GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
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N:o 184

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

HELSINKI
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1

PEDOGEOCHEMICAL PROSPECTING IN GLACIATED[1] TERRAIN

BY

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ABSTRACT

This study was made for the purpose of testing pedogeochemical methods of ore prospecting in glaciated terrain. Presented are the Cu- and Ni- contents of the glacial till in the vicinity of the known ore deposits of Outokumpu, Makola and Kiuruvesi. The longitudinal direction of the anomalies observed runs parallel to the local movement of the glacial ice. The study shows that pedogeochemical prospecting in glaciated regions may be considered as a micro-variation of boulder tracing — as a method of searching for infinitesimal »boulders».

CONTENTS

INTRODUCTION

The Geological Survey of Finland started pedogeochemical prospecting operations in 1952, but even earlier different geochemical prospecting methods had been tried out in Finland and experiments have been continued at

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[1] Received December 19, 1958.

every favorable opportunity (Rankama, 1940, 1941, 1947; Salmi, 1950, 1955, 1956, 1958; Marmo, 1953, 1955, 1958 a and b; Hoffren, 1958; Hyvärinen, 1958; Kauranne, 1951, 1958; Paarma, 1958). In investigating the possibilities of pedogeochemical ore prospecting in Finland, samples have been taken from the vicinity of the following known ore occurrences: the mined-out ore body of Makola, the copper mine under current exploitation at Outokumpu, and the recently discovered copper-nickel prospect at Kiuruvesi. As the location and size of the ore outcrops were known, the sampling grid could be made fairly sparse and the treshold value for the anomaly in the case of the contents obtained could be so determined as to clearly reveal the influence of the ore. It was endeavored to take the samples from along lines drawn perpendicular to the direction of transport of the glacial ice at intervals of 20 to 100 meters. The till samples were taken from the C-layer of soil profile at a depth of 60—110 cm.

MAKOLA

The outcrop of pyrrhotite, pentlandite, chalcopyrite ore of Makola, situated in serpentinite, is some 120 m long and 20 m broad. The mean content of the outcrop of ore was 0.87 % Ni and 1.16 % Cu according to the analyses made during prospecting by the Geological Survey of Finland. The region, which has in places become boggy and where the thickness of the layer of loose earth averages four meters, consists mostly of glacial till. Numerous glacial striae have been measured in the region, but judging from the boulder fan (Saksela, 1949) the direction of transport was approximately N 60°W. The region is partly inhabited, but contamination is not to be feared beyond the immediate vicinity of the mine and the waste area.

The Makola ore was at first quarried as open pit, and therefore it was not possible to obtain samples from directly on top of the outcrop. The samples collected in the surroundings were dried out and sieved and the fraction $\varnothing < 0.012$ mm was separated from the finest sieve fraction by decanting. The grain classes $\varnothing < 0.012$ mm and $\varnothing$ 0.012—0.125 mm thereby obtained were both analyzed chemically, in addition to which the coarser fraction was also studied microscopically and the observation made that, e. g., the chalcopyrite and pentlandite grains had to an extent been preserved in unweathered condition. Fig. 1 shows the Ni-contents of the $\varnothing < 0.012$ mm fraction as obtained spectrographically.

The greatest contents, as will be seen, are accumulated in a fan-shaped area the diameter of which is approximately N 60°W and the tip of which points to the ore outcrop. In drawing the boundaries of the anomaly, twice the background (20 ppm), or, accordingly, 40 ppm, has been taken as the threshold value, while the largest contents are about ten times the back-

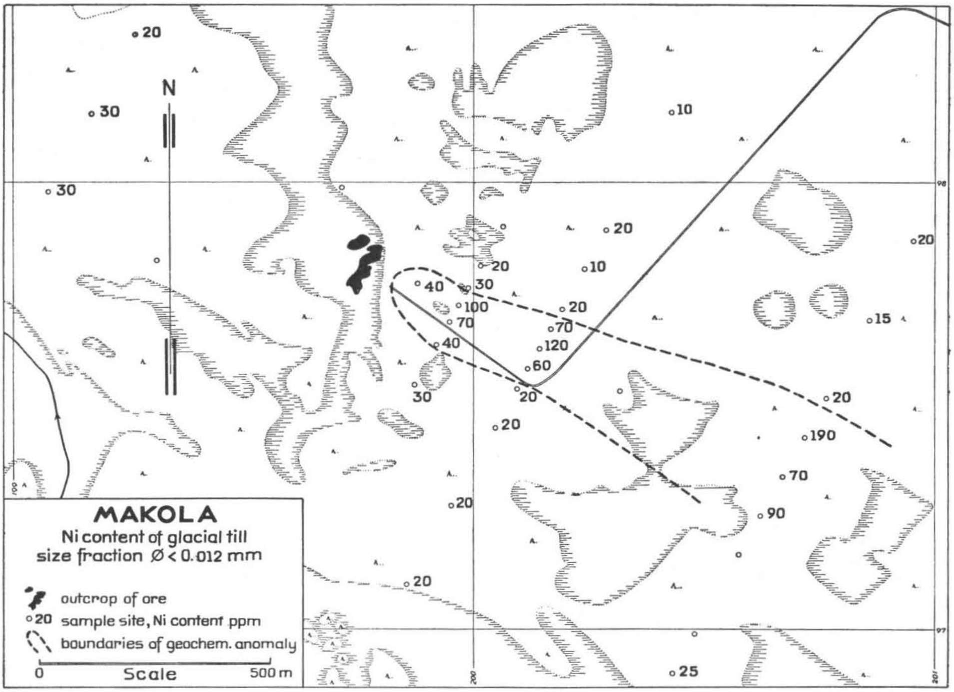

Fig. 1. Ni content of glacial till in Makola.

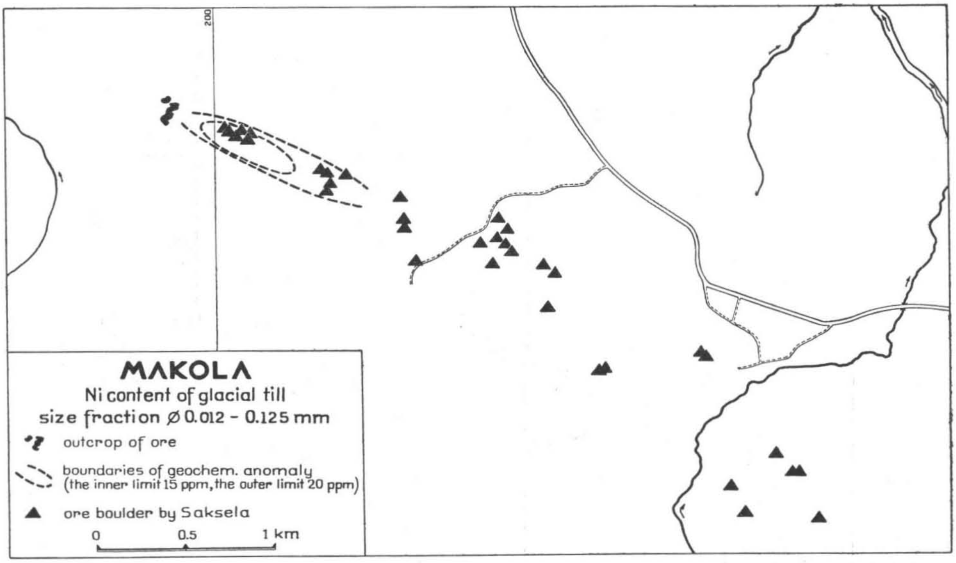

Fig. 2. The boulder fan and Ni content of glacial till in Makola.

ground value. Fig. 2 presents the Ni-contents of the fraction Ø 0.012—0.125 mm obtained on a boulder chart (Saksela, 1949) — in the figure the outer curve being 15 ppm and the inner one 20 ppm. In this fraction the background value may be considered to be 10 ppm, the highest content obtained being six times the background. In the coarser fraction the length of the anomaly that could be followed was about one kilometer, but in the finer fraction the Ni-contents are at this distance at a maximum, the anomaly thus continuing on farther.

OUTOKUMPU

The Outokumpu region is undulating, but the differences in elevation do not exceed 70 m. The overburden in the vicinity of the mine is glacial till, but it alters toward the east and the southeast to beds of sorted sedimentary materials. The pyrrhotite, pyrite, chalcopyrite, sphalerite ore of Outokumpu situated in quartzite crops out in two places to the surface of the rock underneath a moraine mantle 0—5 m thick. The average ore

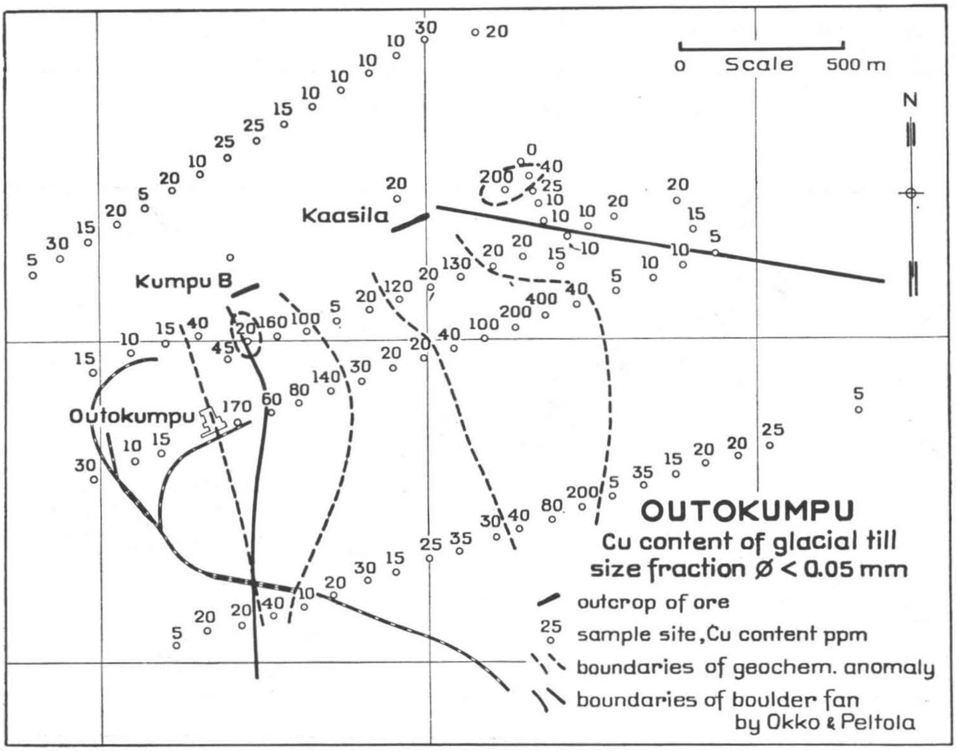

Fig. 3. The boulder fan and Cu-content of glacial till in Outokumpu.

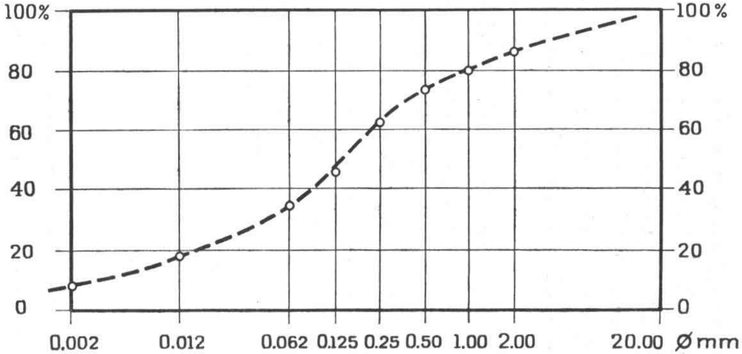

Fig. 4. Grain size analysis of glacial till, Kiuruvesi.

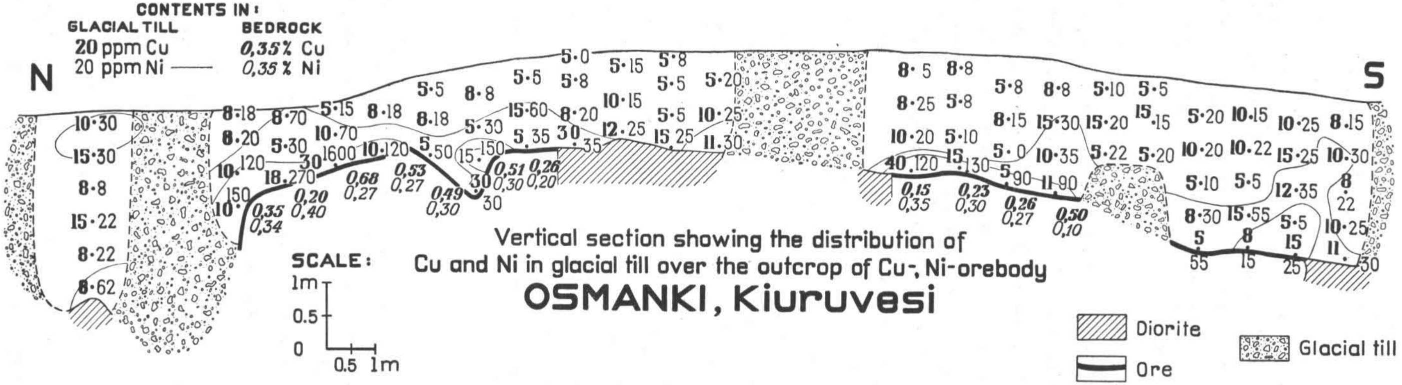

Fig. 5. The distribution of total Cu and Ni in glacial till over the outcrop of orebody, Kiuruvesi.

content is 3.71 % Cu and 1.07 % Zn (Vähätalo, 1953). The area investigated is almost wholly inhabited and accordingly the contamination danger is obvious; in the gathering of samples, however, it was endeavored to practise the utmost care and avoid the taking of contaminated till. Two main striae directions have been met with in the region, namely N 10°W and N 80°W (Repo, 1957; Okko and Peltola, 1958), of which evidently the latter is the older. Fig. 3 shows the copper contents of the grain size fraction $\varnothing < 0.05$ mm of the till.

The background value appears to be about 20 ppm; the threshold value taken for the anomaly is 40 ppm, and the highest content met with is 20 times the background value. As the figure shows, two well defined zones of anomalies were found, the western margin of which runs nearly parallel to the later direction of transport, whereas the eastern margin in each anomaly makes a bend eastward. Evidently the earlier direction of transport first carried the material, which was then hauled off by the new glacial movement along with new material torn off the rock toward the south. The anomaly running from the Kaasila outcrop is still relatively strong as far as a kilometer from the outcrop, continuing apparently farther, whereas the anomaly of Kumpu B appears to terminate about a kilometer from the outcrop. A microscopic examination of the till revealed that even the sulphidic ore minerals had been preserved without weathering, excepting the thin outher crust. The following ore minerals were identified: chromite, hematite, pyrrhotite, pyrite and chalcopyrite (Kauranne, 1951). Sulphidic ore minerals were found in a radius of only about a kilometer from Outokumpu; but oxidic ore minerals, chromite and hematite, which also apparently derived from the rocks of the Outokumpu complex were distributed across at a greater distance as well, although in small amounts.

KIURUVESI

At Kiuruvesi it was possible to investigate the dispersion of heavy metals in the overlying till, for the ore deposit had only jus been discovered and had not yet been quarried, nor had the settlement of the area contaminated the sedimentary deposits. The overburden concealing over the pyrrhotite, pentlandite, chalcopyrite ore was stripped, and it was observed that the outcrop consisted of two parts, being altogether 22 meters broad and, according to geophysical measurements, approximately 200 meters long. It was covered by bed of normal sandy till (Fig. 4), ranging in thickness from one to three meters.

Fig. 5 shows the Cu- and Ni-contents of the ore as well as the variation in the total content of the corresponding metals in the till, as analyzed colorimetrically. The samples were taken from the wall of the excavation, dried, sieved and analyzed for fraction $\varnothing < 0.05$ mm. Also the easily dissolved copper was analyzed according to Holman's (1956) method, but only in those till specimens collected from the surface of the rock could low contents be found.

Even the largest total contents — the analysis was made colorimetrically with dithizone after an HNO_3 extraction lasting one hour — are concentrated in the lowermost samples, the contents rapidly decreasing upward. The nickel appears to extend slightly farther from the surface of the rock, but even it does not extend to an anomalous degree beyond a meter. Also the diffusion is thus quite meaningsless (comp. Hawkes, 1954), even though the soil consists of normal, loose, sandy till. It is not possible to determine the exact age of the overburden, but the region has been free of the continental ice sheet for some 7 500 years (Sauramo, 1940).

SUMMARY

The examples described in the foregoing indicate that in carrying out geochemical prospecting in glaciated areas, attention should be focussed, above all, on the direction of transport of the glacial ice and ore should be sought on the side of the anomaly corresponding to the direction of advance of the ice (Hyvärinen, 1958; Kauranne, 1958) and that pedogeochemical anomalies may be dealt with just like boulder fans (Sauramo, 1924; Grip, 1953; Aurola, 1955; comp. also Hawkes, 1957, p. 278). The transport of ore material in the till took place mainly in the form of mineral grains thanks to the pushing movement of the glacial ice; the chemical diffusion has been extremely slight. Accordingly, special attention must be concentrated on the size of the grains to be analyzed as well as on the extraction methods. Table 1 presents a theoretical calculation of ball-shaped grains with the purpose of investigating the number of pentlandite grains required to achieve a given Ni-content in a 1-gram sample.

Table 1.

Size of grain $\varnothing$ mm	Amount of pentlandite grains in 1 g of sample		Total number of grains in 1 g
	1 ppm Ni	100 ppm Ni	
1	0.002	0.2	723
0.1	2	188	$723 \cdot 10^3$
0.01	1 880	$188 \cdot 10^3$	$723 \cdot 10^6$
0.001	$1\,880 \cdot 10^3$	$188 \cdot 10^6$	$723 \cdot 10^9$

As it will be seen, the finer the class of grain analyzed, the better it is (statistically), though, of course, in practise the lower limit is set by the sieves. The Geological Survey of Finland has ended up by using sieves with silk screen, the diameter of the holes of which is 0.05 mm. So far we have analyzed only the healthy, unaltered C layer of the till, for according to our experience the different metals are enriched in different ways into the upper (A and B) layers of the podsol-profile, and we do not yet know the enriching mechanism. In analyzing the total contents, we have used as the extracting means both fusion and digestion with hot mineral acids.

Acknowledgments. I should like to thank the Geological Survey of Finland for giving me permission to publish the results of this study. I am especially grateful to my chief, Dr. Aarno Kahma, who directed the work from the very start. The Suomen luonnonvarain tutkimussäätiö (Foundation for the Investigation of Finland's Natural Resources) has assisted the carrying out of the research and the development of the methods with grants. The spectrographic analyses were performed by A. Löfgren, M. A., and the figures were drawn by Miss. K. Dahl. The English translation was made by Paul Sjöblom, M. A.

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2

THE PIEDMONTITES OF PIEDMONT (ITALY), KAJLIDONGRI (INDIA), AND MARAMPA (SIERRA LEONE)[1]

BY

VLADI MARMO, K. J. NEUVONEN, and PENTTI OJANPERÄ

ABSTRACT

The piedmontites listed in the title of this paper have been examined optically, chemically, and by X-rays. The results of the examination are presented and discussed.

CONTENTS

INTRODUCTION

During the diamond drilling carried out at the iron mine of Marampa, Sierra Leone, a 50 cm thick sheared zone was pierced. The rock at this point was brownish red, distinctly sheared, and mainly composed of piedmontite and quartz associated with minor amounts of epidote and tremolite. This zone was situated in pyroxene-bearing gneiss, which is in places rich in the specular hematite, mined at Marampa.

The piedmontite of the drill core was kindly supplied by Dr. Bleek, the chief geologist of the Marampa mine, to V. Marmo for closer examination. In addition, piedmontite samples of Piedmont (Italy) and Central India

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[1] Received December 30, 1958.

were obtained from the mineralogical collections of the British Museum, London, by the courtesy of Dr. M. Hey.

The piedmontite samples were analysed chemically by Pentti Ojanperä, chemist of the Geological Survey of Finland.

An X-ray study of these minerals was carried out by K. J. Neuvonen in a search for superstructures among epidote minerals. The samples were optically studied under his guidance by Mr. A. I. Vorma at the Geological Survey of Finland. The diagrams for this paper were drawn by Miss Thyra Åberg.

The present paper was written largely by V. Marmo and its English kindly corrected by Mr. Paul Sjöblom, M. A.

CHEMICAL COMPOSITION

The piedmontites were extracted from the host rocks and purified by using the Clerici solution. The products thus obtained were quite pure.

The purified material was analysed and, thereby, methodically, the common lines of an ordinary silicate analysis have been followed. Some remarks concerning the method should, however, still be made:

1. After removal of the SiO_2, the sulphide metals were precipitated in an acid solution by H_2S.
2. After the precipitation of the sesquioxides, the manganese was removed from the solution as sulphide. Nevertheless, the manganese content of the R_2O_3, Ca, and Mg-precipitates was also checked and the respective corrections made.
3. The total iron and titanium were determined spectrophotometrically on the basis of the color of the ferric chloride complex and the color of the titanium »tiron» complex.
4. The spectrophotometrical determination of the total manganese was carried out from a separate portion.
5. The amount of Mn_2O_3 was calculated according to the oxidation ability of the powdered sample. The oxidation was determined from a separate portion using the ferrous sulphate method. For this the sample powder was disintegrated with a mixture of HCl and H_2SO_4 in a container closed by a Bunsen valve.
6. The water was determined utilizing the Penfield tube.

The results of the chemical analyses are presented in Table 1. All three piedmontites examined fit well in the general formula: $(OH)_2(Ca, Mn^{\cdot\cdot}, Na)_4\ (Al, Fe^{\cdot\cdot\cdot}, Mn^{\cdot\cdot\cdot})_6Si_6O_{24}$, but there occurs a marked variation in the ratio of $Fe^{\cdot\cdot\cdot}$ to $Mn^{\cdot\cdot\cdot}$. Furthermore, the Italian piedmontite is decidedly richer in MnO (2.0 %) than the other piedmontites examined (0.45 % and 0.51 % MnO). If, however, the total manganese content is considered, then all three samples are quite similar, containing 9.32 to 11.56 % ($MnO + Mn_2O_3$). The piedmontite of Sierra Leone is remarkable in respect to its contents of Cu (0.04 %), Pb (0.01 %), and Sn (0.01 %). Whereas the said elements are not detectable in the piedmontite from Italy by the analytical methods used. In the Indian sample, there is, however, 0.01 % Cu [1].

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[1] The occasional Cu and Pb content of piedmontites is well known from earlier studies as well. Doelter (1917) has cited two analyses of piedmontites containing still greater amounts of these metals than the piedmontites of the present paper:

1. From rhyolite of South Mountain, Pa: 0.13 % Cu—0.17 % Pb.
2. From rhyolite of Pine Mountain, Monterey, Maryland: 0.11 % Cu—0.14 % Pb.

The respective Mn_2O_3 and MnO contents of these piedmontites are: 1: 8.15 %—2.285 %; and 2: 6.85 %—1.92 %. Sn was not determined. Unfortunately, there are no optical data available for the piedmontites analysed.

Table 1. Chemical Composition of the Three Piedmontites
Analyst, Pentti Ojanperä

	Piedmontite, Piedmont, Italy			Piedmontite, Kajlidongri, India			Piedmontite, Marampa, Sierra Leone		
	wt %	cation %	Content of the Unit cell	wt %	cation %	Content of the Unit cell	wt %	cation %	Content of the Unit cell
SiO_2 ..	37.54	34.33	$2 \times \left\{ (Ca_{1.76}\ Mn_{0.14}\ Mg_{0.01}\ Ti_{0.03}\ Na_{0.02})\ (Al_{1.88}\ Fe_{0.63}\ Mn_{0.44})\ Si_{3.01}\ O_{12.0}\ (OH)_{0.78} \right\}$	37.16	33.58	$2 \times \left\{ (Ca_{1.96}\ Mn_{0.03}\ Mg_{0.02}\ Na_{0.01})\ (Al_{1.91}\ Fe_{0.40}\ Mn\ _{0.69})\ Si_{2.95}\ O_{12.17}\ (OH)_{0.97} \right\}$	36.82	33.27	$2 \times \left\{ (Ca_{1.94}\ Mn_{0.04}\ Mg_{0.01}\ Na_{0.01})\ (Al_{1.84}\ Fe_{0.49}\ Mn_{0.67})\ Si_{2.99}\ O_{11.96}\ (OH)_{1.01} \right\}$
TiO_2 .	0.54	0.38		0.04	0.03		0.07	0.05	
Al_2O_3	19.80	21.63		19.96	21.26		19.17	20.42	
Fe_2O_3	10.46	7.30		6.47	4.40		8.03	5.46	
Mn_2O_3	7.32	5.16		11.11	7.64		10.80	7.43	
FeO ..	0.00	0.00		0.00	—		0.00	—	
MnO .	2.00	1.57		0.45	0.34		0.51	0.39	
MgO .	0.08	0.11		0.17	0.23		0.04	0.05	
CaO ..	20.47	20.33		22.60	21.88		22.29	21.58	
Na_2O .	0.13	0.23		0.05	0.09		0.10	0.17	
K_2O .	0.01	0.01		0.00	—		0.00	—	
H_2O+	1.46	8.95		1.75	10.55		1.85	11.15	
H_2O-	0.04	—		0.04	—		0.04	—	
CuO ..	0.00	—		0.01	—		0.04	0.03	
PbO ..	0.00	—		0.00	—		0.01	—	
SnO ..	0.00	—		0.00	—		0.01	—	
Total	99.85	100.00		99.81	100.00		99.78	100.00	

X-RAY AND OPTICAL DATA

For the optical and X-ray determinations, both the powder (the same as was used for the chemical analyses) and single crystals were used. All the respective data are compiled in Table 2. As regards the X-ray data, all the piedmontites examined for this study are quite similar. As regards the optical properties, however, remarkable variations occur. Usually, such variations have been attributed to differences in chemical composition. For the samples under consideration, this correlation is not quite clear, as will be discussed in the following pages.

Table 2. Optical and X-ray

Locality	[1] α	[1] β	[1] γ	$\gamma - \alpha$	[3] $2V\gamma$ (observed)	$2V\gamma$ (calc.)
Italy. Ceres, Val di Lanzo, Val d'ala, Piedmont	1.751 (±0.002)	1.781 (±0.002)	1.812 (±0.002)	0.061	86 ± 2°	90.5°
India. Kajlidongri, Tabja State, State, Central India	1.742 (±0.002)	1.767 (±0.002)	1.805 (±0.002)	0.063	81 ± 2°	80°
Sierra Leone. Marampa Mine	1.756 (±0.002)	1.783 (±0.002)	1.823 (±0.002)	0.067	72 ± 3°	80°

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[1] Immersion method in sodium light. By Mr. A. I. Vorma.
[2] Pycnometer.
[3] On the universal stage.

CONTENT AND DIMENSIONS OF THE UNIT CELL

The measured unit cell dimensions are given in Table 2. The axes of the cell were so chosen that the cleavage plain is parallel to (001). Weissenberg photographs calibrated with silicon powder were used for the measurements.

From the cell volume, specific gravity, and chemical analysis, the average content of the unit cell is calculated and presented in Table 1. The calculated contents agree well with the general formula of epidote. According to the structure determined by Ito (1947) for epidote, the Al and Fe (Mn) atoms occupy octahedral positions. In addition, aluminium also occupies centers of oxygen tetrahedra. Byström (Öhman, 1950) observed that in the piedmontite from Jokkmokk, Sweden, there is an excess of trivalent manganese entering the six co-ordinate positions. This same observation is true of the three piedmontites here analysed, wherein the sum of $Fe^{\cdots}$ and $Mn^{\cdots}$ per unit cell exceeds the number of the available octahedral positions. Consequently, some of these atoms must enter into other positions in the structure. In fact, there is a corresponding deficiency of Al atoms in the unit cell, so that some iron (and manganese) evidently is located in the tetrahedral positions. The distribution of these cations in the samples analysed is

Data of Three Piedmontites

[2] $D^{4°}_{25°}$	Space group	Unit cell			
		α and γ	β	Dimensions	$a_0:b_0:c_0$
3.485 (±0.005)	$P2_1/m$	90°	115.43°	$a_0 = 8.87$ Å $b_0 = 5.66$ » $c_0 = 10.15$ » $V_0 = 459.86$ Å^3	1.568:1:1.794
3.486 (±0.006)	$P2_1/m$	90°	115.51°	$a_0 = 8.89$ Å $b_0 = 5.67$ » $c_0 = 10.22$ » $V_0 = 464.80$ Å^3	1.568:1:1.803
3.511 (±0.004)	$P2_1/m$	90°	115.6°	$a_0 = 8.89$ Å $b_0 = 5.67$ » $c_0 = 10.17$ » $V_0 = 461.86$ Å^3	1.568:1:1.794

Piedmontite from Italy	$(Fe_{0.56}\,Mn_{0.44})^{VI}\ (Al_{1.88}\,Fe_{0.07})^{IV}$
Piedmontite from India	$(Fe_{0.31}\,Mn_{0.69})^{VI}\ (Al_{1.91}\,Fe_{0.09})^{IV}$
Piedmontite from Sierra Leone	$(Fe_{0.33}\,Mn_{0.67})^{VI}\ (Al_{1.84}\,Fe_{0.16})^{IV}$

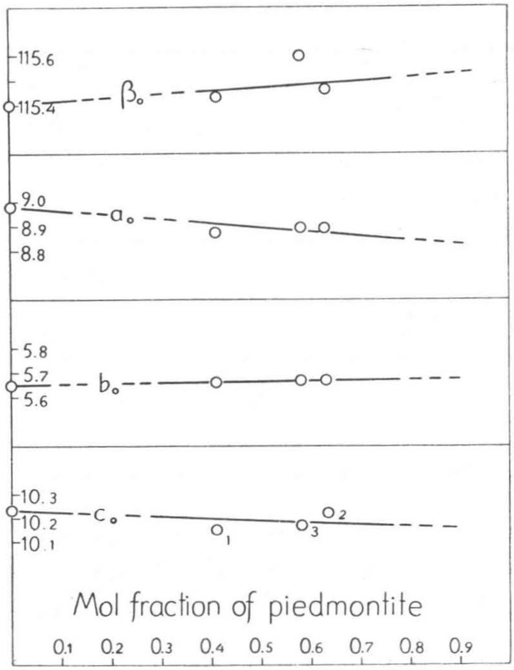
If it is assumed that in pure clinozoisite, pistacite, and piedmontite, the octahedral position is occupied by Al, Fe, and Mn, respectively, the analysed piedmontite samples may be held as solid solutions of pistacite and piedmontite. The mol fraction of the piedmontite end member in the mineral is given by the cation ratio

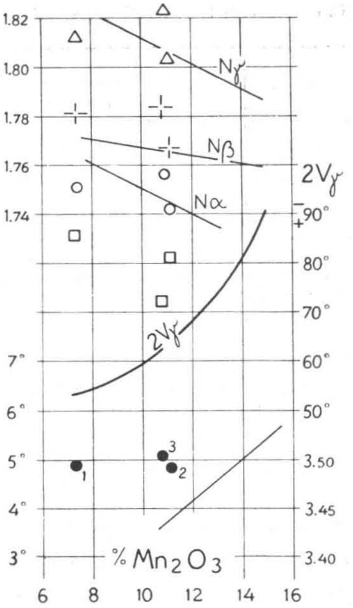
$$\frac{Mn^{\cdots}}{Fe^{\cdots}+Mn^{\cdots}}.$$

In Fig. 1 the unit cell dimensions are plotted against mol fraction of piedmontite in the samples analysed. A straight line connects the measured values to those given for pistacite by Ito (1947). The dimensions measured from piedmontite from Jokkmokk by Byström (Ödman, 1950) do not agree with the values now measured. This might be caused by a different choice of a_0 and c_0 axes. The dimensions of the Italian piedmontite tend to be low, which might be due to a high amount of $Mn^{\cdot\cdot}$ in this mineral.

OPTIC AXIAL ANGLE

According to Short (1933), when Mn_2O_3 and Fe_2O_3 are present in approximately equal quantities, the mineral piedmontite is optically positive, and it should have $2V\gamma$ between 50° and 60°. When the manganese content grows, the optical axial angle will rapidly increase. In Fig. 2, his curve of



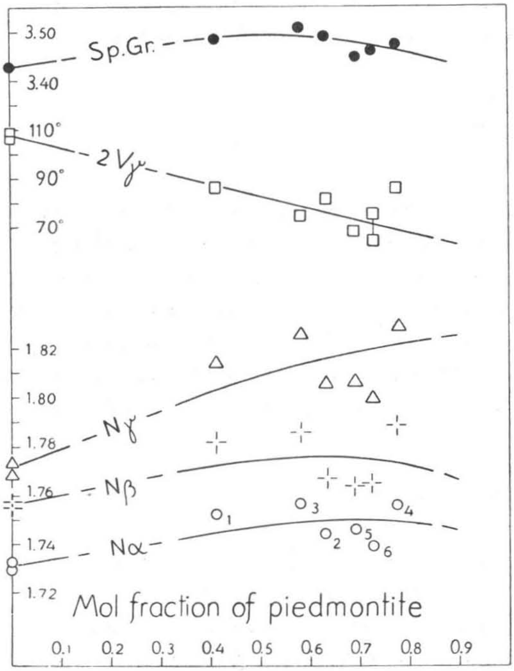


↖ Fig. 1. Unit cell dimensions in the pistacite ($HCa_2FeAl_2Si_3O_{13}$)-piedmontite ($HCa_2MnAl_2Si_3O_{13}$) series.

Numbers in the diagram refer to the samples studied as follows: (1) Italy, (2) India, (3) Sierra Leone, (4) Långban, (5) Tunaberg, and (6) Shadow Lake.

↑ Fig. 2. The refractive indices, optical axial angle, and the specific gravity of the samples analysed on the diagram of Short (1933).

○ = α, + = β, △ = γ, □ = 2V, and ● = Sp. gr. The sample numbers the same as in Fig. 1.



← Fig. 3. Variation of the specific gravity and optic properties in the pistacite ($HCa_2FeAl_2Si_3O_{13}$)-piedmontite ($HCa_2MnAl_2Si_3O_{13}$) series.

○ = α, + = β, △ = γ, □ = 2V, and ● = Sp. gr. The sample numbers the same as in Fig. 1.

2V is reproduced. If the values of the piedmontites discussed in the present paper are plotted on the same graph, there appears a marked discrepancy between the new values and the curve of Short. There the relationship between 2V and the Mn_2O_3-per cent is exactly opposite: The decreasing optical angle is irregularly parallel to the increase of the Mn_2O_3-content.

For the common epidotes, a relationship between 2V and the Fe_2O_3-content has been suggested by Winchell (1951). The 2V-values measured for the now analysed samples, however, do not agree with this curve either. In Fig. 3 the measured 2V-values are plotted against the mol fraction of piedmontite. The optical axial angle around γ decreases together with the increasing tenor of the manganese. The optic axis measured by Short (1933) for piedmontite from Shadow Lake, California, is plotted in the same diagram and fits well the curve drawn. The 2V measured by Malmqvist (1929) for piedmontites from Långban and Tunaberg, Sweden, agree also fairly well with the curve in the diagram. The sample from Långban is exceedingly rich in manganese and does not strictly belong to the piedmontite-pistacite series, which might explain the high value of $2V\gamma$ of the mineral.

REFRACTIVE INDICES

According to Winchell (1951), the refractive indices of the epidote increase together with the iron content of the mineral. From the data given by Larsen and Berman (1934) Short deduced that »it is very probable that Ng and Np curves converge with an increase of Mn_2O_3 but not enough data are available to justify a definite statement». In Fig. 2, the curves of Short are reproduced and the values of piedmontites treated in the present paper are inserted in the graph. There the Sierra Leonean piedmontite clearly stands out from the curves proposed by Short. The piedmontites of Italy and India, on the other hand, fit the curves of Short quite well.

In Fig. 3 the refractive indices are plotted against the mol fraction of piedmontite, including the values of pure pistacite and those of the two Swedish samples of Malmqvist (1929) as well as those of Shadow Lake (Short, 1933). The distribution of the plots is quite uniform in the graph for all the refractive indices, α, β, and γ. The values of the samples from Italy, Sierra Leone, and Långban are high and the indices of the samples from India, Tunaberg, and Shadow Lake are all low. It is notable that no such tendency could be seen in the variation of the optic angle.

SPECIFIC GRAVITY

The highest value for the specific gravity of the piedmontites cited by Winchell is 3.47 (6.43 % Fe_2O_3 and 22 % (Mn_2O_3 + MnO)). Short reported a value of 3.55 (approx. 14.5 % Mn_2O_3); Larsen and Berman gave a value of 3.47.

For the piedmontites of Italy, India, and Sierra Leone, the specific gravitites are: 3.485 (10.46 % Fe_2O_3 and 9.32 % (Mn_2O_3 + MnO)); 3.486 (6.47 % Fe_2O_3 and 11.56 % (Mn_2O_3 + MnO)); and 3.511 (8.03 % Fe_2O_3 and 11.31 % (Mn_2O_3 + MnO)). According to Short, the specific gravity increases together with the Mn_2O_3-content. This tendency, although not very convincing, appears in the piedmontites dealt with in the present paper, too (Fig. 2). According to Winchell, the specific gravity of epidotes increases with the content of Fe_2O_3. Consequently, a more evident parallelity between the specific gravity and the total of (Fe_2O_3 + Mn_2O_3 + MnO) should be expected. The piedmontites of Italy, India, and Sierra Leone are in good agreement with such an assumption. Likewise, when plotted against the mol fraction of piedmontite, a good mutual agreement is found, as seen in Fig. 3. The values of the piedmontite samples from Shadow Lake, Tunaberg, and Långban harmonize well together with those now measured.

DISCUSSION

The somewhat surprising feature revealed by the present study is that the optic properties of the piedmontites of Italy, India, and Sierra Leone do not follow the chemical composition, as they should do in the light of the curves of Short and Winchell. These deviations may be well understood: in the composition of the piedmontites, there are too many factors affecting the optic properties of these minerals. If the epidote proper is concerned, Al_2O_3 and Fe_2O_3 may be taken as the only variable constituents of the mineral. Therefore, in the series indicated by the end members of $HCa_2Al_3Si_3O_{13}$ and $HCa_2FeAl_2Si_3O_{13}$, a distict correlation between the chemical composition, the respective optical properties, and the specific gravity can be found. The diagram of Winchell (1951) mainly represents this series. The same would probably work for the hypothetical series between the end members of $HCa_2Al_3Si_3O_{13}$ and $HCa_2MnAl_2Si_3O_{13}$ as well. By means of the samples analysed, an approximative correlation between the chemical composition and unit cell dimensions, optical properties, and specific gravity was evaluated and presented in Figs. 1 and 3 for the series pistacite-piedmontite ($HCa_2FeAl_2Si_3O_{13}$—$HCa_2MnAl_2Si_3O_{13}$). The curves in the diagrams are drawn to agree with all the points plotted, although some of the samples considered do not strictly belong to the pistacite-piedmontite series. Members of this series should have the cation ratio Al: ($Fe^{\cdots}$ + $Mn^{\cdots}$) = 2, but the samples from Långban (4) and Sierra Leone (3) deviate greatly from this requirement, having too high a content of iron and manganese. This fact explains the high indices measured in these samples. In the same manner the indices are high in the Italian piedmontite because of the deviation from the series owing to the high $Mn^{\cdot\cdot}$ content. The series is evidently best represented by the samples from India (2), Tunaberg (5) and Shadow Lake (6).

In addition to the chemical composition, the optical properties might be affected by differences in the mutual distribution of the atoms in the structure of the mineral, caused by contrasts in the thermal prehistory. No superstructures were detected in any of the piedmontites examined. Consequently, possible structural dissimilarities cannot be of the order-disorder type. The distribution of the atoms Al, Fe, and Mn in the six and four co-ordination positions in the structure offers, however, many more possibilities. Only more elaborate x-ray studies would give information about such differences in the minerals considered.

According to the structural requirements the formula of the epidote minerals is taken as HCa_2 (Al, Fe, Mn) $Al_2Si_3O_{13}$. This gives at least two aluminium atoms per formula or four per unit cell. As mentioned earlier, however, this postulation does not hold and types with an excess of $Fe^{\cdot\cdot\cdot}$ ($\pm Mn^{\cdot\cdot\cdot}$) exist. Consequently, end members of the type $HCa_2Fe_2AlSi_3O_{13}$ or $HCa_2Fe_3Si_3O_{13}$ have to be accepted. Both these formulas have been used. The use of the latter formula, however, is not justified either from a structural or chemical point of view. The end members $HCa_2Al_2AlSi_3O_{13}$, $HCa_2Fe_2AlSi_3O_{13}$, and $HCa_2AlSi_3O_{13}$ were not applied in this discussion since only a small part of the atoms Fe and Mn could be assumed to occupy tedrahedral positions in the minerals analysed. Accepting more aluminium-rich end members, it was possible to handle the specimens as representatives of the binary system pistacite-piedmontite, instead of a more complex ternary system including clinozoisite as one of the end members.

Compiling iron and manganese, the properties of the analyzed piedmontites could be demonstrated in the series clinozoisite—mangano-pistacite, ($HCa_2(Fe, Mn)_2AlSi_3O_{13}$). Thereby it appears that the effect of iron and manganese upon the optical properties is not similar, however. Therefore the authors wanted to use the new data measured to trace the variation of the properties in the series pistacite-piedmontite, e. g., to ascertain the difference between the effects of iron and manganese. It is hoped, that when more data is available, the variation can be explored in greater detail, to provide a basis for the preparation of a diagram of the ternary system as well.

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ON THE LATEGLACIAL FROST PHENOMENA IN SOUTHERN FINLAND [1]

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ABSTRACT

This paper deals with disturbances met with in the stratified drift in Southern Finland. These disturbances, called involutions, are caused by ancient frost-action during the Late-Glacial. The permafrost is a prerequisite to the formation of involutions. Some examples of the sequence of the superficial deposits are presented.

CONTENTS

INTRODUCTION

Knowledge of the influence of frost phenomena on superficial deposits is rather scarce in Finland. In addition to the papers of Helaakoski (1912) and Kokkonen (1926, 1930) little information exists in the literature on the recent influence of frost. Hult (1887) mentions recent solifluction in Lapland and Blüthgen (1942) reports that recent frozen soil is found

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[1] Received February 4, 1959.

in Finland only above the 800-meters contour line. More extensive is the literature dealing with the effect of frost on peat deposits (e. g. Auer, 1920; 1927; Cajander, 1913; Tanttu, 1915).

The literature discussing the recent and/or ancient influence of frost on superficial deposits is fairly extensive elsewhere in the world. The investigations concerning ancient frost-action are mainly restricted to the areas situated outside the last glaciation limit, but so near it, however, that the direct effect of the inland ice on the superficial deposits can be established in many ways. The phenomena that occurred near the border of the Pleistocene ice sheet and restricted mainly to the areas outside the extreme ice border are called »periglacial.» The deformation of the layers of sorted material by the frost-action plays an important role among the periglacial phenomena. Fossil frozen ground is found e. g. in Denmark, chiefly outside the limit of the last glaciation (Nørvang, 1942). In the Netherlands van der Hammen and Maarleveld (1952) have discovered fossil frost phenomena from the Younger Dryas period. It has been suggested that deformations in the superficial deposits caused by the periglacial frost-action could not be found in the area of the latest glaciation (e. g. Poser, 1947, p. 5). Ahlmann (1947, p. 129) and Högbom (1914,p. 376) point out that periglacial frost phenomena could not have occurred in Scandinavia during the retreat of the glacier because the climate had already lost ist high polar features. Recent frost phenomena are met with, however, not only in the Scandinavian mountains (Högbom, 1914, 1926), but also in Gothland (Du Rietz, 1925; Hesselman, 1915), and even in the tropics (Troll, 1947, pp. 600—613). Recently Johansson (1956) reported fossil frost phenomena from Southern Sweden, too.

The deformations caused by the ancient frost-action and called »involutions» by Bryan (1946), Denny (1936), Schafer (1949), Sharp (1942) and others appear most distinctly in stratified and washed material. This is due above all to the structure of these deposits. The involutions become more distinctly evident in the alternation of the layers with a different grain-size from that in clay or in till, in which the structure and grain size are more homogenous.

Some examples of the involutions in the stratified drift in Southern Finland caused mainly by the ancient frost-action will be described in this paper. The observations in question are made in connection with other investigations. Thus they do not include a systematic presentation of the deformations of the sequence of Finnish superficial deposits, but rather scattered observations like sample tests in the areas where the writer has undertaken investigations and mappings. The observations were made in the years 1950—1954 and they are restricted to Southern Finland, mainly between the towns Helsinki and Tampere.

DESCRIPTION OF THE OBSERVATION LOCALITIES

JÄRVENPÄÄ

The involution deposit of Järvenpää is among the most distinct and most characteristic ones I have seen in Finland. The gravel pit where the involutions occur is situated in the rural town of Järvenpää near the Kaitaranta farm (Fig. 1, loc. 1). Figure 2 presents the scheme of the sequence of layers

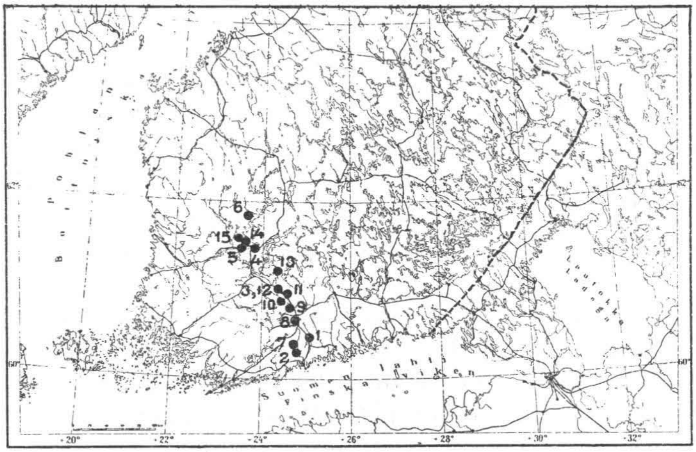

Fig. 1. Observation localities in Southern Finland.

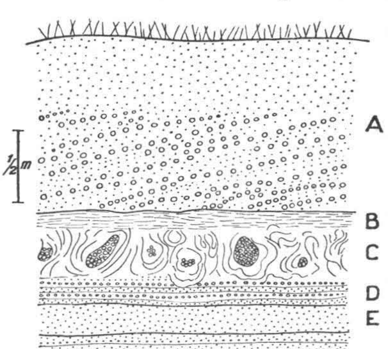

Fig. 2. Schematic picture of the involution layer in Järvenpää. A = littoral accumulation, B = fine sand, C = involution layer, D = sand and fine sand, E = coarse sand, F = stratified sand and fine sand, G = stratified gravel and coarse sand.

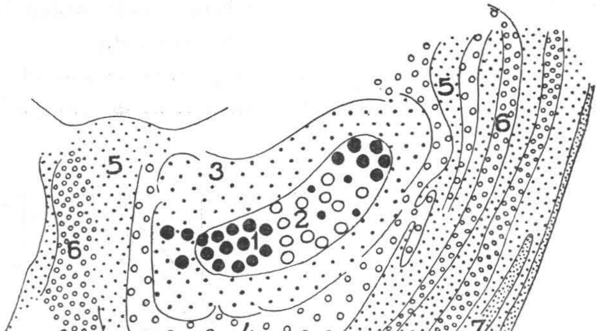

Fig. 3. Detailed picture of one involution in Järvenpää. 1 = small stones, 2 = gravel, 3 = coarse sand, 4 = medium sand, 5 = fine sand, 6 = very fine sand, 7 = silt.

in the observation locality, and Fig. 1 (Plate I) the general view of the southern border of the sand pit where involutions are met with. The surface of the earth inclines gently to the southeast, at most 3—4 degrees. The height of the locality is about 67—68 m above sea level.

A layer of coarse gravel 1.3 m thick lies topmost in the section. The upper part lacks bedding. Layers of coarse sand, gravel and small stones gently inclined to the southeast form the lower part of the bed (Fig. 2,A; Fig. 1, Plate I).

A strongly disturbed layer about 0.7 m thick is situated below the bed A and sharply separated from it (Fig. 2, B—D). Three parts can be distinguished in this bed. A layer of fine sand about 15 cm thick with a very fine microstructure is situated topmost (Fig. 2, B; Plate I, Fig. 2, upper part). The true involution layer of a thickness of 35 cms is situated below this. It consists of strongly deformed and disturbed, bag-like or bulbous involutions of fine sand, sand and gravel (Fig. 2, C; Plate I, Fig. 2, the middle part). This horizon is underlain by layer D, about 20 cm thick (Fig. 2; Plate I, Fig. 2, the lower part), consisting of several horizontally alternating layers of sand and fine sand.

Coarse sand without bedding to a thickness of 20 cm is situated below the involution layer, being quite distinctly distinguished from it (Fig. 2, E; Plate I, Fig. 2, lower border). A layer F about 60 cm thick (Fig. 2) lies below this level, consisting of several horizontal layers of sand and fine sand. The thickness of the layers alternates from 0.5 to 5 cm. The grain size increases in each layer downwards.

Finally coarse sand and gravel is situated lowermost in the section, continuing at least 2—3 m downwards and probably still deeper (Fig. 2, G).

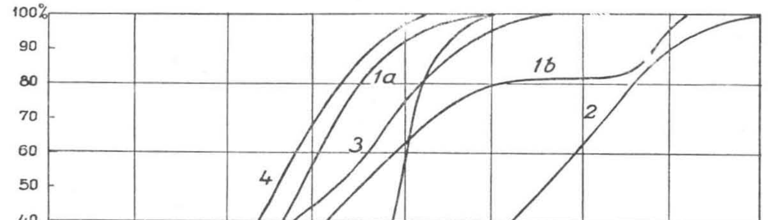

Fig. 4. Cumulative curves of grain size of some involutions. 1a = Järvenpää, below the involutions, 1b = Järvenpää, the involution layer, 2 = Seutula, 3 = Messukylä, Vilusenharju, 4 = Kuru, 5 = Ylöjärvi.

It is distinctly to be observed that in the deformed layer coarser material — sand, gravel, even small stones — is situated almost regularly in the middle part of the involutions (Fig. 3). The layers covering bulbously the interior of the involutions become outwards little by little still finer. The grain size changes thus from silt to coarse gravel. The cumulative curves of the grain size show that the material of the involutions is more poorly sorted and contains coarser material than the layers above and below it (Fig. 4). The parts of the involution bags containing the finest grains are usually turned upwards, thus indicating an intrusion of material from the lower layers. A slight orientation of the involutions can hardly be observed in the direction of the surface slope. As a whole the involutions indicate movements, which are largely vertical. This is shown by the fact that in the vertical sections at right angles to each other the involution bags appear quite similar. If the sliding process or similar horizontal movements had caused these involutions, it should be visible by the dissimilarity of the sections in different vertical planes.

The size of the involutions changes to a considerable extent. The height of the greatest involutions is the same as that of the entire deformed layer while the smallest involutions occur one on top of the other, several at a time. The breadth of the involutions alternates from 20 to 50 cm.

Involutions are met with in the wall of the sand pit at a distance of only about 20 meters, although it seems that similar superficial deposits and inclination relations continue in the locality. The amount and width of the involutions decrease toward the sides.

The emergence of the involution layer from the Baltic Sea may be determined on a large scale. A well developed ancient wave-cut cliff exists ca 63—64 m a. s. l. about 4 m below the involutions. According to recent investigations (Mölder, Valovirta, Virkkala, 1957) this cliff corresponds most nearly to the limit between the Younger Dryas and the Preboreal. In the vicinity of Helsinki this shore-line is the best-developed ancient shore feature, according to Hyyppä (1937), too.

SEUTULA

A small gravel pit is situated by the side of the road from the village of Vantaa to Seutula, about 2 km north of the Nurmijärvi highway (Fig. 1, loc. 2). Distinct deformations of the sequence of layers are met with in the rather coarse material of the gravel pit (Fig. 3, Plate I).

The character of the deformations resembles that of the former observation locality. The upward bending of the layers indicates that material has been heaved up by intrusion from the lower layers into the overlying beds. Most of the coarse grains in the material, small pebbles and gravel, are oriented according to the layers. Their long axes are nearly vertical in the upright layers. The deformed layers seem in places to form a true frost crater (Fig. 4, Plate I), similar to those described by Johnsson (1956) in Southern Sweden.

The grain size composition of the deformed layer ranges from medium sand to coarse gravel, and even to small pebbles (Fig. 4). Medium sand forms, however, the main part. Gravel and small pebbles are situated in it as smaller or larger inclusions. Thin beds of fine sand are found between the layers of coarser material.

The terrain of the observation locality is quite even, its elevation being about 55 m. The interpolated height of the Ancylus shore-line is, according to Hyyppä (1937, Appendix), about 55 m a. s. l. The height of the gravel pit thus corresponds to the Baltic level of the beginning of the Ancylus period.

HÄMEENLINNA, VUORENTAKA

In the village of Vuorentaka, near the town of Hämeenlinna, there is a large gravel pit (Fig. 1, loc. 3). Topmost, there occurs here stratified coarse gravel to a depth of about 2.5 m, obviously a littoral accumulation. Below it lies nearly two meters of fine sand with distinct involutions in the upper part (Figs. 5 and 6, Plate I).

The thickness of the involuted layer is here only 20—40 cm. The breadth of the separated involution bags changes from 5 to 20 cm. The involuted layer extends in the section only about 20—25 meters.

The separated involutions form quite distinct bulbous bags in which outer layers concentrically surround the middle parts of the involutions. Their material is thoroughly fine sand. If the section is made in great features in the middle part of the bag, a thinner neck continues upward from the bulb, as in Fig. 5, Plate I. In other cases the involutions appear as concentric circles.

Horizontal sections differ considerably from the vertical ones (Fig. 6, Plate I). More or less concentric arcs do not indicate any one-sided horizontal force, such as sliding, ice advance or earth creep.

MESSUKYLÄ, VILUSENHARJU

Quite distinct involutions are found in the great gravel pit in the esker of Vilusenharju, which rises about 30 m above its environment. The present surface of the locality slopes about 2—3 degrees and its elevation is 125 m (Fig. 1, loc. 4).

The sequence of the layers of the observation point is the following:
0—180 cm, coarse and stony gravel,
180—230 cm, varved clay,
230—410 cm, involution layer,
410 cm and downwards, stratified medium sand.

The thickness of the involution layer, gravel and clay decreases towards the sides. The sequence of the superficial deposits about 25 m from the former observation point is in the same wall of the gravel pit as follows.
0—70 cm, coarse gravel with stones,
70—100 cm, fine sand and silt corresponding to the varved clay of the former observation point and changing gradually into it,
100—190 cm, involuted layer overlying stratified sand as at the former observation point.

The material of the deformed layer is mainly fine sand and sand (Fig. 4, 3). The entire layer is filled by bends and folds, bags and lenses of various forms and stages. The breadth of the separate involutions ranges from a few centimeters to about 40 cm. Their height is commonly half their breadth. The interiors of the larger involutions consist of smaller bags of some centimeters in size. The inner parts of the involutions are composed of sand, the outer parts and surroundings of fine sand. The involutions appear quite similar in the vertical planes. They are distinctly caused by forces exerted vertically. Fig. 1, Plate II, represents a characteristic detail of the involution layer.

The involutions are visible in the wall of the pit at a distance of about 30—40 m. The thickness of the involution layer decreases towards the sides. At the same time the variableness of the forms and the density of the involutions become less distinct. About 50 m from the first observation point the involuted layer turns into ripple-marked stratified sand and fine sand lacking all disturbances.

TAMPERE, VILLILÄ

A large gravel pit is situated about 9 km from the center of the town of Tampere toward Nokia ca 300 m north of the highway and on top of an esker (Fig. 1, loc. 5). Distinct involutions are met with in the northern wall of the pit rather deep under the surface of the earth and below varved clay. The terrain slopes on the observation locality about 5—6 degrees and its elevation is ca 120 m.

The sequence of the superficial deposits is as follows.
0—50 cm, coarse gravel with small pebbles,
50—170 cm, varved clay,
170—460 cm, stratified sand with alternating darker and lighter sand beds and with thin silt layers,

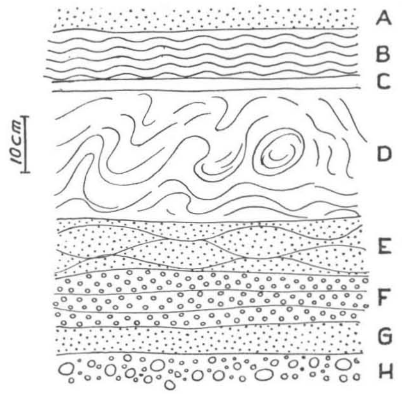

Fig. 5. Schematic picture of the involutions in Villilä, Tampere.

460—520 cm, involution layer composed mainly of sand and fine sand,
520 cm and downwards, chiefly coarse glaciofluvial material with 5—10 cm layers of sand, gravel, and pebbles. The deposit continues to a depth of at least 10—12 m.

A detail of the involutions and of the layers lying nearest to it is shown in Fig. 2, Plate II. The sequence of the layers is in Fig. 5 as follows.

A = the lower part of stratified sand,
B = ripple-marked stratified sand changing gradually into horizontal stratified sand,
C = a thin silt layer,
D = the true involuted layer consisting of several bags and folds the interiors of which are composed of light, medium sand and the outer parts and spaces between the involutions of darker fine sand. The border of the deformed layer is sharp on top, but gradual below.
E = ripple-marked stratified, dark, fine sand,
F = stratified gravel and sand,
G = fine sand and medium sand,
H = underlying coarse glaciofluvial gravel and sand.

Involutions are met with at least at a distance of about 10—15 m in the wall of the pit. It is possible that they are much greater in extent, but collapsed material prevents further observations.

KURU

The observation locality is situated about 1 km south of the church of Kuru alongside the Tampere highway (Fig. 1, loc. 6). The terrain at the observation point seems to be quite even and the elevation of the locality is about 130—135 m.

The sequence of the layers is here as follows.

0—70 cm, coarse sand and gravel,
70—90 cm, homogeneous sand, lacking stratification,
90—100 cm, coarse sand and fine gravel,
100—110 cm, silt,
110—160 cm, fine sand with very fine bedding,
160—200 (230) cm, involuted layer of fine sand (Fig. 3, Plate II),
200 (230)—350 cm, fine sand and coarse silt,
350 cm and below it, stratified material with alternating layers of sand and fine sand.

Involutions here form closed or, at the upper end open bags and lenses situated in places next to each other. The height of the bags ranges from 3 to 10 cm and the breadth from 4 to 20 cm. They have commonly a bulbous structure the interior consisting of coarser material, usually of medium sand. On the outer parts of the involutions are situated some layers of fine sand which are darker than the interior. The involutions appear as concentric circles in cases where the section is not situated in the middle part of the involutions. Likewise the involutions in the section of the horizontal plane are composed of short and bent, concentric arcs of the circle.

The space between the involutions is filled up with fine sand similar to the outer parts of the involutions. The smaller involution bags form larger complexes with a bag-like form, too. They range from 0.5 to 1.2 m in breadth (Fig. 6).

The lower limit of the involutions is not distinct, but bag-like bulges have intruded from the overlying deformed sand into the underlying undisturbed layer of fine sand.

Rather large, sausage-like lenses of sand about 30—40 cm long and 10 cm high are found at one point in disturbed sand (Fig. 7). Fig. 4, Plate II, presents a detail from the structure of the involutions, too.

The grainsize analyses of the involutions do not essentially differ from those of the overlying and underlying layers (Fig. 4,4).

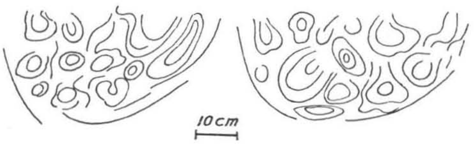

Fig. 6. Complex involutions in Kuru.

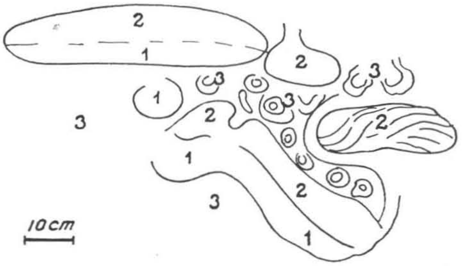

Fig. 7. Sausage-like lenses of the involution layer in Kuru. 1 = fine gravel and coarse sand, 2 = medium sand, 3 = fine sand.

The involuted layers are met with only on the northern wall of the gravel pit at a distance of about 15 m, although similar superficial deposits continue elsewhere, also. Nor do they appear continuously in this section, but in groups, at short distances of 1—2 m.

Investigation of the deformed layers shows that the involutions are caused by vertical movements. The material of the involutions appears as if it had been intruded from the underlying layers upwards.

OTHER OBSERVATION LOCALITIES

Traces of ancient frost-action are met with in numerous other observation localities, too. In the following they are listed very briefly. Only the most important features can be mentioned here.

Nurmijärvi, Paraatinkalliot (Fig. 1, loc. 7), small involutions and deformed lenses in fine sand at a depth of 1.1—1.7 m (Fig. 5, Plate II). The slope of the locality is about 2 degrees and the elevation 75 m, which corresponds nearly to the limit between pollen zones III and IV (Mölder, Valovirta, Virkkala 1957).

Riihimäki, Herajoki (Fig. 1, loc. 8), involution bags composed of sand and fine sand are found in compact fine sand at a depth of 3.1—3.8 m, mainly below sand and gravel deposits, but also below a clay layer of a thickness of about 0.2—0.3 m (Fig. 6, Plate II). The terrain slopes of the observation point about 6—7 degrees and its elevation is 110 m.

Tervakoski, Melkkola (Fig. 1, loc. 9), an indistinctly and irregularly deformed layer of sand consisting of clear involution bags and bent or distorted beds is situated at a depth of 1—1.5 m below thin sand-, gravel- and clay layers. The surface of the earth inclines about 8 degrees and its elevation is 100 m.

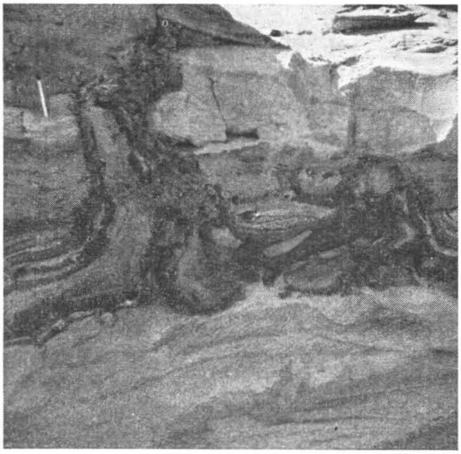

Fig. 8. Sandy wedges of the involution layer in Harjunmäki, Janakkala.

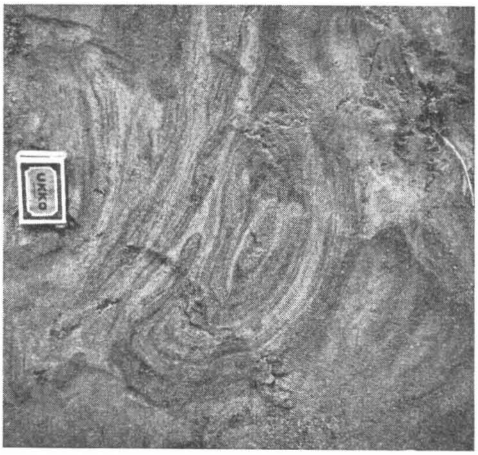

Fig. 9. Detail of the involutions in Pälkäne.

Janakkala, Harjunmäki, the southern shore of Lake Kernaalanjärvi (Fig. 1, loc. 10). Involuted layer is situated 2.5—2.65 m under the surface. It is composed of silt, fine sand, and sand. Small involution bags and wedges occur in the deformed layer. The deformations appear partly as sandy wedges penetrating obliquely upwards from the underlying beds (Fig. 8). The inclination of the terrain is about 3—4 degrees and its elevation about 100 m.

Hämeenlinna, Vanaja (Fig. 1, loc. 11), involution bags and wedges in compact fine sand. The slope is about 5—6 degrees and the height about 110 m a. s. l. The involutions are met with 1.9—2.0 m below the surface.

Hämeenlinna, Ahvenisto esker (Fig. 1, loc. 12), involutions at a depth of 2.0—2.4 m on the steep eastern slope of the esker. Plenty of small, 5—20 cm high involution bags consisting of silt and fine sand are situated in the deformed layer. The elevation of the locality about 120 m.

Pälkäne (Fig. 1, loc. 13) village, site of church, clear involutions which incline partly in the direction of the slope are met with in a small sand pit about 110—115 m a. s. l. and at a depth of 0.7—1.5 m (Fig. 9). The slope of the esker has evidently been an important factor in the origin of the disturbances.

Tampere, Lielahti (Fig. 1, loc. 14), indistinct involutions, lenses and bags of fine sand are met with underlying littoral gravel of a thickness of 0.3 m and below varved silty clay 0.6 m thick. The locality is situated on the northern slope of a great esker. The surface slopes about 5 degrees and the elevation of the locality is 115—120 m.

Ylöjärvi, Pinsiönkangas (Fig. 1, loc. 15), some rather large involution bags and folds exist in the primary glaciofluvial material of a depth of 1.2—1.9 m. The earth's surface has here a slope of 1—2 degrees and the elevations is 147—148 m. Material is very well sorted sand and fine sand (Fig. 4, 5).

Karvia, Kantinkylä, the esker Pohjankangas (Fig. 1, loc. 14), very distinct involutions in fine sand at a depth of 1.6—1.8 m. The terrain is quite even, about 140—150 m a. s. l.

Petäjävesi, 1.3 km northwest from the railway station toward the parish of Multia (Fig. 1, loc. 15), distinct involution bags in fine sand on top of a small esker. The elevation of the locality is about 110 m.

ORIGIN OF THE INVOLUTIONS

Disturbances in the sequence of the superficial deposits similar to the afore-described examples can be caused by different agencies. The pressure and the pushing of the inland ice may be mentioned first among the possible causes of these disturbances (Slater, 1925, 1927). In such a case, however, a distinct orientation of the disturbances and a fold axis should become visible in two vertical sections at right angles to each other (Johnsson, 1956; Sharp, 1942). As mentioned in the foregoing such a fold axis could not be noticed in the examples described. The most thoroughly investigated instances rather indicate a movement in the vertical direction. Where a more or less distinct orientation of the disturbances appears, this is rather in the trend of the greatest slope and not in the direction of the ice movement. No evidence exists that the disturbances were caused by the pressure of the overlying deposits, either (Horn, 1912; Kindle, 1917; Miller, 1922).

Disturbances of this kind have been explained to be due to subaqueous sliding, too (Nørvang, 1946; van Straaten, 1949). These are generally, however, distinctly oriented and in such disturbances a distinct sliding axis ought to be evident. The subaqueous sliding reveals that the movements occurred rather in the horizontal than in the vertical direction. Sliding cannot, however, be wholly excluded from among the causes of the said disturbances. On the contrary, it is probable that subaqueous sliding has occurred at least in the locality of Pälkäne (No. 13) and possibly elsewhere, also, where the inclination of the slope is rather considerable.

The most common reason for the afore-described disturbances is, however, repeatedly and irregularly melting and freezing ground (e. g. von Bahr, 1932; Poser, 1947; Richmond, 1949; Schafer, 1949; Sharp, 1942; Steeger, 1944; Troll, 1944). The influence of frost may have been the main cause of the disturbances of the sequence described in the foregoing.

The presence of abundant water is essential to the formation of the frozen ground. The water can derive from below, where it can be raised farther upward by the capillarity (Högbom, 1905; Lundqvist, 1948; Taber, 1929). The grain size analyses show that a rather large part of the material of the involutions has a diameter less than 0.1 mm. The capillarity increases noticeably in the grain sizes smaller than this. Very thick and coarse-grained deposits exist below the involution layers, however. The present ground water table in these deposits thus lies so deep that the rising of the capillary water from below is impossible. The beds underlying the involutions must thus have been impermeable to water or the ground water table must have been noticeably higher than at present. The topography of the most of the afore-described examples shows, however, that the last-mentioned alternative can be excluded. The only factor which in these circumstances could cause an impermeable base for water seems to be the permafrost. If the conditions during the fossil frost-action are considered, the permafrost in the ground should not at all seem impossible. The continental ice had recently retreated from the region under consideration. Although the air temperature must have been relatively high—otherwise the ice should not have retreated so fast as it really did — the ice temperature must have been very near to zero centigrade (cf., e. g., Ahlmann, 1947). Further pieces of dead ice had become buried in the superficial deposits, where they could have been preserved for a long time (Alden, 1924; Brigham, 1929; Flint, 1932). Even at the present day ice, in all probability from the Ice Age, has been found in the glaciofluvial deposits of Finland (Leiviskä, 1912). It is thus under these circumstances quite plausible that permafrost could have existed in the ground immediately after the retreat of the continental ice.

Water necessary for the frost-action can partly be derived from the rain water seeping through the superficial layers. An impermeable ground is prerequisite in this case; otherwise the water would have penetrated the underlying coarse-grained beds. The existence of the permafrost is required in this case, also.

The permafrost is considered by many investigators to be the primary condition in the origin of the afore-described disturbances in the sequence of the superficial deposits (e. g., von Bahr, 1932; Edelman and others, 1936; Gripp, 1927; Krekeler, 1929; Orwin, 1943; Poser, 1947; Schafer, 1949; Sharp, 1942; Smith, 1949; Soergel, 1936; Steeger, 1944; Washburn, 1950).

Because the layers in question contained water in abundance, it is easier to explain the formation of the afore-described involutions. The inclination conditions ranging from zero to 8° were mentioned previously in the description of the different observation localities. The present slope of the ground surface is thus in many localities rather considerable. The original slope at the time of the frost-action is certainly difficult, even impossible to visualize. A part of the deposits may have been transported away by the later erosion and new accumulations deposited by the littoral process. It can be supposed, however, that the inclination of the ground surface during the frost-action was of the same magnitude as at present. On the sufficiently inclined and frozen ground the layers could have slowly slid during the frost-action down the slope, which could have caused more details and new features in the disturbed sequence of the layers. This earth sliding may not, however, have been very decisive in the examples in question. The main direction in the sliding process is the greatest inclination of the slope. The long axes of the stones are often oriented in this direction, too (Lundqvist, 1948). In most examples the movement occurred, on the other hand, mainly in the vertical direction.

It should be pointed out that in the afore-mentioned examples no remains of ancient ground surfaces have been met with. As reported before, this can be partly caused by the fact that the previously existing layers were destroyed by erosion and replaced by littoral deposits. On the other hand, the origin of the frozen ground does not necessarily require a direct nearness of the ground surface. In addition to the permafrost the freezing of the surficial parts of the ground must have occurred during the formation of the frozen ground. This freezing phenomenon reaches at present in South Finland a depth ranging from 0.1 m to some meters, depending on the severity of the winter, on the snow cover, on the nature of the superficial deposits and their moisture, and on other factors (Keränen, 1923). It may be assumed that during deglaciation the winters were severer than at present and the freezing extended deeper into the ground than now, in the most severe winters and in favorable circumstances perhaps to a depth of many meters.

During the penetration of the winter freezing into the ground, the lower limit between the frozen and unfrozen ground did not always become even, since the frozen ground was reached sometimes deeper and sometimes shallower, depending on the superficial deposits, on their grain size conditions, and on their moisture proportions. Considerable pressure thus developed in the unfrozen ground between the lower limit of the freezing and the upper limit of the permafrost (Johnsson, 1956; Poser, 1947; Steeger, 1944; Washburn, 1950). This pressure strove to exert itself in the direction of the least resistance, accordingly outwards and upwards. As is pointed out, e. g., by Sharp (1942, pp. 130—131), the frozen ground of the unhomogenous material act as pressure centers, which have their effect on the coarser deposits in the environment, too. Hence the finer-grained material have to surround the coarser deposits, which are less plastic and freeze more slowly. The shape of the disturbances accordingly becomes a characteristically bulbous structure, in which the coarsest material exists innermost and the finest grains outermost. The pressure centers recur at relatively equal distances; hence several neighboring centers of the disturbances are produced. Thus many bulbous involutions are repeated in the frozen ground. Smaller involution bags are situated inside the larger ones, showing a freezing center on a minor scale.

In places the involutions are met with in several layers, one on top of the other (e. g., Messukylä, Vilusenharju, loc. 4). This phenomenon is explained by Keilhack (1927, pp. 360—369) as indicating different climatic periods and repeated colder phases, possibly separate Ice Ages. This does not need to be the case with the afore-mentioned examples. The involutions situated higher up are less developed and broken. It is there fore most probable that such an arrangement of the involutions indicates only the gradual downward shifting of the permafrost surface.

The afore-described disturbances of the stratified deposits are therefore understood to have originated during the deglaciation and near the ice border, where permafrost existed in the ground. The involutions were caused by the squeezing of the winter frost near the upper border of the permafrost and by the unequal pressure in unhomogenous material, and their formation was aided by the land sliding owing to the sloping ground.

GEOLOGICAL SIGNIFICANCE OF THE FOSSIL FROST-ACTION

Frost-action may occur only on dry land or at the very most in shallow water. In deeper water the formation of the involutions is not possible. Their presence at different altitudes can accordingly elucidate the oldest shore-line displacement in Finland.

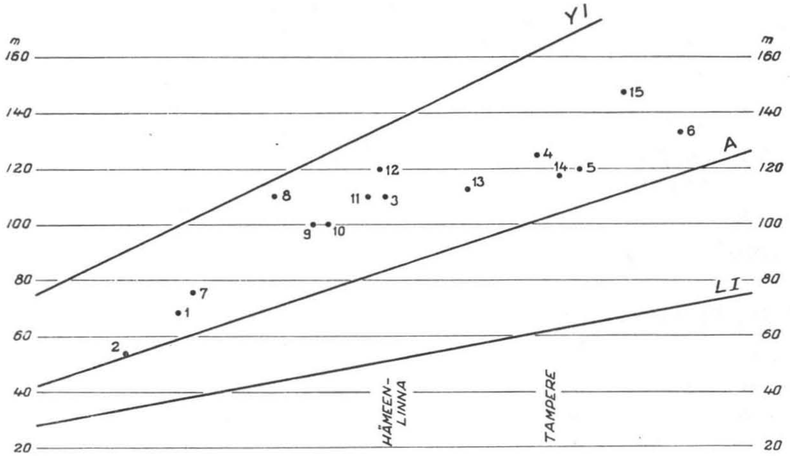

Fig. 10. Relation of the elevations of the involution layers to the most important late-quaternary shore-lines in Southern Finland. Y I = the highest limit of the postglacial Yoldia Sea, A = the transgression of the Ancylus Lake of the Baltic, L I = the highest limit of the Littorina Sea.

By plotting the altitudes of the said involution localities in the generalized shore-diagram (Hyyppä, 1937; Sauramo, 1955), it appears that all the hitherto known occurrences are situated above the shore-line corresponding to the Ancylus transgression maximum of the Baltic (Fig. 10). For the present such features are not known at lower altitudes. This may indicate that the water level of the Baltic should have been at least at this niveau or possibly at a still lower niveau during the origin of the involutions. On the other hand, because the involutions are covered regularly by littoral accumulations and because the highest shore-line is known to be situated at considerably higher altitudes, the afore-mentioned explanation should indicate that a marine transgression occurred in the early Late-Glacial period. It would be untimely to try to estimate the amplitude of this transgression; in any case some ten meters may be in question. It may be mentioned further that several Finnish investigators have described for other reasons a similar transgression in the early Late-Glacial (Hyyppä, 1951; Mölder, Valovirta, Virkkala, 1957; Sauramo, 1955).

The causes of this marine transgression are quite obvious. Noticeable amounts of ocean water were taken up by glacier ice during the glaciation. The sea level was lowered 90 m according to Antevs (1928, pp. 74—82), 85 m according to Daly (1934, p. 46), and according to the estimates of different investigators, 70—102 m (Flint, 1947, p. 437). The water was released back to the oceans during the melting of the ice and transgression was caused. The amount of this transgression in Finland may have been, however, much smaller than the said estimates of the regression of the sea level. A noticeable amount, perhaps the main part of the fixed water, may have returned to the oceans before the melting ice border had reached South Finland.

The existence of this late-glacial transgression is evidenced by the relations of the involuted layers to the other superficial deposits, also. The fossil frozen ground is met with in places underlying varved sediments. Because dry land is a requirement of the frost-action and because the overlying varved sediment indicates rather deep water, the sequence of the superficial deposits thus proves that the transgression postdates the formation of the frozen ground.

The occurrence of the frozen ground below the varved sediments shows that it was formed earlier than the last mentioned. Varved sediments originated in South Finland during the Late-Glacial, with the possible exception of the final phase of the Younger Dryas. Involutions of a corresponding kind as in Finland have been dated in the Netherlands as belonging to the Younger Dryas (van der Hammen, and Maarleveld, 1952). The date of the origin of the fossil frozen ground may be in Finland still older, because the ice retreated during the Younger Dryas at least 100 km to the north from the First Salpausselkä (Mölder, Valovirta, Virkkala, 1957). It is known, on the other hand, that the water level of the Baltic was relatively low in South Finland during the Alleröd period *(op. cit.)*. It is thus possible that the fossil frost involutions originated in South Finland not later than in the Alleröd period and south of the Salpausselkä zone they may be still older, dating from the final phase of the Older Dryas.

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EXPLANATION TO PLATES

PLATE I

Fig. 1. General view of the position of the involution layer in Järvenpää.
Fig. 2. Detail of the involution layer in Järvenpää.
Fig. 3. General view of the involution layer in Seutula.
Fig. 4. Frost crater of the involution layer in Seutula.
Fig. 5. Vertical section of one involution in Hämeenlinna, Vuorentaka.
Fig. 6. Horizontal section of the involution layer in Hämeenlinna, Vuorentaka.

PLATE II

Fig. 1. Involutions in Messukylä, Vilusenharju.
Fig. 2. Involutions in Tampere, Villilä.
Fig. 3. General view of the involutions in Kuru.
Fig. 4. Detail of the involutions in Kuru.
Fig. 5. Detail of the involutions in Nurmijärvi, Paraatinkalliot.
Fig. 6. Detail of the involution layer in Riihimäki, Herajoki.

PLATE I

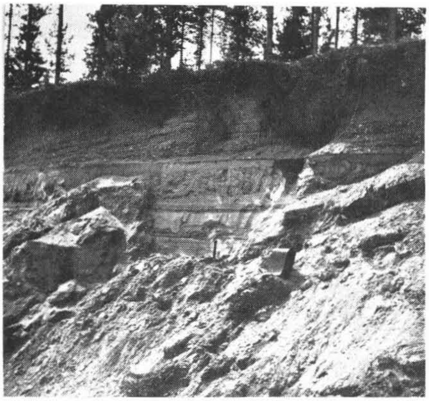

FIG. 1

FIG. 2

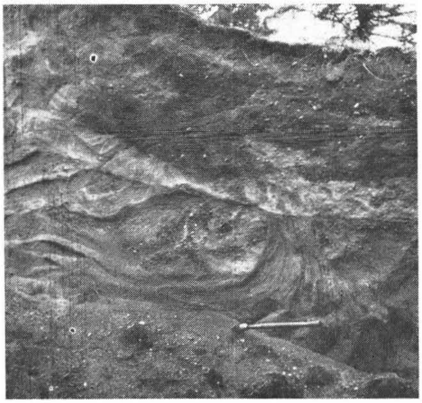

FIG. 3

FIG. 4

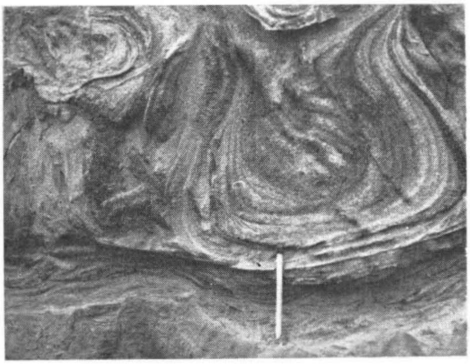

FIG. 5

FIG. 6

K. Virkkala: On the Lateglacial Frost Phenomena

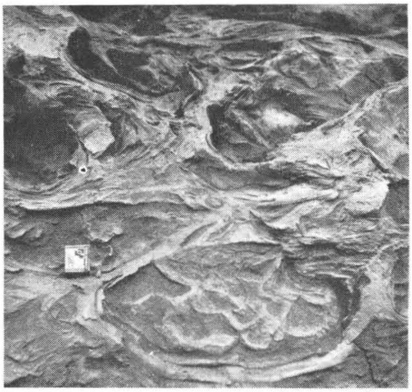

FIG. 1

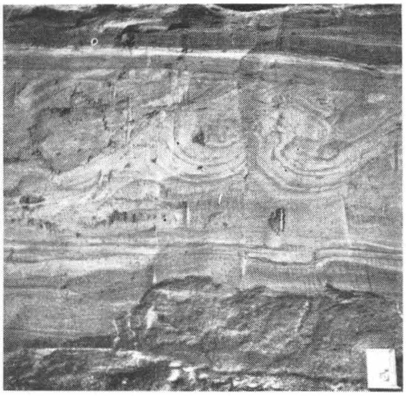

FIG. 2

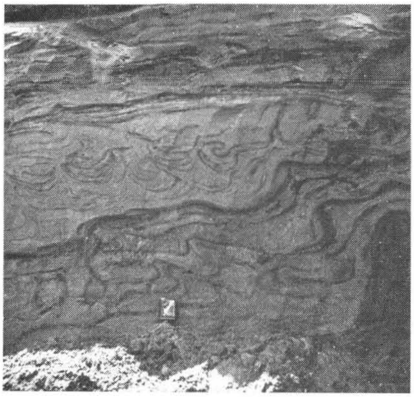

FIG. 3

FIG. 4

FIG. 5

FIG. 6

K. Virkkala: On the Lateglacial Frost Phenomena

4

EXAMPLES OF THE GRANITIZATION OF GNEISSES [1]

BY

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ABSTRACT

Examples of the granitization of biotite-plagioclase gneisses and of quartzite (orthoquartzite) are presented. They show that the granitization is caused by a granitic melt in which the potassium dominates over the sodium. The migmatitic veined gneisses represent an intermediate stage in the granitization process, and they are products of the injection of the granitic melt. This injection is mostly accompanied by metasomatic processes. The granitization of the sodium-poor rocks shows that the migration of the sodium may precede that of the potassium in the migmatite front. In the metasomatic processes the country rock receives the elements that it contains in smaller measure than the granitizing melt, and it gives off the elements whose content is greater in the country rock than in the melt. The new chemical composition approaches the average composition of both participants and, owing to the quantitative superiority of the granitic matter, the end composition approaches that of the granitic melt. Thus the process has the character of a diffusion which tends to homogenize the chemical composition of the rocks in the field of action of this granite. The general character of the granitization process is discussed.

CONTENTS

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[1] Received January 16, 1959.

INTRODUCTION

The term granitization signifies a process whereby a rock is converted into a granitic composition. The original composition of the rock may be varyable but the ultimate result of the process is a granitic rock. Accordingly, the process cannot always be the same in detail. The velocity and the character of the process also depends on the pressure and temperature, on the texture of the rock and on the kinetic factors (Wegmann, 1935, p. 330). To arrive at the general principles of the process would require a great many instances (cf. Härme and Laitala, 1955; Härme, 1958). For this purpose some examples of the granitization of supracrustal rocks are presented in the following.

In the examples described here the active factor has been a granite containing more potassium than sodium. Commonly this granite contains only a slight amount of biotite as femic constituent, and it has a rather aplitic character. The primary granite does not contain so much free silica and thus no pegmatitic melt proper was involved. The granite is often coarse-grained and resembles pegmatite (Härme, 1958, p. 59), for which reason many earlier investigators assumed a granitization caused by the pegmatites to have occurred.

EXAMPLES OF THE GRANITIZATION OF SOME GNEISSES

The author has selected some examples of the granitization of biotite gneisses. These gneisses are stratified and of sedimentary origin and therefore their primary composition varied in different layers. Granitized gneisses occur richly in southern Finland but the difficulty is to get specimens of both undisturbed and granitized gneiss types as far as possible of the same primary composition (preferably of the same layer) from the same outcrop. Only thus is it possible to draw conclusions from the changes in composition during the granitization. Some examples of such cases are presented in Table 1. The b-examples represent granitized rocks and the a-examples are corresponding unchanged or only weakly changed gneiss types.

In all the cases presented here the granitizing factor was microcline granite (cf. Härme, 1958, p. 46) of »ideal granitic» composition (Eskola, 1950, p. 5; 1956, p. 89). Its veins cut the bedding and the schistosity of the gneisses, and they may continue as thin veinlets following the schistosity of the wall rock (see Härme, 1958, Fig. 1, Plate I). In all the examples the gneiss was originally poor in potassium, i. e. the rock contains as potash minerals only biotite and no primary microcline (the rock in example 2a, Table 1, is already partly granitized). Hence the *lit-par-lit* structure produced is of arteritic character. Such migmatites were excellently described by Wegmann and Kranck (1931).

The determinations of the mineral compositions have been made with the point counter method, whereby — in consequence of the inhomogeneity of the gneisses and especially of the granitized types — the degree of inaccuracy is remarkably high and thus these numbers are only approximate values. The observations and conclusions presented herein are, however, based on a larger number of examples than are described and thus they have to some degree a statistical basis.

Table 1. Mineral Composition of Some Gneisses

Constituent	1		2		3			4	
	a	b	a	b	a	b	c	a	b
Quartz	43.1	39.5	46.5	41.2	51.1	20.6	23.8	22.1	26.6
Plagioclase	38.9	10.3	11.5	11.6	36.5	9.7	11.9	55.4	21.8
Microcline	—	34.1	24.2	34.2	—	30.1	43.5	—	16.8
Biotite	18.0	13.4	16.3	10.9	8.2	10.5	5.9	21.1	8.8
Almandite	—	—	—	—	3.4	9.1	9.6	1.3	9.6
Cordierite	—	2.7	1.5	2.1	—	19.7	4.7	—	15.1
Accessories	—	—	—	—	0.8	0.3	0.6	0.1	1.3
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

1. *Päivölä, Degerby*

   a. The plagioclase (An_{30}) is initially sericitized. The quartz has a clear undulatory extinction and some quartz grains are elongated along the schistosity.

   b. The plagioclase (An_{25-27}) is partly sericitized. The microcline contains flame perthite. The quartz has a strong undulatory extinction. The microcline contains roundish inclusions of quartz, which have only a slight undulatory extinction. Similar inclusions occur also in the cordierite, which is incipiently pinitized.

2. *Päivölä, Degerby*

   a. The plagioclase (An_{25-27}) is fresh. The cordierite is strongly pinitized.

   b. The plagioclase (An_{24-25}) is incipiently altered. The microcline contains roundish inclusions of quartz. The cordierite is strongly pinitized.

3. *Palojoki, Nurmijärvi*

   a. The plagioclase (An_{28-32}) is fresh. The rock is even-grained.

   b. The plagioclase (An_{25}) is fresh. Inclusions of drop quartz occur in the microcline as well as in the cordierite, which contains inclusions of biotite and sillimanite, too. The oblong grains of microcline and cordierite are oriented along the schistosity (Fig. 1, Plate I). The cordierite shows a multiple twinning.

   c. The plagioclase (An_{23}) contains myrmekite at the border against the microcline. The microcline contains inclusions of drop quartz (Fig. 2, Plate I). The twinned cordierite is partly pinitized. The almandite encloses needles of sillimanite. Sometimes the garnet occurs as an inclusion in the cordierite. The grains of almandite are oblong in the direction of the schistosity and biotite myrmekite occurs sometimes at the end of the garnet grains (Fig. 3, Plate I; cf. Pehrman, 1936, p. 14).

4. *Palojoki, Nurmijärvi*
   a. The plagioclase (An_{35-40}) is weakly zoned and incipiently sericitized. The almandite occurs as porphyroblast. The texture is granoblastic (Fig. 4, Plate I).
   b. The plagioclase (An_{25-30}) is incipiently sericitized and it contains myrmekite at the border against the microcline. The microcline contains flame perthite and inclusions of drop quartz, which shows only a weak undulatory extinction. Similar quartz inclusions occur in the cordierite, which contains also inclusions of microcline and needles of sillimanite. The biotite is pale-colored and richly fringed by speckles of hematite pigment.

In all the examples the a-types have a crystalloblastic texture. In all the cases investigated the sedimentary rock recrystallized to gneiss during the earlier regional metamorphism before the injection of the granitic melt. In general no strong movements seem to have occurred during the injection. The injection took place chiefly along the previously existing schistosity planes (cf. Edelman, 1956, p. 21). The arteritic veinlets follow the older schistosity (Härme, 1958, Fig. 1, Plate I) and are megascopically discernible by virtue of their color. Microscopically one can see that mostly the coarser microcline metablasts and sometimes also the cordierite grains are oblong in the direction of the schistosity (Fig. 5, Plate I; cf. Edelman, 1956, p. 20).

Sometimes the granitic veinlets are microscopically distinguishable from the older host rock, but often there is no evident line between the light-colored minerals of the veinlets and those of the older host rock (cf. Wegmann and Kranck, 1931, p. 92) and sometimes they are also of the same grain size. Mostly at the contact against the veinlet the gneiss also recrystallized to coarser grain size during the injection process. It is possible that during the injection the gneiss partially remelted (cf. Wegmann, 1935, p. 317) although no plain evidences were revealed in the microscopical investigation.

The metablasts of the garnet usually enclose quartz and biotite as do also the cordierite and the microcline. The main part of the quartz, even in the granitized rock, usually has a strong undulatory extinction whereas the quartz enclosed by cordierite, microcline and garnet is to the greater part drop quartz with no or only faint undulatory extinction.

Considering the changes in the mineral compositions (Table 1), a strong increase of the microcline content is most conspicuous. Commonly the quartz content decreases. Example 4, Table 1, makes a deviation, which may in part be due to the inhomogeneity of the rock and to the inaccuracy of the determinations but also to the fact that in the unchanged a-type the quartz content is considerably lower than in the other examples. The amount of the plagioclase seems to decrease in accordance with the growth of the cordierite content. The amount of the biotite decreases. In some cases the a-type contains a little garnet or cordierite, but in all these examples the granitization has yielded a garnet- and/or cordierite-bearing rock.

In the unchanged gneisses the biotite sometimes encloses small needles of sillimanite, and the almandite contains inclusions of quartz. In the granitized gneisses the microcline, the cordierite, the almandite and partially also the plagioclase contain inclusions of drop quartz, which has no or only a weak undulatory extinction, whereas the other quartz has a distinct undulatory one.

Zircon, apatite and sphene occur as accessory minerals. Granitized rock types contain a little pyrite, too, which mostly does not occur in the unchanged a-types. The granitized rocks sometimes contain a little sillimanite, usually as inclusions in the other minerals, e. g. in the cordierite (Fig. 6, Plate I; cf. Pehrman, 1936, p. 8) and in the garnet.

Table 2. Chemical Analyses of two Gneisses. Analyst, P. Ojanperä

	a		b		Weight norm		
	Weight %	Mol. prop.	Weight %	Mol. prop.		1	2
SiO_2	75.75	1 2606	73.43	1 2220	qu	43.41	35.92
TiO_2	0.28	35	0.34	42	or	10.68	31.85
Al_2O_3	11.77	1154	11.98	1175	ab	25.21	16.94
Fe_2O_3	0.53	33	0.69	43	an	8.56	1.20
FeO	2.64	367	2.81	391	en	4.30	4.78
MnO	0.03	4	0.03	4	fs	4.00	4.09
MgO	1.73	429	1.92	476	cor	1.78	2.42
CaO	1.77	316	0.30	53	mt	0.77	1.00
Na_2O	2.98	481	2.00	323	il	0.53	0.64
K_2O	1.81	192	5.39	572	ap	0.09	0.09
P_2O_5	0.04	3	0.04	3		99.33	98.93
CO_2	0.00		0.00		CO_2	0.00	0.00
H_2O+	0.54		0.64		H_2O+	0.54	0.64
H_2O-	0.10		0.16		H_2O-	0.10	0.16
Total	99.97		99.73			99.97	99.73

a. Biotite-plagioclase gneiss, Päivölä, Degerby.
b. Granitized biotite-plagioclase gneiss, Päivölä, Degerby.

Chemical analyses are made (a and b, Table 2) of unchanged and granitized gneisses corresponding most nearly to examples 1a and b, in Table 1. The potassium revealed by analysis a, Table 2, is included in the biotite, whereas the granitized rock type (b, Table 2) contains microcline in abundance. The percentages of Na_2O and CaO show a decrease. A slight but evident diminution in the relatively high content of silica is verifiable, whereas the alumina remained essentially unchanged. The percentages of magnesium and iron oxides show a faint growth, which may also be due to the inhomogeneity of the rock. It is to be noted that in example 3, Table 1, too, the changes in the mineral composition partly differ from the regularities shown by the other examples.

An example of the granitization of gneiss in the parish of Kirkkonummi is presented in Figs. 7 and 8, Plate II. The gneiss is of sedimentogenous origin and well stratified; and the original compositions of various layers correspond in broad outlines to the compositions of the a-examples in Table 1. The granitizing factor here is so-called Obbnäs granite (Sederholm, 1926, p. 90; Edelman, 1949, p. 37), whose main massif is truly intrusive and of distinct cutting character. East of the body proper a dike of Obbnäs granite transects the gneiss in the manner of a ring intrusion. The Obbnäs granite is rather coarse-grained and contains about twice as much potassium ($K_2O = 5.79$ %) as sodium ($Na_2O = 2.21$). Also in the dike the granite is rather coarse-grained and of cutting character, but at a point in the village of Bondarby, between two cutting portions in the dike line, granitization occurs only in the gneiss. In this instance the texture of the country rock shows that a strong crushing took place before and during the injection of the granitic magma but later than the recrystallization caused by the normal regional metamorphism.

In this granitized gneiss there occur coarse grains of potassium feldspar (mainly microcline) which are true porphyroblasts. Some microcline grains are deformed and show also an undulatory extinction. The quartz has a very strong undulatory extinction; only the drop quartz inclusions in the microcline and in the idiomorphic plagioclase have none or only a weak undulatory one. The rock contains two kinds of plagioclase (An_{27-33}). One is irregular in outline and broken, whereas the other is idiomorphic, zoned (Fig. 11, Plate IV) and often with a sericitized center. The former plagioclase is obviously a primary component of the gneiss, while the latter is of secondary origin and produced by the injected granite melt. The former is commonly of the same grain size as the host rock, whereas the latter occurs in part as porphyroblasts. Both plagioclases have in many cases a clear and more albitic rim against the microcline. Myrmekite occurs in the plagioclase grains at the border against the microcline; that is the case with the secondary plagioclase as well. No aluminum-rich minerals (almandite, cordierite, etc.) were produced in connection with this granitization.

CONCLUSIONS I

In the examples presented in the foregoing (Table 1) the *lit-par-lit* structure is produced by an arteritic granitic melt. This injection was, however, accompanied by metasomatic processes (see Wegmann, 1935, p. 317). The potassium obviously caused an alteration of the plagioclase. This is revealed by the fact that in these examples the myrmekite appears only when secondary microcline occurs, and the myrmekite is in the plagioclase at the very contact against the microcline. The more abundant occurrence of the sillimanite in the granitized rock types shows that the alteration process of the plagioclase released more aluminum than it was possible to associate with (MgFe)O. The common occurrence of the tiny needles of sillimanite in the cordierite as well as sometimes also in the almandite supports this view. Of course it is possible that the gneiss contained a sedimentogenous »excess» of alumina but it seems evident that sillimanite, too, may be partially of metasomatic origin (Buddington, 1948, p. 38; Watson, 1948, p. 161; Saksela, 1953, p. 49; Härme, 1958, p. 57).

In example 4b, Table 1, the microcline contains inclusions of cordierite but on the other hand we know that cordierite may occur as inclusions in microcline (cf. Härme, 1954, p. 39). These facts indicate that the microcline and the main part of the cordierite are formed simultaneously and in such PT-conditions where both minerals may occur together.

Almandite occurs sometimes as inclusions in the cordierite. This reveals that in these migmatites the almandite may be partially older than the cordierite (cf. Wegmann, 1935, p. 322; Pehrman, 1936, p. 22; Metzger, 1945, p. 33). Specimen 3a, Table 1, contains almandite although there occurs no microcline. The same is sometimes true of cordierite, in which case the almandite and the cordierite are yielded through normal regional metamorphism involving utilization of the primary sedimentogenous »excess» of alumina (cf. Pehrman, 1936, p. 22).

In any event it is evident that the granitization process may yield almandite and cordierite (Wegmann and Kranck, 1931, pp. 62, 89, 93; Wegmann, 1935, p. 322; Hietanen, 1943, p. 98; Reynolds, 1947, p. 392; Härme and Laitala, 1955, p. 98; Buddington, 1957, p. 297; Härme, 1958, p. 57). Many investigators (Pehrman, 1936, p. 21; Eskola, 1939, p. 354; Metzger, 1945, p. 32; Mehnert e. a., 1945—48, p. 126; Hietanen, 1947, p. 1026; Ramberg, 1948, p. 563; 1952, p. 57; etc.) have submitted the opinion that the biotite changes into microcline and almandite (or cordierite). This alteration needs, however, additional alumina, which is liberated when potassium replaces the calcium of the anorthite. This alteration of the anorthite takes place at the initial stage of granitization. For some reason — probably owing to the scantiness of the invading potassium or to the slowness of the granitization process (cf. Eskola, 1956, p. 97) — the potassium does not then consume all the alumina of the anorthite to form the potassium feldspar, but the process releases a part of this alumina.

Obviously in the examples presented an alteration of the biotite has taken place, as indicated by the bleached color of the biotite fringed by speckles of hematite. Further evidence is the occurrence of biotite myrmekite in Fig. 3, Plate I (cf. Pehrman, 1936, p. 14). The released (Mg, Fe)O then reacted with alumina and silica to yield almandite and cordierite.

This is indicated also by the decrease in the biotite content in examples 1, 2 and 4, Table 1.

As mentioned, during the granitization of the gneisses a decrease in the quartz content took place, except in example 4 (Table 1), where the a-specimen in itself contained relatively little quartz. The decrease of the quartz seems to have been relatively stronger in the quartz-rich rock, example 1, Table 1, than the decrease of the SiO_2 percentage revealed by the (most closely corresponding) chemical analyses, Table 2. (cf. qu)

In this case the relative SiO_2-content of the country rock has been higher than that of the granitizing melt. On the other hand, the amount of »free silica» in the host rock in example 4, Table 1, may have been relatively lower than that of the granitizing melt, and in this case the amount of the quartz has grown. The investigations published earlier by the author (Härme and Laitala, 1955; Härme, 1958) show similar features. Thus in such granitization the amount of free silica in the country rock approaches that of the granitizing melt. Of course the metasomatism process in itself binds or releases silica and thus changes the amount of the quartz without a migration of the SiO_2 from one side to the other.

The said increase of the sulphide ore minerals during granitization confirms the view that the volatiles play a great role in this kind of metasomatic granitization process.

The example of granitization induced by the Obbnäs granite (p. 46) is in many respects quite interesting. The intrusive and magmatic character of the Obbnäs granite is evident. The granitic dike is situated in a fault zone developed in connection with the intrusion of the Obbnäs granite body. In the place described the movement has not, however, developed any open fissure but has only foliated the gneiss and in that way favored the injection of the granitic magma into the gneiss and the granitization in this zone. The granitization here took place on a relatively small scale but locally it has led to a granitic composition, as is seen in Fig. 8, Plate II. The idiomorphic plagioclase grains occurring in the semi-granitized patches are most likely idioblasts yielded by the invading granitic matter. It seems that also the amount of the plagioclase has increased in the rock although, owing to the inhomogeneity and to the coarseness of the rock, it is not accurately determinable. At the border against the potassium feldspar this idiomorphic plagioclase contains myrmekite, thus showing that, at least partially, this secondary plagioclase crystallized earlier than the potassium feldspar. An interesting feature is that the idiomorphic secondary plagioclase is zoned, although it was not properly surrounded by magma during the crystallization.

The absence of Al-rich metablasts in this case points up the fact that the granitization had not here such a metasomatic character as in the other examples previously described but appeared here only as a relatively rapid impregnation of granitic magma. Thus the granitization here has been rather a physical granitization (Härme, 1958, p. 46), although it is scarcely accompanied by a corrosion of the plagioclase, and other chemical processes.

AN EXAMPLE OF THE GRANITIZATION OF QUARTZITE

On the NW-slope of the Puijo hill (below the ski-jump), near the town of Kuopio, granitic veins transect the quartzite. The potassium dominates over the sodium in the granite, which in regard to its chemical composition corresponds most nearly to the »ideal granites» of Eskola. The quartzite is sedimentogeneous (cf. Wilkman, 1916; Preston, 1954) and to some degree inhomogeneous, but in addition to the quartz it contains few other minerals. Its primary mineral composition is represented approximately by example a, Table 3. The quartzite is evidently schistose and its elongated quartz grains have a strong undulatory extinction.

Close to the cutting veins the quartzite is granitized. The granitic matter was partly injected as a network of veinlets and partly soaked without any visible channel into the quartzite (Fig. 9, Plate III), where it formed separate feldspar grains or lenticular feldspar aggregates whose longitudinal direction follows the schistosity. The manner of the occurrence of these feldspar lenses or streaks shows that the feldspatic matter invaded the rock in the direction of the schistosity (Fig. 10, Plate III).

Table 3. Mineral Composition at various Granitization Stages of the Quartzite. Puijo, Kuopio Town

Constituent	a	b	c	d	e	f	g	h	i	j	k
Quartz	98.2	93.8	79.7	56.1	94.3	92.6	95.2	94.5	70.3	89.4	36.3
Plagioclase .	0.4	5.1	20.0	43.5	4.0	5.4	0.3	0.9	23.1	2.9	10.6
Microcline ...	—	—	—	—	0.6	0.6	3.8	4.6	5.1	7.3	52.4
Sericite	1.4	1.1	0.3	0.4	1.1	1.4	0.7	—	1.5	0.4	0.7
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

a. The quartz has a strong undulatory extinction. The plagioclase is xenoblastic and obviously of detrital origin. The texture is crystalloblastic.

b. The plagioclase is dim and sericitized. It is possibly in part of secondary origin. The quartz has a strong undulatory extinction.

c. The plagioclase is partly zoned (center An_{33}, margin An_{22}), and incipiently altered, in part along twinning lamellae (Fig. 12, Plate IV). The main part of it is of secondary origin. The quartz has a strong undulatory extinction.

d. The plagioclase (oligoclase) is of secondary origin; it encloses grains of quartz which have a strong undulatory extinction.

e. The plagioclase is at least in part of secondary origin, and it contains a little sericite. The microcline occurs as narrow streaks. The quartz has a strong undulatory extinction.
f. The plagioclase (albitic) is strongly altered, and it contains sericite. The quartz has a strong undulatory extinction.
g. The microcline is dim and gives an impression of an altered plagioclase. The quartz has a strong undulatory extinction.
h. The plagioclase is dim and in part weakly zoned. Some ragged grains of the plagioclase occur as inclusions in the microcline. The quartz has a strong undulatory extinction.
i. The plagioclase (albitic) is strongly altered. The microcline contains perthite. Some feldspathic streaks are domated by the microcline and the others by the plagioclase.
j. The plagioclase is weakly zoned and in part altered. It contains myrmekite at the border against the microcline. The microcline (Fig. 13, Plate IV) contains perthite.
k. The plagioclase is dim but in many places has a clear rim against the microcline. At the border against the microcline the plagioclase contains myrmekite. Dim grains of the plagioclase occur as inclusions in the microcline. The bigger microcline grains contain roundish grains of quartz, which show no undulatory extinction. These microcline grains richly contain perthite (in the center of the grain more than at the border), which is partly twinned. Around and between these porphyroblasts of the microcline (Fig. 14, Plate IV) there occur in places a fine-grained groundmass of plagioclase, microcline and quartz. This fine-grained microcline contains only little or no perthite. The fine-grained quartz shows only a weak undulatory extinction whereas the main part of the quartz is medium-grained and has a strong undulatory extinction.

There are no signs to show that any strong movement took place in the quartzite during the granitization. The schistosity and the strong undulatory extinction of the quartz as well as the recrystallization are products of movements and regional metamorphism taken place before the granitization. It is obvious that the quartz, which has a strong undulatory extinction, is the primary component of the host rock in the more strongly granitized places, too.

Example a, Table 3, represents a wholly recrystallized quartzite (orthoquartzite), which has a crystalloblastic texture. This example is unaffected by the granite.

Examples b-k, Table 3, represent the mineral compositions of the several stages of the gradational granitization. Each of these determinations (point counter method) is made of one thin section only but each thin section is made of its own specimen, megascopically selected to represent the various stages of this granitization.

In the microscopical study attention is drawn to the fact that in some places the granitized quartzite contains as secondary feldspar only plagioclase (examples c-d and actually also examples e-f, Table 3). In some other places the feldspar is mainly microcline (examples g-h, Table 3) whereas in the places of the strongest granitization the secondary feldspar is both plagioclase and microcline (examples i-k, Table 3). Thus it seems that the entering of the plagioclase partly precedes that of the microcline whereupon the microcline replaces the earlier plagioclase. This fact is revealed by the occurrence of the myrmekite at the border of the plagioclase grains against the microcline (examples j-k, Table 3). In example i, Table 3, the different feldspar streaks are partly dominated by the plagioclase or by the microcline, thus showing that the sodium and the potassium may soak in, at least to some extent, apart from each other.

CONCLUSIONS II

The granitization of the quartzite is in many respects a very favorable example. Because the host rock, in addition to the quartz, does not essentially contain other minerals, it is suitable to observe, eg., the behavior of the alkalies in the granitization process.

In example k, Table 3, the bigger microcline grains contain abundant perthite, whereas between and around them there occur small grains of plagioclase as well as small grains of microcline, which contain hardly any perthite. It is obvious that the perthite of the large microcline grains is a product of exsolution. In regard to the formation of the small grains of microcline without perthite there are several possibilities:

— they are crystallized later than the large ones and in changed PT-conditions, wherefore in the crystallization not so much of the plagioclase component was included in the microcline as a solid solution and, accordingly, the plagioclase was crystallized separately as small grains;

— the material of the small plagioclase grains was originally partly included in the small grains of potash feldspar, but in the exsolution the plagioclase component — owing to the small size of the microcline grains — was passed outside, where it crystallized alone as small plagioclase grains. It is also to be noted that the large microcline grains contain more perthite in the centre of the grain than at the border.

It is possible, too, that the small plagioclase grains referred to were formed at a later stage, when the potassium replaced the sodium of the plagioclase and through this process the solutions were enriched in sodium, which finally crystallized as plagioclase (cf. Härme, 1958, p. 57).

The granitizing melt contained matter of both sodium and potassium feldspars, but in the migration into the wall rock these alkalies seemed to behave in different ways. At the incipient stage of the granitization they migrated in part in different phases. Since the migration of the sodium at first differed from that of the potassium, it would seem that the potassium followed closely and in part replaced the sodium. Thus there are in this case actually two »fronts»: the low »front» of sodium precedes the proper »migmatite front» of the potassium (cf. De Vore, 1956, p. 46; Ljunggren, 1956, p. 643; 1957, p. 130; Mísař, 1958, p. 287). The plagioclase formed remains partially but the potassium replaces a considerable portion of it. In this case the distance between the successive sodium-rich »front» and the potassium-rich »front» measures only a few centimetres — at the most some ten centimetres.

The different behavior of the alkalies is obviously peculiar to the process. A similar phenomenon had been previously observed in the formation of the orbicular granites. Simonen (1941, pp. 128, 129), in his interesting study of the orbicular granite of Kemijärvi, submits this interpretation: »A metasomatic albitization of the granodiorite, or a soda metasomatism, has proceeded, resulting in the formation of the esboitic nuclei.» This is followed by the potash metasomatism: ».. at a certain stage, potash metasomatism came instead of soda metasomatism.» His explanation (Simonen, 1941, p. 136) of the orbicular granite of Esbo (cf. Sederholm, 1928) is principally the same, and later he writes (Simonen, 1950, p. 37): »The occurrence of orbicular rocks in migmatitic surroundings indicates, however, that the esboitic crystallization is closely connected with metasomatic processes taking place in an approaching migmatite front.»

In the granitization of the quartzite the relative amount of the quartz decreased through the addition of the feldspars. No evidence of the coming of the free silica has been obtained. In example k, Table 3, the main part of the quartz has a strong undulatory extinction and thus obviously no essential recrystallization of the quartz took place during the granitization. Only the inclusions of the drop quartz in the microcline as well as the more fine-grained quartz around the larger feldspar grains do not show an undulatory extinction. They were obviously formed in connection with the crystallization of the microcline. There is no evidence to show whether the silica necessary to the feldspars originated from the granitizing melt or if it was the quartz of the host rock. The latter assumption seems to be the more plausible one because the silica content of the host rock (in the pore solution) has in this case undoubtedly been higher than that of the granitizing melt.

As mentioned in the foregoing, the granitizing matter cuts as veins the quartzite, thus showing that it is arteritic and had intruded in a liquid state. The appearance of the feldspars as single grains or lenticular aggregates without any visible feeding channel points up the fact that the granitizing matter soaked into the quartzite as solutions rather than as a magma proper. The magma had a constant composition, a distinct K:Na-ratio, but in the migration the sodium and potassium have partially behaved in different ways. This, on the other hand, shows that here it was not the magma that infiltrated but rather the magmatic solutions soaked along the intergranular film into the wall rock. The process may continue in this way up to the granitic end product (Eskola, 1950, p. 6; 1956, p. 89; Härme, 1958, p. 59).

CONCLUDING REMARKS

Several investigators have emphasized the increase in the potassium content through granitization. On the ground of the granitization of the quartzite herein discussed, it is obvious that also the amount of the sodium increase when the country rock is originally poor in sodium. In the sodium-poor wall rock the increase in the sodium content together with that of the potassium during granitization takes place in such a way that the process tends to result in a granitic composition where the Na:K-ratio is the same as in the granitizing melt. If the relative amount of sodium is greater in the wall rock than in the granitizing melt, then the sodium-content of the wall rock correspondingly decreases (Härme, 1958, p. 58; cf. Reynolds, 1947, p. 392).

In the granitization of the basic rocks the amount of SiO_2 increases, whereas in the granitization of quartzite or of other rocks relatively rich in quartz the amount of the silica decreases. Goldschmidt (1920, p. 119) supposed that the addition of the alkalies takes place as alkali silicates. Wegmann (1935, p. 346) concluded that the addition of the alkalies takes place at least in part as alkali aluminates. The same seems to be the case in the granitization of quartzite. The alumina content, however, decreases in the granitization of aluminum-rich rocks (see Härme, 1958, p. 50, Table 1a). Evidently the alternative increase or decrease of silica depends on the circumstance whether the amount of »free silica» in the wall rock is lesser or greater than in the granitizing melt. The case of the calcium in this metasomatic process is apparently similar. These features indicate that this granitization process has the character of diffusion which tends to eliminate the differences in the concentrations existing in the field of action of the granitizing melt. There is considerable evidence, however, that a dry diffusion is here not in question (cf. Read, 1951, p. 10) because the active factor of these processes is a melt which contains water and other volatiles. The diffusion is, of course, slow because the wall rock is in a solid state.

The examples cited earlier (Härme and Laitala, 1955; Härme, 1958) and in the foregoing do not show any evident increase of basicity in front of this kind of granitization. On the contrary, at least in the granitization of more basic rocks, the amounts of Fe, Mg and Ca decrease; only the granitization of the quartzite shows an increase of Ca, too. But this increase of Ca speaks in favor of a tendency toward homogenization: the composition of the country rock approaches that of the granitizing melt. The granitization of the quartzite does not show any increase of Fe and Mg. This is quite understandable because the microcline granite is very poor in femic constituents, and it has rather an aplitic character. If a true homogenizing diffusion is involved, it is not likely that the process would cause greater differences in the concentrations, at least on a larger scale. Many factors, however, participate in the process and thus it is quite possible that local concentrations of the cafemic constituents may occur on a small scale.

In general, it is more probable that the Mg, Fe and Ca released in the process migrate in the direction where their concentrations are at a minimum, or, in other words, towards the granitic melt (cf. Ramberg, 1952, p. 263). For example, in the migmatite and granite areas of southern Finland the potassium granite occurs so abundantly (Härme, 1958, p. 59) that it may take in considerable magnesium, iron and calcium without any essential increase of its basicity stage. In these areas the granitization is of a homogenizing character and the microcline granite tends to destroy the older rocks and to drown them in its own abundance rather than bringing about differences in the concentrations.

As is well known, the migration takes place most readily in the direction of the schistosity and therefore the rocks which have an orientated structure respond most readily to granitization. The same is true of the formation of the *lit-par-lit* structure, which developes especially in the direction of the previous schistosity (cf. Wegmann, 1935, pp. 317, 321). It is, however, to be noted that the metablasts, especially the porphyroblasts of the microcline, settle down with their longitudinal direction generally in the direction of the schistosity (Eskola, 1952, p. 129; Edelman, 1956, pp. 20, 37). This is quite a common phenomenon.

The question of a possible change in volume in connection with granitization has still been under discussion. At least in the case of physical granitization (Härme, 1958, p. 46) where the host rock is injected by arteritic veinlets (the *lit-par-lit* structure), we must assume also an increase in volume. The same is the case with the example (p. 46) in which the gneiss has been infiltrated by the magma of Obbnäs granite. Another matter then is that usually the formation of such a *lit-par-lit* structure is accompanied by metasomatic phenomena whereupon the replacement releases some elements, which pass away. The evaluation of the change in volume is in such cases very difficult.

We must assume that in the said granitization of the quartzite more matter invaded into the quartzite than escaped from it, because — at least in the incipient stage of the granitization — no proper substitution of the matter of the host rock took place. The volume must have been dilated even if we assume that the silica of the quartzite is used to form the feldspars. Many field observations indicate that the potassium granite melt acted under high pressure. Wegmann made a large part of his investigations of migmatites in southern Finland, and he has especially emphasized the dilatation of the volume through granitization (Wegmann and Kranck, 1931, p. 62; Wegmann, 1935, p. 348).

In any case we must assume that some of the elements increased in quantity in the wall rock, in connection with granitization, whether through injection or through migration. At least in the case of injection the decrease in the content of some elements may be only relative because the share of the country rock in the total volume decreased.

As already mentioned, the almandite, the cordierite and other Al-rich minerals may form in the gneisses and also in the igneous rocks in connection with the metasomatism caused by the potassium granite. This does not, however, mean that all the garnet- and cordierite-bearing gneisses (kinzigites) of southern Finland have been formed this way. This metasomatic process is only one possibility. Very often minerals rich in alumina are formed in sedimentogenous rocks through regional metamorphism. The primary excess of alumina in the argillaceous rocks makes it possible.

According to Simonen (1953), however, the gneisses of southern Finland are graywackes in origin. The graywackes do not characteristically contain so large an amount of alumina as clays proper (Pettijohn, 1957, pp. 307, 343, 345), but the alumina of the graywackes occur for the greater part in detrital feldspars. This is true also of the gneisses of southern Finland, where the graywackes — as graywackes in general — occur in connection with basic volcanites and thus they also contain basic volcanic material. Obviously the anorthite component of their plagioclase had altered partially in the appropriate PT-conditions of the regional metamorphism (sometimes also without potassium metasomatism) and given alumina to the formation of the Al-rich metablasts. Thus the plagioclase was to some extent a store of aluminum and a regulator in the aluminum economy of the rock during metamorphism. It is also one of the factors which has sometimes caused a decrease in the quartz content during the early stage of the granitization process, i. e., the desilication of Reynolds (1946, p. 392).

In many cases it is true that the gneiss, as a product of the regional metamorphism, contained, e. g., almandite before the injection of the potassium (cf. Pehrman, 1936, p. 22; Kranck, 1937, p. 78). In connection with the potassium metasomatism the alteration of the anorthite component became strong, as is revealed by the fact that Al-rich metablasts occur most abundantly expressly in the migmatite front (Wegmann and Kranck, 1931, pp. 25, 89), at the stage of incipient granitization.

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EXPLANATION TO PLATES

PLATE I

Fig. 1. Grains of microcline (m) and cordierite (c) are oblong in the direction of the schistosity. The almandite (a) contains inclusions of drop quartz, which have no undulatory extinction. Nurmijärvi, Palojoki. Nic. +. Magn. 10 ×.

Fig. 2. Inclusions of biotite (b) and drop quartz (q) (without undulatory extinction) in the microcline. Nurmijärvi, Palojoki. Nic. +. Magn. 45 ×.

Fig. 3. Biotite myrmekite close to the almandite (a). The biotite is partly enclosed by the almandite (right below). Nurmijärvi, Palojoki. Nic. +. Magn. 45 ×.

Fig. 4. The granoblastic texture of a (biotite-plagioclase-quartz-) gneiss unaffected by the granitization. Nurmijärvi, Palojoki. Nic. +. Magn. 25 ×.

Fig. 5. Veined gneiss. The salic bands are bordered by streaks of biotite. Grains of twinned cordierite (c) are oblong in the direction of the schistosity. Tuusula, Lahela. Nic. +. Magn. 10 ×.

Fig. 6. Nodules of sillimanite as inclusions in the cordierite. Loppi, Launonen. Nic. +. Magn. 35 ×.

PLATE II

Fig. 7. Granitized gneiss. The light-colored spots are porphyroblasts of feldspars. The dark bands are original layers. Kirkkonummi, Bondarby. $^1/_8$ natural size.

Fig. 8. Granitized gneiss. The lower part is weakly granitized gneiss whereas the upper part is already of granitic composition. Kirkkonummi, Bondarby. $^1/_5$ natural size.

PLATE III

Fig. 9. Granitized quartzite. The light-colored spots are feldspars. Kuopio, Puijo. Photo: E. Halme.

Fig. 10. Incipiently granitized quartzite. The light-colored narrow streaks and spots are feldspars. Kuopio, Puijo. Photo: E. Halme.

PLATE IV

Fig. 11. A zoned porphyroblast of plagioclase in granitized gneiss. Kirkkonummi, Bondarby. Nic. +. Magn. 30 ×.

Fig. 12. Idioblasts of plagioclase in quartzite. Kuopio, Puijo. Nic. +. Magn. 40 ×.

Fig. 13. Metablasts of microcline in quartzite. Kuopio, Puijo. Nic. +. Magn. 20 ×.

Fig. 14. Porphyroblasts of microcline contain inclusions of drop quartz (without undulatory extinction). The metablasts are surrounded by a fine-grained mass of feldspars and quartz. Kuopio, Puijo. Nic. +. Magn. 10 ×.

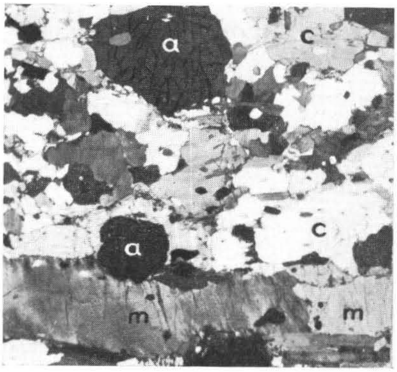


Fig. 1

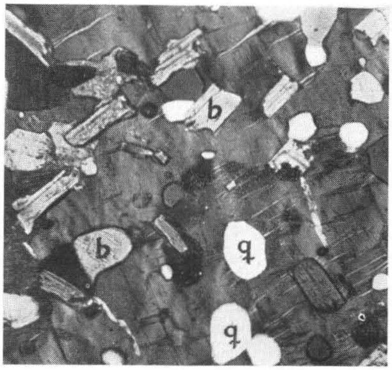


Fig. 2

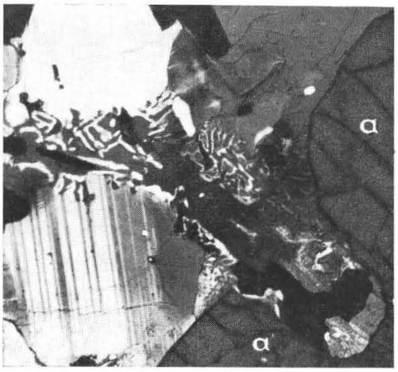
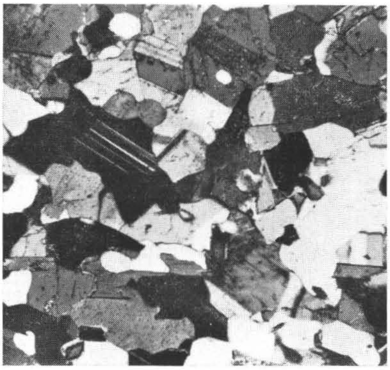


Fig. 3



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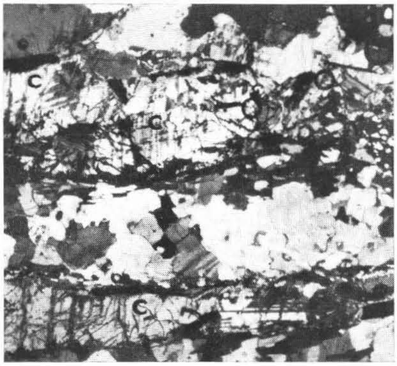

Fig. 5

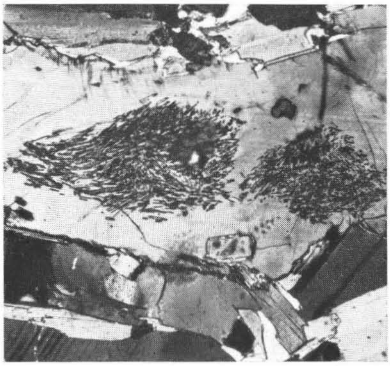

Fig. 6

Maunu Härme: Examples of the Granitization of Gneisses

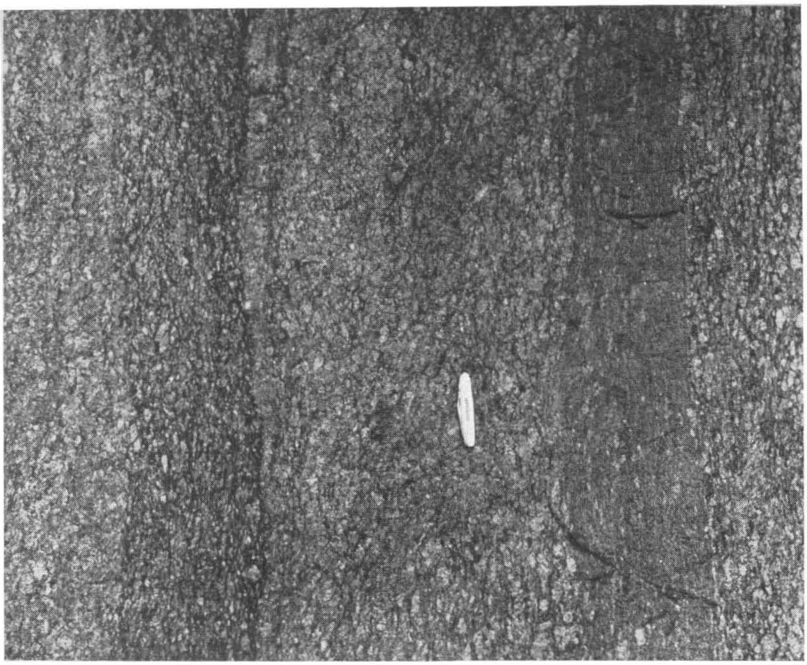

Fig. 7

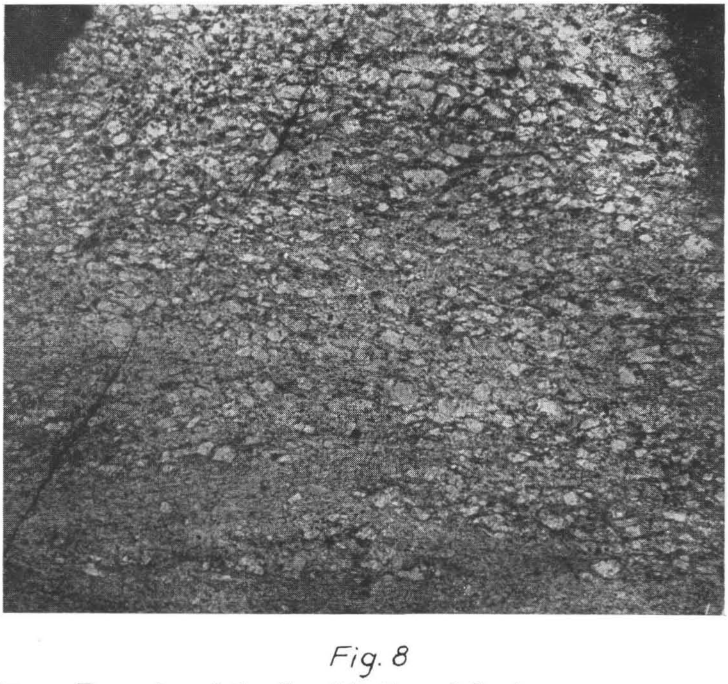

Fig. 8

Maunu Härme: Examples of the Granitization of Gneisses

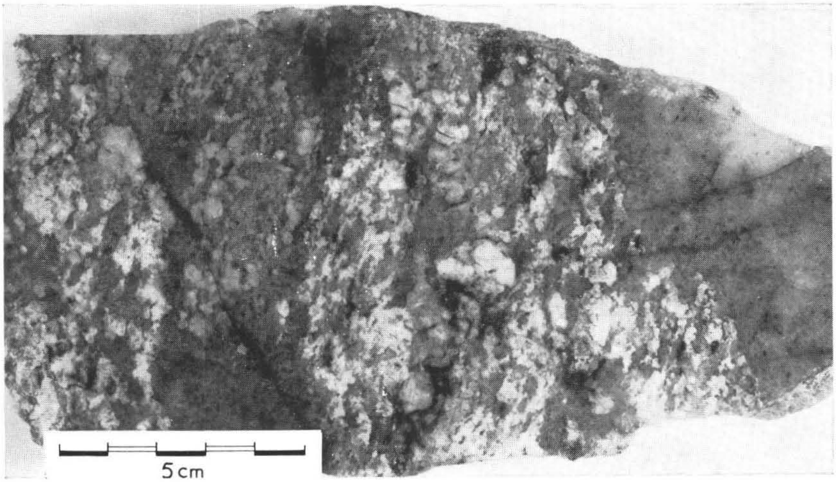

Fig. 9

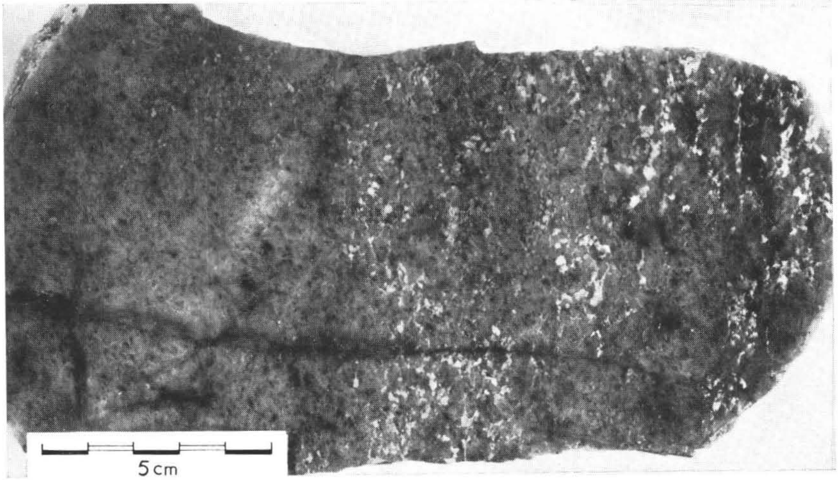

Fig. 10

Maunu Härme: Examples of the Granitization of Gneisses

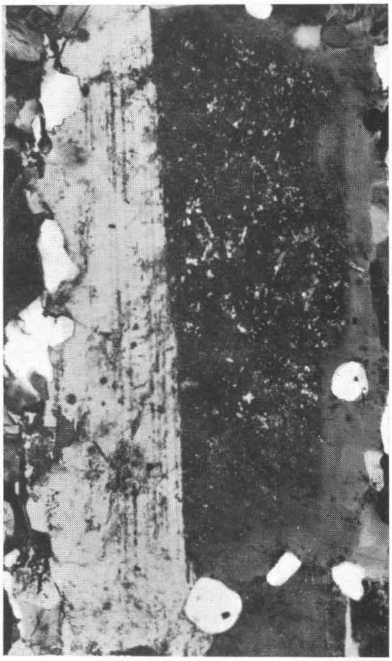

Fig. 11

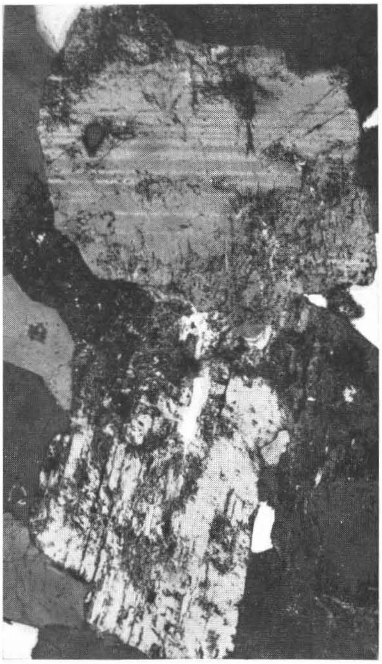

Fig. 12

Fig. 13

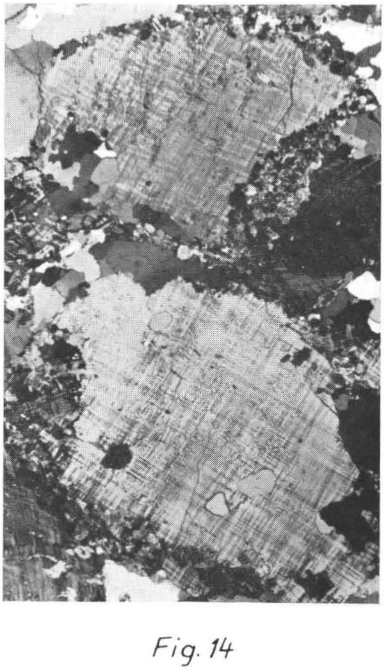

Fig. 14

Maunu Härme: Examples of the Granitization of Gneisses

5

ON THE PETALITE OCCURRENCES OF TAMMELA, SW-FINLAND

BY

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Geological Survey of Finland, Otaniemi

ABSTRACT

This paper deals with the bedrock of the vicinity of petalite occurrences found in the summer of 1957, the petalite deposits themselves and the minerals contained in them, the properties and chemical composition of petalite, x-ray studies, certain Li_2O-contents and the results of drilling in the largest of the occurrences.

CONTENTS

INTRODUCTION

Lithium is a rare metal included among the trace elements. According to Strock (1936, p. 190), the mean Li_2O-content of pyroxenite is 0.0004 %, of gabbro and basalt 0.002 %, of diorite and andesite 0.004 % and of granite

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[1] Received December 31, 1958.

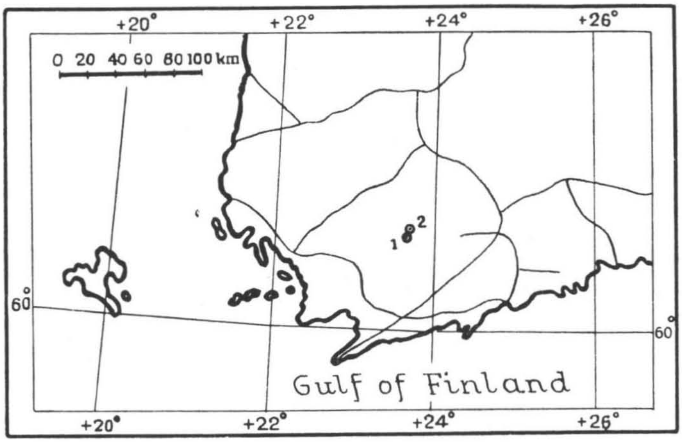

Fig. 1. Locations of petalite in Finland. 1. Somero; 2. Tammela.

and liparite 0.03 %. Sahama (1945, p. 72) has noted the same increase in the Li_2O-content upon going from basic to acid rocks also among the igneous rocks of southern Lapland. Lithium is thus enriched in the final products of the main stage of magmatic crystallization of igneous rocks, especially in granitic rocks. In granite and nepheline-syenite pegmatites, lithium is likely therefore to have become enriched to the extent of forming distinct minerals in them. Lithium deviates from other alkaline metals in that it does not form feldspars. The most important lithium minerals are spodumene, amblygonite-montebrasite, triphylite-lithiophilite, petalite and the Li-micas, lepidolite and zinnwaldite.

Among the lithium minerals mentioned, petalite is quite rare. Quensel (1957, p. 73) mentions petalite as having been met with in seven other localities in Europe besides Varuträsk. Petalite was found in Finland for the first time in 1945, the locality being Luolamäki, in the commune of Somero (Mikkola and Wiik, 1947).

While carrying out pegmatite investigations in the summer of 1957 for the Geological Survey in the commune of Tammela, I ran across a previously unknown petalite occurrence in the village of Kojo 6.2 km NNE from the petalite occurrence of Somero (Fig. 1). The occurrence was on the southern edge of the bog known as Torronsuo, where there is a rock, situated on the northern tip of Härksaari, rising 17 m above the surface of the bog. During the investigation it became known as Hirvikallio. The middle parts of the exposure consist of petalite-bearing pegmatite. In the summer of 1957 geological maps were made of Hirvikallio and its near vicinity, while trench samples were taken for chemical analyses. In the spring of 1958 one hole was drilled through the exposure to a depth of 33.9 m. The mineralogical and chemical studies were carried out during 1957 and 1958 at the laboratories of the Geological Survey.

In mapping the vicinity of Hirvikallio in the summer of 1957, Mr. Erkki Marttila, M. A., found two more small petalite occurrences on Lehtisaari, situated east of Hirvikallio.

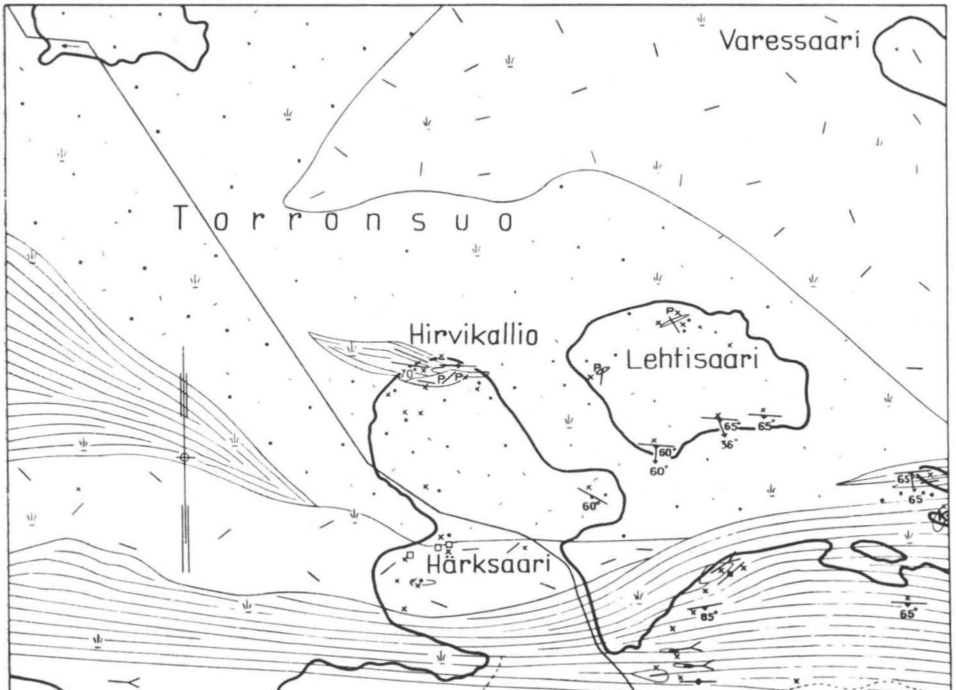

Fig. 2. The geological map of the southern part of Torronsuo, Tammela. 1. Granodiorite; 2. Pegmatite; 3. Garnet-diopside skarn; 4. Mica schist; 5. Uralite-plagioclase porphyrite, partially amphibolite; 6. Petalite; 7. Quarry; 8. Bog; 9. Outcrop; 10. Foliation and lineation.

BEDROCK

The bedrock of Somero—Tammela belongs to the deeply eroded region comprising the base of the Svecofennian mountain chain, where granitic rocks penetrate schists (Simonen, 1956, p. 6). Salient features of the geological map (Simonen, 1955) are the ragged schist belt formed by the amphibolites and the extensive migmatite areas.

The predominating rocks in the vicinity of the Tammela petalite occurrence are granodiorite, pegmatite, mica schist and, in places, uralite-plagioclase porphyrite containing amphibolite (Fig. 2).

The granodiorite occurring in the S-part of the area in the map is even-grained and weakly oriented. Its chief minerals are plagioclase, microcline, biotite and quartz; in some cases, there is hornblende in addition. In the contact zones of the granodiorite and the mica schist there occur in the mica schist granodiorite veins running roughly parallel to the schists and in the granodiorite sharp-edged schist fragments. The granodiorite contains some pegmatite nodules and pegmatite veins penetrating it. In the marginal parts of the pegmatite nodules it is possible to see granodiorite fragments.

On the N-side of the granodiorite there occurs in a narrow zone mica schist containing in many places plagioclase in addition to biotite and quartz. The schistosity is for the most part nearly W-E, sometimes also N45°E. The dip is usually vertical, but sometimes 65°S. The lineation is vertical. In addition to the granodiorite veins there are in the mica schist veins of granite running approximately parallel to the schistosity as well as pegmatite veins penetrating the mica schist. In one exposure on the very eastern edge of the map, the marginal zone of mica schist contains garnet-diopside skarn.

Rocks of volcanic origin comprise a large area in the center of the map. Amphibolite occurs only in the SE-part of Lehtisaari, elsewhere uralite-plagioclase porphyrite. These rocks are unhomogeneous, frequently layered and usually undergo a gradual alteration from one type to another. The plagioclase and uralite porphyrites are in places schistose in a nearly W by E direction, the dip varying from vertical to 60°S. In many instances these rocks have a massive character. The uralite grains vary in size between one and ten mm and the plagioclase grains 2—4 mm, the matrix consisting of fine-grained hornblende and plagioclase, besides which the matrix often contains slight amounts of biotite and carbonate. The same rock often has phenocrysts of both uralite and plagioclase. The pegmatite veins penetrate the porphyrites by cutting the bedding. At Härksaari there occurs in the contact of plagioclase porphyrite and pegmatite a shear zone 40 cm wide and oriented N84°E, 60°N.

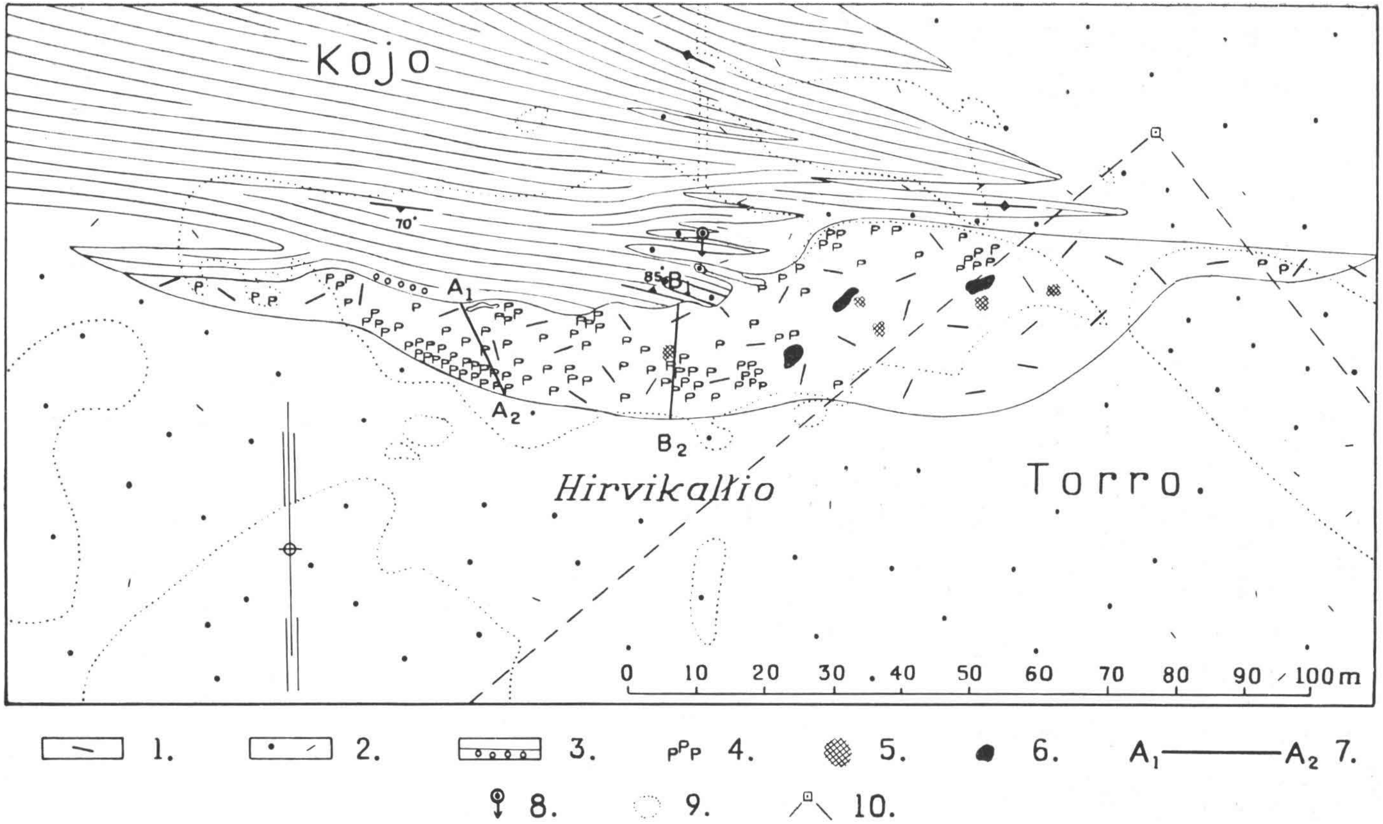

Fig. 3. The geological map of the Hirvikallio petalite occurrence. 1. Pegmatite; 2. Uralite-plagioclase porphyrite; 3. Mica schist, partially cordierite-bearing; 4. Petalite; 5. Microcline; 6. Quartz; 7. Specimen line; 8. Drill hole; 9. Outcrop; 10. Limit of village.

Pegmatites occur at numerous places in the area of the map. At Varessaari there is pegmatite containing aplitic portions, the chief minerals of which are red microcline, quartz, biotite, muscovite and tourmaline. Härksaari is penetrated from the west by pegmatite, the predominating mineral ingredients of which are likewise red microcline, quartz, biotite, muscovite and tourmaline. There are three pegmatite quarries on the Härksaari. The rock of the two quarries to the W consists mainly of graphic feldspar. Mäkinen (1913, pp. 37—96) run across on Härksaari gigantolite, chrysoberyl, heterosite, andalusite, tapiolite and apatite. In the quarry to the east, quartz and microcline are frequently separate, whereupon the largest, fairly pure microcline crystals are likely to be a meter in length. Small pegmatite nodules and veins are met with in the map area at several places, including, for example, Haiponmäki, where white microcline has even been quarried to a slight extent.

PETALITE OCCURRENCES

The most interesting pegmatite from the standpoint of the present study are the petalite-containing pegmatite veins of Hirvikallio and Lehtisaari. The pegmatite of Hirvikallio (Fig. 3) forms a vertical vein about 170 m long and 5—25 m broad, parallel to the prevailing schistosity, situated partly in the uralite-plagioclase porphyrite, partly between it and the mica schist. The contact between the pegmatite veins and the adjacent rocks can be seen well at numerous points. The contact is always sharp. In the S-part of the rock, the pegmatite cuts the porphyrite nearly perpendicularly to the bedding. The contact between the pegmatite and the mica schist follows by and large the schistosity of the mica schist. In the mica schist at a certain point near the contact with pegmatite there are abundant cordierite porphyroblasts. In the N-part of the pegmatite there is a mica schist fragment about 6 m long and 20 cm wide which is bent and cut by the pegmatite.

The largest portion of the Hirvikallio pegmatite vein comprises massive aplite, formed mostly of 1—2 mm long plagioclase grains and in many cases of microcline, quartz and muscovite as well. This rock contains large accumulations of petalite crystals and large microcline crystals and quartz nodules. In a few spots there occurs fine-flaked mica next to the petalite crystals. Especially at the ends of the vein, the aplite is oriented longitudinally. At such points the petalite crystals are also small.

The zoned structure characteristic of pegmatite occurrences (Cameron, 1949, p. 20) is not clearly evident in the Hirvikallio pegmatite. Nevertheless, some quartz nodules measuring several m^2 in size, situated at the E-part of the exposure in the vicinity of which there occurs coarse-grained microcline, may be regarded as cores.

The pegmatite veins of Lehtisaari are situated in the uralite-plagioclase porphyrite (Fig. 2). The petalite-bearing veins on the SW side can be seen running at a width of 2—4 m for 28 m and on the NE side at a width of 2—3 m for a distance of 60 m. On both sides of the last-mentioned vein are small pegmatite veins containing no petalite. The sharp contact between the pegmatite veins and the porphyrite is plainly visible.

The pegmatite vein on the SW-side of Lehtisaari consists mainly of petalite, in addition to which the rock contains small accumulations of quartz and plagioclase, spodumene and a little fine-flaked mica. The pegmatite vein containing petalite on the NE side resembles in structure the Hirvikallio occurrence. Here, too, the fine-grained, plagioclase-rich aplite contains large petalite and microcline crystals. Petalite occurs in the central part of the vein. No petalite is met with at the ends of the vein. In the pegmatite there is a quadrangular porphyrite fragment about 0.5 m long.

PETALITE

Mode of Occurrence. In most cases the petalite consists of accumulations in the aplite situated at the Hirvikallio and on the northeastern occurrence at Lehtisaari, accumulations measuring mostly between 0.5 and 2 m in diameter and in which the size of the petalite crystals crystallized in different directions ranges from 2 to 30 cm. Between the petalite crystals there generally occurs fine-lathed spodumene and, in spots, some muscovite. On the cleavage planes of the petalite crystals occurring on the surface layer of the rock at Hirvikallio there occurs a bluish-black precipitate containing manganese in abundance. No precipitate was found in the drill core. In both of the aforementioned occurrences petalite is met with as separate crystals, in places as much as 50 cm long.

Several portions of the Hirvikallio and the greatest part of the southwestern occurrence at Lehtisaari consists of rock between the petalite crystals of which, measuring 2—10 mm in diameter, there occurs fine-lathed spodumene. The rock in such spots has a net-like structure, the cords being spodumene and the mesh petalite.

Properties. The Tammela petalite is grayish white in color, and in spots just gray. Most usually the petalite is not transparent, but also a bright and transparent type is met with. Crystals elongated parallel to b. The cleavage of the petalite in the (001) direction is exceptionally well developed. In addition, a cleavage has been noted in the following directions: (201), (010), (100), (011) and (101). The luster on the (001) cleavage plane is pearl-like and on the rest glass-like. On the (100) cleavage planes the cleavage in the (001) direction appears as parallel lines. Crystal faces have not been met with in the petalite. $D_{4^\circ}^{20^\circ} = 2.366 \pm 0.003$ (pycnometer).

The refractive indices and optical constants (for Na-light) for the bright, transparent petalite (analyzed) are as follows: $\alpha = 1.506 \pm 0.001$, $\beta = 1.513 \pm 0.001$ and $\gamma = 1.523 \pm 0.001$; $(\gamma - \alpha) = 0.017$. $2V\gamma = 84°$, and $a \wedge \alpha = 9° \pm 1°$. The axial plane $\perp$ (010). Optically positive. In thin section the Tammela petalite in many instances reveals a polysynthetic twinning; the twinning plane is according to (001) as recently pointed out by Mikkola and Wiik (1947, p. 282) and Ginzburg & Guschtschina (1954).

The petalite is seldom altogether pure. The most common impurity occurring is spodumene, but in addition the petalite contains as small inclusions mica, apatite, microcline and plagioclase.

Table 1. Chemical Composition of Petalite from Hirvikallio, Tammela. Analyst A. Heikkinen.

Constituent	wt %	Mol. prop.	Cations per 10 Oxygen atoms
SiO_2	78.00	12 987	Si = 3.981
TiO_2	0.00		
Al_2O_3	17.03	1 671	Al = 1.024
Fe_2O_3	0.01	1	Fe^{3+} = 0.001
FeO	0.03	4	Fe^{2+} = 0.001
MnO	0.00		
MgO	0.00		
CaO	0.00		
Li_2O	4.74	1 586	Li = 0.973
Na_2O	0.07	11	N = 0.007
K_2O	0.05	5	K = 0.003
P_2O_5	0.02	1	P = 0.001
CO_2	0.00		
H_2O+	0.04	22	H = 0.013
H_2O-	0.00		
Total	99.99		

Table 2. Powder Diffraction Data for Petalite from Hirvikallio, Tammela.

d	I/I_0	d	I/I_0	d	I/I_0
6.99	1	2.41	5	1.21	5
5.39	1	2.07	15	1.14	5
4.63	1	1.93	10	1.13	20
4.21	1	1.90	10	1.06	15
3.73	100	1.80	1	1.02	10
3.66	100	1.76	1	1.01	15
3.51	20	1.73	5	1.00	5
3.10	5	1.72	1	0.99	5
3.06	1	1.63	5	0.95	10
2.98	10	1.53	1	0.94	10
2.95	1	1.48	5	0.90	10
2.71	1	1.45	10	0.89	5
2.57	5	1.29	1	0.88	5
2.54	5	1.23	5	0.85	1

Chemical Composition. The purification of the petalite for chemical analysis was performed by centrifuging the material in Clerici's solution. The pure sample was chemically analyzed by Mr. A. Heikkinen, M. A., at the Geological Survey of Finland. The result of the analysis appears in Table 1. This table also lists the number of atoms when the number of oxygen atoms is scaled to ten. The chemical formula of the analyzed petalite, as based on the number of atoms, is:

$$(Li_{.97}\ Na_{.01}\ H_{.01})\ \ Si_{3.98}\ Al_{1.02}\ O_{10}$$

The formula of petalite in Winchell (1947, p. 309) is $Li(Si_4Al)O_{10}$.

X-Ray Examination. A powder pattern was prepared of the material used for analysis. The apparatus used was the Philips Geiger counter x-ray diffractometer, CuK-radiation with Ni-filter. The powder diffraction data are given in Table 2. The intensities are calculated from the areal measurements of the peaks.

MINERAL PARAGENESIS OF PETALITE

In addition to the petalite, the following minerals have been met with in the petalite occurrences of Tammela: microcline, plagioclase, quartz, muscovite, spodumene, heterosite, topaz, apatite, tourmaline and garnet.

Microcline perthite occurs in the aplite of Hirvikallio as small crystals. In addition, it occurs as large crystals, especially in the E-part of the occurrence. In color the microcline is a slightly yellowish pink or a dirty white. The microcline of Hirvikallio has been spectrographically observed to contain rubidium.

Plagioclase is the chief mineral of aplite. It is rich in albite, white, bright in thin section and invariably twinned. When plagioclase occurs as inclusions in petalite, in many cases it contains myrmekite at the edges.

There is little quartz in association with large petalite crystals, but quartz is common in aplite. In sheared portions of pegmatite, quartz occurs in the form of veins running roughly parallel to the schistosity. At a certain point in the margin of the pegmatite at Hirvikallio, quartz occurs to a breadth of 10 cm against the adjacent rock, and in the eastern part of the occurrence there are quartz nodules.

Muscovite is a common mineral in the petalite occurrences of Tammela, but the total amount is small. This is due to the small size of the muscovite flakes. The largest muscovite flakes, measuring 1—2 cm in length, occur in the sheared portions of the pegmatite and in the proximity of large microcline crystals.

Spodumene is a common mineral at Hirvikallio, although the total amount of it is small. It occurs mostly in association with petalite. Most commonly spodumene forms lamellae 0.5—5 mm broad situated at intervals of 0.5—3 cm from each other. The spodumene lamellae are in most cases oriented in the petalite in a (100) direction, but also lamellae in a (001) direction are met with. In the lamellae the spodumene appears in the form of laths which are generally perpendicular to the lengthwise position of the lamellae. The same mode of occurrence is observed for situated between the spodumene laths in the broadest lamellae (Fig. 4). In numerous instances, the spodumene also occurs in the petalite as thin veins of indefinite shape.

Sometimes spodumene occurs as a net-like formation around small petalite crystals, as has been pointed out in the foregoing discussion on petalite. In such cases, the fine-lathed spodumene has sometimes spread within the petalite crystals and in part fill them. Spodumene has also tended to spread over the mica layer surrounding the petalite accumulations and mixed with the plagioclase. In the drill core there has been found, contained within aplite, a piece about 10 cm long consisting of fine-lathed spodumene and slight amounts of quartz, plagioclase and mica.

In the pegmatite vein situated at the SW side of Lehtisaari, fine-lathed spodumene is a common mineral, but in the other petalite occurrence on the Lehtisaari there are only few thin spodumene lamellae.

Heterosite is present in the sheared plagioclase-quartz-muscovite pegmatite situated at the ends of Hirvikallio as dark, purple-red aggregates measuring a few millimeters in diameter. The heterosite is markedly pleochroic: purplish red—dark brownish gray—reddish brown. The mineral has two

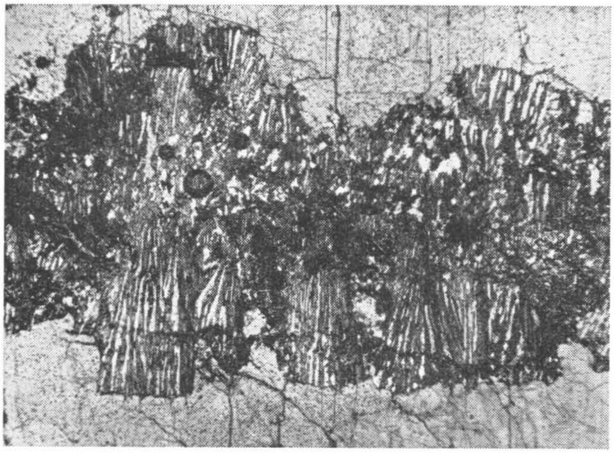

Fig. 4. Spodumene-quartz vein in the Hirvikallio petalite. Grey = petalite; dark = spodumene; light = quartz. One nic. Magn. 22×. Photo Erkki E. Halme.

Table 3. Powder Diffraction Data and Unit Cell Dimensions for Heterosite from Hirvikallio, Tammela.

hkl	d	I/I_0	hkl	d	I/I_0	Unit Cell Dimensions
(020)	4.85	m	(140)	2.16	w	Orthorhombic
(110)	4.27	s	(132)	1.96	w	$a_0 = 4.76$ Å
(021)	3.72	m	(142)	1.73	w	$b_0 = 9.70$ Å
(111)	3.45	s	(133)	1.57	w	$c_0 = 5.83$ Å
(002), (121)	2.91	s	(043)	1.52	w	
(130)	2.68	w	(004), (242), (321)	1.46	m	
(131)	2.43	s	(330)	1.42	w	
(210)	2.31	w	(162)	1.34	w	
(041)	2.22	w				

perpendicular cleavage planes and a straight extinction parallel to them. The powder diffraction data and unit cell dimensions are given in Table 3. The dimensions of the unit cell tally with those of the heterosite of Varuträsk (Quensel, 1957, p. 42), which are in Ångströms: a = 4.77 Å, b = 9.70 Å and c = 5.83 Å.

Small quantities of three, up to now unidentified minerals have been discovered in association with heterosite.

Topaz has been met with in the aplite of Hirvikallio. Usually it has crystallized poikilitically between plagioclase grains. The topaz contains inclusions of plagioclase, quartz and mica. Certain of the topaz grains have begun to alter at the edges and cracks into a fine, mica-like mineral.

Slight amounts of apatite have been found in the Hirvikallio aplite near the contacts with mica schist as well as in the tourmaline-bearing aplite of the drill-core.

There is little tourmaline in the pegmatite of Hirvikallio. It is to be seen in the central part of the aplite exposure as black crystals 2—10 mm long, as well as in certain portions at the contacts of pegmatite and mica schist as zones 1—2 cm thick.

There is some garnet in the Hirvikallio aplite, where it occurs as aggregates of grains 2—3 cm in diameter and which in addition contains quartz.

Li_2O-CONTENT OF HIRVIKALLIO

In addition to the chemical analysis made from pure petalite, two total analyses have been made from the Hirvikallio occurrence as well as several other chemical determinations. All the analyses and separate Li_2O-, CaO-, Na_2O- and K_2O-determinations have been carried out by Mr. A. Heikkinen, M. A. The BeO- and Rb-determinations have been performed by Mr. A. Löfgren, M. A., spectrographically.

Table 1 shows the Li_2O-content of pure, separated petalite to be 4.74 %, which comes quite close to the theoretical value of 4.88 %. In addition to lithium pure petalite contains only slight amounts of other alkalis.

Two trenches were quarried across the occurrence, namely, A_1—A_2 and B_1—B_2 (Fig. 3). Samples were taken separately from both trenches

Table 4. Chemical Composition of General Samples from Hirvikallio, Tammela. Analyst. A. Heikkinen.

Constituent	Sample A_1—A_2 wt %	Sample B_1—B_2 wt %
SiO_2	73.65	73.76
TiO_2	0.00	0.00
Al_2O_3	16.76	16.94
Fe_2O_3	0.14	0.11
FeO	0.12	0.16
MnO	0.06	0.07
MgO	0.00	0.01
CaO	0.34	0.44
Li_2O	2.06	1.54
Na_2O	2.60	3.53
K_2O	3.12	2.69
P_2O_5	0.35	0.42
CO_2	0.00	0.00
H_2O+	0.57	0.51
H_2O-	0.08	0.09
Total	99.85	100.27

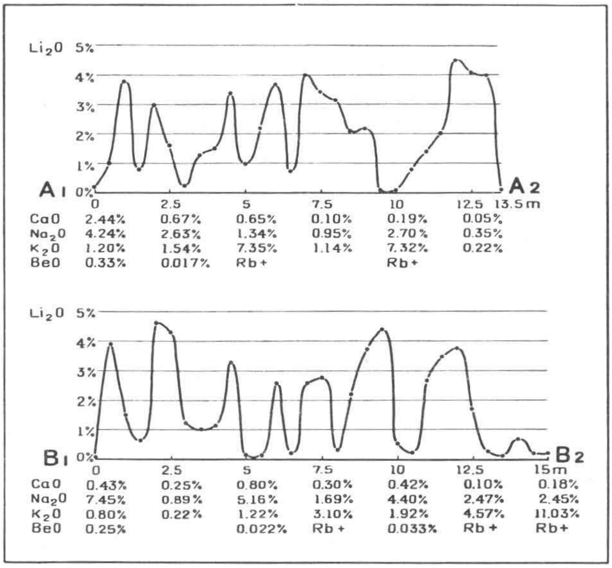

Fig. 5. The Li_2O- and other separate determinations made from trenches A_1—A_2 and B_1—B_2.

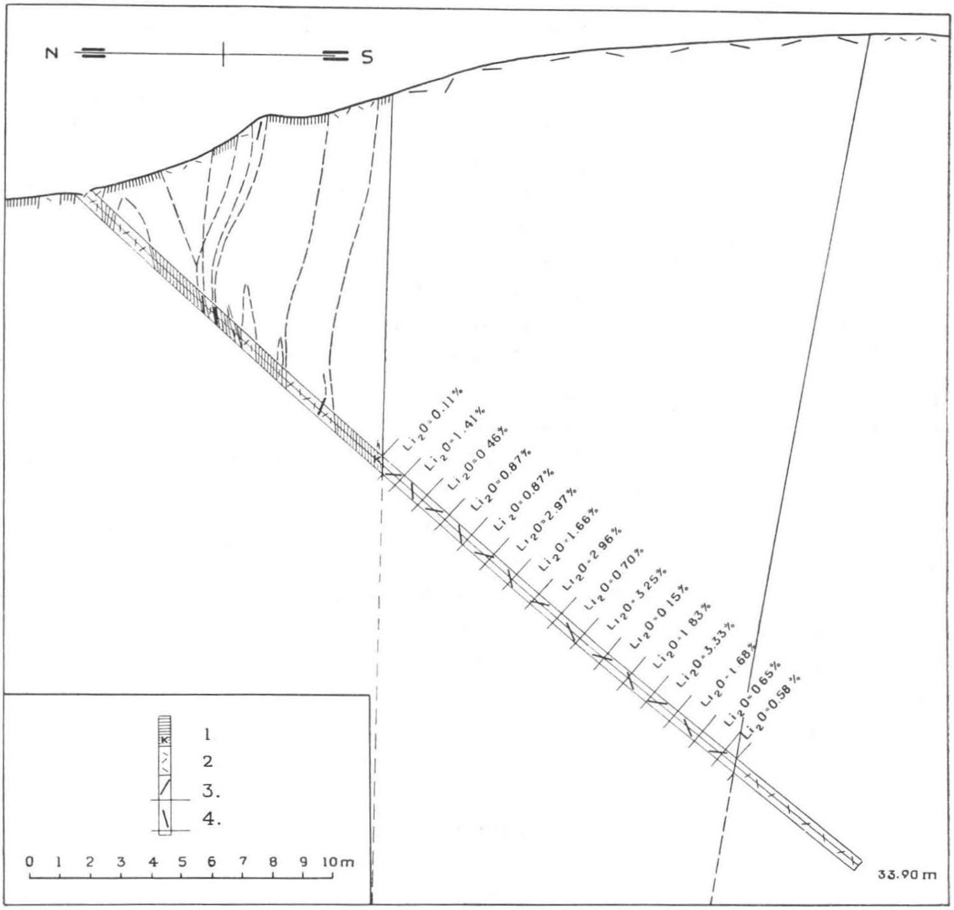

Fig. 6. The drill profile of the Hirvikallio petalite occurrence and the Li_2O-contents determined from the drill-core. 1. Mica schist and skarn (K); 2. Uralite-plagioclase porphyrite; 3. Pegmatite; 4. Analyzed portion.

in such a way that a general sample was taken every two meters, and these samples were then combined and ground to powder. The analyses made from the two general samples are presented in Table 4. These analyses essentially deviate from the analysis of the pure petalite only on account of their larger contents of Na_2O and K_2O and smaller contents of SiO_2. The Li_2O-content of trench A_1—A_2 is 2.06 % and that of trench B_1—B_2 1.54 %.

From the same trenches samples for analysis were further taken at 0.5 m intervals and from all of them the Li_2O-content was determined. From every fifth sample the CaO, Na_2O and K_2O were likewise determined and the occurrence of beryllium and rubidium was investigated. The variation in Li_2O-contents in both trenches is presented graphically in Fig. 5. The other results are also shown in the same figure. It makes evident that the Li_2O-content varies greatly, ranging from values corresponding to pure petalite to a few tenths of a per cent. This is due to the mode of occurrence of petalite in pegmatite. Petalite, after all, occurs most commonly at Hirvikallio as accumulations of crystals 0.5—2 m in diameter, in the aplite in between them of which there are generally only small petalite crystals. The mean value of the Li_2O-contents thus obtained from the A_1—A_2 trench is 2.02 % and from trench B_1—B_2 1.79 %.

The variation in the Li_2O-contents in the interior parts of Hirvikallio may be studied in Fig. 6, presenting the drilling profile. A hole marked in Fig. 3 was drilled through the pegmatite in a N—S direction; penetrating the pegmatite it met the contacts at depths of 12.3 and 24.4 m. The portion of the drill-core consisting of pegmatite was split and cut to pieces for analysis. The Li_2O of the drill-core portions thereby obtained was determined. The contents obtained are marked at the corresponding intervals in the figure. The mean of the Li_2O-contents of the drill-core is 1.47 %.

Calculating the mean of the mean values obtained in the manner described in the foregoing and the Li_2O-contents of the general samples taken from the trenches, we obtain 1.78 % Li_2O. This may be regarded as the mean Li_2O-content of Hirvikallio.

SUMMARY

Hirvikallio is the largest petalite occurrence found up to the present in Finland. It is estimated that to a depth of 25 m it contains over 100 000 ton of pegmatite. According to the mean Li_2O-content of the occurrence, approximately 30—40 % of this amount consists of petalite. Hirvikallio is complemented by the nearby petalite occurrences of Lehtisaari, the one situated on the SW-side quite likely continuing westward. No signs of it, however, have been met with on the Härksaari side.

The mineral paragenesis, petalite and the associated spodumene of the Hirvikallio and Lehtisaari pegmatites indicate the lithium-minerals to have crystallized, in accordance with Quensel (1957, Plate II), at the high temperature phase of the beginning of the pneumatogenic stage following the pegmatitic stage. The pneumatogenic stage is between the inversion point of quartz and the critical point of water, i. e., about 600°—400° C. Montebrasite belonging to the pegmatitic stage, which is associated at Varuträsk (Quensel, 1957, p. 91) with the intermediate zones, has not been met with so far in the petalite occurrences of Tammela.

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ON THE PETROLOGY AND MINERALOGY OF SOME ALKALINE VOLCANIC ROCKS OF TOROR HILLS, MT. MOROTO, AND MORULINGA IN KARAMOJA, NORTHEASTERN UGANDA[1]

BY

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PREFACE

The material that forms the subject of the present investigation is part of the material of volcanic rocks from East Africa collected by two expeditions, *viz.* in 1952 (Prof. Th. G. Sahama, Institute of Geology and Mineralogy, University of Helsinki; Dr. K. J. Neuvonen, of the Geological Survey of Finland; the present author) and in 1954 (Prof. Th. G. Sahama and the author). The material was examined during the period of 1953—1958 in the laboratory of the Institute of Geology and Mineralogy, University of Helsinki.

In his work the author received valuable assistance in different forms from a number of persons and organizations.

The author received advice during all stages of the work from Professor Th. G. Sahama, who also contributed valuable criticism of the manuscript.

For the field work in Uganda the facilities of the staff of the Geological Survey of Uganda, Entebbe, were of a great value. The author would like to mention especially Dr. C. G. B. Du Bois, who kindly gave important assistance during the excursions in the Toror Hills area. During the later stage of his study, in May 1958, the author made a visit to England, where the facilities of the Department of Mineralogy and Petrology, University of Cambridge, of the Department of Geology and Mineralogy, University of Oxford, and of the Department of Mineralogy, British Museum, London, were placed at his disposal. In making some of the optical determinations, the author had the opportunity to use the laboratory of the Mineralogical Institute, Stockholm University, Sweden, permission having been obtained through the courtesy of Professor Sven Gavelin.

For comparative studies, specimens of different kind of volcanic rocks and minerals were placed at the author's disposal by the following gentlemen: Professor C. E. Tilley and Dr. S. O. Agrell, both of the University of Cambridge, Dr. M. H. Hey, of the Department of Mineralogy, British Museum, London, Dr. Henrich Neumann, of the Mineralogisk-Geologisk Museum, Oslo, Norway.

The chemical analyses were made by Dr. H. B. Wiik, Mr. Pentti Ojanperä and Mr. Aulis Heikkinen, all of the Geological Survey of Finland, and by Dr. Aarno

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[1] Received February 27, 1959.

Juurinen, formerly of the Institute of Geology and Mineralogy, University of Helsinki. Dr. K. J. Neuvonen gave advice in interpreting the powder patterns of the feldspars. The maps were drawn by Mrs. K. Mansala. The English text was corrected by Mrs. J. E. Agrell, Cambridge, England.

To cover some of the costs of the investigation, the author received financial aid from the following organizations: Outokumpu Company Foundation (Outokumpu Osakeyhtiön Säätiö), Emil Aaltonen Foundation (Emil Aaltosen Säätiö), University of Helsinki, The British Council.

To all the persons and organizations mentioned, the author expresses his gratitude.

Finally, the author wishes to thank the Geological Society of Finland for accepting this paper for publication in the Bulletin of the Society.

ABSTRACT

The present investigation deals with the petrology and mineralogy of the Tertiary alkaline volcanic rocks of the Toror Hills (natrolite- and analcite-bearing tinguaites), Mt. Moroto (mainly phonolites, trachytes, trachybasalts and olivine-melanephelinites) and Morulinga (olivine-melanephelinite), all in Karamoja, northeastern Uganda. Five rock analyses and nine mineral analyses (nepheline, alkali feldspar, three clinopyroxenes, basaltic hornblende, two olivines, 'iddingsite') are presented. Optical properties, unit-cell dimensions and powder patterns are given for the minerals analyzed. Some optical and X-ray data are presented for a number of other minerals. Special attention is given to the crystallization history of the Toror Hills tinguaites.

CONTENTS

INTRODUCTION

The district of Karamoja is situated in the northeastern part of the Uganda Protectorate (Fig. 1). To the north, the district is bordered by the Republic of Sudan, to the east, by Kenya Colony. The neighboring districts of Uganda are Bugishu and Teso in the south, and Lango and Chua in the west.

The volcanic mountains that form the subject of this investigation occur as three separate well-marked topographical and geological units rising from the slightly undulating Karamoja plain (average altitude ca. 1200 m) in the central part of the district. The Toror Hills (1945 m) (Fig. 2) and especially Mt. Moroto (ca. 3000 m) are relatively large mountains, both in height and in extent, whereas Morulinga (1340 m) (Fig. 3) is a small hill covering an area of ca. 1 km^2. The highest point of the Toror Hills (Toror Peak) is situated at latitude 2°50′ N and longitude 34°12′ E. Mt. Moroto lies between latitude 2°25′ and 2°38′ N and longitude 34°39′ and 34°49′ E. Morulinga is situated at latitude 2°25′ N and longitude 34°28′ E.

PREVIOUS GEOLOGICAL WORK

The Toror Hills and Mt. Moroto represent two of the five large eastern volcanoes of Uganda. The other three volcanoes are Elgon (4321 m), Kadam (= Debasien; 3063 m) and Napak (2530 m). Elgon, Kadam and Moroto lie close to the Karamoja-Turkana scarp of the Eastern Rift Valley. Napak and the Toror Hills lie farther to the west of the scarp. With the exception of Elgon, all are situated within the district of Karamoja.

The first geological observations on Karamoja were presented by Wayland in 1920 [1]. In that paper, the mountains and hills scattered over the southern and central parts of Karamoja were considered to be of volcanic origin. In the first geological sketch map of Uganda, published under the direction of Wayland (1925), the positions of both the Toror Hills and Mt. Moroto are shown. In a report published in 1926 the same author notes the presence of a limestone occurrence in the Toror Hills (Wayland, 1926). Wayland (1929) gives short notes referring to the general petrological character of the eastern volcanic formations of Uganda.

Later, Davies and Bisset visited the area. The name Morulinga is included in the list of the Tertiary volcanic localities given by Davies (1935). The description of Bisset (1935) is noteworthy because it gives a few details of the geology of Mt. Moroto. In fact, this is the only paper available to the author with a description of this volcano. According to Bisset, the three occurrences,

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[1] Wayland, E. J., 1920, Annual report of the Geological Survey of Uganda Protectorate, p. 35. — Not available to the author. Reference from Du Bois (1956).

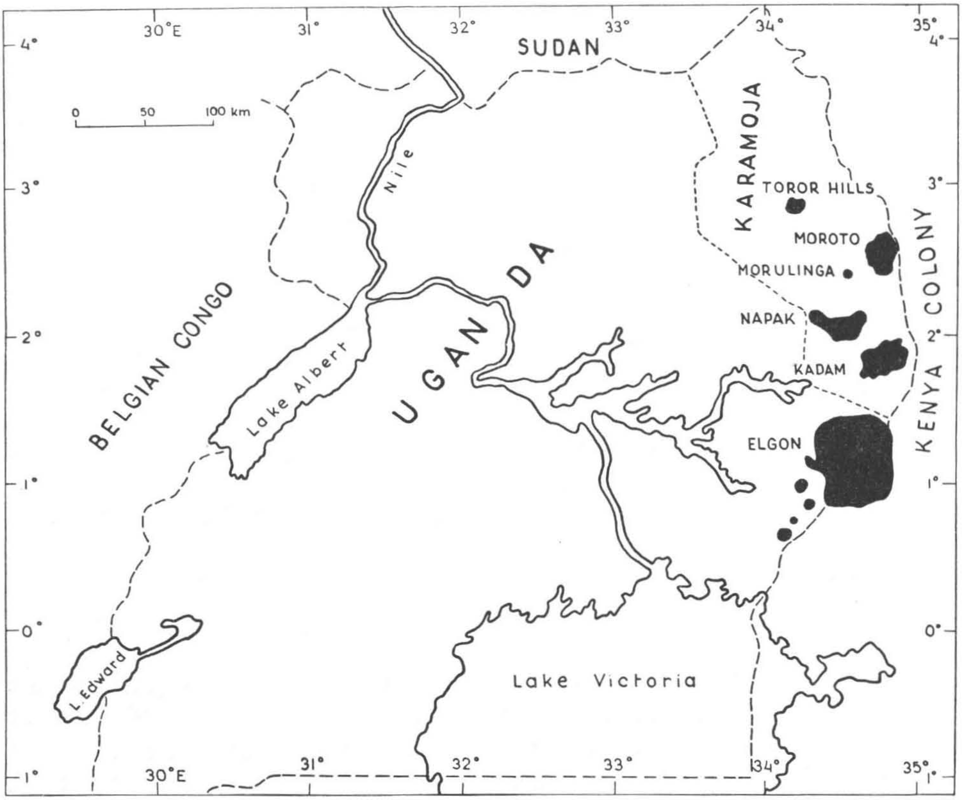

Fig. 1. Map of Uganda showing the District of Karamoja. Black: volcanics of eastern Uganda.

Napak, Moroto and Toror Hills, are regarded as »Essentially similar in character and probable mode of origin, but as having suffered widely varying degrees of denudation. At Moroto the lavas, although doubtless considerably reduced in extent and thickness, still form a continuous mass; at Napak denudation has gone a step further and has cut the original mountain into a number of blocks. Finally at Toror the lavas have disappeared altogether and the intrusion is represented only by dyke and allied types». In addition, Bisset notes the occurrence of limestones at Napak and Toror Hills and remarks upon their similarity to those from Tororo (in Budama, Uganda) and southern Nyasaland (Dixey, Campbell Smith and Bisset, 1937; revised edition 1955).

The descriptions mentioned above give only a few lines to the general character of the geology of the volcanic formations which form the subject of the present study. Recently, however, a very important paper dealing with the geology of the Toror Hills has been presented by Du Bois (1956). The paper gives a detailed description of the field relations and the emplacement

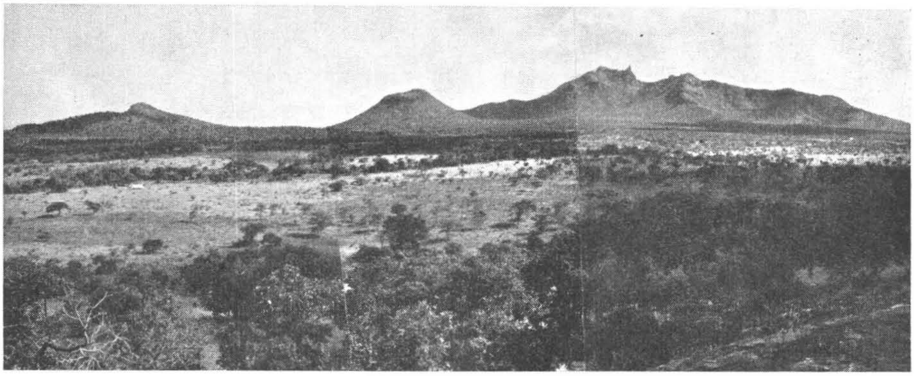

Fig. 2. General view of the Toror Hills from the top of North Amairoi.

of the complex and of the petrology, geochemistry and petrogenesis of the rocks. According to Du Bois, the Toror Hills constitute the eroded remnants of a central-type volcano and its related hypabyssal phases (Fig. 4). The complex is intruded into granitic gneisses of the basement. Structurally, the Toror Hills consist of a central vent choked with phonolitic agglomerate surrounded by an irregular ring-shaped body of intrusive carbonatite of magmatic origin which is, in turn, bounded by a zone of metasomatised country rock. All the rocks are cut by numerous phonolitic and tinguaitic dykes and plugs. The Toror Hills complex is related in time to the other volcanic mountains in northeastern Uganda, especially to the Napak, the volcanic period of which is dated as Middle Tertiary or Pliocene (King, 1949).

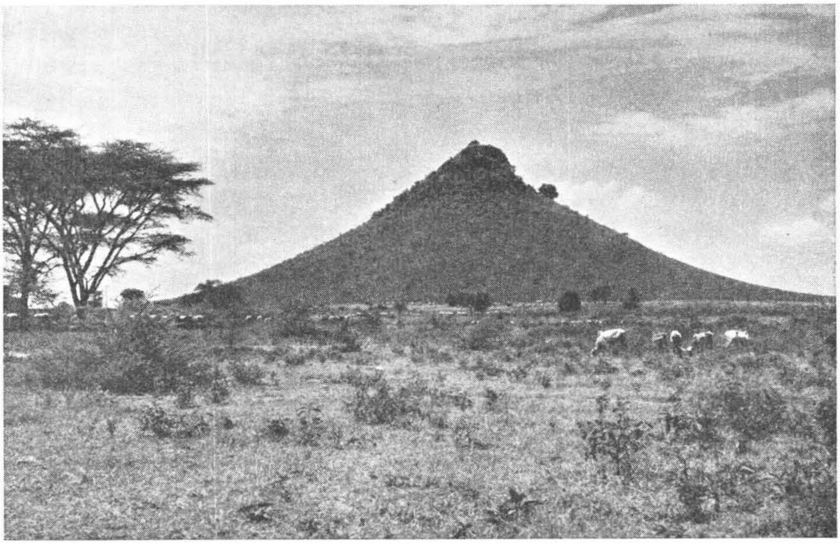

Fig. 3. The volcanic neck of Morulinga from the north.

SCOPE OF THE PRESENT STUDY

The present investigation deals only with the petrology and mineralogy of the volcanic rocks and tries to present some details in addition to those given by Du Bois (1956) of the Toror Hills and to give an idea of the general character of the rocks of Mt. Moroto and Morulinga. No investigations have been made on the field relations of the volcanic complexes and of the surrounding basement complex, nor have the carbonatite of the Toror Hills and the pyroclastic rocks been studied.

The specimens of 1952 carry the symbol FEAE (= the Finnish East African Expedition) and those of 1954 the symbol H (= Hytönen).

METHODS

The separation of the minerals was carried out with Clerici's solution in a centrifuge and with a Frantz-type isodynamic separator. The purity of the minerals to be analyzed was tested by means of powder patterns and under the microscope using immersion liquids of approximately the same refractive indices as those of the minerals. The powder patterns were recorded with a Philips Norelco Geiger Counter diffractometer. For single-crystal X-ray measurements, rotation and Weissenberg photographs were taken. All optical properties are given for sodium light. The accuracy of the refractive indices is ± 0.002. The determinations of the compositions of the plagioclases were based on the refractive indices using the curves of Chayes (1952). For the determination of the specific gravity of the minerals, a pycnometer was used. The results were corrected to correspond to the temperature of 20° C as compared with the specific gravity of water at 4° C. The specific gravities of the bulk rocks were determined from small pieces (ca. 100 grams) that were weighed in air and in water. The planimetric analyses of the rocks were made with the Leitz Integration Stage. The zeolites were identified by powder patterns.

The determinations of the composition of olivine, nepheline and alkali feldspar were made by powder patterns using the method of Yoder and Sahama (1957) for olivine, that of Smith and Sahama (1954) for nepheline and that of Bowen and Tuttle (1950) for the alkali feldspar. The determinations of the alkali feldspars were made from the original samples and from samples heated for 24 hours at 900° C and quenched in water.

The determination of the composition of nepheline needs the following comment. Smith and Sahama *(op. cit.)* recommend that the curves and equations based on data for the synthetic nephelines be used also for the determinations of the K/ (K + Na + Ca) ratio for natural nephelines. However, the curves of the synthetic nephelines deviate in the composition range between 9 and 20 mol. % Ks from almost all the points of the natural analyzed nephelines by having lower Ks contents *(op. cit.*, Fig. 1, p. 442). The deviation

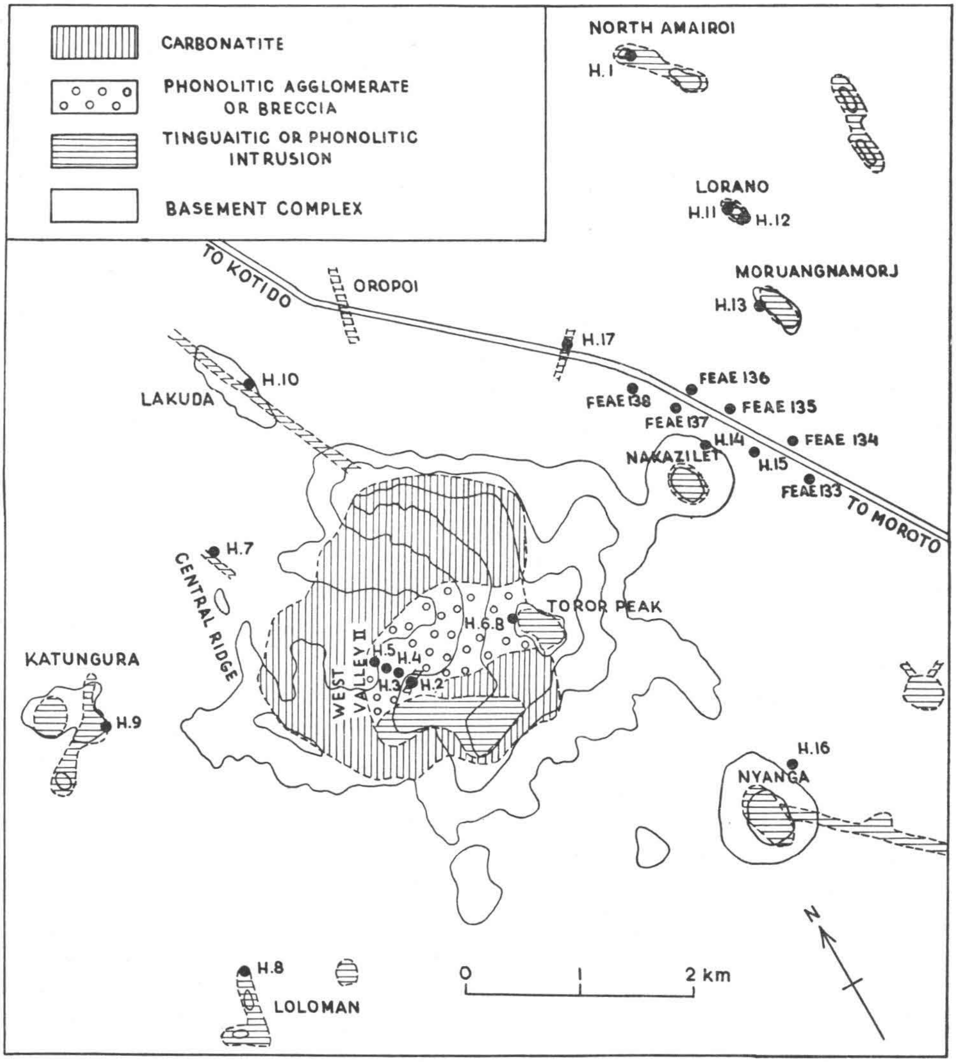

Fig. 4. Geological map of the Toror Hills simplified from Du Bois (1956) showing the location of specimens studied.

is of the order of ca. 3 mol. % Ks. Accordingly, the values for K/(K + Na + Ca) calculated from the equations proposed by these authors will presumably be ca. 3 % too low. The existence of a deviation is supported by the fact, that the composition deduced from powder pattern for the analyzed nepheline of the specimen FEAE No. 134 (Toror Hills) is 15 mol. % Ks, whereas the analysis shows 18 mol. % Ks. In addition, for the nephelines of specimens FEAE No. 116 (Mt. Moroto) and FEAE No. 136 (Toror Hills) the equations give 9 mol. % Ks and 10 mol. % Ks, respectively. These values are lower than is known for any natural nepheline at present. For these reasons, the compositions of the nephelines in the composition range between 9 and 20 mol. % Ks are indicated in this paper by numbers to which 3 mol. % has been added to the value obtained by using the equation of Smith and Sahama. For the values exceeding 20 mol. % Ks, no correction has been made.

TOROR HILLS

The 22 specimens collected by the author in the Toror Hills area are completely crystalline porphyritic undersaturated rocks of the feldspathoidal syenite clan. The dominant minerals are nepheline, alkali feldspar, natrolite, analcite and aegirine-augite. In addition, the following less abundant constituents have been identified: sphene, pectolite, apatite, biotite, apophyllite, fluorite, diopside, zircon, sodalite, calcite, basaltic hornblende and cancrinite.

In mineral composition, the rocks are phonolitic. As, on the other hand, most of the specimens show a well-developed tinguaitic texture and as definitely no extrusive feldspathoidal rocks but only their hypabyssal equivalents are present at Toror Hills (Du Bois, 1956), the term phonolite has not been used and the term tinguaite is adopted here.

The most characteristic feature of the tinguaites of the Toror Hills is the abundance of the zeolites, natrolite and analcite. These two minerals are, at least in part, primary or deuteric constituents of the rocks. Every sample studied contains either natrolite or analcite or both two. Accordingly, the classification of the tinguaites is based on the presence and amount of these minerals and the terms natrolite-tinguaite, natrolite-analcite-tinguaite, analcite-natrolite-tinguaite and analcite-tinguaite have been used here following mainly the terminology proposed by Johannsen (1938).

This classification differs from that of Du Bois *(op. cit.)* who divides the volcanic rocks of the Toror Hills into five classes: phonolite, tinguaite, nephelinite, microfoyaite and trachyte. Of these rock types the nephelinite, the microfoyaite and the trachyte are rare and have been found only in a few localities. The author did not visit these localities and, therefore, no specimens of these rock types were available for the present investigation. The distinction between phonolites and tinguaites in the Toror Hills is based by Du Bois on textural differences. The mode of occurrence is not regarded as essential as is clearly stated in his paper. Accordingly, the term tinguaite is reserved by Du Bois for rocks with a distinct tinguaitic texture whereas the other feldspathoidal feldspar-bearing rocks with the same mineral composition, but without the tinguaitic texture, are called phonolites.

In the following description of the tinguaites of the Toror Hills, each specimen will be described separately. As all the tinguaites contain alkali feldspar and aegirine-augite in the groundmass, the presence of these minerals in any of the specimens will not always be mentioned.

NATROLITE-TINGUAITES

The specimens are as follows: FEAE No. 134, FEAE No. 137, H. 2, H. 8, H. 14 and H. 17. These natrolite-tinguaites have been so named because of the large amount of natrolite recognizable in thin section. The specimen FEAE No. 134 was selected as a type specimen for the natrolite-tinguaites of the Toror Hills. A detailed description of this specimen is given below. At the end of this paragraph, short descriptions of the other five natrolite-tinguaites are given.

Specimen FEAE No. 134 *(with accessory sodalite). Sp. gr. 2.58.* Loose boulder on the Moroto-Kotido road, close to Nakazilet. In hand specimen the rock is compact and porphyritic, without any preferred orientation. The euhedral phenocrysts of nepheline and alkali feldspar are clear and colorless. The aegirine-augite phenocrysts are black or dark green. The groundmass is very fine-grained with a greenish gray color caused by tiny needles of aegirine-augite. In addition, the rock contains numerous vesicles filled with light-colored minerals.

Table 1. Chemical Composition of Natrolite-tinguaite. Specimen FEAE No. 134. Toror Hills, Karamoja, Uganda. Analyst: Kai Hytönen, 1953.

Weight per cent		Molecular norm		Niggli values	
SiO_2	51.73	or	21.5	si	158
TiO_2	0.54	ab	20.5	ti	1.3
Al_2O_3	19.96	ne	38.9	al	35.8
Fe_2O_3	3.11	hl	0.5	fm	15.3
FeO	2.11	Salic	81.4	c	9.0
MnO	0.20			alk	39.9
MgO	0.50	ac	8.6	k	0.18
CaO	2.74	di	8.2	mg	0.15
Na_2O	11.06	wo	0.8	o	0.47
K_2O	3.70	il	0.7	c/fm	0.59
P_2O_5	0.15	ap	0.3	qz	—90
CO_2	0.00				
H_2O+	3.89	Femic	18.6		
H_2O-	0.24				
Cl	0.16				
	100.09	C.I.P.W. symbol: II, 7, 1, 4.			
—O	0.04				
Total	100.05				

Under the microscope, the specimen shows the following mineral composition (the constituents listed in order of relative abundance): phenocrysts; nepheline, aegirine-augite, sphene, alkali feldspar; groundmass; natrolite, potassium feldspar, aegirine-augite, low-albite, sodalite, biotite, zircon; amygdules; apophyllite, apatite, pectolite, fluorite, natrolite, calcite, diopside.

A planimetric analysis of the rock made on five thin sections yields the following composition: nepheline 14 %, aegirine-augite 9 ½ %, sphene ½ %, alkali feldspar ¼ %, groundmass 74 %, amygdules 1 ½ %.

The bulk chemical composition of the rock is given in Table 1. Chemically the rock is a typical tinguaite with a low silica content and low lime and magnesia. It is characterized by the high content of alkalis compared with that of aluminum. The very high water content is especially important.

Nepheline phenocrysts occur as well-developed hexagonal prisms with a length up to 3—5 mm, usually evenly distributed in the groundmass. Sometimes they form aggregates of more than half a dozen individuals with aegirine-augite, zircon and sphene as inclusions. In some places, natrolite is found as secondary alteration product. In addition, a few areas of the nepheline crystals are spotted with a zonally distributed fine dust. Most of the nepheline is unaltered and clear.

Table 2. Chemical Composition and Physical data for Nepheline from Natrolite-tinguaite. Specimen FEAE No. 134. Toror Hills, Karamoja, Uganda. Analyst: H. B. Wiik, 1953.

Weight per cent		Unit cell content		Physical data	
SiO_2	45.04	Si	8.48	Sp. gr.	2.607
TiO_2	0.02	Al	7.38	ε	1.532 — 1.534
Al_2O_3	33.25	Fe^{3+}	0.08	ω	1.534 — 1.539
Fe_2O_3	0.55	Fe^{2+}	0.04	Weighted average: $\omega =$	1.537
FeO	0.28	Mg	0.02	a_0	9.990 ± 0.004
MnO	0.01	Na	5.80	c_0	8.370 ± 0.004
MgO	0.07	K	1.27		
CaO	0.00	O	31.75		
Na_2O	15.89				
K_2O	5.30				
P_2O_5	0.00				
H_2O +	0.08				
H_2O —	0.00				
Total	100.49				

Table 2 summarizes the chemical composition and the physical data for the nepheline. The analysis yields the composition 18.0 mol. % Ks, and shows an excess of Si and a deficiency of alkalis. A deviation of this kind from the ideal nepheline composition seems to be very common among natural nephelines as is found, e. g., by Bannister (1931) and also by Smith and Sahama (1954).

The nepheline is extremely strongly zoned. The zoning is shown by concentric variations of the refractive indices, parallel to the basal pinacoid and prism faces. In some crystals, up to 150 zones may be counted. The thickness of the individual zones varies between 0.02 and 0.002 mm. The refractive index ω of 334 grains was measured by the single variation method thus allowing a weighted average to be calculated. The refractive index ε was determined in the usual way.

The material and the chemical analysis of the nepheline of this specimen have been used by Smith and Sahama (1954) for developing a method for determination of the composition of natural nephelines by X-ray powder pattern.

The phenocrysts of the aegirine-augite form euhedral prisms with the faces (110), (100) and (010). The terminal faces are visible in some crystals. Twinning on (100) is common. The phenocrysts are usually between 0.2 and 1 mm in length, but sometimes they range up to 4 mm.

To avoid contamination from the aegirine-augite of the groundmass, the separation of the aegirine-augite phenocrysts began with a rather coarse material crushed only with a jaw crusher. The phenocrysts, thus liberated from the groundmass, were then separated in the ordinary way and ground down to a grain size suitable for the final purification. Of the material to be used for the chemical analysis, the grain count made under the microscope indicated the total amount of impurities to be approximately 1.8 %. Table 3 presents the chemical composition and physical data for the aegirine-augite.

Table 3. Chemical Composition and Physical Data for Aegirine-augite Occurring as Phenocrysts in Natrolite-tinguaite. Specimen FEAE No. 134. Toror Hills, Karamoja, Uganda. Analyst: H. B. Wiik, 1953.

Weight per cent		Unit cell content		Physical data	
SiO_2	49.49	Si	7.75	Sp. gr.	3.529
TiO_2	0.76	Ti	0.09	α	1.724 — 1.727
Al_2O_3	1.88	Al	0.35	β	1.735 — 1.738
Fe_2O_3	10.00	Fe^{3+}	1.18	γ	1.753 — 1.758
FeO	13.16	Fe^{2+}	1.74	$2V_\gamma$	85° — 94°
MnO	0.83	Mn	0.11	$c \wedge \alpha$	16° — 21°
MgO	4.03	Mg	0.94	a_0	9.747 ± 0.004
CaO	16.07	Ca	2.70	b_0	8.917 ± 0.004
Na_2O	4.11	Na	1.25	c_0	5.313 ± 0.003
K_2O	0.09	K	0.02	$a_0 : b_0 : c_0$	= 1.093 : 1 : 0.596
P_2O_5	0.00	O	24.10	β	105°42′ ± 15′
H_2O+	0.05			Volume	444.5 Å³
H_2O-	0.00				
Total	100.47			Systematic extinctions agree with the space group C2/c	

Calculated composition:	
$NaFe\,(Si, Al)_2O_6$	29.3 mol. %
$(Na, K)\,(Al, Ti)\,(Si, Al)_2O_6$	2.7
$Ca\,(Si, Al)O_3$	33.5
$Mg\,(Si, Al)O_3$	11.7
$(Fe, Mn)\,(Si, Al)O_3$	22.8

The lengths of the unit cell axes were measured both from single crystal photographs and from the powder pattern presented in Table 4. The β^* angle of the reciprocal cell was measured from the *b*-axis zero layer Weissenberg photograph.

The pleochroism of the mineral is relatively strong and shows the scheme typical for aegirine-augite: α (grass green) $>$ β (olive green) $>$ γ (yellowish green). The variations in the optical properties are caused by zoning. The extinction angle $c \wedge \alpha$ varies considerably. Usually, the extinction angle is greater in the core of a phenocryst than in the margin. The optic axial angle varies from positive values in the core to negative values in the margin. This fact, according to Tröger (1952, p. 64), would indicate that the acmite component increases towards the margin.

Sphene occurs unaltered as well-developed and commonly twinned phenocrysts up to 1.5 mm in length.

The alkali feldspar is the least abundant constituent of the phenocrysts occurring in sparing amounts as tabular crystals that range up to 1 mm in diameter. Some of the crystals are completely altered to natrolite.

On the basis of the chemical composition of the bulk rock (Table 1), of the nepheline (Table 2) and aegirine-augite (Table 3) phenocrysts as well as of the result of the planimetric analysis, the chemical composition of the ground-

Table 4. Powder Pattern of Aegirine-augite Occurring as Phenocrysts in Natrolite-tinguaite. Specimen FEAE No. 134. Toror Hills, Karamoja, Uganda. Filtered Iron Radiation with Silicon Standard.

hkl	I	2 $\Theta_{obs.}$	$d_{obs.}$	$Q_{obs.}$	$Q_{calc.}$
110	30	17.215	6.468	0.0239	0.0239
200	10	23.795	4.695	0.0454	0.0454
020	20	25.07	4.460	0.0503	0.0503
220	40	34.835	3.234	0.0956	0.0957
$22\bar{1}$	80	37.73	2.9937	0.1116	0.1114
310	100	38.315	2.9497	0.1149	0.1148
$31\bar{1}$	20	39.025	2.8980	0.1191	0.1192
$13\bar{1}$	20	44.335	2.5655	0.1519	0.1515
[1] $11\bar{2}$ 221	10	44.885	2.5356	0.1555	0.1543 0.1565
[1] $32\bar{1}$ 230	30	45.23	2.5173	0.1578	0.1569 0.1586
330	20	53.415	2.1538	0.2156	0.2154
$33\bar{1}$	20	53.97	2.1333	0.2197	0.2197
$42\bar{1}$	10	54.66	2.1084	0.2250	0.2251
150	20	67.10	1.7515	0.3260	0.3258
$53\bar{1}$	20	73.115	1.6251	0.3786	0.3789
[1] 440	10	73.68	1.6144	0.3837	0.3829
350	10	77.325	1.5494	0.4166	0.4166

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[1] Line broad

mass of specimen FEAE No. 134 was calculated as follows: SiO_2 53.9, TiO_2 0.4, Al_2O_3 20.2, Fe_2O_3 2.8, FeO 1.1, MnO 0.2, MgO 0.2, CaO 0.8, Na_2O 11.2, K_2O 3.9, H_2O+ 5.1, Cl 0.2, total 100.0.

In thin section, natrolite and potassium feldspar form the predominating constituents of the groundmass. The relative abundance of these two minerals is difficult to estimate under the microscope. The calculated chemical composition of the groundmass indicates, however, that the amount of natrolite exceeds considerably that of potassium feldspar. The largest natrolite crystals measure up to 0.05 mm in diameter. Spherulitic occurrence is typical. Due to a very fine-grained pigment, the mineral is rarely quite clear. Usually, natrolite is evenly distributed over the whole groundmass. Prismatic cleavage is common, sign of elongation positive. $\alpha = 1.481$, $\beta = 1.486$,

Table 5. Powder Pattern of Natrolite from Natrolite-tinguaite. Specimen FEAE No. 134. Toror Hills, Karamoja, Uganda. Filtered Copper Radiation with Quartz Standard.

hkl	I	2 $\Theta_{obs.}$	$d_{obs.}$	$Q_{obs.}$	$Q_{calc.}$
220	85	13.59	6.510	0.0236	0.0236
111 130	100	15.095	5.864	0.0291	0.0290 0.0290
040	40	19.08	4.647	0.0463	0.0463
400	30	19.42	4.567	0.0479	0.0479
131	85	20.28	4.375	0.0522	0.0522
311 330	75	20.425	4.344	0.0530	0.0530 0.0530
240	45	21.43	4.143	0.0583	0.0583
420	30	21.665	4.099	0.0595	0.0595
112 151	60	27.97	3.187	0.0985	0.0984 0.0985
511 530	65	28.33	3.148	0.1009	0.1009 0.1009
022 060 202	35	28.81	3.096	0.1043	0.1041 0.1042 0.1045
222 260	60	30.425	2.9354	0.1161	0.1161 0.1162
312 351	100	31.255	2.8594	0.1223	0.1223 0.1224
531	90	31.465	2.8408	0.1239	0.1241
422 460	25	34.965	2.5640	0.1521	0.1520 0.1521
502 152	25	36.765	2.4425	0.1676	0.1674 0.1678
711 730	30	37.34	2.4061	0.1727	0.1728 0.1728
442	15	38.85	2.3160	0.1864	0.1867
003 081 262	20	41.145	2.1920	0.2081	0.2081 0.2083 0.2086
103 181 622 660	20	41.50	2.1741	0.2116	0.2111 0.2114 0.2119 0.2120

$\gamma = 1.494$, $2V_\gamma \sim 61°$. Owing to difficulties encountered in the separation of natrolite, a chemical analysis of the mineral was carried out (analyst: H. B. Wiik) on a material that contained ca. 5 % of aegirine-augite and ca. 13 % of sodalite. The result of the original analysis will not be reproduced here. The approximate composition of natrolite corrected for the admixtures mentioned was calculated as follows: SiO_2 47.4, Al_2O_3 25.6, Na_2O 11.5, K_2O 4.1, H_2O 11.4, total 100.0. Table 5 summarizes the powder pattern of natrolite. The indexing was based on the following unit cell dimensions: $a_0 = 18.270 \pm 0.005$, $b_0 = 18.589 \pm 0.005$, $c_0 = 6.577 \pm 0.003$ Å.

According to the chemical analysis (Table 1), the rock contains ca. 2 to 3 % sodalite. The mineral occurs in the groundmass as rounded crystals, ca. 0.05 mm in diameter, and is colorless, isotropic and contains numerous inclusions. $n = 1.486$. The powder pattern agrees with that of sodalite. $a_0 = 8.875 \pm 0.005$ Å.

The aegirine-augite of the groundmass occurs as small needles. The needles are clear and unaltered, 0.01 — 0.05 mm in length and 0.001 — 0.005 mm in thickness. The optical properties are different from those of the phenocryst aegirine-augite: $\alpha = 1.750 — 1.755$, $\gamma = 1.800 — 1.805$, $c \wedge \alpha = 0° — 4°$. Because of the small grain size, the optical study of the aegirine-augite needles is difficult and, therefore, the data given above are incomplete. According to the diagram presented by Tröger (1952, p. 64), the aegirine-augite of the groundmass contains between 60 and 70 mol. % of the acmite component. The composition obtained is rather close to the composition of the analyzed aegirine-augite from the groundmass of the natrolite-tinguaite, specimen H. 8 (Table 6).

The groundmass contains two alkali feldspars, *viz.* potassium feldspar and low-albite with an approximate abundance ratio of 9 : 1. The potassium feldspar occurs as laths of 0.05 — 0.5 mm length, in trachytoidal arrangement. $\alpha = 1.519$, $\beta = 1.524$, $\gamma = 1.526$. Common Carlsbad twinning. Monoclinic. Composition $Or_{88}Ab_{12}$ ($d_{(20\bar{1})} = 4.206$). The low-albite shows $\alpha = 1.530$ and $\gamma = 1.540$. Composition $Ab_{95}An_5$. $B_{CuK\alpha} = 2\,\Theta_{(1\bar{1}1)} — 2\,\Theta_{(20\bar{1})} = 1.024°$ (Smith and Gay, 1958, p. 754).

Biotite is the least abundant constituent of the groundmass. The mineral is brown, pleochroic, $\gamma \sim \beta = 1.655$.

Of the amygdule minerals listed on p. 16, apophyllite, pectolite and apatite are quantitatively by far the most important. The crystals are often rather large, ranging up to almost the same size as the phenocrysts of the rock. The amount of calcite, fluorite and diopside is small.

Apophyllite fills the interstices of the other constituents of the amygdules. In a few cases, the mineral occurs as one single poikilitic crystal with a diameter of 2 to 3 mm. In thin section, the grains are colorless and somewhat turbid. The interference colors are abnormal with undulating extinction.

Optically uniaxial, positive. The refractive indices are as follows: $\omega = 1.534 - 1.535$, $\varepsilon = 1.537$. Perfect cleavage parallel to (001). The powder pattern agrees with that of apophyllite.

Pectolite forms clear, colorless grains, in average between 0.2 and 0.5 mm in length. Exceptionally, grains of a length of 2 mm and a thickness of 0.5 mm have been recorded. Two perfect cleavages. The angle between these cleavage planes was measured with an optical goniometer as 84.5° ± 0.1°. The optical properties of the pectolite are as follows: $\alpha = 1.604$, $\beta = 1.612$, $\gamma = 1.638$, $2V_\gamma = 54° \pm 2°$. Optical orientation: $a \wedge \beta = 18° \pm 2°$, $b \wedge \gamma = 0°$, $c \wedge \alpha = 14° \pm 2°$. The powder pattern agrees with that of pectolite.

Apatite occurs in the amygdules as clear and colorless hexagonal prisms measuring up to 0.2 — 0.5 mm in length, with refractive indices: $\varepsilon = 1.635$, $\omega = 1.639$.

A little calcite is present in the amygdules. For identification of the mineral, the refractive index was measured ($\omega \sim 1.660$) and the powder pattern obtained.

Some amygdules contain colorless fluorite. The following data were recorded: sp. gr. ca. 3.17, n = 1.435.

In thin section, only a few grains of diopside have been observed and these have an average diameter of ca. 0.2 mm. The optical data are as follows: $\alpha = 1.677$, $\beta = 1.684$, $\gamma = 1.704$, $2V_\gamma = 58°—62°$, $c \wedge \gamma = 45°$. According to the diagram presented by Hess (1949, p. 641), these optical properties correspond to a composition of approximately 85 mol. % diopside and 15 mol. % hedenbergite. The powder pattern agrees with that of diopside.

Specimen FEAE No. 137. *Sp. gr. 2.40.* — Boulder on the Moroto-Kotido road, close to Nakazilet. The rock differs considerably from the type specimen of the natrolite-tinguaites, *viz.* specimen FEAE No. 134. More than half of the bulk of the rock is natrolite. In addition, the amount of calcite is remarkable. Almost a half of the nepheline ($Ne_{85}Ks_{15}$) phenocrysts are replaced by natrolite and calcite (Fig. 1, Plate II) whereas the phenocrysts of aegirine-augite, sphene and alkali feldspar are unaltered. The groundmass is made up mainly of natrolite and alkali feldspar. The rock contains a number of amygdules filled with calcite, natrolite and apatite. In hand specimen, the rock is brownish gray, almost equigranular, with only a few small phenocrysts visible.

Specimen H. 2. *Sp. gr. 2.35.* — Dyke on the S wall of West Valley II, altitude 1630 m. Like specimen FEAE No. 137, this rock is also characterized by a very high amount of natrolite. A planimetric analysis of the rock made on five thin sections gave the following amounts of unaltered Na-sanidine and aegirine-augite phenocrysts: Na-sanidine 11 ½ %, aegirine-augite 4 ½ %. The Carlsbad twinned Na-sanidine laths are well-developed, on the average

2—5 mm and sometimes up to 10 mm in length. The mineral proved to be monoclinic with a composition $Or_{63}Ab_{37}$ ($d_{(20\bar{1})} = 4.161$). $\alpha = 1.520$, $\beta = 1.524$, $\gamma = 1.526$, $2V_{\alpha} = 44°—49°$, optic axial plane perpendicular to (010). All nepheline phenocrysts and a part of the feldspar phenocrysts have been replaced by natrolite. The short, stumpy euhedral phenocrysts of aegirine-augite measure 0.1—0.4 mm in length. The constituents of the groundmass are natrolite, alkali feldspar and aegirine-augite. The amygdules are numerous, filled solely with spherulitic natrolite that is yellowish-brown in color due to pigment. This specimen has the lowest specific gravity recorded of all the tinguaite specimens collected by the author at the Toror Hills. In hand specimen, the groundmass of the rock is gray-green. The feldspar phenocrysts are visible at a distance.

Specimen H. 8. *Sp. gr. 2.47.* — N end of Loloman. This rock is characterized by the absence of phenocrysts of aegirine-augite and unaltered nepheline. The former nepheline phenocrysts are completely replaced by natrolite. A few Carlsbad twinned tabular potassium feldspar phenocrysts are present, partly replaced by natrolite and containing pigment. The potassium feldspar shows $\alpha = 1.517$, $\gamma = 1.525$, $2V_{\alpha} = 62° — 66°$. The mineral is structurally intermediate between microcline and sanidine with a triclinicity $2\,\Theta_{(1\bar{3}0)} - 2\,\Theta_{(130)} = 0.55°$ (CuKα). The texture of the groundmass is typically tinguaitic, grayish green in hand specimen, with spherulitic natrolite as the main constituent. In addition to natrolite, the groundmass contains potassium feldspar and aegirine-augite. The aegirine-augite occurs as a felt of randomly oriented very thin needles either lying between or partly penetrating the other constituents. The needles are quite clear and unaltered, 0.01—0.05 mm in length, almost colorless, only slightly greenish and pleochroic. For closer study, the mineral was separated out of the rock. Table 6 presents the chemical composition and some physical data for the mineral. The chemical analysis was made on very pure material. Due to the small size of the aegirine-augite needles, the optical properties given are incomplete. The rock also contains a few skeletons of an unidentified almost opaque material.

Specimen H. 14. *Sp. gr. 2.49.* — Boulder on the E slope of Nakazilet, altitude 1300 m. The rock shows large phenocrysts of clear nepheline ($Ne_{84}Ks_{16}$) that, on the margins, are partly altered and replaced by natrolite and analcite. The dark green, needle-like phenocrysts of aegirine-augite are heavily zoned and commonly twinned. The rock contains also apatite prisms, up to 1 mm long, and, in addition, some phenocrysts of alkali feldspar and sphene and some few amygdules filled mainly with natrolite, pectolite, apatite and calcite. Spherulitic natrolite is the dominant constituent of the grayish-green groundmass. Compared with natrolite, the amount of analcite is very small and, therefore, the rock has been included in the group of the

Table 6. Chemical Composition and Some Physical Data for the Aegirine-augite of the Groundmass from Natrolite-tinguaite. Specimen H. 8. Toror Hills, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1957.

Weight per cent		Cation per cent		Physical data	
SiO_2	51.82	Si	50.46	Sp. gr.	3.552
TiO_2	1.82	Ti	1.33	α	1.755
Al_2O_3	0.50	Al	0.57	γ	1.805
Fe_2O_3	23.38	Fe^{3+}	17.13	α closely parallel to the c-axis	
FeO	5.71	Fe^{2+}	4.65		
MnO	0.57	Mn	0.47		
MgO	0.64	Mg	0.93		
CaO	4.49	Ca	4.69		
Na_2O	10.29	Na	19.42		
K_2O	0.28	K	0.35		
H_2O+	0.22	Total	100.00		
H_2O-	0.02				
Total	99.74				

Calculated composition:

$NaFeSi_2O_6$	68.5 mol. %
$(Na, K)(Al, Ti) Si_2O_6$	10.0
$CaSiO_3$	9.4
$MgSiO_3$	1.9
$(Fe, Mn) SiO_3$	10.2

natrolite-tinguaites proper. In addition to the zeolites, needles of aegirine-augite and lath-shaped crystals of alkali feldspar are found in the groundmass.

Specimen H. 17. *Sp. gr. 2.41.* — Dyke on the Moroto-Kotido road, ca. 2 km SE from Oropoi. Petrographically the rock resembles the specimen FEAE No. 137. The rock is porphyritic and contains a number of amygdules (natrolite, calcite, some apatite). Nepheline is completely altered to saccharoidal natrolite and calcite. Some of the nepheline phenocrysts have been entirely replaced by one calcite single crystal. Usually calcite forms large poikilitic crystals in both the nepheline and groundmass, ranging up to 2—3 mm in diameter. The aegirine-augite is unaltered, strongly zoned and mantled by rims of fine-grained aegirine-augite needles. A few unaltered alkali feldspar phenocrysts are seen; $2V_\alpha = 40°$; optically monoclinic, optic axial plane perpendicular to (010). Of the constituents of the groundmass, natrolite and alkali feldspar are most important; the natrolite is brownish and very cloudy, due to pigment.

NATROLITE-PECTOLITE-TINGUAITE

Specimen H. 3. *Sp. gr. 2.43.* — S wall of West Valley II, altitude 1535 m. The rock is of an especial mineralogical and petrological interest. In hand specimen, it differs considerably from all the other tinguaites collected by the author on the Toror Hills. The rock is compact, medium-grained, non-por-

Table 7. Natrolite-pectolite-tinguaite. Specimen H. 3. Toror Hills, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1957.

Weight per cent		Molecular norm		Niggli values	
SiO_2	52.50	or	26.4	si	167
TiO_2	0.25	ab	22.5	ti	0.6
Al_2O_3	16.64	ne	25.7	al	31.3
Fe_2O_3	3.95	hl	0.1	fm	14.9
FeO	1.34	Salic	74.7	c	15.8
MnO	0.31			alk	38.0
MgO	0.21	ac	11.1	k	0.24
CaO	4.62	Na_2SiO_3	1.8	mg	0.07
Na_2O	9.39	wo	4.5	o	0.64
K_2O	4.43	di	5.7	c/fm	1.06
P_2O_5	0.12	ap	0.2	qz	—65
CO_2	0.65	cc	1.7		
H_2O+	5.02	il	0.3		
H_2O-	0.27				
Cl	0.02	Femic	25.3		
Total	99.72	C.I.P.W. symbol: II, 6, 1, 4.			

phyritic and slightly ophitic in appearance. It shows aegirine-augite, natrolite, potassium feldspar and pectolite as main constituents. In addition, some nepheline, sphene, calcite, apatite and apophyllite are found.

Table 7 presents the chemical composition of the rock. Chemically it is a typical tinguaite, almost identical with the natrolite-tinguaite of specimen FEAE No. 134 (Table 1). The contents of lime and water are slightly higher, those of alumina and alkalis lower.

The main part of aegirine-augite occurs as branching aggregates or skeletons (Fig. 2, Plate II). Often the aggregates are made of a bunch of radially oriented needle-like grains. Homogeneous euhedral crystals are rare. When present, they are extremely strongly zoned showing a colorless core with an extinction angle up to 20°—25° larger than that in the margin. Accordingly, the core of the crystals is augite or diopside and grades over into green aegirine-augite in the margin. Because of zoning, the refractive indices vary considerably, ranging up to $\alpha = 1.755$ and $\gamma = 1.800$. According to the diagram presented by Tröger (1952, p. 64), these refractive indices would correspond to ca. 65 mol. % of the acmite component. From the chemical analysis of the rock, the average composition of the aegirine-augite was calculated as follows: $NaFeSi_2O_6$ 66 mol. %, $CaMgSi_2O_6$ 7 mol. %, $CaFeSi_2O_6$ 27 mol. %.

Pectolite occurs as slender colorless prisms with random orientation scattered over the whole matrix. The prisms measure 0.05—0.2 mm in length. The norms, presented in Table 7, give an idea of the approximate amount of pectolite present in the rock. The norm shows an excess of 4.5 % $CaSiO_3$ over the diopside molecule and an excess of 1.8 % Na_2SiO_3 over the acmite molecule. In addition to aegirine-augite and pectolite, no other constituents with femic lime and femic soda are present in the rock. Accordingly, the excess of $CaSiO_3$ and of Na_2SiO_3 are due to the pectolite, and the calculation shows that it forms ca. 8 % of the rock.

Natrolite is quantitatively the most important constituent of the rock. It is clear and fresh and shows saccharoidal texture in the matrix and spherulitic texture when replacing the nepheline crystals. In addition, natrolite forms a number of large more or less euhedral crystals with uniform extinction and a diameter up to 0.5—2.0 mm. They enclose, poikilitically, aegirine-augite and pectolite. No potassium feldspar grains are enclosed in the natrolite crystals.

The potassium feldspar occurs in the rock as lath-shaped crystals with a diameter of ca. 0.05—0.2 mm. The mineral is structurally intermediate between microline and sanidine with the triclinicity $2\,\Theta_{(1\bar{3}0)} - 2\,\Theta_{(130)} = 0.52^\circ$ and $2\,\Theta_{(1\bar{3}1)} - 2\,\Theta_{(131)} = 0.50^\circ$ (CuKα). $\alpha = 1.518$, $\gamma = 1.524$.

Only a few large nepheline crystals are found, and these are almost completely replaced by natrolite. Nepheline is absent in the matrix.

Calcite forms poikilitic grains with a diameter up to 1 mm. Apophyllite is present in very sparing amounts.

NATROLITE-ANALCITE-TINGUAITES

Specimen H. 11. *Sp. gr. 2.49.* — Boulder from the top of Lorano. The rock contains abundant white lath-shaped phenocrysts of potassium feldspar (2—6 mm in length) and less abundant phenocrysts of aegirine-augite and of water-clear nepheline in a dark gray-green groundmass. The powder pattern of the feldspar phenocrysts shows diffuse (130) and (131) lines. Accordingly, the mineral seems to be structurally intermediate between microcline and sanidine with varying degrees of triclinicity. $\alpha \sim 1.524$, $\gamma \sim 1.530$, $2V_\alpha =$ 56°—61°. The mineral is zoned and partly altered. The cores of the largest phenocrysts are sometimes completely replaced by natrolite. The margins are commonly unaltered and contain pigment. Polysynthetic twin lamellae are often seen in two directions at right angles to each other. As is stated by Laves (1950) and also by Goldsmith and Laves (1954 b), this characteristic twinning is indicative of inversion from an original monoclinic crystal. In addition, twinning on the Carlsbad law is commonly seen. In a few cases, microcline-perthite occurs in the cores of the phenocrysts. The amount of albite present is too low to be detected in the powder pattern. The nepheline ($Ne_{84}Ks_{16}$) phenocrysts are euhedral and zoned. The aegirine-augite is strongly zoned and commonly twinned. In addition, sphene, biotite and brown hornblende are present in small amounts.

In addition to natrolite and analcite, the groundmass contains the following constituents: nepheline, laths of alkali feldspar and aegirine-augite. The zeolites do not occur together, but separately as in the groundmass of specimen H. 7 (p. 28). The rock contains a number of amygdules filled mainly with nepheline, pectolite, apophyllite and apatite.

Specimen H. 12. *Sp. gr. 2.44.* — E foot of Lorano. In hand specimen, the rock shows abundant white phenocrysts of alkali feldspar and a few, very large amygdules in a grayish green groundmass. The alkali feldspar is structurally intermediate between microcline and sanidine with the following triclinicity: $2\,\Theta_{(1\bar{3}0)} - 2\,\Theta_{(130)} = 0.61°$ and $2\,\Theta_{(1\bar{3}1)} - 2\,\Theta_{(131)} = 0.56°$ (CuKα). $\alpha = 1.518$, $\gamma = 1.525$, $2V_\alpha = 67°—74°$. The largest feldspar crystals measure up to 5 mm in length. The mineral is cloudy due to pigment, has been partly replaced by calcite and contains needles of aegirine-augite as inclusions.

Nepheline ($Ne_{83}Ks_{17}$) is almost completely replaced by a medium-grained mass of natrolite, saccharoidal in texture. The boundaries of the nepheline phenocrysts are sharp showing that nepheline has been present as large euhedra with a length up to 4 mm (Fig. 3, Plate II). In addition to zoned aegirine-augite, euhedral phenocrysts of sphene are found. The groundmass contains saccharoidal or spherulitic natrolite, laths of alkali feldspar, aegirine-augite needles, euhedral microlites of nepheline completely replaced by natrolite, and some analcite and calcite. Natrolite is the main constituent of the amygdules which, in addition, contain apophyllite, analcite, and calcite.

Specimen H. 13. *Sp. gr. 2.39.* — Boulder on the NW slope of Moruangnamorj. In hand specimen, the rock is characterized by tabular white alkali feldspar phenocrysts in contrast to the very dark gray-green groundmass. The alkali feldspar is twinned on the Carlsbad law. Its powder pattern shows diffuse (130) and (131) lines with a triclinicity not accurately measurable. $\alpha = 1.518$, $\gamma = 1.524$, $2V_\alpha = 57°—62°$. It is cloudy due to fine pigment. The nepheline ($Ne_{85}Ks_{15}$) phenocrysts are almost completely replaced by natrolite or, more rarely, by analcite. The aegirine-augite phenocrysts show a strong zoning. The aegirine-augite needles of the groundmass are relatively large (average length 0.1—0.2 mm) as compared with the aegirine-augite of the groundmass in the other rocks of the Toror Hills (Fig. 1, Plate I). Natrolite is the dominant constituent of the groundmass. Compared with natrolite, the amounts of analcite and of the alkali feldspar of the groundmass are low. A few amygdules are filled mainly with natrolite, analcite, calcite and pectolite.

Specimen H. 15. *Sp. gr. 2.54.* — Boulder on the access road at the base of Nakazilet. The rock is characterized by the abundance of natrolite and analcite present in the groundmass, by its large euhedral nepheline phenocrysts and by its numerous white amygdules. In hand specimen, the rock is green, only slightly grayish. The nepheline ($Ne_{85}Ks_{15}$) is zoned, clear and shows abundant fissures filled with natrolite. The euhedral aegirine-augite phenocrysts are zoned and twinned. In addition, sphene, apatite and brown hornblende occur as phenocrysts. As in specimens H. 7 and H. 11, natrolite and analcite do not occur together in the same areas in the groundmass but separately. In the areas rich in analcite, unaltered nepheline microlites are seen. The amygdules contain natrolite, pectolite, apatite, opaque oxide, apophyllite and analcite.

ANALCITE-NATROLITE-TINGUAITES

Specimen H. 1. *Sp. gr. 2.44.* — North Amairoi. The rock is compact with a dense, gray-green groundmass and with relatively few phenocrysts. No amygdules are seen. Alkali feldspar is quantitatively the most important constituent of the phenocrysts. It is very cloudy due to pigment and shows commonly alteration to analcite. The mineral is structurally intermediate between microcline and sanidine with the triclinicity $2\,\Theta_{(1\bar{3}0)} - 2\,\Theta_{(130)} = 0.50^\circ$ (CuKα). $\alpha = 1.514$, $\gamma = 1.520$, $2V_\alpha = 62^\circ$—72°. Nepheline ($Ne_{84}Ks_{16}$) is not abundant. The phenocrysts are euhedral, clear and unaltered. A few strongly zoned phenocrysts of aegirine-augite are seen. In addition, some apatite is present.

The groundmass is medium-grained, equigranular and contains analcite, alkali feldspar and aegirine-augite as main constituents and natrolite, euhedral nepheline and calcite as minor constituents (Fig. 2, Plate I). Most of the nepheline has been replaced by analcite. The alkali feldspar shows lath-shaped crystals in trachytoidal arrangement. The aegirine-augite occurs not as needles but as dark green stumpy crystals.

Specimen H. 7. *Sp. gr. 2.36.* — Boulder on the N end of the Central Ridge, altitude 1320 m. The rock is very rich in analcite and natrolite. The specific gravity of the rock is remarkably low thus indicating the large amount of zeolites present. The rock contains relatively abundant phenocrysts of zoned nepheline and Na-sanidine of composition $Or_{52}Ab_{48}$ ($d_{(20\bar{1})} = 4.142$). $\alpha = 1.522$, $\gamma = 1.529$, $2V_\alpha = 51^\circ$, optic axial plane perpendicular to (010). Most of the nepheline ($Ne_{84}Ks_{16}$) phenocrysts are replaced by natrolite. The few aegirine-augite phenocrysts are strongly zoned, dark green in color. Apatite and sphene are common. Analcite and natrolite are quantitatively the most important constituents of the groundmass and are almost equal in abundance. Usually, the zeolites occur on separate areas with relatively sharp boundaries. Areas rich in natrolite are characterized by the moderate birefringence of natrolite whereas the analcite-bearing areas are almost isotropic between crossed nicols. A few amygdules are filled mainly with natrolite. In hand specimen, the groundmass of the rock is grayish brown, not greenish, like the main types of tinguaites of the Toror Hills.

Specimen H. 16. *Sp. gr. 2.48.* — Boulder on the E slope of Nyanga, altitude 1280 m. The rock is characterized by fissure-like vesicles in subparallel orientation filled with natrolite. The fillings are white and visible at a distance in the field and in hand specimen. The abundant euhedral phenocrysts of nepheline ($Ne_{85}Ks_{15}$) are zoned and clear, 0.2 mm in average length. Sometimes alteration to natrolite is seen. The aegirine-augite phenocrysts rarely show euhedral forms. Usually, the margin is surrounded by fine-grained detritus of aegirine-augite. The mode of occurrence gives an impression of a partial mechanical breakdown of the original aegirine-augite phenocryst. In hand specimen, the aegirine-augite phenocrysts show a subparallel orientation perpendicular to the fissure-like vesicles. A few unaltered alkali feldspar phenocrysts are found. The mineral is K-monalbite with the composition $Or_{44}Ab_{56}$ ($d_{(20\bar{1})} = 4.123$). $\alpha = 1.525$, $\gamma = 1.533$, $2V_{\alpha} = 56°—58°$. In addition, the rock contains a very few large phenocrysts of alkali feldspar, almost completely replaced by natrolite and analcite. The amount of the unaltered portion is very small as compared with the size of the original phenocrysts. The powder pattern shows that this unaltered portion of the feldspar is triclinic with diffuse (130) and (131) lines. The groundmass is very fine-grained, almost aphanitic. It contains analcite and, less abundantly, natrolite. Some calcite and apophyllite is seen in the amygdules.

Specimen FEAE No. 133. *Sp. gr. 2.54.* — Boulder on the Moroto-Kotido road, close to Nakazilet. In hand specimen, the rock shows abundant water-clear, beautifully zoned phenocrysts of nepheline ($Ne_{84}Ks_{16}$) (Fig. 4, Plate II), and needles of dark green, almost black aegirine-augite. The margins of the nepheline phenocrysts are often altered to cancrinite with high birefringence. In addition to nepheline and aegirine-augite, apatite and sphene occur as phenocrysts. No feldspar phenocrysts have been observed. The groundmass is green and contains the following constituents: analcite, saccharoidal natrolite, lath-shaped crystals of alkali feldspar in trachytoidal arrangement, needles of aegirine-augite and some cancrinite. According to the powder pattern taken of the groundmass, analcite, and natrolite are almost equal in quantity. The amygdules contain natrolite, apatite and pectolite.

Specimen FEAE No. 135. *Sp. gr. 2.54.* — Boulder on the Moroto-Kotido road, close to Nakazilet. In hand specimen, the rock shows a large number of small nepheline ($Ne_{83}Ks_{17}$) and aegirine-augite phenocrysts with an average length of 0.5—2 mm and a few large (up to 10 mm) lath-shaped crystals of optically monoclinic alkali feldspar with $2V_{\alpha} = 40°$ and optic axial plane perpendicular to (010). The phenocrysts are fresh and clear. In a few cases, nepheline has been in part replaced by natrolite. Microscopically, sphene is commonly seen as phenocrysts. In the medium-grained groundmass, analcite predominates over natrolite. In addition to the alkali feldspar and the zeolites, pectolite is present as a minor constituent in the groundmass. It occurs as slender colorless prisms, 0.1—0.5 mm in length.

Specimen FEAE No. 136. *Sp. gr. 2.46.* — Boulder on the Moroto-Kotido road, close to Nakazilet. The rock contains large phenocrysts of nepheline and alkali feldspar, approximately equal in quantity. Nepheline is clear and fresh with a composition $Ne_{87}Ks_{13}$. Alteration to natrolite is seen only in the margins. The feldspar phenocrysts are zoned. The cores are sometimes completely replaced by spherulitic natrolite or by poikilitic calcite or by these two. The margins are commonly unaltered, containing pigment. Cross-hatched twinning is not seen. The mineral is structurally intermediate between microcline and sanidine with the triclinicity $2\,\Theta_{(1\bar{3}0)} - 2\,\Theta_{(130)} = 0.15^\circ$ (CuKα), the lines (131) not being measurable. $2V_\alpha = 48^\circ - 57^\circ$. The aegirine-augite phenocrysts are strongly zoned, showing sometimes an almost colorless core. In addition, phenocrysts of brown hornblende and of sphene are seen. The medium-grained groundmass, grayish green in hand specimen, shows a typical tinguaitic texture with laths of alkali feldspar and needles of aegirine-augite in random orientation. Analcite is quantitatively the most important constituent of the groundmass, natrolite being seen only in a few spots. However, spherulitic natrolite is the main constituent in the numerous amygdules.

Specimen FEAE No. 138 *(with accessory sodalite). Sp. gr. 2.54.* — Boulder on the Moroto-Kotido road, close to Nakazilet. The rock contains abundant large phenocrysts of nepheline. The mineral is clear and fresh, but sometimes it shows alteration to natrolite and analcite starting from cracks. The well-developed slender aegirine-augite phenocrysts are zoned. Sphene forms euhedral grains, up to 1.5 mm long. Only a few alkali feldspar phenocrysts are seen. The groundmass, dark green in hand specimen, shows a typical tinguaitic texture under the microscope. It contains the following constituents listed in the order of approximate abundance: analcite, alkali feldspar, aegirine-augite, sodalite, nepheline, natrolite. The amygdules are filled mainly with apatite, natrolite and pectolite.

ANALCITE-TINGUAITES

Specimen H. 5. *Sp. gr. 2.62.* — S wall of West Valley II, altitude 1435 m. The rock is compact, dark gray, not greenish. Under the microscope, it shows abundant euhedral phenocrysts of nepheline ($Ne_{84}Ks_{16}$), alkali feldspar and aegirine-augite. Both nepheline and alkali feldspar are almost unaltered. The feldspar represents a Na-sanidine with the composition $Or_{70}Ab_{30}$ ($d_{(20\bar{1})} = 4.176$). $\alpha = 1.522$, $\gamma = 1.529$, $2V_\alpha = 43^\circ$, optic axial plane perpendicular to (010). Some sphene and apatite are present. The groundmass is very fine-grained and contains the following constituents: alkali feldspar, analcite, needles of aegirine-augite, calcite and sodalite. A few amygdules and cracks are filled mainly with natrolite and calcite.

Specimen H. 6 B. *Sp. gr. 2.60.* — W foot of Toror Peak, altitude ca. 1860 m. This rock is characterized by a low content of zeolite. In hand specimen, the rock differs considerably from the common type of tinguaites of the Toror Hills. It is compact, dark gray, only slightly greenish. The number of phenocrysts is small. The few euhedral nepheline ($Ne_{84}Ks_{16}$) phenocrysts are slightly zoned, fresh and contain aegirine-augite and zircon as inclusions. Also the aegirine-augite phenocrysts are strongly zoned showing an almost colorless core. Euhedral sphene is present. The groundmass is very fine-grained with the following constituents: monoclinic potassium feldspar, nepheline, aegirine-augite and analcite. This is one of the few specimens of tinguaites from the Toror Hills where unaltered nepheline occurs in the groundmass. No amygdules are seen.

Specimen H. 9. *Sp. gr. 2.49.* — Main eastern part of Katungura. This specimen represents the only Toror Hills tinguaite studied in which low-albite occurs as phenocrysts. The white feldspar phenocrysts are visible at a distance and form a contrast to the dark gray-green groundmass. A microscopic study shows that the feldspar phenocrysts are extremely dusty due to pigment. The phenocrysts mostly occur as aggregates of several individuals. The largest phenocrysts measure up to 2 mm in length. Twinning on the albite law is common. The powder pattern shows that the main part of the feldspar phenocrysts (ca. 75 %) is low-albite (Smith and Gay, 1958) with the composition $Ab_{99}An_1$. $\alpha = 1.528$, $\gamma = 1.538$. $B = 1.04°$; $\Gamma = —0.70°$ (CuKα). In addition, the strongest lines of potassium feldspar are seen in the powder pattern. Because of the large amount of low-albite present, it is not possible to conclude whether the potassium feldspar is monoclinic or triclinic. The small optic axial angle ($2V_\alpha = 50°—54°$) would indicate that the potassium feldspar is a sanidine. Examination of the feldspar powder under the microscope with immersion liquid shows that the two phases, *viz.* low-albite and potassium feldspar, occur as separate grains in the rock. Compared with the amount of alkali feldspar phenocrysts, the amount of the nepheline ($Ne_{79}Ks_{21}$) phenocrysts is very small. No phenocrysts of aegirine-augite are seen in the rock. The medium-grained groundmass shows a typical tinguaitic texture, and contains abundant euhedral nepheline, very thin needles of aegirine-augite and small laths of alkali feldspar in a matrix of analcite. Analcite is completely isotropic with $n = 1.488$ and $a_0 = 13.714 \pm 0.003$ Å. A few amygdules are filled mainly with analcite, pectolite and natrolite. Compared with analcite, the amount of natrolite present in the rock is low. In addition, natrolite does not occur in the groundmass but only in the amygdules.

Specimen H. 10. *Sp. gr. 2.47.* — E side of Lakuda = North Ridge. In hand specimen, the rock shows abundant large phenocrysts of nepheline, alkali feldspar and aegirine-augite in a dense gray-green groundmass. In addition, numerous white amygdules are seen.

Table 8. Chemical Composition of Alkali Feldspar (Ba-bearing K-monalbite) from Analcite-tinguaite. Specimen H. 10. Toror Hills, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1958.

Weight per cent		Unit cell content		Physical data	
SiO_2	64.14	Si	11.76	Sp. gr.	2.618
TiO_2	0.02	Al	4.33	α	1.526
Al_2O_3	20.05	Fe^{3+}	0.03	β	1.530
Fe_2O_3	0.23	Fe^{2+}	0.01	γ	1.532
FeO	0.06	Ca	0.03	$2V_\alpha$	56°—63°
MnO	0.00	Ba	0.17	$\gamma // b$	
MgO	0.00	Sr	0.02	$a \wedge \alpha = 5°$	
CaO	0.16	Na	1.97	$c \wedge \beta = 21°$	
BaO	2.36	K	1.64		
SrO	0.21	O	32.10	Monoclinic:	
Na_2O	5.53			$a_0 = 8.362 \pm 0.003$	
K_2O	7.02			$b_0 = 12.977 \pm 0.005$	
H_2O+	0.10			$c_0 = 7.157 \pm 0.003$	
H_2O-	0.00			$\beta = 116° 4' \pm 5'$	
Total	99.88			$a_0 : b_0 : c_0 = 0.644 : 1 : 0.552$	
				Volume 697.7 Å^3	

Calculated composition:

Or	42.9 mol %
Ab	51.3
An	0.8
Celsian	5.0

The alkali feldspar of this rock has been investigated in detail. The phenocrysts are relatively thin and tabular, ranging up to 15 mm in length. They are generally quite fresh and clear and contain sometimes nepheline, aegirine-augite and sphene as inclusions. In addition, a few areas of the feldspar crystals are spotted with fine dust.

To avoid contamination from the alkali feldspar of the groundmass, the separation of the feldspar of the phenocrysts began with a rather coarse material crushed only with the jaw crusher. The phenocrysts, thus liberated from the groundmass, were then collected by hand picking and were ground down to a grain size suitable for final separation. To remove the nepheline, the powder was treated with dilute hydrochloric acid. A microscopic test and the powder pattern showed the material used for chemical analysis to be very pure. The result of the analysis and the physical data are presented in Table 8 and the indexed powder pattern is summarized in Table 9.

The powder pattern does not show any lines of albite nor of microcline. The strong (002) line of albite at $2\,\Theta = 28.06°$ (CuKα) is absent (Donnay and Donnay, 1952; Smith, 1956), the lines (130) and (131) are sharp, well-defined and do not show any sign of triclinicity (Goldsmith and Laves, 1954 a and b; MacKenzie, 1954). In addition, a single crystal b-axis oscillation photograph shows that the mineral consists of a single homogeneous phase (MacKenzie

Table 9. Powder Pattern of Alkali Feldspar (Ba-bearing K-monalbite) from Analcite-tinguaite. Specimen H. 10. Toror Hills, Karamoja, Uganda. Filtered Copper Radiation with Silicon Standard.

hkl	I	$2\,\Theta_{obs.}$	$d_{obs.}$	$Q_{obs.}$	$Q_{calc.}$
110 020 001	4	13.65	6.482	0.02380	0.02366 0.02375 0.02419
$11\bar{1}$	2	15.265	5.799	0.02974	0.02966
$20\bar{1}$	4	21.51	4.128	0.05868	0.05869
111	6	22.845	3.889	0.06612	0.06604
130	18	23.72	3.748	0.07119	0.07116
$13\bar{1}$	3	24.72	3.598	0.07725	0.07716
$11\bar{2}$	9	25.78	3.453	0.08387	0.08404
$20\bar{2}$ 040	50	27.465	3.245	0.09497	0.09488 0.09501
002	100	27.72	3.215	0.09675	0.09675
131	11	30.10	2.9664	0.11364	0.11354
041	12	30.845	2.8963	0.11921	0.11920
201 $13\bar{2}$	8	32.42	2.7592	0.13135	0.13145 0.13154
$24\bar{1}$	6	35.145	2.5512	0.15364	0.15370
112	2	35.51	2.5258	0.15675	0.15680
310 240 150	2	36.57	2.4550	0.16592	0.16542 0.16589 0.16617
$15\bar{1}$	2	37.265	2.4108	0.17206	0.17217
$11\bar{3}$	3	38.845	2.3163	0.18638	0.18680
151	2	41.205	2.1890	0.20869	0.20855
330 $30\bar{3}$ 060	13	41.69	2.1646	0.21342	0.21292 0.21348 0.21377
003	1	42.125	2.1432	0.21771	0.21771
241 $15\bar{2}$	2	42.96	2.1035	0.22600	0.22646 0.22655
$16\bar{1}$ 061	2	44.11	2.0512	0.23768	0.23749 0.23796
222	2	46.68	1.9442	0.26456	0.26415
400 260	2	48.49	1.8758	0.28420	0.28352 0.28465
$35\bar{2}$ 113	3	49.52	1.8391	0.29566	0.29555 0.29594
062 204	9	50.91	1.7921	0.31137	0.31053 0.31240
$15\bar{3}$ $44\bar{2}$ $44\bar{1}$	2	52.465	1.7426	0.32931	0.32931 0.32977 0.32996

and Smith, 1955 and 1956; Smith and MacKenzie, 1955). According to the diagram presented by Bowen and Tuttle (1950), the spacing of the $(20\bar{1})$ line corresponds to a molecular composition of $Or_{47}Ab_{53}$. After heating at 900° C for 24 hours the spacing of the $(20\bar{1})$ line corresponds to ca. $Or_{50}Ab_{50}$. These values are close to the ratio of the potassium component to the sodium component calculated on the basis of the chemical analysis, *viz.* $Or_{45}Ab_{55}$. The unit cell dimensions for a synthetic alkali feldspar of a composition $Or_{45}Ab_{55}$, estimated from diagrams presented by Donnay and Donnay (1952), are as follows: $a_0 = 8.375$, $b_0 = 12.980$, $c_0 = 7.163$, $\beta = 116° 10'$, volume = 700 Å^3. These values are very close to the unit cell dimensions presented in Table 8.

According to the nomenclature proposed by Laves (1952) and revised by Schneider and Laves (1957), the feldspar of specimen H. 10 should be called either K-monalbite or Na-common orthoclase. The former name refers to the monoclinic sanidine-monalbite series disordered with respect to Al-Si and stable at high temperatures. The latter name refers to the unstable intermediate states between the disordered sanidine-monalbite series and ordered low-temperature microcline-albite series. As no triclinicity could be detected, the name Na-common orthoclase is rejected and the name K-monalbite or barium-bearing K-monalbite is adopted here.

The refractive indices correspond to the orthoclase-Na-orthoclase series (Spencer, 1937, Tuttle, 1952). The optical orientation is that of a typical »common orthoclase» whereas the 2V value is intermediate between the values for the orthoclase-low-albite and the low-sanidine-high-albite series (Tuttle, 1952). The amount of barium present in the feldspar seems not to be able to affect notably the optical properties. According to the diagrams presented by Strandmark (1904) and Winchell and Winchell (1951) for the adularia-celsian series, a content of 5 mol. % of $BaAl_2Si_2O_8$ may increase the refractive indices of adularia by an amount of ca. 0.003—0.004. The effect of the barium content is seen in the specific gravity of the mineral. The specific gravity measured (2.618) is clearly higher than that estimated for a common alkali feldspar with a theoretical composition of $Or_{45}Ab_{55}$ (2.597; Tröger, 1952, p. 96).

The aegirine-augite and nepheline ($Ne_{85}Ks_{15}$) phenocrysts are euhedral and zoned. Nepheline is almost unaltered and contains abundant aegirine-augite, sphene and zircon as inclusions. Some of the aegirine-augite phenocrysts range up to 6 mm in length. Sphene is common as phenocrysts. The groundmass is fine-grained and contains analcite, alkali feldspar and aegirine-augite. The amygdules are filled mainly with natrolite and calcite.

SUMMARY OF THE PETROGRAPHY OF THE TOROR HILLS TINGUAITES

A very brief review of the main petrographic characteristics of the Toror Hills tinguaites will now be given.

As has been mentioned above (p. 14), the rocks described in this paper are of hypabyssal origin, mostly with more or less pronounced tinguaitic texture.

With a few exceptions, the rocks are porphyritic, containing nepheline, alkali feldspar and aegirine-augite as the main phenocrysts. In only one specimen (H. 9) was low-albite found as a phenocryst mineral.

The nepheline phenocrysts are more or less zoned, but it has not been possible to determine the range of variation of the composition in the different zones. The X-ray powder diffraction method gives only the overall average and that varies between 13 and 21 mol. % of the potassium component. According to the compilation published by Miyashiro (1951), this range of variation is found in plutonic rocks in general and is also that most common in volcanic rocks. It is to be noted in this connection that the content of the potassium component of nepheline in these rocks is less than that required by the ideal nepheline formula, *viz.* $KNa_3Al_4Si_4O_{16}$ (Hahn and Buerger, 1955). The two chemical analyses of the Toror Hills tinguaites given in this paper (Tables 1 and 7) as well as those presented by Du Bois (1956) show considerable predominance of sodium over potassium. Accordingly, potassium-rich nephelines like those found in the lavas of the Western Rift Valley are not to be expected in the Toror Hills. The nepheline phenocrysts in the analcite-tinguaites are unaltered. In most of the natrolite-bearing tinguaites nepheline is more or less altered and the degree of alteration seems to be clearly dependent on the amount of natrolite present in the rock. Thus, in the most natrolite-rich tinguaites, the nepheline phenocrysts have been almost completely replaced by natrolite. A few exceptions to this rule are found. For instance, the nepheline of the natrolite-tinguaite of specimen FEAE No. 134 is almost unaltered despite the relatively high content of natrolite present in the groundmass of the rock. In the groundmass itself the absence of nepheline is characteristic of the natrolite-bearing tinguaites. In analcite-bearing tinguaites (specimens H. 6B, H. 9, H. 11, H. 15), unaltered nepheline microlites and analcite lie in close association with each other. In two of the natrolite-analcite-tinguaites (specimens H. 11 and H. 15), the zeolites do not occur together in the same areas of the groundmass but separately. In these two rocks, nepheline is found only in analcite-rich areas and not in the natrolite areas.

In eight of the tinguaite specimens from the Toror Hills which have been studied (H. 8, H. 3, H. 11, H. 12, H. 13. H.1, H. 16, FEAE No. 136), the alkali feldspar phenocrysts are triclinic with varying degrees of triclinicity. The high triclinicity corresponding to a complete Si-Al ordered microcline and found, e. g., for pegmatitic microclines, has not been detected among the potassium feldspars of the Toror Hills. In one case only (specimen H. 11), has polysynthetic twinning and perthitic exsolution texture been recorded. In accordance with the varying degrees of triclinicity found, the negative optic axial angle measurements made with the universal stage yielded values lower than that to be expected for a completely ordered microcline. The $2V_\alpha$

values obtained range from 48° to 72°. Accordingly, the triclinic potassium feldspars of the Toror Hills are to be classed among the »intermediate microclines» as defined by Goldsmith and Laves (1954 b). As to the composition of these intermediate microclines studied, the X-ray powder diffraction method used for determination of the composition of the monoclinic alkali feldspars is not strictly applicable to the triclinic potassium feldspars. The method developed by Bowen and Tuttle (1950) and based on the d-value of the line $(20\bar{1})$ would yield for the triclinic potassium feldspars of the Toror Hills compositions not far from the pure $KAlSi_3O_8$. Because the refractive indices determined for the triclinic potassium feldspars of Toror Hills tinguaites are regularly lower than those of the monoclinic alkali feldspars of these rocks, the conclusion may be justified that the triclinic potassium feldspars are more rich in the potassium component than are the monoclinic alkali feldspars of the same rocks. All triclinic potassium feldspar phenocrysts are turbid and contain natrolite and analcite as alteration products.

In contrast to the intermediate microcline, the phenocrysts of monoclinic alkali feldspar are usually fresh and clear. In the alkali feldspar phenocrysts whose composition has been determined by X-ray powder diffraction (specimens H. 2, H. 7, H. 16, H. 5, H. 10), the composition ranges from $Or_{44}Ab_{56}$ to $Or_{70}Ab_{30}$. The optic axial plane was always found to be perpendicular to (010). In Fig. 5 the values for the negative optic axial angle for these alkali feldspars, measured with the universal stage, are plotted against feldspar composition reproduced in simplified form from Tuttle and Bowen (1958, p. 104). As is seen from the figure, all the points for the Toror Hills feldspars lie between the curves representing the high-albite-low-sanidine and the low-albite-orthoclase series, respectively. Accordingly, with respect to the Si-Al order, the monoclinic alkali feldspar phenocrysts of the Toror Hills tinguaites are intermediate between the low-sanidines and the orthoclases proper. According to the terminology proposed by Laves (1952) and Schneider and Laves (1957), these feldspars may be called Na-sanidines to K-monalbites. Of the monoclinic alkali feldspars of the groundmass, only that of specimen FEAE No. 134 has been studied by X-ray powder diffraction. The result, $Or_{88}Ab_{12}$, shows it to be more rich in the potassium component than any of the monoclinic alkali feldspars occurring as phenocrysts.

The two albites of the Toror Hills, one of which (specimen H. 9) is present as phenocrysts and the other (specimen FEAE No. 134) in the groundmass were studied by X-ray powder diffraction method. In Fig. 6 the *B*-values ($= 2\Theta_{(1\bar{1}1)} - 2_{(20\bar{1})}$) for these two albites are plotted against feldspar composition reproduced in simplified form from Smith and Gay (1958, p. 754) and indicate that both albites represent the low-form of the mineral.

Aegirine-augite is the predominating femic constituent in all the Toror Hills tinguaites studied occurring both as phenocrysts and in the ground-

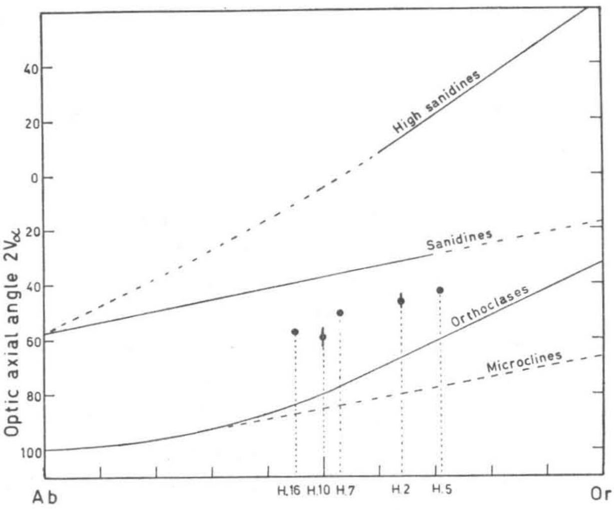


Fig. 5. Optic axial angles of alkali feldspars of the Toror Hills tinguaites plotted against the feldspar composition.

mass. Two chemical analyses of the mineral have been presented, *viz.* of specimen FEAE No. 134 (Table 3) representing the composition of the aegirine-augite phenocrysts and of specimen H. 8 (Table 6) representing the mineral in the groundmass. Comparison of these two analyses with each other reveals a very marked difference in the ratio of the acmite component to the diopside-hedenbergite component. In the aegirine-augite of the phenocrysts (Table 3) the diopside-hedenbergite component predominates over the acmite component whereas in the aegirine-augite of the groundmass (Table 6) the ratio between the two components is the opposite. The difference in chemical composition between the two aegirine-augites is also clearly reflected in the optical properties. Of the optical properties, the extinction angle is most readily observable in thin section and may be regarded as indicative of the content of the acmite component in the mineral. The aegirine-augites occurring as phenocrysts in the Toror Hills tinguaites are always more or less zoned with a light-colored or sometimes even colorless core that gradually changes its color to intense green in the margin. The extinction angle $c \wedge \alpha$ decreases from the core towards the margin. The aegirine-augites occurring in the groundmass of the tinguaites show a virtually straight extinction with a negative sign of elongation. Optical observations which have not been reported in detail in the description of the specimens indicate that the differences in chemical composition between the two analyzed aegirine-augites hold for the Toror Hills tinguaites generally. In other words, the aegirine-augite of the groundmass is regularly more rich in the acmite component than the aegirine-augite that occurs as phenocrysts.

Among the constituents of the groundmass of the Toror Hills tinguaites, natrolite and analcite predominate quantitatively. The relative amount of these two zeolites is very characteristic of the particular specimen and, therefore, it has been used as a basis for a petrographic classification of the rocks studied. A quantitative ratio between natrolite and analcite is very difficult to determine in the fine-grained groundmass. A microscopic study of the rocks in thin section reveals, however, that this ratio varies more or less continuously from specimen to specimen. For that reason, no sharp boundaries between the different tinguaite groups proposed in this paper can be drawn. The mode of occurrence of natrolite and analcite makes their separation for chemical analysis very difficult. Accordingly, no complete analyses of these minerals have been presented. Natrolite has been separated from specimen FEAE No. 134 and chemically analyzed. The material used for the analysis was, however, very impure and, therefore, only a rough estimate of the composition could be made (p. 20). This estimate together with optical and X-ray data presented prove the correctness of the identification of the mineral as natrolite. The identification of analcite was made by powder pattern supported by optical observations.

Very characteristic of the mode of occurrence of natrolite and analcite in the Toror Hills tinguaites is the fact that, where both these minerals are present in large amounts as is the case in some of the natrolite-analcite- and analcite-natrolite-tinguaites, the two zeolites occupy separate areas of the groundmass with sharp boundaries between natrolite-bearing and analcite-bearing areas. The areas are often irregular in form, but sometimes are lenticular showing a poorly developed parallel orientation throughout the thin section. Natrolite occurs in saccharoidal crystallization and often shows a tendency to form incomplete spherulites. In one specimen only (H. 3) it shows poikilitic development forming euhedral crystals the grain size of which exceeds that of the average groundmass. These poikilitic natrolite crystals resemble the porphyroblasts of metamorphic rocks. Both natrolite and, to a lesser extent, analcite represent alteration products of nepheline and feldspar.

In the tinguaites studied, sodalite, pectolite and apophyllite are common constituents. These minerals occur only in small amounts and are never more than accessories. The only exception to this rule is represented by the specimen H. 3 where pectolite is one of the essential constituents of the rock and forms ca. 8 % of the bulk.

The abundance of amygdules is a further characteristic feature of the tinguaites of the Toror Hills. The main constituents of the amygdules are pectolite, apophyllite, natrolite, calcite and apatite.

With the exception of specimen H. 3, the two zeolites, natrolite and analcite, are the only essential constituents of the Toror Hills tinguaites that contain water in appreciable amounts. Accordingly, most of the H_2O+ shown by the chemical analyses of these rocks is contained in the zeolites. The analysis of Table 1 would correspond to ca. 41 weight % of natrolite in that rock. To illustrate further the amount of zeolites in the tinguaites of the Toror Hills, a sample of ca. 200 g of specimen H. 8, ground down to a grain size of 40—200 mesh, was treated with dilute hydrochloric acid and weighed before and after the treatment. In addition to natrolite, the rock contains virtually only aegirine-augite and potassium feldspar, both insoluble in the acid. In the acid treatment, ca. 56 % of the sample was dissolved giving an approximate idea of the amount of natrolite present in that specimen. The analysis of the »average phonolite», made of a composite sample of 10 rocks from the Toror Hills and published by Du Bois (1956) suggests a high average content of zeolites in these rocks. The figure for H_2O+ in that »average phonolite» analysis is very close to that of the analysis of Table 7 (specimen H. 3) of this paper.

The high amounts of zeolites in the Toror Hills tinguaites are also reflected in the low specific gravities of these rocks. The specific gravity of the 22 tinguaite specimens studied ranges from 2.35 to 2.62, averaging 2.48. This average value is intermediate between the specific gravity of natrolite and analcite (ca. 2.25) on the one hand and that of nepheline and alkali feldspars (ca. 2.56—2.62) on the other. The average value for the specific gravity of the Toror Hills tinguaites is remarkably low for holocrystalline compact rocks and is clearly lower than the average specific gravity of phonolites and trachytes, *viz.* 2.55 (Johannsen, 1938; vol. III, p. 77 and vol. IV, p. 126). No value for an average specific gravity of tinguaites could be found in the literature. The lowest values found for the specific gravities of the Toror Hills tinguaites lie very close to the specific gravity of blairmorite (d = 2.388), a porphyritic volcanic rock with 71 % analcite from near Crowsnest Lake, Alberta, Canada (MacKenzie, 1914). The specific gravity of blairmorite is one of the lowest found for feldspathoidal rocks. Chemically, blairmorite is very close to the tinguaites of the Toror Hills. In addition to analcite, it contains the following modal constituents: sanidine 4 %, nepheline 5 %, aegirine-augite 14 % and accessories (melanite, sphene, hematite, calcite) 4.5 %.

CONCLUSIONS

The primary mode of origin of the two zeolites, natrolite and analcite, in a number of different kind of alkaline volcanic rocks has been stated and accepted by several petrologists, e. g., by Washington (1898), Pelikan (1906), Hibsch (1914), MacKenzie (1915), Scott (1916), Marshall (1928) and Bartrum (1936). Also the general abundance of natrolite and analcite in the groundmass of the hypabyssal tinguaites of the Toror Hills suggests that these two minerals are not solely products of secondary alteration of other pre-existing minerals but represent essentially primary or at least deuteric constituents. It is true that both natrolite and analcite occur clearly as alteration products of the nepheline and feldspar phenocrysts and, in a couple of specimens, of the nepheline found in the groundmass. This fact does not, however, contradict the conclusion that the main part of the zeolites is primary. In most of the tinguaites of the Toror Hills, the groundmass is entirely fresh and does not show any signs of secondary alteration. The crystallization of primary zeolites is essentially connected with the presence of water in the solidifying rock melt. The presence of large quantities of water in the eruptions of the other neighboring volcanoes, especially on Elgon, is recorded by Davies (1952, p. 51).

As the tinguaites of the Toror Hills are mainly composed of analcite, natrolite, nepheline, and alkali feldspar, the system alkali-Al_2O_3-SiO_2-H_2O is especially important for the study of the phase relationships in these rocks. Aegirine-augite is virtually the only phase of the tinguaites that is not in this system. In what follows, the rocks of the Toror Hills will be compared with the results of some synthetic experiments made under hydrothermal conditions by several investigators.

Investigations of the phase assemblages in the system Na_2O-Al_2O_3-SiO_2-H_2O at pressures of 1 000 kg/cm^2 (Sand, Roy and Osborn, 1954; Levin, McMurdie and Hall, 1956, p. 270; Barrer and White, 1952; Roy and Tuttle, 1956,; Yoder, 1950) show that natrolite occurs as a stable phase at temperatures below about 290° C. Under the same pressure conditions analcite is stable up to ca. 525° C. At temperatures below 290° C both analcite and natrolite are stable. The minimum temperature of formation of nepheline is 460° C. Below this temperature nepheline decomposes to a hydrated nepheline phase. The experiments made at various pressures on dehydrated natrolite by Koizumi and Kiriyama (1957) suggest that the stability field of natrolite is probably below 200° C. The formation of natrolite appears to be favored by the presence of nuclei or by a low pH. In alkaline media at 200° C, both natrolite and analcite are present.

In the system K_2O-Al_2O_3-SiO_2-H_2O initial work by DeVries and Roy (1953, unpublished; abstracted by Roy and Tuttle, 1956) show the absence of any hydrated phases excepting muscovite. Some preliminary experiments in the silica-poor part of the important system $NaAlSiO_4$-$KAlSiO_4$-SiO_2-H_2O have been made by MacKenzie (1956). These experiments are especially interesting as they give information of the phase relations in the bulk composition range which approximates the composition of natural rocks of the syenite and nepheline-syenite clans. The preliminary experiments of MacKenzie show that analcite and a potassium-rich feldspar coexist as stable phases at 500° C in the presence of water vapor at a pressure of 8000 psi. The experiment was made with a glass of composition $Qtz_{35}Ne_{45}Kp_{20}$. Between 775° and 625° C the compositions of the coexisting alkali feldspar and nepheline were $Or_{46}Ab_{54}$ and $Ne_{93}Ks_7$, respectively.

The study of the phase relationships in the systems mentioned above does not yet cover completely all the compositions, temperatures and pressures required for a direct application of the information obtained to the history of crystallization of the Toror Hills tinguaites. In addition, of course, the Toror Hills rocks contain femic material (mainly aegirine-augite) the presence of which may affect the conclusions drawn from the results of the synthetic work. For the interpretation of the microscopic and other observations, however, the results of the synthetic studies provide some valuable indications.

It is evident that the phenocrysts of aegirine-augite, nepheline and monoclinic alkali feldspar are not in equilibrium with the groundmass but have crystallized at temperatures above the stability range of the two zeolites, natrolite and analcite. Aegirine-augite changes continuously in composition from a diopside-hedenbergite-rich (phenocrysts) to an acmite-rich (groundmass) clinopyroxene. Nepheline phenocrysts show varying degrees of alteration into natrolite. Microscopic observations seem to indicate that the paragenesis nepheline + analcite may be stable if crystallization occurs in a suitable temperature range but, on the other hand, that the paragenesis nepheline + natrolite is entirely unstable in the rocks studied. The monoclinic alkali feldspar phenocrysts, intermediate between low-sanidines and orthoclases, are always clear and unaltered. As soon as the phenocrysts become partly altered, they are no longer monoclinic but show varying degrees of triclinicity. These intermediate microcline phenocrysts are more or less turbid and apparently crystallized originally as a monoclinic alkali feldspar that, on decreasing temperature, became exsolved producing a more potassium-rich triclinic phase and a sodium feldspar. This latter phase has been altered to natrolite.

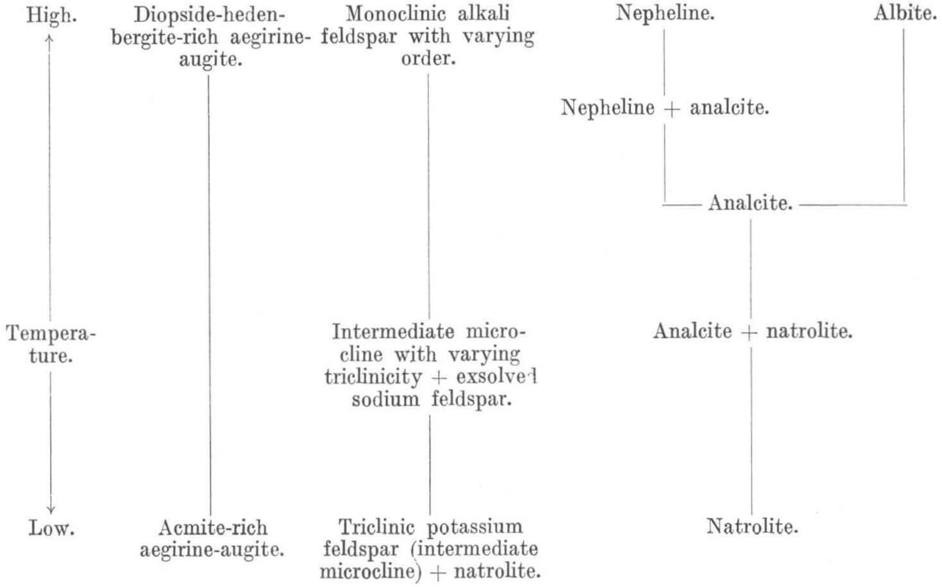
The natrolite-pectolite-tinguaite, specimen H. 3, is an especially good example of the primary mode of occurrence of natrolite. In addition, the rock gives some indications as to the potassium feldspar-natrolite paragenesis. Microscopically, the minerals of the rock show the following order of crystallization: 1) nepheline; 2) aegirine-augite; 3) pectolite; 4) large crystals of poikilitic natrolite; 5) triclinic potassium feldspar and spherulitic natrolite and natrolitization of nepheline. The mode of occurrence of the poikilitic natrolite suggests that this mineral is formed during the main crystallization of the rock and, therefore, must be regarded as a primary constituent. The following points indicate that the poikilitic natrolite crystals do not represent replacement products of nepheline: 1) the alteration product of nepheline is always fine-grained spherulitic or saccharoidal natrolite; 2) the original nepheline crystals are always mantled with well-developed rims composed of aegirine-augite and pectolite. No traces of such rims are seen around the poikilitic natrolite or in any connection with it; 3) the unaltered and altered nepheline crystals contain only a few inclusions of aegirine-augite and no inclusions of pectolite in contrast to the poikilitic natrolite that contains inclusions of these two minerals in abundance. Further, no indications have been found suggesting that the poikilitic natrolite would be an alteration product of some other pre-existing mineral, e. g., of alkali feldspar or analcite.

The mode of occurrence of triclinic potassium feldspar suggests that the mineral is later than the poikilitic natrolite and contemporaneous with the spherulitic natrolite. Apparently, the formation of this feldspar occurred at a very low temperature, probably below the maximum temperature of stability for natrolite. The coexistence of a triclinic potassium feldspar and natrolite suggests that these two minerals are the stable phases in the system $NaAlSiO_4$-$KAlSiO_4$-SiO_2-H_2O at low temperatures in the same manner as analcite and potassium-rich feldspar coexist at 500° C in MacKenzie's (1956) experiment. However, this comparison is somewhat uncertain because the sodium-potassium-silica ratio of the glass used by MacKenzie differs slightly from that of the natrolite-pectolite-tinguaite of the Toror Hills.

In specimen H. 3, the equilibrium is almost completely attained during the later stages of crystallization. The mode of occurrence of aegirine-augite suggests that the mineral originally existed as well-developed euhedral phenocrysts that have been shattered and recrystallized during the later stages of the crystallization of the rock. The alterations were presumably caused by the water- and alkali-rich liquid residuum that yielded the large amount of natrolite. The probability of such changes is further supported by the facts that the aegirine-augite occurs virtually in one generation and that the acmite content of the mineral is rather high as compared with the aegirine-augite phenocrysts of the other tinguaites of the Toror Hills.

In the hypabyssal crystallization of such relatively small dykes and plugs as those formed by the tinguaites of the Toror Hills, the rate of cooling and, accordingly, the temperature of the final crystallization of the groundmass can, of necessity vary within wide limits. For that reason, it is well understandable that in individual rocks phase assemblages will be found that do not correspond to equilibrium conditions but represent intermediate stages in the cooling and consolidation of the rock.

All the information about the history of crystallization of the Toror Hills tinguaites is presented in the following scheme:



MT. MOROTO

The volcanic rocks of Mt. Moroto are represented in this paper by 23 specimens. All specimens were collected from loose boulders in the Nediket valley, ca. 5 km S of Moroto station, at a locality where the Moroto-Kitale road crosses a small river flowing from Mt. Moroto. The locality of the Nediket valley is indicated on the unpublished sketch map of Mt. Moroto given by Messrs. Tanganyika Concessions Limited[1]. The FEAE numbers of the rocks collected are the following:

Phonolites.	
Phonolites with accessory sodalite	112, 122, 127.
Mafic analcite-phonolite	116.
Trachytes.	
Andesitic trachytes	119, 130, 114.
Pyroxene-trachyte	124, 110.
Trachybasalts	123, 128, 126, 129, 125, 118, 117, 131, 120.

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[1] Map No. Mcb/134 C, available at the Geological Survey Department, Entebbe, Uganda.

Analcite-bearing trachydiabase	121.
Phillipsite-clinopyroxene-rock	113.
Olivine-melanephelinites	109, 115, 111.

Short descriptions of these specimens will follow.

PHONOLITES

Specimen FEAE No. 112 *(with accessory sodalite). Sp. gr. 2.67.* — This dark gray rock is porphyritic with anorthoclase-cryptoperthite, fayalite (97 mol. % Fa), aegirine-augite and brown hornblende as phenocrysts amounting to ca. 15 % of the bulk. The fine-grained, compact groundmass with trachytic texture contains potassium feldspar, albite, nepheline, aegirine-augite, sodalite, analcite, brown hornblende, apatite and fayalite. Table 10 presents the chemical composition of the rock. Chemically, the rock is a typical phonolite.

Anorthoclase-cryptoperthite is quantitatively the most important constituent of the phenocrysts. It is euhedral, clear, slightly zoned and shows albite, pericline and Carlsbad twinning. Refractive indices: $\alpha = 1.532$, $\gamma = 1.540$. The powder pattern of the mineral shows two unmixed phases, *viz.* $Or_{14}Ab_{86}$ ($d_{(20\bar{1})} = 4.061$) and $Or_{76}Ab_{24}$ ($d_{(20\bar{1})} = 4.188$), with the sodium-rich phase predominating over the potassium-rich phase. The feldspar was homogenized by heating at 900° C for 24 hours. Powder patterns of the heated sample showed a single phase with the composition $Or_{20}Ab_{80}$ ($d_{(20\bar{1})} = 4.074$).

The aegirine-augite phenocrysts are commonly altered to a granular opaque product the unaltered part of the mineral occurring as a narrow rim that surrounds the opaque core of the crystal. Optical properties: $\alpha = 1.720$—1.730, $\gamma = 1.755$—1.770, $2V_\gamma = 69°$, $c \wedge \gamma = 53°$, green, slightly pleochroic.

The brown hornblende occurs as strongly pleochroic phenocrysts. The crystals are mantled by an opaque alteration rim. $\alpha = 1.682$, $\beta = 1.700$, $\gamma = 1.710$, $2V_\alpha = 75°$, $b \parallel \beta$, $c \wedge \gamma = 12°$.

The main constituents of the groundmass are potassium feldspar and albite. Potassium feldspar shows $\alpha = 1.523$, $\gamma = 1.529$. Composition $Or_{83}Ab_{17}$ ($d_{(20\bar{1})} = 4.201$). From the powder pattern it is not possible to conclude whether the mineral is monoclinic or triclinic. The refractive indices of albite are $\alpha = 1.529$, $\gamma = 1.538$. Composition virtually $Ab_{100}An_0$. Nepheline ($Ne_{85}Ks_{15}$) occurs as very small euhedral or subhedral grains in the interstitial matrix. Sodalite is scattered over the whole area of the groundmass. It forms isotropic, colorless, commonly dodecahedral crystals, 0.05—0.1 mm in diameter. Judging from the chemical analysis of Table 10, the sodalite content of the rock amounts to 4 %. Analcite equals sodalite in quantity. The femic constituents of the groundmass show shattered forms.

Table 10. Chemical Composition of Phonolite with Accessory Sodalite. Specimen FEAE No. 112. Mt. Moroto, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1957.

Weight per cent		Molecular norm		Niggli values	
SiO_2	56.47	or	28.5	si	184
TiO_2	0.22	ab	36.4	ti	0.4
Al_2O_3	19.93	ne	24,3	al	38.3
Fe_2O_3	2.32	hl	0.6	fm	16.8
FeO	3.43			c	5.8
MnO	0.18	Salic	89.8	alk	39.1
MgO	0.27			k	0.26
CaO	1.65	ac	0.8	mg	0.08
Na_2O	9.13	di	5.6	o	0.34
K_2O	4.97	ol	1.2	c/fm	0.34
P_2O_5	0.13	mt	2.0	qz	—70
CO_2	0.00	il	0.3		
Cl	0.20	ap	0.3		
H_2O+	0.80	Femic	10.2		
$H_2O—$	0.10				
	99.80				
— O	0.05	C.I.P.W. symbol: I, 6, 1, 4.			
Total	99.75				

Specimens FEAE Nos. 122 and 127 *(with accessory sodalite). Sp. gr. 2.64—2.65.* — The two specimens are almost identical with the specimen FEAE No. 112. However, no fayalite has been found. The powder pattern of the anorthoclase-cryptoperthite ($2V_\alpha = 59°—62°$) phenocrysts shows two phases, *viz.* $Or_{14}Ab_{86}$ ($d_{(201)} = 4.060$) and $Or_{70}Ab_{30}$ ($d_{(20\bar{1})} = 4.174$), with the sodium-rich phase predominating over the potassium-rich phase.

Specimen FEAE No. 116 *(mafic analcite-phonolite). Sp. gr. 2.79.* — According to planimetric analysis, this porphyritic rock contains the following amounts of phenocrysts: clinopyroxene 22 %, nepheline 4 %, magnetite 7 %. The clinopyroxene is zoned with the faces (100), (010), (001), (111) and (110). The crystals show hour-glass texture and the slightly violet color typical of the titanian variety. Optical properties: $\alpha = 1.702—1.704$, $\beta = 1.709—1.711$, $\gamma = 1.728—1.730$, $2V_\gamma = 53°—56°$, $c \wedge \gamma = 40°$. The subhedral nepheline ($Ne_{88}Ks_{12}$) forms aggregates with an average diameter of 1—2 mm. The mineral is partly altered to analcite. In thin section, a few small phenocrysts of plagioclase are seen. Measurements of the extinction angle in the zone perpendicular to (010) yields the approximate composition $Ab_{65}An_{35}$. The plagioclase is partly replaced by natrolite and calcite. The fine-grained groundmass shows grains of magnetite and needles of slightly violet clinopyroxene in random orientation. The matrix is composed of analcite, natrolite and potassium feldspar. Analcite predominates quantitatively.

TRACHYTES

Specimens FEAE Nos. 119 and 130 *(andesitic trachytes). Sp. gr. 2.75 and 2.72, respectively.* — The rocks are fine-grained and compact. A few phenocrysts of andesine ($\alpha = 1.550$, composition ca. $Ab_{55}An_{45}$) and of zoned clinopyroxene are seen. The groundmass shows trachytic texture with alkali feldspar as the predominating constituent and, in addition, contains oligoclase ($\alpha = 1.540$), clinopyroxene, magnetite, calcite, analcite, apatite and a birefringent product, presumably chlorite. The mafic constituents amount to ca. 15—20 % of the bulk rock.

Specimen FEAE No. 114 *(andesitic trachyte). Sp. gr. 2.60.* — Plagioclase is quantitatively the most important phenocryst mineral. It is strongly zoned and commonly shows alteration to calcite and analcite. Albite, pericline and Carlsbad twinning. $\alpha = 1.540—1.547$, $\gamma = 1.548—1.555$. Composition $Ab_{76-62}An_{24-38}$. According to the powder pattern, the mineral is structurally intermediate between the high and the low temperature series of plagioclases (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.48°$; $B = 0.92°$ ($CuK\alpha$). In addition, the following minerals are present as phenocrysts: almost colorless clinopyroxene ($\alpha = 1.695—1.702$, $\beta = 1.705—1.712$, $\gamma = 1.725—1.730$, $2V_\gamma = 58°—63°$, $c \wedge \gamma = 40°$), basaltic hornblende, apatite and some few crystals of nepheline.

Hornblende amounts to 2—3 % of the bulk rock. In hand specimen the mineral is black. Under the microscope, it shows strong pleochroism with the brownish color typical of basaltic hornblende. The phenocrysts are up to 0.5—1.5 mm in length and often form aggregates of half a dozen individuals. The crystals are mantled with a black rim. The core is usually fresh without inclusions and alteration products. No zoning has been detected. The mineral was separated out of the rock for chemical analysis and X-ray study. Table 11 presents the chemical composition and physical data. The lengths of the crystallographic axes of the unit cell were measured both from single crystal rotation photographs and from powder pattern presented in Table 12. The β^* angle of the reciprocal cell was measured from the zero layer b-axis Weissenberg photograph. The chemical composition of this amphibole is rather close to that of basaltic hornblende (Strunz, 1957). The structural formula may be written as $(Na,K)_{2.0}Ca_{3.7}Mg_{4.4}Fe^{2+}_{3.1}Ti_{0.9}Fe^{3+}_{1.6}\,[(OH)_{3.2}\,Al_{3.7}Si_{12.2}O_{44.9}]$.

The groundmass is fine-grained with trachytic texture. It contains alkali feldspar, slightly greenish clinopyroxene, magnetite, apatite and a greenish interstitial product, probably chlorite.

Specimen FEAE No. 124 *(pyroxene-trachyte). Sp. gr. 2.90.* — According to planimetric analysis, the rock contains the following amounts of phenocrysts: clinopyroxene 18 %, »iddingsite» 4 %, magnetite 4 %. The clinopyroxene phenocrysts are strongly zoned, twinned and show hour-

Table 11. Chemical Composition and Physical Data for Basaltic Hornblende from Andesitic trachyte. Specimen FEAE No. 114. Mt. Moroto, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1958.

Weight per cent		Unit cell content		Physical data	
SiO_2	40.09	Si	12.22	Sp. gr.	3.307
TiO_2	4.09	Ti	0.94	α	1.683
Al_2O_3	10.37	Al	3.73	β	1.702
Fe_2O_3	6.83	Fe^{3+}	1.57	γ	1.712
FeO	11.78	Fe^{2+}	3.00	$2V_\alpha$	69°—72°
MnO	0.35	Mn	0.09	$b \parallel \beta$	
MgO	9.77	Mg	4.44	$c \wedge \gamma$	3°—6°
CaO	11.18	Ca	3.65	$a_0 = 9.898 \pm 0.004$	
Na_2O	2.78	Na	1.64	$b_0 = 18.101 \pm 0.007$	
K_2O	0.92	K	0.36	$c_0 = 5.324 \pm 0.003$	
H_2O+	1.56	OH	3.17	$a_0 : b_0 : c_0 = 0.547 : 1 : 0.294$	
H_2O-	0.10	O	44.86	$\beta = 106° 2' \pm 15'$	
Total	99.82			Volume 916.8 Å^3	
				Systematic extinctions agree with the space group C2/m	

glass texture and the violet color typical of the titanian variety. $\alpha = 1.701$, $\beta = 1.711$, $\gamma = 1.732$, $2V_\gamma = 53°—56°$, $c \wedge \gamma = 45°$.

»Iddingsite» occurs as pseudomorphs after euhedral olivine with uniform extinction. The dark brownish grains measure up to 2 mm in diameter. No unaltered olivine is seen in thin section. The mineral was separated out of the rock for chemical analysis and for X-ray study. Table 13 presents the chemical composition and the physical data. The mineral is orthorhombic. The unit cell was measured from rotation and Weissenberg photographs about all three axes. The reflections are large indicating an incomplete order in the crystal. No reflections of olivine are present. The X-ray photographs indicate the existence of a sub-cell and of a super-cell. Only the strongest lines of goethite are seen in the powder pattern. The single crystal X-ray data given in Table 13 are supported by investigations of »iddingsite» recently made by Ming-Shan Sun (1957) and Wilshire (1958) using X-ray powder diffraction. According to Ming-Shan Sun, »study of iddingsite samples from New Mexico and Colorado - - - shows that goethite is the only crystalline phase and that the other substances shown by chemical analysis to occur in iddingsite are largely amorphous». In the »iddingsite» from Carmel Bay, California, Wilshire shows the presence of the following crystalline phases: goethite, mixed layer smectite(= montmorillonite)-chlorite, quartz and feldspar.

The analysis of »iddingsite» FEAE No. 124 was made of a very pure material. Chemically, the mineral shows marked similarities with the »iddingsites» from several localities studied by Ross and Shannon (1925) and also

Table 12. Powder Pattern of Basaltic Hornblende from Andesitic Trachyte. Specimen FEAE No. 114. Mt. Moroto, Karamoja, Uganda. Filtered Copper Radiation with Quartz Standard.

hkl	I	$2\,\Theta_{\mathrm{obs.}}$	$d_{\mathrm{obs.}}$	$Q_{\mathrm{obs.}}$	$Q_{\mathrm{calc.}}$
110	30	10.52	8.402	0.0142	0.0141
200	10	18.64	4.756	0.0442	0.0442
040	5	19.61	4.523	0.0489	0.0488
041 $23\bar{1}$ 150	10	26.30	3.386	0.0872	0.0870 0.0872 0.0874
240	40	27.17	3.279	0.0930	0.0930
310	100	28.545	3.124	0.1025	0.1025
$32\bar{1}$	10	30.40	2.9378	0.1159	0.1159
330	20	31.85	2.8073	0.1269	0.1269
231	10	32.57	2.7468	0.1325	0.1325
$25\bar{1}$	20	33.01	2.7112	0.1360	0.1360
$16\bar{1}$ 061	10	34.46	2.6004	0.1479	0.1478 0.1481
$12\bar{2}$ 241	5	35.125	2.5527	0.1535	0.1534 0.1539
350	10	37.665	2.3861	0.1756	0.1757
251 $42\bar{1}$	10	38.295	2.3483	0.1814	0.1814 0.1819
321 $30\bar{2}$	10	38.585	2.3313	0.1840	0.1839 0.1842
$36\bar{1}$ 132	20	41.685	2.1649	0.2134	0.2135 0.2140
212 $45\bar{1}$	10	44.89	2.0174	0.2457	0.2454 0.2459
090	10	45.01	2.0124	0.2469	0.2472
510	5	48.00	1.8937	0.2789	0.2793
$38\bar{1}$ 281	5	49.89	1.8264	0.2998	0.2990 0.3004
072 530	10	50.175	1.8166	0.3030	0.3024 0.3037
381	10	55.61	1.6512	0.3668	0.3670
$61\bar{1}$ 501	5	55.97	1.6414	0.3712	0.3710 0.3712
$38\bar{2}$ $62\bar{1}$	5	56.685	1.6225	0.3799	0.3796 0.3802
352 $42\bar{3}$ $15\bar{3}$	5	58.05	1.5875	0.3968	0.3966 0.3967 0.3971

with the »iddingsite» from New Mexico studied by Ming-Shan Sun (1957). The amount of Si in the »iddingsite» of Moroto is lower and that of Al is higher than in any of the »iddingsites» presented in the above papers. In addition, the content of ferric iron is one of the highest recorded for the »iddingsites». The complete absence of ferrous iron is noteworthy. The specific gravity and the refractive indices are highest recorded for »iddingsite.» The

Table 13. Chemical Composition and Physical Data for »Iddingsite» from Pyroxene-trachyte. Specimen FEAE No. 124. Mt. Moroto, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1957.

Weight per cent		Physical data	
SiO_2	17.52	Sp. gr.	3.324
TiO_2	0.42	α	1.855—1.870
Al_2O_3	8.00	γ	1.920—1.940
Fe_2O_3	55.09	$2V_\alpha$	ca. 50°
FeO	0.00		
MnO	0.45	Sub-cell:	
MgO	2.73	a_0	4.59 ± 0.03
CaO	1.29	b_0	9.98 ± 0.03
Na_2O	0.18	c_0	2.99 ± 0.02
K_2O	0.31		
H_2O+	11.52	Super-cell:	
H_2O-	2.22	$2b_0$, $4c_0$	
Total	99.73		

mineral is zoned, as is seen from the variations of the refractive indices. The zoning is seen in thin section in the variation of the birefringence. Dispersion extremely strong. The »iddingsite» single crystals show always straight extinction with respect to the visible crystal faces of olivine. Accordingly, the crystallographic axes of olivine and the axes of the optical indicatrix of »iddingsite» are parallel to each others: »iddingsite» α (goethite b) parallel to olivine b; »iddingsite» β (goethite a) parallel to olivine a; »iddingsite» γ (goethite c) parallel to olivine c.

The medium-grained groundmass contains the following constituents: K-monalbite, clinopyroxene, analcite, magnetite, »iddingsite», apatite and a little sodian plagioclase. K-monalbite with the composition $Or_{47}Ab_{53}$ ($d_{(20\bar{1})} = 4.129$) is quantitatively the most important constituent. The laths show random orientation. $\alpha = 1.524$, $\gamma = 1.530$.

Specimen FEAE No. 110 *(pyroxene trachyte). Sp. gr. 2.93.* — This rock is almost identical with specimen FEAE No. 124.

TRACHYBASALTS

The trachybasalts are represented by 9 specimens. They all contain calcian plagioclase (composition $Ab_{50}An_{50}$ to $Ab_{40}An_{60}$) in abundance. In addition, titanian clinopyroxene, olivine (composition varies between 18 and 42 mol. % Fa) and alkali feldspar occur as essential constituents. Nepheline has not been detected in these rocks. A little analcite or natrolite of secondary or deuteric origin is generally present in the groundmass.

Table 14. Chemical Composition of Trachybasalt. Specimen FEAE No. 123. Mt. Moroto, Karamoja, Uganda. Analyst: Pentti Ojanperä, 1957.

Weight per cent		Molecular norm		Niggli values	
SiO_2	44.74	or	7.6	si	98
TiO_2	2.80	ab	20.1	ti	4.6
Al_2O_3	11.51	an	14.0	al	14.8
Fe_2O_3	4.47	ne	5.8	fm	50.4
FeO	9.19	Salic	47.5	c	26.3
MnO	0.23			alk	8.5
MgO	7.89	di	30.0	k	0.20
CaO	11.23	ol	11.5	mg	0.51
Na_2O	3.19	mt	4.8	o	0.15
K_2O	1.23	il	4.0	c/fm	0.52
P_2O_5	0.63	ap	1.4	qz	—36
CO_2	0.30	cc	0.8		
H_2O+	2.30				
$H_2O—$	0.26	Femic	52.5		
Total	99.97	C.I.P.W. symbol: III, 5, 2, 4.			

Specimen FEAE No. 123. *Sp. gr. 3.01.* — The planimetric analysis of this porphyritic rock shows the following amounts of phenocrysts: clinopyroxene 26 %, olivine 5 %, magnetite 3 %, amygdules (natrolite, calcite, analcite) 1 %. Table 14 presents the chemical composition of the rock.

The clinopyroxene phenocrysts show well-developed crystals with the faces (100), (010), (110), (111) and (001) and average 2—4 mm in length. In thin section, the mineral is non-pleochroic with a slightly brownish violet color typical of the titanian variety. In a few cases, hour-glass texture is seen. Twinning common, zoning very strong. $\alpha \sim 1.697$, $\beta \sim 1.704$, $\gamma \sim 1.728$, $2V_\gamma = 53° — 58°$, $c \wedge \gamma \sim 43°$. The mineral is quite fresh, but contains inclusions of magnetite and hollows filled with the groundmass. The euhedral olivine phenocrysts (31 mol. % Fa) are of the same size as the clinopyroxene phenocrysts, and show alteration along the fissures and cracks to »iddingsite». When unaltered, the mineral shows no inclusions. The magnetite occurs as grains with a diameter up to 0.5 mm.

The groundmass is composed of subparallel oriented laths of plagioclase and of a fine-grained matrix containing magnetite, columns of titanian clinopyroxene, alkali feldspar, analcite and apatite. The salic constituents amount to ca. 70 % of the groundmass. Plagioclase is quantitatively the most important constituent. It is twinned on the albite and Carlsbad laws. $\alpha = 1.557$, $\gamma = 1.565$. Composition $Ab_{45}An_{55}$. According to the powder pattern, the plagioclase is structurally intermediate between the high and low series of plagioclases (Smith and Gay, 1958). $\Gamma = 0.81°$; $B = 0.84°$ (CuKα). The alkali feldspar occurs as very small interstitial anhedral grains. Refractive indices:

$\alpha = 1.525$, $\gamma = 1.531$. According to the powder pattern, the mineral is Na-sanidine with a composition $Or_{51}Ab_{49}$ ($d_{(20\bar{1})} = 4.136$). The amount of zeolites present in the amygdules as well as in the groundmass is rather important, and is reflected in the relatively high water content in the chemical analysis of the rock. In addition, the analysis shows 5.8 % nepheline in the norm. As there is actually no nepheline in the rock, the feldspathoidal molecule $NaAlSiO_4$ present in the norm is presumably due to the zeolites.

Specimen FEAE No. 128. *Sp. gr. 2.98.* — This rock is virtually identical with the specimen FEAE No. 123. The composition of the olivine is 25 mol. % Fa.

Specimen FEAE No. 126. *Sp. gr. 2.96.* — This rock is porphyritic with clinopyroxene as quantitatively the most important phenocryst mineral, the crystals of which range up to 10 mm in length, and are zoned and show hour-glass texture and the violet color typical of the titanian variety. Olivine (33 mol. % Fa) shows alteration along fissures. The zoned plagioclase laths measure up to 2 mm in length. $\alpha = 1.556$—1.559, $\gamma = 1.564$—1.568. Composition $Ab_{46-40}An_{54-60}$. Magnetite is present as relatively large crystals. Calcian plagioclase is the main constituent of the fine-grained groundmass which, in addition, contains magnetite, clinopyroxene, olivine, alkali feldspar and analcite.

Specimen FEAE No. 129. *Sp. gr. 2.89.* — A planimetric analysis of this porphyritic rock shows the following amounts of phenocrysts: plagioclase 12 %, clinopyroxene 18 %, olivine 4 %, magnetite 5 %. The clinopyroxene shows hour-glass texture and a violet color typical of the titanian variety. $\alpha = 1.700$, $\beta = 1.711$, $\gamma = 1.730$, $2V_\gamma = 50°$—$54°$, $c \wedge \gamma = 42°$—$49°$. The euhedral olivine (41 mol. % Fa) is partly altered to a brownish or greenish »iddingsite». The zoned plagioclase phenocrysts show albite, pericline and Carlsbad twinning. The mineral is clear and unaltered. $\alpha = 1.554$, $\gamma = 1.562$. Composition $Ab_{50}An_{50}$. According to the powder pattern, the mineral is structurally intermediate between the high and the low series of plagioclases (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.69°$; $B = 0.85°$ (CuKα). The medium-grained groundmass, dark gray in hand specimen, shows under the microscope the following constituents: calcian plagioclase, titanian clinopyroxene, olivine, magnetite, alkali feldspar, analcite, apatite and natrolite. The amygdules are filled with analcite, natrolite and phillipsite.

Specimen FEAE No. 125. *Sp. gr. 2.86.* — This rock is virtually identical with the specimen FEAE No. 129. The following data have been recorded for the phenocryst minerals: plagioclase, $\alpha = 1.554$, $\gamma = 1.562$, $Ab_{50}An_{50}$. According to the powder pattern, the mineral is structurally intermediate between the high and the low series of plagioclases (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.69°$ and

$B = 0.86°$ (CuKα). Clinopyroxene, $\alpha = 1.699$—1.702, $\beta \sim 1.710$, $\gamma = 1.730$—1.734, $2V_\gamma = 40°$—$53°$, $c \wedge \gamma = 40°$—$44°$. Olivine composition 42 mol. % Fa.

Specimen FEAE No. 118. *Sp. gr. 2.83.* — This rock is virtually identical with the specimens Nos. 125 and 129. However, it contains more amygdules filled with spherulitic natrolite. The following data have been recorded for the phenocrysts: plagioclase (partly replaced by natrolite), $\alpha = 1.553$—1.556, $\gamma = 1.561$—1.564, $Ab_{52-47}An_{48-53}$. According to the powder pattern, the mineral is structurally intermediate between the high and the low series of plagioclases (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.74°$, $B = 0.84°$ (CuKα). Olivine composition 21 mol. % Fa.

Specimen FEAE No. 117. *Sp. gr. 2.82.* — Clinopyroxene is quantitatively the most important phenocryst mineral. It shows hour-glass texture and a slightly violet color. Plagioclase is next in the order of relative abundance. It is unaltered and shows albite, pericline and Carlsbad twinning. $\alpha = 1.557$, $\gamma = 1.565$, $Ab_{45}An_{55}$. According to the powder pattern, the mineral is structurally intermediate between the high and the low temperature series of plagioclases (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.82°$ and $B = 0.82°$ (CuKα). In addition, the rock contains phenocrysts of olivine (30 mol. % Fa) and magnetite. The groundmass is fine-grained and shows trachytic texture with laths of calcian plagioclase in subparallel orientation. Other constituents are alkali feldspar, magnetite and clinopyroxene. No amygdules.

Specimens FEAE Nos. 131 and 120. *Sp. gr. 2.96—2.97.* — These two rocks are virtually identical with each other. Slightly violet clinopyroxene is quantitatively the most important constituent of the phenocrysts. Next in abundance is euhedral olivine (18 mol. % Fa for FEAE No. 131 and 31 mol. % Fa for FEAE No. 120), partly altered. Some magnetite and brown hornblende. The fine-grained groundmass shows trachytic texture. The laths of plagioclase (andesine-labradorite) lie in subparallel orientation, the interstices being occupied by clinopyroxene, magnetite, alkali feldspar with cross-hatched twinning and analcite.

ANALCITE-BEARING TRACHYDIABASE

Specimen FEAE No. 121. *Sp. gr. 2.92.* — The fine-grained subophitic rock contains a few plagioclase phenocrysts. The following minerals have been recorded: plagioclase, clinopyroxene, olivine, alkali feldspar, analcite, brown hornblende, apatite and magnetite.

Table 15. Chemical Composition and Physical Data for Olivine from Analcite-bearing Trachydiabase. Specimen FEAE No. 121. Mt. Moroto, Karamoja, Uganda. Analyst: Aarno Juurinen, 1955.

Weight per cent		Physical data	
SiO_2	34.08	Sp. gr.	3.904
TiO_2	0.04	α	1.747—1.754
Al_2O_3	0.00	β	1.775—1.783
Fe_2O_3	0.27	γ	1.790—1.798
FeO	47.30	$2V_\alpha$	71°
MnO	0.65	d(130)	2.8077
MgO	17.83		
CaO	0.00		
H_2O+	0.05		
H_2O-	0.00		
Total	100.22		
Calculated composition:			
Fe_2SiO_4	59.5 mol. %		
Mn_2SiO_4	0.8		
Mg_2SiO_4	39.7		

Plagioclase is the most important quantitatively. It forms unaltered laths, twinned on the albite and Carlsbad laws and ranging sometimes up to 4 mm in length. $\alpha = 1.559$, $\gamma = 1.568$. $Ab_{40}An_{60}$. According to the powder pattern, the plagioclase is structurally of the low-temperature type (Smith and Gay, 1958) with the following angular separations: $\Gamma = 0.58°$, $B = 0.88°$ ($CuK\alpha$). The alkali feldspar occurs interstitially between the plagioclase laths. The powder pattern indicates the alkali feldspar to be triclinic with the composition ca. $Or_{30}Ab_{70}$ ($d_{(20\bar{1})} = 4.095$). $\alpha = 1.525$, $\gamma = 1.532$. Analcite occurs interstitially between the other constituents of the rock. It is also found as a filling in the amygdules. The zoned clinopyroxene shows a slightly violet color typical of the titanian variety. $\alpha \sim 1.699$, $\beta \sim 1.705$, $\gamma \sim 1.727$, $2V_\gamma \sim 50°$. The hornblende amounts to ca. 1 % or slightly less of the bulk rock. It forms small subhedral grains (0.1—0.2 mm), usually clear. $\alpha = 1.651$, $\beta = 1.663$, $\gamma = 1.674$, $2V_\alpha = 80°—85°$, b // β, $c \wedge \gamma = 18°—22°$, γ greenish brown $\sim \beta$ chocolate brown $> \alpha$ brownish yellow. Magnetite is rather abundant (ca. 5 % of the bulk) and occurs as euhedral grains.

Olivine was separated out of the rock and chemically analyzed. Table 15 gives the result of the analysis of a very pure fraction. The variations of the refractive indices are caused by zoning. The grains are anhedral ranging up to 0.5 mm in diameter. The mineral is partly altered. The material and the chemical analysis of the olivine have been used by Yoder and Sahama (1957) in developing a method for the determination of the composition of natural olivines by X-ray powder diffraction.

PHILLIPSITE-CLINOPYROXENE ROCK

Specimen FEAE No. 113. — The rock is characterized by the abundance of phillipsite present in the groundmass and in the amygdules. It is porphyritic with a few phenocrysts of clinopyroxene and reddish brown »iddingsite». Clinopyroxene is the quantitatively predominating femic constituent. The larger phenocrysts are zoned with a colorless core and a slightly grayish margin. The core is free from inclusions whereas the margin contains opaque inclusions in abundance. The smaller clinopyroxene phenocrysts are rich in the opaque inclusions. Olivine forms well-developed, euhedral, relatively large phenocrysts, almost completely altered to »iddingsite». The groundmass contains columnar grayish clinopyroxene, »iddingsite», magnetite, prisms of apatite and interstitial phillipsite. The amygdules contain phillipsite forming either spherulites or penetration twins, and some natrolite.

OLIVINE-MELANEPHELINITES

Specimens FEAE Nos. 109 and 115. *Sp. gr. 3.10—3.12.* — These two specimens are virtually identical with each other. They show abundant clinopyroxene and a few olivine and nepheline phenocrysts. A planimetric analysis of the specimen FEAE No. 109 yields the following amounts of phenocrysts: clinopyroxene 42 %, olivine 12 %, magnetite 4 %. The clinopyroxene forms twinned and zoned euhedra with almost colorless core and slightly violet margin. $\alpha = 1.683—1.700$, $\gamma = 1.715—1.728$, $2V_\gamma$ varies from 60° in the core to 52° in the margin, $c \wedge \gamma$ from ca. 51° to 45°, respectively. The olivine phenocrysts show a composition of 15 mol. % Fa. The crystals are partly altered to brown »iddingsite». In specimen FEAE No. 115, nepheline occurs as rare anhedral phenocrysts. In specimen FEAE No. 109, the nepheline forms aggregates of a few small crystals; composition $Ne_{79}Ks_{21}$. Magnetite occurs as euhedral crystals. The groundmass is very fine-grained, almost aphanitic, and contains nepheline in addition to clinopyroxene and magnetite. The amygdules are filled with natrolite and phillipsite.

Specimen FEAE No. 111. *Sp. gr. 3.02.* — This porphyritic rock contains phenocrysts of clinopyroxene and olivine. The clinopyroxene is strongly zoned, twinned, almost colorless. It contains magnetite inclusions and hollows filled with natrolite. Olivine (11 mol. % Fa) is often altered into a fibrous birefringent greenish material. The groundmass is very fine-grained and contains the following constituents: anhedral interstitial nepheline, euhedral clinopyroxene, magnetite, pleochroic mica ($2V_\alpha$ ca. 10°) and apatite.

Poikilitic calcite occurs in several spots as irregular grains. The rock contains a few narrow fissures filled with natrolite, nepheline, alkali feldspar and calcite.

SUMMARY OF THE PETROGRAPHY OF THE MT. MOROTO ROCKS STUDIED

Compared with the tinguaites of the Toror Hills, the volcanic rocks of Mt. Moroto described in this paper show considerably more petrographical and chemical variation. As only two analyses (Tables 10 and 14) have been made, the range of the chemical variations can not be illustrated by accurate numerical data or variation diagrams but must be based mainly on microscopic observations. A few of the rocks are representatives of the ultramafic clan (melanephelinites) and the great bulk belongs to the alkali gabbro, to the nepheline syenite and to the syenite clans. Using the Shand's (1947) classification based on saturation with respect to silica, the rocks of Mt. Moroto are undersaturated and belong to the feldspathoidal division. In the olivine-melanephelinites, the alkalies are completely unsaturated and no feldspar is present. In the trachybasalts, trachytes and phonolites, the alkalies are partly saturated. Feldspars are present in abundance. Nepheline, sodalite, analcite or other zeolites do occur as essential yet less abundant constituents. In the classification based on the alumina-saturation principle (Shand, 1947), the volcanic rocks of Mt. Moroto are either subaluminous (trachybasalts and trachytes) or peralkaline (phonolites). The zeolites have not the same importance as in the Toror Hills and can not be used as a basis of the rock classification. As in the other lavas connected with the Eastern Rift Valley, sodium predominates over potassium in the Mt. Moroto rocks. Microscopic study supports the idea that, in contrast to the hypabyssal rocks of the Toror Hills, surface lavas are abundantly represented in Mt. Moroto, in accordance with Bisset's observations (1935) on the geological structure.

Compared with the lavas of Napak and Elgon, the volcanic rocks of Mt. Moroto show a feature which is rather characteristic. Ten of the 23 specimens collected in the Nediket valley at the base of Mt. Moroto are trachybasalts or trachydiabases carrying andesine-labradorite as a main constituent. According to the detailed investigations of Ödman (1930), King (1949) and Davies (1952), no similar trachybasalts or other rocks containing calcian plagioclases are represented among the lavas of Napak and Elgon. In addition, no basaltic rocks are present in the Toror Hills complex (Du Bois, 1956). On the other hand, Davies (1953) has described a tholeiitic basalt and a dolerite occurring in the most northerly part of the Karamoja district, some 9 km E of Loyoro. Further, basaltic rocks including basalts, trachybasalts, basanites and tephrites are common among the volcanic rocks of Turkana in Kenya, the province bordering the district of Karamoja in the east (Smith, 1938).

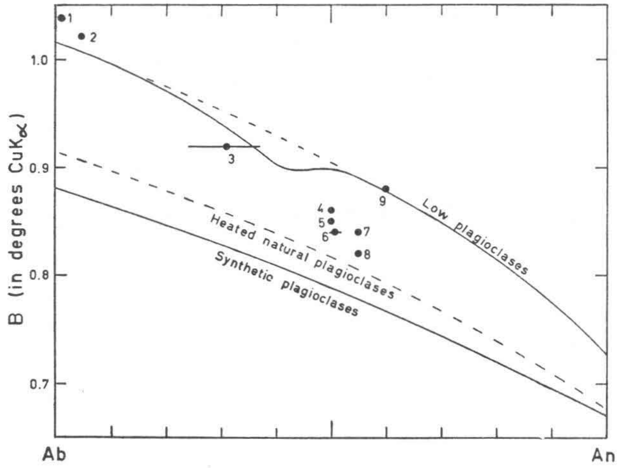

Fig. 6. *B*-values for some Toror Hills and Mt. Moroto plagioclases plotted against the plagioclase composition.

No. 1.	Specimen H. 9.	Toror Hills.
2.	FEAE No. 134.	» »
3.	FEAE No. 114.	Mt. Moroto.
4.	FEAE No. 125.	» »
5.	FEAE No. 129.	» »
6.	FEAE No. 118.	» »
7.	FEAE No. 123.	» »
8.	FEAE No. 117.	» »
9.	FEAE No. 121.	» »

A few remarks may be added as to the main constituents of the Mt. Moroto rocks studied. The angular separations B and Γ in plagioclases were measured in six trachybasalts and one trachyte. The B-values obtained were plotted against the plagioclase composition in Fig. 6. From this figure reproduced in simplified form from Smith and Gay (1958, p. 754), it is seen that the plagioclases of the Mt. Moroto rocks studied are mostly intermediate between the low series of natural plagioclases and the heated natural plagioclases and correspond to natural volcanic plagioclases. Only the plagioclases FEAE Nos. 114 and 121 correspond to the low-temperature plagioclase series. In trachybasalts and trachydiabases, the olivine composition varies relatively greatly, ranging from 18 to 60 mol. % Fa. The compilation published by Yoder and Sahama (1957) reveals a relatively wide range of variation for the olivines in the rocks of the gabbro and alkali gabbro clans. In one phonolite (FEAE No. 112), a virtually pure fayalite was found. The occurrence of fayalite in phonolitic rocks seems to be rather common. The ratio of the diopside component to the hedenbergite component in the clinopyroxenes of the trachybasalts, as deduced from the powder pattern (not reproduced), is not in accordance with the optical properties. Compared with the data given by Zwaan (1954), the powder patterns show that the Mt. Moroto clinopyroxenes would not be very far from the magnesium end of the diopside-hedenbergite series with some pigeonite whereas the refractive indices indicate higher iron contents. This discrepancy, together with the color, optic axial angle and mode of occurrence, indicates that the clinopyroxenes of Mt. Moroto are usually represented by more or less titanian varieties. No chemical analyses of clinopyroxenes from Mt. Moroto have been made. One hornblende (specimen FEAE No. 114) and one »iddingsite» (specimen FEAE No. 124) have been studied in greater detail.

OLIVINE-MELANEPHELINITE OF MORULINGA

The small volcanic neck of Morulinga is represented by two specimens, *viz.* FEAE No. 139 and H. 18, collected from loose boulders on the NE slope of the hill. Megascopically the two specimens are almost identical. The rock is dark gray and porphyritic with no preferred orientation and without vesicles. Black clinopyroxene and greenish olivine phenocrysts, up to 3 mm in diameter, are visible in hand specimen. Microscopic study shows that the groundmass is holocrystalline and contains the following minerals: clinopyroxene, nepheline, olivine, magnetite, pale mica, sodalite and pyrite. In addition, some cracks filled with natrolite are seen. The specific gravity of the rock is 3.13. Planimetric analysis of the rock gave the following amounts of phenocrysts: clinopyroxene 19 %, olivine 3 %. The chemical composition of the rock (specimen FEAE No. 139) is presented in Table 16. The rock is characterized by a complete absence of the feldspars. As salic constituents it contains only feldspathoids, *viz.* nepheline and a little sodalite. The amount of femic constituents exceeds 50 % of the bulk. Accordingly, the name olivine-melanephelinite was adopted here. Both chemically and mineralogically the rock is rather close to the olivine-melanephelinites (or tannbuschites according to the nomenclature proposed by Johannsen, 1938) from the Mittelgebirge of Bohemia (Hibsch, 1898). Further, some of the analyzed melanephelinites and olivine-nephelinites from Napak (King, 1949) show clear mineralogical and chemical similarities with the olivine-melanephelinite of Morulinga. Microscopically, the olivine-melanephelinites of Mt. Moroto, specimens FEAE Nos. 109 and 115, are very similar to the Morulinga rock.

The euhedral clinopyroxene is non-pleochroic and colorless. The violet color typical of the titanian clinopyroxene is lacking. Twinning is common, zoning very strong. The clinopyroxene separated from the specimen FEAE

Table 16. Chemical Composition of Olivine-melanephelinite. Specimen FEAE No. 139. Morulinga, Karamoja, Uganda. Analyst: Kai Hytönen, 1953[1].

Weight per cent		Molecular norm		Niggli values	
SiO_2	42.92	lc	9.7	si	81
TiO_2	2.08	ne	20.7	ti	3.0
Al_2O_3	8.66	hl	0.4	al	9.7
Fe_2O_3	5.99	Salic	30.8	fm	49.1
FeO	5.62			c	30.2
MnO	0.15	ac	4.2	alk	11.0
MgO	11.11	di	41.1	k	0.23
CaO	14.88	ol	9.1	mg	0.64
Na_2O	4.60	Ca_2SiO_4	5.5	o	0.17
K_2O	2.08	ap	0.7	c/fm	0.62
P_2O_5	0.32	cc	0.4	qz	—59
CO_2	0.17	il	2.9		
Cl	0.14	mt	4.6		
S	0.28	pr	0.7		
H_2O+	1.13				
$H_2O—$	0.10	Femic	69.2		
	100.23				
—O	0.10	C.I.P.W. symbol: IV, 2, 3, 2.			
Total	100.13				

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[1] Cl and S by Pentti Ojanperä, 1957

No. 139 was found to be very pure, and the result of the chemical analysis and the physical data of the mineral are summarized in Table 17. The powder pattern of the mineral is presented in Table 18. The unit cell dimensions were determined from the powder pattern. The unit cell dimensions given in Table 17 are close to the values for pure diopside. For comparison the unit cell dimensions of two analyzed clinopyroxenes (Hess, 1949; Kuno, 1955) with almost the same chemical composition are given below:

	Diopside. St. Lawrence Co. N.Y.	Salite. Taga volcano, Wadaki, Japan.
	$Ca_{49}Mg_{47}Fe_4$ 0.4 % Al_2O_3	$Ca_{46}Mg_{43}Fe_{11}$ 4.9 % Al_2O_3
a_0	9.750	9.742
b_0	8.930	8.901
c_0	5.249	5.268
β	105° 50′	106° 5′

The most marked difference in chemical composition between the three clinopyroxenes is in the aluminum content. According to Kuno (1955), the aluminum content affects the values for b_0 and c_0 of the unit cell, but does not appear to affect the values for a_0 and β. The a_0- and the β-values of

Table 17. Chemical Composition and Physical Data for Clinopyroxene from Olivine-melanephelinite. Specimen FEAE No. 139. Morulinga, Karamoja, Uganda. Analyst: H. B. Wiik, 1953.

Weight per cent		Unit cell content		Physical data	
SiO_2	53.13	Si	7.72	Sp. gr.	3.316
TiO_2	0.26	Ti	0.03	α	1.678—1.683
Al_2O_3	2.37	Al	0.41	β	1.685—1.690
Fe_2O_3	0.72	Fe^{3+}	0.08	γ	1.707—1.712
FeO	3.12	Fe^{2+}	0.38	$2V_\gamma$	58°—64°
MnO	0.09	Mn	0.01	$c \wedge \gamma$	39°—44°
MgO	16.95	Mg	3.67	a_0	9.746 ± 0.004
CaO	22.91	Ca	3.57	b_0	8.912 ± 0.004
Na_2O	0.59	Na	0.17	c_0	5.253 ± 0.003
K_2O	0.15	K	0.03	$a_0 : b_0 : c_0 = 1.094 : 1 : 0.589$	
P_2O_5	0.07	P	0.01	β	106° 2′ ± 5′
H_2O+	0.08	O	23.99	Volume 438.5 Å^3	
H_2O-	0.02				
Total	100.46				

Calculated composition:

$NaFeSi_2O_6$	2.0 mol. %
(Na, K) (Al, Ti) Si_2O_6	2.8
$CaSiO_3$	44.5
$MgSiO_3$	45.8
(Fe, Mn) SiO_3	4.9

the three clinopyroxenes are identical within the accuracy of the measurement whereas the values for b_0 and c_0 depend on the aluminum content. The variations in the optical properties are caused by zoning, which is limited to a relatively narrow area in the margin of the crystal. The core shows a uniform extinction. The extinction angle is greater in the margin, the difference ranging up to 5°. According to the diagram presented by Hess (1949, p. 641), the content of $FeSiO_3$ increases in the margin.

Table 19 presents the chemical composition and physical data for the olivine from specimen FEAE No. 139. A grain count made on the material to be analyzed showed it to contain 1.6 % of clinopyroxene. The result of the analysis was corrected for this admixture. That the olivine is zoned is indicated by the variation of the optical properties. In thin section, the zoning may be detected with a rotating (elliptical) mica compensator or with phase contrast optics. The refractive indices and the birefringence increase towards the margin of the olivine crystals indicating a higher fayalite content in the margin. The mineral is partly altered. The material and the chemical analysis of the olivine have been used by Yoder and Sahama (1957) in developing a method for the determination of the composition of natural olivines by X-ray powder pattern.

Table 18. Powder Pattern of Clinopyroxene from Olivine-melanephelinite. Specimen FEAE No. 139. Morulinga, Karamoja, Uganda. Filtered Copper Radiation with Silicon Standard.

hkl	I	$2\,\Theta_{obs.}$	$d_{obs.}$	$Q_{obs.}$	$Q_{calc.}$
020	10	19.92	4.453	0.0504	0.0504
021	5	26.69	3.337	0.0898	0.0896
220	80	27.615	3.227	0.0960	0.0959
$22\bar{1}$	100	29.86	2.9896	0.1119	0.1118
310	80	30.325	2.9448	0.1153	0.1151
$31\bar{1}$	40	30.88	2.8932	0.1195	0.1194
$13\bar{1}$	30	34.985	2.5625	0.1523	0.1523
002	20	35.54	2.5238	0.1570	0.1569
221	50	35.725	2.5111	0.1586	0.1585
$30\bar{2}$ 311	20	39.20	2.2962	0.1897	0.1895 0.1894
112	5	40.75	2.2123	0.2043	0.2043
022	5	41.04	2.1974	0.2071	0.2073
330	20	41.935	2.1525	0.2158	0.2159
$33\bar{1}$	30	42.37	2.1314	0.2201	0.2201
$42\bar{1}$	20	42.895	2.1065	0.2253	0.2253
$14\bar{1}$ 041	20	44.395	2.0388	0.2406	0.2404 0.2407
$40\bar{2}$	10	44.89	2.0175	0.2457	0.2459
$13\bar{2}$	5	46.115	1.9666	0.2586	0.2583
510	10	49.70	1.8329	0.2976	0.2975
150	30	52.18	1.7514	0.3260	0.3262
$22\bar{3}$	40	56.62	1.6242	0.3791	0.3791
440	10	56.985	1.6146	0.3836	0.3838
530	5	58.185	1.5842	0.3984	0.3982
600	10	59.14	1.5608	0.4105	0.4102
350	20	59.665	1.5484	0.4171	0.4174
$62\bar{1}$	10	60.645	1.5257	0.4296	0.4298
521	10	60.855	1.5209	0.4323	0.4328
060	10	62.46	1.4856	0.4531	0.4533
531	30	65.70	1.4200	0.4959	0.4958
$35\bar{2}$	10	66.285	1.4088	0.5039	0.5043
252	10	70.705	1.3312	0.5643	0.5640
512	10	71.16	1.3238	0.5706	0.5710
$26\bar{2}$	5	73.90	1.2814	0.6090	0.6091
004	5	75.20	1.2624	0.6275	0.6277

The groundmass of specimen FEAE No. 139 is very fine-grained, almost aphanitic. In specimen H. 18, the groundmass is coarser and shows automorphic-granular texture. Clinopyroxene predominates quantitatively. Nepheline ($Ne_{73}Ks_{27}$) commonly shows euhedral forms. Completely isotropic sodalite occurs interstitially in the matrix between the other constituents. Mica is present as poikilitic crystals, ranging up to 1 mm in diameter, and is pale brownish, pleochroic. $\beta \sim \gamma = 1.591$. Titanian magnetite with $a_0 = 8.43$ is abundant.

Table 19. Chemical Composition and Physical Data for Olivine from Olivine-melanephelinite. Specimen FEAE No. 139. Morulinga, Karamoja, Uganda. Analyst: Aulis Heikkinen, 1953.

Weight per cent			Physical data	
Original analysis		Corrected[1]		
SiO_2	38.91	38.53	Sp. gr.	3.467
TiO_2	0.04	0.04	α	1.665—1.675
Al_2O_3	0.77	0.74	β	1.680—1.684
Fe_2O_3	0.67	0.67	γ	1.696—1.706
FeO	16.47	16.60	$2V_\gamma$	88°—92°
MnO	0.27	0.27	$d_{(130)}$	2.7786
MgO	42.95	43.15		
CaO	0.36	0.00		
H_2O+	0.00	0.00		
H_2O-	0.00	0.00		
Total	100.44	100.00		

Calculated composition:

Fe_2SiO_4	18.2 mol. %
Mn_2SiO_4	0.3
Mg_2SiO_4	81.5

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[1] Corrected for 1.6 % analyzed clinopyroxene.

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EXPLANATION TO PLATES

PLATE I

Fig. 1. Tinguaitic texture with aegirine-augite needles in the groundmass of natrolite-analcite-tinguaite. Specimen H. 13. NW slope of Moruangnamorj, Toror Hills. Magnification 90 ×. Without analyzer.

Fig. 2. Euhedral nepheline replaced by analcite in the groundmass of analcite-natrolite-tinguaite. Specimen H. 1. North Amairoi, Toror Hills. Magnification 125 ×. Without analyzer.

PLATE II

Fig. 1. Euhedral nepheline phenocryst mostly replaced by natrolite in natrolite-tinguaite. Specimen FEAE No. 137. Moroto-Kotido road, close to Nakazilet, Toror Hills. Magnification 35 ×. Without analyzer.

Fig. 2. Aegirine-augite in natrolite-pectolite-tinguaite. Specimen H. 3. S wall of West Valley II, Toror Hills. Magnification 35 ×. Without analyzer.

Fig. 3. Euhedral nepheline phenocryst completely replaced by natrolite in natrolite-analcite-tinguaite. Specimen H. 12. E foot of Lorano, Toror Hills. Magnification 50 ×. Without analyzer.

Fig. 4. Unaltered zoned nepheline phenocryst in analcite-natrolite-tinguaite. Specimen FEAE No. 133. Moroto-Kotido road, close to Nakazilet, Toror Hills. Magnification 85 ×. Nicols crossed.

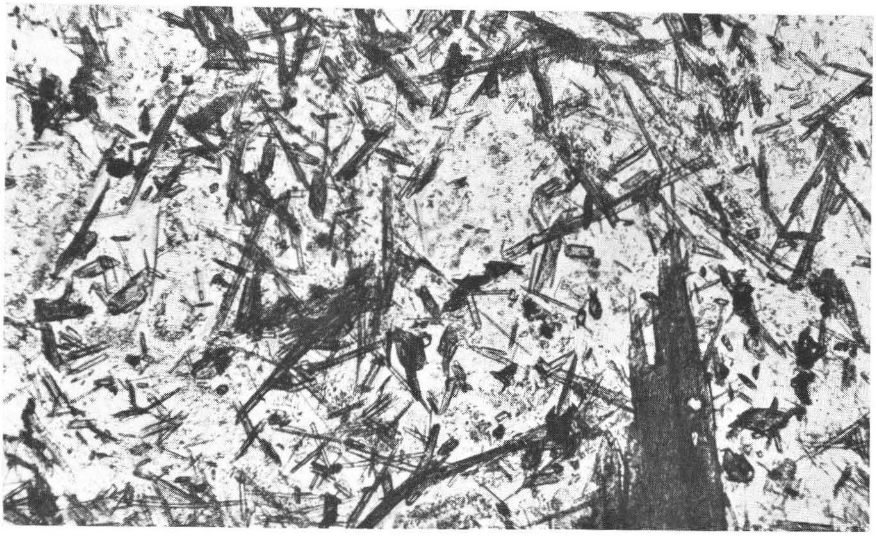

Fig. 1

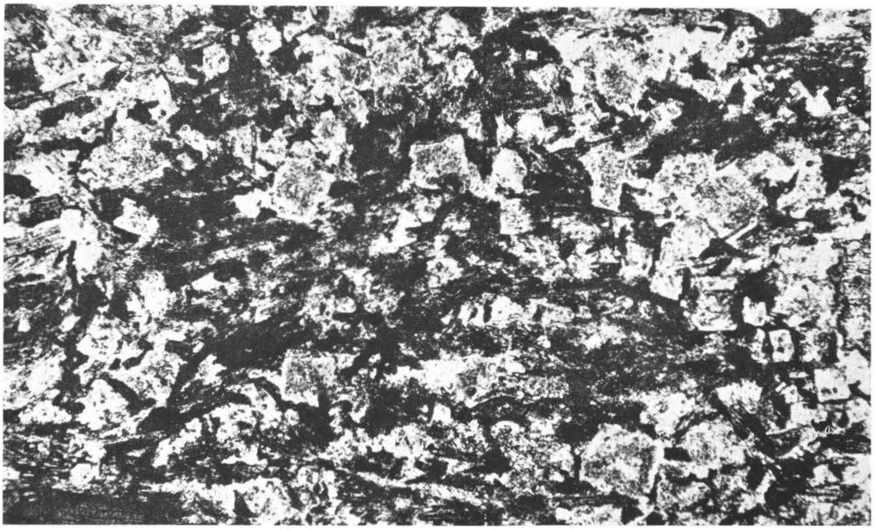

Fig. 2

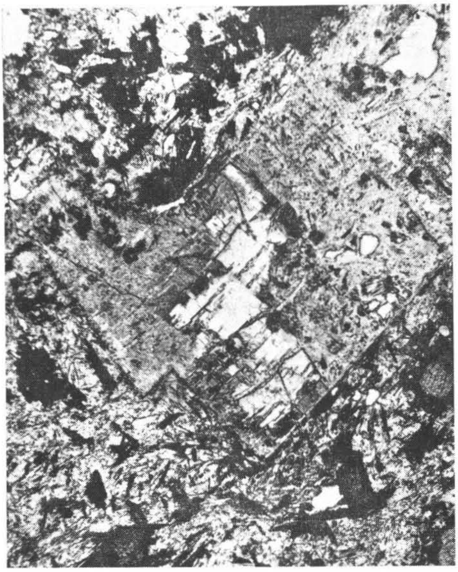

Fig. 1

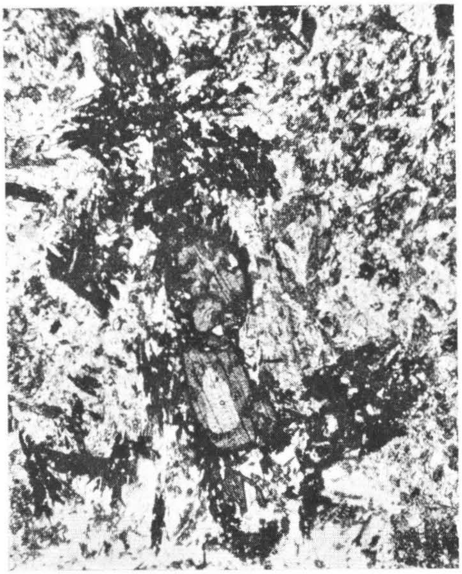

Fig. 2

Fig. 3

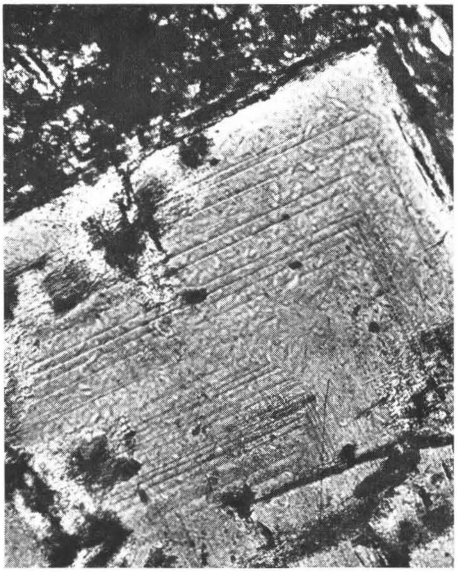

Fig. 4

Kai Hytönen. On the Petrology and Mineralogy of Some Alkaline....

7

ON THE STABILITY OF POTASH FELDSPARS [1]

BY

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In comments on the paper of Ferguson, Traill, and Taylor (1958), MacKenzie and Smith (1959) remark that the authors »make no reference to the mineralogical and petrological observations that are at variance with their conclusions.» This refers to the consequences of the theory of Ferguson et allie regarding the stability of potash feldspars.

This remark is quite correct. The present writer would add, however‘ that in respect to petrological observations a similar criticism could be directed at the conclusions drawn by MacKenzie and Smith. The question about the stability of potash feldspars being of outmost importance for petrology, granite petrology in particular, I feel impelled to partake in this discussion as a granite petrologist.

Heating experiments have indicated that microcline, under elevated temperatures (dry or wet), will become »disordered» and finally transformed into monoclinic sanidine. From this MacKenzie and Smith conclude that »microcline is more stable at a low temperature than any other potassium feldspar, including orthoclase». Theoretically, this conclusion is probably correct. But, does this experimental finding necessarily mean that microcline, at low temperatures, is more stable than orthoclase also in a geological environment, if once formed? We can definitely state that at elevated temperatures (according to Goldsmith and Laves (1954), under hydrothermal conditions, above 525°C), microcline is not stable but will be converted into monoclinic potash feldspar. But there are no experiments or direct observations to prove that at low temperatures orthoclase would be transformed into a microcline. Furthermore, in the sediments, at definitely

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[1] Received March 10, 1959.

low temperatures, the potash feldspar also grows as orthoclase; in addition, microcline has never been obtained by direct crystallization in laboratories.

For cross-hatched, natural microcline a monoclinic ancestry has been claimed by Laves, and it has been suggested that in ancient rocks, the orthoclase would alter, by aging, completely into microcline. This, an assumption, has probably been considered by MacKenzie and Smith, taking the fact that »microcline perthite is the potassium feldspar characteristic of the oldest rocks and of most pegmatites,» as an argument for the nonstability of orthoclase at low temperatures».

On a petrological basis I have shown (Marmo, 1958) that this conclusion is unwarranted, but that in plutonic rocks potash feldspar grows directly as microcline. The synkinematic and latekinematic granodiorites and granites contain exclusively microcline, which in these rocks is by no means always perthitic. To my knowledge, there are no exceptions to this rule. The undoubteadly metasomatically introduced potash feldspar, including the large porphyroblasts in the sediments, and that replacing plagioclase are likewise exclusively microcline, and not only in the pre-Cambrian, but also in the Tertiary rocks.

The postkinematic granites, on the other hand, in most cases examined are true orthoclase granites, including the Alpine pink granites, as well as the pre-Cambrian rapakivi granites. These rocks, however, are usually cut by aplitic dykes and veinlets (e. g. Marmo and Permingeat, 1957), which are much younger than the granites penetrated. These aplites are microcline granites which do not contain any traces of orthoclase. In the orthoclase granites themselves, there are always some patches of microcline present, and this is probably a product of the microclinization of orthoclase. Such patches, however, are never very abundant, and they have probably been formed at an early stage in the evolution of the rock. But in the same granites, there usually occurs a younger generation of microcline of high triclinicity, as well, and it is found in interstices, or it forms minute veinlets cutting orthoclase grains, and it is always definitely younger than other constituents of the rock (Marmo and Permingeat, 1957).

How would these phenomena fit into the theory with respect to the general microclinization of orthoclase? These facts have been always neglected when the stability of potash feldspars has been discussed, and still they are diametrically opposed to the conclusions reached by mineralogists on a crystallographic and speculative basis. Therefore, to quote MacKenzie and Smith, »it is necessary to question the validity of the theory» on which the said writers have based their opinion concerning the geological relation of microcline to orthoclase.

In my recent papers (Marmo and Permingeat, 1957; Marmo, 1958 b), I have tried to explain this phenomenon: If there is some material to be crystallized into potash feldspar, and if the temperature is above that at which microcline still survives, an orthoclase will be produced under all circumstances. If the temperature is below that at which the Al-Si framework of microcline becomes disordered, then there appears an additional factor to govern the modification of the potash feldspar to be crystallized, and this factor is time. If the introduction of material to be crystallized is sufficiently sluggish to permit the formation of a full Al-Si-order in the lattice, then the microcline will be crystallized. Otherwise an orthoclase will be formed. It is possible that, even if the growth of the crystals is slow, the very first seeds of crystals may still be of monoclinic symmetry (Marmo, 1955); but, the accumulation of the additional material being sluggish, the »seeds» will re-arrange themselves, at a very early stage, into a microcline in a form corresponding to the lower free energy, which is theoretically the stable form of potash feldspar at low temperatures. (At this point I agree with MacKenzie and Smith.) If, however, the rate of the accumulation of material exceeds that at which the transformation of monoclinic into triclinic nuclei takes place, then the orthoclase will be likewise formed at the low temperatures, as it does in laboratory experiments, as well as in the younger sediments. If an orthoclase once has been formed, then it seems to be as stable as the microcline at low temperatures, too.

As the matter of fact, such a slow accumulation of potassium does take place in the replacement phenomena; and by the replacement of albite, both Laves (1951), and Wyart and Sabatier (1956) have succeded in producing an artificial microcline. Under the conditions of metasomatism, including granitization, the conditions for the formation of microcline seem to be suitable. Therefore, in the plutonic rocks, the form of potash feldspar occurring is invariably microcline. Because microcline is likewise present in aplites younger than the orthoclase granites, conditions suitable for the growth of microcline may also be obtained in some other environment. A slow infiltration and seeping of hydrothermal solutions may be attributed to the formation of the microcline of these aplites.

This interpretation is certainly inconsistent with the magmatic views, but so are also many other problems bound to the potash feldspars. At present, however, one can not be either a magmatist or a metasomatist. In nature, both magmatic and metasomatic features certainly occur, and no geologist is entitled to accept prejudices in this matter. If some feature does not fit into, say, magmatic views, these must be revised, instead of adopting some artificially found interpretations of the feature.

As to the sodium contents of potash feldspars, I am in full agreement with MacKenzie and Smith. The microclines of plutonic rocks and those of high triclinicity are almost invariably poor in sodium. The sodium-content is somewhat higher in perthitic microclines, and the highest contents occur in orthoclase, which is mostly quite perthitic. This is in agreement with the theory concerning the growth of the primary microcline and the factors determining the modification of potash feldspar to be formed (Marmo, 1958 b).

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