GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 212

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

XXXV

HELSINKI 1964

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GEOLOGICAL SURVEY OF FINLAND
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OBSERVATIONS ON GRANITIZATION AND ITS CHARACTERISTIC IN SOME ROCKS NEAR KEURUU [1])

BY

MAURIZIO FELICI

ABSTRACT

Series of transitions from granodioritic to granitic rock types have been investigated. Interesting features like relative and absolute abundance of minerals, twinning of plagioclase, variation in composition and sources of added material are discussed. Some views are presented on mineral components and their alterations.

CONTENTS

INTRODUCTION

On Vuorelanmäki, near Lihjamo in the Keuruu area of Central Finland, there is an outcrop in which favourable conditions have suggested a detailed investigation of contacts. This outcrop trends roughly NW-SE and its dimensions are about 400 × 50 m.

Surrounding the outcrop are different kinds of rocks: a porphyroblastic granite occurs W, N and S of the outcrop, while a gabbro intrusion,

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[1]) Received December 15, 1962.

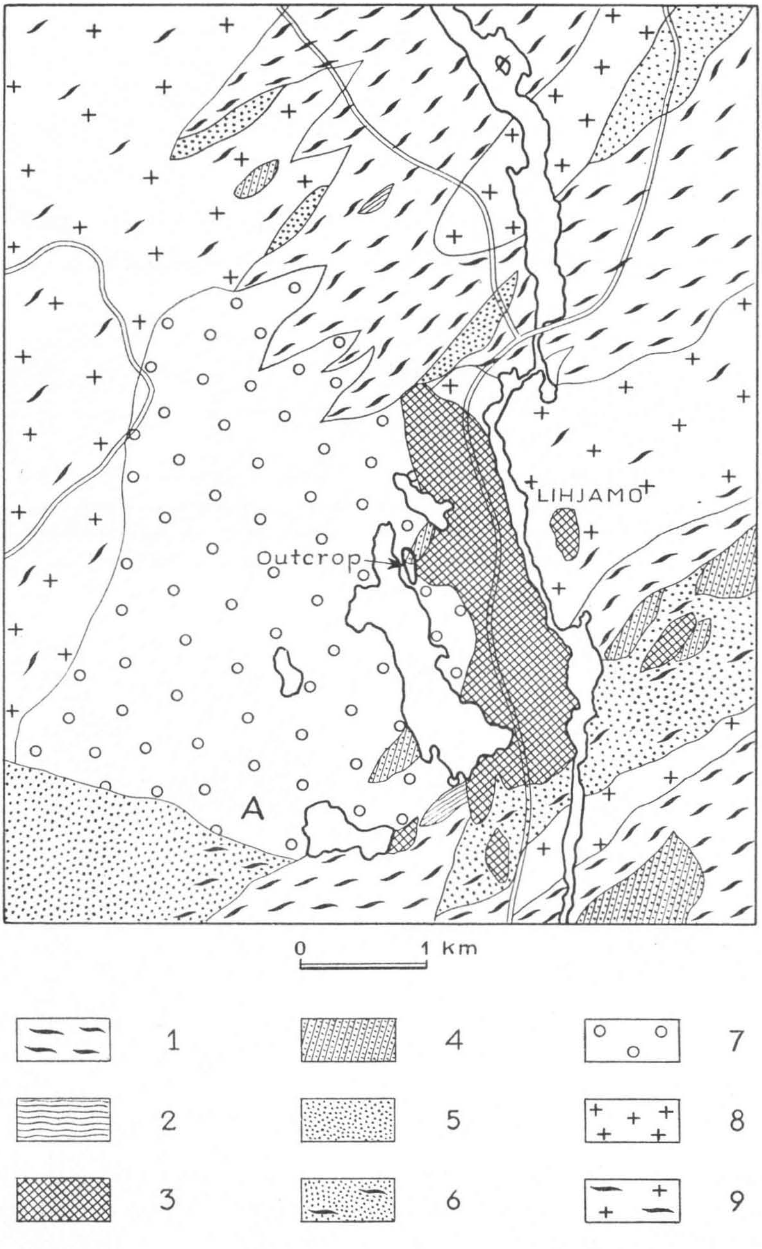

Fig. 1. Geological map of the area in which the outcrop is situated. 1 = granitic gneiss, 2 = amphibolites, 3 = gabbro, 4 = diorite, 5 = quartz diorite and granodiorite, 6 = gneissic quartz diorite and granodiorite, 7 = porphyroblastic granite, 8 = granite gneissic.

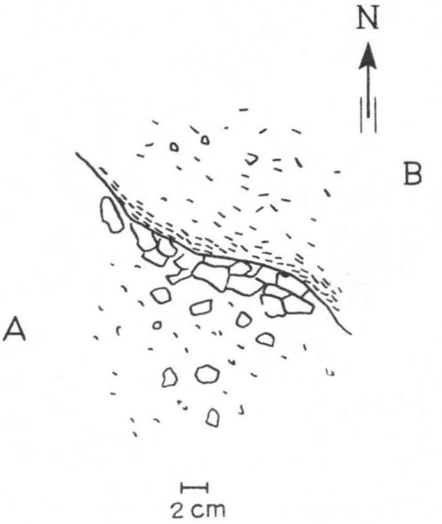

Fig. 2. Detail of a contact between porphyroblastic granite (A) and granodiorite (B). On the granitic side, potash feldspar porphyroblasts occur abundantly near the contact zone; on the other side, there is an enrichment in mafics.

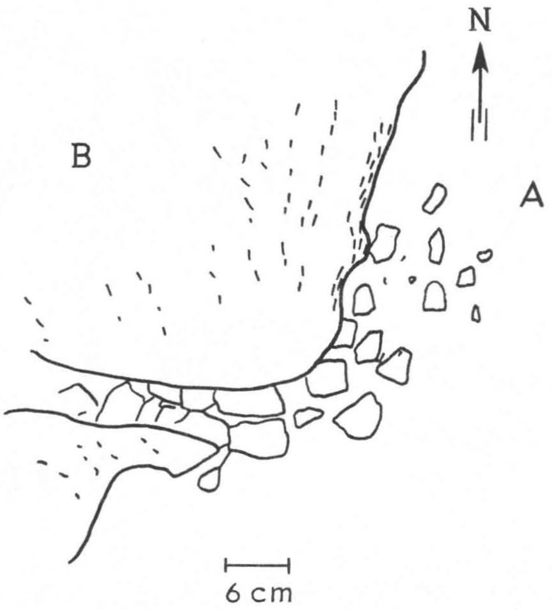

Fig. 3. Contact between porphyroblastic granite and quartz diorite (point A in Fig. 1).

3 000 × 500 m in size lies E of the outcrop. Granitic gneisses, quartz diorites (more or less gneissose) and gneissose granites are also present in the area (Fig. 1).

The rocks of the outcrop with which we are dealing may be classified as quartz diorite, granodiorite and porphyroblastic granite, the latter occurring in stripes and bodies of various size.

Transitions from such granite to granodiorite are often gradual and in such cases megascopic variations display mainly a decrease in the amount of the potash feldspar porphyroblasts, which in the surrounding granodioritic rock become sparser and sparser until they disappear. In other instances, the contact zone is crowded with potash feldspar porphyroblasts on the granitic side and marked by an enrichment in mafic minerals along the contact in the granodiorite (Fig. 2). Such a feature has also been found at a granite — quartz diorite contact (Fig. 3) about 2 km SW of the outcrop (point A, Fig. 1).

In the outcrop the granodiorite contains portions of quartz diorite rich in mafics and devoid of potash feldspar, other mineralogical features being similar to those of the enveloping granodioritic rock (plagioclase An_{45}, green amphibole $c \wedge \gamma = 24°$, pyroxene = augite, $c \wedge \gamma = 42°$). Around these quartz dioritic portions, longish crystals in the enveloping rock are subparallel to

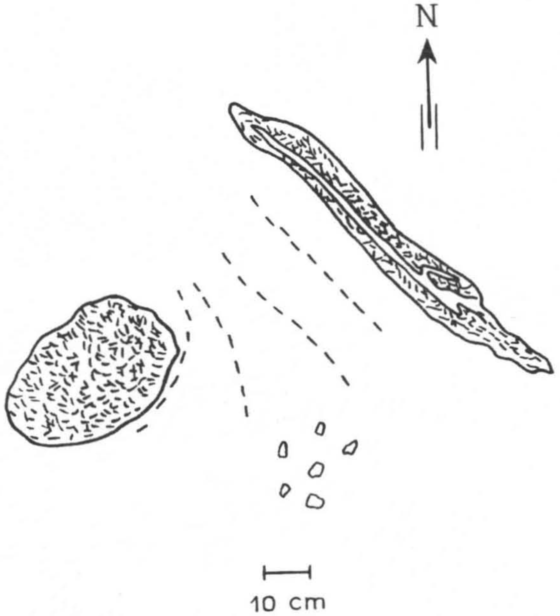

Fig. 4. Quartz dioritic inclusions in granodiorite and crystal arrangement around them.

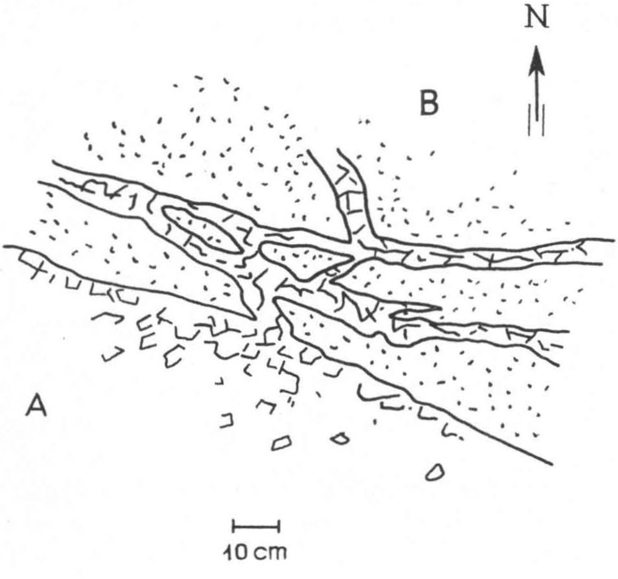

Fig. 5. Pegmatite-like net at the contact granodiorite (B)-porphyroblastic granite (A).

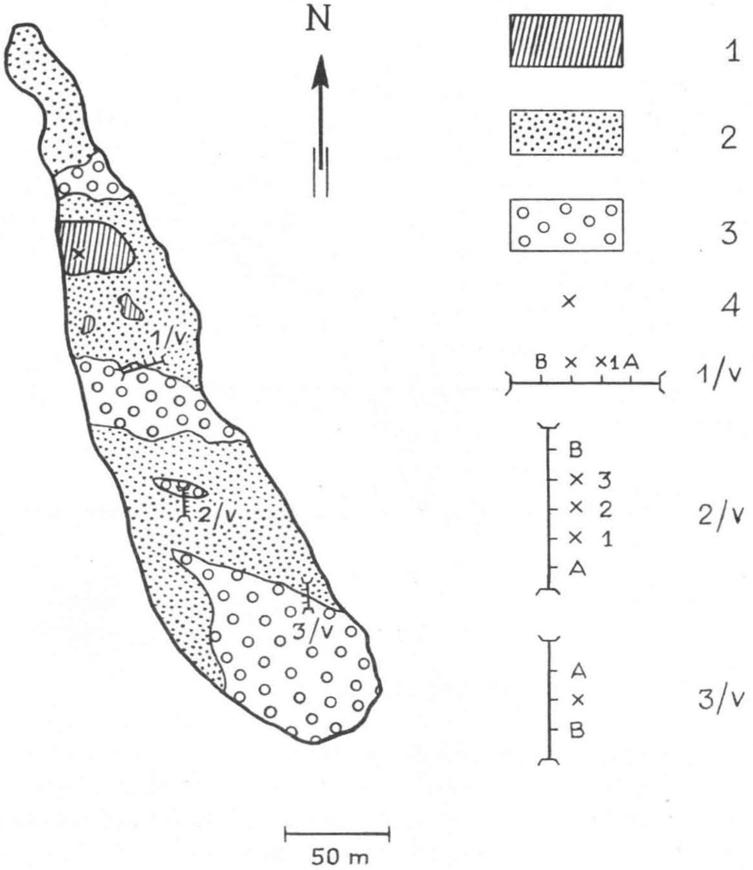

Fig. 6. The outcrop, distribution of rocks and traverses with sampling sites.
1 = quartz diorite, 2 = granodiorite, 3 = porphyroblastic granite.

the contact contours and in one of these portions a vein of salic material can be seen, without any continuation outside the inclusion (Fig. 4).

A pegmatitic net rich in potash feldspar is observed at some contacts between the porphyroblastic rock and the granodiorite (Fig. 5).

Thin sections have been made from samples taken along three traverses, each one three to four meters long, so that three series of sections, each one involving its own traverse, have been studied. The traverses pass across gradual granite — granodiorite contacts. One sample from a quartz dioritic inclusion was taken, too (Fig. 6). In relation to the granite body around the outcrop, traverses from number 1 to number 3 are always nearer to this granite, and gradation from granodioritic to granitic types is seen in the mineral composition of each one of the series.

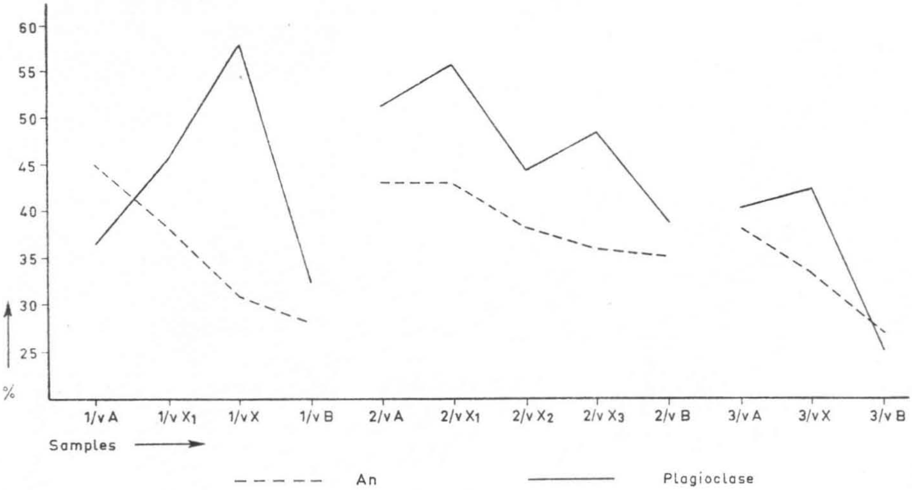

Fig. 7. Variation in the proportion and An content of the plagioclase for the three series of samples.

MINERALS

PLAGIOCLASE

In Fig. 7 the plagioclase and An contents are given for the series. Here it is worth noting the position of the maximum in the plagioclase percentage (intermediate members) and the regular decrease in the An content toward the acid members in each series. The maximum in the plagioclase percentage is not found in the corresponding members of the series, but it approaches to some extent the granodioritic side of the series, from series 1 to series 3. In other words, the maximum in the plagioclase percentage in series 1 is found in sample 1/VX, rather near the granitic side of the series, while in series 2 and 3 the plagioclase percentage maximum is found in samples nearer to the granodioritic side of the series (Table 1, samples 2/VXl, 3/VX). This could perhaps mean differences in the power of the granitizing agents in the different portions of the outcrop, another influencing and/or contrasting factor being the primary composition of the rock in such portions.

The plagioclase of the rocks examined occurs in two generations: the smaller, more idiomorphic, clear individuals are older, while the porphyroblasts are believed to be of younger age. Such porphyroblasts are nearly always zoned, in the main regularly from basic core to acid rims. Zoning can be very regular and multiple (as many as ten zones have been observed); the porphyroblasts also frequently show fractures and deformations of the twinning lamellae and are always more albitic than the older plagioclase.

Table 1. Mineral composition.

Sample	Series 1				Series 2					Series 3			Quartz diorite
	1/VA	1/VX1	1/VX	1/VB	2/VA	2/VX1	2/VX2	2/VX3	2/VB	3/VA	3/VX	3/VB	QD
Minerals in vol. %													
Plagioclase	36.6	45.5	57.9	32.2	51.1	54.3	44.3	48.3	38.7	40.3	42.3	25.0	43.9
Quartz	28.4	15.9	18.8	27.7	22.2	24.8	32.1	23.8	24.6	20.6	18.9	20.6	19.4
Microcline	17.7	21.7	11.7	29.0	13.6	1.8	3.7	6.3	16.6	26.1	25.6	35.8	—
Biotite	12.4	9.8	9.7	6.5	8.5	13.2	13.5	13.4	9.2	9.6	6.6	11.6	23.3
Pyroxene	2.5	1.4	—	—	2.8	0.9	0.7	1.5	—	0.1	0.8	—	5.2
Amphibole	1.7	5.1	0.9	3.8	1.1	3.4	5.4	6.1	8.0	2.8	5.1	6.0	6.0
Chlorite	0.8	—	—	—	0.4	1.2	—	—	—	—	—	—	—
Titanite	0.4	0.2	0.5	—	—	—	—	—	0.3	0.4	0.5	0.1	—
Apatite	0.1	0.2	0.2	0.2	0.1	—	0.2	0.3	0.6	0.2	0.1	0.5	1.0
Magnetite	—	—	0.4	0.7	0.2	0.1	0.2	0.5	2.2	—	—	0.9	1.0
	100.7	99.8	100.1	100.1	100.0	99.7	100.1	100.2	100.2	100.1	99.9	100.5	99.8
Size of slide (mm)	24 × 20	19 × 22	23 × 22	20 × 20	18 × 29	21 × 25	18 × 22	20 × 23	20 × 22	15 × 30	18 × 28	13 × 21	26 × 22
Points examined	2 826	2 277	2 823	2 235	3 132	3 176	2 542	2 726	2 425	2 513	2 751	1 499	2 655

The percentages were obtained using the integration stage.

Moreover, plagioclase porphyroblasts retain the same composition throughout the whole series (An about 30 %) and their abundance in series 1, 2, 3 has a variation diagram similar to the total plagioclase amount, *i.e.*, they are more abundant in intermediate members.

Another feature of the plagioclase porphyroblasts is a clear albitic rim against microcline, as well as myrmekitic structures occurring in some cases abundantly and in well-developed form.

Plagioclase porphyroblasts often include smaller plagioclase grains, which are often included by many other minerals.

In both plagioclase generations a variation is observed in the occurrence of certain twinning laws. While Albite twinning is common in every section of each series, Albite-Carlsbad twinning is more abundant in the granitic types, and Albite-Pericline twinning always becomes less and less frequent until it disappears in the acid members. This was observed in each series. As reported by Gorai (1950) and Turner (1951), the presence and relative abundance of A-twinning (Albite, Pericline, Acline) in plagioclase is indicative in some way of a metamorphic origin.

It may be supposed that the disappearance of the Albite-Pericline twinning is somehow related to certain magmatic-like conditions. The concomitant high abundance of Carlsbad twinning denotes, too, according to Köhler (1949), high crystallization temperatures.

MICROCLINE

It occurs mostly as large porphyroblasts, more frequent in acid members of the series; it shows perthitic structures, sometimes quite conspicuosly. Perthitic structures present are: stringlet, rod, string and patch perthite, the string type being the most frequent (Plates I and II, Figs. 1—3). The microcline is likely to include nearly all the other minerals and this probably means that the microcline (and likewise potash) developed last in the rock. The microcline shows a percentage minimum in the intermediate members, where plagioclase has a maximum.

It must be noted that a certain amount of the microcline is not porphyroblastic and this would suggest the possibility of the existence of a primary microcline, a view supported by the fact that this mineral can occur as inclusions in plagioclase porphyroblasts (which are believed to have preceded the introduction of potassium) and even in microcline porphyroblasts. The abundance of microcline (and quartz) is shown in Fig. 8. The shape of the diagram lines for the microcline roughly displays a concavity in all the three series, a notable feature considering that the percentage of microcline in the thin sections has been affected by a random and irregular

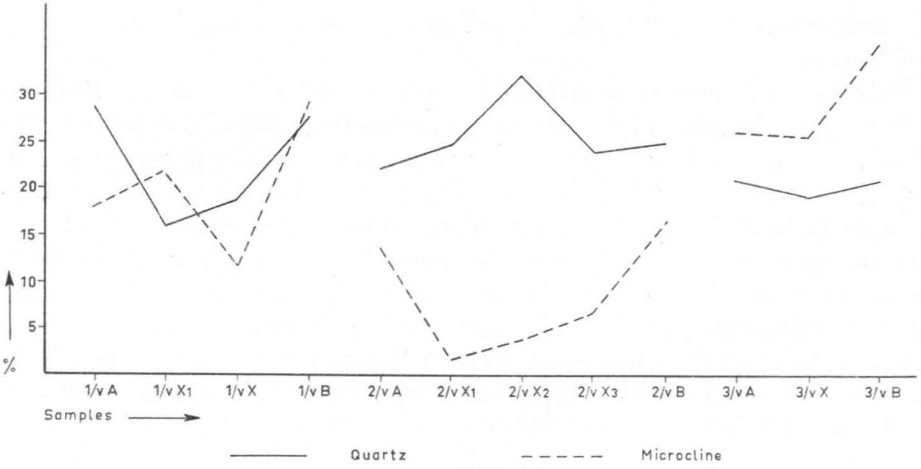

Fig. 8. Variation in the amount of quartz and microcline for the three series of samples.

distribution of the porphyroblasts in the rocks. The microcline sometimes shows Carlsbad twinning, but without preference for any members of the series.

QUARTZ

The variation in the abundance of this mineral does not display any particular feature, nor can it be correlated satisfactorily with the variations in the abundance of other minerals. The quartz shows an allotriomorphic shape, the recrystallized portion is mostly free from the many little inclusions (liquid or gaseous) that the quartz otherwise shows. Such inclusions are often uninterrupted in their alignment through many grains, which means that they have developed after the formation of such grains. The inclusions in the quartz sometimes form an alignment constant in the whole section. The quartz is often fractured and in some cases shows a considerable undulatory extinction. Drop quartz is found in the altered amphibole. The minerals contained in the quartz may be biotite, plagioclase and microcline, and this supports the possible existence, to be discussed later, of several generations of quartz.

PYROXENE

In each series it is absent in the acid member. In any case, even in basic ones it has nearly always been partially altered to amphibole, beginning with pleochroic rims (hornblende, Plate II, Fig. 4) probably owing to the migration of iron from the core (introduction of Fe from the outside and/or removal of lime could have played a part, too), and finally passing into amphibole (Plate III, Fig. 5), which will subsequently be affected by biotitization.

This kind of process is diffuse throughout the series; in addition, its continuity in the acid members (porphyroblastic granite) shows that this body is not the granitizing agent but more probably the primary rock in its more granitized aspect (Plate III, Fig. 6; Plate IV, Figs. 7 and 8). Iron ore is found associated with the pyroxene, probably owing to loss of iron from the pyroxene itself. As regards composition, the pyroxene probably belongs to a diopsidic augite type, not easily distinguished optically from diopside-hedenbergite series, with $c \wedge \gamma = 40°$ (average). This low angle may be due to a primary composition rich in Mg, but more easily could it be considered a consequence of the loss of iron as iron oxide and of the alteration to amphibole.

AMPHIBOLE

Variations in the abundance of amphibole are connected with its secondary derivation from pyroxene, showing maxima where this alteration has been more complete and concomitant with a higher primary abundance of amphibole. We must take into account biotitization of the amphibole, too, an alteration that occurs more often in acid members of the series. Such an alteration, when well developed, is apt to affect the amphibole content so that a decisive regularity cannot be expected in graphs. As regards composition, the amphibole belongs to a hornblende type, with pleochroism α = yellowish, β = green, γ = dark green. Among characteristic features are a sieve structure, presence of drop quartz, and some occurrence of twinning. The minerals included can be biotite, quartz and plagioclase. The extinction angle $c \wedge \gamma$ does not show appreciable variation; mean value = 22°.

BIOTITE

As already mentioned, the biotite is partly derived from amphibole. Such a process, as described by Härme (1958) requires the addition of potash, alumina and water, and it releases silica and (Mg, Fe) O. Abundant basal sections occur in some series, gradually disappearing in acid members, where biotite is richer in associated iron ore. Minerals like quartz, plagioclase and biotite itself may be present. Iron ore shows some enrichment at the granitic end of the series, where its amounts reach values as high as 2 %.

CONCLUSIONS

The portions of the quartz diorite rich in mafics and devoid of microcline that are observed in the outcrop and included in the granodioritic rock most probably represent relicts of a rock that has undergone potash metasomatism and from which actual granodiorite and porphyroblastic granite are derived.

This quartz diorite is by no means, however, the primary rock. The area we are dealing with is situated in the central part of the Svecofennidic Range, where old sediments and igneous rocks have been strongly metamorphosed, even up to the point of more or less complete palingenesis.

Pelitic sediments and volcanic rocks were the primary rocks in the area (more or less metamorphic relicts of them, such as the many amphibolitic rocks and tuffites, are observed here and there) so that they could have been the source of the added material necessary for explaining the widespread occurrence of microcline and alterations like biotitization of amphibole. During the stage of the regional metamorphism of this area, material was mobilized, perhaps with formation of a melt, in which differentiation processes could even have developed. I tentatively attribute to such processes the many small intrusions of gabbro and dioritic rocks of the zone, which are, judging from the manner of their occurrence, most probably younger than the other rocks of this area.

The gabbro body east of the outcrop is one of such intrusions, and its presence near the outcrop means that this place has been a zone of tension or at least of less resistance.

The material the provenance of which was the old sediments (containing K_2O, Al_2O_3, and SiO_2) has concentrated in an area of lower free energy, such as the area in which the outcrop is situated. The result is the granitization of a part of the rocks present in that area. The porphyroblastic granitic rock is believed to belong to such a portion of rocks.

Many minerals of the rocks examined show that the outcrop area has been affected by dynamic actions. The biotite is often deformed (Plate V, Fig. 9); twinning lamellae in the plagioclase are bent (Plate V, Fig. 10) and often the plagioclase is fractured, with fragments showing some dislocation, too; the quartz and microcline are often fractured, the former displaying a strong undulatory extinction.

Taking into account also the fact that some of the minerals occur as inclusions in each other, a picture can be drawn of the probable sequence in the granitization of the rocks examined: because of changed conditions pyroxene begins to change into amphibole and this further into biotite. These alteration products (amphibole and biotite) begin to include older minerals like quartz and primary biotite. Drop quartz in altered amphibole represents the release of silica, according to a reaction similar to the following (Härme 1958):

$$7\,(Mg, Fe)\,O\,.\,8\,SiO_2\,.\,H_2O+K_2O+Al_2O_3+H_2O =$$

$$K_2O\,.\,6\,(Mg, Fe)\,O\,.\,Al_2O_3\,.\,6\,SiO_2\,.\,2H_2O+(Mg, Fe)\,O+2Sio_2$$

An introduction of silica and soda follows: plagioclase porphyroblasts develop more albitic than the older plagioclase generation. Such porphyroblasts can retain or include some crystals of primary plagioclase and microcline. Free silica develops in quartz as grains of medium size which are likely to include biotite and plagioclase, both of primary origin. Subsequently potash was introduced, causing the formation of microcline porphyroblasts. The final stage of soda arrival is probably more or less contemporaneous with the first arrival of the potash, as may be supposed from the perthitic structures in the microcline, structures probably caused by an intergrowth of the two feldspars rather than by unmixing phenomena. There has been a partial replacement, probably depending on fluctuations of P, T and concentrations.

There have also occurred a sericitization of the plagioclase and a chloritization of the biotite, but the position they occupy in the sequence cannot be well defined.

Almost in each mineral species there are several generations. The microcline is primary and porphyroblastic, like the plagioclase as well.

The biotite and amphibole are primary and they are alteration products. Quartz occurs as a primary component, as drop quartz in altered amphibole and as a product of recrystallization of introduced silica. The fluids carrying added materials (Na_2O, K_2O, SiO_2, Al_2O_3) did not have, during the last stage of the process, too strong a penetrating action. This can be inferred from features at some contacts where potash feldspar porphyroblasts accumulate, as if solutions could not have passed through.

CHEMICAL COMPOSITION

Table 2 gives the chemical composition of the granodioritic (2/VA) and granitic rock (2/VB). In the same table are given the molecular norms, atomic percentage and cation number per standard cell (Barth's method) for the same two rocks.

From here it appears that Na, K, Si and Fe^{3+} have been added to and Mg, Ca, and Al removed from the granodioritic rock.

Table 2. Chemical composition.

Sample	Weight %		Cation %			Molecular norm		
SiO_2	64.55	65.47	Si	61.36	62.55	Q	19.38	20.35
TiO_2	0.64	0.88	Ti	0.46	0.64	Or	21.66	29.25
Al_2O_3	15.59	14.26	Al	17.41	16.02	Ab	25.66	28.10
Fe_2O_3	0.85	2.61	Fe^{3+}	0.58	1.84	An	19.86	11.37
FeO	4.00	3.89	Fe^{2+}	3.13	3.10			
MnO	0.09	0.11	Mn	0.06	0.06	Wo	0.37	0.72
MgO	2.21	0.96	Mg	3.08	1.32	En	6.17	2.64
CaO	4.28	2.78	Ca	4.33	2.81	Fs	4.81	3.12
Na_2O	2.80	3.05	Na	5.14	5.63	Ap	0.31	0.45
K_2O	3.60	4.85	K	4.33	5.85	Il	0.92	1.26
P_2O_5	0.17	0.22						
CO_2	0.08	—						
H_2O+	0.60	0.43	P	0.12	0.18	Mt	0.86	2.74
H_2O-	0.04	0.05						
	99.50	99.56		100.00	100.00		100.00	100.00
Cations per Standard cell (Barth's method)				96.52	96.05			
Modal feldspar							64.70	55.30
Normative feldspar							67.18	68.72

When the amounts of modal and normative feldspar are compared, differences appear. This also suggests that material has been added, even if in the granitic rock the difference seems rather large in being due only to this cause.

The rocks examined in this paper belong to the synkinematic group of the Svecofennidic Range. In this connection it is interesting to examine the numbers of cations per standard cell.

In the rocks investigated this value is on the average 96.28, well in accord with the average 96.82 calculated for the group of Sveconfennidic synkinematic rocks (diorites+granodiorites+granites) using the average chemical composition given by Simonen (1948).

A comparison can also be made with the results obtained by Marmo (1955) with Precambrian granites in West Africa. In these granites with a petrochemistry similar to that of Svecofennidic synkinematic granites, the number of cations per standard cell is greater than 95.0. It must be noted finally that synkinematic silicic rocks are rich in soda, whereas the rocks examined in this paper, especially the most silicic ones, have a notably high content of potash. As pointed out by Härme (1958) and others, such synkinematic rocks must have acquired their high potash content secondarily from the outside.

This is another fact that shows how, in the granitization phenomena examined in this paper, an addition of material must definitely have taken place.

SUMMARY

In the rocks examined, which have been affected more or less by granitization, many interesting features are observed. Plagioclase occurs in two generations, the porphyroblastic being the younger; its abundance shows maxima in the intermediate members of the series (zone of gradual transition), whereas the microcline shows minima. There are variations in the relative and absolute abundance of some twinning laws in the plagioclase, notably in the Albite-Carlsbad and Albite-Pericline ones, the former being more abundant in granitic types where the latter disappears. This seems somehow to be associated with magmatic-like conditions. Continuous alteration processes (pyroxene-amphibole-biotite) lead to the assumption that the granitic portion of the rocks should be regarded as the result rather than the agent of the granitization. In this instance material has been added, as can be deduced from the considerations and calculations regarding the alteration processes, the occurrence and abundance of microcline porphyroblasts, the amount of normal and normative feldspar, the number of cation per standard cell and data from the Svecofennidic silicic synkinematic rocks. Added material is believed to have concentrated in an area of lower free energy like that in which the outcrop is situated and its provenance is supposed to be metamorphosed old sediments and volcanic rocks (actually granitic gneisses and amphibolites) probably similar to the rocks of these groups near the outcrop. Elements added are Na, K, Si and Fe^{3+}, while Mg, Ca and Al have been removed.

ACKNOWLEDGMENTS

The author wishes heartly to thank Dr. V. Marmo, Director of the Geological Survey of Finland, for his most valuable criticism and counsel. Thanks are due to Mr. Paul Sjöblom who corrected the English of the manuscript.

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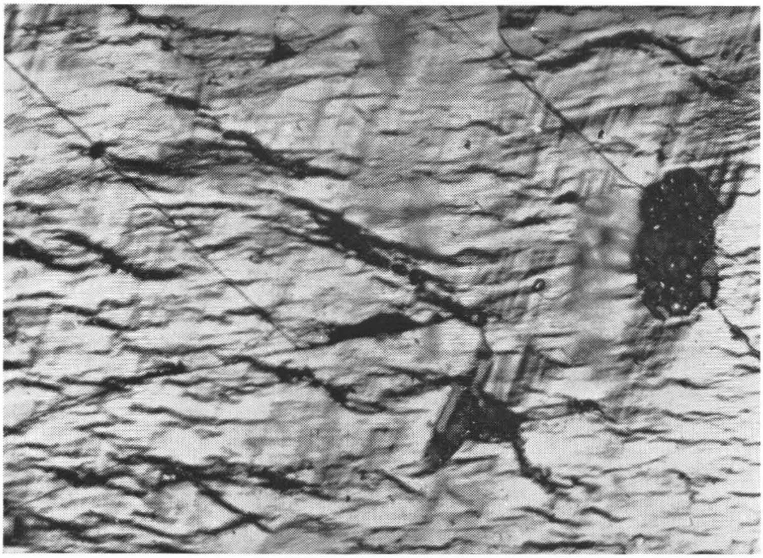

Fig. 1. Perthitic structures in microcline. String perthite and (upper right) stringlet perthite. Section 2/VB, nicols +, 51 ×.

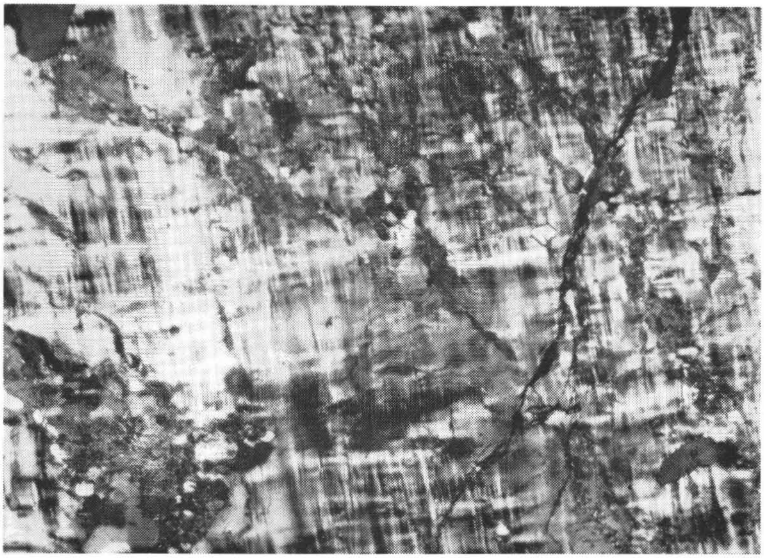

Fig. 2. String perthite in microcline (cross-hatched), passing to patch type Section 3/VB, nicols +, 51 ×.

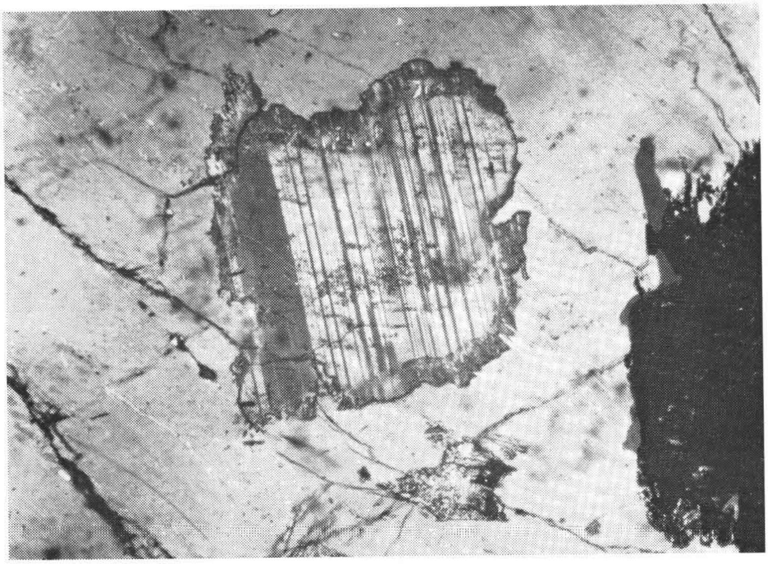

Fig. 3. Stringlet, string and rod perthite in microcline (light gray). Plagioclase is included in microcline and shows myrmekitic structure against it. A perthite string is continuous across the boundary plagioclase-microcline (right). Section 1/VA, nicols +, 51 ×.

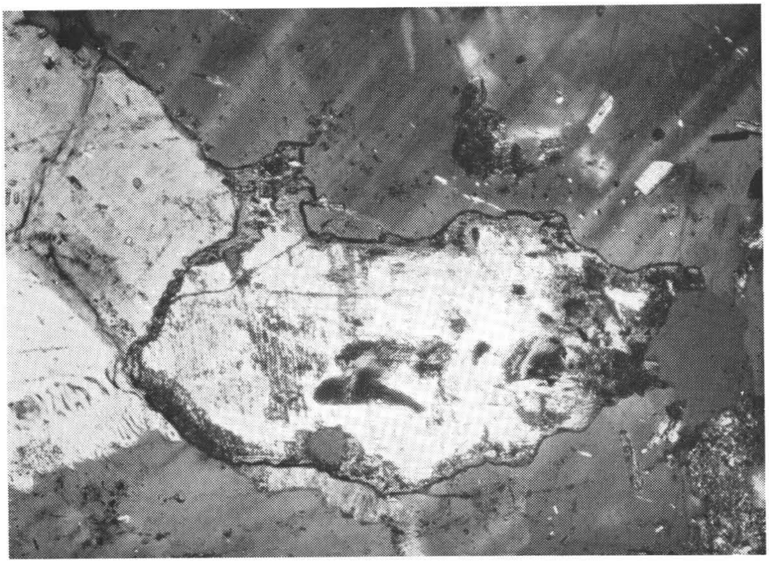

Fig. 4. Pyroxene with amphibole rim (darker). Also visible is a myrmekitic structure in the plagioclase (lower left, light gray) against microcline (gray). Section 1/VA, nicols +, 130 ×.

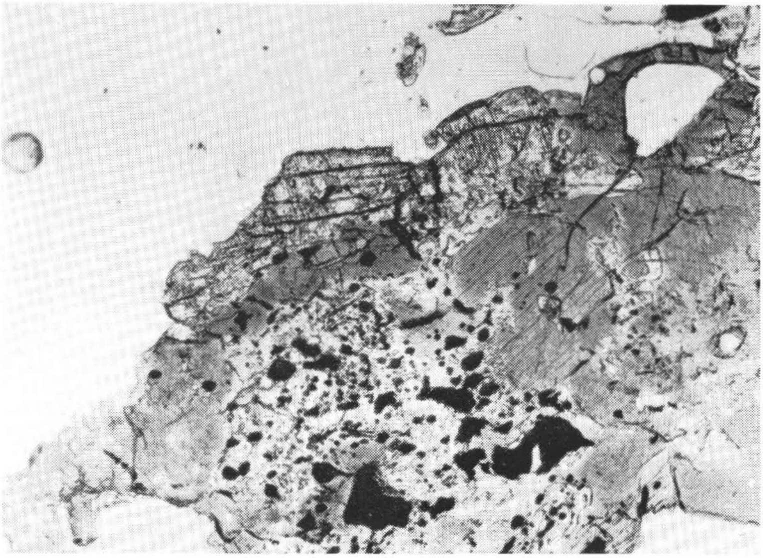

Fig. 5. Pyroxene (light gray) partially altered to amphibole (gray). Iron oxides (dark) are due to loss of iron from the pyroxene. Section 1/VA, one Nic., 130 ×.

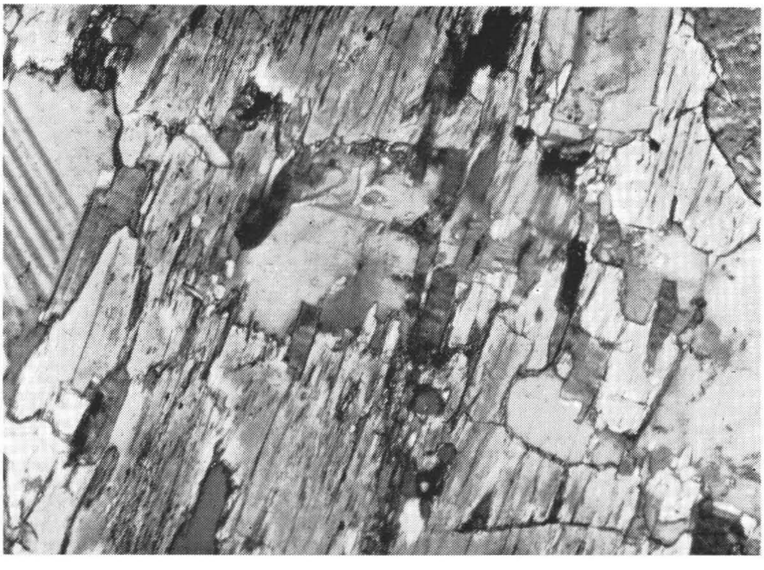

Fig. 6. Amphibole (light gray) with small darker flakes of biotite. Section 2/VB, nicols +, 130 ×.

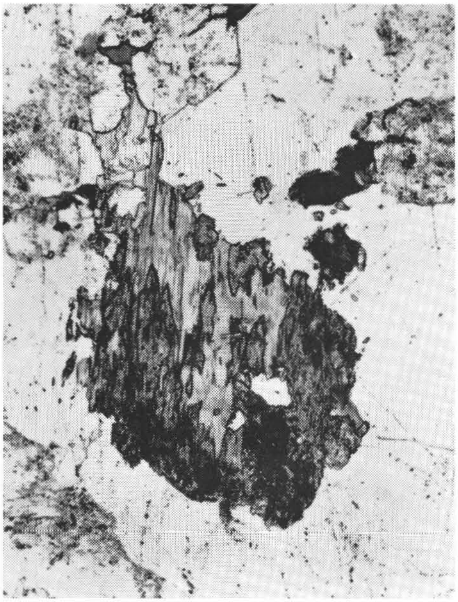

Fig. 7. Biotite (light gray) replacing amphi bole (dark gray). Section 2/VXI, one Nic., 68 ×.

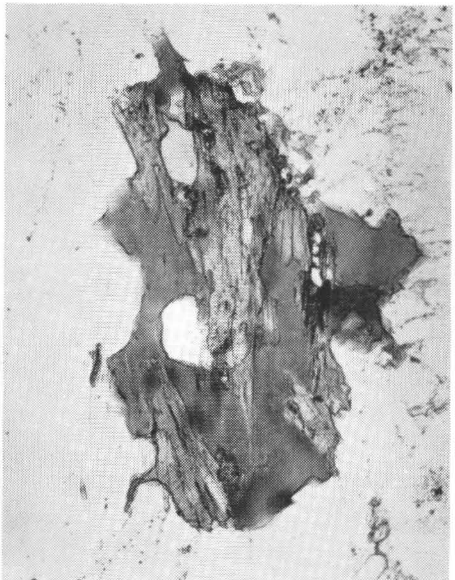

Fig. 8. Biotitization of amphibole (lighter). Section 2/VXI, one Nic., 130 ×.

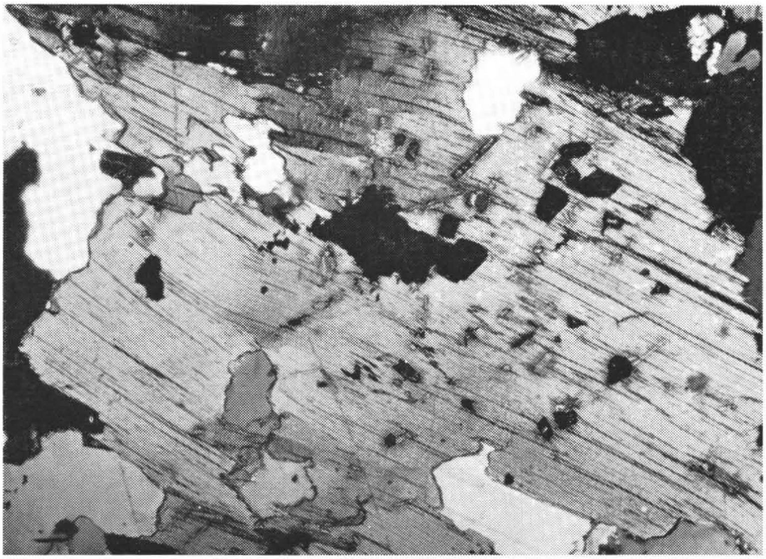

Fig. 9. Large crystal of biotite deformed by dynamic action. Section 3/VB, nicols +, 51 ×.

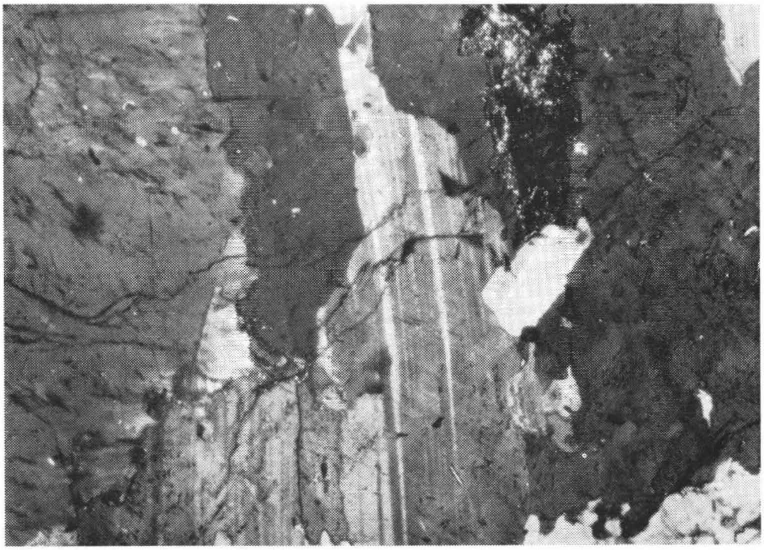

Fig. 10. Plagioclase showing bent twinning lamellae. Section 3/VA, nicols +, 51 ×.

2

ON DIABASE DIKES AT KEURUU, CENTRAL FINLAND [1]

BY

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CONTENTS

INTRODUCTION

Fine-grained diabase cutting various rocks as narrow dikes has been observed in several localities in Finland. Such dikes of diabase have attracted the attention of Finnish geologists for more than a century. Kuhlberg (1869, p. 42) analyzed the diabase of Parainen, and Wiik (1869) studied the fine-grained dikes in the vicinity of Helsinki. In his inaugural dissertation Wiik (1876) also mentioned diabase dikes near the town of Kristiina. He distinguished this fine-grained variety from the olivine-bearing diabase of Satakunta and associated the fine-grained diabases contained in the gneiss granite in the Helsinki area with those in the limestone at Parainen and in the porphyritic granodiorite at Kristiina (Wiik 1876, p. 34). Further, he described pseudomorphs shaped like augite but filled with a mass of dark minerals and common in fine-grained diabase.

After Wiik, similar dikes of fine-grained diabase have been described in detail by several Finnish geologists, among others Frosterus (1893) and Wilkman (1924).

In regard to the »youth» of this kind of rock (perhaps of post-Jotnian age) all the authors mentioned have been unanimous.

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[1] Received May 20, 1963.

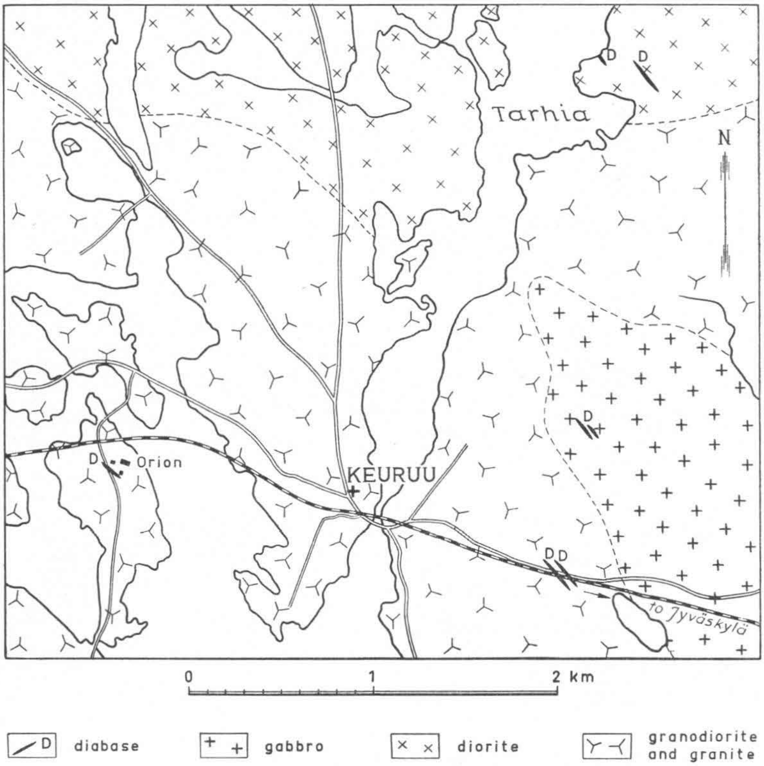

Fig. 1. Geological map of the area around Keuruu, Central Finland.

The present paper deals with the diabase dikes of Keuruu in Central Finland. The petrographic study was made by Marmo and the mineralogical determinations by Mikkola.

PETROGRAPHIC DESCRIPTION

In the surroundings of the church of Keuruu, some 60 km W of the town of Jyväskylä in Central Finland, a porphyritic granodiorite is the most common kind of rock (Fig. 1). East of the church there is an area occupied by gabbro, while diorite occurs to the north (see map of Fig. 1). The petrography of these rocks has been described by Marmo (1963) in the explanatory notes to the geological map sheet of Keuruu.

At several places within the area (illustrated in Fig. 1) there occur narrow diabase dikes. The diabase is fine-grained, often aphanitic, and it has a distinct ophitic texture. The orientation of these dikes remains in all cases nearly the same: strike N 45° W, dip vertical.

Fig. 2. Diabase dike at the railway cutting W of Keuruu. The length of the arrow is 10 cm. The country rock is porphyritic granodiorite. Photo E. Halme.

The diabase is dark grey, often black, and its contacts with penetrated rocks are usually quite sharp. SW of the Orion factories, however, the porphyritic granodiorite is brecciated by the penetrating diabase.

The dikes near Keuruu are usually only a few cm in width. The widest one (3 m in width) has been found on the eastern shore of Lake Tarhianpohja. This dike can be followed some hundred meters, not, however, as a dike but as a channel, owing to the complete wearing off of the diabasic material in the course of time. This erosion took place easily because the said diabase has excellent cleavages in several directions, mainly semiperpendicular to the strike of the dike. At the railway cutting about 1 km E of the church of Keuruu, a determination was made of the cleavage of the diabase (Fig. 2). There are two parallel dikes, one 2 m, the other 90—100 cm in width. The best plane of cleavage had developed vertically to the strike of the dike dipping 30° at N 45° W. Another poorer cleavage plane dips 60° at S 45° E. These directions of cleavage may not be connected with tectonic movements, but are rather to be considered directions of natural cleavage of the rock.

The narrow dikes are homogeneous but when more than 1/2 m in width their margin is finer of grain than the center.

Megascopically the rocks of all the diabase dikes at Keuruu are very similar, but microscopically they display several differences.

Some 2 km WNW of Orion there are two diabase dikes, both about 60 cm in width. The diabase is aphanitic with only a few, strongly resorbed remnants of ophitic plagioclase. Therefore a more accurate determination of the plagioclase was difficult. Zonal structure was observed in one grain, where two individuals were twinned according to the Albite law, one of them zoned with the center showing a lower (ca. 25 % An) anorthite content than the border and the other individual (ca. 60 % An). The groundmass consists of plagioclase, clinopyroxene, amphibole, chlorite, epidote, and biotite, the three firstmentioned minerals, however, in much altered condition. Fresh grains of these minerals appear occasionally. Also small grains of apatite and pyrite have been met with. Glass is not present. Biotite occurs only as a few minute flakes. The dikes penetrate diorite there.

At Lapinsalmi (1 km E of Keuruu) the diabase (in the marginal part of the dike) is much fresher than that described above. Plagioclase (andesine) occurs as numerous fine ophitic laths, plentifully twinned. Only a few of the laths are saussuritized or chloritized and are then cut by narrow veins of pyrite. The pyroxene is often completely chloritized. In some grains of chloritized matter, the plagioclase is still preserved as small remnants, indicating that also a strong chloritization of plagioclase has taken place, particularly in the plagioclase of the extremely fine-grained groundmass.

A thin section made of a sample taken from the central part of the dike described displays a coarse-grained groundmass which consists of plagioclase, uralite and chlorite. Pyroxene does not appear but the shape of some altered grains is that of augite.

The most interesting diabase occurs on the eastern shore of Lake Tarhianpohja. There the phenocrysts consist of ophitic plagioclase (labradorite) together with orthorhombic pyroxene (Fig. 3). The same minerals, together with uralite, chlorite, apatite and small amounts of pyrite, form the groundmass. Biotite occurs also but only as small sporadical flakes. The occurrence of the orthorhombic pyroxene is striking because the diabases described earlier from Finland are not known to contain it, their pyroxene usually being enstatite-augite. Here the orthorhombic pyroxene is pale in colour and rarely pleochroic. In spite of these facts the axial angle 2 Vα is 67°, the optical character negative and nα = 1.695 and nγ = 1.710 (±0.002). The values indicate an iron content of some 30 %. Consequently the mineral is hypersthene.

Fig. 3. Thin section of Tarhianpohja diabase where orthorhombic pyroxene occurs.

ON THE TWIN LAWS OF PLAGIOCLASES

The sections of this kind of altered, fine-grained rock do not offer much possibilites for accurate optical determinations. Yet some attempts were made. The plagioclases at the railway cutting showed consistently the same twin laws, or Albite, Carlsbad, Albite-Carlsbad, and in some cases Pericline lamellae. The anorthite content varied from 54 to 60 %.

One km N of the railway cutting the groundmass plagioclases of the gabbro showed the same twin laws, Albite, Carlsbad and Albite-Carlsbad. The only exception observed was made by a large, clear grain, which here

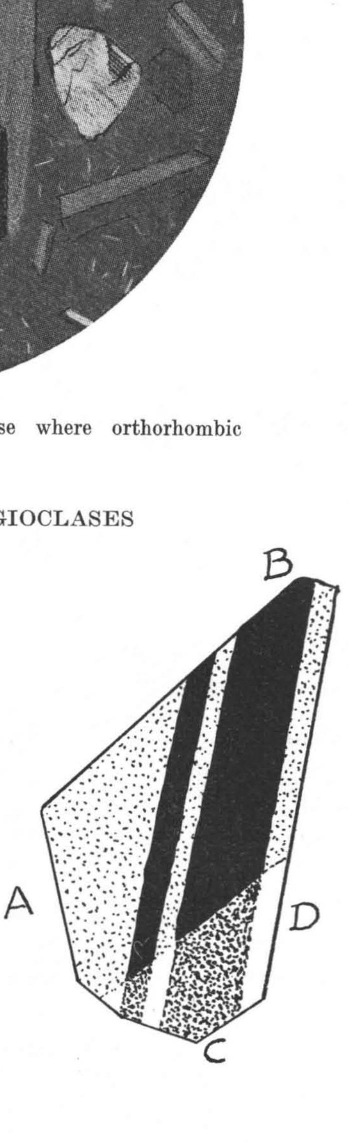

Fig. 4. A plagioclase phenocryst with twin laws Albite, Ala A, and Albite-Ala A.

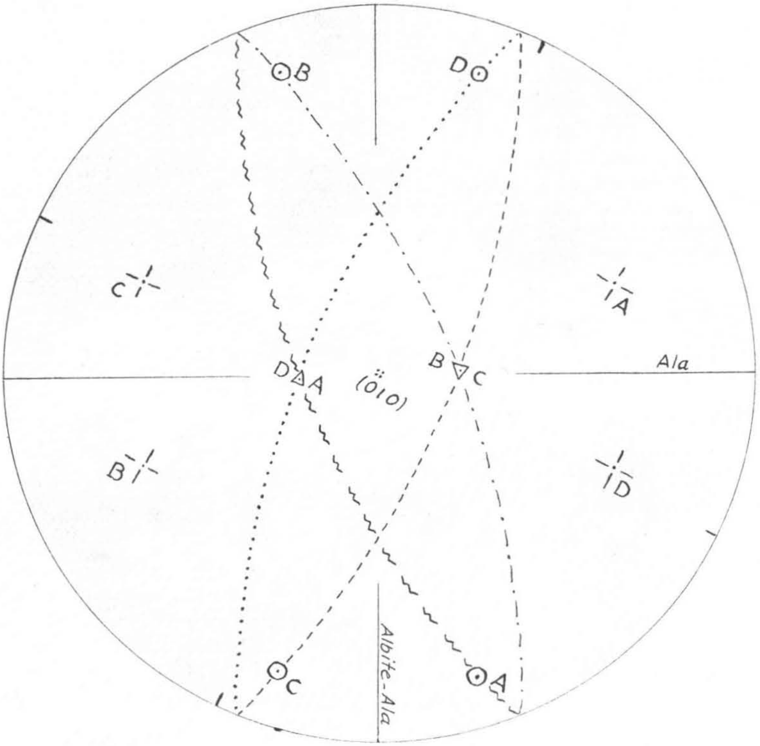

Fig. 5. Twinned plagioclase crystals A, B, C and D on Reinhard Table 3.

deserves a more detailed description. Four individuals (Fig. 4) were twinned according to the following laws:

A B Albite
B C Ala A
C D Albite

Consequently the following relations would be valid:

A D Ala A
A C Albite-Ala A
B D Albite-Ala A

The anorthite content was lower, or 45 %. At this percentage the diagrams on Reinhard Table 2 give symmetrically nearly the same positions for the faces (010) and (021) as well as for (001) and $(0\bar{2}1)$. The first impression was that of the Baveno law, which we have not yet met in Finnish rocks. The angle $\gamma_1 \wedge \gamma_2$ was nearly 0°. The only possible law was consequently Ala; and, the composition face being (100), it was Ala A, which van der

Table 1.

	1. Diabase, Keuruu 2 km WNW from Orion. Anal. M. Tavela	2. Gabbro, Keuruu, Lapinsalmi, Anal. M. Tavela	3. Enstatite-augite diabase, Kalliomäki, Sonkajärvi. Anal. G. A, Aartovaara (Wilkman, 1924, p. 10).	
SiO_2	48.20	49.17		49.87
TiO_2	1.42	1.16		0.94
Al_2O_5	18.04	15.58		11.46
Fe_2O_5	2.07	2.21		3.27
FeO	8.08	7.90		10.04
MnO	0.19	0.20		0.74
MgO	6.56	9.36		6.65
CaO	8.34	8.68		12.68
Na_2O	2.33	2.35		1.18
K_2O	1.05	0.09		0.75
P_2O_5	0.71	0.58		0.13
Cr_2O_5	tr.	0.07		not. determ.
CuO	not. determ.	not. determ.		0.14
ZnO	not. determ.	not. determ.		0.16
S	0.28	tr.		0.19
H_2O+	2.28	1.18		0.20
H_2O-	0.16	0.14		0.48
	99.73	99.67		98.88
Normative minerals:				
Quartz	0.49	0.00		4.58
Orthoclase	6.12	6.67		4.44
Albite	19.39	19.91		10.01
Anorthite	35.70	28.58		23.75
$CaO \cdot SiO_2$	0.52	4.52	$FeCa(SiO_5)2$	14.15
$MgO \cdot SiO_2$	16.40	23.40	$MgCa(SiO_5)2$	17.62
$FeO \cdot SiO_2$	10.69	5.45	$FeSiO_5$	7.77
			$MgSiO_5$	8.43
Magnetite	3.02	3.25		4.74
Ilmenite	2.74	2.13		1.78
Chromite	0.00	0.15		
Apatite	1.68	1.34		0.29
FeO	0.00	2.92	Chalcopyrite	0.32
MnO	0.19	0.20	Sphalerite	0.19
S	0.28	0.00	Pyrrhotite	0.04
H_2O	2.28	1.18		0.68
	99.50	99.70		98.97

al	26	si	112.56	al	21	si	118.56	al	15.5	si	114
fm	45	k	0.24	fm	51	k	0.23	fm	50.0	k	0.30
c	22	mg	0.624	c	21	mg	0.534	c	31.0	mg	0.46
alk	7			alk	7			alk	3.5		
	100				100				100		

Some determinations of the specific gravity of the diabases of Keuruu:

1.	diabase, Keuruu, 2 km WNW from Orion	2.919	(±0,005)
2.	» » S from Orion	2.836	»
3.	» » 1 km E from Lapinsalmi, railway cutting	2.920	»
4.	» » 1 km N from railway cutting	2.854	»

Kaaden (1951) mentions as being rather rare. He has established it »in volcanic rocks as spilitic rocks, trachyandesites, dacites and pyroxene andesites, in plutonic rocks as granite and eucrite».

Fig. 5 shows the relations on Reinhard Table 3.

The thin sections from Tarhianpohja presented again Albite, Carlsbad and Albite-Carlsbad laws rather frequently mixed with Pericline lamellae. Another Ala A case was noticed here among the middle-sized laths. The anorthite content varied from 47 to 60 %.

ABOUT THE CHEMISTRY OF THE KEURUU DIABASE

The diabase of the dike 2 km WNW of Orion was analyzed by Mr. M. Tavela, Mag. Phil., at the Chemical Laboratory of the Geological Survey of Finland. In Table 1 this analysis (1) is given in addition to the chemical analysis (3) of the enstatite-augite diabase from Kalliomäki, Sonkajärvi (Wilkman 1924, p. 10). Anal. 2 in the same table represents the gabbro E of the church of Keuruu. This rock is coarse-grained, and owing to its ophitic texture, the name dolerite would be more appropriate than gabbro. Its main minerals are plagioclase, augite, hornblende and biotite, accessories being apatite, titanite, ilmenite and magnetite. The chemistry of fine-grained diabase is very similar to that of this »dolerite-like gabbro».

At the end of Table 1 some determinations of the specific gravity of the Keuruu diabases are given.

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3

SOME MANGANESE MINERALS FROM INDIA [1]

BY

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ABSTRACT

The chemical composition, unit cell dimensions, powder data and optical properties are given for a manganese-bearing aegirine, two amphiboles and a piemontite from two Indian localities. The values measured are briefly discussed.

CONTENTS

PREFACE

India is known as one of the richest and largest manganese provinces in the world. Many manganese minerals have been originally discovered in and described from various Indian localities. The memoirs of L. L. Fermor (1906,

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[1] Received May 25, 1963.

1909) are classical in this respect. The four minerals described in this paper originate from a collection of rock samples from Indian manganese deposits which one of us (V. K. N.) sent to the mineralogical museum of the newly established Institute of Geology of the University of Turku, Finland. The authors, as far apart geographically as they are, have a common interest in manganese silicates and the minerals now under consideration were so beautiful and showed such a remarkble pleochroism that they simply not could be left unstudied.

NORTH TIRODI MINE, BALAGHAT DISTRICT, MADHYA PRADESH

The Tirodi Mine is the type locality for the alkali amphibole tirodite. The specimen now studied was collected by Mr. D. K. Chatterjee, geologist of the C. P. M. O. Co., Ltd. It comes from pit No. 34 of the mine and it occurred in association with a pegmatite body in contact with the ore zone.

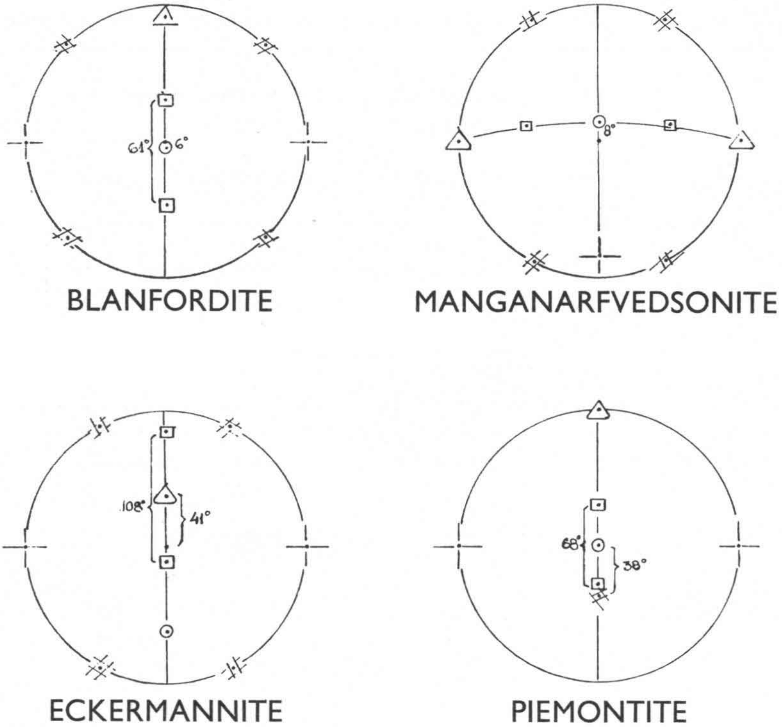
BLANFORDITE

The mineral blanfordite, a manganiferous pyroxene from the Kacharwahi manganese mine in the Nagpur district, was originally discovered and named by Fermor (1906) after W. T. Blandford. Subsequently, similar pyroxenes were reported by him (1909) from four different Indian localities. Although Fermor did not know whether all these minerals were the same as the Kacharwahi pyroxene, he nevertheless included them tentatively under the term blandfordite. Krishnan (1937) reported a blanfordite type of pyroxene from Gangpur State, India. In recent years reports of the mineral have come from the Ponia and Tirodi manganese mines in the Balaghat district (Kilpady and Dave 1958), from Chikla in the Bhandara district (Bilgrami 1956) and from the Nagpur district (Zwaan and van der Plas 1958).

In the North Tirodi mine, blanfordite occurs as anhedral, 1-mm crystals in a granular banded rock together with microcline, manganarfvedsonite (juddite), mangan-chlorite and braunite.

The mineral was purified using Clerici's solution and analysed. The chemical analysis was performed by Mr. Paavo Väänänen, M. Sc., of the Geological Survey of Finland. The composition of the mineral is given in Table 1. The table also lists the unit cell dimensions and the cell content of the mineral. The content of the unit cell is calculated on the basis of the chemical analysis, the cell dimensions given and the specific gravity measured.

The optical properties and the orientation of the mineral are given in Table 2 and in Fig. 1.



BLANFORDITE MANGANARFVEDSONITE

ECKERMANNITE PIEMONTITE

Fig. 1. Optical orientation of blandfordite, manganarfvedsonite, eckermannite and piemontite.

Table 1. The chemical composition and the unit cell of blanfordite from Goldongri, India. Chemical analysis by Paavo Väänänen, M. Sc., of the Geological Survey of Finland.

	wt %	cation %	Unit cell content			
SiO_2	52.86	50.3	Si	7.97 atoms		
Al_2O_5	1.43	1.6	Al	0.26	3.61	The unit cell:
Fe_2O_5	27.61	19.7	Fe^{3+}	3.13		a = 9.64 Å
FeO	—					b = 8.78 »
Mn_2O_5	1.91	1.4	Mn^{3+}	0.22		c = 5.28 »
MnO	0.09	0.1	Mn^{2+}	0.01	0.44	β = 106° 57′
MgO	1.47	2.1	Mg	0.34		Vo = 427.5 $Å^3$
TiO_2	0.75	0.5	Ti	0.09		z = 4
CaO	1.57	1.6	Ca	0.25	3.80	Space group C 2/c
Na_2O	11.94	21.9	Na	3.52		D = 3.516
K_2O	0.15	0.2	K	0.03		Aegirine 80 mol. %
H_2O + ...	0.07	0.4	OH	0.09		Jadeite 10 » »
H_2O	0.02		O	23.80		Johansenite 6 » »
CO_2	n. d.					Enstatite 4 » »
PbO	n. d.					
CuO	n. d.					
P_2O_5	0.04	0.1				
	99.91	100.0				

Table 2. The optical properties of some blanfordites from India.

	Kacharwahi, Fermor (1909)	Chikla, Bilgrami (1956)	Nagpur, Zwaan and van der Plas (1958)	Ponia, Kilpady (1960)	Tirodi, Kilpady (1960)	Tirodi mine, (this study)
α	1.752 rose pink	1.732 rose pink	1.767 rose, pink to violet	1.760 pink with a tinge of brown	1.746 yellow pinkish	1.748 rose pink
β	1.754 bluish lilac	1.756 pale blue	— light bluish lilac	1.768 pale pink	1.748 brown	1.773 violet
γ	1.780 sky blue	1.770 blue	1.799 sky blue	1.780 blue	1.768	1.788 blue
2Vα ..		80°	72°—68°			61°
c : α ..		9°	5°			6°

Table 3. X-ray powder diffraction data for some mangan-bearing minerals from Goldongri, India. Cu-K α-radiation, 114.7 mm diameter camera.

Blanfordite			Manganarfvedsonite			Eckermannite			Piemontite		
hkl	d_{meas}	I	hkl	d_{meas}	I	hkl	d_{meas}	I	hkl	d_{meas}	I
111	6.37	S	020	9.05	W	020	9.05	W	100	8.05	M
020	4.43	MS	110	8.43	S	110	8.43	M	101	5.07	MS
	3.59	VW	040	4.46	MS	200	4.88	W	002	4.60	VW
021	3.34	W	021 220	4.19	W	040	4.46	MS	200	4.00	MS
221	2.989	VS				220	4.13	W	211	3.49	MS
310 311	2.895	S	131	3.38	S	131	3.82	W	102	3.43	W
			240	3.24	W	131	3.38	S	112	2.912	VS
002 202	2.595	MS	310	3.10	S	240	3.27	MS	020	2.840	M
			221	2.97	W	310	3.12	S	013	2.682	S
221	2.462	MS	330	2.78	M	221	2.941	W	211	2.600	S
311	2.254	W	151	2.691	VS	330	2.797	M	202	2.536	M
222	2.196	M	061	2.563	W	151	2.699	VS	121	2.475	W
330	2.123	M	002	2.523	M	061	2.571	W	310	2.417	S
421	2.090	W	350 351	2.321	W	002	2.515	M	222 212	2.309	M
041	2.030	W				351	2.316	W			
402	2.019	W	171	2.259	W	171	2.265	W	122	2.176	W
240	1.981	W	261	2.156	M	171	2.196	VW	403	2.156	W
241	1.930	W		2.063	VW	261	2.156	M	014+	2.123	M
511	1.884	VW		1.642	W		2.054	VW	023	2.085	W
510	1.801	VW	480	1.624	VW	351	2.007	»	203	2.049	VW
150	1.731	M	110	1.600	»	190	1.945	»	400	2.002	VW
	1.689	VW	600	1.578	W	510	1.896	W	213	1.922	VW
313	1.659	VW	551	1.510	»	512	1.676	VW	123	1.881	MS
223 312	1.634	VW	010	1.484	VW		1.645	W		1.828	VW
			661 461	1.423	W	480	1.629	VW		1.653	»
531	1.608	M				111 0	1.598	»	420+	1.634	MS
403	1.590	W		1.336	VW	552	1.576	»		1.588	M
351	1.528	»		1.280	»		1.524	»		1.539	M
	1.504	»				012 0	1.504	»		1.459	M
	1.463	VW				442	1.488	»		1.419	M
	1.393	W				661	1.429	M		1.395	M
	1.372	»					1.397	VW		1.363	VW
	1.326	»					1.367	»	435	1.299	»
	1.284	»					1.336	»		1.240	»
	1.226	»					1.305	»		1.222	»
							1.282	W		1.175	»
							1.043	W		1.160	»
									433	1.134	»

The symmetry and the unit cell dimensions were determined with the single crystal Weissenberg method. The values obtained were checked by calibrated powder photograph measurements (Table 3). The indices of the d-lines and the refinement of the cell dimensions were obtained with the aid of the Wegematic 1 000 electronic computor. Mr. Törn, M. Sc., of the Computing Centre in Turku, drew up the program for computation.

Table 1 shows that the mineral is very rich in sodium and in ferric iron. The aegirine content must thus be high in the mineral analysed. This fact is also observed in the optical properties of the mineral. For comparison, optical values reported for blanfordite by Fermor (1909), Bilgrami (1956), Zwaan and van Plas (1958), and Kilpady (1960) are reproduced in Table 2.

The refractive indices of blanfordite from Chikla are considerably lower than those measured on our mineral. This is in fair accordance with the higher aegirine content of the now analysed mineral. The values measured on the blanfordite from Nagpur by Zwaan and van der Plas (*op. cit.*) indicate a still higher aegirine solid solution in the mineral. The now analysed mineral is, according to the optical properties highly similar to the variaties studied by Fermor (1909) and by Kilpady (1960).

The single crystal X-ray photographs show systematic extinctions in agreement with the space group C 2/c. The cell dimensions do not greatly differ from the values measured for the pure aegirine end member by Yagi (1958). The small manganese content in the mineral does not greatly affect the cell size. Its peculiar pleochroim is the only property of the mineral that readily shows the manganese content.

MANGANARFVEDSONITE (JUDDITE)

This amphibole mineral occurs as small, 1—2 mm grains together with blanfordite in an even-grained, banded microcline gneiss. The mineral has a beautiful pink colour and a strong pleochroism from light rose to violet. After separation in Clerici's solution, the mineral was chemically analysed and investigated optically as well as with X-rays, using powder and single crystal methods.

The chemical composition of the mineral is shown in Table 4; together with the calculated atomic content of the unit cell. The optics of the mineral are illustrated in Fig. 1 and the numerical values given in Table 5.

The powder data of the mineral are listed in Table 3. The d-lines are indexed by means of the cell dimensions obtained on the Weissenberg photographs.

The analysed mineral can be classified as belonging to the eckermannite-arfvedsonite series. The ratio Mg : (Mg+Fe+Mn) = 0.55 indicates that the mineral should be called magnesioarfvedsonite, after the nomenclature

Table 4. The chemical composition and the unit cell of manganarfvedsonite (juddite) from Goldongri, India. Chemical analysis by Paavo Väänänen, M. Sc., of the Geological Survey of Finland.

	wt %	cation %	Unit cell content		atoms	
SiO_2	56.02	48.1	Si	15.78	16.00	
Al_2O_5	2.49	2.5	Al	0.83		The unit cell:
Fe_2O_5	12.92	8.4	Fe^{3+}	2.74	3.97	a = 9.75 Å
FeO	—					b = 17.84 »
Mn_2O_5	2.91	1.9	Mn^{3+}	0.62	4.78	c = 5.28 »
MnO	0.28	0.2	Mn^{2+}	0.07		β = 104°7′
MgO	10.03	12.8	Mg	4.23		Vo = 890.6 Å^3
TiO_2	2.27	1.7	Ti	0.48		Z = 4
CaO	0.46	0.4	Ca	0.14	6.30	
Na_2O	10.46	17.5	Na	5.72		Space group C 2/m
K_2O	1.23	1.3	K	0.44		D = 3.159
H_2O^+	0.90	5.2	OH	1.68		
H_2O^+	0.00		O	45.58		Mg : (Mg+Fe+Mn) = 0.55
CO_2	n. d.					
PbO	n. d.					
CuO	n. d.					
P_2O_5	0.04		P	0.01		
	100.01	100.0				

Table 5. The optical properties of juddite

	Goldongri, India	Chikla, India (Bilgrami, 1955)	Nagpur, India (Zwaan and van de Plas, 1958)
α	1.663 rose	1.667	1.658 yellow
β	1.684 violet	1.683	— carmine — violet
γ	1.697 orange	1.692	1.668 pinkish orange
2 V γ	70°	62°	70°—72°
c : α	8°	(c : γ) 3°	20°—35°
(110) : (110)	56.0°		55.5°

adopted by Deer, Howie, and Zussman (1963). Fermor (1909) described an alkali amphibole from Kacharwahi, India, and named it juddite. The arfvedsonite described in this paper is highly similar to this mineral juddite of Fermor. It also resembles the juddites from other Indian localities described by, *e. g.*, Bilgrami (1955), Zwaan and van der Plas (1958) and Nayak (1961). The similarity of these minerals is observed in the optical properties given in Table 5. The chemically analysed juddite from Chikla, India (Bilgrami 1955) differs from the now analysed mineral mainly by its higher manganese content (7.69 wt % MnO). Bilgrami classified Chiklan juddite with the richterite type of amphiboles. This was justified by the similarity in the optical orientation of the minerals. Judging by its optical characters, Nayak (1961) considered juddite from the Kajlidongri manganese mine to be related to the riebeckite group as described by Miyashiro (1957). The present amphibole resembles arfvedsonite in having the optic axial plane perpendicular to the symmetry plane.

The calcium content of the mineral is considerably lower than that of the juddite from Chikla and lower than observed in most arfvedsonites. The low lime content is, however, in accordance with the ideal end member formula proposed for the arfvedsonite—eckermannite series by Sundius (1945).

THE GOLDONGRI MANGANESE MINE, PANCHMAHAL DISTRICT, GUJARAT STATE

The Goldongri is a small isolated hill which rises to about 60 m from the surface level and is 90 m long and nearly 60 m wide. The only work conserning it in the literature is one by the late Fermor (1909), who visited the place in 1902 and described its geology, mineralogy and petrography to some extent. The ore-body is intricately folded and an interesting variety of minerals occurs here in the contact between the manganese ore and the pegmatite and granite intrusions.

During the years 1957—1959, one of the authors (V. K. N.) visited the Goldongri Mines and collected several interesting mineral samples for investigation. The present paper records the studies on eckermannite and piemontite.

ECKERMANNITE

This alkali amphibole occurs in Goldongri as grains 1 mm in diameter in an even-grained rock with quartz, microcline, oligoclase, apatite, biotite and aegirinic pyroxene. The mineral is characterized by a strange colour, which varies, in thin sections, from violet to light blue.

The purified mineral was chemically analysed. The analysis is given in Table 6. The unit cell content and the calculated formula agree well with the ideal formula of the eckermannite-arfvedsonite series suggested by Sundius (1945): $Na_3(Mg, Fe^{+2})_4(Fe^{+3}, Al)Si_8O_8O_{22}(OH,F)_2$. Some of the trivalent iron is replaced by Mn^{+3} in the analysed mineral. The magnesium content of the mineral is very high and the Mg to (Mg+Fe+Mn) ratio puts the mineral into the eckermannite end of this series. Optically, the mineral now under consideration is similar to the eckermannite mineral originally described by Adamson (1942, 1944) from Norra Kärr, Sweden. The optical properties are given in Table 7. The main difference is observed in the angle of extinction and in the pleochroism. The difference in colour is evidently associated with the manganese content of the Indian mineral. The manganese percentage of this mineral, however, is considerably lower than that of the manganarfvedsonite described above.

Table 6. Chemical composition and the unit cell of eckermannite from Goldongri, India. Chemical analysis by Paavo Väänänen, M.Sc.,of the Geological Survey of Finland.

	wt %	cation %	Unit cell content			
SiO_2	55.27	46.34	Si	15.51	15.89	The unit cell:
Al_2O_5	1.16	1.14	Al	0.38		
Fe_2O_5	8.21	5.18	Fe^{3+}	1.73	9.46	a = 9.80 Å
FeO	—					b = 17.91 »
Mn_2O_5	1.16	0.74	Mn^{3+}	0.25		c = 5.29 »
MnO	0.00					β = 104°6′
MgO	17.75		Mg	7.42		Vo = 900.26
TiO_2	0.29	22.18	Ti	0.06		Z = 4
CaO	4.95	0.18	Ca	1.49	6.28	Space group C 2/m
Na_2O	7.95	4.44	Na	4.32		D = 3.099
K_2O	1.32	12.92	K	0.47		Mg : (Mg+Fe+Mn) = 0.79
H_2O+	0.88	1.40	OH	1.69		
H_2O-	0.00	4.92	O	45.66		
CO_2	n. d.					
PbO	n. d.					
CuO	n. d.					
P_2O_5	0.80	0.56	P	0.19		
	99.73	100.00				

Table 7. Optical properties of (mangan)eckermannite from India and of eckermannite from Norra Kärr, Sweden.

	(Mangan)eckermannite, Goldongri, India	Eckermannite, Norra Kärr, Sweden (Adamson, 1944)
α	1.624 rose	1.636 bluish green
β	1.631 light lila	1.644 light bluish green
γ	1.637 light blue	1.649 pale yellowish green, almost colourless
2 V α	72°	75°
c : α	49°	25°—53°

The calcium content of the mineral is quite high (4.95 wt % CaO). In this respect, it could be grouped with richterites and winchites. The optical properties of the mineral and the orietation of the optic axial plane (Fig. 1 and Table 7) are so closely related to eckermannite that the mineral here is classified with the eckermannite series. It is possible that many of the minerals called winchites have the same kind of relation to juddites as the eckermannite has to arfvedsonite.

The X-ray powder data of the mineral are given in Table 3. The powder lines show a close relationship to juddite.

PIEMONTITE [1])

The mineral occurs as well formed short prisms, about 10 mm in length, together with calcite, plagioclase; spessartite garnet, phlogopite, quartz and oxide ore. The mineral was chemically analysed by Mr. Väänänen. The analysis is given in Table 8. The table includes the unit cell dimensions, the calculated atomic content of the cell and the optical properties of the mineral. The d-spacings of the mineral are given in Table 3.

Table 8. Chemical composition and the unit cell of piemontite from Goldongri, India. Analyst Paavo Väänänen, M. Sc., of the Geological Survey of Finland.

	wt %	Cation %	Unit cell content			
SiO_2	36.40	35.0	Si	5.91	5.96	The unit cell:
Al_2O_5	15.89	18.0	Al	3.05		a = 8.88 Å
Fe_2O_5	11.06	8.0	Fe^{3+}	1.35	2.00	b = 5.66 »
FeO	—					c = 10.16 »
Mn_2O_5	11.99	8.8	Mn^{3+}	1.48	3.92	β = 115.5°
MnO	0.55	0.4	Mn^{2+}	0.08		Vo = 464.67
MgO	0.00					Z = 4
TiO_2	0.11	0.1	Ti	0.01		Space group P 2_1/m
CaO	22.25	23.0	Ca	3.88		D^4_{20} = 3.522
Na_2O	0.41	0.8	Na	0.13	4.02	
K_2O	0.04		K	0.01		The optical properties
H_2O+	0.93	5.9	OH	0.97		(Sodium light)
H_2O-	0.00		O	25.24		α = 1.768 yellow
CO_2	n. d.					β = 1.803 orange
PbO	0.04					γ = 1.843 carmine
CuO	n. d.					2 V α = 68°
P_2O_5	0.13	0.1	P	0.02		c : α = 38°
	99.80	100.0				

The unit cell content readily shows that there is just enough calcium and sodium in the mineral to fill the eight co-ordination sites in the structure (3.88+0.13 = 4.01). On the other hand, there is just enough silicium to occupy the tetrahedral positions (5.96). As usual in epidote minerals, Al evidently occupies the two octahedral sites in the cell which belong to the chain structure of the mineral. The remaining cations per unit cell, *e. g.*, 1.0 Al, 1.4 Fe and 1.5 Mn, divide among themselves the four remaining octahedral positions. This corresponds to the solid solution mixture: 27 mole % clinozoisite, 35 % pistasite and 38 % piemontite ($Ca_2Mn_2AlSi_3O_{12}(OH)$). The piemontite now analysed is thus unusually rich in iron and manganese. In fact, the sum of the pistasite and piemontite is one of the highest ever analysed. This is clearly seen in the refractive indices of the mineral, which

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[1]) The term piemontite — corresponding to the original Italian place name — is used instead of »piedmontite» for the reason given by Deer, Howie and Zussman (1962).

are higher than any values reported earlier. The same holds also for the specific gravity of the mineral. The piemontite from Långban, Sweden (Malmqvist 1929), and that from Chikla, India (Bilgrami 1956), have higher contents of $Fe_2O_3+Mn_2O_3$ but the indeces are lower. The refractivity measured on those minerals must be associated with the considerably higher Mn/Fe ratio. The optical properties measured from the now analysed material do not fit the curves drawn by Deer, Howie and Zussman (1962) because the curves do not take into account the variations in the Mn/Fe ratio. The pistasite ($Ca_2FeAl_2Si_3O_{12}(OH)$) - piemontite ($Ca_2MnAl_2Si_3O_{12}(OH)$) diagram of Marmo, Neuvonen and Ojanperä (1959) cannot be used for comparison since their curves take into account only those piemontites in which the ratio Al to (Fe+Mn) is about 2. In the mineral now analysed this ratio is as low as 1.05.

The unit cell dimensions determined agree well with those reported by Marmo, Neuvonen and Ojanperä (1959). They differ greatly, however, from the values usually listed for piemontite. The deviation cannot be explained by the difference in the choice of the monoclinic unit cell. The calculated density for the pure piemontite, $Ca_2\ Mn_2\ AlSi_3O_{12}\ (OH)$, is as high as 4.2 if computed according to the cell dimensions given in the hand books $a = 8.05$, $b = 5.70$, $c = 9.41$ and $x = 115°42'$). The density calculated from the dimensions now measured is 3.57 and thus agrees well with the density value obtained with the direct pycnometer measurements 3.522. It is most likely, consequently, that there is an error in the earlier cell size determinations.

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4

A PRECAMBRIAN MOLASSE FACIES DEPOSIT IN GHANA [1]

BY

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ABSTRACT

Field and petrographic evidence is described which ends the speculation on the place of a group of arenaceous and rudaceous rocks in the Birrimide cycle. These deposits contain fragments of late-orogenic metasomatic rocks, but at the same time have been partially deformed by the forceful emplacement of a late- to post-orogenic granodiorite.

CONTENTS

INTRODUCTION

Associated with Birrimian rocks in widely separated parts of Ghana are deposits of arenaceous and rudaceous material which have been variously described as of Birrimian, Tarkwaian, or intermediate age (Junner 1940, Hirst and Junner 1946, Murray 1960). There is no reason to suppose that all these deposits are necessarily contemporaneous but some of the occurrences have features in common and the deposit to be described in this account shows clear structural, metamorphic, and petrographic evidence

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of its origin in the late stages of the Birrimide orogeny.1) This deposit, near Bibiani in the Western Region of Ghana, has previously been described as either Tarkwanian (Bates 1939) or latest Lower Birrimian in age (Hirst and Junner 1946).

Table 1. Stratigraphical succession.

Assumed (Hirst and Junner 1946)	Proposed (this paper)
Tarkwaian	Tarkwaian
……… unconformity ………	Arenaceous Group ? ……… unconformity ………
Upper Birrimian	……… unconformity ………
Arenaceous Group	Upper Birrimian
Lower Birrimian	Lower Birrimian

GENERAL DESCRIPTION

The Birrimian of the Bibiani area has a regional NNE-SSW strike and the most prominent feature is the hill range underlain by the metavolcanics and other rocks of the Upper Birrimian. This range is flanked on the east by low-lying ground underlain by phyllites, greywackes, and other minor rocks of the Lower Birrimian (Lobjoit 1963). The Upper Birrimian outcrop has been divided into a Central Zone of metasomatised metavolcanics and two flanking Border Zones of unmetasomatised metavolcanics. The latter are mainly metabasites, though andesitic and more acid rocks are represented, and they have been orogenically metamorphosed (Misch 1949) in the green schist facies. The metasomatites are quartz diorites, grading to adamellites, and are representative of the amphibolite facies (Lobjoit 1963 and manuscript in preparation). This account is concerned only with the eastern flank of the Bibiani hills.

The deposit to be described occurs as a strip from a half to one mile wide on the lower slopes of the hills, where it occupies a position astride the Lower-Upper Birrimian boundary. The rocks have been called the Arenaceous Group by reason of the volumetric superiority of the sandstones over the conglomerates and siltstones.

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1) The term »Birrimide» has been introduced by the writer as a name for the orogeny because the term »Birrimian» has hitherto been used for both the system of geosynclinal rocks and the subsequent orogeny and its accompanying infracrustal rocks. Thus »Birrimian granites» is a distinctly misleading expression.

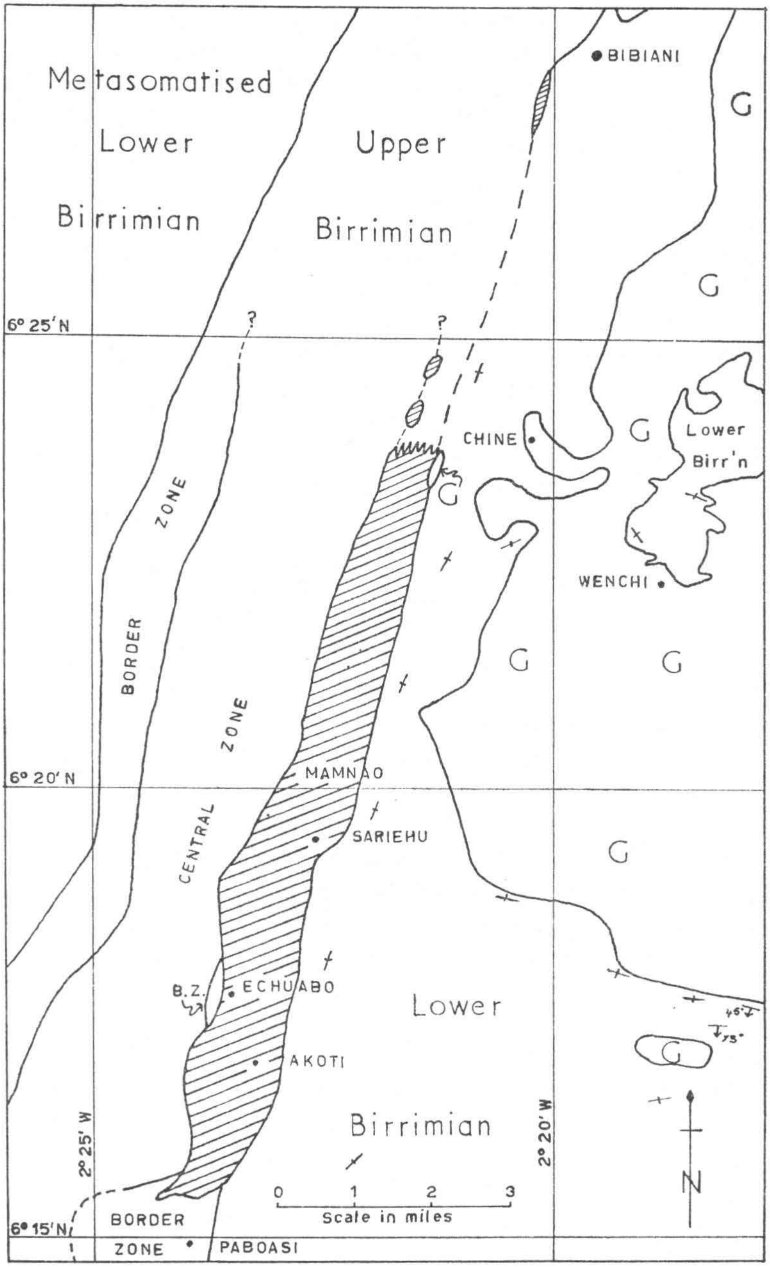

Fig. 1. Sketch map of the Bibiani—Paboasi area showing the essentials of the geology. The Arenaceous Group is shown shaded and the Wenchi granodiorite and its satellites are marked with a G. The small area of Border Zone rocks near Echuabo is labelled B.Z. . The Upper Birrimian near Bibiani has not yet had its zonation mapped and the boundary between Upper and Lower Birrimian from Bibiani to Chine is shown as a broken line to indicate a degree of uncertainty.

In the part of the outcrop between Paboasi and Sariehu these rocks are undeformed and look like typical unmetamorphosed sandstones, *etc.*, though there is some evidence for the beginning of mechanical deformation in the north of this sector. Bates (1939) records pebbles up to 10 inches in length from some of the conglomerates. Pebbles are commonly quartz, pink »quartz porphyry» (a metasomatic rock whose significance is described later), greenstone, phyllite, quartzite, and (?) porphyrite. The rock fragments are more angular than the quartz and quartzite. These conglomerates are apparently restricted to the western margin of the Group outcrop.

The sandstones range in colour from red and brown through purple and mauve (and rare green) to pale cream or buff. Isolated rounded pebbles are often seen and one was identified as a coarse diorite. The less common siltstones are similar to the sandstones though finer grained and usually darker.

Bedding is generally inconspicuous. The strike is parallel to the regional trend of the Birrimian and dips range from 10° to 50° westwards, but eastward dips of 26°, 29° and 45° have also been recorded. Beds, where visible, are from one to five feet thick. Nice (1962) records dips, on a sandstone cliff-face about 50 feet high, ranging from 45° at the base to 10° at the top. On the other hand, he also mentions a twenty-five foot high exposure showing no signs of bedding. A poorly developed graded bedding is the only sedimentary structure that has been seen, and this rarely.

Upper Birrimian metavolcanics are exposed to the east of the Group outcrop near Paboasi, but before Akoti is reached, going north, only Lower Birrimian phyllite is found in this position. Within the Group outcrop phyllite is exposed in its eastern half and diorites or metavolcanics in the western half, giving an impression of interbedding that is later shown to be false. The western margin of the Arenaceous Group is in contact with either the diorites and other rocks of the Central Zone or with the metavolcanics of the Border Zone. The boundary between the two Zones apparently winds »into» and »out of» the Arenaceous Group outcrop. On its eastern margin exposure is scanty in low-lying ground and there is no distinct boundary between the Group and the phyllites and greywackes of the Lower Birrimian. This indistinct eastern boundary is the only main feature of this southern part of the Group outcrop which carries on into the northern part.

At the Mamnao latitude the signs of deformation that were noted in the northern part of the southern sector become the most distinctive feature of the rocks of the Group. Flattened pebbles in the conglomerates are in sharp contrast to the angular fragments in the rocks near Paboasi. Dips from Mamnao northward become high and many are vertical. At the Chine latitude the Group ceases to be a continuous deposit and is represented simply by undeformed sandstones occurring in the area of the Upper Birrimian outcrop. The uncertain boundary between Upper and Lower Birrimian from

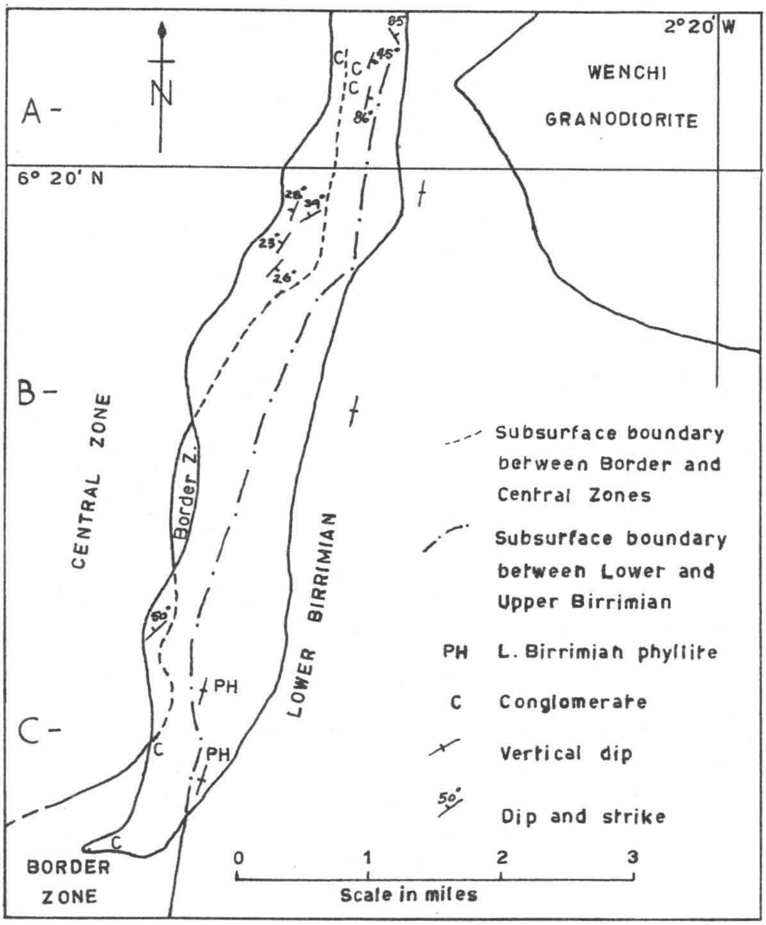

Fig. 2. Geological sketch map of the southern portion of the Arenaceous Group showing some details. The subsurface boundaries have been mapped on the basis of the rocks exposed in the windows. They are therefore postulated boundaries. The letters A, B and C show the positions of the crosssections presented in Fig. 3.

Chine northward is a shear zone and, although Arenaceous Group rocks are undoubtedly present, their identity has generally been lost, certainly past the point where it is possible to distinguish them from other coarser-grained representatives of the Lower Birrimian. Nice (1962) describes examples of the latter from southwest of Chine to near Bibiani and gives sequences on cross-strike traverses indicative of interbedding with phyllitic rocks and green schists. Black carbonaceous fragments are the most common in these greywackes and petrographic examination reveals them as quite distinct from the arenaceous rocks of the Group. This distinction is important as it eliminates contradictory evidence on the nature of the Arenaceous Group.

The Arenaceous Group can thus be divided into three parts:

1. From Paboasi to Sariehu — arenaceous and rudaceous rocks, undeformed and showing low dips.
2. From Mamnao to Chine — gradational from undeformed to deformed and from low dips to high (vertical) dips, from south to north of this sector.
3. From Chine to Bibiani — generally deformed; high or vertical dips; but some undeformed rocks in what would have been the western part of the outcrop.

PETROGRAPHY

The arenaceous rocks are exemplified by specimen WL 88, an analysis of which is given in Table 2. This rock was described in the field (½ mile NNW of Echuabo) as a »gritty sandstone with some moderate quartz pebbles.» In thin section it is seen to consist of fragments of quartz and rock. Feldspar appears to be absent. The quartz (exclusive of the well-rounded pebbles) is mainly sub-angular to sub-rounded and with a low degree of sphaericity. The rock fragments are more rounded and their shape depends upon the rock type: schistose rocks appear as tablets and those with a non-orientated fabric have a moderate sphaericity. The grains fit quite closely together and leave little room for matrix material (now mainly limonite). The rock is a subgreywacke (Pettijohn, 1957).

The quartz and rock fragments are interesting when considered in detail. The quartz is moderately or highly strained, the latter type showing blocky strain shadows and strings of cross-cutting quartz granules. This texture is abundant in the adamellites of the Central Zone. Polycrystalline grains with sutured internal boundaries are common, as is also true among the adamellites. Most of the rock fragments are quartzofeldspathic with more or less sericite and could be groundmass material from quartz- and other porphyries that occur within the Birrimian outcrop in this area. The evidence for provenance from this specimen is suggestive but not conclusive.

Other specimens from the Group differ from WL 88 in often having plagioclase fragments (in the oligoclase-andesine range) and in having rock fragments that are more readily distinguishable. Thus, in a coarse conglomerate near Paboasi, pebbles and fragments of what appears to be pink quartz porphyry are revealed in thin section to be of a type of metasomatic rock that occurs in the Central Zone. The quartz »phenocrysts» are poikiloblasts that replace plagioclase, alkali feldspar, and earlier quartz. The cuspate boundaries are characteristic of this type of replacement, on the small scale,

Table 2. Chemical analysis, expressed as weight per cent and as cation per cent, and epinorm of subgreywacke WL 88, with two selected analyses for comparison.

	1.		2.		3.		4.	5.
SiO_2	77.1	Si	77.25	Qz	61.33	SiO_2	76.84	79.30
TiO_2	0.43	Ti	0.32	Ab	8.95	TiO_2	—	0.22
Al_2O_3	11.1	Al	13.11	Zo	0.44	Al_2O_3	11.76	9.94
Fe_2O_3	2.59	Fe^{3+}	1.95	Musc.	9.45	Fe_2O_3	0.55	1.00
FeO	3.61	Fe^{2+}	3.02	Kaol.	9.13	FeO	2.88	0.72
MnO	0.02	Mn	0.02	Chl.1	6.85	MnO	tr	0.02
MgO	0.67	Mg	1.00	Mt	2.93	MgO	1.39	0.56
CaO	0.26	Ca	0.28	Il	0.64	CaO	0.70	0.38
Na_2O	0.82	Na	1.59	Ap	0.24	Na_2O	2.57	2.21
K_2O	1.06	K	1.35	Cc	0.04	K_2O	1.62	4.32
P_2O_5	0.11	P	0.09			P_2O_5	—	0.05
H_2O+	2.26	H	(7.55)			H_2O+	1.87	0.55
H_2O-	0.32					H_2O-	—	0.41
CO_2	0.02	C	0.02			CO_2	—	—
	100.37		100.00		100.00		100.18	99.68

1. Subgreywacke WL 88, weight per cent.
2. » » », cation per cent.
3. » » », epinorm.
4. Subgreywacke from Pettijohn (1957), Table 54, B.
5. Jotnian sandstone from Simonen and Kouvo (1955) Sandstones in Finland. Bull. Comm. géol. Finlande 168, p. 57.

Analysis of WL 88 was done in the Ghana Geological Survey laboratories.

while the growth »front» of the blast, being roughly oval, gives the appearance of a normal corroded quartz phenocryst in hand specimen. In other rocks examined fragments of quartz-feldspar replacement textures are of fairly frequent occurrence, though never abundant. Rock fragments occur in sizes down to fine sandstone but the siltstones and most of the fine sandstones are basically quartz and plagioclase, the quartz predominating. The subgreywacke texture is present in the fine-grained rocks, where, on average, the fragments are more angular. The limonitic matrix is common. Quartz and feldspar grain boundaries in many rocks have started to break down but this is considered to be a diagenetic effect (Krynine 1940, James 1955, Dapples 1962).

A full list of rock and mineral fragments recorded from these rocks is given below:

Fine quartzofeldspathic material (? groundmass material)
Fine quartz rock (cf. quartzite)
Fine chlorite — opaque — feldspar rock (? metabasalt)
Fine sericitic rock ± preferred orientation
Black slate or carbonaceous phyllite
Phyllite ± crenulation cleavage
Chlorite-rich metabasite
Feldspar porphyry

Quartz-feldspar porphyry
Quartz porphyry
Myrmekite
Pseudo-granophyre (metasomatic rock from Central Zone)
Pseudo-quartz porphyry
Quartz-feldspar replacement material
Plagioclase-microperthite replacement material
Several mesoscopically identified fragments (see the general description)
Quartz and plagioclase.

Although there is a singular paucity of fragments of rocks close to hand (*e. g.*, diorites and metavolcanics) there is no doubt as to the provenance of the fragments which are present, the most significant of which come from the acid metasomatic rocks of the Central Zone.

In thin section the other coarse rocks from the general area of the northern part of the Arenaceous Group outcrop are seen to be greywackes (*sensu stricto*). They have scattered quartzes; some rounded and some more angular, in a highly sericitic matrix. The quartz fragments do not form a very high proportion of the rock. Rock fragments also occur in these greywackes and they are generally dark carbonaceous slate-like tablets or fine sericitic rocks. Where these undeformed fragments occur in a roughly laminated quartz-sericite greywacke matrix it is clear that the rock is in practically its original state and is not simply a deformed subgreywacke.

DISCUSSION

In assessing the origin of the Arenaceous Group the following points are critical:

1. The rocks are subgreywackes.
2. They are undeformed south of Mamnao and generally deformed north of it.
3. Undeformed rocks occur in the western part of the outcrop where deformed ones occur in the eastern part at the same latitude.
4. The Arenaceous Group is in contact with amphibolite facies quartz diorites, *etc.*, (metasomatic) while being itself unmetamorphosed and unmetasomatised.
5. The rocks of the Group contain fragments of metasomatic rocks from the adjacent Central Zone.
6. Central and Border Zone rocks are exposed in the western part of the outcrop and phyllites in the eastern part.
7. In the extreme south of the outcrop Border Zone rocks occur to the east of the Arenaceous Group. Elsewhere only phyllites and greywackes occur in this position.

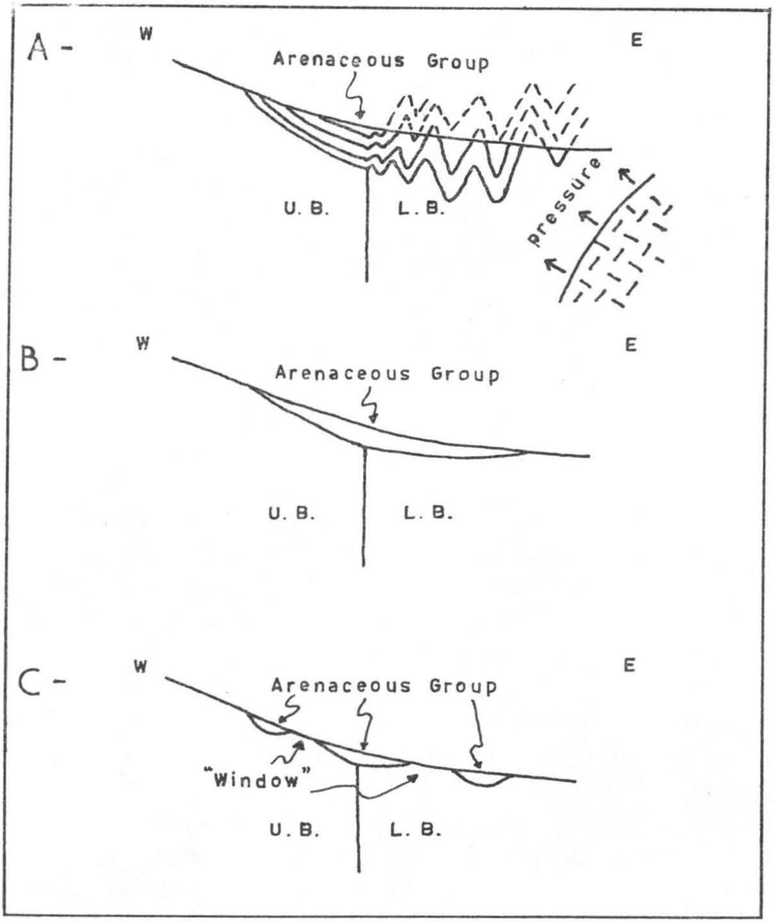

Fig. 3. Cross-sections (not to scale) of the Arenaceous Group and underlying rocks to illustrate the theory described in the text. The letters U. B. and L. B. stand for Upper and Lower Birrimian respectively. In section A the Wenchi granodiorite is shown intruding the Lower Birrimian and so producing the pressure which folded the Arenaceous Group.

There are thus several lines of evidence indicating the nature of the Arenaceous Group. The transgression (point 7.) suggests an origin later than both Lower and Upper Birrimian and their initial metamorphism. This is reinforced by the evidence of the Border-Central Zones boundary in relation to the margin of the Group outcrop. Because the Central Zone is representative of the amphibolite facies, the Arenaceous Group must post-date this metamorphism, for it is itself unmetamorphosed. The metasomatism which gave rise to the rocks of the Central Zone has been identified as belonging to the late-orogenic phase (Lobjoit 1963). This immediately indicates that the Group must have been deposited no earlier than the very latest stages of the orogeny. This is confirmed by the fact that fragments of these metasomatic rocks occur in the rocks of the Group.

The evidence so far suggests that a molasse origin for the Arenaceous Group is the only possible explanation, provided the Group belongs to this orogenic cycle. The fact that the rocks are subgreywackes indicates by this independent line of evidence that a molasse origin is to be considered possible (Pettijohn 1957). The final piece of evidence necessary to date the Group accurately is the deformation. It will be seen that this deformation starts at the same latitude as the southern edge of the granodiorite body centered on Wenchi. This granodiorite shows distinct evidence of having been forcefully emplaced, staurolite and kyanite having been formed in its aureole (Lobjoit 1963 and 1963 a). This forceful intrusion is considered to be the cause of the deformation of the northern part of the Arenaceous Group. That part of the Group which lay over incompetent Lower Birrimian phyllites was readily folded to give higher dips and the crushed fabric. On the other hand, the part of the Group overlying the more resistant metabasites of the Upper Birrimian was to a large extent protected and exists today in its original undeformed state, as does the southern part of the Group, which lay outside the sphere of influence of the deforming forces. In this undeformed part of the outcrop the exposures of Birrimian rocks are interpreted as »windows». The Arenaceous Group is here probably thinner than where folded and the underlying rocks emerge through their cover in places.

There is no evidence indicating whether the Wenchi granodiorite is a late-orogenic or a post-orogenic body; but, since it is clearly a constituent of the Birrimide granitic activity, the Arenaceous Group is equally clearly bracketed between the late-orogenic metasomatism and this yet later stage in the orogeny. The suggested molasse origin thus appears confirmed.

Since these rocks were at the same depth as the roof of the Wenchi granodiorite (or near it) at the time of the formation of the kyanite, it is clear that this depth can be estimated if a satisfactory mean is found between (a) a great depth to account for the presence of kyanite and the fact that the granodiorite is not a rapidly chilled body, and (b) a shallow depth to account for the lack of metamorphism in the Arenaceous Group. On balance a depth of about 4 kilometres can be regarded as reasonable (Lobjoit 1963 a). This means that the Arenaceous Group is merely a fragment of a much greater thickness of molasse facies rocks. The Tarkwaian is the obvious missing series. Junner (1940, p. 13) estimated its thickness as up to 10 000 feet, which is well on the way to the four kilometres suggested. The implications of regarding the Tarkwaian as simply a representative of the late stage of the Birrimide orogeny are many and this is not the place to enter into a discussion of this major point. Suffice it to say that petrographically the rocks are of the kind that would be expected and Junner himself suggests that they are terrestrial in origin (1940, p. 12; see also Arnould 1961, pp. 39—45).

Thus the Arenaceous Group, which appears to be a base to the molasse series, may be simply a local piedmont development (suggested by the metasomatic rock fragments and the situation of the deposit at the junction of two rock series of different weathering properties), later overidden by the major deposit. In the former respect it is similar to rocks described recently from Spitsbergen by Atkinson (1962).

It is to be hoped that further research will shed similar light on the other occurrences of arenaceous and rudaceous rocks in Ghana.

ACKNOWLEDGMENTS

The author wishes to thank the Director of the Geological Survey of Ghana for the opportunity to work in the Bibiani area, and Brian E. Nice for company and disputation in the field.

ADDENDUM

Since this paper was submitted an account by Moon (1963), of similar (though very highly sheared) rocks some 35—40 miles NNE along the strike, has been published. Moon's evidence suggests that Hirst's (1946) alternative is to be preferred but there is the contradictory evidence of transgression of the Group a little to the south (Bates, 1940) and absence of the feldspathisation that affects the adjacent Upper Birrimian. This latter is one of the strongest arguments for the Bibiani-Paboasi Arenaceous Group being post-migmatisation in age and it may well be the deciding factor in determining the age of Moon's grit and conglomerate group.

It is interesting to note that immediately opposite the group outcrop on Moon's map is a string of outcrops of granite believed to be extensive at depth. The Lower Birrimian rocks are described as being more highly metamorphosed near the boundary with the sheared grits and conglomerates. Here, then, may be the clue to the true explanation, in exactly the same way as the deformation caused by the Wenchi granodiorite was the clue to the explanation of the form and presence of the Arenaceous Group.

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5

ON THE OCCURRENCE OF POTASH FELDSPARS OF INFERIOR TRICLINICITY WITHIN THE PRECAMBRIAN ROCKS IN FINLAND [1])

BY

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ABSTRACT

One hundred forty-seven feldspars extracted from porphyroblasts of Precambrian granites, granodiorites and quartz diorites were investigated by X-ray for triclinicity. Some of them were examined also optically. The presence of monoclinic potash feldspars, as well as of others of inferior triclinicity, was indicated to a much larger extent than had been earlier supposed with respect to Finnish Precambrian rocks; but the occurrence of such potash feldspars seems nevertheless to be restricted to particular, comparatively well defined areas. This study should still be considered as being of a preliminary character.

CONTENTS

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[1]) Received August 18, 1963.

INTRODUCTION

One of us has in many connections (Marmo 1956, 1963) stressed the importance of the behaviour and modification of potash feldspar in granitic rocks as providing more than merely a hint as to the origin and evolution of such rocks.

During the geological mapping of certain parts of southern Finland by Matisto (1962), and southeastern Finland by Vorma, as well as somewhat later of Central Finland by Marmo, greatly varying triclinicities were observed among the potash feldspars of the porphyroblastic granodiorites and granites in the areas. These observations were at variance with the earlier views on the Precambrian potash feldspars, but still, with respect to the granite problem, of utmost interest:

— In the 1950's, the opinion steadily grew stronger that the microcline (the prevailing potash feldspar contained in Precambrian rocks) was a product of the microclinization of orthoclase; and Laves (1950) contended on crystallographic grounds that if microcline is cross-hatched, it must have definitely had a monoclinic ancestry. On petrological grounds, Marmo (1956 and 1959) objected, claiming also the possibility of a cross-twinned microcline having a triclinic growth, while admitting, however, that the very initial stage in the growth of the crystals must have been monoclinic.

— Microcline has never been synthetized in laboratories by direct crystallization, but obtained only by a replacement of albite (Laves 1951, Wyart and Sabatier 1956)). Instead monoclinic potash feldspar has always been produced.

— According to recent views (Goldsmith and Laves 1954), microcline cannot exist above 500° C in the presence of water, but will undergo monoclinic modification. On the other hand, a monoclinic potash feldspar may be formed at very low temperatures as well. Therefore, Marmo (1959) suggested that if the temperature of crystallization (in the presence of water) is sufficiently low (probably below 500° C) and the accumulation of materials (growth of the mineral) is sluggish, triclinic potash feldspar will result, as it probably has in the granitized rocks exlusively containing microcline. If the temperature is raised or the crystallization takes place more rapidly, a potash feldspar of inferior triclinicity will be produced, as in the case, for instance, of many younger Alpine granites of marked intrusive character.

If all the suppositions referred to are correct or nearly so, then a thorough study of the potash feldspars of salic plutonic rocks would have a very important bearing on the genetical problems concerning granitic rocks in general. This is the reason why the present study was begun. The senior author (Marmo) is mainly responsible for the petrological part of this work; which is restricted, however, to the occurrences in Central Finland. Vorma is responsible for all the material obtained from southeastern Finland (around and close to rapakivi areas), and also for the optical data. Hytönen is mainly responsible for the triclinicity determinations (and separation) of potash feldspars and, together with Vorma, for the mineralogical and crystallographic considerations.

EARLIER VIEWS ON THE PRESENCE OF ORTHOCLASE IN FINNISH ACID PLUTONIC ROCKS

In earlier geological literature on the Finnish basement, there are some scattered notes on the orthoclase occurring in the oldest Finnish rocks, in which the microcline was always taken as the predominant form of potash feldspar. Seldom, however, is any explanation given on what basis the potash feldspar was considered as being orthoclase. It seems that mostly the extinction angle $\gamma \wedge \perp (010)$ was used for this purpose, but often the lack of typical cross-hatching was probably taken as proof of the presence of orthoclase instead of microcline. In only a few cases was the optical axial angle determined for the potash feldspar occurring in the Precambrian basement of Finland.

In the following, a review of the Finnish literature discussing orthoclase will be made, excluding, however, rapakivi, which has always been considered as containing considerably orthoclase. In this review, the less dependable data will also be included, because recently some of them have been substantiated by X-ray methods.

In his classical work on the Archean formation of the Tampere area, Sederholm (1899) mentioned orthoclase in the following connections: 1. in even-grained granitic pebbles of the conglomerate of the Tampere area the occurrence of orthoclase with patches of microcline (*op. cit.* pp. 26—27), as well as, in conglomerates with pebbles of coarse-grained granite (p. 32), in orthoclase porphyrite (p. 39) and in dacite (p. 41); 2. in porphyrite rich in orthoclase (p. 67); 3. in slightly altered phyllite on the shore of Näsijärvi (p. 87); 4. in porphyritic granite at Lavia, as well as in other similar post-Bothnian porphyritic granites of the area (p. 107) and in the even-grained granites grading into the former (p. 110—111, 115).

In southeastern Finland, in the area covered by the map sheet of Mikkeli, Frosterus (1902 a) claims that the post-Bothnian porphyritic granites are likewise orthoclase-bearing (*op. cit.* p. 42—44). Especial mention is made of the porphyritic granite at Petäjävesi, Central Finland, which is rich in orthoclase and, in the investigator's opinion, comparable with rapakivi granites. This occurrence has been recently re-investigated by Rouhunkoski (1957), who could prove the ample occurrence of orthoclase in this rock.

According to Frosterus (*op. cit.* pp. 44—45), the pegmatite-like younger granites of Asikkala, Heinola and Mäntyharju are likewise orthoclase-bearing. In another paper (Frosterus 1902 b, pp. 4—7), he refers to some medium- and coarse-grained granites as well as certain porphyritic varieties as containing orthoclase besides considerable microcline; similarly, monoclinic potash feldspar should occur in addition to cross-hatched microcline in some augen gneisses belonging to the basement of the Ladogian schists in Carelia (p. 20).

Following the literature in chronological order, the next mention of orthoclase is made by Borgström (1907), who found monoclinic potash feldspar as insets in the granite porphyry dikes at Östersundom.

In the explanatory notes for the map sheet (1 : 400 000) of Tampere, Sederholm (1913) repeats the orthoclase observations which he had made already in 1899. In addition, he mentions the following localities: 1. Orthoclase porphyries at the lakes of Vaavu- and Varvusjärvi (Teisko), at Valkeajärvi (Orivesi) and at Pyljyhuhdanniemi (Messukylä) (*op. cit.* p. 36); 2. Post-Bothnian porphyritic granites on the boundary between the communes of Siipyy and Merikarvia (*op. cit.* pp. 74, 77).

In 1917 there appeared, however, a very remarkable work on the potash feldspars by Mäkinen (1917), a paper which, in its time, was probably the most important contribution to the study of feldspars. According to Mäkinen, the occurrence of orthoclase in Finnish granites, other than rapakivi, is most questionable. Furthermore, Mäkinen claims for all the cross-hatched microclines either an unhatched triclinic or a monoclinic ancestry, thus being secondarily formed possibly from orthoclase. The untwinned insets of potash feldspar occurring in the porphyritic granites of Central Finland have, however, according to Mäkinen, grown primarily as microcline.

The presence of orthoclase in the crystalline schists of Finland was definitely denied by Mäkinen.

The localities mentioned by Väyrynen ((1920, 1923) are of particular interest because they are close to those localities in Central Finland where the special study by the authors of the present paper has been predominantly carried out. Therefore we will take here, as an example, one of the detailed descriptions of Väyrynen (1920, p. 27): In the village of Lahdenkylä (Parish of Ilmajoki), there is a deeply disintegrated alkaline granite, which, when fresh, is grayish green in colour. The granite is coarse-grained and porphyritic, containing lens-shaped insets of orthoclase, up to 6 cm across, in a matrix with black amphibole, albite and quartz. Orthoclase has a homogeneous extinction which is straight on the basal plane, but has an angle of extinction of 5.0° on the clinopinacoids. In addition, the rock contains small grains of highly cross-twinned microcline.

Väyrynen (1923) takes the potash feldspar insets of his younger porphyritic quartz- and granodiorites of southern Pohjanmaa as being partly microclinized orthoclase (*op. cit.* p. 22). In the granitic rocks of his older series, however, the potash feldspar consists always of non-perthitic, highly cross-hatched microcline (*op. cit.* pp. 13 and 19).

The orthoclase-bearing younger rocks of Väyrynen are described as strongly disintegrated and as often containing diopside, hypersthene and biotite. They occur, for instance, around Vähäkyrö, Ilmajoki, Jalasjärvi and Kurikka. Väyrynen has mentioned (p. 43) that similar rock types have been met with by Wilkman at Kiuruvesi, Karttula and Iisalmi (eastern Finland).

The next mention of orthoclase in Finnish Precambrian rocks is made by Eskola (1929) in a paper dealing with potash feldspars in Finnish granites and pegmatites. On p. 54 he mentions orthoclase as occurring in a pegmatite (now a days located in Soviet territory), 1 km W of the mine of Lupikko. As the whole, Eskola considers orthoclase as being very rare in the Archean granites of Finland.

According to Saksela (1935, p. 14), and with the reference to Väyrynen, orthoclase occurs in the granite of Vaasa, which is also mentioned, probably on the basis on the same source, by Laitakari (1942, p. 30).

Edelman (1949) has determined the presence of orthoclase, using the extinction $(\gamma \wedge \perp (010))$, in the granite of Tammo, being there, however, to a large extent transformed into microcline (*op. cit.* pp. 36—37).

Eskola (1952, 1957) has described orthoclase as an essential part of potash feldspars contained in the granulites of Finnish Lapland.

According to Parras (1958, p. 89), the potash feldspar of the charnockites of southern Finland is sometimes unhatched and has a straight extinction.

Excepting Eskola's (1957) determinations, all the identifications of orthoclase listed in the foregoing are based on optical examination. It seems that most of them are based on the measurements of the extinction angle, but, at least in some of the older works, the lack of cross-hatching only may have been used as the criterion, which, however, is very unreliable in drawing conclusions as to the symmetry of a potash feldspar, because there are many descriptions of unhatched microclines in Finland: the potash feldspar of the pegmatites of Tammela (Mäkinen 1912, p. 60), the insets of the porphyritic granites of Central Finland (Mäkinen 1917, pp. 139—140), the potash feldspar of the oligoclase granite of the Otravaara area (Saksela 1923, p. 24), the feldspars of the older series of grano- and quartzdiorites of southern Pohjanmaa (Väyrynen 1923, p. 19), of the pegmatite of Pellotsalo in Impilahti (Eskola 1929, p. 55), etc. The last mentioned microcline is of especial interest, because it has been used by Brown and Bailey (1961) for the refinement of the structure of a maximum microcline.

The information concerning the occurrence of orthoclase in Finnish Precambrian rocks has also been proved by using X-ray methods, and such cases are the following: In the granulites of Lapland the potash feldspar consists to some extent of orthoclase (Eskola 1952 and 1957); the potash feldspar contained in the porphyritic granite at Petäjävesi, Central Finland (Rouhunkoski 1957), and in the porphyritic granodiorites at Nokia, Kuru and Höytämä, southern Finland (Matisto 1962) is orthoclase.

The instances mentioned are all concerned with either synkinematic or postkinematic rocks. The only instance of orthoclase so far met with in the latekinematic granites is that of the granite at Skräddarby in the parish of Sipoo (Härme 1962).

NEW DISCOVERIES OF ORTHOCLASE IN FINNISH PRECAMBRIAN ROCKS

Simultaneously with the work of Matisto (1962), who found the orthoclase in the granodiorites of the areas around Tampere (see above), the authors of the present paper have also had an active interest in this problem, and at the Geological Survey of Finland ample material has been available from different parts of Finland for petrological, mineralogical, and X-ray investigation. The main results of the authors' work are presented in this paper, but, considering the preliminary character of the study from the point of view of the actual regional distribution of orthoclase in Finnish Precambrian rocks and of the petrological significance of these results, no far-reaching conclusions have been drawn.

It should also be mentioned that Dr. Olavi Kouvo, formerly the mineralogist of the Outokumpu mining company and now head of the age-laboratory of the Geological Survey of Finland, has performed, in connection with his work on the radiogenic ages of Finnish rocks, a considerable number of X-ray powder triclinicity determinations for potash feldspars from various parts of Finland — predominantly from Carelian rocks. It is a pleasure for us to acknowledge here our gratitude to Dr. Kouvo for his kindness in placing all his triclinicity data, without any limitations, at our disposal.

The locations of the potash feldspars investigated for the present paper are marked on the map of Fig. 1.

CENTRAL FINLAND

During the geological mapping of the area of the quadrangle sheet 2232 (Keuruu) of the map of Finland, in the village of Keuruu, an acid rock was encountered containing well-formed and partly idiomorphic insets of ortho-

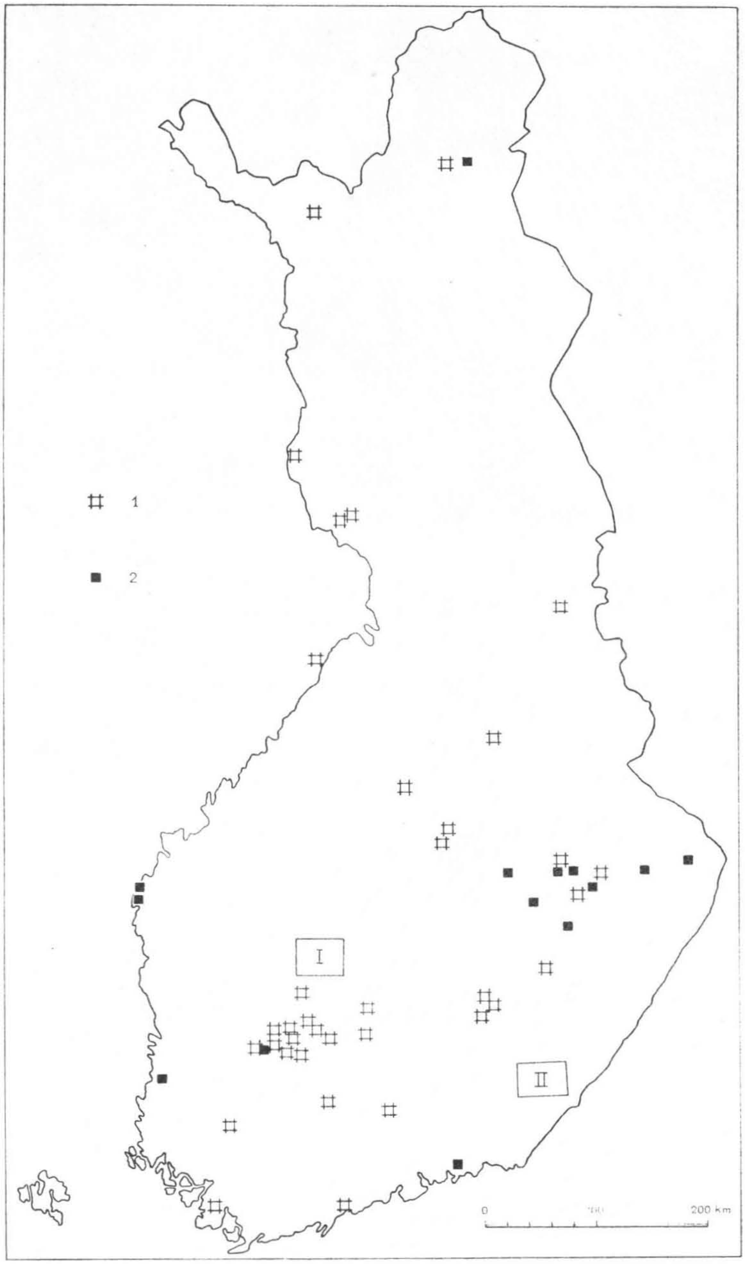

Fig. 1. Location of the samples investigated. 1 = microcline; 2 = orthoclase; I = map sheet 2232; II = map sheet 3134.

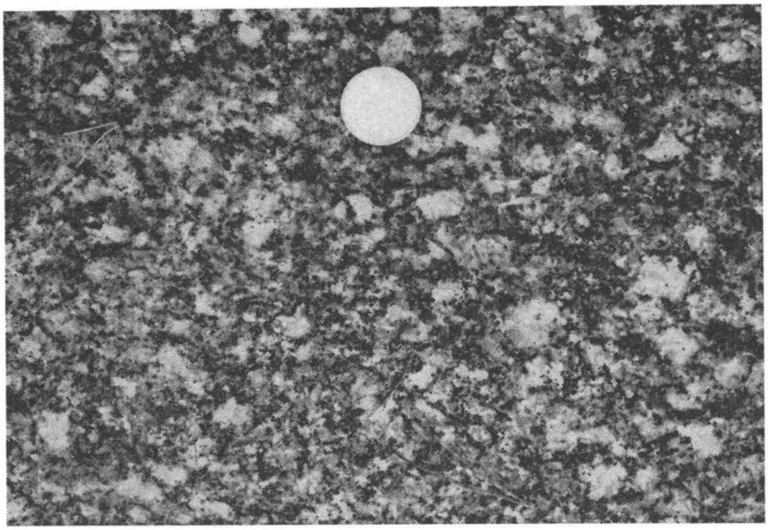

Fig. 2. Granite with orthoclase porphyroblasts at the church of Keuruu, Central Finland.

clase (Fig. 2). This rocks is quite homogeneous and in the yards and along the streets of Keuruu it is well exposed at many places. The potash feldspar is mainly concentrated into insets, the matrix being of granodioritic composition (Table 1). Among the dark minerals, biotite conspicuously prevails

Table 1. Chemical composition and calculated mode of the matrix of monzonitic orthoclase granite. Specimen 449 /LP/62. Keuruu Village. Analyst: Pentti Ojanperä, 1963.

Weight per cent		Calculated mode	
SiO_2	64.79	quartz	23.70
TiO_2	0.54	albite	28.82
Al_2O_3	15.57	anorthite	17.51
Fe_2O_3	1.79	potash feldspar	13.34
FeO	3.68	biotite	11.36
MnO	0.09	magnetite	2.55
MgO	1.52	sphene	1.37
CaO	4.10	apatite	0.34
Na_2O	3.45	rest	0.86
K_2O	3.55		99.85
P_2O_5	0.16		
CO_2	0.00		
H_2O+	0.48		
H_2O-	0.10		
F	0.05		
	99.87		
$-O = F_2$	0.02		
	99.85		

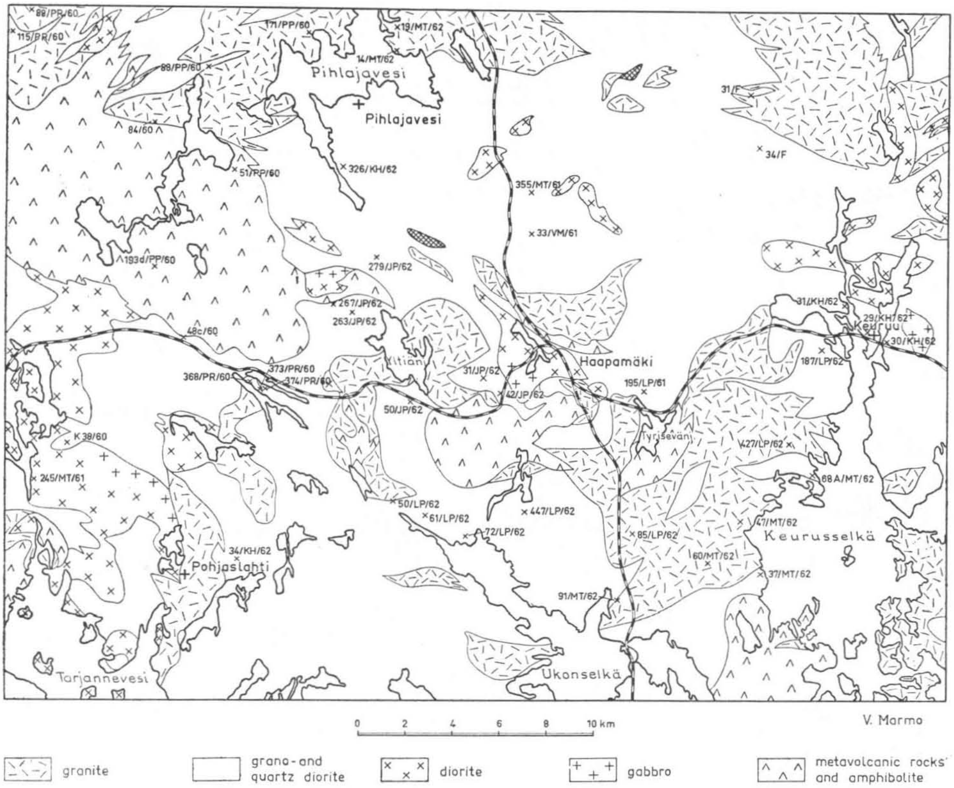

Fig. 3. Location of samples investigated within map sheet 2232 (Fig. 1, area I).

over hornblende. At its contacts this granite seems to grade into orientated and less porphyroblastic varieties. The observation of the presence of orthoclase in it proves to be of very great interest, because, refuting Mäkinen's (1917) negative opinion as to the presence of orthoclase in the rocks of Central Finland, Rouhunkoski (1957) has reported that the rock surrounding a rapakivi-like variety at Petäjävesi, only 25 km E of Keuruu, is gneiss granite containing highly triclinic (triclinicity 0.8—1.0) microcline. The potash feldspar of the rapakivi-like granite, on the other hand, was described as being exclusively orthoclase with a triclinicity of 0.0 and the plagioclase there contains 26—29 % An. In the adjoining granite, the triclinicity varies between the values of 0.0 and 0.8. The insets of the porphyroblastic granodiorite and granite (with plagioclase of An_{20-25}) of the same locality have a triclinicity of 0.0 to 0.3. Furthermore, there occurs a pyroxene quartz diorite with orthoclase insets with a triclinicity of 0.0.

Because also Matisto (1962) had described orthoclase from porphyritic granodiorites some 50 km SW of Keuruu, the senior author of the present

Table 2. The triclinicity values for the predominantly or partially monoclinic potash feldspars extracted from acid rocks in the areas of the map sheets 2232 (Keuruu) and 2214 (Virrat).

Sample	Locality	Triclinicity of an inset	Triclinicity of the potash feldspar of matrix	Remarks
29/KH/62	Keuruu, E of the church	0.0—0.5; max. at 0.0		porphyroblastic granodiorites and granites
30/KH/62	Keuruu, 600 m NWW of the church	0.0—0.75; » » 0.0	no potash feldspar in the matrix	
31/KH/62	Keuruu, Hospital bridge	0.0—0.02; » » 0.0		
34/KH/62	Pohjaslahti, 500 m SW of Tiura	0.0	no potash feldspar in the matrix	
267/JP/62	Sheet 2232/05	0.0—0.2; max. at 0.0	0.0—0.85; max. weak at 0.0	
263/JP/62	Sheet 2232/05	0.0—0.65; max. weak at 0.0	0.35—0.8; max. at 0.8	
279/JP/62	Sheet 2232/05	0.0—0.75; max. weak at 0.0		
50/LP/62	Sheet 2232/04	0.0—0.2; max. at 0.0		
72/LP/62	Sheet 2232/04	0.0—0.75; no max.		
61/LP/62	Sheet 2232/04	0.0—0.75; max. at 0.75		
91/MT/62	Sheet 2232/07	0.0—0.65; max. weak at 0.0		
447/LP/62	Sheet 2232/07	0.0—0.75; max. at 0.75		
536/MT/61	Sheet 2232/05	0.0—0.5; max. at 0.5		
245/MT/61	Sheet 2232/01	0.0—0.75; max. at 0.75		
31/JP/62	Sheet 2232/08	0.0—0.65; max. weak at 0.0		
19/MT/62	Sheet 2232/06	0.0—0.15; max. at 0.0	0.75	
33/JP/62	Sheet 2232/08	0.3	0.0—0.3; max. at 0.0	
355/MT/61	Sheet 2232/09	0.0—0.3; max. at 0.0	0.0—0.75; » » 0.0	
33/VM/61	Sheet 2232/09	0.0—0.8; max. at 0.8		
47/MT/62	Sheet 2232/10	0.2	0.0—0.4; max. at 0.0	
68A/MT/62	Sheet 2232/10	0.45		
37/MT/62	Sheet 2232/10	0.2—0.25	0.0—0.55; no max.	
427/LP/62	Sheet 2232/11	0.0—0.5; no max.		
42/JP/62	Sheet 2232/08	0.0—0.1; max. at 0.0	0.0—0.2; max. at 0.0	
31/F	Sheet 2232/12	0.0—0.85; max. at 0.85		
187/LP/62	Sheet 2232/11	0.0—0.85; no max.		
368/PR/60	Sheet 2232/05 A	0.0—0.45; max. at 0.0		
88/PR/60	Sheet 2232/03 B	0.0—1.0; max. at 0.0		
115 a/PR/60	Sheet 2232/03 B	0.0—0.7; max. weak at 0.7		
374/PR/60	Sheet 2232/05 A	0.0—0.7; max. at 0.7		
326/KH/62	Sheet 2232/06	0.0—0.8; max. at 0.8	0.8	
378/PR/62	Sheet 2214/12	0.0—0.8; max. weak at 0.8		
63/PR/63	Sheet 2214/09	0.0—0.8; max. at 0.8		
436/LP/63	Sheet 2214/05	0.0—0.3		

paper decided to examine the potash feldspars of the whole map sheet more thoroughly, and he was assisted in this work by Hytönen, who also determined the triclinicities of all the samples by X-ray methods.

The distribution of all the samples examined, within the map sheet 2232 (area I of Fig. 1) is shown on the map of Fig. 3. In Tables 2 and 3, a complete list of these triclinicities is given. The determinations have mainly been made for the potash feldspar insets, but, in some cases, the triclinicity of the potash feldspar extracted from the matrix has been determined as well. In the latter case, as shown in Table 2, the potash feldspar consists both of orthoclase and microcline of high triclinicity.

Table 3. The triclinicity values for the triclinic potash feldspars extracted from the insets of acid rocks within the map sheets 2232 (Keuruu) and 2214 (Virrat).

Sample	Locality	Triclinicity	Remarks
50/JP/62	Sheet 2232/05	0.85	see also p. 62
60/MT/62	Sheet 2232/07	0.55	
85/LP/62	Sheet 2232/07	0.75	
373/PR/60	Sheet 2232/05 A	0.7	
14/MT/62	Sheet 2232/06	0.85	
34/F	Sheet 2232/12	0.85	
195/LP/61	Sheet 2232/08	0.85	
171/PP/60	Sheet 2232/06 B	0.8	
88/PP/60	Sheet 2232/03 D	0.9	
3/PR/62	Sheet 2214/12	0.55—0.9	
3A/PR/62	Sheet 2214/12	0.85	
92/PR/63	Sheet 2214/09	0.75	
47/IP/63	Sheet 2214/08	0.85	
Latekinematic aplitic granite and granite porphyry:			
51/PP/60	Sheet 2232/03 C	0.8	granite porphyry
193 d/PP/60	Sheet 2232/02 D	0.8	
84/60	Sheet 2232/03 D	0.9	
K 38/60	Sheet 2232/02 A	0.9	aplitic granite
48 c/60	Sheet 2232/02 C	0.9	

In all the cases microscopically examined the interstitial potash feldspar is highly cross-hatched and thus microcline.

In the case of the orthoclase-granite around the church of Keuruu, the triclinicity has been determined for several insets of each of two samples (Table 4). These determinations have substantiated the microscopical observations indicating microcline, both as patches and grains, within the insets composed predominantly of orthoclase.

Table 4. Triclinicities for separate grains of two samples.

Sample	Locality	Grain	Triclinicity
30/KH/62	Keuruu, 600 m NNW of the church	1	0.0—0.5; no max.
		2	0.0—0.6; max. at 0.0—0.2
		3	0.0—0.75; no max.
		4	0.0—0.85; max. at 0.0—0.2
		5	0.0—0.55; max. at 0.25
31/KH/62	Keuruu, Bridge of hospital	1	0.0—0.1
		2	0.3
		3	0.0—0.1

In order to examine the variation of triclinicities within a single apparently homogeneous rock body, a small quarry was chosen at the railway station of Yltiä. There a very handsome and homogeneous porphyroblastic granodiorite containing some hornblende and traces of molybdenite was

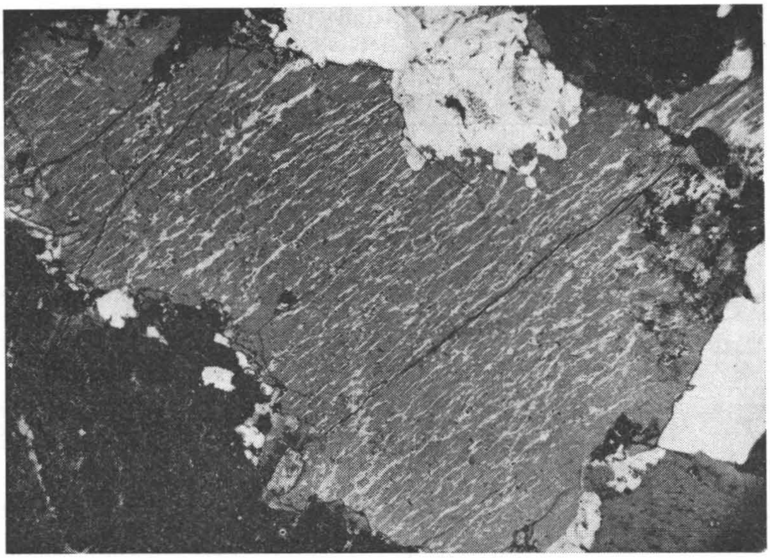

Fig. 4. Porphyroblast of perthitic orthoclase. Granodiorite (42/JP/62. Fig. 1, area I). N +, magn. 40 ×.

quarried for a short while to produce stones used during the building of the railway from Haapamäki to Pori. The first sample taken from this quarry contains large, pinkish gray porphyroblasts of potash feldspar, which have a triclinicity of 0.85 (50/JP/62, Table 3). Later on, three more samples were collected from different parts of the same quarry, and they have the following values of triclinicity:

33—1/KH/62: Mainly 0.9, small part 0.0
33—2/KH/62: 0.0—0.8, no max.
33—3/KH/62: 0.0—0.75, max. at 0.75

This test also supports the view that within an area like that of map sheet 2232, where low triclinicities are conspicuously widespread, true, homogeneous microcline porphyroblasts are uncommon and probably occur patch-wise. Concerning the latekinematic granites of the same area, on the other hand, microcline is the only modification of potash feldspar so far met with.

There is no sense in describing petrologically and mineralogically all the rocks examined and listed in Tables 2 and 3, the rock types of the area having been described by Marmo (1963). As examples, however, the following very brief type descriptions will be given.

Samples 301 and 355/MT/61 (Table 2), porphyroblastic granodiorite, contain insets of sericitic plagioclase with 27 % An (measured using the

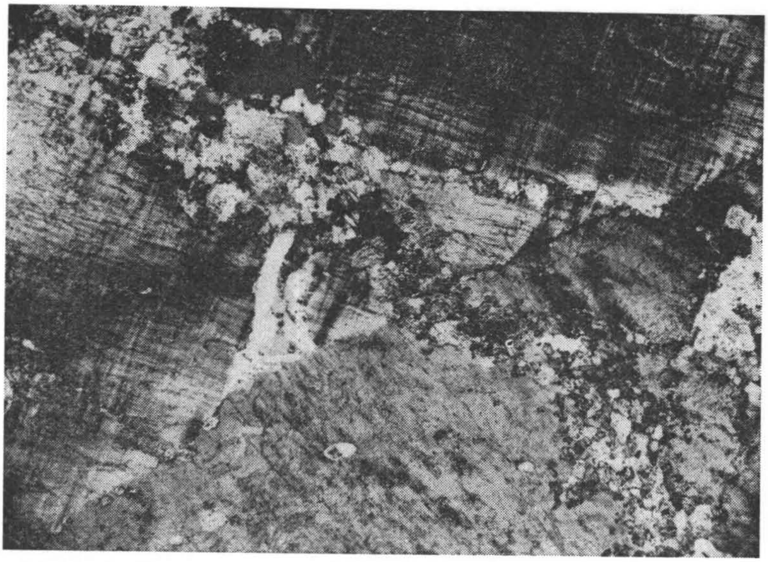

Fig. 5. Orthoclase and microcline occurring in the same porphyroblast. Granodiorite (33/VM/62, Fig. 1, area I). N +, magn. 40 ×.

extinction angle) and of potash feldspar, which is variously and patch-wise cross-hatched. Some grains are clear, homogeneous and ungrated, and they have straight extinction $N\gamma \wedge \perp (010)$. The matrix is made up of a medium-grained mixture of quartz, plagioclase, microcline, hornblende, biotite, epidote, chlorite, and contains a few scales of a very minute, brownish mineral, which, according to X-ray determination, belongs to the vermiculite group. It may be mentioned that in this area this mineral is exceedingly sparse, but still met with in some tens of thin sections.

Sample 43/JP/61 (Table 2) is monzonitic in composition, containing insets of plagioclase (An_{27}) and potash feldspar (triclinicity 0.0—0.7 with max. at 0.0). The matrix is made up of plagioclase, microcline, orthoclase, quartz, biotite, apatite, sphene and epidote.

Sample 33/VM/61 (Table 2) represents one of the most common types of porphyroblastic granodiorites in the area here under consideration. The potash feldspar of the porphyroblasts varies greatly in appearance, and its triclinicity is 0.0—0.8 with a max. at 0.8. In some cases it is a well cross-hatched, slightly perthitic microcline, or in other cases strongly perthitic, like that illustrated in Fig. 4, and obviously orthoclase. In some porphyroblasts, microcline and orthoclase occur together (Fig. 5), and they may contain almost all the minerals occurring in the matrix as inclusions, which indicates clearly that these insets are true porphyroblasts. Around the microcline grains, the occurrence of myrmekite ranging from coarse to fine

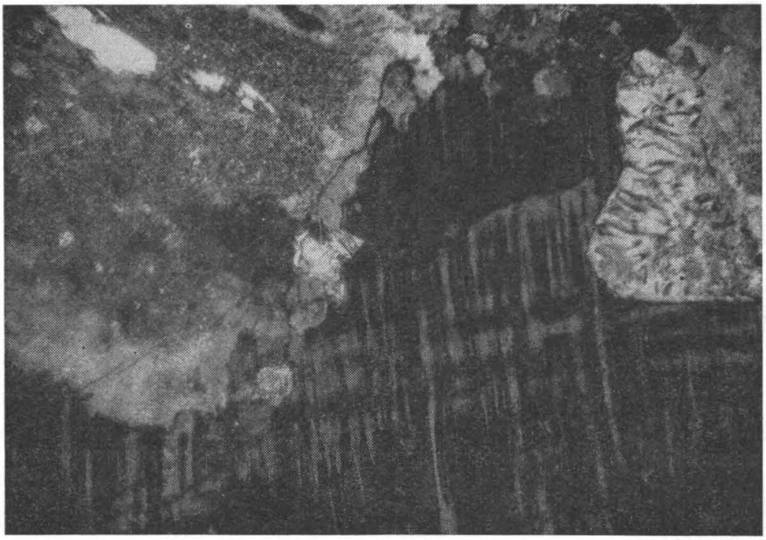

Fig. 6. Microcline with myrmekite in a granite. Pihlajavesi (Fig. 1, area I). N +, magn. 40 ×.

is not uncommon (Fig. 6). Plagioclase (An_{25}), when forming coarse grains, sometimes displays a zonal structure (Fig. 7). The matrix is similar to that of 43/JP/61, but contains ample hornblende in addition.

The area here under consideration is large: more than 1 200 sq. km, and it is an essential part of an area commonly considered as the »granite of Central Finland». Recent geological remapping on a larger scale than earlier, and also in much more detail, has shown, however, that this »granite area» is by no means homogeneous, but contains abundant sedimentary and volcanic strips, as well as plutonic basic and ultrabasic rock types. Furthermore: the granite is there in the minority, for quartz- and granodiorites are the predominant acid rocks, the porphyroblastic varieties of these rock types being conspicuously common.

Such a situation led the authors to wonder whether the orthoclase richness could be typical of porphyroblastic feldspathic rocks also in general. Therefore, in the environment of the area of map sheet 2232, scattered samples of similar rocks were collected and the potash feldspar of porphyroblasts examined for triclinicity. The results are seen in Table 5. They indicate that, excepting the Orivesi sample (37/KH/62), the occurrence of less ordered varieties of potash feldspar is no omnipresent feature of synkinematic porphyroblastic rocks, in addition to which one feature already mentioned (see p. 00) — that in the late kinematic granites the potash feldspar is highly ordered microcline — seems to get support from the data of Table 5.

Fig. 7. Zonig in plagioclase. Porphyroblastic granodiorite (33/VM/62, Fig. 1, area I). N +, magn. 40 ×.

Table 5. Triclinicities of the potash felspar extracted from granites and granodiorites of various localities in Finland.

Sample	Locality	Triclinicity	Rock type	Remarks
A. Porphyroblasts of rocks around map sheet 2232				
26/KH/62	10 km from church of Kuhmoinen towards Jämsä	0.85	granodiorite	
27/KH/62	9 km S of Jämsä	0.8	»	
35/KH/62	Ruovesi, Salonsaari	0.8	»	
B. Porphyroblasts of rocks from remote localities, South Finland				
37/KH/62	Orivesi, 6 km S of the railway st.	0.0—0.65, max at 0.65	granodiotite	matrix 0.0—0.8, no max.
38/KH/62	Orivesi, 1 km S of the road to Siitama	0.85	»	
160 a/JS/62	Loimaa, Kojonperä	1.00	porphyroblastic granite	
C. Latekinematic granites				
282/M-La/50	Kemiö	0.9	pink granite	
—	Espoo, Otaniemi	0.95	migmatizing granite	
—	Enontekiö, Kökkälänjoki bridge ..	0.95	granite	so-called »Hetta granite»
MH/59/47	Kemi	0.9	»	

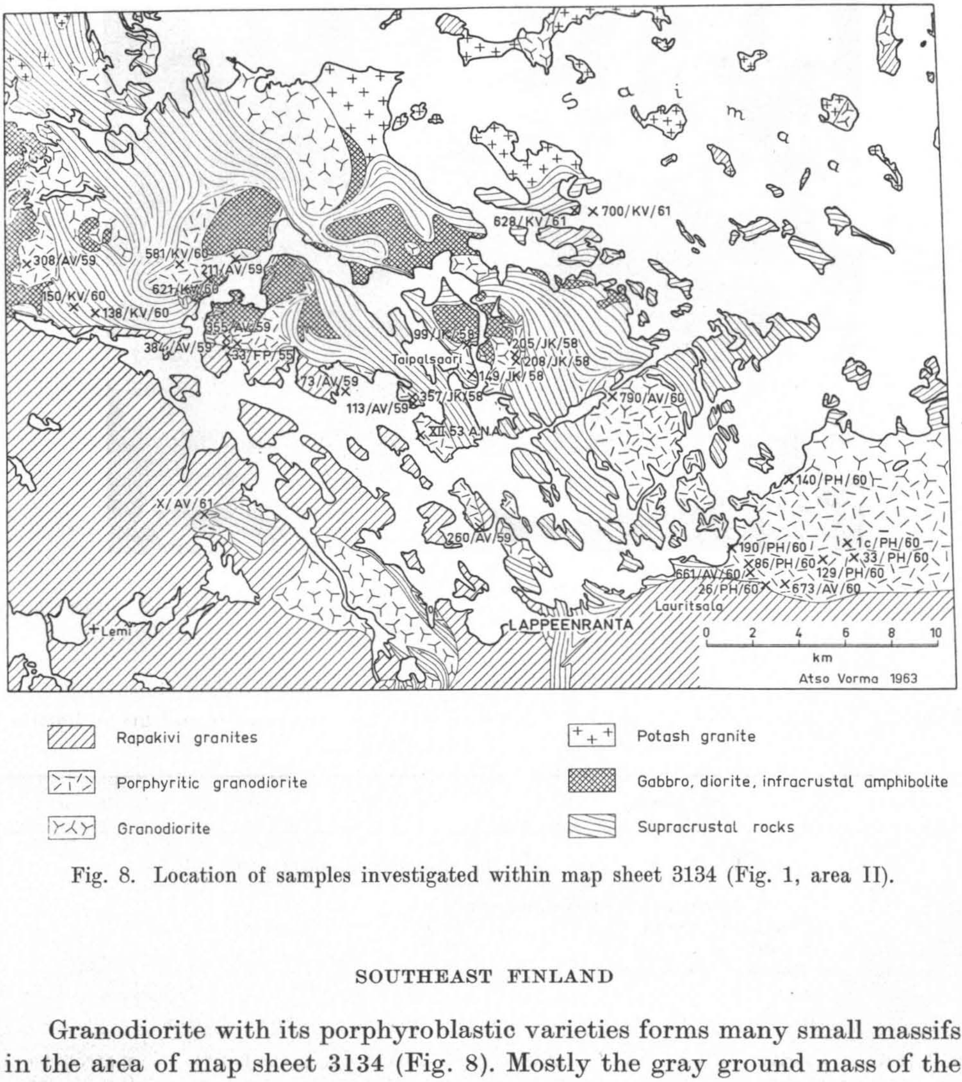

Fig. 8. Location of samples investigated within map sheet 3134 (Fig. 1, area II).

SOUTHEAST FINLAND

Granodiorite with its porphyroblastic varieties forms many small massifs in the area of map sheet 3134 (Fig. 8). Mostly the gray ground mass of the rock is slightly foliated and medium-grained. Potash feldspar insets vary from 0.5 cm to 5 cm (or even to 7 cm) in diameter. In places they are very far apart, in places side by side. Between these two types there occur all the varieties in between. The most common is the variety where insets of 2—5 cm in diameter are at distances of 2—4 cm from each other (Fig. 9). The insets are usually angular, in more basic varieties often also ovoidal in shape.

The gabbros and diorites of the area are regularly older than the granodiorites: occasionally, however, a gradual transition between diorites and granodiorites is seen. The transition from even-grained to porphyroblastic

Fig. 9. Porphyroblastic granodiorite (308/AV/59, Fig. 1, area II).

granodiorite is usually gradual (Fig. 10). Sometimes the transition zone between the even-grained and porphyroblastic varieties is only a few cm in width, but sometimes it is very broad. In some few places the transition from one variety to the other is such that the ground masses have sharp contacts.

In some places, there are remnants of gneisses in the porphyroblastic granodiorite. In the exposure X/AV/61, for example, there are fragments of mica gneiss in the plutonic rock. The fragments are from a few dozen cm to several meters in diameter. The contact against the porphyroblastic granodiorite is sharp. However, the potash feldspar porphyroblasts have grown through the contact surface. The author (A.V.) who has mapped map sheet 3134 regards these potash feldspar insets in granodiorite as autometasomatic in origin.

In some places (*e.g.*, north of Lauritsala) the porphyroblastic granodiorite loses its porphyroblastic character and becomes so rich in potash feldspar as to justify it to be called a potash granite. In some places the gradual increase of potash feldspar proceeds differently, the result being a coarse pegmatitic granite (*e.g.*, at Haikola village, NW of Taipalsaari church).

Usually, the potash feldspar insets are orientated. Completely unorientated varieties have, however, been observed. The potash feldspar insets are sometimes orthoclase, sometimes microcline, though usually having a struc-

Fig. 10. Gradation of a porphyroblastic granodiorite into the even-grained variety (673/AV/60, Fig. 1, area II).

ture between these extremes. The foliation of the rock does not seem to bear any correlation to the triclinicity of potash feldspar insets.

The triclinicities determined by the X-ray powder method are given in Table 6. The regional distribution of monoclinic potash feldspar indicated at the beginning of the work that the orthoclases were a result of the thermometamorphism caused by the rapakivi granite (postkinematic). The results of the investigations of Matisto (1962) on the triclinicities of potash feldspars of the porphyritic granodiorites in the surroundings of Tampere (SW Finland), in an area where no postkinematic granites have been encountered, as well as in the localities of Central Finland described in the foregoing pages, suggest that it is not necessary to take into consideration the influence of rapakivi granites. In addition, the regular occurrence of almost maximum microcline in potash feldspar pegmatite dikes, older than the rapakivi itself, quite near the contact zone of rapakivi granite indicates that the thermal influence of the rapakivi granites on the triclinicity of potash feldspar has been negligible.

Nine specimes of potash feldspar insets have been investigated optically in detail. The results are given in the chapter dealing with the optical properties of the potash feldspars (Table 11).

Table 6. The triclinicity values for potash feldspar from acid plutonic rocks of map sheet 3134 (Lappeenranta), SE-Finland. Most of the specimens investigated are from porphyroblastic granodiorites; the exceptions are mentioned in the column of Remarks.

Sample	Locality	Triclinicity	Remarks
308/AV/59	Hyrkkälä	0.0	
138/KV/60	Lyytikkälä	0—0.2	not porphyroblastic; assumed to belong together with the synkinematic porphyroblastic granodiorites
150/KV/60	»		
211 a/AV/59	Leväinen	0.0	
211 b/AV/59	»	0.95	pegmatitic portion
355/AV/59	Vitsainen	0.0	
384/AV/59	»	0.0	
33/FP/55	»	0.0	
621/KV/60	Haikola	0.0	
581/KV/60	»	0.87—0.94	three specimens
99/JK/58	Vainikkala	0.0	
149/JK/58	Toijala	0.30	
205/JK/58	Nieminen	0.30	very coarse-grained
208/JK/58	»	0.95	some of the insets ovoidal and surrouned by a plagioclase rim
73/AV/59	Summinkisaari	0.0	
73 b/AV/59	»	0.0	
113/AV/59	Taipale	0.8	even- and medium-grained pink granite; gradual change to a gray granodiorite
357/JK/58	»	0.85	
260/AV/59	Kuivaketvele	0.0	
XII/53/A.N.A.	Jauhiala	0.0	gneissose, very rich in potash feldspar
X/AV/61	Hyvärilä	0.54	a large inclusion in the viborgite rapakivi-granite
628/KV/61	Suuri Jänkäsalo	0.87	
700/KV/61	Lammatti	0.95	
790/AV/60	Ampujala	0.95—0.97	very large insets
661/AV/60	Saikkola	0—0.20	fourteen insets investigated
673/AV/60	Ryösölä	0—0.18	three determinations
1 c/PH/60	Saarniala	0.0	» »
33/PH/60	»	0—0.25	» »
19/PH/60	Mentula	0—0.55	may belong to the rapakivi granite
26/PH/60	»	0—0.38	three determinations
86/PH/60	Partala	0.38	» »
129/PH/60	»	0.0	» »
140/PH/60	Ilottula	0.90—0.90	» »
190/PH/60	Muukkola	0.20—0.38	porphyroblastic potash granite (three determinations)

In addition to the foregoing determinations, the $NaAlSi_3O_8$ content in solid solution in potash feldspar was determined. Three microclines and three orthoclases were selected. Both the X-ray powder methods of Goldsmith and Laves (1961) and that of Tuttle and Bowen (1958) were used. The results are given in Table 7. These are in agreement with the data of Goldsmith and Laves (*op. cit.*) showing the average $NaAlSi_3O_8$ content in orthoclase to be somewhat higher than that in microcline.

Table 7. Solid solution of $NaAlSi_3O_8$ in potash feldspar of the porphyroblastic granodiorites, map sheet of Lappeenranta (3134), Southeastern Finland.

Sample	Locality	Triclinicity	$NaAlSi_3O_8$ in solid solution in potash feldspar
790/AV/60	Ampujala	0.95—0.97	3 wt. % (G &L) [1]
140/PH/60	Ilottula	0.90—0.95	4 (G &L)
581 b/KV/60	Haikola	0.94	4 (G &L)
129/PH/60	Partala	0.0	6 (G &L) 4 (T &B) [2]
73 b/AV/59	Summinkisaari	0.0	7 (G &L) 7 (T &B)
661/AV/60	Saikkola	0.0 —0.20	7 (T &B)

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[1] (G & L) refers to method of Goldsmith and Laves, 1961.
[2] (T & B) » » » » Tuttle and Bowen, 1958.

OTHER PARTS OF FINLAND

To obtain a still more widely dispersed network of observations, we include also the results of the triclinicity determinations performed by Dr. O. Kouvo (see p. 56), arranged in three tables (8 to 10) in such a way that

Table 8. The triclinicities of potash feldspars extracted from acid plutonic rocks from various localities. Determined by O. Kouvo.

Locality	Rock type	Triclinicity	Remarks
Siuro, Linnanvuori	porph. granodiorite:		
	porphyroblasts	1.0	
	matrix	1.0	
Tiirismaa	granodiorite, gneiss	0.85	
Varmala	granodiorite	1.00	
Aulanko	»	0.9	
Mikkeli parish, Uusi Kola.	porphyroblasts	1.0	
	matrix	1.0	
Rantasalmi at the railway station of Hiltula	even-grained granite	1.0	
Suomussalmi	granodiorite	0.85	
Savia	gneiss granite	1.0	
Outokumpu	pegmatite granite	0.75	
Murtomäki, between Kajaani and Iisalmi	» »	1.0	
Aavasaksa	granite	1.0	
Kemi, Sompujärvi, road to Alaniemi	»	1.0	possibly late kinematic
Varala	pegmatite	1.0	
Inari, Kyyneljärvi	granite	1.0	late kinematic
Eräjärvi	pegmatite	0.85	
Pyhäsalmi mine	microcline pegmatite	1.0	
Pielavesi, Vaaraslahti	pyroxene granite	1.0	may be nonhomogeneous
Sotkuma	granite gneiss	0.75	two phases
33 km N of Mikkeli	porphyrobl. granite	0.9	may be augen gneiss

Table 9. Triclinicities of potash feldspar extracted from various rock types. Determined by O. Kouvo.

Locality	Rock type	Triclinicity	Remarks
Ylöjärvi, E of Mastosjärvi	trachyandesite	0.75	traces of orthoclase
Kalkku	acid porphyry	0.9	
Aitolahti, 21 km of Tampere	» »	0.75	traces of orthoclase
Kangasala, Mutikko	very coarse graywacke	1.0	
N part of the gulf of Bothnia	feldspar porphyry	1.0	
Kaavi, Rouvantaipale	veins in mica schist	0.75	
Sompujärvi	granite pebble of conglomerate	1.0	
Lemi, Matalaluoto	rapakivi pegmatite	0.75	

Table 10. Triclinicity of orthoclasic potash feldspars extracted from various rock types. Determined by O. Kouvo.

Locality	Rock type	Triclinicity	Remarks
Nokia, Vihola	graywacke, sandy layer	orthoclase	traces of microclinization
Lemi, Varto, 15 km from Lappeenranta	charnockitic rapakivi	»	both insets and matrix
»	pegmatite dyke	»	
Suvasvesi	granite	»	both insets and matrix
Heinävesi	»	»	both insets and matrix
Kuusjärvi, Rikkaranta	pegmatite granite	»	
Kaavinkoski	even-grained granite	0.25	
Korsnäs	red feldspar	0.65,	pegmatite
»	gray	orthoclase	»
Ilomantsi, church	granite	orthoclase + 0.65	
Kuopio, Vuorimäki	granite gneiss	0.4	
Kaavi	granite	0.4	
Eurajoki	rapakivi	orthoclase	
Pyhtää, Saunasaari	ovoids of dark rapakivi	»	
Inari, Kenttäsaaret	quartz diorite	»	
Heinävaara	granite gneiss	0.65	

Table 8 contains acid plutonic rocks with microcline, Table 9 other rocks with microcline and Table 10 rocks containing potash feldspars of inferior triclinicity.

In general, these tables clearly show that orthoclase is uncommon and uncharacteristic outside the areas considered in this paper. On the other hand, the results given in Table 10 give much support to the possibility that still more areas could be found which may be characterized by potash feldspars of inferior triclinicity to much larger extent than suggested by the present results.

Actually, in Table 10, only the following occurrences of orthoclase are those not belonging to the areas either described above or already earlier known to contain orthoclase (rapakivi): Suvasvesi, Heinävesi, Kuusjärvi, Ilomantsi, Kuopio, Kaavi. They all are situated in eastern Finland. Thus, such an »orthoclase area» may well be expected to exist there.

OPTICAL STUDY

A REVIEW ON THE OPTICAL PROPERTIES OF POTASH FELDSPAR

The optical properties of potash feldspars are dependent on the composition, submicroscopically exsolved phases, submicroscopic twinning, and the degree of the Al/Si order.

The influence of chemical composition on optical properties has been treated by Winchell 1925 (K-Na feldspars and K-Ba feldspars), Spencer 1937 and Tuttle 1952 (K-Na feldspars), and Oftedahl 1948 (K-Na-Ca feldspars). Hewlett (1959) studied the influence of Ca, Ba, Sr, and Fe on the optical properties of potash feldspar, emphasizing that the substitution of Na for K causes such a small change in the index of refraction in the range Or_{100} to Or_{60} that its effects normally are masked by the larger change caused by the substitution of minor amounts of Ca, Sr, Ba and Fe. Microcline-microperthite usually contains in natural occurrences less than 5 wt. % $NaAlSi_3O_8$ in solid solution in the potash feldspar phase (Goldsmith and Laves 1961). The influence of this substitution on 2V of microcline can be omitted. In orthoclase-microperthites the corresponding $NaAlSi_3O_8$ content is of the magnitude of 8—10 wt. % (Goldsmith and Laves, *op. cit.*). Tuttle's diagram (1952) shows the axial angle to change a few degrees over the composition range Or_{100}—Or_{90}.

In the light of present knowledge, the influence of the cryptoperthitic exsolutions on the optical properties of potash feldspar is small. During homogenization of cryptoperthites there occur very small changes in the indices of refraction, and the changes in 2V are almost negligible, occasionally possibly some few degrees.

Marfunin (1961, 1962) has shown that submicroscopically twinned triclinic potash feldspar has an axial angle smaller by 2 to 3 degrees than the single crystal potash feldspar with the same degree of order in Al/Si. If the twinning is balanced, the average structure has the axial plane perpendicular to (010), and the optical orientation is monoclinic. Depending upon the nature of the submicroscopic twinning (balanced or unbalanced), the extinction angle $\gamma \wedge \perp$ (010) varies from 0° to 18° for complete order in Al/Si in potash feldspar; at the same time 2Vα remains almost constant (ca. 84°).

The optical orientation and the values of the optical axial angle are the most sensible criteria for the evalution of the degree of Al/Si order of potash feldspars. During sanidinization, 2Vα of microcline changes from 84° (optical axial plane nearly perpendicular to (010)) to 0°. Then the axial angle again opens to 60°, the axial plane becoming parallel to (010).

In the light of the afore-mentioned statements, it is possible to determine with the aid of 2Vα the structural state of potash feldspars provided that the composition of the potash feldspar is approximately known. In this connection it is appropriate to cite the statement of Marfunin (1961): »There are two types of order disorder in potash feldspar: 1) disorder without deviation from monoclinic symmetry (submicroscopically and/or sub-x-ray twinned; balanced twinning) and 2) disorder with change of symmetry into a triclinic one (single crystal). In addition there are all intermediate states between disorder and order of both types (unbalanced twinning). Every state of substitution disorder may be sub-X-ray twinned.» Thus the triclinicity obtained from the X-ray powder pattern value of Goldsmith and Laves (1954) would not be sufficient for the determination of the structural state of the sub-X-ray twinned potash feldspar.

RESULTS OF THE OPTICAL DETERMINATIONS OF THIS STUDY

Twenty of the potash feldspars, which are listed in Tables 2, 4, 5 and 6, were selected for optical study. All are supposed to be representatives of synkinematic porphyroblastic granodiorites or to belong genetically together with them. In addition, the specimen of the only late kinematic potash granite of southern Finland containing orthoclase found to date is included (Härme 1962). According to the X-ray powder diagram, the potash feldspar in the specimens selected is mainly orthoclase. In addition, however, insets of potash feldspar with varying triclinicities were found in these specimens.

Under the microscope, all the potash feldspars studied contain areas without cross-hatching. Such areas change gradually into areas with cross-hatched microcline. The optical investigation reveals that the areas with cross-hatching are much more frequent than could be expected from the triclinicity values obtained by the X-ray powder study. On the other hand, many insets without cross-hatching were found that possess the optics of maximum microcline.

With the U-stage, only those areas of potash feldspar were studied that do not have any cross-hatching. The optical axial angle 2Vα and the extinction angle ($\gamma \wedge \perp$ (010)) for the potash feldspar in synkinematic grano-

Table 11. Optical data of the potash feldspars of synkinematic porphyroblastic granodiorites in Finland.

	Sample	$2V\alpha$	$\gamma \wedge \perp (010)$
CENTRAL FINLAND	26/KH/62	80°	20°
	30/KH/62	80°	18°
	31/KH/62	64°—84°	0°—15°
	32 b/KH/62	80°	16°
	33/KH/62	88°	18°
	34/KH/62	60°—70°	0°—8°
	37/KH/62	75°—85°	—
	19/MT/62	81°	17°
	355/MT/62	64°—73°	18°
	371/MT/62	76°	14°
	50/LP/61	70°—80°	12°
SOUTHEASTERN FINLAND	73/AV/59	60°—68°	0°
	260/AV/59	44°	0°
	384/AV/59	68°	7°
	661/AV/60	—	0°—10°
	138/KV/60 1)	80°	0°
	649/KV/61	76°—86°	18°
	129/PH/60	62°—65°	0°—5°
	1 c/PH/60	65°	0°
	208/JK/58	74°	15°

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1) Even-grained; supposed to belong genetically together with the synkinematic porphyroblastic granodiorites.

diorites are given in Table 11. When any variation in the axial angle occurs, the limits are given. In some places, very weak lamellae have been met with, the adjacent lamellae showing slightly different axial angles. In the cases, when the microscopical and X-ray powder investigation were performed on the same inset, the data obtained are in good agreement with each other. When the two data were from separate insets, even though from the same specimen, the correlation is more variable.

The axial angles vary from —44° to that characteristic of a maximum microcline. Orthoclase in specimen 260/AV/59 has $2V\alpha = 44°$. The locality from which this specimen is derived is situated very near the rapakivi granite contact (some few meters). It is well known that the major portion of the potash feldspar contained in rapakivi granites is orthoclase. The thermometamorphic action caused by the rapakivi granite might be responsible for the very low 2V value of this specimen. On the other hand, one of the authors (A.V.) has investigated optically potash feldspar phenocrysts from rapakivi granites but has never found $2V\alpha$ values as low as 44°.

Specimen 138/KV/59 representents an even-grained potash granite which may belong genetically together with the synkinematic porphyroblastic granodiorites. The feldspar is orthoclase-microperthite: both the extinction angle and the triclinicity determined by the X-ray powder method indicate orthoclase. The value of the axial angle, $2V\alpha = 80°$, however, could be explained if the potash feldspar is assumed to be a submicroscopically and sub-X-ray twinned microcline. All the potash feldspars listed in Table 11, with $2V\alpha$ in the neighbourhood of 60—65° and a straight extinction, might be submicroscopically twinned too, the individual lamellae having structural states between an orthoclase and a maximum microcline.

The optics of the potash feldspar of the orthoclase-bearing latekinematic granite (Härme 1962) is highly variable. The axial angle in the untwinned homogeneous parts of the potash feldspar varies from 63° to 76°. Cross-hatched areas also occur.

CONCLUSIONS

In the present paper the classification of the potash feldspars is mainly based on X-ray triclinicity determinations (method of Goldsmith and Laves), and accordingly the term »orthoclase» for feldspars of inferior triclinicity (0.4 or less) is mainly used.

The list of the triclinicity determinations indicates that orthoclase occurs surprisingly widely within the areas investigated in Central and Southeast Finland, though confined to comparatively well-defined, limited areas. One such area possibly exists also in Eastern Finland.

Further scrutiny of the problem, especially its mineralogical aspects, revealed, however, some new points of view, which made such a conclusion to a certain extent doubtful. This uncertainty is due to the possibility that the disorder determined by X-rays may either indicate a change in symmetry, or it is only the apparent one being caused by the sub-X-ray twinning. In the latter case we could still have a triclinic microcline despite the low triclinicity deduced from the X-ray work. In other words, the distribution of low values obtained for triclinicities might indicate the variations in apparent triclinicities — not, however, that there would necessarily always be a deviation from well-ordered potash feldspars or the presence of monoclinic orthoclases.

Therefore we have to assume that, if the conditions prevailing during the formation of the rocks are to be parallellized with the form of the potash feldspars occurring in them, at least the following two alternatives must be considered:

a) the potash feldspar is either monoclinic or triclinic with all the intermediates;

b) the potash feldspar is microcline of sub-microscopic and sub-X-ray twinning which, in earlier investigations, may not always have been detected or distinguished from the truly monoclinic potash feldspars, but which likewise may give low and variable values for the triclinicities.

Case (a) has been earlier much discussed, and also used for interpretation of the rock formations. Case (b), on the other hand, is in this sense quite new territory (Marfunin 1962), and so far does not give us any hints for petrogenetic interpretations.

Concerning the cases discussed in the present paper, it seems that the monoclinic-triclinic case (a) is of importance, and that the monoclinic phase, which occurs mainly in forming the porphyroblasts, may be the result of the mode of the growth of the porphyroblasts, which did not allow the full ordering of the lattice of potash feldspar. This also explains the ample presence of structurally intermediate potash feldspars.

But there is also evidence to indicate that case (b) is also involved. It seems, furthermore, that the porphyroblasts are also apt to undergo the balanced, sub-X-ray twinning.

We are not yet, however, able to say or predict what the petrological conditions causing the balanced or unbalanced sub-X-ray twinning might be. Perhaps they are close to those affecting the Al-Si ordering of the potash feldspars, but there may also be some additional factors determining whether a sub-X-ray twinned microcline, or a potash feldspar of interior triclinicity will result.

In the opinion of the present authors, a matter has here been encountered that may be of great petrogenetic importance, especially if the problems of the origin of granite and of the formation of other acid plutonic rocks are involved. It is well understood that the results so far obtained are too scanty and thus insufficient for generalizing, but still they do indicate how urgent and important is the thorough study of potash feldspars, not only from the mineralogic, and crystallographic, but also from the petrologic point of view.

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STROMATOLITE STRUCTURES IN PRECAMBRIAN DOLOMITE IN TERVOLA, NORTH FINLAND [1]

BY

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ABSTRACT

A preliminary report is given on the stromatolite structures found in Precambrian Karelidic dolomite.

In his paper »On the Geology of Fennoscandia» Sederholm (1932, p. 15; *cf.* Hausen 1936) observes that in the dolomitic limestones of the Jatulian area along the Kemi River in northern Finland »stromatolitic structures occur which may be of algal origin». This short communication contained no further information about the locality. During recent years a number of other reports on stromatolite structures in the Karelidic belt have been made. Ödman (1957, p. 54; *cf.* Geijer 1931, p. 173) reported on the occurrence of stromatolite structures in the dolomite found in the archipelago of Kalix in northern Sweden. Somewhat later the Russian geologists (Соколов, В. А. 1959; Слодкевич, В. С., Соколов, В. А. и Бутин, Р. В. 1960; Соколов, В. А. и Бутин, Р. В. 1961) described similar structures found in Soviet Karelia.

In recent years a systematic search for stromatolite structures was conducted in Finland until, on the basis of Sederholm's observation (*op. cit.*), occurrences were found (in September, 1963) in the dolomite on the shore of the Kemi river, near the Ollitervo farm in Tervola parish, northern Finland (Figs. 1 and 2, Plate I, and Figs. 3 and 4, Plate II). There the dolomite contains intercalations of quartzite (*cf.* Mikkola 1949, p. 16).

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[1] Received October 14, 1963.

The stratigraphy of the area of supracrustal rocks south of Tervola is described by Härme (1949), who notes that the lowermost member in the sequence of these rocks is quartzite. It is overlain by basic volcanics which gradually pass over upwards into slates. The quartzite is current bedded, and the stratified volcanics as well as the slates exhibit graded bedding. The dolomite occurs as intercalations partly in the upper section of the quartzite and mainly in the transition zone between the stratified volcanics and the slates. The occurrence of the stromatolite structures in the dolomite gives an idea of the conditions that prevailed during the sedimentation of the Karelidic dolomites.

The age of the supracrustal formation of that area is an open question, for no radiometric age determinations of the supracrustal rocks exist. The age of the biotite contained in an intersecting granodiorite occurring in the neighborhood (Nosa, Alatornio parish) has been determined (Wetherill, Kouvo, Tilton and Gast 1962) by the Rb-Sr-method, and the figure is 1780 millions of years.

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EXPLANATION TO PLATES

PLATE I

Fig. 1. Stromatolite structures in the stratified dolomite, section across the bedding. Ollitervo, Tervola. 1/5 natural size. Photo: V. Perttunen.

Fig. 2. Stromatolite structures in the stratified dolomite, section across the bedding. Ollitervo, Tervola. 1/2 natural size. Photo: V. Perttunen.

PLATE II

Fig. 3. Stromatolite structures in the stratified dolomite, section across the bedding. Ollitervo, Tervola. 1/3 natural size. Photo: V. Perttunen.

Fig. 4. Stromatolite structures in the stratified dolomite, the weathered surface of a bed. Ollitervo, Tervola. Natural size. Photo: V. Perttunen.

PLATE I

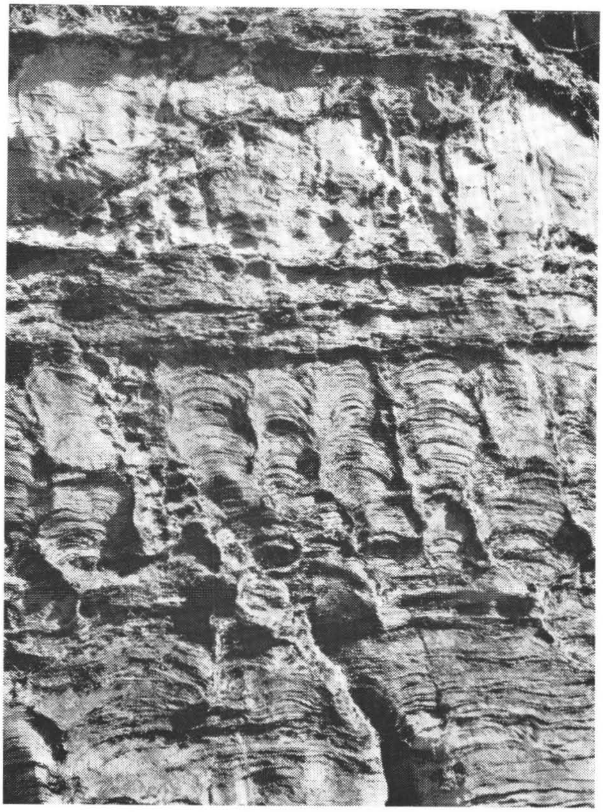

Fig. 1

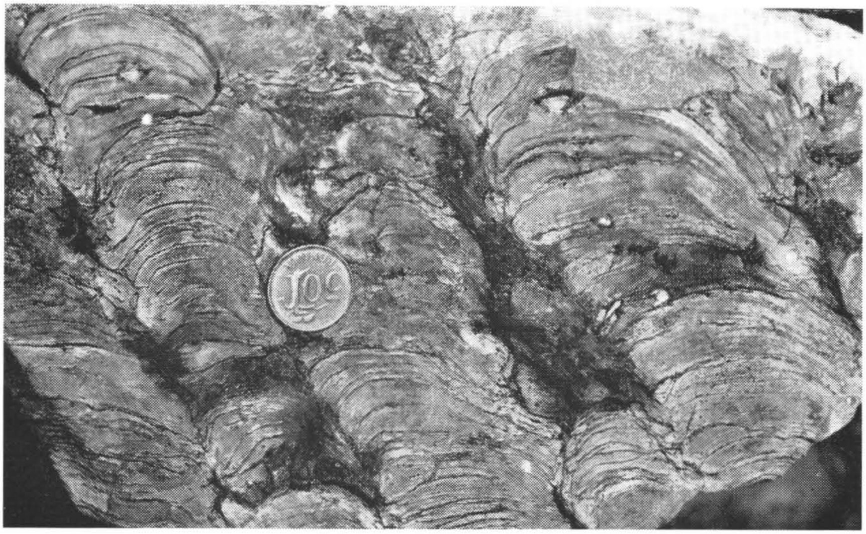

Fig. 2

Maunu Härme and Vesa Perttunen: Stromatolite Structures ...

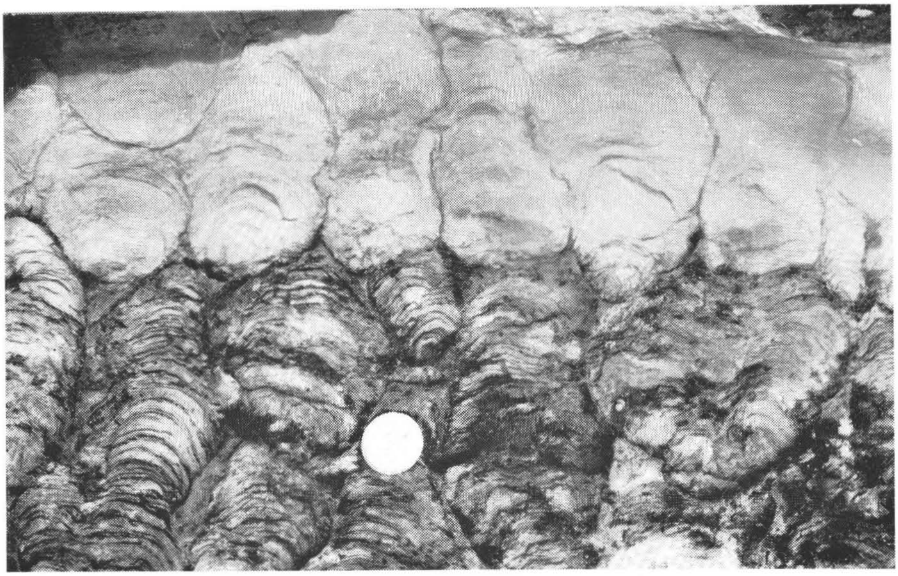

Fig. 3

Fig. 4

Maunu Härme and Vesa Perttunen: Stromtolite structures . . .

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THE HYPERSTHENE GRANITES AND UNAKITES OF CENTRAL FINLAND [1]

BY

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CONTENTS

INTRODUCTION

In 1916 Eero Mäkinen published a study on the Precambrian rocks of Central Ostrobothnia in western Finland which was important in many respects. From this region Mäkinen describes a great number of granites, granodiorites, quartz diorites, diorites, pyroxene diorites, gabbros, norites

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[1] Received September 25, 1963.

and even olivine-bearing norites associated with them, and all containing hypersthene. Such rocks occurring in Central Finland were later described by Hackman (1931) in his report on the rocks of the Nyslott section D2 of the geologic map of Finland, and by Wilkman (1931, 1938) in the descriptions of the rocks of the sections Kajaani C4 and Kuopio C3. Wilkman points out that these hypersthene granites gradually pass over into granites, as well as into a series of granodiorites, diorites and norites containing both monoclinic pyroxenes and hypersthene, and thus form transitional types in a rock series between granites, pyroxene diorites, quartz gabbros and norites. Mäkinen had observed similar relations with regard to the Ostrobothnian hypersthene rocks, the granites there and the more basic diorites.

Hackman and Wilkman describe these hypersthene rocks as hypidiomorphic, granular, rather coarse-grained rocks of a peculiar brown colour: greyish-brown, greenish-brown to reddish-brown, which is highly characteristic of this variety but unusual for granite rocks in general. In his descriptions Hackman called these rocks »gabbro-granites», which had been the field name used during the survey of these peculiar brownish granites.

GEOGRAPHICAL DISTRIBUTION OF THE HYPERSTHENE ROCKS IN CENTRAL FINLAND

Wilkman, in his descriptions of the rocks included in the map sheet Kajaani C4 (1931) and Kuopio C3 (1938), has described a great number of occurrences of hypersthene granites and of pyroxene granodiorites, pyroxene-quartz diorites and pyroxene diorites associated with them. According to the publications cited, about sixty occurrences of these brownish hypersthene rocks are known from Central Finland. They all occur NE of a line drawn on the map from the town of Wiborg (Viipuri) in the southeast to the town of Piteå in the northwest, on the Swedish side of the Gulf of Bothnia. Again, none of these rocks are to be found NE of a line drawn from the town of Sortavala, on the north coast of Lake Ladoga, to the town of Uleåborg (Oulu) on the northeastern coast of the Bothnian Gulf.

The hypersthene granites and related dioritic rocks thus occur in a strip or belt between the two aforesaid lines running NW-SE, which strip is about 100 km broad and has a length of about 450 km.

It may seem that a crack zone of about 450 km in length and about 100 km in breadth represents a rather extreme case. I should therefore mention, as a basis for comparison, that the fault line of St. Andreas, along which the movement of the San Francisco earthquake of 1906 took place, is 600 km long, and that the destructive work of that earthquake could be followed along a line of more than 400 km (Wahl 1917).

Outside this strip, hypersthene granodiorites have been described by Hackman (1923) from the islands of Kakskerta and Kustö, about 15 km southward from Åbo (Turku) in SW Finland (»charnockite»). Further, Laitakari (1942) has described pyroxene diorite containing hypersthene from Lill-Kyro (Vähäkyrö), east of Vasa. Parras (1958) has described a fairly large district of charnockitic rocks from the northwestern part of the province of Nyland (Uusimaa), the centre of this district being about 50 km NW of Helsingfors (Helsinki). The occurrence of hypersthene in trondhjemite NE of Åbo has also been mentioned by Huhma (1959).

From Ojanperä in the parish of Paavola (Raahe), Nykänen (1959) has recently described a small district of coarse-grained granodiorite containing hypersthene. This occurrence is inside the 100 × 450 km strip of rocks dealt with here.

As we have seen, there are in Finland outside the belt of hypersthene rocks mentioned only four occurrences of hypersthene rocks. The principal ones are the charnockite and paracharnockite districts described by Parras (1958), the charnockite district described by Hackman (1923) and the two other small occurrences referred to here.

In contrast, inside the belt, from Ladoga to the northern part of the Bothnian Gulf, there are about 60 localities where hypersthene rocks of the peculiar brown colour occur. The thin section collection of the Geological Survey of Finland contains about 70 sections of these rocks, and six analyses of the hypersthene rocks have been made. It is also noteworthy that several of these occurrences are rock bodies of considerable size: thus the occurrence NW of Savonlinna is about 5 × 20 km, the occurrence of Rautalampi about 6 × 25 km, the occurrence of Konnevesi about 10 × 8 km, the occurrence of Viitasaari about 4 × 12 km and the occurrence of Kiuruvesi about 8 × 17 km, and there is further the large occurrence in southern Kangaslampi, which continues southward across the islands between Äänisvesi and Haukivesi, being mostly covered by the lakes.

The rocks of the large district of hypersthene rocks in southern Nyland described by Parras (1958) and mentioned above are, however, probably not original igneous rocks, but have through later transformations changed to hypersthene rocks, that is »charnockites». The hypersthene rocks of Central Finland are, as clearly indicated by many circumstances with regard to their mode of eruption, their contacts and the apophyses and dykes they send into surrounding rocks, to be considered as primarily igneous rocks. They have thus nothing in common with the »charnockites» with regard to origin, and they are therefore not comparable to what are now generally considered as »charnockites», which recently have been dealt with by most authors as secondary altered rocks. I refer in this respect to the publications by Quensel (1952) and Eskola (Barth, Correns, Eskola 1939). I have therefore in this publication avoided calling the hypersthene granites of Central Finland »charnockites» and applied to these rocks, which I consider to be hypersthene-bearing, originally igneous rocks, the names »hypersthene granites» and »hypersthene diorites».

THE OCCURRENCE OF THE HYPERSTHENE ROCKS ALONG SETS OF STRAIGHT LINES

Some time ago the present author noticed that a great number of the occurrences of the hypersthene rocks of this belt in Central Finland run along straight lines. There are two sets of such lines: a southern one running nearly 45° NW—45° SE, and a northern one running about 25° NW—25° SE. The lines and the principal occurrences of hypersthene rocks are marked on the accompanying map (Fig. 1).

Lines running 45° NW. There are two of these: No. I, a short one, to which belong occurrences (mentioned from SE-NW) at Konnevesi, Sumiainen and Konginkangas; and No. II, which includes some occurrences at Simpele, six occurrences between Savonlinna and southern Kangaslampi, at Rautalampi (southern Vesanto) and at Viitasaari. In all there are on this line eleven occurrences, which, however, partly lie about 5 km on either side of the principal line. We may thus also consider them as belonging to three distinct parallel lines only five km apart. Line II is about 230 km long.

The lines running 25° NW are five in number. No. III, the southernmost and longest of all these lines, extending about 430 km, starts from the SE at Taipalsaari and passes through the occurrences at SW Konnevesi, running NE from Reisjärvi to Vasankari on the shore of the Bothnian Gulf, about 10 km to the NE of the mouth of Kalajoki (river). No. IV is a line about 100 km long. It starts at Kuuslahti, about 10 km SW of Vesanto, and passes through the occurrence at Viitasaari to an occurrence about 12 km south of Haapajärvi church. No. V is 180 km long. It starts at Keitele and passes through the occurrence of Sarjakylä, Haapajärvi (in Haapavesi), Vihanti and Pattijoki. No. VI passes through eight occurrences and is 140 km long, starting in the SE at Väärälahti and passing through the big occurrences of southern Kiuruvesi, the occurrences south of Piippola church and north of Pulkkila and between two occurrences south of Rantsila church, which lie only 5 km apart. In addition, there is an occurrence about 12 km south of the town of Iisalmi, which lies 6 km NE of the line and one between Piippola and Haapavesi about 5 km SW of line VI. No. VII: The occurrences lying farthest to the NE are the two large ones of Sonkajärvi and Varpaisjärvi. The line between them extends only 25 km,

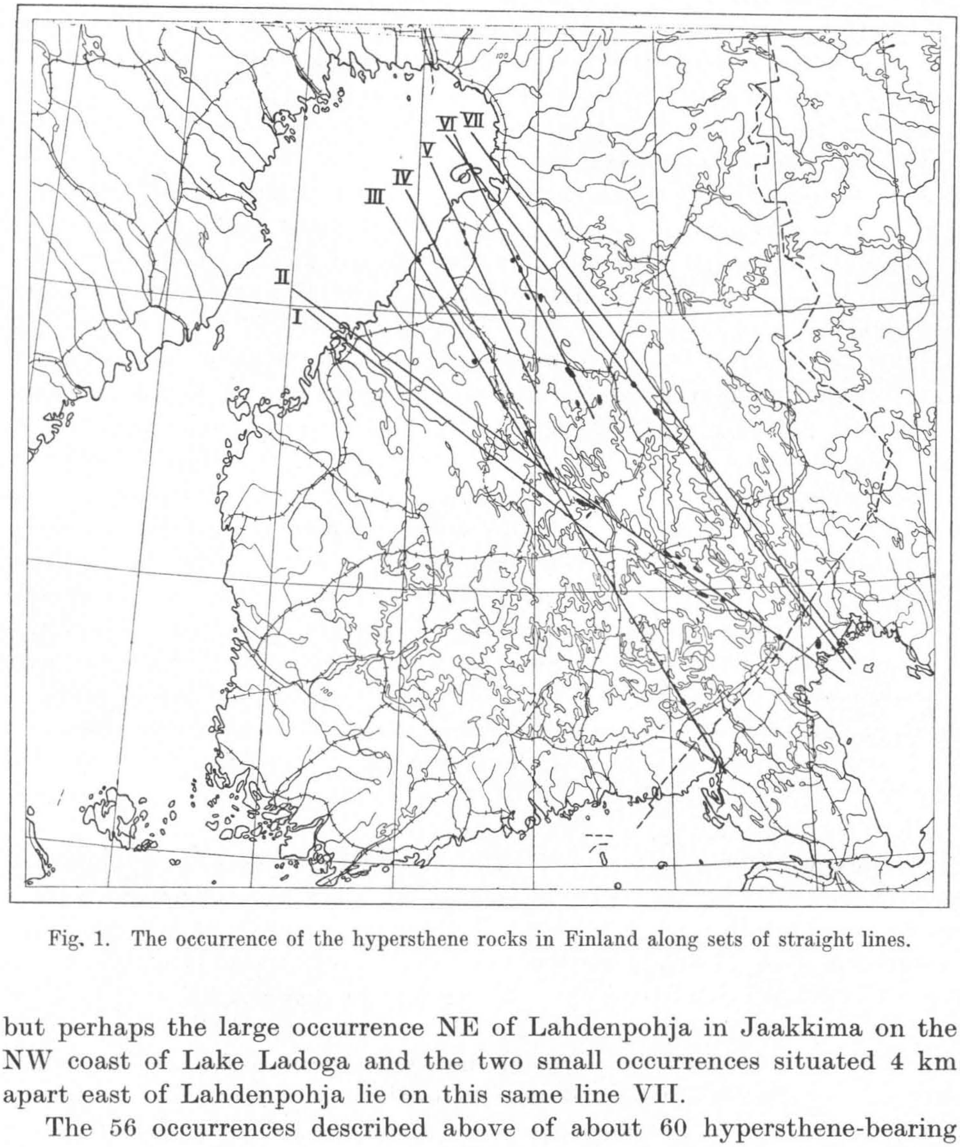

Fig. 1. The occurrence of the hypersthene rocks in Finland along sets of straight lines.

but perhaps the large occurrence NE of Lahdenpohja in Jaakkima on the NW coast of Lake Ladoga and the two small occurrences situated 4 km apart east of Lahdenpohja lie on this same line VII.

The 56 occurrences described above of about 60 hypersthene-bearing rocks lie on or close to two sets of lines running in the directions 45° NW and 25° NW and of which we can follow No. III for a length of about 430 km, No. II for a length of about 230 km, No. V for a length of about 180 km, No. VI for a length of about 140 km and No. IV for a about 100 km. Nos. I and VII are short.

This distribution of the hypersthene rocks along crack-lines inside a comparatively narrow belt about 100 km wide and about 430 km long is of considerable geological interest. It shows that:

1) The hypersthene rocks have erupted along two systems of crack-lines and have consolidated and crystallized at a depth sufficient to allow a coarse-grained hypidiomorphic structure to develop.

2) It is probable that all the hypersthene rocks, the hypersthene granites as well as hypersthene-quartz diorites, diorites and gabbros originally emerged from a common magma, which must have been differentiated before the cracks were formed and the magmas of the present rocks rose in them. Perhaps the process of differentiation resulted in a gain in total volume and a rise in pressure, which in turn caused the production of the cracks in the crack belt.

3) The cracks in this zone may have reached the then surface of the earth, allowing volatiles and water from the magma to escape in large quantities. Thus the remaining magma became poor in water and volatiles, and less mica and hornblende could crystallize from it. Hypersthene was formed instead and the rocks crystallized at a higher temperature than would have been the case had biotite and hornblende rocks been produced.

4) In all the cracks of the zone the magma probably rose and crystallized during a comparatively short time. There are, however, cases where one magma of the zone has penetrated another as a dyke. The rise and crystallization of the magma must therefore have lasted long enough for part of the rocks to solidify to the extent of being able to crack and let portions of the magma form dykes in the portions of rock that crystallized earlier.

It is difficult to say which of the two crack systems, that of 45° NW or that of 25° NW, is older than the other, since all the rock types were intruded in all the crack lines.

5) Meanwhile, we can conclude with certainty that before the hypersthene rocks were produced and crystallized and the cracking of the earth's crust took place, all the gneiss-granites, pre-Bothnian granites and post-Bothnian granites must have been in place as they are at present, and that the age of the hypersthene rocks is younger than that of the post-Bothnian granite.

At Taipalsaari (and at Simpele) the hypersthene rocks are in contact with the rapakivi granite, which penetrates them with apophyses and which has caused contact metamorphism in them. The hypersthene rocks are therefore older than the rapakivi.

6) As much as can now be seen from the distribution of the rocks, the NE borderline of the strip of hypersthene rocks is not crossed by Kalevian or Jatulian rocks, and thus it also seems to form the SW borderline of the Karelian formations. This indicates that the hypersthene rocks along the crack zone are older than the Karelidic rocks.

Moreover, no hypersthene rocks are to be found inside the Karelidic zones.

However, it could also be assumed that during the formation of the Karelidic mountain zone or soon after, the Karelidic earth movements gave rise to the formation of the crack zones westward of them, which were filled with the magmas of the hypersthene rocks.

7) An age determination of two of the hypersthene rocks, pyroxene-granite from Vaarslahti, NE of Pielavesi, and pyroxene granite from Petäjävesi, has recently been published by Wetherill, Kouvo, Tilton and Gast (1962), according to which the age of the hypersthene rocks is 1730—1760 million years. The age of the Svecofennides is given as 1 800 million years.

In an earlier publication by Kouvo (1958) the age of the rapakivi has been given as 1 620 ± 20 million years.

The age of the Karelides is given by Wetherill, Kouvo, Tilton and Gast (1962) as 1 800 million years.

Earlier all Finnish geologists considered the age of the Karelidic movement as less than that of the Svecofennian movement. On account of the age determinations which give about the same age for the Svecofennides and the Karelides, some authors have considered both the Svecofennian and the Karelian mountain chains as belonging to the same period of mountain formation; the Kalevian, Jatulian, and Onegian formations (the whole of the Karelian) would then appear to correspond to a Flysch-period of the Svecofennian.

This is not the place for a discussion of these important age problems. Here I wish only to point out that the hypersthene rocks are about 100 million years younger than the principal part of the remaining rocks of the Svecofennidian movement, and that there are indications that they are older than the Karelidic movement.

Also it must be remembered that an uncertainty of 10 to 15 per cent in the age determination of rocks as old as the Svecofennian would mean an uncertainty of about 180—270 million years, which is a longer period of time than that required by some great mountain chains to form.

It therefore seems premature to state, considering of present methods of making age determinations, that there had not been a period of 200 to 300 million years between the formation of the Svecofennian and the Karelian mountain chains and that the belt of the hypersthene rocks was not formed during this interval.

Besides the hypersthene rocks dealt with above, there occur in Central Finland two other series of igneous rocks from the same period, a series of »unakites» and a series of aplitic granites, which are described in the following.

THE OCCURRENCE OF UNAKITE IN CENTRAL FINLAND

In his descriptions accompanying the map sheets of Kajaani and Kuopio Wilkman (1931, 1938) has described a great number of occurrences of these peculiar albite-orthoclase and epidote-chlorite rocks, which he called »unakites,» since such rocks had earlier been described by Bradley (1874). Mäkinen had earlier described similar rocks from the Ostrobothnian coastal district, calling them »natron syenites».

The name »unakite» was given by Bradley in 1874 to granitic rocks containing orthoclase, epidote, quartz and magnetite, from the Great Smoky Mountains of the Unaka Range of the Blue Ridge between North Carolina and Tennessee. He considered the rocks to be secondary and formed from mica granite or hornblende syenite. Later Phahlen (1904) described similar rocks from Milam's gap on the summit of Blue Ridge, and mentioned as constituents red feldspar and green epidote; quartz, iron oxide, zircon and apatite are mentioned as other constituents. Phahlen (1904) likewise regarded these rocks as secondary after »hypersthene akerite» (probably corresponding to the »hypersthene granodiorites» of this paper).

Similar rocks were also described by Laitakari (1918), who called them »helsinkites», as well as by Asklund (1925) and by von Eckermann (1925), who also applied the name helsinkite to occurrences in Sweden. Later, however, all the Finnish authors who have written about these rocks have called them unakites, considering them to be primary, deep-seated igneous rocks.

In describing the rocks of the Kajaani sheet, Wilkman (1931) gives the following description of the unakites (original text in Swedish): »The principal constituents of this curious type of rock are albite, iron-rich epidote and green chlorite (pennine). Besides these minerals the rock mostly contains microcline, microcline perthite and quartz in varying quantities, and as accessory components magnetite, titanite, ilmenite, apatite and hematite. Common hornblende and biotite do, however, not occur in typical unakites.»

THE CONDITIONS OF CRYSTALLIZATION OF THE UNAKITES, AND VARIOUS OBSERVATIONS ON CERTAIN OCCURRENCES

Wilkman (1931) writes (original text in Swedish): »On account of the content of albite and the minerals epidote and chlorite, which contain water, it is generally believed that these rocks have crystallized from magmas rich in water at a comparatively low temperature (possibly below 440°) and at a high pressure.»

From Isojärvi, in the parish of Oulainen, Wilkman (1931) describeds an unakite porphyry of fluidal structure. The phenocrysts consist of dark red pure albite ($1.5222 < \alpha \leqq 1.526$).

From Ollikkaanvuori, NE Haapavesi, Wilkman describes an unakite (with analysis) which contains slightly more microcline than albite and less than the usual amount of epidote (6 %). The double refraction of the epidote is given as 0.037, corresponding to a Fe_2O_3 content of 27 %.

From Naistenkallio, near Vatjusjärvi in the commune of Haapavesi, he describes a small-grained unakite which is traversed by porphyric unakite veins. This equigranular unakite (analysis) contains 38.7 % albite, 37.7 % epidote and 15 % microcline. The chlorite of these occurrences is green and the quantity is from 1.5 % and 2.4 % to 4.4 %.

Wilkman (1938) describes several unakite occurrences from the Kuopio section characterized by comparatively strong pleochroism of the chlorites (yellowish to grass-green).

Besides these descriptions accompanying the section maps, Wilkman (1928) has published a small monograph on the unakites, in which he deals with occurrences in the North Bothnian coastal districts.

Mäkinen (1916) has, as mentioned, also described the unakites from the district of North Ostrobothnia under the name of natron syenites and kalinatron syenites. He emphasizes that they are hypidiomorphic abyssal rocks and gives two analyses, included in Table 1 shown on p. 92, pointing out that their high sodium and aluminium content depends on the high albite content, and, further, that they contain no anorthite, but that the lime, corresponding to 15—25 % normative anorthite, is not contained in anorthite or plagioclase, instead of which the CaO is almost entirely contained in the epidote.

The present author would in this connection like to point out that the typical sodium unakites are not only free from anorthite but also from biotite, and mostly also from hornblende. All the potash goes into the microcline and no biotite is formed, and all the lime is contained in the epidote, no anorthite being formed. It should also be pointed out that one might expect the formation of anorthite in a magma of the composition of the unakite magma. But in this case no anorthite is formed, while the calcium of the magma, combined with water, enters into a chlorite and/or a serpentine molecule.

As seen from the publications on unakites cited, there are about 30 to 40 unakite occurrences in the NW part of Central Finland. A great many of them lie inside the zone mentioned in the description of the hypersthene granites, but only in the NW part of this belt, most of them in Oulainen, Haapajoki and Haapajärvi, the easternmost occurrences being those situated about 30 km SE of Suonenjoki and about 50 km SW of Kuopio and at

Table 1. Analyses of hypersthene

	I	I a	II	II a	II b	III	III a	III b
SiO_2	67.07	1.1123	69.38	1.151	74.90	68.19	1.137	74.90
TiO_2	1.02	0.0129	0.44	0.006	0.66	0.79	0.010	0.66
Al_2O_3	14.59	0.1428	16.01	0.175	9.35	15.13	0.148	9.75
Fe_2O_3	0.56	0.0035	0.36	0.002	—	1.06	0.0007	—
FeO	5.18	0.0721	2.81	0.039	4.08	3.41	0.047	4.08
MnO	0.06	0.0008	0.01	—	—	0.05	—	—
MgO	1.02	0.0253	1.03	0.026	1.58	0.97	0.024	1.58
CaO	3.99	0.0711	3.34	0.060	2.37	2.00	0.036	2.37
Na_2O	2.10	0.0468	4.66	0.075	3.36	3.17	0.051	3.36
K_2O	2.41	0.0256	1.46	0.016	3.29	4.69	0.050	3.29
P_2O_5	0.21	0.0015	0.21	0.002	—	0.03	—	—
H_2O+	0.76	0.00422	0.52	0.029	—	0.65	—	0.72
H_2O-	0.05	—						
	99.83		100.23		100.00	100.14		100.00
Sp. Grav.	2.718							

An. I	Porphyritic hypersthene granite. Kymönkoski, Viitasaari. Anal. L. Lokka. Map sheet C 3, Kuopio, p. 100.
An. I Mode:	Quartz 27.90, plagioclase (An_{34}) 37.80, microcline 10.94, biotite 9.16, hornblende 3.45, ilmenite 0.91, magnetite 0.70 %,
An. I a	Molecular proportions of I.
An. II	Hypersthene granite. Suurluotonen, Sonkajärvi. Anal. L. Lokka. Map sheet C 4, Kajaani, p. 34.
An. II Mode:	Quartz 2.6 %, albite 39.4, anorthite 16.7, microcline 7.2., hypersthene 5.6 ($FeSiO_3$ 3. 4, $MgSiO_3$ 2.2), biotite 3.2, ilmenite 0.9, magnetite 0.4, apatite 0.6 %.
An. II a	Molecular proportions of II.
An. II b	Molecular percentages of II.
An. III	Porphyric granodiorite, Vasankari, Kalajoki, east coast of Bothnian Gulf. Anal. E. Mäkinen. Bull. Comm. géol. Finl. N:o 47, p. 47.
An. III Norm.	Quartz 24.24, albite 26.72, anorthite 10.01, orthoclase 27.80, corundum 12, hypersthene 6.19, magnetite 1.62, ilmenite 1.52 %.
An. III a	Molecular proportions of III.
An. III b	Molecular percentages of III.

Vieremä, about 20 km N of Iisalmi. But then we have a considerable number of occurrences outside the zone of hypersthene granites, south of its south-western limit. These occurrences form a continuation southward from the unakite occurrences in Oulainen, Haapavesi, Haapajärvi and Viitasaari and are principally situated in the communes of Kivijärvi, Kannonkoski, Karstula, Pylkönmäki, Multia and Konginkangas.

The unakite analyses are given in Table 2. Characteristic of these analyses is that the aluminium content is high, from 17.78 % to 21.10 % in the Finnish occurrences. Also the sodium content is high, from 5.15 % to 7.94 % in the Finnish occurrences. In a few cases the K_2O is higher than the Na_2O. Such unakites contain more microcline than is the case with most granites and diorites.

IV	IV a	V	V a	VI	VI a	VII	VII a	VIII
61.48	1.020	61.85	1.026	60.58	1.0046	75.48	1.2517	77.47
0.92	0.012	1.88	0.024	1.35	0.0168	0.07	0.0009	0.26
16.44	0.161	15.80	0.155	15.43	0.1510	13.24	0.1295	11.00
1.06	0.007	0.32	0.002	2.04	0.0128	0.04	0.0002	1.04
5.11	0.071	8.35	0.116	5.18	0.0731	1.19	0.0166	2.02
0.05	0.001	0.01	—	0.14	0.0020	0.01	0.0001	—
2.93	0.073	1.02	0.025	0.36	0.0833	0.10	0.0025	0.43
5.82	0.104	3.86	0.068	5.95	0.1061	1.25	0.0223	1.02
4.11	0.066	3.78	0.061	2.63	0.0424	4.07	0.0656	2.86
1.33	0.014	2.80	0.030	2.23	0.0237	3.61	0.0383	4.14
0.26	0.002	0.48	0.003	0.23	0.0016	0.15	0.0009	—
0.69	—	0.44	0.024	0.62	0.0344	0.58	0.0322	0.25
0.11	—			0.09	—	0.09	—	—
100.31		99.85		99.83		99.86		100.59
				2.817		2.627		2.67

An. IV	Hypersthene granodiorite from Lahdenpohja, Jaakkima, north coast of Lake Ladoga. Analysed by L. Lokka.
An. IV a	Molecular proportions by V. Hackman. Map sheet D 2, Nyslott, p. 12.
An. V	Equigranular hypersthene granodiorite, Näläntöjärvi, Kiuruvesi. Anal. L. Lokka. Map sheet C 4, Kajaani, p. 137.
An. V Mode:	Quartz 16.7. plagioclase 6.8 (An$_{30}$), microcline 13.9, hypersthene 7.8, biotite 9.1, diopside 1.6, ilmenite 3.6, magnetite 0.4, apatite 0.9 %.
An. V a	Molecular proportions of V.
An. VI	Pyroxene-quartz diorite, Kilpimäki, Rautalampi. Anal. L. Lokka. Map sheet C 3, Kuopio, p. 85.
An. VI Mode:	Quartz 20.48, plagioclase 35.89, microcline 5.03, diopside 11.09, hypersthene 4.83, magnetite 2.38, ilmenite 1.46, apatite 0.50 %.
An. VI a	Molecular proportions of VI.
An. VII	Aplitic granite, Juurikkamäki, Saamaskylä, Leppävirta. Anal. L. Lokka. Map sheet C 3, Kuopio, p. 110.
An. VII Mode:	Quartz 34.94, oligoclase (An$_{13}$) 39.70, microcline 19.90, biotite 4.37, apatite 0.28 %.
An. VII a	Molecular proportions of VII.
An. VIII	Charnockite, St. Thomas Mt., Madras, India.

unakites. The unakite from Ollikkaanvuori contains as much as 45 % microcline. The unakites from Aspetorp in Sweden are also rich in K_2O. These unakites are thus »microcline unakites», whereas the unakite from Hogland (Suursaari) contains only 0.19 % K_2O against 7.71 % Na_2O and is a fairly pure albite-epidote rock. With a falling amount of SiO_2 it seems that the CaO content rises and thus also the amount of epidote. The unakite of Naistenkallio in Haapavesi contains as much as 38 % epidote.

On account of the high content of aluminium oxide and the alkalis the unakites approach the »alkaline rocks» in their composition and we may say that the whole series stands between the »alkaline» and the »alkali-lime» series of rocks.

Table 2. Analyses

	I	II	II a	III	III a	III b
SiO_2	68.77	61.21	1.015	60.10	1.001	67.05
TiO_2	0.60	0.52	0.007	0.40	0.005	0.33
Al_2O_3	14.02	17.78	0.174	19.86	0.195	13.06
Fe_2O_3	2.77	2.96	0.019	3.53	0.022	—
FeO	1.74	1.01	0.014	1.25	0.017	3.42
MnO	0.05	0.05	0.001	—	—	—
MgO	0.93	0.13	0.003	0.90	0.022	1.47
CaO	3.81	1.95	0.035	3.98	0.071	4.76
Na_2O	3.56	5.23	0.084	7.74	0.128	8.57
K_2O	3.10	7.74	0.082	1.90	0.020	1.37
P_2O_5	0.02	0.20	0.001	—	—	—
H_2O+	0.84	0.83	—	0.80	—	—
H_2O-	—	—	—	—	—	—
	100.21	99.61		100.71		
Sp. Grav.						

An. I	Unakite (»helsinkite») boulders from Alfsta, Sweden. Anal. Almström. Geol. Fören. i Stockholm Förh. 47, p. 504.
An. I Mode	Albite 31.03 (An_5), microcline 18.35, epidote 17.24, chlorite 4.34, titanite 0.41, ilmenite 0.68, apatite 0.03 %.
An. II	Unakite, Ollikkaanvuori, NE Haapavesi. Anal. L. Lokka. Map Sheet C 4, Kajaani, p. 122.
An. II Mode	Albite 43, microcline 45, epidote 6, chlorite 1.5, magnetite 3, titanite 1.2, apatite 0.3%.
An. III	Unakite of medium grain, Räsy, Oulainen. Anal. E. Mäkinen. Bull. Comm. géol. Finlande 47 and 73.
An. III a	Molecular proportions of III.
An. III b	Molecular percentages of III.
An. III Norm.	Albite 64.97, anorthite 13.07, orthoclase 13.07, nephelite 1.14, wollastonite 2.78, olivine 1.68, magnetite 2.78, ilmenite 0.76, hematite 1.60.
An. III Mode	Albite 67.07, microcline 11.12, epidote 15.43, chlorite 4.40, titanite 0.98, magnetite 1.62.

THE OCCURRENCE IN CENTRAL FINLAND OF GRANITE VEINS CUTTING SOME OF THE HYPERSTHENE GRANITES AND UNAKITES

In his description of the hypersthene granites of the islands in Haukivesi, Hackman (1931) states that in the islands of Anttosaaret, Paavalinsaari and surrounding islands there occur granite veins of an aplitic character which traverse the hypersthene granite. In describing the unakites of Pulkkila and Rantsila Wilkman (1931) says that they are cut by granite veins.

This granite is fairly small-grained and has what closely resembles a »panidiomorphic» structure (Rosenbusch). One of these granites has been analysed and the analysis is published as No. IX in Table 1.

These granites piercing the hypersthene granites and pyroxene diorites are thus the youngest intrusive igneous rocks of Central Finland. They occur both as dykes and as small fields of igneous rock.

of unakites.

IV	V	VI	VI a	VI b	VII	VIII	VIII a
60.10	58.10	57.52	0.960	64.84	57.27	53.63	0.889
0.68	0.11	0.75	0.010	0.68	0.75	0.65	0.008
18.60	21.10	19.22	0.188	12.71	18.50	18.87	0.185
2.32	3.07	5.11	0.032	—	2.43	6.03	0.038
2.43	0.21	2.02	0.028	6.22	2.22	0.98	0.014
0.07	—	—	—	—	0.09	0.03	—
1.09	0.88	0.68	0.017	1.15	1.06	1.98	0.049
1.45	7.12	4.30	0.077	5.21	4.09	10.10	0.180
5.15	7.71	7.63	0.123	8.31	2.65	4.44	0.072
6.67	0.19	1.26	0.112	0.88	8.69	2.62	0.018
0.13	—	—	—	—	0.61	0.17	0.001
1.13	0.95	1.06	—	—	1.23	0.60	—
—	—	—	—	—	—	—	—
100.16	99.54	99.45	—	—	99.99	100.10	
					2.827	2.664	

An. IV	Unakite (»helsinkite») fine-grained. Streaks in VII (with 0.19 % BaO and 0.15 CO_2). Anal. A. Bygden. Sveriges Geol. Unders. Ser. C N:o 325.
An. V	Unakite (»helsinkite») Suursaari. Anal. A. Laitakari, Bull. Comm. géol. Finlande N:o 51 (1918).
An. VI	Unakite, felsitic. Räsy, Oulainen, Anal. E. Mäkinen, *loc. cit.*, An. III.
An. VI a	Molecular proportions of VI.
An. VI b	Molecular percentages of VI.
An. VI Norm.	Albite 64.45, anorthite 14.73, orthoclase 6.67, diopside 3.67, wollastonite 0.81, hematite 2.24, magnetite 4.18, ilmenite 1.52 %.
An. VI Mode	Albite 64.45, microcline 6.67, epidote 16.03, chlorite 4.24, magnetite 3.97.
An. VII	Unakite (»helsinkite») rich in microcline, Aspetorp, Sweden (with 0.38 % BaO). Anal. A. Bygden. Sveriges Geol. Unders., Ser. C N:o 325.
An. VIII	Unakite rich in epidote, Naistenkallio, Haapavesi. Anal. L. Lokka. Map sheet C 4, Kajaani, p. 123.
An. VIII a	Molecular proportions of VIII.
An. VIII Mode	Albite (An_2) 38.7, microcline 15.6, epidote 37.7, chlorite 2.4, actinolite 2.6, titanite 1.6, magnetite 1.4 %.

THE SIGNIFICANCE OF THE OCCURRENCE OF HYPERSTHENE GRANITES AND UNAKITES IN THE SAME PART OF THE COUNTRY AND THEIR MODE OF ERUPTION

The many occurrences of hypersthene granites and unakites in Central Finland are all situated in the same part of the country, with the hypersthene granites and their more basic differentiation products, the hypersthene granodiorites and various kinds of hypersthene diorites, mostly occurring in the North Ostrobothnian parishes of Pattijoki, Paavola and Kalajoki and in the previously described zone to the southeast.

The unakites do not have quite so wide a spread as the hypersthene granites. They occur not only in the northeastern half of the hypersthene granite belt, but also outside this belt, directly southwest of it, on the northwestern extension of the belt formed by lines I and II (Fig. 1) and southwest of the intersection of lines II and III and, further, about 80 km SW of this intersection, which all are here interpreted as a system of craks. These craks probably opened under the pressure of great masses of magma not containing much water-vapour, from which the hypersthene-bearing rocks were crystallized. The unakites seem to have erupted from magma very rich in water and volatiles. The representatives of these two magmas, which we now find exposed at the surface, never reached the surface of the earth at the time in question, *i. e.* about 1 700 million years ago as shall be explained later on, but crystallized as deep-seated eugranitic »hypidiomorphic granular rocks» at such places along the crack lines and near by in larger cavities which they had managed to open for themselves. The magma of the hypersthene granites and related hypersthene diorites erupting along the crack system has produced some sixty fairly considerable occurrences of hypersthene granites and associated rock types (granodiorites, quartz diorites). This cracking of the rock ground did not, however, cease with the crystallization of the principal part of the hypersthene granites, but continued in some parts of the vicinity of the belt, causing some of the hypersthene granite occurrences to become strongly brecciated. This is especially the case where the important crack-line II crosses the present site of the lake district of Haukivesi (Hackman 1931).

The eruption of the unakites seems, however, to have taken place in quite a different way. They occur mostly at single points. Only in a few places are they described as now forming short, broad dykes or as smaller, narrow, often porphyritic dykes penetrating the already consolidated hypidiomorphic unakite bodies (Hackman 1931). It is probable that the unakites forced their way upwards in the earth's crust by some kind of »stooping» and ultimately also they crystallized when the pressure had increased sufficiently to cause a complete crystallization and, at that time, the formation of deep-seated rocks. Wetherill, Kouvo, Tilton and Gast (1962) recently determined by the potassium-argon method the time at which the crystallization of two of the unakite occurrences took place and found a value of 1 700 million years. This is in accordance with the field observations, according to which the hypersthene granites and also the unakites evolved later than the post-Bothnian granites (about 1 800 million years or more). There is further field evidence that the Wiborg (Viipuri) rapakivi at Taipalsaari cuts the hypersthene granite. Further, it would seem that the hypersthene granites are older than the Karelidian orogenesis since no hypersthene granites or unakites have been observed inside the domains of the remnants of the Karelides. They occur, however, in the old Archean east of the main region of Karelidian rocks in Finland (Wilkman 1928).

THE AGE OF THE HYPERSTHENE ROCKS AND THE UNAKITES

In his descriptions of the hypersthene granites and the unakites in the map sheets of Kajaani and Kuopio, Wilkman (1931, 1938) in some places states that the unakite is older than the hypersthene granite, in other places that the unakite or unakite porphyry cuts the hypersthene granites and there is younger. It is probable that both observations are correct, which would mean that there are unakites which differ somewhat in age of crystallization just as there are hypersthene granites differing in age, and that both groups of rocks belong to the same period, which may be regarded as dating back to about 1 700 million years.

DISCUSSION

The hypersthene rocks crystallized from a magma poor in water and volatiles and this is the principal cause for the mineralogical peculiarities they show.

The unakites crystallized from a magma rich in water and volatiles, and the peculiar character of the minerals formed results from this abundance of water and volatiles present in the unakite magma.

We can say that if the hypersthene granite magma would have acquired and incorporated a certain portion of the unakite magma, the resulting magma would have crystallized as an ordinary biotite granite.

And if the unakite magma would have acquired and incorporated a certain portion of the hypersthene granite magma, it would also have crystallized as an ordinary biotite granite or biotite diorite.

What actually happened is, however, quite different, indeed the very opposite of what was just suggested. Instead of one biotite granite magma, we have in the same district a hypersthene granite — hypersthene diorite magma, unakite magma and an aplite granite magma, each feeding its particular rock sites where the hypersthene rocks and the unakites and, further, the granite veins had crystallized.

There must thus have existed at great depths — possibly at two or three different depth zones — three different magmas, one poor in water, giving rise to hypersthene rocks, the second rich in water, giving rise to unakites, and the third a granite magma. And, since these magmas could theoretically form in certain proportions an ordinary biotite granite magma, we arrive at the possibility that an ordinary biotite granite magma at great depth had split, or »differentiated» into the two peculiar magmas described here and a certain amount of aplite granite magma. This might have happened in two different ways, either of which we may describe as having taken place:

a) By way of crystallization-differentiation at great depths and at high temperature and, consequently, at high pressure. During such crystallization, a crystallization-differentiation took place. Subsequently, after the pressure had fallen sufficiently to produce by melting three magmas from the three kinds of rocks, these magmas later erupted, separately forming the hypersthene granites, the unakites and the granite dykes.

There are, however, certain difficulties with regard to an explanation of the production of the unakite magma by a crystallization-differentation of this kind. Part of the unakites have an equal amount or more of iron oxides, magnesia and lime as the hypersthene rocks, whereas the unakite magma, if produced as a residual magma after the crystallization of the hypersthene rocks, should, in combination with the bulk of the water contained in the primary magma, have less Fe, Mg and Ca than the hypersthene rocks crystallizing first. The formation of these rock magmas can not, therefore, be explained by crystallization-differentiation and a subsequent melting of the products.

b) Or, then, the rocks may be formed by way of a splitting of the parent magma into a hypersthene rock magma poor in water; into an unakite magma rich in water and, as a third member, an acid granite magma containing some water.

In an experimental work on waterless silicate melts Greig (1924) has furnished us with examples of silicate melts splitting up into three liquids. At close to 1 700° he obtained a liquid field of three co-existing liquids: a field A rich in silica, a field consisting of a mixture of A+B and a third field B containing much of the basic component and rich in water. He investigated mixtures SiO_2 and $Mg+Al_2O_3$ of SiO_2 and FeO and obtained three co-existing liquid fields at temperatures close to 1 700°, which corresponds to a depth of about 160 km in the earth's crust.

The occurrence, described in this paper, of three different magmas existing simultaneously might perhaps be interpreted as the splitting up of a magma rich in SiO_2 and FeO at high temperature in three co-existing magmas, one of acid granite composition, one of the composition of hypersthene granite and hypersthene diorite magmas fairly poor in water and one rich in albite and water, together with the components of epidote and chlorite giving rise to the unakites. A continuation of Greig's experiments at high pressure and silicate melts containing water would probably give further insight into the problems dealt with in this paper.

In the present author's view, the rocks dealt with here thus originated as a result of two different kinds of »differentiations», first, magmatic differentation, leading to the differentation from a primary magma of the hypersthene granite magma and the unakite magma, and, second, »crystallization-differentation», leading to the formation from the hypersthene granite magma of the hypersthene granites, hypersthene-quartz diorites, hypersthene diorites, quartz-hypersthene gabbros and norites.

The author is of the opinion that further field work with regard to the occurrence of hypersthene rocks and their situation along crack-lines and near the SW borderline of the Karelidic formations would be desirable. It is therefore the aim of this paper not only to give a record of what is at present known about the rocks dealt with here, but also to inspire further investigations into the problems connected with the hypersthene rocks and unakites of Central Finland.

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8

ON THE INFLUENCE OF SEISMIC VIBRATIONS ON SEDIMENTS [1]

BY

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ABSTRACT

The manganese nodules and magnetic spherules at the sediment-water interface have been explained in a number of ways. A new approach to the problem is presensed in this paper. The influence of seismic waves is discussed in the light of experiments and mathematical theory.

CONTENTS

INTRODUCTION

The following study concerns the influence of earthquakes on sediments on the vertical distribution of particles in sea floor sediments. The scope of the investigation was to see if vibrations might be able to keep larger and lighter particles on the sediment surface and thus be a factor to be remembered in discussion concerning the abundant occurence of manganese nodules, *etc.*, at the sediment—water interfaces.

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[1] Received October 15, 1963.

The seismic vibrations are irregular. It is extremely difficult to simulate the phenomenon in the laboratory exactly as it appears in natural conditions. We were therefore compelled to simplify the actual experiments by assuming that the effect of the vibrations can be studied by treating horizontal and vertical components separately. We are aware that this kind of separation of the components may be rather far from reality, but in so doing we hope to shed some light on the influence of seismic vibrations on sediments. Furthermore we had to change the grain size considerably during the first experiments which we assume to be a rough approximation, but may give further impetus to study similar phenomena.

EXPERIMENTS

The experiments started with a short study of the influence of vertical vibrations on sediments. Except for closer packing, in good agreement with the values obtained by Fraser (1935), no very visible features were observed when the accelerations were well below 1 *g*. Above 1 *g*, a redistribution took place; but as such intensities are rare (Richter 1958), no further studies were made of the vertical component. Instead, we have investigated horizontal vibrations of certain amplitudes and frequencies.

Our method was to shake the sample container with a sediment suspension in a horizontal direction, submerged in a stationary bucket. In order to copy, on a labororatory scale, the influence of horizontal waves, it was necessary to keep the water stationary to cause friction against the suspension surface while the sample container was moving to and fro in the water. This was difficult to achieve with silt and clay suspensions, so we used a sand suspension which only slightly escapes the sediment container. Our sand consisted of rounded quartz grains of the size 400—1 200 μ, with a median size of about 800 μ.

In order to avoid undesired secondary vibrations, the shaker was made as stable as possible. The range of amplitudes was about 1—3 mm while the frequency could be varied between 2 and 30 c.p.s. In order to prevent, or at least to suppress, the formation of turbulent cells, we placed a plastic plate with many furrows at the bottom of the container. The furrows were made about 2 mm deep. The bottom friction could then be assumed to be high. In addition, at each end of the container, two higher lamellae were placed; the outer ones were highest. The lamellae were intended to diminish the effect of the walls during the shaking. After a series of experiments we finally obtained an area in the middle of the container which, at a certain amplitude range, was only very slightly influenced by the walls of the container. This area was used for the experiments.

The acceleration given to the container during a shaking is transferred to the suspension by the bottom. The energy is then transferred from the lower layers upwards as a damped transversal wave motion. However, owing to the friction, the acceleration decreases in the direction against the sediment-water interface. Therefore, the maximal acceleration at the bottom of the suspension is higher than at the sediment surface.

To study the movements of larger grains, glass spherules (~ 8 mm in diameter) of the density 2.53 g/cm^3 were used. Larger quartz grains were also used and they showed the same average results, but on account of their varying shape, roundness and initial orientation in the sediment, the result showed a very large dispersion. The glass balls were mainly placed with their upper surface at a depth of 1 cm below the surface of the sand suspension. The sand suspension was in a rather loose state regarding the porosity of deep-sea deposits (see Hamilton 1960).

The experiments showed that a low frequency and an amplitude not exceeding a few millimeters, giving a low acceleration, the glass balls did not change their position. If the acceleration was increased slightly (~ 0.05 g) the glass balls began to move upwards very slowly, not reaching the surface in less than 10 minutes of shaking. Further experiments with maximal acceleration increased by steps of about 0.05 g, gave travelling times of respectively 5—7 minutes, 1—2 minutes, and 20—60 sec. The last refers to an amplitude of 1.8 mm at a maximal acceleration varying between 0.75 g (given to the container), and about 0.4 g (at the sediment surface). At an amplitude of 3 mm the balls reached the surface after 20—60 sec. if the maximal acceleration of the container was about 0.65 g and the maximal acceleration at the sediment surface about 0.5 g. It is possible that the larger amplitudes are more effective than the shorter ones. The observation of the surface accelerations was difficult and the values are not very reliable.

The spherules of lighter glass (d = 2.23 g/cm ^{3}) moved upwards more rapidly than the heavier ones.

The first effect of the vibration seems to be a closer packing. It affects the grain distribution at the very surface only. If however, the intensity was increased by a certain amount, the packing was not enhanced, but rather the sand was loosened. The minimum acceleration necessary to influence the structure of the sand suspension is not known for this or any other sample. The loosening of the particles at and near the surface indicates that the sediment ceases to move as a body and there is a velocity gradient downwards from the sedimentary surface so that the grains at the surface have a lower velocity than those further down. This difference, indicated by the difference in acceleration between the bottom of the container and the acceleration of the central part of the sediment surface (see above), and the increasing friction and pressure downwards cause an upward movement of larger particles relative to smaller ones.

MOVEMENTS OF SAND DURING SHAKING

Let us assume that during a horizontal shaking of sand in water in a definite direction, the sand particles have the same velocity in each horizontal layer. The size of the sand grains then, can be assumed to be negligible in comparison to the thickness of the layers considered. The Cartesian coordinate system is fixed so that the horizontal x-axis points in the direction of the movements of the sand lying in a certain layer below the sand surface, and the z-axis points vertically upwards.

The following notations are used:

ϱ = density of the medium (water or water-sand mixture)
ϱ' = apparent density of sand in water
x = distance of a particle from equilibrium = elongation
v = velocity of a particle
r = amplitude of the vibrations at $z = 0$
$\sigma/2\pi$ = frequency of the harmonic motion
τ = shear stress in the medium
η = apparent viscosity coefficient
ν = apparent kinematic viscosity coefficient
μ = friction coefficient
p = »hydrostatic» pressure caused by the sand alone
F_μ = friction
F = normal force
z_o = z-coordinate of the sand surface

We assume that the layer $z = 0$ is kept in harmonic motion

$$x = r \sin (\sigma t) \tag{1}$$

and we look for the motion of the upper sand and water layers caused by friction and viscosity.

Friction between two bodies is

$$F_\mu = \mu F \tag{2}$$

where the normal force F_n is the pressure force of the sand

$$F = pA. \tag{3}$$

A is the area of the sand bodies in common. The »hydrostatic» pressure is

$$p = \varrho' g(z_o—z). \tag{4}$$

From (2) and (3) we find the frictional force between gliding layers

$$F = \mu pA. \tag{5}$$

The shear stress is

$$\tau = \partial\, v/\partial z \tag{6}$$

We assume for a moment that

$$\partial v/\partial z > 0.$$

Then the total force acting on a layer between z and $z + dz$ from above is

$$\eta\,\frac{\partial v(z+dz)}{\partial z}\,A + \mu p\,(z+dz)A$$

and from below

$$-\eta\,\frac{\partial v(z)}{\partial z}\,A - \mu p(z)A.$$

These forces give the acceleration $\partial v/\partial t$ to the layers of mass $\varrho\, Adz$. According to Newton's second law we then have

$$\eta\,\frac{\partial v(z+dz)}{\partial z} + \mu p\,(z+dz) - \eta\,\frac{\partial v(z)}{\partial z} - \mu p(z) = \varrho dz\,\frac{\partial v}{\partial t}$$

or

$$\eta\,\frac{\partial^2 v}{\partial z^2} + \mu\,\frac{\partial p}{\partial z} = \varrho\,\frac{\partial v}{\partial t}.$$

Substituting p from (4) and dividing by ϱ we obtain

$$\nu\,\frac{\partial^2 v}{\partial z^2} - \frac{\partial v}{\partial t} = \mu\varrho' g/\varrho\ ,\ (\partial v/\partial z > 0), \tag{7}$$

where $\nu = \eta/\varrho$. In the same way we find for $\partial v/\partial z < 0$

$$\nu\,\frac{\partial^2 v}{\partial^2 z} - \frac{\partial v}{\partial t} = -\mu\varrho' g/\varrho \cdot (\partial v/\partial z < 0). \tag{8}$$

When sand layers are in relative motion, many sand grains are in suspension and make occasional impact on other particles. Thus, the friction between layers is small but the apparent shearing stress large. Therefore, it is reasonable to solve the simpler equation

$$\nu\,\frac{\partial^2 v}{\partial z^2} - \frac{\partial v}{\partial t} = 0 \tag{9}$$

instead of (7) or (8) for which the boundary conditions are highly complicated.

The percentage of sand grains in suspension depends on the velocity gradient. Therefore, the apparent shearing stress (as well as the friction itself) varies with time. Indeed, considering ν as a function of the velocity gradient, the equation (9) will be rather complicated. To simplify the solution we assume ν to be constant. This assumption influences our results somewhat, but the general features should not change appreciably.

For constant ν the solution of (9) reads

$$v = \sigma\, r \exp\{-\sqrt{\sigma/(2\nu)}z\} \cos[\sigma\, t - \sqrt{\sigma/(2\nu)}z]. \tag{10}$$

This solution satisfies the boundary conditions

1° $x = r \sin(\sigma t)$, when $z = 0$,

2° v approaches zero when z approaches + infinity.

The distance of a particle from equilibrium is found from (10)

$$x = r \exp\left[\sqrt{\sigma/(2\nu)}z\right] \sin\, \sigma\, t - \sqrt{\sigma/(2\nu)}z]. \tag{11}$$

The acceleration of a particle is

$$a = -\sigma^2\, r\, \exp\left[\sqrt{\sigma/(2\nu)}z\right] \sin[\sigma t - \sqrt{\sigma/(2\nu)}z]. \tag{12}$$

Formulas (10), (11), and (12) indicate that the particles are in harmonic motion. The amplitude

$$r \exp\left[\sqrt{\sigma/(2\nu)}z\right]$$

decreases upwards. The phase of the motion is retarded. The phase shift is $\sqrt{\sigma/(2\nu)}z$. The energy in the sand will be transferred from lower layers upward as a damped transversal wave motion.

In solving equation (9), we assume that ν is a constant. The solution for ν varying with z can be found, for example, in dividing the sand mass into a certain number of separate layers where ν can be assumed to be constant and combining the different results as an entity.

Friction certainly has a marked effect on our solution. At the very moment $\partial v/\partial z$ becomes zero, the adjacent layers begin to move as a rigid body until the forces opposing the friction increase enough to cause relative movements again.

HORIZONTAL SHAKING WITH PEBBLES IN SAND

For the sake of simplicity it is assumed that the pebbles are spherical and that they are embedded in the sand so far apart that their mutual influences can be neglected. Let us consider the motion of a single pebble. It is mainly linear and rotational to-and-fro motion caused by frictional forces.

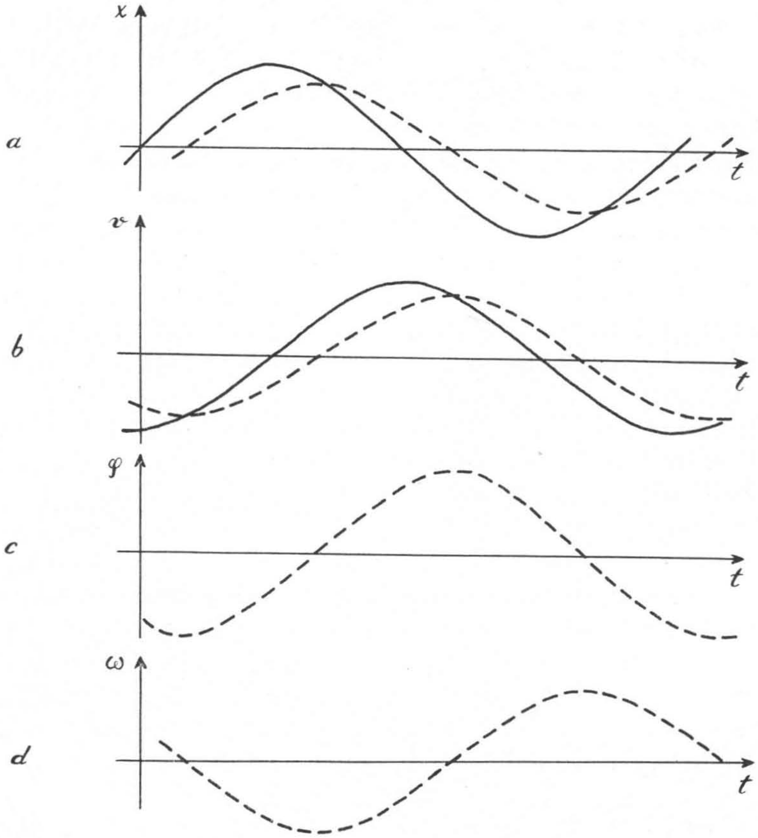

Fig. 1. Curves representing sand motion and pebble motion are plotted as a function of time in full line and dashed line respectively. In a and b the elongations and velocities of the linear motion and, in c and d those of the rotational motion are drawn. The figures are not in scale.

If the density of the pebble is the same as that of the medium, the linear velocity of the pebble will be about equal to the weighted mean velocity of the medium in the vicinity of the pebble. The mean motion of the sand and the motion of the pebble are in phase. The horizontal component of the resultant force acting on the pebble is opposite to the elongation, *i. e.*, to the distance from the equilibrium, of the pebble. The linear velocity shows a phase shift of 90 degrees in advance of the linear motion and the angular velocity is just 180 degrees behind the linear motion. Because of symmetry of the motion in space and time, there will be no vertical net force in this case.

If the density of the pebble is greater than that of the medium and the frequency is not too high, the motion of the pebble will be retarded and slowed down. That is to say, there will be a phase shift between the weighted mean velocity of the medium and the velocity of the pebble, and the amplitude of the motion will be smaller. Also the rotational component will be retarded and the amplitude smaller. In this case, however, we have a contribution to the lifting forces. This can be seen as follows (see Fig. 1). At a specific instant, the sand and the pebble move to the right; at the same time the elongation of both is positive. Then the speed of the pebble is in general greated than that of the sand. Therefore, the pressure in front of the pebble is larger than that of the rear side. When the elongation is positive, then the rotational direction is clockwise. Therefore, the lifting force generated by the friction at the high pressure side outweighs the negative force at the low pressure side. From similar conclusions it can be deduced that there will be a positive net contribution to the lifting force in this case.

In places, the sand has a tendency to become detached from the surface of the pebble because of the relative motion of the sand and pebble so that the sand grains will fall more or less freely. This causes a force because the sand grains remain in the gaps created after the direction of the movement has changed. This force is of importance at longer periods only, when the particles have time to fall down. One force, which depends on the difference of the density of the pebble and its surroundings, must also be considered. Naturally, it can act in either direction. If we assume that the density of the pebble and that of the medium are the same, this force is zero.

According to equation (10)

$$\lambda = 2\pi \sqrt{2\nu/\sigma}.$$

The wave length of the transversal wave motion in sand decreases when the frequency increases. That is to say, when the frequency is low enough, the wave length λ is large compared to the diameter of the pebble. With higher frequencies the wave legth may be of the same magnitude as, or small compared to, the pebble size.

In the case of a small wave length, compared to the pebble size, the above considerations about the lifting force partly lose their effect because the amplitudes of the linear and rotational motions diminish considerably with increasing frequency. Furthermore, it is no more evident that the phase shifts are such that lifting forces will be created.

For low frequencies and long wave lengths, the relative velocities around the pebble will be comparatively large. For, neglecting the influence of the cosine factor, the maximum difference of the velocities just above and below the pebble is, according to equation (10),

$$\Delta v_{\max} = -\partial/\partial z \left\{\sigma r \exp\left[-\sqrt{\sigma/(2\nu)}z\right]\right\} D$$

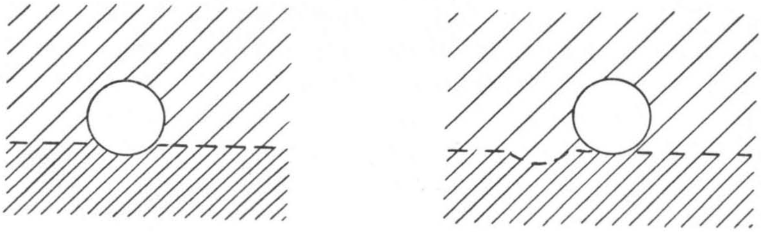

Fig. 2. Sand motion pushes a pebble from the indentation in a denser sand below the pebble.

where D is the diameter of the pebble. Then, according to equation (12),

$$\Delta v_{\max} = (a_o/\sqrt{2\sigma\nu})\, D$$

where a_o is the amplitude of the acceleration. Thus, for constant a_o, the relative velocities in the vicinity of the pebble increase with decreasing frequency of the motion. In this case the amplitude of the motion comparatively large and it may seem that the lifting forces should be large too. Indeed, according to our experiments, this is not true, but on the contrary, the forces are quite small or vanish altogether. The phenomenon can be explained by assuming that the degree of packing decreases because of increased velocity gradients. Hence, the forces around the pebble become more symmetrical thereby cancelling each other to a large extent.

When all the forces mentioned (and the accelerational forces that are horizontal) are taken into account, a net force that acts upwards becomes discernible in most cases.

When consideration is given to the settling of the sand in phases during which sand is moving as a rigid body, the uplifting force increases considerably. This effect is illustrated in Fig. 2. The sideward movement pushes the pebble from the indentation in the harder sand below the pebble.

APPLICATION OF THE EXPERIMENTS TO DEEP-SEA DEPOSITS

No direct observations of the influence of seismic vibrations on deep-sea deposits have been made. Besides, there are no features in the structures or textures which could be proved to be caused by such vibrations. Therefore, this discussion is based only on the probability, considering only what we may expect to find.

Our experiences seem to be valid for an area rather near an epicenter, because the accelerations used in our experiments are rather high and because the amplitudes and frequencies used correspond to those prevailing near epicenters. With increasing distance from such regions, the intensities decrease and therewith also the possibilities of the influence of seismic waves. However, the decrease in intensity does not only depend on the distance from the epicenter; the surface intensities on the light overburden may be considerably higher than on the basement rock outcrop (see, *e. g.*, Neumann 1954). Owing to the resonance effect, large bodies of loose earth tend to shake more severely than rock outcrops. This effect is so pronounced that it may much more than offset the relatively greater ability of hard rocks to transmit the vibrations without absorption of energy (Howell 1959, p. 93). Deep earthquakes are felt over a much wider area than shallow earthquakes of the same intensity, but this is because the maximum intensity is less in proportion to the total energy of the earthquakes owing to the greater depth of focus (*op. cit.*, p. 96).

The highest seismicity of the deepsea occurs in association with arcuate structures of the Pacific type. Moderate and occasionally high seismicity is found associated with structures that are not arcuate, for example around submarine ridges (see further Gutenberg and Richter 1953, and Menard 1960 b). There are difficulties in cataloguing the regions where we may except to find the influence of vibrations on sediments. One such difficulty is that the seismicity on the floor of the oceans is known, geologically speaking for a short time only. Accordingly all the known earthquakes, including those affecting the ocean floors, have hitherto been registred on seismographs on the continents only. Our present picture of the seismicity of the ocean basins may therefore be somewhat incomplete considered in terms of longer periods in the past.

First it must be noted that, as experiments show, coarser grains may travel upwards owing to vibrations of certain intensities. A redistribution of particles in a top layer is possible, however, only in seismic regions of sufficient intensity as far as we can judge from the experiments of horizontal vibrations; the effect of simultaneous vibrations in other directions have not been studied and we cannot decide whether it might change the sphere of influence or not. If the coarser grains are on the surface, smaller intensities are needed to keep them there. This is because rolling, or the friction against the water may be sufficient to allow them to stay at the sediment-water interface or to uncover coarser particles from smaller ones.

The cohesion, the result of the molecular interaction between the solid particles at their points of contact, increases with decreasing grain size owing to the increasing number of contacts between solid particles. Furthermore, pressure and viscosity exert an influence on the particles. For sediments that do not contain any clay, the value of the cohesion is equal or close to zero (Terzaghi 1956, p. 2). Our experiments are therefore applicable only to sandy or silty deposits such as carbonate oozes (especially foraminiferal oozes), siliceous oozes, and organic and inorganic silty and sandy muds.

The effect of seismic vibrations on clay may be assumed to be small because the impulse per clay particle, which tends to displace the particle with reference to its neighbors, is very small compared to the adhesion between the clay particles (Terzaghi 1956, p. 7). Therefore, the clay may be assumed to act as a rigid body during vibrations.

On the other hand, larger particles on a clayey surface might be thrown out of their former holes thus reaching a higher position (*cf.* Richter 1958, Figs. 5—4). Particles laid down on the surface or larger particles may be moved away by the friction against the water during the vibration or by currents initiated by earthquakes. All this might help nodules, slabs, etc. to remain at the surface for longer periods. Silty deposits have the character of forming a metastable structure which may collapse on slight provocation. The collapse is termed spontaneous liquefaction (Terzaghi 1956, p. 3). This problem will not be further dealt with here.

Some irregularities in the distribution of particles of different size and density will be discussed in the following in the light of our experiences.

INVERSE GRADING

The foregoing discussion suggests that during jarring smaller grains on and near the sediment surface have a better opportunity of travelling downwards than larger ones. The larger grains will thus have a tendency to be at the surface, thus giving rise to an inverse grading. If, after additional sedimentation, new vibrations take place, this new layer will probably be inversely graded so that two layers with inversed grading may be found atop each other. A part of such a pile may give the impression of being normally graded. In the interpretation of silty and sandy sequences from, *e. g.*, deep-sea trenches, this possibility should be borne in mind.

THE DISTRIBUTION OF NODULES AND MAGNETIC SPHERULES

The concentration of the manganese nodules varies both regionally and locally (Mero 1960). Judging from photographs and corings, they occur mainly in a layer at the surface one nodule thick; however, they occur also further down but the concentration seems to be considerably less. The bulk of the manganese nodules seems therefore to occur mainly at the sediment-water interface. The nodules must be in contact with the water to accrete. Should the sedimentation rate exceed the rate of growth of the nodules, the nodules will be buried and cease to form. Since the rate of growth of the nodules is estimated to be of the order of 1 mm per 1 000 years

(Pettersson 1945, Buttler and Houtermans 1950) or still less (Goldberg and Arrhenius 1958) while the reported rate of sedimentation for red clays and oozes ranges from 1 mm to several cm per 1 000 years, it raises the question as to why the manganese nodules are still at the interface. Shipek (1960) has discussed the problems regarding the mechanisms of the nodule formations and the factors involved. Here, only an additional factor, the influence of vibrations, will be discussed — this without denying the role of other factors.

The uplifting effect on large particles caused by vibrations may be an additional force helping the nodules to remain at the surface for longer times and giving rise to high concentrations of nodules at the sediment-water interface also in regions with a comparatively high rate of sedimentation. In regions of major seismicity the top layer of the sediment carpet can be involved in seismic vibrations several times before reaching a depth of some centimeters. Therefore, the possibility of a redistribution of the grains must be considered. Also the presence of non-horizontal discontinuities may produce some kind of turbulent cells during the vibrations which disturb the structure and texture of adjacent sediments. Coarse grains could accumulate in convergences at the surface.

The density of the solids in foraminiferal oozes (and muds) is 2.69 g/cm^3 according to Hamilton (1960, p. 372). The density of manganese nodules is about 2.4 g/cm^3 but ranges between 2.0—3.1 g/cm^3 according to Mero (1960), the density of phosphoritic nodules is 2.62 g/cm^3 (Mero 1961, p. 5). Considering that the experiments are carried out with quartz sands ($d = 2.65$ g/cm^3) and glass balls ($d = 2.53$ g/cm^3) the results obtained from our investigations can be applied to foraminiferal deposits containing the nodules here mentioned. However, the comparatively high rate of sedimentation in such areas (*cf.* Olausson 1961) limits the influence of earthquake vibrations to the regions of major seismicity, where the uplifting effect may be able to keep pace with the sedimentation. In the oligotrophic regions where coccolithophorids are responsible for most of the carbonates, the rate of sedimentation is lower and, therefore, also seismic vibrations with a lower frequency and intensity may have an influence on the redistribution of the grains.

The density of the solids in siliceous oozes is 2.1—2.3 g/cm^3 (for diatomite; Baas Becking and Moore 1959). Since manganese nodules may have an equally low density, it is our conclusion that the influence of vibrations on, for example, radiolarian ooze (muds) might be about the same as for foraminiferal deposits.

Larger phillipsite crystals ($d \sim 2.2$ g/cm^3) and consolidated slabs (sensu Menard 1960 a) may be expected to remain at the surface as a result of vibrations. The slabs are assumed to have been exposed on the sea floor for 10^4—10^5 years (*op. cit.*, p. 35).

Phosphoritic nodules are found, for example, on the top and sides of banks, on steep escarpments which appear to be fault scarps, on the walls of submarine canyons, and on the break of the continental shelf (Mero 1961, p. 7). All the localities where phosphorites are found are essentially non-depositional environment (Dietz *et al.* 1942). The nodules must be in contact with the water to accrete. Once buried in sediment, the nodules will cease to enlarge by accretion of colloidal particles (Mero 1961, p. 7). Discontinuous layering occurs within the nodules. This has usually been explained as due to covering by sediment, which would stop the growth process, with subjacent erosion of the unconsolidated sediment cover, exposing the nodules to sea water again. However, this discontinuous layering as well as the growth of the nodules might also be discussed in the light of the uplifting effect or other rearrangements of the nodules owing to seismic vibrations. Such vibrations might be able to separate larger particles from finer particles, to lift them up to the sediment surface or to move them in a horizontal direction (*cf.* Fig. 2).

Most phosphoritic nodules in the California region seem to have been formed during the Miocene (Mero 1961, p. 4). This epoch is known for intense volcanic activity in that area and at the same time likely a seismic one as well.

Another irregularity concerns the distribution of »cosmic» spherules. In some cores their number increases towards the top of the core (Fredriksson 1959). This is also true in older strata (Lower Miocene in age) outcropping at the bottom (Core 90; diagram by Pettersson and Fredriksson 1958). However, the density of these grains is greater than the density of the solids in oozes and clays and, therefore, the uplifting effect of vibrations may be reduced or even reversed.

Burri and Koczy (1956) made an experiment on a clay suspension with a water content of 60 % and with a regular distribution of artificial magnetic spherules. By vertical vibrations produced by a small hammer striking the bottom of a container during 5 days, they finally obtained a distribution of the spherules increasing upwards. The magnitude of the accelerations used is not known.

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