grains. Apparently the formation of this s-plane was simultaneous with the rotation of the chlorite flakes. The degree of rotation varies, but the direction is always the same.

Layers of black fels also occur in the greenstones independent of the ore deposits proper. They are lens-shaped, 1 to about 10 m long in the direction of strike (possibly longer), and 10 to about 40 cm thick. They seem to occur exclusively within tuff beds and have been found in the Venna valley. In places they may be irregularly folded (Fig. 15).

Iron ores

The iron ores consist of magnetite with varying amounts of hematite. The deposits that apparently have been extensively worked are: The Grönli mine (loc. 65), The Kleptjern mine (238), the Nonshaug mine (226), and the Heinberg mine (246).

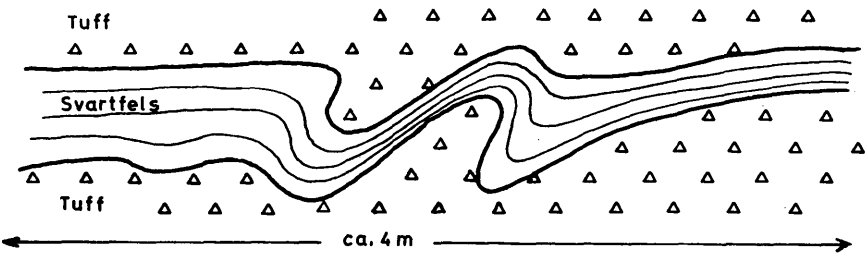

Fig. 15. Layer of black fels in green tuff. Venna valley.

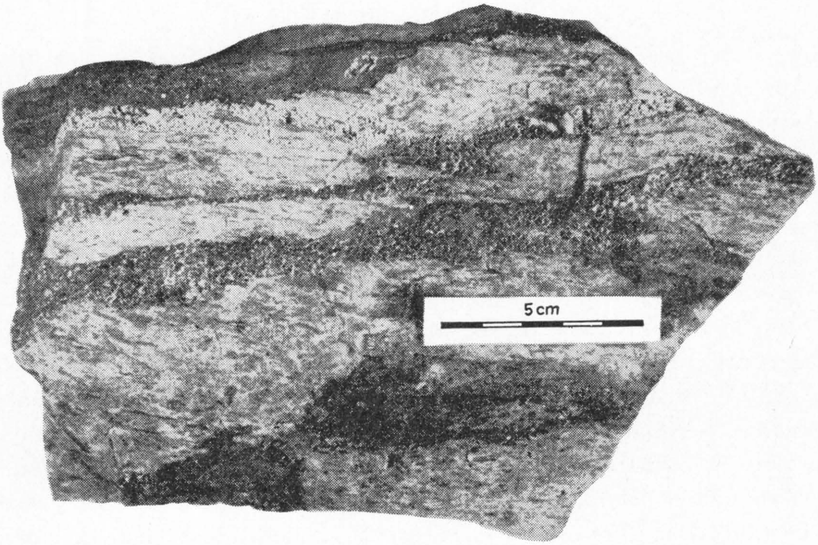

Fig. 16. Layered magnetite-hematite ore. Light-coloured: Quartz-hematite, dark-coloured: Magnetite. Loc. 226.

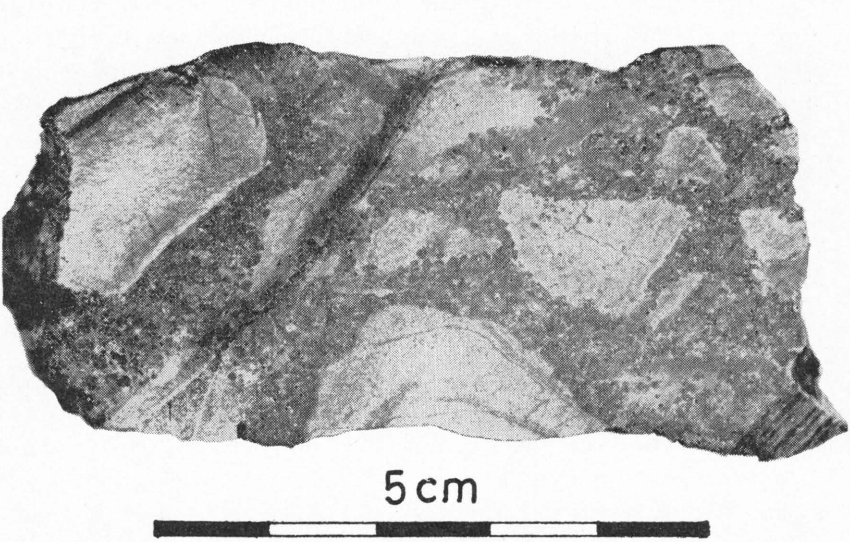

Fig. 17. Agglomerate ore: Lapilli of fine-grained quartz keratophyre. The groundmass consists of magnetite, quartz and albite.

The ore as a rule is layered, but more diffusely and irregularly so than the mixed ores (Fig. 16). The rock commonly consists of alternating layers and bands of *magnetite* and of *quartz—hematite*. The magnetite grains often are euhedral, with sizes up to 1 mm. The ores are possibly complementary supracrustal varieties of the iron oxide content of the quartzitic xenoliths in the greenstones, and are thought to have been formed from iron-rich precipitates of exhalative-sedimentary origin (The origin of the quartz in these rocks is uncertain).

At the Grönli mine a rock type occurs which in the writer's opinion is best interpreted as an *agglomerate ore*. It has light grey-coloured fragments (lapilli) in a dark, magnetite-rich matrix (Fig. 17). The fragments consist of a light-coloured, very fine-grained *quartz keratophyre-like rock*, in places porphyritic, and often with a blastophitic texture. The mineral composition is: Albite, quartz, magnetite, sericite, epidote.—The matrix of the agglomerate is very rich in *magnetite*, as mm-big octahedra, or as small roundish grains. In addition, it contains fine-grained equigranular *quartz*, scattered, somewhat larger grains of *albite*, anhedral, oval-shaped grains of *apatite*, strongly green-coloured *chlorite*, and grains of *calcite*. The lapilli have been elongated. The matrix is not very different from the other iron ores in the area.

This type of ore occurs together with ordinary layered hematite-bearing magnetite ore, and its genesis probably is the same, apart from the simultaneous outbursts of explosive extrusion by the same magma that expelled the iron-rich gases. The interpretation is thus that the groundmass consists mostly of material supplied by exhalative-sedimentary processes, while the rock fragments are volcanic lapilli.

Sedimentation, volcanism, petrogenesis

The lithology of the supracrustal rocks within the map area reflects variations within wide limits in sedimentary environment, as brought out by the stratigraphic sequence: The lower phyllite indicates a quiet period of deposition of clayey sediments. The transition through time from phyllite via greywacke phyllite to the metagreywackes and finally the quartz schist points to a rising of land above sea level in the source area of the sediments, brought about by an early orogenic phase, possibly the so-called Trysil disturbance (Skjeseth 1961), which in the miogeosynclinal regions far to the south and southeast is represented by a depositional break below Etage 3 b. The gap in sedimentation in the miogeosyncline may thus in the eugeosynclinal deposits in the Selbustrand area be represented by a concomitant change in lithologic facies. The quartz schist might conceivably be correlated with the Blue Quartz of the Hardangervidda deposits. The deposition of the quartz schist marks the end of the influence caused by the Trysil disturbance, and a quiet deposition of clays started anew (The upper phyllite). This time, however, the sedimentation did not remain undisturbed for long: An episode of explosive volcanism initiated the extrusion of huge masses of basaltic lavas, viz. the Stören Greenstones.

The occurrence of fossils of 3 b γ age above the greenstone series near Lökken (Blake 1962), and the tentative correlation of the deposition of the metagreywackes and the quartz schist with the Trysil disturbance, of 3 a (?) age, requires that the Stören lavas were extruded through a very short span of time. The inevitable question arising out of this situation is: Could the lavas have poured forth at such a rate as to attain a thickness of up to at least 2500 m (T. Vogt 1945) in the time available according to the correlations outlined above? One feature of the Stören Greenstones points towards an affirmative answer: The scarcity of intercalated sediments in the Stören Group. All through the section the character of the greenstones indicates a submarine extrusion of the lavas, and one might *a priori* have expected the deposition of copious amounts of sediments if the volcanic events took place over an extensive period of time. In that case, it should be possible to single out individual lava flows by the intervening sedimentary beds. As it is, the lavas follow suit one upon the other to form a nearly unbroken and homogeneous pile of extrusive rocks. The Trondheim disturbance (T. Vogt 1945), with its ensuing uplift and erosion of the Stören Greenstones, and the following transgression with the deposition of the Stokvola conglomerate must also have taken place during the interval between the Trysil disturbance and 3 b γ time.

It is generally agreed that the volcanism of lower Ordovician age within the Norwegian mountain chain has been of submarine character. It is further held that the lavas were mainly extruded by fissure eruptions (Vogt 1946). The volcanic processes within the Trondheim Region have been discussed in the literature particularly by C. W. Carstens, T. Vogt, and in later years by H. Carstens, and Chr. Oftedahl.

Within the present area the volcanic processes are characterized by two phases of explosive activity, the earliest of which apparently was the start of the period of volcanism, the second occurring at a late stage of the lower Ordovician basaltic volcanism.—The products of the first explosive phase are not everywhere easily traced in the field, and probably the explosive activity here have been of minor importance as compared to that which gave rise to the formation of the upper pyroclastic horizon. In the interval between the two explosive phases the volcanism has manifested itself by relatively calm effusions of lava, with the formation of pillow lavas, massive lavas, and porphyritic lavas. Even within the lavas one may locally find structures indicating a pyroclastic rather than an effusive origin. To what degree the upper greenstone breccia has been affected by volcanic gases, or to what extent pyroclastic material is present among its fragments, is not certain, but, as previously mentioned, there are indications that such material indeed is a constituent of this rock type.

The sporadic explosive character of the volcanism shows that the magma at times must have been rich in gas and vapour, and it seems reasonable that the explosions have taken place by point eruptions rather than fissure eruption, as the existence of fissures probably would inhibit the accumulation of excessive pressures in the magma.—Present-day volcanic gases are dominantly composed of water vapour, and to all probability this have also been the case with the volcanic gases of the Mostadmarka magmas. These rocks are considered as spilites, and it is commonly held that the spilites acquire their petrological singularities through a high water content of the magma. Other common constituents of volcanic gases are CO_2, SO_2, H_2S, Cl_2, HCl, HF, CO.—By high temperatures water vapour will to some extent be dissociated to H_2 and O_2. This process may render the gas oxidizing. The element of ubiquitous presence most susceptible to red.-ox. processes is iron. The degree of oxidation of iron in igneous rocks have been investigated by, among others, Goldschmidt (1943), G. C. Kennedy (1955), T. C. Phemister (1934), T. Krokström (1937).—Kennedy finds a direct connection between the degree of oxidation of iron in an igneous rock and the water content of the magma: A high content of water vapour yields a high Fe_2O_3/FeO ratio. Concerning the spilites, and the role of water content in their genesis, M. H. Battey (1956) advocates the existence of a more or less continuous rock series from tholeiite to spilite, and sets up the following features as essential to bring about the change from a "dry" tholeiite to a spilite: 1) Increase of Na_2O and decrease in CaO and Al_2O_3, 2) increase in iron, 3) increase in TiO_2, 4) increase in

P_2O_5, 5) decrease in MgO, 6) higher state of oxidation of iron, 7) increase in combined water.

Rock analyses indicate that particularly high ratios Fe^{+3}/Fe^{+2} is prevailing in pyroclastic rocks. For pyroclastic rocks of subaerial deposition this relation is easily explained as due to atmospheric oxidation at elevated temperatures, but for submarine tuffs it seems reasonable that the state of oxidation of iron might be inherited from the magma. This may be corroborated by the fact that basaltic glass, which have been only superficially acted upon by the atmosphere, shows the same feature: A high Fe^{+3}/Fe^{+2} ratio.

In recent years it has become apparent that the degree of oxidation of iron in a rock undergoes practically no change during ordinary regional metamorphism, as brought out by Yoder (1957), James and Howland (1955), and Chinner (1960).

In one of his papers on the Trondheim Region C. W. Carstens draws the conclusion that the process of devitrification of basaltic glass brings about an oxidation of the iron present. After what has been said above on basaltic glass and pyroclastic material, and about red.-ox.-processes in a solid rock during regional metamorphism it seems reasonable to suggest that the degree of oxidation of iron in such rocks is largely determined by processes taking place during the magmatic or late magmatic stages, rather than being a by-product of devitrification.

In a glassy rock iron will be present either in solution in the glass, or as finely dispersed iron oxide minerals. By the process of devitrification it will enter into mineral combinations. In common rocks the essential amount of iron present will be found either as iron oxide minerals (hematite, magnetite), in ilmenite, or as a component of silicate minerals. Provided a substantial part of the iron is in the trivalent state, what minerals may it enter? In the oxide minerals ferric iron enters without any difficulty. Most of the common rock-forming silicates, however, may accommodate a slight amount of ferric iron only, and ferrous iron will be in large excess over ferric. One exception to this is epidote. In this mineral trivalent iron is the representative of the ferroan component. In the writer's opinion epidote may under certain conditions occur as a primary mineral in a rock. This question has for many years been repeatedly discussed, and differences of opinion still seem to exist. It is here suggested that epidote is a primary constituent of some of the greenstones in the sense that it is the first ferroan silicate mineral to crystallize by the devitrification of basaltic glass with a Fe^{+3}/Fe^{+2}

ratio too high for the formation of any of the other common ferroan silicate minerals (amphibole, chlorite, biotite etc.). This epidote formation presumably would take place preferentially in rocks where the iron was in solution in the glass.—Goldschmidt (1916) makes the observation that the greenstones of the Norwegian mountain chain are strongly epidotized and carbonatized even in areas of negligible regional metamorphism.

Whether epidote in any of the rocks described herein are the product of direct magmatic crystallization is uncertain, but the conclusion is offered that in some of the rocks epidote is a primary constituent in the sense given above, and that its presence is not due to the regional metamorphic processes which the rocks have undergone.—A consequence of this view is that epidote cannot in all cases be a critical mineral for the metamorphic facies classification. In this connection it may be mentioned that epidote is practically absent from the sedimentary rocks of the present area, but that it is ubiquitously present in the greenstones.

One of the rocks described has, in addition to its profusion in fine-grained epidote, another feature taken to indicate rapid cooling: One of the varieties of porphyritic greenstone (No. 63, Pl. II, fig. 2) is characterized by the following features: It contains large albite phenocrysts set in a groundmass essentially consisting of fine-grained epidote. The phenocrysts are excessively sericitized, while the groundmass is devoid of K-bearing minerals altogether. The content of sericite is thought to have originated by the decomposition of a primary significant Or—content in the phenocrysts, and supposedly the confinement of sericite to the phenocrysts is caused by the rock cooling so rapidly that the plagioclase phenocrysts were unable to expel the components of potassium felspar by a normal process of exsolution with the formation of two felspars, or by a redistribution of the K to comprise the groundmass also.—A plagioclase with a high K-content is stable only at elevated temperatures, and the observed relations are thus best explained by supposing a rapid cooling of the rock, which in its turn indicates that the epidote of the groundmass may have crystallized directly from a glass.

The synthetization of epidote has as yet not met with undisputed success, and for this reason the stability relations are rather imperfectly known. The role of iron content in determining the stability field of epidote is unknown, and in what direction a varying content of iron displaces the phase boundaries, is, according to the literature, subject to differences of opinion. Rosenqvist (1952) writes:

> This means that iron-bearing epidotes are stable together with a plagioclase of a given composition at a higher temperature than was assumed for clino-zoisite. It is possible, for example, that an epidote rich in iron may crystallize as a primary mineral from a magma, provided this magma has a very high iron III and hydroxyl activity.

Miyashiro and Seki (1958), maintains that an epidote with about 30 mol. per cent Fe-end-member is the stable form at low temperatures, while higher and lower Fe-contents both become stable at higher temperatures. It is thus not possible, on the basis of the composition of the epidote, to draw safe conclusions regarding its occurrence as a primary constituent in the greenstones of the present area.

Stilpnomelane is an iron-bearing silicate mineral which show a great flexibility as far as the state of oxidation of its iron content is concerned. Turner and Hutton (1935), in a discussion of the occurrence of stilpnomelane in metamorphic rocks from Western Otago, make the following observations:

> It is suggested, therefore, that stilpnomelane may arise directly from chlorite, either by the oxidation of the ferrous iron present in the chlorite itself, or by addition of iron from some external source. — — — — A further supply of ferric iron might well be derived from the ferruginous epidote, which show a marked tendency to pass over to clinozoisitic varieties as reconstitution of the schists becomes more complete as the biotite isograde is approached. In many of the sections examined stilpnomelane and chlorite appear to have crystallized independently of one another.

In this connection the remark of the authors about the relation between stilpnomelane and epidote is interesting. This appears to corroborate the view put forward by Miyashiro and Seki.

Leucoxene is always present in the greenstones, and the saussurite gabbros have leucoxene pseudomorphs after skeletal ore grains, probably ilmenite. The formation of leucoxene at the expense of Ti-bearing ore minerals may also be connected with the oxidation of iron (Flinter 1959). The processes of decomposition of the primary ore mineral of the saussurite gabbros are thought to have taken place in the magmatic or late magmatic stage, and to be a consequence of the high water content of the magma.

The metamorphic grade of the rocks places them all in the greenschist facies, represented by the sub-facies quartz-albite-muscovite-chlorite and quartz-albite-epidote-biotite sub-facies. The criteria for the subfacies classification are not clearcut in these rocks, but the metamorphic grade evidently varies within rather narrow limits. Unstable relicts show that not all of the rocks have reached mineralogical equilibrium during the metamorphism.

Ore genesis

For the Leksdal type sulphide ores C. W. Carstens maintained a biochemical sedimentary origin in connection with volcanic activity. The ores were supposed to be formed by precipitation on the floors of isolated sea basins, with a highly reducing environment. In this connection parallels were drawn with the Black Sea, and the conclusion offered that in Lower Ordovician times isolated sea basins were present in the central part of the Trondheim region. It was suggested that the sea water carried iron in solution, which then reacted with hydrogen sulphide to form iron sulphides. In recent times, H. Carstens (1955) has proposed a similar genesis for the iron oxide ores of the northwestern part of the Trondheim region, and offers the opinion that the latter are contemporaneous, or nearly so, with the sulphide ores. The difference in mineralogy, according to H. Carstens, was caused by a difference in chemical environment during sedimentation, induced by the different submarine topography in two separated geosynclinal basins, one of which was isolated from the open sea, as postulated by C. W. Carstens. H. Carstens concludes by assigning the iron oxide ores to an oxide sedimentary facies, and the sulphide ores to a sulphide sedimentary facies.

Several problems of general interest to ore geology are implicit in the short excerpts of the current theory on the genesis of the ore deposits of the Trondheim region given above. Those emerging by the discussion of the ore deposits of the Mostadmarka—Selbustrand area falls within the following categories: 1. Magmatic concentration of iron; 2. Volatile transport of iron; 3. Supracrustal concentration and deposition of iron as sulphides and oxides; 4. Volcanic gases, their composition, and their physico-chemical properties.

The behaviour of iron during differentiation of a magma has been a subject of disagreement between two different schools of thought.

One, whose main proponent was N. L. Bowen, held that iron in a magma will join the early crystallizing minerals, rendering the later differentiates successively poorer in this element. In opposition to this, Fenner and others maintained that iron tends to become concentrated in the melt as differentiation proceeds.—The result of recent research indicate that the antagonistic views are both right (Kennedy 1955, Osborn 1959), in that both lines of magmatic evolution are possible, and that the products of both are found in nature. The physico-chemical parameter that determines in what direction differentiation will proceed in a given case, is the partial pressure of oxygen, or more specific: The rate and direction of change of the oxygen pressure during the crystallization of a magma. The factor thought to determine the oxygen pressure is, beside the PT-conditions, the water content of the magma. According to the above-mentioned hypothesis (Osborn 1959), oxidation of iron in a magma will cause a fixation of iron in the early-forming crystals, in accordance with the view advocated by Bowen. This trend of differentiation is typical of orogenic regions, the magmas of which from other considerations are thought to be relatively rich in water. A Bowen-trend of differentiation apparently has been operating in the evolution of the igneous rocks of the Caledonian mountain chain in Norway, giving rise to the sequence serpentine, gabbro, opdalite, trondhjemite, and granites, and the volcanic derivatives of some of them.

The supracrustal accumulation of iron of magmatic origin leads to problem 2: Volatile transport of iron. The first relation to be considered in this connection is: What compound of iron may enter a gas or vapour phase? It has been postulated that volatile transport of iron is possible only if the iron is present as chloride, preferentially as ferric chloride. Krauskopf (1959) presents a table of the vapour pressures at 600°C of heavy metal compounds that may be involved in such a transport, namely, elemental metals, sulphides, oxides, and chlorides. He concludes:

"For all the metals in table 3 ¹) except arsenic, copper, and gold the chlorides are the most volatile compounds under plutonic conditions." "— — — — so that volatile chlorides must be considered as a possible means of transportation."—Oftedahl (1958), in his hypothesis of exhalative-sedimentary ore genesis, supposes that iron is transported in the form of chloride. Landergren (1958), on chemical grounds, could not

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¹) Including iron.

accept this kind of transport in the presence of water, because of the "scholar example" of an ampholytic reaction: $2FeCl_3+3H_2O = Fe_2O_3 + 6HCl$. His objection, however, is contradicted by experiments made by Stirnemann (1926), in which mixtures of water and $FeCl_3$, contained in closed tubes, were heated to 500—560°C under their own vapour pressures (totalling approximately 8.5 atm). In these experiments iron was transported in the vapour phase and crystallized as hematite in the opposite end of the tube, away from the original position of the $FeCl_3$-solution. Thus iron may be a component of the vapour phase even in the presence of water if the system is under pressures only moderately higher than atmospheric unity, and by temperatures far below that which may be expected to pertain during the process of volatile transport connected with volcanism.

The Lower Ordivician basaltic volcanism of the Norwegian mountain chain is supposed to have been mostly submarine, and hydrostatic pressure in the oceans increases by about 1 atm for every 10 m increase of depth. Even a moderate subaquatic depth will thus yield pressure conditions favourable to the gaseous transfer of iron.

Another question to be answered is: Is there any reason to suppose that Cl was present in the magma causing the volcanism of the present area? Chlorine and some of its compounds are common components of present-day volcanic gases, and it seems reasonable to assume that Cl was present also in the gases of the Lower Ordovician volcanism of the Trondheim region. Correns (1957) maintains that the content of chlorine in the oceans can be explained only by the assumption of a degassing of the earth brought about by volcanism.

If iron in the trivalent state enters the volatile phase this means a small-scale mixing of the Bowen and Fenner trends of differentiation: The oxidized iron instead of entering early-formed crystals remains in solution in the fluid phase in the form of chloride (small-scale because this process in reality is not a differentiation into rock melts of differing compositions, but one taking place between the liquid and the vapour phases of one spilitic magma).

By the submarine eruption the volatiles leave the magma, and the iron chloride enters the marine environment, spontaneously passing from the pneumatolytic transport phase via a hydrothermal to a sedimentary phase. The ore genesis thus is both magmatic and sedimentary: 1) Magmatic: Transport of iron; 2) Sedimentary: Precipitation of iron. The precipitation probably is a process of simple hydrolysis: $Fe_2Cl_6 +$

$6H_2O = 2Fe(OH)_3 + 6HCl$. The ferric hydroxide fell out as a gel in the vicinity of the point of effusion. This implies that the primary precipitation of the iron was in the form of hydroxide, and that the formation of sulphide was a later event. This is not in agreement with the views of C. W. Carstens. In his opinion, the iron remained in solution until it was precipitated as a sulphide gel by H_2S, mainly of biogenic origin. This process was thought to take place under euxinic conditions. For physico-chemical reasons, however, the processes outlined above seem to be in better agreement with the features of the ore-bodies in the present area. This is the case particularly with the ore deposits consisting of thin interbedded layers of pyrite and magnetite. For such a rock to be deposited in the way Carstens advocated one has to suppose rhythmical changes in sedimentary and chemical environment. Such changes could be brought about by changes in the physiography of the sea bottom, but the diastrophic processes responsible for the changes would have to be extremely regular, and with very rapid oscillations. Volcanic eruptions, especially gas-rich phases, would tend to cause circulation in the sea-water, unless the hydrostatic pressure exceeded the critical pressure of water. Euxinic conditions might prevail in an area of volcanic activity provided the depth of the sea was in excess of about 2000 m.

If the volcanic activity itself is held responsible for the physico-chemical conditions in the sea during the period of ore formation, there are no particular requirements to be met regarding sea depth and bottom topography, and it is not necessary to assume the existence of closed basins in the geosyncline.

The composition of gases discharged by volcanic exhalations appears to be temperature dependent. Walker and Walker (1956) classifies fumaroles as follows:

"The emission of the different gases from a magma does not take place simultaneously and uniformly, but in a certain sequence, which is governed by temperature, as per the following classification of fumaroles:

1) Dry fumaroles. From these fumaroles, at temperatures ranging from 750°C down to 250°C issue chiefly superheated steam or water gas, hydrogen, carbon monoxide, methane, and metal chlorides. Other products of minor importance, occurring in variable amount, are chlorine, fluorine, nitrogen and sulphur vapor.

2) Acid fumaroles. In fumaroles of next lower temperature, ranging from 250°C down to 150°C, atmospheric oxygen unites with hydrogen to form water vapor, with methane to form CO_2 and water vapor, with CO to form CO_2, and with sulfur to form SO_2. The water vapor thus produced reacts to some extent with metal chlorides, generating hydrochloric acid, whence the name "acid fumarole."

3) Intermediate fumaroles. At temperatures ranging from 150°C down to 100°C, these give off chiefly water vapor, H_2S, CO_2, and NH_4Cl.

4) Aqueous fumaroles. At temperatures of about 100°C these give off chiefly steam and CO_2."

Similar relations are reported by Lovering (1957) from the fumaroles in the Valley of Ten Thousand Smokes.

To give a valid picture of the processes leading to supracrustal concentration and precipitation of iron in the form of sulphides and oxides (hydroxides) it is important that the physico-chemical conditions governing the several possible processes be looked into. C. W. Carstens expressed the view that most of the iron expelled by the volcanic gases remained in solution in the sea water for some time. This, however, seems unlikely in the light of physico-chemical evidence established after Carstens formed his theory of ore deposition. Baas Becking, Kaplan and Moore (1960), in a work on pH- and Eh-relations in natural environments give values for these parameters in different environments: Marginal marine sediments yield pH-values between 7 and 8, mostly just below 8. Sea water pH is mostly 8 or just above, with no values below 6. Open-sea sediments yield pH-values as for sea water.—No measurements in these environments have given pH-values lower than 5.—The solubility of iron is strongly pH-dependent, and Cooper (1937) gives the following quantitative relations on the solubility of iron in water under differing pH-conditions: "The total quantity of iron in true solution in water after equilibrium has been attained does not exceed

3 $10^{\div 8}$ mg iron per cubic metre at pH 8.5
4 $10^{\div 7}$ „ „ „ „ „ „ pH 8.0
4 $10^{\div 5}$ „ „ „ „ „ „ pH 7.0
5 $10^{\div 3}$ „ „ „ „ „ „ pH 6.0

. . . . These values apply to iron in sea water, in natural waters and in many physiological fluids."—The solubility of ferric iron is insignificant for pH-values above 4. This means that a marine environment is several pH-units too high for ferric iron to remain in solution.—Galliher (1933) states: "The anaerobic phase of the sulfur cycle is carried on in an alkaline milieu. It is a generally recognized fact that anaerobic decomposition tends to raise the pH. Baas Becking, in a written communication to the writer, noted that 'the organisms (of the sulfuretum) occurred exclusively in alkaline waters, pH 7.6—8.6, either fresh water, sea water, or brine.,"—Thus, it seems unreasonable that anaerobic, H_2S-generating bacteria and iron in solution can coexist in the same environment.

C. W. Carstens stated that both H_2S and iron compounds are common volcanic exhalation products. His views necessitate that these components occur simultaneously in the gas exhalations. This seems not to be the case.—No doubt, bacteria do play an important role in the precipitation of chemical sediments, but in the depositional environment of the present ores, contribution of organisms is not required to explain the processes.

The ore genesis in the Mostadmarka area is here tentatively explained in the following way: Iron was transported to the surface as gaseous $FeCl_3$ (or as Fe_2Cl_6, according to temperature). In the sea water, hydrolysis immediately took place, and ferric hydroxide was precipitated as a gel. Circulation in the water brought about by the gas eruptions caused the HCl formed by the hydrolysis to be quickly dissipated so that the local pH was not significantly lowered for any length of time. As the temperature of the gases was gradually lowered with time, their composition changed. The volatile transport of iron compounds would cease to be effective, and at a later stage the fumaroles would bring large amounts of hydrogen sulphide into the water. This H_2S then reacted with the surface layer of the ferri-hydroxide precipitate, reduced it, and converted it to iron sulphides. Below a screening surface layer of sulphide there might remain smaller or larger amounts of unaffected ferric hydroxide, which in turn gave rise to iron oxides like magnetite or hematite. Periodical volcanic activity might thus lead to the formation of alternating layers of sulphide and oxide minerals. Sulphidation of ferric hydroxide by H_2S is described by Allen, Crenshaw, Johnston and Larsen (1912).

The Leksdal type ores are often dark-coloured to nearly black. The dark colours have been ascribed to the content of carbonaceous matter, usually amounting to 1—2 per cent. Will a C-content of this order of magnitude give a dark-coloured pyrite rock? In similar ores from Finland of Precambrian age, the black-coloured constituents were found to be a water-bearing mixture of sulphur and various ferrous sulphides (Saksela 1960). The present writer has failed to find similar investigations on Norwegian sulphide ores of Caledonian origin mentioned in literature. However, it is not unreasonable that some black-colouring of the Norwegian ores may be caused by similar components as that of the Finnish ores.

The occurrence of magnetite in the iron oxide ores is ascribed to a subsequent partial reduction of ferric iron. Such processes are known to take place in sediments that have been covered, and the partial reduction is thought to have occurred soon after the original ferric hydroxide had become covered by later deposits.

There is a remarkable regularity in the geographical distribution of the ore types within the present area, and this reflects the geological conditions. Sulphide minerals are practically lacking in the pyroclastic rocks, and no sulphide deposits are known in the Venna valley, where the Upper Pyroclastic horizon dominates the bedrock. In the tuffs and the volcanic breccias grains of magnetite are the common accessory mineral even outside the ore deposits proper. The reason for this distribution may be that the volcanic phases giving rise to the pyroclastic deposits have yielded sufficient material in a sufficiently short time to cover the precipitates of ferric hydroxide before the latter had been affected by H_2S-exhalations, which played a role as volcanic gases only late in the history of the individual fumaroles. Explosive volcanism of great intensity may even have repeatedly destroyed the fumaroles before their "coming of age," so that the temperatures of the exhalations rarely were permitted to reach the stage of H_2S-generation. This implies that the different conduits of the gas exhalations were relatively short-lived, and that the individual ore deposits have been formed during geologically speaking very short episodes in the sequence of events of the Lower Ordovician volcanism.

In the views expressed above, emphasis has been placed on an inorganic-chemical interpretation of the ore genesis, while the contribution of organisms apparently has been rejected. This may be the place to state that one cannot exclude the possibility of bacterial influence on the processes involved in the precipitation of the ore components. Needless to say, the presence in the sea-water of H_2S-generating bacteria may have yielded H_2S capable to react with the precipitated ferric hydroxide in exactly the same way as that eventually produced by the same fumaroles which exhaled the iron-bearing gases. The view that iron remained in solution in the sea-water is thus not a *conditio sine qua non* for the participation of organisms in ore formation. However, the obvious way of classifying chemical precipitate rocks and mineral deposits like the ones considered here, is by using physico-chemical criteria, and include in one group rocks of the same composition without regard to whether their genesis was being organically or inorganically controlled. This view has been emphasized by Krumbein and Garrels (1952).

Structural geology

Structural data are presented on the geological map, and on the diagrams A—E (Pl. IX). The diagrams are lower hemisphere equal-area projections.

The central parts of the Trondheim region have, to the author's knowledge, not been made subject to detailed systematic structural investigations, and the present area is far too small for general conclusions on a regional scale to be drawn.

The Mostadmarka—Selbustrand area is bordered to the north by the so-called Hell Syncline (H. Carstens 1960) (Fig. 18). In the present area the linear structures have a general direction of plunge between NNE and E. Provided the Hell area does comprise a syncline (i. e. the tectonic setting is not one dominated by overthrusts), it is to be expected that the linear structures NE-wards along the syncline will show decreasing angles of plunge, eventually to pass the horisontal and attain SW- to W-erly plunges, like the map (fig. 18) demonstrates the plunge of the synclinal trough to be in its eastern parts.—One question arises in this connection: Have the linear structures primarily been formed with their present angle of plunge (40—60°), or have later deformations tilted them away from their original positions? It seems reasonable to suppose that linear structures like fold axes, stretching etc. caused by tangential stresses during orogeny will have fairly flat-lying concentration maxima. The high-angle plunges of linear structures within the present area could conceivably be the result of deformation with directions deviating from the main direction of stress of the Caledonian Orogeny in the region (NW—SE).

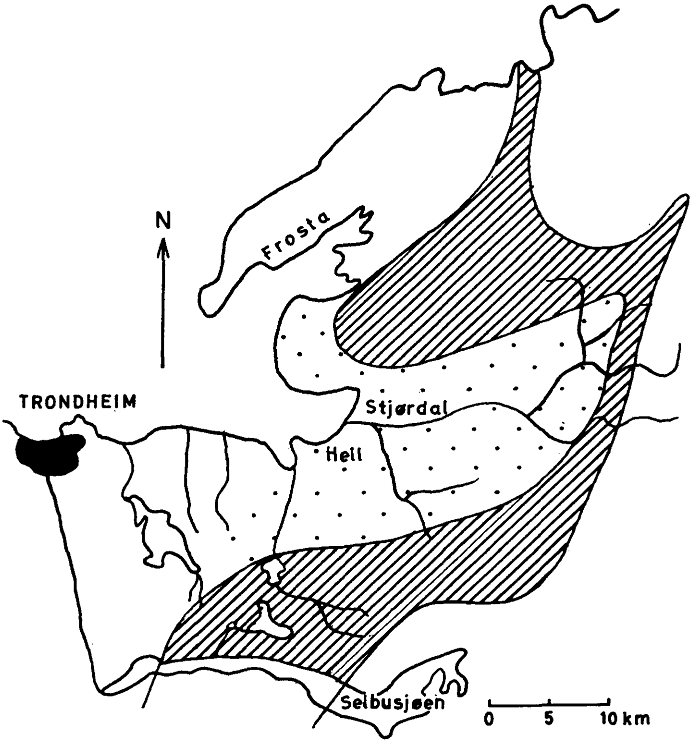

Fig. 18. The Hell Syncline (H. Carstens 1960). Geology after Th. Vogt (1946). Cross-ruled: Stören Group, stippled: Hovin Groups.

The structural diagrams show no clear-cut correlation between the different structure elements. The reason for this probably is that the different rock types are dominated by different structural elements, so that a rock displaying distinct lineation frequently will not show any measurable planar structure, and *vice versa*. There is fairly good agreement between the diagrams, *e.g.* the two lineation maxima and the fold axes concentration maximum all lie on a great circle in the stereographical projection, *i.e.* they are co-planar. This plane has a direction of strike of *c.* 170° and an easterly dip of *c.* 45° (Pl. IX, E). By tilting the plane 45° about its direction of strike, the maxima for linear structures come to lie in the horisontal plane. The direction of 170° is in close correspondence with the direction of strike of secondary small-scale folds and flexures, observable in the upper phyllite of the Lower

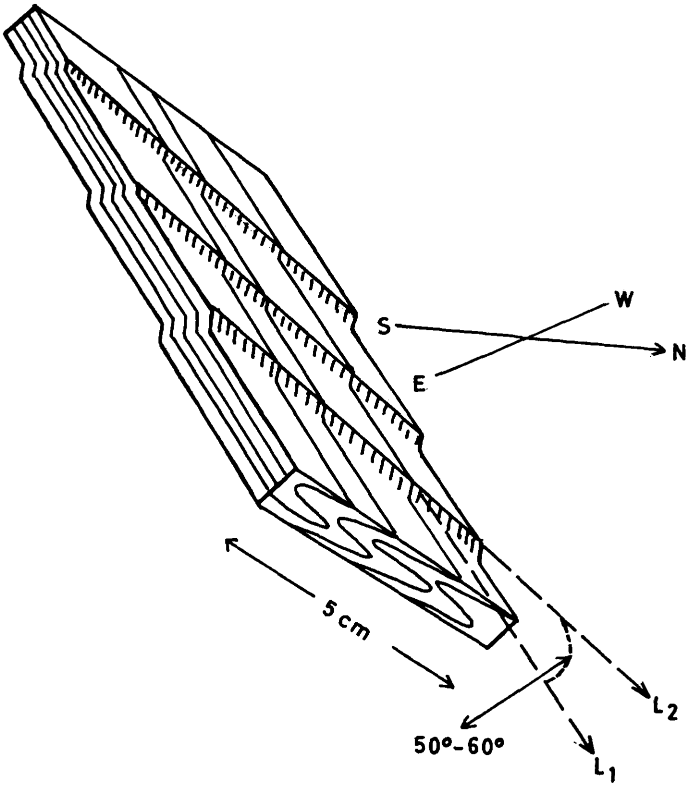

Fig. 19. Phyllite with two directions of folding (L_1 and L_2). Upper Phyllite, Lower Sedimentary Unit.

Sedimentary Unit, and in the conglomerate horizon of the Upper Sedimentary Unit. The types of folding have been mentioned in the petrographic descriptions of the rocks. The relations between primary and secondary folds are shown in the sketch (Fig. 19) of a hand specimen of the upper phyllite. Here, the first folds, whose fold axis is designated L_1, has been refolded by small-scale flexures of low amplitude with axis L_2. Similar relations are revealed by the stretched conglomerate from the Upper Sedimentary Unit (Fig. 20): The direction of stretching, L_1, has been flexured about the axis L_2.

The geological map reveals a variation of direction of strike from nearly N—S in the southern and southeastern parts of the area to nearly E—W along the Greenstone Unit- Upper Sedimentary Unit boundary.

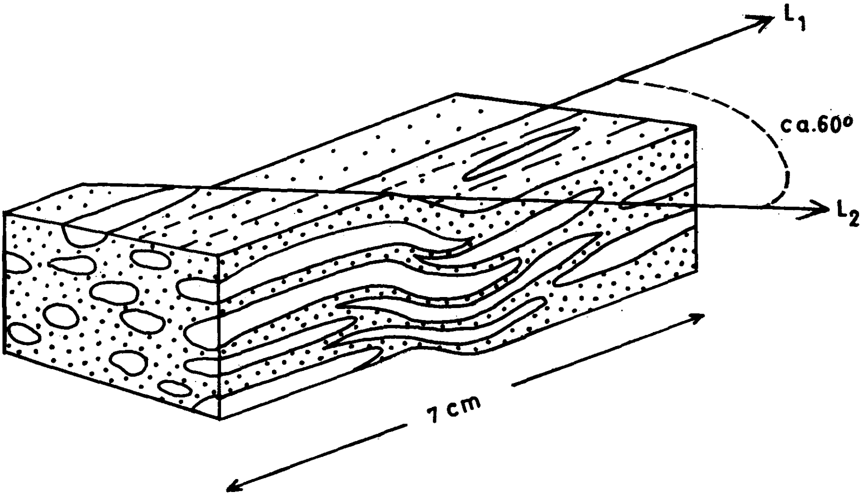

Fig. 20. Stretched conglomerate (L_1), flexured by folds in a second direction (L_2). Lower part of Upper Sedimentary Unit.

The bending thus exhibited is not easily correlated with any specific process of deformation, but it is thought to post-date the formation of schistosity and primary linear structures, and may be related to the formation of the secondary linear structures.

The quartz schist of the Lower Sedimentary Unit has, as previously mentioned, been subject to transverse faults, with NW—SE trends. This faulting may be related to a large-scale bending, as they show movements in the same sense of direction.

Sammendrag

Området Mostadmarka—Selbustrand hører i geologisk henseende til det sentrale Trondheimsfelt. Berggrunnen består vesentlig av svakt metamorfe suprakrustal-bergarter av ordovicisk alder, deriblant tallrike små svovelkis- og jernmalmforekomster. I tillegg til suprakrustalene opptrer mindre intrusjoner av gabbro, trondhjemitt og kvarts-keratofyr.

De suprakrustale bergarter hører til disse stratigrafiske hovedledd: Rørosgruppen (eldst), Størengruppen, og Undre Hovingruppe (yngst).

Rørosgruppens bergarter finnes i feltets sydlige og syd-østlige deler, og bare de øverste ledd i gruppen er representert. Bergartene er, regnet fra eldst til yngst: Fyllitt, gråvakkefyllitt (fyllittisk skifer med klastisk tekstur), gråvakke, kvartsskifer, og en øvre fyllitt. Den intense foldning av bergartene, med resulterende repetisjon av lagrekken gjør at mektighetsanslag er meget usikre, men det antas at de representerte ledd av Rørosgruppen tilsammen utgjør en mektighet på ca. 300 m, regnet fra gråvakkefyllitt og oppover.

Over Rørosgruppen følger Størengruppen. Denne bergartsserien domineres av grønnsteiner og -skifre, som er metamorfe basaltiske lavaer og pyroklastiske bergarter. Underordnet opptrer også epiklastiske bergarter: Kvartsitter, kvartsskifre, fyllitt og grønnsteinsbreksje. Stratigrafien innen Størengruppen er i området slik: Underst en tynn agglomerat-tuff-horisont. Over denne følger lavabergarter i en mektig serie sammensatt av varietetene: Massive finkornete grønnsteiner, putelavaer, og porfyrittiske grønnsteiner (med fenokrystaller av albitt). Disse leddene innen Størengruppen dominerer lithologien i feltets sentrale områder: Vottafjell, Strandbygdfjell, Langjon, Klimpan, og Vennafjell. Høyere oppe i lagrekken kommer en markert pyroklasthorisont, med grønne, meget grove tubbbreksjer, lapillituffer, og finkornete tuffer, dels i form av devitrifiserte glasstuffer. Dette leddet strekker seg i feltet fra Foldsjøen i vest, og østover langs Vennadalen. Fordelingen av de enkelte varieteter, grovt materiale i vest, finere materiale lenger østover, tyder på at materialtransporten har foregått fra vest mot øst. Nord for Vennadalen, over den pyroklastiske horisont, er igjen lavabergartene dominerende, men også her finnes hyppig pyroklastiske strukturer. Her finnes også et lag av kvartsskifer. Mot toppen av Størengruppen opptrer en grønnsteinsbreksje, som inneholder spredte bruddstykker av rød «jaspis», og for øvrig kantete bruddstykker av grønnsteinsmateriale. Breksjen oppfattes som et forvitringsresiduum av de underliggende vulkanittene. Størengruppens mektighet innen feltet anslås til maksimalt ca. 1000 m for den NE-lige dels vedkommende; det er vanskelig å avgjøre om et lignende tall kan anvendes for grønnsteinsavdelingen lenger vest.

Undre Hovingruppe har ikke vært gjenstand for detaljert kartlegging. Materialet som er innsamlet derfra er: Flere varieteter av et polymikt konglomerat som utgjør basallagene i denne gruppen. Noen av disse varietetene synes å inneholde ikke så lite pyroklastisk materiale i grunnmassen. Konglomeratene viser en tildels meget intens strekning av bollematerialet. Over, og nord for, konglomerathorisonten følger tuffer og tuffitter av andesittisk til dacittisk sammensetning. Høyere oppe domineres denne gruppen av en lys grønn- til gråfarget lagdelt og skifrig sandstein.

Intrusive bergarter opptrer i små felter innen Røros- og Størengruppen, men ble ikke funnet i Undre Hovingruppe. Som det eldste ledd blant intrusivene oppfattes saussurittgabbro, som innen feltet bare er påvist i Størengruppen. De sure intrusiver består for det meste av kvartskeratofyrbergarter, som opptrer både i Røros- og Størengruppen. Trondhjemitt er funnet et enkelt sted i feltet, som en intrusjon i Rørosgruppens kvartsskifer.

De tallrike malmforekomstene opptrer alle i Størengruppen, og de inndeles i svovelkisforekomster, blandete svovelkis-magnetittforekomster og hematitt-magnetittforekomster. Forekomstene er uten økonomisk interesse.

Malmforekomstenes genetiske forhold diskuteres i lys av for det meste nyere litteratur. C. W. Carstens hypoteser for Leksdalskisenes dannelse blir imøtegått på enkelte punkter, men hans hovedoppfatning, at malmene er syngenetiske vulkanske ekshalasjonsprodukter knyttet til den submarine basaltiske vulkanismen, styrkes ved en slik konfrontasjon med relevante arbeider publisert etter at teorien ble fremsatt (1919.)

De synspunkter om malmdannelsen som er fremsatt i dette arbeidet kan sammenfattes slik: — Jern er ved fumarolevirksomhet ført ut i havet som klorid i gassform. I havet inntrådte spontan hydrolyse, med utfelling av ferrihydroksyd som en gel. Ettersom ekshalasjonenes temperatur avtok, endret også gassenes kjemiske sammensetning seg, slik at de istedenfor jernklorider inneholdt bl. a. store mengder H_2S. Denne gass reagerte med det utfelte hydroksyd, og omdannet dette helt eller delvis til jernsulfid, som i sin tur ga opphav til dannelse av svovelkis. På denne måten oppsto svovelkisforekomstene av Leksdaltypen, og de lagdelte svovelkis-magnetitt-malmene.—De rene jernmalmene antas å være «Leksdalsforekomster» som er stanset i sin utvikling på halvveien, ved at de av en eller annen grunn ikke har vært utsatt for påvirkning av svovelvannstoff-gasser.

Feltets tektonikk er bare summarisk behandlet. Strukturmålingene gir et bilde som tolkes slik: Bergartene er blitt foldet om NE—SW-strykende akser, og denne deformasjon har ført til dannelse av linjestrukturer i samme retning. Senere har berggrunnen i området vært utsatt for lateralt stress, i retningen E—W. Denne fasen har ført til at bergartene er blitt foldet påny, denne gang om en akse ca. N—S. Som et resultat av dette opptrer kulminasjoner og depresjoner langs strøkretningen. Disse strukturtrekk er av slike dimensjoner at det kartlagte område bare utgjør den østlige skjenkel av en kulminasjon. — Mindre forkastninger opptrer i Rørosgruppens kvartsskifer, og en forkastning finnes i Størengruppen like NW for Strandbygdfjell. — Enkelte trekk ved grensen mellom Størengruppen og Undre Hovingruppe tolkes som indisier på at det har foregått bevegelser langs denne, uten at retning og omfang av bevegelsene er brakt på det rene. Der synes ikke å være noen betydelig tektonisk diskordans mellom de to avdelinger.

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Abbreviations:

DKNVS: Det Kongelige Norske Videnskapers Selskab.
GFF: Geologiska Föreningens i Stockholm Förhandlingar.
NGT: Norsk Geologisk Tidsskrift.
NGU: Norges Geologiske Undersøkelse.
SGU: Sveriges Geologiska Undersökning.

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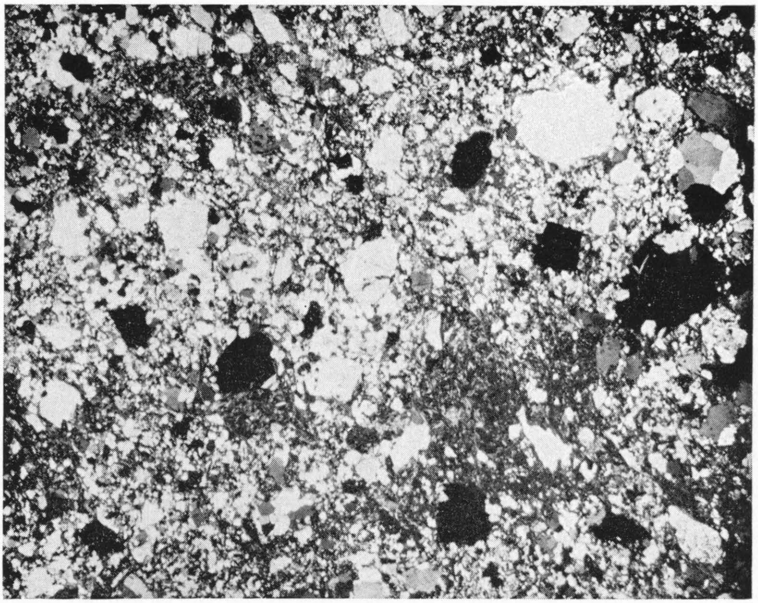


Plate I, fig. 1. Metagreywacke with small quartzite fragments. Loc. 30. Nicols crossed, 12x.

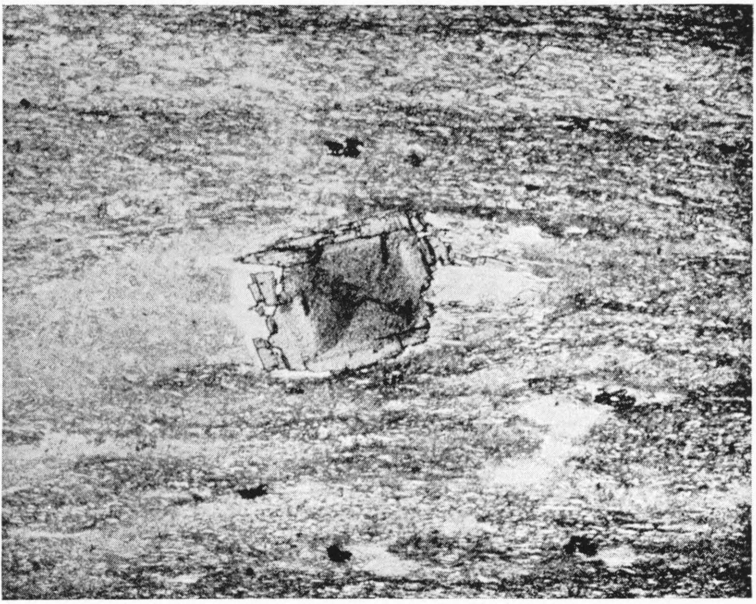


Plate I, fig. 2. Greywacke phyllite with siderite porphyroblast. Loc. 14. Ordinary light, 12x.

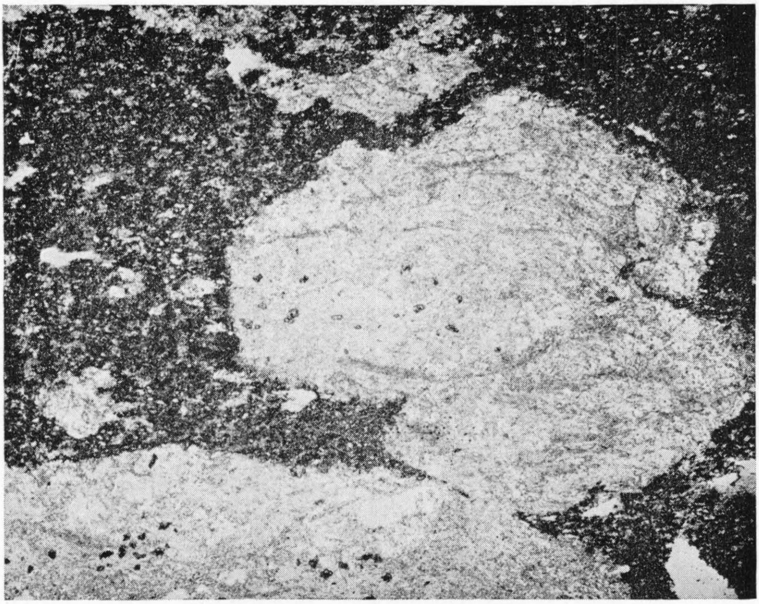


Plate II. Porphyritic greenstone, same as fig. 5. Strongly sericitized phenocrysts of albite; groundmass consisting mostly of epidote. Ordinary light, 12x.

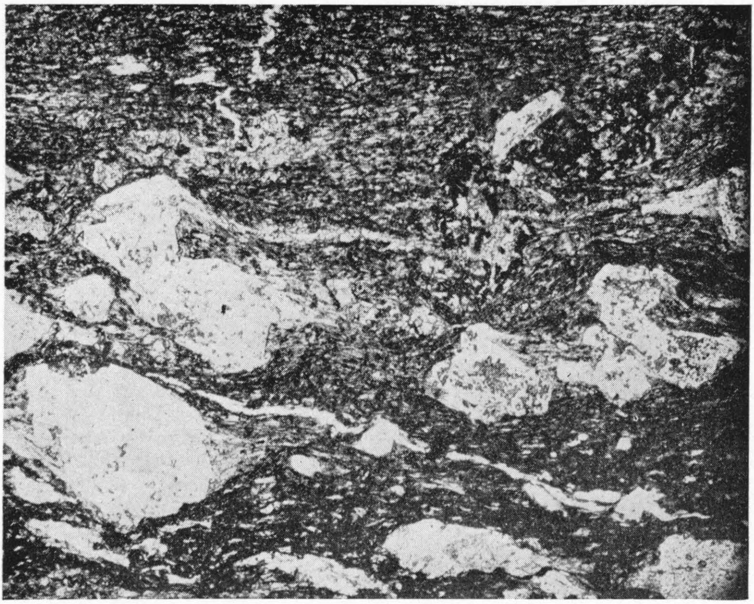


Plate III, fig. 1. Porphyritic greenstone with fluidal texture. Loc. 123. Ordinary light, 12x.

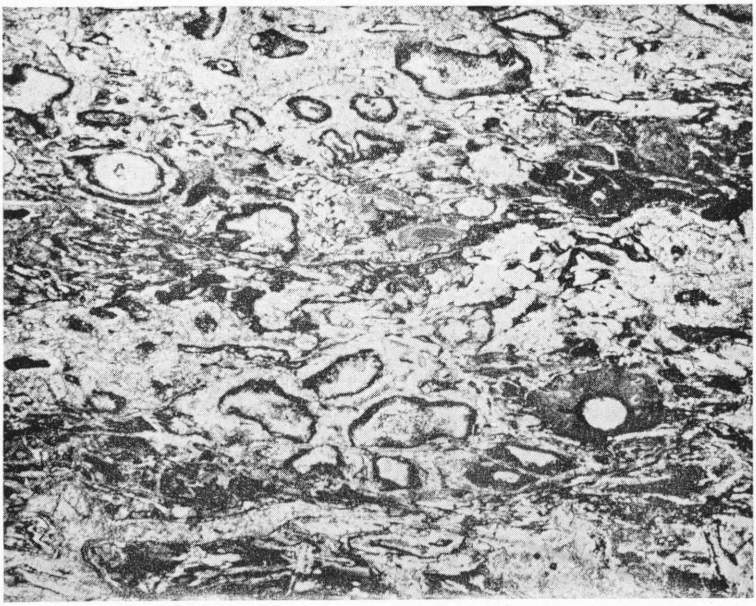


Plate III, fig. 2. Devitrified green tuff with albite-filled vesicles. The texture is delineated by leucoxene. Loc. 151. Ordinary light, 12x.

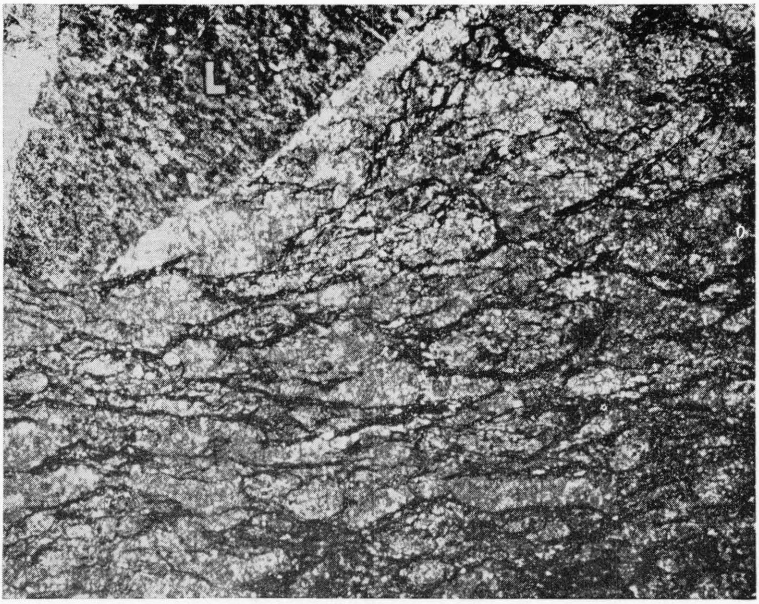


Plate IV, fig. 1. Green tuff with lapillo fragment (L). Texture delineated by leucoxene. Loc. 311. Ordinary light, 12x.

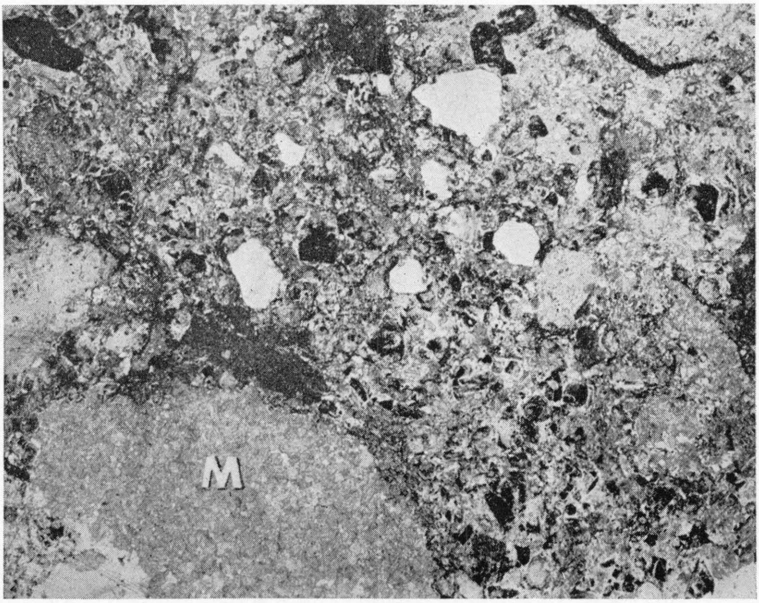


Plate IV, fig. 2. Greenstone conglomerate (Stokvola cgl.) with pebble of marble (M). In the groundmass grains of quartz (white) and strongly epidotized plagioclase grains (nearly black). Loc. 236. Ordinary light, 5x.

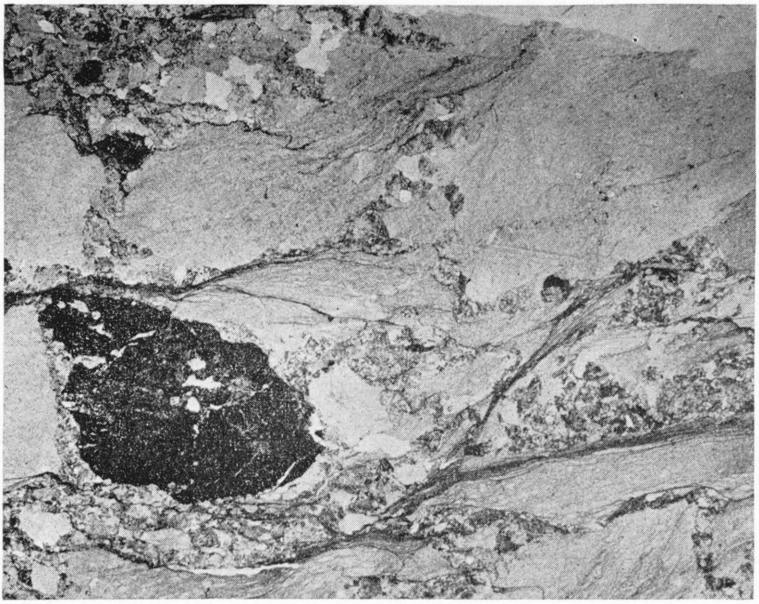


Plate V, fig. 1. Strongly deformed light grey-coloured conglomerate. One pebble of greenstone (dark), other pebbles of quartzite and quartz keratophyre. Loc. 229. Ordinary light, 5x.

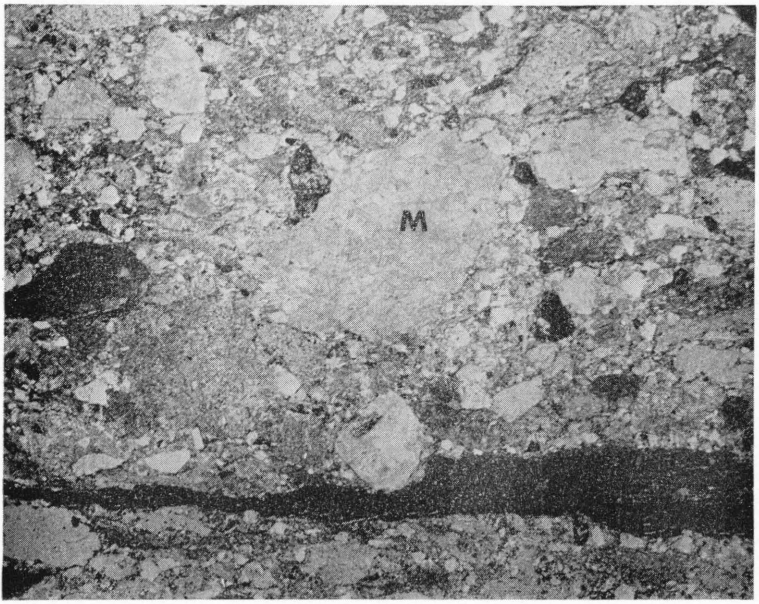


Plate V, fig. 2. Conglomerate layer in dacitic tuff. One pebble of marble (M). Dark-coloured zone consists of fine-grained epidote. Ordinary light, 4.5x.

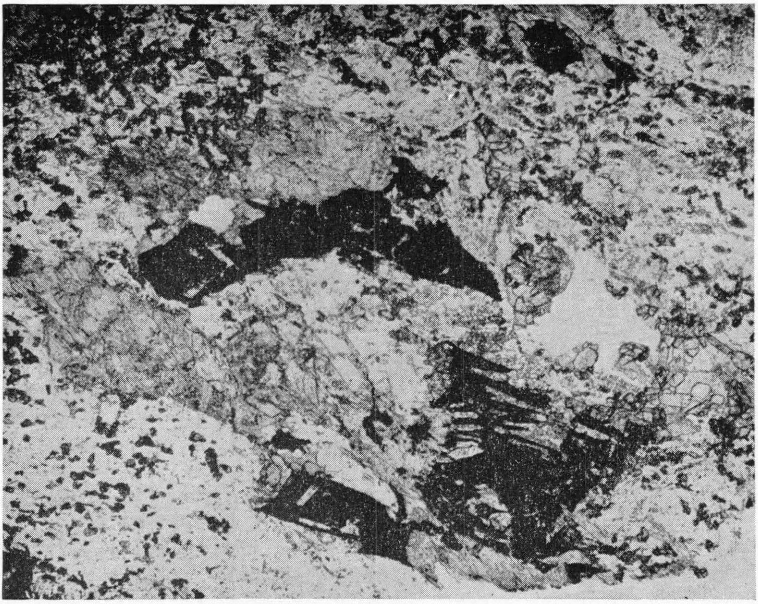


Plate VI, fig. 1. Saussurite gabbro, with skeletal pseudomorphs of leucoxene after ore grains, and with uralitic amphibole. Loc. 58. Ordinary light, 12x.

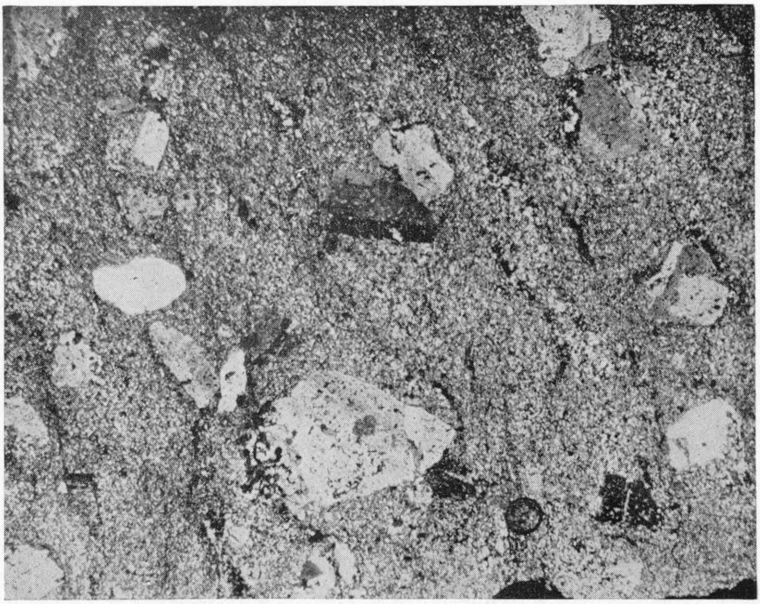
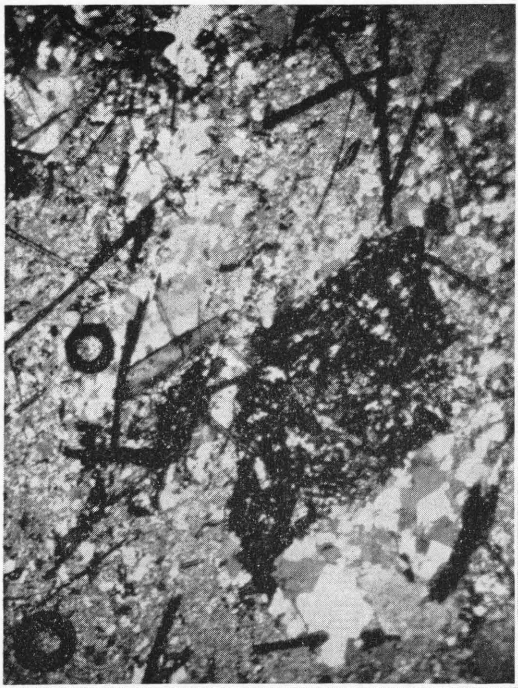
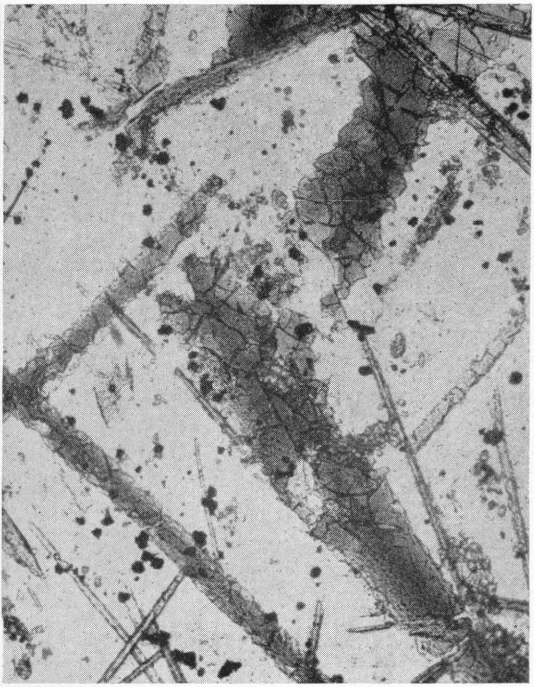


Plate VI, fig. 2. Porphyritic quartz keratophyre. Phenocrysts of albite and quartz, the latter embayed and rounded by absorption. Loc. 309. Nicols crossed, 12x.





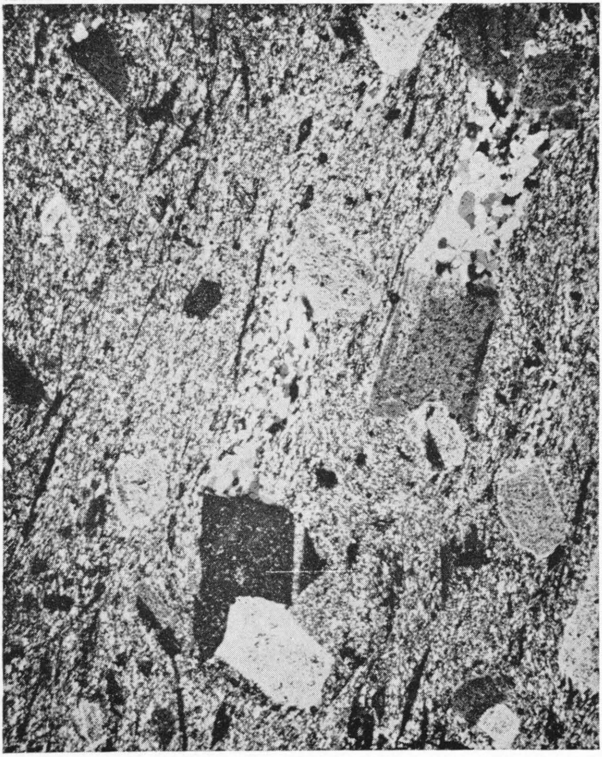


Plate VII, fig. 1a. Amphibole-stilpnomelane quartz keratophyre. Dark laths of stilpnomelane and amphibole. Rhomb-shaped pseudomorph-like aggregate of fine-grained magnetite and leucoxene. Nicols crossed, 33x.

Plate VII, fig. 1b. Same rock. Ordinary light, 110x.

Plate VII, fig. 2. Porphyritic amphibole-stilpnomelane quartz keratophyre. Phenocrysts of albite, in the groundmass laths of stilpnomelane and amphibole. Note coarse-grained quartz between some of the phenocrysts. Loc. 220.
Nicols crossed, 12x.

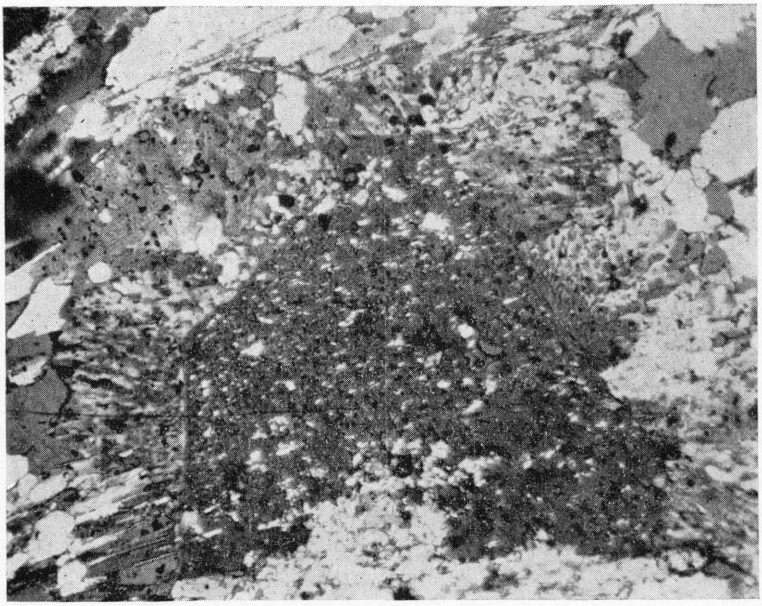


Plate VIII, fig. 1. Quartz-albite symplectite in garnet-bearing quartz keratophyre. Loc. 15. Nicols crossed, 70x.

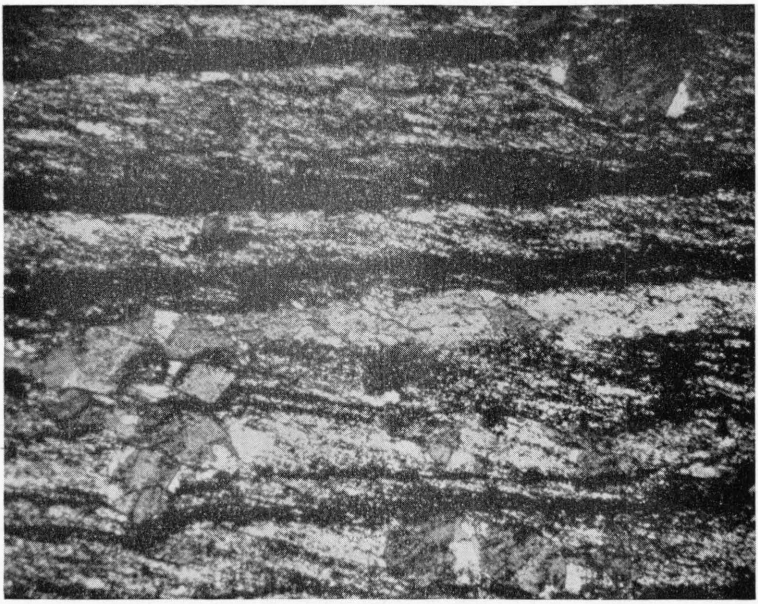
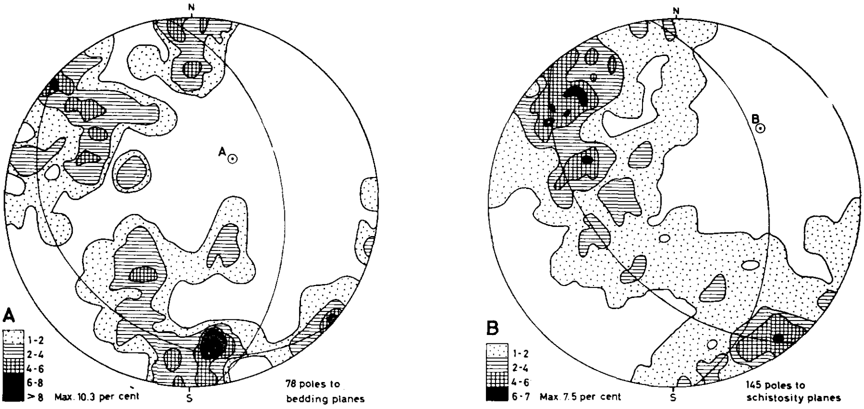
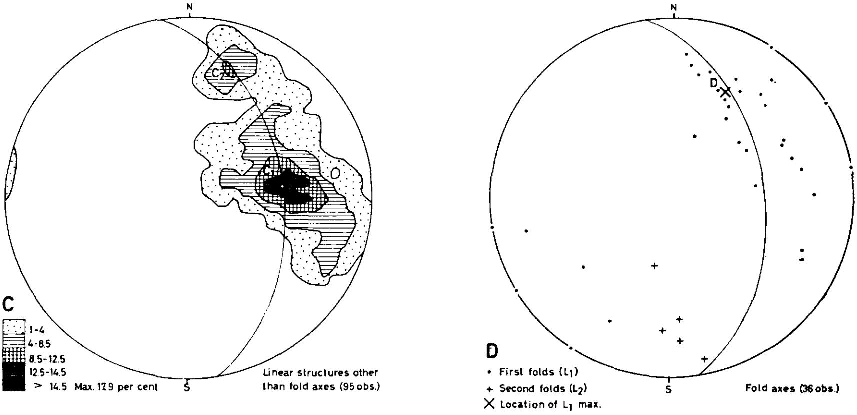


Plate VIII, fig. 2. Black fels with alternating quartz- and magnetite-rich laminae, and rotated chlorite porphyroblasts. Note secondary planar structure, oblique to lamination. Loc. 64. Ordinary light, 37x.



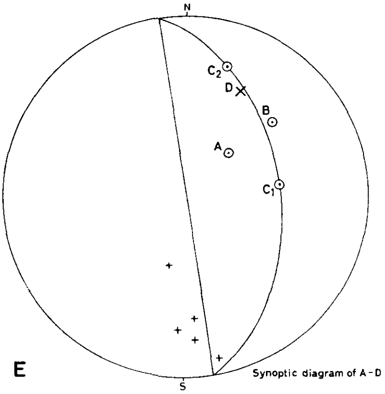
78 poles to bedding planes

145 poles to schistosity planes



Linear structures other than fold axes (95 obs.)

Fold axes (36 obs.)



Synoptic diagram of A - D

Plate IX. Structure diagrams. Equal area projection, lower hemisphere.

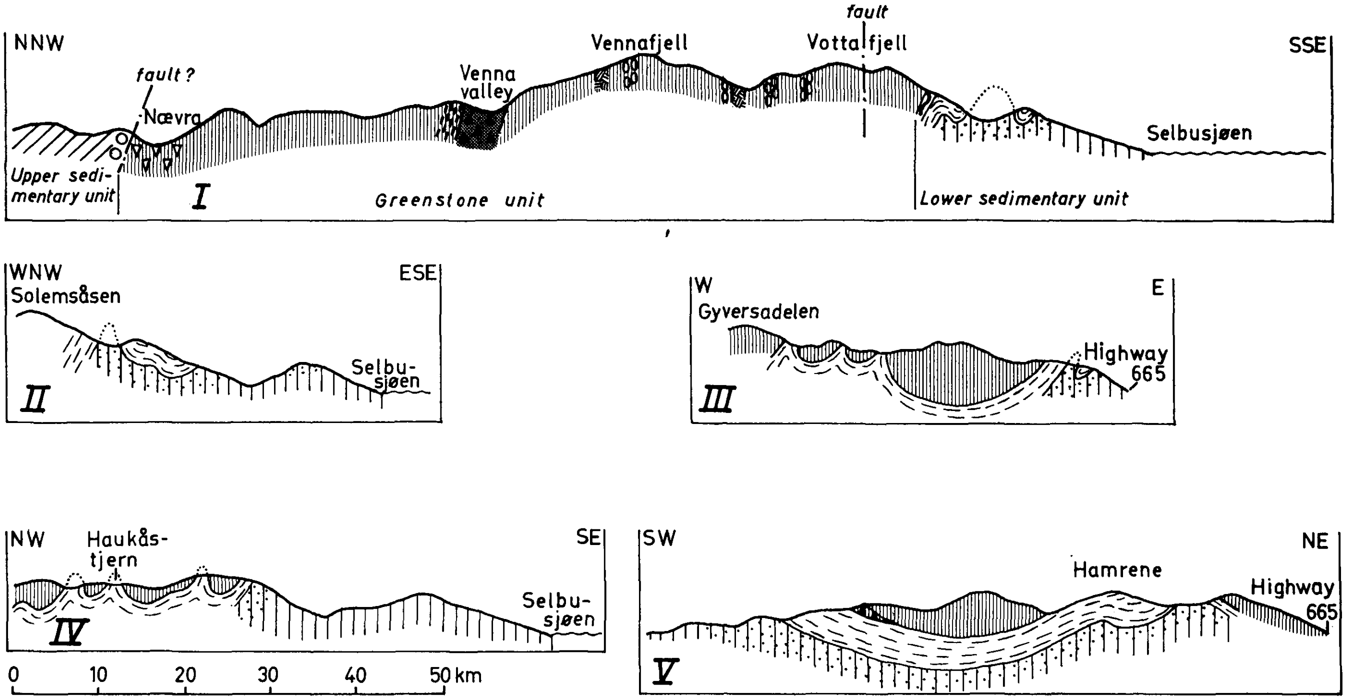
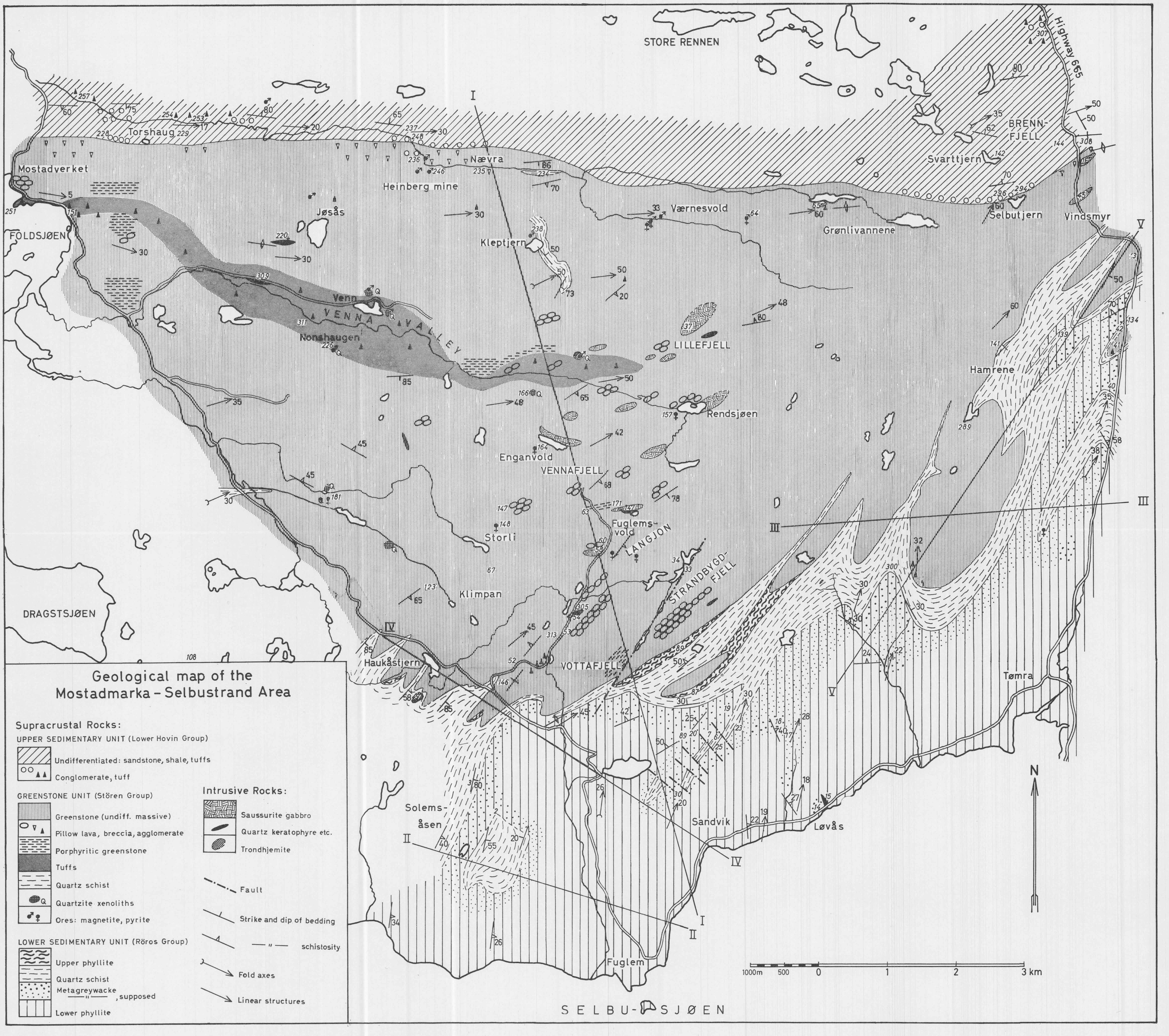


Plate X. Profiles.

PL. XI



Geological map of the Mostadmarka – Selbustrand Area

S. Hammerstads Boktrykkeri, Oslo.