GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 172

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

XXIX

HELSINKI
NOVEMBRE 1956

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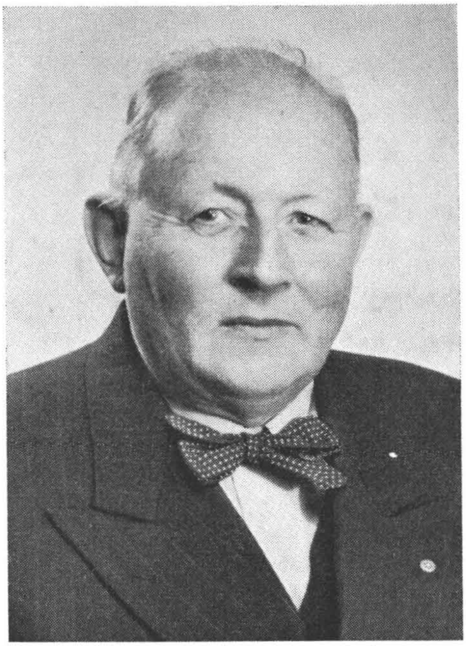

ADOLF METZGER 60 YEARS

February 26, 1956

World War I caught young Adolf Metzger, fresh from his matriculation examination in 1914, and gave him a full-time job until the end of hostilities. In 1918 Metzger came to Finland and began his studies in geology. As early as 1920 he participated in an investigation of the Porkonen—Pahtavaara area as a temporary research assistant on the staff of the Geological Survey during the summer months. There he already proved his ability and fitness for work in practical geology. In 1921 he earned his Mag. Phil. degree and in 1924 defended his thesis »Die jatulischen Bildungen von Suojärvi in Ostfinnland» being awarded the degrees of Phil. Lic. and Ph. D. in 1926.

His university studies completed, Metzger carried out his main work in the field of practical geology, chiefly in the service of the limestone industry. Circumstances prevented Metzger from publishing more than a few treatises on his favorite subject, paleontology. The results of his scientific investigations in the Parainen limestone area are compiled in the brilliant work »Zur Geologie der Inseln Ålö und Kyrklandet in Pargas-Parainen, S. W. Finnland». Here as well as in his other works Metzger visualizes the geological formations in three dimensions and presents them most lucidly to the reader.

At Åbo Akademi Metzger has performed great work as a university teacher, first as docent and later as associate professor.

The Geological Society of Finland congratulates Prof. Adolf Metzger on his anniversary, as a member of the society and scientist full of vigour and initiative, and wishes him many more years of activity for the benefit of applied and theoretical geology in Finland.

1

MINERALS OF THE KATUNGITE FLOW[1]

BY

K. J. NEUVONEN

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ABSTRACT

The chemical composition of the bulk rock and of olivine, melilite, perovskite, and kalsilite extracted from a lava specimen from Katunga volcano in Uganda are presented. The optical properties and density values of the minerals are given.

The Katunga volcano, surrounded by basement rocks in south-western Uganda about twenty kilometers south of the known Bunyaruguru volcanic field, was discovered in 1932 by A. D. Combe, who later published a description of the volcano (Combe, 1937). The lava specimens collected by Combe were examined by Holmes (1937) and he gave to the potassium-rich olivine-melilite rock the name katungite, after the volcano.

Bannister and Hey (1942) discovered the rare natural polymorph of $KAlSiO_4$, kalsilite, in some volcanic lavas of the Bunyaruguru field. Later Holmes (1950) detected this mineral also in the katungite flow of the Katunga volcano.

To obtain material for mineralogical studies of the kalsilite-nepheline series, Professor Sahama organized an expedition to the kalsilite field in East Africa. The present author was fortunate to join this expedition. The Katunga volcano was visited and two large specimens, FEAE Nos. 38 and 39, were collected from the main lava flow. Both specimens derive from the middle part of the flow about 300 m. apart. The specimens are almost identical and similar to those described by Holmes (1937). Olivine and melilite occur as phenocrysts in an aphanitic groundmass and zeolite minerals fill numerous small vesicles and fine veinlets in the rock.

One of the samples collected (FEAE No. 39) was crushed and ground for mineral separations and a portion of the homogenized powder was chemically analyzed by Mr. Aulis Heikkinen of the Geological Survey of Finland.

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[1] Received June 13, 1955.

This analysis is given in Table 1 together with the katungite analysis (C. 4407) made by A. W. Groves, published in the paper of Holmes (1937). The analyses are given both in weight and in cation per cent. Molecular norms are calculated from the analyses and given in Table 1. In the calculation »normative perovskite» was formed instead of ilmenite, and a part of the wollastonite was calculated to form »normative monticellite» because of the low SiO_2 content of the rock.

Chemically, the two samples analyzed are very much alike. The most significant difference observed is a higher calcium content in sample C. 4407, yielding more monticellite in the normative composition. The greater normative acmite for sample FEAE No. 39 is caused by the higher alkali-aluminium ratio in the composition of this sample.

Table 1. Chemical composition of the katungite flow

	Weight %		Cation %		Molecular norm				
	1	2	1	2		1		2	
SiO_2	36.20	35.37	31.83	31.01	Kp	13.55		13.65	
Al_2O_3	5.63	6.50	5.83	6.72	Ne	3.94	17.49	6.51	20.16
Fe_2O_3	8.33	7.23	5.52	4.77	Wo	15.11		16.26	
FeO	5.46	5.00	4.01	3.67	En	13.27	Di 30.21	15.16	32.52
MnO	0.22	0.24	0.16	0.19	Fs	1.83		1.10	
MgO	13.82	14.08	18.11	18.40	Fo	16.87	Ol 19.20	16.23	17.43
CaO	15.21	16.79	14.33	15.77	Fa	2.33		1.20	
Na_2O	1.74	1.32	2.97	2.24	Mo	3.55		6.24	
K_2O	4.02	4.09	4.51	4.55	Ac	6.61		0.28	
H_2O+	2.99	2.78	8.77	8.13	Mt	5.80		7.05	
H_2O-	0.84	1.15	—	—	Ap	1.73		1.47	
CO_2	0.10	0.09	0.12	0.11	Pv	6.40		5.10	
ZrO_2	n. d.	none	—	—	Cc	0.24		0.22	
TiO_2	4.84	3.87	3.20	2.55	Py	—		0.86	
P_2O_5	0.88	0.74	0.64	0.55	H_2O	8.77	82.51	8.13	
Cl	n. d.	0.02	—	0.03				rest 0.54	79.84
F	n. d.	0.16	—	0.44		100.00		100.00	
S	n. d.	0.35	—	0.57					
Cr_2O_3	n. d.	0.01	—	0.01					
V_2O_3	n. d.	0.03	—	0.02					
NiO	n. d.	0.19	—	0.13					
BaO	n. d.	0.25	—	0.08					
SrO	n. d.	0.04	—	0.02					
CuO	n. d.	0.06	—	0.04					
Li_2O	n. d.	none	—	—					
	100.28	100.36 −0.24 100.12	100.00	100.00					

1. Katungite lava, Katunga, FEAE No. 39. Analyst, A. Heikkinen.
2. Katungite lava, Western Flow, C. 4407, Katunga. Analyst, A. W. Groves (Holmes, 1937).

Mo = molecular normative monticellite, Pv = normative perovskite. In norm calculations barium and strontium are added to calcium, while manganese and nickel are added to magnesium. Cl, F, Cr, V, and Cu are omitted in the norm calculation.

Also the mineral composition of the samples now studied is similar to those described by Holmes. The amount of the phenocryst minerals and their mutual ratios, however, are different, as seen in the modes listed in Table 2. The mode of sample C. 4407 is quoted from Holmes (1937) and that of sample FEAE No. 39 was measured from four thin sections using the point counting technique of Chayes (1948).

Table 2. The mode of the two analyzed katungite samples

	FEAE No. 39	C. 4407
Olivine	16.0	21.3
Melilite	18.9	40.0
Perovskite	2.0	7.2
Zeolites, etc.	6.4	10.2
Groundmass	56.7	22.3
	100.0	101.0

Olivine, melilite, perovskite, and kalsilite were extracted from the powdered sample FEAE No. 39 using Clerici's solution and centrifuge. The pyroxene, occurring as small needles in the groundmass, was too tiny in grain size to enable separation. Zeolite minerals, occurring in abundance both in the groundmass and as vesicle fillings, were not extracted for chemical analysis.

The olivine extracted from the rock sample was chemically analyzed by Dr. H. B. Wiik. This analysis is reported in Table 3. The table also lists the number of cations scaled to four oxygen atoms. The calculated mole fraction of forsterite is 0.85 and agrees well with the optical properties given in Table 3. The measured density 3.38 is somewhat high for such a composition. Monticellite was not observed in the thin sections studied, but the analysis of the olivine shows some calcium in the mineral.

Table 3. Olivine from katungite lava FEAE No. 39, Katunga, Uganda

	wt %	cation %	cations per 4 O atoms	
SiO_2	39.83	32.97	0.987	0.991
TiO_2	0.19	0.12	0.004	
Al_2O_3	1.10	1.07	0.032	2.002
Fe_2O_3	0.07	0.04	0.001	
FeO	11.43	7.91	0.237	
MnO	0.26	0.18	0.005	
MgO	45.96	56.71	1.697	
CaO	1.11	0.98	0.029	
P_2O_5	0.02	0.02	0.001	
H_2O+	0.00	—		
H_2O-	0.06	—		
	100.03	100.00		

84.9 mole % forsterite

$\alpha = 1.658$

$\beta = 1.676$

$\gamma = 1.695$

$2V\gamma = 88°$

$D^{4°}_{20°} = 3.380 \pm 0.007$

Table 4. Melilite from katungite lava FEAE No. 39, Katunga, Uganda

	wt %	cation %	cations per 7 O atoms	
SiO_2	43.12	39.72	1.987	69.6 mole % åkermanite
Al_2O_3	4.59	4.98	0.249	8.6 % iron-åkerm.
Fe_2O_3	0.00	0.00		1.3 % gehlenite
FeO	2.13	1.64	0.082	20.5 % alkali-mel.
MnO	0.09	0.07	0.004	100.0
MgO	10.18	13.98	0.699	
CaO	35.75	35.29	1.765	Interior:
Na_2O	2.16	3.32	0.166	$\omega = \varepsilon = 1.639$
K_2O	0.69	0.81	0.039	Margins:
TiO_2	0.11	0.08	0.004	
P_2O_5	0.14	0.11	0.006	$\omega = 1.6383$, $\varepsilon = 1.6355$
H_2O+	0.81	—	(0.125)	$D^{4^\circ}_{20^\circ} = 2.981$
H_2O-	0.06	—		
	99.83	100.00		

Melilite was extracted from the powdered rock sample and the pure fraction was analyzed by Dr. Wiik. This analysis, published earlier by the writer (Neuvonen, 1955), is reproduced in Table 4 together with the optical properties and density value of the mineral. The mineral is unusually rich in potash. This must be caused by the high potash content of the rock. In the interior of the melilite crystals, birefringence is almost zero, causing a deep inky blue interference color. The margin of the crystals has a higher birefringence (0.003) and a normal color. The refractivity of the interior is about $\omega = \varepsilon = 1.639$, and that of the marginal zone, $\omega = 1.6383$, $\varepsilon = 1.6355$. The lower indices of the margin are evidently due to a higher alkali or lower iron content.

Perovskite was also separated from the sample and chemically analyzed. This analysis, made by Mr. Heikkinen, is given in Table 5 which reports index of refraction and density values of the mineral as well. The calculated cation ratio does not fully agree with the theoretical composition of perovskite. This might be caused by rare earth metals not determined in the analysis. The index of refraction of the mineral varies from 2.375 to 2.390 and the mineral shows moderate birefringence. The index measurements were performed with the dispersion method, using mixtures of sulphur and selenium.

Kalsilite was observed in only one of the thin sections cut. It occurs as small (0.05 mm) irregular grains poikilitically enclosing groundmass minerals. Consequently, the rock had to be ground very fine and the separation and purification of the kalsilite proved difficult. It was further observed, during the separation procedure, that the mineral absorbed the Clerici solution employed. Material so contaminated showed higher indices of refraction and a higher density than the pure mineral. Organic heavy liquids, however, could not be used effectively because of the high moisture content of the

Table 5. Perovskite from the katungite lava, Katunga, Uganda

	wt %	cation %	cations per 3 O atoms	
SiO_2	0.62	0.71	0.014	
TiO_2	60.16	51.74	1.019	
Al_2O_3	0.05	0.07	0.001	0.933
Fe_2O_3	0.95	0.81	0.016	
FeO	0.53	0.51	0.010	
MnO	tr.			
MgO	0.00	0.00		
CaO	37.00	45.34	0.893	
Na_2O	0.23	0.51	0.010	
K_2O	0.11	0.16	0.003	
P_2O_5	0.16	0.15	0.003	
H_2O+	0.14	—	(0.011)	
H_2O-	0.00	—		
	99.95	100.00		

$n = 2.375 — 2.390$

$D^{4^\circ}_{20^\circ} = 4.146 \pm 0.005$

pulverized material. Therefore, all the kalsilite fractions obtained were either contaminated with thallium or contained foreign mineral grains. An extract slightly contaminated with thallium was subjected to a chemical analysis. This analysis, made by Mr. Heikkinen, is reported in Table 6. The material contains 4.24 weight per cent Tl_2O and is not as pure from foreign mineral grains as desired. Omitting thallium, the cation numbers calculated to four oxygen atoms agree with the composition of the kalsilite minerals analyzed earlier (Bannister and Hey, 1942, Bannister, Sahama, and Wiik, 1953, and Sahama, 1954). The sum of cations Na, K, and Ca shows, however, a deficiency in respect to the cations Al and Si. This same tendency is observable in the earlier analyses quoted and in many nepheline analyses as well (Sahama, 1953). Likewise is the small excess of Si, observed in the present analysis, common to most of the nepheline and kalsilite analyses made.

To check the alkalis of the kalsilite of the sample, the Na_2O and K_2O were determined from another kalsilite fraction, not notably contaminated with thallium but containing quartz grains as impurities. As mentioned by Holmes (1937), the Katunga lava contains fragments of basement rocks. The quartz grains in the kalsilite fraction might originate from such a source. Unfortunately, the fraction was not sufficient for a total analysis. Mr. Heikkinen determined the alkalis with the following results: Na_2O = 2.80 %, and K_2O = 21.55 %. The amount of quartz in the fraction was determined, counting a large number of grains under the microscope, to be 2.60 %. If the alkali content of the quartz-free material is calculated, it turns out to be nearly the same as that reported in Table 6 for the uncontaminated material, namely:

Na_2O = 2.87 %, K_2O = 22.13 %, cation ratio Na/K = 0.020

Table 6. Kalsilite from Katunga lava flow, Uganda

	wt %	cation %	wt % recalc.[1]	cations per 4 O atoms		
SiO_2	37.16	34.82	38.81	1.026		
Al_2O_3	25.58	28.25	26.73	.833		
Fe_2O_3	4.73	3.34	4.94	.098		
FeO	0.43	0.34	0.45	.010	0.992	$\omega = 1.543$
MnO	0.04	0.03	0.04	.001		$\varepsilon = 1.539$
MgO	1.21	1.69	1.26	.050		
CaO	0.75	0.72	0.78	.021		$D_{20^\circ}^{4^\circ} = 2.623 \pm 0.007$
Na_2O	2.78	5.05	2.90	.149	0.918	Na/K = 0.020
K_2O	21.24	25.39	22.18	.748		
TiO_2	0.51	0.37	0.54	.011		
H_2O+	1.02	—	1.07	(0.094)		
H_2O-	0.29	—	0.30			
Tl_2O	4.24	—	—			
	99.98	100.00	100.00			

The high content of ferric iron increases the refractive indices as mentioned by Sahama (1954). The now analyzed kalsilite from Katunga is rich in ferric iron but the measured indices are relatively low. The effect of the iron might be counterbalanced by unoccupied cation sites (due to deficiency of alkalis) in the structure of this mineral.

The most abundant mineral in the groundmass of the Katunga lava in the samples studied is a fibrous mineral having an index of refraction about 1.537, and which was assumed to be ashcroftine (kalithomsonite, Hey and Bannister, 1933) by Holmes (1937). This mineral was not separated or studied more closely because of the small grain size. For the same reason it was not possible to extract pyroxene for analysis, although it occurs abundantly in the groundmass of one of the specimens (FEAE No. 39) studied. From the pyroxene needles (0.02 by 0.005 mm) the following optical properties were measured:

$\alpha = 1.680$ $2V\gamma = 78^\circ$
$\gamma = 1.705$ $c\wedge\gamma = 47^\circ$

Color pale green, pleochroism very weak.

According to the optical properties the mineral belongs near to the diopside-hedenbergite series.

Among the zeolite minerals filling the vesicles of the rock, natrolite and phillipsite are most common. The optical properties of these minerals vary greatly in different parts of the sample and within one single crystal and were not systematically studied.

To Dr. H. B. Wiik and Mr. Aulis Heikkinen, M. A., the author expresses his sincere thanks for the chemical analyses made.

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[1] Composition of the uncontaminated mineral.

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2

CANCRINITE-IJOLITE FROM IIVAARA, KUUSAMO, FINLAND[1]

BY

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ABSTRACT

Sulphatic cancrinite-bearing ijolite from Iivaara has been analysed. The sulphatic cancrinite, the nepheline and the pyroxene have been separated from this specimen. The chemical composition and optical properties of these minerals are given.

The alkaline district of Iivaara is situated about 25 km south-east of the central village of Kuusamo. There are outcrops mainly on the summit of Iivaara and on the western slope of Ahvenvaara, which lies south of Iivaara. A few outcrops are also to be found on Penikkavaara, situated about 3 km south-east of the top of Iivaara. The valley east of the said hills is covered with thick moraine.

When the author examined the earlier specimens taken from Iivaara, one of them contained a mineral that had a low refringence. This proved to be a sulphatic cancrinite (Lehijärvi, 1955). In addition, the author has separated and analysed the nepheline and the pyroxene from this specimen and also made a rock analysis.

The rock is fine-grained and grey in colour. Because suitable outcrops are lacking, it is not easy to decide how it is connected with normal ijolite. A planimetric analysis of the rock, made on the Leitz Integration Stage, gave the following results:

Nepheline	36.4	Cancrinite	2.6
Pyroxene	37.0	Apatite	2.3
Sulphatic cancrinite	20.8	Sphene	0.9

Johannsen (1950) has grouped Family 2125 P on the basis of the feldspathoids. No species of rock in which the predominant feldspathoid is

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[1] Received February 3, 1956.

cancrinite has been reported. In the species of rock herein considered the minerals belonging to the cancrinite group are among the main ones of the species, even though they are not the predominant feldspathoids. The term cancrinite-ijolite for this species of rock may, however, be justified. The chemical composition, Niggli numbers, weight norms and one-cation molecular norms (Eskola, 1954) of this specimen are given in Table 1.

Table 1. The chemical composition, Niggli numbers, weight norms and one-cation molecular norms of cancrinite-ijolite from Iivaara

	%	Mol.prop.	Niggli numbers		Cation %	O in oxides	Weight norm		One-cation molecular norm	
SiO_2	43.03	7 160	*si*	95.8	37.55	75.10	lc	4.05	Lc	3.60
TiO_2	0.58	73	*qz*	—109.0	0.38	0.76	kp	7.74	Kp	7.92
Al_2O_3	20.34	1 995	*ti*	0.96	20.91	31.37	ne	36.58	Ne	40.62
Fe_2O_3	2.26	142	*al*	26.7	1.49	2.23	an	10.26	An	9.575
FeO	3.30	459	*fm*	22.9	2.46	2.46	th	0.64	Th	0.69
MnO	0.14	20	*c*	24.2	0.10	0.10	nc	1.49	Nc	2.19
MgO	3.84	952	*alk*	26.2	4.99	4.99				
CaO	10.14	1 808	*mg*	0.55	9.48	9.48	ac	6.56	Ac	5.96
Na_2O	10.02	1 616	*k*	0.17	16.95	8.47	fs	5.36	Fs	4.36
K_2O	3.18	338	*o*	0.17	3.54	1.77	en	9.52	En	9.98
P_2O_5	1.62	114	*c/fm*	1.06	1.19	2.97	wo	12.29	Wo	11.18
H_2O+	0.60	(333)			(1.75)	—	ap	3.83	Ap	3.165
$H_2O—$	0.10	(56)			—	—	il	1.11	Il	0.76
SO_3	0.36	45			0.23	0.69		99.43		100.00
Cl	trace	—			—	—	H_2O+	0.60		
CO_2	0.62	141			0.73	1.46	$H_2O—$	0.10		
	100.13	—	—		100.00	141.85		100.13		
					—O for (OH)	1.75				
					O	140.10				
					(OH)	3.50				
					Anions pro 100.00 cations	143.60				

Under the microscope the sulphatic cancrinite looks grey because of the pigment therein. Some of the crystals are hexagonal in form. The carbonatic cancrinite that appears as inclusions is pigmentless, but contains some grains of opaque ore (Fig. 1a). In connection with the sulphatic cancrinite there occasionally occurs some zeolite, probably natrolite The chemical composition of the sulphatic cancrinite is as follows: SiO_2 35.40, Al_2O_3 29.78, Fe_2O_3 0.05, FeO 0.42, MgO 0.13, CaO 1.12, Na_2O 20.77, K_2O 0.96, Cl 0.35, SO_3 4.70, CO_2 2.16, H_2O+ 4.17 (Lehijärvi, 1955).

The nepheline is grey in colour. Under the microscope hexagonal forms may be observed. The inclusions consist of needles of pyroxene, which are sometimes ranged in accordance with the crystal outlines (Fig. 1c). The chemical composition and physical properties of the nepheline are given in Table 2 (Smith and Sahama, 1954).

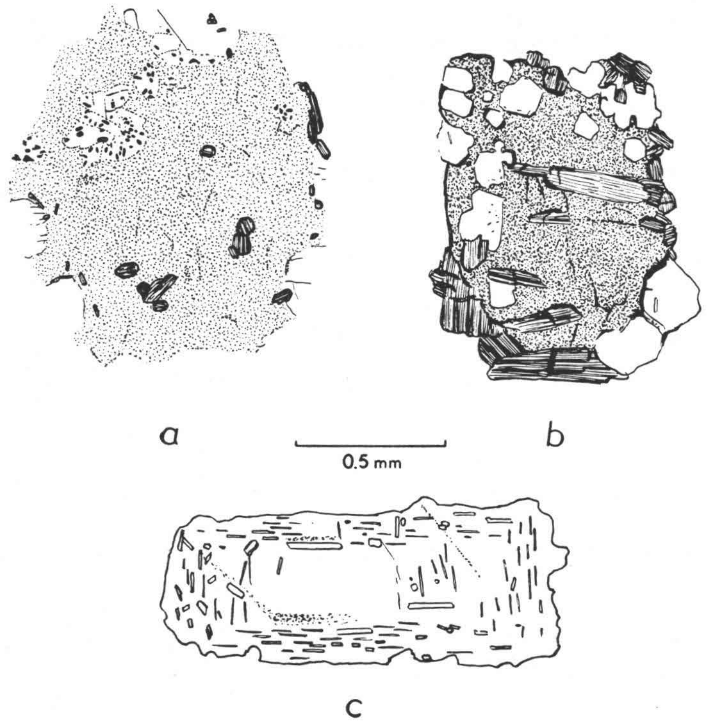

Fig. 1. (a) Dotted = sulphatic cancrinite, lined = pyroxene, white with black spots = carbonatic cancrinite. (b) Dotted = sphene, lined = pyroxene, white = nepheline. (c) Nepheline with needles of pyroxene. — Drawn by Toini Mikkola.

Table 2. Chemical composition and physical properties of nepheline from Iivaara

	%	Mol. prop.	Atomic prop. (O = 32)		
			Si	8.06	
SiO_2	41.48	6 903	Al	7.80	8.05
TiO_2	0.00	—	Fe^{3+}	0.18	
Al_2O_3	34.12	3 346	Fe^{2+}	0.04	
Fe_2O_3	1.26	79	Mg	0.03	
FeO	0.29	40	Na	6.10	7.60
MnO	0.00	—	K	1.50	
MgO	0.12	30	Unit cell dimensions [1]		
CaO	tr.	—			
Na_2O	16.20	2 613	a_0 =	10.001	±0.003
K_2O	6.06	643	c_0 =	8.379	
H_2O+	0.17	94	ω =	1.545	
H_2O-	0.00	—	ε =	1.541	
	99.70	—	Sp.gr. =	2.638 (pycnometer)	

The pyroxene takes a prismatic shape or appears as grains that have no definite form. The edges are usually darker green in colour than the

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[1] Determined by Sahama.

centre. Sometimes there are several zones. The analysis in Table 3 has been made of a material in which the different zones have not been separated.

Table 3. Chemical composition of pyroxene from Iivaara

	%	Mol. prop.	Atomic prop. (O = 24)			
SiO_2	51.92	8 640	Si	7.80		8.03
TiO_2	0.64	80	Al	0.23		
Al_2O_3	1.34	131	Ti	0.07	0.88	4.02
Fe_2O_3	7.15	448	Fe^{3+}	0.81		
FeO	6.36	885	Fe^{2+}	0.80	3.14	
MnO	0.28	39	Mn	0.03		
MgO	10.31	2 557	Mg	2.31		
CaO	19.73	3 518	Ca	3.17		3.86
Na_2O	2.28	368	Na	0.66		
K_2O	0.15	16	K	0.03		
H_2O+	0.05	28				
H_2O-	0.00	—				
	100.21	—	—			

Table 4 shows some optical properties of this pyroxene, determined by Mrs. Toini Mikkola, M.A., and its specific gravity.

Table 4. Physical properties of pyroxene from Iivaara

Refractive index		Axial angle		Specific gravity
Dark	Light	Dark	Light	
$\alpha = 1.700$	$\alpha = 1.697$	$2V\gamma = 84°$	$2V\gamma = 58°$	3.415
$\gamma = 1.734$	$\gamma = 1.724$	$2V\alpha = 78°$	$2V\gamma = 70°$	(pycnometer)

Sphene and apatite are found as accessory minerals. Sphene generally appears as idiomorphic crystals, but occasionally takes an indefinite form and contains inclusions of idiomorphic crystals of nepheline and pyroxene (Fig. 1b). Apatite appears as indefinite grains.

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3

THE FIBROUS SEPIOLITE DEPOSIT OF STANSVIK [1]

BY

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ABSTRACT

In the old iron mine of Stansvik there has been found a fibrous mineral, which, on the basis of optical, chemical and X-ray investigations, has been ascertained to be hydrous magnesium-trisilicate, parasepiolite. Parasepiolite occurs at Stansvik in a fissure vein with quartz in limestone and it has directly crystallized from hydrothermal solutions. The chemical composition of the Stansvik sepiolite may be reduced to the formula $2MgO \cdot 3SiO_2 \cdot 4H_2O$.

CONTENTS

INTRODUCTION

The present study deals with a fibrous, asbestiform mineral found in the old, abandoned iron mine of Stansvik situated within the city limits of Helsinki (Fig. 1). Since the case history of the Stansvik »asbestos» is closely linked to the past of this little, forgotten mine, I should like to touch on its earlier phases briefly.

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[1] Received April 9, 1956.

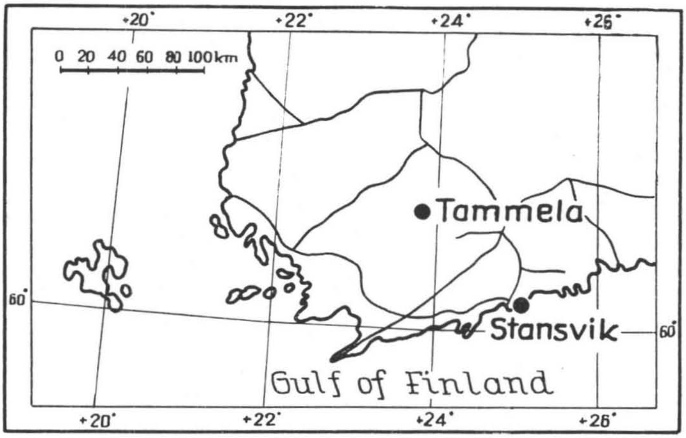

Fig. 1. Locations of sepiolite discoveries in Finland.

The iron ore occurrence of Stansvik was known quite a long time ago. Mining operations were carried on there in 1766 and in 1796—97. When, early in the 19th century, Finland was politically cut off from her mother country of Sweden, which made uncertain further imports of Swedish iron ore, concern over the situation led to an energetic search for deposits of iron ore in this country. Old mines were also investigated and their resources of ore re-evaluated. Toward such an end Nils Gustaf Nordenskiöld worked the old Stansvik mine experimentally in the years 1838—39. The results were not satisfactory. It was decided that commercial mining operations there would be unprofitable.

In the same year that test working at Stansvik was terminated, a paper was published by Sokolowski (Kazakowa, 1911) that listed a large number of different minerals found there. At the conclusion of his exposition Sokolowski tells about a long cavity occurring in the limestone at Stansvik, which contained an abundance of soft minerals, notably bole, mountain leather, disintegrated feldspar and iron-bearing clay.

From the standpoint of the present investigation, the description of the same limestone cavity given by the Finnish researcher Holmberg (1857, p. 5) is even more interesting. Holmberg writes: »Bei der Sprengen in einer Grube wurde ziemlich grosse Kalkhöhle angetroffen, die manche interessante Mineralier theils lose, theils in einer an ihrem Boden gebildeten sandsteinartigen Masse aufzuweisen. Zu diesen gehören: Quartz, roth, kristallisiert in einfachen Rhoemboädern (selten). Amethist-Kristalle. Kalkspath, schön kristallisiert auf Braunkalk und mit rothem Talk überzogen. Asbest und Bergleder.»

Numerous mineral specimens have been sent from Stansvik to the museums and mineralogical collections of many universities and corresponding institutions. Sample No. 1609 in the mineralogical collections of the University of Moscow was studied by Kazakowa, who mentioned (Kazakowa, 1911, p. 679) that the site where it was picked up was precisely the limestone cavity described by Sokolowski. The location in question is thus the Stansvik mine. Kazakowa gives the following description of the sample she had studied:

The mineral is easy to cut with a knife. Pure fragments shine like silk and are soft and flexible. The structure is mixed fiber. Ground to a fine powder the mineral dissolves in a strong concentration of sulphuric acid.

The chemical composition of the sample analyzed by Kazakowa is shown in Table 1.

Table 1. Chemical composition of palygorskite from Stansvik. Analyst, A. B. Kazakowa (Kazakowa, 1911)

	No. 1609 %
SiO_2	54.85
Al_2O_3	12.62
Fe_2O_3	tr.
FeO	—
MnO	—
MgO	9.48
CaO	—
H_2O+	11.73
H_2O-	9.30
	100.00

Kazakowa's view was that the fibrous mineral from Stansvik is β-palygorskite (see Fersmann, 1908, p. 270). Its chemical composition was given by Kazakowa according to the following formula: $H_{20}Mg_2Al_2Si_7O_{20}$.

In Kazakowa's analysis attention is arrested by the high Al_2O_3 quantity. In this respect, it deviates markedly from the fibrous material investigated by the present author, which, it would seem, derives from the same »limestone cavity» at Stansvik as Kazakowa's sample No. 1609.

The idea of re-investigating the Stansvik asbestos was proposed by Prof. A. Laitakari. He reported to this author in 1952 that the schoolboys Ilkka Laitakari and Paunu Oivanen had found good asbestos in the old mine of Stansvik and brought back samples. The same year, I went to investigate the mine for the first time; and on that same excursion I collected the main part of the material on which the present study is based. The fibrous mineral that I found there has now been ascertained to a high degree of certainty to be sepiolite or Meerschaum. No mineral possessing the chemical composition of palygorskite has yet been met with at Stansvik.

Previously sepiolite had been found in ancient Finnish rocks only once. The site of the find is Salmistonmäki, in the parish of Tammela (Holmberg, 1858; Nordenskiöld, 1863). See Fig. 1. According to Holmberg, the sepiolite of Tammela originated as a result of alteration of augite rock. The sepiolite of Tammela was analyzed by Fersmann. The result of the analysis is presented in Table 8, p. 34.

The discoverer of the Tammela sepiolite was evidently the ore prospector E. J. Westling, on whose report on his travels during the years 1824—39 (unpublished) Holmberg's description (1858, p. 145) of this first sepiolite find in Finland was based.

REVIEW OF EARLIER SEPIOLITE INVESTIGATIONS

It is necessary to consider first some sepiolite occurrences abroad and to review earlier sepiolite investigations before entering into a detailed examination of the sepiolite met with at Stansvik.

In Asia Minor, in Anatolia, near the town of Eskishehr, there is situated the best known and commercially most valuable of sepiolite deposits. According to the descriptions in geological literature, sepiolite is met with there mainly in secondary occurrences as nodules of varying size, which contain, in addition to sepiolite, fragments of serpentine, magnesite, and hornblende rock (e. g., Ladoo and Myers, 1951, p. 311). Serpentine-sepiolite rock seldom found in its primary site in Anatolia (Doelter, 1914, p. 382).

Serpentine rocks seem very often to occur as the parent rock of sepiolite. Such an observation has been made, e. g., in Croatia, at Agramer Gebirge, where Tucán (1915, p. 73) has found narrow veins containing fibrous mineral exclusively in serpentine rocks. The mineral is not, however, serpentine asbestos, as might be supposed, but fibrous sepiolite. This fact Tucán has established by, among other means, chemical analyses. The sepiolite investigated by Tucán corresponds in its chemical composition to parasepiolite.

There are sepiolite occurrences that deviate from the foregoing, as in cases where it is met with in veins together with quartz, sometimes opal. Examples of this kind of occurrence are to be met with in Grant County, New Mexico. Sterret (1908, p. 469) describes such an occurrence as follows: »The meerschaum occurs in veins, lenses, seams, and balls in the limestone . . . The veins are filled with chert, quartz, calcite, clay and meerschaum.»

The sepiolite found in serpentine rock is explained to have originated as an alteration product of serpentine. Magnesite is also likely to change into sepiolite, though certain investigators, like Weinschenk, regard it as less probable (see Doelter, 1914, p. 382). Impure opal with a high content of magnesium also belongs among the minerals likely to produce sepiolite through alteration. This view has been presented by e. g. Ladoo and Myers (1951, p. 311).

In studies dealing with the mineral, little attention has been paid to its origin compared to the voluminous exchange of opinions provoked by the occurrence of sepiolite sometimes as an amorphous, sometimes as a crystalline mineral. The amazingly large variations in the water content of the mineral arrived at in chemical analyses have also excited the interest of investigators. The aforementioned occurrence in Anatolia, Asia Minor, has been considered, especially in the past, a typical representative of amorphous sepiolite. Similar sepiolite is reported to have been met with elsewhere, as, for instance, in Greece and Bosnia. Detailed investigations have, however, proved that, e. g., the Anatolian sepiolite is not altogether amorphous material, there are also distinctly crystalline parts (Doelter, 1914, p. 374). Microscopic examination has revealed that the crystalline samples consist of extremely thin fibers.

The aforesaid observation as to the simultaneous occurrence of amorphous and crystalline sepiolite in the same deposit was first made by Vernadsky as early as 1901. According to this researcher (Vernadsky, 1901), it is possible to distinguish two species of sepiolite: crystalline and amorphous. Vernadsky designated the crystalline variety as α-sepiolite and the amorphous variety as β-sepiolite. In regard to chemical composition, the α-sepiolite, according to him, is more hydrous, containing four molecules of water, while the chemical formula for β-sepiolite includes only two molecules of water. The chemical constituents of the two varieties of sepiolite might thus be represented as follows:

$$\alpha\text{-sepiolite: } 2MgO \cdot 3SiO_2 \cdot 4H_2O$$
$$\beta\text{-sepiolite: } 2MgO \cdot 3SiO_2 \cdot 2H_2O$$

Fersmann (1908) has the same view of the chemical composition of α-sepiolite and β-sepiolite as Vernadsky. He calls the variety with a higher H_2O content (α-sepiolite) parasepiolite. Also according to Fersmann, a crystalline structure is characteristic of parasepiolite. Upon being heated to a temperature of 100° C, it loses one-half of its water content. If, after heating, the temperature again falls, parasepiolite absorbs water from the air.

The capacity of sepiolites to absorb water has been the object of study by, e. g., Zambonini (1908). According to his investigations, both varities of sepiolite readily absorb moisture from the air, and in this respect there appears to be no difference between α-sepiolite and β-sepiolite.

Doelter (1914, p. 379) points to observations made by him in microscopically investigating a number of sepiolite varieties regarded as amorphous and notes that, in his view, they are not pure gels but that between gelatinous and crystalline sepiolites there exist states of gradual transformation, which evidently signifies a slow alteration of gelatinous sepiolites into crystalline sepiolites.

Investigations carried out by colloid-chemical means to differentiate gelatinous and crystalline sepiolites have proved to be very unreliable. As examples might be cited the studies by Michel (1913) and Tucán (1915), of which only the final results are presented in the following. Michel took numerous samples from different sepiolite occurrences, ground the samples into a fine powder and examined the staining of amorphous and crystalline parts and observed their reaction to different stains. According to Michel's observations (1913, p. 165), there were two substances mixed up in the sepiolite occurrences represented in his material, of which the crystalline part gave an oxyphilic reaction and the amorphous part a basophilic reaction.

Tucán, for his part, offers a precise report on his corresponding staining tests, carried out with the sepiolite of Agramer Gebirge. His conclusion is that his material is distinctly basophilic but nevertheless crystalline (Tucán, 1915, p. 76).

The somewhat uncertain observations made by Doelter and Michel in regard to the mixed structure (partly amorphous, partly crystalline) of some sepiolites would none the less seem to fit in with the results obtained in regard to the variations in the H_2O content of the mineral. Mention has already been made of the fact that according to certain researches, there are varieties of sepiolite containing only two molecules of water, even though the great majority of analyses result in a higher content of H_2O. Kauffman (1943) has described this circumstance, pointing to the sepiolite analyses collected by Schaller (1936). This material comprises altogether 56 analyses, of which, according to Schaller, 34 represent α-sepiolite, 15 β-sepiolite and 7 an intermediate variety. The following explanation of Schaller is very interesting (Kauffman, 1943, p. 515): »It was, however, subsequently found that the specimens with the low water content had been heated varying degrees, some as high as 200° C, before analysis. These analyses were recalculated and all but four agreed with the alpha form of four molecules of water. This seems to indicate that there is actually only one type of sepiolite and it contains four molecules of water.»

Unfortunately, Schaller and Kauffman have not explained exactly how the recalculations were carried out. It would likewise have been interesting to know what the sepiolite was like represented by the four analyses that, in spite of all efforts at correction, had to be left outside the parasepiolite concept. Observation regarding the treatment of the material before analysis is wholly justified. Even a modest rise in temperature suffices to lower the water content of sepiolite. Special attention will be devoted in the present study to this matter.

GEOLOGICAL LOCATION OF THE STANSVIK SEPIOLITE DEPOSIT

The island of Degerö (Laajasalo), where the Stansvik sepiolite occurrence is situated, belongs to an area of ancient rocks of early Archaean age. Along

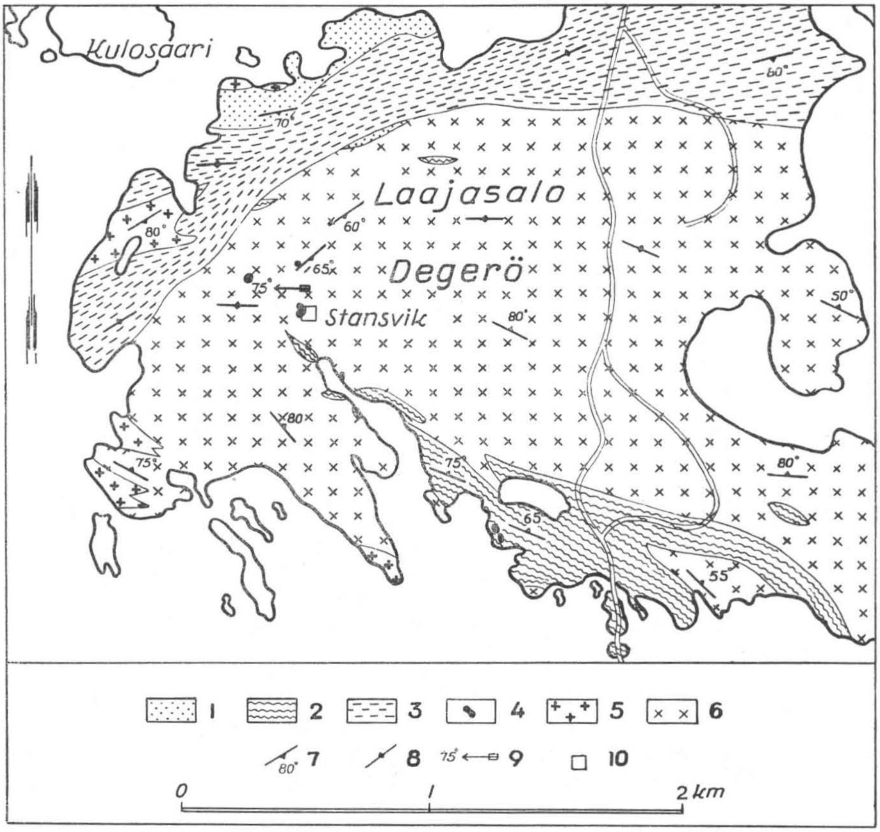

Fig. 2. Geological map of the Degerö (Laajasalo) area according to Seitsaari (1943). 1. Leptite. 2. Cordierite gneiss. 3. Amphibolite. 4. Skarn rock. 5. Microcline granite. 6. Quartz diorite. 7. Strike and dip of schistosity. 8. Vertical schistosity. 9. Folding axis. 10. The old iron mine of Stansvik.

the northern side of the island (Fig. 2) there runs a zone of leptite and amphibolite in an east-westerly direction, and along the southern side a parallel cordierite-gneiss zone. Between the two schist zones there is situated a gneissose quartz diorite, in which here and there can be seen lenses of cordierite gneiss as xenoliths. The microcline granite occurring to a slight extent on the W and S shores of the island of Degerö is clearly younger than the leptite, amphibolite and quartz diorite.

Iron ore deposits are very often associated with the Svecofennides of southern Finland. Such deposits have also been met with on the island of Degerö. They are associated with the skarn formations occurring in the W and S parts of the island.

No sepiolite is visible on the surface. The only occurrence so far discovered is situated inside the Stansvik mine, about 6 m below the present ground surface. At this depth there is a fissure vein in the wall of the mine in which there still remains an aggregate of fissure filling about 16 cm thick containing in thin stripes dense, brown steatite[1], quartz and a mineral that subsequent investigation proved to be sepiolite.

Under the fissure vein there is more steatite (Fig. 3), though of a coarser variety than between the sepiolite stripes, and there appears in it clearly recognizable, even though steatized, skarn minerals, diopside and andradite garnet (Fig. 4). Altered skarn is thus in question.

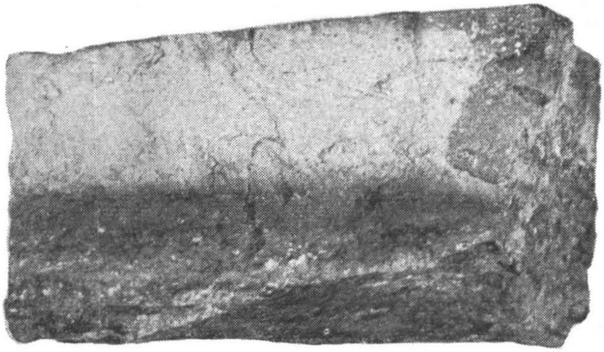

Fig. 3. Steatite at the lower side of the fissure vein. Natural size. Photo A. Matisto.

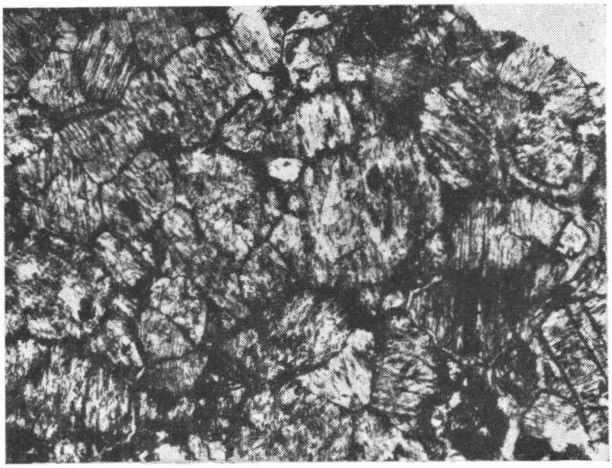

Fig. 4. Steatitized pyroxene rock. Nic. //. Magn. 20 ×. Photo A. Matisto.

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[1] The present author has designated as steatite a dense, soapstone-like alteration product of pyroxene-andradite skarn rocks, the chemical composition of which closely resembles that of talc or pyrallolite poor in CaO. The steatite of Stansvik differs from talc in respect to both structure and its higher content of H_2O. Terminalogically the definition of steatite given by Hess (1933) has been followed most nearly: »Steatitization may be defined as that process of hydrothermal alteration of an ultrabasic which in its final stages results in the formation of talcose rock.»

Above the sepiolite-quarz fissure vein there is first a thin sheet of skarn rock and overlying it calcitic limestone, white and very clean. The fissure vein is thus situated in a skarn zone in limestone.

GENERAL DESCRIPTION OF THE SEPIOLITE FISSURE VEIN

The sepiolite-quartz veins form a fissure vein of varying thickness, the dip of which is 30° E. This fissure vein can be followed in the wall of the mine for a distance of about 2 m, after which it narrows and finally terminates in a surface of slickensiding. Figs. 5 and 6 are from the upper part of this fissure vein. Between the sepiolite veins there is soapy, dense, chocolate-brown steatite.

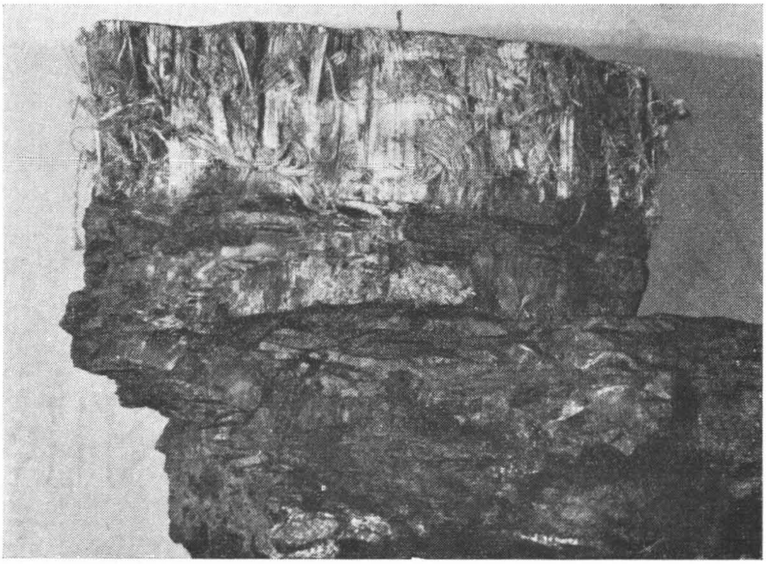

Fig. 5. Cross-fiber sepiolite and quartz veins from the upper part of the fissure vein. Natural size. Photo A. Matisto.

All the sepiolite veins are not identical in structure. The most general type is so-called *cross-fiber sepiolite*, in which the sepiolite fibers lie at right angles to the wall of the vein, with the length of the fibers being most often equal to the width of the vein. In the middle of certain cross-fiber sepiolite veins there is a thinner or broader band of a pulpy, gleaming white mineral. The present author has termed this mineral *mass-fiber sepiolite.* A third sepiolite type is *mountain leather* or *slip-fiber sepiolite,* which is met with in contacts between the fissure vein and the country rock. In mountain leather the sepiolite fibers lie parallel to the fissure.

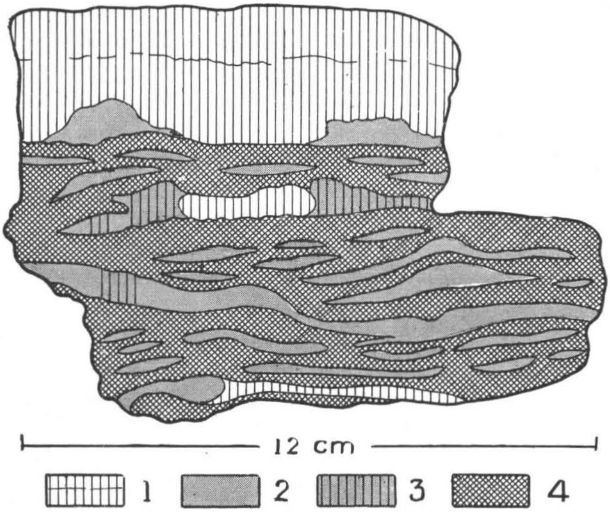

Fig. 6. Sketch of the upper part of the fissure vein. 1. Cross-fiber sepiolite. 2. Quartz. 3. Quartz with fibrous habit. 4. Steatite.

Cross-fiber sepiolite. Most of the sepiolite at Stansvik is of the cross-fiber type. At its best this sepiolite type occurs in the upper part of the fissure vein, where the greatest measured length of the fibers has been 28 mm. The shortest fibers are less than one mm in length. In color cross-fiber sepiolite is greyish white, with a silky shine. The fibers are exceedingly fine, flexiable and remarkably tough. After heating to incandescence the color is a glowing white, but the fibers are brittle.

Sepiolite fibers very seldom extend unbrokenly across the vein from edge to edge. More generally, there is a distinct seam running through the middle of the vein, and usually an abundance of pigmented impurities have collected in it. Several seams of this kind can be observed in the same vein, which indicates that the fissure has continually opened and shut. In Fig. 5 the seam is seen only as a thin line in the middle of the vein, whereas in Fig. 7 the middle seam is much as 2 mm thick.

Mass-fiber sepiolite. Mass-fiber sepiolite has been met with chiefly in the middle seam of cross-fiber sepiolite veins, but also as veins or small nodules in steatite fissures. In studying mass-fiber sepiolite under the microscope, it can with certainty be observed to be anisotropic and formen felt-like fibers. Pure, dry mass-fiber sepiolite floats for some time on water.

Slip-fiber sepiolite = mountain leather. This sepiolite type appears to have derived from cross-fiber sepiolite through the effect of tectonic movements. Between the fibers running parallel with the fissure there is some steatite and sometimes pellet-like grains of black manganese-bearing hematite.

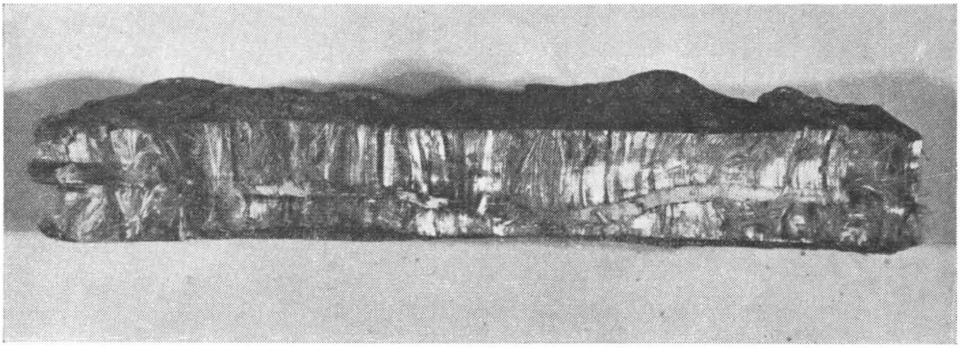

Fig. 7. Cross-fiber sepiolite vein. In the middle seam of the vein there occurs mass-fiber sepiolite. On the left side of the picture mass-fiber sepiolite has been replaced by quartz. Natural size. Photo A. Matisto.

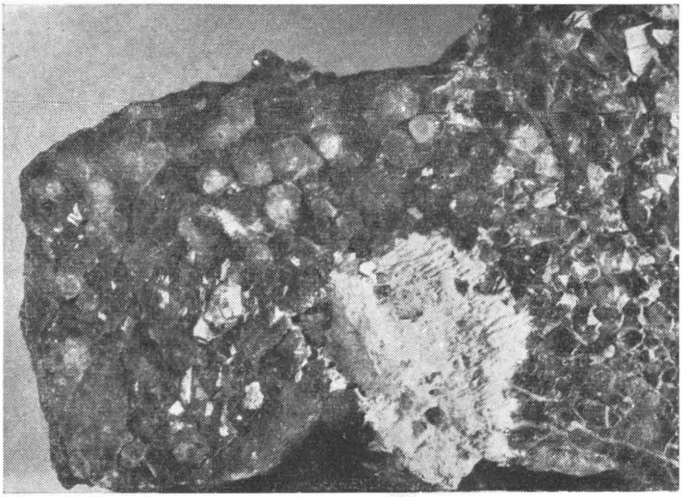

Fig. 8. Sepiolite-quartz vein viewed from on top. Among the quartz crystals amethyst occurs. In the middle of the picture there is white sepiolite, remaining from the replaced sepiolite vein. Natural size. Photo A. Matisto.

Quartz veins. The quartz veins mostly run parallel with the sepiolite veins, but in addition there are quartz veins that are younger than these and penetrate them. The very youngest quartz generation comprises peculiarly shaped quartz crystals, which occur in the way shown in Fig. 8 in fissures that have opened up between the sepiolite veins. The two youngest generations of quartz differ very often from the quartz veins of earlier origin also in color.

A certain geologically interesting phenomenon is connected with the development of the fissure vein. The mass-fiber and cross-fiber sepiolites

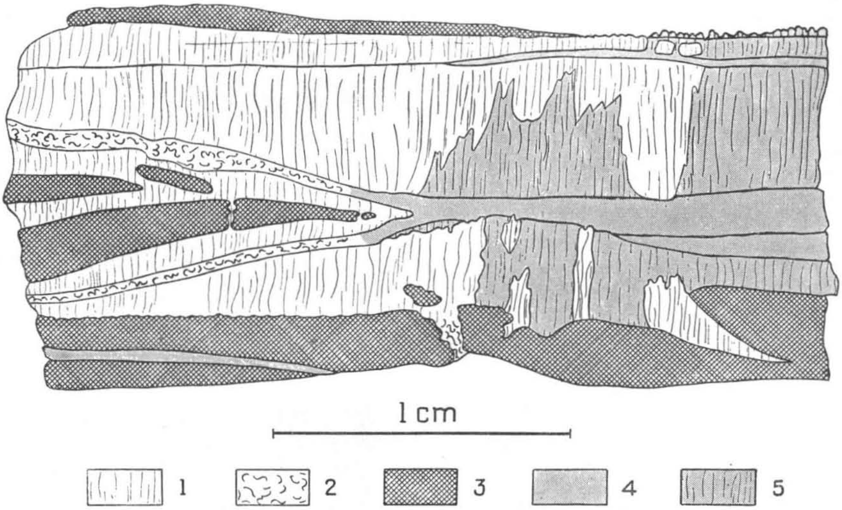

Fig. 9. Replacement of the sepiolite vein by quartz. 1. Cross-fiber sepiolite. 2. Mass-fiber sepiolite. 3. Steatite. 4. Quartz. 5. Quartz pseudomorphs after cross-fiber sepiolite.

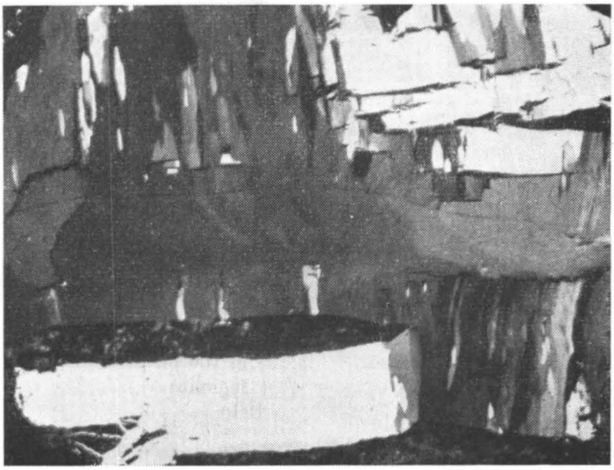

Fig. 10. Sepiolite veins replaced by quartz. The quartz is fibrous in habit. Nic. +. Magn. 20 ×. Photo A. Matisto.

have partly altered into quartz, evidently under the influence of silicic acid solutions coming from outside. This phenomenon is easy to observe under the microscope (Fig. 9). The silicic acid solutions have traveled to their present situation either along open fissures or then along sepiolite veins. In the latter case, the first phase is a thin seam of quartz in the contact of the sepiolite vein. Gradually quartz crystals begin to penetrate deeper and deeper into the center of the sepiolite vein, until the sepiolite vein has been totally silicified. The crystal habit of sepiolite is still visible in the shape of the quartz crystals. An evident replacement is thus involved. Quartz veins originating by replacement are shown also in Fig. 10.

Quartz appears to replace mass-fiber sepiolite more easily than the cross-fiber variety. Also this circumstance is illuminated by Fig. 7, in which mass-fiber sepiolite replaced by quartz is to be seen in the middle of the cross-fiber sepiolite, while the latter shows no sign of replacement. Apparently silicic acid solutions flow more easily through the felt-like mass-fiber sepiolite, with the result that this sepiolite type is sooner subject to a replacement process.

Steatite. In association with the sepiolite and quartz veins there is a chocolate-brown, dense, soapy stone, which the present author has referred to as steatite. It is not a pure talc; this is proved by the *d*-values obtained by X-ray examinations; only one of them is close to talc, but the majority are foreign to talc. There are numerous indications, however, that the said steatite is an alteration product principally of two skarn minerals, diopside and andradite garnet. The alteration product is a talcose, hydrous magnesium silicate. It does not belong to the sepiolite group on account of its low content of H_2O.

The steatite in between the sepiolite-quartz veins is very dense in structure. In examining a thin section of it under the microscope, scales 1/40 mm —1/50 mm long may be noted in places within an indefinite mass. These scales are arranged partly perpendicularly and partly radially in relation to each other, with the result that the rock has to some extent a net-like texture. The elongation of these miniature scales is positive (like antigorite). In addition to the foregoing there are some larger scales in steatite whose elongation is negative. Talc flakes are rare, but some have nevertheless been observed with certainty.

The steatite sitiated below the fissure vein is more coarse-grained than the former and less altered. This steatite area has portions in which altered idiomorphic crystals of diopside can plainly be observed. In the same type of steatite rock there is practically unaltered garnet, the species of which could be determined by the X-ray method. Andradite is in question. The pyroxene, despite its well-preserved outward appearance, has wholly changed into a soft, white mass. In addition to diopside and garnet some extremely small grains of carbonate are also met with.

Table 2. Chemical analyses of steatite from Stansvik

Anal. No.	SiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	H_2O+	H_2O-	Total
1522	56.12	0.52	6.45	3.20	0.08	27.14	5.16	0.52	99.79
1523	56.66	0.70	6.62	3.05	0.11	27.09	5.62	0.48	100.33

1522. Dense steatite from the side of sepiolite vein. Analyst, P. Ojanperä.
1523. Brown, more coarse-grained steatite under the fissure vein. Analyst, P. Ojanperä.

The steatites derived from diopside and andradite were investigated by the X-ray method, taking a specimen of separated »diopside» and another of a steatite that, to judge the form of the pseudomorphs, appeared to have derived wholly from andradite. The *d*-values obtained were in both case practically the same, as the following table shows:

»Diopside» *d*	»Andradite» *d*
9.8 (m)	9.8 (m)
9.61 (w)	—
3.13 (m)	3.16 (m)
2.50 (w)	2.495 (w)

(m = medium; w = weak)

In this connection, attention should be called also to the chemical analyses (Table 2). The dense steatite taken from between the sepiolite veins is chemically of the same kind of hydrous magnesium silicate as the coarser and less altered steatite under the fissure vein. Minerals of originally very different composition have metamorphosed into steatite of the same composition. The brown color of the steatite results mainly from the high iron content occurring as an impurity.

OPTICAL PROPERTIES OF THE SEPIOLITE

The optical properties of the sepiolites of Stansvik have been investigated both by the immersion method and from thin sections. The cross-fiber, mass-fiber and mountain leather sepiolites have proved to be distinctly anisotropic. No amorphous sepiolite has been met with. All the sepiolite varieties met with at Stansvik have an extinction parallel to the length of the fibers. The refractive indices for the cross-fiber sepiolite are: $\alpha = 1.491$, $\gamma = 1.509$. The refractive indices for the mass-fiber sepiolite are higher (Table 3).

In geological literature numerous different values have been given for the refractive indices of sepiolite. The following refractive indices are from Kauffman's study (Kauffman, 1943). To Table 3 have been added the corresponding optical determinations for the Stansvik sepiolite.

The fibrous sepiolite presented first in Table 3 is, according to Kauffman, a typical parasepiolite. Its refractive indices ($\alpha = 1.490$, $\gamma = 1.505$) are very close to the values that Larsen and Berman (1934) as well as Caillere (1933) obtained for fibrous sepiolite. Caillere regards the sepiolite studied by him as parasepiolite, reporting its chemical composition in the same way as Kauffman ($2MgO \cdot 3SiO_2 \cdot 4H_2O$).

Table 3. Optical properties of sepiolite

Specimen	Color	Extinction	Elongation	α	γ	Optic sign	2 V
1	white	parallel	positive	1.490	1.505	neg.	40°—60°
2a	white	parallel	positive	1.498	1.506	neg.	50°
2b	white	parallel	positive	1.519	1.526	neg.	—
3	white	parallel	positive	1.506	1.526	neg.	—
4	white	parallel	positive	1.499		—	—
5	gray	parallel	positive	1.531		—	—
6	white	parallel	positive	1.491	1.509	—	—
7	white	parallel	positive	1.508	1.524	—	—

1. Hydrous magnesium silicate, Yavapai Co., Arizona (Kauffman, 1943).
2a. Fibrous sepiolite (Larsen and Berman, 1934).
2b. Sepiolite (Larsen and Berman, 1934).
3. Fibrous sepiolite (Daly, 1935).
4. Fibrous sepiolite (Caillere, 1933).
5. Fibrous sepiolite (Bradley, 1930).
6. Fibrous sepiolite (cross-fiber), Stansvik.
7. Fibrous sepiolite (mass-fiber), Stansvik.

The cross-fiber sepiolite of Stansvik clearly belongs to the parasepiolite group on the basis of its refractive indices. On the other hand, the mass-fiber variety differs considerably in its refractive index values as compared with Kauffman's determinations or the cross-fiber sepiolite of Stansvik. In Table 3, however, similar refractive index values are presented for fibrous sepiolite, so that the mass-fiber sepiolite of Stansvik does not appear in this respect to be a notable exception.

According to Winchell (1947) the specific gravity of sepiolite is 2. It is nevertheless evident that the specific gravity varies somewhat, depending on the chemical composition of the mineral. The determination of the specific gravity of the cross-fiber sepiolite of Stansvik was made from a sample kept for a period of about one year at room temperature (averaging + 21° C), with a specific gravity of 2.15—2.17 being obtained.

CHEMICAL EXAMINATION

Analytic results. The simultaneous occurrence in the same fissure vein of two in appearance quite different types of sepiolite led to a suspicion that these minerals might differ also in their chemical composition. This assumption appeared to be supported by the view held in geological literature that the water content of sepiolite is controlled by the structure of this mineral (see p. 18). The different optical properties of the cross-fiber and mass-fiber sepiolite of Stansvik also gave food for thought.

For the investigation of the chemical composition of the sepiolites, analyses were made of all three sepiolite types of Stansvik, noting especially the content of water and variations in it. Before analysis the samples were kept for about one year at room temperature (21° C). The amount of water that was released during 16 hours at 110° C was designated as H_2O—.

Table 4. Chemical analyses of sepiolite

	No. 1525			No. 1509			No. 1716			No. 1524		
	%	Mol. prop.	Mol. %	%	Mol. prop.	Mol. %	%	Mol. prop.	Mol. %	%	Mol. prop.	Mol. %
SiO_2	54.82	9 128	35.29	55.47	9 236	35.58	55.22	9 194	35.82	52.80	8 791	35.72
TiO_2	0.00	—	—	0.00	—	—	0.00	—	—	0.00	—	—
Al_2O_3	0.10	10	0.04	0.24	24	0.09	0.24	24	0.09	0.50	49	0.19
Fe_2O_3	0.85	(53)	—	0.83	(52)	—	1.39	(87)	—	5.44	(341)	—
FeO	0.07	10	0.04	0.11	15	0.06	0.21	29	0.11	0.43	60	0.24
MnO	0.60	85	0.33	0.67	95	0.37	0.54	76	0.30	0.42	59	0.24
MgO	24.28	6 022	23.28	23.25	5 766	22.21	23.62	5 858	22.82	22.29	5 528	22.46
CaO	0.00	—	—	0.00	—	—	0.00	—	—	0.00	—	—
P_2O_5	0.00	—	—	0.00	—	—	0.00	—	—	0.00	—	—
CO_2	0.00	—	—	0.00	—	—	0.00	—	—	0.00	—	—
H_2O+	9.18	5 096	19.70	9.48	5 262	20.27	9.72	5 395	21.02	9.44	5 240	21.29
H_2O-	9.94	5 517	21.33	10.02	5 562	21.42	9.17	5 090	19.83	8.80	4 885	19.85
	99.84	25 868	100.01	100.07	25 960	100.00	100.11	25 666	99.99	100.12	24 609	99.99

1525. Cross-fiber sepiolite (separated sample), Stansvik. Analyst, P. Ojanperä.
1509. » » (general sample), » » » »
1716. Mass-fiber sepiolite (general sample), » » » »
1524. Slip-fiber sepiolite (= mountain leather), » » » »

The results of the analyses are presented in Table 4. With the exception of Analysis No. 1525, made from cross-fiber sepiolite separated under the microscope, all the other analyses represent general samples from corresponding sepiolite types. The difference in chemical composition between separated and unseparated cross-fiber sepiolite is not great but nevertheless noticeable (the samples are from precisely the same place in the vein). The SiO_2 content in the separated sample is lower and approaches the theoretical value of SiO_2 in parasepiolite, while, at the same time, the MgO content increases and practically achieves the theoretical MgO value of parasepiolite, from which it deviates only 0.06 %. The higher SiO_2 content of the general sample results from the penetration of the quartz between the fibers and also from the replacement phenomenon described in the foregoing (p. 24).

All four sepiolite analyses correspond very closely. The higher iron content of slip-fiber sepiolite is a consequence of the hematite occurring in mountain leather as an impurity. The occurrence of pellet-like grains of ore in slip-fiber sepiolite was observed already in the microscopic examination. The abundance of Fe_2O_3 in the analysis correspondingly lowers the figures for SiO_2 and MgO.

Table 5 shows the theoretical weight and molecule percentages of parasepiolite, together with the corresponding values for the sepiolite of Stansvik. Kauffman (1943) points out in his report that, according to Surfleet and Porter, a true sepiolite should have a gravimetric ratio of magnesia to silica equivalent to 1 : 2.24, the allowable limits being between 2.21—2.28.

Table 5. Theoretical chemical composition of parasepiolite and composition of Stansvik sepiolite. (Only the three chief components of the Stansvik sepiolite have been taken into account)

	No. 1		No. 1525		No. 1716		No. 1524	
	%	Mol. %	%	Mol. %	%	Mol. %	%	Mol. %
SiO_2 ...	54.13	33.34	54.82	35.29	55.22	35.82	52.80	35.72
MgO ...	24.22	22.22	24.28	23.28	23.62	22.82	22.29	22.46
$H_2O\pm$ ..	21.65	44.44	19.12	41.03	18.89	40.85	18.24	41.14
	100.00	100.00	98.22	99.60	97.73	99.49	93.33	99.32

1. Theoretical parasepiolite = $2MgO \cdot 3SiO_2 \cdot 4H_2O$.
1525. Cross-fiber sepiolite, Stansvik.
1716. Mass-fiber sepiolite, Stansvik.
1524. Slip-fiber sepiolite, Stansvik.

Calculating this ratio from Table 5 for the theoretical parasepiolite with three decimals, is $MgO/SiO_2 = 1 : 2.235$. In the purest cross-fiber sepiolite of Stansvik the MgO/SiO_2 weight ratio is $1 : 2.257$, which corresponds wholly to the requirements set for parasepiolite.

In comparison with theoretical parasepiolite, the sepiolite of Stansvik is poorer in H_2O. The lack of water is most conspicuous in the slip-fiber sepiolite, in which the total content of water falls 3.41 % below the theoretical value. From the standpoint of the present sepiolite investigation, it is interesting to note that no such difference in bulk composition or water content can be detected between mass-fiber and cross-fiber sepiolite that on this basis these types could be deemed different varieties. On the contrary, the analyses convincingly show that two identical minerals are in questions.

Since, however, both the optical determinations and the X-ray investigations reveal, as will subsequently be seen, that some kind of difference does exist between mass-fiber and cross-fiber sepiolite, it was considered necessary to investigate the dehydration of both types at low temperatures. These studies give at the same time a good picture not only of the dehydration of sepiolites but also of their astonishing capacity to absorb water. In the following is a description of these studies.

Dehydration and absorption of water. The test was carried out with both mass-fiber and cross-fiber sepiolite. A small amount of each sepiolite was weighed, 421 mg of the former and 710.2 mg of the latter. The samples were not pulverized. Then the samples to be studied were placed into a platinum crucible at room temperature (+ 21° C), after which the crucibles were placed into a oven, with a minimum temperature of +30° C. After three hours the samples were taken out of the oven, were cooled in an desiccator (on top of silica gel) for 20 minutes and then weighed. The loss in weight was calculated in percentages. Thus the treatment of sepiolite samples was continued 10° C intervals up to a temperature of 190° C. At it was not

Table 6. Loss in weight of mass- and cross-fiber sepiolite by heating from 21° C to 190° C

Temperature in C	Mass-fiber %	Cross-fiber %	Temperature in C	Mass-fiber %	Cross-fiber %
21°	—	—	120°	9.19	9.46
30°	1.43	1.38	130°	9.19	9.52
40°	3.37	3.15	140°	9.33	9.57
50°	5.23	4.84	150°	9.38	9.66
60°	7.55	7.07	160°	9.41	9.70
70°	8.46	8.22	170°	9.48	9.77
80°	8.74	8.79	180°	9.57	9.83
90°	8.98	9.04	190°	9.64	9.94
100°	9.07	9.24	Total H_2O ..	18.89	19.15
110°	9.17	9.41			
H_2O-	9.17	9.41	H_2O+	9.72	9.74

possible to achieve higher temperatures with the oven, the remaining water content was removed by ignition. The transfer of the platinum crucibles to the desiccator and from the desiccator to the scales was carried out as fast as possible, for sepiolite absorbs moisture from the air in very little time. The results of the test are seen in Table 6.

The dehydration of sepiolites between the temperatures of 21° C and 190° C is represented also by the accompanying diagram (Fig. 11). The dehydration of sepiolites is exceedingly strong between the temperatures of 21° C and 70° C, slowing down thereafter upon the temperature's rising close to 100° C. At temperatures of 110° C—130° C, dehydration practically ceases. Upon a further rise in temperature, small amounts of water are lost up to 190° C, the highest point that could be reached with the oven at my disposal. The remaining water (H_2O+) is lost only by ignition.

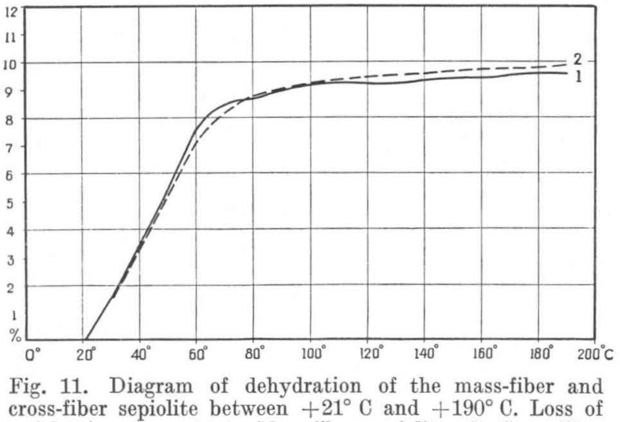

Fig. 11. Diagram of dehydration of the mass-fiber and cross-fiber sepiolite between +21° C and +190° C. Loss of weight in per cent. 1. Mass-fiber sepiolite. 2. Cross-fiber sepiolite.

In general 110° C is regarded in chemical analyses as the limit in determining H_2O+ and H_2O-. In the tests performed the amount of water lost from the mass-fiber sepiolite at 110° C is 9.17 % or 48.5 % of the total content of water, and from the cross-fiber sepiolite at 110° C 9.41 % or 47.7 % of the total content of water (within the same time interval). The slight difference to be noted in the dehydration curves of the sepiolites (Fig. 11) probably is due to the difference in the mechanical structure of mass-fiber and cross-fiber sepiolites. Felt-like sepiolite loses and correspondingly also absorbs water (see Table 7) more readily than denser sepiolite consisting of longer fibers.

The test series described in the foregoing proved that these minerals undergo strong dehydration already at a very low temperature. As a result of this observation, an important question from the standpoint of the present study may now be made: What is the lowest temperature at which dehydration takes place? The initial temperature of +21° C used in the test surely is not the lowest.

In order to be able to answer this interesting and difficult question in some degree at least, one would be obliged to return to Table 6 and the diagrams in Fig. 11. At a temperature of 30° C, maintained for a period of three hours, the weight of each sepiolite sample diminished an average of 1.4 %, or, calculated in relation to the amount of water removed from the sepiolites at temperatures under 110° C, more than 15 %.

In Fig. 11 both curves between the temperatures of 21° C and 60° C are practically linear. Although, for lack of observations, dehydration curves have not been continued below a temperature of 21° C, it may nevertheless be assumed that no radical changes in the direction of the curves would take place immediately. If the amount of water considered as dehydrating between +8° C and +21° C is calculated by extrapolation, 2 %—2.5 % should be obtained. This percentage reveals fairly accurately the amount of water lacking in the sepiolite analyses as compared with the theoretical water content of α-sepiolite (see Table 5). Evidently the water content of sepiolites achieves its maximum value only at +4° C, where the density of water is highest. Upon studying sepiolite samples at ordinary room temperature, dehydration has evidently already taken place in them.

The following test of the absorption capacity of sepiolites was made: When the sepiolite samples had been heated to 190° C and immediately

Table 7. Absorbtion of water

	18 h	24 h	42 h
Mass-fiber sepiolite	77.3 %	84.2 %	86.0 %
Cross-fiber sepiolite	41.9 %	48.2 %	54.5 %

thereafter weighed, they were allowed to remain in an open platinum crucible in the room with the scales. The crucibles were now weighed after 18, 24 and 42 hours. Table 7 contains the results obtained in this test. The percentages reveal the amount of water absorbed as compared with the amount lost from the samples at 190° C.

Both sepiolite types thus absorb water from the air. The absorption process is swifter, however, in the case of mass-fiber sepiolite. At the end of the test only 14 % of the primary amount of water (at +21° C) was lacking from the mass-fiber sepiolite, whereas the corresponding deficiency in the case of the cross-fiber type was still 45.5 %. Obviously, both types of sepiolite would have absorbed more water upon continuation of the test and thereby gradually begun to revert to the water content prevailing in the samples before heating.

When sepiolite fibers were heated to a red glow and thereafter their absorption capacity was studied, it was ascertained that they no longer absorbed water. The samples had burned »to death» and the sepiolite fibers had become very fragile. Upon the crystal water's having been lost, the structure of the sepiolite had thus changed.

On the basis of the tests carried out, two kinds of water may be distinguished in sepiolite:

1. Constitution water, consisting of one-half the whole amount of water in sepiolite. The constitution water is totally lost only after the mineral has been heated to a high temperature, whereupon the lattice of the sepiolite

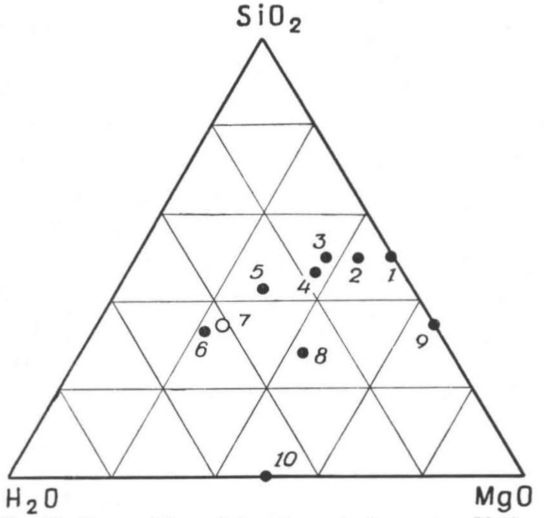

Fig. 12. Composition of the phases in the system MgO—SiO_2—H_2O. 1. Enstatite. 2. Anthophyllite. 3. Talc. 4. Steatite from Stansvik. 5. β-sepiolite (theoretical composition). 6. Parasepiolite (theoretical composition). 7. Fibrous sepiolite from Stansvik. 8. Serpentine. 9. Forsterite. 10. Brucite.

at the same time breaks down. Upon the constitution water's having been lost, the sepiolite no longer absorbs water.

2. Pore water (H_2O below 110° C) or »zeolitically» bound water. This water is also essential in the water content of sepiolites, although it is not constitution water proper. Sepiolite may contain varying amounts of pore water, depending on the conditions prevailing during analysis, but the total amount of pore water is nevertheless limited. Sepiolite may lose and re-absorb pore water without damage to the crystal structure of the mineral being caused.

Triangular diagram. The chemical composition of the Stansvik sepiolites and steatite is also presented by means of a triangular diagram (Fig. 12). In this diagram have been marked the places of all the minerals originating in the system $MgO—SiO_2—H_2O$. The sepiolites of Stansvik are situated in the diagram, as might be expected, near the theoretical point of parasepiolite. The diagram also makes evident that the SiO_2/MgO ratio in the Stansvik sepiolite and in the theoretical parasepiolite is approximately the same, but a slight difference appears in the H_2O content. The Stansvik sepiolite is situated much closer the α-sepiolite than the β-sepiolite. Calculating the chemical formula[1] of the mineral according to the purest sepiolite of Stansvik, the following form is obtained: $2MgO \cdot 3SiO_2 \cdot 3.4H_2O$. The formula shows, as is natural, a deficiency of water in comparison with theoretical parasepiolite.

The triangular diagram also reveals the relation between the chemical composition of pure talc and that of Stansvik steatite. As mentioned in the foregoing (p. 25) the steatite of Stansvik is not to be regarded as a talc, but as a talc-like hydrous magnesium silicate. This fact is also made clear by the diagram. The steatite of Stansvik is richer in water than talc, but the SiO_2/MgO ratio is about the same in both.

Other sepiolite analyses. In conclusion a comparison should be made between the chemical composition of the Stansvik sepiolite and certain sepiolite analyses published in the literature.

Table 8 includes ten sepiolite analyses, four made in Finland and six abroad. It will be noted that the analyses correspond closely. Especially interesting is the observation that the total H_2O content is of the same magnitude in all the analyses, varying within the narrow limits between 18.24 % and 19.60 %. If the analysed samples had contained both gelatinous and crystalline parts in arbitrary proportion, the likelihood is that there would have been greater variations in the total amount of water.

The difference between the theoretical total water content of parasepiolite and that obtained in the analyses is due mainly to dehydration

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[1] For the formula the molecule percentages were calculated in such a way that half the Al_2O_3 molecule per cent was added to SiO_2 and the other half to MgO. To MgO have also been added the FeO and MnO molecule percentages occurring to a slight extent in the analyses.

Table 8. Chemical analyses of sepiolites

	1	2	3	4	5	6	7	8	9	10
SiO_2	54.83	53.10	52.86	54.37	55.34	54.74	55.47	54.82	55.22	52.80
Al_2O_3	0.28	0.32	2.51	—	1.81	1.80	0.24	0.10	0.24	0.50
Fe_2O_3	0.45	1.56	—	0.98	0.43	—	0.83	0.85	1.39	5.44
FeO	—	—	—	—	—	1.03	0.11	0.07	0.21	0.43
MnO	—	—	0.59	—	—	tr.	0.67	0.60	0.54	0.42
MgO	24.51	24.87	22.91	24.13	22.95	22.68	23.25	24.28	23.62	22.29
CaO	0.55	1.27	0.80	0.59	0.24	tr.	0.00	0.00	0.00	0.00
Na_2O	0.35	—	—	—	—	—	n.d.	n.d.	n.d.	n.d.
K_2O	0.03	—	—	—	—	—	n.d.	n.d.	n.d.	n.d.
H_2O+	10.74	10.11	8.49	10.19	10.20	6.83	9.48	9.18	9.72	9.44
H_2O-	8.18	8.42	10.71	9.41	8.60	12.72	10.02	9.94	9.17	8.80
	99.92	99.65	100.00	99.67	99.57	100.00	100.07	99.84	100.11	100.12

1. Fibrous sepiolite, Yavapai Co., Arizona (Kauffman, 1943).
2. Fibrous sepiolite, Madagascar (Caillere, 1933).
3. Parasepiolite, Tempelstein, Moravia. Analyst, F. Kovař (Doelter, 1914).
4. Sepiolite, Coulommiers (Caillere, 1933).
5. Sepiolite, Durango, Mexico (Foshag, 1928).
6. Parasepiolite, Tammela, Finland. Analyst, A. Fersmann (Doelter, 1914).
7. Fibrous sepiolite (parasepiolite), Stansvik, Finland. Analyst, P. Ojanperä.
8. Fibrous sepiolite (parasepiolite), Stansvik, Finland. Analyst, P. Ojanperä.
9. Fibrous sepiolite, Stansvik, Finland. Analyst, P. Ojanperä.
10. Fibrous sepiolite (mountain leather), Stansvik, Finland. Analyst, P. Ojanperä.

taking place in sepiolites already at very low temperatures. The deficiency in relation to the theoretical value is 2 %—3 %. This fits in well with what was said earlier regarding the pore water of sepiolites and its permanence under the conditions of analysis.

X-RAY EXAMINATION

The X-ray pattern of the Stansvik sepiolite was made by the powder method, using CuK_α radiation. The interplanar spacing values are given in Table 9. The corresponding determinations published by Kauffman (1943) and Longhambon (1937) are included in the same table.

Both the said investigators have, according to the reports, had fibrous sepiolite as research material, which is a relevant matter in comparing the respective *d*-spacings. Longhambon has noted (see Kauffman, 1943, p. 517) that the first three lines are characteristic of the fibrous sepiolite variety, namely: 12.2 (±0.15) Å, 7.5 (±0.10) Å and 6.7 (±0.05) Å. In the X-ray pattern of the cross-fiber sepiolite of Stansvik (Table 9, No. 3), all three of these lines are observed, but in that of the mass-fiber sepiolite (No. 4) only the first, strongest line (12.09 Å), although in other respects the X-ray pattern of this mineral is wholly comparable to that of other sepiolites. The lack of the second and third line from the X-ray pattern of the mass-fiber sepiolite of Stansvik would seem to indicate the less developed stage

Table 9. Interplanar spacings of sepiolite

1		2		3		4	
I	*d*	I	*d*	I	*d*	I	*d*
s	12.3		12.2	s	12.09	s	12.09
w	7.52		7.6	w	7.49		—
w	6.75		6.7	w	6.79		—
w	5.02		5.05		—	w	5.03
w	4.51		4.50	m	4.52	m	4.52
s	4.29		4.30	s	4.30	s	4.30
m	3.97		—		—		—
m	3.75		3.83	m	3.75	m	3.75
w	3.53		3.60		—		—
m	3.35		3.40	m	3.34	m	3.35
m	3.18		3.22	m	3.18	m	3.18
w	3.04		—		—	w	3.05
w	2.82		—		—	w	2.81
w	2.67		—		—		—
m	2.61		2.61	m	2.61	m	2.61
s	2.55		—	m	2.56	s	2.55
m	2.44		2.45	w	2.43	w	2.43
w	2.38		—	w	2.38	w	2.40
m	2.25		2.28	w	2.23	m	2.25
w	2.18		—		—	w	2.16
w	2.11		2.09		—		—
w	2.05		—		—	w	2.05
w	1.95		1.97		—		—
w	1.86		1.88		—	w	1.87

Intensity: s = strong, m = medium, w = weak.

1. Fibrous sepiolite, Yavapai Co., Arizona (Kauffman, 1943).
2. Sepiolite, D'Ampandrandava (Lonchambon, 1937).
3. Fibrous sepiolite (parasepiolite), Stansvik, Finland.
4. Fibrous sepiolite (mass-fiber sepiolite), Stansvik, Finland.

of crystallization, perhaps, of this sepiolite type, even though gelatinous parts have not been noted in it.

The results given by the X-ray examination support the concept arrived at previously that the cross-fiber sepiolite of Stansvik belongs — together, possibly, with mass-fiber sepiolite — expressly to the group of parasepiolites.

ORIGIN OF THE SEPIOLITE-QUARTZ FISSURE VEIN

Fissures are created in the rock crust particularly in connection with tectonic movements. From the standpoint of the origin of ores and other commercially valuable mineral deposits, fissure formations are of considerable significance, inasmuch as solutions have used the fissures as transportation channels (see Schneiderhöhn, 1941).

Into the fissure of the Stansvik limestone there have been introduced hot silicic acid solutions containing magnesium. Under the influence of these solutions the skarn rock has been metasomatically steatitized. In examining samples under the microscope, the diopside can be observed to begin to undergo change before the andradite.

According to laboratory tests carried out by Bowen and Tuttle (1949), it may be assumed that steatitization takes place at temperatures exceeding 350° C, whereupon at a certain pressure talc originates from the system $MgO—SiO_2—H_2O$. When in the tests the temperature was lowered to 350° C or below and the pressure of water vapor was 1 055 kg/cm^2, there was created in place of talc a product with the optical and roentgenographic properties of sepiolite. Talc and sepiolite may, accordingly, derive from the same solution and originate under very much the same conditions. The production of talc presupposes, however, a higher temperature.

After the steatitization of the skarn rocks of Stansvik, the first sepiolite veins were apparently formed directly from the solutions, whose temperature had by then fallen to 350° C or below. No evidence whatsoever was obtained in the studies that the sepiolite might be an alteration product of some previously formed mineral. Upon the origin of the first sepiolite veins or immediately thereafter, movements took place along the fissure, as a consequence of which the sepiolite veins were deformed and pressed together into slip-fiber sepiolite or mountain leather. The effect of tectonic movements is also to be observed in the structure of the steatite situated immediately on both sides of the contact. The original structure of the steatite has totally been destroyed and the skarn altered into a dense, soapy mass.

After the formation of the mountain leather, a new phase begins in the development of the fissure vein, characterized by a continuous opening up, dilatation, of the fissures. Ever new fissures of varying breadth open up side by side in the steatitized zone, filling up thereafter in turn. Cross-fiber sepiolite now starts to crystallize out of the solution. The fibers grow from the edges of the veins toward the center, as proved by the seam observable in the center of the veins. Often seams that have grown together have opened anew and the sepiolite continued its crystallization process. Thereby more than one seam is likely to have developed in cross-fiber sepiolites.

The third sepiolite generation consists of mass-fiber veins and small sepiolite grains. It is hard to understand why at this stage there should have formed sepiolite of mass-fiber structure and not the same type as in the preceding stage. Chemical analyses do not give an answer to the question. Decreasing temperature may, however, play some role in the process.

The final stages in the filling up of the fissures begin with the formation of pure quartz veins. The quartz veins may be divided into three generations. The oldest veins are parallel to the sepiolite veins, while the younger ones deviate from the general direction and penetrate the quartz veins of an earlier stage. The formation of quartz veins terminates in idiomorphic quartz crystals which are met with in the fissures opening up between the sepiolite veins.

The sepiolite veins are quite generally silicified. This process took place during the stage of formation of the first and second quartz generations. As the quartz replaced the sepiolite, quartz pseudomorphs formed after sepiolite.

The filling up of the Stansvik fissure took place under hydrothermal conditions. Upon the temperature's steadily falling, first sepiolite and then quartz veins successively formed. The progress of events might briefly be illustrated as follows:

A. Stage of Mg-metasomatism, temperature over 350° C
Steatitization of the skarn rocks

B. Hydrotermal stage, temperature under 350° C
  I. Sepiolite veins
    1. Slip-fiber sepiolite
    2. Cross-fiber sepiolite
    3. Mass-fiber sepiolite
  II. Quartz veins
    1. Parallel quartz veins
    2. Intersecting quartz veins
    3. Idiomorphic quartz crystals

The history of the evolution of the Stansvik fissure vein proves convincingly that fibrous sepiolite forms directly by crystallizing out of solutions. Such a manner of origin for sepiolite appears to be rare in nature, considering that previously it has not attracted appreciable attention; at least, in geological literature sepiolites are quite generally described as alteration products of magnesian minerals.

SUMMARY

In the altered limestone at Stansvik there is a tectonic fissure, in which one meets with fibrous sepiolite and quartz. On the basis of its mode of occurrence and appearance, the sepiolite of Stansvik may be divided into three groups: slip-fiber, cross-fiber and mass-fiber types, listed in order of age. In chemical composition all the sepiolite types considered in the present study are the same variety of fibrous sepiolite, the chemical formula of which may be written as follows:

$$2MgO \cdot 3SiO_2 \cdot 4H_2O$$

The sepiolite of Stansvik has derived from hydrothermal solutions at a temperature evidently under 350° C. No evidence has come to light to indicate that this mineral might be an alteration product of earlier magnesian minerals.

The sepiolite veins are in many cases silicified. It could be observed that a distinct replacement of sepiolite fibers is involved.

The sepiolite of Stansvik contains both constitution water and pore water. The pore water belongs as a fundamental part to the water content of sepiolite, but it varies in amount, depending on the temperature. The quantity of pore water attains its theoretical value at apparently a temperature of +4° C, when it equals the content of constitution water. The studies support the view already held by Schaller (1936) that in reality there exists only one type of sepiolite and that it contains four molecules of water.

The possibilities of commercially exploiting the Stansvik fibrous mineral deposit have been investigated by quarrying on an experimental basis. These tests have proved that the fissure vein containing sepiolite fibers is small in size and lacking in economic value.

ACKNOWLEDGMENTS

During the period of my investigation I received valuable assistance from several of my colleagues, to all of whom I hereby express my gratitude.

The chemical analyses as well as the water content determinations of the sepiolites were carried out by Mr. Pentti Ojanperä, M. A., at the Chemical Laboratory of the Geological Survey of Finland. The X-ray examinations were performed by Messrs. K. J. Neuvonen, Ph. D., and Kai Hytönen, M. A., at the Institute of Geology of the University of Helsinki. The refractive indices of the sepiolites were determined by Mrs. Toini Mikkola, M. A.

I wish to extend thanks to Professor Aarne Laitakari, head of the Geological Survey, for proposing the subject of this study. Mr. Paul Sjöblom, M.A., deserves thanks for his conscientious English translation.

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4

SOME NEW DATA ON THE BLUE-GREEN HORNBLENDE FROM THE TAMPERE SCHIST BELT [1]

BY

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ABSTRACT

The manner of crystallization and composition of a zoned blue-green hornblende are described and discussed. Some DTA data for the blue-green hornblende from the Tampere area and for »common green» hornblende are presented.

CONTENTS

A ZONED BLUE-GREEN HORNBLENDE FROM ORIVESI

The present author has earlier (Seitsaari, 1951, 1953, 1954b) described blue-green hornblendes occurring in the Tampere schists and discussed the physical conditions under which they have crystallized. These hornblendes have a comparatively high Fe_2O_3/FeO ratio and usually contain more water than hornblendes in general.

Recently, a blue-green hornblende with more or less zoned crystals was found in hornblende schist at Hieta, about 1 km S.W. of the Orivesi railway station. Many crystals have almost colourless or faintly greenish kernels, which sometimes show rather distinct boundaries, but for the most part occur as indefinite spots (Figs. 1 and 2). The two different parts of the crystals are homoaxial and the mode of occurrence of the kernels is similar

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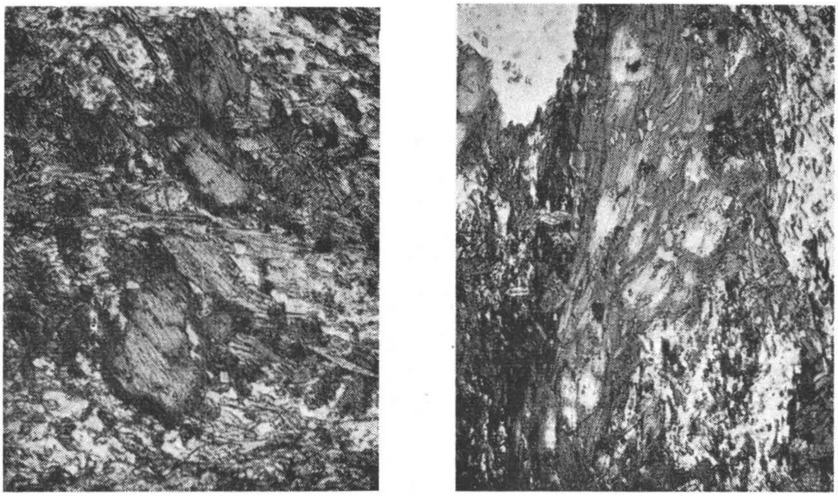
[1] Received April 24, 1956.



Figs. 1 and 2. Mode of occurrence of the light hornblende in the blue-green hornblende. Nic. //. Magn. 32 ×.

to that of cummingtonite in hornblende. The extinction angle $\gamma \wedge c$ of both parts is 18° ± 1°. The axial angles $2V_\alpha$ are: 66° ± 2° (blue-green hornblende, five determinations); 78° ± 2° (light hornblende, four determinations). All the determinations were made on the U-stage. The refractive indices are: $\alpha = 1.657$, $\gamma = 1.679$ (pure blue-green hornblende); $\alpha = 1.636$, $\gamma = 1.659$ (light hornblende, the minimum figures). The determinations were made by the immersion method on the U-stage. An attempt was made to separate the two types of hornblende by means of the Clerici solution in a centrifuge. Pure blue-green hornblende was obtained within a range of specific gravity from 3.19 to 3.22. Within a range from 3.10 to 3.19, amphibole double grains and transitional forms between the two types were mainly obtained, but the amount of the lightest variety was not enough for a chemical analysis. 1.7 g of the fraction was within a range of specific gravity from 3.10 to 3.15, and this part of the fraction was used for an analysis. The chemical composition of the two fractions is presented in Table 1.

The blue-green hornblende is very similar to another hornblende described earlier (Seitsaari, 1953; see Table 2, Anal. 5), with the exception of a lower H_2O content. The composition of the lighter fraction differs from the former by a lower content of Al_2O_3, Fe_2O_3, FeO, and H_2O, and by a higher content of SiO_2 and MgO. Because the amount of the lightest homogeneous kernels in the said fraction was rather small, the differences in the above points between the kernel and the blue-green hornblende will be much greater. Both analyses show practically the same CaO content. It is probable that the kernels are not far from tremolite-actinolite.

Table 1. Chemical composition of the zoned hornblende from Hieta, Orivesi

	1		2	
	%	Mol. prop.	%	Mol. prop.
SiO_2	45.70	7 609	49.36	8 218
TiO_2	0.51	64	0.40	50
Al_2O_3	10.46	1 026	8.45	829
Fe_2O_3	5.56	348	3.83	240
FeO	9.26	1 289	8.27	1 151
MnO	0.38	54	0.35	49
MgO	12.77	3 167	14.08	3 492
CaO	12.20	2 175	12.10	2 158
Na_2O	1.38	223	1.05	169
K_2O	0.09	10	0.24	25
H_2O+	1.89	1 049	1.62	899
H_2O-	0.03	—	0.06	—
	100.23	—	99.81	—
$ox \left(= \frac{2\,Fe_2O_3}{2\,Fe_2O_3 + FeO}\right)$	0.35		0.29	

1. Blue-green hornblende. Analyst, P. Ojanperä.
2. Lighter fraction of the hornblende. Analyst, P. Ojanperä.

Thus a calcian amphibole rich in Fe_2O_3 and Al_2O_3 has crystallized around another calcian amphibole that is conspicuously poorer in these constituents. It is known that, in the metamorphic recrystallization of hornblende, the higher the temperature the richer is the resulting hornblende in Al and ferric iron. In the present instance, a rise in temperature thus seems to have happened during the crystallization, because the shell of the amphibole double grains can hardly be older than the kernel. The author has earlier (Seitsaari, 1953, 1954b) pointed out that the blue-green hornblende crystallized under the conditions of the amphibolite facies (in the sense of Turner). Evidently it holds true of the case under consideration. It was suggested, too, that the blue-green hornblende crystallized under a low shearing stress *(op. cit.)*. There is evidence of a similar stage of metamorphism from another rock in another part of the Tampere area (Seitsaari, 1954a). The mineral parageneses concerned and their genesis were discussed in more detail in the said papers. It is possible that in considerable parts of the Tampere schist belt a late, probably slight, rise in temperature has occurred, as proved by the metamorphic evolution of certain fine-grained schists with minerals sensitive to variation in the physical conditions. During the increase of temperature the shearing stress seems to have been low. With this stage the regional metamorphism of the area was fulfilled.

DTA DATA FOR SOME BLUE-GREEN AND GREEN HORNBLENDES

Because a high degree of oxidation of the iron in hornblende is usually accompanied by a low water content, it was supposed (Seitsaari, 1953, p. 92)

Table 2. Chemical composition of some blue-green hornblendes from the Tampere schist belt (Seitsaari, 1953, pp. 88—9)

	2		3		5	
	%	Mol. prop.	%	Mol. prop.	%	Mol. prop.
SiO_2	42.00	6 993	40.12	6 680	45.28	7 539
TiO_2	0.48	60	0.80	100	0.46	58
Al_2O_3	12.72	1 248	18.68	1 832	9.20	902
Fe_2O_3	7.27	455	5.43	340	5.85	366
FeO	16.02	2 230	13.20	1 837	9.65	1 343
MnO	0.08	11	0.29	41	0.24	34
MgO	7.34	1 820	6.36	1 577	12.93	3 207
CaO	10.29	1 835	11.10	1 979	12.82	2 286
Na_2O	1.13	182	1.29	208	0.64	103
K_2O	0.59	63	0.58	62	0.42	45
H_2O+	2.40	1 332	2.32	1 287	2.68	1 487
H_2O-	0.00	—	0.24	—	0.00	—
	100.32	—	100.41	—	100.17	—
$ox \left(= \frac{2\,Fe_2O_3}{2\,Fe_2O_3 + FeO}\right)$	0.29		0.27		0.35	

Column numbers 2, 3 and 5 refer to Tables 2, 3 and 5, respectively, in the said paper.

that in the blue-green hornblendes hydroxyl groups may to some degree substitute for oxygen in the silicon-oxygen chains with a contemporaneous removal of a corresponding amount of Si. An example of such a substitution in another mineral group is the $3CaO \cdot Al_2O_3 \cdot 3SiO_2$—$3CaO \cdot Al_2O_3 \cdot 6H_2O$ solid solution series to which the minerals hydrogrossular (Hutton, 1943), plazolite (Pabst, 1937), and hibschite (Pabst, 1942) belong. Machatschki (personal communication, 1954) does not accept this idea, but

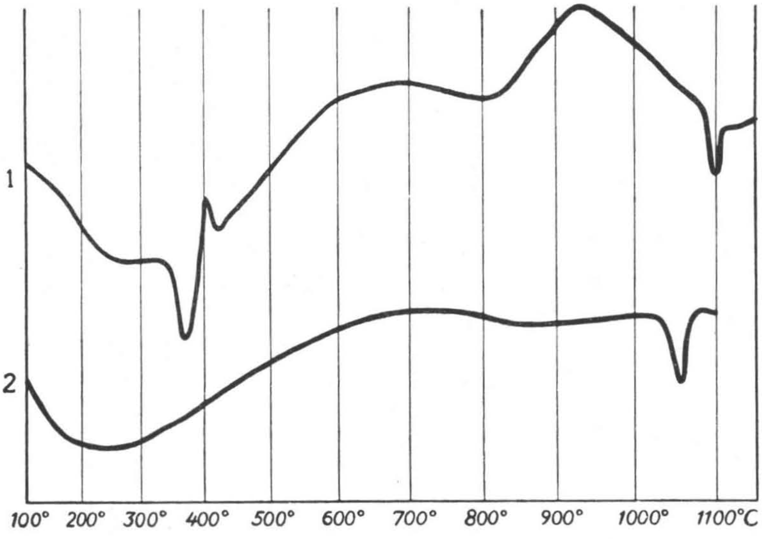


Fig. 3. (1) DTA curve for Hornblende 2, Table 2. (2) Type DTA curve for the other blue-green hornblendes and some »common green» hornblendes.

suggests that H_2O molecules may have been introduced to the vacant positions in the lattice.

An attempt was made to solve by means of differential thermal analysis whether there is water in different positions in the blue-green hornblende. Three hornblendes previously described by the author (Seitsaari, 1953; the analyses are quoted in Table 2) and the hornblende described in the present paper (Table 1, Anal. 1) were investigated. For comparison, differential thermal analyses were made of four »common green» hornblendes with an Al_2O_3 content between 6.5 % and 13 %, and with a moderate degree of oxidation of the iron (ox = 0.16—0.20) and a moderate H_2O content (1.7 %, on an average). The DTAs were carried out by Dr. Bengt Collini, Uppsala. A heating rate of about 10° C/min. was used.

It was found that almost all curves of both the blue-green and the »common green» hornblendes were practically identical. There was one distinct exception only, namely, Hornblende 2 (Table 2). The DTA curve for this hornblende is reproduced in Fig. 3, Curve 1. Curve 2 in Fig. 3 will represent a type curve for all the other hornblendes, because each deviation from it is small.

In all the curves, endothermic peaks were visible between 1 030° C and 1 150° C. These peaks obviously mean a structural disintegration of the hornblendes, at which monoclinic pyroxene, hematite, and some plagioclase are formed and water is released (Wittels, 1952).

In Curve 1, however, another endothermic peak appears at 370° C. The hornblende in question (Table 2, Anal. 2) is rich in total iron and in Fe_2O_3. If there are abundant trivalent ions in the hornblende structure, the hydroxyl groups may be partly replaced by oxygen (Warren, 1930). Thus a high degree of oxidation of the iron is usually accompanied by a low water content and, consequently, the higher the total Fe content, the greater might be the decrease of the water content, when the degree of oxidation of the iron is high. In Hornblende 2 the water content, however, exceeds the average water content of »common green» hornblendes, and thus some »extraordinary» water may be present in it. The endothermic peak at 370° C presumably means the removal of that part of water. To the author's knowledge, observations of low driving-off temperatures of water like this do not exist in respect to the amphibole group. The hydroxyls representing the ordinary combined water are not broken-down until by continued heating at about 800° C (Barnes, 1930). The low driving-off temperature of a part of the water seems to indicate the presence of H_2O molecules in this hornblende. This would agree with Machatschki's opinion (personal communication, 1954) of the character of the excessive water.

However, the foregoing case is just an individual one. It is explicable that hornblendes with a high MgO content, like Hornblende 5 (Table 2), may have a higher hydroxyl content (i. e. a higher content of »ordinary» water) than the Fe-rich types, although the figure for *ox* were high. But neither did Hornblende 3 (Table 2), which is rather similar to Hornblende 2, give any endothermic peak at low temperatures. For additional criteria, a great number of water-rich blue-green hornblendes must be investigated.

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5

A COMPARISON OF THE MINOR BASE METAL CONTENTS OF SOME FINNISH GALENAS [1]

BY

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ABSTRACT

The galenas from the occurrences of older Svecofennian rock zones and from those within the rapakivi areas in Finland have been compared by ore-microscopic and spectrographic methods. Subsequently, based mainly on compositional differences in minor base metal contents, two distinct groups of galenas have been distinguished and their genetical relations discussed.

CONTENTS

INTRODUCTION

When investigating the galena-bearing ores that exist within the rapakivi areas, the present author has repeatedly observed the striking purity of these galenas as regards sulphosalts of bismuth, antimony or silver, which on the other hand almost without any exception are associated with the galenas in the ores existing within the zones of older rock formations in southern Finland. This behavior has suggested a closer study of the matter, especially because the said differences, if typical enough, might be used as criteria when resolving the origin of galena-bearing boulders met with during prospecting operations going on in southern Finland.

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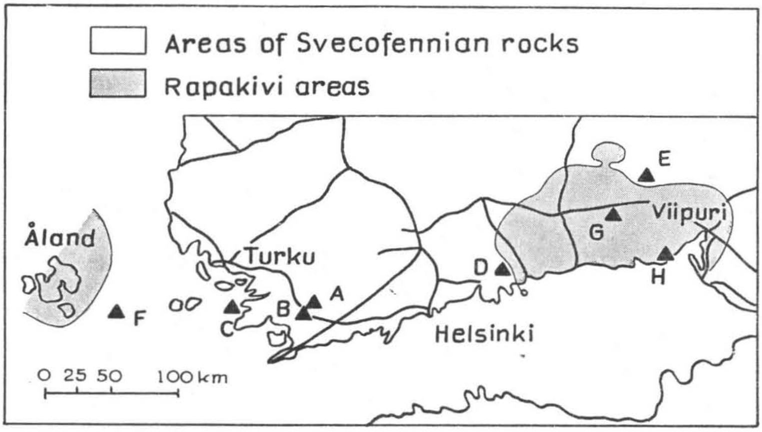

Fig. 1. Locations of the specimens studied. The area marked as Svecofennian in age includes smaller portions of younger pre-Cambrian rocks, e. g. sandstones and diabases.

With this purpose in mind, galenas from eight Finnish localities, shown in Fig. 1, have been studied by ore-microscopic and spectrographic methods, while their unit-cell dimensions have been controlled by X-ray diffraction patterns.

THE MATERIAL AND THE CHARACTER OF THE OCCURRENCES STUDIED

In collecting the material, attention was paid to finding specimens rich in galena in each locality studied. A study of the material selected under the ore-microscope revealed in order of abundance the ore minerals as explained in the following list, which also includes brief information on the occurrences concerned.

A. Orijärvi Mine, Kisko. Level +215, in the neighbourhood of the shaft. Polished section No. 4189[1]. Sphalerite, galena, tetrahedrite associated with galena, Bi-sulphosalt (probably cosalite), chalcopyrite and pyrrhotite.

The area containing the ore deposits forms the central area for the famous study on the mineral facies classification and Fe-Mg-metasomatism by Eskola (1914). Up-to-date facts on the ores mined are given in the report by Varma (1954a). The ore-bearing zones in the Orijärvi region exist in extensive formations of supracrustal schists, which have undergone strong metamorphic and metasomatic changes. The main ore mineral constituents include pyrite, pyrrhotite, chalcopyrite, sphalerite and galena. Apparently galena is very scarce and occurs mainly in association with soft ores, i. e. in the zones of tremolite-actinolite skarns. In addition to the main ore minerals, there occur some tetrahedrite and occasionally other sulphosalts associated with galena.

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[1] The number refers to the collection of the Geological Survey of Finland.

B. Aijala Mine, Kisko. Eastern drift, level +88. Polished section No. 2874. Galena, pyrite, chalcopyrite and some tetrahedrite as well as small grains of Sb-sulphosalts (probably meneghinite and jamesonite) associated with galena.

The area is about midway between Helsinki and Turku in southern Finland. The deposits of Aijala as well as those of the nearby Metsämonttu mines are situated in the same formation of metamorphosed supracrustal schists that enclose the Orijärvi deposits. The mineral composition of the Aijala ore is as follows: Pyrite, pyrrhotite, chalcopyrite, sphalerite, arsenopyrite and galena. In addition there occur different kinds of Bi-, Sb-sulphosalts, sometimes as separate grains, but mainly as inclusions or exsolution products in galena. The sulphosalts in especial are relatively abundant in the old quarries close to the Aijala and Metsämonttu mines (Kaitaro and Vaasjoki, 1950). The following analytical figures illuminate the main metal contents of the ore of Aijala (Varma, 1954b): 1.8 % Cu, 0.6 % Zn, 0.02 % Pb, 0.001 % Bi, 0.5—1.0 g/ton. Au, 13—16 g/ton. Ag.

C. Attu, Parainen. Polished section No. 3916. Galena, pyrrhotite accompanied by some marcasite, chalcopyrite and pyrite. Some small spots of tetrahedrite occur in galena. Obviously the presence of argentite in galena, as noted earlier by Pehrman (1931), is more or less occasional because the mineral was not found again in the course of the present study.

The ore mineralizations in Attu are associated with tremolite-bearing quartzose rocks belonging to the series of supracrustal schists analogous to those in the Orijärvi region. According to Pehrman (1931), the ore mineral composition is as follows: Pyrite, pyrrhotite, sphalerite, galena and chalcopyrite. In places galena occurs as coarse grains obviously having been the latest mineral to crystallize.

D. Koskenkylä, Pernaja. Polished section No. 2235. Arsenopyrite, sphalerite, galena, chalcopyrite, pyrite and pyrrhotite. Very few and small grains of tetrahedrite associated with galena.

Veins of silver-bearing galena and other sulphides have been known since the 17th century in the area. These veins occasionally may contain a number of different ore minerals like arsenopyrite, loellingite, pyrite, sphalerite, pyrrhotite and some tetrahedrite associated with galena (Vaasjoki, 1953). These small prospects, however, have not been sufficient for mining on a commercial scale.

E. Hyvärilä, Lemi. Polished section No. 3969. Sphalerite, galena, pyrrhotite and chalcopyrite. The galena shows no inclusions or exsolution products of sulphosalts.

The ore mineralizations are situated in amphibolitic schists as impregnations and separate lumps. The chief ore mineral is pyrrhotite. Besides, there occur sphalerite and smaller amounts of galena and chalcopyrite. An analysis of the ore shows the following metal contents: 3.74 % Zn, 1.10 % Pb, 0.06 % Cu, 72.4 g/ton. Ag. The occurrence is situated over a distance of more than half a kilometer from the contact of rapakivi and, as Pääkkönen (1947) has insisted, the emplacement of ore mineralizations into the enclosing schists seems to be older than the emplacement of rapakivi in the area.

F. Södö, Sottunga. Polished section No. 3972. The specimen is relatively pure galena accompanied by small amounts of chalcopyrite and pyrite. In places the covellite replaces the chalcopyrite. No sulphosalts have been observed either in the galena or as separate grains.

In Sottunga and in the neighbouring eastern archipelago of Ahvenanmaa there occur several small ore prospects, which so far have been without economic significance. In Södö the ore mineralizations, which mainly contain pyrite and pyrrhotite associated with some galena, exist as vein-like formations in surrounding granitic rocks. Feldspars and quartz occur as gangue minerals.

G. Luotola, Luumäki. Polished section No. 3612. The specimen is composed almost entirely of galena. As additional ore material very small amounts of replacement series chalcopyrite→chalcosite→ covellite were observed.

The whole area is occupied by typical rapakivi rocks belonging to the Viipuri massif. In these rocks the galena occurs, besides quartz and feldspars, as fillings in narrow fissures (Kulonpalo, 1947). The occurrences have no commercial value.

H. Säkkijärvi. Polished section No. 3615. The composition of this specimen is exactly the same as that from Luumäki (G). Here also the replacement chalcopyrite-chalcosite-covellite was identified. When etched 1 minute by HNO_3 (1 : 1) a clear twinning texture, apparently caused by post-mineral movements, was developed in the individual galena grains.

The most prominent galena-bearing veins within the rapakivi formations in Finland have been found at Lamminniemi and Muhunniemi in the parish of Säkkijärvi (Remmler, 1886). In addition to the galena some chalcopyrite and pyrite are usually associated with the vein material. The thickness of the veins varies from a few millimeters to about five centimeters. Veins very similar to those in Säkkijärvi have been met with in rapakivi also in the parishes of Virolahti and Inkeroinen (Kulonpalo, 1947). As a gangue mineral in all these veins there occurs mainly quartz, in places, as in Inkeroinen, followed by fluorite.

After the specimens were crushed, handpicked under the binocular microscope, and ground to pass 50 Din (= 0.125 mm), the final concentration of galena from each specimen was made by magnetic and gravitational methods. The effectiveness of purification was controlled by examination of polished sections made from the concentrates. The impurities, composed of some silicate constituents and other sulphides than galena, amounted to less than 1 % in the final concentrates used for analysis.

SPECTROGRAPHIC DETERMINATIONS

The contents of Bi, Sb, Ag, Cd, Zn and As were analyzed from each concentrate spectrographically with the result as seen in Table 1. According to the ore-microscopic determinations of the concentrates, the contents of Cd, Zn and As are due to impurities. E. g. the presence of zinc and cadmium in Concentrate B is caused by sphalerite, which was no longer separable. The presence of arsenic (D) is similarly caused by unseparable arsenopyrite.

According to the distribution of elements Ag, Sb, Bi and Sn the specimens analyzed form two distinctly different groups: The said elements are always present, sometimes in considerable amounts, in the specimens A, B, C, D and E, whereas in the specimens F, G and H the said elements practically taken are absent. The two last-mentioned specimens are from the typical occurrences of galena, which have been found in fissure fillings in the areas occupied by rapakivi. Obviously very similar in character is the specimen F, although the occurrence concerned has been found in fissures occurring in older, Svecofennian formations. The incorporation of ore-bearing solutions in these fissures, however, may have taken place later or by the emplacement of rapakivi, which now reaches the intimate periphery of the occurrences under consideration.

On the basis of the ore-microscopic study, the content of antimony is caused by the presence of Sb-sulphosalts, mainly by tetrahedrite, which was observed in all other specimens, except F, G and H. Bi-sulphosalts, probably cosalite, have been met with only in the specimens of Orijärvi. Silver minerals like argentite have not been observed in any specimen studied, although silver may follow the tetrahedrite in some degree. Accordingly, in all other instances, except the specimen from Orijärvi, bismuth and at least part of silver seems to enter the galena structure. That is a result reached earlier in several studies on lead ores and which has involved different explanations. Ramdohr (1938) has suggested that a mineral schapbachite (matildite) $AgBiS_2$, having a cubic cell at temperatures over 215° C enters as an unlimited solid solution into the galena structure; at temperatures lower than this, the mineral schapbachite, while then of orthorhombic symmetry, unmixes from the galena. It is, however, hard to believe that any schapbachite molecules exist in the galenas here under consideration, because the unit-cells of all the said galenas show no deviations from the normal size. Therefore, instead of assuming a high temperature α-schapbachite, one might accept Ramdohr's (1938) other proposal, which sets forth that silver and bismuth may enter the galena structure without forming independent species, if the amounts of the said elements are too small.

Tin seems to be an essential trace element in the galenas represented by the Concentrates A, B, C and E. To make sure that tin is not caused by small amounts of sphalerites, which possibly still exist as impurities in the aforementioned concentrates, the content of the element was checked in sphalerites from the Orijärvi, Aijala and Lemi occurrences, which all are spectrographically analyzed. In these cases the result was negative, i. e. the content of tin in the sphalerites was very slight and not comparable

Table 1. Results of spectrographic determinations

Concentrate	Locality	Element in ppm						
		Ag	Sb	Bi	Sn	Cd	As	Zn
A	Orijärvi Mine, Kisko	>1 000	60	4 000	50	+	+	—
B	Aijala Mine, Kisko	800	1 000	900	80	150	+	—[1]
C	Attu, Parainen	200	1 200	90	150	+	+	—
D	Koskenkylä, Pernaja	600	120	200	5	+	1 000	—
E	Hyvärilä, Lemi	1 000	60	high	high	+	+	—
F	Södö, Sottunga	50	+	4	+	+	+	—
G	Luotola, Luumäki	20	+	2	+	+	+	—
H	Lamminniemi, Säkkijärvi	15	60	2	+	+	+	—
			+>60	high = over 4 000	+ <3	+<100	+<100	—

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[1] Very low.

to the contents represented in Table 1, even if the Analysis D is considered. As pointed out by Oftedahl (1940), in ores of contact metamorphic high temperature formations tin seems to enter into galena, whereas in ores of hydrothermal low temperature formations it enters into the sphalerite structure. Subsequently, the tin as analyzed from the Concentrates A, B, C, E and even D probably indicates relatively high temperature conditions prevailing during the formation of corresponding galenas. According to the statements made earlier in this paper, that is well consistent with the geological characters of corresponding occurrences, which all, during their formation or later, have been subjected to strong metamorphic processes at different orogenic phases in the Archean time.

CONCLUDING REMARKS

1. The galenas from the sulphide ore deposits of the older rock zones in Finland usually seem to be accompanied by the elements Ag, Bi and Sn, which most probably are not caused by foreign minerals, but exist as trace elements in the galena structure.

2. The content of Sn in the galenas of the aforementioned occurrences refer to relatively high temperature conditions by the formation of the galena-bearing paragenesis of the ore.

3. The obviously very simple composition of the galenas in typical occurrences of the rapakivi areas and in Sottunga might refer to slowly cooling, in composition less complicated hydrothermal solutions, which have circulated under relatively static tectonic conditions. This is in good agreement with the field evidence, which reveals that the galena-bearing veins in both these areas most probably belong to post-tectonic fissure fillings, perhaps even younger than pre-Cambrian in age.

4. The replacement chalcopyrite → chalcosite → covellite is typical of the galena-bearing veins of the rapakivi areas and of Sottunga. If hypogene in origin this phenomenon also emphasizes the low hydrothermal properties of the mineralizing solutions that have formed the veins now in question.

Acknowledgments — The spectrographic determinations were made by Mr. Oiva Joensuu, M. A., in the Department of Geology, University of Chicago. The separation of galenas from the ore specimens was carried out by Mr. Lauri Hyvärinen, M. A., and Mr. Kalevi Kauranne, M. A., from the Geological Survey of Finland. To these gentlemen the author's sincere gratitude is due.

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6

ON CENTRAL COMPLEXES WITH RADIAL LAMPROPHYRIC DIKES [1]

BY

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ABSTRACT

Three areas with radial, prevalently lamprophyric dikes (Åva, Finland; Shap, England; and Spanish Peaks, U. S. A.) are briefly reviewed. Many common characteristics, e. g. an identical differentiation tendency, are found, apparently owing to the same kind of intrusion mechanics. Some structural problems are discussed in greater detail. The general formation of radial and ring fractures during the intrusion has been induced by sideward and upward pressure, respectively. At a slow intrusion the latter is less potent and hence it favours the formation of radial fractures. The special petrological character of the rocks, which form the complex sequence of the central and dike intrusions, is suggested to be a result of common petrogenetic processes.

CONTENTS

INTRODUCTION

It is always an important observation that the same geological structures occur in connection with rocks of similar petrological character, especially because it suggests common factors operating in different structural and petrological processes. It is purposed in this paper to compare three areas,

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[1] Received April 28, 1956.

where radial, prevalently lamprophyric dike systems associated with mainly granitic central stocks are known, viz. Åva (Åland Islands; Kaitaro, 1953), Shap (Lake District, Northern England; Harker and Marr, 1891, and Grantham, 1928) and Spanish Peaks (Colorado, U. S. A.; Knopf, 1936). These are, so far as is known, the only localities of this kind in the whole world. In spite of their quite different surroundings and ages (Åva, late pre-Cambrian; Shap, middle Palaeozoic; and Spanish Peaks, Tertiary), they have surprisingly much in common. This cannot be a mere chance, especially if we take into consideration the rather unusual character of these provinces. The author is interested in this problem because he has had an opportunity to make a detailed mapping in the Åva area, which has perhaps the best and most regular example of the radial lamprophyric dikes. For the purpose of making comparative studies, a two weeks' visit was made to the Shap area in May, 1954.

The main purpose of this study is to present the structural development of these complexes with special attention to the Åva area. This is because the other areas mentioned are less familiar to the author. Only in case some other of the areas offers in any respect a more intelligible common feature for solving some distinct problem, it will be used.

Because of the comparative nature of this study, a short summary of the general character of each area will be given to facilitate the personal judgment of the reader. The description of the Shap area is based, except on the studies mentioned, on the personal informations of Dr. Grantham and the field observations of the author.

The Åva area (Kaitaro, 1953). The composite central stock about 4—6 km in diameter has intruded into the pre-Cambrian rocks with a general vertical dip. It represents a complex sequence of separate intrusions formed by porphyritic biotite granite and monzonite. It is immediately surrounded by concentric granitic ring intrusions, which are best developed on the eastern side. Aplites and pegmatites are locally very common. Especially in the border zone and in all ring intrusions a subparallel arrangement of feldspar phenocrysts is frequent. The structure of the country rock follows in the map pattern almost completely the shape of the intrusion, but the contact is in many details discordant and granite dikes cut the country rock. Lamprophyres, up to 2 m in width, have a generally radial arrangement around the central stock. Diabase and quartz-porphyry dikes are definitely of later age than the lamprophyres. The former form a rather regular parallel swarm in a S. W.—N. E. direction, while the latter are somewhat younger and considerably more variable in their direction.

The Shap area (Harker and Marr, 1891, and Grantham, 1928). Porphyritic biotite granite forms an oval area averaging two km in diameter. Granite has broken through slightly folded Ordovician volcanics and Silurian sediments. The latter occur only along the southern contact. The intrusion has given rise to a general outward dip in its immediate vicinity. A sideward bending is also commonly visible in the map pattern. The southern and eastern contacts are due to a fault. On the western side, where the contact zone is better displayed, there occurs a network of irregular veins along the margin. Potash feldspar phenocrysts generally have at least a subparallel arrangement. These phenocrysts are often slightly rounded and especially in the »early basic» (a basic variety of the Shap granite) surrounded by a narrow white rim. This feature (rapakivi structure) is quite peculiar also to the Åva area. There are many microscopic evidences of movements during the solidification of the granite. Aplites and miarolitic cavities are relatively common, while pegmatites are rare. Both cognate and country rock inclusions occur. Especially the rounded inclusions of »early basic granite» are most striking and they may be of considerable size. The Shap granite has intruded in at least three stages. The radial character of the dike system has been emphasized by Harker and Marr. The granite area itself is well displayed, but most dikes, especially in the immediate vicinity to the south of the granite are exposed only in river beds without visible contacts against the country rock, which makes the determination of their direction very difficult. They may, moreover, be even more radial than shown on the map of Harker and Marr. Most dikes in the immediate vicinity are also composite, i. e. quartz porphyry has intruded into the same fracture with the lamprophyric material. Only in some cases do they form separate dikes there.

The Spanish Peaks area (Knopf, 1936). Two stocks close to each other have intruded into the nearly horizontal eastern limb of a broad asymmetrical syncline, which is formed by sedimentary rocks of late Cretaceous to early Eocene age. The greater is composite (granite porphyry and granodiorite porphyry) and arcuated in shape and about 7 km in length (ring intrusion?). It is to be noted that the more basic rock is the younger member of the stock. Their habit is only partly plutonic. The other stock is of syenodiorite and transected by both granite porphyry and diorite porphyry dikes. These stocks have caused only a very slight or imperfect doming. The dikes are essentially radial only near the stocks; furthermore, they tend to be perpendicular to the axis of the syncline. The great majority of the dikes are lamprophyres, but there are also some other dikes in the same system. The sequence of central and dike intrusives is complicated, consisting of at least seven pulses.

PETROLOGY

To characterize the rocks of these three areas, their Niggli values have been calculated from the original analyses given (Table 1). The first group in each area is formed by central or major intrusives, the second by dike rocks. From the Shap area no recent data on dike rocks are available. The only analyses of lamprophyres, apparently associated with the Shap granite, are very old (Bonney and Houghton, 1879). The mica traps of the Kendal and Sedbergh districts are, besides, more than 10 km south of the Shap area, so that the data are not of very much use. The general character of these dikes will, however, be given. They are mostly minettes with a greater range of *si* than those of the lamprophyres of Åva. The figures for *k* and *mg* are in most rocks very high, both generally exceeding 0.6. The content of carbonate is commonly very considerable.

The last three rocks designated as dike rocks of the Spanish Peaks are petrologically related to the major intrusives. The others are lamprophyres in a chemical sense, basic dike rocks relatively rich in alkalies (Eskola, 1954). Among them there are rocks called trachydolerites, but they may grade

Table 1

si	*al*	*fm*	*c*	*alk*	*k*	*mg*	
							Åva area
144	25	43	19.5	13	0.46	0.50	Monzonite, Österskär
384	41.5	15.5	11	32	0.51	0.32	Granite, Quarry, Åva
101	20	49.5	19	11.5	0.48	0.59	Pyroxene minette, the southern shore of Långö
106	21.5	45	21	12.5	0.56	0.57	Contact facies of the dike above
116	24	45.5	18	12.5	0.57	0.53	Hornblende minette, Inre Björkskär, S. W. of Långö
121	24	41.5	21.5	13	0.62	0.49	Biotite-hornblende vogesite, the western shore of Åva
124	31	41	13.5	14.5	0.49	0.53	Hornblende vogesite, northwestern Åva
132	25	40	22.5	12.5	0.44	0.41	Hornblende kersantite, Hällholm, Brändö
							Shap area
262	35	27	14	24	0.43	0.50	»Early basic granite»
304	38.5	21	12.5	28	0.48	0.51	Granite, Stage I
323	39	19	12	30	0.50	0.49	Granite, Stage II
346	44	11	7.5	37.5	0.59	0.47	Granite, Stage III
							Spanish Peaks area
193	30.5	31.5	16.5	21.5	0.35	0.42	Pyroxene syenodiorite, West Peak
396	43.5	10.5	7.5	38.5	0.40	0.26	Granite porphyry, East Peak
92	15.5	52	24.5	18	0.44	0.62	Monchiquite (olivine fourchite), Bear Creek
94	16.5	49.5	24	10	0.36	0.66	Olivine fourchite, Pitou dike
97	16.5	51.5	22.5	9.5	0.34	0.63	Teschenite, Bear Creek
98	16	50.5	24	9.5	0.44	0.65	Olivine fourchite, Unfug dike
98	18.5	45.5	26	10	0.34	0.60	Olivine fourchite, border facies of dike, upper Bear Creek
100	17.5	51	22.5	9	0.26	0.60	Analcitic olivine trachydolerite, Bear Creek
101	18	51	22	9	0.26	0.60	Analcitic olivine trachydolerite, Bear Creek
107	16	48	23.5	12.5	0.50	0.60	Biotite-augite vogesite, Walsen dike [1]
118	17.5	49	20	13.5	0.64	0.63	Augite minette, South Fork of Trujillo Canyon
133	15	50.5	19	15.5	0.66	0.68	Shonkinite, Walsenburg—La Veta road
158	29	32.5	19.5	19	0.37	0.37	Microsyenodiorite, Bear Creek
166	31.5	30	19.5	19	0.32	0.43	Augite syenodiorite porphyry, Bear Creek
332	42.5	15	11	31.5	0.38	0.41	Granite porphyry, Devil Stairway dike

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[1] The corresponding values given by Knopf (1936, p. 1766) are 111, 16, 48, 23, 13, 0.50 and 0.61, respectively. Knopf gives Niggli values for no other rocks.

to lamprophyric rocks. Also the lamprophyres proper have a heteromorphic variation. The variable rock names signify changes in their petrography rather than in their chemistry. In the case of hardly any other rock types is the terminology so miscellaneous and confused as in that of the lamprophyres, but such a variable petrography is also very peculiar. Some analyses

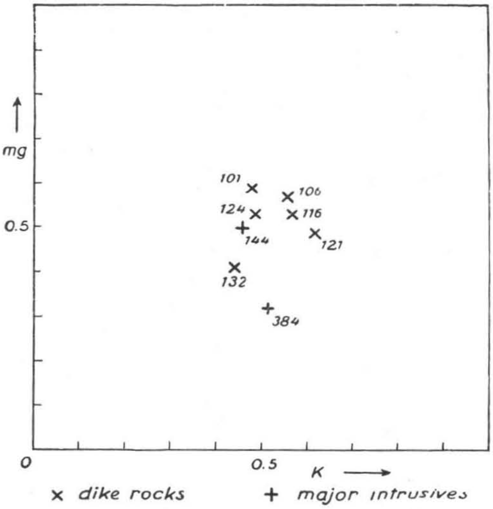

Fig. 1. *k*—*mg* diagram in the rocks of the Åva area. The figures indicate *si* values.

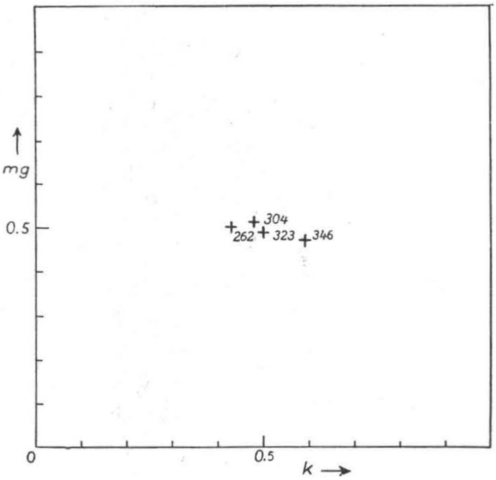

Fig. 2. *k*—*mg* diagram in the rocks of the Shap area.

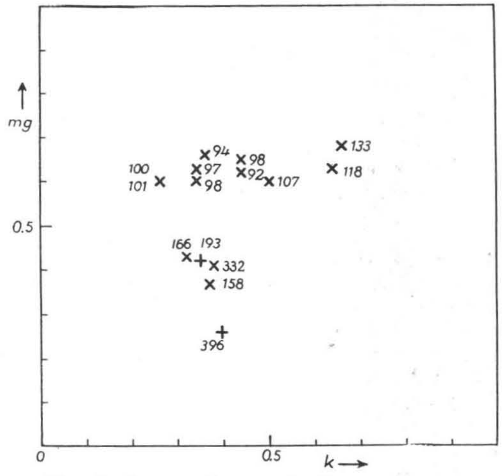

Fig. 3. *k*—*mg* diagram in the rocks of the Spanish Peaks area.

of isolated and outlying stocks are omitted (Table 15, Anal. IX and XVII in Knopf, 1936).

In spite of the great diversity of the rock types, especially when including dikes, all these provinces have many common petrological features. The most striking peculiarity is the high content of alkalies, with potassium predominant even in basic rocks. This character is most clearly shown by the *k*—*mg* diagrams (Figs. 1—3). This is the most elucidative graphical way to plot the data on all rocks. They show that *k* and *mg* mostly do not change very much with the value of *si*. The dike intrusives of the Spanish Peaks are exceptional in this respect. In the Shap area the uniformity is very remarkable in this case.

In the *k—mg* diagram of the rocks of the Spanish Peaks (Fig. 2), the points fall into two general groups. The one (*mg* and also $k < 0.5$) is formed by the central intrusives (*si* = 193, 396) and petrologically related dike rocks (*si* = 158, 166, 332) and the other by the lamprophyres ($mg > 0.5$). Both trachydolerites (*si* = 100, 101) falling into the same point may be considered as soda lamprophyres. In the *k—mg* diagram of the Åva intrusives this grouping is not so pronounced, but the general tendency is the same.

Even though the *k* and *mg* ratios show some differences, we notice that in the variation diagrams the central intrusives in all these areas have the

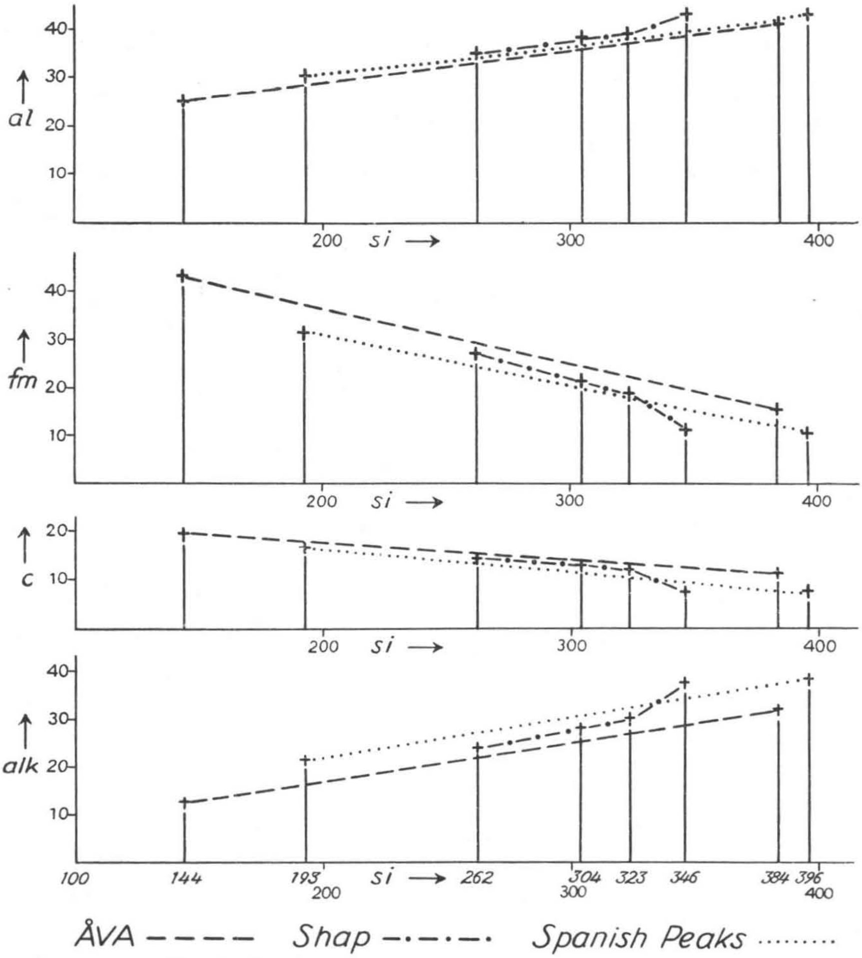

Fig. 4. Variation diagrams of the major intrusives.

same differentiation tendency (Fig. 4).[1] Resemblance in the general tendency cannot be greater, although the rocks are not identical. If the analyzed rocks were more numerous, they would be more nearly identical, especially within granitic rocks. The variation diagrams of the dike rocks are not given, because the values show departures of very heterogeneous character and generally do not form any single variation series (Table 1).The amplitude on the *si* values is usually also too small for drawing any general conclusions.

The central intrusives of the Åva area represent magmas rich in potassium (Kaitaro, 1953, and Table 1), while the rocks of the Spanish Peaks form transitions to the calc-alcalic series, e. g. the granite porphyry of the West Peak represents an engadinite-granitic magma type (Burri and P. Niggli, 1945). The intrusives of Shap represent in many respects an intermediate position. The Shap granite is commonly called adamellite, but the Niggli values of Stage I are the only ones that more closely agree with those of the adamellitic magma type (Burri and P. Niggli, 1945). It is to be noted that the *mg* values of the Spanish Peaks lamprophyres are generally higher than those of Åva.

GEOCHEMICAL AND RELATED CHARACTERISTICS

The most remarkable geochemical feature of all the rocks considered is the relatively high content of BaO (Table 2). The amount of SrO is also always conspicuous, if it is accurately determined, but the material available is mainly of the dike rocks of the Spanish Peaks area. The use of barium and strontium as kinds of geochemical guide elements has been suggested by Knopf (1936), but he has not taken into consideration that also certain petrological processes may cause enrichment of these elements. A high barium content is a general characteristic of lamprophyres (Eskola, 1954), but here it is common in central intrusives, too. According to Eskola, lamprophyres are also characterized by high contents of TiO_2 and P_2O_5. The present data agree with this statement, because the lowest values of this component are generally those of the dike rocks petrologically related to central intrusives. Among the central intrusives the basic rocks are relatively rich in both TiO_2 and P_2O_5.

Titanium and phosphorus are mostly contained in sphene and apatite, respectively, but e. g. in the syenodiorite of West Peak there is no sphene and a relatively large amount of TiO_2 may be present in the pyroxene. Another common accessory in all these rocks is zircon. Allanite is also characteristic, but it is found in the Åva area only in pegmatites. In the

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[1] In the Spanish Peaks area the dike rocks related to the central intrusives are omitted in the diagram, not that they would not fit in a general differentiation, but in principle no dike rocks have been included. As can easily be seen from the corresponding Niggli values in Table 1, they agree relatively well with the general tendency.

Åva and Shap rocks fluorite is common but has not been found in the Spanish Peaks area.

Table 2. Geochemical characteristics

		Åva %	Shap %	Spanish Peaks %
Central intrusives	TiO_2	0.29—1.64	0.29—1.08	0.34—1.74
	P_2O_5	0.19—0.82	0.25—0.41	0.08—0.74
	BaO	0.08—0.23	0.05—0.09	0.09—0.14
Dike rocks	TiO_2	1.60—2.16		0.38—2.41
	P_2O_5	0.77—1.82		0.08—1.34
	BaO	0.10—0.62		0.10—0.56
	SrO	0.05—0.24		0.04—0.20

STRUCTURAL DEVELOPMENT

Both the pre-intrusive structure and character of the country rock have been quite different in each case. Therefore it cannot be supposed that, except in their distinct eruptive character, the structural relation of the intrusives to the surroundings would be quite the same in all these areas. The internal structure again depends more on the character of the intrusive itself. In this respect it must be noted that the level of intrusion is an important factor.

In an earlier study the author (Kaitaro, 1953) has paid special attention to the ring intrusions and emphasized the upward pressure of the magma. The comparison with the Spanish Peaks area shows, on the other hand, the upward pressure and the doming caused by it not to be essential to the development of the common structure, a radial dike system. Because there is no apparent dome structure in the Åva area, the walls dip steeply inwards rather than outwards. The structural pattern indicates by its symmetrical relation to the centre of the complex that the structures have been formed by a force inside the central intrusives. The other structural process for solving the space problem in such an active intrusion is forcing apart the wall rock. If, keeping this in mind, we look at the maps of the Åva and Shap areas, we will find that this may be a solution for the radial fractures, which concurrently or later have been opened and injected by the magma available. These fractures or fracture echelons should in most cases be located to pass through the igneous centre, where the crust was already perforated by magma. The sideward pressure caused by recurring intrusions will continually be in force. By the gradual growth a space for the central intrusions is thus formed, although very slowly, owing to the great strength of the surrounding rocks. The map pattern of the structure of the surroundings of the Åva area shows that it has definitely been a very important process and that the most space needed may have been produced in this way.

Have then the ring intrusions and radial dikes been formed by quite different processes? In many areas, e. g. in Åva, they are so closely related that this is improbable. It is rather to be thought that the conditions have changed in favour of one or the other. The upward movement of the intrusions may have a different character with regard to velocity and continuity. A very forceful, quick and sudden upward movement leads to the formation of shear fractures and ring intrusions (Anderson, 1936), because the strain relaxes upwards, i. e. in the most easily yielding way in this case. When the intrusive force is smaller and in action for a longer period, the overlying rock mass begins after the first uplift to move aside by gravitation. Because of the small compressibility of the surrounding rocks, this is a very slow process. It contributes to increase the force pushing sidewards and tangential tension, causing radial fractures. If the intrusion is very quick, e. g. explosive, there is no possibility of compensation in that way.

In this process the place and level of the intrusions and its relation to the surrounding structure apparently are of relatively less importance. The deep-seated intrusions, the pressure of which is comparatively small in relation to the static load, favour, however, a radial fracture system.

CONCLUDING REMARKS

The structure of rocks can tell very much of their petrogenesis, but only of those processes that have been working in the last stages. The structures produced by earlier petrogenetic processes may be very obscure. Magma, meaning in a broader sense hot, mobile rock material, may have a long history behind it before the final emplacement, which has brought about the rock structures visible at present.

To discuss the petrogenetic development as a whole, it is necessary to know the nature of parent magma, or generally speaking parent material, and the character of the differentiation process. The differentiation of these rocks, especially the basic rocks with a high content of alkalies, cannot be very simple. By normal crystallization and gravitational differentiation, it is not possible to bring about such products. The dike rocks do not belong to the same variation series as the central intrusives. This becomes evident in the *k—mg* diagrams of Åva and the Spanish Peaks (Figs. 1 and 3). The intricate variation of the Niggli values of lamprophyres (Table 1) is apparently a product of several differentiation processes, the last of which is a possible diffusion between the dike and its wall rock (Bederke, 1947).

In the Åva area the lamprophyric dikes favour the radial fractures, while the ring fractures are filled by granite or monzonite material. This could mean that the intrusion of lamprophyric magma and the opening of radial fractures inherent to shrinkage in consolidation and cooling of rocks have been simultaneous. According to Neumann (1947) this is the reason for the formation of the late-magmatic basic dikes. In all these areas described the frequency of dikes generally decreases away from the central intrusives, in accordance with this theory. The lamprophyric dikes have intruded before the final cooling, while the dike rocks formed later on have a more basaltic composition. The contamination has taken place, not only at the present level, but all the way the magma has moved into its final position (Kaitaro, 1953).

The relation between lamprophyres and granite in time and space has often been emphasized and it is evident also in this case. The displacement of granite has resulted in the formation of radial fractures, which have opened by the cooling of the rocks. The material available below the granite has intruded into the fractures. The variable composition may be only partly explained by different primary material. Later on there have been many changes between the dike material and its foreheated wall rock, e. g. assimilation and diffusion. Repeated granite or other intrusions may also make the conditions favourable to metasomatism or metamorphism, which necessarily are not the same in different parts of the dike. In these conditions it is not surprising that the composition of the dikes is so variable and that they do not form any single variation series. Because in most areas there are no major intrusives with the same or even similar composition, the lamprophyric dikes may be formed in the fractures opened by cooling in the manner discussed, which explains both the mode of occurrence and the special character of these rocks. This agrees with Bederke's idea (1947) that the association with granites is more mechano-tectonic than material. The lamprophyric dikes are generally formed at a relatively low temperature, even lower than the formation of granite. This is caused by the large amount of volatile constituents, such as H_2O and CO_2, in the lamprophyric magma. The lamprophyre problem has recently been discussed in more detail by Eskola (1954).

This paper is not concerned merely with the lamprophyre problem. I am not, however, prepared at present to solve the problem of petrogenesis of central eruptives. Nearly two years ago I wrote a draft for it; but, reading it now, I put it again in my deskdrawer, since it does not in some points agree with my present view. I have not abandoned all hope of finding a better solution to the whole problem. I will take the advice given by Horace in *De Arte Poetica* that a literary product should ripen for nine years before being brought out. Prof. Eskola, in one of his lectures, recommended this advice also to the geologists. The foregoing treatment is, however, so largely based on plain facts that I think my point of view on the matter will not change. I consider the matter so important that it is worth publishing separately. Everyone may, however, try to explain the facts presented in the foregoing according to his own personal view.

The purpose of this paper in a petrological sense is to emphasize the importance of local conditions, of which the character of the surroundings is not as significant for the petrogenetic development of magma as the mechanics of intrusion. This idea has been discussed by Smyth (1913). Some years before him Harker (1909) pointed out the importance of mechanical conditions to the origin of alkalic rocks. This has developed the fundamental truth that nonorogenic areas are favourable to the formation of magmas rich in alkalies as has been emphasized by P. Niggli and others. The tectonic environment is recognized as an active factor in the evolution of rock provinces.

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7

KOMPLEXE ERZLAGERSTÄTTEN [1]

VON

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AUSZUG

Neue Deutungen altbekannter Lagerstätten, ihre weitverbreiteten Gründe. Einer der Hauptgründe ist aber die Komplexheit der Lagerstätten und die verschiedenartige Beurteilung der in ihnen vorliegenden heterogenetischen Vorgangsreihen. Verschiedene Beispiele. Bedingungen zum Zustandekommen komplexer Lagerstätten und ihre Systematik.

In den letzten Jahren kann man immer mehr die Beobachtung machen, dass langbekannte und oft untersuchte Lagerstätten und Lagerstättentypen, deren Genesis seit langem beinahe übereinstimmend anerkannt wurde, in neuen Arbeiten völlig anders gedeutet werden. Es ist naheliegend, dies zunächst mit der Vermehrung unserer Einzelkenntnisse, mit der Verfeinerung der apparativen und instrumentellen Untersuchungsmethoden und der experimentellen Kenntnisse in Zusammenhang zu bringen. Es steigt dadurch ja die Mannigfaltigkeit und Kompliziertheit der erkannten Phänomene gewaltig an, und die vorher so einfachen Linien der genetischen Vorgänge werden unklarer und verschwommener und vieldeutiger. Damit wird aber auch das vorher so klare Schema der Klassifikation und Systematik der Lagerstätten immer mehr verwischt und zuletzt immer unsicherer, ja ganz problematisch. Wenn man aber von einer höheren Warte aus und auf Grund ausgedehnter Kenntnisse zahlreicher Lagerstätten und ihrer Literatur diese modernen Arbeiten aufmerksam analysiert, bemerkt man fast durchweg folgende Erscheinungen: in den neuen revolutionären Arbeiten werden ganz neue Methoden verwandt, oft ganz neue Anschauungen zu Grunde gelegt, die aber selbst noch weitgehend umstritten sind (Granitisation z. B.); es sind junge Forscher, oft aus ganz anderen Ländern und Schulen, die

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wenig ältere und ausländische Literatur kennen und auch wenig von ihr halten. Es ist naheliegend, dass der ältere Forscher, der ja auch seine sorgfältigen unantastbaren Beobachtungen vorzuweisen hat, die neueren Erkenntnisse zu sehr ignoriert und die Arbeiten ablehnt. Die jüngeren Forscher übersehen aber auch die älteren Befunde zu sehr und schieben ihre neuen gleichfalls unantastbaren Erkenntnisse in den Vordergrund. Der Ältere ist konservativ und bevorzugt einfache systematisch gegliederte Deutungen, der Jüngere ist revolutionär und gefällt sich oft in bizarren Erklärungen, ohne sehr auf die systematische Klassifikation zu achten. Beide reden also aneinander vorbei und der Fernerstehende ist in Verzweiflung, dass die Lagerstättenkunde so schlecht genetisch unterbaut sei.

Ich glaube, dass in vielen, ja den meisten Fällen dieser Art beide Generationen von Forschern ein sehr einfaches und weitverbreitetes Phänomen übersehen, das geeignet ist, alte und neue Beobachtungen und Ansichten miteinander zu verbinden und mit einem Schlag die meisten genetischen Kontroverse aus der Welt zu schaffen. Die Lagerstättenforscher gingen fast alle seither von dem Gedanken aus, dass jede Lagerstätte so sehr nur durch einen Grundvorgang entstanden sei, dass dieser das gesamte Lagerstättenschema beherrscht und man infolgedessen die Lagerstätte in der Klassifikation nur an einer einzigen Stelle einordnen kann. Dabei ist es in der Petrographie schon seit vielen Jahrzehnten bekannt, dass zahlreiche Gesteine hintereinander von mehreren genetischen Vorgängen ergriffen wurden, die alle ihre Spuren hinterlassen haben: es sei an die Sederholm'schen deuterischen Effekte erinnert, an diagenetische Veränderungen der Sedimente und vor allem an alle metamorphen Gesteine, die ja »per se» Zeugen mehrerer Vorgänge sind und Relikte, Neubildungen dynamischer und chemischer u. a. Art enthalten. Auch in der Lagerstättenkunde sind die von mir schon 1925 (Schneiderhöhn, 1925, 1926) aufgestellten »Übergangslagerstätten» und die noch viel früher besonders von W. Lindgren (1928) verfolgten Umwandlungen der Nebengesteine schon lange bekannt. Das sind aber alles gewissermassen Angelegenheiten der »inneren Politik» (Eskola) der Genesis, wo im Laufe der Entwicklung sich allmählich die äusseren und inneren Zustandsbedingungen im System selbst änderten und infolgedessen neue Genesen, mit oder ohne Verdrängung der alten, enstanden. Viel zu wenig, z. T. aber noch garnicht beachtet wurden diejenigen Fälle, bei denen auf einer älteren Erzlagerstätte jüngere erzbildende Vorgänge aufgetreten sind, die mit dem älteren Erzvorgang genetisch garnichts zu tun haben, die gewissermassen »zufällig» an dieser Stelle der Erdrinde später einmal aufgetreten sind. Mengenmässig können sie alle Bereiche umfassen, von Spuren bis zur Dominanz. Wenn sie bei einer eingehenden und moderneren Differentialdiagnose besonders beachtet werden, ist der jüngere Forscher, der sie entdeckt hat, naturgemäss geneigt, ihnen den grösseren Wert beizulegen und demgemäss stuft er die ganze Lagerstätte darnach an einer ganz anderen Stelle in die Systematik ein. Der Ältere aber sieht nach wie vor seine älteren Paragenesen als die Hauptsache an und bleibt bei seiner früheren Meinung. Der wissenschaftliche Streit ist da, aber man sieht auch die Trugschlüsse, die beide begehen.

Ich möchte solche Lagerstätten als komplexe Lagerstätten bezeichnen und verstehe darunter solche, deren heutiger Inhalt aus mehreren heterogenetischen Vorgangsreihen stammt. Manchmal sind die jüngeren Vorgänge noch in weiterem Sinne mit den älteren blutsverwandt, wenn sie auch an anderer Stelle im System stehen. Vielfach fehlen aber auch alle systematischen und kausalgenetischen Beziehungen.

Im folgenden seien nur einige charakteristische Beispiele erwähnt.

Sehr umstritten sind in den letzten Jahren die *Lagerstätten des Schwefelkreislaufs*. Ich nannte sie so in meiner Kupferschieferarbeit (Schneiderhöhn, 1921) und verstand darunter sedimentäre Sulfidlagerstätten mit Pyrit und Kupfersulfiden, seltener Zinkblende und Bleiglanz, die schichtig in bituminösen Schwarzschiefern und Mergeln, d. h. ehemaligen marinen Gyttja- und Faulschlammgesteinen liegen. Sie sind unter dem Einfluss des von anaëroben Bakterien des Schwefelkreislaufs entwickelten Schwefelwasserstoffs aus den im Meerwasser gelösten Spurenmetallen ausgefällt worden. Die Herkunft der im Wasser gelösten Metalle war dabei zunächst offengelassen worden, sie konnten aus den Festlandssedimenten ausgelaugt sein, oder von älteren Festlandslagerstätten ins Meer transportiert sein, oder aus einer tieferen weitreichenden Grundwasserzirkulation stammen oder endlich aus submarinen Thermalquellen oder Exhalationen, die in das Meerwasser eintraten, zugebracht sein. Für die systematische Zurechnung der Sulfidlagerstätte ist diese Herkunft der Metalle offenbar gleichgültig und bedingt höchstens verschiedene Untertypen. Nun sind in den letzten Jahren die altbekannten grossen und seit Jahrhunderten wirtschaftlich sehr wichtigen Lagerstätten, Kupferschiefer (Eisenhut u. Kautzsch, 1954), Rammelsberg a. Harz (Kraume u. Mitarb., 1955), Meggen a. d. Lenne (Ehrenberg, Pilger u. Schroeder, 1954) genauer untersucht worden, wobei vor allem auf die fazielle Stellung der Nebengesteine und auf die regionale und lokale Tektonik genauer geachtet wurde. Im Rammelsberg und in Meggen wurden in der Schichtenfolge der Umgebung geringmächtige aber verbreitete Einlagerungen keratophyrischer Tuffe gefunden, die tektonisch zu eigenartigen Aufwölbungen gehören. Darunter nimmt man mit guten Gründen einen höheren Aufstieg einer hochplutonischen Intrusivmasse an. Aus ihr sollen hydrothermale Lösungen oder Exhalationen hochgetiegen, ins Meerwasser übergetreten sein, den Metallgehalt dort vermehrt und zu einer lokal erheblichen Sulfidbildung beigetragen haben. Als Fällungsmittel wird z. T. noch bakterieller Schwefelwasserstoff, zum grossen Teil aber juveniler Schwefelwasserstoff von den jüngeren Autoren angenommen. Der Erzbestand, der in Meggen praktisch ganz im ursprünglichen Zustand vorliegt, im Rammelsberg stark metamorph überprägt ist, ist aber ursprünglich ganz epithermal und noch tiefertemperiert, mit überwiegenden Geltexturen und zahllosen Spuren der von mir so aufgefassten »vererzten Bakterien» erfüllt. Es ist nun ein Trugschluss, solche Lagerstätten als »hydrothermal» zu klassifizieren, wie es einige der jüngeren Autoren tun. Sie sind und bleiben vielmehr echte syngenetischsedimentäre marine Ausfällungen, an deren Erkenntnis nur neu ist, dass sehr wahrscheinlich der grösste Teil der Metalle submarin-exhalativ zugeführt wurde.

Ganz ähnliche Verhältnisse liegen bei der grössten türkischen Kupferlagerstätte Ergani Maden in Kurdistan vor, die ich selbst eingehend untersuchen konnte (Schneiderhöhn, 1954b). Die Erzlager zeigen ganz ähnliche Paragenesen und Strukturen, aber im Liegenden, wo heute schon die Zufuhrspalten aufgeschlossen sind, sind in ihnen höherthermale Paragenesen vorhanden, aus denen neuere Forscher den irrtümlichen Schluss zogen, dass das ganze Lager auch oben höherthermal, d. h. hydrothermal enstanden sei. Es liegt dafür garkein Anzeichen im Lager selbst vor. Nur kommen ab und zu auch im Lager einige höherthermale Mineralien vor, die aber aus leichter Metamorphose stammen. Sonst sind aber die durchgängigen Gel- und Bakterienstrukturen herrschend.

Wahrscheinlich gehören zu diesem Typus auch die primären Teile der grossen australischen Lagerstätte Mt. Isa in Queensland und der grossen finnischen Lagerstätte Outokumpu. Beide sind allerdings von jüngeren metamorphen Vorgängen stark überprägt.

Im vorhin schon genannten Kupferschiefer in Deutschland liegen die Verhältnisse etwas anders. Für den westlichen Teil (bei Sontra; Messer, 1955) ist gerade jetzt in der neuesten Arbeit der Primärvorgang wie er früher entwickelt wurde bestätigt worden. Für den Hauptteil in Sachsen (Eisenhut u. Kautzsch, 1954) wurde aber neuerdings gezeigt, dass neben dem gleichfalls anerkannten syngenetisch-sedimentären Hauptbildungsvorgang spätere hydrothermale Zufuhren kamen aus Randspalten des Kupferschieferbeckens gegen die alten Gebirge. Die Lösungen stiegen bis zum Kupferschieferflöz hoch, verbreiteten sich im erzführenden bituminösen Kupferschiefer und reicherten die Erze, z. T. noch mit anderen Paragenesen an. — Hier kam also zu der ursprünglichen syngenetisch-sedimentären Erzschicht eine jüngere mit der ersten Lagerstätte garnicht mehr im Kausalzusammenhang stehende hydrothermale Erzzufuhr hinzu und hat stellenweise den Charakter und den Reichtum der Erze merklich verändert.

Eine ganz ähnliche, wenn auch mengenmässig sehr verschiedene Entwicklungsreihe liegt möglicherweise bei den ja immer sehr umstrittenen kalkalpinen *Blei-Zinkerzen* in den bayrischen Nordalpen und den österreichisch-italienischen Südostalpen vor. Sie wurden bis vor kurzem mit guten Gründen beinahe allgemein als hydrothermal aufgefasst (Schneiderhöhn, 1941, 1944, 1952), wenn auch ein zugehöriger Magmatismus bis heute noch nicht erkannt bezw. allgemein anerkannt ist. Ich fasste sie deshalb vor einigen Jahren als »sekundärhydrothermal» auf, d. h. als ausgelaugte und höher wieder abgesetzte Fraktionen aus älteren vielleicht variskischen tiefergelegenen Lagerstätten. Neuerdings fanden nun jüngere Autoren (Taupitz, 1954) in manchen Teilen besonders nordalpiner Lagerstätten Anzeichen einer syngenetisch-sedimentären, an schwache bituminöse Lagen der kalkalpinen Trias geknüpfte Vererzung und wollen nun die ganzen Lagerstätten so auffassen. Allerdings sind auch nach ihnen wechselnde Mengen der Erze teils hydrothermal umgelagert, teils hydrothermal zugeführt. Das ursprüngliche Problem des fehlenden Magmatismus ist durch diese Auffassung keinen Schritt weitergekommen, dafür ist aber ein neues Problem aufgetaucht, nämlich die Erklärung eines Metallgehalts in solchen warmen Riff-Faziesmeeren und seine selektive Ausfällung. Aber einmal alles als richtig angenommen: dann ist es zweifellos ein Trugschluss, nun auf einmal alle diese Lagerstätten als syngenetisch-sedimentär zu klassifizieren. Höchstens handelt es sich um komplexe Typen aus zwei garnicht miteinander zusammenhängenden Vorgangsreihen.

Ähnliche Probleme bieten ja von jeher die anderen grossen metasomatischen Blei-Zinklagerstätten in Karbonatformationen dar. Für die nordafrikanischen Lagerstätten, besonders in Tunis und Algerien habe ich vor einigen Jahren (Schneiderhöhn, 1954a) den sekundär-hydrothermalen Ursprung aus dem variskischen Untergrund propagiert und habe bei nordafrikanischen und französischen Fachgenossen viel Beifall gefunden. Einen gleichen Ursprung möchte ich vorerst bei den oberschlesischen und den Mississippi-Missouri-Erzen annehmen, wo aber noch dahingehende Einzeluntersuchungen ganz fehlen. Dieser Schluss basiert also nur auf der seitherigen Literatur. Jedenfalls ergeben sich auch bei diesen grossen und wirtschaftlich so überaus bedeutenden Lagerstätten zahlreiche Anhaltspunkte, dass sie nicht durch einen Vorgang entstanden sind, sondern dass es komplexe Lagerstätten sind.

Auch in einer ganz anderen Gruppe grosser und wirtschaftlich wichtiger Lagerstätten genügt es meines Erachtens heute nicht mehr, dass wir sie monogenetisch auffassen. Es sind dies die Gruppen, die ich als »*Aride Konzentrationslagerstätten in terrestrischen Schuttbecken*» zusammengefasst habe (Schneiderhöhn, 1931, 1932, 1937, 1944). Es sind Kupferlagerstätten (Nordrhodesia, Katanga, Westrand des Ural u. A.), Uran-Vanadium-(Carnotit)-Lagerstätten in den Red Beds in USA (Schneiderhöhn, 1944), Blei-Zink-Lagerstätten im Buntsandstein der Nordeifel in Westdeutschland

(Schneiderhöhn, 1953), um nur einige grosse zu nennen. Ich habe sie seit vielen Jahren (Nordrhodesia nach eigenen Untersuchungen) als sygenetisch-sedimentär aufgefasst, z. T. mit nachträglichen inneren Umlagerungen und z. T. auch mit schwachen späteren durch Überlagerung bedingten epimetamorphen Umbildungen. Die amerikanisch-englische Schule hat die Kupfererze immer epigenetisch-hydrothermal aufgefasst, trotzdem garkeine Indizien dafür vorliegen und zahlreiche negative Indizien dagegen sprechen. Die jüngere Schule in Nordrhodesia ist aber jetzt (Brummer, 1955) meiner Meinung beigetreten. Bei den Carnotiterzen sind die Meinungen geteilt, neuerdings scheint die sedimentäre Meinung zu überwiegen.

Die Nordeifeler Blei-Zinkerze fasste ich immer gegen die meisten dortigen Geologen und Bergleute als sedimentär auf. Sie wollten sie hydrothermal erklären, es fehlte allerdings immer der zugehörige Magmatismus. Neuerdings ist man an das Problem von der Seite der Metallbilanz herangegangen. Es wurde gezeigt, dass die alten Blei-Zink-Gänge im devonischen Untergrund, die zahlreich dort vorhanden sind, nicht hinreichen, um bei der Abtragungshöhe des Rheinischen Schiefergebirges über dem Buntsandstein die Metallmenge der gut bekannten Lagerstätten im Buntsandstein zu erklären. Nun liegen aber unter dem Buntsandstein noch zahlreiche alte Erzgänge, aus denen sekundär-hydrothermal zusätzliche Metallmengen heraufgebracht werden konnten, sodass die grossen Metallmengen sich zwanglos erklären lassen, ebenso auch noch andere Erscheinungsformen, die bei der rein sedimentären Entstehung Schwierigkeiten bieten.

Denselben Mischtypus aus sedimentärer und sekundär-hydrothermaler Entstehung möchte ich nun auch für die vorgenannten Kupfer- und Uran-Vanadiumerze annehmen. Wenn man sie in ähnlicher Weise als komplexe Lagerstätten ansieht, fallen auch bei ihnen viele Schwierigkeiten weg, die bei einseitiger Deutung so oder so immer bleiben.

Auch die grösste *Goldlagerstätte*, die wir kennen, der Witwatersrand bei Johannesburg, bietet seit Jahrzehnten jeder monogenetischen Erklärung Schwierigkeiten. Gold und das neuerdings zur grossen Bedeutung gelangte Uran sind wohl zweifellos sedimentäre Seifenbildungen (Ramdohr, 1954). Es sind aber infolge der enormen Sedimentüberdeckung und der mehrfachen magmatischen Durchwärmungen interne Umlagerungen und Umbildungen eingetreten, »pseudo-hydrothermale» Wirkungen, die auch dieser Lagerstätte einen völlig komplexen Charakter verliehen haben.

In vielfach gesteigertem Masse und in mannigfachster Abstufung bis zu den grössten Intensitäten sind ja nun solche spätere Umbildungen und Umlagerungen im gesamten metamorphen Grundgebirge erfolgt. Die mobilisierte Gebirgsfeuchtigkeit, die aus den Mineralien freigesetzten Wasser-, Kohlendioxyd-, Chlor-, Sulfat- etc. Lösungen, weitere vagabundierende magmatische Zufuhren etc., erzeugten zahlreiche Stoffumlagerungen, Neu-und Umkristallisationen. Am empfindlichsten dafür sind die Erzmineralien, die den meisten Silikaten dabei vorauslaufen. Es ist eigentlich eine Selbstverständlichkeit und in zahlreichen Arbeiten auch nachgewiesen, dass in diesen Lagerstätten des metamorphen Grundgebirges vom kleinsten bis zum grössten Umfang Mobilisationen, Stoffumlagerungen und Neuabsätze von Erzen auch an vorher erzfreien Orten vorkommen. Oft kann man nur an Relikten, meist sogar nur im Mikroskop die alten Bestände und Gefügearten erkennen. Naturgemäss kommen für solche Bildungen das kristalline Grundgebirge aller alten Schilde und dazu noch die stärksten orogenetischen Zonen aller grossen Orogenesen in Betracht.

Damit sind also eigentlich alle Lagerstätten in solchen Gebieten als komplex zu bezeichnen, soweit sie prätektonisch sind. Es können die jüngeren Merkmale kaum erkennbar, aber auch ganz stark und geradezu dominierend sein. In Fennoskandia und auch in Russland hat man dies schon lange erkannt, aber in USA, Kanada, Zentralafrika, Indien, Australien will man auch heute oft noch aus dem gegenwärtig vorliegenden Mineralbestand und seinem Gefüge auf die Primärvorgänge schliessen und beachtet fast nirgends den total komplexen Charakter der Lagerstätten.

Zum Schluss seien die Bedingungen kurz zusammengefasst, die für das Zustandekommen komplexer Lagerstätten günstig sind (die amerikanischen Lagerstättenforscher würden es wohl als »complex control» bezeichnen). Komplexe Lagerstätten sind ja dadurch gekennzeichnet, dass auf früheren Paragenesen sich jüngere Sulfide, Oxyde und Gangarten niederschlagen, meist nur durch Reaktionswirkung, seltener und meist untergeordnet durch Verdrängung, wobei im allgemeinen eine bedeutende Vermehrung der Metalle sowohl mengenmässig als auch prozentual eingetreten ist.

Es gibt dabei günstige chemisch-mineralogische und auch günstige formal-geologische Bedingungen. Chemisch-mineralogisch günstig ist das Vorhandensein älterer sulfidischer Mineralien, vor allem der unedlen Metalle von Fe, Pb, Zn, besonders von Gelpyrit, weiter bituminöse und kohlige Substanzen, Imprägnationen und verkohlte Pflanzenreste. In geringerem Masse wirken zersetzte Alkaligesteine, Karbonate insbesondere mit Fe und Mn, zersetzte Fe-Mg-Silikate, auch dünne Tonlagen zwischen mächtigeren klastischen oder karbonatischen Gesteinen. Die Wirkung dabei beruht teils auf Wechselreaktion, teils auf Adsorption.

Formal-geologisch günstige Bedingungen bieten dünne Zwischenlagen, Bänder, Linsen mit den oben genannten Mineralien dar, wenn sie Grundwasser oder tiefere vadose und magmatische Wasserhorizonte führen und wenn sie aufstauend wirken; auch poröse erzimprägnierte Lagen sind günstig, auf denen Lösungen leicht und zusammenhaltend zirkulieren können, wo jedenfalls Wechselreaktionen leicht und durch lange Zeiten langsam stattfinden können.

Geotektonisch sind dafür lange persistente Ruhezeiten günstig, also terrestrische Denudationsflächen oder marine Emersionsflächen, auch langsam sinkende Geosynklinalen, aber vor der beginnenden Faltung.

Was nun die Systematik solcher Lagerstätten anlangt, so habe ich schon 1919 in meinem Frankfurter Habilitationsvortrag (Schneiderhöhn, 1919) betont, dass man strenggenommen nicht Lagerstätten, sondern nur lagerstättenbildende Vorgänge klassifizieren kann. Wenn auf einer Lagerstätte ein weitaus überwiegender Hauptvorgang sich ausprägt, ist ihr Platz in der Systematik ja unumstritten. Eine Lagerstätte aber, die zu beinahe gleichen Teilen aus heterogenen Vorgängen gebildet und aus heterogenetischen Mineralien besteht, an einer Stelle einzuordnen, ist prinzipiell falsch. Sie steht an zwei (oder mehr) Stellen mit Anführungen der jeweils charakteristischsten Paragenesen und mit Verweis auf den anderen Teil und mit der stark betonten Bemerkung, dass es sich dabei um eine komplexe Lagerstätte handelt.

Die Systematik und Darstellung der Erzlagerstätten wird dadurch zwar wesentlich komplizierter und umständlicher, aber der Charakter und die Genesis der Einzellagerstätte wird dadurch auch klarer und richtiger, und das liegt ja nicht nur im Interesse der Wissenschaft, sondern auch im Interesse des Bergbaus.

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8

A FERROAN-ALUMINIAN ANTHOPHYLLITE FROM KEMIÖ, FINLAND [1]

BY

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ABSTRACT

A quartz-feldspar schist (leptite) containing anthophyllite and garnet is described. The anthophyllite is extremely rich in iron and aluminum. The chemical composition, optical properties, X-ray powder pattern, and unit-cell dimensions of the anthophyllite are presented. The garnet is an almandite poor in the pyrope component. The chemical and powder X-ray data for the garnet are given.

CONTENTS

THE ROCK

In 1952, during field work in the parish of Kemiö, southwestern Finland, the author found a large area of quartz-feldspar schist (leptite) containing, as a rule, sillimanite and/or garnet. The constituents of the rock are quartz, oligoclase, microcline, varying amounts of fibrous sillimanite and almandite garnet, and biotite.

When sillimanite is absent, the rock often contains only little or no microcline. At a locality about 5 km S. S. W. of the church of Kemiö and 0.7 km N. N. E. of Lake Bjensböleträsket, anthophyllite occasionally occurs in the microcline-free rock within a small area. The main minerals of this rock are quartz, plagioclase ($An_{12\text{-}18}$), garnet, anthophyllite, and biotite,

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[1] Received May 30, 1956.

ilmenite being the most common accessory (see p. 81). The garnet grains are surrounded by aureoles consisting of mere light minerals, but the largest crystals (up to 9 mm in diameter) are associated with light bands parallel to the foliation, which mainly consist of sericitized plagioclase and are more coarse-grained than the main part of the rock. The larger prisms of anthophyllite (up to 4 mm in length) form groups, while the smaller crystals are dispersed. The anthophyllite is often largely altered to a mixture of serpentine, chlorite, and iron oxide. Although both garnet and anthophyllite can always be seen in the same hand specimen, they often occur, however, somehow apart from each other, so that certain thin sections lacked one or the other mineral.

THE ANTHOPHYLLITE

The pleochroism of the anthophyllite was so strong and its refractive indices so high, considerably higher than the maximum figures for anthophyllite given by Winchell (1951) and Tröger (1952), that an accurate investigation of the mineral was desirable. Unfortunately, quite unaltered anthophyllite was found in one small specimen only, and thus material for a chemical analysis was obtained somewhat less than 0.8 g. The separation was carried out by means of the Clerici solution in a centrifuge and of an isodynamic magnetic separator. The amount of impurities (garnet and oxidic ore) was estimated at 2 %.

The determination of the constituents was made by the methods commonly used. Thus the FeO content was determined by the HF-H_2SO_4 method. For the determination of combined water by means of the Penfield tube, the powder was mixed with the equal amount of lead peroxide. The chemical composition and optical properties of the anthophyllite are given in Table 1.

In Rabbitt's extensive study (1948) of the anthophyllite series, no undeniable anthophyllite is presented that would have an equally high Fe content as the anthophyllite under consideration (MgO/FeO = 1/4). The original analysis of gedrite (from Gèdres, Pyrenees Mountains; Rabbitt, 1948, p. 274), sometimes referred to in literature, shows 45.83 % FeO, 4.13 % MgO, and the proportion of FeO from the total FeO + MgO about 87 %. But this analysis dates back to 1836, and such a high FeO content »has not been duplicated in any of the many analyses made on gedrite from Gèdres since that time» (Rabbitt, 1948, p. 286). Rabbitt mentions two anthophyllites with a higher weight-% of Al_2O_3 than in the Kemiö anthophyllite, but regards one of them as doubtful. However, the substitution of Al for Si in the Kemiö anthophyllite is higher than ever reported, exceeding 1/4 of the number of the Si atoms. The comparatively high content

Table 1. Chemical composition and optical properties of anthophyllite, Kemiö, Finland. Analyst, P. Ojanperä

	%	Mol. prop.	Atomic prop. (O = 24)			
SiO_2	36.65	6 102	Si	5.69		8.00
TiO_2	0.73	91	Al	3.58	2.31	
Al_2O_3	19.55	1 918			1.27	7.33
Fe_2O_3	4.32	271	Ti	0.08		
FeO	30.03	4 180	Fe^{3+}	0.51		
MnO	0.18	25	Fe^{2+}	3.90		
MgO	4.45	1 104	Mg	1.03		
CaO	0.05	9	Mn	0.02		
Na_2O	1.70	274	Ca	0.01		
K_2O	0.00	—	Na	0.51		
H_2O+	2.14	1 188	K	—		
H_2O-	0.06	—	OH	2.22		2.22
	99.86	—		—		

$mg = 0.19$

$\alpha = 1.683$
$\gamma = 1.713$
$\gamma \wedge c = 0°$

(+) 2V = 85°
= 85°
= 87°

Pleochroism:

α very pale brownish yellow
β brownish green
γ greyish green

The refractive indices were determined by the immersion method, 2V and $\gamma \wedge c$ on the U-stage. The three figures for 2V resulted from the measurements for three different crystals.

of Na_2O is in agreement with Rabbitt's data (1948), according to which Na_2O increases with increasing FeO and Al_2O_3.

It is true that the analysis was made of a small amount of material and thus the figures obtained are, perhaps, somewhat inaccurate. The analyst, however, is very well acquainted with silicate analysis, and small inaccuracies do not affect essentially the figures.

The birefringence of the mineral (0.030) is rather high. Tröger (1952) holds the opinion that the birefringence of the aluminian anthophyllites increases with an increase of Fe, but according to Rabbit (1948, p. 296) no significant effect of Al on the birefringence is apparent and, in general, the birefringence would decrease with increasing Fe. The high birefringence observed might also indicate some variation in the composition of the mineral.

The powder X-ray diagram was taken by means of a Norelco Geiger Counter X-ray Diffractometer (North American Philips Co., Inc.) using unfiltered CoK_α radiation. The anthophyllite powder was mixed with a silicon standard. The run was made within a range of $2\theta = 10°—90°$ with a speed of 1°/2 inches.

Table 2. Powder pattern of anthophyllite, Kemiö, Finland

hkl	*d* (obs.)	*Q* (obs.)	*Q* (calc.)	Intensity
210	8.264	0.0146	0.0147	7
400	4.658	0.0461	0.0463	1
040	4.471	0.0500	0.0499	1
420	4.120	0.0589	0.0587	1
231	3.646	0.0752	0.0753	2
250	3.344	0.0894	0.0896	2
440	3.223	0.0963	0.0962	4
610	3.054	0.1072	0.1072	10
521	2.883	0.1203	0.1204	2
251	2.825	0.1253	0.1253	2
630	2.752	0.1320	0.1322	2
351	2.676	0.1396	0.1397	3
161	2.574	0.1509	0.1509	5
202	2.550	0.1538	0.1542	3
451	2.502	0.1598	0.1600	3
551	2.319	0.1859	0.1860	2
502	2.158	0.2148	0.2150	1
561	2.131	0.2202	0.2203	1
661	1.992	0.2521	0.2521	3
751	1.979	0.2554	0.2554	2
851	1.830	0.2987	0.2988	2
961	1.618	0.3821	0.3823	1
10,6,1	1.513	0.4370	0.4372	2
0,12,0	1.492	0.4495	0.4495	1
6,11,0	1.441	0.4818	0.4818	2
11,0,2	1.425	0.4922	0.4926	1
4,12,0	1.421	0.4956	0.4957	2
8,10,0	1.418	0.4971	0.4972	1
11,6,1	1.417	0.4978	0.4979	1
12,6,1	1.331	0.5644	0.5644	2

The powder pattern of the anthophyllite is presented in Table 2. The indices have earlier been listed by Johansson (1930) as well as Warren and Modell (1930). Some indices were found by calculation on the basis of the values $a_o = 18.60$ Å, $b_o = 17.90$ Å, and $c_o = 5.30$ Å. The extrapolation was extended against $\sin^2\theta = 1$. The unit-cell dimensions of the Kemiö anthophyllite obtained from the powder pattern are as follows:

$$\left.\begin{array}{lcr} a_o & = & 18.596\ \text{Å} \\ b_o & = & 17.899\ \text{Å} \\ c_o & = & 5.295\ \text{Å} \end{array}\right\} \pm 0.005\ \text{Å}$$

$$\text{Unit-cell volume} = 1\,762.5\ \text{Å}^3 \pm 2\ \text{Å}^3$$

Rabbitt (1948, p. 305) reports the unit-cell dimensions determined by the Weissenberg method for nine analyzed anthophyllites. The variation ranges are: $a_o = 18.50$ Å—18.58 Å; $b_o = 17.66$ Å—18.10 Å; $c_o = 5.28$ Å—5.31 Å. In b_o the range is thus considerable, but if the figures for b_o are plotted against the variable contents of SiO_2, Al_2O_3, $FeO(+Fe_2O_3)$, and MgO, no systematic variation can be seen. In a_o and c_o the variation range is much smaller, and if these bear some relation to the chemical composition, this is impossible to detect from the data given. In earlier reports of the unit-cell dimensions of anthophyllite (Warren and Modell, 1930, p. 165; Johansson, 1930, p. 36) the figures for a_o and c_o are in good agreement with each other and near the corresponding minimum figures given by Rabbitt. The b_o values deviate from each other, but lie within the aforementioned range. When the unit-cell data for the Kemiö anthophyllite, which is extremely rich in Fe and Al, are compared with the data cited, it is seen that the figure for a_o is a little higher than ever reported, that for c_o seems to exceed most of the corresponding earlier figures, and the figure for b_o represents an average among the most different figures reported for b_o.

THE GARNET ASSOCIATED WITH THE ANTHOPHYLLITE

After the composition of the anthophyllite was known, it might be assumed that the composition of the garnet regarded as almandite does not deviate greatly from it, and the garnet was also investigated. The small sample from which only unaltered anthophyllite could be obtained was rather poor in garnet. After the separation of the anthophyllite, a garnet fraction measuring 0.5 g was obtained, but it still contained considerable amounts of an oxidic ore mineral. This impure garnet fraction was analyzed by Mr. Pentti Ojanperä, M. A., with the following result: SiO_2, 35.23 %; TiO_2, 5.12 %; Al_2O_3, 16.23 %; Fe_2O_3, 3.77 %; FeO, 37.22 %; MnO, 0.35 %; MgO, 2.32 %; CaO, 0.00 %; total 100.24 %. Na_2O, K_2O, and H_2O were not determined. The high TiO_2 content indicated that the ore mineral was probably ilmenite, the amount of which would be about one-tenth of the whole. In a powder X-ray diagram of the garnet fraction, five ilmenite peaks could be recognized. For comparison, a powder X-ray diagram of another almandite garnet was taken. If the whole TiO_2 content and the equivalent amount of FeO are subtracted from the total and the rest is recalculated to 100 %, the approximate composition of the garnet will be as follows:

SiO_2	38.9 %
Al_2O_3	17.9 »
Fe_2O_3	4.2 »
FeO	36.0 »
MnO	0.4 »
MgO	2.6 »
	100.0 %

It may be seen that the garnet is an almost pure almandite ($mg = 0.10$), although the Fe_2O_3 content is too high. The calcian components are entirely lacking. The refractive index $N_D = 1.821 \pm 0.003$ is in agreement with the said composition.

Table 3. Powder diffraction data for the garnet

hkl	*d* (obs.)	Intensity
400	2.878	4
420	2.575	10
422	2.351	4
510	2.260	4
521	2.104	4
440	2.032	2
611	1.870	7
444	1.664	3
640	1.599	5
642	1.541	4
800	1.440	3

$a_0 \sim 11.51$ Å

The powder X-ray diagram of the garnet was taken using unfiltered CuK_a radiation, and the powder was mixed with a silicon standard. The run was made within a range of $2\theta = 70° — 10°$ with a speed of 1° per 1/2 inch. The figures for the *d*-spacings (with a maximum error of ± 0.005), corresponding to prominent peaks, are given in Table 3. The figure for $a_0 \sim 11.51$ Å was obtained. For a fairly pure almandite-pyrope with $mg = 0.16$, Fleischer (1937, p. 757) has reported $a_0 = 11.495$ Å.

It is surprising that two different minerals, the compositions of which are so nearly similar to each other, apparently occur together. It is true that the garnet is somewhat richer in Fe than the anthophyllite, but an almandite-pyrope with the same FeO/MgO ratio as in the anthophyllite should have been formed as well in this rock, and all the other differences are very small. Now, most of the garnet seems to occur in the bands of sericitized plagioclase, which do not contain anthophyllite. The bands, often lenticularly formed, are more coarse-grained than the main part of the rock and give the impression of segregation. The sericitization of the plagioclase seems to indicate that considerable amounts of water have been present during this stage of metamorphism. If an excess of water is present, anthophyllite, according to Yoder (1952, p. 609), would be unstable at all temperatures. However, it is unknown whether this is valid also in respect of ferroan anthophyllites. Because almost all the anthophyllite has altered to more hydrous minerals, the anthophyllite seems to have crystallized during an earlier stage and is no longer stable. The garnet is quite unaltered. It would be expected that garnet had formed as an alteration product of the anthophyllite, because the compositions of the two minerals are so closely related to each other. Such an alteration, however, could not be observed.

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9

POSTMAGMATIC POTASH METASOMATISM OF GRANITE [1]

BY

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ABSTRACT

Since the technics in the study of phase equilibria at high pressures and temperatures enabled determining the temperature-concentration diagram of the »haplogranitic» system and the common granites proved closely to agree in composition with the »eutectoid» minimum in the said system, the long pursued aim of demonstrating granite to be the last product of the crystallization-differentiation of magmas has been reached. The composition of many late-kinematic granites, however, deviates systematically from the minimum-melting composition in being two or three pt. richer in potash. I have earlier termed such rocks ideal granites. Quite commonly, on the other hand, replacement of some soda by potash is known to have taken place in granites and other silicate rocks. The complex pegmatites carrying trace element minerals as well as feldspars in mineral veins have been known as hydrothermal crystallizations, and several »geologic thermometers» indicate that crystallization in granites has continued well below the minimum temperature of magmatic crystallization. According to Schneiderhöhn the pneumatolytic-pegmatitic ore deposits mark the divide between magmatic and hydrothermal crystallization. Szadeczky-Kardoss has recently found that the bonding potentials (mol. weights divided by ionic volumes) of minerals decrease in the crystallization series of rocks and ores but, just at the said divide, suddenly jump high up again, as a diluted aqueous solution takes the place of concentrated magmatic solution. It is clear, therefore, that the crystallization of granites, also, must continue within the hydrothermal range. Hereby the potash metasomatism and the excess of potash in the ideal granites can be accounted for. The source of the potash is so far problematic, but may possibly be sought in the general upward migration of specifically light atoms in the earth's crust.

CONTENTS

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[1] Received May 26, 1956.

Oppia ikä kaikki
(Life is all learning)

A Finnish proverb

INTRODUCTION

In my last paper on the granite problem (Eskola, 1955) I made some concessions to the transformist views modifying my earlier stand that the metasomatic changes in the process of granitization as well as the primary crystallization of granite are entirely results of the action and the materials of granitic magma (Eskola, 1950). I wrote (1955, p. 127) that rocks like clay or diorite, which in themselves contain most of the granite elements, may be converted to granite by potash metasomatism alone, and furthermore (1955, p. 128) that alkali silicates brought by diffusion into rocks may effect changes in the minerals and, at proper proportions of the substances added, lead over to hydrothermal granitization.

What I have to add in the present paper concerns the mechanism of the replacement, and more especially potash metasomatism, in rocks that have crystallized as granite primarily from the magma. Such postmagmatic hydrothermal alteration has been well known long and elucidated e. g. by such an old master of petrology as Fr. Becke. Now it appears, however, by application of the principles of ionic and bonding potential, as recently developed by Szadeczky-Kardoss (1953, 1954, 1955), that the hydrothermal potash metasomatism is of great importance in the development of granites. Being a well defined process and one distinguishable from the primary magmatic crystallization, the hydrothermal postmagmatic metasomatism opens new avenues for study bound to throw more light upon the most important of all petrochemical problems, the granite problem. Among other things, we find at once a satisfactory explanation for the comparatively low temperatures of the granite genesis found by means of many recently detected geological thermometers (Barth, 1955; Kullerud and Neumann, 1953). Thus, also, we find that these and other similar results are not contradictory to other experimental results, such as those of Tuttle (1948) and Tuttle and Keith (1954) proving beyond doubt that the granites have originated by crystallization from granitic magma at the temperature of the granitic minimum.

EUTECTOID GRANITE AND IDEAL GRANITE

In my earlier papers (Eskola, 1950, 1952) I had designated the typical microcline granites characterized by a preponderance of potash over soda as ideal granites. It had been my assumption at that time, before 1950, that this might be the composition of the granites crystallized from the lowest melting residual granitic magma, corresponding to the minimum temperature of the granitic system Qu—Or—Ab—H_2O, with some An, mafic and accessory minerals. Soon after that time the exhaustive experimental investigation by Bowen and Tuttle (Adams, 1952; Bowen, 1954) had proceeded far enough to prove that the minimum-melting mixture in the system, which may be called (though the name is not quite correct) eutectoid granite, is that which contains almost equal amounts of soda and potash feldspars, as established by the experimental investigation. The same composition was found by plotting the normative quartz-orthoclase-albite proportions from 571 rock analyses on a triangle first published in Adams (1952). The ideal-granitic potash-rich composition was found

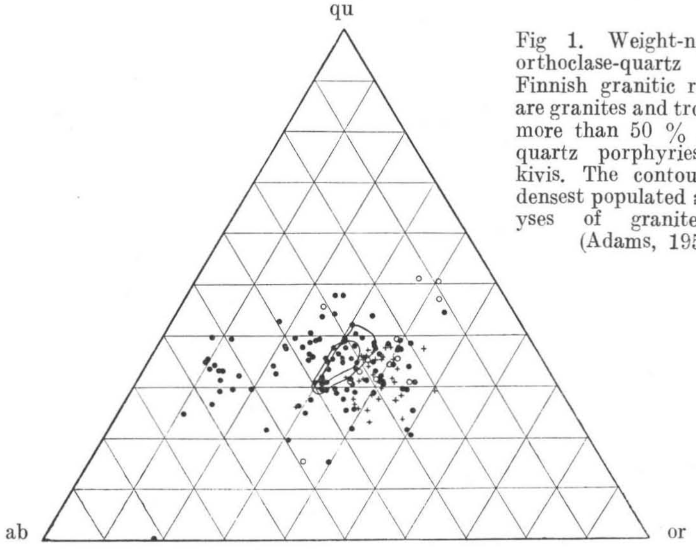

Fig 1. Weight-normative albite-orthoclase-quartz ratios of 149 Finnish granitic rocks. Black dots are granites and trondheimites (with more than 50 % Ab); open circles quartz porphyries; crosses rapakivis. The contours delineate the densest populated areas of 571 analyses of granites and syenites (Adams, 1952, p. 41).

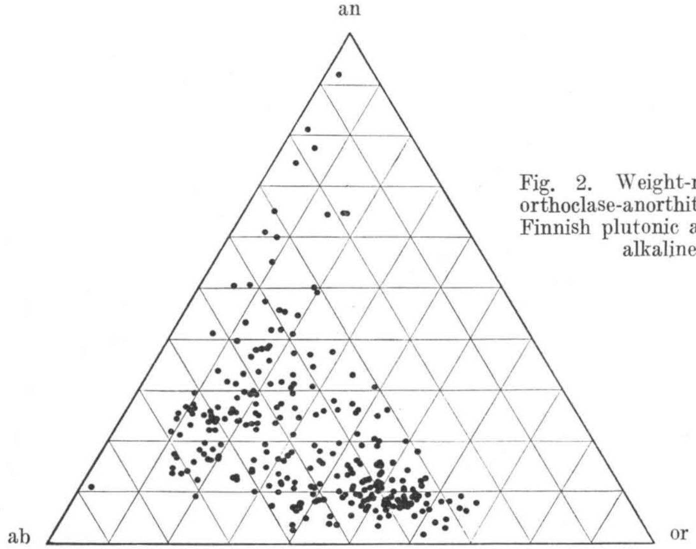

Fig. 2. Weight-normative albite-orthoclase-anorthite ratios of 266 Finnish plutonic and volcanic calc-alkaline rocks.

to be very common among the late-kinematic and postkinematic granites of Finland. Plotted on the triangle, their analyses concentrate at some distance toward the or-qu side from the minimum-melting composition (Fig. 1). In the ab-or-an triangle the points likewise deviate toward the or-an side (Fig. 2). Analyses showing the range of variation in both groups are listed in Table 1.[1]

Table 1. Compositional variations of late-kinematic granites

	Oxides, weight per cent					Cation numbers per cent			
	Eutectoids		Ideal granites			Eutectoids		Ideal granites	
	1	2	3	4		1	2	3	4
SiO_2	71.20	72.82	73.91	71.75	Si	68.12	68.48	69.79	68.29
Al_2O_3 ...	14.02	13.66	13.90	14.13	Al	15.81	15.16	15.46	15.85
Fe_2O_3 ...	0.48	1.30	0.14	1.12	Fe^{3+} .	0.34	0.92	0.10	0.80
FeO	1.15	1.22	1.01	1.37	Fe^{2+} .	0.92	0.95	0.80	1.09
MnO	0.01	0.03	tr.	—	Mn ...	0.01	0.02	—	—
MgO	0.10	0.39	0.28	0.64	Mg ...	0.14	0.55	0.40	0.90
CaO	0.74	1.10	1.00	1.59	Ca	0.76	1.11	1.01	1.62
Na_2O ...	4.40	3.45	2.42	2.25	Na ...	8.15	6.29	4.43	4.15
K_2O	4.57	5.00	6.53	5.85	K	5.57	6.02	7.86	7.10
TiO_2	0.09	0.55	0.22	0.18	Ti	0.06	0.39	0.18	0.14
P_2O_5	0.14	0.14	—	0.07	P	0.11	0.11	—	0.06
H_2O + ..	0.37	0.35	0.19	0.83		—	—	—	—
H_2O — ..	—	0.06	—	—		—	—	—	—
	100.27	100.07	99.93	99.78		99.99	100.00	100.00	100.00
	Weight-normative ratios					One-cation mol.-normative ratios			
	1	2	3	4		1	2	3	4
ab	39.85	28.16	28.56	22.17	Ab ...	43.64	32.25	24.23	23.99
or	29.19	34.16	42.34	40.52	Or ...	29.83	34.61	43.00	41.04
qu	30.96	37.68	35.10	37.31	Qu ...	26.53	33.14	32.77	34.97
ab	54.83	45.43	31.99	30.95	Ab ...	56.28	46.86	33.31	33.17
or	39.76	46.10	60.28	56.20	Or ...	38.28	44.88	59.10	56.75
an	5.41	8.47	7.73	12.85	An ...	5.25	8.25	7.59	10.07

1. Granite, eutectoid, magmatic, central type. Northeast of Ylivieska church. Analyst, L. Lokka. Wilkman, 1931, p. 146.
2. Nattanen granite, eutectoid, magmatic, almost postkinematic. Analyst, E. Nordensvan. Mikkola, 1928, p. 9.
3. Perniö granite, ideal-granitic, magmatic. Sillanpää, Kisko. Analyst, P. Eskola. Eskola, 1914, p. 17.
4. Hangö granite, ideal-granitic, metasomatic, with relict layering. Skarvkyrkan, Archipelago of Hangö. Analyst, Laura Hezner. Sederholm, 1912, p. 304.

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[1] As emphasized by Niggli, Barth, and Eskola (see **Eskola**, 1954), the rock analyses should be presented in ionic percentage. This form would be preferable also in illustrating the alkali ratios, for their differences would be accentuated, as the numbers of the univalent alkali ions are twice the numbers of the two-cation alkali molecules (oxides). The triangle projections (Figs. 1 and 2) are, however, made on the weight norm basis to make them comparable with the presentation practiced by the Geophysical Laboratory. The differences in the percentage figures are not very great either, as appears from Table 1 in which analyses of two eutectoid and two ideal granites are presented in weight and cation number percentages for comparison and the feldspar ratios likewise in weight-normative as well as one-cation molecular normative (= ionic) percentage ratios.

The ideal-granitic composition is quite common in granites of similar tectonic position all over the world within orogenic zones of different ages from the Alpine to the Svecofennian. These granites invariably belong to the younger granites of their orogenic zones, as also do the eutectoid granites, but a general difference in their mode of occurrence can be revealed. Recently Dr. Ahti Simonen reminded me about the difference between the late-kinematic granites of the coastal districts of southwestern Finland and of Central Finland including Middle Ostrobothnia and the Tampere tract. The following features are characteristic of both these districts.

In the coastal districts the late-kinematic granites are potash-rich ideal granites in the sense just defined. They are partly cross-cutting magmatic granites with no relict structures and partly metasomatic granites conformable with the schists they penetrate forming migmatites which, in many districts, pass over into granites showing relict layering and nebulitic structures such as the Hangö granite (Table 1, Anal. 4; cf. Eskola, 1955, p. 129) but — and this is important — they are nowhere connected by gradual transitions with more basic rocks. They are all palingenic and, what indeed is a natural consequence of being palingenic, they are no end products of differentiation.

In the central districts there are eutectoid as well as ideal granites and the former ones display differentiation. Gradual transitions to granodiorites, quartz-diorites, gabbros, and peridotites do occur in connection with them. Palingenic granites occur in these districts as well.

Here I shall not enter upon a discussion concerning the deeper meaning of this different behaviour and only wish to point to the regular relation of the compositions. As seen from the analyses, the ideal granites contain one or two per cent more potash and about the same amount less soda than the eutectoid granites.

For many years I was not able to think of anything but that the compositional difference between the two groups would be due to circumstances connected with their primary magmatic crystallization. My former students who attended my lectures in the years 1950—52 will remember my fruitless efforts to find an explanation for the deviation of the granites containing 5 to 7 per cent potash and 3 to 1.5 per cent soda from the eutectoid granites with almost equal amounts of soda and potash.

In the fall of 1952 I consulted Dr. Bowen in this matter, asking whether he could explain why the composition with that pertinent excess of potash is so common among the late granites. I quote with his permission a few passages from his two letters. On December 12, 1952, Dr. Bowen wrote:

> »It is true that the binary minima in the alkali feldspar series are as given in our earlier papers but this does not mean that the ternary minima for mixtures with added quartz need have Ab and Or in the same ratio and in fact they do not. The position of the ternary or »granitic» minimum for various pressures of H_2O is shown

in the sketch diagram herewith. (Later published in Adams, 1952.) You will note that the ratio of Or to Ab is always greater than in the binary minimum. There is, however, as in the feldspars, a tendency for the minimum to be at higher Or : Ab ratio at low pressures of H_2O than at high pressures. — — — I must confess that I have not succeeded in evaluating the significence for natural rocks of the fact that the low water content minimum (low pressure) is richer in potash relative to soda than the higher water content minimum.»

In another letter, dated January 18, 1953, Dr. Bowen wrote:

»— — — We have found the peculiar fact that although water aids the crystallization of the purely alkali feldspars it does not aid much when some lime is added to the alkali feldspars, in fact it even seems to have an opposite effect. Still the addition of An to alkali feldspars — cannot fail to push the low-melting liquids to Or-richer compositions since the »eutectic» between Or and An is practically at Or, whereas that between Or and Ab is richer in Ab than in Or.»

As appears from the last lines, Dr. Bowen's statement that an increase in the An component in the plagioclase »cannot fail to push the low-melting liquid to Or-richer compositions» is founded merely upon the apparent fact mentioned by him and not upon direct experiment. Analyses of potash-rich granites do not show any corresponding systematic increase in the An-ratio which, to be sure, is low in all true granites. Therefore the existence of the potash-rich ideal granites cannot be due to greater An-contents of plagioclase. Nor can one think it possible that the unexpected increase of the Or-ratio with decrease of water pressure would be the cause as even at atmospheric pressure Ab is a little higher in percentage than Or.

It is clear from a comparison of analyses of eutectoid and ideal granites that the latter can be derived from the former merely by replacement of some soda with potash in their alkali feldspars. As nowadays every petrologist well knows, replacement of plagioclase by potash feldspar is in fact an exceedingly common phenomenon in granites and other silicate rocks. Its great importance was established especially by Drescher-Kaden (1948); related phenomena have been described by a great number of petrologists. As a rule this replacement is explained to have taken place in the hydrothermal temperature range, although it is not easy to see why it could not have commenced in the pneumatolytic stage. The crystallization of potash feldspar phenocrysts of porphyritic granites is today attributed by many investigators to the same metasomatic replacement. Therefore the phenocrysts are now called porphyroblasts. Even such a phenomenon as myrmekitization has been classed with phenomena resulting from potash metasomatism, instead of Becke's very plausible theory that myrmekitization is quite a contrary phenomenon: partial replacement of potash feldspar by plagioclases, whereby the separation of the myrmekite quartz also appears to find its natural explanation. Although Becke's explanation of myrmekite, at least in most cases, appears more plausible to me, there is no escape from the conclusion that an addition of potash by metasomatism of already crystallized plagioclase is a widely spread phenomenon in granites.

But it is hard to get an idea about the quantitative importance of the several kinds of potash metasomatism observed in granites. Also it is in many cases impossible to demonstrate that the change has not been mediated by granite magma as I conceive to be the case of the granitization of leptites and mica-schists to yield late-kinematic granites in the Finnish Archean.

LIQUID-MAGMATIC AND HYDROTHERMAL ORE DEPOSITS

A new impetus in the forming of my ideas about the genesis of granites was given during Professor H. Schneiderhöhn's lecture on physicochemical and atomic physical factors in the genesis of ore deposits of the magmatic series, held on 6th April 1956 in Helsinki.

In the well-known Schneiderhöhn classification the pneumatolytic-pegmatitic deposits (Schneiderhöhn, 1941) make the divide between the liquid-magmatic and the hydrothermal series. Beginning from the early magmatic, first crystallized magmatic minerals and proceeding finally to the evaporates and products of weathering (clay minerals), the series denotes a fall of crystallization temperatures from about 1 200° C to temperatures prevailing on the earth's surface. All the way there are two parallel series, one comprising oxidic and sulphidic ore minerals and the other silicatic, carbonatic, sulphatic and halogenidic rock and gangue minerals. In both groups the divide between the high temperature and low temperature part is the critical temperature of the respective liquids from which the minerals crystallized. It is thus theoretically quite sharply defined, though it is different for different compositions.

As each assemblage of ore minerals occurs together with its gangue or enclosing rock minerals, the ore and the gangue or the rock must in any pair have been formed approximately at the same temperature. Both parallel series indicate the sequence of crystallization of the mineral assemblages, the ores and the rocks, at gradually lower temperatures, and the passing over to the hydrothermal stage does not mean any discontinuity in the fall of temperature, though, from the study of the ore deposits as well as from petrological research it proves to be very sharp.

In his lectures on the ore deposits of magmatic parentage Professor Schneiderhöhn reviewed the investigations of Elemer Szadeczky-Kardoss on the application of atomistic principles in the study of the crystallization of ores and rocks. The works of this Hungarian geologist had been unknown to me earlier, but I at once understood that his ideas would greatly influence our understanding of rock crystallization, especially also the problem of granitization. I proceeded avidly to study his writings, and the result of this is the present note.

WICHTE, IONIC POTENTIAL AND BONDING POTENTIAL AS FACTORS OF GEOCHEMICAL MIGRATION INCLUDING GRANITIZATION

Szadeczky-Kardoss (1953a) first discusses the role of the specific gravity of ions, i. e. the atomic weight divided by the ionic volume. In terms of the weight A and radius r this is

$$\varrho = \frac{3\,A}{4\,\pi\,r^3}$$

The new German name »Wichte» which I shall use in this form in the sequel, has been coined for this ratio. In analogy to the ionic wichte the atomic wichte is calculated from the atomic radius. Both are periodic functions of the atomic numbers. For the eight major elements of the earth's crust the numeric values of the wichtes are calculated as follows (Table 2).

Table 2

	Ionic wichtes	Atomic wichtes
8 O	1.66	0.60
11 Na	5.83	0.85
12 Mg	12.33	1.42
13 Al	34.76	2.20
14 Si	113.60	4.18
19 K	3.97	0.76
20 Ca	8.04	1.27
26 Fe (Fe^{2+})	23.31	6.99
(Fe^{3+})	44.29	

The migration of the elements is regarded by Szadeczky-Kardoss (1953, 1955) as a kind of gravitative differentiation, which may take place not only in the liquid state but also in solid rocks, and not only by way of soaking liquid phases through solid phases but also in perfectly solid crystalline matter by means of a metabolism, a replacement of heavier downward-moving ions by lighter ones rising upward. Recent experimental investigation has shown that diffusion through crystalline matter is extremely slow, whereas the intergranular film between individual crystals affords easier pathways for liquid solutions. At least in the upper crust, where granitization takes place, infiltration along fissures is, as advocated by Korzhinsky (1954), probably still more important than diffusion.

The anion wichtes are of course smaller than the cation wichtes, but both are always greater than the atomic wichtes and, moreover, rise with the increase of valence. Elements with a variable valence and ensuing ionic wichte, especially those occurring now as cations and now as neutral atoms or even as anions, like carbon, are bound to perpetual migration and can never reach a lasting stable condition.

The mobilities of the elements do not stand in any direct relation to the ionic or atomic radii, as earlier believed by many investigators.

The ionic wichtes are believed by Szadeczky-Kardoss (1953b, 1954a) to control the entire dynamics of the geochemical migration. Especially important is the upward wandering of the lighter ions constituting the granites; it may be asked what role the Szadeczky-Kardoss principle has played in the greatest geochemical migration, the differentiation of the sial crust from a primary homogeneous earth.

As introduced by Cartledge in 1928, the concept of ionic potential was applied only to the cations. It is the quotient of valence by the ionic radius and it has been characterized as the intensity of the positive charge upon the surface of the ion. Szadeczky-Kardoss uses also the anionic potential of complex as well as of simple anions, and furthermore the bonding potential of crystalline phases, *i. e.* minerals. This is calculated from the potentials of the composing ions according to the following equation:

$$\Psi = \frac{\Psi_1 a + \Psi_2 b + \Psi_3 c + \ldots.}{a + b + c + \ldots.}$$

where Ψ_1, Ψ_2, Ψ_3 are the ionic potentials and a, b, c the numbers of the ions.

For difficulties arising in calculating the potentials for lattices, including infinite chains, nets and three-dimensional structures, reference may be made to Szadeczky-Kardoss, 1954b. Moreover, other difficulties must arise from the fact that the crystal-chemical bonds, either ionic, homoeopolar, metallic, or van der Waals bonds, each of which involves its different radii and consequently different bonding potentials, are in real crystals often mixed in varying proportions. These cannot always be determined. In general the potentials cannot therefore be calculated quite exactly, and many other factors, like the presence of mixed crystals and camouflaged trace elements, etc., affect the values, so the possible error in the potentials may be as great as 10 or 20 pt. But even so the relative values generally agree well with what may be expected on other grounds.

The ionic and bonding potentials representing the free energies of the minerals are shown to elucidate, in a surprising manner, many petrologic and geochemical processes and relations.

The crystallization takes place in the order of decreasing bonding potentials, which primarily indicate the energy levels. The idioblastic series of minerals likewise corresponds to a series of decreasing bonding potentials. This fact enables one to calculate the exact position of a mineral in the idioblastic series. The bonding potential affords a measure of the mineral's »power of crystallization» and, also, its hardness and melting point. The latter, again, is one of the decisive factors for the sequence of crystallization.

Table 3. Bonding potentials (Szadeczky-Kardoss, 1955b)

Bonding potential ↑	Liquid-magmatic		Residual-magmatic		Metamorphic and evaporatitic	
3.4						
					Kyanite	3.32
					Rutile	3.09
					Staurolite	3.06
					Zircon	3.05
3.0	Pt-metals 2.9—3.0		Corundum	3.00		
					Sphene	2.76
					Garnet	2.71
					Topaz	2.65
					Epidote	2.60
			Hematite	2.68		
	Chromite 2.53		Claudetite	2.64		
2.5	Ti-V-magnetite 2.50		Baddeleyite	2.54		
	Magnetite 2.47		Cassiterite	2.47		
	↙	↘			Beryl	2.26
			Uraninite	2.27		
	Forsterite 2.17		Molybdenite			
	Fayalite 2.07		Realgar	2.15		
2.0	↓	↓	Arsenopyrite	1.97	Tourmaline	1.9
			Wolframite	1.93		
			Pyrrhotite	1.76	Hornblende	1.75
			Sphalerite	1.71	Sericite	1.73
			Chalcosite	1.60	Kaolinite	1.67
	Hypersthene 1.60				Chlorite	1.64
	Diopside 1.54		Galenite	1.53		
1.5	Amphibole 1.51	Anorthite 1.52				
	Biotite 1.47	Albite 1.41			Biotite	1.47
	↘	↙ Orthoclase 1.38			Ankerite	1.4
	Quartz 1.38		Cinnabarite	1.29	Calcite	1.33
			Scheelite	1.22		
					Cryolite	1.28
					Strontianite	1.18
			Argentite	1.10	Fluorite	1.13
					Baryte	1.04
1.0						
					Gypsum	0.85
					Halite	0.79
					Nitratite	0.70
0.6					Sylvinite	0.65

In Table 3 are collected the numeric values of the bonding potentials of some important minerals of magmatic, metamorphic and sedimentary rocks and of magmatic ore minerals (Szadeczky-Kardoss, 1955a, 1955b). As appears, the magmatic minerals arrange themselves exactly in the sequence of Bowen's reaction series, while the metamorphic minerals represent the idioblastic series of Becke. The ore minerals, again, arrange themselves in the sequence of crystallization as indicated by microscopic study.

The liquid-magmatic stage of crystallization ends within the crystallization period of the pegmatites, the graphic intergrowth of feldspars and quartz representing the last portion that crystallized directly from liquid magma. The corresponding sulphidic ore parageneses can be located within the series, and in both series the hydrothermal stages follow after the pegmatitic-pneumatolytic stage.

The decrease of temperature, of course, proceeds continually, and the crystallization of quartz and feldspars continues without break, while in the values of the bonding potentials we see a striking discontinuity. Pegmatite minerals like topaz and beryl, many tourmalines and most other trace-element minerals have crystallized in the hydrothermal stage at moderate temperatures, yet they show potentials that place them above olivine, the phase of the Bowen series first separated. This fact undoubtedly means that topaz would crystallize very early and at a very high temperature if it could crystallize from concentrated silicate solutions, i. e. magmas. Instead it separates from diluted hydrothermal solutions at low temperatures. The same is the case with metamorphic minerals like kyanite, staurolite, garnets, or epidote. The crystallization of quartz and feldspars, micas, etc., on the other hand, continues in the hydrothermal range without break.

CONSEQUENCES FOR GRANITIZATION

We know (Landes, 1933, and others) that the so-called complex pegmatites have crystallized in the hydrothermal stage from very diluted solutions. Their tenor of trace-element minerals is one indication of this: compounds of lithium, scandium, lanthanides, columbium, tantalum, caesium, uranium, etc. Still, the most abundant minerals of the complex as well as of the simple pegmatites are feldspars, quartz, and micas. Adularia, albite, and quartz are found in low-temperature mineral veins. All this means that the crystallization of granite also can continue below the critical point of water.

Then we know the products of alteration in granites: epidote, chlorite, kaolin. Sericitization of plagioclase means potash metasomatism, but the potash may have been derived from the minute amounts contained in the plagioclase (Abelson, 1955, p. 143). For our present problem still more important, however, is the potash-feldspathization of plagioclase. It is not probable that this potash metasomatism had taken place entirely at the hydrothermal stage but certainly below the temperature of the granitic minimum, somewhere at 640° (Tuttle, 1955).

An important indication of the extremely small concentration of silicatic compounds in the aqueous solutions left after the magmatic crystallization is the fact that the pegmatite veins are followed by pure quartz veins. The solubility of silica amounts to some 0.30 pt. in maximum and it is also aoluble in superheated steam. According to Morey (Adams, 1950, p. 30) »at maximum observed solubility (500° C, 1 000 bars) the gas contains 0.26 per cent silica by weight and has a partial pressure of 2.62 bars». At still higher pressures of water vapor the solubility increases further, being at 500° C and 2 000 bars 0.5022 pt (Morey and Hesselgesser, 1951). The feldspars are less soluble. Morey and Chen (Abelson, 1955, p. 143) found that water held at 350° C for 103 days with finely powdered orthoclase contained a total amount of 268 parts dissolved material per million of water, or 0.000268 per cent. The albite is just a little more soluble.

At such small solubilities of the feldspars in superheated water hydrothermal granitization in a large scale might appear improbable. The various facts mentioned, however, leave little doubt of its reality and we must find some reasonable explanation for it. Now, the bonding potentials shed much new light upon the problem. The crystallization of quartz and feldspars continues with their low potentials in the hydrothermal region and no discontinuity can be inferred in their rate of separation. The amount of the trace element minerals is, in the total masses of the »complex» pegmatites, still minimal as compared with that of quartz and feldspars. Therefore we must conclude that the hydrothermal solutions could not be as diluted as the experiments would indicate.

A fact hinted at by Roy and Tuttle (1956) will, I think, account for both the hydrothermal crystallization of pegmatites after the magmatic stage and the excess of potash in the ideal granites. The fact I mean is that, although the solubility of water in granitic magma has proved to be limited to some 10 per cent in maximum, the solubility of water in alkali silicate melts is unlimited. As early as 1917 Morey and Fenner (quoted by Roy and Tuttle, 1956) found that in the system $H_2O—K_2O—SiO_2$ there is complete solubility between anhydrous silicate compositions and water. In 1948 Tuttle and Friedman found that a similar relationship exists in the system $H_2O—Na_2O—SiO_2$. Due to the unlimited solubility there is a continuous transition from alkali silicate melts containing a little water to aqueous solutions.

The composition of many surely magmatic granitic rocks, on the other hand, proves that the granitic magmas may contain excessive alkali silicates in solution. As the composition of such a granitic magma containing potash in excess over the minimum-melting ratio belongs to the orthoclase field, potash feldspar will be the first mineral to crystallize out of such a magma. This in fact is what has happened in granite porphyries and quartz porphyries with idiomorphic orthoclase phenocrysts and also in porphyritic granites, such as rapakivi.

There is little hypothetical in this assumption, although it is not possible to tell exactly how such an excess of potash in the granite magma has originated. We may assume, e. g., that the gradual slow upward migration of potash in the mantle and crust of the globe will introduce a limited percentage of potash into the granitic magma wherever this is formed by crystallization differentiation or by anatexis.

As appears from the results of Szadeczky-Kardoss (see Table 3), orthoclase's place in the idioblastic series is at the low end. It is therefore not reasonable to think of perfectly euhedral orthoclase or microcline crystals as porphyroblasts. Their idiomorphism can only be accounted for by an early crystallization from the magma. Savolahti in his academic thesis has independently explained the rapakivi structure in this manner, and thus can many porphyritic granites possibly be explained.

Crystallization of granites in the geosynclines, while kinematic processes continue, may lead to differentiation and depletion of the excessive orthoclase component, and eutectoid granites will result, while under the quieter conditions of the latest postkinematic, or akinematic, granite genesis, ideal granites will form directly from the magma.

But when the magma crystallizes, yielding eutectoid granite, possibly with some excess of potash over the minimum ratio, the potassium silicate will separate out as a liquid phase, remaining in the rock as a pore solution in the residual water; and now the potash metasomatism continues. Albite is converted into orthoclase or microcline and the residual soda is leached out, finally finding its way as sodium carbonate solutions in thermal wells.

The rather constant K/Na ratio of the ideal granites, which is the most difficult feature to account for by this hypothesis, could possibly be a consequence of the constant slow upward migration of the potassium ions. In rocks with excessive alumina the potash metasomatism would first bring about potash feldspar instead of aluminous minerals like garnet or cordierite and eventually result in alkali minerals, in granites first as acmite or riebeckite. The fact that such alkali rocks rarely form may be due to the presence of much excessive alumina in the sialic rocks, but also to the slowness of the addition of potash.

Potash metasomatism can, of course, befall any ancient rocks as well as granite just crystallizing. Granitic-looking rocks may thus originate. The genesis of true granites of ideal-granitic or eutectoid composition cannot be expected, however, unless the original rock was the product of magmatic crystallization. It may then have been formed through metasomatism mediated by granitic magma, as I have concluded earlier (Eskola, 1950). Otherwise the composition would be expected in most cases to deviate appreciably from the eutectoid, and even from the ideal-granitic range. Such clearly non-eutectoid, granitic-looking rocks are not rare, but mostly they differ also texturally so much from true granites that, in petrographic descriptions, they only rarely are called granites. The inference presented in the foregoing that alkali silicates could be dissolved in granitic magma is not new, and it is also corroborated by experiment, as appears from the following quotation from Roy and Tuttle (1956, p. 176):

> »A second important feature of silicate-water systems discovered in the early studies of Morey was the fact that all compositions between water and anhydrous silicate melts were possible in compositions containing alkalies. This information can be extrapolated to granite compositions and thus affords a reasonable mechanism for producing the large zoned pegmatites associated with granitic rocks as well as a source of hydrothermal solutions which carry the ores and give rise to granitization in the wall rocks of batholiths. The studies of Goranson on granitic-water mixtures appear to contradict the above concept as he has shown that only a limited amount of water will dissolve in a granite liquid at pressures expected in the earth's crust. However, the mixtures he used did not contain an excess of alkalies as granites and rhyolites commonly do, and, therefore, his results cannot be used to prove that continuous solubility is impossible and that the early theories are incorrect.»

THE PRESENT STAND OF THE GRANITE CONTROVERSY

This note turned out to another »interim report» of my conclusions in the granite problem and it may be concluded with my evaluation of the situation on the granite controcersy just now. Stated briefly, it seems to me that the dispute for this time is nearing its end, and a happy end, too, inasmuch as both sides can regard themselves as winners.

In the last yearbook of the Geophysical Laboratory of the Carnegie Institution (Abelson, 1955) one finds the laconic statement that the long awaited work of Tuttle and Bowen, »The origin of granites in the light of experimental studies in the system $NaAlSi_3O_8$—$KAlSi_3O_8$—SiO_2—H_2O», was ready for publication in December 1955. So much is known already about the results of these studies that the magmatic origin of true granites is settled, or, as Roy and Tuttle (1956, p. 176) modestly put it: »The fact that granites do have compositions grouped near this minimum (in the »haplogranitic» system Albite—Orthoclase—Quartz—Water) strongly supports the magmatic theories for the origin of granites». The evidence is physicochemical and based upon experimental research. Additional evidence from petrological and geological observation in nature has been given i. a. by Eskola (1950).

Evidence of a metasomatic origin of some granites has recently also come from experimental laboratory work proving that magmatic crystallization in certain cases can bring forth replacement of atoms yielding granitic compositions. The review by Roy and Tuttle (1956) well illustrates the magnificent development of the methods of high pressure-high temperature investigation. By means of the newest hydrothermal bombs, investigations may be carried out under pressures as high as those at the very deepest levels of the earth's crust. Practically all the minerals occurring in the deep-seated plutonic and metamorphic rocks can now be synthetized and their stability conditions have been established in many cases already. Crystallization in the subsolidus field as well as the migration of ions in hypercritical solutions are being studied and, in general, metamorphic and metasomatic phenomena, including those leading to the formation of granites, are open to experimental investigation. Some physico-chemical principles, the application of which brings forth further elucidation to the problem of strictly hydrothermal granitization, have been pointed out in this note. One such principle is that of the bonding potential as set forth by Szadeczky-Kardoss, equally applicable to the illustration of magmatic as well as metamorphic and metasomatic crystallization of mineral masses.

Heretofore the metasomatic or transformistic theories of granitization have been almost entirely based upon field observation and many of their advocates have taken a somewhat supercilious attitude towards those who regard the general physicochemical principles as the foundation of all petrology. Therefore, many of the transformists have fallen to exaggeration and many of them apparently still indulge in too lofty speculations. It is to be hoped that the situation might now improve, as the only thing that the transformists are still expected to admit is the truth in Read's saying that »there are granites and granites». Practically all the advocates of the magmatic view already admit the reality of metasomatic granitization, whereas some of the transformists still seem to believe that all granites are of a metasomatic origin.

My own opinions in the granite problem have changed somewhat from those presented in the paper of 1950. I do not regard this a shame and indeed feel fortunate in being able to learn some new things and to correct my mistakes. One must learn as long as one is living. This is the meaning of the saying adopted as the motto of this article.

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