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On the geology of the copper  
ore field in the Virtasalmi area,  
eastern Finland

by Lauri Hyvärinen

Geologinen tutkimuslaitos • Otaniemi 1969



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ON THE GEOLOGY OF THE COPPER ORE FIELD  
IN THE VIRTASALMI AREA,  
EASTERN FINLAND

WITH 44 FIGURES AND 17 TABLES IN TEXT AND ONE APPENDED MAP

BY

LAURI HYVÄRINEN

GEOLOGINEN TUTKIMUSLAITOS  
OTANIEMI 1969

## ABSTRACT

The rocks of the Virtasalmi area were studied petrographically. Geophysical maps were used in determining the structure of the poorly exposed bedrock and the rock boundaries. The conditions prevailing during the metamorphism were determined by comparing the mineral associations of the crystalline schists to experimental data on the stability limits of minerals. Plutonic rocks are presented as a suite from the peridotites to the trondhjemites. The classification is mainly based on the mineral compositions of the rocks.

Numerous copper deposits of the same general type were discovered in the district. The occurrence at Hällinmäki was studied in detail. The crystallization of the sulfides is discussed in the light of recently published experiments.

## PREFACE

Since 1963 the Exploration Department of the Geological Survey of Finland has been carrying out extensive regional surveys in and around the district of Virtasalmi. The present author has been responsible for the geological investigations conducted in this area. As a result of the prospecting work, the copper deposit now being mined at Hällinmäki was discovered, along with numerous other, smaller occurrences of copper ore. The field work involved in the present study was done mainly in the years 1964—1967. The microscopic and other laboratory studies were performed during the winters of 1957—1967 and the manuscript written in 1968. During both the field operations and the writing stage, I received valuable support and advice from my immediate superior, Dr. Aarno Kahma, head of the Exploration Department, to whom I express my warmest thanks.

One gratifying feature of the operations carried out in surveying the Virtasalmi district was the cooperative spirit that prevailed among the persons engaged in the work. Thus it gives me pleasure to acknowledge the support given me by my colleagues the late Erkki Lyytikäinen, and Jouni Pekkarinen and Eric Lindberg. To Dr. Toivo Siikarla I am indebted for processing and interpreting the results of the geophysical research.

My heartiest thanks are also due to Drs. Erkki Aurola, Raimo Lauerma and Atso Vormaa, with whom I had many fruitful discussions in the field and from whom, in addition, I received many worthy suggestions with regard to my manuscript.

I am grateful to Dr. Vladi Marmo, director of the Geological Survey of Finland, who kindly accepted my manuscript for publication in the series *Bulletin de la Commission Géologique de Finlande*.

Further, I wish to express my appreciation to the many persons who helped me at various stages of this work. The English translation was done by Mr. Paul Sjöblom, M. A. Dr. Brian Gulson helped me in revising the technical terminology.

Finally, I must acknowledge with gratitude the financial aid given me by Suomen Kulttuurirahaston Mikkelin läänin rahasto (The Mikkeli Section of the Cultural Foundation of Finland).

Otaniemi, May 1969

*Lauri Hyvärinen*



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## INTRODUCTION

The incentive for the survey project in the Virtasalmi district (Fig. 1.) was given by ore specimens sent to the Geological Survey of Finland by local inhabitants. Up to the year 1960 about a dozen such specimens had been received from the area. All the specimens had come from boulders, the northernmost of which were located in the area between southeastern end of Virmasjärvi and Narila (Fig. 2.). The rest of the boulders were located in a tract about 50 kilometers long, farther to the south-east. Previously, fairly extensive investigations had been carried out in the near surroundings of some of these boulders (Pääkkönen 1954). The provenance of the boulders could not be traced at that time.

It was in 1960 that my attention was drawn to the fact that the special, reddish-brown color of the garnet of the skarn boulders containing copper minerals was the same in every boulder. Microscopic and x-ray investigations proved that the garnet in all the specimens had a similar chemical composition. It therefore seemed possible that the ore boulders originated in the same place or in occurrences that had been created under the same kind of geological conditions.

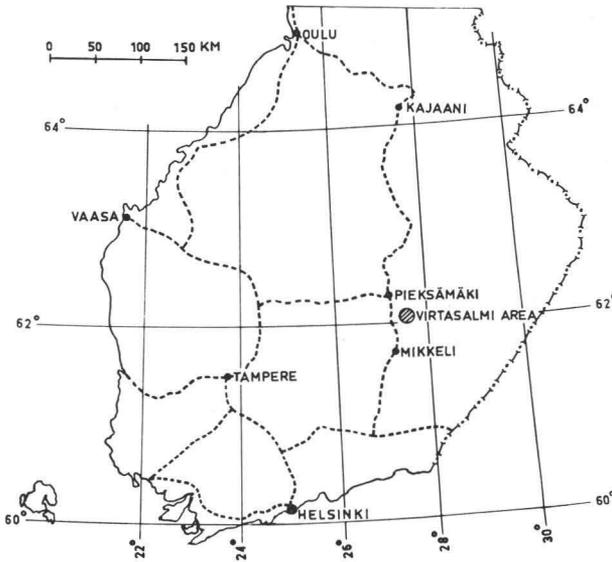


FIG. 1. The location of the exploration area.

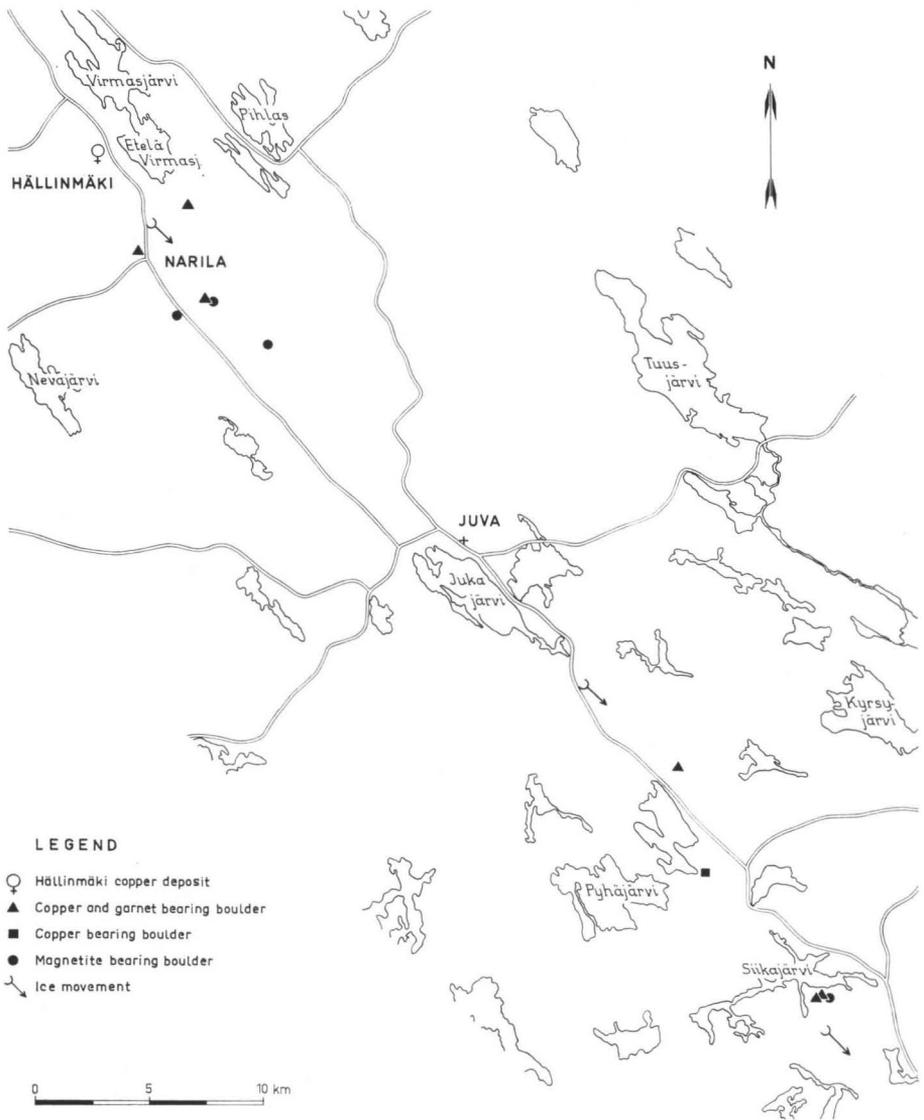


FIG. 2. The ore boulders known in the communes of Virtasalmi and Juva up to the year 1960.

In order to trace the source of the ore boulders, pedogeochemical methods were used in addition to the boulder investigations. No copper could be found in the fine-grained portions of the till in the surroundings of the boulders located to the southeast of Narila. Closer to Narila and approaching the southeast end of Virmasjärvi in a northwesterly direction, the copper content of the fine-grained till increased

appreciably. Thus, on the basis of both the boulder investigations and the preliminary geochemical analyses, it became clear that the field work should be principally concentrated in the vicinity of Narila and the area to the north of it. In the years 1960—1962, aeromagnetic maps of the Virtasalmi survey area were completed. On the basis of these maps, previous field observations (Pääkkönen 1954) and new reconnaissance, it became possible to delineate the critical area. It was now decided to subject the magnetic anomaly sequence extending from Narila to Pieksämäki to detailed study.

Geological mapping and geophysical measurements were started in the vicinity of Narila in the summer of 1963. Up to the end of 1968 geophysical measurements had been performed over an area of some 350 square kilometers. In 1964 systematic geochemical studies and an intensified search for boulders were added to the survey program. In October 1964 the present author discovered the Hällinmäki copper deposit during geological mapping operations. Since then many special investigations have been carried on this site as well as in the areas of other copper occurrences.

The present study deals with the petrography, mineralogy and structure of these ore deposits and the surrounding bedrock. The genesis of the rocks and ore deposits is discussed in the light of recently published experiments on silicate and sulfide systems.

## GEOLOGICAL INTERPRETATION OF THE GEOPHYSICAL MAPS

The Virtasalmi area is included in the general geological map published by Frosterus (1903) on a scale 1 : 400 000. Because of the small scale of the map and the research conditions prevailing at the time, the area had to be re-mapped.

The appended map covers an area of approximately 200 square kilometers. On it the outcrops observed, some 1 000 in number are marked with crosses. The tectonic and geological observations concerning the outcrops are indicated on the map next to each outcrop. The mapping has generally been done on a scale of 1 : 2 000, but in the quartz diorite and mica gneiss areas a scale of 1 : 10 000 or 1 : 20 000 has been used. A large part of the outcrops has been bound to a stake-line system laid out for the geophysical survey. There is an average of five outcrops per km<sup>2</sup>. Owing to the scanty occurrence of outcrops and their uneven distribution, there are broad stretches of the territory surveyed without a single exposure. The accumulation of the till in the form of drumlins almost completely prevents detailed photogeological interpretation of the structural features of the bedrock. Drawing a sufficiently detailed and reliable geological map of the kind generally needed in ore prospecting is thus hardly possible on the basis of outcrop observations alone.

Geophysical measurements carried out on the ground greatly facilitate both the mapping of the bedrock and the interpretation of its structure. Such measurements have been performed along lines laid out 50 or 100 meters apart, the distance between observation points being 20 m. An average of about 700 points per km<sup>2</sup> has been measured electromagnetically and magnetically, the corresponding number gravimetrically investigated being c. 670. The results of the measurements have been incorporated in maps drawn on the scale of 1 : 2 000. These maps have served as the basis for drawing combination maps on a scale of 1 : 10 000, the latter again being reduced in scale to 1 : 20 000 and 1 : 50 000 (Siikarla 1967).

In order to facilitate and increase the accuracy of the interpretation of geophysical maps, measurements have been made of the density and susceptibility values of the different rocks. The values presented in the following are those determined by Siikarla (1967).

The general geological structure of the Virtasalmi district takes shape even on a magnetic map. However, according to the laboratory studies the same susceptibility categories include different kinds of rocks and, on the other hand, the same variety of rock is apt to yield varying values. For example, the amphibolites, have susceptibility with values ranging from  $74 \times 10^{-6}$  to  $9\,700 \times 10^{-6}$  cgs/cm<sup>3</sup>. The determination of the boundaries of rocks and the structural features of the bedrock solely on the basis of a magnetic map would thus lead to misinterpretations. Gravimetric maps likewise have their flaws in this respect because a number of different rocks may belong to the same density class. Used together, the various geophysical maps suitably complement each other, with the result that a reliable picture of the bedrock emerges.

By electromagnetic methods, conductible sulfide- and graphite-bearing gneisses possessing the same magnetic intensity can be easily differentiated from non-conductible amphibolites without any sulfide content. It has therefore been possible to differentiate, for example, the sulfide-bearing gneisses running northwest from Narila from the amphibolites occurring in the area (see appended geological map). In places the amphibolites, too, are apt to be conductible owing to their sulfide content. In addition to the aforementioned techniques, it is possible to distinguish them from the sulfide-bearing gneisses by gravimetric methods. For example: the sulfide-bearing amphibolite located on the west side of Tervalampi (density  $\approx 3.0$  g/cm<sup>3</sup>) is conductible, but its separation from the sulfide- and graphite-bearing gneisses (density  $\approx 2.8$  g/cm<sup>3</sup>) of the surroundings has been possible by means of the gravimetric maps.

It has been relatively easy to follow the long limestone zone of the Virtasalmi district with the aid of geophysical maps by virtue of the fact that the limestones are not magnetic ( $4\text{--}30 \times 10^{-6}$  cgs/cm<sup>3</sup>) and, moreover, are located in the minimum zones of the gravimetric map. The last-mentioned phenomenon is due to the fact that the limestones have a lower density than the other rocks in the area, besides which they are covered by a thick, weathered, surface layer. Diamond drilling on the site of the Hällinmäki copper deposit, for example, revealed the existence of an intensely weathered layer not less than 50 to 60 meters thick, overlying the limestone sequence.

The periodotites and gabbros are generally characterized by great densities ( $\approx 2.9$  g/cm<sup>3</sup>) and low susceptibility values ( $66\text{--}263 \times 10^{-6}$  cgs/cm<sup>3</sup>). It is thus difficult in areas with few exposures to determine the boundaries between basic plutonic rocks and weakly magnetic amphibolites. The form of the anomalies, however, has helped. The form of the positive gravimetric anomalies induced by plutonic rocks is generally roundish, whereas the anomalies induced by amphibolites are long and narrow.

The large diorite and quartz-diorite massifs are magnetically non-homogeneous. According to laboratory findings, for instance, the susceptibility values of these rocks vary between the limits of  $19 \times 10^{-6}\text{--}4\,054 \times 10^{-6}$  cgs/cm<sup>3</sup>. Nevertheless, it is possible by means of gravimetric procedures to distinguish the dioritic massifs from the amphibolites owing to lower density of the dioritic rocks (e.g., average density of the dominant quartz diorite is 2.74 g/cm<sup>3</sup>).

Differentiating gneisses lacking sulfides and quartz diorites by geophysical means is exceedingly difficult. Hence, in areas lacking outcrops, inaccuracies are liable to occur in determining the boundaries between these rocks.

Determining the attitude of the regional fold axis on the basis of the minor fold axis observations alone has not been possible. A fairly reliable idea of the strike and plunge of the regional axis can be obtained by comparing the fold axis, foliation and bedding observations with the data recorded on the geophysical maps. This is particularly true of the central portion of the area studied.

A magnetic and gravimetric anomaly zone extends from Narila to Litmanen (Siikarla, 1967, appended maps). In the southern part, the anomalies consist of two parallel belts. The gravimetric anomaly, furthermore, has a break in its northwestern section. A break can also be observed on the detailed magnetic map in the zone formed by anomalies exceeding 1 000  $\gamma$ . A positive gravimetric anomaly is created by a diopside amphibolite of higher density than its surroundings and a magnetic anomaly by one possessing higher susceptibility values. The regional axis would appear, according to the geophysical maps, to be nearly horizontal in the middle of the anomaly zones, plunging, on the basis of fold axis observations, 40—60° SE at the southeastern end and 40° NW at the northwestern end. The form of the anomalies and the fold axis observations would thus indicate an anticline. The break in the gravimetric anomaly and the weakening of the magnetic anomalies at their northwestern end is probably due, according to the fold axis observations, to the culmination of the regional axis, with the result that underneath the diopside amphibolite there has become exposed a lighter rock, mica gneiss. In the light of the foregoing data, the regional axis thus probably runs in a NW-SE direction and lies in a nearly horizontal position. The folding appears to be isoclinal except in the southwestern corner of the area, where it seems to be rather gentle (Fig. 3).

In the gravimetric map (Siikarla 1967), between Narila and Vuorenmaa there is an arch-shaped positive anomaly, which opens up to the southeast. The fold axis observations at the end of the arch and the form of the anomaly are suggestive of a synclinal structure.

#### GEOLOGICAL OUTLINE

Characteristic of the bedrock of the Virtasalmi district are zones of supracrustal rocks trending mainly in a northwesterly direction. Extensive massifs of plutonic rocks occurring in the middle of the district and in its northern section penetrates the schist complex while smaller massifs of these rocks are generally concordant with the bedding.

The proportions of the different rocks in the Virtasalmi district, measured with a planimeter from the geological map, are as follows:

mica gneiss .....	36 %
diopside gneiss .....	1 »
amphibolites .....	12 »
limestone .....	2 »
quartz diorite and diorite .....	48 »
gabbro and peridotite .....	1 »
	<hr/>
	100 %

About one-half of the bedrock consists in the area of stratified rocks. They are highly metamorphosed rocks, the original character of which is generally hard to

determine. The foliation is in most cases parallel to the bedding or banding except at the hinges of the folds, where it often intersects the bedding. The dip of the foliation — with the exception of the southwestern corner of the district — is steep,  $70^{\circ}$ — $90^{\circ}$  N—NE. The strike and the dip of the foliation have been affected locally by intrusion of plutonic rocks, which form massifs elongated parallel to the foliation.

Aside from the general isoclinal folding upon NW striking subhorizontal axes, discussed in the foregoing chapter, a great deal of minor folding occurs in the area. It probably represents drag folding on the limbs of the major folds. The axes of the minor folds usually have a steep or vertical plunge. The lineations, which occur as mineral parallelism on the foliation planes, in places follow, and in other places completely deviate, from the directions taken by the axes of the minor folds.

The contact relations between the rocks indicate that the supracrustal rocks are the oldest in the Virtasalmi area. No basal formations have been observed in the area. The investigation of the stratigraphy of the supracrustal rocks has proved exceedingly difficult because of the high grade of metamorphism and intensive folding. The inter-relations of the rocks, although not completely solved are in agreement with the data reported from southwestern Finland (Härme, 1954).

If the interpretation of the syncline and anticline (pp. 11—12) is assumed correct the chronological sequence of the supracrustal rocks from top to bottom is as follows (Fig. 3):

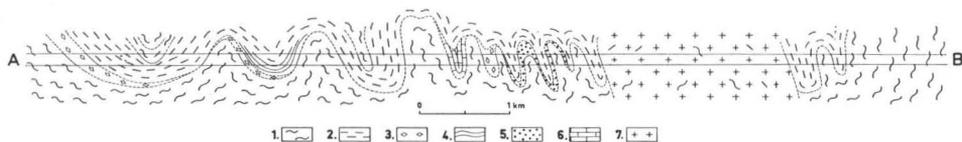


FIG. 3. Sketch showing the stratigraphy of the Virtasalmi area. A—B is a profile from the appended geological map. 1 = mica gneiss, 2 = amphibolite, 3 = pyroxene gneiss, 4 = hornblende gneiss, 5 = quartz-feldspar gneiss, 6 = limestone and 7 = quartz diorite.

- garnet-, cordierite- and sillimanite-bearing gneisses;
- amphibolites, the major part of which consists of diopside amphibolites with intercalations of hornblende gneiss, uralite-plagioclase porphyrite and garnet skarn. The copper occurrences of the district are invariably located in the garnet skarn;
- diopside- and quartz-feldspar gneisses with limestones;
- garnet-, cordierite- and sillimanite-bearing gneisses with graphite-bearing intercalations;
- unknown base of sedimentation.

The plutonic rocks are peridotites, gabbros, diorites, quartz diorites, trondhjemites and pegmatites. The peridotites and gabbros, and diorites and gabbros, are closely

associated and often grade into each other. The contact between the quartz diorites and more basic plutonic rocks is often clearly defined by a brecciated zone. The quartz diorites have been observed to grade into trondhjemites. The pegmatites and aplites intersect all the other rocks and are not folded.

## SUPRACRUSTAL ROCKS

### Metasediments

#### Mica gneisses

The structure of the mica gneisses is mostly that of veined gneiss. This is strongly folded and the primary structural features are rarely discernible. In the proximity of amphibolites, the mica gneisses usually contain intercalations of quartz-feldspar gneiss (p. 18), amphibolite (often fragmented), hornblende gneiss or diopside gneiss. In the contact zone also the amphibolites generally contain narrow mica gneiss intercalations.

The texture of the mica gneiss is generally granoblastic, although lepidoblastic and porphyroblastic textures do occur. The grain size of the mica gneiss averages 0.2 to 0.5 mm, except in the veined-gneiss varieties, in which the grains measure between 1 and 2 mm. In graphite-bearing varieties, the grain size changes from coarser (0.2 mm) in the graphite-poor beds to finer (0.02 mm) in beds rich in graphite.

TABLE 1

Mineral compositions of the mica gneisses in the Virtasalmi area. Determined by the point counting method. Vol. %.

	1.	2.	3.	4.	5.	6.	7.	8.
plagioclase .....	26.5	45.4	35.8	22.7	33.1	39.3	13.8	4.0
microcline .....	—	—	31.5	—	—	—	20.3	—
quartz .....	15.8	29.1	17.2	38.8	33.3	24.3	28.1	—
biotite .....	55.4	24.9	8.2	33.9	27.8	35.4	21.1	57.2
garnet .....	—	—	—	—	4.6	0.5	—	36.5
cordierite .....	—	—	—	3.6	—	—	—	—
sillimanite .....	—	—	5.9	—	—	—	—	—
graphite .....	—	—	0.4	—	—	—	6.7	—
accessories .....	2.3	0.6	1.0	1.0	1.2	0.5	10.0*	2.3
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

\*) mostly sulfides

- |  |                   |
|--|-------------------|
| 1. Mica gneiss                               | 176/EJL—64/Vrs    |
| 2. » »                                       | 41A/HA—66/Jv      |
| 3. Sillimanite-bearing, arkosic mica gneiss  | P 186/EJL—64/Vrs  |
| 4. Cordierite-bearing mica gneiss            | 102/HA—66/Jv      |
| 5. Garnet-bearing mica gneiss                | 161/HA—66/Jv      |
| 6. » » » »                                   | R 103Vrs/63.50 m  |
| 7. Graphite-bearing mica gneiss              | R 104Vrs/81.30 m  |
| 8. Garnet-bearing biotite-plagioclase schist | R 107Vrs/124.20 m |

The chief minerals of the mica gneiss are plagioclase, quartz and biotite, though their proportions vary greatly (Table 1). Common mica gneiss minerals are also muscovite, sericite, garnet and graphite. In addition microcline, sillimanite and cordierite may occur. Accessory minerals are apatite, zircon, rutile, sphene, epidote and calcite.

The anorthite content of the plagioclase is 20—25 per cent, but in the graphite-bearing mica gneisses the composition of the plagioclase is more basic (An 25—35 per cent). In the graphite-bearing varieties particularly, the plagioclase is brownish and altered to sericite.

The biotite is brown, strongly pleochroic, and in places, showing incipient alteration to chlorite. Partial chemical analyses (total iron as FeO + MgO) of certain whole rocks and their constituent minerals, biotite and garnet, are given in Table 2. In this table the FeO + MgO contents of the biotites from the mica gneisses are the same, whereas in the garnet-bearing mica gneisses they vary somewhat. However, with the exception of the sillimanite-bearing mica gneisses (which are generally richer in MgO and poorer in FeO), the biotites from the mica-gneisses have essentially the same composition.

With the exception of the graphite-bearing mica gneiss varieties, garnet is usually present in mica gneisses. In general, it occurs as porphyroblasts measuring 1 to 2 mm in diameter — and in rare cases over 5 mm. The large garnet crystals are sometimes surrounded by a zone almost devoid of biotite (Fig. 4). The garnet crystals regularly contain inclusions of other minerals. In the narrow, mica-schist-like intercalations

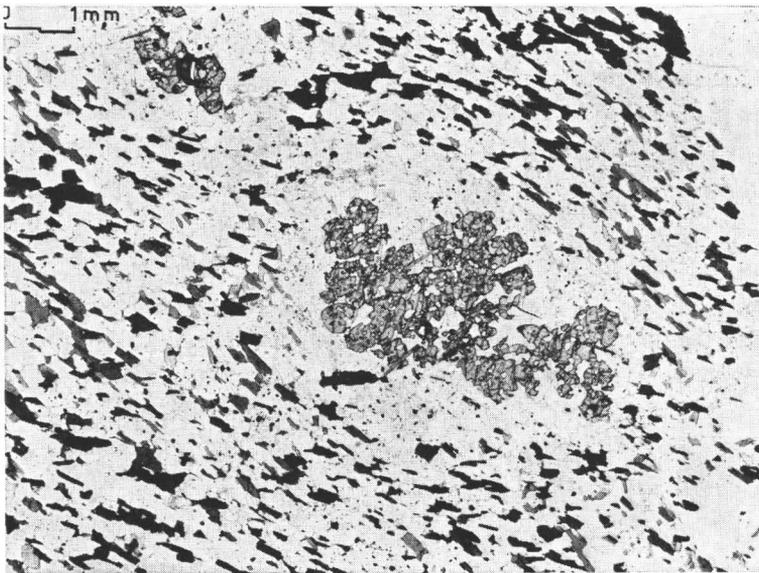


FIG. 4. Garnet-bearing mica gneiss. One Nicol. Photo E. Halme.

of the diopside amphibolite, there is often considerably more garnet than in mica gneiss in general (see Table 1). The composition of the garnet is the same in different parts of the district, even though the bulk composition of the rock may vary (Table 2).

Cordierite has been observed only in the mica gneiss inside the amphibolite arch on the west side of Tervalampi. The cordierite occurs as elongated crystals 1 to 2 mm long and parallel to the schistosity and contains tiny quartz or biotite inclusions. Sillimanite needles sometimes occur along the edges of the grains. A microprobe study performed by K. Laajoki revealed the cordierite (specimen 102/HA—66/JV) to contain  $\text{FeO} = 8.3$  weight per cent and  $\text{MgO} = 9.6$  weight per cent.

Sillimanite is not abundant; nevertheless, it is distributed over the entire district. It generally occurs in mica gneisses as cluster of aligned needles, parallel to the foliation. In most instances, the sillimanite is intergrown with muscovite and biotite (Fig. 5).

Potassium feldspar occurs only either in the immediate proximity of quartz-feldspar gneiss layers, in the light veinlets of veined gneiss or in graphite-bearing mica gneisses.

The original textures of the mica gneisses of the Virtasalmi district have generally been destroyed, as typical graywacke features have been observed in only one case (sample R 1/Vrs—64/194.00). But if the mineral compositions of the mica gneisses obtained from the point counting analyses are placed on the triangular diagram of Krumbein and Sloss (1951, p. 130), it would appear as if all the mica gneisses occurring

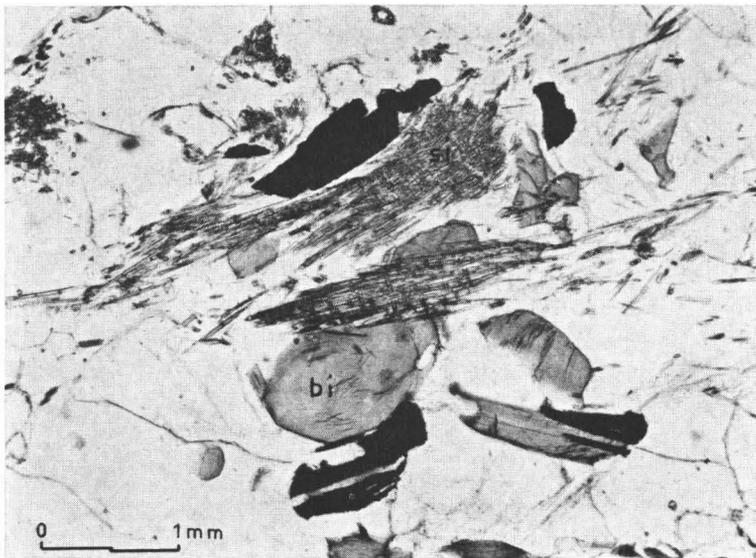


FIG. 5. Biotite scale (bi) pierced by a cluster of sillimanite needles (si). Pale minerals are quartz and feldspar. One Nicol. Photo E. Halme.

in the Virtasalmi area had originally been graywacke slates (Fig. 6). The chemical determinations also support this view. The mica gneisses of the Virtasalmi district, however, contain iron and, especially, magnesium in greater concentrations (Table 2) than do graywackes in general (Simonen 1953).

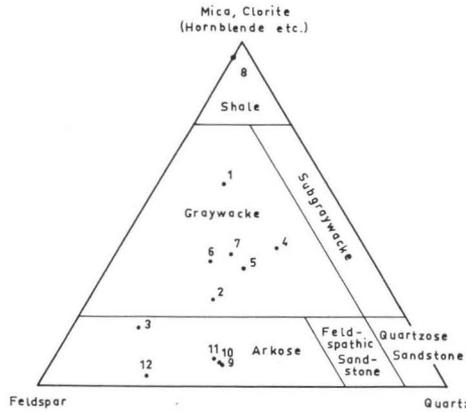


FIG. 6. Point counting analyses in Krumbein's and Sloss's (1951) triangular diagram. Nos. 1—8 denote mica gneisses, whose exact mineral composition appears in Table 1 and Nos. 9—12 denote quartz-feldspar gneisses, whose exact mineral composition appears in Table 3.

TABLE 2

Iron and magnesium contents of the mica gneisses and their constituents biotite and garnet. Rock analyses by P. Ojanperä, and mineral analyses by K. Laajoki. Wt %. Determinations of  $a_0$  and  $n$  by P. Kallio.

No.	Rock		Biotite		Garnet		CaO	Al <sub>2</sub> O <sub>3</sub>
	FeO*)	MgO	FeO*)	MgO	FeO*)	MgO		
1	7.66	2.60	19.1	12.2	32.4	4.0	—	—
2	5.80	2.19	18.1	11.3	32.1	4.5	—	—
3	5.70	2.82	17.8	10.5	32.4	4.8	2.2	21.0
4	7.09	3.52	17.0	11.9	32.4	4.9	—	—
5	5.93	1.80	18.3	9.4	32.2	4.1	—	—
6	6.52	2.03	20.6	9.0				
7	6.37	3.64	18.3	10.0				
8	9.40	4.97	17.5	10.4				
9	5.36	2.82	17.8	10.4				
10	5.25	2.65	18.4	10.3				
11	5.87	2.81	13.7	13.1				
12	7.33	3.15	16.2	11.4				
13	2.59	2.06	9.9	16.9				

$a_0$  and  $n$  values of the garnet of sample 1 are 11.55 Å and 1.80.

Composition of the garnet (No. 3)

almandine	75
pyrope	17
spessartite	5
grossularite	3

\*) total iron

100 %

- 1. Garnet-bearing mica gneiss
- 2. » » »

161/HA—66/Jv, in Table 1, No. 5  
R 106Vrs/30.00 m

3. Garnet-bearing mica gneiss	R 106Vrs/41.15 m
4. » » » »	Narila
5. » » » »	199 b/JP—64/Jv
6. » » » »	R 3Jv/44 m
7. Gordierite-bearing mica gneiss	102/HA—66/Jv, in Table 1, No. 4
8. Mica gneiss	176/EJL—64/Vrs, in Table 1, No. 1
9. » »	Narila
10. » »	R 7Jv/89.80 m
11. Sillimanite-bearing mica gneiss	R 3Jv/55.55 m
12. » » » »	R 7v/52.60 m
13. » » » »	P 186/EJL—64/Vrs, in Table 1, No. 3

The mica gneisses gradually change into veined gneisses in the Virtasalmi district. In the initial stages there appear in the mica gneiss small lenses of quartz and plagioclase surrounded by biotite. The idiomorphic plagioclase crystals contain inclusions of biotite and quartz. The myrmekite is present. In the later stages the lenses increase in size, whereupon they merge and finally produce veined gneiss, the main veinlets of which consist of biotite and quartz + plagioclase. When the last mentioned veinlets are 10 to 20 cm thick, tiny microcline grains are likely to occur between the plagioclase and quartz crystals measuring one or two centimeters in diameter. The microcline replaces the plagioclase.

#### Quartz — feldspar gneisses

The quartz-feldspar gneisses generally occur as relatively narrow and slightly reddish intercalations of varying thickness in the mica gneiss or amphibolite (Fig. 7). The thickest quartz-feldspar gneiss beds (100 meters) are found close to the copper

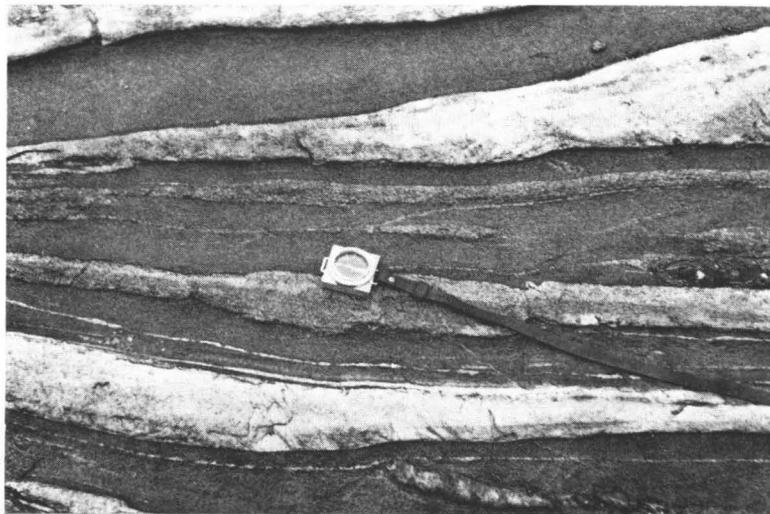


FIG. 7. Quartz-feldspar intercalations in mica gneiss.

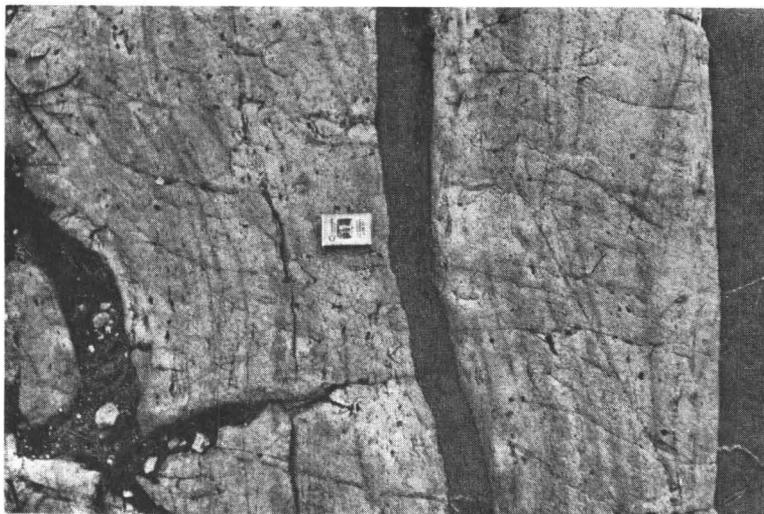


FIG. 8. Banded quartz-feldspar gneiss. The dark rock is mica gneiss.

occurrence of Lari. According to the data yielded by diamond drillings, in addition to the observations made of exposures and boulders, quartz-feldspar gneiss occurs mainly on the northeastern and eastern sides of the great limestone horizon.

The quartz-feldspar gneisses are in some places banded (Fig. 8), which is apparently due to bedding. No features indicative of volcanic origin have been observed. The texture is granoblastic. In the fine-grained types the grain size is 0.1—0.2 mm and in the coarse-grained ones, 1—2 mm. The fine-grained and coarse-grained types occur in some cases as alternating beds. The different beds vary somewhat in mineral composition.

The principal minerals of the quartz-feldspar gneisses are quartz, plagioclase and potassium feldspar. In addition to these minerals, biotite and muscovite are generally present. Accessory minerals are apatite, epidote, zircon, sphene, graphite, garnet and calcite. The mineral composition of some quartz-feldspar gneisses is presented in Table 3.

The plagioclase ( $An_{20-25}$ ) is generally — especially in the coarse-grained varieties — brownish and sericite-bearing. The potassium feldspar is microcline exhibiting a cross-hatched pattern; it apparently replaces the plagioclase, particularly in the coarse-grained variety. In such cases, myrmekite is also present. The microcline generally occurs in definite bands. The biotite is often partially chloritized.

The nature of the accessory minerals contained in the quartz-feldspar gneisses depends on the rocks surrounding them. The quartz-feldspar gneisses occurring in association with diopside gneisses and hornblende gneisses contain as accessory minerals sphene, epidote, hornblende, scapolite, carbonate and graphite. The quartz-

TABLE 3

Mineral composition of quartz-feldspar gneisses. Determined by the point counting method. Vol. %.

	1.	2.	3.	4.	5.
plagioclase .....	42.4	49.0	47.3	68.8	69.9
quartz .....	41.6	41.4	39.0	24.9	22.0
potassium feldspar .....	9.0	3.9	6.1	3.5	6.2
biotite .....	6.5	3.5	2.0	2.8	1.9
chlorite .....	—	1.4	—	—	—
accessories .....	0.5	0.8	5.6	—	—
	100.0	100.0	100.0	100.0	100.0

1. 3/JP—64/JV	Marked in Fig. 6 by	No. 9
2. R 100Vrs/136.90 m	»	» 10
3. R 103Vrs/34.40 m	»	» 11
4. 11/JP/65	»	» 12
5. 11/JP/65	pegmatitic dike penetrating	4

feldspar gneisses associated with mica gneisses contain garnet and mica minerals. This indicates that the original arkosic material (Fig. 6) was partly mixed with either calcareous mud or clay.

The quartz-feldspar gneisses are penetrated by numerous pegmatitic veins. Their contacts against the gneiss are generally indistinct. The middle portion of the veins have a coarser grain size than the margins. They are almost identical in composition to the host rock of quartz-feldspar gneiss (Table 3).

### Diopside gneisses

The diopside gneisses usually occur as intercalations in the mica gneiss in the proximity of limestone and diopside amphibolites. In texture they are granoblastic. The grain size averages between 0.3 and 0.6 mm, but the extensive diopside gneiss zone on the northwest side of Narila is composed of a rock coarser than usually, the diameter of the grains there measuring from 3 to 5 mm.

The principal minerals are quartz, plagioclase and diopside. The mineral compositions of the diopside gneisses are presented in Table 4. Accessory minerals present are epidote, sphene, potassium feldspar, biotite and ore minerals. The coarse variety contains an abundance of actinolite. The composition of the polysynthetically twinned plagioclase varies from 30 to 55 per cent An. In places it has undergone epidotization.

The index of refraction of the diopside has been measured from two samples from Hällinmäki:  $\alpha = 1.677$  and  $\alpha = 1.679$ . According to Tröger's (1959) tables, they contain 15—20 Mol. per cent of the hedenbergite component. The actinolite generally alters to diopside. The actinolite has  $\gamma = 1.646$ , so that, according to Tröger, it contains c. 30 Mol. per cent of the Fe component.

The generally accepted view is that diopside gneisses in the Finnish Precambrian are of sedimentary origin, being alteration products of carbonaceous sandstones (Simo-

TABLE 4  
Mineral composition of diopside gneisses. Determined  
by the point counting method. Vol. %.

	1.	2.
Quartz .....	38.6	2.7
Plagioclase .....	18.0	21.2
Diopside .....	38.8	71.1
Accessories .....	4.6	5.0
	100.0	100.0

1. Diopside gneiss 93/JP—64/Vrs

2. Diopside gneiss 93a/JP—64/Vrs

nen 1956; Härme 1960; Vorma 1965). The diopside gneisses of the Virtasalmi district may be depicted as having formed in the same manner as all the transitional forms between the mica gneisses and the quartz-feldspar gneisses can be observed in the field.

### Amphibolites and garnet skarns

The rock marked on the geological map in green is principally diopside amphibolite, which nevertheless contains an abundance of other rocks as intercalations. Among these, the amphibolite intercalations are a common occurrence. Numerous limestone beds are associated with the lower portions of the diopside amphibolite layers, as are the quartz-feldspar gneisses. Characteristic of the diopside amphibolites of the Virtasalmi district are zones composed almost wholly of garnet, and which may be hundreds of meters long and many meters wide. Hypersthene- and cumingtonite-amphibolite intercalations are sporadically distributed throughout the diopside amphibolite. In the proximity of the plutonic massifs, the diopside amphibolite is penetrated by numerous diorite or quartz-diorite dikes of varying thickness and these are usually concordant with the bedding.

The diopside amphibolite with its intercalations is strongly folded (Fig. 9).

#### *Diopside amphibolites*

The essential minerals of the diopside amphibolite are diopside, plagioclase and hornblende in addition to epidote or scapolite. The constituents combine to form various mineral associations typical of skarn rocks. It is noteworthy that the composition of the minerals varies according to the mineral associations. The diopside amphibolite is granoblastic in texture and the average diameter of the grains is 0.6—0.9 mm.

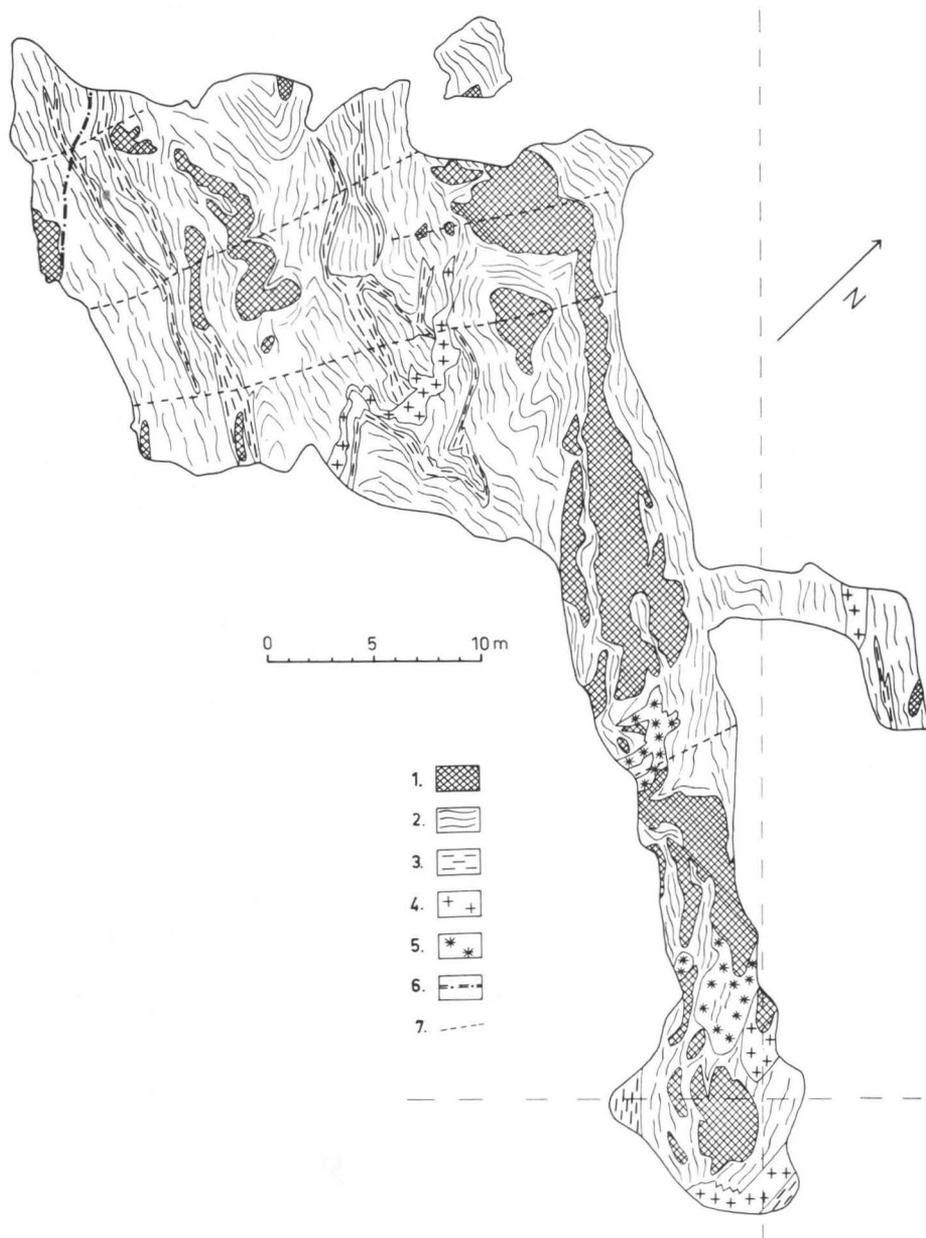


FIG. 9. Folded diopside amphibolite with garnet skarn zones, Hällinmäki. 1 = garnet skarn, 2 = diopside amphibolite, 3 = amphibolite, 4 = diorite, 5 = pegmatite, 6 = biotite amphibolite and 7 = fissure.

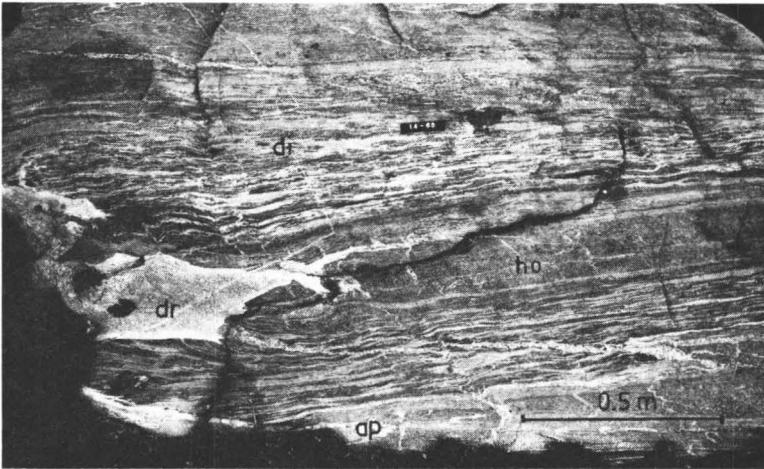


FIG. 10. Diopside amphibolite in which dark hornblende-bearing (ho) and light diopside-bearing (di) bands alternate. In addition, the figure shows diorite (dr) and aplite (ap). Hällinmäki. Photo E. Halme.

The diopside amphibolite is banded: between the dark bands, there occur light-colored lenses or bands (Fig. 10). The proportions of these dark and light layers vary considerably. The principal minerals of the dark portions are hornblende and plagioclase. The hornblende is a green variety, whose  $\gamma = 1.688$ . The anorthite content of the polysynthetically twinned plagioclase is 55–65 per cent. The diopside is pale and nearly colorless with  $\alpha = 1.690$  and contains, according to Tröger's (1959) tables, c. 40 Mol. per cent of the hedenbergite component. When the dark bands contain abundant hornblende with little or no diopside, the accessory minerals are generally apatite and sphene. If, on the other hand, the diopside content is abundant, the accessory minerals include, in addition to apatite and sphene, epidote, calcite, scapolite, vesuvianite, quartz and microcline. The mineral compositions of the dark and the light bands are presented in Table 5.

The principal constituents of the light-colored portions, plagioclase and diopside, have the same optical properties as those in the dark bands. The hornblende is very dark:  $\alpha = 1.680$  and  $\gamma = 1.696$ . It is an alteration product of pyroxene. The light portions contain the same accessory minerals as do the dark diopside-bearing ones. However, the abundance of epidote and scapolite is sometimes so great that they constitute the principal minerals (Table 5). Epidote is generally most abundant in the proximity of garnet zones. The pistacite content of the epidote averages 20 Mol. per cent; *e.g.*, one grain of epidote from Hällinmäki yielded the values  $2V_x = 85^\circ$ ,  $\gamma - \alpha = 0.028 - 0.30$  and  $X \wedge c = 2^\circ$ . These values correspond to c. 20 Mol. per cent of pistacite (Tröger 1959).

TABLE 5

Mineral composition of dark and light bands in diopside amphibolite. Determined by the point counting method. Vol. %.

	1.	2.	3.	4.	5.	6.	7.
Plagioclase .....	29.4	68.6	43.7	83.6	24.3	7.2	55.9
Diopside-hedenbergite ....	2.4	6.9	52.4	11.6	32.8	60.2	—
Hypersthene .....	—	—	—	—	—	—	29.9
Hornblende .....	65.4	23.8	1.8	1.8	—	—	7.2
Scapolite .....	—	—	—	—	29.8	—	—
Epidote .....	—	—	—	—	3.5	12.3	—
Others .....	2.8	0.7	2.1	3.0	9.6	0.3	7.0
Opaque minerals .....	—	—	—	—	—	20.0	—
	100.0	100.0	100.0	100.0	100.0	100.0	100.0

1. Dark band	263/JP/64/Vrs
2. » »	17/BS/65/Jv
3. Light band	R 4/22.05 m, Hällinmäki
4. » »	R 9/74.06 m, Hällinmäki
5. » » scapolite-bearing	90/JP—64/Vrs
6. » » epidote-bearing	R 7/66.38, Hällinmäki
7. » » hypersthene-bearing	R 107/110.17 m

Refractive index measurements indicate the scapolite to be almost pure meionite ( $\omega = 1.587$  and  $\varepsilon = 1.552$ ; Tröger 1959). Its contacts are sharp against the other minerals excepting plagioclase. In this case the boundary of the minerals is indefinite and the margins of the plagioclase grain contain more albite than the centres.

### Garnet skarns

The diopside amphibolite of the Virtasalmi area has in many places garnet skarn zones composed almost entirely of garnet. These zones are situated mainly in the lower part of the diopside amphibolite layers. The garnet layers were stretched to boudins during the folding (Fig. 11).

The garnet skarn zones are important from the standpoint of ore geology, for the ore deposits discovered in the Virtasalmi area are intimately associated with them. The biggest garnet skarn zones occur in the copper deposit at Hällinmäki, being from 100 to 200 meters long and 10 to 20 meters wide. As the mineral assemblages and compositions from the various parts of the area are similar, a description of the most thoroughly investigated garnet skarn zone, that of Hällinmäki, will be given.

The principal minerals contained in the garnet skarn zones are garnet and hedenbergite. In addition, there are varying amounts of epidote, plagioclase, scapolite and magnetite (Table 6). Accessory minerals are sphene, apatite, carbonate and sericite.

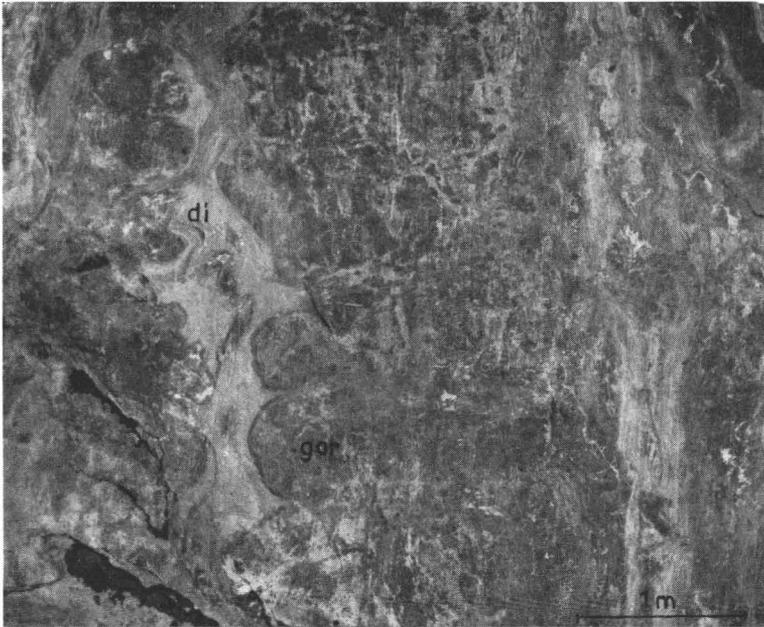


FIG. 11. Boudinaged garnet layer (gar) and diopside amphibolite (di). Hällinmäki. Photo E. Halme.

TABLE 6

Mineral composition of garnet-skarn zones at Hällinmäki. Determined by the point counting method. Vol. %.

	1.	2.	3.	4.	5.
Hedenbergite .....	21.9	29.2	1.3	9.7	67.3
Garnet .....	11.4	32.5	60.4	29.0	16.8
Epidote .....	29.7	14.1	3.2	—	—
Plagioclase .....	34.7	3.7	4.2	1.4	—
Scapolite .....	—	—	—	32.8	—
Quartz .....	—	—	—	26.8	—
Opaque minerals .....	—	20.4	30.9	—	14.5
Accessories .....	2.3	0.1	—	0.3	1.4
	100.0	100.0	100.0	100.0	100.0

1. Garnet skarn R 22/54.90 m  
 2. » » R 19/101.33 m  
 3. » » R 3/25.17 m  
 4. » » R 3/44.15 m  
 5. » » R 7/66.62 m

In addition to the almost monomineralic layers, garnet also occur as small lenses or disseminations. The garnet occurs throughout the area as idiomorphic or roundish crystals, the sizes of which vary according to the mode of occurrence. In the zones composed of the garnet, the crystals are generally one to three centimeters in diameter.

In the lenses or disseminations, garnet occurs as small grains averaging between two and three millimeters in diameter. Where the amount of garnet is small the diameters of the grains, are also small, averaging less than 1 mm.

The large garnet crystals generally contain inclusions of pyroxene, epidote and plagioclase. The ore minerals usually occur between the garnet grains or in fractures. Wherever there is an abundance of ore material, netlike structures are formed, in which the ore minerals separate the silicate minerals from each other.

The garnet crystals are in places anomalously anisotropic. The crystals in such cases exhibit twinning, which may be due to some internal strain in the crystals (Deer, Howie, Zussman 1962, Vol. 1, p. 81). It is possible that it has been caused by shear movements.

The chemical composition of the garnet has been analyzed from drill hole 65 between 250.50 and 250.60 m (Table 7). The density obtained for the same garnet sample is  $d = 3.87$ .

TABLE 7

Chemical composition of the garnet. Hällinmäki, drill hole 65/250.50—250.60 m. Analyzed by Marie Šmejkalová. Determinations of  $a_0$  and  $d$  by P. Kallio

	Analysis		Unit cell contents		
	I	II			
SiO <sub>2</sub> .....	36.48	36.58	Si .....	24.31	$a_0 = 11.97 \text{ \AA}$ $V_0 = 1715.06 \text{ \AA}^3$ $d = 3.87$
TiO <sub>2</sub> .....	0.50	0.50	Ti .....	0.25	
Al <sub>2</sub> O <sub>3</sub> .....	6.80	7.13	Al .....	5.48	
Fe <sub>2</sub> O <sub>3</sub> .....	21.94	21.84	Fe <sup>+3</sup> .....	11.0	
FeO .....	3.33	3.33	Fe <sup>+2</sup> .....	1.85	
MnO .....	0.56	0.56	Mn .....	0.31	
MgO .....	0.00	0.00	Ca .....	21.47	
CaO .....	30.22	30.00	O .....	97.47	
Na <sub>2</sub> O .....	0.03	0.03			
K <sub>2</sub> O .....	0.02	0.02			
H <sub>2</sub> O .....	0.06	0.06			
	99.94 %	100.06 %			

Cu = 0.00 %, Sr = 0.00 %, Y<sub>2</sub>O<sub>3</sub> = 0.00 %, ZrO<sub>2</sub> = 0.01 % and BaO = 0.00 %  
Analyzed by V. Hoffren

Almandine .....	7.7
Andradite .....	71.3
Grossularite .....	19.7
Spessartite .....	1.3
	<hr/> 100.0 %

In determining the indices of refraction of the garnet, it has been observed that several deviating values can be obtained within the range of a single grain in the Hällinmäki (Aurola 1966). Microscopically however, the garnet generally appears to be homogeneous. Zoned garnet has been observed at a depth of 259.55 meters in drill hole 25, and it has been noted that the central portion of the crystals is lighter in color than the marginal zones. A partial microprobe analysis (J. Siivola) of the dark border

area and the lighter center of one crystal showed the chemical compositions to be essentially the same, although there are variations in the Fe content:

	CaO	FeO + Fe <sub>2</sub> O <sub>3</sub>	Al <sub>2</sub> O <sub>3</sub>	Total
Dark portion . . . . .	31.7 %	29.5 %	5.8 %	67.0 %
Light portion . . . . .	31.4 %	26.6 %	5.8 %	63.8 %

The garnet of Hällinmäki is mainly an andradite and it shows irregular variations in different parts of the formation. According to differences in the indices of refraction (1.810—1.845) and in the values of  $a_0$  (11.94—11.98 Å), the variations in the composition of the garnet are mainly between the grossularite and the andradite components. The almandine and the spessartite contents in the garnet are low and their variations are also slight.

The other principal constituent in the garnet skarn is diopside-hedenbergite. The composition obtained for one pyroxene sample from Hällinmäki was 85 Mol. per cent of the hedenbergite component, since  $\alpha = 1.718$ ,  $2V_z = 61^\circ\text{--}63^\circ$  and  $Z \wedge c = 47^\circ$  (Tröger 1959). The pleochroism is Z = dark bluish green and Y = yellowish green.

The composition of the clinopyroxene changes sharply when the diopside amphibolite alters into garnet skarn. The pyroxene of the diopside amphibolite contains 40 to 45 Mol. per cent of hedenbergite, whereas the pyroxene of the garnet skarn contains 80 to 90 Mol. per cent of hedenbergite. In association with the ore mineralization, the hedenbergite has altered in some places to hornblende (Fig. 12).

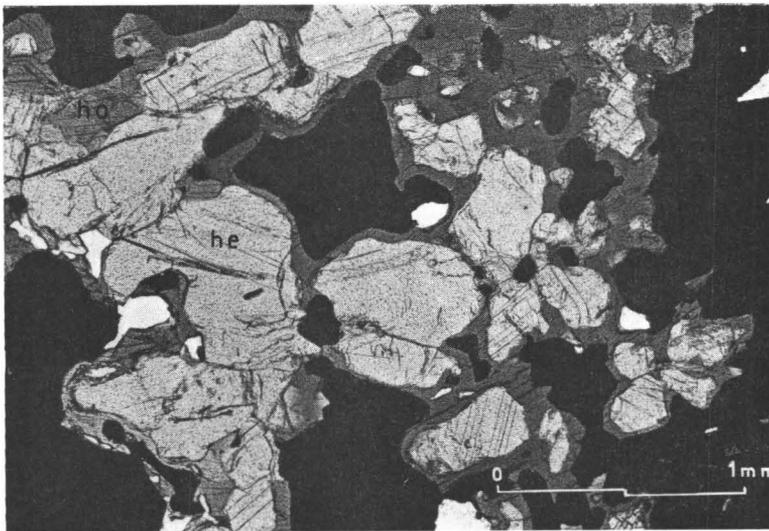


FIG. 12. Hornblende seam (ho) between hedenbergite (he) and ore (black).  
R 25/50, 60 m. One Nicol. Photo E. Halme.

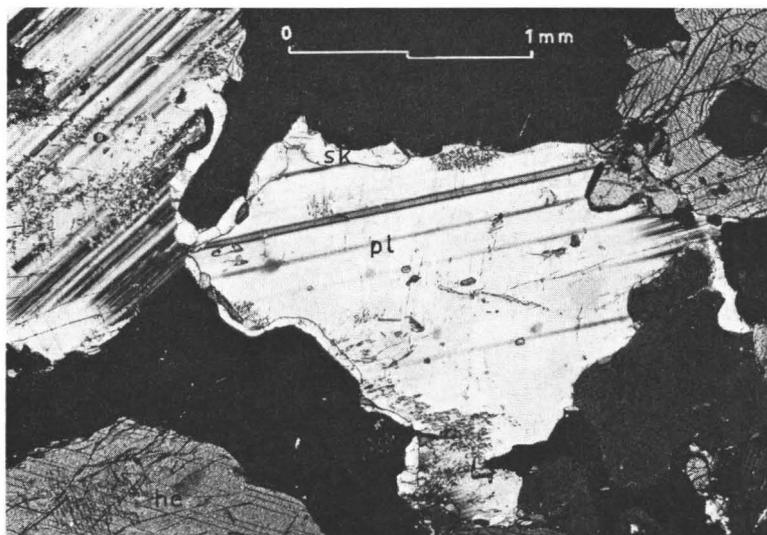


FIG. 13. Scapolite seam (sk) between plagioclase (pl) and ore (black); he = hedenbergite. R 3/29,35 m. Nic +. Photo E. Halme.

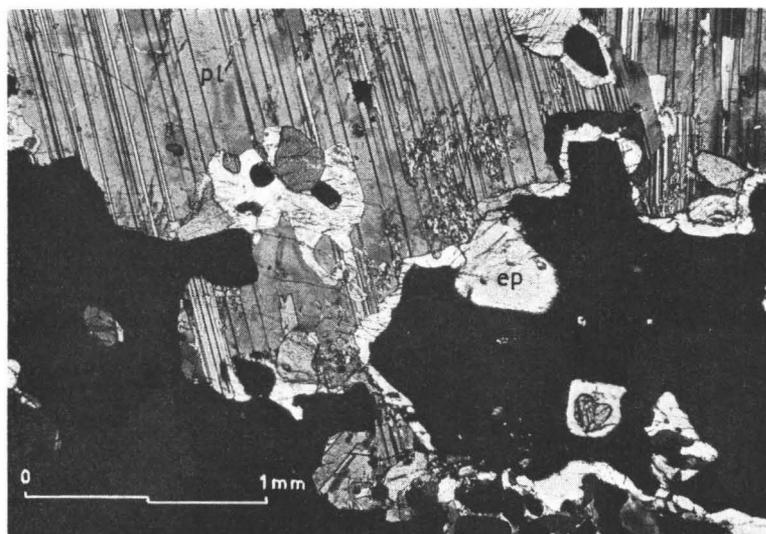


FIG. 14. Epidote seam (ep) between plagioclase (pl) and ore (black). R 3/29,55 m. Nic +. Photo E. Halme.

The plagioclase contains 60 to 70 per cent of anorthite. In conjunction with the ore mineralization, it is partly altered to epidote and/or scapolite (Fig. 13 and 14). Sphene generally occurs in the garnet zones in smaller amounts than in the diopside amphibolite.

Scapolite and epidote are often present in barren garnet skarn zones. In such cases, the scapolite and epidote were produced before the ore mineralization and as an event completely separate from the latter process. This is to be seen in, for example, drill hole 16 (231.00 m), where the ore veins penetrate the epidote (Fig. 15).

According to microscopic determinations ( $\omega-\epsilon$ ), the scapolite is dominantly meionite (>70 per cent), just as it is in the diopside amphibolite, (p. 24). The composition of the epidote varies, on the other hand, according to the  $2V_x$  determinations (Tröger 1959), from 30 Mol. per cent to 15 Mol. per cent. The epidote generated during the process of ore mineralization is richer in pistacite than the epidote generated during the regional metamorphism. However, according to Deer *et al.* (Vol. 1, p. 204), the  $2V_x$  values are unreliable in determining the composition of epidote. The compositions are thus subject to error, but they nevertheless reveal that epidotes created in different ways differ in composition.

Magnetite forms in places in the garnet skarn layers compact pockets a few meters long. The magnetite lenses are broken during the folding.



FIG. 15. Ore vein pierces garnet skarn. Ore is black. pl = plagioclase, gar = garnet and ep = epidote. R 16/231.00 m. One Nicol. Photo E. Halme.

*Amphibolites*

The amphibolites occur as intercalations of differing thickness in the diopside amphibolite. Contacts between the amphibolite and diopside amphibolite are diffuse. The thickest amphibolite layer, measuring between 100 and 150 meters, occurs on the northeastern side of Hällinmäki. The northeastern edge of the zone gradually changes into quartz diorite or diorite containing amphibolite fragments. The amphibolites are brecciated with folding and diorite or quartz diorite has in many instances penetrated the fissures.

In texture the amphibolite is granoblastic, the average diameter of the grains being 0.6—0.8 mm. Its principal mineral constituents are plagioclase and hornblende, along with a small amount of biotite. Accessory minerals are apatite and sphene. The relative contents of plagioclase and hornblende vary considerably (Table 8).

TABLE 8  
Mineral composition of amphibolite. Determined by the  
point counting method. Vol. %.

	1.	2.
Plagioclase .....	62.7	27.6
Hornblende .....	31.8	68.3
Biotite .....	4.5	3.9
Accessories .....	1.0	0.2
	100.0 %	100.0 %

1. R 5/35.00 m, Hällinmäki  
2. R 4/40.00 m, Hällinmäki

*Biotite amphibolites*

The bedding of the diopside amphibolite is transgressed by biotite amphibolite dikes, which are folded at the same time as the diopside amphibolite. The plutonic rocks of the area are still younger than the biotite amphibolite. The biotite amphibolite dikes range in thickness from 10 centimeters to 2 meters.

The texture of the biotite amphibolite is lepidoblastic. The average grain size of the plagioclase crystals is 0.4 mm and that of the biotite and the hornblende 0.7 to 0.9 mm. The principal mineral constituents are hornblende, biotite, plagioclase and, in one instance, anthophyllite. Accessory minerals are apatite, sphene, zircon and ore grains. The relative amounts of the various minerals vary appreciably in the biotite amphibolite samples taken from different localities (Table 9).

The anorthite content of the plagioclase is 50 per cent. The pleochroism of the biotite is  $Z = Y =$  dark brown and  $X =$  pale brownish yellow and its indices of refraction are  $\gamma = \beta = 1.626$ . The pleochroism of the hornblende is  $Z =$  green and

TABLE 9

Mineral composition of the biotite amphibolites of Hällinmäki. Determined by the point counting method. Vol. %.

	1.	2.	3.
Plagioclase .....	15.2 %	—	6.8 %
Biotite .....	24.2 %	19.5 %	6.3 %
Hornblende .....	56.3 %	80.3 %	64.0 %
Anthophyllite .....	—	—	22.7 %
Accessories .....	4.3 %	0.2 %	0.2 %
	100.0 %	100.0 %	100.0 %

1. Biotite amphibolite R 4/27.58 m  
 2. » » R 16/213.10 m  
 3. » » R 6/100.60 m

X = yellowish brown. The indices of refraction are  $\alpha = 1.626$  and  $Z \wedge c = 18^\circ$ . The pleochroism of the anthophyllite is X = colorless and Z = pale brown, and the refractive indices are  $\alpha 1.644$ ,  $\gamma = 1.670$ , and  $2V_z = 63^\circ$ . These optical measurements indicate that the anthophyllite contains 35 to 40 Mol. per cent of the  $Fe^{+2}$  component (Tröger 1959).

#### *Hypersthene amphibolites*

Narrow hypersthene-bearing amphibolites have been observed in the middle of the long diopside amphibolite zone on the west side of Hällinmäki. The chief minerals of this rock are plagioclase ( $An_{35-40}$ ), hypersthene and hornblende (Table 5). The hypersthene and hornblende occur as independent layers. The hypersthene occurs as grains c. 0.1 to 0.5 mm in diameter. Its  $\gamma = 1.724$  and  $Z =$  greenish and  $X =$  reddish. According to Tröger's Tables (1959), it contains c. 40—45 per cent of the  $Fe^{+2}$  component. The hornblende is dark green and its  $\gamma = 1.700$ . The hypersthene-bearing portions contain no diopside.

#### *Cummingtonite amphibolites*

On the north side of Narila, the diopside amphibolite contains a cummingtonite-bearing intercalation. The amphibolite is quite fine-grained, the diameter of the grains being c. 0.01 mm. The cummingtonite accounts for approximately 50 per cent of the components, the remainder being feldspar, with very minor amounts of biotite. The cummingtonite is usually twinned and its  $Z \wedge c = 17^\circ$ .

#### *Genesis*

The origin of the diopside- and hypersthene-bearing amphibolites of southwestern Finland are, according to Härme (1960), either basic volcanics or sediments

mixed up with partially weathered, basic volcanic material; in some places these sediments have also been carbonaceous. Metzger (1945) has distinguished two types among the diopside amphibolites of the Parainen region, one of which he regards as marl by origin and the other as magmatic. According to Eskola (1915), the diopside amphibolites of Orijärvi are metamorphic marls and clayey limestones. An analysis of the diopside amphibolite from Hällinmäki is compared in Table 10 with the diopside amphibolites analyzed by Metzger (1945) and Eskola (1915). The chemical composition of the samples 1, 2 and 4 are so alike that the rocks could have been derived from a similar type of material.

TABLE 10  
Chemical composition of diopside amphibolites.

	1.	2.	3.	4.		
SiO <sub>2</sub> .....	48.25	48.40	51.37	49.61	la)	
TiO <sub>2</sub> .....	0.63	0.63	1.35	0.56	Rb <sub>2</sub> O .....	0.00
Al <sub>2</sub> O <sub>3</sub> .....	14.00	14.02	7.97	15.21	SrO .....	0.02
Fe <sub>2</sub> O <sub>3</sub> .....	2.03	2.04	6.26	0.89	ZrO <sub>2</sub> .....	0.00
FeO .....	6.59	9.93	11.50	8.77	BaO .....	0.01
MnO .....	0.16	—	—	0.05		
MgO .....	8.19	7.62	5.82	5.02		100.37
CaO .....	18.46	12.75	12.03	16.32		
Na <sub>2</sub> O .....	1.20	3.21	2.98	1.20		
K <sub>2</sub> O .....	0.28	1.12	0.02	1.36		
P <sub>2</sub> O <sub>5</sub> .....	0.05	—	—	0.19		
CO <sub>2</sub> .....	0.00	—	—	0.86		
H <sub>2</sub> O <sup>+</sup> .....	0.47	0.40	0.24	0.55		
H <sub>2</sub> O <sup>-</sup> .....	0.03	—	—	—		
	100.34	100.12	99.84	100.59		

1. Diopside amphibolite. Virtasalmi, Hällinmäki. Analyzed by Aulis Heikkinen (1) and V. Hoffren (la)
2. Diopside amphibolite. Storgård, Parainen. Metzger (1945, p. 41).
3. Diopside amphibolite. Parainen. Metzger (1945, p. 41).
4. Diopside amphibolite. Vetjo, Kisko. Eskola (1915, p. 49).

In most instances, the diopside amphibolite in the Virtasalmi district is so highly metamorphosed as to make recognition of original structures uncertain. In the southern parts of the zone the diopside amphibolite shows a distinct banding of alternating hornblende- and diopside-rich beds. This banding suggests a sedimentary origin for the diopside amphibolites.

Since the diopside amphibolite clearly contains intercalations of material of volcanic origin (*cf.*, p. 37), the diopside amphibolites could have formed in the manner described by Härme (1960), *i.e.*, with the volcanic material and calcareous mud being deposited simultaneously. The possibility cannot be overlooked that diopside amphibolites were, at least in some places, marls. Notably in connection with the mica gneiss intercalations associated with the lower parts of the diopside amphibolite layers, the diopside amphibolite may have originally been marl, especially in view of the fact that structures of rocks of volcanic derivation have not been noticed there.

The amphibolite, hypersthene amphibolite and cummingtonite amphibolite were probably derived from a similar type of material.

The derivation of the biotite amphibolite is uncertain. One possibility is, however, that the biotite amphibolites were produced from the material that flowed into fissures in the calcareous sediments. The fissures could have been formed, for instance, through slumping in sedimentary state. The appreciable variation in the composition of the biotite amphibolite might thereby be attributed to the filling in of the fissures opened at different times with varying material.

Irregular hornblende lenses are scattered throughout the diopside amphibolite in the Hällinmäki area, as well as at the contact between this rock and diorite. Metzger has described (1945) similar hornblende lenses in diopside amphibolite. He attributes this phenomenon to segregation of the dark and the light portions by shear movements. Mikkola (1955) explains the ultrabasics in diopside amphibolite zones of the Orijärvi area as having formed by migration resulting from penetrative movements. It is also my view that penetrative movements could have caused the segregation of the light and the dark portions, although the heterogeneity of the diopside amphibolite is probably a sedimentary feature.

The presence of ore minerals in economic quantities in the garnet skarn zones of the Virtasalmi area permits their designation as »skarn ores».

There is a number of hypotheses to explain skarn ores. The general view is that the skarn ores were produced metasomatically by a reaction of metal halogenides and silica with limestone (Goldschmidt 1911, p. 214 and Eskola 1946, p. 267).

Another hypothesis is that skarn rocks and ores originated by the recrystallization of sedimentogenous rocks under conditions of regional metamorphism. According to Barth (1928, p. 1905), skarn rocks of Kristiansund were formed metasomatically during a period of regional metamorphism at the contact of the limestone and gneiss or pegmatite. Magnusson (1953, p. 44) suggests that a reaction, under increasing temperature conditions, between iron minerals (limonite, siderite, hematite) and silica and carbonates (calcite, dolomite) can produce ferriferous skarn minerals. Banas (1965, p. 60) has observed that in addition to the skarn generated during the older regional metamorphism, a new contact metamorphic skarn phase could form.

The author's view is that the skarn zones of the Virtasalmi area formed mainly as a result of regional metamorphism, with the sedimentogeneous rocks recrystallizing in the manner described by Magnusson (1953, p. 44). As the garnet skarn zones underwent ore mineralization, there was a simultaneous superimposition of the second skarn phase.

The history of the skarn rocks begins with the deposition of the clay or volcanic material. Simultaneously, limonite, siderite, calcite and possible silica were sporadically distributed throughout the sedimentary material. It is theoretically possible that the principal minerals of the garnet zones, andradite garnet and hedenbergite, can be derived from such material. Andradite-garnet may form by the following reaction:  $3\text{CaCO}_3 + \text{Fe}_2\text{O}_3 + 3\text{SiO}_2 \rightarrow 3\text{CaO} \cdot \text{Fe}_2\text{O}_3 \cdot 3\text{SiO}_2 + 3\text{CO}_2$ . Hedenbergite may

form by the following reaction:  $\text{CaCO}_3 + \text{FeCO}_3 + 2\text{SiO}_2 \rightarrow \text{CaO} \cdot \text{FeO} \cdot 2\text{SiO}_2 + 2\text{CO}_2$ . Andradite and hedenbergite do not, however, form quite so simply as these reactions indicate. For example: diopside does not form directly from its source materials, dolomite and quartz, as water also takes part in the reaction in the first stage, generating tremolite and calcite. When the temperature and the pressure rise to  $540 \pm 20^\circ\text{C}$  at 2 000 bars, the tremolite, calcite and quartz react to produce diopside (Winkler, pp. 23—45). If the source materials are ferriferous, ferroactinolite is generated instead of tremolite, and upon the reaction's continuing, hedenbergite develops instead of diopside (Deer, Howie and Zussman, Vol. 2, p. 68).

The magnetite lenses (p. 29) in the garnet skarn zones have probably originated also by regional metamorphism. Of the plutonic rocks of the Virtasalmi area, the magnetite apparently crystallized during the crystallization of the diorites and quartz diorites (*cf.*, p. 52). At least the magnetite analyzed from the quartz diorites deviates from the magnetite of the garnet skarn zones of Hällinmäki with respect to  $\text{TiO}_2$ , V and Cr so markedly that the last-mentioned magnetite can scarcely be regarded on this basis, either, as having derived from quartz diorite (Table 11).

TABLE 11  
Content of certain trace elements in the magnetites  
of the Virtasalmi area.

	1. E-shore of Litmanen, Virtasalmi	2. 27/JP-64, Virtasalmi	3. Hällinmäki, Virtasalmi
TiO <sub>2</sub> .....	1.01 %	0.78 %	0.3 %
V .....	0.26 %	0.45 %	0.02 %
Cr .....	0.12 %	0.05 %	0.002 %
Ni .....	0.007 %	0.018 %	0.012 %
Co .....	0.006 %	0.007 %	0.008 %

- 1 and 2. The magnetites have been separated from the quartz diorite; analyzed by P. Ojanperä and A. Löfgren. By courtesy of Dr. J. Haapala.
3. The magnetite analyzed has been taken from a lens containing nearly compact magnetite. By courtesy of Dr. R. Frietsch.

The fractures and cracks produced during the folding of the garnet zones became filled with ore-bearing solutions. The solutions reacted mainly with the plagioclase of country rock, producing epidote and scapolite (p. 29). Thereby a metasomatic skarn phase was created. Thus, the same zones have undergone two separate skarn-forming processes.

### Limestones

Limestone zones occur abundantly in the Virtasalmi district, especially in or near the contacts of diopside amphibolites and mica gneisses. In the light of the observations made from the drill cores and boulders, all the known limestone occurrences are alike. In most cases, these zones are only few meters thick. However, one

limestone clearly visible on the geological map is over 10 km long and from 300 to 400 meters wide. There is also a fairly large limestone occurrence on the northeastern side of the copper ore deposit at Hällinmäki (the name of which used to be Karsikumpu): only its southeastern end is known (Aurola 1966 and Virkkunen 1967).

The limestone of Hällinmäki may be classified mineralogically as a diopside-wollastonite-limestone. The main characteristic of such limestones is their variety of Ca-silicate minerals. In addition to diopside and wollastonite, they also contain epidote, scapolite, garnet, olivine, humite minerals, serpentine minerals, sphene, amphibole minerals, micas, quartz and occasionally chalcopyrite and magnetite grains.

In the southwest, the limestone zone is bounded by diopside amphibolite and in the northeast by amphibolite or diorite. Within the zone occur intercalations of diopside amphibolite and garnet skarn or, in certain instances, quartz feldspar gneiss. These intercalations vary in thickness from a few to several dozen meters. The contact zone between the limestone and the other rocks are, on the average, only about 1 cm thick. The contact zones between the limestone and the diorite and quartz dikes appear to have formed metasomatically, whereas the contact zones between the limestone and the schists may be attributed to reactions between the solid phases. The minerals of the contact zones are generally andradite garnet and hedenbergite. In addition there can also be wollastonite and epidote.

The typical limestone is almost pure calcite, as its MgO content, according to five analyzed drill holes, ranges from 0.10 to 1.76 per cent (this is equivalent to 0.2—3.68 per cent of  $MgCO_3$ ). For instance, drill hole 24 has yielded the following compositions for limestones (Table 12):

TABLE 12  
Chemical composition of the limestone. Drill hole 24. Hällinmäki. Aurola 1966.

Hole No. 24 depth interval	Length m	insol- uble	$R_2O_3$	CaO	MgO	Ignition loss	%
36.31—38.50 . . . .	2.19	2.92	0.44	53.6	0.30	42.6	99.86
38.50—43.40 . . . .	4.90	3.22	0.40	53.6	0.27	42.4	99.89
43.40—46.00 . . . .	2.60	9.48	0.78	50.2	0.14	39.0	99.60
46.00—48.00 . . . .	2.00	9.58	1.34	50.6	0.10	38.1	99.72
48.40—55.20 . . . .	6.80	6.98	0.74	51.2	0.47	40.4	99.79

Results of analyses calculated as carbonates

$CaCO_3$	$MgCO_3$	Silic.	%
96.1	0.6	3.4	100.1
95.8	0.6	3.6	100.0
88.4	0.3	11.0	99.7
86.4	0.2	13.1	99.7
97.7	1.0	9.1	100.8

The calcite is, on the whole, coarse, the grain size usually averaging 5 mm and in places, as much as 1 cm. In color it is generally white, but salmon-red has also been observed. The small amount of dolomite occurs within the calcite grains as tiny myrmekite inclusions.

The usual silicate mineral of the limestone is a monoclinic pyroxene of the diopside-hedenbergite group, and it generally occurs as grains measuring 0.2 mm in diameter in coarse-grained carbonate.

There is a regularity in the variation of the composition of the pyroxene, depending on its mode of occurrence. If the pyroxene occurs as tiny grains between calcite crystals or as inclusions in the calcite, it is invariably pure diopside. However, if the limestone contains other silicate minerals besides pyroxene such as epidote, garnet, quartz, wollastonite and scapolite, hedenbergite component increases.

Wollastonite is common in the limestone. It occurs either as dissemination or as clusters of several centimeters in diameter. Wollastonite is further found at the contacts between the limestone and country rock as veins about one centimeter wide. Here the splinters are aligned parallel to the contacts. Between the splinters, tiny quartz grains and diopside can frequently be seen. The wollastonite splinters are in most instances 3 mm long. Their  $2V_x = 41^\circ$ .

The scapolite has the composition almost of pure meionite. In some instances, it is markedly altered and is very fine-grained. The limestone generally contains well-crystallised sphene crystals as long as 3 mm. The epidote in the limestone is rich in pistacite.

Olivine and humite minerals are minerals rarely met with in the limestone of Hällinmäki. They occur as tiny grains, which have undergone partial or even total serpentinization. According to Virkkunen (1697), the forsterite content of olivine varies in the different parts of the limestone formation between the values of Fo 55 per cent and Fo 83 per cent. The outermost part of the kelyphite rim around certain of the olivine grains is composed of iron oxides and humite.

As far as their mineral composition and structure is concerned the garnet skarn zones occurring in the limestone are identical to the garnet skarn zones occurring in the diopside amphibolite. In my opinion, the latter zones formed during the period of regional metamorphism by recrystallization, and the garnet skarn zones observed in the limestone probably formed in the same manner.

## Metavolcanics

### Hornblende gneisses

The hornblende gneisses occur as intercalations of varying thickness in the diopside amphibolite. Although the grain-size varies considerably, some sorting, into layers, of grains measuring between 0.1 and 0.2 mm and others of 0.3 to 0.6 mm

in diameter, is noticeable. The texture of the hornblende gneiss is often blastoporphyr-  
ic.

The proportions of the principal minerals, plagioclase, quartz, biotite and horn-  
blende, vary considerably (Table 13). Accessory constituents are sphene, apatite and  
zircon. The plagioclase is often zoned and averages 35 to 40 per cent An. The  
hornblende is green and its  $\gamma = 1.693$ .

TABLE 13  
Mineral composition of hornblende gneiss (1.) and  
biotite-hornblende gneiss (2.). Determined by the point  
counting method. Vol. %.

	1.	2.
Quartz .....	30.5	9.0
Plagioclase .....	56.1	55.8
Biotite .....	8.8	26.9
Hornblende .....	4.0	1.9
Accessories .....	0.6	6.4
	100.0 %	100.0 %

1. 3/RS—65/Vrs, Hällinmäki

2. 166/HA—66/Jv

The porphyritic hornblende gneiss probably represents tuffaceous material. The  
hornblende gneisses occurring in the proximity of the mica gneisses and quartz-  
feldspar gneisses may, on the other hand, represent carbonaceous sandstones, for  
although they resemble the hornblende gneiss in mineral composition, no phenocrysts  
have been observed in them.

### Uralite-plagioclase porphyrites and metabasites

In places the diopside amphibolite contains thin uralite-plagioclase porphyrite  
intercalations, 0.1 to 0.0 m thick (Fig. 16).

The principal minerals contained in the uralite-plagioclase porphyrite are horn-  
blende and plagioclase. In addition, there is a small amount of biotite. The plagioc-  
lase and hornblende porphyroblasts are approximately 2 mm long, which contrasts  
with the finer (0.4 mm) groundmass. The anorthite content of the plagioclase is 35  
to 40 per cent and the crystals are zoned. The hornblende is green hornblende and,  
according to the index of refraction, it is of the same kind as that in the hornblende  
gneiss.

Metabasite dikes intersect the diopside amphibolites at Hällinmäki (Fig. 17).  
They underwent folding at the same time as the diopside amphibolite.

The metabasite is granoblastic, the average diameter of the crystals being 0.2—  
0.3 mm. The main minerals, hornblende and plagioclase, are similar to these in the

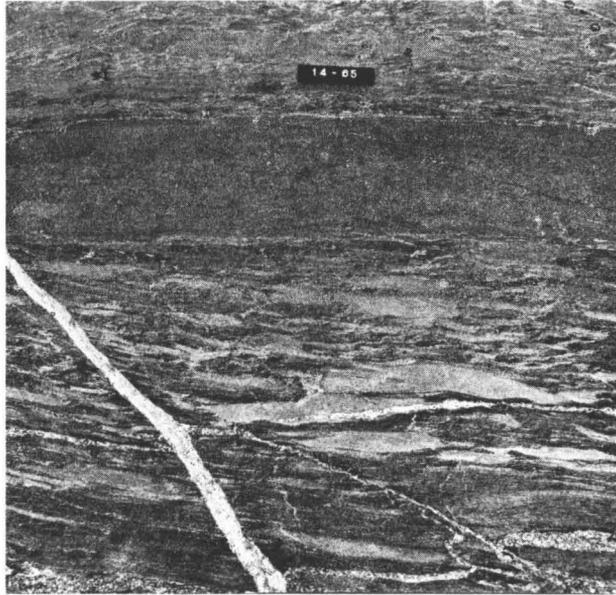


FIG. 16. Uralite-plagioclase porphyrite layer situated between diopside amphibolite and amphibolite. Hällinmäki.  
Photo E. Halme.

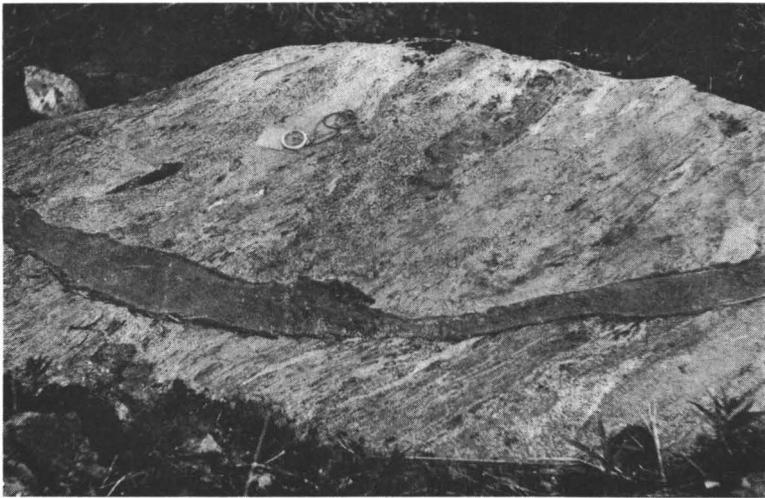


FIG. 17. Metabasite dike intersecting diopside amphibolite. Hällinmäki.

uralite-plagioclase porphyrite. The borders contain slightly more hornblende than the middle of the dike. Only a few flakes of biotite are present. In addition, there is a slight amount of apatite. The metabasite dikes probably represent dikes of volcanic origin, although primary structures are no longer visible. The relation of the dikes to the biotite amphibolites is a matter of conjuncture.

### Metamorphism of the supracrustal rocks

The mineral assemblages of the supracrustal rocks of the Virtasalmi area suggest a high degree of metamorphism. In order to delineate the temperature and pressure conditions prevailing during the metamorphic process the rocks shown in the following tabulation were considered:

Mica gneisses and veined gneisses	Quartz-feldspar gneisses	Diopside amphibolites	Limestones
plagioclase An <sub>20-35</sub>	plagioclase An <sub>20-25</sub>	plagioclase An <sub>55-60</sub>	wollastonite
quartz	quartz		quartz ±
biotite	biotite	calcite ±	calcite
microcline ±	microcline	diopside	diopside
muscovite ±	muscovite ±	hornblende	olivine ±
cordierite ±		hypersthene ±	humite ±
sillimanite ±		cummingtonite ±	serpentine ±
almandine ±	almandine ±	andradite ±	andradite ±
	scapolite ±	scapolite ±	scapolite ±
	epidote ±	epidote ±	epidote ±
	calcite ±		

### Mica gneisses, veined and quartz-feldspar gneisses

According to Turner and Verhoogen (1960, p. 548), the two gneiss groups belong to the sillimanite-almandine-muscovite subfacies of the almandine-amphibolite facies, in which the typical mineral assemblage for the pelitic rocks is, for example: quartz-almandine-muscovite-biotite-plagioclase.

Within the amphibolite facies, the following reactions are important, and these may take place under almost the same conditions (Winkler 1967, p. 181):

muscovite + quartz → potassium feldspar + Al<sub>2</sub>SiO<sub>5</sub> + H<sub>2</sub>O and muscovite + biotite + quartz → cordierite (almandine) + potassium feldspar + H<sub>2</sub>O

The following P—T values (see Figure 18) have been established for these reactions (Winkler 1967):

600 ± 10°C . . . . .	1 000 bars	H <sub>2</sub> O pressure
630 ± 10°C . . . . .	2 000 bars	»
690 ± 10°C . . . . .	4 000 bars	»

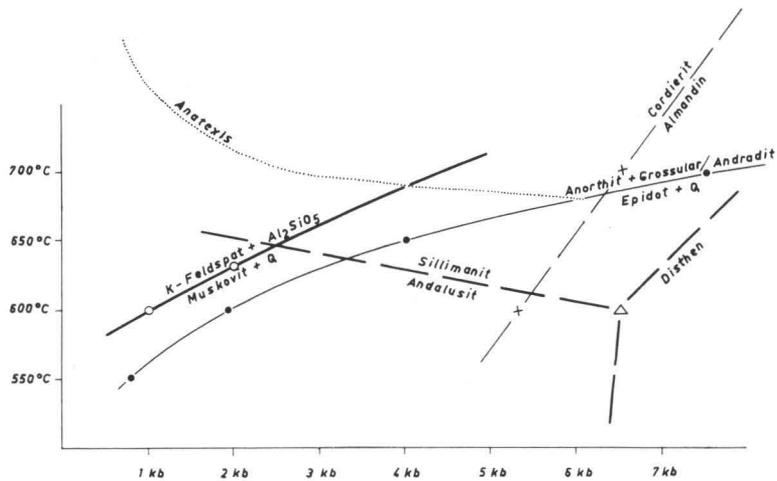


FIG. 18. Temperature and pressure conditions required for metamorphism. The curves have been drawn according to data reported by Winkler (1967) and they have been dealt with in more detail in the text.

The upper stability conditions for the coexistence of muscovite and quartz are c. 725°C and c. 5 000 bars  $H_2O$  pressure. Above this temperature and pressure they begin to melt, irrespective of the composition of the plagioclase in the gneiss. The stability field of the muscovite + quartz is limited by two boundaries: the phase boundary muscovite + quartz/potassium feldspar +  $Al_2SiO_5$  +  $H_2O$  and the P— $T_{H_2O}$  curve of the eutectic melt forming in the system, potassium feldspar—quartz— $H_2O$ . In examining the melting of natural rocks containing plagioclase, the last-mentioned curve is replaced by the P—T curve of the beginning of anatexis. The two curves run parallel, but the curve of anatexis is at a lower temperature (Winkler 1967, 176—186 and 206—214).

The partial melting of the graywackes begins, according to Winkler (1967, pp. 192—223), with sufficient  $H_2O$  pressure at 650 to 700 degrees, or under the P—T conditions of the amphibolite facies. Since the graywackes and all the other sediments of the Virtasalmi area primarily accumulated in water, it may be assumed that a sufficiently high  $H_2O$  pressure also prevailed.

The anatexis of the graywackes produces a melt that has the eutectic composition in the system Q—Ab—An—Or— $H_2O$ . With constant pressure prevailing, the composition of the eutectic melt also depends on the ratio Ab/An. This ratio further affects the temperature of the beginning of the melting process (von Platen 1965, p. 363).

If the light veinlets of veined gneisses are to be regarded as having originated through partial melting of mica gneisses, then the melt produced must have been granodioritic or granitic, for the muscovite and/or biotite would in such a case have

reacted in conjunction with the quartz to form a potassium feldspar component (Winkler 1967, pp. 209—214). This reaction would have taken place at a temperature of 700°C. As the light veinlets of veined gneisses in the Virtasalmi area contain little or no microcline, they can hardly be regarded in the light of the foregoing discussion as having originated through partial melting, particularly in view of the fact that the minerals occurring in the light veinlets are not in eutectic relation.

In the authors view, the veined gneisses of the Virtasalmi area may have originated through metamorphic differentiation. In the series of events, the plagioclase of the mica gneisses was first mobilized, and its composition did not change during the process (Mehner 1962). Following this stage the quartz (and/or the potash feldspar depending on the composition of the rock) became mobilized. Upon crystallization, they replaced the plagioclase.

Schmidt (1932) explains the alternation of light and dark layers in tectonites to be due to the differing translation efficiency (*Translationsfähigkeit*) of the minerals: the more mobile crystals become separated from the less mobile material and thereafter compose layers of their own. It is in this way that the veined gneisses of the Virtasalmi area may also have formed.

Because no melting had taken place in the gneisses of the Virtasalmi area, the temperature had not attained the curve of anatexis (Fig. 18). No andalusite or disthene has been detected; hence the metamorphism seems to have occurred in the stability field of sillimanite in the system  $\text{Al}_2\text{SiO}_5$ . In the mica gneiss the muscovite alters partially into sillimanite (p. 16). Thus the temperature probably reached, or approached, the phase boundary muscovite + quartz/potash feldspar +  $\text{Al}_2\text{SiO}_2 + \text{H}_2\text{O}$ , but it was not exceeded, as muscovite is still present.

In the light of the foregoing discussion, the metamorphism of the mica gneisses and quartz-feldspar gneisses of the Virtasalmi area appears to have taken place at a temperature of 640°—680° under a pressure of 2 500—4 000 bars (Fig. 18). On the other hand, in the quartz-feldspar gneiss, adjacent to the diopside amphibolite, epidote occurs in equilibrium with quartz. According to Nitsch and Winkler (1965), the epidote reacts with the quartz to form anorthite and some grossularite/andradite solid solution plus some hematite and  $\text{H}_2\text{O}$ . The reaction is reversible and the following data have been provided for the equilibrium:

550°C .....	800 bars	$\text{H}_2\text{O}$ pressure
600°C .....	1 900 bars	»
650°C .....	4 000 bars	»
700°C .....	7 500 bars	»

If it is assumed that the temperature had not exceeded 680°C, the maximum pressure which can be obtained by interpolations is c. 6 100 bars.

Under specific conditions, cordierite can exist only below a certain pressure, and Mn-poor almandine, only above a certain pressure. According to Winkler (1967, p. 183), the subfacies of the muscovite-free amphibolite facies has only a small

pressure field in which cordierite and almandine can coexist. The following pressure limits define the Mg-Fe cordierite/almandine boundary in the presence of biotite, quartz and  $\text{Al}_2\text{SiO}_5$ :

$$\begin{array}{ll} 5\,300 \pm 200 \text{ bars} & 600^\circ\text{C} \\ 6\,500 \pm 200 \text{ »} & 700^\circ\text{C} \end{array}$$

However, Winkler suggest that even a slight amount of the spessartine component will substantially reduce the lower pressure boundary for formation of almandine.

The cordierite in the mica gneisses of the Virtasalmi area appears to have formed according to the reaction muscovite + biotite + quartz  $\rightarrow$  cordierite + potassium feldspar +  $\text{H}_2\text{O}$  described. However, compared to almandine garnet, cordierite is rare.

According to Eskola (1946), in the typical amphibolite facies, potassium feldspar does not ordinarily coexist with andalusite (sillimanite), cordierite or anthophyllite, but if they do occur together a higher temperature subfacies must be involved. Moreover, the diadochy of the  $\text{Mg}^{+2}$  and  $\text{Fe}^{+2}$  ions is not complete under the conditions of the amphibolite facies, for the occurrence of cordierite and almandine-garnet depends on the chemical bulk composition of the rock. In the garnet-bearing mica gneisses of the Virtasalmi area, the weight percentage relations are, on the average,

$\frac{\text{Fe}^{+2}}{\text{Fe}^{+2} + \text{Mg}^{+2}} = 0.77$ , in cordierite-bearing mica gneiss 0.69, in mica gneiss 0.71 and in sillimanite-bearing mica gneiss 0.70 (Table 2). This may explain why there is overwhelmingly more garnet than cordierite; in other words, the rock was originally so rich in iron that under the P—T conditions in question, no cordierite was crystallized.

The differences between the various gneiss types are often negligible, so that on the basis of chemical composition alone, the formation of mineral assemblages

TABLE 14

Chemical composition of the cordierite-garnet gneiss, Näärinki, Juva. Sample 2/JP—64/Jv. Rock analysis by P. Ojanperä and mineral analyses by K. Laajoki.

	Rock	Garnet	Cordierite	Biotite
$\text{Al}_2\text{O}_3$ .....	17.55 %	20.9 %	—	—
$\text{Fe}_2\text{O}_3$ .....	1.03 %	—	—	—
FeO .....	6.74 %	—	—	—
MgO .....	3.47 %	5.1 %	10.4 %	11.4 %
FeO(total) .....	—	32.0 %	7.6 %	16.8 %
MnO .....	—	1.9 %	—	—
CaO .....	—	1.2 %	—	—
$\text{Fe}^{+2}$ *) .....	0.74	—	—	—
$\text{Fe}^{+2} + \text{Mg}^*$ )				

\*) total iron

of the mica gneisses of the Virtasalmi area cannot be described. Some five to ten kilometers beyond the study area, the cordierite-garnet gneisses are common. The bulk composition of one of them, as well as the composition of its various minerals (Table 14) closely resemble the compositions of the mica gneisses and their different minerals presented in Table 2. Such very similar compositions in both rocks and minerals suggests that different P—T conditions prevailed in the Juva area (abundant cordierite) compared to those in the Virtasalmi district (lacking in cordierite).

The above data suggest that, the metamorphism of acid schists in the Virtasalmi district took place at a temperature of  $650 \pm 30^\circ\text{C}$  and under a pressure of  $5\,000 \pm 1\,000$  bars (Fig. 18).

### Diopside amphibolites and limestones

The criteria for the beginning of the amphibolite facies are fulfilled with the first appearance of diopside (Winkler 1967). The carbonaceous rocks in the Virtasalmi area consistently contain diopside. Tremolite is only present in the diopside gneisses and here it generally alters to diopside. Thus, the mineral assemblages of the carbonaceous rocks are indicative of the amphibolite facies or its higher temperature subfacies.

The presence of wollastonite signifies, in Eskola's (1946) view, the higher subfacies of the amphibolite facies. According to Winkler (1967, p. 35), wollastonite is absent from rocks representative of high-grade regional metamorphism, and quartz and calcite coexist. The partial pressure of  $\text{CO}_2$  in the fluid phase increases to such an extent in the carbonaceous rocks during the period of regional metamorphism that a temperature of  $700^\circ$  to  $800^\circ\text{C}$  would be required before the wollastonite reaction could take place (Winkler, 1967). Temperatures of this order are unknown in regional metamorphic terrains.

However, wollastonite does coexist with carbonate and quartz in the Virtasalmi area, indicating that the wollastonite reaction was arrested. During the initial stage of regional metamorphism the partial pressure of  $\text{CO}_2$  evidently was so slight that the wollastonite reaction began. At a later stage, the partial pressure of  $\text{CO}_2$  increased and wollastonite reaction halted.

The reactions that took place in the carbonaceous rocks during the process of metamorphism may have prohibited the formation of cordierite in the adjacent mica gneisses. For instance, in the wollastonite reaction  $\text{CO}_2$  and  $\text{H}_2\text{O}$  will form, and if they cannot escape or if they are released very slowly, the gas pressure naturally increases. In the vicinity of carbonaceous rocks, therefore, the pressure is liable to reach a point where it exceeds the limits of the stability field of cordierite. This may have happened in the area of Virtasalmi.

## PLUTONIC ROCKS

The synkinematic plutonic rocks of the Virtasalmi area include peridotites, gabbros, diorites, quartz-diorites and trondhjemites. To a limited extent, the various plutonic rocks are regionally arranged: Trondhjemites and their granodioritic varieties occur mainly in the northwest corner of the area. Peridotites and gabbros are present in more than usual abundance in a NNE direction from Narila as well as between the Ankele quarry and Luomanen. The gabbros generally occur in the crystalline schists, but they also lie fairly close to the large quartz- diorite massifs.

The quartz-diorite massifs are heterogeneous in that they often contain fragments of schist (Fig. 19) and basic plutonic rocks.

With the exception of the acid dikes, the plutonic rocks of the Virtasalmi area have been folded with the schists. The thin diorite dikes that have concordantly penetrated the schists, in places show a boudinage structure. Aplite and pegmatite dikes intersect all the other rocks occurring in the study area.

### Peridotites and gabbros

The peridotites and gabbros are intimately associated in the field as peridotites do not form independent intrusions. These rocks mainly form tongues or lenses in crystalline schists or in the more acid members of intrusive suite. The occurrences

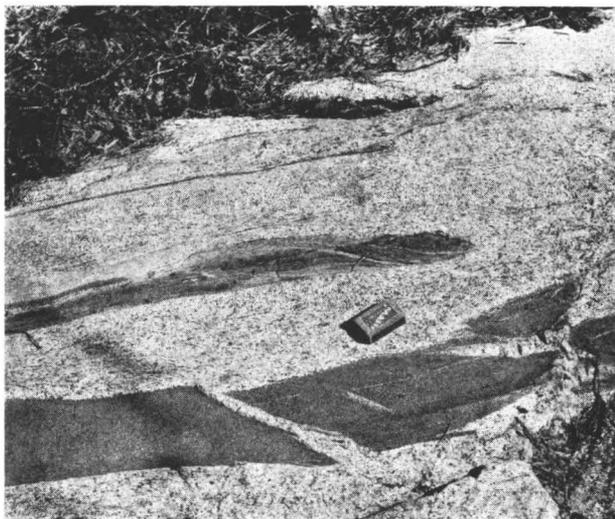


FIG. 19. Diopside amphibolite brecciated by quartz diorite.  
E shore of Litmanen.

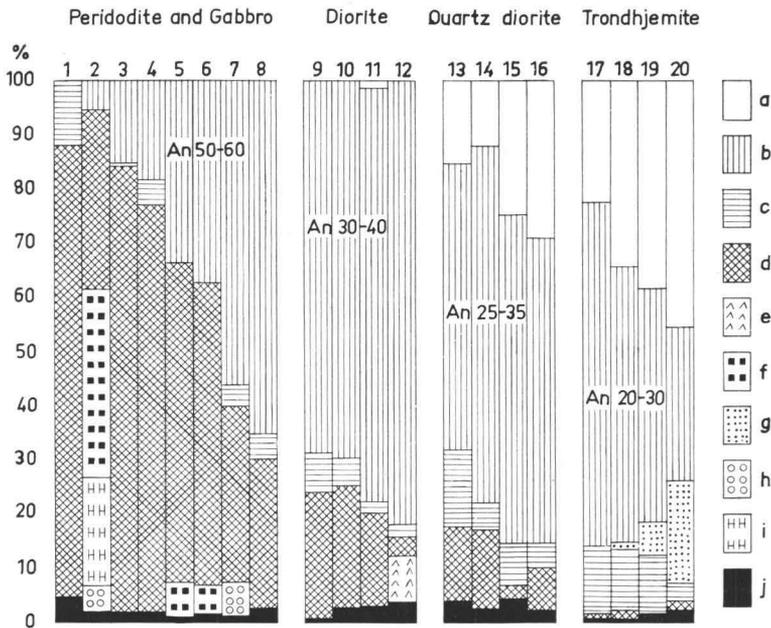


FIG. 20. Mineral composition of plutonic rock in the Virtasalmi district. Determined by the point counting method. Vol. %. a = quartz, b = plagioclase, c = biotite, d = hornblende, e = cummingtonite, f = augite, g = potassium feldspar, h = olivine, i = hypersthene and j = accessory minerals. 1 = hornblendite (P 187/ER—64), 2 = peridotite (Narila/Hulkkonen), 3 = gabbro-peridotite (R 143/32.75 Karhuniemi), 4 = gabbro-peridotite (R 80/73.90 Hällinmäki), 5 = gabbro (R 6/18.55 Hällinmäki), 6 = gabbro (R 7/147.50 Hällinmäki), 7 = gabbro (K 27/HA—66), 8 = gabbro (K 10B/HA—65), 9 = diorite (R 3/88.65 Hällinmäki), 10 = diorite (R 3/27.75 Hällinmäki), 11 = diorite (R 3/92.95 Hällinmäki), 12 = diorite (R 3/145.30 Hällinmäki), 13 = quartz diorite (R 81/81.83 Hällinmäki), 14 = quartz diorite (60/JP—64), 15 = quartz diorite (60/JP—64), 16 = quartz diorite (R 76/312.87 Hällinmäki), 17 = trondhjemite (R 51/95.70 Hällinmäki), 18 = trondhjemite (21K/HA—66), 19 = granodioritic trondhjemite (186/HA—66) and 20 = granodioritic trondhjemite (34/HA—66).

are limited in aerial extent: they have an average length of about 0.5 km and breadth of 100 to 200 meters.

No sharp boundaries can be drawn between the peridotites and gabbros, for the percentage of the dark minerals diminishes steadily as the rocks change from peridotite to gabbro (Fig. 20).

The grain size of the peridotites is usually 3 to 5 mm, and that of the gabbros, 2 to 3 mm. Both rocks are represented by coarse-grained varieties which have grain size of 1 cm. The marginal portions of the massifs generally exhibit mineral paralle-

lism, whereas the middle portions have a massive structure. Depending on the degree of mineral foliation, the texture is either hypidiomorphic or blastohypidiomorphic.

The peridotites are hornblende-dominated rocks, additional constituents being plagioclase, biotite, olivine, hypersthene, augite and serpentine (alteration product of olivine). Accessory minerals are apatite, sphene, chalcopyrite, pyrrhotite, pentlandite and ilmenite. In addition to these minerals the gabbros contain cummingtonite.

On the basis of optical determinations the minerals of both the peridotites and the gabbros are similar in composition. Thus, in both rocks, olivine has a  $2V_x = 85-90^\circ$ , in augite the  $Z \wedge c = 40-45^\circ$ , and in orthopyroxene  $2V_x = 67-73^\circ$ . These data correspond to a fayalite content of c. 15 to 25 Mol. per cent in the olivine and a c. 25 Mol. per cent  $Fe^{+2}$  component in the orthopyroxene (Tröger 1959). Values of  $Z \wedge c = 18-19^\circ$  and  $2V_x = 80^\circ$  were recorded for the cummingtonite and this corresponds to a  $Fe^{+2}$  component of c. 40-45 Mol. per cent. In many instances there is a homoaxial intergrowth between hornblende and cummingtonite. The An content of the plagioclase averages 50-60 per cent An. The plagioclase crystals are often normally zoned, and in such cases the center is usually c. 5 per cent richer in An than are the marginal portions. In certain plagioclase crystals of the peridotites, however, the marginal portions are more basic (c. 5 per cent An) than the center.

One feature of the peridotites and gabbros of the Virtasalmi area is that if they contain pyroxene, they have no primary biotite. Also the hornblende is slightly different in pyroxene-bearing peridotites from that in the biotite-bearing ones. For example: in the pyroxene-bearing peridotites of the Hulkkonen massif, the hornblende is light brown and its  $Z \wedge c = 16^\circ$  and  $2V_x = 88^\circ$ . The hornblende of the biotite-

TABLE 15  
Chemical analyses of the plutonic rocks in the Virtasalmi area.

	1.	2.	3.
SiO <sub>2</sub> .....	49.64	51.40	67.79
TiO <sub>2</sub> .....	1.24	1.43	0.86
Al <sub>2</sub> O <sub>3</sub> .....	12.94	19.00	15.25
Fe <sub>2</sub> O <sub>3</sub> .....	2.81	3.92	2.12
FeO .....	6.18	5.76	2.88
MnO .....	0.15	0.15	0.05
MgO .....	9.74	3.85	1.36
CaO .....	12.62	8.05	3.75
Na <sub>2</sub> O .....	2.40	4.80	4.20
K <sub>2</sub> O .....	0.36	0.34	0.90
P <sub>2</sub> O <sub>5</sub> .....	0.15	0.57	0.18
CO <sub>2</sub> .....	0.00	0.00	0.10
H <sub>2</sub> O <sup>+</sup> .....	1.46	0.74	0.42
H <sub>2</sub> O <sup>-</sup> .....	0.05	0.02	0.06
	99.74 %	100.03 %	99.92 %

1. Gabbro. Analyst: P. Ojanperä.
2. Diorite. » »
3. Quartz diorite. Analyst: A. Heikkinen.

bearing hornblendite in the same massif is green and its  $Z \wedge c = 18^\circ$  and  $2V_x = 84^\circ$ . At the present time, nothing can be said of the distribution of the pyroxene- and biotite-bearing peridotites and gabbros.

### Diorites

The diorites occur as small, irregularly-shaped, intrusives associated with the gabbros or located in the quartz diorite. Their marginal portions are often foliated, whereas their centers are massive. In texture the diorites are either hypidiomorphic or blastohypidiomorphic, depending on the degree of mineral elongation.

Mineralogically the diorites contain hornblende, plagioclase, biotite and, in some case, cummingtonite (Fig. 20). Accessory minerals are quartz, apatite, sphene, zircon, magnetite, ilmenite, chalcopyrite and pyrrhotite.

The average An content of the plagioclase is 30—40 per cent. Normally zoned plagioclase crystals are also present, the central zones being c. 5 per cent richer in An than the marginal ones. The hornblende is green in color and has a  $2V_x = 68—70^\circ$  and  $Z \wedge c = 17—18^\circ$ . The cummingtonite has a  $Z \wedge c = 20^\circ$  and  $2V_z = 80^\circ$ , indicating that it is similar in composition (40—50 Mol. per cent  $Fe^{+2}$  component) to the cummingtonite of the gabbro.

Sharp boundaries cannot be drawn between the diorites and gabbros, the proportion of dark minerals in the rock decreases fairly evenly as the composition changes from gabbros to diorites and then to quartz diorites (Fig. 20). The main difference between the diorites and quartz diorites, as the name suggests, lies in the greater abundance of quartz in the quartz diorites. Besides the variation in the composition of the feldspars, the most marked difference between gabbros and diorites is that latter contain 2—4 Vol. per cent of magnetite; this mineral is absent from peridotites and gabbros. The ilmenite contains magnetite exsolutions. The pyrrhotite observed in diorites does not contain pentlandite exsolutions as does the pyrrhotite of peridotites and gabbros.

### Quartz diorites

Quartz diorite is the main plutonic rock found in the Virtasalmi district. It forms elongated massifs trending in most cases, parallel with the foliation. These massifs are over 10 km long and several km wide. Only in the central portions of rather large areas are there small tracts where the rock is massive. Elsewhere the quartz diorite exhibits a mineral foliation. In the best preserved, massive varieties, the texture is hypidiomorphic, but in the gneissose varieties the structure is, by contrast, blastohypidiomorphic. The quartz diorite is an even-grained rock, in which the grain size of the principal minerals is between 2 and 8 mm.

The quartz diorite contains fragments of more basic members of the intrusive suite and crystalline schists and also exhibits local variations in composition. The contacts between the inclusions and host quartz diorite are generally sharp. The quartz-feldspar gneiss fragments, however, are sometimes roundish and lacking sharp boundaries, indicating that the quartz diorite had partially assimilated quartz-feldspar gneiss fragments. The quartz diorite brecciates the diorites and other more basic rocks. The quartz diorite massifs are, therefore, largely mixed rocks, but there do occur broad areas of apparently homogeneous quartz diorite.

The essential mineral composition of the quartz diorite is plagioclase, quartz, hornblende, biotite while the accessory constituents are apatite, sphene, zircon, magnetite and ilmenite (Fig. 20). The plagioclase has an average anorthite content of 25—35 per cent. The central portions of the zoned crystals are usually sericitized and saussuritized. Hornblende has a  $Z \wedge c = 18^\circ$  and  $2V_x = 70^\circ$ . Magnetite and ilmenite account for 2—4 Vol. per cent. The ilmenite contains magnetite exsolutions.

In marginal parts of the massifs on the northern side of Narila and the north-western side of Litmanen, there are porphyrite varieties of quartz diorite. A distinct lineation particularly illustrated by the potassium feldspar porphyroblasts is a common feature of this quartz-diorite variety. The reddish potassium feldspar porphyroblasts (up to 3—4 cm diameter) may be angular or rounded, usually contain perthite and in some instances are mantled by plagioclase. The porphyroblasts increase in abundance toward the center of the porphyritic quartz-diorite massifs, being at the maximum 1—2 phenocrysts/1 dm<sup>2</sup>.

### Trondhjemites

Trondhjemite occurs in greatest abundance in the northwest of the Virtasalmi area where it is intimately associated with the quartz diorite. In places the trondhjemite is present as narrow zones between the diorite and the schists (Fig. 21).

The trondhjemite invariably exhibits a mineral lineation, its texture is blastohypidiomorphic and has a grain size resembling that of the quartz diorite. The trondhjemite consists mainly of plagioclase, quartz and biotite with minor amounts of hornblende. Granodioritic varieties also occur, containing the added constituent, microcline. Accessories are apatite, sericite, epidote and zircon.

The composition of the plagioclase is generally 20 per cent An, but crystals containing 30 per cent An have also been observed. The plagioclase is occasionally altered to sericite and epidote. In the granodioritic varieties, the microcline apparently replaces plagioclase and myrmekite has been noted. Myrmekite is not characteristic of primary plutonic rocks in Finland (Härme 1958, p. 58), suggesting that the granodioritic varieties of trondhjemite probably represent secondary alterations of the rock.



FIG. 21. Trondhjemite zone (trj) situated between diorite (dr) and amphibolite (af). Further, between the trondhjemite and the diorite is narrow band of quartz diorite (qvdr). The figure shows that all the rocks underwent simultaneous folding. Hällinmäki.

Photo E. Halme.

### Pegmatites and aplites

Pegmatites and aplites are not common in the Virtasalmi area. They intersect all the other rocks and are not participants in the folding. The pegmatites are not everywhere of the same age, for they have observed to intersect each other. However there may not be very great age difference between the intersecting pegmatites, as there seems to be only one intrusive suite in the area.

The principal minerals of the pegmatites are plagioclase and quartz. In addition, there is some muscovite and, in places, tourmaline. The plagioclase is a slightly sericitized oligoclase. The main minerals in the aplite are microcline and quartz in addition to small amount of turbid oligoclase and minor muscovite flakes. At Hällinmäki analcite and barite have been observed in the pegmatite dikes associated with the copper ore.

### Classification and chemistry of the plutonic rocks

The classification of the plutonic rocks of the Virtasalmi area is based on criteria such as mineralogical composition, composition of the plagioclase, the amount of

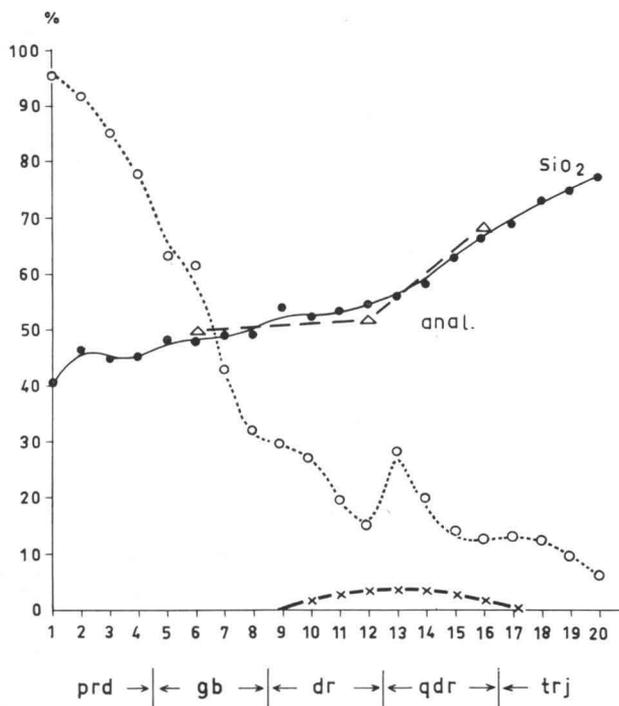


FIG. 22. SiO<sub>2</sub> contents of the plutonic rocks of the Virtasalmi area as calculated from the point counting analyses.

- = SiO<sub>2</sub> Wt %, according to point counting analyses,
- △—△—△ = SiO<sub>2</sub> Wt %, according to chemical analyses,
- = Vol. % of dark minerals,
- ×—×—× = Vol. % of magnetite + ilmenite.

prd = peridotite, gb = gabbro, dr = diorite, qdr = quartz diorite and trj = trondhjemite. The numbers correspond to the rocks shown in Fig. 20.

mafic minerals, and SiO<sub>2</sub> content (Fig. 22). On the basis of this classification it is possible to derive a suite of rocks ranging from peridotites to trondhjemites.

According to Simonen (1960, p. 51), the plagioclase content of the rocks of the granodiorite province is highest in the diorites (or »quartz gabbros«, to use Simonen's terminology) and quartz diorites. According to his definition, the Virtasalmi district may be designated as a granodiorite province, inasmuch as the dominant rock types, the diorites and quartz diorites, contain plagioclase in greater amounts than do the other members of the suite (Fig. 20).

An analysis of each of the gabbro, diorite and quartz diorite from the Virtasalmi area is given in Table 15. These analyses may be compared with the average compositions of the petrographic province (Table 16).

TABLE 16  
H<sub>2</sub>O-, K<sub>2</sub>O- and Na<sub>2</sub>O-contents of the different rock provinces.

	Gabbro			Diorite			Quartz diorite		
	H <sub>2</sub> O %	K <sub>2</sub> O %	Na <sub>2</sub> O %	H <sub>2</sub> O %	K <sub>2</sub> O %	Na <sub>2</sub> O %	H <sub>2</sub> O %	K <sub>2</sub> O %	Na <sub>2</sub> O %
Virtasalmi area .....	1.51	0.36	2.40	0.76	0.34	4.80	0.48	0.90	4.20
Trondhemite province									
Simonen, 1960 .....	2.35	0.55	1.66	0.92	1.82	2.95	0.87	2.00	3.62
Granodiorite province									
Simonen, 1960 .....	1.76	0.72	1.73	0.92	1.82	2.95	0.87	2.00	3.62
Charnockite province									
Simonen, 1960 .....	*) 1.44	1.17	3.04	—	—	—	0.81	1.89	4.50
Granite province									
Simonen, 1960 .....	1.51	1.00	3.08	1.52	1.83	2.93	0.93	2.47	3.63

\*) hornblende-rich noritic gabbro.

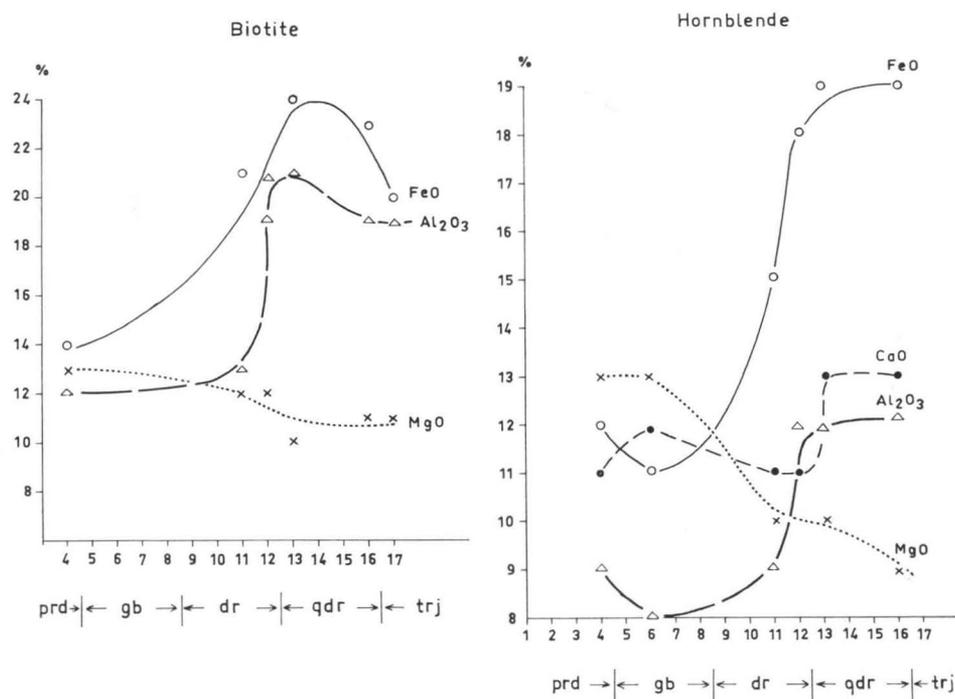


FIG. 23. Variations in the composition of biotite and hornblende in the plutonic rocks of the Virtasalmi area. The numbers correspond to the rocks shown in Fig. 20. prd = peridotite, gb = gabbro, dr = diorite, qdr = quartz diorite and trj = trondhemite. FeO = total iron content. The minerals were analyzed by J. Siivola with a microprobe.

As Table 16 reveals, a very low  $K_2O$  content appears to be characteristic of the plutonic rocks of the Virtasalmi area in all classes, accompanied by an above average  $Na_2O$  content in the diorite and quartz diorite.

These differences in composition disallow the comparison of the Virtasalmi rocks with Simonen's averages. The plutonic rocks of the Virtasalmi area must be regarded, however, as plutonic rocks of the trondhjemite series in view of their almost total lack of potassium feldspar.

A feature characteristic of the plutonic rocks of Virtasalmi, furthermore, is the fact that the peridotites and gabbros are very weakly magnetic, whereas the quartz-diorite massifs are in places strongly magnetic. This difference in magnetic properties is due to the abundance of magnetite in the diorites and quartz diorites.

The suite of the plutonic rocks of the Virtasalmi area from peridotites to trondhjemites is also reflected in the changes in the composition of the hornblende and the biotite (Fig. 23). The magnesium content of the hornblende and biotite decreases with increasing of the silica, whilst the iron and aluminum contents increase as the composition of the rock changes from peridotite to quartz diorite. The iron and aluminum contents decrease, however, as the rock changes to trondhjemite. This change may well be due to the fact that when magnetite crystallized simultaneously with the quartz diorite, the composition of the residual magma changed so much with respect to the iron content that the biotite of the trondhjemite became poorer in iron than that of the quartz diorite.

## THE COPPER DEPOSITS

Almost without exception, the copper deposits in the Virtasalmi area are associated with the garnet skarns bordering on the extensive quartz diorite massif located on the southwestern side of Virmasjärvi. The majority of the observed copper-bearing boulders are situated either within the massif or along its margins or on its southeastern side (Fig. 24). The copper-bearing skarn deposits are usually of small dimensions.

In addition to the copper-bearing skarn boulders, several copper- and nickel-bearing gabbro boulders have been found in different parts of the area. The peridotite-gabbro massif of Hulkkonen contains some chalcopyrite and pyrrhotite with exsolutions of pentlandite. Because the nickel and copper contents of this massif do not occur in economic quantities and as they are not related to the copper skarn ores, they will not be discussed in any more detail.

As the copper skarn occurrence at Hällinmäki is the most thoroughly investigated deposit, only data pertaining to it will be presented along with minor details of the Karhuniemi and Lari deposits.

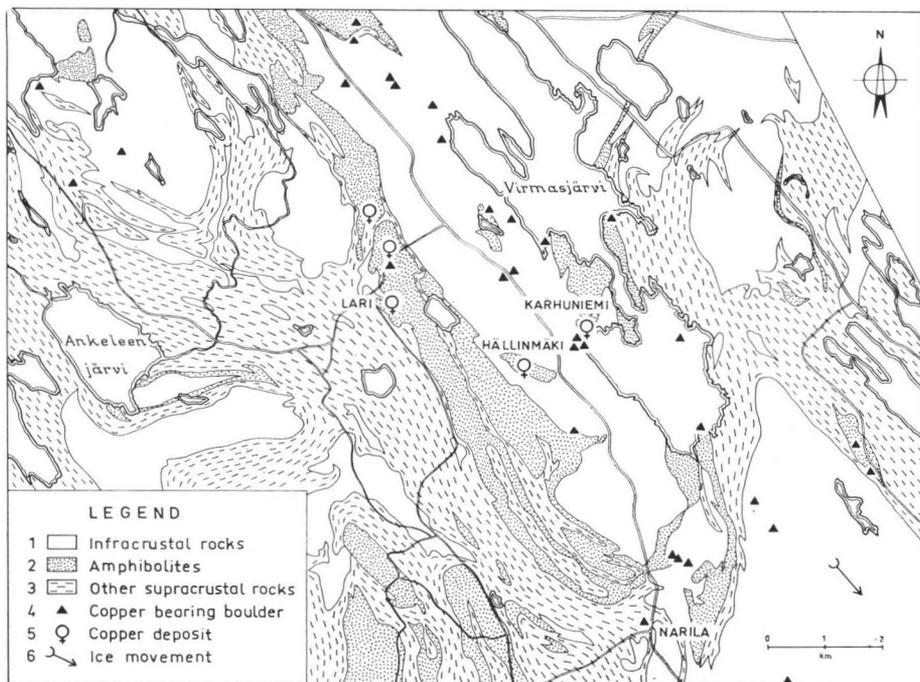


FIG. 24. Location of known copper deposits and boulders in the Virtasalmi area.

### The Hällinmäki deposit

#### The bedrock of Hällinmäki

The copper deposit of Hällinmäki is situated on the southwestern edge of a formation containing schists (c. 750 to 800 m long and 200 to 250 m wide) which is bounded, and to some extent also disrupted, by diorites. The general strike of the foliations of the schists varies between  $50^{\circ}$  and  $60^{\circ}$  W, and the dip is mainly  $65^{\circ}$ — $75^{\circ}$  NE. There occurs in the schists intense drag folding, the axes of which plunge, as an average  $60^{\circ}$ — $70^{\circ}$  to S  $10^{\circ}$ — $20^{\circ}$  E.

In general, the schists occur with the amphibolite and limestone horizons lying along the northeast margin and the diopside amphibolite horizons (with their garnet skarn intercalations) in the southwestern part (Fig. 25).

At Hällinmäki there also occur plutonic rocks, which are both conformable to the foliation and transgress the schists sequences (Fig. 26). Diorite is present in the greatest amounts of the plutonic rocks. Gabbro is not abundant and it occurs especially at the southwestern margin of the ore occurrence. Quartz diorite and trondhjemite occur in smaller amounts than do the other plutonic rocks, and they have been encoun-

tered for the most part in the southeastern localities. With the exception of acid dikes, the plutonic rocks of Hällinmäki have undergone deformation at the same time as the drag folds were formed.

The zone containing the ore (c. 500 m long and 2 to 30 m wide) is situated at the contact between the diopside amphibolite and diorite (Figs. 25, 26 and 27). As Figs. 26 and 27 show, the ore zone is discontinuous and is divided into several ore bodies (I—V). The separate ore bodies are almost vertical and appear to run parallel to the axes of the drag folds. At least on a small scale, this is actually the situation, for the tops of the small drag folds are in many cases mineralized in line with the axis, whereas the surrounding rock is completely barren of ore minerals.

According to an estimate made by Geological Survey of Finland in 1966, the Hällinmäki copper deposit contains approximately two million tons of copper ore of average copper content 0.95 per cent.

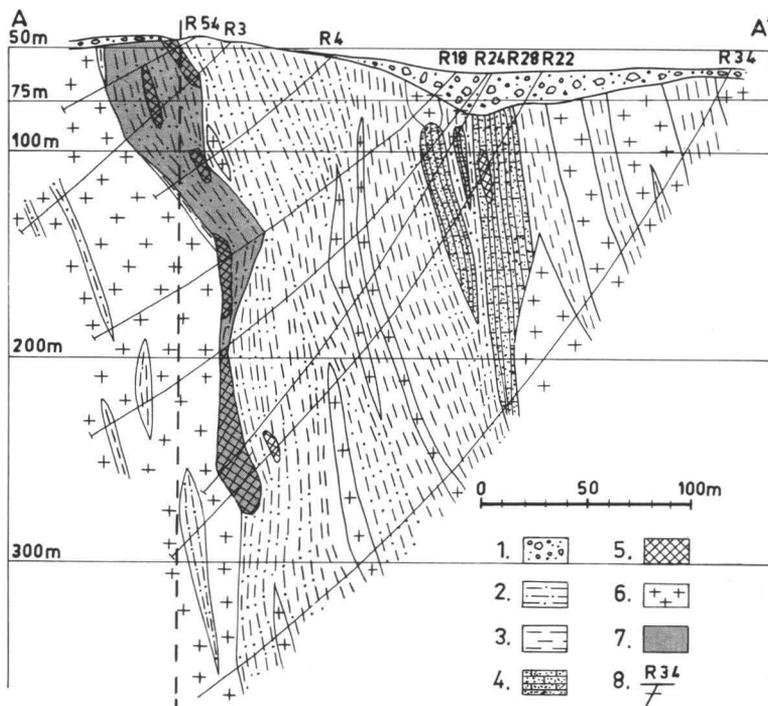


FIG. 25. Cross-profile of the Hällinmäki deposit. Section 55.385 in Fig. 27 and A-A' in Fig. 26. 1 = till, 2 = diopside amphibolite, 3 = amphibolite, 4 = limestone, 5 = garnet skarn, 6 = plutonic rocks, 7 = ore zone and 8 = drill hole.

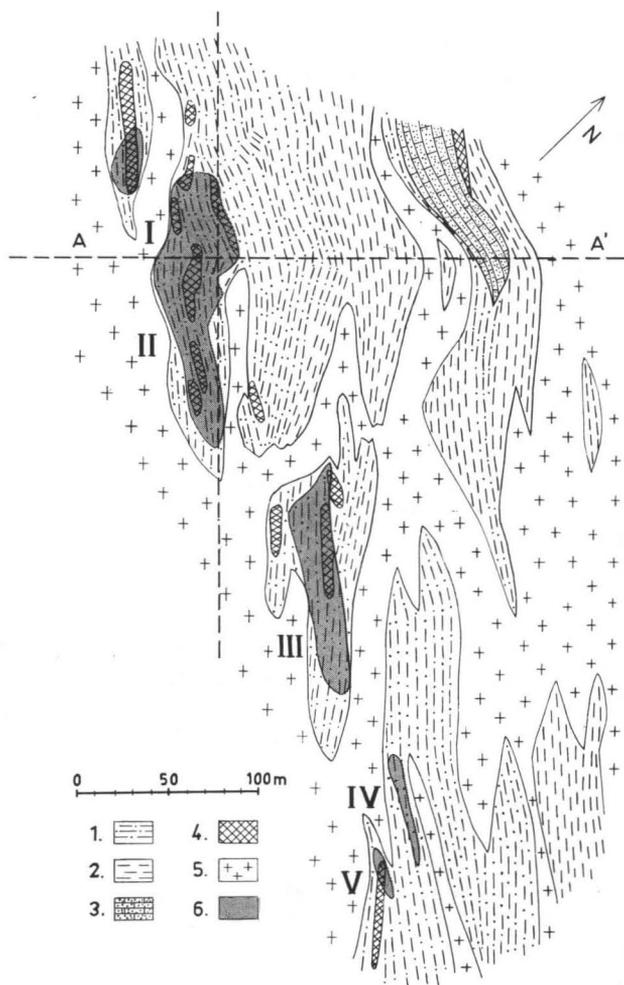


FIG. 26. Geological map of the Hällinmäki deposit. Horizontal plane 75 m in Fig. 25. 1 = diopside amphibolite, 2 = amphibolite, 3 = limestone, 4 = garnet skarn, 5 = plutonic rocks and 6 = ore zone. I = ore body.

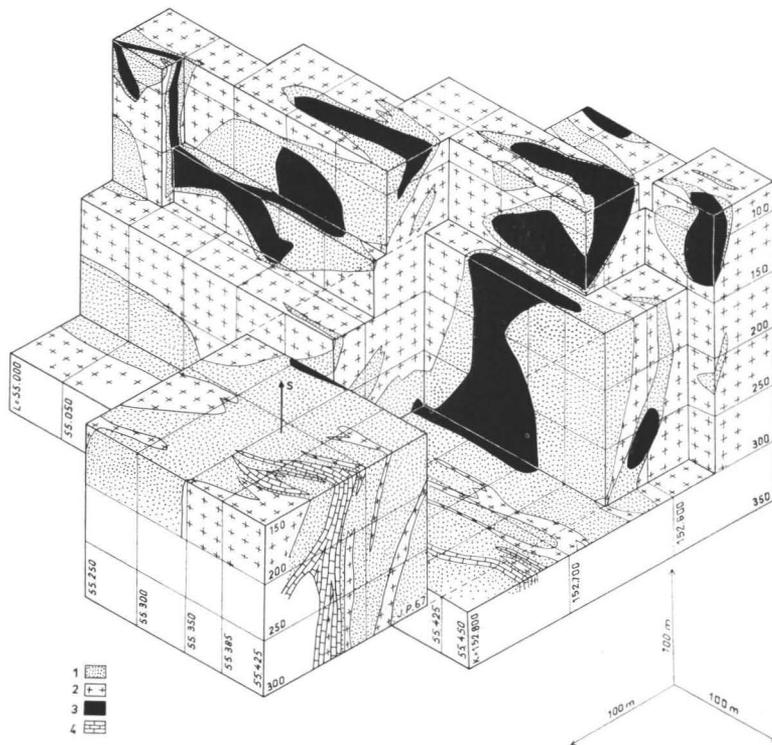


Fig. 27. Block diagram of the copper occurrence of Hällinmäki. 1 = amphibolite and garnet skarn, 2 = diorite, 3 = ore and 4 = limestone. Drawn by Jouni Pekkarinen.

### Chemical composition of the ore zone

Copper is the only commercially significant metal in the Hällinmäki ore. Its distribution in the mineralized zone is presented in the two diagrams shown in Fig. 28. One of them depicts the distribution of the copper according to specimens taken from drill cores (694 m) and the other surface specimens (570 in all). The striking similarity of the two diagrams (Fig. 28) implies an even distribution of copper throughout the ore.

The trace amounts of cobalt and nickel vary in concentration depending upon fluctuations in the copper content. The highest nickel content is 0.07 per cent and the highest analyzed cobalt content 0.03 per cent. In drill hole 7, the ore has an average Ni/Co of 2.5 and an average Ni/Cu of 0.2. In the peridotite-gabbro massif of Hulkkonen, the Ni/Cu = 2.2. Wilson and Andersson (1959) suggest that the Ni/Co ratio is characteristic of each ore deposit, and they have observed that this ratio is higher in ores associated with peridotites than with gabbros; i.e., the ratio

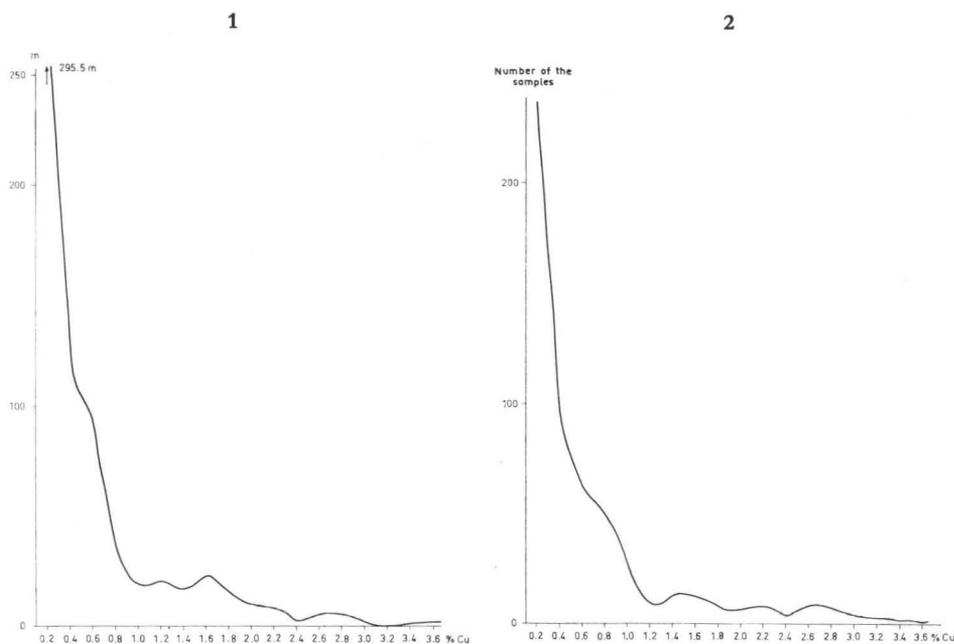


FIG. 28. Distribution of copper in mineralized zone. The diagrams were drawn on the basis of the specimens taken from drill cores analyzed (1) and the samples taken from the top of the ore (2). The samples were collected at a distance of 1 m from each other.

decreases as the  $\text{SiO}_2$  content increases. According to Häkli (1963), the average Ni/Co ratio of the ores in Finland associated with peridotites is 23.3, that of pyroxene gabbros 14.1 and of amphibole gabbros, 14.8. Häkli further notes that the Ni/Cu average of the ores associated with peridotites is 6.4 whilst that of pyroxene gabbros is 2.1 and in amphibole gabbros it is 2.9. Häkli also suggests that the sulfide phase of the copper ore of Hällinmäki is derived from the same magma which gave rise to the diorites, quartz diorites and trondhjemites.

The nickel, cobalt and copper contents of the whole rocks and sulfides were also investigated in the diorite occurring on the southwest side of the ore. The diorite averages 39 ppm Co and 22 ppm Ni whilst the gabbro contains 42 ppm Co and 69 ppm Ni (Fig. 29). Goldschmidt (1937) reports that diorites contain on the average 32 ppm Co and 40 ppm Ni, and gabbros 70 ppm Co and 158 ppm Ni. According to this specification, the Ni and Co contents of the Virtasalmi plutonic rocks — excepting the Co contents of the diorite — are subaverage. The copper content of the diorite is extremely variable (Fig. 29). At distance greater than 60 to 70 m from the ore body the diorite contains less than 50 ppm copper and the Cu/Co is below 1. At distances less than 60 m, the copper content of the diorite increases to an average of 110–130 ppm about 10 m from the ore body. Thus on the southwest side of the ore, there

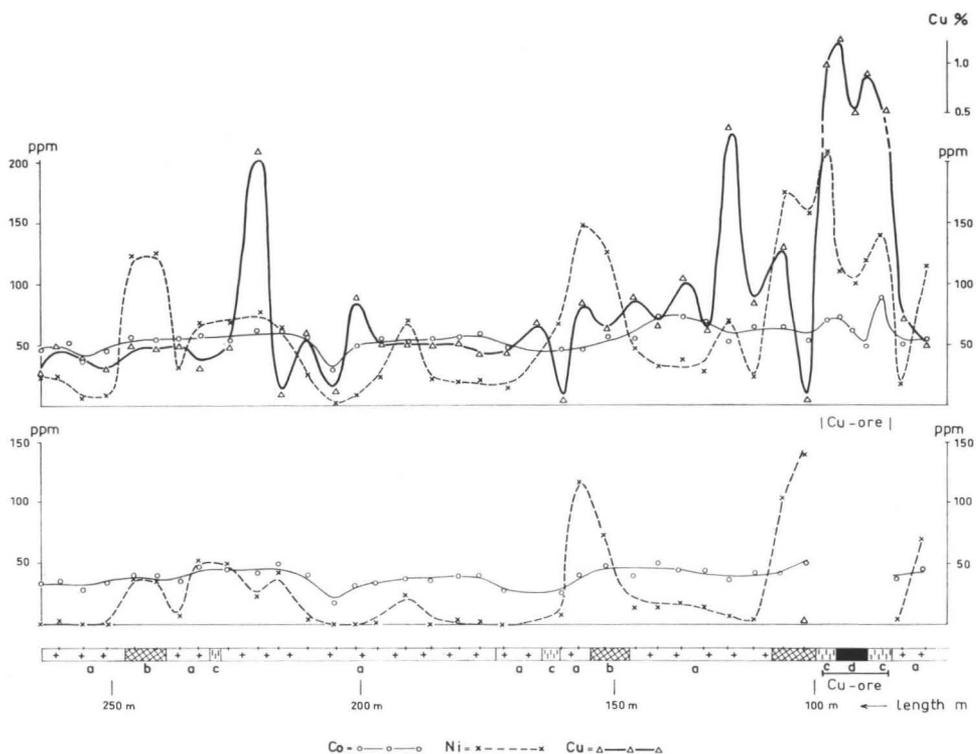


FIG. 29. Cu, Ni and Co contents of the Hällinmäki ore (drill hole 7) and the diorite and gabbro located on its northwestern side. The lower figure represents the Co and Ni contents of the silicate phase, and the upper on the total Co, Ni and Cu contents of the rock. a = diorite, b = gabbro, c = diopside amphibolite and d = garnet skarn. The analyses were done by E. Kontas with an atomic absorption spectrometer.

is a copper aureole, in which the Cu/Co ratio is from one to two. In the ore body this ratio is over a hundred.

From the analytical results of drill hole 3 silver occurs in the Hällinmäki ore in amounts of 1 to 5 g per ton. In places where the copper content of the ore rises over 2 per cent, the silver content likewise increases slightly, reaching an average of 7 to 8 g per ton. Gold is sporadically distributed throughout the ore in amounts of 0.1 g per ton. Molybdenite also occurs in trace amounts: 0.01–0.04 per cent  $\text{MoS}_2$ . The zink concentration does not exceed a few fractions of a per cent.

Portions with the highest concentrations of sulfur register 3 to 4 per cent, but the mean sulfur content of ore remains below 1 per cent. Iron (soluble in HCl) generally occurs in amounts of 2 to 5 per cent although in magnetite-bearing portions the iron content is likely to be nearly 30 per cent.

## Ore types

The ore minerals at Hällinmäki occur either as disseminated ore or brecciated ore. The two types differ markedly with respect to the size, amounts, numbers and textures of the ore minerals.

### *Disseminated ore*

The disseminated ore occurs mainly in diopside amphibolite, where it is either evenly distributed or arranged in layers; most of the layers are diopside-bearing. Where the diopside-bearing beds are folded, the disseminated ore often contains small accumulations of sulfide minerals at the apices of the minor folds. The average grain size of the ore minerals is 0.1—0.4 mm.

The main ore minerals are chalcopyrite, pyrite, pyrrhotite and magnetite. The chalcopyrite content amounts to over 70 Vol. per cent of the ore minerals (Fig. 30). Cubanite accounts for a few per cent of the ore minerals usually and has an irregular distribution. If the grain size of the ore minerals exceeds the average, the cubanite may increase to 20 Vol. per cent. In poorly disseminated ore there is an inverse relationship between the cubanite and pyrite contents. Cubanite exsolution bodies are common in chalcopyrite.

Pyrrhotite accounts for an average of 5 to 6 Vol. per cent of the disseminated ore. Magnetite occurs very irregularly; generally its content is below 1 Vol. per cent. As an exception there are nearly compact, 0.1—1.0 meter thick lenses, in which the magnetite content increases to more than 60 Vol. per cent.

Trace amounts of sphalerite and pentlandite occur as exsolution bodies in chalcopyrite and pyrrhotite, respectively.

### *Brecciated ore*

The brecciated ore (Fig. 31) occurs chiefly in garnet skarn zones or their immediate proximity. The sulfides often form mesh-like structures, from 5 to 10 cm across, which surround garnet or amphibolite fragments. The ore minerals also occur as lenses, 1 to 3 cm thick, which contain in many cases small fragments of the wall rock. The contacts of the lenses and the wall rock are not always sharp as they may be framed by disseminated ore to a depth of 0.5 to 1.0 cm. Sulfides also penetrate the narrow fractures in the fragments.

In the breccia-type ore the main mineral constituents are chalcopyrite, cubanite and pyrrhotite with minor amounts of pyrite and magnetite. In rich ore, chalcopyrite commonly accounts for one-half of the ore mineral content and cubanite 30 to 40 Vol. per cent. The pyrrhotite content varies from 5 to 10 Vol. per cent. There is an irregular

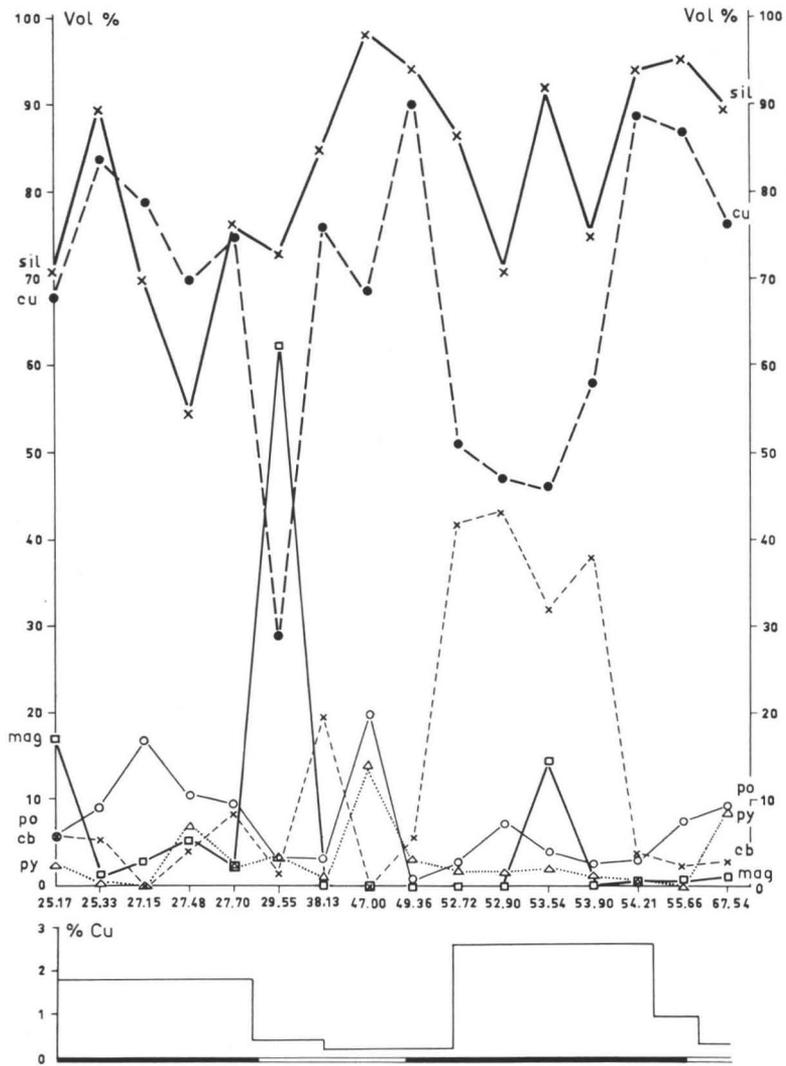


FIG. 30. Interrelations of the ore minerals and their alterations according to the variations in copper contents. The %-figures for the ore minerals represent their interrelations, whereas the % for the silicates signifies the proportion of the whole rock taken up by the silicate phase. sil = silicates, cu = chalcopyrite, mag = magnetite, po = pyrrhotite, cb = cubanite, py = pyrite, ——— = garnet skarn and — = diopside amphibolite. Determined by the point counting method. Hällinmäki, drill hole 3.

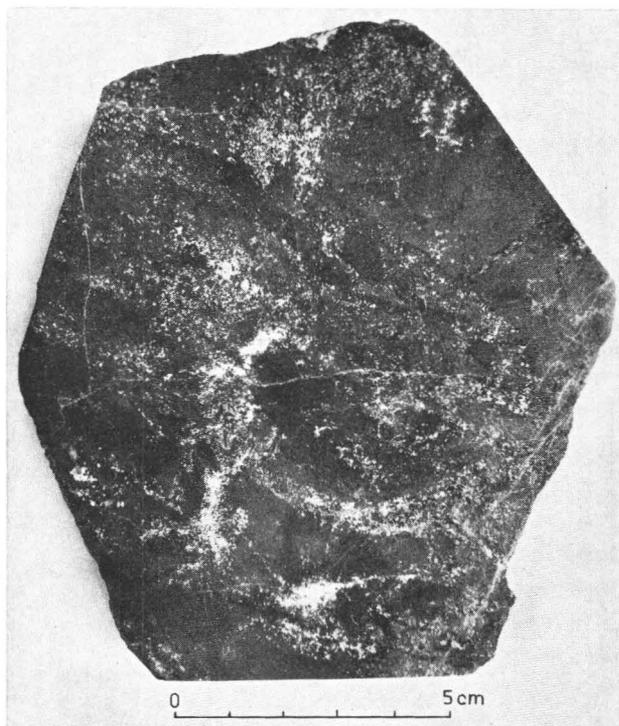


FIG. 31. Brecciated ore. Hällinmäki. Photo E. Halme.

distribution of magnetite, whilst pyrite occupies between 1 and 5 Vol. per cent of the total ore mineral content.

The mineral composition of the brecciated ore is more variable than that of the disseminated ore. In addition to the minerals cited above, the following also occur in the brecciated ore of the Hällinmäki copper deposit: bornite, mackinawite, pentlandite, bravoite, sphalerite, molybdenite, linneite minerals, gersdorffite and millerite. These minerals occur only in microscopic amounts and they usually form exsolution bodies in the chalcopyrite or the pyrrhotite.

Especially in the lenses of the brecciated ore, the mineral grains are considerably coarser than those in the disseminated ore: they range up to 3 to 4 centimeters in diameter. The exsolution bodies are only 0.01 to 3 mm long, except in the case of cubanite, where lamellae as long as 3 or 4 cm have been observed.

## Ore Minerals

### *Chalcopyrite and cubanite*

Chalcopyrite is the principal ore mineral in the Hällinmäki ore. In the dissemi-

nated ore it usually occurs as isolated grains, which rarely contain inclusions or exsolution bodies. In contrast, the chalcopyrite of the brecciated ore contains an abundance of inclusions and exsolution bodies, of which the most important are cubanite and pyrrhotite.

At Hällinmäki cubanite is only present in ore body II (Fig. 26) where it is thought to be an exsolution from chalcopyrite. The elongate cubanite lamellae often extend across the chalcopyrite crystals from edge to edge and the largest may be 3 to 4 cm

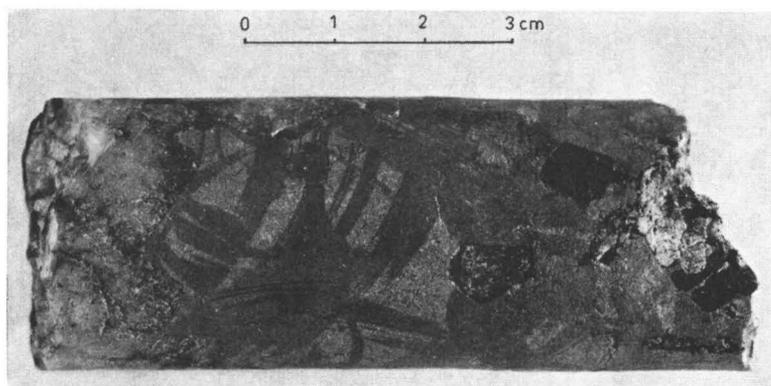


FIG. 32. Exsolution lamellae of cubanite (cb) in chalcopyrite. The figure also includes magnetite crystals. Photo E. Halme.

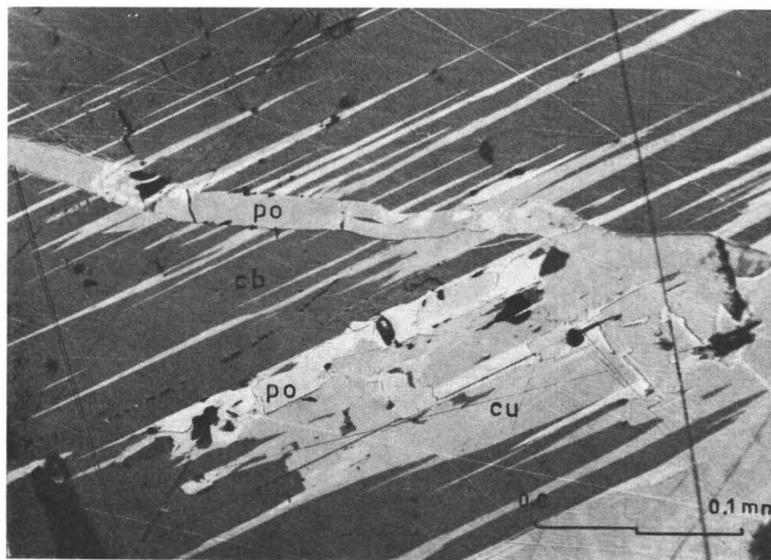


FIG. 33. Pyrrhotite (po) intersects cubanite (cb) and chalcopyrite (cu). The chalcopyrite exhibits twinning. Hällinmäki. Nic partly +. Photo E. Halme.

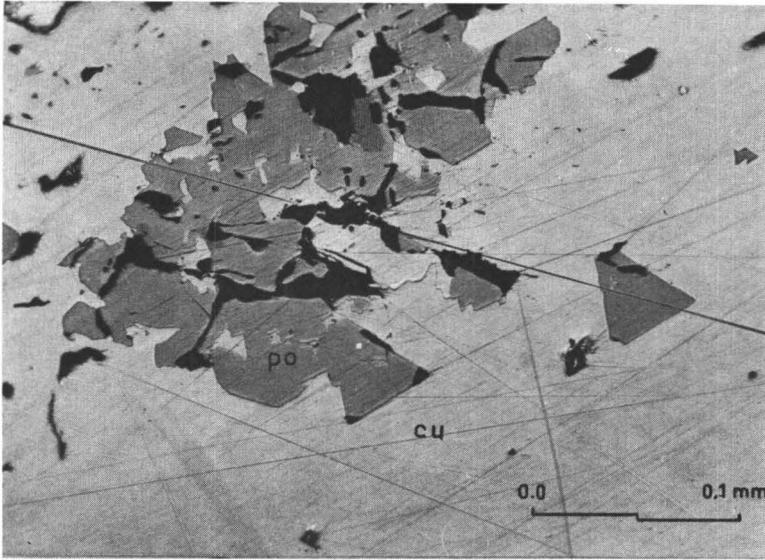


FIG. 34. Exsolution intergrowth of chalcopyrite (cu) and pyrrhotite (po). Nic partly +. Photo E. Halme.

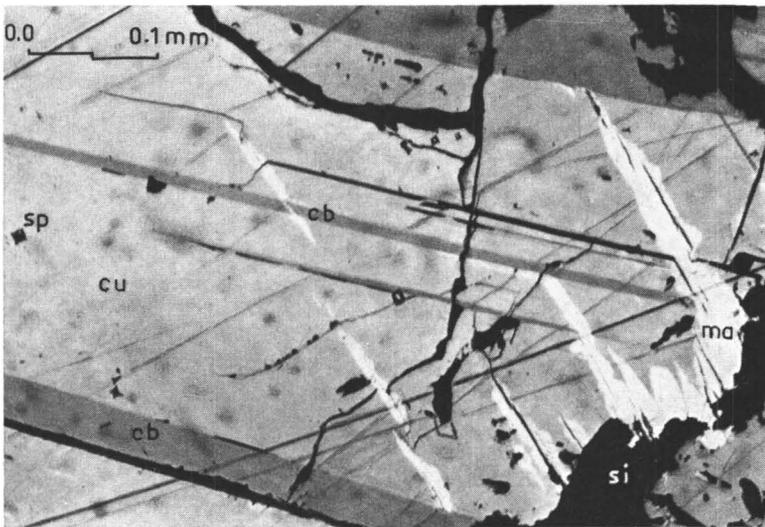


FIG. 35. Slivers of mackinawite (ma) intersect cubanite lamellae (cb). cu = chalcopyrite, si = silicates and sp = sphalerite stars. Hällinmäki. Nic partly +. Photo E. Halme.

long and 0.5 to 1.0 cm wide (Fig. 32). The cubanite also forms clusters of equidimensional exsolution bodies. Often associated with the cubanite lamellae are irregularly or regularly shaped inclusions of pyrrhotite. The cubanite lamellae and elongate pyrrhotite blebs generally run parallel, but in certain instances the pyrrhotite may intersect the cubanite lamellae (Fig. 33). The regular shape of the pyrrhotite bodies suggests that they are also exsolution bodies in chalcopyrite (Fig. 34).

In addition to cubanite and pyrrhotite, pentlandite and sphalerite have likewise unmixed from the chalcopyrite. Sphalerite (whose most common form is as stars) and mackinawite (which occurs as either irregular needles or starlike shapes resembling sphalerite) intersect the cubanite (Fig. 35).

The chalcopyrite is invariably twinned. The lamellae are in many instances lancet-like and biconcave in form, and they are intersected by cubanite lamellae (Fig. 36). However, the cubanite may also be twinned (Figs. 36 and 37).

Cubanite, pyrrhotite, pentlandite and sphalerite have thus unmixed from chalcopyrite, cubanite being the first mineral to exsolve. The unmixing is a later development than the twinning of the chalcopyrite, for according to Yund et al. (1966) the twinning takes place at  $547^{\circ} \pm 5^{\circ}$  C and the cubanite unmixes from the chalcopyrite at approximately  $500^{\circ}$ C (see also p. 73).

The coarse cubanite lamellae often contain bands of pyrrhotite and pentlandite up to 3 to 4 mm long and 0.01 to 0.02 mm wide (Fig. 38). They have apparently undergone the unmixing almost simultaneously, as they intersect each other. The pentlandite is a cobalt pentlandite (Fig. 39). The bands of pyrrhotite also contain

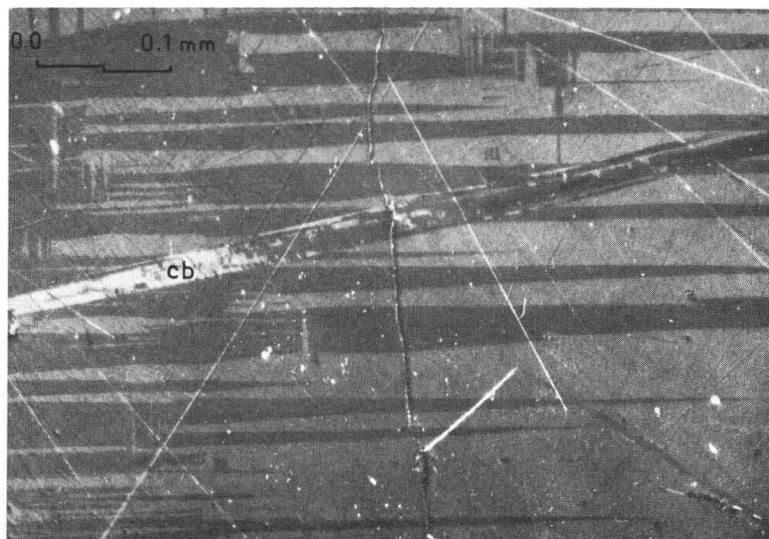


FIG. 36. Cubanite lamellae (cb) intersects twinned lamellae of chalcopyrite. The cubanite is likewise twinned. Hällinmäki. Nic +. Photo E. Halme.

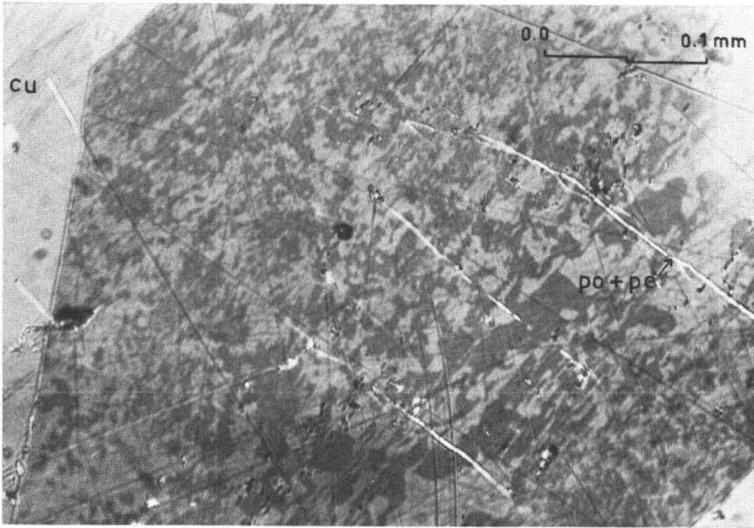


FIG. 37. Twinned cubanite. cu = chalcopyrite and po + pe = pyrrhotite and pentlandite exsolution bodies, which cut across the twinned lamellae of cubanite. Hällinmäki. Nic +.

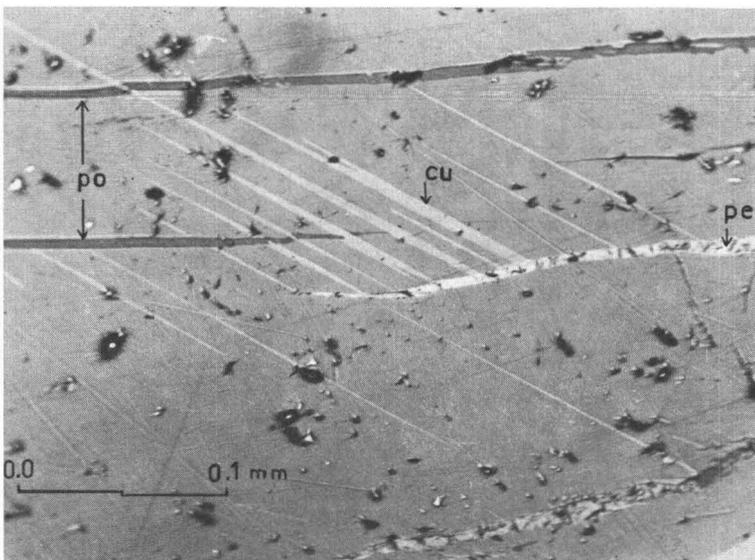


FIG. 38. Exsolution bodies of cubanite. po = pyrrhotite, pe = pentlandite and cu = chalcopyrite. Hällinmäki. Nic partly +. Photo E. Halme.

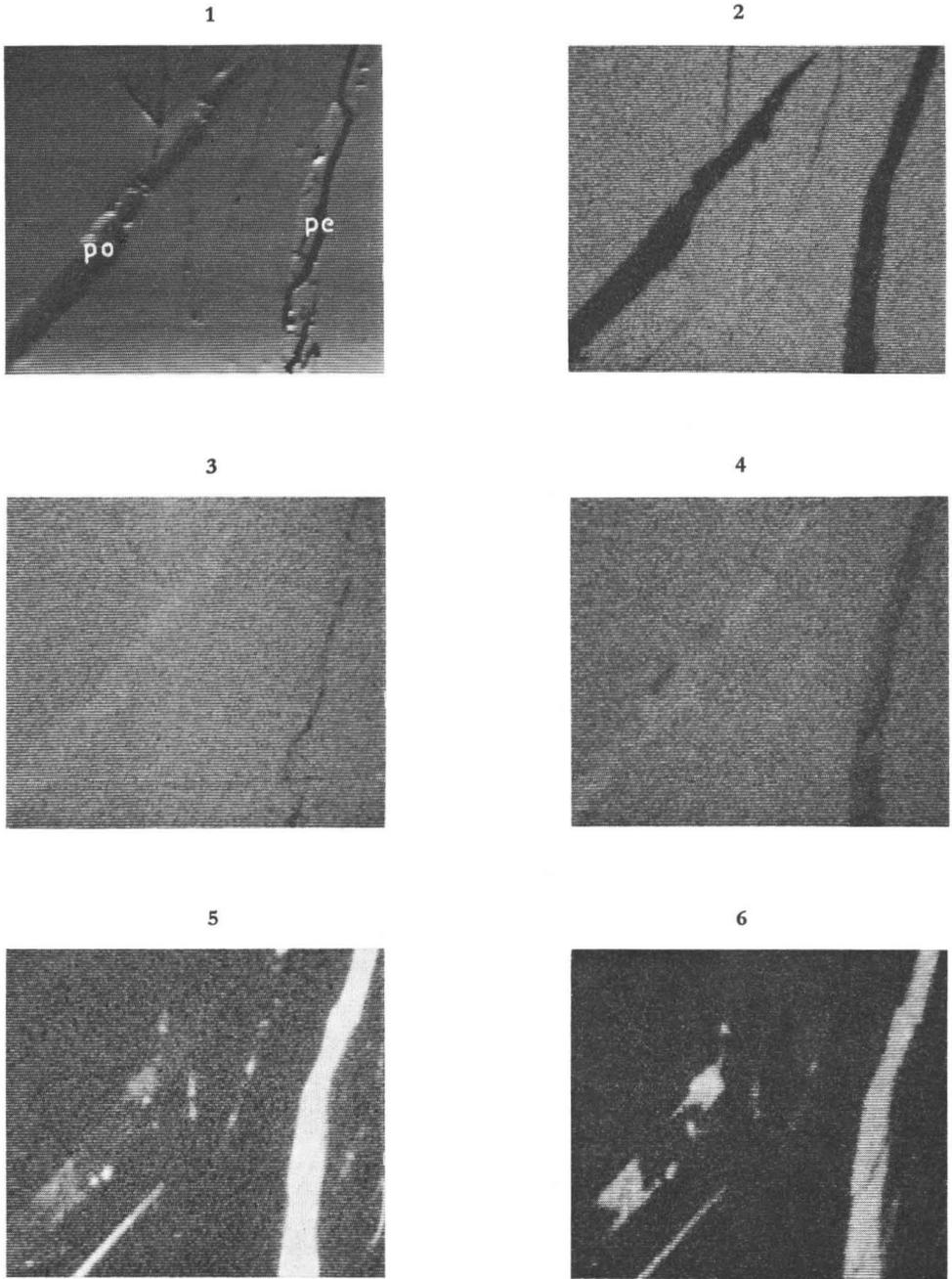


FIG. 39. 1 = Electron image. Pyrrhotite (po) and pentlandite (pe) bands in cubanite, 2 =  $\text{CuK}_{\alpha 1}$ , 3 =  $\text{SK}_{\alpha 1}$ , 4 =  $\text{FeK}_{\alpha 1}$ , 5 =  $\text{NiK}_{\alpha 1}$  and 6 =  $\text{CoK}_{\alpha 1}$ , Hällinmäki. Magnification 210  $\times$ . Analyzed by J. Siivola.

other Ni-Co-Fe sulfides as indicated by the distribution of these metals in the exsolution bodies of cubanite (Fig. 39).

If there is an abundance of cubanite, the large cubanite lamellae contain thin, elongated inclusions of chalcopyrite. These are also probably exsolution bodies and they intersect the pyrrhotite and pentlandite bands (Fig. 38). First the pyrrhotite and the pentlandite unmixed from the cubanite and then the chalcopyrite.

In places, the chalcopyrite and the cubanite contain 0.02 to 0.06 per cent of cobalt.

### *Bornite*

Bornite occurs at both the northwest end and the middle of the Hällinmäki ore deposit as scattered grains a few tenths of a mm in diameter. At the southeastern end of the occurrence, a little beyond the ore body V, the bornite content increases slightly with increasing grain size: the larger crystals are up to 1 to 2 cm in diameter. These large crystals in many instances contain idiomorphic grains of pyrite. Pyrrhotite and bornite do not occur together.

### *Pyrrhotite*

Pyrrhotite is a common mineral in the Hällinmäki ore but it has a variable composition (Table 17). The exsolution bodies in the pyrrhotite are pentlandite and sphalerite. They are not evenly distributed, for at the edges of the pyrrhotite there are considerably more exsolution bodies than at the center of the grains. Wherever the pyrrhotite is the principal mineral and there is only a little chalcopyrite, the latter mineral is generally present as tiny inclusions in the margins of the pyrrhotite grains. Ore body III occasionally contains pyrrhotite-rich pockets of this type.

When the pyrrhotite is associated with cubanite it has generally unmixed from the chalcopyrite. As Table 17 shows, the pyrrhotite associated with cubanite contains more exsolution bodies and smaller amounts of Ni + Co than where there is no cubanite. The  $d_{102}$  values, the Ni + Co contents and the exsolution bodies, show a clear correlation: the  $d_{102}$  values diminish as the Ni + Co contents increase and the quantity of exsolution bodies decreases. Outside the ore body II the  $d_{102}$  of the pyrrhotite is  $2.058 \pm 0.001 \text{ \AA}$ . This value corresponds therefore to the  $d_{102}$  values of the pyrrhotites contained in the cubanite-free parts. In the Hällinmäki ore the pyrrhotite generally contains Cu 1 Wt per cent.

Around the pyrite-bearing veins cutting the pyrrhotite, the latter alters into pyrite. The alteration begins at the margins of the grains and the fissures. Magnetite crystals created during regional metamorphism (*cf.*, p. 34) and located next to chalcopyrite, are occasionally altered to pyrrhotite around their margins.

TABLE 17

Co and Ni contents of the pyrrhotite, chalcopyrite and cubanite of the Hällinmäki II ore body. Minerals were analyzed by K. Laajoki with a microprobe. The  $d_{102}$  values were determined by P. Kallio

Thin section No.	Pyrrhotite	Chalcopyrite	Cubanite	Remarks
8707 depth 96 m	$d_{102} = 2.054 \text{ \AA}$ $\left. \begin{array}{l} \text{Co} = 0.43 \% \\ \text{Ni} = 0.46 \% \end{array} \right\} 0.89 \%$ $\text{Ni/Co} = 1.07$	$\text{Co} = 0.05 \%$ $\text{Ni} = 0.00 \%$	Ore does not contain cubanite	Very few exsolution bodies occur in pyrrhotite
8804 depth 200 m	$d_{102} = 2.057 \text{ \AA}$ $\left. \begin{array}{l} \text{Co} = 0.28 \% \\ \text{Ni} = 0.36 \% \end{array} \right\} 0.64 \%$ $\text{Ni/Co} = 1.29$	$\text{Co} = 0.06 \%$ $\text{Ni} = 0.00 \%$	Ore does not contain cubanite	As above
8652 depth 15 m	$d_{102} = 2.067 \text{ \AA}$ $\left. \begin{array}{l} \text{Co} = 0.21 \% \\ \text{Ni} = 0.28 \% \end{array} \right\} 0.49 \%$ $\text{Ni/Co} = 1.33$	$\text{Co} = 0.06 \%$ $\text{Ni} = 0.00 \%$	$\text{Co} = 0.06 \%$ $\text{Ni} = 0.07 \%$ $\text{Ni/Co} = 1.17$	Some exsolution bodies in pyrrhotite
8729 depth 70 m	$d_{102} = 2.074 \text{ \AA}$ $\left. \begin{array}{l} \text{Co} = 0.13 \% \\ \text{Ni} = 0.11 \% \end{array} \right\} 0.25 \%$ $\text{Ni/Co} = 0.84$	$\text{Co} = 0.04 \%$ $\text{Ni} = 0.00 \%$	$\text{Co} = 0.04 \%$ $\text{Ni} = 0.00 \%$	Pyrrhotite contains abundant exsolution bodies
8549 depth 35 m	$d_{102} = 2.081 \text{ \AA}$ $\left. \begin{array}{l} \text{Co} = 0.03 \% \\ \text{Ni} = 0.04 \% \end{array} \right\} 0.07 \%$ $\text{Ni/Co} = 1.33$	$\text{Co} = 0.02 \%$ $\text{Ni} = 0.00 \%$	$\text{Co} = 0 \%$ $\text{Ni} = 0 \%$	Pyrrhotite contains great abundance of exsolution bodies

### *Mackinawite and pentlandite*

Mackinawite is commonly found in the Hällinmäki ore in microscopic amounts. It is usually present in the chalcopyrite, where it forms a flame-like pattern averaging 0.1 to 0.2 mm in length and 0.01 mm in width. They are often clustered around, or very close to, fissures in the chalcopyrite. The fissures contain, inter alia, pyrite. According to microprobe determinations, the mackinawite of Hällinmäki contains on average, 1 per cent Co and 5 per cent Ni (K. Laajoki).

As with mackinawite, pentlandite is a mineral that occurs generally but in trace amounts. It is most abundant in the pyrrhotite, where it occurs as exsolution bodies a few tenths of a mm across and roundish or flame-like in shape. It has also been observed as exsolution bodies in cubanite and in chalcopyrite. In general, the exsolution bodies in the pyrrhotite are located on the margins of the grains. The pentlandite is cobalt pentlandite, in which there is, on the average, 17 or 18 per cent Co and 20 to

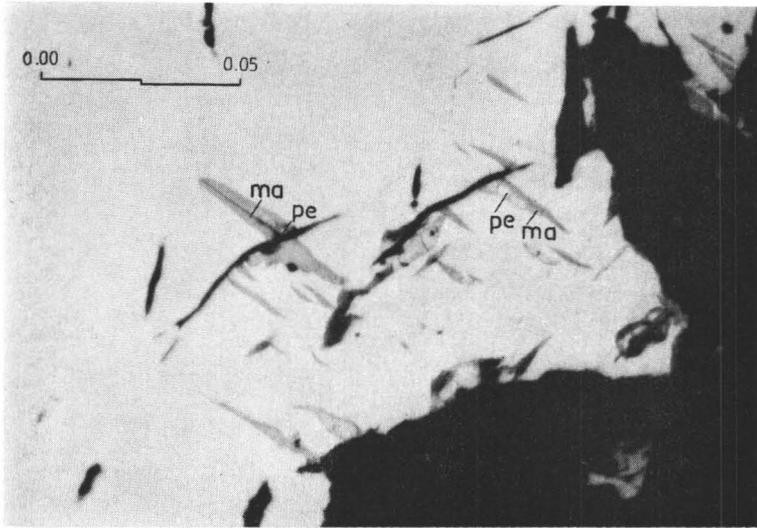


FIG. 40. In the middle of the mackinawite (ma) needles are remains of pentlandite (pe). Hällinmäki. Nic partly +. Photo E. Halme.

23 per cent Ni (microprobe determination by J. Siivola). The Ni/Co is thus slightly greater than one.

Genetically the mackinawite and the pentlandite appear to be closely related. In places, the pentlandite at the margins of the pyrrhotite grains has changed into mackinawite. Antun, El Goresy and Ramdohr (1966) have observed similar alterations. More rarely, the pentlandite bands in the cubanite change to mackinawite. In chalcopyrite the centers of the mackinawite needles often contain remnants of pentlandite (Fig. 40) or the margins of the pentlandite needles are only slightly changed to mackinawite.

### *Pyrite*

Pyrite is present in only small amounts at Hällinmäki, having formed either by primary crystallization or as an alteration product of pyrrhotite. As a primary product, pyrite has crystallized in the thin chlorite-quartz veins penetrating the other ore minerals of the deposit, or in their immediate proximity. Magnetite, millerite and linneite crystallized simultaneously with the pyrite. Close to the quartz-chlorite veins the pyrite replaces the chalcopyrite in a graphic-like texture. Where pyrite crystallized in the veins penetrating the cubanite, some solution of cubanite from around the pyrite crystals has taken place (Fig. 41). Microprobe determinations indicate that the Hällinmäki pyrite is a very pure iron sulfide (J. Siivola).

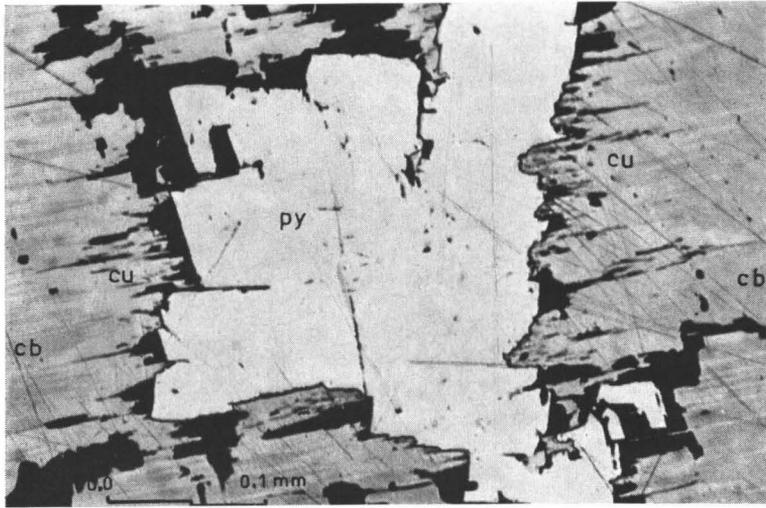


FIG. 41. Part of a large cubanite lamella. A silicate dike (black) has pierced the cubanite (cb) of which pyrite (py) has crystallized. Enveloping the pyrite is a chalcopyrite rim (cu). Hällinmäki. One Nicol.

#### *Linneite minerals and millerite*

Only small amounts of the linneites and millerite occur in the Hällinmäki ore, either by replacing chalcopyrite or by filling the spaces between idiomorphic pyrite crystals. Microprobe analyses show that the Hällinmäki ore contains two minerals belonging to the linneite group, of which one is violarite and the other a mixture of violarite and siegenite. The latter contains cobalt and nickel in nearly equal amounts. The millerite is cobalt-bearing.

#### *Sphalerite*

Sphalerite is widely spread throughout the Hällinmäki deposit, but the amounts are invariably quite small. It occurs only as exsolution bodies in the chalcopyrite and pyrrhotite.

#### *Molybdenite*

Molybdenite is unevenly distributed. In general it occurs as separate flakes, but in some instances there are chalcopyrite inclusions in its margins. Wherever the copper content exceeds 1 per cent, molybdenite is absent.

*Gersdorffite*

Only a few idiomorphic crystals of gersdorffite, measuring 0.2 mm in diameter, occur in the chalcopyrite.

*Magnetite and ilmenite*

Magnetite is unevenly distributed. In places it forms compact pockets a few meters long. In such pockets the roughly idiomorphic magnetite crystals have an average diameter of less than 1 mm. They generally contain minor quantities of ilmenite. Ilmenite crystals are also scattered among the magnetite crystals. The magnetite in the pockets developed during the regional metamorphism, (see page 34).

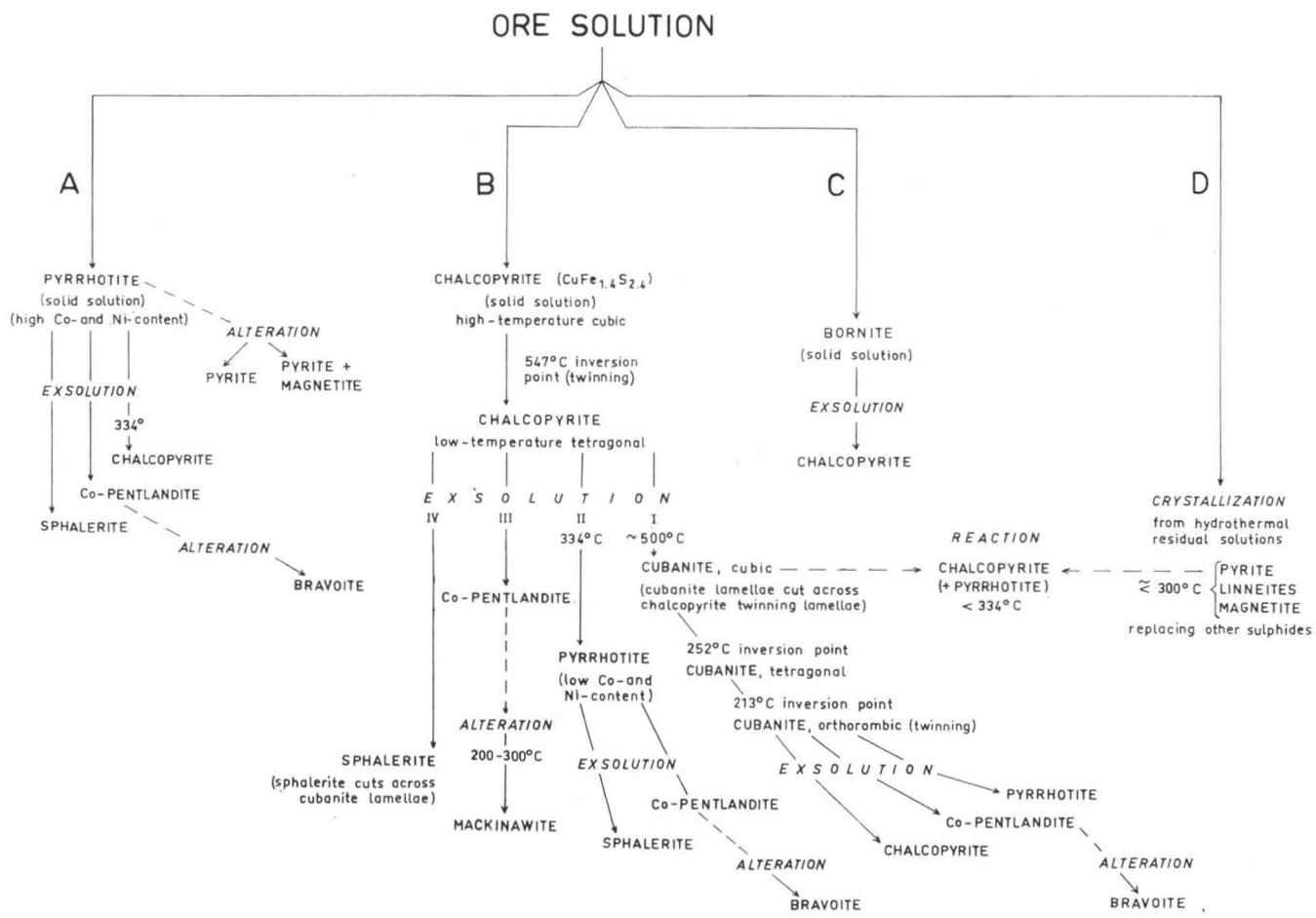
The magnetite that crystallized with the pyrite-veins contains no ilmenite. In the quartz-chlorite veins, or their close proximity, magnetite crystals exceeding 1 cm in diameter have been observed.

**Crystallization, exsolution and alteration of the sulfide minerals**

In the elucidation of the crystallization of the sulfide phase of the Hällinmäki occurrence, the results of the experimental investigations (Arnold 1962, Clark 1966, Kullerud 1962 and 1964, Sugaki 1965 and Yund et al. 1966) were considered in conjunction with the ore-microscopic observations. It should be noted that the experimental work is based on dry systems and as such must be applied, with reservations, to natural systems. The pressure of water lowers the liquidus temperature so that under exceedingly high water pressure, the temperature may be reduced to the extent that exsolution textures no longer form. The abundant exsolution phenomena in the ore minerals of the Hällinmäki deposit suggest, that a high temperature prevailed during the crystallization of the sulfide phase.

According to Kullerud (1964), there occurs in the Cu-Fe-Ni-S system above a temperature of 813°C a state of liquid immiscibility. Such high temperatures probably could not have existed, however, during the crystallization of the Hällinmäki sulfide phase. Microscopic observations indicate that, pyrrhotite, chalcopyrite, and bornite were the first minerals to crystallize. After the crystallization of these minerals, pyrite, the minerals of the linneite group and the magnetite crystallized from the hydrothermal residual solutions. These minerals replace pyrrhotite, chalcopyrite and bornite. In determination of the crystallization sequence, chalcopyrite replaces the pyrrhotite. It has not been possible to determine the crystallization sequence of pyrrhotite and the bornite or of chalcopyrite and bornite. The crystallization temperature of bornite had, however, probably been lower than that of other minerals (*cf.*, p. 74).

Fig. 42. Crystallization of the ore solution of the Hällimäki ore.



The crystallization of the sulfide phase of the Hällinmäki deposit may be illustrated graphically (Fig. 42). The figure represents the crystallization sequence of the different sulfides from left to right. Gersdorffite and molybdenite are not included in the diagram because it is difficult to determine their relation to the other. The temperatures given in the diagram are from the experimental works referred to and as such are evidently too high.

*A. Solid solution of pyrrhotite.* The solubility of copper in pyrrhotite exceeds 3 Wt per cent at 700°C and as much as c. 2 Wt per cent at 600°C (v. Gehlen and Kullerud 1962). This is one of the reasons why pyrrhotite cannot be used in the Hällinmäki copper ore as a geological thermometer, because the thermometer is based on a pure Fe-S system (Arnold 1962). Besides, chalcopyrite and pyrrhotite are a stable mineral pair in the Cu-Fe-S system only below 334°C (Yund et al. 1966). Moreover, the high Ni + Co content of the pyrrhotite may lead to errors in determining the crystallization temperature of the pyrrhotite by means of its  $d_{102}$  value, for on p. 67 the correlation between the  $d_{102}$  values, Ni + Co contents and exsolution bodies were noted. This correlation probably reflects the drop in temperature following the crystallization of the sulfide phase: the higher the Ni + Co content and the lower the  $d_{102}$  value, the more rapid has been the drop in temperature.

At the southeastern end of the Hällinmäki deposit, there are lenses of pyrrhotite, which has a typically high Ni + Co content, along with smaller amounts of exsolutions of chalcopyrite, pentlandite and sphalerite. Since the pyrrhotite and the chalcopyrite form a stable mineral pair only below a temperature of 334°C, the unmixing of these two minerals had to take place below this level.

The solid solution of pyrrhotite thus originally included copper, zinc, nickel and cobalt. The high Ni + Co contents and small quantities of exsolutions bodies suggest that the cooling process was fairly rapid.

*B. Solid solution of chalcopyrite.* Characteristic of high-temperature (cubic) chalcopyrite is the fact that it dissolves abundant pyrrhotite and at its maximum solubility, corresponds approximately to  $\text{CuFe}_2\text{S}_3$  (Yund et al. 1966). Calculated according to point counting analyses of the ore minerals of the ore body II of Hällinmäki (Fig. 26), the solid solution of the chalcopyrite would give the formula,  $\text{CuFe}_{1.4}\text{S}_{2.4}$ .

The high-temperature cubic chalcopyrite inverts at  $547^\circ \pm 5^\circ\text{C}$  into low-temperature tetragonal chalcopyrite (Yund et al. 1966). Yund et al. (1966) have further observed that when synthetic chalcopyrite is cooled below the inversion temperature, twinning takes place in the chalcopyrite. The twinning in the Hällinmäki chalcopyrite could also have formed in this way. In the solid solution of chalcopyrite and cubanite, there occurs a miscibility gap below 500°C (Yund et al. 1966), which means that below this temperature cubanite must unmix from the chalcopyrite.

After the unmixing of cubanite from the chalcopyrite, exsolution of pyrrhotite, cobalt pentlandite and sphalerite took place. The temperature at which the pyrrhotite

unmixes from the chalcopyrite is 334°C (Yund *et al.* 1966). Also according to microscopic observations, the pyrrhotite unmixes later than the cubanite. Sphalerite stars cut across the cubanite lamellae and pyrrhotite inclusions. The pentlandite exsolution bodies likewise cut across the cubanite (*cf.*, p. 64).

Further unmixing from the cubanite has taken place (Fig. 38). The cubanite is cubic above 252°C, tetragonal at 252° to 213°C, and orthorhombic below 213°C (Yund *et al.* 1966). Ramdohr (1960, p. 581) has observed inversion lamellae in cubanite and attributes it to hexagonal-orthorhombic inversion. Yund *et al.* (1966), however, regard Ramdohr's hexagonal cubanite as tetragonal. The pyrrhotite, cobalt pentlandite and chalcopyrite possibly unmix below 213°C, for they cut across the cubanite lamellae. Yet, there is still the possibility that the unmixing took place before the formation of the lamellae.

From the pyrrhotite exsolutions in the chalcopyrite have further unmixed cobalt pentlandite and sphalerite. Exsolution bodies are abundant and the Ni + Co contents of the pyrrhotite are very low.

*C. Solid solution of bornite.* Bornite is capable of dissolving chalcopyrite from 6.6 to 40.6 mol per cent between the temperatures of 375° and 525°C (Sugaki 1965). The value of  $a_0$  of bornite is a function of the amount of dissolved chalcopyrite. At Hällinmäki the bornite has an  $a_0 = 10.95 \text{ \AA}$ , corresponding to fairly pure bornite (Sugaki 1965). Originally, the bornite phase apparently contained little chalcopyrite, as few chalcopyrite exsolution bodies have been observed. The low content of chalcopyrite exsolution bodies indicates a fairly low crystallization temperature for the bornite solid solution; hence its position in the C column of the model seems to be justified.

*D. Ore minerals crystallized from hydrothermal solutions.* The hydrothermal solutions also contained siliceous material, for the ore minerals of this group have also partially crystallized in the quartz-chlorite veins (p. 69). Violarite is stable, according to Kullerud (1962), only below a temperature of 300°C. Below 334°C pyrite and cubanite react, producing chalcopyrite and pyrrhotite (Yund *et al.* 1966). It was observed that around the pyrite crystals in the quartz-chlorite veins penetrating the cubanite, there formed a rim of chalcopyrite; this implies that the pyrite crystallized below 334°C at Hällinmäki.

Bornite and pyrite form a stable mineral pair between 568° and 228°C (Yund *et al.* 1966), and since the bornite contains idiomorphic pyrite crystals in the proximity of the quartz-chlorite veins, this suggests that the pyrite crystallized at  $> 228^\circ\text{C}$ . All the ore minerals of this group thus seem to have crystallized at  $\sim 300^\circ\text{C}$ .

The alterations in the sulfide minerals are at least in part due to the hydrothermal solutions mentioned previously (p. 67). For example, the alteration of pyrrhotite to pyrite in the quartz-chlorite vein penetrating the pyrrhotite suggest that the alteration was caused by the hydrothermal solution. The alteration of the pentlandite exsolution

bodies into bravoite may also be attributed to the increase in the S vapor pressure. This alteration took place, however, at very low temperature, for bravoite and pyrite are stable together only below 137°C (Kullerud 1962).

Mackinawite has only been observed in chalcopyrite and pyrrhotite. In the Hällinmäki deposit, the mackinawite is an alteration product of pentlandite, possibly induced by the presence of hydrothermal solutions. According to Clark (1966, p. 302), the upper stability limit of mackinawite, in which Ni + Co = 6 Wt per cent, is between 200° and 230°C. It is in this temperature range that the alteration of the Hällinmäki pentlandite into mackinawite also probably took place.

Besides the alterations as a result of hydrothermal solutions, the sulfides of Hällinmäki also show changes associated with weathering phenomena. Thus the alteration of pyrrhotite into pyrite and magnetite may be explained as the result of the action of acid solutions flowing downward (Ramdohr 1960, p. 559). Ramdohr describes the phenomenon as occurring in the following way:  $6 \text{ FeS} + 40 = \text{Fe}_3\text{O}_4 + 3\text{FeS}_2$ . The alteration of the chalcopyrite into chalcocite, beginning at the edges of the grains in cleavages, is weathering phenomenon, too, for it has been observed only in the surface parts of the ore.

### The Karhuniemi deposit

The copper deposit of Karhuniemi is located at the southern end of the diopside amphibolite zone about one kilometer to the northeast from Hällinmäki. Owing to the sparseness of outcrops, the structure of the local bedrock has been elucidated by using material obtained from drill cores and geophysical maps (Siikarla 1967).

As the block diagram in Fig. 43 shows, the diopside amphibolite at Karhuniemi is broken up by plutonic rocks. The parts of the schist sequence separated by quartz diorite and diorite make an arch in the southern section of the area and possibly form the end of a fold. The diopside amphibolite contains intercalations of garnet skarn. The rocks are the same varieties occurring at Hällinmäki.

At Karhuniemi, too, the ore minerals are either associated with garnet skarns or situated in their immediate proximity. They do not, however, form any extensive, uniform mineralized zone but only isolated pockets. Further, as the ore minerals consist in places almost exclusively of iron sulfides, the Karhuniemi occurrence is not of economic importance.

The main minerals are magnetite, pyrrhotite and chalcopyrite along with minor pyrite, pentlandite, sphalerite and mackinawite. Magnetite is the most abundant ore mineral. Its banded structure suggests that the magnetite originated wholly during the period of regional metamorphism. It is brecciated by sulfides.

The sulfides occur either as brecciated or as disseminated ore. Pyrrhotite, which contains pentlandite and sphalerite exsolution bodies to a small extent, occurs in

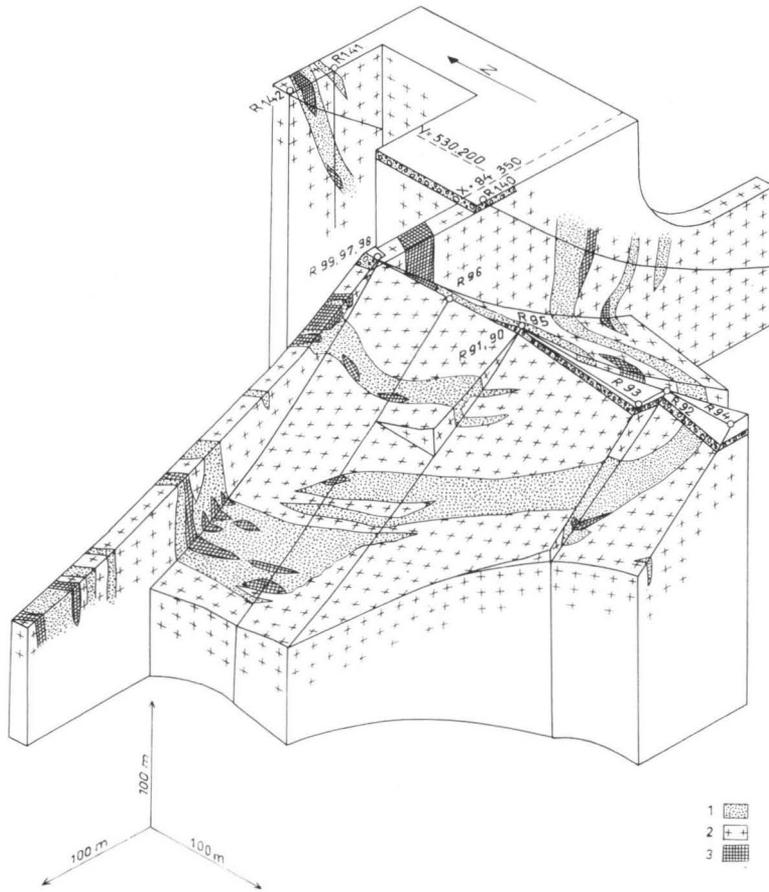


FIG. 43. The bedrock of Karhuniemi. 1 = diopside amphibolite, 2 = diorite and quartz diorite and 3 = garnet skarn. The block diagram was drawn by J. Pekkarinen.

larger amounts than do the other sulphides. The  $d_{102}$  values and Ni and Co contents for the pyrrhotite are as follows:

	$d_{102}^1)$	Ni Wt % <sup>2)</sup>	Co Wt % <sup>2)</sup>	Ni/Co
R 144/12.65 m . . . .	2.062	0.10	0.26	0.39
R 99/48.30 m . . . .	2.058	0.22	0.28	0.80
R 51/51.15 m . . . .	2.056	—	—	—
R 91/45.15 m . . . .	2.061	—	—	—

<sup>1)</sup> Analyst: P. Kallio

<sup>2)</sup> Analyst: K. Laajoki

The Ni/Co of the bulk ore is slightly higher, which may be due to the pentlandite unmixing from the pyrrhotite:

	Cu Wt %	Ni Wt %	Co Wt %	Ni/Co
R 91/ 43— 46 m ..	0.32	0.01	0.01	1.0
R 97/161—166 m ..	0.15	0.03	0.03	1.0

The pyrrhotite commonly alters into pyrite. The alteration is probably in part due to weathering and also to the increase in S-pressure induced by the hydrothermal solutions, as at Hällinmäki (*cf.*, p. 74).

Chalcopyrite generally occurs in rather small amounts. Ordinarily, it occurs as tiny inclusions at the edges of pyrrhotite grains. Wherever the chalcopyrite is abundant, it contains small amounts of sphalerite and/or pentlandite exsolution bodies. In some rare instances, pentlandite has altered to mackinawite.

### Lari's deposit

Lari's copper deposit is situated approximately three kilometers to the north-west from the Hällinmäki field. The deposit was found under a layer of peat about 20 centimeters thick by a dog named Lari, that had been trained to locate occurrences of sulfide ore by his sense of smell. In this area, too, the ore minerals are contained in garnet skarn, which occur as intercalations in diopside amphibolite. In the close proximity of the garnet skarn is quartz diorite (Fig. 44 and appended map).

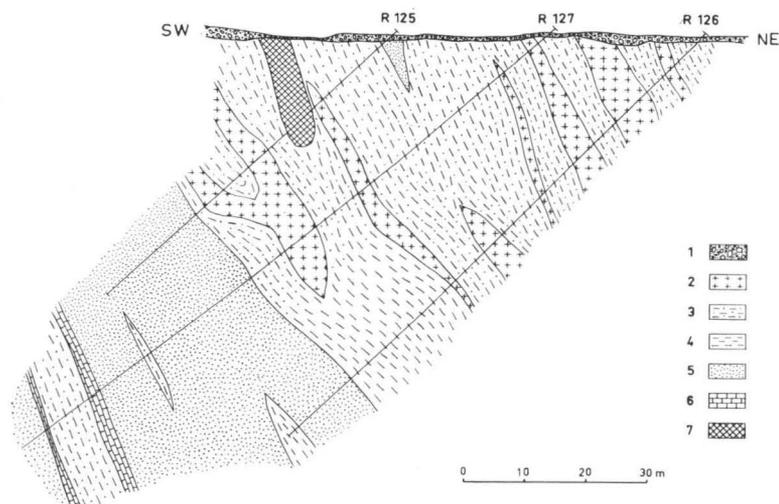


FIG. 44. Copper deposit of Lari. 1 = till, 2 = quartz diorite, 3 = diopside amphibolite, 4 = amphibolite, 5 = quartz-feldspar gneiss, 6 = calcite and 7 = garnet skarn.

The garnet skarn contains abundant ore minerals. The occurrence is very small, however; on the basis of geophysical maps and drillings tests, it is estimated to contain only few hundred tons of ore with an average content of 1.5 % Cu.

The main minerals are chalcopyrite and pyrrhotite, with minor cubanite, pyrite, pentlandite, mackinawite, sphalerite and magnetite. Chalcopyrite is the most abundant of the ore minerals, and it contains many pyrrhotite exsolutions and a few cubanite, pentlandite and sphalerite exsolution bodies. Some of the pentlandite has altered into mackinawite (*cf.*, p. 69). Unlike the Hällinmäki occurrence, the cubanite does not contain other exsolution bodies or inclusions of ore minerals. Pyrite and magnetite occur in small amounts; but they rarely coexist.

The main part of the pyrrhotite did not unmix from the solid solution of chalcopyrite but crystallized directly from the ore solution. The cooling of the pyrrhotite was probably fairly rapid, as indicated by the paucity of pentlandite and sphalerite exsolution bodies and fairly high Ni and Co contents:

	$d_{102}^1)$	Ni Wt % <sup>2)</sup>	Co Wt % <sup>2)</sup>	Ni/Co
R 125/19.77 m . . . .	2.063	—	—	—
R 125/22.15 m . . . .	2.068	0.14	0.23	0.61

<sup>1)</sup> Analyst: P. Kallio

<sup>2)</sup> Analyst: J. Siivola

The alteration of pyrrhotite into pyrite in Lari's copper deposit is evidently a phenomenon associated with weathering, judging by the fact that in conjunction with the crystallization of the hydrothermal solutions, no alteration of the pyrrhotite appears to have taken place.

## SUMMARY

The supracrustal rocks of the Virtasalmi area were originally aqueous deposits — that is, graywackeous and other sands, clay, limestone and ooze. During sedimentation volcanic actions also took place, with the result that volcanic material was intermixed with the sediments.

The regional metamorphism represents the conditions of the amphibolite facies. The temperature did not, however, rise high enough for the silicic schists to undergo partial melting. The alternation of light and dark bands in the veined gneiss may be due to differences in translation efficiency of the various minerals in response to shear movements; the more mobile crystals detached themselves from the less mobile ones to form separate layers. The segregation of different minerals likewise took place for the same reason in the diopside amphibolites, with the hornblende forming separate pockets. The diopside amphibolite contained iron-rich intercalations, which altered during the regional metamorphism into garnet-skarn zones

containing magnetite lenses. During folding, these zones hardened and became brittle while the surrounding diopside amphibolite remained in a plastic state. The mica gneisses occurring in the proximity of the calcareous sediments or as intercalations in them do not in general contain cordierite. This mineral is abundantly present, however, in mica gneisses of similar composition located a few kilometers from the diopside amphibolite and limestone zones. With the onset of regional metamorphism, so much  $\text{CO}_2$  and  $\text{H}_2\text{O}$  formed as reaction products of the calcareous sediments that the pressure of the gas phase increased considerably. As a result, the stability field of cordierite was probably exceeded in the proximity of the diopside amphibolite and limestones. During the initial stage of regional metamorphism, the partial pressure of  $\text{CO}_2$  was so slight that the wollastonite reaction was triggered, but as the metamorphism of the calcareous rocks continued, the partial pressure of  $\text{CO}_2$  increased and the reaction halted.

With the exception of the aplite and pegmatite veins, the plutonic rocks (peridotites, gabbros, diorites, quartz diorites and trondhjemites) of the Virtasalmi district are synorogenic. These rocks are characterized by a fairly high  $\text{Na}_2\text{O}$  content and low  $\text{H}_2\text{O}$  and  $\text{K}_2\text{O}$  contents. A further characteristic of these rocks is the fact that the magnetite did not crystallize with the peridotites and gabbros but with the diorites and quartz diorites.

The folding seems to be isoclinal. The regional axis is subhorizontal. The minor folding upon subvertical axes seems mostly represent drag folding on the limbs of the major folds.

Neither the base of the sedimentation nor the basal formations are known. The lowest member appears to be mica gneiss, which is overlain in the southwestern part of the area by pyroxene gneisses and limestones and in the northeastern part by quartz-feldspar gneisses. Moving upward in the stratigraphic sequence, there follow the amphibolites and finally, once more, mica gneisses.

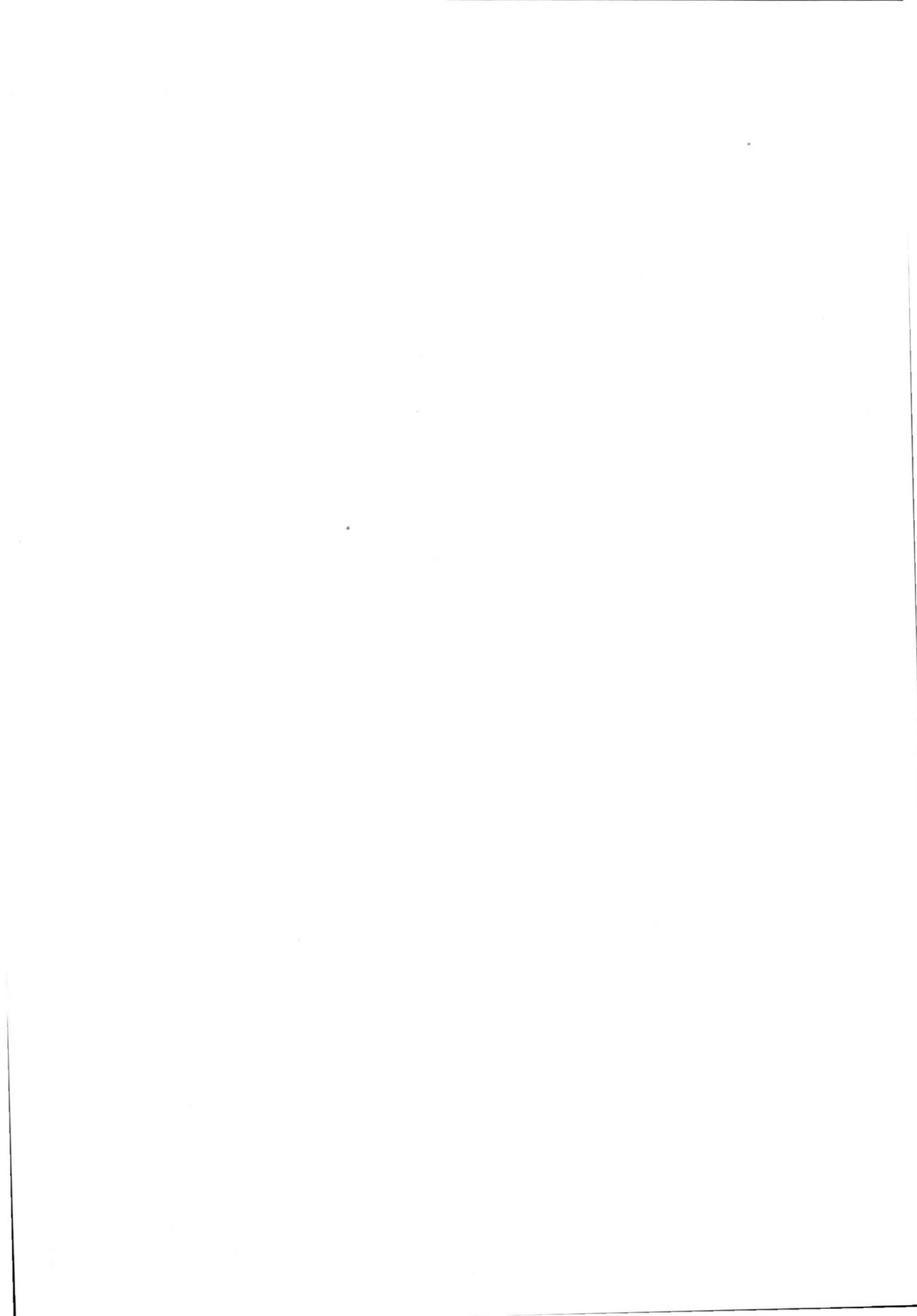
The following general features have been observed in the copper-bearing skarn ores of the Virtasalmi district: They invariably occur either inside or in the immediate proximity of garnet skarn layers contained in diopside amphibolite at contacts of plutonic rocks. The sulfides form brecciated or disseminated ores, the chief minerals of which are chalcopyrite and/or pyrrhotite, and, occasionally cubanite. With the exception of pyrite, there are only trace amounts of the other sulfides, which generally occur as exsolution bodies. The Au and Ag contents are insignificant. The brittle garnet skarn zones became fractured during the folding. The fissures became filled with ore solutions during the folding. The effect of the solutions on the wall rock (second skarn phase) was slight. As the temperature decreased, solid solutions of pyrrhotite, chalcopyrite and bornite crystallized. The ore minerals exhibit many exsolution textures. After the crystallization of the solid solutions, hydrothermal ore solutions remained, subsequently crystallizing mainly as pyrite. In places, these solutions reacted with previously crystallized sulfides such as pyrrhotite resulting in the formation of pyrite.

With respect to their mode of occurrence and mineral assemblages, the copper-bearing skarns of the Virtasalmi area may be grouped, according to Schneiderhöhn's (1949, p. 58) classification, among contact-pneumatolytic copper deposits. Also in Cissarz's (1965, pp. 128—136) classification, they would fit into the category of contact-pneumatolytic copper deposits. Furthermore, according to Cissarz's scheme, the ore mineralization took place during the late-orogenic stage. The sequence of events appears to be genetically related to the origin of the quartz diorites.

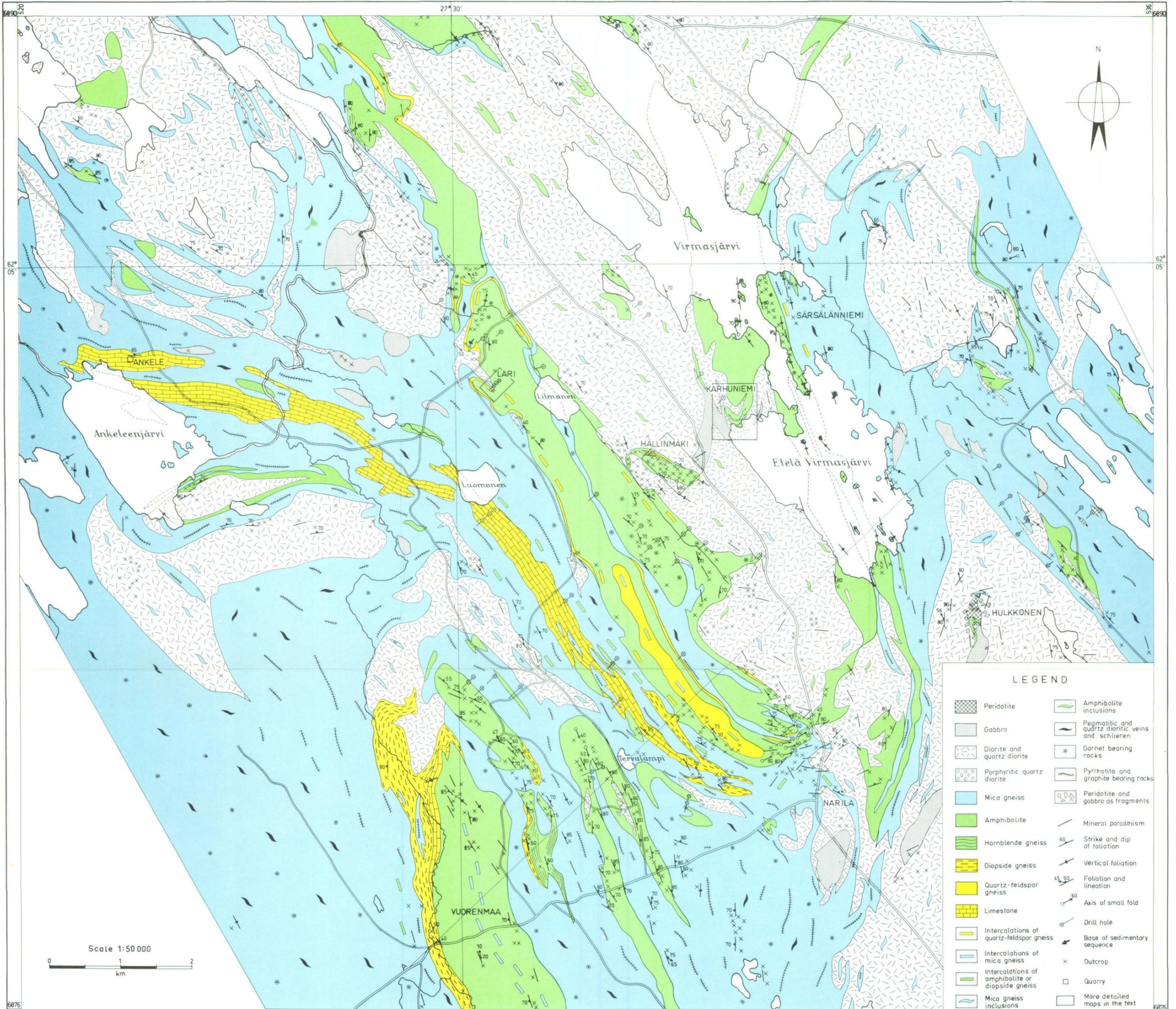
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**LEGEND**

	Peridotite		Amphibolite inclusions
	Gabbro		Pegmatitic and quartz dioritic veins and schieren
	Diorite and quartz diorite		Garnet bearing rocks
	Porphyritic quartz diorite		Pyrrhotite and graphite bearing rocks
	Mica gneiss		Peridotite and gabbro as fragments
	Amphibolite		Mineral parallelism
	Hornblende gneiss		Strike and dip of foliation
	Diopside gneiss		Vertical foliation
	Quartz-feldspar gneiss		Foliation and lineation
	Limestone		Axis of small fold
	Intercalations of quartz-feldspar gneiss		Drill hole
	Intercalations of mica gneiss		Base of sedimentary sequence
	Intercalations of amphibolite or diopside gneiss		Outcrop
	Mica gneiss inclusions		Quarry
			More detailed maps in the text

**GEOLOGICAL MAP OF THE VIRTASALMI AREA**

by  
LAURI HYVÄRINEN  
1969

