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by Juhani Paakkola

Geologinen tutkimuslaitos · Otaniemi 1971

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# THE VOLCANIC COMPLEX AND ASSOCIATED MANGANIFEROUS IRON FORMATION OF THE PORKONEN-PAHTAVAARA AREA IN FINNISH LAPLAND

BY

JUHANI PAAKKOLA

WITH 55 FIGURES AND 9 TABLES IN THE TEXT AND TWO MAP APPENDICES

GEOLOGINEN TUTKIMUSLAITOS
OTANIEMI 1971

Helsinki 1971. Valtion painatuskeskus

## PREFACE

This study forms part of the research work on the Karelian initial magmatism and associated ore-forming processes carried out by a team in the Department of Geology at the University of Oulu.

Geological mapping and other field work was performed during the summers of 1965—67 by the present author in collaboration with the former mining company, Otanmäki Oy, which amalgamated with the Rautaruukki Oy steel company at the beginning of 1969. Minor revisions and photography were carried out in the summers of 1968 and 1969. Microscope examinations and other laboratory work were performed at the Department of Geology in the University of Oulu. Results of a large-scale geophysical survey, diamond drilling and other kinds of prospecting activity were at my disposal, but not all the details can be published at present, for reasons of industrial security. The chemical analyses were carried out by the Department of Chemistry of the Geological Survey of Finland.

During my work I was advised and encouraged in many ways by my distinguished teacher and superior, Professor Juhani Seitsaari, and I deeply appreciate the privilege of having been a disciple of his.

Many discussions with the other members of our team, Associate Professor, Tauno Piirainen and Mr. Risto Piispanen, Phil. Lic., have been very fruitful and instructive.

The chief geologist of Rautaruukki Oy, Mr. Heikki Paarma, M. A., was kind enough to put the company's very important research material at my disposal, and showed his active interest in the problems of the area during numerous animated discussions. Equally important and beneficial was the attitude of my director in the field work, Mr. Juhani Nuutilainen, Ph. D.

My close friend, Mr. Juha Sojakka, B. A., who accompanied me during the field work was an incomparable source of assistance.

Professor Herman Stigzelius, Director of the Geological Survey of Finland, kindly arranged for the publication of this paper as an issue of the Bulletin de la Commission géologique de Finlande.

I received welcome financial support from the Council of Natural Sciences, the Outokumpu Company Foundation and the Foundation of Savings Banks of Northern Finland.

Mrs Maija Outakoski drew the maps and illustrations, and Mr. Matti Kortesluoma, B. A. took some of the photographs.

Mr. J. J. Lavin, B. A., of the English Department at the University of Oulu kindly corrected the English of the manuscript.

To all the individuals and organisations that have aided me in this work, I wish to express my deepest gratitude.

Oulu, August 1970

*Juhani Paakkola*

ABSTRACT

**Paakkola, Juhani** 1971: The volcanic complex and associated manganiferous iron formation of the Porkonen-Pahtavaara area in Finnish Lapland. *Bull. Comm. géol. Finlande N: 247.* 83 pages, 55 figures, 9 tables and two map appendices.

The volcanics and associated manganiferous iron formation form a part of the Kittilä greenstone formation. The volcanic complex consists of lava-born rocks with closely associated pyroclastic and chemical sediments. The lava-born rocks are for the most part albite-amphibole-(epidote) and albite-chlorite rocks with various transitional forms, the rest being albite-biotite, albite-paragonite and ultrabasic rocks. Some are very rich in carbonates. Also hypabyssal intrusives, similar in mineral composition to those mentioned above, are met with. The pyroclasts are tuffaceous schists and volcanic breccias. The chemical sedimentary rocks are jaspilites, carbonaceous schists and sulphide schists with local iron and manganese enrichments.

The igneous rocks belong to the spilite-keratophyre association representing an independent, primary, spilitic magma. The occurrence of the ultrabasic lava-born rocks is due to early liquation of the spilitic magma at moderate or great depths. The iron and manganese of the iron formation, precipitated under oxide, carbonate or sulphide facies, derive from submarine volcanic emanations. The facies distribution was controlled by the bepth of the sea.

The volcanic complex and iron formation belong to the preflysch phase in the evolution of the Lapponian geosyncline and represent the initial magmatism of the Karelidic orogeny.

CONTENTS

# INTRODUCTION

This paper presents geological descriptions of an exceptionally well-preserved volcanic complex and associated manganiferous iron formation of Precambrian age. The aim of this study is to delineate the general features of the geology of the area, to deal with the problems of the magmatic evolution and the nature and origin of the iron formation. Attention is directed to the mutual relationship between the spilitic volcanism and ore formation.

## Location and physiography of the area

The area described in this paper is located in the eastern part of Kittilä parish, Finnish Lapland (Fig. 1). The geologically mapped area forms a gently curved arch, about 40 km in length and 5 km in breadth. The landscape is characterized by gently sloping hills and precipitous ridges that at some places have got fjeld-like features. The summit of Porkonen, the highest peak on the mapped area, rises 406 metres above sea level and 150 metres above the adjacent Lake Vesmajärvi. Owing to its height the area has become a watershed, from which the creeks, controlled by faults, lead waters to east and west. The creek sides generally have become swampy, and the area is surrounded by extensive open moors in the east and partly in the west, too.

## Review of the glacial history

The peculiar character of the glacial history of the area, described in detail by Tapio Mikkola (1967), is reflected even by the outcrop relations.

Central Lapland was already an accumulation area in the early stages of the Scandinavian ice sheet. So it is characterized by a scarcity of glacial erosion forms and by a richness of weathered rocks in situ, local till material, glaciofluvial accumulation and erosion forms, and by the shore lines of ice -dammed lakes. None of the roches moutonnée which are so typical of the landscape of Southern Finland exist here. Instead there are lots of rocky bottoms belonging to ancient glacial rivers, and deeply weathered rocks in situ. The pyroclasts especially and some of the chemical sediments are strongly weathered and broken, thus causing serious difficulties in tectonic observations. At many places the local moraines yield quite reliable information about the underlying bedrock. This is shown by means of trenching.

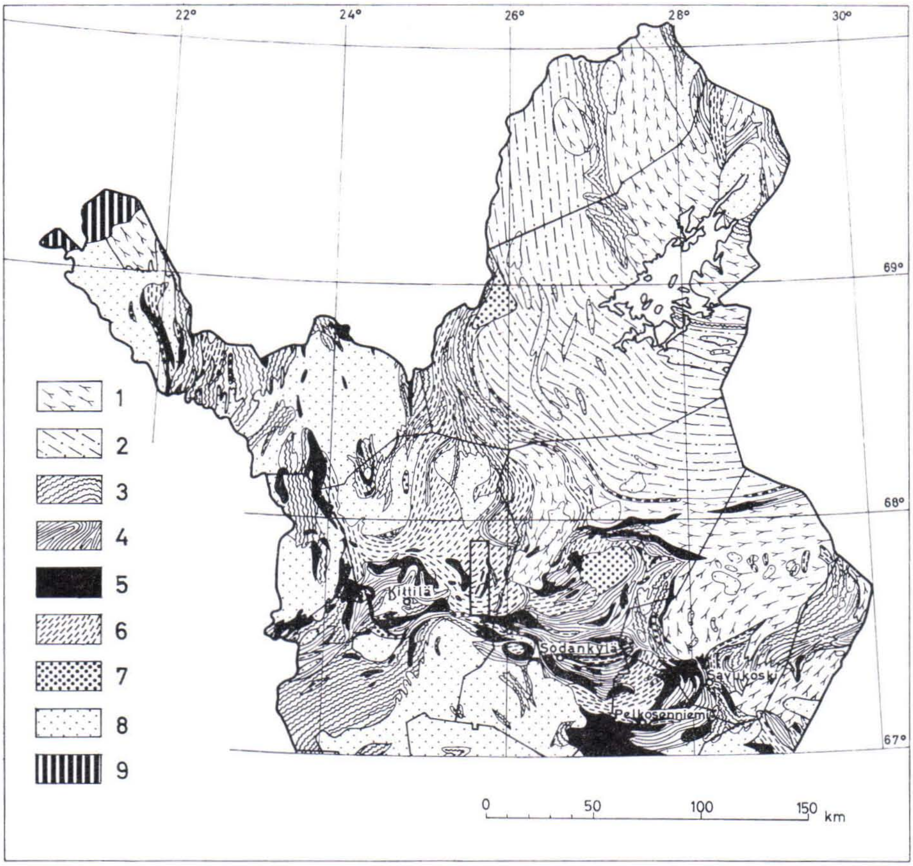


FIG. 1. Geological map of Finnish Lapland. The location of the Porkonen-Pahtavaara area is shown by the rectangel with heavy contours. 1 = granite gneiss, 2 = granulite, 3 = mica gneiss, 4 = phyllite and mica schist, 5 = quartzite, 6 = metabasalt, amphibolite and hornblende gneiss, 7 = basic intrusive, 8 = acid intrusive, 9 = Paleozoic schist in Caledonian mountain chain. (Generalized after the geological map of Finland drawn by Ahti Simonen, 1960).

## Previous works [1])

The centre part of the mapped area was previously known as the Porkonen-Pahtavaara ore field. Prospecting activity on this field started more than one hundred years ago. In 1864 committee surveyor P. W. Aurén found iron ores from the field which was two years later studied by the Governement Office of Mining under instructions from the Senate. The investigations were broken by the hard dearths of 1867—1868.

E. Sarlin and A. von Julin, who were sent by the Geological Commission, mapped the area both magnetically and geologically in 1900, but without any further results.

[1]) Based mainly on the article by Liisa Välitalo (1967).

The first serious attempt to take advantage of the ores was made by J. G. Norman in 1908. The area was mapped magnetically by Mossberg and ore geological studies were made by Otto Witt. On the basis of these studies Witt estimated the ore reserves of the area to be 88 million tons containing 40 per cent iron on an average.

In 1920 the Geological Commission sent an expedition to the field led by Eero Mäkinen. According to Mäkinen's calculations the iron ore reserves of the Porkonen-Pahtavaara ore field were 69 million tons averaging 40 per cent iron.

The next attempt with commercial aims was that of Suomen Malmi Oy in the years 1939—1940. These investigations were led by H. Rajahalli, and he made a final evaluation of 42 million tons of iron ores averaging 35 per cent iron.

Three years later, in 1943, the ores were studied by Vuoksenniska Oy under the leadership of Herman Stigzelius. Stigzelius stated that mining of the ores could not be profitable.

Partly in connection with the above-mentioned prospecting operations and partly independently scientific investigations were carried out, too. In an unpublished report Mäkinen described two kinds of rocks from the area: volcanic greenstones and leptites associated with the ores.

The first proper geologic treatise on the area was that of Victor Hackmann (1925). In this paper Hackmann shows that what Mäkinen look to be leptites, are in fact quartzites impregnated or intercalated with thin beds of iron ores. The ore minerals are hematite, and magnetite as its alteration product. The primary ore material he declares to have been deposited as hydrous iron oxide intercalating with sand and mud beds. On the basis of the original map by Mäkinen Hackmann compiled the first published geological map of the area (Fig. 2). Two years later Hackmann (1927) published a quite comprehensive and detailed study on the bedrock of the whole Kittilä area.

Mikkola (1941) based his ideas of the Porkonen-Pahtavaara area more on the above-mentioned studies of Hackmann than on personal field observations. Kaitaro (1949), on the other hand, was concentrated to a restricted area.

## Recent works

In the late 1950's the mining company Otanmäki Oy began to direct attention to Finnish Lapland as a potential manganese ore province on the initiative of the head geologist of the company, Mr. Heikki Paarma. This idea was supported by the manganese -bearing ore samples sent by the public from different parts of Lapland (see, Paakkola, 1967).

A totally new stage in the prospecting history of the Porkonen-Pahtavaara area began in 1963 when Asser Siitonen sent a small, black, weathered sample to Otanmäki Oy. The manganese content of the sample was interestingly high and

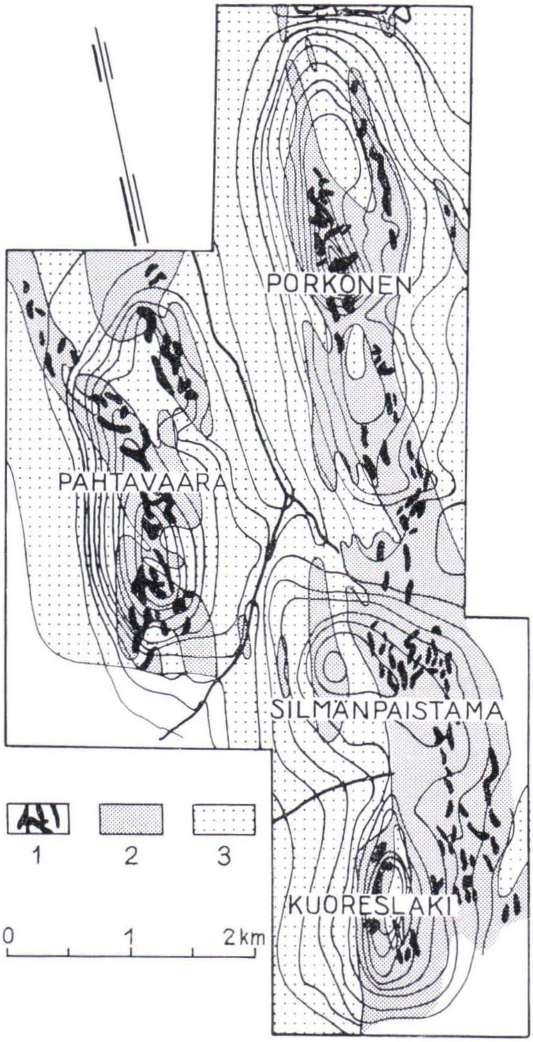


FIG. 2. Geological map of the Porkonen-Pahtavaara ore field. Compiled by Hackman (1925) after the original map of Mäkinen. 1 = iron ore, 2 = quartzite, 3 = greenstone. The interval between contourlines is 10 metres- Scale about 1 : 50 000.

resulted in the start of progressively expanding prospecting work. Boulder tracing, bedrock mapping, trenching, diamond drilling and geophysical surveying started in summer 1964. The greatest prospecting activity was in summer 1965, and it continued right up to summer 1968.

At the very beginning, the prospecting work was led by Väinö Makkonen, soon followed by Juhani Nuutilainen under the supervision of Heikki Paarma.

The bedrock mapping revised by the present author, was carried out by several persons.

Silvennoinen (1966) and Partio (1966) based their unpublished Master Theses, dealing with Petäjäselkä and Pahtavaara areas respectively, on the field works, and the present author used them as the basis of his unpublished Licentiate Thesis (Paakkola, 1969) which forms the major part of the content of this paper.

The stratigraphic problems of the schist area of central Lapland were elucidated by Kaarlo Mäkelä in his unpublished Master Thesis (1966) and Licentiate Thesis (1968).

## GENERAL GEOLOGIC SETTING

The rock crust of Finland is a part of the Baltic Shield, which in the west and north is bounded by the Caledonian mountain range, and in the south-east by the Russian platform. According to Simonen (1960) the Precambrian metamorphic rocks of Finland can be divided into three main units:

— the granite gneiss and granulite complex in eastern and northern Finland

— the Svecofennian schist belt in western and southern Finland.

— the Karelian schist belt in eastern and northern Finland.

Formerly it was supposed that the Svecofennian and Karelian schist belts belonged to two independent orogenic cycles, but nowadays it is generally recognized that they both belong to the same orogenic cycle representing different stages in the evolution of the geosyncline and different location within it. The evolutionary phase of that orogenic unit is represented by the Karelian schists only, no more by the so-called Svecofennian ones.

The Karelian schists in northern Finland form several separated basins that have many features in common, but many differences exist, too. One of these basins is the schist area of central Lapland consisting mainly of quartzites, mica schists and basic volcanics, all together known as the Lapponian series. The Lapponian series is overlain by coarse-grained molasse accumulation of the Kumpu-Oraniemi series. The geologic succession of the schist area of central Lapland is not known in detail. The main lines of the geology of central Lapland were drawn by Mikkola (1941) and in general there is no reason to doubt the correctness of them, even though it seems likely that at least a part of the aluminous schists in the Oraniemi area do not belong to the molasse facies suggested by Mikkola.

As already stated, this paper deals mainly with the Lapponian volcanics and associated manganiferous iron formation. These volcanics and hypabyssal intrusives, forming the centre of the schist area of central Lapland (Fig. 1), were previously known as the greenstones of Kittilä. They belong to a spilitic suite and extend from ultrabasic to acid members. The grade of metamorphism is rather low except for the marginal zones in the north where greenstones change to amphibolites. No true clastic sediments are found with the greenstones in large areas. In the north-eastern part the greenstones intercalate with quartzites and mica schists forming the lowest horizon of the Lapponian series. In the south-west the greenstones grade to the flysch sediments of the Lapponian series overlain by the molasses of the Kumpu-Oraniemi series.

The rocks of the iron formation are intimately interfingered with the volcanics and they do not depend genetically on the clastic sediments.

## PETROGRAPHY OF THE ROCKS

### Lava-born extrusives

Various extrusives form the largest group among the rocks of the volcanic complex. Due to the capriciousness of the outcrops all the kinds of extrusives are marked with the same symbol on the map. It would not have been possible to distinguish between different types, especially when taking into consideration the scale of the map and the dimensions of the single rock units.

In the following, the lava-born extrusives are grouped and described on the basis of mineral compositions which also reflect quite clearly the chemical composition of the rocks.

#### Nonfeldspathic rocks

The chemical character of the nonfeldspathic extrusives is obviously ultrabasic. Their amount on the mapped area is very small, but they form large masses in central Sodankylä, east from the Porkonen-Pahtavaara area.

A m p h i b o l e - c h l o r i t e rocks, which are so abundant in Sodankylä, are found only at one place on the mapped area. It is situated about 1 km WNW from Lake Kuolajärvi. The rock is rather homogeneous, slightly schistose and greenish grey in colour. The microscopic structure is shown in Fig. 3. The main minerals are amphibole and chlorite, accessories are opaque ore, epidote and albite. The mineral proportions are shown in Table 1.

Amphibole crystals are short and tabular. The average Ø is 0.15 mm. The optical properties vary: $-2V = 73°-80°$ and $c \Lambda Z = 17°-24°$, the colour is slightly greenish and pleochroism almost invisible.

Chlorite is colourless or slightly greenish, birefringence is very weak, the highest interference colour is abnormal greyish brown of the first order, and the elongation is positive.

Small grains, measuring less than 0.05 mm in diameter, of albite, opaque ore and epidote occur as accessories.

A m p h i b o l e - e p i d o t e rock, found on the western slope of Pahtavaara, is quite basic in composition, too. The structure of this rock proves quite obviously

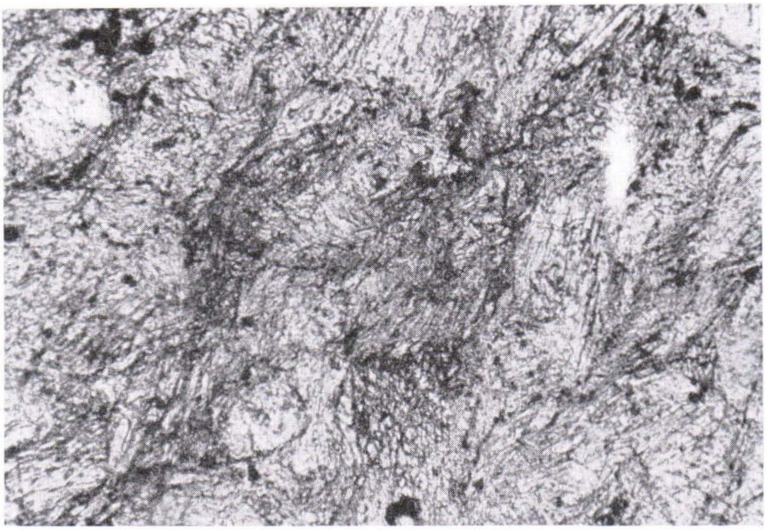

FIG. 3. Structure of a lava-born amphibole-chlorite rock. Thin section, without analyzator. 100 x.

TABLE 1

Mineral compositions of the nonfeldspathic extrusives, in percentages by volume.

| Minerals | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| Amphibole | — | 45.5 | 56.0 | 73.0 | 69.5 | 55.1 |
| Chlorite | — | 48.0 | 38.0 | 20.0 | 30.1 | — |
| Serpentine | 19.7 | — | — | — | — | — |
| Talc | 36.7 | — | — | — | — | — |
| Carbonate | 41.0 | 2.8 | — | — | — | — |
| Ore minerals | — | 3.7 | 5.5 | 3.0 | — | — |
| Apatite | — | — | 0.5 | — | — | — |
| Albite | — | — | — | 4.0 | — | — |
| Epidote | — | — | — | — | — | 41.9 |
| Accessories | 2.6 | — | — | — | 0.4 | 3.0 |

1. Soapstone-like rock, Kuolavaara, Kittilä. (Anal. 1, Table 7, p. 52).
2. Amphibole-chlorite rock, Sattasvaara, Sodankylä. (Anal. 3, Table 7).
3. Amphibole-chlorite rock, Kummitsoiva, Pelkosenniemi. (Anal. 4, Table 7).
4. Amphibole-chlorite rock, Keikkuma-aavan saari, Sodankylä. (Anal. 2, Table 7).
5. Amphibole-chlorite rock, Rovanpää, Kittilä. (Anal. 5, Table 7).
6. Amphibole-epidote rock, Pahtavaara, Kittilä.

its volcanic origin (Fig. 4). The mineral composition of the rock is: amphibole and epidote as main members, and albite, chlorite, sphene and opaque ore as accessories. The mineral proportions are shown in Table 1.

FIG. 4. Lava-born amphibole-epidote rock. The dark laths consist of saussurite, the bright grains are epidote, and the groundmass consists of small-grained amphibole. Thin section, without analyzator. 30 x.

Amphibole is lathy or acicular, ragged and without crystal forms. It is colourless or very slightly greenish. Birefringence is about 0.018, $c \Lambda Z = 10°—16°$, and $—2V = 76°—85°$.

Epidote occurs in a twofold manner; either in dark saussurite laths or as anhedral grains. It is probable that the dark saussurite laths are pseudomorphs after basic plagioclase.

A m p h i b o l e - e p i d o t e - c h l o r i t e rocks are more abundant than the two types described above. They are found at Porkonen and Petäjäselkä, and as inclusions in a serpentine rock at Nolppio. These rocks are generally even- and fine-grained, homogeneous and unfoliated. A few porphyritic variants are found (Fig. 5).

The laths shown in the picture consist of saussurite and the large phenocrysts are filled by fine-grained chlorite mass bounded by a narrow zone of colourless amphibole. The brown matrix consists of very fine grained amphibole and chlorite. It is obvious that the primary rock in this case has been a basic, porphyritic, lava-born rock, bearing phenocrysts of euhedral pyroxene grains and anortite-rich plagioclase laths in an aphanitic or vitreous matrix.

As mentioned already the main area where the ultrabasic and basic extrusives exist is in the eastern part of the basin of central Lapland. Their volcanic origin is proved by pillow structures, amygdules and agglomeratic structures (Figs. 26 and 27 on p. 35).

FIG. 5. Porphyritic amphibole-epidote-chlorite rock. Explanation in the text. Thin section, without analyzator. 30 x.

The volcanic nature of the soapstone-like rocks of Kuolavaara is not indisputable, but the present author is disposed to describe them in this connection, because they completely conform with their surroundings and they have been folded together with other kinds of volcanics. They occur as beds a few metres thick between other lavas and tuffaceous schists. The colour of the fresh rock is light yellowish brown, the weatherd surface is dark brown and rough. The schistosity of the rock is distinct. The rock contains various amounts of magnesite, talc, serpentine and opaque ore minerals.

## Feldspar-bearing rocks

Amphibole-epidote-albite rocks and associated amphibole-epidote-chlorite rocks, amphibole-chlorite-albite rocks and amphibole-albite rocks form the largest group among the rocks described in this chapter. They are generally greenish-grey coloured, unfoliated and homogeneous or at some places weakly brecciated. This kind of breccia is obviously of tectonic origin, and the crack filling chlorite-epidote-ore mass is generated from the rock itself.

The microscopic structure is most usually pilotaxitic (Fig. 6) or porphyritic (Fig. 7).

Amphibole is generally colourless or weakly greenish, $c \Lambda Z = 12°—22°$ and $—2V = 72°—85°$. It occurs as small needles or ragged lamellae, often as small inclusions in albite phenocrysts.

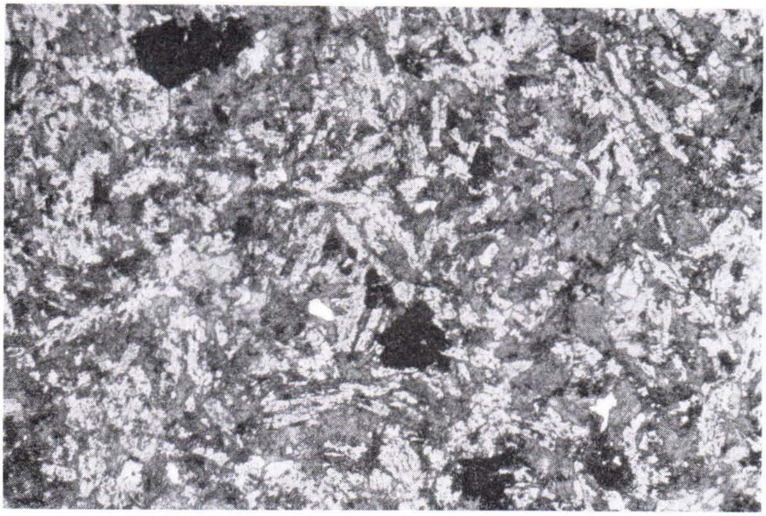

Fig. 6. Pilotaxitic amphibole-albite rock. The pale laths are albite, the grey mineral is amphibole and black magnetite. Thin section, without analyzator. 30 x.

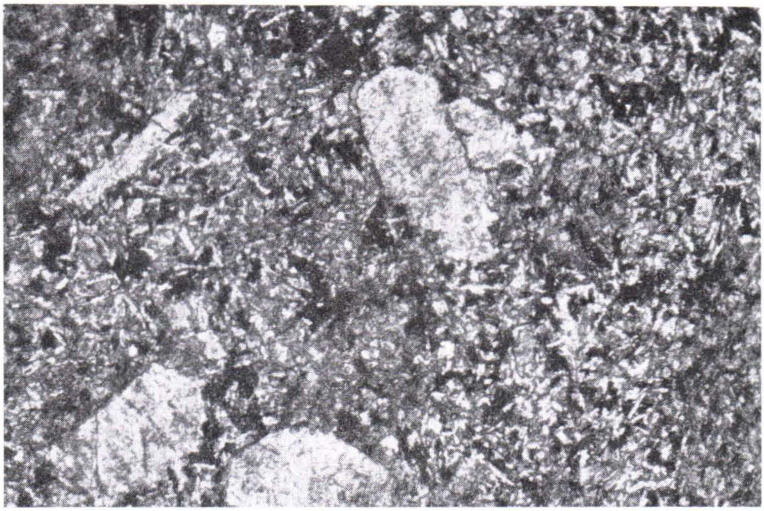

Fig. 7. Porphyritic amphibole-epidote-albite rock. Thin section without analyzator. 30 x.

The composition of plagioclase varies within a fairly limited range, $An_{3-6}$, determined by U-stage and immersion method. It occurs in a threefold manner: as phenocrysts, as subophitic laths or as anhedral grains between the amphibole lamellae.

Epidote is a clinozoisitic one, referring to the optic properties: $-2V = 85°-90°$ and birefringence very low.

FIG. 8. Porphyritic albite-chlorite rock. Thin section, without analyzator. 30 x.

Chlorite is an alteration product of amphibole in most cases, but it may exist independently as crack filling material, too. The colour is weakly greenish, and sometimes it is very slightly pleochroic, $Z \perp (001)$. The highest interference colour is abnormal blue of the first order, birefringence about 0.004.

Sphene exists either as small wedge-shaped prisms or as brown, round granules (insect egg sphene).

The mineral compositions of the rocks of this category are shown in Table 2.

The amount of the plagioclase-bearing extrusives characterized by sheet silicates as dominating mafic minerals is smaller than that of the amphibole and epidote rich ones. There seems to be no difference in the areal distribution of these two types of rock.

On the basis of the mineral compositions the plagioclase-bearing extrusives, rich in sheet silicates, can be grouped as follows: albite-chlorite rocks, albite-chlorite-epidote rocks, albite-biotite rocks, and albite-paragonite rocks.

The external appearance of the chlorite-rich rocks is quite similar to that of amphibole-rich types, and frequently it is hardly possible to distinguish between these rocks megascopically. The biotite-bearing varieties are generally dark and the paragonite-bearing types light. Like the amphibole-bearing ones the extrusives, characterized by sheet silicates, are fine-grained or aphanitic, homogeneous, weakly or not at all schistose, and often slightly brecciated.

The microscopic structure of albite-chlorite rocks is pilotaxitic, bostonitic, or porphyritic (Fig. 8). Some types are scoriaceous (Fig. 9).

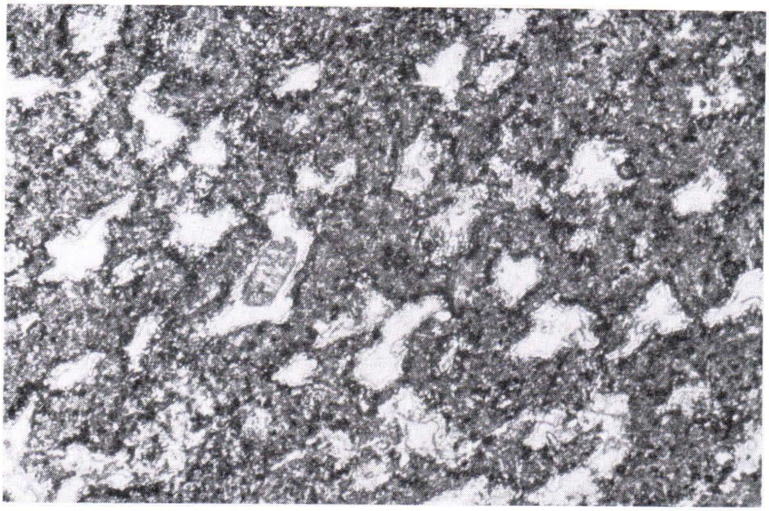

FIG. 9. Scoriaceous albite-chlorite rock. The amygdules are filled by chlorite and biotite and bordered with bright rims of fine-grained albite. Thin section, without analyzator. 30 x.

As can be seen from the pictures, plagioclase ($An_{3-6}$) occurs as microlitic laths or as phenocrysts, sometimes as small anhedral grains varying in shape and size. Bright and pure grains are met only seldom. The phenocrysts are regularly turbid and they enclose other constituents of the rock, especially chlorite.

The optical properties of chlorite are distinctly different from those of the chlorites in amphibole-rich extrusives. Pleochroism is distinct; $X$ = colourless, $< Z$ = dark green, $X \perp (001)$, birefringence = 0.003, $-2V = 3°-6°$. The refractive indices are: $X = 1.647 \pm 0.002$, $Z = 1.650 \pm 0.002$.

Epidote is similar to the epidotes of amphibole-bearing extrusives.

There is a distinct difference between amphibole-rich and chlorite-rich extrusives in the amount of opaque ore minerals which are very abundant in the last mentioned rocks. The quantitative relation of sphene is reverse.

Another marked difference between amphibole- and chlorite-rich extrusives is in the amount of carbonate minerals. The chlorite-rich extrusives almost without exception contain carbonates, while the occurrence of carbonates in the amphibole-rich extrusives is more irregular. This phenomenon will be discussed later.

The occurrences of albite-biotite rocks are few in number, but they are found for instance at Petäjäselkä, Pahtavaara and Ketunliesukuusikko. Except for the dark colour, the external appearance of albite-biotite rocks is quite similar to the other extrusives, and the only essential difference when compared to other sheet silicate-rich extrusives is the existence of biotite instead of chlorite and paragonite.

In some extrusives at Petäjäselkä, Pahtavaara and Kuolavaara the sheet silicate is paragonite, identified by X-ray (Partio, 1966). These albite-paragonite

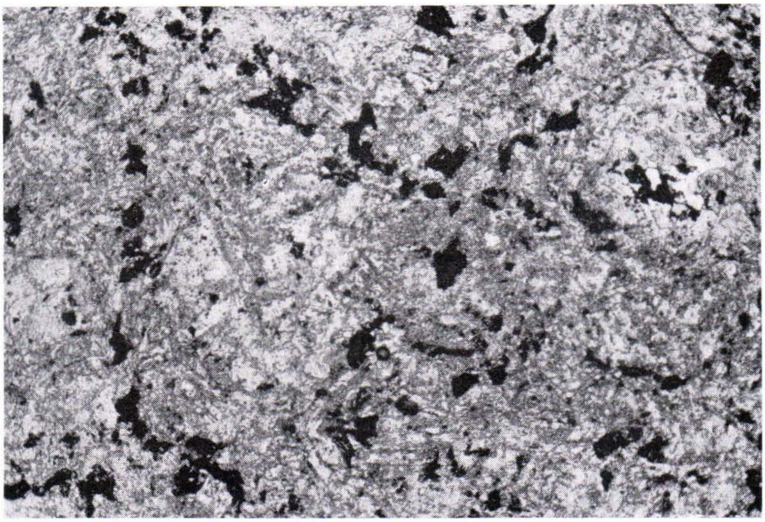

FIG. 10. Lava-born albite-magnetite-paragonite rock. Thin section, without analyzator. 30 x.

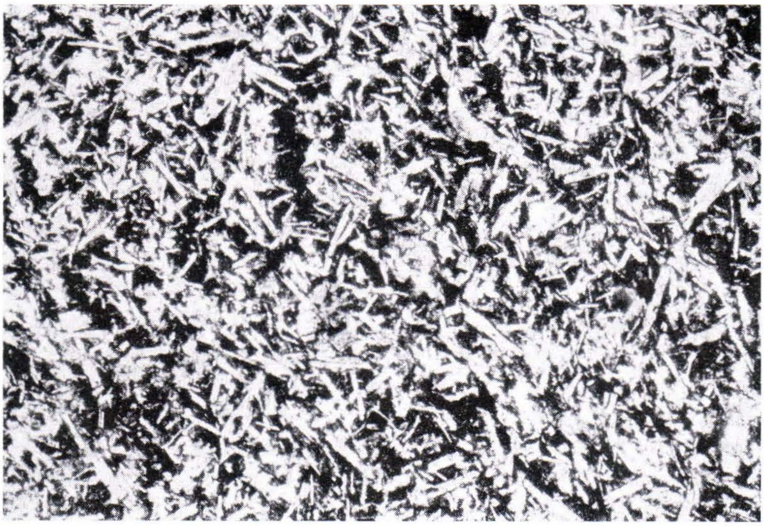

FIG. 11. Pilotaxitic albite-magnetite rock. Thin section, without analyzator. 30 x.

rocks are light grey or yellow, and the microscopic structure is similar to other extrusives (Fig. 10).

Paragonite is weakly greenish in thin section. It occurs as small flakes in and between albite laths. $X \perp (001)$, $-2V = 42°—46°$. The refractive indices, determined by immersion method, are $X = 1.568 \pm 0.002$, $Z = 1.598 \pm 0.002$, $Z—X = 0.030$.

Albite occurs as small laths crowded with flaked paragonite and small, acicular apatite prisms with parallel orientation.

Like the other sheet silicate-bearing extrusives the paragonite-bearing rocks are characterized by an abundance of opaque iron oxides and an almost total lack of sphene. The tenor of carbonate is variable.

The amount of paragonite is variable, too, and very close relatives to albite-paragonite rocks are albite-magnetite rocks. The other constituents of these rocks are paragonite, chlorite, carbonate and apatite. Of special interest is a black, 3–4 metres thick bed of albite-magnetite rock found at Pitslomakuru containing round quartz pebbles, 1—5 mm in diameter. The microscopic structure of the rock is pilotaxitic (Fig. 11). The rock consists of albite laths with magnetite and fine grained interstitial leucoxene. Accessories are paragonite and chlorite. The round amygdules are filled with pure, granular quartz.

The mineral compositions of the sheet silicate rich extrusives are shown in Table 3.

## Pillow lavas and flow lavas

The photographs (Figs. 3—11) clearly show that we have a series of lava-born extrusives to deal with. Further evidence for the volcanic origin of these rocks is provided by pillow structures and flow structures found at many places in the mapped area and outside of it.

Quite well preserved pillow lavas are found at Petäjäselkä and at Vesmaselkä on the eastern side of the River Karjakkojoki (Fig. 12).

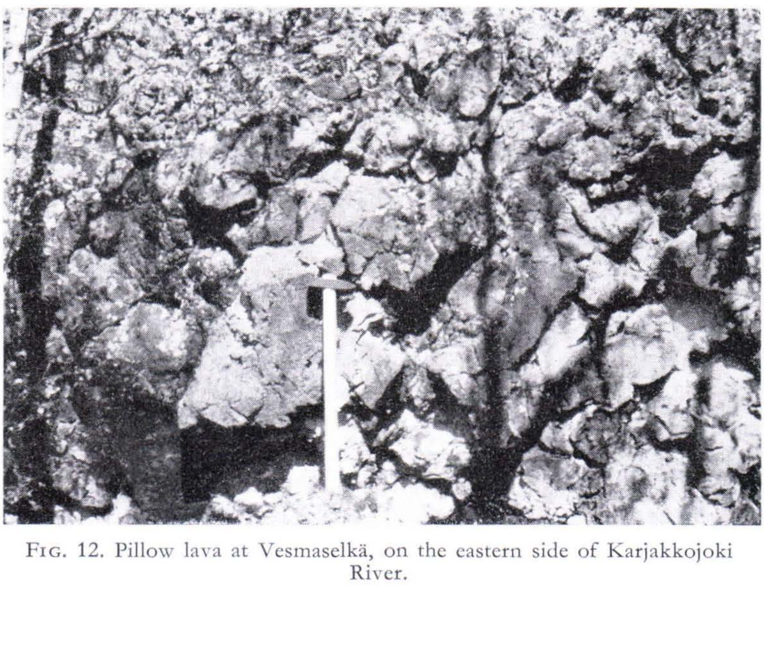

FIG. 12. Pillow lava at Vesmaselkä, on the eastern side of Karjakkojoki River.

At Petäjäselkä, on the western slope of Puolalaki, the pillows are 40—50 cm bread and 15—20 cm thick. There is an intensive radial cracking in the core of the pillows. The rock is very typical albite-amphibole rock. No marked difference in mineral composition has been observed between the core and marginal zones of the pillows.

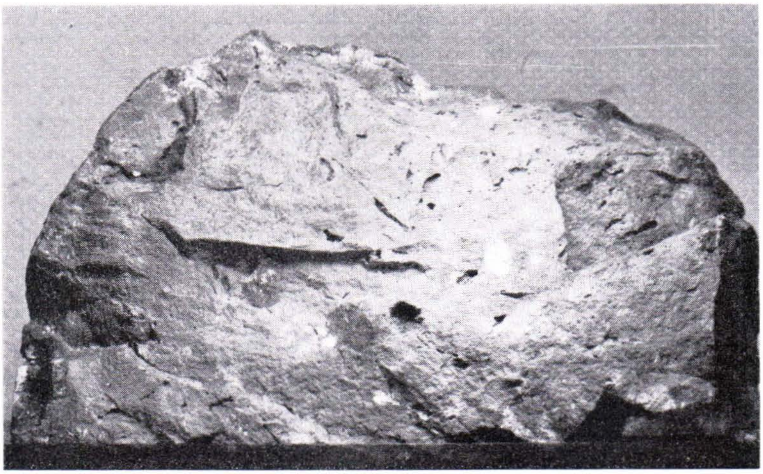

FIG. 13. A pillow with a small contraction cavity in the centre and round or elongated vesicles especially in the marginal zone.

FIG. 14. Flow structure in a lava-born albite-epidote-amphibole rock at Vesmaselkä, about 100 m west from the junction of the road to Vesmajärvi.

The form of the pillows at Vesmaselkä is preserved better than at Petäjäselkä. Most of the pillows look like round or oval loaves. The diameter of the pillows is about 35 cm on an average, and the thickness is about 15—20 cm. The contraction cavities are generally filled by calcite. Round or elongated vesicles, filled by chlorite, epidote or calcite, characterize the marginal zone of the pillows (Fig. 13). The rock is an albite-epidote-amphibole one, and the inserted rock between the pillows is either the same kind of albite-epidote-amphibole rock as in the pillows or calcitic limestone.

Flow structures are found at many places at Petäjäselkä, Pitslomakuru, Silmänpaistama, Pahtavaara and Vesmaselkä. The flow at Vesmaselkä (Fig. 14) is a typical albite-epidote-amphibole rock described on p. 15.

### Carbonate-rich extrusives

Almost every type of extrusive contains carbonate minerals to a greater or lesser extent. The existence and nature of the carbonate minerals in the extrusives is such a variable and capricious phenomenon that it deserves a summarizing description of its own.

The carbonate minerals found in extrusives are calcite, dolomite, ankerite and siderite, identified by X-ray powder camera and immersion method. No direct relationship between the host rock and the kind of carbonate has been observed.

The existence of carbonate in the rock is easily verified even megascopically on the weathered surface of the rock because the carbonates have been corroded very strongly and the surface is thus uneven and rough. The colour of the weathered surface depends on the composition of the carbonate. The weathered crust of dolomite may be several centimetres thick and the colour is red-brown, the weathered crust of calcite is light grey and that of siderite is brick-red. The colour of the weathered surface of mangano-siderite varies from dark brown to black depending on the Fe/Mn proportion.

The richness of carbonates in the bedrock can easily be observed even on areas covered by till, due to the luxuriant growth of the vegetation. This fact has been used as a guide in the following of carbonate rich horizons both from the ground and air (Paarma *et al.*, 1968).

Most extrusive rocks contain small amounts, less than 5 per cent, of carbonate minerals irrespective of the mineral content of the rock. In these cases carbonate exists as irregular aggregates in the interstices or replacing other minerals. It seems possible that a part of this carbonate is primary but that a marked part of it is secondary is proved by the replacement features.

The average carbonate content of those extrusives with amphibole as the predominating mafic mineral is distinctly lower than the carbonate content of the extrusives rich in sheet silicates (*cf.* Tables 2 and 3). Especially rich in carbonate are the amygda-

TABLE 2

Mineral compositions of lava-born amphibole-(epidote-chlorite-)albite rocks, in percentages by volume.

| Minerals | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
|---|---|---|---|---|---|---|---|---|---|
| Amphibole ................ | 59.2 | 44.4 | 33.2 | 28.0 | 55.0 | 51.6 | 58.4 | 28.0 | 52.0 |
| Epidote .................. | — | 1.6 | 12.2 | 10.0 | 3.0 | 0.8 | 11.6 | 4.8 | 1.3 |
| Chlorite .................. | 6.0 | 10.0 | 8.2 | 21.0 | 5.0 | 1.0 | — | 23.0 | 4.5 |
| Albite .................... | 29.8 | 34.8 | 14.3 | 22.0 | 27.0 | 33.2 | 18.4 | 27.0 | 35.0 |
| Sphene .................. | — | 8.8 | 3.5 | 8.0 | 5.5 | 10.7 | 10.2 | 3.0 | — |
| Carbonate ................ | 5.0 | — | 28.0 | — | 2.3 | 3.6 | 1.4 | 3.2 | — |
| Opaque .................. | — | 0.9 | 0.6 | 1.0 | — | 0.3 | — | — | 2.2 |
| Quartz .................. | — | — | — | — | 1.0 | — | — | 11.0 | — |
| Biotite .................... | — | — | — | — | 0.8 | — | — | — | — |
| Apatite .................. | — | — | — | — | 0.4 | — | — | — | — |

1. Amphibole-albite rock, Petäjäselkä, Kittilä. (Anal. 9, Table 7, p. 52).
2. Amphibole-albite rock, Pahtavaara, Kittilä.
3. Amphibole-epidote-albite-carbonate rock, Pahtavaara, Kittilä.
4. Amphibole-epidote-chlorite-albite rock, Sotkajärven saari, Kittilä. (Anal. 7, Table 7).
5. Amphibole-albite rock, Aihikkoselkä, Sodankylä. (Anal. 10, Table 7).
6. Amphibole-albite rock, Haurespää, Kittilä.
7. Amphibole-epidote-albite rock, Vesmaselkä, Kittilä.
8. Amphibole-chlorite-albite rock, Harrilompolo, Kittilä. (Anal. 14, Table 7).
9. Amphibole-albite rock, 5.5 km north from Kersilö, Sodankylä. (Anal. 11, Table 7).

TABLE 3

Mineral compositions of lava-born rocks characterized by sheet silicates, in percentages by volume.

| Minerals | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| Albite .................... | 35.8 | 39.0 | 57.3 | 21.0 | 64.8 | 41.9 | 42.7 |
| Chlorite .................. | 0.6 | 13.6 | 29.5 | 3.1 | 12.4 | — | 2.3 |
| Biotite .................... | 39.4 | 31.3 | 9.5 | 7.9 | — | — | — |
| Paragonite ................ | — | — | — | — | — | 34.8 | 5.6 |
| Carbonate ................ | 8.6 | 9.2 | 0.4 | 65.8 | 15.6 | — | — |
| Sphene .................. | — | 1.8 | 0.2 | — | — | 10.0 | 48.4 |
| Opaque .................. | 15.6 | 5.1 | 9.1 | 2.2 | 7.2 | 7.4 | |
| Apatite .................. | — | — | — | — | — | 2.1 | — |

1. Albite-biotite rock, Ketunliesukuusikko, Kittilä.
2. Albite-biotite-chlorite rock, Kuolavaara, Kittilä. (Anal. 16, Table 7, p. 52).
3. Porphyritic albite-chlorite-biotite rock, Ketunliesukuusikko, Kittilä. (Anal. 15, Table 7).
4. Carbonate-albite-biotite rock, Silmänpaistama, Kittilä.
5. Albite-chlorite-carbonate rock, Petäjäselkä, Kittilä.
6. Albite-paragonite rock, Pahtavaara, Kittilä.
7. Albite-magnetite rock, Pitslomakuru, Kittilä.

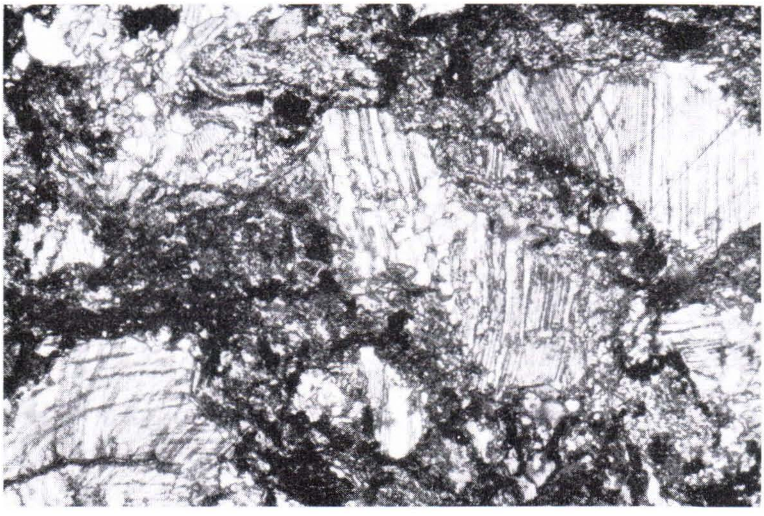

Fig. 15. Dolomite mesh of the sheet silicate net in the matrix of the carbonate breccia at the western end of Kuolavaara. Thin section, without analyzator. 30 x.

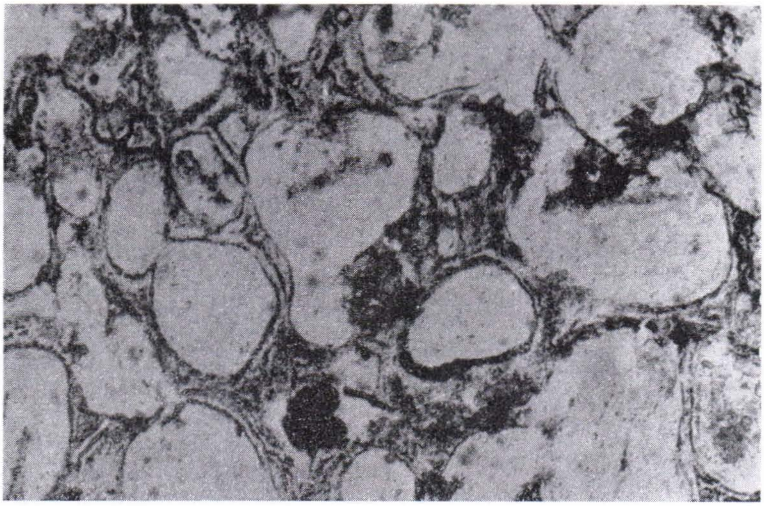

Fig. 16. Partly spherulitic albite nodules filling the mesh of the albite-magnetite net in the matrix of the carbonate breccia at the western end of Kuolavaara. Thin section, without analyzator. 30 x.

loidal extrusives, such as the amygdaloidal albite-biotite carbonate rock on the western slope of Silmänpaistama, the carbonate breccia of Kuolavaara, the carbonate rich extrusives of Petäjäselkä, and the horizon of carbonate rich extrusives on the southern slope of Kuolavaara.

These rocks are schistose and more or less heterogeneous caused by the uneven distribution of carbonates and silicates in the rock.

The carbonate-bearing extrusives are best exposed at the western end of Kuolavaara, where they crop out as a carbonate breccia. The carbonate breccia of Kuolavaara is peculiar and controversial in external appearance, having features of both igneous and sedimentary rocks, thus leading to contradictory interpretations. The carbonate-rich extrusives and breccias of Kuolavaara are split into three different parts. The northernmost of them is strongly schistose and folded, striking east and dipping almost vertically. The rocks of the south-western part are only weakly schistose and the dip of the obscure bedding is 10°—15° to the south-west. This south-western part in particular contains several strongly brecciated zones. The brecciating rock is very rich in dolomite with a net-like matrix of biotite, chlorite and iron oxide streaps. The mesh of this silicate net consists of dolomite (Fig. 15). At some places the brecciating rock becomes hololeucocratic resembling the albite-carbonate felses so common in Lapland. The minerals of these light parts are albite, dolomite, quartz and magnetite. The microscopic structure can be quite conspicuous. The distribution of carbonate is uneven. In the carbonate-poor parts fine grained albite and magnetite form a net with meshes of coarser spherulitic albite (Fig. 16). The result is thus a variolite-like rock.

The brecciated rocks are varying extrusives and tuffaceous schists. The amount of the brecciated fragments varies irregularly from zone to zone, generally it does not exceed 20 per cent of the volume. The commonest type in the fragments is very fine-grained chlorite-albite-paragonite schist impregnated with iron oxide dust. In the schist fragments the carbonate is lacking almost entirely and the contact between

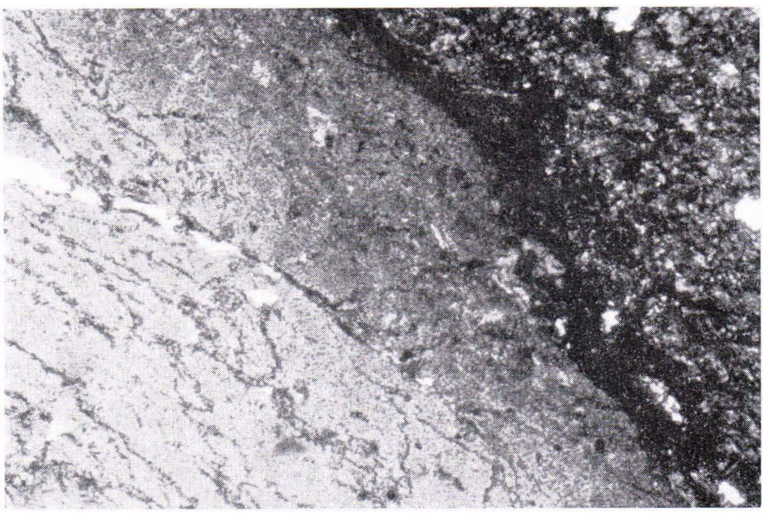

FIG. 17. Contact between chlorite-albite-paragonite rock fragment and brecciating carbonate-rich matrix. Thin section, without analyzator. 30 x.

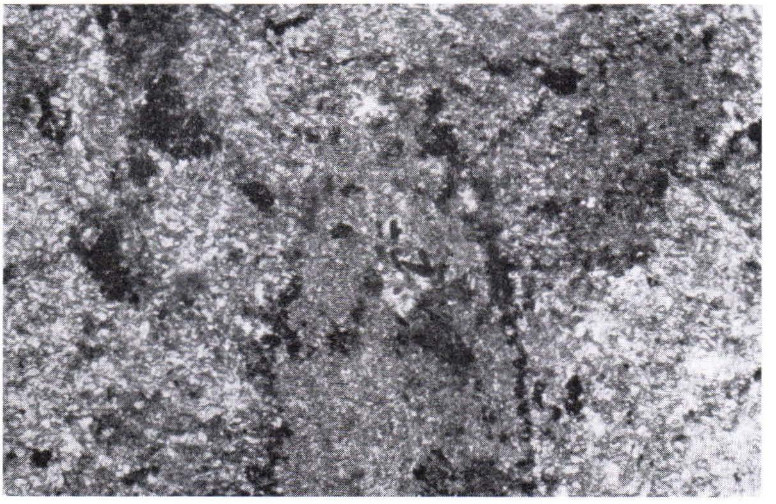

FIG. 18. Ghost-like remnants of albite-biotite-magnetite rock fragments in the carbonate-rich matrix. Thin section without analyzator. 30 x.

the fragments and the carbonate-rich matrix is sharp. Around the fragments there is a 0.2—1 mm thick zone enriched in biotite and iron oxides. In the fragments the contact effect is restricted to a 0.2—0.8 mm thick zone and this zone is also enriched in biotite and fine dispersed iron oxide dust (Fig. 17).

The amount of true lava-born rocks among the fragments is very scanty, and the existence of them is restricted to a few narrow zones in slightly deformed parts. The lava-born rock fragments are fine-grained, obscure amygdaloidal albite-biotite magnetite rocks with oval or elongated amygdules of chlorite and carbonate. The contact between the fragments and the matrix is either sharp or diffuse, and some of the fragments have been assimilated into the matrix almost completely, existing as ghost -like remnants only (Fig. 18).

The question of the original nature of this breccia remains open. The obscure bedding and the zoning of the breccia would indicate sedimentary origin, but the microscopic structures of the brecciating material shown in the pictures (Figs. 15—18) could also be taken as indicating an eruptive origin.

## Hypabyssal intrusives

Besides the extrusive igneous rocks there are numerous hypabyssal intrusives in the mapped area. The amount of them is not known exactly because of the outcrop relations, but all the reliably observed hypabyssal rocks are shown on the map. These intrusives belong to the same spilitic suite as the extrusives and their chemical compositions range within equal limits (see, Table 7, p. 52). Most of the intrusives

TABLE 4

Mineral compositions of hypabyssal intrusives, in percentages by volume.

| Minerals | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
|---|---|---|---|---|---|---|---|---|---|
| Albite | 42.8 | 33.0 | 26.0 | 32.0 | 31.9 | 42.4 | 69.0 | 72.4 | 59.5 |
| Pyroxene | 51.2 | — | — | — | — | — | — | — | — |
| Amphibole | — | 42.0 | 40.0 | 38.5 | 53.1 | 33.3 | — | — | — |
| Epidote | 2.2 | 15.0 | 24.0 | 9.5 | 8.9 | 0.4 | — | — | — |
| Biotite | — | 0.5 | 3.5 | 7.5 | — | — | 11.3 | — | — |
| Chlorite | 1.3 | 5.5 | 3.0 | — | — | 11.8 | — | — | 1.0 |
| Carbonate | — | — | — | — | — | 6.5 | — | 11.1 | 18.0 |
| Sphene | — | 2.5 | 2.3 | 1.1 | — | 4.3 | — | — | — |
| Apatite | — | 0.3 | — | 0.4 | — | — | — | — | — |
| Opaque | 2.5 | 1.2 | 1.0 | 11.0 | — | 1.3 | 8.7 | 13.8 | — |
| Quartz | — | — | — | — | — | — | 6.2 | — | 20.0 |
| Accessories | — | — | — | — | 6.1 | — | 4.8 | 2.7 | 1.5 |

1. Albite-pyroxene rock, Vesmaselkä, Kittilä.
2. Albite-amphibole rock, Holkkuavaara, Kittilä. (Anal. 13, Table 7, p. 52).
3. Albite-amphibole rock, Holkkuavaara, Kittilä. (Anal. 8, Table 7).
4. Albite-amphibole rock, 6 km north from the church, Kittilä. (Anal. 6, Table 7).
5. Albite-amphibole rock, Petäjäselkä, Kittilä.
6. Albite-amphibole rock, Haurespää, Kittilä. (Anal. 12, Table 7).
7. Albite-biotite rock, Mustavaara, Kittilä.
8. Albite-quartz-carbonate rock, Mustavaara, Kittilä.
9. Albite-quartz-carbonate rock, Petkula, Sodankylä. (Anal. 17, Table 7).

conform with their surroundings, differing from the extrusives only in structure. True intrusive, penetrating contacts are found at Pahtavaara, Porkonen and southeast of the Lake Kuolajärvi.

On the basis of chemical composition the hypabyssal intrusives can be divided into three categories:

— ultrabasic rocks: serpentinites and peridotites
— basic and intermediate rocks: albite gabbros and diabases
— acid rocks: albitites

Mineral compositions of hypabyssal intrusives are shown in Table 4.

### Serpentinites and peridotites

The ultrabasic intrusives are either independent serpentinite intrusions or ultrabasic parts of gravitatively differentiated layered intrusions.

Independent serpentinite intrusions exist especially in the northern parts of the mapped area at Nolppio and north of the river Rourajoki. The varying mineral composition of these rocks consists of serpentine, talc, chlorite, and magnesite. The accessories are magnetite and, very occasionally, colourless amphibole. The relation of the serpentinite intrusion of Nolppio to the surrounding albite-epidote-amphibole rock is distinctly intrusive, but no signs of contact effects have been observed.

The ultrabasics from the bottom of a differentiated sill at Vesmaselkä, between the Lakes Karijärvi and Vesmajärvi, differ drastically from the serpentinite of Nolppio. The ultrabasic part of the sill consists of serpentine pseudomorphs after olivine and monoclinic pyroxene. The accessories are magnetite and colourless amphibole.

X-ray determinations and optical properties of the serpentines indicate that there are three different species of serpentines: fibrous chrysotile, flaked antigorite and weakly pleochroic, nearly isotropic bastite.

The crystal forms of pyroxene are well developed. The optical properties: colourless, $+2V = 54°—60°$, $c \Lambda Z = 42°—56°$, indicate that the pyroxene is a common augite. Some of the pyroxene grains have been altered to colourless amphibole in the margins.

On the western slope of Petäjäselkä there are numerous boulders of altered peridotite consisting of serpentine minerals, with accessory talc, monoclinic colourless amphibole and magnetite.

## Albite gabbros and diabases

To the east, the Vesmaselkä peridotite described above gradually changes to albite-pyroxene gabbro, and this in turn gives way further east to albite-amphibole gabbro. Pyroxene bearing rocks have not been found at any other place in the area.

The microscopic structure of albite-pyroxene gabbro is hypidiomorphic. The rock is middle- and even-grained. The colour of the rock is greenish grey. The main minerals are augite (colourless, $+2V = 54°—60°$, $c \Lambda Z = 40°—46°$) and turbid, strongly twinned plagioclase ($An_{2-4}$). The accessories are epidote, chlorite, leucoxene, apatite and magnetite.

Most of the hypabyssal intrusives are varieties of albite-amphibole gabbros and diabases. They contain in various proportions albite, hornblende and epidote as main minerals; and chlorite, leucoxene, apatite and magnetite as accessories.

Plagioclase, $An_{4-8}$, exists as turbid, tabular, equidimensional grains that are generally strongly twinned according to the albite law. The average diameter of albite grains is about 0.3 mm.

Amphibole exists as tabular grains or as ragged lamellae. It is only weakly pleochroic: $Z$ = light green, $X$ and $Y$ = almost colourless. Some of the amphibole grains have been altered to chlorite in the margins. $c \Lambda Z = 14°—18°$ and $—2V = 70°—84°$. Meriläinen (1961, p. 41) presents an analysis of a similar amphibole from a hypabyssal albite-amphibole-epidote rock, not far from the area now under consideration, and he states that the amphibole is actinolitic hornblende. It is obvious that most of the amphiboles in the hypabyssal intrusives are uralitic.

Epidote is clinozoisite-rich and it occurs as well-shaped crystals or as grain aggregates especially in the shear zones.

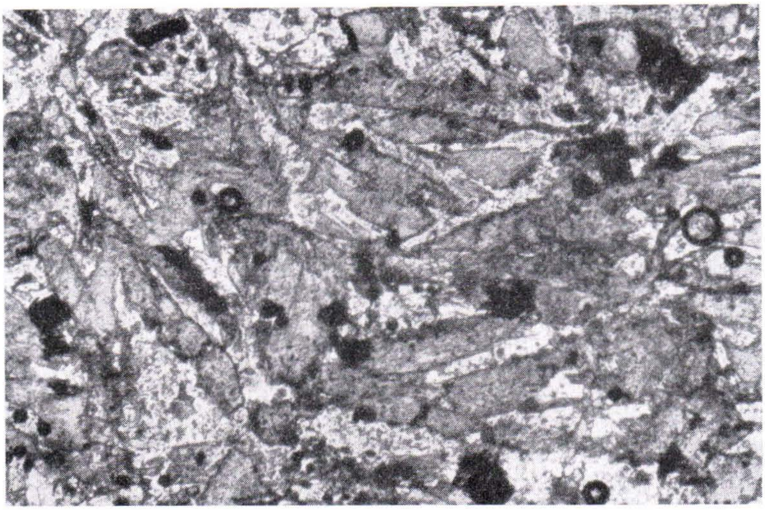

FIG. 19. Blasto-ophitic albite-amphibole diabase. Thin section, without analyzator. 30 x.

Chlorite is weakly pleochroic: $Z$ = pale green, $Y$ = colourless. It occurs as an alteration product of amphibole or as crack filling mass in the shear zones.

Leucoxene is very dark, near lyopaque, and occurs as wedge -shaped grains with small magnetite inclusions.

The microscopic structure of the rock is hypidiomorphic, although uralitization has damaged the primary form of pyroxene grains to some extent. Some of the rocks are blasto-ophitic (Fig. 19). There is no difference in the mineral compositions of blasto-ophitic and other hypidiomorphic albite-gabbros.

The amount of the hypabyssal intrusives characterized by sheet silicates is small. They have been found at two places only: Nilimaa and Ketunliesukuusikko. At both places the rock outcrops as one separate exposure, on account of which the extent of the rock and its relation to the surroundings remained open.

The rock of Ketunliesukuusikko is schistose albite-biotite-chlorite rock resembling veined gneiss. In addition to albite and chlorite veins there are veins of hematite, a few mm. thick, and lenticular accumulations of hematite. The microscopic structure of the rock is obscure and heterogeneous giving some impression of cataclase.

Plagioclase is turbid albite, $An_{3-4}$. The primary albite laths are granulated more or less completely. A conspicuous feature is the existence of thin parallel laths of quartz in the albite.

Biotite is strongly pleochroic: $Z = Y$ = dark brown $X$ = light brown. It exists as quite large piles or as small-flaked veins together with hematite.

Chlorite is distinctly pleochroic: $Z = Y$ = dark green $X$ = pale green, $X \perp (001)$. The highest interference colour is anomalous ultra blue of the first order, and $-2V = 8°—10°$. Chlorite occurs as large, Ø 5—20 mm, nests or as crack filling material.


FIG. 20. Albite-chlorite-carbonate rock, brecciated by chlorite. Photo M. Kortesluoma.

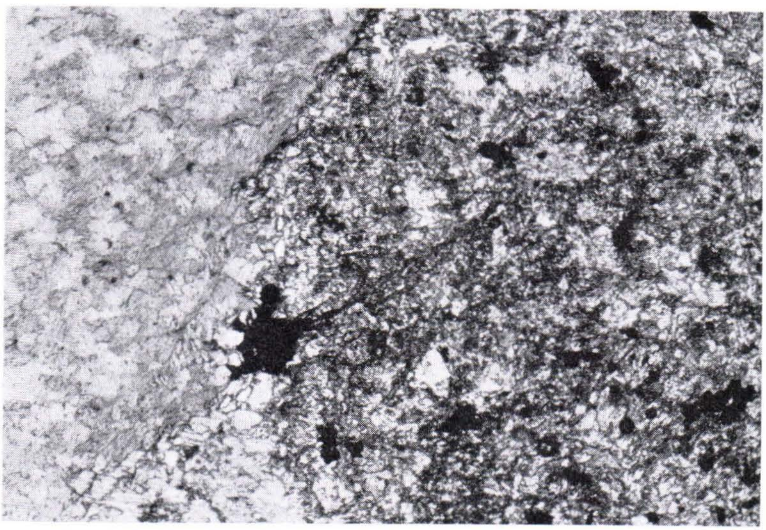

FIG. 21. Albite-chlorite-magnetite rock (right) brecciated by chlorite (left). Thin section from the rock shown in Fig. 20. Without analyzator. 30 x.

Hematite forms lenticular, small-flaked aggregates or thin veins together with biotite.

At Nilimaa, NW from Kuolajärvi, chlorite brecciates a pilotaxitic albite-chlorite-carbonate rock (Fig. 20). Optical properties of the chlorite are similar to those of the chlorite at Ketunliesukuusikko. The brecciated rock contains albite, $An_{3-4}$, almost colourless chlorite, carbonate and magnetite. The microscopic texture of the rock is very typical of the lava-born extrusives (Fig. 21).

Three kilometres southwest from the SW corner of the mapped area, at Mustavaara, there is an albite-amphibole gabbro that differs distinctly in external appearance from the rocks described above in this chapter. This gabbro is medium-grained, homogeneous and unfoliated. The colour of the rock is very dark, almost black.

The microscopic structure is hypidiomorphic. The main minerals are plagioclase, hornblende and epidote; accessories are chlorite, magnetite, apatite and sphene.

Plagioclase, $An_{2-4}$, is very bright, despite numerous inclusions of small-grained amphibole and epidote.

Amphibole is distinctly but not strongly pleochroic: $Z$ = blue green, $Y$ = brownish green, $X$ = almost colourless, $c \Lambda Z = 16°—19°$, $—2V = 76°—84°$. It occurs as elongated, partly ragged lamellae, that are commonly twinned according to the augite law.

Chlorite occurs as small-flaked aggregates, and its optical properties do not differ from those of the chlorites found in other albite gabbros.

Among the outcrops of Mustavaara there are some outcrops of hypidiomorphic albite-biotite rock, too. Relations of this rock to the albite-amphibole gabbro remained indistinct owing to the lack of favourable outcrops.

The microscopic structure of the rock is hypidiomorphic. The mineral content of the rock is presented in Table 4.

Plagioclase, $An_{4-6}$, occurs as tabular grains, twinned according to the albite law. The grains are turbid because of small chlorite inclusions.

Biotite occurs as thin rows of flakes between the albite grains forming beautiful graphic intergrowths with magnetite. The optical properties of biotite are the common ones.

The rock gives some impression of cataclase as a whole.

## Albitites

True albitites have not been found in the mapped area, but only in connection with the gabbro of Mustavaara. At Mustavaara, among the outcrops of albite-amphibole and albite-biotite rocks there are a few outcrops of pale brown albite-quartz-carbonate rock that very much resembles clastic sediment in external appearance.

The microscopic structure of the rock is granular-hypidiomorphic (Fig. 22). Plagioclase and quartz have crystallized partly simultaneously and they form obscure graphic intergrowths.

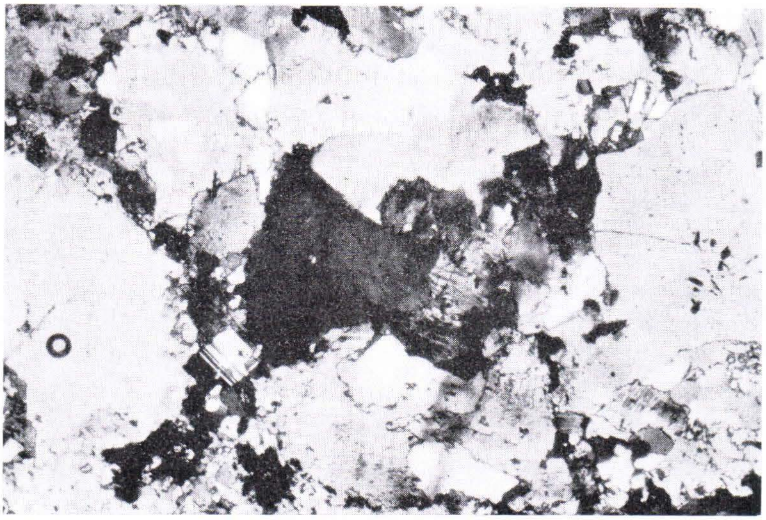

FIG. 22. Granular-hypidiomorphic albitite. Thin section, without analyzator. 30 x.

## Pyroclastic sediments

Pyroclastic sediments form approximately 30 per cent of the rocks on the mapped area.

### Tuffaceous schists

Most of the pyroclastic rocks are fine-grained, bedded and sometimes graded tuffaceous schists. The grain size of these rocks is so small that indentification of minerals, even with the help of a microscope, is a very hard and often completely hopeless task. Tuffaceous schists contain in various proportions chlorite, albite, quartz, carbonate minerals, mica minerals and opaque ore minerals. The amount of micas is very low. The boundaries between the beds are either sharp or diffuse. Intercalated jaspilites are met at some places (Fig. 23).

The fine-grained tuffaceous schists very much resemble phyllites in their external appearance, due to the richness of chlorite. That is why Hackmann (1927) and Mikkola (1941) call them phyllites and present a description of the phyllite from Kuolajärvi as an example. It is possible that a part of the fine-grained tuffaceous schists has the character of flysch sediments, i. e, they are weathering products of volcanics, but true aluminium-rich clays they have never been, as can easily be seen from Table 8, Anal. 1—2 p. 62.

The amount of coarse-grained varieties is not large. The microscopic structure of them is clastic, and bedding is distinct (Fig. 24). They resemble greywackes, and it is not always possible to decide whether they are pyroclastic or true clastic sediment.

FIG. 23. Intercalated beds of jaspilite and graded tuffaceous schist at the southwestern slope of Kuolavaara.

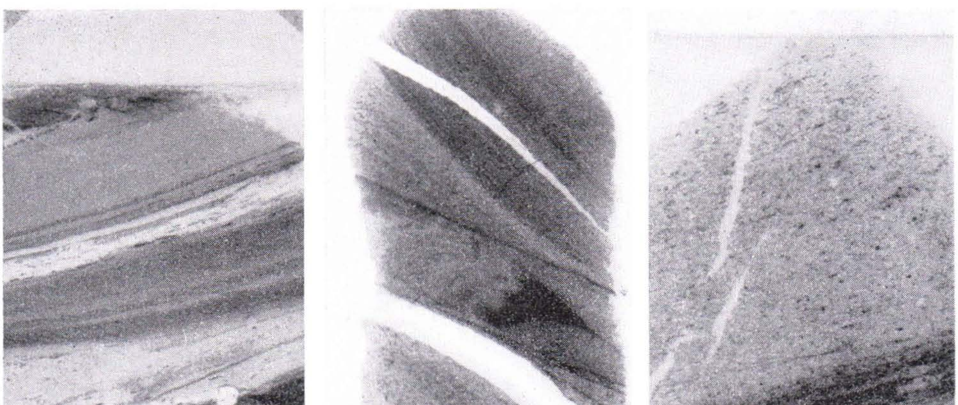

FIG. 24. Bedded tuffaceous schists with various grain size. Thin sections, without microscope. 3.3 x.

### Graphite-bearing schists

Many of the fine-grained tuffaceous shists are graphite-bearing. Graphite-bearing beds intercalate with graphite-lacking ones, and the silicate minerals are identical in both cases. Hackmann (1927) and Mikkola (1941) called them black schists, but both field observations and microscopic studies support the contention of the present author that they could not be of sapropelic origin. At some places there are mangano-siderite and jaspilite beds between the graphite bearing beds, and even parts of these mangano-siderite and jaspilite beds are graphite-bearing. This phenomenon is the most serious argument against the sapropelic origin of graphite. The origin of graphite will be discussed later.

Besides graphite these fine-grained pyroclasts may contain considerable amounts of iron sulphides, thus changing to sulphide schists which are described later in connection with chemical sediments.

### Volcanic breccias and agglomerates

Volcanic breccias, including breccias of an eruptive character are found at many places at Petäjäselkä, Vesmaselkä, Pahtavaara, Pitslomakuru and Selkämäntypää.

In the breccias of an eruptive character the matrix is some kind of lava-born rock, in most cases albite-chlorite or albite-magnetite rock. The brecciated material can be any of the lava-born rocks, pyroclastic rocks or sedimentary rocks described above. These breccias are only seldom polymictic.

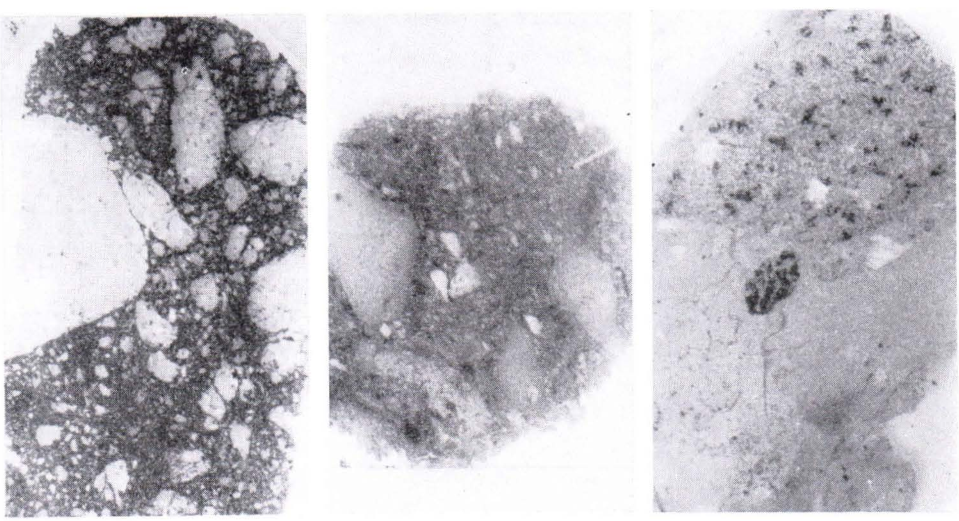

FIG. 25. Different kinds of volcanic breccias. Thin sections, without microscope. 3.3 x.

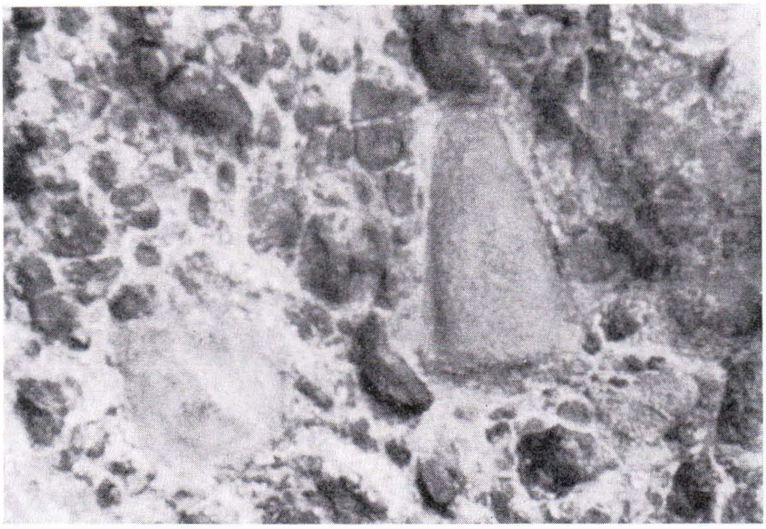

FIG. 26. Agglomeratic amphibole-clorite rock at Sakattipahta on the eastern bank of Kitinen River in the Sodankylä commune. About 1/10 of natural size.

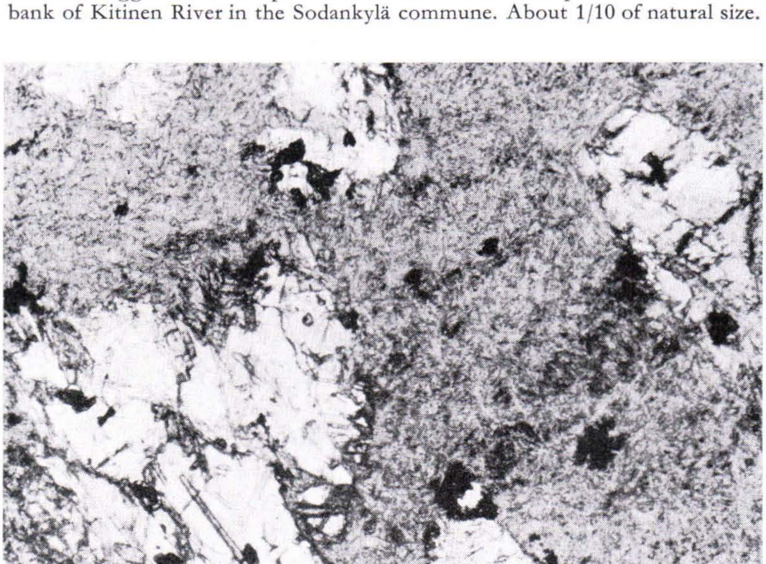

FIG. 27. Structure of agglomeratic amphibole-chlorite rock rich in amygdules filled by carbonate and amphibole lamellae. Thin section from the smaller one of the two large light pebbles shown in Fig. 26. Without analyzator. 30 x.

Most of the true volcanic breccias are polymictic containing fragments of all the kinds of rocks met in the complex. The size of the fragments varies from a few millimetres to several centimetres. Fragments are angular and sharp edged, but some rounded forms may occur, thus giving the rock a conglomeratic appearance.

The matrix consists of chlorite, albite and quartz. The amount of carbonate minerals in the matrix varies greatly. Accessories are sphene, dusty magnetite and epidote.

Examples of the appearance of the volcanic breccias are shown in Fig. 25.

Of special interest in the theoretical sense are the ultrabasic agglomerates described by Mikkola (1941) found at Sodankylä, 40—50 kilometers east from the mapped area. The best outcrop of these agglomerates is located on the eastern bank of the river Kitinen at a place called Sakattipahta. The Sakattipahta agglomerate consists mainly of fragments of dark amphibole rock ranging from a few centimetres to several decimetres in diameter. The rock is almost monomictic, only two pebbles with deviating appearance are found (Fig. 26). The two lighter pebbles shown in the picture differ from others only in containing more numerous and larger amygdules filled by carbonate (Fig. 27).

## Chemical sediments

It is not always possible to draw any sharp boundary between chemical and pyroclastic sediments. Pure variants of both groups are met, but at many places there are rocks containing both chemically precipitated and pyroclastic material.

On the basis of the chemical composition these chemical sediments can be divided as follows:

— jaspilites and associated oxide iron ores,
— mangano-siderite schists and other carbonaceous schists,
— sulphide schists

### Jaspilites

Jaspilites are iron-poor or iron-rich. Iron-poor jaspilites are either banded or brecciated. The main mineral of iron-poor jaspilites is fine-grained, granoblastic quartz (Fig. 28). The accessories are unevenly distributed iron oxide dust, euhedral magnetite, rhombohedral siderite altered partly or totally to goethite, and microlitic apatite prisms. Graphite is found occasionally. The colour and structure of jaspilites varies greatly. Iron-free jaspilites are pale, often almost white. Owing to the amount and kind of the acessory minerals the colour of jaspilites may be yellowish (goethite), reddish (hematite), dark grey or black (magnetite and graphite).

Different kinds of breccias are the commonest structure types found in jaspilites. Most usual is a breccia that consists of angular, white fragments, cemented together with black jaspilite. At some places fragments have a subparallel orientation (Fig. 29). The proportion of fragments and matrix varies. The volume of light fragments is generally larger than that of the dark matrix, but opposite proportions are met.

Besides the brecciated jaspilites there are some banded types. In the banded jaspilites banding is due to the alternating dark and light bands from a few millimetres

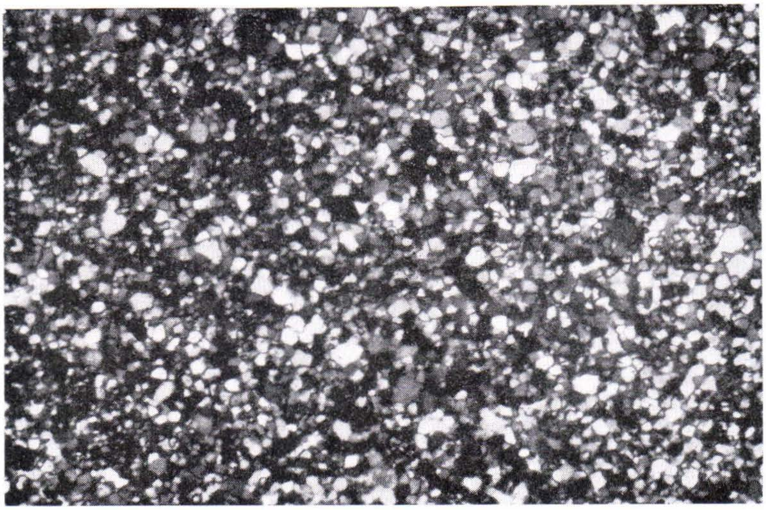

FIG. 28. Granoblastic jaspilite. Thin section, with analyzator. 30 x.


FIG. 29. Jaspilite breccia, consisting of light fragments with parallel orientation and dark matrix penetrated by secondary quartz veins. Photo M. Kortesluoma.

to severals centimetres thick. Under the microscope the dark bands are exactly similar to the dark matrix of the brecciated jaspilites and analogously the light bands are similar to the light fragments of the breccias. The banding is obviosly a primary sedimentary feature. Such an interpretation is supported by the colloform structures consisting of rhythmically alternating dark and light bands (Figs. 30 and 31). A logical corollary of this is that the breccias have originally been banded sediments. The brecciation has been a prefolding phenomenon. This is supported by some interesting

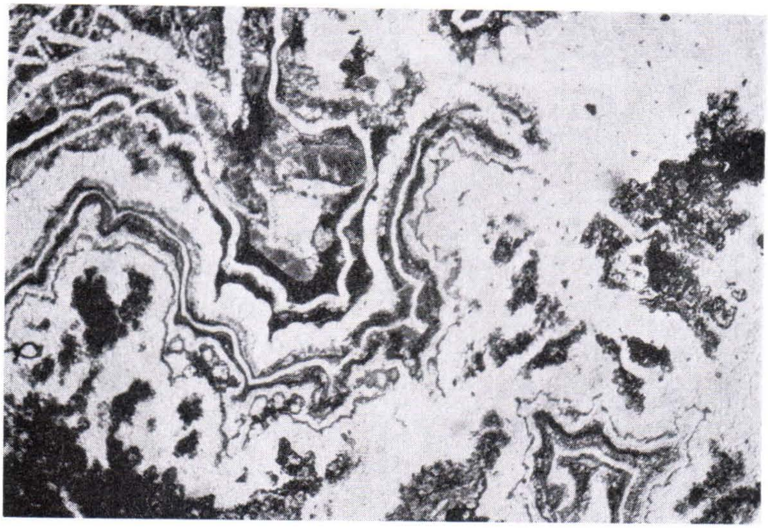

Fig. 30. Colloform structure in a partly recrystallized jaspilite. Thin section, without analyzator. 30 x

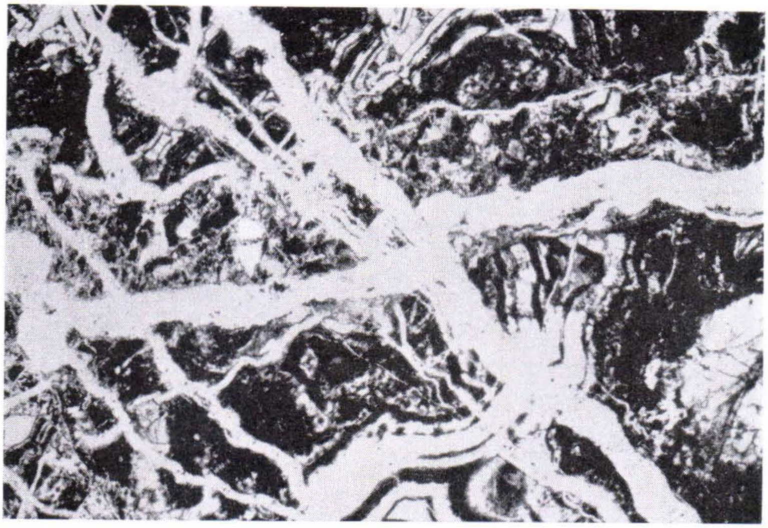

Fig. 31. Colloform dark jaspilite, partly deformed mechanically and penetrated by secondary quartz veins. Thin section, without analyzator. 30 x.

structures met at Kuolavaara. In the western part of Kuolavaara there is a low outcrop of finely banded jaspilite. At one place the structure of the rock strikingly resembles current bedding at first sight. But closer inspection reveals that the structure is not due to the currents but to the gravity gliding and slumping of the beds, and it is probable that this gravity gliding even caused the brecciation (see, Swarbrick, 1967). At another place at Kuolavaara the jaspilite beds were rolled during gliding thus

FIG. 32. Cross section of a jaspilite tube due to gravity gliding and rolling in a pre-consolidation stage. Photo M. Kortesluoma.

forming tube-like structures (Fig. 32). It must be emphasized that taking into account the nature of the deformation and the movements connected with the regional folding, these slumping and rolling structures cannot be of tectonical origin. As will be stated later the movements connected with folding have been differential. Plastic deformation due to recrystallization is lacking. Cleavage in the rock, shown in Fig. 32, is demonstrated by narrow, subparallel quartz-filled cracks. It is clear that all the breccias are not caused by gravity gliding only. Part of them are due to other movements in the earth's crust connected with volcanic activity and even with folding.

S e r i c i t e j a s p i l i t e s, found for instance at Kuolavaara and near the dam on the river Karjakkojoki, differ distinctly from the jaspilites described above. The weathering crust of sericite jaspilite at Karjakkojoki is yellowish and partly rusty due to the pyrite content of the rock. The freshly cut surface is light grey, very much resembling true flint. The almost horizontal bedding is obscure and it is cut by vertical cleavage.

Under the microscope bedding and transversal cleavage are clearly visible. The main constituent of the rock is very fine grained quartz. The average diameter of the quartz grains is less than 0.01 mm. The size of sericite flakes is a little bit larger. Sericite flakes have gathered to lenticular aggregates parallel to the cleavage. In the same manner pyrite, together with bright large quartz grains and biotite flakes, forms lenticular nests, parallel to the cleavage. The basal planes of biotite are regularly perpendicular to the cleavage.

Due to the small grain size of the rock it was impossible to determine the quantitative mineral content of the rock. The chemical composition of the rock is presented in Table 8, Anal. 12 p. 63.

## Oxide iron ores

At some places the iron content of jaspilite rises so high that it can be graded as iron ore. These oxide iron ores are of both banded and brecciated type.

In the banded ores quartz bands and ore bands alternate with either sharp or diffuse boundaries. The thickness of the ore bands is generally 1—2 cm, although

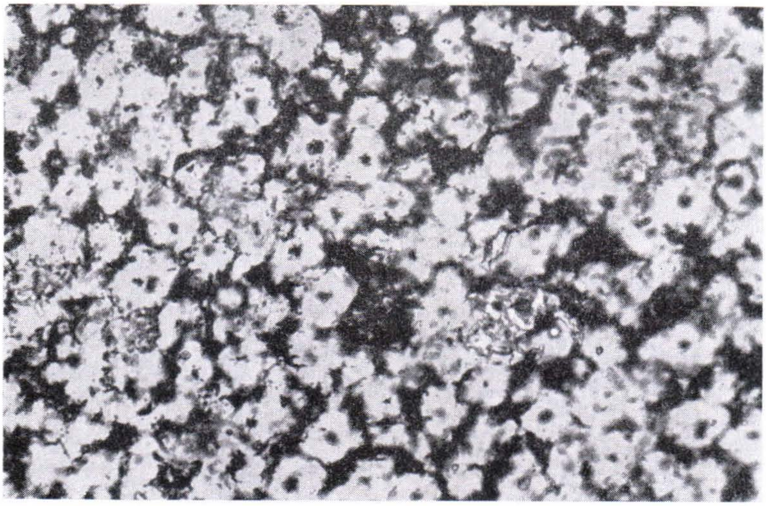

FIG. 33. Nodular structure in a quartz band. White is quartz, black is magnetite. Thin section, without analyzator. 30 x.

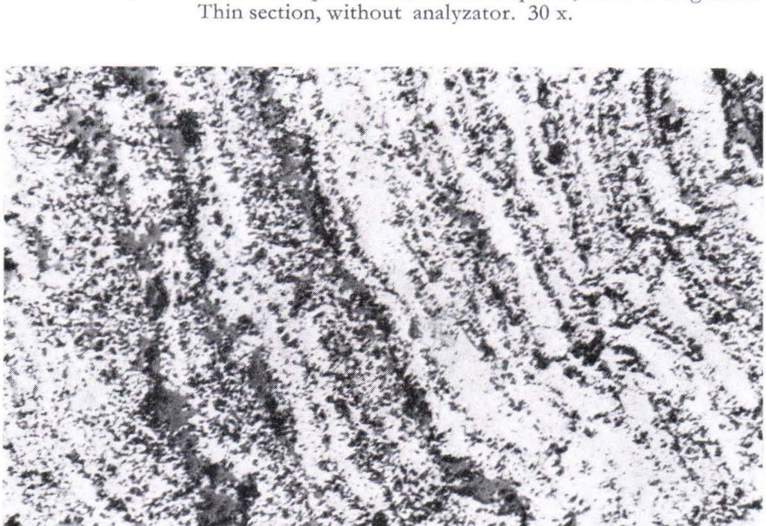

FIG. 34. Microbanded magnetite (white) with minor intercalated quartz (grey) bands. Polished section. 90 x.

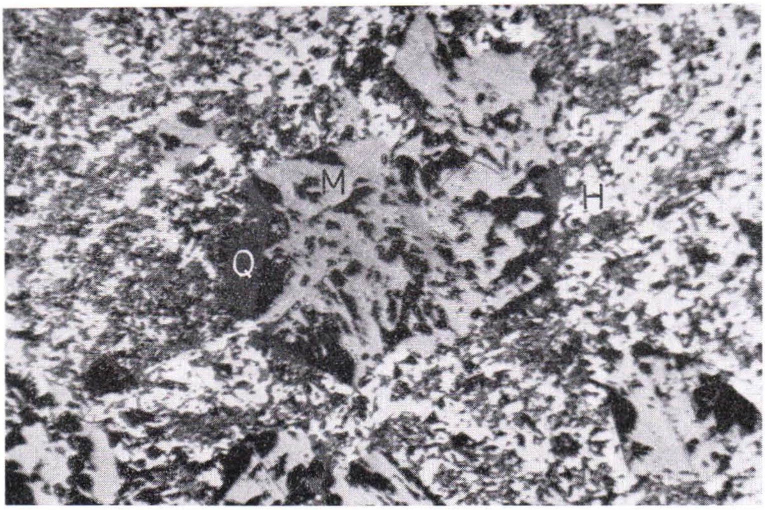


FIG. 35. Euhedral magnetite (M) crystals surrounded by coarse-grained quartz (Q) in a fine-grained hematite (H) band. Polished section. 90 x.

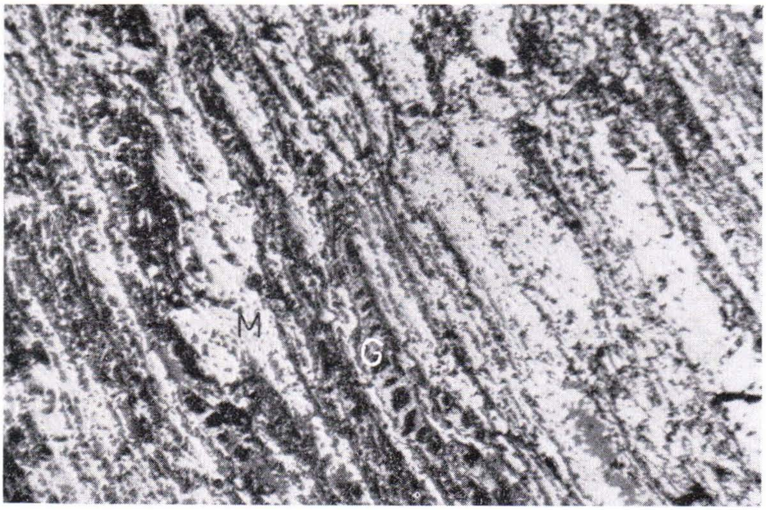


FIG. 36. Thin goethite (G) bands intercalated with magnetite (M) bands. Polished section. 90 x.

ore bands up to 10 cm thick are occasionally met. The thickness of the iron-bearing strata varies from a few metres to several dozen metres. Exceptionally the thickness may rise to one hundred metres. There is no homogeneous ore stratum of any great extent, but the ores occur as hundreds of restricted bodies. The length of the ore bodies is at least four or five times the thickness.

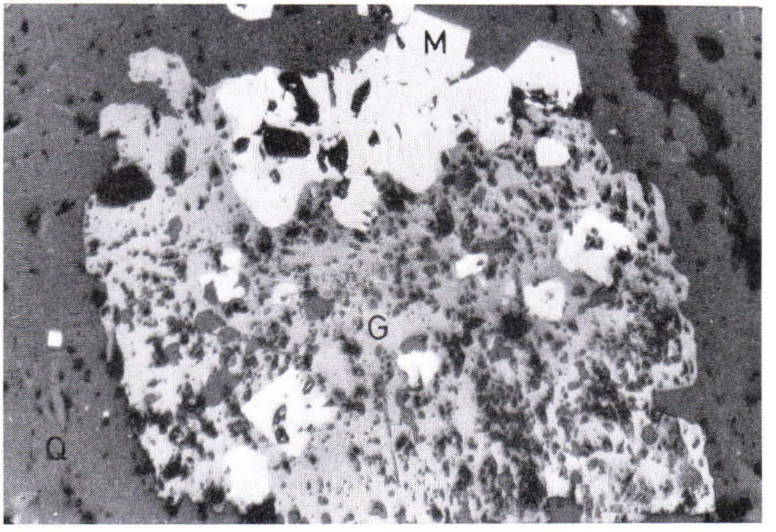


FIG. 37. A lenticular goethite (G) nest, with euhedral magnetite (M) crystals, in quartz (Q) matrix. Polished section. 30 x.

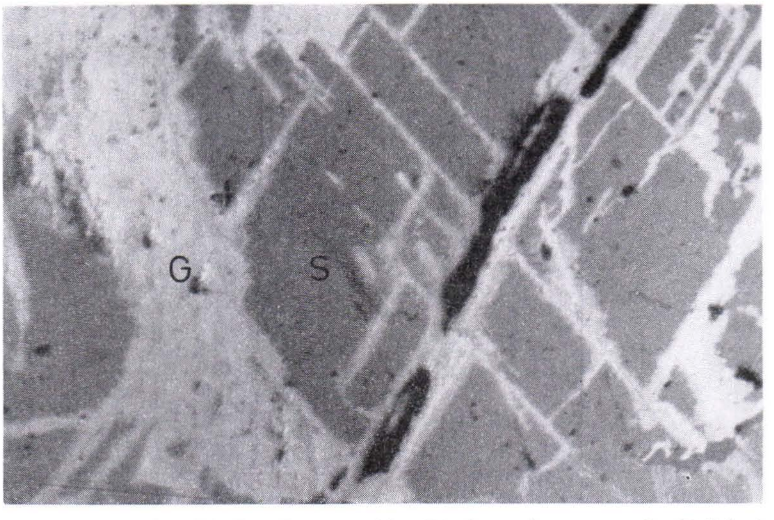


FIG. 38. Siderite (S) altered to goethite (G) along cleavage cracks. Polished section. 300 x.

The intercalated quartz bands are generally thicker than the ore bands thus causing the low grade of the ores. The quartz bands are generally similar to the iron-poor jaspilites, but at some places nodular structures (Fig. 33) in quartz bands are found.

The ore bands consist mainly of magnetite, which can be either coarse-grained and compact or fine-grained and microbanded (Fig. 34). It is very common for compact magnetite bands to send apophyses to the iron poor bands.

FIG. 39. Intercalated bands of jaspilite, siderite, magnetite, goethite, and hematite at the SW end of Haurespää. Photo M. Kortesluoma.

Hematite is met occasionally but it is of no quantitative significance. Hematite bands often contain idiomorphic magnetite crystals surrounded by clean coarse-grained quartz (Fig. 35). Magnetite may exist even as compact bands in fine grained hematite-quartz bands.

More common than hematite is goethite. It exists in several manners. Very common are thin goethite bands among magnetite bands (Fig. 36). Goethite may form lenticular nests with idiomorphic magnetite grains, too (Fig. 37), or it occurs as the alteration product of siderite. The alteration of siderite begins from the cleavage cracks and can lead to complete disappearance of siderite (Fig. 38).

Though siderite belongs essentially to carbonate facies it is not infrequently found in oxide facies as well. Well-shaped siderite rhombohedra and siderite-quartz veins, both of secondary origin, are very common also in oxide iron ores. At some places, e.g. on the southern slope of Haurespää, siderite exists as thin bands intercalating with magnetite, goethite, hematite and quartz (Figs. 39 and 40).

The estimates on the tonnage of the oxide iron ores of the Porkonen-Pahtavaara area vary, and no reliable inventory, based on diamond drilling, has been made. Even the information about the quality of the ores in a commercial sense is insufficient. The available analyses show that the quality of the ores is not very high (Table 5, p. 44).

Not all the minerals belonging to the iron formation have yet been identified with absolute certainty. By means of an X-ray powder camera Partio (1966) identified

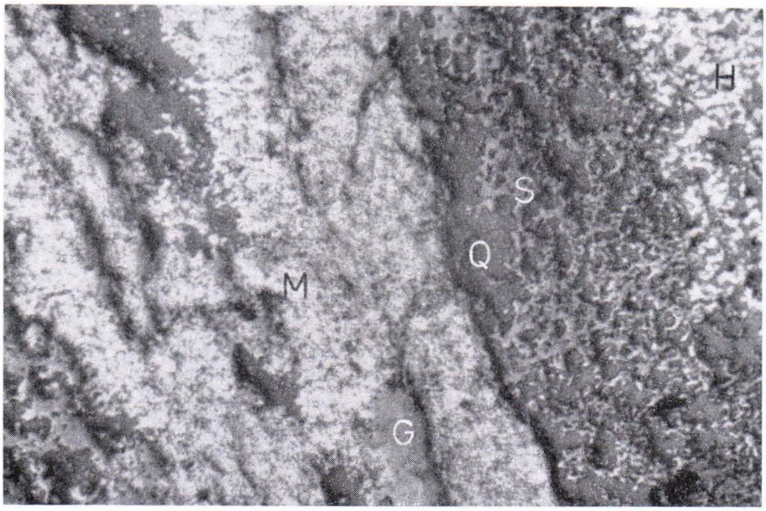


FIG. 40. Intercalated bands of quartz (Q), magnetite (M), goethite (G), siderite (S) and hematite (H). Polished section from the sample shown in Fig. 39. 30 x.

TABLE 5

Analyses of oxide iron ores from the Porkonen-Pahtavaara area.

35 samples from different parts of the area, analyst, O. N. Heidenreich (Välitalo, 1967, p. 50).

| | Haurespää | Porkonen | Pahtavaara | Kuoreslaki |
|---|---|---|---|---|
| Fe | 29.8 —52.6 | 40.1 —65.7 | 41.0 —61.7 | 21.8 —49.3 |
| P | 0.17— 1.78 | 0.79— 0.85 | 0.21— 0.79 | 0.13— 0.89 |
| S | 0.02— 0.09 | 0.02— 0.08 | 0.03— 0.04 | 0.04— 0.10 |
| Mn | 0.09— 0.58 | 0.17— 0.38 | 0.18— 0.20 | 0.21— 0.51 |
| insoluble | 23.3 —55.9 | 21.5 —41.9 | 23.8 —39.8 | 28.7 —47.1 |

7 samples from Porkonen and Pahtavaara, analyst, E. Ingman (Hackman, 1925, p. 9).

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| Fe | 36.7 | 42.4 | 41.5 | 47.0 | 34.8 | 43.9 | 42.4 |
| P | 0.47 | 0.15 | 0.48 | 0.26 | 0.25 | 0.24 | 0.25 |
| S | 0.70 | 0.04 | 0.04 | 0.04 | 0.03 | 0.04 | 0.06 |

8 samples from Pahtavaara, analyzed in the Chemical Laboratory of Rautaruukki Oy.

| | | | | | | | | |
|---|---|---|---|---|---|---|---|---|
| Fe | 38 | 30 | 14 | 38 | 15 | 32 | 27 | 24 |
| Mn | 0.3 | 0.5 | 0.2 | 0.1 | 0.1 | 1.2 | 0.3 | 0.6 |

minnesotaite from Pahtavaara and he suggested that the barrandite mentioned by Mikkola (1941, p. 175) was in fact this mineral. The present author, however, found from Pitslomakuru an acicular, brownish mineral with a marked tendency to form »garben» (Fig. 41). X-ray studies of this mineral revealed it to belong to the variscite — strengite series. The existence of greenalite is likewise supported by X-ray studies.

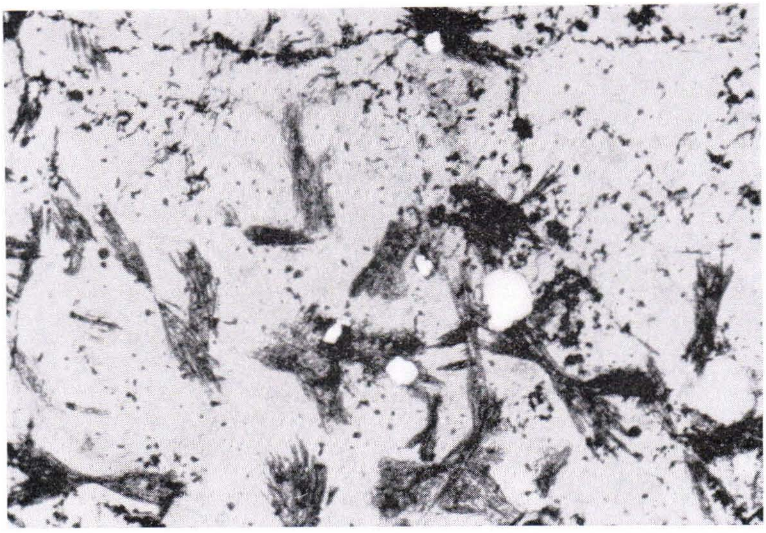

FIG. 41. Barrandide »garbens» in jaspilite. Thin section, without analyzator. 30 x.

A special treatise on the minerals of the iron formation in the Porkonen-Pahtavaara area is under preparation by the author and many of the mineralogical notes of the present paper are to be considered as preliminary ones only.

## Mangano-siderite schists

Besides the oxide iron ores mangano-siderite schists are the main components of the iron formation in the Porkonen-Pahtavaara area. Mangano-siderite schists, varying in quality and quantity, are found together with tuffaceous schists and/or jaspilites at Petäjäselkä, Pitslomakuru, Silmänpaistama, Kuolavaara, Keulakkopää, Karjalehto, Säynäjäjärvi and Riesiövaara. The two last mentioned places are outside the area presented on the map.

FIG. 42. Jaspilite fragments cemented together with mangano-siderite and graphite-bearing tuffaceous schist. Photo M. Kortesluoma.

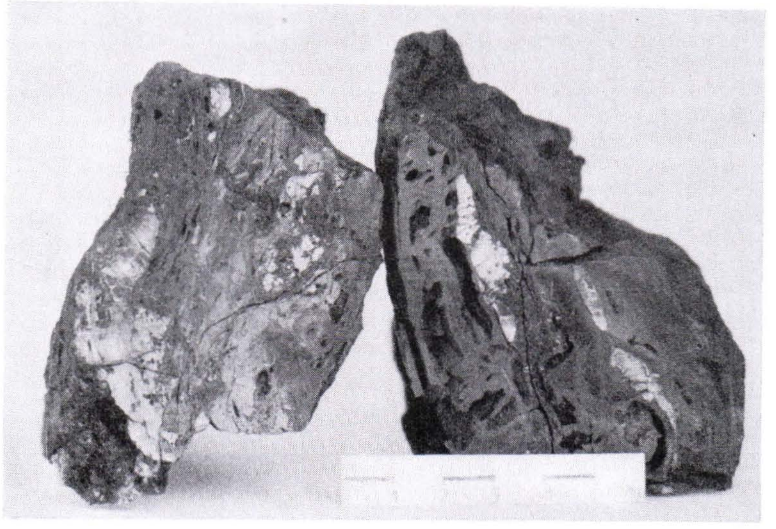

FIG. 43. Partly weathered breccias consisting of jaspilite, mangano-siderite and graphite-bearing tuffaceous schist. Photo M. Kortesluoma.

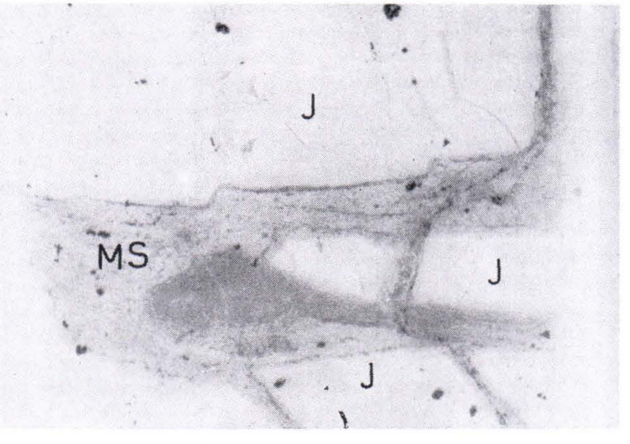


FIG. 44. Jaspilite (J) fragments cemented together with mangano-siderite (MS). Thin section, without microscope. 3.3 x.

The colour of the weathering crust varies from dark brown to black depending on the manganese content of the schist. The manganese rich- types become black and the manganese-poor siderites develop a reddish brown weathering crust. The thickness of the weathering crust varies from 0.5 cm to 5 cm. The freshly cut surface ranges from light grey to almost black depending on the mineral composition of the schist.

Mangano-siderite schists only seldom form independent larger units. Usually they intercalate as beds a few centimetres thick with tuffaceous schists and/or jaspilites.

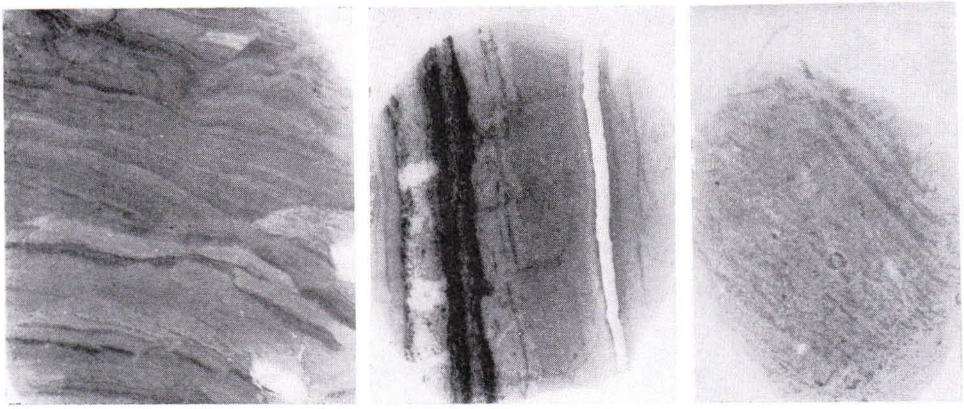

FIG. 45. Bedded mangano-siderite schists. Thin sections, without microscope. 3.3 x. Photo M. Kortesluoma.

In connection with movements the rigid jaspilite has splittered and resulted in the formation of tangled breccias (Figs. 42—44). In these schists and even in the breccias mangano-siderite shows distinct banding (Fig. 45) and the grain size is very small.

The main minerals of these schists are mangano-siderite, quartz, chlorite, mica minerals, albite and magnetite in varying proportions. Graphite is found at some places.

Mangano-siderite exists in a twofold manner: as a fine-grained material or as large, well-shaped porphyroblastic rhombohedra. It is very common to find bands of pure mangano-siderite a few millimetres thick between silicate and quartz bands. Oolitic structures are met occasionally in these bands (Fig. 46). As already stated mangano-siderite also occurs as narrow veins together with quartz.

Two analyses of mangano-siderite from separated samples were made during the early stages of prospecting activity (Table 6). As can easily be seen from the table, both of the samples remained impure due to the small grain size of the separated rock.

Under the microscope mangano-siderite is pleochroic: E = colourless, O = brown. The refractive indices are: $\varepsilon = 1.617 + 0.002$, $\omega = 1.840 \pm 0.002$. Reflectance is very high, hence mangano-siderite can easily be distinguished from other carbonate minerals in thin sections.

A chlorite-like mineral exists mainly between the mangano-siderite bands or around fragments in breccias. The mineral is generally very fine-flaked and quite strongly pleochroic: $X$ = light green, $Y = Z$ = yellowish brown. Elongation is positive and the extinction angle is 1°—3°. Refractive indices are: $X = 1.646 \pm 0.002$, $Z = 1.650 \pm 0.002$; —2V is very small. Optical properties and X-ray powder camera determinations support the belief that the mineral is greenalite.

The occurrence of mica minerals shows the largest variation in the mineral composition of mangano-siderite schists. At some places they are totally lacking but at

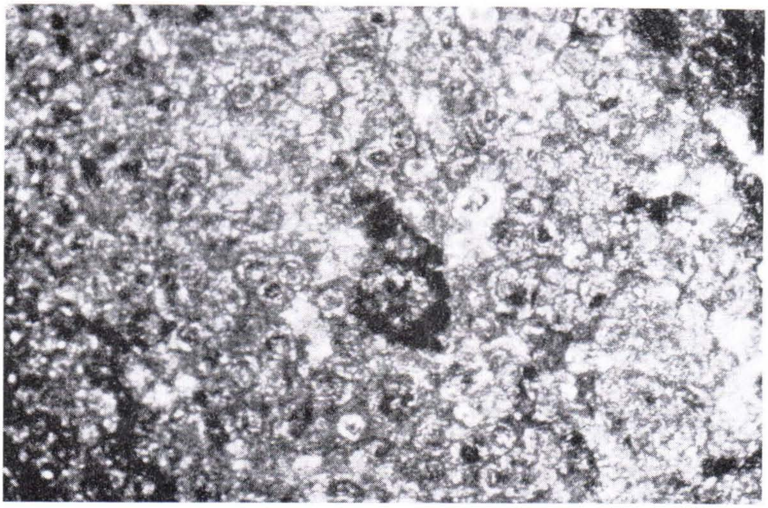

FIG. 46. Obscure oolitic structures in mangano-siderite. The specimen is the same as in the middle of Fig. 45. Thin section, without analyzator. 200 x.

TABLE 6

Chemical composition of mangano- siderite from Petäjäselkä, Kittilä.

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 2.80 | n.d. |
| $TiO_2$ | 0.11 | » |
| $Al_2O_3$ | 0.50 | » |
| $Fe_2O_3$ | 1.90 | » |
| FeO | 39.42 | 33.57 |
| MnO | 13.23 | 15.15 |
| MgO | 3.35 | 2.84 |
| CaO | 1.42 | 2.18 |
| $Na_2O$ | 0.16 | n.d. |
| $K_2O$ | 0.09 | » |
| BaO | n.d | 0.01 |
| $P_2O_5$ | 0.13 | n.d. |
| $CO_2$ | 37.19 | 31.76 |
| $H_2O^+$ | 0.04 | n.d. |
| $H_2O^-$ | 0.02 | » |
| Insoluble | — | 10.69 |

1. Analyst, Pentti Ojanperä, (Hytönen *et al.*, 1966).
2. Analyst, Aulis Heikkinen, (Silvennoinen, 1966).

other places they may exist in large amounts. Micas are generally small-flaked, but sometimes colourless mica, either muscovite or paragonite, forms large porphyroblasts reaching a diameter of 0.5 mm. Biotite cannot always be distinguished optically from stilpnomelane.

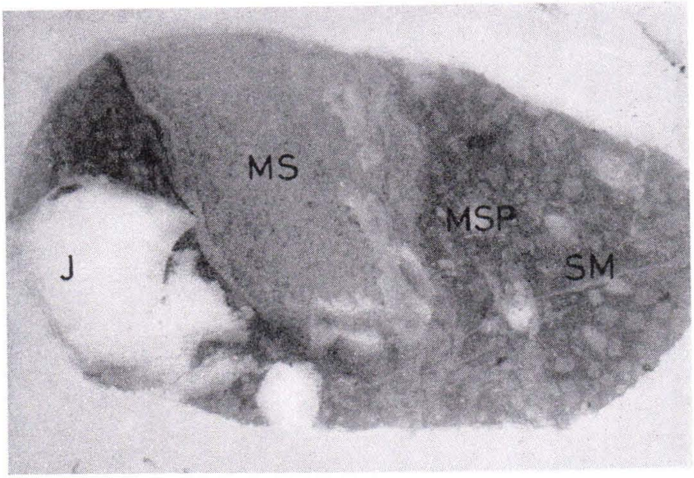


FIG. 47. Jaspilite (J) and mangano-siderite (MS) fragments cemented together with stilpnomelane mass (M) rich in mangano-siderite porphyroblasts (MSP). Thin section, without microscope. 3.3 x. Photo M. Kortesluoma.

Stilpnomelane, identified by the X-ray powder camera method, may at some places form the main constituent of the mangano-siderite schist. It exists as a very fine-flaked mass with a lot of manganosiderite porphyroblasts (Fig. 47). The stilpnomelane mass is cut by cracks filled with colourless mica. The mangano-siderite porphyroblasts have grown through and over these cracks.

### Other carbonaceous schists

Especially at Petäjäselkä the mangano-siderite schists are intimately associated with schists rich in dolomite. These dolomite-bearing schists show obscure bedding and they consist of small-grained dolomite, colourless mica, quartz and albite. The amount of dolomite does not generally exceed 50 per cent of the volume. Opaque iron oxide and occasionally even iron sulphide occur as accessories.

Dolomite from Pitslomakuru differs distinctly from the dolomites of Petäjäselkä. The rock is yellowish grey and shows indistinct bedding. The bedded, fine-grained part of the rock is brecciated by remobilized coarse dolomite. Chemical analyses of the rock are presented in Table 8, p. 62. Other minerals are very fine-grained quartz and small albite laths. Ore dust exists as an accessory.

Calcitic limestone is found at the dam on the river Karjakkojoki and as local boulders about two kilometers down stream from the dam. The structure and the external appearance of the rock very much resembles that of the dolomite from Pitslomakuru, but the rock can be distinguished from dolomite by the colour of the weathering crust which is grey in limestone and reddish in dolomite. The mineral is confirmed to be calcite by X-ray and immersion methods.

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The rock is brecciated. Remobilized coarse-grained calcite together with quartz brecciates fine-grained bedded limestone. Besides calcite the fine-grained part contains quartz and iron oxide dust.

### Sulphide schists

At some places there are sulphide schists alternating with graphite-bearing schists and it is not possible to draw any sharp border between these rocks. On the other hand, alternation with mangano-siderite schists is just as common. The amount of sulphide schists is not large. The best known occurrence is on the eastern flank of Pahtavaara, where it has been verified by diamond drilling. Sulphide schist outcrops are also met on the southern slope of Kuolavaara.

Sulphide schists are similar to the graphite-bearing tuffaceous schists. The essential difference is in the amount of sulphide minerals. The sulphides are almost entirely iron sulphides, mainly pyrrhotite and lesser pyrite. Black, sooty and magnetic melnikowite, identified by the X-ray powder camera method is found in samples from weathered parts of the sulphide schists. Small amounts of chalcopyrite, sporadically distributed, are found occasionally at some places.

## Clastic sediments

### Impure dolomites

Impure dolomite, rich in quartz and mica minerals, occurs west from Jänesvaara, near the shore of the river Karjakkojoki. The colour of the rock is grey and it shows distinct bedding, at some places even current bedding. Under the microscope the rock is surprisingly homogeneous. It consists of small-grained dolomite and very fine-grained quartz. Besides dolomite and quartz the rock contains various amounts of colourless mica and sporadically distributed biotite. Opaque ore minerals occur as accessories.

### Coarse grained quartzites

Coarse grained quartzites are found only at a few places in the mapped area. The most interesting of these occurrences is that by the Kutuoja creek, northeast of Lake Kuolajärvi, the only place where the contact between Lapponian and Kumpu-Oraniemi series is visible. The Lapponian volcanics of that occurrence are tuffaceous schists. The contact is almost vertical and strongly mylonitized and hence it is impossible to observe where the tuffaceous schist ends and the coarse grained quartzite begins. It is possible that the mylonite zone represents a weathered, undenudated crust of the Lapponian volcanics, thus giving evidence for supposed hiatus between Lapponian and Kumpu-Oraniemi series (see p. 77).

This mylonite breccia consists of colourless mica, very fine-grained quartz and albite. Accessories are sphene, chlorite and opaque ore minerals. Mylonite is penetrated by a few narrow siderite-quartz veins and pure albite veins. The mylonite zone is about 1.5 m thick; it changes quickly to coarser, less mylonitized, distinctly clastic sediment. The amount of colourless mica decreases and the amount of albite increases. By inference, some of the mica may be paragonite. The major part of the sediment is grey coloured and partly schistose consisting of clastic grains of quartz and albite cemented together with colourless mica.

Between Kuolavaara and Keulakkopää, some 250 m southwest from the junction of the road to Lomajärvi, there is a dark, clastic quartzite in a weathered outcrop. The relation of this quartzite to surrounding volcanics is unclear. The rock shows weak bedding and it consists of weakly or moderately rounded bright quartz grains with undulating extinction, and small irregular turbid albite grains. The matrix consists of dark or brown hydrous iron oxide and colourless mica.

Quite similar quartzite is found at Keulakkopää, but it is more schistose and contains fragments of jaspilites and tuffaceous schists besides quartz and albite grains.

The Värttiövaara quartzite is light coloured and coarser than the rocks described above. Contact with the volcanics is not visible, but it is possible to deduce its position to within an accuracy of a few metres both on the northern and western slopes of the hill. The quartzite forms a sheet a few dozen metres thick that dips 5°—10° to the south. There is no visible cleavage in the quartzite.

At Mantovaara, a couple of kilometres south from Värttiövaara, the quartzite takes on conglomeratic features. The pebbles of a sample, taken at random, are only slightly rounded. The rock of the pebbles is albite-paragonite rock, very rich in paragonite. Both the pebbles and the matrix, which consists of colourless mica, contain plenty of idiomorphic magnetite crystals and ragged tourmaline grains.

## PETROLOGICAL ASPECTS

This chapter deals with some petrological problems designed to elucidate the general view of the geological evolution of the area.

### The chemical composition and the spilitic nature of the greenstones

In his paper concerning the bedrock of Kittilä, Hackmann (1927) divides the greenstones of the area into two groups: metabasites and amphibolites. According to him metabasites consist of fine-grained or aphanitic, more or less schistose basic effusives which are basic plagioclase porphyrites or uralite porphyrites, fine-grained tuffites, volcanic breccias and agglomerates. Because of the variolitic structure and the peculiar metamorphism which is characterized by the alteration of pyroxenes to

TABLE

Chemical compositions and Niggli numbers

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 |
|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ ... | 36.25 | 43.83 | 41.19 | 42.67 | 45.26 | 46.24 | 47.07 | 49.11 | 52.89 |
| $TiO_2$ .. | 0.41 | 0.90 | 0.88 | 0.80 | 0.72 | 3.70 | 3.86 | 2.00 | 1.02 |
| $Al_2O_3$ .. | 0.54 | 7.13 | 6.59 | 7.10 | 8.08 | 12.44 | 13.55 | 14.11 | 12.99 |
| $Fe_2O_5$ .. | 3.19 | 4.24 | 4.00 | 3.80 | 1.27 | 5.49 | 3.59 | 2.32 | 0.63 |
| FeO ... | 4.43 | 10.76 | 6.05 | 8.24 | 14.84 | 11.33 | 8.93 | 10.15 | 8.12 |
| MnO .. | 0.18 | 0.17 | 0.25 | 0.11 | 0.31 | 0.18 | 0.16 | 0.24 | 0.19 |
| MgO .. | 30.75 | 18.33 | 26.55 | 24.47 | 14.42 | 5.60 | 8.53 | 6.96 | 10.48 |
| CaO ... | 1.80 | 9.26 | 6.04 | 6.55 | 8.54 | 8.12 | 6.56 | 9.10 | 7.10 |
| $Na_2O$ .. | 0.09 | 1.10 | 0.24 | 0.30 | 0.22 | 3.50 | 3.58 | 2.70 | 1.48 |
| $K_2O$ ... | 0.01 | 0.36 | 0.15 | 0.28 | 0.18 | 1.51 | 0.30 | 0.72 | 2.05 |
| $P_2O_5$ .. | 0.02 | — | — | 0.26 | 0.07 | 0.18 | 0.01 | 0.09 | 0.07 |
| $CO_2$ ... | 14.83 | — | 1.47 | — | 0.00 | — | — | — | 0.00 |
| $H_2O^+$ .. | 6.18 | 3.51 | 6.16 | 4.96 | 5.16 | 1.10 | 3.83 | 2.18 | 2.39 |
| $H_2O^-$ .. | 0.02 | 0.13 | 0.17 | 0.14 | 0.04 | 0.18 | 0.04 | 0.17 | 0.05 |
| | [1]) 98.70 | 99.72 | 99.74 | 99.68 | 99.11 | 99.57 | 100.01 | 99.85 | 99.46 |
| si ...... | 65 | 67 | 72 | 73 | 82 | 108 | 112 | 117 | 129 |
| al ..... | 1.6 | 6.5 | 6.9 | 7.2 | 17.1 | 17.2 | 19.1 | 19.7 | 19.0 |
| fm ..... | 92.7 | 61.2 | 82.8 | 79.8 | 65.2 | 52.0 | 55.2 | 49.6 | 55.0 |
| c ...... | 5.2 | 15.3 | 9.7 | 12.1 | 16.5 | 20.5 | 16.9 | 23.2 | 19.1 |
| alk .... | 0.5 | 17.0 | 0.6 | 0.9 | 1.2 | 10.3 | 8.8 | 7.5 | 7.1 |
| k ...... | 0.06 | 0.21 | 0.29 | 0.38 | 0.36 | 0.22 | 0.05 | 0.15 | 0.48 |
| mg .... | 0.84 | 0.69 | 0.82 | 0.79 | 0.60 | 0.38 | 0.55 | 0.50 | 0.69 |

[1]) In addition: Cr = 0.336, Ni = 0.396, Co = 0.0088, V = 0.0290, Cu = 0.0004.

1. Soapstone-like lava-born rock. Kuolavaara, Kittilä. Anal. P. Ojanperä.
2. Lava-born amphibole rock. Keikkuma-aavan saari (the »island» of the marsh of Keikkuma-aapa), NNW of Koitilainen fell, Sodankylä. Anal. H. Lönnroth. (Mikkola, 1941, p. 239).
3. Lava-born amphibole-chlorite rock. Sattasvaara hill, N of the river Sattasjoki, Sodankylä. Anal. H. Lönnroth (Rankama, 1939, p. 8).
4. Lava-born amphibole-chlorite rock, Kummitsoiva hill, Pelkosenniemi. Anal. H. Lönnroth (Mikkola, 1941, p. 238).
5. Lava-born amphibole-chlorite rock. Rovanpää, Kittilä. Anal. P. Ojanperä.
6. Hypabyssal albite diabase. 6 km N of the church, close to highway, Kittilä. Anal. L. Lokka. Hackman, 1927, p. 21).
7. Lava-born albite-amphibole-chlorite-epidote rock. Island of Lake Sotkajärvi, Kittilä. Anal. L. Lokka (Hackman, 1927, p. 18).
8. Hypabyssal albite diabase. Holkkuavaara, Kittilä. Anal. L. Lokka. (Hackman, 1927, p. 22).

amphiboles, chlorite and biotite, and by the albitization of basic plabioclase, he calls these basic effusives spilites which would not represent any independent suite of magma. The hypabyssal members of metabasites change at places to gabbroic or ultrabasic rocks.

7

of the greenstones of central Lapland.

| 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 |
|---|---|---|---|---|---|---|---|---|
| 49.95 | 53.36 | 51.21 | 50.58 | 51.07 | 53.31 | 49.19 | 60.54 | 49.86 |
| 3.05 | 1.20 | 1.91 | 2.10 | 2.15 | 1.20 | 2.03 | 0.45 | 2.48 |
| 11.69 | 13.15 | 12.85 | 12.89 | 12.79 | 16.39 | 12.43 | 11.85 | 13.15 |
| 1.82 | 1.20 | 3.53 | 2.61 | 0.85 | 2.74 | 1.43 | 0.10 | 2.55 |
| 12.49 | 9.00 | 8.48 | 10.56 | 9.84 | 11.19 | 10.29 | 0.81 | 10.58 |
| 0.19 | 0.17 | 0.17 | 0.32 | 0.18 | 0.21 | 0.14 | 0.05 | — |
| 6.25 | 8.19 | 6.30 | 6.80 | 7.66 | 3.11 | 3.27 | 3.82 | 7.05 |
| 8.06 | 7.62 | 6.72 | 7.18 | 6.44 | 1.04 | 7.35 | 5.94 | 7.98 |
| 2.85 | 3.84 | 4.75 | 3.58 | 2.90 | 5.87 | 3.62 | 6.66 | 3.17 |
| 0.46 | 0.47 | 0.78 | 0.42 | 0.22 | 0.34 | 1.24 | 0.30 | 0.59 |
| 0.18 | — | 0.25 | 0.14 | trac. | 0.38 | 0.31 | 0.17 | — |
| 0.97 | — | 1.04 | — | 1.51 | 0.27 | 5.81 | 8.46 | — |
| 1.83 | 1.44 | 1.80 | 2.42 | 4.01 | 3.72 | 2.79 | 0.46 | 2.30 |
| 0.09 | 0.09 | 0.02 | 0.24 | 0.10 | 0.04 | 0.01 | 0.06 | 0.29 |
| 99.88 | 99.73 | 99.81 | 99.84 | 99.72 | 99.81 | 99.91 | 99.67 | 100.00 |
| 130 | 131 | 136 | 141 | 143 | 164 | 188 | 403 | — |
| 17.8 | 19.1 | 20.1 | 21.1 | 21.0 | 29.5 | 28.0 | 46.4 | — |
| 53.5 | 51.0 | 49.0 | 47.0 | 54.4 | 49.3 | 40.8 | 5.2 | — |
| 20.7 | 20.1 | 17.2 | 21.4 | 16.4 | 3.0 | 14.9 | 4.4 | — |
| 8.0 | 9.8 | 13.7 | 10.5 | 8.2 | 18.2 | 16.3 | 44.0 | — |
| 0.10 | 0.07 | 0.10 | 0.08 | 0.04 | 0.04 | 0.18 | 0.03 | — |
| 0.42 | 0.59 | 0.47 | 0.34 | 0.59 | 0.28 | 0.08 | 0.0 | — |

9. Lava-born albite-amphibole rock, Petäjäselkä, Kittilä. Anal. A. Heikkinen. (Silvennoinen, 1966),
10. Lava-born albite-amphibole rock. Aihikkoselkä, Sodankylä. Anal. H. Lönnroth. (Mikkola, 1941, p. 245).
11. Lava-born albite-amphibole rock. 5.5 km N of the village of Kersilö, Sodankylä. Anal. H. Lönnroth. (Mikkola, 1941, p. 247).
12. Hypabyssal albite diabase. Haurespää, Kittilä. Anal. P. Ojanperä.
13. Hypabyssal albite diabase. Holkkuavaara hill, Kittilä. Anal. L. Lokka (Hackman, 1927, p. 22).
14. Lava-born albite-chlorite-amphibole rock. Harrilompolo, Kittilä. Anal. L. Lokka. (Hackman, 1927, p. 23).
15. Lava-born albite-chlorite -biotite rock. Ketunliesukuusikko, Kittilä. Anal. P. Ojanperä.
16. Lava-born albite-biotite-chlorite rock. Kuolavaara, Kittilä. Anal. P. Ojanperä.
17. Hypabyssal albitite. Petkula, Sodankylä. Anal. H. Lönnroth (Mikkola, 1941, p. 254).
18. Average composition of the Lapland greenstones, proposed by Mikkola (1941, p. 260).

Mikkola (1941) divides the extrusives of central Lapland into two groups: Kittilä greenstones and more basic amphibole-chlorite rocks. Although these two groups of rocks differ from each other in areal distribution, amphibole-chlorite rocks being situated in the east and greenstones in the west, Mikkola believes that they belong together genetically.

On the basis of eight analyses Mikkola calculated the average composition of the greenstone magma of central Lapland (presented in Table 7, Anal. 18). According to Mikkola the greenstone magma is characterized by a high Na/Ca ratio and a richness of volatiles. On the origin of the greenstone magma he states: »In accounting for the ultimate source of the greenstones lies the very crux of matter, and as yet we have no adequate explanation concerning the genesis of such magma».

Meriläinen (1961) deals with several hypabyssal albite diabases and albitites of the Kittilä area. According to him albite diabases exist as simple or differentiated sills; primary magma is olivine basaltic. Albitites have crystallized after albite diabases from magmatic liquids and solutions.

Eskola (1963, p. 187) calls attention to the relatively high carbonate content of the spilites of central Lapland. Because, according to him, it is most probable that carbonate is juvenile, he considers the albite-carbonate rocks or »karjalites» a certain kind of carbonatites. In another connection (Eskola, 1963, p. 173) he still emphasizes however the autometasomatic origin of the spilites thus ignoring the fact that karjalites were primary products of magmatic crystallization.

The richness of carbonates in Lahn-Dill weilburgites is explained by Lehman (1941) as being product of assimilation of carbonate-rich sediments by a greenstone magma. The amount of carbonate rocks in the lowest part of Lapponium is so scarce that the hypothesis of assimilation can hardly be accepted in accounting for the ultimate source of carbonates. Agreeing with Eskola's opinion, refered to above, the present author is disposed to consider these extremely carbonate-rich members of the central Lapland spilites certain kind of carbonatites which are not associated with alkalic rocks (*cf.* Bailey, 1961).

In order to get a more detailed picture of the chemical composition of the central Lapland greenstones the present author gathered together the previously published analyses from the literature and got the Department of Chemistry of the Geological Survey of Finland to do some new ones (Table 7). Fig. 48 represents the variation diagram of the greenstones of central Lapland constructed on the basis of the calculated Niggli numbers.

The rocks numbered 1—5 are all ultrabasic extrusives and there are no intrusive serpentinites or other intrusive rocks among them. Rock N:o 5 is considered to represent an intermediate type between ultrabasic and basic rocks. The rocks numbered 6—16 are basic or intermediate hypabyssal rocks which form the majority of the greenstones in central Lapland. Rock N:o 17 represents albitites, the monst acid rocks of the suite.

Without further discussion, we can state that we are dealing with a very typical group of rocks that belong to the spilite- keratophyre association, in the sense that Turner and Verhoogen (1960) use the expression. The terms spilite and spilitic are really kaleidoscopic, having many different shades of meaning, and the discussion concerning the spilites has continued for more than one hundred years, yielding an enormous »spilitic literature». The aim of the present author is not to refer to this

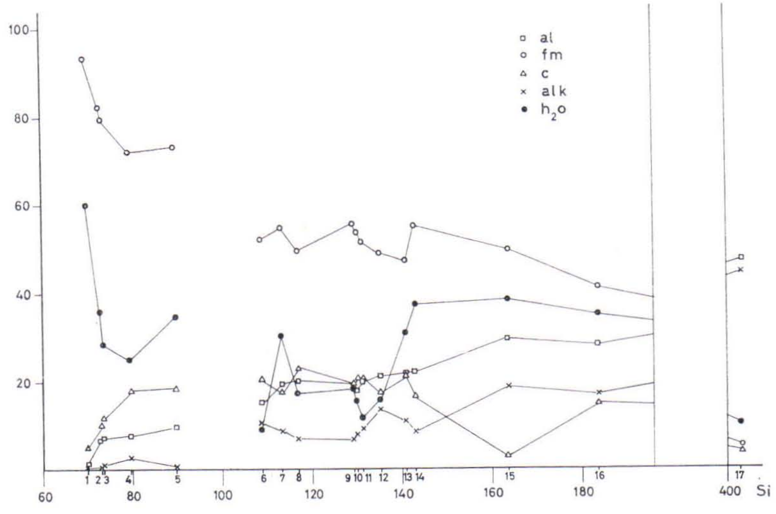


FIG. 48. Variation diagram of the greenstones of central Lapland in terms of Niggli numbers. The numbers of the specimens refer to the analyses presented in Table VII.

literature in detail, but some of the most important points of the spilite problem essential to the topics of the present investigation, will be discussed.

In his very thorough and comprehensive paper Vallance (1960) deals with all kinds of rocks called spilites in the literature, but Vallance himself will by no means define exactly what he understands by the term spilite. According to him a most important chemical difference between spilites and basalts is the large dispersion in the contents of the chemical components of the spilites compared with that of the basalts (Vallance, 1960, p. 37). The same feature appears very clearly from the variation diagram of the central Lapland greenstones, too. Quite similar is the statement of Amstutz (1968, p. 749) ». . . spilites are volcanic rocks of widely variable mineralogical and chemical composition. Albite, chlorite, epidote, calcite and iron oxides are the major constituents observed, with occasional abundance of pyroxene. This composition and its variability may be the only real property that spilites do not have in common with basalts.» On the origin of spilites Amstutz states: »The mode of formation may be understood as resulting from a transfer and differentiation of constituents in a separate aqueous phase during primary crystallization. The hydrous nature of the melt may be normally a resultat of primary differentiation or in places may be caused by contamination from surrounding rocks or by some unknown processes in a hydrated part of the earth's mantle».

With regard to the spilites of central Lapland we can summarize the following characteristic features:

1. The spilites of central Lapland are subaqueous extrusives or hypabyssal intrusives representing the initial magmatic activity of the Lapponian geosyncline in the Karelian zone.

2. The chemical and mineralogical composition varies within very large limits; from ultrabasic serpentinites and amphibole-chlorite rocks to acid keratophyres and albitites.

3. Intrusive serpentinites, sedimentogeneous jaspilites and spilitic pillow lavas form the so-called Steinmann trinity, so common in orogenic regions all over the world (cf. Bailey and McCallien, 1960).

4. In close association with the spilites of central Lapland there are oxide iron ores, manganiferous carbonate iron ores and minor sulphide iron ores forming an iron formation of exhalative-sedimentary origin.

## Ultrabasic extrusives and liquid immiscibility

All the features mentioned above are familiar and characteristic of the spilites described in the literature. The present author will call special attention to the ultrabasic members of the spilites of central Lapland.

The existence of ultrabasic lavas has been doubted on the basis of laboratory experiments. It has been emphasized that the melting range of peridotite is about 1 700°C, and that the temperature of lavas does not exceed 1 200°C. These data have led to the simple conclusion that there cannot exist any ultrabasic lavas (see, Hess, 1938, p. 328). However, Drever *et al.* (1961) confirm the existence of ultrabasic liquids at both high and low temperatures. It should therefore be noticed that the above-mentioned melting range of peridotite, 1 700°C, concerns a dry rock. The existence of volatiles, especially water, radically lowers the melting range of peridotite. Rocks of serpentine composition melt almost completely at 1 300°—1 400°C in laboratory experiments under water pressures corresponding to the deeper parts of the oceans (Barth, 1962). Further, Seifert and Schreier (1968) found that enstatite and forsterite melt at 550°—700°C under the pressure of 1—5 Kb when they contain excess water and small amounts of alkalies.

There seems to be no theoretical reason to doubt the existence of ultrabasic extrusives in central Lapland. Examples of similar occurrences are not lacking. Marmo (1949) describes picrites, associated with extrusive and hypabyssal spilites, in the volcanic complex of Suoju, Sovjet Karelia. Bailey and McCallien (1953) state that many of the serpentinites of Turkey are submarine lava flows. Petraschek (1959) writes that alpinotype peridotites and serpentinites may exist as submarine extrusives. Later Belostotskiy and Kolbantsov (1970) put forth a whole series of indications that the ultrabasites among the Dinarian ophiolites are formed from a primary peridotite magma which was in a melted stage at the beginning of the process leading to their formation.

The question of the primary origin of the ultrabasic extrusives so generally associated with spilites, as stated already by Dewey and Flett (1911), has remained open. Concerning this problem Marmo (1949, p. 59) states briefly: »Die auffallend basische Zwischenmasse konnte, wohl zufolge des hohen Wasserhaltes, nach der hauptsächlichen Kristallisation der Variolen flüssig bleiben, ganz wie die Picrite im ganzen Komplex». In the geological literature known to the present author this is the only tentative to explain the existence of ultrabasic extrusives. For instance the compilation on ultrabasic rocks, edited by Wyllie (1967), contains only one treatise on the ultrabasic volcanics (Gass, 1967) which, however, does not include any comments on the existence of ultrabasic extrusives.

Concerning the ultrabasic material as a whole Wyllie (1967) states that the material of the ultrabasic rocks generates from the upper mantle. After leaving the mantle the ultrabasic material can be in various physical stages. It is naturally possible that such material under certain conditions can reach the surface of the earth and appear as submarine extrusives. A consequence of this way of thinking is that different variants of extrusives would represent melting products of different levels of the earth's crust, because fractional crystallization cannot yield any ultrabasic liquid. Even though the fractional melting may be a very important factor in producing different kinds of magmas, it cannot satisfactorily explain the wide ranges and the capriciousness of variation of the chemical composition of the central Lapland spilites. What is the level in the earth's crust that produces albite-rich keratophyres?

The present author suggests that liquid immiscibility of silicate melts offers the most natural and the most satisfactory explanation of the origin of the ultrabasic extrusives in central Lapland.

The theory of liquid immiscibility in silicate melts, created by the founders of petrology F. Zirkel and H. Rosenbusch, was long a generally accepted explanation of the differentiation of the igneous rocks until Bowen (1928, pp. 7—19), the leading advocate of fractional crystallization, summarized the evidence for and against liquid immiscibility as a mechanism for igneous differentiation. Bowen stated that there is little or no evidence to support the theory of liquid immiscibility in silicate melts. Since then, relying upon Bowen, the great majority of petrologists have considered fractional crystallization and partial melting, supplemented by the possible assimilation, to be the only important mechanisms in igneous differentiation.

The founder of the experimental study of liquation was J. H. L. Vogt (1917) who was the first to prove the immiscibility of silicate and sulphide melts, a fact the significance of which has never since been questioned.

The immiscibility of heavy metal oxide and silicate melts was stated by Fischer (1950). He showed experimentally that such a liquation depends on the following points:

1. on the existence of volatiles, in special cases on fluorine;
2. on the phosphorus content of the melt;
3. on the excess of alkalies to calcium;
4. on the existence of heavy metals in a high degree of oxidation.

The correctness of Fischer's theses was proved by Philpotts (1967) both experimentally and by field observations. In principle the same mechanism is suggested by

8 17 576—76

Piirainen and Piispanen (1967) as an explanation of some scarn iron ores. Likewise Piirainen and Juopperi (1968) explain the titanium iron ore at Porttivaara, Northern Finland, to be a product of liquation. Nuutilainen (1968) does not use the term liquation in explaining the origin of iron ores associated with albite gabbros and albitites in the Misi area, Northern Finland, but his scheme of the magmatic evolution (Nuutilainen, 1968, p. 88) brings a clear image of liquation into one's mind. Bodart (1968) states in his study on some South Norwegian albitites that in a protomagma of gabbroic composition, after the extraction of ultrabasic phase by crystallization, liquation occurred which led in one trend to albitites, in another to ore phase. Further on the ore phase split into two parts: iron ore and apatite veins.

Thus we can state that liquation of sulphide and oxide melts from silicate melts is a generally accepted fact, but the same phenomenon between two silicate melts will not be admitted. However, both experiments and field observations show consistently that this phenomenon is, not only possible, but in some cases even a very important factor in the differentiation of igneous rocks, as stated by Fenner (1948). Although Bowen and Andersen (1914) discovered the immiscibility in the binary system $MgO—SiO_2$ in the compositions rich in $SiO_2$, Bowen (1928, p. 11) did not hestitate to present Greig as the first who has revealed experimentally examples of the immiscibility in silicate melts. Greig (1927) demonstrated that silica shows only partial miscibility with melts of CaO, MgO, FeO and $Fe_2O_3$, but he did not expect it to be a factor of any importance in the differentiation of igneous rocks. Despite the negative results of Bowen and Greig the idea of liquid immiscibility did not die. Several experiments proved the existence of liquid immiscibility in silicate melt under certain circumstances. Kracek (1930) shows that the oxides of alkalies are miscible with silica in all proportions but that the oxides of alkaline earths, Ca, Sr, and especially Mg, show a large gap in the miscibility with silica (Fig. 49). Grigoriev (1935) succeeded in getting two immiscible silicate melts, basic and alkalic ones. Spencer (1938) states liquid immiscibility in the system $KAlSi_3O_8—NaAlSi_3O_8$. Friedman (1950) demonstrates that addition of water to the system $Na_2O—SiO_2$ causes liquation. Later on Roedder (1951) states liquid immiscibility in the system leucite-fayalite-silica at 1 100°C (Fig. 50). Van Groos and Wyllie (1968) discovered liquid immiscibility on a large field of composition in the join $NaAlSi_3O_8—Na_2CO_3—H_2O$. The observed phases of the system at temperatures over 725°C are albite, cancrinite, Na-carbonate, silicate-rich liquid, Na-carbonate-rich liquid and gas. Addition of water decreases the critical temperature.

Liquid immiscibility is used by many geologists as an explanation for certain features observed on the field. Grout (1918) explained gabbro and granophyre at Duluth as the results of liquid immiscibility in a primary magma. Asklund (1925) gets into liquation in the explanation of the differentiation of igneous rocks in eastern Götaland in Sweden. Tanton (1928) interprets the sharply bounded ovals of glass in a partly devitrified obsidian at Agate Point, Ontario, as proof of separation by liquid immiscibility. Geijer (1935) considers that the phosphate-rich iron ores at Kiruna in

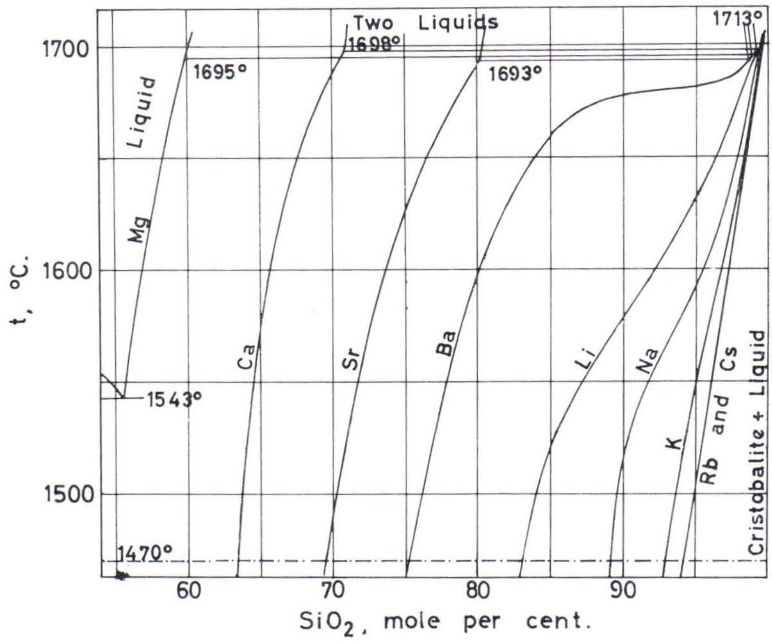


Fig. 49. Diagram to illustrate the periodicity of the cristobalite melting point curves for the alkaline earth and alkali silicate mixtures. Quoted from Kracek (1930).

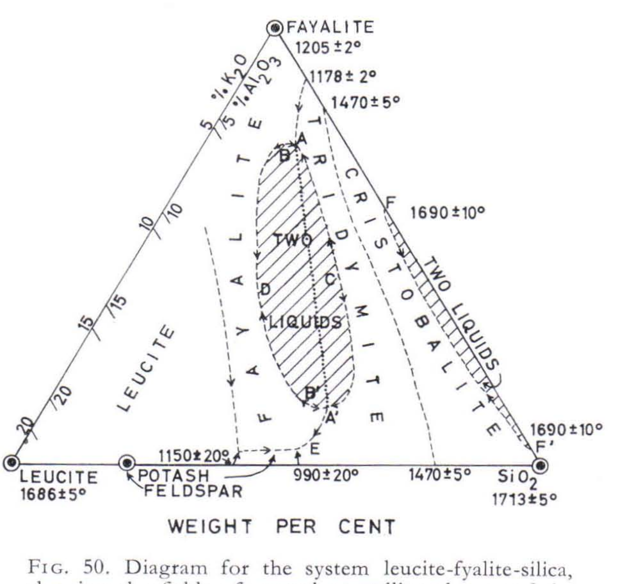


Fig. 50. Diagram for the system leucite-fyalite-silica, showing the fields of several crystalline phases and the areas of immiscibility with the liquidas surface (shaded). Quoted from Roedder (1951).

Sweden, closely associated with sodic syenites, are formed by liquid immiscibility. Schneiderhöhn (1962) accepts the concept of Geijer. Further, Loewinson-Lessing (1935) suggests, in a study on variolites from Yalguba, that after the crystallization of pyroxenes the residual melt split into two immiscible melts: varioles consisting of oligoclase, quartz and magnetite, and a surrounding matrix cosisting of bytownite, olivine and magnetite. Holgate (1954) explains the existence of siliceous xenoliths in basalts by the immiscibility of acid and basic melts in the compositions outside the field of normal effusives. Drever (1960) interprets globules found by him from a picritic intrusion in West Greenland as the result of liquid immiscibility. Wahl (1963) explains the charnockites and unakites of Central Finland to result from deep-seated liquid immiscibility. Roedder and Coombs (1967) interpret the inclusions of granites in Ascension Island as the result of immiscibility between dense saline fluid and silicate magma.

Theoretically, the immiscibility of two liquids is explained by the diagram in Fig. 51. The diagram, modified after Bowen (1928), Grigoriev (1935) and Barth (1962), represents a binary system A—B with two immiscible liquids. A melt with a composition of x is cooled. At 1, drops of liquid 2 begin to separate. At the temperature CD liquids C and D are in equilibrium. At this temperature crystals of B begin to separate from both liquids and this goes on until the eutectic point E is reached. The immiscibility gap has, in principle, the shape of an oval; in multicomponent systems that of an ellipsoid with two critical points. The lower critical point is reached only when the exsolution area is wholly over the stability fields of the crystalline phases. Thus two liquids immiscible at a certain temperature and pressure, may be miscible again at a lower temperature and pressure, presuming that the liquids have not been separated from each other due to differences in the viscosity, density or other properties of the melts.

The diagrams in Figs. 48 and 52 show consistently that if there is any gap in the miscibility of the primary greenstone magma of central Lapland it has to lie on the composition interval between analyses 5 and 6. In the opinion of the present author the diagrams really support the interpretation that the ultrabasic effusives of central Lapland have resulted from liquid immiscibility of the volatile rich spilitic parent magma.

The areal distribution of the ultrabasic and more acid members of the central Lapland spilites is very interesting and may not be forgotten in accounting for the ultimate source and the evolution of the primary magma. The ultrabasic phase exists mainly in the east and the more acid rocks in the west. The demarcation line is rather sharp but some intermixing does exist. According to the field observations it seems very probable that the ultrabasic phase as a whole is slightly younger than the more ordinary greenstone phase.

Taking into account all the facts and assumptions presented above we can recreate the following account for the origin and and evolution of the spilitic greenstones in central Lapland:

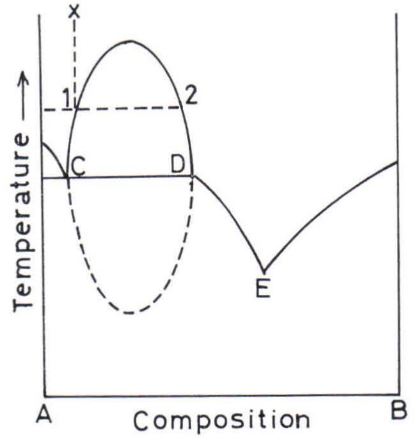


FIG. 51. Schematic diagram of a binary system with two immiscible liquids. Further explanations in the text.

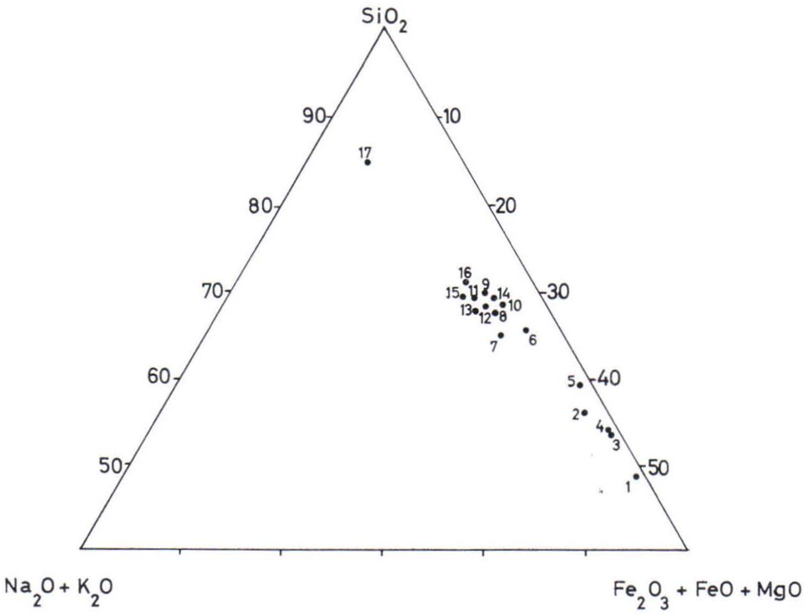


FIG. 52. Variation diagram of the greenstones of central Lapland. The numbers of the specimens refer to the analyses presented in Table 7.

The first phase in the eugeosynclinal development of the Lapponian geosyncline was the formation of deep-reaching fractures. The forces that caused fracturing achieved also partial or complete melting of the upper mantle (see, Sugimura, 1968,

p. 564). The pressure gradient forced the volatiles to migrate toward fracture zones and thus caused the enrichment of volatiles to the magma. According to Yoder and Tilley (1962, p. 465) the hydrous minerals can be a natural reservoir of volatiles at depth. Volatiles, especially water and carbon dioxide caused liquation which led to an ultrabasic phase in one trend and to a spilitic phase in another. Both of the phases were still volatile-rich, but it seems likely that water was enriched to the ultrabasic phase and carbon dioxide to the spilitic phase. The carbon dioxide of the spilitic phase prevented the crystallization of calcic plagioclase (Piispanen,1970) and led to the formation of various albite-bearing carbonate-rich extrusives. The ultrabasic phase, due to its higher density, remained longer on deeper levels. Orogenic movements shut the eruption fissures in the west and thus led to the extrusion of the ultrabasic phase in the east. It seems most probable that independent serpentinite intrusions may also generate from this ultrabasic phase.

The hypabyssal parts of the spilitic branch (Anals. 6, 8, 12 and 13 in Table 7), were differentiated by fractional crystallization which at some places led to the formation of albitites, Anal. 17 (see, Hackman and Wilkman, 1926; Ödman, 1939; Mikkola, 1941; Meriläinen, 1961; Paakkola, 1964; Nuutilainen, 1968; Piirainen, 1968 and 1969).

The present author is well aware of other opinions on the origin and formation of spilitic rocks than presented in this paper, but, for practical reasons, they have not been reviewed and discussed in this study, especially, because, (on the basis of their field relations, structures and textures) the spilites of central Lapland hardly can be interpreted as metasomatic or other than primary magmatic rocks.

## The nature of the tuffaceous schists

There is no absolutely sure criterion for distinguishing between true pyroclasts and sediments produced by weathering of volcanics. Rapid weathering does not cause any noticeable change in chemical composition and sedimentation on the sea bottom under influence of turbidity currents will give rise to quite similar structures for both clastic and pyroclastic sediments. Thus, on the basis of field and microscope observations only, it would be impossible to distinguish between these two groups of rocks. The prerequisite for weathering is the emergence of the weathering object above sea level. In the evolution of a geosyncline this is reached by rising of an island arc, and at once it means beginning of the flysch facies, indicated by sedimentation of graywackes. Sound proof of the existence of flysch would be a hiatus between volcanics and flysch.

Even though there are plenty of graywackes in western Kittilä (Mäkelä, 1968), the present author is inclined to the opinion that most of the tuffaceous schists in eastern Kittilä are of true pyroclastic origin. The chemical analyses (Nos. 1 and 2 in Table 8) do not disprove this assumption.

### The origin of graphite in the tuffaceous schists

As stated earlier (p. 34) graphite is very common in tuffaceou schists but more rare in jaspilites. These graphite bearing schists have been called black schists (Hackman, 1927; Mikkola, 1941).

The black schists of Finland are usually declared to be organogenic, saprogelic, sediments (Laitakari, 1925; Eskola 1932; Peltola, 1960). On the otner hand, as early as 1941 Mikkola made an allusion to the close association of volcanics, jaspilites and black schists, but he paid no attention to the character of these black schists which differ quite clearly from normal sapropelic or pelitic sediments. The present author thinks, however, that the surroundings and the whole geological milieu must be taken into account when investigating the origin of any rock or mineral. Hence, because the base material of graphite-bearing schists is rather of pyroclastic than sapropelic nature (Anals. 3 and 4 in Table 8), one must not automatically suppose a biogenic origin for the graphite.

In accordance with Marmo and Metzger (1953), Preston (1954) suggests that part of the graphite met in the bedrock of Finland, especially in connection with limestones, is a reduction product of carbon dioxide liberated in the metamorphism of limestone.

Piispanen (1964) suggests that graphite in the scapolite-bearing rocks at Svappavaara, North Sweden, is inherited from juvenile carbon dioxide (see, formula 7 below).

Miyashiro (1964) shows that the fugacity of oxygen in graphite-bearing schists, in metamorphism, depends on the partial pressure of carbon dioxide and temperature. In other words, graphite partly controls the oxidation/reduction relatioship.

Vaasjoki (1966), continuing the scrutiny on the basis of Miyashiro, shows that there are possibilities for the following reactions:

(1) $2H_2O = 2H_2 + O_2$

(2) $C + O_2 = CO_2$

(3) $S_2 + 2H_2 = 2H_2S$

(4) $2H_2S + 2Me = 2MeS_2 + 2H_2$

(5) $CO_2 + 2H_2 = C + 2H_2O$

Silvennoinen (1966) deals with the origin of graphite in graphite-bearing schists from Petäjäselkä, and shows, inclining to the view of Pavlides and Milton (1962), that carbon dioxide is liberated when siderite alters to magnetite in accordance with the following formula:

(6) $3FeCO_3 = Fe_3O_4 + 2CO_2 + CO$,

whence follows the appearance of graphite according to formula (2), or after Partington (1951):

(7) $2CO = C + CO_2$

The starting point of Silvennoinen is that the formation of graphite would be due to a diagenetic or metamorphic reaction. One must agree that Silvennoinen's way of thinking is logical and the reactions brought up are possible, but the following points are in contradiction to the hypothesis for the diagenetic or metamorphic origin of graphite:

1. Although graphite-bearing schists commonly contain siderite, magnetite supposedly in reaction (6) is met only occasionally; on the other hand, where siderite has altered more largely to magnetite, graphite is not found at all.

2. Iron sulphides of graphite-bearing schists in this milieu are most probably authigenic (see, Wright, 1965).

3. Graphite-bearing and graphite-free beds alternate irregularly: the thickness of graphite-bearing beds varies from a few millimetres to several metres.

TABLE

Chemical compositions of tuffaceous and sedimentary

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 58.16 | 59.45 | 55.60 | 52.82 | 49.02 | 49.69 |
| $TiO_2$ | 1.77 | 4.60 | 2.68 | 0.86 | 0.36 | 0.81 |
| $Al_2O_3$ | 16.63 | 13.23 | 12.90 | 14.46 | 6.95 | 9.45 |
| $Fe_2O_3$ | 1.55 | 1.47 | 2.12 | 0.17 | 3.80 | 8.24 |
| FeO | 6.52 | 7.70 | 10.91 | 15.69 | 19.99 | 12.65 |
| MnO | 0.14 | 0.18 | 0.15 | 0.02 | 2.62 | 2.04 |
| MgO | 4.95 | 3.49 | 7.29 | 2.98 | 2.89 | 2.54 |
| CaO | 1.99 | 0.63 | 0.02 | 0.86 | 0.38 | 0.21 |
| $Na_2O$ | 2.98 | 5.15 | 1.32 | 5.43 | 0.17 | 0.17 |
| $K_2O$ | 0.97 | 2.42 | 1.12 | 1.04 | 0.36 | 2.33 |
| $P_2O_5$ | 0.16 | — | — | — | 0.14 | 0.10 |
| $CO_2$ | — | — | — | — | 10.25 | 8.74 |
| $H_2O^+$ | 3.90 | 1.44 | 5.12 | 0.71 | 3.53 | 3.33 |
| $H_2O^-$ | 0.13 | 0.15 | 0.17 | 0.12 | 0.01 | 0.08 |
| | 99.85 | 99.91 | C = 0.19 | S = 6.84 | 100.47 | 100.38 |
| | | | 99.53 | Cu = 0.04 | | |
| | | | | Ni = 0.01 | | |
| | | | | 102.02 | | |
| | | | | —O = $S_2$ 1.71 | | |
| | | | | 100.31 | | |

1. Fine-grained tuffaceous schist. Petäjäselkä, Kittilä. Anal. A. Heikkinen (Silvennoinen, 1966, p. 34).
2. Fine-grained tuffaceous phyllite-like schist. Kuolajärvi, Kittilä. Anal. L. Lokka. (Hackman, 1927, p. 47).
3. Black carbonaceous schist with ilmenite scales. Kersilö, Sodankylä. Anal. H. Lönnroth. (Mikkola, 1941, p. 206).
4. Black schist interbedded with pyritiferous layers (sulphide schist of this paper). Sattasköngäs rapid in Sattasjoki River, Sodankylä. Anal. L. Lokka. (Mikkola, 1941, p. 208).
5. Mangano-siderite-schist. Silmänpaistama, Kittilä. Anal. P. Ojanperä.

For the reasons mentioned above it seems most probable that graphite is authigenic, formed contemporarily with sulphides according to formulas (4) and (5) under strongly reducing conditions on the bottom of a fairly deep sea. The components in the reactions would be mainly of volcanic origin.

## The nature of jaspilites and iron formation

Mikkola (1941) was the first to state that the »quartzites» in the Porkonen-Pahtavaara area are chemical sediments, and he supposed that silica had precipitated as cinter from hot springs associated with volcanic activity. On the basis of the trace element determinations by Sahama (1945) Kaitaro (1949) suggests an exhalative-sedimentary origin for jaspilites and associated oxide iron ores.

8

schists associated with central Lapland greenstones.

| 7 | 8 | 9 | 10 | 11 | 12 |
|---|---|---|---|---|---|
| 22.42 | 50.44 | 96.36 | 95.64 | 93.80 | 89.56 |
| 0.27 | 0.16 | 0.00 | 0.00 | 0.00 | 0.29 |
| 3.65 | 7.47 | 0.21 | 0.28 | 0.47 | 4.12 |
| 2.82 | 0.67 | 1.29 | 2.63 | 3.39 | 1.73 |
| 29.52 | 2.18 | 0.84 | 0.22 | 2.01 | 0.82 |
| 9.85 | 0.07 | trac. | trac. | trac. | 0.01 |
| 4.65 | 6.76 | 0.00 | 0.00 | 0.00 | 0.62 |
| 1.80 | 10.70 | 0.00 | 0.00 | 0.00 | 0.34 |
| 0.03 | 1.78 | 0.13 | 0.23 | 0.22 | 0.14 |
| 0.30 | 1.59 | 0.00 | 0.00 | 0.00 | 1.34 |
| 0.38 | 0.13 | 0.00 | 0.00 | 0.00 | 0.02 |
| 23.52 | 16.32 | — | — | — | 0.29 |
| 1.60 | 1.43 | — | 0.58 | 0.08 | 0.62 |
| 0.03 | 0.01 | — | — | — | 0.02 |
| 100.34 | 99.71 | — | — | — | — |
| | loss. of ign. | 0.45 | 0.27 | | |
| | | 99.28 | 99.85 | 99.97 | 99.86 |

6. Mangano-siderite schist. Pitslomakuru, Kittilä. Anal. P. Ojanperä.
7. Mangano-siderite schist. Petäjäselkä, Kittilä. Anal. A. Heikkinen. (Silvennoinen, 1966, p. 20).
8. Impure dolomite. Jänesvaara, Kittilä. Anal. P. Ojanperä.
9. Jaspilite. Northern part of Pahtavaara, Kittilä. Anal. E. Ståhlberg (Hackman, 1925, p. 19).
10. Brecciated jaspilite. Eastern slope of Pahtavaara, Kittilä. Anal. E. Ståhlberg (Hackman, 1925, p. 19).
11. Homogeneous jaspilite. Kuoreslaki, Kittilä. Anal. E. Ståhlberg (Hackman, 1925, p. 19).
12. Sericite jaspilite. The dam of Karjakkojoki River, Kittilä. Anal. P. Ojanperä.

9 17576—70

Refering to the features presented in the description of the rocks there can hardly be any doubt of the chemical-sedimentary character of these rocks, and the present author accepts Kaitaro's idea of the exhalative-sedimentary origin of the formation.

Jaspilites, oxide iron ores, mangano-siderite schists and sulphide schists represent in common a very typical example of Precambrian iron formations, as found all over the world.

Owing to their economic importance the bedded iron deposits have been extensively studied and the literature concerning them is enormous. Fortunately many comprehensive résumé articles have been published during recent years (e.g., Gross, 1965; James, 1966). Much is known about the mode of occurrence and the nature of these rocks, but there are still disagreements regarding their origin and the ultimate source of iron (and manganese).

The terminology concerning bedded iron deposits is confusing. The names of the rocks vary from place to place and from age to age. Precambrian iron-rich supracrustal rocks are known as iron formation, itabirite, taconite, quartz banded iron ore, specularite schist, hematite or mangetite quartzite, calico rock, jacutinga, and probably by many other names. The general name for post-Precambrian iron-rich sediments is ironstone, but they are also known as Minette ore, Clinton-type ore, Lorraine ore, oolitic iron ore, black band ore and perhaps by many other names with local connotations. The problems of the nomenclature and classification of iron formations are so involved that there is no reason to make any great effort in that direction in the present context.

The term »iron formation» is used by the present author as a purely descriptive one for all the rocks of sedimentary origin that contain 15 % or more iron (see, James, 1966).

Among the best known iron formations of the world the following should be mentioned (for references, see, Gross, 1965, and James, 1966):

Lake Superior, the prototype of iron formations, with its subprovinces, the iron formations of the Labrador trough and Greenville belt, Michipicoten, and the Helen iron range in North America;

Minas Gerais and Matto Grosso in South America; occurrences in the Transvaal, Liberia, Morocco, Tunisia, United Arab Rebublic, and beneath the Red Sea in Africa; large iron deposits in different parts of Australia; Singbhum and others in India; the carbonaceous iron formations of the Alps; the oolitic iron ores of Western Europe; the Lahn Dill area in Germany; Krivoy Rog, Olenogorsk and Kursk in South Russia; the quarz banded iron ores of Central Sweden; Sydvaranger in Eastern Finnmark, Norway; the Porkonen-Pahtavaara area, Jussarö, Jauratsi and Ilomantsi in Finland; Kostamus in Soviet Karelia.

The chemical-sedimentary nature of iron formations in commonly agreed, allthough the idea of hydrothermal replacement has spokesmen, too (e.g., Dorr, 1954; Doetsch, 1960; Friedrich, 1961; Nakhla, 1961; Basta and Amer, 1969; McLeroy, 1970).

The precipitation of iron (and manganese) is controlled by the terms of pH and Eh, under the conditions of oxide, carbonate, silicate or sulphide facies (Krumbein and Garrels, 1952; Huber and Garrels, 1953; James, 1954 and 1966; Huber, 1958; Pavlides and Milton, 1962).

The question of the source of iron (and manganese) in iron formations is, perhaps, the most controversial among iron formation discussions. Current ideas on this problem can roughly be grouped as follows (excluding the theories of secondary enrichment):

1. Iron and manganese generate from volcanic emanations (Niggli, 1950; Goodwin, 1958, 1962, and 1964; Gruss, 1958; Oftedahl, 1958; Page, 1958; Bugge, 1960; Aquirre and Menech, 1964; Shatskiy, 1964; Schweigert, 1965; Dzotzenidze, 1966; Sorem and Gunn, 1967; Ferencic, 1969).

2. Iron and manganese have been leached from sediments by volcanic emanations (Marmo, 1958).

3. Iron and manganese have been leached from sediments by ground waters heated by volcanism (Degens and Ross, 1969).

4. Iron and manganese have been leached from volcanics or clastic sediments by sea water on the bottom of the sea (Borchert, 1960; Davidson, 1964; Harder, 1964).

5. Iron and manganese generate from epicontinental weathering solutions (Woolnough, 1941; Hoenes and Tröger, 1948; Guild, 1953; James, 1954; Alexandrov, 1955; Hough, 1958; Huber, 1959; Pavlides and Milton, 1962; Lepp, 1963; Lepp and Goldich, 1964; Govett, 1966; Schultz, 1966).

6. Iron is produced by weathering of pyritiferous schists (Allen 1949, Zirkl 1965).

The arguments presented for every one of the groups seem to be so reliable that it is quite obvious that there must be more than one source for the iron (and manganese) of the iron formations. Just as clear is the fact that the depositional environment and the lithological associations of different iron formations vary within a rather large range. The classification of iron formations should not be made on the basis of external characteristics only. Such descriptive terms as Lake Superior type, Algoma type, Clinton type, Minette type, or non-oolitic type iron formation are often very useful in a practical sense, but so far as the terms have no genetical meaning they must be treated as temporary ones only. The final goal must be a genetic classification.

Taking into account the source of iron and the geological environment then at least the following genetic groups of iron-rich sedimentary formations can be distinguished;

1. Exhalative-sedimentary iron formations in association with initial magmatism in eugeosynclinal environments.

2. Marine sedimentary, oolitic or non-oolitic, iron formations in miogeosynclinal environments.

3. Continental lateritic iron formations generated by weathering.

4. Sea, bog and spring formations especially in arctic glaciated areas.

When investigating the source of manganese and iron in the Porkonen-Pahtavaara iron formation the most remarkable and conspicuous feature is the total lack of proof of continental weathering and associated sedimentation contemporaneous with precipitation of the iron and manganese. Within a couple of dozen kilometres east from the iron formation there are arenaceous and argillaceous sediments intermixed with volcanics, but it is extremely doubtful that iron and manganese would have been transported such a long way and then precipitated as numerous small, separate occurrences. The small size and uneven distribution of single ore bodies proves that the source of the metals cannot be far away.

For the purpose of testing the theory of leaching the present author carefully studied, both microscopically and chemically, volcanics from the immediate vicinity of the iron formation sediments as well as volcanics from some distance away. No signs of leaching could be observed. The iron and manganese content of volcanics varies, of course, but the variation was similar both in the vicinity and at a distance from the iron formation. Thus the most probable explanation is that iron and manganese generate directly from the emanations of initial volcanism.

Further, it is probable that the formation of these metalliferous emanations has a causal connection with the internal development of the initial volcanism and especially with the liquation of the magma. According to Ovchinnikov (1960) magma is a microheterogenic ionic-electronic liquid that contains iron and other metals even in the atomic stage. The metals can separate by liquation caused by the decrease in temperature and then be carried away passively by the gas bubbles. Even though the present author prefers the ionic stage of iron and manganese in emanations, they cannot have been in the form of halogenides as suggested, for example, by Schneiderhöhn (1962), because halogenides would have caused a decrease of pH by hydrolysis and this would have prevented the precipitation of carbonates. Rather than halogenides carbon dioxide and water were the major agents in carrying iron and manganese from magma to the sea water (see, Martin and Piwinskii, 1969).

## Primary versus secondary minerals of the iron formation

Distinguishing between minerals of primary sedimentary, diagenetic or metamorphic origin in the iron formation is a very difficult task. The basic fact is that oxide iron ores have been precipitated under oxide facies conditions, carbonaceous iron-manganese ores under carbonate facies and sulphide iron ores under sulphide facies conditions. No separate silicate facies is to be observed. The iron silicates of probable sedimentary origin occur mainly together with the minerals of carbonate facies.

The present author tentatively suggests that the primary minerals of oxide facies are hematite, magnetite and goethite. Kaitaro (1949) who knew only the ores of oxide facies in this iron formation, emphasized the role of metamorphism in the final formation and location of the ores. On the question of primary minerals Kaitaro

(1949, p. 144) states: »Darum ist es für plausibel zu halten, dass eisenreiche Lösungen abwechselnd mit Kieselsäureexhalationen ins Meer getreten sind und Eisen wahrscheinlich in Form von Eisenhydroxyden, vielleicht aber teilweise als Eisenkarbonate ausgefällt worden ist». Figures 35 and 37 show clearly that a part of the magnetite is an alteration product of both hematite and goethite, but only a part.

Besides hematite and goethite siderite has been suggested as being a progenitor of magnetite, too (LaBerge, 1964). Manganosiderite schists, especially those which are richest in manganese and iron, contain up to ten per cent magnetite, and this magnetite might be an alteration product of mangano-siderite, but the intimate alternation of thin magnetite, siderite and goethite beds on the southern slope of Haurespää proves the authigenity of every one of these minerals. Formation of magnetite as a pure precipitation product has been ignored, and suggested that hematite or iron hydroxide are the only primary minerals of oxide facies (Govett, 1966), but during recent years arguments, based on experiments and field observations, for the formation of magnetite by precipitation have been presented by several authors (Friedman, 1954; James, 1954; Hegemann and Albrecht, 1954/55; Goodwin, 1956; Huber, 1958; Cumberlidge and Stone, 1964; Stewenson and Jeffery, 1964; Klein, 1966). The stability fields of magnetite and siderite in terms of pH and Eh overlap, and the alternation of siderite and magnetite beds may indicate rhytmic alternation of the precipitation conditions or simply rhythmic change in carbon dioxide concentration.

With the exception of minnesotaite the iron silicates occur together with the minerals of carbonate and sulphide facies. Chamosite and greenalite are commonly thought to be of primary or at least of diagenetic origin, but it has been suggested that stilpnomelane is formed by alteration of volcanic material (e.g., LaBerge, 1966 a, 1966 b).

## The distribution of iron and manganese

Iron has been precipitated under every one of the different facies, but noticeable manganese content is met only in sediments belonging to carbonate facies. The amount of manganese in oxide facies is insignificant ranging from 0.1 to 1.2 per cent while that of iron ranges from 14 to 65.7 per cent. No oxide manganese minerals have been found except for some weathered boulders, and manganese in oxide facies is fixed to carbonates.

The iron and manganese content of manganosiderite schists varies from place to place. Because of the difficulties in separating manganosiderites the variation range in their composition has not been made clear but the total analyses of manganosiderite-bearing schists, taking into account the variation in the amounts of other iron-bearing minerals, support the opinion that the iron manganese proportion is quite constant.

The highest known manganese and iron contents in the manganosiderite schists are at Silmänpaistama where 15.8 % Mn and 23.8 % Fe were observed in a 2 m thick bed, but this is exceptional. Generally the manganese content of these schists is less than 5 %. Taken as large units, without any attempt to discern possible profitable ores, the different manganosiderite occurrences are surprisingly homogeneous with regard to the total manganese and iron contents as shown in Table 9.

In sulphide facies no alabandite is met and occasional manganese is fixed to mangano-siderite also here.

The average tenors of iron and manganese presented in Table 7 is not at all exceptional among the iron formations. According to Lepp and Goldich (1964) the average total iron content of 2 200 different samples is 26.7 % Fe. Because no systematic analysis is made of the oxide iron ores of the Porkonen-Pahtavaara area, it is impossible to say what is the average iron content here but a rough estimation gives a value of a little less than 30 per cent Fe which would be in close accordance with the value given by Lepp and Goldich.

Not all the iron formations are manganiferous and the manganese content is quite inconstant. Gustavsson (1960) describes a manganese-bearing siderite ore at Rubben, in the Troms district, Norway, which contains 5—8 % manganese and 30—40 % iron. According to the description given by Goodwin (1964) the Helen Iron Range shows many similarities to the Porkonen-Pahtavaara area and contains 1.99 per cent manganese on an average. The carbonate-rich part of the Ironwood iron formation in Michigan and Wisconsin contains 24—39 per cent iron and 0.93—1.33 per cent manganese (Huber, 1959). The sideritic iron formation in Cuyuna, Minnesota contains 12.9 per cent manganese on an average (James, 1966). At Målselv, in the Troms district, Norway, there is a manganese-bearing siderite ore containing about 40 per cent iron and 5—6 per cent manganese (Landmark, 1952). According to Lepp (1968) the North Range of the Trommald iron formation in the Cuyuna district has a mean manganese tenor of 2.43 per cent. According to Magnusson (1953) the manganese content of the quarz-banded iron ores in Central Sweden seems to be insignificant,

TABLE 9

Average manganese and iron contents in different occurrences of mangano-siderite schists in the Porkonen-Pahtavaara area. Analyzed in the Chemical Laboratory of Rautaruukki Oy.

| Place | Number of anal. | Mn % | Fe % | Fe/Mn |
|---|---|---|---|---|
| Puolalaki | 179 | 4.4 | 18.8 | 4.3 |
| Kerolaki | 126 | 4.9 | 19.7 | 4.0 |
| Silmänpaistama | 304 | 4.5 | 20.2 | 4.5 |
| Kuolavaara | 55 | 2.7 | 22.3 | 8.3 |
| Keulakkopää | 279 | 4.2 | 22.0 | 5.2 |
| Riesiövaara | 35 | 3.0 | 22.0 | 7.3 |
| Säynäjäjärvi | 20 | 3.5 | 18.2 | 5.2 |

but that of skarn iron ores varies from 1 to 12 per cent. The eulysites, magnetite-poor skarns, contain from 2 to 10 per cent manganese.

We can conclude that the manganese content of the manganese-bearing parts of iron formations ranges from 1 to 16 per cent the average value lying between 4 and 5 per cent. It is outside the scope of this study to deal with manganese ores proper which in many cases are analogous in regard to origin with iron formations, but it can be stated that carbonate manganese ores are poor in iron and rich in manganese. According to Gruss (1958) the mean tenors of iron and manganese in exhalative-sedimentary manganese carbonate ores in the Liassic occurrences from the Alpine district of Berchtesgaden und Salzburg are 3.30—8.80 per cent Fe and 17.36—29.34 per cent Mn. The factors controlling the distribution and concentrations of iron and manganese in carbonates are not known in detail and it would be a very intresting subject for further studies.

## The factors controlling facies distribution

Although the tectonics and manner of folding of the area are not known in detail (see pp. 72—75), we can roughly conclude that different facies do not lie side by side but are rather one on top of another. The sequence from bottom to top is oxide, carbonate and sulphide facies. This probably means that in the early stage of initial volcanism which led to the formation of ores the sea was quite shallow and oxidizing conditions were dominant. In the course of time the bottom of the geosyncline submerged unevely and more quickly than it could be filled by volcanics and chemical sediments. A consequence of the submergence was the decrease of oxidation potential and a change to carbonate facies conditions or even to sulphide facies conditions for parts of the deepest troughs, but only locally and strictly temporally.

According to Cumberlidge and Stone (1964) and Shunzo (1966), metamorphism does not always cause any large change either in bulk composition or in the oxidation stage of the sediment. Thus the facies distribution is to be interpreted as a primary feature, especially because the grade of metamorphism has been low (see p. 72).

## The role of metamorphism

The mineral assemblages and microscopic structures of the rocks described in the petrographic part of this study clearly show that the grade of metamorphism in the area has been very low. With regard to the minerals of igneous rocks it is impossible to say which of the minerals are of primary, autometasomatic or metamorphic origin. It is obvious that there are mineral assemblages belonging to each of the above-mentioned categories.

The same difficulty also arises with the sedimentogeneous rocks. The low grade of metamorphism does not offer sufficient qualification to distinguish between primary, diagenetic and metamorphic minerals. So, for instance, the possible primary mineral paragenesis siderite-chlorite can, according to Bubenicek (1969), be formed in diagenesis from limonite, quartz and organic material in accordance with the equation:

6.05 limonite + 1 quartz + reducing organic substance = 4.72 siderite + 4.26 chlorite.

However, in the case now under consideration the oolitic structures of siderite and its alternation with quartz and iron oxides suggest the primary nature of siderite and, as already stated, some iron silicates such as chamosite and greenalite can also be produced by chemical precipitation.

In every case the mineral parageneses of the rocks indisputably show that the grade of metamorphism does not exceed that of the greenschist facies. Recrystallization of the hypabyssal rocks was usually very weak and hence mineral parageneses formally of higher grade were preserved in them.

In the classification of Turner and Verhoogen (1960), the metamorphic facies of the Porkonen-Pahtavaara area lies between the greenschist facies and zeolite facies.

## STRUCTURAL FEATURES

In the creation of a general geological view of the area, tracing the tectonic evolution has been the hardest task. This is caused firstly by the rapid rhythm of the alternation of rocks in both vertical and horizontal directions, a phenomenon very characteristic of a volcanic milieu. Single rocks only seldom form beds thicker than a few dozen metres.

Therefore the connections of the profiles got by diamond drilling cannot be performed in a uniform manner. In addition finding and following the lead horizons, a necessity for resolving fold structures, is a very difficult task. Even the weathered character of the outcrops rendered the observation and measurement of tectonic elements difficult.

Resolving the problems of tectonics and stratigraphy is a composite task: they both depend on each other in such a manner that an error in the one leads automatically to an error in the other. On the other hand, one of the leading goals in geological research is the creation of a logical and, above all, truthful picture of the geological evolution of the subject so far as is possible on the basis of available information. Hence the present author regards it as important to present at least a schematic and generalized view of the tectonic evolution of the area.

Folding, which began in the early stage of the volcanism, was caused by a compression roughly perpendicular to the N—S direction. The undermost parts of the sequence were folded more strongly than the younger rocks (Mäkelä, 1966 and 1968). However, there are local differences in this relationship. As an example one could mention the clastic molasses of the Kumpu-Oraniemi series folded together with

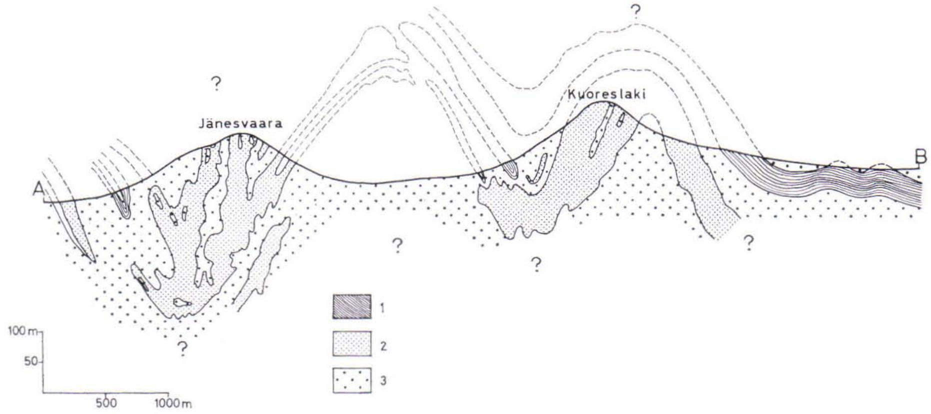


FIG. 53. Schematic cross section of the fold structures in the Jänesvaara-Kuoreslaki area. 1 = horizon of tuffaceous schists; 2 = horizon of jaspilites; 3 = greenstones.

volcanics at the Kutuoja creek. At other places the same bed of volcanics is quite flat-lying and only weakly deformed as seen a few hundred metres NW from Kutuoja by the road leading to Rovanpää.

There are reasons to suppose that the oxide iron ores and associated jaspilites represent the deepest and therefore the most strongly folded horizon. A conspicuous feature is the location of synclinoriums and anticlinoriums in the same places where the iron ores are situated. It seems that the iron ores have played a significant role in the location of fold elements, exactly the opposite of Kaitaro's idea (1949). Taking into consideration the well-preserved primary structures, it would have been no opportunities for any large scale tectonoplastic migration, and still less for any kind of material movement by metasomatism or metamorphic differentiation.

A prerequisite for the formation of folds in the oxide iron ore horizon was east-west directed, steep, faulting. These faults are seen for instance in the dislocation lines of Pitslomakuru and Lomakuru. The block between these two strong fault lines represents the deepest and most strongly folded horizon in the area. Owing to the subsequent uprising and denudation it is now on the same erosion level as its younger surroundings. At Haurespää and Kuoreslaki this horizon occurs as an upraised, at some places intensively folded, synclinorium with almost vertical dipping beds in the middle. To the south of Haurespää folds become more gentle with the axis dipping 20—25° to SSW. In the northern part of Haurespää the axis of the synclinorium dips gently towars the north with the consequence that the flanks of the synclinorium separate from each other on the recent level of erosion at the southern end of Kuoreslaki. The eastern flank of this synclinorium (Fig. 53) locates the mangano-siderite occurrence of Silmänpaistama. Further west, at Pahtavaara, this horizon forms an anticlinorium which further on, at Jänesvaara, changes to a synclinorium. The axis of the system dips 40°—70° northward. There is no reason to suppose that the iron ore-jaspilite horizon would have formed any homogeneous, coherent, widespread bed on the contrary this horizon consists of numerous small-scale, lenticular, trough-filling occurrences of jaspilites and oxide iron ores, the location of which is limited to the immediate vicinity of cracks and fissures utilized by emanations. This means that any iron ore horizon does not necessarily exist on the bottom of the syncline between Porkonen and Pahtavaara. The geophysical maps, which unfortunately cannot be published here, clearly show that Pahtavaara and Jänesvaara are connected with each other by a narrow bipartite ore-free zone. A contradictory feature is the change of the synclinorium of Jänesvaara to an anticlinorium, consisting of younger material, at the Karjakkojoki dam. This anticlinorium consists of small, very gentle, folds with the axial plunge of 10°—15° to the south.

The anticlinorium is cut by the dislocation zone of Lomakuru. The formation of this dislocation has caused a kind of feather jointing and shearing at Kuolavaara and part of the beds have been tilted to an almost vertical position, probably by drag folding. Farther south the schist beds lie in a rather flat position striking west and dipping 10°—40° to the south.

The above thoughts on the structure of the area may not be considered a final report, but only a rough working hypothesis and a starting point for a thorough tectonic analysis which would be a very interesting, but at the same time very difficult, subject of research. The aim of this chapter is to show that the character of folding and deformation differs considerably from what is usually described to be characteristic of the metamorphic bedrock of Finland.

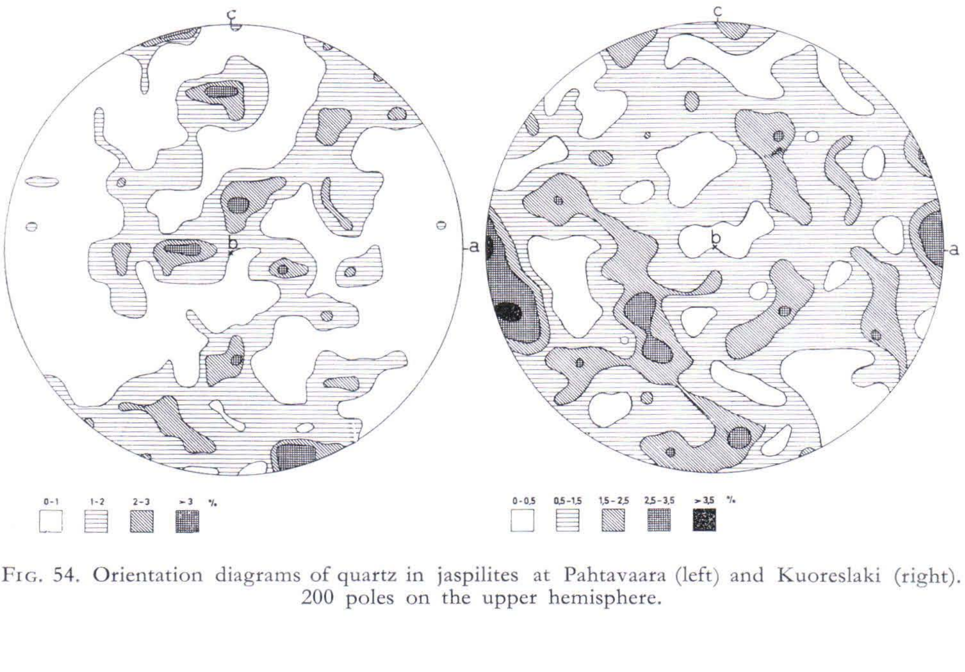


FIG. 54. Orientation diagrams of quartz in jaspilites at Pahtavaara (left) and Kuoreslaki (right). 200 poles on the upper hemisphere.

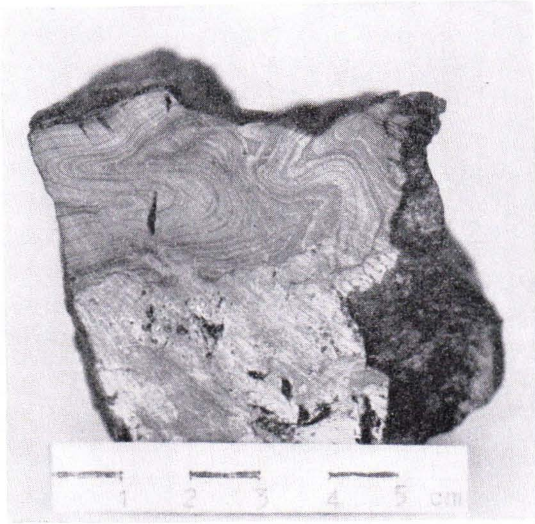


FIG. 55. Isoclinal plastic folding with axial plane cleavage in a tuffaceous schist penetrated by a quartz-siderite vein. The specimen is cut perpendicular to the axial plane.

Folding has not been accompanied by high, or even moderate, grade of metamorphism and recrystallization. The movements within the folding have been differential. This is proved by well-preserved primary structures and by random orientation of jaspilites (Fig. 54).

In the gently deformed schists it is easy to see the nonaffine deformation (in the sense of Metz, 1957) along two swarms of parallel glide planes (Fig. 22) which by further development changes to an axial plane cleavage (Fig. 55).

## STRATIGRAPHIC POSITION OF THE VOLCANIC COMPLEX IN THE SCHIST AREA OF CENTRAL LAPLAND

Among the subareas of the Karelian schist belt in Finland the stratigraphy and the age relations of the rocks in the schist area of central Lapland are the least known.

The first geologic mapping of the schist area of central Lapland was done in the years 1898—1907, and the map was published by the Geographic Society of Finland. Further investigations were made by V. Hackman in the years 1920—1924 especially in the central parts of Kittilä, and the results were published in 1927.

The scheme of Hackman was based on the stratigraphic concepts of that time, created mainly by Sederholm for the Karelian schist belt. Five years after Hackman's paper Sederholm (1932) put forward a stratigraphic scheme radically different from that of Hackman. Sederholm's opinion was that Hackman's pre-Jatulian schists, including the quarz-banded iron ores of the Porkonen-Pahtavaara area and the greenstone formation of Kittilä, do not belong to the Karelian cycle, but form an independent, older, Lapponian cycle. Thus, according to Sederholm, only the youngest, Jatulian, sediments of Hackman, that is, the Kumpu quartzites and their basal formations, would belong to the Karelian cycle.

Sederholm's stratigraphic scheme for the southern parts of Karelian schist belt had been under critical discussion since 1898. The discussion was especially concerned with the age relations of Kalevian and Jatulian and the existence of the discordance between them.

Mikkola (1941) rejected the terms Kalevian and Jatulian, because the schist belt of central Lapland differs from the more southerly Karelian areas so greatly that the results obtained within the latter were not easily applicable to Lapland. In his interpretation of the stratigraphy of central Lapland Mikkola was very cautious and did not present his ideas in the form of a scheme. In any case he divided the rocks of central Lapland into three stratigraphic units which were, beginning from the oldest: the complex of gneiss granite and granite gneiss, the Lapponium, and the Kumpu-Oraniemi series. The Lapponium, lying discordantly on the gneiss granite complex, consist of sedimentogeneous schists and volcanics which are mainly younger than sediments. Lapponium is overlain by the Kumpu-Oraniemi-series consisting of coarse-grained quarzites and conglomerates.

Inclining to Mikkola's opinion Sahama (1945) presented the following scheme of the stratigraphy of central Lapland:

| | | |
|---|---|---|
| | Younger granites | |
| | Granulite formation | |
| Younger | Hetta-granites | |
| | Kumpu-Oraniemi series | |
| | | Volcanics etc. |
| | Lapponium | |
| | | Quartzites and schists |
| | Sillimanite gneisses of Western Lapland | |
| Older | Gneiss granites and granite gneisses | |
| | Tuntsa-Savukoski series. | |

The scheme is in need of a couple of comments. First, according to the unpublished radiometric datings, made by O. Kouvo during last years, the granulites of Finnish Lapland (see Fig. 1) seem to be of Svecofenno-Karelian age. Secondly, according to the field observations of the present author, at least a part of the Hetta granites are penetrated by the basic intrusives of the Lapponium.

During the 1960's Mäkelä, participating in the prospecting work of the Outokumpu Company, had a chance to investigate large areas in western Kittilä. In his Licentiate Thesis Mäkelä (1968), inclining towards the view of Piirainen (1968), calls lowest orthoquarzites and aluminium-rich sediments an evolutionary-transgressive sediment series which is separated from the following revolutionary-transgressive sediment series by initial-magmatic volcanism. The latter series is further overlain by flysch and molasse sediments.

The distinctive features of the volcanism of central Lapland's supracrustal series are so characteristic of the initial magmatism in the sense of Stille, Aubouin and others that they hardly need further discussion. Thus we can conclude that the volcanic complex represents the preflysch phase in the eugeosynclinal succession. The sediments underlying the volcanic complex are of foreland or of miogeosynclinal character. The volcanic complex is overlain by flysch sediments especially in western Kittilä. The sedimentation was ended by accumulation of coarse-grained molasse quartzites and conglomerates.

Largely on the basis of earlier works, the present author suggests a tentative scheme for the stratigraphy of central Lapland.

| PHASE | CHARACTERISTIC ROCKS<br>Sediments | <br>Igneous rocks | SERIES |
|---|---|---|---|
| | | Late-orogene granites<br>(Nattanen granites) | |
| Molasse | Conglomerates and coarse-grained arkose quartzites | | Kumpu-Oraniemi |
| | The main phase of folding | Synorogene granites<br>(Hetta granites) | |
| Flysch | Conglomerates,<br>mica schists and<br>phyllites | Gabbros and<br>dolerites | |
| Preflysch | Pyroclasts, jaspilites,<br>carbonaceous schists and<br>iron formation | Spilitic<br>greenstones | Lapponium |
| Miogeosyncline and foreland | Aluminous mica schists, mica quartzites, arkose quartzites, orthoquartzites | | |
| | Discordance | | |
| Formely cycle | Paragneisses | Orthogneisses | Basement |

The schist area of Central Lapland gives an impression of an unfinished stage in the development of an orogenic cycle. The development of the geosyncline has gone rather far but the orogenic movements with associated metamorphism and acid plutonism have remained incomplete.

The lithological differences between the subareas of the Karelian schist belt in Finland are so conspicuous that there is hardly any reason to consider that this belt would represent a homogeneous geological unit neither in lithological nor in time-stratigraphic sence. In this respect very little has been set forth almost until recent days.

As yet there are no absolute age determinations of the initial volcanism of different basins of the Karelian schist belt in Finland, and may be the accuracy of the age determinations is not good enough to support the opinion that different basins may really have been developed at different times. Naturally there are similarities between different basins in great lines of their geological development, but the conspicuous local differences prove for different, and variable, conditions and probably for different times of the formation of the basins in the manner which is well known from the stories of the development of younger orogenic belts.

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# OUTCROPS AND THE MOST IMPORTANT TECTONIC OBSERVATIONS OF PORKONEN - PAHTAVAARA AREA, FINNISH LAPLAND. (An overlay for the bedrock map).

by Juhani Paakkola, 1970.

Scale 1: 50 000

0 1 2 km

## EXPLANATION TO THE SYMBOLS

Outcrop.

Outcrop.

20 40 Schistosity and lineation.

Vertical schistosity.

Bedding.

Vertical bedding.

Fold axis.

# BEDROCK MAP
## OF PORKONEN-PAHTAVAARA AREA, FINNISH LAPLAND.

by Juhani Paakkola, 1970.

Scale 1:50 000

0 1 2 km

Base-map reduced from incomplete aerial photographic maps scale 1:20 000. Coordinates inaccurate in regard to topographic figures.

LEGEND

- Ultrabasic rocks.
- Extrusives.
- Extrusives, rich in carbonate.
- Hypabyssal intrusives.
- Tuffaceous schists, partly graphite bearing.
- Jaspilites.
- Oxide iron ores, connected with jaspilites.
- Manganosiderite schists, connected with jaspilites and tuffaceous schists.
- Carbonate rocks, mainly dolomitic.
- Coarse grained quartzites, partly conglomeratic.
- Pillow lavas.


*Juhani Paakkola:* The volcanic complex and associated manganiferous iron formation of the Porkonen—Pahtavaara area in Finnish Lapland


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