GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 204

SUOMEN GEOLOGISEN SEURAN JULKAISUJA
MEDDELANDEN FRÅN GEOLOGISKA SÄLLSKAPET I FINLAND
COMPTES RENDUS DE LA SOCIÉTÉ GÉOLOGIQUE DE FINLANDE

XXXIV

HELSINKI 1962

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GEOLOGICAL SURVEY OF FINLAND
OTANIEMI, FINLAND

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PROFESSOR ESA HYYPPÄ AT SIXTY

On June 14, 1962, Prof. Esa Hyyppä, head of the Department of Superficial Deposits of the Geological Survey of Finland, reached the age of sixty.

Prof. Hyyppä was born in the commune of Lapua, South Pohjanmaa where he also attended school. After matriculating in 1922, he enrolled in the University of Helsinki the following year. During his student years, he worked as a faculty assistant in the Department of Mineralogy and Geology and did field research in Quaternary geology in the regions of Satakunta, the Karelian Isthmus and Uusimaa. In 1932 he passed his examinations for degrees of Ph. Cand. and Ph. D. His academic thesis was entitled »Die postglazialen Niveauverschiebungen auf der karelischen Landenge.» Since 1933 he has served as docent of Quaternary geology and since 1961 as docent of geology and paleontology in the University of Helsinki.

Having completed his thorough studies, Dr. Hyyppä started his official career in the service of the Geological Commission in 1933 as a junior assistant geologist, later achieving the rank of senior assistant geologist. In 1938 he was appointed to the office of State Geologist as a specialist in the geology of superficial deposits, and when the Department of Superficial Deposits was established as part of the Geological Survey, he took charge. At the beginning of the current year the ranking of the departments of the Geological Survey was raised, and Dr. Hyyppä's position as departmental head earned him the title of professor. As creator of the Department of Superficial Deposits and by virtue of his efforts to develop it, Prof. Hyyppä has made and is continuing to make valuable contributions to research in the field of superficial deposits in this country.

Prof. Hyyppä has made numerous trips abroad for purposes of study, scientific research and attendance at learned congresses. In the years 1928—29 he was a member of Prof. V. Auer's expedition to Tierra del Fuego as an assistant geologist, in 1933 he did postgraduate work in Stockholm, and in 1938—39 in Yale University as well as in 1959 in Clark University he did research work in the United States. He is a member of the Groundwater Commission of the Geodetic-Geophysical Union, and he is serving as a permanent member of the commission appointed by the International Association on Quaternary Research (INQUA) to draw up a map of the Quaternary geology of western Europe and to deal with problems of Holocene geology. In 1939 he was awarded an honorary doctorate by Colby College (Maine, U.S.A.).

In his special field Prof. Hyyppä has published numerous distinguished studies, which have won him international recognition. They deal with the stages of the Baltic Sea, the evolution of the forests and climate of Finland as well as glacio-geological phenomena, in connection with which he has, *inter alia*, elucidated the directions of transport of ore boulders. Noteworthy, furthermore, are his studies on the technological utilization of the superficial deposits of Finland, groundwater relations and the problem of gold in Lapland. At present Prof. Hyyppä is engaged in producing a synthesis of his investigations into the evolutionary history of the Baltic Sea.

Prof. Hyyppä has participated in the activities of the Geological Society of Finland to a significant extent. At meetings of the society he has frequently given talks, numerous papers of his have been published in the society's series, and he has held the posts of secretary, vice-president and president. At present he is a member of the Publications Board. The Geological Society of Finland extends hearty congratulations to a deserving member.

1

ON THE ROLE OF FAULTING, A REPLY [1]

BY

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ABSTRACT

The paper discusses the role of faulting in the geology and geologic maps of the Precambrian basement of Finland. It is a reply to M. Saksela (1961).

The many systems of topographic trenches in the Baltic shield have long since been regarded as surface expressions of (high-angle) fractures and faults. In Finland, however, they were rather a problem of gecmorphology than one of crustal deformation. Apart from a few exceptions, it was believed that displacements along the fractures are too small to be considered, for instance, in the 1 : 100 000 reconnaissance maps published by the Geological Survey. The smoothly curved and deeply interfingered figures — resembling profiles of single and branched flames — which in these maps describe the areas of the different rocks, give the impression of a thoroughly plastic deformation.

Exceptionally careful geologic and geophysical investigations in the Orijärvi region (Tuominen, 1957) resulted, however, in a sharply different picture. Numerous fault systems split the area into a mosaic of polygonal fields or blocks. Granitic and other rocks formed along some fault zones add to the prominence of the mosaic structure.

The topographic trenches of the Orijärvi area generally coincide with faults. However, some marked faults are not visible in the topography. They are older than the regional amphibolite facies and may be called »relict»

[1] Received August 25, 1961.

faults (Tuominen, 1961). In other cases the breccias and mylonites following single faults range from rocks of the amphibolite facies to recent gouge. These faults are old but have been repeatedly active in later times. In addition, innumerable postmetamorphic faults split the complex into still smaller blocks.

In spite of the intense deformation *en bloc*, the large regional folds resulting from the earlier more plastic stages of deformation are also visible in the Orijärvi map (Tuominen, 1957, Pl.IV) and may be studied in certain better preserved blocks (Tuominen, 1961).

The discrepancy between the »flame-figure» maps and the »cubistic» Orijärvi map raises the question which of the two styles is in better accordance with nature. In the subsequent discussion it has been agreed rather unanimously that faulting affects markedly the structural picture of the old basement (Kranck, 1957; Marmo, 1959; Härme, 1960, 1961; Wegmann, 1961).

A voice of warning, however, has been raised against such ideas. Saksela (1961, p. 262) writes:

»In diesem Zusammenhang sei auf die von Tuominen (1957) im Massstab 1 : 100 000 entworfene Karte über das Orijärvigebiet aufmerksam gemacht, die meiner Meinung nach ein ziemlich irreführendes Bild von dem Einfluss der Bruchlinien auf das Kartenbild gibt (Abb. 3). Keinesfalls kann die Spaltentektonik, auch wenn der Massstab viel grösser wäre, zu solch einem Kartenbild führen, wo die von allen Gesteinen gebildeten Teilgebiete ausnahmslos als geradlinig begrenzte kantige Figuren vorkommen. Es wäre zu wünschen, dass ein derartiger »kubistischer» Stil in den kommenden Übersichtskartierungen nicht ohne weiteres angenommen würde, denn m. E. kann sie wohl hauptsächlich im künstlerischen Sinne von Bedeutung sein.»

Using part of the Suomussalmi sheet of the General Geological Map of Finland, 1 : 400 000 (Matisto, 1958), Saksela (*op. cit.*, Abb. 2.) tries to show that the fractures of the basement have not had any deforming effect on the general style of the small scale geologic maps. His Abb. 2. corresponds to an area of some 2 800 to 2 900 km^2. The total number of recorded outcrops in this area is somewhat more than 400, which means one outcrop per every 7 km^2, on the average (Matisto, *op.cit.*, Appendix). No geophysical maps have been available. The area consists mainly of pre-Karelian »granitegneiss». Other rocks are found in small scattered spots only. On the map they form elegantly curved single or branched flames, with wedges extended kilometers beyond the outcrops — or the outcrop — observed.

It is evident, therefore, that these fashionably flame-shaped figures which define the general style in Saksela's Abb. 2. are highly hypothetic and thus prove nothing in the question concerned.

For comparison, it might be reminded that the Orijärvi map (Tuominen, 1957, Pl. IV), representing an area of only 580 km^2, summarizes a field material of records obtained from tens of thousands of outcrops and hundreds of drill holes, combined with about 500 000 magnetic and 200 000 electromagnetic observations. The many black-and-white outcrop maps, geophysical maps, structural diagrams and other data presented in the paper (Tuominen, *op.cit.*) unquestionably show the faulted nature of the complex.

Similar trends in the small-scale aeromagnetic, gravity and geologic maps of Finland suggest that the major shear zones of the Orijärvi area belong to certain crisscrossing patterns of deformation found all over the country. These patterns, which seem to display all degrees of deformation from plastic to nonplastic, may be of major tectonic significance and deserve serious attention.

The plastic flame-shaped figures — so popular in the Finnish geologic maps — seem to descend from the peel thrust nappes and ophiolitic »fishes» seen in Wegmann's well known stereograms illustrating the Karelidic range. Saksela (1930, 1934) was the first who, in this meaning, applied them to geologic quadrangle maps. In most of the later maps, however, they have just been an easy way to satisfy the doctrine of plastic deformation and the demands of fashion. Little evidence has ever been presented to support such uniform dominance of flame pattern in the basement areas. The maps of Saksela *(op. cit.)*, for instance, are based on far too scattered observations to be of any consequence in this connection.

The type of plastic deformation as well as the intensity of deformation *en bloc* varies from area to area. The task of geologic mapping is to find out the true picture, not to follow any predestinated uniform »style». In this respect I agree with the warning of Saksela.

Professor Eskola has kindly read the manuscript and suggested a few changes which have been taken into account.

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2

ON METAMORPHISM AND STRUCTURES IN THE AHLAINEN AREA, SOUTHWESTERN FINLAND [1]

BY

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ABSTRACT

Geology of a highly metamorphic Archaean area is briefly described. Some aspects on the mineralogical and chemical evolution of veined gneisses, the origin and relative age of quartz dioritic rocks, and tectonics are presented.

CONTENTS

[1] Received August 25, 1961.

INTRODUCTION

In the summers of 1958 and 1959 the present author was commissioned by the Outokumpu Company to investigate an area in the archipelago of the Gulf of Bothnia and at the neighbouring coast. The area is situated in the parishes of Ahlainen and Merikarvia, about 30 km north of the town of Pori (Fig. 1). Its length in the east-westerly direction is about 17 km and the breadth varies between 6 km and 12 km.

The only field work ever done in the area was carried out when the old map sheet of Tampere of the general geological map of Finland was prepared (Sederholm, 1903, 1913). Because of the small scale of this map the picture of the rocks given by it is greatly generalized.

Complementary field and laboratory work is still necessary and the present paper is rather a preliminary report of the results.

MAIN GEOLOGICAL FEATURES OF THE AREA

The rocks of the area may be divided into five main groups as follows (see map, Fig. 2):

— veined gneisses
— hornblende-biotite gneisses (with pyroxene-bearing varieties)
— mica gneisses
— intrusive abyssal rocks (quartz diorite, norite-charnockite series, trondhjemite)
— dike rocks

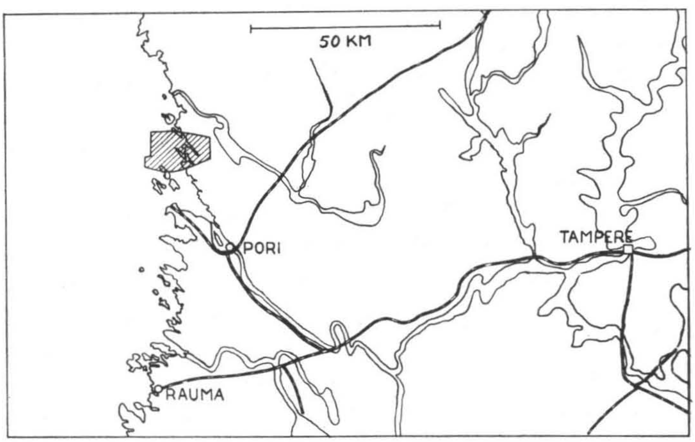

Fig. 1. Location of the Ahlainen area in southwestern Finland.

On the map in Fig. 2 the above grouping is chiefly followed and details are omitted. The intrusive massifs in the western and middle parts of the area actually contain numerous small portions of veined gneiss; on the other hand, within the zone of veined gneiss more homogeneous types of gneiss are not uncommon.

The main zone of the veined gneisses is in the inner archipelago, while the other types of gneiss lie east of them mainly within the coastal area. All the gneisses are characterized by strong deformation and metamorphism. No primary structures have been preserved, and changes in composition, too, are obvious or very probable. The mica gneisses may be fairly homogeneous in appearance at many places; the hornblende-biotite gneisses are more variable.

The western and southwestern parts of the area, close to the open sea, consist of intrusive rocks, the most part of which is biotite-hornblende quartz diorite. At a few places it grades into biotite trondhjemite. These rocks sometimes contain pyroxene and approach charnockite. Furthermore, there is some granodiorite and very occasionally microcline granite.

The longish intrusive massif within the gneisses in the middle of the area consists almost entirely of a pyroxene-free quartz diorite.

The small portion of intrusive rocks in the eastern end of the area is closest to the norite-charnockite series. The rocks range from peridotite to plagioclase-rich noritic gabbro.

Characteristic of most of the intrusive rocks is a variability in appearance and in mineral composition from place to place. Heterogeneity and gradations into various gneisses are common.

The dike rocks are both basic and acid. The basic dikes are of two main age groups, viz., older amphibolite dikes and younger non-metamorphic diabases. One of the diabase bodies, situated in the eastern part of the area, is fairly extensive and seen on the map. Among the acid dikes (excluding pegmatites) there are both potash-rich and soda-rich types.

Tectonically the area may be divided into several parts (see map, Fig. 3) in which the directions of schistosity may greatly deviate from each other. One of the »blocks» also differs quite conspicuously from the others by the direction of lineation, as well as by a gentle dip of the schistosity (in part).

In the following some select points will be treated in more detail.

VEINED GNEISSES

The vein material consists of plagioclase (about An_{30} with slight variations) and quartz. The older component of the rock is mostly represented by abundant biotite-rich schlieren only. In the schlieren, and in the veins at

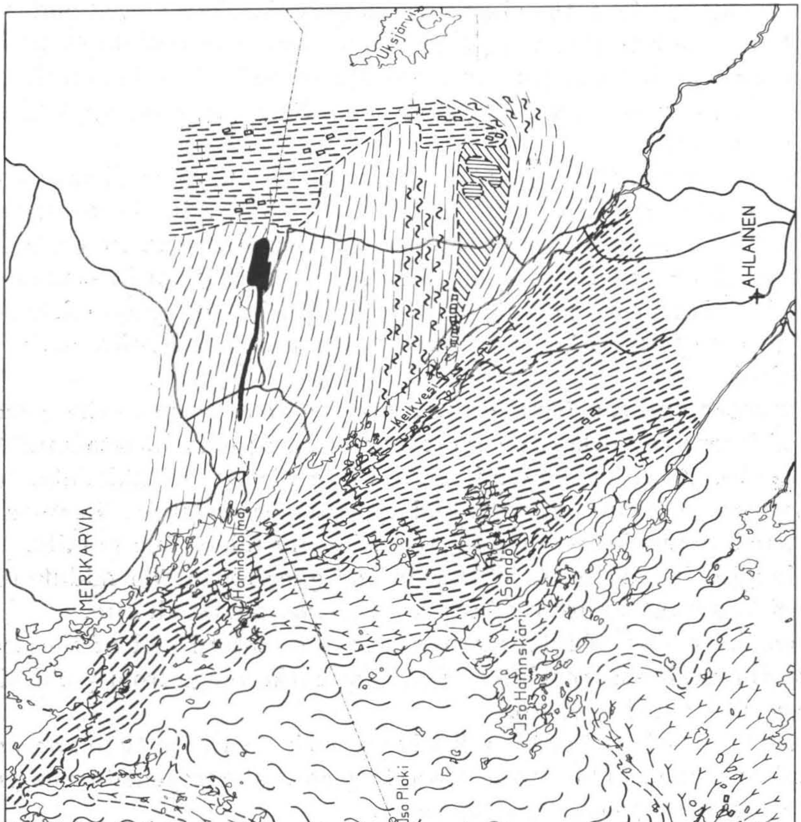

Fig. 2. Petrographic map of the Ahlainen area.

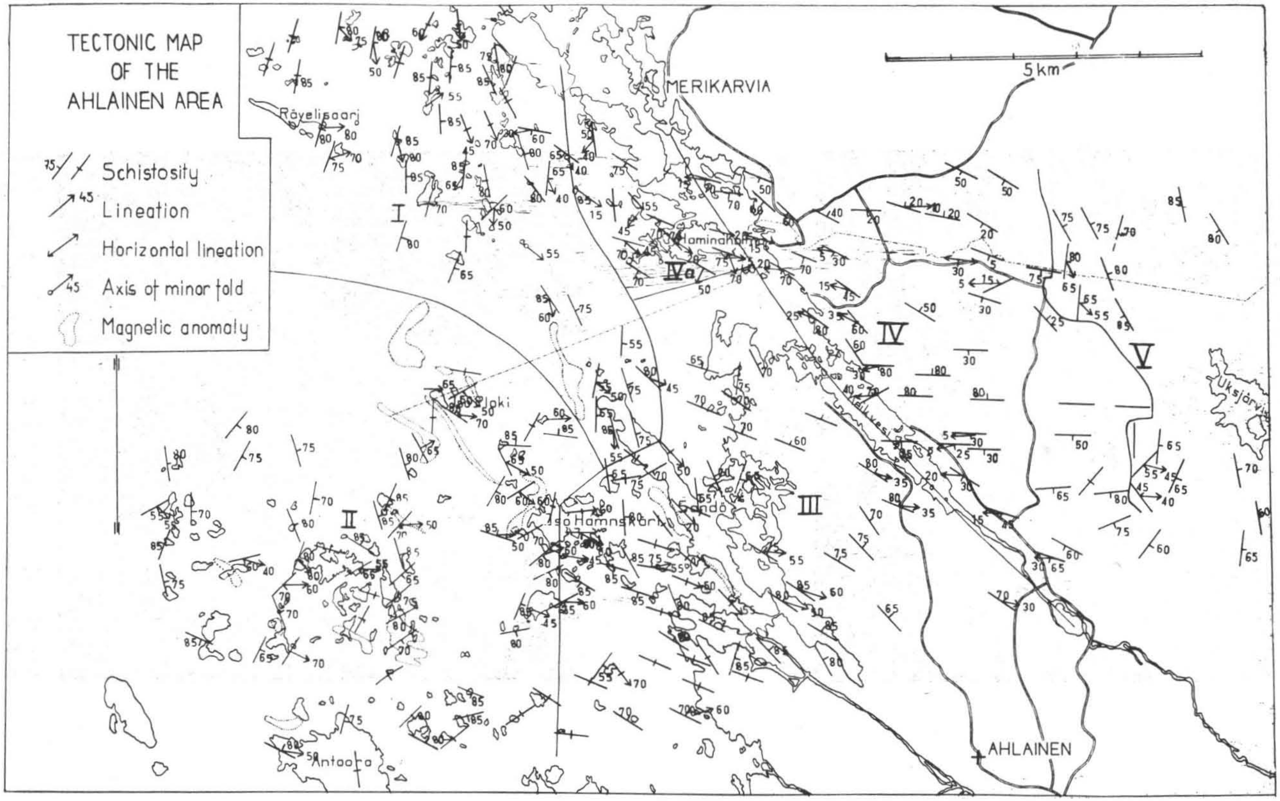

Fig. 3. Tectonic map of the Ahlainen area. For numbers I—V see text.

Fig. 4. Veined gneiss. Iso Ploki, Ahlainen. Photo: Vormisto.

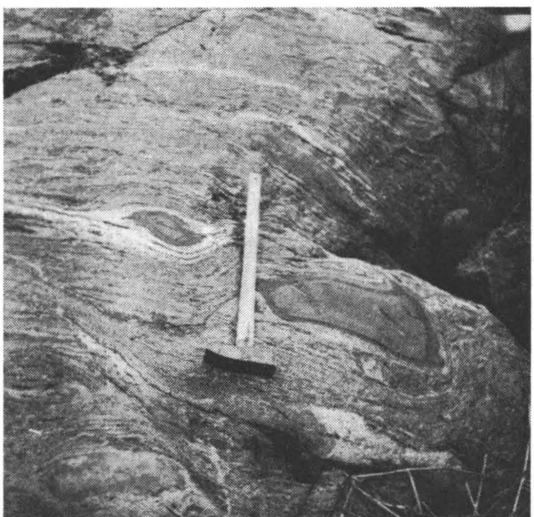

Fig. 5. Veined gneiss. 1 km south of Iso Hamnskäri, Ahlainen. Note the dark margins of the inclusions. Photo: Vormisto.

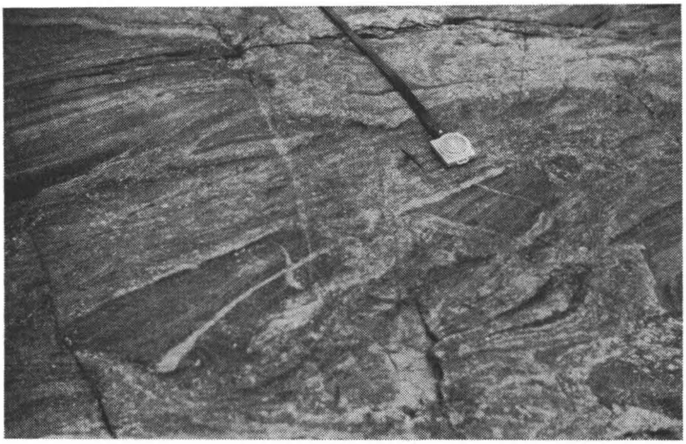

Fig. 6. Veined gneiss. 1.8 km southeast of Iso Hamnskäri, Ahlainen. Photo: Seitsaari.

borders of the schlieren, almandite is frequently found. In addition there are strips of biotitized amphibolite with more definite boundaries.

At places the amount of the vein material greatly increases and the veined gneiss grades into a light, gneissose rock with »ghost-like» remnants of the older component. Because potash feldspar is lacking, the composition of the rock approaches that of trondhjemite.

Quite typical of the veined gneiss is the occurrence of numerous sharply-bordered fragments originating from brecciated portions of older rocks. The fragments are predominantly basic (from amphibolite and gabbro to hornblenditic peridotite), the amphibolite fragments being the most usual. The size of the fragments varies greatly, but most of them are less than 40 cm in diameter. The form is also variable, sometimes angular, sometimes roundish or longish, etc. Typical appearance of veined gneiss is shown in Figs. 4, 5 and 6.

The most important and interesting of the fragment types is a cummingtonite-amphibolitic one. It is similar in composition to some basic, metamorphic dikes (p. 20), and it shows phenomena of metamorphic differentiation and metasomatism comparable to those in basic dike fragments from eastern Sweden described by Gavelin (1960, pp. 245—50). These two areas resemble each other in other respects too, although the primary characteristics of veined rocks have been preserved better in the Swedish instance.

The grain size of the fragments is about 0.15—0.20 mm. The mineral composition in the centre (= main part) of the fragments is: plagioclase (An_{58-73}), cummingtonite, hornblende, biotite, and some quartz. The relative

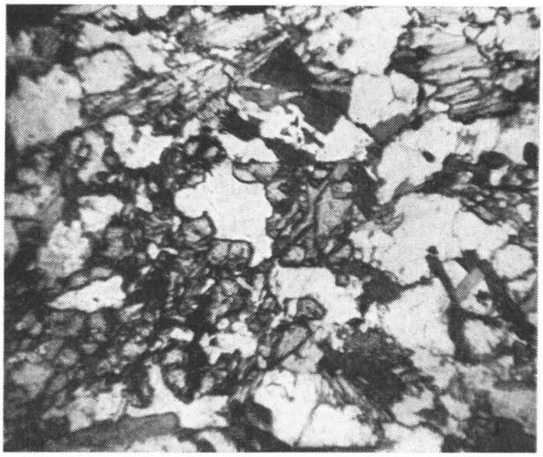

Fig. 7. Garnet-bearing fragment of basic dike rock. One Nic. Magn. 21×. 1.5 km west of Iso Hamnskäri, Ahlainen.

amounts of the two amphiboles vary considerably, but cummingtonite is always abundant, whereas hornblende is sometimes almost lacking. Some almandite is often present (Fig. 7) and very occasionally a little orthorhombic pyroxene. Among accessory constituents ore minerals, usually pyrrhotite and ilmenite, are the most abundant. In Table 1 the chemical and mineral composition of the centre of two fragments is presented. Although no almandite is reported from No. 1, which may represent the most typical composition, a slight excess of Al_2O_3 may indicate the presence of some garnet in the analyzed sample. It must be noted that point counter determinations of more or less heterogeneous rocks like these often give somewhat erroneous results.

Almost without exception the fragments have a dark, narrow rim against the veined gneiss (see Fig. 5). In the rim biotite is greatly enriched and cummingtonite is lacking, as is most often hornblende too. Plagioclase is more sodic (about An_{40-45}) than in the main part of the fragment, the amount of quartz has increased and almandite is fairly abundant. Plagioclase is occasionally observed to occur as porphyroblasts. Sometimes biotite-rich and biotite-poor zones may occur as alternate, indistinct bands through the fragment. In a few cases the whole fragment approaches the rim, but the centre is not as rich in biotite and the plagioclase is labradorite.

In the author's opinion, not only the outer shell but also the centre of the fragments has undergone an alteration. Both the absolute and relative contents of Al_2O_3 are distinctly higher than, say, in a basic dike rock, from which the fragments may originate quite commonly. Therefore it is supposed

Table 1. Composition of the middle part of the basic dike fragments in veined gneiss from the archipelago west of the island of Iso Hamnskäri, Ahlainen. Analyst, Ensio Lindqvist.

	1		2	
	%	Mol. prop.	%	Mol. prop.
SiO_2	51.92	8 645	50.86	8 468
TiO_2	0.80	100	1.10	138
Al_2O_3	19.56	1 919	21.51	2 110
Fe_2O_3	0.27	17	0.04	3
FeO	9.24	1 286	9.19	1 279
MnO	0.12	17	0.11	16
MgO	5.86	1 453	5.83	1 446
CaO	7.84	1 398	7.80	1 391
Na_2O	2.09	337	1.06	171
K_2O	0.80	85	1.11	118
P_2O_5	0.23	16	0.53	37
S	0.70	218	n. d.	—
H_2O+	1.36	755	1.37	760
H_2O-	0.00	—	0.02	—
	100.16	—	100.53	—
Less O for S	0.35			
	99.81			

Niggli numbers	1	2
si	132	130
ti	1.5	2.1
al	29.4	32.3
fm	42.7	42.0
c	21.4	21.3
alk	6.5	4.4
k	0.20	0.41
mg	0.52	0.53
c/fm	0.50	0.51
qz	+6	+12

Mineral composition	1	2
Plagioclase	54.6 %	60.8 %
Quartz	4.8	2.3
Orthorh. pyroxene	0.3	—
Hornblende	10.3	2.4
Cummingtonite	19.9	18.5
Biotite	7.9	7.9
Almandite	—	4.3
Ore minerals	1.7	2.4
Accessories	0.5	1.4
	100.0	100.0

that the Al_2O_3 content has increased while the CaO content has decreased. The alkali ratio has probably slightly changed in favour of potassium. In the rim an influence of introduced material is obvious. Although the present veined gneiss does not contain potash feldspar, it seems to have contained K_2O enough to biotitize the amphiboles. As a hypothesis to explain the fair excess of Al_2O_3, indicated especially by almandite, it is suggested that plagioclase, while becoming more albitic by the introduced Na_2O, is also enriched in potash feldspar in solid solution. In fact potash feldspar then replaces anorthite, as it does in the formation of microcline granite by granitization. Excess of Al_2O_3 is caused by removing CaO and retaining Al_2O_3, which is bound by (Fe, Mg) O to form garnet, as suggested by Härme (1958, 1959). Biotitization of any amphibole and pyroxene occurring in common igneous rocks liberates more or less (Fe, Mg)O and requires Al_2O_3, although there may be varying cases due to the highly varying composition of common hornblende and biotite. The process in the middle part of the fragments may be pictured as follows: A considerable part of hornblende, the most probable dark mineral in a metamorphic basic dike rock (see p. 23), has been replaced by cummingtonite. Thereby CaO has been released and removed from the whole fragment. Some K_2O from the veined gneiss and some Al_2O_3 liberated in the rim have been introduced. Al_2O_3 may be also obtained, when hornblende is replaced by cummingtonite. Thus Al_2O_3 has been disposable to form biotite and almandite from amphibole material (and K_2O).

Härme and Laitala (1955) have described alteration of a gabbro fragment in migmatitic microcline granite. In this example also a zoned structure has formed and the compositions of the different parts of the fragment are principally similar to these described in the present paper. This is reasonable if the effective constituent introduced from the outside is K_2O in both cases, in spite of the different alkali proportions in the surroundings. The above hypothesis may be proved only by determining the K_2O content of plagioclase in numerous series of altered rocks. The area now under consideration is not appropriate to such an investigation, because indisputable primary compositions do not exist.

In the veined gneiss there are also some small portions of breccia in which the fragments lie side by side and consist of basic or ultrabasic, usually coarse-grained rocks. These breccias are of interest from ore-geological point of view and the author will discuss the subject in a later paper.

In addition to the numerous basic fragments, mica gneiss is encountered as inclusions in the veined gneiss, though not abundantly. The inclusions consist of plagioclase (An_{30-40}), quartz, biotite, and some almandite. Keeping in mind the alteration phenomena of the basic fragments, the mica gneiss may also be of more basic origin. As a whole the origin of the old material of the veined gneisses remains unknown. There is often a little graphite in the biotite-rich schlieren, but it is also met with in the fragments of basic dikes as well as in ultrabasic inclusions, and thus cannot give any clear information of the origin. If there has been, say, acid schists before the formation of veined gneiss, their bulk may well form a part of the present vein material.

HORNBLENDE-BIOTITE GNEISS

The area predominantly characterized by hornblende-biotite gneiss lies east of the zone of veined gneiss.

The usual type of rock chiefly consists of plagioclase (An_{30-37}), quartz, and hornblende and biotite in various proportions. Occasionally cummingtonite as well as aggregates of serpentine and chlorite are observed. The aggregates are probably alteration products of pyroxene.

Sometimes the rock contains no amphibole thus grading into mica gneiss. In the southern part of this gneiss zone there are pyroxene-bearing varieties here and there. Both monoclinic and orthorhombic pyroxene is found, the former being more common. The anorthite content of the plagioclase may increase up to 45 %.

The gneiss in the easternmost part of the area much resembles the hornblende-biotite gneisses described in the foregoing. Pyroxene seems to be a little more common here and plagioclase may be a little more calcic. On the other hand, the rock has often been deformed more strongly and then contains secondary microcline and much biotite, while hornblende may be lacking. In the same zone there are portions of mica gneiss, in which pseudomorphs after pyroxene (see p. 16) are sometimes observed.

The appearance and general character of all these gneisses resemble each other to a high degree. They invariably contain plagioclase phenocrysts (usually some millimetres in diameter) and in addition small accumulations of larger plagioclase grains, often with quartz. The phenocrysts are sometimes observed to be more acid (calcic oligoclase) than the rest of plagioclase. Strong deformation is common and indicated by bent or broken lamellae of plagioclase crystals, mortar texture, etc. Mainly in the western part of these gneisses there are portions of veined gneiss. Strips of a darker gneiss, parallel to the schistosity and often with indistinct boundaries, are quite common and much resemble xenoliths in gneissose intrusive rocks. The boundary against the longish quartz diorite zone immediately to the west is far from sharp. The whole mass seems to have been in a fairly mobile state.

MICA GNEISS

The area immediately east of the bay of Keikvesi and farther to the north is mainly occupied by mica gneiss.

The main minerals of the mica gneiss are plagioclase (An_{16-33}), biotite, and quartz, in addition to which considerable amounts of almandite are often present. In a zone through the mica gneiss area (see map) there occurs secondary microcline fairly abundantly. The zone is parallel to the schistosity and some narrow zones are also met with elsewhere. Microcline often occurs as porphyroblasts up to 2 cm in diameter. Aggregates of certain alteration products are not uncommon. Several kinds of aggregates are observed, as follows: serpentine; serpentine and chlorite; serpentine, chlorite, and carbonate; and biotite, chlorite, and carbonate. The serpentine-bearing aggregates probably originate from pyroxene. When serpentine does not exist, hornblende is a plausible primary mineral and is even seen in a few cases.

In appearance and texture the mica gneiss resembles the hornblende-biotite gneiss (p. 15) in all essentials and gives the impression of a thoroughly altered rock. During the alteration process the bulk composition may also have been changed considerably. For example, the presence of garnet hardly proves any primary excess of alumina (see p. 14, and Härme, 1958, 1959). In a broad basic dike, highly metamorphic and hardly recognizable, the rock is now an almandite-bearing, fine-grained mica gneiss containing some hypersthene. Plagioclase and quartz commonly occur as numerous narrow streaks, between which the rock is markedly finer-grained and rich in biotite. A veined gneiss may sometimes have evolved in this way. The origin of the present mica gneiss remains unknown.

PERIDOTITE AND NORITIC GABBRO

The small portion of noritic gabbro with associated peridotite is situated east of the bay of Keikvesi within the zone of mica gneiss. Contact against the host rock is nowhere exposed. The peridotite occurrence in the eastern part of the gabbro area has been split up by gabbro and forms two small separate stocks. The gabbro between the stocks is schistose, the strike being approximately parallel to the contacts.

The constituents of the peridotite are hornblende, augite, olivine, and orthorhombic pyroxene, with a little plagioclase and biotite. Poikilitic hornblende crystals, about one cm in diameter, enclose the other main minerals.

The noritic gabbro consists of plagioclase (An_{45-48}), augite, orthorhombic pyroxene, some biotite, and a little quartz and hornblende. A microphotograph of the rock is presented in Fig. 8 and the chemical composition in Table 2. The rock differs from a typical norite mainly in the relative amounts of the pyroxenes the amount of augite being usually larger. The Al_2O_3 content is comparatively low.

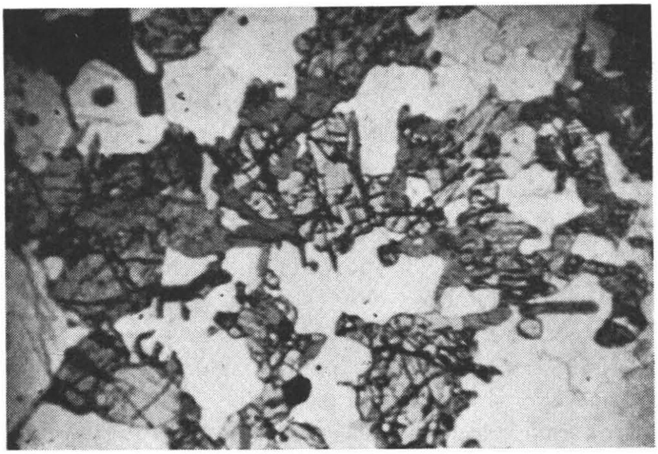

Fig. 8. Noritic gabbro. One Nic. Magn. 13×. 1.7 km east of the bay of Keikvesi, Ahlainen.

Table 2. Chemical composition of noritic gabbro, 2 km east of the bay of Keikvesi, Ahlainen. Analyst, Ensio Lindqvist.

	%	Mol. prop.	Niggli numbers	
SiO_2	51.94	8 614	si	124
TiO_2	0.60	75	ti	1.1
Al_2O_3	16.94	1 657	al	23.9
Fe_2O_3	1.50	94	fm	49.0
FeO	9.71	1 351	c	20.5
MnO	0.16	23	alk	6.6
MgO	7.40	1 835	k	0.12
CaO	7.98	1 423	mg	0.54
Na_2O	2.52	406	c/fm	0.42
K_2O	0.51	54	qz	—2
P_2O_5	0.17	12		
H_2O+	0.49	272		
H_2O-	0.06	—		
	99.98	—		—

The texture of the noritic gabbro is nearly hypidiomorphic. In the western end of the gabbro area the rock, however, is strongly deformed and gneissose. In spite of the deformation the mineral composition shows no essential changes. Hornblende and biotite have increased, but most pyroxene is still present. At the same time a small portion of peridotite has been brecciated altogether.

GNEISSOSE QUARTZ DIORITE AND ASSOCIATED ROCKS

Together with the veined gneiss, gneissose quartz diorite is the most important rock in the archipelago. Heterogeneity of the rock and its transitions into various gneisses were emphasized in the general description of rocks (p. 7). The heterogeneous appearance is caused by several factors. Dark xenoliths are often abundant and may be largely incorporated in the quartz diorite. The grain size and the relative amounts of minerals vary considerably. Some varieties are porphyritic with plagioclase phenocrysts. In the finer-grained type the schistosity is well-developed and straight, whereas in the coarse-grained rock it is less apparent and plastically sinuous. Last, there are transitional forms with diffuse, narrow veins between the quartz diorite and the veined gneiss (Figs. 9 and 10).

The main minerals are plagioclase (usually An_{31-35}), biotite, hornblende, and quartz. On an average, biotite may be more abundant than hornblende. Plagioclase is occasionally more calcic, up to An_{48}. Cummingtonite is observed in a few cases. There are also hornblende-free types, the character of which varies from biotite quartz diorite to biotite trondhjemite according to the amount of biotite and the anorthite content of plagioclase f(rom more than An_{40} to less than An_{30}). At places a decrease of the total amount of the two dark minerals also leads to a trondhjemitic composition.

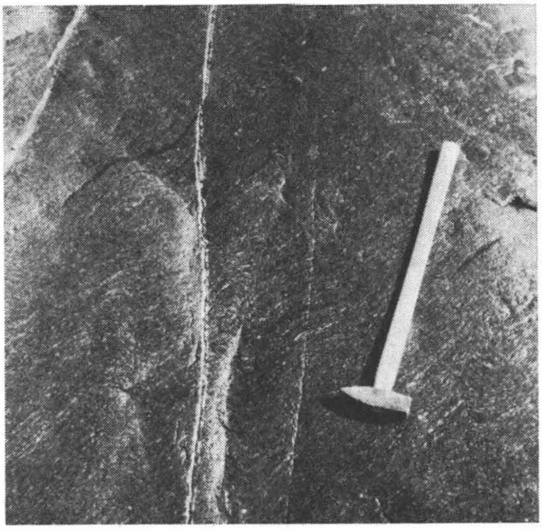

Fig. 9. Transitional form between quartz diorite and veined gneiss. 2.6 km north of Iso Ploki, Merikarvia. Photo: Vormisto.

Fig. 10. The same rock as in Fig. 9. Photo: Seitsaari.

At a few places, mainly near the eastern margin of the large quartz diorite area in the west, there are microcline-bearing varieties. Microcline commonly occurs as porphyroblasts and seems to have been introduced secondarily. Some granodioritic rock has evolved in this way, as well as a little porphyritic microcline granite.

At the southwestern margin of the area pyroxene-bearing charnockitic varieties are met with. They are granoblastic in texture like other intrusive rocks. The chemical and mineral composition of the typical example is presented in Table 3. The rock is similar to quartz dioritic charnockite in the classification of Simonen (1960, pp. 38, 45). The figure of *mg* is remarkably high, even higher than in the noritic gabbro (Table 2), although it may be higher than usual in the last-mentioned rock too. The total amount of pyroxenes is rather small. Around anhedral grains of microcline there is abundant myrmekite. In part the rock is a little richer in dark minerals and microcline is quite an accessory constituent.

Some monoclinic pyroxene is encountered in quartz dioritic and granodioritic rocks close to the open sea in the northwest.

Table 3. Chemical and mineral composition of quartz dioritic charnockite, Antoora, Ahlainen. Analyst, Ensio Lindqvist.

	%	Mol. prop.	Niggli numbers		Mineral composition	
SiO_2	66.06	10 955	si	245	Plagioclase (An_{35-40})	61.8
TiO_2	0.31	39	ti	0.9	Microcline	4.8
Al_2O_3	17.31	1 694	al	37.9	Quartz	17.8
Fe_2O_3	0.01	1	fm	26.3	Hypersthene	1.9
FeO	3.45	480	c	19.2	Diopside	2.7
MnO	0.06	8	alk	16.6	Hornblende	0.3
MgO	2.75	682	k	0.22	Biotite	10.5
CaO	4.82	859	mg	0.58	Accessories	0.2
Na_2O	3.61	582	c/fm	0.73		
K_2O	1.52	161	qz	79		
P_2O_5	0.10	7				
H_2O+	0.54	300				
H_2O-	0.04	—				
	100.58	—		—		100.0

DIKE ROCKS

Dikes of both basic and acid composition are mainly met with in the western part of the area, where excellent exposures in small rocky islands make it possible to observe all the details. Both kinds of dikes are metamorphic, with the exception of some diabases. The largest occurrence of diabase lies in the eastern part of the area. Only the metamorphic, amphibolitic, basic dikes will be described in the present paper, according to its scope.

The amphibolite dikes may be divided into three groups, the intrusion of which has taken place at different times. Most of the dikes are 0.1—1 m in breadth, but certain ones were encountered the breadth of which exceeds 10 m.

The dikes of the oldest group have been fractured and most often highly brecciated by veined gneiss. Connection between the fragments, however, has not always been lost to such a degree that the origin of the fragments would have become uncertain. Sequences of dike fragments in the gneiss are shown in Figs. 11—14. Interestingly enough, in quartz diorite there are also recognizable sequences of dikes, the fragments being often largely assimilated.

The fractured but yet more or less coherent dikes are qualitatively similar to the separate fragments described on p. 11. The ratio hornblende/cummingtonite is, however, distinctly in favour of hornblende. The two amphiboles are often unevenly distributed and cummingtonite may be predominant in some streaks and spots. Varying amounts of biotite are invariably present. Almandite is never found in the main part of the dike

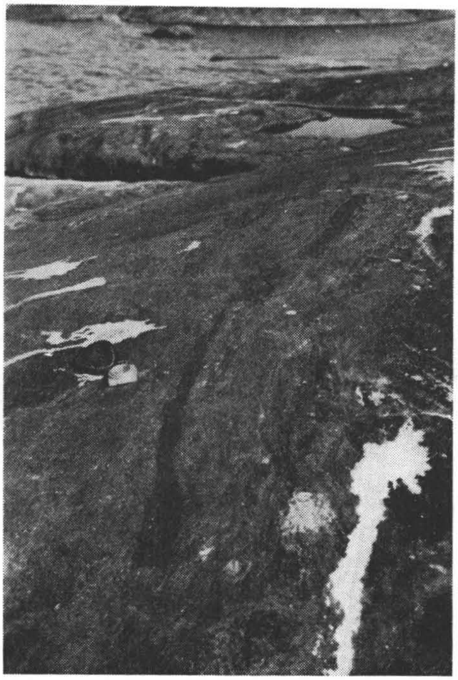

Fig. 11. Fractured amphibolite dike in veined gneiss. Iso Ploki, Ahlainen, Photo: Seitsaari.

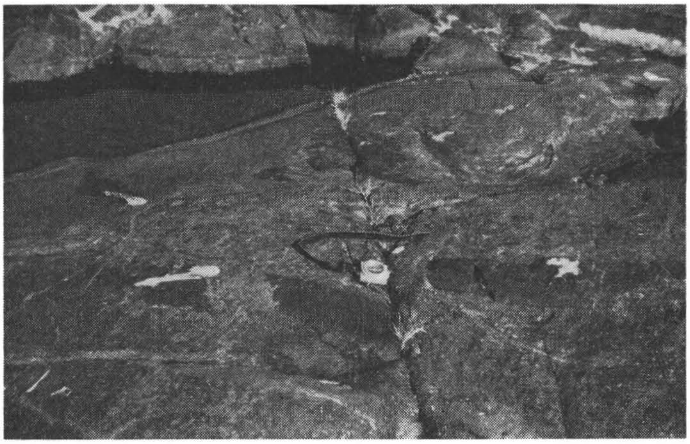

Fig. 12. Fractured amphibolite dike in veined gneiss. Iso Ploki, Ahlainen. Photo: Seitsaari.

Fig. 13. Fractured amphibolite dike in homogenized veined gneiss. 0.7 km northwest of Iso Hamnskäri, Ahlainen. Photo: Järvinen.

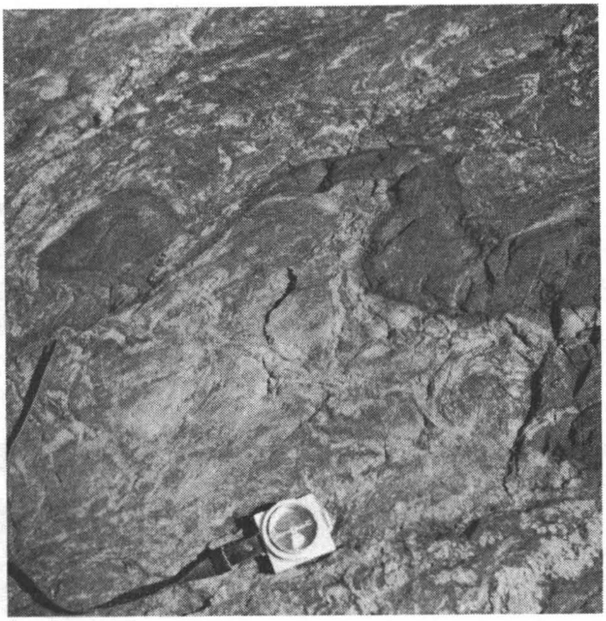

Fig. 14. Fractured and folded amphibolite dike in veined gneiss. 1.2 km southeast of Iso Hamnskäri, Ahlainen. Photo: Järvinen.

Table 4. Mineral composition of amphibolite dikes.

	1		2		3	
Plagioclase	An_{70}	22.5 %	An_{43-50}	39.1 %	An_{52}	19.2 %
Quartz		—		2.2		4.8
Hornblende		44.4		44.6		65.8
Cummingtonite		14.5		8.3		4.2
Biotite		16.4		4.3		0.2
Ore minerals		2.1		1.3		5.7
Accessories		0.1		0.2		0.1
		100.0		100.0		100.0

1. Fractured dike of the oldest group. About 3 km NE of the island of Iso Ploki, Merikarvia.
2. Dike of the middle group. Western shore of the cape of Sandö, Ahlainen.
3. Fine-grained dike of the youngest group. 0.6 km NW of the island of Iso Hamnskäri, Ahlainen.

rock, but may be seen with more abundant biotite in a marginal rim against the veined gneiss. Mineral composition of a fairly melanocratic type is presented in Table 4, No. 1. When the amount of plagioclase is greater, the anorthite contant is usually lower, down to 45 %.

Dikes of the second group are fairly coherent when occurring in the veined gneiss and they cut the older structures of the latter. But again in this case the last mobile phase of the quartz dioritic rocks seems to have been younger than the dikes, in consequence of which these are frequently more discontinuous in quartz diorite. In the veined gneiss a new direction of schistosity approximately parallel to the dikes often appears (see further p. 26), and the movements have been effective enough to change some parts of the dikes into veined gneiss — an additional evidence of the presence of a mobile phase under this stage.

The youngest dikes are wholly coherent and much finer in grain than the dikes of the other groups, even when being several metres thick. Still after this stage movements have taken place along the dikes, which are foliated parallel to the contacts and at places penetrated by thin veins of pegmatite and aplite in the same direction. In the immediate vicinity of the dikes the schistosity may have turned approximately parallel to the dike.

Field photograph of a dike of this group and microphotograph of the same rock are presented in Figs. 15 and 16, respectively.

Examples of mineral composition of the two last groups of dikes are presented in Table 4, Nos. 2 and 3. Minor amounts of cummingtonite are always present, but biotite may be lacking in the youngest dikes.

When the dikes of the second group follow any definite direction, as they do mainly in the northwestern part of the area where most of them

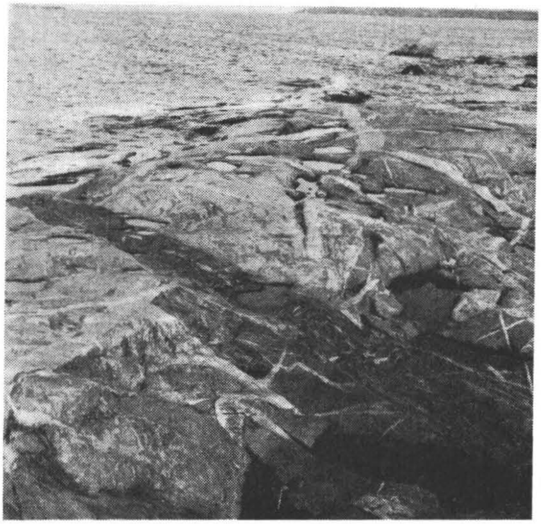

Fig. 15. Coherent amphibolite dike. 0.6 km northwest of Iso Hamnskäri, Ahlainen. Photo: Vormisto.

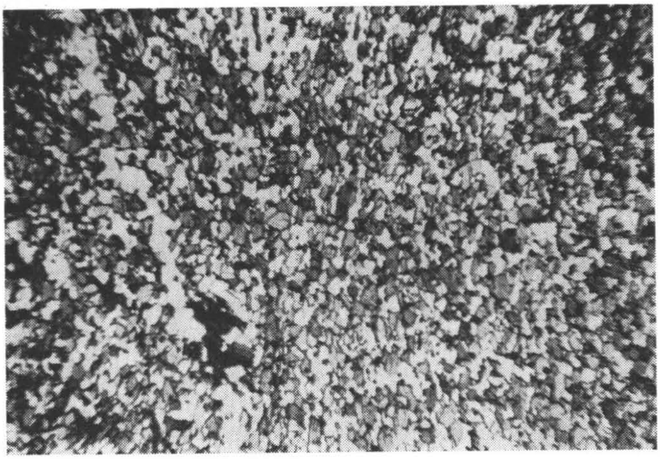

Fig. 16. Microphotograph of the dike in Fig. 15. One Nic. Magn. 50×.

have a north-southerly trend, the youngest dikes are usually observed to follow the same direction. Acid, trondhjemitic or fine-grained granitic, dikes seem to be younger than the young amphibolite dikes, but clearly belong to the same stage of dike formation with the amphibolite dikes mentioned.

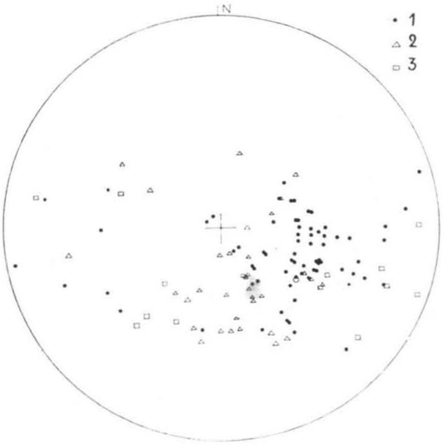

Fig. 17. Lineation in areas I, II, III, and IV a in stereographic projection. Lower hemisphere. (1) Areas II and III. (2) Area I. (3) Area IV a.

TECTONICS

Although the area under consideration is fairly small, it may be divided into several parts with marked differences in the position of schistosity planes and/or in the direction of lineation.

The lineation appears as streaks of minerals and tiny wrinkles, and minor folds in veined gneiss have principally the same direction. In a part of the area (IV on the map in Fig. 3) folding on a larger scale is occasionally recognizable and the gently-plunging fold axis is parallel to the local direction of lineation.

The different parts of the area are numbered from I to V on the map (Fig. 3) and may be briefly described as follows:

(I) Schistosity is fairly uniform and approximately in the north-southerly direction. Most lineation has a medium steep plunge to the SSE but at random directions also occur (Fig. 17). The boundaries against the other parts are not sharp. At many places in the vicinity concordant schistosity is seen.

(II) Schistosity is more variable than anywhere in the whole area. An ESE direction of lineation with a plunge of 50°—55° is predominant (Fig. 17).

(III) Schistosity is almost uniformly in the NW—SE direction and dips steeply to the northeast. Some variation, however, is seen in the northwestern part of this zone. Lineation is nearly parallel to the lineation in zone II. The boundary between (II) and (III) is not sharp.

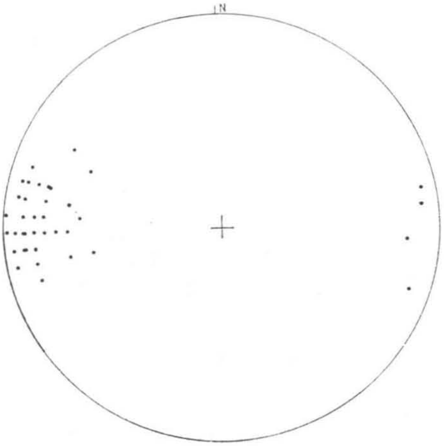

Fig. 18. Lineation in area IV in stereographic projection. Lower hemisphere.

(IV) This part borders with sharp straight-lined boundaries upon the others and rock boundaries are identical with tectonic ones (see Fig. 2). The block has obviously been disconnected by faults from the neighbouring blocks, but the slip direction of the faults could nowhere be detected.

In the southern part of the block the schistosity strikes about east-westerly, in the northern part usually in the direction N 60°—70° W. Proceeding northwards from the southern end the dip becomes more and more gentle to the south, but then suddenly almost vertical again. At this place there is a narrow but long bog in the terrain and there is obviously a fault here, although no difference is seen in the rock on either side. In the northern part of the block the schistosity dips gently again, with the exception of the coastal area, and here dips to the north are observed too.

The lineation within this block markedly deviates from all the others by plunging gently (usually 0—15°) to the west (Fig. 18).

(IV a) Scistosity is concordant with that in the adjoining part of block IV, but the directions of lineation are quite at random (Fig. 17). The rock is of the same type as in (III), but pyroxene-bearing gneisses are entirely lacking.

(V) Style of deformation and the rock very markedly differ from those in the adjoining area (IV). The schistosity strikes approximately north-southerly and the lineation plunges fairly steeply to the southeast, as it does west of block IV.

In the author's opinion, the present schistosity in the whole western part of the area investigated has been essentially caused by the direction of the amphibolite dikes at each place (see p. 23). Dikes are mainly met with in the areas I, II, III, and IV a. In (I) the coherent dikes have the most uniform trend and the strike of schistosity differs from the direction of dike at each place by 10°, on an average (45 observations). In the other areas mentioned the deviation is much larger, from 31° in (II) (43 observations) to 41° in (III) and (IV a) (36 observations), and the dikes are more at random. Renowed movements have utilized the planes of anisotropy when approximately parallel to each other, as in area I, but not so much in the other areas. When the dikes are at random, schistosity may have developed in many different directions. Development of a new direction of schistosity also depends upon the mobility of the rock to be deformed. The mobility has obviously been higher in the western part of the area investigated (I and II). But this stage of deformation has not totally obliterated the old features. The strike of schistosity is often transverse to the longish areas of magnetic anomaly (Fig. 3) which may be caused by sulphide-bearing minor intrusives. (Unfortunately, exposures are very scanty at these places.) They have been broken and brecciated, but connection between the parts has not been lost totally. It is worth noting that the general trend of parallel structures in this part of Finland is about NW—SE or the same which is indicated by many zones of magnetic anomaly in the Ahlainen area. Thus the transverse direction is a new one. In part the zones causing anomalies have been re-oriented in the new direction.

CONCLUDING REMARKS

Metamorphic facies. In the zone of veined gneiss the amphibolite facies is quite predominant, but on either side of it there is also evidence of the conditions of the granulite facies, further indicated by the charnockitic rocks. At places mineral assemblages of the two facies alternate. Because conditions in a highly deformed area like this may change suddenly, it is difficult to decide whether, say, »dryness» of rocks has caused assemblages of the granulite facies, as suggested especially by Yoder (1952, 1955).

The mica gneiss block, in particular, seems to have undergone a thorough alteration, the conditions prior to it being shown by wholly altered pyroxenes. The albite-epidote-amphibolite facies, to follow Turner's terminology, has not been completely realized, because the plagioclase is not albitic. Primarily, however, it has probably been much more calcic, so that considerable amounts of anorthite may have been leached away in any case. Combination almandite + microcline does not prove a low temperature, whereas the aggregates of serpentine, chlorite, and carbonate are typical

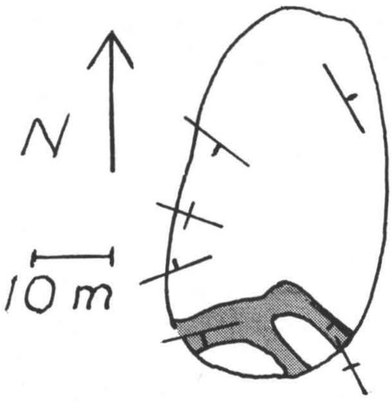

Fig. 19. Fragments of quartz diorite in veined gneiss. White: quartz diorite; hatched: veined gneiss. Sketch of a small island 1 km north of Iso Hamnskäri, Ahlainen.

of an alteration at low temperatures. Recrystallization has obviously not always reached equilibrium. The assemblage in the gneissose noritic gabbro within the mica gneiss area (p. 17) is surprising. It seems to have nothing in common with the surrounding gneiss, in spite of quite a concordant deformation.

Structural evolution. Dikes of different age show that there has been alternation of stages of higher and lower plasticity of rocks. During a late stage of deformation there has still occurred brecciation of the quartz diorite among the veined gneiss (Fig. 19). The phenomenon shows a different state of plasticity of the two rocks during this stage. The contacts of amphibolite dikes seem to have been utilized by movements to the last (Fig. 20).

It has been admitted for a long time that penetrative movements are able to cause mobilization. But a mobile phase, on the contrary, may greatly facilitate movements. Heterogeneity of rocks is obviously an important factor. Reactions between different parts of a rock may be promoted by flow caused by differential stress (Gavelin, 1960, p. 266).

A style of deformation of Precambrian rocks predominantly characterized by faults and brecciation was put forward by Tuominen (1957). Although the present author regards such a picture as exaggerated, movements of rigid blocks are indeed not uncommon (see also Kranck, 1957).

Origin and relative age of the quartz dioritic rocks. The quartz diorite and associated rocks are distinctly migmatitic and in this respect greatly

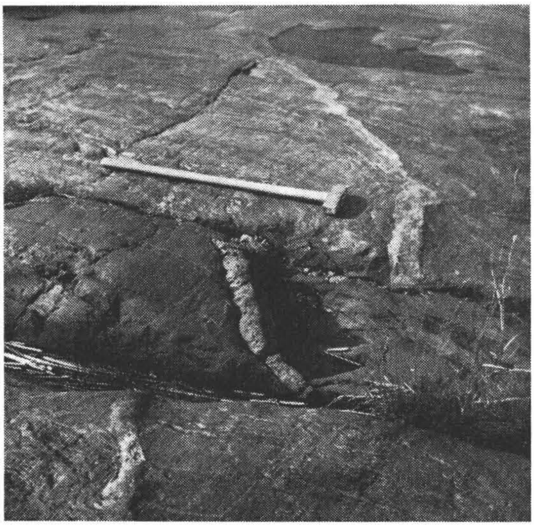

Fig. 20. Faults along contacts of amphibolite dike. 0.7 km northwest of Iso Hamnskäri, Ahlainen. Photo: Järvinen.

resemble the migmatitic microcline granite formed by granitization. It seems that metasomatism has played an important role in their genesis. For example, the fractured but yet recognizable older dikes in the quartz diorite do not indicate any forceful intrusion. A subject like this belongs, however, to the general »granite controversy» and opinions both *pro et contra* will always be presented. In Finland, Hietanen (1947, pp. 1072—3) has interpreted certain diorite trondhjemites as metasomatic, strong differential movements being an important factor in their evolution. The present author set forth related ideas in 1951 and 1955.

The quartz dioritic rocks are younger than most of the amphibolite dikes, in consequence of which they would belong to the late-kinematic stage of the Svecofennian orogeny. Sederholm (1932), Hietanen (1947), and Simonen (1960) consider quartz diorites, trondhjemites, and related rocks to belong to the first cycle of the orogeny, whereas intrusion of such rocks may occur in both cycles according to Saksela (1953). Also in Väyrynen's opinion (1954), quartz diorite and trondhjemite represent the first cycle; but charnockite does not. In the present instance there is obviously no difference in age between the quartz diorite and charnockite. As a whole, the process of deformation and metamorphism seems to be more complicated than is usually assumed.

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THE RAPAKIVI PROBLEM AND THE RULES OF IDIOMORPHISM IN MINERALS [1]

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ABSTRACT

This paper includes observations and data gathered from the literature on rapakivis, in addition to which entirely new material is elucidated and the significance of certain previously described but generally overlooked observations emphasized. Finally, the course of crystallization and the genesis of the structure of rapakivis are considered. Part II of the paper weighs the validity of the rules governing the idiomorphism and order of crystallization of minerals. This has proved essential to an understanding of the structure of rapakivis, while, on the other hand, the critical study of rapakivis has provided a splendid opportunity to test the rules governing the idiomorphism and order of crystallization of minerals.

CONTENTS

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PREFACE

I owe debt of gratitude to Prof. Vladi Marmo, director-general of Geological Survey of Finland, and to Prof. Ahti Simonen, chief of the Petrological Department, in particular, for their magnanimety in dealing with this study. May thanks to the Geological for publishing my paper. I am grateful to the University of Helsinki for the use of its facilities while serving on the faculty and to the State Board of Natural Sciences for financial aid.

PART I

THE RAPAKIVI PROBLEM

Rapakivis, according to SEDERHOLM (1891) and ESKOLA (1930), are among the most nearly typical hypidiomorphic igneous rocks in existence. They are of a medium or coarse grain, but fine-grained varieties of these rocks are also known; the former are rarely porphyric [1]). The magma from which the rapakivis have solidified was rich in K and — in proportion to the Mg — in Fe, and it represented a relatively dry granitic magma. This magma is of the so-called rapakivi type (NIGGLI 1923). The solidification of this magma started with the crystallization of the potash feldspar and the quartz, the later differentiations having been enriched from the plagioclase and the mafic minerals (SAVOLAHTI 1956, 1956 a; TUTTLE and BOWEN 1958; TURNER and VERHOOGEN 1960; SIMONEN 1960 a).

The rapakivis have attracted widespread attention because they include quite odd-looking types of rock. One of them is the so-called typical rapakivi, which has large, often reddish-brown, ovoid potash feldspar grains, usually surrounded by greenish gray plagioclase shells. This kind of fabric is noteworthy in that it appears to contradict the hypidiomorphic order of idiomorphism and crystallization of igneous rocks. That is why explaining the fabric of rapakivis has proved difficult and involved much hard work. The term »rapakivi» itself derives from the fact that many types of the rock have a tendency to disintegrate (r a p a u t u a in Finnish), although not nearly all the types share this tendency — nor is it even necessarily characteristic of any single type everywhere. The causes of the disintegration and the history of the research into the matter have been extensively dealt with by ESKOLA (1930).

The generally accepted view is that the rapakivi magma thrust up from a considerable depth to a relatively high level, even up to the surface. After that the magma crystallized under extremely peaceful, post-tectonic conditions. Such a mode of origin, coupled with the »rapakivitic» composition

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[1]) The term *porphyric* is here used as an arbitrary coinage, deriving directly from »porphyry,» to distinguish it from the less specific adjective »porphyritic,» which derives obviously from »porphyrite.»

of the magma, is a prerequisite to the development of a rapakivi texture. ESKOLA (1961) holds the view that a slight Al-deficiency in the chemical composition of rapakivi is further necessary to produce the plagioclase-mantled potash feldspar ovoids. The present author, however, has seen beautiful plagioclase-mantled potash feldspar ovoids also in rapakivi containing an excess of Al; and they have, moreover, been described in the literature. Again, the development of a rapakivi texture need not depend on the time factor, for, in principle, there is no reason why rapakivi magmas could not have burst forth and rocks warranting the label of rapakivi have formed at different times and in different places (SAVOLAHTI 1958).

1. THE SIZE OF RAPAKIVI MASSIFS AND THE GRAIN SIZE OF THE ROCKS OCCURRING IN THEM

Of the existing rapakivi massifs, the rapakivi area of Viipuri is unquestionably the largest (Fig. 1). This may be stated even though it need not — and, apparently, does not — represent a single continuous eruption. The grain size of the main rock types of this area, viborgite and pyterlite, is larger on the average than the grain size of the rapakivis in other rapakivi areas. This view was held by SEDERHOLM (1891) already in his early day, and later investigators have concurred with him. Next in grain size ranks the rapakivi of the Aland Islands. It is, however, hard to determine whether the rapakivi area of the Aland Islands exceeds that of Laitila in size, for

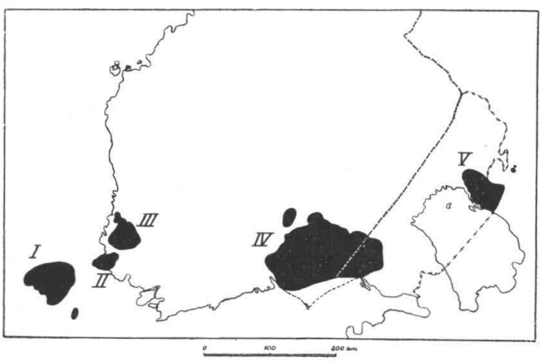

Fig. 1. The East Fennoscandian rapakivi areas (SAHAMA 1945). I — The Aland area. II — The Vehmaa area. III — The Laitila area. IV — The Viipuri area. V — The Salmi area. On the northern side of the Viipuri area lies the separate Ahvenisto area, on the eastern side of which, again, is the Suomenniemi area, representing a northerly bulge of the Viipuri area.

part of the former area is submerged and its limits are not wholly known. But, according to the gravitational investigations of PESONEN (1930), the Aland Islands have quite a large negative gravitational anomaly. HACKMAN (1934) is of the opinion that this is caused by the great thickness of the local rapakivi massif — and as a massif it would, accordingly, be larger than that of Laitila. The rapakivi of the Laitila area is coarser of grain than that of Vehmaa, but so is the rapakivi area of Laitila the more extensive. The rapakivi areas of Ahvenisto and Suomenniemi adjoin and are nearly equally large. The rock of the former is the coarser. The explanation for this is, according to general opinion, supported by the available evidence, that the rapakivi massif of Suomenniemi consists of a thin sheet, for its contacts with the country rock are gently sloping, as described by FROSTERUS (1902) and, most recently, PIPPING (1956).

The size of the rapakivi massifs, on the one hand, and the grain size of their rocks, on the other, seem to be interdependent in direct proportion. This kind of interdependence requires crystallization from the same kind of magma on the same level and under the same conditions. According to SZÀDECZKY-KARDOSS (1958), the texture of igneous rocks depends principally on the depth at which crystallization took place, on the shape and size of the igneous body, on the distance to the country rock, on the viscosity and composition of the magma, and, finally, on the possibility of gravitational differentiation. In certain cases, later block movements are likely to have disrupted this order. In addition, each rapakivi area has small, independent rapakivi sheets and dikes whose rocks do not, perhaps, seem to fit into this system, but, on closer scrutiny, certainly prove to do so.

The dependence of the rock's grain size on the size of the massif in which it crystallized is, at the same time, one proof of the igneous origin of the rock (*cf.* SEDERHOLM 1891, p. 19). This view can evidently be applied to other igneous rocks besides rapakivi.

2. OBSERVATIONS OF THE CONTACT ZONES OF THE RAPAKIVI MASSIFS

a) BRECCIAS AT THE CONTACTS OF THE RAPAKIVI MASSIFS

The contacts of rapakivi massifs against their country rock are sharp, and eruptive breccias are found here and there along their margins. In these breccias, fragments of country rock occur in arbitrary positions, which mode of occurrence is deemed proof of the magmatic origin of the brecciated rock. Fig. 2 shows such a beautiful eruptive breccia, caused by rapakivi, the dark angular fragments being amphibolite. Breccias along the margins of

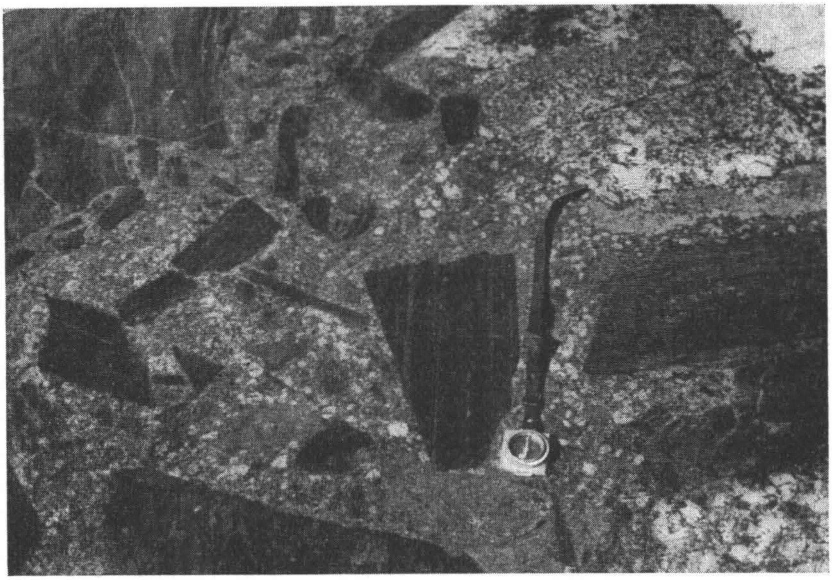

Fig. 2. Rapakivi brecciates amphibolite. Lyöttilä peninsula, Iitti (LONKA 1957). Photo: M. Lehijärvi.

rapakivi massifs are not, however, so common as might be expected. At the contacts of many so-called porphyric granites and certain other granites as well, intrusive breccias are far more frequently met with, being, in addition, much broader. As early as 1891, SEDERHOLM wrote about the scarcity of breccias associated with rapakivis, having, according to WAHL (1925), even exaggerated the matter. Later on, SEDERHOLM (1928) changed his mind somewhat. At Pyhäjärvi and Kiuruvesi, in Oulu province, the marginal zones of many granite masses contain very broad and beautiful breccias. The present author has nowhere run across breccias of the same magnitude in association with rapakivis. Considered thoughtfully, that is only as it should be, if rapakivis are actually postorogenic or, perhaps, as ESKOLA (1957) writes, anorogenic, for granite erupting under strong kinematic conditions must naturally have had greater force in brecciating its country rock than that erupting under peaceful conditions.

b) MARGINAL VARIETIES OF RAPAKIVI MASSIFS

Observations have been made in many places that rapakivi is finer-grained near the contact of its country rock than farther from the contact (*cf.* ESKOLA 1961, p. 120). Phenomena of this kind are described from the western boundary of the Aland rapakivi area (SEDERHOLM 1890). Near the contact, the porphyric rapakivi type is depicted as finer of grain, more porphyric and more micrographic in texture than farther from the contact, while in places it approaches the quartz porphyries in texture.

Similar phenomena have been reported from the borders of the Suomenniemi rapakivi area, on the northern side of the Viipuri rapakivi massif (FROSTERUS 1902 and WAHL 1925). In the summer of 1956, the present author had an excellent opportunity to acquaint himself with the geology of this area under the guidance of F. Pipping, who, then an undergraduate, was preparing an academic thesis on the rapakivi of Suomenniemi. Visible at the western border of this rapakivi area, on the western shore of Kotajärvi, is a sharp contact between the rapakivi and its country rock, consisting of garnet-bearing, fine- and medium-grained granite. The contact, which inclines approximately 15° N70°W, can be seen for a distance of about 250 meters on the face of the precipitous rock. Nearest the contact is a porphyric, fine-grained variety, which grades over little by little into normal coarse-grained rapakivi. To the northeast from the northwesternmost bay of Korpijärvi, in the Suomenniemi area, sharp contacts with a gentle slope may also be observed. There, too, nearest the contact occurs a porphyric variety of rapakivi, which gradually changes, first into a fine- and medium-grained, then into a coarse-grained rapakivi (Plate I, Fig. 1). According to PIPPING's (1956) description, the contact varieties are more micrographic in texture also in the Suomenniemi area than the other rapakivi types.

About one kilometer to the northwest from the Lammi ferry, a very interesting and important contact can be seen. In June of 1957, the present author had the opportunity of becoming acquainted first-hand with the geology of this locality under the guidance of Messrs A. Lonka and M. Lehijärvi, the former of whom has since earned his licentiate and the latter his doctorate in geology. The visit was occasioned by the fact that Mr. Lonka was just completing an academic thesis principally dealing with the so-called Jaala—Iitti hornblende-rapakivi intrusion located there. According to Lonka's researches, no great textural difference exists between so-called porphyric rapakivi and pyterlite, only their respective grain sizes differing somewhat; nor do they differ in age; and, as in the Jaala—Iitti area, they grade over, one into the other. SIMONEN (1960 a) supports this observation. On the present author's recommendation, it was decided to try to find and expose a contact between viborgite and either porphyric rapakivi or pyterlite. Success was rather easily achieved, for Mr. Lonka was able to suggest a likely spot, which proved to be the one just mentioned, northwest of the Lammi ferry. After we had removed the moss and turf covering the rock, we saw a porphyric rapakivi with textural features of pyterlite, which graded over for a distance of some five meters into viborgite. Proceeding from the

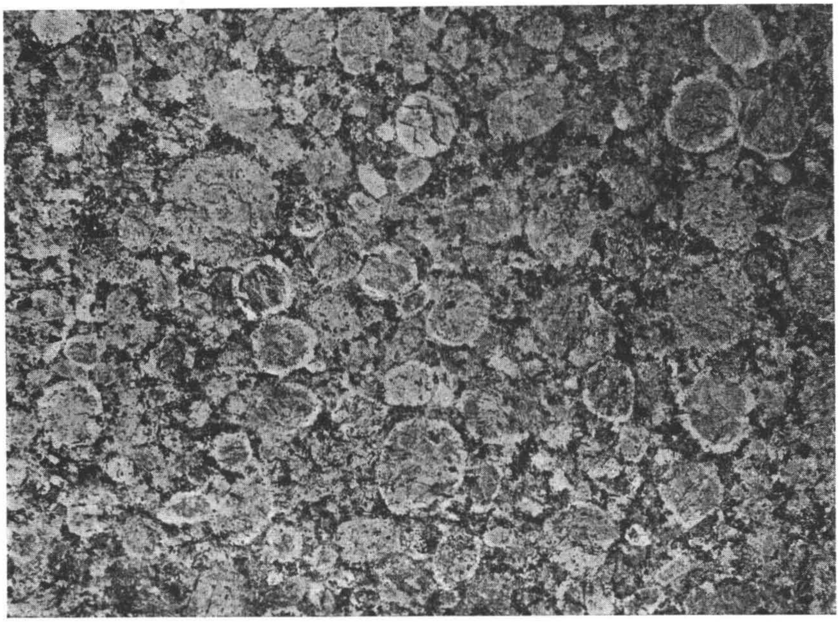

Fig. 3. Viborgite. C:a 2/5 nat. size, Kivijärvi, Lemi, Viipuri area (SEDERHOLM 1928).

viborgite, the plagioclase rings surrounding the potash feldspar grains gradually decreased in number, although some could be found also in the porphyric type, too. With the disappearance of the plagioclase rings, the viborgite also gradually underwent a transition into a rapakivi having a porphyric texture. The significance of the observation described is nowise diminished by the fact that rapakivis of the type in question also have other types of contacts both between themselves and with their country rock. This the perceptive reader will, perhaps, already have noted.

Corresponding to these phenomena, the Ahvenisto region (SAVOLAHTI 1956, 1956 a) has an abundance of rapakivi sheets and dikes in different positions and of different thicknesses, especially in the gabbro-anorthosite surrounding the rapakivi. The rock in them is finer of grain than the principal rapakivi type of the massif, with which type they are directly associated. The sheets referred to differ also among themselves in grain size, and this difference depends on the thickness of the dike or sheet. Belonging among these phenomena, in a way, perhaps, is Fig. 2 (Pl. III), from the Ahvenisto region, where a fault has brought forth a chilled contact, which might otherwise have remained unnoticed.

The marginal varieties of the Aland, Suomenniemi and Ahvenisto areas, described in the foregoing, have been interpreted — quite rightly — as

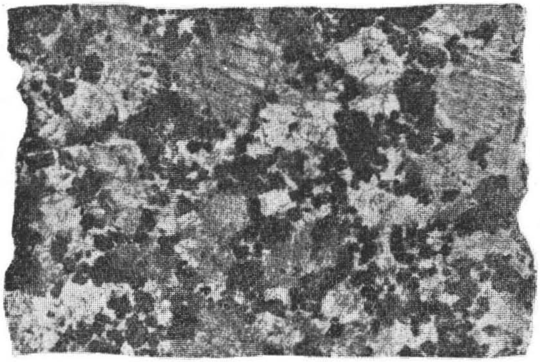

Fig. 4. Pyterlite (Surface of rock polished). C:a 2/3 nat. size. Pyterlahti, Viipuri area (Wahl 1925).

having developed by the rapakivi mass' having crystallized against the colder country rock; *i.e.*, the marginal part solidified faster than the main part. Thereby, a chilled contact resulted, and it provides further evidence of a magmatic origin. The porphyric rapakivi and pyterlite met with in the vicinity of the Lammi ferry, in the Jaala—Iitti area, lie between the bedrock and the viborgite. Accordingly, the porphyric type and the pyterlite constitute marginal varieties, which came about at the margin of the magmatic mass as it crystallized, while the viborgite, again, represents a rock type produced by the solidification of the central portion of the same great massif. Thus did two texturally quite different types of rock simultaneously evolve from the same magma. The viborgite is shown in Fig. 3 and the pyterlite in Fig. 4. The difference in texture between the two is considerable. The principal difference in respect to composition is that the pyterlite is richer in potassium than the viborgite. It is, therefore, only a consequence of potassium autometasomatosis taking place during the crystallization stage (Rein 1955, and Mehnert 1960). The figures are not, to be sure, from the point in question, nor do they provably represent rocks directly associated with each other, but derive from different places in the Viipuri rapakivi area. At any rate, however, they give a good idea of the textural differences between viborgite and pyterlite here and elsewhere. The afore-described contact gives us quite a valuable hint in interpreting the texture of viborgite.

c) THE GREEN MARGINAL VARIETY OF HORNBLENDE RAPAKIVI

Near the southeastern boundary of the Ahvenisto massif, in the village of Nurmaa, there occurs a green marginal variety of hornblende rapakivi,

Table 1. The mode of the hornblende rapakivi (1) and its green marginal variety (2) met with at Nurmaa, Mäntyharju.

Minerals	1	2
Potash feldspar	44.9	46.6
Quartz	28.2	22.6
Plagioclase	15.3	14.7
Olivine	—	2.3
Pyroxenes	—	
Hornblende	6.2	9.1
Biotite	4.3	3.5
Accessories	1.1	1.2

which has been interpreted as having evolved through autometasomatosis and autometamorphosis during the crystallization of the magma mass (SAVOLAHTI 1956, 1956 a). It, too, represents a kind of chilled contact. The accompanying table shows the mineral composition of both the hornblende rapakivi and its green marginal variety.

Hornblende rapakivi is a greenish brown-gray rock. It contains roundish potash feldspar phenocrysts 1 cm in diameter. The potash feldspar grains in many cases exhibit a distinct quadrille structure, and they contain 22.5 % perthite by volume. The quartz occurs both as euhedral and as anhedral grains. The composition of the plagioclase is An_{34} and at the margins of the grains against the potash feldspar under An_{20}. The dark greenish-brown hornblende is poikilitic, containing other mineral grains, and it penetrates the plagioclase. The biotite occurs in the same way.

At the southwestern end of this intrusion, where the contact zone can be observed — because trenches had been dug there during World War I —, the hornblende rapakivi undergoes a gradual transition about five meters from the contact into a green marginal variety. The transition takes place over a distance of about 10—20 cm, and the contact of the marginal variety against the country rock — which in this case consists of gabbro-anorthosite — is as sharp as a knife. The composition of the country rock is completely immaterial to the occurence of the marginal variety, for occurences of the same kind are met with in the Ahvenisto area against, for instance, granite and migmatite. The same impression of the significance of the country rock is also given by the literature (SOBRAL 1913, HACKMAN 1934). But important to the formation of the green marginal variety, on the other hand, is that wollastonite (wo) is included in the normative mineral composition calculated from the chemical analysis of the erupted rapakivi magma.

The green hue is induced by the weed texture formed by hornblende occurring as fine bands; the weed texture has a green appearance (SAVOLAHTI

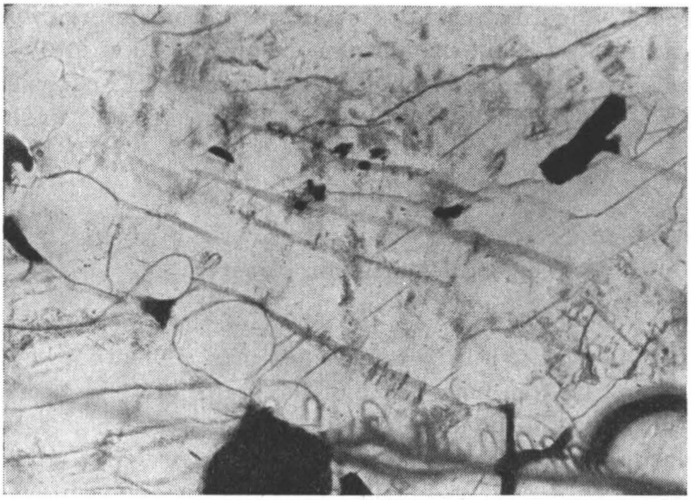

Fig. 5. Weed texture formed by hornblende around and inside quartz- (round), potash feldspar- (large grain in middle) and plagioclase- (in lower part of picture) grains. Without nicols, × 85. Nurmaa, Mäntyharju, Ahvenisto area (SAVOLAHTI 1956).

1956, 1956 a). The hornblende has worked itself around the mineral grains very intensively as hair-thin bands and penetrated them by entering cracks. This is clearly shown by Fig. 5. Since some of the cores of large potash feldspar grains have escaped penetration by »weeds,» reddish potash feldspar grains can occasionally be found in the rock. The marginal variety differs from hornblende rapakivi in many other respects besides its green color. It has, for instance, a different mineral composition, for it contains iron-rich olivine (fayalite) and also pyroxenes as new minerals. The hornblende grains are more poikilitic than in the central part of the intrusive. The perthite of the potash feldspar occurs as thinner streaks than in the hornblende rapakivi, and its quadrille structure is not so conspicuous as in the rock of the central portion. Myrmekite occurs abundantly at the margins of plagioclase grains against the potash feldspar, and the micrographic texture formed by the quartz and the potash feldspar is common. Apatite and zircon appear to occur as larger grains here than in the rock of the central portion. All the phenomena listed seem to fit in neatly with the conception that the border zone of the intrusive started to crystallize first and solidified faster than the central portion, and that, as the central portion crystallized, there took place in the marginal area autometamorphic as well as autometasomatic changes in texture and composition, together with a growth of minerals. Autometasomatism is revealed, *e.g.*, by the modes (Table 1) of the hornblende rapakivi and its green marginal variety, where the increase in the amount of potassium is obvious. According to the observations of REIN (1955) and MEHNERT (1960), the enrichment of potassium into the marginal parts of a massif from acid rocks in general is natural. Also, the enrichment, for instance, of mafic minerals takes place in certain cases in the marginal variety.

THE FAYALITE OF THE GREEN MARGINAL VARIETY OF HORNBLENDE RAPAKIVI

The average length of the fayalite grains of the green marginal variety of the hornblende rapakivi of Nurmaa is 0.5 mm. Sometimes one may find grains as much as 1.5 mm long. Some of the grains are beautifully euhedral, some plainly allotriomorphic. Fig. 2 (Plate I) gives an example of each mode of occurrence. Such modes of occurrence suggest that, upon crystallizing, the fayalite possessed powerful »form energy» but solified at a fairly late stage. The fayalite is relatively fresh in its occurrence, though in places at its margin it is covered with a yellowish brown alteration product. Table 2 presents the chemical analysis and the physical properties of this fayalite (SAVOLAHTI 1951).

Table 2. Chemical composition of the fayalite of the green marginal variety of hornblende rapakivi.[1]) Nurmaa, Mäntyharju.

	Weight per cent	Molecule ratio	Oxygen ratio	Metal ratio	Metal atoms
SiO_2	29.85	0.497	0.994	0.497	0.99
TiO_2	0.23	0.003	0.006	0.003	0.01
Al_2O_3	0.71	0.007	0.021	0.014	0.03
Fe_2O_3	1.48	0.009	0.027	0.018	0.04
FeO	64.87	0.903	0.903	0.903	1.80
MnO	1.82	0.026	0.026	0.026	0.05
MgO	1.25	0.031	0.031	0.031	0.06
CaO					
H_2O+	0.04	0.002	0.002	0.004	0.01
H_2O-	0.07				
Σ	100.32		2.010		

(Metal atoms TiO_2 to MgO: $\}$ 1.99)

$$\frac{4}{2.010} = 1.990$$

[1]) (Impurities under 1 %) ($Fo_3\ Fa_{94}\ Te_3$)

The analyzed fayalite has the following physical properties:

Specific gravity = 4.4 (with a pycnometer)

$$\left.\begin{array}{l}\alpha = 1.821\\ \beta = 1.863\end{array}\right\}\text{pleochroism weak}$$

$\gamma = 0.052$ (with a Berek compensator)
$2V\alpha = 49°$
010 — cleavage occurs

The occurence of iron-rich olivine (fayalite) in conjunction with quartz is natural, although magnesium-rich olivine has no permanence with quartz. Fayalite is a low-temperature form in the olivine series and a permanent form in lieu of pyroxene when the Fe-component among the minerals exceeds 87 %. This is indicated by experimental studies with silicate melts as well as by thermo-chemical tests performed by SAHAMA and TORGESSON (1949). An analysis of fayalite contained in granite pegmatite (Rockport, Mass.) with a total absence of magnesium has been published by BOWEN, SCHAIRER and POSNJAK (1933). The fayalite of the quartz fayalite gabbro of eastern Greenland has been described by DEER and WAGER (1939). In respect to chemical composition and optical properties, the fayalite of Nurmaa very closely corresponds to it. According to YODER and SAHAMA (1957), the Greenland fayalite's $d_{130} = 2.8286$. Fayalite has been met with, in addition, in obsidians, ryolites, trachytes, fonolites, sodalite-nepheline-syenites and quartz syenites. GREIG, MERWIN and SHEPHERD (1933) have managed to produce fayalite out of FeO and SiO_2 by applying hot steam. This serves as evidence to support the view that the fayalite of Nurmaa also formed at a low temperature, not having crystallized until autometasomatic and autometamorphic stages had been reached.

This rock also contains pyroxenes to some extent. It would be interesting to know their compositions. Does, for instance, an increase in the Al-content cause a lowering of the temperature of formation?

THE AMPHIBOLE OF THE GREEN MARGINAL VARIETY OF HORNBLENDE RAPAKIVI

A chemical analysis has also been made of the amphibole of the green marginal variety of the hornblende rapakivi, together with a determination of its physical properties (Table 3) (SAVOLAHTI 1951). The analysis has been calculated according to FOSLIE (1945). Rapakivi amphibole has previously been investigated by POPOFF (1928). In pyterlite he met with blue-green hornblende having a small optic axial angle. Elsewhere, too, in the rapakivi area of Viipuri, according to POPOFF, there exists the same kind of amphibole. SAHAMA (1947) has investigated a rapakivi amphibole at Uuksunjoki, Salmi, which corresponds to ferrohastingsite in composition. The amphibole of Nurmaa differs from this mineral mainly in that it is richer in Mg.

Table 3. Chemical composition of the amphibole of the green marginal variety of hornblende rapakivi. [1]) Nurmaa, Mäntyharju.

	Weight per cent	Molecule ratio	Oxygen ratio	Metal ratio	Metal atoms
SiO_2	39.78	0.662	1.324	0.662	6.35
Al_2O_3	10.14	0.099	0.297	0.198	1.90
TiO_2	2.32	0.029	0.058	0.029	0.28
Fe_2O_3	2.96	0.019	0.057	0.038	0.36
FeO	25.68	0.357	0.357	0.357	3.42
MnO	0.31	0.004	0.004	0.004	0.04
MgO	3.58	0.089	0.089	0.089	0.85
CaO	10.55	0.188	0.188	0.188	1.80
Na_2O	1.62	0.026	0.026	0.052	0.50
K_2O	1.60	0.019	0.019	0.038	0.36
H_2O + ...	1.18	0.065	0.065	0.130	1.25
H_2O —	0.02				
F	0.69	0.036	0.036	0.036	0.35
	100.43		2.520		
—O	0.29		0.018		
Σ	100.14		2.502		

$$\frac{24}{2.502} = 9.592$$

Si	6.35	$Z = 8 \begin{cases} \text{Si } 6.35 \\ \text{Al } 1.65 \end{cases}$	
Al	$1.90 \begin{cases} 1.65 \\ 0.25 \end{cases}$		
Ti	0.28	Y = 0.89	xy = 5.20
Fe***	0.36		
Fe**	3.42		
Mn	0.04	X = 4.31	
Mg	0.85		
Ca	1.80		
Na	0.50	W = 2.66	
K	0.36		
OH.........	1.25	OH, F = 1.60	
F	0.35		

[1]) (Impurities contained in analysis material, 0.3 %)

The analyzed amphibole has the following properties:

Specific gravity = 3.40 (with a pycnometer)
$b//\beta$; $c \wedge \gamma = 23°$
$\alpha = 1.687$; pale greenish brown
$\beta = 1.703$; brownish green
$\gamma = 1.716$; bluish green
$2V\alpha = 60°$
$\gamma—\alpha = 0.028$ (with a Berek compensator)

THE BIOTITE OF THE GREEN MARGINAL VARIETY OF HORNBLENDE RAPAKIVI

The biotites of rapakivis are very rich in iron. WAHL (1925) proposes for them the term »monrepite.» The rapakivi biotites also contain considerable fluorine. In his spectrochemical studies, RABINOWITSCH (1957) noted that rapakivi biotites contain relatively abundant Fe, Ti, Ca, Ba, Sn and Li. The biotites found in rapakivis have fairly high indices of light refraction, which, it is believed, are due to the high contents of iron and fluorine (*cf.* p. 65). These are characteristic features also of the biotite of the green marginal variety of the Nurmaa hornblende rapakivi (Table 4). This mineral contains impurities to a slightly greater extent than the ones described earlier, owing to the fact that it is scaly. Thus, upon being separated with heavy liquids, the impurities are left either underneath or on top of the scales rather than being eliminated from the main fractions. ESKOLA (1949) has published a chemical analysis of biotite from fresh rapakivi situated in Soukainen, Laitila. The percentages by weight in it are: Sio_2 34.72, TiO_2 3.75, Al_2O_3 11.09, Fe_2O_3 6.12, FeO 28.96, MnO 0.45, MgO 2.44, CaO 0.79, Na_2O 0.53, K_2O 7.57, P_2O_5 0.01, H_2O— 0.75, F 0.27, V_2O_5 0.01 and the total amounting to 99.90. The biotite of Soukainen yields the figures $\gamma = \beta = 1.682$ (dark brown).

Table 4. Chemical-composition of the biotite of the green marginal variety of hornblende-rapakivi. Nurmaa, Mäntyharju. Analyst, E. Lindqvist.

	Weight per cent				
SiO_2	35.60	Si	5.64		$Z = 8.00$ { Si = 5.64, Al = 2.29, Fe = 0.07 }
Al_2O_3	12.28	Al	2.29		
TiO_2	1.90	Ti	0.46		$Y = 0.94$
Fe_2O_3	4.65	Fe***	0.55	{ 0.07, 0.48 }	
FeO	29.47	Fe**	3.91		
MnO	0.30	Mn	0.04		$X = 4.69$
MgO	3.14		0.74		
CaO	1.92	Ca	0.32		
Na_2O	0.89	Na	0.27		$W = 1.94$
K_2O	6.67	K	1.35		
P_2O_5	0.11	P	0.02		
H_2O + ...	3.11	OH	3.30		$(OH),F,P_2O_5 = 3.79$
H_2O —	0.00	F	0.47		
F	0.93				
	100.97				
—O	0.39				
	100.58				

The following properties of the analyzed biotite have been determined:

$$\gamma = \beta = 1.695 \quad \text{(black-brown)}$$
$$2V\alpha = 20^\circ$$
$$\text{Specific gravity} = 3.27 \quad \text{(with a pycnometer)}$$

d) THE INFLUENCE OF THE BIOTITE RAPAKIVI CONTACT IN THE AHVENISTO AREA

At the western border of the Ahvenisto massif on the northern side of Kesiönjärvi, there may be seen a contact of reddish brown biotite-rapakivi, the dip of which is 50°W (in some spots less steep) (SAVOLAHTI 1956). The country rock consists of a fine-grained (0.2 mm) biotite-plagioclase gneiss, which is clearly schistose (N 30°E, 80°W) (Pl. II). The mineral composition of this rock about 5 cm from the contact, immediately next to the metamorphosed marginal zone, is shown in Table 5, No. 1. It contains sericite and epidote, in the main, as accessory minerals. There is less potash feldspar than quartz. The occurrence of pyroxene evidently is due to thermometamorphosis induced by the rapakivi. A microscopic examination gives the impression that in this rock pyroxene is evolving at the expense of other minerals. The composition of the plagioclase in this rock is An_{32} ($\gamma = 1.553$, $\alpha = 1.544$) and the refractive index of the biotite $\gamma = \beta = 1.646$.

At the contact against the rapakivi, the biotite-plagioclase gneiss has conspicuously metamorphosed across a breadth of 1—5 cm. Its schistosity has totally disappeared, and it has become greenish and coarser of grain (0.6 mm) than before (Plate II, Fig. 1). (See also SAVOLAHTI 1956 a, Fig. 14.) The mineral composition of this variety is shown in Table V, No. 2. A portion

Table 5. Modes of biotite-plagioclase gneiss (1) and metamorphosed biotite-plagioclase gneiss at contact of biotite rapakivi (2). Kesiönjärvi, Ahvenisto area.

<table>
<tr><th></th><th>1</th><th>2</th></tr>
<tr><td>Plagioclase</td><td>46.9</td><td>37.1</td></tr>
<tr><td>Quartz</td><td rowspan="2">2.8</td><td>6.3</td></tr>
<tr><td>Potash feldspar</td><td>26.7</td></tr>
<tr><td>Biotite</td><td>31.6</td><td>8.3</td></tr>
<tr><td>Hornblende</td><td rowspan="2">12.5</td><td rowspan="2">10.8</td></tr>
<tr><td>Chlorite</td></tr>
<tr><td>Clinopyroxene</td><td rowspan="2">4.4</td><td rowspan="2">10.1</td></tr>
<tr><td>Orthopyrexene</td></tr>
<tr><td>Accessories</td><td>1.8</td><td>0.7</td></tr>
<tr><td></td><td>100.0</td><td>100.0</td></tr>
</table>

Table 6. Chemical analyses of biotite-plagioclase gneiss (1) and metamorphosed biotite-plagioclase gneiss at contact of biotite rapakivi (2). Kesiönjärvi, Ahvenisto area. Analyst, Mr. E. Lindqvist.

Weight percentage	1	2	Normative compositions	1	2	Niggli values	1	2
SiO_2	49.98	58.98	Q	—	9.3	al	27	30
Al_2O_3	18.09	14.05	C	0.9	1.2	fm	47	38
TiO_2	0.86	0.60	or	14.5	26.5	c	12	12
Fe_2O_3	1.03	2.29	ab	35.0	32.0	alk	14	20
FeO	11.70	8.96	ne	1.2	—	si	125	183
MnO	0.14	0.09	an	21.5	6.5	k	0.28	0.47
MgO	5.41	2.07	hl	0.4	0.2	mg	0.43	0.25
CaO	4.68	3.56	ol	23.1 $\begin{cases} \text{fo } 11.3 \\ \text{fa } 11.8 \end{cases}$	—	qz	—351	+3
Na_2O	4.19	3.58	en	—	5.6			
K_2O	2.45	4.41	fs	—	10.4			
P_2O_5	0.00	0.53	mt	1.4	3.0			
H_2O+	1.40	0.64	il	1.2	0.8			
H_2O-	0.08	0.04	ap	—	2.0			
F	0.18	0.60	fr	0.8	2.5			
Cl	0.10	0.03						
	100.29	100.43						
—O	0.10	0.26						
	100.19	100.17						

of the plagioclase grains (An_{27} $\gamma = 1.549$, $\alpha = 1.543$) have grown, and former plagioclase grains have remained within the grown grains as inclusions. The crystal inclusions have beautiful twin lamellae, but they are indistinct, on the other hand, in the enveloping grains. Potash feldspar frequently occurs either within the plagioclase or around it as indeterminate streaks, and it also forms myrmekite along the edges of the plagioclase. This rock contains both rhombic pyroxene ($2V\alpha = 75°$, $\alpha = 1.732$, $\beta = 1.744$, $\gamma = 1.752$) and monoclinic pyroxene ($2V\gamma = 0°$, $\alpha = 1.708$, $\beta = 1.712$, $\gamma = 1.735$, $c \wedge \gamma = 34°$ (varies)). The pyroxenes occur as independent crystals, but also together as intergrown grains. The monoclinic pyroxene generally occurs in the middle portions of grains as vague splotches. Twin lamellae occur in the rhombic pyroxene. Some of them, perhaps, represent an exsolution structure, but to what extent is harder to say. Around the pyroxenes there regularly occurs a narrow green amphibole ring ($2V\alpha = 76°$, $c \wedge \gamma = 24°$, $\alpha = 1.672$, $\beta = 1.690$, $\gamma = 1.697$). Biotite often occurs in conjunction with quartz, and its index of refraction ($\gamma = \beta$) $= 1.662$, *i.e.*, much greater than in the case of the biotite of the biotite-plagioclase gneiss. The accessory minerals are principally apatite and ore.

The chemical analysis of both rocks is given in Table 6. The metamorphosis has resulted in the growth of SiO_2, K_2O and Fe_2O_3 and a decrease in Al_2O_3, FeO, MgO, CaO, Na_2O and H_2O. Fluorine and phosphorus have become enriched into the metamorphosed gneiss, whereas the chlorine appears to have migrated farther into the country rock, unless it had been included in the rock already. The determinations are parallel ones.

As further proof of the fact that the metamorphosis was certainly caused by the rapakivi, it may be mentioned that this rapakivi contains biotite-plagioclase gneiss fragments with a metamorphosed gneiss margin a few centimeters thick, like that described in the foregoing. This is not the only place where such a contact effect induced by biotite rapakivi can be seen in the Ahvenisto area (see also p. 76). The mineral assemblage that was created as the biotite-plagioclase gneiss metamorphosed seems to represent a higher facies than the mineral assemblage of the original country rock. This, for its part, provides evidence of the relatively high temperature at which the rapakivi solidified, as well as of the slightness of the rapakivi's graniticizing influence on its country rock. The reason may have been either extremely peaceful conditions or the magma's poverty of fugitive constituents during crystallization.

3. THE OCCURRENCE AND SIGNIFICANCE OF SHEAR ZONES AND FAULTS IN RAPAKIVI AREAS

The rapakivi area of Laitila lies in the valley of the Kokemäki river, where, as is well known, block movements have occurred. According to the general view, currently accepted, the sandstone of Satakunta found there has been preserved because it is situated in a fault trough (SIMONEN 1960).

The Rääveli lakes, which are located on the northern and northwestern sides of the Ahvenisto area, lie, according to FROSTERUS (1902), in a fault zone; and SAVOLAHTI (1956 a) has described shear and crushed zones of nearly all the different varieties of rock occurring in the area. They are to be met with also in the great rapakivi intrusive in the middle of the massif. The rapakivi of Ahvenisto, it has been observed, has undergone the most far-reaching metamorphosis in the cataclastic rock of the shear and crushed zone on the northwestern side of Kuijärvi. The strike of the zone is here N 55° W and its dip 70° E, and it comprises numerous bands varying between five and fifteen centimeters in breadth and recurring at diverse intervals. The rock contained in these bands is fine-grained, dark gray, partially streaked and schistose, and it includes broken quartz grains, muscovite and chlorite, but hardly any other minerals. Feldspars remain only at a few less sheared points. Fig. 1 (Plate III) shows the microscopic texture of this mylonitic rock. The rapakivi character is seen to have completely vanished and the rock to have taken on a typical cataclastic texture. Associated with this same shear and crushed zone, moreover, is a brecciated zone with angular fragments of rapakivi, the brecciating material consisting partly of fine-grained rock like that described in the foregoing and partly of pegmatite containing quartz, feldspar and fluorite. Of these minerals, each may be predominant here and there in the pegmatite.

Some three kilometers to the east from Enonvesi, there is a shear and crushed zone with a strike of N 60°W and dip of 60°W which runs through a rapakivi dike between fifty and a hundred meters broad intersecting an area of gabbro-anorthosite. The rapakivi has totally lost its texture in this roughly 20-cm-broad shear zone (Plate IV, Fig. 2) and turned into augengneiss. It contains eye-shaped potash feldspar grains 0.5—1 cm long with the quadrille texture of microcline plainly evident and the mafic minerals (biotite) situated around them. CARSTENS (1924, 1925) has described similar occurrences from the Trondhjem area, though there nearly the entire rock is sheared and crushed so that its origin is difficult to determine. SCHERMERHORN (1959) has described similar phenomena from the contact zones of the Sátão granite and applied to them the therm »protoclastic mylonitization.»

There are also zones in the rapakivis where the shearing is so slight that one finds only densely recurring and intersecting slickensides. A good example in the Ahvenisto area is the zone on the eastern side of Enonvesi (SAVOLAHTI 1956 a), with which is associated a more than ordinarily marked disintegration, probably resulting from this dislocation (see p. 54).

At the western border of the Ahvenisto massif, between it and Viilajärvi, on the southern side of the Heinola—Ahvenisto highway, one may see on the steep wall and on top of the rock what should be interpreted as a fault. There the rock contains a 5—10-cm-broad, vertical, straight, fine-grained and pale reddish brown aplitic vein, which runs in the rapakivi for a distance of about 20 meters from the contact toward the center of the massif. The rapakivi is markedly finer-grained between the contact and the vein than on the other side of the vein (Plate III, Fig. 2). A displacement had thus taken place along the vein. The aplitic vein is divided through its center by a thin, coarse-grained pegmatoid stripe. The part of the aplitic vein on the side of the country contact of the pegmatoid stripe is finer of grain than on the other side. This may indicate that the one side originated during the displacement movements from smaller-grained rapakivi and the other side from coarser-grained rapakivi. Evidently, it is necessary to conclude that the coarser-grained rapakivi had risen upward. The upward movement need not, to be sure, have been very great, since the contact is so near — and if the displacement happened to have taken place at a point where the transition of the contact variety to a coarser-grained rock was most rapid.

In such a case, the fault would merely have cut short this gradual transition. True, at the western margin of the Ahvenisto massif, one scarcely notices that the marginal zone is finer-grained than the principal rock. Accordingly, it may be stated also that this fault has brought a chilled contact to the fore here, too.

On an excursion organized in the autumn of 1958 by the Geological Survey to the area covered by the Hamina map sheet of surficial deposits, which had just been published, Dr. K. Mölder, the guide, expressed the view — based on his personal researches — that the Viipuri rapakivi area has blocks depressed during displacement movements where, perhaps, preglacial »moro» still existed. This would explain, at least in part, the fact that precisely the same type of rapakivi has been subjected to the »moro» disintegration process in some places and not in others. The fault in the rapakivi described in the foregoing certainly supports Mölder's view.

The faults and shear zones occurring in the rapakivi areas are significant for many reasons. First, in certain cases, they offer two natural explanations for the capricious behavior of rapakivi in disintegrating. One is that the moro was preserved on top of sunken blocks of the rock, and the other is that the dislocation and crushing of the rock brought about a greater local tendency to disintegration than otherwise.

Second, different rapakivi types are sometimes met with occurring close together in apparently mutually incongruous association. Segmental movements taking place in the rapakivi areas may explain in a natural way some of these modes of occurrence. Fig. 2 (Pl. III) also reveals that sharp contacts may exist between rapakivi types which are quite contemporaneous and have the same derivation but differ only in that one has risen higher than the other. Thus, the displacement on both sides of the boundary results in rocks plainly differing even megascopically, even though they had crystallized simultaneously from the same magma. The remaining modes of occurrence are attributable partly to the rock's serving as a base or roof and partly to later intrusions and to the rock's becoming an inclusion.

Third, the existence of the crushed and shear zones and of the numerous slickenslides occurring outside them testifies — at least in part — also to the fact that the attempts to interpret the cleavages in the rapakivi of the Vehmaa area magmatectonically have not succeeded (Kanerva 1928). It is apparent that measuring the rapakivi cleavages would not be sensible unless it were first possible to distinguish between the cleavages and the slickensides. The latter need have no relation to the cleavages but they do not have any marked effect on statistics. For example, the slickenside direction N 60° W and perpendicular direction N 30° E as well as N 30° W would form a strong maximum on cleavage diagrams. But such maxima do not necessarily tell us a thing about the formation of rapakivi.

Fourth, the rock shown in Fig. 2 (Pl. IV) and its genesis also have a broader theoretical significance, which might be mentioned in this connection, although the question does not bear directly on the present paper. In this case, we have an augen gneiss with microcline porphyroblasts deriving from potash-rich rock with a so-called ideal granitic composition (ESKOLA 1948). On the other hand, gneisses containing potash feldspar porphyroblasts are frequently interpreted as products of granitization. It is apparent that the rock evolving through metamorphosis is in this case no richer in potash or more »granitic» than the biotite rapakivi from which it derived. The original texture and minerals of the rapakivi have only broken up and largely become lost. The end product is a new rock with a new texture and recrystallized minerals. The potash feldspar, at least in this case, crystallized, because it favors movements, into larger grains and grain clusters than the other augen gneiss minerals. This observation is an interesting one and provokes questions and arguments. Has, for instance, sufficient attention been paid to the original chemical composition of the so-called graniticized rocks before it has been claimed that they have originated as a product of granitization?

4. ON THE DISINTEGRATION OF RAPAKIVI

ESKOLA (1930) has written at length on the disintegration of rapakivis to »moro,» as he puts it, and at the same time presented a thorough historical review of the research into the problem of the disintegration of rapakivis. Here it would suffice, to begin with, simply to cite in brief the causes of the disintegration mentioned by him and his conclusions. The disintegration of rapakivi appears to require the simultaneous occurrence of several circumstances. The chief prerequisite seems to be that the marginal surfaces between the mineral grains are straight and fairly smooth. The disintegration is caused by the expansion and shrinkage of the mineral grains as the temperature changes according to the weather. And after the disintegrating process has started, chemical disintegration likewise starts, inducing, e.g., also an alteration of the biotite. The dislocation of rock segments in certain vertical layers promoted the disintegration process, perhaps, and in the horizontal layers it is likely to be promoted by the very weight of the rock itself.

Let us now go on.

The mineral constituents of the rock shrink as the temperature decreases, and this shrinkage is not always in direct proportion to the change in temperature, for it proceeds in jumps. For example, quartz shrinks as the temperature falls from 575° to 500°C by 2.4 % (volume) but below this temperature considerably less (DAY, SOSSMAN and HOSTETTER 1914). There is reason to believe that the fundamental crystallization of rapakivis took place at a temperature above the normal and apparently above 575° C. This may be stated for the very reason that rapakivis often contain hexagonal bipyramidal quartz crystals, though quartz is not the only mineral constituent of rapakivis that may be assumed to have crystallized at a relatively high temperature (*cf.* p. 76). Accordingly, the quartz contained in rapakivi would have shrunk in sudden jumps considerably as the rock cooled, and naturally the whole rapakivi, too, would have done the same to a notable expent. Shrinkage, of course, causes cleavages and cracks and also breaks minerals apart — or at least loosens their bonds —, especially since different minerals do not shrink equally at the same temperature ranges, but some more and some less. Such a high crystallization temperature seems to have increased the tendency of rapakivis to disintegrate considerably. It may also be one of several reasons for the disintegration of porphyric granites. The marked shrinkage of minerals during the cooling down from a high temperature after crystallization may likewise be, at least partially, the cause of the disintegration undergone by the olivine diabase of Satakunta, as reported by ESKOLA (1930), since plagioclase shrinks to an unusual extent, too, at a temperature of about 1 050°C (DAY, SOSSMAN and HOSTETTER 1914).

The structure of rapakivi is certainly another factor in causing disintegration, for there are fine-grained, granitic, partly micrographic types that never disintegrate and the minerals contained in rapakivis have fairly straight boundaries against each other. It may be stated that the crystallization conditions in rapakivis often favor such a structure and, accordingly, the creation of a structure susceptible to disintegration.

A third factor is dislocation of the rock, which explains, *e.g.*, the fact that in the case of some rapakivi massifs the rock has disintegrated in its marginal zones more than elsewhere (SAVOLAHTI 1958) and that in the proximity of certain crushed and shear zones more disintegrating rapakivis are to be found than elsewhere (*cf.* p. 51). This would further explain the tendency of the marginal zones of certain porphyritic granite masses to disintegrate.

The chemical and mineralogical composition of the rock is a fourth possible and noteworthy factor. Examples are known of pyroxene-bearing rocks having a tendency to disintegrate.

These four main factors cause the disintegration of rapakivis as well as other rocks. In certain cases, one of them operating alone is likely to bring on disintegration in a susceptible rock. But if the several factors happen to operate simultaneously, the rock will, of course, undergo more drastic disintegration than otherwise. That there are several factors bearing on the tendency of rapakivis to disintegrate and that these factors vary abruptly from place to place even in the same rock provide an explanation to the capriciousness of the disintegration process.

However, there still remains the question as to why rapakivi disintegrates more readily than many other granites. It appears possible that the chief reason is the high crystallization temperature of rapakivi and the consequent extraordinary and marked shrinkage of the crystallized minerals in conjunction with the cooling of the rock. Another possible reason is that the crystallization of rapakivis often tends to take place under conditions favoring the development of a structure susceptible to disintegration (see p. 79). When a rock has for these reasons become so susceptible, the effects of dislocations act to promote disintegration more easily and to a higher degree than in the case of other rocks. The effect of a dislocation may even fail to become evident if a rock had not previously had a tendency to undergo disintegration. The tendency of a rock's chemical and mineralogical composition to promote disintegration is far stronger in many other rocks than in rapakivis. Sulfide ores are a good example. However, their chemical and mineralogical composition can complicate the capricious disintegration occurring in rapakivis.

The weather in Finland is not particularly conducive to the disintegration of rocks, and its action on all rocks is the same, as may be observed on comparing rocks situated fairly close together.

5. OBSERVATIONS ON THE MINERALS, PETROGRAPHY AND FABRIC OF RAPAKIVIS

Very much has been written on the fabric and petrography of rapakivis. The subject has attracted, *e.g.*, Sederholm (1891, 1916, 1928), Popoff (1897, 1928), Wahl (1925), Eskola (1928) and Savolahti (1956, 1956 a). In this chapter, it is intended to deal with observations of petrographic importance, in the present author's opinion, gathering them from old publications and presenting some new ones. Special attention has been given to such observations as cast light on the genesis of rapakivis.

a) THE OCCURRENCE OF QUARTZ IN RAPAKIVIS

The quartz in rapakivis crystallized, according to many investigators, as high quartz, and it is known as such on account of the hexagonal bipyramidal forms of its grains. This feature proves that the rapakivis crystallized at a relatively high temperature. The inversion temperature of high-low quartz rises as the pressure increases to 10 000 bars from 572°C to 815°C (Yoder 1950). All the rapakivi investigators have likewise observed that quartz occurs in two generations in rapakivis. Popoff (1928) distinguishes four different quartz generations in rapakivis. The first and principal generation, in his view, is idiomorphic quartz. The second, appearing like the first as independent grains, is xenomorphic quartz. The third and fourth quartz generations occur mainly as inclusions in potash feldspar, often in ovoid grains. The surfaces of the third-generation grains are more or less convex, but their idiomorphic character is still in evidence. The fourth generation is so-called *aussenkonkave Quarzeinschlüsse:* these quartz grains are indefinite in shape, with curved sides and rounded corners, though some with straight sides and sharp corners are also to be met with.

Fig. 1 (Plate IV) shows a potash feldspar phenocryst from the granite-porphyry of the Ahvenisto area. During the crystallization phase, magma inclusions resembling the type last described, or Popoff's fourth generation, was apparently left within it, to undergo later crystallization in the same way as the main mass. The form of the »magma drop» resembles the *aussenkonkave* quartz inclusions. Vogt (1921) has observed such an occurrence. Popoff (1928) regarded *aussenkonkave* quartz as primarily formed, simultaneously with potash feldspar.

Erdmannsdörffer (1949) took the position that *aussenkonkave* quartz inclusions are partly metasomatic. In this case, the inclusion in the potash feldspar is a primary one and of later crystallization than the potash feldspar. There is no reason to doubt that the potash feldspars of all the rapakivi types have inclusions formed the same way, although proving it except by generalizing from instances like this is difficult.

Fig. 1 (Plate V) presents two quartz phenocrysts from the granite-porphyries of the Ahvenisto area. Both are beautifully euhedral and both contain indefinite magma inclusions and tongues. These inclusions and tongues have undergone a more far-reaching crystallization than the main mass, for their crystals are larger than those of the main mass. The increase in the grain size, perhaps, had a catalyzing effect on the quartz grain, one promoting crystallization. Owing to the otherwise beautiful idiomorphic shape of the grains, the tongues cannot be the product of corrosion — though, naturally, corrosion can also occur (Foster 1960). Some of them are inclusions, besides, so that any argument favoring corrosion is made even less tenable. In this case, the inclusions are primary and younger — that is, had crystallized later — than the parent crystal. On the other hand, this does not prove that rapakivis do not also contain secondary quartz and other secondary minerals, too. (See p. 88).

It is quite probable that rapakivi minerals other than those just mentioned may likewise contain inclusions having a similar genesis.

b) POTASH FELDSPARS OF RAPAKIVIS

A generally held view is that orthoclase occurs in the rapakivis and microcline in the other granities of Finland. This conception is principally based on reports published by MÄKINEN (1917) and ESKOLA (1929). In describing the petrography of rapakivis, SEDERHOLM (1891) applies to the potash feldspar both terms, orthoclase and microcline. In using different words to designate the potash feldspar of different rapakivi types, he could scarcely have intended them as synonyms. According to WAHL (1925), viborgite contains both orthoclase and microcline, while pyterlite contains microcline. SAVOLAHTI (1956) describes the quadrille structure exhibited by a number of the potash feldspars found in rapakivi types in the Ahvenisto area. This structure was reported by him to appear more plainly in certain types than in others, and within the same type, even in a single grain, more plainly at the margin than in the middle.

Many researchers, especially during the decade of the 1950's, have studied potash feldspar roentgenographically and described their various structural forms. According to LAVES (1952), there are two stable modifications of potash feldspar: monoclinic sanidine, which is stable at a high temperature, and triclinic microcline, which is stable at a low temperature. The inversion temperature from one form to the other ought to be 650°—750° C, but, according to GOLDSMITH and LAVES (1954 a), it could be under 500° C. The symmetry difference between these two forms is understood to be due to the fact that they differ in respect to their Al—Si distribution. In microcline it is orderly, whereas in sanidine it is in a state of disorder. According to LAVES (1952) and GOLDSMITH and LAVES (1954 a, 1954 b), there exists a more or less continous process of conversion from the structure of monoclinic sanidine to that of triclinic microcline. The degree of conversion has been referred to as the triclinicity of potash feldspar (GOLDSMITH and LAVES 1954 b), *e.g.*, it serves as a measure of the degree of order in the Al—Si arrangements. Its value in the case of sanidine is 0 and in that of microcline 1. It has been observed both optically and roentgenographically that the degree of triclinicity varies in potash feldspar grains of the same rock and even in different parts of the same potash feldspar grain. GOLDSMITH and LAVES (1954 a) maintain that orthoclase does not represent a uniform phase but serves as a group term. Thus, it embraces monoclinic sanidines that are changing into triclinic microcline. In LAVES' (1952) view, the degree of conversion into microcline depends on the conditions of formation, involving mainly the factors of time and temperature. Orthoclase thus represents an unstable phase at lower temperatures, where it undergoes conversion to microcline. It should be mentioned, further, that potash feldspar, according to GOLDSMITH and LAVES (1954 a), has a strong tendency to crystallize as a high-temperature or disordered form (metastable) at temperatures within the stability range of microcline. Nevertheless, the monoclinic form probably represents, in general, a higher-temperature phase than the form being converted to microcline, and triclinic microcline the form of the lowest teperature.

A lecture on the potash feldspars contained in rapakivis was delivered in 1957 to the Geological Society of Finland by NEUVONEN, and a condensation of it has appeared in print (NEUVONEN 1957). The lecture included a report on the results of his roentgenographic studies of potash feldspars occurring in rapakivis. These results are in agreement with the observations concerning the mode of occurrence of potash feldspars discussed in the foregoing. It would be pertinent to consider the main results at this point. The degree of triclinicity of the potash feldspar contained in rapakivis varies in different grains as well as within a single grain or ovoid. Various rapakivi types differ somewhat as regards the distribution of the degree of triclinicity. In viborgitic rapakivis, monoclinic or weakly triclinic potash feldspar is the most common. In porphyric types, all the degrees of triclinicity are equally represented. In pyterlitic and even-grained varieties, there is a weak distribution maximum of 0.0—0.3 and another maximum of 0.75 triclinicity. In the rapakivis of southwestern Finland, the degree of triclinicity of potash feldspar is lower than in the rapakivis of the Viipuri area. In the quartz and granite porphyries occurring in a veinlike manner, there prevails a high 0.6—1.0 degree of triclinicity, but also a smaller maximum at the monoclinic end. In the granites of the surrounding bedrock, according to NEUVONEN, only potash feldspars with a high degree of triclinicity are generally to be found. In his lecture, he did not attempt to interpret variations in the degree of triclinicity. On the other hand, he did interpret the extraordinary albite content (approx. 30 %) of the perthitic potash feldspar of rapakivi as bespeaking a relatively high crystallization temperature — as ought to be done according to experimental investigations and potash feldspar studies carried out with other rocks. Of this perthite content, a portion cannot be seen under the microscope on account of the thinness of the veinlike albite streaks. The lecturer further noted that the perthite of the potash feldspar of rapakivi had unmixed in two stages, microperthite being the product at a high temperature and cryptoperthite at a low temperature. No mention was made of the An-content of the exsolved perthite. Describing the perthites of the potash feldspars of granites, SCHERMERHORN (1959) states their plagioclase composition to be An_{18}. WAHL (1925) reports the chemical composition of the orthoclase of viborgite to be as follows:

Orthoclase	65.7 %
Albite	30.1 %
Anorthite	4.2 %

The perthite of pyterlite, according to WAHL (1925), contains slightly less oligoclase than the perthite of viborgite. The oligoclase in pyterlite is also slightly richer in albite than is the oligoclase in the perthite of viborgite, and both are considerably richer in albite than are the plagioclase shells of viborgite (*cf.* p. 78).

The triclinicity of the potash feldspar of viborgite, according to the foregoing, was smaller than that of pyterlite and porphyric rapakivis. If we look upon the latter two as marginal varieties of the former (*cf.* p. 83), they had already crystallized when the middle portion was still undergoing solidification. On the other hand, the temperature gradient caused them to be the whole time at a lower temperature than the middle portion. Moreover, they remained at a lower temperature longer than the main mass. For these reasons, the potash feldspar of viborgite must have a smaller degree of triclinicity than the other rapakivis of the Viipuri area. The potash feldspars of the rapakivi of southwestern Finland, again, proved to have a slightly lower degree of triclinicity than the rapakivis of the Viipuri area. Since a small mass cools after crystallization considerably faster than a large one, the degree of triclinicity of the potash feldspar grains of a small mass has not the time to grow so large as those of a large mass. The two maxima observed in the changes in the degree of triclinicity of the potash feldspars of quartz and granite porphyries are due to the fact that, as I see it, the lecturer included in this group not only the quartz- and granite-porphyries but also the porphyry-aplites (see KANERVA 1928 and SAVOLAHTI 1956, 1956 a). However, they have quite a different crystallization history. The quartz and granite porphyry veins were produced by an independent injection of magma into a cold environment, with the result that the temperature of the injected magma could have varied and the cooling taken place rapidly. The porphyry aplites, again, represent a portion separated from the rapakivi magma during its crystallization at a stage when most of it had already solidified. The crystallization temperature of the magma forming the porphyry aplite in each case had corresponded to the crystallization phase of the main magma at the time involved, but not exceeded it. Likewise, their cooling had proceeded slowly in comparison with the former rocks. This conclusion is suggested by the circumstance that the porphyry aplites, certain small — and evengrained rapakivi granites, slight occurrences of aplites and pegmatites are the only formations belonging among the rapakivis that lack the contact phenomena associated with cooling (see pp. 38, 83 and *cf.* ESKOLA 1961, p. 120). Therefore, they had erupted while the main mass had still been warm. For another thing, their mode of occurrence is such, *i.e.*, in the marginal parts of the rapakivi massifs, whether radially or parallel to the contact (SAVOLAHTI 1956, 1956 a), that it indicates an eruption from the main mass somewhere farther. Certain small- and even-grained rapakivis are also vein-like formations, which may also be late differentiates of the main mass.

Wahl (1925) distinguishes two potash feldspar generations. Eskola (1928), Eckermann (1937) and Savolahti (1956, 1956 a) report features of potash feldspars supporting this view.

Finally, it may be iterated that the abundant perthite content of potash feldspars in rapakivis and the comparatively low degree of triclinicity evidently signify a fairly high crystallization temperature.

Simonen (1961) has likewise concluded that rapakivis have crystallized at a high temperature, as compared with granites. His arguments are based on hypotheses which have been challenged, in Finland, too, by well-known, convincing experimental observations (Winkler and Platen 1958 and Winkler 1961) and theoretical reasoning (Winkler and Platen 1959 and Savolahti 1959). Until the justified doubts raised by the counter-arguments have been removed, the results naturally cannot be trusted. No attempts to do so have yet appeared in the literature.

c) THE OCCURRENCE OF PLAGIOCLASE IN RAPAKIVIS, AND ITS COMPOSITION

Plagioclase is quite an idiomorphic mineral in rapakivis. The thin section of many rapakivis shows it to be the most idiomorphic of the principal minerals contained in the rock (see p. 91). Wahl (1925) distinguishes two different plagioclase generations. Savolahti (1956 a) speaks of two-part plagioclase grains, of which the central part is idiomorphic and the marginal part, sharply bounded against it, contains quartz inclusions. In granite- and quartz-porphyries, which are hypabyssal and, in part, almost effusive rocks, the plagioclase is present, in addition to the potash feldspar and quartz, as independent phenocrysts. This indisputably proves, among other things, that two feldspars crystallized out of the rapakivi magma at a fairly early stage, or right after the composition of the magma had reached the two-feldspar field.

Various alteration products are present in the plagioclase of rapakivis in varying amounts, sometimes very small, sometimes quite large. Plagioclase phenocrysts contained, *e.g.*, in quartz and granite porphyries are, as is well known, highly sericitized, epidotized and/or partially phrenitized (Eskola 1933). Perhaps the reason was rapid cooling and the subsequent altering effect of the volatiles. Myrmekite, a micrographic texture and albitization are also met with in rapakivis. Phenomena of the last-mentioned type have been extensively and thoroughly described by, *e.g.*, Schermerhorn (1959) from the granites of northern Portugal.

The composition of the main portion of the plagioclase contained in rapakivis generally ranges between An_{25} and An_{40} (SAVOLAHTI 1958). According to WAHL (1925), the composition of the plagioclase shells enveloping the potash feldspar ovoids found in viborgite approximates 30 An, while an analysis of the plagioclase contained in the ovoids yielded the following composition:

Orthoclase	8.7 %
Albite	61.2 %
Anorthite	30.1 %
Total	100.0 %

The present writer has met with plagioclase containing over 40 % An in some varieties of rapakivi. On the other hand, the An-content of plagioclase in certain late-stage differentiates is consistently smaller than in others. In contact with potash feldspar, the An-content of plagioclase has frequently decreased — as TUTTLE (1952) reports on granites. Still, according to SIMONEN (1960), for instance, the plagioclase of the so-called microcline-granites of Finland contain between 12 % and 25 % An. Accordingly, a clear difference exists between the An-contents of the plagioclase contained in these different granitic rocks. This difference is all the more remarkable in the light of SAHAMA's (1945) observation: »When compared with other granitic rocks the rapakivi granites . . . contain more SiO_2 and K_2O, and less MgO, CaO, as well as Al_2O_3.» SAVOLAHTI (1958) surmises that, in the crystallization conditions of rapakivis, plagioclase (An_{25-40}) has a greater tendency to crystallize than plagioclases richer in albite. Thus, the chemical composition of the rock would not directly determine the composition of the plagioclase crystallizing at any given time from the magma. Since the quartz and potash feldspar crystallized in rapakivis at rather high temperatures compared to the microcline granites, it is natural to interpret the high anorthite content of the plagioclase of rapakivis as also having been brought about by the high crystallization temperature.

According to WENK (1958), there are »discontinuities» to be noted in plagioclase series of metamorphic origin. By this he means that the plagioclases of differing composition are not equally prevalent, some being quite common and others being totally absent. Furthermore, he holds the view that plagioclase can be a better facies indicator than aluminium-rich mafic minerals. DE WAARD (1959) has published observations supporting this view. All the indications, therefore, are that the plagioclase series is not ideally isomorphic — not, at least, under all circumstances — but that there are disturbances in it. The internal energy of the plagioclases in formation does not apparently change in a direct line as a function of their composition, whereupon the chemical composition of a rock does not alone determine the composition of the plagioclase.

The crystal structure of plagioclase has received much studious attention, along with structural modifications and variations in its properties. EITEL (1958) discusses the literature of the subject at length, expressing the opinion that certain of the results arrived at are important to geological thermometry. LAVES (1954) reports that the plagioclases An_5—An_{17} exhibit a structure of submicroscopic unmixing of An_5 and An_{30} fractions (peristerite structure). Na-oligoclase would, according to this evidence, be less stable and, perhaps, less ready to crystallize than, *e.g.*, plagioclase An_{30}. Evidently, Ca-rich oligoclase is also more stable at high temperatures than the Na-rich variety, although Na-oligoclase is likewise stable there. Thus, it crystallizes alongside potash feldspar more readily than Na-rich plagioclase. DE VORE (1956) has discussed the problem of the stability of the atomic arrangements in the plagioclase series from the viewpoint of structural bond energies. In his view, the most probable plagioclase compositions that can be ordered are those with 33.3, 50.0, 66.6 and 75.0 An. The discontinuity of the plagioclase series at about the An_{20}—$_{30}$ region is also noticeable in changes of the optical properties (SMITH 1958). It is certain that the plagioclase series is not ideally isomorphic.

It may be stated, therefore, that plagioclase, too, crystallized in rapakivis at a relatively high temperature, providing further proof of the magmatic nature of rapakivis, and that the common crystallization range of Ca-rich oligoclase and potash feldspar is apparently wider than experimental investigations with Ab-Or mixtures indicate.

d) THE BIOTITES OF RAPAKIVIS AND ACID ROCKS

The biotites of rapakivis are in many cases quite dark of color and have high refractive indices, compared to the indices of biotites contained in other acid rocks. The very dark hue of the biotites of rapakivis is often brownish, but it is also sometimes greenish. ROY (1949) has observed that in hydrothermal experiments up to 400° C, biotite is green. Table 7 presents the refractive indices (γ—β) of the biotite contained in various rapakivis and other acid rocks. Fig. 6 shows the values of Table 7 arranged on a diagram. They clearly reveal the variation in the refractive indices of different rocks. WAHL (1925) understands it to be due to the high iron content of the biotite met with in rapakivis. According to SAHAMA (1945), rapakivi biotites also have a high content of fluorine. An examination of the table and the figure reveals, however, that the fluctuations in the iron and fluorine contents do not suffice alone to explain the variations in

Table 7. The refractive indices of biotite from acidic eruptive rocks.

No.	Name and location of rock	γ—β	
1	Rapakivi, Suurkuukka quarry, on northern side of Tani, Lappee. (3352, No. of collection) ..	1.688	fm = 15, mg = 0.20
2	Typical Lappee granite, north of Hytti depot, Lappee (3357)	1.671	fm = 23, mg = 0.08
3	Viborgite, Kausala (from excursion in 1957) ..	1.691	
4	Viborgite, stone quarry of Kymi Oy, Kuusankoski (from excursion in 1957)	1.683	
5	Rapakivi, 5 km west of Lappeenranta (402, Steinheil's collection)	1.685	
6	Viborgite, northwest of Lappeenranta (excursion in 1955)	1.692	
7	Spotted granite, Toivarila, alongside Kanalampi highway, Lappee. (3358)	1.668	fm = 13, mg = 0.17
8	Sinkko granite, from near Värtölä, Lappee (excursion 1955)	1.678	
9	Fine-grained rapakivi, stone quarry of Lahti, Artjärvi (excursion 1955)	1.687	
10	Granite of Aland Islands, from island on western side of Lemström canal. (503)	1.659	
11	Rapakivi, Patteri, Boxo, Saltvik, Aland Islands (7884)	1.671	
12	Rapakivi, Rödglasskär, Kumlinge, Aland Islands (7885)	1.687	
13	Granite of Taivassalo, northern quarry, Helsinki, Taivassalo, Vehmaa (7971)	1.673	fm = 15.5, mg = 0.08
14	Fine-grained granite inclusion in Taivassalo granite, northern quarry, Helsinki, Taivassalo, Vehmaa (7962)	1.673	
15	Rapakivi granite (spotted granite), Lahdinko, 500 m west of Torkkila, Vehmaa (9973)	1.689	
16	Rapakivi, Eurajoki station, Turku province (404)	1.675	
17	Rapakivi, Soukainen, Laitila	1.683	
18	Biotite rapakivis, Ahvenisto area (Savolahti 1956)	1.672—1.692	fm = 14, mg = 0.15
19	Green marginal variety of hornblende rapakivi, Nurmaa, Mäntyharju (Savolahti 1956)	1.695	fm = 25, mg = 0.19
20	Hornblende rapakivi, Hietaniemi, Mäntyharju (Savolahti 1956)	1.697	fm = 25, mg = 0.15
21	Porphyry aplite, 1 km east of Korpijärvi, commune of Heinola (Savolahti 1956)	1.697	fm = 17, mg = 0.12
22	Rapakivi, from the first large cut through rock on the road running west from the station of Kavantsaari (3078)	1.690	
23	Rapakivi,Uuksunjoki,area of Salmi(Sahama 1947)	1.730	—2V = 22°
24	Rapakivi, Koirinoja, area of Salmi (Sahama 1945)	1.699	
25	Bodom granite, Vanhakylä, Åkerby, Espoo (6153)	1.687	
26	Coarse-grained granite, Kokemäki (519)	1.678	
27	Syenite-granite, Rautalampi (6078)	1.680	
28	Coarse-grained porphyric granite, Suomenniemi, Lavia (33)	1.670	
29	Biotite granite, so-called older porphyric granite, Korppi W, Orivesi (58)	1.663	
30	Gray porphyric granite, Kurjala shore, Leppävirta (669)	1.658	
31	Porphyric granite, on northern side of Suur-Pyhäkalajärvi, Toivarila, Lappee (3341)	1.668	

Table 7. (Continued).

No.	Name and location of rock	$\gamma-\beta$	
32	Porphyric granite, railroad cut on western side of Korppi, Orivesi (697)	1.661	
33	Porphyric granite, Jämsä (5501)	1.697	
34	Porphyric granite, Suonsalmi, Noukkala, Hirvensalmi (5505)	1.677	
35	Typical coarse porphyric granite, northeast of inn at Taipale (6963)	1.660	
36	Vaasa granite, from airfield, by main track, Kauhava (1)	1.650	fm = 1.2—12 mean value of granite analysis (SIMONEN 1960)
37	Granite, Vihijärvi, Toivakka (5542)	1.668	
38	Granite, Hirvijärvi, Joutsa (5497)	1.654	
39	Granite, western side of Tiirismaa, Hollola (6091)	1.659	
40	Granite, Finska Stenindustri AB, Kumlinge (567)	1.659	
41	Kuru granite, Kuru (SIMONEN 1960)	1.668—1.671	
42	Gray granite, Paloisvuori, Iisalmi	1.652	
43	Hanko granite, Hanko (21)	1.670	fm = 11, mean value of microcline granite analysis (SIMONEN 1960)
44	Microcline granite, Hanko (SIMONEN 1960)	1.675	
45	Microcline granite, Hanko (SIMONEN 1960)	1.671	
46	Microcline granite, Kirkkonummi (SIMONEN 1960)	1.669	
47	Microcline granite, Inkoo (SIMONEN 1960)	1.659	
48	Microcline granite, Ahvenisto area (SAVOLAHTI 1956)	1.658—1.665	
49	Microcline granite, Haveri, Viljakkala (5631)	1.665	
50	Microline granite, Labbis, Quarry, Vestanfjärd (5862)	1.658	
51	Microcline granite, from soil of Östergård, Kiila, Kemiö (5860)	1.649	
52	Nattanen granite, Nattanen (8742)	1.635	
53	Granite I, Suojainen, Ylöjärvi (532)	1.652	
54	Granite I, commune of Pielisjärvi (663)	1.640	
55	Granodiorite, 4 km from church of Mäntyharju to the south (SAVOLAHTI 1956)	1.662	fm = 13—22, mean value of granodiorite analysis (SIMONEN 1960)
56	Granodiorite, Rajala, Nokia (SIMONEN 1952)	1.655	
57	Granodiorite, Heinlampi, Nokia, (SIMONEN 1960)	1.655	
58	Granodiorite, Koijärvi (SIMONEN 1948)	1.651	
59	Granodiorite, Värmälä, Teisko (SIMONEN 1952)	1.666	
60	Quartz diorite, Juolavesi, Mäntyharju (SAVOLAHTI 1956)	1.657	
61	Quartz diorite, from between Vuohijärvi and Kelesjärvi, Jaala (SAVOLAHTI 1956)	1.648	
62	Quartz diorite, Työläjärvi, Ylöjärvi (SIMONEN 1952)	1.648—1.652	fm = 20—30, mean value of quartz diorite analysis (SIMONEN 1960)
63	Quartz diorite, Kalvola (Simonen 1960)	1.648	
64	Quartz diorite,Paarlahti N,Teisko(SEITSAARI 1951)	1.666	
65	Trondhjemite (Uusikaupunki granite), Uusikaupunki (533)	1.658	fm = 13—20, mean value of trondhjemite analysis (SIMONEN 1960)
66	Granodioritic trondhjemite, Haidus, Uusikaupunki (HIETANEN 1943)	1.657	
67	Quartz-dioritic trondhjemite, Uskela, Salo (LEHIJÄRVI 1957)	1.658	
68	Quartz-dioritic trondhjemite, Halikko (LEHIJÄRVI 1957)	1.665	
69	Granodioritic charnokite, Uusikaupunki (HIETANEN 1943)	1.657	

Table 7. (Continued).

No.	Name and location of rock	γ—β	
70	Quartz-dioritic charnokite, Nummi (PARRAS 1958)	1.673	
71	Quartz-dioritic charnokite, Vihti (PARRAS 1958)	1.664	
72	Quartz-dioritic charnokite, Vihti (PARRAS 1958)	1.663	
73	Quartz-dioritic charnokite, Lohja (PARRAS 1958)	1.662	
74	Quartz-dioritic charnokite, Vihti (PARRAS 1958)	1.660	
75	Quartz-dioritic charnokite, Merimasku (HIETANEN 1947)	1.651	
76	Quartz-dioritic charnokite, Vihti (PARRAS 1958)	1.650	
77	Quartz-dioritic charnokite, Vihti (PARRAS 1958)	1.648	

the refractive indices, although their influence is considerable. The high refractive index values of biotite have one further explanation. HELLNER and EULER (1957) have carried out experimental studies under hydrothermal conditions and noted that the refractive indices γ and β of the biotites contained in rocks rise when the temperature exceeds 500° C. For another thing, they have observed that the axis angle (2Vα) of biotite increases when the temperature begins to rise above 650° C. It is known, of course, that optically biotite often has just a single axis — or almost that. It is probably not merely a matter of chance that rapakivis include known exceptions: *e.g.*, the axis angle (2Vα) of the Uuksunjoki biotite is 22° (SAHAMA 1947), and that (2Vα) of the biotite of the green marginal variety of the Nurmaa hornblende rapakivi 20° (SAVOLAHTI 1956 a). Lack of transparency (WAHL 1925) interferes with the observation and determination of the axis angle — perhaps that is why so few of them are known. It may be that the biotites of rapakivis in general represent modifications of a higher temperature than the biotites of many other granites. According to SMITH and YODER (1956), the crystal structure of micas is a function of composition, pressure and temperature, and variation of structure in the natural micas cannot, therefore, be attributed solely to changes in pressure and temperature.

The refractive index of the rapakivi biotites shown in the table and the figure varies — also, it goes without saying, as a function of composition — so that as the Mg/Fe ratio of the rock decreases, the refractive index increases. On the other hand, the present author has reached the conclusion that the biotites of a greenish hue in rapakivis have lower refractive indices than others and that in those types where green biotite occurs, there often also takes place chloritization of the biotite.

SCHWARTZ (1958) writes that under hydrothermal conditions brown biotite alters first into green biotite as the refractive index decreases and then into chlorite. This would explain, at least partly, certain rather low refractive index values for biotite in rapakivis. The figure includes other

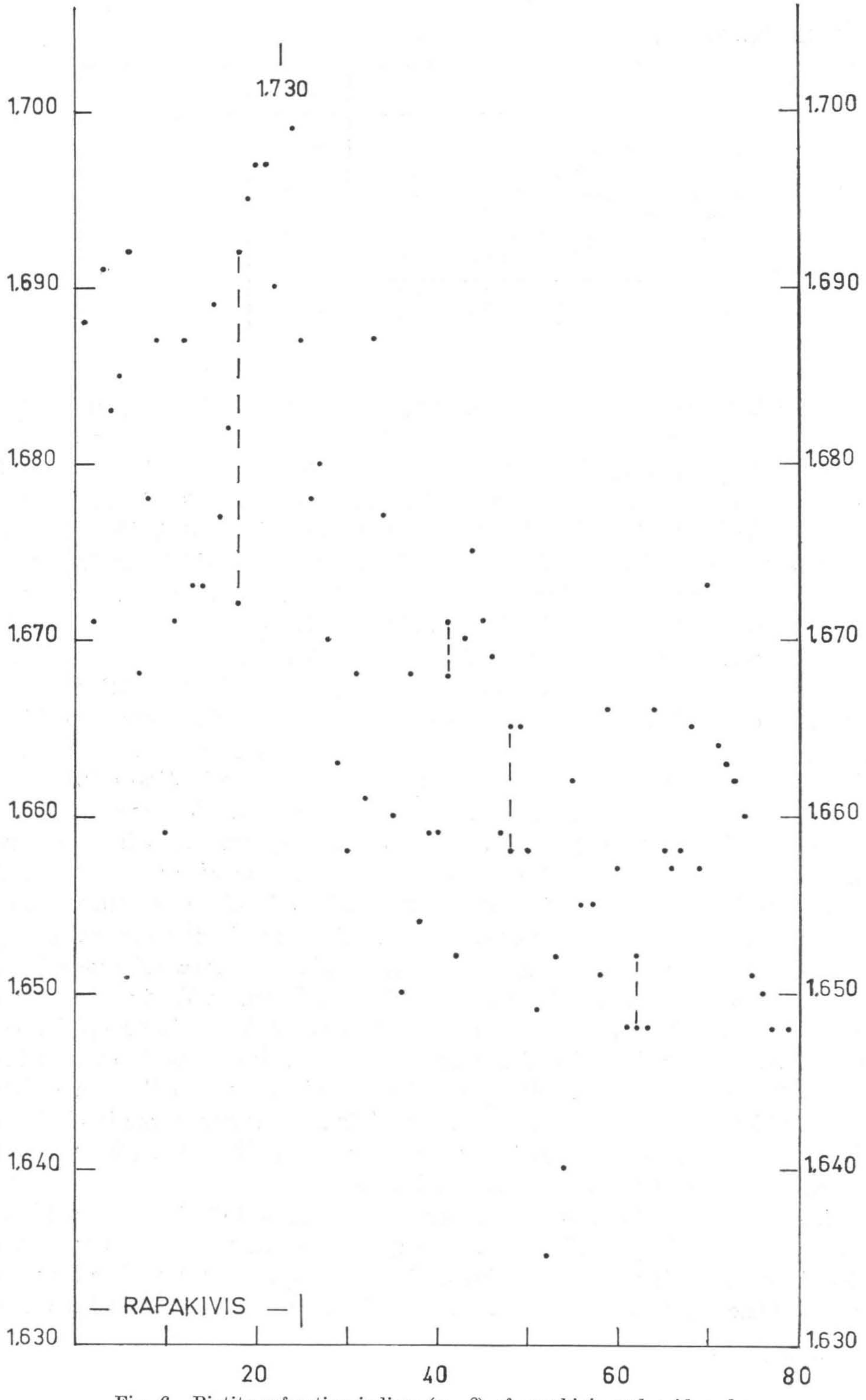

Fig. 6. Biotite refractive indices (γ—β) of rapakivis and acid rocks.

relatively high refractive indices for biotite, as in the cases of a number of porphyric granites and rocks of the charnockite series. Perhaps the reason is the same, in part, as in the case of the rapakivis.

If and when the occurrence of high refractive indices and large axis angles in rapakivis also signifies a high crystallization temperature, then it also bolsters the conception of the rapakivis' relatively high crystallization temperature and magmatic genesis. It would seem that certain of the porphyritic granites solidified at a fairly high temperature. But what was lacking to a certain extent in their case was the peaceful conditions that prevailed during the crystallization of the rapakivis.

e) MAGMATIC INCLUSIONS IN RAPAKIVIS

In the rapakivis there also occur rather small, fine- and small-grained, roundish granite inclusions, which are regarded as magmatic inclusions.

In the Ahvenisto area, there are to be found here and there fine-grained, granitic, to some extent aplitic, inclusions measuring 1—30 cm in length, which always occur sharply bounded by rapakivi-granite. These inclusions are usually ellipsoid or shaped like drops of moisture, with one end sharper than the other. They are generally in a vertical position (the dip of the lineation: 60°—90°), and their sharp end is often pointed downward and the blunt end upward. This phenomenon indicates an upward motion during crystallization or thereafter. The inclusions often contain sparse, large (between 1 and 3 cm long) potash feldspar phenocrysts, which in many cases have a vague, ragged marginal zone. This marginal zone, which contains other minerals as inclusions, is often sharply bounded by the middle portion of crystal. Quartz also occurs as phenocrysts measuring 1—4 mm, some of them idiomorphic, with other minerals sometimes appearing as inclusions in the marginal areas. In thin section, plagioclase crystals somewhat larger than the basic mass are to be seen, now and then. These plagioclase grains likewise have a ragged marginal zone containing inclusions and sharply bordering on an idiomorphic crystal in the middle. Biotite is present in the inclusions, not only as small, independent crystals in the main mass but also as large clusters of crystals. The main mass of the inclusions is distinctly micrographic in fabric. The pigment of the potash feldspar in the main mass seems to grow more intense on proceeding from the margin of the inclusion toward its center. Furthermore, the plagioclase is more sericitized at the center of the inclusions than at the edges, while the biotite is more chloritized and less abundant (SAVOLAHTI 1956, 1956 a). In some instances, it is to be observed that the quartz grains of the inclusions have an even extinction, whereas the quartz components of the country rock have a somewhat undulating extinction. The afore-described inclusions, which occur in all the rapakivis, are generally regarded as magmatic segregations deriving from the rapakivi magma. The mode of occurrence described in the foregoing indicates that for some reason the inclusions had crystallized rapidly (or a large number of crystal nuclei had evolved locally) at quite an early stage and that these crystal clusters had taken on the shape of moisture drops in response to movements. Later, part of the crystals continued to grow and metamorphosed as the magma around them solidified.

Granitic inclusions are also present in the hornblende rapakivis of the Ahvenisto area. In form and structure, they are like those in biotite rapakivi-granites. In addition, they have the interesting feature that, in occurring in the green marginal varieties of hornblende rapakivi, they are green in color and are likely also to contain fayalite (Savolahti 1956 a).

Besides the inclusions considered in the foregoing, the rapakivi also contains other kinds of inclusions, especially in the marginal portions of the intrusive; evidently, they are fragments of the country rock, which must be scrupulously kept apart from the others. Some of these fragments are angular and some rounded. In many cases, they are rimmed by a fine-grained band 1—4 cm wide, poor in mafic minerals and grainy of texture.

f) THE OVOID STRUCTURE AND THE SHELLS ENVELOPING THE OVOIDS

According to Wahl (1925), the Viipuri rapakivi area has the greatest abundance of so-called typical rapakivi, *i.e.*, the type of rock in which the potash feldspar grains are ovoid in shape and enveloped by a shell of oligoclase. Such rapakivis are also quite common in the area of the Aland Islands, but there the potash feldspar ovoids, in their plagioclase shells, are appreciably smaller. In the other rapakivi areas of Finland, on the other hand, potash feldspar ovoids enveloped in oligoclase shells are met with seldom. It is noteworthy that the so-called typical rapakivi occurs in greater abundance wherever there is a large rapakivi massif and *vice versa*. It raises the question: has this correspondence the same causal relation as grain size bears to the size of a massif (see p. 36)?

Sederholm (1891 and 1928), Popoff (1928) and Wahl (1925) have extensively described potash feldspar ovoids enveloped by plagioclase shells. Wahl (1925) applied to the »typical» rapakivi the term »viborgite» and called the other main type of the rock »pyterlite.» Sederholm (1928) did not endorse these terms, but because of their brevity I shall use them here. Typical pyterlite is lacking in the potash feldspar ovoids enveloped in ologioclase shells; but it should be emphasized that between these two types there exists no sharp line of demarcation, for there is a gradual decrase in the amount of oligoclase-mantled potash feldspar ovoids (see p. 39). Furthermore, the degree of roundness of the potash feldspar grains varies, especially in pyterlite. There are types in which some of the potash feldspar grains are angular and some ovoid. There are other types in which round potash feldspar grains are not present appreciably. Moreover, in both types some principal mineral constituent — usually potash feldspar — may to a certain extent occur as grains of extraordinary size, giving the rock a porphyric appearance.

The potash feldspar ovoids in viborgite are generally ellipsoid in shape, but they are not often ideal, for at some point in the ovoid the plane running parallel to the crystal surface tends to be visible. In certain cases, the ovoid is likely to have become concave by shaping itself according to the contours of its neighbor. The ovoids vary in size: in the Viipuri area (WAHL 1925), their diameter is usually 3—4 cm, but a diameter of 6—8 cm often occurs, and the largest met with there — a real rarity — was no less than 27 cm. Potash feldspar ovoids need not represent a single potash feldspar grain but may also be twinned (usually, according to Karlsbad's law), or they may represent a potash feldspar group or a fine-grained rock having the mineral composition of granite or aplite (see p. 78). The potash feldspar of the ovoids may in some cases be quite rich in perthite (perthite ball).

Most of the ovoids occur enveloped in a plagioclase shell — but not all of them. The quantity of ovoids in shells varies. In some instances, large balls have a shell and small ones do not (WAHL 1925). The thickness of the plagioclase shell varies from ovoid to ovoid, and there are ovoids that are only partly enclosed by a shell. According to SEDERHOLM (1928), the thickness of plagioclase shells ranges from 2 to 5 mm. The shell may consist of just one plagioclase grain, but it may also be made up of several small plagioclase grains. WAHL (1925) has counted 60 grains in the cross-section of a single shell. Furthermore, around an ovoid there may be numerous concentric plagioclase rings, situated one after another at short intervals.

Series of concentric inclusions occur in the potash feldspar ovoids as such. The inclusions are usually quite tiny biotite and/or hornblende crystals and/or round quartz grains. There may be plagioclase grains present, too, in the form of slightly larger inclusions. In addition, so-called *aussenkonkave* quartz occurs within the ovoids as inclusions. According to POPOFF (1928), the biotite, plagioclase and quartz inclusions in the potash feldspar ovoids of rapakivis are arranged in conformity to definite rules. FRASL (1954) and SCHERMERHORN (1956), among others, have observed similar features in granite, and their explanation of the zoning and the orderly arrangement is that they are products of crystallization from the magma. SEDERHOLM

(1891 and 1928) writes: »The cause of this egg- or, perhaps better expressed, ovoidal shape of the phenocrysts seems very mysterious. Every thought of resorption in so great a measure and with such regularity appears a priori untenable. Moreover, neither the outer ring, nor the inner rings ever show any cavities which might be ascribed to corrosion, but have always regular shapes. The ovoidal shape seems thus to have existed during all the time of the growth of the crystal.»

Wahl (1925) reports that occasionally dark ovoids enclosed in a plagioclase shell are met with. He also states that idiomorphic quartz grains in pyterlite have been observed to envelop potash feldspar ovoids. The present writer has seen potash feldspar shells around plagioclase grains near Sääksjärvi in the commune of Iitti. This rapakivi is fine-grained and, in addition to the afore-mentioned phenocryst, it contains potash feldspar ovoids 2 cm in diameter, which are enclosed in an oligoclase shell, as well as quartz phenocrysts. At Ratula, Artjärvi, there occurs a hornblende- and epidote-bearing rapakivi which is very poor in quartz. In this rapakivi I have seen hornblende rings around potash feldspar ovoids. In the rapakivi area of Iitti—Jaala, plagioclase shells are met with enclosing both quartz and potash feldspar grains. There I have also run across potash feldspar rings around both biotite and plagioclase grains.

g) THE CRYSTALLIZATION ORDER OF MINERALS IN RAPAKIVIS

The oligoclase-mantled potash feldspar ovoids in rapakivis, the order of idiomorphism of the minerals and, on the other hand, the so-called normal order of crystallization of the minerals have appeared to be in mutual contradiction. In his by now classical work on rapakivis, Sederholm (1891) writes that it is hard for him to accept the fact that many of the known rapakivis have quite a different order of crystallization from other granites, but that only thereby can the remarkable contact relations of such different minerals be rationally explained. Later, Sederholm (1916, 1928) came out with a number of fresh observations regarding the odd crystallization order of rapakivi minerals. *E.g.*, he describes biotite and quartz intergrowths crystallized at a late stage. But these observations did not attract sufficient attention, and Sederholm himself (1928) remarked: »The writer has, however, no opportunity of going very deeply into this subject at present (meaning the subject of rapakivis), but must restrict himself to some suggestions in general discussion of the explanations which have been hitherto given of the peculiar texture of the rapakivi». This indicates the strength of the position held by Rosenbusch's (1882) rules governing idiomorphism and crystallization order (see p. 94).

WAHL (1925) describes and seems to stress the occurence of the principal minerals of all the rapakivis in two different generations. The present author suspects this to be due to the fact that thereby it is apparently possible to avoid a conflict with the rules governing the order of crystallization. This presumption is based on the conviction that such a brill'ant observer as Wahl could not have failed to notice the discrepancy himself.

An investigation of the rapakivi area of Ahvenisto has brought to light considerable new data at odds with the normal order of crystallization (SAVOLAHTI 1956, 1956 a). In this area there are quartz porphyries, for instance, with a chemical composition corresponding, within a range of variation, to that of rapakivis. The crystallization of these rocks had abruptly been interrupted, and the phenocrysts contained in them represent the course of crystallization at the time it was interrupted. Then, there are in the same area so-called biotite rapakivis which are igneous rocks and which were formed when the magma solidified to completion under peaceful conditions. Their mineral composition thus represents the composition of a rock formed out of rapakivi magma without differentiation. Moreover, the area has so-called porphyry aplites, which have been ascertained, on structural-geological grounds, to be a late differentiate of biotite rapakivi magma. The separation could be judged to have taken place after the biotite rapakivi magma had erupted to its present location and the main portion of it had solidified into biotite rapakivi. Knowing the mineral components of these three rocks, it is possible for us to follow in each rock, on the basis of the variations in the content of any one mineral and the proportionate content of the different minerals, the course of the crystallization in the rapakivi magma. It has more recently come to my attention that MOORHOUSE (1956) has used a somewhat similar method in observing the crystallization of apatite in igneous rocks. I am prepared to take the tentative stand that this kind of method, in its various forms, is much more useful than merely noting the degree of idiomorphism — assuming that the aim is to follow the course of differentiation in the magma. Table 8 gives the measurement results. It will be seen that the mafic minerals and the plagioclase have become enriched into the porphyry aplite, or the final differentiate (there are only very small amounts of aplites and pegmatites, so they need not be taken into account, as in the case of rapakivis in general). On the other hand, they show that potash feldspar and quartz at first crystallized out of the magma in greater amounts, relatively, than the biotite and plagioclase did. This is revealed by a comparison of the mineral composition of phenocrysts in the granite-porphyries with the mineral composition of other types.

On the other hand, the same differentiation tendency is also evident in the chemical composition of the afore-mentioned rocks of the Ahvenisto

Table 8. Mineral composition of certain rapakivi types of the Ahvenisto area, per cent by volume.

	1	2	3	4
Potash feldspar	26	60	56.4	47
Quartz	13	29	34.9	26
Plagioclase	1	7	6.2	19
Mafic minerals	—	4.5	2.5	7.5
Groundmass	60	—	—	—
Plagioclase perthite in potash feldspar	—	21.5	23.5	9
Ratio potash feldspar: plagioclase	26	8.6	9.1	2.5
Ratio quartz: plagioclase	13	4.1	5.6	1.4

1. Granite porphyry (phenocrysts + groundmass; average),
2. biotite rapakivi (average),
3. biotite rapakivi and
4. porphyry aplite (average).

area when the values calculated from them are placed in a normative triangular diagram (albite + orthoclase + quartz), (Bowen and Tuttle, see Adams 1952). The chemical composition, to be sure, does not vary nearly so much as the mineral composition. This is but natural, for during the course of crystallization the composition of the minerals undergoing crystallization changes. This phenomenon prevents to some extent any rapid change in chemical composition. *E.g.*, the potash feldspar crystallizing at the beginning of the process is richer in albite and anorthite than the product of later crystallization. The magma cannot undergo impoverishment of its Na and Ca constituents so rapidly as it would if the potash feldspar did not contain them in the first place in such abundance. But, on the other hand, the composition points of all the rapakivis (see Eskola 1961) fall in such a way that in the triangular diagram the eutetic field referred to remains between these composition points and the tip of the plagioclase angle. This alone justifies concluding that under conditions of equilibrium the plagioclase cannot begin to crystallize first from a magma with a rapakivi composition.

In the modal composition triangle (feldspar (F), quartz (Q) and mafic minerals (M) presented by Chayes (1952), the modal composition points of the rapakivis likewise fall in such a way that the maximum field of the granite composition lies between the M-angle and the composition points of the rapakivis.

A great difficulty in understanding the crystallization of rapakivis was that, according to Bowen and Tuttle (1950), in the system $NaAlSi_3O_8$—

$KAlSi_3O_8$—H_2O only one feldspar crystallizes while two different feldspars did not evolve until the subsolidus area was reached, nor did the pressure cause the crystallization temperature to fall that far, either. Thus, two feldspars could not have simultaneously crystallized in rapakivis. The discrepancy had to be only apparent rather than real, considering that in quartz and granite porphyries there may occur potash feldspar and plagioclase phenocrysts simultaneously apart from each other in the same rock. YODER, STEWART and SMITH (1957) report that under an H_2O pressure of 5 000 bars, in the case of certain compositions, two feldspars crystallize directly. This observation is supported by BOWEN and TUTTLE (1958). SAVOLAHTI (1958) noted that the An-content of the plagioclase of rapakivis is large compared to the An-content of granites and surmised that such plagioclases have a greater tendency to crytsallize than the ones rich in Ab. In addition, he remarked that, as a matter of principle, the exsolution and homogenization phenomena of two isomorphic minerals revealed nothing — particularly when a third component was present — about the conditions under which these minerals crystallize as a single homogeneous and as two separate minerals, inasmuch as unmixing and crystallizing are governed be different laws. It may also be asked whether possibly experimental conditions, such as capillary walls, could have a selective effect in themselves toward promoting, at least in some cases, the formation of embryonic crystals and what kind of an effect it might be.

An import contribution to the question of two feldspars is made by an experimental study carried out by WYLLIE and TUTTLE (1961) on a system involving two volatile components. According to this study, under an H_2O-pressure of 2 750 bars, granite begins to melt at a temperature of 670° C. When the system under this pressure includes the additional factor of 4.3 weight per cent HF, the melting of granite starts already at 605°; and when the HF amounts to 8 % by weight, the melting correspondingly starts at 595° C. It is noteworthy that the additional HF at first brings down the crystallization temperature quite steeply, but that further additions cause relatively little reduction. The rapakivis, as is well known, contain abundant fluorine compared to many granites, so that the probability of two feldspars' crystallizing is greater in the case of rapakivis than in that of other granites. On the other hand, WYLLIE and TUTTLE (1959) have investigated the effect of carbon dioxide on the melting of granite and feldspar. They have noted that carbon dioxide reduces their melting temperature much less than H_2O does. However, a greater abundance of carbonate is met with on the average in alkali rocks than in granites. (See, *e.g.*, PECORA (1956) and his bibliography). Is it for this reason that examples are often to be found in alkali rocks of their having crystallized in a one-feldspar field.

The crystallization of rapakivi magmas starts with the crystallization of potash feldspar and/or quartz. I have recently made a number of new interesting petrographic observations that lend support to this conception. There is reason to describe them in this connection.

The collections of the Institute of Geology of the University of Helsinki include a quartz-porphyry sample from Suursaari which contains blackish red feldspar, blackish gray quartz and greenish plagioclase phenocrysts. But, in addition, it contains potash feldspar grains enclosed in plagioclase shells 2—3 mm thick. The shells must have solidified later than the middle portions. Moreover, the shells cannot be the product of unmixing or some other autometamorphic phenomenon, since the rock has a very dense mass (*cf.* NEUVONEN 1957 and p. 78); for if the shell had been an exsolution product, the main mass would have solidified at the same time with a coarser grain.

The fine-grained porphyritic rapakivi of the paving-stone quarry owned by the city of Lahti near Sääksjärvi, in the commune of Iitti, a rock classified by the present writer among porphyry aplites, has a sparse content of potash feldspar phenocrysts. These phenocrysts measure about 2 cm in diameter, some of them are beautiful ovoids, and they are enclosed in plagioclase shells 2—3 mm thick. It is noteworthy that the grain size of the rock itself is approximately 1 mm. These thinly and irregularly scattered phenocrysts can hardly be of autometamorphic origin, but are intratelluric; nor can they be considered exsolution products, either.

The writer (SAVOLAHTI 1956, 1956 a) observed that porphyry aplites had not been met with in the area of Suomenniemi, and that this was due to the thinness of the massif, which caused an unusually rapid cooling. Subsequently, however, there has come to light in the middle of the Suomenniemi massif one narrow, nearly horizontal porphyry aplite layer. This porphyry aplite is peculiar. Its general mass appears almost ophitic in microscopic texture, like diabase. This is because the rock is very rich in plagioclase (oligoclase), and the plagioclase laths in it are situated crosswise (ophitically). It is natural that this rock should be rich in plagioclase, for the crystallization of the small massif had proceeded far before the injection of the residual magma; and the differentiation of the magma has, of course, likewise been carried far.

Another interesting example is provided by the porphyry aplites, which are late differentiates of the rapakivi magma. The biotite of a certain porphyry aplite in the Ahvenisto area seems to have crystallized at a fairly late stage, though the rock itself is a late differentiate. It forms large poikilitic grains, in which the pale minerals are inclusions (Plate VI, Fig. 1). The biotite appears to be a late product of crystallization, even at this stage. In this connection, it is pertinent to remark that large biotite grains of this type have apparently been regarded as resorbed phenocrysts in many rapakivi types where it is not possible for them to be such (see also p. 86).

In the Suomenniemi area, on the right side of the Savitaipale—Mäntyharju road, near the provincial border, there is a quartz porphyry dike, in the cleavages of which I have seen biotite crystallized as thin veins (1 mm and less). The same exposure also contained biotite crystallized around quartz grains. All these things suggest late crystallization. Are both cases to be regarded as belonging to the same crystallization phase or not? The biotite in the cleavages, at any rate, represents a late crystallization. It need not, perhaps, belong to a magmatic crystallization phase, for it could also belong to a phase following the principal magmatic crystallization. Previously, I had seen in cleavages of the granite-porphyry dikes of the Ahvenisto area some quartz veins a few millimeters broad.

Fig. 2 (Plate VII) is of the hornblende rapakivi of the Jaala—Iitti area. LONKA (1957) writes of it as follows: »The quartz grain is roundish, but its distinctly angular contour suggests an originally idiomorphic quartz grain. The dark ring around it contains hornblende in the main, but the constituents include biotite as well as pyroxene ($c \wedge \gamma = 46°$). The case in question points to crystallization order. A portion of the quartz must have crystallized before the mafic minerals did, as is true of the rapakivis (SAVOLAHTI 1956).» Within the dark ring there need not invariably be a uniform quartz grain, for there may also be a cluster of quartz grains. SEDERHOLM (1916) has described biotite as a filling of miarolitic cavities, in conjunction with fluorite, in the rapakivi of the Aland Islands. It may conceivably be argued in some quarters that also the structures described in the foregoing are likely to represent fillings of miarolitic cavities and that the dark ring had crystallized first, after which it had been filled with quartz. Perhaps in some cases the crystallization had taken place in this way, but not in all cases — not, at least, where such a ring is around a hexagonal bipyramidal quartz grain. Nor need all the dark rings and the clusters of quartz grains within them represent fillings of miarolitic cavities, for the quartz porphyries also include quartz phenocrysts that have evolved from quartz clusters. But, on the other hand, the crystallization of the dark minerals must have taken place much later if they crystallized on the walls of the miarolitic cavities. That the quartz would have thereafter filled the centers of the the cavities is also natural, for at such a late stage of crystallization the magma must surely possess a so-called eutectic composition, which means that quartz, too, is certain to crystallize abundantly. In certain cases, autometamorphic and autometasomatic processes are likely also to contribute to the genesis of formations of this kind (*cf.* p. 88).

6. CIRCUMSTANCES PROVIDING EVIDENCE OF THE MAGMATIC ORIGIN OF RAPAKIVIS

ECKERMANN (1937)[1]) deserves to be mentioned here first, having specifically sought proof of the magmatic origin of rapakivi, though, to be sure, most of the evidence collected by him was already known to SEDERHOLM (1891). 1) Known is the grading over of rapakivis into rocks with a hypabyssal fabric — and even farther (see pp. 38—48). 2) Known are the common occurrence of miarolitic cavities and their wholly unordered and unoriented fabric. 3) At Tyysterniemi, in the rapakivi area of Viipuri, near Lappeenranta (HACKMAN 1934), the author has been shown a streaked variety of rapakivi near a contact. The streaks seem to bear witness to a magmatic origin. Calcite and fluorite had also become enriched into the same rock. These minerals appeared to occur mainly as cavity fillings. 4) Likewise, the rapakivis form beautiful eruptive breccias, in which fragments of country rock are situated arbitrarily. 5) The contacts of the rapakivis are sharp and the rapakivi lacks the relict features of the earlier rock from which it has possibly evolved through »granitization.» 6) The regional metamorphism occurring in the areas where rapakivi occurs appears to be independent of the rapakivi as regards time and place (SAVOLAHTI 1956, 1956 a). 7) The contact metamorphism induced by rapakivi is slight, but it involves the formation of minerals, such as pyroxene and even pigeonite (see p. 48). Since contact zones of this kind are relatively narrow and their color does not differ sharply from that of the country rock, not many of them are known. However, the present author has seen one, for instance, in the rapakivi area of Laitila, near the parish church (see also TRÜSTEDT 1907). The fact that the contact metamorphosis has been slight, again, need not have been due to a low crystallization temperature but to the scantiness of the volatiles. At the same time, it demonstrates how small are the migration of materials and the diffusion of atoms into solid rock in the proximity of even a large mass of magma undergoing peaceful crystallization (*cf.* also ESKOLA 1961 p. 121). 8) The way the rapakivis intersect the basic rocks in the Ahvenisto area also indicates their magmatic origin. Other magmatic features can also be pointed out. 9) The size of the grains composing rapakivis seems to depend on the size of the crystallized massif. 10) The properties of the main minerals found in rapakivis — potash feldspar, quartz, plagioclase and biotite — indicate that they had crystallized at a fairly high temperature.

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[1]) Reading for the first time the Jotnian and sub-Jotnian literature of Sweden, the author gained the impression that, particularly as regards basic rocks, many matters had been treated earlier and better there than in the corresponding literature of Finland. In this report, however, the author has chosen his citations principally in the light of his personal experience.

Numerous circumstances providing evidence of the magmatic origin of rapakivis have been listed here, without forgetting additional ones. It must, however, be admitted that not nearly all of them are altogether unambiguous, but in combination they can hardly admit a different interpretation from that offered in this report. It thus frees us from taking into account the results arrived at in studies devoted to the phenomenon of »granitization,» which have brought to light structures like those exhibited by rapakivis (see also p. 81).

7. EXCERPTS FROM THEORIES CONCERNING THE ORIGIN OF RAPAKIVIS

The remarkable appearance of rapakivis, with their potash feldspar ovoids enclosed in oligoclase shells, has excited the interest of many researchers. Any number of geologists from abroad who have visited Finland and seen samples of rapakivi — or who have otherwise become acquainted with the rock — have become fascinated and indulged in theorizing about the formation of the potash feldspar ovoids and the oligoclase shells in which they are enveloped.

It was J. J. SEDERHOLM (1891) who coined the term »rapakivi,» which has become current in all civilized languages, in his classical and still quite serviceable work on the subject. This volume includes descriptions of all the rapakivi areas known at the time and of the various types of rock met with in them, together with the structures. The point of departure rests on a basis of sufficient evidence — as it would seem even when read in this day and age — that the rapakivis crystallized from a magma and that they form a series in which several of its members completely correspond to volcanic types of the granite class in respect to structural formation and geological mode of occurrence, while, on the other hand, they approach plutonic rocks of the same class in fabric without wholly matching the latter type. In SEDERHOLM's view, the ovoid structure evolved during the course of the entire growth period of the crystals and is not a product of corrosion caused by the magma. The reason why the grains are not flat-surfaced crystals but round ones is, as he sees it, that their impure composition impeded their idiomorphic growth. VOGT (1906) and HAKER (1909) thought that such a structure is produced upon the rapakivi magma's crystallizing in the eutectic field of the granite magma while the states of super- and subsaturation varied. In 1928 SEDERHOLM explained anew his conception of the genesis of the oligoclase shells enclosing the potash feldspar ovoids, basing it on the great viscosity of rapakivi magma, which naturally retards diffusion. This would account for the rounded form of the potash feldspar grains and for the production of local enrichments of sodium and calcium around the ovoids, whereupon oligoclase is enabled to crystallize as a shell around the potash feldspar. The Ahvenisto area has yielded evidence (SAVOLAHTI 1956, 1956 a) that the older rapakivi types in the same differentiation series are richer in potash feldspar and quartz than are the younger ones. In other words, the potash feldspar and the quartz separated at an early stage as the magma crystallized. Moreover, the composition point of the principal rapakivi types falls outside the eutectic field of the granite magma, and it does not fall on the normal extension of the crystallization course of the basaltic magma but in between the eutectic field and the potash feldspar, whereupon the eutectic field cuts the crystallization course. From this it naturally follows that the crystallization of the rapakivi magma begins with the formation of potash feldspar and quartz. It is later on that the oligoclase shells crystallize around the potash feldspar. TUTTLE and BOWEN (1958) accept this latter conception and add certain reflections of their own based on experimental investigations, particularly as bearing on the crystallization course. According to GATES (1953), the potash feldspar crystallizes in granites before the sodic feldspar. The potash feldspar contains, as a surplus, sodic feldspar matter, which unmixes from the former when the massif cools and then recrystallizes. As this researcher reasons it, the plagioclase-mantled potash feldspar cores of rapakivi appear to be a »unique variety of perthite» formed under conditions that had been favorable to the thorough penetration of the cores by the plagioclase material. In the light of the material presented, it would seem possible that such unmixing may take place in certain instances. HUTCHINSON (1956) reaches pretty much the same conclusion as GATES in an extensive and interesting report, which has convinced the present author that it holds water. NEUVONEN (1957) has expounded the structure of viborgite according to the same principle. This principle does not, however, provide an explanation to the structure of viborgite, for the proportionate amounts of potash feldspar ovoids with and without plagioclase shells vary from place to place, the changes taking place gradually. Why should there be a ring around one potash feldspar ovoid and none around its neighbor? For another thing, the albite- and anorthite-contents of mantled and unmantled potash feldspar grains do not vary within as broad limits as they should if the ring had become detached from the one and not from the other. In the third place, the great divergency among the ovoids indicates, at least, that the mantle had not derived from them (p. 69). In the fourth place, it has been demonstrated that, in certain instances, the shell had enveloped the potash feldspar also by crystallizing directly from the magma (pp. 72). According to HOLMQUIST (1939), the mantled potash feldspar grains within the same rapakivi are richer in plagioclase components than the grains lacking a mantle. This result is in agreement with WAHL's (1925) results. If such is the case, how, then, can one understand the unmixing and the migration of the sodic feldspar out of the grains?

WEGMANN (1938) describes a granite from Helene Harbour in South Greenland that near the contact with its country rock undergoes a structural transition so as to resemble rapakivi to a remarkable degree. The general mass of the rock becomes finer of grain, the quartz grains idiomorphic, the potash feldspar partially ovoid and the biotite darker. He compares the marginal portion to the rapakivi of western Finland or the Aland Islands and the central portion to the granites of the Onas-Obbnäs type. The present author has seen quite identical phenomena in connection with numerous porphyritic granites in central Finland. And there is reason to believe that such a process does take place, but the author is inclined to believe that the formation of this structure is somewhat different from that of the so-called typical rapakivi structure, although a structure comparable to it does evolve also in the manner described in the foregoing.

The genesis of the rapakivi structure otherwise than magmatically has been propounded by, among others, BACKLUND (1937) and PRUCKA (1957). It is quite possible that structures resembling rapakivi may form to some extent through metasomatic and granitizing processes. The autometamorphic and autometasomatic phenomena associated with the formation of rapakivi structures at least make it a likelihood (see pp. 85—92).

In addition to the names already mentioned, the genesis of rapakivis has been studied by HOLMQUIST (1901), POPOFF (1897), WAHL (1925) and WOLFF (1932), as well as others.

8. THE FORMATION OF RAPAKIVIS

The magmas of the rapakivis and of the granites differ, according to the prevailing view, in composition, though it must be admitted that the composition of certain porphyric granites is very nearly that of the rapakivis. SAHAMA (1945) has written: »When compared with other granitic rocks the rapakivi granites — as . . . previously known — contain more SiO_2 and K_2O, and less MgO, CaO as well as Al_2O_3.» NIGGLI (1923) applies to such a magma the term »rapakivitic magma type,» although the type described by him corresponds more nearly to the so-called tirilites (HACKMAN 1934). SAHAMA (1945) has calculated the mean of the analyses made of the rapakivis of eastern Fennoscandia. Rare types occurring in small amounts affect this mean disproportionately. SAHAMA (1945) has also prepared a so-called standard mixture of rapakivis and had it analyzed. The differences in the size of the various rapakivi areas has been taken into account, but the effect of types occurring as small segregations in some areas has not been deducted. This analysis of a standard mixture nevertheless shows clearly that F and Zr are present in rapakivis abundantly compared to granites. It is important to notice that the composition point of rapakivi magma in composition triangles falls outside the eutectic field on the side of the quartz and potash feldspar. Accordingly, the eutectic field will separate the composition points of the rapakivis from both the plagioclase and the mafic minerals. On the other hand, the differentiation course cannot travel through the eutectic field of the differentiating magma. It is not possible, at least, in known experimental cases, nor does such an assertion appear to be well taken from the theoretical standpoint, either.

The rapakivis erupted in connection with block movements of the earth's crust (SEDERHOLM 1891 and RAMSAY 1890) and are situated, according to a general view, in areas where block movements had taken place and where there are shear zones. Such zones of weakness in the earth's crust are permanent — they will remain throughout the ages. The grand old man of Finnish geology, PENTTI ESKOLA, has a pet saying that wounds in the body of the earth mother never heal. As a consequence, magmas have been able to inject themselves into such zones of weakness and they have erupted repeatedly after the first eruption had succeeded in penetrating the crust. Thus, rapakivis of different ages have been met with in the very same area in Finland. And in conjunction with the rapakivis various basic rocks are also found, some of which are older and some younger than the rapakivis (SAVOLAHTI 1956 a). SEDERHOLM (1928) wrote that it is questionable whether the rapakivis and the basic rocks occurring in association with them have any genetic relationship. But in the same year there appeared a work written by that splendid master of experimental research BOWEN (1928), in which the theory of crystallization differentiation was convincingly set forth. This theory buried the rapakivi concept germinating in the mind of SEDERHOLM (1928).

But gradually new observations were made that supported SEDERHOLM's (1928) hypothetical reasoning that the rapakivis and the basic rocks were not related. ECKERMANN (1938) has studied the late differentiates of the basic rocks of the Jotnian areas in Sweden — which rocks are also to a large extent found in association with rapakivis — and he observed them to be rich in Na. He remarked that sodic residual magma,» occurring as veins and dikes within basic rocks of doleritic composition, are a fairly common feature. Undoubtedly they are in most cases exponents of late differential action within the cooling mother magma and may, consequently, be used to gain additional knowledge of the laws governing the closing stages of magmatic activity». ESKOLA (1948) points out that the late differentiates of the olivine diabase of Satakunta found in association with the rapakivis of Laitila are rich in Na and not in K, as ought to be the case. SAVOLAHTI (1956 a) has demonstrated that the final crystallization of the gabbro-anorthosites of Ahvenisto are also Na-rich and that there had been a distinct time interval between the eruptions of these rocks and the rapakivis, that some of the basic rocks are younger than the main portion of the rapakivis and that the residual part of the rapakivi magma had become impoverished in its content of quartz and potash feldspar in undergoing differentiation.

In addition to field observations, also experimental studies show that magma containing more potassium and silicic acid than is present in the eutectic composition of granitic magma cannot be produced through crystallization differentiation out of the parent basaltic magma. The impressive volume by TUTTLE and BOWEN (1958) includes a clear report on these research projects. The argument advanced here does not, of course, deny crystallization differentiation in general, for it also influences the differentiation of rapakivis, though in a different direction and not toward or proceeding from the basaltic magma. The course of differentiation of those porphyritic and microcline granites in which there is a potassium and quartz content exceeding the eutectic composition might be the same (ESKOLA 1956).

The genesis of such exessively potassium-rich granites, or so-called ideal granites, has been explained by means of »postmagmatic potassium metasomatosis» (ESKOLA 1956). This hypothesis is based on the studies of SZÁDECZKY-KARDOSS (1955) on »trans-vaporization» — studies of which he has a number of others, too, bearing on related questions. The observations made of the rapakivis indicate that postmagmatic potassium metasomatosis is not capable of explaining the composition, texture and formation of the rapakivis (see also p. 82). The pyterlite types are richer in K than the viborgite types (WAHL 1925), and, together with porphyritic rapakivi types, they represent, at least in some localities, marginal varieties of viborgite. The green marginal variety of hornblende-rapakivi is richer in potash feldspar than the hornblende-rapakivi itself (see pp. 42). The potassium is also enriched into the country rocks of the rapakivis during crystallization (see pp. 49), although in relatively small amounts. A classical example of potassium metasomatosis is the rapakivi contact on the island of Bucholm described by SEDERHOLM (1923). To be sure, the metasomatic effect appears to be more extensive there than it really is, owing to the fact that the contact is nearly parallel to the surface of the rock (SAVOLAHTI 1951). These observations, chosen to serve as examples, enable us to decide that, if and when potassium has migrated metasomatically during the crystallization of the rapakivi magma, its migration was in the direction of the margins of the rapakivi massif and also to a certain extent into the country rock. No examples of a contrarywise migration are known in connection with rapakivis. Nor can it be imagined that the potassium had come »in ewigen Teufen,» when the rapakivi masses had been limited in depth and had, in part, occurred as fairly thin layers. Where could the potassium have come from, when the potassium contained in the already stationary rapakivi magma mass had been streaming outward and, on account of the early crystallization of the potash feldspar, it had otherwise, too, been deprived of potassium the whole time? There exist examples of metasomatosis of potassium to the marginal parts of massifs and to their country rocks. To this the reader may still protest that the postmagmatic metasomatosis of potassium, or the arrival of potassium, could have taken place in the lower part of the rapakivi magma mass in the direction of the magma. The country rock may have been rich in potassium and/or the physico-chemical conditions had been such that the diffusion of potassium had taken place in such a fashion. This could, indeed, have happened in certain instances. But there is no reason to suppose that there had been potassium-rich rocks everywhere underneath the rapakivi masses, since there are no observations to indicate this. The present author, at least, has no knowledge of examples in connection with rapakivis that would show that K-diffusion had taken place in the direction of the rapakivi massif even in its basal contacts, although he does know of gently inclined contacts where the rapakivi is on top. If, in spite of everything, potassium diffusion had taken place in certain instances towards the rapakivi in the basal contacts on account of the somewhat different physico-chemical conditions prevailing there, it can scarcely be conceived that this diffusion could compensate for that taking place in every other direction, let alone that it could have resulted in an excess many times over, as would be inescapable along this road. In general, knowing the facts, it is difficult to imagine where the tremendous quantity of potassium could come from which would be needed to make some large rapakivi massif sufficiently rich in potassium if its formation had to be explained by means of postmagmatic potassium metasomatosis.

How the rapakivi magma had evolved deep down in the earth before its eruption is not an easy question to answer. The rapakivi area of Viipuri, which is at least 12 000 km^2, is a colossal massif. Rapakivis are known, according to Wahl (1925), round the world, *e.g.*, in Sweden, Norway, Russia, Siberia and Korea. The magma eruptions had been tremendous, having taken place in conjunction with block movements, faultings, or in the same way as in the case of basalts. Are the foregoing, too, primary magmas? The mode of origin of the rapakivi magmas can be conjectured only hypothetically. They could be, *e.g.*, so-called parent magmas, magmas produced through anatexis, magmas segregated in a liquid state, magmas produced through potassium metasomatosis and the like, or magmas that had evolved in some other, still unknown way.

As for their chemical composition, rapakivis contain fluorine in greater abundance than other granites. On the other hand, many rapakivi students are of the opinion that the rapakivi magma had been a relatively dry and viscous magma. If such had been the case, we may ask whether it had been dry and viscous in the first place or whether volatiles had evaporated from it after its injection into a high horizon. Whatever the answer to the first question might be, we can, at any rate, answer the second affirmatively. For example, in the country rock of the rapakivi along the northern margin of the Ahvenisto massif, I have run across fluorspar veins of a violet color following the cleavages. The contacts of these veins, to be sure, are not always sharp, for the fluorspar has impregnated the country rock, giving it, too, a violet tinge. Since fluorspar segregations are also met with at the core of the rapakivi massif in conjunction with the rapakivi pegmatites (Savolahti 1956 a), we cannot therefore be sure at what stage of the magma crystallization the first-mentioned veins had appeared. But at Tyysterniemi, in the northern part of the Viipuri rapakivi area, the present writer has been shown how, *e.g.*, fluorspar and calcite have become enriched into the fine-grained (induced by cooling) and evidently magmatically banded marginal zone of the rapakivi. Fluorspar and calcite appear in many cases to occur in this locality as if they were filling material in the miarolitic cavities. In the present instance, F and CO_2 had probably become concentrated at a relatively early stage, for otherwise they would have become enriched into that part of the intrusion where the rapakivi begins to turn fine-grained.

If the rapakivi magma had been relatively dry during its crystallization, it must have also been fairly viscous. From these properties it must follow that the rate of diffusion of this magma must also have been fairly low.

Contact zones of the rapakivis are described on p. 38. The descriptions make it clear that the rapakivis are in many instances finer of grain in the immediate proximity of their contacts than in the massif as a whole and that the change in the grain size is a gradual one. See also Jaeger (1957). This change in grain size illustrates the temperature gradient during the course of crystallization in that particular section of the rapakivi massif. In the cases where no appreciable change in the grain size can be perceived, there could not, of course, have been any noteworthy temperature gradient in the rapakivi massif during crystallization, but a nearly even temperature prevailed throughout the entire massif. At that time, the temperature gradient had apparently been on the country rock side. Its shifts there is naturally brought about by the limited heat conductivity of the country rock in relation to the amount of heat to be conducted. The larger the massif, the less will the dissipation of heat be felt. A low temperature gradient, or its absence altogether, would indicate, again, that cooling had been quite slow and the crystallization period very long (see p. 38). The cooling process is further retarded by the great volume of heat dissipated when the minerals crystallize. The amount of heat involved is so enormous that under these conditions it certainly cannot escape on a sufficient scale and at a sufficiently rapid rate. Accordingly, it has the effect of greatly retarding the crystallization process, sometimes perhaps even interrupting it. This is a logical conclusion, since the amount of heat escaping during crystallization is — it can be said — on quite a different order of magnitude than that escaping during the cooling stage.

As the potash feldspar and the quartz crystallize, the composition point of the rapakivi magma migrates in the compostion triangle Ab—Or—SiO_2 close to some temperature isobar toward the eutectic field, where the main part of the crystallization takes place. Thus, the temperature interval in which the principal crystallization of the magma having a rapakivi composition takes place is very narrow. The temperature interval is narrow even when taking into account the changes occurring during crystallization in the concentration of the volatiles contained in the magma, changes which affect the pressure and thereby the crystallization temperature (TUTTLE and BOWEN 1958). (See also YODER and TILLEY 1957). If and since the tendency of rapakivis to disintegrate is largely due to the high crystallization temperature, it would presuppose that the principal crystallization of the rapakivis had taken place at a temperature above 500°—575° C, after which a considerable change in the volume of the rock is induced by the inevitable decrease in the rock's temperature.

The next question is this: What part of the rapakivi minerals crystallized before the magma settled in place and what part thereafter? It has been emphasized that the rapakivi minerals had crystallized in two generations, namely, euhedral and anhedral (*cf.* p. 91). It is tempting to imagine — and the writer has heard about the idea having been aired — that the minerals of the first, euhedral generation are intratellural and had thus crystallized before the eruption of the magma, and that the minerals of the second generation had crystallized after the eruption. Thus the interruption in the crystallization process brought about by the eruption would separate these generations from each other. This, however, is not the case, for there are many arguments against such a conception.

Here are a few examples. Viborgite and pyterlite, which grade over into each other (see p. 39), are two structurally quite different rocks, yet each contains both mineral generations. It is not reasonable to think that the euhedral generation in both cases is intratellural, for then the structural differences of the rocks would not then be comprehensible. The mineral generations of these rocks are not separated by the event of the eruption.

The marginal varieties of the green hornblende rapakivis of the Ahvenisto area (SAVOLAHTI 1956, 1956 a) contain potash feldspar phenocrysts that are quite as large as in the hornblende rapakivi itself. But the hornblende rapakivi is otherwise coarser of grain than its marginal variety, and these rocks have other structural differences as well. Accordingly, only the potash feldspar phenocrysts and the comparable mineral grains can be intratellural, for the differences in structure between these two rocks developed after the eruption and are due to the fact that one is a contact variety of the other, just as was the case in the foregoing.

The biotite rapakivi of the Ahvenisto area is finer of grain when it occurs as straighter veins in gabbro-anorthosites than otherwise, for it had congealed more rapidly. Both of them, however, have two different mineral generations, so that the euhedral generation cannot be exclusively intratellural, for there is more of it at the core.

The contact varieties of the Suomenniemi rapakivi massif (see p. 39) also support the point of view here presented.

In all these examples, mineral grains of both generations have certainly evolved during the crystallization that took place after the eruption, although the proportional quantity of the generations has varied. For example, a graphic texture invariably occurs to a greater extent in the marginal zones than at the core of the massif. On the other hand, in the intratellural phenocrysts there remain magma tongues and inclusions that had in many cases crystallized into a coarser grain than had the general mass of the rock (*cf.* p. 90). These inclusions and the mineral grains contained in them are principally anhedral, but on account of their coarseness of grain they are likely to have become crystallized at a fairly early stage, perhaps to some extent before the eruption. Thus, both the euhedral and anhedral mineral grains are likely to belong to both generations, although the first of these involves euhedral ones in the main. These relations are liable, furthermore, to be mixed up by the factor of corrosion.

It is now time to present a few features of the formation of plagioclase-mantled potash feldspar ovoids. The results of research into this matter have been considered in a historical sequence in the foregoing, so that this aspect of the subject will no longer be touched upon in this chapter. Inasmuch as the larger the rapakivi massif is, the more one meets with feldspar ovoids and plagioclase shells around them, this correlationship must have some bearing on the genesis of the ovoids and their mantles, and such a correlation does exist. The larger the magma massif in the process of crystallization, the slower does it cool (TURNER and VERHOOGEN 1960), for when minerals crystallize the amount of heat escaping increases much faster as the size of the massif grows than the surface through which the heat is conducted out of the massif, *e.g.*, the total area of the massif. Likewise, the larger the body in process of cooling, the more does the heat flowing out of it raise the temperature of the surroundings, and the lower does the temperature gradient round it become and the less heat generally escapes from it. Thus are brought about especial and deviating, if ideal, crystallization conditions — since it should be borne in mind, furthermore, that quite peaceful conditions had to prevail during the crystallization of rapakivis. The result is that during the principal stage of crystallization of a large massif, the outflow of heat is hardly felt as a factor causing the temperature to fall; this is true at least at the core of the massif. The crystallizing process thus takes places at a fairly high temperature, quite slowly, in conditions of extreme equilibrium. Processes comparable to autometamorphism and autometasomatism may take place and do take place already during the principal stage of crystallization, and thereafter, usually to a greater extent — it may be said in abundance —, for the principal stage of crystallization terminates at a high temperature and the cooling afterwards, too, is slow. This is the basis from which the origin of the potash feldspar ovoids must be examined.

On account of the slow and balanced crystallization, the ovoid shape of the rapakivi feldspar represents in part, perhaps in the main, only a tendency toward maximum symmetry and, at the same time, a minimum content of free energy — which tendency is known in many guises (*e.g.*, ESKOLA, 1946). This may be said because the tendency towards symmetry is manifest also by the roundish growth of the crystal faces, this being a common phenomenon — one example being the diamond. MARSHALL and JOENSUU (1961) hold the view that the crystal habit of lead glance depends on the temperature, its cubic form occurring at the lowest temperatures and the octahedral form at other temperatures.

It might also be imagined that to some extent the slowness of the diffusion promotes roundish growth, for the concentration of atoms needed in growing diminishes more rapidly at the points where the growth of the crystal is rapid, that is, in recesses of the crystal, than elsewhere. Such changes in the concentration tend to cause changes in the rates of crystallization. This may happen even when the crystallization process is slow, for diffusion slows down along with the rate of crystallization. When crystallization takes place in conditions of equilibrium, even a very small difference in concentration has its effect.

The present writer does not want to deny the significance of corrosion in the formation of ovoids, for changes in pressure and temperature may play a special part, but ovoids can certainly form — and do form — otherwise as well. Causes of corrosion exist in rapakivis, but there are less signs of its occurrence. (FOSTER 1960). The main crystallization stage may have begun under sub-cooled or super-saturated conditions, and the temperature may have risen on account of the rapid crystallization. Later changes in the rate of crystallization may also, perhaps, have induced changes in the crystallization temperature. Possible changes in the concentrations of volatiles contained in the magma, again, bring about, in turn, changes in the crystallization temperatures of minerals by affecting the pressure. Fig. 1 (Plate VII) shows two plagioclase phenocrysts contained in granite porphyry. The ragged margin seems at first glance to have been caused by corrosion. But this margin is a separate band and it has a sharp, straight border against the beautifully euhedral crystal in the center. In addition, mineral grains of the groundmass have become attached to the marginal band. The present writer maintains that this margin is not a product of corrosion, but evolved during the effusion period or thereafter simultaneously with the groundmass and that the sharp boundary against the intratellural central crystal is evidence of an eruption, signifying an interruption in the crystallization process. It has already been shown that not all the tongues in the phenocrysts are products of corrosion, either, but represent magma inclusions left inside the crystal during crystallization.

Autometamorphic processes also have a significance of their own in the genesis of ovoids. Their significance is apparently one that increases in direct proportion to the size of the crystallizing rapakivi massif. It is noteworthy, moreover, that they form new structures.

A portion of the plagioclase-mantled potash feldspar ovoids, again, crystallized around the ovoids directly out of the magma. There are cases where the plagioclase shell had crystallized around its parent crystal even before the magma erupted, *i.e.*, intratellurally. This is proved, then, by those plagioclase-mantled potash feldspar phenocrysts that have been found in hypabyssal and effusive rocks with the rapakivi composition, in addition to potash feldspar, quartz and plagioclase phenocrysts. On the other hand, plagioclase shells — like independent plagioclase crystals, too — may begin to crystallize almost from the very beginning of the main crystallization stage, if the conditions are favorable, and they are likely to keep on crystallizing throughout the main crystallization stage and, perhaps, even after that. This is indicated, *e.g.*, by the fluctuations in anorthite content in the independent plagioclase and, on the other hand, the plagioclase shells of the rapakivi. This conception is based on the plagioclase composition determinations made from a number of rapakivi samples from the Viipuri and Aland Island areas. These determinations have been made by the immersion method and, although the differences are not large, they are nevertheless clear, inasmuch as the same liquids were used under identical conditions and at the same time. It is intended to continue with these determinations. It has become surprisingly clear that within the same sample the anorthite content of plagioclase has been larger in part of the plagioclase shells, smaller in some other part, and equally large as in independent plagioclase grains. Under peaceful conditions, the rate of formation of crystal nuclei in a viscous magma is very low, if the prevailing temperature is not altogether suitable. In such cases, the potash feldspar ovoids could have promoted the genesis of the plagioclase crystal nuclei alongside them both catalytically and by bringing about the most suitable concentration for the crystallization of the neighboring plagioclase as elsewhere in the magma. The occurrence of numerous concentric plagioclase shells around the same ovoid is readily explained by means of change in the local concentrations induced by slow diffusion, but a part of these structures may have been formed also by autometamorphic processes. Deciding which of them is a different matter. In considering concentration changes in rapakivi magmas, in addition to the rate of crystallization, the viscosity of the magma, the rate of diffusion, the abundance of crystal nuclei, etc., it is also important to bear in mind that the rapakivis often contain 50—60 % potash feldspar by weight. Accordingly, changes in the rate of crystallization of potash feldspar affect the concentrations most, or, in other words, the crystallization of the potash feldspar and that of the other minerals take place alternately.

The inclusions encircling the potash feldspar ovoids as zones formed by other minerals as well — such as biotite and quartz — are to some extent magmatic, but are in part also likely to be autometamorphic.

In undertaking to deal with the autometamorphic processes occurring in rapakivis [1]), it is pertinent to call attention, first of all, to the final paragraph of the work by TUTTLE and BOWEN (1958) on the subject:» In summary, the experimental results presented here demonstrate that many changes have taken place in: (1) the compositions of the minerals, (2) the polymorphic forms of the minerals, and (3) the textural relations

[1]) The term autometamorphosis has been applied so extensively that it embraces all the changes in the form, structure and composition of the minerals which the crystallizing magma and its own residual solutions produce in situ during the course of crystallization. However, not all the ocurrences have been regarded as autometasomatic where a mineral has replaced another of diffreent composition.

among the minerals of the granites, syenites, and nepheline syenites between the time of final magmatic crystallization and observation under the petrographic microscope. A few of these changes have been described here, but many more complex reactions undoubtedly remain to be discovered. For example, probably the myrmekitic intergrowths, the altered cores of plagioclase, and the development of muscovite will be found to owe their origin to these post-crystallization reactions.» In agreement with the foregoing, it might be stressed that it is not always easy to decide in what way any particular mineral and its texture have developed. More minerals may have evolved during autometamorphosis than, for instance, a routine microscopic examination can reveal.

GOODSPEED (1959) writes: »The textural features of magmatic rocks are, in general, the result of a crystallization sequence from an initially fluid magma to a nearly solid medium. The magma is usually limited by the boundaries of the intrusion and with falling temperature solidification and crystallization proceed from the borders inward. In the late magmatic or deuteric stage of crystallization the residual magmatic solutions may replace some of the earlier formed pyrogenetic minerals thus producing textural features somewhat similar to those found in metasomatic rocks.»

Next, we shall consider an old observation on the color of rapakivi types which the present author has discovered to be valid. According to POPOFF (1928), the fundamental feature of pyterlite is its intense, dark brownish-red color, which immediately distinguishes it from the paler viborgite. How to explain this? First of all, as is well-known, potash feldspar as occurring in rapakivis contains abundant iron. (The writer has met with iron in abundance also in the plagioclases of a certain rapakivi type in the Ahvenisto area.) According to HOLMQUIST (1939), the potash feldspar of rapakivi contains iron as Fe_2O_3 in an amount of 0.3—1.1 % by weight. The alkali feldspars crystallized at a high temperature appear to be able, in general, to absorb considerable iron into their lattice. SHIMIZU and KUNO (1960) have investigated the anorthoclase phenocrysts of trachyte, which, they report, contains 0.99 % Fe_2O_3. In pyterlite this iron has exsolved into hematite pigment and gives the rock an intensive color. But in viborgite, apparently, there had occurred something else, too, following the unmixing. At the core of the massif, there apparently had prevailed conditions favoring autometamorphosis to such an extent that the exsolved iron recrystallized into other minerals, biotite and hornblende. This was evidently brought about by the same factors as caused the variation in triclinicity. Thus, some of the inclusions of the potash feldspar ovoids in viborgite, which are present in abundance compared to the potash feldspar grains of pyterlite, may be autometamorphic. This is not, however, the only reason for the abundant occurrence of the inclusions. It is certainly not easy to distinguish these autometamorphically formed grains from the perhaps slightly altered mineral grains that had crystallized directly from the magma. Also the grains formed autometamorphically apparently tend to grow in the direction of the crystal faces, for it is in any case easier to grow in that direction than in any other. This, then, was one example of autometamorphosis.

Perhaps the most conspicuous example of the importance and extent of the autometamorphic phenomena to be observed in rapakivi may be seen in Fig 2. (Plate VI). The three ragged mineral crystals appearing there — and having nearly a simultaneous extinction between crossed nicols — are of potash feldspar. As inclusions they contain plagioclase grains posssessing a small, pale albitic marginal zone, as is to be expected. These three ragged potash feldspar grains occur as inclusions in a large mineral grain appearing pale grey in the picture, which is also potash feldspar. Here, accordingly, one potash feldspar grain has in part displaced another, previously crystallized potash feldspar grain. It may be that these three grains had originally been one and the same grain. Perhaps the earlier grain had been in an unfavorable position in relation to the later grain; perhaps the later grain had been energetically more favorably situated under the prevailing conditions than the earlier one; perhaps they had had a difference in composition or/and perhaps some other factor accounts for the substitution. Peaceful, balanced crystallization conditions evidently bring it about that a maximum degree of equilibrium, if it might so be termed, is achieved during the course of crystallization even for seemingly trivial reasons. Perhaps processes and structures of this kind are produced precisely on that account. Potash feldspar is probably not the only rapakivi mineral that displaces itself. To what extent does displacement of this type occur? Evidently, one should not scorn the significance of this phenomenon, and efforts ought to be made to delve into it by taking advantage of, *e.g.*, the thinking of Schermerhorn (1960). It is possible that entire mineral groups vanish, to be replaced by others.

Among these processes belong also the variations in the triclinicity of potash feldspar and, possibly, corresponding phenomena in other minerals — quite as do the exsolution phenomena, like, for instance, perthite.

Sederholm (1916) describes myrmekite textures as occurring also in rapakivis, although to a lesser extent than in other rocks, regarding them in part as products of a late stage of magmatic crystallization and in part as metamorphic. In addition to ordinary quartz-plagioclase myrmekite present in rapakivis he describes myrmekite-like biotite-quartz intergrowths as well as hornblende grains, at the cores of which there are quartz and fluorspar grains. Included are also such phenomena as the chloritization of biotite, the albitization of plagioclase, *etc.* — easily recognized signs of secondary mineral formation. In connection with biotite undergoing chloritization, the author has often met with unusual zirkon and fluorspar grains, many of them of quite conspicuous size. And he is inclined to venture the tentative assertion in this connection that these minerals had not crystallized until the chloritization process had started. The mode of occurrence, which may also be noticed in other rocks, supports this contention. The fact that some mineral has replaced another can be easily seen if traces of the other mineral are still left. To what extent have minerals in rapakivis replaced others so as to cause the replaced mineral to vanish totally?

The greatest obstacle to understanding the rapakivi fabric has been the fact that the plagioclase occurring in these rocks has been too euhedral; and since the potash feldspar, moreover, replaces the plagioclase at a certain stage, there evolves a fabric which easily gives a false picture of the course of crystallization. Matters related to this problem will be discussed in the next chapter. Here we may already affirm that this idiomorphic feature is surely due mainly to the great idiomorphic energy of plagioclase, in the main, in opposition to other minerals. For another thing, the idiomorphic energy of minerals and their stability inter-relations evidently change during crystallization. These changes apply particularly to the inter-relations between potash feldspar and plagioclase. Such changes are quite as natural an occurrence during the course of crystallization as the transitions in the composition of minerals and the alterations in their crystal structures. In the understanding of autometamorphic phenomena, also knowledge of stability inter-relations is important.

According to Kanerva (1928) and Savolahti (1956, 1956 a), the megascopic appearance, fabric and mineral composition of porphyry-aplites varies in the veins of these rocks to a considerable extent, even within the same massif. These veins are a differentiation product deriving from the rapakivi magma at a late stage in its crystallization. Their composition and texture naturally depend principally on the degree of crystallization and differentiation of the magma, or the moment of injection of the porphyry-aplites. This moment of injection need not have been the same for all, as it varied even within the same massif. When it is taken into consideration, moreover, that an individual erupting porphyry-aplite magma could have been super- and unsaturated in very different ways in relation to different minerals at the moment of injection, the varying properties of the porphyry-aplite veins become understandable.

The so-called porphyric granites did not undergo quite the same crystallization conditions as the rapakivis. That is one reason for the difference between them. The transformation of the marginal parts of the porphyric granites to resemble the rapakivi texture is a major factor in the autometasomatic processes which are given impetus by the difference in crystallization conditions.

The presence of even a slight amount of fluorine in the rapakivi magma has a strong effect on reducing the crystallization temperatures of the minerals.

Through the action of purely metamorphic and metasomatic processes, structures and rock types quite like the structure of rapakivi can evolve — and do evolve —, for the processes referred to differ nowise in marginal cases from the autometamorphic, autometasomatic, and even magmatic processes. This must not be construed in any way to be an admission that the rapakivi in Finland had formed in the first-mentioned way. I wish to add here a statement by TURNER and VERHOOGEN (1960): »This does not exclude the possibility that some large feldspar of some granites are porphyroblasts of postmagmatic origin.» For such could be — and in part is — the case with rapakivis, too. These porphyroblasts are of a late- and postmagmatic origin, but autometamorphic and autometasomatic processes were the cause. It was not essential for potassium or any other constituent to be introduced from the outside.

PART II

THE RULES OF IDIOMORPHISM IN MINERALS

1. INTRODUCTION

SEDERHOLM (1891) in his immortal work on rapakivis, writes: »Der Zweck dieses Aufsatzes wäre aber doch erfüllt, wenn ich verständlich gemacht hätte, dass mit diesen Gesteinen viele Fragen von allgemainerem geologischen Interesse verknüpft sind, und dass man vor allem hier eine ungewöhnlich gute Gelegenheit hat, die verschiedenen Structuren, welche eine grosse granitische Magmamasse bei ihrer Erstarrung ernimmt, in ihren vielfachen Beziehungen zu einander zu studieren.» Therein lies one reason why this part of the present study has been written, for there has been a really good opportunity to examine these rules in connection with rapakivis. Another reason is that phenomena are encountered in rapakivis that are difficult — even impossible — to explain without a critical examination and re-evaluation of the rules governing idiomorphism and the crystallization order in minerals.

ROSENBUSCH (1882) and BECKE (1913) formulated the rules governing idiomorphism and crystallization in rocks very nearly as they have been applied up to now and are still being applied. And the rule drawn up by the former deals with the order of crystallization of minerals in magma rocks, when the rules of the latter govern the structures and the idiomorphism of the minerals in metamorphic rocks. These rules must be considered in combination, for they form a totality and between them there is no sharp boundary. Accordingly, the consideration of both belongs here, although in the present study only magmatic phenomena have been previously considered.

There are petrologists who are of the opinion that it is useless to place stock in ROSENBUSCH's (1882) rules and their premises, because so many exceptions to them are known and so much criticism has been levelled against them. But it must be remembered that many of these exceptions were known already at that time and that attention to them began to be drawn, together with the criticism of the rule. ROSENBUSCH (1882, 1923) himself was aware of these exceptions already at the time of formulating his rule and certainly did not forget their existence during his lifetime. Nevertheless, the rule and its premises gained a very prominent place in petrology. They still continue to hold this place. It may be said that all the hypotheses concerned with the petrology of magmatic rocks as well as the theories of their genesis are based in some way on this rule and on the premises supporting it. It is also certain that numerous research scholars consider the rule to be quite serviceable in spite of the exceptions to it. Indeed, there are those who in quite recent years have taken up arms to defend it. In the second edition of one of the best petrological textbooks — TURNER and VERHOOGEN 1960 — a work well known and highly esteemed also in this country, pronounces favorable judgment on the matter. It does make the reservation, to be sure, that the implications involved can be discussed adequately only in the light of experimental data. Even those who deny the validity of this rule and the premises on which it is based lean on them unconsciously in formulating their own petrological concepts.

As for the rules advanced by BECKE (1913) concerning the minerals of metamorphic rocks, this criticism has not been so voluminous and vociferous as in ROSENBUSCH's case nor has it been aimed against them as a whole; however such criticism (*e.g.*, ESKOLA 1939 and TURNER and VERHOOGEN 1960) has not been opposed.

These rules and their premises have been applied, and are still being applied, to both magmatic and metamorphic rocks as axiomatic. May I now be so bold as to ask the reader to consider the justification of such application.

2. A BRIEF REVIEW OF THE CRITICISM AIMED AGAINST ROSENBUSCH'S AND BECKE'S RULES

SHAND (1949) has thoroughly criticized the rule governing the crystallization order of magmatic rocks and the premises on which it is based, and in conjunction with it he refers to earlier criticism. Many a petrological celebrity flits by in his critical paper. The faults, shortcomings and exceptions that burden ROSENBUSCH's (1882) rule and its premises are conspicuously brought to the fore in discussion. SHAND reaches the conclusion: »The Rosenbusch rule has no theoretical or experimental justification, and even on the basis of observation it cannot be upheld. Yet the strange fact remains that the rule seems to be applicable to a large number of rocks.»

What follows is a defense of the latter remark and the presentation ends quite conventionally, thereby disappointing the reader who had been led to expect a sensation. Obviously, even numerous exceptions to a long-honored and widely accepted rule that had been expounded with great authority will not suffice — even though they should — to nullify it and cause its rejection. It is only human.

In quite recent years, a number of good studies dealing with and/or criticizing the rules governing the order of crystallization of magmatic rocks have been published, *e.g.*, MOORHOUSE (1956), SCHERMERHORN (1958) and GOODSPEED (1960). Among them, MOORHOUSE (1956) is extremely outspoken in writing: »More than most sciences, the study of rocks has been burdened with an incubus of outworn ideas, some of which are merely useless baggage, and others are downright misleading. Among the latter, one which receives regular reincarnations in textbooks is the 'Rosenbusch Order of Crystallization,' in spite of the fact that it was superseded long ago by Bowen's Reaction Series» . . . Sternly put, but certainly behind those words are observations of the critic's own and of others, though, it is true, the cases involved are individual. I shall not here undertake to consider the writer's premises in detail, for everybody with pretentions to being a petrologist owes it to himself, if only on account of the foregoing citation, to become personally acquainted with the paper in question. Against this outpouring of MOORHOUSE's (1956), SCHERMERHORN (1958) ardently took up arms, which he handled with skill and, to some extent, with accurate aim. But only a couple of years later, he would no longer have wielded his weapons in the same way (SCHERMERHORN 1960), or then the present writer has read too much between the lines in this later publication. From the standpoint of the present study, there is no real point to trying to decide what matters in each of the foregoing papers are well taken, for such an undertaking would lead into a jungle of details and axioms — the result would be a separate volume of a considerable work. In no event did the rejoinder help. MOORHOUSE (1959) threw the rule under consideration out of his illuminatingly written textbook on petrography without further ado — as he was justified in doing. But it is not at all certain that, in so doing, he was also able to get rid of the ballast that had accumulated in petrological literature under the influence of the rule during the course of time.

GOODSPEED (1959) correctly affirms that textural features alone are not always adequate for petrogenetic interpretations. This is stated in a study in which he gives numerous illuminating examples of the textures of magmatic, deuteric, metasomatic and metamorphic rocks. He considers the textural features of metamorphic and metasomatic rocks as very similar. The present writer cannot agree with him in all his conclusions because, for instance, he places too much weight on the metasomatic and comparable processes while endeavoring to prove his result. Not nearly always do the magmatic textural features differ from metasomatic and metamorphic ones — a circumstance that would also be worth keeping in mind.

3. ON THE RULES GOVERNING IDIOMORPHISM AND THE ORDER OF CRYSTALLIZATION OF MINERALS IN GENERAL

Before we begin to consider the rules applied to the idiomorphism and the order of crystallization of minerals, we would do well first to examine the question from a quite natural point of view. The reliable fundamental laws of physics and chemistry inform us how the matter stands in a broad sense. Proceeding from these laws, we may state quite generally: The chemical composition of the material providing the point of departure and the pressure (P) and temperature (T) conditions prevailing in it determine — in addition to the individual crystallization tendency of the minerals themselves — when each mineral starts to crystallize and in what order the crystallization of different minerals takes place; on the other hand, the individual properties and crystallizing force of the minerals determine the idiomorphism of the minerals and the other properties comparable to it at that particular moment, under the PT-conditions prevailing just then and in the medium where the crystallization happens to take place at any given time.

The original material may be solid rock, magma, a solution containing water or gas and/or some intermediary material. The individual crystallization tendency of a mineral depends, again, primarily on the rates of formation and crystallization of the nuclei and on the tendency of the mineral to create super- and unsaturated states in magma at any given moment. Properties comparable to the euhedral character of a mineral are, *e.g.*, its capacity to acquire inclusions and its ability to thrust aside alien obstucting material. The properties of minerals comparable to idiomorphism are all those bearing on the structure, strength and energy content of the lattice. The medium may be and often is heterogeneous. The term crystallization force, again, embraces the minerals' form energy, growth energy, the energy required to thrust aside the intervening material and the energy to appropriate inclusions. The following example illuminates the appearance of these forces: Staurolite mica schists, viewed megascopically, exhibit staurolite crystals possessing a very pronounced euhedral character, while the crystal faces are well-developed and appear to be flawless. But microscopically examined, these crystals are in many cases quite full of tiny mineral inclusions, which are concentrated in spots or distributed throughout the crystal. The section lines of the crystal faces are straight and clear enough, *i.e.*, the euhedral appearance of the crystal is good megascopically, but this line is broken by small inclusions. The staurolite crystals have a tendency to acquire a beautiful crystal form, but they at the same time have a conspicuous faculty for absorbing foreign inclusions. There are again, other mineral crystals, which are not euhedral but do not contain inclusions, either. One can see that they have thrust aside the medium in spite of their anhedral form.

The foregoing determination constitutes, in a way, the general rule which governs the crystallization and formation of minerals under all possible circumstances. To be sure, it should be added that the pressure (P) can, of course, vary from place to place but it can also be monotonously unvarying. It can and does cause displacements, whereupon the pressure thus also carries with it the effects which tectonic movements bring to bear in the mineral formation.

The rules governing the euhedral form and the crystallization of minerals should be watched all the time with this general rule in mind. That is why it has primarily been discussed at this point.

4. THE RULES OF ROSENBUSCH AND BECKE

It was in 1882 that Rosenbusch formulated the order of crystallization of igneous rocks, which was soon generally accepted and applied as a rule. In due time, this rule became known by his name. It states that minerals become separated from the rock melt consecutively in four groups: 1) First, the ore minerals and accessory minerals (magnetite, hematite, ilmenite, apatite, zircon, spinel, sphene), 2) then the magnesium- and iron-bearing silicate minerals (biotite, amphiboles, pyroxenes, olivine), 3) next, the feldspars and feldsparthoid minerals (feldspars proper, nepheline, leusite, melilite, sodalite, hainite), 4) and, finally, free silicic acid. Rosenbusch's rule is based on three principles, which are easy to appreciate on any examination of effusive and hypabyssal rocks — as its creator must have done. According to the first, the automorphism of minerals in rocks depends on their age. In other words, the earlier a mineral crystallizes from the rock melt, the more nearly flawless will its outward appearance be, for the freer will have been the space in which the crystallizing process will have taken place. According to the second, there are fewer inclusions in minerals of earlier crystallization than in those of later crystallization, and in general the mineral present as an inclusion is older than the mineral surrounding it.

According to the third, in rocks containing both large and small crystals, the large ones had begun to crystallize earlier than the small ones. These principles and the rule governing the order of crystallization determine the crystallization of igneous rocks.

ROSENBUSCH's rule concerning the order of crystallization is attended by an abundance of exceptions. This is as it should be, for every magma begins to crystallize and continues to crystallize in a way that depends on its chemical composition and on the pressure and temperature conditions prevailing in the magma chamber. Since there exist magmas of very diverse compositions and the conditions attending them are different, the igneous rocks have solidified from them must naturally have also had numerous different orders of crystallization, and ROSENBUSCH's rule is only one among several.

The principles on which ROSENBUSCH's rule leans appear at first sight to admit only a single interpretation, but the fact of the matter obviously is not like that. First of all, the time of crystallization does not alone determine — though it might do so — the idiomorphism of the mineral in magmatic rocks, for the crystals do not crystallize in a free space but within a medium which is a magma — even if a viscous one. This viscous magma can also successfully resist the formation of crystal forms. If the different minerals crystallizing out of the magma happen to possess form energies of different potency during the crystallization process, then these differences might, at least in theory, affect the order of idiomorphism of the minerals. Rapakivi magma is viscous.

For another matter, the abundance of inclusions does not exclusively depend on the age of crystallization, either, for in metamorphic rocks porphyroblasts of different minerals are known to be present in the same rock, and these porphyroblasts in many cases are likely to contain inclusions in different amounts, as, *e.g.*, in the case of garnet and staurolites present in the same rock. The explanation is this: they differ in their ability (or energy) to thrust aside obstructing minerals or in their capacity to enclose foreign minerals as inclusions. It is natural that in magmatic crystallization, especially of a late stage, this phenomenon should also become conspicuous. Naturally, different minerals vary in the way they enclose magma inclusions.

No more need the occurrence in the same rock of large and small crystals, if they happen to represent different minerals, signify that the larger ones had begun to crystallize first. A difference in size between different minerals can be also due to the fact they have had different rates of growth. If, on the other hand, the same mineral is involved, then the occurrence in the same rock of crystals of two distinctly different sizes signifies, in many cases, also a difference in age. (See also SCHERMERHORN 1959, pp. 489—493.)

Into the rule bearing his name, ROSENBUSCH incorporated the re-examination of numerous observations made by himself and many others of the textures of magma rocks and the relations between the minerals contained in them. A study of his writings and extensive textbooks, mentioned herein, leads to a conviction that the initial impulse at the formulation of the rule had been given by the beautiful porphyric, or corresponding, texture of effusive and hypabyssal rocks. As the pheńocrysts contained in them are generally more idiomorphic than the minerals in the ground mass, the relation between the crystal forms and order of crystallization of the minerals appears so clear. The exceptions to the idiomorphism of phenocrysts can be explained on the basis of corrosion — as ROSENBUSCH (1882) did. Perhaps, during the initial stage of its birth in Rosenbusch's mind, the rule embraced only effusive and hypabyssal rocks. To be sure, when Rosenbusch propounded the rule, it had expanded to embrace all the igneous rocks, including the plutonic rocks. At the same time, the rule had noticeably, though not to a great extent, been affected by the fabric of plutonic rocks. This dualism is still to be perceived in it, and it brings in special complications in applying the rule. The rule hereby obtained has evolved, together with its premises, into an axiom of a fundamental character in the theoretical treatment of igneous rocks.

The structure of metamorphic rocks is governed by rules of a different kind. According to BECKE (1913), crystallized schists have a crystalloblastic structure. The distinguishing marks of a typical crystalloblastic structure are: 1) no mineral in a rock can crystallize before the others and all the minerals can be present as inclusions in the others, 2) the development of crystal forms is relatively rare; if they do evolve, the crystal forms are simple and they in many cases exhibit only one face,parallel to the cleavage, or one zone formed by faces, 3) skeleton crystals are lacking, 4) the development of the crystal forms is retarded as the crystallizing force of the mineral diminishes, and the minerals may accordingly be placed in a row, where the one coming after invariably, upon touching the one in front, receives the latter's form, 5) parallel structures also form during dynamic metamorphism and, above all, in a direction perpendicular to the maximum pressure, 6) the minerals lack a zoned structure, or, then, other laws than those applying to consolidation in a rock are observed, 7) the crystal inclusions in the minerals in most cases do not adhere to the zonal structure, but rather to either the growth pyramid or the old structure of the crystal, 8) the crystalline schists are of utmost compactness, and they lack miarolitic structural features.

According to Becke's rules, the essential feature of crystalline schists is the fact that the crystallization in them takes place the whole time in solid rock, that the growing mineral is bounded by numerous other minerals, which resist its growth in various ways, and that the form of the mineral is determined by the mutual adjustment between it and the neighboring grains. The at-first-sight far-reaching applicability of Becke's rules is based on the correct usage of the central term appearing in them — crystallizing force. According to ESKOLA (1939), the structural analysis of minerals contained in metamorphic rocks indicates that they do not always crystallize simultaneously, although this cannot always be perceived (TURNER and VERHOOGEN 1960). Neither is the conception of the simplicity of the crystal forms and the indices of their faces always so self-evident, for it is also affected generally by the PT-conditions prevailing during metamorphosis. Mineral crystals richer in form invariably evolve during high temperature facies. Also temperature areas in which magmatic and, on the other hand, metamorphic rocks originate, can overlap, with consequences affecting the crystal form of the minerals. Nor does the absence of zonal structures — or a difference between them —, which a reversed zoning characterizes, represent a structural feature characteristic only of metamorphic rocks. Zoning may be absent in the minerals, *e.g.*, plagioclase, of igneous rocks, or it may be in reverse, too. Good examples of such reversed zoning are the plagioclase grains of the porphyritic diabase found at Träskby, in the commune of Sipoo, where the marginal zone is about 10 % richer in An than the core and which SAASTAMOINEN (1956) has described.

The origin of Becke's rules, too, resembles that of Rosenbusch's rule. Quite clearly, the idea was generated by the examination, in the first place, of slightly metamorphic rocks, in which relicts of a previous structure still remain. It should be remarked further that Becke himself applied his rules in a more limited sense than they are applied nowadays. This is noted also by TURNER and VERHOOGEN (1960). The method of applying the rules is also of the same kind as in the case of Rosenbusch's rules. They are applied in the petographic and petrological study of metamorphic rocks as axiomatic to a fundamental degree.

5. THE VALIDITY AND APPLICABILITY OF ROSENBUSCH'S AND BECKE'S RULES

There thus exists a general system governing the idiomorphism and order of crystallization of minerals, which we attempted to describe briefly in the foregoing in the form of a brief rule. As considered in the present connection, it makes no difference if our rule does harbor certain flaws. The main thing is that there does exist such a general system that embraces the whole of this group of phenomena. Nor is it important that this rule as such cannot easily be applied in the petrographic classification of rocks. But its existence is essential in examining the applicability of the rules mentioned in the heading above.

Then, we have the rules of Rosenbusch and Becke, which are applied in petrographic research in determining the genesis of rocks and in their classification. It should further be emphatically reiterated that these rules are applied axiomatically. This last statement calls for some elucidation. Nowadays the term hypidiomorphic is defined in such a way that it includes both Rosenbusch's rule and its premises. [1]) Correspondingly, Becke's rules define the crystalloblastic structure of the term. The applicability of these terms is as follows: If a rock has a hypidiomorphic structure, it is said to be an igneous rock; if a rock has a crystalloblastic structure, it is said to be a metamorphic rock. This practice, if anything, signifies the axiomatic applicability of these rules. Such it is, regardless of whether other premises have been resorted to in carrying out a determination or not.

In the third place, we must examine the mode of origin of Rosenbusch's and Becke's rules. The beginning of both can be traced to a narrow category of phenomena, which was later enlarged to include a much larger group of phenomena than was initially the case.

Rosenbusch's rule at first was concerned with effusive and hypabyssal rocks, after which it was enlarged to embrace plutonic rocks as well. To be sure, the rule does not take into account, insofar as it relates to effusive and hypabyssal rocks, the fact that the order of crystallization in these cases, too, had depended on the composition of the magma and on the conditions of crystallization and that it may even have been quite reversed. Even ROSENBUSCH (1923) was to some extent aware of this. Nor does it take into account the fact that also the magma acts as a medium and sometimes a viscous one, so that the differences in the form energy of minerals may become and do become evident already at the magmatic stage. Moreover, in expanding to embrace also plutonic rocks, Rosenbusch's rule received features from the structures of plutonic rocks, too. Consequently, it exhibits a dualism, which adds to the difficulties of applying it. That the origin of the rule was as described here can be ascertained from Rosenbusch's own writings. In all the editions of his petrographic textbooks, he argues in favor of the applicability of his rule also to plutonic rocks. ROSENBUSCH (1923) contends, for instance, that the phenomena of the metasomatic period of eruptive rocks are not such as to form structures but to produce changes, *e.g.*, they have a disturbing influence — but the effects of these disturbing factors, of which he mentions

[1]) Hereafter the present writer will refer to the rule and its premises in combination as Rosenbusch's rules.

quite a number, are generally slight (see, for example, the rapakivis and SCHERMERHORN 1960).

In respect to their formulation and application, Becke's rules are quite comparable to Rosenbusch's. They initially applied distinctly to comparatively slightly metamorphosed rocks with a relict structure, while subsequently their range of applicability expanded to embrace all the metamorphic rocks. Moreover, Becke himself employed his rules in a narrower sense than they are applied nowadays. Also TURNER and VERHOOGEN (1960) write about the expansion of the range of applicability of Becke's rules.

As applied at the present day, the rules of Rosenbusch and Becke extend throughout the range covered by the general rule mentioned in the beginning, *i.e.*, they govern, at least in principle, the idiomorphism, order of crystallization and, at the same time, structure of the minerals of all magmatic and metamorphic rocks. Now it may be asked whether the extension of these rules from some special case so as to embrace the entire groups of rocks to which the said special case belongs, is possible and warranted. Since the question relates to rules applied axiomatically, we must be far more severe than otherwise. The reason is that axioms determine what sort of theory can be formulated. It is well-known, if not always borne in mind, that it is possible to construct an altogether arbitrary system if only the axioms are suitable chosen. Such, however, is probably not the way to go about it even in the field of petrology. Therefore, we shall proceed in the following to scrutinize these axioms.

In order for these axiomatically applied rules to have validity, they must fulfill some conditions. The first condition is that they admit no exceptions within the narrow field in which they originated. It is to be noted immediately that Becke's rules, after subsequent rectifications, which are still perhaps being carried out, fulfill this condition. It is quite as easy to perceive that Rosenbusch's rules do not satisfy the conditions. The exceptions to them are too numerous nor do the rules take sufficiently into account the factors evidently bound up in the overall rule. Since, critically applied, Rosenbusch's rules nevertheless do provide indications of the structure of effusive and hypabyssal rocks and their crystallization, we shall not concern ourselves further with their shortcomings in relation to the aforementioned rocks. Moreover, erroneous interpretations are relatively easy to avoid in dealing with these rocks. Within such a framework, the application of Becke's and Rosenbusch's rules need not lead into error in the classification of rocks, either, since rocks always have other structural features in ample measure to facilitate interpretation — in the former instance, relict structures and in the latter clear magmatic structural features besides the hypidiomorphic. Both rules are thus applicable within their original range.

For another thing, the rules under consideration must be applicable within the enlarged range into which they have subsequently been carried also without exceptions. The rules should there be unambiguous. That is to say, every structure which has a hypidiomorphic appearance must have originated through a magmatic crystallization process, and every structure that has a crystalloblastic appearance must have formed metamorphically (and metasomatically) through a process of re-crystallization. On the other hand, we must require that rocks possessing a structure of a hypidiomorphic and crystalloblastic character do not evolve except as defined by the rules. In other words, our requirement is that metamorphic and/or metasomatic re-crystallization does not lead to the formation of structures resembling the hypidiomorphic and that magmatic processes do not create structures resembling the crystalloblastic. This thought can be given a slightly different significance by stating that our prerequisite is that the PT-area where the crystallization of magmatic rocks occurs does not even in part extend over to the area where the metamorphic re-crystallization process takes place. Such a demand must be made if the rules be intended to embrace all the rocks. How is it?

The facies classification table published as late as 1939 by ESKOLA places alongside the metamorphic sanidinite facies the magmatic diabase facies, and correspondingly alongside the pyroxene-hornfels facies the gabbro facies, as well as alongside the amphibolite facies the hornblende gabbro facies. Later on, the magmatic facies were removed, not because they were erroneous but because they are comparable to the corresponding metamorphic facies and nearly identical minerologically with them. From this alone we can draw the conclusion that metamorphic and, on the other hand, magmatic rocks also form under the same PT-conditions. This is not how it should be, for under identical PT-conditions the identical rock invariably evolves out of matter of the same composition regardless of the road travelled. This, of course, necessarily presupposes that a state of equilibrium is reached. It is not nearly always possible to fall back on the counter-argument that igneous rocks always crystallize over a long temperature interval, with the temperature usually diminishing and that it is precisely the change in temperature that brings about the typical magmatic structural features. The rapakivis are the one sure exception to this, for their temperature gradient could scarcely have existed during their principal crystallization. The decrease in temperature had been so slow that a state of equilibrium had characterized the crystallization conditions throughout. As for the rocks, such as viborgite, of the great rapakivi masses, it has not yet been possible to ascertain in detail what part of the grains of any special mineral had crystallized directly out of the magma and what part had crystallized out of it autometamorphically and autometasomatically. It does not represent the line of demarcation between the so-called mineral generations. Evidently always when a large or a deeper-seated magmatic mass crystallizes under peaceful conditions, the process takes place in a state of equilibrium. Accordingly, autometamorphic and autometasomatic processes take place for a long period and on a large scale, and they act to form structures and do not merely represent disturbances. Rosenbusch (1882) himself was familiar with the structure of diabase and the great structural differences between the gabbros. These structural differences are apparently due solely to differences prevailing in the crystallization condition, differences which do not always have an effect even on the order of crystallization. On the other hand, some of the structural features produced by autometamorphosis and autometasomatism in rapakivis and other comparable magmatic rocks nowise differ from the structural features found in corresponding metamorphic rocks. Precisely the same is true of the structures of metamorphic rocks. There are differences in the structure of rocks of the same composition depending on the degree and mode of metamorphosis. The structural features produced by metamorphosis and metasomatism in many cases resemble the structural features of rocks originating through magmatic processes. Savolahti (1956 a) has presented Fig. 14, in which a beautiful amphibole margin is to be seen around the pyroxene grains — a margin that in some instances occurs homoaxially with the pyroxene. Such a homoaxial character is customarily held to be a clear magmatic structural feature in, for instance, the diorite of Jyväskylä. It is also noteworthy that in the Kiuruvesi region there are rocks that can be proved to have been products of metamorphic and metasomatic processes and that exhibit the features almost of porphyritic granites. Many other examples could be mentioned, but the matter is so self-evident that it would not be necessary in this connection. Suffice it only to observe that, according to an old and established concept, which is also brought out in textbooks, no sharp boundary divides magmatic from metamorphic rocks, but that rocks formed by different processes are likely to have very similar structural features. Take, for instance, the volcanic rocks and many sedimentary rocks that have undergone a far-reaching metamorphosis. Still, this is not taken sufficiently into account, although it is known, when the rules governing idiomorphism and the order of crystallization have been and are being applied.

The fundamental argument of this part is that logic categorically forbids extending the rules of Rosenbusch and Becke to embrace the total range of rocks if even a single exception exists. Accordingly, petrology cannot have such axioms in that range.

We may therefore affirm that, unfortunately, Rosenbusch's and Becke's rules are not valid when a rock lacks quite distinct structural relicts from a previous structure or wholly self-evident other proof of magmatic origin than the so-called hypidiomorphic structure and structural features having comparable value as evidence. The reason is that magmatic (autometamorphic and autometasomatic), metamorphic and metasomatic processes produce quite diverse structures and that all of them also produce quite identical structures. This may also be said: Dividing and restricting the general rule herein presented, in the way Rosenbusch's and Becke's rules did, can in no event be admissible or justified.

The present writer is convinced that the axiomatic application of Rosenbusch's and Becke's rules is largely to blame for, *e.g.*, the controversy involving granites versus granites. This is not to say, of course, that blame does not lie elsewhere, too. The controversy will not be settled until the false axioms discussed in this paper and the other factors contributing to stir up the battle are rejected.

Recently, I read these three sentences: »He is a magmatically-thinking scientist and tries to find observations that fit into his theories. I myself also started my geological research as a magmatist. Later on, however, the progress in petrology and my field observations caused me to change my mind in many respects.» If false axioms and false logic are rejected in petrology, these sentences should be revised to read: »I shall explain magmatic phenomena as magmatic, and metamorphic and metasomatic ones as metamorphic and metasomatic.»

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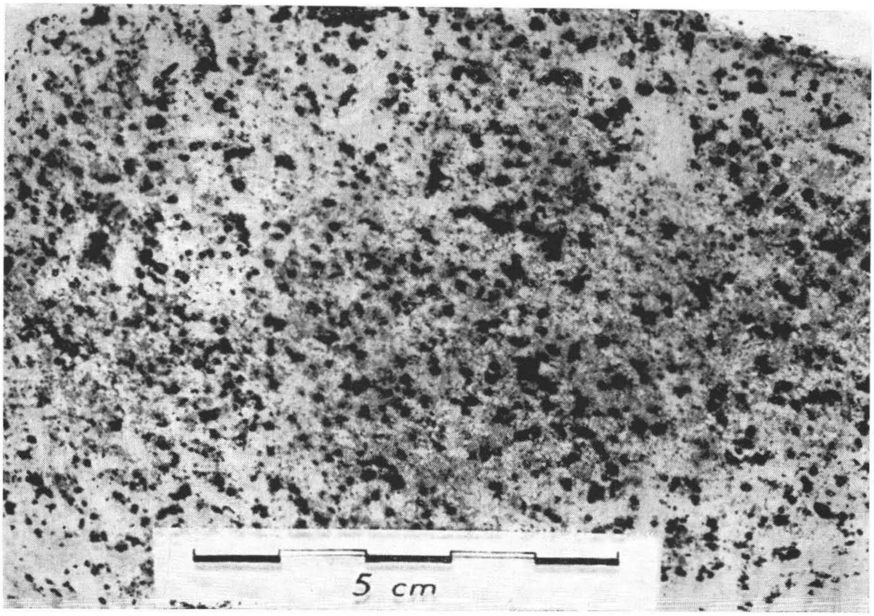

Fig. 1. Contact variety of rapakivi from southern margin of Suomenniemi massif. Laivolahti. Mäntyharju (PIPPING 1956).

Fig. 2. On left, euhedral fayalite crystal with rounded corners and iddingsitized in cracks, contained in hornblende. On right, quartz with xenomorphic fayalite grain plainly situated against it. Crossed nicols. Nurmaa. Mäntyharju. Ahvenisto area.

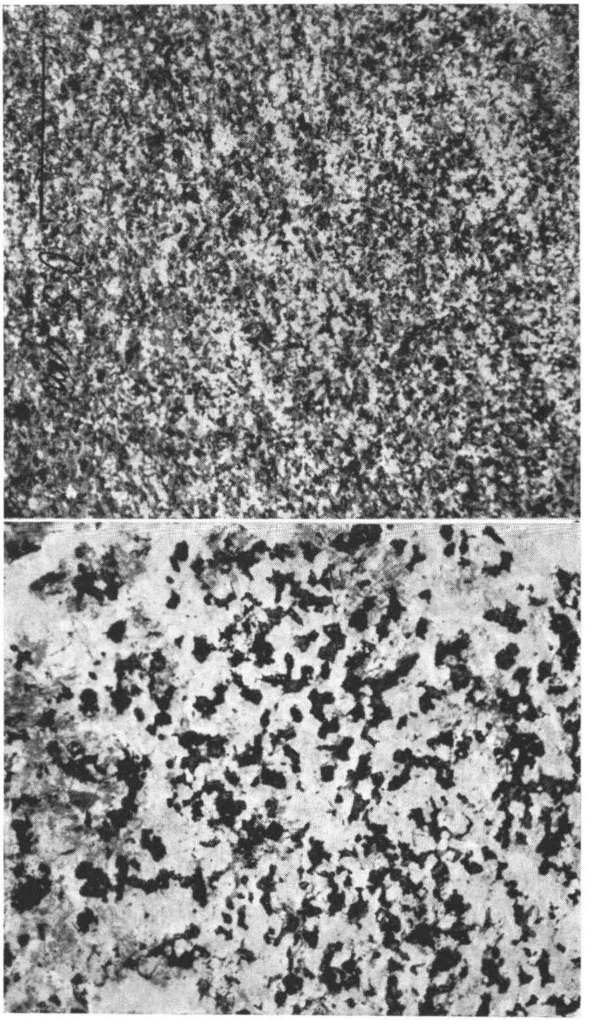

Fig. 1. Above, biotite plagiclase gneiss (crossed nicols). Below, the same rock metamorphised by rapakivi (without nicols). Commune of Heinola. Ahvenisto area.

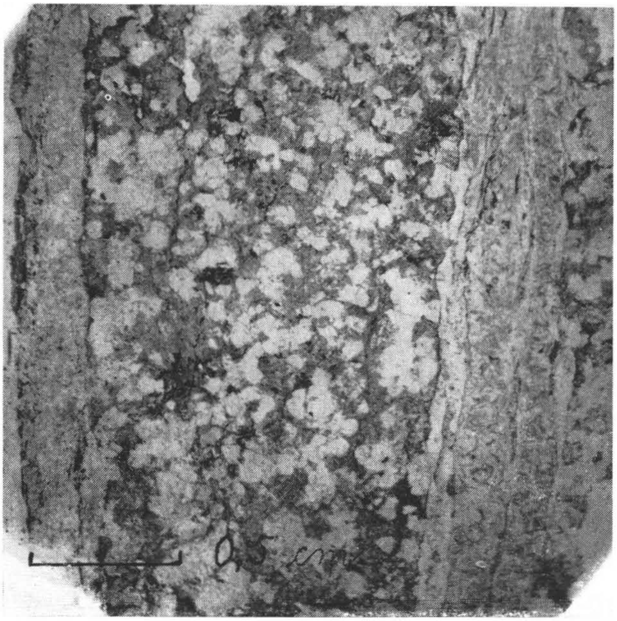

Fig. 1. Mylonitized rapakivi. Without nicols. NW side of Kuijärvi, commune of Heinola. Ahvenisto area.

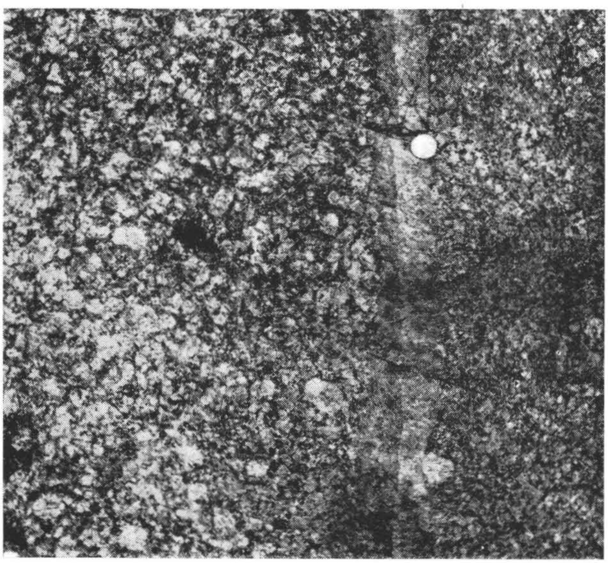

Fig. 2. Fault along fine-grained streak. Rapakivi at left is much coarser-grained than on right side. C:a 1/7 net. size. Viilajärvi E, Heinola commune. Ahvenisto area.

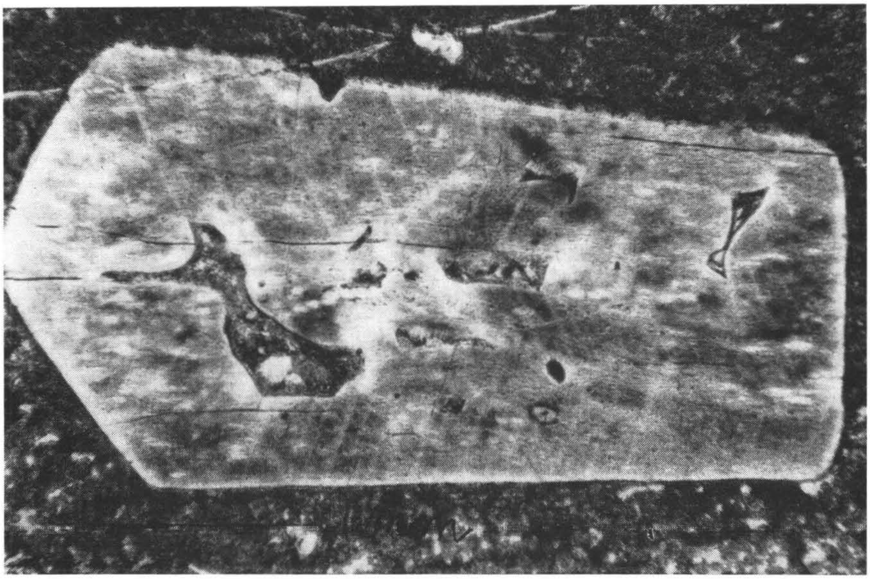

Fig. 1. Within a potash feldspar phenocryst of granite porphyry there has remained a magma inclusion which crystallized later. Without nicols, 30 ×. Heinola commune. Ahvenisto area.

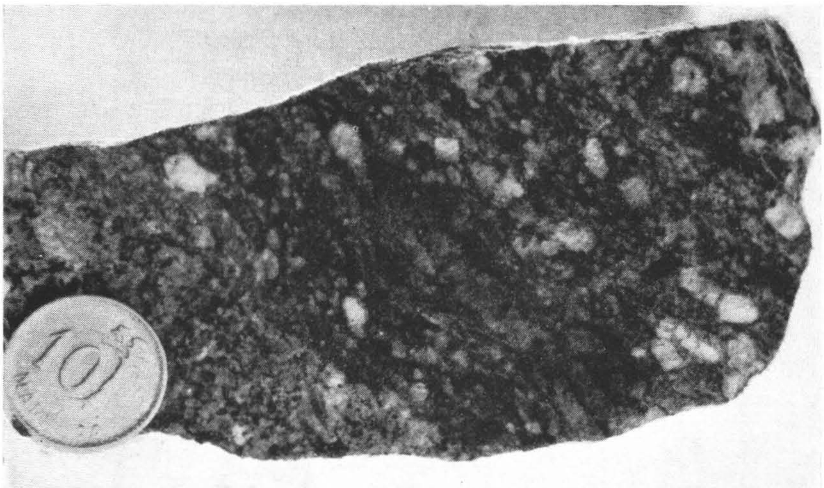

Fig. 2. Eye gneiss, metamorphosed out of rapakivi. Enovesi E. Mäntyharju. Ahvenisto area.

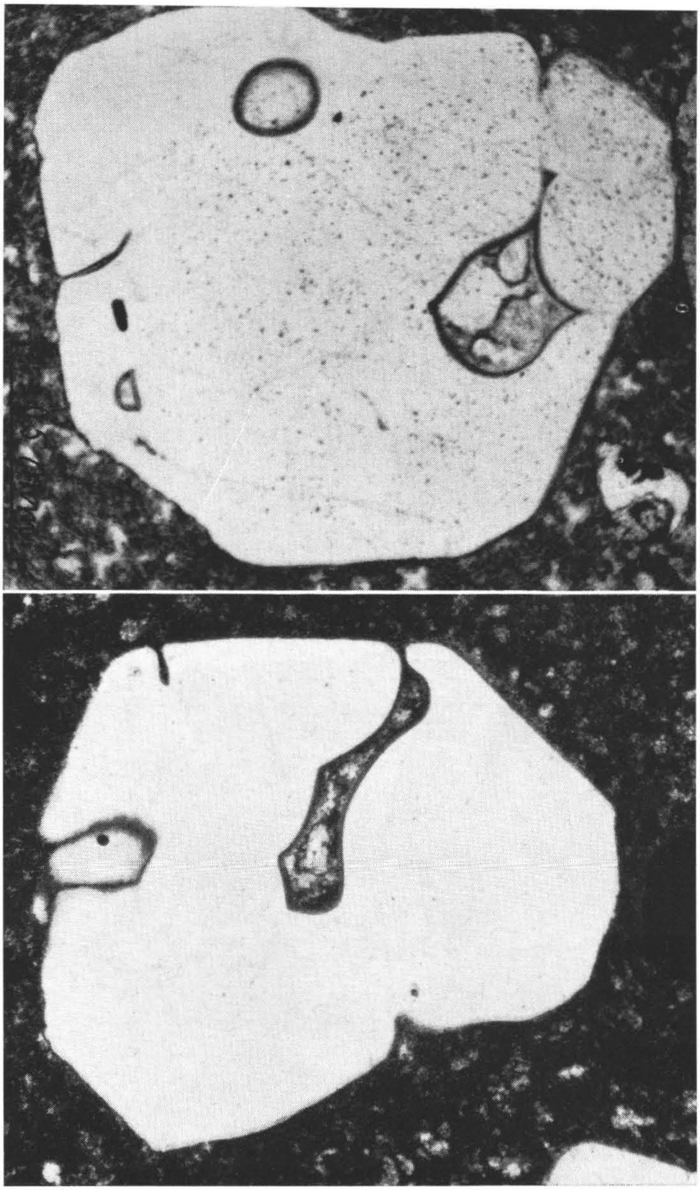

Fig. 1. Two quartz grains in granite porphyry. Within them are magma inclusions which crystallized later. Crossed nicols, 50 ×. Heinola commune. Ahvenisto area.

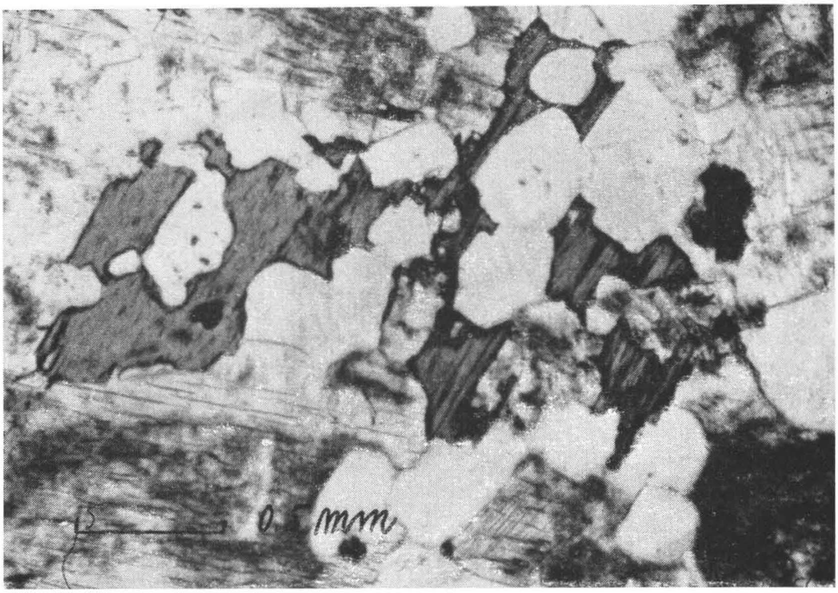

Fig. 1. Poikilite biotite grain porphyry aplite (without nicols, 40 ×). Heinola commune. Ahvenisto area.

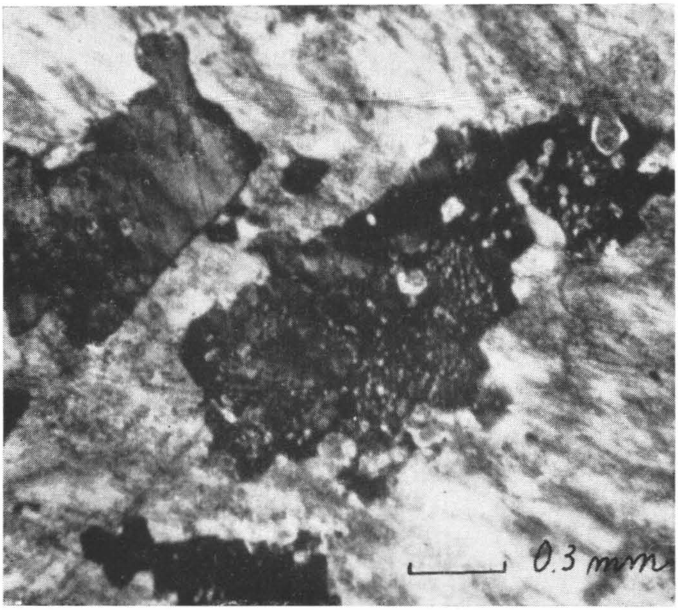

Fig. 2. Potash feldspar grain replaces and encloses other potash feldspar grains with plagioclase inclusions in rapakivi. Crossed nicols. Mäntyharju commune. Ahvenisto area.

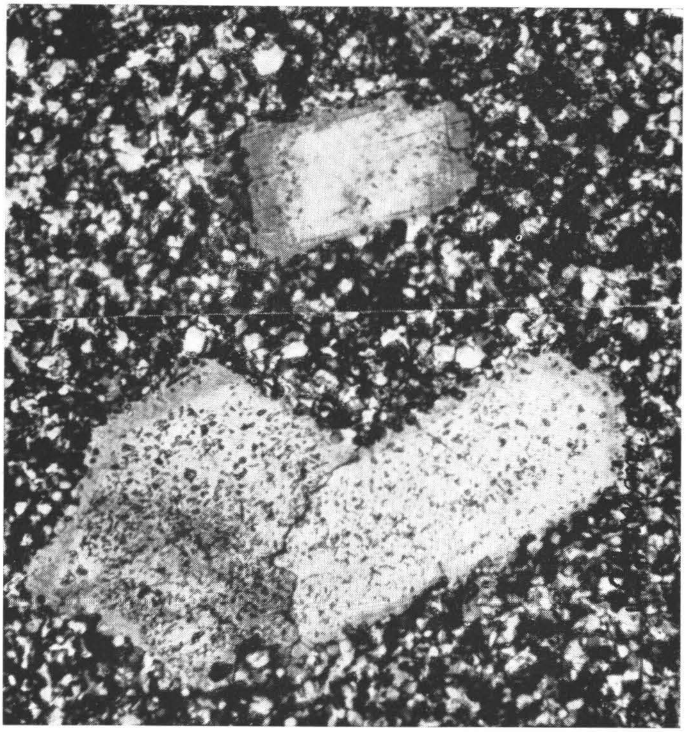

Fig. 1. Two plagioclase phenocrysts, the rims formed later, occurring as inclusions in granite porphyry. Crossed nicols, 50 ×. Heinola commune. Ahvenisto area.

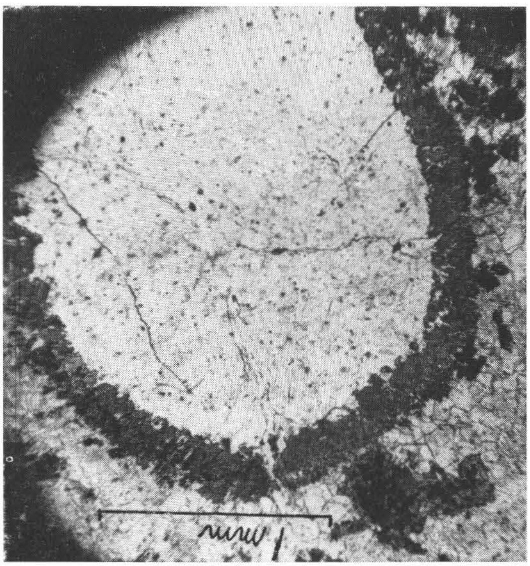

Fig. 2. Mafic minerals surround quartz (without nicols). Jaala commune. Area of Jaala—Iitti.

4

AN EXAMPLE OF ANATEXIS [1])

BY

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ABSTRACT

An example of partial anatexis in a xenolith of plagioclase-quartz-biotite gneiss is described. In the gneiss the mobilizate (plagioclase and quartz) formed segregated veinlets and patches, but at the contact against the granite it exudated into the granite. The anatexis was caused by intrusive granite rich in potassium. The melting temperature was obviously higher than the eutectic temperature at the corresponding pressure in the system Ab—Or—Q—H_2O. Comparisons with some field and laboratory studies are made. The mobilization process and the mobilization temperature are discussed.

CONTENTS

INTRODUCTION

Sederholm (1907, p. 110) originally defined the migmatites as mixed rocks consisting of two rock components of different ages. The younger

[1]) Received April 17, 1962.

component could be »either formed by the resolution of material like the first or by an injection from without». These facts as well as the diversity of the composition of the older rock component indicate processes with a great variety of combinations. The literature concerning these processes is abundant but the amount of detailed evidence collected in the field is smaller. However, simple instances studied in detail may give us material and ground for further studies.

In southern Finland the rich migmatization is mostly connected with the potash granite and the metasomatic processes caused by it.

PETROGRAPHY

In the village of Skräddarby, in the commune Sipoo, about 29 kilometers E from Helsinki, the Precambrian rock is a potash granite migmatite. The older component is a plagioclase-quartz-biotite gneiss (paragneiss). In places the gneiss is of a well preserved banded type, sometimes veined, and in places only biotite-rich nebulitic remnants of this gneiss exist in the granite. The granite is coarse of grain, like many potash granites in southern Finland, and resembles pegmatite (*cf.* Härme, 1958, p. 59; Simonen, 1960, p. 64). It behaves here like an intrusive granite.

In a highway cut the granite contains xenoliths of the aforementioned gneiss. In Figs. 1 and 2, Plate I, a fragmentary inclusion of irregular form is surrounded by a biotite-rich rim of nearly equal thickness. In addition, some pale-colored patches, lenses and veins of a somewhat larger grain size occur in the gneiss xenolith. The compositions of the different portions of the xenolith are presented in Table 1. Owing to the inhomogeneity of the rock, the determinations made with the point counter method are approximate only.

Table. 1. Mineral composition of various portions of the migmatitic xenolith. Skräddarby, Sipoo.

	1	2	3	4	5	6
Plagioclase	48.8	50.4	53.8	52.6	37.6	24.8
Quartz	24.7	33.1	33.4	43.7	16.3	11.4
Biotite	24.1	12.2	5.3	3.3	44.3	63.4
Almandite	0.1	4.1	5.6	0.2	0.5	0.1
Carbonate	0.1	0.0	0.6	0.2	0.0	0.0
Accessories	2.2	0.2	1.3	0.0	1.3	0.3
	100.0	100.0	100.0	100.0	100.0	100.0
Plagioclase	An_{35}	An_{32}	An_{37}	An_{35}	An_{32}	An_{32}

1. The gneiss in the center of the xenolith (Fig. 3, Plate II). The rock is even-grained and the texture crystalloblastic. Apatite, oxidic iron ore, zircon and sphene occur as accessories. Inclusions of drop quartz occur in some plagioclase grains. The largest plagioclase grains are about 1 mm in diameter.
2. A streak of coarser grain in the gneiss xenolith (incipient segregation). Apatite, oxidic iron ore, zircon and sphene occur as accessories. The plagioclase contains somewhat more inclusions of drop quartz than the gneiss proper (1). The almandite contains inclusions of drop quartz. Some small grains of the almandite occur as inclusions in the plagioclase. The biotite is partially bleached, whereat iron oxide pigment occurs between the flakes. The largest plagioclase grains are about 1.3 mm in diameter.
3. A patch of coarser grain in the gneiss xenolith (segregation). Apatite, sphene, zircon and oxidic iron ore occur as accessories. The plagioclase contains inclusions of drop quartz, as in 2. The almandite is often accompanied by the biotite. A part of the biotite is bleached and then the oxide pigment occurs between the flakes. The largest plagioclase grains are about 3 mm in diameter.
4. A vein of coarser grain in the xenolith (segregation vein). Apatite, zircon and oxidic iron ore occur as accessories. The plagioclase contains inclusions of drop quartz as in 2 and 3. The biotite is altered, as in 2. The largest plagioclase grains are about 4 mm in diameter.
5. The gneiss close to the segregation vein (the composition presents a belt about 1 cm broad beside the vein). Immediately against the vein, the rocks is very rich in biotite (Fig. 4, Plate II). Zircon, sphene, oxidic iron ore and apatite occur as accessories. Inclusions of drop quartz occur in some plagioclase grains, as in 1. Immediately against the segregated vein the biotite is bleached (Fig. 5, Plate III), in which case some iron oxide is present between the flakes. The largest plagioclase grains are about 4.2 mm in diameter.
6. The biotite-rich rim of the xenolith (restite). Zircon, oxidic iron ore and sphene occur as accessories. Inclusions of drop quartz in the plagioclase occur sparsely, as in 1. A small part of the biotite is bleached, whereat some pigment of iron oxide occurs between the flakes.

Elsewhere in the same outcrop the gneiss is in places conspicuously veined and also contains cordierite and in some cases secondary potash feldspar.

The chemical composition of the potash granite is presented in Table 2. The specimen analyzed was taken from a homogeneous portion at a distance of about ½ meter from the contact of the gneiss. The contact between the granite and the biotite-rich rim is rather sharp, but the ends of the xenoliths (in the direction of the schistosity of the xenolith) disappear as biotite-rich nebulitic schlieren into the granite (*cf.* Fig. 6, Plate III). Owing to its coarseness the planimetric composition of the potash granite is not determinable.

The potash granite consists of potash feldspar, quartz, plagioclase and biotite. Almandite and cordierite occur in varying amounts in the granite, too. Near the xenoliths the potash granite contains more quartz, plagioclase, almandite and biotite

Table 2. Potash granite. Sipoo, Skräddarby. Analyst A. Heikkinen.

Weight per cent		Mol. prop.	Weight norms		Niggli values	
SiO_2	71.21	11 868	qu	23.97	si	392.5
TiO_2	0.13	16	or	53.38	al	49.1
Al_2O_3	15.16	1 486	ab	15.77	fm	5.7
Fe_2O_3	0.16	10	an	2.67	c	3.5
FeO	0.43	59	en	0.92	alk	41.7
MnO	0.01	1	fs	0.45	qz	125.7
MgO	0.37	92	cor	1.32	k	0.76
CaO	0.59	105	mt	0.23	mg	0.53
Na_2O	1.87	301	il	0.24	o	0.13
K_2O	9.05	960	ap	0.09	c/fm	0.61
P_2O_5	0.04	3				
CO_2	0.09					
H_2O+	0.42					
H_2O-	0.07					
Total	99.60					

than the other portions, which also are more homogeneous. The potash feldspar is somewhat perthitic. The main part of it (about 3/4) gives the triclinicity value 0.0—0.2, and in only a small portion the triclinicity rises up to 0.8.

CONCLUSIONS

The primary structures show that the gneiss is of sedimentary origin. The feldspar of the gneiss was primarily just plagioclase, and the potassium was included in the biotite. This fact as well as the general character of this plagioclase-quartz-biotite gneiss show that the gneiss cannot be the source of the potash granite (see p. 119; *cf.* Härme, 1958, p. 59, 1959, p. 42; Simonen, 1960, p. 96). Here the potash granite is imported material, and evidently the emplacement of the granite took place as an intrusion of a melt.

The field observations show that near the contacts the potash granite is not any more of the primary composition but had already assimilated material from the surrounding gneisses.

The fact that the schistosity is more marked in the rim than in the center of the xenolith is rather an apparent phenomenon. The orientation of mica flakes is the same in the rim and in the center, but the lack of salic components in the rim makes the orientation more conspicuous (Fig. 2, Plate I).

It is obvious that the first recrystallization of the gneiss took place in the regional metamorphism before the emplacement of the potash granite. In this recrystallization the sedimentogenous excess of alumina made possible the formation of almandite and/or cordierite.

The slight amount of carbonate derives most likely from the potash granite. This and many other, earlier observations show that the potash granite contained water and other volatile constituents (*cf.* Härme and Laitala, 1955, p. 98).

The low degree of the triclinicity of the potash feldspar speaks in favor of a relatively high crystallization temperature (*cf.* Heier, 1957, p. 468; Marmo, 1958, p. 360).

The biotite-rich rim around the gneiss inclusion must have been originally of the same composition as the center. In connection with the intrusion of the potash granite, the PT-conditions changed in the xenolith. The rise of the temperature together with the volatiles of the granite caused a partial anatexis of the salic components (plagioclase and quartz) of the gneiss. At the border of the gneiss xenolith, the mobilization was strong and the main part of plagioclase and quartz exudated into the surrounding melt. Thus the biotite-rich rim (sample 6, Table 1) is a restite (*cf.* Mehnert, 1951, p. 182, 1959, p. 175; Gavelin, 1953, p. 47; Ljunggren, 1957, p. 124) or basic behind (*cf.* Read, 1957, p. 353).

The formation of the veins and patches of a larger grain size (samples 2, 3 and 4, Table 1) in the gneiss xenolith belongs plainly to the same mobilization phase. In the middle of the xenolith the mobilized salic matter segregated into patches and veinlets.

The bleached biotite in the segregated portions as well as at the border against the segregations indicates that the biotite gave material to the formation of the almandite occurring in the mobilizate. Obviously a part of the potassium of the altered biotite also lowered the melting temperature in the mobilization process (*cf.* p. 122).

DISCUSSION

The principal processes occurring in the migmatites are granitization and metamorphic differentiation. I have earlier, together with Laitala (Härme and Laitala, 1955), described a basic (gabbro) xenolith affected by the potash granite. The most conspicuous feature in that case was the migration of the potassium into the xenolith, whereat the sodium content of the xenolith also slightly decreased. In the present example a silicic gneiss xenolith is in the potash granite and the essential phenomenon in this case is the mobilization and segregation of the salic components (plagioclase and quartz) and their filtration into the potash granite. In the former example the potash granite was not able to cause anatexis in the basic xenolith and the character of the process was only that of a slow diffusion (*cf.* Härme, 1959, p. 53). In the present instance the thermal action and the volatiles of the intrusive potash granite really melted (anatexis) the salic components. The process was here more rapid than in the former case and did not attain any noticeable diffusion of the potassium into the xenolith. In addition, the gneiss contained biotite and therefore the gradient of the potash concentration between the gneiss and the potash granite was not so steep as in the former instance (see p. 122).

Several investigators stressed the rich occurrence of Al-rich minerals either in the rest (*eg.*, Mehnert, 1959, p. 176) or at the border against the migmatite front (Wegmann and Kranck, 1931, p. 62; Pehrman, 1936, p. 23; *cf.* Härme, 1959, p. 47). In the present example the determinations of the mineral compositions (Table 1) are, owing to the inhomogeneity of the rocks, not accurate enough but the example suggests that the almandite also formed in the mobilization process. Megascopically, it is often to be seen that the Al-rich minerals in the veined gneisses occur either in the veins and/or in the biotite-rich spaces between the veinlets (Figs. 9 and 10, Plate V; *cf.* Härme, 1958, Fig. 6, Pl. II, 1959, Fig. 5, Pl. I).

In the field work it has been verified in many cases how a quartz-feldspar gneiss (leptite gneiss) poor in biotite gradually passes over into a garnet- and/or cordierite-bearing gneiss. In mapping this phenomenon has caused great difficulties in the classification of rocks according to sedimentary origin. Figs. 7 and 8, Plate IV, illustrate this phenomenon well. In Fig. 7, Plate IV, the silicic leptite gneiss passes out of sight into the granite, and only some biotite-rich streaks are left in the granite. In the course of the process (Fig. 8, Plate IV) the biotite changes in the granite into almandite or cordierite (*cf.* Härme, 1959, p. 47). This explains why almandite and cordierite are so common in the potash granites of southern Finland (Härme, 1958, p. 57).

I have earlier (Härme, 1959) described examples of veined gneisses where the vein material largely derives from potash granite on the outside. In such cases the biotite-rich spaces between the veinlets are common. (Figs. 9 and 10, Plate V). Principally the process is similar to the present example: in the injection of the granitic matter the salic components (plagioclase and quartz) of the gneiss were mobilized and moved into the veinlets while biotite remained in the spaces. Primarily the composition of the injected veinlets was typically rich in potassium and poor in sodium. The partial mobilization of the wall rock (through anatexis or through the hydrothermal solutions) caused the composition of the veinlets to change so as to approach the normal granite composition. Obviously the gneiss was at that stage quite plastic.

The garnet- and cordierite-bearing gneisses of southern Finland are often called kinzigites, as proposed by Wegmann (Wegmann and Kranck, 1931). On the other hand, German geologists have stressed that the original kinzigite is a restite (*cf.*, *eg.*, Mehnert, 1959, pp. 174 and 176). Granitization and anatexis are, however, often closely connected. In the foregoing cases the

Table 3. Average (of 18 analyses) chemical composition of the potash granite (Simonen, 1960, p. 68).

Weight per cent		Mol. prop.	Weight norms	
SiO_2	73.04	12 173	qu	30.37
TiO_2	0.22	28	or	35.83
Al_2O_3	14.03	1 375	ab	23.48
Fe_2O_3	0.61	38	an	5.23
FeO	1.16	161	en	0.89
MnO	0.02	3	fs	1.29
MgO	0.36	89	cor	0.95
CaO	1.14	203	mt	0.88
Na_2O	2.78	448	il	0.43
K_2O	6.07	644	ap	0.15
P_2O_5	0.08	5		
H_2O+				
H_2O-	0.51			
Total	100.02			

biotite-rich rim around the gneiss xenolith, the biotite-rich spaces between the granitic veinlets of the veined gneisses and the biotite-rich nebulitic schlieren in the potash granite are all restites. But they are all by-products of different stages of the granitization process, too.

I want here to emphasize that the potash granites of southern Finland have two principal modes of occurrence. When it is truly intrusive then it is usually coarse-grained and resembles pegmatite (Härme, 1958, p. 59, 1959, p. 42), and it shows no orientation in the texture. On the other hand, when it is a product of metasomatism then it is often porphyritic and of oriented texture, and the orientation follows that of the paleosome. In the former case its composition is nearer to the primary magmatic composition than in the latter case.

In the average chemical composition of potash granite (Table 3; Fig. 1) the potash content is essentially higher than in eutectoid granite composition (*cf.* Simonen, 1960, p. 70). The most analyzed potash granites have been of medium-grained (mostly more or less porphyritic) types and products of granitization. Those compositions no longer correspond to the primary composition of the potash granite melt (*cf.* p. 123).

It is not likely that the composition of the potash granite (Table 2) had been formed, at least solely, through anatexis. In that case the source rock would have had to be extremely rich in potassium, and such rocks are not known to exist in sufficient measure for the huge amounts of potash granite in southern Finland (see Härme, 1958, p. 59). The known gneisses in the area are mostly rich in sodium.

Partial anatexis in natural systems has been experimentally studied by Winkler and von Platen (1961 a and b). They obtained the result (1961 b,

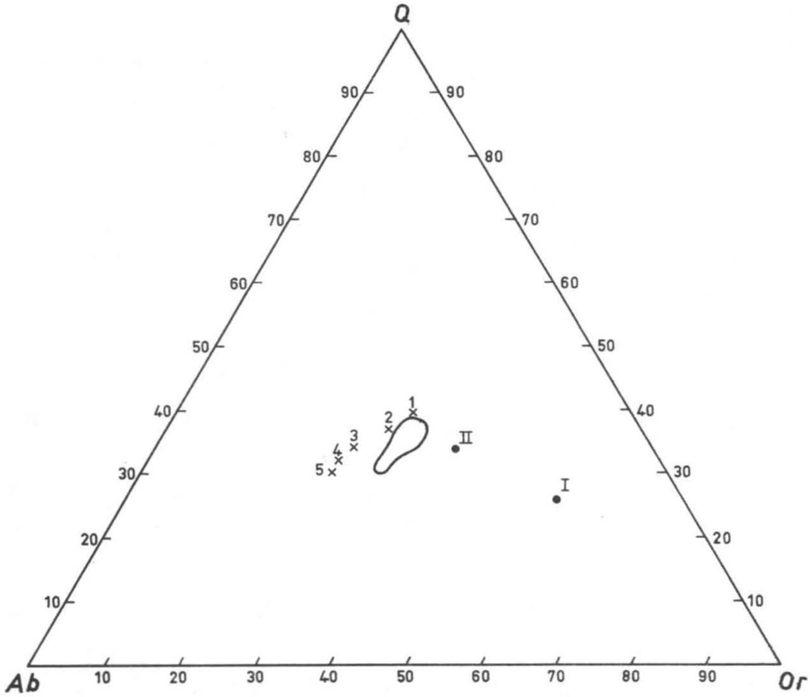

Fig. 1. Weight-normative albite-orthoclase-quartz ratios. I. Potash granite (Table 2); II. Average composition of the potash granites (Table 3) (Simonen, 1960). Crosses: the positions of eutecticums for various pressures of water-vapor, 1. 500 kg/cm², 2. 1 000 kg/cm², 3. 2 000 kg/cm², 4. 3 000 kg/cm², 5. 4 000 kg/cm² (Tuttle and Bowen, 1958, p. 75). The outlined area: the concentration maximum of granitic rocks.

p. 251) that aplitic, granitic, granodioritic and tonalitic melts are possible. The anatexis begins at temperatures between 670—740°C and at a water-vapor pressure of 2 000 atm. with the formation of a potash-rich aplitic melt. The rising temperature then causes a fractionating melting of the plagioclase, which continuously changes the composition of the anatectic melt into a granitic, granodioritic and tonalitic composition. The ratio albite: potash feldspar of the source rock is decisive to the petrochemical character of the final stage of the anatectic melt. Accordingly Winkler and von Platen conclude (1961 b, p. 252) that when this ratio is high (>0.4) the formation of great masses of aplitic melts rich in potassium are not to be expected because even a rise of about 10—40°C in the temperature changes the melt from an aplitic into a granitic composition. Therefore it is also difficult to imagine the formation of such a potash-rich melt on great scale (Table 2) from the sediments now prevailing in southern Finland (Simonen, 1953; Härme, 1959, p. 55).

On the other hand, it is just as difficult to imagine its formation solely from basaltic magma according to the classic theory of magmatic differentiation. Such huge amounts of potash granite presuppose multiple amounts of basic igneous rocks as products of differentiation, and they do not exist on such a scale in southern Finland. This theory alone is not adequate to explain the formation of potash granite on such a great scale.

The potash granite is supposed to be late-kinematic in age. The occurrence of its coarse-grained variety seems, especially in southern Finland, to have a connection with block movements. On the ground of the high potassium content and of the usually low primary quartz content the composition of the original potash granite melt resembles that of the alkaline rocks. It is thus possible, too, to think that the potash granite belongs chronologically rather to the early stages of postorogenic alkaline magmas than to the late phases of synorogenic silicic magmas. To the composition and to the manner of occurrence of the potash granites I hope to be able to return in another paper.

COMPARISON WITH SOME FIELD STUDIES

Ljunggren (1957, p. 118) verified in veined gneiss at the border of the granitic bands corresponding rims rich in mafic minerals (biotite and epidote). He assumed the metamorphic differentiation to have taken place in such a way that »the biotites and epidotes have been driven out from certain layers and have been enriched along the borders of the new layers.» He further described (p. 122) the potash feldspar as having »taken an active part in the development of the present banded gneiss.» In his view, the potash feldspar in the light-colored bands is a product of potash metasomatism which caused an intense activation of these bands.

A strong marginal alteration of a xenolith in Revsund granite is reported by Lundegårdh (1960, p. 59 and Fig. 33). According to him and Gavelin (Gavelin, 1955, p. 93) the Revsund granite is a product of potash metasomatism. Gavelin (1953, p. 47), too, further presumed that a selective mobilization takes place in connection with metasomatism.

Many beautiful examples of partial anatexis are presented by German geologists (*eg*. Scheumann, 1937, p. 405; Hoenes, 1940; Wimmenauer, 1950; Mehnert, 1951, 1953, p. 5, 1957, *cf*. 1959, p. 142, 1960, p. 335). Hoenes (1940, pp. 209 and 214) specifically called attention to the anatectic mobilization that took place in connection with potash metasomatism. — Obviously a similar phenomenon figures in the case reported by Saksela (1933, p. 53 and Fig. 14).

As early as 1913 Sederholm emphasized (p. 130) that the veins and veinlets formed through anatexis in the coastal area of southern Finland are undoubtedly in close connection with the migmatitic potash granite. This is one possibility, but it does not exclude the other possibility: a mobilization (anatexis) without the action of an outside melt.

COMPARISON WITH LABORATORY STUDIES

In the present example the mineral composition of the segregated portions (samples 2, 3 and 4, Table 1) is quite interesting. Taking into consideration the inhomogeneity of the whole xenolith, it is to be noted that in the segregated portions the ratio plagioclase:quartz is nearly constant. This fact speaks in favor of melting rather than of precipitation from the solutions.

According to the experimental studies of Tuttle and Bowen (1958, p. 52) in the ternary system $NaAlSi_3O_8$—SiO_2—H_2O at the pressure 2 000 kg/cm^2, the eutectic point is at about 750°C, corresponding to the composition 63 % Ab—37 % Q. This composition is near the ratio plagioclase:quartz in our present example (samples 2, 3 and 4, Table 1).

In the quaternary system $NaAlSi_3O_8$—$KAlSi_3O_8$—SiO_2—H_2O (Tuttle and Bowen, 1958, p. 54) at the pressure 2 000 kg/cm^2, the eutectic point is at about 680°C and the corresponding composition about 39 % Ab—27 % Or—34 % Q.

In our present instance the composition of the segregated plagioclase is about An_{30-35} (optical determinations) which should mean that in the anatexis the temperature might have been higher than the ternary minimum (*cf.* Winkler and von Platen, 1960, p. 301, 1961 a, p. 56). On the other hand, the anatectic material was composed essentially of plagioclase and quartz but, in addition to Ca, of the plagioclase also other elements had an effect on the melting in the gneiss. The gneiss contained biotite and its elements (*eg.*, K and Fe, even if in minimal amounts) correspondingly may have caused a decrease in the melting temperature.

In the present example the composition of the potash granite (Table 2, according to the norms 17 % Ab—57 % Or—26 % Q) deviates essentially from the eutectoid granite composition towards the field of the potash feldspar (Fig. 1). The same is revealed by the studies of Simonen (1960, p. 69). It is still, however, to be noted that the analyzed potash granites presented by Simonen are mostly products of potash metasomatism or at least they (the intrusive types, *cf.* p. 119) have more or less assimilated the country rocks, and thus the analyses presented do not represent the primary composition of that granite. In all probability the primary composition was particularly rich in potash feldspar and poor in plagioclase as well as in quartz, too.

Laboratory studies show (Tuttle and Bowen, 1958, p. 53) that in the system $KAlSi_3O_8$—SiO_2—H_2O the melting temperatures are of the same order of magnitude as in the system $NaAlSi_3O_8$—SiO_2—H_2O. According to the experiments of Winkler and von Platen (1961 b, p. 250): »In high-grade metamorphic rocks a partial anatectic melt is formed at a temperature of $700 \pm 40°C$ and at 2 000 atm. of H_2O-pressure whenever quartz and alkali-feldspar are constituents of the rock. If only plagioclase and quartz are the leucocratic minerals that temperature is a little higher . .».

The foregoing gives hints of the temperature of the potash granite during the anatexis caused by it, and simultaneously it shows that the temperature was, most likely, higher than the eutectic temperature in the pure quaternary system Ab—Or—Q—H_2O.

Owing to its high temperature the potash granite melt was able to cause a partial anatexis in the country rock, whereat the exudate (plagioclase and quartz) changed the composition of the potash granite melt somewhat so as to bring it closer to the eutectoid granite composition. Therefore, despite the decreased temperature, the melt was maintained, and the main part of it may have crystallized at a lower temperature.

The partial anatexis of the gneiss was a relatively rapid process and no essential diffusion of the potassium took place. First, in the changed PT-conditions the slower potash metasomatism may have made progress and continued in the hydrothermal stage (*cf.* Eskola, 1956, p. 95; Mehnert, 1959, p. 180). Owing to its greater local abundance the potash granite may then, through potash metasomatism, yield such granitic compositions as contain more potassium than the eutectoid granite (*cf.* Härme, 1958, p. 59, 1959, p. 53; Simonen, 1960, p. 71).

Recently Orville (1960, p. 104) has proved that when solutions containing alkali ions are simultaneously in contact with alkali feldspars at two different temperatures, the potassium tends to migrate toward the low-temperature and the sodium toward the high-temperature region. It has been previously emphasized by several researchers that the concentration gradient is a requisite condition for the diffusion of the alkalies in the rocks. The thermal gradient may, however, have a noticeable effect on the attainment of equilibrium. Further experiments with this hydrothermal ion exchange will surely be very significant in the study of the mechanism of alkali metasomatism.

Acknowledgments — The author wishes to express his sincere thanks to Mr. A. Heikkinen, M. A., for the chemical analysis and to Dr. K. Hytönen for the X-ray determination of the triclinicity of the potash feldspar. Credit is due to Mr. Paul Sjöblom, M. A., for correcting the English of the manuscript. The Finnish State Board of Natural Sciences (Valtion luonnontieteellinen toimikunta) supported this work by means of a research grant which is gratefully acknowledged.

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Fig. 1. Xenolith of plagioclase-quartz-biotite gneiss in coarse-grained potash granite. The xenolith is irregular in form but surrounded by a rim rich in biotite. Sipoo, Skräddarby. 1/15 natural size.

Fig. 2. The uppermost portion of the xenolith presented in Fig. 1. Some pale-colored patches and streaks in the gneiss are segregations (plagioclase and quartz). Sipoo, Skräddarby. About 1/4 natural size.

Fig. 3. Plagioclase-quartz-biotite gneiss in the middle of the xenolith. The schistosity is weak but discernible. Sipoo, Skräddarby. Diagonal nicols. Magn. 5 x. Photo: E. Halme.

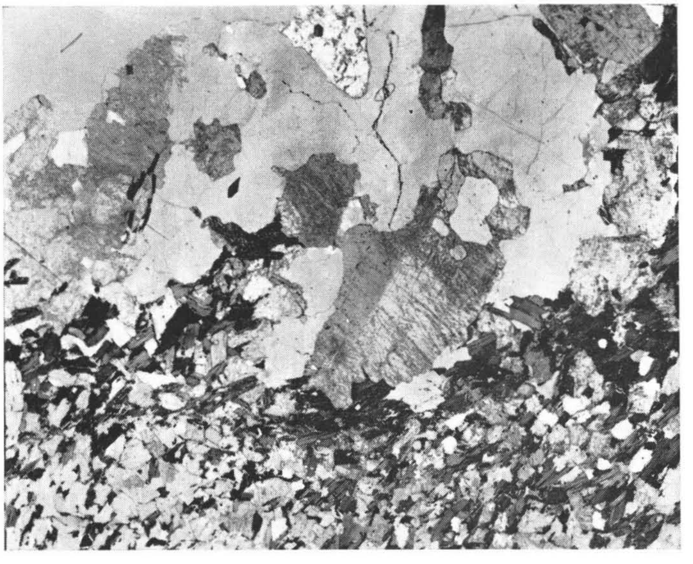

Fig. 4. Contact between the gneiss and the segregated vein. At the contact the gneiss is rich in biotite. Sipoo, Skräddarby. Diagonal nicols. Magn. 5 x. Photo: E. Halme.

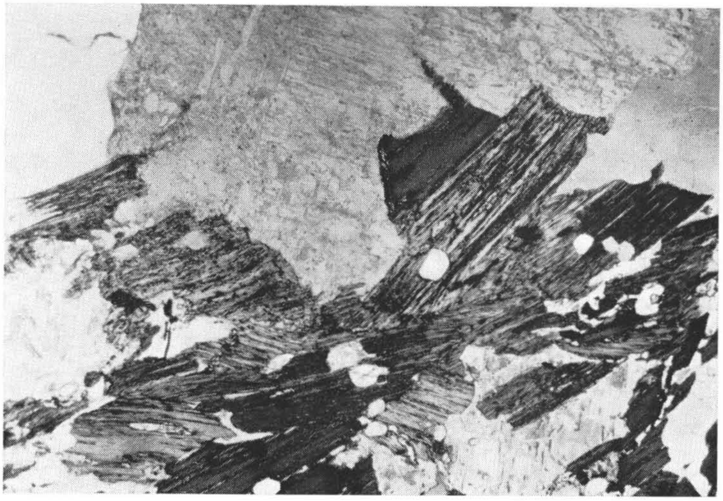

Fig. 5. Bleached biotite at the very contact against the segregated vein. Pigment of iron oxide occurs between the flakes of the altered biotite. Sipoo, Skräddarby. Diagonal nicols. Magn. 20 x. Photo: E. Halme.

Fig. 6. Xenoliths of the veined gneiss disappear as schlieren into the coarse-grained potash granite. Sipoo, Skräddarby. About 1/10 natural size.

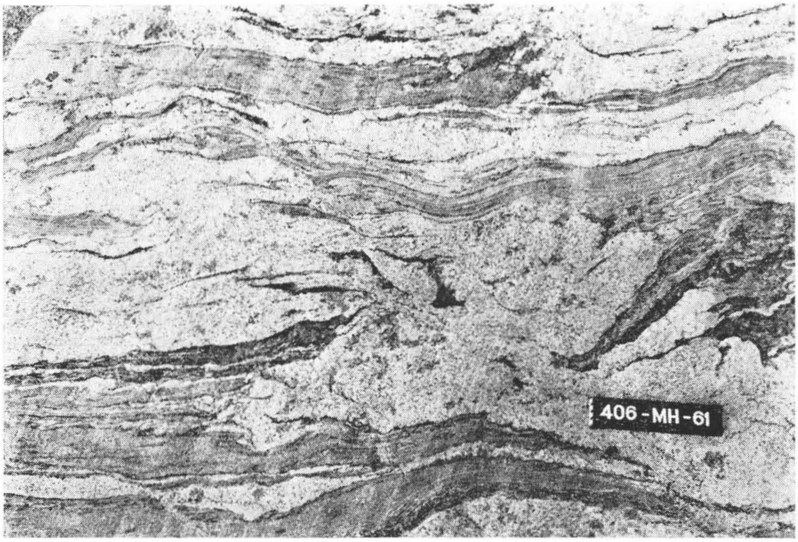

Fig. 7. Leptite gneiss relatively poor in biotite disappears into the granite. Only a few biotite-rich streaks are left in the granite. Seutula airport. About 1/5 natural size.

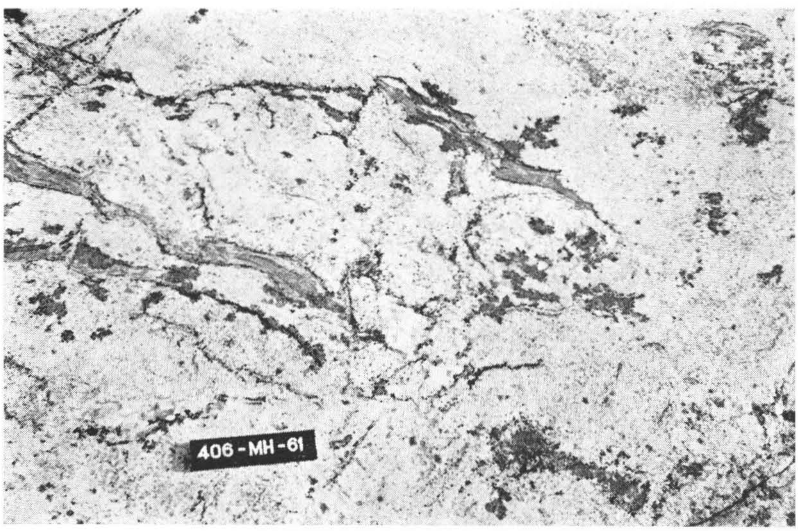

Fig. 8. Biotite-rich remnants of leptite gneiss change into cordierite in the granite. The dark roundish dots are cordierite. Seutula airport. About 1/7 natural size.

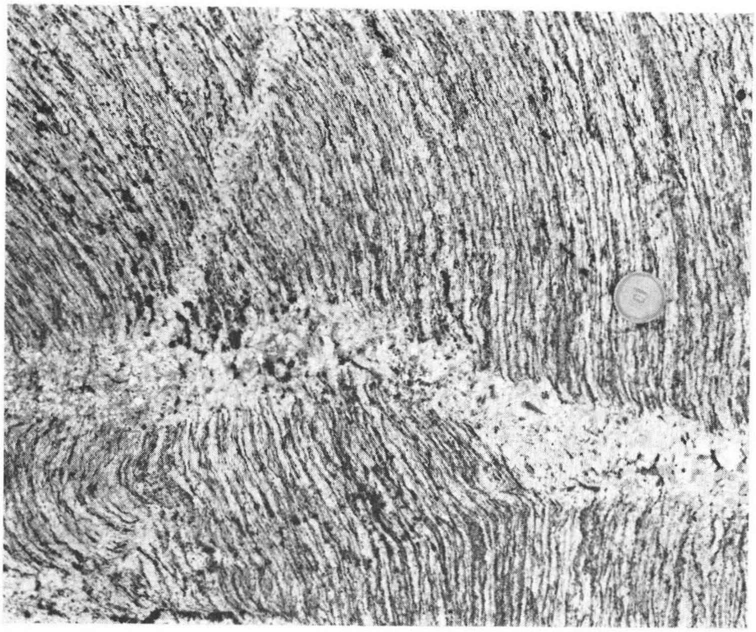

Fig. 9. Veined gneiss. Thin streaks of biotite exist as remnants of the gneiss between the granitic veinlets. Degerby, Päivölä. About 1/3 natural size.

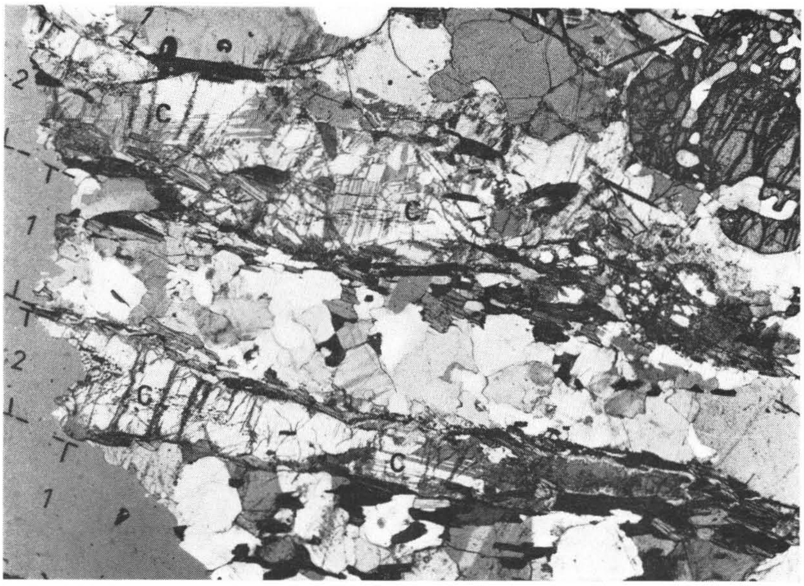

Fig. 10. Veined gneiss. Granitic veinlets (1) alternate with streaks (2) of biotite (black) and cordierite (c). Tuusula, Lahela. Diagonal nicols. Magn. 14 x. Photo: E. Halme.

5

THE QUARTZITE AREA OF TIIRISMAA [1])

BY

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ABSTRACT

The petrography and structural features of the Tiirismaa quartzite area are described.

CONTENTS

INTRODUCTION

The Tiirismaa quartzite occurrence begins some 5 km northwest of the town Lahti (Fig. 1) and continues as a belt 7 km long westward along the northern slope of Salpausselkä. The quartzite area comprises three consecutive hills separated by rather deep valleys: Rautakankare, Tiirismaa

[1]) Received April 20, 1962.

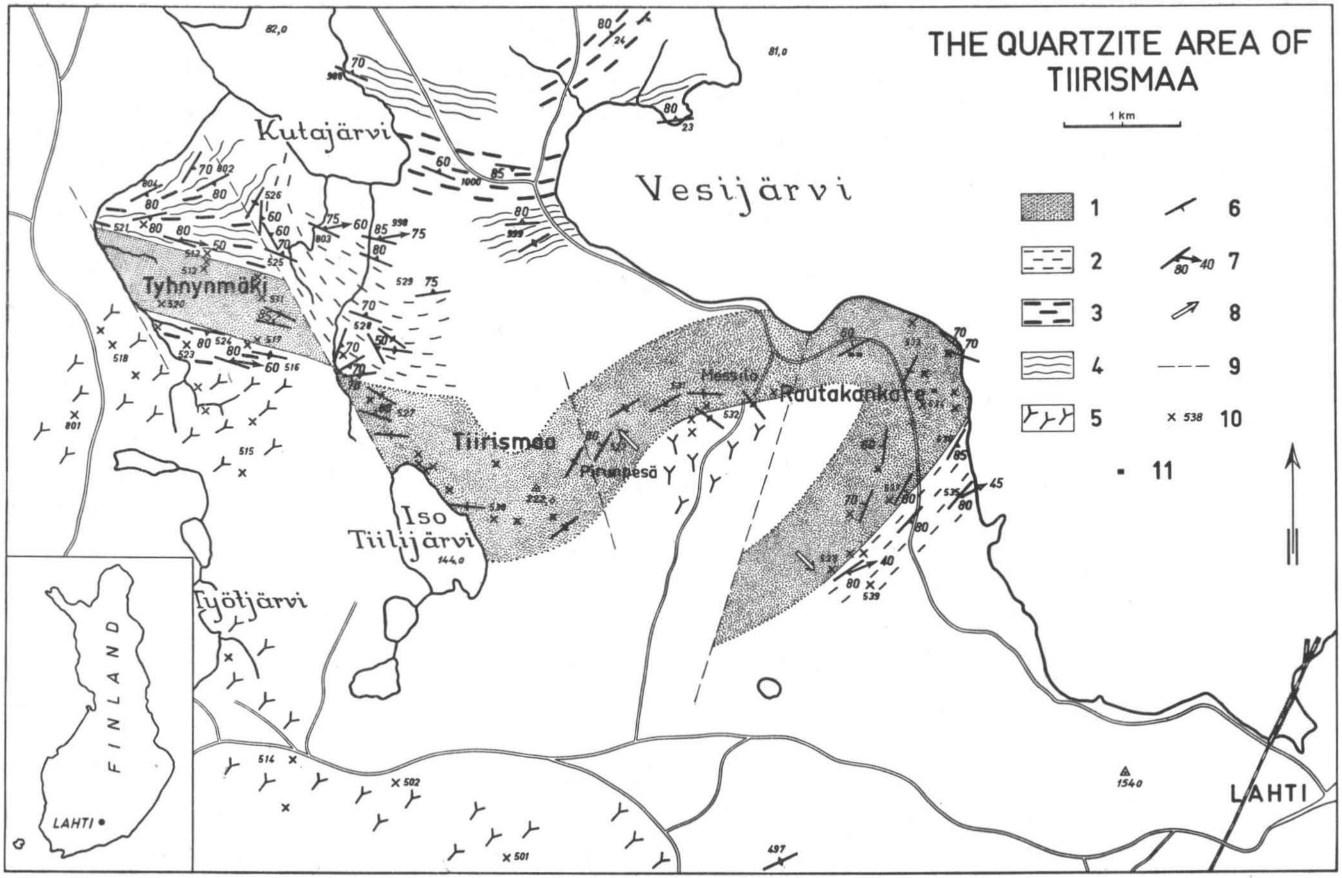

Fig. 1. Geological map of the Tiirismaa area. 1. Orthoquartzite. 2. Acid gneiss. 3. Amphibolite and hornblende gneiss. 4. Mica gneiss. 5. Quartz diorite and granite with microcline phenocrysts. 6. Strike and dip of stratification. 7. Foliation and lineation. 8. Top of stratification. 9. Fault. 10. Outcrop. 11. Quarry.

and Tyhnynmäki. The highest summit of Tiirismaa rises more than 140 m above the water level of adjacent Vesijärvi [lake].

Eskola and Nieminen (1938) described petrologically and geologically the Tiirismaa quartzite area, and its petrofabric was studied by Hietanen (1938). The present author carried out investigations in the Tiirismaa area in connection with the geological mapping of the map sheets of Kärkölä (2133) and Lahti (3111). During this work such new exposures were discobered as warrant revisions in the former conception of the mapped area. In the summer of 1961 I remapped the area with the aid of a grant from the Finnish Scientific Research Council (Valtion luonnontieteellinen toimikunta). An aerial map on a scale of 1: 20 000 was used as the base map.

QUARTZITE

The orthoquartzite of Tiirismaa is mainly reddish in color, owing to its hematite pigment. Marked metamorphism has almost completely destroyed the clastic texture. The quartz grains are sometimes much elongated. The grain size averages 1—5 mm. The boundary line between the grains is dented in places, and in some places granulated quartz fills the spaces between larger grains. The accessories consist mainly of sillimanite, which in all probability derived from primary kaolin during metamorphism. It generally forms coherent fascicles between the quartz grains and thus occurs as fibrolite; but sometimes it also appears as small needles within the quartz grains. In places the sillimanite may form large crystals. Another significant accessory is magnetite, which usually occurs arranged in definite horizons and produces distinct bands in the rock. This bandedness no doubt represents the original stratification of the rock. In places minute amounts of sericite, tourmaline and zircon are found.

Some pegmatite veins, which intersect the quartzite, are met with in the shore cliffs of Vesijärvi in Rautakankare.

Arkose quartzite forms thin intercalations here and there in the belt of acid gneisses which are limited by the orthoquartzite proper. The most remarkable of these occurrences is the one northeast of Tyhnynmäki at observation point 525. The grain size of this arkose quartzite is 1—3 mm. Potash feldspar is ragged, with abundant quartz inclusions. Sillimanite forms large nodules. In addition, it contains cordierite, colorless mica, magnetite and apatite. The mineral assemblage of the intercalating gneiss is: quartz, plagioclase, potash feldspar, hornblende, biotite, titanite, magnetite, apatite, calcite and zircon.

Although the Tiirismaa quartzite is strongly metamorphosed, it has retained features of its original sedimentary character. The banded structure

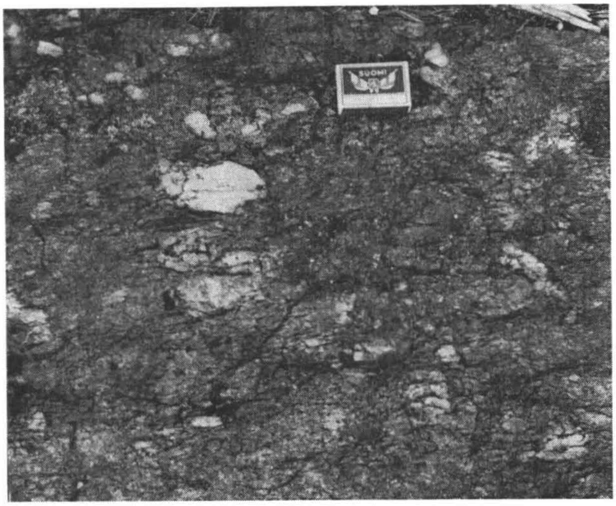

Fig. 2. Quartz pebbles in quartzite which contains abundant sillimanite. Tiirismaa, Pirunpesä. Observation point 522.

is caused by magnetite-rich layers. This, no doubt, represents the original bedding, and the bedding marked on the map was measured from the banded rock. In Pirunpesä pure quartz is present as some 5-cm-thick nodules in coarse-grained reddish quartzite containing abundant sillimanite (Fig. 2). The bed containing quartz nodules is a few meters thick and can be followed for long distances, and its strike conforms with that of the banded rock.

Some steep walls along the southeastern slope of Pirunpesä bear ripple marks (Fig. 3). They occur on the surfaces of the layers and, accordingly, could be true ripple marks and not tectonic ones. They are, however, so poorly preserved that the top direction cannot be measured with certainty. The current bedding occurring in many places yields, on the other hand, measurable top directions. At Pirunpesä the trend is northwest, and at Rautakankare at observation point 538, southeast. The quartzite of Rautakankare in many places shows cracking conformable to the current bedding (Fig. 4).

At the western end of the Tyhnynmäki area the quartzite contacts are evident on both sides. The quartzite here forms a 500—600 m thick bed, which, if compared with the original bedding, is vertically situated. Toward the east the contact with quartzite can be drawn near Messilä and at the southeastern border of the occurrence at Rautakankare with great reliability, but in both places only on one side of the bed. It seems plausible that the thickness of quartzite bed is the same also in the eastern parts of the area at Tiirismaa and Rautakankare as at Tyhnynmäki. At its eastern end the quartzite bed has turned back whereat the same horizon of acid gneisses

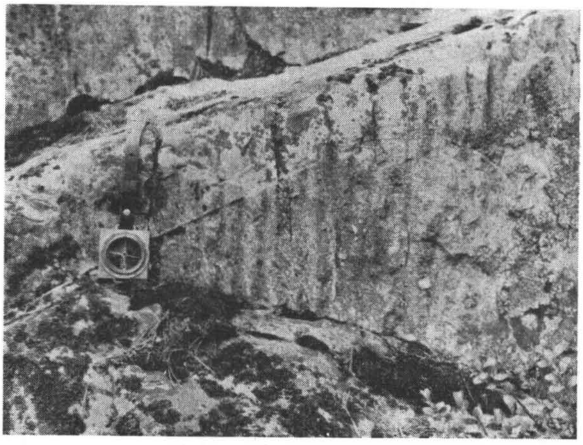

Fig. 3. Ripple marks. Tiirismaa, Pirunpesä. Observation point 522.

Fig. 4. Cracking conformable to current bedding. Rautakankare. Observation point 537.

which occurs at Tyhnynmäki, north of the quartzite, is met with at the southeastern quartzite border at Rautakankare.

A long fracture valley runs across Iso Tiilijärvi [lake] in the direction N 30° W and intersects the quartzite. Another distinct fracture valley appears also E of Messilä in direction N 20° E and at Pirunpesä in direction N 20° W. Lack of suitable outcrops prevents the determination of possible faults.

ACID GNEISS

Heterogeneous acid gneiss occurs east and northeast of Tyhnynmäki. It is strongly tectonized (Figs. 6, 7, 8 and 9). Light minerals are quartz and plagioclase, sometimes also potash feldspar. Of the mafic constituents, hornblende and biotite often occupy different layers. Occasionally pyroxene-bearing layers are met with. The abundance of epidote and titanite is a characteristic feature, and they, too, are usually concertrated in certain layers.

The mineral assemblage of gneiss at observation point 998 is: quartz, plagioclase (An_{50}), potash feldspar, biotite, hornblende, ore minerals, titanite, epidote, calcite and apatite. The biotite and hornblende are quite pale in color. Hornblende is scarcer and it appears mostly in layers by itself. Potash feldspar occurs in layers rich in quartz and mica, but hardly in layers bearing hornblende.

At observation point 526 the gneiss is strongly tectonized and heterogeneous. Light and dark, and greenish and reddish layers alternate on the weathered surface. The mineral composition of the reddish layer is: quartz, chlorite, epidote, plagioclase (albite), potash feldspar, muscovite, ore minerals and apatite. The dark fragments, seen in Fig. 7 contain quartz and magnetite and very little colorless mica. The light spherical fragments contain quartz and potash feldspar and some grains of colorless amphibole. One dark gray layer was observed to contain quartz grains ca. 1—3 mm in diamter and nearly colorless pyroxene as well as intergrown with it, some hornblenfe and, in many cases, rather abundant epidote. Between the large grains there are small grains of plagioclase (An_{60}) and quartz, and in such places also abundant titanite ore minerals, pyroxene, amphibole, epidote and calcite.

The mineral composition of the most common type of gneiss at observation point 528 is shown in Table 1, No. 1. Most significant of the accessories is titanite, which also appears as large ragged crystals full of plagioclase inclusions. Other accessories are epidote, apatite and tourmaline. The dark fragments contain mostly potash feldspar and smaller amoints of quartz and plagioclase among the light minerals. Their biotite and hornblende are almost colorless. Magnetite is abundant and causes the dark color of these fragments.

Table 1. Mineral composition of the rocks in Tiirismaa area. Integration stage determinations (Leitz).

	1	2	3	4	5	6	7	8
Quartz	43.8	30.6	40.9	88.5	3.0	2.6	1.7	15.2
Potash feldspar	21.0	10.9	39.9	—	—	—	3.3	4.6
Plagioclase	12.2	46.5	14.2	9.5	48.5	47.3	72.8	51.9
(An %)	(60)	(55)	(60)	(65)	(40)	(40)	(25)	(25)
Biotite	17.3	7.1	1.6	—	—	—	2.2	8.1
Hornblende	—	—	—	—	47.2	46.1	17.8	16.8
Ore minerals	4.4	3.7	3.3	—	—	3.5	—	—
Accessories	1.3	1.2	0.1	2.0	1.3	0.5	2.2	3.4
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

1. Acid gneiss. W end of Tiirismaa. Observation point 528.
2. » » Rautakankare. Observation point 539.
3. » » » »
4. Quartzitic pebble in acid gneiss. Rautakankare. Observation point 539.
5. Amphibolite. Tyhnynmäki. Observation point 523.
6. » » » » 513.
7. Quartz diorite. Tiirismaa, Messilä. Observation point 532.
8. » » » » »

Accessories are titanite, epidote and apatite. Epidote-rich layers are frequent at this observation point. They usually contain abundant titanite and calcite.

In Rautakankare, southeast of the quartzite, there occurs an acid gneiss the mineral composition of which is presented in Table 1, No. 2. The plagioclase there is partly sericitized. The biotite is pale in color. Titanite, prehnite and apatite occur as accessories. As is seen in Fig. 5, this gneiss includes paler and darker layers. The pale layer is richer in quartz and its mineral composition is shown in Table 1, No. 3. Its potash feldspar/plagioclase proportion is in reverse if compared with the dark part, the mineral composition of which is shown in Table 1, No. 2. A distinctly stratified quartzite pebble is to be seen in Fig. 5. The size of the quartz grains there is about 1—3 mm. The mineral composition of this pebble is shown in Table 1, No. 4. Accessory minerals are biotite, apatite, magnetite and titanite. At observation point 539 coarse-grained gneiss occurs as a bed some meters thick between quartzite and acid gneiss. Its main constituents are quartz, potash feldspar, plagioclase, cordierite, muscovite and, to some extent, sillimanite, biotite, chlorite, ore minerals, prehnite and apatite. At observation poinst 535 and 536 on the shore of Vesijärvi the rock is similar to the above except that in places it contains quite abundant epidote.

As mentioned, the quartzite is limited by acid gneisses in places where entirely quartzitic beds occur, as in the northern part of Tyhnynmäki and southeastern part of Rautakankare. These acid gneisses lie stratigraphically upon quartzite. Particularly at observation point 539 in Rautakankare, the

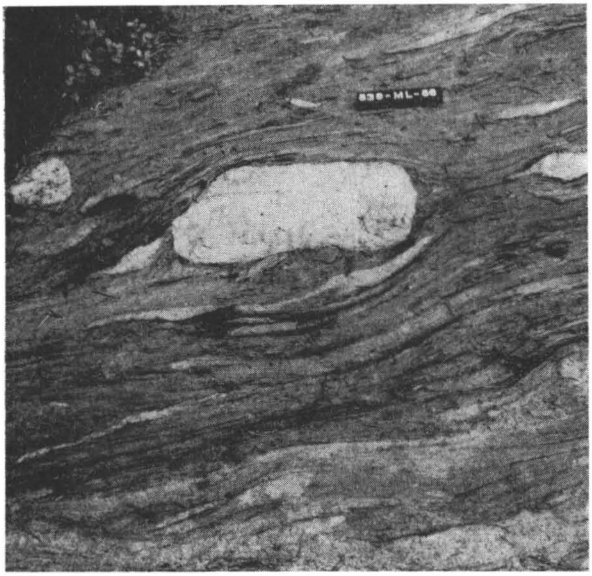

Fig. 5. Quartzitic cobbles in acid gneiss. Rautakankare. Observation point 539.

Fig. 6. Mica-rich plastic material has intruded between the cobbles of quartzitic material. W end of Tiirismaa. Observation point 528. Scale 1:6.

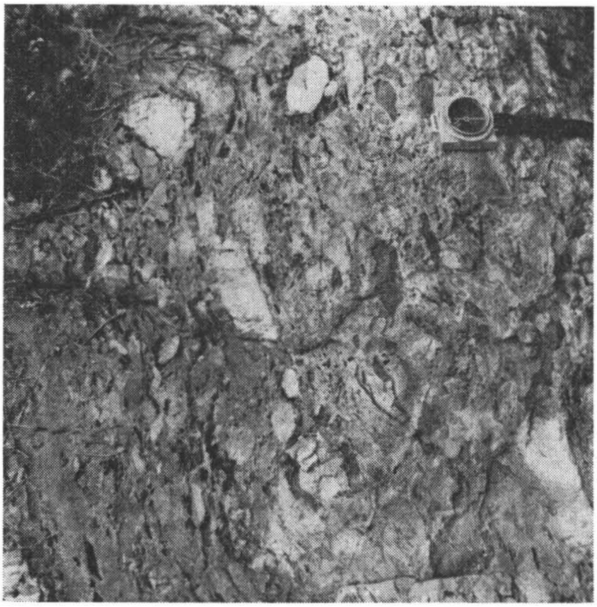

Fig. 7. Pale and dark fragments in acid gneiss. N of Tyhnynmäki. Observation point 526.

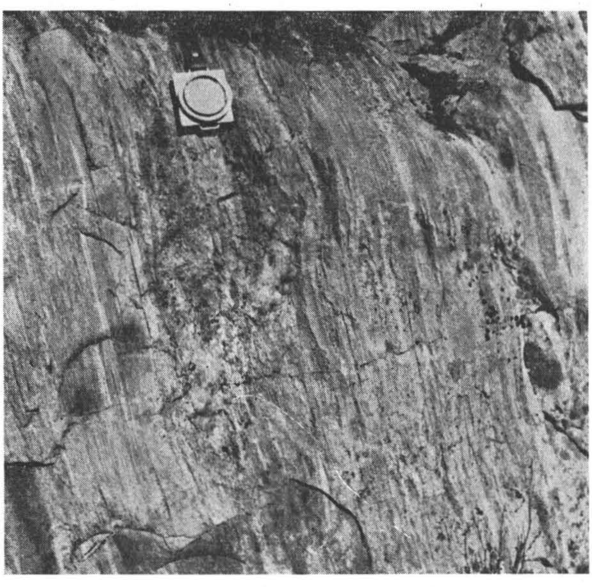

Fig. 8. Acid gneiss, perpendicular to the schistosity. Some 400 m S of the S end of Kutajärvi. Observation point 998.

Fig. 9. Horizontal surface of the same outcrop as in Fig. 8. Scale 1:10.

acid gneiss contains, in the zone close to the contact with quartzite, both rounded and angular cobbles, some of which are clearly stratified quartzite, some again of the vein-quartz type (Fig. 5). Eskola and Nieminen (1938) hold this formation to be a conglomerate. As shown in Fig. 5, the gneiss material is quite heterogeneous where layers rich and poor in mica and also wholly quartzitic layers alternate. The gneiss is strongly tectonized and part of the strata are broken. Fig. 6. of observation point 528 at the western end of Tiirismaa shows how the more plastic mica-rich material intruded between the cobbles formed of quartzitic layers. Also observation point 526 north of Tyhnynmäki shows pale and dark fragments (Fig. 7) in acid gneiss. Both are quartzite, and the dark fragments contain abundant magnetite. Fig. 8 is from observation point 998 south of Kutajärvi, showing a vertical wall perpendicular to the schistosity in acid gneiss. Owing to the heterogeneity of the gneiss material, dark and pale layers are seen there. In the photograph taken from the horizontal surface of the same outcrop (Fig. 9), cross-sections of these layers are spherical. There are also large and small angular fragments of a material resembling vein-quartz.

AMPHIBOLITE

Amphibolite occurs on both sides of the quartzite at the western end of the Tyhnynmäki area. At the southern border of the quartzite, the nearest outcrops are of heterogeneous granite and amphibolite, whose mineral composition is presented in Table 1, No. 5. (Eskola and Nieminen (1938) described a green schist from this area. In the thin sections I investigated, I found only one with some 5 % of chlorite and epidote, but even there the composition of the plagioclase was An_{40}, as it is in the amphibolite. North of Tyhnynmäki the quartzite is bordered by an amphibolite similar to that in the foregoing (Table 1, No. 6)

BIOTITE-HORNBLENDE GNEISS

As for the area north of Tyhnynmäki, where the outcrops adjacent to the quartzite were amphibolite, the main part is fine-grained, dark biotite-hornblende and biotite-plagioclase gneiss. It is quite typical of it that, even in the same thin section, hornblende and biotite alternate as mafic minerals. Sometimes also titanite is arranged in layers. At observation point 519 this gneiss displays pale layers some cm thick where the predominant minerals are quartz and plagioclase (An_{35}) and accessories are potash feldspar, biotite and ore minerals. Occasionally such an acid layer may contain slightly more biotite and perhaps hornblende, both nearly colorless. Accessories are titanite, apatite and ore minerals. The skarn-like nodules, which appear in places,

Fig. 10. Quartz diorite containing phenocrysts of potash feldspar. About 1 km S of observation point 497. Okeroinen.

are composed of andradite, diopside, calcite, plagioclase, epidote, titanite and quartz.

Observation point 516, south of Tyhnynmäki, displays biotite-hornblende gneiss.

Observation point 999, between Lakes Kutajärvi and Vesijärvi is mica gneiss, where biotite in some layers has completely chloritized. Abundant tourmaline occurs in places in this gneiss.

QUARTZ DIORITE AND GRANODIORITE

Some exposures at the southern border of the Tiirismaa quartzite near Messilä are of gneissose hornblende-quartz diorite, the mineral composition of which is presented in Table 1, Nos. 7 and 8. The Pb^{207}/Pb^{206} age for zircon is 1885 m.y. (Kouvo, personal communication). The contact between the quartzite and quartz diorite is not visible. This quartz diorite is in all probability associated with the great massif south and southwest of Tiirismaa, the closest outcrops of which occur some 4 km from Messilä. This rock shows a distinct parallel texture, and potash feldspar appears there as phenocrysts (Fig. 10). The groundmass is, however, of a quartz diorite similar to that near Messilä.

South of Tyhnynmäki the rock is a coarse-grained, weakly schistose granodiorite, where potash feldspar occurs sporadically as phenocrysts. In places the mineral composition may be entirely granitic, yet it differs in appearance distinctly from the migmatitic microcline granite of the surroundings.

SUMMARY

Quartzites are very rare in the Svecofennidic belt of southern Finland, the most significant of them being the Tiirismaa area described in the foregoing.

According to the terminology of Pettijohn (1957), the Tiirismaa quartzite is an orthoquartzite. Its red color, caused by hematite pigment, indicates a warm climate during the weathering of the rock material (Eskola, 1961). Original sedimentary structural features remain preserved in places in this quartzite, as bedding, current bedding and probably also ripple marks. The top direction could be measured by means of current bedding, and accordingly the quartzite lies lowermost with acid gneisses overlying it. This gneiss is strongly tectonized and displays clearly how layers rich and poor in mica and also wholly quartzitic layers have been broken and form elongated, sometimes spherical fragments. Bedded structure is clearly to be seen in some

of these spherical fragments (Fig. 5). As to their mineral composition, they are arkoses or arkose quartzites. Some of the pebbles are of the vein-quartz type. No such pebble was found as could be identified with certainty to be of the sillimanite-bearing orthoquartzite of Tiirismaa.

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ON THE SIGNIFICANCE OF SOME TEXTURAL AND COMPOSITIONAL PROPERTIES OF THE MAGNETITES OF TITANIFEROUS IRON ORES [1])

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ABSTRACT

This study concerns magnetites of titaniferous iron ores situated in association with subsilicic rocks in diverse formational environments. As shown by the results obtained, the textural features, the contents of V and TiO_2 and the ratio $FeO : Fe_2O_3$ of these magnetites offer useful criteria for correlations tending to reveal the primary genetic character of the respective occurrences. In the light of previous reports, the opinion has been advanced that the granular aggregates of magnetite and ilmenite in the titaniferous iron ores of magmatic origin form as a result of the exolution of originally high-temperature cubic solid solutions. The possibility of using the ratio $FeO : Fe_2O_3$ for relative geologic thermometry has been proposed.

CONTENTS

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[1]) Received May 20, 1962.

INTRODUCTION

The material presented in this paper was investigated during the years 1954—56 with the aim of producing a compositional study on different titaniferous iron ore prospects and deposits known in Finland. The magnetite component of these ores, however, did arouse a special interest in the material, which was then extended to include some foreign specimens as well. A more detailed estimation of the results and observations was postponed by some years, though re-determinations and complementary investigations of the material were carried out in line with the opportunities. The break in the work has been only fortunate, however, as in the meantime several new, meritorious investigations on the magnetites of titaniferous iron ores have become available. Among these especially the publications by Ramdohr (1953), Vincent *et.al.* (1954), Buddington *et.al.* (1955), Vincent (1957 and 1960) and Wright (1959 and 1961) as well as the results of experimental works performed by several other investigators have supplied the present authors with many new data and ideas.

The particular aim of this paper is elucidate the properties and constitution of titaniferous magnetites and on the basis of the facts presented, to use this mineral for genetic interpretation of titaniferous iron ore deposits. In view of the latter aspect, in especial it has seemed necessary, in regard to at least the Finnish samples, to describe briefly the main geological character of the respective occurrences. In this connection, it must be pointed out that the specimens designated as being from the USSR are from occurrences situated in the part of Karelia that belonged to Finland before World War II.

The main part of this work has been completed by using the facilities of the Geological Survey of Finland. The first author is responsible for collecting the material, performing the X-ray determinations, carrying out the geological study and writing the text. The second author has done the analytical work and calculated the values for the tables and diagrams.

DESCRIPTION OF THE LOCALITIES AND THE ORE MICROSCOPIC OBSERVATIONS

In the following, the most characteristic ore microscopic properties of every specimen investigated during the present study will be presented. Further, a short description of the lithologic association of the occurrences involved will be given according to the available data.

ATTU

A titaniferous iron ore mineralization of small dimensions occurs on the Jermo peninsula on the shore of the Finnish Gulf in southwestern Finland. The mineralization is associated with differentiated series of schistose gabbroic rocks (Pehrman, 1927). Among these rocks, Pehrman has distinguished the following varieties: noritic hornblende-gabbro, garnet-bearing olivine-hornblende-gabbro, cummingtonite-hornblende-gabbro, pegmatitic hornblende-gabbro and hornblende-gabbro. The ore concentrations mainly follow the hornblende-gabbro zones.

The specimens here studied represent a high-grade ore in which magnetite is the dominating constituent. Ilmenite occurs as large anhedral individuals or as lamellae in the magnetite. The lamellae are arranged in two sets, showing a difference in size and direction. In places the lamellae of ilmenite in magnetite seem to be connected with the adjacent ilmenite grains (Pl. I, Fig. 1) — obviously in a way similar to that recently described by Wright (1961) exhibited by samples from Schmoo Lake. These phenomena, as pointed out by Wright, may support the view that granular aggregates of titaniferous magnetite and ilmenite have originated by a process of exsolution.

The presence of streaky or spotted spinels following the octahedral directions of the magnetite is a common feature. Concentrated mostly within the ilmenite lamellae, there occur in places abundant half-opaque flecks, which are probably composed of rutile.

OTANMÄKI

The titaniferous ore deposit of Otanmäki in eastern central Finland is at present being mined. The general geology of the occurrence and of the surrounding areas have been described by Pääkkönen (1956) and Paarma (1954 and 1960). The main ore-mineralogical properties and genetical features of the ore have been described by the first author (Vaasjoki, 1947) and the mineralogy still later by Ramdohr (1956). The Otanmäki deposit is associated with a basic rock belt stretching about 30 kilometers in length. The zone is dominated by amfibolitic rocks accompanied by some gabbroic and anorthositic rock varieties. The deposit comprises numerous ore lenses which may reach a length of 200 meters and a width of 30—50 meters. In the high-grade ore, chlorite, hornblende and some plagioclase occur as silicatic material. On the basis of the available data, it seems that the crystallization of the ore had occurred in conditions corresponding to those of the amfibolite facies (Vaasjoki, 1947).

The samples studied in this connection belong to the high-grade ore. The essential ore minerals are magnetite and ilmenite, which appear as separate grains. Some pyrite and chalcopyrite have been met with as accessories, in addition to which some högbomite has been recognized in places (Vaasjoki, 1947; Ramdohr, 1956). In the magnetite component of the ore, irregularly shaped flecks of ilmenite are common while the lamellae of ilmenite are rare. Small streaks and spots of spinel occur following the octahedral directions of the magnetite.

In the ilmenite there usually occur two generations of exsoluted hematite. Magnetite, which appears as fine streaks like the exsoluted hematite, has been first identified in this ore by Ramdohr (1956). In places a pronounced twinning of the ilmenite has been observed.

RIUTTAMAA

The small, non-economic titaniferous iron ore concentrations are emplaced in a zone of gabbroic rocks, which for a distance of four kilometers enclose another occurrence of titaniferous iron ore — the Susimäki deposit. Accordingly, the main geology in Riuttamaa is analogical to that of the Susimäki region, which will be demonstrated later on in this paper.

The specimens studied represent an average ore. The magnetite and ilmenite occur as a granular aggregate, although lamellae of ilmenite are met with in the magnetite, too. The magnetite is spinelliferous in a way similar to the foregoing examples. This is to be seen in Pl.I, Fig. 2, which further shows that the amount of spinel diminishes while approaching the border of the magnetite grain. On the outermost border of the ilmenite grains, there usually occur small, half-opaque grains which may be assumed to be rutile.

TAHAWUS, SANFORD LAKE, U.S.A.

During his stay in the U.S.A., the first author had an opportunity in the spring of 1952 to visit the locality by the courtesy of the National Lead Co. The general geology of the area has been described by Buddington in several connections (*e.g.*, 1939 and 1952) and the ores especially by Balsley (1943). Particular mention has to be made of the fact that two main types of the ore have been distinguished by means of the titanium iron ratio in the ore; the one is characterized by the ratio Ti : Fe $>$ 1, and the other by the ratio Ti: Fe $<$ 1. In this study, the latter ore type has been analyzed.

In both ores, magnetite and ilmenite occur as granular aggregates accompanied by some sulfide minerals as accessories. In the ore with Ti : Fe $>$ 1, relatively abundant ilmenite lamellae occur in the magnetite, whereas in the ore with Ti : Fe $<$ 1 the amount of ilmenite lamellae is considerably smaller. The fine ilmenite network in the magnetite of Tahawus is shown in Pl. II, Fig. 1, which also reveals spinels obviously belonging to systems of two different generations.

ROUTIVARE, SWEDEN

The geology and ores of the Routivare region, northern Sweden, have been described by Peterson (1893) and Sjögren (1893). Both of these investigators agree that the titaniferous iron ore belonged as a member to a formation of basic rocks, which later faulted and underwent regional metamorphosis as indicated by zones of strong alterations. Ramdohr has similarly emphasized the properties which later tectonic movements and metamorphism have developed in the Routivare ore (1945). The magnetite and ilmenite in the ore occur as granular aggregates. The magnetite contains ilmenite as small irregular flecks or as lamellae. As in the foregoing samples in this study, spinel is abundantly present in the magnetite, whereby högbomite has often been identified.

SUSIMÄKI

The titaniferous ore of Susimäki, in the commune of Vampula, southern Finland, have been described by Palmunen (1925), who considers that the ore-bearing area is characterized by rocks produced by magmatic differentiation. Starting from the most basic member, he established the following rock series: Ilmenite-magnetite-olivinite, peridotite, hornblende-ilmenite-gabbro, hornblende-biotite-gabbro. This series is associated with a gabbroic rock unit covering an area of several square kilometers and including also the occurrence of Riuttamaa, as referred to earlier. The ilmenite-magnetite-olivinite corresponds to the richest ore in the area. One of the most typical phenomena in the rock, called peridodite by Palmunen, is the presence of reaction rims, i.e., synanthetic textures (Sederholm, 1916) around the olivines. These formations generally show intergrowths of silicatic material with ore oxides (Pl. II, Fig. 2) as earlier demonstrated by the first author (Vaasjoki, 1955). In the said connection, the first author has also proposed that the peridotite in question should be defined as gabbro.

In the specimens studied, the ilmenite predominates in relation to the magnetite. A few sulfide grains, mainly chalcopyrite, have been noted here and there in the specimens. The magnetite is characterized by abundant spinel. In some cases a faint texture indicating a kind of heterogeneity in the magnetite is visible as shown in Pl. III, Fig. 1. This texture has not been satisfactorily understood as yet by the authors. Tentatively it is suggested that the structure might indicate the presence of ulvite (see page 154). Presumably on the basis of microscopic observations, Ramdohr (1956) has proposed a similar possibility as regards the Susimäki magnetite.

ULVÖ, SWEDEN

The occurrences of titaniferous deposits in Ulvö belong to wider formations known by the name of Nordingrå massive. According to the investigations made by Sopral (1910) and Mogensen (1946), the ores occur as almost horizontal sheets in association with post-Jotnian dolerite dikes. Mogensen, especially, suggest that the ore be regarded as a highly ferriferous variety of dolerite. Subsequently, the ore conceivably represents a relatively rapidly cooled high-temperature formation.

The specimen here investigated is a typical high-grade ore in which the magnetite shows a dense network formed by an intergrowth of magnetite and exsoluted ulvite. However, the mineralogy of the ore, which has been thoroughly described by Mogensen (1946), needs no further comment in this connection.

VUORIJÄRVI, U.S.S.R.

The titaniferous iron ore prospects are situated in Karelia at about 66.5° northern parallel. According to the investigations by Hackman (1925), the region is characterized by calcareous alkaline rocks. The bedrock is mainly covered by a heavy overburden and indications of titaniferous iron ore have been established by boulders which occur as accumulations in situ on the hillsides of Niskavaara and Tuohivaara, near Vuorijärvi. Hackman *(op.cit.)* like Palmunen (1925) has suggested that ore concentrations in the bedrock have probably been formed by contact metamorphic processes in the calcareous wall rocks.

The main mineralogical composition of the specimens studied from the Vuorijärvi region comprises perovskite, ilmenite, magnetite, spinel and rutile, listed in the order of abundance. It seems that the perovskite has been the latest to crystallize, replacing both iron oxide minerals mentioned. Here and there occur some pyrite as an accessory ore mineral.

The abundant spinel contained in magnetite appears in two systems following as usually the octahedral directions. Irregularly shaped lamellae or flecks of ilmenite are characteristic of the magnetite. In Pl. III, Fig. 2, the mutual relations of magnetite, ilmenite and perovskite are shown from a specimen of Niskavaara. The ore of Tuohivaara is richer in rutile. According to a recent study by Verhoogen (1962), the presence of the combination magnetite + perovskite in undersaturated rocks is quite understandable because of a low silica activity and high temperature, which conceivably prevailed under contact metamorphism. Obviously the conditions of crystallization in Niskavaara and Tuohivaara have been somewhat different in character, as in the latter case the formation of rutile is more pronounced than the formation of perovskite. This possibly reveals a higher silica activity as the titaniferous mineralization in the Tuohivaara zone was formed.

VÄLIMÄKI, U.S.S.R.

The deposits of the Välimäki region are situated on the northern shore of Lake Ladoga, NE of the town of Sortavala. According to Blankett (1896), the ore bodies are associated with dike-like formations composed of basic rock, which he has called peridotite. On the basis of a detailed study, Blankett characterizes this rock as magnetite-diallage-olivinite, which alters into diallage-amfibolite towards the dike borders. The dikes measure 50—300 meters in width and occur in connection with an extensive, strongly altered massive of gabbro and gabbro-diorite. The titaniferous iron ores are present as pockets or as vein-like formations in the central and more basic parts of the dikes. The gabbro-gabbrodiorite massive, including the dikes, borders on schists extending to the northern side of Lake Ladoga. The contacts between the basic dikes and gabbro-gabbrodiorite rocks are somewhat diffuse. Blankett has considered *(op.cit.)* that this circumstance may indicate a plasticity of the gabbro massive at the time of intrusion of the basic dikes.

The specimens studied represent a high-grade ore in which magnetite is the dominating ore mineral. Besides individual grains, ilmenite occurs commonly as lamellae in the magnetite, which also contains spinel in the customary way. Sulfides like pyrite and chalcopyrite have been observed as minor accessories.

SPECIFIC GRAVITIES AND X-RAY DETERMINATIONS

From each of the specimens studied, a fraction of magnetite was prepared by using a magnetic and heavy liquid separation. From the fractions obtained, the specific gravity was determined by using the temperature-controlled pycnometric method. The values obtained are presented in connection with the analytical results of the corresponding fractions (Table 2).

For testing a possible change in the lattice parameters of the magnetites studied, a sample of each fraction was run through an X-ray spectrometer. The measurements made by using reflections (220), (311), (400) and (333) show, however, that only the sample of Ulvö magnetite gave a distinctly deviating value for a_0. All the others, even that from Susimäki (see page 154), revealed values that may be regarded as normal for magnetites, *i. e.*, a_0 shows varieties slightly on either side of 8.400 Å.

SOME REMARKS ON THE ANALYTICAL RESULTS AND THE OCCURRENCE OF VANADIUM

The results of chemical analyses show (Table 2) that the amounts of SiO_2 generally are smaller than 1 %, except analysis 7, which represents the sample from Susimäki.

The amount of alumina usually varies between 1—3 %, except the samples from Otanmäki and Niskavaara (Vuorijärvi), which contain less than 0.5 % Al_2O_3. Apparently the amount of alumina for the most part depends on the amount of (Fe, Mg, Al)-spinel in a sample as controlled by microscopic observations. These indicate, for example, that the (Fe, Mg, Al)-spinel content in the samples of Otanmäki and Niskavaara is less pronounced.

Magnesium and manganese may substitute the two valency iron in the ilmenite, which then would be considered as a mixed crystal including some of components geikielithe and pyrophanite. On the other hand, the elements in question may, in some degree, also incorporate the spinels.

On the basis of the well-known similarities between the ionic properties of the vanadium and iron, the vanadium could be expected mainly to incorporate the magnetite in the titaniferous iron ores, a phenomenon which is confirmed also by the analytical results of the present study. It might be further suggested that the relative amounts of vanadium and iron in a titaniferous magnetite may reveal a property related to the genetic character of the ore formation concerned. For estimation of this possibility, the relationships between the vanadium and iron contents in the magnetites of the present study have been illustrated by using the vanadium module (Table 1) proposed by Carstens (1939). As defined by Carstens, the vanadium module is the percentage of vanadium contained by the total iron in a magnetite.

Table 1. The Vanadium Modules in the Magnetites Analyzed.

N:o	Location	V. Mod.	W. %		N:o	Location	V. Mod.	W. %	
			Fe	FeO				Fe	FeO
1	Attu	1.38	65.4	31.82	6	Susimäki	1.22	63.2	33.50
2	Otanmäki	0.85	70.7	30.92	7	Ulvöen	1.42	56.6	44.22
3	Riuttamaa......	0.58	63.9	31.60	8	Vuorij. N.......	0.17	66.4	25.08
4	Tahawus	0.80	67.5	31.56	9	Vuorij. T.......	0.18	62.5	28.20
5	Roudivare	0.54	66.3	31.02	10	Välimäki	0.78	64.4	31.90
					11	Jacupiranga [1]) ..	0.24	58.0	17.00

[1]) The specimen was put at the authors' disposal by courtesy of Dr. Valto Veltheim, Geological Survey of Finland, and it probably represents one of the usual titaniferous ore types of Jacupiranga District, Brazil. The original specimen has not been separated and was analyzed as such mainly as an instance of magnetite including abundant ilmenite as lamellae and irregular flecks.

The magnetite of Ulvö indicates the highest vanadium module, although followed closely by those from Attu and Susimäki. The modules for the magnetites of Otanmäki, Tahawus and Välimäki, being quite similar to each other, form the next category. Somewhat lower values are indicated by the vanadium modules of the magnetites of Riuttamaa and Routivare, and the lowest values by those of the magnetites representing the occurrence of the Vuorijärvi area.

The aforementioned instances thus reveal that the magnetites showing vanadium modules of high or moderate value are all of occurrences designated as basic dikes or which are associate with basic eruptive rock series. The last-mentioned usually indicate varying effects of metamorphism. The occurrence of Vuorijärvi, with a magnetite showing a particularly low vanadium module, represents formations that probably originated by a process of metasomatic replacement (*cf.* Hackman, 1925 and Palmunen, 1925). Accordingly, the opinion that in the titaniferous iron ores there exists an interdependence between the vanadium and iron contents in a magnetite, on the one hand, and the genetic character of the occurrence, on the other hand, seems to gain at least some support. The vanadium module obviously is a convenient term to illustrate the interdependence in question.

GENETICAL EVALUATION OF THE TEXTURAL AND COMPOSITIONAL FEATURES OF THE MAGNETITES STUDIED

The chemistry of the specimens of the present study mainly concern the system determined by compounds FeO, TiO_2 and Fe_2O_3. Corresponding to this system, the main mineralogical species observed are Fe_3O_4, $FeTiO_3$, Fe_2TiO_4 and Fe_2O_3. Which ones of these species are present especially has

Table 2. Chemical Compositions. Normative Minerals and

Component	1 Attu		2 Otanmäki		3 Riuttamaa		4 Tahawus		5 Routivare	
	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.
SiO_2	0.13	2.2	0.05	0.8	0.90	15.0	0.07	1.2	0.14	2.3
TiO_2	5.31	66.5	0.92	11.5	7.71	96.5	4.26	53.3	5.05	63.2
Al_2O_3	1.85	18.2	0.16	1.6	1.47	14.4	1.39	13.6	2.16	21.2
Fe_2O_3	58.20	364.5	66.70	417.7	56.20	351.9	61.50	385.1	60.30	379.6
V_2O_3	1.33	8.9	0.88	5.9	0.55	3.7	0.80	5.3	0.53	3.5
Cr_2O_3	0.01	0.1	—	—	n.d.	—	0.02	0.1	0.05	0.3
FeO	31.82	442.9	30.92	430.4	31.60	439.9	31.53	438.9	31.02	431.8
MnO	0.08	1.1	0.03	0.4	0.17	2.4	0.09	1.3	0.10	1.4
NiO	0.02	0.3	0.01	0.1	0.01	0.1	0.02	0.3	0.02	0.3
CoO	0.02	0.3	0.02	0.3	0.02	0.3	0.02	0.3	0.02	0.3
MgO	0.31	7.7	0.30	7.4	0.68	16.9	0.53	13.2	0.79	9.6
CaO	—	—	—	—	0.04	0.7	0.04	0.7	0.09	1.6
	99.08		99.99		99.35		100.27		100.27	
Normative Minerals	%		%		%		%		%	
Magnetite	88.57		98.10		84.36		90.53		90.80	
Ilmenite	8.77		1.56		11.35		7.40		5.25	
Hematite										
Ulvite									0.29	
Spinel	0.79		0.22		0.38		1.82		1.47	
Pyroxene	0.22		0.08		1.50		0.12		0.23	
Rutile	0.69				1.73		0.36		2.19	
Perovskite										
	99.04		99.96		99.32		100.23		100.23	
Spec. gr.	4.976		5.136		4.974		5.027		4.899	

depended on the effect of certain fundamental factors determined by the general conditions of crystallization of the lithologic environment including the titaniferous occurrence. From this point, particularly, the factors depending on the temperatures of formation as well as the effects determined by variations in the oxidation-reduction relationships during the process of crystallization have commanded much consideration and led to a number of interesting investigations. Among these, the conception by Foslie (1928) has to be considered as of pioneering importance to the problems at hand. According to him, in many cases the exsolution of ilmenite from magnetite might well be a result of subsequent oxidation of a Fe_3O_4—Fe_2TiO_4 solid solution as follows:

$$Fe_2TiO_4 + Fe_2O_3 \rightarrow Fe_3O_4 + FeTiO_3$$

Specific Gravities of Some Titaniferous Magnetites.

6 Susimäki		7 Ulvö		8 Niskavaara		9 Tuohivaara		10 Välimäki		11 Jacupiranga	
W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.	W %	1 000 x m.q.
2.15	35.8	0.91	15.2	0.53	8.8	0.12	2.0	0.08	1.3	0.27	4.5
6.10	76.4	20.26	253.6	1.54	19.3	8.84	110.6	6.68	83.6	9.71	121.5
2.00	19.6	1.49	14.6	0.45	4.4	1.28	12.6	2.65	26.0	3.15	30.9
53.10	332.5	27.60	172.9	67.10	420.2	58.00	363.2	56.60	354.4	64.10	401.4
1.13	7.5	1.12	7.5	0.16	1.1	0.16	1.1	0.72	4.2	0.20	1.3
0.07	0.5	0.06	0.4	—	—	0.02	0.1	0.01	0.1	0.06	0.4
33.50	466.3	44.22	615.5	25.08	349.1	28.20	392.5	31.90	444.0	17.00	236.6
0.20	2.8	0.38	5.4	0.57	8.0	0.51	7.2	0.01	0.1	0.46	6.5
0.02	0.3	0.03	0.4	—	—	0.08	1.1	0.03	0.4	0.11	1.5
0.02	0.3	0.03	0.4	0.02	0.3	0.03	0.4	0.03	0.4	0.03	0.4
2.15	53.3	2.89	71.7	4.34	107.6	2.75	68.2	1.47	36.5	5.18	128.5
0.05	0.9	0.15	2.7	0.33	5.9	0.09	1.6	0.04	0.7	0.04	0.7
100.49		99.14		100.12		100.08		100.22		100.31	

%	%	%	%	%	%
79.09	41.68	76.89	84.33	83.93	49.77
3.84		2.03	11.80	12.38	15.49
					29.98
11.41	52.82				
2.52	2.16	19.50	1.81	3.71	4.48
3.59	1.52	0.88	0.20	0.13	0.45
	0.90		1.61		
		0.80	0.22		
100.45	99.08	100.10	99.97	100.15	100.17
4.826	4.732	4.942	4.990	4.993	4.733

As remarked by Foslie, the ulvite component could only exist under conditions where there was no excess of Fe_2O_3 present.

The problem of the stability of ulvite was revealed in a new light when Mogensen showed (1946) that this mineral exists in the magnetites of the titaniferous iron ore occurrence in Ulvö, Sweden. Since then, Ramdohr (1953) has observed ulvite in several samples and has emphasized that the mineral is relatively common, though usually occurring in forms and dimensions difficult to detect. Besides the microscopic observations, the presence of ulvite may be suggested by an excess of ferrous iron in the analysis, or it may be revealed with X-rays by measuring the lattice parameters of a titaniferous magnetite.

The chemical interdependence of the main compounds of the magnetites analyzed in the present study have been demonstrated by means of the

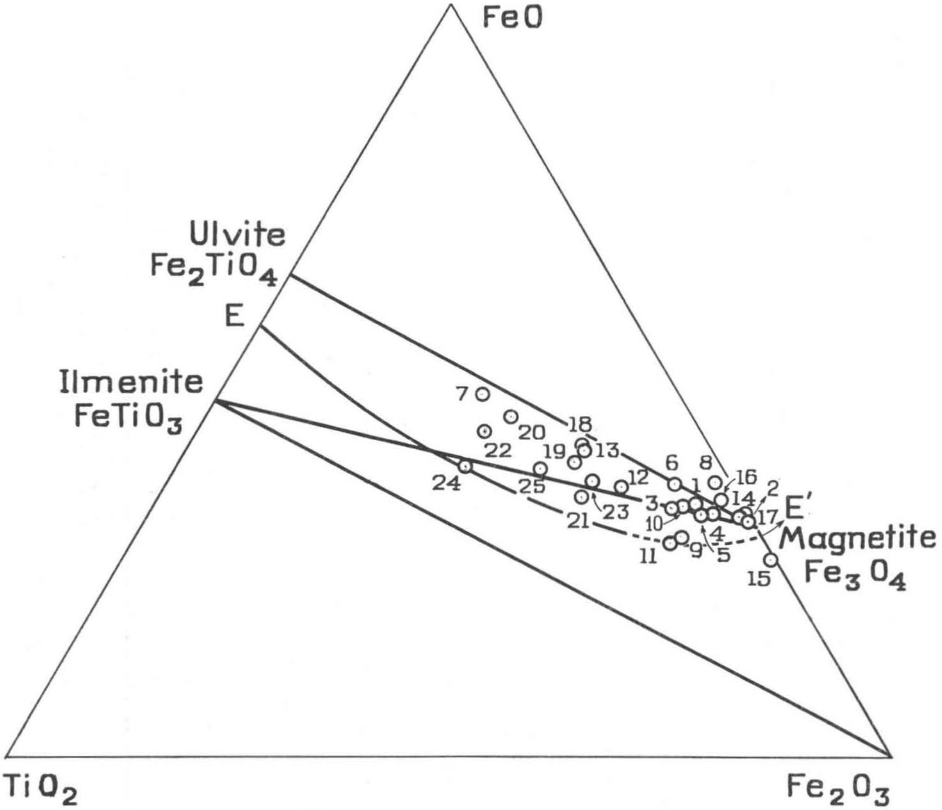

Fig. 1. $FeO—TiO_2—Fe_2O_3$ compositional diagram. The numbers refer to analyses of magnetites from various occurrences as follows:

1. Titaniferous magnetite. Attu, Finland. Analyst A. Heikkinen.
2. » » Otanmäki, Finland. Analyst A. Heikkinen.
3. » » Riuttamaa, Finland. Analyst A. Heikkinen.
4. » » Tahawus, Sanford Lake, U.S.A. Analyst A. Heikkinen.
5. » » Routivare, Sweden. Analyst. A. Heikkinen.
6. » » Susimäki, Finland. Analyst. A. Heikkinen.
7. » » Ulvö, Sweden. Analyst. A. Heikkinen.
8. » » Niskavaara, Vuorijärvi, U.S.S.R. Analyst. A. Heikkinen.
9. » » Tuohivaara. Vuorijärvi, U.S.S.R. Analyst. A. Heikkinen.
10. » » Välimäki, U.S.S.R. Analyst. A. Heikkinen.
11. » » Jacupiranga (*cf.* p. 149, footnote of Table 1). Analyst. A. Heikkinen.
12. » » Dopma, Norway. Anal. D. 10. Carstens, H (1957, p. 33).
13. » » » » » D. 18. » »
14. » » Tafjord, Øyen, Norway. Anal. M. 331. Gjelsvik (1957, p. 491).
15. » » Fiskå, Norway. » M. 852. » »
16. » » Verkshaugen, Norway. » M. 868. » »
17. » » Fiskå, Norway. » M. 1046. » »
18. » » Røddal, Norway. » SM. 1106. » »
19. » » Skaergaard, Greenland. » 1. Vincent and Phillips (1954, p. 12).
20. » » » » » 2. » » » »
21. » » » » » 4. » » » »
22. » » » » » 5. » » » »
23. » » » » » 6. » » » »
24. » » » » » 7. » » » »
25. » » » » » 8. » » » »

diagram shown in Fig. 1. The plots of the diagram correspond to weight percentages, which have been corrected in regard to the MgO-, MnO- and V_2O_3- contents in the respective analyses. To afford an opportunity for direct comparison, analyses of some Norwegian titaniferous magnetites (Carstens, 1957; Gjelsvik, 1956 and 1957) as well as of Skaergaard magnetites (Vincent and Phillips, 1954) have been plotted with the results of the present study in Fig. 1. The curve E—E′ in this figure has been reproduced from an analogous diagram constructed by Vincent and Phillips (1954), who have superimposed on their diagram the stability fields of compounds in the $FeO—Fe_2O_3—TiO_2$ dry thermal system at their melting points. The curve E—E′ represents the trace of a valley in the ternary system and joins the $FeTiO_3—Fe_2TiO_4$ eutectic at 1320°C and the $Fe_3O_4—Fe_2O_3$ eutectic. Apparently Fig. 1 shows that all the analysis points here considered are situated on the probable thermal surface sloping downwards towards line E—E′ from the crest-line representing the solidus between the pure compounds $Fe_2TiO_4—Fe_3O_4$. Although the eutectic $Fe_3O_4—Fe_2O_3$ has not been determined, as pointed out by Vincent and Phillips (1954), it now seems, on the basis of the trend of the plotted points, that the eutectic may exist close to hypothetical point E′.

In the connection just referred to, Vincent and Phillips have suggested further that the magnetites from earlier, high-temperature rocks of the layered series at Skaergaard would lie nearer the crest-line and the magnetites of the later series nearer the valley. Considering the lithologic properties of the occurrences of the magnetites analyzed during the course of the present study, the aforementioned statement by Vincent and Phillips seems to be justified.

On the basis of the investigations of Skaergaard and the Schmoo Lake materials, Wright has recently (1961) made an interesting contribution to the study of the problems under discussion, promoting especially the idea that granular aggregates of ilmenite and titaniferous magnetite are the result of exsolution from an originally high-temperature cubic solid solution. This proposal is most acceptable, as the existence of originally high-temperature cubic solid solutions in the series of titaniferous magnetites seems to be supported by several theoretical conceptions and evidenced by observations made of ores that originated at various temperatures of formation. *E.g.*, Vincent *et. al.* (1957) proposed that »at high temperatures fairly extensive solid solution probably exists over much of the magnetite-ulvöspinel-ilmenite compositional field, and homogeneous titanomagnetites of widely varying compositions and magnetic properties may persist without unmixing in some rapidly cooled lavas.» This conclusion is appropriately supported, *e.g.*, by recent investigations of minerals in carbonatite lavas of Nyiragongo (Sahama, 1961), in which the magnetite included is homogeneous in texture while the composition corresponds to that of an ulvite. Along with the foregoing statements, the following main observations of the present study should serve as appropriate reminders at this juncture:

— In association with dolerites and diabases, dense intergrowths of magnetite and ulvite characterize the textures of the titaniferous iron ores, as shown by the specimen of Ulvö.

— In association with peridoditic and gabbroic rocks, granular aggregates of ilmenite and magnetite represent a dominating feature in the texture of the titaniferous iron ore. The magnetite, however, may show a considerable content of titanium, whereby exsoluted lamellae of ilmenite in the magnetite may be abundant.

— In occurrences of a distinctly metamorphic character, the texture of granular aggregates of magnetite and ilmenite is most pronounced. Even the magnetite formed, *e.g.*, those from Otanmäki, indicate a surprisingly small content of titanium, and even this is preferably incorporated in small amounts of exsoluted ilmenite in magnetite.

All the textural phenomena, as demonstrated, display this tendency: starting from a homogeneous solid solution, the unmixing of magnetite and titanian components apparently results in granular aggregates as the final product in conditions of slow cooling. As extreme instances of this development, the titaniferous magnetites of lavas on the one hand, and those from occurrences of a metamorphic field on the other, might be cited.

In addition to the effects caused by the velocity in the change of temperature, the oxidation-reduction relationships have had an important role in the process of exsolution, as emphasized in several earlier investigations on the titaniferous magnetites. Fig. 2 has been drawn to illustrate the rate of oxidation as indicated in various analyzed samples, though an analogical tendency should be estimated by means of Fig. 1. The values on the abscissa-axis of Fig. 2 represent the variation of Fe_2O_3 in weight percentages, as analyzed from a sample. The values on the ordinate-axis represent the variation of FeO in respective samples. Line A demonstrates the trend of the ratio $FeO : Fe_2O_3$ of magnetite having a theoretical composition. The circles in the figure represent the ratio $FeO : Fe_2O_3$, obtained directly on the basis of weight per cent results of an analysis. The crosses represent the ratio $FeO : Fe_2O_3$ of a normative magnetite as reported in Table 2.

The amounts of FeO closest to the theoretical line A are shown by analysis 8 (Vuorijärvi, Niskavaara) and analysis 2 (Otanmäki). Each of these samples shows low values of TiO_2, which means that these magnetites are relatively free of titanian components. A considerable excess of FeO is present in analyses 6 (Susimäki) and 10 (Jacupiranga). The respective contents of TiO_2 in these samples are 6.10 % and 9.71 %. In the former case, the content of titanium is in all probability mostly included in ulvite, though

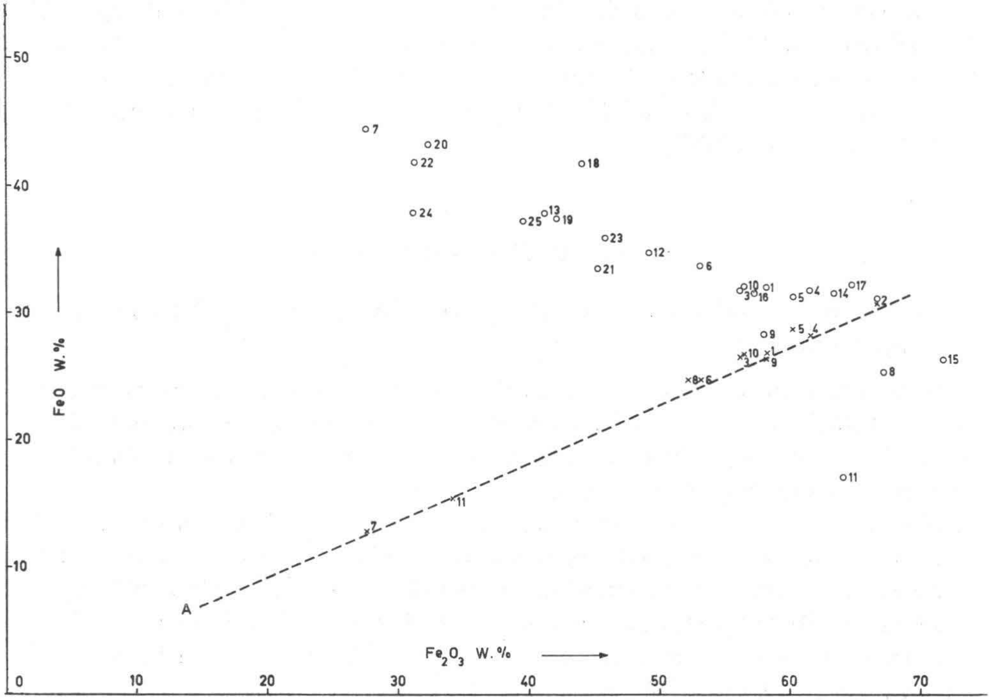

Fig. 2. Variation of the ratio FeO : Fe_2O_3 in the magnetite analyses referred to in Fig. 1.

this mineral has not been unequivocally indicated by the Susimäki samples in the course of the present study (*cf.* Ramdohr, 1956). In the latter case, the content of titanium is caused by a large number of small exsolution laths of ilmenite in the magnetite. The most pronounced deviation from the theoretical value is shown by analysis 7 (Ulvö, Sweden), as is to be expected. In another words, at higher temperatures the partial pressure of oxygen apparently had been at a minimum, while the iron had dominantly been in a ferrous state, maintaining this state in final crystallization because of rapid cooling. On the other hand, slow cooling favored the oxygen toward entering the system in increasing amounts, while the change of two-valency iron into three-valency iron became more intence. Actually this is what had been suggested by earlier conceptions, which imply that the ulvite characterized by two atoms of ferrous iron, is formed at higher temperatures, whereas under the oxidizing conditions of a lower temperature area the ilmenite is the stable form.

The rate of oxidation of iron in magnetites, or, which is the same thing, the ratio FeO : Fe_2O_3, seems to have a certain trend, as might be revealed by Fig. 2. In regard to the lithologic environment of the occurrences, including the magnetites analyzed, the ratio decreases in direction dolerite → gabbro-peridotite → gabbro → amfibolite. This behavior suggests that the ratio $FeO : Fe_2O_3$ might be used as an indicator of relative temperatures of formation in an analogical way as presupposed by geologic thermometer based on the variation of the TiO_2-contents of titaniferous magnetites (Buddington *et. al.*, 1955).

SUMMARY OF CONCLUSIONS

In the course of the present study, the following essential conclusions have been reached:

1. The vanadium content of magnetites of titaniferous iron ores associated with subsilicic rock formations depends on the genetic character of the respective occurrences. The vanadium module obviously is a convenient term to represent this interdependence.
2. The process of exsolution of the magnetite and titanian iron components has depended on the velocity in the change of temperatures as well as on changes in the oxidation-reduction relationships during crystallization.
3. The ratio $FeO : Fe_2O_3$ in the magnetites of titaniferous iron ores associated with subsilicic rock formations obviously might be used as an indicator of relative temperatures of formation.
4. The microscopic textures of the specimens studied have supported the opinion that the granular aggregates of magnetite and ilmenite in titaniferous iron ores are a result of the exsolution of a high-temperature cubic solid solution as proposed especially by Wright (1961). Observations revealing that the ilmenite lamellae in magnetite have an intimate connection with adjacent grains of ilmenite provide special evidence (Pl. I, Fig. 1) of such a process.

Acknowledgments — The authors wish to express their gratitude to the Finnish State Board of Natural Sciences for supporting this work by means of a research grant.

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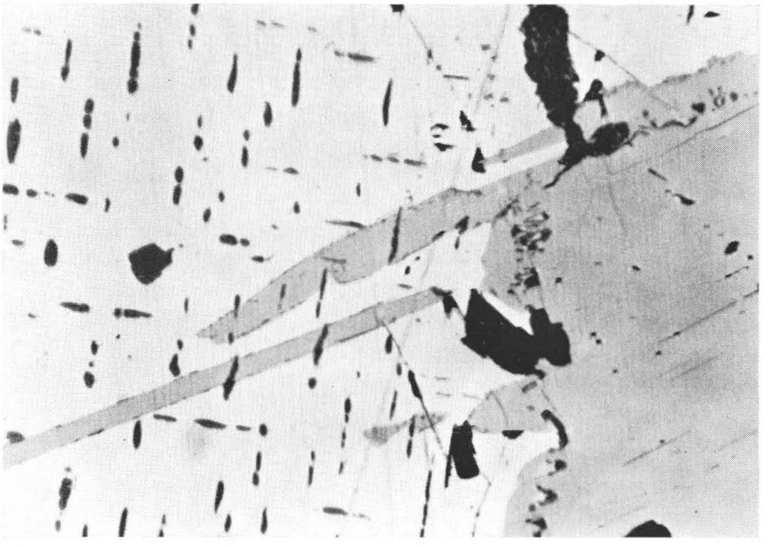

Fig. 1. Ilmenite (grey) and magnetite (light). Lamellae of ilmenite in magnetite are partly connected with ilmenite grains. Elongated or spot-like small grains (black) in cubic arrangement are spinel. Attu. Reflected light, 250 x.

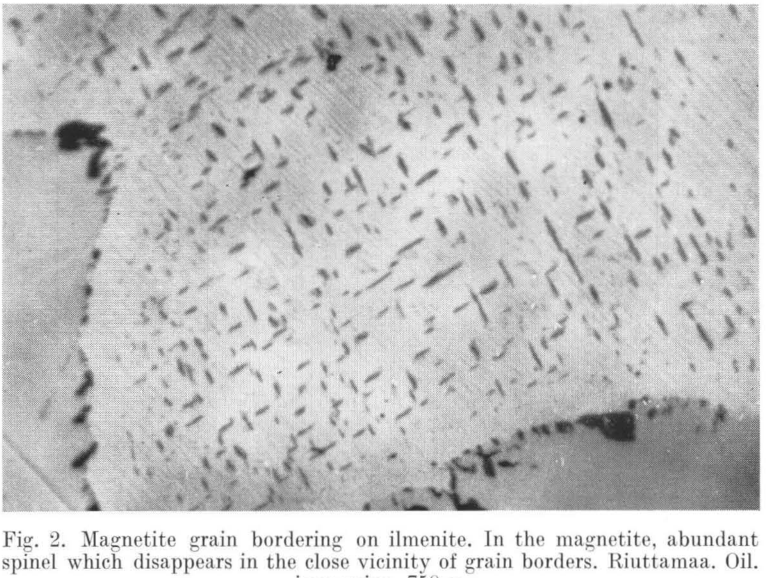

Fig. 2. Magnetite grain bordering on ilmenite. In the magnetite, abundant spinel which disappears in the close vicinity of grain borders. Riuttamaa. Oil. immersion, 750 x.

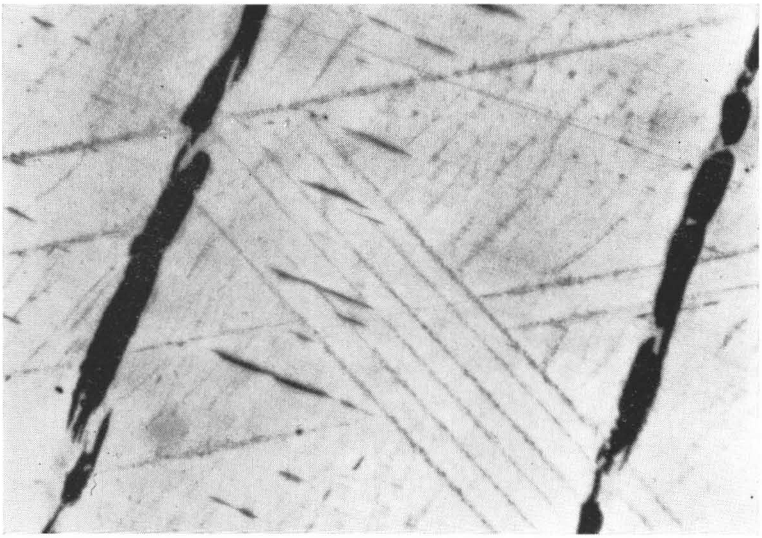

Fig. 1. Grain of magnetite crossed by two sets of faint lamellae of ilmenite, along which abundant darker spots probably composed of rutile are weakly visible. Dark and elongated bodies forming two sets in cubic arrangement are spinel. Tahawus. Oil. immersion, 750 x.

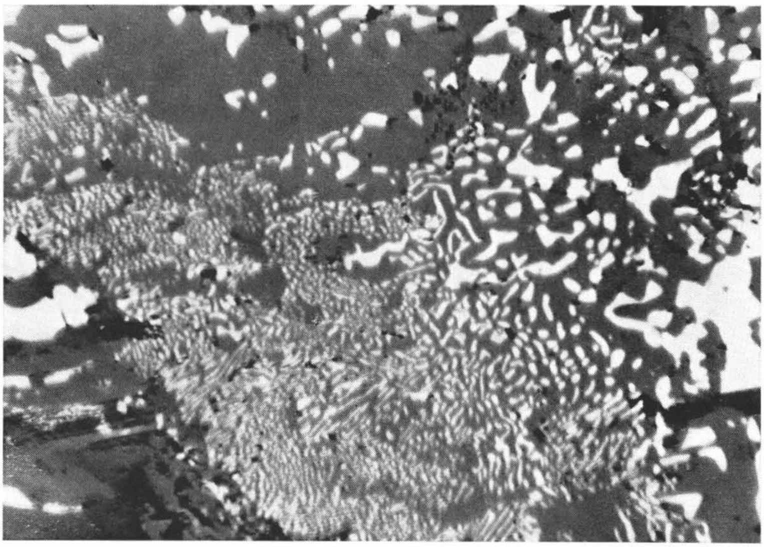

Fig. 2. Graphic intergrowth of ore oxide and silicates in connection with synantetic rims around olivine grains (not visible in the figure). Susimäki. Reflected light, 250 x.

O. Vaasjoki and *Aulis Heikkinen:* On the Significance of Some . . . .

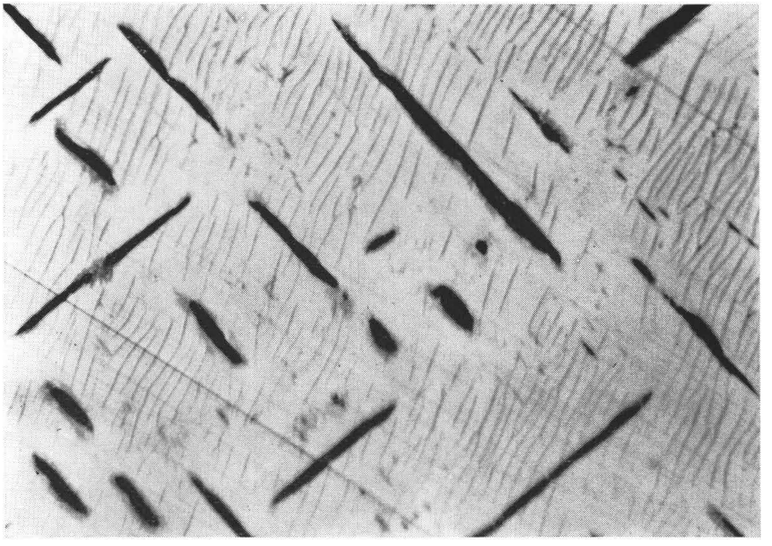

Fig. 1. Grain of magnetite. Elongated dark bodies of spinel form two sets in cubic arrangement. A faint, worm-like texture which has not been interpreted (*cf.* page 146). Susimäki. Oil. immersion, 1 200 x.

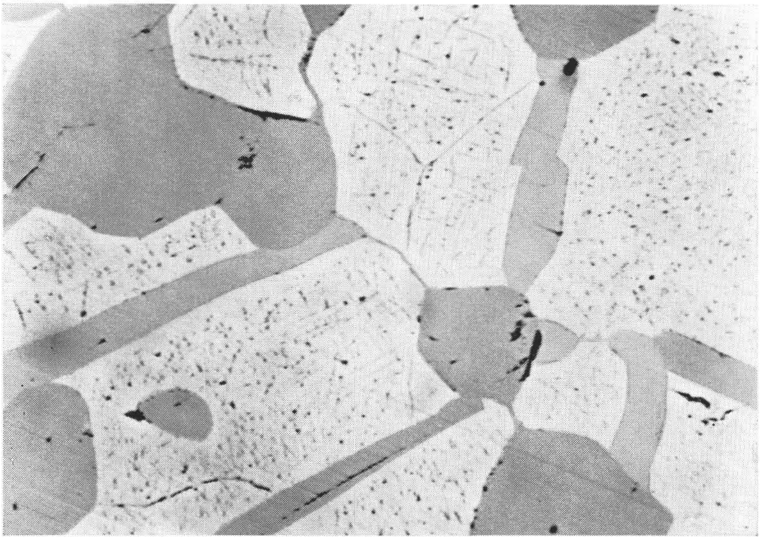

Fig. 2. Aggregate of magnetite (light), rounded grains of perovskite (darker grey) and lamellae of ilmenite (lighter grey). Abundant spinel in magnetite. Vuorijärvi, Niskavaara. Reflected light, 250 x.

7

INVESTIGATIONS ON THE DISTRIBUTION OF POLLENS IN AN EXTENSIVE RAISED BOG [1])

BY

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ABSTRACT

The author has investigated the distribution of pollen in an extensive, treeless raised bog in western Finland. Surface samples were gathered from hummocks and hummock ridges as well as adjacent muddy hollows, and considerable differences were observed in their pollen contents, both relatively and absolutely. *Betula nana* could be distinguished from the arboreal species of birch by the sizes of the respective pollen grains. Also the pollen grains contained in the series of samples taken from the peat deposits were studied. The interpretation of the pollen diagram was facilitated by taking note of the relative amounts of different pollens in the surface samples.

CONTENTS

[1]) Received May 31, 1962.

INTRODUCTION

It is well known that plants produce pollen in extremely varying quantities — anemophilous plants many times more than entomophilous plants. But even within these two main categories of plants, considerable differences may be observed between different species.

Sarvas (1955 a) has studied the production of pollen grains by forest trees and obtained the following counts: *Pinus* 278, *Picea* 94, *Betula verrucosa* 124 and *Betula pubescens* 116 per square millimeter.

The distribution of the pollen of anemophilous plants is affected quite fundamentally by its specific gravity. Studies in this field have been carried out by, among others, Pohl (1937 a). As examples, the specific gravity of certain pollen grains might by cited: *Pinus silvestris* 0.39, *Picea exelsa* 0.55, *Alnus glutinosa* 0.75, *Betula verrucosa* 0.81 and *Corylus avellana* 1.01.

The rate of descent of pollens in the air has been reported by Dyakowska (1936): *Picea exelsa* 6.8, *Quercus robur* 4.0, *Pinus silvestris* 3.7, *Tilia cordata* 3.2, *Betula alba* 2.9, *Corylus avellana* 2.9 and *Alnus glutinosa* 2.8 cm per second.

Many kinds of investigations into the distribution of pollens have been conducted. The results indicate that a great portion of the wind-borne pollen grains never gets beyond the immediate locality (local distribution), a considerable quantity is distributed across a radius of 500 meters (distribution over near surroundings), a portion travels between 500 meters and 10 km (short flight), while a small quantity is borne from ten to a hundred kilometers (long flight) and a few scattered grains are borne more than 100 km (exceptionally long flight). In the last-mentioned case, the pollen grains travel to alien surroundings or descend in a locality where the corresponding species flowers considerably earlier or later (Bertsch, 1942 and Firbas, 1949).

Hesselman (1919) has investigated the flight of pollens from a couple of ships out in the Gulf of Bothnia, one thirty and the other fifty-five kilometers off the Swedish coast. In the first investigation, 16.2 pollen grains per square millimeter were counted, seven grains being *Picea*, 6.7 *Betula* and 2.4 *Pinus*. In the second, the count was 8.8 grains per sq. mm, 4.1 being *Picea*, 3.6 *Betula* and 1.1 *Pinus*.

Sarvas (1955 b) later undertook similar studies by boarding a lightship situated at a distance of 20 km south off the Helsinki coast. The result: there was a greater frequency of certain pollens on the ship than in woods on shore. The investigation was concerned with pollen grains of four species, and the results were reported as pollen frequency per mm^2 in 24 hours. In the following, the first number indicates the frequency registered in the woods and the second that on board ship: *Betula pubescens* (939—1869), *Picea abies* (302—454) and *Pinus silvestris* (1228—1365). In respect to *Betula verrucosa*, the situation was contrarywise (723—589). According to the investigations, pollens may be borne long distances (20—50 km), at least over water. The combined count for the birches was highest, for *Picea* the lowest. On the other hand, Sarvas (1956) in studying the possibilities of obtaining birch seeds of maximum purity in nature reached the conclusion that a distance of 400 meters between a seedling stand and a natural stand would be sufficient to secure a nearly pure seed. The pollen production of a seedling stand of birch proved to be 150 grains per mm^2, of which only five per cent, or eight grains, had been borne a distance of 400 meters or more.

Erdtman (1921) carried out pollen studies of surface samples from southern Sweden and observed, *e.g.*, that in *Sphagnum* samples the *Pinus* pollen were »over-represented.» This was apparently due to the pine-dominated forests occurring over an extensive area and not to the pollen output of the few pines growing in the bog. On the other hand, the pollen frequencies of alder and birch approximated the proportionate representation of the species in the vicinity.

L. von Post (1924) also investigated the surface samples of the bog and reached the conclusion that the tree pollens in them corresponded in general to the composition of the surrounding forests. *Picea* pollen appeared, however, to be well represented and, particularly in places where the forest was sparse or was totally absent, considerably over-represented.

Aario (1940) investigated the relative pollen contents in surface samples collected from bogs in different forest zones in Lapland. He ascertained that *Pinus* pollen grains were generally over-represented. Thus in samples taken from the birch zone of Lapland *Pinus* accounted for 33 % of the pollens and on the tundras 72 %, although the distance to pine forests was between 100 and 130 km. The percentage of *Picea* pollen was five on the tundras, two in the birch zone, four in the pine zone and only slightly higher, or six, in the spruce zone. *Alnus* occurred sparsely (1 %) in the forest zone of Lapland at a distance of more than 100 km from the localities where the alder grew. Also *Tilia*, *Ulmus* and *Quercus* occurred sporadically at distances of several hundred kilometers from the closest habitat of the plants as a result of long-distance transportation.

On the basis of the absolute quantity of pollen grains, tundra can be distinguished, according to Aario, from forest zones by the frequency of tree pollens. Tundra yielded on the average 69 grains of pollen per 50 mg of peat. The same amount of peat from zones overgrown with trees yielded between 1 415 and 5 200 grains of tree pollen.

Aario states that the absolute quantity of pollens in peat depends on its degree of humification: *i.e.*, the pollen frequency of highly decomposed peat is generally greater than that of less humified peat. The rate of growth of the peat, however, is the most determining factor. The pollen frequency is highest in slowly growing species of peat.

Lüdi (1947) has carried out investigations of particular interest from the standpoint of the present paper of pollens contained in surface samples taken from the shore bogs of Katzenzee near Zürich. The majority of the samples were collected from areas grown over with different kinds of trees or at distances of a few dozen or hundred meters away. The greatest distance was approximately 400 meters. The result obtained was that at the greatest distances *Picea* became greatly enriched. Also *Pinus*, *Corylus* and *Quercus* became somewhat more abundant at the most distant points, whereas *Betula* decreased rapidly as the distance grew and *Alnus* ceased to be represented altogether at the farthest distance investigated. *Betula* and *Alnus* proved to be quite dominant in the immediate locality of their occurrence.

As the present author has been obliged to move about in bogs of different types and sizes, his attention has frequently been drawn to the great variety of conditions under which pollen grains are borne to bogs and subsequently imbedded in peat deposits. The conditions prevailing in extensive treeless bogs, in especial, differ from those of small and forest-covered bogs. The bog investigated by Lüdi most closely approximates the latter type. The present author has concerned himself with the following questions: 1— To what extent and in what proportions do the pollens of the species of trees represented in surrounding forests spread across the surface of an extensive treeless bog? 2— What part does the pollen of a bog's own vegetation play in the pollen statistics? 3— Have the different areas of a bog's surface any influence on the occurrence of pollens?

When a bog that appeared suitable for the purpose was found at Pirttikylä, in the raised bog area of western Finland, surface samples were collected there, together with a series of samples from the peat deposits as well, in the hope that the results yielded by the surface samples would provide additional light in interpreting the pollen diagram.

THE BOG INVESTIGATED — ITS NATURE AND SURROUNDINGS

Sanemosse bog of Pirttikylä, which is a typical western Finnish raised bog, is located about six kilometers to the NE from the parish church of Pirttikylä and some 40 km to the S from Vaasa. Its altitude is about 35 m above sea level. The area of the bog is about 1 200 ha and it is almost circular. Its diameter varies at different points between 2.5 and 3.5 km (Fig. 1). From its eastern margin the bog extends northeast as a narrow bay.

The western and southern margins of the bog consist of clayey soil, like the bottom. The northern margin and the southern half of the eastern margin consist of fine sand and medium sand while the northern half of the eastern margin consists of till.

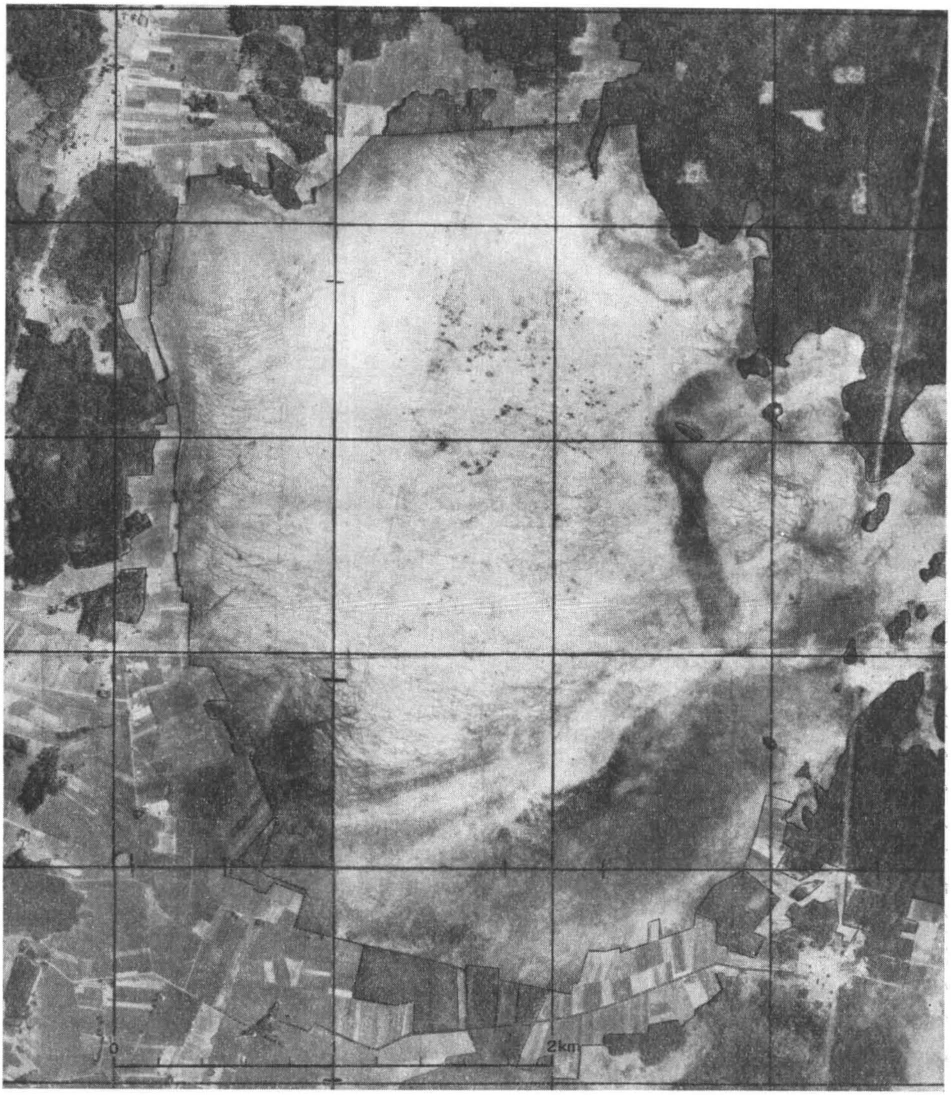

Fig. 1. Aerial photograph of Sanemosse and its surroundings. By courtesy of the Surveyor General's Office.

Sanemosse is nearly treeless and in type it principally represents a *Sphagnum fuscum—Eriophorum vaginatum* bog with hummock ridges (Fig. 2). The hummock ridges and hummocks are rather low and in places form the concentric rings typical of a raised bog, as revealed by the aerial photograph. In places the intermediate areas consist of *Sphagnum—Eriophorum*

Fig. 2. *Sphagnum fuscum-Eriophorum vaginatum* bog with hummock ridges. From point XVIII NE. Photo M. Salmi.

bog with *Scheuchzeria*. The vegetation covering the hummock ridges is dominated by *Sphagnum fuscum*, in addition to which mention should be made of various *Cladonia* species, the subshrubs *Andromeda*, *Calluna*, *Rubus chamaemorus* as well as *Betula nana*. The last-mentioned is the most common of the subshrubs and also, from the standpoint of the present paper, the most interesting. It is met with throughout nearly the entire bog in varying quantities. *Betula nana* grows in greatest abundance among the pines and subshrubs found in the marginal parts and, most particularly, in the southern part of the bog (Fig. 3). It occurs most sparsely or not at all in the most watery parts of the bog, running eastward from the center. Around the middle of line A XXX (Fig. 10), *Myrica gale* abounds. At the center of the bog and to the north and northeast, large pools occur in spots (Fig. 1), while there are muddy hollows without vegetation here and there in the bog.

At the eastern margin of the bog, there are extensive stretches of wet treeless *Carex* and *Sphagnum* bog, appearing in the aerial photograph as dark areas. The most prevalent of the former type is the *Carex lasiocarpa* bog. In places it is mesotrophic, so that *Equisetum limosum* and *Menyanthes* are the most common species of grass. The principal species of the treeless *Sphagnum* bogs are *Sphagnum parvifolium*, *S. balticum*, *S. papillosum* and *S. fuscum*. In addition, this type of bog has a sparse representation of various *Carex* species, whereas *Eriophorum vaginatum* and, among the subshrubs, *Oxycoccus* and *Andromeda* are quite abundant; but *Betula nana* is a rare occurrence.

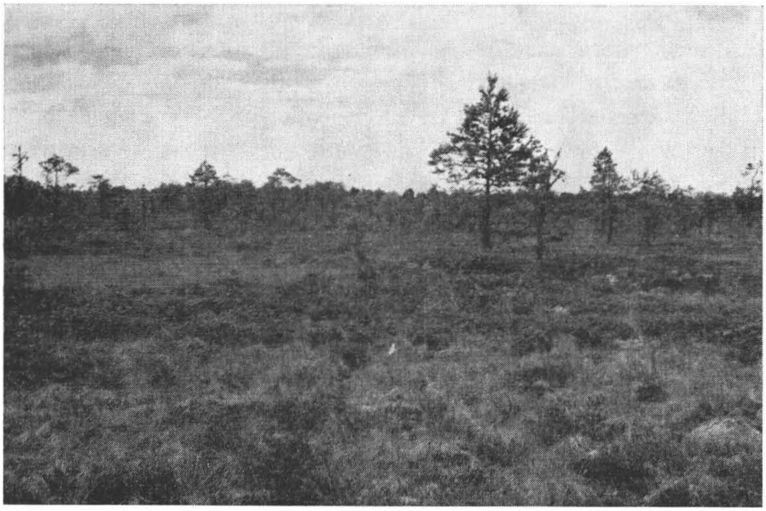

Fig. 3. Pine bog with *Betula nana* and various subshrubs. Point A XXII. Photo M. Salmi.

The marginal parts of Sanemosse have been brought under cultivation by burning the surface of the peat. The eastern margin of the bog is largely in a natural state, the northeastern till margin, for instance, *in toto*. The cultivated marginal parts have been separated by ditches from the parts in a natural state (Fig. 1). From the western margin a number of ditches measuring 300—500 meters in length have been dug toward the center of the bog. They can be seen in the aerial photograph, and trees, chiefly pine and birch, have sprung up along the edges of the ditches.

In the south and the southwest, Sanemosse is bounded by a broad, tilled open space, which grows hay, rye and oat crops. At the northern margin of the bog there are a couple of small bays of the bog which have been brought under cultivation. The areas drained by ditches along the western margin are mainly pasture land.

According to Ilvessalo (1960), in the environs of Pirttikylä spruce-dominated forests account for 51—60 %, pine-dominated ones 21—30 % and birch-dominated ones 11—20 % of the total. Locally, the dominance of spruce appears to be quite marked in the immediate surroundings of Sanemosse. In this connection, mention should be made of some of the present author's observations in the proximity, mostly, of the investigation lines drawn across the bog for research purposes (shown in Fig. 10).

At the northern margin of the bog, the trees growing in the clay and sand near the starting point of line A consist 90 % of spruce, 4 % of pine and 6 % of birch. The proportions of the various tree species along the western edge of the bog, at the starting points of lines A X and A XX, have been observed to be as follows. In the case of the former, 95 % of the trees are spruce and 5 % birch, while in the latter, the figures are 90 % spruce, 4 % pine and 6 % birch. At the eastern end of lines A X and A XX, the species growing out of the till are 80 % spruce, 15 % pine and 5 % birch, while on the sandy soil at the eastern end of line A XXX, spruce accounts for 95 % and birch for 5 % of the trees.

The forests representing direct extensions of the stands bordering the bog differ considerably in composition from the foregoing. They form around the margins of the bog a zone twenty to thirty or more meters broad consisting of 50 % pine, 30 % spruce and 20 % birch. In places, the spruce is missing altogether. The share of the pine in such places amounts to 95—98 % of the trees, the remainder being birch and alder *(Alnus incana)*. In the woods along the southern margin of the bog near the cultivated fields, pine and birch are equally prevalent. Lines A X and A XX run along the eastern border of the bog in the proximity of »islets» of till. Along the former, at a distance of between 700 and 500 m from the eastern end of the line, pine and birch grow in equal abundance, though an occasional alder occurs, too. Two hundred meters farther to the east, along the line, there is a pine bog with subshrubs on which no other tree but the pine grows; but 100 m from the edge of the bog, in addition to the pine, birch accounts for 5 % of the trees. The line terminates at the edge of the forest, where there first occurs a narrow zone of pine and birch and a scattering of alder. Farther back one meets with spruce-dominated woods. At the eastern end of line A XX, the composition of the forest approximates that described in the foregoing. Some 500—600 m from the eastern margin of the bog, birch and pine occcur in equal abundance, but quite at the terminal point of the line, the ratio changes to about 9—1 in favor of the pine, with an alder here and there thrown in for good measure. Spruce-dominated forest begins immediately behind the stand bordering the bog.

In its natural state, therefore, the bog does not support any stands of trees, which are limited to the margins, the strips lining the ditches dug in the bog from its western margin and the ditches dug to drain the land for purposes of farming, the boggy stretches covered with pines and subshrubs at the southern and eastern margins, and the islands and their surroundings, with their slight patches of trees. *Picea* is the dominant species growing out of the mineral soil beyond the area of the bog, while *Pinus* lords it over the woods lining its margin. *Betula* is in spots quite conspicuously represented, especially along the ditches; but *Alnus* occurs only sparsely in the company of *Betula*.

THE RESEARCH MATERIAL AND THE METHODS USED

The research material was collected from investigation lines — shown in Fig. 10 — laid down across the bog. Line A runs just about N—S approximately 400—700 m from the western margin of the bog, while lines A X, A XX and A XXX run at right angles to it a kilometer apart.

Along each line surface samples were taken at intervals of 200 m. The samples were cut with a knife from spots in the bog covered with peat moss; but if next to the sampling site there happened to be a muddy hollow lacking a moss covering, a parallel specimen was peeled off its surface. Thus a total of 77 surface samples was collected from the bog, fifteen of them having been taken from muddy hollows.

In addition to the surface samples, a series was taken with a drill at depth intervals of 10 cm from top to bottom of the entire peat deposit at the crossing point of lines A and A XX. The thickness of the peat deposit was 3.4 m, but the drilling was continued to the underlying water sediments up to a depth of 5.2 meters. Thus a total of 52 specimens was taken from the profile.

The samples were dried in laboratory air, after which 100-mg pieces were weighed for investigation. The surface samples consisted of slightly humified *Sphagnum* peat or nearly unhumified peat moss. The sample for testing was always sought from similar spots, ones where live and dead moss lay adjacent. The samples taken from muddy hollows represented a slightly higher degree of humification (H_3), but the sample material as a whole reveals only minor differences in this respect.

Since the samples proved to contain pollen grains in scanty measure, the weighed portions of peat were treated by the acetolysis method developed by Erdtman (1934, 1936), just as were the specimens taken from the peat profile. The inorganic matter contained in the water sediments was removed from the specimens by the HF-method (Assarsson and Granlund, 1924), after which the residue was treated in the manner described in the foregoing.

An effort was made in the study to separate *Betula nana* from the arboreal birch pollens. For this purpose, investigations were performed with recent pollens from three species of birch, obtained from the Botanical Museum of the University of Helsinki. These grains were likewise treated with acetolysis. Accordingly, it became possible to make comparisons between all the specimens comprising the research material.

In the pollen analyses, attention was given not only to arboreal pollens (AP) but also to non-arboreal pollens (NAP). The share of the latter is reported in terms of the number of grains per 100 grains of AP. Moreover, the absolute quantities of pollen per 100 mg of peat (PF/100 mg) are given separately for AP and for NAP. The pollen grains of *Betula nana* are reported as percentages of the arboreal birch pollens.

Table 1. The size of the recent pollen grains of different *Betula* species, per cent.

Species	Diameter in μ																		
	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36
B. pubescens ..											0.5	0.5	2	17	18	32.5	13	16	0.5
B. verrucosa ..						0.5	1	4.5	12	33	32	14	3						
B. nana	0.5	1	1	12	27	33	19	1	2	2	1	0.5							

THE SIZE OF THE POLLEN GRAINS OF DIFFERENT *BETULA* SPECIES

Sanemosse is practically treeless. *Betula nana* occurs sparsely nearly throughout the bog, and in spots along the margin it is the main subshrub of the pine-grown boggy stretches. Accordingly, it was important during the study to separate its pollen from that of the other *Betula* species in order to ascertain the distribution of arboreal birch pollens in the bog. The comparative study took in the following species: *Betula pubescens*, *B. verrucosa* and *B. nana*.

Microscopically 200 pollen grains of each species of birch were counted. The results are presented in Table 1, which gives the distribution on the basis of size of the pollen grains of the different species percentually. Fig. 4 presents the same results graphically.

The pollen grains *B. nana* are distinctly the smallest. They measure mostly 21—24 μ. The commonest size is 23 μ. The grains of *B. verrucosa* pollen are larger, being predominantly 26—29 μ, though the greatest number are in between, or 27—28 μ. The pollen grains of *B. pubescens* are larger than

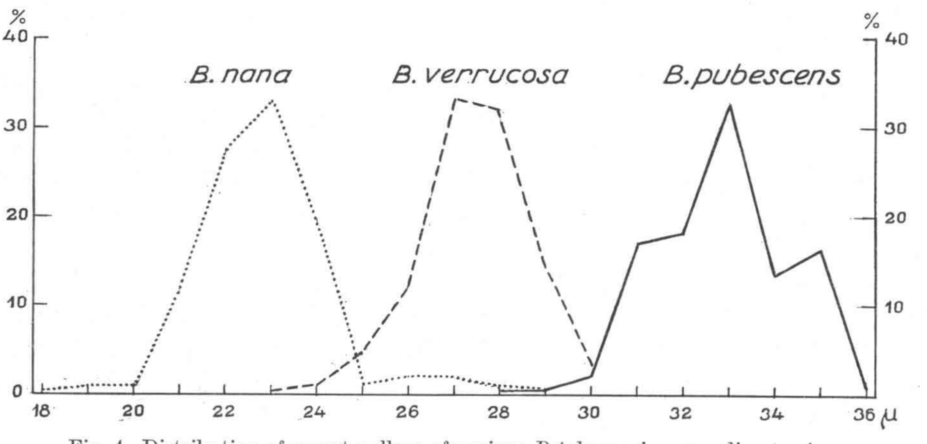

Fig. 4. Distribution of recent pollens of various *Betula* species according to size.

Table 2. The size of recent pollen grains of different *Betula* species according to Kujala (1946).

Species	Diameter in μ												Pollen
	20	21	22	23	24	25	26	27	28	29	30	31	
	Number of pollen grains												
B. verrucosa		1	1	7	8	3	3	1	1				25
B. pubescens, Southern Finland ..						1	9	21	19	9	1		60
B. pubescens, Regio alpina							1	13	12	16	5	1	48
B. nana	1	5	14	8	1								29

those of the other species investigated. The most significant portion of them vary between 31 and 35 μ, while 33 μ is the commonest size.

The differences in size between the pollen grains of different species are distinct. The *Betula* species investigated share the feature that the main portion of the pollen grains, or 91—96 %, is concentrated in four or five successive size classes and that the peak of each species consistently accounts for 32.5 and 33 % of the total.

Inasmuch as it was desired for the purposes of the present study to separate the *B. nana* pollen from that of the other species of birch, all the grains measuring less than 24 μ among the *Betula* pollens found in the peat specimens have been classed as representing *B. nana.*

Kujala (1946) has also investigated the size of the pollen grains of the birches growing in Finland. His results are presented in Table 2. Comparing the different species, one will note that *B. nana* grains are the smallest and those of *B. pubescens* the largest. However, the corresponding species in Kujala's marerial are distinctly smaller than the ones included in the present study. The differences are evidently due to a difference in the method of preparing the specimens for measurement. Kujala treated his pollens in alcohol, where the grains become round without swelling. The acetolysis method used in connection with the present study, on the other hand, results in a swelling of the pollen grains (Wenner, 1947; Erdtman, 1954). Eneroth (1951) has treated recent *Betula* pollens in 10 % KOH. His values are likewise smaller than those obtained in connection with the present investigation. Moreover, they are also somewhat smaller than those obtained by Kujala.

THE OCCURRENCE OF POLLENS IN SURFACE SAMPLES

THE FREQUENCY OF DIFFERENT POLLENS

Figs. 5—8 show the percentages of arboreal pollens met with in the surface samples collected along the investigation lines. Reading from the top in the second of the figures, one will observe the proportions of the arboreal

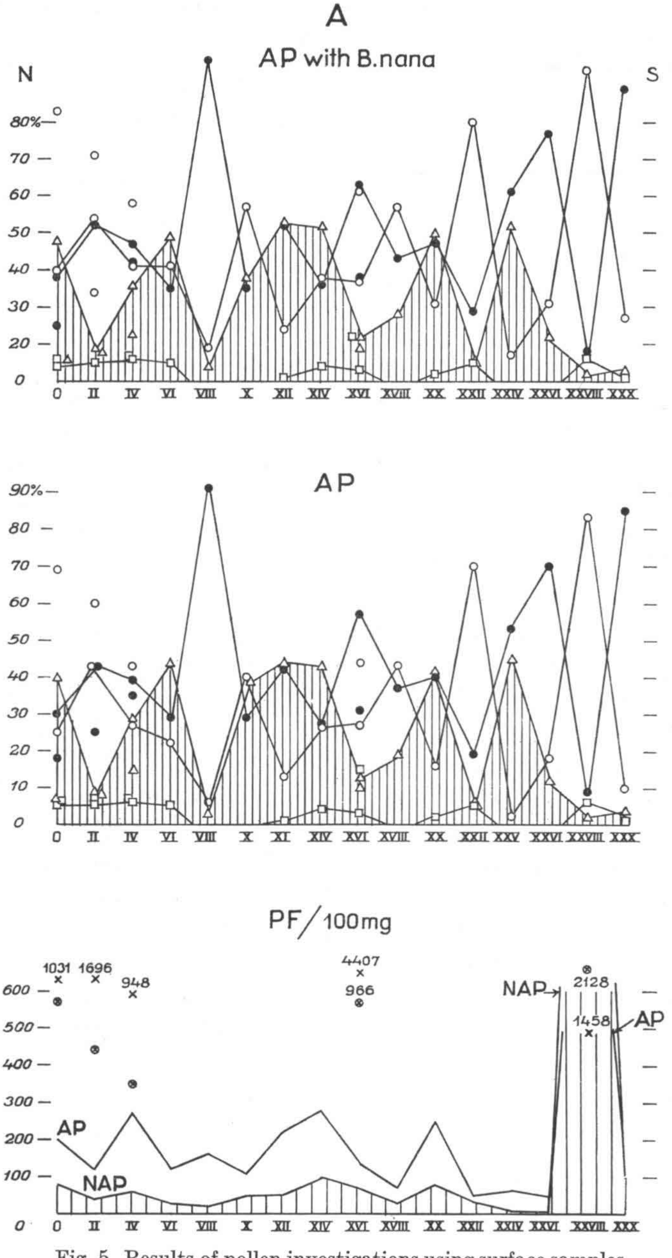

Fig. 5. Results of pollen investigations using surface samples taken from along line A. The uppermost diagram presents the percentages of AP with *Betula nana* and the next the percentages after *Betula nana* had been separated from the rest. The bottom drawing shows the absolute pollen frequencies (PF) of AP and NAP per 100 mg of peat. *B. nana* is included among NAP. x = AP and ⊗ = NAP in terms of PF/100 mg of sample taken from muddy hollow.

pollens after *Betula nana* has been removed from the pollen grains of other species of birch. The differences among these figures are slight.

As a general feature of the figures, the proportions of the arboreal pollens vary rather considerably from point to point in the same line and the variations along the different lines differ notably. Common to the pollen proportions of all the lines, with a few exceptions, however, is the predominance of *Pinus*. This becomes clearly evident when *Betula nana* has been separated from the other *Betula* species. Along lines A X (Fig. 6) and XX (Fig. 7) the strength of *Pinus* is most conspicuous. These lines run across the middle of the bog, farther from the forests and the small patches of woods growing in the bog than the rest. The proportion of *Pinus* is highest in the middle of the bog, where in most spots it exceeds 50 %. But near the margins of the bog *Picea* and *Betula* are in most places on a par with it, and in some spots even exceed *Pinus* in abundance. Along lines A (Fig. 5) and A XXX (Fig. 8), which run near the edge of the forest, the proportions of the different pollens vary quite irregularly. However, at the eastern end of line A XXX *Pinus* dominates. There one meets with mesotrophic *Carex* bog. In these damp parts of the bog there are no stands of trees, but the distance to the pine-grown stretch of bog to the south is only about 200—300 meters. Its influence is reflected in the pollen statistics. The terminal point of the line is in the woods, where pine and birch grow on equal terms, though the birch trees are largely young, not all of them yet probably fertile. No great differences exist here between the proportions of *Pinus* and *Betula* pollen grains. The proportion of spruce pollen approximates that of the other two species. A spruce forest occurs there some twenty or thirty meters beyond the woods lining the margin of the bog.

The occurrence of *Picea* is most clear-cut along lines A X and XX and least so along line A. Its proportion along the former two is quite high in the marginal parts of the bog — it rises to its maximum value at a distance of between 800 and 1 000 m from either end and then rapidly decreases toward the middle. Excepting the peak values, spruce accounts for ca. 40 % of the pollen total in the marginal parts of the bog, but in the middle the figure is only 20—25 %. The conspicuous occurrence of spruce in the bog at a distance of about a kilometer from its margin is probably to be explained this way: pine and birch form the woods immediately bordering the bog and their pollen grains are therefore prominently represented in the area close by, whereas the share of the more distant spruce forest becomes proportionally more important in the pollen statistics as that of *Betula* diminishes farther out in the bog. The light pollen grains of the spruce are borne, accordingly, in relatively great abundance for a distance of about a kilometer, but the proportion of *Picea* decreases rapidly past that distance in comparison with the even lighter *Pinus* pollens.

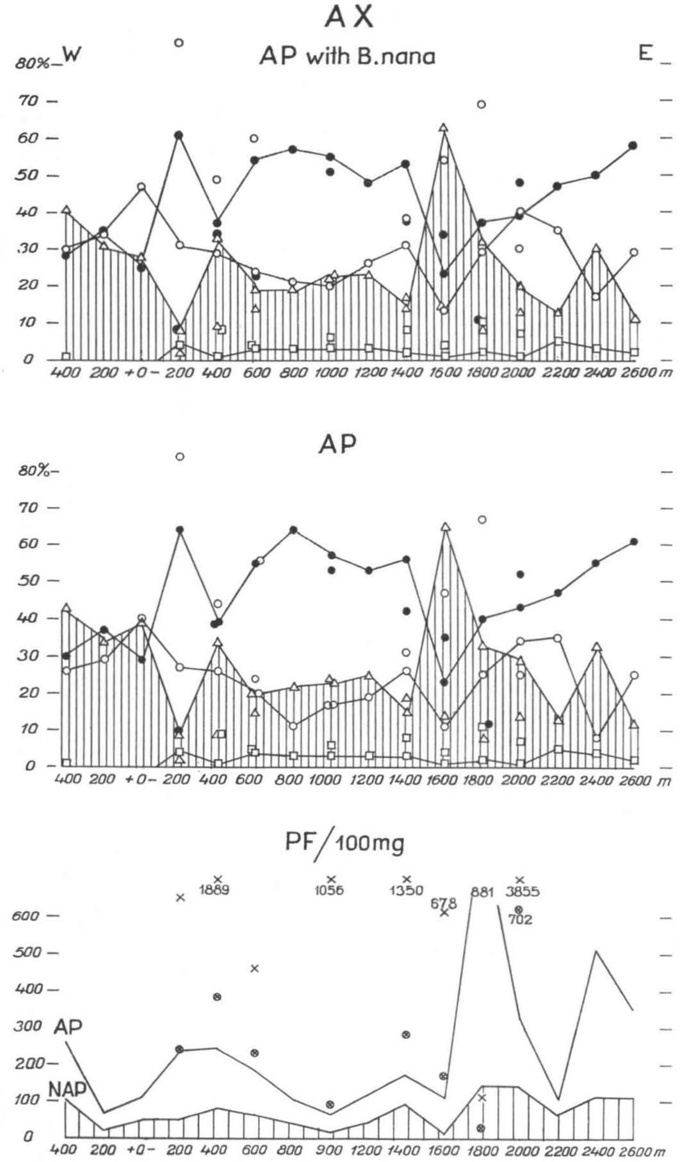

Fig. 6. Results of pollen investigation carried out with surface samples collected along line A X. Explanations in Fig. 5.

The presence of birch in the surface samples from the bog is indefinite and variable. However, along lines A X and A XX the *Betula* pollens decrease rapidly in abundance toward the middle of the bog. *Betula* generally occurs in smaller amounts than *Picea* and *Pinus*, just as the birch is sparsely present

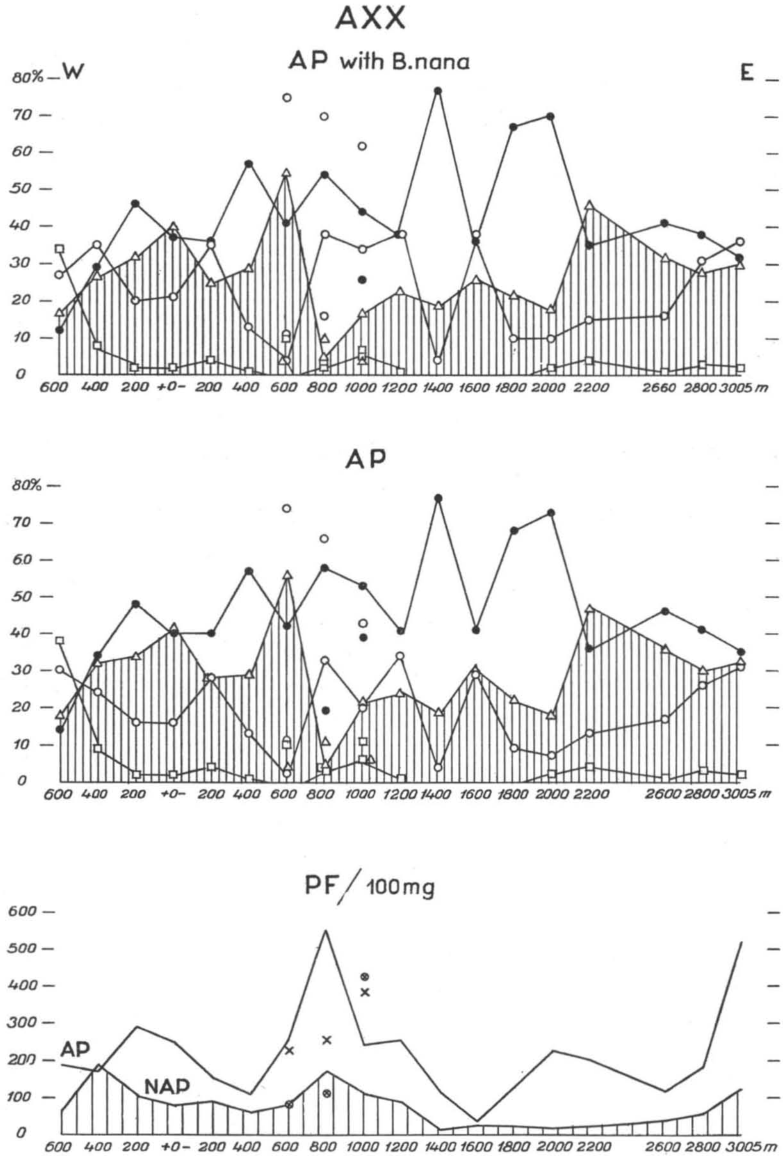

Fig. 7. Results of pollen investigation carried out with surface samples collected along line A XX. Explanations in Fig. 5.

in the bordering woods. At the western end of line A XXX, however, *Betula* exceeds the other arboreal pollens. This is explained by the fact that 70 % of trees growing in that area are birches, the rest being pines. At the — 1 000-m and—1 200-m points along the same line, the birch pollens again exceed the others in amount. At these points the bog type may be designated as wet mesotrophic *Carex* bog. At other points as well where the surface of the bog is wet, *Betula* pollens seem to be most abundant, as evidenced by those specimens taken along lines A, A X and A XX that happen to come from muddy hollows (Figs. 5—7). With only a single exception, the share of *Betula* in specimens collected from muddy hollows is higher — in many cases considerably so — than in the specimens taken from adjacent hummocks or hummock ridges, where usually *Pinus*, sometimes *Picea*, takes precedence. The relative abundance of *Betula* pollens in muddy hollows and in wet boggy spots, which during the trees' flowering time in early summer are submerged, is apparently due to the sorting effect of the water. The *Betula* pollen grains, the specific gravity of which is greater than that of *Pinus* and *Picea* and which lack the air-sacs of the latter two, sink rapidly after falling into the water down to the bottom of the hollows. With their air-sacs, the *Pinus* and *Picea* pollen grains, on the other hand, are capable of floating for some time; and the wind will drive some of them to the edges of the hollows. The *Betula* — as well as *Alnus* — pollens thus appear enriched within the hollows, whereas the wind-blown *Pinus* and *Picea* pollens are concentrated along the edges of the hollows.

This phenomenon is illustrated by the observations made by the author at Paukkaneva, in the commune of Nurmo. From a pool about 8 m long, 4 m wide and 0.5 m deep at the deepest point and from the surrounding area, five samples were taken for investigation (spots marked in Fig. 9). The pool was bordered by a *Sph. cuspidatum* zone 20—30 cm broad, which was only a few centimeters above the water level of the pool. From it rose *Sphagnum fuscum—Cladonia* hummock ridges to a height of 30—50 cm on different sides of the pool. On one of the long sides grew a few pines.

The pollen counts showed that *Pinus* dominated in all the samples, which is understandable in view of the local conditions, for the distance to the nearest forest, consisting of spruce, pine, birch and, to a minor extent, alder, is about half a mile. *Betula*, from which *B. nana* has been separated, ranks distinctly second in the samples taken from the middle of the pool and near one side of it. In the *Sph. cuspidatum* zone the share of *Betula* equals that of *Picea* or is lower. Also *Alnus* is at its most abundant in the middle of the pool.

The absolute amounts of pollen contained in 100 mg of peat vary greatly in different specimens collected at Paukkaneva. The highest counts were yielded by the samples Nos. 3 and 4, *i. e.*, those taken from the side of the pool opposite to the pine-grown hummock ridge. The direction in which the wind happens to be blowing during the flowering season naturally determines to which side the pollen grains are borne.

As is well known, large bodies of water, like lakes and the sea, concentrate in their sediments the pollen grains of conifers, particularly *Pinus*, which

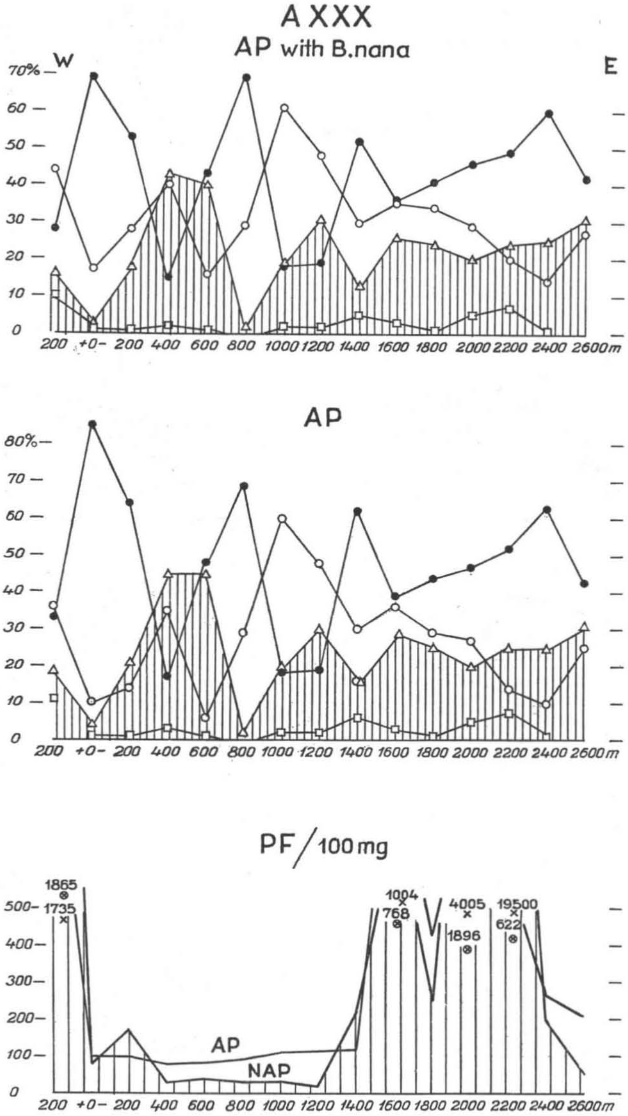

Fig. 8. Results of pollen investigation carried out with surface samples collected along line A XXX. Explanations in Fig. 5.

with their good flying and floating properties are borne great distances off shore. Observations of this phenomenon are mentioned by the present author, among others, as evidenced by the sediments of Murronneva, in the commune of Vähäkyrö, about 45 km N from Sanemosse (Mölder and Salmi,

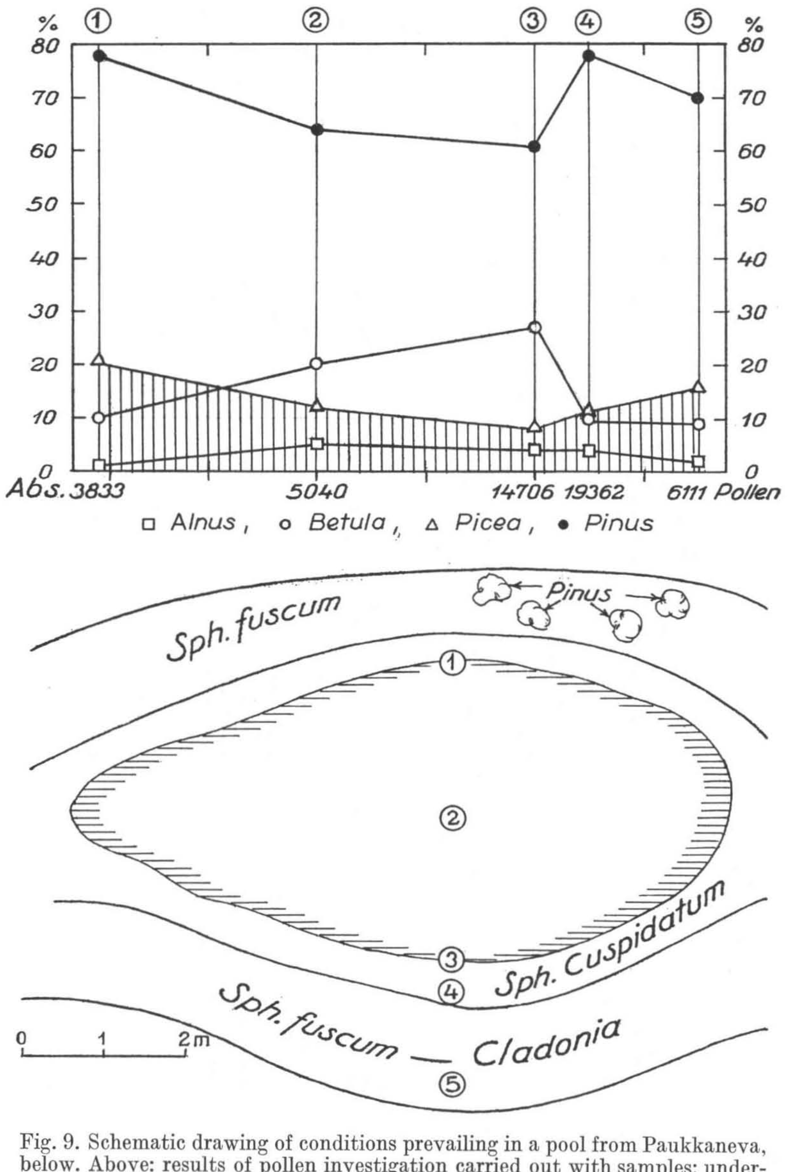

Fig. 9. Schematic drawing of conditions prevailing in a pool from Paukkaneva, below. Above: results of pollen investigation carried out with samples; underneath diagram, absolute PF/100 mg of peat.

1955). In addition, these pollens are known to collect in great abundance on the shores of bodies of water during the flowering season of conifers.

The sorting and enrichment of pollens evidently take place also in hummocks and hummock ridges within different horizons. The large *Picea* and

Pinus grains with their air-sacs become caught in the upper part of the *Sphagnum* covering more readily than the roundish *Betula* and *Alnus* grains, which sink lower than the former when, for instance, washed by rain waters. Accordingly, among the pollen grains of the same year's crop, *Betula* and *Alnus* and other pollens of comparable size and shape sink lower down in the moss than the *Pinus* and *Picea* pollens. But since the same process is repeated year after year, this need not cause disturbances in the pollen statistics of the peat layers as does the sorting performed by the water in the bog pools.

Alnus grains are only sparsely present in the surface samples from Sanemosse; but alders account for only a minor part of the woods bordering the bog. *Alnus* pollen occurs in greatest abundance in places along the investigation lines where alders grow in the immediate vicinity. Notewothy, above all, is the western end of line A XX, where the frequency of *Alnus* exceeds that of the other species, comprising 34 % of the total. The point lies in a field and the edge of the forest is about 80 m away. Birch and pine make up 40 % of the trees and spruce 20 %, but numerous alders grow there, too. Also near the last point at the eastern end of the same line there is a cluster of a few alders, but the *Alnus* pollen there amounts to only two or three per cent of the total. The closed-in spot where the alders are situated prevents their pollen from spreading beyond the immediate locality in any appreciable abundance. The forest there consists 90 % of pine and 10 % of birch, and *Betula, Pinus* and *Picea* are fairly evenly matched in the pollen statistics. The local occurrence of *Betula* is conspicuous, though the species is only slightly represented among the trees. At the western end of line A XXX, *Alnus* pollen accounts for 11 % of the total count. This point likewise is situated in an open field and about fifty meters from a patch of woods consisting of smallish pines and birches as well as a scattering of alders. *Alnus* pollen is absent at the starting point of line A X. A search of the bog margin failed to lead to the finding of any alders among the forest trees. Along all the lines, the proportion of alder diminishes toward the center of the bog, and in places its pollen is missing altogether. For a distance of half a mile along the center of line A XX, no *Alnus* pollen whatsoever has been met with. Here the distance to the alders growing along the eastern margin of the bog is approximately one kilometer. There is another gap in *Alnus* pollen about a km from the western end of the line. On the basis of these observations, it may be stated that the local occurrence of *Alnus* pollens is quite abundant for a distance of between 100 and 200 m from the trees, after which there takes place a steep decline in its frequency so that beyond a km the species occurs only incidentally.

It should be added that the surface samples from Sanemosse yielded some grains of the pollen of certain rare species, as listed by number per 100 grains of AP in the following. Along line A X the count was: at point — 600 *Quercus*

1, — 1 800 *Corylus* 2, — 2 000 *Corylus* 1 and *Ulmus* 1 (all specimens taken muddy hollows); along line A XX: at point + 400 *Tilia* 1, — 800 *Fagus* 1, — 1 000 *Corylus* 1 and — 1 800 *Tilia* 1 (*Fagus* and *Corylus* from muddy hollows); and along line A XXX at point + 200 *Tilia* 1. A special feature in the occurrence of the pollens mentioned is the fact that, with the exception of *Tilia*, they occur only in muddy hollows. Furthermore, it is noteworthy that pollens of mixed oak forests have been met with principally in the treeless central part of the bog. *Tilia* represents an exception in this respect, too. In any case, the pollen grains of rare species as occurring in Sanemosse must be considered the result of fortuitous long-distance flights, for the northern limit of the present distribution of the plants in question lies scores and even hundreds of kilometers from Sanemosse — with the exception of *Tilia*, which is likely to grow fairly near the bog.

THE ABSOLUTE POLLEN FREQUENCY

In connection with the present study, attention was also given to the absolute amounts of pollen. The results are shown in the diagrams of Figs. 5—8, the arboreal pollens (AP) being represented separately from the non-arboreal ones (NAP), with *B. nana* included among the latter.

Along line A (Fig. 5) the absolute AP count varies little on the whole. However, at point XXVIII at the southern end of the line, the pollen count is exceptionally high. The point lies in a field cultivated by burning the natural vegetation, at a distance of about 100 m from woods consisting of 70 % birch and 30 % pine. Here the AP count is 1 458, as compared with counts ranging between fifty and 280 elsewhere along the line. Birch scores highest among the arboreal pollens, with 83 %, pine scoring second with a very modest count. The great abundance of *Betula* is clearly local. The absolute NAP count is still higher at the same point, or 2 128, although elsewhere along the line it is as low as between 5 and 92. The high NAP score is primarily due to *Rumex*, which accounts for 87 % of the non-arboreal pollens. *Rumex acetosella* grows in abundance in the burnt-over clearing with its acid soil. According to Pohl (1937 b), the pollen output of this plant is very prolific. The abundance of NAP is also partly due to the fact that *Myrica gale* grows nearby.

The AP count is notably higher in the muddy hollows than in the adjacent *Sphagnum*. This is brought about primarily by the birch. NAP is likewise high in the muddy hollows. The pollens include *Cyperaceae*, *Gramineae* and various grasses as well as, in small amounts, *Ericaceae* and *B. nana*.

Line A X (Fig. 6) generally resembles the foregoing in respect to absolute pollen counts. NAP remains at a low figure, however, right down the line; but AP is abundant at points —1 800 and —2 400, where *Pinus* in each case is in the majority. At these points pine is the predominant species, while the spruce takes over at some distance beyond the eastern terminus of the line. *Ericaceae* pollens are the most prevalent among the NAP, but at the western end of the line *Cyperaceae* pollens are commonly met with in the area of wet *Sphagnum—Carex—Eriophorum* bog, as at points —1 600 and —1 800. The points at the western end of the line also have a sparse occurrence of *Gramineae* pollens. Along line A XX (Fig. 7) the absolute numbers of pollen grains are low and vary little. At points —800 and —3 000 the AP count rises above the general level.

The greatest variation in the absolute quantities of pollen is met with along line A XXX (Fig. 8). At point +200, at the western end of the line, which is situated in a hay field and in the immediate proximity of which pine and birch trees grow in fairly equal abundance, interspersed with a few alders, while spruce is met with at some distance away, both the AP and NAP counts are high. Among the latter, *Rumex*, *Gramineae*, *Epilobium* and *Compositae* pollens are most abundantly represented. Also east of the middle of the bog, the absolute quantities of pollen are high in the wet, mesotrophic area grown over with *Carex*. There *Ericaceae*, *Cyperceae*, *Gramineae* and *Rumex* pollens are common among the NAP, but from point —1 600 eastward the commonest is *Myrica gale*. At the point in the line mentioned, *Myrica gale* grows abundantly, but it affects the absolute pollen counts only locally and spreads at most 100 m.

Fig. 10 presents a general picture of the absolute quantities of pollen in Sanemosse. It makes clear that the arboreal pollens are chiefly distributed over an area extending only a few hundred meters from the forest and that the limit of 200 grains falls on the average 300—400 meters from the forest's edge. The point of land jutting into the bog in the vicinity of point A XX has been created by a ditch extending 200—300 m from this point and along the edges of which trees grow, scattering pollen around the locality. In this case, too, the local effect can be plainly seen. Further, over 200 AP grains are found in two separate areas, one of which is situated near the western margin and the other in the eastern part of the bog. The explanation seems to lie in the flight capacity of spruce pollen, which, as pointed out in the foregoing, is capable of flying over the woods bordering the bog. Wind conditions are also likely to affect the occurrence of these separate areas. It is possible that winds blow more freely and strongly from the fields stretching across to the SW and SE margins of the bog than from other directions and thereby cause a considerable spread of pollens across the bog. In the middle of the bog there is a narrow area where the AP count is below 100 grains per mg of peat. The distance even to the bordering woods is over a km, which indicates that of the pollens of ordinary forest species, only a small portion flies in the Sanemosse region farther than one km. *Pinus* occurs in greatest

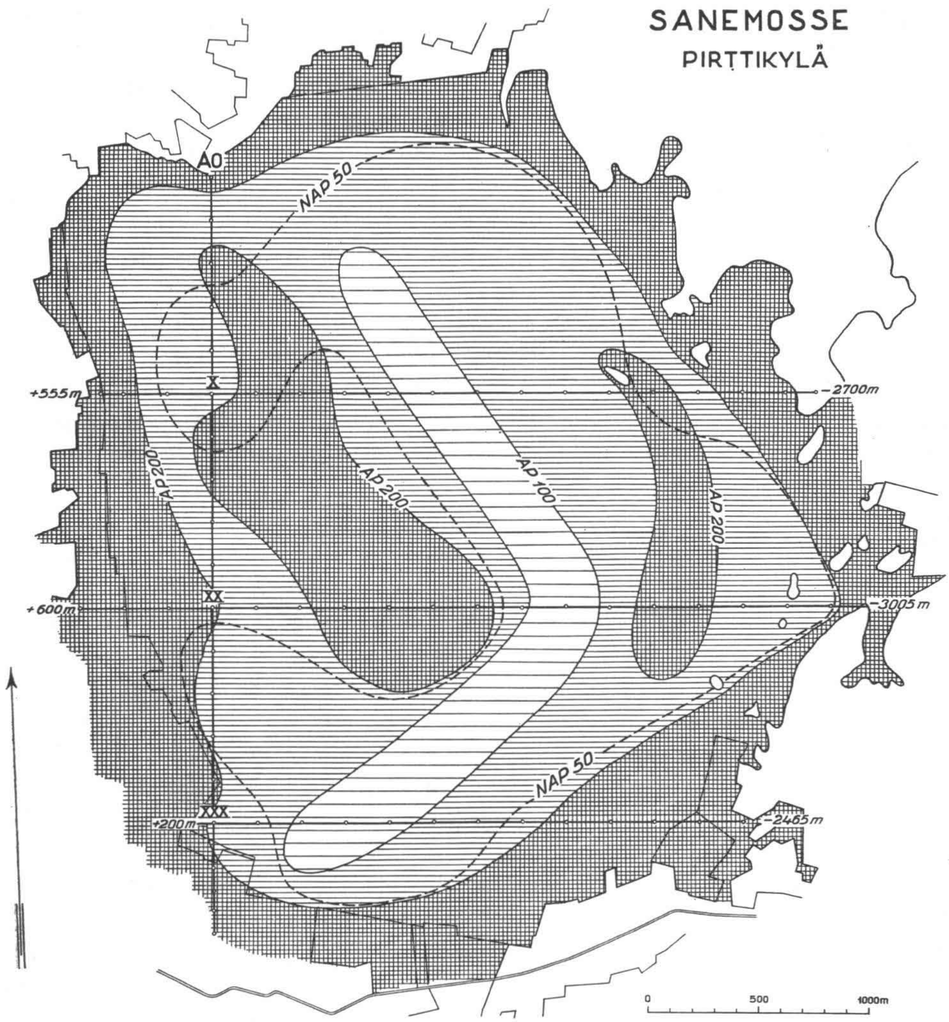

Fig. 10. Investigation lines and points of Sanemosse, together with absolute pollen frequencies, recorded separately for AP and NAP. Numbers show pollen density PF/100 mg of peat.

abundance — the grains being lightest in weight tend to be carried farthest. In the pollen statistics of the middle of the bog, *Picea* is greatly underrepresented in comparison to its share of the trees growing along the margin of the bog.

The distribution of NAP over the bog is slight, indeed, scarcely worthy of notice. More than 50 grains of NAP per 100 mg of peat occur only in the margins of the bog. Counts of less than 50 grains have been made over a broad area in the middle of the bog, where *B. nana*, *Rubus chamaemorus* and *Eriophorum* are commonest, representing local production. The boundaries along the different lines between the margin and the middle of the bog are sharp. The highest grain counts are met with in the fields, in the pine-grown stretches of the bog with subshrubs and in the mesotrophic *Carex* stretches.

THE PROPORTION OF *Betula nana* AMONG BIRCH POLLENS

Fig. 11 gives the percentage of *Betula nana* among the pollens of arboreal birches found in surface samples.

Along line A the share of *Betula nana* varies between six and twenty-six per cent. Its percentage is lowest at the beginning of the line. Arboreal birch grows there in the woods bordering the bog, but *Betula nana* occurs only sparsely in the bog near the beginning of the line. At point VI the percentage of *B. nana* is highest. At this point no birch in the tree class grows in the bog and the distance to the bordering woods is some 500 m. Though *B. nana* grows in only moderate numbers on the spot, the aforementioned circumstances have the effect of bringing the share of *B. nana* conspicuously to the fore in the pollen count. In the center of the line its proportion approaches the values at the beginning of the line. Between points XII and XXIV the pollens of birches growing along the ditches and the margin of the bog have evidently spread to the vicinity of the line, with the result that the proportional share of *B. nana* is slight. The part of the line approaching its terminus runs across a stretch of pine bog with rich *B. nana*. The large share of this dwarf birch is locally evident here in spite of the fact that around this portion of the line there also grows arboreal birch to the extent of 20—30 %, in addition to pine.

Along lines A X and A XX the occurrence of *B. nana* is very much the same. At the western margin of the bog, it registers the lowest count and in the middle, where arboreal birches and trees of any species are generally lacking, the highest. The share of *B. nana* decreases again at the eastern margin of the bog. Along line A X the maximum proportion of *B. nana* is 24 % of the birch pollens and it occurs at a point where the bog represents a treeless type with hummock ridges. There is a sparse growth of *B. nana* on these ridges, but since the places where arboreal birches grow are at quite a distance, along the margin of the bog, the share of the local, if sparse *B. nana* is evident in the total pollen material. The effect of the pine-grown stretch with abundant *B. nana* situated at the eastern end of the same line is likewise in evidence. Along line A XX, proceeding eastward from the —600-meter point there occurs water-logged bog with hummock ridges and at point —1 600 pine bog with rich *B. nana*. The entire area is likewise marked by an abundance of *B. nana* pollen, compared with other species of birch.

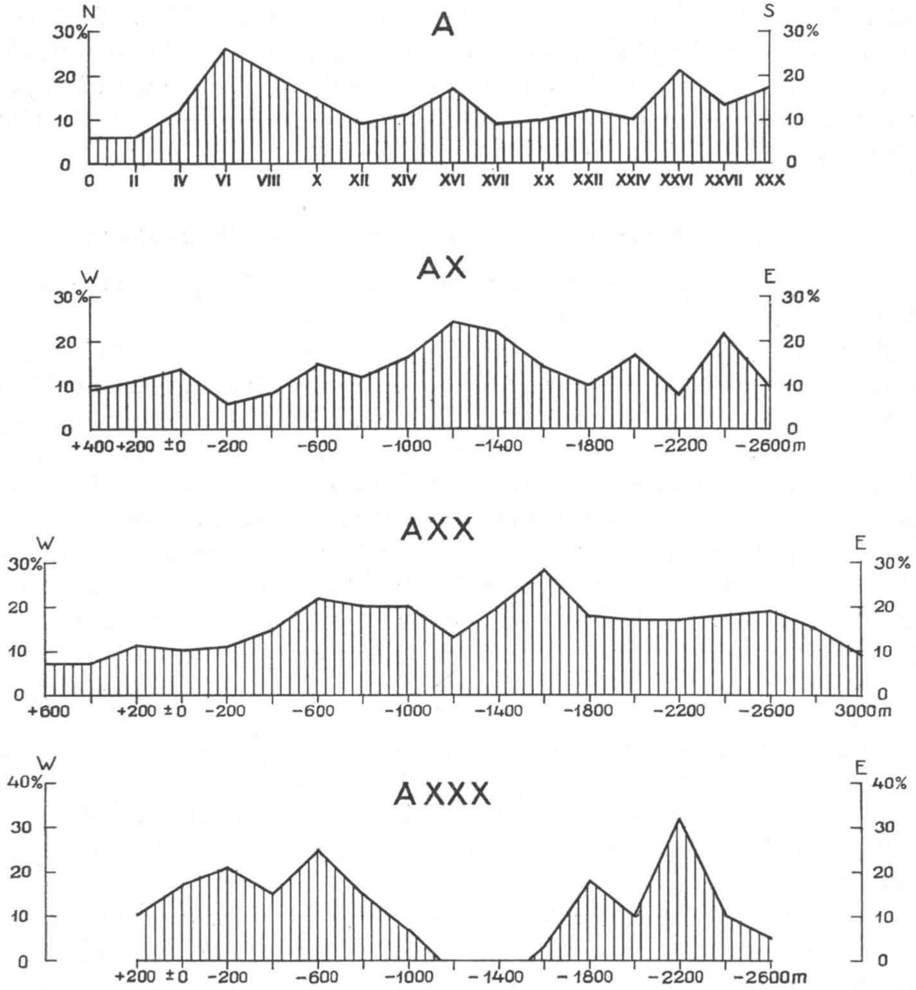

Fig. 11. Percentages of *Betula nana* pollen among all birch pollens along investigation lines of Sanemosse.

Farther east there occurs *Carex* bog in which *B. nana* grows sparsely, but as a treeless stretch it has a relatively high proportion of dwarf birch.

Line A XXX clearly deviates from the others. In its western part the proportion of *B. nana* is remarkably high (10—25 %) in spite of the fact that near the line arboreal birches grow. However, the area consists of pine bog with a rich occurrence of *B. nana*. Accordingly, this is locally reflected by the pollen statistics, as was also noted in the southern part of line A. In the central part of the line, no *B. nana* whatsoever has been found. The area consists either of wet *Sphagnum papillosum* bog or mesotrophic *Carex* bog where *B. nana* does not grow. The same types of bog are to be found 200 m west and east of the point under consideration, and there one may find a sparse occurrence of *B. nana* pollen, or 3—7 %. Judging by these circumstances, *B. nana* pollens spread only about 100—200 m from the spots where the species grows. At point —2 200 the share of *B. nana* is 32 %. The area consists of pine bog with subshrubs, but only 200 m away is the edge of the forest, where both pine and, sparsely, birch are represented. The latter, however, is unable to cut down the share of the locally prominent *Betula nana*, for 200 m appears to be about the transportation limit of arboreal pollens in any quantity. Closer to the forest's edge, the proportion of *B. nana* declines.

In muddy hollows the share of *B. nana* among birch pollens is approximately the same as it is in specimens taken from hummock ridges and plant-covered spots elsewhere as well. Although the absolute pollen count rises in them considerably, the relative positions of arboreal birches and *B. nana* remain just about unchanged.

THE POLLENS IN PEAT LAYERS

At point XX along line A in Sanemosse, a series of samples was taken from the peat deposit for pollen analyses. *Betula nana* was separated from the other birch species by the method described in the foregoing. Otherwise, the pollen diagram (Fig. 12) has been made in the usual way. Attention has also been paid to the occurrence of NAP, and their absolute numbers were counted along with those of AP from all the samples investigated at depth intervals of ten centimeters. The series of samples presented in Fig. 12 includes in its lower part allochtonous peat, silt clay and, at the bottom, clay. Since the allochtonous peat contains salt-water diatoms, just as do the sediments below, it can be concluded that the bog began to form immediately after the region had risen out of the sea as a consequence of the land upheaval.

Judging by what has previously been brought to light by pollen investigations carried out in the near environs of Sanemosse (Brandt, 1949), the paludification did not begin there until the beginning of the sub-Atlantic period, considerably after the general spread of the spruce, which time falls in the pollen diagram around the four-meter stage, the sediment being silt clay.

In the lower part of the pollen diagram, *Pinus* is by far the most common species up to the turn of the sub-boreal and sub-Atlantic periods. The point referred to must be situated, on the basis of the powerful rise of the spruce and the general decline of the mixed oak forest, at about the 3.7-m-level in

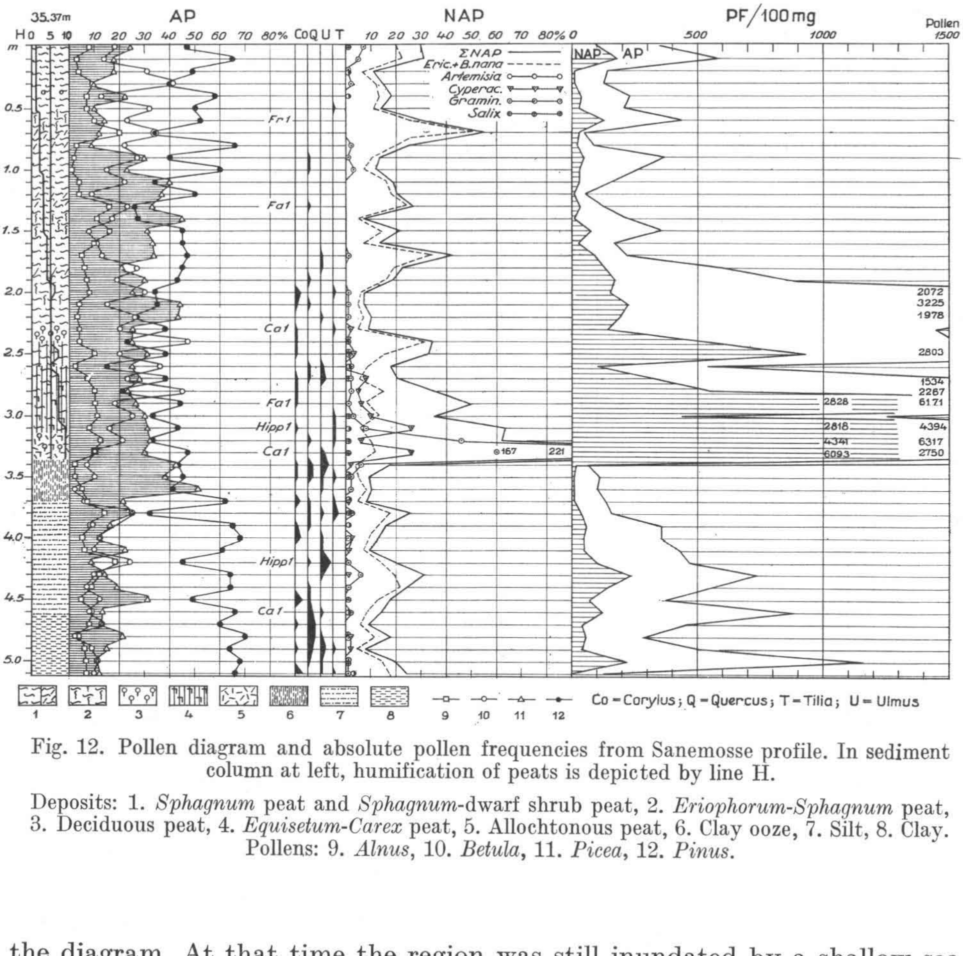

Fig. 12. Pollen diagram and absolute pollen frequencies from Sanemosse profile. In sediment column at left, humification of peats is depicted by line H.

Deposits: 1. *Sphagnum* peat and *Sphagnum*-dwarf shrub peat, 2. *Eriophorum-Sphagnum* peat, 3. Deciduous peat, 4. *Equisetum-Carex* peat, 5. Allochtonous peat, 6. Clay ooze, 7. Silt, 8. Clay. Pollens: 9. *Alnus*, 10. *Betula*, 11. *Picea*, 12. *Pinus*.

the diagram. At that time the region was still inundated by a shallow sea, but soon afterward paludification got under way. The prominence of pine on the levels of clay and silt clay, which belong to the sub-boreal, possibly also to the Atlantic, forest-historical period, is exceptional. The pollens of the periods referred to are generally deciduous. The strong occurrence of *Pinus* must in this instance be attributed to enrichment brought about by the sea, as mentioned in the foregoing. *Picea* is also abundantly present for the same reason. In peat *Picea* pollen is ordinarily never met with, or only incidentally, at the corresponding stage.

Proceeding from the beginning of the sub-Atlantic period upwards, first coniferous pollens and slightly afterwards also *Betula* pollens alternate pretty much on a par all the way up to the 2.0-meter level, after which *Pinus* dominates the diagram. At the same stage, the mixed oak forest pollens, which are present in surprising quantity so high in the diagram, suddenly diminish. Only *Quercus* and *Tilia* are met with incidentally higher up. After the stage in question, also the absolute quantity of arboreal pollens declines steeply, the peat turns predominantly to *Sphagnum*, and its degree of humification lessens. Obviously, these changes have a common cause. The factor is either local, connected with the evolution of Sanemosse, or general, affecting a broader area. *Picea* occurs quite prominently in the diagram up to 3.7—1.0 m, but then declines in approaching the modern era.

The occurrence of NAP is relatively abundant throughout the diagram. Clear and sharp peaks in its segment, however, warrant closer scrutiny. Of these the one appearing at 3.2—3.4 m stands apart from the rest. In it *Gramineae (Phragmites)* pollens predominate, but also *Cyperaceae* is abundant compared to the quantities met with elsewhere. Among grasses mention should be made of *Comarum*, *Filipendula ulmaria*, *Lycopus* and *Umbelliferae*, and among subshrubs and shrubs, *Ericaceae* and *Hippophaës*. There is also a fair abundance of *Equisetum* and *Polypodiaceae* spores, which have not been included in the NAP category. The strong occurrence of NAP here thus is associated with the shore stage. The other NAP peaks are due chiefly to *Ericaceae* and *Betula nana* and probably represent dry periods in the development of Sanemosse. Particularly from the 2.5-m level upwards, the share of these pollens among the NAP is notable, whereas in the littoral zone it is quite unimportant.

Brandt (1949) has discovered while carrying on investigations in the surroundings of Sanemosse the existence in the bogs of six so-called recurrence levels, representing the change of a dry period to a moist one, following the general appearance of *Picea*. In addition, there is a seventh, dating back slightly prior to the advent of *Picea en masse*. Brandt's studies were based on *Sphagnum* species determinations and variations in kinds of peat. He took note of the fact that in most cases the pollens of *Picea* — and also *Betula* — become more abundant while those of *Pinus* decline following the recurrence levels.

The pollen diagrams also reveal a certain amount of variation in the kinds of peat in Sanemosse, likewise suggesting recurrence levels. It is due to variations in the occurrence, on the one hand, of subshrub and *B. nana* pollens and, on the other, of types of peat, paralleled also by the variations described by Brandt in the pollen quantities registered by spruce, birch and pine. Thus the joint peaks of *B. nana* and *Ericaceae* pollens are in many cases accompanied by the occurrence of dwarf shrub peat — both phenomena indicating a dry stage in the evolution of the bog. At these points also *Pinus* is prominent. By contrast, *Picea* and *Betula* increase at higher levels, evidencing a wet phase. Such recurrence levels are to be observed at the following depths, starting from the top: 0.2, 0.7, 1.3, 1.8 and 2.5 m, most distinctly, but also at 3.0 and 3.9 m, or numerically at seven levels, as Brandt has demonstrated. The use in the present study of NAP to define the recurrence levels thus presupposes the separation of *B. nana* pollen from that of other birch species, a measure not, however, usually undertaken in connection with pollen investigations.

The fact that the proportions of *B. nana* and *Ericaceae* pollens among the NAP between 3.5 and 2.5 m in the diagram are smaller than elsewhere, particularly the upper part of the diagram, is probably due to seashore conditions as well as the well-known dampness of the climate at the beginning of the sub-Atlantic period. Accordingly, the subshrubs and *B. nana* had not thriven in the bog in the same degree as later, when the climate became drier. Under the influence of the land upheaval and the thickening of the bog, Sanemosse also grew progressively drier with the approach of the modern era.

The absolute pollen frequency varies in the diagram within a wide range. This observation applies to both AP and NAP counts. At the level of the water sediments in the lower part of the diagram, the AP rises at short intervals to three peaks, which the NAP grains follow; but the quantities of each category steeply decline as the water level sinks. The values are at a minimum at the contact of the silt and the allochtonous peat.

In the allochtonous peat, which represents a juncture when the locality has risen out of the sea, first the NAP count (6093 grains per 100 mg) and immediately thereafter the AP count (6 317 grains) reaches a high maximum, in terms of absolute numbers, in the diagram. Both categories remain high except for a few brief dips at the 2.4-m level, where the AP plunges steeply, and at 1.8 m, where the AP likewise takes a dive. Proceeding upward from these points, the NAP count remains low with the exception of a couple of humps. By contrast the AP count varies considerably and in its segment four distinct surges can be observed.

At the stage when the region of Sanemosse rose out of the sea and the process of paludification started, the undermost peat layers bear witness particularly to the formation of *Phragmites*-sumpf, which Brandt (1949) has met with on the nearby seacoast even at the present day. The luxuriant plant life covering the exceedingly wet shore bog of that past period produced an abundance of both AP and NAP. Arboreal pollens also flew over from the surrounding forests and through the agency of water became enriched in the allochtonous peat. It is apparent that the damp conditions also preserved the pollen grains better than the subsequent dryish surface of the bog was able to. Even higher up the wet bog type continued to prevail, represented by a grassy sedge bog that produced *Equisetum—Sphagnum—Carex* peat. Also at this level the pollen counts are high. The first signs of drying appear at the depths of 2.4—2.6 m, containing dwarf shrub and deciduous peats. There the NAP count drops under 200 grains, and the AP count likewise drops to a corresponding low at 1.8 m, which is featured by the occurrence of dwarf shrub peat. The bog was apparently already then approximately the same size as it is today and the prevailing conditions corresponded to those of the present.

It is interesting to remark that the absolute pollen frequency decreases at higher levels, too, whenever the peat indicates a dry period in the evolution of the bog and the *B. nana* and *Ericaceae* pollens achieve their common peaks. This observation takes the form of steep drops in the course of the pollen segment — and involved are both AP and NAP. This quantitative variation in the pollen diagram thus corresponds to the observations made of the occurrence of recent pollens in the muddy hollows and hummocks of Sanemosse, as reported in the foregoing. Accordingly, variations corresponding to those found in the pollen proportions of surface samples are to be seen also in the diagram illustrating the pollen contents of the peat layers.

A CRITICAL EXAMINATION OF THE RESULTS

In the surface samples from Sanemosse, *Pinus* dominates the pollen statistics. This is particularly true of the samples taken from the middle of the bog. The result is most clearly in evidence in the material from the central parts of lines A X and A XX, which are situated farther from the sphere of influence of the forests. This demonstrates that *Pinus* pollen, as compared to other arboreal pollens, fly best for distances of about 1.5 km and, as is known, even farther, though it is not possible within the scope of the present study to wander so far afield. In the distribution of *Picea* pollen, at a distance of approximately one km from the edge of a spruce-dominated forest there is a steep drop to be observed. The distribution of *Betula* pollens is still more limited, for at a distance of 400 m from the woods lining the margin of the bog, the birch pollens account for less than 20 % of the total. This is best brought to the fore by the material from line A X. Inasmuch as the percentage of arboreal *Betula* is very low, except for a few spots, in relation to that of coniferous trees, the conspicuous local occurrence of the species among the pollens warrants notice. On the basis of this study of the Sanemosse bog, the spread of *Alnus* pollen mainly in the near proximity of stands of the species appears likewise to be clear. At the western end of line A XX there is an open space 80 m from growing alders where pollen grains of these species make up 38 % of the total — but 200 m farther on the figure drops to 9 %, 400 m away to 2 % and at a distance of 1 000 m down to zero. The absolute count for 100 mg of peat correspondingly falls: 73 grains, 16, 5 and 0. At the eastern end of the same line there is a cluster of alders in a confined forest space, but only 20—30 m away *Alnus* pollen has a frequency higher than 2 %, and 100 mg of peat contains only 11 pollen grains representing the species. At a distance of a km from this spot, in the bog, not a grain of *Alnus* pollen is to be found. Accordingly, between 100 and 200 meters seems to be the principal radius of distribution for *Alnus* pollens.

The pollen production of *Pinus* is three-fold compared to that of *Picea* and over two-fold compared to *Betula* (Sarvas, 1955 a), so this, too, ought to be considered in evaluating the occurrence of pollens. The results of the survey of the dissemination of pollen over the Sanemosse area largely correspond to the specific gravities presented by Pohl (1937 a), as reported in the foregoing. Thus, the lightest *Pinus* and *Picea* pollen grains are borne the farthest. There seems to be a discrepancy between *Alnus* and *Betula* insofar as the lighter *Alnus* pollen would be expected to spread more readily than the *Betula*. *Alnus incana*, however, is generally a smaller tree than *Betula*, and this holds true of the Sanemosse vicinity, too. Furthermore, it grows there sparsely in comparison with the birch and the *Alnus* stands are in confined spots. These circumstances obviously prevent gaining a true picture of the distribution of *Alnus* pollens. Nevertheless, according to Lüdi (1947), *Alnus* pollen grains do spread less readily than *Betula*. Noteworthy, too, is the fact that the sample taken from a muddy hollow at point —600 along line A XX contains 10 % *Alnus*, amounting to 23 pollen grains, and that the adjacent hummock yields not a single grain of the species. This signifies that *Alnus* pollens are borne distances of considerably over a kilometer, but in relative terms their occurrence is so sparse that they make a showing in the statistics only fortuitously, except where muddy hollows are involved — there, as elucidated in the foregoing, the main mass of coniferous pollens has altered the original pollen composition through the sorting action of water.

The results reported by Hesselman (1919) and Sarvas (1956 b) concerning the long-distance flights of pollen grains and those by von Post (1924) on the distribution of *Picea* pollen do not support those obtained from Sanemosse. On the other hand, the investigations conducted by Aario (1940) in Lapland yielded fairly parallel observations. Particularly closely do the results reported by Sarvas (1956) in his study of seedling *Betula* stands and by Erdtman (1921) in his investigation of the distances traveled by birch and alder pollens agree with those obtained in carrying out the present study.

The pollens of shrubs, subshrubs and grasses spread in the main only about the immediate surroundings of the plants. Thus, *Betula nana* and *Myrica gale* have been observed to distribute their pollen around the immediate locality but not to any notable extent beyond a distance of a hundred meters, notwithstanding the fact the places they grow in are mostly wide open. Both species have a strong local influence on the pollen statistics. This has a certain signficance as far as *B. nana* is concerned, for the pollen of this species is usually included in pollen studies among arboreal *Betula* species. As *B. nana* has been set apart from the birch trees proper in the present study, the proportions registered by this species have been observed to run high in comparison with other birches. In the middle of the treeless stretch of bog and in the pine-grown stretches featured by the presence of subshrubs, the species accounts for between 22 and 32 % of the total pollen count. In samples from wet types of bog, where *B. nana* does not grow any closer than 300—400 m away, its pollen is totally absent — which demonstrates the strictly local character of its distribution. Comparing the two uppermost diagrams in Figs. 5—8, one will note that *B. nana* nevertheless fails to upset the pollen statistics appreciably, at least in the surface samples; this is evidently true also of the stratigraphic investigation, except for, perhaps, late-glacial times, during which *B. nana* played a prominent role. Thus, *e.g.*, Hyyppä (1941) has noted that in the sediment dating from the early Yoldia stage in Kuusamo between 45 and 50 % of the *Betula* pollens were of the *B. nana* type; and, likewise according to Hyyppä (1936), some 30 % of the birch pollens contained in the Dryas strata on the Karelian Isthmus consist of *B. nana*. As reported by Aario (1940), *B. nana* does not appreciably figure in pollen research in northern Finland. Surprisingly insignificant, too, according to Eneroth (1951), is the representation of *B. nana* in the pollen diagrams from northern Sweden.

The distribution of grass pollens is of still more limited local character than that of shrubs and subshrubs. The reason for this is the tininess of the pollen grains and, in many cases, the small pollen output. *Rumex acetosella*, which grows in profusion in the cultivated fields of the southern part of Sanemosse and the pollen production of which is remarkably prolific, spreads pollen abundantly around the local area, but at a distance of only 100 m the occurrence of pollen grains of this species is quite fortuitous. Open spaces promote the spread of grass pollens. Accordingly, the proximity of the cultivated fields at the southern end of line A and at the western and eastern ends of line A XXX contributes to the heavy representation of NAP, notably *Gramineae* pollen.

The influence of pools in sorting the pollens is apparently responsible for the sharp *Betula* upthrusts in the pollen diagrams. Comparably, the samples taken from the edges of bog pools lead to increases in the proportions of *Pinus* and *Picea* pollens. Also the absolute quantities of pollen grains seem to be remarkably high in places where pools are located, compared to samples taken from adjacent hummocks and hummock ridges. This would suggest that pollen grains are preserved better in pools and muddy hollows than in the relatively dry conditions prevailing in the hummocks. The abundant occurrence of *Betula* pollens, in particular, in samples taken from pools and muddy hollows is probably in part due to the fact that the resistance of birch pollens is lower than that of coniferous pollens on the dryish surface of the bog, where, accordingly, the latter appear to be enriched in moss-covered areas.

The absolute quantities of pollen in Sanemosse are surprisingly low in the middle of the bog: *e.g.*, along line A XXX the count is about 100 grains per 100 mg of peat, while corresponding values are met with along other lines as well as in the pollen profile. The values cited are below those obtained by Aario (1940) as averages from surface samples collected on tundras. He holds the AP value of 69 per 50 mg of peat as characterizing tundra also in pollen diagrams. This, however, does not jibe with the results obtained at Sanemosse. It would therefore seem that local conditions also affect the absolute quantities of pollen quite fundamentally.

The interest commanded by the pollen diagram of Sanemosse is concentrated mostly in the NAP and absolute pollen counts. The figures yielded are particularly high for the littoral stage.

Since *Ericaceae* and *B. nana* have been combined in the same category, their fluctuations reveal a conformity with the variations in the occurrence of different types of peat insofar as the peaks of the former coincide with the appearance of dwarf shrub peat, *i.e.*, dry horizons. These horizons evidently correspond to Brandt's (1949) recurrence levels, which he has observed both at Sanemosse and in nearby bogs. Thus, they would seem to be regional and not bound up exclusively with the evolution of Sanemosse. The depressions in the absolute pollen curve occurring in conjunction with dry periods — representing a decrease in the pollen content of the samples — agree well with the results obtained from the surface samples collected at Sanemosse. Accordingly, the frequency of pollen is lower in moss-covered areas than in wet boggy stretches and muddy hollows and pools.

Attention is further drawn to the fact that among the pollens contained in surface samples collected from Sanemosse, *Picea* is clearly under-represented in relation to the occurrence of spruce in the forests surrounding the bog. *Pinus*, on the other hand, is over-represented, especially in the middle of the bog. *Betula* is also locally well represented in the marginal parts of the bog, as well as in the middle, considering the occurrence of birch in the sphere of the bog. The same may be said of *Alnus*.

When the pollen diagram of Sanemosse is examined with this in mind, it may be concluded, especially on the evidence of the surface samples, that the frequency of spruce does not nearly come up to what the occurrence of the species in the terrain surrounding the bog would seem to presuppose. Thus *Picea* is distinctly under-represented and *Pinus*, again, over-represented, together with *Betula* and *Alnus*. The overall impression is that *Picea* is under-represented in the pollen diagrams.

As a final conclusion, it may be stated in the light of the research done with material from Sanemosse that in conducting any pollen investigation one should avoid taking a sample series — in the event that a single series must suffice — from the central part of an extensive bog. It would give an exceedingly erroneous picture of the history of the forests occurring in the surrounding country and of the evolution of the plant life as well as of the absolute quantities of pollens. The relative frequencies are liable to be confounded by the pollens from local stands of trees and other plants growing in the bog. Variations in the surface of the bog also have a decisive influence on the pollen composition and on the absolute numbers of pollen grains.

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8

RADIOCARBON DETERMINATIONS FROM THE BOG PROFILE OF LAPANEVA, KIHNIÖ, WESTERN FINLAND [1])

BY

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ABSTRACT

Three radiocarbon determinations have been carried out at the C 14-Laboratory of the Department of Superficial Deposits of the Geological Survey of Finland from a bog profile taken from Lapaneva, Kihniö commune, western Finland. The lowermost sample, representing the point of the pre-boreal *Betula* maximum, yielded the result 9 850 ± 320 years B.P. The determination from the point of the boreal *Pinus* maximum yielded the value 8 100 ± 200 y. B.P., and the one from the point marking the general spread of *Picea* 4 100 ± 100 y. B.P. With the exception of the last, the values correspond to the conception prevailing earlier regarding the postglacial geochronology. However, the emergence of *Picea* as a common genus in western Finland has been believed to be a later occurrence. At the sampling point where the C-14 determination was made, there was a carbon layer, indicating a bog fire. Accordingly, the age arrived at may conceivably be too great. Consequently, the author is leaving the date of the emergence of *Picea* as a common genus open. It will be up to later research to fill the gap.

CONTENTS

[1]) Received June 2, 1962.

INTRODUCTION

When the C 14-Laboratory built for the Dapartment of Surficial Deposits of the Geological Survey of Finland began to be used in 1960, the author had on hand for radiocarbon determinations a series of samples collected in the fall of 1958, on the basis of investigations carried out in 1952, from a bog known as Lapaneva situated in the commune of Kihniö, western Finland. The geographical position of Lapaneva is 62°15′ N. Lat., 23°18′ E. Long. and the altitude of the sampling site 163.1 m above sea level. This bog appeared to be suitable because the preliminary investigation proved its pollen diagram to be clear-cut and organic deposits to have begun to form in the area as early as the pre-boreal period, which is by no means a common phenomenon in these parts. Furthermore, the thickening process in this barren watershed bog had been so slow that it was possible to dig up the sample series with a spade all the way down from the bottom.

In 1958 Lapaneva was still in a natural state, representing a treeless *Sphagnum—Eriophorum vaginatum* bog; but since then it has had ditches dug into it for the exploitation of the peat. The walls of the sampling pit made it possible to examine the bog profile more closely. In this connection attention was drawn to layers of carbon at a depth of about a meter. From the vertical wall of the pit a series of samples was taken, starting from the bottom and proceeding to the surface, by pressing plastic boxes into the peat and cutting it loose with a knife. The boxes measured 8 × 12 × 40 cm.

After the series of samples had been taken, Mr. Lauri Aaltonen, engineer in charge of Suo Oy's peat-digging operations in Aitoneva, a bog next to Lapaneva, reported having salvaged part of the stem of a large pine found underneath the peat layers. The top of the stem was still in place *in situ*, and the bottom part was likewise still left. It had been badly burned on one side. The fire that had damaged the tree might presumably have had some connection with the carbon layers of Lapaneva. A large part of the stem of the pine had been so well preserved that planks, out of which souvenirs were later made, could be sawed out of it. The thickness of the bottom of the trunk was about half a meter, while the height of the tree was estimated to have been between 27 and 30 meters. On the basis of the annual growth rings, I calculated the age of the tree to have been about 100 years.

The pollen analyses incorporated into the present study were performed by Miss Ester Uussaari, Mag. Phil., of the Department of Superficial Deposits, and the radiocarbon determinations were carried out in the C 14-Laboratory of the same department by Mr. V. Hoffrén, Mag. Phil., and Mr. Aulis Isola, Lic. Phil. The diagrams were drawn by Mrs. Lyyli Orasmaa and the manuscript was translated from the Finnish into English by Mr. Paul Sjöblom, M. A. I would like to acknowledge my debt to all the persons mentioned.

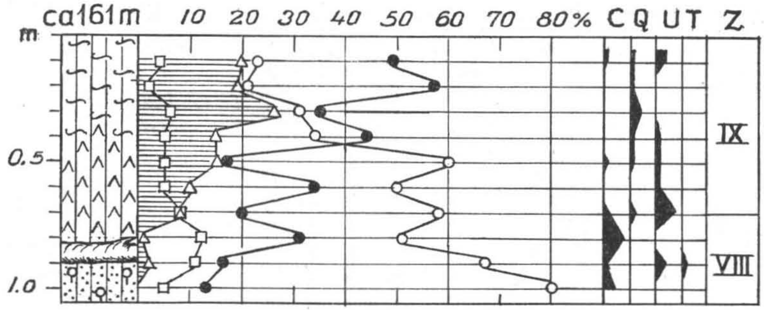

Fig. 1. Pollen diagram from locality where ancient pine was discovered in the bog of Aitoneva. Explanations of signs in Fig. 2.

DETERMINING THE AGE OF THE PINE FOUND IN AITONEVA

The location of the pine trunk found in the bog profile of Aitoneva is shown in Fig. 1. The trunk lay on sandy ground containing the remains of deciduous trees. It was covered by a peat layer consisting in the lower part of coniferous *Carex* and in the upper part of *Sphagnum—Carex* peat. Peat to a depth of fully half a meter had been removed from the area. A series of samples was taken from a point near the top of the pine where the stem measured about ten centimeters in diameter. The altitude of the surface of the bog was about 161 m above sea level.

The division of the pollen diagram into forest-historical zones can be carried out on the basis of *Picea* and mixed oak forest pollens without difficulty. Thus, the lower part of the diagram belongs to the sub-boreal and the part above the depth of 0.7 m to the sub-Atlantic stage. In the light of the diagram, the pine trunk may be said to have become covered with peat a short time before the general spread of *Picea*, revealed here by the strengthening of the spruce pollen sector, beginning at 0.8 m, slightly before the turn of zones VIII and IX.

The radiocarbon determination carried out from the bottom part of the pine trunk gave the result

$$4\,350 \pm 100 \text{ y. B.P.}$$

The number of the determination at the C 14-Laboratory of the Geological Survey is Su-5.

The evidence indicates that the pine perished in a forest fire at the end of the sub-boreal period but remained standing for a long time afterward. Only after the locality began to turn into a bog did the dead tree fall. Peat deposits must have covered the pine very quickly, for it has been preserved in remarkably good condition for about 4 300 years.

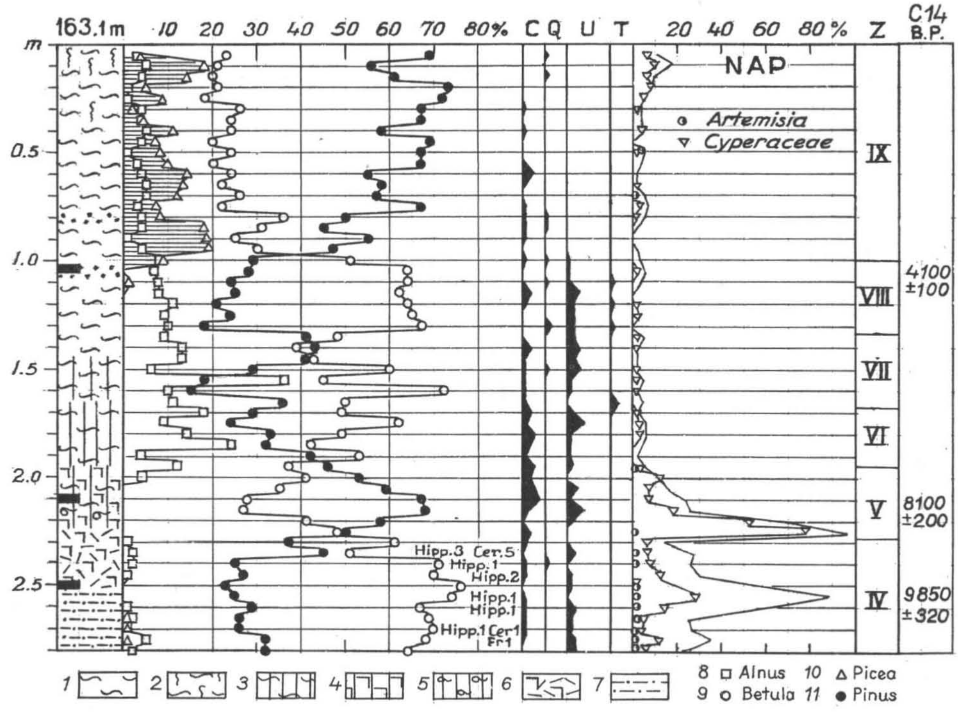

Fig. 2. Pollen diagram from sampling site in Lapaneva. Black rectangles at left indicate spots where samples were taken for C 14-determinations. 1 = *Sphagnum* peat, 2 = *Eriophorum-Sphagnum* peat, 3 = *Sphagnum-Carex* peat, 4 = *Equisetum-Carex* peat, 5 = *Bryales-Carex* peat, 6 = *Equisetum*-allochtonous peat, 7 = silt, 8 = *Alnus*, 9 = *Betula*, 10 = *Picea*, 11 = *Pinus*, C = *Corylus*, Q = *Quercus*, U = *Ulmus* and T = *Tilia*.

DATING THE PEAT PROFILE OF LAPANEVA

The Lapaneva series of samples was taken some two or three kilometers to the NE from the sampling site of Aitoneva by the buried pine. The bog stretches lengthwise in the NW—SE direction, and it covers an area of about 125 hectares. The series of samples was dug up from the middle of the extensive southern part of Lapaneva, more precisely stated about 20 m westward from point A VIII in the network of lines drawn up in the bog investigated in 1958.

Fig. 2 shows the depth of the profile, the arrangement of layers and the pollen diagram made from the sample series. Lowermost, overlying the till, is 25 cm of silt, followed by 35 cm of allochtonous peat, which contains principally remains of *Equisetum*. The allochtonous peat turns into *Carex—Sphagnum* peat after a transitional stage of *Equisetum—Bryales—Carex* peat, and around the depth of 1.5 m the peat turns to *Sphagnum*, with *Eriophorum* as an accessory constituent at the surface.

According to the diatom analysis, the undermost silt specimen contains *Nitzchia tryblionella* (1 %), which is a brackish water species. Otherwise, the diatoms contained in the sample belong among fresh-water flora. At the level of the allochtonous peat, the diatoms represent species characteristic of a lake undergoing a filling-in process. To judge by the macro-fossils, the bottom sediments likewise contain only fresh-water flora, signifying that in the area of Lapaneva the lake stage dates back to the pre-boreal period. The filling-in gradually led to paludification. The fact that abundant *Hippophaës* pollens have been noted in the silt under the microscope points to the proximity of the seashore. Calculating on the basis of the relation diagram drawn up by Virkkala (1959) from Satakunta, the turn of zones IV and V, or the position of the water level indicating the end of the Yoldia stage, was approximately at the elevation of 154 m in the Lapaneva area, *i. e.*, six or seven meters below the level at which the filling-in of the bog took place.

The zone division of the pollen diagram is clear-cut. It is questionable, however, whether the part of the profile below 2.6 m belongs to Zone IV or Zone III. The sparse occurrence of NAP, however, does not speak for the latter. For this reason, it has been classified as belonging to Zone IV. The diagram is characterized by three *Betula* periods. The oldest of them, the pre-boreal one, has the highest frequency of birch pollen, or 70—75 %. The second, and most instable, *Betula* period coincides with the initial and middle stages of the Atlantic period. It weakens toward the end of the period but surges upward again at the beginning of the sub-boreal to its third period of dominance. There are two *Pinus* periods, of which the older, the boreal is clear and strong. Continuing throughout the sub-Atlantic period and growing stronger toward the modern era, the second *Pinus* period lays its stamp prominently on the upper part of the diagram. The *Alnus* sector is particularly strong during the Atlantic period, starting with the end of which it weakens toward the modern era. The rise of *Picea* begins slightly before the transition from Zone VIII to Zone IX. The abundance of mixed oak forest pollens is a surprising feature of this watershed bog. The variations in their frequency emphasize the zone division made on the basis of pollens. The *Corylus* and *Ulmus* sectors are strong and nearly continuous from Zone IV to the end of the sub-boreal. Evidently both plants have grown in the region since the beginning of the pre-boreal period. In respect to *Corylus*, this conception is further bolstered by the fact that in a bog situated a couple of kilometers from Aitoneva to the west sub-fossil nuts have been found, which will be dealt with in detail in a subsequent paper. The frequency of NAP is highest at the contact of silt and allochtonous peat in the lower part of the diagram and at the contact of allochthonous peat and telmatic peat slightly higher up; at these points sharp peaks are registered by the pollens, and in both cases *Carex* is the genus in the majority. Otherwise, the frequency of NAP is low.

Three radiocarbon age determinations have been carried out from the points in the Lapaneva profile shown in Fig. 2. For the purpose, a flat piece of peat three cm thick and otherwise measuring 8 × 12 cm was used. The depths from which the samples were taken and the age determinations obtained from them are as follows:

No. of sample in C 14-Laboratory	Depth	Age in years B.P.
Su— 6	ca. 103 cm	4 100 ± 100
Su— 3	» 210 »	8 100 ± 200
Su—15	» 250 »	9 850 ± 320

Radiocarbon age determinations have been used with the aim of verifying certain points of reference generally occurring in the pollen diagrams of Finland and particularly, in the case at hand, in the western part of the country. Accordingly, the oldest possible organic deposit of Lapaneva was selected for investigation. It indicates the beginning of the filling-in process in the area of the bog and corresponds in the pollen diagram to the time of the *Betula* maximum, dating back to the latter half of the per-boreal period. Another significant niveau coincides with the boreal *Pinus* maximum, which corresponds to the time of the maximum height of the water level of the Ancylus lake in the pollen diagram. Moreover, it closely approximates the beginning of the *Alnus* rise (Ao), regarded as important in Sweden (G. Lundqvist, 1957). The third important niveau is the beginning of the general spread of *Picea,* which in western Finland has customarily been dated at the end of the sub-boreal period, often at the point where it turns into the sub-Atlantic.

CONCLUSION

Finland has so far produced few C 14-determinations for publication. Okko (1960) has published a determination (St 527) carried out at the C 14-station is Stockholm from a sample taken near Helsinki, which on the basis of pollen analysis was classified as belonging to Zone VII of the Atlantic period. The result 2 600 ± 115 y. B.C. corresponds to the pollen-chronological dating. Korpela (1962) has published the result of a determination made at Copenhagen's C 14-station from a specimen taken from a peat layer discovered in northern Finland between two beds of till. The age obtained was >35 000 years, signifying an interglacial peat layer.

The C 14-Laboratory of the Geological Survey of Finland recently delivered a report on thirteen of the determinations it has carried out to the geochronological laboratory of Yale University (Hyyppä, Hoffrén, Isola, 1962). The determinations here presented were included in the report, with the exception of the oldest. Additional comparative material is available from elsewhere, especially Sweden, but before launching upon comparisons in that direction, let us take a look at geochronological timetables drawn up in this country by Sauramo (1958) and Hyyppä (1960).

The oldest of the determinations included in the present study agrees well with Sauramo's dating of the pre-boreal Yoldia *Betula* maximum. The age arrived at by the C 14-determination for the boreal *Pinus* maximum, which, as mentioned in the foregoing, corresponds to the maximum of the Ancylus-transgression, coincides with the Ancylus-transgression in Sauramo's chronology as well, so the datings likewise jibe at this point. Hyyppä's chronological table (1960), which has not been divided into forest-historical zones, also agrees with the C 14-determinations discussed. By contrast, at the point marked by the topmost specimen, there appear to be considerable discrepancies, particularly in comparing Sauramo's chronology. The Lapaneva sample, which was taken at the point marking the emergence of *Picea* as a common genus, yielded the result 4 100 $\pm$ 90 y. B.P. In Sauramo's chronology, the dating of the corresponding point is approximately 3 200 y. B.P. and in Hyyppä's chronology it is 3 800 y. B.P. The problem posed requires further elucidation.

It was observed in the foregoing that at the point in question in the Lapaneva profile, carbon occurs as evidence of an obvious bog fire. Thus, it is possible that at that point peat had burned to a depth of between ten and twenty centimeters, which is quite a plausible estimate in the case of an unchecked bog fire. The peat of the layer containing carbon could therefore be older than the pollen diagram would indicate. A comparison of the rise in the *Picea* pollen sector of Lapaneva with Fig. 1 and many other diagrams from western Finland as well as Sauramo's (1958) general diagram reveals it to be strangely sharp. This could well indicate that the point in question truly lacks peat. This could also signify that the Lapaneva bog fire and the forest fire in which the great pine of Aitoneva perished were not simultaneous events but that the latter occurred much earlier than the former. The difference between their ages would thus in reality be greater than what the C 14-determination indicates. The occurrence of numerous fires during the dry sub-boreal time is quite conceivable. There is a carbon layer, after all, also at the depth of 0.8 m in the Lapaneva profile testifying to this likelihood.

Virkkala (1950) and Hyyppä (1950) have reached the conclusion on the basis of the land uplift and of archeological findings that the spruce began to take its place among the common trees of western Finland about 1 800 B.C., and Hyppä (1936) previously arrived at the same estimate in regard to the region around the far end of the Gulf of Bothnia. According to Sauramo's geochronology, the emergence of *Picea* does not take place until about 1 000—1 200 B.C. He bases his dating on the result obtained by Fromm (1938) from the sediments of the Ångerman river regarding the emergence of *Picea* and assumes that the event was contemporaneous on both sides of the Gulf of Bothnia. The same dating has been reached by C 14-determinations in many Swedish localities, also, for example, in Lapland (G. Lundqvist, 1957). The present author (Salmi 1955), has investigated the bronze sword unearthed at Panelia, in the commune of Kiukainen. The sword was found at the boundary between fine and coarse detritus, which in the pollen diagram coincides with the emergence of *Picea*. According to archeologists, this type of southern German sword dates back to between 1 400 and 1 100 B. C. On the strength of this evidence, the conclusion was drawn that the spruce emerged as a common genus in the region of Kiukainen between 1 000 and 1 200 B.C. This dating seems convincing, but the possibility remains that the sword had sunk into sediment after falling into water. In that event the date of the emergence of *Picea* would have to go back farther. Judging by the character of the sediments in the locality, it does not appear likely, however, that the depth to which the weapon sank could have been as great as the difference between the dates for the emergence of *Picea* obtained at Kiukainen and Lapaneva would presuppose.

Since highly divergent results regarding the date of the emergence of different investigators and different research methods, the possibility should be taken into account that the spruce took its place among common trees in different localities at different times, as in the case of Värmland, Sweden, according to J. Lundqvist (1957). On the basis of pollen analyses and C 14-determinations (G. Östlund), he has demonstrated that the emergence of *Picea* in the area of Falun, in the northeastern part of the region referred to, took place in about 1 000 B.C. Proceeding southwest from that area, the spruce made its general advent later and later, so that the date of the event at Töcksmark, near the Norwegian border, some 250 km from Falun, has been set at 210 B.C. A picture illustrating this phenomenon has been published by G. Lundqvist (1957, p. 18).

The general appearance of the spruce in western Finland could have followed a similar pattern, *i.e.*, one marked by a progressively later emergence in a southwesterly direction. Since an element of uncertainty is bound up with the material from Lapaneva used in the C 14-determinations, it would be sensible to leave the date of the emergence of *Picea* there open for the time being, or until numerous supplementary datings will have cleared up the matter.

In Sweden, according to G. Lundqvist (1957), the age of Ao from Adak, Lapland, as determined by the C 14-method, was 8 575 $\pm$ 125 y. B.P., and Wenner (1960) reports the age of the same horizon in the sediments of the Indal river to be 9 100 $\pm$ 120 y. B.P. The differences are considerable and the latter, in particular, deviates to a remarkable degree from the value of the Lapaneva boreal *Pinus* maximum. A comparison raises doubts as to whether the features of corresponding pollen diagrams are generally synchronous over extensive areas. The question can be answered only on the basis of a considerable body of age-determination material.

According to Hyyppä (Hyyppä, Hoffrén, Isola, 1962) a specimen taken from the half-way point of the pre-boreal deposits in the bog profile of Vävarsbacka, near the town of Porvoo, was given the C 14-age of 9 600 $\pm$ 350 y. B.P. (Su-13). A similar analysis was made of a sample of his taken from Pölläkkälä, on the Karelian Isthmus (now Soviet territory), which represents the latter half of the pre-boreal. The result obtained was 9 850 $\pm$ 300 y. B.P. (Su-8). Accordingly, these samples and the undermost one from Lapaneva have yielded practically the identical result. The Kihniö area, according to this, had become deglaciated prior to the time mentioned, for before turning into bog the bottom of Lapaneva was covered with a silt deposit that pollen analysis dates back to the first half of the pre-boreal.

Hyyppä (1960) has published a relation diagram of the land uplift and the shore displacement that has taken place in the Baltic region since the Ice Age. Certain of the most important stages of the evolutionary history of the Baltic Sea are marked in it according to a chronology based on the author's own observations and datings derived from various other sources. From the standpoint of the present study, special interest attaches to the age of Yoldia I, namely, approximately 8 000 B.C. The age of the oldest peat of Lapaneva arrived at by a radiocarbon determination is 7 900 $\pm$ 320 B.C. — which, taking into account the margins of error, may coincide with the age of Y I — and at most about 400 years younger.

Lapaneva is situated on the 80-m Littorina isobase, according to Hyyppä (1960). At the corresponding point in the relation diagram, the altitude of Y I is c. 218 m, while the oldest peat of Lapaneva is situated at 160.5 m above sea level. Thus, in about 8 000 B.C. the Lapaneva area ought to have been covered by a sea 58 m deep. Assuming that the C 14-age determination of the oldest peat of Lapaneva is correct — as appears to be firmly supported by the other determinations discussed in the foregoing — dry land should nevertheless have existed there at that time and the paludification should have started. This somewhat contradictory result can only be explained by the fact that the age of Y I must be older than has been assumed. The age difference need not be very great, for the rate of land uplift near the center of uplift was in those times considerably more rapid than at present. In Denmark, according to Iversen (1960), the turn of the younger Dryas and the pre-boreal dates back to 8 300 B.C. On the basis of the elevation of the oldest peat of Lapaneva, the age 9 850 $\pm$ 320 y. B.P. would thus fall in the relation diagram into the latter half of the pre-boreal. The oldest peat deposits of Lapaneva, Vävarsbacka and Pölläkkälä also date back to this same time, according to pollen analyses, as pointed out in the foregoing.

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GEDANKEN ZUR TEKTONIK DER »LAPPLÄNDISCHEN GRANULITE» [1]

VON

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Im nördlichen Lappland zieht ein breiter Zug granulitischer Gesteine (Granat-Quarz-Feldspat-Gneise = Granulite, Hypersthen-Granulite und Gesteine, die zu den Charnokiten hinüberleiten) in einer 40—60 km breiten und nicht ganz 400 km langen, schön geschwungenen Zone vom Kaledonischen Gebirge am Rastegaissa durch die Tundren Finnlands und des Murmansk-Gebietes. Diese Zone endet etwa 60 km westlich der nördlichen Ausläufer des Imandra-Sees. Eine Fortsetzung findet sich östlich von Kandalaschka.

Die wasserarmen Gesteine zogen schon früher die Aufmerksamkeit der Geologen auf sich (Jernström, 1874); sie wurden von Sahama (1936), Sahama u. Mikkola (1936), Mikkola (1941, Blätter Sodankylä und Tuntsajoki) und Meriläinen (1958 [2]), bearbeitet. Eine umfangreiche Beschreibung mit Analysen, Aufschluss-und Dünnschliff-Bildern gab Eskola (1952). Das Murmansk-Gebiet fand eine neue zusammenfassende Darstellung in der Geologie der USSR, Band 27, Teil 1 (Granulit-Gebiete S. 99 ff.). Der grösste Teil des Ausstriches ist auch in Karten im Masstab 1: 1 Mio. (russischer und norwegischer Anteil) und im Masstab 1: 400 000 (finnischer Anteil) dargestellt. Nur die Kartierung des nördlichen und westlichen Teiles des finnischen Granulit-Gebietes ist noch nicht veröffentlicht. Scheumann (1961 u.a.) hat sich in letzter Zeit mehrfach zu den »Granuliten» Lapplands geäussert. Er hebt besonders die Unterschiede zu den klassischen Granuliten Sachsens hervor, die in der geringeren Durchbewegung bestehen. Die sächsischen Gesteine sind stärker durchbewegt und

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[1]) Eingegangen den 16. August, 1962.

[2]) Frau T. Mikkola hatte die Freundlichkeit, die nur in finnisch erschienene Arbeit von Meriläinen für mich zu übersetzen. Ich danke ihr für die grosse Mühe, der sie sich unterzogen hat.

haben einen engen Zeilenbau. Dieser tritt beim Anblick der lappländischen Granulite wenig in Erscheinung. Sahama konnte aber zeigen, dass die Gefügediagramme beider Gesteine auf gleichen Baustil hindeuten; aber auch er hebt die petrologischen Unterschiede hervor. Wenn ich hier bei dem eingeführten Namen »Granulit-Komplex» bleibe, so ist neben dem auch von Scheumann nicht bestrittenen Vorkommen echter Granulite in der Serie für mich ausschlaggebend, dass ich den von Scheumann vorgeschlagenen Namen »Granat-Gneise» für zu farblos halte und keinen anderen Namen ohne eingehende petrographische Untersuchung der ganzen Granulit-Frage vorschlagen möchte.

Der Verfasser hatte in den Jahren von **1941** bis **1943** mehrfach Gelegenheit, teils allein, teils in Begleitung von Prof. Laitakari und Prof. Sahama, denen herzlichst für die erwiesene Hilfe gedankt sei, das Gebiet in allen Richtungen zu durchqueren. Wenn die Ergebnisse der Reisen nur flüchtige Beobachtungen sind und eine petrographische Ausarbeitung nicht erfolgen konnte, da die gesammelten Gesteinsproben verloren gegangen sind, so hat mich die Arbeit von K. Meriläinen doch veranlasst, meine Gedanken kurz zusammenzustellen. Eine irgendwie erschöpfende Behandlung des Problems ist nicht beabsichtigt. Ein weiterer Anlass zur Veröffentlichung war die Bearbeitung der Blätter Skandinaviens für die Internationale Geologische Karte von Europa und Diskussionen mit zahlreichen Fachgenossen, mit denen gemeinsam die Tektonische Karte von Europa zu bearbeiten ich die Freude hatte.

Die Gesteine des Granulit-Gebietes kann man mit Meriläinen (auch Scheumann, 1961; Eskola, 1952) als veränderte Sedimente auffassen, die jetzt, ihres ursprünglichen Wassergehaltes beraubt, als echte Granulite, Cordierit-Granat-Gneise (mit Biotit, z.T. auch Sillimanit) und Pyroxen-Granulite vorliegen. Übergänge zu Pyroxen-Dioriten sind nicht selten. Basische bis ultrabasische Anteile kommen vor und nehmen nach Osten merklich zu. Die Pyroxengesteine häufen sich am Aussen- (Süd-) Rand. Aber auch gegen den Innen- (Nord-) Rand nimmt die Basizität zu. Das Parallelgefüge und die Streifigkeit der Gesteine ist an vielen Stellen nicht sehr ausgeprägt. Im kleinen sind Typen mit intensiver Durchbewegung am Aussen- (Süd-) Rand häufiger als im Inneren oder am Innen- (Nord-) Rand des Granulit-Gebietes. Die Pyroxengesteine kommen auch quergreifend vor, sind also wenigstens teilweise jünger als die Hauptmasse der Granulit»- Gesteine (Tafel I). Die Bezeichnung »Granulit» ist für die so bezeichneten Gesteine nicht ganz korrekt (Scheumann, 1961). Ich benutze trotzdem den Ausdruck »Granulit-Komplex».

Im Norden schliesst sich an den Granulit-Komplex mit langsamem Übergang ein Gebiet mit Hornblende-Gneisen, Granat-Amphiboliten und gewöhnlichen Amphiboliten an. In gewissem Abstand vom Granulit-Komplex nehmen die migmatischen Erscheinungen stark zu. Die Amphibolite können dann als Schollen in einer pegmatischen hellen Grundmasse schwimmen, was besonders schön an der Staumauer am Jäniskoski des Pasvik-Flusses zu sehen war (Tafel II, Fig. 2—4). In den Hornblende-Gneisen sitzt ein grosses Massiv von Charnokiten, die als Intrusion aufgefasst werden. Weiter gegen Norden gehen die Hornblende-Gneise allmählich in die im Grundgebirge gewöhnlichen Biotit-Gneise über, die vielfach völlig migmatisiert sind. Ihnen liegen, jeweils durch eine Diskordanz getrennt, die präkambrischen Serien von Kirkenes und Petsamo auf; die Petsamo-Formation ist von der allgemeinen Granitisierung nicht mehr erfasst worden, die Varanger-Formation nur in geringem Masse.

Die Petsamo-Formation wurde in der russischen Literatur neuerdings in das Ordovizium eingestuft. Diese Einstufung beruht auf den Funden *Collenia*- und *Gymnosolen*-ähnlicher Fossilien in den Kalklagen und einem Vergleich mit ähnlichen Funden in den Kalken der Porsanger-Serie, die von Holtedahl (1918) in das Ordovizium gestellt wurden. Holtedahl hat aber 1932 das präkambrische Alter dieser Serie klar erkannt und Gründe dafür angegeben (Gerölle in den eokambrischen Konglomeraten u.a.). Man kann wohl unter diesen Umständen eine Einstufung der Petsamo-Serie in das Präkambrium (jüngeres Proterozoikum?) auch weiterhin vertreten.

Bei eingehender Untersuchung der Räume im Norden des Granulit-Bogens wird man sicher noch viele Unregelmässigkeiten und Abweichungen vom allgemeinen schüsselförmigen Bau finden. Grosse, etwa NS-streichende Diaphtorese-Zonen sind dem Verfasser an vielen Stellen bekannt.

So ist unter dem ehemaligen Turisthotel von Liinahamari am Petsamo-Fjord eine solche Diaphtorese-Zone vorzüglich aufgeschlossen gewesen. Die Überscherung des postkinematischen Kristallin-Gefüges und die Umbildung der Gneise in Quarz-Biotit-(Chlorit-) Schiefer ist ausgezeichnet zu beobachten (Tafel III, Tafel II, Fig. 1).

Zuerst stellt sich ein scharf geregeltes, neues »s» ein. Es bilden sich Lagengneise. Bei weiterer Annäherung an die Störungszone verschwinden die eingeschichteten hellen Gesteine. Biotit mit einer Neigung zur Vergrünung bildet neben Quarz den Hauptbestandteil des scharf geregelten und sehr weichen Gesteines.

Alle Anzeichen deuten auf einen wichtigen grossen Bewegungshorizont hin. Er streicht etwa NNW und ist vielleicht eine Fortsetzung des weiter unten behandelten karelischen Risses.

Die Scherzone ist in grosser Tiefe entstanden, da sich bei der Durchbewegung Biotite gebildet haben. Die pt-Bedingungen dürften nicht wesentlich von denen abgewichen sein, die bei der Auskristallisation der umgebenden Gneise bestanden haben.

Der ganze grosse Raum nördlich des Granulit-Bogens dürfte mit den Granuliten zusammen mehrfach einheitlich verformt worden sein. Allerdings sind die nördlichen Anteile unter anderen Bedingungen (wasserreich) aus- und umkristallisiert als die Granulite. Dies gilt nicht mehr für die Petsamo- und die Varanger-Formation, die deutlich jünger als ein grosser Teil der Durchbewegung, Deformation und Metamorphose der älteren Serien sind.

Ganz anders liegen die Verhältnisse im Süden und Westen des Granulit-Bogens. Bis auf wenige Kilometer an die Granulite heran finden wir den »Karelischen Bau» unversehrt vor. Der Granulit-Bogen stösst dabei an ganz verschiedene Glieder des Karelikums. Im Osten sind es vorwiegend die Grundgebirgsgneise (Migmatit-Gneise), das Basement der Kareliden, die »Bjelomoriden», im Westen die eingefalteten suprakrustalen Gesteine des Karelikums. In der Mitte wechselt in der Nähe der Granulit-Grenze die Zusammensetzung. Teils stossen die Amphibolite und Quarzite, teils das präkarelische Grundgebirge an den Granulit-Bogen.

Im Gegensatz zur Nordgrenze, an der man einen langsamen Übergang von den Granuliten bis zu den Biotit-Migmatit-Gneisen zu erkennen glaubt, liegen an der Südgrenze des Granulit-Komplexes zwei sehr verschiedene Baustile nebeneinander. Im Süden und Westen die unruhigen Synklinalen des Karelischen Baues, im Osten und Norden der grosszügigere und viel einheitlicher gefügte Granulit-Bogen. Allerdings scheint nach den Untersuchungen von Meriläinen der innere Bau der Granulite nicht so ruhig zu sein, wie das bisher vorliegende Kartenbild vermuten lässt. In einiger Entfernung von dem Südrand glaubt er einen Sattel- und Muldenbau mit wechselndem Achsenfallen zu erkennen. Wenn dies richtig ist, so mutet die glatte Kurve der Südgrenze, die zwei Gebiete mit einem dem Schlingenbau ähnlichen Grossgefüge trennen würde, umso fremdartiger an.

Man kann an dieser Südgrenze ein generelles Einfallen gegen das Innere der Granulite sowohl in den überscherten Teilen der Kareliden und Bjelomoriden wie in den »Granuliten» selbst feststellen. Dies ist auch von Mikkola und Sahama (1936), Mikkola (1941) und Eskola (1952) schon klar herausgestellt.

Vielfach schalten sich zwischen Granulit und den südlichen Kareliden-Bjelomoriden-Komplex noch stark ausgeschieferte Amphibolite und Hornblendegneise ein, die den Südrand weithin begleiten. Diese Gesteine schmiegen sich in ihrer Verformung — nach dem Kartenbild und dem äusseren Anschein der Gesteine zu urteilen — der Grenze beider Zonen an.

Besonders deutlich wird das Auflager der Granulite an der Salnyje-Tundra, wo die Granulite muldenförmig den alten Gneisen des Südkomplexes aufliegen (Charitonow, 1958, S. 623). Ich wiederhole mit kleinen Veränderungen die dort gegebene Abbildung (Abb. 1). Es ist für mich nicht zweifelhaft und wohl auch von allen Bearbeitern nie ernstlich bezweifelt worden, dass der Granulit-Komplex auf dem südlichen Kristallin liegt.

Die gegenseitigen Altersbeziehungen des Süd- und Nordkomplexes lassen mehrere Deutungen zu. Da die Grenzzone den Bau der Kareliden rücksichtslos überschneidet, muss sie j ü n g e r als die Faltung der Kareliden sein, in die sie sich in ihrem Verlauf und in der Art der Verformung kaum einordnen lässt.

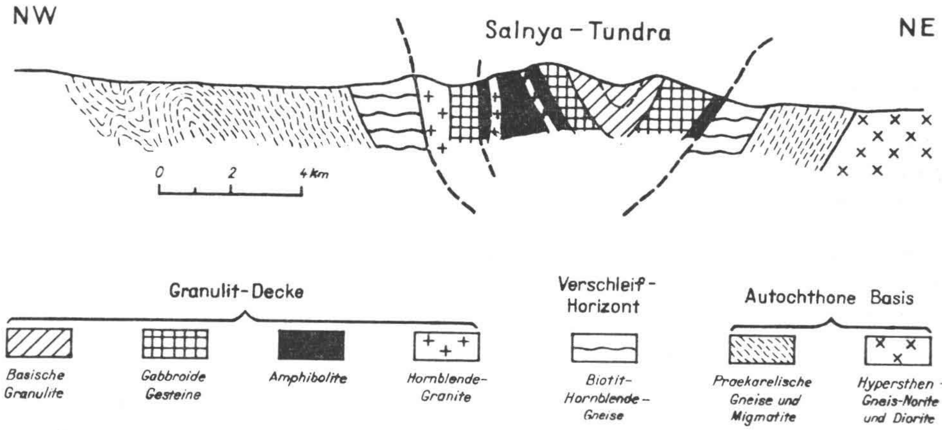

Abb. 1. Profil durch die Salnyje-Tundra (nach E. H. Wolodin aus Charitonow ,1958, S. 623, Abb. 56).

Die Schichten des Nordkomplexes selbst können jünger oder älter als die Kareliden sein. Da sich bis jetzt im ganzen nördlichen Raum bis zum Eismeer keine sicheren Äquivalente der Kareliden finden liessen, kann eine solche Deutung nicht grundsätzlich abgelehnt werden. Die von mir behandelte Südgrenze des Nordkomplexes könnte das Auflager einer (oder mehrerer) jüngerer Serien auf ein Basement sein; der ungewöhnlich ruhige Verlauf der Grenze würde dem ruhigen Verlauf des Auflagers auf ein Basement entsprechen, wie wir es so oft unter intensiv gefalteten Serien finden (z.B. *Hyolithus*-Schiefer des Kambriums vor den Kaledoniden auf älterem Gneis). Gegen diese Deutung sprechen allerdings gewichtige Gründe. Der Versuch von Väyrynen (1938), die Petsamo-Formation als karelisch anzusehen, ist wohl allgemein abgelehnt worden. Da die Petsamo-Formation bestimmt jünger als die Gneiskomplexe unter ihr ist, würde eine solche Gleichsetzung einwandfrei besagen, dass der Nordkomplex älter als karelisch ist. Leider ist eine haltbare Beweisführung für diese Ansicht nicht so einfach. Auch die Einstufung der Kjev-Serie der östlichen Kola-Halbinsel ist noch zu wenig gesichert, um Gleichsetzung sicher vorzunehmen. Auch ihre Gleichsetzung mit den Kareliden würde auf ein präkarelisches Alter der Schichten im Nordkomplex hindeuten.

Vom Standpunkt der Petrofazies aus muss man wohl — entsprechend der allgemeinen Erfahrung — die Granulit-Serie tief ins Grundgebirge einstufen, tiefer als die Kareliden, die vielfach kaum die Amphibolit-Facies erreicht haben. Dem entspricht auch, dass im Inneren der Schüssel, welche die Granulite bilden, auf den Granuliten wieder Gesteine der Amphibolit-Facies — mit einer Übergangszone von Granat-Amphiboliten — liegen, die den ganzen Norden aufbauen. Auch die Migmatisierung ist im wesentlichen unter pt-Bedingungen der Amphibolit-Facies erfolgt.

Dieser Auffassung entspricht auch die von Scheumann und seinen Mitarbeitern entwickelte Vorstellung über die Genese der »Granulite» Lapplands. Sie sind nach ihm in ständig zunehmender Metamorphose aus den Migmatit-Gneisen des Nordgebietes entstanden. Seine erst 1955 entwickelte Auffassung, die Cordierit-Zeilen (u.a.) als diaphtoretische Neuprägungen anzusehen, hat er (1961 mit Mitarbeitern) zu Gunsten einer nur in einer Richtung aufsteigenden Metamorphose aufgegeben. Meine Feldbeobachtungen stimmen am Nordrand (»Innen»-Rand) des Granulit-Komplexes ganz mit der letzteren Auffassung von Scheumann überein. In diesem Raum spricht nichts für die Annahme einer absteigenden Metamorphose. Anders liegen die Verhältnisse am Südrand (Aussenrand), wo meiner — allerdings nur auf makroskopischen Beobachtungen gegründeten — Ansicht nach eine retrograde Metamorphose (Granulit-Facies → Amphibolit-Facies) unter Wasserzuführung auf z.T. sehr eng gestellten Bewegungsbahnen nicht zu übersehen ist.

Hinzu kommt noch, dass nach den Feststellungen Polkanow's und seiner Mitarbeiter die Gesteine des Nordens (einschliesslich ihrer Hauptdeformation) älter als die des Südens sind: Saamiden mit mehr als 2 000 Mio. Jahren gegen Kareliden mit 1 700 Mio. Jahren im Süden. Ob die Vereinigung der Granulite mit den Bjelomoriden 1 950—2 100 Mio. Jahre, die Polkanow vorgenommen hat, wirklich haltbar ist, könnte nur ein Studium jeder einzelnen Altersbestimmung zeigen. Mir scheint, als ob hier z.T. der Einfluss jüngerer Deformation älterer Gesteine ein Alter, das zu jung ist, vorgetäuscht hat. Auf jeden Fall, selbst wenn man die Altersdeutungen von Polkanow und Gerling voll anerkennt, bleibt ein höheres Alter der Granulite gegenüber den Kareliden bestehen.

Es ist in Zusammenfassung dieser und einiger weniger wichtiger anderer Überlegungen (Parallellisierungen in der östlichen Kola-Halbinsel, generelle Zusammenhänge mit den Saamiden, die auf den Granuliten liegen) mit grösster Wahrscheinlichkeit anzunehmen, dass längs der Südgrenze der Granulite Älteres auf Jüngerem liegt. Ich wiederhole ferner, dass die Deformation an der Grenzzone jünger als die Faltung der Kareliden ist.

Wenig wichtig scheint mir eine Diskussion zu sein, ob hier eine Überschiebung oder Überfaltung vorliegt. Auf jeden Fall müsste bei einer Überfaltung der Mittelschenkel so ausgequetscht sein, dass der Charakter der tektonischen Zone dem einer Überschiebung nahe steht. Die Linsen von Granat-Amphibolit, die sich örtlich noch einigermassen gut erhalten haben, können als Äquivalente der nördlich an die Granulite anschliessenden Granat-Amphibolite und damit als eine Art Mittelschenkel aufgefasst werden. Die Serpentinite usw. stammen wohl aus grösserer Tiefe (vgl. auch die Ausführungen über die Montsche-Tundra).

Die Fortsetzung dieser Granat-Zone nach Osten ist noch zu erörtern, zumal sich daraus noch einige interessante Zusammenhänge zu ergeben scheinen. Die Betrachtung des Kartenblattes Murmansk 1: 1 Mio (vgl.Abb. 2, in der dies Blatt verwertet ist) zeigt, dass genau dort, wo die Pyroxengranulite und Charnokite der Salnyje-Tundra, nach einem nordwestlichen Verlauf, welcher der Muldenflanke der NNW-streichenden Salnyje- (Decken-) Mulde entspricht, ihr Nordost-Ende finden, nach einer kurzen Unterbrechung am Noto-See wieder Gesteine einsetzen, welche weiter westlich den Übergang der Granulite in die Migmatitgneise bilden; sie ziehen nach

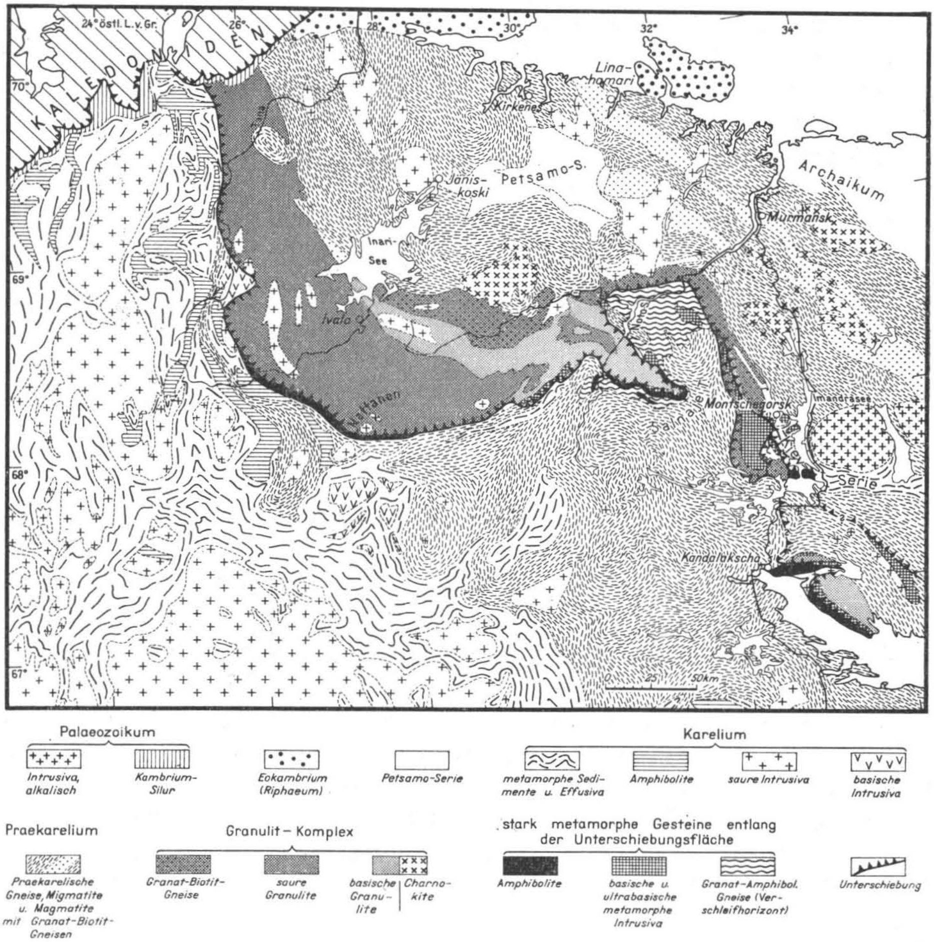

Abb. 2. Übersicht über den »Granulit»-Komplex Lapplands und seine Umgebung.

Osten in EW-Richtung weiter. Ich nehme an, dass ihre Südgrenze die Fortsetzung der tektonischen Zone ist, welche im Westen den Südrand des Granulit-Komplexes bildet; ich sehe — mit anderen Worten — den »ungegliederten Komplex der Granat-Biotit- und Biotitgneise mit untergeordneten Pyroxen- und Amphibolit-Varietäten und ihren Migmatiten» als das Dach des Granulitkomplexes an, das hier noch allein erhalten ist. Die Zusammensetzung dieses Daches ist von der des eigentlichen Granulit-Komplexes nicht so grundsätzlich verschieden, wie man es nach der getrennten Darstellung auf allen Karten vermuten sollte. Die Schilderungen von Scheumann (1961) zeigen dies deutlich. Der Granulit-Komplex ist in der Bucht östlich der Salnyje-Tundra abgetragen.

Vor den Biotit-Granat-Gneisen liegt in grösserer Ausdehnung ein anderer Komplex, der als »Hornblende- und Biotit-Hornblende-Gneis mit untergeordneten Biotit-, Biotit-Granat-Varietäten und Amphiboliten» bezeichnet wird. Dies Gestein begleitet immer den Südrand der »Granulite» und ist wohl mit den ausgeschieferten Hornblende-Gneisen, die ich weiter im Westen unter den Granuliten fand, identisch. Ich halte diese Gesteine für einen Verschleif-Horizont, in dem Granulite und Gneise des Kareliden-Basement (Bjelomoriden) bei der Südbewegung des Granulit-Komplexes erneut durchbewegt worden sind. Diese Gesteine würden hier in $\pm$ flacher Lagerung (bzw. als »dünne Decke») unter der Bewegungszone erhalten geblieben sein und einen Schluss auf den Untergrund der Granulite weiter im Westen gestatten (vgl. auch das wenig weiter südlich gelegene Profil GDE der Karte der Kola—Halbinsel, Blatt Murmansk 1: 1 Mio.).

Die Granat-Gneise, die das Dach der Granulite bilden, biegen nach längerem EW-Verlauf nach S um, und unsere tektonische Zone geht in eine sehr steil gestellte NS-laufende Störungszone über, die schon von unseren russischen Kollegen ausführlich bearbeitet worden ist.

Es ist die Störungszone, an der die grossen Komplexe ultrabasischer Gesteine in der Montsche-Tundra eingeklemmt sind. Die Störungszone, welche die ultrabasischen Gesteine in E und W einschliesst, fällt sehr steil nach E ein. Sie leitet schliesslich nach 100 km Länge wieder zu den Gesteinen des »Granulit-Komplexes» am Südende des Imandra-Sees und in einer noch nicht ganz erkennbaren Weise zu dem »Granulit-Komplex» im E nach Kandalakscha hin. Bezeichnenderweise treten am Südufer der Kandalakscha-Bucht nach der Karte wieder jene Gesteine auf, die ich für Gesteine eines Schleifhorizontes an der Basis der Granulite halte und die sonst den Bjelomoriden fremd sind.

An diesem Lineament ist der »Granulit»-Komplex um 150 km verschoben worden, im E der Linie relativ nach S, im W von ihr ebenso relativ nach N. Die Störungszone der Montsche-Tundra entspricht einer riesigen Blattverschiebung.

Ich deute den Bewegungsvorgang hier ähnlich, wie es Stille (1953) für die Deutung der Tektonik am S-Rand der Karpathen ausgeführt hat (Danubischer Riss, S. 164 ff, vgl. Abb. 3). Der Block der Bjelomoriden und Kareliden im Westen der N—S-Störung hat sich nach N vorgeschoben und daher auf einem Bereich von mehr als 250 km die Granulite unterfahren. Die Unterschiebung ist im Raum zwischen Salnyje-Tundra und Montsche-Tundra mit einem Verschiebungsbetrag von etwa 100 km sichtbar. Es handelt sich also um gewaltige Verschiebungen, die mit den Bewegungen am Danubischen Riss der Karpathen in einem höheren Stockwerk durchaus zu vergleichen sind. Ich nenne den Riss Karelischer Riss. Er hat möglicherweise Fortsetzungen nach N bis an den Rand des Eismeeres, z.B. am Petsamo-

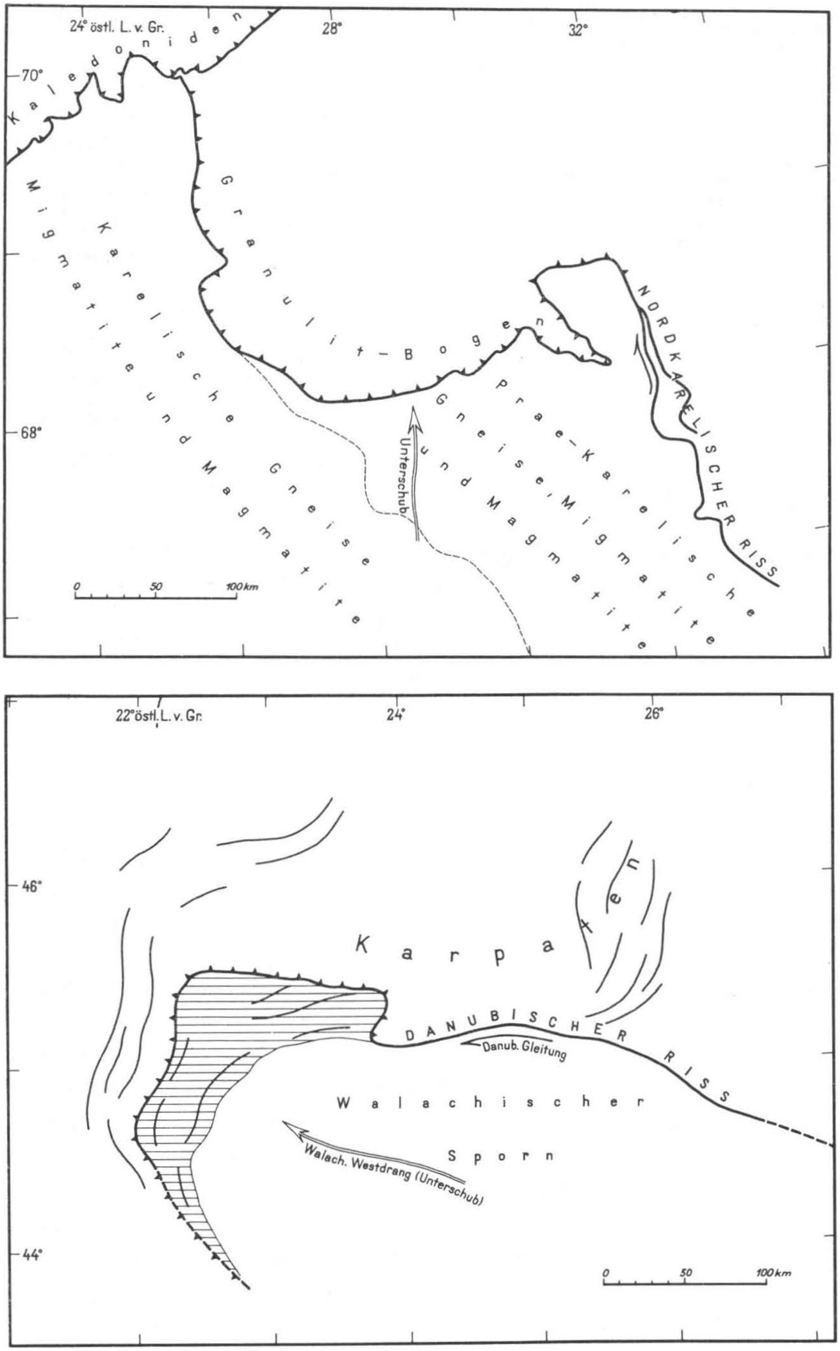

Abb. 3. Vergleich des Granulit-Komplexes mit dem Karpathen-Bogen in der Deutung von Stille.

Fjord. Die Bewegungen an dieser senkrechten Blatt-Verschiebung sind aber geringer als im Bereich der Montsche-Tundra.

Es scheint mir ferner wichtig, dass am Innenrand der »Granulite» und in den anschliessenden Amphiboliten und Migmatit-Gneisen junge Mobilisationen häufiger als sonst in diesem Raum des Baltischen Schildes auftreten. Sie sind fast nie von einer intensiveren Durchbewegung erfasst worden (vgl. Taf. II, Fig. 2—4, aber auch Taf. I). Man könnte in ihnen Mobilisate aus der unterschobenen und damit in andere pt-Bedingungen geratenen Masse sehen, ähnlich wie es Stille für den Vulkanismus im Innenbogen der Karpathen dargelegt hat.

Die für das Grundgebirge junge Bewegung sowohl an der Überschiebung wie auch am Karelischen Riss erfolgte unter pt-Bedingungen, die der Amphibolit-Facies entsprechen. Sie müssen jünger als die karelidischen Faltungen sein und könnte vielleicht mit den Deformationen zusammenhängen, welche die Varanger-Formation vor der Ablagerung der Petsamo-Formation erlitten hat.

Die Unterschiebung könnte mit den tief in das Erdinnere eingreifenden Bewegungslinien verglichen werden, die wir rezent als Herde der Tiefbeben kennen. Wir hätten dann zum ersten Mal die Möglichkeit, eine solche Bewegungszone in grösserer Tiefe, natürlich immer noch nicht so tief wie die tiefsten Erdbebenherde, zu studieren.

ZUSAMMENFASSUNG

Der Lappländische Granulit-Komplex ruht mit einer tektonischen Deformationsfläche auf einem aus Karelikum und seinem Basement aufgebauten Kristallin. Die Bewegung an dieser Fläche ist jünger als der karelische Bau.

Die Granulite sind vermutlich bedeutend älter als das Karelikum und vermutlich auch älter als die Bjelomoriden.

Die tektonische Bewegungsfläche unter dem Granulit-Komplex wird im Osten von einer gewaltigen Blattverschiebung abgeschnitten, in deren Bereich auch die Ultrabasite der Montsche-Tundra zu stellen sind. Die Blattverschiebung hat vielleicht weniger bedeutende Fortsetzungen bis zum Eismeer am Petsamo-Fjord.

Der Zusammenhang mit der Blattverschiebung legt die Deutung der Bewegungen unter dem Granulit-Komplex als eine Unterfahrung nahe; ein Vergleich mit der Unterfahrung unter dem Karpathenbogen, für die Stille die gleiche Deutung gegeben hat, ist in der Art wie in dem Ausmass der Bewegung möglich (Danubischer Riss — Karelischer Riss).

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ERLÄUTERUNGEN ZU DEN TAFELN

TAFEL I

»Granulit» Blöcke an der Strasse Ivalo—Inari bei km 22. Diese Aufnahmen stammen vom Nordrand des Granulitbogens. Bilder von mehr äusseren Teilen in Eskola (1952).

Fig. 1. »Granulit» mit ausgesprochenem Parallelgefüge. Dunkle und helle Schlieren deuten das frühere Migmatit-Gefüge an (vgl. Taf. III). Granat in Zeilen und als Rand um dunklen Einschluss. Jüngere Verschiebung mit hellem Metatekt gefüllt (Überprägung bei der Unterschiebung?).

Fig. 2. »Granulit» mit ausgesprochenem Lagen- und Zeilenbau. Grosse Granate in hellen Partien gehäuft. Jüngere Bruchzonen mit Umwandlungen.

Fig. 3 u. 4. Dunkler »Pyroxengranulit» mit viel Pyrit und Biotit-Säumen als Quergriff im »Granulit» mit verwaschenem Parallel-Gefüge.

TAFEL II

Fig. 1. Nördl. Liinahamari (vgl. Taf. III). Fast nur noch Biotit-Material mit z.T. neugesprossten oder regenerierten Feldspäten, die dem Gestein ein Perlgneis-Gepräge verleihen.

Fig. 2—4. Migmatisierte und mit Metatekten durchschwärmte Amphibolite. Felsen unterhalb der Talsperre Jäniskoski.

Fig. 2. Eingeschlichtete, helle Metatekte und Quergriffe, wohl einer späteren Injektions-Phase.

Fig. 3. Parallel-Gefüge mit deutlicher synkinematischer Überscherung. In den Boudinage-Räumen pegmatitische Neukristallisation.

Fig. 4. Das gleiche wie Fig. 3, aber am unteren Rand sehr helles, deutlich posttektonisches Metatekt.

TAFEL III

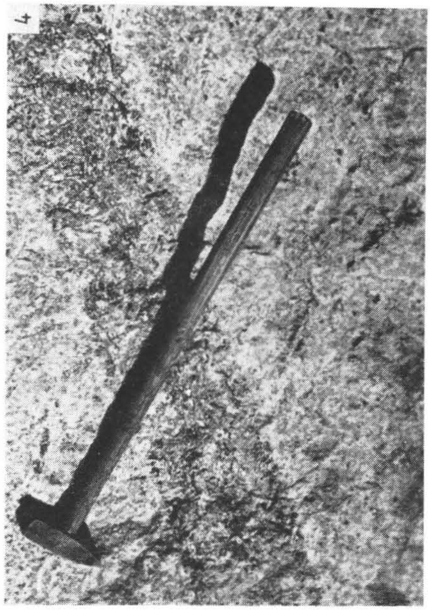
Aufschlüsse nördlich des Turist-Hotels Liinahamari am Hafen am Südende des Petsamo-Fjordes. Jüngere, NNW-streichende Störungszone mit Bewegungen in Amphibolit-Facies.

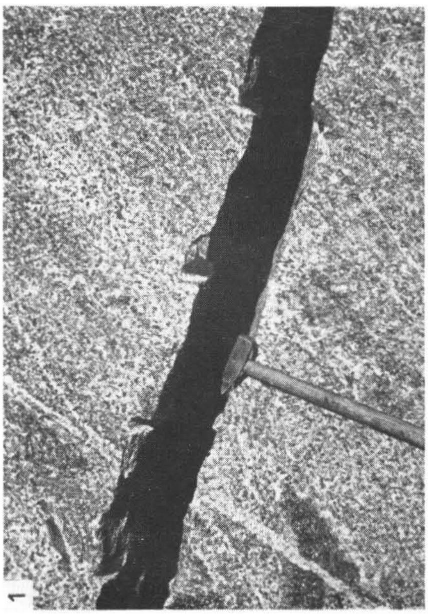
Fig. 1. Migmatitgneis (Cordierit-Biotit-Granat-Gneis) mit reichlich hellem, hier postkinematischem Metasom (Metallstab 2 m).

Fig. 2. Beginnende Zerbrechung und Einregelung des Metasoms.

Fig. 3. Biotitschiefer der Scherzone. Metasome teils aufgebraucht, teils völlig eingeschlichtet. Biotite leicht vergrünt.

Fig. 4. Einschlichtung verstärkt (vgl. auch Taf. II, Fig. 1)





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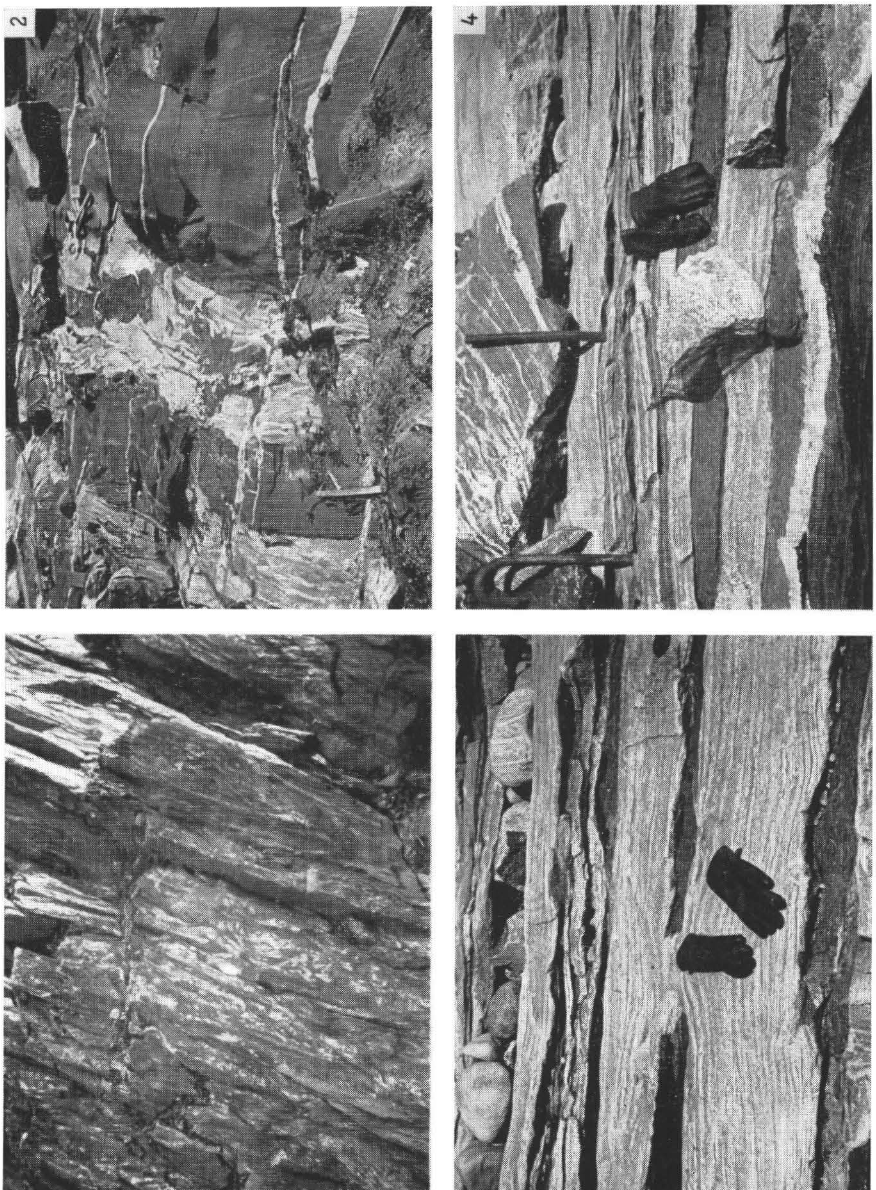

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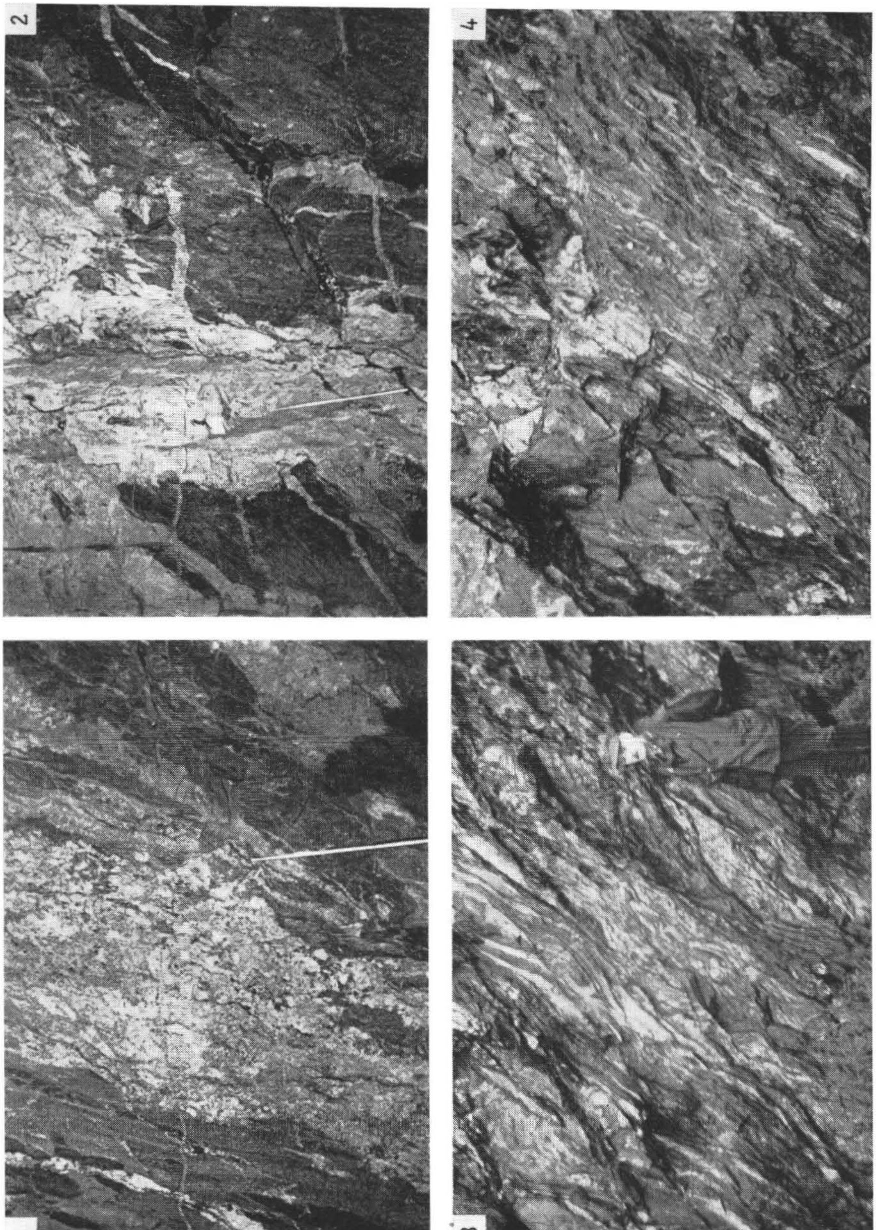

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