GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 188

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

HELSINKI
MARS 1960

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GEOLOGICAL SURVEY OF FINLAND
OTANIEMI, FINLAND

Helsinki 1960. Valtioneuvoston kirjapaino

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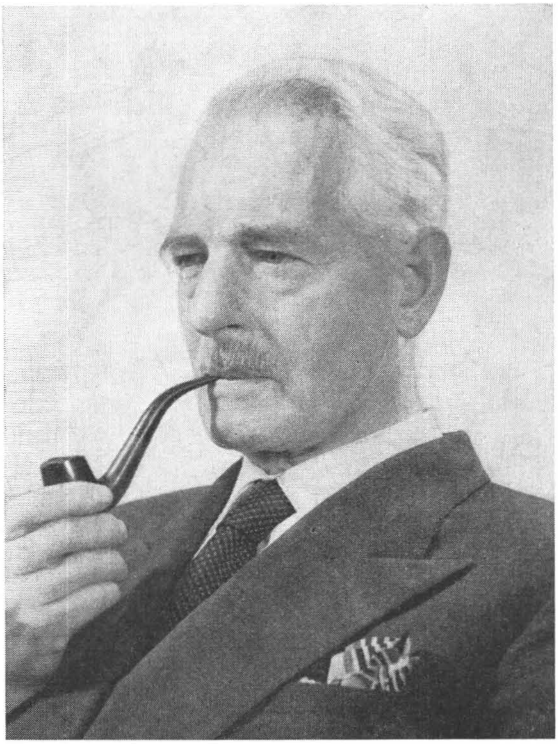

AARNE LAITAKARI AT SEVENTY

December 12, 1960

»He who values even rocks is everywhere surrounded by riches.»

Per Lagerqvist

Aarne Laitakari is privileged to look back upon a long and distinguished career developing geological research in Finland. His interests as a mere schoolboy marked out the course of his academic studies. He was still an undergraduate when he came out with his first publications, among them a handbook on minerals, done in collaboration with Pentti Eskola. Following a number of shorter mineralogical publications, there appeared in 1921, to terminate his academic studies, his doctor's thesis on the petrology and mineralogy of the Parainen limestone.

In addition to his research activities, Laitakari took on various teaching duties and focussed attention with especial interest on ore prospecting and other practical aspects of geology. His attitude could be expressed in these words: »Let nothing concerning rocks and their utilization be foreign to me.» An excellent opportunity to adhere to this principle of his opened up to him when he was engaged as instructor in mineralogy and geology by the Institute of Technology in 1922. He held this position until 1935, serving at the same time as docent of mineralogy in the University of Helsinki. During this period, in spite of the demands on his time made by his teaching duties, he produced, *inter alia*, the following publications: »Die Graphitvorkommen in Finnland und ihre Entstehung» (1925), »Über das jotnische Gebiet von Satakunta» (1925), »Über Mineralbildung längs schmalen Spalten» (1929), »Verdrängungen in Sulphidmineralien von Pitkäranta und Outokumpu» (1931) and »Orijärven vismuttimineraaleista» (1934). Subsequent research has distinguished among the minerals described in the last-mentioned work a new mineral species, already detected by Laitakari, and it has accordingly been named after Laitakari.

It was in 1937 that Laitakari was appointed head of the Geological Board, which evolved into the Geological Survey of Finland we know today. Previously the institution had mainly concentrated on scientific research and on making geological maps of the country, but upon Laitakari's taking charge its program began to include in an ever-increasing degree measures required by practical geology, specifically ore prospecting. Impetus was given principally by the development of the copper deposit of Outokumpu into a productive mining enterprise, the existence of large deposits of nickel ore in Petsamo, found in 1920, and the discovery of numerous ore prospects calling for investigation in various parts of the country. A feature of Laitakari's term as head of the Geological Survey was the development of the institution in every respect as well as its expansion to meet practical needs, and during his term the staff alone grew tenfold. Notably in the work of educating the public in geological matters, Laitakari's personal contribution has been significant; his countless articles in different newspapers and periodicals have stimulated many a citizen to search for ores and useful minerals, and his »Everyman's Book on Rocks» (Jokamiehen kivikirja) has literally become worn ragged in the hands of people.

In spite of the financially difficult years following the war, Laitakari succeeded in drawing government attention too to the necessity of modernizing the Geological Survey. Supporting the reform measures was the new heavy industry, created to carry out the war reparations program, for it clearly showed that Finland sorely needed resources of industrial raw material of her own and that, accordingly, more serious attention than ever should be devoted to ore prospecting. The Geological Survey was reorganized in 1945, its staff was substantially enlarged, the research facilities were augmented and in 1956 it moved to new headquarters at Otaniemi. The new headquarters building of the Geological Survey ranks as one of the most up-to-date in the world, and it will serve as Laitakari's most important and enduring monument for later generations.

During Laitakari's term the general geological mapping of Finland was nearly completed on the scale of 1: 400 000, and in addition a more detailed mapping on a scale of 1: 100 000 by now covers broad areas of southern Finland and Ostrobothnia. Furthermore, in 1951 a systematic aerogeophysical exploration of the entire country was started, and the maps drawn up as result also serve the general ore prospecting program. The direct search for ores conducted in conjunction with these basic research operations has aroused faith in the potentialities of the Finnish mining industry as several new ore deposits have been discovered in different parts of the country. In addition to the exploratory activities concentrating on ores containing the basic metals, Laitakari has laid particular emphasis on the need to look for useful rocks and minerals in general and to exploit them as a raw material of domestic industries. He is to be credited more than anybody else with having fathered, for instance, the concept »native gem,» and he exercised a fundamental influence on the birth of the small-scale industry based on native stones of ornamental value.

The great scope of the reforms he sought to carry out made such heavy demands on his time that he was obliged to relegate his scientific activities proper to the background. The saying »resting on his laurels» does not, however, apply to Laitakari. Having reached the age of seventy and left behind his official career of more than forty years' duration, Laitakari enters his emeritus-sanctum with his accustomed punctuality in order to resume his long-interrupted literary labors. Wishing him the best of success in these tasks, the Geological Society of Finland congratulates an active member on this anniversary of his. At all stages Laitakari has worked for the best interest of the society, supporting its activities now as secretary, now as chairman, now as member of a number of committees.

LAITAKARITE A NEW Bi-Se-MINERAL [1]

BY

ATSO VORMA

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ABSTRACT

This paper describes a new Bi-Se-mineral discovered at the Orijärvi mine in SW-Finland. The chemical composition is Bi_4Se_2S. X-ray investigations show the mineral to be isostructural with joseite (Bi_4 $(Te,S)_3$). The crystal symmetry class is D_{3d}—$\bar{3}m$ and space group $R\bar{3}m$, R32 or R3m. Its hexagonal cell dimensions are: $a_0 = 4.225$ Å and $c_0 = 39.93$ Å.

This new mineral is named laitakarite after Professor Aarne Laitakari, director of the Geological Survey of Finland.

CONTENTS

INTRODUCTION

The selenides are exceedingly rare minerals compared with the tellurides and sulfides. It is for this reason that we seldom find any new literature about the selenide minerals. In the last few years the selenides have been treated rather extensively by Earley (1950) in his paper dealing with the

[1] Received November 11, 1959.

synthesis and appearance of selenide minerals. His description of the Bi-selenide group is confined to the rhombic guanajuatite (isostructural with bismuthinite), rombohedral paraguanajuatite and the artificial compound Bi_2Se_3 (isostructural with the rhombohedral tellurobismuthinite and tetradymite).

In Finland Prof. Aarne Laitakari has investigated and described (1934) the Bi-selenides of the Orijärvi mine. In 1932 were found at a depth of 60 m very small mineral grains resembling bismuthinite in the shaft. In the following year Laitakari found in the mine dump material of the same kind contained in quartz-anthophyllite-cordierite rock; and later on he ran across some more grains in the same mine. He treated 25 kg of specimens on a shaking table and electromagnetic separator and obtained a concentrate of 0.5 kg. By means of ore microscopic examinations and chemical analysis he came to the conclusion that the concentrate contained guanajuatite, native bismuth, bismuthinite, and chalcopyrite.

Laitakari was not quite sure that the mineral regarded as guanajuatite was definitely guanajuatite. At his suggestion the author has reinvestigated the material on the selenides of Orijärvi. During this work it became clear that the mineral regarded previously as guanajuatite was a new species, the structure of which was different from that of guanajuatite.

OCCURRENCE

The Orijärvi region belongs to the Svecofennian schist zone and is characterized by amphibolites and cordierite-anthophyllite rocks intercalated with both skarn rocks and pure limestones.

The bismuth- and selenium-bearing minerals of Orijärvi occur in quartz-anthophyllite-cordierite-biotite rocks as thin veinlets, the thickness of which varies from 0.1 mm to 2 mm. According to Eskola (1914) the cordierite-anthophyllite rocks have been formed by the action of magnesian metasomatism. The rock is characterized by radiating aggregates of anthophyllite, measuring from 0.5 to 3 cm in diameter. Between the aggregates occur granoblastic quartz (diameter from 0.2 to 2 mm), allotriomorphic cordierite (diameter under 5 mm), which is only very slightly pinitized, and biotite poor in iron (diameter from 0.25 to 1 mm). There are also in the host rock disseminated sulfide and selenide grains. The greatest part of the ore minerals occur, however, in the aforementioned veins. The ore mineral association in these veins is: laitakarite (91.3 %, Leitz's integration stage), chalcopyrite intergrown with laitakarite (4.6 %), native bismuth (2 %), chalcopyrite, sphalerite, molybdenite, silver (Bi-bearing), pyrite, and galena.

The diameter of the grains of laitakarite varies between 0.5 and 2 mm. In some places laitakarite is graphically intergrown with fine-grained chalcopyrite. Chalcopyrite also forms individual grains (diameter under 0.5 mm). Native bismuth occurs both as individual grains (diameter under 0.5 mm) and as a graphic intergrowth in laitakarite. Sphalerite, molybdenite and pyrite occur as individual grains measuring up to 0.5 mm in diameter. Silver has been found as individual grains, the largest being 1 mm in diameter. It has also been found as a thin (0.05 mm) lamella in laitakarite. (Spectrographically the silver has been found to contain some Bi). Galena occurs as inclusions in laitakarite and has been found only as very small (<0.1 mm) allotriomorfic grains. Besides the minerals mentioned, it is suspected that there are also some grains of bismuthinite, Au-telluride and Bi-tetrahedrite in the veins. Owing to their small grain size and scarsity, it has not been possible to identify these minerals with certainty.

The Bi-Se-minerals occur at a distance of only a few meters from the actual Orijärvi ore. Se-minerals, silver, galena and sphalerite usually belong in the hydrothermal formations in siliceous rocks, while bismuth, molybdenite, chalcopyrite and pyrite are more »ubiquitous» minerals, which may be formed already at the pneumatolytic stage. It seems evident that the main formations of these veins occured at the high hydrothermal stage and at a lower temperature than the main sulphide ore of Orijärvi.

PHYSICAL PROPERTIES

Laitakarite forms plates and sheets with a foliated structure owing to an exceedingly easy cleavage parallel to the plane of the sheets. The colour is galena white with a high metallic lustre. Old surfaces are lead grey. The folia are soft, flexible and inelastic. The specific gravity is 7.93. This value represents the average of six determinations with a range of ±0.08. The chemical analysis shows the material to be impure. As impurities there occur galena, chalcopyrite, sphalerite and silicates (Table 3). When these impurities are taken into consideration, the corrected value would be 8.12.

On a polished surface the colour is white. Compared with native bismuth the colour is greyish. The mineral is moderately anisotropic with polarization colours as follows:

Nicols completely crossed: dark grey, slightly brownish

Nicols not completely crossed: clearly brown.

The grinding hardness is greater than that of native bismuth and less than that of galena.

The Vickers microhardness (Measured with Reicherts Mikrohärteprüfer) of laitakarite and, for the sake of comparison, those of native bismuth, galena and tetrahedrite were measured: [1]

[1] The measurements have been made with a load of 50 gr.

Laitakarite	Mh =	50 kg/mm^2
Bismuth	Mh =	15 kg/mm^2
Galena	Mh =	80 kg/mm^2
Tetrahedrite	Mh =	310 kg/mm^2

The reflectivity in air determined with a photometer ocular (Leitz, Mikroskop Photometer), is,

green	48 %
orange	46 %
red	46 %

and determined with a photoelectric cell it is 52 % (white light). Galena with the reflectivity value of 43.8 % given by Moses (Ramdohr 1955, p. 495) has been used as the basis in this determination.

The etch-reactions gave the following results

1: 1	HNO_3	quickly etches dark grey
1: 1	HCl	negative
20 %	KCN	negative
20 %	$FeCl_3$	stains blue grey
40 %	KOH	negative
5 %	$HgCl_2$	negative
30 %	H_2O_2	negative

These results differ only slightly from those reported by Thompson (1949) concerning the minerals of the tetradymite group.

X-RAY INVESTIGATION

The powder patterns were recorded with a Debye-Sherrer camera (diameter 114.6 mm) using filtered CuKα radiation (CuKα_1 = 1.54050 Å, CuKα_2= 1.54434 Å and CuKα = 1.5418 Å). The data for the inteplanar spacings-were calibrated with a silicon standard. The results for laitakarite are given in Table 1. The reflections with asterisks were recorded with a Debye-Sherrer camera, with a diameter of 57.3 mm, without silicon standard, because some of these reflections change to meet the reflections of silicon, and some are too weak to be recorded with a 114.6 mm camera. In column 2 the calculated spacings, the estimated intensities, the rhombohedral (pqr)-indices and the hexagonal (hikl)-indices are represented. On the right side of the table the spacings and intensities of joseite (Bi_4 (Te, S)$_3$) are represented.

The structural identity of laitakarite with joseite is evident. There are some slight differences in intensities between these two minerals. Also there are some reflections in joseite, especially in the back-reflection area, which are missing in laitakarite and vice versa, but as a whole the structural identity is clear.

The single crystal photographs were also taken with a Weissenberg camera. The crystal was rotated about the hexagonal axis [1100]. The Weissenberg photographs (equi-inclination method) were taken of the zero, first and second layer lines.

The crystal symmetry class proved to be D_{3d}—$\bar{3}$m. There are no systematic extinctions beyond the rhombohedral conditions. Only planes with (—h+k+l) = 3 n reflected. The possible space group is R$\bar{3}$m, R32 or R3m.

c_0 is determined from (0001)-reflections using the method of Taylor and Floyd (1950) plotting c_0 against $1/2(\frac{\cos^2\theta}{\sin\theta} + \frac{\cos^2\theta}{\theta})$ and extrapolating to zero value of the abovementioned function. The reflections used were those in which l = 27, 30, 33, 39, 42, 48 and 51. 2θ varies between 63.7°—159.5°. The extrapolation curve proved to be a straight line and c_0 = 39.93 Å ± 0.02 Å.

$a_0$has been calculated from the zero layer line Weissenberg photograph as an average of reflections in the back reflection area. 2θ was between 127.4°—158.6° and a_0 turned out to be 4.225 Å ± 0.002 Å.

The corresponding rhombohedral unit-cell dimensions are: a_{rh} = 13.53 Å ±0.007 Å, α = 17° 58′±04′ and V_{rh} = 205.8 Å^3±0.3 Å.

As seen in Table 1, this new mineral is isostructural with joseite. Therefore Peacock's presumption concerning the structure of joseite can be extended to apply to laitakarite. Peacock writes (1941, p. 92): »In spite of shortgage in Te and S, the ideal number of atoms in the rhombohedral cell is clearly seven. Tellurbismuth (Bi_2Te_3) and terradymite (Bi_2Te_2S) contain five atoms in the rhombohedral cell; these lie on vertical rows, giving a 15-layer structure (Harker, 1934). The natural presumption that joseite has a similar arrangement of seven atoms, giving a 21-layer structure, is supported by a rough estimate of the resulting length of the vertical period. If the layers are spaced as in tetradymite the joseite structure would have c_0 = 38.1 Å. The presumed kind of arrangement is also indicated by the fact that most of the strong diffractions have l-indices (hexagonal notation) which are divisible by seven.» As seen in Table 1 laitakarite has the same kind of variation in the intensities of reflections as joseite.

In Table 2 the unit-cell dimensions and space groups of the minerals of the tetradymite group (together with rhombohedral bismuth selenides) are represented. As seen, the space group is in every species probably the same; R$\bar{3}$m. a_0-axes are comparable, and c_0-axis is in every species a multiple of 5.5 Å.

Table 1. Interplanar Spacings of Laitakarite and Joseite

Laitakarite	Joseite
From Orijärvi mine SW-Finland	From British Columbia (Peacock 1941)
a_o 4.225 Å ± 0.002 Å	a_o 4.25 Å
c_o 39.93 Å ± 0.02 Å	c_o 39.77 Å
a_{rh} 13.53 Å ± 0.007 Å	a_{rh} 13.48 Å
α 17°58′ ± 0°04′	α 18°08′
V_{rh} 205.8 Å³ ± 0.3 Å³	V_{rh} 207.2 Å³

d_{obs}	d_{calc}	I_{obs}	(pqr)	(hikl)	d_{obs}	I_{obs}
6.66	6.665	vw	(222)	(0006)		
4.425	4.437	ms	(333)	(0009)	4.39	m
3.586	5.599	ms	(110)	$(01\bar{1}2)$	3.62	m
3.434	3.435	vvw	(211)	$(10\bar{1}4)$		
3.319	3.327	m	(444)	(000, 12)	3.31	w
3.072	3.080	vvs	(322)	$(10\bar{1}7)$	3.08	vvs
2.573	2.577	w	(443)	$(01\bar{1}, 11)$	2.58	m
2.246	2.251	vs	(554)	$(01\bar{1}, 14)$	2.24	s
2.112	2.113	vs	$(10\bar{1})$	$(11\bar{2}0)$	2.11	s
2.058	2.062	w	(655)	$(10\bar{1}, 16)$	2.05	w
			(665)	$(01\bar{1}, 17)$	1.971	w
1.900	1.901	m	(777)	(000, 21)	1.898	m
1.821	1.822	vw	(766)	$(10\bar{1}, 19)$	1.823	vw
1.781	1.783	vw	(543)	$(11\bar{2}, 12)$	1.783	vw
1.741	1.748	s	(331)	$(02\bar{2}, 7)$	1.747	m
1.655^{+}	1.662	vvw	(888)	(000, 24)	1.657	w
			(877)	$(10\bar{1}, 22)$	1.619	vw
1.538	1.544	ms	(644)	$(20\bar{2}, 14)$	1.540	m
1.475^{+}	1.479	vvw	(009)	(000, 27)		
1.410	1.413	ms	(876)	$(11\bar{2}, 21)$	1.412	m
1.341	1.344	ms	(421)	$(21\bar{3}7)$	1.348	s
1.330	1.329	vw	(10, 9, 9)	$(10\bar{1}, 28)$		
			(9, 8, 7)	$(11\bar{2}, 24)$	1.305	vw
1.247^{+}	1.249	mw	(653)	$(12\bar{3}, 14)$	1.248	m
1.220^{+}	1.220	vvw } d	$(2\bar{1}\bar{1})$	$(30\bar{3}0)$	1.223	vw
1.213	1.209	w	(754)	$(21\bar{3}, 16)$	1.212	m
1.175^{+}	1.176	vvw	(522)	$(30\bar{3}9)$	1.178	vw
1.125^{+}	1.125	vw	(10, 10, 8)	$(02\bar{2}, 28)$		
1.055^{+}	1.056	w(d)	$(20\bar{2})$	$(22\bar{4}0)$		
			(12, 11, 10)	$(11\bar{2}, 33)$	1.046	vw
1.027^{+}	1.027	w	(966)	$(30\bar{3}, 21)$	1.028	w
0.996^{+}	0.999	w(d)	(430)	$(13\bar{4}7)$	1.007	m
	0.995		(14, 11, 10)	$(31\bar{4}, 35)$		
			(13, 12, 11)	$(11\bar{2}, 36)$	0.984	w
0.957^{+}	0.956	w(d)	(743)	$(31\bar{4}, 14)$	0.958	w
0.943^{+}	0.941	vw	(11, 8, 8)	$(30\bar{3}, 27)$	0.945	w
0.924^{+}	0.923	vw(d)	(9, 7, 5)	$(22\bar{4}, 21)$		
0.900^{+}	0.903	vvw(d)	(511)	$(40\bar{4}7)$	0.909	m
0.827^{+}	0.830	w(d)	(520)	$(32\bar{5}7)$	0.838	m
	0.827		(11, 10, 7)	$(13\bar{4}, 28)$		
			(16, 15, 14)	$(11\bar{2}, 45)$	0.818	m
0.796^{+}	0.795	w(d)	(17, 16, 16)	$(10\bar{1}, 49)$		

Table 2. Comparison of Crystal Data of Laitakarite and the Minerals of the Tetradymite Group

		ao	co	co/ao	arh	α	Space group
Laitakarite	$Bi_4 (Se,S)_3$	4.225	39.93	9.451	13.53	17°58′	R$\bar{3}$m, R32, R3m
Joseite A [1]........	$Bi_{4+x}Te_{1-x}S_2$	4.25	39.77	9.36	13.48	18°08′	R$\bar{3}$m
Joseite B [2]........	$Bi_{4+x}(Te,Se)_{2-x}S$	4.34	40.83	9.41	13.84	18°02′	R$\bar{3}$m
Paraguanajuatite [3]..	$Bi_2 (Se,S)_3$	4.08	54.7	13.41	—	—	
Syntetic [4]	Bi_2Se_3	4.125	28.56	6.92	—	—	R$\bar{3}$m
Tetradynamite [5] ...	Bi_2Te_2S	4.22	29.49	6.99	10.13	24°02.5′	R$\bar{3}$m
Tellurbismuth [6]	Bi_2Te_3	4.384	30.45	6.946	10.46	24°11.5′	R$\bar{3}$m
Wehrlite [7].........	$Bi_{2+x}Te_{3-x}$	4.53	29.91	6.75	10.29	24°51′	R$\bar{3}$m
Hedleyte [8].........	Bi_7Te_3	4.47	119.0	26.6	39.76	6°26′	R$\bar{3}$m

[1] [2] Peacock (1941).
[3] [4] Strunz (1957).
[5] [6] [7] [8] Thompson (1949).

CHEMICAL DATA

A chemical analysis was made by Mr. Paavo Väänänen (Geological Survey). Results are given in Table 3.

Table 3. Chemical Analysis of Laitakarite from Finland. Analyst: Päävo Väänänen

	Wt-%	Atom relations	PbS	$CuFeS_2$	ZnS	$Bi_4 (Se, S)_3$
Bi [1]	78.28	0.3746	—	—	—	0.3746
Pb	0.78	0.0038	0.0038	—	—	—
Ag	0.71	—	—	—	—	—
Cu	0.26	0.0041	—	0.0041	—	—
Zn	0.14	0.0021	—	—	0.0021	—
Se	15.50	0.1963	—	—	—	0.1963
S	3.28	0.1023	0.0038	0.0082	0.0021	0.0882
undissolved	0.93	—	—	—	—	—
	99.88					

[1] A Spectrographical analysis gave: Bi, Se, S, Pb, Ag, Cu, Zn, Fe, Si, Al, Mg and traces of Te. (Mr. Arvo Löfgren).

The analysed material contained less than 4 per cent impurities, namely galena, calcopyrite sphalerite and silicates. In the table the impurities have been subtracted from the total and the remainder has been calculated as laitakarite.

The specific gravity (7.93 or corrected value 8.12) and the volume (205.8 Å) of the rhombohedral unit-cell have been presented previously. By means of these values the molecular weight of the unit-cells are 902.6 and

1006 with specific gravities 7.93 and 8.12 respectively. Based on these values the atom content of the unit-cell can be calculated (Table 4).

Table 4. The Content of the Unit-Cell of Laitakarite

	Atom relations	Weight per cent	Weight per cent (corrected for impurities)	Atom relations	The number of atoms in unit cell with M = 982.6	The number of atoms in unit cell with M = 1006
Bi	0.3746	78.28	81.03	0.3877	3.81	3.90
Se	0.1963	15.50	16.04	0.2031	2.00	2.04
S	0.0882	2.83	2.93	0.0913	0.90	0.92
Impurities ..		3.27	—		—	—
		99.88	100.00		6.71	6.86

It is clear that there are in the unit-cell 7 atoms: 4 Bi-, 2 Se- and 1 S-atoms. The chemical formula is

$$Bi_4Se_2S$$
$$\text{or } Bi_4(Se, S)_3$$

As seen, this mineral is almost completely devoid of Te. The corresponding Te-mineral joseite is usually lacking in Se. Se-contents have been reported only in joseites from Brazil. In these joseites the highest Se-content is less than 2 per cent.

NOMENCLATURE

The existence of a natural selenium mineral corresponding in composition to Bi_4Se_2S is proposed.

The mineral is named in honor of Professor Aarne Laitakari, director of the Geological Survey of Finland, in recognition of Professor Laitakari's many accomplishments in the field of mineralogy.

Acknowledgments — The author wishes to thank Dr. K. J. Neuvonen of the Geological Survey of Finland for his abundant advice during this work. Thanks are also due to Mr. L. Hyvärinen, M. A., for determing the microhardness and the reflectivity.

To Mr. Paavo Väänänen, M. A., and Mr. Arvo Löfgren, M. A., of the Geological Survey, the author is idebted for the chemical analyses. The author also wishes to express his sincere thanks to Mr. Paul Sjöblom, M. A. for correcting the English of this manuscript.

APPENDIX [1]

After the manuscript of this paper was completed, the author's attention was called to the Mineralogical Journal (published by the Mineralogical Society of Japan) Vol. 2, No. 6, Aug., 1959, which contains an article written

[1] Received March 4, 1960.

by A. Kato on a new mineral, ikunolite, closely resembling laitakarite. According to a chemical analysis the formula of ikunolite is $Bi_{3.79}S_{2.76}Se_{0.24}$, which is close to $Bi_4(S, Se)_3$ with S : Se = 92 : 8. The mineral belongs to the rhombohedral system with $a_{rh} = 13.28$ Å, $\alpha = 18°00'$ and Z = 1. The space group is $R\bar{3}m$. Ikunolite resembles joseite A and laitakarite in structure, crystallography and physical properties.

A. Kato suggests in his paper that a diadochy between sulphur and selenium probably takes place in ikunolite. Ikunolite contains about two per cent selenium. Laitakarite contains about 16 per cent. Accordingly, it is appropriate to name the selenium-rich end member of the mineral $Bi_4(Se, S)_3$ as laitakarite and the sulphur-rich member as ikunolite.

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REFERENCES

DONNAY, J. D. H. and NOWACHI, W. (1954) Crystal Data. Geol. Soc. Amer. Memoir. 60.

EARLY, J. W. (1950) Description and synthesis of the selenide minerals. Amer. mineralogist 35, 337.

ESKOLA, PENTTI (1914) On the petrology of the Orijärvi region in Southwestern Finland. Bull. Comm. géol. Finlande 40.

FRONDEL, CLIFFORD (1940) Redefinition of Tellurobismuthite and Vandestite. Am. J. Sci. 238. 880.

GARRIDO, J. and R. FEO (1938) Sur les sulfotellurides de bismuth. Bull. Soc. Franc. Min. 40.

HARCOURT, G. ALAN (1942) X-ray Patterns of Ore Minerals. Am. mineralogist 27, p. 63.

HARKER, D. (1934) The crystal structure of the mineral tetradymite. Zeits. Krist. A Vol. 89, pp. 175—181.

LAITAKARI, AARNE (1934) Über Orijärvi Bi-Mineralien. Suomen kemistilehti 9, pp. 90—92.

NEUMAN, H. and SELLEVOLL, M. A. (1955) X-ray powder patterns for mineral identification. Norsk. Vid. Ak. Mat. Nat. kl. N:o 3.

PALACHE, BERMAN and FRONDEL (1944) The System of Mineralogy, New York.

PEACOCK, M. A. (1941) On joseite, gruenlingite, oruetite. Univ. Toronto. Studies. Geol. Ser. 46, pp. 83—105.

RAMDOHR, PAUL (1955) Die Erzmineralien und ihre Verwachsungen. Berlin.

STRUN., HUGO (1957) Mineralogische Tabellen. Leipzig.

TAYLOR, A. and FLOYD, R. W. (1950) Precision Measurements of Lattice Parameters of Non-Cubic Crustals. Acta Cryst. 3, 285.

THOMPSON, R. M. (1949) The telluride minerals and their occurrence in Canada. Am. mineralogist 34, pp. 342—382.

UYTENBOGAARDT, W. (1951) Tables for Microscopic Identification of Ore Minerals. Princeton.

VORMA, A. (1959) Laitakariitti, Uusi Bi-Se-mineraali Orijärveltä. Geologi N:o 2. Helsinki.

2

ON LINEATION IN GNEISSES AND SCHISTS [1]

BY

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ABSTRACT

The following discussion in the formation of linear structures in metamorphic rocks is an attempt to explain the mechanics of the formation of orientated patterns in high metamorphic rocks and their use in field work.

INTRODUCTION

The author believes that lineation, like schistosity in general, is essentially the result of differential internal transport in connection with rock failure. The expression, differential transport, means that the movement is the sum of a great number of small dislocations, which together may produce a large-scale displacement. This process cannot as yet be treated mathematically. For practical purposes the most satisfactory presentation of the process of rock failure is still Mohrs' hypothesis, which, in this simplified form, as used in the following pages, gives a perspicuous picture of the conditions in rock deformation, even if the presentation is not mathematically exact.

The reason for presenting this paper is the obvious reluctance of many field geologists to use linear structures for tectonic interpretations, in spite of the excellent treatment of the subject for example, Sander, E. Cloos, Weiss, Metz and others.

MECHANICS OF THE FORMATION OF LINEAR STRUCTURES

Ernst Cloos has presented a comprehensive discussion of linear structures observed in rocks (1946), with a list of works on the subject published up to 1945. An excellent up-to-date treatment of the theory of the formation

[1] Received September 15, 1959.

of orientated structures is found in Fairbarn & Chayes book on Petrofabrics (1954). The present short paper is an attempt to elucidate the tectonic significance of linear structures. The mechanical aspect of the formation is somewhat overemphasized, because it is easier to evaluate it in the field than the still imperfectly understood recrystallization process under directional stress.

Lineation, like schistosity in general, is controlled by the stress conditions in the rock and the internal movement (transport) during its crystallization. A discussion of their origin can therefore be based on our knowledge of rock failure under compressive stress, obtained by laboratory experiments. We know that at a certain compression, two sets of joints are formed in the rock (Mohrs lines) cutting each other along lines perpendicular to the compressive force and parallel to the medium stress component. Progressive compression will produce slip movement along at least one of these sets of joints (shearing).

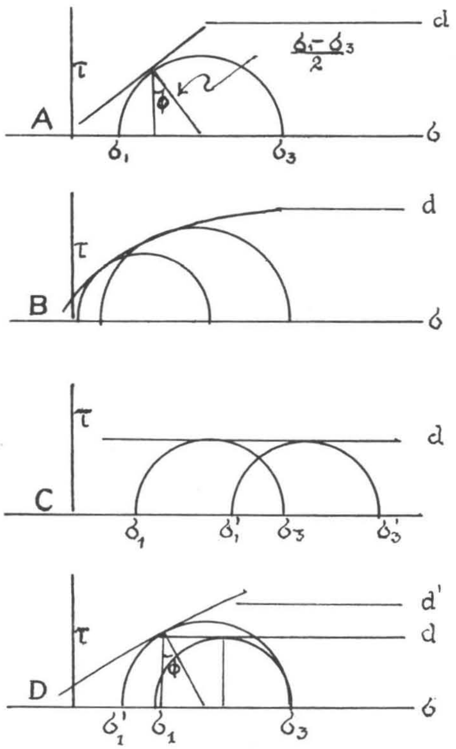

Fig. 1.

A. The stress conditions in the deformation of a rock composed of brittle constituents. The envelope curve of deformation by shearing forms an angle with the curve of flow deformation (d).
τ — tangential stress.
σ_1 — the smallest stress component, σ_3 — the biggests tress component.
ϕ — the angle of internal friction.

B. The stress conditions in the deformation of a rock composed of ductile constituents. The shear envelope curve gradually passes into the flow envelope curve.

C. Stress conditions in progressive flow deformation. The difference between the maximum and minimum stress components stays constant. There is no change in internal friction.

D. The stress conditions in a deformation with the building up of local resistance.
σ_1 and σ_3 — are the stress components in original flow deformation. In progressive deformation σ_1 increases owing to an increase in the internal friction. When it reaches σ'_1 shearing takes place.

This process is illustrated by Mohrs diagrams, (Fig. 1) showing the relation between the shearing stress (τ) and the biggest and the smallest component of the stress perpendicular to the shear plane (σ_1, σ_3). The breaking points of the rock at different stresses is illustrated by the envelope curves of ultimate strength (d). Above this curve the rock will at least temporarily loose some of its cohesivirt because of fracturing. The angle ø is the angle of internal friction. In the case of ideal shearing it is about 45°.

Above a certain stress, solid material loses its shearing strength and starts to flow, which means that the deformation (internal transport) proceeds without any further change of cohesivity (Griggs & Handin, 1942, 1957). The angle of internal friction remains constant; in the case of ideal flow 0°. The curve of deformation is then a line parallel to the transversal stress axes (Fig. 1, C).

Experiments carried out with different types of rocks show that the change from shear to flow takes place in different ways according to the physical properties of the rock. If the material is brittle, consisting of minerals like quartz and feldspar, the envelope curve is a straight line up to the point of flow. The curve of flow forms a sharp angle with the envelope curve of the shear (Fig. 1, A) and the angle of friction is nearly the ideal 45° up to this point. Again, if the rock is plastic, consisting of minerals like calcite, chlorite and mica, the curve will gradually change from oblique to parallel with the transversal axes. The angle of friction gradually approaches 0° (Fig. 1, D). From Diagram 1, A it is also easy to see that an increase in the difference between the biggest and smallest stress component (σ_2—σ_1) (that is, increasing degree of directed stress) will favor the formation of shear, whereas hydrostatic pressure will favor flow.

Deformation in natural rocks is in most cases a mixed process including simultaneously both ruptural and continuous movements (shear-slip and flow). After the rock has broken up along the shear joints the fragments will be displaced by slippage and rotation, because of the lower cohesionty between the fragments in the fractured rock than between the grains in the original rock. This may produce a »cataclastic» flow, each fragment striving to move in the direction of smallest resistance, or generally perpendicular to the main stress component. We can also say that we have flow in the rock as a whole but shearing in the smaller portions of it.

The formation of deformation banding and secondary foliation in tectonites is the result of a rearrangement of the fragments of the rock according to the principle of least resistance. If a rock consists of a mixture of brittle and ductile material, such as, for example, feldspar, quartz and chlorite, the internal friction at cataclastic flow will be greater if these components are in random distribution than if the plastic and brittle components are arranged in different layers. In the latter case the movement will take place essentially along the plastic layers. The brittle portion of the rock will stay as immobile as possible. It might also be pointed out that the same principle would apply if the rock contains a certain portion of molten material.

The molten portion would tend to be segregated in separate bands. Schmidt's hypothesis (1932) of metamorphic differentation because of differences in the physical properties of the mineral components is based on a similar way of thinking.

Recrystallization plays, of course, an important role in the formation of tectonites. Brittle minerals, like quartz and feldspar tend to crystallize in protected positions (stress minima), increasing locally the shearing strength. Formation of chlorite and mica will have the opposite effect.

Rock deformation is a complicated process with alternation of shear and flow and an intermittent change of stress pattern in the rock (Fig. 1, D). The deformation can be visualized as a hold and let go process.

The structure of a tectonite shows traces of both shear and flow movement. Under the microscope we can usually see converging structural planes. Two positions usually dominate. One of them represents the plane of the differential flow (schistosity). It mostly displays itself rather as a lining up of mineral grains than as a pronounced slip plane. The other is a shear slip, situated obliquely in relation to the first-named and in a position of around 40 degrees to the direction of compression. Both the structural directions can be seen in Fig. 2, an Alpine augen-gneiss from Splügen in Switzerland.

The converging structural planes can be studied also in artificially deformed rocks. The Department of Geological Sciences of McGill University has a collection of micro-photographs taken by Frank Adams during his

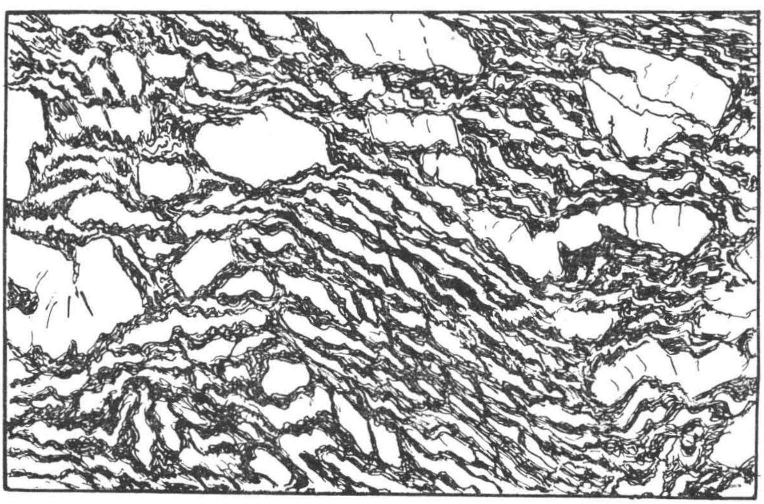

Fig. 2. Augen gneiss, Splügen, Switzerland. White = feldspar and quartz, Dark = mica and chlorite. Natural size.

classic experiments in rock deformation (Adams, 1917). Many of them lend themselves admirably to the illustration of the points discussed in the foregoing. (Fig. 1—2, Plate I.)

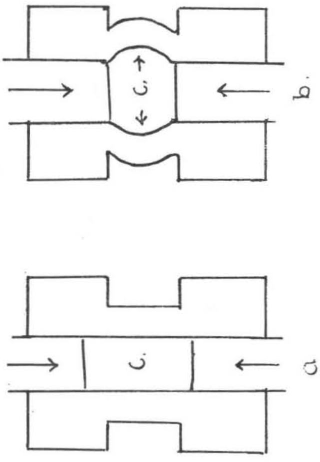

Fig. 3. Steel cylinder used in the experiments of F. Adams. a-cylinder and test core before, b-after the test.

Fig. 3 shows the cross-section of the type of steel cylinders used in the experiments. The thinner ring on the cylinder permits a certain amount of cataclastic flow perpendicular to the direction of compression. At the yield point a differential movement takes place in the rock cylinder used. Fig. 1—2, Plate I, show a norite from Sudbury before and after the test. The deformation structure in Fig. 2, Plate I, shows a close resemblance to the structure in many flaser gneisses, of the type shown in Fig. 2. The dominating feature is the lining up of dark and light components in the plane of flow or in the plane of internal transport. Shear joints dominantly sloping to the left in the picture, are well developed, particularly in the feldspar bands. The direction of movement was from left to right. The second set of shear joints is only indistinctly developed. The shearing oblique to the plane of flow is obviously the result of a packing of material against the more resistant portions of the rock.

Such a packing will produce a b-lineation. The linear structure is partly the result of the intersection of two or several slip and shear slip planes, partly rotation of mineral grains. Another kind of b-lineation, often found in metamorphic schists, is produced by microfolding and crenulation. The deformation is practically continuous, with only very small slip movements. There was, however, a continuous change in the stress pattern; we cannot speak about flow deformation.

Formation of a b-lineation is a retarding process during the internal transport in the rock, causing intermittent changes of the stress conditions and also of the position of the main stress axes.

a-lineation or the lineation subparallel to the direction of transport [1], forms where shearing plays a minor role or none at all. The stress conditions remain fairly unchanged during the internal transport. The most extreme case is viscose flow, forming flow banding. In solid material formation of a-lineation is a high degree of streamlining of the material. We might use

[1] The deviation of the a- and b-lineations from the ideal position relative to the movement, which actually always occurs, will not be discussed in this paper. The reader referred to, for instance, Weiss' discussion of the subject (1956). Mostly the deviation may be overlooked for practical purposes in the tectonic analyses on a larger scale.

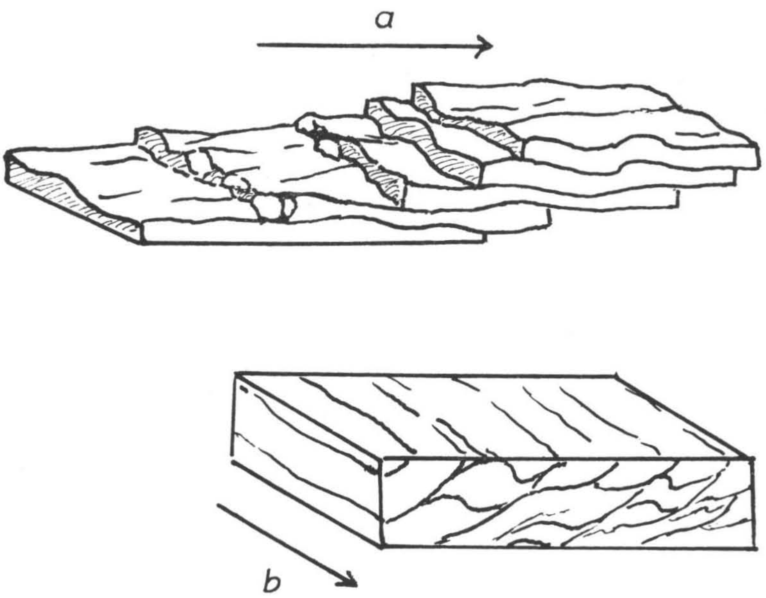

Fig. 4. The most common types of a- and b-lineation in gneisses and schists.

as a comparison the formation drumlinoids under an ice flow, as a contrast to marginal morains. It is fairly obvious that such a deformation will often produce an elongation or stretching in the direction of transport and the stress conditions may intermittently correspond to an extension.

We assume the formation of an a-lineation if the internal transport in the rock takes place without any build-up of temporary resistance. The movement in the rock under such conditions may be accelerated. The Mohrs diagram will be of the type shown in Fig. 1 C.

A schematic illustration of a- and b-lineations, of the types most common in crystalline rocks, is shown in Fig. 4.

FOLDING AND LINEATION

Folding is the manifestation of rock deformation most often mentioned in geological field reports. As a rule the formation of folds during the process of metamorphism is associated with the formation of lineation. We can distinguish between two genetically different types of folding, namely: 1. ordinary folding consisting of the bending of the layers (symmetrical folds, flexure folds) and 2. axial plane slip folding (shear folds and flow folds). The first type of folding includes extension, compression or torsion; and its formation therefore requires a considerable amount of energy. The second type is formed by a simple gliding and takes place with the absorption of relatively considerable less energy. The latter is the most common type of folding in the plastic zone of deformation, as has been pointed out by Goguel (1954). It is also found in soft unmetamorphosed and unconsolidated material (Hubbert, 1939). The extreme case is flow folding in viscose material. We can be fairly safe in assuming that small folding in gneisses and schists, formed during metamorphic conditions in most cases starts out as a shear slip folding, and gradually with increasing compression grades into flow folding, with differential gliding perpendicular to the compression (Kranck, 1959).

As might be expected, lineation in ordinary folds always is a b-lineation.

a-lineation, again is generally found in connection with flow folding, owing to their formation without the build-up of resistance.

Particularly in highly metamorphic gneisses, axial plane slip deformation may produce a b-lineation, if recrystallization takes place contemporaneously to the movement building up local resistance. The common flow folding or wild folding in gneisses therefore often shows good axial lineation perpendicular to the general direction of transport. This lineation may be quite uniform and over large agreas gives rise to surprisingly constant linear structures in spite of an extremely complicated folding pattern in the rock.

STRETCHING

Stretching has often been described in connection with both a-lineations and b-lineations. It is a normal result of the mechanics of flow. On the other hand, stretching generally does not run parallel to the b direction (McIntyre, 1951). In the cases where it does occur there must have been a slip movement parallel to the b-axes (»three-dimensional deformation»). A stretching in the axial direction is often connected with a differential rotation, a process that can be compared to the rolling out of soft clay. Most rodding structures seem to be formed in this way.

A linear orientation of elongated crystals like hornblende is, so far as the author knows, connected as a rule with a b-lineation, and may be the result of a rotational deformation of this type.

Ernst Cloos' well-known paper about deformed oolites (1954) from the Appalachian folding in Pennsylvania describes a special case of stretching subparallel to the direction of transport caused by axial plane slip. The mechanics of it may produce a stretching at an oblique angle to the schistosity. The angle will decrease with increasing intensity of deformation. Some highly stretched conglomerat, as, for example, the Bygden conglomerate in Norway, have been explained in the same way (Strand, 1945).

FIELD RELATIONSHIP

The observations about linear structures in different types of rocks and in different types of tectonic styles generally confirm the theoretical considerations discussed in the foregoing pages.

The classic example of a uniform axial lineation is found in the Penninic zone of the Swiss Alps, where low-angle linear schistosity following the plunge of the folding axes, extends from the French border south of Geneva to the Austrian border in the east. It is the result of the packing of the Penninic nappes against the threshold of the Herscynian massifs of Aar and Aigiulles Rouges. The lineation is particularly well developed in the frontal (northern) parts of the nappes. In the central parts the conditions are more complicated, depending on movements in different directions (Venk, 1955).

Similar conditions are found in the crystalline sections of the Northwest Highlands of Scotland (Wilson, Ramsay, McIntyre and others), in the thrust zones of the Appalachians; in the Caledonian folding in Scandinavia and in Precambrian mountain built zones. The b-lineation is particularly well developed in the marginal parts of big nappes and thrust sheets.

In Precambrian terrain high-grade metamorphic rocks (granulite facies) often show a strong b-lineation in comparatively narrow zones of intense movement. One good example is the southern and western edge of the Morin anorthosite body north of Montreal (Osborn, 1949) where both the anorthosite itself and also the interlayered gneisses show an extremely strong lineation in spite of the fact that the material (quartzite, norite) contains no needly or flaky minerals. The lineation is produced by miniature shearing and microfolding invisible in hand specimens (Fig. 5). The structure is probably caused by low-angle overthrust movements.

Whereas it is easy to list examples of b-lineations in metamorphic rocks, examples of definitely proved a-lineations are not very many. To be sure, we have plenty of a-lineation in soft low grade metamorphosed sediments, but less convincing cases in well-crystallized material. The steep, often vertical, lineation found in connection with tight, isoclinal folding, often seems to belong to this type. True mylonites mostly show a-lineation.

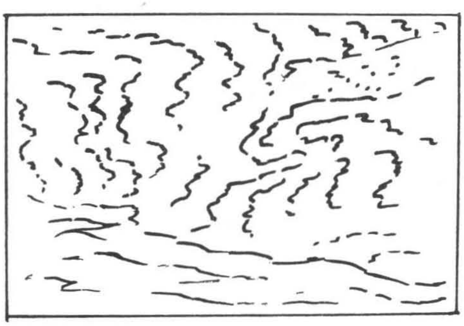

Fig. 5. Microfolding in quartzite with strong b-lineation. St. Jerome. P. Q. Canada. The section is cut perpendicular to the lineation. The structural pattern is formed by contorted «platy quartz.» Magn. 8 +

LINEAR STRUCTURES IN MIGMATITIC GNEISSES

With increasing granitization the linear structures usually become less well developed. This observation is interesting because of the information about rock melting provided by recent experimental work (Tuttle and Bowen, 1958). The fact that quartz and alkali feldspar melt at temperatures slightly over 600° means that a liquid phase may often form at deep level metamorphism.

The progress of melting of solid rocks has been studied in microscopic slides in experiments carried out at McGill university (Kranck and Oja, 1959). The melting begins with the formation of a liquid phase along the grain boundaries between alkali feldspar and quartz. At that stage the cohesivity of the rock will rapidly decrease. Tuttle and Bowen state that up to an amount of about 50 % liquid material there probably would not be any complete homogenization of the rock or separation of melt and solid. The rock would on other words react to directional stress as a solid body with low cohesivity. As the confining pressure at the same time would be high, the type of deformation to be expected is flow, and if lineation occurs, it would be of a-type (unless it is later superimposed). If the amount of liquid material becomes high enough, no lineation will form. As a matter of fact, we find in many migmatic gneisses very vague linear structures. However, we also have to reckon with migmatic gneisses in which the granitic material was injected, mostly in connection with tectonic movements. They form above the zone where partial melting takes place. Here the original rock still has enough strength to produce a good linear structure. Only the granitic portion does not show lineation. Such are the conditions in many veined gneisses in Precambrian areas.

The disappearance of lineation within a gneiss complex thus may indicate a considerable amount of melting in the zone in question. The presence of lineation, on the other hand, obviously proves that the rock was solid at the time of deformation, not necessarily that it never was molten. This can be proved only by the presence of primary supercrustal structures or largescale tectonic structures of a type that does not form in a molten material.

HOW TO DISTINGUISH BETWEEN a- AND b-LINEATION

The field geologist is probably more interested in how to recognize the different kinds of lineation that in the theory of how they form. Lineation is an exceedingly important tool for the interpretation of structures, but it can be used successfully only if we know what kind of a lineation we work with. In many field reports the linear structures are found to be far too

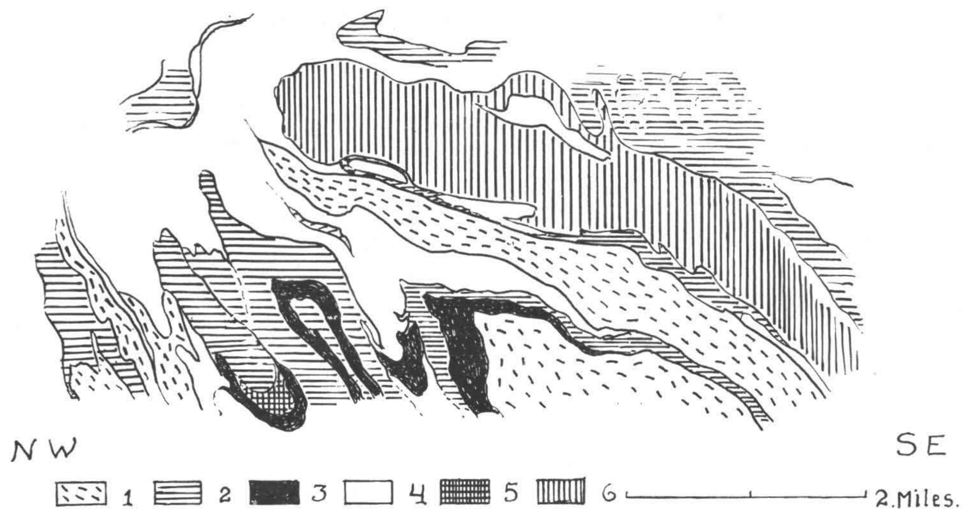

Fig. 6. Structural profile through the central parts of the bedrock of the Clarandon-Dalhousi Map Sheet in southern Ontario. 1 = granite, 2 = limestone, 3 = basic metamorphics, 4 = mica schists, 5 = peridotite, 6 = gabbro.

complicated to be of any practical interest or too difficult to find. These complications are due to the factors explained in this paper, namely the formation of different kinds of lineation in the same rock.

There is today hardly any foolproof way to distinguish between a- and b-lineation, unless we can connect them with larger scale structures which show the direction of transportation in the rock: Some general rules, however, can be pointed out.

b-lineation is, at least in crystalline rocks, more common than a-lineation, and it is safer to regard a lineation as a b-lineation until the contrary has been proved.

Proceeding from the mode of formation discussed in the foregoing, b-lineation is mostly large dimensional (occurring in more coarse-grained rocks) as compared with a-lineation. If a- and b-lineation occur together in the same area, we will probably find that the b-lineation is of a lower angle and more strongly developed.

A promising way to distinguish between different types of lineation is applying petrofabric analyses. Although much progress has been made in this field, the number of systematic studies of field localities is too small to permit definite conclusions concerning the relationship between the type of deformation and the petrofabric diagram. However, the method has already been proved applicable, and will be more so in the future.

The best way to find out if the lineation corresponds to the axial direction is to draw profiles using the lineation as fold axes. Fig. 6 shows, for example, a section through the Clarandon Dalhousie map sheet of Southern Ontario (Smith, 1956). In itself this rough sketch gives convincing proof that the lineation marked on the map is an axial lineation. The tectonic style is the same as has been described by Wegmann (1929) from the Karelian orogeny in eastern Finland and the one found in the Haliburton Bancroft area in Ontario.

REFERENCES

ADAMS, FR. D. and BANCROFT, J. A. (1917) On the amount of internal friction developed in rocks during deformation and on the relative plasticity of different types of rocks. J. Geology 25.

BOWEN, N. L. see Tuttle.

CLOOS, ERNST (1946) Lineation, a ceritical review and annotated bibliography. Geol. Soc. Amer. Mem. 18.

—»— (1947) Oolite deformation in the South Mountains fold, Maryland. Bull. Geol. Soc. Amer. 58.

FAIRBARN, H. W. (1954) Structural petrology of deformed rocks. 2nd ed. Reading, Mass.

GOGUEL, JEAN (1954) Introduction a l'étude méchanique des déformations de l'écorce terrestre. 3e éd. Mém. Carte géologique détaillée de la France.

GRIGGS, D. (1942) Strength and plasticity. Handbook of physical constants. Geol. Soc. Amer. Spec. Papers 36, pp. 107—130.

HANDIN, J. and HAGER, REX V. (1957) Experimental deformation of sedimentary rocks under confining pressure: Tests at room temperature on dry samples. Bull. Amer. Assoc. Petroleum Geologists 41.

HEWITT, D. F. (1955) Geology of Dungannon and Mayo Townships, Ontario Dept. of Mines, Vol. 64, pt. 8.

HUBBERT, M. K. (1951) Mechanical basis for certain familiar geologic structures. Bull. Geol. Soc. Amer.. 62.

KRANCK, E. H. (1957) On folding-movements in the zone of the basement. Geol. Rundschau 46.

KRANCK, E. H. and OJA, R. (1960) Experimental studies on anatexis. To be presented at the XXI International Geol. Congress, Copenhagen.

McINTYRE, D. B. (1950) Note on lineation, boudinage and recumbentfolds in the Struan Flags (Moine), near Dalnacardoch, Pertshire (Scotland). Geol. Mag. 87.

OSBORN, F. F. (1949) Coronite, labradorite, anorthosite and dykes of andesine anorthosite, New Glasgow, P. Q. Trans, Roy. Soc. Canada, 3rd ser., Vol. 43, Sect. 4.

SCHMIDT, W. (1932) Tektonik und Verformungslehre. Berlin.

SMITH, B. L. (1956) Geology of the Clarandon — Dalhousie Area. Ontario Dept Mines, Vol. 65, pt. 7.

STRAND, TRYGVE (1944) Structural petrology of the Bygdin conglomerate. Norsk Geol. Tids. 24.

TUTTLE, O. F. and BOWEN, N. L. (1958) Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$—$KalSi_3O_8$—SiO_2—H_2O. Geol. Soc. Amer. Mem. 74.

WEGMANN, E. (1929) Beispiele tektonischer Analysen des Grundgebirges in Finnland, Bull. Comm. géol. Finlande 87, p. 98—127.

WENK, E. (1955) Eine Strukturkarte der Tessineralpen. Schweiz. Miner. Petr. Mitt., B. 35.

WEIS, L. E. (1958) Fabric analysis of a triclinic tectonite and its bearing on the geometry of flow in rocks. Am. J. Sci., Vol. 253.

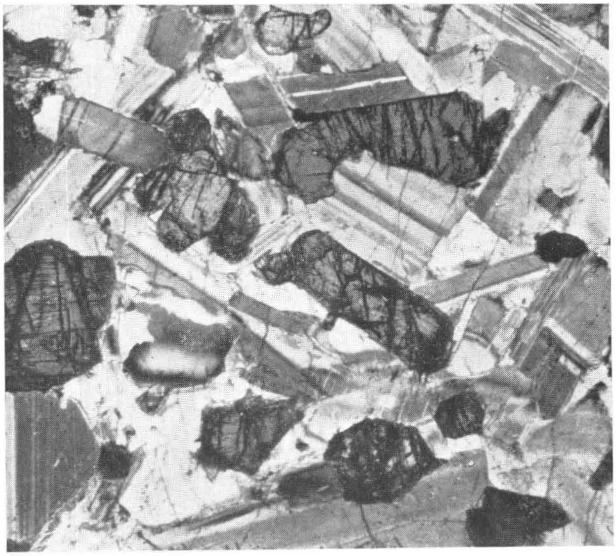

Fig. 1. Undeformed norite from Sudbury. Photograph from Frank Adams Collection, Mc Gill University, Magn. 25 ×.

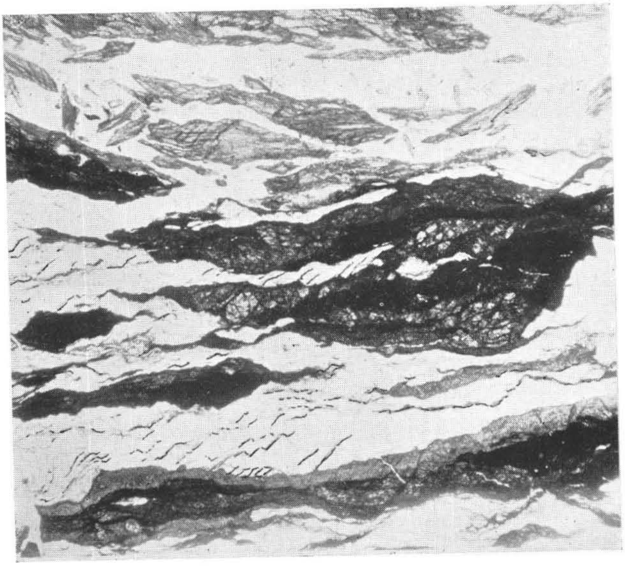

Fig. 2. Norite from Sudbury, artificially deformed. Flow foliation and secondary shear jointing. Photograph from Frank Adams Collection. Mc Gill University, Magn. 25 ×.

E. H. Kranck: On Lineation in Gneisses and Schists

3

RADON MEASUREMENT IN URANIUM PROSPECTING [1]

BY

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ABSTRACT

This is a preliminary report on radon measurements made in connection with uranium prospecting work in the Koli area in North-Karelia. The activated carbon method using a very simple apparatus was employed. A great number of radon anomalies have been found, ranging from 3 to $3000 \cdot 10^{-10}$ curies/liter in intensity. The origin of radon in soil air is discussed and some examples of different types of radon anomalies are presented. A definite connection between a radon anomaly and uranium ore has been established in two cases.

CONTENTS

[1] Received February 2, 1960.

INTRODUCTION

Uranium prospecting was initiated by Atomienergia Oy in 1956. As no economic uranium deposits were known in Finland at that time, an extensive exploration program was started, using air- and carprospecting, by selecting exploration areas on the basis of geological analogies and by encouraging amateur prospectors to look for radioactive minerals. In the fall of 1957 the company received uranium bearing samples from North-Karelia. This discovery led to immediate exploration activity which has already given positive results and is still being continued.

In the uranium exploration work in the area (called the Koli area) conventional methods, radiometric prospecting and radiometric mapping have been used. Magnetic and electromagnetic, as well as seismic measurements, have been applied in some cases. Boulder tracing — an important method in all ore exploration in glaciated areas (SAURAMO, 1924; SAKSELA, 1949; AUROLA, 1955) — has been carried out successfully using scintillometers.

Radon measurement was started in the Koli area in the fall of 1958. Most work was concentrated on the measuring of radon in soil air, but the radon contents of waters, especially springs, have been determined also. The work had a purely practical objective, namely to locate uranium ores, or at least to limit a potential ore field. The interpretation of anomalies is based on field observations. Only calibration and some control measurements have been carried out in the laboratory. Model tests have not been made as yet.

THE KOLI AREA

The area is situated in North-Karelia, on the western shore of Lake Pielinen (Fig. 1). Its direction is from NW to SE; it has a length of about 35 km and an area of about 300 km². Radon measurements have been made in three parts of the area: Riutta in the southern part, Paukkajanvaara in the middle and Herajärven kannas in the northern part.

Fig. 1. Location of the exploration field.

The relief of the Koli area is fairly high (Fig. 2). The highest point, Koli, at the northern edge of the area, has an elevation of 347 m above sea level and the lakes range from 90 to 150 m above sea level. Ridge systems with a NW to SE direction, cut in some places by transverse valleys, are characteristic of the relief. The valleys between the ridges are sometimes swampy, and sometimes covered by lakes, but apart from these valleys the terrain is generally rather

Fig. 2. A part of the Koli area. Herajärven kannas in the foreground. Foto Iffland, SMY.

dry. The bedrock is fairly well exposed by comparison with other parts of Finland, but most of the terrain is covered by ground moraine. In addition, some places in the southern part of the area are covered by gravel.

The bedrock of the Koli area belongs to the marginal zone of the so-called Karelian formation, resting on an older gneissose-granite basement (Frosterus and Wilkman, 1920; Wegman, 1928; Väyrynen, 1933). The rocks belong to the quartzite sequence, dominated by orthoquartzite, but including also mica-quartz-schists, conglomerates and feldspar-quartzites. The quartzites have an average strike of NW—SE 30° W. They are frequently, and in different directions, penetrated by steeply dipping basic rocks (so-called metabasites, or uralite diabases) and by some pegmatite dikes. In places all the rocks of the area are strongly deformed. As a result of the sectional deforming movements blocks of the gneissose-granite basement can be seen within the quartzite sequence in many places.

Uranium mineralization in the rocks of the Koli area seems to be rather common. It usually occurs as fissure fillings and as impregnations in sedimentary rocks of the contact zones of the metabasites. It also extends into the metabasites, especially in their tectonized parts.

The most important uranium mineral is pitchblende, but the brightly colored secondary minerals are also common. The uranium mineralization is, in most cases, monometallic and alteration of the bedrock is small. The only metals following uranium to any extent seem to be iron and vanadium.

EXPERIMENTAL METHOD

INTRODUCTION

Most field measurements of radon have been made with ionisation chambers (Běhounek, 1927; Hatuda, 1954; Alekseev et al. 1957; Budde 1958). A basically different approach to radon measurement is presented by Cadudal (1955) and Pradel (1956). In their method the radon to be measured is adsorbed on activated carbon and the radioactivity in the carbon is counted after the short-lived daughter elements of radon have attained equilibrium.

A comparison of the different methods and the physical background of the active carbon method have been presented by Landström (1960). Essentially the active carbon method is based on the counting of the RaC β-activity with a max. energy of 3.15 Mev. RaC attains equilibrium with radon by an »apparent half-life» of about 40 min. The activity curve has a broad maximum between 3 and 6 hours. In practice no correction is needed, if the counting is done during this time interval.

In Sweden the active carbon method has been applied in the prospecting work of AB Atomenergi (Landström, 1960). As a result of a visit by Mr. Landström to Finland in 1958 the same method was also adopted in this country as a means of prospecting for uranium.

APPARATUS AND PROCEDURE

The apparatus is essentially the same as that introduced by Landström. Only some modifications were made in the routine work.

The samples of soil air were taken from about 70 cm depth using a steel tube with a diameter of 20—24 mm. The lower end was perforated and provided with a sharp bit. After the steel tube, which had to be of very resistant material, had been struck into the ground, the pumping system was connected (Fig. 3). About 10 liters of soil air were pumped through the system using a Frinet laboratory acid pump with improved valves (not shown in the figure). The flow was 1.5 to 2.5 l/min and the air was passed first through a $CaCl_2$ desiccator and then an adsorption tube filled with 100 ml of active carbon. The active carbon used was Norit PK brand, particle size 1 to 3 mm. The carbon was re-used after standing for about one month. In the systematic routine work the sampling was performed by a team of 4—5 persons, taking daily from 100 to 150 samples. The sampling lines or profiles were, if possible, oriented transversely to the strike of the geological structure and placed from 20 to 50 m apart. The distance between individual points was usually 5 m.

Fig. 3. The soil air pumping system.

The carbon samples in stoppered glass bottles were carried to a field laboratory and counted after standing from 3 to 6 hours. A well-type counting cup and Tracerlab scaler were used. Where no electric power was available, a battery-operated scaler (Hoffman Countmaster) had to be relied upon. A slight escape of the adsorbed radon from the bottles was observed, but did not affect the results. An »apparent» half-life of 50 to 60 hours, instead of the theoretical 91.8 hours, was measured. A small percentage of the radon was always lost during the counting of the samples, but this did not immediately affect the γ-active daughter elements.

CALIBRATION AND TESTING OF THE METHOD

The method was calibrated by preparing several solutions with known contents of radium [1]. The solutions were sealed in glass containers. After having attained equilibrium, the radon in the solution was quantitatively pumped into active carbon and the samples counted after the usual 3—4 hours delay.

[1]) The radium was obtained from the Radiochemical Center, Amersham.

The same apparatus was also used in the determination of the radon content of natural waters.

The two counting systems had the following sensitivities:

Tracerlab	19 700	cpm/μ-curie
Countmaster	49 200	»

The unit eman = 10^{-10} curies/liter will be used throughout this paper.

With a shielded background of about 10 cpm the limit of sensitivity, by pumping 10 liters of air and counting 1 to 3 minutes, may be estimated to be about 10^{-10} curies/liter or 1 eman. E.g. counting 3 minutes with the Tracerlab system implies a random counting error of 13 % for a radon content of 1 eman. The sensitivity could no doubt be increased by modifying the counting geometry. On the other hand it was soon realized, that the real and significant anomalies in the area were rather pronounced, from 3 to 1 000 emans and in exceptional cases as high as 3 000 emans.

The short-term and seasonal variations in the radon content of soil air have been measured and their causes discussed by several authors (HATUDA, 1954; BUDDE, 1957; ALEKSEEV et al. 1957). As the present method and its sampling techniques were not accurate enough to distinguish small variations of the »background» level, only some statements with regard to anomalies will be made.

To check our results, several lines were measured twice and certain points several times. As may be seen from Fig. 4, 5 and 7 the general tendency is reasonably well reproduced, although values on single sampling points often vary considerably (e. g. Fig. 4, x = 5100). Actually the sampling of the soil air most probably was the step with the least accuracy, the aforementioned fluctuations in radon concentration and the intake of atmospheric air being the greatest sources of error. Evidently more reliable and reproducible results would have been obtained if a sampling depth of more than 70 cm had been used. The choice was, however, very much a practical one. In the stony moraine of the Koli area sampling at greater depth would have been very difficult and in many places quite impossible by the simple procedure used.

One further remark should be made in this connection. Measurements were sometimes carried out in moraine containing uranium ore boulders, which were found to cause limited, but often intense radon anomalies (Fig. 6). Naturally the reproducibility of measurements of such anomalies with high concentration gradients is poor.

The adsorption of radon on active carbon has been studied and the adsorption coefficient determined by BECKER and STEHBERGER (1929) and by GÜBELI and STAMM-

BACH (1951). At the extremely small concentrations involved the usual Langmuir adsorption equation is reduced to a direct proportionality.

$$\frac{x}{m} = \gamma \cdot p \quad (T = \text{const., equilibrium conditions})$$

where x = amount of adsorbed radon

m = mass of adsorbent

p = partial pressure of radon in the gas phase

γ = adsorption coefficient

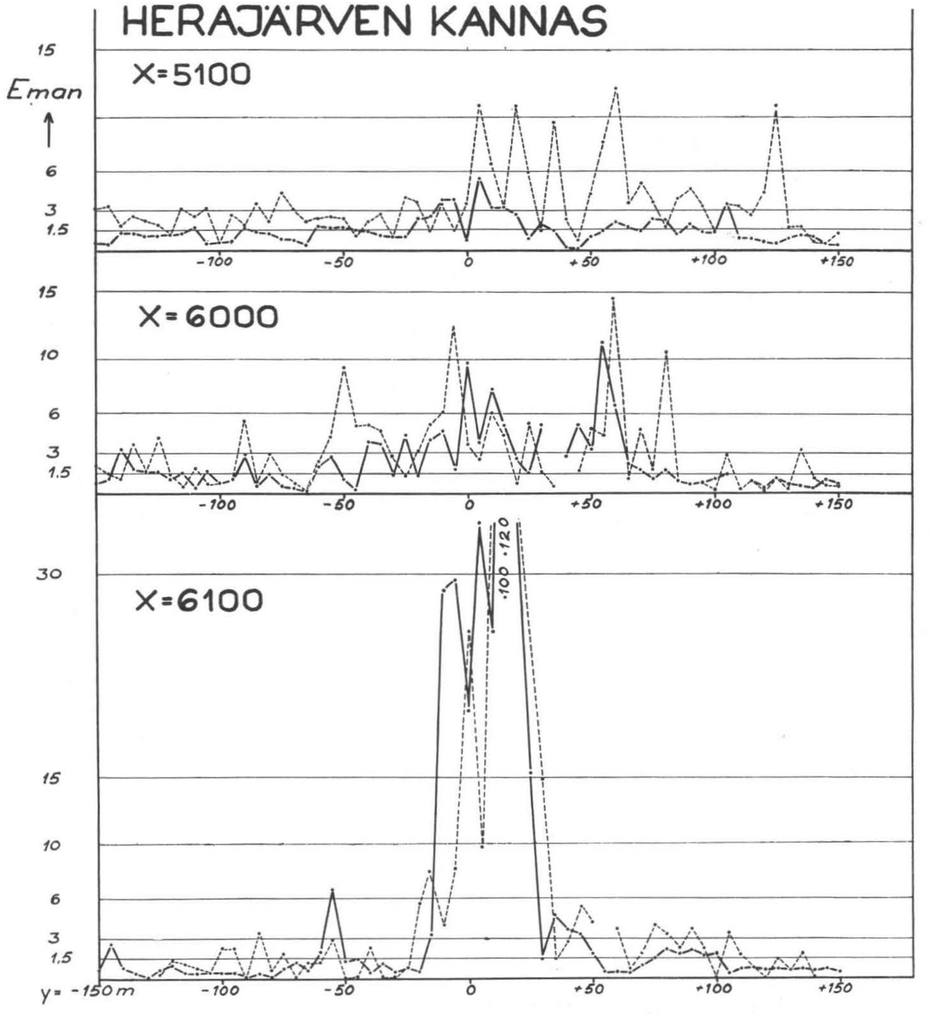

Fig. 4 Typical radon anomaly from the exploration field. Dashed line = measured in winter. Full line = measured in summer.

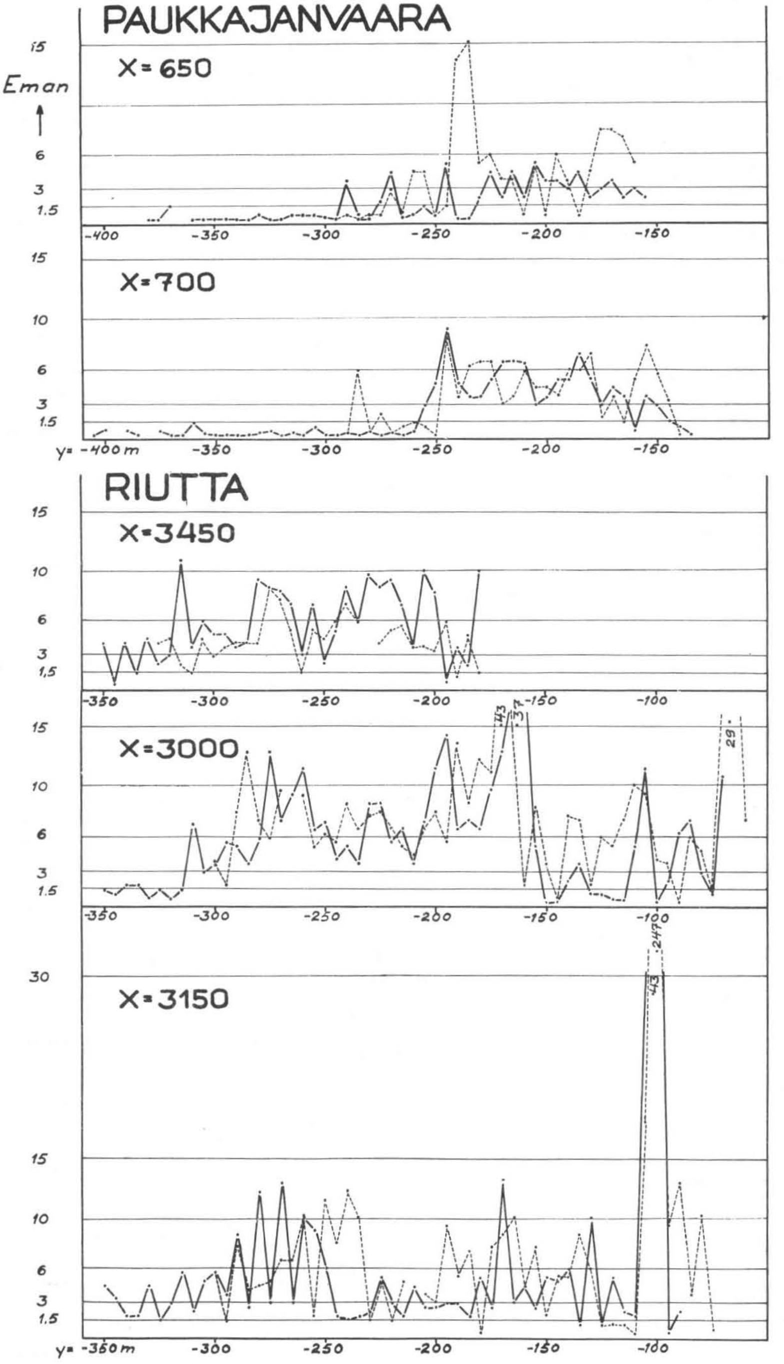

Fig. 5. Typical radon anomaly from the exploration field. Dashed line = measured in winter. Full line = measured in summer.

The adsorption coefficient is strongly temperature dependent, as was observed also during this investigation. A »chromatographic» advancement of the radon front with streaming air could be observed in the carbon column, the velocity of the front at a given rate of air flow depending on temperature.

The completeness of adsorption in the present system was tested by connecting several adsorption tubes in series. The first tests in the field, as well as a considerable number of the routine measurements, were carried out in winter. More than 98 % adsorption on the first carbon was observed with as much as 3 000 emans at a temperature of —5°C.

Later several laboratory measurements were carried out at different temperatures with 0.1—0.4 μ-curies of radon. They showed that with new and dry carbon the adsorption of radon was practically complete (>99 %), in any case, below 20°C. Prolonged usage and especially moisture were observed to decrease markedly the adsorptive capacity of the carbon, causing serious radon leakage at higher temperatures, but the capacity could be restored by drying at 120°C.

RESULTS

GENERAL REMARKS

The immediate radon source is radium (Ra^{226}). Most of the radium is fixed in uranium minerals, where it is generated by the decay of uranium, but a small part of it may be leached out by water and redeposited in soils and sediments. Although the redeposited amounts are small, their emanating power is relatively high and they may cause strong anomalies far from the related uranium deposit. The emanating power of different uranium minerals varies greatly (GILETTI and KULP, 1955). Probably by far the greater part of the radon in soil air originates in the weathered secondary minerals on the surfaces of the outcrops, in fissures, or in an aureol of redeposited radium.

The migration of radon from its sources may occur by diffusion, and/or by convection involving air or water. The diffusion constant of radon in various soils has been measured and theoretical models have been calculated by ALEKSEEV et al. (1957) and BUDDE (1958). The calculations, which have been partly confirmed experimentally, indicate that diffusion is very dependent on the porosity and humidity of the soil. Five centimeters of clay is practically impermeable to radon, but even in more porous soils, purely diffusive transportation takes place only over some meters distance.

Applications of radon measurement to uranium prospecting have been reported by PRADEL (1956) ALEKSEEV et al. (1957) and LANDSTRÖM (1960). Measurements have also been made in the vicinity of known uranium deposits.

In the Koli area a great number of radon anomalies were found and the problem was to find the correct interpretation of these anomalies. Typical examples are presented in Fig. 4 and 5. The anomalies have been broadly divided into two classes:

1. Peak anomalies, usually elongated narrow local anomalies of more than about 30 emans. The most intense examples are about 3 000 emans.
2. Small extensive anomalies with a rather constant radon content, normally 3 to 10 emans. They often accompany a peak anomaly, but may also occur separately.

In some cases the origins of the anomalies have been deduced, but other cases remain still to be clarified. According to our field observations, the following factors evidently may be causes of radon anomalies:

1. A nearby uranium-radium mineralization, or an associated radio-active aureol (redeposited radium).
2. Radon flowing out of fissures in the rock.
3. Uranium bearing boulders in the moraine.
4. Radon dissolved in ground water which rises to the surface in springs.
5. Radium redeposited from ground water, for instance in the neighbourhood of springs.
6. Topography; soil layers with different permeability etc.

Approximately the same causes are also given by ALEKSEEV et al. (1957) and others. Naturally in most cases several factors are responsible, but one may dominate. Examples of different types of anomalies will be presented later.

PRESENTATION OF RESULTS

The results of the radon measurements have been presented in the following three ways:

1. By radon profiles along sampling lines.
2. By maps, on which the radon concentrations are marked as numbers or suitable symbols.
3. By taking the average of the radon concentrations of several (usually 5) adjacent points on a profile, and drawing an »smoothed» iso-radon map. Averages of adjacent points are taken for instance at every second point.

The drawing of iso-radon maps based on single radon measurements by the present method is not recommended because the map tends to become too complex and the iso-radon curves usually cannot be reproduced.

FIELD EXAMPLES

1. In the Riutta field an anomaly has been found (Fig. 6), where an exceptional number of uranium ore boulders in the ground moraine is the source of the radon. Two pits were dug in the ground about 20 m apart.

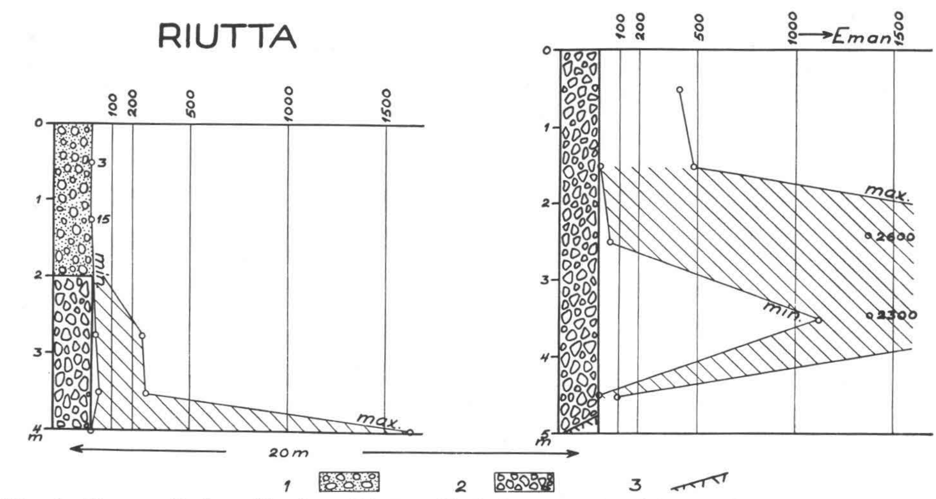

Fig. 6. Two vertical profiles from Riutta. Shaded area = the difference between maximum and minimum radon contents observed at each depth. 1. Gravel 2. Ground moraine 3. Solid rock.

The pit in the right-hand figure was dug to the rock, but the low radon content in the moraine just above the rock indicates that the radon does not originate in fissures of the rock. The left-hand figure shows a section of the other pit, where a barren layer of gravel has been deposited on the uranium bearing moraine. The radon content of the moraine shows great variation, obviously reflecting the irregular distribution of uranium bearing boulders. In the gravel the radon content is relatively small (from 3 to 20 emans), but increases with depth illustrating the restricted range of radon diffusion.

2. In the Myllykorpi field in Herajärven Kannas, high concentrations of radon in soil air and water have been encountered. The soil was 0.5—2.0 m of coarse moraine. Fig. 7 presents a typical radon profile, with a peak anomaly of about 3 000 emans, from this field. Radon was shown to flow from fissures in the rock, which were later exposed, but other possible radon sources should also be considered.

It is interesting to note, that this anomaly was known, before the radon measurements were made, as a γ-anomaly of about 10—20 times background during winter, whereas in summer the γ-anomaly disappeared completely. Obviously in winter the diffusion of radon through the frozen ground is very slow, so that the γ-active daughter elements accumulate. Previous observations of radon accumulation in winter are referred to by FAUL (1954). In summer the radon freely diffuses out of the coarse ground moraine and no γ-activity accumulates. It may also be noted that, so far, no appreciable uranium mineralization has been found in the vicinity of this anomaly. Investigations at this interesting locality will be continued.

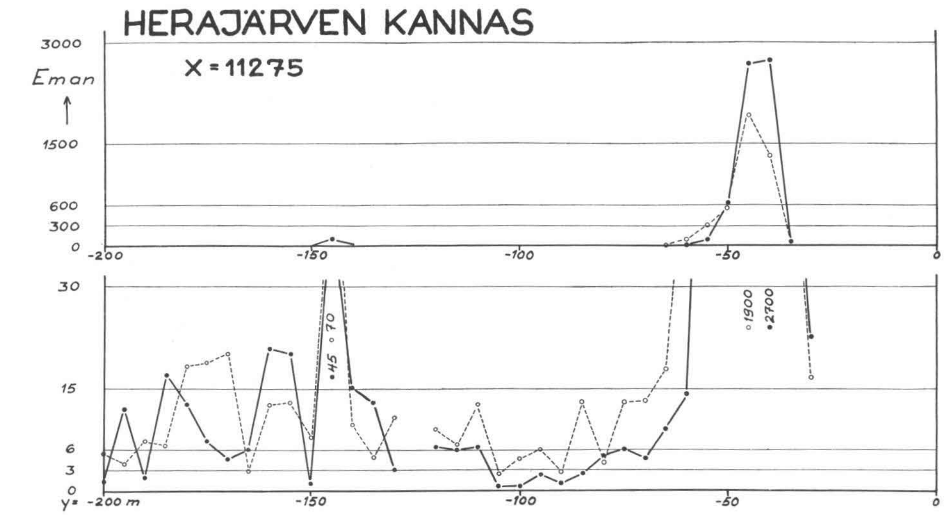

Fig. 7. Radon profile from the northern part of Herajärven Kannas. Dashed line = measured in November 1958. Full line = measured in December 1958.

3. Radon, dissolved in water, is carried through rock fissures by intramontane water or through the soil by ground water. The significance of surface water in radon transportation is small, as radon rapidly escapes from water exposed to the atmosphere. In areas, where the soil air shows an anomalous radon content, the spring waters usually also contain radon. Table 1 presents some values of the radon content of waters of the Koli area.

Table 1

Date of measurement	Number and type of water source	Location	Radon content of water in emans
July 1959	No. 1 pit in ground	Riutta x = 3440	620
» »	No. 2 drill-hole	» x = 3420	860
» »	No. 12 spring	» x = 3200	140
Oct. 1958	No. 65 spring	Paukkajanvaara	3 480 [2]
» »	No. 68 drill-hole	» [1]	1 020
July 1959	No. » »	»	780
» »	No. 82 »	» [1]	700
Oct. 1958	No. 89 spring	Herajärven Kannas x = 11150	690
July 1959	No. 87 drill-hole	» » x = 11300	380
Oct. 1959	No. 120 spring	» » x = 1900	103
Sept. 1959	No. 126 »	» » x = 7450 [3]	580
» »	No. 127 »	» » x = 7350 [3]	210

[1] in the vicinity of the Mårtensson ore, but not intersecting the ore.
[2] very small flow.
[3] shown in Fig. 8.

It may be mentioned, that most of the springs listed in the table were found by scintillometer as a result of their γ-activity. In one case (No. 89) the uranium content of the water was fluorimetrically determined to be 2.6 mg/l. Obviously small amounts of radium are also present and deposited from the water, because a rather high residual radon anomaly was observed in the vicinity of the spring even when the spring dried up.

4. Small extensive anomalies in the Koli area are common, indicating radon contamination of the soil air. In many cases they are caused by radioactive boulder trains etc. However, in some cases small anomalies have been encountered in places where the soil coverage is as much as 30 m of gravel. (e. g. Fig. 5, x = 700 and x = 3450). In one case also other indications give good reasons for assuming that the small anomalies are caused by deeply buried uranium deposits. So far it has not been possible to investigate this case further. In principle, anomalies of this type may be assumed to originate from a limited radon source at some depth, or in most cases from radium redeposited in the soil over large areas.

5. In Herajärven Kannas an elongated uranium mineralization zone has been found, directed NW—SE and dipping 30°—45° to W (Fig. 8 A). The mineralized zone contains several small ore pockets with uranium contents of 0.3 %—0.4 % U. The zone is located in a forested hillside, gently sloping to SW, and the outcrop of the ore is covered by 0.5 to 1.5 m of ground moraine. Down the slope the moraine layer becomes thicker. The radon anomaly is shown by smoothed iso-radon curves in Fig. 8. The highest values (up to 600 emans) were obtained just above the outcrops, which were indeed found by locating the anomalies. Eastwards from the outcrop the radon content of soil air decreases rapidly, but to the west there is a more extensive anomaly. The relatively barren region in the center is located in an elevated area. Peak anomalies in addition to those found above the outcrops also occur at springs on the lower part of the hill. The shape of the anomaly not immediately above the outcrops may be explained as follows:

a. Water carries radon and radium from the known outcrops.
b. Radon diffuses upwards from the mineralization, which continues to the west with a dip of 30° to 45°.
c. There are, as yet undiscovered, mineralizations to the west parallel to the known outcrops.

6. The Mårtensson ore in Paukkajanvaara is situated at the contact between sedimentary rocks and a metabasite dike (Fig. 9). The ore is like a bent sheet and dips steeply to the north. The average U-content of the ore is 0.2 %. The section in Fig. 9 shows the western continuation of the ore. The picture shows a radon- and a γ-profile across the ore zone. Downhill, to the right in the picture, the soil was wet and radon samples were not

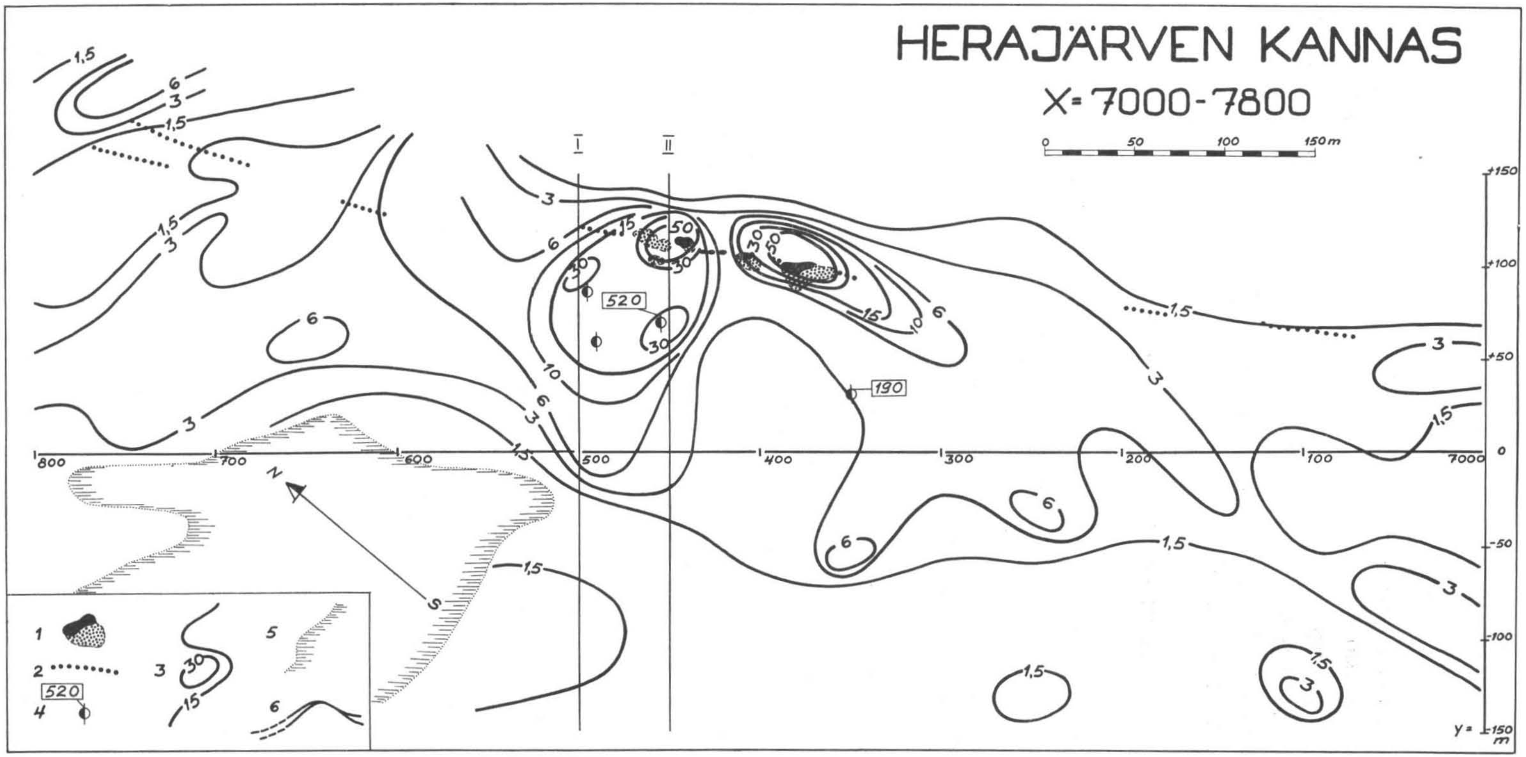

Fig. 8 A. Isoradon-map from Herajärven Kannas (X = 7 000—7 800). 1. Uranium ore lenses, the outcrop and its known continuation at depth. 2. Uranium mineralization. 3. Smoothed isoradon-curves. 4. Spring; radon content of water in emans. 5. Swamp. 6. Surface of ground and rock.

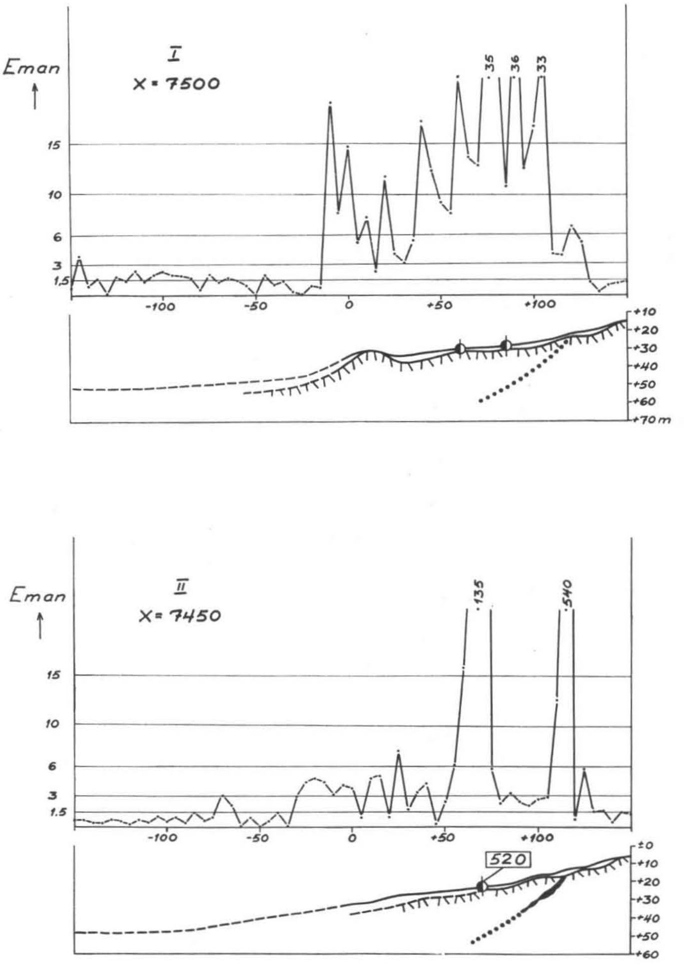

Fig. 8 B. Two profiles from the map on page 36. See explanations in Fig. 8 A.

obtained at all points. The peak anomaly, about 30 emans, was obtained just above the ore. It is assumed that the anomaly is caused by radon diffusing through fissures in the rock. The γ-anomaly coincides with the radon peak and the smaller one is probably caused by radioactive elements adsorbed on peat.

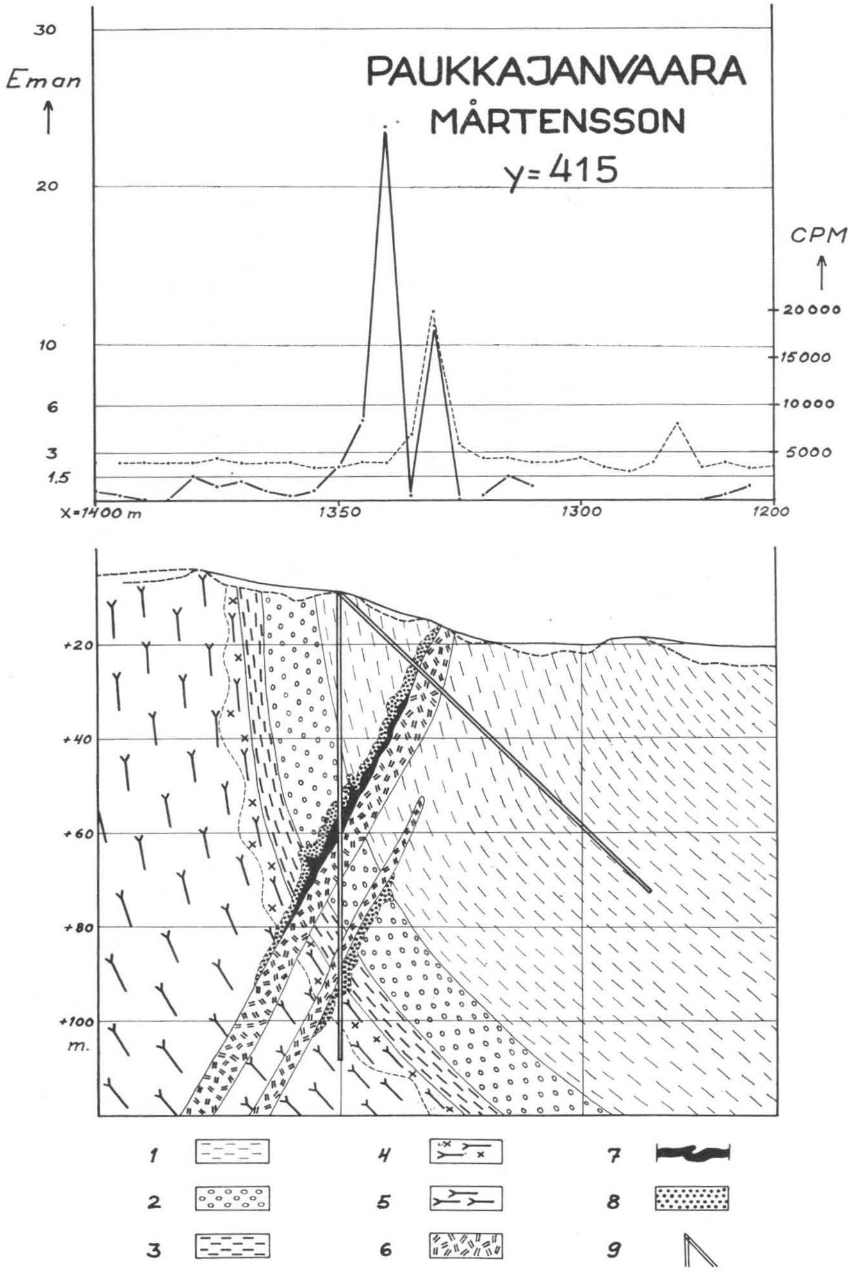

Fig. 9. Section of the Mårtensson-ore, Paukkajanvaara. Full line = radon content of soil air. Dashed line = γ-activity on surface (arbitrary units). 1. Orthoquartzite. 2. Conglomerate. 3. Mica-quartz-schist. 4. Micagneiss. 5. Gneissose-granite. 6. Metabasite. 7. Uranium ore. 8. Uranium mineralization. 9. Drill-hole.

SUMMARY

Radon measurement has been used as a systematic exploration method of uranium prospecting in the Koli area. The active carbon method has provided a means of rapid and inexpensive sampling, which has made possible the investigation of large areas with untrained personnel and with reasonable cost.

As a result of the measurements a large number of anomalies were found. The interpretation of limited peak anomalies has, as a rule, been easy, whereas the explanation of the extensive small anomalies in many cases has proved to be difficult.

The development of radon anomalies is a complex process, and many factors must be taken into account: hydrogeological conditions, the radium content of soil and water, redeposited radium, the topography, the porosity of the soil etc. It is believed, that investigation of these factors will lead to a better understanding of radon anomalies.

The importance of radon measurements in the Koli area may be estimated as follows.

1. Radon measurements may limit the potential ore field. Tests in barren areas have given low radon contents of soil air (between 0.5 to 1.5 emans) as well as waters.
2. In favourable circumstances it has been relatively easy to correlate a radon anomaly with a uranium mineralization. In many cases this connection is, however, obscure. To make predictions as to the quality of the uranium mineralization from consideration of the radon anomalies, does not appear to be justified.

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REFERENCES

ALEKSEEV, V. V., GRAMMAKOV, A. G., NIKONOV, A. I., TAFEEV, G. P. (1957) Radiometric methods in the prospecting and exploration of uranium ores. Book 2 (In Russian). State scientific-technical publishers of literature on geology and mineral resources conservation, Moscow. English translation: AEC-tr-3738 (Book 2), United States Atomic Energy Commission.

AUROLA, ERKKI (1955) Über die Geschiebeverfrachtung in Nordkarelien. Geotekn. julk. 56, Helsinki.

BECKER, K. H. und STEHBERGER, J. (1929) Über die Adsorption der Radiumemanation. Ann.Phys. (5) 1. p. 529—542.

BÊHOUNEK, F. (1927) Über die Verhältnisse der Radioaktivität im Uranpecherzbergbaurevier von St. Joachimsthal in Böhmen. Physik. Z. 28, p. 444—446.

BUDDE, E. (1958) Bestimmungen des Beweglichkeitskoeffizienten der Radiumemanation in Lockergesteinen. Z. Geophys. 24, p. 96—105.

—»— (1957) Rn-Messungen auf Flusspatgängen. Geol. Bavarica Nr. 35, p. 39—58.

CADUDAL (1955) Dosage du radon dans les mines d'uranium par la méthode de prélévement sur charbon. J. phys. radium 16, p. 479—482.

FAUL, H. (1954) Nuclear Geology, p. 138—143. John Wiley & Sons, Inc., New York.

FROSTERUS, B. and WILKMAN, W. W. (1920) Suomen geologinen yleiskartta, Lehti D. 3. Joensuu, Vuorilajikartan selitys. Suomen geol. toimikunta, Helsinki.

GILETTI, B. J. and KULP, J. L. (1955) Radon leakage from radioactive minerals. Am. Mineralogist 40, p. 481—496.

GÜBELI, O. und STAMMBACH, K. (1951) Zur Adsorption von Radon an Aktivkohle und Silicagel. Helv. Chim. Acta 34, p. 1257—63.

HATUDA, Z. (1954) Radioactive method for geological exploration. Mem. Coll. Sci., Univ. of Kyoto, ser. B, Vol. XXI, N:o 2, article 9.

LANDSTRÖM, O. (1960) Application of radon measurements to uranium prospecting. No AE-7, AB Atomenergi, Stockholm.

PRADEL, J. (1956) La prospection de l'uranium par le radon. Bull. d'inform. scient. et techn. C. E. A., 8—9, p. 1—9.

SAKSELA, MARTTI (1949) Malminetsintä. Gummerus Oy. Jyväskylä.

SAURAMO, MATTI (1924) Tracing of glacial boulders and its application in prospecting. Bull. Comm. géol. Finlande 67.

WEGMAN, C. E. (1928) Über die alpine Tektonik der jüngeren Faltungen in Ost-Finnland. Fennia 50 N:o 16.

VÄYRYNEN, H. (1933) Über die Stratigraphie der Karelischen Formation. C. R. Soc. géol. Finlande 6, Bull. Comm. géol. Finlande 101, p. 54.

4

PALÄOBOTANISCHE UNTERSUCHUNG ÜBER EINEN NÖRDLICHEN FUNDORT SUBFOSSILER TRAPA NATANS L. IN SÜD-POHJANMAA[1]

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AUSZUG

In dieser stratigraphischen und paläobotanischen Arbeit wird der nördlichste zur Kenntnis gelangte ehemalige Standort von Wassernuss *(Trapa natans* L.), das Moor Valmossa im Kirchspiel Evijärvi beschrieben. Die Wassernuss ist an dieser Stelle gegen Ende der postglazialen Wärmezeit rund 500 Jahre gewachsen. Die wichtigsten Voraussetzungen des Gedeihens der Pflanze sind die klimatisch günstige Zeit, der topographisch geschützte Standort und die edaphisch befriedigten Verhältnisse gewesen.

INHALT

[1] Eingegangen den 20. Januari, 1960.

EINLEITUNG

Der Verfasser fand im Jahre 1948 in Gyttjaproben des Moores Valmossa einige makroskopischen Überresten von Wassernuss *(Trapa natans L.)*. Dieser neue Fundort von *Trapa* liegt in Süd-Pohjanmaa, im Kirchspiel Evijärvi, in unmittelbarer Nähe der Kirche (Abb. 1). Die geographische Lage des Moores beträgt etwa 63° 38′ n. Br. und 23° 38′ ö. L. Das Moor Valmossa ist also der nördlichste bekannte ehemalige Standort dieser quartärgeologisch und paläobotanisch interessanten Pflanze in Fennoskandien. Das im Kirchspiel Siilinjärvi auftretende Moor Mikansuo, dass von jeher als nördlicher Fundort der *Trapa* bekannt ist (Lumiala 1940), liegt etwa drei Meilen weiter südlich als das Moor Valmossa.

Obwohl die Makroflora der Moore von Pohjanmaa in erster Linie als Ergebnisse der Arbeit Backmans (1941, 1943, 1948 und 1950) recht gut bekannt ist, hat man subfossile Überresten von *Trapa* dort zuvor nicht gefunden. Dies ist als ein ganz natürlicher Sachverhalt angesehen worden, weil Pohjanmaa in der Litorinazeit, der Zeit der weitesten Verbreitung von *Trapa*, grösstenteils vom Meere bedeckt gewesen ist. Daher hat man nicht einmal zu erwarten verstanden, dass Subfossilien dieser Pflanze in der Küstenzone des Bottnischen Meerbusens vorkämen (Aario 1932, S. 154).

Weil das Moor Valmossa in der Peripherie des ehemaligen Verbreitungsgebietes von *Trapa* lag, war es zu erwarten, dass dessen Subfossilflora die zur Zeit des Auftretens und des Schwindens der Pflanze bestehenden Verhältnisse empfindlich widerspiegelte. Bei den Untersuchungen galt denn auch die Aufmerksamkeit ausser dem geologischen Aufbau des Moores auch dem Beschaffen eines vielseitigen Makrofossilienmaterials sowie der Klärung der Bedingungen, die auf die Entwicklung des Moores und seiner ehemaligen Flora haben einwirken können.

Trapa ist in postglazialer Zeit in den Gewässern der südlichen Hälfte Finnlands sehr häufig vorgekommen; man hat sie nämlich in über fünfzig verschiedenen vorzeitlichen Seen angetroffen. Der nördlichste rezente Standort der Pflanze liegt heute in Lettland, in dem See Klaucanu (Apinis 1940), so dass sich die nördliche Grenze ihrer Verbreitung rund 750 Kilometer südwärts verschoben hat. Es erhebt sich die Frage, welche Ursachen auf eine so starke Grenzverlegung hingewirkt haben. Offenbar lässt sich keine in jedem einzelnen Falle wirksam gewesene Ursache ausmachen. Die Beschaffenheit des Standortes beruht auf dem Zusammenwirken vieler Faktoren, auf einer Gesamtheit, von der wir nur eine verhältnismässig lückenhafte Vorstellung haben. Wir kennen nicht genügend alle Teilfaktoren und ihre Bedeutung, nicht einmal für die gegenwärtigen Pflanzenbestände (Jalas 1950, S. 179). Im folgenden wird jedoch versucht, einige Hauptfaktoren zu betrachten und zu erschliessen, welche von ihnen auf das Schicksal von *Trapa* und zugleich auch anderen wärmezeitlichen Pflanzenarten in Finnland haben einwirken können.

Die mit der Untersuchung verbundenen Makrofossilienanalysen hat der Verfasser im Laboratorium der Geologischen Forschungsanstalt ausgeführt. Mag. phil. Ester Uussaari hat mir mit Pollenanalysen und Lic. phil. Kyllikki Salminen mit Diatomeenbestimmungen geholfen. Die Übersetzung der Arbeit ins Deutsche hat Dr. phil. Marta Römer besorgt. Die sorgfältig ins reine gearbeiteten Zeichnungen sind von Frau Lyyli Orasmaa ausgeführt worden. Allen obengenanten Helferinnen möchte ich meinen besten Dank zum Ausdruck bringen.

AUFBAU UND STRATIGRAPHIE DES MOORES

Valmossa ist ein etwa 150 ha umfassendes Hochmoor, in dem Schlenken und Stränge miteinander abwechseln und dessen nördlichem Ende der Weiher Kuikkalampi ein Nachlass vom vorzeitlichen Binnenseestadium des Moores ist. Der östliche Rand des Moores stösst an den See Evijärvi. Unter den Pflug genommene Moränenböden umgeben die übrigen Ränder des Moores (Abb. 2). Die Höhe des mittleren Moorteiles wechselt zwischen 64.5 und 65.4 m ü. d. M.

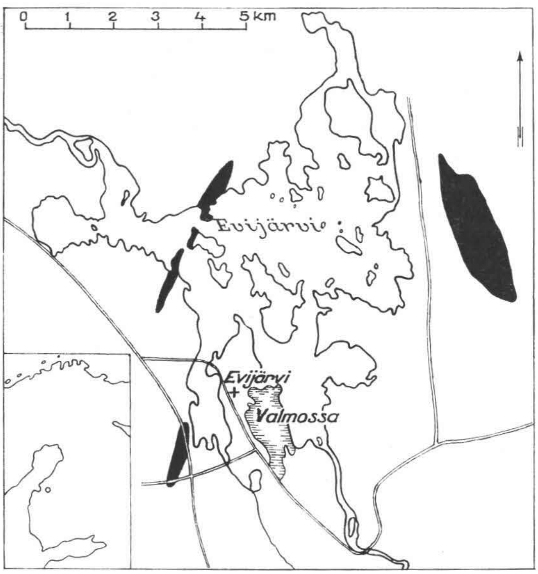

Abb. 1. Das Moor Valmossa liegt am südlichen Teil des Sees Evijärvi, am östlichen Rand der Halbinsel Kirkkoniemi. Die schwarzen Linsen vertreten Amfibolitvorkommen im Gebiet.

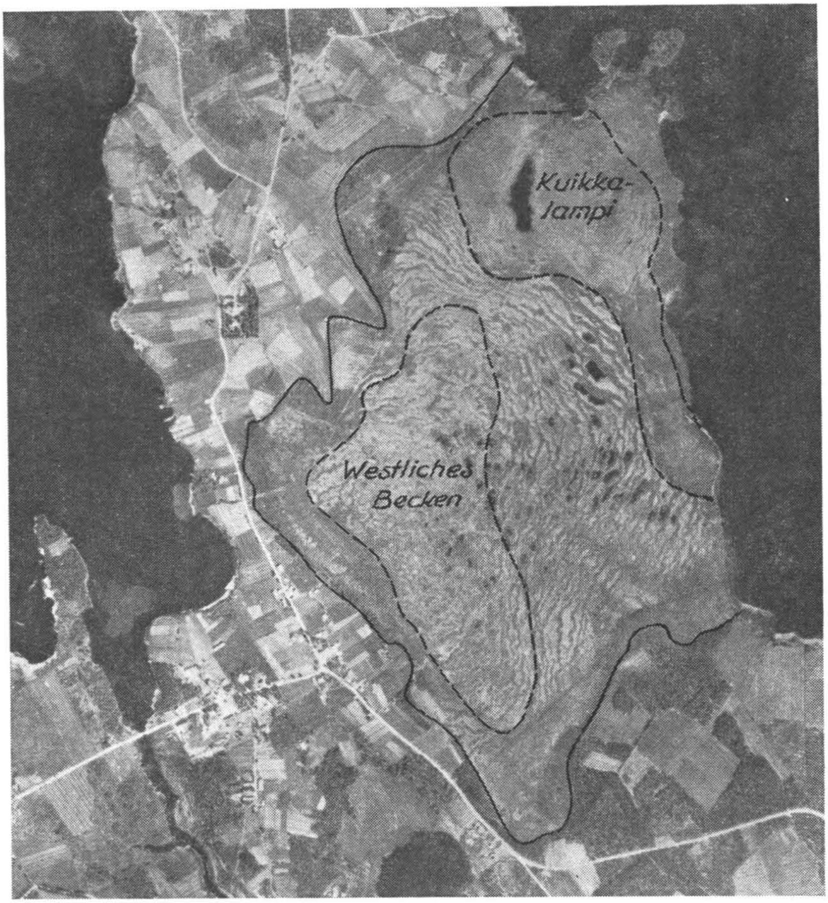

Abb. 2. Das Moor Valmossa nebst Umgebung, von der Luft aus gesehen. Die durch gestrichelte Linien unterschiedenen Gebiete zeigen die Lage der limnischen Sedimenten in Untersuchungsgebiet. Veröff. mit Genehmigung von Landesvermessungsamt in Finnland.

Der Felsgrund des Gebietes besteht aus zu der suprakrustischen Bildungen gehörenden Gesteinen, vorwiegend Biotitplagioklasgneisen (Laitakari 1942). In diese Formation gehen basische Effusivgesteine ein, von denen dunkelgrauer Amphibolit in drei gesonderten Vorkommen den See Evijärvi umgeben (Abb. 1).

Die häufigste Bodenart im Gebiet ist Moräne. Sie bildet schwach gewundene Rücken und Buckel, deren Zwischenräume meistens von Torfschichten bedeckt sind. Das Moor Valmossa ist in einer derartigen moränengründigen Senke entstanden. Der das Moor unterlagernde Rücken trennt es in zwei Teile (Abb. 2 und 3). Die Senke im westlichen Teil des Moores ist als »westliches Becken» und der östliche Teil des Moores oder der an das Ufer des Evijärvi reichende Moorteil als »Ufervermoorung» bezeichnet worden.

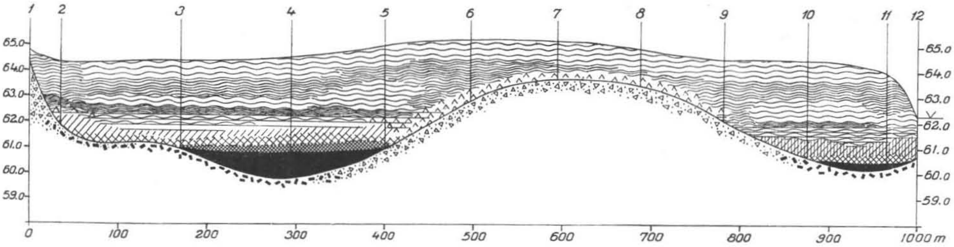

Abb. 3. Querschnitt durch das Moor Valmossa. Erläuterung der Erdarten in Abb. 4.

Die Senken in den Becken enthalten Tonsedimente, in die reichlich Salzwasserdiatomeen eingehen, wie *Campylodiscus echeneis*, *Cocconeis scutellum*, *Diploneis Smithii*, *Grammatophora oceanica*, *Melosira moniliformis*, *Rhabdonema arcuatum* und *R. minutum*. Diese Diatomeen kennzeichnen die Ablagerungen des Litorinameeres (Hyyppä 1937, S. 37), und besonders in den Litorinaablagerungen von Pohjanmaa kommen sie häufig vor (Mölder 1946). Die den Ton überlagernde Gyttja enthält dagegen Arten, die ausschliesslich in süssen und kleineren Gewässern gedeihen. Die Tonsedimente sind demgemäss zur Zeit des Meeresstadium und die Gyttjen in dem Süsswassersee entstanden.

Der zwischen Ton und Gyttja deutlich zu erkennende Kontakt, der im Gebiet des westlichen Beckens in einem Niveau von 61 m liegt, bezeichnet die Abschnürung des Evijärvi vom Litorinameer sowie das schnelle Sinken des Wasserspiegels im Zusammenhang mit dem Abschnürungsvorgang.

Am Grunde des vom Meere befreiten Sees setzte sich zunächst Feindetritusgyttja ab, die wenig organischen Stoff enthielt. Das Moor Valmossa war damals eine Einbuchtung des Evijärvi, und dieser selbst war ein Durchflusssee im Gewässersystem des Flusses Ähtävänjoki geworden.

Der Kontakt zwischen Fein- und Grobdetritusgyttja erweist ein neues Sinken des Wasserstandes. Das westliche Becken begann sich zu einem lagunenartigen Einschnitt des Evijärvi zu gestalten, und dieser Einschnitt stand durch einem im nordöstlichen Teil des Beckens gelegenen Sund mit dem eigentlichen See in Verbindung. Dieses zweitmaligen Absinken des Wassers bewirkte das Aufkommen einer üppigen und artenreichen Vegetation. Durch sie begann sich das Becken auszufüllen. Der frühere See bedeckte sich nach der Zuwachsung mit Seggenweissmoor, das die eigentliche Vermoorungsphase im Becken einleitete. Der mit Baumresten durchsetzte Torf auf dem Seggentorf bezeugt eine trockene Periode in der Entwicklung des Moores. Eine bruchmoorartige Vegetation strebte von den Rändern des Moores gegen seine Mitte, aber eine starke Sphagnum-Torfbildung verhinderte die Bewaldung des ganzen Moores.

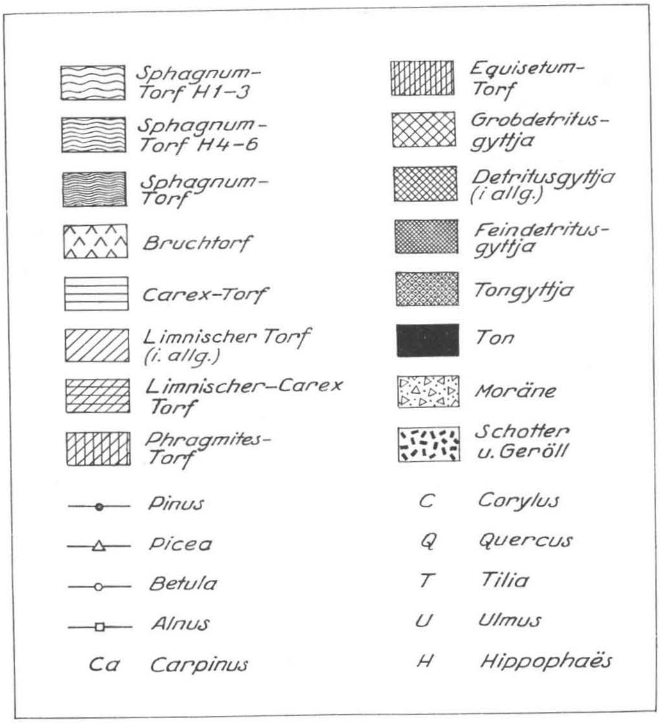

Abb. 4. Die in der Querschnittzeichnungen, der Pollen- und Makrofossiliendiagrammen benutzten Erdart- und Holzartbezeichnungen. H bedeuet den Huminositätgrad, wobei H 1 nichthumifizierten und H 10 völlig humifizierten Torf bedeutet.

DATIERUNG DER MOORABLAGERUNGEN

Das Pollendiagramm in Abb. 5 stellt die Lagerfolge im mittleren Teil des westlichen Beckens dar. Der Pollen von *Hippophaës rhamnoides*, in 4.1 und 4.3 m Tiefe gefunden, bedeutet nach Salmi (1948, S. 19) bei Vorkommen in der Moorablagerungen die Nähe des Meeresufers. Die besagten Stellen im Diagramm bezeichnen die Zeit, in der die höchsten Geländestellen in der Gegend des Evijärvi aus dem Meere aufzusteigen begannen. *Hippophaës* gedieh offenbar an den Ufern dieser gesonderten Inseln, wie gegenwärtig auf Åland (Palmgren 1912) oder im Schärenhof von Vaasa (E. J. Valovirta 1937).

Pollen von edlen Laubbäumen sowie Birke und Erle kennzeichnen die Gegend des Abschnürungsstadiums, den Kontakt zwischen Ton und Gyttja. Fichte fehlt in diesem Niveau. Das einheitliche Auftreten ihrer Pollenkurve setzt an der Grenze zwischen Fein- und Grobdetritusgyttja ein und erstarkt zum erstenmal im limnotelmatischen Kontakt. Darauf sinkt die Fichten-

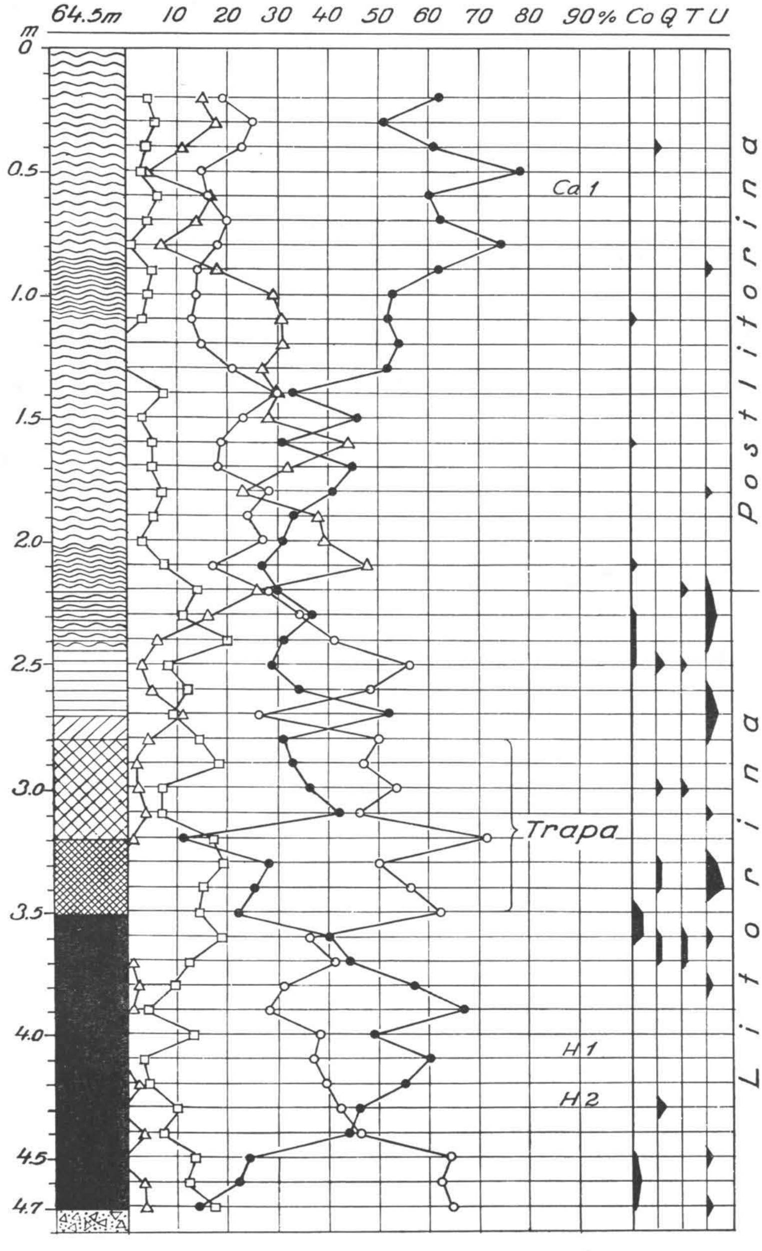

Abb. 5. Pollendiagramm aus der mittleren Gegend des westlichen Beckens. Zeichnenerklärung in Abb. 4.

kurve, wonach sie im unteren Teil des *Sphagnum*-Torfes auf ihr postglaziales Maximum erreicht. Diesem Anstieg der Fichtenkurve entsprechen das Abnehmen der edlen Laubbäume sowie ein deutlich Sinken der Kurven von Birke und Erle. Die Fichte erlangte eine bleibende Stellung als Holzart der Wälder, und in der Schichtenfolge des Moores war der *Sphagnum*-Torf die vorherrschende Bodenart.

Das höchste Ufer des Litorinameeres liegt nach dem von Hyyppä (1950) dargestellten Isobasensystem in der Gegend der Kirche von Evijärvi 97 m ü. d. M. Die Höhe des Abschnürungsniveaus des westlichen Beckens beträgt 61 m ü. d. M. Sie entspricht dem Litorinameerstadium, wie die Diatomeen bereits erwiesen haben. Das Abschnüren des westlichen Beckens vom Litorinameere geschah zu einem Zeitpunkt, als sich das Meeresufer, infolge der Landhebung, 36 m unter die Litorinagrenze verschoben hatte. In der von Hyyppä (1950, S. 29) dargestellten Uferverschiebungskurve fällt diese Höhe, 61 m ü. d. M., in die Zeit um das Jahr 2000 v. Chr.

Das einheitliche Auftreten von Fichte beginnt oberhalb des Abschnürungsniveaus (Abb. 5), also etwas später als der oben angegebene Zeitpunkt. Nach den Untersuchungen von Hyyppä (1935, 1950) beginnt das einheitliche Auftreten der Fichte um die Einbuchtung des Bottnischen Meerbusens sowie in Pohjanmaa, in der Gegend von Lapua, um 1800 v. Chr. Ebenso hat Virkkala (1950) auf Grund seiner Beobachtungen in N-Satakunta und durch Prüfen vieler Pollendiagramme der Küste des Bottnischen Meerbusens als Zeit des Hervortretens von Fichte auch die um 1800 v. Chr. ermittelt. Die Pollenuntersuchungen von Valmossa erweisen, dass das einheitliche Auftreten der Fichte auch am Evijärvi um 1800 v. Chr. einsetzt.

Das Ansteigen der Fichtenpollenkurve auf ihren höchsten Gipfel bezeugt nach Okko (1949, S. 99) das Aufhören des Litorinameeres in Mittel-Pohjanmaa. Die damalige Uferlinie liegt um 40 m über dem gegenwärtigen Meeres-

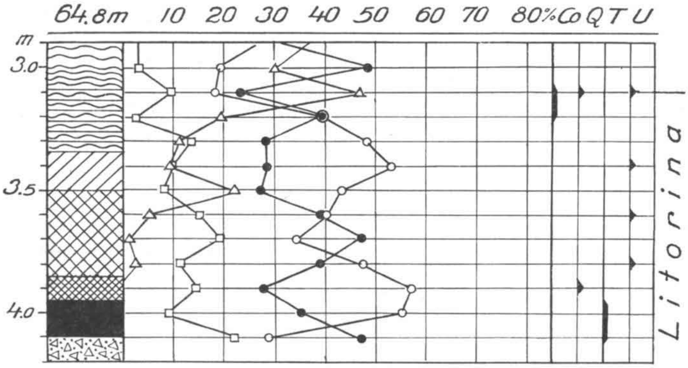

Abb. 6. Pollendiagramm aus dem südlichen Ende des westlichen Beckens. Zeichnenerklärung in Abb. 4.

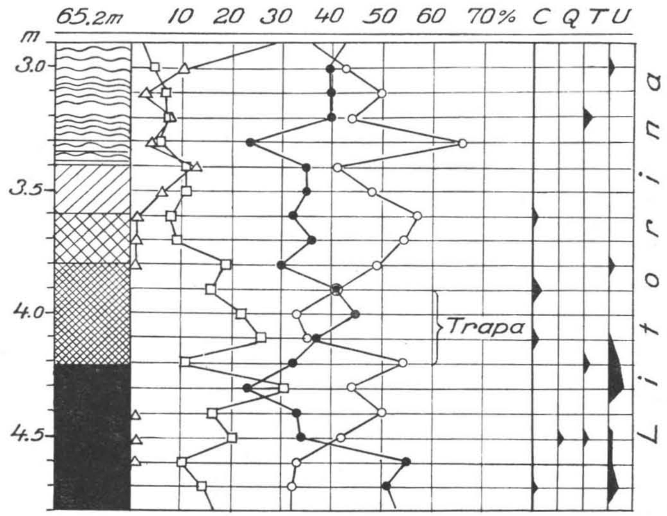

Abb. 7. Pollendiagramm aus dem nördlichen Ende des westlichen Beckens. Zeichnenerklärung in Abb. 4.

spiegel. Diese Höhe entfällt in der obengenannten Uferverschiebungskurve von Hyyppä ungefähr auf das Jahr 1200 v. Chr. Dieser Zeit entspricht im Diagramm (Abb. 5) der starke Anstieg der Fichtenkurve in 2.2 m Tiefe.

Die Teildiagramme (Abb. 6—7) lassen erkennen, dass das Abschnürungsniveau ziemlich genau bei 61 m liegt. Das Einsetzen der einheitlichen Fichtenkurve fällt in allen Diagrammen in die Grenze zwischen Fein- und Grobdetritusgyttja. Das erste Ansteigen der Fichtenkurve erscheint in den mittleren und nördlichen Teilen des westlichen Beckens im limnotelmatischen Kontakt (Abb. 5 und 7). Am südlichen Ende des Beckens liegt dieser Kontakt etwas höher (Abb. 6), so dass das südliche Ende später zugewachsen ist als die nördliche, wo die eng aufeinanderfolgenden Schichten eine ununterbrochene Zuwachsung bezeugen.

Tabelle 1. Zusammenstellung einiger spätpostglazialen Vorgänge. Z = Waldgeschichtliche Zonengliederung. Ostseegeschichte nach Sauramo (1958) und Hyyppä (1937)

Alter	Z	Klimaperioden	Ostsee nach Sauramo	Ostsee nach Hyyppä	Ereignisse in Valmossa
1 000	IX	Subatlantikum	Mya	Post-Litorina	Maximum der Fichtenpollenkurve
± 0			Limnaea		
1 000					
2 000	VIII	Subboreal			Trapa natans
3 000	VII	Atlantikum	Litorina	Litorina	Abschnürung des westl. Beckens

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Tabelle 2. Die subfossile Flora des Moores Valmossa

Arten	Westliches Becken S-Ende	Westliches Becken Mittl. Teil	Westliches Becken N-Ende	Ufervermoorung N-Ende	Ufervermoorung Mittl. Teil	Ufervermoorung S-Ende
WASSERPFLANZEN:						
Alisma plantago-aquatica	+	+	+	+	+	+
Batrachium sp.	—	+	+	—	—	—
Equisetum limosum	+	+	+	+	+	+
Hippuris vulgaris	+	—	+	—	+	+
Iris pseudacorus	—	+	+	—	+	+
Lysimachia thyrsiflora	+	+	+	+	+	+
Myriophyllum alterniflorum	+	+	—	—	+	—
Najas tenuissima	+	+	+	—	—	—
Nuphar luteum	+	+	+	+	+	+
Nymphaea alba	+	+	+	+	+	+
Phragmites communis	+	+	+	+	+	+
Potamogeton cf. gramineus	+	—	+	+	—	—
— natans	+	+	+	+	+	+
— pectinatus	—	—	—	+	—	—
— sp.	+	—	+	—	—	—
Rumex hydrolapathum	+	+	+	—	+	—
Sagittaria sagittifolia	+	+	+	+	—	—
Scirpus lacuster	+	+	+	+	+	+
Sparganium minimum	+	+	+	—	+	+
— ramosum coll.	+	+	+	+	—	—
— simplex	+	—	+	+	—	—
— sp.	+	—	—	—	—	—
Trapa natans	—	+	+	—	—	—
Typha sp.	—	+	+	+	—	—
UFERPFLANZEN:						
Bidens sp.	—	—	—	—	+	—
Calla palustris	—	+	+	+	+	—
Caltha palustris	—	+	—	—	—	—
Carex canescens	—	+	+	+	—	+
— diandra	—	+	+	+	+	+
— lasiocarpa	—	—	+	+	+	—
— pseudocyperus	+	+	+	+	+	+
— rostrata	+	+	+	+	+	+
— vesicaria	—	—	+	—	+	—
— sp.	—	—	+	—	—	—
Cicuta virosa	+	+	+	+	+	+
Comarum palustre	+	+	+	+	+	+
Dryopteris sp.	—	+	+	—	+	—
Filipendula ulmaria	—	+	—	—	+	—
Lycopus europaeus	—	+	+	+	+	—
Menyanthes trifoliata	+	+	+	+	+	+
Peucedanum palustre	—	+	+	—	+	—
Scirpus silvaticus	—	—	—	—	+	—
Stellaria cf. uliginosa	—	—	+	—	—	—
Trollius europaeus	—	—	+	—	—	—
Viola palustris	+	—	—	—	—	—

Tabelle 2 (Forts.)

Arten	Westliches Becken			Ufervermoorung		
	S-Ende	Mittl. Teil	N-Ende	N-Ende	Mittl. Teil	S-Ende
MOORPFLANZEN:						
Andromeda polifolia	+	+	+	+	+	+
Carex limosa	—	—	—	+	+	—
Eriophorum vaginatum	+	+	+	+	+	—
Oxycoccus sp.	—	+	—	+	—	—
Scheuchzeria palustris	—	—	+	+	—	—
Vaccinium sp.	+	—	—	—	—	—
Bryales	+	+	+	+	—	—
Sphagnales	+	+	+	+	+	+
BÄUME UND STRÄUCHER:						
Alnus glutinosa	+	+	+	+	+	+
— incana	—	+	+	+	+	+
Betula alba	+	+	+	+	+	+
— nana	—	+	—	—	—	—
Hippophaës rhamnoides	—	+	+	—	—	—
Picea excelsa	—	—	—	—	—	+
Pinus silvestris	+	+	+	—	+	+
Salix sp.	+	+	+	—	—	+
TIERE:						
Coleoptera	+	+	+	+	+	+
Cristatella	+	+	+	—	+	+
Daphnia	+	+	—	—	—	+
Nephelis	+	+	+	+	—	+
Piscicola	+	+	+	+	+	+

MAKROFOSSILIENGEHALT DER MOORABLAGERUNGEN

Die im Gebiet Valmossa gefundenen subfossilen Pflanzenarten sind in Tabelle 2 angegeben. Die Nomenklatur richtet sich nach der von Hiitonen (1933).

Die Tabelle enthält insgesamt 61 Pflanzenarten, von denen 24 Wasserpflanzen, 21 Uferpflanzen, 8 Moorpflanzen und 8 Bäume und Sträucher sind.

Bei Ausführung der Pollenanalysen sind neben Holzarten (Obs. Carpinus, Abb. 5) Vertreter vieler anderen Pflanzen angetroffen worden. Von diesen sind in Tabelle 2 *Hippophaës* und *Typha* aufgenommen worden, denn diese zwei Arten haben wesentlich zur Vegetation der frühesten Entwicklung des Moores gehört.

In der Tabelle sind auch die zum Tierreich gehörenden makroskopischen Überresten angegeben. Deckflügel und Glieder von Käfern *(Coleoptera)* sind in den Moorablagerungen häufig. Der Statoblast des Moostierchens *(Cristatella)* mit seinen vielen Widerhaken ist charakteristisch. Die Überwinterungsorgane des Wasserflohes *(Daphnia)*, die Sättel (Ephippia), sind verhältnismässig häufig und leicht erkennbar. Die Eierkokons und -kapseln der Egel *(Piscicola* und *Nephelis)* sind gemeine Makrofossilien limnischer Gyttjen.

Die Mengen der Makrofossilien wechseln beträchtlich in den verschiedenen Teilen des Moores und seinen verschiedenen Schichten. Die in der Horizontalrichtung hervortretenden Schwankungen der Pflanzenarten sind aus Tabelle 3 zu ersehen. In ihr sind die Pflanzen einiger Vertikalprofile angeführt. Die Profile 1—6 sind aus dem Gebiet des westlichen Beckens und die Profile 7—10 aus der Ufervermoorung gebohrt worden. Am meisten Arten enthält das an der Seite des westlichen Beckens erbohrte Profil (3). Dem Gebiet der Ufervermoorung (Profil 8), ist ebenfalls eine reichhaltige Flora entnommen worden. Am wenigsten Arten gibt es im südlichen und mittleren Teil des westlichen Beckens (Profile 1—2).

Tabelle 3. Die Artenfrequenzen der Vertikalprofile des Moores Valmossa, auf die ökologischen Gruppen verteilt

Profil	Wasser-pflanzen	Uferpflanzen	Moorpflanzen	Bäume u. Sträucher	Arten zus.
1	8	2	4	4	18
2	8	6	3	2	19
3	13	14	2	5	34
4	13	10	4	2	29
5	9	8	3	4	24
6	7	7	3	3	20
7	10	7	3	2	22
8	8	14	4	2	28
9	7	5	3	6	21
10	8	6	2	4	20

Als kurze Zusammenfassung über die Unterbringung der Pflanzenarten und ihr Verhältnis zu den verschiedenen Schichten des Moores wird noch Tabelle 4 betrachtet. Die Makrofossilien, die Wasserpflanzen vertreten, entfallen in den Lagerfolgen der Valmossa hauptsächlich auf die limnischen Schichten. Wasserpflanzen mit untergetauchten Blättern *(Batrachium sp., Myriophyllum alterniflorum* und *Najas tenuissima)* erscheinen in Feindetritusgyttja. Wasserpflanzen mit schwimmenden Blättern *(Potamogeton natans* und *Trapa natans)* entfallen auf Fein- und Grobdetritusgyttja. *Sagittaria sagittifolia* zählen ebenfalls zu den Arten im Übergangsgebiet zwischen diesen Gyttjaarten. Der grösste Teil der Wasserpflanzen, besonders die Helophyten, kommen in Grobdetritusgyttja vor, die sich in flacherem Wasser abgesetzt hat. *Equisetum limosum, Phragmites communis* und *Lysimachia thyrsiflora,* im limnischen Torf dominierende Arten, erscheinen zum Teil schon in Grobdetritusgyttja sowie oberhalb des limnotelmatischen Kontaktes in dem vorwiegend Seggen führenden, als telmatisch anzusprechenden Torf.

Tabelle 4. Die Beziehung der Makrofossilien zu den verschiedenen Moorschichten

Arten	Zuwachsungsphase			Vermoorungsphase		
	Fein-detr.	Grob-detr.	Limn. Torf	Telm. Torf	Semit. Torf	Terr. Torf
WASSERPFLANZEN:						
Alisma plantago-aquatica	—	—	—			
Batrachium sp.	—					
Equisetum limosum		—	—	—		
Hippuris vulgaris		—				
Iris pseudacorus		—				
Lysimachia thyrsiflora		—	—	—		
Myriophyllum alerniflorum	—					
Najas tenuissima	—	—				
Nuphar luteum	—	—				
Nymphaea alba	—	—				
Phragmites communis		—	—	—		
Potamogeton natans	—	—				
Rumex hydrolapathum		—				
Sagittaria sagittifolia	—	—				
Scirpus lacuster	—	—				
Sparganium minimum		—				
— ramosum	—	—				
— simplex		—				
Trapa natans	—	—				
UFERPFLANZEN:						
Calla palustris		—	—	—		
Carex canescens		—	—	—		
— diandra						
— lasiocarpa		—	—	—	—	
— pseudocyperus		—				
— rostrata		—	—	—	—	—
— vesicaria		—				
Cicuta virosa		—				
Comarum palustre		—	—	—	—	—
Dryopteris sp.		—				
Filipendula ulmaria		—				
Lycopus europaeus		—				
Menyanthes trifoliata			—	—	—	
Peucedanum palustre		—	—			
MOORPFLANZEN:						
Andromeda polifolia				—	—	—
Carex limosa						—
Eriophorum vaginatum						—
Oxycoccus sp.					—	—
Scheuchzeria palustris						—
Vaccinium sp.					—	
Bryales		—	—	—	—	
Sphagnales			—	—	—	—
BÄUME UND STRÄUCHER:						
Alnus glutinosa	—	—	—	—	—	
— incana		—	—	—	—	
Betula alba	—	—	—	—	—	—
— nana			—		—	—
Picea excelsa			—	—	—	—
Pinus silvestris			—	—	—	
Salix sp.			—	—		

Helophyten sowie eigentliche Uferpflanzen sind in ihren Vorkommensweisen einander ähnlich. Sie wachsen am besten in flachem Uferwasser, und deswegen erscheinen beider Überreste in Grobdetritusgyttja.

Die eigentliche Moorpflanzen finden sich in semiterrestrischem sowie terrestrischem Torf, wie es auch zu erwarten ist. Doch kommen gewisse Braunmoose und einige *Sphagnum*-Moose häufig schon unterhalb des limnotelmatischen Kontaktes vor.

Das Auftreten von Birke und Erle sowie überhaupt von Pflanzen festen Bodens ist nicht allein von den Moorboden abhängig, sondern ihre Vorkommen beruhen oft auf der Nähe der Uferlinie, was im folgenden Abschnitt des näheren darzulegen sein wird.

DIE MAKROFOSSILIEN ALS INDIKATOREN FÜR DIE MOORENTWICKLUNG

Die Lagerfolge im Moore Valmossa ist sehr deutlich. Auf dem Grunde des vom Meere befreiten Beckens, auf Ton, setzte sich zunächst Feindetritusgyttja ab. Infolge der im Wasserhaushalt eingetretenen Wandlungen sanken stets grössere und reichlichere Pflanzenteilchen auf den Boden des Beckens ab, erfüllten ihn mit Grobdetritusgyttja und letztens mit limnischem Torf. Nach der Zuwachsung des Beckens folgten aufeinander seggenreicher, stellenweise holzreicher und sphagnumreicher Torf, wodurch die verschiedenen Phasen der eigentliche Vermoorung bezeichnet sind (vgl. Abb. 3). Die verschiedenen Schichten sind also unter bestimmten Verhältnissen entstanden und in jeder einzelnen Schicht hinterlassenen Pflanzen vertreten den jeweils herrschenden Pflanzengesellschaftstyp. Die im folgenden darzustellenden Makrofossiliendiagramme zeigen die Aufeinanderfolge der Schichten in den verschiedenen Teilen des Moores und ferner, welche Pflanzen jeweils Komponenten jeder Schichten gewesen sind.

Die Diagramme sind im Anschluss an die Darstellungsweise Sellings (1939, S. 470—475) gezeichnet worden. Die vegetativen Makrofossilien sind schräg schraffiert und die übrigen Makrofossilien als schwarze Felder vermerkt. Die ausgesparten Teile des Birkenfeldes bedeuten die Mengen der Fruchtschuppen und die schwarzen Teile die der Samen. Die Bruchzahl bei einer Pflanzenart bedeutet, in welchem Verhältnis im Vergleich mit dem am rechten Rand des Diagrammes gegebenen Massstab der das Auftreten der Pflanzen vetretende Abschnitt gezeichnet ist.

Die ersten Makrofossilien kommen im Kontakt zwischen Ton und Feindetritusgyttja vor, im Abschnürungsniveau. Das Auftreten der *Trapa natans* im unteren Teil der Feindetritusgyttja (Abb. 10) bedeutet, dass das westliche Becken sich bei beginnender Sedimentation der genannten Gyttja als eigener Süsswassersee vom Meere getrennt hat. Die letzten Anzeichen von *Trapa* finden sich in oberen Teil der Grobdetritusgyttja. Auf die Sedimentation dieser Gyttjen sind rund 500 Jahre vergangen. Demgemäss ist *Trapa* im Becken des Moores Valmossa in der subborealen Zeit von 2000—1500 v. Chr. gediehen. Als Begleiter von *Trapa* aufgetreten sind *Alisma plantago-aquatica*, *Najas tenuissima*, *Potamogeton natans*, *Rumex hydrolapathum*, *Sagittaria sagittifolia*, *Scirpus lacuster*, *Sparganium ramosum* und *S. minimum*. Von den mit *Trapa* gleichzeitig gewachsenen Arten nennen wir *Calla palustris*, *Caltha palustris*, *Carex diandra*, *C. pseudocyperus*, *Cicuta virosa*, *Filipendula ulmaria*, *Lycopus europaeus*, *Lysimachia thyrsiflora* und *Peucedanum palustre*. Diese Arten erscheinen alle auf dem nördlichsten rezenten Standort von *Trapa natans* im Klaucanu-See in Lettland (vgl. Apinis 1940).

Die Analysen über die Trapa-Gyttja des Moores Valmossa erweisen diese als schwach sauer (pH 5.4—6.8). Die Grobdetritusgyttja, in der *Trapa* am reichlichsten vorkommt, enthält in reichlichem Masse zersetzte organische Substanz. Die Begleiter der Pflanze weist auf sehr eutrophe Verhältnisse hin. Der Elektrolytgehalt des Beckens ist vorwiegend beim Absetzen von Grobdetritusgyttja bedeutend gross gewesen. Die Tiefe des Wassers hat damals zwischen 30 un 80 cm gewechselt. Apinis (op. cit., S. 78) hat durch Analysen erwiesen, dass an den heutigen Standorten von *Trapa* die Reaktion des Wassers schwach sauer oder neutral ist, dass es in gelöstem Zustande geringe Mengen zahlreicher Salze sowie in reichlichem Massa organische Substanz enthält, weswegen der See als eutroph-dystropher Typ anzusehen ist. Die Verhältnisse auf den gegenwärtigen Standorten von Trapa in Lettland ähneln also stark denen im Becken des Moores Valmossa gegen Ende der Litorinazeit.

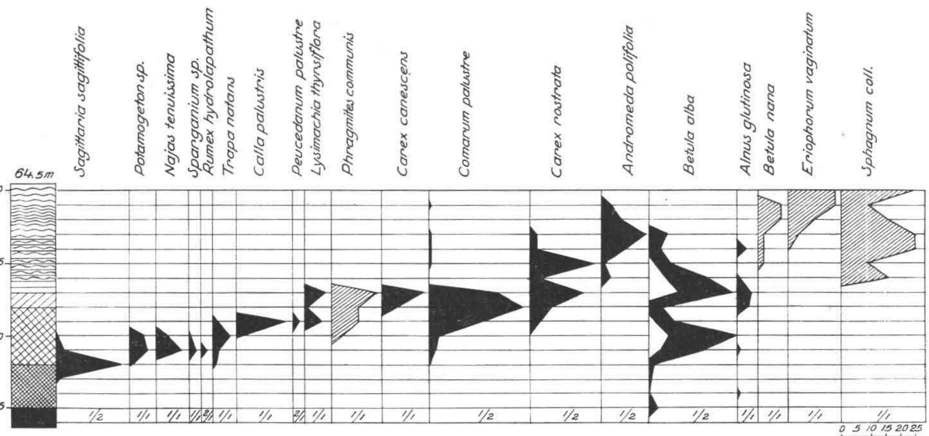

Abb. 8. Makrofossiliendiagramm aus der mittleren Gegend des westlichen Beckens. Zeichnenerklärung in Abb. 4.

Die Makrofossiliendiagramme erweisen, dass die Feindetritusgyttja wenig Pflanzenarten enthält. Bei der in flachem Wasser entstandenen Grobdetritusgyttja sind die Gesamt- und Artenmengen der Pflanzenarten bedeutend grösser. Besonders der Anteil der Helophyten nimmt vom unteren Teil der genannten Gyttja an zu. Dies hat an dem schnellen Sinken des Wasserspiegels gelegen. Die flachen Flanken des am Grunde des Moores sich erstreckenden Moränenrückens haben vielen in flachem Uferwasser gedeihenden Pflanzen gute Standorte geboten.

Auch die makroskopischen Überreste von Birke *(Betula alba)* haben die Schwankungen des Wasserspiegels im Becken registriert. Selling (1939, S. 474) hat nämlich erwiesen, dass die Pflanzen festen Bodens der Uferlinie nachgehen. Im Makrofossiliendiagramm ist die Transgression des Wassers als Abnahme dieser Pflanzen und die Regression umgekehrt als Zunahme zu erkennen. Die Schmalheit des Birkenfeldes in der Feindetritusgyttja des Moores Valmossa bezeugt ein Stadium ziemlich tiefen Wassers, die starke Ausweitung des Feldes in der Grobdetritusgyttja ein schnelles Sinken des Wasserspiegels. Das im limnischen Torf gelegene Minimum des Birkenfeldes bedeutet eine neue positive Verschiebung der Uferlinie, also eine Transgression des Wassers. Die Wassermenge des Beckens hatte jedoch infolge der Landhebung wie auch der Verlandung abgenommen. Der Anstieg des Wasserspiegels erschien denn auch als Überschwemmung, die den Rand des Birken- und auch des Erlenbestandes weiter auf den festen Boden verschob. Eine eigentliche Wasservegetation vermochte diese Transgression nicht nennenswert zu beleben, aber in flacherem Wasser gedeihende Pflanzenarten *(Phargmites, Equisetum* und *Lysimachia)* erlangten zur Zeit des Birkenminimums (im limnischen Torf) ihre grösste Uppigkeit.

Das endgültige Zuwachsen des Beckens ist am Schwinden der Wasserpflanzen in der Lagerfolge zu erkennen. Auf das Zuwachsen folgte eine trockene Periode. Der Birkenbestand vermochte sich wieder auf das Moor

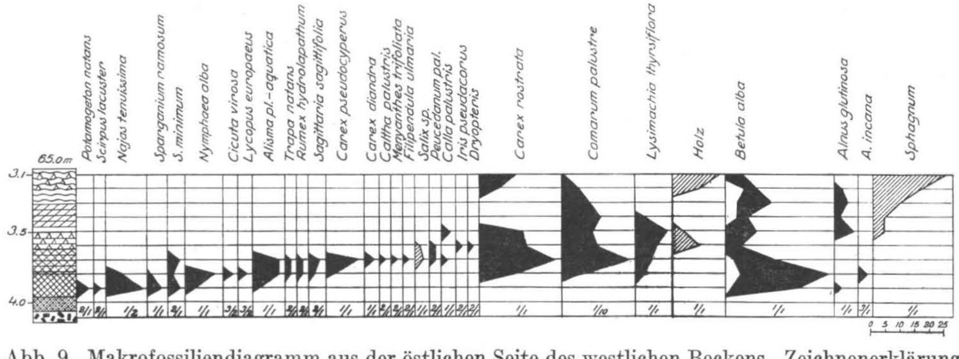

Abb. 9. Makrofossiliendiagramm aus der östlichen Seite des westlichen Beckens. Zeichnenerklärung in Abb. 4.

vorzuschieben (vgl. Abb. 8 und 10). Eine starke *Sphagnum*-Bedeckung brachte jedoch die Birkeninvasion zum Stocken. Das Birkenfeld verschmälert sich entschieden in dem Niveau, in dem das *Sphagnum*-Feld einsetzt.

Die Erlenvorkommen in den Makrofossiliendiagrammen begleiten den Gang des Birkenfeldes. Der erste Gipfel der Erle fällt in Grobdetritusgyttja, wie auch der der Birke. Nach dem Zuwachsen des Beckens erreicht Erle zusammen mit Birke ein deutliches Vorkommen. Diesen Holzarten erschien ein starker Wettbewerber erst im Zuwachsungsstadium des Beckens und später mit beginnender eigentlicher *Sphagnum*-Bedeckung. Auf Grund der Pollendiagramme kann geschlossen werden, dass die Fichte bei den besagten Stadien eine bedeutende Stellung im Bestande der Wälder in der Umgebung des Moores erlangte. Als vitale Holzart frischer Böden verdrängte die Fichte Birkenbestände, und die offenbar von Schwarzerle *(Alnus glutinosa)* beherrschten Moorrandwaldungen wurden von Fichte besetzt, als das Klima zu Beginn der subatlantischen Zeit für diese Holzart günstig wurde. Dieser waldgeschichtliche Vorgang ist in den Pollendiagrammen als starker Anstieg der Fichtenpollenkurve zu erkennen. Das gleichzeitige Abnehmen von Schwarzerle und edlen Laubbäumen beruht zum mindesten nicht unmittelbar darauf, dass das Klima für diese Holzarten sehr nachteilig geworden wäre, sondern die Fichte hat die früheren Lebensräume der Laubbäume erheblich eingeschränkt.

In den Makrofossiliendiagrammen ist eine gewisse Gruppierung der Pflanzenarten wahrzunehmen, die am deutlichsten in der Flora der Grobdetritusgyttja hervortritt (Abb. 10 und 11). In ihr lassen sich eine untere, eine mittlere und eine obere Gruppe sowie ferner die das ganze Gyttja-

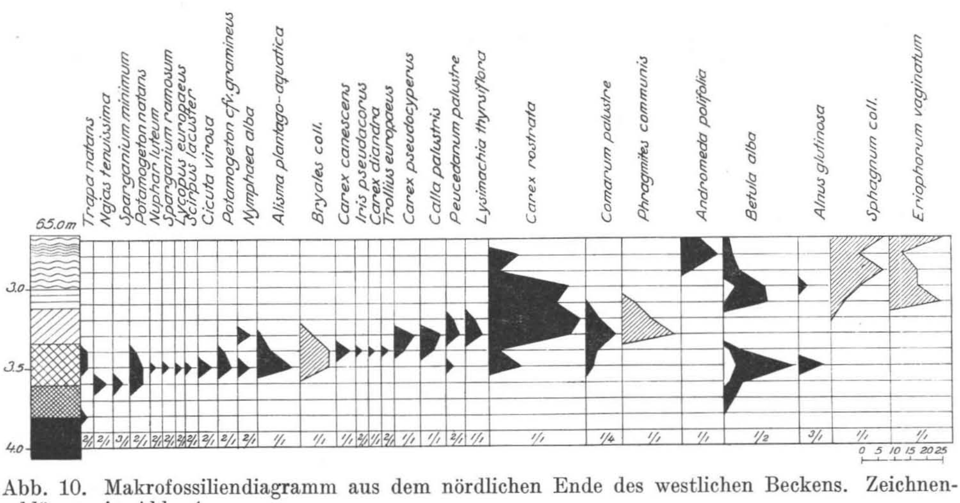

Abb. 10. Makrofossiliendiagramm aus dem nördlichen Ende des westlichen Beckens. Zeichnenerklärung in Abb. 4.

sediment umfassenden Arten als eigene Gruppe unterscheiden. Diese Pflanzengruppierungen vertreten verschiedene Sukzessionen in der Entwicklung des Beckens. Doch ist anzuführen, dass nicht die gleichen Arten in den verschiedenen Teilen des Beckens diese Gruppen bilden. So können besonders Wasserpflanzen, wie *Trapa*, an einer Stelle des Beckens durch Sinken des Wasserspiegels ihre grösste Frequenz erreichen, gleichzeitig aber irgendwo anders im Gebiet des Beckens schwinden.

Die Entwicklung der Anfangsphase des Moores lässt sich auf Grund eines am flachen Rand des Moränenrückens erbohrten Profils und des ihm entsprechenden Makrofossiliendiagrammes eingehender verfolgen (Abb. 9). Auf Geröll lagert makrofossilleerer gyttjahaltiger Schluffton. Ihn bedeckt eine 16 cm starke Gyttjaschicht, die in gyttjahaltigen Seggentorf mit groben Pflanzenresten übergeht. Auf diesem liegt holzrestführender Seggentorf, der seinerseits in limnischen Torf mit reichlichen *Carex rostrata* und *Comarum palustre* übergeht. In der Lagerfolge ist zuoberst holzführender *Sphagnum*-Torf zu sehen, der sich jedoch bald in reinen *Sphagnum*-Torf wandelt.

Der Makrofossiliengehalt beginnt mit einer Reihe von sechs eigentlichen Wasserpflanzen *(Potamogeton natans, Scirpus lacuster, Najas tenuissima, Sparganium ramosum, S. minimum* und *Nymphaea alba)*. Die Reihe endet mit einem aus zwei Uferpflanzen *(Cicuta virosa* und *Lycopus europaeus)* bestehenden Vorkommen in einer Tiefe, in der die Birke einen grossen Gipfel erreicht. Wieder folgt eine hydrophytische Reihe *(Alisma, Trapa, Rumex hydrolapathum* und *Sagittaria sagittifolia)*, die ihrerseits in einem aus vielen Uferpflanzen bestehenden üppigen Pflanzenbestand *(Carex pseudocyperus, C. diandra, Caltha, Calla, Filipendula, Dryopteris* und *Iris)* endet. Die Holz enthaltende Bodenart im Minimum des Birkenfeldes ist Akkumulationstorf. Die Holzstückchen sind darin glattgeschliffen, und sie erscheinen ebenso wie das vom Überschwemmungswasser auf das Ufer

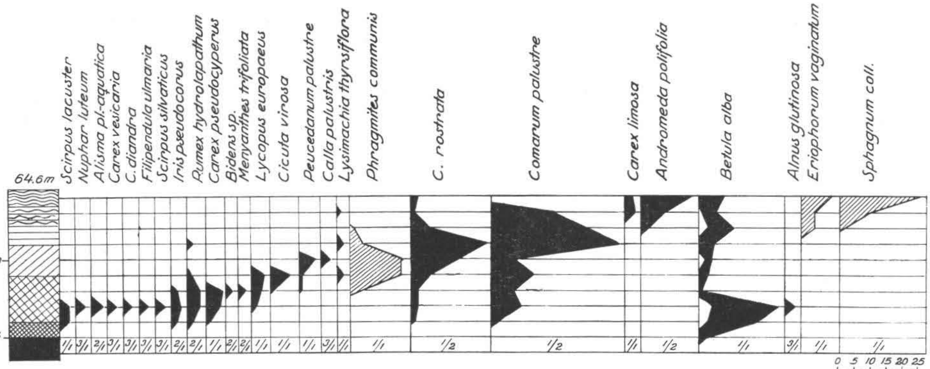

Abb. 11. Makrofossiliendiagramm aus der Gegend der Ufervermoorung. Zeichnenerklärung in Abb. 4.

getriebene Material. Dieses Holzmaterial im Torf bezeugt also kein Bewalden des Moores, sondern umgekehrt eine Überschwemmung, die eine Moorrandwaldung vernichtete und zugleich den Waldrand weiter auf den Festboden verschob. Die durch das Diagramm vertretene flachwässerige Uferzone hat sehr empfindlich auf die in den Höhenverhältnissen des Wassers eingetretenen Veränderungen reagiert und zugleich bald den Ufer-, bald den Wasserpflanzengesellschaften günstige Verhältnisse geboten.

ÜBER DEN ANTEIL DER STANDORTSFAKTOREN AN DER VORZEITLICHEN FLORA UND VEGETATION

KLIMATISCHE FAKTOREN

Jede Pflanzenart hat ihre jeder nacheiszeitlichen Periode entsprechenden Verbreitungsgebiete gehabt (Kalela 1949). Die Grenzen dieser Gebiete haben sich letztens nach dem jeweils bestehenden Klima gerichtet. Der Einfluss des Klimas ist am deutlichsten in der ehemaligen Verbreitung der sog. wärmebedürftigen Pflanzenarten zu erkennen. Solche Pflanzenarten, die eine verhältnismässig hohe Temperatur erfordern, haben sich kaum weit über ihr gegenwärtiges Wohngebiet hinaus nach Norden ausbreiten können, wenn nicht das Klima auch dort ihrem Wärmebedarf entsprochen hat. Auf der anderen Seite hat man sich daran zu erinnern, dass die Pflanzen im allgemeinen die Möglichkeit gehabt haben, sich durch ihre Diaspora gleichmässig auszubreiten über das Gebiet, das klimatisch für sie geeignet gewesen ist (Jalas 1950, S. 180).

Ein Vergleich zwischen den einstigen und gegenwärtigen Wohngebieten und Standorten der im Makrofossiliengehalt des Moores Valmossa vorkommenden Pflanzenarten erweist, dass sich in der Verbreitung von *Trapa* die grösste Wandlung vollzogen hat. Sie ist in der späteren Litorinazeit in den Gewässern der südlichen Hälfte Finnlands ebenso häufig aufgetreten wie gegenwärtig in Südeuropa. Die Überreste von *Trapa* sind denn auch der überzeugendste Beweis für das warme und den Pflanzen günstige Klima, das in der Litorinazeit in Finnland bis nach Süd-Pohjanmaa geherrscht hat.

Neben dem in weitem Raume bestehenden allgemeinen Klima ist das örtliche, das Mikroklima, besonders für das Auftreten klimatisch empfindlicher Pflanzenarten von grosser Bedeutung. Die an der Peripherie ihres Verbreitungsgebietes gewachsenen Arten, wie *Trapa natans*, haben mikroklimatisch und edaphisch günstige Standorte aufgesucht, auf denen sie zusammen mit den übrigen Arten haben fortkommen können.

Zur Zeit des Auftretens der Wassernuss war das Moor Valmossa eine seichte, am Unterlauf des Ähtävänjoki gelegene lagunenartige Bucht des

Evijärvi. Schützende bewaldete Moränenböden umgaben es auf drei verschiedenen Seiten. In einem derartigen Becken mit flachem Wasser erwärmte sich dieses schon zeitig im Frühjahr, bot dadurch einen guten Standort und verbürgte für das Reifen der Früchte eine hinreichend lange Vegetationsperiode.

Das Abkühlen des Klimas nach beendeter Litorinazeit wurde am verderblichsten für die an der äusserten Grenze ihres Verbreitungsgebietes gewachsenen Arten. Das Verzögern der beginnenden Vegetationsperiode, der oft darauf folgende kurze und kühle Sommer sowie die Winterkälte, bei der die seichten Gewässer bis auf den Grund gefroren sind, haben wahrscheinlich als entscheidende Faktoren in Finnland auf das Geschick nicht nur der Wassernuss, sondern auch anderer einjähriger Wassepflanzen (*Najas*-Arten) eingewirkt.

EDAPHISCHE FAKTOREN

Der Einfluss edapischer Faktoren ist in der gegenwärtigen Pflanzendecke in der Umgebung des Untersuchungsgebietes auf mancherlei Weise zu erkennen. Saksela (Saxen 1928) hat in der Nachbargemeinde von Evijärvi, in Teerijärvi, den üppigen Hainwiesen und einzelnen Anbaugebieten, die inmitten trockenen und kargen Heidebodens auftreten, Aufmerksamkeit zugewandt. Diese örtliche Uppigkeit ist vor allem auf den verhältnismässig hohen und leichtlöslichen Kalkgehalt der im Felsgrund anstehenden Amphibolitvorkommen (vgl. Abb. 1) zurückzuführen. Die Pflanzen nehmen zwar nicht unmittelbar Kalk aus dem Felsgrund auf, sondern aus Moräne, deren Gesteine nach der Untersuchung Sakselas die Beschaffenheit des sie unterlagernden Felsgrundes deutlich widerspiegeln.

Soweit die genannten Amphibolitvorkommen in der Umgebung des Moores Valmossa auf die Üppigkeit der Vegetation einwirken, muss dieser Einfluss infolge der Kleinheit der Vorkommen verhältnismässig gering und auf beschräktem Raume erkennbar bleiben. Die westlich der Kirche von Evijärvi gelegene Bucht (Abb. 1) scheint indessen ständig dem nahen Amphibolitvorkommen entstammendes Material aufzunehmen. Maristo (1941, S. 88) führt nämlich an, dass die besagte Bucht, in der die Tiefe des Wassers gegenwärtig etwa 1 m und pH 6.8 ausmacht, in ihrer Üppigkeit durchaus von dem eigentlichen Hauptsee abweicht. Neben dominierendem *Scirpus lacuster* gibt es dort ein einziges grosses *Typha angustifolia*-Vorkommen sowie in sehr reichlichem Masse runde *Sparganium ramosum*-Ringsiedlungen, wie sie anderswo im See nicht anzutreffen sind. Maristo hat denn auch diese kleine Bucht des Evijärvi der eutrophen *Typha-Alisma*-Gruppe angeschlossen.

Die in der Umgebung des Evijärvi gegenwärtig örtlich erscheinende Üppigkeit zeigt sich deutlich auch in der ehemaligen Flora. So ist *Najas* *tenuissima*, die von Backman (1950, S. 28) als anspruchsvollste Art im Makrofossiliengehalt Finnlands angesehen wird, im westlichen Becken des Moores Valmossa sehr häufig gewesen. Sie ist in 13 Bohrungspunkten angetroffen worden. Auch viele andere eutrophe Pflanzenarten, wie *Carex pseudocyperus* (14), *Sparganium ramosum* (12), *Sagittaria sagittifolia* (11), *Lycopus europaeus* (10), *Iris pseudacorus* (6) und *Rumex hydrolapathum* (6), sind besonders im westlichen Becken ziemlich häufig vorgekommen.

Die edaphischen Verhältnisse sind in dem Entwicklungsstadium des Sees, bei dem sich Grobdetritusgyttja im Becken absetzte, am günstigsten gewesen. Das westliche Becken hatte eine verhältnismässig geringe Ausdehnung und seichtes Wasser, dessen Konzentration, Elektrolytgehalt im Laufe des Sommers zunahm. Die reichhaltige Vegetation steigert ihrerseits die Eutrophie des Beckens, denn als Zersetzungsprodukte von Pflanzenteilchen geraten ständig Pflanzennährstoffe zurück in das Wasser. Die geringe Tiefe des Wassers wiederum beschleunigt den Zersetzungsprozess, denn das Wasser des flachen Beckens erwärmt sich im Sommer bis auf den Grund und fördert dadurch die Tätigkeit der für das Zersetzen der Pflanzenreste sorgenden Mikroorganismen (vgl. Steinecke 1940, S. 165—166).

Der Kreislauf der Pflanzennährstoffe blieb bis gegen Ende der Litorinazeit lebhaft. Das Moor Valmossa hat ebenso wie auch manches andere in der Wärmezeit aus dem Meere aufgestiegene Becken in Pohjanmaa zur Zeit seines Binnenseestadiums in dieser Hinsicht eine günstige Stellung eingenommen.

Das Abkühlen des Klimas nach beendeter Wärmezeit bewirkte ein Verlangsamen in der Tätigkeit der Mikroorganismen. Der Zersetzungsprozess der Pflanzenreste nahm ab. Stets grössere und unzersetzte Pflanzenteilchen begannen in schnellem Tempo die Becken zu füllen. Nach ihrer Zuwachsung stockte der Kreislauf der Pflanzennährstoffe. Je mächtiger das eigentliche Torflager im Moore wurde, desto mehr Nährstoffe wurden an den Torf gebunden und der Reichweite der Pflanzen entzogen. Der Standboden verarmte, und eine unmittelbare Folge davon war eine starke Expansion oligotropher *Sphagnum*-Vegetation auf die früheren üppigen Standorte.

Bei Betrachtung der Faktoren, die auf die Eutrophie der vorzeitlichen Vegetation des Moores Valmossa eingewirkt haben, kan neben den örtlichen Faktoren auch das weite Gebiet berücksichtigt werden, aus dem die Wassermassen des Evijärvi eingezogen sind. Das Moor Valmossa lag zur Zeit seines Binnenseestadiums am Unterlauf des Ähtävänjoki als lagunenartige Bucht des Evijärvi. Zum Niederschlagsgebiet des Ähtävänjoki gehören basische Kalksteingebiete, wie die Kalksteinvorkommen von Vimpeli, von denen das Wasser durch den Lappajärvi in den Evijärvi fliesst. Als die allgemeine Vermoorung noch nicht bis zu seiner gegenwärtigen Ausdehnung vorgeschritten war, speiste das aus dem Lappajärvi kommende nährstoffreiche und wahrscheinlich nur schwach saure Wasser die Vegetation des Evijärvi und seiner Buchten. Linkola (1933 b) hat eine Untersuchung ausgeführt, deren Gegenstand die Vegetation dreier zu einem und demselben Gewässersystem gehörenden, aber in verschiedenen Höhen gelegenen Seen gewesen ist. Die Arbeit erweist, dass die am Unterlauf des Seensystems gelegenen Gegenden oft fruchtbarer als die höher gelegenen sind. Nach Linkola dürften die günstigeren Verhältnisse der unteren Glieder eines Seensystems in hohem Grade, öfters sogar ausschliesslich durch folgende sekundäre Faktoren veranlasst sein: 1. durch Schlammtransport, welcher besonders im Frühling die Feinerde nicht nur in Bächen und Flüssen sondern auch durch kleinere Seen weiterführt, und eine darauf folgende Sedimentation des Transportmaterials, dessen tonige Bestandteile hauptsächlich die niedriger liegenden Seen nach und nach eutrophieren; 2. durch allmähliche Ausfällung der Humuskolloide, die eine Abnahme der Humussäuren und zugleich der Azidität des Seewassers in den unteren Seen bedingt; 3. durch Verdunstung, die die Nährsalzkonzentration des Wassers nach dem unteren Lauf der Seereihen zu erhöht.

Die Wasser- und Uferpflanzen sind nicht allein von der Beschaffenheit des Beckengrundes, sondern auch von dem diese Pflanzen umgebenden Wasser und seiner Konzentration abhängig. Die eutrophe Flora konnte daher nur gedeihen, solange nährstoffreiches Wasser unbehindert Zugang fand. Nachdem die Becken zugewachsen und die Mooroberfläche infolge der Torfbildung über den Grundwasserstand gestiegen war, gelangte das Wasser nicht mehr in Reichweite der genannten Pflanzen. Da sich auch im Klima zugleich eine deutliche Wandlung in schlechterer Richtung vollzog, schwand auch schon aus diesem Grunde die wärmezeitliche Üppigkeit der Pflanzenwelt im Gebiet des Moores Valmossa.

DIE TIEFENVERHÄLTNISSE DES WASSERS, DER MORPHOLOGISCHE BAU DER UFERLINIE UND DIE HIMMELSRICHTUNGEN IN IHREM EINFLUSS AUF DIE FLORA

Die Tiefenverhältnisse des Sees sowie der morphologische Aufbau des Ufers sind von sehr grosser Bedeutung für die Entwicklung der Vegetation im allgemeinen (vgl. Donat 1926, S. 50, Maristo 1941, S. 206—210, Jaatinen 1950, S. 204—208). Dieser Sachverhalt tritt auch bei den Makrofossilienuntersuchungen deutlich hervor. Bei den in tiefem Wasser entstandenen Sedimenten, wie Ton, Tongyttja und oft auch Feindetritusgyttja, ist der Makrofossiliengehalt spärlich und oligotroph, auch wenn der Nährstoffgehalt und die pH-Stufe des Wassers eine anspruchsvolle Vegetation zu befriedigen vermöchten. Bei Flachwasserablagerungen, wie Grobdetritus-gyttja und limnischem Torf, steigen die Artenfrequenz und die Gesamtmenge der Makrofossilien erheblich hoch.

Das Gefälle des Litoralgebietes ist ebenfalls von beträchtlicher Bedeutung. Auf einem flachrandigen Ufer erreicht die Vegetation quantitativ wie auch qualitativ ihre grösste Üppigkeit. Auf einem steilrandigen Ufer bleibt die Vegetation spärlich, einerlei ob es sich um einen oligotrophen oder eutrophen Seetyp handelt (Jaatinen 1950, S. 206).

Bei Erforschung der Flora eines Vorzeitsees ist der horizontale Aufbau der Uferlinie genau zu beachten. Maristo (1941, S. 209) hat bemerkt, »je grösser die Uferentwicklung, d. h. je geschlängelter die Uferlinie eines Sees ist, desto bessere Entwicklungsmöglichkeiten stehen der Wasservegetation zu Gebote, vorausgesetzt, dass die edaphischen Faktoren überall die gleichen sind». Dies spiegelt sich deutlich auch in der Subfossilflora. Das westliche Becken sowie der nördliche Teil der Ufervermoorung des Moores Valmossa sind geschützte kleine Buchten und innerste Buchtteile gewesen, in denen sich die Vegetation in Ruhe hat entwickeln können. Ausserdem haben die vielen Inseln und Halbinseln des Sees die Vegetation vor der Wirkung der Wellen geschützt. So treten *Trapa natans* und *Najas tenuissima* nur im Gebiet des westlichen Beckens auf (vgl. Tabelle 2). Der vorzeitliche Standort von *Trapa* im Becken des Moores Valmossa entspricht denn auch durchaus der Schilderung Eberles (1926, S. 166) über ihre Standorte in Mitteleuropa, »die Standorte der Wassernuss sind stille Wasser, seeartige Bildungen, meist im Bereich der grossen Ströme, sehr viel seltener Seen und Teiche abseits der grossen Wasserstrassen. Unter diesen stehen in Deutschland die Standorte im Rheingebiet und an der Elbe im Vordergrund.»

Auch die Himmelsrichtungen haben ihren eigenen Anteil an der Frequenz und der Üppigkeit der Flora. Aario (1933, S. 98) hat diesem Sachverhalt bei seinem Erforschen der gegenwärtigen Vegetation des Sees Nurmijärvi Aufmerksamkeit zugewandt. Am nördlichen Ende des Sees ist die Vegetation üppig, aber an seinem südlichen Ende sehr spärlich. Dies ist durch das hohe und bewaldete Kleinhügelland am Südufer bedingt. Morgens bleibt das Südufer lange im Schatten, und abends verschwindet die Sonne früh hinter dem Wald. Die Topographie der Umgebung des Moores Valmossa sowie die Verteilung der Makrofossilien erweisen, dass die Verhältnisse dort ähnlich wie heute am Nurmijärvi gewesen sind. Die anspruchsvollen Pflanzenarten sind im nördlichen Beckenteil gut gediehen, aber im südlichen fehlen sie. Besonders das Fehlen von Überresten der Wassernuss im südlichen Teil des Beckens ist ein Beweis für den wichtigen Anteil der Himmelsrichtungen an der Vegetation des Vorzeitsees.

LITERATURVERZEICHNIS

AARIO, L. (1932) Pflanzentopographische und paläogeographische Mooruntersuchungen in N-Satakunta. Fennia 55.

—»— (1933) Vegetation und postglaciale Geschichte des Nurmijärvisees. Ann. Bot. Soc. 'Vanamo' 3.

APINIS, A. (1940) Untersuchung über die Ökologie der Trapa L. I Teil. Systematische Zugehörigkeit, Geschichte, Standortsverhältnisse und die Verbreitung der Pflanze in Lettland. II Teil. Bedingungen der Keimung und Entwicklung der Pflanze. Acta Horti Bot. Univ. Latviensis 13.

AUER, V. (1921) Zur Kenntnis der Stratigraphie der mittelösterbottnischen Moore. Acta Forest. Fennica 18.

—»— (1924) Die postglaziale Geschichte des Vanajavesi-Sees. Bull. Comm. géol. Finlande 69.

—»— (1925) Investigations of the ancient flora of Häme. Comm. Inst. Quaest. Forest. Finlandiae 9.

—»— (1928) Über die Einwanderung der Fichte in Finnland. Comm. Inst. Quaest. Forest. Finlandiae 13.

AUROLA, E. (1938) Die postglaziale Entwicklung des südwestlichen Finnlands. C. R. Soc. géol. Finlande 11. Bull. Comm. géol. Finlande 121.

BACKMAN, A. L. (1941) Najas marina in Finnland während der Postglazialzeit. Acta Bot. Fennica 30.

—»— (1943) Ceratophyllum submersum in Nordeuropa während der Postglazialzeit. Acta Bot. Fennica 31.

—»— (1948) Najas flexilis in Europa während der Quartärzeit. Acta Bot. Fennica 43.

—»— (1950) Najas tenuissima (A. Br.) Magnus einst und jetzt. Sos. Sci. Fennica Comm. Biol. X. 19.

BRANDT, A. (1948) Über die Entwicklung der Moore im Küstengebiet von Süd-Pohjanmaa am Bottnischen Meerbusen. Ann. Bot. Soc. 'Vanamo' 23.

DONAT, A. (1926) Die Vegetation unserer Seen und die »biologischen Seentypen». Ber. Deutsch. Bot. Ges. 44.

EBERLE, G. (1926) Einige über die Wasser- oder Spitznuss (Trapa natans). Natur und Museum 6.

—»— (1927) Die Entwicklung der Wassernusspflanze (Trapa natans L.) von reifen Frucht bis zum Auftauchen der Blattrosette. Natur und Museum 1—12.

HIITONEN, I. (1933) Suomen Kasvio. Helsinki.

HYYPPÄ, E. (1935) Kittilän lusikkalöydön geologien iänmääräys. Suomen Museo.

—»— (1937) Post-glacial changes of shore-line in South-Finland. Bull. Comm. géol. Finlande 120.

—»— (1950) Lapuan ja Pattijoen muinaissuksien geologinen ikäys. Suomen Museo.

JAATINEN, S. (1950) Bidrag till kännedom om de Åländska sjöarnas strandvegetation. Acta Bot. Fennica 45.

JALAS, J. (1950) Zur Kausalanalyse der Verbreitung einiger nordischen Os- und Sandpflanzen. Ann. Bot. Soc. 'Vanamo' 24.

KALELA, A. (1949) Mistä ja milloin Suomi on saanut kasvistonsa? Suomen Luonto 8.

LAITAKARI, A. (1942) Suomen geologinen yleiskartta. Lehti B 3. Vaasan kivilajikartan selitys. Suomen Geologinen Toimikunta, Helsinki.

LINKOLA, K. (1933 a) Regionale Artenstatistik der Süsswasserflora Finnlands. Ann. Bot. Soc. 'Vanamo' 3.

LINKOLA, K. (1933 b) Über die Flora und Vegetation in den Seen Ylä-, Keski- und Ala-Vekarainen im Kirchspiel Sulkava, Mittel-finnland. Ann. Bot. Soc. 'Vanamo' 3.

LUMIALA, O. V. (1940) Zwei Moorprofile aus Siilinjärvi, dem nördlichsten bekannten Fundort fossiler Trapa natans in Finnland. Bull. Comm. géol. Finlande 15.

MARISTO, L. (1941) Die Seetypen Finnlands auf floristischer und vegetations-physiognomischer Grundlage. Ann. Bot. Soc. 'Vanamo' 15.

Mölder, K. (1946) Zwei Tonprofile aus Süd-Pohjanmaa. C. R. Soc. géol. Finlande 19. Bull. Comm. géol. Finlande 138.

OKKO, V. (1949) Suomen geologinen yleiskartta. Lehti B 4. Kokkola. Maalajikartan selitys. Geologinen tutkimuslaitos.

PALMGREN, A. (1912) Hippophaës rhamnoides auf Åland. Acta Soc. Fauna et Flora Fennica 36.

SALMI, M. (1948) Die Ancylustransgression in dem Moore Hangassuo im Süd-Finnland. C. R. Soc. géol. Finlande 21. Bull. Comm. géol. Finlande 142.

SAURAMO, M. (1941) Die Geschichte der Welder Finnlands. Geol. Rundschau 32.

—»— (1958) Die Geschichte der Ostsee. Ann. Acad. Sci. Fennicae, A, III. Geol.-Geogr. 51.

SAXEN (SAKSELA), M. (1928) Über den Einfluss des Gesteingrundes auf die Vegetation. Beobachtungen aus einem Moränengebiete im Mittleren Österbotten. Fennia 50.

SELLING, O. (1938) Entwicklungsgeschichtliche Studie im Molken-See mit besonderer Rücksicht der Frequenzwechsel der Makrofossilien. Geol. Fören. i Stockholm Förh. 60.

STEINECKE, F. (1940) Der Süsswassersee. Die Lebensgemeinschaften des nährstoffreichen Binnensee. Studieb. Deutsch. Lebensgem. I. Leipzig.

VALOVIRTA, E. J. (1937) Untersuchungen über die säkulare Landhebung als pflanzengeographischer Faktor. Acta Bot. Fennica 20.

—»— (1950) Über die Strandhöhenzonen als Hilfsmittel bei der Schärenflora im Landhebungsgebiet. Acta Bot. Fennica 47.

VALOVIRTA, V. (1957) Vesipähkinästä (Trapa natans L). Geologi 1.

VIRKKALA, K. (1950) Kuusen yleistymisen ajankohta Länsi-Suomessa. Terra 1.

5

SERPENTINITE OF PAHTA-AUTSI, FINNISH LAPLAND [1]

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INTRODUCTION

Mikkola and Sahama (1936) have described the area south-west of the granulite belt of Finnish Lapland. Within this region there are several scattered minor ultrabasic bodies, embedded in different gneisses. These bodies have been examined by the authors cited, who also mentioned among them the group of eight small outcrops of ultrabasic rocks occurring in the gorge of Pahta-autsi, at the southern end of the lakelet of Pahtajärvi (Fig. 1). This occurrence is situated at the uppermost course of the Repojoki River, approx. six km east-southeast of the sharp bend in the Finnish Norwegian frontier on the fjeld of Peltotunturi.

The gorge of Pahta-autsi is narrow and deep, and it

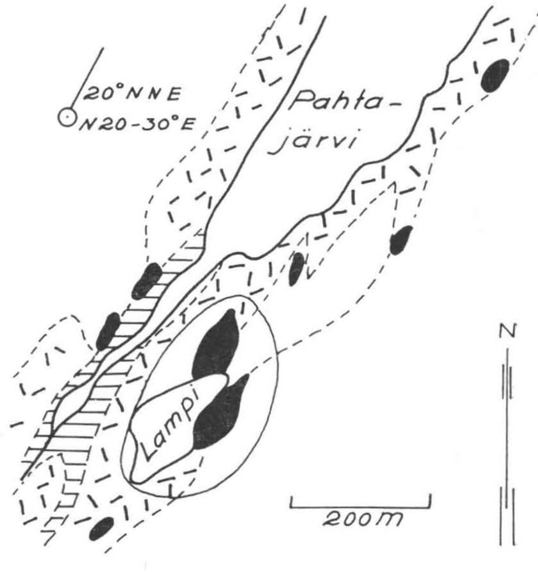

Fig. 1. Ultrabasic outcrops at Pahta-autsi, according to Mikkola and Sahama (1936).

[1] Received Dezember 23, 1959.

runs through the flat and smooth, elevated fjeld terrain. From the bottom of the gorge, the ultrabasic bodies rise as hillocks, comparatively high in the topography, but not exceeding the height of the walls of the gorge.

According to Mikkola and Sahama, the ultrabasic bodies outcropping along the southeastern wall of the gorge consist, mainly or exclusively, of serpentinite; those occurring along the northwestern side are mainly made up of amphibole. Furthermore the aforesaid researchers wrote (op. cit. p. 36): »Here it is quite obvious that the greatest elongation of the bodies is the same as the direction of the pronounced rodding structure in the gneisses, which has a rather gently-dipping position.»

In 1959 this remote locality was visited by the present writer. He carried out a detailed study of the serpentinites on the southern side of the Pahtajärvi lakelet (the encircled area in Fig. 1). The results of this re-examination are presented in Fig. 2 and in the following pages.

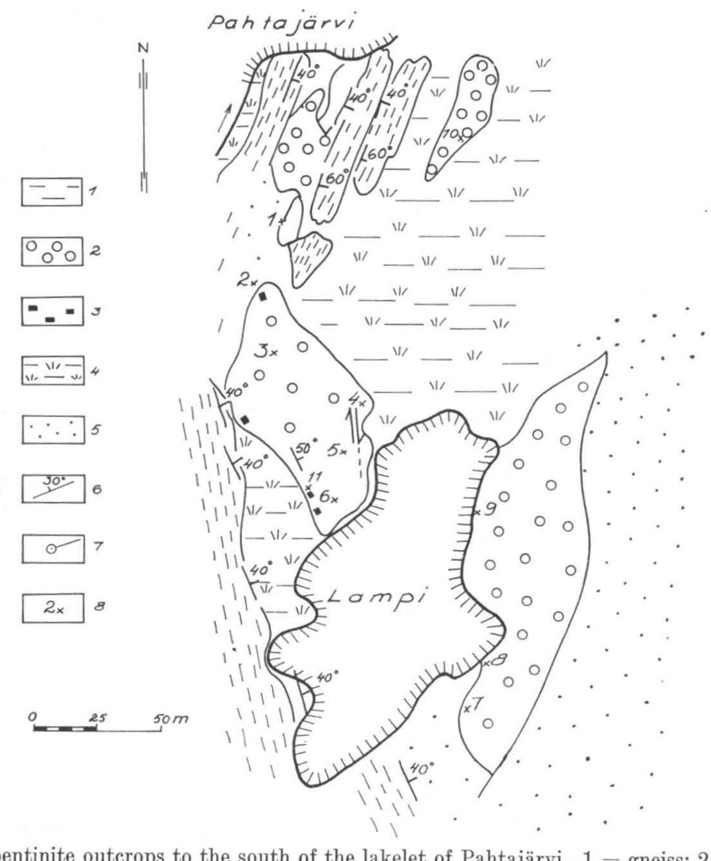

Fig. 2. Serpentinite outcrops to the south of the lakelet of Pahtajärvi. 1 = gneiss; 2 = serpentinite; 3 = abundant magnetite veinlets; 4 = swampy cover; 5 = masked by overburden; 6 = dip and strike; 7 = folding axis; 8 = observation point.

GEOLOGY

The gorge of Pahta-autsi is mainly made up of a quite nonhomogeneous gneissose rock consisting of alternating layers and bands of amphibolite, hornblende gneiss, andalusite-bearing mica gneiss, and quartzose aplitic rock. The thickness of the single bands varies from a few millimeters up to 1 meter.

Approx. 1 km to the south of the area illustrated in Fig. 2, there occurs a pink, aplitic microcline granite. It forms dykes which traverse the gneiss.

The general strike of this mixed gneiss is roughly S-N, but there are ample local deviations of up to 20 degrees on both sides of the northern direction.

The dip is gentle. In general it is easterly and averages 30—40°, but locally the position of the beds is subhorizontal (Fig. 3).

The gorge itself is definitely a tectonic rupture zone. The vertical walls of the gorge serve as excellent geological sections; and owing to the steepness of the walls, the ultrabasic bodies have actually been exposed. The outcrops outside the gorge are few, and therefore it is difficult to detect the structure of the folding of the strata there. The general direction of the folding axis there seems to be N20°E, and the pitch varies from 10 to 20 degrees.

In the strata, the serpentinite bodies occupy places where the gneiss has been slightly bent, thereby opening up space for serpentinites. To the northwest of the small lake shown in Fig. 2 on the western side of the serpentinite body, the lower serpentinite contact seems to conform perfectly well with the local structure of the gneiss, and the dip of the contact is 40° towards the east. Northwards from this place, the strike of the strata turns towards

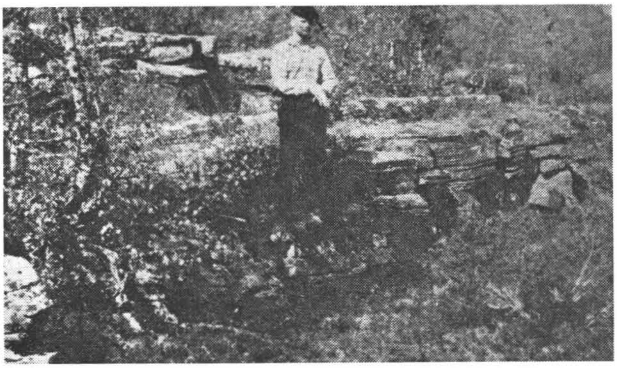

Fig. 3. Subhorizontal position of the gneissose beds. The southern shore of the lakelet of Pahtajärvi.

the northeast; but southwards from the exposed contact, the strike of the gneiss is southeasterly.

In the northern part of Fig. 2, the gneiss seems to be inside the serpentinite. The westerns contact of this gneiss tongue is visible. It is the upper contact of the serpentinite body and shows the conformity of the serpentinite to the gneiss. Furthermore the contact steepens southwards from 40 to 60 degrees.

Elsewhere, the upper contacts of the serpentinite body are masked by drift or bog. In the serpentinite body itself, some kind of bedding likewise occurs, and this seems to be parallel with the layers of gneiss.

From these observations it may be deduced that the serpentinite bodies shown in Fig. 2 belong to the same lense shaped mass which shows up as a concordant body between gneiss layers.

Its continuation northwards, from the area shown in the map of Fig. 2, is hardly expectable. Towards the southeast, however, the form of the serpentinite body is entirely masked, and therefore also its continuation in this direction is unknown. There the last serpentinite outcrop is exposed in the wall of the gorge. Along the gorge, the nearest outcrop consists of gneiss. In the direction of the strike of serpentinite, however, the nearest outcrop, which is also made up of gneiss, is far outside the gorge. Therefore, only the minimum dimensions of the serpentinite body can be given. It is at least 200 meters in length, and the width of its horizontal section is more than 80 meters. The actual thickness of the serpentinite body, as constructed from the field data of Fig. 2, approximates 50 or 60 meters.

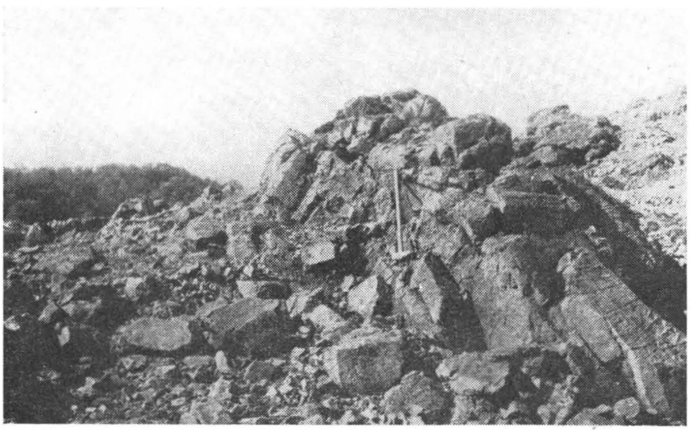

Fig. 4. The top of the serpentinite hillock (point 4 of Fig. 2). The form of disintegration is mainly dictated by the semi-northerly stretching joint.

The highest serpentinite hillock is to the northwest of the lakelet (Fig. 2). This is entirely barren except for a few typical serpentinite plants growing along cracks and minor depressions in the rock. Especially the top of the hillock is void of any kind of vegetation. There the rock has broken along the joints rending south-north (points 4 and 5 of Fig. 2), and the top resembles some kind of quarry (Fig. 4).

PETROLOGY OF THE SERPENTINITE

On the average, the serpentinite is fairly homogeneous and, as mentioned, it forms some kind of »bedding», mainly represented by parallel fracturing of the rock concordant to the schistosity of the embracing gneisses.

The serpentinite consists mainly of antigorite and chrysotile with a few needles of tremolite. In some of the »beds» minute grains of olivine have been met with. In most parts of the bodies, olivine occurs sporadically, but in patches may be rather abundant, as, for instance, in some thin sections made from the serpentinite outcropping on the eastern shore of the lakelet as well as on the top of the western outcrop (point 5 of Fig. 2), where calcite is also abundant. The olivine forms minute, wellrounded grains. There is no sign of any kind of serpentinization of the olivine, which has obviously crystallized together with the antigorite.

At the southern end of the northern exposures (point 1 of Fig. 2) occurs a true soapstone consisting of abundant calcite and antigorite, which contains tiny needles of tremolite. At point 3, prisms of anthophyllite penetrating the antigorite have been seen under the microscope.

The petrologically fairly uniform serpentinite is usually cut by several narrow zones filled by chlorite and chrysotile. These zones have probably been produced by a slight shearing. Narrow chrysotile veinlets likewise occur throughout the whole serpentinite mass, and sometimes they have been developed into asbestos, though not of any good quality.

The chlorite zones mentioned may appear in very different directions. At the northwestern end of the serpentinite body on the western shore of the lakelet, the shear-zones have produced a peculiar structure resembling that of pillow lavas (Fig. 5). The thickness of the beds here is approx. 1 meter, and they

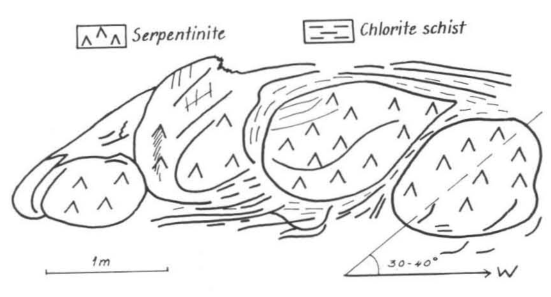

Fig. 5. Pillow-like structures of the serpentinite. Point 3 of Fig. 2.

consist of large »pillows» which are slightly elongated, the long axis pitching 30 to 40 degrees in a subeastern direction. These bodies also consist mainly of antigorite.

A well-foliated chlorite embraces the »pillows» there and separates them from each other. The chlorite portions are 5 to 10 cm thick. Also the »pillows» themselves are often penetrated by shear fractures filled with chlorite.

Well-developed chlorite shear planes occur elsewhere within the serpentinite bodies as well. One of these is shown in Fig. 2 by reverse double arrows, near the northwestern shore of the lakelet (between points 4 and 5). Another shearzone is seen at point 6 in the same figure. There the shearzone, filled with chlorite, is 40 cm thick, and it is subparallel to the »bedding» of the serpentine. Along the upper margin of the chlorite, there occurs a 10-cm-thich portion of coarse calcite.

MAGNETITE VEINLETS

Especially along the western edge of the central part of the serpentinite body here considered, there occur peculiar veinlets consisting of magnetite (Fig. 6). Among them, two types are to be distinguished:

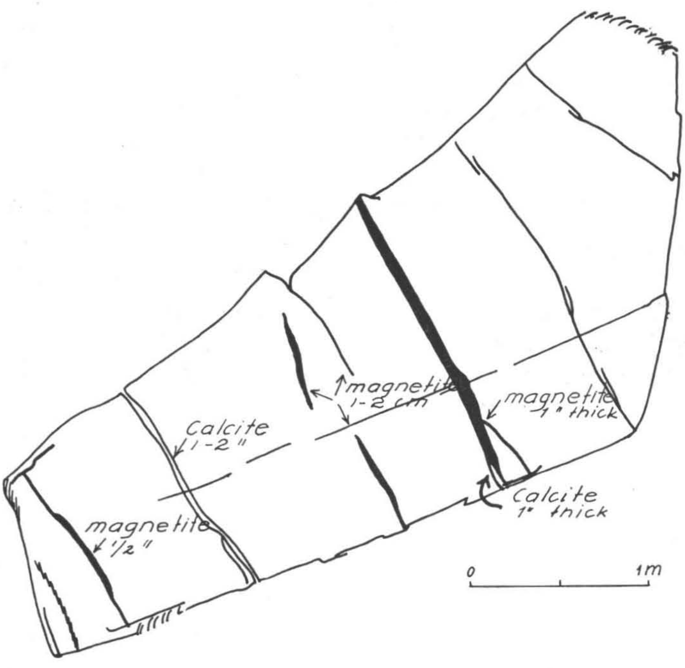

Fig. 6. Magnetite and calcite veinlets in a vertical section at point 11 of Fig. 2.

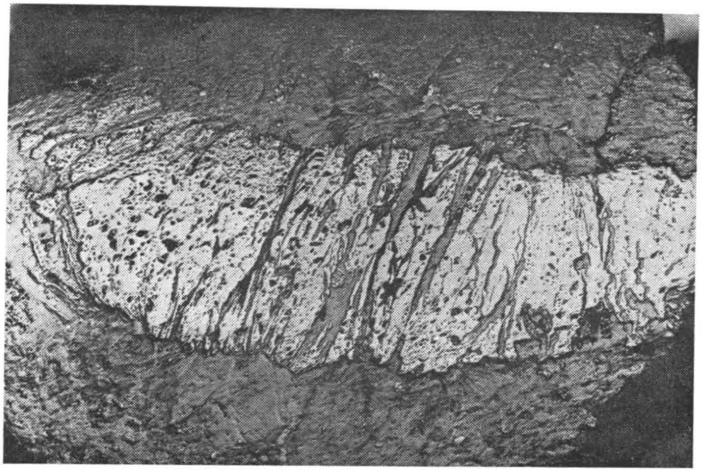

Fig. 7. Magnetite (light) veinlet in antigorite mass. Gray fibres in magnetite are asbestos. Point 11 of Fig. 2. Polished section in ordinary light. The thickness of the veinlet is 1 cm.

1. Those filling the fractures and containing well-developed octahedra of magnetite adhering to the walls of the fissures.

2. Those filled by magnetite, which forms peculiar »fibres» perpendicular to the walls of the vein, and occurring in the same fashion as do the asbestos serpentine fibres in similar veinlets. These veinlets vary from a few mm up to one inch in thickness. The fibrous-looking magnetite is mostly pure, but sometimes it displays some kind of tube-like structure with a silicate core. Magnetite of this kind may also occur alongside and intermixed with fibres composed of chlorite and serpentine.

This magnetite has been examined by Dr. O. Kouvo of the Outokumpu Co., and, using X-rays, he could not find anything anomalous in its structure. The lattice is normal, as is the chemical composition of the magnetite. Also in the polished sections the magnetite is quite normal. In the veinlets richest in serpentine, however, it may also be seen under the microscope that the minute grains of magnetite which have adhered to the asbestos fibres are octahedral. Therefore it is assumed that the magnetite there has accumulated in the veins, pre-occupied by something fibrous, obviously chrysotile, along the surfaces of which the magnetite crystals were formed.

In Fig. 7, a polished surface of such a magnetite-rich vein is seen. The magnetite contains abundant quartz inclusions and gray fibres of serpentine.

SERPETINE ACCOMPANYING MAGNETITE VEINLETS

As mentioned in the foregoing, the fibrous-looking magnetite often occurs intermixed with fibrous serpentine. In some cases there also occurs, together with magnetite a filling composed of massive, pale olive-green serpentine, which has been examined more closely by Dr. O. Kouvo and Mr. Y. Vuorelainen, at the Research Laboratory of the Outokumpu Co. They found that, despite its megascopical appearence, which resembles that of chrysotile, the structure of this serpentine is that of antigorite, characteristically marked by d 24.3.0 = 1.563 Å, and by the presence of the line of d 060 = 1.541 Å, both absent in the lattice of chrysotile or lizardite.

The experimental work of Roy and Roy (1954) shows that the difference between chrysotile and antigorite is possibly not only structural but appears also in the presence of $R^{III}{}_2O_3$ (R = preferably Al, but may be Fe^{III} as well) in antigorite. This was observed by Marmo (1958) for the antigorite serpentinites of Sierra Leone too. Thus, in regard to the antigorite of the magnetite veinlets of Pahta-autsi the presence of Al or Fe^{III} would likewise be expected.

According to microscopic examinations, the serpentine here in question was comparatively pure. Still an attempt was made to purify it by removing all the material lighter than bromoform. From the heavier fraction, the magnetite was removed using the magnetic franz. The remaining portion of serpentine was used for the spectrochemical determination of aluminium, according to which it contained 1.22 % Al. This like-wise indicated antigorite and was in good agreement with the structural determinations of this serpentine.

Furthermore, the antigorite of the magnetite veinleits was found to contain 0.007 % Cu; 0.2 % Ni; 0.02 % Co; 0.0002 % Ag; 0.4 to 0.5 % Cr; and 0.05 % Mn (all determined spectrochemically).

CHEMISTRY

Chemical analysis of the rock (Table I) revealed that the serpentinite of Pahta-autsi is a true ultrabasic rock, but unexpectedly rich in water. Moreower, compared with serpentinites examined abroad and in Finland in particular, this rock is distinctly shorter in silica than are similar rocks in general.

The analyzed sample contained but few minute and rounded grains of olivine, and still, according to the calculation of chemical analysis, it should contain as much as 42.7 % olivine. On the other hand, there appears to be a surplus of water (9 %), which excludes the possibility of there occurring much fresh olivine. As a matter of fact, the total water-content of the rock (13.55 %) is close to the amount that according to Hess, is necessary to establish the conditions where serpentine magma can exist (13 % H_2O).

Yet, it should be pointed out that, chemically, the transversion of olivine into chrysotile means an addition both of silica and water: $3\ \underset{\text{olivine}}{(2\ MgO \cdot SiO_2)} + SiO_2 + aq \rightarrow 2\ \underset{\text{serpentine}}{(3\ MgO \cdot 2\ SiO_2 \cdot aq)}$. Thus, if a rock consists, practically, of serpentine, no deficiency of silica should appear.

Table 1. Chemical Analysis of the Serpentinite of Pahta-autsi.
Analyst: Aulis Heikkinen

Weight per cent		Calculated mineral composition	
SiO_2	34.39	Olivine	42.70
TiO_2	0.01	Chrysotile	36.79
Al_2O_3	0.10	Al-serpentine	0.52
Fe_2O_3	5.46	Magnesite	1.25
FeO	2.88	Chromite	1.57
MnO	0.08	Magnetite	7.66
MgO	41.38	H_2O-surplus	9.55
CaO	0.07	NiO	0.35
Na_2O	0.21	TiO_2, Na_2O etc.	0.24
K_2O	0.00	Total	100.64
P_2O;	0.02		
CO_2	0.55		
H_2O+	13.55		
H_2O-	0.55		
Cr_2O_3	1.03		
NiO	0.35		
S	0.01		
Total	100.64		

Haapala (1936) has carried out an extensive study of the Karelian serpentinites. Among those described by him, only one is chemically in some way comparable with that of Pahta-autsi. The serpentine rock from Outokumpu (130-m level, hanging wall), which according to Haapala (op. cit. p. 34) »was completely serpentinized,» is also exceptionally rich in water, and slightly deficient in silica:

SiO_2	37.26 %	618 Mol
Al_2O_3	0.48 »	5 »
MgO	38.11 »	945 »
H_2O+	12.95 »	719 »
Cr_2O_3	0.28 »	2 »
NiO	0.24 »	3 »

Furthermore, this rock is likewise short in alumina, but contains both chromium and nickel.

The chromium-content of the serpentinite of Pahta-autsi appears to consist entirely of chromite; chrome-bearing magnetite may also be present, but so far only the magnetite of veinlets has been examined, and it does not contain chromium.

Thus the chemical analysis strongly supports the view that the serpentinite of Pahta-autsi is closely related to the actual peridotites, but it differs from them on account of its exceptionally high water-content.

The scanty occurrence of alumina in the rock — though it occurs in the antigorite extracted from magnetite veinlets — may indicate that it does not belong to the primary composition of the rock.

CONCLUSIONS

From all the geologic and petrologic evidence described in the foregoing, the following assumptions seem to be warranted:

1. The serpentinite of Pahta-autsi is of an intrusive character, and it was emplaced into the space opened by a gentle bend of the strata.

2. It has been emplaced in the form of serpentinite. The scanty grains of olivine occurring there are not relics but had crystallized contemporaneously with antigorite at the spots of silica deficiency in agreement with the experimental findings of Yoder (1952) and in accordance with the interpretation of the present author for the Sierra Leonean serpentinites (Marmo 1958).

3. The formation of fibrous magnetite veinlets is a phenomenon in which the magnetite evolved as very fine crystals which adhered to the thin serpentine fibres filling the veins.

4. There is no evidence to show the source of the material which intruded as serpentinite. This material, however, must have been comparatively widespread within the region to which the occurrence of Pahta-autsi likewise belongs. This is indicated by the ample occurrence of similar minor bodies of serpentinite not only within the gorge but also in its environment. In addition to those mentioned by Mikkola and Sahama (1936), at least two new occurrences of similar bodies have been discovered by the present writer within the area between Pahta-autsi and Huuva-autsi, which is another gorge, situated about six km to the east from the former.

REFENENCES

HAAPALA, PAAVO (1936) On serpentine rocks in northern Karelia. Bull. Comm. géol. Finlande 114.

MARMO, VLADI (1958) Serpentinites of Central Sierra Leone. Bull. Comm. géol. Finlande 180, p. 1.

MIKKOLA, ERKKI and SAHAMA, TH. G. (1936) The region to the southwest of the »Granulite Series» in Lapland, and its ultrabasics. Bull. Comm. géol. Finlande 115, p. 357.

ROY, DELLA and ROY, RUSTUM (1954) Synthesis and stability of minerals in the system $MgO—Al_2O_3—SiO_2—H_2O$. Am. Mineralogist 39, p. 147.

YODER, H. S. JR. (1952) The $MgO—Al_2O_3—SiO_2—H_2O$ system and the related metamorphic facies. Am. J. Sci. (Bowen Vol.), p. 569.

6

ANALYSES OF CERTAIN POLLENS FOUND IN VOISALMENSAARI, NEAR LAPPEENRANTA[1]

BY

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ABSTRACT:

While mapping postglacial formations in the region of Lake Saimaa, southeastern Finland, certain herbaceous pollens of a special character were found. It seems likely that these pollens may date from the Stone Age in Finland, and accordingly may be used as a criterion in archaeological studies on the distribution of Stone Age culture in Finland.

CONTENTS

INTRODUCTION

Numerous European publications dealing with forest history, have focussed attention, in drawing up pollen diagrams, on NAP flora present not only in old but also younger deposits. On the other hand, in the diagrams of Finnish researchers herbaceous pollens have in most cases been presented only from the Boreal period and still older deposits (cf. Donner 1951, 1952, Salmi 1948, 1959) inasmuch as the object of the studies has been late-glacial forest history associated with old marine or lake stages, in connection with

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which the herbaceous pollens of the post-glacial period have been considered to be of secondary importance.

In performing pollen analyses in the autumn of 1958 under the direction of Prof. V. Auer at the Department of Geology and Paleontology, Helsinki University, my attention was arrested by the fact that in the series of specimens collected from the island of Voisalmi, near the town of Lappeenranta (Fig. 1), there were present in the part representing the postglacial climatic stage — besides certain common herbaceous pollens — grains of *Plantago*, *Artemisia*, *Chenopodiaceae* and *Rumex* pollen. All of them are so-called bare mineral soil species (Fries 1951, p.p 44—45), the occurrence of pollen grains of which during the postglacial period has been observed in, e. g., Denmark, Sweden, Germany and Great Britain to be associated with neolithic settlements (cf., e. g., Iversen 1941 a and b, Troels-Smith 1942, 1943, Inge Müller 1946, Firbas 1949, Fries 1951, 1958, Florin 1958, Godwin 1956). In subsequently studying my material, I have again encountered the said pollens in only three series of specimens collected from spots situated close together on the island of Voisalmi, where pollen form clearly groups.

THE FIELD OBSERVATIONS, LABORATORY STUDIES AND CORRELATION OF THE RESULTS

In the eastern part of island of Voisalmi there is a longitudinal esker, the steep sides of which have been worn by ancient bodies of water and which exhibit numerous distinct shore features. At the eastern margin of the esker, on the side toward Lake Saimaa, there are two steep esker kettles, which in ancient times had contained esker ponds. Nowadays they are embogged, and drainage ditches have been dug in the one situated farther north. The slope of the kettles on the side of the lake is as if cut at the point of the former ponds, which are thus in a direct line of vision at one spot over the waters of Saimaa. The surface of the bogs is at present on the same level as the lake, but the ancient waterline of the ponds has left its mark about half a meter higher on the sandy slope. A short series of specimens has been collected with a

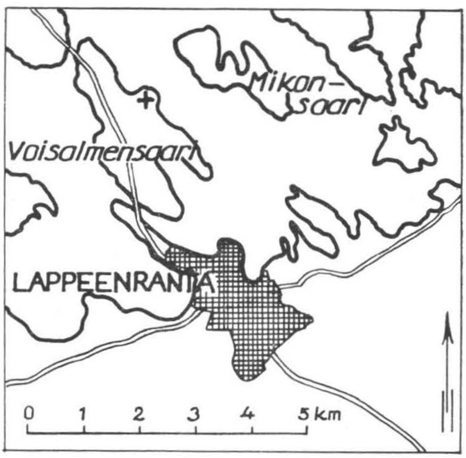

Fig. 1. The Voisalmi insel at Lappeenranta.

drill at the site on the northern esker bog where it had an outlet in to Lake Saimaa. A similar series was drilled from then eighboring bog to the south. A profile for purposes of comparison was made a year later in order to obtain a more accurate forest-historical dating.

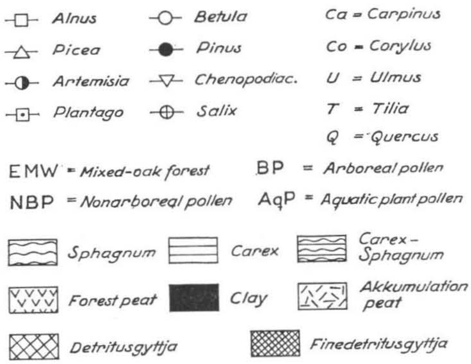

Fig. 2. Symbols used in the pollen diagrams.

The series of specimens have been studied in part at the University's Department of Geology and Paleontology and in part at the Geological Survey. The KOH method was applied in making the preparations, and the spceimens were studied at intervals of 5—6 cm. In each specimen 150—200 grains of arboreal pollen were counted.

Comparative Diagram 1 (Fig. 3) was drawn up on the basis of the series of specimens taken from the middle of the southern bog, comprising peat and ooze, in order to determine the position of these pollens, which may be designated as »cultural pollens», in the forest history. The diagram is a typically east-finnish and deviates somewhat from other published southern Finnish diagrams in that spruce pollen occurs as a low percentage but nevertheless forming a uniform, gentle curve even during the period when the rare, deciduous trees reached their maximum (cf. e. g., Hellaakoski 1936, Hyyppä 1937, 1942, 1950 and Sauramo 1941). Picea is, so far as is known, an earlier arrival in eastern than in southern Finland (Auer 1929). The quantity of rare deciduous trees is also fairly small and their common, strong habitat was of short duration. On the other hand, the lower part of the diagram represents a typical southern Finnish species (cf. e. g., Sauramo 1958, p. 44). The other zones used for the pollendiagrams in the present investigation are the same as for southern, (Sauramo 1949, 1958, Donner 1951, Mölder, Valovirta, Virkkala (1957), southeastern Finland (Donner 1952), the Kuopio district (Donner 1957) and for the Hyrynsalmi district in northeastern Finland (Kanerva 1956). This zonation has been generally accepted in other parts of Europe (Jessen 1935, Firbas 1949, Godwin 1943, Fries 1951). The high *Betula* maximum thus represents the Pre-Boreal period, which ends at the beginning of the Boreal period, when the pine became the dominant species of tree. At the same time *Corylus* appears to be on the rise. The rapid rise of *Alnus* may be regarded as the start of the Atlantic period and the period of the more abundant occurrence of rare deciduous trees appears to correspond to the VII forest-historical period (Sauramo 1949, Donner 1952, 1957). The rational border of *Tilia*

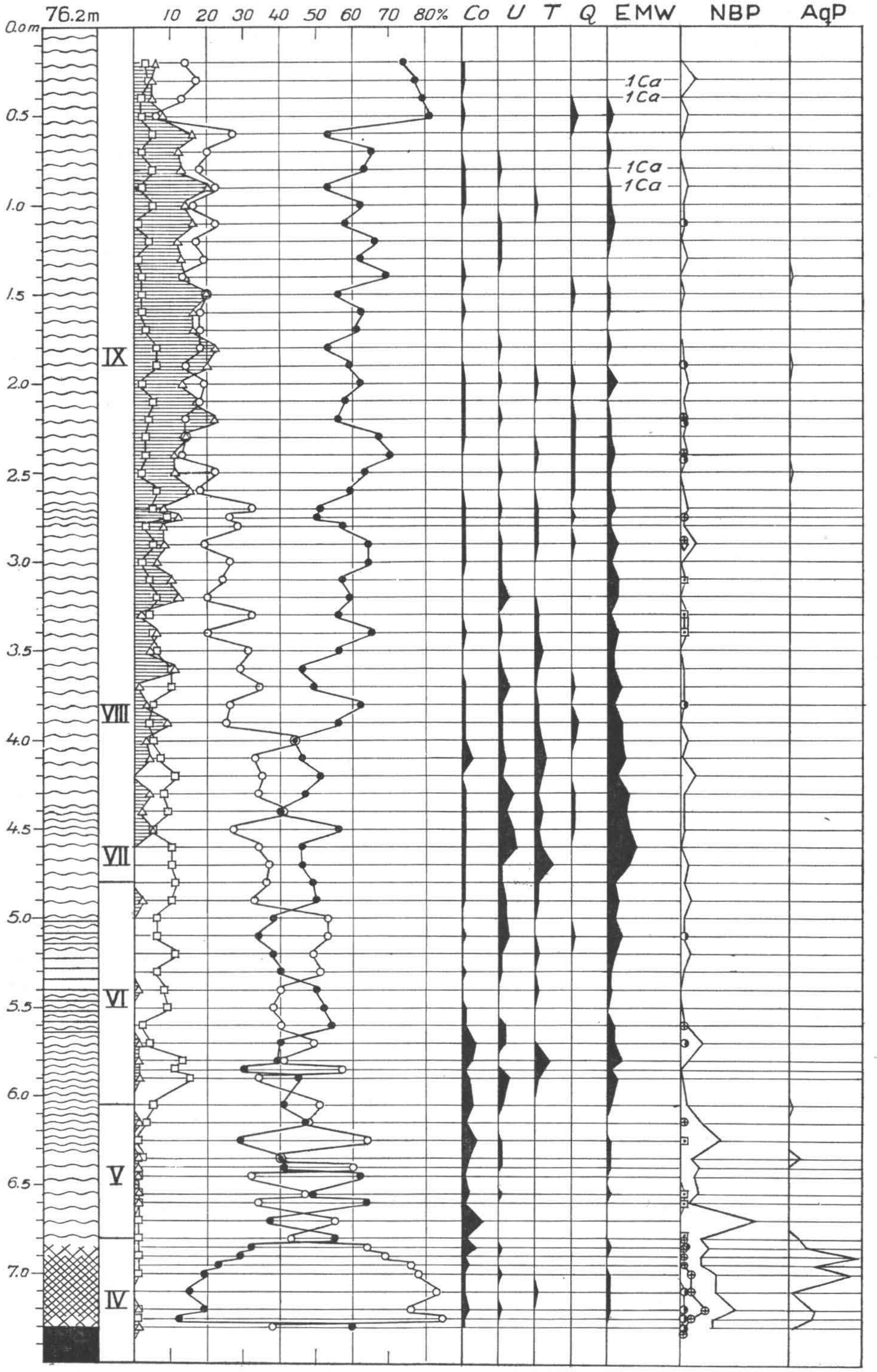

Fig. 3. Pollen diagram 1.

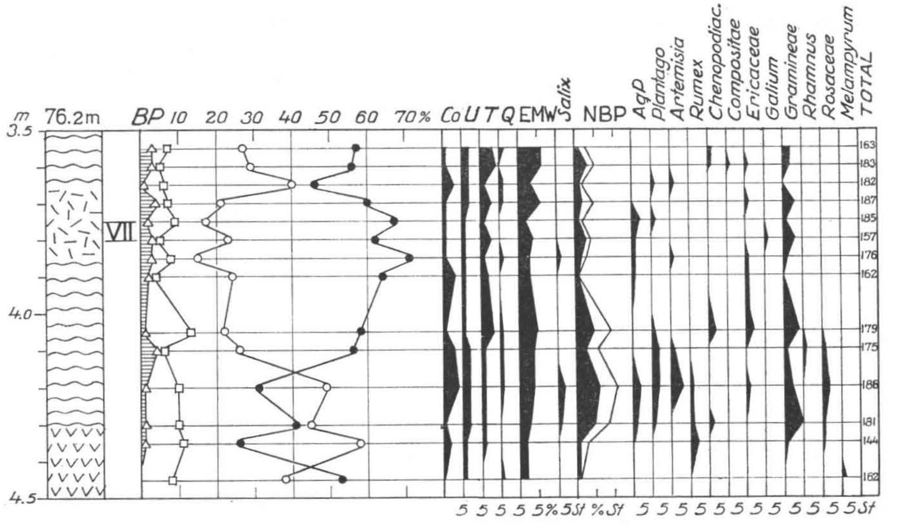

Fig. 4. Pollen diagram 2.

marks the beginning of zone VII. The deterioration in climatic conditions seems to have taken place gradually, and the strong growth of bog peat indicates a cold and damp climate. No clear boundary between the two latter zones had demonstrated on the area investigated.

In Profile 2 (Fig. 4) the common curve of the herbaceous pollens rises somewhat from its low Boreal-stage position. In addition to the ordinary

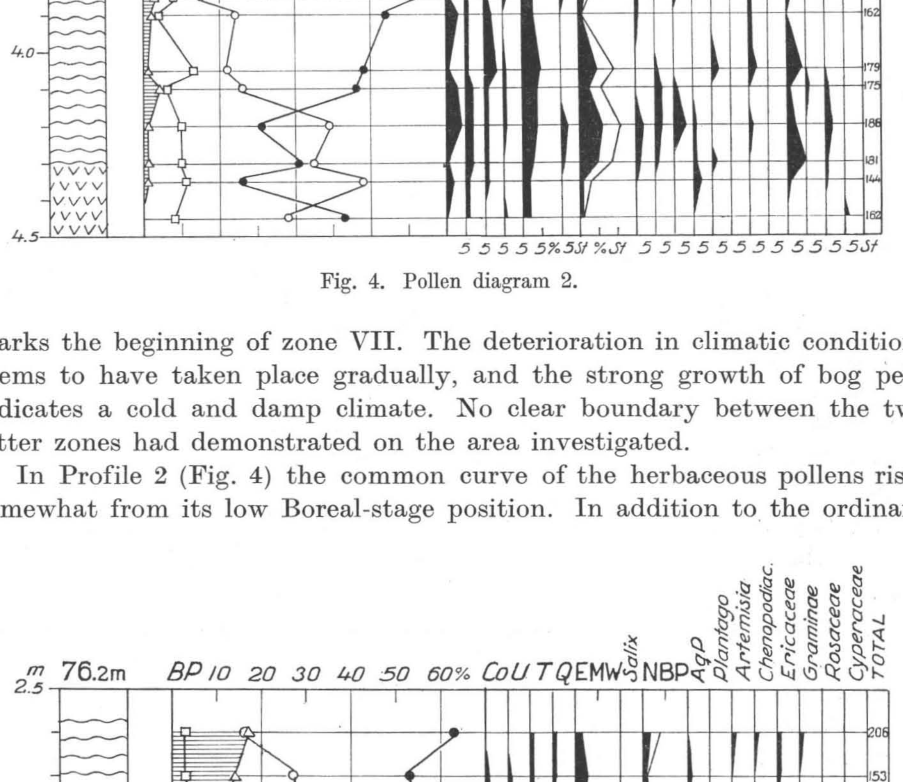

Fig. 5. Pollen diagram 3.

herbaceous pollens, there occur in it the aforementioned cultural pollens during the period of the abundant occurrence of rare deciduous trees. These pollen grains are present also in Diagram 3 (Fig. 5) — and, judging from the abundance of Picea, in a notably younger layer. The *Plantago* occurring in both diagrams has been identified as *P. major*.

In comparing Profiles 2 and 3 with the diagram representing the forest history of the locality dealt with, one will clearly observe them to be of different ages. Diagram 2 falls into a warm period and the first occurrences of *Plantago major* in the middle of the best warm period. The rest of the pollens regarded as »cultural» are grouped likewise. On the other hand, profile 3 falls after the central warm period proper in the main diagram, belonging apparently to the Sub-Boreal period or perhaps into the beginn of Sub-Atlantic period. Up to the present, however, it has been difficult to distinguish the various post-glacial zones clearly, since the forest history of the region has not yet been studied on a sufficient scale.

The first occurrence of cultural pollens takes place, according to Diagram 2, during the best warm period. This corresponds in our chronology to a period dating back about five thousand years. Archeologically it coincides with the Comb-Ceramic stages of the Neolithic Age (Luho 1948).

On the basis of finds in this country, we know that before the decisive spread of the spruce, settlement had extended near the area under discussion. Finds identified as dating back to the Stone Age point to a widespread settlement of the shores along the southern parts of Lake Saimaa which are closely associated with the Suursaimaa stage studied by Hellaakoski (1922) in that the Stone Age settlements were concentrated on the level of the transgressive shore of ancient Lake Saimaa or slightly below it (Meinander 1947). The closest Stone Age find was made on the sandy beach in the center of the island of Mikonsaari, situated roughly one kilometer to the northeast from the locality of the present investigation. It was here that Ailio in 1931 found Stone Age relicts at an elevation of 80.5 meters, or, in other words, 4.5 m above the present level of Lake Saimaa. Among the relicts were fragments of pottery which, dated archeologically, belong to stylistic stage II. 2. Moreover, farther to the east, at a distance of five or six kilometers from the island of Voisalmi, near the mouth of the Saimaa canal, Pälsi found in 1938 at a level 1.3 meters below the Suursaimaa stage designated by Hellaakoski ceramic objects which in respect to their ornamentation have also been identified as representing stylistic stage II: 2. In addition, far to the north, pottery representing in style the same stage has likewise been unearthed at Vaateranta on the shore of Taipalsaari. Worthy of mention, too, are the Stone Age finds made at Kärenlampi, in the nearby village of Rutola, along the Kivijärvi watercourse, which is assumed to have been the discharge channel of ancient Lake Saimaa.

Before the river Vuoksi evolved, the island of Voisalmi was divided into two parts separated by a shallow channel at the time the level of Lake Saimaa was at its maximum height. The eastern esker formation made up one part, and the esker kettles on its eastern side were ponds the waterlevel of which was determined by the level of the surrounding lake. It is conceivable that during certain recurrent floods the ponds were connected directly by water with the lake. In this part of the island bordered by the esker there is still an abundance of game birds, and its highest point towers above all the other islands in the vicinity. Thus it must have served as an admirable lookout point for Stone Age men, and on account of the esker kettles, again, it offered an incomparable shelter against both enemies and the wind. Situated close by a great body of water, the kettles, with their ponds, provided a protected haven for the boats of Stone Age men, and the locality offered splendid opportunities for fishing, either in the ponds or in great Lake Saimaa, which in those times must surely have abounded with fish.

CONCLUSIONS

In the light of the foregoing, it may be assumed with full reason that the aforementioned grains of the pollen of plants characteristic of bare mineral soil, are signs left on the vegetation of that limited area by Neolithic man. During the height of the best warm period, man took advantage of the kettles protected by high gravel slopes. The foreshadowing occurrence of *Rumex* pollen in Diagram 2 may be taken to indicate that people have dwelt in the near vicinity. The occurrence slightly higher up of cultural pollen in richer variety, comprising *Plantago*, *Artemisia* and *Chenopodiaceae*, would suggest that the locality was visited more frequently or settled more definitely at a later date, slightly after the optimum of the warm period. In Diagram 3 *Plantago* no longer constitutes a uniform curve but occurs sporadically. On the other hand, the curve representing *Artemisia* is continuous, breaking off slightly above more humified layer occurring in the bog peat. Presumably during this period, apparently belonging to the Sub-Boreal period, the locality no longer commanded the interest of man. Settlement shifted elsewhere, for during the Sub-Atlantic period these pollens did not appear in the bog deposits. The kettles in the esker were for a long time forgotten by man, the reason largely being that, on account of the new outlet provided by the river Vuoksi, the waterlevel of Lake Saimaa sank considerably and the easy connection between the ponds and the lake was lost. The filling in of the ponds became more rapid, too, in these times, with the growth of bog peat gaining strength.

Acknowledgements — In conclusion I wish to thank the Geological Society of Finland and the Geological Survey of Finland for giving me the possibility to publish this study. I should like to thank Prof. V. Auer Ph. D. at whose suggestion this study was started and Doc. Esa Hyyppä Ph. D. for his valuable advice also. I wish to express my special thanks to J. Donner Ph. D., and R. Tynni M. A. for helping to control my pollen determinations. The English translation was made by Paul Sjöblom M. A.

REFERENCES

AUER, V. (1929) Über die Einwanderung der Fichte in Finnland. Comm. Inst. Quaest. Forest. Finl., ed. 13.

DONNER, J. (1951) Pollen-analytical studies of Late-Glacial deposits in Finland. C. R. Soc. géol. Finlande 24; Bull. Comm. géol. Finlande 154.

—»— (1952) On the early Post-Glacial shore-line displacement in southeast Finland. Ann. Acad. Scient. Fenn., Ser. A. Tom. III. 29.

—»— (1957) The Post-Glacial shore-line displacement in the Kuopio district. Ann. Acad. Scient. Fenn. Ser. A. Tom. III. 50.

FIRBAS, F. (1949) Spät- und nacheiszeitliche Waldgeschichte Mitteleuropas nördlich der Alpen. Erster Band. Allgemeine Waldgeschichte. Jena.

FLORIN, MAJ-BRITT (1958) Pollen analytical evidence of prehistoric agriculture at Mogetorp Neolithic settlement. In Sten Florin: Vråkulturen. Stockholm 1958.

FRIES, M. (1951) Pollenanalytiska vittnesbörd om senkvartär vegetationsutveckling särskilt skogshistoria i nordvästra Götaland. (Deutsch. Abzug: Pollenanalytische Zeugnisse der spätquartären Vegetationsentwicklung, hauptsächlich der Waldgeschichte, im nordwestlichen Götaland, Südschweden). Acta Phytogeogr. Suec. 29. Uppsala.

—»— (1958) Vegetationsutveckling och odlingshistoria i Varnhemstrakten. (Deutsch. Abzug: Vegetationsentwicklung und Siedlungsgeschichte im Gebiet von Varnhem). Acta Phytogeogr. Suec. 39. Uppsala.

GODWIN, H. (1943) Coastal peat beds of the British Isles and North Sea. J. Ecol. 31.

HELLAAKOSKI, A. (1922) Suursaimaa. (Deutsches Referat). Fennia 43, 4.

—»— (1936) Das Alter des Vuoksi. Bull. Comm. géol. Finlande 115.

HYYPPÄ, E. (1937) Post-Glacial changes of shore-line in South Finland. Bull. Comm. géol. Finlande 120.

—»— (1942) Beiträge zur Kenntnis der Ladoga- und Ancylustransgression. C. R. Soc. géol. Finlande 15; Bull. Comm. géol. Finlande 128.

—»— (1950) Lapuan ja Pattijoen muinaissuksien geologinen ikäys. Suomen Museo LVII.

IVERSEN, J. (1941 a) Landnam i Danmarks Stenalder. En pollenanalytisk Undersøgelse over det første Landbrugs Indvirkning paa Vegetationsudviklingen. Land Occupation in Denmark's Stone Age. A pollenanalytical study of the influence of farmer culture on the vegetational development. Danm. Geol. Unders. II. R. II. R. Nr. 66. København.

—»— (1941 b) The influence of prehistoric man on vegetation Danm. Geol. Unders. IV. R. 3. Nr. 6. København.

JESSEN, K. (1935) Archaelogical dating in the history of North Jutland's vegetation. Acta archaeol. 5. København.

KANERVA, R. (1956) Pollenanalytische Studien über die Spätquartäre Wald- und klimageschichte von Hyrynsalmi on NO-Finnland. Ann. Acad. Scient. Fenn., Ser. A. Tom. III. 46.

LUHO, V. (1948) Suomen kivikauden pääpiirteet. Helsinki.
MEINANDER, C. F. (1947—1948) Vehmersalmen Roikanmäen kivikautinen asuinpaikka. Suomen Museo LIV.
MÜLLER, INGE (1946) Der pollenanalytische Nachweis der menschlichen Besiedlung im Federsee und Bodenseegebiet. Planta 35.
MÖLDER, K., VALOVIRTA, V., VIRKKALA, K. (1957) Über Spätglazialzeit und frühe Postglazialzeit in Südfinnland. Bull. Comm. géol. Finlande 178.
SALMI, M. (1948) Die Ancylustransgression in den Moore Hangassuo im Süd-Finnland. Bull. Comm. géol. Finlande 142.
—»— (1959) Konnunsuo. Turvegeologinen tutkimus. Konnunsuo bog. A peatgeological investigation. Suo 3.
SAURAMO, M. (1941) Die Geschichte der Wälder Finnlands. Geol. Rundschau 32, 4/5.
—»— (1949) Das dritte Scharnier der fennoskandischen Landebung. Soc. Scient. Fennica. XXVII. B. 4.
—»— (1958) Die Geschichte der Ostsee. Ann. Acad. Scient. Fenn., Ser. A. Tom. III. 51.
TROELS-SMITH, J. (1942) Geologisk Datering af Dyrholms-Fundet. Kgl. Danske Vidensk. Selsk. Arkaeol.-Kunsthist. Skr. I.
—»— (1943) Geologiske Dateringer af bopladser i Aamosen. Nordiske Forntidsminder 3 : 3, 147.

7

A STATISTICAL STUDY OF STONE ORIENTATION IN GLACIAL TILL [1]

BY

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ABSTRACT

The long axes of elongated stones in glacial till are more or less distinctly oriented in a certain plane, and on this plane in a certain direction, generally parallel to the movement of glacial ice. This phenomenon has been widely used in studying many geological problems. The basis of the phenomenon has also been studied hydrodynamically in the laboratory. The purpose of this paper is to show how such a phenomenon may be studied by simple statistical methods.

INTRODUCTION

In 1884, H. Miller observed that the long axes of stones in glacial till were arranged parallel to the direction of ice movement. K. Richter in 1932 was the first to make use of this phenomenon in studying the directions of ice movements. Subsequently, stone-orientation analysis has been used succesfully by many workers (Holmes, 1941; Lundqvist, 1948; Seifert, 1954; Glen, Donner and West, 1957, for example). In this paper the terminology employed by Glen, Donner and West is used.

Jeffery (1921), when studying particle motion in a viscous fluid in laminar motion, concluded that »particles will tend to adopt that motion which corresponds to the least dissipation of energy». This theory has been tested and applied by geologists and it may be especially valid for motion in the shear plane, for example in a vertical section of till parallel to the ice movement. A thick ice mass may be considered to behave as a fluid (mixture of ice and stony material) under laminar motion (Demorest, 1942).

[1] Received March 4, 1960.

The parallel orientation of stones appears to be caused predominantly by stone collisions and by the scraping of stones along bedrock surfaces or between moving-ice layers. A previously deposited layer of glacial till may be disturbed by new ice movements and the earlier orientation partly destroyed, resulting in a differrent orientation. During the melting stage the till is thoroughly wet and flows easily if deposited on inclined ground. Solifluxion and the pressure of dead ice masses also cause flow. The orientation may be destroyed by slumping of dead ice during the melting stage, by frost action and also by tree roots. The final orientation is thus a function of several factors.

Despite the lack of knowledge of the fundamental causes of orientation, orientation analysis has frequently been used in the solution of many kinds of problems, for example prospecting (Hyyppä, 1948), directions of ice movements (Okko, 1949; Virkkala, 1951) and till stratigraphy (Virkkala, 1951; Järnefors, 1952; West and Donner, 1956).

MATHEMATICAL TREATMENT

In an orientation distribution the maxima are the most distinctive features, but these are only a part of the orientation, although the most important for the practical geologist. It would be possible to study only the maximum, its situation, direction or its height and to compare it with that of anorther orientation. It would be possible to investigate, for example, the central positions (arithmetic mean, median, mode), variability (standard deviation, range) and skewness, etc., but any one taken alone incompletely describes the orientation distribution.

For the of comparing the distributions statisticians have developed methods of which the chi-square test is widely used. This» χ^2-test is accomplished by comparing the observed data expressed as frequencies in various categories or groups with theoretical or expected results in the same categories or groups. The value of the chi-square is computed, based on the difference between the observed and theoretical frequencies as follows $\chi^2 = \sum \left[\frac{(O-E)^2}{E}\right]$ » (Arkin and Colton, 1958).

In the foregoing formula: O = observed value, E = expected value. In the following text, degrees of freedom (df) and probability (p) are given for every χ^2 calculated. In this paper a significance level $p < .05$ (5 %) is used. Probability values less than .05 indicate that the observed and expected distributions are not samples from the same population. When in the following several distributions are compared, the arithmetic mean is taken as expected value and deviations of values of the samples from those of the mean computed.

Chi-square may be defined as »an index of dispersion, devised to measure the deviation of the sample from the hypothetical population distribution, provided with tables of its probable occurrence in sampling» (Snedecor, 1957) and »the coefficient or criterion of comparison (χ^2) is a kind of composite weighed variance» (Aitken, 1952).

The value of the computed chi-square may be interpreted by reference to the tables of χ^2 (Arkin and Colton, for example), where the probability of the chi-square value obtained is shown as a function of the degrees of freedom. This value indicates the probability that the difference between the actual and the theoretical values for the stated chi-square could have arisen by chance or by sampling variation. A low probability value indicates that the differences are not accidental or do not arise through sampling variation; a large probability value indicates that the differences could have arisen by chance or through sampling variation. To use the χ^2-test the frequency of every cell must be at least five (according to some investigators 10). The χ^2-test is succesfully used in orientation studies by Harrison (1957). The interested reader is referred to Chayes' (1949) very illustrative description of this statistical method and its application in the solution of geological problems.

RELIABILITY OF ORIENTATION ANALYSIS

Measurement is commonly made by compass and the direction of the longest axis is read to the nearest 5-degree interval. The measurements may be affected by such magnetic errors as variable declination, local magnetic anomalies, magnetic boulders in the pit or magnetic tools used in the pit. These errors may be evaluated. Frequently it is not easy to determine which axis has had the greatest effect in determining the orientation (Fig. 1, C). The measurement is thus somewhat subjective. The measurement should not be made before the whole stone is seen, because the longest axis of the visible part may not be the true long axis of the stone (Fig. 1, A and B). Sometimes when the stone is displaced the hole is partly filled with fine sandy material, and it may be impossible to determine the true direction of the long axis. Small stones are sometimes aligned along the surfaces of larger stones and it may be impos-

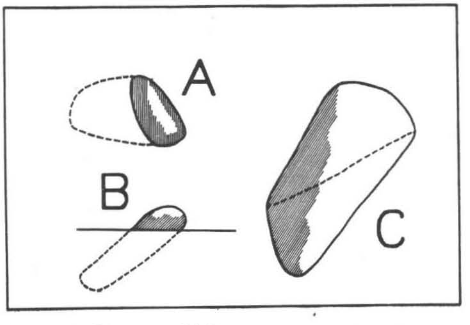

Fig. 1. Stones which may cause erroneous measurements

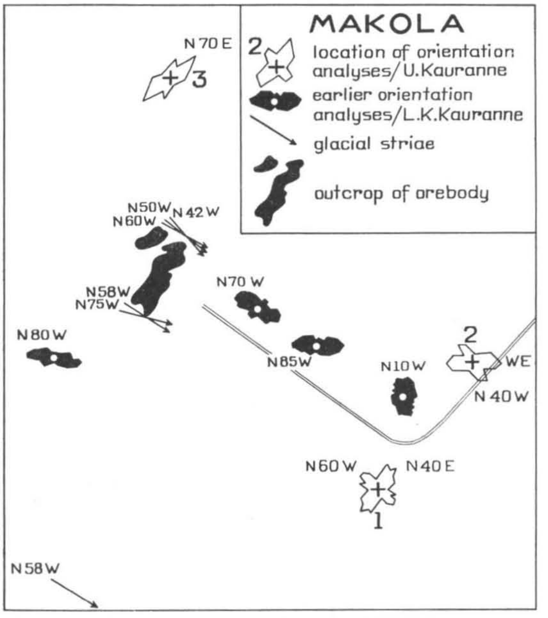

Fig. 2. Location of orientation analyses in Makola, Central Finland

sible to know whether their orientations should be measured or not.

Previously some factors which cause or destroy orientation have been mentioned. The orientation is a function of several variables which do not equally effect the whole layer of glacial till. Results obtained in the same area or in the same pit may not always be comparable. The till fabric itself may cause difficulties.

When evaluating the results of orientation analyses it should be noted that the intensity and direction of the maxima depend both on the reliability of measurement and consistency of the till fabric. Sampling techniques may make independent investigation of these factors difficult.

An attempt was made to assess the effect of the operator, but owing to the inconsistency of the orientation in that particular area the results were inconclusive. It would be possible to make this test in some other area where the orientation is consistent.

Table 1. Stone orientation consistency in glacial till

Pit	Number of Measurements	χ^2	df	p <
	A B			
1	100 + 100	12.8	13	.50
2	100 + 100	5.5	12	.95
3	100 + 100	8.6	12	.80

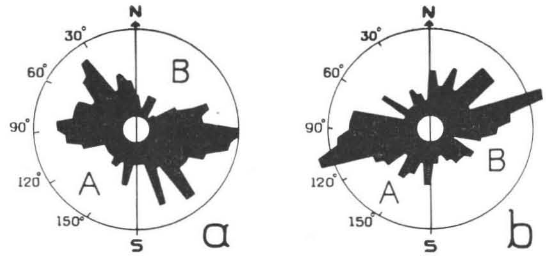

Fig. 3. Orientation analyses: diagram a, pit 1, groups A and B, $p < .50$; diagram b, pit 3, groups A and B, $p < .95$

The reliability of measurement owing to c o n s i s t e n c y was easier to study. U. Kauranne investigated three pits in Makola (Fig. 2) and collected the results in two groups as follows: the first measurement in group A, the second measurement in group B, the third again in group A and so on. The results of this investigation are presented in Table 1 and Fig. 3.

THE NATURE OF ORIENTATION

In the Kotanen area a larger study of the effects of location was made. From a trench three meters deep 2 000 stone directions were measured. The position of these measurements, made in groups of 100 (4×25), and spaced at 100 cm horizontal and 50 cm vertical intervals, is shown in Fig. 4.

The results were grouped first into four levels and the following results calculated ($\chi^2 = 110.7$ df $= 105$ p $< .35$). All the levels could be considered as samples from the same orientation distribution. When the orientations were studied closer it was seen that the uppermost level had a much greater influence on the χ^2-value than did the three others. Therefore the uppermost level was compared with the sum of the others and the obtained values ($\chi^2 = 57.2$ df $= 35$ p $< .025$) indicate that the orientation of this level differs distinctly from that of the three others. The boundary between the lower and upper beds is situated between 1.5 and 2 meters. The stone orientation is thus not constant in vertical section (Fig. 5).

When the orientation of 600 stones from the N-wall of the trench in the lower bed of till in Kotanen were compared with that of 600 stones from the S-wall, the calculation ($\chi^2 = 23.3$ df $= 35$ p $< .95$) showed that both groups were from the same distribution. When the horizontal interval between measurements does not exceed several meters the orientation may be considered constant.

If measurements are made on the same field but spaced some tens or hundreds of meters apart, the orientation varies. When the results from

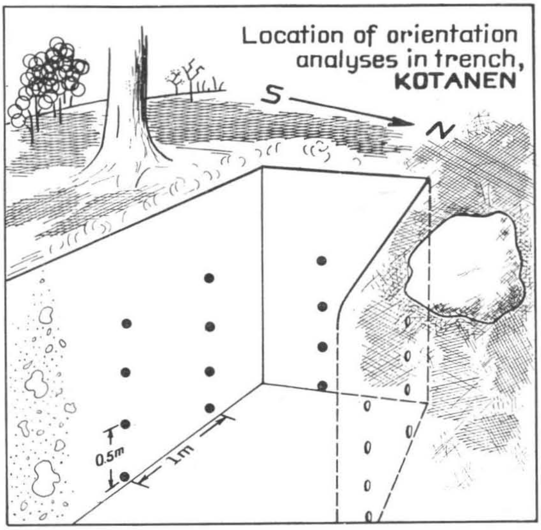

Fig. 4. Location of orientation analyses in trench, Kotanen, Central Finland

the three pits in Makola were compared with each other, values ($\chi^2 = 89.3$ df $= 34$ p $< .0005$) were obtained. The orientation of the stones in these pits must thus be different (Fig. 2).

Further studies on the effect of size on orientation were not made. Some comparison experiments made at Kotanen may, however, be mentioned. One hundred orientations of boulders with a long axis of more than one meter in length were measured and the results compared with the orientation of the stones in the upper bed of till. Calculation ($\chi^2 = 16.2$ df $= 13$ p $< .30$) indicates that the two groups may be considered to be from the same distribution.

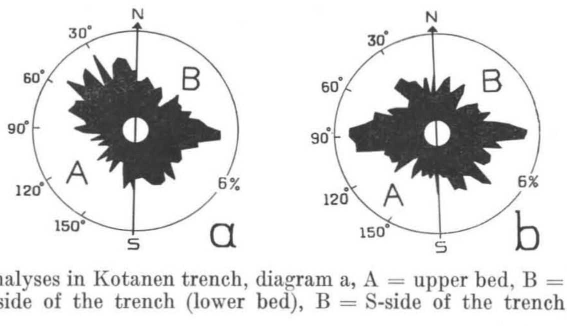

Fig. 5. Orientation analyses in Kotanen trench, diagram a, A = upper bed, B = lower bed; diagram b, A = N-side of the trench (lower bed), B = S-side of the trench (lower bed)

THE SHAPE OF ORIENTATION DISTRIBUTION

If the stones in glacial till were randomly oriented the orientation-frequency curve should be a line parallel to the base (Fig. 6). When such a random orientation was compared with a natural orientation in glacial till (Makola pit 3) the result ($\chi^2 = 91.7$ df $= 17$ p $< .0005$) indicates, as does the picture below (Fig. 6), that there is a preferred orientation in the till.

An attempt to produce experimentally a random orientation was made by drying about 30 kg of glacial till, which was then poured to the ground from a height of 50 cm. The orientation of the stones was then studied by measuring the long-axis directions of 300 stones. The resultant distribution was compared with the random distribution and the calculations ($\chi^2 = 16.7$ df $= 35$ p $< .99$) show that the experimental distribution is random.

Some comparisons between an orientation formed by a fluid mechanism and the orientation of stones in glacial till were also made. According to Jeffery (1922) the angular velocity of ellipsoidal particles in a viscous fluid under laminar motion is directly proportional to the angle which the long

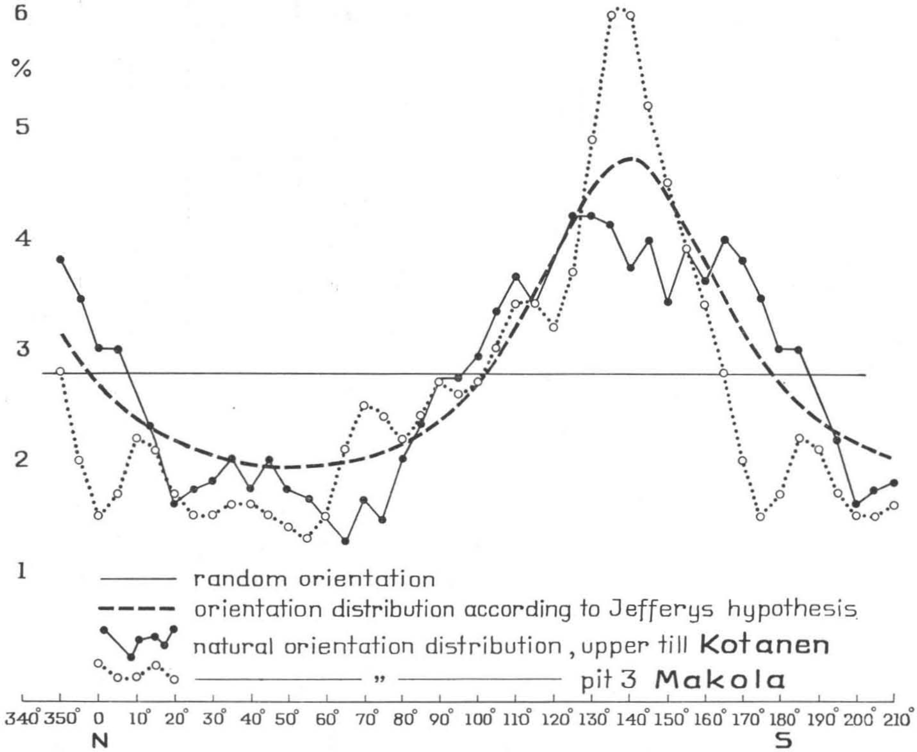

Fig. 6. Orientation of ellipsoids in a viscous fluid under laminar motion (according to Jeffery) and random orientation compared to preferred orientations of stones in glacial till.

axis of the ellipsoid makes with the parallel plane. The rotation velocity is thus at a minimum when the long axis lies in the parallel plane; but it is never zero because the shorter axis, which is then in the shear plane, also effects the rotation. Fig. 6 shows a comparison of such an orientation (calculated to an axial ratio of 2 : 1) with natural orientations.

When the orientation distributions were compared, using the chi-square test, and taking the theoretical distribution as expected, the following results were obtained: Kotanen/Jeffery ($\chi^2 = 29.4$ df $= 17$ p $< .05$) and Makola/Jeffery ($\chi^2 = 15.6$ df $= 16$ p $< .50$). The Kotanen/Jeffery difference is significant at the 5 % level. The latter comparison (Makola/Jeffery), which is shown also in Table 2, shows that the stone orientation at Makola pit 1 may be considered to follow the fluid mechanical orientation laws.

Table 2. Comparison of a natural orientation distribution (Makola pit 1) with that based on Jeffery's hypotesis

Class	O	E	O'	E'	Δ	Δ^2	Δ^2/E'
181—190	12	6.0	12	6.0	6.0	36.00	6.00
191—200	7	6.8	7	6.8	0.2	0.04	—
201—210	12	8.0	12	8.0	4.0	16.00	2.00
211—220	10	10.0	10	10.0	—	—	—
221—230	12	12.8	12	12.8	0.8	0.64	0.05
231—240	15	16.6	15	16.6	1.6	2.56	0.15
241—250	11	20.4	11	20.4	9.4	88.5	4.3
251—260	28	22.2	28	22.2	5.8	33.6	1.5
261—270	20	20.4	20	20.4	0.4	0.16	0.01
271—280	16	16.6	16	16.6	0.6	0.36	0.02
281—290	14	12.8	14	12.8	1.2	1.44	0.11
291—300	3	10.0					
301—310	11	8.0	14	18.0	4.0	16.00	0.9
311—320	6	6.8	6	6.8	0.8	0.64	0.09
321—330	6	6.0	6	6.0	—	—	—
331—340	7	5.6	7	5.6	1.4	1.96	0.35
341—350	5	5.6	5	5.6	0.6	0.36	0.06
351—360	5	5.6	5	5.6	0.6	0.36	0.06
	200						15.6

In the foregoing table, one cell frequency is only three and therefore two cells must be combined and a new classification made. E represents the original measurement, E' the new values after the new classification, O = the distribution according to Jeffery and O' = the same after the new grouping. Δ = O' —E'.

Nearly all the orientation distributions of glacial tills have several maxima and these partly overlap, thus producing skewed peaks in the diagram. The cause of these additional maxima is not known. West and Donner (1956) considered them to have resulted from small changes in the ice movement. The maxima that are not right angles to the largest maximum may thus be caused by some unknown factors. That maximum which lies at right angles to the large maximum may be caused by the rolling of stones, which is often obvious between layers of till (Seifert, 1954). Seifert has treated the results of orientation analyses by first calculating a sliding average, thus smoothing the diagram, and then splitting the diagram into two symmetrical maxima. The justification for this treatment is questionable but it may help in the solution of problems of ice movements in an area where the general directions are known.

THE ACCURACY OF ORIENTATION ANALYSIS

The accuracy of orientation analysis is of interest in practical work. For a study of accuracy, as a function of the number of measurements, the results collected in the Kotanen area were used. The measurements there were made in groups of 25 and by addition new groups of 50, 100, 200 and 500 were formed. From these the situation and distribution of the maxima were observed. The arithmetic mean and standard deviation were calculated according to following formula (Vahervuo, 1946):

$$\delta = \sqrt{\frac{\Sigma n(x—K)}{N}}$$

n = cell frequequency
N = number of observations (Σ n)
K = arithmetic mean
x = values of sample (0—180; 5° classes)

Table 3. Accuracy of orientation analysis in the Kotanen area

1	2	3	4
25	170°	269°	36°
50	155°	281°	33°
100	105°	275°	22°
200	60°	271°	12°
500	10°	270°	1°

1 number of measurements
2 range of distribution
3 arithmetic mean of values
4 standard deviation

The distribution of maxima may be considered to be normal and thus 66 % of the occurrences will deviate from the mean by less than the standard deviation and only 5 % will deviate by more than twice the standard deviation. The example from the Kotanen area indicates that the measurement of not less than 300—400 long axes would give a result which at the 5 % level would keep within class limits (5°). A qualitative picture of the orientation may be obtained with a smaller number of measurements. This is indicated in the following diagram (Fig. 7), in which the first

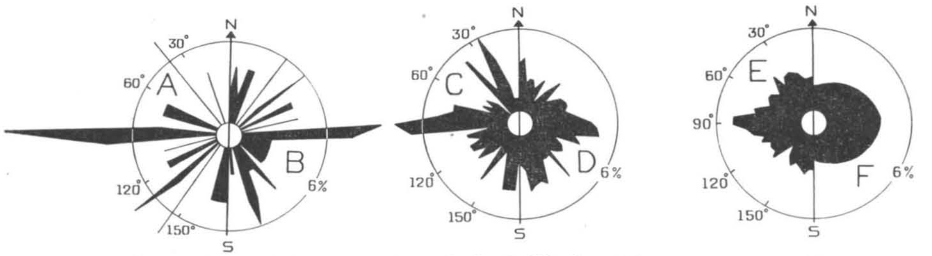

Fig. 7. Orientation analysis at Kotanen (lower bed of till). A = 25 measurements, B = 50, C = 100, D = 500, E = 1 500, F = orientation distribution according to Jeffery's hypothesis.

groups of 25, 50, 100, 500, 1 000, 1 500 orientation measurements of stones in the lower bed of glacial till at Kotanen, with the theoretical orientation accoding to Jeffery's hypothesis, are presented.

SUMMARY

The long axes of elongated stones in glacial till have obtained a preferred orientation. Most stones lie approximately in a parallel plane (Glen, Donner and West, 1957) and within this plane the stones may be oriented parallel to the ice movement or transverse to it. Some may be concentrated in other directions also. In this paper the preferred orientations were studied by common statistical methods, and some new relationships and characteristical features found. The χ^2-test may be succesfully used in evaluating results obtained by direction measurements and may therefore become a valuable tool in the solution of scientific and practical problems.

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REFERENCES

ARKIN, H. and COLTON, R. R. (1958) Tables for statisticians Barnes & Noble, Inc. New York.

AITKEN, A. C. (1957) Statistical mathematics Oliver and Boyd LTD. Edinburg.

CHAYES, F. (1949) Statistical analysis of two-dimensional fabric diagrams. In H. W. Fairbairn: Structural petrology of deformed rocks, p. 297—307. Addison-Wesley, Cambridge Mass.

DEMOREST, M. (1942) Glacier thinning during deglaciation. Part I Glacier regimens and ice movements within glaciers. Am. J. Sci. 240, p. 30.

DONNER, J. J. and WEST, R. G. (1957) The Quaternary geology of Brageneset, Nordaustlandet, Spitsbergen. Norsk Polarinstitut Skrifter 109.

GLEN, J. W., DONNER, J. J. and WEST, R. G. (1957) On the mechanism by which stones in till become oriented. Am. J. Sci. 255, p. 194.

HARRISON, P. W. (1957) New technique for three-dimensional fabric analysis of till and englacial debris containing particles from 3 to 40 mm. in size. J. Geol. 65, 1, p. 98.

HOLMES, C. D. (1941) Till fabric. Bull. Geol. Soc. America 52, 9, p. 1 299.

HYYPPÄ, E. (1948) Tracing the source of the pyrite stones from Vihanti on the basis of glacial geology. C. R. Soc. Géol. Finlande XXI. Bull. Comm. géol. Finlande 142, p. 97.

JEFFERY, G. B. (1922) The motion of ellipsoidal particles immersed in a viscous fluid. Royal Soc. London Proc., Ser. A 102, 715, p. 161.

JÄRNEFORS, B. (1952) A sedimentpetrographic study of glacial till. Geol. Fören. i Stockholm Förh. 74, 2, p. 185.

LUNDQVIST, G. (1948) Blockens orientering i olika jordarter. Sver. Geol. Undersökning Ser. C 497.

OKKO, V. (1949) Explanation to the map of surficial deposits. Sheet B 4. Kokkola. Geologinen tutkimuslaitos. Helsinki.

SAURAMO, M. (1924) Maalajikartan selitys. Lehti C 4. Kajaani. Geologinen tutkimuslaitos. Helsinki.

SEIFERT, G. (1954) Das mikroskopische Korngefüge des Geschiebemergels als Abbild der Eisbewegungen, zugleich Geschichte des Eisabbaues in Fehmarn, Ost-Wagrien und dem Dänisschen Wohld. Meyniana Bd 2, p. 124.

SNEDECOR, G. W. (1957) Statistical Methods. 5: th ed. Ames, Iowa, U.S.A.

VAHERVUO, T. (1946) Johdatus matemaattis-tilastollisiin menettelytapoihin. Porvoo.

WEST, R. G. and DONNER, J. J. (1956) The glaciation of East Anglia and the East Midlands. A differentiation based on stone orientation measurements of the tills. Quart. J. Geol. Soc. London 122, p. 69.

VIRKKALA, K. (1951) Glacial geology of the Suomussalmi area East Finland. Bull. Comm. géol. Finlande 155.

8

ON HIGH-TEMPERATURE ALKALI FELDSPARS OF SOME VOLCANIC ROCKS OF KENYA AND NORTHERN TANGANYIKA [1]

BY

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ABSTRACT

The chemical composition, unit-cell dimensions and optical properties of six high-temperature alkali feldspars from Kenya and Northern Tanganyika are given. The results are compared with previous data.

CONTENTS

INTRODUCTION

Among the material collected by the Finnish East African Expedition (Prof. Th. G. Sahama, Dr. K. J. Neuvonen and Dr. Kai Hytönen) in 1952 from Kenya and Northern Tanganyika, there were a score of specimens with alkali feldspar phenocrysts. All these specimens belonged to Tertiary or younger alkaline volcanic rocks and thus offered excellent material for the study of natural high-temperature alkali feldspars.

[1] Received March 21, 1960.

A closer examination of the specimens took place in 1959 of the laboratory of the Institute of Geology and Mineralogy, University of Helsinki.

To obtain an idea about the material at hand, the composition of the feldspars was determined by means of X-ray powder patterns, using the method of Bowen and Tuttle (1950). It turned out that the compositions ranged from Or_{15} to Or_{40}, a part being monoclinic, and another triclinic. For the detailed investigation six specimens were then selected so that their compositions were distributed evenly over the range Or_{15}—Or_{40}. Accordingly, the following FEAE specimens were chosen: 145, 161, 173, 183, 195 and 200. To avoid contamination of the alkali feldspars of the groundmass, the specimens were crushed only with a jaw crusher. The phenocrysts were then hand picked and ground down to a grain size suitable for final separation, which was carried out with Clerici's solution. To remove the rest of the nepheline, the powder was treated with diluted hot hydrocloric acid. The material thus obtained was quite pure.

BRIEF PETROGRAPHICAL DESCRIPTION OF THE SPECIMENS

FEAE 145. Kapiti Phonolite, Stony Athi River. The groundmass of the rock is dark bluish-grey and very fine-grained (average grain size 0.03 mm). It contains large phenocrysts of alkali feldspar (1—5 cm in length) and somewhat smaller phenocrysts of nepheline. In addition to feldspar and nepheline, the groundmass is composed of aegirine, basaltic hornblende (which is probably katophorite), analcite, magnetite, chlorite and apatite.

FEAE 161. Tranchyte, Timau-Nanyuki road, Lorimenji River. The specimen labelled 161 contains a number of loose feldspar crystals only. More detailed data concerning the rock are lacking.

FEAE 173. Trachyte, Nairobi-Arusha road, NE of Lassarkartarta. Viewed as a hand specimen, the rock has a bluish-grey tint. It is compact and porphyritic. The phenocrysts of alkali feldspar are rather large (0.5 cm in length) and clear. Under the microscope they show distinct zonig, especially in the margins of the phenocrysts. In addition to feldspar the rock contains microphenocrysts of basaltic hornblende and aegirine-augite. The groundmass is coarse-grained (grain size 0.5 mm) and is composed of feldspar, aegirine-augite, basaltic hornblende, sphene, magnetite and apatite. Zeolites are almost completely lacking. Only a small amount of analcite has been noticed and identified by the X-ray method.

FEAE 183. Trachyte, Arusha-Moshi road, Kikafu River. The rock closely resembles specimen 173. Its groundmass is somewhat less coarse-grained (average grain size about 0.1 mm), and instead of aegirine it contains augite. In addition, the groundmass is characterized by iddingsite, chlorite, sodalite and apatite. In some amygdules a fibrous zeolite is seen. It s probably natrolite.

FEAE 195. Tuff, NE rim of Menengai Crater. This specimen contains the most beautiful water-clear feldspar crystals encountered in the whole material. The groundmass is composed of brown, porous material non-identifiable under the microscope, in which there are embedded small crystals of feldspar, olivine and clinopyroxene. In addition, some volcanic glass can be found.

FEAE 200. Phonolite, Sergoit River, Eldoret-Sergoit road. The rock shows feldspar laths in the compact dark-grey groundmass. Some nepheline phenocrysts are also visible. The groundmass is fine-grained and composed of needle-like potassium-rich feldspar, nepheline and analcite, which is very abundant and replaces the nepheline and feldspar along the cracks. In addition, the groundmass contains basaltic hornblende, aegirine-augite, apatite, biotite and magnetite.

CHEMICAL COMPOSITION

The content of the unit cells is calculated from the chemical analysis, cell volume and density and presented in Table 1, together with the chemical composition. A distinct difference in the composition of the feldspars of phonolites and trachytes is seen at once. The feldspars of the former are pure alkali feldspars, while those of the latter contain, in addition to alkalis, variable amounts of calcium. (Specimens 145 and 200 are, it will be recalled, phonolites, and 161, 173 and 183 trachytes). This is probbaly due to differences in the crystallization temperature, which in trachytes, as a rule, is higher than in phonolites.

Characteristic of all the feldspars investigated is their relatively high Ba content. As far as can be judged from the few data available, Ba seems to be distributed equally between the feldspars of phonolites and trachytes.

Table 1. Chemical Composition and Unit Cell Content of the Six Alkali Feldspars.
Analyst: H. B. Wiik.

FEAE	145	161	173	183	195	200
SiO_2	65.61	62.74	65.16	64.08	66.40	65.42
Al_2O_3	20.48	22.88	20.95	21.18	19.84	20.30
Fe_2O_3 [1]	0.20	0.20	0.26	0.47	0.32	0.20
CaO	0.0	3.11	1.27	2.39	0.22	0.0
Na_2O	7.03	7.57	7.92	7.90	6.91	6.16
K_2O	6.26	2.94	4.02	3.47	6.46	7.41
BaO	0.32	0.21	0.35	0.14	0.01	0.48
H_2O-	0.00	0.02	0.05	0.00	0.01	0.00
H_2O+	0.03	0.13	0.08	0.10	0.05	0.10
Total	99.93	99.80	100.06	99.73	100.22	100.07

[1] Total iron.

FEAE	145	161	173	183	195	200
Si	11.76	11.25	11.67	11.49	11.87	11.73
Al	4.33	4.84	4.42	4.48	4.18	4.29
Fe	0.03	0.03	0.04	0.06	0.04	0.03
Ca	0.00	0.60	0.24	0.46	0.04	0.00
Na	2.44	2.63	2.75	2.75	2.40	2.14
K	1.45	0.67	0.92	0.79	1.47	1.69
Ba	0.02	0.01	0.02	0.01	0.00	0.03
O	32.02	32.06	32.13	32.03	32.07	31.89
Or	36.8	17.1	23.4	19.7	37.6	43.8
Ab	62.7	67.3	70.0	68.6	61.4	55.4
An	0.0	15.3	6.1	11.5	1.0	0.0
Cels.	0.5	0.3	0.5	0.2	0.0	0.8

X-RAY AND OPTICAL DATA

The material analyzed was used for X-ray study. The powder patterns were recorded with a Philips Norelco Geiger Counter diffractometer using CuKα radiation. All the runs were made from $2\theta = 13°$ to $2\theta = 57°$, with the internal silicon standard at a rate of 0.125° per minute. The readings of the peak positions were taken at the mid-point of peak width at approximately two-thirds of peak height to minimize the interference of the α_2 peak. The calibration correction of 2θ was found to vary from —0.04° to + 0.03°. Peaks owing to only one reflection were used to derive the values of the reciprocal-lattice parameters. The conversion of the reciprocal cells to direct cells was carried out with the aid of the scheme given by Donnay and Donnay (1952).

Table 2. Unit Cell Parameters, Optical Data and Densities of the Six Alkali Feldspars.

FEAE	145	161	173	183	195	200
a_o (Å)	8.298	8.219	8.250	8.234	8.308	8.322
b_o (Å)	12.950	12.912	12.930	12.925	12.969	12.970
c_o (Å)	7.160	7.130	7.145	7.138	7.158	7.161
α	90.64°	92.69°	92.05°	92.73°	—	—
β	116.37°	116.39°	116.34°	116.41°	116.28°	116.23°
γ	90.04°	90.28°	90.12°	90.10°	—	—
V (Å^3)	689.3	677.3	682.5	679.4	691.6	693.3
α	1.527	1.535	1.5305	1.5335	1.5255	1.5255
β	1.533	1.5415	1.537	1.5405	1.530	1.531
γ	1.535	1.5445	1.540	1.5435	1.531	1.5325
2 Vα	52°	63°	55°	61°	49°	49°
$c \wedge b$	21°	22°	16°	21°	14°	15°
D (gm/cm^3)	2.591	2.631	2.615	2.622	2.583	2.578

In Table 2 the unit cell parameters and optical properties are presented, together with densities. The refractive indices are determined by the immersion method in sodium light to an accuracy of 0.0005. Because of the zoning, there is some fluctuation in the values of the refractive indices, the magnitude of which is about $\pm$ 0.001. The 2Vα angles were measured conoscopically, the results being correct within $\pm$ 1 degree. For the determination of density, a pycnometer was used.

The dimensions of the unit cells obtained in this investigation agree closely with the data given by Donnay and Donnay (1952). Some slight discrepancy is nevertheless to be seen. Perhaps the most outstanding difference is in the location of the transition point between monoclinic and triclinic feldspars. According to Donnay and Donnay (1952) monoclinic feldspar changes to triclinic at room temperature when its composition is Or_{33}. The present study shows however that the change of symmetry takes place at Or_{37}, i. e., at the same point as reported by MacKenzie and Smith (1956). The discrepancy is probably due to the difference in the degree of Al—Si order, which is known to vary considerably in natural feldspars.

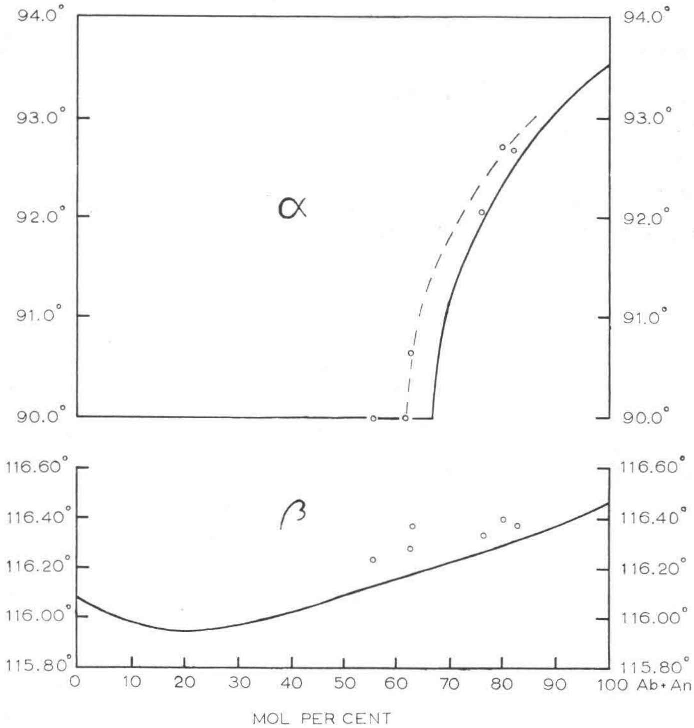

Fig. 1. The α- and β-angles of the six anorthoclases as a function of the composition. Solid lines: according to Donnay and Donnay. Circles and dashed line: according to the present study. The feldspars in order of the growing Ab-content: 200, 195, 145, 173, 183 and 161.

In Figure 1 the α and β angles are plotted against the composition. As is seen, β shows systematically slightly higher values than those recorded by Donnay and Donnay (1952), but the deviation is small and obviously due to a difference in composition.

As regards Figure 2, where the d-values of the $20\bar{1}$-reflections are plotted versus composition, one point os of interest. The Or-content of the three component feldspars can be determined quite accurately with the aid of $20\bar{1}$-reflection, using the diagram of Bowen and Tuttle (1950). This means that the effect of sodium and calcium upon the unit cell dimensions is very nearly the same. Even though a feldspar should contain as much anorthite as that of specimen FEAE 161 (15.3 %), its d-value, corresponding to $20\bar{1}$-reflection, still lies in the curve, within the limits of error.

The An-component of such a feldspar is easily found by measuring the indices of refraction and comparing the results with the corresponding diagram of the plagioclases. This has been done in Figure 3. As is seen, the refractive indices of the anorthoclases are virtually identical with those of plagioclases.

The values of the $2V\alpha$ angles of the six feldspars show them to belong to anorthoclase-sanidine-cryptoperthites. In fact, the observed angles are somewhat higher than those of the anorthoclase-sanidine-cryptoperthites proper, so that the studied feldspars are structurally intermediate between the anorthoclase-sanidine and low-albite-orthoclase series, respectively. In addition, the feldspars are not unmixed, even though they belong to the part of the anorthoclase-sanidine series, where cryptoperthites commonly exist. The X-ray powder patterns show only one $20\bar{1}$-reflection for each

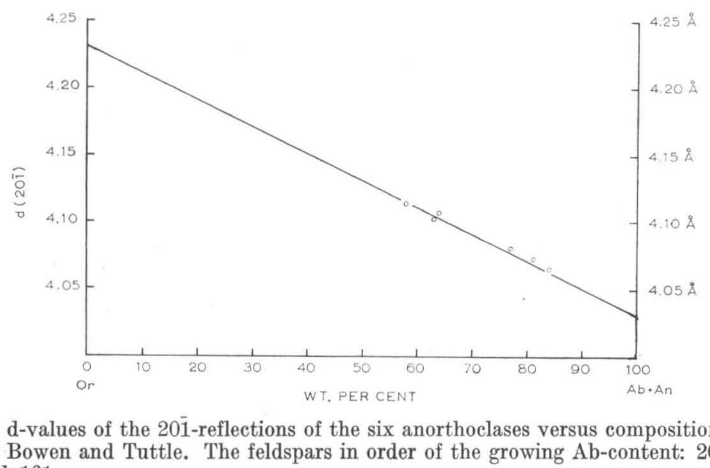

Fig. 2. The d-values of the $20\bar{1}$-reflections of the six anorthoclases versus composition. Solid line: according to Bowen and Tuttle. The feldspars in order of the growing Ab-content: 200, 195, 145, 173, 183 and 161.

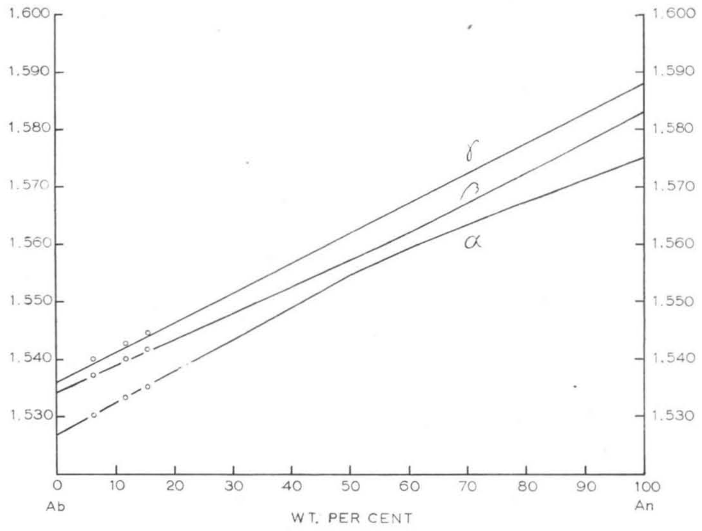

Fig. 3. The indices of refraction of the three An-bearing anorthoclases compared with those of the high-temperature plagioclases. As is seen, the agreement is perfect. The feldspars in order of the growing An-content: 173, 183 and 161.

specimen. On the other hand, in oscillation photographs, taken by the method of MacKenzie and Smith (1955), some reflections can be seen to have changed to the diffuse streaks parallel to the layer lines, indicating the existence of the pericline type superstructure. This is noticeable, above all, in the monoclinic feldspars 195 and 200, and in the triclinic feldspar 145.

The monoclinic feldspars have $b//\gamma$, and the axial plane $\perp$ (010). The triclinic feldspars are optically» pseudo monoclinic», i. e., they have $\gamma \perp (010)$, or very nearly so. The optical orientation shows, however, some fluctuation from grain to grain, even in the same specimen, and α has been noticed to differ slightly from the (010) plane in a few cases.

GROUNDMASS FELDSPARS

After the composition of the feldspars of phenocrysts was determined, attention was paid to the groundmass feldspars. Specimens 145, 173, 183 and 200 were used for this study. The separation of the groundmass feldspars was carried out taking special care to avoid the contamination caused by the phenocryst feldspars. The material obtained was not, however, pure enough for chemical analyses and, consequently, the composition of the feldspars was determined by means of the $20\bar{1}$-reflection.

An outstanding difference in the groundmass feldspars of the phonolites and trachytes was established. In the groundmass of the former, only one potassium-rich monoclinic feldspar is encountered. The feldspars of specimens 145 and 200 have the compositions Or_{82} and Or_{90}, respectively.

In the trachytes the groundmass feldspars have become unmixed, in specimen 173 into two, and in 183 into three components, the compositions of which are: Or_{82} and Or_{11}, and Or_{96}, Or_{78} and Or_{6}, respectively. By heating at 1 000 C° both the last-mentioned feldspars were homogenized, which was established by powder patterns, where only one $20\bar{1}$-peak was seen. According to the positions of the $20\bar{1}$-reflections the composition of the homogenized feldspars turned out to be in specimen 173: Or_{44}, and in specimen 183: Or_{47}.

It is obvious, from the petrographical investigation, that the difference in the mode of occurrence of groundmass feldspars in phonolites and trachytes is mainly due to the different degree of zeolitization. The phonolites studied often contain abundant analcite, while in trachytes this mineral is almost completely lacking. During the cooling, the groundmass feldspars have become unmixed both in the phonolites and trachytes into potassium- and sodium-rich components. In the phonolites the sodium feldspar has then analcitized, while in the trachytes both components have survived.

SUMMARY

From the data obtained from the investigated alkali feldspars, the following conclusions can be drawn.

There exists a distinct difference between the phenocryst feldspars of the phonolites and trachytes. The feldspars of the former are pure alkali feldspars, while those of the latter contain, in addition, anorthite.

The refractive indices of An-bearing anorthoclases are practically the same as those of plagioclases containing an equal amount of anorthite. This can be explained by the strong influence Ca has upon the refractive indices; in this respect Na and K behave nearly identically.

The triclinic anothoclase-sanidine-cryptoperthites are optically »pseudo monoclinic.» This is probably due to the stress prevailing in the crystal lattices, foreshadowing the incipient unmixing.

The An-component has scarcely any influence upon the d-value of the $20\bar{1}$-reflection of the anorthoclases, which is determined almost completely by the Or-content of the feldspars.

The groundmass feldspars have become unmixed in both the phonolites and trachytes studied. The albite component of the former has then analcitized, whereas in the latter it has remained unchanged.

Acknowledgments — The author is greatly indebted to Prof. Th. G. Sahama, who placed the excellent material used in this investigation at his disposal. During all stages of the work Prof. Sahama offered valuable advice and, in addition, contributed penetrating criticism of the manuscript. Dr. K. J. Neuvonen cheerfully gave counsel in intepreting the oscillation photographs. Dr. Kai Hytönen offered valuable assistance in different forms during the investigation. The chemical analyses were made by Dr. H. B. Wiik, of the Geological Survey of Finland. From the Outokumpu Company Foundation (Outokumpu Osakeyhtiön Säätiö) the author received a grant to cover some of the costs of the investigation. Mr. Paul Sjöblom, M. A., corrected the English text. To all the persons mentioned the author expresses his gratitude.

REFERENCES

Bowen, N. L. and Tuttle, O. F. (1950) The system $NaAlSi_3O_8$—$KAlSi_3O_8$—H_2O. Geology Vol. 58 No. 5.

Campbell Smith, Walter (1931) A classification of some rhyolites, trachytes, and phonolites from part of Kenya with a note on some associated basaltic rocks. Quart. J. Geol. Soc. Vol. 87. Part 2.

Chayes, F. (1952) Relations between composition and indices of refraction in natural plagioclase. Am. J. Sci.

Donnay, Gabrielle and Donnay, J. D. H. (1952) The symmetry change in the high-temperature alkali-feldspar series. Am. J. Sci. Bowen Volume.

Eitel, Wilhelm (1958) Structural conversion in cristalline systems. Bull. Geol. Soc. Am. Special Paper 66.

Goldsmith, Julian, R. and Laves, Fritz (1954 a) The microcline-sanidine stability relations. Geochim. et Cosmochim. Acta. Vol. 5.

—»— (1943 b) Potassium feldspare structurally intermediate between microcline and sanidine. Geochim. et Cosmochim. Acta. Vol. 6.

Hytönen, Kai (1959) On the petrology and mineralogy of some alkaline volcanic rocks of Toror Hills, Mt. Moroto, and Morulinga in Karamoja, Northeastern Uganda. C. R. Soc. géol. Finlande 31. Bull. Comm. géol. Finlande 184,6.

Laves, Fritz (1952 a) Phase relations of alkali feldspars I. Introductory remarks. J. Geology Vol. 60.

—»— (1952 b) Phase relations of alkali feldspars II. The stable and pseudo-stable phase relations in the alkali feldspar system. J. Geology Vol. 60.

MacKenzie, Scott (1952) The effect of temperature on the symmetry of high-temperature soda-rich feldspars. Am. J. Sci. Bowen Volume.

—»— (1954 The orthoclase-microcline inversion. Mineralog. Mag. Vol. 30.

MacKenzie, Scott and Smith, J. V. (1955) The alkali feldspars. I Orthoclase microperthites. Am. Mineralogist Vol. 40.

—»— (1956) The alkali feldspars III. An optical and X-ray study of high-temperature feldspars. Am. Mineralogist Vol. 41.

Oftedahl, Christoffer (1950) Note on »pseudo-monoclinic» plagioclase. J. Geology Vol. 58.

Pulfrey, William (1947) The geology and mineral recources of Kenya. Bull. Imp. Inst. Vol. XLV No. 3.

Schairer, J. F. (1950) The alkali-feldspar join in the system $NaAlSiO_4$—$KAlSiO_4$—SiO_2. J. Geology Vol. 58.

SCHOEMAN, J. J. (1951) A geological reconnaissance of country between Embu and Meru. Geol. Survey Kenya. Rept. No. 17.

SHACKLETON, R. M. (1946) Geology of the country between Nanyuki and Maralal. Geol. Survey Kenya, Rept. No. 11.

SIKES, H. L. (1939) Notes on the geology of the country surrounding Nairobi. Geol. Survey Kenya.

SMITH, J. R. (1958) The optical properties of heated plagioclases. Amer. Miner. Vol. 43.

—»— and MACKENZIE, W. S. (1955) The alkali feldspars II. A simple X-ray technique for the study of alkali feldspars. Am. Mineralogist Vol. 40.

—»— (1958) The alkali feldspars IV. The cooling history of high-temperature sodium-rich feldspars. Am. Mineralogist Vol. 43.

TUTTLE, O. F. (1952) Optical studies on alkali feldspars. Am. J. Sci. Bowen Volume.

—»— and BOWEN, N. L. (1950) High-temperature albite and contiguous feldspars. J. Geology Vol. 58.

WINCHELL, A. N. and WINCHELL, H. (1951) Elements of optical mineralogy. Part II, fourth edition, John Wiley & Sons Inc., New York.

9

THE AGE OF THE SUBFOSSIL ROOTS OF *ALNUS GLUTINOSA*, TUUSULA, SOUTHERN FINLAND [1]

BY

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ABSTRACT

In southern Finland, about 28 km north of Helsinki, there occur subfossil roots of black alder *(Alnus glutinosa)*. Stratigraphic studies have indicated that the swampy black alder forest was submerged by a Baltic transgression either during the Ancylus or the Echeneis stage. The radiocarbon age of the subfossil roots, determined at the C^{14}-Station of Stockholm (St 527), is 4 600 ± 115 B. P. At this time the Baltic Sea had retreated to a level 25 meters below the basin where the roots are located. By means of the roots it can be established that black alders grew at the site at the end of the Atlantic period. A comparison with the dated diagram from Norrland, Sweden, shows great similarities between the pollen contents of peat from both regions around the year 4 600 B. P.

INTRODUCTION

In the commune of Tuusula, some 28 km north of Helsinki, there is a hill, known as Mätäkivenmäki, built of glaciofluvial matter. The summit of the hill reaches an altitude of 80 meters above sea level. After deglaciation the hill was submerged up to the top. Gradually the water level lowered as is shown by the shore deposits and raised beaches met with on the summit and the slopes. The most prominent of the raised beaches is a wave-cut cliff, the foot of which lies at an elevation of 61.5 meters, on the southern slope of the hill. Also on the northern slope there occurs a cliff on the same level. In addition, less distinct raised beaches are met with on the eastern slope at an elevation of 59.5—62 meters. The location of these beaches is shown in sketch maps published by Sauramo (1936, 1958).

[1] Received March 2, 1960.

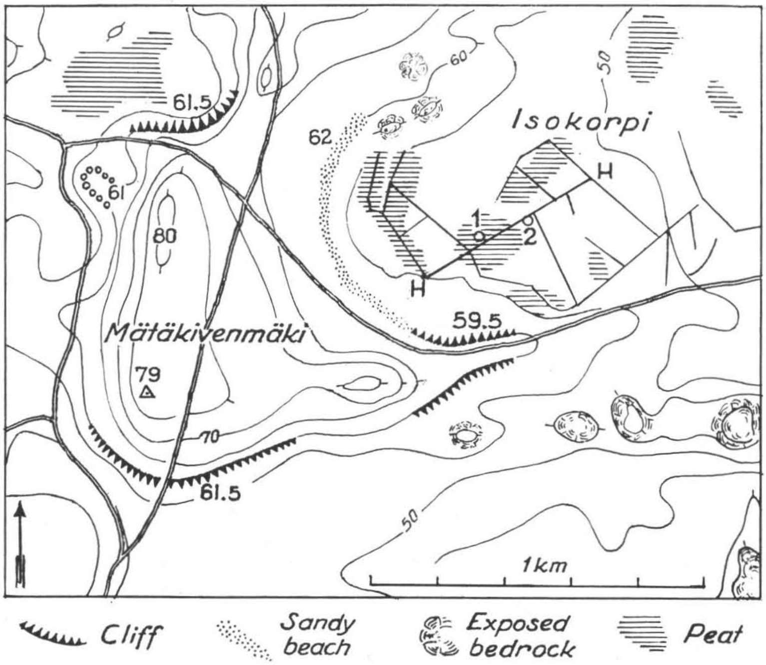

Fig. 1. The location of the Isokorpi bog. The sites of the samples are indicated by the numbers 1 and 2. No. 1 also denotes the site of the diagram by Sauramo (1958), reproduced in Fig. 2.

According to earlier studies (Rudeberg, 1925; Sauramo, 1936; Hyyppä, 1950) the raised beaches described in the foregoing denote the highest limit of the Ancylus transgression in Tuusula. Recent investigations (Sauramo, 1954; 1958; Virkkala, 1959; see also Hyyppä, 1937, Appendix II) indicate that waters never reached the cliff level during this transgression. Hence the cliffs must have been formed during an earlier phase of the Baltic.

Sauramo (1958, p. 144) was of the opinion that the cliffs were formed in the beginning of the Boreal period, when the Yoldia stage culminated in the Echeneis transgression. The elevation of this transgression is, according to Sauramo *(op. cit.)*, 61.5 m in Tuusula. According to Virkkala the Mätäkivenmäki cliffs should be older: they most probably date from the turn of the late-glacial into the post-glacial period (Virkkala, 1959, p. 89).

Both Sauramo's and Virkkala's conclusions are based on stratigraphic sequences. Among the pollen analyses in Sauramo's material the pollen diagram of Isokorpi (Sauramo, 1936; 1958, p. 137—145) is of interest in this connection. Isokorpi (korpi = spruce and deciduous tree peat-bog) is located in Ruotsinkylä, on the northeastern side of Mätäkivenmäki (Fig. 1). In the 1920's, the Isokorpi bog was drained, exposing thin roots of trees, embedded in sand. Determinations by V. Kujala (1924; Sauramo, 1936; 1954) showed that these roots are from the black alder *(Alnus glutinosa)*. J. Donner has confirmed the results of Kujala by microscopic determination (Sauramo, 1954, p. 215). When studying the local stratigraphy Sauramo established that the roots are *in situ* and represent remnants of a black alder community. He concluded that the alder forest was submerged by a Baltic transgression (Sauramo, 1936, 1958). Determinations by Donner (1954) showed that the pollen of *Alnus* encountered in the zone IV [1] in the Hanaböleträsk bog, about 5 km southeast of Ruotsinkylä, is from *Alnus glutinosa*, and thus confirmed the assumption that black alder flourished in the Tuusula area during that time.

Sauramo showed that the formation of peat started in the Isokorpi basin in the beginning of the pollen-analytical zone V b, or before the Ancylus transgression took place. According to Sauramo this transgression coincides with the zone V c. As peat was formed throughout V c in the Isokorpi basin, the Ancylus transgression did not reach the localities, or the elevation of 51 m at Ruotsinkylä. As a consequence, Sauramo concluded, black alders must have grown here prior to the Echeneis transgression, which preceded the Ancylus stage in the history of the Baltic.

In 1953 Professor Sauramo led an undergraduate excursion headed by Mr. Sten Florin, Associate Professor, and Mrs. Maj-Britt Florin, instructor, from the University of Uppsala, Sweden. I had the opportunity to take part in this excursion. Sauramo's interpretation of the locality started an animated discussion. Mrs. Florin collected some samples for diatom analysis, the results of which she reported orally to me during my visit to Uppsala in January, 1960.

In the summer of 1959 I had the pleasure of making several field trips in the company of Professor Hans Reinhard, Greifswald, DDR. During one trip we visited Isokorpi. We collected three soil samples and a representative sample of the roots. Miss Kyllikki Salminen, Lic. phil., of the Geological Survey of Finland, determined the pollen flora and the diatoms contained in the samples. Through the friendly mediation of Professor Gösta Lundqvist, the roots were dated by the radiocarbon method at the C 14-station in Stockholm, Sweden. The determination was financed by the State Scientific Board.

LOCATION OF THE SAMPLES

The samles were collected from the ditch marked H, as shown in Fig. 1.

SAMPLE NO. 1 came from the bottom of the bog, underneath the peat layer. The sample consists of detritus mud containing a large amount of sand. Sauramo (1958, p. 142) has published a profile and a pollen diagram from the same locality (Fig. 2). According to the profile the altitude of the

[1] In the present paper the zonation proposed by Jessen is used. (cf. Sauramo 1958).

site is 50.9 meters above sea level. The mud represents the youngest limnic sediment in this locality; upwards a continuous layer of peat is met with. According to Sauramo *(loc. cit.)* the formation of *Carex*-peat started in the pollen-analytical zone V b, and continued until the forest spread over the bog surface, which happened in the first part of zone VII. The forest was probably dominated by deciduous trees as indicated by the abundance of pollen of *Betula*, *Alnus*, *Tilia* and *Corylus* in the diagram. The deposition of sand on the bottom of the basin coincides with zone V a (Fig. 2). The diatoms of the sand represent a flora typical of great lakes (Sauramo, 1958, p. 142).

SAMPLE NO. 2 exemplifies the layer of sand in which the alder roots are embedded. The locality is situated about 100 meters northeast of the site of sample No. 1, in ditch H (Fig. 1). The elevation of the ground here is 52.5 meters (according to the drainage map prepared by the Forest Research Institute: see also Sauramo, 1958, Fig. 34 on page 138). The drain, here 1.5 meters in depth, has been dug in sand. The sand is rather loose, is brown in color and contains much fine sand and humus matter. No bedding was to be observed in the wall of the drain. The uppermost layer, about 30 cm thick, is lighter in color and drier than the underlying sand. The zone of the roots begins about 50 cm below the surface and continues downward at least to the bottom of the drain. That gives a minimum thickness of one meter. In the upper part of the zone the roots are darker and more decayed than in the lower, moist bed of sand. The dark roots tend to break up into pieces under the lightest touch, whereas it is easy to collect 10—15 cm-long specimens from the lower bed. In general the rind of the roots is better preserved and harder than the soft wood matter. The thickness of the roots varies between 1—5 mm.

SAMPLE NO. 2 was collected at a depth of 70 cm, or at an elevation of 51.8 meters. The root sample was collected from the same layer between the depths 70—120 cm.

SAMPLE NO. 3 is from the same place as sample No. 2, but from nearer the surface, at the depth of 40 cm. The sand is similar to that in sample No. 2, but no roots are be observed in the upper sand layer.

POLLEN AND DIATOM ANALYSES

The results of the pollen analyses, done by Miss Kyllikki Salminen, are shown in Table I.

The pollen flora of sample No. 1 is in accord with the pollen diagram by Sauramo (Fig. 2). The Isokorpi basin was covered by shallow waters in the beginning of the Boreal period, zone V. The percentage of *Pinus* is so high that corresponding values are not known from previous periods (comp. the

Table 1.

Arboreal pollen	1	2	3	Non-arboreal pollen	1	2	3
Alnus	1	—	3	Artemisia	—	3	1
Betula	22	25	38	Chenopodiaceae	—	1	—
Corylus	1	4	5	Compositae excl. Artemisia	—	1	—
Picea	1	1	5	Cyperales	—	1	2
Pinus	74	66	49	Ephedra	—	1	1
Quercus	—	3	—	Ericales	—	—	1
Ulmus	1	1	—	Gramineae	2	5	1
	100	100	100%	Hippophae	1	2	—
				Ranunculaceae	3	—	—
				Rosaceae	1	1	—
				Salix	2	1	2
				Varia	1	3	1
					10	19	9
				Σ NAP	10	19	12

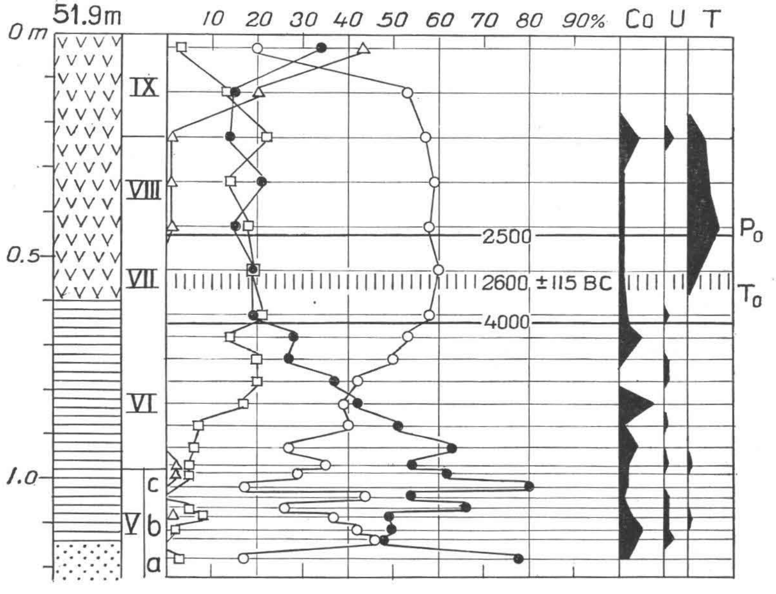

Fig. 2. Vertical profile and dated pollen diagram from the Isokorpi basin, dammed up by beach sand (according to Sauramo 1958). The lower and upper limits of zone VII are drawn according to the generalized diagram by Sauramo *(op. cit.)*. In addition, the absolute age of the subfossil roots of alder, determined to be 2 600 years B. C. by the radiocarbon method, is marked in the diagram.

generalized diagram by Sauramo 1958, Fig. 8 on p. 44). Deposition of the sand forming the substratum of the black alder forest seems to have occurred during the Boreal period. The abundant occurrence of pollen of deciduous trees, however, points to a slightly younger climatic phase. *Pinus* is dominant in sample No. 3, too, but the share of deciduous trees is somewhat higher than in sample No. 1. As much as 5 % of the pollen in sample No. 3 is made up of *Picea*, whereas the corresponding figure for samples No. 1 and No. 2 is 1 %.

The abundance of non-arboreal pollen (NAP) in relation to arboreal pollen (AP) is rather uniform, ranging 10—19 NAP/100 AP. The relatively high amount of NAP may indicate the influence of a nearby shore (Donner, 1957). Most of the NAP species are even at the present day the first to spread out on bare ground recently freed from waters. A special feature of the NAP group is the ocurrence of pollen of *Ephedra* in samples No. 2 and No. 3. Pollen of *Ephera* has been identified in Finnish samples only recently (Tynni, 1959; Salmi, 1959). Tynni has found this pollen from samples collected in southern Finland. His investigations show that *Ephedra* pollen is met with in pollen zones I—IV, possibly in zone V, too (Tynni, *op. cit.*, p. 125). Miss Ester Uussaari, Mag. phil., of the Geological Survey of Finland, has recently found pollen of *Ephedra* in samples collected from several localities, including northern Finland (Salmi, 1959). Salmi's conclusion is that »pollen of *Ephedra* is likely to be found in ancient sediments throughout Finland» (*op. cit.*, p. 12).

Providing that the pollen of *Ephedra* in the beach sand at Isokorpi is in primary position, this Eurasian steppe shrub must have prospered in Finland as late as the Boreal period. However, the pollen may occur only by chance as the beach sand itself is redeposited, and thus its position would be secondary. Only one of the findings of *Ephedra* originates from an organogenic deposit. This deposit has been dated into zone V (Tynni, *op. cit.*, p. 125) but the zonation seems to be somewhat uncertain. Considering the facts that the till of Finland contains an abundance of secondary pollen, also met with in clays (Heinonen, 1958), and that the interglacial deposits near Leningrad include pollen of *Ephedra* (Znamenskaia, 1959, p. 401), one must be extremely careful in weighing the stratigraphic position of the *Ephedra*.

The results of the diatom analyses are shown in Table II. All three samples contain rather numerous diatoms. The majority of species in all the samples belong to the group typical of shallow fresh water *(Fragilaria, Navicula cocconeiformis, N. tuscula, Opephora Martyi, Tabellaria)*. A few species belong to the so-called Ancylus-flora, which flourishes in large bodies of clear fresh water *(Amphora ovalis, Campylodiscus noricus, C. noricus* var. *hibernica, Cymatopleura elliptica, Diploneis domblittensis, Epithemia Hyndmanni, Gyrosigma attenuatum, Melosira arenaria* and *Stephanodiscus astraea)*. Although the abundance of the species is relatively low, the occurrence of this group in all three samples is probably due to the fact that the fresh waters of the Ancylus Lake did reach the 51—52 meter level at Ruotsinkylä. Obviously the waters were shallow and the connections to the open lake slight; hence no distinct Ancylus-flora could develop. During the excursion in 1953 (p. 3)

Table 2.

Species	Sample No. 1	2	3
Achnanthes calcar	—	1	2
» Clevei	—	1	—
» » var. rostrata	—	—	1
» flexella	1	1	1
» Östrupi	—	1	1
Amphora ovalis	2	1	—
» » var libyca	2	2	—
» » var. pediculus	3	7	1
Anomoeneis sphaerophora	1	—	—
Campylodiscus noricus	—	1	—
» » var. hibernica	1	1	—
Caloneis latiuscula	1	1	—
» silicula	1	—	—
Cocconeis diminuta	—	1	2
» placentula var. euglypta	2	4	3
Cymbella Ehrenbergi	1	—	—
» prostrata	—	1	1
» sp.	6	4	2
Cymatopleura elliptica	1	1	1
» solea	1	—	—
Diploneis domblittensis	1	1	2
» elliptica	1	—	—
» Mauleri	—	1	—
» Smithi	1	—	—
Epithemia argus	1	—	—
» Hyndmanni	2	1	1
» intermedia	1	—	—
» sorex	1	—	—
» turgida	2	—	—
» » var. granulata	2	—	—
» » var. Westermanni	1	1	1
» zebra	1	—	—
» » var. porcellus	4	1	—
» » var. saxonica	2	5	3
Fragilaria sp.	40	33	45
Gomphonema sp.	2	—	—
Gyrosigma acuminatum	2	—	—
» attenuatum	1	1	1
Mastogloia Grevillei	1	1	—
» Smithi	—	1	—
» » var. amphicephala	1	—	—
» » var. lacustris	1	—	—
Melosira arenaria	—	1	1
» islandica subsp. helvetica	7	4	—
Navicula cocconeiformis	1	26	12
» cuspidata	3	—	—
» gastrum	1	—	—
» platystoma	1	1	1
» pupula	1	—	—
» radiosa	1	1	—
» tuscula	1	10	3
Neidium iridis	1	—	—
Opephora Martyi	2	2	14
Pinnularia sp.	2	1	—
Rhopalodia gibba	2	1	—
» gibberula	2	—	—
Stauroneis anceps	1	—	—
» phoenicenteron	1	—	—
Stephanodiscus astraea	1	1	—
Surirella biseriata	1	—	—
Synedra capitata	1	—	—
» ulna	2	—	—
Tabellaria fenestrata	2	2	—
» flocculosa	—	1	—
	125	125	99

Mrs. Maj-Britt Florin took a sample of the sand surrounding the alder roots. The sample contains a diatom flora typical of small fresh water lakes. She sent the list of species to Professor Sauramo (oral communication by Mrs. Florin).

In addition to the fresh water flora a few indicators of slight salinity are met with in the samples *(Diploneis Smithi, Mastogloia Smithi, M. Smithi* var. *amphicephala, Epithemia turgida* var. *Westermanni).* They are most abundant in sample No. 1. As has been pointed out (e. g., Hyyppä, 1937) the occurrence of this group of diatoms among the Ancylus-flora is rather common. It may be due either to the fact that the salinity of the Baltic diminished gradually, or that these diatoms have been redeposited by shore action. Providing that this group of diatoms in the samples indicates a slight salinity, the water in the basin was somewhat more saline during the early formation of peat than at the time the beach sand was deposited. The evidence presented in this paper is insufficient to support such a conclusion. However, my opinion is that the material presented here suffices to confirm the results of Sauramo (1958) and Virkkala (1959) concerning the isolation of the Isokorpi basin during the Boreal period, or in the pollen zone V. Neither the pollen analysis nor the study of the diatoms showed any essential differences among the samples. On the contrary, the microflora of samples No. 2 and No. 3 resemble each other to such an extent that the sand containing roots of *Alnus* and the overlying sand were deposited under the same conditions.

THE AGE OF THE ROOTS OF ALNUS

The root specimens collected at the site of sample No. 2 were dated by the CO_2-method using a proportional counter at the C^{14}-Station of Stockholm (Östlund, 1956). The age of the roots is (St 527)

$$4\,600 \pm 115 \text{ years B. P.}$$

Comparing this age to the generalized pollen diagram for Finland, dated by Sauramo (1955, p. 9; 1958, p. 44), it is to be seen that the year 2 600 B. C. places itself in the upper part of the pollen zone VII, or at the end of the Atlantic period. This period coincides with the Littorina stage in the development of the Baltic. The shore line of this stage is well established in the Helsinki region (Aario, 1935; Hyyppä, 1937; Sauramo, 1958). At the end of the Atlantic period the shore line was located some 25 meters below the Isokorpi basin. Thus it is impossible that the black alders could have been destroyed by a Baltic transgression. Evidently the trees rotted in a regular way. The decay has proceeded so far that only the deepest parts of the roots are left. The decay of these remains was probably retarded by a lack of air in the water-saturated sand. As pointed out in the description of the sample sites (p. 4), no signs of a transgression were to be observed in the field.

DISCUSSION

According to radiocarbon age determination a black alder vegetation flourished in the Isokorpi basin around the year 2 600 B. C., or at the end of the pollen-analytically established Atlantic period. It is known that black alder forests were common on shores and bogs in southern Finland at that time (Auer, 1952, p. 241).

The pollen diagram from Isokorpi, dated by Sauramo (1958, Fig. 38 p. 142; Fig. 2) shows that in the basin dammed up by beach sand the formation of peat began at the time of zone V. The basin was an open *Carex*-bog until a forest spread out on the bog at the time of pollen zone VII. The formation of *Carex*-peat stopped and wood peat was formed. The wood peat contains considerably pollen of deciduous trees. Among these *Alnus* comprises 20 % of the total tree pollen. Most probably black alder grew in this part of the basin, too. The comparison gives an important result, the absolute age of pollen zone VII. It also confirms Sauramo's dating of the generalized diagram. In the Isokorpi basin the rational border for *Picea* P_0 lies slightly above zone VII, and is, accordingly, younger. The rational border for *Tilia* T_0 is situated in pollen zone VII, approximately where the formation of wood peat begins. The minimum age of T_0 in the Helsinki region is thus 2 600 years B. C.

The conclusions stated in the foregoing have significance in giving the first absolute reference niveau to the Finnish pollen chronology. True, there is an earlier radiocarbon dating. The age of the lowermost peat layer at Varrassuo on the First Salpausselkä was determined at Yale, but the result 8 030 $\pm$ 140 years B. P. (Barendsen, Deevey and Gralenski 1957) must be regarded as too young. Most probably secondary impurities, such as younger humus matter dissolved in groundwater, have been mixed into the primary organogenic matter. This may also be true in the case of the subfossil roots of alder, but a comparison of the age with the stratigraphy of the Isokorpi bog indicates that the result of the age determination can be relied upon.

In the generalized diagram by Sauramo P_0 is located slightly above pollen zone VII, whereas T_0 falls within zone VII proper.

In addition to Sauramo's diagram only one diagram has been co-ordinated with the absolute calendar, namely that by Donner (1957, p. 29) from the Kuopio district, eastern Finland. In Donner's diagram P_0 is dated 2 000 years

B. C., whereas T_0 is of the same age as in the Isokorpi basin, 2 600 years B. C. The spruce and the linden became common in the same order and at the same time both in southern Finland and in the Kuopio district. Thus the absolute age of pollen zone VII is approximately the same in both areas.

Knowing the absolute age of zone VII in Finland makes possible a comparison with Sweden. The generalized diagram from Norrland by E. Fromm has been supplied with absolute ages by Lundqvist (1957, p. 17). The year 2 600 B. C. in the diagram by Fromm is denoted by pollen of rare deciduous trees and *Alnus* together with pollen of *Betula* and *Pinus*. The rational border of *Picea* P_0 lies much higher up than in the Finnish diagram. Although a comparison of two areas rather far apart tends to be vague, the pollen flora embedded in peat of Norrland and southern Finland in the times around the year 2 600 B. C. resemble each other to such a degree that no shifting of vegetational zones can be distinguished.

SUMMARY

1. The Isokorpi bog at Tuusula, lying at an elevation of 52 m, was isolated from the Baltic Basin in the beginning of the Boreal period, at a time when the Ancylus stage of the Baltic Sea had begun. From the beach sand redeposited in the Isokorpi basin during this stage, two grains of *Ephedra* pollen have been found.

2. The *Carex*-bog filling the basin turned into a black alder *(Alnus glutinosa)* forest growing on the peat-bog. The black alder vegetation spread out over the former sandy beaches. The age of the subfossil roots of black alder, embedded in the beach sand, is 4 600 $\pm$ 115 years B. P., determined by the C^{14}-method. In pollen chronology this date falls within pollen zone VII in the Isokorpi diagram.

3. A comparison of diagrams from southern Finland, the Kuopio district and Norrland, Sweden, showed that at the end of the Atlantic period, or zone VII, the pollen floras in all the regions are rather similar to each other. No shifting of vegetational zones could be established.

4. The subfossil roots of *Alnus glutinosa* in the Isokorpi basin, Tuusula, do not provide evidence of a Baltic transgression in the locality. At the time black alders grew in the basin the shore line lay at a level some 25 meters below Isokorpi.

REFERENCES

AARIO, L. (1935) Die postglazialen Niveauverschiebungen im mittleren Uusimaa mit Berücksichtigung ihrer Beziehungen zu steinzeitlichen Wohnplätzen. Ann. Acad. Sci. Fennicae, Ser. A 44, 1.

AUER, VÄINÖ (1951) Peat lands. *In* Suomi, a general handbook on the geography of Finland. Fennia 72: 15.

BARENDSEN, G. W., DEEVEY, E. S. and GRALENSKI, L. J. (1957) Yale natural radiocarbon measurements III. Science 126, 908—919.

DONNER, JOAKIM (1954) Measurements of pollen of *Alnus glutinosa* and *A. incana.* C. R. Soc. Géol. Finlande XXVII: 5. Bull. Comm. géol. Finlande 166.

—»— (1957) The post-glacial shore-line displacement in the Kuopio district. Ann. Acad. Sci. Fennicae, Ser A III, 49.

—»— (1958) The late-glacial period and its correlation with the retreat of the ice in Finland. C. R. Soc. Géol. Finlande XXX: 5. Bull. Comm. géol. Finlande 180.

HEINONEN, LEO (1958) Studies on the microfossils in the tills of the North European glaciation. Ann. Acad. Sci. Fennicae, Ser. A III, 52.

HYYPPÄ. ESA (1937) Post-glacial changes of shore-line in South Finland. Bull. Comm. géol. Finlande 120.

—»— (1950) Helsingin ympäristö. Maaperäkartan selitys. Geologinen tutkimuslaitos. Helsinki.

KUJALA, V. (1924) Tervaleppä *(Alnus glutinosa L,* Gaertn.) Suomessa. Referat: Die Schwarzerle *(Alnus glutinosa L,* Gaertn.) in Finnland. Comm. Inst. Forest. Fenniae 7.

LUNDQVIST, G. (1957) C 14-analyser i svensk kvartärgeologi 1955—57. English summary. Sveriges Geol. Unders. Ser. C, 557.

RUDEBERG, G. (1925) Om nivåförändringarna i sydvästra Finland. Summary: On late-Quaternary changes of level in SW Finland. Geogr. Ann. 1925, H. 4.

SALMI, MARTTI (1959) Imatra stones in the glacial clay of Vuolenkoski. Bull. Comm. géol. Finlande 186.

SAURAMO, MATTI (1936) Der von der Ancylustransgression begrabene Erlenwald in Ruotsinkylä. Helsinki.

—»— (1954) Das Rätsel des Ancylussees. Geol. Rundschau 42, H. 2.

—»— (1958) Die Geschichte der Ostsee. Ann. Acad Sci. Fennicae, Ser. A III, 51.

TYNNI RISTO (1959) Über das Vorkommen von *Ephedra distachya L.* während des Spät- und Postglazials in Finnland. Arch. Soc. »Vanamo» 13: 2.

VIRKKALA, K. (1959) Maaperäkartan selitys. Lehti 2043, Kerava. Suomen geologinen yleiskartta 1: 100 000.

ЗНАМЕНСКАЯ, О. М. (1958) Стратиграфическое положение мгинских морских отложений (Znamenskaia, O. M.: The stratigraphic position of marine deposits in the valley of the Mga river).
Доклады акад. наук СССР 129, 2.

ÖSTLUND, GÖTE (1956) Tekniska data om metoden. *In* Åldersbestämning med hjälp av koi—14. Ymer 76.

10

TELESCOPING OF MINERAL FACIES IN GRANITES [1]

BY

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ABSTRACT

Higher- and lower-temperature minerals and textures may occur telescoped in rocks. In rocks formed at high temperatures but recrystallized at lower temperatures a low-temperature mineral assemblage may be found still showing a high-temperature fabric; though the mineral assemblage may be equilibrial it is in disequilibrium or disharmony with the texture: two or more mineral facies (representing stages of rock evolution) have been telescoped. In such rocks the low-temperature assemblage pseudomorphoses an older higher-temperature assemblage of which the original texture, when not obliterated during low-temperature recrystallization, may be the only relic left. This is the case in many albite granites. In granites all albite is secondary.

INTRODUCTION

Granites often show recrystallization effects resulting from adjustments to later lower-temperature stages in their history. Both the rock-building minerals and the types of textures of the more important rocks are relatively restricted in number. If a simple dichotomy is applied, dividing minerals and textures into higher-temperature (HT) and lower-temperature (LT) forms, HT and LT rocks may be distinguished. In originally HT rocks recrystallization may have led to the formation of LT minerals and textures, to the extent of obscuring and even destroying its HT characters. This may be schematically represented as follows:

	Texture	*Minerals*	
HT rocks	HT	HT LT	Increasing LT Recrystallization
	LT	LT	
LT rocks	LT	LT	

[1] Received March 20, 1960.

Severe LT recrystallization may entirely degrade an HT rock to an LT one. It takes much more recrystallization however to develop an LT texture in an HT rock than to form LT minerals in it while preserving the HT fabric. An LT texture does not form in HT rocks without concomitant recrystallization of HT to LT minerals (exception: cold cataclasis).

The HT or LT character of minerals may be expressed in internal structure, crystal morphology, and/or composition (except quartz all rock-building minerals have variable compositions since they are solid solutions). For instance, in a quartz bipyramid consisting of low quartz we have LT material in HT morphology. A sodium-rich monoclinic potash feldspar with a small optic axial angle in sanidine habit (HT) may have unmixed and inverted to perthitic microcline with a large optic axial angle (LT) retaining the HT crystal habit. In the plagioclase series anorthite is at the HT end but there is also an HT/LT division along the series.

The replacement of HT by LT minerals may be expressed in polymorphic changes and in chemical substitutions of varying magnitudes, grading from ion exchange in isomorphous series (as in the plagioclases) and the introduction of hydroxyl (as in the pyroxene-hornblende relationship) to larger compositional changes (e. g. biotite-chlorite). Higher temperatures of crystallization cause a larger extent of solid solution in many mineral series with limited isomorphous substitution, resulting in progressive unmixing during sufficiently slow cooling, and even expulsion of the guest phase, as in the series homogeneous potassium-sodium feldspar → perthite → expulsion of perthitic plagioclase.

Crystallization and recrystallization in rocks are governed by physical conditions and chemical composition, and the PTX conditions in the range of rock formation in the crust as expressed in the resulting mineral assemblages are classified in Eskola's Mineral Facies System. Rocks may be described as being composed of the minerals a, b, c, . . . (with the compositions p, q, r, . . .) with the spatial (crystallographic, textural, structural and geological) relations x, y, z, . . . Such descriptions are no end in themselves but offer the starting point in deducing the former states of the actual rocks. They contain the clues for explaining rock formation in terms of evolution. Here textures form a useful and necessary complement to the mineral facies concept. In the unravelling of the history of granites, especially albite granites, the study of textures is indispensable.

TELESCOPING OF MINERAL FACIES AND TEXTURES

The mineral facies concept is strictly based on the phase equilibria that may be reached in mineral assemblages but is often most useful in the study of disequilibrial rocks bearing the imprint of several facies. In such rocks internal equilibrium has either not been attained or else surpassed in their later evolution, i. e., two or more mineral facies are superposed in the same rock, or *telescoped*. In assemblages that have reached equilibrium the traces of their former history have been obliterated and their later history has failed to find expression. This history can only be deduced petrographically from non-equilibrial features such as relics and hysterogene products, as has been indicated by Eskola (1939, p. 341—343).

Even though certain sets of conditions or stability relations may only have been active for a short while they may have left their imprint on the rock in passing, as in the successive zones of zoned crystals.

Most common among the non-equilibrial assemblages is the case of rocks crystallized in a higher facies which have adjusted themselves to lower temperatures, sometimes under (directed) pressure, but in which the LT reactions have not proceeded to completion so that relics of the HT stage remain. These relics may be mineral and/or textural. The first stage is the formation of LT minerals in a rock that still retains an HT fabric. At later stages the HT texture itself may become supplanted by an LT fabric. This process may have been arrested in an intermediate stage in which the LT fabric has but partly replaced the HT one so that the two textures are telescoped in the rock. In granites it is common to find LT minerals with varying amounts of HT relics in an HT fabric on which LT textures have been superposed to varying extents. For instance, many granites contain eu- to subhedral hornblende and/or biotite and plagioclase with sub- to anhedral potash feldspar and anhedral quartz; the plagioclase may be zoned from andesine to albite while the potash feldspar may be microcline with monoclinic or less triclinic relics; late minerals, such as muscovite, albite, epidote-zoisite and later generations of quartz, may replace the older minerals to varying extents, in forms that grade from inter- and intragranular veinlike shapes (continuous or in stringers of discrete grains in contact with each other or separated) following crystal boundaries or cleavage directions and dislocations within older crystals to idioblastic crystals transecting the grain boundaries of the older minerals. Thus the older minerals (hornblende, biotite, plagioclase, monoclinic potash feldspar, quartz) form a hypidiomorphic framework on which have been superposed the blastic textures of the replacement minerals while late recrystallization and pseudomorphic replacement (such as decalcification of plagioclase, triclinization of potash feldspar, chloritization of biotite, etc.) alter the original minerals but leave the framework intact. While the HT minerals become changed by inversion, recrystallization and pseudomorphism the HT texture is altered by aggressive LT neocrystallization.

The telescoping of mineral facies and textures means that the rocks in question during their formation covered a PT interval greater than the individual ranges of the mineral facies and that their constituents responded through recrystallization to the changing conditions. This depends on the stability range of the component minerals and especially their sensitivity to subsequent changes under favourable physical conditions and in the presence of activating solutions.

Although we may not find signs of LT processes in HT rocks, such rocks have undoubtedly reached the LT stage during cooling. This is dependent on a variety of factors, such as rate of cooling (HT assemblages may be frozen), presence of residual water (a powerful promoter of LT recrystallization), of catalysts, etc.

A few examples may be given from mafic rocks in which the distinction between HT and LT minerals and textures is often easier to make than in granites and allied rocks.

Dolorites and basalts are HT rocks and may consist only of HT minerals (olivine, pyroxene, labradorite) in an HT consolidation fabric (ophitic, intergranular). In such rocks there has been no adjustment to the later stages of cooling once their minerals had separated, and no later metamorphism. They are in Eskola's Diabase Facies. Many dolerites and basalts however contain lower-temperature minerals, such as hornblende, biotite, epidote-zoisite, actinolite, albite, serpentine, chlorite and calcite. Generally these minerals stand in reaction relationships to the HT minerals. These alterations may occur in the course of cooling (autometamorphism) or be due to later (external) metamorphism, and represent LT facies. Eskola states: »Bei postmagmatisch umgewandelten Gesteinen findet man bisweilen zwei Mineralfazien in demselben Gestein vertreten, wie bei den Pyroxenspiliten. Diese kristallisierten zuerst etwa in der Diabasfazies mit Plagioklas und Augit, sind aber später in die Grünschieferfazies übergegangen, die Amphibolfazies überspringend. Der primäre Pyroxen blieb als instabiles Relikt erhalten, aber der Plagioklas wurde albitisiert, und Chlorit, Serpentin, Kalzit wurden neugebildet.» (Eskola, 1939, p. 343). Intermediate facies may be represented by hornblende, biotite, plagioclase between labradorite and albite.

If a basalt or dolerite, i. e., an HT rock with HT minerals in an HT fabric, has been spilitized, then the resulting spilite or albite diabase consists of LT minerals in an inherited HT texture. In intermediate stages there may be left-relics of the HT minerals, such as pyroxene. If a basalt, dolerite, spilite or albite diabase becomes metamorphosed at low temperatures to a greenschist, then the resulting completely recrystallized LT rock, e. g. an albite-epidote-actinolite-chlorite-sphene assemblage, contains LT minerals in an LT texture. When this process has not run to completion there are left relics of HT minerals and textures, and these may persist during subsequent processes of transformation. A spilite, for example, may be contact-metamorphosed to a blastophitic albite amphibolite.

In gabbros a similar sequence may be found. Certain gabbros and norites consist of olivine, pyroxene and calcic plagioclase in a pyrogenic fabric and were formed in Eskola's Gabbro Facies. Many gabbros however carry other lower-temperature minerals in addition, such as hornblende (Hornblende Gabbro Facies) and may show recrystallization effects owing to cooling and/ or metamorphism, e. g., saussuritization, uralitization, etc. (Epidote Amphibolite Facies, Greenschist Facies; these are facies named after metamorphic assemblages for their temperatures are below the level of magmatic consolidation). While most of these rocks consist of HT minerals in HT fabrics a certain amount of LT minerals have formed and several facies have been telescoped. These change may take place under various conditions.

From the mutual relationships of minerals formed in different facies the nature of the changes in crystallization conditions may be deduced. Pyroxene crystals rimmed and corroded by primary hornblende in gabbros or other igneous rocks indicate the transition from Gabbro to Hornblende Gabbro Facies but as the two minerals do not form a mix-crystal series it cannot be decided whether the change was gradual or abrupt. However, if the plagioclase crystals in the same rock are continuously zoned from, say, labradorite cores to oligoclase rims, then it may be concluded that the change was gradual. A sharp core-and-rim structure in the plagioclase on the other hand points to a break in crystallization conditions, such as might be due to eruption of a crystallizing magma into higher crustal levels causing relief of pressure and rapid consolidation.

Similarly, a hiatus is indicated by those porphyritic textures in which minerals that form phenocrysts recur in the groundmass in much smaller sizes, as in certain lavas and dyke rocks. The phenocrysts often consist of high-temperature minerals relative to those in the matrix.

Such breaks indicate abrupt facies changes and a high degree of telescoping. In many cases however the facies changes are gradual.

TELESCOPING IN GRANITES

LT minerals in an HT fabric define a mineral assemblage that pseudomorphoses an older higher-temperature assemblage. The fact that LT minerals may occur in an HT texture and have formed by recrystallization from HT minerals is of a certain importance in the study of granites where this combination is frequently met with and may lead to apparent contradictions in the interpretation of their mode of formation. Before concluding that a rock which consists of LT minerals crystallized at low temperatures it must be established that these minerals did not replace HT minerals, i. e. whether they are primary or secondary.

The minerals that make up granites are mostly rather sensitive to LT recrystallization, and HT minerals have generally given way to LT ones. The minerals plagioclase, potash feldspar and quartz have distinct HT and LT modifications (the HT—LT transition temperatures are generally different) but in granites they are virtually restricted to LT forms. The high quartz bipyramids of paragenetically early crystallization found in many granites always consist of low quartz owing to the ease with which the phase transformation takes place and only their shape is indicative of HT origin. The place of sanidine is taken by orthoclase and microcline. Plagioclase in high-temperature form (Köhler, 1941) as known from volcanites is absent in granites. Transitional optics however are shown by the potash feldspar and plagioclase from some high-level granites, owing to sufficiently rapid cooling.

Moreover, recrystallization textures caused by the aggressive growth of LT minerals in response to falling temperatures may tend to veil the original HT texture in granites. In the end stages of the consolidation of a hydrous granitic magma its residual portions become enriched in volatiles, resulting in a highly corrosive siliceous alkalic residual fluid which pervades the crystal aggregates formed, following intergranular pathways. Petrographically this is evident from the attitude of replacement minerals grown across or along primary grain boundaries, cleavage planes or dislocations. Such textures are blastic since they developed in the solid at the expense of existing crystals but in general they differ considerably from the common metamorphic fabrics. These textures are likewise different from the fabrics produced by the breakdown of HT minerals resulting in replacement or pseudomorphism by LT minerals, such as muscovite-zoisite development in plagioclase or chloritization of biotite. Secondary reaction textures are most pronounced in potassic granites (late albite, quartz, muscovite, etc.) and decrease in intensity and extent, within the granite (s. l.) series, towards the quartz diorites.

Thus the textural complexity of many granites, in contrast with their mineralogical simplicity, becomes understandable: the later phases in their crystallization history tend to obscure the earlier phases. It is rare to find granites with relics of HT minerals; they generally consist of LT minerals (of various temperature levels) in a hypidiomorphic HT fabric on which LT fabrics have been superposed to varying degrees.

The telescoping of HT and LT features has led some authors to assume LT origins for granites in general (i. e., by granitization in low mineral facies). Recognition of telescoping should help to clarify the complex crystallization sequences of many granites.

THE HT-LT TRANSITIONS IN GRANITE MINERALS

The HT—LT transitions are functions of temperature, pressure and composition and vary from mineral to mineral.

Thus the high-low quartz inversion, at atmospheric pressure taking place at 572° C., shifts to 815° C. at 10 000 bars (Yoder, 1950). The tridymite-high quartz transition point lies at 870° C. at atmospheric pressure. In granites the quartz bipyramids may be taken to have formed between about 900° and 600° C. (compare Schermerhorn 1956a, p. 99), most probably in the Gabbro Facies.

Quartz bipyramids in granite porphyries (and volcanic rocks) are often rounded and embayed by the groundmass. This shows that the early quartz became unstable later during rock crystallization. Hence quartz bipyramids in granite are commonly found only where spared from resorption by inclusion in slightly later minerals (often plagioclase) which afforded a protective covering. They are thus armoured relics. In rapidly cooled high-level granites, quartz bipyramids are more common than in deep-seated rocks. For example, Jacobson et al. (1958, p. 21) state that quartz bipyramids are frequent in many of the amphibole granites of the Younger Granite Ring Complexes in Northern Nigeria while universal in the closely associated granite porphyries. These granites are high-level instrusions; they may contain olivine and pyroxene which are likewise unstable HT relics incompletely converted into amphibole and biotite.

However, most of the quartz in granites is generally in anhedral grains representing one or more later generations so that its shape usually does not provide clues to the temperature of formation.

For plagioclase the transition temperatures between the HT and LT modifications are not yet well known (sodic plagioclase inverts to an LT form at about 700° C.) and HT plagioclase is generally unknown from granites. Tuttle and Keith (1954) described partially transformed HT oligoclase from a Tertiary granite from Skye.

As regards potash feldspar, Laves (1952) concluded that there exist two stable modifications of this feldspar, monoclinic sanidine stable at high temperatures and triclinic microcline stable at low temperatures, and suggested a transition temperature of 650—700° C. This value was later considered to be too high, and the temperature of transition may be as low as about 500° C. according to Goldsmith and Laves (1954 a). The same authors (1954b) believe that orthoclase is not a single phase but composed of extremely small triclinic units in contrast with truly monoclinic sanidine. It could be an unstable transition phase. Other authors (Tuttle and Bowen, 1958) oppose this view. Goldsmith and Laves (1954b) state that there is a strong tendency for potash feldspar to crystallize metastably as the monoclinic HT modification, which could be expected to take place readily in the stability field of microcline.

Tuttle and Keith (1954) measured moderate optic axial angles in the alkali feldspar in a Skye granite, with a variability which they think is due to partial transformation from sanidine optics to LT forms owing to a sufficiently rapid cooling of the magma. The fact that monoclinic potash feldspar with small to moderate optic axial angles (»anorthoclase») is occasionally found in granites indicates that potash feldspar may crystallize as an HT mineral in granites. Also, at higher temperatures potash feldspar takes up more plagioclase in solid solution (compare, Bowen and Tuttle, 1950; Tuttle and Bowen, 1958) so that the amount of soda, whether still dissolved or already unmixed as perthitic plagioclase, could give information as to crystallization temperatures, provided the perthite has not undergone any secondary modifications.

HT TEXTURES IN GRANITES

In view of the difficulties in evaluating a possible HT mode of formation of the LT quartz, plagioclase and potash feldspar commonly found in granites, the information provided by the textural aspects of the rocks must not be neglected.

The spatial relationships between the essential constituents are generally expressed in the usual pyrogenic hypidiomorphic texture of granites, namely euhedral to subhedral hornblende, biotite and plagioclase with subhedral potash feldspar and anhedral quartz (with minor early euhedral quartz). In porphyritic granites the potash feldspar megacrysts, though often showing euhedral accretion patterns expressed in zoning and other features, are anhedral towards included and adjacent older constitiuents. Other magmatic features are the preferential distribution of euhedral primary accessories in early main constituents (Schermerhorn, 1956a, c; 1958), the orientation of biotite and plagioclase inclusions in potash feldspar parallel to its growth planes (Frasl, 1954; Schermerhorn, 1956 a, b, c), primary flow textures, etc. These fabrics can only have developed in fluid, i. e., magmatic surroundings, and are therefore indicative of high temperatures, at least as high as the lower boundary of the Hornblende Gabbro Facies (which boundary lies below the high-low quartz inversion point).

THE FORMATION OF ALBITE GRANITES

Among the most controversial granites are the albite granites. It is as well to note here that by albite granites are meant those granites in which the separate plagioclase crystals consist of albite. Most granites contain albite, in perthite, myrmekite, albite rims, etc., but these secondary textural features are of minor importance compared with the first-order, essential textural units of granites, the large primary crystals of potash feldspar, plagioclase, quartz and the femics.

Albite granites, then, consist essentially of quartz, microcline and albite. The problem is whether the albite is p r i m a r y, which would entail a low-temperature, non-magmatic origin for the rock, or s e c o n d a r y, replacing earlier-formed, more calcic plagioclase. No undisputed magmatic albite is known while decalcification or albitization of originally more calcic plagioclase, i. e., soda (auto)metasomatism, is a well-known phenomenon, widespread in many kinds of different plagioclase-bearing rocks.

The quartz-microcline-albite assemblage is LT. The very presence of oligoclase in granites points to facies higher than the Albite-epidote Amphibolite Facies (oligoclase may and is of course often present in low facies as an unstable relic.) This assemblage in a granite, judged by itself, would therefore indicate a non-magmatic origin for the rock. On the other hand, many albite granites show the typical hypidiomorphic consolidation texture of magmatic granites; they are composed, femics apart, of tabular plagioclase (albite) crystals with potash feldspar and quartz in the interstices, corroding plagioclase, and they commonly also show other pyrogenic features such as mentioned in the preceding section.

The textural relations indicate therefore that the plagioclase separated relatively early, before the potash feldspar and quartz, as is generally the case in granites. We would thus have early crystallized LT albite. Here then is a paradox: while the fabric points to HT, magmatic crystallization, the mineral assemblage indicates an LT, non-magmatic origin. From the foregoing it is clear that the evidence is not really conflicting and that there is no contradiction in the mode of formation of such rocks: they consist of an LT assemblage pseudomorphosing and replacing an HT assemblage of which the texture is the only relic, and the albite is replacive. I. e., the primary plagioclase is now pseudomorphosed by secondary albite. Similarly, biotite may be replaced by chlorite pseudomorphs.

This is supported by textural and other evidence. In many granites it can be verified under the microscope, both in secondarily zoned plagioclase (with irregular albitic margins formed by later decalcification) and in serially studied samples, that albite replaces more calcic plagioclase, a process varying in its results from thin albite rims to complete albite pseudomorphs after primary plagioclase.

This process of albitization of plagioclase, best called d e c a l c i f i c a t i o n (Schermerhorn, 1956a, b, p. 345) since it tends towards the end stage of albite but may not have reached it, can often be followed step by step as it were. Some thin sections from a granite massif may show all or many primary plagioclase grains consisting of homogeneous albite, occasionally with vague cores of relict oligoclase, while other sections may display more calcic plagioclase with albitic margins. The habit, size, textural relations, etc., of the primary plagioclase remain the same but its anorthite content changes from place to place. The degree of decalcification may vary from crystal to crystal in the same specimen, and from outcrop to outcrop in the same pluton. This in itself constitutes conclusive evidence that the process is of post-magmatic origin. When run to completion however all primary plagioclase in a granite massif may consist of albite; in that case we have the texture of the rock to go by in deducing its previous character.

In normal zoned plagioclases the zoning shows an euhedral accretion pattern and corrosion outlines may cut across it, evidencing that it is primary growth zoning. Decalcification zoning however follows the outlines of the plagioclase crystals, even when irregular owing to corrosion (e. g. embayment by quartz or potash feldspar), which proves it to be of secondary origin. Moreover, it is generally confined to those parts of plagioclases that are in contact with adjacent potash feldspar crystals so that part at least of the replacing sodic material stems from perthitic exsolving of potash feldspar (the relationships between decalcification, perthitization and triclinization are discussed in another paper). Myrmekite may occur in decalcified patches or rims.

The process of decalcification may be autometamorphic, taking place at the end of, or after, magmatic consolidation in the presence of residual fluids, or it may be externally induced by later metamorphism.

Anorthite is unstable at low temperatures and plagioclase, more calcic than albite, is not stable under the conditions of post-magmatic (re)crystallization in granites. It may survive however as unstable relics. Decalcification is essentially an adjustment of plagioclase formed at higher temperatures to later lower temperatures, in the presence of Na and Si.

In a general way it can be asserted from the petrographical study of minerals and their textures in granites that all albite in granites is secondary, formed either by decalcification of initially more calcic primary plagioclase or, to a rather lesser extent, by replacement of potash feldspar (albite rarely replaces other minerals in granites). This is corroborated by the experimental feldspar diagram (Bowen and Tuttle, 1950, Tuttle and Bowen, 1958).

In an albite granite with igneous texture, then, the albite is non-magmatic but it cannot be concluded that the granite itself is non-magmatic for it was earlier a rock with plagioclase more calcic than albite and such plagioclase is known to separate from magmas.

In a general way albite granites stand in the same relationship to granites as albite diabases to dolerites or spilites to basalts; they are altered LT facies of the HT rocks.

The evidence of telescoping in pure albite granites is now shown only by the occurrence of LT minerals in a persisting HT fabric.

There is one class of albite granites however in which the texture is granoblastic so that from microscopic evidence alone (unless there are relict igneous textures such as the various types of preferential and oriented distributions of minerals in other minerals referred to in the foregoing) nothing can be said about its primary or secondary origin. Such granites have been recast by metamorphism; they may, or may not, have been igneous rocks. It would seem better to call such rocks not granites but granoblastic albite gneisses.

REFERENCES

BOWEN, N. L., and TUTTLE, O. F. (1950) The system $NaAlSi.O_8$—$KAlSi_3O_8$—H_2O. J. Geology 58, p. 489—511.

ESKOLA, P. (1939) *in* BARTH, T. F. W., CORRENS, C., ESKOLA, P., Die Entstehung der Gesteine. Ein Lehrbuch der Petrogenese. Julius Springer, Berlin, p. 263—407.

FRASL, G. (1954) Anzeichen schmelzflüssigen und hochtemperierten Wachstums an den groszen Kalifeldspaten einiger Porphyrgranite, Porphyrgranitgneise und Augengneise Österreichs. Jb. Geol. Bundesanstalt 97, p. 71—132.

GOLDSMITH, J. R., and LAVES, F. (1954 a) The microcline-sanidine stability relations. Geo. et Cosmo. Acta 5, p. 1—19.

—»— and — (1954 b) Potassium feldspars structurally intermediate between microcline and sanidine. Geo. et Cosmo. Acta 6, p. 100—118.

JACOBSON, R. R. E., MACLEOD, W. N., and BLACK, R. (1958) Ring-complexes in the Younger Granite Province of Northern Nigeria. Geol. Soc. London, Mem. 1, p. 1—72.

KÖHLER, A. (1941) Die Abhängigkeit der Plagioklasoptik vom vorabgegangenen Wärmeverhalten. Min. Petr. Mitt., 53, 24—49.

LAVES, F. (1952) Phase relations of the alkali feldspars. J. Geology 60, p. 436—450 and p. 549—574.

SCHERMERHORN, L. J. G. (1956 a) Petrogenesis of a porphyritic granite east of Oporto (Portugal). Min. Petr. Mitt. 6, p. 73—115.

—»— (1956 b) The granites of Trancoso (Portugal): a study in microclinization. Am. J. Sci. 254, p. 329—348.

—»— (1956 c) Igneous, metamorphic and ore geology of the Castro Daire-São Pedro do Sul-Sátão region (Northern Portugal). Com. Serv. Geol. Portugal 37, p. 1—617.

—»— (1958) The paragenesis of accessory minerals in igneous rocks. Econ. Geol. 53, p. 215—218.

TUTTLE, O. F., and BOWEN, N. L. (1958) Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8$—$KAlSi_3O_8$—H_2O. Geol. Soc. America. Mem. 74, p. 1—153.

—»— and KEITH, M. L. (1954) The granite problem: evidence from the quartz and feldspar of a Tertiary granite. Geol. Mag. 91, p. 61—72.

YODER, H. S. (1950) High-low quartz inversion up to 10,000 bars. Trans. Amer. Geophys. Union 31, p. 827—835.

11

POLLUCITE FROM LUOLAMÄKI, SOMERO, FINLAND [1]

BY

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ABSTRACT

Pollucite has been found in a small lithium pegmatite for the first time in Finland. The mineral is described and its chemistry discussed. Petalite from the same locality is described and the distribution of the alkali metals in the pegmatites considered. The alteration of the minerals indicates that the crystallization took place at a temperature of around 450° C.

CONTENTS

[1] Received March 3, 1960.

INTRODUCTION

In summer of 1958 the Ore Department the of Geological Survey of Finland remapped and investigated the old pegmatite occurrences in the region of Tammela and Somero in Southern Finland. In this connection, one of us (A. Vesasalo) found a quartz-like but somewhat strange looking mineral from the Luolamäki petalite pegmatite previously described by Toini Mikkola and H.B. Wiik (1948). Later this mineral was identified as pollucite. Thus the first known Finnish petalite occurrence also became the first known occurrence of pollucite.

The Luolamäki pegmatite is situated in the commune of Somero, 500 meters NNW from the Pelto-Koivula farm in the village of Pajula. Geologically the area belongs to the Svecofennian range, having been mapped and described by Ahti Simonen (1956). The immediate vicinity of the

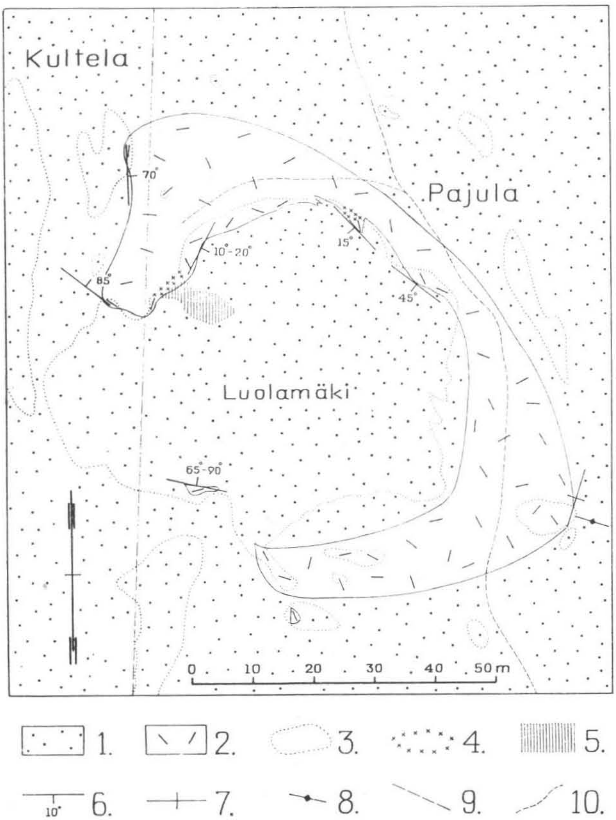

Fig. 1. The geological map of the Luolamäki pegmatite occurrence. 1, diorite. 2, pegmatite. 3, outcrop. 4, quarried. 5, pit. 6, strike and dip of contact. 7, vertical contact. 8, foliation. 9, village boundary. 10, path.

pegmatite is built of a quartz diorite complex cut by pegmatite and aplite veins. Plagioclase An_{40}, hornblende, and biotite are the main constituents of the diorite. The grain size is about 0.5—1.0 mm. Apatite, epidote, titanite, zircon, and oxide ore are met with as accessories.

THE MODE OF OCCURRENCE

The Luolamäki pegmatite is observed in places on the gentle slopes of a small hill. It can be concluded that the pegmatite forms a flat cake or dike dipping about 10° southward (Fig. 1). The pollucite and the petalite are found in an ancient quarry cave located in the northwest corner of Luolamäki. The thickness of the pegmatite varies considerably. It measures about three meters in the cave and is less than two meters in the NW corner, where the pegmatite has been excavated.

The contact against the upper diorite is sharp. Only some minor offsets have been observed. The lower contact is more irregular and often steep. The grain size of the pegmatite is generally moderate. Usually the individual crystals do not exceed 5 cm. At both quarries, however, the grains grow larger and some zoning is observable.

Against the upper contact a fine-grained aplite rock 2—3 cm thick is met with. It is composed of quartz in majority and potassium feldspar. The aplite is followed by a 30—50 cm thick band of quartz, microcline, and black tourmaline. The crystals measure about 10 cm in diameter. Below this zone the rock becomes still coarser, the individual crystals measuring up to 30 cm. The main minerals are graphic microcline perthite, tourmaline, and quartz. Muscovite and albite also occur but in minor quantities and in a somewhat smaller grain size. In association with the albite some crystals of columbite and turbid beryl are found.

The quarry excavated in the core part of the pegmatite, evidently more than fifty years ago, is about 12 meters long, 5 meters broad and 2 meters high. Judging from the old dump around the cave, the innermost part of the pegmatite was composed of quartz mixed with petalite and pollucite. The largest crystals are as large as 1.5 meter in diameter and the main constituents are quartz, petalite, pollucite, and spodumene. In places these minerals are cut by medium-grained albite-muscovite rock with some tourmaline.

POLLUCITE

OCCURRENCE AND ASSOCIATION

Pollucite occurs in Luolamäki as two large crystals (Fig. 2 and Fig. 1, Plate I). The abundance of the pollucite in the dump indicates that the

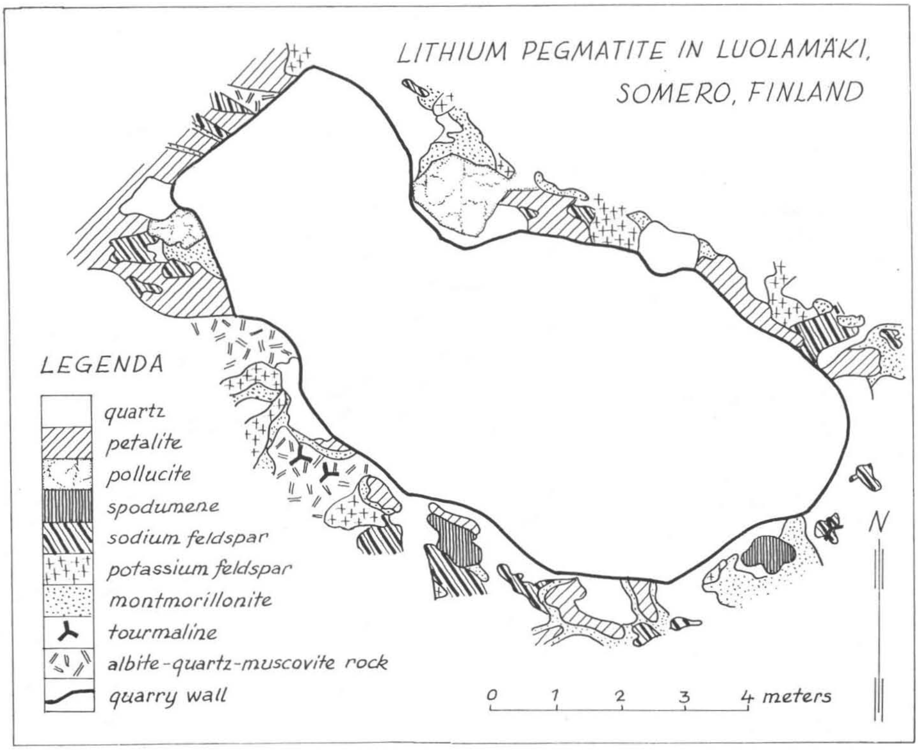

Fig. 2. Sketch of the excavation.

mineral was present in the excavated part of the pegmatite. No crystal face has ever been observed on the mineral. Usually the pollucite from Luolamäki is dim with a faint pink or green luster. Some almost water-clear pieces have been found though.

The mineral is generally intersected by numerous veins of varying composition. Most abundant are cryptocrystalline muscovite veins very much like those described by Quensel (1938; 1957) from the Varuträsk pollucite. These veins are less than one millimeter in breadth. Spodumene also occurs as very thin veinlets in the same way as it does in Varuträsk (Fig. 2, Plate I). A symplektitic mixture in these veins is observed to consist of spodumene and uncertain eucryptite, as indicated by the X-ray powder photographs taken. Some few and narrow veins of petalite have also been noted in the pollucite of Luolamäki.

Apatite inclusions are abundant. They are seen in any thin section as prisms almost one centimeter long and also as tiny microscopic crystals. Spectroscopical investigation of this apatite reveals no rare elements to be present.

Against petalite the contacts of the mineral are sharp and clear. Generally, however, the mineral is surrounded by a 1—2 cm broad seam of a white or light green cryptocrystalline mixture of kaolinite and montmorillonite. This band is in places much thicker and evidently originates as an alteration product from the pollucite.

Microcline occurs in pollucite as inclusion-like nodules and as narrow veins. The same types of veins and nodules are observed in the altered part of the mineral too.

CHEMICAL COMPOSITION AND PHYSICAL PROPERTIES

Carefully selected water-clear pollucite material was chemically analysed by Dr. H.B. Wiik and the alkalies were determined by Mr. Pentti Ojanperä, M.A., using a Beckman-type flame photometer. The composition is given in Table 1.

Table 1. Chemical composition and physical properties of pollucite, Luolamäki, Somero, Finland. Analysts H. B. Wiik and Pentti Ojanperä (alkalis), from the Geological Survey of Finland.

	wt %	Cation prop.	Physical properties
SiO_2	45.60	7 589	
TiO_2	0.003		
Al_2O_3	17.60	3 452	
Fe_2O_3	0.01	1	Sp. gr. 2.832 ± 0.004
FeO	—		
MnO	0.03	4	N_C = 1.5142
MgO	0.00		
CaO	0.00		N_D = 1.5172 ± 0.0005 20° C
Na_2O	1.65	532	
K_2O	0.01	2	N = 1.5223
Li_2O	0.009	6	
Rb_2O	0.73	78	Unit Cell: Ia3d
Cs_2O	32.43	2 301	a_0 = 13.688 ± .003 Å
P_2O_5	0.41	58	
H_2O+	1.91	2 120	
H_2O-	0.00		
	100.392		

Power diffraction data of the chemically analysed material were obtained by running the sample with a Phillips diffractometer and employing silicon (a_0 = 5.4305 Å) as an internal standard. The data are given in Table 2.

Table 2. X-ray powder data, Pollucite, Somero, Finland. Ni filtered Cu-radiation. Values corrected with internal Si-standard ($a_0 = 5 \cdot 4305$ Å).

Ø	sin Ø	d	I	hkl	N
7.96	.1385	5.56	2	211	6
9.20	.1599	4.82	1	220	8
12.21	.2115	3.65	36	321	14
13.05	.2258	3.417	100	400	16
14.62	.25241	3.053	1	420	20
15.34	.26455	2.911	45	332	22
16.03	.27614	2.792	1	422	24
16.71	.28753	2.681	3	431	26
17.94	.30802	2.503	4	521	30
18.59	.31879	2.418	25	440	32
20.31	.34710	2.221	12	532	38
20.86	.35609	2.165	1	620	40
22.46	.38204	2.018	4	631	46
22.96	.39009	1.976	3	444	48
23.96	.40562	1.899	2	640	52
24.42	.41342	1.863	12	552	54
26.30	.44307	1.738	15	732	62
26.77	.45041	1.710	5	800	64
27.20	.45710	1.685	2	741	66
27.64	.46376	1.660	1	644	68
28.51	.47731	1.614	1	822	72
28.94	.48389	1.592	1	831	74
29.80	.49697	1.550	3	752	78
30.23	.50347	1.530	5	840	80
31.46	.52190	1.4758	2	921	86
32.24	.53347	1.4435	1	930	90
33.06	.54552	1.4120	5	932	94
33.43	.55092	1.3981	1	844	96
34.62	.56813	1.3558	1	1011	100
36.16	.59004	1.3054	5	765	110
36.95	.60112	1.2814	1	855	114
37.31	.60613	1.2708	1	864	116
37.68	.61125	1.2601	1	1033	118
39.16	.63149	1.2197	4	1051	126
39.56	.63689	1.2094	1	880	128
40.66	.65157	1.1822	4	1053	134
41.40	.66131	1.1647	1	875	138
42.47	.67520	1.1408	1	884	144
43.58	.68937	1.1173	3	1055	150
45.01	.70723	1.0891	1	1073	158

Weighed average for $a_0 = 13.688 \pm 0.003$ Å

The reflections observed are in good agreement with the data published by Fleischer and Ksanda (1940) for pollucite. The reflections observed fit well with the extinctions of the space group Ia3d, which has been accepted for pollucite by Strunz (1936) and Naray-Szabo (1938). The content of the unit cell calculated from the powder data listed and based on the analysis given is shown in Table 3.

Table 3. The content of the unit cell of pollucite, Luolamäki, Somero, Finland.

$a_0 = 13.688$ Å; $V_0 = a^3{}_0 = 2.56460 \cdot 10^{-21}$ cm³
$D^{4^0}_{20^0} = 2.832 \pm 0.004$
Weight of the unit cell $= V_0 \cdot D = 7.26295 \cdot 10^{-21}$ gr

	wt %	atoms per unit cell	
SiO_2	45.60	Si	33.20
Al_2O_3	17.60	Al	15.10
FeO + MnO	0.04	Fe	0.03
$Na_2O + K_2O$	1.66	Na	2.33
$Rb_2O + Li_2O$	0.74	Rb	0.35
Cs_2O	32.43	Cs	10.07
H_2O	1.91	H_2O	4.65
(P_2O_5)	0.41	P	0.25
		O	96.08

THE GENERAL FORMULA OF POLLUCITE

The Pollucite from Luolamäki is strictly isotropic under the microscope The powder photographs do not indicate any departure from the cubic symmetry either. Professor Fritz Laves from Zürich was kind enough to take some photographs of the mineral with a high resolution camera, but no additional reflections were observed. All the reflections and systematic extinctions found fit well with the space group Ia3d proposed for pollucite by Strunz (1936). According to Naray-Szabo (1938) pollucite is only pseudo-cubic and belongs actually to the tetragonal space group I 4/acd. The proposal is based upon the fact that 16 Al and 32 Si atoms occupy a common 48-fold place (g) in the assumed structure of pollucite. As pointed out by Naray-Szabo himself, the symmetry of the mineral depends on the type of arrangement of these atoms. If they are ordered in specific sites the structure is tetragonal, but it is cubic if the atoms are statistically in random position. All the pollucites described in the literature are isotropic. Consequently, the disordered cubic structure has to be assumed to be valid for the natural mineral. Chemically, the random distribution of aluminium and silicium in the structure must mean that the Al to Si ratio is not strictly fixed but allows for a considerable variance in the composition of the mineral. Thus a variations is seen in the chemically analysed pollucites. Many types of general formula have been proposed to demonstrate this variation. The formula given by Richmont and Gonyer (1938) $Cs_{14+x}Al_{14}(Al_xSi)_{34}O_{96} \cdot 4-9H_2O$ allows a slight variation in the Al to Si ratio but does not consider the Na-content of the mineral. In this sense the general formula proposed by Nell (1944)

$Cs_{16-x}Na_{x-y}Al_{16-y}Si_{32+y}O_{96} \cdot xH_2O$ is better. The assumed end-members in this formula are:

Pollucite end-member $Cs_{16}\ Al_{16}\ Si_{32}\ O_{96}$
Analcite end-member $Na_{16}\ Al_{16}\ Si_{32}\ O_{96} \cdot 16\ H_2O$
Excess SiO_2 framework $Si_{16}\ Si_{32}\ O_{96}$

Investigation of the natural and synthetic analcites performed by Saha (1959) points out that the amount of water in the mineral varies with the content of Si. The ratio H_2O to Si tends to equal 0.5. In pure analcite, in which Si = 32 and H_2O = 16, all the sites usually assumed to be occupied by water are used up. Consequently, when the number of atoms exceeds 32 and the number of H_2O-molecules exceeds 16, additional places must be found for water. As pointed out by Saha (1959), the only sites available are those occupied by Na in the analcite, although they are somewhat too small for the purpose. If we assume that such a possibility exists, the composition of the natural pollucite can be expressed by means of the hypothetical end-members given in Table 4.

Table 4 Hypothetical end-members in the natural pollucite mineral.

	Type and number of the positions			
	(b) 16	(c) 24	(g) 48	(h) 96
Pollucite end-member	16 Cs	—	16 Al + 32 Si	96 0
Analcite end-member	16 H_2O	16 Na	16 Al + 32 Si	96 0
Excess of SiO_2	16 H_2O	16 H_2O	48 Si	96 0

These end-members give rise to the following general formula for the mineral:

$Cs_{16-x-y}Na_x\ Al_{16-y}\ Si_{32+y}\ O_{96} \cdot (2y+x)\ H_2O$, in which x and y denote analcite and the excess SiO_2, respectively. This formula differs from the formula proposed by Nel (1944) only in respect to the slightly higher water content.

Coombs (1954) has proposed as many as three different modifications for analcite. The theoretical reasoning of Beattie (1954) shows that there is a possibility for an order — disorder inversion in analcite. The isomorphism between analcite and pollucite exists evidently in the disordered structure. This would mean that at lower temperatures an exolution takes place. In Luolamäki, analcite has been observed on several occasions. Spectrographically, no cesium has been detected in this mineral. As will become clear in a later connection, pollucite and analcite grow simultaneously or little later than petalite. This analcite can not be the cubic disordered high-temperature form of the mineral, since it shows clear birefringence, although the X-ray powder photograph shows a very slight broadening of some of the higher reflections only. The co-existence of pollucite and this analcite indicates that a clearly defined immiscibility gap occurs between these minerals.

The pollucite from Luolamäki and from Varuträsk contains small amounts of phosphorus, as seen from the analyses given. In addition, at least the mineral from Luolamäki always includes small grains of apatite about 1—5 mm in diameter. It is quite possible that the pollucite originally contained phosphorus and other elements now in apatite in the same manner as was observed in viséite by McConnel (1952).

PETALITE

The petalite of Luolamäki has been described in detail by Mikkola and Wiik (1948). The properties of the mineral are checked and given in Table 5.

Table 5. Petalite, Luolamäki, Somero, Finland. Chemical composition according to Mikkola and Wiik (1948). Analyst H. B. Wiik. Alkalis rechecked by Pentti Ojanperä.

	wt %	Physical properties
SiO_2	77.47	
Al_2O_3	17.12	
Fe_2O_3	0.24	Sp. gr. 2.412 (Mikkola and Wiik)
MnO	tr.	$\alpha = 1.5047$ ⎫
MgO	0.00	$\beta = 1.5098$ ⎬ $\pm$ 0.0005 at 20° C
CaO	0.13	$\gamma = 1.5158$ ⎭
Li_2O[1])	4.35	$2V\gamma = 83°$
Na_2O	0.37	Polysynthetically twinned,
K_2O	0.13	(001) as twinning plane
Rb_2O	0.01	Unit cell:
P_2O_5	tr.	$a_0 = 7.61$ Å
F	—	$b_0 = 5.13$ Å
H_2O+	0.25	$c_0 = 11.74$ Å Pa or P2/a
H_2O-	0.03	$\beta = 122.4°$
	100.10	

[1]) The original values for the alkalis are Li_2O 3 · 95, Na_2O 0.78, K_2O 0.30.

Based on the values from Table 5, the unit cell content of the petalite is calculated and given in Table 6.

Table 6. Unit cell content of the petalite from Luolamäki, Somero, Finland.

Vo = 423.90 Å³ $D_{4°}^{20°} = 2.412$

Weight of the cell = Vo · D = 1.022.447 · 10^{-21} gr

	wt %	atoms per unit cell		
SiO_2	77.47	Si	7.94 atoms	
$Al_2O_3 + Fe_2O_3$	17.12	Al	2.07	
CaO	0.13	Ca	0.1	2.05
Li_2O	4.35	Li	1.79	
Na_2O	0.35	Na	0.07	
$K_2O + Rb_2O$	0.14	K	0.01	
$H_2O +$	0.25	H_2O	0.17	
		O	19.92	

Petalite occurs in the cave as very large crystals, in many cases water-clear. The outer margins of the crystals are always milky white, owing to initial alteration. The same feature can be seen along the cracks and more open cleavage fractures, too. The mineral is also frequently surrounded by an alteration zone about 3 cm broad.

OTHER CONSTITUENTS

SPODUMENE

Spodumene is met with as an even-grained, sugar-white intergrowth with quartz. The optical properties of the mineral are given in Table 7.

Table 7. Optical properties of spodumene from Luolamäki, Somero, Finland

$\alpha = 1.655$	$2V\gamma = 70^\circ$
$\beta = 1.663$	$c \wedge \gamma = 32^\circ$
$\gamma = 1\ 670$	

The grain size of the mixture may vary considerably but the quartz-to-spodumene ratio remains about constant all over the pegmatite. In places clevelandite albite occurs with spodumene and quartz. No alteration zone similar to that around petalite and pollucite is observed in connection with spodumene.

POTASSIUM FELDSPAR

Potassium feldspar, rich as microcline perthite in the outer zones of the pegmatite, is rarely found in the core part of the occurrence.

Potassium feldspar occurs in the core as narrow veins cutting through the pollucite and as small nodules met with in pollucite and in the alteration zones around the pollucite (Fig. 1, Plate II). These nodules are usually small, about 1—4 mm in diameter and spherical. It is seen under the microscope that the grain is a single crystal of microcline, commonly quite rich in alteration products, mainly kaolinite. All the grains are heavily corroded on the outer surface. 2V of the mineral is about 90° and $\gamma = 1.525$ $\alpha = 1.518$.

In addition to these round microcline crystals, larger aggregations of potassium feldspar are also found in the alteration zone around pollucite and petalite. Optically the mineral in these aggregates is similar to the small spheres.

To study the distribution of the alkali metals in the pegmatite, in which such minerals as pollucite and petalite occur, a suite of alkali bearing minerals was investigated. Mr. Pentti Ojanperä, M.A., from the Geological Survey of Finland, performed the alkali determinations utilizing Beckman's recording flame photometer. The results are listed in Table 8.

Table 8. Alkali content of some minerals from the Luolamäki pegmatite, Somero, Finland.

	Weight per cent				
	Li_2O	Na_2O	K_2O	Rb_2O	Cs_2O [1]
Pollucite	0.009	1.65	0.01	0.73	32.43
Petalite	4.35	0.37	0.13	0.01	—
Coarse muscovite	0.41	0.64	9.26	1.45	—
Coarse microcline perthite	0.004	1.92	12.22	1.51	—
Large microcline aggregate in the montmorillonite zone	0.055	0.92	13.02	0.77	—
Small microcline spheres in the montmorillonite zone	0.01	0.85	13.23	2.98	—
Clevelandite	0.003	11.16	0.06	0.00	—
Beryl	0.56	1.07	0.16	0.01	—

[1]) Cesium content was found to be low in all the samples except pollucite and was not determined with the flame photometer.

The amount of rubidium found in small microcline nodules within the alteration zone of pollucite is unusually high. The only feldspar known to the authors to contain still more rubidium is the microcline perthite from Varuträsk, which has 3.3 wt % Rb_2O (Adamson, 1942).

ALBITE

Albite occurs in the core part of the pegmatite, forming cutting veins together with quartz and muscovite. Most of the tourmaline found in the pegmatite appears in connection with these clevelandite veins, too.

The mineral consists of sugar-white or water-clear lathes typical of the variety generally called clevelandite. It was found to be very poor i alkali metals other than sodium (Table 8) and a very low content of Ca was also noted. The specific gravity of the mineral was measured to be 2.59 ± 0.01, $\gamma = 1.539$ $\beta = 1.532$ and $\alpha = 1.528$ $2V\gamma = 72°$.

It is assumed that the albite-quartz-muscovite rock is one of the latest among the primary constituents of the pegmatite. It cuts through the petalite and pollucite crystals but is definitely older than the secondary alteration products of these minerals.

Black t o u r m a l i n e, d = 3.185 ± 0.005, and, rarely, c o l u m b i t e, d = 6.91 ± 0.01, are met with in connection with the albite-quartz-muscovite rock. B e r y l with d = 2.698 is observed only outside the cave proper.

LAUMONTITE AND ANALCITE

Laumontite is met with in minor amounts as a fine-grained powder on the surface of clevelandite. The following optical properties are measured for the mineral:

$\alpha = 1.513, \ \beta = 1.519, \ \gamma = 1.525$

$2V\alpha = 48°, c \wedge \gamma = 42°$

Analcite occurs in the petalite and on the cleavage plane of this mineral. It is evidently formed as a replacement of petalite at a stage before the main low temperature alteration reactions. The mineral shows a weak birefringence with $\omega = 1.488$, $\varepsilon = 1.484$. The x-ray powder photographs taken show a slight broadening of the higher reflections, indicating a departure from the cubic symmetry.

ALTERATION OF POLLUCITE AND PETALITE

As mentioned earlier, the pollucite and petalite in Luolamäki are surrounded by a zone of alteration products. In general these alteration products are similar to those described by Quensel (1937, 1938, and 1945) from Varuträsk. In places the textural features of the parent mineral are still recognizable (Fig. 2, Plate II).

Identification of the components of the mixture meets with difficulties. It is noted, however, that a montmorillonite-type mineral with an expansive lattice was one of the main constituents. The composition of this phase must vary considerably since the index of refraction changes from $\alpha = 1.518$ to $\alpha = 1.540$. The color of the mineral varies also from pure white to light green. Kaolinite, or a mineral of the kaoline group is also frequently present in the mixture ($\alpha = 1.540$, $\beta = 1.555$). According to the x-ray powder photographs taken, serpentine like chlorite is met with in the mixture. Chlorite occurs also as mixed lattice layers in the montmorillonite.

The alteration seam around the petalite is commonly rich in quartz. The zone surrounding the pollucite does not contain quartz except in small quantities.

Barrer and White (1951) crystallized petalite and Barrer and McCallum (1953) grew pollucite from gels under hydrothermal autoclave conditions. They did not, however, observe any formation of montmorillonite or kaoline at the water pressures and temperatures applied. It is considered, though, that the alterations met with in Luolamäki are caused by the action of hydrothermal solutions. The reactions may be approximated as follows:

$$\underset{\text{pollucite}}{2CsAlSi_2O_6} + 5H_2O \rightarrow \underset{\text{montmorillonite}}{Al_2(OH)_2Si_4O_{10} \cdot 4H_2O} + \underset{\text{in solution}}{Cs_2O}$$

or $$\underset{\text{petalite}}{2LiAlSi_4O_{10}} + 5H_2O \rightarrow \underset{\text{montmorillonite}}{Al_2(OH)_2Si_4O_{10} \cdot 4H_2O} + \underset{\text{quartz}}{2SiO_2} + \underset{\text{in solution}}{Li_2O}$$

The formation of quartz during the replacement of petalite explains the abundance of the mineral in the alteration zone around petalite. The presence of kaolinite in the seams is explained by a further dehydration of montmorillonite. The co-existence of montmorillonite and kaolinite fixes the temperature of the alteration rather precisely. According to Roy and Osborn (1954), aluminium montmorillonite is obtained in the Al_2O_3 — SiO_2 — H_2O system at temperatures from 200° to 420° under varying H_2O pressures, and kaolinite was synthesized between 150° and 405°. The upper limit for the formation of Mg-montmorillonite in the system MgO — Al_2O_3 — SiO_2 — H_2O is, according to Roy and Roy (1955), as high as 480° C.

No reaction seam occurs in connection with spodumene. This must indicate that it was stable under the conditions which caused alteration in petalite and pollucite. This is to be understood since natural spodumene is considered to be a low temperature form among the lithium alumino silicates (Roy, Roy, and Osborne, 1950). The presence of strange ions necessary for spodumene formation might, according to Isaacs and Roy (1958), also control the stability of the mineral. The invariable mixture of quartz in spodumene in Luolamäki gives rise to the view that the present mixture was originally petalite that later changed over to a spodumene-quartz rock.

CONCLUDING REMARKS

The pegmatite of Luolamäki is unique among the large number of pegmatite bodies met with in Tammela and Somero. The only one known to contain lithium minerals in considerable quantities in the region is the recently found petalite pegmatite in Hirvikallio, Tammela, described by Vesasalo (1959).

No explanation for the concentration of Li and Cs in Luolamäki can be given. It seems evident, however, that an unusual concentration of volatiles took place during the later stage of crystallization. Consequently, the crystallization of pollucite and petalite become possible but the high pressure in the hydrothermal solutions readily caused alterations of the minerals to hydrated aluminium silicates with a very low content of lithium and caesium.

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REFERENCES

ADAMSON, O. J. (1942) The feldspar group. Minerals of the Varuträsk Pegmatite XXXI. Geol. Fören. i Stokholm Förh. 64, p. 19.

BARRER, R. M. and McCALLUM, N. (1953) Hydrothermal chemistry of Silicates. Part IV. Rubidium and caesium aluminosilicates. J. Chem. Soc. 821, pp. 4029—4035.

BARRER, R. M. and WHITE, E. A. D. (1951) The hydrothermal chemistry of silicates. Part I. Synthetic lithium aluminosilicates. J. Chem. Soc. 283, pp. 1267—1278.

BEATTIE, J. R. (1954) The structure of analcite and ion-exchanged forms of analcite. Acta Cryst., Vol. 7, pp. 357—359.

COOMBS, D. S. (1955) X-ray observations on wairakite and non-cubic analcite. Mineralog. Mag., Vol. 30, pp. 699—708.

FLEISCHER, MICHAEL and KSANDA, C. J. (1940) Dehydration of pollucite. Am. Mineralogist, Vol. 25, pp. 666—672.

ISAACS, THELMA and ROY, RUSTUM (1958) The α—β inversion in eucryptite and spodumene. Geoch. et Cosmoch. Acta, Vol. 15, pp. 213—217.

McCONNEL, DUNCAN (1952) Viséite, a zeolite with the analcime structure and containig linked SiO_4, PO_4, and H_xO_4 groups. Am. Mineralogist, Vol. 37, pp. 609—617.

MIKKOLA, TOINI and WIIK, H. B. (1947) Petalite, a mineral new to Finland. Soc. géol. Finlande 20, pp. 282—286. C. R. Bull. Comm. géol. Finlande 140;

NARAY-SJABA,v. ST. (1938) Die Structure des Pollucites $CsAlSiO_6$ x H_2O. Zeitschr. f. Krist., Vol. 99, pp. 277—282.

NEL, H. J. (1944) Pollucite from Karibib, Southwest Africa. Am. Mineralogist, Vol. 29, pp. 443—452.

QUENSEL, P. (1937) Minerals of the Varuträsk Pegmatite IV. Petalite and its Alteration Product, Montmorillonite. Geol. Fören. i Stockholm Förh. 59, pp. 150—156.

—»— (1938) Minerals of the Varuträsk Pegmatite XIII. Pollucite, its Vein Material and Alteration Products. Geol. Fören. i Stockholm Förh. 60, pp. 612—634.

—»— (1945) Minerals of the Varuträsk Pegmatite XXXVI. Further alteration products of pollucite. Geol. Fören. i Stockholm Förh. 67, pp. 549—554.

—»— (1957) The paragenesis of the Varuträsk pegmatite. Arkiv för Mineralogi och geologi, Vol. 2, pp. 9—125.

RICHMONT, W. E. and GONYER, F. A. (1938) On pollucite. Am. Mineralogist, Vol. 23, pp. 783—789.

ROY, DELLA M. and ROY, RUSTUM (1955) Synthesis and stability of minerals in the system MgO—Al_2O_3—SiO_2—H_2O. Am. Mineralogist., Vol. 40, pp. 147—178.

ROY, RUSTUM and OSBORN, E. F. (1954) The system Al_2O_3—SiO_2—H_2O. Am. Mineralogist., Vol. 39, pp. 853—885.

ROY, RUSTUM, ROY, DELLA M. and OSBORN, E. F. (1950) Compositional and stability relationships among the lithium alumino-silicates eucryptite, spodumene and petalite. J. Amer. Ceram. Soc., vo. 33, pp.

SAHA, PRASENJIT (1959) Geochemical and X-ray investigation of natural and synthetic analcites. Am. Mineralogist, Vol. 44. 300—313.

SIMONEN, AHTI (1956) Explanation to the map of rocks. Sheet—2024 — Somero. Geological map of Finland. Helsinki.

STRUNZ, HUGO (1936) Die chemische Zusammensetzung von Pollucite. Zeitschr. f. Krist., Vol. 95, pp. 1/8.

VESASALO, ARVO (1959) On the petalite occurences of Tammela, SW-Finland. C. R. Soc. géol. Finlande 31. Bull. Comm. géol. Finlande 184; pp. 59—74.

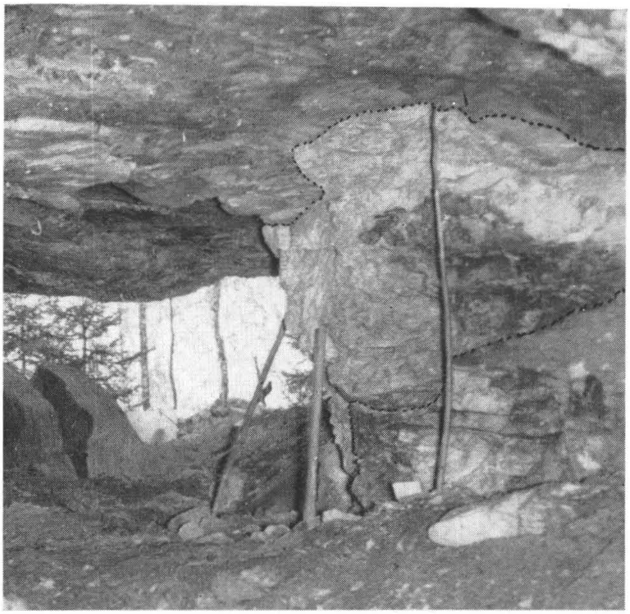

Fig. 1. Pollucite crystal on the cave wall. Photo A. Vesasalo

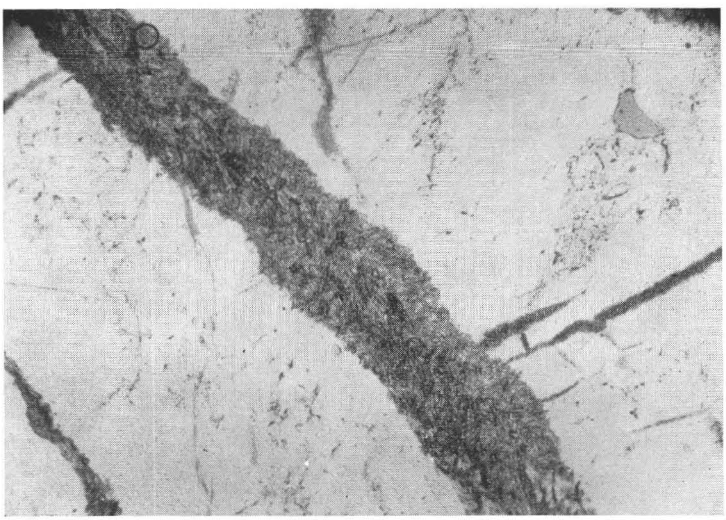

Fig. 2. Spodumene vein in pollucite. Nicols II. Photo Erkki Halme. Magn. 40 x.

K. J. Neuvonen and *Arvo Vesasalo:* Pollucite from Luolamäki, Somero, Finland

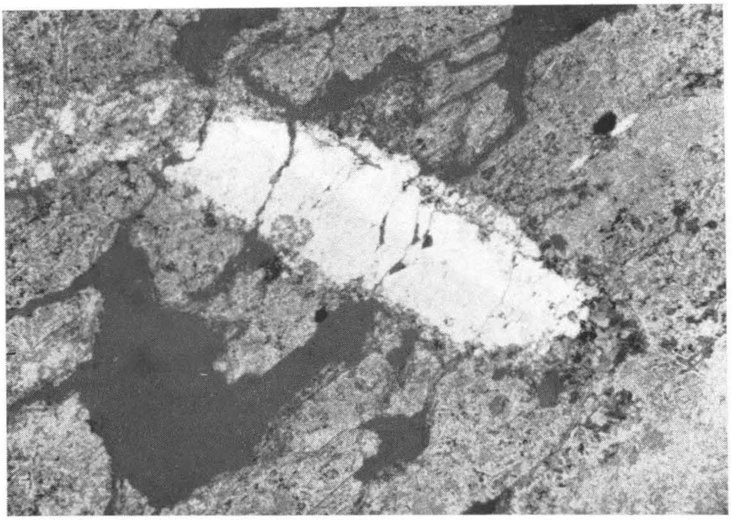

Fig. 1. Microcline nodule in montmorillonite. Nicols +. Photo Erkki Halme. Magn. 25 x.

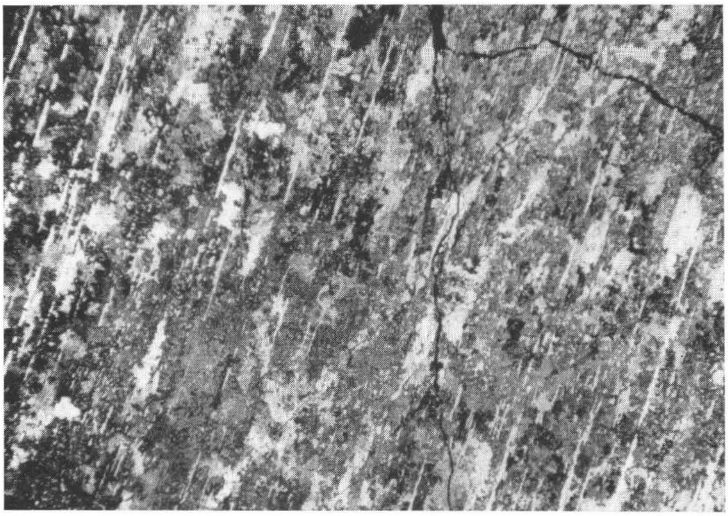

Fig. 2. Altered petalite. Nicols +. Photo Erkki E. Halme. Magn. 10 x.

K. J. Neuvonen and *Arvo Vesasalo:* Pollucite from Luolamäki, Somero, Finland

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OSTSEESTADIUM WÄHREND DER ALLERÖDZEIT IN ASKOLA, OST-UUSIMAA, (SÜDFINNLAND)[1]

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AUSZUG

Der Artikel befasst sich mit der auf der Pollendatierung von anorganischen Sedimenten basierenden spätglazialen Stratigraphie von Moorbecken. Die in Askola festgestellten Sedimentwechsel in den Grundablagerungen der Moore werden Uferbeobachtungen gegenübergestellt. Den Resultaten gemäss ist der Wasserstand des marinen Ostsee-Stadiums in der Allerödzeit in Askola verhältnismässig niedrig, nämlich etwa 56 m gewesen.

EINLEITUNG

Die postglaziale Uferverschiebung im südlich von den Höhenrücken Salpausselkä gelegenen Kirchspiel Askola ist schon früher untersucht worden. Das Litorina I-Ufer liegt bei Monninkylä in Höhe von etwa 30.5 m (Hyyppä 1937 a, Virkkala 1953). Über boreal- und präborealzeitliche Entwicklungstadien der Ostsee haben u.a. Sauramo (1954; 1958) und der Verfasser (1956) geschrieben. Luho (1956) hat die die Askola angetroffenen, auf die Präborealzeit zurückgehenden, steinzeitlichen Siedlungen der den Untersuchungsergebnissen von Sauramo entsprechenden Uferverschiebungskurve gegenübergestellt. Alle diese Vorgänge sind postglazial, und bei ihrer Pollendatierung standen ausser Tonsedimenten auch organische Ablagerungen zur Verfügung.

In Askola sind auch ältere anorganische Sedimente als die obengenannten untersucht worden (Tynni, 1956 op.c., Mölder, Valovirta und Virkkala, 1957). Die Pollenzusammensetzung solcher Sedimente können durch Pollen unterschiedlichen Alters kompliziert werden, was den Gebrauch der Pollen-

[1] Eingegangen am 31. März 1960.

diagramme zur Datierung und Aufklärung der Uferverschiebungen erschwert. Es sollen nun Vermutungen über den Anteil der in den fraglichen Sedimenten möglicherweise vorkommenden primären und sekundären Pollenbestände sowie das Pollendiagramm der Schichtenfolge eines im südlichen Askola, in Nietoonkylä, gelegenen Moors »Nietoonsuo» und daran anschliessende Beobachtungen über die Uferverschiebung dargelegt werden. Für diesen anorganischen Schichtenfolgenanteil sind in dem Gebiet zahlreiche Vergleichsobjekte gefunden worden.

HERKUNFT DES POLLENBESTANDS

Bei der Pollendatierung von spätglazialen anorganischen Sedimenten positiv wirkende Faktoren:

a) Das Untersuchungsgebiet ist Grundgebirgebereich, wo interglaziale Ablagerungen selten sind oder im allgemeinen fehlen. In Südfinnland sind derartige Sedimente bis jetzt nocht nicht untersucht worden. Aus den Moränen ist in solchen Verhältnissen vermutlich nicht so viel Interglazialpollen in den Ton geschwemmt worden wie im Bereich der jungen Sedimentgesteine, wo Interglazialablagerungen in grosser Zahl angetroffen werden. Ignatius (1953) hat das entsprechende Problem in Ostkanada behandelt.

b) Die prozentualen Verhältnisse der Baumpollen u.a. in der Präborealzeit sind in den organischen und anorganischen Ablagerungen weitgehend ähnlich. Man darf annehmen, dass ihre Verhältnisse auch in den älteren Sedimenten im grossen und ganzen von der Zusammensetzung des Sediments unabhängig sind.

c) Die Floratypen der Krautpollen sind nicht an die Sedimentfazies gebunden. Offenbar ist in dem präboreal abgesetzten Ton die Zusammensetzung der Krautpollen eine andere als in dem darunterliegenden älteren Tonsediment.

d) In den vertikalen Tonprofilen werden klar ausgeprägte, die Vegetationsfloren widerspiegelnde Pollen angetroffen. Die Wald- und Steppenvegetationsfloren treten in gleicher Reihenfolge auf wie die spätglazialen Klimaveränderungen.

e) Der fremde, nicht zu den Floratypen passende Pollen kann grossenteils durch Fernflug angefallen sein, statt dass der interglazial wäre. Fernflugpollen hat sich auch im Inlandeis ansammeln können, von wo er mit dem Schmelzwasser sich in den benachbarten Sedimenten ausgebreitet hat.

f) Die den spätglazialen Sedimenten anzutreffenden *Ephedra distachya*-Pollen (Tynni, 1959; Salmi, 1959) dürften als Leitfossilien der Steppe gelten können.

Auf die Pollendatierung von spätglazialen anorganischen Sedimenten negativ einwirkende Faktoren:

a) Dem Rückzug des Eisrandes hat sukzessiv eine lokale Tundrazone folgen können (vgl. Hyyppä, 1933; Donner, 1958), die von den klimatischen Verhältnissen unabhängig war.

b) Der grosse *Alnus*- und *Betula*-Anteil der spätglazialen Sedimente kann z.T. durch interglaziale oder interstadiale, bisher noch unbekannte Ablagerungen (vgl. Iversen, 1936; Hyyppä, 1937 b) und Moränen bedingt sein, in deren Unterschichten nach Heinonen (1957) Laubbaumpollen vorherrschen.

Die Maxima der spätglazialen *Betula* und Krautpollen in den Sedimenten sind wahrscheinlich grösstenteils primär und stammen aus der Vegetation der südfinnischen Verlandungen. Die Salpausselkä-Rücken bildeten u.a. in der Jüngeren Dryaszeit ein Archipel, wo ausser offener Vegetation auch schütterer Birkenwald wuchs (Donner 1951). Die reichen *Artemisia*-Vorkommen weisen auf Fernflug und Inselvegetation der näheren Gegenden hin (vgl. Erdtman, 1949).

LAGERFOLGE DES MOORS »NIETOONSUO».

Der untere Abschnitt des beigefügten Pollendiagramms, (Abb. 1), von 1.5 m Tiefe an abwärts, umfasst Ton-, Sand-, Bänderton- und Feinsandablagerungen mit gut abgegrenzten Fazies. Im unteren Ton ist die Bänderung nur im Mittelteil zu sehen, wo die Warven etwa 3 mm dick sind. Nach der Sedimentationskurve von Ignatius (1958) wird die Warven des vor einem normal zurückweichenden Inlandeis abgesetzten Tons mit wachsendem Abstand vom Eis sukzessiv nach oben immer dünner. Dann kann im oberen Abschnitt der Schichtenfolge eine Mikrostruktur möglich sein. Aus bohrungstechnischen Gründen ist diese nicht immer wahrzunehmen, und möglicherweise kommen im oberen Abschnitt des Tonsediments vom »Nietoonsuo» auf eine Längeneinheit mehr Warven als im mittleren. Die scharfen Faziesgrenzen weisen darauf hin, dass die Lagerfolge Lücken hat.

In dem Pollendiagramm, Spalte A, sind die Anteile der Baum- und Strauchpollen angegeben. Im oberen Abschnitt der anorganischen Schichtenfolge herrscht ein ähnliches Betula-Maximum wie weiter oben im Schlamm der Präborealzeit. Der Ton unterhalb der Sandschicht weist im oberen Teil ein *Betula*-Maximum auf, in 195 cm Tiefe aber ein schwaches *Pinus*-Maximum, von wo abwärts der *Betula*-Anteil wieder steigt. *Alnus* erreicht einen verhältnismässig hohen Wert (17 %) in 185 cm Tiefe, und *Picea* tritt durchweg in geringer Menge auf. Die hohen *Alnus*- und *Corylus*-Mengen lassen eine sekundäre (interglaziale oder interstadiale) Herkunft oder Fernflug vermuten, desgleichen die *Carpinus*-Vorkommen, die nach Heinonen verhältnismässig häufig in den finnischen Moränen anzutreffen sind.

Die Spalten D und E zeigen Anreicherung von *Graminae*- und *Ranunculaceae*-Pollen im oberen homogenen Ton. Erst weiter unten bildet der Krautpollen eines zwischen Grob- und Feinsandschihten liegenden Tonlagers ein kräftiges Maximum. Es enthält reiche *Artemisia*-Vorkommen, die nur für diese Tonablagerung charakteristisch sind. Ebenso ist der gemeinsame NBP-Anteil stellenweise mehr vorherrschend als der BP-Anteil (Spalte F). Der Pollenbestand, der ein *Ephedra distachya*-Vorkommen einschliesst, zeigt

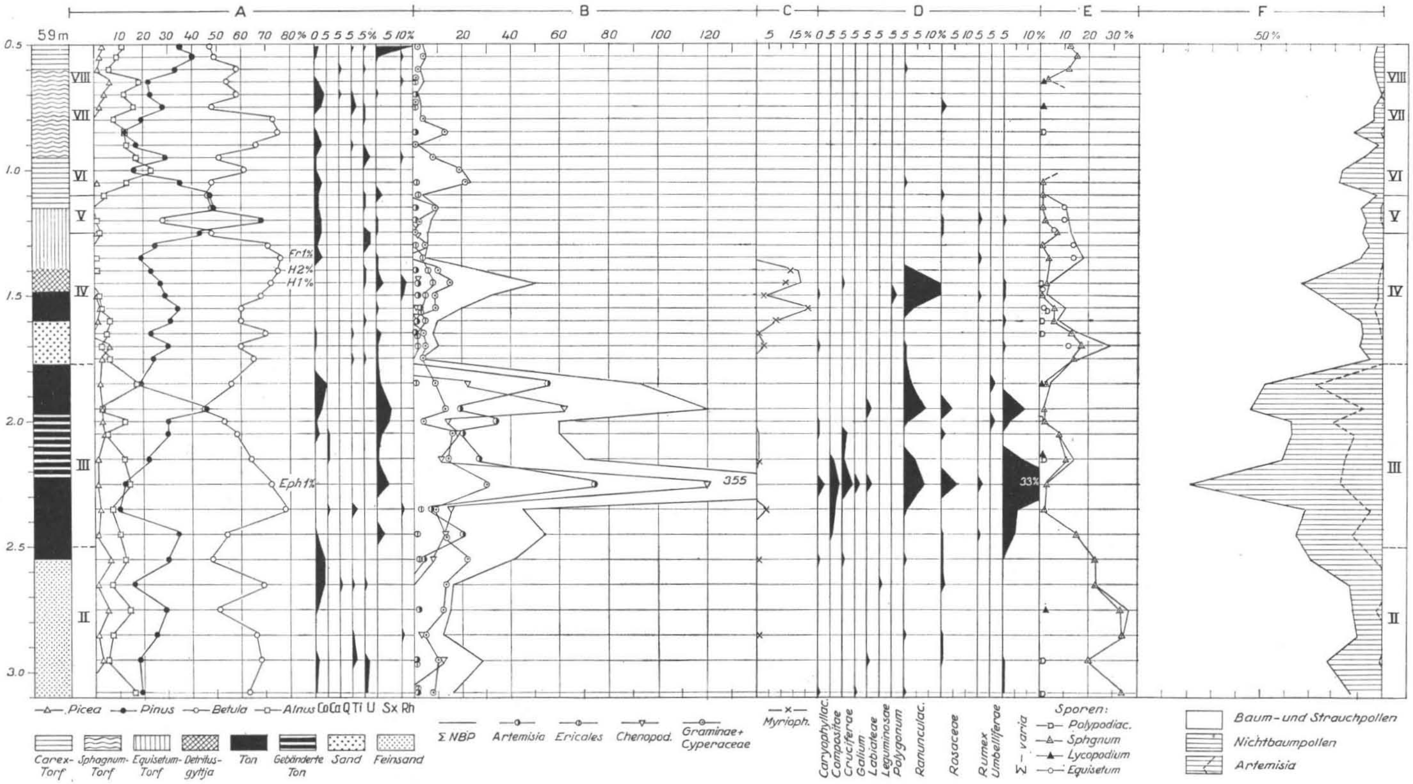

Abb. 1. Pollendiagramm des Moors »Nietoonsuo»

also Steppenvegetation an. Der Pollenfloratyp des Tonsediments erinnert an die Flora der Älteren und Jüngeren Dryaszeit. Der oberhalb vom Sand liegende Ton ist präboreal. Unterhalb von dieser Ablagerung wären Sedimente der Jüngeren Dryas zu erwarten, und wenn in dem Sandlager keine lange, Jüngere Dryas- und Allerödzeit einschliessende Lücke enthalten ist so entspricht dieses von Krautpollen charakterisierte Tonlager der Jüngeren Dryaszeit. Die Bedeutung des Krautpollenbestands für die Datierung wird besonders durch den ziemlich allgemein in Europa angetroffenen Pollenbestand entsprechender Art aus der Jüngeren Dryaszeit erhöht.

In der unter dem Ton liegenden Feinsandschicht weicht der Pollenbestandtyps von dem vorigen gänzlich ab, und die Pollendichte ist im Feinsand viel grösser als im Bänderton. Wahrscheinlich sind die Pollenverhältnisse im Feinsand dadurch bedingt, dass in der Zeit, wo der Pollen sich ansammelte, ein die Waldvegetation begünstigendes Klima herrschte. Ein solches Stadium war nach den in Dänemark und Südschweden durchgeführten Untersuchungen die Allerödzeit. Wenn eine Allerödablagerung in Frage steht, ist die Pollenanreicherung entweder auf Fernflug, auf das Anwachsen der sekundären Komponente, was unwahrscheinlich ist, oder auf die Ausbreitung der Wälder in den naheliegenden Gebieten zurückzuführen. Im ersteren Falle hatten sich die Nadelbaumpollen in der Feinsandschicht angereichert, aber dass darin ein *Betula*- Maximum anstelle von *Pinus* auftritt, zeigt, dass die Dichte des Pollenbestands nicht durch Fernflug bedingt ist. Das relativ günstige Klima im Alleröd hat die Bewaldung der Verlandungen in den benachbarten Gebieten gefördert. Die unten dargelegten Gesichtspunkte über das zur Zeit der Entstehung der Feinsandschicht herrschende, relativ niedrige Ostsee-Stadium stehen mit dieser Ansicht im Einklang.

In dem kleinen, im ausgewaschenen Moränengebiet gelegenen Becken des Moors »Nietoonsuo» hat in der Zeit, wo auf der Moräne der gut sortierte Feinsand abgelagert wurde, die Höhe des Wasserspiegels die Schwellenhöhe des Sedimentbeckens, 57.5 m, nicht wesentlich überschritten, weil im Sediment die tonigen Bestandteile fehlen. In der Schichtenfolge fehlen die breitgebänderten Tone, deren Zustandekommen die Nähe eines Eisrandes voraussetzt. In der Nähe vom Inlandeis hätten die Schmelzwasserströmungen infolge grösserer Stromgeschwindigkeit gröbere Sedimente auch in tieferem Wasser absetzen können. Auf ein niedriges Ostsee-Stadium weist auch der Umstand hin, dass in erheblich höheren Becken keine Feinsandablagerungen aus der gleichen Zeit gefunden worden sind. Im Nahgebiet findet sich auch keine über 62.5 m hohe Erhebung, von welcher der Sand herstammen könnte. In der genannten Höhe liegt jedoch ein gut ausgebildetes steiniger Uferwall, und der Sand kann infolge einer Auswaschung desselben abgesetzt worden sein. Dies hätte eine Wassertiefe von 5 m vorausgesetzt. Den in der Nähe befindlichen Lagerfolgen gemäss reichte die allerödzeitliche Litoralzone auch bis zu 56 m Höhe. Weitere Untersuchungen dürften exaktere Werte von der Höhe des Wasserspiegels in der Allerödzeit ergeben.

Nach den von Dr. K. Mölder kontrollierten Kieselalgenuntersuchungen beträgt der Anteil der im Feinsand angetroffenen Salzwasser-Diatomeen 1—2 %, und umfasst die Arten *Cocconeis scutellum*, *Coscinodiscus*, fragm. und *Thalassionema nitzschioides*. Obwohl der Anteil der halinen Formen so gering ist, ist er bezeichnend für die Sedimentationsverhältnisse des spätglazialen Meerstadiums in einem geschützten Becken. Die Kieselalgenflora der spätglazialen Sedimente umfasst grösstenteils Süsswasserformen. U.a. der Diatomeenfloratyp, den Mölder (1956) in den Bändertonen bei Jokela angetroffen hat, kommt grösstenteils da und dort in Finland in den kleinen Gewässern vor.

Nach zahlreichen russischen Forschern sowie auch nach Hyyppä (1943), Mölder (1944) und Sauramo (1954) bestand in der Spätglazialzeit eine Meerverbindung zwischen dem Weissen Meer und dem Ostseebecken.

Das Litoralsediment des niedrigen Meerstadiums in der Allerödzeit ist im »Nietoonsuo» in der Jüngeren Dryas von Bänderton überdeckt worden. Im Anfangsstadium hat die Sedimentation im Meer stattgefunden, wie auch aus dem Diatomeendiagramm des nahen Moors »Pikkusuo» ersichtlich ist, wo die bis zu 16 % ausmachenden Salzwasser-Kieselalgen die Arten *Amphora giganthes*, *Campylodiscus echeneis*, *Coscinodiscus*, fragm. und *Nitzschia punctata* umfassen. Dass die Korngrösse des Sediments feiner wird, weist auf eine beträchtlichere Transgression nach der Allerödzeit hin.

Später in der Jüngeren Dryaszeit verschwinden die halinen Kieselalgenformen aus den meisten untersuchten Schichtenfolgen des Gebiets, was ein Zeichen für den Beginn des Baltischen Eissee-Stadiums ist. In den seltenen Fällen, wo in den Ablagerungen des oberen Teils der Jüngeren Dryaszeit haline Diatomeen angetroffen worden sind, handelt es sich wahrscheinlich um Sekundarität infolge von Durchmischung der Schichten beim Bohren. Dies hat sich u.a. bei Kolbenbohrerkontrollen ergeben.

Die einheitliche Tonsedimentation der Jüngeren Dryaszeit ist in Askola im allgemeinen gegen Ende dieser Zeit abgeschlossen. In vielen untersuchten Becken, so wie auch im Becken des »Nietoonsuo» ist schlecht sortierter, pflanzliche Zellen enthaltender Sand abgelagert. In der hierauf folgenden Sedimentation von homogenem Ton, insbesondere in dem schon früher veröffentlichten Diagramm des Moors »Vajakkaneva» (Tynni, 1956), gibt der beträchtliche Anteil der Salzwasserdiatomeen einen Hinweis auf den Anbruch des eigentlichen Yoldiameerstadiums. In der entsprechenden Ablagerung des »Nietoonsuo» sind die Salzwasserkieselalgen sehr selten, und möglicherweise hat die Sedimentation auch in einem flacheren Yoldia-meer-Stadium stattgefunden als im »Vajakkaneva». Es möge erwähnt werden, dass das Diagramm des Moors »Vajakkaneva» auch abweichend von den Resultaten des Verfassers datiert worden ist (Mölder, Valovirta und Virkkala, 1957), nach erneuter Kontrolle bleibt der Verfasser aber bei seiner ursprünglichen Datierung.

Der Abfluss des Baltischen Eissees in der Jüngeren Dryaszeit hat wahrscheinlich die nach dem Zonenwechsel III/IV entstandenen, schlecht sortierten Sandschichten hervorgebracht. Sie sind in Askola in Moorschichtenfolgen von 40—68 m angetroffen worden, aber die bedeutendsten liegen im Dorf Nietoo in 62.5 m und im Dorf Nalkkila in 65 m Höhe, in den gleichen Höhenlagen wie die dort angetroffenen Ufer. Dieses geneigte Höhenniveau dürfte dem Wassenspiegel des Yoldiameeres nach dem Abfliessen des Baltischen Eissees entsprechen. Im Dorf Nietoo trifft die Höhe des Yoldiameeres nach dem Abfluss mit der Höhenzone des allerödzeitlichen Meerstadiums zusammen.

Nach Sauramo (1933; 1958) machte das plötzliche Absinken des Baltischen Eissees, BIII, am Ausgang des Dryaszeit auf das Niveau des Yoldiameeres 27—28 m aus. Ufer des Baltischen Eissees aus der Zeit vor seinem Ablauf sind in Askola wegen der einschränkenden Höhenverhältnisse keine angetroffen worden, aber auf höheren Isobasen, in Halkia in Pornainen befindet sich in 97 m Höhe ein mächtiger Steinwall, und in Numminen in Mäntsälä findet man in etwa 100 m Höhe dem gleichen Stadium entsprechende Wälle. Die Bestimmung dieser Wälle hat Dr Virkkala freundlicherweise bestätigt. In Pornainen kommen auch andere Wälle u.a. 4 und 9 m tiefer vor. Als Gradient dieser Ufer ist 1.4 m/km ermittelt worden, welcher Wert noch weiterer Kontroll bedarf.

Nach Sauramo (1958) beträgt die Höhe des BIII am I Salpausselkä bei den Städten Lahti und Hyvinkää 140 m und der Ufergradient 1.5 m/km. Der zur Uferisobasis des Salpausselkä senkrechte Abstand von den Isobasis des Litorina I 31 m in Askola beträgt etwa 34 km, und wenn der Gradientwert 1.5 m/km wäre, müsste der dem BIII entsprechende Wert in Askola 89 m sein, also nicht weit von dem ein Absinken von 27-28 m voraussetzenden Niveau. Der in dem Untersuchungsgebiet festgestellte Gradientwert von 1.4 m/km stimmt überein mit den Uferbeobachtungen aus Pornainen und dem auf der Isobasis des Litorina I 31 m in Askola vor dem Abfluss anzunehmenden Niveau von 92—93 m.

Von dem niederen Stand der Allerödzeit, im »Nietoonsuo» 56 m und auf der Isobasis von L I 31 m etwa 65 m, hat das Ostsee-Stadium der Jüngeren Dryaszeit sich dann auf 92.5 m gehoben. Die Pollenflora in der oberen Partie der Tonsedimente aus der Jüngeren Dryas weist in der Regel eine Zunahme der *Pinus*-Pollen auf, was vermutlich ein Zeichen für ein

Transgressionmaximum ist, wo sich die schwimmfähigen *Pinus*-Pollen im Stadium des offenen Baltischen Eissees anreicherten.

Mit den Höhenschwankungen des Ozeans in der Allerödzeit und den diese begrenzenden Zeitperioden in Gebieten, wo die Erdkruste trotz der Vergletscherung ruhig geblieben ist, befassen sich die umfangreichen Untersuchungen von Auer in Feuerland und Patagonien, wo ausser den Pollendatierungen noch die bei Vulkanausbrüchen aufgeschütteten Aschenschichten sowie die Radiokohlen-Datierungen Anhaltspunkte für die Stratigraphie bieten. Nach Auer (1959) war der Meeresspiegel in der Allerödzeit niedriger als in der Älteren und Jüngeren Dryaszeit. Vom Transgressionsmaximum der Älteren Dryaszeit, etwa 18 m, war der Wasserspiegel zunächst auf etwa 12 m gesunken. Am Ausgang der Allerödzeit begann das in etwa 9 m Höhe liegende Meeresufer erneut sich zu heben und erreichte in der Jüngeren Dryaszeit sein Maximum in etwa 14 m Höhe, welchen Hochstand dann auch keine jüngere Transgression mehr erreicht hat. Zwischen diesen beträchtlicheren Transgressionen trat noch eine schwächere Transgression auf. Vergleicht man dieses Resultat mit den Verhältnissen in Askola, so trifft man von der Allerödzeit an Schwankungen des Wasserstandes, in der gleichen Reihenfolge, aber in verschiedenem Ausmass, welcher letztere Umstand auf die örtlichen Bewegungen der Erdrinde zurückzuführen ist.

LITERATURVERZEICHNIS

AUER, V. (1959) The Pleistocene of Fuego-Patagonia. Part III: Shoreline Displacements. Ann. Acad. Sci. Fennicae, Ser. A, III, 60.

DONNER, J. (1951) Pollen-analytical studies of Late-Glacial deposits in Finland. C. R. Soc. géol. Finlande 24; Bull. Comm. géol. Finlande 154.

—»— (1958) The Late-Glacial period and its correlation with the retreat stages of the ice in Finland. C. R. Soc. géol. Finlande 30; Bull. Comm. géol. Finlande 180.

ERDTMAN, G. (1949) Palynological aspects of the pioneer phase in the immigration of the Swedish flora. II. Identification of pollen grains in Late Glacial samples Mt. Omberg, Ostrogothia. Svensk Botan. Tid. 43.

HEINONEN, L. (1957) Studies of the microfossils in the tills of the North European glaciation. Ann. Acad. Sci. Fennicae Ser. A, III, 52.

HYYPPÄ, E. (1933) Das Klima und die Wälder der spätglazialen Zeit im Bereich der karelischen Landenge. Acta Forest. Fennica 39.

—»— (1937a) Post-Glacial changes of shore-line in south Finland. Bull. Comm. géol. Finlande 120.

—»— (1937b) Bemerkungen über G. Branders Aufsatz »Ein Interglazialfund bei Rouhiala in Südostfinnland», und zwei neue Tonfunde auf der karelischen Landenge. C. R. Soc. géol. Finlande 10; Bull. Comm. géol. Finlande 119.

—»— (1943) Itämeren historia uusimpien Itä-Karjalassa suoritettujen tutkimusten valossa. Vortragsreferat: Östersjöns historia i ljuset av de nyaste forskningarna i Öst-Karelen. Terra 55, H. 3/4.

IGNATIUS, H. (1953) Late-Glacial History in North-Central Quebec, Ontario, Canada. Palynology Symposium. New Haven.

—»— (1958) On the rate of sedimentation in the Baltic Sea. C. R. Soc. géol. Finlande 30; Bull. Comm. géol. Finlande 180.

IVERSEN, J. (1936) Sekundäres Pollen als Fehlerquelle. Danmarks Geol. Undersøgelse IV Rk. 2, 15.

LUHO, V. (1956) Die Askola-Kultur. Die frühmesolithische Steinzeit in Finnland. SMYA 57.

MÖLDER, K. (1944) Das Karelische Eismeer im Lichte der fossilen Diatomeenfunde. C. R. Soc. géol. Finlande 16; Bull. Comm. géol. Finlande 132.

—»— (1956) Die Diatomeenflora der Bändertone bei Jokela in Südfinnland. Arch. Soc. Vanamo 10: 1.

MÖLDER, K., VALOVIRTA, V. und VIRKKALA, K. (1957) Über spätglazialzeit und frühe Postglazialzeit in Südfinnland. Bull. Comm. géol. Finlande 178.

SALMI, M. (1959) Imatra stones in the glacial clay of Vuolenkoski. Bull. Comm. géol. Finlande 186.

SAURAMO, M. (1933) Yoldiameri entisten ja nykyisten tutkimusten valossa. Terra 45: 1.

—»— (1954) Das Rätsel des Ancylussees. Geol. Rundschau 42.

—»— (1958) Die Geschichte der Ostsee. Ann. Acad. Sci. Fennicae Ser. A, III, 51.

TYNNI, R. (1956) Rannansiirtymisistä Askolassa. Referat: Über postglazial Uferverschiebungen in Askola, Ost-Uusimaa (Südfinnland). Terra 2.

—»— (1959) Über das Vorkommen von Ephedra distachya L. während des Spät- und Postglazials in Finnland. Arch. Soc. Vanamo 13: 2.

VIRKKALA, K. (1953) Altitude of the Littorina limit in Askola, southern Finland, C. R. Soc. géol. Finlande 26; Bull. Comm. géol. Finlande 159.

13

ON THE STRIATIONS AND GLACIER MOVEMENTS IN THE TAMPERE REGION, SOUTHERN FINLAND [1]

BY

K. VIRKKALA

ABSTRACT

This article deals with glacier movements in the Tampere region, mainly on the basis of striations and the till fabric. Glacier movements of three different ages can be observed in the area. The oldest were westerly. The strongest, central movement was one from the northwest toward the southeast. The youngest movements, which were strongly controlled by the topography, took place first from the west, and afterwards from the sector in between the north-northwest and the east-northeast on the northern side of the Tampere esker, as well as between the west-southwest and south-southwest on the southern side of the esker.

CONTENTS

INTRODUCTION

By the Tampere region is meant in the present connection sheet 2 123 of the general map published by the Surveyor-General's Office.

Glacial striations were investigated in the Tampere area as early as last century, when the first mapping of superficial deposits was carried

[1] Received April 26, 1960.

out in this country on the scale 1: 400 000. In the explanatory text accompanying the map of superficial deposits, Sauramo (1924) brought together the observations that had been made of the striations. He pointed out that they occurred in the area ».... in nearly all directions except from the east and south.» The text does not, however, deal with the significance of the varying striae directions or with their age sequence.

Later on Helaakoski (1943) made a study of the movements of the continental ice sheet in the Tampere region, chiefly concentrating on the vicinity of the southern end of Näsijärvi. According to Helaakoski, the striae here comprise five different categories, between which they do not occur. The striae groups become progressively younger to the extent that they turn from the west via the north toward the northeast.

A detailed study of the striae directions was carried out by the present author in the area during the years 1955—1958. The observations were made in conjunction with the mapping of superficial deposits.

RELIEF

The elevation of the ground often played a decisive part in glacier movements. Even very slight differences in elevation have been observed to have caused, under favoring conditions, changes in the movement of the ice (Demorest, 1938, Holmes, 1937). Especially when the ice had considerably thinned down from its maximum thickness, its movements tended to conform to the main features of the local topography. The valley systems have been observed to have in many cases fundamentally influenced the directions of flow of the ice (Chamberlin, 1888; Goldthwait, 1941). On the other hand, thick glacier ice is not appreciably affected by small local differences of elevation (Virkkala, 1951).

The absolute relief of the area is shown in Figure 1. Farthest to the west, the surface of Kulovesi is 57 meters above sea-level, while the highest peak in the area, Vuoreksenvuori, to the south of the town of Tampere, rises to an elevation of 193 meters. Certain other hilltops in the area also reach an elevation of at least 180 meters. The mean elevation of the area is 112 meters.

Local differences in elevation throughout most of the region are less than thirty meters. Along the shores of lakes, at the edges of the Tampere esker as well as in watersheds, they are to a large extent under ten meters. Only in exceptional places are the differences more conspicuous, as in the vicinity of Kulovesi and eastward from there along the banks of the river Nokia, where they are 50—70 meters. The esker of Pyynikki in Tampere rises about 80 meters above the waters of Pyhäjärvi — and, if the depth

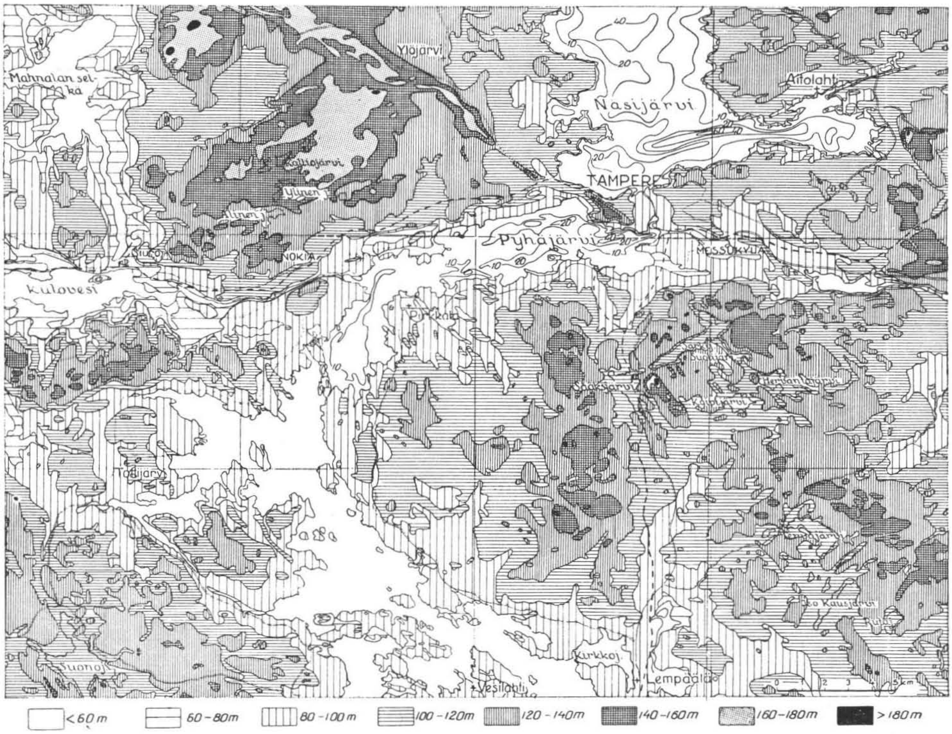

Fig. 1. Absolute relief as well as depth curves of Näsijärvi and of the northern part of Pyhäjärvi.

of the lake at the margin of the esker be taken into account, the difference is about 120 meters. Likewise, Vuoreksenvuori soars 73 meters above the surface of the pond at its foot, and Aitovuori is 78 meters above Näsijärvi. in certain other small areas too, the differences in elevation are of comparable magnitude.

A certain prominent fracture valley runs from Aitolahti to Tampere in a west-southwesterly direction, continuing via the northern part of Pyhäjärvi and the Nokia river to Kulovesi. It is joined by another valley system, to which belong the deep hollows in the northern part of Pyhäjärvi and their extension eastward. The third fracture valley, located to the southeast from the church of Ylöjärvi, is buried for the most part under superficial deposits.

SUPERFICIAL DEPOSITS

Till is the most prevalent kind of deposit in the region. It comprises more than half, or 51.5 %, of the total land area. The coarser sorted deposits, gravel, sand and fine sand account for about 8.3 % of the land area. The most prominent of the formations is the imposing esker chain running through Tampere across the northern part of the area. Its direction appears to conform to the fracture valleys and lake hollows signifying the greatest differences in elevation, as mentioned in the preceding chapter. The stretches of fine sand are situated mainly along the margins of this esker chain. The small occurrences of sand and fine sand met with in the western part of the area are mainly shore deposits.

Tracts of clay and silt comprise 28.4 % of the total land area. They are mainly confined to the shores of large lakes or the bottoms of shallow and gently sloping valleys extending from them. They have likewise been encountered, however, at considerably higher elevations as slight occurrences at the bottoms of bogs.

Tracts of peat and gyttja comprise 7.5 % of the total land area. The most extensive bogs are situated in the watershed regions. Numerous peat deposits are under cultivation along lakeshores.

Rock outcrops occur fairly evenly throughout the entire area. They account for 4.3 % of the total land area. In certain spots, however, they are uncommonly abundant, as in the northern parts of the area as well as around the northern ends of Näsijärvi and Pyhäjärvi.

STRIAE OBSERVATIONS

Most of the rock outcrops in the region have been studied for the purpose of making striae observations. The number of such observations is considerable. All told striations have been observed in more than 500 different places. Striae directions have been calculated in more than 1 300 cases.

The fundamental part of the striae observations made in the region may be seen in Fig. 2. The percentual distribution of the striations among the different directions is shown in Fig. 3. The arrows indicate the direction in which the glacier ice had advanced, and the azimuths represent the directions of advance. On the shores of Näsijärvi the youngest striations were made by glacier movements advancing in quite opposite directions to those in the Pyhäjärvi basin. In many, though not nearly all, cases it has been possible to tell from the microtopography of the rock from which direction the ice had advanced.

The clearest and most general of the striae directions in the region is approximately northwesterly. It is the direction of the majority of the observations made throughout the region. About 35 % of the striae observed are situated between the azimuths 310°—320°, while the corresponding figure between 300°—330° is approximately 55 %.

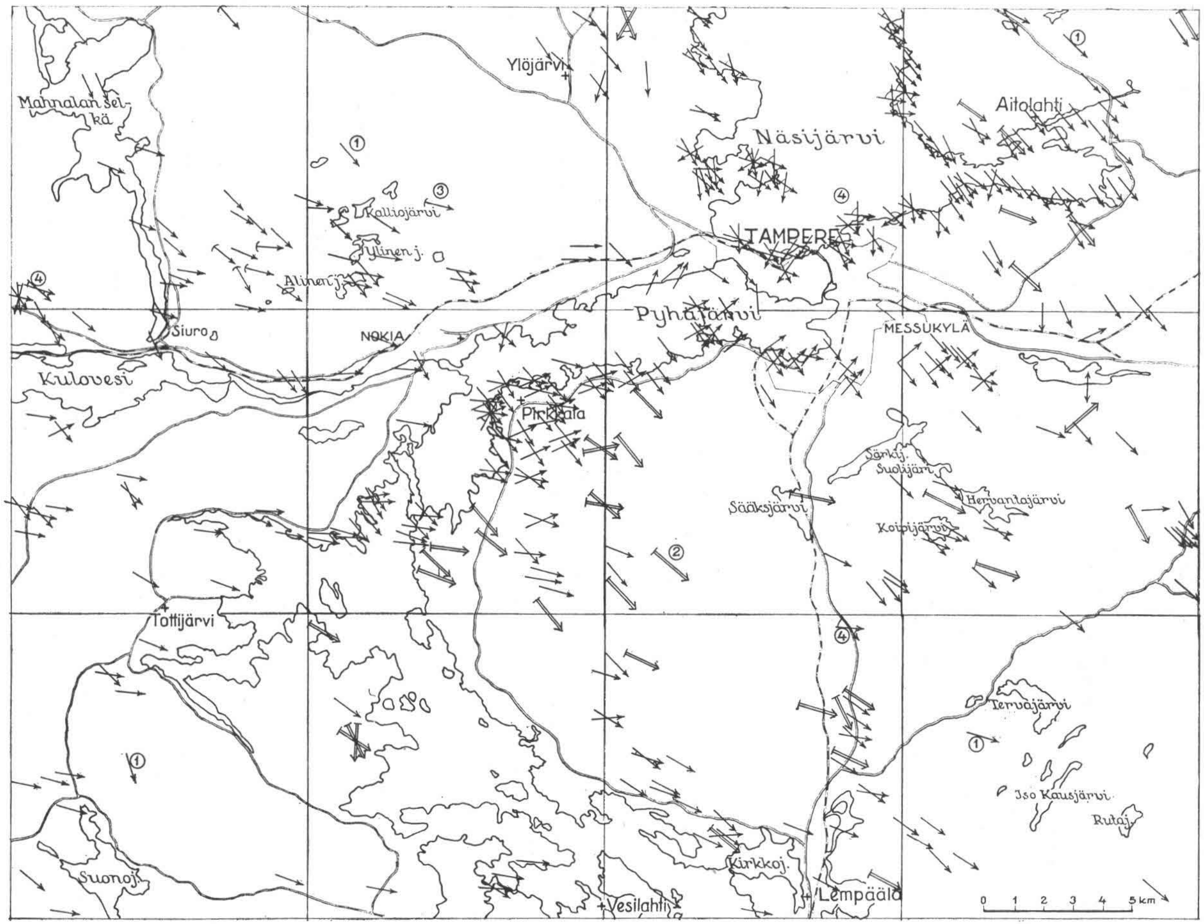

Fig. 2. Summing up of the striae observations and till fabric analyses: 1) stria, 2) till fabric analysis, 3) trend of stoss-side of polished rock, 4) varying striation trend, the extreme values of which are marked by arrows.

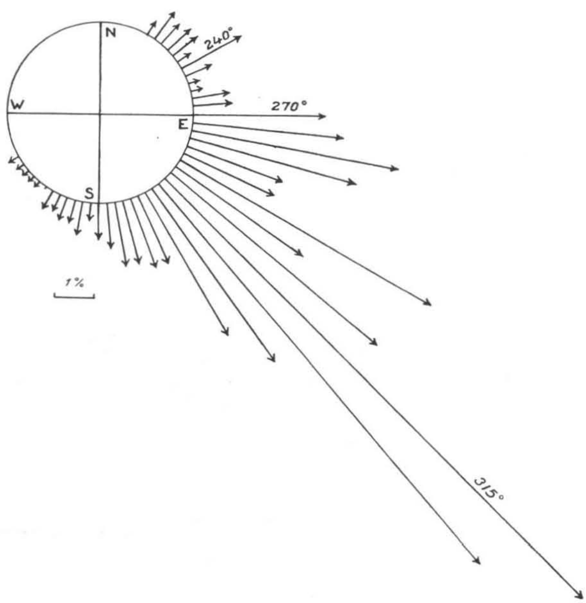

Fig. 3. The percentual distribution of the striae in the region among the different directions. Arrows indicate directions of glacier movement and the numbers of degrees of the azimuths of the directions.

The stoss-sides of the polished surfaces were abraded mostly by glacier ice advancing likewise from the northwest. According to the striations, this direction thus represents the most powerful direction of movement and erosion of the ice sheet. Moreover, in places the forms of the terrain and the trends of the lakes, islands and bays as well as the course of the waterways and the orientation of the valleys reveal the same northwest—southeast direction. Most clearly is this to be noticed in the surroundings of the southern end of Pyhäjärvi. Elsewhere the general trend of the terrain appears to be determined more by the character of the bedrock and the fracture valleys than by glacial erosion.

The rapid decrease in the number of northwesterly striations northward and westward (Fig. 3) indicates that the ice sheet creating them had shifted relatively little in its directions of movement. Even the greatest local differences in elevation in the region do not appear to have caused notable changes of direction in its movement. This would indicate that the glacier ice had at this stage been exceedingly thick.

Certain exceptions can, however, be noted in the general trend of the northwesterly striations. Where the polished surfaces of rocks form slopes, even as little as only couple of meters high, the trend of the striations is likely to vary considerably. On one shore cliff of Näsijärvi the striae on the stoss-side have a trend of 315°, while the striae on the rather steep lee-side gradually shift to a direction of 345°. Similar observations have been made in several other places.

But even on quite even surfaces there is likely to be a diversity of directions among the otherwise parallel northwesterly striations. The extreme values of northwesterly striae on one Näsijärvi shore cliff are 295° and 330°, and on another 285° and 325°. On the rather steep stoss-sides there have been observed northwesterly striations varying between 280° and 350°. Most abundant in the aforementioned examples have been observations representing the mean between the extremes, however. From the mean trend the number of striations rapidly and regularly decreases toward the extreme values. All the striae groups of this description have been included among northwesterly ones, since their means approximate 315°, as in the examples cited in the foregoing.

In addition to striations and erosion trends on the stoss- sides, the rock surfaces exhibit certain other northwesterly erosion marks as well. At some ten observation sites in the region the striations are accompanied by crescentic fractures (Harris, 1943). They occur particularly in dense, fine-grained varieties of rock. The crescentic fractures open up in the directions of both the advance and the retreat of the glacier ice (Okko, 1950). More common than the foregoing on the rock surfaces are small crescentic gouges (Plate I, Figs. 1 and 2). They occur quite commonly also on granitic rocks.

The second largest group in the striation diagram Fig. 3 is composed by the approximately westerly or west-northwesterly striae. Their placing in the diagram as a clear maximum of their own suggests that they must be considered as the product of a separate glacier movement. Of all the striae observations 12.4 % are situated between azimuths 270° and 285°. As mentioned in the foregoing, the northwesterly and westerly striations are likely to overlap in part in their transition zone.

In not nearly all the cases has it been possible to determine the mutual age relationship of the westerly and the northwesterly striations. This has been impossible for the simple reason that the two striae directions have not been present on the same rock surface. This has in many instances not been possible even when the westerly and northwesterly striae cross. The surfaces of the rocks are frequently so weathered that the finer features of the striae have disappeared. In such cases it is often hard to decide which trend crosses the other.

In many cases, again, the age relationship appears to be obvious. This is especially so when northwesterly striae have been met with on the stoss-side of rocks sloping toward the west. The northwesterly striations are certainly younger than the westerly ones. The northwesterly erosion in such cases destroyed the western striations without being able to change the original form of the polished surface of the rock. Fig. 4 schematically

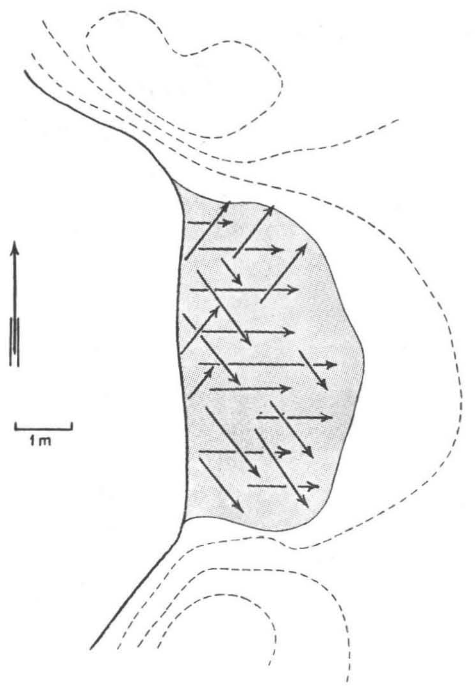

Fig. 4. Schematic drawing of the striations on a rock situated on the shore of Pyhäjärvi. The direction of erosion of the stoss-side of the polished rock is westerly, as is also that of the prevailing striae on it. The northwesterly striae are situated on the southern side of the rock, while the southwesterly striae are situated on the northern side of the rock. The broken line represents approximate elevation curves.

represents a certain rock from the shore of Pyhäjärvi the stoss-side of which slopes westward. The westerly striations prevail elsewhere than in the southern part of the exposure, where the northwesterly striae predominate. In the northern part of the rock the northwesterly striae are lacking, for in respect to the NW-direction it is in a protected position. For the same reason the scattered and youngest southwesterly striations are situated in the northern part of the outcrop.

From certain other observations as well it may be judged that the numerous westerly striations antedate the northwesterly ones. In Figs. 1 and 2 (Plate I) the fractures formed according to the northwesterly trend cut across the westerly striations and are accordingly younger. Likewise, it may be barely noticed in Fig. 3 (Plate I) that the northwesterly striae cut across the westerly ones and must therefore be younger.

However, opposite age relations can also be observed in the region between the westerly and northwesterly striations. In Fig. 4 (Plate I) it may be noticed indistinctly that the finest, western striations cut across the very coarse northwesterly ones. Moreover, the northwesterly striae have been asymmetrically eroded. Their NE-side has been worn lower and to a more gentle slope, for it was confronted with erosion coming from the west. Furthermore, there are a couple of observations from the region in which fine westerly striae continue along the bottoms of well-developed northwesterly grooves.

There have thus been observed in the region western striae which are both older and younger than the strongest, northwesterly striations. The former observations occur, to be sure, in far greater abundance and throughout the entire region under study. On the other hand, the westerly striations of younger age than the northwesterly trend appear to be confined to a rather narrow area around Näsijärvi and the northern end of Pyhäjärvi. Since older westerly striae also occur there, westerly striae of both ages are likely to appear on outcrops situated quite close together. On the whole, of the oldest westerly striae only the coarsest ones are left; the fine striations worn away in the course of time. On the other hand, the younger westerly striations are on the average conspicuously finer and sharper than the older westerly striae.

The relatively abundant occurrence of the oldest westerly striations in the region indicates that the glacial erosion between these and the northwesterly striations cannot have been very great, at least locally. This does not, however, say anything about how great the erosion of the bedrock had been before the earliest striae were made.

Differences can also be observed in favorable instances on polished rock surfaces between westerly striae of different ages in respect to their position. The younger westerly striae are generally situated on the northwestern stoss-sides in the same places as the prevailing northwesterly striae too or in the western parts of the stoss-sides. In certain cases it may be noticed that the older westerly striae tend more to be situated on the lee-side in respect to the northwesterly erosion, that is, on the eastern sides of the polished rocks. But they occur also in the western parts of polished surfaces according to local circumstances. On the other hand, both older and younger westerly striae have not been so far observed on the same polished rocks.

Besides the westerly ones, striations younger than the northwesterly ones are to be found in the region representing many other directions. In the surroundings of Näsijärvi they are largely northerly but vary in direction from north-northwesterly to east-northeasterly. The small erosion forms on the rocks in many cases reveal that the ice really did advance from the northeast and not from the opposite direction. All told 8.5 % of all the striae observations in the area represent the directions between north-northwest and east-northeast.

The northern striae in the surroundings of Näsijärvi have been noted to be younger than the younger westerly striae. In many cases the fine north-northeasterly striae cut across both the northwesterly and westerly striations. In Fig. 4 (Plate I) the relatively fine north-northeasterly striae likewise indistinctly cut the westerly and northwesterly striae. An example of the greatly varying striae directions of the southern part of Näsijärvi is provided by Fig. 5 (Plate I). The north-northeasterly glacier movement formed in its lower part a weakly developed erosion facet, in which — even in the picture herein reproduced — very fine east-northeasterly striae, representing the very youngest variety, can be seen. The original stoss-side, which is still visible, conforms to the northwesterly erosion. Northerly striations revealing several directions are also to be observed in Fig. 6 (Plate I). Fine northeasterly striae continue there at the bottom of broad northwesterly striae. They are likewise cut by northerly striae, but the relative age of the latter and the northeasterly striations is not clear.

Certain other circumstances provide further evidence that the northerly-northeasterly striations are the youngest in the vicinity of Näsijärvi. The striae are situated most generally on the northern and northeastern flanks of the rocks, though there is of them on rocks sloping in other directions too. Furthermore, the finest of the northwesterly striae have in many instances been worn away, so that only the coarsest of them are left. The northeasterly glacier movement has in some cases worn the SW-side of the northwesterly striae to a more gentle slope, for it was the stoss-side of the ice advancing from the northeast. On the other hand, the NE-sides of the northwesterly striae have in such cases remained steeper. The striae have thus in cross-section received a somewhat asymmetrical shape.

Althought the northerly striae vary in direction quite considerably, considering the Näsijärvi vicinity as a whole, they generally occur parallelly on the surface of the same rock. This is evidenced by Figs. 3—6 (Plate I). Rarely does one run across a rock in which all the possible intermediate degrees from north-northwest to east-northeast occur. In such cases the striations on the rock tend to form one general hodgepodge, in which quite coarse single striae and broken-up ones too occur. It is possible that all these striae are not of glacial origin.

In the vicinity of the northern end of Pyhäjärvi and its extension, the trend of the youngest glacier movement was quite opposite to what it was on the shore rocks of Näsijärvi (Figs. 2 and 4). The last movements of the glacier ice took place in this area from in between south-southwest and west-southwest.

This southwesterly erosion trend is in many places on the shore rocks of Pyhäjärvi quite clear and pronounced. Though northwesterly facets prevail there too, numerous smaller ones inclined southwestward are present besides. Southwesterly striations in many instances predominate, especially on the southwestern sides of polished rocks. But they are met with generally across the whole length and breath of the polished surfaces, whereupon the older northwesterly striae are dimly outlined underneath the southwesterly ones. It has been occasionally noted that southwesterly striae extend along the base of broad northwesterly grooves or they have worn northwesterly striations into asymmetric shapes. Certain other miniature erosion forms on rocks likewise indicate that the southwesterly striae in the region are unquestionably younger than the northwesterly ones. They further indicate that the ice sheet had actually advanced approximately from the southwest — and not from the opposite direction.

What the age relationship of the southwesterly striae might be to the aforementioned (p. 195) younger westerly ones cannot be determined for certain. Presumably they are approximately of the same age. The younger westerly striae would in that case represent the extreme values of the southwesterly movement. Other observations, again, would seem to prove that the younger westerly striae are distinctly older than the southwesterly ones.

In how extensive an area southwesterly striae occur, it is difficult to say. Rocks in which striae occur unweathered are met with only on the northern end of Pyhäjärvi and its extension. Farther in the interior weathering has generally destroyed the finer striae. Only the coarsest and strongest northwesterly striations are consistently visible on the rocks.

Southwesterly striae between azimuths 210° and 260° have been met with in about eight per cent of the cases.

It may seem strange that in as limited an area as the one under investigation striations should be encountered which have been produced by glacier ice advancing from opposite directions. Since glacier movements occur at right angles to the margin, one must assume that during the period when the last striae were created the ice made great detours; this would have had to happen to make such movements at the glacier margin possible. The reasons for this circuitous movement must apparently be sought in the local topography. Reference has been already made (p. 189) to the great fracture valleys met with in the region — evident in the relief and in the depth curves of the lakes. These fracture valley system could not but greatly affect the final movements of the edge of the greatly thinned-down glacier. They produced in the ice a deep bay tapering toward the west-northwest which gradually changed proximally into a crevasse and into which the mighty Tampere esker was deposited. The ice had not yet even at this stage wholly lost its power of movement. Its movements were now directed at right angles to the glacier margin. In the surroundings of Näsijärvi the movement thus took on by and large a north-northeasterly direction. This trend was further strengthened by the largely north-south trend of the basin of Näsijärvi. The local topography began, however, to affect this north-northeasterly movement to an ever-increasing extent.

Especially the aforementioned deep bay of Aitolahti tended to turn it east-northeastward. This makes understandable the youngest striations and their very diverse trends.

The younger westerly movement thus had to take place on the northern side of the esker chain already before the formation of this glacial bay in the region. The striae observations do indicate that the younger westerly striae are older than the striations with a trend between north and northeast.

South of Tampere the youngest movements of the ice sheet likewise had an initial trend from west to east. This trend was probably decisively influenced by the west-east orientation of the northern part of Pyhäjärvi and the fracture valley situated in its extension. The glacier's edge was at this stage still outside the region. When the edge had retreated to the region and the aforementioned topography-controlled glacial bay had formed, the final movements of the ice sheet became directed at right angles to its margin and, accordingly, on the average from the southwest to the northeast. The topography is likely to have affected the movements of the ice more and more as the margin of the retreating ice got closer to the region and the glacier thinned down by melting.

Outside the basins of Näsijärvi and Pyhäjärvi the youngest striations described in the foregoing have not been encountered, whether northerly or southwesterly in trend. To some extent this may be due to the fact that there are fewer outcrops in the interior and above all that they are considerably more weathered on the surface than the rocks cropping out along the lakeshores. The lack of southwesterly and northerly striations outside the lake basins is primarily due to the fact that the most recent glacier movements did not take place in those directions there because the necessary topographic factors were not present beyond the immediate surroundings of the lake basins.

TILL FABRIC

Glacier movements can be traced on the basis of not only striations but also the orientation of stones in till. In the light of numerous studies, it is known that a noteworthy proportion of the stones in till tend to lie in such a position that their long axis parallels the movement of the ice which deposited the till (Holmes, 1941; Hyyppä, 1948; Richter, 1936).

The till in the region is a relatively coarse material. Sand and gravel are the dominant constituents in material consisting of grain sizes under 20 mm. Fig. 5 shows the mechanical composition of certain typical samples of till from the region, together with the mean mechanical composition of the local till on the basis of some forty samples.

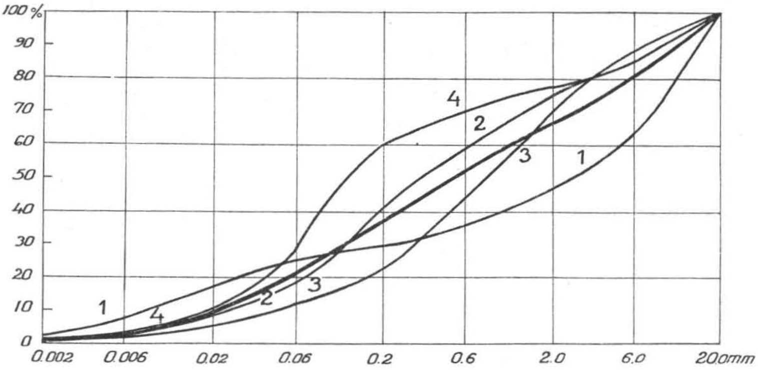

Fig. 5. Cumulative curves of the till occurring in certain parts of the region. The thick line represents the average mechanical composition of the till in the region on the basis of some forty analyses.

Also coarser grain sizes are common in the till. Surface boulders, in particular, are quite numerous in many places; quite commonly they even form boulder fields as the result of solifluction (Fig. 6). Besides boulders the till also contains stones of different sizes in abundance, primarily in its inner parts. In many places they constitute the chief portion of the entire mass formed by the till. Angular rocks of various sizes are usually wedged into each other, producing a formation hard to penetrate. Since, moreover, the fundamental constituents of the spaces in between the rocks are mainly sand and gravel — and thus quite permable —, the iron and humus compounds precipitated from the circulating solutions of water have further cemented the till.

Under such conditions a till fabric analysis does not give the best possible result. There are often too many stones; there has not been room enough for all the stones to arrange themselves in the manner described in the foregoing. It is known on the basis of pebble counts that the stones in the till have in many cases been transported only a short distance, which is also indicated by their prevailing angularity. On account of the short-distance transport there was not time for the numerous stones to become oriented in conformity with the glacier movement. A considerable part of the fabric analyses of the local till does not therefore give any clear-cut picture.

Certain fundamental features can, however, be observed in the orientation of the till stones in the locality under investigation. The results of the fabric analyses are presented in Fig. 2, along with the striae observations. The figure elucidates only those analyses in which a distinct orientation has been ascertained. It may be noticed from the picture that the trends corresponding to the striations occur also in the arrangement of the pebbles in the till. The northwesterly trend, which was perceived to represent the

Fig. 6. A boulder field originating as the result of solifluction in a level valley.

direction of the most powerful abrasive action of the ice sheet, is likewise the one occurring most frequently in the till fabric. It is thus also the most prominent trend exhibited by the accumulations of glacial drift in the region.

The pebbles in the till reveal trends corresponding to other striae directions too. On the eastern shore of the middle part of Pyhäjärvi two indistinct till beds have been encountered in a certain till section. The orientation of the upper bed is plainly northwesterly. In the lower bed, on the other hand, the most distinct trend is approximately from the west. This observation bolsters the conception gained from the arrangement of the striations that the westerly movement of the glacier was the earliest in the region. Here and there in other localities too the westerly orientation is to be perceived in the till. This is, however, relatively weak and indistinct, being partly due apparently to the fact that the subsequent, northwesterly glacier movement disturbed and mixed up and rearranged the oldest till fabric.

In one locality in the surroundings of Näsijärvi, a northerly orientation has also been observed in the till fabric. This indicates that the last movements of the continental ice sheet were in some spots of sufficient strength to produce accumulations or at least to rearrange previously deposited till. On the other hand, the till fabric reveals no southwesterly trend.

The till fabric in the region thus provides evidence of the same glacier movements as the striations. Thereby the conclusions to be drawn on the basis of the striae observations regarding the glacier movements in the region under investigation are confirmed by the till fabric analyses.

SUMMARY

Glacier movements of three different ages may be perceived. All the movements date, however, from the same, or last, period of glaciation.

The earliest movement noted came roughly from the west. Evidences of it can be found throughout the entire region studied and its direction varied very little.

The most conspicuous direction of movement in the region was from the northwest toward the southeast. This movement is borne witness to everywhere in the region by the strongest and clearest evidences left behind by any of the three. Its direction likewise varied very little. The local differences in elevation, which amount to less than 80 meters, did not appreciably affect its direction of movement. From this it may be judged that the glacier producing both the westerly and northwesterly striations was extremely thick. It is possible that the earlier westerly movement corresponds to the stage of advance of the ice sheet across the region and the northwesterly movement the period prior to the achievement of the maximum extent of the final glaciation. The position of the ice divide would thus have shifted in the Scandinavian mountains (Ljungner, 1949). It would have been situated during the opening stage of glaciation farther south and shifted by the time of the maximal glaciation a few hundred kilometers farther north.

The youngest glacier movements varied in the region quite considerably in direction. The topography got to control the directions of movement, inasmuch as by the recession stage the ice sheet had conspicuously thinned down. At first the youngest movements of the ice took place in conformity with the basins of Aitolahti and the northern end of Pyhäjärvi from the west toward the east. The margin of the glacier was then at some distance from the region. As the retreating ice margin approached the region, a large glacial bay was created through the influenc eof the topography in the locality of the Tampere esker; it was in the tapering proximal part of this bay that the mighty esker evolved by deposition. The last glacier movements took place on the northern side of the esker largely from the north, or from the sector lying between north-northwest and east-northeast. Their great diversity of direction may be explained by the influence of the topography on the glacier movements.

On the southern side of the esker the last movements took place between the southwest and the south, in the area nearest the esker partly from the west-southwest as well. The topography determined the local direction of the final movements of the ice here likewise.

The last movements of the ice sheet thus took place in the region under consideration by and large at right angles to the mighty esker chain occurring here. Similar observations have been made elsewhere in Finland too — in the Kainuu region (Virkkala, 1951) and northern Karelia (Repo, 1957). Corresponding relations are also known from Canada (Norman, 1939). This appears to be the case commonly in other regions too where imposing

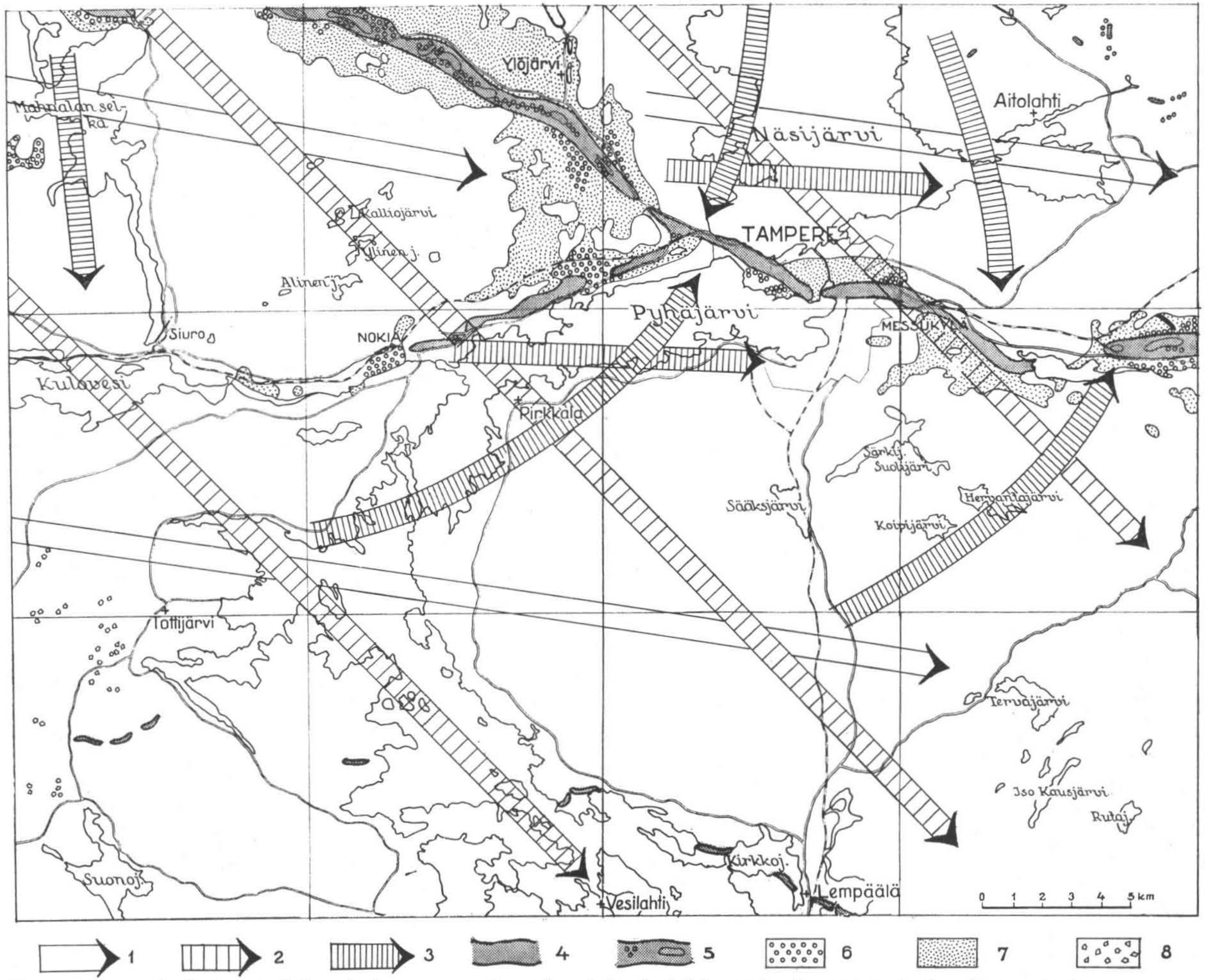

Fig. 7. The main features of the glacier movements and certain glacial forms in the region: 1) the older westerly movement, 2) the strongest, northwesterly movement, 3) the youngest direcitons of movement of the ice in the region, 4) esker, 5) knob -and-hill terrain and an esker summit plain, 6) sand beyond an esker, 7) fine sand, 8) ablation moraine.

esker chains occur and where conditions permit the making of sufficient observations of the glacial striae.

A summing up of certain glacial forms and glacier movements in the region is presented in Figure 7.

An analysis of the till fabric in the region brings to light the same dominant features as the striations in regard to the directions of movement of the continental ice sheet. This in spite of the fact that certain unfavorable circumstances hamper the performance of till fabric analyses in the region.

REFERENCES

CHAMBERLIN, T. C. (1888), The rock-scorings of the great ice invasions. U. S. Geol. Survey, 7th Ann. Rept.

DEMOREST, M. (1938), Ice flowage as revealed by glacial striae. J. Geol. 46.

GOLDTHWAIT, R. P. (1941), Wisconsin glaciation of the Presidential Range, New Hampshire. Geol. Soc. Am., Bull. 52.

HELAAKOSKI, A. R. (1943), Mannerjäätikön liikuntosuunnista Pohjanmaan rannikolla ja Tampereen ympäristössä. Referat: Über die Bewegungsrichtungen des Inlandeises an der Küste von Pohjanmaa und in der Umgebung von Tampere. Fennia 67.

HOLMES, C. D. (1937), Glacial erosion in a dissected plateau. Am. J. Sci. 29.

—»— (1941), Till fabric. Geol. Soc. Am., Bull 52.

HYYPPÄ, E. (1948), Tracing the Source of the Pyrite Stones from Vihanti on the Basis of Glacial Geology. Bull. Comm. géol. Finlande 142.

LJUNGNER, E. (1949), East-west Balance of the Quaternary Ice Caps in Patagonia and Scandinavia. Geol. Inst. of Upsala Bull. 33.

NORMAN, G. W. H. (1939), The South-Eastern Limit of Glacial Lake Barlow-Ojibway in the Mistassini Lake Region, Quebec. Trans. Roy. Soc. Canada, Ser. 3, Vol. 23, sec. IV.

OKKO, V. (1950), Friction Cracks in Finland. Bull. Comm. géol. Finlande 150.

REPO, R. (1957), Untersuchungen über die Bewegungen des Inlandeises in Nordkarelien. Bull. Comm. géol. Finlande 179.

RICHTER, K. (1936), Ergebnisse und Aussichten der Gefügeforschung im pommerschen Diluvium. Geol. Rundschau 27.

SAURAMO, M. (1923), Studies on the Quaternary Varve Sediments in Southern Finland. Bull. Comm. géol. Finlande 60.

—»— (1924), Suomen geologinen yleiskartta. Lehti B 2. Tampere. Maalajikartan selitys. Geologinen Komissioni. Helsinki.

VIRKKALA, K. (1951), Glacial Geology of the Suomussalmi Area, East Finland. Bull. Comm. géol. Finlande 155.

EXPLANATIONS TO THE PLATE I

Fig. 1. Crossing striae in the northwestern part of the region from the shore of Mahnalanselkä. Pencil points west, compass northwest. Latter direction is younger.

Fig. 2. Crossing striae and crescentic gouges from shore of Pyhäjärvi. Pencil points west. Westerly striations are cut by crescentic gouges and striae of younger, northwesterly trend.

Fig. 3. Crossing striae from shore of Näsijärvi. Direction shown by compass 275°, by film-roll 315° and by pencil 20°—30°. Age relationship of striae is only faintly evident in figure. The westerly striations are the oldest, the north-northeasterly the youngest.

Fig. 4. Weaker westerly striae (book, 280°) cut across older and very coarse northwesterly striations (direction of exposure of film). Direction of erosion of polished rock is northwesterly. In direction of pencil (20°) are youngest, north-northeasterly striations.

Fig. 5. Varying striae directions from shore of Näsijärvi. Knife points in direction 305°, sheath 330°, pencil 20° and compass 60°. North-northeasterly movement wore weakly developed facet on northwestern stoss-side, where, in lower part of picture, can be seen distinct north-northeasterly as well as fine, east-northeasterly striae of the youngest variety.

Fig. 6. Crossing striae from eastern side of Tampere. Pencil points in direction 315°, matchbox 360° and focal direction of photo 40°. Weaker and youngest northeasterly striations continue at the base of coarser and older northwesterly striations.

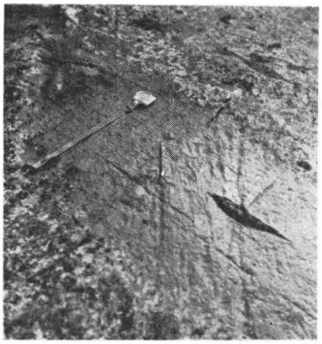

Fig. 1

Fig. 2

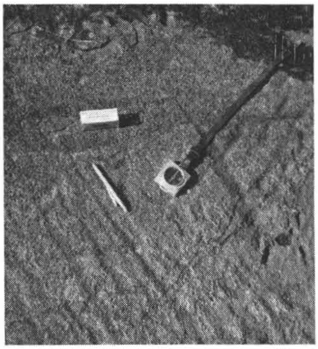

Fig. 3

Fig. 4

Fig. 5

Fig. 6

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